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A comparative study of contrasting structural styles in the range-front region of the northeastern Arctic National Wildlife Refuge, northeastern Brooks Range, Alaska

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A COMPARATIVE STUDY OF CONTRASTING STRUCTURAL STYLES IN THE RANGE-FRONT REGION OF THE NORTHEASTERN ARCTIC NATIONAL WILDLIFE REFUGE, NORTHEASTERN BROOKS RANGE, ALASKA

A

THESIS

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A COMPARATIVE STUDY OF CONTRASTING STRUCTURAL STYLES
IN THE RANGE-FRONT REGION
OF THE NORTHEASTERN ARCTIC NATIONAL WILDLIFE REFUGE,
NORTHEASTERN BROOKS RANGE ALASKA

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ABSTRACT

The range front of the northeastern Brooks Range in the Arctic National Wildlife Refuge (ANWR) is defined by anticlinoria cored by a 'basement' complex of weakly metamorphosed sedimentary, volcanic and intrusive rocks. These anticlinoria are interpreted to reflect horses in a northward-propagating regional duplex between a floor thrust at depth in the 'basement' complex and a roof thrust near the base of the cover sequence. Lateral variations in the geometry of these range-front anticlinoria reflect changes in lithology and deformational style of both the 'basement' and its cover.

Two distinct structural geometries are displayed along the range front of northeastern ANWR. To the east, the large range-front anticlinorium is interpreted to reflect multiple horses of Cenozoic age within the stratified, slightly metamorphosed sedimentary and volcanic rocks of the pre-Mississippian 'basement'. During Cenozoic thrusting, these mechanically heterogeneous rocks deformed primarily via thrusting and related folding with minor penetrative strain. The Mississippian and younger cover sequence shortened via both thrust duplication and detachment folding above a detachment in the Mississippian Kayak Shale.

In contrast, to the west the pre-Mississippian rocks consist primarily of the mechanically homogeneous Devonian Okpilak batholith. The batholith was transported northward during Cenozoic thrusting and now forms a major topographic and structural high near the range front. The batholith probably shortened during thrusting as a homogeneous mass via penetrative strain. Because the Kayak Shale is thin to absent in the vicinity of the batholith, Mississippian and younger rocks remained attached to the batholith and shortened via penetrative strain and minor imbrication.

These two range-front areas form the central portion of two regional transects through northeastern ANWR. General area-balanced models for both transects suggest that the amount of total shortening is governed by the structural topography and the geometry of the basal detachment surface. While the structural topography of northeastern ANWR is

reasonably well-constrained, the geometry of the basal detachment is not. Given a range in reasonable basal detachment geometries, shortening in both transects ranges from 16% to 61%. Detailed balanced cross sections based on subsurface and surface geologic data yield 46-48% shortening for both transects.

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ACKNOWLEDGEMENTS

This dissertation represents over five years of my life. When I began the study in 1986 I, like many other Ph.D. students, did not truly appreciate the commitment this single project would represent in time, energy and money. And now that it is almost done, I cannot honestly say "It's all been worth it." But I have learned much and seen parts of Alaska I will never forget.

During these five years, many people and organizations have influenced the course of this project in a variety of ways. Without the sometimes superhuman efforts of Wes Wallace, Keith Crowder and Keith Watts of the Tectonics and Sedimentation Research Group, this study would not have even been started, much less completed. These three individuals managed to organize and fund an ongoing major academic research project in a very remote area. Their financial, logistical and moral support have been essential and is greatly appreciated.

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Once the field work was over, the real work began of actually writing the dissertation. I originally intended this dissertation to be a collection of published and submittable papers. Although this goal may superficially appear straight forward and attainable, it is definitely not the easiest way to write a thesis. The components of a standard thesis are fairly easy to recognize regardless of subdiscipline; however, not everyone agrees as to what constitutes a good structural geology paper. The as yet unpublished parts of this dissertation are therefore the result of many compromises. I would like to thank all the members of my committee, David Stone, Mary Keskinen, Keith Watts, Keith Crowder, Gil Mull and Hans Avé Lallemant, for the many hours they spent reading drafts of the dissertation and making comments. Although we did not necessarily agree, the final dissertation has benefited from all their efforts.

And lastly, I would like to thank my immediate 'family,' which has changed and grown considerably over the last five years. The makeup of that family has fluctuated over the years with the addition and subtraction of various feline and canine dependents, all of whom have done what they could to make those five years enjoyable. The recent arrival of a new and entertaining family member, Theresa, has undoubtedly complicated their lives as well as my own.

However, I especially wish to thank the one constant member of my family--my husband, Wes Wallace. He has provided five years of scientific, financial and emotional support, as well as a wonderful daughter. On top of it all, he has survived co-authoring three papers with me. This dissertation would have been impossible without him.

CHAPTER 1: INTRODUCTION AND STATEMENT OF PROBLEM

1.1 Introduction

The structural geometry and evolution of foreland fold-and-thrust belts have recently become the focus of increased interest to structural geologists, both in academia and industry. This interest has been spurred partly by major strides in concepts regarding the evolution of fold-and-thrust belts, as well as by recognition of the interrelationship between the structural evolution of a foreland fold-and-thrust belt, the subsidence and depositional history of the foreland basin, and the subsequent uplift history of the basin. An additional impetus to study foreland fold-and-thrust belts is the increasing awareness of the hydrocarbon potential of these areas. As traditional hydrocarbon exploration targets become fewer and smaller, explorationists are looking for more subtle and structurally complex prospects. Foreland fold-and-thrust belts can be highly productive, and provide the structural control for many of the major oil fields of the world, including those of the Persian Gulf and the Rocky Mountain Overthrust of the western U.S.A. and Canada (North, 1985).

This study focuses on a portion of a remote foreland fold-and-thrust belt in Alaska--the northeastern Brooks Range of the Arctic National Wildlife Refuge (ANWR) (Figure 1.1). As with other mountain belts, the problems posed by this particular fold-and-thrust belt are both specific to the region and of more general interest. In this chapter I will identify the various questions that are addressed by this study. In order to meaningfully define these questions, it is necessary to outline briefly the regional geology of the northeastern Brooks Range and summarize the current state of knowledge regarding the structural evolution of the specific part of the range covered by this dissertation. I will then discuss the questions addressed by this study under two broad categories: those issues that deal primarily with increasing our understanding of the geologic evolution of the

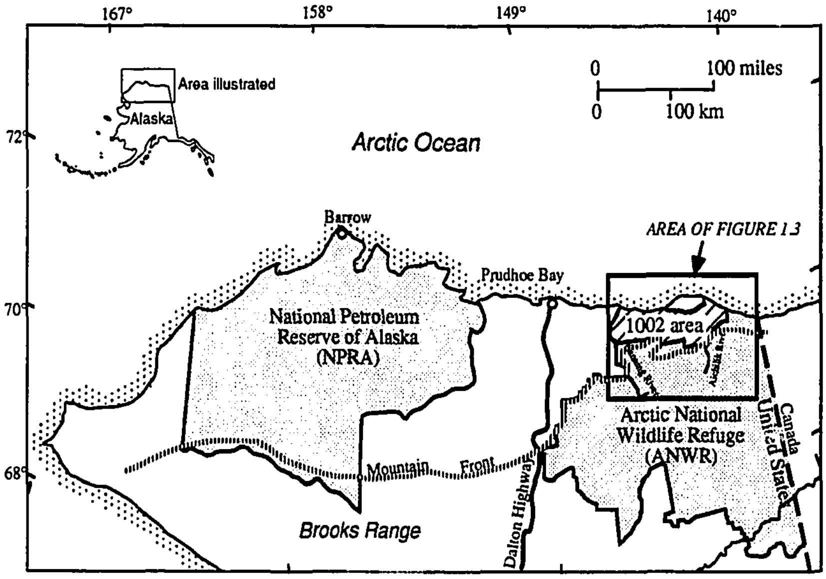


Figure 1.1. Generalized map of northern Alaska showing location of the Arctic National Wildlife Refuge (ANWR) with respect to other geographic and geopolitical features.

northeastern Brooks Range and Arctic Alaska, and those questions that delve primarily into the origin and structural evolution of fold-and-thrust belts in general. At the end of this chapter, I will briefly outline the geographic area of the study and the organization of the remainder of the dissertation.

1.2 Regional geology of the northeastern Brooks Range

Most of the shortening in the main axis of the Brooks Range is the result of the collapse of a south-facing late Paleozoic to early Mesozoic passive continental margin during a Middle Jurassic to Early Cretaceous north-vergent thrusting event (Mull, 1982; Mayfield and others, 1983; Oldow and others, 1987b). This deformational event resulted in hundreds of kilometers of northward displacement, and the formation of a complex series of stacked thrust sheets containing age-equivalent but lithologically distinct rocks. In contrast, the northeastern Brooks Range is a Cenozoic, still active salient of the Brooks Range proper (Grantz and others, 1983; Moore and others, 1985a; Carter and others, 1986; Wallace and Hanks, 1990), involving parautochthonous rocks similar in stratigraphy to those preserved in the undeformed foreland of the Brooks Range, the North Slope subsurface (Reiser, 1970).

The parautochthonous stratigraphy of the northeastern Brooks Range can be divided into three major packages, each having a different depostional history and a different structural response to Cenozoic shortening (Figure 1.2; Bird and Molenaar, 1987; Wallace and Hanks, 1990). The oldest and structurally lowest package is an assemblage of heterogeneous, weakly metamorphosed pre-Mississippian sedimentary rocks with minor volcanic rocks and granitic intrusive rocks (Reiser and others, 1980). These pre-Mississippian rocks form the cores of the regional anticlinoria that are the dominant regional structures of the northeastern Brooks Range (Figure 1.3; Bader and Bird, 1986).

The overlying Mississippian and younger cover sequence can be divided into two general stratigraphic packages (Figure 1.2). Mississippian through Lower Cretaceous rocks of the Ellesmerian sequence were deposited on a south-facing passive continental

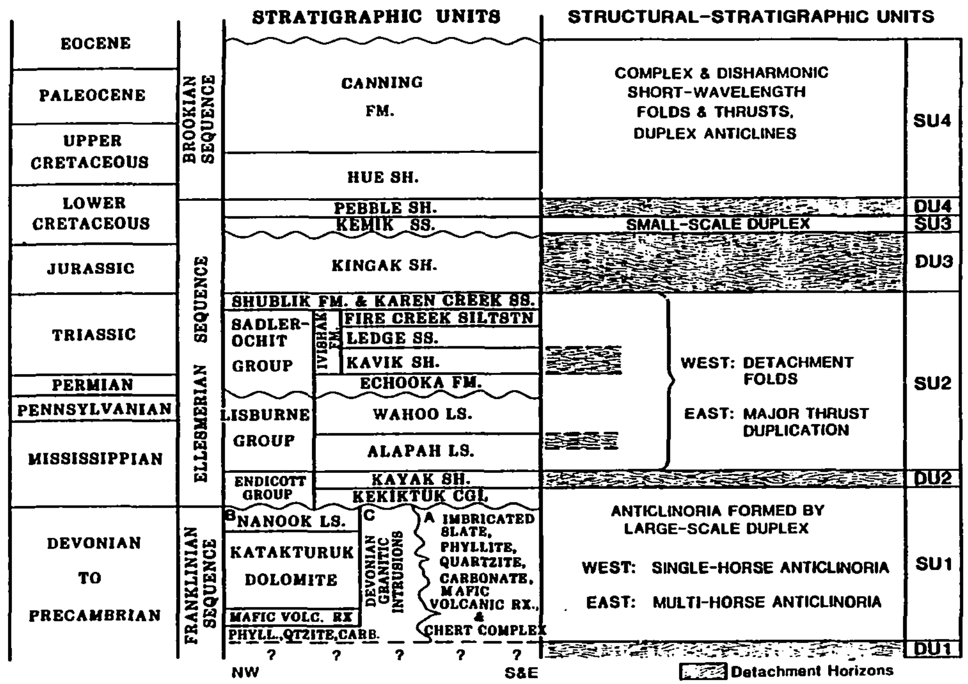
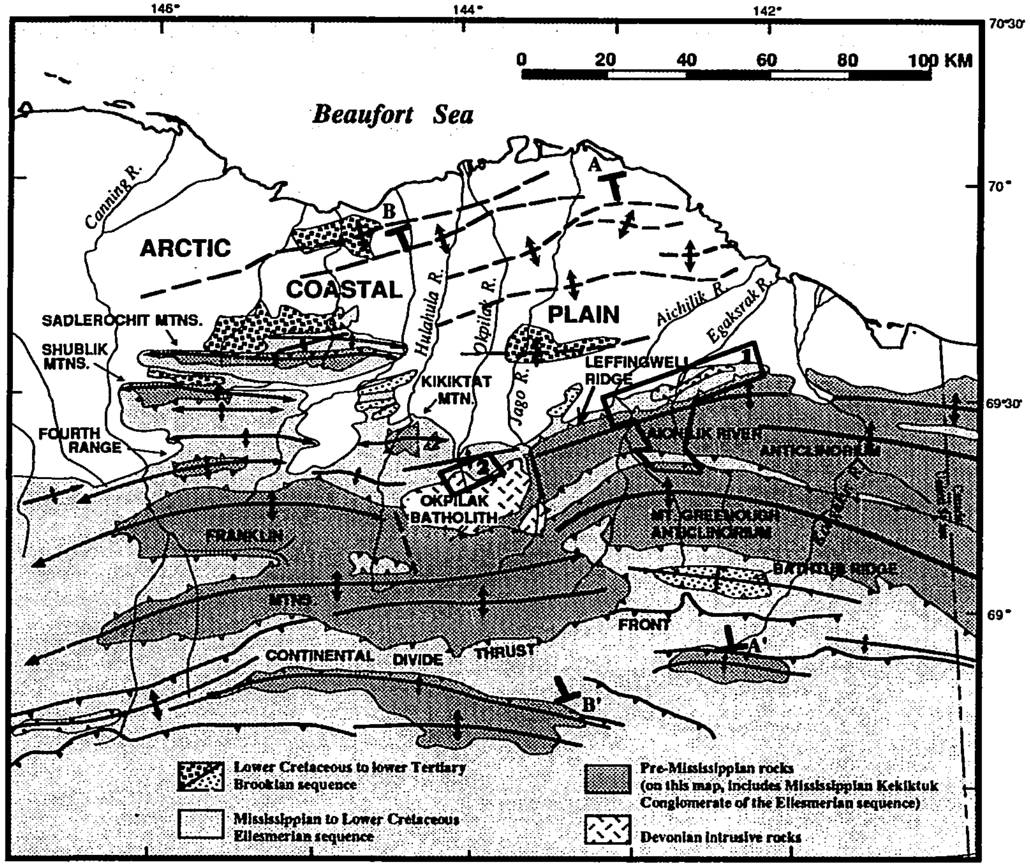


Figure 1.2. Stratigraphic column of rocks exposed in the northeastern Brooks Range.

Figure 1.3. Generalized tectonic map of the northeastern Brooks Range showing the location of the two study areas. 1: Aichilik and Egaksrak Rivers area; 2: northern margin of the Okpilak batholith. Cross section A-A': Aichilik River transect; cross section B-B': Okpilak batholith transect. The distribution of the major structural-stratigraphic units and structural features of the northeastern Brooks Range as shown on this map was modified from Brosgé and Reiser, 1965; Brosgé and others, 1976; Bader and Bird, 1986; Clough and others, 1987. Solid teeth on thrust faults indicate older-over-younger thrust faults that duplicate stratigraphic section; open teeth indicate detachment surfaces along which there has been slip but no disruption of the normal stratigraphic succession.



146°
144°
142°
70°30'
70°
69°30'
69°
0 20 40 60 80 100 KM
Beaufort Sea
Canning R.
ARCTIC
COASTAL
PLAIN
A
B
Hulahula R.
Okpilak R.
Jago R.
Aichilik R.
Egaksrak R.
SADLEROCHIT MTNS.
SHUBLIK MTNS.
FOURTH RANGE
KIKIKTAT MTN.
LEFFINGWELL RIDGE
AICHILIK RIVER
ANTICLINORIUM
OKPILAK BATHOLITH
MT. GREENOUGH ANTICLINORIUM
FRANKLIN
MTNS.
BATHTUB RIDGE
FRONT
CONTINENTAL DIVIDE THRUST
A'
B'
1
2
Lower Cretaceous to lower Tertiary Brookian sequence
Mississippian to Lower Cretaceous Ellesmerian sequence
Pre-Mississippian rocks (on this map, includes Mississippian Kekiktuk Conglomerate of the Ellesmerian sequence)
Devonian intrusive rocks

margin, and consist of platform carbonate rocks and primarily marine clastic rocks (Bird and Molenaar, 1987). In the northeastern Brooks Range, the Ellesmerian sequence is separated from the underlying pre-Mississippian rocks by an angular unconformity (Reiser, 1970; Bader and Bird, 1986; Reiser and others, 1980; Robinson and others, 1989) and is well-preserved on the flanks, and locally over the crests, of the regional anticlinoria (Figure 1.3).

Overlying the Ellesmerian sequence is the Brookian sequence, a thick sequence of deep marine to non-marine Lower Cretaceous to Tertiary clastic rocks shed from the growing Brooks Range to the south (Figure 1.2; Mull, 1985; Molenaar and others, 1987). These rocks are only locally exposed in structural lows in the northeastern Brooks Range and in poor exposures in the coastal plain north of the range front, and do not occur in the field areas of this dissertation.

1.3 Questions specific to northeastern Alaska

While the geology of the entire northeastern Brooks Range is poorly understood, published information on the northeastern corner of the northeastern Brooks Range east of the Okpilak batholith (Figure 1.3) is especially sparse. In the past, both stratigraphic and structural studies have focussed on the northwestern portion of the northeastern Brooks Range, specifically in the vicinity of the Sadlerochit and Shublik Mountains and the northern portions of the Franklin Mountains (e.g., Reiser, 1970; Armstrong and Mamet, 1975; Oldow and others, 1987a; Ziegler, 1989). This is undoubtedly because the rocks exposed here are the closest exposed equivalents to those encountered in the subsurface at Prudhoe Bay. In addition, the northwestern portion of the northeastern Brooks Range in general has better exposure than range-front areas east of the Okpilak batholith. However, that part of the range front between the Okpilak batholith and the U.S./Canada border is areally quite extensive, comprising approximately one third of the range-front region of the Alaskan portion of the northeastern Brooks Range. Consequently, any attempt to understand the structural evolution of the northeastern Brooks Range must consider this region. However, mapping east of the Okpilak batholith has been primarily at a reconnaissance level at 1:250,000 scale (Reiser and others, 1980) with few detailed structural studies (e.g., Sable, 1977).

In recent years, increased government and industry interest in the resource potential of the northeastern Brooks Range has emphasized how little is actually known about the region. The foredeep basin north of the range front of the northeastern Brooks Range (the coastal plain or '1002 area' of ANWR, Figure 1.1) is, at the time of writing, actively under consideration by the U.S. government for oil leasing and exploration. Because the stratigraphy and structures exposed at the range front probably extend into the subsurface to the north, understanding the relationship between stratigraphy and the geometry and evolution of the exposed structures may be crucial in developing exploration strategies for the subsurface to the north.

Only the lowest two stratigraphic sequences (the pre-Mississippian sequence and the Ellesmerian sequence, Figure 1.2) are extensively exposed east of the Okpilak batholith. Each of these two sequences has its own structural and stratigraphic problems relevant to understanding the structural evolution of the northeastern Brooks Range. Some of these questions can be uniquely addressed in the northeastern part of the range.

1.3.1 Stratigraphy and structure of the pre-Mississippian rocks and their role in the evolution of the circum-Arctic region

The pre-Mississippian rocks of the northeastern Brooks Range and northwestern Canada are a very diverse and poorly dated assemblage of rocks (Reiser and others, 1980) whose depositional setting, structural style and tectonic history are poorly understood. Although age control in these rocks is generally poor, most workers have considered them to range from Proterozoic to Devonian in age (e.g., Reiser and others, 1980; Norris and Yorath, 1981). These pre-Mississippian rocks also have a complex deformational history, which further complicates understanding their stratigraphic and structural history.

Regionally, Lerand (1973) correlated the pre-Mississippian rocks of the northeastern Brooks Range with lower Paleozoic sedimentary rocks in the Canadian Arctic Islands that were deformed during the Franklinian orogeny (Figure 1.4). Emphasizing this correlation, Lerand then proposed the term 'Franklinian sequence' to refer to the entire pre-Mississippian sequence in the northeastern Brooks Range. However, this correlation has not yet been convincingly documented and may be overly simplistic. Norris and Yorath (1981) have suggested that a major unconformity at the base of the Cambrian sequence in both northwestern Canada and northeastern Alaska divides the entire pre-Mississippian sequence into a lower Paleozoic succession and an older Proterozoic sequence. If true, referring to the entire pre-Mississippian sequence in the northeastern Brooks Range as the 'Franklinian sequence' would be erroneous and misleading. On that basis, I have avoided using the term 'Franklinian sequence' throughout this dissertation.

The correlation between the lower Paleozoic rocks of the northeastern Brooks Range and similar age rocks in the northwestern Yukon Territories of northwestern Canada is becoming better established (Norris and Yorath, 1981; Lane and Cecile, 1989; Lane and others, 1991). The lower Paleozoic rocks in northwestern Canada are thought to represent basinal deposits of a west-facing Proterozoic to early Paleozoic continental margin and are correlated with lower Paleozoic rocks of the Selwyn Basin (Cecile, 1988). However, the stratigraphic relationship between these basinal deposits of the northwestern Yukon and the lower Paleozoic rocks of the Canadian Arctic Islands remains unclear, and the two areas are separated by a major fault system of unknown displacement (Figure 1.4).

The Proterozoic sequence in the northeastern Brooks Range of Alaska and Canada is even less understood than that of the overlying lower Paleozoic rocks. Norris and Yorath (1981) have suggested that these older rocks may be correlative with Proterozoic rocks of the Inuitian sequence of the Canadian Arctic Islands. However, there is little stratigraphic work presently available to support this hypothesis.

The relationship betwen the pre-Mississippian rocks of the northeastern Brooks Range with similar age rocks in the main axis of the Brooks Range is also poorly constrained (Figure 1.4). While correlation of the pre-Mississippian rocks of northeastern Alaska with those of the northwestern Yukon is facilitated by a continuous outcrop belt

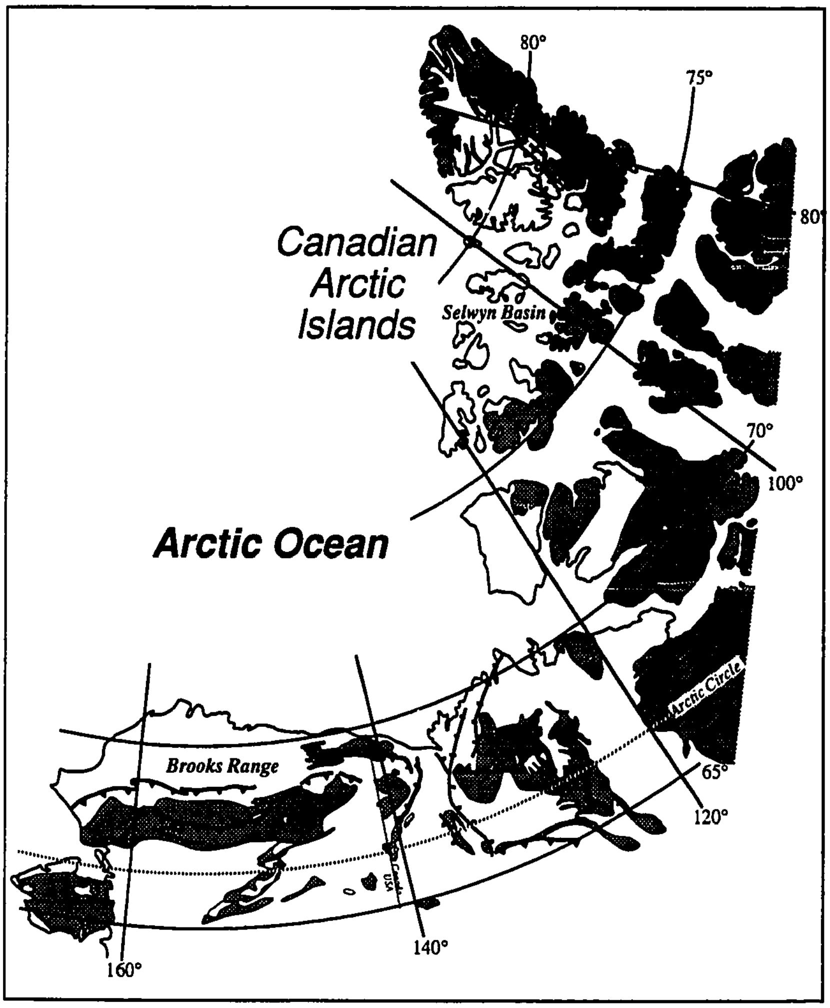


Figure 1.4. Distribution of pre-Mississippian rocks in the Canadian and Alaskan arctic regions, based on Okulitch and others, 1989.

between the two areas, the pre-Mississippian rocks of the main axis of the Brooks Range are separated from those of the northeastern Brooks Range by exposures of the younger cover sequence. In addition, restoration of the Mesozoic structural shortening between the two outcrop belts suggests that the pre-Mississippian rocks of the central Brooks Range may have originated hundreds of kilometers south of their present location (Oldow and others, 1987b). The stratigraphy and structural style of the two groups of pre-Mississippian rocks are also quite different. To date, there is little convincing data correlating the two groups before Devonian time, when both sequences were overlain by Devonian clastic rocks (Mull, 1982; Reiser and others, 1980; Moore and Nilsen, 1984; Anderson, 1991)

As with the stratigraphy, the pre-Mississippian structural history of the older rocks is poorly understood. Norris and Yorath (1981) suggested that in northwestern Canada at least one deformational event pre-dated the unconformity at the base of the lower Paleozoic sequence. However, the existence of this Proterozoic event, much less its style and vergence, has yet to be well documented in either northeastern Alaska or northwestern Canada. In the northeastern Brooks Range of Alaska, the pre-Mississippian rocks have experienced at least two deformational events of Late Silurian to Middle Devonian age (Grantz and others, 1990). These deformational events resulted in an angular unconformity at the base of the overlying Mississippian and younger cover sequence. However, the vergence and style of these two early Paleozoic events are poorly constrained. The earlier, Late Silurian to Early Devonian event appears to have involved an early component of shortening followed by an extensional event resulting in formation of isolated non-marine basins. A later, Middle (?) Devonian event deformed these extensional basins. Bedding that is generally upright and south-dipping within the pre-Mississippian rocks and apparently truncanted by the sub-Mississippian angular unconformity throughout the northern part of the northeastern Brooks Range has led many workers to suggest that at least one of these early Paleozoic events was north-vergent (e.g., Hubbard and others, 1987; Lane and others, 1991). However, pre-Mississippian-age mesoscopic structures indicating south-vergence have been reported from pre-Mississippian rocks of the Franklin

Mountains and British Mountains (Oldow and others, 1987a; Avé and Oldow, 1987). These structures have been interpreted to reflect south-vergence during the Devonian orogeny and correlated with south-vergent structures of the Late Devonian to Early Mississippian Ellesmerian orogeny of the Canadian Arctic Islands (Oldow and others, 1987a). However, the extent of these south-vergent structures, the age of the rocks deformed by them and the age of the deformation have not been convincingly documented at this time.

Although this is primarily a structural study, it is necessary to have at least a provisional stratigraphy in order to identify and properly interpret map-scale structures within the pre-Mississippian rocks. Thus, part of this study required developing a generalized stratigraphy for the pre-Mississippian rocks of the study area. This stratigraphic information, in conjunction with observations regarding the pre-Mississippian structural style of the pre-Mississippian rocks of the study area, provides a means of comparing the pre-Mississippian rocks of the northeastern Brooks Range of Alaska with those of northwestern Canada and the rest of the circum-Arctic region. The result of this study, in conjunction with similar studies elsewhere in the northeastern Brooks Range of the U. S. and Canada, could provide valuable clues regarding the Proterozoic to early Paleozoic depositional and tectonic history of northern Alaska and Canada.

The tectonic history of the pre-Mississippian sequence in northeastern Alaska also has implications for the origin of the Canada basin, one of the more enigmatic and least understood of the world's ocean basins. Due to extensive ice cover, thick basin sediments and poorly defined magnetic anomaly patterns, interpretation of magnetic lineations has not yielded a unique solution to the question of the origin of the basin. Different models abound for the opening of the Canada basin, constrained by relatively limited geophysical and geological data (see Nilsen, 1981, or Lawver and Scotese, 1990, for a summary). The inaccessibility and relative unknown nature of the geology of the circum-Arctic regions have exacerbated this problem. Thus, the origin of Arctic Alaska remains controversial and somewhat model-dependent. A better knowledge of the stratigraphy and structural history of the pre-Mississippian rocks of the northeastern Brooks Range could help constrain these models, both by providing a basis for correlation with pre-Mississippian rocks elsewhere in the circum-Arctic, and by shedding light on the pre-Mississippian structural evolution of northeastern Alaska and northwestern Yukon (e.g., Oldow and others, 1987a; Lane and Cecile, 1989; Lane and others, 1991).

1.3.2 Cenozoic structural evolution of the northeastern Brooks Range

The northeastern Brooks Range is a relatively young and still actively growing portion of the Brooks Range, as suggested by deformed Tertiary sediments and active seismicity (Grantz and others, 1983; Moore and others, 1985a; Carter and others, 1986). However, both the direction of tectonic transport during this deformation and its driving mechanism are not understood. One of the most striking features of the northeastern Brooks Range of both Alaska and northwestern Canada is the obvious arcuate nature of the fold-and-thrust belt (Figure 1.4). This arcuate trend may have been the result of several different combinations of the timing and transport direction of deformation (Wallace, 1990). In order to understand the Cenozoic structural evolution of the entire northeastern Brooks Range, the origin of the arcuate trends and, ultimately, the driving force for the deformation, it is necessary to at first determine how various parts of the range evolved. This requires understanding how each stratigraphic sequence behaved during Cenozoic deformation, what factors controlled that structural behavior, how each sequence may have influenced the behavior of the other sequences and the amount of Cenozoic tectonic shortening. Pre-Mississippian rocks have played an important role in this Cenozoic thrusting event. Which structures are pre-Mississippian in age, which are Cenozoic in age, and the role of the pre-Mississippian structures in Cenozoic deformation are all questions critical in unravelling the Cenozoic deformational history of the region. Detailed structural studies such as this one can provide information on the Cenozoic structural style and transport direction for a limited area of the fold-and thrust belt. However, similar studies throughout the northeastern Brooks Range of Alaska and Canada are necessary in order answer the broader tectonic questions that affect the origin of the entire range.

1.4 Questions relevant to the structural evolution of fold-and-thrust belts in general.

1.4.1 Influence of stratigraphy on the geometry of fold-and-thrust structures

Studies of other foreland fold-and-thrust belts suggest that lateral changes in the character and/or extent of a regional detachment horizon can influence the structural style of an entire structural/stratigraphic package. For example, unconformities and lateral facies changes within a detachment horizon can control the lateral extent of that detachment horizon, resulting in changes in the deformational style of the overlying rocks both along strike and towards the foreland (e.g., the Keuper Salt of the Jura Mountains, described in Laubscher, 1972 and Bachmann and others, 1982). Structural termination of a detachment horizon can result in ramps in the basal detachment surface (e.g., the Salt Range Formation of Pakistan, McDougall and Hussain, 1991; Moussouris and Davis, 1989). In the northeastern Brooks Range, a major detachment horizon in a thick shale near the base of the Ellesmerian cover sequence has permitted the cover sequence to deform independently of the underlying pre-Mississippian rocks. However, the thickness and lithologic character of this detachment horizon vary widely across the region. Changes in the structural style of the overlying cover sequence appear to correspond to some of these stratigraphic variations. Study of these variations in structural style may provide some indication as to how changes in the lithology and thickness of a siliciclastic detachment horizon has influenced the structural style of the overlying package of rocks.

1.4.2 Influence of 'depositional basement' on the character of fold-and-thrust deformation

Foreland fold-and-thrust belts generally form in previously undeformed stratified rocks and, as they evolve, advance into the deposits of the adjacent synorogenic foredeep basin (e.g., Canadian Rockies, Bally and others, 1966, and Price, 1981; Alps, Laubscher, 1972, and Butler and others, 1985). The depositional basement that underlies these cover rocks generally is not involved in deformation in the range-front region, although these older rocks may predominate in the interior of the mountain belt (Hatcher and Williams, 1986). Involvement of depositional basement near the leading edge of a fold-and-thrust belt is unusual, and when it occurs may provide significant insights into the evolution of the fold-and-thrust belt.

In many mountain belts, depositional basement consists of significantly older crystalline and polydeformed igneous and high-grade metamorphic rocks and is generally considered synonymous with mechanical basement. However, depositional basement and mechanical basement may not always coincide. In some mountain belts, depositional basement may consist of previously deformed, perhaps weakly metamorphosed, stratified rocks. In the latter instance, although these rocks may be similar to the overlying cover sequence in their stratified character, their previous deformation has resulted in lithologic discontinuities and structural characteristics that may have profound effects on both their later deformation and that of the cover.

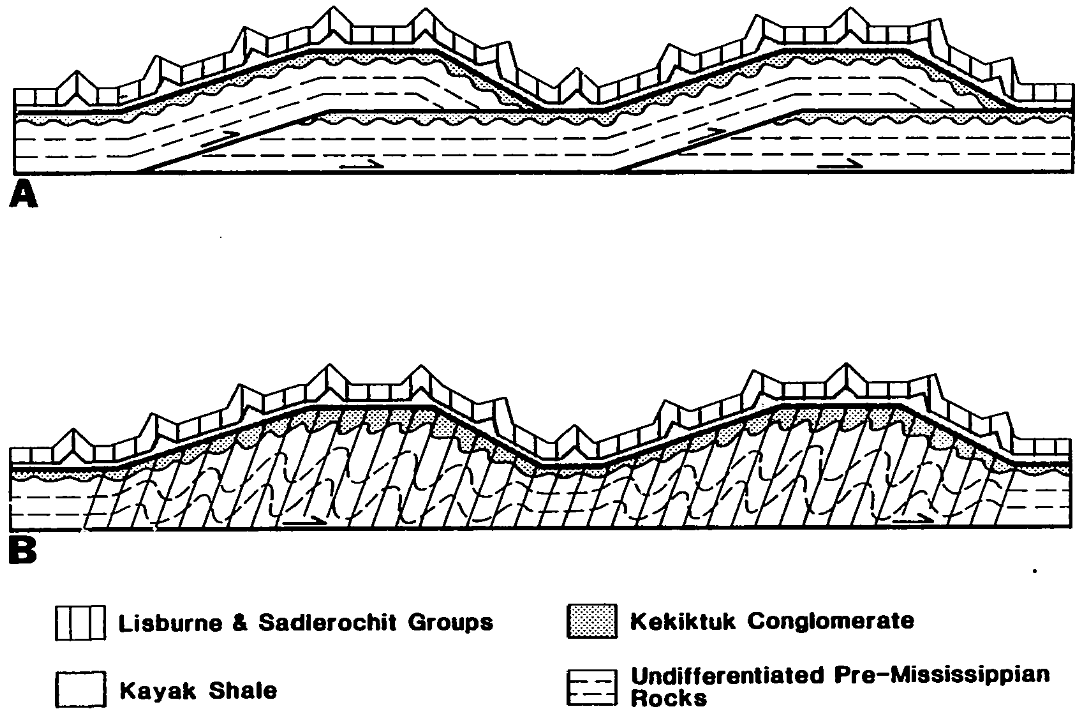
The northeastern Brooks Range is unusual in that previously deformed and weakly metamorphosed sedimentary and volcanic rocks of the depositional basement (the pre-Mississippian sequence) were involved in foreland fold-and-thrust belt deformation and structurally elevated immediately adjacent to deposits of the foreland basin. The reasons for the involvement of depositional basement in the foreland fold-and-thrust belt of the northeastern Brooks Range, and how these older, multiply-deformed rocks have responded to Cenozoic deformation, are two major questions that need to be addressed in order to understand the structural evolution of the region.

The behavior of depositional basement during Cenozoic thrusting in the northeastern Brooks Range is further complicated by the fact that at least two different types of depositional basement are involved. Most of the depositional basement in the eastern portion of the northeastern Brooks Range consists of stratified, weakly-metamorphosed sedimentary and volcanic rocks (Reiser and others, 1980). This assemblage is lithologically very heterogeneous and contains many potential detachment horizons. However, the pre-Mississippian sequence of the northeastern Brooks Range also includes the Okpilak batholith, a large Devonian granitic batholith that has been involved in the range-front deformation (Sable, 1977; Reiser and others, 1980; Hanks and Wallace, 1990). This batholith is mechanically very homogeneous, with no obvious internal detachment horizons. The contrast in the Cenozoic structural behavior between the stratified rocks and the batholith may provide valuable information on how the character of depositional basement rocks may control their structural behavior in foreland fold-and-thrust deformation.

The behavior of stratified depositional basement during fold-and-thrust deformation has been partially addressed by local detailed and regional studies in the western part of the northeastern Brooks Range along the Canning River (Figure 1.3). These studies have resulted in two different and apparently contradictory models for the mode of deformation of the depositional basement during the Cenozoic formation of the northeastern Brooks Range (Figure 1.5). Model A, based on regional studies and mapping of map-scale structures, proposes that most of the Cenozoic shortening in the pre-Mississippian rocks was accommodated by large-scale thrust duplication of the depositional basement in a regional duplex between a floor thrust at depth in the depositional basement and a roof thrust in a shale near the base of the overlying Ellesmerian cover sequence (Rattey, 1985; Namson and Wallace, 1986; Kelley and Foland, 1987; Leiggi, 1987; Wallace and Hanks, 1990). Model B is based on local detailed analysis of mesoscopic structures within the pre-Mississippian sequence which suggests that Cenozoic shortening within the depositional basement is accommodated via the development of penetrative mesoscopic and microscopic structures (Oldow and others, 1987a).

A detailed structural analysis of a relatively large region involving different types of depositional basement (such as this study) is critical in order to determine which of these two models most accurately describes the deformational style of depositional basement in the region. In the process, it should be possible to determine if the composition of the depositional basement could play a role in determining which mechanism of shortening dominated during shallow-level thrusting.

Figure 1.5. Conceptual end-member models for the mode of shortening of the pre-Mississippian rocks and the Kekiktuk Conglomerate during Cenozoic thrusting. For simplicity and clarity, the attitude of the pre-Mississippian rocks with respect to the unconformity surface, an artifact of pre-Mississippian-age deformation, has not been included. The dashed lines within the pre-Mississippian rocks represent arbitrary markers that were horizontal prior to Cenozoic deformation, and do not represent bedding. Model A represents a north-vergent regional duplex, where Cenozoic shortening of the pre-Mississippian rocks is accommodated by thrust duplication. The floor thrust of the duplex is at depth in the pre-Mississippian rocks, and the roof thrust is in the Mississippian Kayak Shale. Pre-Cenozoic horizontal markers remain horizontal with respect to the unconformity during this style of deformation. Model B represents a scenerio where Cenozoic shortening within the pre-Mississippian rocks is accommodated primarily by strain and mesoscopic structures, and results in folding of the pre-Cenozoic horizontal markers. Steeply dipping solid lines within the pre-Mississippian rocks and Kekiktuk Conglomerate represent Cenozoic-age cleavage.


A
B
Lisburne & Sadlerochit Groups
Kekiktuk Conglomerate
Kayak Shale
Undifferentiated Pre-Mississippian Rocks

1.4.3 Use of models to determine the gross geometry of a foreland fold-and-thrust belt

Balanced cross sections have become a standard technique used in illustrating and studying fold-and-thrust belts worldwide, and are useful at all scales of structural analysis, from regional tectonic synthesis to strain analysis (e.g., Dahlstrom, 1969; Price, 1981; Boyer and Elliot, 1982; Woodward and others, 1986). Such sections are becoming more popular in petroleum exploration, where abundant well and seismic data can be utilized to extrapolate structures to greater depths and lateral extents.

Unfortunately, not all fold-and-thrust belts readily lend themselves to construction of well-constrained balanced sections, generally due to a lack of seismic, well, and/or surface data. These fold-and-thrust belts are generally in remote areas where both academic research and industry exploration are hampered by high costs and/or politically or geographically hostile environments. This is especially unfortunate because construction of a balanced cross section often identifies important structural and stratigraphic constraints on the geometry of surface and subsurface structures. A balanced cross section can therefore highlight those areas needing further study as well as what additional types of data might be useful in order to refine the structural interpretation of the region.

The northeastern Brooks Range offers a good opportunity to investigate the most effective means of constructing reasonable balanced cross sections in remote, relatively unknown areas. Most of the northeastern Brooks Range is mapped at only a reconnaissance level, with only a few local detailed studies (including this study). Seismic data, critical in evaluating the subsurface portion of the fold-and-thrust belt, are publicly available only for the foreland basin immediately north of the range front, are of poor quality, and do not continue into the range. The questions of how surface structures in the mountains compare with those in the subsurface to the north, and how much shortening has occurred across the entire fold-and-thrust belt, are important from both a regional and petroleum exploration perspective. However, due to the lack of good seismic data and widespread detailed surface information, few constraints are available for the construction of detailed balanced cross sections.

This lack of information allows multiple structural interpretations, and could lead to a time-consuming process of constructing multiple detailed balanced cross sections in order to explore a variety of ideas regarding the gross geometry of the orogenic wedge. In this study, I have attempted to make this process more efficient and cost effective by exploring the range of possible orogenic wedge geometries by area-balancing simple models of the fold-and-thrust belt. These models illustrate the range of possible structural solutions simply and rapidly, and make it possible to explore a variety of different geometries. These models also serve to illustrate a number of key regional controls on the geometry of the fold-and-thrust belt, which can serve as a focus for future research.

1.4.4 The tectonic setting of continental fold-and-thrust belts.

Continental fold-and-thrust belts commonly form on the outer margins of a collisional zone between two lithospheric plates, resulting from attempted subduction of a passive continental margin. The sedimentary wedge of the subducting margin is reactivated as an orogenic wedge and transported toward the interior of the continent, generally preceded by a foreland sedimentary basin (Figure 1.6 A)

The northeastern Brooks Range is an example of an unusual continental fold-and-thrust belt that has developed on a narrow subducting continental fragment bounded by two passive continental margins (Figure 1.6 B). The growing Cenozoic fold-and-thrust belt is collapsing a late Paleozoic and early Mesozoic passive continental margin, and transporting these sedimentary rocks across a relatively narrow continental fragment towards a younger, Cretaceous-age passive continental margin. Information on the relationship between the timing of the thrusting, subsidence of the foreland basin and the geometry of the orogenic wedge may help in understanding the influence of the extent and thickness of the continental crust on the dynamics of superimposed fold-and-thrust deformation worldwide. This study, while not addressing this issue directly, will provide information on the geometry of the fold-and-thrust belt that could eventually be applied to this problem.

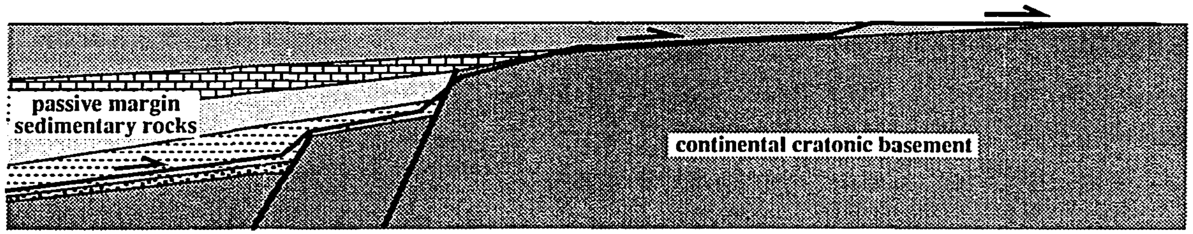


Figure 1.6 A. Schematic cross section of a 'classic' continental margin prior to thrusting. In this situation, passive continental margin sediments are incorporated into the orogenic wedge. These frequently deeper-water sediments are thrust onto thinner and shallower-water sediments of similar age that were deposited higher up on the continental margin, sometimes on cratonic basement. Depositional thinning of the passive margin sedimentary rocks towards the continental interior often aids in the development of the characteristic wedge-shape of a classic orogenic belt. The orogenic sole fault becomes progressively shallower towards the foreland, where it either intersects the surface and/or loses slip and dies out in foredeep deposits.

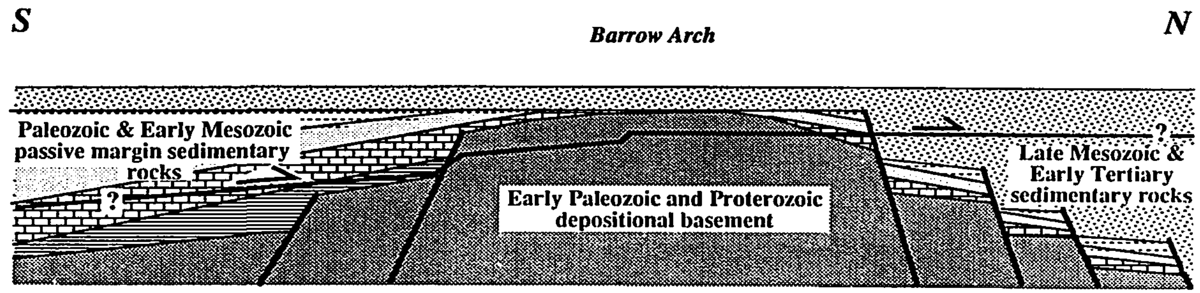


Figure 1.6 B. Schematic cross section of a continental fragment with two passive continental margins of different ages and opposite orientations prior to a collapse by thrusting. Passive margin sediments of one margin may be thrust onto the shallow water sediments related to the second passive margin. This model may represent the northeastern Brooks Range. North-derived sedimentary rocks of a Paleozoic and Early Mesozoic passive continental margin are preserved on the south side of the Barrow Arch, and possibly in isolated basins on the north side of the arch. Clastic rocks of the younger Cretaceous passive continental margin were derived from the south and thicken rapidly on the north side of the arch. Cenozoic thrust faulting in the northeastern Brooks Range post-dates formation of both margins.

1.5 Area and approach of study

This study focuses on the structural geometry and evolution of the eastern range-front region of the Arctic National Wildlife Refuge (ANWR) of the northeastern Brooks Range. Two different areas along the range front (Figure 1.3) in this portion of ANWR are included in the study. The two areas represent two of the three structural provinces of the northeastern Brooks Range (see Wallace and Hanks, 1990 and Chapter 2) and differ greatly in both the lithology and Cenozoic structural style of the pre-Mississippian rocks, as well as in the structural style of the Mississippian through Triassic cover sequence. Comparison of the structural evolution of these two areas can help elucidate the influence of stratigraphy on the structural style both within these two structural provinces and throughout the northeastern Brooks Range.

The two study areas form the central portion of two regional transects across the two structural provinces of northeastern ANWR. Detailed mapping of both areas at 1:25,000 scale was conducted via helicopter-placed spike camps during four field seasons (mid-June to early August; 1986, 1987, 1988 and 1989). This detailed information was combined with published regional surface geologic data and published seismic data from the ANWR coastal plain and used to constrain alternate structural models and detailed balanced sections of each transect. The models and cross sections were then used as a basis for comparing and contrasting the structure of the two transects and the structural provinces they represent.

The first and larger study area comprises the central portion of the Aichilik River transect (A-A', Figure 1.3). It is an east-west trending, irregularly 'H' shaped area across the northern anticlinorium between the Aichilik and Egaksrak rivers (area 1, Figure 1.3). The northern limb of the 'H' extends for approximately 24 miles (38 km) along Leffingwell Ridge. The cross-bar of the 'H' extends south from Leffingwell Ridge across the pre-Mississippian rocks of the core of the anticlinorium to the next outcrop belt of Mississippian and younger cover rocks to the south. The southern limb of the 'H' is approximately 7.5 miles (12 km) long and parallels the southern belt of Mississippian and younger rocks on the southern limb of the anticlinorium. Approximately 140 square miles (360 square kilometers) were mapped at a scale of 1:25,000 during three field seasons (1986, 1987 and 1988). These detailed maps are available as Alaska Division of Geological and Geophysical Surveys (ADGGS) Public Data Files (Hanks, 1987, 1988, 1989). A simplified geologic map of the entire area is included in chapter 3 as figure 3.5. This area has relatively gentle topography with elevations ranging from 1200 to 4000 feet (360 m to 1200 m), with vertical relief rarely exceeding 2,000 feet (600 meters). In general, the area has only fair exposure, as is typical of thrust front regions where topography is often modest. Outcrops vary from good to nonexistent, with the best exposures being along ridge crests and stream cuts. Pre-Mississippian rocks are generally poorly exposed, limiting what could be learned from these rocks in this area.

The second and smaller area comprises the central portion of the second regional transect, the Okpilak batholith transect (B-B', Figure 1.3). It is located along the range front to the west along the northern margin of the Okpilak batholith (area 2, Figure 1.3). This area is considerably steeper than that along Leffingwell Ridge, with elevations ranging from 3000 to 7000 feet (900 to 2100 meters). Exposures vary from fair to excellent, with the granite and limestones being the best exposed. Approximately 30 square miles (78 square kilometers) were mapped at 1:25,000 scale during two field seasons (1988 and 1989). A detailed geologic map is included in this dissertation as Appendix A. A simplified geologic map of the area is included in chapter 4 as figure 4.2.

1.6 Organization of dissertation

The dissertation consists of a series of papers that are published or will be submitted for publication. Each paper is included in the dissertation as a chapter. Some overlap in the content of the chapters is unavoidable, since each chapter must stand alone as a separate paper.

Chapter 2 is a summary of the regional setting and structural provinces of the northeastern Brooks Range and has been published in the American Association of Petroleum Geologists (AAPG) *Bulletin* under the title "Structural provinces of the northeastern Brooks Range, Arctic National Wildlife Refuge, Alaska" (Wallace and Hanks, 1990). This paper was coauthored with W. K. Wallace and summarizes the influence that stratigraphy has had on the overall structural geometry of the northeastern Brooks Range, as well as providing an abbreviated discussion of the structural evolution of the region. I contributed the field observations and conclusions regarding the structure of northeastern ANWR, constructed the structure contour maps of the sub-Mississippian unconformity, and developed a model for the style of Cenozoic deformation within the pre-Mississippian rocks of the Aichilik River and Okpilak batholith transects. In addition, I helped write and revise the manuscript.

Chapter 3 focuses on a detailed discussion of the structural geometry and evolution of the eastern study area along Leffingwell Ridge. I plan to submit this paper to the Geological Society of America *Bulletin* under the title "Thin-skinned thrusting in non-crystalline basement rocks: an example from the northeastern Brooks Range, Alaska." This paper documents the influence of the stratified pre-Mississippian rocks on the Cenozoic formation of the anticlinorium and illustrates how changes in the detachment horizon near the base of the cover sequence influenced the structural geometry of the overlying rocks. This chapter also includes a discussion of area-balanced models and how they can be used to provide constraints on the geometry of the orogenic wedge in this area.

Chapter 4 is a short but detailed summary of the structure of the northern margin of the Okpilak batholith, and includes a discussion of the structural evolution of the Okpilak batholith transect. This chapter has been published in *Geology* under the title "Cenozoic thrust emplacement of a Devonian batholith, northeastern Brooks Range: Involvement of crystalline rocks in a foreland fold-and-thrust belt" (Hanks and Wallace, 1990). This paper was co-authored with W. K. Wallace, who provided a regional perspective on the problem, assisted in the gathering and interpretation of the field data, and reviewed the manuscript. I did the majority of the field work, interpretation and writing.

Chapter 5 discusses the use of area-balanced models in determining regional constraints on the gross geometry of a fold-and-thrust belt, focusing on the uses of these models in areas of little data. The Okpilak batholith transect is used as an example. I plan to submit this paper to AAPG *Bulletin* in the near future.

Chapter 6 compares the two transects and summarizes the conclusions of the study. This chapter discusses the two main points of the dissertation: how the composition of the pre-Mississippian sequence influenced its structural style during Cenozoic deformation and how the stratigraphy of the detachment horizon near the base of the overlying Ellesmerian cover sequence influenced the Cenozoic structural style of the Ellesmerian sequence.

Detailed maps of the eastern study area along Leffingwell Ridge with accompanying stratigraphic descriptions and preliminary field observations have been published as ADGGS Public Data Files (Hanks, 1987, 1988, 1989) and are available from ADGGS, 794 University Ave., Fairbanks, Alaska 99709. 1:125,000 scale balanced cross sections of the Aichilik River and Okpilak batholith transects are also available as a ADGGS Public Data File (Hanks, 1990). A detailed map of the Okpilak batholith area is included in the dissertation as an appendix, as is a brief summary of the stratigraphy and a compilation of the area-balanced models developed for both transects.

CHAPTER 2: REGIONAL GEOLOGIC SETTING[1]

2.1 Abstract

The dominant Cenozoic structures of the northeastern Brooks Range are anticlinoria cored by pre-Mississippian rocks, reflecting a regional north-vergent duplex with a floor thrust in the pre-Mississippian sequence and a roof thrust in the Mississippian Kayak Shale. The number of horses forming each anticlinorium and the structural style of the overlying Mississippian and younger cover sequence varies regionally, providing a basis for dividing the northeastern Brooks Range into structural provinces. In the western province, each anticlinorium contains a single horse, and shortening above the Kayak Shale was accommodated mainly by detachment folds. To the north in the Sadlerochit Mountains, the Kayak Shale is depositionally discontinuous and rocks elsewhere separated by this detachment deformed together. In the eastern province, each anticlinorium contains multiple horses, and shortening above the Kayak Shale was accommodated largely by thrust duplication of Mississippian through Triassic rocks. In the narrow central province, the Devonian Okpilak batholith was detached from its roots, internally shortened along shear zones and by penetrative strain, and transported northward. Because the Kayak Shale is locally absent, the Mississippian and younger cover sequence deformed with the batholith, in part penetratively.

East-northeast trends formed where pre-Mississippian rocks were not involved in deformation, and probably are normal to the direction of Cenozoic tectonic transport. East trends formed where pre-Mississippian rocks were involved in deformation, and probably reflect a pre-Mississippian structural grain. At any given location, east trends generally post-date east-northeast trends, reflecting a drop over time of the basal detachment into pre-

[1] Chapter 2 contains the complete text and figures of the manuscript, *Structural provinces of the northeastern Brooks Range, Arctic National Wildlife Refuge, Alaska,* by W. K. Wallace and C. L. Hanks, as published in the American Association of Petroleum Geologists *Bulletin, v. 74, no. 7, pp 1100-1118, 1990.*

Mississippian rocks.

2.2 Introduction.

Well-exposed structures in the northeastern Brooks Range fold-and-thrust belt (Figure 2.1) may provide insights into the evolution of similar structures elsewhere in the world, as well as offering clues to the factors that control their geometry. In addition, the northeastern Brooks Range includes the nearest well- exposed analogs to structures that may underlie the Arctic coastal plain immediately to the north, the most promising area for onshore hydrocarbon exploration remaining in North America.

The stratigraphy of the northeastern Brooks Range has had a significant influence on the geometry of structures formed during deformation, as is true in many other fold-and-thrust belts (Woodward and Rutherford, 1989). The interlayering of strata of differing thickness, lithology, and structural competency has resulted in a structural stratigraphy in which particular stratigraphic intervals display a specific structural style. Several different structural provinces can be defined in the northeastern Brooks Range based upon lateral variations in structural style (Figure 2.2). These lateral variations commonly correspond with lateral variations in stratigraphy.

Recent discussions of the structural geometry and evolution of the northeastern Brooks Range have dealt mainly with the western part of the region (for example, Kelley and Foland, 1987; Leiggi, 1987; Oldow and others, 1987a). In this paper, we will illustrate the variations in structural geometry that exist over a much larger region, and will argue that lateral changes in stratigraphy influence the style of deformation. The objective of this paper is to provide a regional overview of the structure of the northeastern Brooks Range, and an interpretation of its structural evolution that incorporates the influence of variations in stratigraphy on the structural geometry of the fold-and-thrust belt. This overview and interpretation are based primarily on our own detailed geologic studies throughout the northeastern Brooks Range, complemented by studies of specific structural problems by graduate students at the University of Alaska. However, it is not our intention

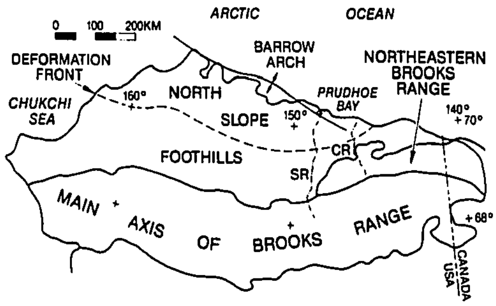


Figure 2.1. Map of northern Alaska, showing the major physiographic and tectonic provinces, including the northeastern Brooks Range. SR = Sagavanirktok River, CR = Canning River.

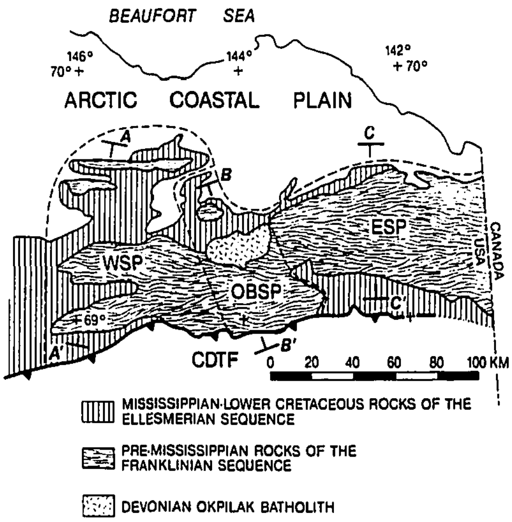


Figure 2.2. Generalized geologic map of the northeastern Brooks Range showing the major structural provinces. Brackets indicate approximate locations of schematic cross sections shown in figure 5. WSP = western structural province, OBSP = Okpilak batholith structural province, ESP = eastern structural province, CDTF = Continental Divide thrust front.

to do more than summarize the results of these studies here. Rather, we seek in this paper to provide a conceptual and testable regional structural interpretation that will serve as a framework for future, more detailed papers, and for further detailed structural and stratigraphic studies.

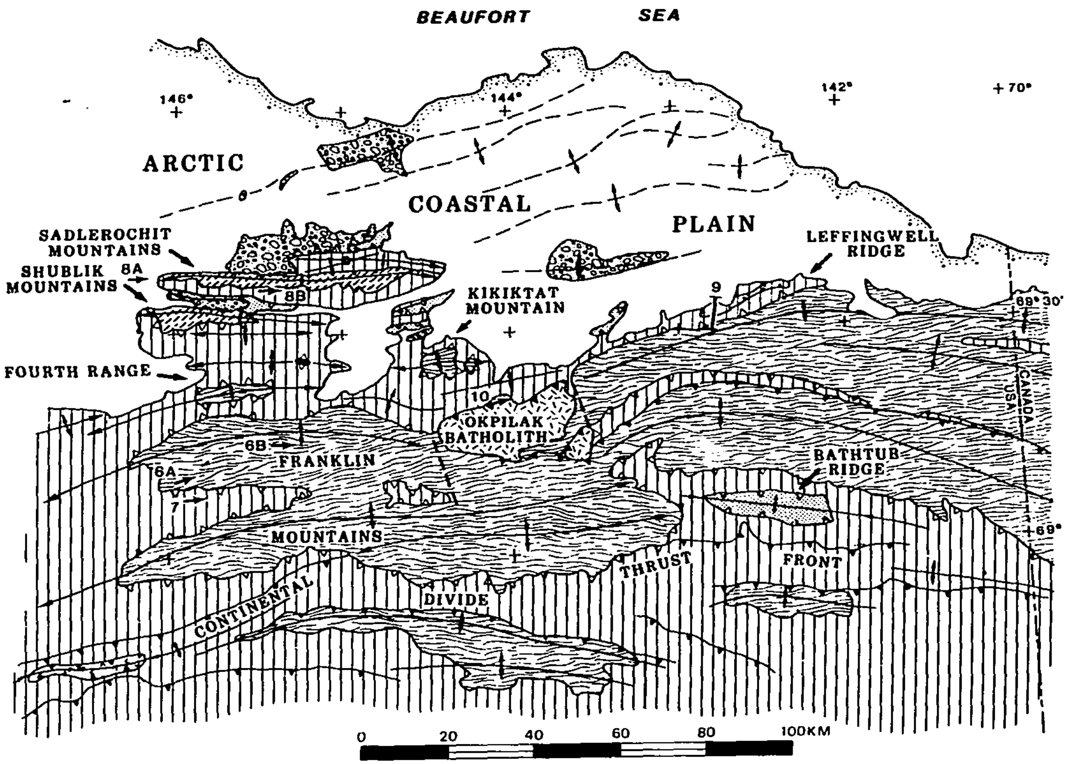
2.3 Regional geologic setting

The northeastern part of the Brooks Range differs in several important respects from the remainder of the Brooks Range. The main axis of the Brooks Range trends eastward from the Chukchi Sea to the Canadian border (Figure 2.1). The northeastern Brooks Range, however, projects north of the mountain front of the western and central Brooks Range, forming a prominent northward-convex arcuate topographic salient that contains the topographically highest parts of the Brooks Range.

The stratigraphic sequence exposed in the northeastern Brooks Range (Figures 2.3, 4) is analogous to the autochthonous stratigraphic sequence that characterizes the North Slope petroleum province, in the subsurface of the Arctic coastal plain (Reiser, 1970). Three major depositional sequences have been identified in the northeastern Brooks Range (Figure 2.4; Bader and Bird, 1986; Bird and Molenaar, 1987). The Proterozoic to Devonian Franklinian sequence consists of a lithologically heterogeneous assemblage of rocks that were strongly deformed and weakly metamorphosed during one or more pre-Mississippian deformational events. A major angular unconformity marks the top of the Franklinian sequence. The overlying Mississippian to Lower Cretaceous Ellesmerian sequence consists of marine carbonate and clastic strata deposited on the flanks of a relative continental high to the north, now absent due to formation of the northern Alaska continental margin by rifting in Early Cretaceous time (Grantz and May, 1983). The Lower Cretaceous to Cenozoic Brookian sequence consists of detritus derived from the Brooks Range to the south.

The rocks of the main axis of the Brooks Range were shortened by hundreds of kilometers during Middle Jurassic to Early Cretaceous time in a dominantly north-vergent

Figure 2.3. Generalized tectonic map of the northeastern Brooks Range showing the distribution of the major structural-stratigraphic units and structural features of the northeastern Brooks Range (modified from Brosgé and Reiser, 1965; Brosgé and others, 1976; Bader and Bird, 1986; Clough and others, 1987). Structural-stratigraphic units are as identified in Figure 2.2. Solid teeth on thrust faults indicate older-over-younger thrust faults that duplicate stratigraphic section; open teeth indicate detachment surfaces along which there has been slip but no disruption of the normal stratigraphic succession.



BEAUFORT SEA
146°
144°
142°
+70°
ARCTIC
COASTAL
PLAIN
SADLEROCHIT MOUNTAINS
SHUBLIK MOUNTAINS
8A
8B
KIKIKTAT MOUNTAIN
LEFFINGWELL RIDGE
9
69°30'
FOURTH RANGE
10
OKPILAK BATHOLITH
6B
6A
7
FRANKLIN
MOUNTAINS
BATHTUB RIDGE
USA
CANADA
69°
FRONT
THRUST
DIVIDE
CONTINENTAL
0
20
40
60
80
100KM

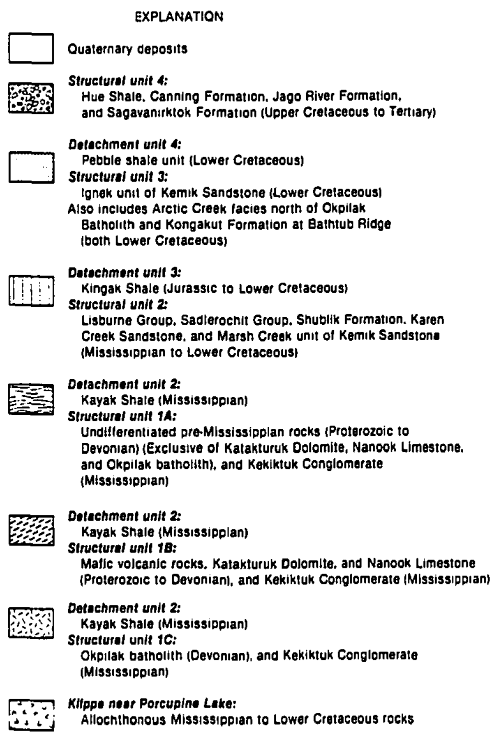


Figure 2.3. (cont.)

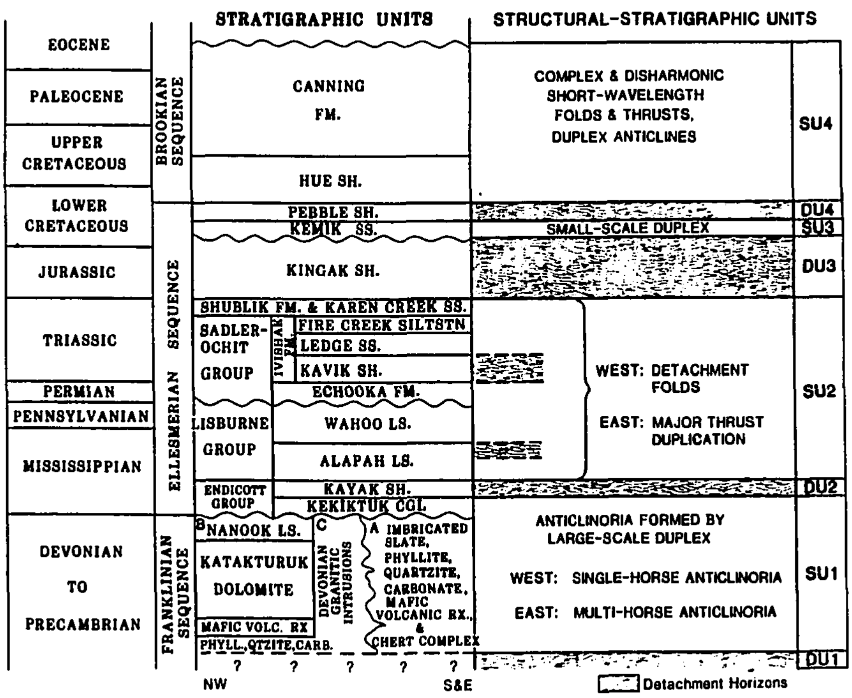


Figure 2.4. Structural stratigraphy of the northeastern Brooks Range. A generalized stratigraphic column is shown to the left, and the structural-stratigraphic units and their dominant characteristics are shown to the right. Thicknesses are not to scale. Detachment horizons are shown with a wavy line pattern, which extends only partly across column for minor detachment horizons. SU = structural unit (competent), DU = detachment unit (incompetent). The distribution of sub-units (A,B, and C) of structural unit 1 is shown in Figure 2.3.

fold-and-thrust belt (Mull, 1982; Mayfield and others, 1983; Oldow and others, 1987b). However, to the north, the rocks of the northeastern Brooks Range have been shortened by less than a hundred kilometers (Namson and Wallace, 1986; Leiggi, 1987; Hanks and Wallace, 1990). The deformational style of the northeastern Brooks Range appears superficially to be much simpler than that of the rest of the Brooks Range (Reiser, 1970; Mull, 1982). Its structure is dominated by a series of anticlinoria, each with a central "core" of pre-Mississippian rocks and a Mississippian and younger cover deformed into shorter wavelength folds. In the northeastern Brooks Range, the Mississippian and younger cover rocks display neither the obvious sense of north-vergence nor the abundance of exposed thrust faults characteristic of Upper Devonian and younger rocks in the main axis of the Brooks Range.

The northeastern Brooks Range is also unusual in that it contains extensive exposures of pre-Upper Devonian rocks, which otherwise are restricted mainly to the interior of the orogen, in the southern Brooks Range. Widespread exposure of such deep structural levels is unusual so close to the foreland of an orogenic belt. Mississippian and younger rocks thin northward onto the Barrow arch, a structural high that has persisted from at least Mississippian time to the present and that trends east-northeast toward the range front of the northeastern Brooks Range (Figure 2.1; Mull, 1982; Grantz and May, 1983). Northward migration of the deformation front onto this structural high during formation of the northeastern Brooks Range probably accounts for the involvement of pre-Upper Devonian rocks. The age of rocks exposed in the northeastern Brooks Range contrasts markedly with that of rocks exposed in the foothills along strike to the west, which consist of deformed Cretaceous rocks that were deposited in the foredeep of the Brooks Range and deformed late during the evolution of the main axis of the range.

2.3.1 Boundaries of the northeastern Brooks Range.

The arcuate salient of the northeastern Brooks Range is bounded to the north by a well defined topographic range front that coincides approximately with the northern boundary of exposures of Jurassic and older rocks (Figures 2.1, 2.2 and 2.3). West of the Canning River, the range front trends generally southwestward to its intersection with the east-trending range front of the central Brooks Range at the Sagavanirktok River. East of the border with Canada, the range front trends generally southeastward, bounding a region of extensive exposures of pre-Mississippian rocks to the southwest. These range fronts are probably the surface expression of ramps along thrust faults at depth (Vann and others, 1986). Until the direction of tectonic transport at the time of formation of the different range fronts is determined, it is uncertain whether the various segments of range front mark frontal or oblique ramps.

The southern boundary of the northeastern Brooks Range is here considered to be the northern boundary of closely spaced and distinctly north-vergent folds and imbricate thrust faults in Mississippian and younger rocks (Continental Divide thrust front, Figure 2.3). This structural style is characteristic of the northern part of the main axis of the Brooks Range but is not typical of the northeastern Brooks Range. The boundary is gradational, but appears to lie within a relatively narrow zone near the continental divide of the eastern Brooks Range (Wallace and others, 1988). This boundary can be traced eastward from the range front of the central Brooks Range, and corresponds approximately with the northern boundary of the eastern part of the "disturbed belt", as used by Brosgé and Tailleur (1970, 1971).

We confine our discussion primarily to the central part of the northeastern Brooks Range, between the Canadian border and the area drained by the Canning River, because our observations are restricted to this area. We are uncertain to what extent our interpretations apply to the parts of the northeastern Brooks Range to the east and west, which appear to be regions of transition to other major tectonic provinces.

2.3.2 Age of deformation.

The main axis of the Brooks Range formed during a Middle Jurassic to Early Cretaceous major shortening event, and was modified in mid-Cretaceous to early Tertiary time by uplift and continued shortening of lesser magnitude, accompanied by deposition and deformation in the foredeep to the north (Mayfield and others, 1983; Mull, 1985). The northeastern Brooks Range formed significantly later than the major shortening event in most of the rest of the Brooks Range. Mid-Cretaceous foredeep deposits like those preserved in the western and central foothills are present only as isolated remnants in the northeastern Brooks Range (Figure 2.3; Bathtub Graywacke and Arctic Creek facies; Detterman and others, 1975; Molenaar and others, 1987; Camber and Mull, 1987; Decker and others, 1988). U-Pb and K-Ar isotopic ages indicate a metamorphic and subsequent cooling event at about 61-59 Ma in the southern part of the Okpilak batholith and an adjoining stock, south of the range front in the central portion of the northeastern Brooks Range (Dillon, 1987; Dillon and others, 1987). Apatite fission-track analyses to the south from Bathtub Ridge indicate an uplift age of about 62 Ma, whereas to the north, analyses from west of the Sadlerochit Mountains and east of the Shublik Mountains indicate uplift between about 45 and 32 Ma (O'Sullivan, 1988). Apatite fission-track analyses from the Okpilak batholith indicate uplift between about 42 and 31 Ma (O'Sullivan, 1989). Upper Cretaceous to Lower Tertiary rocks are exposed near the range front of the northeastern Brooks Range, and have been deformed along with older rocks (Reiser and others, 1971; Reiser and others, 1980; Kelley and Foland, 1987).

