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**STRATIGRAPHY AND RECURRENT TECTONISM,
CAT CREEK AND RELATED STRUCTURES,
CENTRAL MONTANA**

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ABSTRACT

The Cat Creek anticline, scene of Montana's first major oil discovery, is a Laramide foreland structure east of the Big Snowy and Judith Mountains. About 60 miles long, the anticline trends ESE and has a steep northeast flank and a gentle southwest one. Arranged along the crest of the fold are eight smaller domes en echelon, and numerous northeast-striking en echelon normal faults. Underlying the steep limb of the Cat Creek anticline is the Cat Creek fault, a southwest-dipping high-angle reverse fault. The Cat Creek fault is part of a system of faults that strike east-west, ENE-WSW, and ESE-WNW. These faults reflect a long history of recurrent tectonism that involved at least three, and probably five or more separate episodes of displacement.

The Cat Creek fault is interpreted as forming part of the northern boundary of a Proterozoic failed rift, the Montana aulacogen. The fault separates Archean crystalline rocks, on the north, from slightly metamorphosed sedimentary rocks of the Proterozoic Belt Supergroup on the south. Proterozoic displacement was down to the south along high-angle, listric normal faults.

Abrupt thinning of the Flathead Sandstone and Emerson Formation in wells south of the Cat Creek and Gumbo Ridge faults suggests reverse faulting and uplift of the southern block during Cambrian time. This displacement was opposite to the inferred Proterozoic normal displacements.

A second episode of reverse faulting on the Cat Creek and Gumbo Ridge faults took place near the end of the Devonian Period. The blocks south of the faults were elevated 100 to 600 feet, and Devonian strata were largely eroded from the uplifted blocks prior to Mississippian sedimentation. This compressional deformation may have been related to the Antler orogeny west of Montana.

The Cat Creek region was stable tectonically from Kinderhookian through early Chesterian (Mississippian) time. An east-trending trough (Big Snowy trough) developed south of the Cat Creek fault during Chesterian and early Pennsylvanian time.

A major episode of normal faulting took place after Atokan (Pennsylvanian) and before Middle Jurassic time. Most faults in the study area underwent displacements of several hundred feet to more than 1,500 feet down to the south. Pennsylvanian and Chesterian strata were eroded from the uplifted blocks. Limestones formed prominent scarps under the arid climate prevailing at the time. The eroded terrain gradually was overlapped by Middle Jurassic marine sediments.

Relatively little tectonic activity occurred in the study area from Middle Jurassic through the end of Cretaceous time. Intense compressional deformation in western Montana (Sevier orogeny) induced only mild warping of the foreland. Clastic sediments were shed from rising mountains in western Montana and deposited in the study area. Initial uplift of the Cat Creek anticline may have commenced during the Campanian Stage (Late Cretaceous, Judith River Formation), but the main deformation of the Laramide orogeny in central Montana was early Tertiary. Reverse and left-lateral displacement along pre-existing faults, with forced folding of sedimentary cover, were the principal effects of Laramide deformation.

Oil has been produced from only three out of eight domes along the Cat Creek anticline. Hydrocarbons probably were generated mainly during Cretaceous time, prior to the Laramide orogeny. Pre-Laramide structures trapped hydrocarbons along the eastern part of the Cat Creek fault, but permitted them to escape up-dip along the central and western portions of the fault.

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INTRODUCTION

The Cat Creek anticline attracted early attention as the scene of the first significant oil discovery in Montana. Cat Creek is a Laramide foreland structure on the central Montana plains east of the Big Snowy Mountains. It is an asymmetrical, compound anticline about 60 miles long having 9 separate areas of closure. Drilling shows that the steep northeast flank is underlain by a fault (Cat Creek fault) that has undergone several reversals of displacement through geologic time. These reversals are mentioned in extant literature, but nowhere are they discussed systematically. Good surface exposures, abundant well data, and available seismic reflection profiles make Cat Creek an excellent place to study a case history of recurrent tectonism, with implications for petroleum accumulation.

Three major episodes of displacement (Late Devonian, post-Pennsylvanian/pre-Jurassic, and post-Cretaceous) are demonstrated for the Cat Creek fault, and two more (Proterozoic and Late Cambrian) are strongly indicated. In addition, warping or flexure along the Cat Creek fault influenced Mississippian, Pennsylvanian, Jurassic and Cretaceous sedimentation. Subsurface mapping reveals that the Cat Creek fault is part of a network of generally east-trending, recurrently active faults, some of which have no surface expression.

Only three of the nine areas of structural closure on the Cat Creek anticline have yielded oil in 70 years of test drilling. Several explanations of why production is so restricted have been put forward. This study suggests that pre-Laramide structural deformation was the key to localizing hydrocarbons to certain portions of the anticline.

This report focuses on the tectonic history of the Cat Creek area, as deduced from stratigraphic trends. A second report (Nelson, 1993) deals with structural geology of Laramide folds and faults in the Cat Creek area.

Location and geologic setting

The study area is a rectangle that covers Townships 12N through 20N and Ranges 20E through 40E (fig. 1). This rectangle is approximately 126 miles east-to-west and 54 miles north-to-south, and has an area of about 6,800 square miles. Eastern Fergus, nearly all of Petroleum, western Garfield and a small part of northern Rosebud Counties are included.

Central Montana is situated in the northern Great Plains, east of the main ranges of the Rocky Mountains. The study area is mainly rolling grasslands at elevations of 2,500 to 4,000 feet above sea level. Steep bluffs, deep coulees and badlands occur along major streams, such as the Musselshell River. Several isolated groups of mountains rise 2,000 to 4,500 feet above surrounding plains; among them are the Big Snowy and Judith Mountains at the western edge of the study area and the Bearpaw and Little Rocky Mountains north of the area (fig. 1).

The plains of central Montana are underlain by about 5,000 to 9,000 feet of Phanerozoic sedimentary rocks that rest upon a basement of dominantly Archean crystalline rock of the Wyoming province. To the west is the fold-thrust belt of the Sevier orogeny (Late Jurassic-Eocene); to the northeast is the Williston basin, an intracratonic basin. In central Montana, the Central Montana uplift (also called Big Snowy anticlinorium) is a compound structure, comprising several domal and anticlinal structures, among them the Cat Creek anticline (fig. 1). Some components of the Central Montana uplift, such as the Cat Creek and Devil's Basin anticlines and Porcupine dome, are fault-bounded uplifts of crystalline basement. Other components of the Central Montana uplift, such as the Judith and Moccasin Mountains, are cored by igneous intrusions. Both the fault-bounded uplifts and the igneous activity were products of the early Tertiary Laramide orogeny.

Previous research

The earliest geologic maps that cover part of the Cat Creek area are those of Bowen (1914, 1915, and 1920). A more detailed map of the eastern part of the Cat Creek anticline was published by Lupton and Lee (1921). Reeves (1927) mapped a large area including the Cat Creek and Devil's Basin anticlines at a scale of 1:125,000. Johnson and Smith (1964) remapped much of the same area at a scale of 1:63,360. All of these maps are accompanied by reports that describe outcropping Cretaceous rocks and make some mention of structure. Porter (1990) mapped the Eagle Sandstone and older rocks in the Winnett 30 X 60-minute quadrangle (scale 1:100,000).

Other geologic mapping in the report area is indicated in fig. 2. Most of the relatively large-scale mapping was conducted in the southwestern quarter of the area, where outcrops are more numerous and structure more complex than elsewhere. No published maps at scales larger than 1:500,000 are available for much of the northern and eastern portions of the study area. In these poorly-mapped areas the surface bedrock is largely flat-lying Upper Cretaceous and Cenozoic rock.

Numerous maps and reports deal with various aspects of subsurface geology in the study area and adjacent regions. These publications will be cited at appropriate points in the text.

Methods of study

For this report the bedrock geology of the Cat Creek oil field (West, Mosby and East domes) was mapped at a scale of 1:24,000 (map is published in Nelson, 1993). In addition, many outcrops along the Cat Creek anticline west of the oil field were visited, and small areas were mapped geologically.

Subsurface mapping made use of oil-test well records on public file at the Montana Oil and Gas Conservation Division in Billings. Additional well records were viewed at the offices of Cenex, Inc., and water-well records were examined at the MBMG in Butte. All available records (more than 1,000) were examined for the area outlined in fig. 1. Among these records, approximately 700 were deemed sufficiently reliable to be used in constructing isopach and structure maps and cross-sections. Reliable records are those that include geophysical logs and/or sample studies and core descriptions made by geologists. Drillers' logs of oil-test holes and water wells were used only where no nearby wells having better data were available, and where the drillers' logs were stratigraphically consistent with better logs from the area. Subsurface interpretation of Cretaceous rocks was based on direct comparison of lithologies from outcrop to core and sample studies. For sub-Cretaceous rocks interpretations are based on comparison of well records with published descriptions of the various units in their type areas.

STRUCTURAL GEOLOGY

The focus of this paper is the tectonic history of the Cat Creek area, as deduced mainly from stratigraphic evidence. Details of structural geology of the area are presented in Nelson (1993). A summary of structural geology is presented here to provide background for the discussion of tectonic history.

The study area contains a system of interconnected faults that strike east-west, ENE-WSW, and ESE-WNW (fig. 2). Most of these faults do not reach the surface; they are known primarily from subsurface information, including wellbore penetrations of fault surfaces, displacements of rock at depth, and seismic reflection profiles. Some buried faults are expressed at the surface as monoclines, asymmetrical anticlines and belts of small en echelon faults. In general, competent Paleozoic rocks underwent brittle deformation (shearing), whereas the shale-dominated Mesozoic rocks near the surface exhibited ductile behavior under stresses of the Laramide orogeny (post-Cretaceous). Many faults that were active during the Laramide orogeny underwent one or more earlier episodes of movement. Some faults, such as the Armells and North Judith faults, were active in pre-Laramide but not Laramide time.

The name Cat Creek fault is used here for the fault that underlies the steep northeast limb of the Cat Creek anticline. It is the best-documented subsurface fault in the study area, and it exhibits the greatest displacements. The Cat Creek fault is a slightly listric, southwest-dipping high-angle fault. During the Laramide orogeny it underwent oblique slip, with as much as 4,000 feet of vertical separation plus nearly one mile of left-lateral strike-slip. Four boreholes penetrate the fault surface, as evidenced by repeated section. The fault also is apparent on (proprietary) reflection seismic profiles. The belt of en echelon domes and normal faults along the crest of the Cat Creek anticline are products of left-lateral wrenching along the Cat Creek fault.

The East Cat Creek fault extends eastward from the eastern terminus of the Cat Creek fault in southern Garfield County (fig. 2). Closely-spaced wells along the western part of the East Cat Creek fault indicate displacements of 700 to 1,200 feet down to the north in Lower Cretaceous rocks. In Upper Cretaceous rocks at the surface, the fault is expressed as a north-facing monocline accompanied by a belt of northeast-trending en echelon faults. A seismic profile published by Stephens and Rodenbaugh (1985) portrays the East Cat Creek fault as an upward-branching zone of vertical and south-dipping faults. Thus, the East Cat Creek fault, like the Cat Creek fault, underwent left-lateral oblique slip during the Laramide orogeny.

The Gumbo Ridge fault lies south of and strikes parallel to the East Cat Creek fault in northernmost Rosebud County (fig. 2). The name is taken from the Gumbo Ridge oil field, which lies near the fault. Abrupt thickness changes in Paleozoic units in wells along the fault indicate possible displacements beginning during the Cambrian Period. Cretaceous rocks show little vertical offset, but a belt of en echelon faults and anticlines has been mapped at the surface (Schulte, 1959). These structures indicate that Laramide displacement of the Gumbo Ridge fault was predominantly left-lateral, with little or no dip-slip component.

The Sand Creek fault as named by Plawman (1983) and Stephens and Rodenbaugh (1985), is a southeast-dipping that begins just east of the Cat Creek fault and strikes ENE, curving to an eastwest heading and joining the Cohagen fault (fig. 2). Seismic and borehole data indicate that the Sand Creek fault was a listric normal fault before Middle Jurassic time and a reverse fault during the Laramide orogeny. The Laramide throw on Lower Cretaceous rocks in the subsurface is 650 to 900 feet down to the north. Surface structure along the Sand Creek fault is poorly documented, although Plawman (1983) stated that a complex pattern of en echelon faults appears on (unpublished) structure maps of the Eagle Sandstone (Upper Cretaceous) along the fault. Evidently, this is another left-lateral oblique fault.

The Cohagen fault trends ESE near the village of Cohagen in eastern Garfield County (fig. 2). The Cohagen fault joins the Sand Creek fault at its western end. Drilling indicates Laramide displacement of 700 to 1,100 feet down to the northeast, and pre-Jurassic displacement of 400 to 600 feet in the opposite direction.

The Weldon fault was first identified at the surface near Weldon in McCone County, Montana by Collier and Knechtel (1939), and was extended into the subsurface by Edmiston and Foster (1969). The Weldon fault strikes northeast; its southwestern end extends into the study area (fig. 2). The only documented displacement of the Weldon fault within the study area is post-Pennsylvanian, pre-Jurassic throw of about 1,000 feet down to the southeast. In McCone County, northeast of the study area, the Weldon fault was active in Devonian (Sandberg and Mapel, 1967), Pennsylvanian and Jurassic (Edmiston and Foster, 1969), and post-Paleocene time (Collier and Knechtel, 1939).

The Rattlesnake Butte fault zone trends eastward through southern Petroleum County and intersects the Cat Creek and East Cat Creek faults in southwestern Garfield County (fig. 2). It is a belt of northeast-trending, en echelon faults. Near its eastern end the fault zone curves toward the northeast and is accompanied by a southeast-facing monocline (Johnson and Smith, 1964). The en echelon fault zone and monocline are interpreted as products of Laramide left-lateral oblique displacement on a fault in crystalline basement (Nelson, 1993).

The Flat Willow fault zone is slightly northwest of and parallel to the Rattlesnake Butte fault zone (fig. 2). En echelon faults at the surface (Johnson and Smith, 1964) are indicative of left-lateral displacement on a fault in crystalline basement during the Laramide orogeny. Neither the Rattlesnake Butte nor the Flat Willow fault zone show evidence of pre-Laramide activity.

The Armells fault, in the northwest part of the study area, trends eastward and may intersect the Cat Creek fault (fig. 2). The Armells fault has no surface expression, and is known only from borehole data. Post-Mississippian, pre-Middle Jurassic throw of as much as 800 feet down to the south is indicated.

The North Judith fault lies a few miles south of the Armells fault and forms a graben with the latter (fig. 2). Borehole data indicate pre-Jurassic throw of 400 to 600 feet down to the north. Jurassic and younger rocks apparently are not displaced.

The Crystal Peak fault is south of and parallel with the North Judith fault (fig. 2). Subsurface data reveal a post-Pennsylvanian, pre-Middle Jurassic displacement of 400 to 700 feet down to the north. Laramide movement on the Crystal Peak fault was in the opposite direction; Cretaceous rocks are downthrown 200 to 400 feet to the south.

STRATIGRAPHY

Precambrian rocks

Precambrian rocks of central Montana are concealed by sedimentary cover except in small areas of the Big Snowy, Little Rocky and North Moccasin Mountains. Additional information comes from deep well borings and from (proprietary) seismic reflection profiles. These data show two types of Precambrian rocks in central Montana. The older type comprises plutonic and high-grade metamorphic rocks, chiefly granite, gneiss, schist and amphibolite. These older rocks are assigned to the Wyoming province and yield Archean radiometric ages of 2,500 Ma and older (Hedge et al., 1986). Younger rocks consist of unmetamorphosed or slightly metamorphosed quartzite, siltite, argillite and carbonate rocks assigned to the Proterozoic Belt Supergroup, approximately 800 to 1,500 Ma (Hedge et al., 1986).

Data on Precambrian rocks in and near the study area are presented in Table 1 (below) and in fig. 3.

Table 1.
Precambrian rocks in central Montana

	Site	Rock type	Assignment	References
1.	Little Rocky Mts.	Schist, gneiss, quartzite, amphibolite	Pre-Belt (1710-1750 ma)	Knechtel, 1959; Hedge et al., 1986
2.	North Moccasin Mts. 20-18N-18E	Biotite schist (in xenoliths)	Pre-Belt	Lindsey, 1982
3.	Mobil, F34-30G Federal Dovetail, 30-18N-29E	Granite	Pre-Belt	Tooke Eng. log
4.	Arco, 1-32 Salzgeber, 32-15N-35E	Granite gneiss	Pre-Belt	Core chips
5.	Arco, #1-33 B.N. Coastal, 33-15N-35E	Granite gneiss	Pre-Belt	Cuttings
6.	General Petr., #5-25 N.P., 25-16N-38E	Granite	Pre-Belt	Cuttings
7.	Ohio Oil Co., #1 N.P., 9-10N-39E	Garnet-mica schist, metaquartzite	Pre-Belt	Cuttings
8.	Big Snowy Mts. 4-11N-19E	Argillite with thin carbonate beds	Belt (Newland Fm.)	Reeves, 1931; Lindsey, 1980
9.	Black Eagle, #1-35 Delaney, 35-17N-23E	Quartzite	Belt	C.A. Balster, oral 10
10.	United Petr. & Mng., #1 Woodworth, 20-14N-24E	Argillite, quartzite	Belt	Company log
11.	Marmik, Mlekush 1, 12-12N-26E	Argillite, quartzite	Belt	Cuttings
12.	Champlin, State Rattler 1, 16-13N-28E	no data	Belt	Company
13.	Marathon, B.N. 1, 9-11N-24E	Coarse-grained granite	Pre-Belt	Company log
14.	Richfield, #A-1 N.P., 5-10N-25E	Quartzite, argillite	Belt	Company log
15.	Texaco, Zoerb 6, 18-10N-27E	Argillite, siltite	Belt	Cuttings
16.	Superior, Copulos 71-22, 22-2N-21E	Mica schist	Pre-Belt	Company
17.	USGS, Madison Test Well 3, 35-2N-27E	Gneiss, 2,780 my	Pre-Belt	Blankenagel et al., 1979; Peterson, 1984

Note: The word "Company" means a list of "tops", a letter, a drilling summary sheet, or similar oil-company document other than a geologist's sample study or lithologic log. The entry "Core chips" or "Cuttings" means that I personally examined samples from the public collection at the Montana Oil and Gas Conservation Division in Billings, Montana.

Plotting these data on a map (fig. 3) suggests that the Cat Creek fault forms a boundary between Archean crystalline basement rocks on the north and Belt Supergroup rocks on the south. Seismic reflection profiles that I have viewed substantiate the presence of several thousand feet of layered rocks below the Flathead Sandstone in the Cat Creek oil field south of the Cat Creek fault. Pre-Flathead layered rocks are absent or considerably thinner north of the Cat Creek fault on these profiles. Shepard and Precht (1989, p. 456) stated, "Seismic data (in approximately T11N, R31E, Rosebud County) indicate 2,000 to 4,000 feet of Belt sediments occur draped over an irregular crystalline basement surface and overlain with slight angularity by the Middle Cambrian Flathead Sandstone." Hence, the Cat Creek fault evidently underwent displacement down to the south during deposition of the Belt Supergroup.

Sauk Sequence

The Sauk Sequence of Sloss (1963) comprises sedimentary rocks of Cambrian and Lower Ordovician (Canadian) age. The Sauk Sequence rests unconformably on Precambrian basement and is overlain unconformably by the Tippecanoe Sequence. In the study area the Sauk Sequence comprises the Cambrian Flathead Sandstone and the Cambrian to lower Canadian Emerson Formation.

Flathead Sandstone

The Flathead Sandstone is the oldest Paleozoic unit in central Montana. It was named by Peale (1893) for Flathead Pass in the Bridger Range north of Bozeman, Montana. The name Flathead Sandstone is used by nearly all geologists in central and western Montana; equivalent strata in eastern Montana are assigned to the lower part of the Deadwood Formation.

The Flathead Sandstone was completely penetrated by 8 wells in the study area and partially penetrated by 3 wells. Four additional wells completely penetrated the Flathead within 10 miles of the southern border of the area. Sample studies and core descriptions wells indicate the Flathead to be dominantly sandstone, composed of clear to light orange or pink quartz sand and granules. Fine grains generally are subangular to subrounded, whereas medium and coarser grains are well rounded, frosted and pitted. Some grains have silica overgrowths. Minor constituents include pyrite, chlorite (?) and black opaque grains. Occasional partings or thin interbeds of greenish-gray to reddish-brown, well-indurated, silty shale are present. Such lithologies are characteristic of the Flathead Sandstone in mountain ranges north and west of the study area (Blixt, 1933; Knechtel, 1959; Lindsey, 1980 and 1982).

The age of the Flathead in central Montana is Middle Cambrian. The unit is time-transgressive, becoming younger eastward across the state (Hanson, 1957; Lochman-Balk, 1972). The Flathead lies unconformably on the Belt Supergroup or older Precambrian rocks throughout Montana.

Thickness of the Flathead in the study area varies from zero (in one well) to more than 212 feet (fig. 5). Abrupt local thickness variations are evident between closely-spaced wells, as in T15N, R35E, southwestern Garfield County, where two wells a little more than a mile apart encountered 97 and 186 feet of Flathead. In T13N, R28E, southern Petroleum County, one well penetrated 143 feet

of Flathead overlying the Belt Supergroup, and another well about 1 1/2 miles south drilled through 212 feet of Flathead without reaching Precambrian rocks.

Emerson Formation

The Emerson Formation was named by Knechtel (1956) for Emerson Gulch in the Little Rocky Mountains, about 30 miles north of the report area. Knechtel defined the Emerson as an interval of 950 to 1,100 feet of interbedded shale, limestone, dolomite, and intraformational conglomerate overlying the Flathead Sandstone and underlying the Ordovician Bighorn Dolomite (fig. 4). Many geologists who work with subsurface data in central Montana use the name Emerson Formation. Equivalent rocks west of the study area are divided into several alternating shale and limestone formations (Reeves, 1931; Blixt, 1933; Lindsey, 1980 and 1982). Because I was unable to distinguish the western formations in well records, I referred the entire interval to the Emerson Formation.

The Emerson in the study area is composed dominantly of shale with interbeds of carbonate rock; carbonate beds increase in number and thickness upward (fig. 4). Shale in the Emerson is largely greenish-gray, but some is gray, reddish-brown and purple. It is typically calcareous, micaceous and glauconitic; much of it is silty to sandy, particularly in the lower portion. Thin interbeds of siltstone and sandstone are common in the lower Emerson, intertonguing with the Flathead Sandstone. Limestone beds in the Emerson are largely light gray, micritic to finely fragmental, and argillaceous. Wavy and rippled laminations and nodular bedding are described in cores. Dolomite is common in the upper Emerson. Intraformational conglomerate, composed of flat pebbles of limestone oriented at all angles to bedding and set in an argillaceous matrix, has been described in cores. Similar "flat-pebble conglomerates" are common in the Emerson and equivalent units throughout central and western Montana.

The age of the Emerson and its western equivalents is Middle Cambrian through lowermost Ordovician (Knechtel, 1956 and 1959; Hanson, 1957; Lochman-Balk, 1972). Closest to the study area, Lindsey (1980) reported Lower Ordovician trilobites and conodonts from the uppermost Emerson-equivalent rocks in the Big Snowy Mountains. No lithologic break or disconformity is known at the Cambrian-Ordovician boundary; however, the Ordovician portion of the Emerson may have been eroded beneath Middle Ordovician strata in parts of the study area.

The Emerson Formation grades downward and intertongues with the Flathead Sandstone. The contact is time-transgressive, becoming younger eastward (Hanson, 1957; Lochman-Balk, 1972).

Thickness of the Emerson in the study area varies from 50 to 1245 feet (fig. 5). The Emerson is less than 800 feet thick in all of the wells drilled south of the Cat Creek fault. North of the fault, and in the northeastern part of the study area, the Emerson is thicker than 800 feet in all but one well out of six.

Abrupt changes in thickness of the Flathead and Emerson Formations between two closely-spaced wells suggest that the Cat Creek fault was active during Sauk sedimentation. The wells are located about 4 1/2 miles apart on opposite sides of the fault (fig. 6). The #1-15 Delaney well southwest of the fault drilled through 800 feet of Emerson Formation directly overlying quartzite of the Belt Supergroup. The basal 30 to 40 feet of the Emerson are sandy in this well, but the Flathead Sandstone is absent. Northeast of the Cat Creek fault the #1 Sharptail well penetrated 1,105 feet of Emerson Formation and 187 feet of Flathead Sandstone without reaching Precambrian rocks. Correlation of

geophysical-log markers within the Emerson indicates that nearly all of the thickness difference between the two wells is in the basal part of the formation. This evidence suggests that the northeast side of the Cat Creek fault was downthrown during deposition of the Flathead and lower Emerson.

The Emerson Formation is anomalously thin - only 50 feet - in the Conoco No. 1 Kincheloe well located in Sec. 10, T12N, R32E, Rosebud County. This well penetrated 197 feet of Flathead Sandstone below the Emerson without reaching basement (fig. 7). The Kincheloe well is about 2 miles south of the Gumbo Ridge fault. North of the Gumbo Ridge fault and 3 miles northeast of the Kincheloe well, the Arco No. 1-36 State well reached total depth 40 feet into the Emerson Formation. A direct comparison of the Emerson between the Kincheloe and State wells is not possible, but in the next nearest wells that completely penetrated the Emerson north of the Gumbo Ridge fault, the formation was 1,134 and 1,245 feet thick (fig. 5). These data suggest that the south side of the Gumbo Ridge fault was uplifted either during deposition of the Emerson, or after deposition of the Emerson and before that of the Red River Dolomite (Middle Ordovician).

Tippecanoe Sequence

The Tippecanoe Sequence (Sloss, 1963) comprises rocks that range in age from Champlainian (Middle Ordovician) through Early Devonian. In the study area the sequence contains Ordovician and Silurian rocks assigned to the Winnipeg, Red River and Interlake Formations. The Tippecanoe Sequence is separated from the Sauk Sequence below and the Kaskaskia Sequence above by regional unconformities.

Winnipeg Sandstone

The Winnipeg Sandstone is a discontinuous unit that ranges up to about 70 feet thick in the study area (fig. 4). It is more commonly present in wells to the east, closer to the Williston basin. The Winnipeg is composed of white to light gray, fine to coarse-grained quartz sandstone with dolomitic cement. As in the Flathead Sandstone, fine grains are subangular, whereas medium and coarse grains are well-rounded, frosted and pitted. The coarsest sand is concentrated at or near the base. The Winnipeg lies unconformably upon the Emerson Formation. Age of the Winnipeg is late Chazy through middle Trentonian (Foster, 1972).

Red River Dolomite

The name Red River Dolomite, derived from Canada, is generally used in the subsurface of eastern and central Montana for the unit that is called the Bighorn Dolomite in outcrops. Foster (1972) assigned the Red River Dolomite and the overlying Stony Mountain and Stonewall Formations to the Bighorn Group in the subsurface of eastern Montana. Some wellsite geologists have identified the Stony Mountain Formation in wells in the eastern part of the report area; but I was unable to differentiate this unit from the Red River. Regional studies by Fuller (1961) and Foster (1972) indicate that the Stony Mountain and Stonewall Formations lose their identities near the eastern edge of the study area, grading laterally into dolomite identical to the Red River. The name Red River, as used

here, refers to the entire interval of light-colored dolomite between the Winnipeg Sandstone and the Interlake Formation (fig. 4). The Red River of this report may include time-equivalents of the Stony Mountain and Stonewall Formations farther east.

The Red River typically is a light-colored, fine-grained massive dolomite. Colors range from white and light gray to cream and light brown. The texture is micritic to medium fragmental, most commonly very finely to finely fragmental. Crinoid fragments, corals and pellets are present. The Red River is commonly dense, but may contain occasional pinpoint and vuggy porosity, yielding shows of oil. In general, the rock becomes coarser-grained, more crystalline and porous toward the base, and in some cases the lower portion is sandy. The Red River is an important hydrocarbon reservoir rock in the Williston basin, but to date no commercial production has been obtained from this unit in the study area.

On gamma-ray logs the Red River is characterized by nearly uniform low radioactivity, reflecting the scarcity of shale interbeds in this unit (fig. 8). The Red River contrasts markedly with the shaly Emerson Formation below. It also contrasts with the overlying Interlake Formation and Devonian rocks, which contain numerous shale partings that impart a jagged pattern to the gamma log (fig. 8). Distinguishing the Red River from overlying carbonate units on other types of geophysical logs is difficult. Where a gamma-ray log is not available, a good sample log is essential for accurate identification of the Red River.

The Red River grades into and may intertongue with the Winnipeg Sandstone. Where the Winnipeg is absent the Red River overlies the Emerson Formation unconformably.

The age of the Red River is mid-Trentonian through Richmondian in the Williston basin (Foster, 1972). The uppermost Red River, as defined here, may include post-Richmondian strata.

Thickness of the Red River in the report area varies from 30 to about 530 feet. The unit is thinner than 200 feet in the western half of the study area and thickens northeastward into the Williston basin. To the east, where the Red River is part of a conformable sequence, the increase in thickness is gradual and steady. To the west, where the upper contact is erosional, large thickness differences are observed between adjacent wells. Some of these differences, however, may be due to incorrect placement of the top of the Red River.

Interlake Formation

Originally described in Manitoba, the Interlake Formation occurs through most of the Williston basin, and was identified on the logs of about 10 wells in the northeastern corner of the study area (fig. 9). Cuttings of the Interlake are composed of white, cream and light brown, dense dolomite that is dominantly micritic. Gibbs (1972) reported that the Interlake contains alternating thin beds of finely laminated micritic dolomite and fine to medium fragmental dolomite. These features are seldom evident in cuttings or on geophysical logs. Except for its finer texture and its more frequent presence of argillaceous or silty interbeds, the Interlake lithologically is similar to the Red River. On gamma-ray logs, the Interlake shows more numerous shale breaks than the Red River; the base of the Interlake tends to be especially shaly (fig. 8).

The contact between the Red River and Interlake appears to be gradational in the study area. The Ordovician-Silurian boundary is thought to lie close to

the Red River-Interlake contact, (Fuller, 1961; Gibbs, 1972). In the study area the Interlake thickens northeastward from a feather-edge to a little over 200 feet.

Lower Kaskaskia Sequence

The Kaskaskia Sequence, as originally defined by Sloss et al. (1949) and Sloss (1963), comprises all strata of Middle Devonian through Chesterian (Late Mississippian) age. In the northern Great Plains region of the United States an important regional unconformity occurs within the Kaskaskia Sequence, dividing it into upper and lower parts. As recognized in the Williston basin by Gerhard et al. (1991), the dividing unconformity in central Montana is at or very close to the top of the Devonian System. The Lower Kaskaskia Sequence in the study area contains four Upper Devonian formations: Souris River Formation, Duperow Dolomite, Birdbear Formation, and Three Forks Formation.

Souris River Formation

The Souris River Formation was defined in Bottineau County, North Dakota, by the Williston Basin Nomenclature Committee of the American Association of Petroleum Geologists; however, no type section was then designated. A standard section was proposed by Sandberg and Hammond (1958) in the Mobil Oil Co. Birdbear No. 1 well in Dunn County, North Dakota. The Souris River Formation is widely recognized by geologists who work with subsurface data in the Williston basin and adjacent areas. The Maywood Formation, as described in outcrops in the Little Rocky Mountains by Knechtel (1959), probably is equivalent.

The Souris River Formation was identified on the logs of about 10 wells, all in the northeastern part of the report area. Sample studies indicate the unit to be composed largely of greenish-gray, bluish-gray and brownish-gray, micritic, argillaceous dolomite. Interbeds of greenish-gray dolomitic shale and mudstone are common, and increase in abundance toward the base of the unit. Some sample studies report white to cream-colored, dolomitic anhydrite. On gamma-ray logs the Souris River is characterized by maximum natural radioactivity at the base, decreasing upward in response to an upward decrease in shale content (fig. 10). Both the lithologic content and the log response of the Souris River, as found here, compare closely with those reported and illustrated by Sandberg and Hammond (1958) and Sandberg (1961).

The Souris River lies unconformably on either the Silurian Interlake Formation or the Ordovician Red River Dolomite. Age of the Souris River is Frasnian (Late Devonian), according to Sandberg and Mapel (1967). Within the study area, the Souris River thickens northward from a feather-edge to a maximum of approximately 90 feet.

Duperow Dolomite

The earliest published reference to the Duperow Dolomite was in Sandberg and Hammond (1958), who established a standard section in the Mobil Birdbear well, which also contains the standard section of the Souris River Formation. The name

Duperow is widely used in the subsurface in the Williston basin and vicinity. The Duperow is the subsurface equivalent of the Jefferson Dolomite in outcrops.

The Duperow is present in most of the study area, except south of the Cat Creek fault where the Duperow, along with the remainder of the Devonian System, is largely eroded beneath Mississippian rocks (fig. 11).

The Duperow typically consists of light to medium brown or brownish-gray, finely fragmental dolomite. Micritic dolomite and medium to coarse fragmental dolomite are present in lesser amounts. Portions of the Duperow display intergranular porosity; small oil shows have been reported. The dolomite is commonly argillaceous and contains inclusions of anhydrite. Thin interbeds or laminae of light to medium gray dolomitic shale occur throughout the unit. Limestone and thin beds of anhydrite are noted in the lower Duperow in some sample studies.

On gamma-ray logs the Duperow exhibits overall low radioactivity, punctuated by numerous thin "spikes" of high radioactivity representing shale interbeds (fig. 10). Some shale spikes can be correlated among nearby wells, but none appear to be regionally continuous. Shale spikes are the chief means of differentiating the Duperow from the Red River Dolomite on gamma-ray logs. Neutron, density and resistivity logs are of lesser value in distinguishing the Duperow from the Red River. The Duperow and Interlake Formations are closely similar on gamma logs, but in most cases the distinctive Souris River Formation intervenes between Interlake and Duperow. In cuttings, the Duperow is typically brown, whereas the Interlake and Red River are typically white, light gray, or cream. However, all three formations contain light brown, fine-grained dolomite and are difficult to distinguish on sample logs.

The Duperow apparently is conformable with the underlying Souris River. Where the latter is absent, the Duperow unconformably overlies the Interlake or Red River Formation.

Sandberg and Mapel (1967) determined the age of the Duperow to be Frasnian (Late Devonian).

North and east of the Cat Creek fault, the Duperow Dolomite is 125 to 280 feet thick. It is thickest near the northern edge of the study area. South of the Cat Creek fault the Duperow is thin or absent as a result of uplift and erosion that took place near the end of the Devonian Period.

Birdbear Formation

The Birdbear Formation was named by Sandberg and Hammond (1958) for the Mobil Oil Co. Birdbear No. 1 well in Dunn County, N.D., and the type section consists of cores from this well. Many wellsite geologists in Montana still use the older name, Nisku Formation, for this interval of rocks. However, Sandberg and Hammond (1958) and Sandberg (1961) reported that the Nisku Formation proper is confined to Canada, although it is somewhat similar in lithology and age to the Birdbear Formation.

The Birdbear was identified in most wells in the northern part of the study area (fig. 11). It is absent due to uplift and erosion south of the Cat Creek fault, and absent due to nondeposition in the southeastern portion of the area, south of T15N.

The Birdbear consists of limestone and dolomite intercalated with shale (fig. 4, 10). Carbonate rocks may be white, reddish-brown, or greenish-gray, but most are described as light gray or brown. They consist of abundant coarse

fossil fragments in a micritic to finely fragmental groundmass. Intergranular and vuggy porosity are common, but no hydrocarbon shows have been reported. Carbonate rocks are slightly to highly argillaceous. Shale in the Birdbear is gray, greenish-gray, and black, rarely red; and soft, pyritic, and dolomitic. It lacks silt, and commonly is described as "waxy".

With its many widely persistent shale interbeds, the Birdbear is readily distinguished from adjacent formations on nearly all types of geophysical logs (fig. 10). The contact of the Birdbear with the Duperow is sharp, but apparently conformable. This contact is drawn at the base of the lowest thick shale interbed. In the far northern part of the study area, the Three Forks Formation conformably overlies the Birdbear. Elsewhere, the Birdbear is unconformably overlain by relatively massive limestone and dolomite of the Mississippian Lodgepole Formation.

Sandberg and Mapel (1967) reported the age of the Birdbear as late Frasnian (Late Devonian). The Birdbear is consistently 60 to 80 feet thick in the study area, except where the upper portion has been truncated by pre-Lodgepole erosion.

Three Forks Formation

The Three Forks Formation was named by Peale (1893), for Three Forks, Montana. Sandberg (1962) redescribed the type section along the Gallatin River at Logan, Montana. The Three Forks occurs only in the northern part of the report area (fig. 11), where it is composed largely of shale, with dolomite beds in the middle portion (fig. 4, 10). Sample studies (cursory) and one core description indicate shale of the Three Forks to be mostly greenish-gray, soft and pyritic, containing laminae and lenses of light gray to brown argillaceous dolomite. The medial dolomite is white to light gray, micritic and contains streaks of shale. These rock types are similar to those found at the type locality (Sandberg, 1962). The Three Forks is easily distinguished from adjacent units in well logs as a result of its much greater content of shale (fig. 10).

The Three Forks conformably overlies the Birdbear and is unconformably overlain by the Bakken Formation or, where the Bakken is absent, by the Lodgepole Limestone. Maximum observed thickness of the Three Forks in the study area is about 98 feet. Southward thinning and disappearance of this unit is the result of pre-Lodgepole erosion. The Three Forks is of Famennian (very Late Devonian) age (Sandberg and Mapel, 1967).

Upper Kaskaskia Sequence

The Upper Kaskaskia Sequence, as recognized here, is the same as identified in the Williston basin by Gerhard et al. (1991). The included rocks range from latest Devonian(?) through late Chesterian age, and comprise the very thin Bakken Formation, the Madison Group (dominantly carbonate rock), and the Big Snowy Group (dominantly siliciclastic rock with thin carbonate beds). The boundary between the Kaskaskia Sequence and the Absaroka Sequence above is controversial in central Montana, but is placed at the top of the Big Snowy Group by most workers.

Bakken Formation/Cottonwood Canyon Member

A thin unit of dark gray to black, fissile shale occurs at the base of the Lodgepole Limestone in many wells in the study area. This shale unconformably overlies units ranging from the Three Forks Shale to the Red River Dolomite and is conformably overlain by the Lodgepole. Very high natural radioactivity characterizes this shale on gamma-ray logs. It is less strongly expressed on other types of geophysical logs, and commonly is overlooked in sample logs on account of its thinness. The shale rarely exceeds 5 feet thick and typically is 1 to 3 feet. The maximum thickness is 32 feet in the Marmik Mlekush #1 well in 12-12N-32E, near the southern border of the study area.

Dark shale at this stratigraphic position is widespread in Montana and neighboring areas. Knechtel et al. (1954) described a 1- to 1.5-foot black shale containing Kinderhookian conodonts in the Little Rocky Mountains and named it the Little Chief Canyon Member of the Lodgepole Limestone. The name Little Chief Canyon subsequently has fallen into disuse. The Cottonwood Canyon Member of the Madison Limestone in the northern Bighorn Mountains (Sandberg and Klapper, 1967) is at least a partial temporal and lithologic equivalent of the black shale in the study area. Another likely equivalent is the Bakken Formation, originally defined as a subsurface unit in the Williston basin (Nordquist, 1953). In the central Williston basin the Bakken reaches a maximum thickness of about 125 feet and apparently is conformable to the underlying Three Forks Formation (Kume, 1963). Westward the Bakken thins to a feather-edge and progressively oversteps older units unconformably (Sandberg and Mapel, 1967; Peterson, 1984; Peterson and MacCary, 1987). The lower contact of the Cottonwood Canyon also is unconformable; it locally overlies Precambrian rocks in Wyoming (Sandberg and Klapper, 1967). Age of the Bakken and Cottonwood Canyon is latest Devonian to early Kinderhookian, based on conodont faunas (Sandberg and Klapper, 1967; Sandberg and Mapel, 1967).

The Bakken/Cottonwood Canyon in the study area represents the initial marine deposits succeeding the major Late Devonian episode of uplift and erosion.

Madison Group

The Madison Group originally was named the Madison Limestone by Peale (1893) in southwestern Montana. In the subsurface of central Montana, the Madison currently is ranked as a group containing three formations: the Lodgepole Limestone (oldest), the Mission Canyon Limestone, and the Charles Formation (fig. 4). Overall, the Madison Group in the study area is composed dominantly of carbonate rock, contrasting with siliciclastic units above and below. The group contains numerous geophysical log markers that are widely traceable and useful in regional subsurface mapping (Peterson, 1984; Brown et al., 1984).

Lodgepole Limestone. The Lodgepole Limestone takes its name from Lodgepole Canyon in the Little Rocky Mountains north of the study area (Collier and Cathcart, 1922). The Lodgepole was identified in the logs of all wells that reach sufficient depth in the study area.

The Lodgepole is composed mostly of limestone, but dolomite is present, particularly in the lower portion. Carbonate rock is mostly light to medium gray or brownish-gray, less commonly dark gray. The Lodgepole is slightly darker in color, overall, than the rest of the Madison Group. Most of the Lodgepole is

micritic to finely fragmental, although intervals of medium to coarse, porous oolitic and bioclastic limestone are observed in many wells. A fairly persistent zone of vuggy and coarsely crystalline limestone occurs about 200 feet above the base of the Lodgepole. Also, brown fragmental and vuggy dolomite is commonly present in the basal 50 to 100 feet. Otherwise, the Lodgepole is dense, hard, and drills more slowly than the Charles and Mission Canyon. The formation is argillaceous and siliceous throughout, and it contains numerous closely-spaced laminae or partings of medium to dark gray calcareous shale. Chert nodules are present throughout, but are concentrated in the basal portion, where they comprise up to 50% of the rock. Quartz crystals and quartz sand were noted in the basal Lodgepole in a few sample studies.

The numerous shaly laminae impart a finely ragged appearance to gamma-ray logs of the Lodgepole (fig. 15). This log signature contrasts with that of Devonian strata, which contain a number of thick and well-defined shale interbeds; and with the log signature of the Mission Canyon, which has uniformly low radioactivity. The greater density and lower permeability of the Lodgepole, as compared to the Mission Canyon, is reflected by higher and less variable resistivity, neutron and density-log readings in the Lodgepole. These log characteristics are consistent throughout the study area and enable the boundaries of the Lodgepole to be picked with confidence on all but a few of the oldest electrical logs.

The contact of the Lodgepole to the underlying Bakken Formation (Cottonwood Canyon) apparently is conformable. In wells where the Bakken is absent or too thin to be recognized, the Lodgepole unconformably overlies strata ranging from Ordovician through Devonian.

Study of conodonts, foraminifera and megafossils indicate that the Lodgepole ranges from early Kinderhookian through middle Osagian (Lower and Middle Mississippian) in central and eastern Montana (Sando, 1976). Thickness of the Lodgepole within the area of study varies from 560 to 690 feet (fig. 16).

Mission Canyon Limestone. The Mission Canyon Limestone was named by Collier and Cathcart (1922) for exposures at Mission Canyon in the Little Rocky Mountains north of the study area. The Mission Canyon in outcrops is a massive, commonly vuggy or cavernous limestone, in contrast to the thin-bedded and shaly Lodgepole Limestone below (Knechtel, 1959; Lindsey, 1980 and 1982; Goddard, 1988).

The Mission Canyon is present in the subsurface throughout the study area. Lithologically the unit is largely limestone, but considerable dolomite occurs in the lower 1/3 and also near the top (fig. 4). The typical color is light to medium brownish-gray. Textures vary from micritic to coarsely fragmental (bioclastic); overall, the Mission Canyon is coarser-grained than the Lodgepole. Much of the Mission Canyon is crinoidal, oolitic zones are common, and pisolites are present in the lower portion. Pinpoint and vuggy porosity occur sporadically throughout the formation. Although some cuttings are described as "chalky" or "earthy", the clay content is low and shale partings are unusual.

At the top of the Mission Canyon is a persistent zone of fragmental dolomite that displays vuggy, fracture, and intergranular porosity. Oil shows and asphaltic residues have been observed here in a number of wells. The Mission Canyon is an important oil and gas-producing unit in other areas of Montana, but to date no commercial production has been established from the Mission Canyon in central Montana.

The Mission Canyon has uniformly low natural radioactivity and a sharply variable profile on resistivity, neutron and density logs (fig. 15). The porous dolomite at the top of the Mission Canyon can be identified readily on most resistivity, neutron, and density logs. This upper porous zone is generally at the base of the lowest thick anhydrite bed of the overlying Charles Formation. Because the upper Mission Canyon is viewed as a prospective pay zone, this interval generally is sampled and logged with extra care, facilitating placement of the formational contact. However, in the northeastern part of the study area, the anhydrite beds that distinguish the Charles from the Mission Canyon are absent; porous, vuggy dolomite occurs at several positions within the upper half of the Madison Group. The Charles and Mission Canyon were not differentiated in this area.

Both the upper and lower contacts of the Mission Canyon are sharp and conformable. Sando (1976) determined the age of the Mission Canyon in central Montana to be approximately the upper half of the Osagian Epoch. The Mission Canyon ranges from 410 to 565 feet in the report area, but it is thicker than 500 feet in only a few wells. Typically it is 440 to 490 feet thick (fig. 17).

Charles Formation. The Charles Formation was proposed as a subsurface unit by Seager (1942). It takes its name from the type well, which was begun in 1940 as the Arro Oil and Refining Co. #2 Charles, and completed in 1942 as The California Co. #4 Charles. This well is located on Mosby dome of the Cat Creek anticline, in the SE $\frac{1}{4}$ NW $\frac{1}{4}$, Sec. 21, T15N, R30E. The original definition of the Charles was modified by Perry and Sloss (1943), and again by Nordquist (1953). Nordquist defined the base of the Charles as the base of the thick anhydrite section that overlies porous carbonate rock at the top of the Mission Canyon Limestone. Nordquist's definition is the one employed here. Correlations of the Charles were made directly from the log of the type well, using the formation boundaries delineated by Nordquist.