On the coastal plain to the north, rocks as young as Pliocene display steep dips as a result of folding (Reiser and others, 1971). Seismic reflection data on the coastal plain and offshore show that fold-and-thrust deformation extends to the continental slope north of the northeastern Brooks Range (Grantz and May, 1983; Craig and others, 1985; Bruns and others, 1987; Clough and others, 1987; Kelley and Foland, 1987). The seismic data has been interpreted to show that rocks are more deformed beneath a prominent Eocene unconformity, but the unconformity is itself deformed, indicating that deformation continued later. Surficial geologic evidence of recent uplift onshore, Quaternary structures offshore, and active seismicity indicate that deformation continues to the present in and north of the northeastern Brooks Range (Carter and others, 1986; Grantz and others, 1983). An arcuate deformation front north of the northeastern Brooks Range is defined by the northern limit of young deformation and active seismicity (Moore and others, 1985a). In summary, several lines of evidence indicate that deformation in and north of the northeastern Brooks Range occurred episodically throughout the Cenozoic and has continued to the present. Unfortunately, Cretaceous and younger rocks have been removed by erosion throughout most of the northeastern Brooks Range, and as a result it is difficult to date particular structures except in a relative sense.

2.3.3 Models for the structural evolution of the northeastern Brooks Range.

Early interpretations of the structural evolution of the northeastern Brooks Range focussed on the apparently small amount of shortening and lack of exposed thrust faults, and hence attributed creation of the regional anticlinoria mainly to broad folding and vertical uplift (Reiser, 1970; Mull, 1982). More recent interpretations have focussed on the fold-and-thrust geometry of the region, and called upon northward thrust displacement of Franklinian sequence rocks above an unseen detachment at depth to form the regional anticlinoria (Rattey, 1985; Namson and Wallace, 1986; Kelley and Foland, 1987; Leiggi, 1987).

The manner in which Cenozoic shortening has been accommodated in both the Franklinian sequence and its Ellesmerian cover is currently a subject of debate. This debate centers on two possible end-member models, which are illustrated in Figure 2.5. In model A, a regional imbricate thrust system is inferred to exist mainly within the Franklinian sequence, bounded below by a floor thrust at depth and above by a roof thrust at or near the base of the Ellesmerian sequence. This thrust system constitutes a "duplex", which consists of a system of contractional faults that bound a series of imbricate thrust slices, or "horses", and that include a floor thrust, a roof thrust, and imbricate faults that connect them (Boyer and Elliot, 1982; Mitra, 1986; Mitra and Boyer, 1986). The basis for application of this model to the northeastern Brooks Range is that the geometry of the anticlinoria is consistent with what would be expected as a result of imbrication on faults branching from a detachment at depth. A roof thrust is suggested because faults only

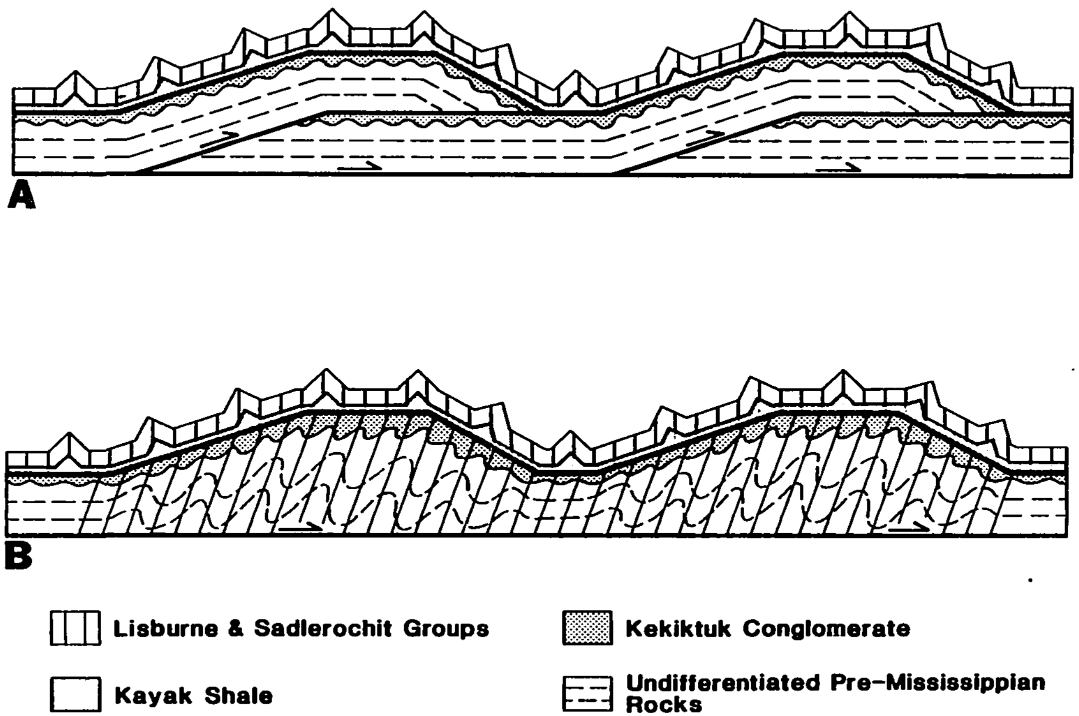


Figure 2.5. End-member conceptual models for the mode of shortening of the pre-Mississippian rocks and the Kekiktuk Conglomerate during Cenozoic thrusting. Dashed lines within pre-Mississippian rocks represent arbitrary markers that were horizontal prior to Cenozoic deformation, and are not necessarily parallel with bedding. Steeply dipping solid lines within pre-Mississippian rocks and Kekiktuk Conglomerate represent cleavage.

locally cut from the Franklinian sequence up-section through the Ellesmerian sequence, and the Ellesmerian cover displays a different structural style than the Franklinian sequence rocks contained in the anticlinoria. In this model, most Cenozoic shortening of the Franklinian sequence is accommodated by thrust duplication. Several variations of this model have been proposed, which differ mainly in the degree of Cenozoic imbrication within the Franklinian sequence and in the role of the Kayak Shale (Figure 2.4), near the base of the Ellesmerian sequence, as the roof thrust of the duplex (Rattey, 1985; Namson and Wallace, 1986; Kelley and Foland, 1987; Leiggi, 1987).

In contrast, model B (Figure 2.5) requires that most of the Cenozoic shortening of the Franklinian sequence be accommodated by mesoscopic and microscopic strain, folds, and faults. Northward transport of the Franklinian sequence en masse above a basal detachment is likely, but not required. In this model, the basal unit of the Ellesmerian sequence, the Kekiktuk Conglomerate (Figure 2.4), must accommodate the same amount of shortening as the underlying Franklinian sequence by penetrative strain and minor folds and faults. Rocks in and above a roof thrust above the Kekiktuk Conglomerate are detached from the dominantly pre-Mississippian rocks below, and so have shortened in a different way, as in model A. Local observations by Avé Lallement and others (1987) and Oldow and others (1986, 1987a) suggest that penetrative structures within both the Franklinian sequence and the Kekiktuk Conglomerate reflect significant Cenozoic internal shortening, thus supporting an important role for model B in the structural evolution of the northeastern Brooks Range.

The actual structures of the northeastern Brooks Range probably are not the product of deformation following either one of these end-member models alone. In the remainder of this paper, we will argue that most shortening in the region has been accommodated according to model A, although the total shortening probably includes a component accommodated according to model B. Quantification of the relative importance of each end-member model in particular areas must await further studies. Furthermore, modifications to the end-member models are required to account for the actual structures in some areas.

2.4 Structural provinces of the northeastern Brooks Range.

The northeastern Brooks Range can be divided into two major structural provinces, a western province and an eastern province (Figure 2.2). These two provinces are separated by a narrow province that contains the Okpilak batholith. The regional structural provinces are defined based primarily on 1) whether anticlinoria contain single or multiple horses of Franklinian sequence rocks and 2) the deformational style of the rocks of the overlying Ellesmerian sequence. Within a given province, these two factors are controlled largely by the composition and rheology of the Franklinian sequence rocks and the stratigraphy of the overlying Ellesmerian sequence.

The size and distribution of anticlinoria are well illustrated on a structure contour map of the unconformity surface that separates the Franklinian and Ellesmerian sequences (Figure 2.6a). This map illustrates the present geometry of the unconformity surface extrapolated both above and below the present ground surface. As seen in this map and in schematic cross sections (Figure 2.7), each anticlinorium in the west is defined by a single structural high, whereas each anticlinorium in the east is a composite of multiple highs. This lateral variation in structural expression is one basis for division of the northeastern Brooks Range into structural provinces.

The stratigraphy of the region may be subdivided into four "structural units" separated by detachment horizons, mainly in shales (Figure 2.4) (Rattey, 1985; Namson and Wallace, 1986; Kelley and Foland, 1987; Leiggi, 1987; Wallace and Hanks, 1988a & b). Each of these structural units displays a distinctive structural style that has been determined largely by its stratigraphic character, with lateral changes in stratigraphy resulting in corresponding changes in structural style. The internal deformation of each structural unit has occurred independently of the other structural units by decoupling between the units along the intervening detachment horizons. Deformation in the different structural units is related mainly by the fact that any shortening of lower structural units must somehow be accommodated in higher structural units, and that any structural relief created in lower structural units must be reflected in overlying structural units.

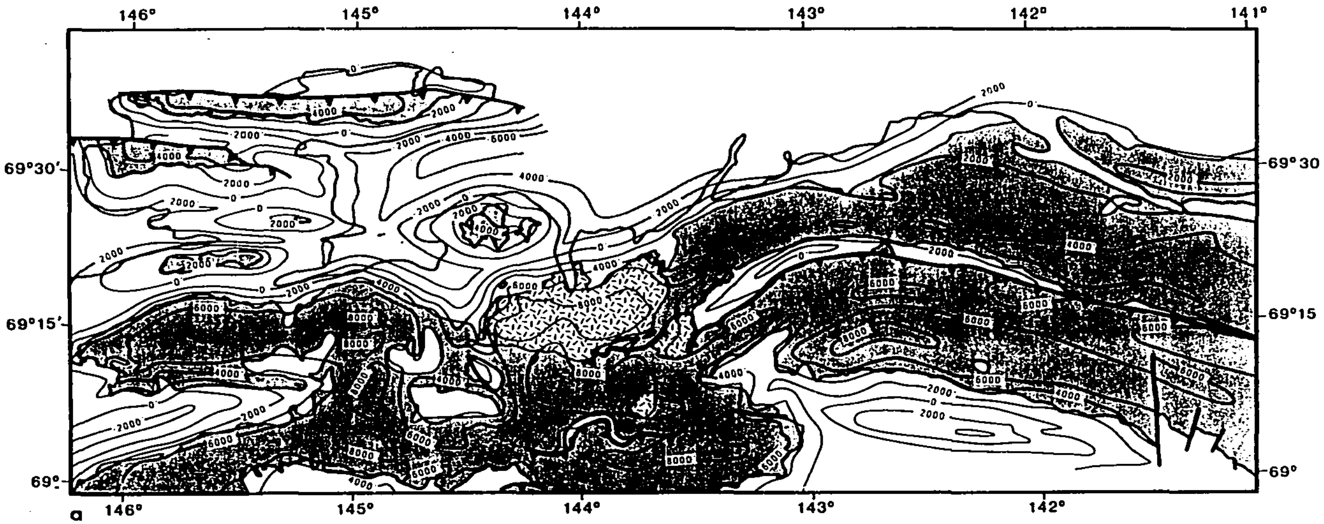


Figure 2.6. (A). Structure contour map on the sub-Mississippian unconformity, with a contour interval of 2000 feet. These contours were derived from the geologic map of Bader and Bird (1986) by extrapolation into the subsurface using published thicknesses and by assuming a minimum elevation of the surface over areas of exposed pre-Mississippian rocks.

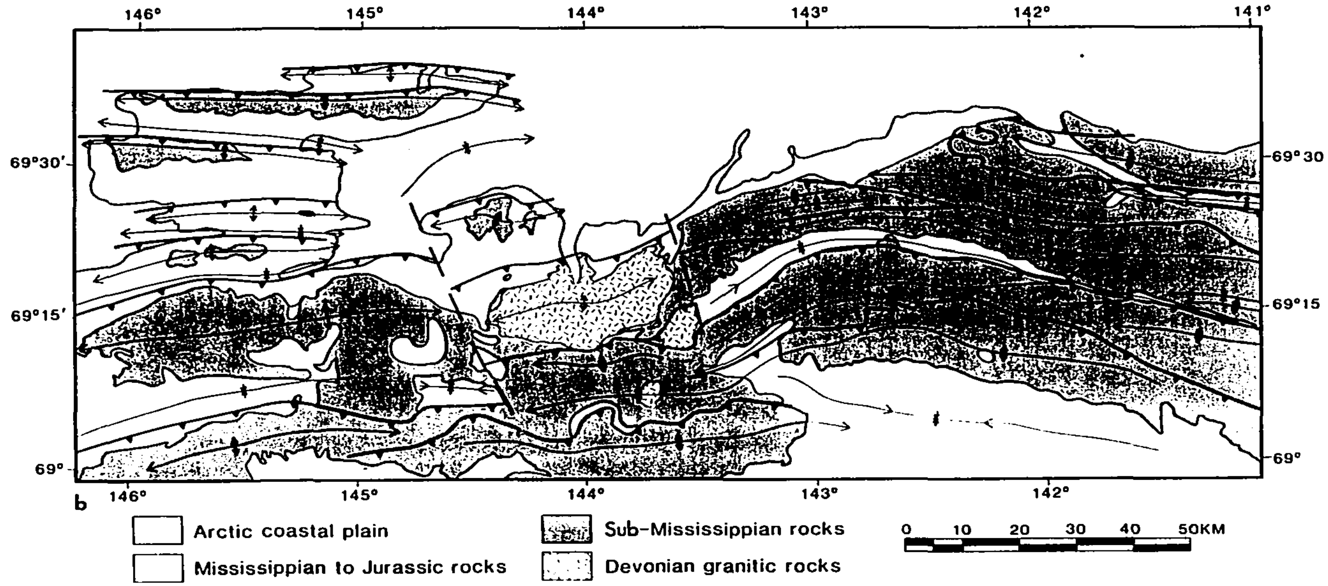


Figure 2.6. (B) Map showing traces of Cenozoic folds and faults that deform the sub-Mississippian unconformity. Crests of anticlines and troughs of synclines were determined from a more detailed version of the structure contour map in (A). The faults are based on mapped faults within the pre-Mississippian rocks, boundaries between major lithologic packages in the pre-Mississippian rocks, and abrupt changes in structural relief on the detailed structure contour map.

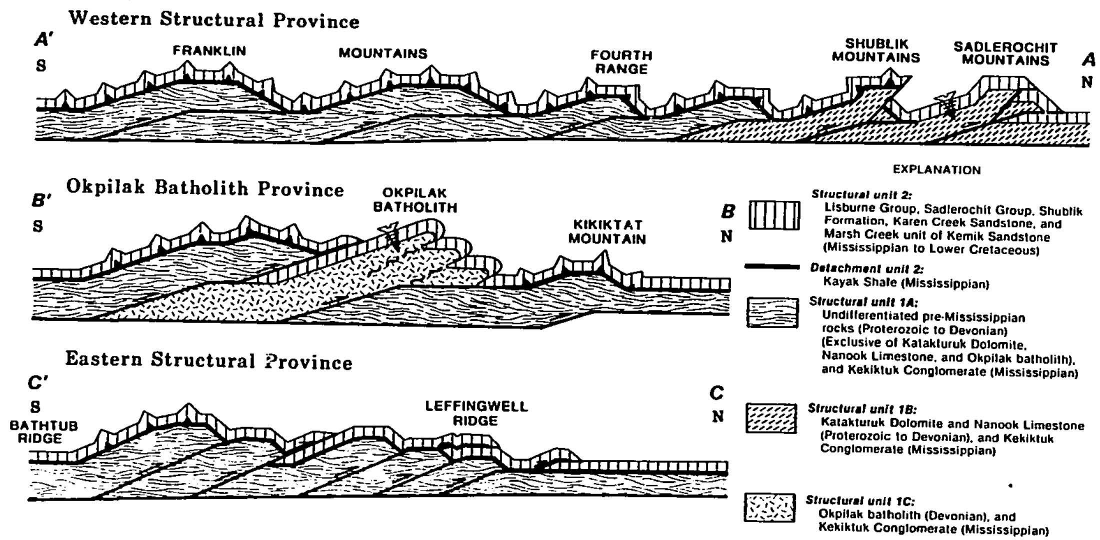


Figure 2.7. Cross sections illustrating schematically the structural geometry in each of the structural provinces of the northeastern Brooks Range. The approximate locations of the sections are shown in Figure 2.2. The sections are not to scale, and are intended to represent specific structures only schematically. Screws indicate places where little or no slip has occurred across horizons that elsewhere are major detachment horizons.

We are concerned mainly with structures developed in the two lowest structural-stratigraphic units (Figure 2.4). The lower unit (structural unit 1) consists of a heterogeneous assemblage of pre-Mississippian rocks and the unconformably overlying Mississippian Kekiktuk Conglomerate, which varies considerably in thickness up to a maximum of about 60 m. The upper and lower units are separated by a detachment in the Mississippian Kayak Shale, a thick (up to 400m), carbonaceous, and highly fissile shale that is present throughout most of the region. Carbonates of the Mississippian to Pennsylvanian Lisburne Limestone and terrigenous clastic rocks of the Permian to Triassic Sadlerochit Group comprise most of the overlying unit (structural unit 2). Higher stratigraphic units have been removed by erosion throughout most of the northeastern Brooks Range.

2.4.1 Western structural province.

The western structural province is characterized by a series of east-trending anticlinoria containing pre-Mississippian rocks in their cores (Figures 2.3 and 2.6). The province can be subdivided into two parts based on the length, width, and amplitude of these anticlinoria. Two broad, west-plunging anticlinoria of equal amplitude occur to the south, in the Franklin Mountains. These anticlinoria are about 90 km long, 25 km wide, and 2.5-3 km in amplitude. At least 4 smaller, doubly plunging anticlinoria occur to the north (the Fourth Range north to the Sadlerochit Mountains), defining the front ranges of the northeastern Brooks Range. The anticlinoria in the front ranges are 35-65 km long, 5-13 km wide, and 1.25-2.5 km in amplitude.

The single structural high defining each anticlinorium in the western structural province is interpreted to mark a single horse in a duplex bounded by a floor thrust within the pre-Mississippian Franklinian sequence, and a roof thrust in the Kayak Shale (Detachment units 1 and 2, respectively) (Figures 2.4 and 2.7). To the north, in the Sadlerochit and Shublik Mountains, a single coherent thrust slice is exposed within each anticlinorium and consists of a stratigraphically intact and structurally competent sequence of massive carbonates. To the south, where the Franklinian sequence is lithologically heterogeneous and displays considerable internal deformation, the geometry of the anticlinoria supports the interpretation that each marks only a single horse (Figure 2.7). Here, a structural form surface marked by the Mississippian Kekiktuk Conglomerate and the underlying sub-Mississippian unconformity defines a simple anticlinorium geometry characterized by planar limbs with a moderately dipping, short forelimb, flat crest, and long, gentle backlimb (Figure 2.8). This geometry suggests that each anticlinorium is a fault-bend fold (Suppe, 1983) resulting from northward displacement of a horse over a footwall ramp (Namson and Wallace, 1986). Although we interpret this to be the dominant mechanism by which shortening was accommodated and the major structures formed, some shortening within each of the horses would be expected in the heterogeneous and commonly structurally weak pre-Mississippian rocks. Oldow and others (1986, 1987a) and Avé Lallemant and others (1987) have suggested, based on local observations, that both the Franklinian sequence and the overlying Kekiktuk Conglomerate were pervasively shortened during Cenozoic deformation by mesoscopic and microscopic strain, folding, and thrust faulting. They further suggest that a significant part of this shortening was accommodated in the Kekiktuk Conglomerate by small-scale thrust duplication above a detachment at the sub-Mississippian unconformity. Our regional observations indicate that significant intra-horse shortening occurred locally, but that on the whole the Kekiktuk Conglomerate displays little internal deformation and only minor and local detachment from or imbrication with the pre-Mississippian rocks. This suggests that Cenozoic shortening internal to each horse is subordinate to that accommodated by thrust emplacement of the horse. Not surprisingly, minor structures involving the Kekiktuk Conglomerate have been observed mainly on the steep north limbs of the anticlinoria (Ziegler, 1988, 1989; Ziegler and Wallace, 1988).

In contrast with the Kekiktuk Conglomerate, the Kayak Shale acted as an extremely effective detachment horizon. Penetrative cleavage and crenulation, and complex tight to isoclinal folds and imbricate faults are common within the unit, and it varies considerably in

Figure 2.8. (A) Photograph looking east-northeast at the south limb of the northern major anticlinorium in the Franklin Mountains. Smooth surface is dip slope marking thin veneer of Mississippian Kekiktuk Conglomerate on sub-Mississippian unconformity surface. This form surface illustrates the simple geometry typical of the anticlinoria.

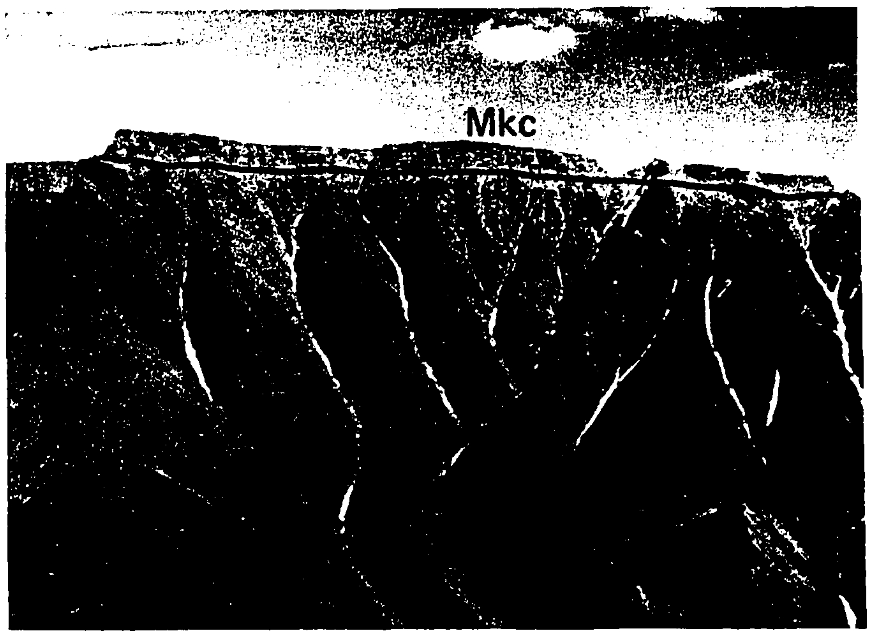


Figure 2.8. (B) Closer view of plateau at crest of anticlinorium, in upper left of (A). A tabular sheet of Kekiktuk Conglomerate (Mkc) overlies the sub-Mississippian unconformity surface. Beds beneath the unconformity dip moderately to the south and display numerous tight chevron folds with south-dipping axial surfaces.

structural thickness. The overlying structurally competent limestones of the Lisburne Group have been deformed into folds that are "second-order", or parasitic, with respect to the "first-order" anticlinoria, and an order of magnitude smaller (about 0.8-2.5 km in wavelength and 0.75-1.25 km in amplitude). Sets of smaller parasitic folds (third-order and higher) are also present. The folds range from box folds to tight chevron-shaped folds (Figure 2.9). They are mostly upright to steeply inclined and display no consistent sense of vergence, though axial surfaces commonly dip toward the center of an underlying anticlinorium. The folds are most consistently upright and display the shortest wavelength and greatest amplitude in the narrow synclinoria between anticlinoria. The Kayak Shale is thickened in the cores of anticlines and thinned in synclines in both first- and second-order folds. We interpret the folds above the Kayak Shale to be detachment folds (Jamison, 1987), formed by slip in and flow of the Kayak Shale, and buckling of the overlying Lisburne Group.

The thick (450-850 m) and relatively competent Lisburne Group acts as the dominant structural unit between the Kayak Shale and overlying detachment horizons in Jurassic or Cretaceous rocks (Figure 2.4). Thus, the geometry of folds in the Sadlerochit Group and Shublik Formation is controlled largely by second- and third-order folds within the underlying Lisburne Group. Local detachment occurs in the Sadlerochit Group along the Kavik Shale, resulting in folding in overlying rocks.

South of the Shublik Mountains (Figure 2.3), only local thrust faults cut up-section from within or below the Kayak Shale into rocks overlying the Kayak Shale. Displacement on these faults is up to 1.5 km (Ziegler, 1989), but generally is less. These are mainly out-of-the-syncline thrusts, associated with tight detachment folds.

Significant changes in structural style occur to the north, in the Sadlerochit and Shublik Mountains (Figures 2.3 and 2.7). In the Sadlerochit Mountains the Kayak Shale is depositionally very thin to absent, and no detachment folds occur within the overlying Lisburne Group and overlying rocks (Figure 2.10; Imm and Watts, 1987; Wallace and others, 1987; Robinson and others, 1989). Instead, these rocks deform with the underlying pre-Mississippian rocks up to the level of the next detachment horizon, either

Figure 2.9. View looking east at detachment folds in the Mississippian-Pennsylvanian Lisburne Group, showing traces of selected axial surfaces within the synclinorium separating the two major anticlinoria in the Franklin Mountains.

Figure 2.10. (A) Photograph looking east at the west-plunging western end of the Sadlerochit Mountains anticlinorium. North limb of the anticlinorium is truncated by the range-front fault. Note absence of detachment folds in strata of the Lisburne (lPMl) and Sadlerochit (TrPs) Groups in the long south limb of the anticlinorium. Asymmetric anticline-syncline pair in strata of the Lisburne Group in the middle of the anticlinorium accommodates minor displacement on a fault in the unconformably underlying Katakturuk Dolomite (pCk).

Figure 2.10. (B) Photograph looking east at sub-Mississippian unconformity in central Sadlerochit Mountains. Strata of the Lisburne Group (lPMl) directly overlie more steeply south-dipping strata of the Nanook Limestone (DpCn); Kekiktuk Conglomerate and Kayak Shale are absent, and Lisburne Group remains structurally coupled to the Nanook Limestone.

the Jurassic to Lower Cretaceous Kingak Shale (where present) or the Lower Cretaceous pebble shale unit (Figure 2.4). Local small-displacement (<100 m) faults cut across the sub-Mississippian unconformity and flatten within the Lisburne Group, generally near the contact between the Alapah and Wahoo Formations (Imm and Watts, 1987; Wallace and others, 1987). Similarly, high-angle reverse faults that define the range-fronts of both the Sadlerochit and Shublik Mountains also cut across the sub-Mississippian unconformity, probably flattening up-section in the Kingak Shale or the pebble shale unit. These faults are late out-of-sequence structures with displacements of up to several kilometers (Leiggi, 1987; McMullen, 1989; Meigs, 1989; Meigs and Wallace, 1987; Rogers, 1989; Rogers and Wallace, 1987; Wallace and others, 1987; Till and Wallace, 1988).

Thus, rocks above and below the Kayak Shale detachment have been shortened by similar amounts (Namson and Wallace, 1986), but by different means: Duplexing below and detachment folding above (Figure 2.7). The Kayak Shale detachment horizon, which forms the roof thrust of the duplex, terminates depositionally to the north in the Sadlerochit Mountains. Thus, the structural units above and below the stratigraphic position of the Kayak Shale detachment horizon are structurally "pinned" in the Sadlerochit Mountains, resulting in their deformation as a single structural unit (Figure 2.7). There are no detachment folds in the Sadlerochit Mountains, nor any evidence of significant thrust duplication of the structural unit immediately overlying the stratigraphic position of the Kayak Shale. Thus the shortening that was accommodated to the south by duplexing below the Kayak Shale must also have been accommodated above the Kayak Shale entirely to the south of the Sadlerochit Mountains. However, the floor thrust of the duplex, the detachment within the pre-Mississippian sequence, continues to the north of the Shublik Mountains (Figure 2.7). Northward slip along this surface has resulted in continued deformation to the north, including formation of the faulted anticlinorium in the Sadlerochit Mountains, but in a duplex with a roof thrust in the pebble shale instead of the Kayak Shale.

A duplex in which shortening above the roof thrust is accommodated directly above the duplex has been referred to as a "passive roof duplex" (Banks and Warburton, 1986) or a "low taper triangle zone" (McMechan, 1985). The duplex in the western province differs from the examples presented by these authors in that faulting plays only a minor role in shortening above the roof thrust. This probably is because the Kayak Shale is such an extraordinarily effective detachment horizon, allowing shortening to be accommodated mainly by detachment folding. This mode of deformation would require significant local variation in the amount and sense of displacement within the Kayak Shale detachment, so that study of minor structures in the Kayak Shale offers a potential means of testing our interpretation.

We interpret the western structural province to conform closely with model A (Figure 2.5), in which the majority of shortening in the pre-Mississippian rocks is accommodated by large-scale thrust duplication in a duplex. Other similar interpretations differ from ours in that they call for multiple imbrications of pre-Mississippian rocks in each anticlinorium (Kelley and Foland, 1987; Leiggi, 1987), place a roof thrust along or below the sub-Mississippian unconformity instead of in the Kayak Shale (Kelley and Foland, 1987; Leiggi, 1987), and place a major thrust boundary between the Franklinian and Ellesmerian sequences in the Sadlerochit Mountains (Kelley and Foland, 1987). We attribute only a minor role to shortening by the penetrative strain and small-scale structures required by model B (Figure 2.5), in contrast with the interpretation of Oldow and others (1986, 1987a) and Avé Lallemant and others (1987).

2.4.2 Eastern structural province.

In contrast with the western structural province, the eastern structural province consists of only two broad (15-30 km wide) and arcuate, northward-convex anticlinoria (Figure 2.2 and 2.3). Structures similar to those exposed in the front ranges of the western province probably continue along strike north of the eastern province, but their structural relief is considerably less and hence they are confined to the subsurface of the coastal plain (Bruns and others, 1987; Clough and others, 1987; Kelley and Foland, 1987). As in the western province, the exposed broad anticlinoria of the eastern province contain pre-

Mississippian rocks of the Franklinian sequence in their cores, with the flanks of the folds defined by Mississippian through Triassic rocks of the Ellesmerian sequence. In most places, the Mississippian and younger rocks have been eroded from the crests of the anticlinoria.

In contrast with the western structural province, each anticlinorium in the eastern province consists of several subordinate structural highs, probably marking separate thrust slices of pre-Mississippian rocks (Figure 2.6). These thrust slices correspond with distinct, thrust-bounded lithologic packages within the pre-Mississippian rocks (Reiser and others, 1980; Hanks, 1987, 1988, 1989), the boundaries of which are marked locally by imbrications of the Kekiktuk Conglomerate within the pre-Mississippian rocks.

In contrast with the structural style characteristic of the western structural province, several major thrust faults have cut up-section from or across the Kayak Shale (Figures 2.3 and 2.7) and there is no evidence in the northern part of the eastern structural province of major detachment folding above the Kayak Shale. Pre-Mississippian rocks were thrust northward over Mississippian through Triassic rocks at the boundary between the southern and northern anticlinoria (Figures 2.3 and 2.7) (Reiser and others, 1980). On the northern limb of the northern anticlinorium, there was major thrust duplication of much of the Ellesmerian sequence, as indicated by a large klippe of Kayak Shale through Triassic Karen Creek Sandstone resting structurally upon the Kingak Shale (Figure 2.11) (Hanks and Wallace, 1987; Hanks, 1987, 1988). This indicates that both the Kayak and Kingak Shales acted as major detachment surfaces in the eastern structural province, but with shortening between the detachment horizons being accommodated by thrust duplication rather than detachment folding. The Kayak Shale is present throughout the eastern structural province. However, in the northern part of the province, the Kayak Shale is thinner and siltier than elsewhere, and a major carbonate interval, which is discontinuous and of variable thickness, occurs in its upper part (Hanks, 1987, 1988). The presence of this competent carbonate interval may have facilitated the formation of ramps, resulting in shortening of the rocks above the Kayak Shale detachment by thrust duplication rather than detachment folding. In the southern part of the eastern province, the Kayak Shale consists of a thick

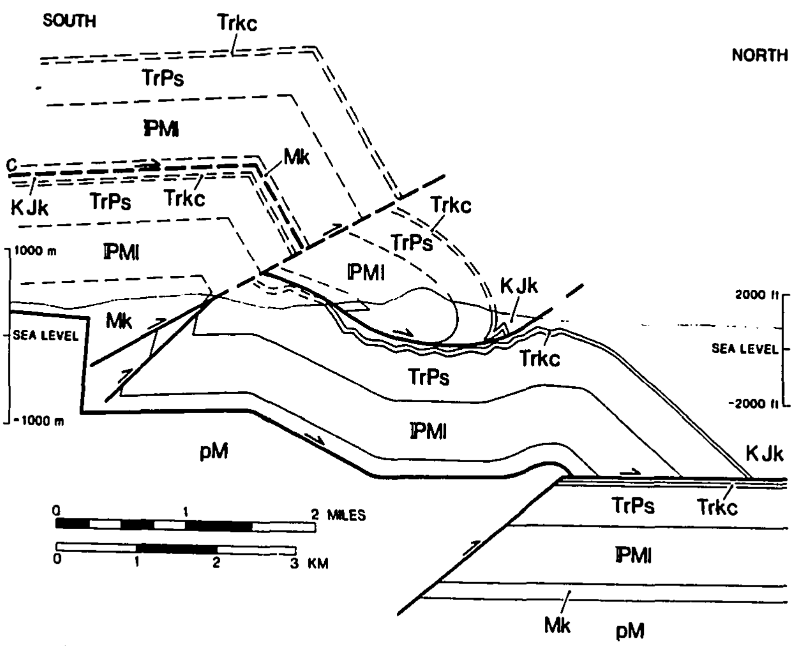


Figure 2.11. Structure section across Leffingwell Ridge. Illustrates the thrust duplication of Mississippian through Triassic strata typical of the eastern structural province. pM = undifferentiated pre-Mississippian rocks, Mk = Kekiktuk Conglomerate and Kayak Shale, lPMl = Lisburne Limestone, TrPs = Sadlerochit Group, Trk = Karen Creek Sandstone, KJk = Kingak Shale.

shale sequence with only minor carbonates. Here, shortening above the Kayak Shale is accommodated by a combination of thrust faults and detachment folds.

In summary, we suggest that each anticlinorium of the eastern province is an anticlinal stack consisting of multiple horses in a duplex (Figure 2.7). As in the western province, the Kayak Shale has acted as the roof thrust. In the eastern province, however, shortening above the roof thrust has been by thrust duplication accompanied by relatively minor detachment folding, probably due to changes in the character of the Kayak Shale. This interpretation is a modified version of model A (Figure 2.5), in which each anticlinorium reflects emplacement of multiple horses, but shortening is still accommodated mainly by thrust duplication and not penetrative strain within the pre-Mississippian rocks.

2.4.3 Okpilak batholith structural province.

A narrow province separating the eastern and western provinces corresponds with the location of the Okpilak batholith (Figures 2.2, 2.3, and 2.6). The batholith is of Devonian age, as indicatedby U-Pb zircon isotopic dating (380 ±10 Ma, Dillon and others, 1987). It is overlain depositionally by either the Kekiktuk Conglomerate or Lisburne Group, and the Kayak Shale is thin to absent (Sable, 1977; Dillon and others, 1987; Watts and others, 1988). The batholith is a major topographic and structural high, and the overlying Mississippian and younger rocks dip away from the batholith on its east, west, and north flanks (Figure 2.6a) (Reiser and others, 1971, 1980; Sable, 1977; Bader and Bird, 1986). Structures along strike with and immediately to the south of the batholith are deflected northward around the east and west flanks of the batholith, suggesting that the batholith may have acted as a structural "buttress" during or after formation of those structures (Figures 2.3 and 2.6b). However, the northern limb of the anticlinorium that defines the range front of the eastern structural province extends undeflected to the north of the batholith, and connects westward with the northern limb of a regional anticlinorium of the western structural province. To both the east and west, this structure is interpreted to reflect displacement over a ramp from a detachment within the pre-Mississippian sequence up to the Kayak Shale, suggesting that the change in structural relief immediately north of the Okpilak batholith may have a similar origin (Figure 2.7).

Pre-Mississippian rocks are involved in various fold-and-thrust structures to the north of the batholith, including an anticlinorium at Kikiktak Mountain and structures at considerable depth (over 7 km) beneath the coastal plain (Figure 2.3; Bruns and others, 1987; Clough and others, 1987; Kelley and Foland, 1987). The involvement of pre-Mississippian rocks in these structures indicates that deformation occurred above a detachment at depth within the pre-Mississippian sequence. Existence of such a detachment to the north of the batholith requires that the batholith itself be underlain by a detachment at depth, above which it was displaced to the north (Figure 2.7). Fault displacement of the batholith is further supported by the fact that granitic rocks along its north flank display strong mylonitic fabrics and in places are imbricated and structurally overlie Ellesmerian sequence rocks (Figure 2.12; Sable, 1977; Hanks and Wallace, 1989 and 1990). Mylonitic fabrics and local shear zones also occur across the width of the batholith (Sable, 1977; Hanks and Wallace, 1990), indicating some internal deformation. The topographic and structural elevation of the batholith (Figure 2.6a) implies that the thrust sheet containing the batholith is thicker than the duplexes involving pre-Mississippian rocks both to the east and west, suggesting that the basal detachment beneath the batholith formed at greater depth than equivalent detachments to the east and west. Some structural thickening may also have occurred within or beneath the batholith during or after its structural emplacement by pervasive internal strain and/or imbrication.

Where they are exposed along the northern flank of the batholith, Ellesmerian sequence rocks resting nonconformably on the batholith have deformed along with the batholith. Structures in the Ellesmerian sequence rocks do not appear to reflect any significant detachment at or near the unconformity, probably because the Kayak Shale is thin to absent over the batholith. Instead, shear zones cut from the granite across the unconformity and into the cover rocks, where slip is accommodated in folds and penetrative fabrics (Hanks and Wallace, 1989 and 1990). Penetrative fabrics in the cover rocks may also reflect accommodation of shortening by penetrative strain in the underlying

Figure 2.12. Photograph looking down-plunge to the east along the northern margin of the Okpilak batholith. Granitic rocks (Dg) are duplicated on a south-dipping shear zone in the foreground. To the east, the same shear zone places rocks of the Okpilak batholith over rocks of the Lisburne (lPMl) and Sadlerochit (TrPs) Groups in the overturned south limb of a syncline. To the north, rocks of the batholith and the Lisburne Group have been thrust over rocks of the Sadlerochit Group. Leffingwell Ridge, which marks the range front in the eastern province, is in the background to the east.

batholith.

The structural high including the Okpilak batholith and an area to its south is inferred to be bounded to the east and west by mostly unexposed north-northwest trending transverse faults (Figures 2.3 and 2.6) (Sable, 1977; Reiser and others, 1971, 1980; Bader and Bird, 1986). These faults could reflect local differences in the way shortening was accommodated in the vicinity of the Okpilak batholith as compared with the adjacent portions of the western and eastern structural provinces. In turn, these differences in accommodation may reflect lateral ramps marking a change in the depth to the basal detachment to both the east and west.

In summary, our interpretation of the Cenozoic deformation of the Okpilak batholith, in which the pre-Mississippian rocks are relatively homogeneous and isotropic, consists of a combination of models A and B (Figure 2.5). The batholith probably has been transported northward above a detachment at depth, but some shortening has also been accommodated by pervasive strain and imbrication within the batholith itself. Since the Kayak Shale is absent in the vicinity of the batholith, the Ellesmerian sequence remained structurally attached to the batholith during Cenozoic deformation. Any internal shortening and structural thickening that occurred within the batholith during thrusting would have to be accommodated within the structurally attached sedimentary cover. This coupling would account for at least some of the increase in penetrative fabrics in Ellesmerian sequence rocks near the batholith, and for the shear zones that cut from the granite into the cover rocks.

2.5 Structural trends and the directions of tectonic transport.

In the western structural province, folds and faults occur on two distinct structural trends. Structures which involve pre-Mississippian rocks, including the regional anticlinoria themselves, are consistently east-trending (Figures 2.3 and 2.6). However, east-northeast trends are displayed by structures that formed entirely above the Kayak Shale or higher detachment horizons, and thus do not involve pre-Mississippian rocks. East-trending structures involving pre-Mississippian rocks locally have overprinted east-northeast trending structures in Mississippian and younger rocks, a relationship that is particularly apparent in the eastern Sadlerochit Mountains area (Fig. 2.13).

It is tempting to assume that these two trends formed during temporally distinct deformational events, with the changes in trend reflecting different directions of tectonic transport (Leiggi, 1987). However, an alternative interpretation is suggested by two observations. First, pre-Mississippian rocks are consistently involved in the east-trending structures. Second, east-northeast trends are displayed by some of the youngest structures in the fold-and-thrust belt, near the northern front of deformation in the northern part of the coastal plain and offshore (Figure 2.3), where pre-Mississippian rocks apparently are not involved (Craig and others, 1985; Bruns and others, 1987; Clough and others, 1987; Kelley and Foland, 1987). Bedding and structures in the pre-Mississippian rocks trend eastward with respect to the sub-Mississippian unconformity, thus demonstrating that an east-trending structural grain was established in pre-Mississippian time. This relationship is particularly well displayed in the Sadlerochit and Shublik Mountains (Robinson and others, 1989). Reactivation or preferential formation of structures along this pre-Mississippian grain during Cenozoic deformation would result in east-trending structures, not necessarily reflecting the true direction of tectonic transport.

Superposition of the pre-Mississippian-controlled east-trend on structures which formed earlier, but at a higher structural level, would be a natural consequence of the normal evolution of a fold-and-thrust belt. As deformation migrates toward the foreland of a fold-and-thrust belt, detachment occurs at progressively deeper structural and stratigraphic levels at any given location. In the northeastern Brooks Range, this sequence would lead to formation of structures above the Kayak Shale detachment that would reflect the true direction of tectonic transport, but that later would be overprinted by east-trending structures when the basal detachment dropped into the pre-Mississippian sequence. The absolute age of overprinting would vary from place to place depending on when the detachment surface dropped into the pre-Mississippian sequence at any particular location.

East-northeast trends are the oldest-formed structures where multiple trends are

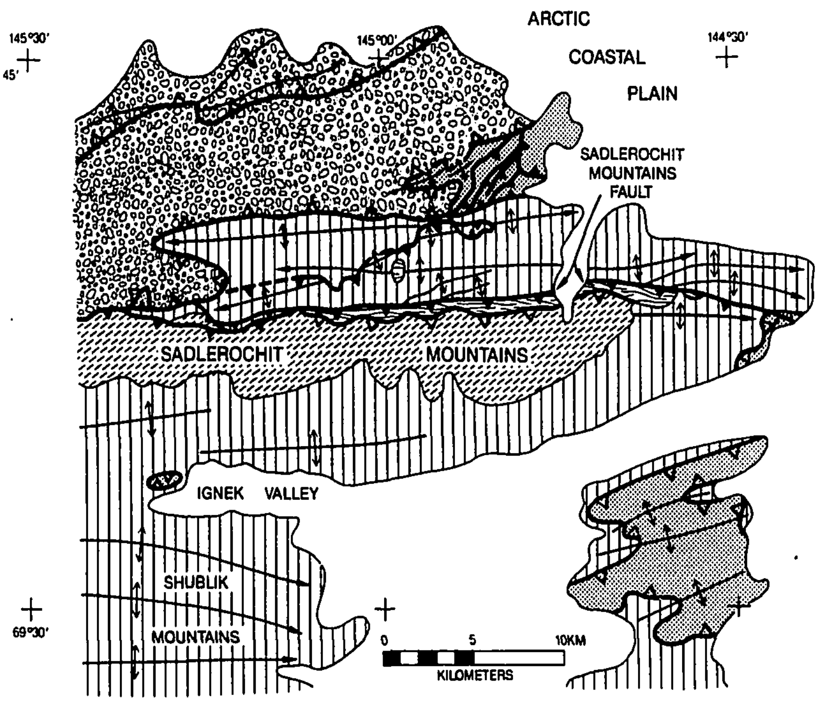


Figure 2.13. Simplified geologic map of the eastern Sadlerochit Mountains area, modified from Bader and Bird (1986), Meigs (1989), Kelley and Foland (1987), and Robinson and others (1989). Unit patterns as in Figure 2.3. East-northeast trending structures are most prominent in the Cretaceous and younger rocks to the lower right (Arctic Creek facies) and to the north. These early-formed structures were preserved only in structural lows following formation of east-trending anticlinoria in the Shublik Mountains, Sadlerochit Mountains, and northeastern Sadlerochit Mountains (north of the Sadlerochit Mountains fault). Pre-Mississippian rocks core each of these anticlinoria. Mississippian through Lower Cretaceous rocks north of the Sadlerochit Mountains fault display both east-northeast and east-trending structures. The east-trending Sadlerochit Mountains fault truncates some of the east-northeast trending structures, and formation of the anticlinorium north of the fault led to folding of earlier east-northeast trending structures north of the anticlinorium.

observed, yet the most recently formed structures near the leading edge of the fold-and-thrust belt also trend east-northeast (Figures 2.3 and 2.13). Consequently, we do not interpret the east-trends to represent a temporally and directionally distinct deformation event, but rather to reflect control by the pre-existing structural grain in the pre-Mississippian rocks. We infer that the dominant direction of tectonic transport during deformation of the northeastern Brooks Range was to the north-northwest, normal to the east-northeast trending structures. Although deformation probably occurred episodically throughout Cenozoic time up to the present, this interpretation requires no significant change in tectonic transport direction from episode to episode.

In the eastern structural province, the pattern of structural trends is more complex (Figures 2.3 and 2.6), and the factors controlling the trends remain uncertain. The pre-Mississippian rocks display generally eastward structural and lithologic trends. Multiple east-trending horses within each anticlinorium appear to be marked by structural highs, fault-bounded lithologic packages, and faults involving rocks of the Endicott Group (Figure 2.6) (Reiser and others, 1980).

Although structures involving pre-Mississippian rocks are generally east-trending within the anticlinoria, the trends of the anticlinoria themselves, and of the Mississippian and younger rocks on their flanks, are more complex. The regional anticlinoria of the eastern province are strongly arcuate, with their western portions trending east-northeast and their eastern portions trending west-northwest (Figure 2.3). The trends of the anticlinoria are defined by a composite of multiple horses, rather than being defined by a single horse. This is well illustrated in the northern anticlinorium, where the east-northeast trending range front at Leffingwell Ridge appears to flank a series of east-trending, left-stepping, en echelon structural highs corresponding with horses in the core of the anticlinorium (Figure 2.6). The cause of the overall arcuate trend of the anticlinoria is unknown. It may be due to some combination of 1) buttressing by the Okpilak batholith, 2) lateral variations in the direction of tectonic transport (a radial pattern?), 3) lateral variations in the amount of shortening (decreasing to the east?), and/or 4) multiple deformation events with different transport directions. If a north-northwest direction of tectonic transport is assumed, structures in the western, east-northeast trending portion of the arc would be related to frontal ramps, whereas the west-northwest trends in the east would be related to oblique ramps.

2.6 Implications for the subsurface of the ANWR coastal plain.

Franklinian and Ellesmerian sequence rocks comparable to those exposed in the northeastern Brooks Range extend northward beneath the Arctic coastal plain of the Arctic National Wildlife Refuge (ANWR). Consequently, structures similar to those exposed in the northeastern Brooks Range probably also are present in the subsurface of the coastal plain, although their depth of burial indicates a northward decrease in shortening and resulting structural relief. The structural provinces defined in the northeastern Brooks Range are likely to extend northward beneath the coastal plain, as suggested by lateral changes in structural geometry and relief determined from seismic data (Bruns and others, 1987; Clough and others, 1987; Kelley and Foland, 1987).

Pre-Mississippian rocks probably form the cores of broad subsurface anticlinoria defined by single or multiple horses. Significant stratigraphic differences are likely in the pre-Mississippian rocks of different horses, reflecting both pre-Mississippian and Cenozoic structural displacements, and the tendency for bounding structures to form along major lithologic boundaries. Individual horses in the anticlinoria most likely trend eastward, reflecting control by a pre-Mississippian structural grain. However, where an anticlinorium is defined by multiple horses, it may trend east-northeast, normal to the Cenozoic transport direction, but be a composite of smaller, east-trending, left-stepping en echelon highs.

The choice of a model for formation of the regional anticlinoria (model A or B, Figure 2.5) has important implications for prediction of the distribution and quality of reservoir in the Kekiktuk Conglomerate (where it is not missing beneath younger unconformities). If most of the Cenozoic shortening within the pre-Mississippian rocks and overlying Kekiktuk Conglomerate has been accommodated by thrust faulting, as in model A, the Kekiktuk Conglomerate would lie in the footwall of each horse and reservoir potential could still be preserved. However, if shortening has been accommodated mainly in mesoscopic and microscopic structures, as in model B, porosity within the Kekiktuk Conglomerate probably has been destroyed.

Detachment folds or major thrust duplications may be expected to occur above the Kayak Shale. The occurrence of thrust duplications may be favored by ramping due to the local presence of thick carbonate intervals within the Kayak Shale, or the thinning and coarsening of shale within the unit. Major thrust duplication of Ellesmerian sequence rocks above the Kayak Shale, as seen in the eastern structural province, has important implications for trap geometry and the vertical distribution of reservoir intervals. Where the Kayak Shale is depositionally absent over pre-Mississippian topographic highs, pre-Mississippian and overlying rocks may be expected to deform together as a single structural unit and faults will tend to cut up-section to a higher detachment horizon. Reverse faults with moderate to steep dips and that cut across the sub-Mississippian unconformity may occur where the pre-Mississippian rocks are particularly competent and/or the Kayak Shale is missing.

Subsurface observations west of ANWR and surface observations in the Sadlerochit Mountains suggest that progressively more section has been eroded beneath the Lower Cretaceous unconformity in a generally north-northeastward direction, ultimately resulting in complete truncation of the Ellesmerian sequence (Bird and others, 1987). Consequently, structurally and economically significant rocks of the Ellesmerian sequence may be absent to the north, particularly north of the western province. However, the presence of Jurassic rocks in and adjacent to the coastal plain in eastern ANWR (Reiser and others, 1980) indicates that a significant change must occur to the east in the relatively simple geometry inferred for the unconformity. The presence of the Kayak Shale and the preservation of a significant thickness of Ellesmerian sequence rocks above it would be indicated in the subsurface if distinctive structures of the types that are seen above the Kayak Shale in surface exposures were observed in seismic reflection data.

Cretaceous and Tertiary foredeep deposits of the Brookian sequence are structurally separated by a major detachment in the pebble shale unit from underlying rocks of the Franklinian and Ellesmerian sequences (Figure 2.4). Folds and faults have formed within the Brookian foredeep sequence above this detachment, and structures have been superimposed upon them by deformation of underlying rocks. In particular, major highs and lows are likely to be controlled mainly by the geometry of the anticlinoria formed in pre-Mississippian rocks, and structures within the foredeep sequence will have formed to accommodate displacement on major faults cutting up-section from the Franklinian and Ellesmerian sequences. Structures are likely to be progressively younger with increasing structural depth.

The amount of Cenozoic structural shortening will be difficult to determine using seismic interpretation of the stratigraphically and structurally complex foredeep deposits. However, it may be easier to estimate the amount of shortening that has occurred within the underlying Ellesmerian and Franklinian sequence rocks. Applying this approach to a combination of seismic reflection data from the coastal plain and surface data from the northeastern Brooks Range may provide a basis for an estimate of the shortening that has been accommodated by the foredeep deposits in the subsurface of the ANWR coastal plain.

2.7 Conclusions.

The northeastern Brooks Range may be divided into two major structural provinces separated by a narrow third province. These provinces are defined based on the structural characteristics of both pre-Mississippian rocks and their Mississippian and younger cover. Anticlinoria containing pre-Mississippian rocks are the dominant structural element in the northeastern Brooks Range, and their geometry is controlled by a north-vergent regional duplex with a floor thrust in the pre-Mississippian sequence, and a roof thrust generally in the Kayak Shale.

In the western structural province, each anticlinorium contains a single horse, and shortening above the Kayak Shale is accommodated mainly by detachment folds. In the northern part of the western structural province, the Kayak Shale is depositionally absent in the Sadlerochit Mountains. In this area, rocks normally separated by the Kayak Shale detachment deform together, there are no detachment folds, and faults originating in the pre-Mississippian sequence cut up-section to higher stratigraphic levels.

In the eastern structural province, each anticlinorium contains multiple horses, and shortening above the Kayak Shale is accommodated mainly by thrust duplication of Mississippian through Triassic rocks. The presence of thick carbonate intervals within the Kayak Shale may favor the ramping of thrust faults to higher detachment horizons.

In the Okpilak batholith structural province, the batholith has been detached from its roots and transported northward. The Kayak Shale is thin to absent in the vicinity of the batholith, and the overlying Ellesmerian sequence rocks have remained structurally coupled to the batholith during thrusting. Shortening within the batholith prior to or during thrusting has resulted in imbrication and formation of penetrative fabrics within both the batholith and the overlying Ellesmerian cover sequence. The Okpilak batholith province is inferred to be bounded to the east and west by transverse faults. These faults may have developed due to lateral differences in the way shortening has been accommodated, and may reflect lateral ramps marking changes in the depth to the basal detachment from the eastern and western structural provinces.

East-northeast trends have formed where pre-Mississippian rocks were not involved in deformation, and probably are normal to the direction of tectonic transport. East trends have formed where pre-Mississippian rocks were involved in deformation, probably controlled mainly by pre-Mississippian structural trends. At any given location, east-trends generally post-date east-northeast trends, reflecting a drop of the basal detachment into pre-Mississippian rocks over time.

Variations in structures similar to those observed in the northeastern Brooks Range are likely where rocks of the Franklinian and Ellesmerian sequences have been involved in deformation in the subsurface of the coastal plain to the north.

2.8 Acknowledgments.

The results presented here are part of a continuing program of geological research in the northeastern Brooks Range by the University of Alaska Fairbanks (UAF). This work was supported by grants from Amoco, ARCO, BP Exploration (Alaska), Chevron, Conoco, Elf Aquitaine, Exxon, Japan Oil, Marathon, Mobil, Murphy, Phillips, Shell, and Texaco. Additional support was provided by the University of Alaska Fairbanks, to Wallace by the Petroleum Research Fund (administered by the American Chemical Society), and to Hanks by the Geological Society of America and Sigma Xi. Our work would not have been possible without the cooperation of the Alaska Division of Geological and Geophysical Surveys (ADGGS) and the U.S. Fish and Wildlife Service.

Our work has greatly benefitted from detailed studies by and conversations with UAF graduate students and faculty, including A. Anderson, W. Camber, R.K. Crowder, T. Imm, D. Knock, B. (Till) McMullen, A. Meigs, P. O'Sullivan, E. Pavia, J. Rogers, K. Watts, and J. Ziegler, and members of ADGGS, including J. Decker, J. Dillon (deceased), C.G. Mull, and M. Robinson. We appreciate the comments of two anonymous reviewers, and particularly the careful review by John S. Oldow. W.K. Wallace would like particularly to thank M. Mangus for providing an excellent introduction to the geology of the northeastern Brooks Range and Jay Namson for introducing him to some new ways of looking at fold-and-thrust structures.

CHAPTER 3: THE AICHILIK RIVER TRANSECT[2]

3.1 Abstract

In the northeastern Arctic National Wildlife Refuge (ANWR) of the northeastern Brooks Range, Cenozoic-age anticlinoria expose non-crystalline 'basement' rocks immediately adjacent to much younger foredeep deposits. The anticlinoria probably reflect multiple horses in a regional north-vergent duplex with a floor thrust at depth in pre-Mississippian 'basement' rocks and a roof thrust in a Mississippian shale at the base of the younger stratified cover sequence. Mesoscopic and map-scale structural analysis and modeling of regional shortening suggest that the majority of Cenozoic shortening in the basement rocks was accommodated by thrust duplication in the duplex, not by the development of penetrative mesoscopic structures. East-northeast Cenozoic structural trends in the Mississippian and younger cover sequence probably reflect a north-northwest transport direction during thrusting. Regional east-west Cenozoic structural trends within the pre-Mississippian rocks may partially reflect an inherited pre-Mississippian structural grain and/or pre-Mississippian structures reactivated during Cenozoic deformation.

General area-balanced models of northeastern ANWR suggest that the amount of Cenozoic shortening is controlled primarily by the depth to the basal detachment surface in the foreland, the structural topography of the region, the geometry of the basal detachment surface and the depth to the brittle/ductile transition, where the basal detachment surface might be expected to flatten. All of these variables cannot be constrained in northeastern ANWR at the present time due to insufficient data. Given a reasonable set of assumptions regarding the geometry and behavior of the basal detachment surface, shortening in northeastern ANWR could range from 16% to 61%. Currently available regional and detailed structural data suggest that an intermediate figure, 46%, is probably the most

[2] Chapter 3 contains the complete text and figures of the manuscript, *Thin-skinned thrusting in non-crystalline basement rocks: an example from the northeastern Brooks Range, Alaska,* to be submitted to the Geological Society of America *Bulletin.*

accurate.

3.2 Introduction

Many geologists use the term 'basement' very loosely, depending upon their area of interest. In most orogenic belts, 'basement' refers to metamorphosed and frequently polydeformed crystalline rocks that underlie a younger stratified cover sequence. These older 'basement' rocks are generally exposed in the core of the orogenic belt where deformation has involved ductile deformation at deep crustal levels (Hatcher and Williams, 1986). This mode of deformation contrasts with that of the cover sequence, which is usually involved in fold-and-thrust style deformation adjacent to the foreland.

However, the term 'basement' also is often used to mean 'depositional basement' or 'economic basement.' This usage is especially common in the petroleum industry where age, temperature, or lithology may limit the depths at which hydrocarbons can be reasonably expected or produced. In these cases, and in exposed mountain belts, 'basement' therefore actually refers to the oldest rocks exposed in an orogenic belt that are acting as 'depositional basement' to the overlying rocks of interest. These older rocks may not even be igneous or high-grade metamorphic rocks, but instead stratified sedimentary rocks separated from the overlying cover sequence by a major regional unconformity. The older sequence may have been deposited in a different tectonic setting and have a more complex structural history when compared to that of the overlying younger cover sequence.

In most foreland fold-and-thrust belts, basement and depositional basement are synonymous, consist of old crystalline rocks, and commonly are not involved in thrusting immediately adjacent to the foredeep (e.g., the Canadian Rockies (Bally and others, 1966; Price, 1981) and the southern and central Appalachians (Hatcher, 1981)). In the northeastern Brooks Range, however, the term 'basement' has been commonly used to describe previously deformed, low-grade, sedimentary and volcanic rocks that underlie the younger sedimentary cover sequence and foredeep deposits (e.g., Mull, 1982; Leiggi, 1987; Bird and Molenaar, 1987; Hubbard and others, 1987). Although these rocks can be considered 'depositional basement' to the overlying sediments, the usage of the unmodified term 'basement' can lead to erroneous assumptions by workers not familiar with the region as to the pressures and temperatures these rocks have experienced as well as the style of deformation they have undergone.

In the Arctic National Wildlife Refuge (ANWR) of the northeastern Brooks Range (Figure 3.1), the low-grade polydeformed rocks of the 'depositional basement' also have been intimately involved in the young and actively growing fold-and-thrust belt that has formed the present-day range. These low-grade rocks occupy an anomalous structural position in that they are structurally elevated with respect to immediately adjacent, and considerably younger, foredeep deposits. This is in sharp contrast with the structural behavior of similar age rocks in the main axis of the Brooks Range, where they are not structurally involved in the deformation immediately adjacent to the foreland. The degree and manner in which these older rocks are involved in the foreland fold-and-thrust deformation of the northeastern Brooks Range have been the topic of much debate, with mechanisms ranging from thrust duplication in a regional duplex (e.g. Rattey, 1985; Wallace and Hanks, 1990) to shortening of the older rocks primarily via the development of penetrative structures (e.g., Vann and others, 1986; Oldow and others, 1987a).

The involvement of depositional basement in the range-front region of the northeastern Brooks Range implies either a correspondingly deep orogenic sole thrust immediately adjacent to the foredeep or a shallow depth to depositional basement in the foreland region, or a combination of the two. A structural study of the northeastern Brooks Range may not necessarily indicate what the total stratigraphic thickness of the cover sequence may have been at the time of deformation, but it may shed some light on the geometry of the orogenic wedge, the depth and geometry of its basal detachment horizon, and the conditions under which it developed. This in turn could provide new insights into the nature and causes of the variability in the geometry of orogenic sole faults both within and between actively growing fold-and-thrust belts.

This paper is divided into two sections. The first describes the structural style of both the depositional basement and its sedimentary cover in a northeastern portion of the

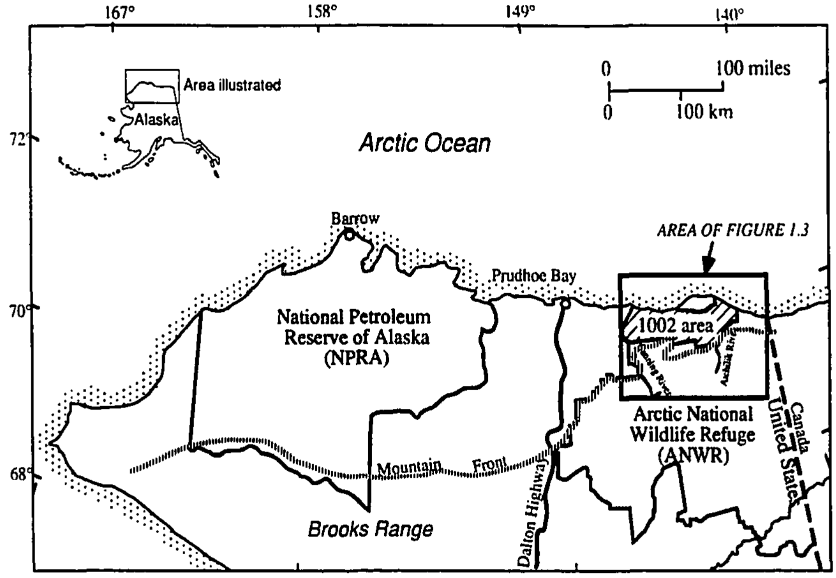


Figure 3.1. Generalized map of northern Alaska showing location of the Arctic National Wildlife Refuge (ANWR) with respect to other geographic and geopolitical features.

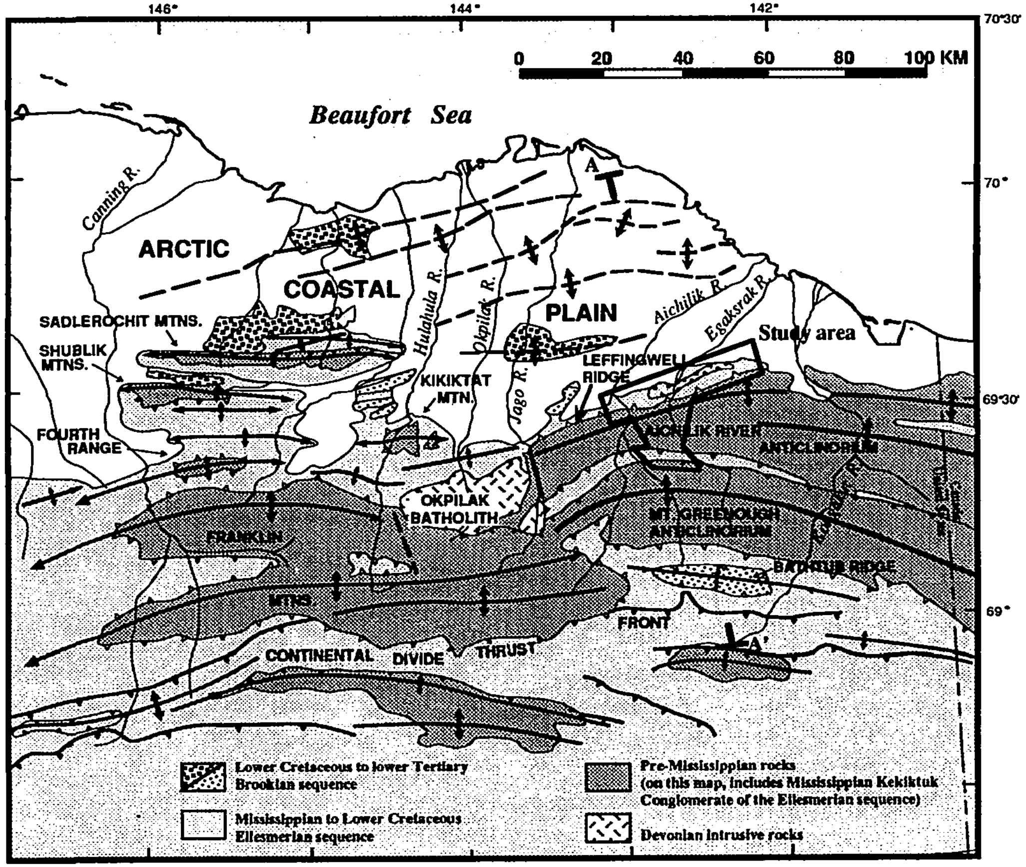
northeastern Brooks Range. I provide evidence that, during Cenozoic thrusting, the older rocks deformed primarily by thrusting and related folding. Penetrative structures played a minimal role in accommodating Cenozoic shortening in these older rocks. In the second part of the paper, I use a series of simple area-balanced models to explore some of the different possible geometries of the orogenic sole thrust, and examine the effect of these different geometries on the amount of regional shortening in the area. These models are then refined into a balanced cross section which serves to summarize my preferred version of the structural geometry and evolution of the region.

3.3 Regional Setting

The northeastern Brooks Range is a prominent topographic and structural salient with respect to the main east-west axis of the Brooks Range (Figures 3.1 and 3.2). The main axis of the Brooks Range formed during Late Jurassic to Early Cretaceous time by the collapse of a wide, and undoubtedly complicated, late Paleozoic and early Mesozoic south-facing continental margin (Mull, 1982; Mayfield and others, 1983). The collapse of this margin resulted in the emplacement of a series of large, internally deformed thrust sheets in the main axis of the Brooks Range, with each sheet consisting of age-equivalent strata deposited on different parts of the continental margin (Mull, 1982; Mayfield and others, 1983). The autochthonous rocks upon which these sheets were emplaced are preserved in the subsurface of the North Slope, and exposed in the northeastern Brooks Range (Reiser, 1970). The main axis of the Brooks Range has undergone subsequent Late Cretaceous and Tertiary uplift which is probably related to deformation in the foredeep to the north (Oldow and others, 1987b; Blythe and others, 1990; O'Sullivan and others, 1990).

Unlike the main axis of the Brooks Range, where the majority of the shortening occurred in Late Jurassic to Early Cretaceous time with only minor amounts of shortening during Late Cretaceous and Early Tertiary time, the majority of the deformation and shortening in the northeastern Brooks Range appears to be Cenozoic in age. In the central and western Brooks Range, Cretaceous and younger foredeep deposits were only mildly

Figure 3.2. Generalized tectonic map of the northeastern Brooks Range, modified from Wallace and Hanks (1990). Solid teeth on thrust faults indicate older-over-younger thrust faults that duplicate stratigraphic section, open teeth indicate detachment surfaces along which there has been slip but no duplication of stratigraphic section. Line A-A' is location of area-balanced models (Figure 3.13) and the regional balanced cross section along the Aichilik River transect (Figure 3.16). The outlined area is shown in detail in figure 3.7.


146°
144°
142°
70°30'
70°
69°30'
69°
0 20 40 60 80 100 KM
Beaufort Sea
Canning R.
ARCTIC
COASTAL
PLAIN
A
A'
Hulahula R.
Okpilak R.
Jago R.
Aichilik R.
Egaksrak R.
Study area
SADLEROCHIT MTNS.
SHUBLIK MTNS.
FOURTH RANGE
KIKIKTAT MTN.
LEFFINGWELL RIDGE
AICHILIK RIVER
ANTICLINORIUM
OKPILAK BATHOLITH
MT. GREENOUGH ANTICLINORIUM
FRANKLIN
MTNS.
BATHTUB RIDGE
FRONT
CONTINENTAL
DIVIDE
THRUST
Lower Cretaceous to lower Tertiary Brookian sequence
Mississippian to Lower Cretaceous Ellesmerian sequence
Pre-Mississippian rocks (on this map, includes Mississippian Kekiktuk Conglomerate of the Ellesmerian sequence)
Devonian intrusive rocks

deformed by the advance of the fold and thrust belt (Mull, 1985; Oldow and others, 1987b), whereas in the northeastern Brooks Range, the Late Cretaceous and Tertiary foredeep deposits have been extensively involved in the thrusting, with deformation migrating north of the range-front and encroaching onto the modern day continental margin of northern Alaska (Figure 3.2; Reiser and others, 1971; Bader and Bird, 1986; Bruns and others, 1987; Grantz and others, 1987). Apatite-fission track ages indicate uplift ranging from 60 my at Bathtub Ridge in the south to 45 my west of the Sadlerochit Mountains and 25 my along Leffingwell Ridge (Figure 3.2; O'Sullivan, 1988 and pers. comm.). A 61 ± 10 Ma lead-loss event and a 59 ± 2 Ma K/Ar cooling age on recrystallized biotite from a granitic batholith and adjacent stock, and apatite fission track ages of 42 and 31 my from the batholith all probably reflect a Tertiary deformation event (Dillon and others, 1987; O'Sullivan, 1989; Hanks and Wallace, 1990). Deformed clastic foredeep deposits as young as Pliocene in age (Reiser and others, 1971) and active seismicity north of the range-front of the northeastern Brooks Range suggest that deformation in the region may be continuing today (Grantz and others, 1983, 1987; Moore and others, 1985a).

The autochthonous stratigraphy of the North Slope and northeastern Brooks Range can be divided into three distinct, unconformity-bounded depositional sequences (Figure 3.3; Reiser, 1970; Mull, 1982). The oldest and structurally lowest rocks exposed in the northeastern Brooks Range are an assemblage of lithologically heterogeneous, variably deformed and slightly metamorphosed Proterozoic to Devonian sedimentary and volcanic rocks[3] (Reiser, 1970; Reiser and others, 1980; Moore and others, 1985b). The stratigraphy and age of the pre-Mississippian rocks throughout the northeastern Brooks Range are poorly understood, as is their relationship to potentially correlative strata in Canada and the rest of the circum-Arctic region. Leffingwell (1919) first defined a pre-

[3] This pre-Mississippian sequence commonly has been referred to as the 'Franklinian sequence,' a term first used with respect to these rocks by Lerand (1973). This term refers to Early Paleozoic rocks deformed during the Franklinian orogeny of the Canadian Arctic Islands. The pre-Mississippian rocks of the northeastern Brooks Range include both Proterozoic and lower Paleozoic rocks, have not been definitively correlated with the supposedly coeval rocks of the Canadian arctic, and may represent a totally different stratigraphic and structural history. Consequently I have refrained from using the term 'Franklinian sequence' in this paper.

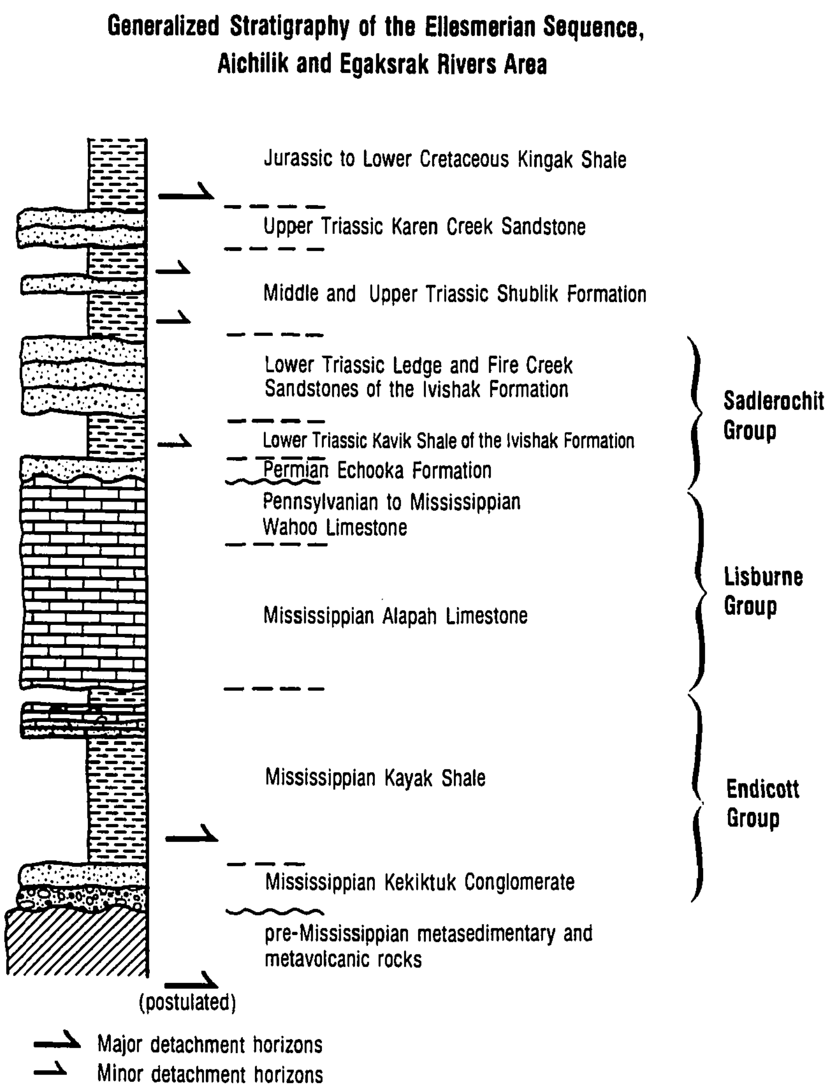


Figure 3.3. Generalized Ellesmerian Sequence stratigraphy of the northeastern Brooks Range.