Sample studies from wells in the report area show the Charles to be composed largely of limestone and dolomite, similar to those of the Mission Canyon. Intercalated with the carbonate rocks are thick beds of anhydrite, which is white to light gray or brown, translucent, finely crystalline, and shaly. The thickest interval of anhydrite is at the base of the Charles. The basal anhydrite is over 100 feet thick in the type well and in several other wells, but becomes thin or absent toward the northeastern corner of the report area. Other widely traceable intervals of anhydrite, 5 to 20 feet thick, occur in the middle and upper portions of the Charles. In the northeastern part of the study area anhydrite is absent, and the Charles cannot be separated from the Mission Canyon. Shale interbeds are numerous and commonly are associated with anhydrite. Shales of the Charles are gray to black, greenish-gray, and occasionally red. They are typically calcareous and blocky to moderately fissile. Many of them produce widely traceable markers on gamma-ray and resistivity logs (fig. 15).

Zones of intergranular, vuggy and fracture porosity are common in the Charles, particularly near the middle of the formation. Shows of oil and asphaltic residue have been reported in numerous wells from these porous intervals.

As noted before, the base of the Charles is picked at the bottom of the lowest thick bed of anhydrite, which normally occurs just above the porous, oil-stained dolomite at the top of the Mission Canyon (fig. 4, 15). This contact is conformable.

Fossils of the Charles indicate early Meramecian age (Sando, 1976).

The thickness of the Charles Formation in the study area ranges from 0 to 700 feet. The Charles is locally thin or absent as a result of post-Mississippian erosion north of the Cat Creek and Sand Creek faults (fig. 18). Within the area shaded in fig. 18, the Jurassic Ellis Group directly overlies the Charles. South of the faults, where the Charles is fully preserved, its thickness is 500 to 700 feet, generally increasing westward.

Big Snowy Group

The Big Snowy Group takes its name from the Big Snowy Mountains and, (as currently defined) comprises three formations of Chesterian (Late Mississippian) age: Kibbey, Otter, and Heath (fig. 4). The Big Snowy Group consists of shale with numerous thin interbeds of limestone and dolomite, and lesser amounts of siltstone, sandstone, anhydrite, gypsum and coal. The Big Snowy contains numerous geophysical log markers that can be traced for long distances across the study area (fig. 19). Such markers are very helpful in detecting thickness and facies changes of various intervals, and in locating faults within the group.

Kibbey Formation. The Kibbey Formation originally was named by Weed (1899) for a schoolhouse in southeastern Cascade County, Montana. The Kibbey occurs in the subsurface south of the Cat Creek and Sand Creek faults, and also in small areas north of the faults (Plate 1).

The Kibbey is divisible into three members in the study area. These are (1) a lower shale member, (2) the Ray Limestone Member, and (3) an upper sandstone member (figs. 4, 19). A similar division of the Kibbey was described by Harris (1972) in his regional study of the Big Snowy Group in central Montana.

The (informal) lower shale member is commonly 80 to 100 feet thick and consists mainly of pink, orange and rusty-red shale and siltstone, locally intercalated with greenish-gray shale and siltstone. These rocks are blocky to slightly fissile and commonly are dolomitic. A small proportion of very fine-grained sandstone is present. Thin dolomite beds and some anhydrite near the base of the member are indicated in some sample logs.

The Ray Limestone Member (Rawson, 1967), commonly called "Kibbey limestone", is widespread in the study area, and generally is 10 to 40 feet thick. It is a dolomitic limestone, mottled in shades of gray, brown and reddish-brown. The Ray is dense and micritic, very argillaceous to silty, and commonly contains inclusions or interbeds of anhydrite. It is easily recognized on most electrical logs by its high resistivity (fig. 19).

The (informal) upper sandstone member of the Kibbey also includes siltstone, dolomitic or calcareous shale, and beds of argillaceous to silty limestone and dolomite. The sandstone is white, gray, or light red, and occurs most prominently near the top of the member. Typically it is poorly sorted, fine to coarse-grained, and composed of quartz along with smaller amounts of chert, cemented by calcite, dolomite, or anhydrite. Siltstone and shale of the upper member are similar to those of the lower member, while the carbonate beds resemble the Ray Member in lithology. The upper member is about 100 to 180 feet thick.

The Kibbey is readily recognized in sample studies, many driller's logs and most geophysical logs (fig. 19). Resistivity logs are most useful for

correlation. The base of the Kibbey is placed at the point of abrupt change from dominantly carbonate rock of the underlying Charles Formation to red shale and siltstone of the Kibbey (fig. 19).

The Charles-Kibbey contact has been a subject of dispute. Some authors, including Sloss (1952), Easton (1962), Sando (1976), and Smith and Gilmour (1979), interpreted this contact as unconformable throughout central Montana, and stated that it represents a large hiatus. A temporal gap, however, cannot be verified because no fossils useful for biostratigraphic zonation have been identified in the Kibbey. Other workers, including Harris (1972), Maughan and Roberts (1967), and Maughan (1984) reported the Charles-Kibbey contact to be gradational or possibly intertonguing in parts of central Montana. My study of well records indicates the Kibbey-Charles contact to be sharp but conformable in most of the report area. The upper Charles contains several distinctive gamma-ray log markers that can be traced throughout the study area (fig. 15). Truncation of these markers was noted in only a few small areas.

The Kibbey is not fossiliferous, but most workers have inferred a Chesterian age based on biostratigraphic evidence from units above and below. The overlying Otter Formation, which intergrades with the Kibbey, contains a Chesterian fauna (Easton, 1962). Maughan (1984) cited fossil evidence from southwestern Montana and an intertonguing relationship between the Charles and Kibbey as evidence of Meramecian age for at least part of the Kibbey.

The Kibbey is thin or absent in the northern half of the study area as a result of post-Mississippian, pre-Middle Jurassic uplift and erosion. Where fully preserved, the Kibbey ranges from about 200 to 325 feet thick (fig. 20).

Otter Formation. The "Otter Creek shales" were named by Weed (1892) for Otter Creek in southeastern Cascade County, Montana; the name later was revised to Otter Formation. The Otter is present south of the Cat Creek and Sand Creek faults in the study area, as shown by numerous well records.

The Otter in the study area can be divided into three informal members: a lower shale, a middle carbonate, and an upper shale (figs. 4, 19). Several thin carbonate beds are widely traceable on resistivity logs and serve as excellent marker beds.

The lower member of the Otter is 40 to 50 feet thick in most wells. This member consists dominantly of shale that may be light to dark gray, olive-gray, ochre, light red, or green. Bright green or "apple green" shale is diagnostic for the Otter Formation. The colors typically are streaked or mottled. Most of the shale is soft and blocky to slightly fissile, and it contains little silt or sand. A minor amount of dark gray to black, brittle or flaky shale occurs in the lower member of the Otter in the western part of the study area.

The middle member of the Otter is generally 70 to 100 feet thick and comprises mostly carbonate rock, with occasional interbeds of anhydrite and shale or mudstone. Limestone and dolomite are dominantly light-colored, typically mottled white, light gray or brown, and greenish-gray. These rocks are largely micritic and are classified as lime mudstone or wackestone by some wellsite geologists. Fossils and fossil fragments are common, particularly ostracods and forams. Oolitic and pelletal limestone, glauconite and anhydrite also are present. Harris (1972) reported that many limestones of the middle Otter are algal or stromatolitic, and others contain abundant pellets and pisolites. The upper member of the Otter varies from about 150 to more than 300 feet thick in the report area. This unit is largely shale that is gray to bright green, soft,

blocky to slightly fissile, and partly calcareous. Numerous limestone and dolomite beds 2 to 10 feet thick are found in the upper member; a number of these are widely traceable on geophysical logs (fig. 19). Near the western edge of the report area the number and thickness of carbonate beds in the upper Otter increase; limestone intervals as thick as 50 feet are indicated on some logs.

As previously reported by Easton (1962), Harris (1972) and Smith and Gilmour (1979), the Otter-Kibbey contact is gradational through an interval of about 10 to 30 feet. I placed the contact at the highest occurrence of sandstone, as indicated by sample studies and by gamma-ray or spontaneous-potential logs. The top of the Kibbey sandstone also usually produces a small "shoulder" on resistivity logs (fig. 19).

Mega-fossils and ostracods indicate a Chesterian (Late Mississippian) age for the Otter Formation (Easton, 1962).

Thickness range of the Otter in the report area is about 260 to 600 feet (fig. 21). Only small erosional remnants are preserved north of the Cat Creek and Sand Creek faults. Pre-Jurassic erosion also removed the Otter in some areas south of the Cat Creek fault near the western edge of the study area. In the southeastern part of the area, the Otter is locally truncated beneath the Tyler Formation (Mississippian-Pennsylvanian).

Heath Formation. The Heath Formation was named by Scott (1935) for the village of Heath on the north flank of the Big Snowy Mountains, at the western edge of the report area. The Heath is characterized by dark-colored shale and limestone, in contrast to the light-colored (commonly green) shale and carbonate rocks of the Otter. Shale of the Heath is largely dark gray or brown to black, although some olive-gray and greenish-gray shale also is present. Cores of the Heath reveal intervals of alternating dark and light-colored shale, particularly near the base, where the Heath grades into the Otter. Red and variegated shale and claystone has been logged in the Heath in some wells; but some of this material probably has caved from overlying formations. Dark shales of the Heath tend to be fissile, hard and splintery, and frequently are described as carbonaceous or petroliferous.

Limestone and dolomite occurs in the Heath as numerous beds mostly less than 5 feet thick (fig. 19). These rocks are predominantly dark brownish-gray to black, dense, micritic and argillaceous. Commonly they grade to calcareous shale. Thin beds of light-colored, micritic, marly limestone also occur, mainly in the lower Heath. Light limestones are ostracodal; dark limestones contain numerous brachiopods and trepostome bryozoans.

An interval of siltstone to fine-grained sandstone occurs in the upper Heath in the western part of the study area. This rock is white to dark brownish-gray and is composed mainly of subangular to subrounded quartz grains. Cores show it to be laminated. Oil shows are common and some core samples are saturated with oil; however, no commercial production has been achieved from the Heath in the study area.

Thin coal beds occur in the lower part of the Heath in many wells. Gypsum or anhydrite occurs approximately 120 feet above the base of the Heath in several wells north of the Big Snowy Mountains. This gypsum probably is the Loco Ridge Gypsum Bed of Derkey et al. (1985).

Among several widespread, distinctive marker beds in the Heath Formation, the most notable is a bed of black petroliferous shale that occurs an average of about 80 feet above the base of the Heath. Correlation of logs (figs. 22 and 23)

indicates that this is the Cox Ranch Oil Shale Bed of Derkey et al. (1985). The Cox Ranch contains near its middle a 5-to 15-foot zone of extremely low electrical resistivity, between intervals of high resistivity representing carbonate beds (fig. 19). Wellsite geologists in the Gumbo Ridge area refer to the Cox Ranch as the "U" marker. On gamma-ray logs, the Cox Ranch has very high radioactivity, generally sending the log off-scale. In cores the Cox Ranch consists of black, thinly laminated, highly petroliferous shale thinly intercalated with dark-colored limestone. The Cox Ranch appears to be continuous throughout the study area, except where it is eroded at the base of the Tyler Formation or Jurassic rocks. In agreement with previous geologists (Easton, 1962; Maughan and Roberts, 1967; Smith and Gilmour, 1979; Derkey et al., 1985), I found the Heath-Otter contact to be gradational and possibly intertonguing. Black shales typical of the Heath commonly alternate with greenish-gray shales typical of the Otter through an interval of 50 feet or more. Consistent placement of this contact based on color is not possible. I arbitrarily placed the Heath-Otter contact at the base of limestone bed "E", which occurs near the middle of the transition zone (fig. 19). This limestone bed is prominent on resistivity logs and also can be identified on many gamma ray-density and neutron logs. In those few areas where bed "E" is absent or poorly developed, its position can be closely approximated with reference to adjacent marker beds.

The age of the Heath Formation is late Chesterian (Late Mississippian) on the basis of abundant faunal evidence (Scott, 1942; Easton, 1962; Maughan and Roberts, 1967).

Thickness of the Heath in the report area varies from 0 to about 360 feet (fig. 24). The Heath is thickest in the southwestern part of the study area where it is locally conformable with the overlying Tyler Formation. The area of conformable contact extends from the north flank of the Big Snowy Mountains in T15N, R21E, eastward to T13 and 14N, R32E. North, south and east of this belt, the Heath is partially or completely eroded at the base of either the Tyler Formation or Middle Jurassic strata.

Absaroka Sequence

The Absaroka Sequence was defined by Sloss et al. (1949) and Sloss (1963) as comprising strata ranging in age from Early Pennsylvanian through Early Jurassic. In large areas of the North American craton the sub-Absaroka unconformity is very prominent, but geologists disagree on the existence of an unconformity and on placement of the Mississippian-Pennsylvanian systemic boundary in central Montana. My subsurface study suggests that the contact is conformable in some areas and unconformable in others.

Amsden Group

The Amsden Formation originally was described near Sheridan, Wyoming by Darton (1904), and was revised to a group by Maughan and Roberts (1967). As used by Maughan and Roberts, the Amsden Group contains the Tyler Formation (oldest), the Alaska Bench Limestone, and the Devil's Pocket Formation (figs. 4, 19). These rocks range in age from late Chesterian through Morrowan (Early

Pennsylvanian) and probably Atokan (Middle Pennsylvanian) in the study area. My usage differs from that of many petroleum geologists in Montana, to whom Amsden refers to the Alaska Bench and Devil's Pocket Formations only.

Tyler Formation. The Tyler Formation is the most controversial unit in the study area. Researchers differ about the definition, age, and depositional environment of the Tyler. Some workers reject the Tyler altogether and assign its strata to other formations. By whatever name, these rocks attract interest as reservoir rocks for most of the oil produced in central Montana outside of the Cat Creek field.

The name "Tyler sand" first was used informally by Freeman (1922). The name fell into disuse until Mundt (1956) formally proposed the Tyler Formation. Freeman's description of the type section is unclear. Mundt described what may be called a reference section in Sections 21, 22, and 27, T14N, R21E, Fergus County. Most geologists who work with subsurface data in central Montana have accepted the Tyler Formation. However, several stratigraphers and paleontologists who work mainly on outcrops in the Big Snowy Mountains reject the Tyler (Gardner, 1959; Easton, 1962; Williams, 1983; Horner, 1985; Horner and Lund, 1985). These workers maintain that no lithologic or temporal break exists between the Heath Formation and the strata called Tyler. They assign "lower Tyler" rocks to Heath and "upper Tyler" rocks to the Cameron Creek Formation.

I found the Tyler to be a mappable subsurface unit in most of the study area. This is especially true on the southeast, where a pronounced disconformity separates the Tyler from the Big Snowy Group. Near the Big Snowy Mountains, however, the Tyler-Heath contact becomes gradational and probably is conformable. An interval of transitional lithologies several tens of feet thick occurs between distinctly "Tyler" and "Heath" rocks in this area. The area of gradational Tyler-Heath contact encompasses most of the outcrops belt. Hence, one's view of the Tyler depends largely upon where one has studied it.

The Tyler Formation has been divided into three members. The lower Stonehouse Canyon Member was named by Maughan and Roberts (1967). The medial Bear Gulch Limestone is variously classified as a lentil, tongue, or member; and either overlies or intergrades laterally with the Stonehouse Canyon Member (Norton, 1956; Smith and Gilmour, 1979; Maughan, 1984; Horner, 1985). The upper Cameron Creek Member originally was named as a formation by Gardner (1959), and revised to a member of the Tyler by Willis (1959). Some researchers who reject the Tyler Formation designate the Cameron Creek as a separate formation, overlying the Heath.

Thickness of the Tyler Formation in the study area ranges from zero to about 475 feet. The upper part of the Tyler was truncated by pre-Zuni (Jurassic) erosion in large areas (fig. 26). North of the Cat Creek, Sand Creek and Weldon faults the Tyler was completely removed. Even in areas where the Tyler is fully preserved, its thickness varies greatly. Thickness changes of 100 to 200 feet within a mile are not unusual. The thickest Tyler is confined to narrow belts that run east-west or northeast-southwest. One of these belts is near the southeast corner of the study area, another is in the Gumbo ridge area, and a third is in the southwestern part of the mapped area (fig. 26). These belts of thick Tyler represent the filling of paleovalleys eroded into the Big Snowy Group. Outside of paleovalleys and away from areas of pre-Zuni erosion, the Tyler thickens westward overall. It is as thin as 40 feet in T13N, R37E, and thickens to 250-300 feet in the southwestern part of the map area.

Stonehouse Canyon Member.—The Stonehouse Canyon Member—is composed of interbedded dark-colored shale, siltstone, sandstone, minor carbonate rock, and local thin coal. Shale of the Stonehouse Canyon is mostly dark gray and greenish-gray or olive-gray to black, fissile and well laminated, pyritic, and more or less silty or sandy. Shale is carbonaceous, containing abundant coal fragments, plant fragments and, less commonly, gilsonite or asphaltic residue. Coaly plant remains and lack of disseminated kerogen help distinguish the Stonehouse Canyon from the Heath.

Cores of the Stonehouse Canyon reveal alternating layers of (1) black shale, having paper-thin laminae, and (2) medium gray, slightly fissile shale or mudstone. Siltstone and sandstone of the Stonehouse Canyon are white to medium gray, the coarser-grained rocks being lighter in color. Most sandstone is very fine-grained, but locally it is fine to medium-grained, particularly in the eastern part of the report area. Sandstone is quartzose and well sorted, and the grains generally are subrounded to subangular. Silica overgrowths on quartz grains may make the grains appear angular. Other constituents are mica and glauconite. Little interstitial clay is present; the sand is cemented by calcite or dolomite. In cores, thin Tyler sandstone commonly displays horizontal laminations, whereas thick sandstone is crossbedded. Thin coal beds are present in the Stonehouse Canyon Member. Limestone, a minor lithology, generally is dark gray or brown, micritic, and highly argillaceous.

In the northwestern part of the study area the lower part of the Tyler contains much red and variegated shale, similar to that in the Cameron Creek Member. Division of the Tyler into members is difficult in this area.

The Stonehouse Canyon contains a much larger proportion of sandstone, siltstone, and silty shale than does the Heath. In most of the study area the Stonehouse Canyon contains much sandstone, whereas the Heath lacks sandy beds. The Heath contains sandstone in the western part of the study area, but Heath sandstones are intercalated with black, kerogen-rich shales that contrast with the gray and variegated silty shales of the lower Tyler. Fragments of coal and plant fossils are common in the Tyler but rare in the Heath.

The lenticular bedding of the Stonehouse Canyon contrasts with the laterally persistent bedding of the Heath. Many shale and limestone beds in the Heath can be traced for miles, and some can be correlated across most of the study area. In comparison, correlation of beds within the Stonehouse Canyon often is difficult even among closely neighboring wells. Sandstones of the Stonehouse Canyon are lenticular in cross section and sinuous to gently meandering in plan view.

Recognition of the difference of bedding character is a key to mapping the Tyler-Heath contact. Truncation of "layer-cake" Heath strata beneath complexly intercalated Tyler clastics is evident throughout the southeastern quarter of the study area (fig. 22). In some wells the Heath is entirely eroded and the Tyler rests directly on the Otter Formation (fig. 24). Similar conditions occur in the western part of the study area north of R14N and also along the southern edge of the area. The disconformity at the base of the Tyler in the Sumatra area has been noted by many geologists (Krantzler, 1966; Jensen and Carlson, 1972; Maughan and Roberts, 1967; Maughan, 1984 and 1989; Stanton and Silverman, 1989).

The Tyler-Heath contact is gradational and probably conformable within the east-trending belt where the Heath is thickest in the southwestern part of the study area (fig. 24).

Correlation of logs here reveals no truncation of the laterally persistent beds of the upper Heath. A transition is seen from dark, kerogen-rich shales and limestones below to lighter gray silty shales and sandstones above. However, the lower Tyler contains more dark-colored shale here than elsewhere, and the upper Heath includes siltstone and sandstone. The contact was mapped at the top of the highest widely traceable marker on geophysical logs (fig. 22, 23). This position appears to coincide with the Tyler-Heath disconformity away from the belt of thick Heath.

Gamma-ray logs, in conjunction with resistivity logs and sample studies, are the best tools for locating the Tyler-Heath contact.

Thickness of the Stonehouse Canyon varies from 0 to more than 300 feet within the study area. Regionally it is thickest in the area where the Heath is thickest and the contact is gradational. Elsewhere, the member is thickest where paleovalleys are incised into the Big Snowy Group. Sandstones in the Stonehouse Canyon Member are best developed within the paleovalleys, but there, as elsewhere, they are lenticular.

The age of the Stonehouse Canyon Member is unresolved. Maughan and Roberts (1967, p. B20-21) reported Morrowan (Early Pennsylvanian) spores from near the top of the member, and transitional Chesterian-Morrowan spores from the lower part. However, Scott (1973), Cox (1986), and Horner and Lund (1985) have assigned the diverse fauna of the Bear Gulch Limestone to the late Chesterian Epoch. The Bear Gulch partly overlies and partly is a lateral facies of the Stonehouse Canyon. Taken together, these findings point to Late Mississippian age for the lower part of the Tyler and a possible Early Pennsylvanian age for the upper part. Conodont biostratigraphy by Davis (1983), however, indicates the Tyler to be entirely of Mississippian age.

Bear Gulch facies. The Bear Gulch Limestone Member or Lentil was named by Norton (1956) for exposures in Bear Gulch, Sec. 35, T14N, R21E, Fergus County. Subsequent workers variously classified the Bear Gulch as a member, a tongue, or a lentil in the Heath, Tyler, or Cameron Creek Formation. Williams (1983) identified three separate limestone bodies, and named them the Bear Gulch, Becket and Surenough beds.

In this report the name Bear Gulch is applied to any of several bodies of limestone that overlie or occur within the Stonehouse Canyon Member of the Tyler Formation. No attempt was made to delineate individual lenses of limestone as Williams (1983) did. The Bear Gulch is not really a member (it is inside another member), a bed (it is not a single discrete layer) or a lentil (it is several separate lentils). In this report I am referring to the Bear Gulch as a facies.

All Bear Gulch limestone is similar in lithology and in geophysical log character. It is dark gray or dark brown to nearly black, and micritic, and dense. It has a high organic (kerogen) content and is commonly described as petroliferous. The limestone is very argillaceous and contains laminae and interbeds of hard, black, fissile shale. Typically, the shale content is greatest at the base and gradually decreases upward, so that gamma-ray radioactivity decreases and electrical resistivity increases upward (fig. 19). Both types of logs have ragged profiles, reflecting the numerous shale interbeds. Several feet of black, highly organic shale commonly occurs at the base of Bear Gulch lentils; these shales have sufficiently high radioactivity to send most gamma-ray logs off-scale (fig. 19). The overall lithology of the Bear Gulch is

similar to that of portions of the Heath Formation, which is why some geologists assign this unit to the Heath.

Thicknesses of Bear Gulch limestone lentils vary from a few feet to nearly 200 feet. The thickest lenses are in the southwestern part of the report area, where the Tyler-Heath contact is conformable. The Bear Gulch conformably overlies and probably intertongues laterally with the Stonehouse Canyon Member.

The Bear Gulch facies contains a rich and diverse marine fauna. Fossil fishes from this unit are particularly noteworthy (Horner and Lund, 1985). Scott (1973) reported Namurian A (Chesterian) conodonts from the Bear Gulch; Cox (1986) described Late Chesterian palynomorphs (citing identification of Chesterian cephalopods to W.B. Saunders, personal communication). The Chesterian age assignment was accepted by Horner and Lund (1985), who worked extensively with the vertebrate and invertebrate megafauna. These ages conflict with Pennsylvanian ages reported by Maughan and Roberts (1967) for Stonehouse Canyon strata that underlie or are laterally equivalent to the Bear Gulch facies.

Cameron Creek Member. The Cameron Creek Member of the Tyler Formation (Cameron Creek Formation of some workers) is composed of variegated shale and claystone, containing thin beds of limestone and occasional lenses of sandstone. Shale and claystone are various hues of red, orange, yellow, lavender, green and gray. Most are soft, somewhat bentonitic, and lack fissility or lamination. Loose grains of fine to medium quartz sand are commonly reported in well cuttings. Limestone in the Cameron Creek is generally light gray to pink or reddish-brown, micritic to very finely crystalline, and dolomitic. Limestone beds generally are less than 10 feet thick and they occur most commonly near the top of the unit. Some limestone beds can be traced for distances of several miles, but none are regionally persistent.

Thickness of the Cameron Creek varies from 0 to more than 135 feet in the study area.

The lower contact of the Cameron Creek appears to be unconformable in most, if not all, of the study area. The unconformity is indicated by sharp truncation of Bear Gulch limestone lentils at the base of the Cameron Creek. This is seen even in the area of thickest Bear Gulch, where the Tyler-Heath contact is conformable. The sub-Cameron Creek unconformity is difficult to define in the northwestern part of the study area, where variegated shale occurs throughout the Tyler. Lateral variability and the fact that variegated shale is prone to caving in wellbores renders interpretation problematic.

A Morrowan (Early Pennsylvanian) age of the Cameron Creek Member is supported by studies of fusulinids (Easton, 1962) and brachiopods (Maughan and Roberts, 1967). Conodont study by Davis (1983), however, indicated that the Mississippian-Pennsylvanian boundary lies within the overlying Alaska Bench Limestone.

Alaska Bench Limestone and Devil's Pocket Formation

Overlying the Tyler Formation on the flanks of the Big Snowy Mountains are the Alaska Bench Limestone (older) and the Devil's Pocket Formation. The Alaska Bench and Devil's Pocket are difficult to tell apart in the subsurface; I mapped them as a single unit. Most petroleum geologists in Montana use the name Amsden for both formations.

The Alaska Bench Limestone was named by Freeman (1922), who described the unit in cursory fashion and did not designate a type section. Mundt (1956) formally defined the Alaska Bench Limestone and established a type section near the western edge of my study area. Gardner (1959b) gave the name Devil's Pocket Formation to an interval of strata overlying the Alaska Bench just beyond the southwestern corner of my study area.

The Alaska Bench was described as a unit of limestone with lesser amounts of dolomite and thin interbeds of red and gray mudstone (Mundt, 1956; Gardner, 1959b; Maughan and Roberts, 1967). According to Mundt, the limestone weathers gray, pink, and red; but fresh limestone (as seen in well cuttings) characteristically is brown. The Devil's Pocket is reported to consist of thin-bedded dolomite and limestone intercalated with shale, siltstone, and quartzose sandstone, the proportion of sandstone increasing upward. The shale interbeds are red according to Maughan and Roberts (1967) and Smith and Gilmour (1979); but Maughan (1989, and written communication, 1990) stated that they are mostly green.

Well records in the study area show a unit of limestone and dolomite with interbeds of red and varicolored shale overlying the Tyler Formation (fig. 19, 22 and 23). Carbonate rocks are white, pink, red, and light brown, and micritic to finely fragmental with scattered large bioclasts. Solution cavities, fracture porosity and intergranular porosity are visible in cores. The upper part of the unit, directly beneath the sub-Jurassic unconformity, underwent extensive dolomitization and dissolution. This vuggy dolomite yields oil at Mosby dome in the Cat Creek oil field and in the Rattlesnake Butte oil field. Shale interbeds are generally less than three feet thick and separate carbonate intervals 5 to 20 feet thick.

These rocks fit published descriptions of the Alaska Bench Limestone. However, Edwin Maughan (written communications, 1990 and 1991) asserts that the oil-producing rocks are largely Devil's Pocket Formation. Strata that fit published descriptions of the Devil's Pocket were logged in only about 10 wells in the report area. In these wells sandy dolomite and limestone are interbedded with sandstone in layers 5 to 20 feet thick and with thinner layers of variegated shale and claystone.

Although Mundt (1956), Maughan and Roberts (1967) and Maughan (1989) reported the lower contact of the Devil's Pocket to be unconformable, no evidence for such an unconformity was found in this study. Nor could a consistent Alaska Bench-Devil's Pocket contact be picked from available logs. A more comprehensive study is needed to clarify the relationship between these units.

The contact of the Alaska Bench/Devil's Pocket with the Tyler Formation is gradational and intertonguing. Progressing upward, limestone beds become thicker and more numerous at the expense of shale beds. I placed the contact arbitrarily at the point above which limestone becomes dominant. The Alaska Bench/Devil's Pocket is unconformably overlain by Middle Jurassic rocks throughout the study area. Thickness of the interval therefore primarily reflects pre-Zuni Sequence erosion. The Alaska Bench/Devil's Pocket is thickest in the southwestern part of the study area, commonly being over 200 feet and reaching a maximum of 303 feet. The interval also exceeds 200 feet in places along the crest of the Cat Creek anticline, and also in the triangle outlined by the Weldon fault, Cohagen fault, and eastern border of the study area. North of the Cat Creek, Sand Creek, and Weldon faults the Alaska Bench/Devil's Pocket is entirely eroded.

The age of the Alaska Bench Limestone is reported to be Morrowan (Early Pennsylvanian) on the basis of fusulinids (Gilmour, 1967; Maughan and Roberts,

1967). Fusulinids from the Devil's Pocket indicate Atokan (lower Middle Pennsylvanian) age (Henbest, 1954; Easton, 1962). Conodont biostratigraphy by Davis (1983) places the Mississippian-Pennsylvanian boundary within the Alaska Bench.

The depositional environment of the Alaska Bench was interpreted by Gilmour (1967) and Davis (1983) as cyclic (eustatically controlled?) alternation between normal shallow marine shelf and intertidal to supertidal conditions. The Devil's Pocket was interpreted as subtidal to intertidal carbonates alternating with quartz sand derived from coastal dune fields. South of the report area the Devil's Pocket grades upward into thick aeolian sandstones of the Tensleep and Quadrant Formations (Maughan, 1989).

Zuni Sequence

The Zuni Sequence of Sloss (1963) comprises Middle Jurassic through Upper Cretaceous strata, separated from the Absaroka Sequence below and the Tejas Sequence above by major regional unconformities. In the study area the Zuni Sequence is represented by Middle Jurassic through uppermost Cretaceous rocks.

Nesson Formation (?)

The Nesson Formation was defined by Nordquist (1955) in the subsurface of Williams County, North Dakota. Rayl (1956) mapped the Nesson in central Montana, including the northern portion of my study area. As described by Nordquist, the Nesson contains three members: the basal Poe Member of anhydrite, red shale and claystone; the Picard Member of gray, green and red shale; and the Kline Member of limestone and dolomite, with minor sandstone and shale.

I tentatively identified the Nesson Formation in six wells in the northeastern part of the study area (plate 2). In these wells, an interval of variable proportions of shale, siltstone, sandstone and anhydrite underlies the Tampico Member of the Piper Formation. Sample logs for most of these wells are cursory. The best sample description of Nesson (?) in the report area is in the Ralph Lowe #1 Sandquist well in Sec. 28, T16N, R36E. The Nesson (?) in this well is about 70 feet thick, and consists of about 35 feet of sandstone overlying about 35 feet of interbedded shale and anhydrite. The sandstone is white, fine to medium-grained, and partly conglomeratic, containing interbeds of dark gray silty shale. The anhydrite is white and it is interbedded with greenish-gray to dark gray shale. In other wells, 10 to 140 feet of similar lithologies occur between the Tampico and Mississippian strata. Although these rocks are not typical for the Nesson, they are even less typical for the Piper.

Rayl (1956) identified as Nesson thin deposits of dolomitic siltstone and sandstone beneath typical Piper strata in two wells south of the Cat Creek fault in the western part of my study area. I interpret these strata as part of the Piper Formation, apparently confined to small areas on the downthrown blocks of faults.

The Nesson unconformably overlies Triassic and older rocks in the Williston basin and is conformably overlain by the Piper. The age of the Nesson is inferred to be Early or early Middle Jurassic (Imlay, 1980).

Ellis Group

The Ellis Formation was named by Peale (1893) and raised in rank to a group by Cobban (1945). As currently used in central Montana, the Ellis Group comprises (in ascending order) the Piper, Rierdon and Swift Formations (fig. 4). The Nesson Formation traditionally is not considered part of the Ellis Group, although a case could be made for its inclusion, because the Nesson is lithologically similar to and conformable to the Ellis Group.

Piper Formation. The Piper Formation was named by Imlay et al. (1948) for the small village of Piper on the northern flank of the Big Snowy Mountains. The type section is at the western edge of the report area in Sec. 17, T14N, R20E. Subsequent studies show that the type section is not representative of the Piper in most of Montana; yet the Piper is generally accepted as a surface and subsurface formation.

The Piper in the subsurface is readily divided into three members: the Tampico Shale Member (oldest), the Firemoon Limestone Member, and the Bowes Shale Member (fig. 29).

Nordquist (1955) defined the Tampico Shale Member of the Piper Formation in a well in Valley County, Montana, about 50 miles north of my study area. The Tampico consists largely of greenish-gray, olive-gray, reddish-brown and maroon soft shale and claystone, with greenish-gray the dominant color. Light to medium brown, micritic, argillaceous limestone occurs in beds less than 5 feet thick in the upper part of the Tampico. Anhydrite or gypsum occurs as stringers in shale and as discrete beds, most of which are in the lower part of the member.

In a few widely scattered wells the Tampico contains a basal conglomerate or rubble zone of fragments of limestone, dolomite, and other lithologies derived from Paleozoic units, commonly intermixed with sand grains. Also locally, a dolomitic sandstone or siltstone or sandy dolomite as much as 32 feet thick occurs at the base of the Tampico. The sand is mostly fine-grained subangular quartz, but scattered larger rounded quartz grains are present. The largest area of basal sandstone is immediately south of the Cat Creek fault in T16N, R25 and 26E. Two smaller areas (representing one or two wells each) are farther west, between the Cat Creek and Crystal Peak faults (Plate 2). Rayl (1956) noted this basal sandstone in two wells and assigned it to the Nesson Formation.

The Tampico is present in most of the eastern half of the report area and in a broad strip south of the Cat Creek fault (Plate 2). In this area its thickness ranges from a feather-edge to about 125 feet (fig. 30). The Tampico was not deposited north of the Cat Creek fault, nor in a large area in the southwestern corner of the report area. At the margins of these areas the Tampico is overlapped by younger Jurassic units (Plate 2).

The lower contact of the Tampico is the pre-Jurassic unconformity, except in the few wells where the Nesson Formation (?) is present. Picking the base of the Tampico can be difficult where it overlies the Cameron Creek Member of the Tyler, because both Tampico and Cameron Creek are composed largely of variegated shale and claystone. The age of the Tampico was reported as Bajocian to Bathonian (Middle Jurassic) by Peterson (1972), and as Bajocian by Imlay (1980).

The Firemoon Limestone Member was named by Nordquist (1955) for the Murphy Corp. #1 Firemoon well in Valley County, Montana. Wellsite geologists informally call the Firemoon the "Piper lime". The Firemoon is an excellent subsurface

marker on geophysical logs and reflection seismic profiles, and it is widely used as a structural datum.

The Firemoon characteristically is a single bed of light to medium brown or yellowish-brown limestone. It is largely micritic, but contains numerous fossil fragments, and some portions are oolitic. The Firemoon is argillaceous to sandy and commonly is logged as "earthy" or "chalky". Finely disseminated pyrite produces a speckled appearance, prominent in well cuttings. Occasional lenses or inclusions of variegated claystone and of anhydrite are present.

The Firemoon conformably overlies the Tampico Member, and overlaps the Tampico slightly in areas where the latter pinches out (plate 2 and fig. 31). The Firemoon is 40 to 50 feet thick across wide areas, reaching a maximum of about 55 feet. It thins and pinches out north of the Cat Creek fault and in the southwestern part of the area. The Firemoon is either of Bathonian (Middle Jurassic) age (Peterson, 1972), or slightly older: Bajocian to Bathonian, (Imlay, 1980).

Small quantities of oil have been produced from fractured (?) Firemoon Limestone in Mosby dome on the Cat Creek anticline and at other scattered localities in central Montana.

The Bowes Shale Member was named by Nordquist (1955) for the Bowes oil field in Blaine County, north-central Montana, where production was obtained from this unit. The Bowes is composed mainly of shale and claystone, but it also includes siltstone, thin limestone beds and anhydrite. The shale and claystone are various shades of maroon, brick-red, orange, brown and greenish-gray. These rocks are soft to firm, blocky to slightly fissile, and partly bentonitic. Limestone beds, lithologically similar to the Firemoon, are common in the lower part of the Bowes. White to pink finely crystalline anhydrite also is reported in the lower part of the member.

The contact of the Bowes to the Firemoon is sharp and conformable. The Bowes is identified by its predominant red color and by its characteristic pattern on resistivity logs (fig. 29).

The Bowes overlaps the Firemoon Member north of the Cat Creek fault and in the southwestern part of the area (plate 2, fig. 32). The Bowes in turn pinches out and is overstepped by the Rierdon Formation. Away from these areas, the Bowes is typically 60 to 90 feet thick, reaching a maximum thickness of 115 feet. The age of the Bowes Member is Middle Jurassic: either upper Bathonian (Peterson, 1972) or lower Bathonian (Imlay, 1980).

Rierdon Formation. Cobban (1945) named the Rierdon Formation and described its type section in western Teton County, Montana. The name Rierdon is used by most geologists in central Montana; equivalent beds in Wyoming are assigned to the lower part of the Sundance Formation (Imlay, 1945 and 1954).

The Rierdon is a unit of gray to greenish-gray calcareous shale, marl and argillaceous limestone. In much of the study area it contains a basal interval about 25 to 40 feet thick consisting of two or more thin limestone beds separated by shale (fig. 29). The basal limestone beds are white, greenish-gray or brownish-gray, dense, micritic and fossiliferous. Ostracods, pelecypod fragments and oolites are common. The lowermost bed is very sandy and grades to calcareous sandstone. Shale of the basal Rierdon is medium gray to greenish-gray, sandy and strongly calcareous. The remainder of the Rierdon is composed of interbedded shale, marl and limestone; the proportion of limestone increases upward (fig. 29). Shale is dominantly light to medium gray and greenish-gray, but

reddish-brown shale is present locally in the lower portion. The shale is typically blocky to slightly fissile, silty, glauconitic and strongly calcareous. Limestone in the middle and upper Rierdon is light to medium gray and brownish-gray, micritic, highly argillaceous, and "chalky" or "earthy". Glauconite grains and fossil fragments, including pelecypods and foraminifera, are mentioned in many sample studies.

The Rierdon Formation produces a distinctive pattern of resistivity logs in most of the report area. Two or more sharp "spikes" represent the basal limestone-shale member. Above the basal member resistivity gradually increases upward in ragged fashion, reflecting the upward increase of limestone and decrease of shale. At the top the resistivity decreases abruptly, producing the so-called "Rierdon shoulder" (fig. 29).

I picked the Rierdon-Piper (Bowes) contact at the bottom of the basal member, as discussed above. Most wellsite geologists place the contact here; but some assign part or all of the basal member to the Piper. The lithology of the basal member conforms much more closely to published descriptions of the Rierdon than the Piper. However, local red shales in the lower Rierdon suggest that the Rierdon and Bowes intertongue in places.

The Rierdon is present throughout the study area, except in a small area to the southwest, where the Rierdon pinches out and is overstepped by the Swift Formation (plate 2, fig. 33). The Rierdon is thin adjacent to this area, and also north of the Cat Creek fault. Elsewhere the Rierdon typically is 160 to 190 feet thick, being thickest in the central portion of the study area and thinning slightly in the east-central portion.

Peterson (1972) reported the age of the Rierdon as Callovian (late Middle Jurassic), whereas Imlay (1980) indicated that the Rierdon is late Bathonian to Callovian.

Swift Formation. The Swift Formation was named by Cobban (1945); its type locality is in western Pondera County, Montana. The name Swift is in current usage in most of Montana, except the south-central portion where equivalent rocks are assigned to the upper part of the Sundance Formation (Balster, 1980).

The Swift Formation is composed of interbedded sandstone, siltstone and shale. In most of the report area, the Swift coarsens upward from largely shale at the base to sandstone at the top (figs. 4, 29). Sandstone of the Swift is white to light gray and greenish-gray, and very fine- to medium-grained. Quartz is the major component, but glauconite is conspicuous in samples; its abundance helps to characterize the Swift. Cores show the sandstone to be crossbedded and to contain numerous laminae of dark gray silty shale. Generally the cement is calcite. Shales and siltstones of the Swift are typically light to dark gray and greenish-gray; less common are red, purple and ochre shale. Most shale is silty and slightly fissile, and noncalcareous to slightly calcareous. Bentonitic shale and calcareous shale are more common in the lower part of the unit. Minor amounts of argillaceous limestone and marl occur in the lower 40 to 50 feet.

An upward-coarsening profile is evident on most gamma-ray, spontaneous-potential and electrical resistivity logs (fig. 29). The Swift-Rierdon contact generally is conformable and gradational through an interval of several tens of feet. Noncalcareous gray siltstones and shales of the middle Swift grade downward to greenish, calcareous shales or marls and occasional thin shaly limestones typical of the Rierdon. The transition point (as shown by sample

studies) commonly is 20 to 50 feet above the "Rierdon shoulder". However, I have followed the custom of most wellsite geologists and picked the Swift-Rierdon contact at the top of the "shoulder". Peterson (1957b) argued that the contact should be placed above the "shoulder"; but the horizon that Peterson recommended as the contact cannot be identified on most logs in the report area.

Cobban (1945), Peterson (1957a and b and 1972) and Imlay (1954 and 1980) reported an unconformity at the base of the Swift. I found evidence of a pre-Swift unconformity only near the western edge of the area where the Piper and Rierdon are thin. The lower Swift is largely sandstone in this area, and the Rierdon is partially or completely truncated beneath.

The Swift is present in the subsurface throughout the study area and crops out near the southwestern corner. The formation is thinnest -locally less than 100 feet - in the western part of the study area, where it directly overlies pre-Jurassic strata (plate 2, fig. 34). The Swift thickens gradually and steadily toward the northeastern corner of the report area.

The age of the Swift is Oxfordian (Late Jurassic; Peterson, 1972; Imlay, 1980).

Sandstone at or near the top of the Swift is the principal oil-producing interval in the East and Mosby domes of the Cat Creek oil field. Minor production has been obtained from the Swift on West dome and in scattered wells near the southern edge of the study area.

Morrison Formation

The Morrison Formation was named for Morrison, Colorado (near Denver), and extends through a huge area of the western United States. In central Montana the Morrison is composed largely of varicolored claystone and shale containing lenticular bodies of siltstone and sandstone (fig. 4, 36). Also present are thin beds of limestone and, near the top, coal. Sandstone tends to be most abundant near the middle of the Morrison. The informal drillers' term "Brindley sand" is applied to oil-producing sandstone of the middle Morrison in the Cat Creek oil field.

Shale and claystone of the Morrison are mottled and variegated in shades of gray, greenish-gray, brown and red. Grays and greens generally predominate, in contrast to the dominant reds of the overlying Kootenai Formation. Morrison mudstones generally are blocky to slightly fissile, soft and more or less bentonitic. They are commonly silty and sandy, and grade laterally and vertically to siltstone and sandstone.

Siltstone in the Morrison is mostly gray and greenish-gray, argillaceous, massive or slightly laminated, and partially calcareous.

Sandstone of the Morrison is light gray, having a "salt and pepper" appearance. Grain size is very fine to coarse. Sand is composed mainly of quartz and chert, along with lesser amounts of glauconite and pyrite. Much interstitial clay is present, and the cement is typically calcite. Sandstone occurs as lenticular bodies that commonly reach 20 to 60 feet thick. Porosity, permeability and grain size of Morrison sandstones normally decrease upward from a sharp, erosional basal contact. This upward-fining profile contrasts with the upward-coarsening profile of the Swift Formation.

Limestone beds in the Morrison are lenticular and rarely exceed 10 feet thick. Limestone is mostly white to medium brown, micritic to finely fragmental

and argillaceous. Limestone occurs throughout the Morrison, but mainly in the lower portion.

Coal and black carbonaceous shale beds up to at least 5 feet thick occur sporadically at or near the top of the Morrison throughout the report area, and are useful in locating the top of the formation. Upper Morrison coal is widespread in central and west-central Montana and formerly was mined in the Great Falls and Lewistown coal fields (Vine, 1956; Silverman and Harris, 1967; Walker, 1974).

The Morrison exhibits much lateral variability. Bed-for-bed correlations within the Morrison seldom are possible over distances of more than a few thousand feet. Even on modern geophysical logs, accurate placement of the boundaries of the Morrison can be difficult. Sample descriptions are valuable because they indicate presence of coal, color changes of shales, and compositional changes of sandstones that characterize the Morrison.

The Morrison-Swift contact generally is conformable and gradational through an interval of about 5 to 20 feet. I placed the contact at the highest occurrence of glauconitic sandstone typical of the Swift. The Morrison-Swift contact is locally disconformable, as in the Gumbo Ridge area (T12 and 13N, R33E), where light gray, very fine-grained massive sandstone more than 100 feet thick occupies channels eroded into the Swift (fig. 37).

Thickness of the Morrison ranges from about 70 feet near the northeast corner of the study area to about 320 feet near the western border. The thickest Morrison extends in a tongue east-southeast from the southern edge of the Judith Mountains toward Winnett. An isopach map of the Morrison is not presented here, because occasional difficulty was encountered in locating the Kootenai-Morrison contact.

The age of the Morrison is Late Jurassic, encompassing part or all of the Kimmeridgian, Portlandian and Tithonian Stages (Peterson, 1972; Imlay, 1980).

Oil is produced from lenticular sandstones of the middle Morrison ("Brindley sand") in the East and Mosby dome areas of the Cat Creek oil field. Production from the Morrison is largely incidental to development of larger reserves in the underlying Swift.