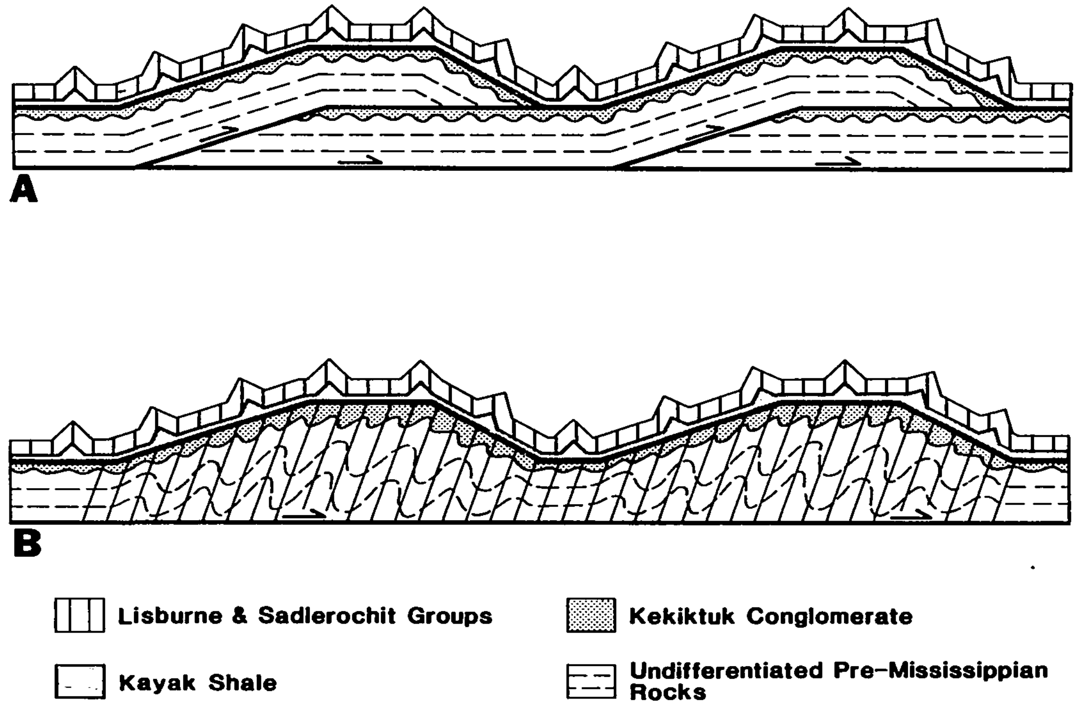
Mississippian unit, a quartzose sandstone/semischist, as the Neruokpuk Formation and suggested it was pre-Cambrian in age. Rare fossils indicate that at least a part of the pre-Mississippian sequence is Cambrian to Ordovician in age (Brosgé and others, 1962; Dutro, 1970; Dutro and others, 1972; Reiser and others, 1980; Moore and Churkin, 1984). Potentially correlative strata immediately east of the Canada/U.S.A. border have been dated as Lower Cambrian through Early Silurian in age (Lane and Cecile, 1989; Lane and others, 1991). Stratigraphic studies in the Sadlerochit and Shublik Mountains (Blodgett and others, 1986) indicate that the distinctive platform carbonates exposed there are Proterozoic through Early Devonian in age. Recent radiometric dating of mafic sills in metasandstones and pelites structurally underlying these platform carbonates has yielded a Rb/Sr isochron age of 801 ± 20 Ma and a Nd/Sm isochron age of 704 ± 38 Ma (Clough and others, 1990), suggesting that these igneous rocks and the metasedimentary rocks they intrude are also Proterozoic in age.

The pre-Mississippian metasedimentary and metavolcanic rocks of the northeastern Brooks Range have undergone at least one pre-Mississippian deformational event that resulted in the development of penetrative structures. South of Bathtub Ridge (Figure 3.2), Ordovician (?) isoclinally folded cherts and phyllites are unconformably overlain by Middle Devonian clastic rocks (Reiser and others, 1980; Anderson, 1991). Although the Middle Devonian section is generally underlain by a thrust fault, these Middle Devonian clastic rocks lack the penetrative fabrics and steep dips of the older Ordovician rocks, and are assumed to be separated from them by an angular unconformity. The Middle Devonian clastic rocks are also separated from overlying Mississippian clastic rocks by a slight angular unconformity, suggesting the possibility of a second pre-Mississippian, post-Middle Devonian deformation. The pre-Middle Devonian rocks locally have also been intruded by Devonian plutonic rocks (Sable, 1977; Reiser and others, 1980; Dillon and others, 1987).

Throughout most of the northeastern Brooks Range, the pre-Mississippian rocks are the depositional basement for overlying Mississippian through Lower Cretaceous carbonate and clastic sedimentary rocks of the Ellesmerian sequence (Figure 3.3). The base of the Ellesmerian sequence is a regional angular unconformity, with the underlying pre-Mississippian rocks generally upright and dipping steeply to the south (e.g., Reiser and others, 1980; Robinson and others, 1989). The Ellesmerian sequence was deposited on a south-facing passive continental margin, with its siliciclastic sediments derived from a source terrain north relative to the present-day rifted margin of northern Alaska (Mull, 1982; Mayfield and others, 1983). The Ellesmerian sequence is in turn overlain by middle Cretaceous through Tertiary southerly-derived clastic sedimentary rocks of the Brookian sequence (Figure 3.3). The source material for these marine and non-marine siliciclastic rocks was progressively eroded from the growing Brooks Range fold and thrust belt to the south and deposited in a foredeep. These deposits eventually filled in the foredeep and spilled over the modern continental margin of Arctic Alaska to the north (Mull, 1985; Molenaar and others, 1987).

The structure of the northeastern Brooks Range is dominated by a series of anticlinoria cored by pre-Mississippian rocks and overlain by variably deformed Mississippian and younger rocks (Bader and Bird, 1986; Figure 3.2). The origin of the anticlinoria and the mode of deformation of the pre-Mississippian rocks during Cenozoic thrusting have been the subject of some controversy. Recent regional and detailed structural studies by various workers (Rattey, 1985; Namson and Wallace, 1986; Leiggi, 1987; Kelley and Foland, 1987; Ziegler, 1989; Wallace and Hanks, 1990) have suggested that most of the Cenozoic shortening within the pre-Mississippian sequence has been by thrust duplication, with only minor amounts of shortening via penetrative structures. In this model, the pre-Mississippian-cored anticlinoria in northwestern ANWR are interpreted to reflect horses in a regional duplex between a floor thrust in the pre-Mississippian rocks and a roof thrust in a shale near the base of the overlying Ellesmerian sequence, the Mississippian Kayak Shale (Figure 3.4 A). In this interpretation, the Kayak Shale acted as a very effective detachment horizon, permitting the overlying Mississippian and younger cover sequence to deform independently of the underlying pre-Mississippian rocks, primarily by detachment folding. The basal sandstones and conglomerates of the Ellesmerian sequence, the Mississippian Kekiktuk Conglomerate, generally remained

Figure 3.4. Conceptual end-member models for the mode of shortening of the pre-Mississippian rocks and the Kekiktuk Conglomerate during Cenozoic thrusting. For simplicity and clarity, the attitude of the pre-Mississippian rocks with respect to the unconformity surface, an artifact of pre-Mississippian-age deformation, has not been included. The dashed lines within the pre-Mississippian rocks represent arbitrary markers that were horizontal prior to Cenozoic deformation, and do not represent bedding. Model A represents a north-vergent regional duplex, where Cenozoic shortening of the pre-Mississippian rocks is accommodated by thrust duplication. The floor thrust of the duplex is at depth in the pre-Mississippian rocks, and the roof thrust is in the Mississippian Kayak Shale. Pre-Cenozoic horizontal markers remain horizontal with respect to the unconformity during this style of deformation. Model B represents a scenerio where Cenozoic shortening within the pre-Mississippian rocks is accommodated primarily by strain and mesoscopic structures, and results in folding of the pre-Cenozoic horizontal markers. Steeply dipping solid lines within the pre-Mississippian rocks and Kekiktuk Conglomerate represent Cenozoic-age cleavage.


A
B
Lisburne & Sadlerochit Groups
Kekiktuk Conglomerate
Kayak Shale
Undifferentiated Pre-Mississippian Rocks

attached to the unconformably underlying pre-Mississippian rocks, and now defines the upper surface of the broad antiformal structures formed in the pre-Mississippian rocks.

In contrast to this interpretation, local detailed mesoscopic structural studies in the pre-Mississippian rocks of northwestern ANWR by other workers have been interpreted to indicate that most of the Cenozoic shortening in the pre-Mississippian rocks was accommodated by the development of mesoscopic structures and penetrative fabrics (Figure 3.4 B; Oldow and others, 1987a). As in the previous interpretation, rocks above the Kayak Shale are interpreted to have been structurally detached from the underlying pre-Mississippian rocks and deformed independently during Cenozoic deformation. However, this interpretation requires that Cenozoic shortening of the Mississippian Kekiktuk Conglomerate be accommodated by significant penetrative strain on a regional scale and/or detachment and duplexing above the pre-Mississippian rocks. This type of behavior by the Kekiktuk Conglomerate has yet to be documented on a regional scale.

In both interpretations, the Mississippian Kayak Shale plays a major role in influencing the structural behavior of the overlying carbonate and clastic rocks of the Ellesmerian sequence, with regional variations in its lithology and thickness key factors in its effectiveness as a regional detachment horizon (Wallace and Hanks, 1990). Where the Kayak Shale is thick, shaly and incompetent, the overlying Mississippian and younger rocks are deformed into generally upright and tight detachment folds (e.g., Canning River region, Namson and Wallace, 1986). Where the Kayak Shale is present, but has a significant component of carbonate and siltstone, the overlying carbonate and clastic rocks have deformed via thrust duplication (e.g., Leffingwell Ridge, Hanks, 1987). Where the Kayak Shale is depositionally thin to absent, structurally competent carbonates of the Mississippian to Pennsylvanian Lisburne Group depositionally overlie the pre-Mississippian rocks and have deformed with them as a single structural unit (e.g., Sadlerochit Mountains, Wallace and Hanks, 1990; and the Okpilak batholith, Hanks and Wallace, 1990).

The northeastern Brooks Range between the Canning River and the Canada/U.S.A. border can be divided into three structural provinces based on regional variations in the structural behavior of both the pre-Mississippian rocks and the Ellesmerian cover sequence (Figure 3.5; Wallace and Hanks, 1990). In defining these provinces, structural interpretations of the Cenozoic behavior of the pre-Mississippian sequence drew heavily on the duplex model of figure 3.4 (model A). In the western structural province, each pre-Mississippian-cored anticlinorium is defined by a single major horse in the regional duplex. The overlying Ellesmerian sequence has deformed primarily by detachment folding, except in the Sadlerochit Mountains, where the Ellesmerian sequence has deformed with the underlying pre-Mississippian sequence because the intervening Kayak Shale is depositionally absent. In the Okpilak batholith structural province, the pre-Mississippian rocks include a Devonian granitic batholith. During Cenozoic thrusting, the rocks of the batholith deformed by thrusting and penetrative strain. The Kayak Shale is depositionally thin to absent in the vicinity of the batholith, and the overlying Ellesmerian sequence has remained attached to the batholith during thrusting, deforming by the development of penetrative fabrics and local folds and thrusts (Hanks and Wallace, 1990). In the eastern structural province, the regional anticlinoria are defined by multiple horses, and the overlying Ellesmerian sequence has deformed via thrust duplication and detachment folding. This paper documents the structural style of the eastern structural province and presents an interpretation of the geometry and sequence of Cenozoic deformation in that area.

3.4 A detailed structural analysis of the eastern structural province: the Aichilik River transect

The eastern structural province, as defined by Wallace and Hanks (1990), consists of that part of the northeastern Brooks Range between the Jago River and the Canada/U.S.A. border, bounded to the south by the Continental Divide Thrust Front (Figure 3.2 and 3.5). As part of a regional structural analysis of the northeastern Brooks Range by University of Alaska geologists, I conducted detailed (for the region) mapping and structural analysis of the northern portion of a transect across this province. This

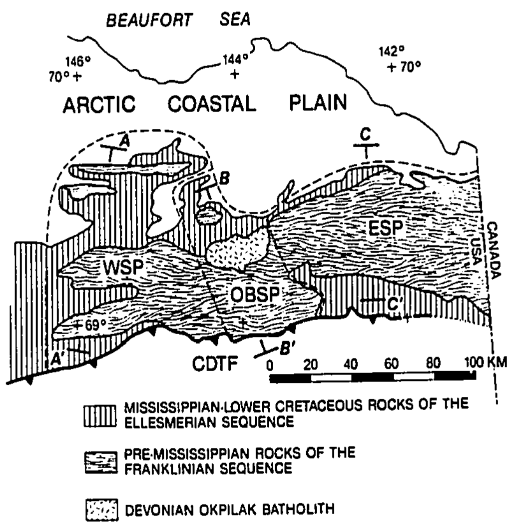


Figure 3.5. Generalized geologic map of the northeastern Brooks Range showing the major structural provinces, as defined by Wallace and Hanks (1990). WSP = western structural province, OBSP = Okpilak batholith structural province, ESP = eastern structural province, CDTF = Continental Divide thrust front.

detailed information was integrated with published regional maps and seismic data from the coastal plain to the north in order to derive a structural interpretation of the province. The transect extends from the coastal plain in the north through Leffingwell Ridge and Bathtub Ridge to south of the Continental Divide Thrust Front (A-A', Figure 3.2).

In general terms, the structure of the eastern structural province is characterized by two broad and arcuate anticlinoria cored by pre-Mississippian rocks, with the limbs defined by Mississippian through Triassic rocks of the Ellesmerian sequence (the Aichilik River anticlinorium in the north, the Mt. Greenough anticlinorium in the south, Figure 3.2). Detailed mapping and structural analysis documented in this paper focussed on the Aichilik River anticlinorium and the northern flank of the Mt. Greenough anticlinorium between the Aichilik and Egaksrak Rivers.

3.5 Stratigraphy

3.5.1 Pre-Mississippian rocks of the depositional basement

In the eastern structural province, the pre-Mississippian rocks consist of a variety of low-grade metasedimentary rocks and minor metavolcanic rocks. Age control is very sparse, especially in the north (Reiser and others, 1980). In the Aichilik River anticlinorium and the northern margin of the Mt. Greenough anticlinorium, the pre-Mississippian sequence consists predominantly of heterogeneous, multiply-deformed and slightly metamorphosed carbonate and clastic rocks. These rocks have been described in detail by Reiser and others (1980) and Hanks (1989) and are summarized in figure 3.6 and Appendix A. For the purposes of this paper, the pre-Mississippian rocks of the Aichilik River anticlinorium have been divided into four lithologic packages, each of which consists of several distinct and mappable units.

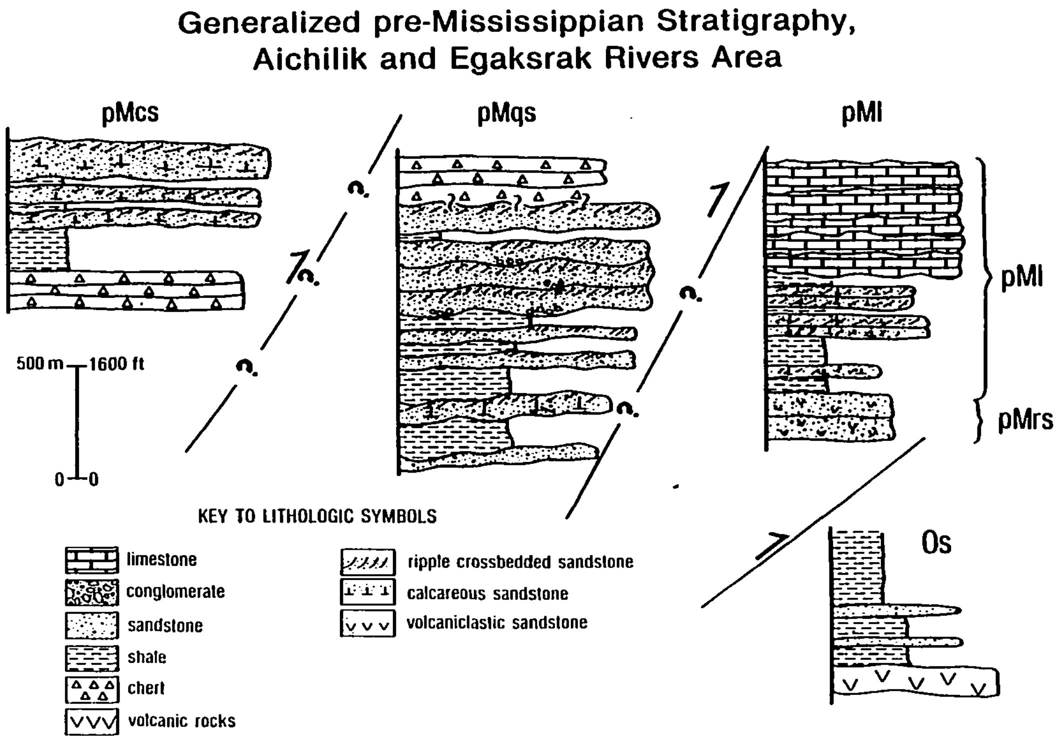


Figure 3.6. Generalized stratigraphy of the pre-Mississippian rocks of the Aichilik and Egaksrak Rivers area, northeastern Brooks Range, based on field observations by the author. Thicknesses represent the observed structural thicknesses and are therefore only approximate. It was not possible to determine the true stratigraphic thickness due to poor exposure, lack of fossil control and structural disruption.

Ordovician siltstone and shales (Os)

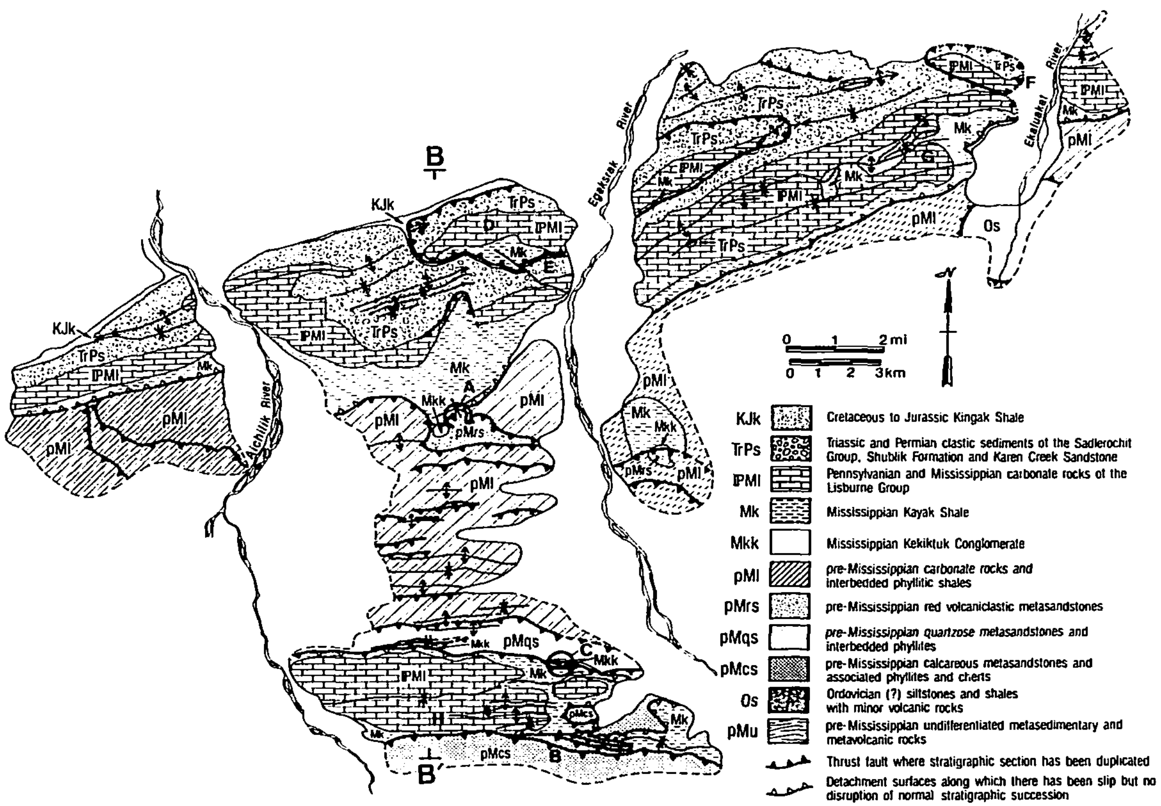
The structurally lowest lithologic package consists of shales, siltstones and minor mafic volcanic rocks that are found at the eastern edge of the study area (Os, Figures 3.6 and 3.7 A). Reiser and others (1980) correlated these rocks with shales bearing Ordovician graptolites near the U.S.A./Canada border. This unit may be correlative with the lower Paleozoic succession recognized by Lane and others (1991) along the USA/Canada border.

Pre-Mississippian carbonates (pMl)

The Ordovician clastic rocks are structurally overlain by a distinctive carbonate succession (pMl, Figures 3.6 and 3.7) that is exposed primarily in the northern half of the Aichilik River anticlinorium. The most obvious and best exposed constituents of this succession are thickly bedded partially or totally recrystallized black limestones and dolomites. Local relict textures suggest that these carbonates originally may have contained peloids and oncolites. Sedimentary structures, where preserved, suggest that some of the rocks were deposited as turbidites. While individual beds typically are 1-2 meters thick, they occur as part of a thick amalgamated sequence of similar beds with no intervening shales. These amalgamated carbonate intervals can be as thick as 80 to 100 meters and define most of the ridge tops in this area. Without age control or a known stratigraphy, however, it is unclear exactly how many of these carbonate sequences exist. It not known if there is only one amalgamated bedded carbonate interval that has been structurally repeated (as suggested by figure 3.6), or if there is a series of similar amalgamated intervals with intervening shales in a relatively intact, but internally structurally disrupted, stratigraphic sequence.

These thick, cliff-forming black carbonate rocks appear to stratigraphically overlie and/or interfinger with rippled sandy limestones/limy sandstones and brown argillites (Figure 3.6). The finer grained rocks are often more highly deformed than the massive ridge-forming carbonate rocks and are apparently the locus of structural disruption in the

Figure 3.7. (A) Generalized geologic map of Aichilik and Egaksrak Rivers area, as outlined in Figure 3.2. Letters A-G refer to specific locations mentioned in text.


B
B'
KJk
TrPs
IPMI
Mk
Mkk
pMl
pMrs
pMqs
pMcs
Os
A
B
C
D
E
F
G
H
Aichilik River
Egaksrak River
Ekaluakat River
0 1 2 mi
0 1 2 3 km
KJk Cretaceous to Jurassic Kingak Shale
TrPs Triassic and Permian clastic sediments of the Sadlerochit Group, Shublik Formation and Karen Creek Sandstone
IPMl Pennsylvanian and Mississippian carbonate rocks of the Lisburne Group
Mk Mississippian Kayak Shale
Mkk Mississippian Kekiktuk Conglomerate
pMl pre-Mississippian carbonate rocks and interbedded phyllitic shales
pMrs pre-Mississippian red volcaniclastic metasandstones
pMqs pre-Mississippian quartzose metasandstones and interbedded phyllites
pMcs pre-Mississippian calcareous metasandstones and associated phyllites and cherts
Os Ordovician (?) siltstones and shales with minor volcanic rocks
pMu pre-Mississippian undifferentiated metasedimentary and metavolcanic rocks
Thrust fault where stratigraphic section has been duplicated
Detachment surfaces along which there has been slip but no disruption of normal stratigraphic succession

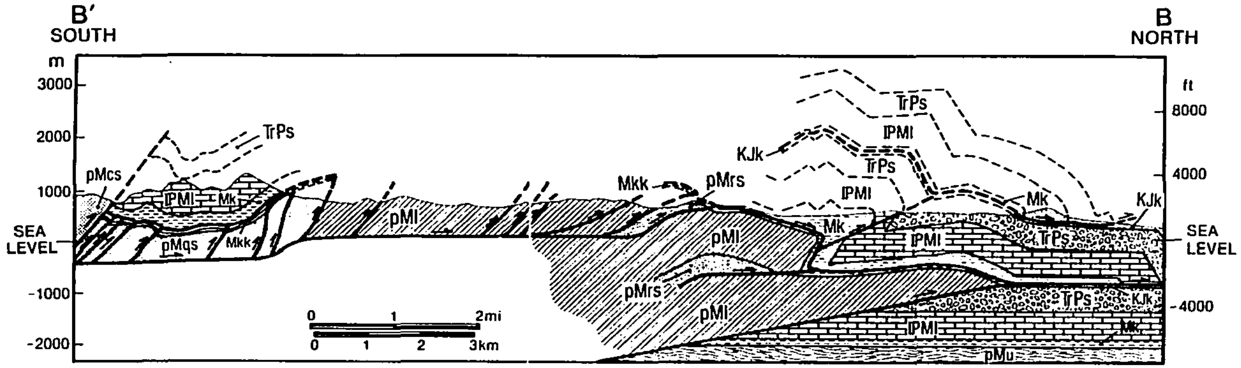


Figure 3.7. (B) North-south cross section through Aichilik River anticlinorium along section line B-B'.

sequence. The thickness and dominance of the black limestones and dolomites decreases eastward, possibly reflecting a facies change or stratigraphic thinning in that direction. The total thickness of the entire carbonate succession, including the massive black carbonates, the thinner-bedded rippled carbonates and the argillites, appears to vary greatly throughout the area, probably due both to stratigraphic variations and structural complexities, but may be 1000 meters (3200 feet) or more in the vicinity of the Aichilik River.

The age of this carbonate succession is uncertain. It was considered by Reiser and others (1980) to be pre-Cambrian in age, possibly because the carbonate sequence structurally underlies sandstones and Early Cambrian limestones exposed in the Mt. Greenough anticlinorium. It is also similar to pre-Mississippian carbonates exposed in the Fourth Range which Reiser considered to be deep-water equivalents of the Cambrian through Devonian Nanook Limestone and of the pre-Cambrian Katakturuk Dolomite (Reiser, 1970). The carbonate sequence of the Aichilik River area may also be correlative with similar carbonates along the USA/Canada border that are thought to be Early Cambrian to Proterozoic in age (Lane and Cecile, 1989; Lane and others, 1991).

Pre-Mississippian red volcaniclastic sandstones (pMrs)

In the northern part of the study area, and to the east on Redwacke Creek (Reiser and others, 1980), the carbonate sequence (pMl) locally overlies red-weathering sandstones containing varying amounts of volcanic debris (pMrs, Figures 3.6 and 3.7). This unit varies laterally to a great degree, ranging from fine-grained, well-sorted sandstones and thin shales to rare volcaniclastic breccias and possible recrystallized tuffs. The unit is highly deformed, with bedding difficult to discern. It is not clear whether the contact between these red-weathering sandstones (pMrs) and the overlying carbonate rocks (pMl) is depositional or structural in nature. The age of this unit and its relationship to the volcanic rocks exposed at Whale Mountain to the south are not known (Reiser and others, 1980).

Pre-Mississippian quartzose sandstones (pMqs)

In the southern part of the Aichilik River anticlinorium, the carbonate succession is structurally overlain by a predominantly siliciclastic sequence (pMqs, Figures 3.6 and 3.7). This clastic succession is dominated by a thick interval of quartz-lithic sandstones and local chert-granule conglomerates. The sandstones are generally well-bedded and highly rippled, with interbeds of maroon and tan siltstones. Feeding trails and burrows are preserved on the bedding surfaces of some siltstones. This sandstone package appears to be underlain by distinctive gray and white bedded cherts with interbeds of shale and maroon rippled sandstone and a heterogeneous package of siltstones, shales and rippled sandy limestones. The quartz-lithic sandstone unit is structurally overlain by a thick succession of gray, black and white bedded cherts.

The quartz-lithic sandstone unit (pMqs) was considered by Reiser and others (1980) to be correlative with the Neruokpuk Formation sensu strictu as defined by Leffingwell (1919). The occurrence of the trace fossils *Oldhamia* and *Planolites* in shales associated with a similar quartz-lithic sandstones in Canada have led Lane and Cecile (1989) to suggest that these rocks are Early Cambrian in age. I have recovered possible *Planolites* from quartz-lithic sandstones of the pMqs west of the Egaksrak River, but these samples have not yet been positively identified.

Pre-Mississippian calcareous siliciclastic rocks (pMcs)

The relationship of the quartz-lithic sandstones of pMqs to the calcareous sandstones and shales (pMcs) exposed on the northern flank of the Mt. Greenough anticlinorium to the south is unclear (Figures 3.6 and 3.7). These latter rocks consist of rippled fine-grained calcareous sandstones with interbedded black shales. The sandstones are occasionally bioturbated or contain sedimentary structures suggesting deposition via turbidity currents (scoured bases, Bouma sequences), while the shales are thick with local pebbly mudstone horizons. Reiser and others (1980) considered this sequence to be Early

Cambrian in age based on Cambrian trilobites from a limestone at the base of structurally overlying volcanic rocks immediately south of the study area in the Mt. Greenough anticlinorium. This unit is more calcareous and shale-rich than the sandstone succession to the north (pMqs). However, bedded cherts and maroon shales similar to those of pMqs are also associated with these more calcareous clastic rocks. The contacts between the calcareous sandstones, cherts and shales are all structurally disrupted, so that the original stratigraphic relationships between the various rock types of pMcs and between the two different clastic units (pMcs and pMqs) are not clear. The apparent lithologic differences between two clastic units in the two anticlinoria may reflect facies changes within a related sequence, or the two units may be stratigraphically unrelated and structurally juxtaposed.

3.5.2 Mississippian through Triassic rocks of the Ellesmerian sequence

Both the northern and southern flanks of the Aichilik River anticlinorium are defined by Mississippian through Triassic rocks of the Ellesmerian cover sequence (Figure 3.7). The stratigraphy of these rocks is similar to that of coeval rocks of the North Slope subsurface, with lithologic variations due to facies changes (Figure 3.3; Reiser, 1970).

The base of this sequence is a dominantly siliciclastic succession, the Mississippian Endicott Group (Mkk, Mk, Figure 3.3 and 3.7). In the Aichilik and Egaksrak Rivers area, the Endicott Group can be divided into two components--a basal siliciclastic succession and an upper succession dominated by carbonates. At the base of the siliciclastic succession, the Mississippian Kekiktuk Conglomerate overlies the pre-Mississippian sequence. Where exposed, this sub-Mississippian surface is an angular unconformity. The Kekiktuk Conglomerate characteristically is laterally discontinuous, highly variable in thickness (0 to 30 meters) and lithology, and only locally exposed. Where present, it generally consists of quartz and chert granule conglomerates and medium-grained sandstones, with local siltstone or shale interbeds.

The Kekiktuk Conglomerate is overlain by a thick interval of shale, the Mississippian Kayak Shale. The roof thrust of the regional duplex is localized near the base of this horizon. The Kayak Shale varies considerably in character and thickness within the map area. On the southern limb of the Aichilik River anticlinorium, the shale is fairly thick (>100 meters) with isolated 20-30 cm thick beds of bioturbated sooty siltstones containing abundant macerated plant fragments. On the northern limb of the anticlinorium, the shale is generally thinner (<50 meters) and siltier than in the south. The exact thickness of the shale portion of the Endicott Group on both limbs of the anticlinorium is difficult to ascertain and variable because of structural thickening and thinning.

The upper part of the Endicott Group is composed of carbonate rocks interbedded with shales (Figure 3.3). In the north along Leffingwell Ridge, this carbonate succession locally comprises up to 50% of the Endicott Group, reaches thicknesses of 200 meters, and consists of orange- to tan-weathering crinoidal grainstones and local biohermal coral buildups with interbedded black shales. These carbonate rocks are laterally highly variable in both facies and thickness. The thickness of this carbonate succession decreases to the south to less than 15 meters, and generally consists of orange-weathering, medium-grained calcareous sandstones, sandy limestones and grainstones.[4]

The color, lithology, and lateral variability in facies of the carbonate rocks within the upper portion of the Endicott Group distinguishes them from the overlying limestones of the Lisburne Group. The laterally variable and orange-weathering carbonate succession of the Endicott Group is abruptly overlain by the gray peloidal wackestones and packstones of the Mississippian and Pennsylvanian Lisburne Group, a regionally extensive carbonate platform sequence (lPMl, Figures 3.3 and 3.7). The Lisburne Group is up to 615 meters thick in this area, and acts as the dominant structural component in the Ellesmerian

[4] This carbonate interval in the upper part of the Mississippian Kayak Shale occupies approximately the same stratigraphic position as the Itkilyariak Formation of the North Slope subsurface as described by Mull and Mangus (1972). However, the Kayak carbonates of Leffingwell Ridge bear little resemblance to the type section of the Itkilyariak Formation as exposed in the Sadlerochit Moutains. In the type section, the Itkilyariak occurs as a thin and discontinuous mixed carbonate and siliciclastic horizon between the Lisburne Group and the underlying pre-Mississippian carbonates with no intervening Kayak Shale. The type section of the Itkilyariak Formation therefore could be correlative with either the Kekiktuk Conglomerate or the Kayak limestones. Further study of this interval is obviously needed, and presently is part of a Ph.D. dissertation at UAF by D. LePain. For more information, see LePain and Crowder, 1991.

sequence. The detailed stratigraphy of the Lisburne Group is described by Watts and others (in review).

The Lisburne carbonate platform sequence is unconformably overlain by a complex Permian and Triassic clastic succession derived from the north (TrPs, Figures 3.3 and 3.7). This clastic sequence consists of three main parts in the eastern ANWR. The majority of the Permian and Triassic sedimentary sequence consists of the Permian and Triassic Sadlerochit Group, a thick (approximately 300 meters) sequence of shallow marine sandstones and shales (Crowder, 1990; Figure 3.3). The Sadlerochit Group is overlain by the Triassic Shublik Formation, a thick (approximately 175 meters) dominantly shallow marine shale with thin limestones and phosphatic sandstones. The Shublik Formation is in turn overlain by a distinctive shallow marine sandstone approximately 45 meters thick, the Triassic Karen Creek Sandstone. These Permian and Triassic clastic rocks are in turn overlain by a thick Jurassic and Cretaceous shale, the Kingak Shale (KJk, Figure 3.3 and Figure 3.7). The stratigraphy of these rocks is discussed in more detail by Detterman and others (1975).

3.6 Structural observations

The regional Cenozoic structural style is, as mentioned previously, that of large, thrust-related anticlinoria cored by pre-Mississippian rocks, with the flanks of the anticlinoria defined by Mississippian and younger rocks. The structural behavior of the Ellesmerian sequence rocks in northeastern ANWR during Cenozoic thrusting can be determined, at least locally, by study of the flanks of the Aichilik River anticlinorium. In contrast, although the pre-Mississippian rocks exposed in the core of the Aichilik River anticlinorium are highly deformed at both map and outcrop scale, it is difficult to easily ascertain whether individual structures are pre-Mississippian or Cenozoic in age. This problem arises because most of the Mississippian and younger rocks are eroded from the core of the anticlinorium. However, regional observations, local examples of involvement of Mississippian rocks in the Cenozoic deformation, and trends of mesoscopic structures all yield clues as to which structures in the pre-Mississippian rocks are related to the Cenozoic formation of the anticlinorium.

At least two distinct deformational episodes are preserved in this part of the Aichilik River anticlinorium. D1 structures are preserved only in the pre-Mississippian rocks, and include mesoscopic and map-scale structures related to folding and thrusting. This D1 event appears to pre-date the sub-Mississippian unconformity and is pre-Mississippian in age. D2 structures occur in both the pre-Mississippian rocks and in the Ellesmerian cover sequence and are probably related to the formation of the Aichilik River anticlinorium during Cenozoic time. The nature and scale of development of D2 structures depend upon location and rock type. The geometry and distribution of both D1 and D2 structures will be discussed in more detail in the following sections.

3.6.1 Structural style of the Mississippian and younger rocks

The shale near the base of the Ellesmerian sequence, the Mississippian Kayak Shale, acted as a major detachment horizon during Cenozoic formation of the Aichilik River anticlinorium, permitting the overlying limestones of the Carboniferous Lisburne Group and the siliciclastic rocks of the Permian and Triassic Sadlerochit Group, Shublik Formation and Karen Creek Sandstone to deform independently of the underlying pre-Mississippian rocks. Where exposed, the Kayak Shale commonly is highly deformed and structurally thickened, displaying tight to isoclinal, upright to overturned-to-the-north mesoscopic folds and a related axial planar cleavage. Deformation is particularly strong where Cenozoic thrust faults have cut up-section from the pre-Mississippian rocks and flattened in the Kayak Shale (e.g., Figure 3.7 A, location A; Figure 3.8). In these locations, bedding and D2 mesoscopic structures such as fold axes and axial planar cleavage within the deformed Kayak Shale generally trend east-northeast, suggesting a north-northwest direction of tectonic transport (Figure 3.9 A).

In contrast with the Kayak Shale detachment horizon, the underlying Mississippian Kekiktuk Conglomerate generally appears to have deformed with the pre-

Figure 3.8. Photograph looking west-southwest at pre-Mississippian rocks that have been imbricated in the footwall of a Cenozoic thrust during formation of the Aichilik River anticlinorium. The Cenozoic faults appear to offset the sub-Mississippian unconformity, but disappear upsection within the overlying Mississippian Kayak Shale, suggesting that they merge with a detachment within that unit.

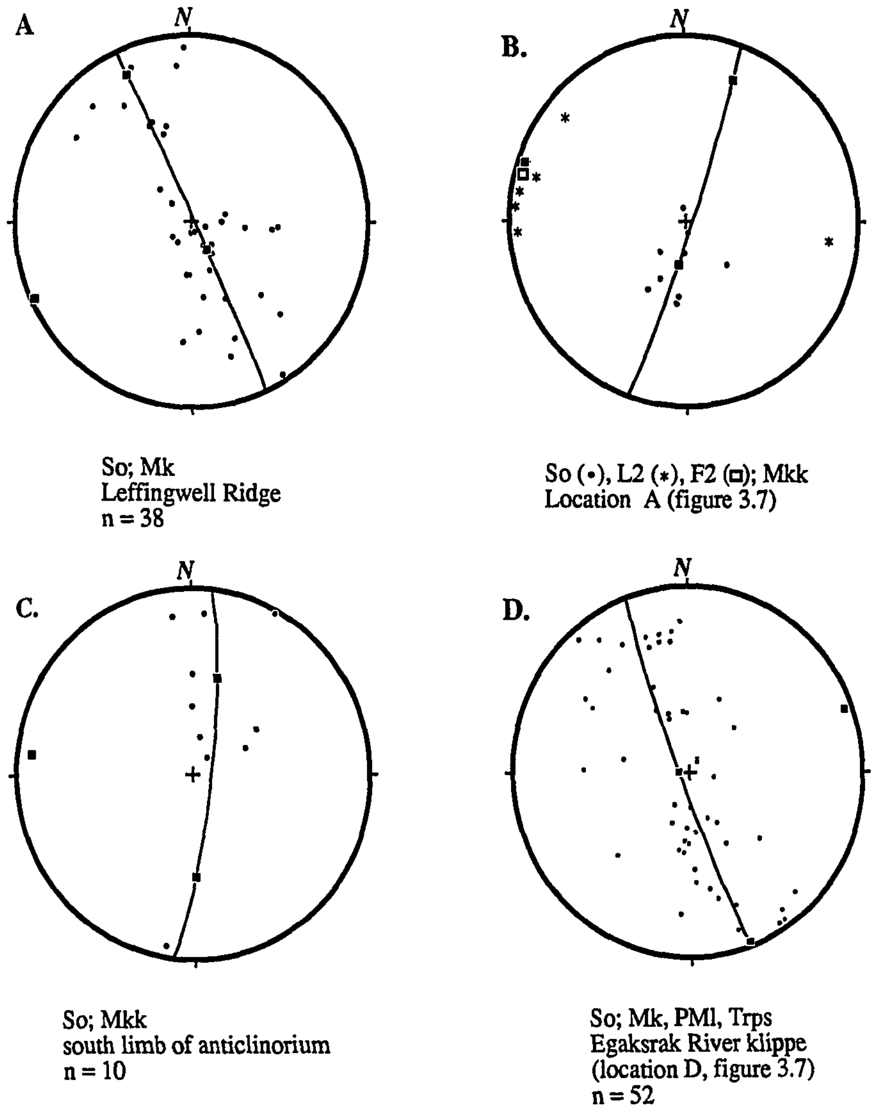
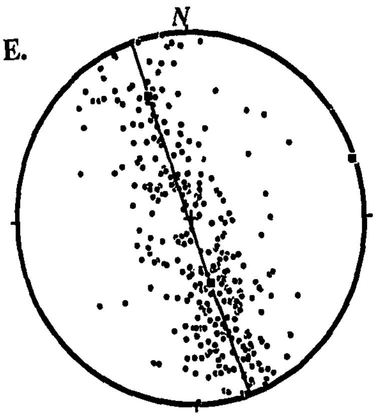


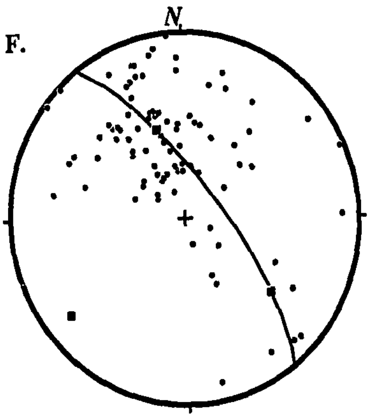
Figure 3.9. Lower hemisphere, equal area stereographic projections of poles to bedding and D2 structures from Ellesmerian sequence rocks of the Aichilik River anticlinorium. Mk = Mississippian Kayak Shale; Mkk = Mississippian Kekiktuk Conglomerate; PMl = Pennsylvanian to Mississippian Lisburne Group; TrPs = Triassic to Permian Sadlerochit Group.

E.



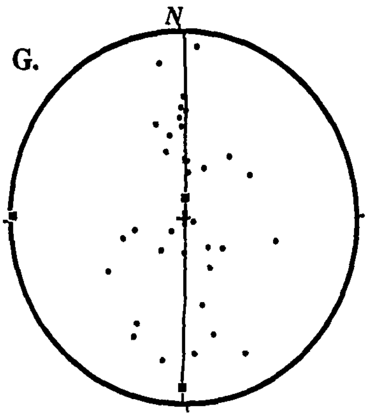
So; Mk, PMl, TrPs
Leffingwell Ridge
n = 328

F.



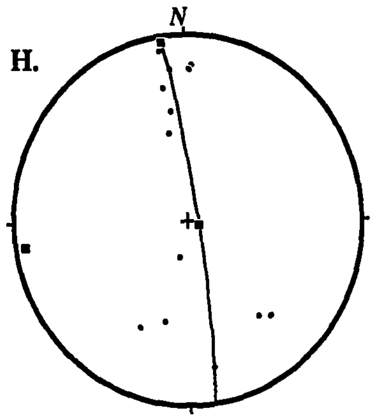
S1; Mk, PMl, TrPs
Leffingwell Ridge
n = 89

G.



So; Mk, PMl, TrPs
south limb of anticlinorium
n = 36

H.



S1; Mk, PMl, TrPs
south limb of anticlinorium
n = 13

Figure 3.9. (cont)

Mississippian sequence during Cenozoic formation of the anticlinorium. Where present, the Kekiktuk Conglomerate has remained attached to the underlying pre-Mississippian sequence, providing a clear marker that defines the geometry of the sub-Mississippian unconformity (e.g., Figure 3.7 A, locations A, B and C). Depending upon whether the Kekiktuk Conglomerate is in the footwall or hanging wall of the Cenozoic thrust fault, the lithology of the conglomerate, and the lithology and degree of deformation within the underlying pre-Mississippian sequence, the Kekiktuk Conglomerate varies from undeformed to mildly deformed. D2 structures include extension fractures, slightly stretched pebbles or a semi-pervasive spaced cleavage. Deformation within the Kekiktuk Conglomerate appears to be greatest in the footwall of Cenozoic thrusts and where the conglomerate is thin and/or has interbeds of shale or finer-grained sandstone. In all instances, however, the Kekiktuk Conglomerate does not appear to be detached from the underlying pre-Mississippian rocks, with no instances of thrust duplication or detachment folding of the Kekiktuk Conglomerate above the pre-Mississippian sequence. Bedding orientation and mesoscopic structures in the Kekiktuk Conglomerate show either east-west, or west-northwest trends (Figure 3.9 B & C), probably reflecting the influence of the D1 structural grain within underlying pre-Mississippian rocks on the orientation of these D2, Cenozoic, structures. The underlying pre-Mississippian rocks, however, are everywhere more highly deformed and show more mesoscopic shortening than the overlying Kekiktuk Conglomerate, reflecting shortening during the pre-Mississippian D1 deformational event.

Above the Kayak Shale, the thick limestones of the Carboniferous Lisburne Group have acted as the main structurally competent interval of the Ellesmerian sequence, and consequently have controlled the deformational style of the overlying Permian and Triassic clastic rocks. Although the Ellesmerian sequence has been eroded from over most of the anticlinorium, a large klippe on the north flank of the anticlinorium between the Aichilik and Ekaluakat rivers (Figure 3.7 A, location D) suggests that the Ellesmerian sequence above the Kayak Shale has been shortened primarily by large-scale thrust duplication. This klippe (here named the Egaksrak River klippe) preserves a stratigraphically complete sequence from the upper portions of the Mississippian Kayak Shale through the Triassic

Karen Creek Sandstone. The klippe is laterally extensive, extending approximately 22 km (13 miles) from west of the Egaksrak River to west of the Ekaluakat River (see Figure 3.7 A). In general, the klippe is preserved in two synclinal lows north of Leffingwell Ridge proper. In the main body of the klippe, the rocks dip steeply to the north. However, on both the western and eastern ends of the klippe, the rocks within the klippe are deformed into large tight map-scale anticlines overturned to the north, reflecting the structural geometry of the hanging wall rocks adjacent to the floor thrust.

The stratigraphy of the klippe is the same as that of Leffingwell Ridge and includes well-developed carbonate buildups in the upper portions of the Kayak Shale. These rocks are thrust over the Karen Creek Sandstone of Leffingwell Ridge proper. Two major detachment horizons were active during emplacement of the klippe: the Mississippian Kayak Shale at the base of the thrust sheet and the Jurassic-Cretaceous Kingak Shale at the top of the thrust sheet. Where these shales are exposed, very little of either unit has been incorporated into the klippe itself. In the case of the Kayak Shale, siltstones and well-developed limestones of the upper part of the Kayak Shale are preserved, but not a significant portion of the underlying shale. This suggests that the detachment surface is located in the upper portion of the Kayak Shale. Little of the Kingak Shale is preserved in the klippe, implying that the upper detachment is located near the base of the Kingak Shale. These two horizons are prominent detachment horizons elsewhere in the northeastern Brooks Range (Kelley and Foland, 1987; Wallace and Hanks, 1990).

The klippe itself was folded after its emplacement into a map-scale, east-northeast trending syncline by the formation of Leffingwell Ridge and related structures. This fold probably developed in two phases. Initially, the klippe was isolated north of Leffingwell Ridge by the development of the Aichilik River anticlinorium, of which Leffingwell Ridge is the northern limb (Figure 3.7 B). Thus, Leffingwell Ridge forms the southern limb of the map-scale east-northeast-trending syncline in which the klippe is preserved. The northern limb of this map-scale syncline formed later, during the development of large folds north of both the klippe and Leffingwell Ridge. These younger folds could be due to either detachment folding of the Lisburne Group above the Kayak Shale, or emplacement of another Cenozoic horse of pre-Mississippian rocks at depth north of the Aichilik River anticlinorium. Regardless, the resulting map-scale syncline in which the Egaksrak River klippe is preserved trends east-northeast, as is clearly seen both on the map (Figure 3.7 A) and in the stereographic projection of bedding data (Figure 3.9 D), and probably reflects a north-northwest tectonic transport direction. Other minor structures cutting the klippe also indicate a similar transport direction and are probably related to the formation of the anticlinorium. These minor structures include a north-northwest trending high-angle tear fault west of the Egaksrak River and an out-of-sequence thrust fault and related fold west of the Ekaluakat River (see Figure 3.7 A, locations E and F).

The preserved lateral extent of the Egaksrak River klippe implies that a significant portion of the Ellesmerian sequence in this region was involved in this style of deformation. However, because most of the Ellesmerian sequence is eroded from the crest of the Aichilik River anticlinorium, the footwall cut-off for the thrust sheet has not been preserved, and consequently the amount of shortening represented by the klippe is not known. Since the footwall cutoff must have been located in the Ellesmerian sequence eroded from the crest of the anticlinorium, the minimum amount of shortening represented by the klippe is 5 miles (8 km), with a maximum amount of shortening of greater than 10 miles (16 km).

Local detachment folding of the Lisburne Group and overlying horizons also played a role, albeit a relatively minor one, in the overall Cenozoic shortening of the Ellesmerian sequence in northeastern ANWR. The geometry of the map-scale detachment folds is controlled by and most visible in the thick carbonates of the Lisburne Group. These folds are generally tight to open, upright to slightly overturned to the north and have wavelengths of 1-2 km and amplitudes of 100+ meters. Detachment folds within the Ellesmerian sequence are present on both the northern and southern limbs of the anticlinorium (Figure 3.7 A, locations G and H). These detachment folds within the Lisburne Group control the first-order folding within the overlying Permian and Triassic siliciclastic rocks, with second-order detachment folding within the clastic sequence facilitated by detachments in the Triassic Kavik Member of the Ivishak Formation and Shublik Formation (Figure 3.3).

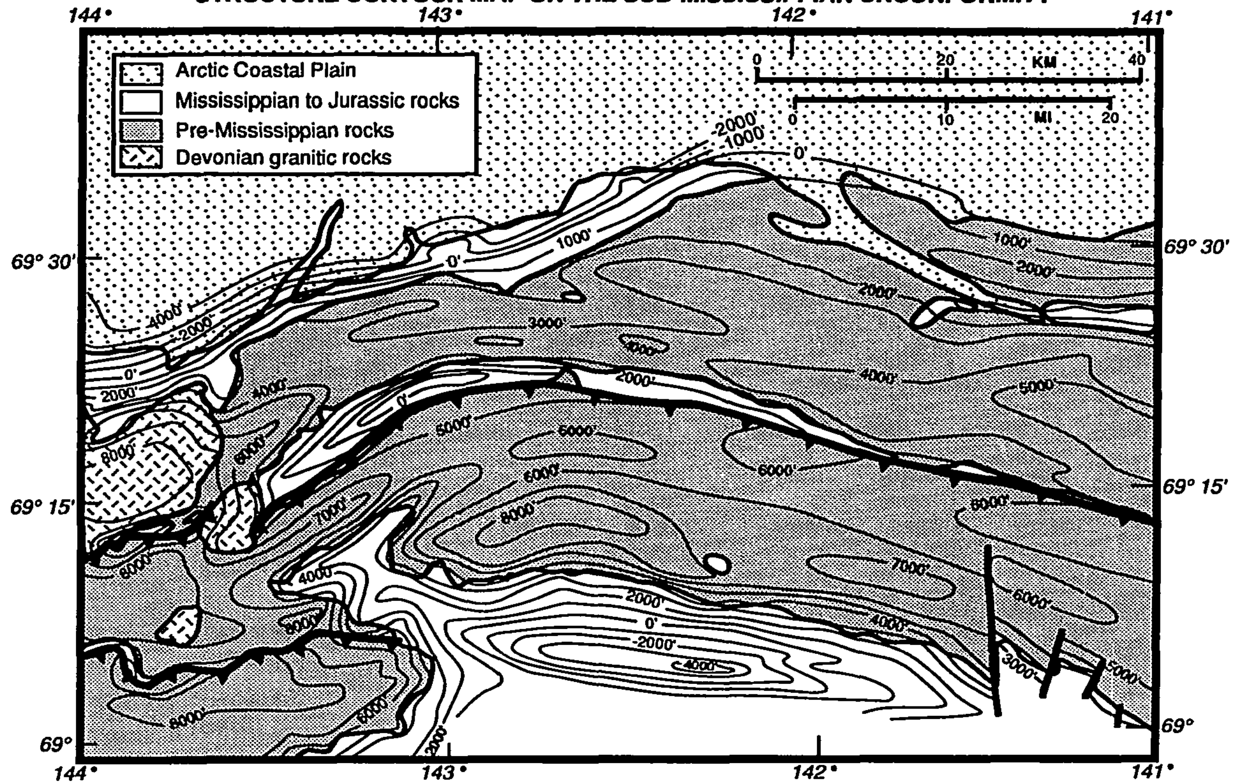
Slaty cleavage in shales and a spaced dissolution cleavage in the limestones have both developed locally, and are probably partially related to the detachment folding. Both map-scale and mesoscopic scale fold axes and mesoscopic fabrics within the Ellesmerian sequence rocks of the northern limb (Leffingwell Ridge) of the Aichilik River anticlinorium display east-northeast trends (Figures 3.9 E and F). These trends are interpreted to reflect a north-northwest tectonic transport direction during Cenozoic deformation. East-west trends seen in the Ellesmerian sequence rocks above the Kayak Shale on the southern limb of the anticlinorium (Figure 3.9 G and H) probably reflect reorientation of these northwest-directed structures by later reactivation of older, pre-Mississippian structural trends in the underlying pre-Mississippian rocks as progressively deeper structural levels were involved in Cenozoic thrusting (see following section).

3.6.2 Structural style of the pre-Mississippian rocks

The structural style of the pre-Mississippian rocks during both D1 and D2 deformation are recorded in both the development of mesoscopic and map-scale D1 and D2 structures. The structural relief of the pre-Mississippian rocks in the core of the Aichilik River anticlinorium with respect to Cretaceous and Tertiary foredeep deposits beneath the coastal plain to the north clearly indicates that the pre-Mississippian rocks were involved in the Cenozoic D2 thrusting event and the formation of the anticlinorium (Figure 3.10 A). However, as mentioned earlier, absence of preserved Ellesmerian sequence rocks across most of the anticlinorium renders it difficult to determine the absolute age of the multiple generations of mesoscopic and map-scale structures in the pre-Mississippian rocks. The problem is compounded by the fact that the Kayak Shale has acted as a remarkably effective detachment horizon, permitting most of the Ellesmerian sequence to deform totally independently of the underlying pre-Mississippian rocks. Thus, establishing the relationship between structures across the Kayak Shale is difficult. The most useful unit for determining the absolute timing of various structures within the pre-Mississippian sequence, the Mississippian Kekiktuk Conglomerate, is only locally exposed over most of

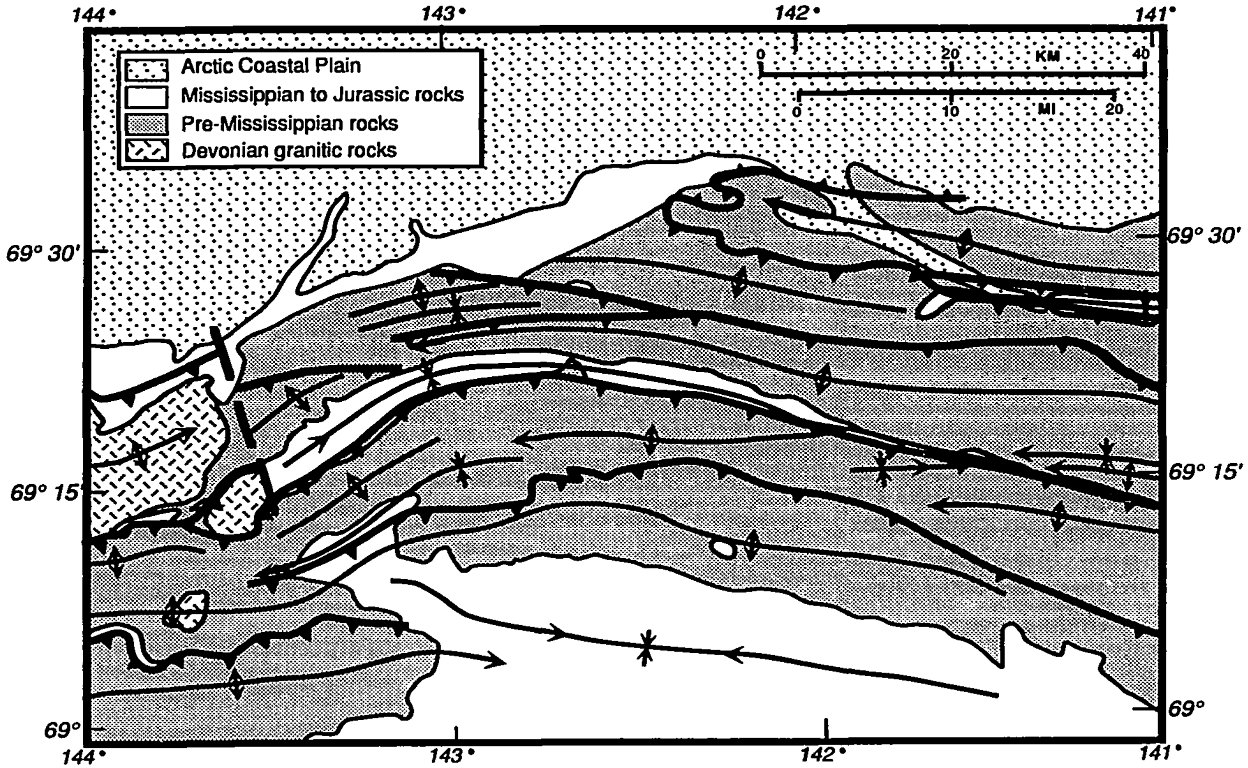
Figure 3.10. (A) Structure contour map of the sub-Mississippian unconformity in the exposed regions of eastern ANWR, with a contour interval of 1000 feet. Data derived from Bader and Bird (1986) by extrapolation into the subsurface using published thicknesses and by assuming a minimum elevation of the unconformity surface over areas of exposed pre-Mississippian rocks.

STRUCTURE CONTOUR MAP ON THE SUB-MISSISSIPPIAN UNCONFORMITY



(B) Map showing interpreted traces of Cenozoic folds and thrust faults that deform the sub-Mississippian unconformity in eastern ANWR. Crests and troughs of folds are based on the structure contour map. Thrust faults are based on abrupt changes in structural relief on the structure contour map and on mapped regional geologic relationships illustrated on 1:250,000 maps by Reiser and others (1980) and Bader and Bird (1986), including post-Mississippian faults within the pre-Mississippian sequence, boundaries between major lithologic packages in the pre-Mississippian rocks, and regional map patterns of the exposed unconformity surface.

CENOZOIC STRUCTURES AFFECTING PRE-MISSISSIPPIAN ROCKS



the map area due both to its thin and discontinuous depositional geometry and to recent erosion.

Many of the larger D2 structures within the pre-Mississippian rocks are best seen at a regional scale where they can be tied to isolated exposures of Ellesmerian sequence rocks. In the following discussion, I will first present evidence of the scale and geometry of D2, and possibly D1, regional structures within the pre-Mississippian rocks. I will then discuss the geometry, relative age and distribution of mesoscopic structures preserved in the pre-Mississippian rocks related to the D1 and D2 events.

3.6.2-1 Regional D1 structures

In order to distinguish D1 and D2 structures at both a regional and mesoscopic scale, it is necessary to determine which structures pre-date and which post-date the sub-Mississippian unconformity. However, the sub-Mississippian unconformity is either eroded or buried in the subsurface over much of the northeastern Brooks Range. A structure contour map of the unconformity surface would illustrate the present geometry of that surface. This geometry combined with published mapping can then be used to infer the geometry of Cenozoic structures that have deformed the unconformity surface.

Such a structure contour map is illustrated in figure 3.10 A. This map was constructed at 1:250,000 scale by using the mapped elevation of the unconformity surface where it is exposed (from Bader and Bird, 1986) and projecting the elevation of the unconformity surface above and below the present erosion surface. In order to project the geometry of the unconformity surface below the present day erosion surface in areas where Ellesmerian sequence rocks are now exposed, the thicknesses of the exposed Ellesmerian sequence rocks were estimated using the available published maps and stratigraphic information (Armstrong and Mamet, 1975; Detterman and others, 1975; Reiser and other, 1980) and sections I measured specifically for structural thicknesses along Leffingwell Ridge.

In order to project the unconformity surface above the present erosion surface, I assumed that the Mississippian Kekiktuk Conglomerate has been barely eroded in areas where pre-Mississippian rocks are now exposed. This assumption seems reasonably valid in the Aichilik River anticlinorium, where part of or the entire Ellesmerian sequence is preserved in several isolated inliers and peaks. In the Mt. Greenough Anticlinorium, scattered local high peaks also preserve mapped remnants of Kekiktuk Conglomerate (Reiser and others, 1980). In both anticlinoria, where the unconformity is eroded I used the elevation of the highest peaks in each township and range. Using this sampling density, the effect of present-day erosion (e.g., rivers) is minimized. Thus, in the resulting structure contour map, the minimum elevation of the unconformity lies just above the highest peaks. A combination of erosional topography on the unconformity surface during deposition of the Kekiktuk Conglomerate, and present day erosion is a recognized source of error in this technique. However, the amount of observed Mississippian-age erosional relief on the unconformity rarely exceeds 70 meters (270 feet) (LePain, pers. comm.) and is negligible at the scale of the structure contour map. The amount of present day erosion is difficult to quantify, but would have to be consistently greater than 500 to 600 meters (1500 to 2000 feet) over relative small distances (less than 10 km (6 miles)) to have a significant impact on the topography illustrated on the structure map. This seems highly improbable. Therefore I conclude that this technique for determining the approximate elevation of the unconformity surface where it is not exposed is relatively valid.

This resulting structure contour map of the unconformity surface should reflect the geometry of map-scale Cenozoic structures within the pre-Mississippian rocks (Figure 3.10 A). Multiple highs and lows on the sub-Mississippian unconformity surface characterize the cores of both the Aichilik River and Mt. Greenough anticlinoria of northeastern ANWR. The lows commonly coincide with regionally mapped post-Mississippian thrust faults within the pre-Mississippian sequence, boundaries between different pre-Mississippian lithologies that may reflect post-Mississippian thrust faults, and/or local remnants of Ellesmerian sequence rocks (Reiser and others, 1980). Thus, the structural topography of the sub-Mississippian unconformity in northeastern ANWR can be interpreted to reflect multiple horses in the regional, Cenozoic-age duplex within the pre-Mississippian rocks, with the highs indicating hangingwall antiforms, and the lows reflecting thrust faults and associated synforms between horses (Figure 3.10 B). These inferred structures correspond well with the geometry of mapped Cenozoic structures as defined by the map pattern of the exposed sub-Mississippian unconformity (Figure 3.2 and 3.10 B). Good examples of this correlation include the anticline/syncline/anticline triplet defined by the unconformity surface in the western part of the Mt. Greenough anticlinorium, and the multiple east-trending thrusts and related folds involving the Ellesmerian sequence in the northeastern part of Aichilik River anticlinorium. The apparent truncation of these inferred structures by other Cenozoic structures (e.g., the multiple highs in the core of the Mt. Greenough anticlinorium by the Cenozoic thrust fault along the northern flank of the anticlinorium) may reflect the interaction of pre-Mississippian-age D1 structures and Cenozoic D2 structures, or multiple episodes of Cenozoic deformation. Intersection of inferred Cenozoic thrust faults in the pre-Mississippian sequence with the sub-Mississippian unconformity surface indicates flattening of the Cenozoic thrust fault into the regional roof thrust, the Mississippian Kayak Shale. It should be noted that orientation of bedding within the pre-Mississippian sequence will not necessarily reflect these Cenozoic structures. Bedding orientation will in most cases be controlled by the more pervasive pre-Mississippian D1 structures. For example, the regional syncline in which the Whale Mountain volcanics are preserved is a significant and regional Cenozoic structural high (Figure 3.10 A and B). This structure probably represents a pre-Mississippian syncline that was later incorporated into a Cenozoic hanging wall anticline.

West of the Egaksrak River, the overall trend of both the Aichilik River and Mt. Greenough anticlinoria is east-northeast, as is the trend of map-scale and mesoscopic Cenozoic folds and faults within the Ellesmerian sequence rocks. East of the Egaksrak River, the regional trend of the anticlinoria becomes east-west, and eventually south-southeast near the Canadian border. However, map-scale Cenozoic structures within the pre-Mississippian rocks of the cores of the anticlinoria generally trend uniformly east-west (Figure 3.10 A and B). In the following discussion of map-scale and mesoscopic structures, I will present evidence that suggests that this difference in trend west of the Egaksrak River between the regional anticlinoria, the Ellesmerian sequence and the map-scale Cenozoic structures within the pre-Mississippian rocks could reflect the reactivation of D1 pre-Mississippian-age structures during Cenozoic formation of the anticlinoria. I will also argue that the orientation of the D2 map-scale Cenozoic structures in the older rocks reflect a pre-Mississippian structural grain.

3.6.2-2 Map-scale and mesoscopic structures

The pre-Mississippian rocks are highly deformed at both map and mesoscopic scale, with lithology and related competency contrasts having a strong influence on their structural style. Map-scale structures within the pre-Mississippian carbonate succession (pMl) are characterized by north-vergent folds and thrust faults in the more rigid and thick-bedded limestones and dolomites above a detachment in the underlying shales and sandy limestones (Figure 3.7 A & B). The map-scale deformational style of the siliciclastic sequences (pMqs and pMcs) is less clear because of their limited exposure in the map area. Some structural repetition of the quartz-lithic sandstone portion of the quartzose siliciclastic sequence (pMqs) can be seen immediately north of the southern flank of the Aichilik River anticlinorium (Figure 3.7 A & B), suggesting that this sandstone interval may have deformed above a detachment in an underlying shaly horizon. Both the carbonate and siliciclastic successions are also highly deformed at a mesoscopic scale. Most of this small-scale deformation is localized in the shales and siltstones, with relatively few mesoscopic structures developed in the thicker-bedded and competent units. Both sequences exhibit at least two generations of structures.

The first deformational event, D1, is represented by a pervasive slaty cleavage, S1, that is commonly well-developed in the siltstones and shales of both the siliciclastic and carbonate sequences. This fabric is generally sub-parallel to bedding and south-dipping, although some steeply north-dipping S1 surfaces are present in the southern half of the Aichilik River anticlinorium where bedding is also steeply north-dipping and overturned to the south. (However, these north dips are probably a result of later D2 Cenozoic folding.) Identifiable F1 fold hinges are relatively uncommon, are restricted to shale horizons and are generally isolated isoclinal fold hinges with no obvious sense of vergence (Figure 3.11). Locally, the thick-bedded carbonate rocks of pMl also have a poorly developed dissolution cleavage. Generally, however, penetrative structures within both the competent carbonates and sandstones are relatively uncommon. Throughout the study area, the thick-bedded carbonates do display several sets of calcite-filled extensional fractures, but they were not studied in sufficient detail to determine their relationship to D1 and subsequent deformations.

Poor exposure and detachment of most of the competent carbonate and sandstone horizons from the underlying shales preclude any direct correlation of S1 cleavage in the shales with the dominant map-scale folds in the carbonates and sandstones. Stereographic projections of poles to both bedding and S1 surfaces in the carbonate rocks (pMl, Figures 3.12 A & C) and quartz lithic sandstone unit (pMqs, Figures 3.12 B & D) define girdles about east-trending subhorizontal axes, F1b. F1b folding of the S1 mesoscopic fabrics could reflect rotation of earlier formed thrusts, folds and related cleavage above younger thrust faults during a progressive D1 deformational event. Alternatively, F1a folding of the S1 fabrics could reflect a distinct later deformational event.

Mesoscopic structures related to a second deformational event, D2, are also generally localized in shales of both the carbonate and siliciclastic sequences but are less well-developed than S1. S2 generally is a south-dipping spaced fracture cleavage (Figure 3.11) and/or an anastomosing semi-pervasive shear surface. The spaced fracture cleavage is commonly axial planar to mesoscopic F2 open folds. These F2 folds are generally open to tight and overturned to the north, with amplitudes and wavelengths of less than 20 cm. Both types of S2 surfaces generally strike east-northeast, and dip moderately to the south (Figures 3.12 E & F). Stereographic projections of S2 suggest that this surface has been folded about a west-southwest-trending, subhorizontal to moderately plunging fold axis, corresponding to the orientation of the observed F2 folds. This open folding of S2 fabrics could be due to local diffraction of S2 in different lithologies and resulting fanning of the

Figure 3.11. Outcrop photograph of D1 and D2 mesoscopic structures in a shale within pMqs. D1 is represented by relatively rare F1 isoclinal folds, with an associated well-developed and areally extensive S1 axial planar cleavage. D2 is characterized by open, upright F2 folds with an associated spaced fracture cleavage, S2. North is to the left.

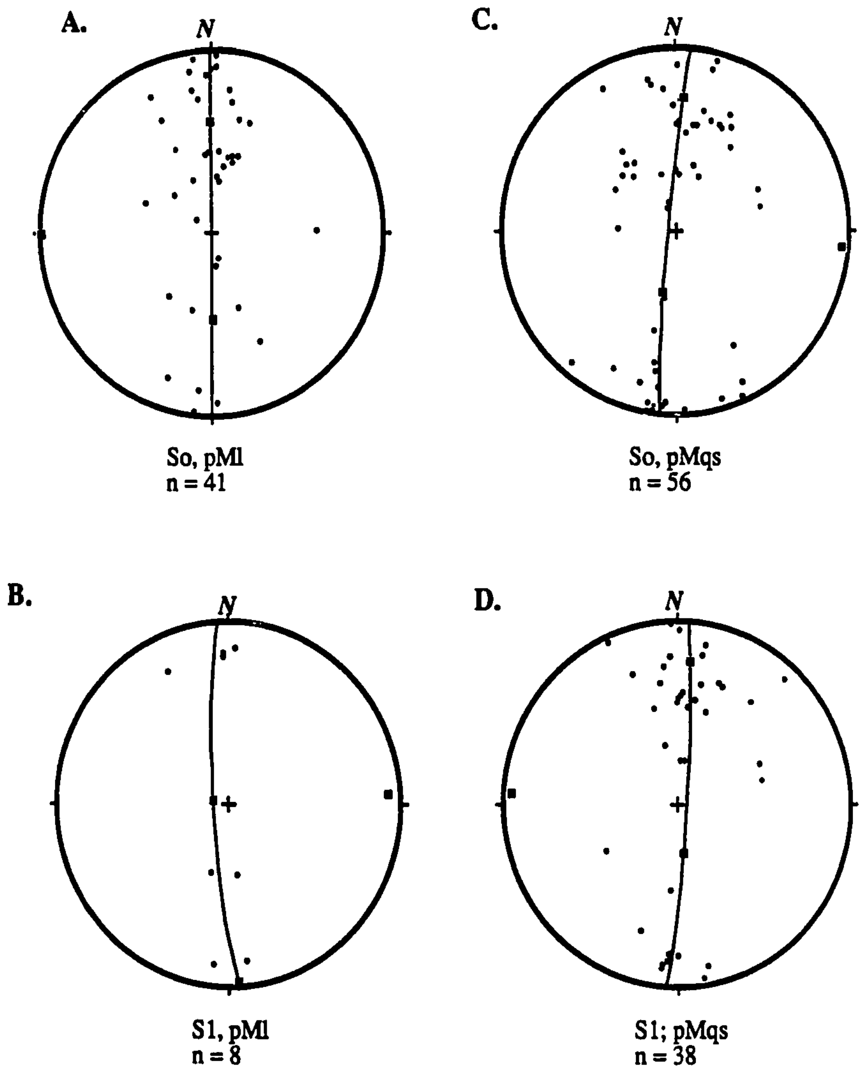
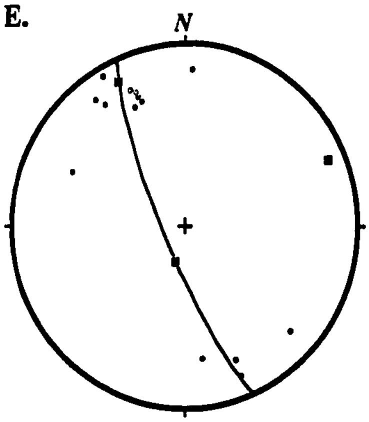


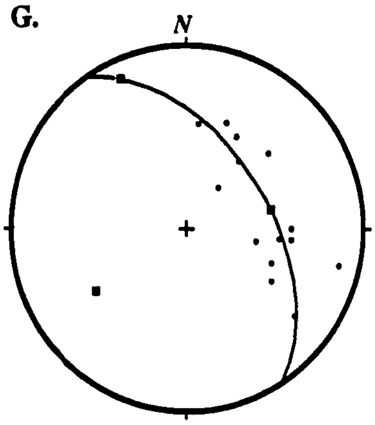
Figure 3.12. Lower hemisphere, equal area stereographic projections of poles to bedding and D1 and D2 planar fabrics in pre-Mississippian rocks of the Aichilik River anticlinorium.

E.



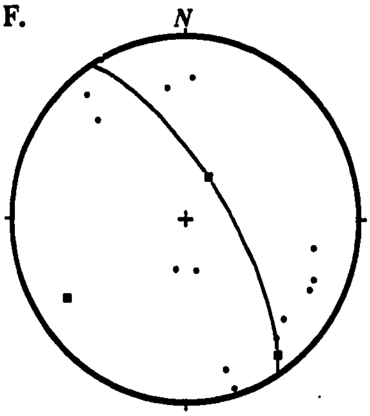
S2; pMl
n = 14

G.



So; pMl
location A (figure 3.7)
n = 14

F.



S2; pMqs
n = 14

Figure 3.12. (cont)

orientation of the fabric. Alternatively, D2 could have been a progressive deformational event, with earlier formed mesoscopic structures folded by subsequent map-scale structures.

However, D2 structures do not appear to have reoriented bedding or D1 fabrics throughout the study area (Figures 3.12 A - D). This may be partially due to the fact that D2 structures are not widely developed throughout the pre-Mississippian rocks of the study area and where present, are generally restricted to detachment horizons.

The absolute age of D1 and D2 throughout most of the study area is uncertain because the Mississippian Kekiktuk Conglomerate has been eroded. In two areas, however, the Kekiktuk Conglomerate has been deformed with pre-Mississippian rocks during thrusting and the age of structures relative to the Kekiktuk Conglomerate can be determined. In the northern portion of the Aichilik River anticlinorium, a thrust fault places the red volcaniclastic sandstone (pMrs) with a carapace of Kekiktuk Conglomerate over pre-Mississippian carbonate rocks (pMl) which have a thin cover of Kekiktuk Conglomerate and Kayak Shale (Figure 3.7 A, location A; Figure 3.8). The pre-Mississippian carbonate rocks in the footwall of this demonstrably post-Mississippian, most likely Cenozoic, thrust fault display strong brittle deformation in the form of numerous D2 anastomosing shear fractures. What bedding has been preserved in the carbonate rocks (pMl) was totally reoriented (Figure 3.12 G). The overlying Kekiktuk Conglomerate in the footwall displays a variety of D2 fabrics including spaced cleavage and open folds. The pre-Mississippian red sandstones (pMrs) of the hanging wall do not have as well-developed D2 fabrics as those seen in the carbonates of the footwall. The overlying Kekiktuk Conglomerate does display some D2 structures, including minor intraformational folding, extension fractures and stretched pebbles.

This same location suggests that north-directed D1 thrust faults may have been reactivated during D2 deformation. Transport on the D2, Cenozoic-age, thrust fault has been relatively small (Figure 3.7 B), implying that the red sandstones (pMrs) were already structurally juxtaposed with the carbonate rocks (pMl) by the time the Mississippian Kekiktuk Conglomerate was deposited on both units. The generally south-dipping and upright bedding with south-dipping cleavage in both pre-Mississippian units suggests that this D1 deformation was north-vergent. D2 thrusting appears to have reactivated this D1 structure, with the red sandstones (pMrs) structurally elevated with respect to carbonate rocks (pMl) for at least 8 km (5 miles) along strike (Figure 3.7 A).

A second example of post-Mississippian deformation of pre-Mississippian rocks is in the southern portion of the study area, on the northern margin of the Mt. Greenough anticlinorium (Figure 3.7 A, location B and figure 3.7 B). In addition to the Whale Mountain thrust fault, which places pre-Mississippian sedimentary rocks (pMcs) over Carboniferous Lisburne Group, a small duplex in the footwall of the main fault consists of small horses of the chert component of the calcareous siliciclastic unit (pMcs), each horse having a thin cover of Kekiktuk Conglomerate. The thrust faults bounding these horses appear to flatten into a roof thrust in the Mississippian Kayak Shale. D2 mesoscopic structures within the Kekiktuk Conglomerate associated with these horses are limited to minor fracturing and occasional shearing, in contrast to the D1 isoclinal folds in the underlying pre-Mississippian cherts and sandstones.

3.7 Structure--Discussion/Interpretation

Two distinct generations of structures in the pre-Mississippian rocks indicate that there have been at least two deformational events in the Aichilik and Egaksrak Rivers area. D1 resulted in a penetrative cleavage and isoclinal folds within the shaly lithologies. These fabrics and mesoscopic folds share the same east-west trend as the map-scale folds and thrust faults that affect the majority of the pre-Mississippian rocks, suggesting that these larger structures also formed initially during D1 deformation.

In contrast, D2 was a far less intense, more brittle and more localized event than D1. D2 fabrics, although scattered throughout the core of the anticlinorium, are concentrated and best developed in areas where deformation appears to post-date the sub-Mississippian unconformity. In some of these locations, the Kekiktuk Conglomerate has remained attached to the underlying pre-Mississippian rocks and clearly defines the style and location of post-Mississippian thrust faults. In the footwall of these thrust faults, post-Mississippian deformation has totally disrupted and reoriented bedding in the pre-Mississippian rocks and D2 structures are well-developed. However, in most of these areas, D2 structures overprint more penetrative D1 structures, indicating that the total amount of shortening in the pre-Mississippian rocks far exceeds that which can be accounted for by the D2 deformational event alone.

The east-northeast trends of the D2 mesoscopic structures also suggest that these structures developed during Cenozoic formation of the Aichilik River anticlinorium. The Cenozoic structural trend, as indicated by the attitudes of bedding, mesoscopic structures and map-scale structures within the overlying Ellesmerian sequence, also strikes east-northeast in the map area, suggesting a north-northwest transport direction (Figures 3.2, 3.7 A, 3.9 and 3.10 B). A north-northwest transport direction during Cenozoic deformation is also suggested by minor tear and thrust faults along Leffingwell Ridge (Figure 3.7 A).

However, the vast majority of the pre-Mississippian rocks in the study area show little evidence of D2 structures or the associated east-northeast trends. To the contrary, D1 mesoscopic structures in the pre-Mississippian rocks generally display east-west trends (Figure 3.12 A-D). These D1 mesoscopic structures in conjunction with the observed D1 map-scale structures suggest that the pre-Mississippian rocks were deformed by north-directed thrusting and related folding during pre-Mississippian time. The main Cenozoic effect, if any, on these early map-scale structures may have been reactivation of faults and tightening of some of the folds, as seen at location A, figure 3.7 A. Regional east-west trending highs and lows within the pre-Mississippian cores of the anticlinoria (Figure 3.10 A) also suggest that east-west trending D1 map-scale structures may have controlled the location of large horses during Cenozoic D2 deformation. In the pre-Mississippian rocks, D2 east-northeast trends are generally observed at the mesoscopic scale, as are the D2 structures themselves.

These observations and interpretations suggest that Cenozoic mesoscopic structures are not pervasive in the pre-Mississippian sequence of the Aichilik River anticlinorium.

Cenozoic shortening in this area therefore was not accommodated penetratively, as suggested by Vann and others (1986) and Oldow and others (1987a). To the contrary, D2 mesoscopic structures are localized along probable Cenozoic thrust faults, implying that most of the Cenozoic shortening within the pre-Mississippian rocks was via thrust duplication.

3.8 Construction of a balanced cross section and its implications

A balanced cross section across northeastern ANWR can provide additional constraints on more regional aspects of the Cenozoic structure, such as the depth to and geometry of the orogenic sole fault, as well as the amount of tectonic shortening. If a reasonable range of Cenozoic tectonic shortening for this area can be established, it may help in evaluating whether Cenozoic shortening was accommodated in the pre-Mississippian rocks primarily by thrust duplication or also by penetrative structures.

Although the amount of published detailed geologic information about the northeastern Brooks Range has grown considerably in the past few years (e.g., Bird and Magoon, 1987; Robinson and others, 1989; Wallace and Hanks, 1990), the detailed structural data required for a well-constrained balanced cross section are still somewhat sparse. Seismic data are publicly available for the coastal plain north of the range front of northeastern ANWR (Bruns and others, 1987), but consist of only a few widely separated lines that do not extend into the mountains. In addition, the data as presented in these lines are of poor quality and reproduced at a very small scale. Other than this study, detailed mapping and structural analysis have not been published for the exposed portions of the Aichilik River transect, although work is in progress in the southern part of the transect (e.g., Wallace and others, 1988; Anderson and Wallace, 1990; Homza and Wallace, 1991). Regional mapping at 1:250,000 is the only other published geologic information (Reiser and others, 1980) for this area.

Because of the general lack of detailed information from northeastern ANWR and the resulting wide range in the possible structural interpretations, a single balanced cross section is difficult to construct with confidence. In addition, detailed balanced cross sections are time-consuming to construct, and it would be not be time-effective to explore all the different possible structural geometries of northeastern ANWR using that method alone. However, the available regional data, sparse as they are, do provide constraints for the construction of a series of simple area-balanced models across the entire region. These regional models can then be used to explore the range of gross structural geometries and resulting tectonic shortening that is possible in northeastern ANWR given the known constraints. Since each simplified cross section is essentially a wedge model of the fold-and-thrust belt, a range of different wedge geometries can be evaluated relatively quickly. The model that best fits the available detailed and regional information can then be used as a basis for constructing a more detailed balanced cross section.

3.8.1 Area-balanced models of northeastern ANWR--the constraints

Seventeen generalized wedge models were constructed across northeastern ANWR, from the coastal plain in the north, through the study area in the Aichilik River anticlinorium to the Continental Divide Thrust Front south of Bathtub Ridge (A-A', figure 3.2; Appendix B). Four representative models are illustrated in figure 3.13. Each model illustrates a different possible structural geometry, based on varying certain regional constraints as discussed below.

The sub-Mississippian unconformity surface was used as a regional datum to define the upper surface of each wedge. This surface was drawn based on publicly available regional surface geologic data and on the published and interpreted seismic reflection data from the coastal plain to the north (Figure 3.14; Reiser and others, 1980; Bader and Bird, 1986; Bruns and others, 1987; Wallace and Hanks, 1990). Since this surface is perhaps the best constrained element of the fold-and-thrust belt, it remains constant in the different models.

Based on the published seismic reflection data (Figure 3.14), a pin line was chosen in the subsurface north of the Leffingwell Ridge, in the vicinity of the Niguanak High.

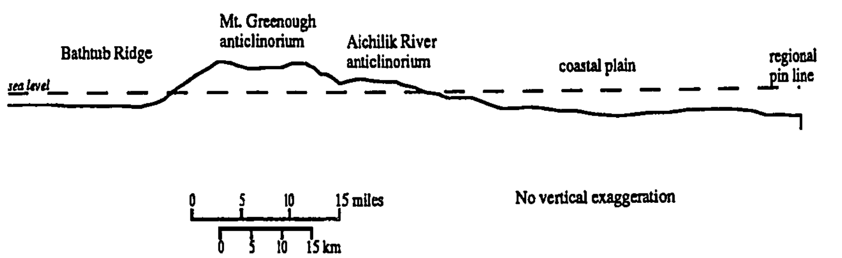


Figure 3.13. Examples of generalized area-balanced structural models for the Aichilik River transect (A-A', Figure 3.2). The structural topography of the upper surface of the wedge (the sub-Mississippian unconformity), the depth to the basal detachment surface at the pin-line, and the location of the pin-line are held constant for all of the models and are illustrated in the first figure. The dip of the basal detachment surface (with respect to the unconformity surface), the depth to the brittle/ductile transition, and the presence of ramps in the basal detachment surface vary from model to model.

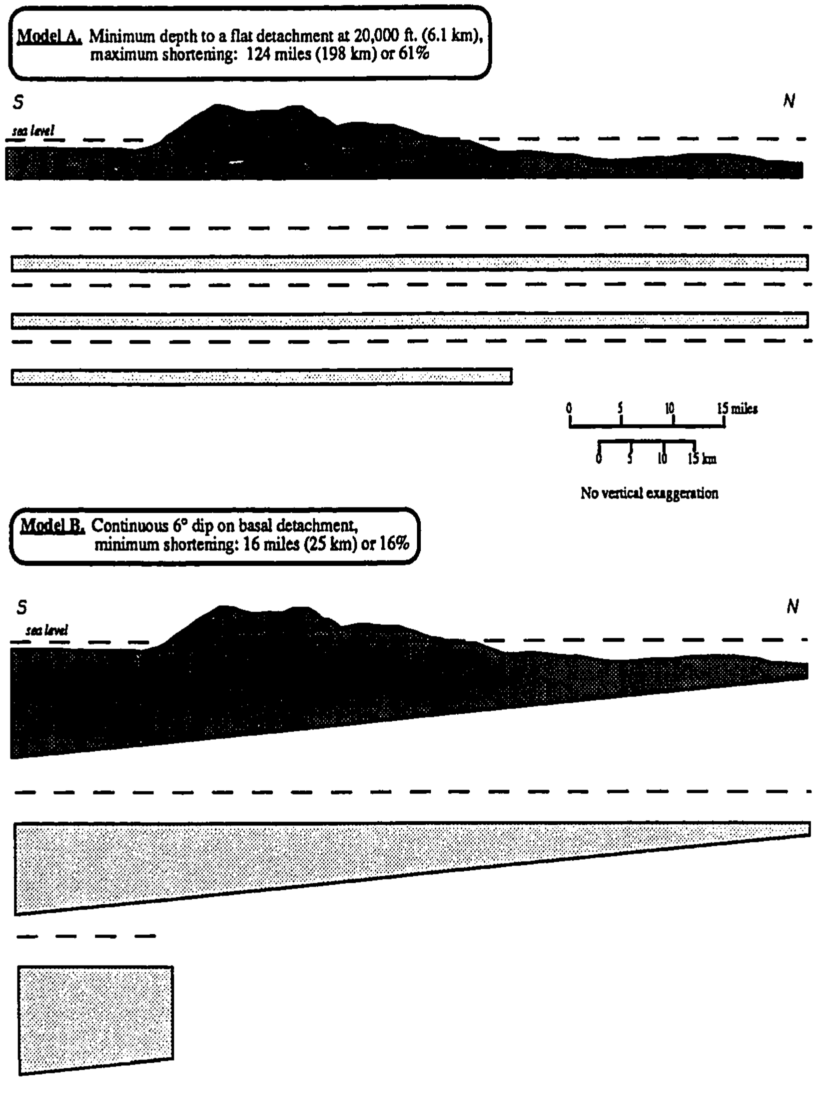


Figure 3.13. (cont)

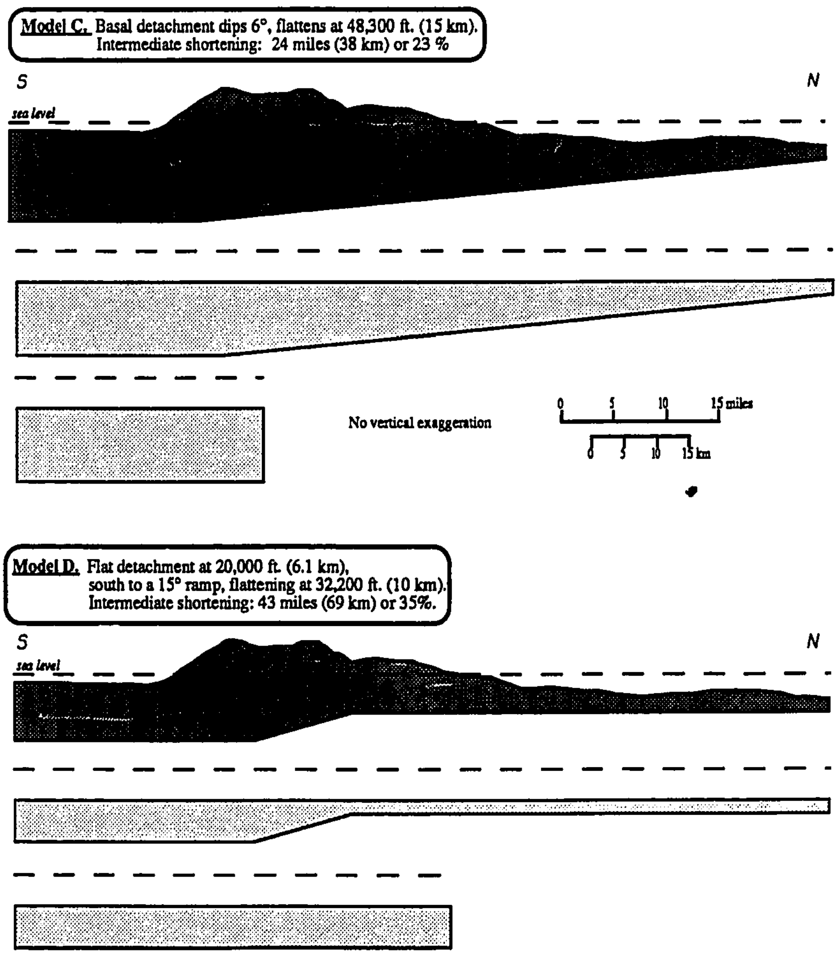
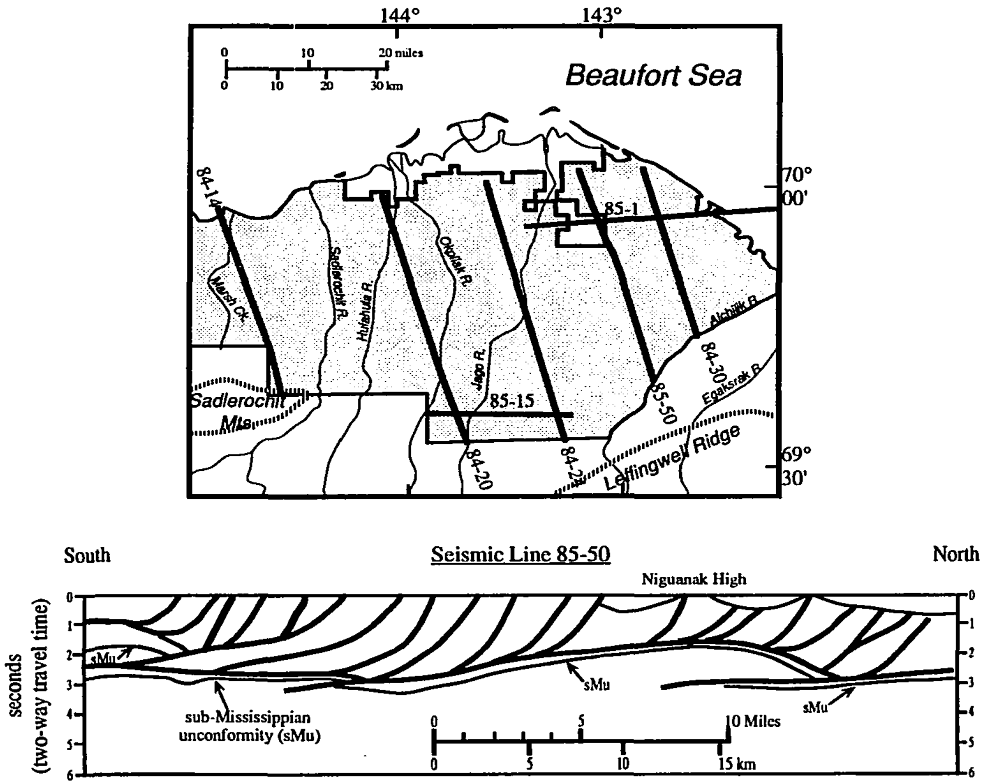


Figure 3.13. (cont)

Figure 3.14. Seismic line 85-50 that parallels the northern portion of the Aichilik River transect (A-A', Figure 3.2). Interpretation by Bruns and others (1987).


144°
143°
Beaufort Sea
0 10 20 miles
0 10 20 30 km
70° 00'
69° 30'
84-14
85-1
85-15
85-50
84-30
84-20
84-2
Sadlerochit Mts
Leffingwell Ridge
Marsh Ck.
Sadlerochit R.
Hulahula R.
Okpilak R.
Jago R.
Aichilik R.
Egaksrak R.
South
Seismic Line 85-50
North
Niguanak High
seconds (two-way travel time)
sMu
sub-Mississippian unconformity (sMu)
0 5 10 Miles
0 5 10 15 km

Cenozoic deformation has undoubtedly continued north of this point, but seismic data are discontinuous and not available to the north. The pin line was therefore placed in the northernmost structural low, where structural thickening is at a minimum, and the models balanced from that point southward. Until more and better seismic data become available, this is the farthest north point suitable for a pin line, and therefore remains fixed in all the models.

The depth to the basal detachment surface at the pin line was also inferred using the published seismic reflection data (Figure 3.14), based on the maximum structural relief on the mapped sub-Mississippian unconformity surface adjacent to the pin line. Two-way travel time to the surface interpreted by Bruns and others (1987) to be the sub-Mississippian unconformity was converted to depth using their time/depth conversion chart. Assuming that the structural relief observed on the unconformity surface is due to structural duplication in a fault-bend fold-style structure (Suppe, 1983), the detachment surface can be inferred to underlie the unconformity at the pin line by a depth equal to that amount of structural relief. This gives a minimum depth to the basal detachment surface under the coastal plain of 20,000 feet (6.1 km). (A deeper detachment depth is possible using other fault-bend fold-style geometries, but with less slip on the faults.) Because the detachment depth at the pin line is partially constrained by the available subsurface data, it too remains fixed in each of the models.

Observations on the slope of basal detachment horizons in other fold-and-thrust belts suggest that basal detachment surfaces can dip up to 6° under an actively growing wedge (Davis and others, 1983). Due to the lack of resolution in the seismic data, the slope of the basal detachment surface south of the pin line is not constrained beneath the coastal plain. Since there are no seismic data available from the range itself, the depth and slope of the basal detachment surface south of the range front are also unknown. In the area-balanced models, the dip of this surface with respect to the sub-Mississippian unconformity surface from the pin line was varied over a range of values from 0° to 6° (e.g., models A and B, Figure 3.13).

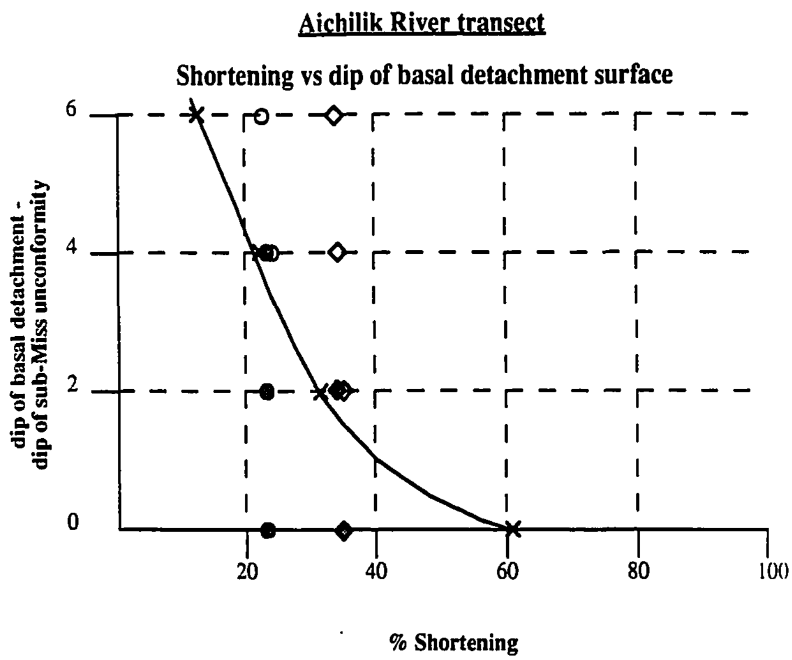
It seems likely that the basal detachment horizon would not maintain a constant dip indefinitely. Evidence provided by deep earthquakes in continental areas suggests that brittle behavior is limited to the upper crust, with ductile deformation mechanisms predominating in the aseismic lower crust (Chen and Molnar, 1983). The depth to this brittle/ductile transition in continental regions generally is thought to vary from 10 to 50 km (Bott, 1982; Suppe, 1985), with the exact depth of the transition dependent on the age and composition of the crust, the geothermal gradient and the fluid pressure. Young, hot crust will have a shallower depth to the brittle/ductile transition (<20 km) than will old, cold crust (~25 km or greater) (Chen and Molnar, 1983; Suppe, 1985). This transition from brittle to ductile deformation could provide a zone of low strength in the lower crust that could facilitate the development of a detachment surface and associated crustal nappes, or 'crustal flakes,' during compressional deformation (Chen and Molnar, 1983; Dewey and others, 1986). Such a zone would be, in effect, a crustal-scale detachment horizon, permitting high-angle brittle faults in the overlying brittle upper crust to flatten at some depth in the more ductile and weaker lower crust.

There is no direct evidence as to the depth of the brittle/ductile transition in northeastern ANWR. It must be at least as deep as the inferred depth of the basal detachment surface at the pin line in the foreland, or 6.1 km (20,000 ft). The depth to the brittle/ductile transition, and thus where the basal detachment horizon would theoretically flatten, was varied in the area-balanced models (e.g., model C and D, Figure 3.13), However, the brittle/ductile transition has not been placed any deeper than 15 km because both conodont alterations indices and apatite fission track studies (O'Sullivan, 1988; Watts and Harris, pers. communication) suggest that the geothermal gradient at the time of thrusting in the northeastern Brooks Range was probably greater than 20°C/km. A geothermal gradient of this value would result in a shallow depth to the brittle/ductile transition (Suppe, 1985).

The geometry of the basal detachment surface was also modified in several of the models by the incorporation of relatively steep ramps (e.g., model D, Figure 3.13). These ramps were located south of the abrupt change in structural relief of the unconformity surface, on the assumption that this change in structural relief would be due to the hanging wall ramp. The exact location of the footwall ramp within the pre-Mississippian rocks south of this point, while arbitrary, was consistent in all models. A dip of 30° was used for all ramps, based on the maximum theoretical dip for thrust ramps in fault bend folds (Suppe, 1985). Ramps connected the upper detachment surface (that either was flat or dipped gently south of the pin line) to a lower flat basal detachment surface. The depth to this lower detachment surface was varied from 7 to 15 km.

Figure 3.13 illustrates four representative models that were constructed using the available regional surface and subsurface data, with the slope of the basal detachment surface, the presence or absence of ramp, and the depth to the brittle/ductile transition as variables. The complete range of variables used in all 14 models is illustrated on figure 3.15. Figure 3.15 also graphically illustrates the effect of these variables on the amount of shortening obtained by area-balanced retrodeformation of these generalized cross sections. It should be noted that these cross sections are area-balanced, but not line-balanced. The line length of the upper surface of the wedge (in this case the sub-Mississippian unconformity) has not been held constant, but has been allowed to lengthen as required by the variables incorporated in that model. In effect, this allows for shortening of the sub-Mississippian unconformity by various mechanisms, including pervasive strain and/or thrust duplication.

These simple area-balanced models illustrate the five regional factors that control the gross structural geometry and amount of shortening across the Aichilik River transect. These are 1) the structural topography of the sub-Mississippian unconformity along the line of the cross section; 2) the depth to the basal detachment surface at the pin line in the foreland; 3) the dip of the basal detachment surface with respect to the sub-Mississippian unconformity; 4) the presence of ramps in the basal detachment surface, and 5) the depth to the brittle/ductile transition, where the basal detachment surface would be expected to flatten. Based on reasonable variations in just the last three factors, the regional shortening along the Aichilik River transect ranges from 16-61%, corresponding to a total shortening for the entire transect of 25 km (16 miles) to 198 km (124 miles). Shortening within the Aichilik River anticlinorium alone could range from 17 km (10 miles) to 30 km (18 miles).



✕ Constant slope on basal detachment

◇ Constant slope on basal detachment, flattening at brittle/ductile transition at 10 km.

○ Constant slope on basal detachment, flattening at brittle/ductile transition at 15 km.

◆ Constant slope on basal detachment, south to a 15° ramp, flattening at brittle/ductile transition at 10 km.

● Constant slope on basal detachment, south to a 15° ramp, flattening at brittle/ductile transition at 15 km.

Figure 3.15. Graph of the dip of the basal detachment horizon vs. the calculated shortening, Aichilik River transect. This graph illustrates the range of possible shortening along the transect as determined by the area-balanced models.

3.8.2 A balanced cross section of the Aichilik River transect

Which of the generalized area-balanced models (Figure 3.13 and Appendix B) best fits the available data on the Aichilik River transect? The published seismic data from the coastal plain are not good enough to resolve the dip of the basal detachment south of the pin line, and no seismic data are available from the range itself. Thus it is necessary to look for other means of constraining the geometry of the basal detachment surface.

From both the structure contour map of the sub-Mississippian unconformity, and from the topography of this surface in the models (Figures 3.10 A and 3.13), we can see that the sub-Mississippian unconformity maintains a relatively constant structural elevation over large portions of the transect. Based on the structure contour map (Figure 3.10 A), the transect can be further divided into three parts: the subsurface north of the range front, the Aichilik River anticlinorium and the Mt. Greenough anticlinorium. Each segment displays a relatively constant structural relief that is significantly higher than that of the segment immediately to the north. The unconformity surface of the Aichilik River anticlinorium is elevated approximately 14,000 feet (4.3 km) with respect to the unconformity surface under the coastal plain to the north. The contrast in elevations of the unconformity between the Mt. Greenough anticlinorium and the Aichilik River anticlinorium is not nearly as dramatic, with the unconformity of the Mt. Greenough anticlinorium consistently 2000-3000 feet (650-1000 meters) higher than that of the Aichilik River anticlinorium to the north. The abrupt changes in structural relief from one segment to the next could reflect major hanging wall ramps above the basal detachment surface, with the corresponding footwall ramps in the pre-Missippian rocks located further to the south. The relatively flat sub-Mississippian unconformity surface within each segment could reflect a relatively flat basal detachment horizon between the ramps.

However, the absolute change in the structural relief related to hanging wall ramps along the transect is probably not great. The greatest change in structural relief is from the coastal plain to the Aichilik River anticlinorium (14,000 feet or 4.3 km), but part of this structural relief is probably due to structural thickening within the pre-Mississippian rocks of the anticlinorium. In addition, the metamorphic grades across the entire transect are relatively low, generally no higher than lower greenschist facies, even in the core of the Mt. Greenough anticlinorium. Conodont alteration indices do vary along the entire transect. CAI's range in value from 3 to 4.5 along Leffingwell Ridge and at Bathtub Ridge (corresponding to temperatures of 145-250°C) to highs of 5 to 6 on the southern flank of the Aichilik River anticlinorium (corresponding to temperatures >300°C) (K. Watts, 1991). This variability in the CAI's and the location of the Lisburne Limestone immediately above the Kayak Shale roof thrust suggests that the CAI's may be partially reflecting hydrothermal activity, and therefore not necessarily a reliable indicator of the depth to the basal detachment. However, the low metamorphic grade and CAIs do suggest that neither the pre-Mississippian rocks nor the Ellesmerian sequence were buried to any great depth, and that rocks from increasingly deeper depths are not exposed as we go south along the transect. These observations imply that, although there are ramps in the basal detachment surface, the change in structural level across each ramp and the absolute change in the depth to the basal detachment from the pin line in the north to the southern end of the transect are probably not great.

Using these observations and assumptions, model D (Figure 3.13) may be the most reasonable model for the transect given the present state of knowledge. This model incorporates a relatively flat basal detachment extending south from the pin line in the coastal plain with a footwall ramp of relatively low structural relief. Models incorporating variations on this geometry (varying ramp heights and varying depths to the lower detachment horizon) suggest that the total amount of shortening across the transect is relatively high, i.e. >35% (Figure 3.14).

Given this general model for the geometry of the fold-and-thrust belt, what mechanism of Cenozoic shortening within the pre-Mississippian rocks should be used in the detailed balanced cross section? There are several lines of evidence that suggest that most of the Cenozoic shortening within the pre-Mississippian sequence of northeastern ANWR was by thrust duplication and not development of penetrative strain and mesoscopic structures. The stepwise changes in structural relief of the sub-Mississippian unconformity surface in the subsurface of the coastal plain and the flat crests and long gentle backlimbs of folds in that surface are both most easily interpreted as large folds related to thrusts with a ramp/flat geometry (Figure 3.13 and 3.14). Such folds would require the pre-Mississippian rocks to shorten primarily by thrust duplication, with little internal shortening and penetrative strain. In the range to the southwest along the Canning River, the geometry of similar anticlinoria and the limited strain and detachment of the Kekiktuk Conglomerate immediately overlying the sub-Mississippian unconformity also suggest thrust duplication with a fault-bend fold geometry (Namson and Wallace, 1986; Wallace and Hanks, 1990). And finally, the detailed structural observations along the exposed portions of the transect outlined in this paper suggest that the majority of Cenozoic shortening in the pre-Mississippian rocks has been by thrust duplication and not penetrative strain. The observed D2 mesoscopic structures within the Aichilik River anticlinorium are only locally developed and non-penetrative, and could not alone account for >35% shortening.

Since the area-balanced models deal solely with the gross structural geometry of rocks beneath the sub-Mississippian unconformity, the models provide no special insights into the behavior of the Ellesmerian cover sequence during Cenozoic thrusting. However, the same amount of regional shortening accommodated by Cenozoic deformation within the pre-Mississippian sequence must also be accommodated by the overlying Ellesmerian sequence rocks. The mode of Cenozoic shortening above the sub-Mississippian unconformity surface along the Aichilik River transect can be inferred from the structural geometry of the Mississippian through Jurassic rocks exposed in the range and visible in the seismic data from the coastal plain. The large thrust displacement indicated by the Egaksrak River klippe on Leffingwell Ridge suggests that most of the shortening in the Ellesmerian sequence of the Aichilik River anticlinorium was by thrust duplication, with only minor detachment folding. North of Leffingwell Ridge, the available seismic data show no evidence of either fold or thrust detachment of the Ellesmerian sequence from the underlying pre-Mississippian rocks, assuming the Ellesmerian sequence is present (Bruns and others, 1987). Therefore, I have assumed that the Kayak Shale ceased to act as an effective detachment horizon north of Leffingwell Ridge, and the Ellesmerian sequence remained attached to the pre-Mississippian rocks and deformed with them as one structural unit. Thrusts ramping from the regional basal detachment horizon at depth in the pre-Mississippian sequence flattened in the next available detachment horizon up-section, the Jurassic to Lower Cretaceous Kingak Shale. This is consistent with the structural behavior of the Ellesmerian sequence along strike to the west in the Sadlerochit Mountains, where the Kayak Shale is thin or missing (Wallace and Hanks, 1990). In contrast, the structure of the Ellesmerian sequence exposed on the southern flank of the Aichilik River anticlinorium (Figure 3.7) and north of Bathtub Ridge (Reiser and others, 1980; Homza and Wallace, 1991) suggests that both detachment folding and thrust duplication are important in the shortening of the Ellesmerian sequence of the Mt. Greenough anticlinorium and to the south. This change in deformational style corresponds to an increase in thickness of the Kayak Shale.

Based on these observations, the detailed balanced cross section (Figure 3.16) incorporates a basal detachment surface that remains flat at a depth of 20,000 feet (6.1 km) from the pin line in the coastal plain and steps deeper to the south via a series of small ramps to a second flat detachment at 23,000 feet (7.0 km) and a third at 31,000 feet (9.5 km). The footwall ramps are located based on line-length restoration of hanging wall ramps (at the northern margins of both the Aichilik River anticlinorium and Mt. Greenough anticlinorium) south to their corresponding footwall ramps, using the sub-Mississippian unconformity surface as a datum. In this cross section, the pre-Mississippian sequence deforms by thrust-duplication, not penetrative strain. The Cenozoic-age horses in the duplex within the pre-Mississippian sequence were reconstructed using fault-bend fold geometry, with the entire cross section balanced using the sub-Mississippian unconformity as a datum and maintaining line length. As illustrated in the cross section, shortening within the Ellesmerian sequence varies according to location: north of the range front it remains attached to the underlying pre-Mississippian rocks, deforming with them as one structural unit; in the vicinity of the Aichilik River anticlinorium, shortening is primarily by thrust duplication between detachment horizons in the Kayak and Kingak shales; and in the

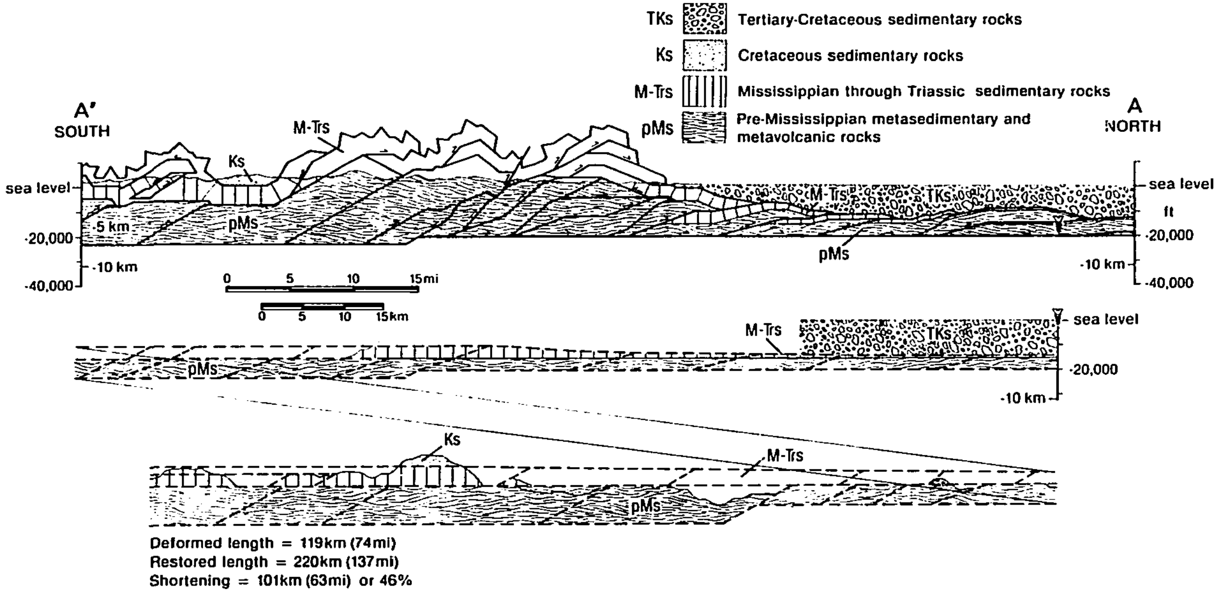


Figure 3.16. Balanced cross section along the Aichilik River transect. Cross section location is shown on figure 3.2 as A-A'.

Mt. Greenough anticlinorium, shortening is by thrust duplication and detachment folding. Restoration of this cross section indicates 46% shortening, or 101 km (63 miles) of total shortening, of an undeformed length of 137 miles (220 km). This amount of shortening does not include the contribution made by mesoscopic Cenozoic structures. Although this contribution is probably minor, it would increase the total amount of shortening.

3.9 Summary and conclusions

North-northwest-directed Cenozoic thrusting in the Aichilik and Egaksrak rivers area of northeastern ANWR resulted in the formation of large anticlinoria cored by pre-Mississippian metasedimentary and metavolcanic rocks, with the limbs defined by the Mississippian to Jurassic Ellesmerian cover sequence. During this thrusting event, a thick shale near the base of the cover sequence, the Mississippian Kayak Shale, acted as a regional detachment horizon, permitting overlying Mississippian through Triassic clastic and carbonate rocks to deform independently of the underlying pre-Mississippian rocks. These cover rocks deformed primarily by thrust duplication, with minor detachment folding.

A basal conglomerate for the cover sequence, the Mississippian Kekiktuk Conglomerate, underlies the Kayak Shale, and remained attached to the pre-Mississippian sequence during Cenozoic deformation. The conglomerate thus helps define the geometry of Cenozoic structures within the pre-Mississippian rocks. These older metasedimentary and metavolcanic rocks retain evidence of an earlier, pre-Mississippian event in the form of D1 penetrative fabrics with a dominantly east-west striking, south-dipping orientation. D2 structures have reactivated some of these D1 structures during Cenozoic time, as is suggested by local D2 imbrication of different pre-Mississippian units that were already juxtaposed at Mississippian time. Although D2 structures are scattered throughout the pre-Mississippian sequence and share share the northeast-southwest, south-dipping orientation of Cenozoic D2 structures in the overlying Ellesmerian sequence, D2 structures within the pre-Mississippian rocks D2 structures are only locally well-developed, generally non-penetrative, and almost always associated with Cenozoic faults. During Cenozoic thrusting, regional and detailed field evidence suggests that the pre-Mississippian sequence deformed primarily via thrust duplication in a regional, northward-migrating duplex, with a roof thrust in the Mississippian Kayak Shale and a floor thrust at depth in the pre-Mississippian sequence, possibly at the brittle-ductile transition.

Based on the observations in this study, there appears to be no compelling evidence that penetrative strain played a major role in the formation of the regional anticlinoria in northeastern ANWR, as suggested by Vann and others (1986) and Oldow and others (1987a). To the contrary, this study suggests that the Cenozoic-age mesoscopic structures within the pre-Mississippian sequence were too scattered and non-penetrative to account for the total amount of regional shortening within the region. The observations do suggest that thrust duplication of the pre-Mississippian sequence during Cenozoic thrusting is sufficient to account for the amount of Cenozoic shortening in the region.

This conclusion is plausible given the nature of the pre-Mississippian sequence in northeastern ANWR. In this part of ANWR, the pre-Mississippian sequence consists of an assemblage of heterogeneous, low-grade metasedimentary and minor metavolcanic rocks with numerous shale horizons. These shales would act as good detachment horizons during deformation, thus favoring formation of faults and related folds. This deformational style contrasts markedly with that seen immediately to the west, where the pre-Mississippian rocks consist of a homogeneous granitic batholith (Sable, 1977; Reiser and others, 1980; Hanks and Wallace, 1990). Here, penetrative structures probably have accommodated much of the Cenozoic shortening within the batholith, probably due to the lack of potential detachment horizons in the mechanically homogeneous granitic rocks. Thus, the structural behavior of the pre-Mississippian sequence is strongly controlled by its lithology and the presence of potential detachment horizons.

The deformational behavior of the pre-Mississippian rocks illustrates the hazards of using the term 'basement' too loosely. Although the pre-Mississippian rocks of the northeastern Brooks Range can be considered 'depositional basement' to the overlying Ellesmerian sequence, for the most part they do not share the lithologic properties of cratonic basement and have thus behaved differently during thrusting. Application of the term 'basement' to the pre-Mississippian rocks of the northeastern Brooks Range may have led to many misconceptions as to the nature of the sequence and its structural behavior.

Why were the pre-Mississippian rocks of the northeastern Brooks Range involved in Cenozoic thrusting immediately adjacent to the foredeep, and not in the central and western Brooks Range? Several factors may have contributed to this unusual circumstance. The sub-Mississippian unconformity may have been shallower in the northeastern Brooks Range than elsewhere in the range, reflecting an Ellesmerian cover sequence that is thinner because it is higher up on the passive continental margin and/or over some other type of structural high. A structural high cored by pre-Mississippian rocks, the Barrow Arch, extends the length of the North Slope (Figure 3.1). This high is far north of the Brooks Range range front along most of its length, but trends east-southeast into the range front region of the northeastern Brooks Range. Thus the Cenozoic fold-and-thrust belt may have prograded up the southern flank of this major high, resulting in the involvement of rocks that are older than those seen elsewhere near the leading edge of the fold-and-thrust belt.

Other factors that could contribute to the involvement of pre-Mississippian sequence in foreland deformation include an anomalously deep orogenic sole fault in the foreland and the non-crystalline, heterogeneous nature of the sequence. However, the suggested depth to the basal detachment horizon in the coastal plain as interpreted from the available seismic data (6.1 km or 20,000 feet) is not unusually deep when compared to the depth to the orogenic sole fault in the foreland of other fold-and-thrust belts (for e.g., ~5 km in the Canadian Rockies, Thompson, 1981; ~7 km in the central and southern Appalachians, Hatcher, 1981). The absolute depth to the basal detachment horizon alone can not, therefore, be called on to explain why pre-Mississippian rocks are involved adjacent to the foredeep.

However, as mentioned earlier, the heterogeneous low-grade metasedimentary rocks of the pre-Mississippian sequence of the northeastern Brooks Range have numerous potential detachment horizons. These horizons of potential structural failure could result in an overall structurally 'weak' package of rocks that would fail at lower differential stresses than homogeneous, higher-grade rocks at the same depth, especially if fluid pressures in the shales were high (Hubbert and Rubey, 1959; Suppe, 1985). This effect could be augmented, of course, by a basal detachment horizon deep enough to encounter these older rocks. The role of other factors that could influence the manner in which the pre-Mississippian rocks deformed, including the geothermal gradient at the time of thrusting, fluid pressure in the pre-Mississippian sequence, and the presence of a major lithologic and/or rheologic change at depth (the brittle/ductile transition or a major compositional change), remains unclear.

Additional constraints on the Cenozoic structural evolution of northeastern ANWR could be provided by addressing poorly understood aspects of the region that affect the structural interpretation. More information regarding the geothermal gradient during Cenozoic deformation and the subsequent uplift history of the region would aid in determining the depth to the orogenic sole fault, as well as clarify the timing of the growth of the various anticlinoria with respect to the evolution of the foreland basin. More detailed stratigraphic and structural studies within the pre-Mississippian sequence elsewhere in the northeastern Brooks Range would help further document the nature of the pre-Mississippian deformational event and the character of the Cenozoic structures throughout the region. Detailed strain and mesoscopic structure studies within the Ellesmerian sequence throughout the region, specifically within the Kekiktuk Conglomerate, the Lisburne Group and the sandstones of the Sadlerochit Group, would provide information as to the amount of shortening accommodated by the development of mesoscopic structures during Cenozoic thrusting. This information could greatly aid in refining the current balanced cross sections.

In conclusion, the Cenozoic fold-and-thrust belt of the northeastern Brooks Range provides a good example of the role of non-crystalline, heterogeneous 'depositional basement' in the structural evolution of a foreland fold-and-thrust belt. These older rocks have deformed primarily by thrusting and related folding, despite their age and previous deformational history. This structural style is due to their lithology and the conditions of deformation: they are essentially low-grade, lithologically heterogeneous stratified rocks deformed at relatively low pressures and temperatures. These rocks were at relatively shallow depths in the foreland of the advancing fold-and-thrust belt, and thus eventually were able to be incorporated in the deformation at the range front.

3.10 Acknowledgments

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CHAPTER 4: THE OKPILAK BATHOLITH TRANSECT[5]

4.1 Abstract.

Involvement of crystalline rocks in thrusting near the foreland basin of a fold-and-thrust belt is relatively uncommon. In the northeastern Brooks Range, the Devonian Okpilak batholith was thrust northward and structurally elevated above adjacent foreland basin deposits during Cenozoic fold-and-thrust deformation. The batholith may have acted initially as a regional structural buttress, but a drop in the basal detachment surface to greater depth south of the batholith resulted in northward transport of the batholith. Shortening within the batholith was accommodated by (1) the development of discrete thrust slices bounded by ductile shear zones, or (2) simple shear and development of penetrative mesoscopic and microscopic fabrics throughout the batholith, or both. The Mississippian Kayak Shale, a regional detachment horizon at the base of the overlying cover sequence, is depositionally thin or absent adjacent to the batholith. Thus, most of the cover sequence remained structurally coupled to the batholith during thrusting and was shortened by the development of penetrative structures.

4.2 Introduction.

Intimate involvement of crystalline rocks in thrust faulting is common in the interior of mountain belts, reflecting structural detachment at deep levels (Boyer and Elliot, 1982; Coward, 1983). Involvement of crystalline rocks tends to decrease toward the foreland as the orogenic sole fault propagates toward the surface and as deformation moves into the foreland basin where basement is flexurally depressed (Price, 1981; Boyer and Elliot, 1982). The Okpilak batholith of the northeastern Brooks Range is unusual because it has

[5]Chapter 4 contains the complete text and figures of the manuscript, *Cenozoic thrust emplacement of a Devonian batholith, northeastern Brooks Range: Involvement of crystalline rocks in a foreland fold-and-thrust belt*, by C. L. Hanks and W. K. Wallace, as published in *Geology, v. 18, no. 5, p. 395-398, 1990.*

been intimately involved in fold-and-thrust deformation adjacent to a foreland basin. The batholith is an isolated crystalline body within a stratified sequence, and thus offers a good opportunity to observe the structural response of a large mass of crystalline rock to foreland fold-and-thrust deformation. The character of deformation within the crystalline rocks and their cover may be compared directly with that of adjacent stratified rocks that display typical fold-and-thrust structures.

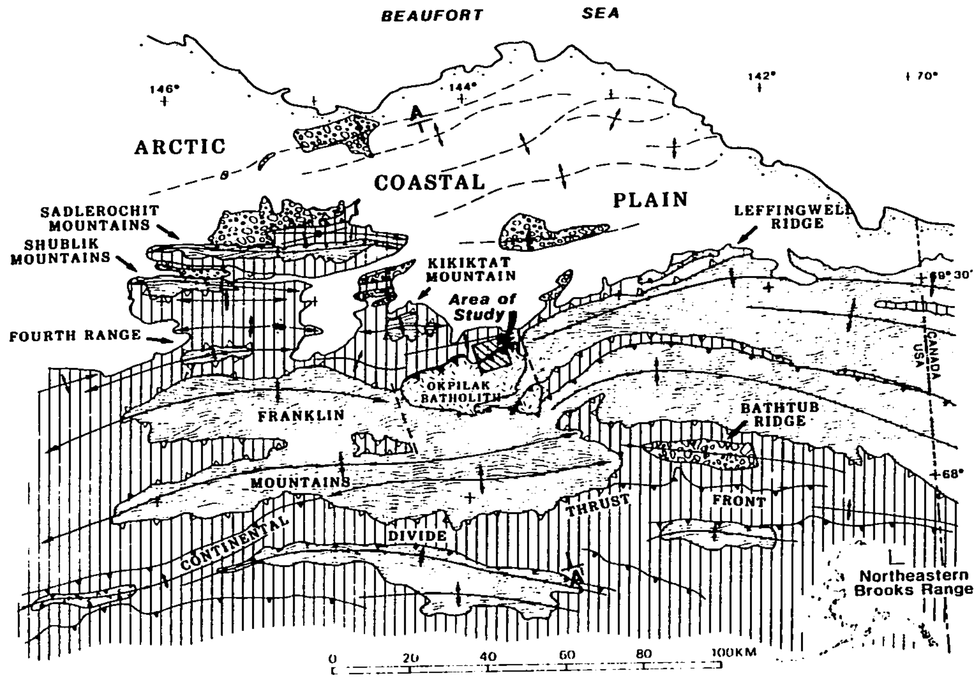
4.3 Regional setting.

The northeastern Brooks Range is a Cenozoic fold-and-thrust belt involving para-autochthonous rocks (Fig. 4.1; Reiser, 1970; Kelley and Foland, 1987; Wallace and Hanks, 1990). The oldest and structurally lowest rocks exposed are heterogeneous, weakly metamorphosed Proterozoic to lower Paleozoic sedimentary and volcanic rocks of the Franklinian sequence (Reiser et al., 1980). These rocks are separated by an angular unconformity from overlying northerly derived Mississippian to Lower Cretaceous carbonate and clastic rocks of the Ellesmerian sequence (Bird and Molenaar, 1987). The Ellesmerian sequence is in turn overlain by the Brookian sequence, made up of Lower Cretaceous and younger sediments that were shed from the rising Brooks Range to the south (Mull, 1985; Molenaar et al., 1987).

All three sequences were involved in the Cenozoic fold-and-thrust deformation that formed the dominant structures of the northeastern Brooks Range. These structures are anticlinoria cored by pre-Mississippian rocks, with Mississippian and younger rocks of the Ellesmerian sequence on the limbs (Bader and Bird, 1986). The anticlinoria are interpreted to reflect horses in a regional-scale duplex, with a floor thrust at depth in the Franklinian sequence and a roof thrust in the Mississippian Kayak Shale, near the base of the Ellesmerian sequence (Namson and Wallace, 1986; Wallace and Hanks, 1990).

The Okpilak batholith and its satellite stocks are the only large intrusions in the northeastern Brooks Range (Sable, 1977; Bader and Bird, 1986). A U-Pb zircon analysis of the granite yielded upper and lower concordia intercepts of 380 ±10 Ma and 61 ±10 Ma,

Figure 4.1. Generalized tectonic map of the northeastern Brooks Range. Solid teeth on thrust faults indicate older-over-younger thrust faults that duplicate stratigraphic section, open teeth indicate detachment surfaces along which there has been slip but no disruption of normal stratigraphic succession. Line A-A' is location of cross section in figure 4.4. Map modified from Wallace and Hanks, 1990.



Lower Cretaceous to lower Tertiary Brookian sequence

Mississippian to Lower Cretaceous Ellesmerian sequence

Pre-Mississippian metasedimentary and metavolcanic rocks of the Franklinian sequence. Pattern includes Mississippian Kekiktuk conglomerate of the Ellesmerian Sequence

Devonian Okpilak batholith

respectively (Dillon et al., 1987). The older age and intrusive and depositional relations indicate that the Okpilak batholith is Devonian and predates the unconformity at the base of the overlying Ellesmerian sequence. Near the batholith, the Ellesmerian sequence includes the Mississippian Kekiktuk Conglomerate and Kayak Shale, the Mississippian to Pennsylvanian Lisburne Group, and the Permian to Triassic Sadlerochit Group. The Ellesmerian sequence has been eroded over most of the batholith, but unconformably overlies the batholith around its margins (Sable, 1977; Dillon et al., 1987). Here, the Kayak Shale is depositionally thin or absent and the Lisburne Limestone directly overlies either the Kekiktuk Conglomerate or the batholith itself, suggesting that the batholith was a topographic high at the time of deposition (Watts et al., 1988).

4.4 Evidence for Cenozoic thrusting of the Okpilak batholith.

4.4.1 Regional Observations.

The batholith is one of the highest parts of the entire Brooks Range, reaching elevations above 2700 m. It is structurally higher than rocks of the Ellesmerian sequence along its northern margin, which in turn are structurally higher than the deformed Cretaceous and Tertiary foreland basin deposits that underlie the coastal plain to the north. Published seismic data and interpretations (Bruns et al., 1987, Pl. 4) indicate that the batholith is approximately 9000 m structurally higher than pre-Mississippian rocks in adjacent parts of the coastal plain. It also is approximately 1800 m above pre-Mississippian rocks to the east. Uplift of the batholith relative to both other pre-Mississippian rocks and the young, deformed foreland basin deposits implies that the batholith was involved in the Cenozoic deformation.

East of the batholith, Leffingwell Ridge marks the north limb of an anticlinorium that formed as a result of Cenozoic thrusting (Fig. 4.1; Reiser et al., 1980; Wallace and Hanks, 1990). This range-front structure continues uninterrupted north of the batholith, requiring that the batholith also be underlain by a thrust fault. Pre-Mississippian rocks are exposed in the core of a thrust-related anticlinorium north of the batholith (Kikiktak Mt., Fig. 4.1; Bader and Bird, 1986), again requiring that a detachment pass through or beneath the batholith.

There is an apparent increase in the intensity of mesoscopic folding and axial planar cleavage in Triassic and older rocks of the Ellesmerian sequence approaching the northern margin of the batholith (Sable, 1977; this study). This indicates that a post-Triassic dynamothermal metamorphic event was spatially associated with the batholith. A 61 ±10 Ma lead-loss event and a 59 ±2 Ma K/Ar cooling age on recrystallized biotite from the granite (Dillon et al., 1987), may reflect the structural emplacement and uplift of the batholith that led to the formation of these structures.

4.4.2 Field Observations.

The northern margin of the batholith is marked by ductile shear zones that dip moderately to the south. In the eastern part of the map area, a complete but northward-overturned stratigraphic sequence is exposed beneath the batholith in the footwall of a shear zone and includes the Kekiktuk Conglomerate, the Lisburne Limestone, and the basal formation of the Sadlerochit Group, the Echooka Formation (Fig. 4.2). Stratigraphic continuity is preserved, but both the granite and the sedimentary rocks display penetrative fabrics. The intensity of these penetrative fabrics decreases both to the north and up stratigraphic section. Penetrative mesoscopic structures include well-developed south-dipping schistosity at a low angle to bedding, north-directed S-C protomylonitic to mylonitic penetrative fabrics, and north-trending stretching lineations (Fig. 4.3). The granitic rocks show microscopic evidence of deformation under lower greenschist facies conditions (Simpson, 1985), with brittle deformation of feldspars, ductile deformation and recrystallization of quartz grains, and recrystallization and kinking of micas. Calcite within the carbonate rocks has been ductilely deformed and recrystallized.

The intensity of the penetrative mesoscopic fabric increases westward where the overturned sequence forms the attenuated south limb of a major overturned syncline (Fig.

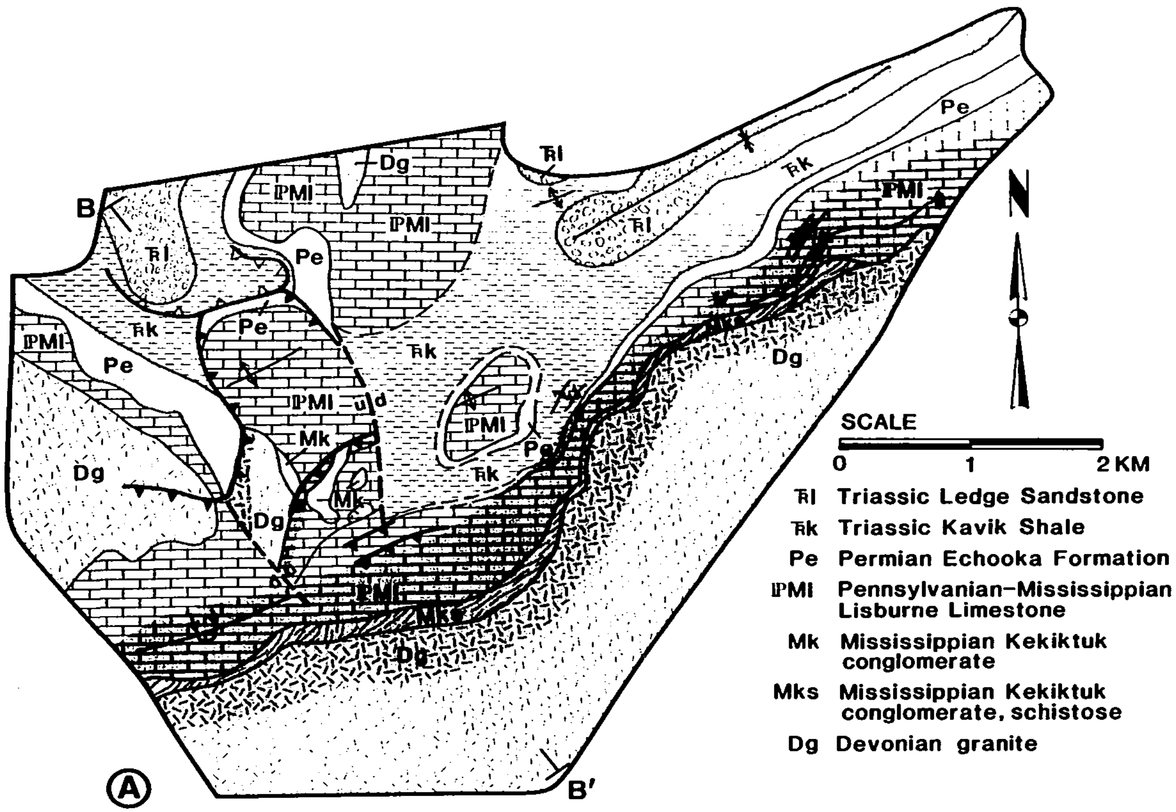


Figure 4.2. (A) Geologic map of the northern margin of Okpilak batholith between McCall Creek and the Okpilak River. Shaded areas: ductile shear zones displaying mylonitic fabrics.

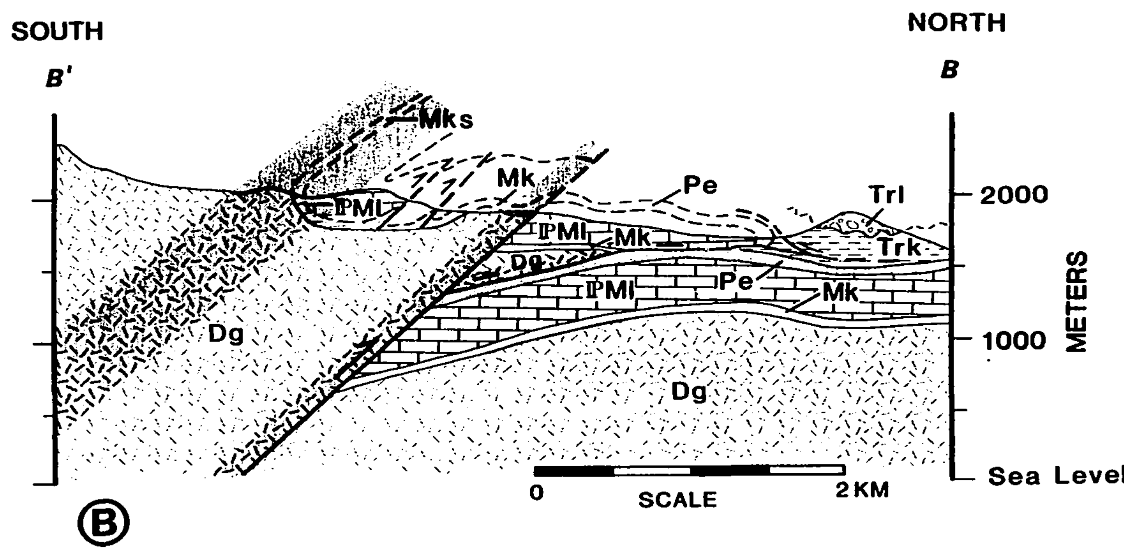


Figure 4.2. (B) Cross section of the northern margin of Okpilak batholith between McCall Creek and the Okpilak River. Shaded areas: ductile shear zones displaying mylonitic fabrics.

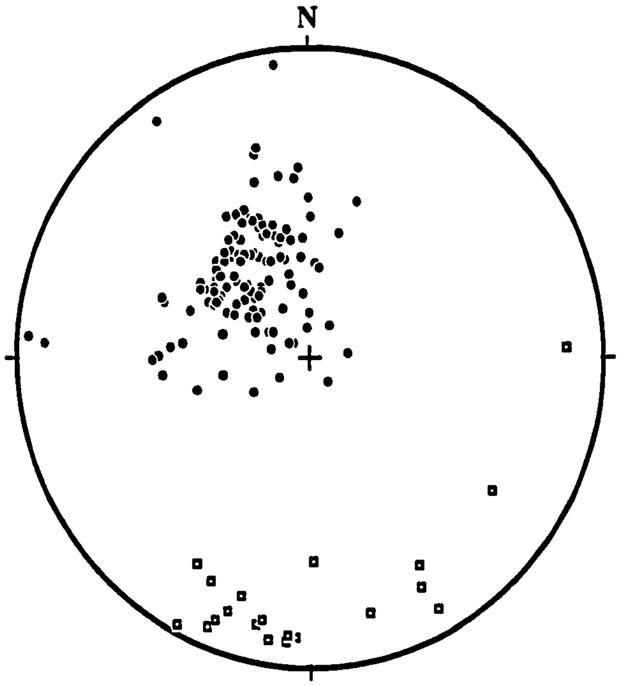


Figure 4.3. Equal area, lower hemisphere projection of poles to schistosity and cleavage within the northern margin of the Okpilak batholith, Kekiktuk Conglomerate, and adjacent Lisburne Limestone (circles) n = 124; stretching lineations in the Lisburne Limestone (squares) n = 22.

4.2). The Lisburne Limestone and Echooka Formation locally form tight overturned folds in the core of the syncline. Where thinning is greatest, bedding appears to have been totally transposed, and north-trending sheath folds have formed in the Lisburne Limestone. In the western part of the study area, at least two more shear zones lie north of the shear zone that marks the north flank of the batholith to the east (Fig. 4.2). One of these places granite over rocks of the Sadlerochit Group.

Pervasive south-dipping foliations and local planar zones of non-coaxial high strain also occur across the width of the batholith south of the study area (Sable, 1977; this study). This suggests that there has been significant shortening within the batholith itself, and that displacement on shear zones is not restricted to the northern margin of the batholith.

4.5 Discussion.

Regional and field data have been used to construct an area-balanced cross section across the batholith that offers clues to the degree and manner of structural involvement of the batholith in Cenozoic fold-and-thrust deformation (Fig. 4.4). Seismic data from north of the batholith (Bruns et al., 1987, Pl. 4, line 84-20) provide constraints on the possible geometry of the pre-Mississippian rocks under the coastal plain. The data indicate that the minimum depth to the basal detachment surface under the coastal plain is approximately 10 km. The abrupt increase in elevation of the batholith relative to pre-Mississippian rocks to the north and east suggests the presence of frontal and lateral ramps. This would require that the batholith was displaced from its original location above a detachment surface at a depth greater than 10 km. The absolute depth of this lower detachment surface cannot be constrained at present, but the geometry of Cenozoic horses in pre-Mississippian rocks south of the batholith suggests a depth of 15 km or more. Restoration of the deformed section suggests a total of 115 km of shortening over 245 km of undeformed length, or 47%.

Cenozoic deformation of the batholith contrasts markedly in style with that of

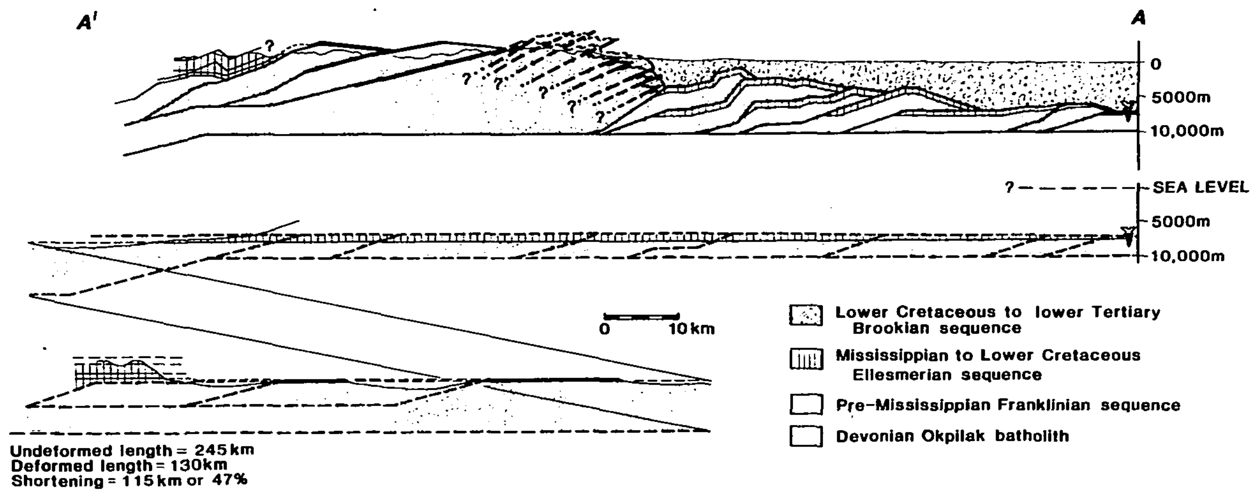


Figure 4.4. Balanced cross section through the Okpilak batholith, along line A-A' in Figure 4.1.

adjacent lithologically heterogeneous and stratified pre-Mississippian rocks. These rocks were deformed in a regional duplex with multiple horses of pre-Mississippian rocks between a floor thrust at depth and a roof thrust in the overlying Mississippian Kayak Shale (Wallace and Hanks, 1990), in a style similar to that north and south of the batholith (Fig. 4.4). The difference in deformational style between the granite and other pre-Mississippian rocks may be due to the homogeneous, competent nature of the batholith and, at least locally, to the lack of a well-defined overlying detachment horizon. The batholith lacks internal incompetent layers or zones that would serve as structural detachment horizons, and at least initially behaved as a homogeneous, relatively rigid body. Consequently, the batholith may have acted initially as a structural buttress, delaying thrusting in the vicinity of the batholith, as suggested by the northward-concave arcuate trend of structures around its southern margin (Fig. 4.1; Bader and Bird, 1986). A subsequent drop in the basal detachment surface may have allowed eventual northward displacement of the batholith. As this detachment propagated northward, the batholith may have deformed either as a series of discrete fault slices bounded by shear zones or by homogeneous simple shear at the mesoscopic and microscopic level. Our preliminary observations suggest a combination of the two.

Because of the lack of a detachment horizon in the Kayak Shale, the batholith and its cover were coupled, and the cover was shortened largely by development of penetrative fabrics. This mode of shortening contrasts markedly with the detachment folding or thrust faulting of the Lisburne Group and overlying rocks that occurs above the Kayak Shale where it is thick and continuous (Wallace and Hanks, 1990).

The Okpilak batholith is a good example of the way an isolated body of crystalline rocks may deform in a foreland fold-and-thrust belt. This study suggests that (1) a drop in the basal orogenic detachment surface may be needed to allow deformation of the structurally competent, homogeneous crystalline body; (2) onset of deformation may be delayed within the crystalline mass, resulting in deflection of structural trends around its margins; (3) deformation within the crystalline body may occur on ductile shear zones bounding discrete fault slices, by homogeneous simple shear at the mesoscopic and microscopic scale, or both; and (4) an isolated crystalline body may be a topographic high during deposition of overlying sediments, thus controlling the location of potential detachment horizons (e.g., shale or salt.). This in turn may control the eventual deformational style of the cover sequence over the crystalline body.

4.6 Acknowledgements.

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CHAPTER 5: RAPID EVALUATION OF THE REGIONAL GEOMETRY AND SHORTENING OF A FOLD-AND-THRUST BELT[6]

5.1 Abstract

Simple area-balanced models of a portion of the northeastern Brooks Range illustrate five regional constraints on the gross geometry of a foreland fold-and-thrust belt. These factors include: the structural topography of the upper surface of the orogenic wedge, the depth to the basal detachment surface at the pin-line, the dip of the basal detachment surface, the presence of major ramps in the basal detachment surface, and the depth at which the basal detachment horizon flattens (if it does). Area-balanced models incorporating variations in these factors can be quickly constructed and illustrate a range in the possible regional geometry and tectonic shortening for the fold-and-thrust belt. This method provides a rapid means of evaluating a variety of different geometries of the orogenic wedge before developing a more detailed balanced cross section. In the northeastern Brooks Range, the models alone suggest that the absolute range in possible shortening is from 17-57%, corresponding to a total tectonic shortening for the region of 31 km (19 miles) to 203 km (126 miles).