Kootenai Formation

The Kootenay Formation (note spelling) originally was defined in Alberta (Cairnes, 1908); the name Kootenay Group is currently applied to an interval of coal-bearing strata of Late Jurassic to probable Early Cretaceous age in Alberta and British Columbia (Gibson, 1985). In Montana the spelling changes to Kootenai and the name refers to an interval of red and variegated shale and sandstone. The Kootenai (Montana) is younger than the Kootenay (Canada). The Kootenai is equivalent to the Cloverly Formation in southern Montana and north-central Wyoming, and to the Lakota Formation in the Black Hills (Balster, 1980).

In the study area the Kootenai can be divided into three members (figs. 4, 36):

1. The basal Pryor Sandstone Member, commonly called the Third Cat Creek sand.
2. A middle sandstone and shale member that includes the Second Cat Creek sand.
3. An upper shale member that contains minor sandstone.

The total thickness of the Kootenai Formation in the report area ranges from about 175 feet near the northeastern corner to 575 feet at the western edge of the area adjacent to the Judith Mountains. A map of the combined thicknesses of the Kootenai and Morrison (fig. 38) illustrates the regional eastward thinning of this interval of non-marine strata. A belt of thick Kootenai and Morrison extends east-southeast from the Judith Mountains, south of and approximately parallel with the Cat Creek fault.

The age of the Kootenai is reported as late Aptian to early Albian (late Early Cretaceous) by Balster (1980), Ballard et al. (1983), and Anna (1986). Although a large time gap is indicated between the Morrison and Kootenai, there is no evidence of angularity between the two units.

Pryor Sandstone Member (Third Cat Creek sand). The Pryor Conglomerate Member of the Cloverly Formation was named by Hares (1917) in Carbon County, southern Montana. Roberts (1965) extended usage of the name Pryor to the sandstone at the base of the Kootenai Formation near Livingston, Montana. In both areas the Pryor is a coarse, commonly conglomeratic sandstone, which contains abundant chert grains and fills channels eroded into the Morrison Formation. The sandstone that is informally known as the Third Cat Creek sand in central Montana is similar in lithology to the Pryor Member of southern Montana and occurs at the same stratigraphic position.

Some wellsite geologists in the Cat Creek area refer to the Pryor as the Lakota Sandstone, a name derived from the Black Hills. The Lakota Sandstone and overlying Fuson Shale formerly were recognized as approximate equivalents of the Kootenai in Montana. Waage (1959) abandoned the name Fuson Shale in the Black Hills and redefined the Lakota Formation as equivalent to the entire Kootenai. Usage of the name Lakota for just the basal sandstone of the Kootenai therefore is inappropriate.

In the Cat Creek oil field the Pryor (Third Cat Creek) Sandstone is 80 to 140 feet thick. The sandstone is white, gray, or brown and very fine to coarse-grained (generally medium to coarse). It is poorly sorted; the grains are well rounded to subrounded. Clear quartz grains and gray to black chert grains are the major components. Also present are pyrite and coal fragments (derived from the Morrison?), especially near the base. The sandstone is commonly massive, but may contain one or more shale breaks.

The Pryor Sandstone is very porous and permeable and contains large quantities of fresh water. This unit is the source of water for many ranch and town wells and for injection wells in the Cat Creek field. Water production as high as 10,000 barrels per day has been reported.

Outside of the Cat Creek oil field the Pryor Sandstone exhibits similar lithology and is variable in thickness. Comparison of logs of closely-spaced wells suggests that the Pryor intertongues with overlying shales. In some wells the Pryor is coarsest at the base and becomes finer upward, but in other wells the vertical variation in grain size is inconsistent. A "blocky" pattern commonly is shown on gamma-ray logs, differing from the fining-upward pattern shown by most Morrison sandstones. The Pryor becomes shalier and finer-grained eastward across the study area.

The lower contact of the Pryor Member normally is sharp and probably is disconformable, as reported by other workers (Vine, 1956; Silverman and Harris, 1972; Walker, 1974). The contact is easy to recognize where subjacent Morrison

strata include coal, carbonaceous shale and greenish-gray mudstone. Where the Pryor Member directly overlies sandstone of the Morrison, identification of the contact can be problematic. The Pryor tends to be coarser-grained than sandstones of the Morrison, but this criterion is not always reliable.

Middle sandstone and shale member. The middle member of the Kootenai Formation is composed of interbedded and intertonguing sandstone and shale, along with lesser amounts of other rock types. Sandstones in this interval are generally similar to the Pryor Sandstone, except that they are finer-grained and thinner. Large amounts of chert grains, poor sorting, and plenty of interstitial clay are usual. Sandstone bodies range from a few feet to over 100 feet thick and are interbedded with shales. As many as 4 or 5 sandstones may be encountered in a single well. Abrupt lateral variations imply much lateral intertonguing of shale and sandstone.

The name Second Cat Creek sand commonly is applied to the highest well-developed sandstone encountered in the middle member. The Second Cat Creek probably comprises a number of separate sandstone bodies, all at more or less the same stratigraphic position.

Shale or claystone of the middle member typically is brick-red to reddish-brown, mottled and variegated with gray, green and ochre. Most of it is soft, blocky to slightly fissile, and variably silty or sandy. All gradations between claystone and siltstone are present. Other lithologies of the middle member of the Kootenai include white to light green bentonite, and light yellow or brown, micritic, argillaceous to silty limestone and dolomite. These rock types commonly occur just above the Pryor Member in the Cat Creek oil field and vicinity.

Both the upper and lower contacts of the middle member are interpreted as gradational and intertonguing. Thickness of the member is highly variable.

The Second Cat Creek sand, along with the First Cat Creek sand (Fall River Sandstone), is the principal oil-producing unit in the West dome of the Cat Creek oil field. Minor production has been obtained from sandstones of the middle Kootenai Formation in Mosby and East domes and in the Rattlesnake Butte oil field.

Upper shale member. The upper shale member of the Kootenai, sometimes called Fuson Shale by wellsite geologists, is generally 70 to 150 feet thick in the report area. This member is largely reddish-brown and maroon blocky shale and claystone, variegated or mottled with other colors. Lenses or thin discontinuous beds of gray siltstone and sandstone occur sporadically in the upper member. These are similar in lithology to siltstones and sandstones of the middle member. White to light gray bentonitic claystone and lenses of light gray micritic limestone are common at or near the base of the member.

Colorado Group

The name Colorado Shale or Colorado Group long has been applied in Montana to the thick succession of marine Cretaceous rocks, dominantly dark gray shale, between the Kootenai Formation and the Telegraph Creek Formation. Early mappers

such as Lupton and Lee (1921) and Reeves (1927) treated the 1,900-foot-thick Colorado Shale as a single formation, although they noted a number of members and beds useful for mapping. Cobban (1951 and 1953) and Reeside and Cobban (1960) correlated most of the formations of the Black Hills standard section into the Cat Creek area. Johnson and Smith (1964), in mapping the Cat Creek area, adopted the Black Hills terminology and classified the units as members within the Colorado Shale (formation). In this report I am following Porter (1990) and classifying the Colorado as a group, divided into formations.

The Colorado Group contains many marker beds that can be traced through most or all of the report area on geophysical logs, particularly on resistivity logs (fig. 39). These markers include bentonite beds, which are time markers. Other markers are peaks of high resistivity that represent thin sandstone, siltstone and concretionary layers. These markers greatly facilitate mapping of the Colorado Group and recognizing faults therein.

Fall River Sandstone (First Cat Creek sand). The Fall River Sandstone was named by Russell (1927), and the type section near Fall River, South Dakota was described by Russell (1928). Cobban (1951) correlated the Fall River Sandstone with the First Cat Creek sand in the Cat Creek oil field. Johnson and Smith (1964) referred to this unit as the basal sandstone member of the Colorado Shale. The term "Dakota Sandstone" is widely used by wellsite geologists, but is inaccurate, because the type Dakota Sandstone of northeastern Nebraska is substantially younger than the Fall River Sandstone (Russell, 1928). The Fall River is equivalent to the Flood Member of the Blackleaf Formation in the Sweetgrass arch area of western Montana (Cobban et al., 1959 and 1976).

The Fall River consists of white to light gray, very fine to medium-grained quartzose sandstone. It is generally cleaner and better sorted than sandstones of the Kootenai Formation. Chert grains are present, but are less abundant than in the Kootenai. Cores reveal numerous laminae and thin interbeds of greenish-gray to dark gray silty, fissile shale and siltstone.

In the Cat Creek oil field and elsewhere, the Fall River Sandstone can be divided into two units of roughly equal thickness. The lower unit consists of porous, permeable sandstone that serves as the reservoir for petroleum. Grain size, porosity and permeability typically increase upward, so that the main pay zone is near the top of the unit, about 50 feet below the top of the Fall River. The name First Cat Creek sand commonly is applied only to the lower unit of the Fall River. The upper unit of the Fall River is very fine sandstone and siltstone containing numerous shale interbeds and having little or no permeability.

The Fall River can be mapped reliably only from geophysical logs (fig. 39). Picking the top of the Fall River on the basis of well cuttings is difficult because lithologically similar siltstones and fine sandstones occur in the lower part of the overlying Thermopolis Shale. A structure map such as that of Reeves (1927), contoured on top of the First Cat Creek sand as identified by drillers, is not reliable.

The Fall River-Kootenai contact is disconformable, but generally concordant. In only a few wells is scouring evident at this contact; maximum erosional relief is approximately 80 feet. The Fall River-Kootenai contact is easy to pick on drillers' logs and sample logs where the upper Kootenai consists of red shale, as is commonly the case. The contact is difficult to pick where the uppermost Kootenai is sandstone. In such cases a change from clean, well-

sorted Fall River sandstone to clay-rich, poorly sorted, chert-rich Kootenai sandstone sometimes can be recognized. The Fall River averages about 90 feet thick across much of the report area, thinning to about 55 feet near the western edge (fig 40). Locally the Fall River thickens where it fills channels eroded into the Kootenai Formation. The maximum observed thickness is 163 feet.

The Fall River in central Montana is assigned a mid-Aptian age (upper part of Lower Cretaceous) by Ballard et al. (1983) and Anna (1986 a and b).

Thermopolis Shale. The Thermopolis Shale was named for Thermopolis, Wyoming by Lupton (1916). In Montana, the Thermopolis was mapped by Collier and Cathcart (1922) and Knechtel (1959) in the Little Rockies, Knechtel and Patterson (1956) in the Hardin district, and Roberts (1965) near Livingston. The Thermopolis is a thick unit of dark gray shale with thin interbeds of siltstone, sandstone, and bentonite. In the report area the Thermopolis is 550 to 650 feet thick, and ranges in age from middle to late Albian (Anna, 1986 a and b).

On geophysical logs the Thermopolis Shale can be divided into five members (figs. 4 and 39). These members also can be identified in many outcrops, but in general they cannot be mapped separately at the surface because of the poor quality of exposures (Porter, 1990, and personal communication).

The basal siltstone member of the Thermopolis is commonly known among petroleum geologists as the "Dakota silt". This unit is distinctive and widely traceable on geophysical logs (Rice, 1976). The basal siltstone is composed of light to medium gray quartzose siltstone, interbedded and interlaminated with lesser amounts of dark gray silty shale. Toward the east, where the unit thickens, the siltstone intergrades with very fine-grained sandstone. Small amounts of white bentonite and of ironstone (probably concretions) also are reported in well cuttings. The signature of this unit on resistivity logs is consistent throughout the report area. The upper contact is very sharp and distinct, and provides an excellent structural datum. The lower contact is also sharp and conformable.

The basal siltstone member thickens steadily from about 45 feet at the western edge of the study area to 120-135 feet in R37-39E; it thins slightly at the eastern border of the area (fig. 41).

The Skull Creek Shale Member was established by Collier (1922) near Osage, northeastern Wyoming. This unit (at formation rank) has been mapped throughout the Black Hills and Powder River basin. Cobban (1951) correlated the Skull Creek Shale to the Mosby dome area; Johnson and Smith (1964) mapped the Skull Creek as a member of the Colorado Shale. Porter (1990) recognized the Skull Creek as a member of the Thermopolis Shale in central Montana. As mapped by Porter, the Skull Creek included the basal siltstone member of this report.

In the report area the Skull Creek Member is composed dominantly of dark gray to black clay shale and silty shale, containing occasional laminae and thin interbeds of light gray siltstone. The shale is finely carbonaceous and pyritic, platy or fissile. Thin beds of white bentonite occur near the top. Calcareous shale, limestone concretions, and *Inoceramus* fragments occur in well cuttings. Dark reddish-purple ironstone concretions are conspicuous in outcrop (Porter, 1990). The Skull Creek has uniformly low resistivity on electric logs, contrasting with high resistivity of the enclosing units (fig. 39).

The Skull Creek Member is thinnest, about 65 feet, along the south-central edge of the report area (fig. 42). It thickens to about 120 feet at the northeast corner of the area and 140 feet near the northwest corner. Toward the

northwest, the upper part of the Skull Creek intertongues with the lower part of the Cyprian Sandstone Member.

The Cyprian Sandstone Member of the Thermopolis Shale was named by Knechtel (1959) during a study of the Little Rocky Mountains, about 25 miles north of my study area. The Cyprian Member in the Cat Creek area correlates directly to a reference geophysical log published by Knechtel. Petroleum geologists use the name Muddy Sandstone or, occasionally, Newcastle Sandstone for this sandstone. Cyprian is used here instead of Muddy or Newcastle because correlation of the Cyprian with the Muddy and Newcastle has not been established.

The Cyprian consists of light to medium gray siltstone and very fine-grained sandstone, which are glauconitic, micaceous, partly calcareous, and interbedded with dark gray silty shale. Johnson and Smith (1964) described subrounded to well-rounded dark-colored quartzite pebbles at the top of a sandstone that probably is the Cyprian in Secs. 18 and 19, T13N, R25E. Knechtel (1959) observed well-rounded chalcedony pebbles up to 2 inches in diameter in the upper part of the Cyprian in its type area. Porter and others (1993) also reported that chert pebbles are widely present at the top of the Cyprian in central Montana. Resistivity logs in the study area indicate that the Cyprian coarsens upward; the lower contact being gradational to the Skull Creek Member, and the upper contact being sharp (fig. 39).

The Cyprian is absent or poorly developed in the eastern 1/3 of the study area, and is about 15 to 25 feet thick in the central 1/3 (fig. 43). Near the western edge of the area the Cyprian thickens rapidly to as much as 120 feet and intertongues in complex fashion with the upper part of the Skull Creek Member.

The unnamed sandy member of the Thermopolis Shale is the thick middle unit of the formation, as designated by Johnson and Smith (1964). This member is composed of dark gray to black shale that is noticeably siltier and sandier than the Skull Creek and Shell Creek Members. Laminae and thin interbeds of siltstone to fine sandstone are numerous, and many are widely traceable on geophysical logs. The thickest siltstone/sandstone intervals are 5 to 15 feet thick, and in many cases they coarsen upward like the Cyprian Member. Sample studies indicate that sandstones have a "salt-and-pepper" composition of quartz, chert and glauconite grains, in common with much of the Colorado Group. The sandy member also contains several widely traceable bentonite beds (fig. 39).

The sandy member is 200 to 250 feet thick in the eastern part of the report area and gradually thickens westward to 350-380 feet (fig. 44). The contact to the Cyprian Sandstone Member, below, is sharp and is considered to represent a regional lowstand surface and sequence boundary (Porter and others, 1993).

The Shell Creek Shale Member was named (as a formation) by Paull (1957) and further described by Eicher (1960) in the Bighorn basin, Wyoming. Following Porter (1990), the Shell Creek is classified as a member of the Thermopolis Shale herein. Lithology and stratigraphic position of the Shell Creek Member are essentially the same as those of the type Shell Creek Shale.

Lithology of the Shell Creek is similar to that of the Skull Creek: dark gray, greenish-gray, and black shale, pyritic and glauconitic, and moderately to highly fissile. Little siltstone is present. Brown to black ironstone concretions have been observed in outcrops and occasionally in well cuttings. Thin but widespread bentonite beds occur at the base and near the middle of the unit. Resistivity of the Shell Creek is low and uniform (fig. 39). The lower contact is conformable and is mapped at the base of the lower bentonite bed.

The Shell Creek Member ranges from 80 to 135 feet thick in the report area, except in the southwestern corner, where the upper part of the Shell Creek

intertongues with the overlying Mowry Shale (fig. 45). The Shell Creek is 45 to 80 feet thick in the area of intertonguing.

Mowry Shale. The Mowry Shale was named by Darton (1904) near Buffalo, Wyoming. It is a widespread and distinctive unit that extends across large areas of Colorado, Utah, Wyoming, and Montana. In the Cat Creek area the Mowry was classified as a member of the Colorado Shale by Reeves (1927), Cobban (1951 and 1953), and Johnson and Smith (1964), and as a formation in the Colorado Group by Porter, (1990). The Mowry is equivalent to the Bootlegger Member of the Blackleaf Formation in the Sweetgrass arch area of western Montana (Cobban et al., 1959 and 1976).

In outcrops of central Montana, as elsewhere, the Mowry is composed of interbedded hard, siliceous shale and fine-grained thinly bedded sandstone. The shale is dark gray when fresh, and weathers light silvery-gray. Fish scales are abundant in both shales and sandstones of the Mowry. Sandstones are very fine to fine-grained and exhibit horizontal and ripple laminations. Bentonite beds occur at the top, base, and at several positions within the Mowry (Porter, 1990).

The Mowry is not particularly distinctive in well cuttings, although some wellsite geologists observe its abrasive, siliceous character and abundant fish remains. On electric logs the Mowry is readily identified as an interval of high resistivity, punctuated by many low-resistivity spikes that represent the bentonite beds (fig. 39). Bentonite beds are 1 to 3 feet thick and are spaced 15 to 25 feet apart throughout the Mowry. The Mowry thickens from a little over 40 feet at the northeast corner to as much as 180 feet in the southwestern part of the report area (fig. 46). The unit becomes sandier southwestward, and thickens at the expense of the Shell Creek in the southwestern part of the area. In effect, the upper part of the Shell Creek becomes sandy toward the southwest and intergrades with the Mowry. Bentonite beds can be traced directly from the shaly facies (Shell Creek) into the sandy facies (Mowry), documenting the intertonguing relationship.

A late Albian age was assigned to the Mowry by Ballard et al. (1983) and Anna (1986), who placed the top of the Lower Cretaceous Series at the top of the Mowry. Cobban and Kennedy (1989) reassigned the Mowry to the early Cenomanian Stage of the Upper Cretaceous Series on the basis on ammonites.

Belle Fourche Shale. The type locality of the Belle Fourche Shale is near the Belle Fourche River on the western flank of the Black Hills in Crook County, Wyoming (Collier, 1920). The Belle Fourche was traced to the Mosby dome area by Cobban (1951 and 1953), and mapped as a member of the Colorado Shale by Johnson and Smith (1964). Porter (1990) mapped the Belle Fourche as a formation, and her classification is followed here.

The Belle Fourche is the oldest bedrock unit that crops out in the Cat Creek oil field. The upper part of the Belle Fourche is exposed along the Musselshell River in the Mosby dome area (Nelson, 1993).

Lithologically, the Belle Fourche is similar to the sandy member of the Thermopolis Shale. Dark gray silty to sandy shale, with occasional siltstone and sandstone interbeds, predominates. Several bands of light gray to orange-weathering septarian limestone concretions are prominent in outcrops. Two thick bentonite beds in the lower to middle portion of the Belle Fourche (fig. 39) can be traced throughout the study area on geophysical logs. Regional cross-sections by Rice (1976) indicate probable correlation of the two bentonites with the Soap

Creek and "L" bentonites, which Knechtel and Patterson (1956) identified in the Hardin district, southern Montana.

The Mosby Sandstone Member, at the top of the Belle Fourche, was named for Mosby Post Office by Lupton and Lee (1921) and assigned to the Belle Fourche by Rice (1984). The Mosby is a widespread and useful marker both in surface mapping and on geophysical logs. Near Mosby Dome the member contains two prominent ledge-forming sandstones separated by a shaly interval. The sandstone ledges are both 1 to 3 feet thick and the medial shale is 10 to 12 feet thick. An interval of 6 to 8 feet of dark gray sandy shale, overlying the upper sandstone ledge, also is considered part of the Mosby. Away from Mosby Dome, only the upper sandstone bed is persistent. Sandstone of the Mosby is light gray, very fine to fine-grained, partly calcareous, and displays hummocky crossbedding in outcrops.

An unnamed sandy interval occurs 30 or 40 feet below the Mosby Member both in outcrops and subsurface.

The Belle Fourche Shale thickens steadily eastward across the study area from less than 180 feet to a little over 500 feet (fig. 32). Ballard et al. (1983) and Anna (1986) both assigned the Belle Fourche Shale in central Montana to the Cenomanian Stage of the Late Cretaceous.

Greenhorn Formation. The Greenhorn Formation was defined by Gilbert (1896) near Pueblo, Colorado, and subsequently has been traced through large areas of Kansas, South Dakota, Wyoming and Montana. It is an interval of calcareous shale, marl and limestone that contrasts with noncalcareous shales above and below. The Greenhorn was first identified in the Cat Creek area by Cobban (1951), and later mapped by Johnson and Smith (1964) and Porter (1990). The Cone Member of the Marias River Formation in the Sweetgrass arch area is equivalent to the Greenhorn Formation in lithology and age (Cobban et al., 1959 and 1976).

In outcrops the Greenhorn is a strongly calcareous shale that is dark gray when fresh and weathers to soft, light gray soil. Lenses of dark gray micritic limestone occur at the base, and lenses of laminated calcareous siltstone or sandstone are found near the top. A persistent bentonite bed occurs a few feet above the base, and several others directly overlie the Greenhorn. In sample studies the Greenhorn is logged as calcareous shale that contains limestone and fossil fragments, bentonite, and siltstone. On electric logs the Greenhorn appears as an interval of moderate resistivity between low-resistivity spikes that represent bentonite beds (fig. 39).

The Greenhorn thickens from less than 10 feet near the eastern edge of the report area to 55 feet at the western edge (fig. 48). At Mosby Dome, where it crops out, the Greenhorn is 20 to 25 feet thick.

Age of the Greenhorn in central Montana is early Turonian. The Greenhorn is much thinner and represents a shorter span of time here than in the Black Hills (Anna, 1986).

Carlile Shale. The Carlile Shale was named by Gilbert (1896) near Pueblo, Colorado and, like the Greenhorn, extends through much of the Great Plains and eastern Rocky Mountain region. Cobban (1951 and 1953) identified equivalents of the Carlile at Mosby Dome; Johnson and Smith (1964) mapped the Carlile as a member of the Colorado Shale. Porter (1990) raised the Carlile in rank to a formation. The Carlile is largely, if not entirely, equivalent to the Ferdig Member of the Marias River Formation on the Sweetgrass arch (Cobban et al., 1959 and 1976).

The Carlile Shale crops out in an ovoid belt that surrounds Mosby and East Domes, and also along the crest of West Dome near Cat Creek village (Nelson, 1993).

Johnson and Smith (1964) described five distinct lithologic units in the Carlile; Porter (1990) and I regard these units as members. The first (basal) unit consists of 7 to 9 feet of dark gray, soft, noncalcareous clay shale that weathers orange-brown and contains several bentonite beds less than 1 foot thick. The second unit consists of 30 to 35 feet of very dark gray, fissile shale that becomes sandy near the top, and characteristically erodes to rolling surfaces that support dense stands of junipers but no grass. The third unit consists of 18 to 20 feet of dark gray, moderately fissile, silty shale that contains abundant dark red ironstone nodules arranged in distinct layers. This unit erodes nearly bare, sparsely vegetated surfaces. The fourth unit consists of roughly 100 feet of dark gray shale that coarsens upward from clay shale at the base to silty shale and siltstone near the top. Large septarian limestone concretions that weather yellowish-gray to orange are numerous. The fifth (upper) unit consists of about 150 feet of shale similar to the fourth unit, but its concretions weather light gray rather than yellow. The upper unit coarsens upward to a point 30 to 50 feet below the top, and then fines upward to the top of the Carlile.

Porter (1990) identified Unit 1 as the Fairport Member, Units 2 and 3 together as the Blue Hill Member, Unit 4 unnamed, and Unit 5 as the Sage Breaks Member. The Fairport and Blue Hill Members were named in Kansas and the Sage Breaks Member was named in the Black Hills. Units 1, 2 and 3, together, closely resemble the type section of the Pool Creek Member of the Carlile Shale, as described by Knechtel and Patterson (1962) in an exposure near Belle Fourche, South Dakota. Unit 4 is at the position of the Turner Sandstone Member in the Black Hills, but lacks the sandstone content typical of the Turner. My preferred classification is: Units 1, 2, and 3 to the Pool Creek Member, Units 4 and 5 to the Sage Breaks Member.

On electric logs the Pool Creek Member appears as a 35-to-60-foot interval of upward-increasing resistivity (fig. 39). Resistivity of the Sage Breaks increases irregularly upward to within a few tens of feet of the top, then decreases. The top of the Carlile forms a distinctive resistivity "shoulder".

The Carlile Shale thickens gradually and steadily from about 200 feet at the western edge of the study area to 395 feet at the eastern edge (fig. 49). Near the northern edge of the area, the upper part of the Carlile contains sandstone lenses that may be equivalent to the Bowdoin sand (informal) of northern Montana (Anna, 1986).

Anna (1986) indicated the Carlile Shale in central Montana to be of late Turonian age. Ballard et al. (1983) reported the age of the Carlile as late Turonian and early Coniacian.

Niobrara Shale. The Niobrara Formation was named in northeastern Nebraska by Meek and Hayden (1862). It is a unit of calcareous shale, marl and chalky limestone that covers large areas of Colorado, Kansas, Nebraska, South Dakota, Wyoming, and Montana. The Niobrara was first recognized in the Cat Creek area by Cobban (1951 and 1953) and it was mapped as the Niobrara Member of the Colorado Shale by Johnson and Smith (1964). Porter (1990) mapped the Niobrara as a formation in the Colorado Group.

As exposed in the Cat Creek area, the Niobrara differs considerably in lithology from the type Niobrara, and is more similar in lithology to the Kevin Shale Member of the Marias River Formation as described by Cobban et al. (1959 and 1976) in western Montana. A good case could be made for using the name Kevin instead of Niobrara here, but I am deferring to priority and calling the unit Niobrara Shale.

The Niobrara crops out in an ovoid belt that surrounds the entire Cat Creek oil field (Nelson, 1993). The outcrop belt is broad on the gently dipping southwest flank of the Cat Creek anticline, and narrow on the steeply-dipping northeast flank.

In the Cat Creek area the Niobrara contains a lower noncalcareous unit about 200 feet thick. This unit is dominantly composed of dark gray and olive-gray, soft, slightly fissile clay shale and silty shale. It is much less silty than the Carlile and contains smaller and fewer septarian concretions. Thin beds of bentonite are common in the lower part.

A persistent zone of sandy limestone concretions at the base of the Niobrara was informally named the Sage Hen bed by Lupton and Lee (1921). Subsequent mappers, including Reeves (1927), Johnson and Smith (1964), Porter (1990) and Nelson (1993) all recognized the Sage Hen as a persistent and useful marker in outcrops. This bed has not been identified in the subsurface, but is believed to lie within the interval of bentonitic shale near the base of the Niobrara.

The upper calcareous unit of the Niobrara comprises 150 to 200 feet of soft, olive-gray shale that produces a distinctive pale orange soil. This shale is strongly calcareous. Tiny light gray calcareous specks, visible in hand lens, give rise to the informal name "white speck zone". Gypsum crystals, masses of calcite crystals, and small septarian limestone concretions are additional features of the upper Niobrara. The upper calcareous unit of the Niobrara has slightly higher resistivity on logs than the lower noncalcareous member (fig. 39).

The contact of the Niobrara to the Carlile is gradational and conformable. This contact was mapped at the base of the Sage Hen bed in outcrop and at the top of the "Carlile shoulder" on resistivity logs.

The Niobrara is 275 to 350 feet thick in most of the study area (fig. 50). No consistent thickness trend is apparent. Extreme thickness values are 175 and 395 feet.

Gill and Cobban (1973), Ballard et al. (1983), and Anna (1986) all reported the age of the Niobrara in central Montana as Coniacian to mid-Santonian.

Montana Group

The Montana Group comprises five Upper Cretaceous formations: the Telegraph Creek Formation, Eagle Sandstone, Claggett Shale, Judith River Formation, and Bearpaw Shale. The Montana Group differs little in overall lithology from the Colorado Group; both groups contains the same types of shale and sandstone, and in nearly the same proportions. Thus, the names Montana and Colorado Groups are not particularly useful in denoting lithologic relationships, but are names mainly of historical value.

Telegraph Creek Formation.—The Telegraph Creek Formation was named by Thom (1922) for a stream in southern Yellowstone County, Montana, and subsequently has been mapped in the Cat Creek area (Johnson and Smith, 1964; Porter, 1990; Nelson, 1993). The Telegraph Creek is lithologically similar and probably equivalent in age to the Gammon Shale as described in the Black Hills area (Robinson et al., 1964). The main distinction is that the Telegraph Creek is siltier and sandier than the Gammon.

The Telegraph Creek crops out in a narrow belt, offset by numerous en echelon faults, along the steep northeast flank of the Cat Creek anticline. On the southwest flank of the anticline the dips are gentle and the outcrop belt wide (Nelson, 1993).

The Telegraph Creek is a formation of silty mudstone, siltstone and fine-grained sandstone. Mudstone and siltstone are medium to dark gray and olive-gray; sandstone is light gray. These rocks commonly occur in coarsening-upward cycles 20 to 50 feet thick. Such cycles are evident in outcrops, particularly along Dog Creek and Sage Hen Creek southeast of East Dome. The cycles also are apparent on gamma-ray and resistivity logs (fig. 39). Each cycle commences with silty shale at the base, and grades upward through siltstone to fine-grained shaly sandstone. Horizontal and ripple laminations are common. Dark red-brown, dense ironstone concretions occur in several intervals, and are easily recognized in float. One to three widespread beds of creamy-white bentonite also are present.

The lower contact of the Telegraph Creek is gradational in most of the report area. Near the east-central edge of the area, logs of several wells indicate the Telegraph Creek fills a channel eroded into the upper Niobrara. Thickness of the formations varies in seemingly irregular fashion, from about 90 to over 400 feet (fig. 51). The Telegraph Creek is of late Santonian age (Gill and Cobban, 1973; Ballard et al., 1983; Anna, 1986).

Eagle Sandstone. The Eagle Sandstone was named for Eagle Creek, about 30 miles northeast of Fort Benton, Montana (Weed, 1899). The Eagle is recognized widely in the Sweetgrass arch and in central and southern Montana. The Eagle gradually loses its identity eastward as it intergrades with the Telegraph Creek Formation.

Near the Cat Creek oil field the Eagle crops out as either a single sandstone ledge or as two sandstone ledges separated by shale that is largely soil-covered. The ledges are steeply tilted north of West Dome and nearly horizontal south of West Dome. Eastward, near Mosby and East Domes, the Eagle Sandstone intergrades with shale and siltstone of the Telegraph Creek, and is not a mappable unit (Nelson, 1993).

Lithologically the Eagle is an interval of sandstone that is interbedded with mudstone and siltstone similar to the Telegraph Creek Formation. Sandstone of the Eagle is white to light gray or brown, very fine to medium-grained, rarely coarse, generally fairly clean and well-sorted. It contains chert grains, glauconite and calcite cement. Thicker sandstone bodies display planar and ripple lamination, crossbedding, and massive bedding. Large calcareous sandstone concretions are common in outcrop. Horizontal, inclined and vertical burrows of genus *Ophiomorpha* are abundant.

Sandstone bodies of the Eagle can be traced for many miles in outcrop or by means of well logs, but regionally these sandstones all are lenticular and either wedge out or grade laterally into siltstone and mudstone. The Eagle is thickest (up to 300 feet) and contains the greatest proportion of sandstone at the

northwestern corner of the study area (fig. 51). It becomes thin and shaly in the central and eastern areas, but where present, generally is 100 to 180 feet thick. The Eagle again thickens noticeably near the east-central edge of the area.

Claggett Shale. The Claggett Shale was named by Hatcher and Stanton (1903) for Fort Claggett, which was located where the Judith River enters the Missouri. The name Claggett is in general usage in central Montana.

The Claggett is a nonresistant unit that underlies a strike valley between hogbacks of the Eagle and Judith River Formations on the northeast flank of the Cat Creek anticline (Nelson, 1993). The best exposures are in bluffs along the Musselshell River and in cut banks along Dog Creek and Sage Hen Creek, southeast of East Dome.

The Claggett consists of dark gray to olive-gray, soft and moderately fissile clay shale and silty shale. Large septarian limestone concretions are scattered throughout. Numerous bentonite beds that range from a few inches to about 2 feet thick are present in the basal 45 to 80 feet of the Claggett (fig. 39). The lowest of these is the Ardmore Bentonite Bed, a regional marker that Rice (1976) used as structural datum for subsurface cross-sections throughout the Montana plains. The base of the Ardmore Bentonite was taken as the base of the Claggett Shale in subsurface mapping, although most logs and outcrops indicate a gradational change from clay shale in the lower Claggett to sandstone or siltstone of the Eagle or Telegraph Creek.

The Claggett Shale thickens from less than 150 feet near the eastern border of the study area to more than 500 feet near the northwestern corner (fig. 52). Gill and Cobban (1973) reported the age of the Claggett to be mid-Campanian, and the radiometric age of the Ardmore Bentonite to be about 79 m.y.

Judith River Formation. The "Judith River beds" were named by Meek and Hayden (1856), and the Judith River Formation was defined more thoroughly by Stanton and Hatcher (1905). The type locality is near the mouth of the Judith River northwest of the study area. This formation has been mapped by many workers throughout central and northern Montana and southern Alberta.

Sandstone beds of the Judith River produce steeply tilted hogbacks along the northeast flank of the Cat Creek anticline (Nelson, 1993). Less resistant strata between the sandstone beds are well exposed near the Musselshell River, Long Coulee, and Dog Creek, and roadcuts along State Rt. 200.

The Judith River is composed of variable proportions of sandstone, siltstone, mudstone and shaly coal. Individual rock units are lenticular and laterally variable. Sandstone in the Judith River is light gray, very fine to medium-grained, and has a "salt and pepper" appearance due to abundant chert and other dark grains. Generally a substantial amount of interstitial clay is present. Bedding is highly variable, from irregular laminations to crossbedded and massive sandstone. Some sandstone layers are calcareous and have concretionary bedding. Calcareous sandstone concretions up to 10 feet in diameter are conspicuous in the lower Judith River on the north flank of the Cat Creek anticline.

Siltstone of the Judith River is light to medium gray, bentonitic, carbonaceous, and laminated to massive. Mudstones occur in various shades of gray, greenish-gray, olive and brown. Most mudstones are soft, bentonitic, and massive to weakly fissile. Carbonized plant fragments, stringers of lignite, and

silicified wood are common. Beds of very shaly lignite or lignitic claystone as thick as 8 feet crop out north of Mosby and West domes. Coal beds appear to have greater lateral continuity than most other rock units in the Judith River.

The Judith River grades into the Claggett Shale through an interval of 50 feet or more. The upper portion of the Judith River intertongues with the overlying Bearpaw Shale. I mapped the top of the Judith River at the point above which more than 50% of the gross interval is dark gray shale typical of the Bearpaw.

As thus defined, the Judith River ranges in thickness from 140 feet at the southeastern corner of the study area, to 640 feet in a well near the northwestern corner (fig. 53). A belt of thick Judith River extends east-southeast, parallel with and just north of the Cat Creek fault. Another lobe of thick Judith River roughly follows the northern edge of the study area.

The age of the Judith River is mid-Campanian (Anna, 1986) or late Campanian (Gill and Cobban, 1973).

Bearpaw Shale. The Bearpaw Shale was named for the Bearpaw Mountains by Hatcher and Stanton (1903), and further defined by Stanton and Hatcher (1905). The Bearpaw is recognized across most of northern and central Montana and in Alberta.

The Bearpaw crops out in a broad, rolling and gullied area northeast of the steep flank of the Cat Creek anticline (Nelson, 1993). The outcrop belt is largely soil-covered. A few large exposures of the shale are available in cut banks and deep ravines near the Musselshell River and Long Coulee.

The Bearpaw is lithologically similar to the Claggett Shale. Dark gray, soft and moderately fissile clay shale and silty shale compose most of the Bearpaw. Occasional thin intervals of siltstone and fine-grained sandstone are present. Septarian limestone concretions up to 5 feet in diameter occur in layers associated with silty or sandy zones. Silty concretionary layers form small hogbacks on the north flank of the Cat Creek anticline just east of the Musselshell River. Beds of creamy-white bentonite 1 to 2 feet thick crop out in the lower Bearpaw, and also can be identified in well logs (fig. 39). None of the bentonites seem to be regionally widespread.

The contact between the Bearpaw and the Judith River is either sharp or gradational through an interval a few feet thick. These formations intertongue on a large scale across the study area. A widespread sandstone 10 to 30 feet thick occurs 110 to 160 feet above the base of the Bearpaw. This "stray" sandstone coarsens upward from a gradational base to a sharply defined top (fig. 39). The "stray" sandstone may to be a tongue of the Judith River, but it was mapped as part of the Bearpaw in this study.

The Bearpaw is totally eroded from the crest of the Cat Creek anticline and is partially eroded in large areas elsewhere. Where fully preserved, the Bearpaw ranges from 775 to nearly 1250 feet (fig. 54). The Bearpaw Shale is of Maastrichtian age (Gill and Cobban, 1973; Ballard et al., 1983; Anna, 1986).

Fox Hills Sandstone. Originally defined in South Dakota, the Fox Hills Sandstone was mapped in the Winnett-Mosby area by Johnson and Smith (1964). Equivalent units west of the study area are the Lennep Sandstone and the Horsethief Formation (Balster, 1980). The Fox Hills and younger units were not examined in detail for the present study because they do not crop out in the Cat Creek oil field, and are penetrated by relatively few wells.

The Fox Hills is a unit of light to medium gray, very fine to medium-grained sandstone. The sandstone is "salt and pepper" and contains much interstitial clay. Bedding is variable; many portions are crossbedded. Deformed bedding, which may represent syndepositional slumping or dewatering of sediment, was observed near Mosby dome. Occasional interbeds of shale and siltstone are present. The lower contact of the Fox Hills is gradational to the Bearpaw through an upward-coarsening interval several tens of feet thick.

Thickness of the Fox Hills is 63 to 141 feet according to Johnson and Smith (1964), or 60 to 120 feet according to Frahme (1979). Age of this unit is Maastrichtian (very late Cretaceous) (Gill and Cobban, 1973; Anna, 1986.)

Hell Creek Formation. The Hell Creek Formation was defined by Brown (1907) and named for Hell Creek in northeastern Garfield County. Johnson and Smith (1964), Frahme (1979), and Stoner and Lewis (1980) mapped the Hell Creek in portions of the present study area. Lithology of the Hell Creek, as described by the above authors, in sample studies from wells, and as personally observed at a few outcrops, is similar to that of the Judith River Formation. Clay-rich, poorly sorted "salt and pepper" sandstone occurs as lenticular bodies, intertonguing with bentonitic siltstone, mudstone and claystone, which is banded and variegated in dull shades of gray, green, and brown. Discontinuous beds of shaly lignite or lignitic claystone are common in the Hell Creek. I observed a claystone containing fossil roots beneath a lignite in the lower Hell Creek northeast of Mosby dome. The Hell Creek is well known for dinosaur fossils.

The upper part of the Hell Creek is eroded in most of the study area. Johnson and Smith (1964), citing unpublished studies, gave the thickness as 350 to 550 feet in Garfield County. Frahme (1979) reported a thickness of 200 to 280 feet in the same area. The lower contact is either gradational or locally disconformable (Frahme, 1979). Age of the Hell Creek is late Maastrichtian, or latest Cretaceous (Gill and Cobban, 1973; Anna, 1986).

Tejas Sequence

The Tejas Sequence (Sloss, 1963) comprises strata of Tertiary age. In the study area the Tejas Sequence is represented only by the Fort Union formation.

Fort Union Formation. The Paleocene Fort Union Formation overlies the Hell Creek in the northeastern part of the study area. No information on this unit was obtained from wells, because it is generally behind surface casing and not logged. Frahme (1979) and Stoner and Lewis (1980) mapped the Ft. Union in Garfield County. Frahme reported the lower Tullock Member to be 150 to 175 feet thick, and the upper Lebo Member 75 feet thick with the top eroded. The Tullock consists of light-colored sandstone, siltstone, and mudstone, containing abundant plant remains and thick, widespread coal beds. The Lebo Member was described by Frahme as being largely cross-bedded, channel-fill sandstone.

Quaternary deposits

Quaternary deposits were not systematically examined during this study, but Pleistocene terrace gravels and Holocene alluvium were mapped in the Cat Creek oil field by Johnson and Smith (1964) and Nelson (1993). Terrace gravels are commonly 5 to 10 feet thick and consist of well-rounded pebbles and small cobbles in a matrix of sand. Clasts include abundant limestone and chert, probably from the Big Snowy and Judith Mountains, along with locally-derived fragments of sandstone and limestone concretions. Occasional cobbles and boulders of granite and other plutonic rocks up to 2 feet in diameter are scattered on upland surfaces. The closest source of such rocks is in the Little Rocky Mountains.

Johnson and Smith (1964) mapped seven terrace levels in the Musselshell Valley near Cat Creek. The highest terrace is 635 feet above the altitude of the river. Porter (1990) mapped many high-level terrace gravels west of the oil field. Alluvium along the Musselshell River and its tributaries is composed of crudely stratified clay, silt, sand, and rock fragments.

No structural deformation of Quaternary deposits was found during this study or reported by other authors. Horizontal gravel terraces truncate steeply dipping Cretaceous rocks on the flank of the Cat Creek anticline along the Musselshell River (fig. 55).

TECTONIC HISTORY

Precambrian

The Belt Supergroup in central Montana occupies an east-trending salient bounded on the north by the Cat Creek fault (fig. 3). This salient is interpreted as a failed rift or aulacogen, called the Montana aulacogen by Shepard (1987), that developed during Proterozoic time. The aulacogen ran perpendicular to the main basin of Belt sedimentation, which extended northward through Idaho, western Montana and adjacent parts of Canada (fig. 3).

The southern boundary of the Montana aulacogen is a series of high-angle faults collectively called the Perry line (Winston, 1986). The Perry line separates Archean crystalline rocks on the south from Belt rocks on the north, and it extends from the Highland Range south of Butte to the Bridger Range north of Bozeman (fig. 3). Close to the Perry line the Belt contains coarse synorogenic conglomerate (LaHood Formation) that intergrades northward were finer-grained clastics (McMannis, 1963; Schmidt and Garihan, 1986; Winston, 1986). The Pass fault in the Bridger Range is the easternmost exposed segment of the Perry line. It was a normal fault downthrown to the north during Proterozoic time and was reactivated as a reverse fault during the Laramide orogeny (Lageson, 1989). The Pass fault connects with a buried, seismically-defined fault that underlies the northeast-trending Battle Ridge monocline in the Crazy Mountains basin (Garrett, 1972). Fault and monocline extend toward the Big Snowy Mountains. The Precambrian normal fault may have acted as a buttress against Laramide thrusting that produced the near-surface monocline (Garrett, 1972).

The northern border of the Montana aulacogen is concealed by younger rocks along most of its length. The Volcano Valley fault in northern Meagher County may be one segment (Winston, 1986). Eastward the boundary passes between the Big

Snowy Mountains, where the Belt crops out (Lindsey, 1980) and the North Moccasin Mountains, which contain xenoliths of pre-Belt schist (Lindsey, 1982). The northern boundary, projected farther east, intersects the Cat Creek fault (fig. 3).

Faults in the Cat Creek area resemble faults formed in other rifts worldwide. Rift-bounding faults are commonly listric normal faults (Harding, 1984; Shelton, 1984; Bretagne and Leising, 1991). Proprietary seismic profiles and borehole data suggest that the Cat Creek fault is listric. The Sand Creek fault was interpreted as listric on published seismic profiles (Plawman, 1983; Stephens and Rodenbaugh, 1985). Also, the "dogleg" form of the Montana aulacogen in map view is shared with other ancient failed rifts, such as the Reelfoot rift and Rough Creek graben in the central Mississippi Valley (Howe and Thompson, 1984; Nelson, 1991). Reactivation of ancient rift-bounding normal faults under later compression is common. This has occurred in the Reelfoot/Rough Creek rift (Howe and Thompson, 1984; Bretagne and Leising, 1991; Nelson, 1991), the southern Oklahoma aulacogen (Perry, 1989), and the Midcontinent rift system in Minnesota (Craddock, 1972).

Sauk Sequence

Limited data available suggest that the south side of the Cat Creek fault was uplifted in an episode of reverse faulting during the late Cambrian (Table 2). The best evidence is from two closely-spaced wells that straddle the western part of the fault. A cross section (fig. 6) shows that the Flathead Sandstone and lower part of the Emerson Formation thicken substantially north of the fault. The upper Emerson, defined by electric-log markers, appears to maintain nearly constant thickness across the fault, implying tectonic quiescence. Possible reactivation of the fault with the south side downthrown (normal faulting) is suggested by the thicker section of uppermost Emerson (above marker A) beneath the sub-Tippecanoe unconformity in the southern well (fig. 6).

Uplift of the block south of the Gumbo Ridge fault is strongly suggested by the abnormal Cambrian section in the Conoco # 1-10 Kincheloe well drilled just south of the fault (fig. 7). The Kincheloe well penetrated only 50 feet of Emerson Formation overlying more than 200 feet of Flathead Sandstone. The Emerson is more than 1,000 feet thick in wells north of the Gumbo Ridge fault (fig. 5). However, the control wells are too far apart to demonstrate conclusively that the thickness changes are due to concurrent faulting.