5.2 Introduction

Construction of balanced cross sections has become a standard technique used in illustrating and studying fold-and-thrust belts worldwide, and is useful at all scales of structural analysis, from regional tectonic synthesis to strain analysis (e.g., Dahlstrom, 1969; Price, 1981; Boyer and Elliot, 1982; Woodward and others, 1986). Such sections are widely used in petroleum exploration where well and seismic data can be utilized to

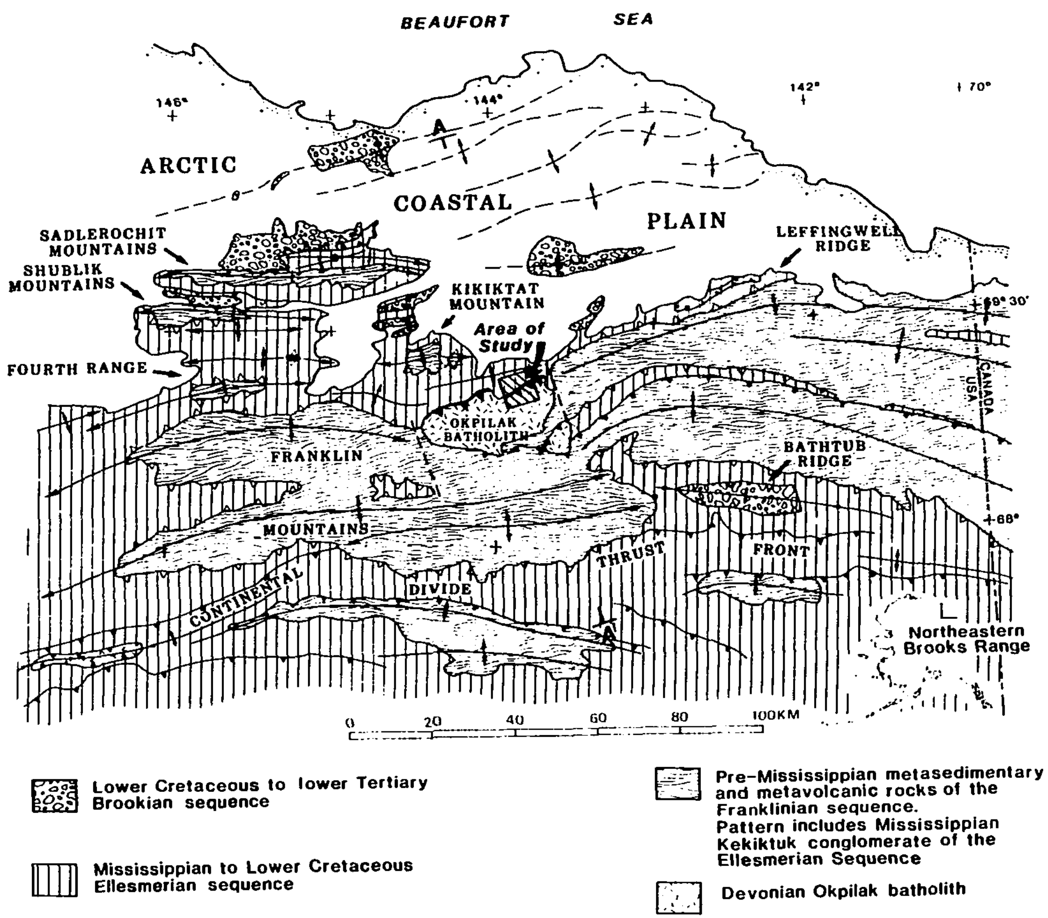
[6] Chapter 5 contains the complete text and figures of the manuscript, *Rapid evaluation of the regional geometry and shortening of a fold-and-thrust belt: an example from the northeastern Brooks Range, Alaska*, by C. L. Hanks. This manuscript has been submitted to the American Association of Petroleum Geologists *Bulletin*.

constrain the geometry of surface and subsurface structures and hence identify future areas of research and/or exploration. However, not all fold-and-thrust belts readily lend themselves to construction of well-constrained balanced sections, generally due to a lack of seismic, well, and/or surface data. These fold-and-thrust belts are generally in remote areas where both academic research and industry exploration are hampered by high costs and/or politically or geographically hostile environments.

The northeastern Brooks Range is one such area (Figure 5.1). The foreland basin immediately north of this portion of the Brooks Range is a focus of much industry interest as it is an unexplored region of high potential that lies only 160 km (100 miles) east of the largest oil field in North America, Prudhoe Bay. It is also of considerable academic interest for a variety of reasons, including its key position in many models for the opening of the Arctic Ocean basin (see Lawver and Scotese (1990) for review). However, most of the northeastern Brooks Range is mapped at only a reconnaissance level (Bader and Bird, 1986; Reiser and others, 1971 and 1980). Reflection seismic data are publicly available only for a portion of the foreland basin, are of fairly poor quality, and do not continue into the range. Construction of detailed balanced cross sections across the region would help address a variety of structural questions, including how the geometry of surface structures in the mountains compare with those in the subsurface to the north, and how much shortening has occurred across the entire fold-and-thrust belt. Both of these questions are important from both a regional and petroleum exploration perspective, but the lack of good seismic data and detailed surface information places severe limits on the construction of well-constrained, detailed balanced cross sections.

Important information can be gained despite a lack of detailed data by the construction of a series of generalized area-balanced cross sections that model the fold-and-thrust belt as a simple, internally homogeneous wedge. These models can be constructed relatively quickly using a minimum of regional surface and subsurface data and an inexpensive computer drafting program. The resulting area-balanced models serve to illustrate and clarify the general constraints on the gross geometry of the fold-and-thrust belt and the

Figure 5.1. Generalized tectonic map of northeastern Brooks Range. Solid teeth on thrust faults indicate older-over-younger thrust faults that duplicate stratigraphic section, open teeth indicate detachment surfaces along which there has been slip but no disruption of normal stratigraphic succession. Line A-A' is location of cross section in figure 5.4. Map modified from Wallace and Hanks, 1990.



BEAUFORT SEA
146°
144°
142°
70°
ARCTIC
COASTAL
PLAIN
A
SADLEROCHIT MOUNTAINS
SHUBLIK MOUNTAINS
FOURTH RANGE
KIKIKTAT MOUNTAIN
Area of Study
OKPILAK BATHOLITH
LEFFINGWELL RIDGE
69° 30'
CANADA
USA
BATHTUB RIDGE
FRANKLIN
MOUNTAINS
68°
CONTINENTAL
DIVIDE
THRUST
FRONT
A'
Northeastern Brooks Range
0
20
40
60
80
100KM
Lower Cretaceous to lower Tertiary Brookian sequence
Mississippian to Lower Cretaceous Ellesmerian sequence
Pre-Mississippian metasedimentary and metavolcanic rocks of the Franklinian sequence. Pattern includes Mississippian Kekiktuk conglomerate of the Ellesmerian Sequence
Devonian Okpilak batholith

possible range of regional shortening across the region, and help to define areas of future research. Construction of these generalized cross sections also allows the rapid comparison of a variety of different interpretations without the large investment of time required to construct the equivalent detailed cross sections. The model that best fits both the regional data and the available surface and subsurface geologic data can then serve as a basis for construction of a more detailed balanced cross section.

5.3 Regional geologic setting

The northeastern Brooks Range is a Cenozoic fold-and-thrust belt (Reiser, 1970; Kelley and Foland, 1987; Wallace and Hanks, 1990) involving parautochthonous rocks equivalent to those of the North Slope subsurface (Figure 5.1; Reiser, 1970). The structurally lowest and oldest exposed rocks are heterogeneous and weakly metamorphosed Proterozoic to lower Paleozoic sedimentary and volcanic rocks locally intruded by Devonian granitic bodies (Reiser and others, 1980). The stratigraphy and age of these rocks are poorly constrained, due to a complex deformational history and a general lack of fossil control. These rocks were deformed during both pre-Mississippian and Cenozoic deformation, with the pre-Mississippian deformation probably including both fold-and-thrust style and penetrative ductile deformation, depending upon the lithology and structural position of the pre-Mississippian rocks (e.g., Oldow and others, 1987a; Lane and Cecile, 1989; Hanks and Wallace, 1990). These rocks display a metamorphic overprint that ranges up to lower greenschist facies. This metamorphism is generally of pre-Mississippian age, although a second, younger overprint is locally present.

The pre-Mississippian rocks are truncated by a regionally extensive sub-Mississippian angular unconformity and overlain by northerly derived Mississippian to Lower Cretaceous carbonate and clastic rocks of the Ellesmerian sequence (Bird and Molenaar, 1987). The Ellesmerian sequence is in turn overlain by the Brookian sequence, which consists of Lower Cretaceous and younger sedimentary rocks that were shed from the rising Brooks Range to the south (Mull, 1985; Molenaar and others, 1987). The Brookian sequence is only exposed in isolated structural lows in the mountains, in local areas along the mountain front, and in the subsurface of the coastal plain.

All three sequences have been involved in the north-vergent Cenozoic deformation that formed the dominant structures of the northeastern Brooks Range. These structures consist of large east-trending anticlinoria cored by pre-Mississippian rocks, with Mississippian and younger rocks of the Ellesmerian sequence defining the limbs (Bader and Bird, 1986). The anticlinoria have been interpreted by some workers to be fault-bend folds in a regional-scale duplex, with a floor thrust at depth in the pre-Mississippian rocks and a roof thrust in the Mississippian Kayak Shale, near the base of the Ellesmerian sequence (e.g., Rattey, 1985; Namson and Wallace, 1986; Leiggi, 1987; Kelley and Foland, 1987; Wallace and Hanks, 1990). Others have suggested that these regional folds reflect ductile deformation of the pre-Mississippian rocks during Cenozoic deformation, with the majority of the shortening within the pre-Mississippian rocks having been accommodated by the development of penetrative fabrics (Oldow and others, 1987a). Evidence supporting both sides of the controversy is discussed in Oldow and others (1987a) and Wallace and Hanks (1990). In contrast, the Ellesmerian sequence is generally structurally detached near its base from the underlying pre-Mississippian rocks along a detachment horizon in the Kayak Shale. Cenozoic faults cutting up-section from the pre-Mississippian rocks generally flatten in this horizon and do not affect the Mississippian and younger rocks, which have shortened either by detachment folding or thrust duplication (Wallace and Hanks, 1990).

This paper discusses a cross section along a transect across the Okpilak batholith in the north-central part of the northeastern Brooks Range (Figure 5.1). The Okpilak batholith and its satellite stocks are the only large (500 km^2) intrusive bodies in the northeastern Brooks Range (Sable, 1977; Bader and Bird, 1986). The Okpilak batholith itself is a prominent topographic and structural high, reaching elevations of 2700 meters (9000 feet) immediately adjacent to deformed Upper Cretaceous and Tertiary deposits of the foredeep in the coastal plain to the north. Adjacent stratified pre-Mississippian rocks to the east are at substantially lower elevations (1500 meters/4000 feet). Although the batholith is demonstrably Devonian in age (Dillon and others, 1987), metamorphic and cooling ages (Dillon and others, 1987), apatite fission-track ages (O'Sullivan, 1989), the present structural elevation of the batholith, and Cenozoic thrust-related structures involving pre-Mississippian rocks north of the batholith (Kikiktak Mountain, Figure 5.1) all indicate that the batholith was transported to the north and uplifted during Cenozoic thrusting. Detailed structural studies on the northern margin of the batholith suggest that, during Cenozoic thrusting, the batholith deformed primarily by the formation of penetrative fabrics and thrust imbrication on semi-ductile to ductile shear zones (Hanks and Wallace, 1990).

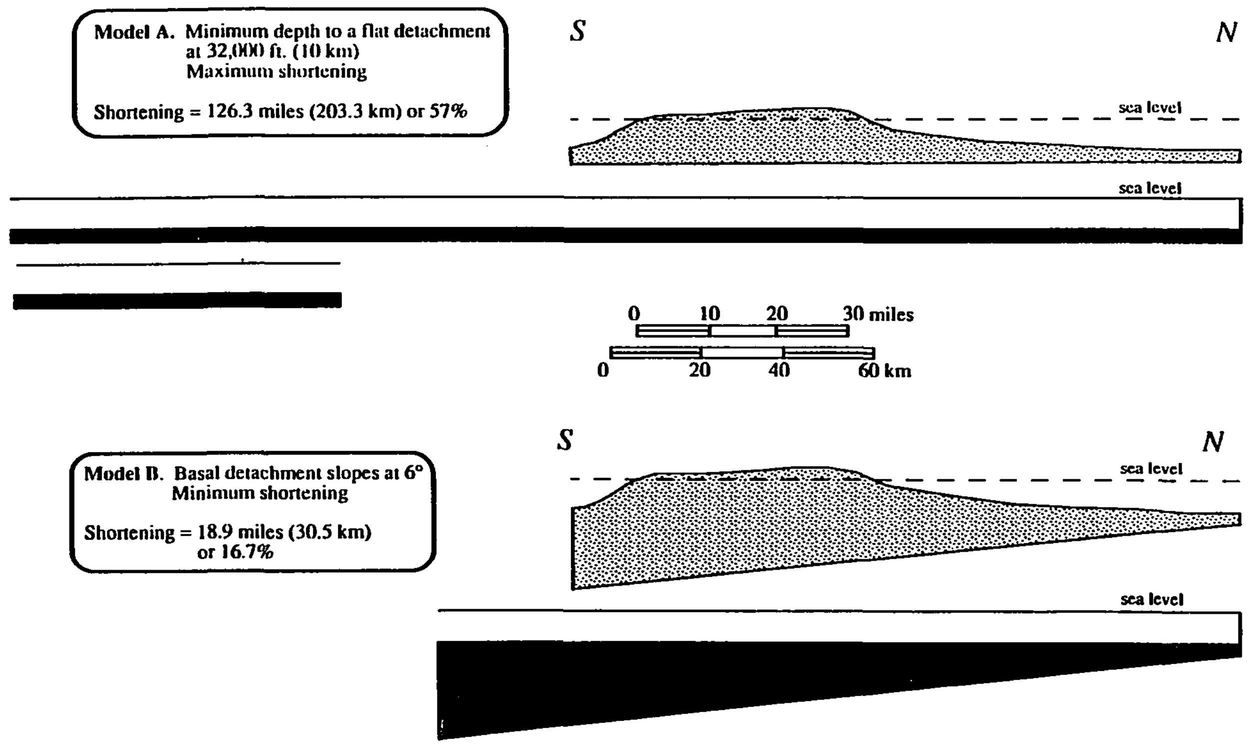
The Okpilak batholith is also one of the few places in the northeastern Brooks Range where the Kayak Shale, which serves as the regional detachment horizon near the base of the Ellesmerian sequence, is depositionally missing. Consequently, the Ellesmerian sequence remained attached to the batholith during Cenozoic thrusting, and has deformed primarily via the development of penetrative structures and minor folds and faults.

5.4 Area-balanced models

Before attempting to construct a detailed balanced cross section across the Okpilak batholith, a series of generalized area-balanced models were constructed in order to: 1) evaluate the range of possible regional shortening, 2) determine the primary regional factors that might control the gross geometry of the fold-and-thrust belt and the overall amount of shortening, 3) determine how well these factors can be constrained and which should be addressed in future studies, and 4) choose the best model to serve as a basis for a more detailed balanced cross section.

These objectives can be attained by constructing a series of models (selected examples shown in Figure 5.2) showing the range of possible orogenic wedge geometries. Each wedge model is bounded by a datum defining structural topography above and a basal detachment below. Shortening may be determined by area balancing each model. Since the models are simply area-balanced wedges, they are fairly easy and rapid to construct and balance. The range of possible wedge geometries is controlled by a number of factors,

Figure 5.2. Generalized area-balanced models of a transect across the Okpilak batholith of the northeastern Brooks Range. The structural topography of the upper surface of the wedge (the sub-Mississippian unconformity), the depth to the basal detachment surface at the pin-line, and the location of the pin-line are held constant for all the models. The dip of the basal detachment surface (with respect to the unconformity surface) , the depth to the brittle/ductile transition, and the presence of any ramps in the basal detachment surface are varied from model to model.


Model A. Minimum depth to a flat detachment at 32,000 ft. (10 km)
Maximum shortening
Shortening = 126.3 miles (203.3 km) or 57%
S
N
sea level
sea level
0 10 20 30 miles
0 20 40 60 km
S
N
Model B. Basal detachment slopes at 6°
Minimum shortening
Shortening = 18.9 miles (30.5 km) or 16.7%
sea level
sea level

Model C. Basal detachment slopes 6°, flattening at a brittle/ductile transition at 15 km (49,000')

Shortening = 76.5 miles (123.2 km) or 44.7 %

Model D. Basal detachment slopes 6°, flattening at a brittle/ductile transition at18 km (59,000')

Shortening = 52.6 miles (84.7 km) or 35.7%

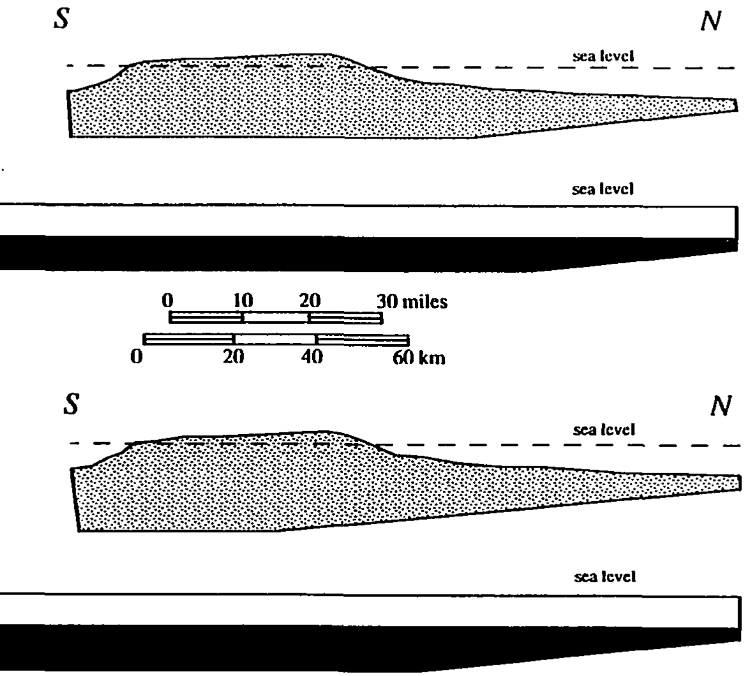


Figure 5.2. (cont)

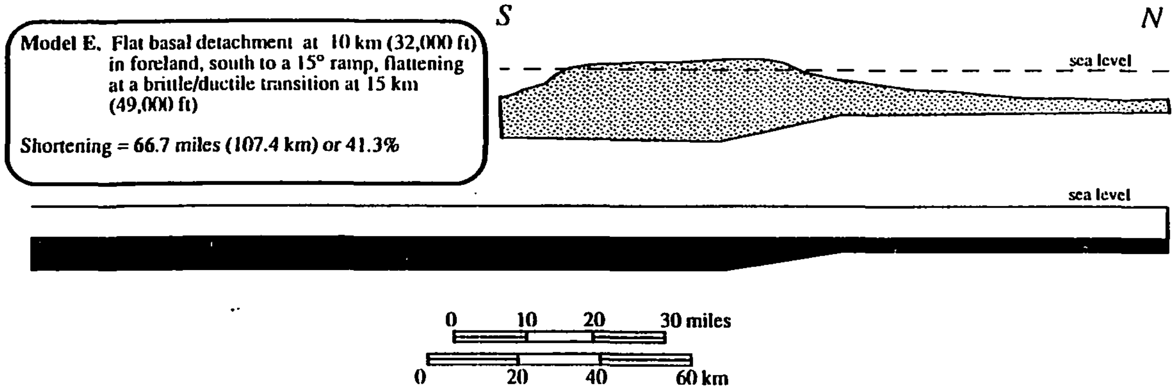


Figure 5.2. (cont)

which can be constrained to varying degrees.

As seen in figure 5.1 and discussed above, the lowest structural level exposed in the northeastern Brooks Range consists of pre-Mississippian rocks. The regional anticlinoria cored by pre-Mississippian rocks form the dominant structures of the region, and control the distribution and gross structural geometry of all stratigraphically higher strata. Ellesmerian and Brookian sequence rocks are structurally separated from the underlying pre-Mississippian rocks by a major regional detachment surface in a shale near the base of the Ellesmerian cover sequence. Consequently, in order to evaluate the gross geometry of the orogenic wedge in the northeastern Brooks Range (specifically the geometry of the basal detachment surface), only the pre-Mississippian rocks need be included in the models. Thus, the sub-Mississippian unconformity surface was used as the regional datum that defines the upper surface of each wedge. This surface was drawn based on regional surface geologic data and the published seismic reflection data from the coastal plain to the north (Bader and Bird, 1986; Bird and Magoon, 1987; Wallace and Hanks, 1990). Since the topography of this surface is perhaps the best constrained feature of the fold-and-thrust belt, it was held constant in all of the models.

Based on the published seismic reflection data, a pin line was chosen in the subsurface north of the batholith. Cenozoic deformation has undoubtedly continued beyond this point, as suggested by deformed Cenozoic sediments and active seismicity (Grantz and May, 1983; Craig and others, 1985; Moore and others, 1985a; Carter and others, 1986; Bruns and others, 1987; Clough and others, 1987; Kelley and Foland, 1987), but the amount of shortening is minor and the seismic data are discontinuous to the north. The pin line was therefore placed in the northernmost structural low shown by the seismic data, where structural thickening is at a minimum, and the models were restored from that point. Until more and better quality seismic data become available, this is the most suitable pin line and was used for all of the models. This pin line defines the northern limit of each model (Figure 5.2).

The depth to the basal detachment surface at the pin line was also inferred from the published seismic reflection data, based on the maximum structural relief on the mapped sub-Mississippian unconformity surface adjacent to the pin line. Two-way travel time to the surface was converted to depth using a time/depth conversion chart (Bruns and others, 1987). The minimum depth to the basal detachment surface under the coastal plain is inferred to be 10 km, assuming that the structural relief is due to structural duplication in a fault-bend fold-style structure (Suppe, 1983), with the difference between the height of the unconformity at the pin line and at the top of the nearest structure representing the thickness of a 'stiff' layer of pre-Mississippian rocks. The depth to the basal detachment horizon could be deeper if it is assumed that the structural relief does not represent the total thickness of the thrust slice of pre-Mississippian rocks. Because the minimum depth is constrained by at least some data and the simplest assumption, it too was held constant in all of the models.

Observations on the dip of basal detachment horizons in other fold-and-thrust belts suggest that basal detachment surfaces can dip towards the hinterland by up to 6° under an actively growing wedge (Davis and others, 1983). Due to the lack of resolution in the seismic data for the coastal plain, the slope of the basal detachment surface south of the pin line is not constrained in the northeastern Brooks Range. The dip of this surface with respect to the sub-Mississippian unconformity surface from the pin line was varied from 0° to 6° in the area-balanced models (e.g., models A and B, Figure 5.2).

The basal detachment surface may also include one or more relatively steeper segments, or ramps (e.g., model E, Figure 5.2). Ramps were incorporated in some of the models (Figure 5.3), although the location, dip, and height of actual ramp(s) are unknown. An arbitrary ramp location and dip were chosen for those models incorporating a ramp, but remained constant from model to model. Ramp height was varied (Figure 5.3).

Is it reasonable to assume that a basal detachment surface will maintain a constant dip under a growing wedge, or will it flatten at some horizon? Evidence provided by deep earthquakes in continental areas suggests that brittle behavior is limited to the upper crust, with aseismic lower crust indicative of ductile deformational mechanisms (Chen and Molnar, 1983). The depth to the brittle/ductile transition in continental regions generally is thought to vary from 10 to 50 km (Bott, 1982; Suppe, 1985), with the exact depth of the transition dependent on the age, composition, and thickness of the crust, the geothermal gradient, and the fluid pressure. Young, hot crust will have a shallower depth to the brittle/ductile transition (<20 km) than will old, cold crust (~25 km or greater) (Chen and Molnar, 1983; Suppe, 1985). This transition from brittle to ductile deformation could define a zone of low strength in the lower crust that would facilitate the development of crustal nappes, or 'crustal flakes,' during contractional deformation (Chen and Molnar, 1983; Dewey and others, 1986). Such a zone would be, in effect, a crustal-scale detachment horizon, permitting brittle faults in the overlying brittle upper crust to flatten at some depth in the more ductile and weaker lower crust. However, the depth to the brittle/ductile transition, and hence the detachment, would depend on the geothermal gradient and fluid pressure at the time of thrusting, and the type of rocks involved in the deformation (Bott, 1982; Suppe, 1985).

For the purpose of the area-balanced models, the basal detachment surface was allowed to flatten at a hypothesized brittle/ductile transition. The depth at to this transition in the northeastern Brooks Range is not presently known and is, therefore. also a variable in the models (e.g., models C and D, Figure 5.2). If the brittle/ductile transition defines the maximum possible depth of the basal detachment horizon, it must be at least as deep as the depth to the basal detachment surface at the pin line in the foreland, or 10 km. Apatite fission-track studies (O'Sullivan, 1988) and conodont color-alteration indices (Watts and Harris, personal communication) suggest that the geothermal gradient at the time of thrusting in the northeastern Brooks Range was greater than 20°C/km. Consequently, the brittle/ductile transition was not placed any deeper than 18 km in the models.

Figure 5.2 illustrates five representative area-balanced cross sections that incorporate the constraints and variables discussed above. These generalized models illustrate the major regional geometric factors that control the overall wedge geometry and amount of shortening across any thrust belt, and along the Okpilak batholith transect in particular. These factors are: 1) the structural topography of the deformed sub-Mississippian unconformity surface along the entire cross section, which defines the upper surface of a deformed wedge, 2) the depth to the basal detachment surface at the pin line in the foreland,

3) the dip of the basal detachment surface with respect to the sub-Mississippian unconformity, 4) the presence of major ramps in the basal detachment surface, and 5) the depth at which the basal detachment surface flattens (due to whatever cause). In balancing the models, the line length of the upper surface of the wedge (in this case the sub-Mississippian unconformity) was not held constant, but was allowed to lengthen during reconstruction as required by the variables incorporated in that model. In effect, this allows for shortening of the sub-Mississippian unconformity during deformation by various mechanisms, including penetrative strain and/or thrust duplication.

Figure 5.3 is based on the results of area-balancing 12 generalized cross sections and graphically illustrates the effect of the different variables on the amount of shortening. The greatest range in possible shortening values is seen in the wedges with the simplest geometry, a constant slope on the basal detachment (models A and B, Figures 5.2 and 5.3). The most shortening occurs in the model with the lowest basal detachment dip (0°, model A) and the least shortening in the model with the greatest basal detachment dip (6°, model B). Dips between 0° and 6° yield shortening values that define a smooth curve between the two end-member values (Figure 5.3). Modifying this simple wedge geometry by flattening the basal detachment at some depth results in shortening values that consistently fall to the right of this curve, indicating greater shortening for a given detachment dip (e.g., models B, C, and D, Figure 5.3). Modifying the simple wedge geometry by addition of a ramp to a detachment of constant dip results in shortening values that fall to the left of the curve, indicating less shortening for a given detachment dip (e.g., models A and E, Figure 5.3).

5.5 A detailed balanced cross section through the Okpilak batholith.

A particular area-balanced model can be chosen as the basis for a more detailed cross section by incorporating into the interpretation regional and detailed observations regarding structural geometries. For example, the geometry of the sub-Mississippian unconformity in the subsurface north of the batholith can be interpreted to reflect pre-Mississippian rocks deformed into fault-bend folds above a relatively flat basal detachment horizon. The

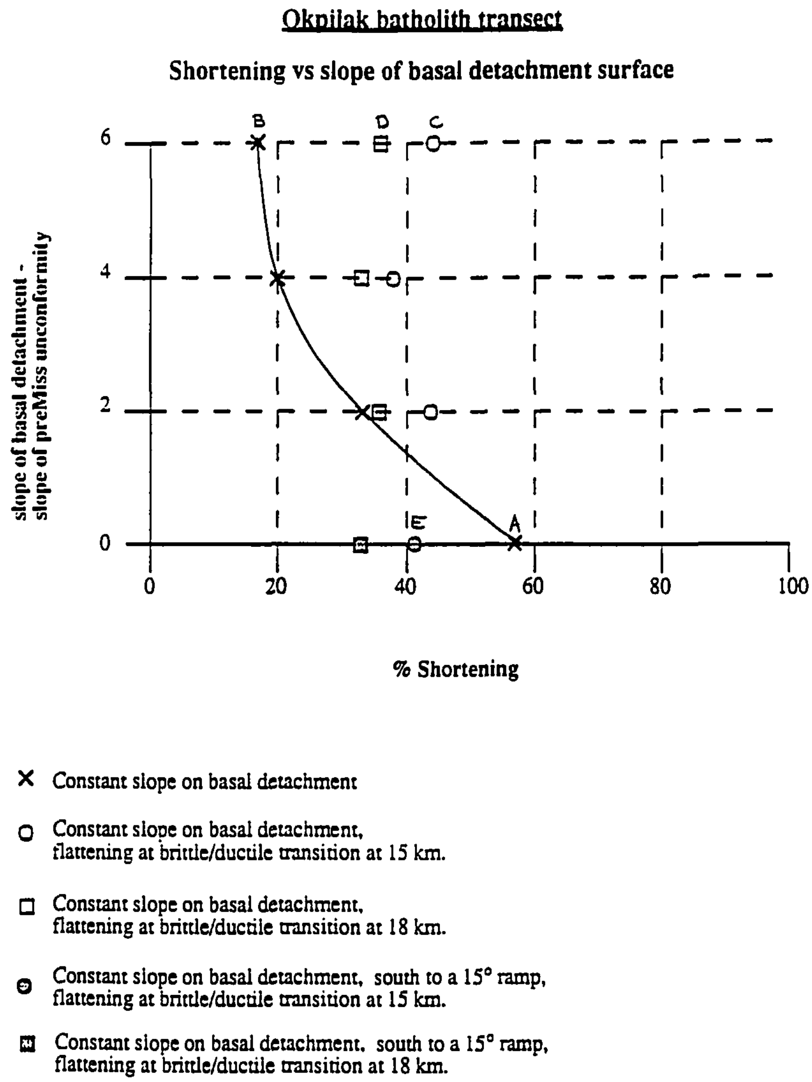


Figure 5.3. A graph of the dip of the basal detachment horizon vs. shortening illustrates the range of possible shortening along the cross section as determined by the area-balanced models.

dramatic and abrupt southward increase in structural relief at the range front suggests the presence of a significant hangingwall ramp that coincides with the exposed northern margin of the batholith (Figure 5.1). The structural relief remains relatively constant for a considerable distance to the south, but decreases near the south end of the cross section. This suggests a relatively flat detachment beneath the batholith, with a footwall ramp to the south. Based on these observations, the detailed balanced cross section (Figure 5.4) includes a flat basal detachment horizon with a major ramp, essentially a refinement of model E (Figure 5.2). Also incorporated in the detailed balanced cross section are field-based observations on how the batholith itself deformed during Cenozoic thrusting.

Determination of the depth at which the basal detachment flattens into the brittle/ductile transition in this detailed section remains a significant problem. The only constraints on the depth to the brittle/ductile transition are that: 1) it must be significantly deeper than 10 km (the depth to the basal detachment horizon in the foreland), 2) it must be consistent with a geothermal gradient that is reasonable for the region, the tectonic setting, and granitic rocks, 3) it must be consistent with the generally low metamorphic grade seen in the vicinity of the batholith, and 4) it must be consistent with the known structural geometries and low metamorphic grade seen south of the batholith. At this time, the geothermal gradient during Cenozoic thrusting is poorly constrained, the P/T history of the pre-Mississippian rocks that make up the majority of the exposed rocks south of the batholith is not known (except that the rocks are low grade), and the Cenozoic structural geometry of the pre-Mississippian rocks south of the batholith is speculative at best. Research in all of these areas could potentially yield critical observations that would further constrain the geothermal gradient, the depth to the brittle/ductile transition, and thus the balanced cross section.

Despite all of these uncertainties, the resulting detailed balanced cross section (Figure 5.4) does provide some clues as to how the batholith as a whole may have behaved during Cenozoic thrusting and may explain certain regional structural trends in the vicinity of the batholith. Northward transport of stratified pre-Mississippian rocks above a relatively shallow basal detachment horizon and emplacement of those rocks against the southern

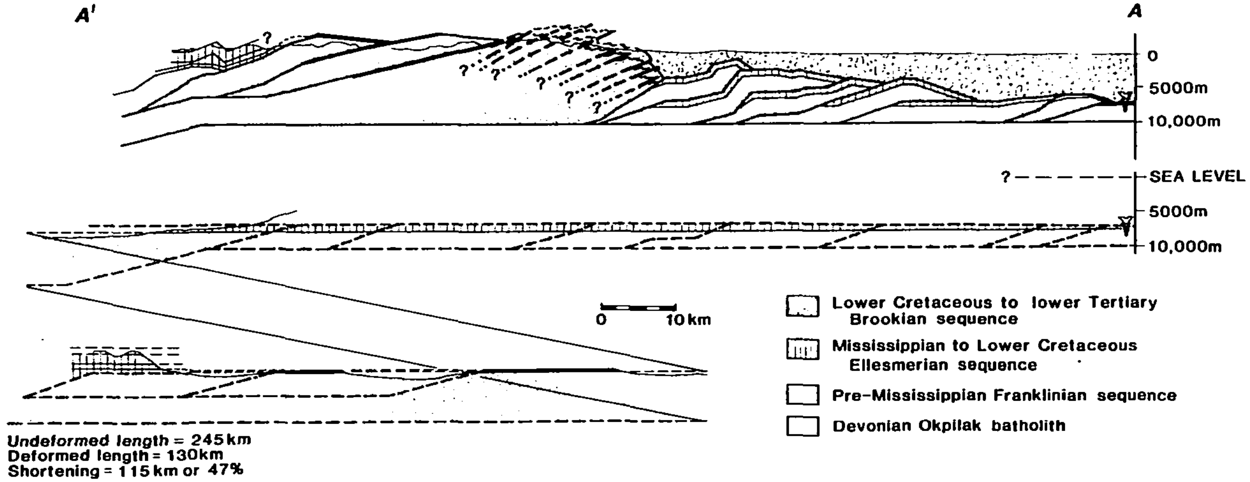


Figure 5.4. Balanced cross section across Okpilak batholith, along line A-A' in Figure 5.1. (From Hanks and Wallace, 1990.)

margin of the batholith early in the Cenozoic deformational history of the region may have caused deflection of regional structural trends south of the batholith as the batholith acted as a structural buttress (Figures 5.1 and 5.4). Eventual northward transport of the batholith itself may have been due to the activation of a deeper detachment horizon in the vicinity of the batholith, with the older thrust sheets being transported 'piggyback' on the deeper detachment. Alternatively, the arcuate trends south of the batholith could reflect folding of the older structural trends as these structures were deformed over lateral footwall ramps bounding the batholith. Immediately prior to or during transport of the batholith, shortening within the batholith was accommodated by the development of discrete thrust slices bounded by ductile shear zones and the development of penetrative mesoscopic and microscopic fabrics throughout the batholith (Hanks and Wallace, 1990). The activation of a deeper basal detachment horizon was probably confined to the vicinity of the batholith, resulting in separation from the shallower basal detachment horizons to the east and west by lateral ramps. The northern margin of the batholith probably marks a hangingwall ramp that reflects a footwall ramp in the basal detachment surface from approximately 18 km to 10 km depth. The cause of this ramp is unknown, but the presence of stratified pre-Mississippian rocks north of the batholith at Kikiktat Mountain (Figure 5.1) suggests that it may be due to a northward change in lithology from homogeneous granitic rocks to heterogeneous and stratified metasedimentary and metavolcanic rocks. The location of the footwall ramp in the detailed section (Figure 5.4) is based on restoration of the deformed pre-Mississippian rocks north of the batholith.

5.6 Conclusions

Generalized area-balanced cross sections across the Okpilak batholith of the northeastern Brooks Range pinpoint certain key regional constraints on the geometry of the fold-and-thrust belt wedge and amount of shortening possible in the region. These constraints are: 1) the structural topography of the deformed sub-Mississippian unconformity along the entire transect, 2) the depth to the basal detachment surface at the pin line, 3) the dip of the basal detachment surface with respect to the unconformity as measured from the pin line, 4) the presence of major ramps in the basal detachment surface, and 5) the depth to the brittle/ductile transition, where the basal detachment horizon might be expected to flatten. Some of these factors are better constrained in the region than others, and can be held constant in the models (e.g., factors 1 and 2); the others are less well constrained and can be considered variables. Based on reasonable variations in just these five regional factors, the regional shortening in this area ranges from 17-57%, corresponding to a total shortening for the entire transect of 31 km (19 miles) to 203 km (126 miles); or a northward transport of the batholith ranging from 11 km (7 miles) to 86 km (54 miles).

The model that best fits the regional and detailed observations regarding structural geometries and speculations regarding the geometry of the basal detachment horizon can then be refined into a detailed balanced cross section. This cross section incorporates rigid deformation of the stratified pre-Mississippian rocks north and south of the batholith by faults and related folds, with the batholith itself deforming via the development of more penetrative structures. The dip of the basal detachment horizon remains fixed at 0°, with a major footwall ramp in the southern part of the cross section. The amount of shortening represented by this section is 47% or 115 km (69 miles), including northward transport of the batholith by 58 km (36 miles). This value falls well within the range of shortening values predicted by the area-balanced models.

The models used in this study do not by any means exhaust either the possible range in or controls on wedge geometries in a fold-and-thrust belt such as the northeastern Brooks Range. For example, these models do not incorporate a dip toward the hinterland of the upper datum (in this case, the sub-Mississippian unconformity) in the restored section, which would obviously affect the pre-deformational geometry and thus the total amount of shortening. The models also do not incorporate other potential causes for flattening of the basal detachment in the hinterland, such as a major lithology change in the basement, nor do the models incorporate changes in dip or depth of the basal detachment horizon during deformation. All of these factors could change the range in shortening and geometry of the orogenic wedge, and could be investigated using this technique if warranted or desired.

Constructing generalized area-balanced models across a region can be very useful in areas with limited surface and/or subsurface data. The area-balanced models provide a rapid and visual means of exploring and illustrating the key regional constraints on the gross geometry of a regional cross section and how assumptions regarding those constraints control the final wedge geometry and thus the range in possible shortening in the region. This in turn pinpoints what types of data and/or observations are needed to refine the model further and construct a better-constrained and more detailed balanced cross section.

5.7 Acknowledgments

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CHAPTER 6: DISCUSSION AND CONCLUSIONS

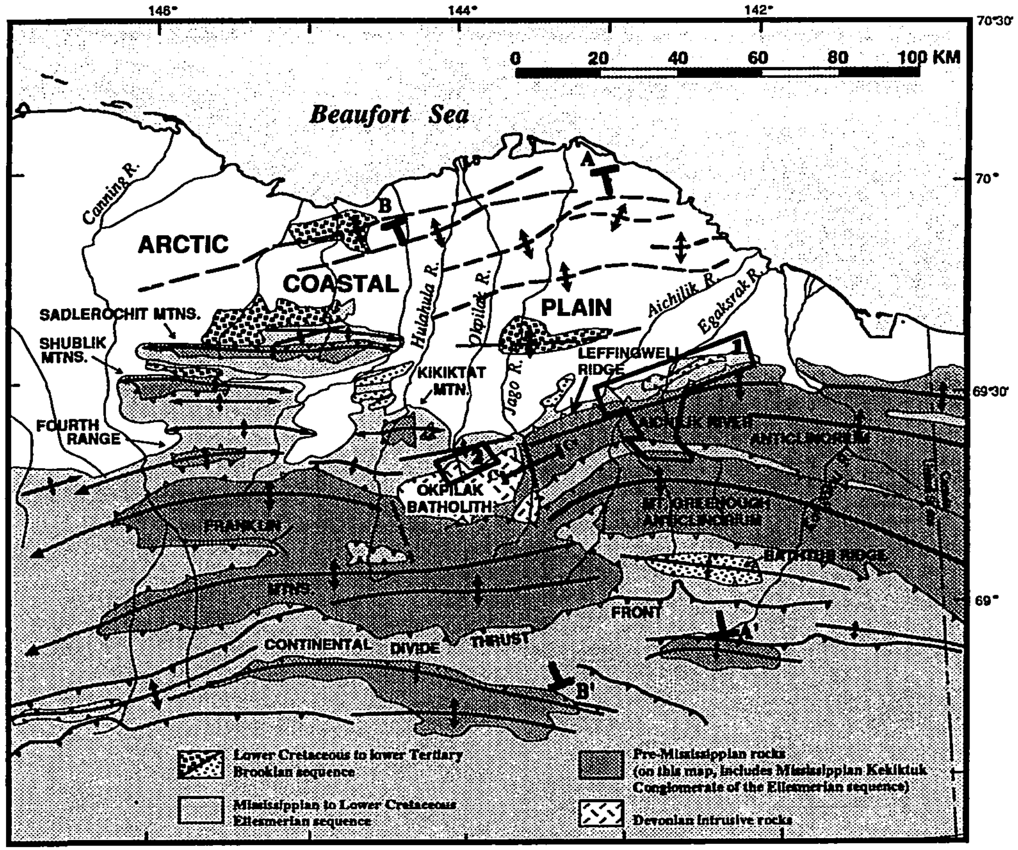
This dissertation addresses several major questions applicable to the external regions of mountain belts world-wide, focussing on a particular fold-and-thrust belt, the northeastern Brooks Range. The data, analysis, and conclusions for two representative transects have been presented in the preceding chapters. In this chapter, I will briefly compare the structural characteristics and interpretations of the two transects. I will then discuss how these two transects provide some possible answers to the various questions outlined in chapter 1.

6.1 A comparison of the Aichilik River and Okpilak batholith transects

The Aichilik River and Okpilak batholith transects are two parallel transects through northeastern ANWR that cross the structural strike of the region (Figure 6.1). The two transects are approximately 27 miles (45 km) apart, have pin lines in approximately the same structural position in the coastal plain, and are of approximately the same length. The main differences between the two transects are the lithology and structural style of the pre-Mississippian rocks in the area covered by each transect, and the lithology and structural style of the Mississippian through Triassic Ellesmerian cover sequence. These differences are summarized in Table 6.1.

Balanced cross sections were drawn parallel to each transect (Figures 6.2 and 6.3). Both cross sections were line-length balanced using the sub-Mississippian unconformity surface as the datum, except for the Okpilak batholith, which was area-balanced. Interpreted seismic data from the coastal plain to the north of each transect (Bruns and others, 1987) provided subsurface information for the northern portions of both transects. These data were used to infer the depth to and topography of the sub-Mississippian unconformity surface and the depth to the basal detachment horizon under the coastal plain. My interpretation of the Cenozoic structural style of both the pre-Mississippian rocks and

Figure 6.1. Tectonic map of the northeastern Brooks Range showing the locations of the two regional transects. A-A': Aichilik River transect; B-B': Okpilak batholith transect; C-C': schematic strike section illustrated in Figure 6.7. Areas 1 and 2 refer to the detailed study areas documented in this dissertation: area 1: Aichilik and Egaksrak Rivers area (chapter 3); area 2: northern margin of the Okpilak batholith (chapter 4).



146°
144°
142°
70°30'
70°
69°30'
69°
0 20 40 60 80 100 KM
Beaufort Sea
Canning R.
ARCTIC
COASTAL
PLAIN
Hulahula R.
Okpilak R.
Jago R.
Aichilik R.
Egaksrak R.
A
B
A'
B'
SADLEROCHIT MTNS.
SHUBLIK MTNS.
FOURTH RANGE
KIKIKTAT MTN.
LEFFINGWELL RIDGE
OKPILAK BATHOLITH
AICHILIK RIVER
ANTICLINORIUM
MT. GREENOUGH ANTICLINORIUM
FRANKLIN
MTNS.
FRONT
CONTINENTAL DIVIDE THRUST
Lower Cretaceous to lower Tertiary Brookian sequence
Mississippian to Lower Cretaceous Ellesmerian sequence
Pre-Mississippian rocks (on this map, includes Mississippian Kekiktuk Conglomerate of the Ellesmerian sequence)
Devonian intrusive rocks

Table 6.1. Table comparing the lithology and structural style of the primary lithotectonic elements of the Aichilik River and Okpilak batholith transects

		Aichilik River transect	Okpilak batholith transect
Remaining Ellesmerian sequence rocks			
	Cenozoic structural style	Thrust duplication of Lisburne Group through Karen Creek Sandstone between detachments in the Mississippian Kayak and Jurassic Kingak Shales.	Pervasive strain and mesoscopic structures with minor thrust faulting and related folding.
	Lithology	Thick carbonates of the Lisburne Group overlain by clastic rocks of the Sadlerochit Group, Shublik Formation and Karen Creek Sandstone.	Thick carbonates of the Lisburne Group overlain by clastic rocks of the Sadlerochit Group.
Kayak Shale			
	Cenozoic structural style	Tight to isoclinal folds with related pervasive axial planar cleavage.	Not applicable
	Lithology	In north: Silty shale with thick, laterally discontinuous carbonate buildups in upper portion. In south: Shale with minor calcareous sandstones and sandy limestones in upper portion.	Depositionally thin to absent
Kekiktuk Conglomerate			
	Cenozoic structural style	Minor mesoscopic structures	Pervasive strain and mesoscopic structures with minor thrust faulting and related folding.
	Lithology	Discontinuous lenses of quartz sandstones and granule conglomerates.	Coarse-grained quartz sandstones and granule conglomerates of highly variable thickness.
pre-Mississippian rocks			
	Cenozoic structural style	Thrusts and related folds with locallized minor mesoscopic structures	Pervasive strain and mesoscopic structures with minor thrust faulting along ductile shear zones.
	Lithology	Stratified, slightly metamorphosed sedimentary and volcanic rocks	Granitic rocks of the Okpilak batholith

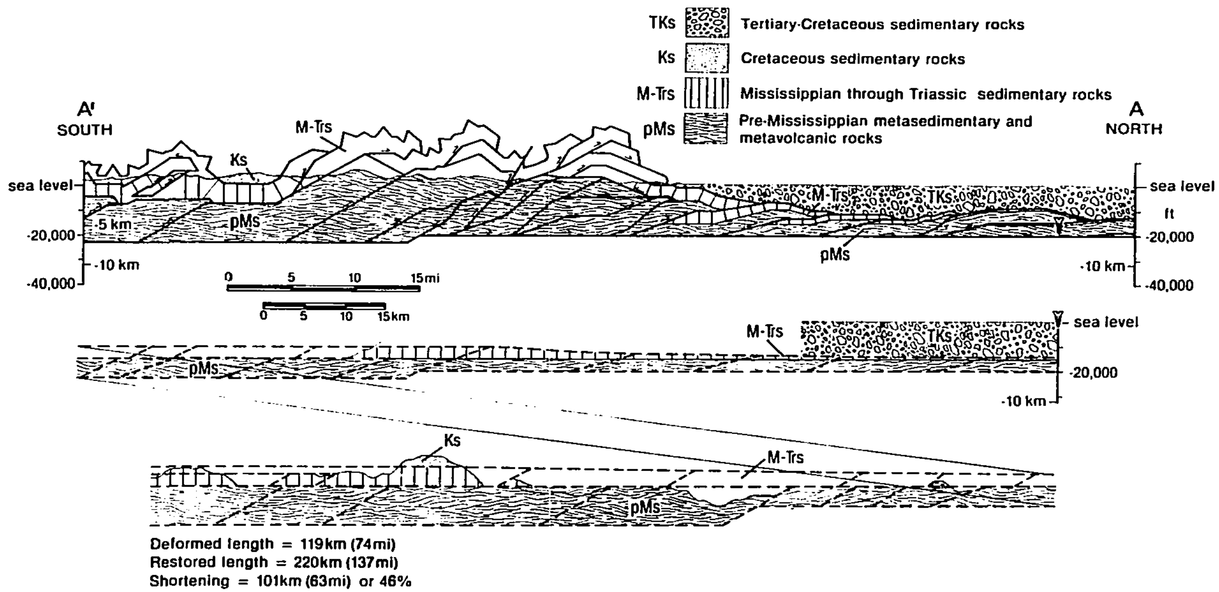


Figure 6.2. Balanced cross section of the Aichilik River transect. Cross section location is shown on figure 6.1 as A-A'.

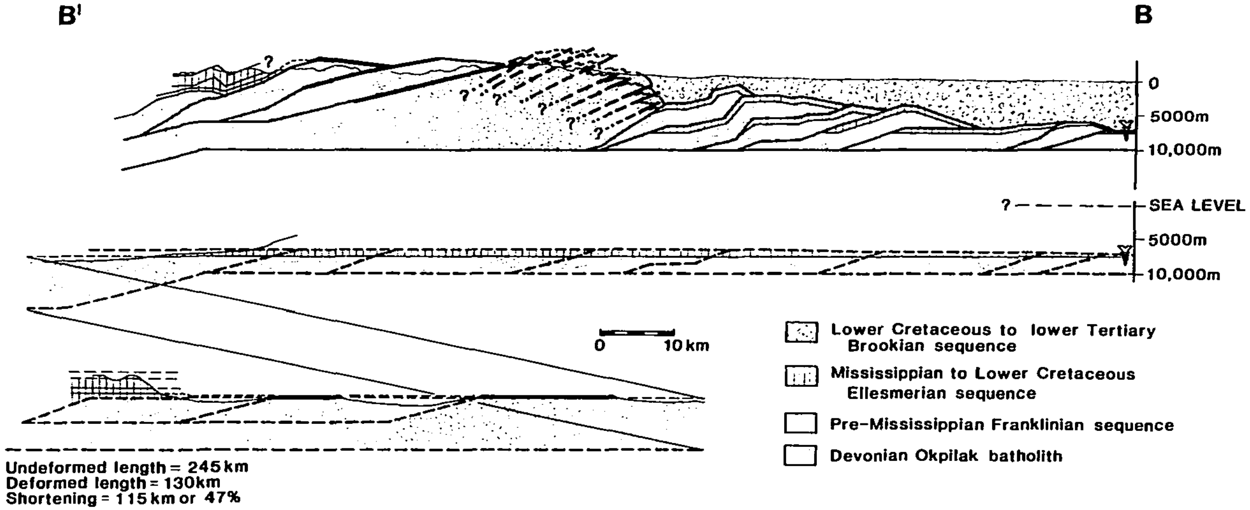


Figure 6.3. Balanced cross section of the Okpilak batholith transect. Cross section location is shown on figure 6.1 as B-B'.

the Ellesmerian cover sequence was based on my surface observations and mapping along portions of both transects. Published mapping, primarily Reiser and others (1980) and Sable (1977), provided data in those areas of the two transects I was unable to visit. The main differences and similarities between the two balanced cross sections are summarized in Table 6.2. The significance of these similarities and differences between the two transects is summarized in the remainder of this chapter.

6.2 Questions successfully addressed by this study

At least three of the main questions raised in chapter 1 have been successfully addressed during the course of this study. Other questions could not be answered for a variety of reasons, including the limited areal extent of the area, limited time and limited exposure. And, as with any study, some of the study's results created more questions than they answered. In this section, I will focus on how my observations and conclusions regarding these two very different transects can answer some of the major questions posed in chapter 1.

6.2.1 How has 'depositional basement' behaved during Cenozoic deformation of northeastern ANWR?

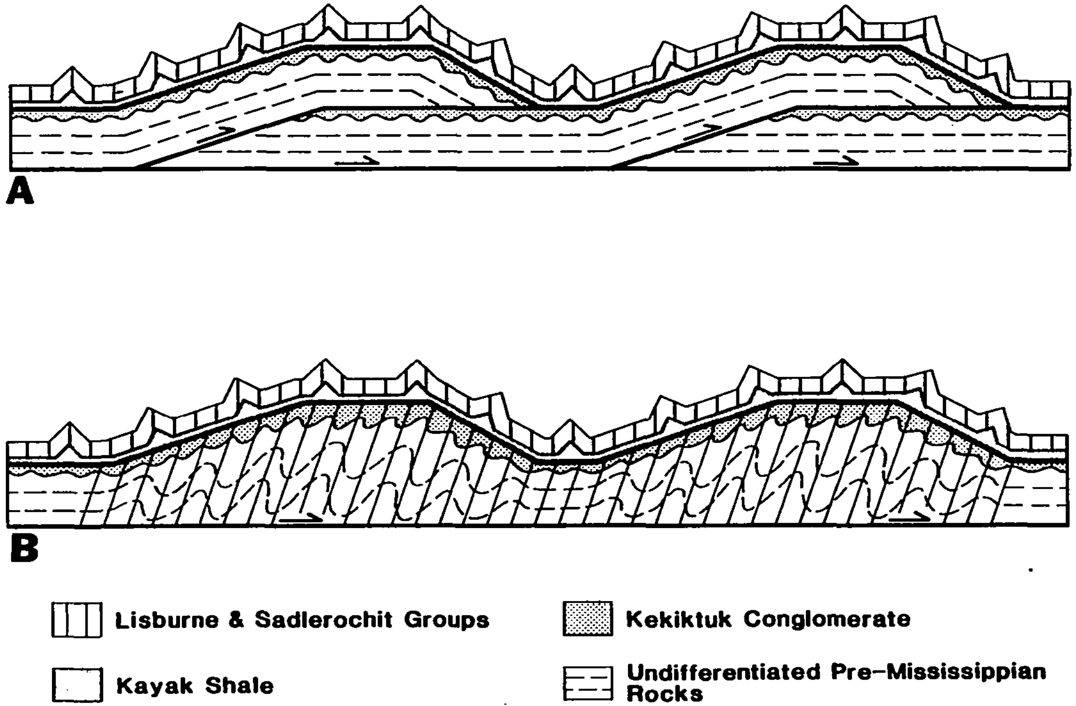
One of the major questions regarding the structural evolution of northeastern ANWR is the role that the pre-Mississippian depositional basement has played in Cenozoic deformation. Two competing, but not necessarily mutually exclusive, hypotheses have been proposed (Figure 6.4). One model suggests that shortening was via thrust duplication (Model A), while the other model involves shortening primarily by penetrative strain (Model B).

A number of workers have suggested that the majority of Cenozoic shortening within the pre-Mississippian sequence has been accommodated by thrust duplication in a

Table 6.2. Table comparing the balanced cross sections of the Aichilik River and Okpilak batholith transects.

	Aichilik River transect	Okpilak batholith transect
Detailed balanced section		
Method of balancing	line balanced on pre-Mississippian unconformity	line balanced on pre-Mississippian unconformity, except for Okpilak batholith, which is area balanced.
Depth to basal detachment horizon at pin line	-20,000 feet (-6.1 km)	-32,000 feet (-9.8 km)
Maximum postulated depth to basal detachment horizon	-31,000 feet (-9.5 km)	-48,000 feet (-14.6 km)
Deformed length	74 miles (119 km)	80 miles (129 km)
Restored length	137 miles (220 km)	153 miles (247 km)
Shortening	63 miles (101 km) or 46%	73 miles (118 km) or 48%
Area-balanced models		
Range of shortening	16-61%	17-57%

Figure 6.4. Conceptual end-member models for the mode of shortening of the pre-Mississippian rocks and the Kekiktuk Conglomerate during Cenozoic thrusting. Model A represents a north-vergent regional duplex, where Cenozoic shortening of the pre-Mississippian rocks is accommodated by thrust duplication. The floor thrust of the duplex is at depth in the pre-Mississippian rocks, and the roof thrust is in the Mississippian Kayak Shale. Model B represents a scenario where Cenozoic shortening within the pre-Mississippian rocks is accommodated primarily by strain and mesoscopic structures. Dashed lines within pre-Mississippian rocks represent arbitrary markers that were horizontal prior to Cenozoic deformation, and do not represent bedding. Steeply dipping solid lines within the pre-Mississippian rocks and Kekiktuk Conglomerate represent cleavage.


A
B
Lisburne & Sadlerochit Groups
Kekiktuk Conglomerate
Kayak Shale
Undifferentiated Pre-Mississippian Rocks

northward-propagating regional duplex. This duplex developed between a floor thrust at depth in the pre-Mississippian rocks and a roof thrust in the Mississippian Kayak Shale of the overlying Ellesmerian sequence (Figure 6.4 A; Rattey, 1985; Namson and Wallace, 1986; Leiggi, 1987; Kelley and Foland, 1987; Ziegler, 1989; Wallace and Hanks, 1990). In this model, shortening via the development of penetrative strain and mesoscopic structures is minor. In the past, evidence for this mechanism of shortening has come primarily from regional and local studies in the Canning River area of western ANWR, where the basal conglomerate of the Ellesmerian cover sequence, the Mississippian Kekiktuk Conglomerate, is believed to have remained structurally coupled to the pre-Mississippian rocks during Cenozoic deformation. In this region, the basal conglomerate defines a broad, relatively continuous folded surface across each anticlinorium that is interpreted to reflect a Cenozoic fault-bend fold geometry within the underlying pre-Mississippian rocks (Namson and Wallace, 1986; Wallace and Hanks, 1990). This geometry is also supported by the structure contour map of the sub-Mississippian unconformity in the Canning River area, which suggests that each anticlinorium in northwestern ANWR consists of a single broad structural high (Figure 6.5 A; Wallace and Hanks, 1990). A detailed structural study of the northern flank of one of these anticlinoria also suggests that most of the Cenozoic shortening within the pre-Mississippian sequence is via thrust duplication, with the contribution by penetrative strain and mesoscopic structures to the total amount of Cenozoic shortening being relatively minor (Ziegler, 1989).

In contrast, Oldow and others (1987a) have proposed that most of the Cenozoic shortening in the pre-Mississippian sequence has been accommodated by the development of penetrative mesoscopic structures, as in figure 6.4 B. Evidence for this mechanism of Cenozoic shortening is based on a detailed structural study of two small areas on the northern flanks of the two largest anticlinoria of northwestern ANWR. They noted that second-generation penetrative fabrics within the pre-Mississippian rocks were similar in orientation to fabrics within the Mississippian Kayak Shale. Using this observation, they

Figure 6.5. (A) Structure contour map of the sub-Mississippian unconformity in the exposed regions of eastern ANWR, with a contour interval of 1000 feet. Data derived from Bader and Bird (1986) by extrapolation into the subsurface using published thicknesses and by assuming a minimum elevation of the unconformity surface over areas of exposed pre-Mississippian rocks.

STRUCTURE CONTOUR MAP ON THE SUB-MISSISSIPPIAN UNCONFORMITY

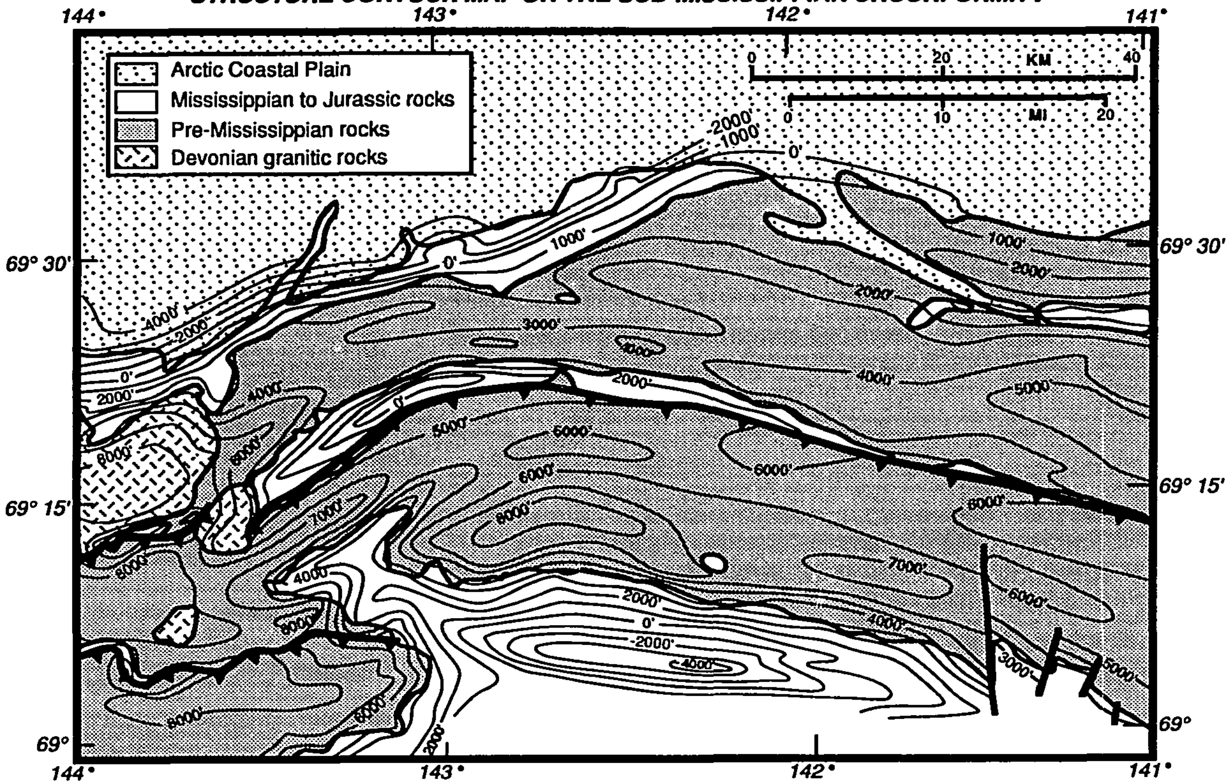
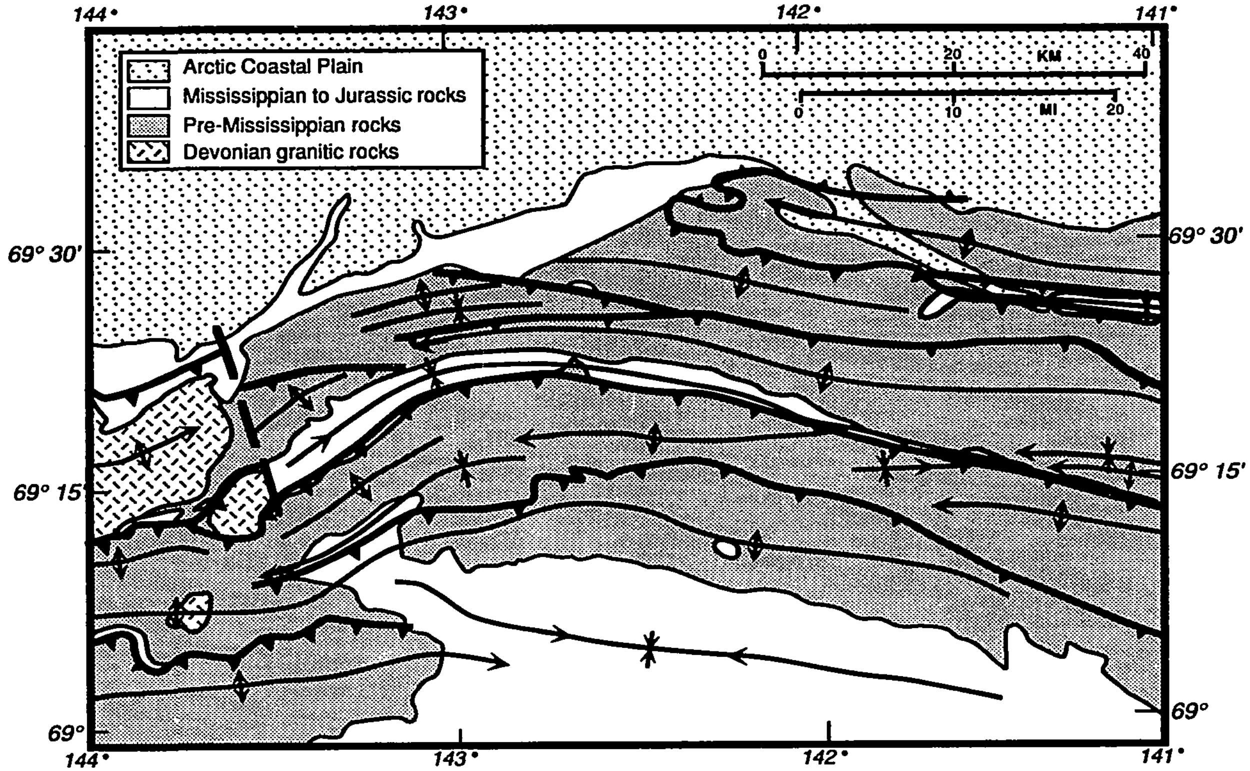


Figure 6.5. (B) Map showing interpreted traces of Cenozoic folds and thrust faults that deform the sub-Mississippian unconformity in eastern ANWR. Crests and troughs of folds are based on the structure contour map. Thrust faults are based on mapped post-Mississippian faults within the pre-Mississippian sequence, boundaries between major lithologic packages in the pre-Mississippian rocks, regional map patterns of the exposed unconformity surface (Bader and Bird, 1986) and abrupt changes in structural relief on the structure contour map.

CENOZOIC STRUCTURES AFFECTING PRE-MISSISSIPPIAN ROCKS



surmised that these D2 fabrics were Cenozoic in age, and represented the major mechanism of shortening within the pre-Mississippian sequence. They speculated that Cenozoic shortening within the basement sequence could be on the scale of hundreds of kilometers, which would represent >60% total shortening. Unfortunately, their study did not extend to the Kekiktuk Conglomerate, but they suggested that the basal conglomerate was detached from both the underlying pre-Mississippian rocks and overlying Kayak Shale, and was complexly shortened by both penetrative strain and duplex formation.

My work suggests that Cenozoic deformation within the pre-Mississippian sequence of northeastern ANWR was not solely by thrust duplication or by penetrative strain. In the Aichilik River area, the evidence suggests that most of the Cenozoic shortening within the pre-Mississippian sequence was by thrust duplication, with the development of penetrative structures playing a relatively minor role. Although penetrative structures probably related to Cenozoic formation of the Aichilik River anticlinorium are locally present, they are not abundant throughout the area, and appeared to be concentrated near Cenozoic structures. However, the two major anticlinoria of the Aichilik River area (the Aichilik River anticlinorium and the Mount Greenough anticlinorium) both appear to be more complex than similar anticlinoria in northwestern ANWR. Each anticlinorium in northeastern ANWR is composed of multiple thrust-bounded horses in a duplex of Cenozoic age (Figure 6.2 and 6.4 A), in contrast to the anticlinoria formed by individual thrust-bounded horses as seen in northwestern ANWR (Wallace and Hanks, 1990).

In contrast to this mode of deformation in the Aichilik River area, Cenozoic shortening within the Okpilak batholith appears to have been dominated by the development of penetrative strain and mesoscopic structures, with some additional shortening accommodated by thrust duplication on ductile and semi-ductile shear zones (Figure 6.3). Cenozoic penetrative structures are pervasive within the granitic rocks of the northern margin of the batholith (Sable, 1977; Hanks and Wallace, 1990), as well as common throughout the main body of the batholith (Wallace, pers. comm.; Sable, 1977). Thrust duplication also occurs on the northern margin of the batholith, although total displacements on these thrust faults are probably less than 2-3 km (Figure 6.3).

This contrast in Cenozoic structural styles within the pre-Mississippian rocks along the range front of eastern ANWR suggests that the structural response of the pre-Mississippian sequence during Cenozoic deformation was controlled, at least partially, by the lithologic character of the rocks involved. The pre-Mississippian rocks of the Aichilik and Egaksrak Rivers area are mechanically heterogeneous, highly stratified low-grade metasedimentary rocks, with numerous shale horizons that could act as good detachment surfaces during a shortening event. This mechanical heterogeneity resulted in Cenozoic shortening by thrusting and related fold development, with penetrative mesoscopic structures forming only immediately adjacent to the Cenozoic thrust faults. The granitic rocks of the Okpilak batholith, on the other hand, are mechanically very homogeneous and lack any structurally weak horizons that could act as detachment horizons. Thus, Cenozoic deformation within the batholith proceeded via the development of penetrative structures throughout most of the batholith, with only relatively small-scale thrust displacements on ductile shear zones.

The balanced cross sections of each transect (Figures 6.2 and 6.3) illustrate another implication for the change in structural style and structural relief along the eastern range front of ANWR. These balanced cross sections suggest that, although the total amount of shortening in each transect was similar (46-48%), the depth to the basal detachment surface was different for each transect. Specifically, the maximum depth to the basal detachment horizon beneath the Okpilak batholith was considerably deeper (14.6 km) than that under the Aichilik River transect (9.5 km) (Table 6.2). The footwall in northeastern ANWR has at least four different basal detachment levels, based on the interpretation of the regional and local structural geometries discussed in this dissertation and illustrated in the balanced cross sections. This footwall geometry is illustrated schematically in figures 6.6 and 6.7. In order to accommodate the change in depth to the basal detachment horizon between the two transects, a major lateral ramp in the footwall must connect the two different basal detachment levels. This lateral ramp probably is approximately parallel to the Jago River.

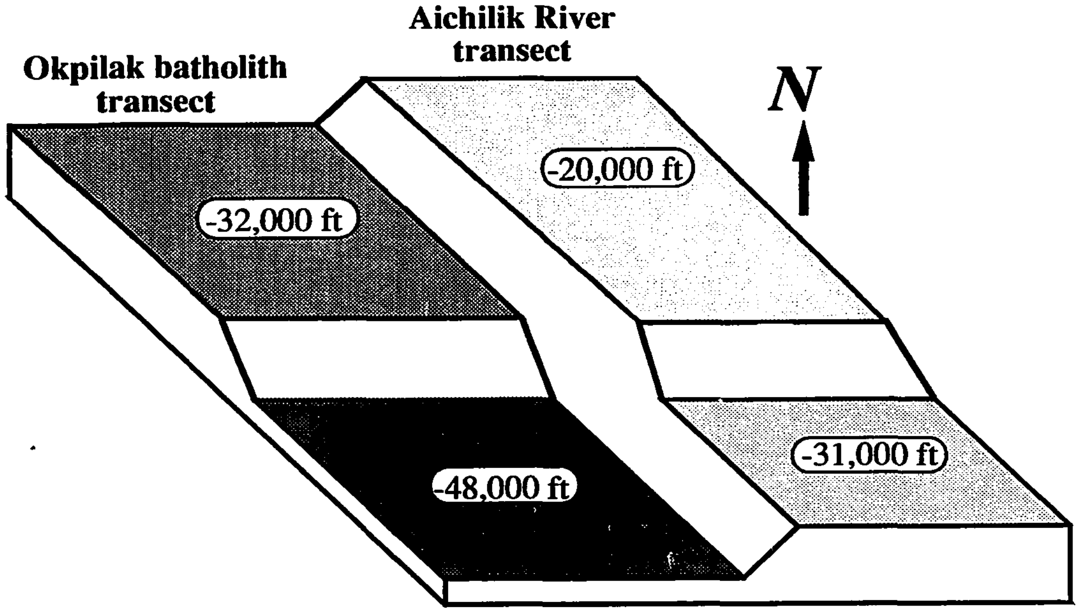


Figure 6.6. Schematic illustration of the footwall geometry of northeastern ANWR, as viewed from the southwest.

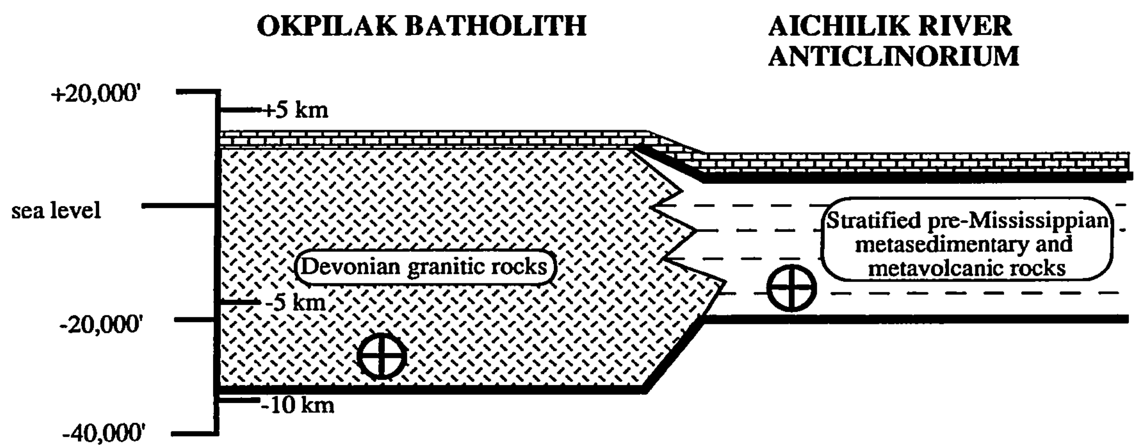


Figure 6.7. Schematic cross section along strike illustrating the relationship between the basal detachment surface of the Okpilak batholith transect with that of the northern Aichilik River transect. This cross section is drawn at the latitude of the batholith (C-C' on Figure 6.1), and illustrates the lateral ramp that probably connects the two basal detachments. The rocks in the hanging wall immediately over the footwall ramp probably have been modified by a high-angle fault with both variable dip-slip and strike-slip motion.

The abrupt east-west change in structural topography at the Jago River (as illustrated on the structure contour map, Figure 6.5 A) also would reflect the change in thickness of the hanging wall across the lateral ramp. The geometry of this lateral ramp in the hanging wall is probably more complicated than is illustrated in figure 6.7 due to variations in the amount and manner of Cenozoic shortening within the hanging wall across the ramp. Variations in the mode of Cenozoic shortening between the Okpilak and Aichilik River transects require that the regional hanging wall be modified by a high-angle fault of variable displacement that parallels and overlies the footwall ramp. No such fault disrupts the Ellesmerian rocks of the northern flank of the Aichilik River anticlinorium (Leffingwell Ridge), which continue relatively uninterrupted north of the batholith (Figure 6.1). This implies that displacement on this high-angle fault in the hanging wall terminates at the range front.

Although the change in depth to the basal detachment horizon explains the difference in structural relief within and between the two transects, the reason for the changes in the depth to the basal detachment horizon across the region remains uncertain. The balanced cross section across the batholith suggests that the presence of the batholith itself may have resulted in the activation of a second, deeper detachment horizon in the vicinity of the batholith. This may have been due to the higher competency of the granitic rocks and/or a lower geothermal gradient in the vicinity of the batholith at the time of thrusting. Isolated CAI values from the vicinity of the batholith suggest that the geothermal gradient in this area was higher than that of the surrounding region. However, these high CAI values also may reflect hydrothermal activity along the shear zones.

The change in lithology from granitic rocks in the west to stratified pre-Mississippian rocks in the east may also account for the lateral ramp between the northern portions of the two transects (Figure 6.7). However, the cause of the frontal footwall ramps in the central portions of both transects that correspond to the hanging wall ramps that define the topographic range front is unclear. The deep basal detachment surface beneath the batholith may have ramped to a shallower detachment level once the northern margin of the batholith was penetrated, resulting in a frontal footwall ramp (Figure 6.8).

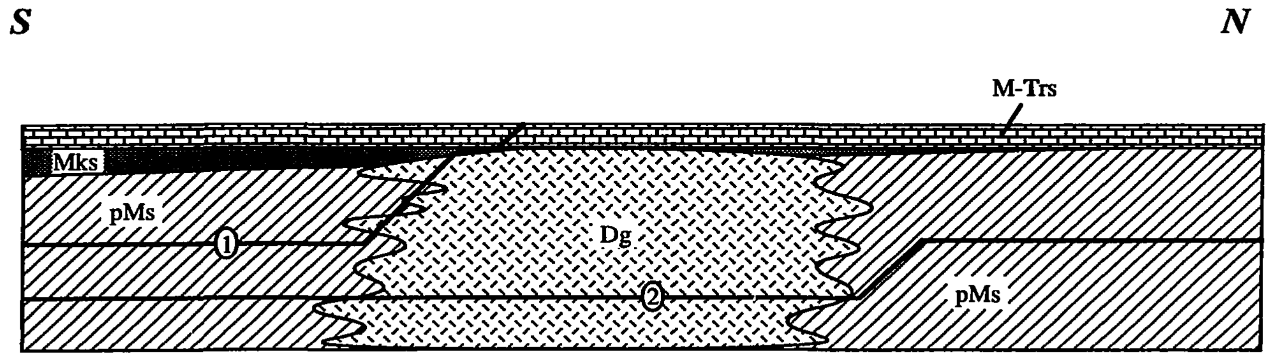


Figure 6.8. Schematic restored north-south cross section of the Okpilak batholith transect illustrating the possible influence of lithology on the location of footwall ramps. Fault 1 predates fault 2. Abbreviations used in this figure: pMs: pre-Mississippian sedimentary rocks; Dg: Devonian granitic rocks; Mks: Mississippian Kayak Shale; M-Trs: Mississippian through Triassic sedimentary rocks of the Ellesmerian sequence.

Stratified pre-Mississippian rocks exposed in the core of a small anticlinorium (Kikiktak Mountain, Figure 6.1) north of the batholith do suggest that the batholith ceases to be a major element in the pre-Mississippian rocks north of its present-day margin. However, this does not provide conclusive evidence that the frontal hanging wall ramp at the present-day northern margin of the batholith coincides with the original northern intrusive margin of the batholith.

Along the Aichilik River transect, there is no direct evidence of a major lithologic change within the pre-Mississippian rocks north of the topographic range front (Leffingwell Ridge, Figure 6.1). A slight decrease in structural elevation from the northern margin of the Mt. Greenough anticlinorium to the southern part of the Aichilik River anticlinorium suggests a minor hanging wall ramp associated with the volcanic rocks of Whale Mountain (Figures 6.2 and 6.9). A similar, albeit greater, change in structural relief from the northern margin of the Aichilik River anticlinorium to the coastal plain suggests that it too could be due to a hanging wall ramp. Whether the corresponding footwall ramp formed in response to a major change in lithology of the pre-Mississippian rocks (Figure 6.9) remains speculative since the lithology of the pre-Mississippian rocks in the subsurface of the coastal plain is presently unknown.

A better understanding of the relative age and timing of motion on these various basal detachments could help illuminate the reasons behind their varying depths and ramps. Ongoing apatite fission track studies in the region may eventually permit some speculation on the relative timing of movement along the various postulated detachment levels, but at present the age relationship between them remains uncertain.

In general, these observations and conclusions suggest that the lithology and rheology of depositional basement play a major role in its deformational style. This appears to be especially significant in the relatively shallow external regions of a mountain belt, such as the range front of the northeastern Brooks Range, where deformation occurs under relatively low pressure and temperature conditions.

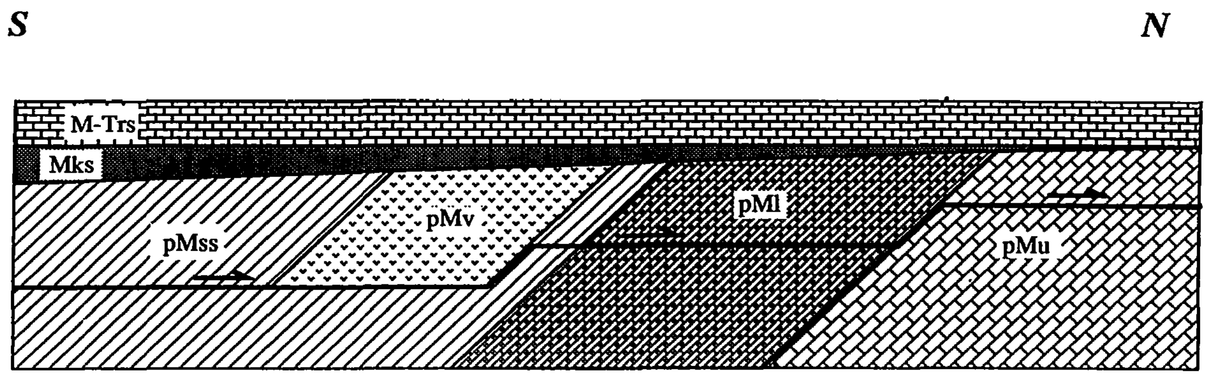


Figure 6.9. Schematic north-south cross section along the Aichilik River transect illustrating the possible linfluence of lithology on the location of footwall ramps as discussed in text. The abbreviations used in this figure are: pMss: pre-Mississippian sandstones and shales of the Mt. Greenough anticlinorium; pMv: pre-Mississippian volcanic rocks of Whale Mountain; pMl: pre-Mississippian limestones of the Aichilik River anticlinorium; pMu: unknown pre-Mississippian rocks of the subsurface of the coastal plain; Mks: Mississippian Kayak Shale; M-Trs: Mississippian through Triassic sedimentary rocks of the Ellesmerian sequence.

6.2.2 How has the stratigraphy of the Ellesmerian sequence influenced its deformational style?

The deformational style of the Mississippian and younger Ellesmerian sequence also varies along the length of the range front of northeastern ANWR and appears to correspond to changes in the stratigraphy of the regional detachment horizon near the base of the cover sequence, the Mississippian Kayak Shale. Along the Aichilik River transect, the Kayak Shale has acted as an effective detachment horizon during Cenozoic thrusting, permitting the entire overlying Ellesmerian sequence to deform independently of the underlying pre-Mississippian sequence and attached basal Ellesmerian conglomerate, the Mississippian Kekiktuk Conglomerate (Figure 6.2, Table 6.2). In the northern exposed portion of the Aichilik River transect at Leffingwell Ridge, the Mississippian through Triassic cover sequence appears to have shortened primarily via thrust duplication between detachment horizons in the Mississippian Kayak Shale and the Jurassic to Cretaceous Kingak Shale, with detachment folding playing only a minor role. This structural style appears to correspond to a relatively thin and silty Kayak Shale with a thick carbonate interval in its upper portions. Further south along the Aichilik River transect in the vicinity of Bathtub Ridge, detachment folding plays an increasingly important role in the Cenozoic shortening of the Ellesmerian sequence, corresponding to a thicker and less silty Kayak Shale lacking a significant carbonate interval.

In contrast, the Ellesmerian sequence along the Okpilak batholith transect exhibits a very different deformational style (Figure 6.3, Table 6.2). In the vicinity of the batholith, Cenozoic shortening within the Ellesmerian sequence has been via the development of penetrative mesoscopic structures and minor thrust faults and associated folds. This style of deformation is totally unlike that of the Ellesmerian sequence both to the east and to the west, but is very similar to the Cenozoic deformational style of the underlying batholithic rocks. This change in structural style of the Ellesmerian sequence probably can be best explained by the observation that the regional detachment horizon, the Mississippian Kayak

Shale, appears to be depositionally thin or absent in the vicinity of the batholith (Sable, 1977; Watts and others, 1988). Thus the Ellesmerian sequence remained structurally coupled to the batholith during Cenozoic thrusting and shared with the batholith the development of penetrative fabrics and mesoscopic structures.

The nature of the lateral change in structural style of the Ellesmerian sequence from thrust duplication in the east to penetrative deformation in the west over the batholith has not been addressed in this study. Qualitative observations along Leffingwell Ridge and along the southern limb of the Aichilik River anticlinorium suggests that there may be a gradual increase in the severity of penetrative deformation and/or the development of small-scale structures in the Lisburne Group westward toward the Jago River and the batholith (Sable, 1977; W. K. Wallace and K. F. Watts, pers. comm.). However, as in the case of the underlying pre-Mississippian rocks, differences in the style of deformation and the amount of local shortening of the Ellesmerian sequence of the two transects may also have resulted in an 'accommodation fault' of variable displacement overlying and parallel to the Jago River. Since most of the Ellesmerian sequence has been eroded between the batholith and the core of the Aichilik River anticlinorium, the existence of such a fault within the now-eroded Ellesmerian cover sequence is speculative.

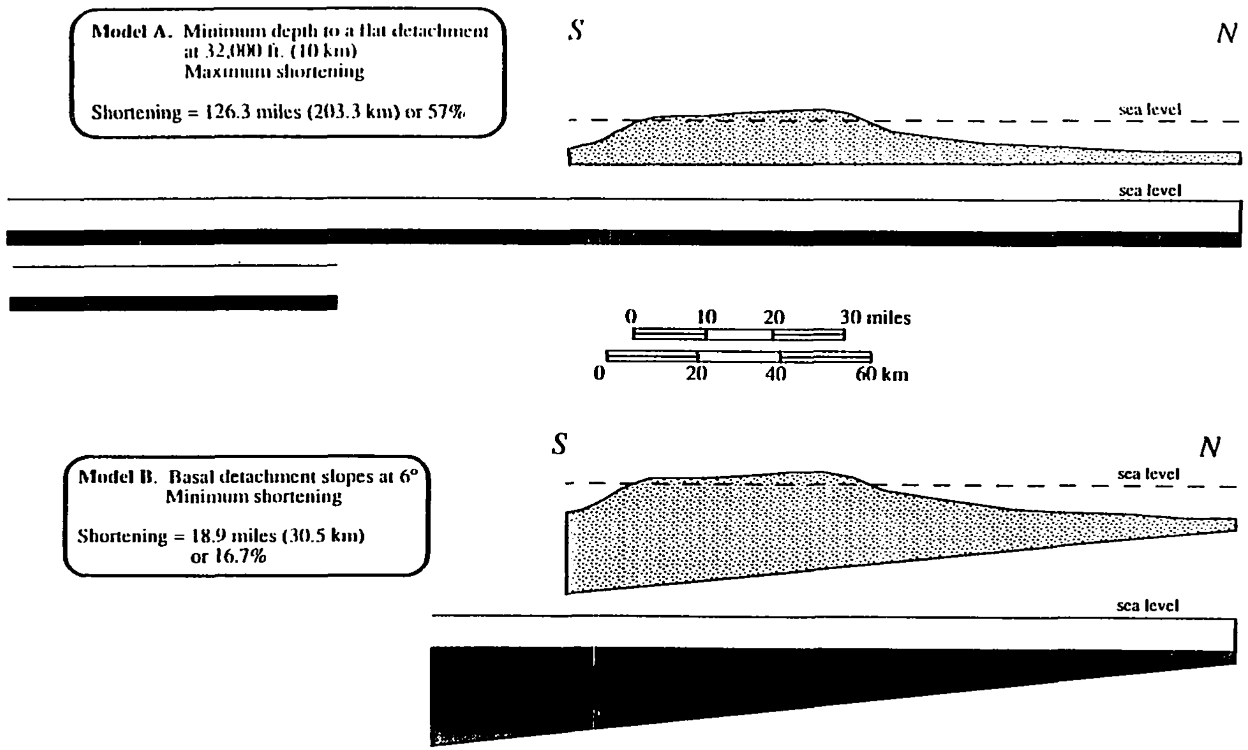
These observations confirm the importance of the character of regional detachment horizons in determining the structural style of the overlying stratified sequence during a thrusting event. Similar radical changes in structural style are associated with a major change in the stratigraphy of the detachment horizon in other foreland fold-and-thrust belts (e.g., the Keuper Salt horizon of the Jura Mountains, Laubscher, 1972; Bachmann and others, 1982). Loss of an effective regional detachment horizon can have a number of causes, such as structural truncation, non-deposition, erosion or lithologic changes due to stratigraphic variations. In the instance of the Kayak Shale of the northeastern Brooks Range, the batholith probably was a topographic high during deposition of the Kayak Shale, resulting in little or no shale deposition in the vicinity of the batholith (Watts and others, 1988). Northward onlap of the Mississippian Kayak Shale on paleotopography on the sub-Mississippian unconformity in the Aichilik River area may also have led to a stratigraphic thinning and coarsening of the Kayak Shale towards the north and northeast, along with the development of extensive carbonates in its upper portions. Both of these stratigraphic changes in the Kayak Shale appear to have resulted in major changes in the Cenozoic structural style of the overlying Ellesmerian sequence. Stratigraphic studies to document further these changes in the Kayak Shale are now in progress (LePain, pers. comm.).

6.2.3 Use of balanced cross sections in areas of limited surface and subsurface data

Historically, balanced cross sections have been constructed in areas with abundant surface and subsurface data (e.g., Bally and others, 1966; Price, 1981). These cross sections are generally viewed as a well-constrained 'best-fit' geologic solution and as such are often one of the end products of a study. However, in many remote and/or little studied areas, the detailed structural data needed for the construction of balanced cross sections are limited, may be poor in quality or totally lacking. In these regions, constructing a balanced cross section may be more useful as a method of developing and evaluating alternative structural solutions for an area, rather than as a summary of a definitive geologic 'solution.' In an area with little data, balanced sections may also help define which of the many unanswered questions are most critical in determining the best structural solution for the region.

Because both surface and subsurface data are limited in northeastern ANWR, balanced cross sections were constructed at two different scales for both transects during the course of this study. A series of generalized area-balanced models were constructed for each transect (e. g. Figure 6.10, Appendix B). As outlined in chapters 3 and 5, for each transect the topography of the sub-Mississippian unconformity surface along the transect and the depth to the basal detachment horizon at the pin line were held constant in all the models. The dip of the basal detachment surface south of the pin line, the depth to the

Figure 6.10. Examples of generalized area-balanced structural models for the Okpilak batholith transect. The structural topography of the upper surface of the wedge (the sub-Mississippian unconformity), the depth to the basal detachment surface at the pin-line, and the location of the pin-line are held constant for all of the models illustrated here. The dip of the basal detachment surface (with respect to the unconformity surface), the depth to the brittle/ductile transition, and the presence of ramps in the basal detachment surface vary from model to model.


Model A. Minimum depth to a flat detachment at 32,000 ft. (10 km)
Maximum shortening
Shortening = 126.3 miles (203.3 km) or 57%
S
N
sea level
sea level
0 10 20 30 miles
0 20 40 60 km
Model B. Basal detachment slopes at 6°
Minimum shortening
Shortening = 18.9 miles (30.5 km) or 16.7%
S
N
sea level
sea level

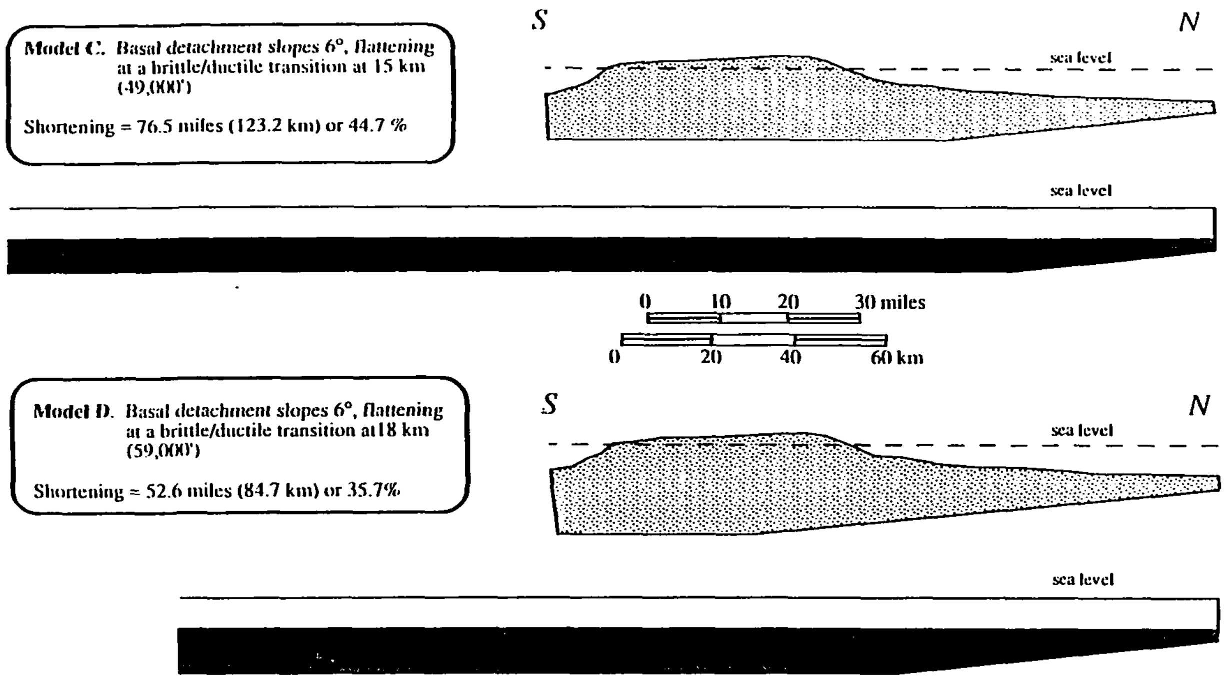


Figure 6.10. (cont.)

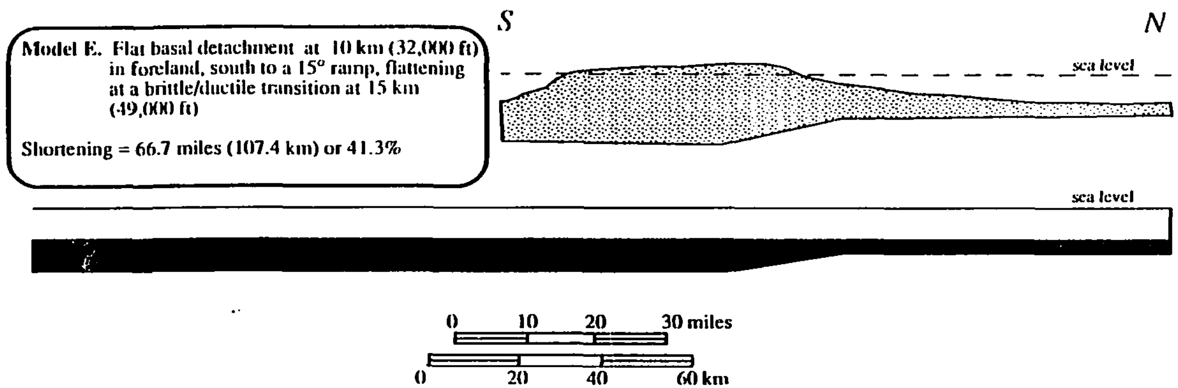


Figure 6.10. (cont.)

brittle/ductile transition (where the basal detachment might be expected to flatten), and the presence of ramps in the basal detachment surface were varied between models. By varying these parameters, a variety of different wedge geometries and their corresponding amounts of total shortening was examined for each transect. The most reasonable model was then chosen as a basis for construction of a more detailed balanced cross section of each transect (e.g., Figures 6.2 and 6.3). Each balanced cross section incorporated the available detailed surface geologic data from that transect, as well as my interpretation of the Cenozoic deformational mechanism of the pre-Mississippian rocks. This interpretation was based on my detailed field studies in portions of each transect (see chapters 3 and 5).

Construction of the area-balanced models points out several regional factors that influence the overall geometry of the fold-and-thrust belt in northeastern ANWR. The most important factor appears to be the geometry of the basal detachment surface, specifically, its slope, the depth at which it flattens (if it does) and the presence or absence of any ramps. Geologic influences on the geometry of this basal surface could include:

The geothermal gradient at the time of deformation. The geothermal gradient at the time of thrusting could control the depth to the brittle/ductile transition, which in turn may influence the depth at which the basal detachment horizon flattens (Chen and Molnar, 1983; Suppe, 1985).

The lithology of the rocks involved in thrusting. Lithology could influence both the geothermal gradient and the depth to the brittle/ductile transition (Suppe, 1985). Abrupt lateral changes in lithology could also result in ramps in the basal detachment surface.

Nature of the basal detachment horizon. If the basal detachment horizon is limited to a specific lithologic horizon or boundary (such as a shale horizon or a boundary between basement and suprabasement rocks), the geometry of the basal detachment horizon could be controlled by the geometry or stratigraphy of that horizon.

Unfortunately, the nature of most of these factors is unknown in northeastern ANWR. Indirect lines of evidence, such as the observed surface geometry of the wedge, must be used to infer the geometry of the basal detachment surface. However, these simple models do emphasize several areas of potentially fruitful study in little known regions such as the northeastern Brooks Range, such as the geothermal gradient at the time of thrusting and the composition and structure of the upper and middle crust.

In addition, these simple area-balanced models also help determine the possible range of Cenozoic shortening in northeastern ANWR, regardless of the specific model used for the Cenozoic deformational style within the pre-Mississippian rocks. This information could be useful in evaluating regional structural and tectonic models, such as those for the opening of the Arctic basin. Shortening values also can be useful in paleogeographic reconstructions of the Mississippian through Lower Cretaceous Ellesmerian passive continental margin and/or modeling of the pre-deformational foreland basin geometry.

Thus, where constraints are few, use of simple area-balanced models in a foreland fold-and-thrust belt can both provide valuable information regarding the possible range of tectonic shortening and help define those questions that need to be addressed in order to develop a better constrained structural interpretation of the fold-and-thrust belt. Using one of these simple models as a base, a more detailed balanced cross section can be constructed that illustrates the preferred structural interpretation of the region.

6.2.4 Implications of this study for pre-Mississippian stratigraphy, structural style and the evolution of the circum-Arctic region.

Pitfalls abound when discussing the potential significance of the pre-Mississippian rocks of the northeastern Brooks Range in the stratigraphic and structural evolution of the circum-Arctic region. The age of the lower Paleozoic and Proterozoic rocks in the northeastern Brooks Range is poorly constrained in most parts of the region, and the stratigraphic relationship of the various lithologic and structural packages complicated by both pre-Mississippian and Cenozoic deformational events. As part of this study, I developed a working interpretation of the structural stratigraphy for the stratified pre-Mississippian rocks of a local area of the northeastern Brooks Range--south of Leffingwell Ridge in the core of the Aichilik River anticlinorium (Figure 6.1). Although this stratigraphy has no age control, it may provide a framework for further stratigraphic studies in the pre-Mississippian rocks in nearby parts of the northeastern Brooks Range, and a basis to begin correlating these rocks with similar pre-Mississippian rocks in northwestern Canada.

As discussed in chapter 1, the pre-Mississippian structural history of the northeastern Brooks Range is poorly understood. As suggested by Norris and Yorath (1981), evidence of several different pre-Mississippian deformational events may be preserved in different pre-Mississippian stratigraphic sequences. The number of pre-Mississippian deformational events and their structural style and sense of vergence remain poorly understood, although at least one detailed study in the Franklin Mountains of western ANWR has suggested that Devonian deformation was penetrative and south-vergent (Oldow and others, 1987a).

In the Aichilik and Egaksrak rivers area of northeastern ANWR, the stratified pre-Mississippian rocks of the Aichilik River anticlinorium have undergone a pre-Mississippian event of unknown age that is interpreted in this study to have resulted in north-vergent thin-skinned thrusting, related folding and development of penetrative axial planar cleavage. This deformation resulted in the D1 structures discussed in chapter 3. No south-vergent, pre-Mississippian-age structures were observed in the pre-Mississippian rocks in this area. How can this observation of north-vergence in the pre-Mississippian rocks of northeastern ANWR be reconciled with the south-vergent observed structures in pre-Mississippian rocks of western ANWR?

One possible explanation for the apparent contradiction is that the vergence and/or age of one or the other of the pre-Mississippian structural events were misinterpreted. For example, the north-vergent pre-Mississippian structures of northeastern ANWR actually could be Cenozoic in age, or the south-vergent structures of western ANWR might not really be south-vergent. However, it seems more likely that both sets of observations are correct, and that a geologic explanation for the apparent contradiction should be sought.

The absolute age of the pre-Mississippian deformational events in these two parts of ANWR is poorly constrained. This leaves open the possibility that there were two distinct pre-Mississippian events, but structures related to the two events were not equally well preserved everywhere in ANWR. This opens a wide variety of possible scenarios for consideration. For example, if the pre-Mississippian rocks of the Aichilik River anticlinorium are Proterozoic in age, as suggested by Reiser and others (1980), a north-vergent deformational event could have predated a south-vergent Devonian event. In this scenario, the Devonian event may have been non-penetrative in northeastern ANWR, so that few mesoscopic structures were preserved. Lack of age control and a well-defined stratigraphy, or overprinting by the north-vergent Cenozoic event may have obscured any south-vergent map-scale structures.

Alternatively, the south-vergent deformation documented by Oldow and others (1987a) in the Franklin Mountains could actually be pre-Devonian in age and predate a penetrative north-vergent Ellesmerian orogeny. If so, all evidence of this pre-Devonian south-vergent event has been obliterated in the Aichilik River area by the later north-vergent Ellesmerian event, since the only structures preserved there are north-vergent.

A third possibility is that the Devonian Ellesmerian orogeny was two-sided, with both north and south vergence, and was subsequently telescoped during Cenozoic shortening. This interpretation could explain the predominance of north-vergence in the northern portions of the northeastern Brooks Range (e.g., Sadlerochit and Shublik Mountains (McMullen, 1989 and Wallace, pers. comm.) and Aichilik River anticlinorium (this study)) and presence of south-vergent structures in the southern portions of the range (e.g., Franklin Mountains (Oldow and others, 1987a) and possibly the southern part of Mt. Greenough anticlinorium (pers. obs.)). I personally find this explanation appealing, but have no concrete evidence to support it over the others.

Finally, the south-vergent structures documented by Oldow and others (1987a) in the Franklin Mountains could represent local south-vergence in an overall north-vergent orogen of either early Paleozoic or Proterozoic age. This would explain why south-vergence appears to be preserved only in certain areas. However, Oldow and Avé Lallemant have found more evidence of south-vergence in other parts of the range (pers. comm.) which makes this possibility seem less plausible.

Obviously, my observations on the pre-Mississippian structural style of northeastern ANWR have not provided conclusive evidence for the structural style and vergence of all pre-Mississippian deformational events in the northeastern Brooks Range. Quite to the contrary, this study seems to have raised more questions than it answered with respect to the stratigraphy and structural history of the pre-Mississippian rocks. In order to unravel the complex pre-Mississippian structural history, we will need a better understanding of the pre-Mississippian stratigraphy, more and better age control and more detailed studies focussing on the structural history of these polydeformed rocks throughout the range. Hopefully this study will provide a starting point for such work.

6.3 The remaining unanswered questions

Because the scope of this dissertation is necessarily limited, several regional questions posed in chapter 1 remain unanswered. First, what can this study reveal regarding the driving mechanism of deformation in the northeastern Brooks Range and its relationship to the main axis of the Brooks Range? Involved in this question is the origin of the arcuate regional trends in the northeastern Brooks Range in ANWR and northwestern Canada. Secondly, can general implications for the structural evolution of a mountain belt be learned from this unusual example of a mountain belt that has formed on a narrow continental fragment bounded by two opposing passive continental margins of different ages (e.g., Figures 6.11 and 6.12)?

In isolation, this dissertation is not broad enough in scope to answer either of these

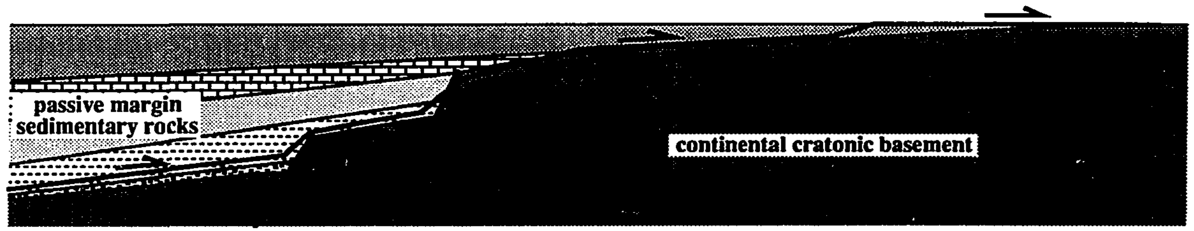


Figure 6.11. Schematic cross section of a 'classic' continental margin prior to thrusting. In this situation, passive continental margin sediments are incorporated into the orogenic wedge. These frequently deeper-water sediments are thrust onto thinner and shallower-water sediments of similar age that were deposited higher up on the continental margin, sometimes on cratonic basement. Depositional thinning of the passive margin sedimentary rocks towards the continental interior often aids in the development of the characteristic wedge-shape of a classic orogenic belt. The orogenic sole fault becomes progressively shallower towards the foreland, where it either intersects the surface and/or loses slip and dies out in foredeep deposits.

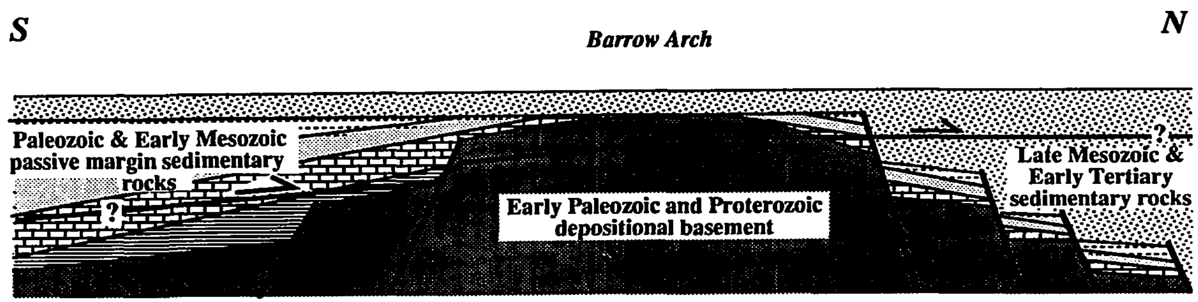


Figure 6.12. Schematic cross section of a continental fragment with two passive continental margins of different ages and opposite orientations prior to a collapse by thrusting. Passive margin sediments of one margin may be thrust onto the shallow water sediments related to the second passive margin. This model may represent the northeastern Brooks Range. North-derived sedimentary rocks of a Paleozoic and Early Mesozoic passive continental margin are preserved on the south side of the Barrow Arch, and possibly in isolated basins on the north side of the arch. Clastic rocks of the younger Cretaceous passive continental margin were derived from the south and thicken rapidly on the north side of the arch. Cenozoic thrust faulting in the northeastern Brooks Range post-dates formation of both margins.

questions specifically. With respect to the tectonic origin of the northeastern Brooks Range and its relationship to the main axis of the Brooks Range, the transect approach used in this study could be applied to future structural, geophysical and/or geochronological studies. For example, in order to address the question of the driving mechanism of deformation in the northeastern Brooks Range, it would be necessary to determine the direction of tectonic transport throughout the entire northeastern Brooks Range and how that direction may have changed with time, as well as the nature, depth and regional behavior through time of the orogenic basal detachment horizon. Two obvious areas where transects would be useful are in the far western parts of ANWR, where the salient of the northeastern Brooks Range meets the main axis of the Brooks Range, and in northwestern Canada, where the structural trends are north-south. Specific studies that should be incorporated into these transects include: 1) a detailed study of mesoscopic structures and strain in the Ellesmerian sequence in order to determine regional variations in the tectonic transport direction; 2) detailed structural, stratigraphic and metamorphic studies of the pre-Mississippian sequence in order to better understand the pre-Cenozoic history of the pre-Mississippian rocks, and possibly to constrain the nature and composition of the basement underlying them; 3) apatite fission track and other geochronologic studies of the both the Ellesmerian sequence and pre-Mississippian rocks of each transect in order to determine the timing of uplift of the major structures and, by inference, the timing of motion on the basal detachment; and 4) other thermal history studies, such as vitrinite reflectance and conodont alteration, aimed at determining the geothermal gradient at the time of Cenozoic deformation.

These detailed studies would also provide additional information that could aid in answering the second question, i.e., how has the structural and tectonic evolution of the northeastern Brooks Range fold-and-thrust belt and associated foreland basin been influenced by the fact that this Cenozoic thrust belt has developed on a narrow continental fragment bounded by two rifted margins? Specifically, what are the consequences of a fold-and-thrust belt migrating across a passive continental margin that faces in the direction of tectonic transport (e.g., Figure 6.12)? Modeling the structural and depositional evolution of the foreland basin would be particularly relevant in addressing this question. Such modeling would require compilation and analysis of geophysical information and stratigraphic data from the subsurface of the coastal plain north of the range front. Although this information is both sparse and largely unavailable to the public at the present time, the result of such a study could provide valuable insight into the influence of the composition, thickness and dimensions of the continental crust on the evolution of the superimposed fold-and-thrust belt and its associated foreland basin.

6.4 Conclusions

The Cenozoic fold-and-thrust belt of the northeastern Brooks Range provides a good example of the important role played by lithology and structural stratigraphy in the structural evolution of a foreland fold-and-thrust belt. Although regional amounts of shortening are similar for the two transects in this study, the Cenozoic structural style exhibited in each is quite different. In the pre-Mississippian rocks of the 'depositional basement,' stratified rocks of the Aichilik River transect deformed primarily by thrusting and related folding, while lithologically homogeneous granitic rocks of the Okpilak batholith transect deformed primarily by penetrative strain. In the Mississippian through Triassic cover rocks, the structural style during Cenozoic deformation apparently was largely controlled by the thickness and lithologic character of the regional detachment horizon near the base of the cover sequence, the Mississippian Kayak Shale. Where the shale is thin or absent, the cover shared the same deformational style as the underlying 'depositional basement.' Where the shale is thick and shaly, the overlying cover sequence deformed by detachment folding and thrusting. Where the shale is thin, silty and has thick carbonate horizons in its upper part, the cover sequence deformed mainly by thrust duplication.

This study also illustrates how simple area-balanced models can be used to evaluate the regional shortening and geometry of a fold-and-thrust belt. This technique could be especially useful in remote or little studied areas where structural data are sparse or poor in quality. In such areas, models can be used to develop and evaluate a variety of regional structural solutions. A model can then be selected as a basis for a more detailed balanced cross section illustrating the preferred structural interpretation of the region.

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APPENDIX A:
DESCRIPTION OF STRATIGRAPHIC AND MAP UNITS

The following map and stratigraphic units were recognized in the field and used during mapping (see Hanks 1987, 1988, 1989 and Plate 1 for maps). Not all of the map units are relevant to the overall structural interpretation of the area and therefore have not been described in great detail in the text of the dissertation. The following descriptions and thicknesses are based primarily on my personal observations, with supplementary information from Reiser and others (1980). Additional and more detailed descriptions of the rocks exposed east of the Okpilak batholith can be found in Reiser and others (1980), Sable (1977), Armstrong and Mamet (1975) and Detterman and others (1975).

AICHILIK AND EGAKSRAK RIVERS AREA

Quaternary alluvium

Primarily colluvium: poorly sorted rock debris, sand, ailt, and clay. Local terrace deposits of moderately-sorted gravels.

Mississippian through Cretaceous rocks

Figure A.1 summarizes the Mississippian through Cretaceous stratigraphy of the Aichilik and Egaksrak rivers area.

Jurassic and Cretaceous Kingak Shale.

Black, fissile shales with occasional thin beds of siltstone and fine-grained sandstone. Only locally exposed north of Leffingwell Ridge and along the Aichilik River. Highly internally deformed, and total thickness undetermined.

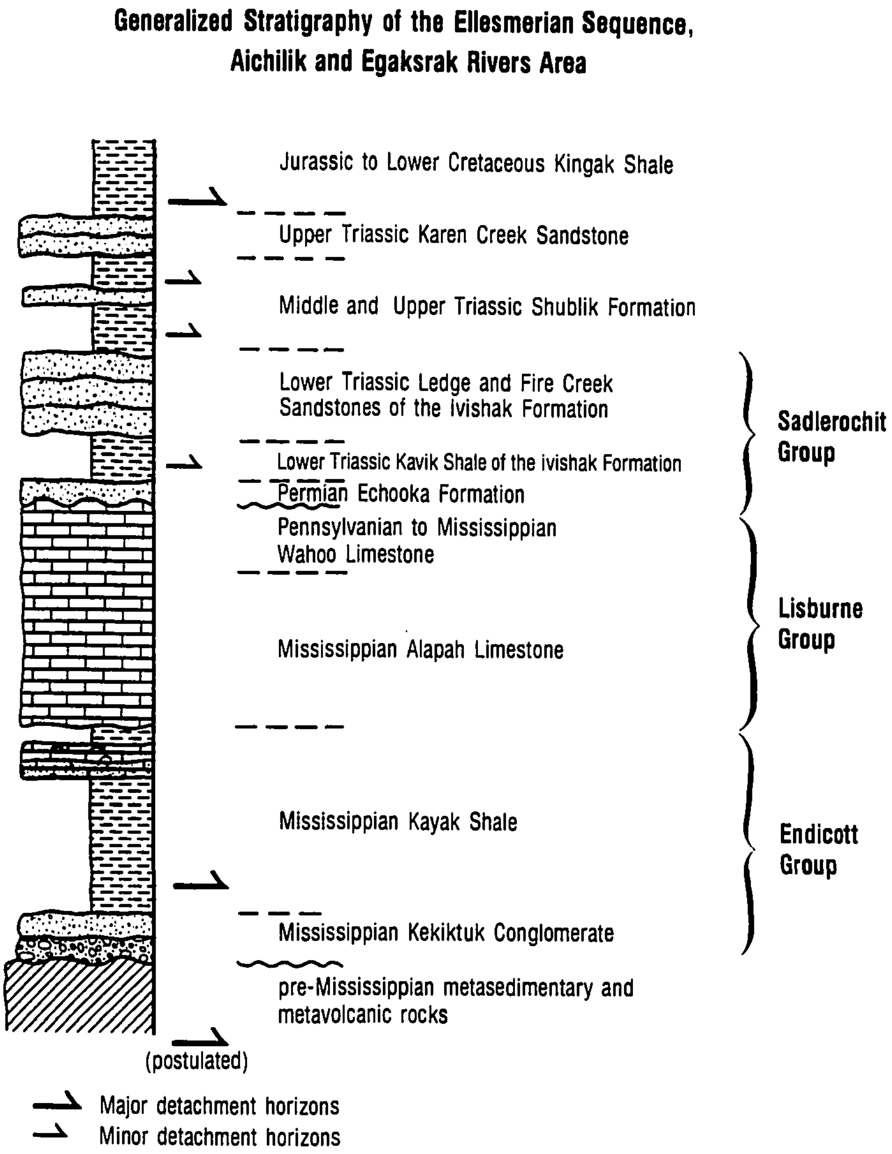


Figure A.1. Generalized stratigraphy of the Ellesmerian sequence, northeastern Brooks Range. Not to scale.

Triassic Karen Creek Sandstone.

Dark gray-weathering, sooty gray, fine- to medium-grained quartz lithic phosphatic sandstone. Highly bioturbated with few preserved sedimentary structures. Occasional hummocky cross stratification. Approximately 30 to 45 meters thick along Leffingwell Ridge.

Triassic Shublik Formation.

Black, thinly bedded, phosphatic shale, siltstone, fine-grained sandstones and minor black fossiliferous limestones. Locally internally deformed. Possibly up to 150 meters thick along Leffingwell Ridge.

Triassic Shublik sandstone.

Dark gray-weathering, gray medium- to fine-grained, phosphatic sandstones with interbedded black shales. Individual sandstone beds up to 50 cm thick and highly bioturbated. This sequence of sandstones appears to occur in the middle of the Shublik Formation and is generally less than 15 meters thick. Well exposed and an effective map unit west of the Egaksrak River.

Triassic Ledge Sandstone/Fire Creek Siltstone (Undifferentiated) Members of the Ivishak Formation.

Tan- to orangish tan-weathering, fine- to medium-grained pyritic quartz sandstone and siltstone, with local grit and pebble conglomerates. Massive with few sedimentary structures, except for local low-angle crossbedding. Bedding generally <.5 meters thick. Forms prominent ridges of frost-riven rubble. Approximately 65 to 90 meters thick immediately west of the Aichilik River.

Triassic Kavik Member of the Ivishak Formation.

Brown, thinly bedded siltstone and shale. Poorly exposed in study area. Approximately 130 meters thick along Leffingwell Ridge.

Permian Echooka Formation.

Reddish-brown, thinly-bedded (<40 cm) calcareous bioclastic limestones, calcareous sandstones and bioturbated siltstones. Approximately 50-100 meters thick along Leffingwell Ridge.

Pennsylvanian to Mississippian Wahoo Formation.

Light gray- to buff-weathering skeletal and oolitic grainstones and bryozoan packstones. Contains prominent orange-weathering, dolomitized horizons toward upper part of section. Approximately 200 meters thick along Leffingwell Ridge.

Mississippian Alapah Formation.

Light gray-weathering peloidal packstones, with local cross-bedded sandy grainstones near base. Commonly forms rubble-covered talus slopes. Approximately 400 meters thick along Leffingwell Ridge.

Mississippian and Pennsylvanian Lisburne Group (Undifferentiated)

Massive limestones of the Mississippian Alapah Formation and Pennsylvanian Wahoo Formation that cannot be differentiated because of poor exposure. Total thickness approximately 600 meters.

Mississippian Endicott Group carbonates.

In the north along Leffingwell Ridge, thin-bedded skeletal grainstones with locally well-developed coral boundstones. Can be very fossiliferous, with abundant corals, brachiopods and crinoids. Along Leffingwell Ridge, thickness varies laterally from < 20 meters to > 100 meters. On the south limb of the Aichilik River anticlinorium between the Aichilik and Egaksrak rivers, consists of orange-weathering, moderately well-bedded (<.5 meters) calcareous sandstones and sandy limestones. Thickness laterally variable from < 50 meters to > 100 meters. In this area, this unit was considered part of the Itkilyariak Formation by Mull and Mangus (1972). Both along Leffingwell Ridge and the south limb of the Aichilik River anticlinorium, contact with the underlying Kayak Shale is gradational. Where the upper part of the carbonate interval is well exposed, it is overlain by a black shale horizon, which is in turn overlain by the Alapah Limestone. See LePain and Crowder (1991) for a more detailed descriptions.

Mississippian Endicott Group--Kayak Shale.

Black, fissile and locally highly deformed shales and siltstones with minor thin-bedded (<30 cm), black-weathering, grey fine- to course-grained quartz lithic sandstones. Local abundant macerated plant fragments. Sequence has a higher percentage of siltstone along Leffingwell Ridge. Along the eastern end of Leffingwell Ridge, at Redwacke Creek, the Kayak Shale is less than 200 meters thick (personal observation), though part of this thickness is probably due to structural thickening. On the southern limb of the Aichilik River anticlinorium between the Egaksrak and Aichilik rivers, the Kayak Shale is considerably more shaly and has significantly less siltstone. Total thickness in parts of this area is greater than 250 meters, but this thickness probably also incorporates structural thickening. For more detailed descriptions, see LePain and Crowder (1991).

Mississippian Kekiktuk Conglomerate.

White- to light grey-weathering, white to grey quartz and chert pebble and granule conglomerates and coarse sandstones. Bedding thickness and geometry highly variable. Thickness highly variable, ranging from 0 to 30 meters.

Pre-Mississippian rocks

Figure A.2 summarizes the overall structural stratigraphy of the pre-Mississippian rocks of the Aichilik and Egaksrak rivers area. In the following descriptions, the general

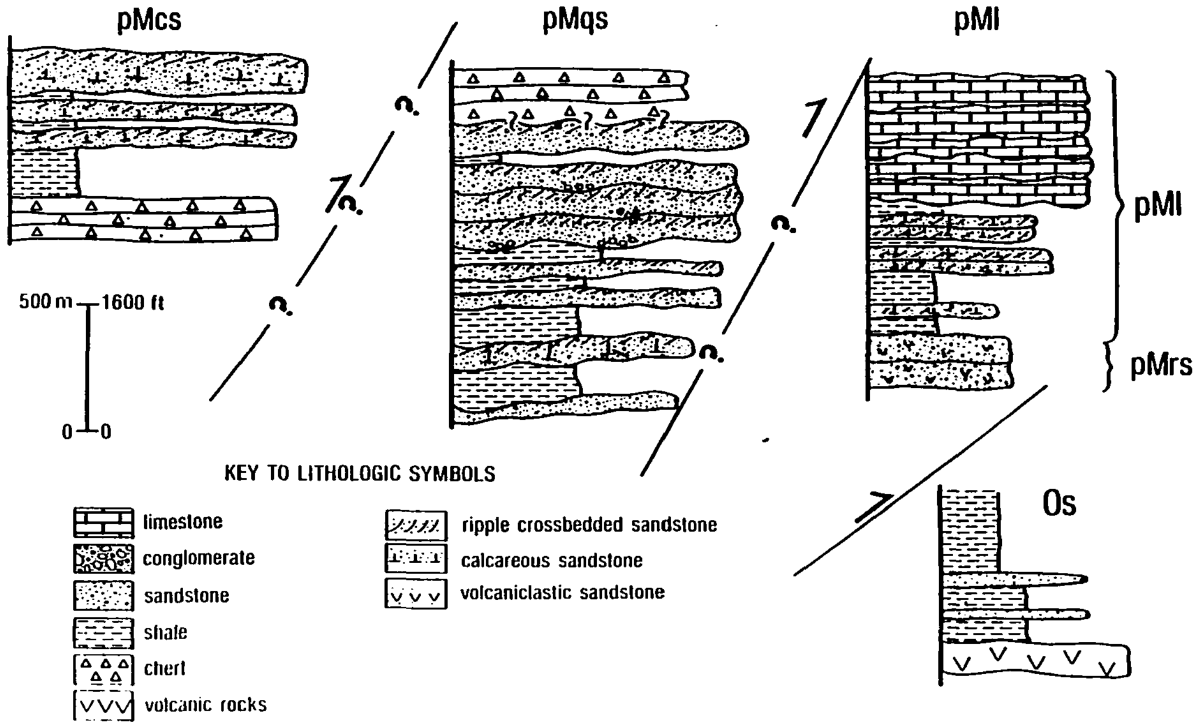
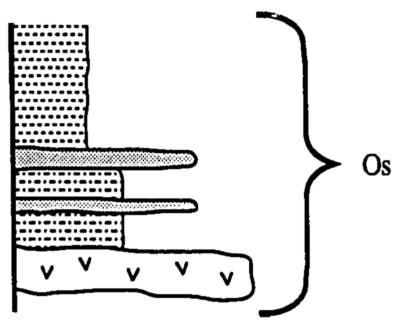


Figure A.2. Generalized stratigraphy of the pre-Mississippian rocks of the Aichilik and Egaksrak Rivers area, northeastern Brooks Range.

pre-Mississippian units used in the discussion of the structural geology of the area in chapter 3 have been further subdivided into the various map units used in the original mapping (see Hanks, 1989). In many instances, these units correspond to a unit recognized by Reiser and others (1980) and is noted. Thicknesses are only approximate--there has been much structural disruption of the sequence and structural thickening is likely. These units are described in a possible age succession, with the youngest rocks described first. However, this age succession is not supported by fossil evidence in the field area and is based on the consistent order of vertical successions of units observed in the field, and correlations with similar sequences described by Reiser and others (1980) and Lane and others (1991) where there is some age control.

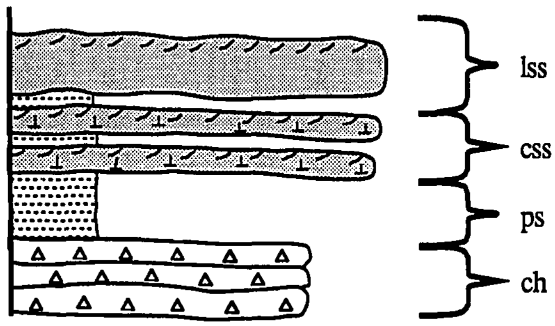
Ordovician siltstone and shales (Os)



Os--Ordovician sedimentary rocks

Black shales, silicified siltstones and fine-grained sandstones mapped as Os by Reiser and others (1980). Sandstones are relatively thin (<30 cm thick), are pyritic and have bioturbated bedding surfaces. Siliciclastic sedimentary rocks are underlain by highly altered mafic volcanic agglomerates. Total thickness of this unit is unknown due to structural disruption, but probably exceeds 100 meters. This unit is associated with maroon and green phyllites (ph of Reiser and others, 1980) that has been identified as Ordovician in age (Lane and Cecile, 1989 and Lane and others, 1991) along the USA/Canada border.

Pre-Mississippian calcareous siliciclastic rocks (pMcs)



lss--lithic sandstones.

Moderate- to thickly-bedded (>20 cm) grayish-black weathering fine- to medium-grained lithic sandstones interbedded with black shales. True thickness indeterminate due to probably structural thickening, but probably greater than 200 meters. Probably equivalent to 'Cs' as mapped by Reiser and others (1980). Considered Early Cambrian in age, but true age not well-documented. Probably early Paleozoic in age.

css--calcareous quartz sandstones.

Thin- to moderately-bedded (<50 cm) tan-weathering and fine-grained quartzose sandstones. Commonly ripple-laminated, but locally has scoured bases and fine upwards. Local rare trace fossils preserved on bedding planes. Interbedded with highly deformed black shales. Thickness of black shales generally <50 cm. Thickness of the entire unit is indeterminate due to structural thickening but probably greater than 200 meters. In map area, possibly equivalent to Css and pCn of Reiser and others (1980). True age not well-documented, probably early Paleozoic.

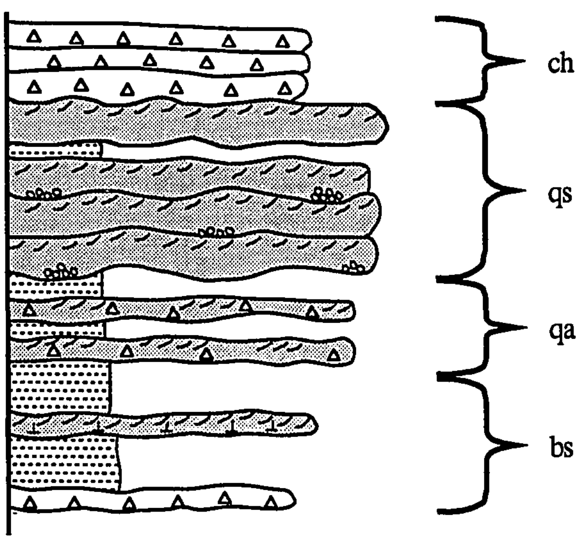
ps--black phyllitic shales.

Black phyllitic shales with thin intervals of fine-grained, dark grey lithic sandstones and local pebbly mudstones. Highly deformed and structurally disrupted, and thickness indeterminate. Probably equivalent to 'Cp' by Reiser and others (1980). Early Cambrian age not well-documented, but probably early Paleozoic in age.

ch--cherts.

Thin-bedded (10-15 cm), grey, black and white cherts. Estimated thickness <100 meters, but may be more or less due to extreme structural disruption. Possibly equivalent to 'Ccp' of Reiser and others (1980) and the same unit as 'ch' recognized in the pre-Mississippian quartzose sandstone (pMqs) described above. True age not documented, possibly early Paleozoic or Proterozoic.

pre-Mississippian quartzose sandstones (pMqs)



ch--cherts.

Thin-bedded (10-15 cm), grey, black and white cherts. Estimated thickness <100 meters, but may be more or less due to extreme structural disruption. Possibly equivalent to 'Ccp' of Reiser and others (1980). True age not documented, possibly early Paleozoic or Proterozoic.

qs--quartz sandstones.

Greenish-brown weathering, fine- to coarse-grained quartz lithic sandstones. Well-bedded, with beds generally <50 cm thick. Individual beds locally have scoured bases andfine upwards. Locally ripple-laminated. Interbeds of maroon and tan slates have occasional bioturbation on bedding surfaces. These slates are generally thin (<1meter) but locally exceed 10 meters in outcrop thickness Estimated thickness of entire quartz-lithic sandstone sequence is greater than 275 meters, but may be more or less due to structural complications. Mapped in part as 'pCn' by Reiser and others (1980). True age unknown, possibly early Paleozoic.

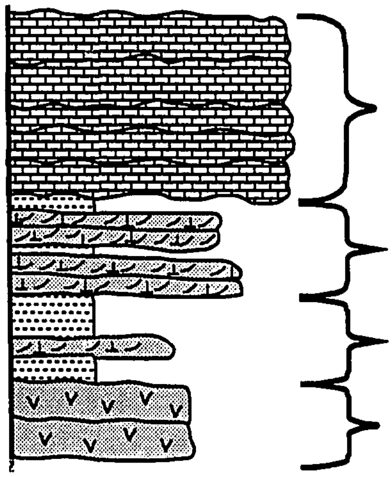
qa--cherts with interbedded siltstones and shales.

'Cherts' are thin-bedded (<30 cm) white- and pink-weathering silicified quartzites (?) with thin, maroon-weathering, ripple-laminated fine-grained sandstones and siltstones at tops of beds. Interbedded tan-weathering slaty shales are <50 cm thick. Estimated thickness of total sequence is <100 meters, but may be more or less due to structural complication. Possibly early Paleozoic in age. Locally, possibly equivalent to 'pCpq' of Reiser and others (1980).

bs--brown-weathering shales and argillites.

Brown-weathering shales and argillites. Locally contains thin- to moderately- bedded (<30 cm) intervals of ripple-laminated calcareous limestones and local maroon and green argillites with thin green cherts. Highly deformed, thickness indeterminate. Possibly equivalent to 'pCpa' and 'pCal' of Reiser and others (1980). Age uncertain--possibly early Paleozoic or Proterozoic.

Pre-Mississippian carbonates (pMl)



bl--black massive bedded carbonates.

Black-weathering black limestones and dolomites that form prominent ridges in the western part of the Aichilik River anticlinorium. These carbonates are thickly bedded, with individual beds up to 2 meters thick. Local relict textures suggest that these rocks were originally pelloidal and/or oncolitic, but now commonly intensely recrystallized with extensive networks of calcite-filled fractures. Sedimentary structures include large-scale cross bedding (10-20 cm high) and local partial Bouma sequences. Individual beds limestones and dolomites commonly occur as thick amalgamated sequences of similar beds with no intervening shales. Approximately 80 to 170 meters thick. Mapped as 'pCl' by Reiser and others (1980). True age unknown, possibly early Paleozoic or Proterozoic.

<u>rl--rippled-laminated sandy limestones and calcareous sandstones.</u>

Tan- and orangish-weathering, thin-bedded rippled calcareous clastic rocks with thin interbeds of black shale. Quartz is dominant detrital component. Local beds of massive recrystallized limestone. Often highly deformed, estimated thickness less than 200 meters. Mapped as 'pCls' and 'pClr' by Reiser and others (1980). True age unknown, possibly early Paleozoic or Proterozoic

<u>ba--brown, black and tan argillites and shales</u>.

Brown, black and tan argillites and shales with occasional thin beds of ripple-laminated, tan sandy limestones. Highly deformed, estimated thickness <200 meters. True age unknown, possibly early Paleozoic or Proterozoic

<u>rss--red-weathering lithic sandstones</u>.

Fine-grained, probably volcaniclastic, red-weathering sandstones. Generally few sedimentary structures. Occasional thin (<5cm) interbeds of shale. Local volcanic (?) agglomerates and green silicified tuffs (?). Highly deformed and thickness indeterminate. Locally mapped as 'pCv' by Reiser and others (1980). True age unknown, possibly early Paleozoic or Proterozoic.

<u>OKPILAK BATHOLITH AREA</u>

The stratigraphy in the Okpilak batholith area is essentially the same as that in the Aichilik and Egaksrak rivers area (see figure A1.1). The Ellesmerian sequence is more highly deformed, with the most of the Triassic rocks missing due to recent erosion. The Mississippian Kayak Shale, a prominent shale and major detachment horizon at the base of the Ellesmerian sequence in the Aichilik and Egaksrak rivers area, is generally missing, or very thin, in the region around the Okpilak batholith. This study focused on the northern margin of the batholith. The entire Ellesmerian sequence is not exposed in this area, and those units that are exposed are highly deformed and commonly internally strained.

Consequently, it was not possible to determine true stratigraphic thicknesses of the Ellesmerian sequence units in this area. More detailed discussion of each unit, including stratigraphic thicknesses and descriptions of those units not encountered in the study area, are available in Sable (1977).

Quaternary alluvium

Alluvium and glacial deposits.

Mississippian through Triassic rocks

Triassic Ledge Sandstone Member of the Ivishak Formation.

Dark-weathering sandstones that form resistant rubble-covered ridges. Poorly exposed, stratigraphic thickness unknown. Detachment folded above Kavik Member of the Ivishak Formation.

Triassic Kavik Member of the Ivishak Formation.

Dark brown to black phyllitic slates and siltstones. Highly deformed, stratigraphic thickness unknown, possible detachment horizon.

Permian Echooka Formation.

Rusty-weathering fossiliferous calcareous fine-grained sandstones and siltstones. Possibly highly strained and internally deformed, true stratigraphic thickness unknown.

Pennsylvanian to Mississippian Lisburne Group (undifferentiated).

Grey-weathering mudstones and crinoidal grainstones. Highly deformed and internally strained, true stratigraphic thickness unknown.

Mississippian Kekiktuk Conglomerate.

Dark-grey quartzose medium- to coarse-grained sandstone to pebble conglomerate. Local dark chert and/or argillite clasts, local fossil hash. Thickness varies from less than 10 meters to greater than 25 meters.

Mississippian Kekiktuk Conglomerate (?), schistose.

Yellow-weathering quartz schist with micaceous and/or clay matrix. Highly variable in thickness from <5 meters to 10 meters.

Pre-Mississippian rocks

The only pre-Mississippian rocks encountered in the Okpilak batholith area during field work for this dissertation was the Devonian Okpilak batholith. Complete descriptions of the various primary igneous phases of the batholith, not recognized or used as map units during this study, are summarized in Sable (1977).

Devonian granite of the Okpilak batholith

Medium-grey weathering granite to quartz monzonite. Local coarse-grained feldspar porphyry with K-spar phenocrysts up to 4 cm in length. Micaceous foliation and S/C mylonites locally well-developed.

APPENDIX B:
AREA BALANCED MODELS OF
THE OKPILAK BATHOLITH AND AICHILIK RIVER TRANSECTS

Area-balanced models of both transects were used to evaluate the effects of the geometry of the orogenic sole fault on the possible range in tectonic shortening. In the process, it was possible to see the effect of a variety of regional factors on the geometry of the entire orogenic wedge.

The area-balanced models for each transect were initially constructed by modifying a detailed balanced section that was in progress. The detailed balanced section of the transect (at 1:125,000 scale) was reduced to 8 1/2" x 11". The form of the sub-Mississippian unconformity surface was then used as a template for the various area-balanced models. (The form of the sub-Mississippian unconformity surface could also be attained by drawing a profile (of the appropriate orientation) across the structure contour map of the northeastern ANWR (Figure 3.10 A) and adding the subsurface geometry of the surface from seismic data.) This template, and the depth to the basal detachment horizon (on the right, or northern, end of the section) remained constant for all models.

At this point, two slightly different methods were used to construct the actual models. For the Aichilik River transect, a copy of the template was scanned into a Macintosh IIci computer. Using the graphics program Canvas 2.1, this template was used as a base to construct a variety of different wedge models with different basal detachment geometries. The area of each of the resulting deformed wedges was calculated via Canvas and a corresponding restored section with the same area constructed. The program was used to calculate the length of both the deformed and restored sections. From this information, the amount of shortening could be determined for each model.

For the Okpilak batholith transect, a new copy of the template was used for each model, with the geometry of the basal detachment surface drawn in by hand. The deformed wedge was then input into a Mac SE computer using a Kurta IS/One digitizing tablet. A graphics program, MacDraft, was then used to calculate the area of the wedge, and construct a restored version of the wedge of the same area. The program was used to calculate the length of both the deformed and restored sections. From this information, the amount of shortening could be determined for each model.

The results using the two procedures are comparable. The method using a scanned image and Canvas is definitely preferable, and probably yields more reliable and accurate results. However, a MacII is not always available and the other method, albeit more tedious, also works quite well.

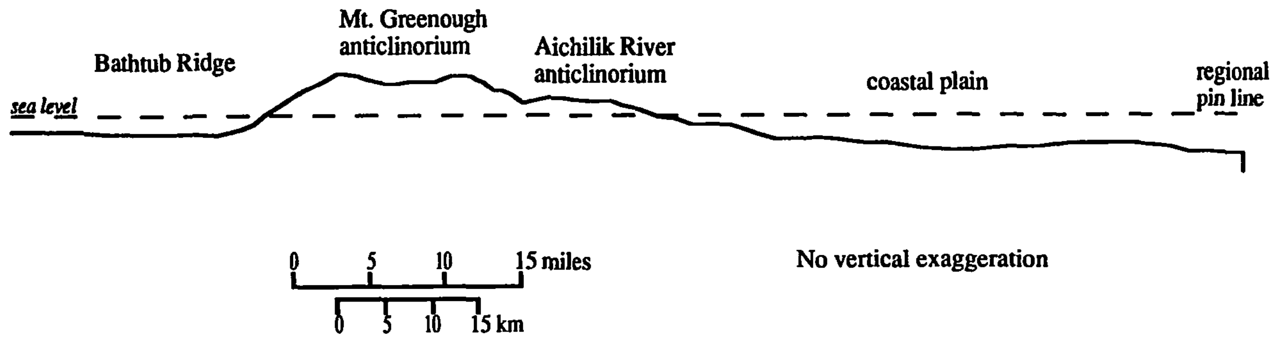
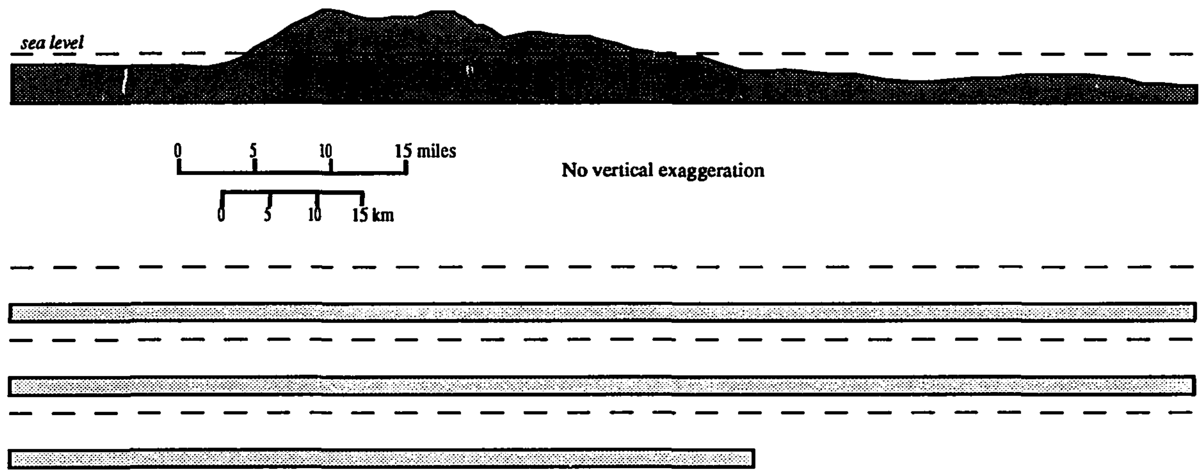


Figure B.1. Area-balanced models of the Aichilik River transect. This figure represents the form of the deformed sub-Mississippian unconformity surface used in all the following area-balanced models.

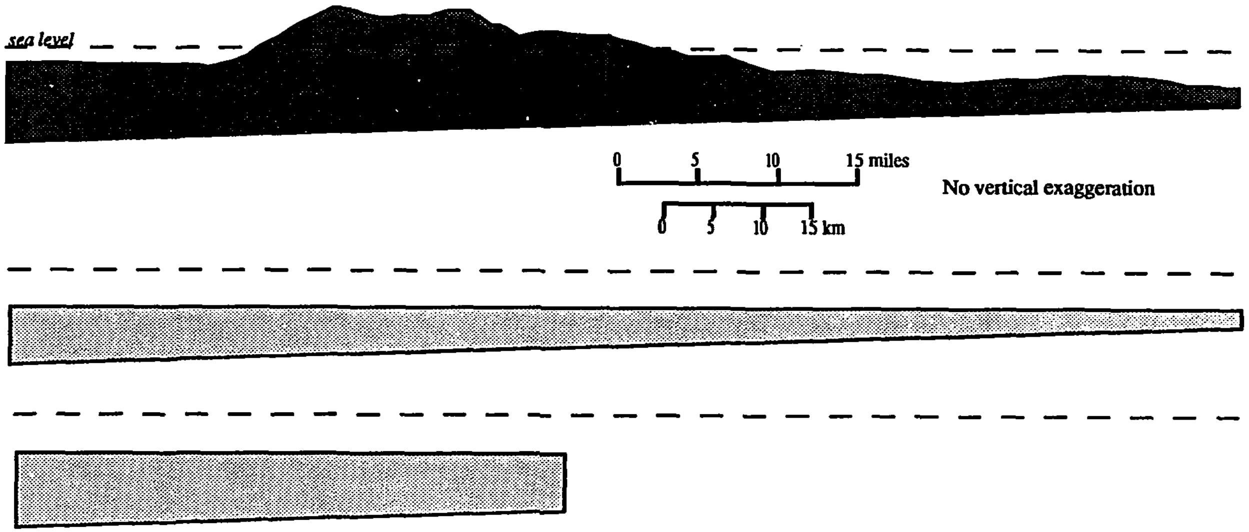


Deformed section--Length = 81 miles (131 km)
Depth to basal detachment at pin line = 20,000 ft (6.1 km)

Restored section--Length = 205 miles (329 km)
Depth to basal detachment at south end of section = 20,000 ft (6.1 km)
Dip of basal detachment held at 0°

Shortening = 124 miles (198 km) or 61.0 %

Figure B.1 (cont.). Model A: Flat basal detachment

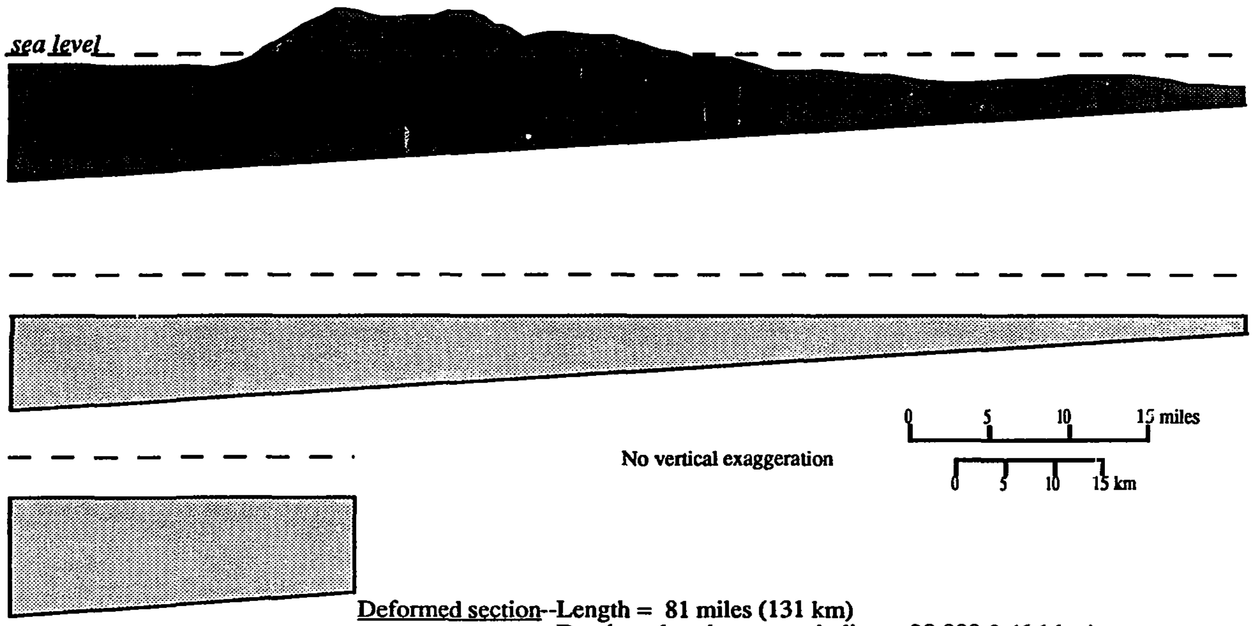


Deformed section--Length = 81 miles (131 km)
Depth to detachment at pin line = 20,000 ft (6.1 km)

Restored section --Length =117 miles (189 km)
Depth to basal detachment at south end of section = 41,000 ft (12.5 km)
Dip of basal detachment = 2°

Shortening = 36 miles (58 km) or 31%.

Figure B.1 (cont.). Model B: 2° dip on basal detachment surface

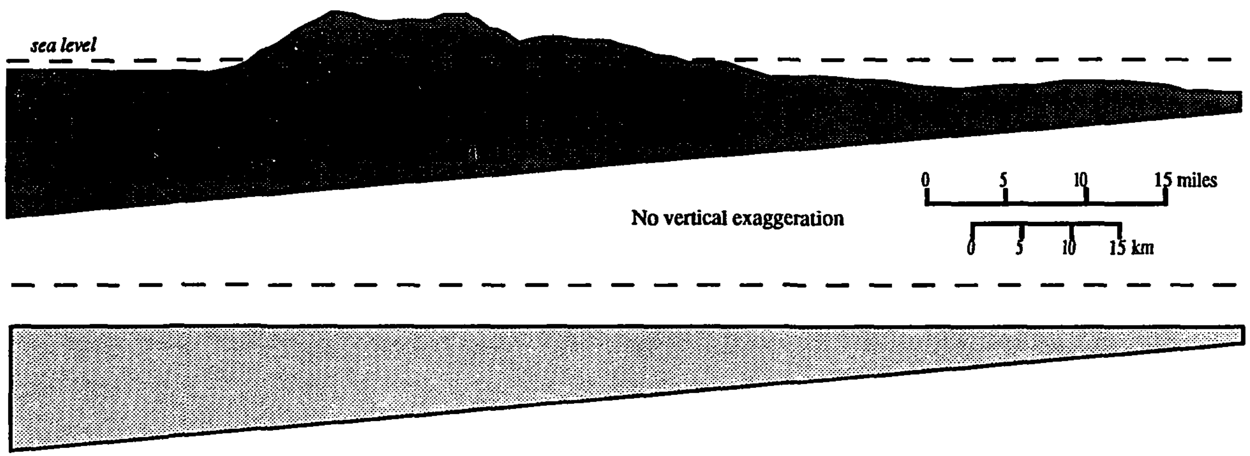


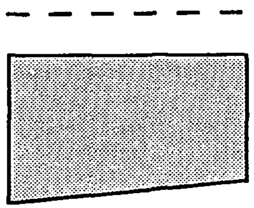
<u>Deformed section</u>--Length = 81 miles (131 km)
Depth to detachment at pin line = 20,000 ft (6.1 km)

<u>Restored section</u> --Length =104 miles (167 km)
Depth to basal detachment at south end of section = 56,800 ft (17.3 km)
Dip of basal detachment = 4°

<u>Shortening</u> = 23 miles (36 km) or 22%.