Other structures in the region may have been active at about the same time. The Deadwood Formation (equivalent to Flathead and Emerson) thins across the crest of the Cedar Creek anticline in southwestern Montana (Clement, 1986). The Deadwood also thins across the Nesson Anticline in western North Dakota (Gerhard et al., 1991). Both anticlines, like Cat Creek, are underlain by seismically-defined high-angle faults. Cambrian displacement on the high-angle listric Cedar Creek fault would have been reverse (Table 3).

West of the report area in the South Moccasin Mountains, a photograph by Miller (1959, Plate 3C) shows the Devonian Jefferson Formation overlying Emerson-equivalent rocks with an angular unconformity. This indicates an episode of deformation some time between Late Cambrian and Late Devonian time.

No regional orogenic events are known that might be related to Cambrian displacements of faults on the craton in Montana and North Dakota.

Table 2 - Episodes of displacement on faults in study area

Fault name	Proterozoic	Sauk	Mid-Kaskaskia	Post Absaroka pre-Zuni	Laramide
Cat Creek	Normal ?	Reverse ?	Reverse	Normal	Left-lateral/reverse
East Cat Creek				Normal	Left-lateral/normal
Gumbo Ridge		Reverse	Reverse	Anticlinal warping	Left-lateral
Sand Creek				Normal	Reverse/left-lateral
Cohagen				Normal ?	Reverse ?
Weldon				Normal ?	inactive
Rattlesnake Butte				inactive	Left-lateral/oblique
Flat Willow				inactive	Left-lateral
Box Elder					Left-lateral/oblique
Armells				Normal ?	inactive
North Judith				Normal ?	inactive
Crystal Peak				Normal ?	Reverse ?

Table 3. Comparison of tectonic episodes involving Cat Creek and Cedar Creek anticlines

	Cat Creek (this study)	Cedar Creek (Clement, 1986)
Proterozoic	Normal faulting	No data
Cambrian	Reverse faulting	No data
Post-Silurian, pre-Middle Devonian	No data	Reverse faulting
Late Devonian	Reverse faulting	Reverse faulting
Late Mississippian- early Pennsylvanian	Inactive (?)	Normal faulting
Post-Pennsylvanian, pre-Middle Jurassic	Normal faulting	Normal faulting (two episodes)
Laramide	Reverse/left-lateral faulting	Reverse faulting

Tippecanoe Sequence

The Winnipeg, Red River and Interlake Formations in the study area comprise a sequence of carbonate rocks bounded above and below by unconformities. An isopach map (fig. 9) shows thickness of the Tippecanoe Sequence along with the erosional limit of the Interlake Formation and several faults. Regionally, the Tippecanoe thickens northeastward, toward the Williston basin. Thickness changes of Tippecanoe rocks on opposite sides of the Cat Creek, Sand Creek, Cohagen and Weldon faults are too slight to suggest any syndepositional movements along these structures. A large thickness change, however, occurs across the Gumbo Ridge fault near the Garfield-Rosebud county line. Between two wells drilled three miles apart the Red River thickens from 106 feet south of the fault to 529 feet north of the fault (fig. 7). Note that the Red River is unconformably overlain by the Devonian Duperow Formation north of the fault and by the Mississippian Lodgepole Limestone south of the fault. This relationship implies pre-Mississippian uplift and erosion of the southern block. Whether any uplift occurred during the deposition of the Tippecanoe Sequence cannot be determined from available data.

No faulting or other tectonic deformation during deposition of the Tippecanoe Sequence is reported in adjacent areas.

Lower Kaskaskia Sequence (Devonian)

An isopach map of the Devonian rocks (fig. 11) demonstrates that displacement occurred along the Cat Creek and Gumbo Ridge faults near the end of the Devonian Period (Table 2). Devonian rocks are thin or absent south of the faults, but immediately north of the faults Devonian strata are 150 to 300 feet thick. North of the faults Devonian rocks are unconformably overlain by either the thin Bakken Formation (Devonian-Mississippian) or by the Lodgepole Limestone (Kinderhookian). South of the faults, where Devonian rocks are absent, the Bakken or Lodgepole directly overlies the Red River Dolomite (Ordovician). Closely-spaced pairs of wells straddling the faults show that the thickness changes are due to actual displacements of the faults rather than regional warping or tilting.

Cross sections further illustrate the relationships. One section shows logs of two wells drilled less than a mile apart on opposite sides of the Cat Creek fault (fig. 12). In the northern well the Three Forks Formation is preserved, whereas in the southern well all of the Three Forks and the upper part of the Birdbear were eroded before the Lodgepole Limestone was deposited. The cross section indicates that 80 to 90 feet of Devonian strata were eroded from the uplifted southern block prior to Lodgepole deposition. Another section (fig. 13) compares logs of two wells farther east along the Cat Creek fault. Here 225 feet of Devonian rocks, along with about 50 feet of the Ordovician Red River Dolomite, were eroded from the southern block prior to Mississippian sedimentation. About 275 feet of fault displacement is implied here.

Even greater displacement is indicated by a cross-section across the Gumbo Ridge fault (fig. 7). The interval from the base of the Lodgepole to the base of the Red River is 685 feet thick north of the fault and 105 feet thick south of the fault. Subtraction yields 580 feet of uplift of the southern block before Lodgepole deposition. In this case, however, it is possible that some of the displacement was pre-Devonian.

Given that the Cat Creek fault dips southwest, its Devonian displacement was reverse. This is the same as the postulated Cambrian displacement, but opposite to the inferred normal displacement during Proterozoic (Belt) sedimentation. The dip of the Gumbo Ridge fault is unknown.

Thickness patterns indicate that the region north of the Cat Creek and Gumbo Ridge faults was tilted northward during early to middle Frasnian (early Late Devonian) time. The Souris River Formation thins southward and pinches out, and is overlapped by the Duperow Dolomite (fig. 14). The Duperow also thins southward, reflecting continued northward tilt. The Birdbear Formation is fairly constant in thickness where its top is not eroded, suggesting relatively uniform subsidence during its deposition. Northward tilting resumed during or after deposition of the Three Forks Formation (Famennian; late Devonian) and before the onset of Lodgepole sedimentation. The result was southward beveling of the Birdbear and Three Forks Formations at the unconformity at the base of the Lodgepole (fig. 14).

Elsewhere, reverse faulting took place at the same time on the Cedar Creek fault in southeastern Montana. The block northeast of the fault was raised and tilted northeastward, with maximum displacement of more than 800 feet (Clement, 1986). This is one of several tectonic episodes shared by the Cat Creek and Cedar Creek faults (Table 3). Also, warping or minor faulting near the end of the Devonian Period is indicated by isopach patterns in northern Montana and along the Nesson anticline (Brown et al., 1984). The Weldon fault in McCone County, northeast of the study area, underwent Late Devonian movement with the northwest side downthrown (Sandberg and Mapel, 1967). In the South Moccasin Mountains the Lodgepole Limestone overlies Devonian and older rocks with an angular unconformity (Miller, 1959).

The Late Devonian intracratonic faulting corresponds closely in time with the culmination of the Antler orogeny on what was then the western margin of North America. The Antler orogeny was an episode of compressional folding and thrust faulting chiefly in Nevada and southern Idaho (Nilson and Stewart, 1980; Speed and Sleep, 1982). Central Montana lay in the foreland hundreds of miles east of the Antler orogenic belt. How compressive stress was transmitted across such a distance is unanswered.

Upper Kaskaskia Sequence (Mississippian)

Madison Group

Thickness trends of the Lodgepole Limestone (fig. 16) are subtle and irregular. The Lodgepole thins slightly onto the blocks south of the Cat Creek and Gumbo Ridge faults, which were uplifted during Late Devonian time. Comparison of logs of closely-spaced wells shows the Lodgepole to be 40 to 60 feet thinner south of the faults than north of the faults. Most, if not all of the thinning takes place in the lower part of the Lodgepole (fig. 12, 13). Thus, the blocks south of the faults probably were still elevated at the onset of Mississippian sedimentation. No thickness changes attributable to faulting can be detected in middle and upper Lodgepole strata, signifying that the area was tectonically stable by that time.

Mission Canyon isopachs (fig. 17) show no trends that can be related to faults or other structures in the study area. Similarly, the isopach map of the Charles Formation (fig. 18) does not suggest contemporaneous faulting. The

Charles was partially eroded beneath Middle Jurassic strata in most of the area north of the Cat Creek, Sand Creek and Cohagen faults. This erosion took place after the Mississippian Period.

Big Snowy Group

Except in a few small areas, the Big Snowy Group is eroded north of the Cat Creek, Sand Creek, Cohagen and Weldon faults (figs. 20, 21, and 24; plate 1). This erosion reflects post-Mississippian tectonic activity that will be discussed later in this chapter.

In the area where the Kibbey Formation is fully preserved the Kibbey is thinnest along the southern edge of the study area and thickest adjacent to the Cat Creek fault (fig. 20). The Kibbey also is thick near the junction of the Sand Creek and Cohagen faults and along the western edge of the study area. The isopach pattern suggests that slight downwarping occurred near these faults. The magnitude and extent of such warping cannot be defined because the Kibbey is absent from most of the region north of the faults. A complete section of Kibbey is preserved in only one well north of the faults mentioned. In this well (The California Co. # 1 Peterson, northeast of Mosby dome), the thickness of the Kibbey is nearly the same as in nearby wells southwest of the Cat Creek fault. This finding indicates that little or no displacement occurred along this portion of the Cat Creek fault during deposition of the Kibbey.

The Otter Formation is thickest along a belt that trends ESE from the western edge of the study area (fig. 21). The belt of thick Otter Formation is 10 to 20 miles south of and nearly parallel with the Cat Creek, Sand Creek and Weldon faults. The Otter thins on the north flank of this belt, but it thickens again close to the faults. Overall, the Otter is thickest to the west. The isopach pattern implies growth of a trough south of the faults during deposition of the Otter. This downwarp is the Big Snowy trough (also called Central Montana trough) of previous authors (Smith and Gilmour, 1979; Brown et al., 1984; Peterson, 1984). The role of the Cat Creek fault and other faults during Otter deposition cannot be assessed because the Otter is absent (due to post-Mississippian uplift and erosion) north of the faults.

In several areas the Otter undergoes abrupt, unexpected thickness changes. One such area is on Kootenai dome, where intervals of missing section in boreholes signify normal faulting (fig. 25). Wells that encountered faults in the Otter are indicated with the symbol "F" in fig. 21, and were not used in isopach contouring. These wells are: Alpha Petroleum #3 Federal (Sec. 7, T16N, R26E), Arco #1-32 Salzgeber (Sec. 32, T15N, R35E), and Arco #1-33 Burlington-Northern Coastal (Sec. 33, T15N, R35E). None of these wells nor any directly adjacent wells encountered faults in strata other than the Otter. Thus, faulting in the Otter may reflect syndepositional slumping, possibly triggered by seismic activity related to tectonic faults in the area.

One well encountered abnormally thick Otter Formation. The 612 feet of Otter logged in the Cardinal #1 Northern Pacific well (Sec. 35, T12N, R29E) is a maximum for the study area, and 150 to 250 feet thicker than in nearby wells. Most of the excessive thickness in the Cardinal well is in the upper part of the Otter. There is no evidence of repeated section. Local syndepositional slumping or tectonic subsidence provides a possible explanation for the over-thickened Otter.

The isopach pattern for the Heath Formation (fig. 24) somewhat resembles that of the Otter Formation. The Heath is thickest along the Big Snowy trough in the western half of the study area. Within the belt of thick Heath the contact to the overlying Tyler Formation is gradational and probably conformable. Outside of this belt, the Heath-Tyler contact is unconformable, the Tyler filling deep paleovalleys eroded into or through the Heath. Pre-Tyler erosion explains abrupt local changes in thickness of the Heath in the eastern half of the study area.

In summary, subsidence was greatest close to the Cat Creek, Sand Creek and Cohagen faults during deposition of the Kibbey. The Big Snowy trough appeared during Otter deposition, and persisted through deposition of the Heath Formation. The trough transects the crustal block that was uplifted near the end of the Devonian Period.

A corresponding episode of normal faulting formed a graben along what is now the crest of the Cedar Creek anticline during Chesterian time (Clement, 1986). Normal faulting also took place along the Nesson anticline near the end of the Mississippian Period (Gerhard et al., 1991). Displacements in both places are small in magnitude.

The events discussed above suggest that by late Mississippian time the compression responsible for Devonian reverse faulting had relaxed, allowing uplifted blocks to collapse. The Antler orogeny had run its course by the end of Mississippian time.

Absaroka Sequence

The isopach pattern of the Tyler Formation (fig. 26) strongly suggests tectonic control of sedimentation. Several areas where the Tyler is thickest correspond with present-day anticlines and fault zones. As Cooper (1956) and Fanshawe (1978) previously hypothesized, these present-day highs evidently were lows during Tyler deposition, and were inverted during the Laramide orogeny.

Three areas of thick Tyler are infilled paleovalleys that trend east or northeast (fig. 26). One northeast-trending paleovalley crosses the southeast corner of the study area. This feature lacks obvious structural control.

The Gumbo Ridge paleovalley (southwestern Garfield and northwestern Rosebud County) extends from Alice dome to the Sumatra anticline just south of the study area. It connects with a larger complex of paleovalleys that follow the Sumatra anticline westward (Stanton and Silverman, 1989). At its north end the Gumbo Ridge paleovalley widens slightly along the East Cat Creek fault.

A third Tyler paleovalley, here called the Rattlesnake Butte paleovalley, runs westward from the Rattlesnake Butte oil field (T13N, R28E) to the northern part of T13N, R24E, where it bends southwest. The east-west segment of this paleovalley coincides with the Rattlesnake Butte fault zone and the Flat Willow anticline and fault zone. The southwest-trending segment crosses the crest of Spindletop dome (a feature better described as an asymmetrical anticline).

Two areas of thick Tyler, not related to paleovalleys, also are on structural highs. One follows the crest of the Cat Creek anticline, and the other is on the southwest side of the Cohagen fault near Freedom dome.

In summary, crustal blocks south of the Cat Creek, East Cat Creek and Cohagen faults apparently moved downward during Tyler deposition, while a syncline or graben developed along the Rattlesnake Butte-Flat Willow-Spindletop trend. Thus, in every case Tyler displacement or folding was opposite in sense

to Laramide deformation. The amount of displacement or downwarping may not have been large. Subtle depressions could have controlled the trends of paleovalleys that eroded deeply during the pre-Absaroka regression.

The Big Snowy trough also persisted during Tyler deposition. The Tyler is thicker in the trough than elsewhere and apparently is conformable to the Heath. Moreover, the marine Bear Gulch facies is best developed within the trough. The presence of the deep Rattlesnake Butte paleovalley within the trough is perplexing. How could a subaerial valley be cut below the level of continuously-deposited Heath and Tyler strata on both sides? Either the Tyler-Heath contact here is not really conformable, or else the valley was not cut subaerially. Perhaps it is a submarine channel analogous to the Mississippian submarine channel-fill in the Illinois basin described by Sedimentation Seminar (1969).

The Alaska Bench/Devil's Pocket Formation in the study area is a sub-Zuni erosional remnant. Thickness of the interval therefore reflects post-Absaroka events. No subdivisions of the interval, usable in interpreting contemporaneous tectonism, were identified. The relatively uniform lithology of the Alaska Bench/Devil's Pocket, as compared to the highly variable Tyler, suggests general stability during Alaska Bench/Devil's Pocket deposition.

Pre-Zuni hiatus

A major unconformity separates rocks of the Absaroka Sequence from those of the Zuni Sequence throughout the report area. A pre-Zuni paleogeologic map (Plate 1) shows that basal Zuni Sequence (Middle Jurassic) strata overlie units ranging from the Madison Group to the Devil's Pocket Formation. Maximum stratigraphic relief on the unconformity is more than 1,500 feet.

A major episode of normal faulting took place during this long interval of erosion. Large displacements occurred on the Cat Creek, East Cat Creek, Sand Creek, Cohagen and Weldon faults (Table 2). In every case, crustal blocks north of the faults were uplifted and eroded, while the southern blocks were downthrown and protected. Displacements also occurred on several newly identified faults near the western edge of the study area. These faults are the Armells, Crystal Peak, and North Judith faults (Table 2, Plate 2). Together, these east-trending faults outlined a graben that probably merged with the western end of the Cat Creek fault. Pre-Zuni displacements were opposite in direction to displacements that occurred during the Laramide orogeny (Table 2).

The largest displacements were along the Cat Creek fault. The entire Amsden and Big Snowy Groups and the upper part of the Madison Group were eroded north of the fault, whereas south of the fault the Alaska Bench/Devil's Pocket Formations are preserved (plate 1). Pairs of wells that straddle the fault show that pre-Zuni displacement was confined to a zone less than one mile wide and coincides with Laramide reverse faults (figs. 27, 28). Maximum pre-Zuni displacement was 1,550 feet on the north side of Kootenai dome (fig. 27). Throw was greater than 1,000 feet from Bear Creek dome eastward to West dome, and diminished to about 700 feet at Mosby dome (fig. 28).

The pre-Zuni outcrop pattern (plate 2) also indicates that an anticline developed 5 to 8 miles south of and subparallel with the Cat Creek fault. This anticline turns sharply northward near its eastern end. The Otter, Heath, and Tyler Formations underlie the pre-Zuni unconformity along the fold crest; pre-Zuni rocks dip northward from the crest of the anticline toward the Cat Creek fault. Thus, the Alaska Bench/Devil's Pocket is preserved in a northward-

thickening wedge between fault and anticline. Farther east in the Cat Creek oil field, pre-Zuni strata dip southward (away from the fault) beneath the unconformity.

Pre-Zuni displacements of the Sand Creek, Cohagen, Weldon and East Cat Creek faults were 200 to 1,000 feet. Pre-Zuni rocks dip southward in the triangular area north of the intersection of the Cat Creek and Sand Creek faults (plate 1). They also dip southward in the angle south of the junction of the Sand Creek and Cohagen faults. The triangular block east of the intersection of Cohagen and Weldon faults is tilted westward. The Armells, North Judith, and Crystal Peak faults all have pre-Zuni throws of 500 feet or more.

An east-trending pre-Zuni anticline passes through the Gumbo Ridge area, approximately following the Garfield-Rosebud County line. The Tyler Formation subcrops beneath Jurassic strata at the crest of the fold (plate 1). The anticline partly coincides with the Gumbo Ridge fault zone, which underwent both Paleozoic and Laramide movements. Good well control near Gumbo Ridge precludes any significant post-Absaroka, pre-Zuni fault displacements here. Pre-Zuni normal faulting is documented in several areas of Montana outside the study area. Borehole data indicate widespread normal faulting during and after deposition of Pennsylvanian rocks south and west of the Big Snowy Mountains (John Grauman, personal communication, 1990). On the Cedar Creek anticline (Table 3) at least two periods of normal faulting occurred between Pennsylvanian and Middle Jurassic time (Clement, 1986). The first was mid-Permian (during deposition of the Opeche Formation) and the second was post-Early Triassic (Spearfish Formation). The Weldon fault in McCone County also underwent two episodes of displacement, but they were at different times and opposite to one another. First, the northwest side of the Weldon fault was uplifted after Middle Pennsylvanian time, later the northwest block was lowered during or shortly after deposition of the Early to Middle Jurassic Nesson Formation (Edmiston and Foster, 1969).

Thus, two or more tectonic stress regimes may have operated in central and eastern Montana during the long post-Absaroka, pre-Zuni interval. Without more precise information on the timing of pre-Zuni faulting, relating these displacements to regional orogenic events is difficult. The best-known orogenic event in the western U.S. during the interval in question was uplift of the Ancestral Rockies, which appears to have been an episode of compressive block-faulting. How compressional events in the Ancestral Rockies might have triggered extensional tectonics in Montana is beyond the scope of this report.

Zuni Sequence

Ellis Group

Following the long interval of pre-Zuni erosion, the sea transgressed into the northeastern part of the study area in early Middle Jurassic time. Initial deposits comprise siliciclastics and evaporites of the Nesson Formation (?) (Plate 3). These sediments probably accumulated on a shallow shelf or in lagoons, the clastic portion being derived from pre-Jurassic sedimentary rocks such as the Kibbey Formation.

With continuing transgression the Tampico Member of the Piper Formation was laid down throughout the eastern half of the report area and in portions of the western half. Two large areas received no Tampico sediments (Plate 2, fig. 35). One was the block north of the Cat Creek fault, which was elevated prior to

deposition of the Tampico. The other large area of nondeposition lay in the southwestern quarter of the study area. This was the eastern end of Belt Island, a Jurassic upland first described by Imlay et al. (1948).

Belt Island was a lobate upland that extended as far east as T12N, R30E during deposition of the Tampico Member. Along the margins of the island the Tampico pinches out and is progressively overstepped by younger units of the Ellis Group (plate 2; fig. 35). The eastern end of the island sloped gently; the Tampico thins gradually here and outliers (deposited in isolated lagoons?) are present. The northern margin of Belt Island is sharp and nearly linear, suggesting fault control; yet closely-spaced wells provide no evidence for faults. The northern edge of Belt Island (plate 2) closely corresponds with the northern limit of the Alaska Bench/Devil's Pocket (plate 1). The Alaska Bench/Devil's Pocket probably was more resistant to erosion than the shale-dominated Tyler Formation north of Belt Island. The outlines of Belt Island may have been controlled more by differential erosion under an arid climate than by tectonic activity.

A scarp existed along the Cat Creek fault during early Middle Jurassic time. Massive limestones of the Madison Group on the upthrown northern block evidently formed a plateau under the arid conditions (Peterson, 1972) then prevailing. South of the fault, the shaly Tyler Formation and thin-bedded, shaly Alaska Bench/Devil's Pocket Formations eroded more readily than the Madison Group. Local deposits of sand eroded from the upthrown block were incorporated into the lower part of the Tampico along the base of the fault scarp. This sand resembles that from the Kibbey Formation, which is now largely stripped from the northern block.

A smaller deposit of basal Tampico sandstone occurs in the graben just north of the Crystal Peak fault (plate 2). The floor of the graben between the Armells and North Judith faults apparently was tilted toward the south; the Tampico was deposited only along the southern edge of this graben. The block between the North Judith and Crystal Peak faults probably was folded in such a way that the thickest Tampico accumulated near the southern edge.

The East Cat Creek fault influenced Tampico sedimentation also. The Tampico is less than 20 feet thick south of the fault and more than 50 feet thick north of the fault. The southern block, where the Tampico is thin, was downthrown prior to Jurassic sedimentation, and was capped by Alaska Bench/Devil's Pocket. The northern, upthrown block was capped by the nonresistant Tyler Formation. Perhaps an inverted fault scarp developed, with the Tyler on the upthrown block being eroded to a valley while the Alaska Bench/Devil's Pocket on the upthrown block became a ridge. A thicker Tampico section then accumulated in the topographic low.

The Sand Creek, Cohagen and Weldon faults exerted little influence on Tampico sedimentation, although all these faults underwent large pre-Tampico displacements and juxtapose rocks of contrasting erodability. Possibly, these fault scarps were eroded for a longer period of time than scarps farther west.

The Firemoon Member of the Piper Formation reflects continuing transgression of the sea. Oolites, anhydrite and a diverse fauna signify deposition on a shallow, stable shelf where circulation was locally restricted (Rayl, 1956). The Firemoon overlaps the Tampico along the edges of Belt Island and on the eastern end of the block north of the Cat Creek fault (plate 2, fig. 35).

Distribution of the Bowes Member of the Piper reflects further transgression which gradually submerged most of Belt Island and the block north of the Cat Creek fault. The study area apparently was tectonically quiet and subsidence

rates fairly uniform. The earlier topography of scarps was gradually being buried.

A minor regression may have occurred at the onset of Rierdon sedimentation, as suggested by the thin, sandy oolitic limestone beds at the base of the Rierdon. The remainder of the Rierdon reflects a long-lasting transgression followed by a brief regression (Rayl, 1956; Peterson, 1972). Nearly uniform thickness and lithologic sequence of the Rierdon in the eastern half of the study area (fig. 33) indicates stability and uniform subsidence. The Rierdon thins toward the west, but it overlaps the entire crustal block north of the Cat Creek fault and all but a small portion of Belt Island (plate 2, fig. 35). The upper part of the Rierdon was eroded near these uplands when the sea withdrew just prior to deposition of the Swift Formation (late Callovian to early Oxfordian time).

Northeastward thickening of the Swift Formation (fig. 34) reflects renewed downwarping of the Williston basin during the Oxfordian Age. The overall upward-coarsening profile of the Swift is consistent with a transgression followed by progradation into the basin. Increasing sand content westward, particularly in the lower Swift, points to a primary sediment source on still-emergent parts of Belt Island west of the report area. Studies by Fox (1982) in southwestern Montana and Porter (1989) in the Great Falls area confirm that Belt Island was an important source of Swift clastics. The Swift and equivalent strata represent environments such as shallow offshore shelves, tidal flats, tidal channels, and lagoons (Kvale, 1986; Porter, 1989).

Morrison and Kootenai Formations

The Morrison and Kootenai Formations comprise an interval of nonmarine deposits sandwiched by marine rocks. During a gradual and perhaps intermittent regression, shallow marine sediments of the Swift were overlapped by Morrison coastal-plain deposits. Mountains began to rise in western Montana in response to the Sevier orogeny. Detritus from these mountains was carried eastward across the study area and deposited in a variety of fluvial-channel, overbank, swamp and lacustrine environments. The climate was subtropical, arid to semi-arid, and marked by pronounced seasonality. Many streams probably were intermittent. As the region aggraded, internal drainage-basins developed, within which upper Morrison peat (coal) accumulated (Silverman and Harris, 1967; Peterson, 1972; Walker, 1974; Kvale, 1986).

After a brief hiatus, renewed uplift of the Sevier orogenic belt brought an influx of coarse sand that now comprises the Pryor Member of the Kootenai (Third Cat Creek sand). The climatic, paleotectonic and environmental setting during the remainder of Kootenai sedimentation was basically the same as during Morrison deposition (Silverman and Harris, 1967; Walker, 1974; Kvale, 1986). Kootenai sedimentation was brought to a close by onlap of the interior Cretaceous sea.

The combined Morrison/Kootenai isopach map (fig. 38) illustrates the westerly source and overall east-southeasterly transport of these sediments. The combined Morrison/Kootenai is thickest along a narrow belt that runs a short distance south of and subparallel with the Cat Creek fault. The sediment belt crosses the fault near its eastern end and continues eastward, now paralleling the East Cat Creek fault. The isopach pattern suggests structural downwarping along these faults. Thick basal Morrison sandstones that are incised into the Swift Formation in the Gumbo Ridge area (fig. 37) further imply local uplift.

These sandstones occur on the southern, structurally high side of the East Cat Creek fault.

Isopachs of the gross Morrison/Kootenai interval cannot reveal details of Late Jurassic-Early Cretaceous paleotectonic events. Detailed isopach and lithofacies mapping of these strata might provide a more definitive story.

Marine Cretaceous rocks

The Cretaceous Period was a tectonically quiet time in central Montana. The only local tectonic activity that is suggested is mild upwarp of the Cat Creek anticline and corresponding downwarp of the Blood Creek syncline during deposition of the Judith River Formation. No evidence was found for displacements on any faults in the study area during deposition of marine Cretaceous strata. Tectonic events here apparently were limited to mild, regional warping that was mostly independent of mapped structures.

Many Cretaceous units are either bounded by bentonite beds or their boundaries maintain consistent positions relative to regionally traceable bentonite beds. Bentonites are volcanic ash deposits and thus represent time lines. Some formation or member contacts involve intertonguing, and cross bentonite-defined time lines. Among time-transgressive contacts are: (1) Skull Creek-Cyprian contact in northwestern part of study area, (2) Shell Creek-Mowry contact in southwestern part of area, (3) Telegraph Creek-Niobrara contact locally disconformable, (4) Telegraph Creek-Eagle contact throughout the report area, and (5) Claggett-Judith River contact.

The Fall River Sandstone represents the basal transgressive sand of the interior Cretaceous seaway. Its thickness is fairly uniform regionally across the study area, but local large fluctuations in thickness are apparent (fig. 40). Most of these appear to reflect Fall River paleochannels eroded into the Kootenai Formation. Edwin Maughan (written communication, 1990 and 1991) characterizes the sub-Fall River disconformity as a ravinement surface. Some local thickness changes in the Fall River, however, may simply be due to incorrect placement of the Fall River-Kootenai contact.

The basal siltstone member of the Thermopolis Shale thickens and coarsens gradually toward a NNE-trending belt located in the eastern part of the study area (fig. 41). The consistent internal stratigraphy of this unit suggests that the upper and lower contacts are both time lines. The isopach pattern suggests that the primary sediment source for this unit lay south of the report area.

The Skull Creek Member of the Thermopolis is thinnest in the east-central part of the report area, and thickens toward the northeast corner (fig. 42). It thickens more dramatically and becomes sandy toward the northwest, where it intertongues with the overlying Cyprian Sandstone Member. A minor sediment source to the northeast and a major one to the northwest are thereby indicated. The lower contact of the Skull Creek is believed to represent a time line. The upper contact is a time line that is consistently positioned with respect to bentonite beds except in the area where the Skull Creek and Cyprian intertongue. In the area of intertonguing, the Skull Creek represents a shorter time interval than elsewhere.

The Cyprian Sandstone is thickest and coarsest along an elongate east-trending lobe north of the Cat Creek fault in the northwestern part of the study area (fig. 43). The sandstone becomes very thin or absent in the eastern half of the area. Thus the Cyprian was derived from a source northwest of the area.

The lower part of the Cyprian grades and intertongues southeastward with shales of the Skull Creek. This evidence indicates that the Cyprian is separate from the Newcastle and Muddy Sandstones, units of similar age in southern Montana. The upper contact of the Cyprian is consistently positioned with respect to bentonite beds in the overlying sandy member, and is assumed to be a time line.

The sandy member of the Thermopolis Shale is thickest on the west and thins gradually and steadily eastward across the study area. It thickens slightly near the northeast corner (fig. 44). The isopach pattern of the sandy member resembles that of the Skull Creek Member. A major western source and a lesser one to the northeast are implied. The sandy member represents a uniform interval of time in the study area; both contacts are closely bracketed by regionally traceable bentonites.

The Shell Creek Member of the Thermopolis shows inconsistent thickness trends (fig. 45). It is 100 to 125 feet thick in most of the area, except near the southwest corner, where the upper part of the Shell Creek intertongues with siltstone and sandstone of the Mowry. The lower contact is a bentonite bed and a time marker. The upper contact also is a bentonite except in the area where Shell Creek and Mowry intertongue, where the contact steps downward across time lines.

The Mowry Shale thickens steadily and its grain size increases toward the southwest corner of the study area (fig. 46). It continues to thicken southwest of the study area, reaching a maximum of over 700 feet in south-central Montana (Anna, 1986a). The Mowry sediment is believed to be largely volcanic in origin (Rubey, 1931). Evidently the volcanoes were located in the Sevier orogenic belt in southwestern Montana. The Mowry contains several regionally persistent bentonite beds that indicate little or no time-transgressive character, except in the southwestern corner of the area where the base of the Mowry is older than elsewhere.

The Belle Fourche Shale thickens steadily toward the east-southeast across the study area (fig. 47). Grain size does not change appreciably. The isopach pattern suggests a southeastern source or, alternatively, more rapid subsidence of the southeastern part of the study area. In either case the source area was remote. Both contacts of the Belle Fourche are bentonite-defined and presumed to be time lines.

The bentonite-bounded Greenhorn Formation thickens very gradually and steadily westward without becoming noticeably coarser (fig. 48). Distant sources and very broad regional warping are again suggested.

The Carlile Shale thickens eastward gradually and steadily (fig. 49). A slight increase in sandstone in the upper part of the Carlile occurs toward the north. Both contacts of the Carlile, as mapped in subsurface, are bentonites; this unit represents a constant interval of time throughout the study area.

Thickness of the Niobrara shows no clear trends (fig. 50). In general, the Niobrara thickens westward. Local thinning near the eastern edge of the report area is the result of erosion at the base of the Telegraph Creek Formation. Elsewhere, the contacts of the Niobrara appear to be consistent time lines with reference to bentonite beds.

Because of complex intertonguing, the Telegraph Creek Formation and Eagle Sandstone were isopached as a single unit (fig. 51). The thickness of the interval does not vary much across the study area (fig. 51). The upper boundary is a time-marker (base of Ardmore Bentonite), whereas the lower contact is slightly time-transgressive. Sandstone content increases strongly toward the northwest and less strongly toward the northeast. West of the study area, the

Eagle continues to thicken, and its facies grades from marginal marine to terrestrial (Roberts, 1965; Gill and Cobban, 1973). Thus the principal source of Eagle clastics lay in the Sevier orogenic belt in western Montana. A secondary northeastern source is indicated by the isopach pattern, particularly by the southwest-trending lobe of thicker clastics in T17N, R41E and southwestward, eroded into the Niobrara. Shurr and Rice (1986) attributed facies changes in the Eagle Sandstone to contemporaneous displacement along the Cat Creek fault, but I found no evidence for that assertion.

The Claggett Shale thickens westward across the study area (fig. 52) with no significant change in lithology. The source probably lay far to the west of the study area. The lower contact of the Claggett is a time horizon (base of Ardmore Bentonite Bed), but the upper contact crosses time lines as the shale intertongues with sandstone of the Judith River.

The Judith River Formation is thickest along a belt that lies just northeast of and runs parallel with the Cat Creek anticline (fig. 53). Possibly, the belt of thick sediment reflect initial uplift of the anticline and downwarping of the Blood Creek syncline. Evidence is not definitive, because the Judith River is eroded from the crest and most of the southern flank of the anticline. The lower contact of this formation is time-transgressive, but the upper contact appears to maintain consistent position with respect to bentonites in the lower Bearpaw Shale.

The limited extent of the Bearpaw Shale, Fox Hills Sandstone, and Hell Creek Formation prevent drawing any conclusions as to tectonic activity that might have occurred during deposition of these units.

In summary, there is no evidence for structural influence on sedimentation of any marine Cretaceous unit in the study area except for (perhaps) the Judith River Formation. My mapping does not substantiate claims of paleotectonic control of Cretaceous sedimentation in central Montana put forward by Anna (1986a and b) and Shurr and Rice (1986). Those authors hypothesized contemporaneous movements along a series of northeast- and northwest-trending fault zones or "lineaments" similar to those proposed by Thomas (1974). Maps used by Anna and Shurr & Rice are smaller in scale, cover thicker stratigraphic intervals, use larger contour intervals, and are based on fewer control points than the maps presented here. In other words, their data base was thinner and more generalized than mine. The existence of long, linear fault zones is particularly called into question. My mapping shows that subsurface faults in central Montana exhibit a variety of trends, and commonly curve and bifurcate.

The absence of Cretaceous deformation in the study area is somewhat surprising in view of intense deformation then in progress in western Montana. Evidently stresses from the Sevier orogeny were not transferred effectively to the foreland during the Cretaceous. Lack of Cretaceous foreland deformation contrasts with well-defined earlier (Cambrian, Late Devonian, Late Mississippian, pre-Zuni) reactivation of foreland faults that were much farther from any possible triggering orogenies.

Tertiary (Laramide orogeny)

The main events of Laramide folding and faulting in the study area took place during the early part of the Tertiary Period. These events will be summarized briefly here; details are presented in Nelson (1993).

The Laramide orogeny induced oblique left-lateral/reverse displacements on many faults in central Montana. Some faults underwent nearly pure strike-slip, others underwent nearly pure dip-slip, and still others underwent a combination of the two (Table 2). Master faults, most of which pre-dated the Laramide orogeny, are high-angle to vertical fractures in crystalline basement. They offset competent Paleozoic and lower Mesozoic sedimentary strata, but generally die out upward into folds and en echelon fault zones in the shaly marine Cretaceous rocks at and near the ground surface. Monoclines and asymmetrical anticlines overlie dip-slip faults, whereas belts of en echelon domes and faults signify strike-slip motion on the master faults at depth.

Faults that underwent mainly strike-slip displacement with little or no dip-slip include the East Cat Creek, Gumbo Ridge, Rattlesnake Butte, and Flat Willow fault zones (Table 2). Most of these faults strike east-west. Faults that underwent dip-slip displacement with no apparent strike-slip component are the Cohagen fault, and also the Cedar Creek fault in southeastern Montana (Clement, 1986; Table 3). The Cat Creek fault exhibits large components of both dip-slip (reverse) and strike-slip (left-lateral) displacement, and this fault strikes southeast along most of its length. The overall pattern of faulting outlined above is consistent with deformation in a compressional stress field in which the principal stress axis was oriented NE-SW (Nelson, 1993).

PETROLEUM GEOLOGY

Cat Creek oil field

Discovered in 1920, the Cat Creek oil field was the first significant oil field developed entirely in the state of Montana. The field contains three separate structural traps and has multiple pay zones. The traps are in East, Mosby and West domes, which lie en echelon along the crest of the Cat Creek anticline. (Some authors recognize a fourth dome, Antelope dome, which is a high point near the east end of Mosby dome).

History

The Cat Creek anticline was mentioned first in Bowen's (1914) reconnaissance for coal in the Musselshell Valley. In 1919 and 1920 Charles Lupton, Wallace Lee and L.R. Van Burgh geologically mapped West, Mosby and East Domes. Lupton joined with others to form the Frantz Corporation, which leased nearly all of the structurally enclosed area and commenced its first wildcat well late in 1919. This well, the Frantz No. 1 Charles on Mosby dome in the SW $\frac{1}{4}$ SE $\frac{1}{4}$ NW $\frac{1}{4}$, Sec. 21, T15N, R30E, was completed for an initial production of 10 BOPD from the Second Cat Creek sand in February, 1920. Depth of the discovery well was approximately 1,000 feet. In May, 1920 a well on West dome was completed for 350 BOPD from the First Cat Creek sand. After cleaning out this well, the output reportedly jumped to nearly 2,000 BOPD. Development of the Cat Creek field proceeded apace, and the population of Winnett, the nearest rail terminal, boomed to nearly 2,000 (compared to about 200 today). Petroleum County, with Winnett as the county seat, was formed out of the eastern part of Fergus County in 1922 (Hadley, 1956; King, 1990).

Early development on West and Mosby domes focused almost entirely on the First and Second Cat Creek sands. The Third Cat Creek sand (Pryor Member of Kootenai Formation) proved to be water-bearing in most wells. Small amounts of oil were produced from fractured zones in the Mowry and Thermopolis Shales. The Frantz Company and others that formed during the boom rapidly were absorbed by competitors. By the early 1930's the Continental Oil Co. controlled most of the production in the Cat Creek field.

Sub-Cretaceous production was established in October, 1945 when the Schrock No. 1 Fifer well was completed for 275 BOPD in the Jurassic Swift Formation. This well was located on Mosby dome in the NW $\frac{1}{4}$ NW $\frac{1}{4}$ NW $\frac{1}{4}$, Sec. 21, T15N, R30E. Swift production soon was extended from Mosby dome to East dome. Many wells targeted for the Swift also recovered oil from the younger Brindley sand in the Morrison Formation. The Farmers Union Central Exchange (Cenex) controlled most of the new production, and eventually (1969) bought out Conoco's remaining interests at Cat Creek (King, 1990). Today all production from the Cat Creek field is shipped via pipeline to the Cenex refinery at Laurel, Montana.

Deeper production was established in 1967 at Mosby dome in porous dolomite of the Alaska Bench/Devil's Pocket Formation (commonly called "Amsden"). A few wells also have been completed in the Jurassic Rierdon and Piper Formations. Through the years strata as deep as the Cambrian Emerson Formation have been tested, but the Amsden is the oldest unit that yields commercial production.

Secondary recovery by water injection began at West dome in 1959, and later was applied throughout the field. Water is obtained from the Third Cat Creek sand. At the end of 1988 there were 16 injection wells serving 68 producing wells. Cumulative production from all pay zones in the Cat Creek field is listed as 23,487,920 barrels of oil (Montana Oil and Gas Conservation Division, Annual Review, 1989).

Reservoirs

Reservoir rocks in the Cat Creek oil field are sandstones in the Fall River, Kootenai, Morrison, and Swift Formations and dolomite in the Alaska Bench/Devil's Pocket Formation. Minor production has been achieved from fractured strata in the Mowry and Thermopolis Shales and the Rierdon and Piper Formations.

Production in the Fall River Sandstone (First Cat Creek sand) is confined to the lower 40 to 50 feet of the formation. The upper part of the Fall River is very fine-grained, silty sandstone that is essentially impermeable. The lower Fall River has a porosity of 10 to 26% and permeability of 10 to 150 md (Bartow-Shepard, 1985). The First Cat Creek sand is essentially a tabular unit, continuous throughout the field.

The Second Cat Creek sand comprises a series of lenticular sandstone bodies in the middle part of the Kootenai Formation. No maps of individual sand bodies have been published. Net pay thickness ranges from 10 to 60 feet, porosity from 0 to 20%, and permeability from 0 to 393 md (Bartow-Shepard, 1985). The basal Pryor Sandstone Member (Third Cat Creek sand) has high porosity and permeability, but is water-bearing in most wells.

The Brindley sand in the middle part of the Morrison Formation is a lenticular unit having porosity as high as 33% and permeability as high as 553 md (Bartow-Shepard, 1985). Production from the Swift Formation is from sandstone at or near the top of the formation. Typical porosity of the Swift pay zone is

11 to 25%, and permeability is 12 to 162 md (Bartow-Shepard, 1985). The Swift is continuous throughout the field, but yields oil only in East and Mosby domes.

The deepest production in the Cat Creek field is from dolomite at the top of the Alaska Bench/Devil's Pocket Formation at Mosby dome. Dolomite in cores displays intergranular porosity, fracture porosity, and solution cavities. Dissolution beneath the pre-Zuni unconformity may be the major cause of permeability. Occurrence of suitable reservoir rock at this position probably is sporadic.

Source

The source of oil in the Cat Creek field is an enigma. Geologists familiar with oil in central Montana have suggested to me possible source rocks ranging from the Precambrian Belt Supergroup through the marine Cretaceous shales. While several units have favorable source-rock properties, none are known that fit the character of oil produced at Cat Creek.

The unit most promising from the standpoint of total organic carbon (TOC) content and thermal maturity is the Heath Formation, specifically the Cox Ranch Oil Shale Bed. Samples of the Cox Ranch Shale from the Big Snowy Mountains yielded TOC of 5.5 to 19.9% (Derkey et al., 1985). In the Williston basin Williams (1984) reported TOC as high as 9% in shales from unspecified positions within the Heath Formation. The Cox Ranch Bed is the probable source of oil produced from the Tyler Formation in the Gumbo Ridge area and from numerous large fields (Sumatra, Stensvad, Jim Coulee, etc.) immediately south of the report area. In these fields Tyler production is generally confined to areas where reservoir-quality sandstones are directly in contact or in close proximity to the Cox Ranch. This relationship was identified by Krantzler (1966) and confirmed by Maughan (1984). In the Gumbo Ridge area (discovered in 1975), the productive Tyler sands are directly in contact with the Cox Ranch Bed.

Oil produced from the Tyler Formation, however, differs greatly from oil produced at Cat Creek. Cat Creek oil has an API gravity of about 48, a pour point below 5 degrees F., and a gasoline/naphtha content of 58%. Oil from the Jim Coulee field, a typical Tyler/Heath field, has a gravity of 31, a pour point of 70 degrees F., and a gasoline/naphtha content of only 22% (French, 1985). Thus the Cat Creek oil has a much larger proportion of the lightweight, volatile components than the oil from Jim Coulee. If both oils were derived from the Heath, why is Cat Creek oil, which migrated to reservoirs hundreds of feet above the Heath, lighter than Tyler oil that remained virtually in situ?

The Bakken Formation is a major source of oil in the Williston basin, having an average TOC of 3.84% and a maximum of over 10% (Williams, 1974). The Bakken/Cottonwood Canyon in central Montana is too thin to be a likely source of oil. Other Paleozoic rocks from the Williston basin analyzed by Williams (1974) contained TOC of less than 1%, with typical values of 0.2 to 0.4%. No published analyses are available for any rocks below the Heath in central Montana.

Some samples of marine Cretaceous shales from the Williston basin contain several percent of TOC, but are not sufficiently mature to generate oil (Williams, 1974). Cretaceous shales in Montana are a source of gas, not oil. Chemical and isotopic studies indicate that Cretaceous gas is biogenic and that these shales have not matured up to the "oil window" (Rice, 1991).

The possibility that oil in the Cat Creek field was derived from Precambrian source rocks cannot be tested with available data. The same is true

of the speculation that Cat Creek oil is abiogenic and migrated upward from great depth along the fault zone.

The source of oil at Cat Creek remains undetermined. Considering thermal maturity, a pre-Cretaceous source seems probable.

Trap

The trapping mechanism at Cat Creek is primarily structural. Oil is produced from reservoirs within structural closure at East, West and Mosby domes. Stratigraphic trapping plays a role in production from the lenticular Second Cat Creek, Morrison, and possibly Swift sands. The Alaska Bench/Devil's Pocket accumulation is partly stratigraphic; dissolution at the sub-Zuni unconformity created the reservoir.

Shales and claystones provide seals to complete the traps. The Morrison and Kootenai Formations and Thermopolis Shales effectively seal the underlying reservoir rocks. These shales behaved in ductile fashion under deformation; they do not maintain open fractures.

En echelon faults seem to exert little influence on trapping, aside from offsetting the reservoir rocks. In most cases, displacements are smaller than thicknesses of pay zones. Hadley (1956, p. 99) wrote, "There is complete sympathy between producing wells drilled on the same dome and supposedly on different fault blocks. If wells in part of the field are shut in, the wells still pumping will show increased production to offset that 'lost' from the shut-in wells. Dye placed in some wells has later shown up in adjacent wells on supposedly separate fault blocks."

Timing

A critical factor in oil accumulation is the timing of oil generation with respect to formation of the trap. If oil is generated before the trap is ready, oil may migrate out of the area. Similarly, a trap formed during one tectonic episode may be destroyed in a later one.

The oil in the Cat Creek field probably was generated before the Cat Creek anticline formed. Maximum depth of burial and, hence, maximum maturity of source rocks presumably was attained in Late Cretaceous time, before the anticline began to form. Once the fold began to rise, overburden was stripped from its crest.