Figure B.1 (cont.). Model C: 4° slope on basal detachment



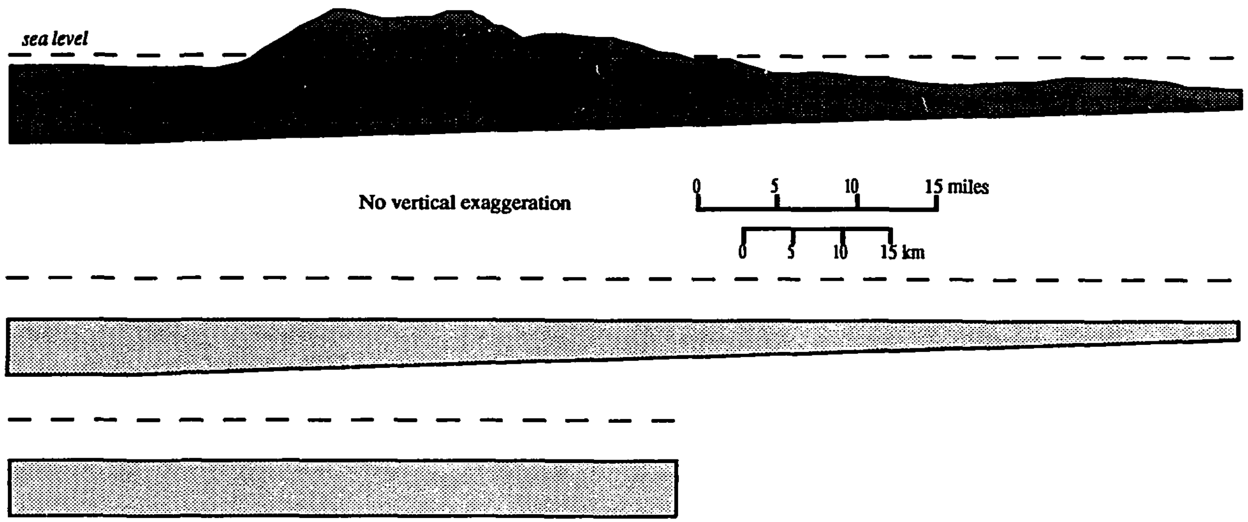


<u>Deformed section</u>--Length = 81 miles (131 km)
Depth to detachment at pin line = 20,000 ft (6.1 km)

<u>Restored section</u> --Length = 97 miles (156 km)
Depth to basal detachment at south end of section = 60,800 ft (18.5 km)
Dip of basal detachment = 6°

<u>Shortening</u> = 16 miles (25 km) or 16%.

Figure B.1 (cont.). Model D: 6° dip on basal detachment surface

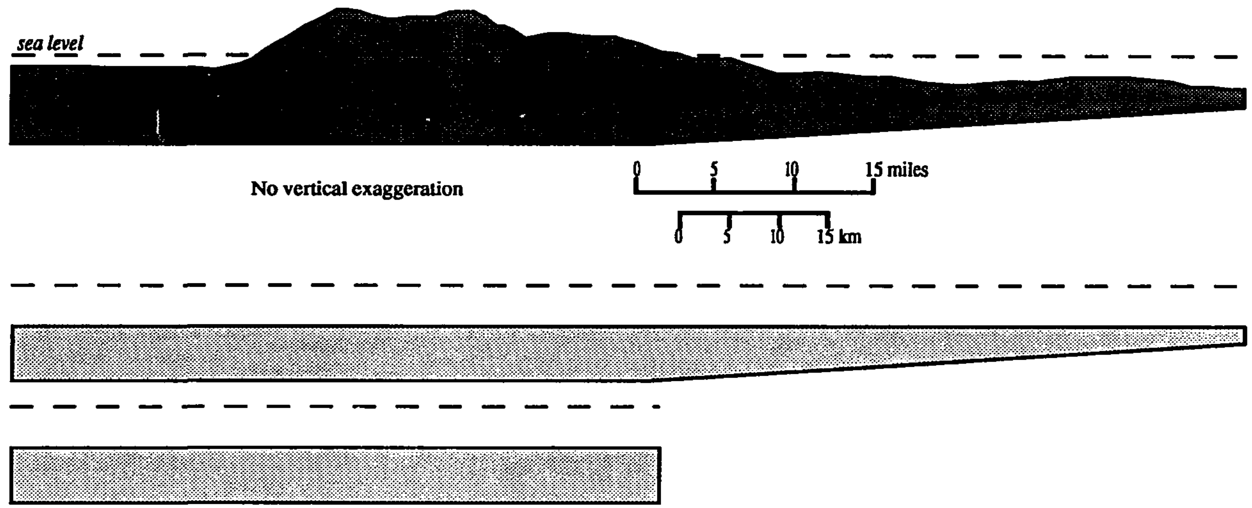


<u>Deformed section</u>--Length = 81 miles (131 km)
Depth to detachment at pin line = 20,000 ft (6.1 km)

<u>Restored section</u> --Length = 126 miles (203 km)
Depth to basal detachment at south end of section = 32,200 ft (10 km)
Dip of basal detachment = 2°

<u>Shortening</u> = 45 miles (72 km) or 35%.

Figure B.1 (cont.). Model E: 2° dip on basal detachment surface, brittle/ductile transition at 10 km

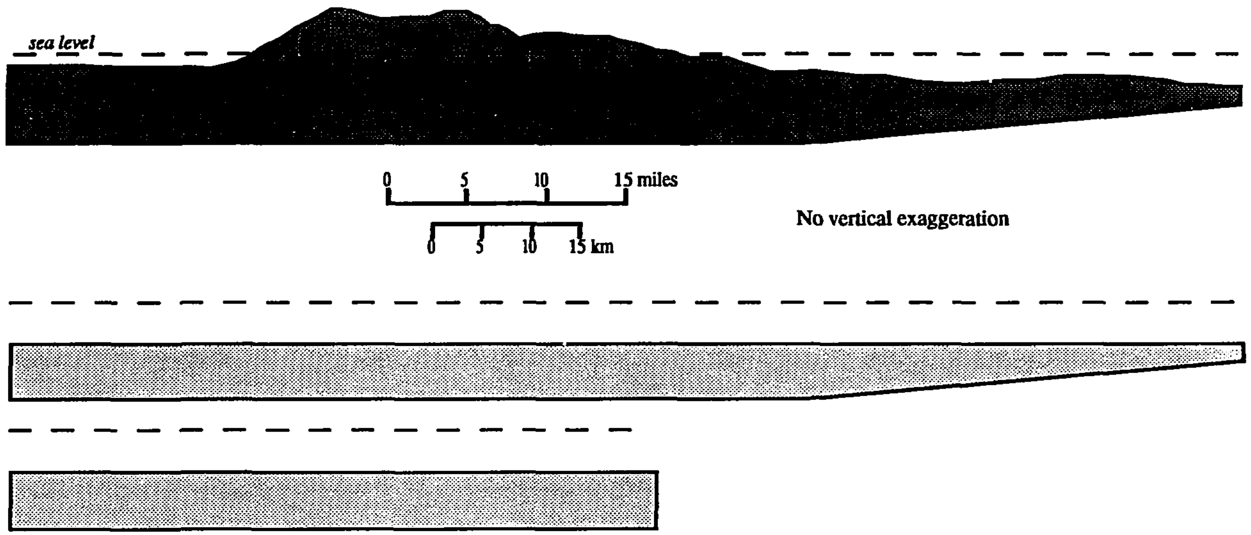


<u>Deformed section</u>--Length = 81 miles (131 km)
Depth to detachment at pin line = 20,000 ft (6.1 km)

<u>Restored section</u> --Length = 124 miles (199 km)
Depth to basal detachment at south end of section = 32,200 ft (10 km)
Dip of basal detachment = 4°

<u>Shortening</u> = 43 bmiles (68 km) or 34%.

Figure B.1 (cont.). Model F: 4° dip on basal detachment surface, brittle/ductile transition at 10 km

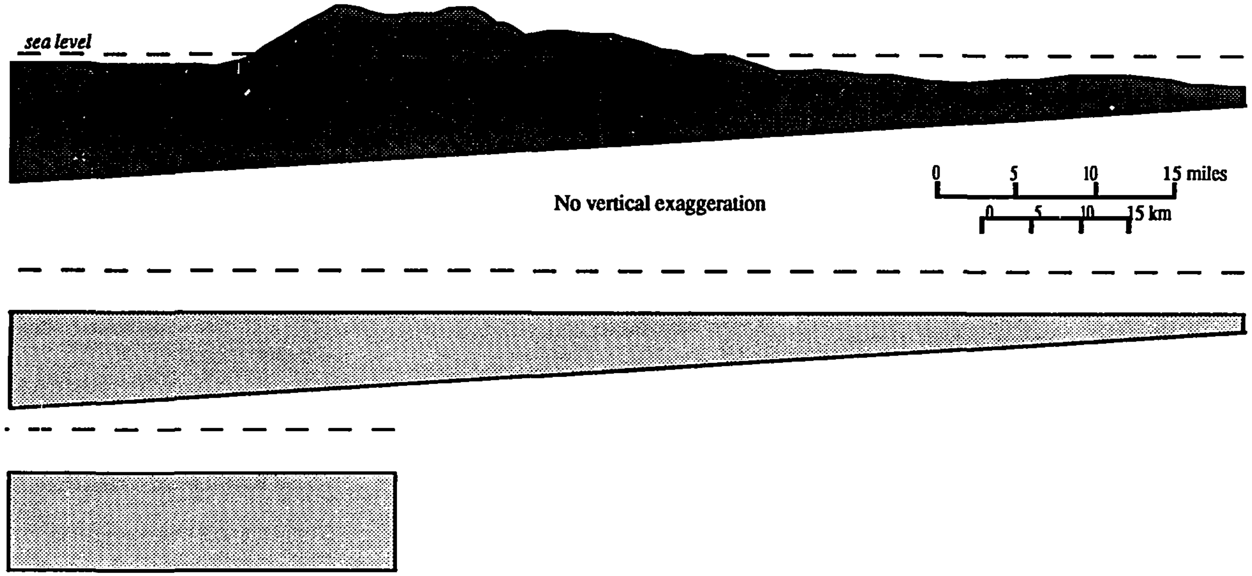


Deformed section--Length = 81 miles (131 km)
Depth to detachment at pin line = 20,000 ft (6.1 km)

Restored section --Length = 124 miles (199 km)
Depth to basal detachment at south end of section = 32,200 ft (10 km)
Dip of basal detachment = 6°

Shortening = 43 miles (68 km) or 34 %.

Figure B.1 (cont.). Model G: 6° dip on basal detachment surface, brittle/ductile transition at 10 km

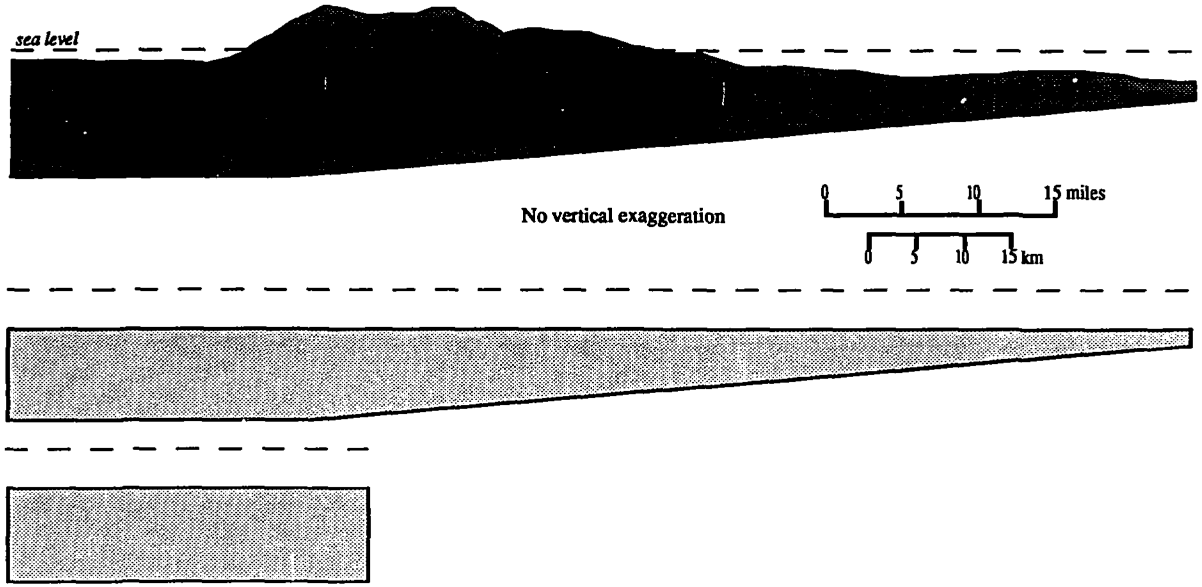


<u>Deformed section</u>--Length = 81 miles (131 km)
Depth to detachment at pin line = 20,000 ft (6.1 km)

<u>Restored section</u> --Length = 107 miles (172 km)
Depth to basal detachment at south end of section = 48,300 ft (15 km)
Dip of basal detachment = 4°

<u>Shortening</u> = 26 miles (41 km) or 24%.

Figure B.1 (cont.). Model H: 4° dip on basal detachment, brittle/ductile transition at 15 km

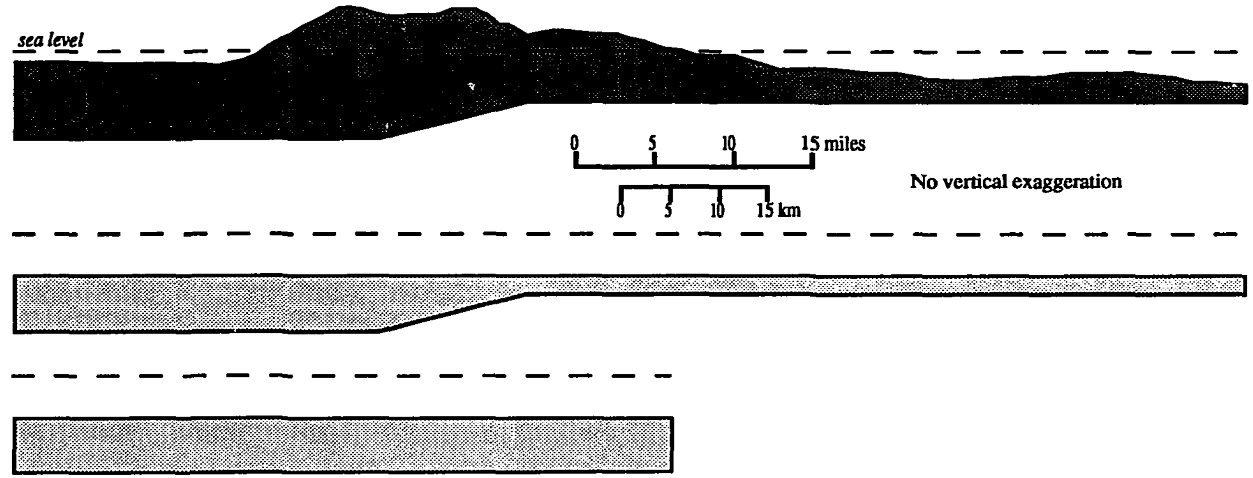


Deformed section--Length = 81 miles (131 km)
Depth to detachment at pin line = 20,000 ft (6.1 km)

Restored section --Length = 105 miles (169 km)
Depth to basal detachment at south end of section = 48,300 ft (15 km)
Dip of basal detachment = 6°

Shortening = 24 miles (38 km) or 23%.

Figure B.1 (cont.). Model I: 6° dip on basal detachment, brittle/ductile transition at 15 km

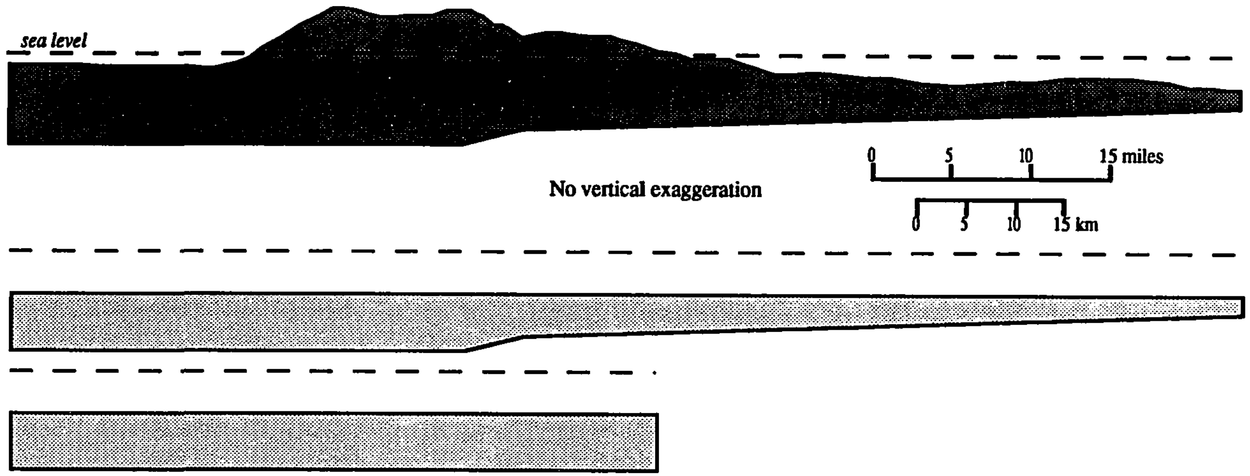


Deformed section--Length = 81 miles (131 km)
Depth to detachment at pin line = 20,000 ft (6.1 km)

Restored section --Length = 124 miles (200 km)
Depth to basal detachment at south end of section = 32,200 ft (10 km)
Dip of basal detachment = 0°

Shortening = 43 miles (69 km) or 35%.

Figure B.1 (cont.). Model J: 0° dip on basal detachment surface, brittle/ductile transition at 10 km, 15° ramp

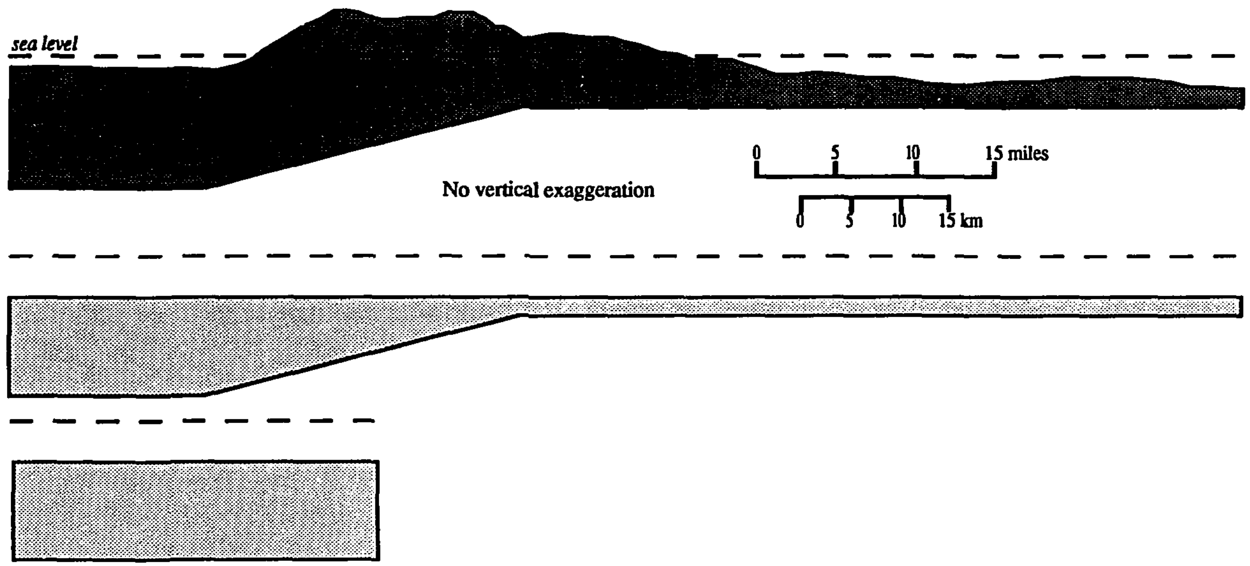


Deformed section--Length = 81 miles (131 km)
Depth to detachment at pin line = 20,000 ft (6.1 km)

Restored section --Length = 124 miles (199 km)
Depth to basal detachment at south end of section = 32,200 ft (10 km)
Dip of basal detachment = 2°

Shortening = 43 miles (68 km) or 34%.

Figure B.1 (cont.). Model K: 2° dip on basal detachment surface, brittle/ductile transition at 10 km, 15° ramp

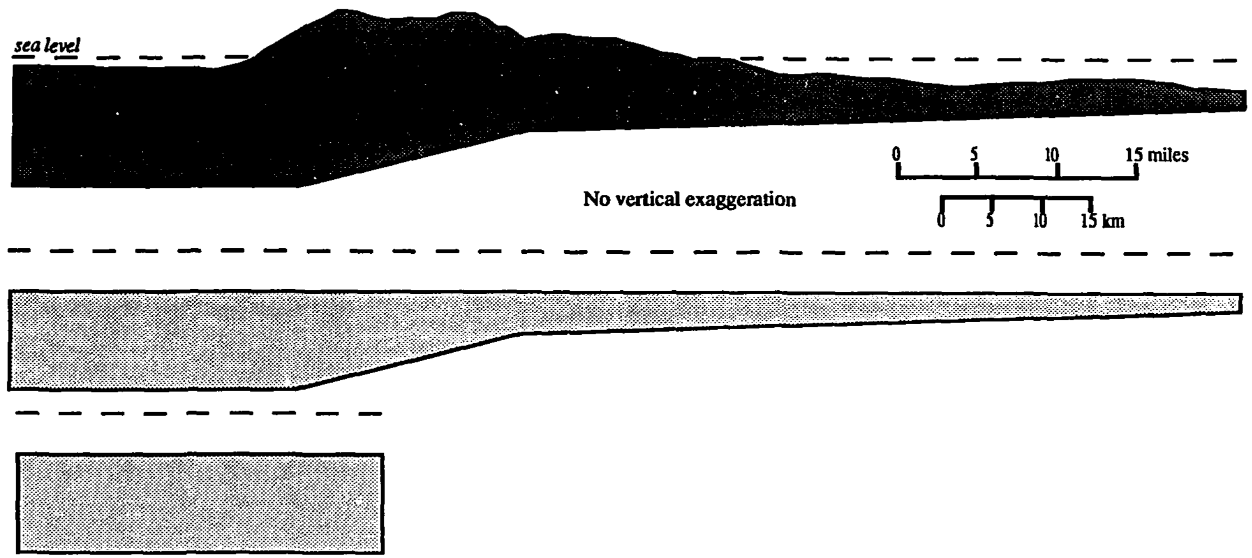


<u>Deformed section</u>--Length = 81 miles (131 km)
Depth to detachment at pin line = 20,000 ft (6.1 km)

<u>Restored section</u> --Length = 105 miles (169 km)
Depth to basal detachment at south end of section = 48,300 ft (15 km)
Dip of basal detachment = 0°

<u>Shortening</u> = 24 miles (38 km) or 23%.

Figure B.1 (cont.). Model L: 0° dip on basal detachment surface, brittle/ductile transition at 15 km, 15° ramp



<u>Deformed section</u>--Length = 81 miles (131 km)
Depth to detachment at pin line = 20,000 ft (6.1 km)

<u>Restored section</u> --Length = 105 miles (169 km)
Depth to basal detachment at south end of section = 48,300 ft (15 km)
Dip of basal detachment = 2°

<u>Shortening</u> = 24 miles (38 km) or 23%.

Figure B.1 (cont.). Model M: 2° dip on basal detachment surface, brittle/ductile transition at 15 km, 15° ramp

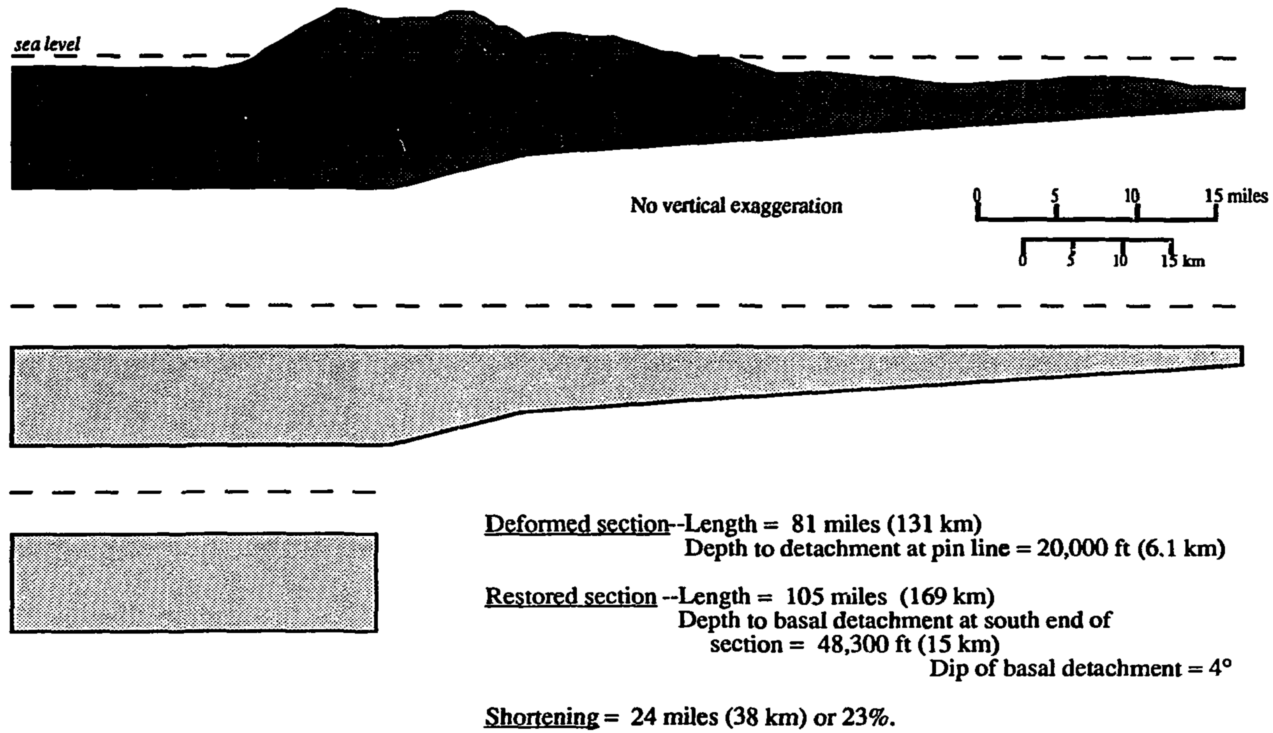


Figure B.1 (cont.). Model N: 4° dip on basal detachment surface, brittle/ductile transition at 15 km, 15° ramp

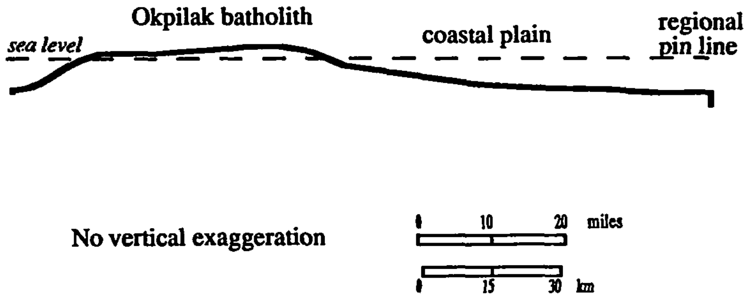
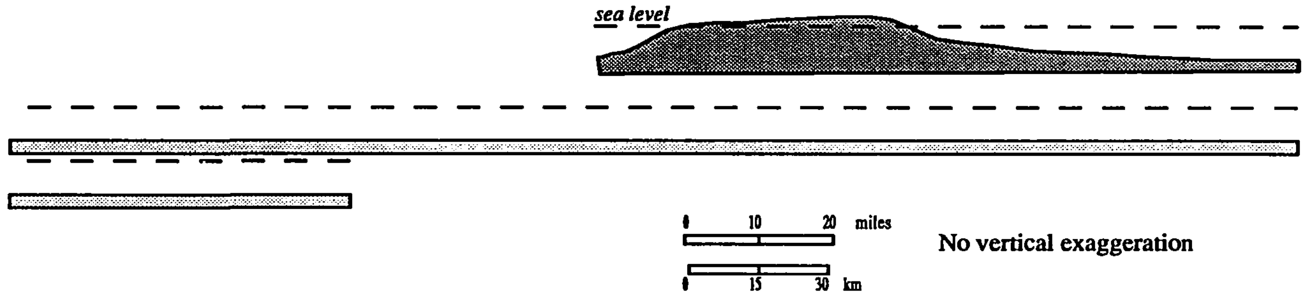


Figure B.2. Area-balanced models of the Okpilak batholith transect. This figure represents the form of the deformed sub-Mississippian unconformity surface used in all of the following area-balanced models.

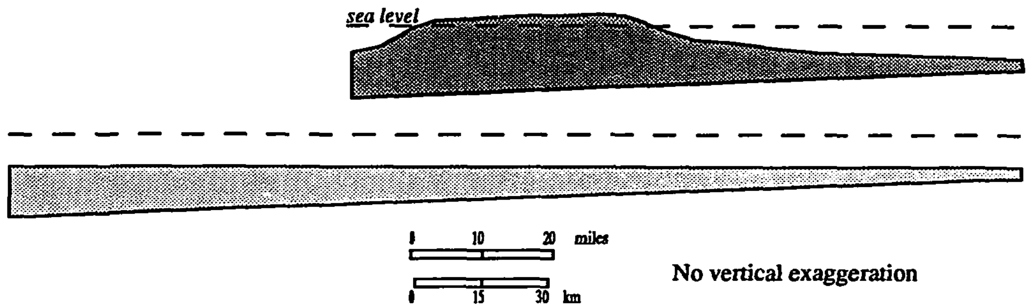


<u>Deformed section</u>--Length = 94.7 miles (151.7 km)
Depth to unconformity at pin line = 24,000 ft. (7.3 km)
Depth to detachment at pin line = 32,000 ft (9.8 km)
Slope on basal detachment = 0°

<u>Restored section</u>--Length = 221.1 miles (353.9 km)
Depth to unconformity held horizontal at 24,000 ft (7.3 km)
Depth to basal detachment at south end of section = 32,000 ft (9.8 km)
Slope on basal detachment held at 0°

<u>Shortening</u> = 126.3 miles (202.2 km) or 57%

Figure B.2 (cont.). Model A: Flat detachment at 32,000 ft. (9.8 km)

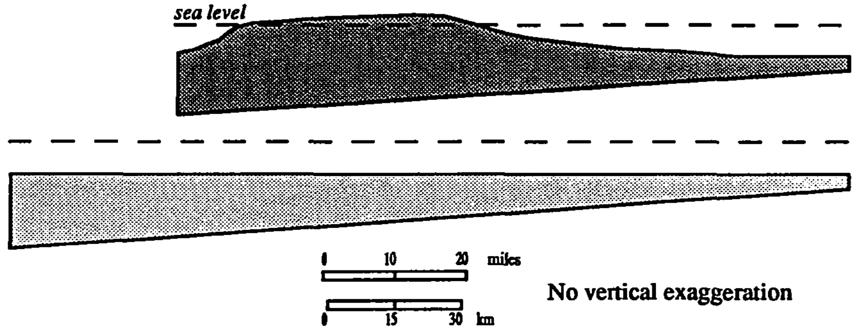


<u>Deformed section</u>--Length = 94.7 miles (151.7 km)
Depth to unconformity at pin line = 24,000 ft (7.3 km)
Depth to detachment at pin line = 32,000 ft (9.8 km)
Slope on basal detachment = 2°

<u>Restored section</u>--Restored length = 143.2 (229.2 km)
Depth to unconformity held horizontal at 24,000 ft (7.3 km)
Depth to basal detachment at south end of section = 64,000 ft (9.8 km)
Slope on basal detachment held at 2°

<u>Shortening</u> = 48.4 miles (77.5 km) or 33.8%

Figure B.2 (cont.). Model B: Constant 2° dip on basal detachment

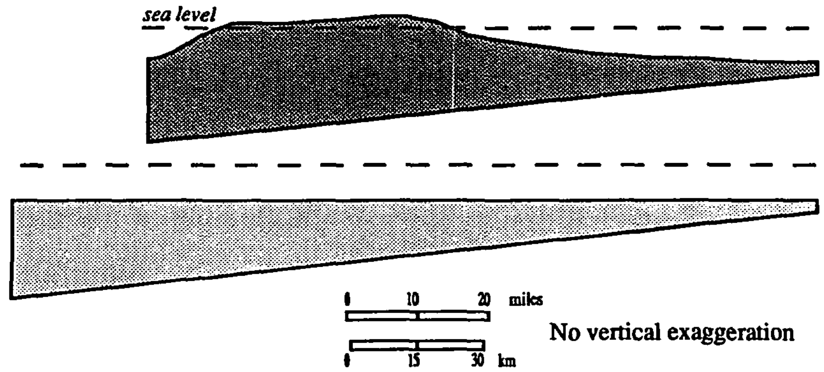


Deformed section--Length = 94.7 miles (151.7 km)
Depth to unconformity at pin line = 24,000 ft (7.3 km)
Depth to detachment at pin line = 32,000 ft (9.8 km)
Slope on basal detachment = 4°

Restored section--Length = 118.6 miles (189.9 km)
Depth to unconformity held horizontal at 24,000 ft (7.3 km)
Depth to basal detachment at south end of section = 82,700 ft (25.2 km)
Slope on basal detachment held at 4°

Shortening = 23.9 miles (38.2 km) or 20.1%

Figure B.2 (cont.). Model C: Constant dip of 4° on basal detachment

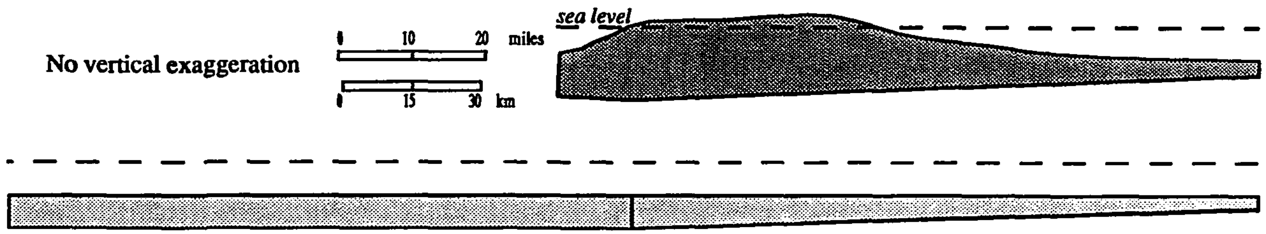


Deformed section–Length = 94.7 miles (151.7 km)
Depth to unconformity at pin line = 24,000 ft (7.3 km)
Depth to detachment at pin line = 32,000 ft (9.8 km)
Slope on basal detachment = 6°

Restored section--Length = 113.7 miles (182 km)
Depth to unconformity held horizontal at 24,000 ft (7.3 km)
Depth to basal detachment at south end of section = 100,500 ft (30.6 km)
Slope on basal detachment held at 6°

Shortening = 18.9 miles (30.3 km) or 16.7%

Figure B.2 (cont.). Model D: Constant 6° dip on basal detachment

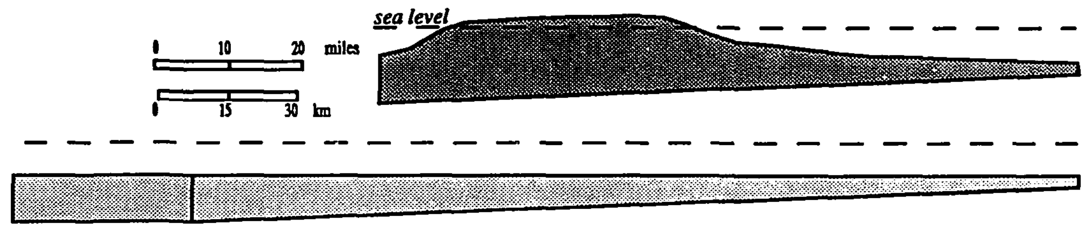


Deformed section--Length = 94.7 miles (151.7 km)
Depth to unconformity at pin line = 24,000 ft (7.3 km)
Depth to detachment = 32,000 ft (9.8 km) at pin line, flattening at 49,000 ft (15 km)
Slope on basal detachment = 2°

Restored section--Length = 169.8 miles (271.9 km)
Depth to unconformity held horizontal at 24,000 ft (7.3 km)
Depth to basal detachment at south end of section = 49,000 ft (15 km)
Slope on basal detachment held at 2°

Shortening = 75 miles (120.2 km) or 44.2%

Figure B.2 (cont.). Model E: Constant 2° dip on basal detachment, brittle/ductile transition of 49,000 ft (15 km)



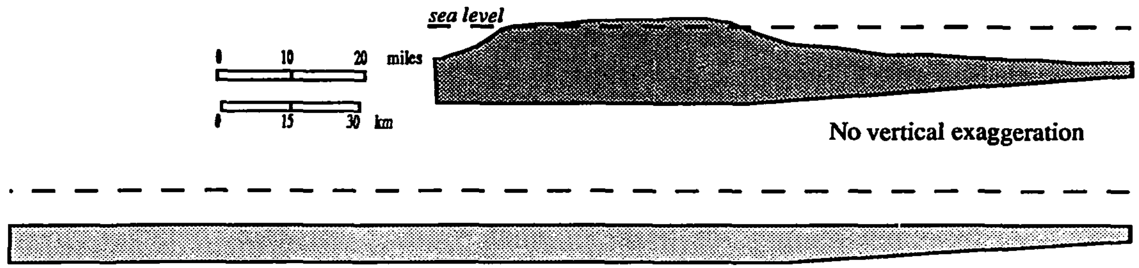
No vertical exaggeration

<u>Deformed section</u>--Length = 94.7 miles (151.7 km)
Depth to unconformity at pin line= 24,000 ft (7.3 km)
Depth to detachment = 32,000 ft (9.8 km) at pin line,
flattening at 59,000' ft (18 km)
Slope on basal detachment = 2°

<u>Restored section</u>--Length = 144.6 miles (231.5 km)
Depth to unconformity held horizontal at 24,000 ft (7.3 km)
Depth to basal detachment at south end of section = 59,000 ft (18 km)
Slope on basal detachment held at 2°

<u>Shortening</u> = 49.8 miles (79.8 km) or 34.5%

Figure B.2 (cont.). Model F: Constant 2° dip on basal detachment, brittle/ductile transition at 59,000 ft (18 km)

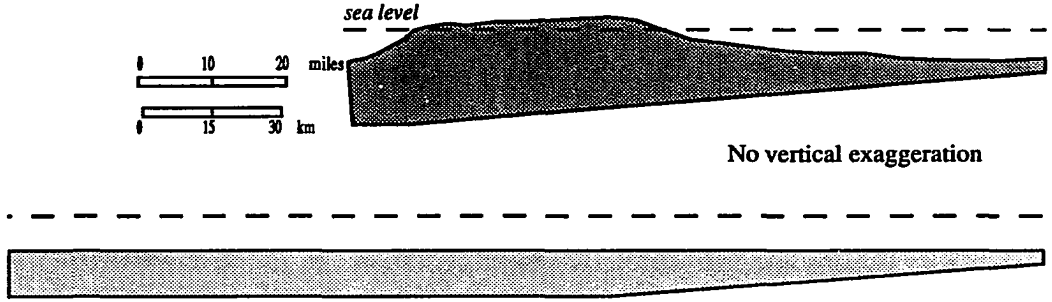


Deformed section--Length = 94.7 miles (151.7 km)
Depth to unconformity at pin line = 24,000 ft (7.3 km)
Depth to detachment = 32,000 ft (9.8 km) at pin line, flattening at 49,000 ft (15 km)
Slope on basal detachment = 4°

Restored section--Length = 152.3 miles (243.8 km)
Depth to unconformity held horizontal at 24,000 ft (7.3 km)
Depth to basal detachment at south end of section = 49,000 ft (15 km)
Slope on basal detachment held at 4°

Shortening = 58.2 miles (93.3 km) or 38.3%

Figure B.2 (cont.). Model G: Constant 4° dip on basal detachment, brittle/ductile transition of 49,000 ft (15 km)

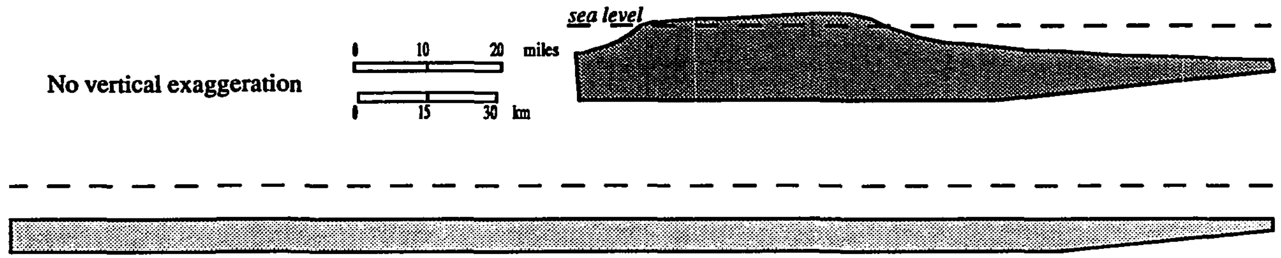


<u>Deformed section</u>--Length = 94.7 miles (151.7 km)
Depth to unconformity at pin line = 24,000 ft (7.3 km)
Depth to detachment = 32,000 ft (9.8 km) at pin line, flattening at 59,000 ft (18 km)
Slope on basal detachment = 4°

<u>Restored section</u>--Length = 140.4 miles (224.7 km)
Depth to unconformity held horizontal at 24,000 ft (7.3 km)
Depth to basal detachment at south end of section = 59,000 ft (18 km)
Slope on basal detachment held at 4°

<u>Shortening</u> = 45.6 miles (73 km) or 32.5%

Figure B.2 (cont.). Model H: Constant 4° dip on basal detachment, brittle/ductile transition at 59,000 ft (18 km)

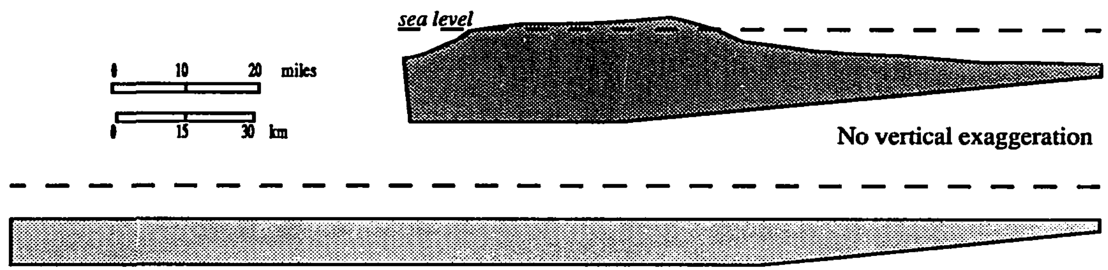


<u>Deformed section</u>--Length = 94.7 miles (151.7 km)
Depth to unconformity at pin line = 24,000 ft (7.3 km)
Depth to detachment = 32,000 ft (9.8 km) at pin line, flattening at 49,000 ft (15 km)
Slope on basal detachment = 6°

<u>Restored section</u>--Length = 171.2 miles (274.2 km)
Depth to unconformity held horizontal at 24,000 ft (7.3 km)
Depth to basal detachment at south end of section = 49,000 ft (15 km)
Slope on basal detachment held at 6°

<u>Shortening</u> = 76.5 miles (122.5 km) or 44.7%

Figure B.2 (cont.). Model I: Constant 5° dip on basal detachment, brittle/ductile transition at 49,000 ft (15 km)

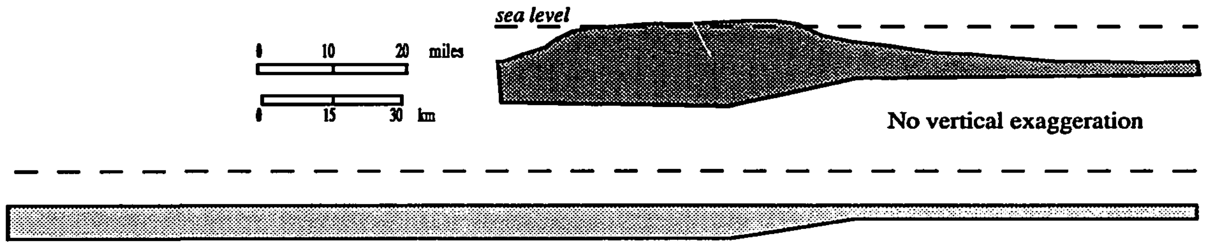


<u>Deformed section</u>--Length = 94.7 miles (151.7 km)
Depth to unconformity = 24,000 ft (7.3 km)
Depth to detachment = 32,000 ft (9.8 km) at pin line, flattening at 59,000 ft (18 km)
Slope on basal detachment = 6°

<u>Restored section</u>--Length = 147.4 miles (236 km)
Depth to unconformity held horizontal at 24,000 ft (7.3 km)
Depth to basal detachment at south end of section = 59,000 ft (18 km)
Slope on basal detachment held at 6°

<u>Shortening</u> = 52.6 miles (84.3 km) or 35.7%

Figure B.2 (cont.). Model J: Constant 6° dip on basal detachment, brittle/ductile transition of 59,000 ft (18 km)

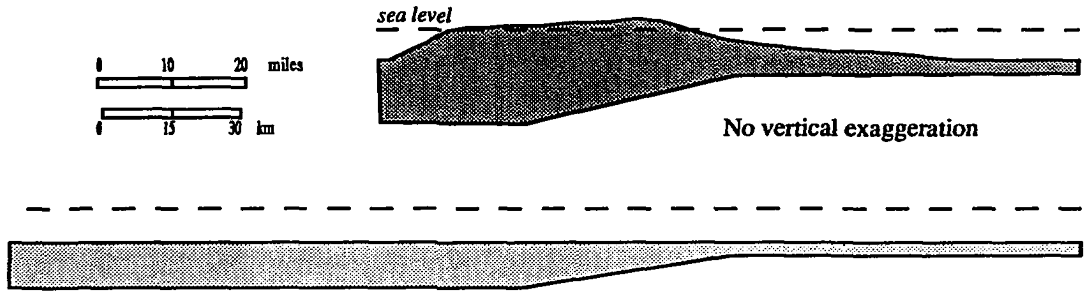


<u>Deformed section</u>--Length = 94.7 miles (151.7 km)
Depth to unconformity at pin line = 24,000 ft (7.3 km)
Depth to detachment = 32,000 ft (9.8 km) at pin line,
flat south to 15° ramp,
flattening at 49,000 ft (15 km)
Slope on basal detachment = 0°

<u>Restored section</u>--Length = 160.8 miles (258.4 km)
Depth to unconformity held horizontal at 24,000 ft (7.3 km)
Depth to basal detachment at south end of section = 49,000 ft (15 km)
Slope on basal detachment held at 0°

<u>Shortening</u> = 66.7 miles (106.7 km) or 41.3%

Figure B.2 (cont.). Model K: 0° dip on basal detachment surface, brittle/ductile transition at 49,000 ft (15 km), 15 ramp

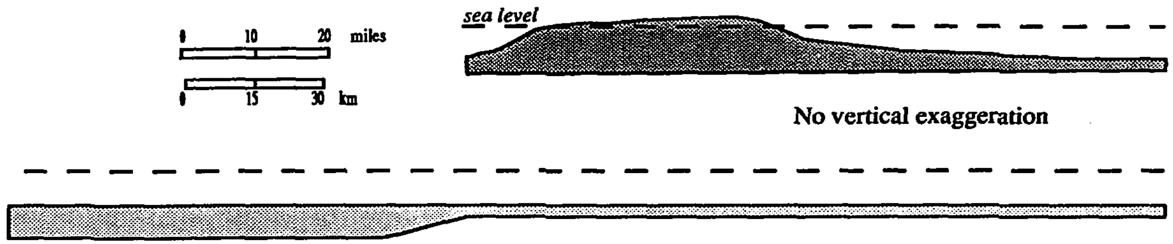


Deformed section--Length = 94.7 miles (151.7 km)
Depth to unconformity at pin line = 24,000 ft (7.3 km)
Depth to detachment = 32,000 ft (9.8 km) at pin line,
flat south to 15° ramp,
flattening at 59,000 ft (18 km)
Slope on basal detachment = 0°

Restored section--Length = 145.3 miles (232.6 km)
Depth to unconformity held horizontal at 24,000 ft (7.3 km)
Depth to basal detachment at south end of section = 59,000 ft (18 km)
Slope on basal detachment held at 0°

Shortening = 50.5 miles (80.9 km) or 34.8%

Figure B.2 (cont.). Model L: Flat basal detachment, brittle/ductile transition at 59,000 ft (18 km), 15° ramp



<u>Deformed section</u>--Length = 94.7 miles (151.7 km)
Depth to unconformity at pin line = 24,000 ft (7.3 km)
Depth to detachment = 32,000 ft (9.8 km) at pine line,
flat south to 15° ramp,
flattening at 49,000 ft (15 km)
Slope on basal detachment = 0°

<u>Restored section</u>--Length = 156.5 miles (259.6 km)
Depth to unconformity held horizontal at 24,000 ft (7.3 km)
Depth to basal detachment at south end of section = 49,000 ft (15 km)
Slope on basal detachment held at 0°

<u>Shortening</u> = 61.8 miles (98.9 km) or 39.5%

Figure B.2 (cont.). Model M: Flat basal detachment, brittle/ductile transition of 49,000 ft (15 km), 15° ramp

APPENDIX C:
AGE AND TEMPERATURE INFORMATION

Palynology Samples

Six shale samples were analyzed by J. Utting of the Canadian Geological Survey to determine if palynomorphs and/or other organic matter were present, what age determinations were possible, what were possible environments of deposition and in order to the thermal history of the area using the Thermal Alteration Index (TAI). The locations of these samples are shown on Figure C.1. The results of this analysis are summarized in Table C.1. Figure C.2 summarizes the relationship of TAI values to absolute temperature (North, 1985).

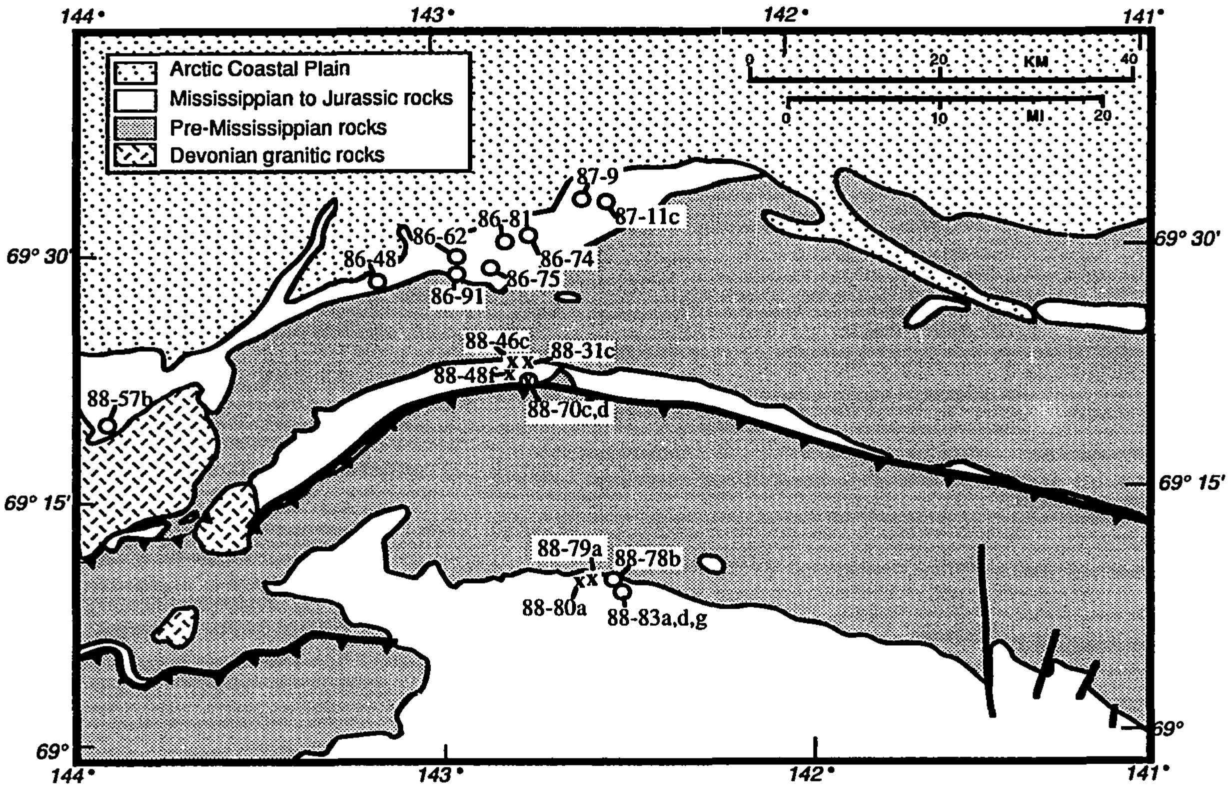


Figure C.1. Tectonic map of the northeastern Brooks Range showing the location of palynology and conodont samples collected during this study.

Table C.1. Table summarizing palynology results.

Sample	Unit	Location	Field description	Palynological results
88LR-31C	Mississippian Kayak Shale	Lat. 69°, 24.12' N, Long. 142°, 49.10' W; Demarcation Point B4, T1N, R38E, NW S 33, elevation ~ 3200'	Sample taken ~80 meters above Mississippian Kekiktuk Conglomerate float	Black amorphous, exinous, woody and coaly fragments. Some black trilete sporses. TAI = 5
88LR-46 C	'pCqs' or Mississippian Kayak Shale?	Lat. 69°, 24.28' N, Long. 142°, 49.90' W; Demarcation Point B4, T1N, R38E, NE S 32, elevation ~ 2800'	Highly deformed shale in possible fault contact with surrounding pre-Miss. rocks.	Rare woody and coaly fragments?
88LR-48F	Mississippian Kayak Shale	Lat. 69°, 24.20' N, Long. 142°, 50.8' W; Demarcation Point B4, T1N, R38E, NC S 32, elevation ~ 3050'	Sample collected ~27 m. above last exposed pMiss; may be only 5 m. above unconformity surface	Black amorphous, exinous, woody and coaly fragments. Some black trilete spores TAI = 5
88LR-70D	Mississippian Kayak Shale (limestone portion)	Lat. 69°, 22.6' N, Long. 142°, 41.6' W; Demarcation Point B3, T1S, R39E, SL 9 & 4 elevation ~ 2100'	Small exposure of deformed shales and thin bioclastic limestones, possibly in footwall of Whale Mt. thrust, above roof thrust of Cenozoic duplex in pMiss rocks.	Black amorphous, exinous, woody and coaly fragments. Some black trilete spores TAI = 5

Sample	Unit
88LR-79A	Mississippian Kayak limestones
88LR-80A	Mississippian Kekiktuk Conglomerate

Table C.1. (cont).

Location	Field description	Palynological results
Lat. 69°, 10.72' N, Long. 142°, 34.38' W; Demarcation Point A3, T3S, R39E, SE S 13 elevation ~ 3300'	Deformed section of Kayak limestones ~50 m. below base of Lisburne Group.	Black amorphous, exinous, woody and coaly fragments. Some black trilete spores TAI = 5
Lat. 69°, 11.02' N, Long. 142°, 39.4' W; Demarcation Point A3, T3S, R39E, E S 15 elevation ~ 2700'	Shale underlying conglomerate. No more than 4 m. above the unconformity surface.	Black amorphous, exinous, woody and coaly fragments. Some black trilete spores TAI = 5

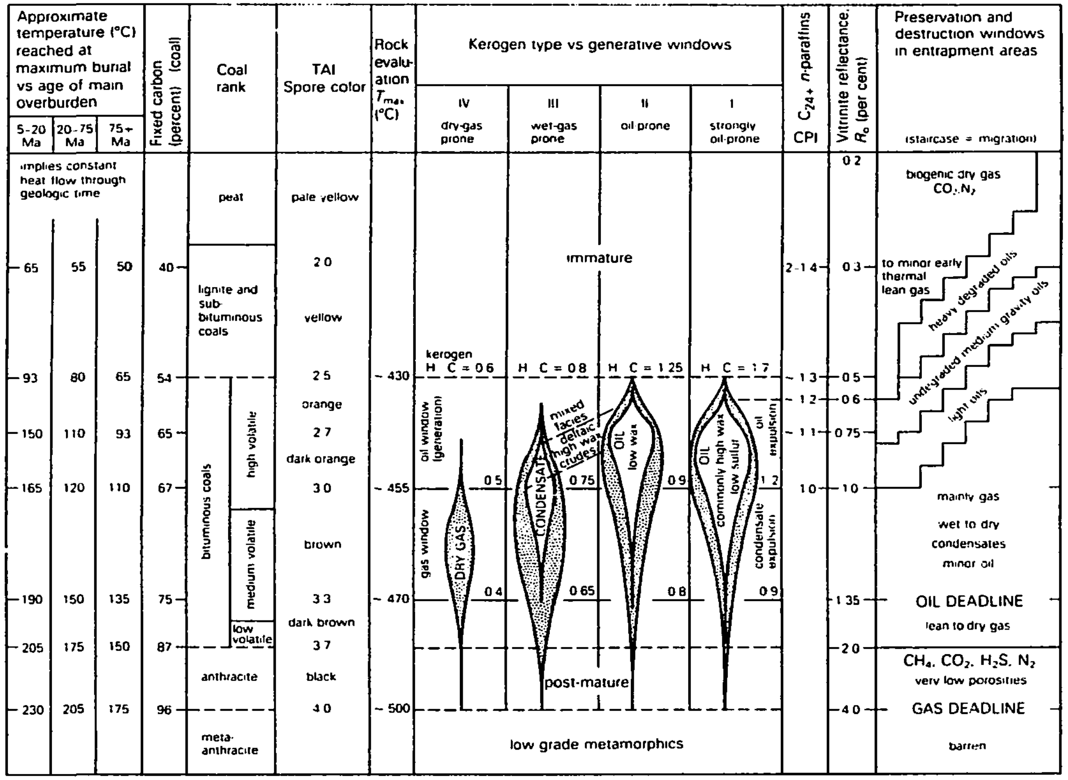


Figure C.2. Chart summarizing the relationship of TAIs to absolute temperature (from North, 1985)

Conodont Samples

Thirty-one carbonate samples from pre-Mississippian carbonate rocks and Mississippian and Pennsylvanian rocks of the Lisburne Group were analyzed by Dr. A. Harris of the United States Geological Survey in order to assess the age, depositional environment and thermal history of the samples. Not all of the samples submitted for analysis yielded conodonts--all seven samples from the pre-Mississippian carbonate succession were barren and ten samples from the Lisburne Group failed to yield conodonts. Fourteen samples of Lisburne Group and Kayak limestones from the area on the northern margin of the batholith, Leffingwell Ridge, south limb of the anticlinorium and immediately north of Bathtub Ridge yielded recognizable, datable conodonts from which a CAI value could be determined. The poor conodont recovery may have been due in part to incomplete processing of the samples, and some of the samples will probably be redone in the future. Table C.2 is a summary of the age, paleoenvironment and CAI results for each sample. The locations of the samples are shown on Figure C.1; Figure C.3 summarizes the relationship of CAI values to absolute temperature (Rejebian and others, 1987).

Table C.2. Table summarizing conodont results.

Sample	Unit	Location	Field description	Conodont results
86LR-75	Lisburne Limestone	Lat. 69°, 32' N, Long. 142°, 45' W; Demarcation Point C3, T2N, R38E, S 14 elevation ~ 1750'	Base of Lisburne Group in the hanging wall of small thrust south of Leffingwell Ridge between the Aichilik and Egaksrak rivers	Early Morrowan age (Early Pennsylvanian) Normal marine, shallow water CAI = 2 or 3 (70-130°C)
86LR-81	Alapah Limestone	Lat. 69°, 32' N, Long. 142°, 45' W; Demarcation Point C3, T2N, R38E, S 14 elevation ~ 1750'	Sample from lowest exposure of Alapah lst. in Egaksrak River klippe west of Egaksrak River.	Middle Meramecian to late Wolfcampian (Late Mississippian to middle Early Permian), probably Late Mississippian. Normal marine, shallow, high energy water CAI = 4 (200°C)
87LR-9	Alapah lst.	Lat. 69°, 32.6' N, Long. 142°, 39' W; Demarcation Point C3, T2N, R39E, S 7 elevation ~ 1400'	Sample from lowest exposure of Alapah lst. in Egaksrak River klippe east of Egaksrak River.	Middle Meramecian-late Wolfcampian (Late Mississippian to middle Early Permian), probably Late Mississippian. High energy marine water CAI = 3 (130°C)

Table C.2. (cont).

Sample	Unit	Location	Field description	Conodont results
87LR-11c	Wahoo (?) lst.	Lat. 69°, 34.05' N, Long. 142°, 33.5' W; Demarcation Point C3, T3N, R39E, S 33 elevation ~ 1300'	Wahoo limestone from the east end of the Egaksrak River Klippe	Morrowan-Atokan (Early-Middle Pennsylvanian) Normal marine, high energy, shallow water CAI = 4 (200°C)
86LR-48	Wahoo lst	Lat. 69°, 28.5' N, Long. 143°, 7.3' W; Demarcation Point B4, T1N, R37E, S 30 elevation ~ 2800'	Top of Wahoo limestone along Leffingwell Ridge.	late Morrowan-late Wolfcampian (late Early Pennsylvanian to middle Early Permian, but probably Atokan) Warm, normal marine, platform or shelf CAI = 4 (200°C)
86LR-62	Wahoo lst	Lat. 69°, 30' N, Long. 142°, 55.4' W; Demarcation Point B4, T2N, R37E, S 25 elevation ~ 2800'	Top of Wahoo lst along Leffingwell Ridge east of the Aichilik River.	Morrowan to early Atokan Normal marine, platform CAI = 4 (200°C)

Table C.2. (cont).

Sample	Unit	Location	Field description	Conodont results
88LR-83A	Alapah lst, poss. Kayak Shale?	Lat. 69°, 09.9' N, Long. 142°, 34.2' W; Demarcation Point A3, T3S, R39E, E 1/2, SE 1/4 of S 24 elevation ~ 3100'	From basal Alapah lst in imbricate zone underlying intact Lisburne sequence immediately north of Bathtub Ridge.	Kinderhookian to early Osagean (Early Mississippian) CAI = 5 to 5.5 (300 - 350°C)
88LR-83D	Alapah lst	Lat. 69°, 09.8' N, Long. 142°, 34.4' W; Demarcation Point A3, T3S, R39E, E 1/2, SE 1/4 of S 24 elevation ~ 3100'	From basal cherty Alapah lst of intact Lisburne sequence immediately north of Bathtub Ridge.	Devonian to Permian CAI = ~5
88LR-83G	upper (?) Alapah lst ?	Lat. 69°, 09.4' N, Long. 142°, 34.6' W; Demarcation Point A3, T3S, R39E, NE 1/2, S 25 elevation ~ 3500'	From base of upper massive grainstone/ packstone section of Lisburne immediately north of Bathtub Ridge.	middle Meramecian to late Wolfcampian (Late Mississippian to early Early Permian) CAI = 5 to 5.5 (300 to 350°C)
86LR-74	Alapah lst	Lat. 69°, 32' N, Long. 142°, 45.1' W; Demarcation Point C3, T2N, R38E, S 16 elevation ~ 1800'	From base of Alapah lst of the Egaksrak River klippe.	Middle Meramecian to Chesterian (probably early Chesterian) Shallow, high energy, normal marine water CAI = ~3

Table C.2. (cont).

Sample	Unit	Location	Field description	Conodont results
86LR-91	Alapah lst	Lat. 69°, 30' N, Long. 142°, 54.9' W; Demarcation Point B4, T2N, R37E, S 25	From base of Alapah lst in small hills south of Leffingwell Ridge, immediately east of the Aichilik River.	middle Meramecian to Chesterian (but not latest Chesterian). Shallow water platform CAI = 3.5 (160°C)
88LR-57B	Alapah lst (?) or Kayak limestones	Lat. 69°, 20.35' N, Long. 143°, 56.1' W; Demarcation Point B5, T1S, R34E, WC S 21 elevation ~ 6000'	From deformed lower Alapah lst (?) north of Okpilak batholith; granite is both structurally above and below this location.	Kinderhookian to early Osagean (Early Mississippian) CAI = 4.5 (250°C)
88LR-70C	Kayak Shale	Lat. 69°, 22.5' N, Long. 142°, 41.6' W; Demarcation Point B3, T1S, R39E, boundary of S 4 & 9 elevation ~ 2100'	From Kayak Shale structurally overlying Cenozoic imbricates within the pre-Miss rocks.	Probably Kinderhookian to Meramecian (Mississippian) CAI = 5 (300°C)
88LR-78B	Alapah lst	Lat. 69°, 10.5' N, Long. 142°, 34.4' W; Demarcation Point A3, T3S, R39E, E 1/2, N 1/4 of S 24 elevation ~ 3000'	From basal Alapah lst immediately north of Bathtub Ridge	No conodonts recovered but 8 ichthyoliths found. post-Devonian CAI = ~5 (300°C)

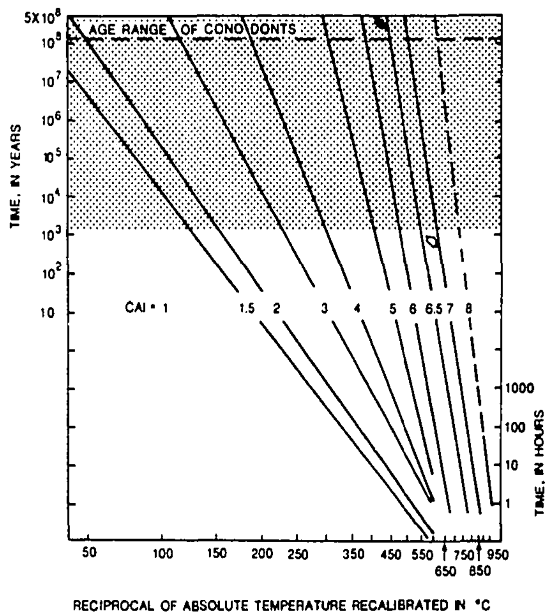


Figure C.3. Graph summarizing the relationship of CAIs to absolute temperature (from Rejebian and others, 1987).

PLEASE NOTE:

Oversize maps and charts are filmed in sections in the following manner:

LEFT TO RIGHT, TOP TO BOTTOM, WITH SMALL OVERLAPS

The following map or chart has been refilmed in its entirety at the end of this dissertation (not available on microfiche). A xerographic reproduction has been provided for paper copies and is inserted into the inside of the back cover.