Only three out of nine en echelon domes along the Cat Creek anticline have yielded commercial quantities of hydrocarbons. All six unproductive domes have been tested to Mississippian rocks, three have been tested to Cambrian, and one to Precambrian. A few tests yielded shows of oil, but water was the only recoverable fluid. Nonproductive domes all have closure, and the seal (marine Cretaceous shale) is intact except at Kootenai dome. No significant differences in stratigraphy or character of reservoir rocks are known between producing and nonproductive domes.

Perry (1959, p. 59) postulated that strong artesian water flow in central Montana flushed oil out of many domes along the Cat Creek anticline. According to Perry, the numerous en echelon faults at East, Mosby and West domes blocked water flow and prevented flushing. Perry's hypothesis falls short for two reasons. First, en echelon faults are present in the unproductive domes as well as in the productive ones (Reeves, 1927; Johnson and Smith, 1964; Porter, 1990;

Nelson, 1993). Although the density of faults decreases westward, Oiltana and Brush Creek domes contain nearly as many faults as West, Mosby and East domes. According to Perry's model, Oiltana and Brush Creek domes should yield oil. Second (as cited above), Hadley (1956, p. 99) showed that en echelon faults in the Cat Creek oil field do not block migration of fluids.

I propose that pre-Laramide deformation was a key to oil accumulation. Many large anticlines and domes in central Montana have yielded only dry holes. Among them are Alice dome, Freedom dome and the giant Porcupine dome southeast of the study area. Fanshawe (1978 and 1985) proposed that those three domes were structural depressions before the Laramide orogeny. Hence, any oil migrated away before the traps were formed. My work confirms Fanshawe's hypothesis of structural inversion at Alice and Freedom domes. Extending Fanshawe's model to the Cat Creek anticline, my mapping shows that no trap existed west of West dome before Laramide deformation. On the western part of the Cat Creek anticline strata below the sub-Zuni unconformity dip northward, toward the Cat Creek fault (Plate 1, fig. 57). Any oil generated in sub-Zuni strata could have escaped up-dip toward the south before the Cat Creek anticline formed. However, at East, Mosby and West domes the sub-Zuni strata dip away from the Cat Creek fault. Oil generated here thus migrated toward the fault. Assuming the fault formed a seal, this oil was trapped. When the domes developed during the Laramide orogeny, this oil re-migrated up-dip and along fractures into its present position in Alaska Bench through Fall River reservoir rocks.

Gumbo Ridge area

Oil fields in the Gumbo Ridge area (south-central part of study area) include Gumbo Ridge, Breed Creek, Studhorse, Rattler Butte, and Bean Blossom (fig. 56). Oil in these fields is produced from lenticular channel-form sandstones in the lower part of the Tyler Formation. Productive sandstones are in direct contact or close proximity to the probable source rock, the Cox Ranch Oil Shale Bed in the Heath Formation. Because Tyler sandstone bodies are lenticular and are enclosed in impermeable shales, oil in these sandstones could not escape during Laramide deformation.

Krantzler (1966) observed that Tyler production in central Montana is generally restricted to areas where reservoir sandstones are directly in contact with source rocks in the Heath Formation. Krantzler's model is further verified by the absence of Tyler production in the Cat Creek field, where Tyler sandstones are not in close proximity to the Cox Ranch Bed.

Hagar and Goodman (1985) published two reflection seismic profiles from the Gumbo Ridge oil field to illustrate how Tyler plays can be identified through the use of seismic data.

Rattlesnake Butte field

The Rattlesnake Butte oil field (fig. 56) is situated in a small area of structural closure along an anticline within the Rattlesnake Butte fault zone. Oil is produced from the dolomitized upper portion of the Alaska Bench/Devil's Pocket (Amsden) Formations directly beneath the pre-Zuni unconformity. Porosity probably developed as a result of pre-Zuni subaerial weathering (Maughan, 1989). Laramide left-lateral wrenching along the Rattlesnake Butte fault zone provided fracture pathways for oil to migrate upward from presumed source rocks in the

Heath to reservoir rocks in the Amsden. The anticline also is a product of Laramide wrenching.

Summary

Successful oil exploration in central Montana depends on locating areas where traps existed continuously since oil was generated. The Cat Creek oil field owes its existence to fortuitous coincidence of pre-Laramide and Laramide structural traps at the same site. Traps in the Gumbo Ridge area are stratigraphic, and were not affected by structural reversals. The Rattlesnake Butte field also appears to contain stratigraphic traps not affected by structural movement. Production from the Tyler Formation in large fields south of the study area also taps stratigraphic traps in an area where structure was inverted (Cooper, 1956; Sonnenberg, 1956; Fanshawe, 1978 and 1985).

The explorationist in central Montana should seek areas where (1) no structural reversals have occurred, or (2) pre-Laramide and Laramide structural traps coincide, or (3) traps are not of structural origin. Similar guidelines apply in other parts of the North American craton where structural reversals have taken place.

CONCLUSIONS

High-angle faults in cratonic settings are prone to repeated reactivation under different stress regimes. Faults are susceptible to reactivation because fractured, crushed rock is weaker than surrounding intact rock. Each additional episode of displacement pulverizes more rock, setting the stage for the next episode. The Cat Creek fault is an example of a fault that underwent normal, reverse, and oblique-slip displacement under a succession of stress fields.

Intracratonic stress fields can affect large regions and rejuvenate faults having a variety of orientations. For example, the Late Devonian event caused movement along the WNW-trending Cat Creek fault, east-trending Gumbo Ridge fault, ENE-trending Weldon fault and NNW-trending Cedar Creek fault. Such diverse orientations do not readily fit tectonic models calling for conjugate fault sets formed under dominantly horizontal stresses. Specifically, the findings of this study call into question tectonic models such as those of Stone (1969), Thomas (1974), Brown (1978), and Brown and Brown (1987), which divide the craton into mosaics of diamond-shaped blocks outlined by lengthy shear zones or "lineaments". Although some faults in the study area coincide with lineaments defined by the authors named above, other faults do not fit these models or conform to a conjugate system. This study suggests that most of the faults in the study area were formed during a Proterozoic rifting event and later reactivated under a variety of stress regimes.

The relationship of intracratonic faulting to continental-margin tectonism also poses problems. Some major events in central and eastern Montana (Cambrian reverse faulting, post-Pennsylvanian/pre-Jurassic normal faulting) relate to no known marginal tectonism. For other events, the time correspondence varies widely. Late Devonian reverse faulting in Montana coincides with the climax of the Antler orogeny far to the west. However, the intense Sevier orogeny, which was closer to central Montana than the Antler orogeny, induced no significant

intracratonic faulting for at least 80 million years. I make no attempt to explain these discrepancies here, but will leave them to inspire further study.

Whatever may be its cause, recurrent tectonism concerns the petroleum geologist. Many of the most promising Laramide structural highs in central Montana were structural lows at the critical time of hydrocarbon migration. Successful exploration here, and in other regions subjected to recurrent tectonism, requires identifying traps that have survived all tectonic activity that occurred after the emplacement of hydrocarbons.

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FIGURES

FIGURE CAPTIONS

Fig. 1. Location map, showing study area and major structural features of Montana.

Fig. 2. Structural features of study area. Faults are drawn as they are inferred to exist at top of Precambrian crystalline basement, with symbols indicating sense of most recent displacement.

Fig. 3. Belt basin and Montana aulacogen, showing surface and subsurface control points.

Fig. 4. Stratigraphic column.

Fig. 5. Thickness of Sauk Sequence and presence of Belt Supergroup.

Fig. 6. Log section near Bear Creek dome, illustrating reverse displacement of Cat Creek fault during deposition of Sauk Sequence. Flathead Sandstone and lower part of Emerson Formation are greatly thinned in the well southwest of the fault and thickened in the well northeast of the fault. Line of section shown in fig. 5.

Fig. 7. Log section across Gumbo Ridge fault indicating Paleozoic displacement. Abnormally thin Emerson Formation south of the fault suggests Late Cambrian or Early Ordovician uplift of the southern block. Additional post-Tippecanoe, pre-upper Kaskaskia uplift is evident from truncation of these rocks beneath upper Kaskaskia Sequence south of the fault. Line of section shown in fig. 5.

Fig. 8. Reference log for Tippecanoe Sequence: Gas Producing Enterprises #1 Burlington-Northern, SE $\frac{1}{4}$ NE $\frac{1}{4}$, Sec. 5, T15N, R40E.

Fig. 9. Thickness of Tippecanoe Sequence.

Fig. 10. Reference log for lower Kaskaskia Sequence: Pubco-Empire #16-12 State, NW $\frac{1}{4}$ SW $\frac{1}{4}$, Sec. 16, T18N, R33E.

Fig. 11. Thickness of lower Kaskaskia Sequence and extent of Souris River, Birdbear and Three Forks Formations.

Fig. 12. Log section across Cat Creek fault near Bear Creek dome, Sec. 17, T17N, R24E (see fig. 11 for line of section). Southwest side of fault was uplifted after deposition of the Three Forks Formation (Upper Devonian) and before deposition of the Lodgepole Limestone (Lower Mississippian). Dashed lines in Lodgepole represent geophysical log markers correlated across the fault. The lower part of the Lodgepole thins slightly on the upthrown side of the fault.

Fig. 13. Log section across Cat Creek fault near Oiltana dome (see fig. 11 for line of section). The southwest side of the fault was uplifted, and the entire lower Kaskaskia Sequence and upper part of the Tippecanoe Sequence were eroded prior to deposition of the Lodgepole Limestone. As in fig. 12, basal Lodgepole strata thin slightly onto the upthrown block.

Fig. 14. North-south cross section in eastern part of study area (see fig. 11 for line of section). Souris River Formation (Upper Devonian) onlaps erosion surface at top of Tippecanoe Sequence toward the south. After deposition of the Three Forks Formation the region was tilted northward, and Devonian strata were bevelled by erosion prior to Lodgepole deposition. The crustal block south of the Gumbo Ridge fault was upthrown during the same time (very late Devonian to earliest Mississippian).

Fig. 15. Reference log for Madison Group: Arco #1-12 Fitch, NW $\frac{1}{4}$ NE $\frac{1}{4}$, Sec. 12, T15N, R35E. Small x symbols represent widely traceable geophysical-log markers.

Fig. 16. Thickness of Lodgepole Limestone.

Fig. 17. Thickness of Mission Canyon Formation.

Fig. 18. Thickness of Charles Formation.

- Fig. 19.** Reference log for Big Snowy and Amsden Groups: Cenex #5-26 U.S.A. Madison, SW $\frac{1}{4}$ NW $\frac{1}{4}$, Sec. 26, T15N, R30E, on Mosby dome. Members and beds in upper-case letters are formal; those in lower-case letters are informal. Letters in quotation marks designate widespread geophysical-log markers.
- Fig. 20.** Thickness of Kibbey Formation.
- Fig. 21.** Thickness of Otter Formation.
- Fig. 22.** East-west log section illustrating lateral variation in Heath and Tyler Formations. Line of section on fig. 24.
- Fig. 23.** North-south log section illustrating lateral variation in Heath and Tyler Formations. Line of section appears on fig. 24.
- Fig. 24.** Thickness of Heath Formation.
- Fig. 25.** Log section comparing Otter Formation in two wells about one mile apart. Missing section in left-hand log indicates a normal fault having about 45 feet of throw.
- Fig. 26.** Thickness of Tyler Formation.
- Fig. 27.** Cross section of Cat Creek fault near Bear Creek dome. The Humble #1 Delaney well, southwest of the fault, encountered a normal section of the Amsden, Big Snowy and Madison Groups below Middle Jurassic strata. The Black Eagle #1 Sharptail well began drilling in the hanging wall of the Cat Creek fault, penetrated the fault, and encountered Jurassic rocks directly overlying the Madison Group in the footwall. This evidence indicates that the block northeast of the fault was uplifted and eroded after Amsden and before Jurassic sedimentation. This displacement was opposite to subsequent Laramide (post-Cretaceous) reverse faulting.
- Fig. 28.** Cross section of Mosby dome and Cat Creek fault. Southwest of the fault Zuni Sequence (Jurassic) rocks rest on the Amsden Group. Northeast of the fault Jurassic rocks overlie the Otter Formation (Big Snowy Group). Normal displacement of the Cat Creek fault, with pre-Zuni erosion of the upthrown northeast block, is thereby indicated.
- Fig. 29.** Reference log for the Ellis Group: Tricentrol U.S. #22-15 Federal, SE $\frac{1}{4}$ SW $\frac{1}{4}$ SE $\frac{1}{4}$, Sec. 22, T15N, R30E.
- Fig. 30.** Thickness of the Tampico Shale Member of the Piper Formation.
- Fig. 31.** Thickness of the Firemoon Limestone Member of the Piper Formation.
- Fig. 32.** Thickness of the Bowes Shale Member of the Piper Formation.
- Fig. 33.** Thickness of the Rierdon Formation.
- Fig. 34.** Thickness of the Swift Formation.
- Fig. 35.** North-south log section of Ellis Group, from north side of Cat Creek fault to Belt Island. Line of section shown on plate 2.
- Fig. 36.** Reference log for Morrison and Kootenai Formations: Utah Southern Oil Co. #23 Rhens-Government, NE $\frac{1}{4}$ SE $\frac{1}{4}$, Sec. 9, T15N, R29E.
- Fig. 37.** Log section showing local development of thick channel-fill sandstone in lower part of Morrison Formation in the Gumbo Ridge area.
- Fig. 38.** Thickness of Morrison and Kootenai Formations (combined).
- Fig. 39.** Reference log for Cretaceous System: Montana Power Co. #12-22 Federal, NW $\frac{1}{4}$ SW $\frac{1}{4}$, Sec. 22, T16N, R29E.
- Fig. 40.** Thickness of Fall River Sandstone.
- Fig. 41.** Thickness of basal siltstone member of Thermopolis Shale.
- Fig. 42.** Thickness of Skull Creek Member of Thermopolis Shale.
- Fig. 43.** Thickness of Cyprian Sandstone Member of Thermopolis Shale.
- Fig. 44.** Thickness of sandy member of Thermopolis Shale.

- Fig. 45.** Thickness of Shell Creek Member of Thermopolis Shale.
- Fig. 46.** Thickness of Mowry Shale.
- Fig. 47.** Thickness of Belle Fourche Shale.
- Fig. 48.** Thickness of Greenhorn Formation.
- Fig. 49.** Thickness of Carlile Shale.
- Fig. 50.** Thickness of Niobrara Shale.
- Fig. 51.** Thickness of Telegraph Creek Formation and Eagle Sandstone (combined)
- Fig. 52.** Thickness of Claggett Shale.
- Fig. 53.** Thickness of Judith River Formation.
- Fig. 54.** Thickness of Bearpaw Shale.
- Fig. 55.** Photograph (looking west) showing a horizontal, undeformed Quaternary gravel terrace truncating steeply dipping Cretaceous strata on the flank of the Cat Creek anticline. The view is of the west bluff of the Musselshell River where it bisects Mosby dome. Cretaceous rocks in the scene range from Telegraph Creek Formation (left) through lower part of Bearpaw Shale (right).
- Fig. 56.** Oil fields in study area.
- Fig. 57.** Diagram illustrating migration paths (arrows) of hydrocarbons being generated during Cretaceous Period prior to formation of Cat Creek anticline. Along the western part of the anticline (upper diagram) pre-Jurassic strata dip northeast, and hydrocarbons escaped up-dip before the anticline formed. In the area of West, Mosby and East domes pre-Jurassic rocks dip southwest, so hydrocarbons moving up-dip were trapped against the Cat Creek fault (lower diagram). During the Laramide orogeny oil re-migrated upward along fractures and bedding planes into suitable reservoir rocks at the crest of the domes.

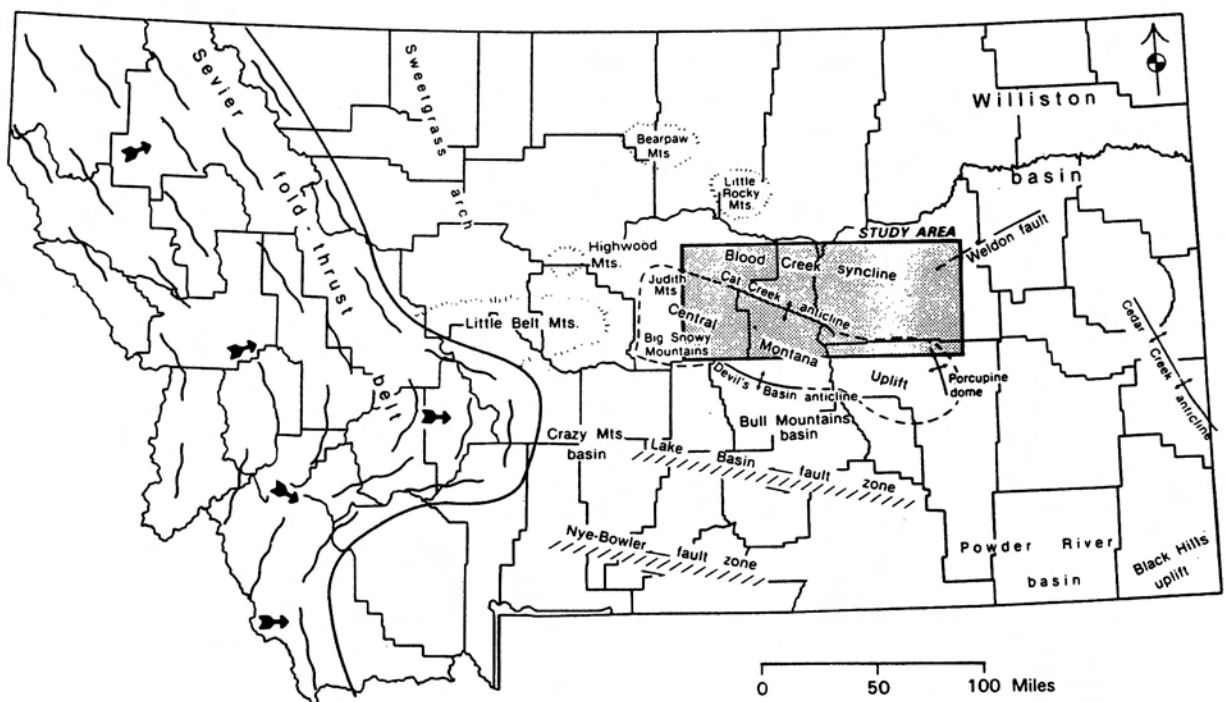


Figure 1—Major structural features of Montana.

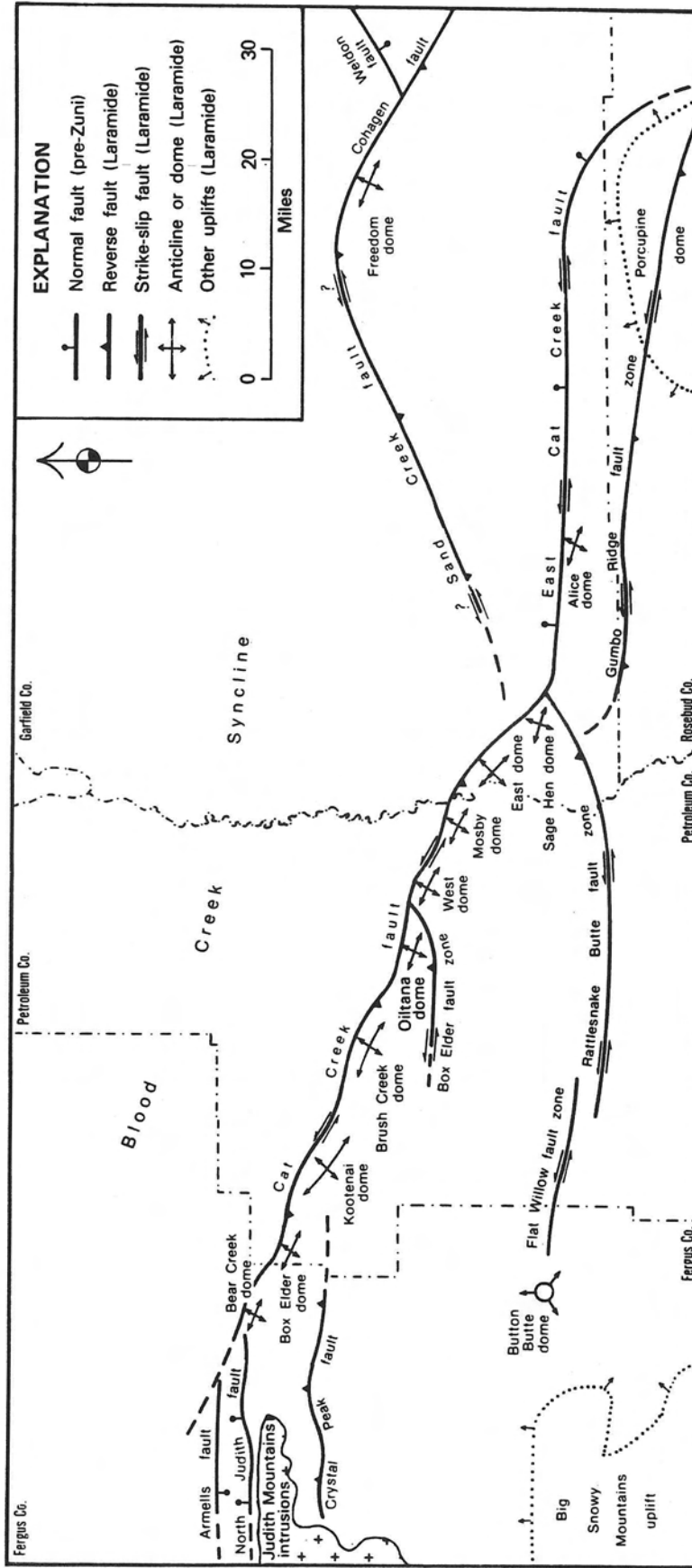


Figure 2—Structural features of the study area. Faults are drawn as inferred to exist at top of Precambrian crystalline basement, with symbols indicating sense of most recent displacement.

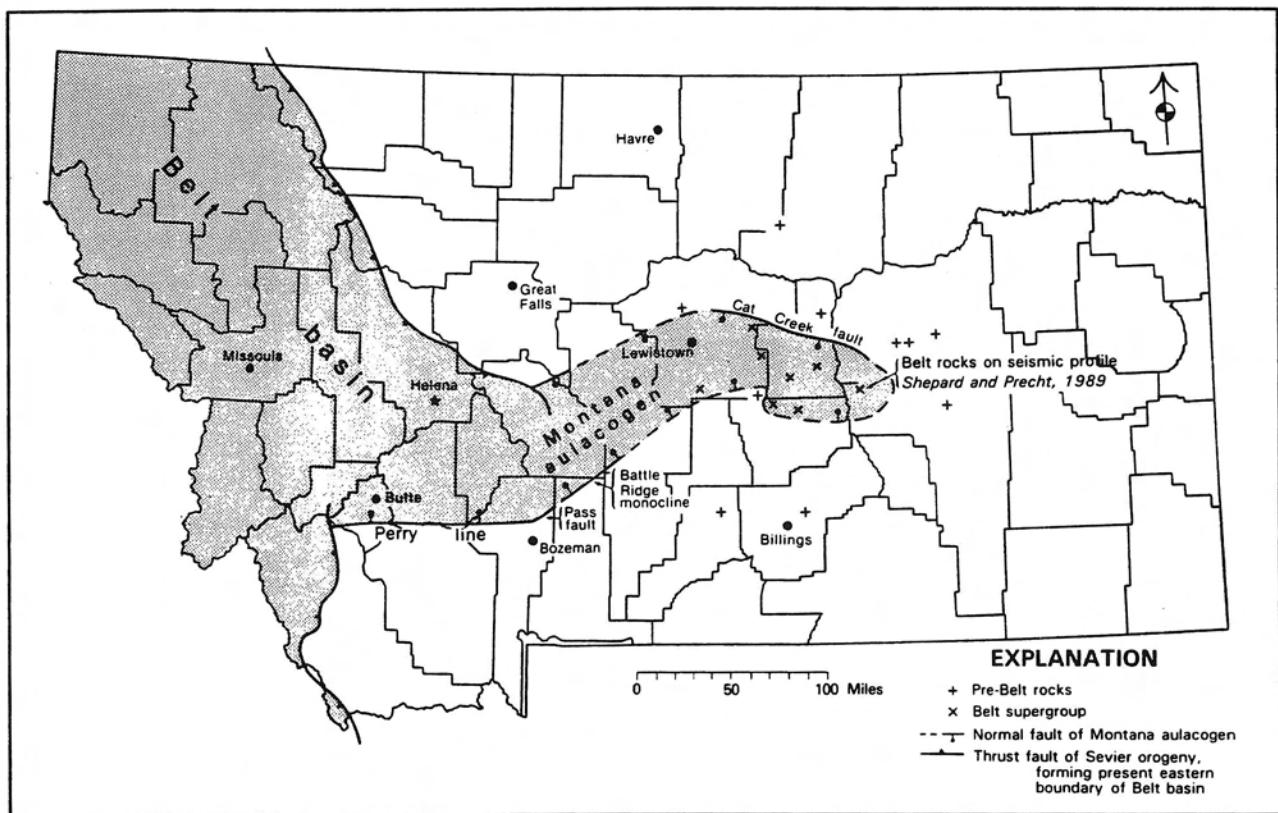
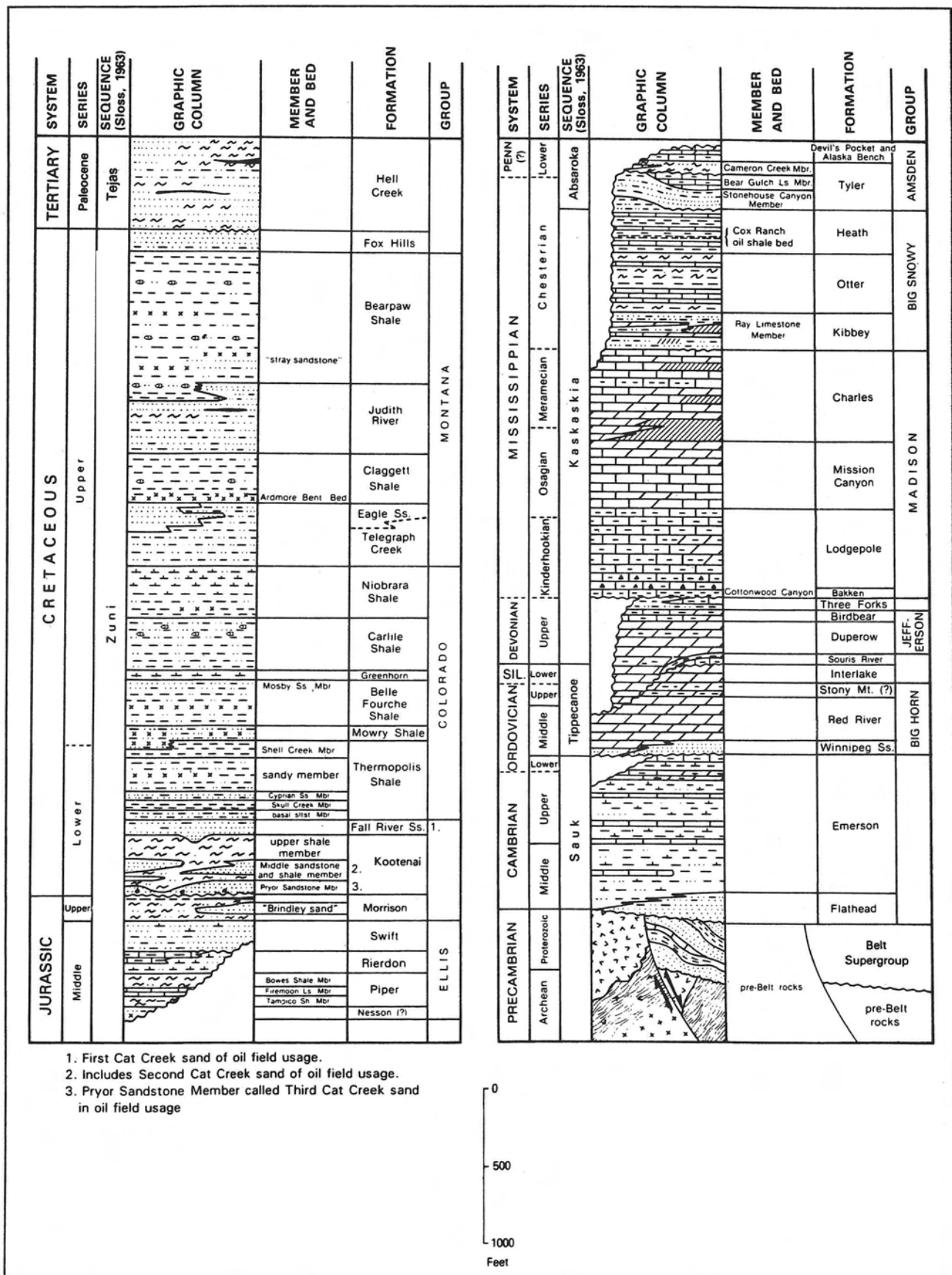


Figure 3—Belt basin and Montana aulacogen, with surface and sub-surface control points.



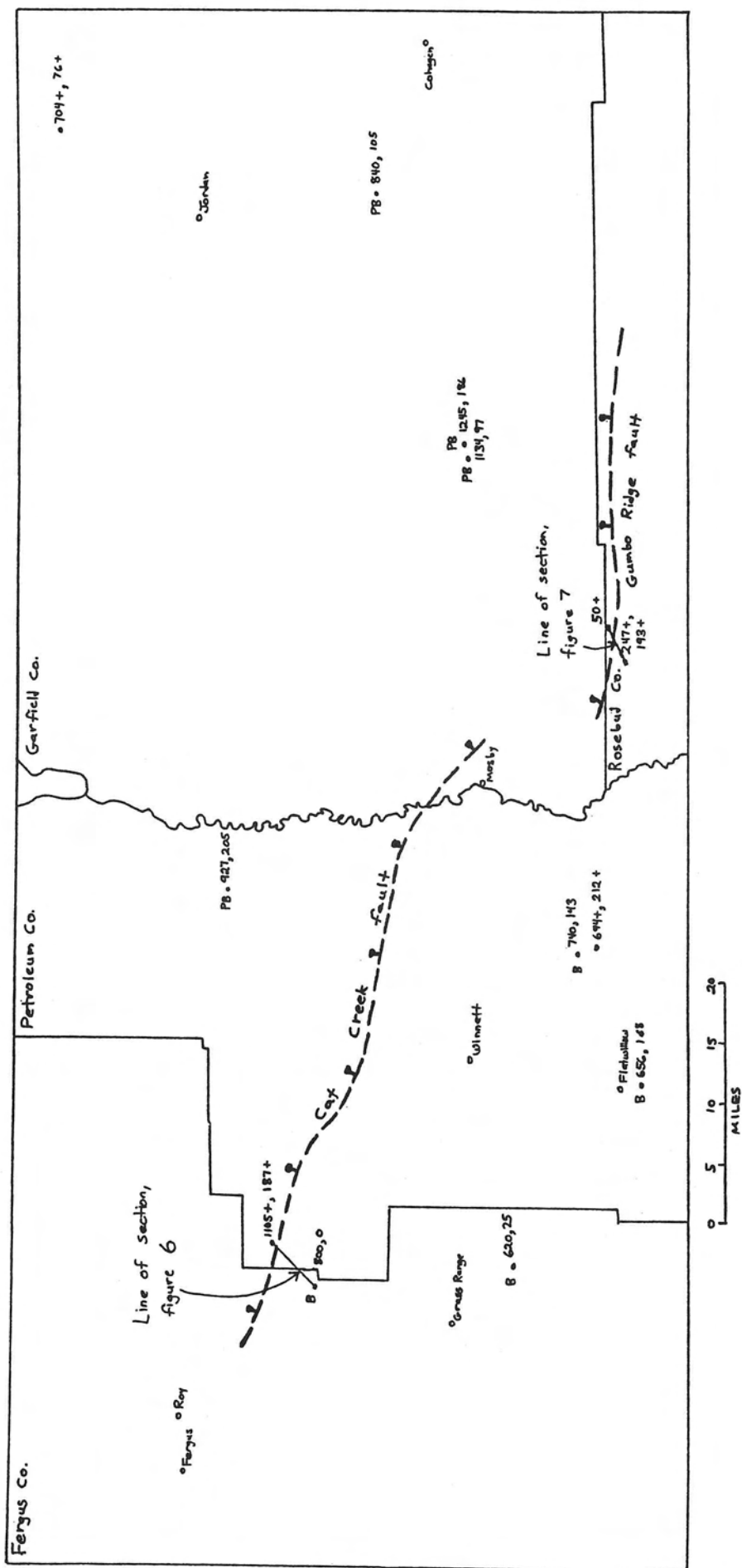


Figure 5. Thickness of Sauk sequence (Flathead Sandstone and Emerson Formation) and Precambrian rocks encountered in deep wells.

- 740, 143 Control well, with thickness of Sauk sequence in feet (first number) and thickness of Flathead Sandstone in feet (second number). B= Belt Supergroup; PB= pre-Belt rocks.

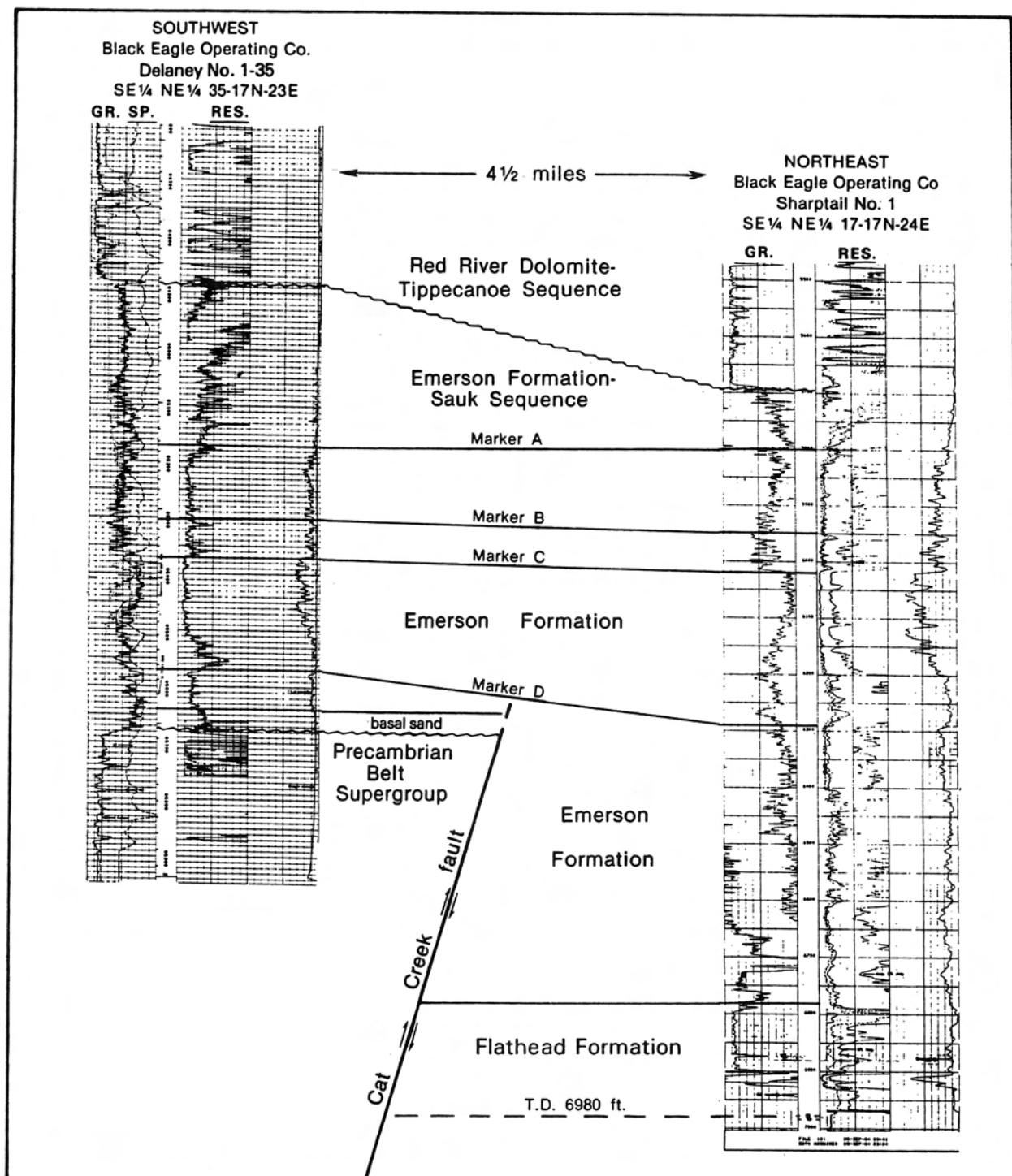


Figure 6—Log section near Bear Creek dome illustrating reverse displacement of Cat creek fault during deposition of Sauk sequence (Upper Cambrian—Lower Ordovician). (See Figure 11 for line of section.) The Flathead Formation and lower part of the Emerson Formation are greatly thinned in well southwest of fault and thickened in well northeast of fault.

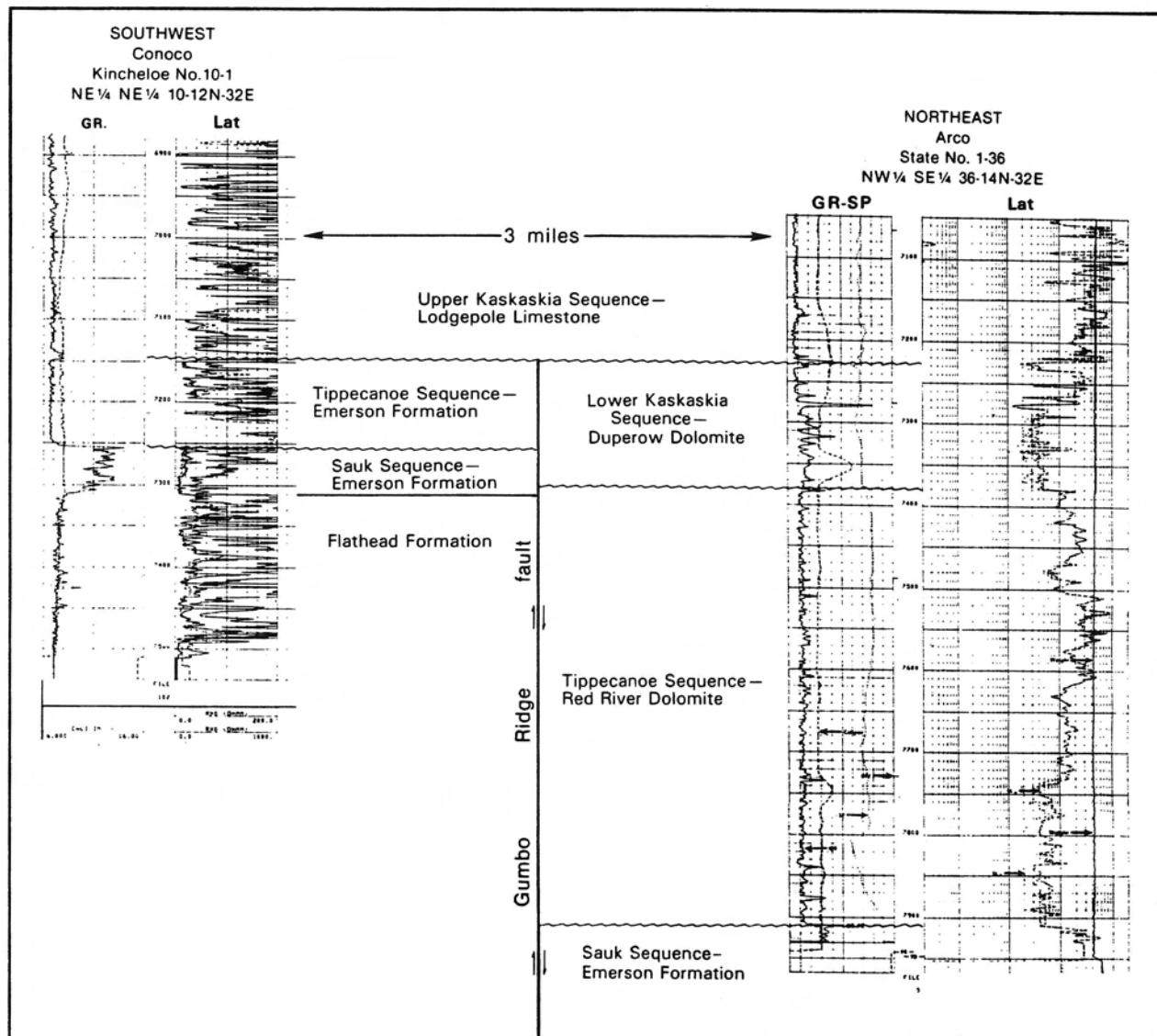


Figure 7—Log section across Gumbo Ridge fault indicating Paleozoic displacement. (See Figure 11 for line of section.) Abnormally thin Emerson Formation (Cambrian) south of fault suggests Late Cambrian or Early Ordovician uplift of southern block. Additional post-Tippecanoe/pre-upper Kaskaskia uplift is also evident from truncation of strata south of fault.

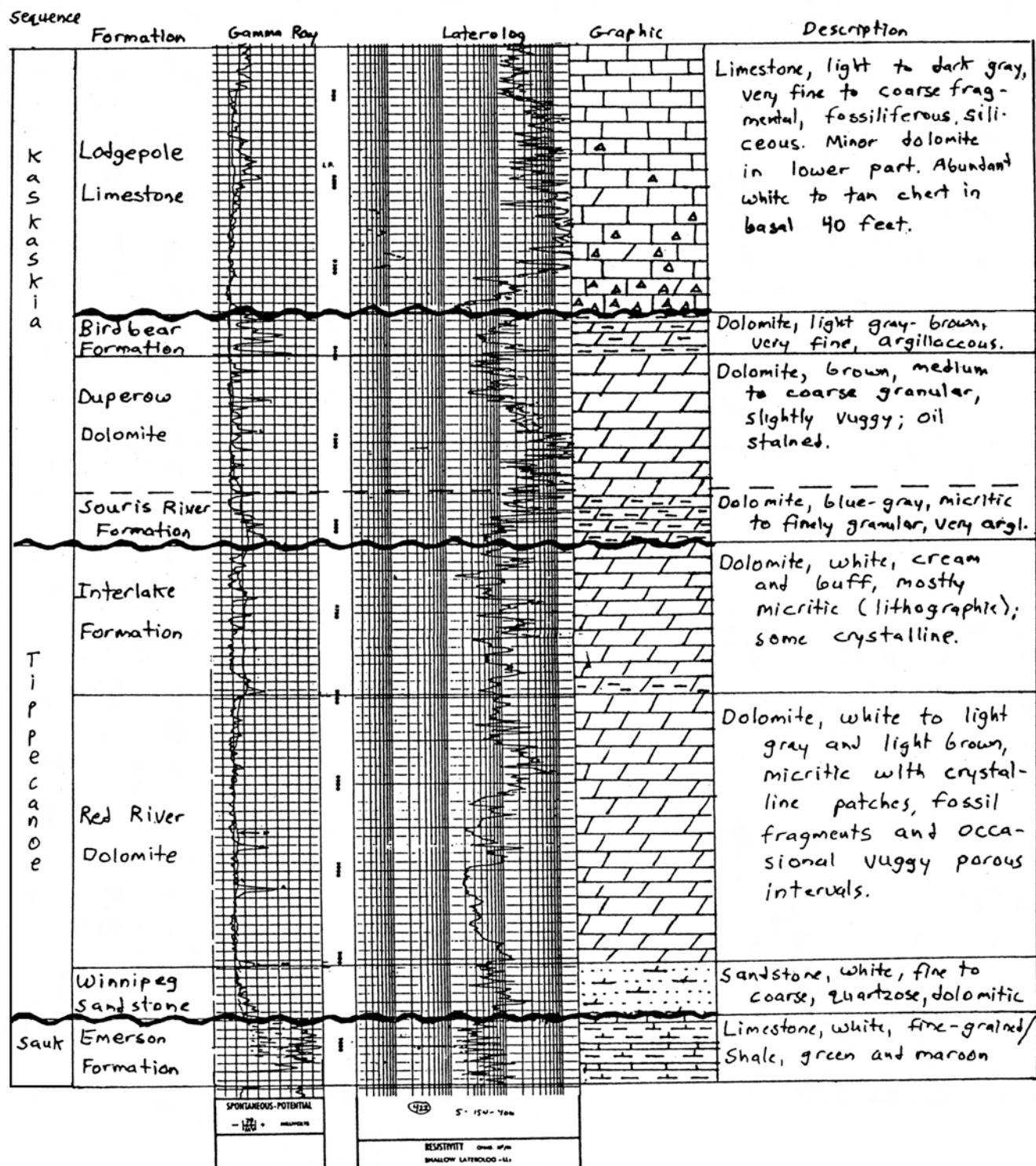


Figure 8. Reference log for Tippecanoe sequence: Gas Producing Enterprises # 1 Burlington-Northern, SE $\frac{1}{4}$ NE $\frac{1}{4}$, Sec. 5, T15N, R40E, Garfield County.

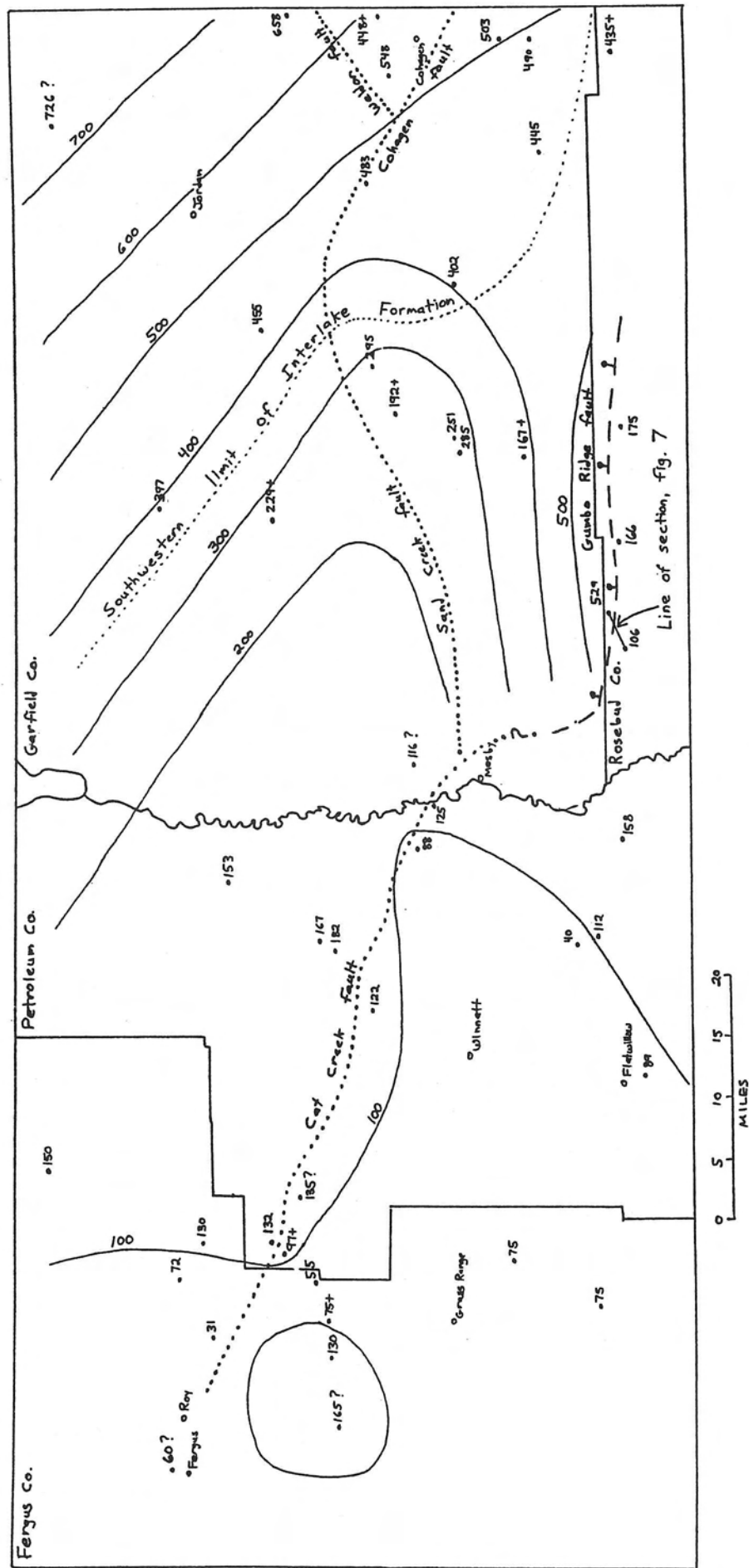


Figure 9. Thickness of Tippecanoe sequence (Winnipeg, Red River, and Interlake Formations).

- 75 Control well, with thickness of Tippecanoe sequence in feet.
- 200 Isopach line; contour interval 100 feet.
- 1-1 Fault active during Tippecanoe interval; ball and bar on downthrown side.
- Inactive fault.

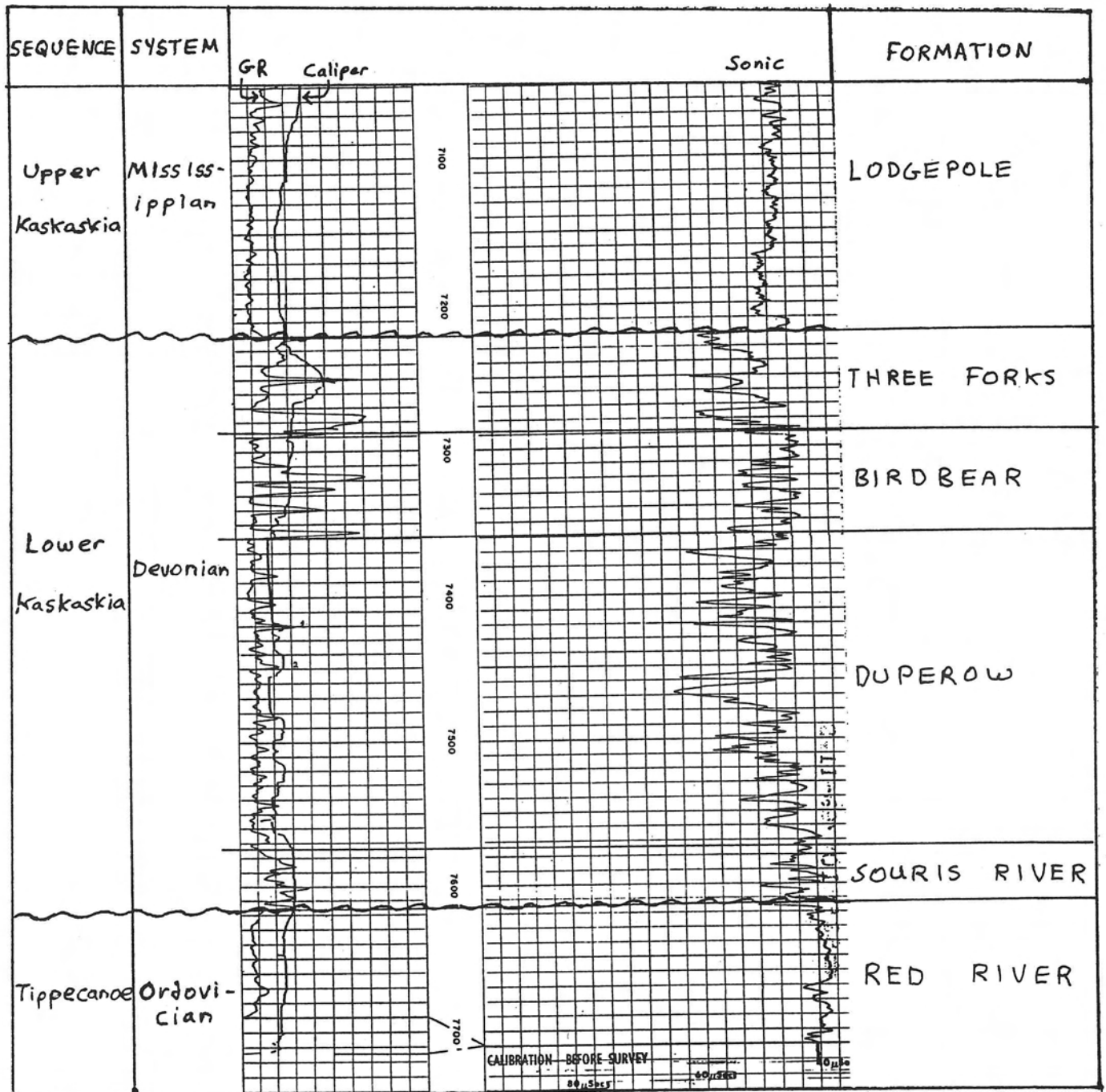


Figure 10. Reference log for Lower Kaskaskia sequence (Devonian System): Pubco-Empire # 16-12 State, NW $\frac{1}{4}$ SW $\frac{1}{4}$, Sec. 16, T18N, R33E, Garfield County.

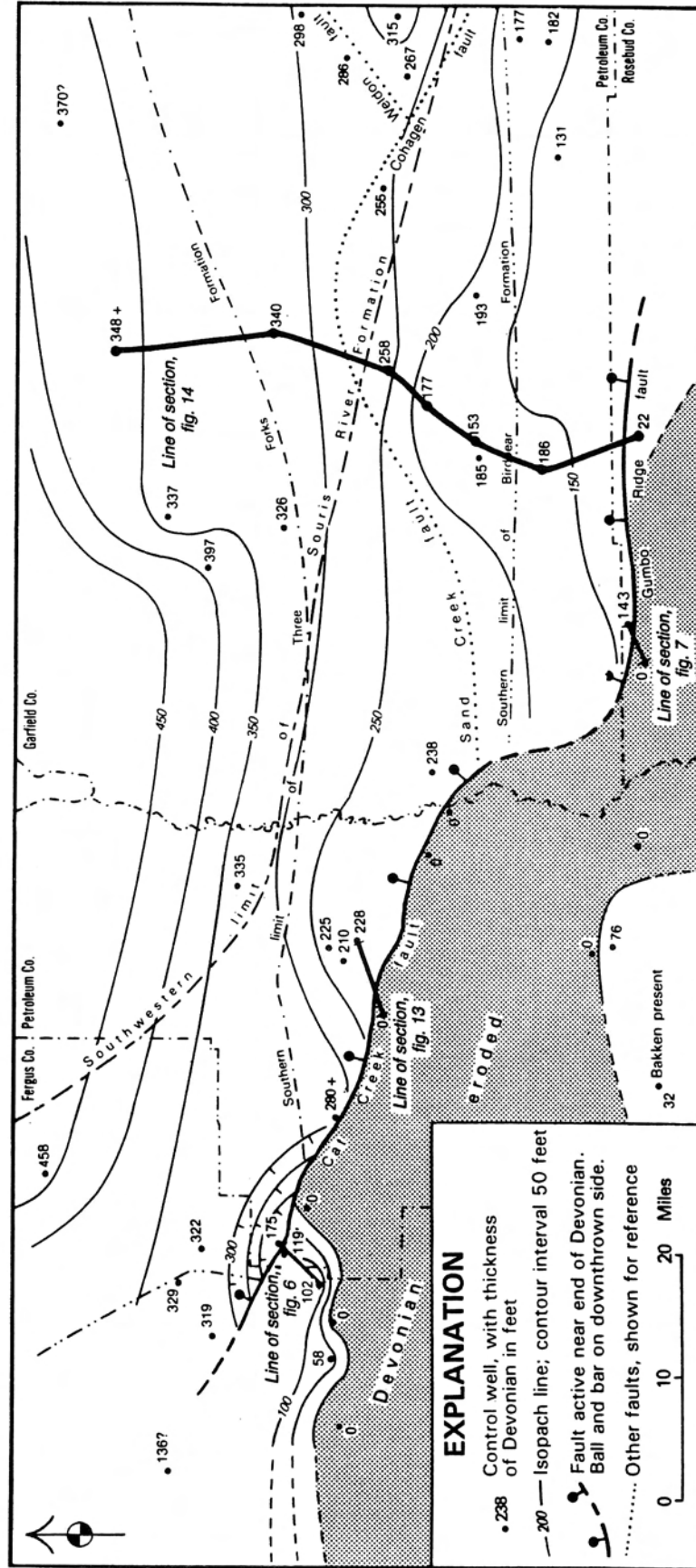


Figure 11—Thickness of lower Kaskaskia sequence (Devonian). Erosional or depositional limits of formations are indicated. The crustal block south of the Gumbo Ridge and Cat Creek faults was uplifted near the end of the Devonian Period. Devonian rocks were largely eroded before Mississippi sedimentation.

Black Eagle #1 Sharptail
SE $\frac{1}{4}$ NE $\frac{1}{4}$, Sec. 17, T 17N, R 24E

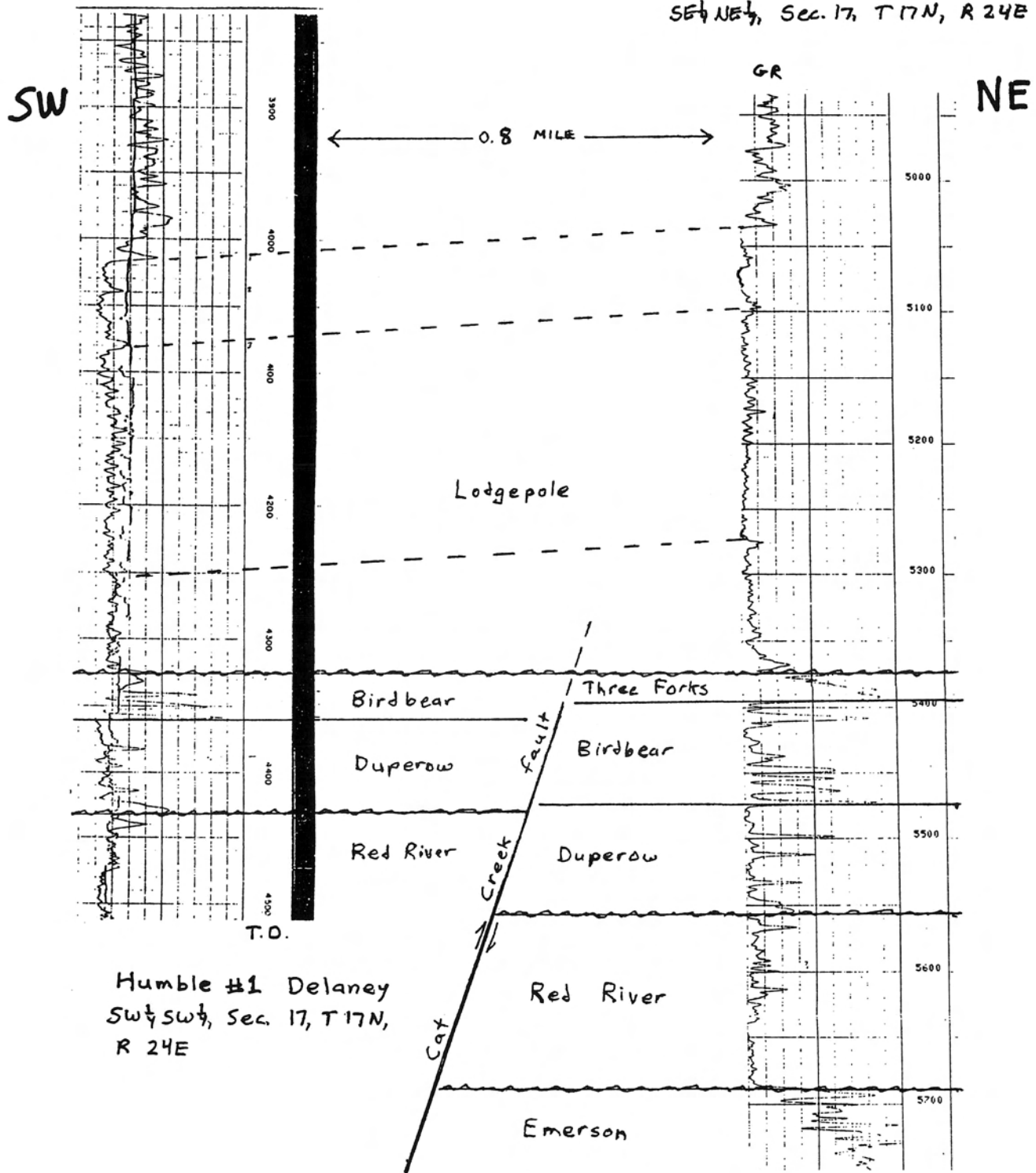


Figure 12. Log section across Cat Creek fault in western Petroleum County, illustrating Late Devonian displacement down to the northeast. Line of section on Plate 1.

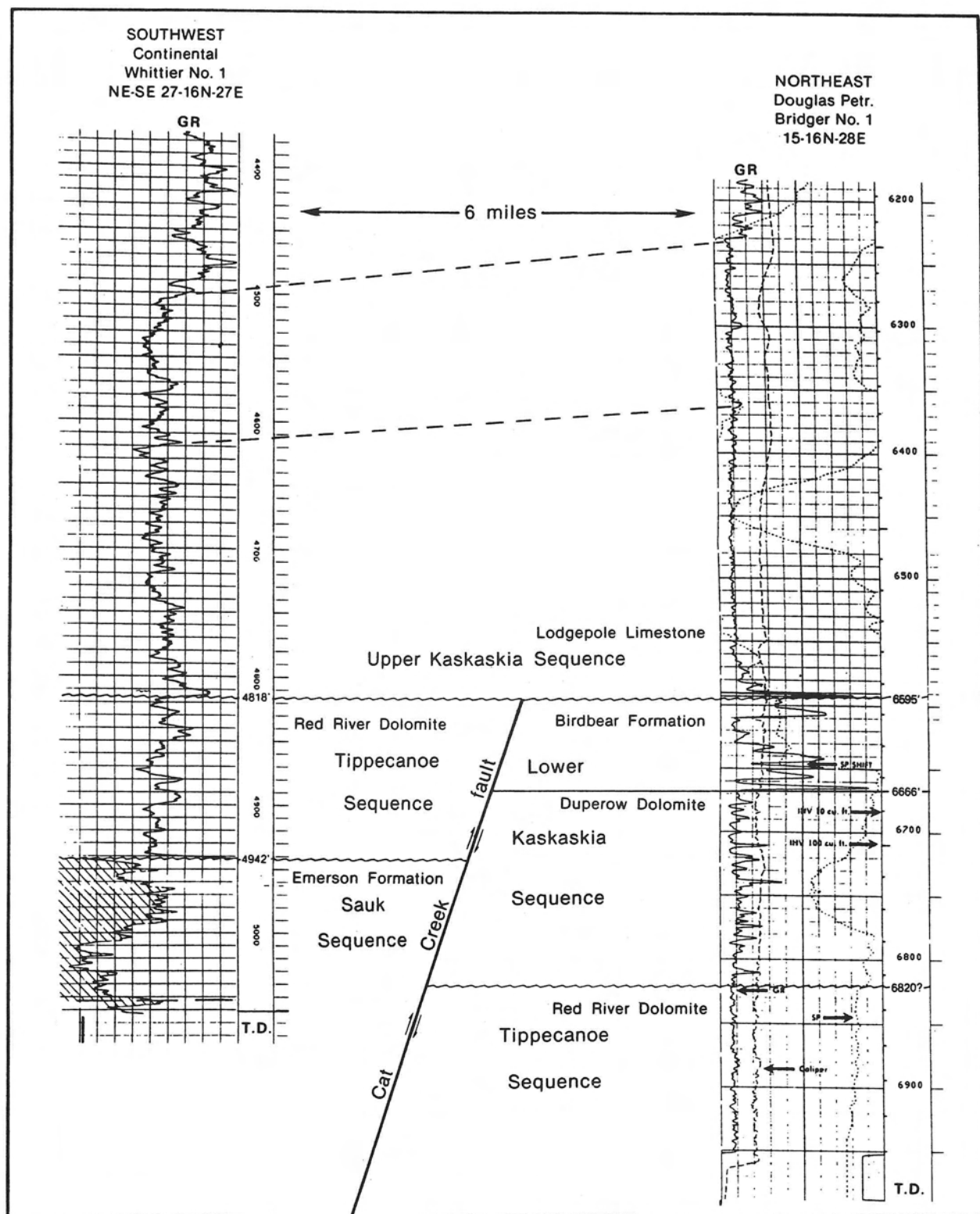


Figure 13—Log section across Cat Creek fault near Oiltana dome. (See Figure 11 for line of section.) Reverse displacement of Cat Creek fault raised the southwestern block; lower Kaskaskia sequence was eroded before deposition of upper Kaskaskia sequence. Timing of fault movement was near the end of the Devonian.

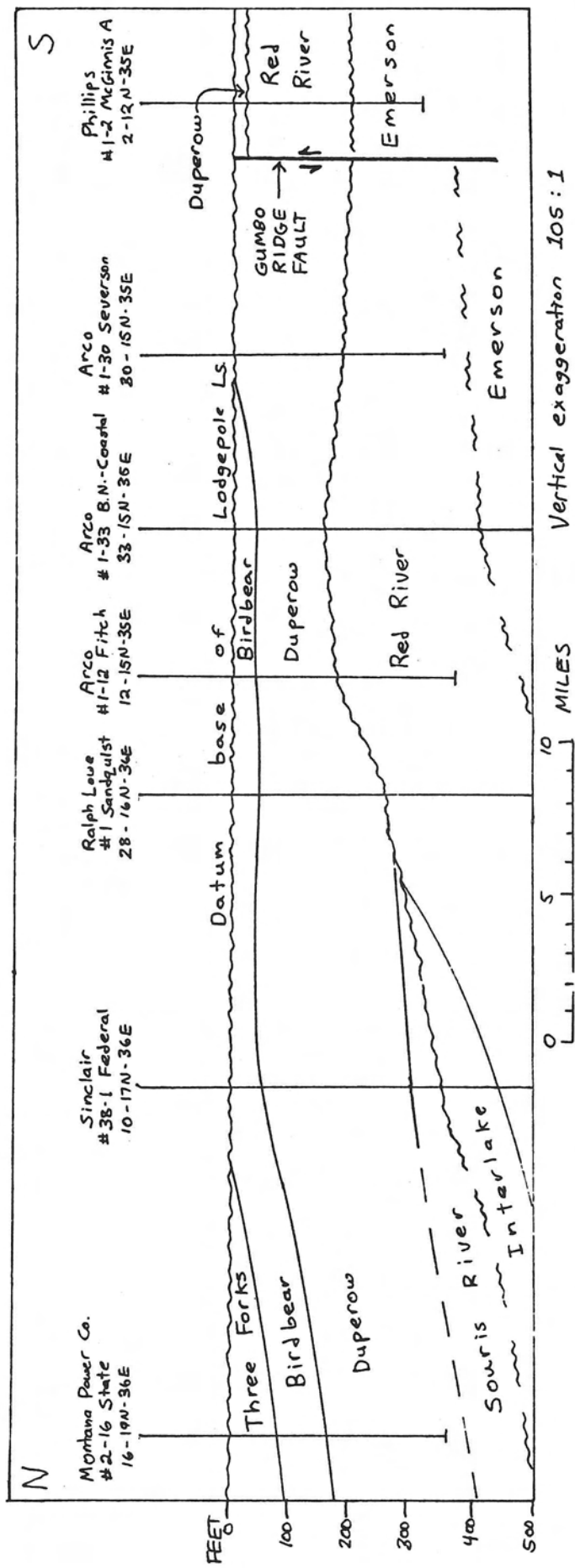


Figure 14. Cross section illustrating southward onlap of Devonian strata onto Tippecanoe sequence, and beveling of Devonian rocks at base of Lodgepole Limestone. Line of section on fig. 11.

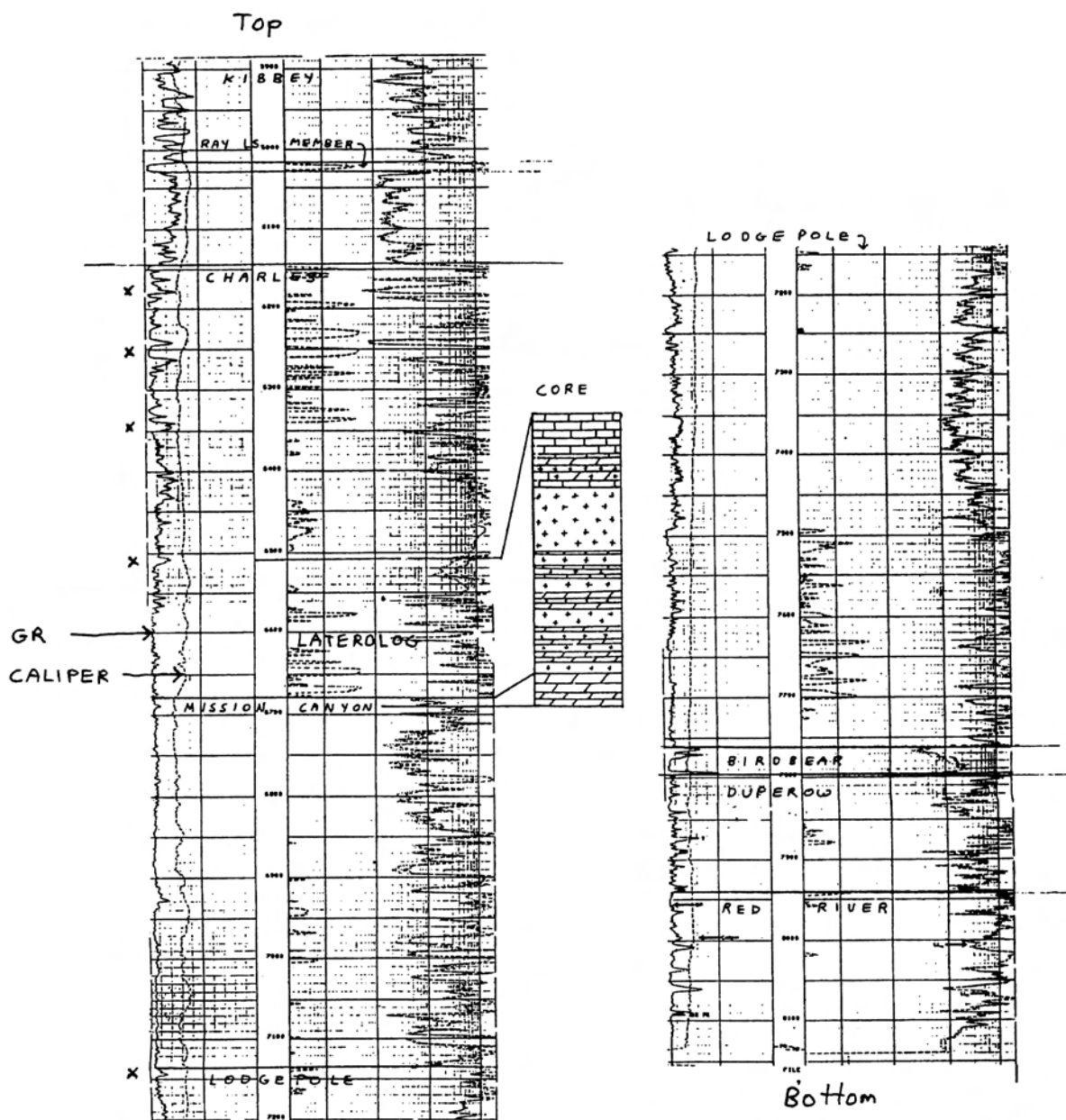


Figure 15. Reference log for Madison Group: Arco # 1-12 Fitch, NW $\frac{1}{4}$ NE $\frac{1}{4}$, Sec. 12, T15N, R35E, Garfield County. X symbols designate widely traceable log-correlation markers.

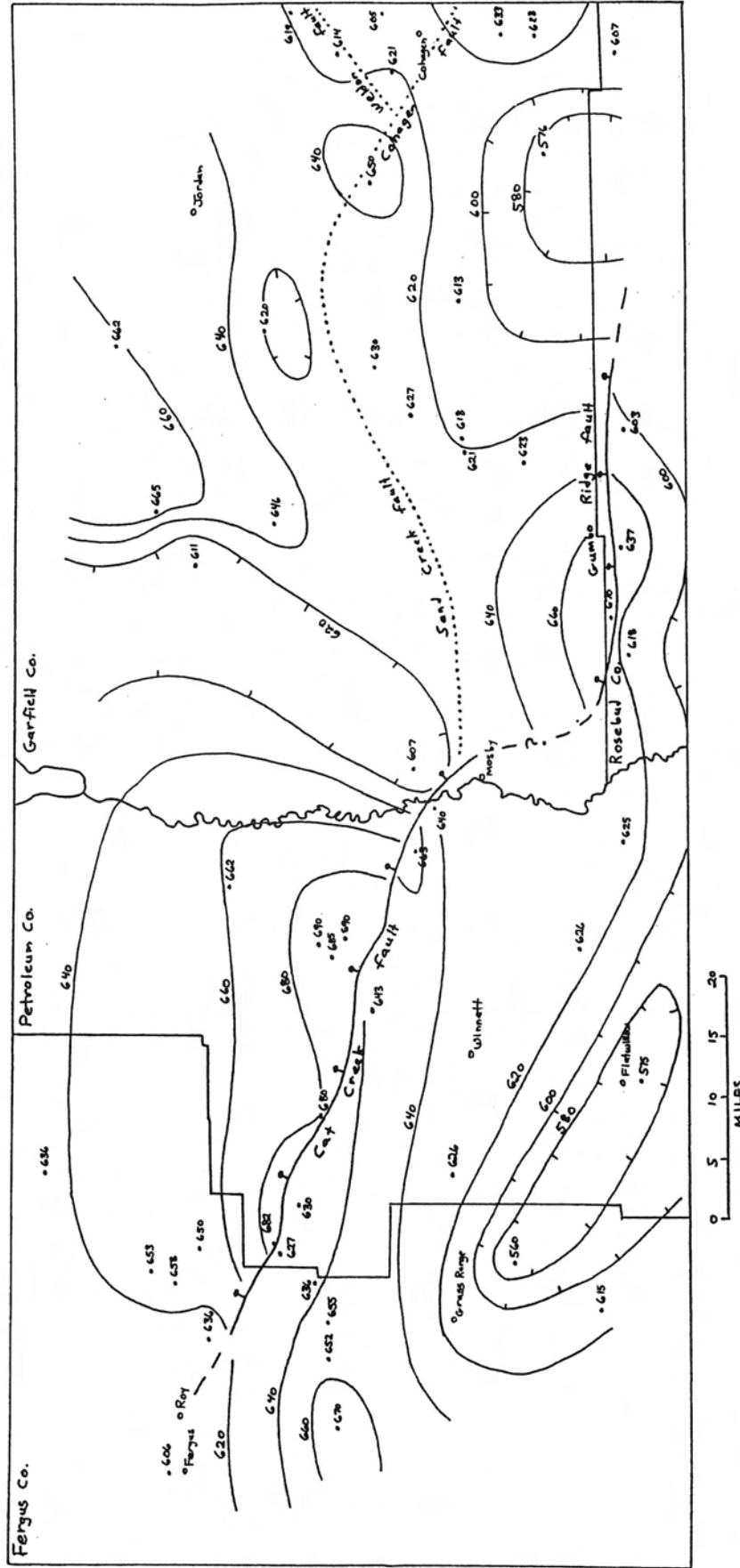


Figure 16. Thickness of Lodgepole Limestone.
 • Control well, with thickness in feet.
 — Isopach line; contour interval 20 feet.
 — Fault active near end of Devonian Period; ball and bar on downthrown side.
 Inactive fault.

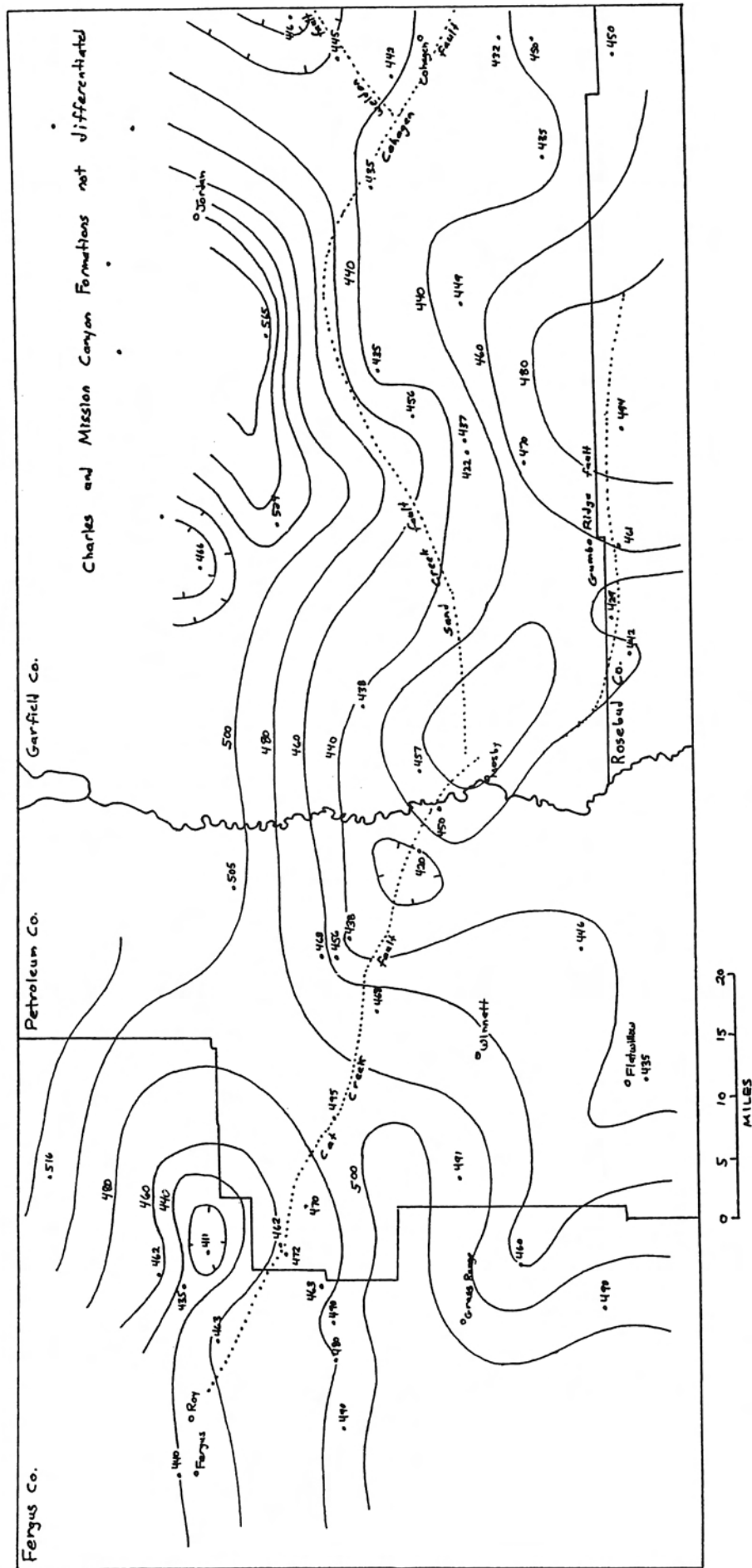
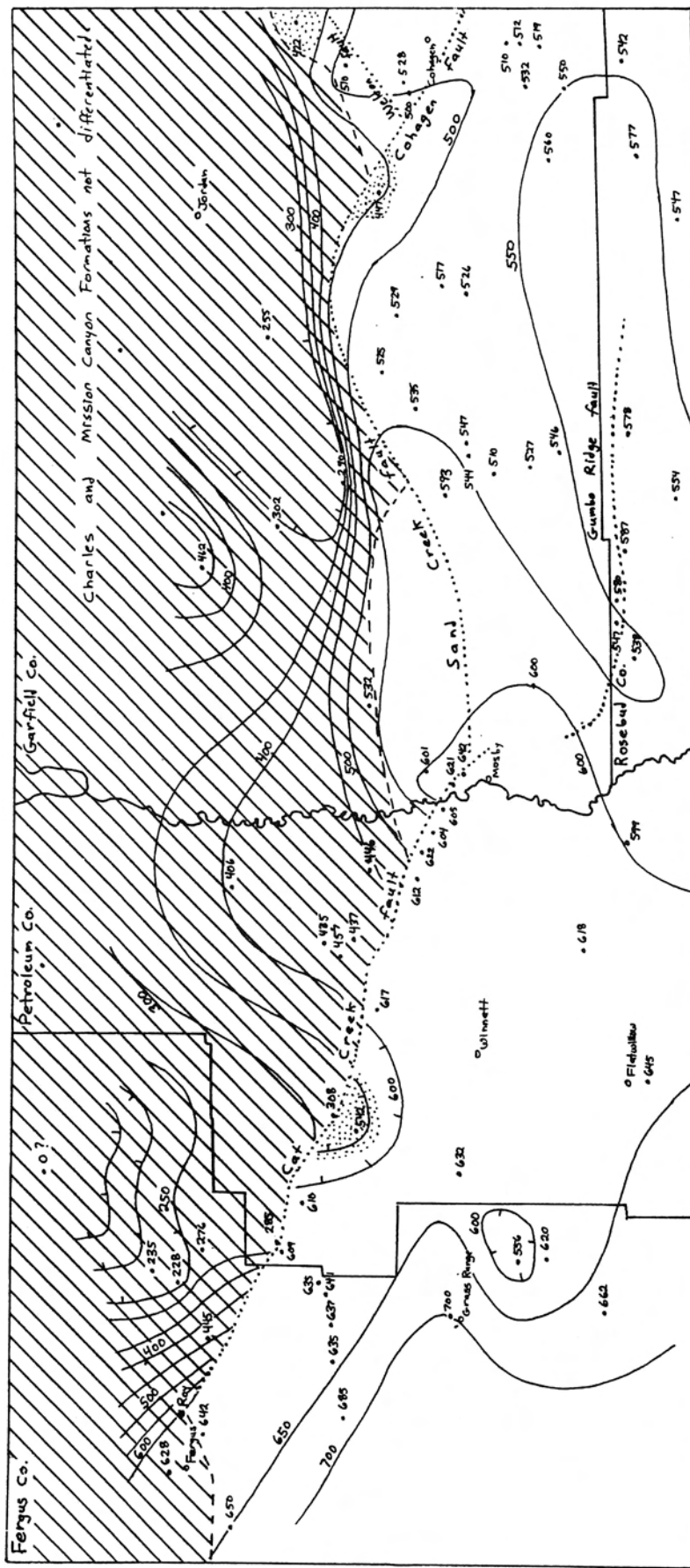


Figure 17. Thickness of Mission Canyon Limestone.
 468 Control well, with thickness in feet.
 480 Isopach line; contour interval 20 feet.
 Inactive fault



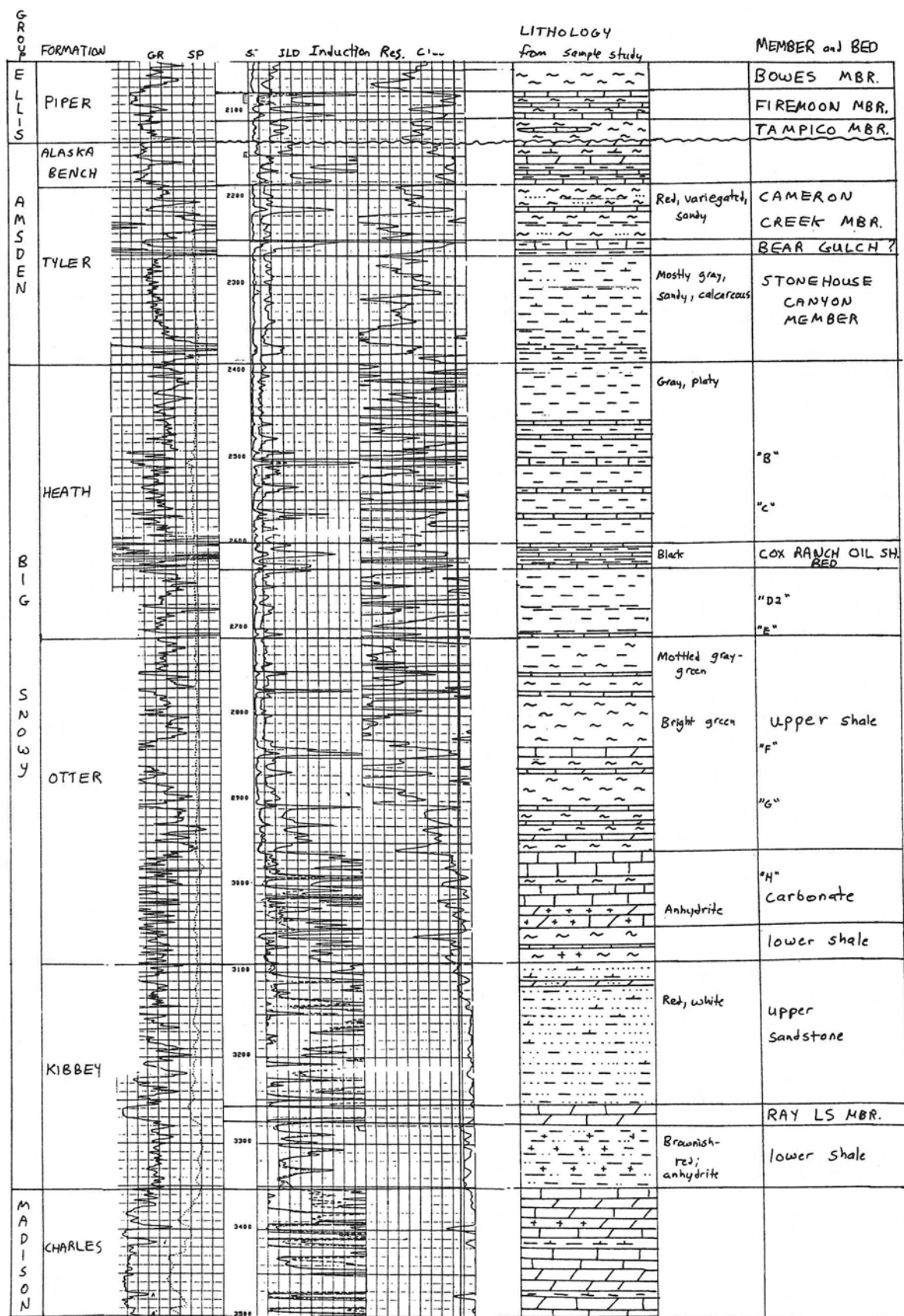


Figure 19. Reference log for Big Snowy and Amsden Groups: Cenex # 5-26 USA Madison, SW $\frac{1}{4}$ NW $\frac{1}{4}$, Sec. 26, T15N, R30E, Garfield County. Members and beds in upper-case are formal, those in lower-case are informal. Letters in quotes, such as "B", designate widely traceable log-correlation markers.

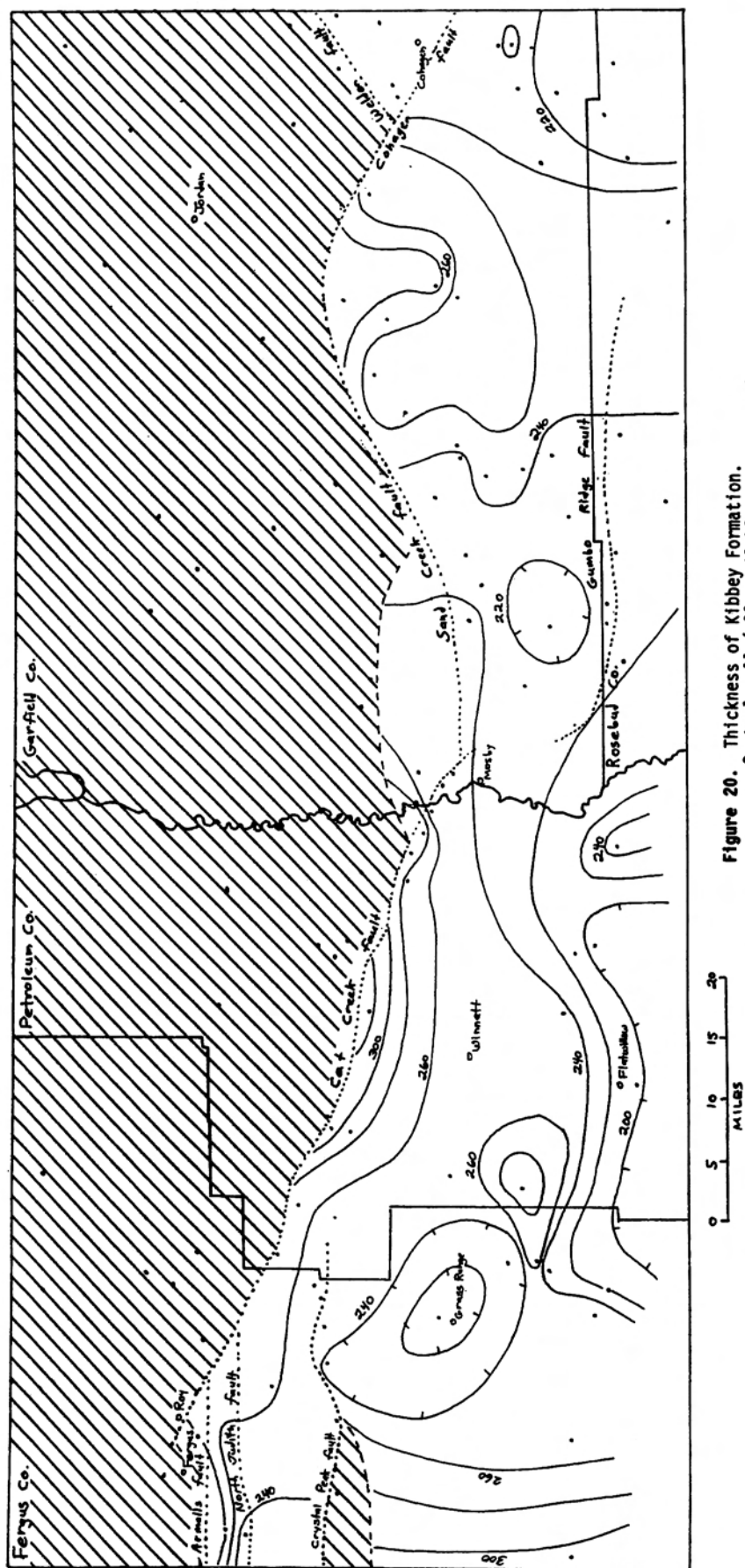


Figure 20. Thickness of Kibbey Formation.