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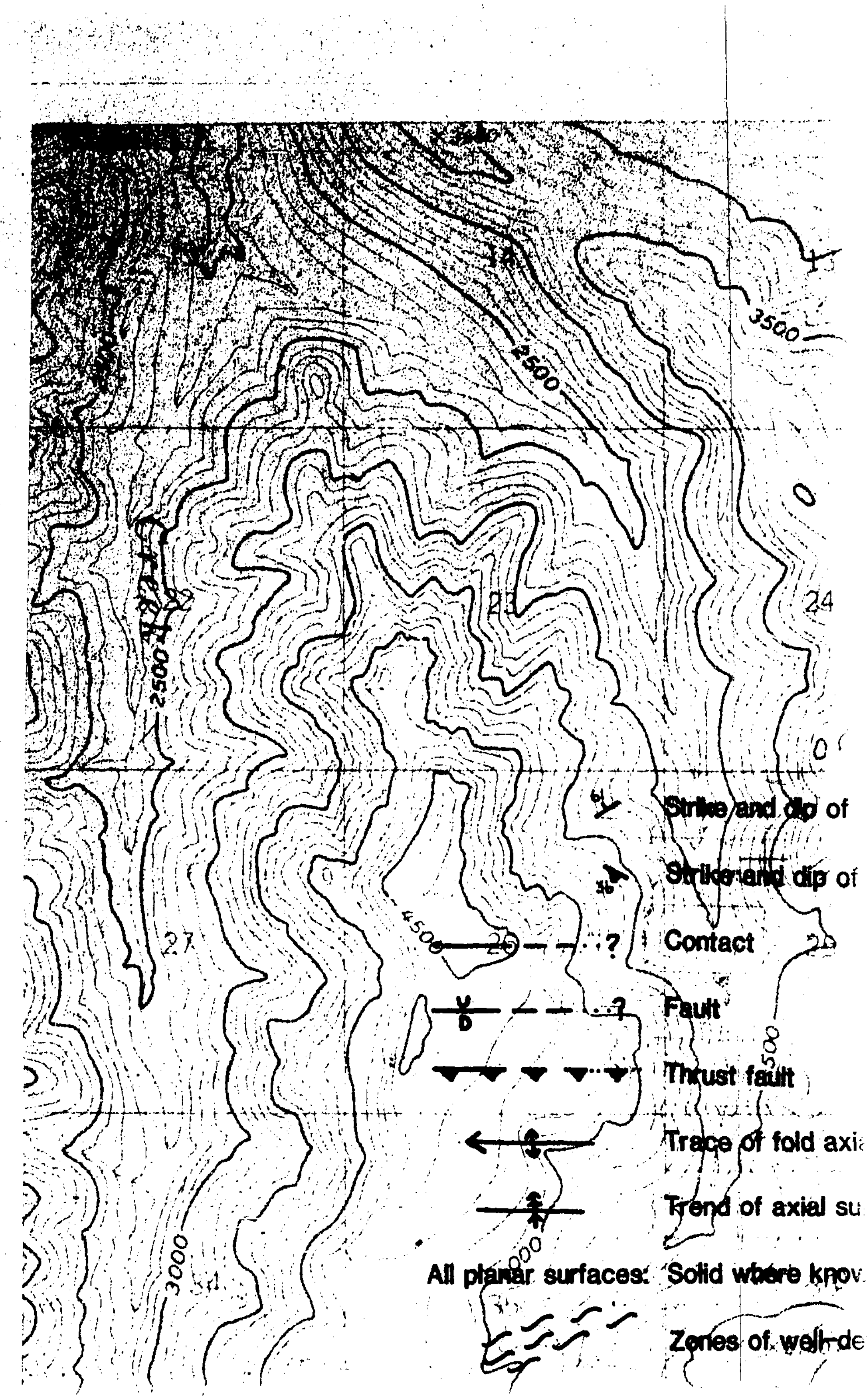
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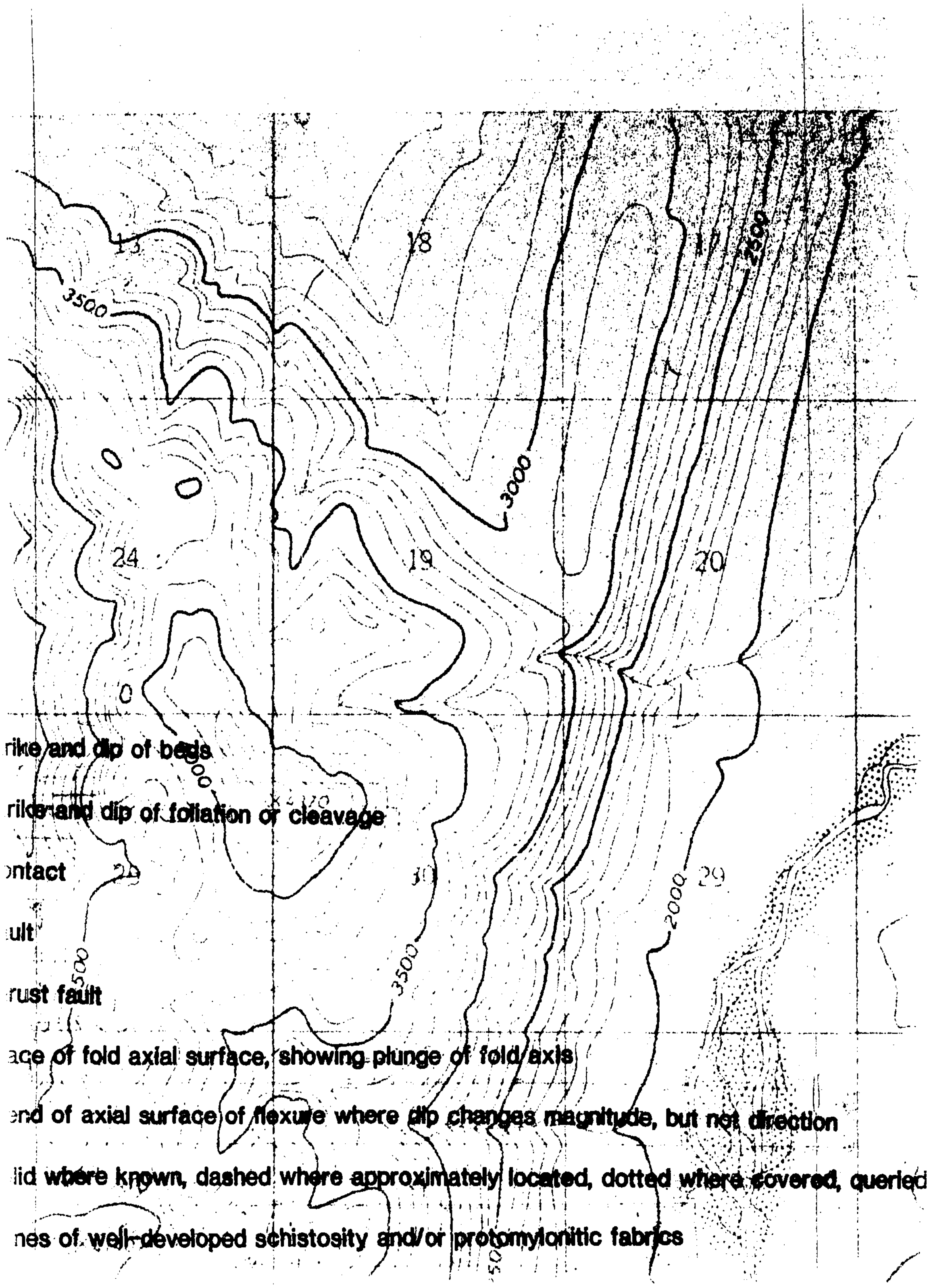
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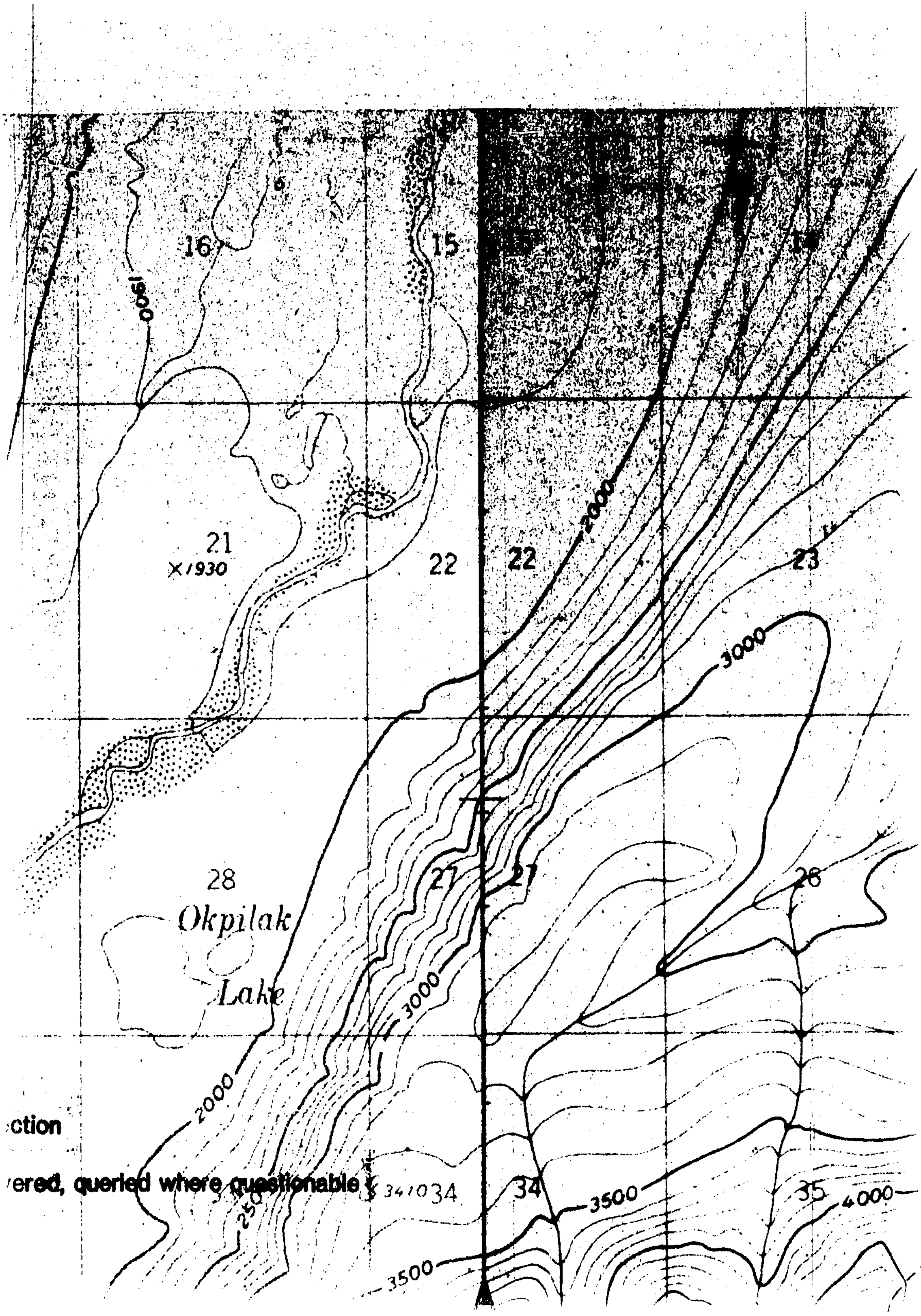
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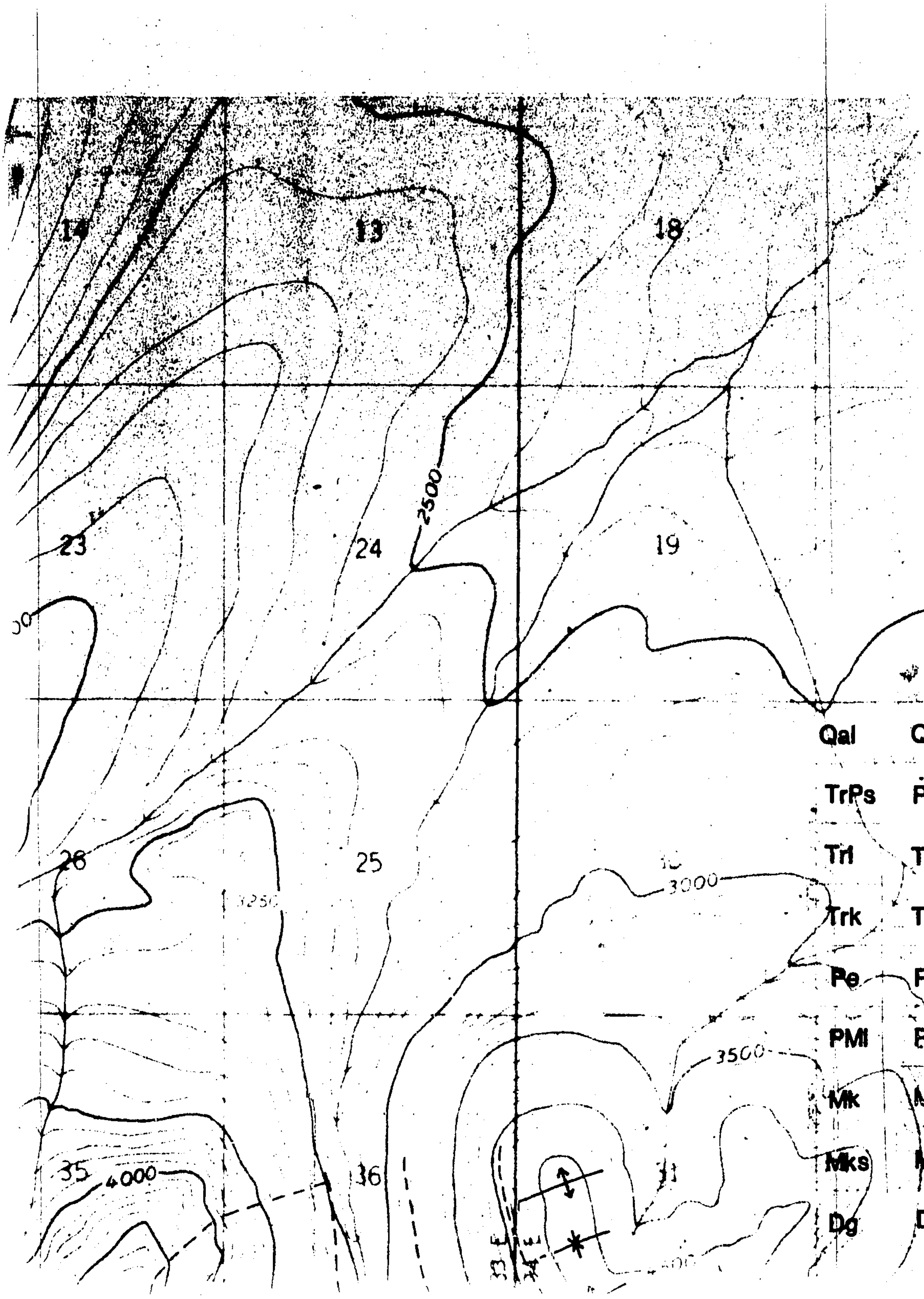
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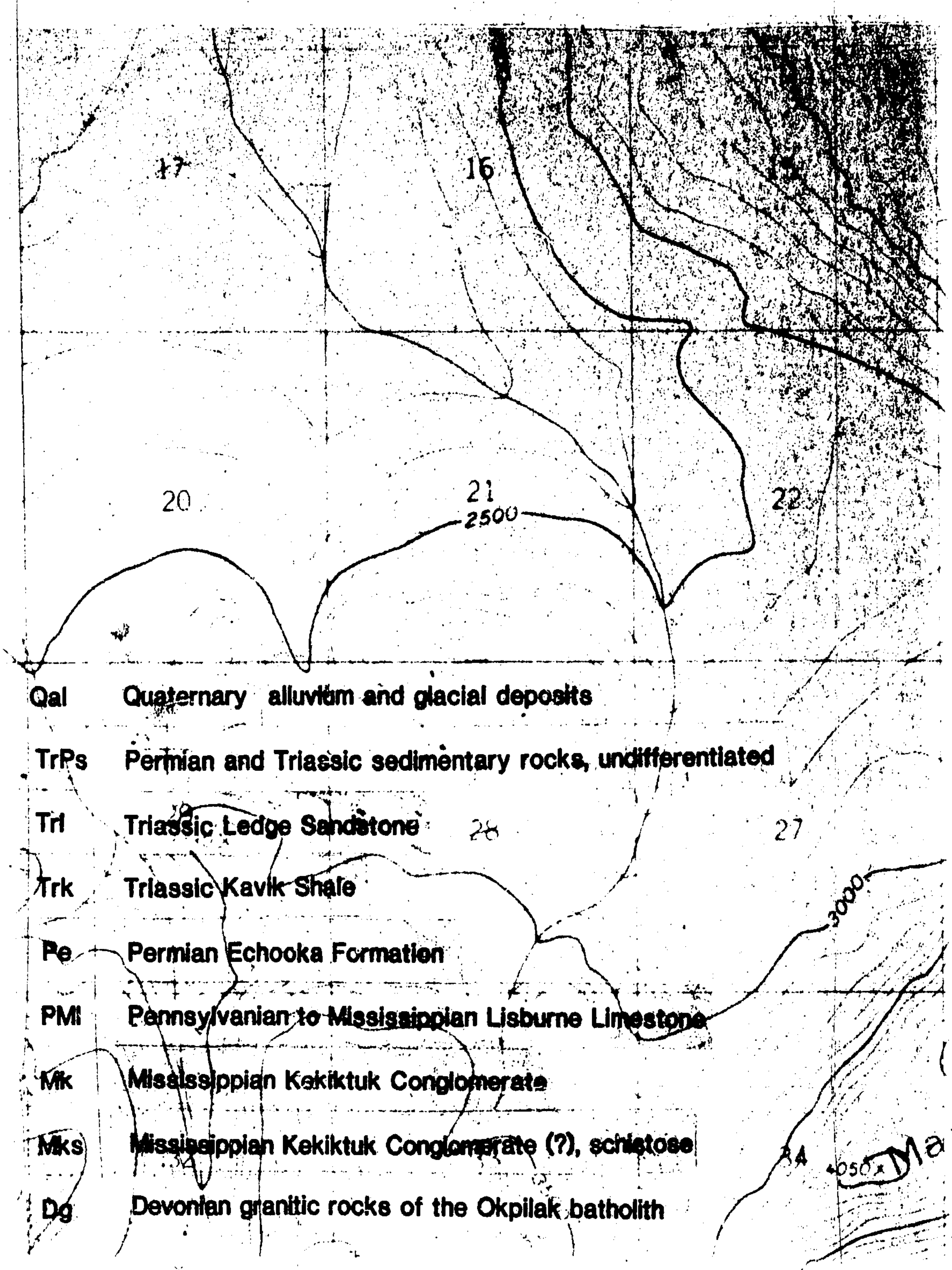

Geologic map and cross sections
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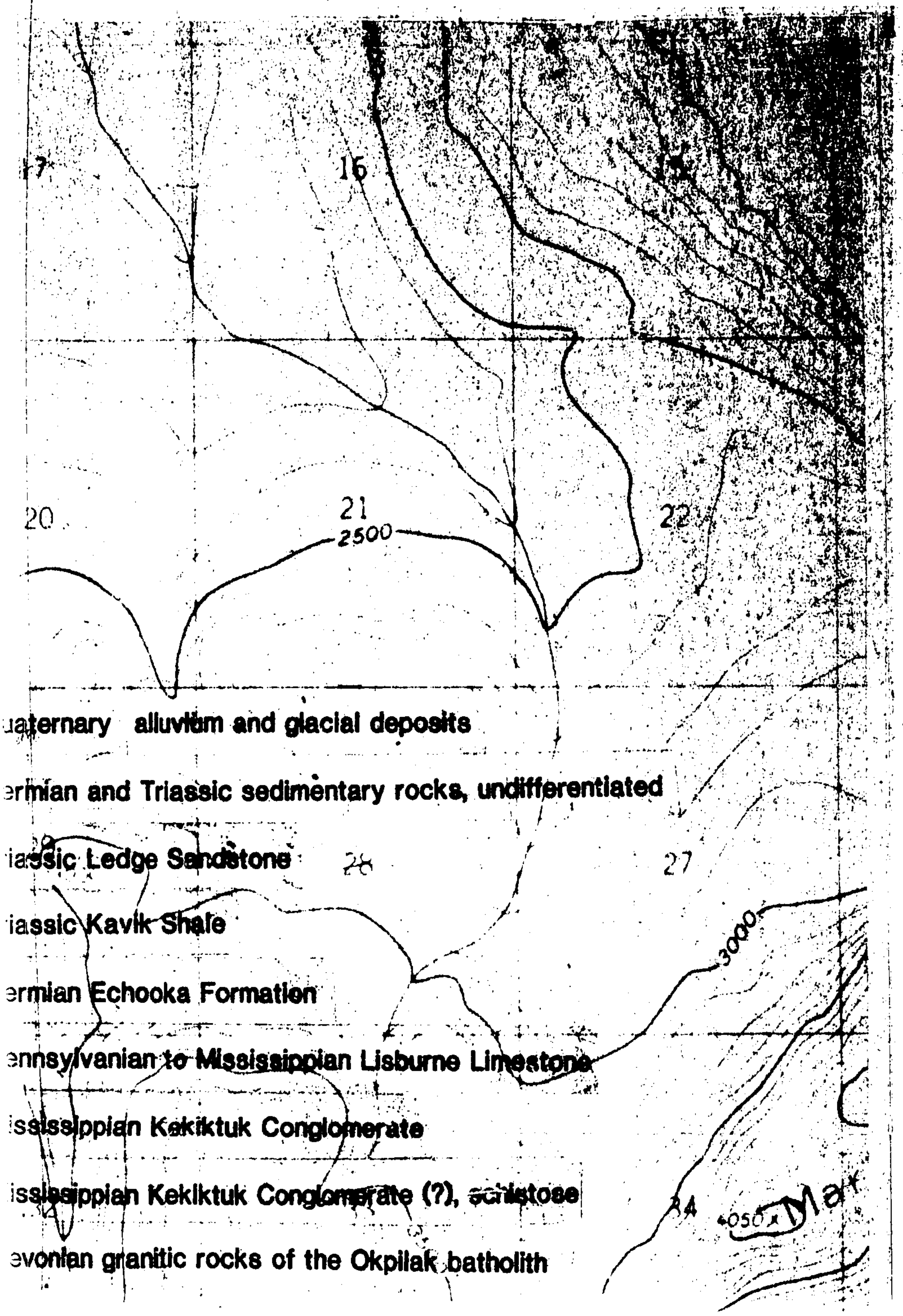

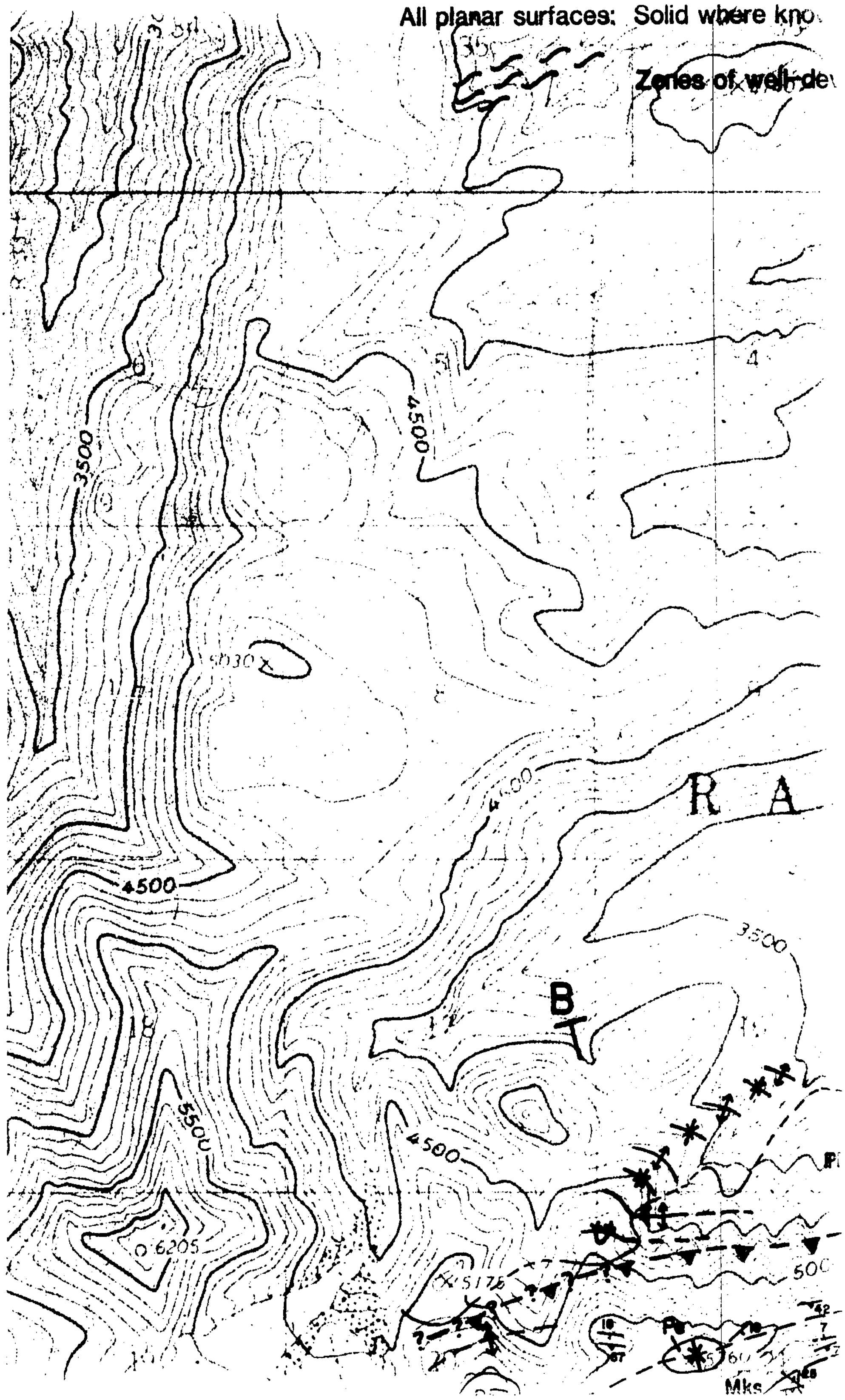
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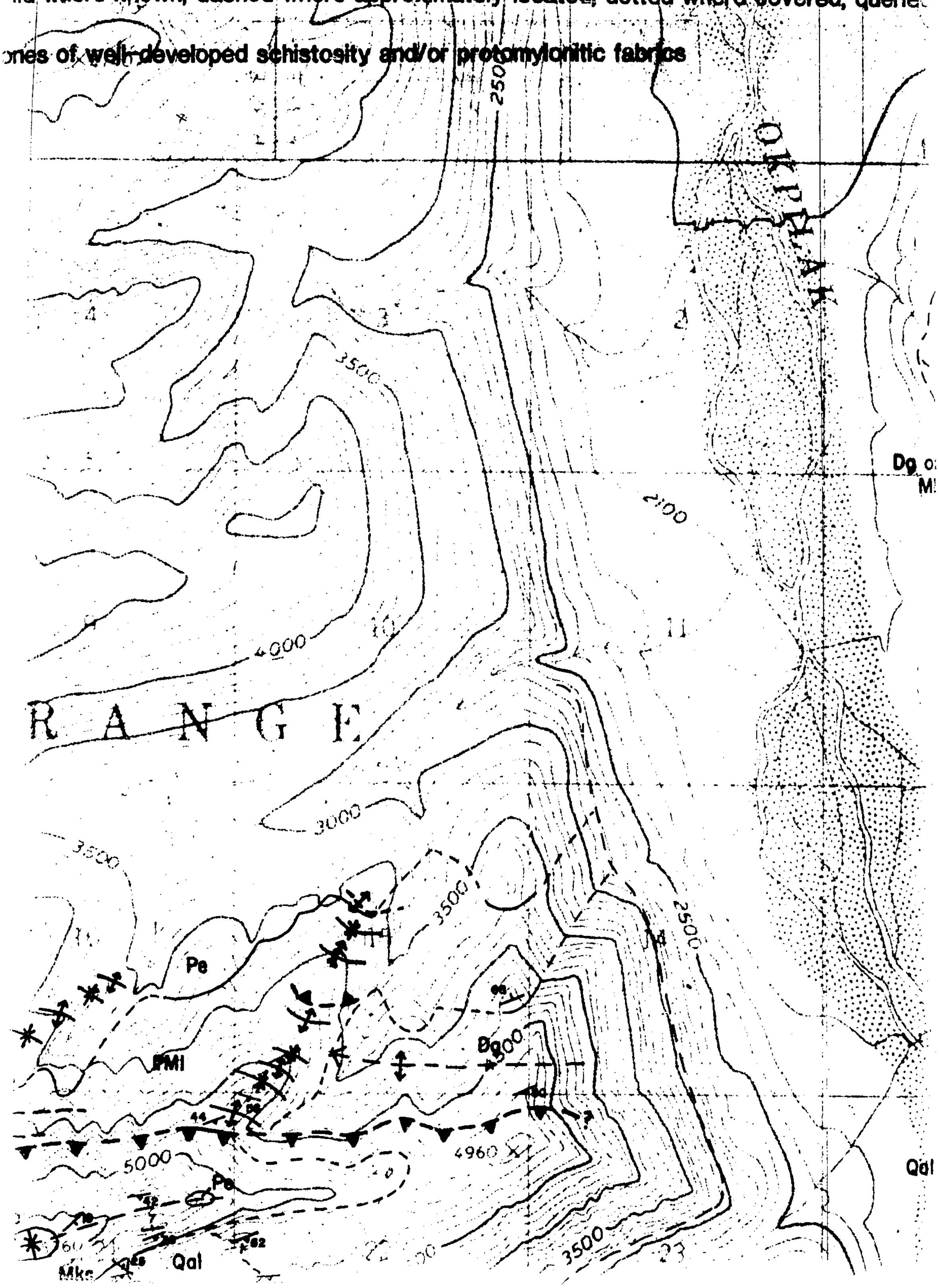

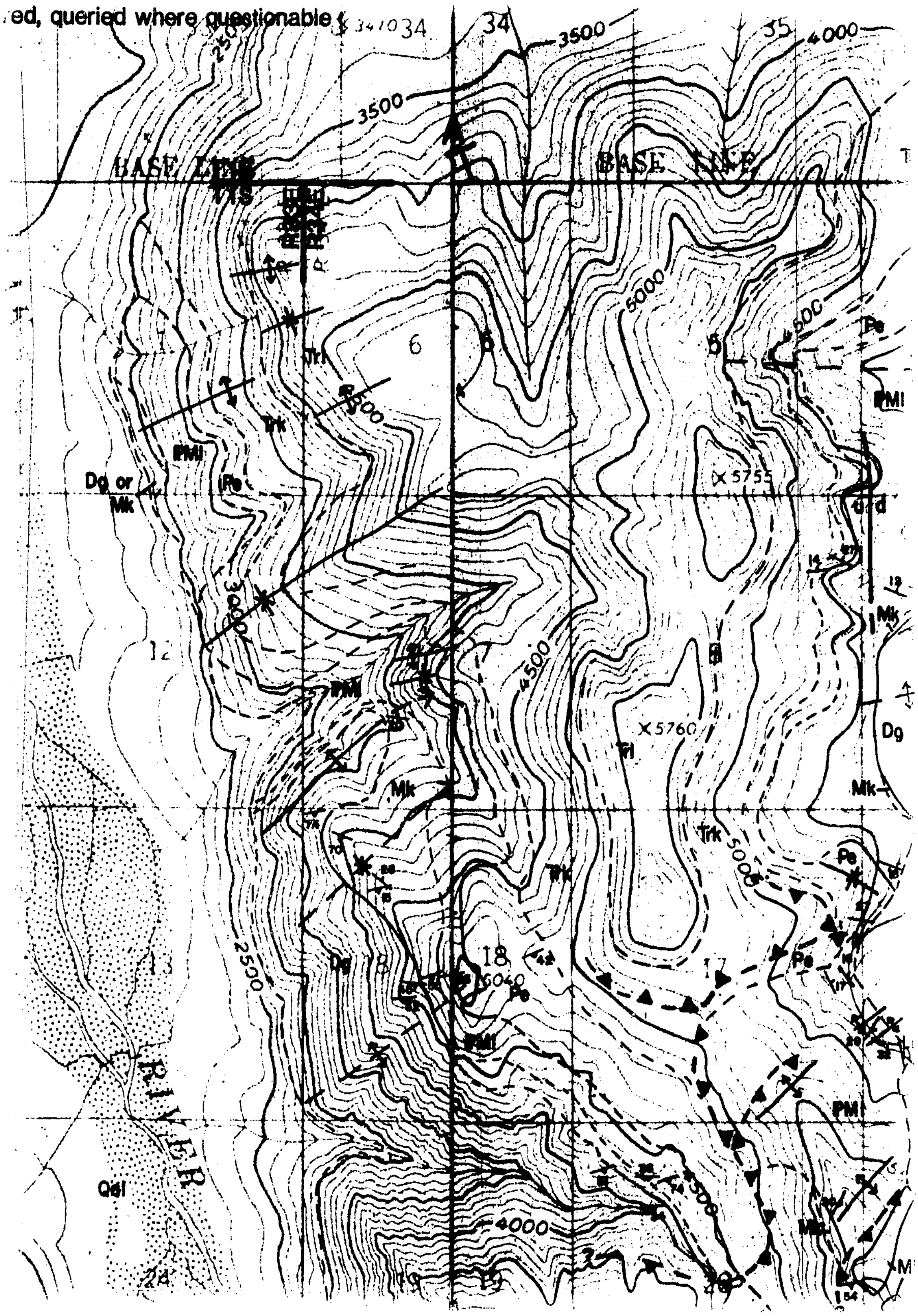

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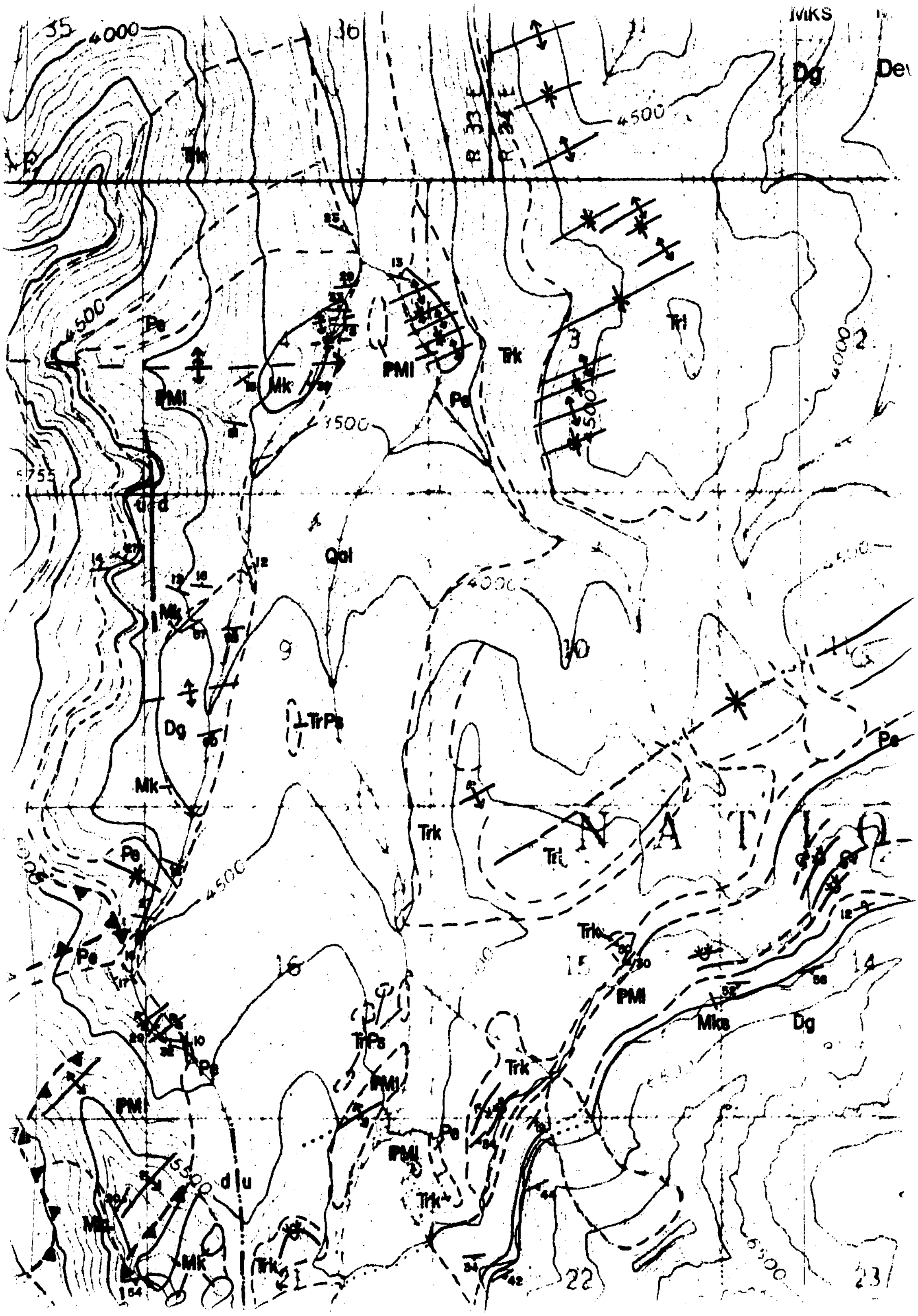
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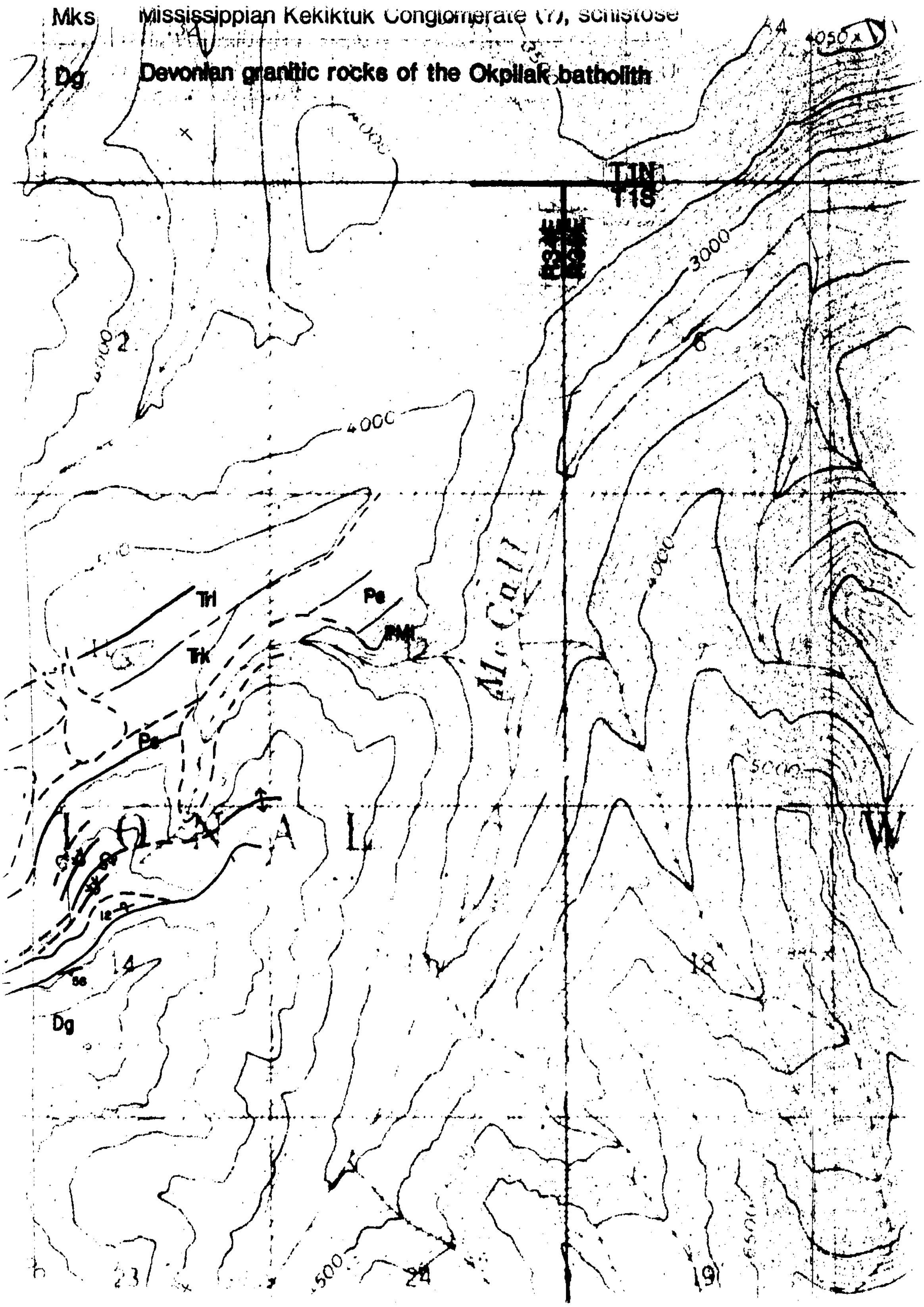
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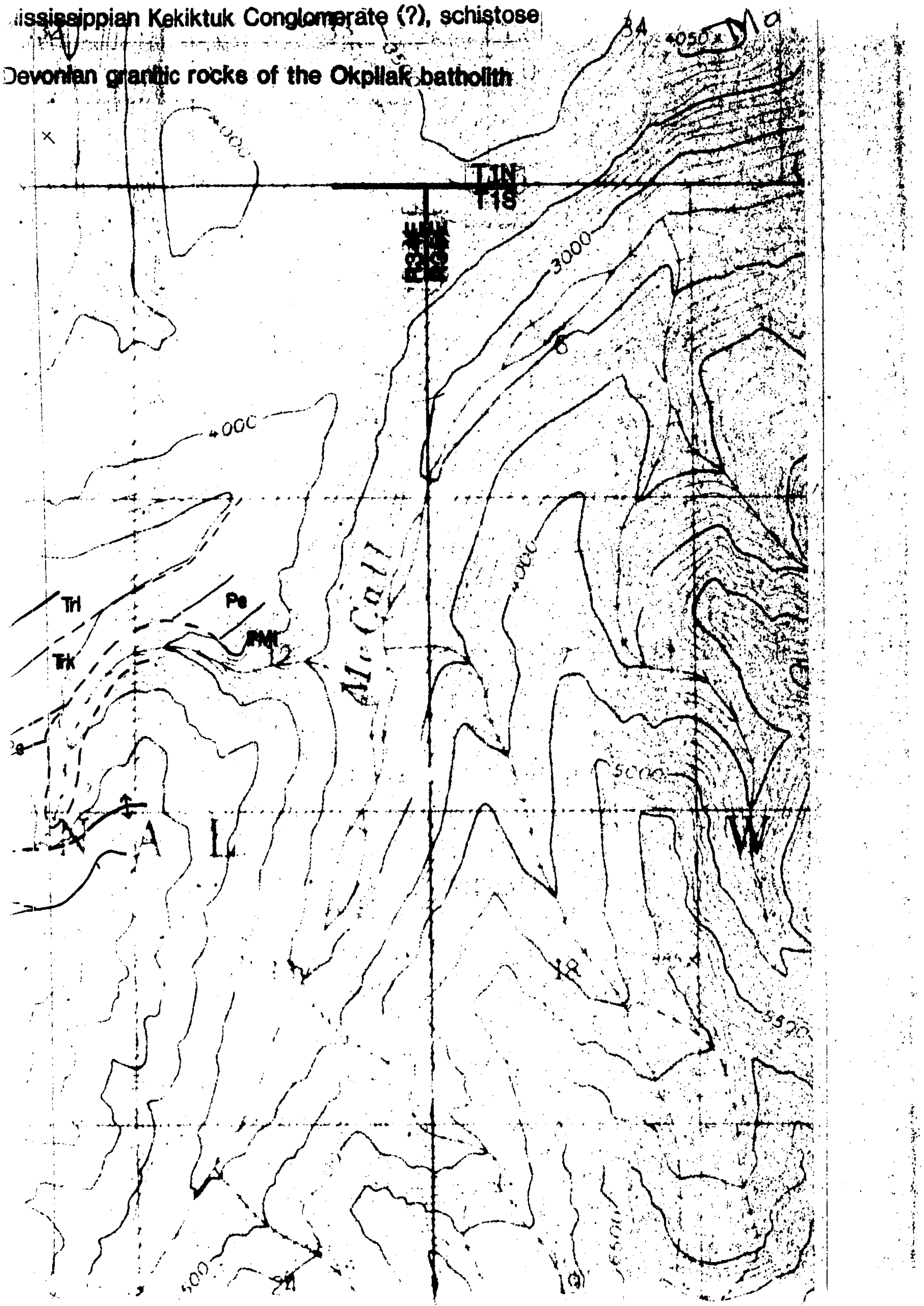
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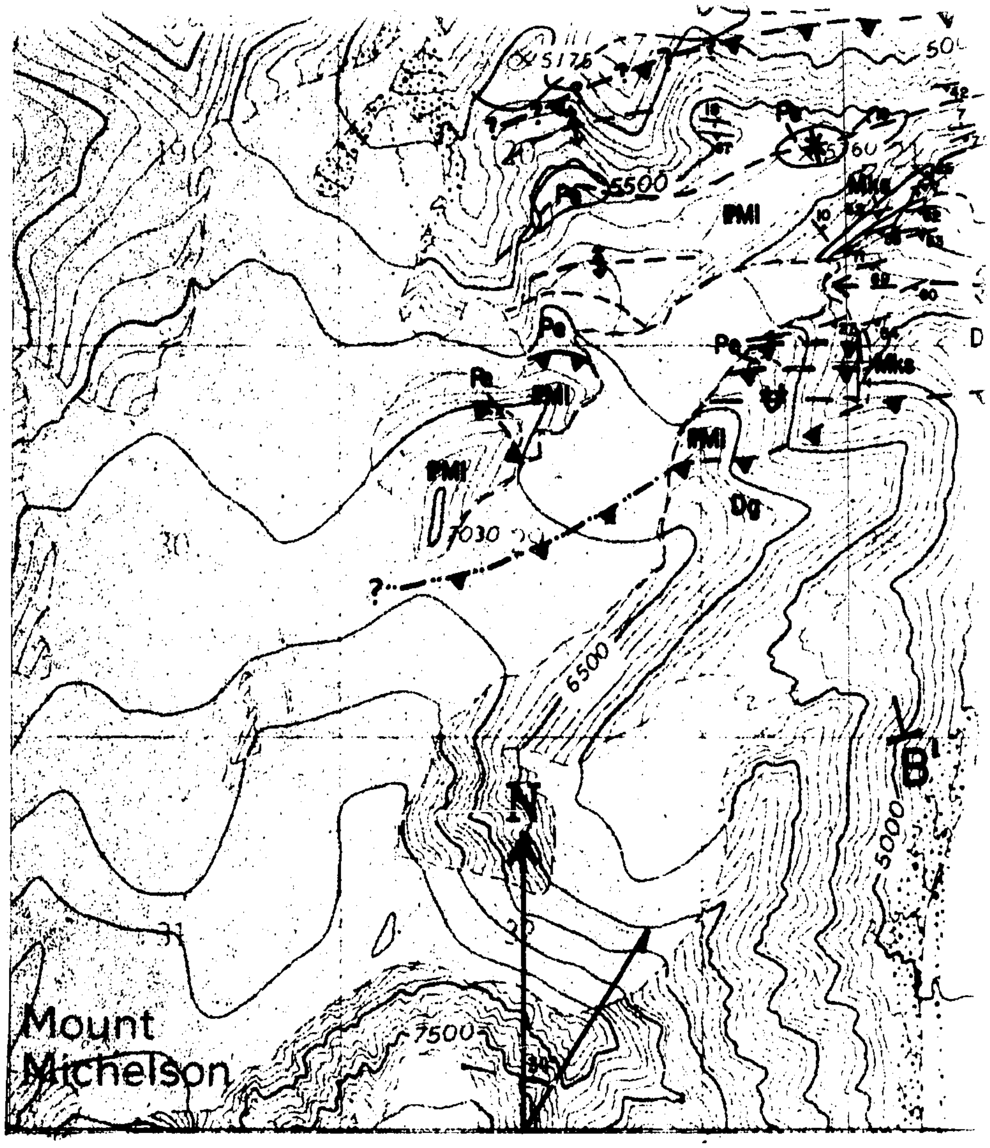
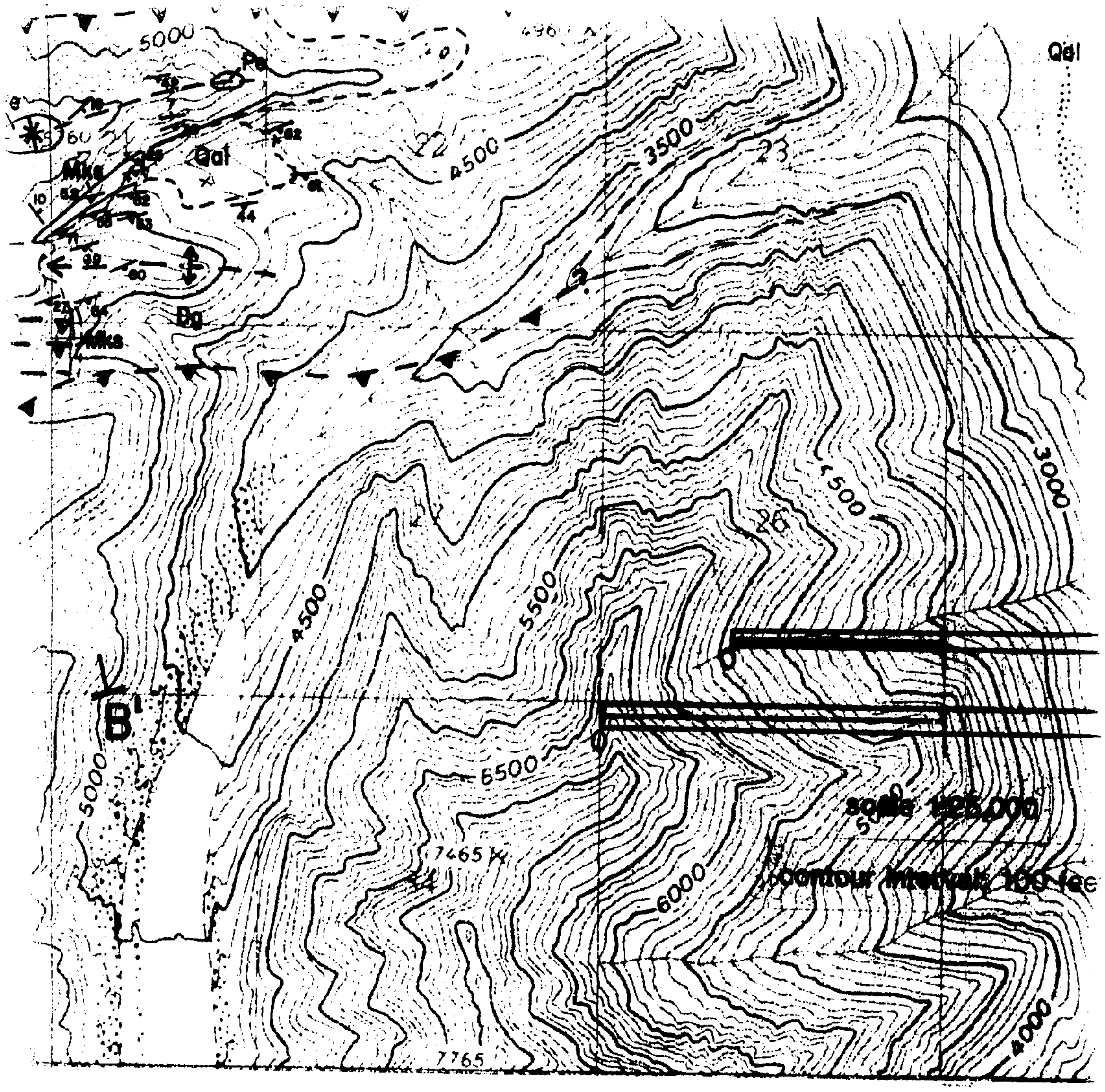


Approximate mean declination, 1955



Geologic map

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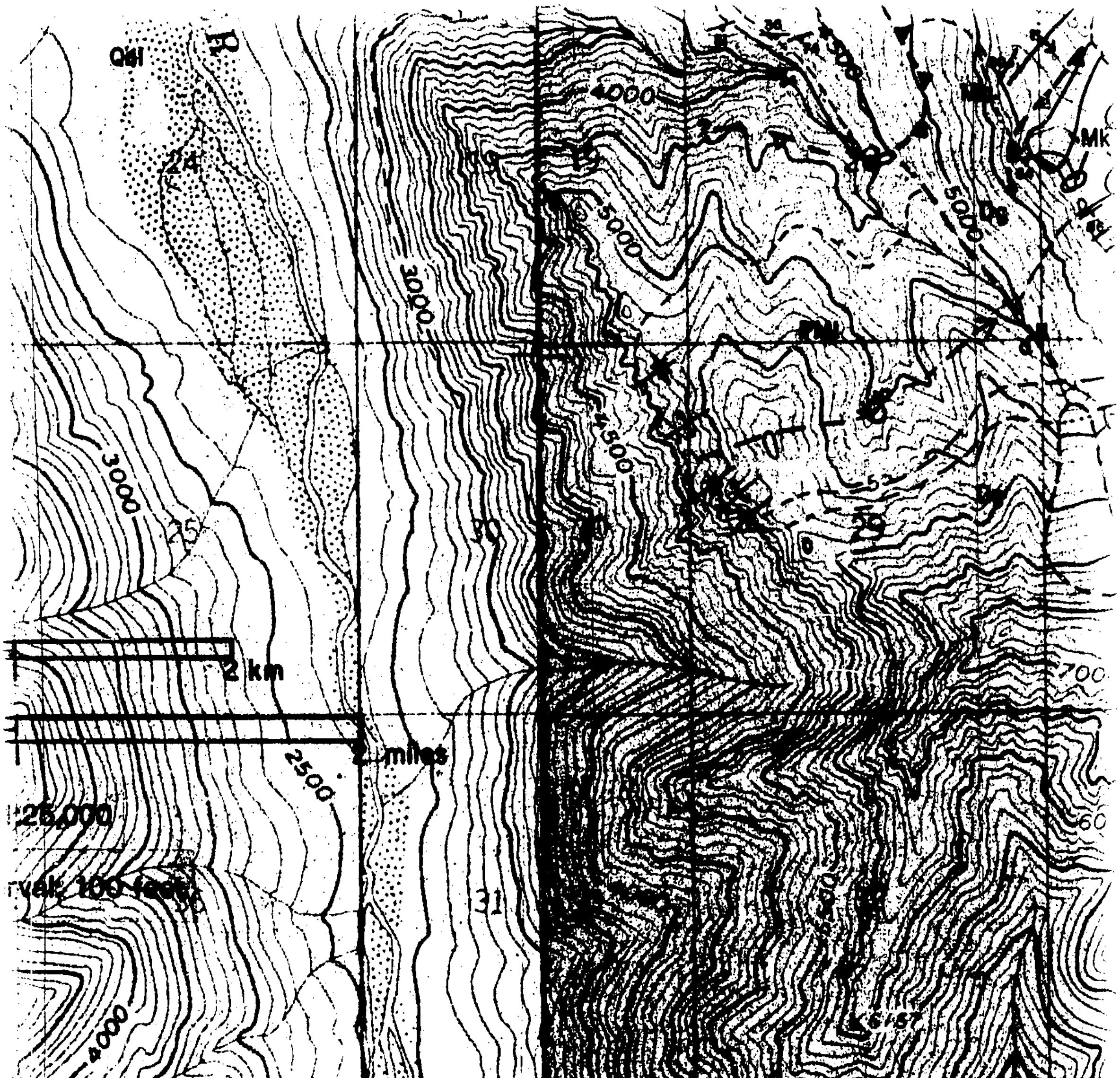

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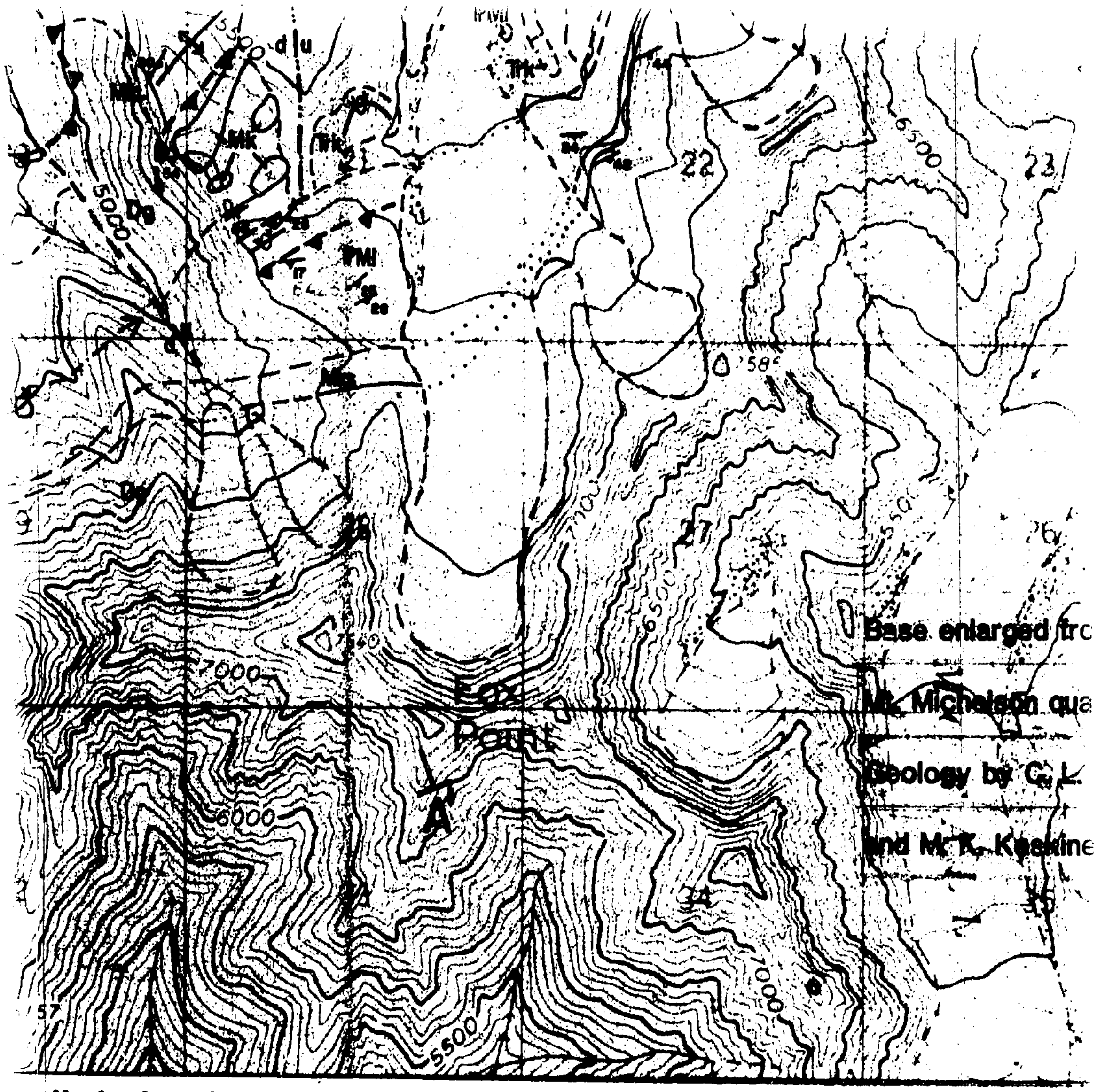

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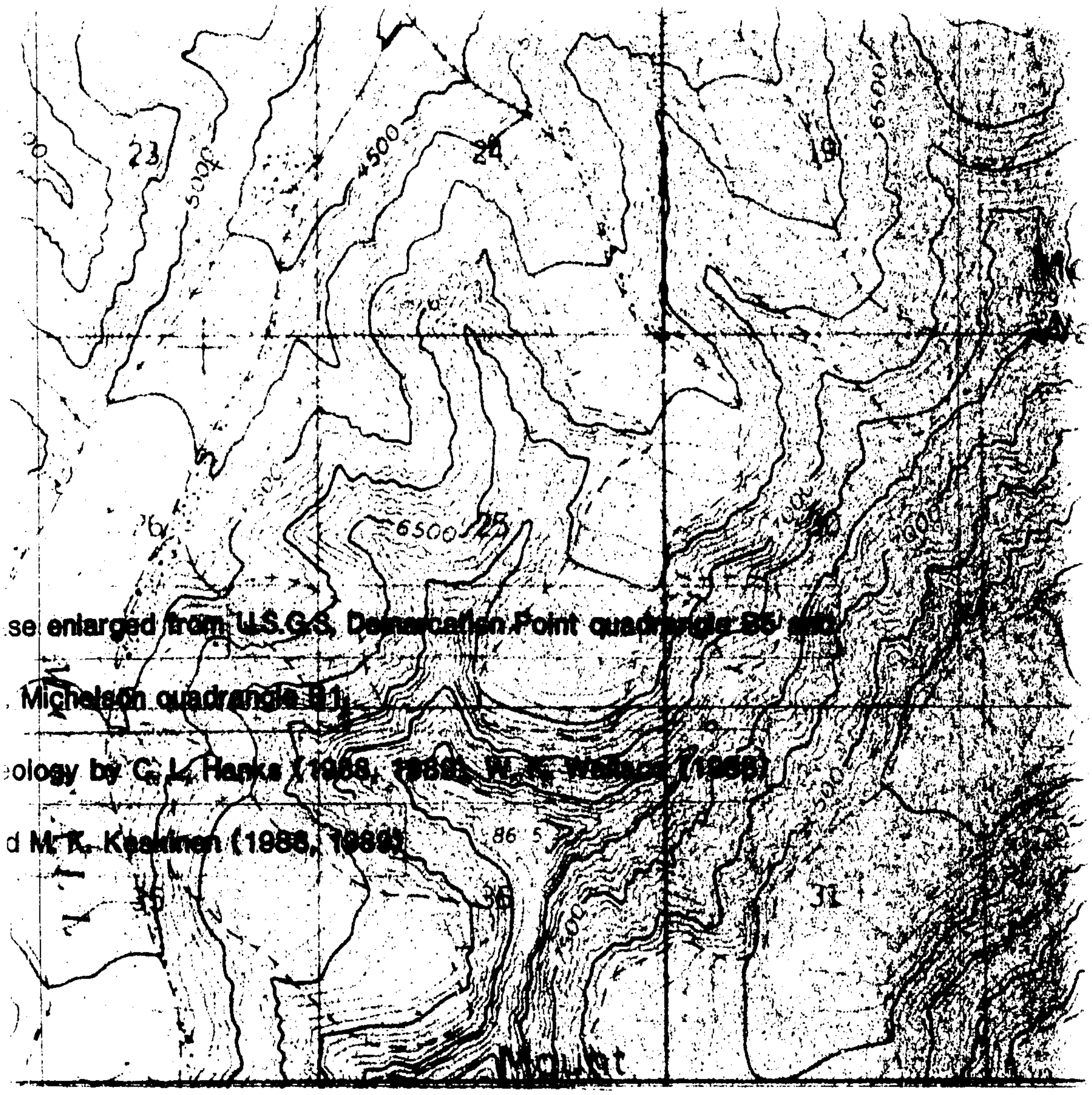
Brooks Range, Alaska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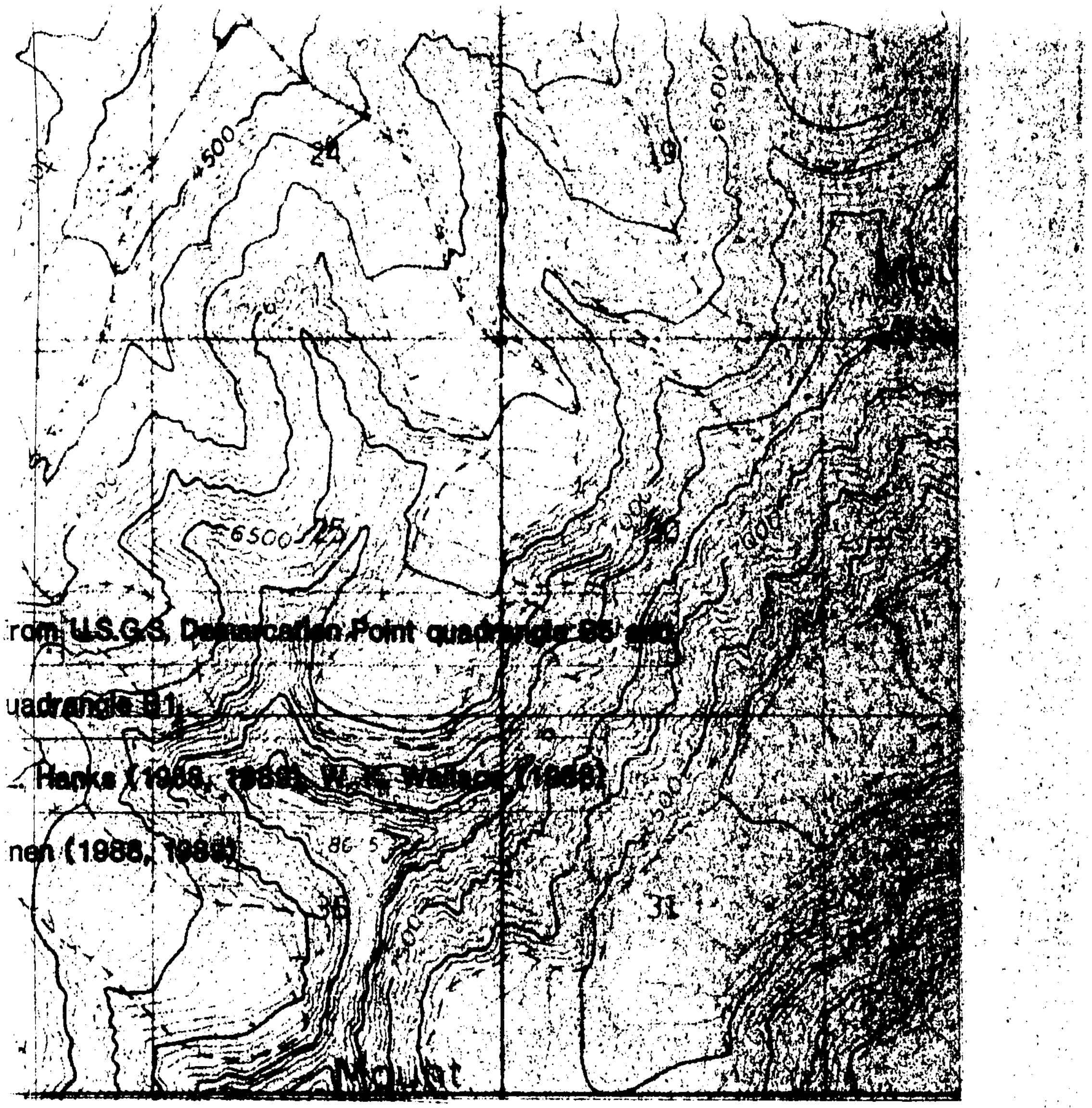

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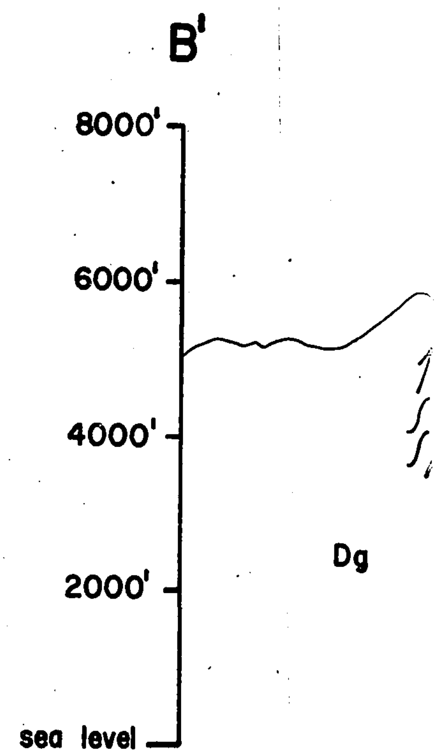

Geologic ma

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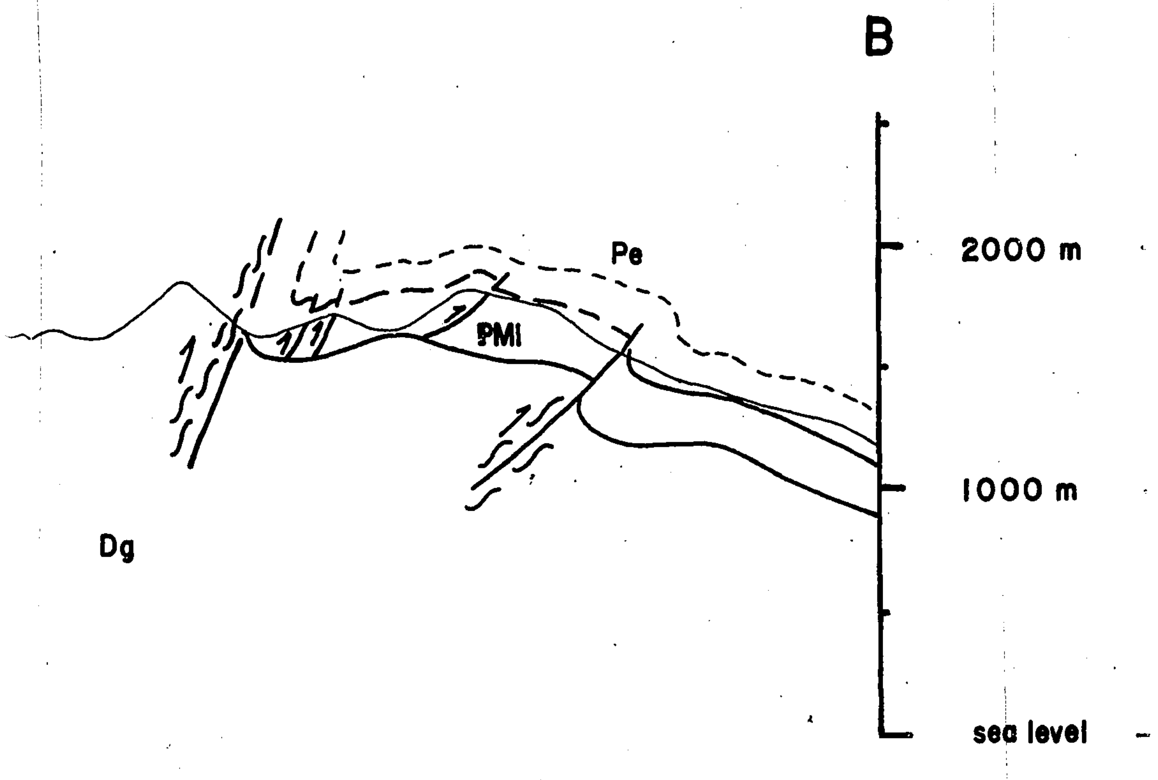

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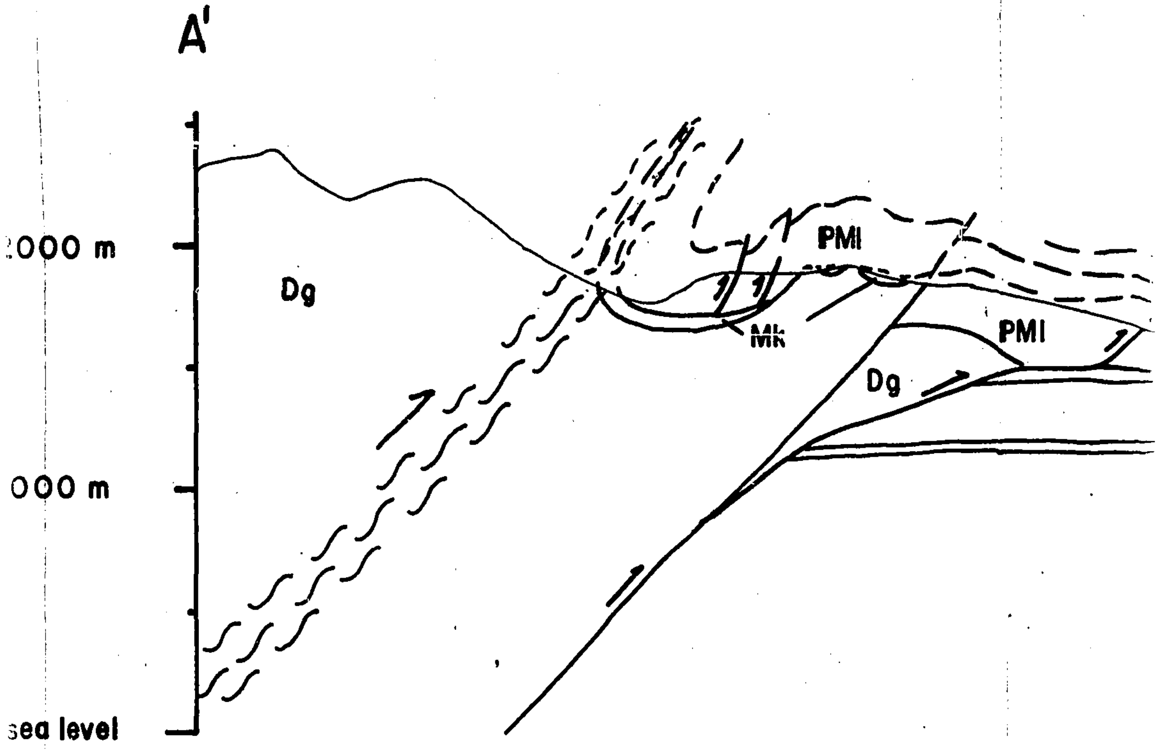

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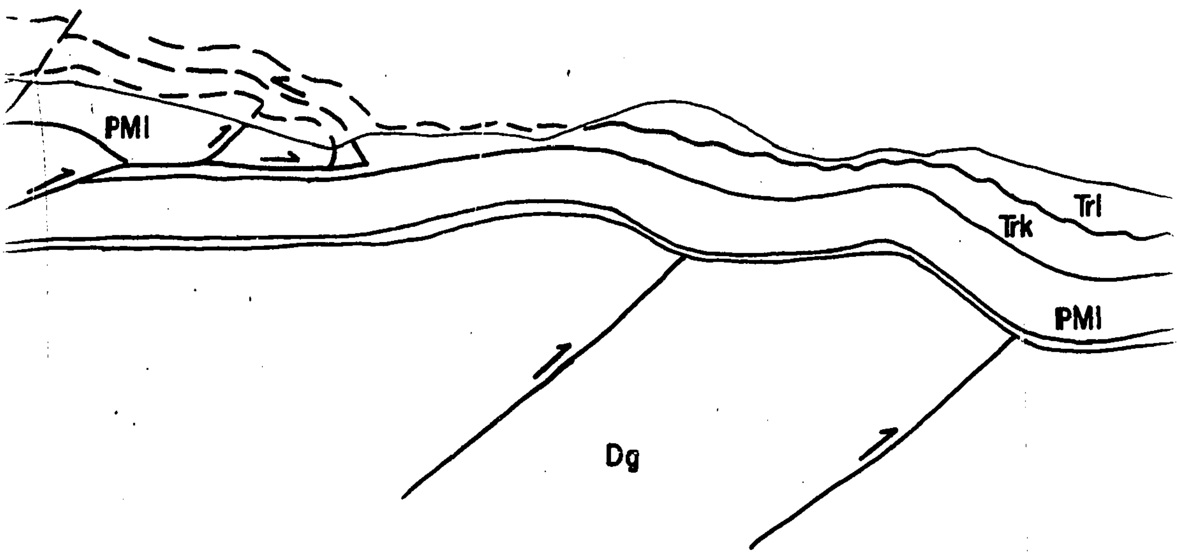
Brooks Range, Alaska



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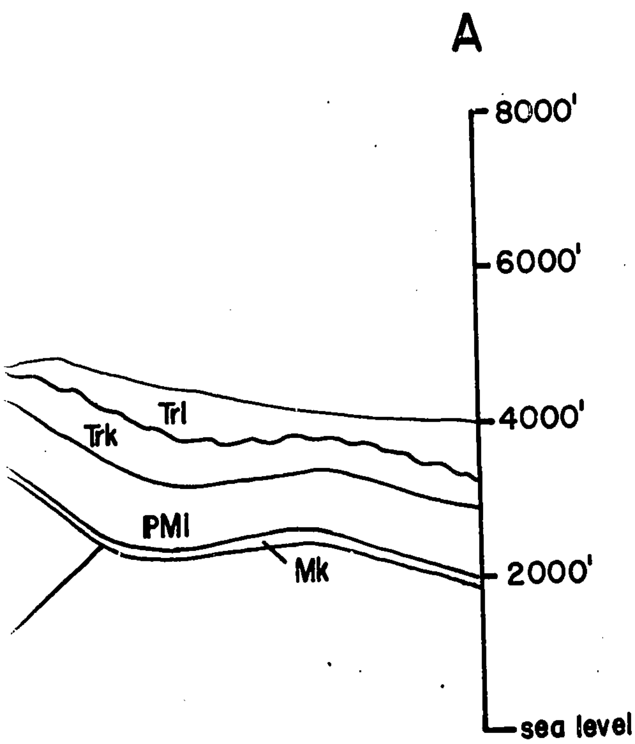
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