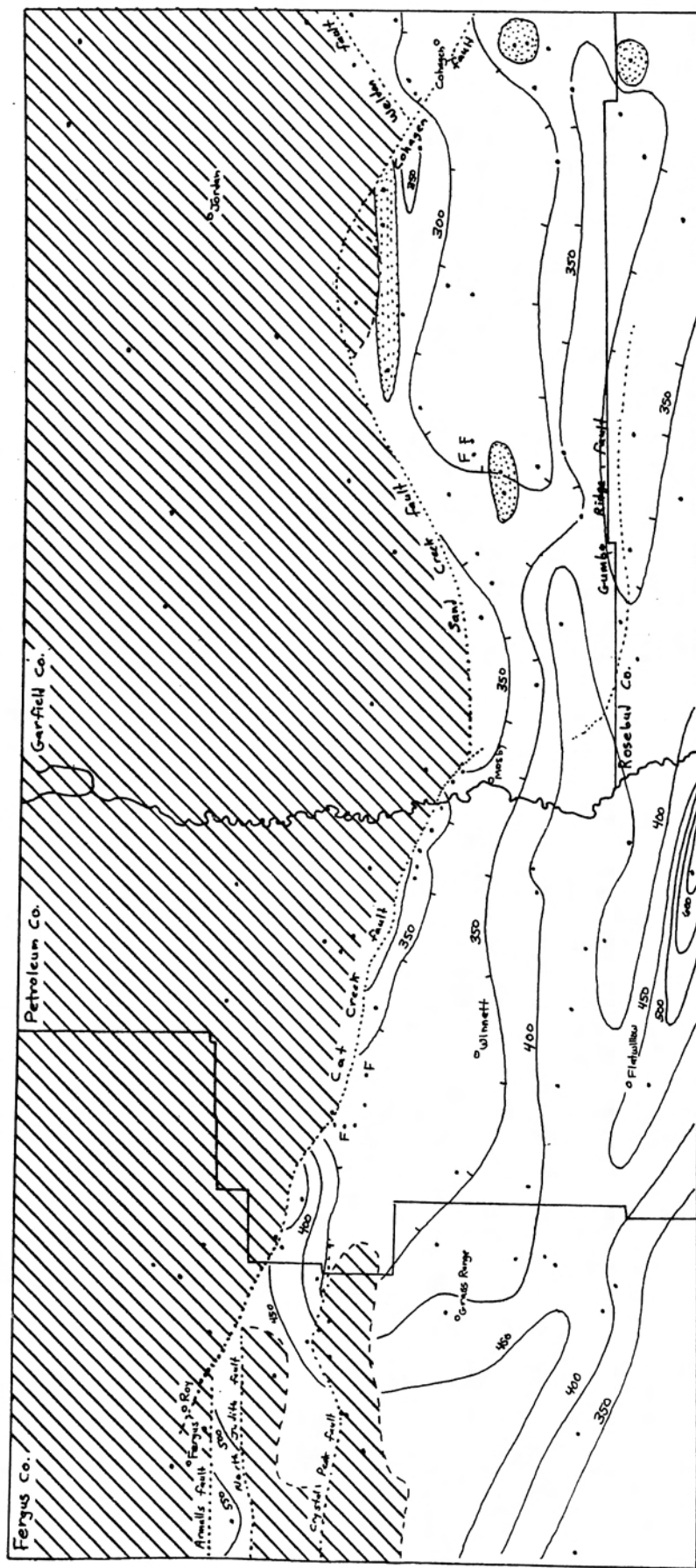


Figure 21. Thickness of Otter Formation.
 • Control well (all available records used).
 --- Isopach line; contour interval 50 feet.
 - - - Fault(s) in Otter.
 ▨ Otter eroded beneath Tyler Formation.
 ▩ Otter eroded beneath Jurassic strata.
 Inactive fault.

MBMG 256

INSERT FIGURE 22 (tabloid)

MBMG 256

INSERT FIGURE 23 (tabloid)

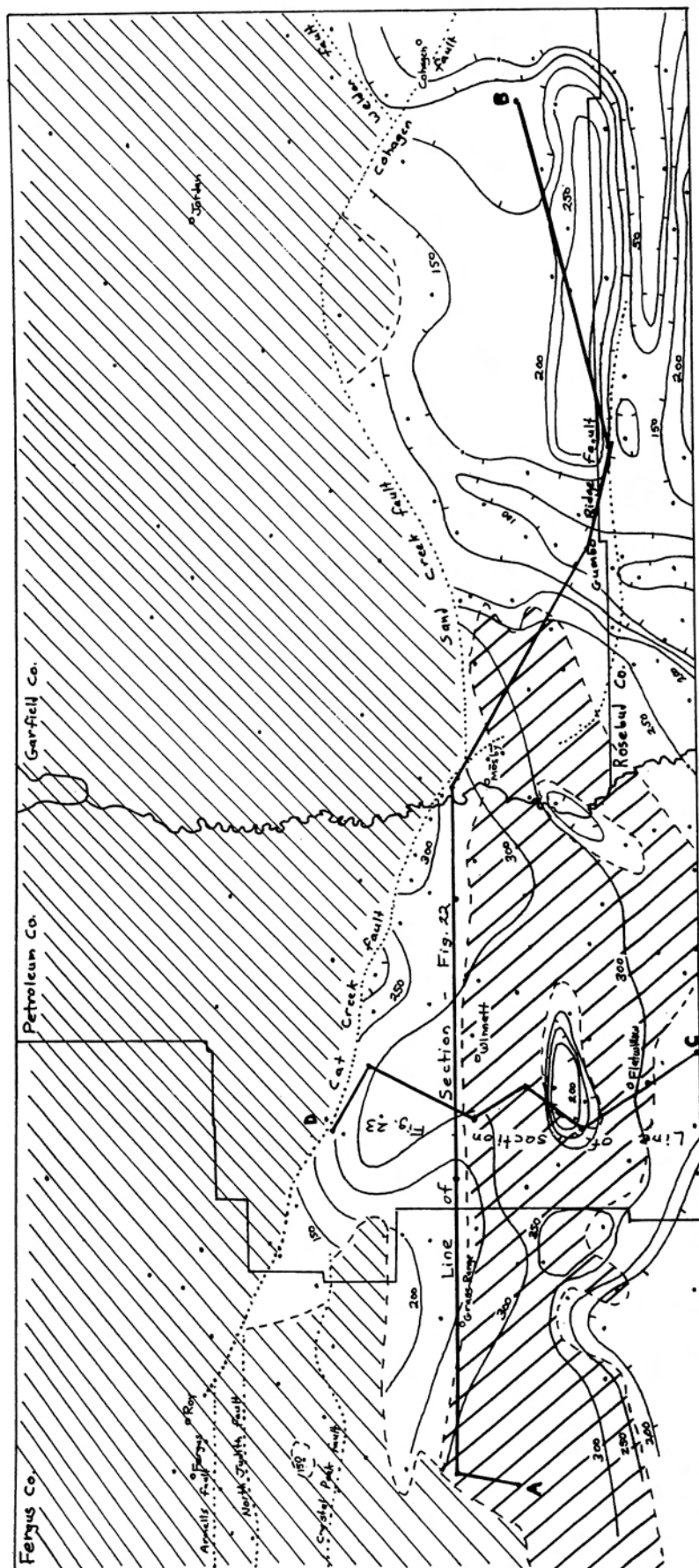


Figure 24. Thickness of Heath Formation.
 • Control well (selected records used).
 --- Isopach line; contour interval 50 feet.
 ~~~~~ Tyler/Heath contact conformable.  
 [Hatched pattern] Heath eroded beneath Jurassic strata.  
 ..... Inactive fault.



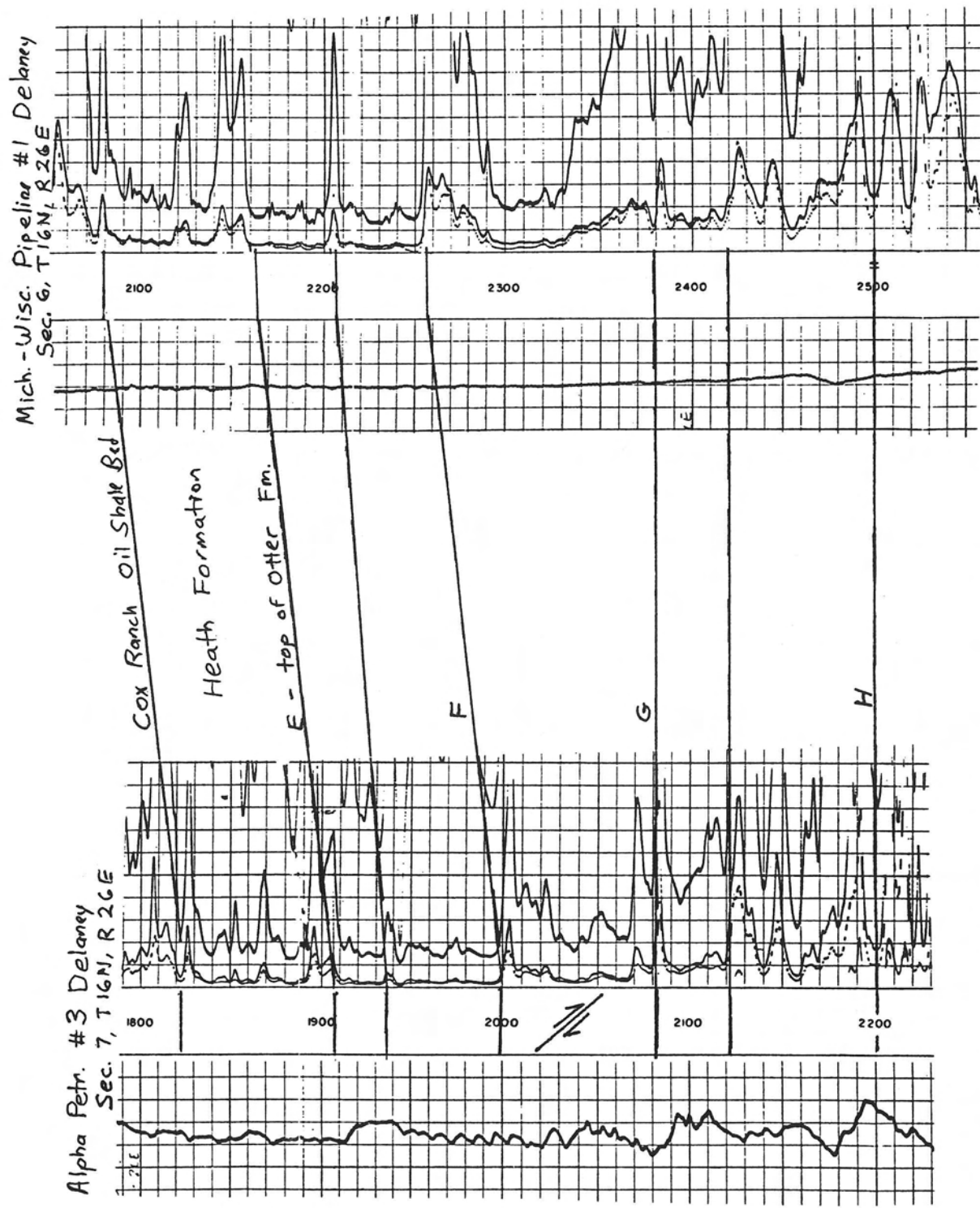


Figure 25. Electric logs of two wells about 1 mile apart, with marker beds correlated. Missing section in left-hand well indicates a normal fault with about 45 feet of throw in the Otter Formation. Line of section on Plate 1.





MBMG 256

INSERT FIGURE 26 (tabloid)



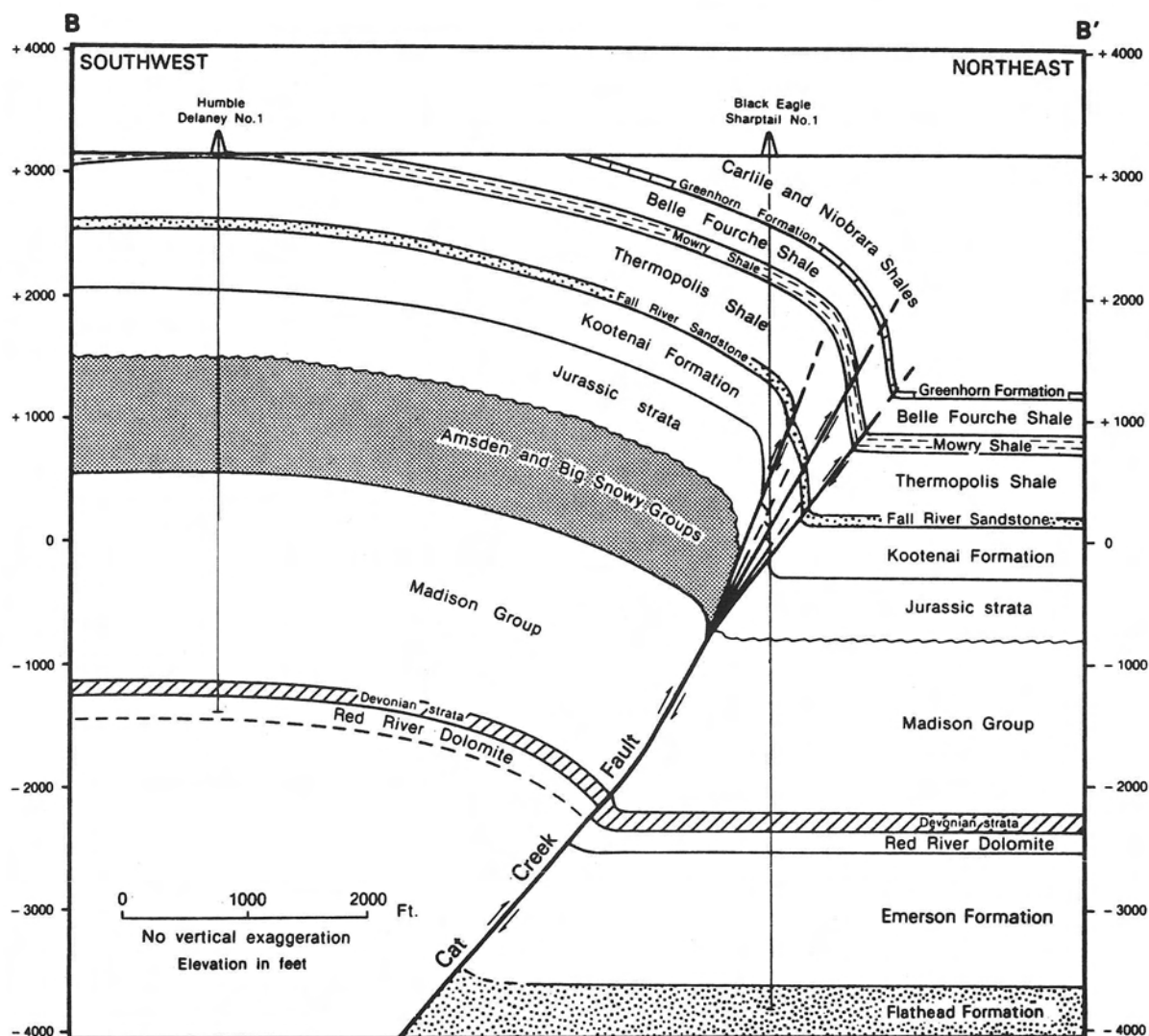


Figure 27—Cross section of Bear Creek dome in Section 17, T.17N., R.24E. (See Plate 1 for line of section.) The Black Eagle–Sharptail well penetrated steeply dipping Cretaceous strata, with several small reverse faults in the Kootenai Formation. Jurassic and older strata in the well are nearly horizontal. The Cat Creek fault is inferred to bifurcate and die out upward by ductile thinning of Cretaceous shales. Two pre-Cretaceous episodes of movement of the Cat Creek fault are indicated. Slight thinning of Devonian rocks on the southwest side of the fault implies reverse faulting and erosion of Devonian beds on the uplifted block prior to deposition of the Madison Group. Post-Absaroka/pre-Zuni sequence normal faulting is demonstrated by erosion of Big Snowy and Amsden groups from northeast side of the fault prior to Jurassic sedimentation.



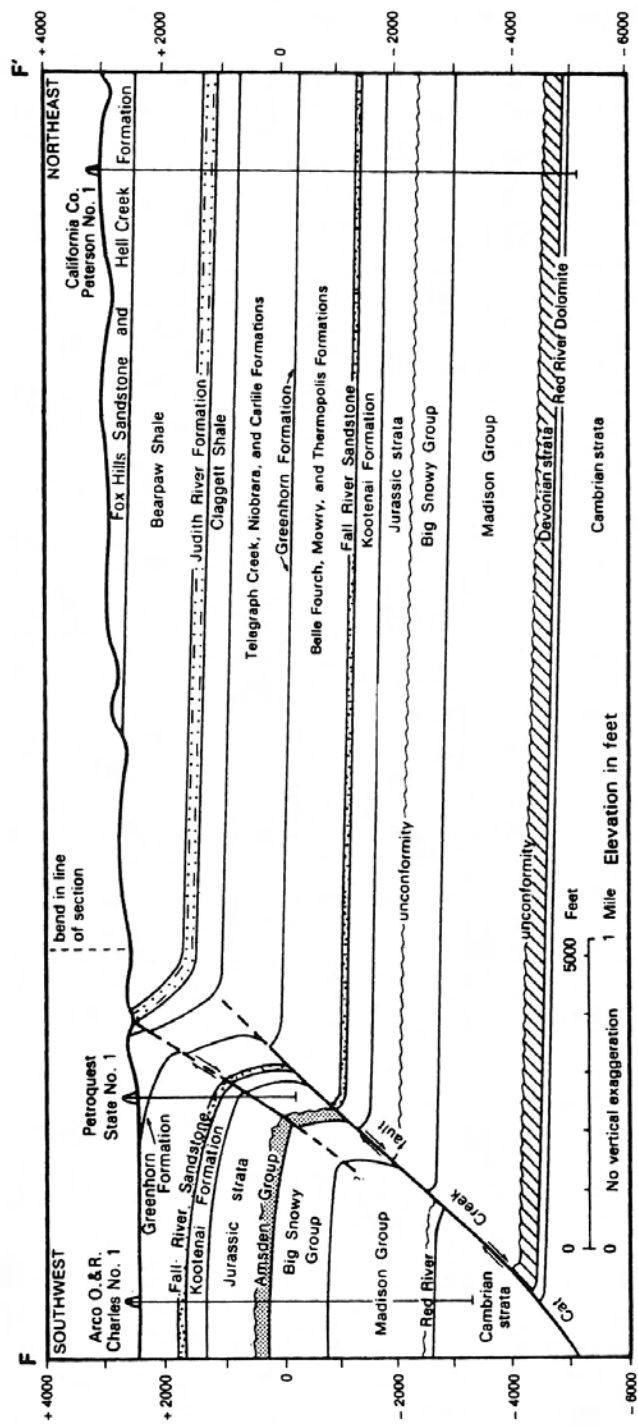


Figure 28—Cross section of Mosby dome and Cat Creek fault. (See Plate 1 for line of section.) The Petroquest-State No. 1 well drilled through steeply dipping strata and a small reverse fault, which is probably in the hanging wall of the main Cat Creek fault. The small fault may reach the surface, displacing Judith River strata. The section shows evidence for two pre-Laramide displacements of the Cat Creek fault. The southwest block was uplifted and eroded near the end of the Devonian Period; reverse faulting was involved. Later post-Absaroka/pre-Zuni normal faulting resulted in erosion of Amsden Group and part of Big Snowy Group from northeast block prior to Jurassic deposition.





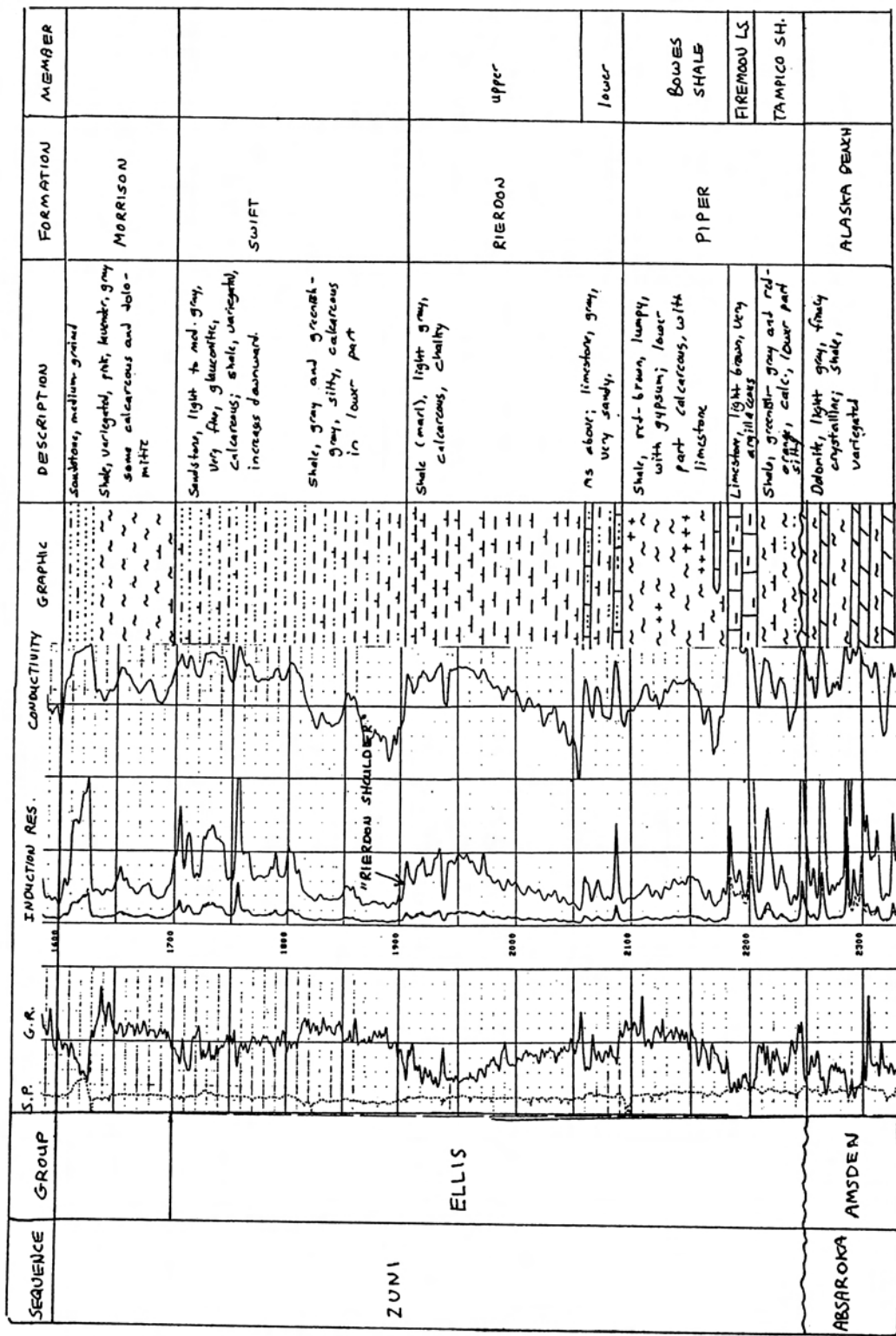


Figure 29. Reference log for Ellis Group: Tricentrol U.S. # 22-15 Federal, SE $\frac{1}{4}$  SW $\frac{1}{4}$  SE $\frac{1}{4}$ , Sec. 22, T15N, R30E, Petroleum County.



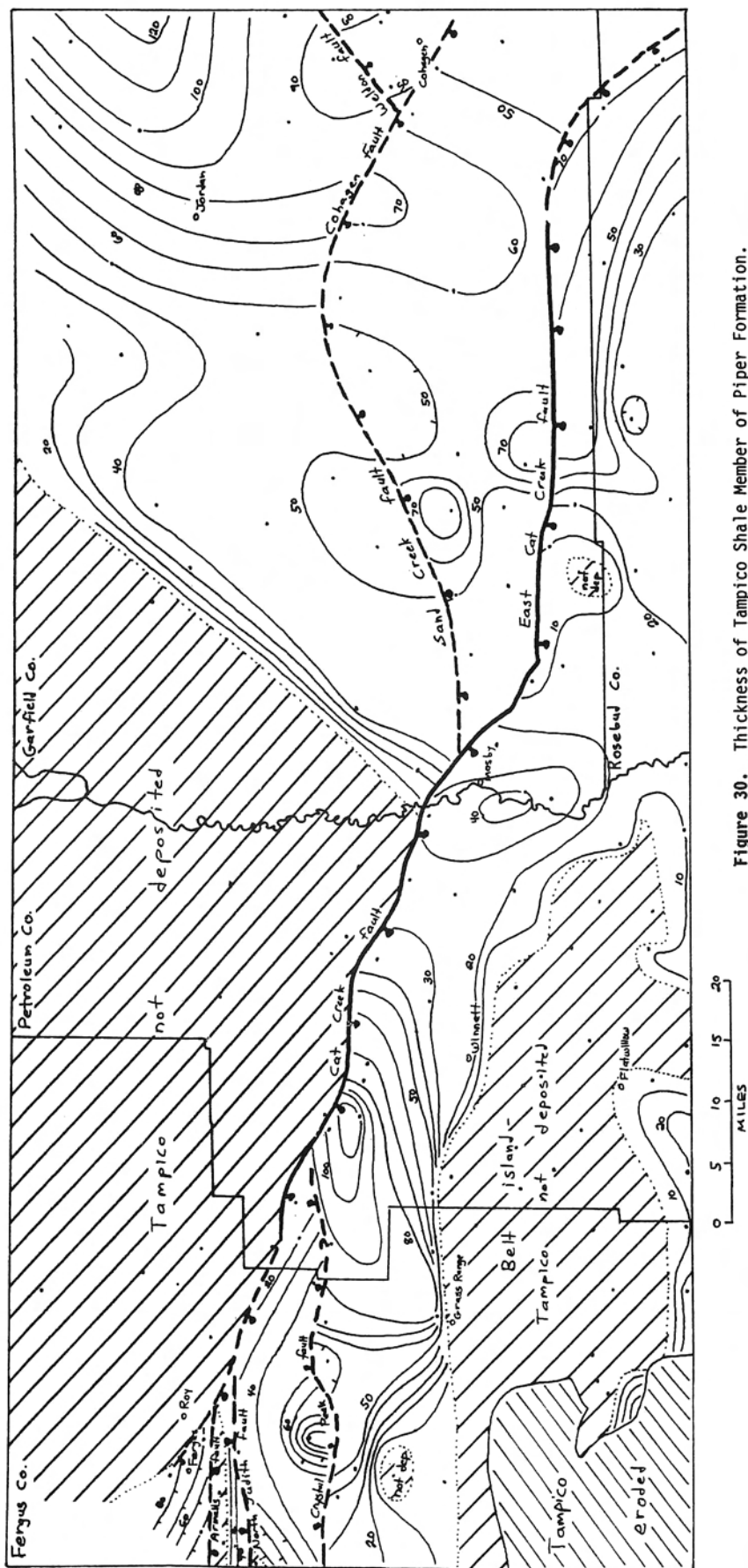


Figure 30. Thickness of Tampico Shale Member of Piper Formation. Control well (selected records used). Isopach line; contour interval 10 feet. Pre-Jurassic fault.



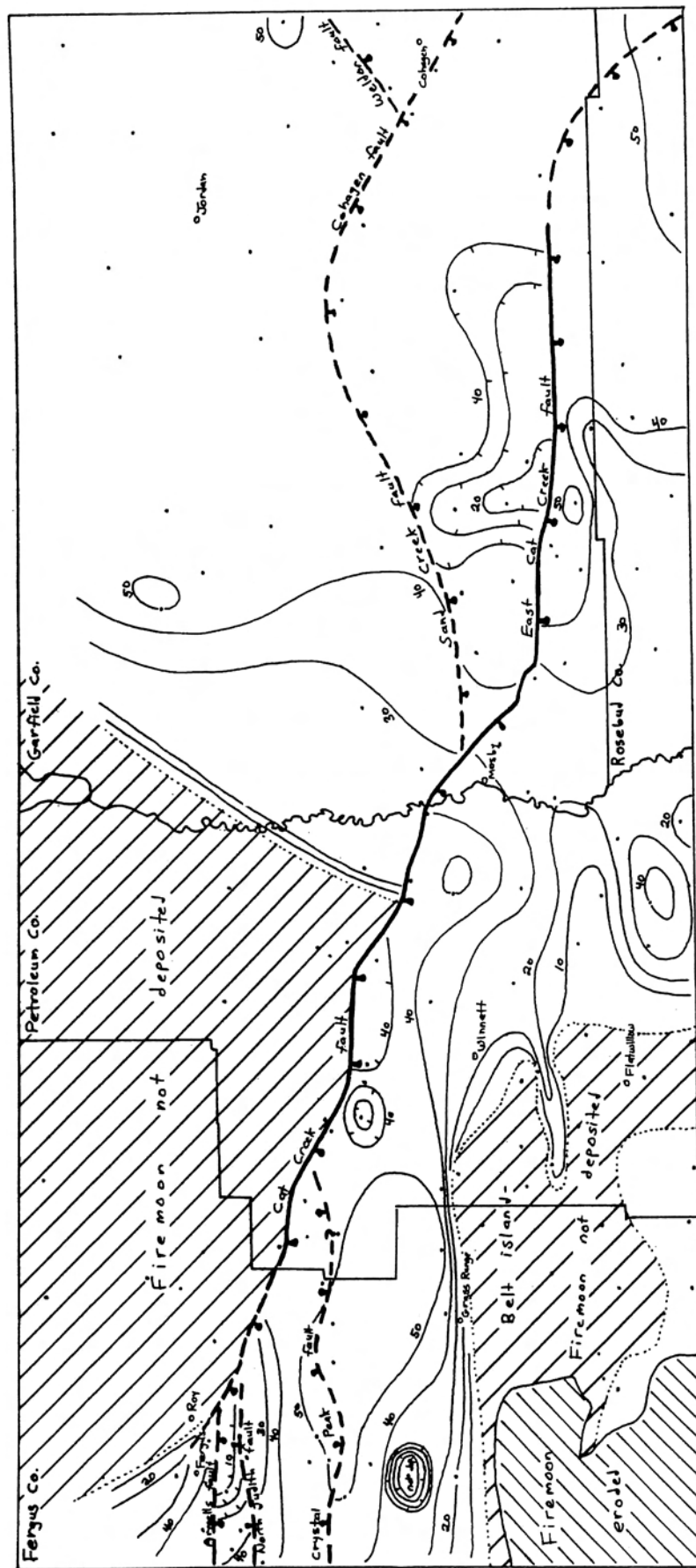


Figure 31. Thickness of Firemoon Limestone Member of Piper Formation.  
 . Control well (selected records used).  
 --- Isopach line; contour interval 10 feet.  
 — Pre-Jurassic fault.



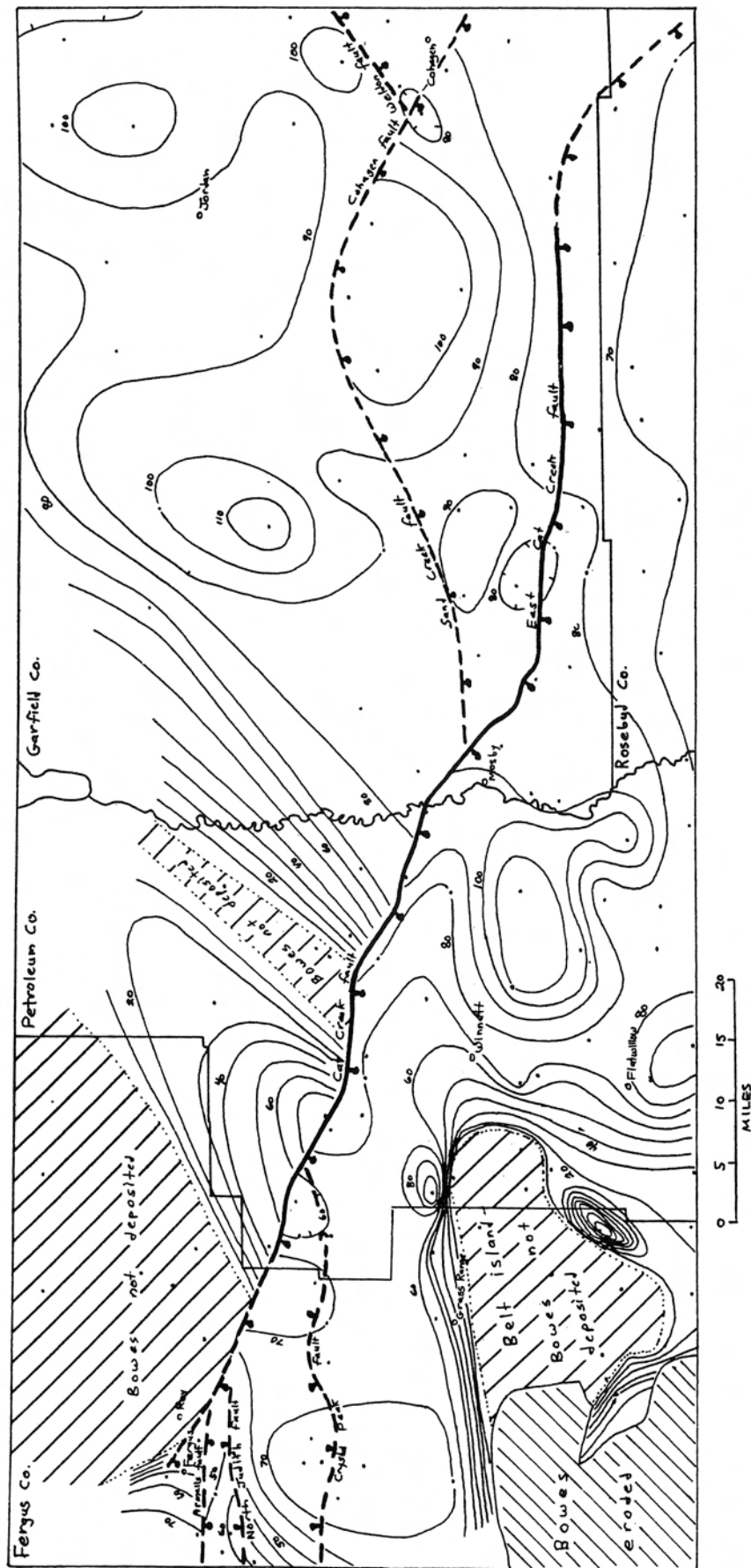


Figure 32. Thickness of Bowes Shale Member of Piper Formation.  
 • Control well (selected records used).  
 — Isopach line; contour interval 10 feet.  
 1-2 Pre-Jurassic fault.





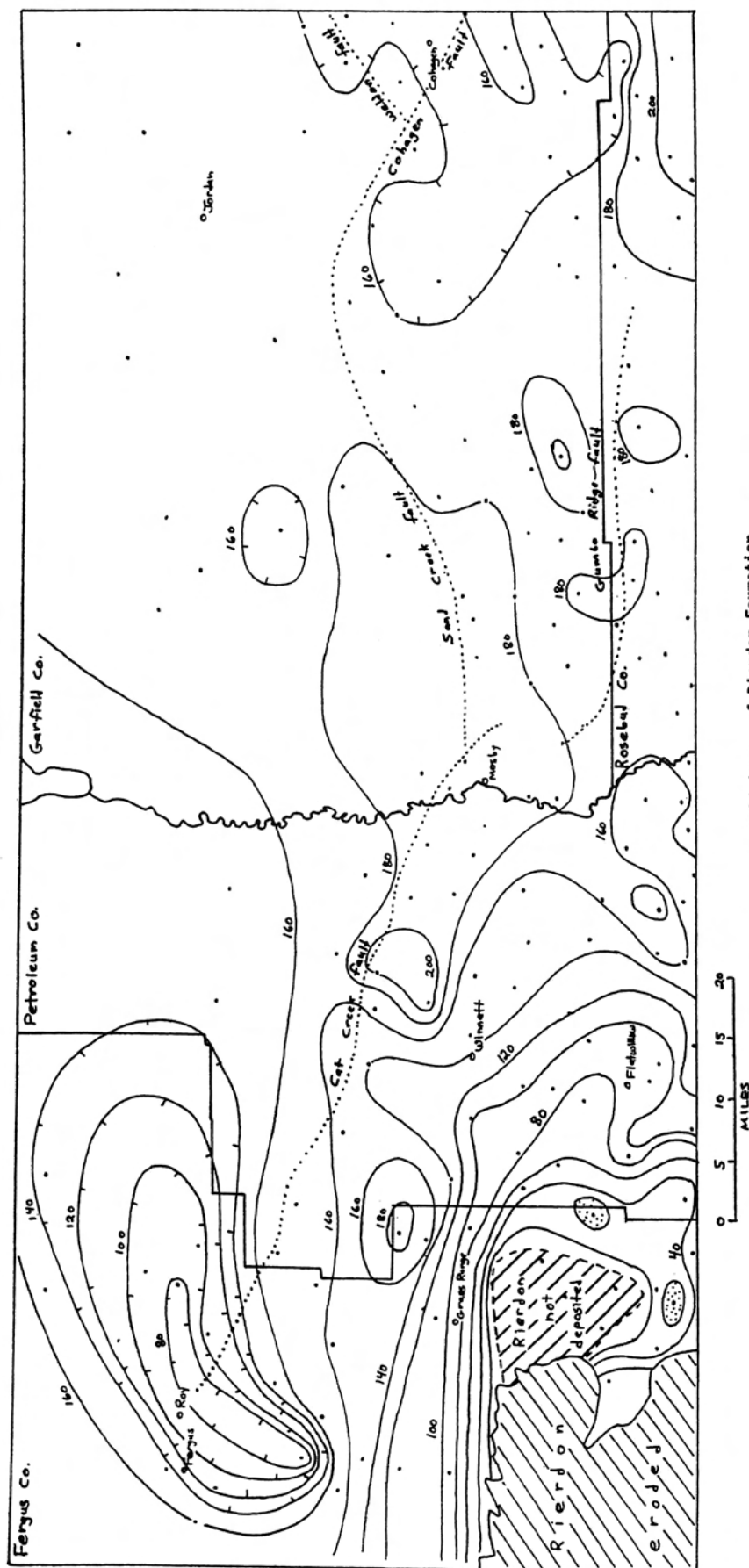
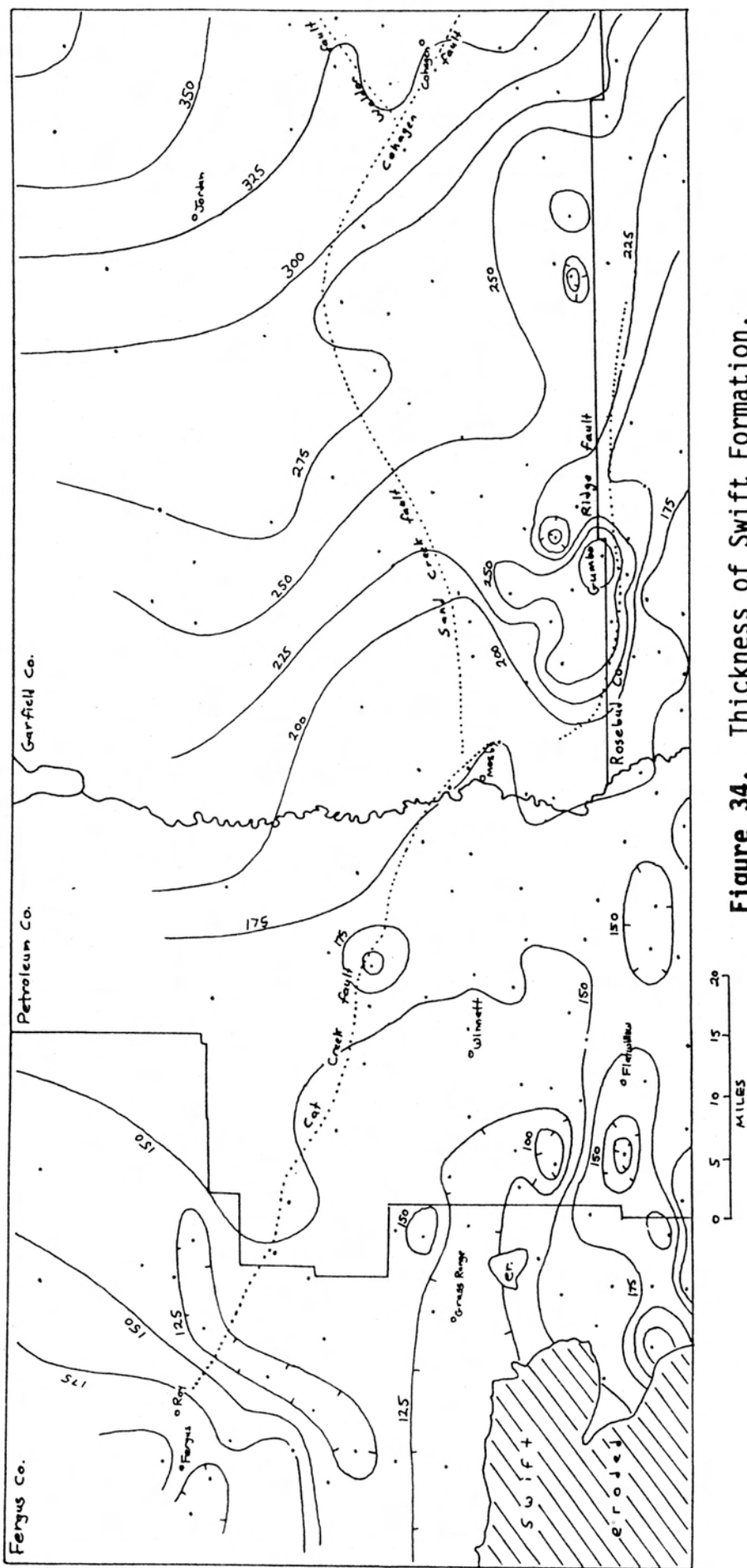


Figure 33. Thickness of Rierdon Formation.  
 \* Control well (selected records used).  
 --- Isopach line; contour interval 20 feet.  
 ..... Inactive fault  
 [Stippled area] Rierdon eroded beneath Swift Formation.





**Figure 34.** Thickness of Swift Formation.  
 • Control well (selected records used).  
 — Isopach line; contour interval 25 feet.  
 ..... Inactive fault.



MBMG 256

INSERT FIGURE 35 (tabloid)





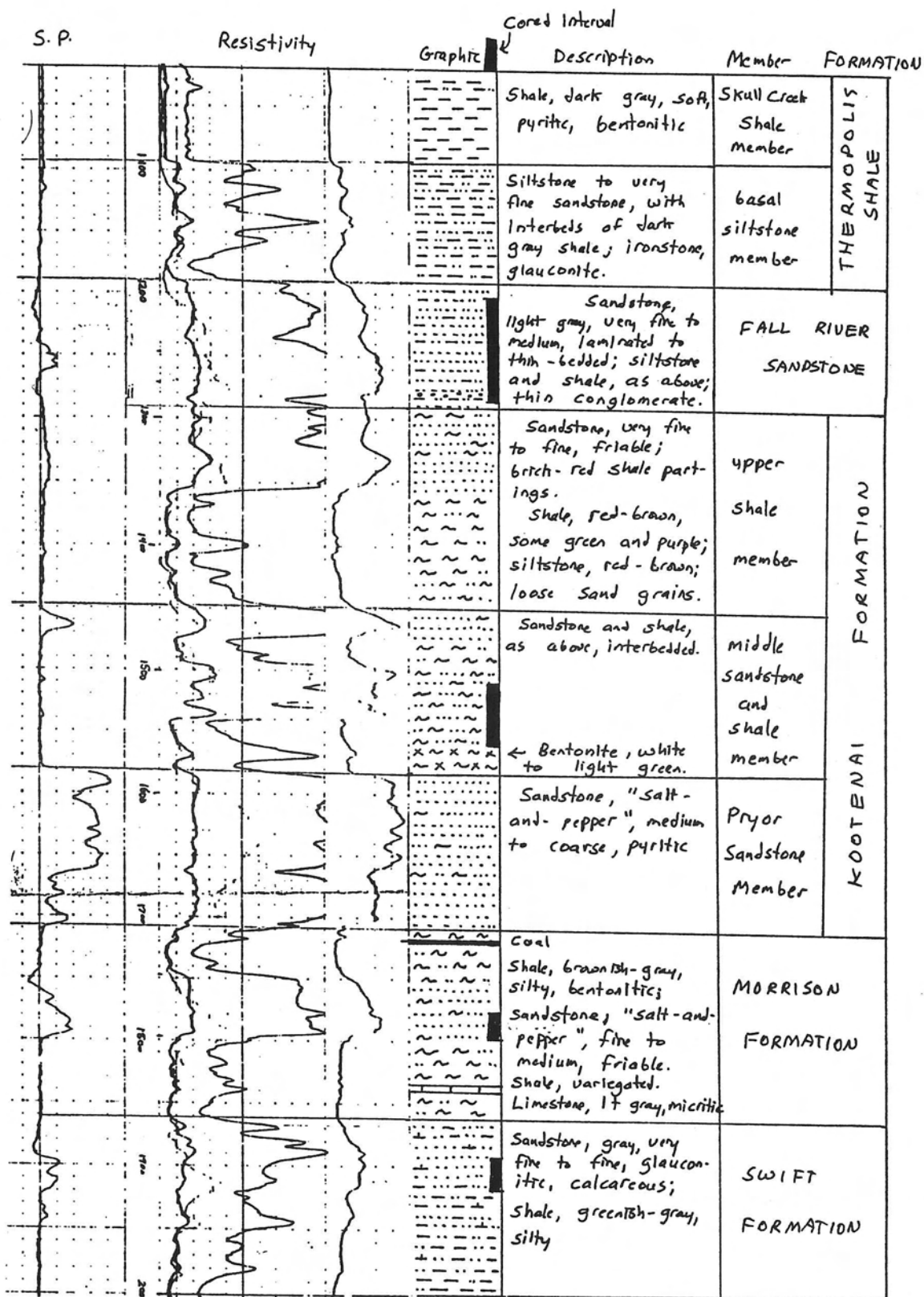


Figure 36. Reference log for Kootenai and Morrison Formations:  
Utah Southern Oil Co. # 23 Rhens-Government, NE $\frac{1}{4}$  SE $\frac{1}{4}$ , Sec. 9,  
T15N, R29E, Petroleum County.



E.C. Brown et al., Grebe #1  
SW-NE 33-12N-32E

Vaughey and Vaughey, Grebe #1  
SE-SW 34-12N-32E

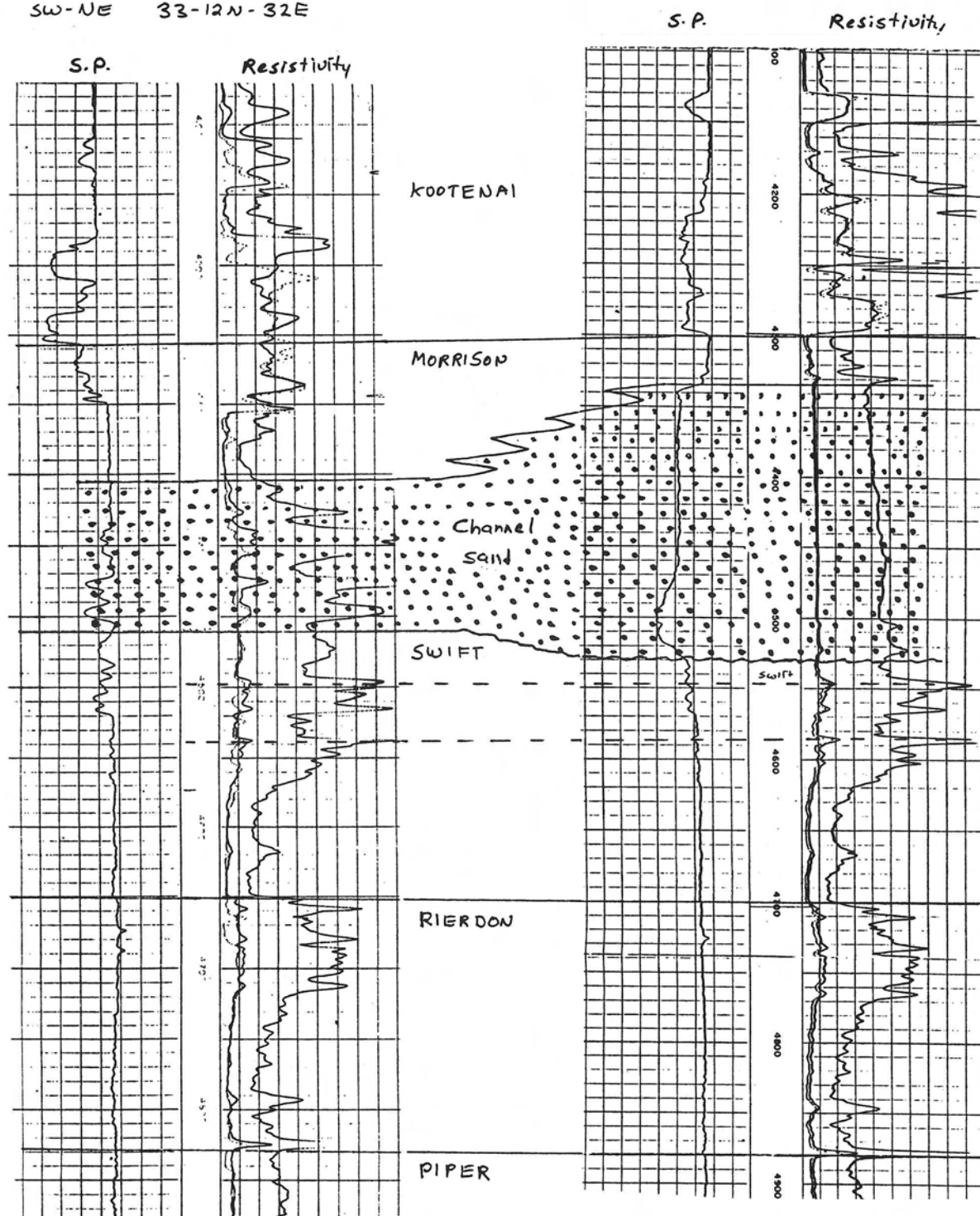


Figure 37. Log section showing thick channel sand in Morrison Formation in Gumbo Ridge area.



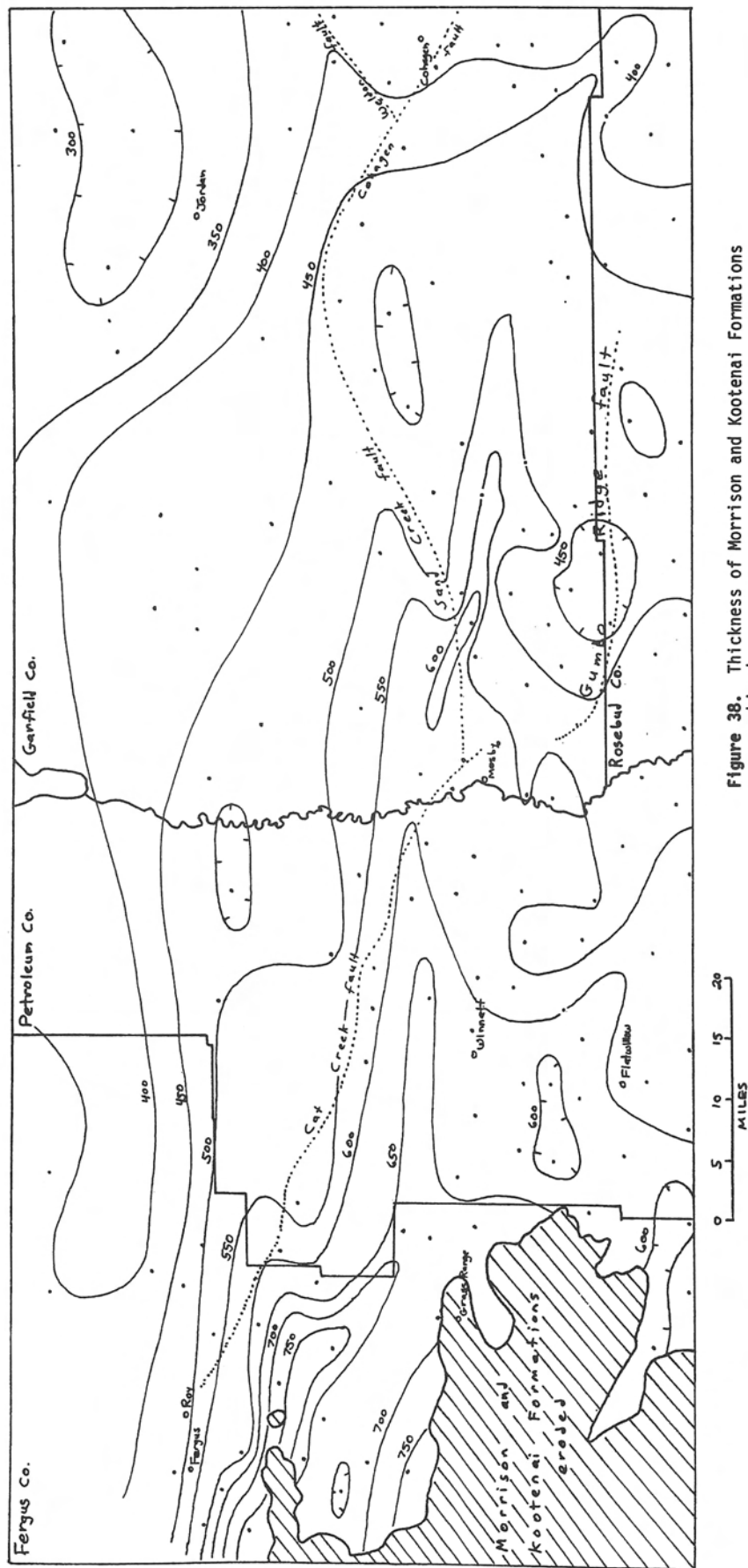


Figure 38. Thickness of Morrison and Kootenai Formations combined.  
 Control well (selected records used).  
 Isopach line; contour interval 50 feet.  
 Inactive fault.



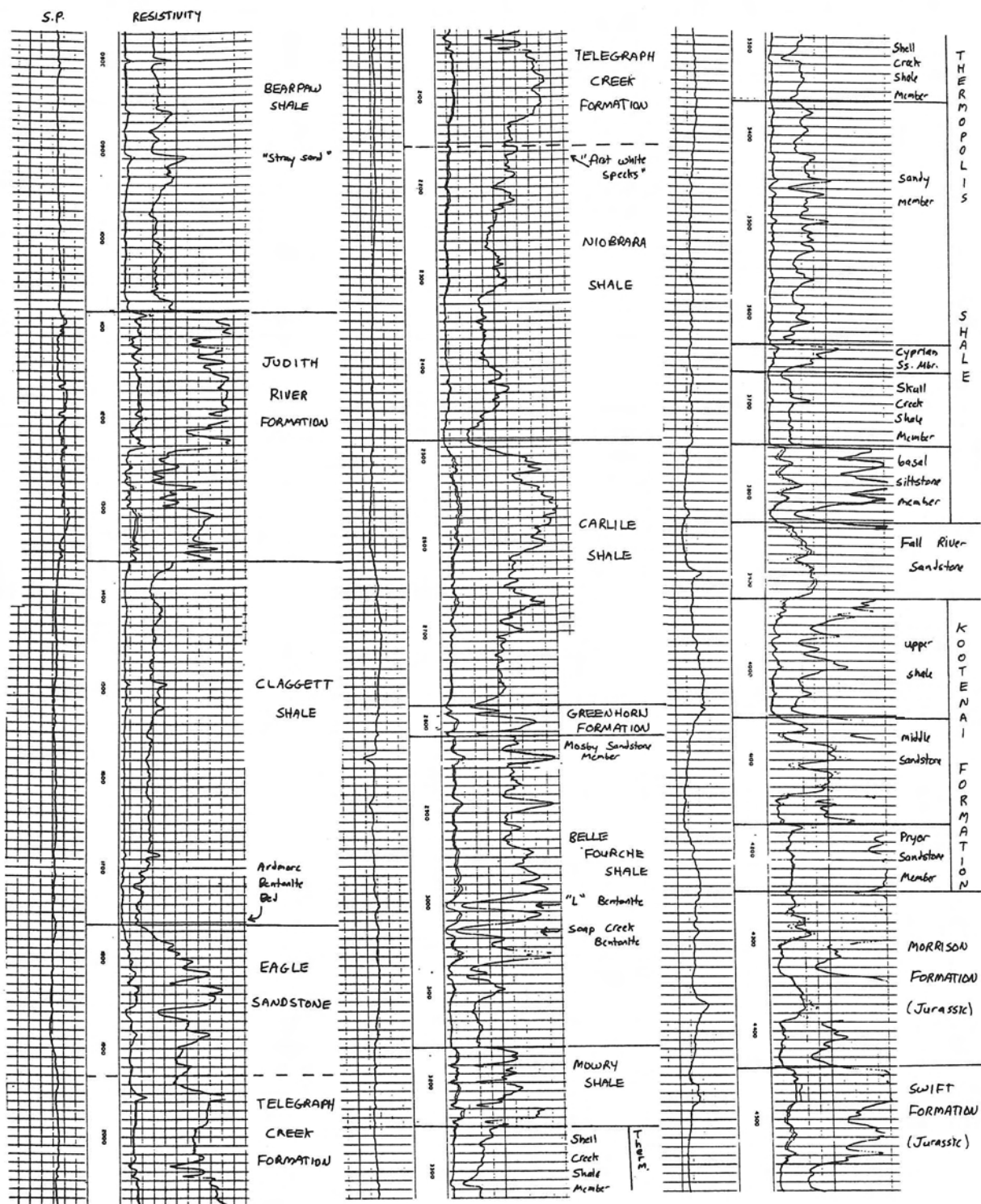


Figure 39. Reference log for Cretaceous System: Montana Power Co.  
# 12-22 Federal, NW $\frac{1}{4}$  SW $\frac{1}{4}$ , Sec. 22, T16N, R29E, Petroleum  
County.





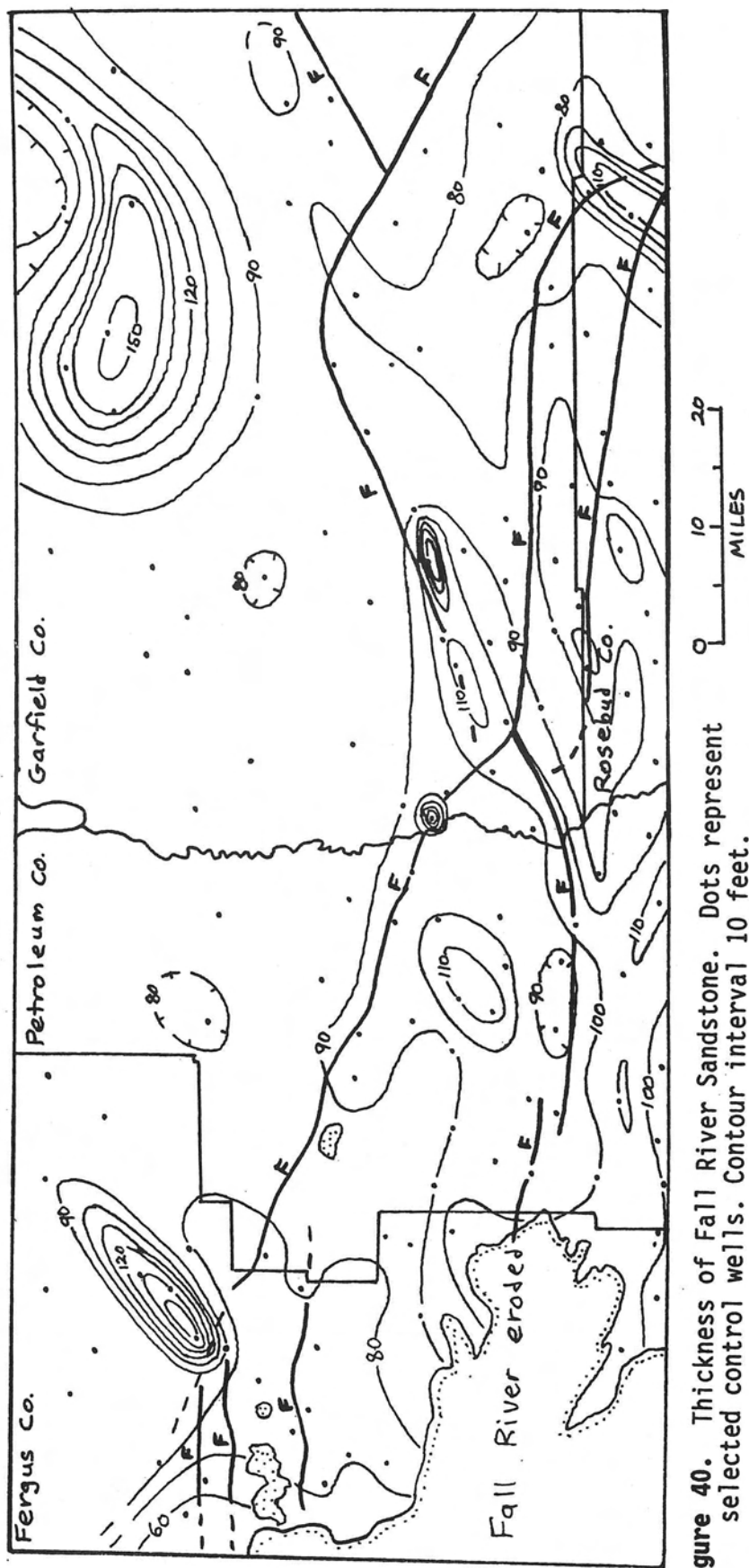


Figure 40. Thickness of Fall River Sandstone. Dots represent selected control wells. Contour interval 10 feet.

F — Fault.



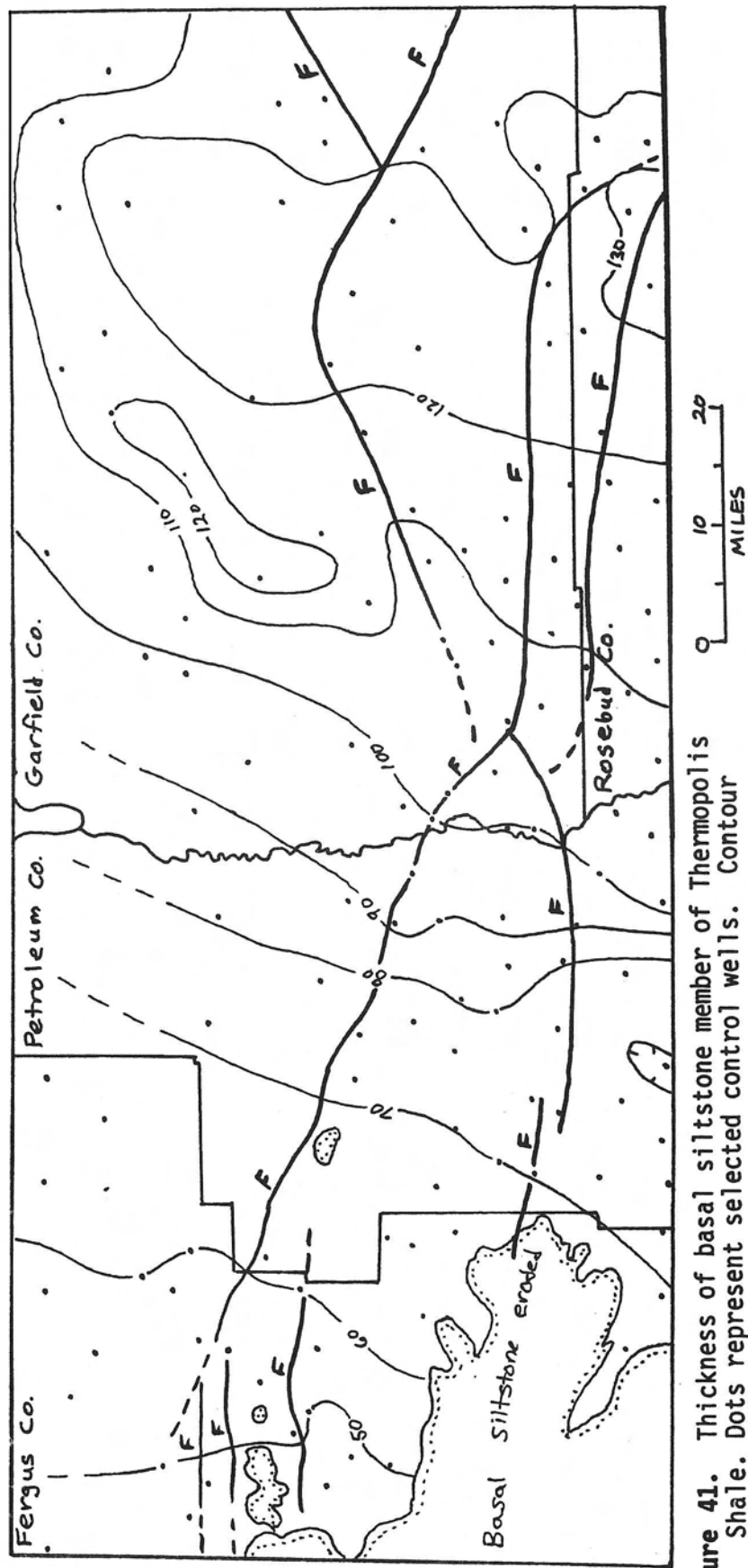


Figure 41. Thickness of basal siltstone member of Thermopolis Shale. Dots represent selected control wells. Contour interval 10 feet.

F Fault.



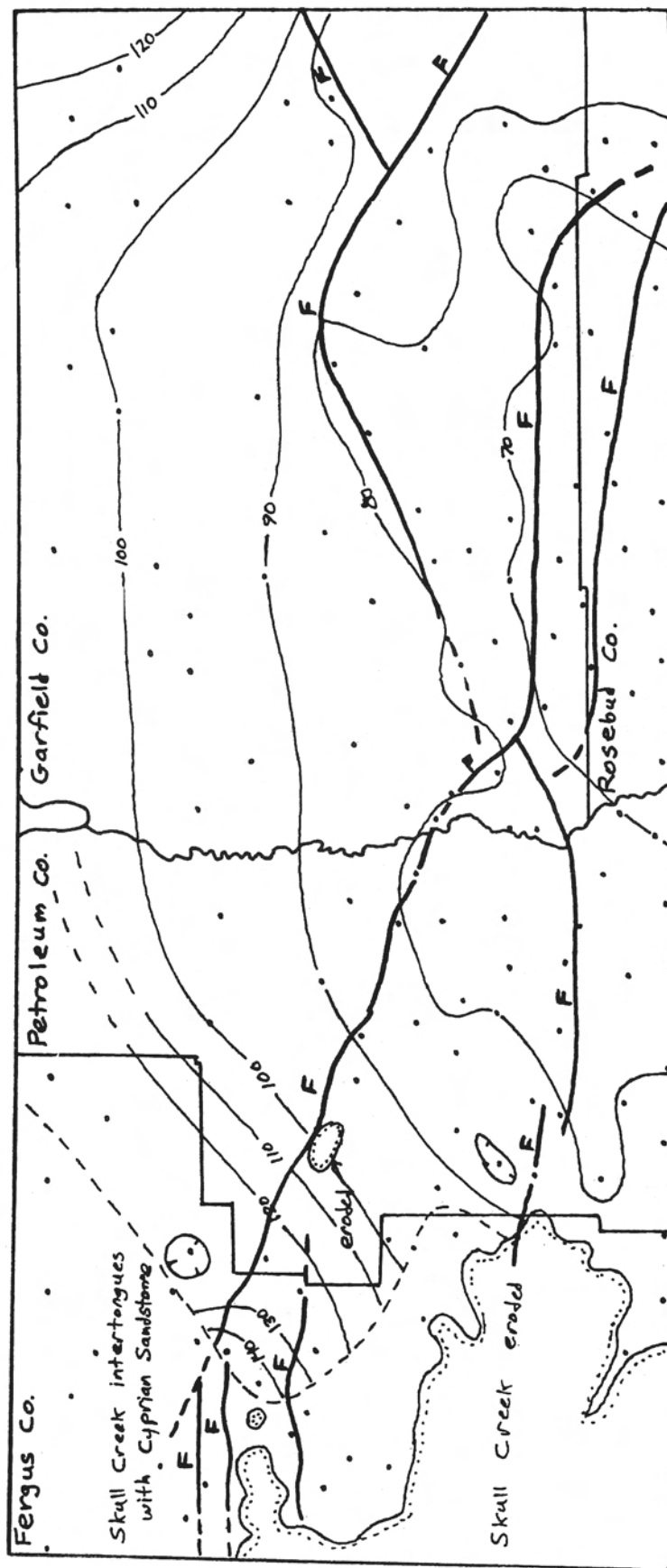
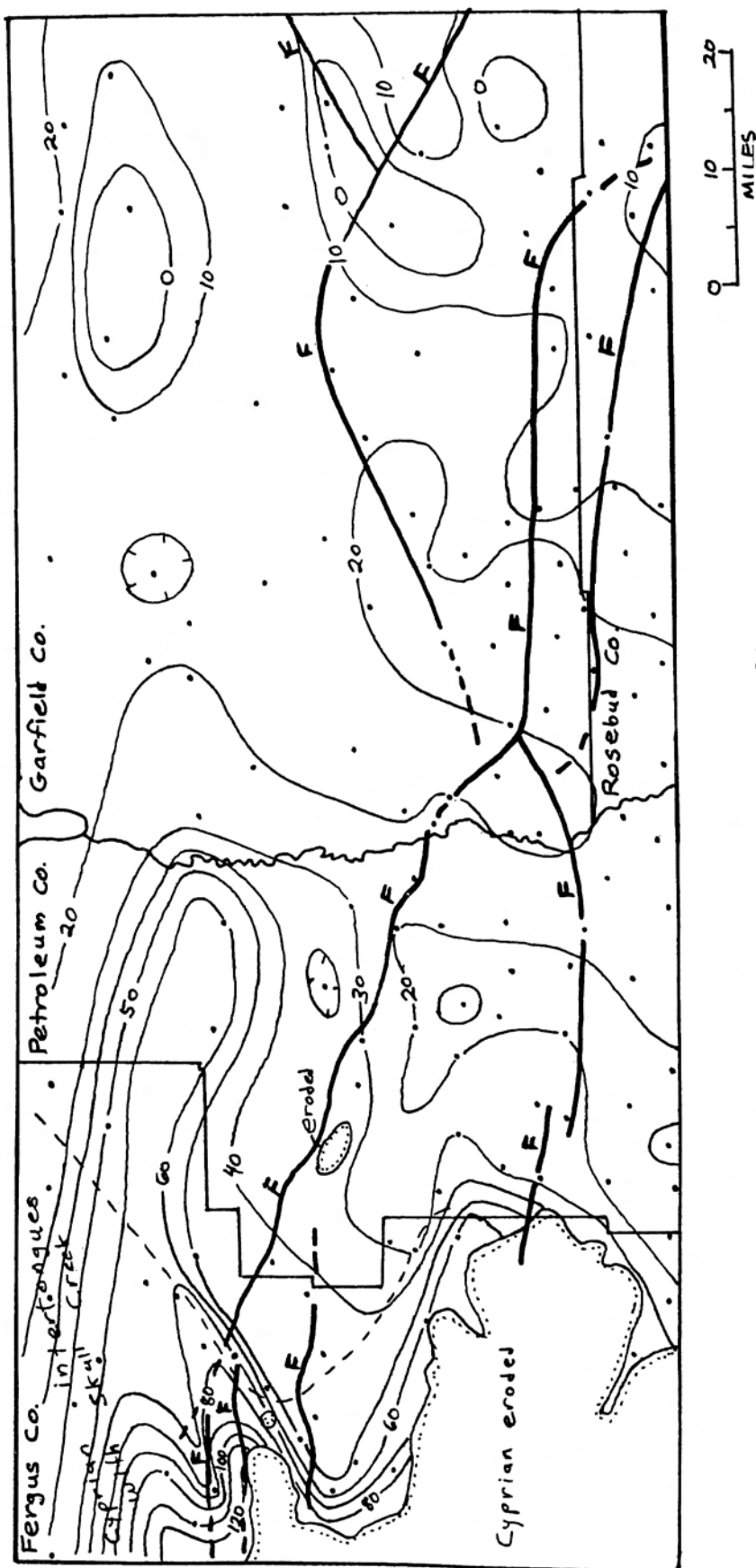


figure 42. Thickness of Skull Creek Shale Member of Thermopolis Shale. Dots represent selected control wells. Contour interval 10 feet.  
 F Fault.







**Figure 43.** Thickness of Cyprian Sandstone Member of Thermopolis Shale. Dots represent selected control wells. Contour interval 10 feet.  
**F** — Fault.



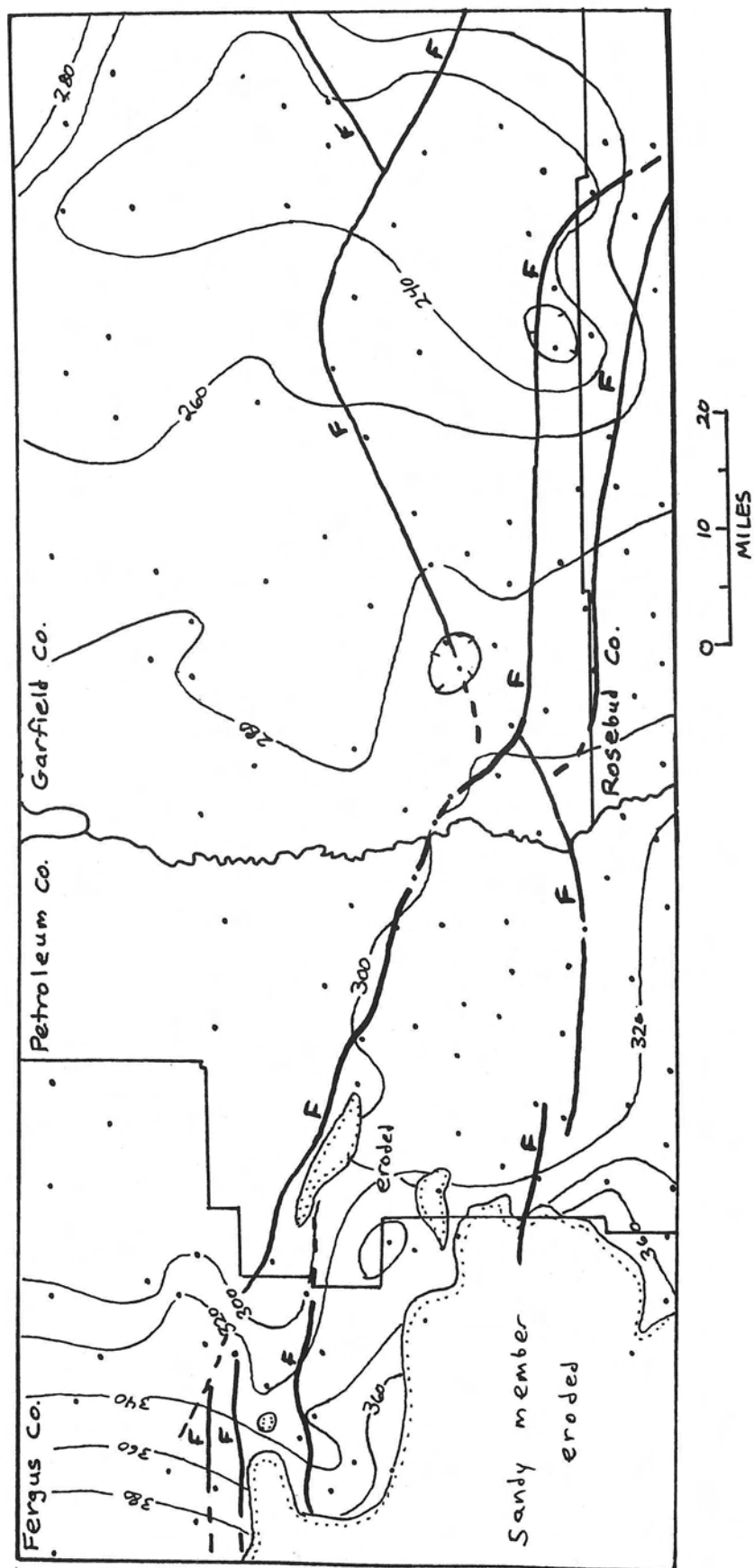
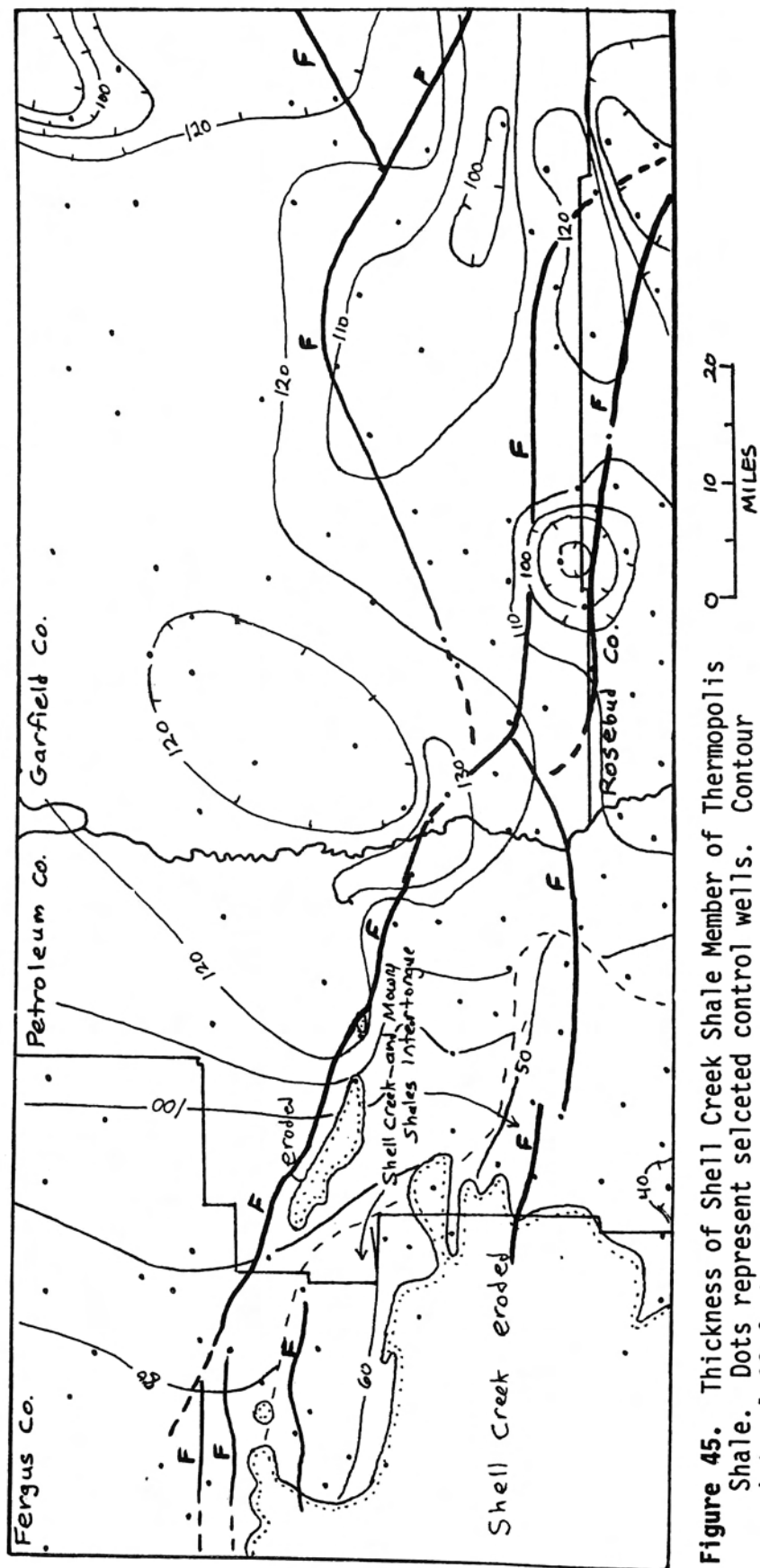


Figure 44. Thickness of unnamed sandy member of Thermopolis Shale. Dots represent selected control wells. Contour interval 20 feet.  
 F Fault.





**Figure 45.** Thickness of Shell Creek Shale Member of Thermopolis Shale. Dots represent selected control wells. Contour interval 10 feet.

**F** Fault.



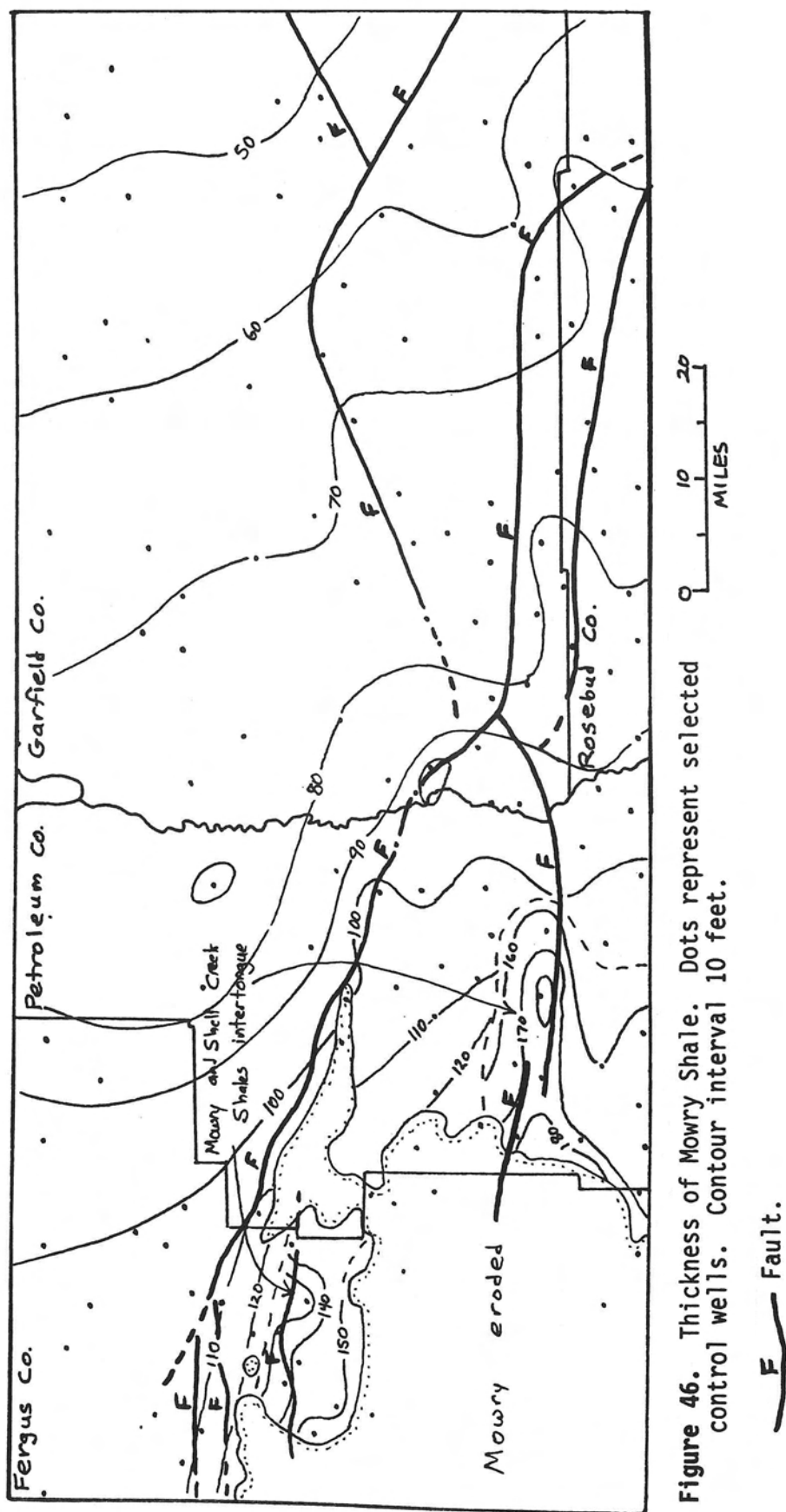


Figure 46. Thickness of Mowry Shale. Dots represent selected control wells. Contour interval 10 feet.

F — Fault.





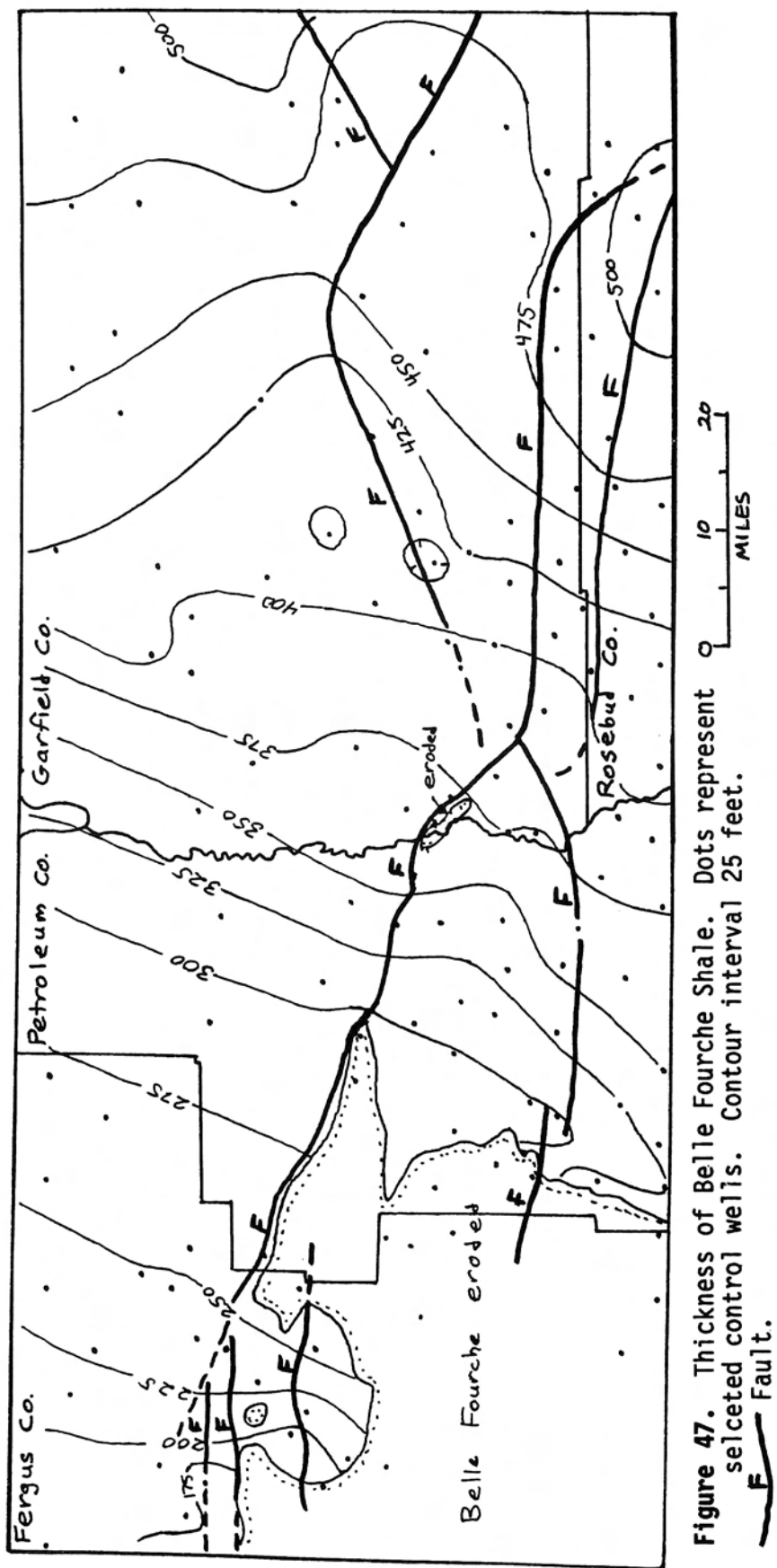


Figure 47. Thickness of Belle Fourche Shale. Dots represent selected control wells. Contour interval 25 feet.

F Fault.



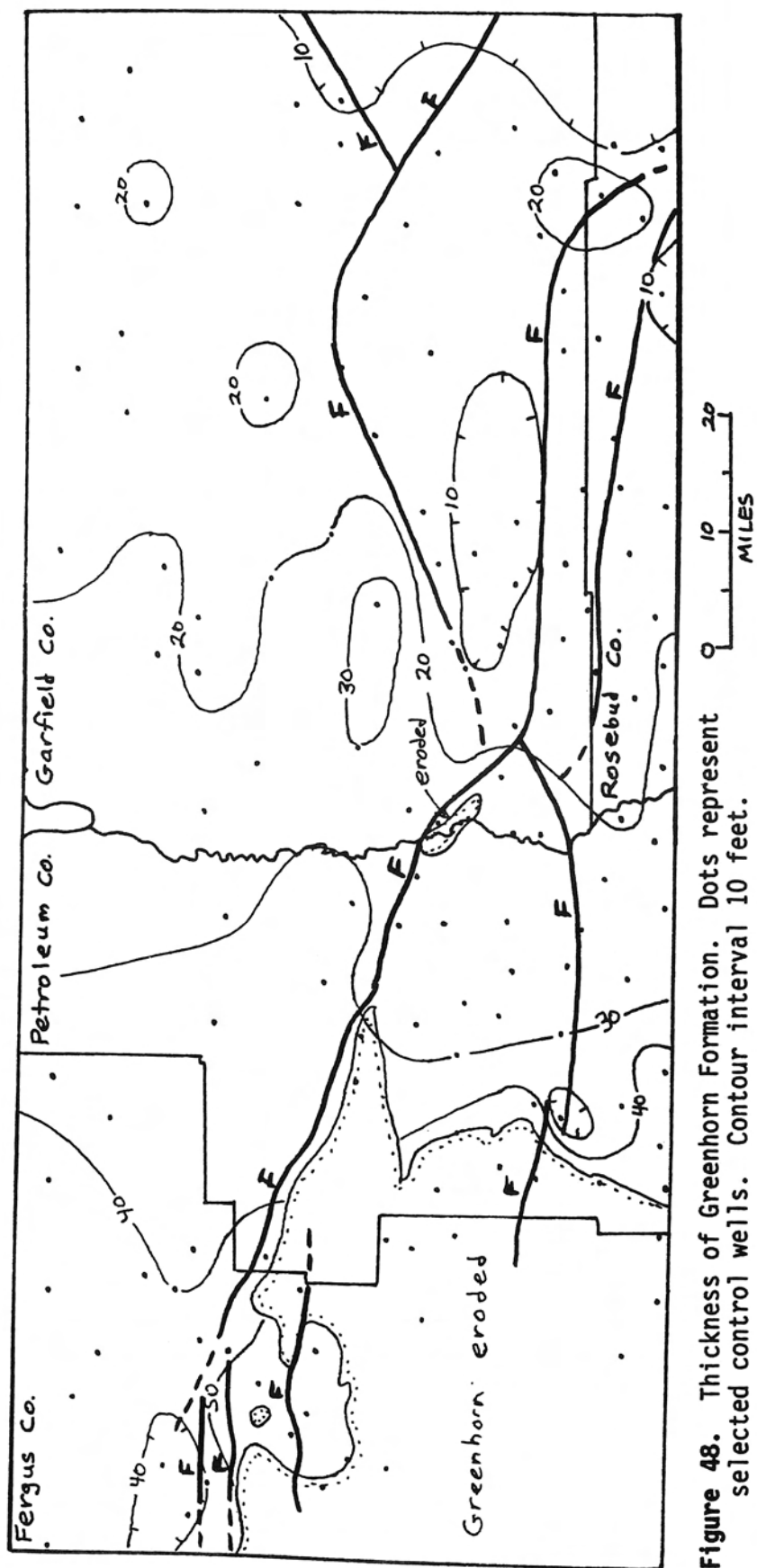


Figure 48. Thickness of Greenhorn Formation. Dots represent selected control wells. Contour interval 10 feet.

**F** Fault.



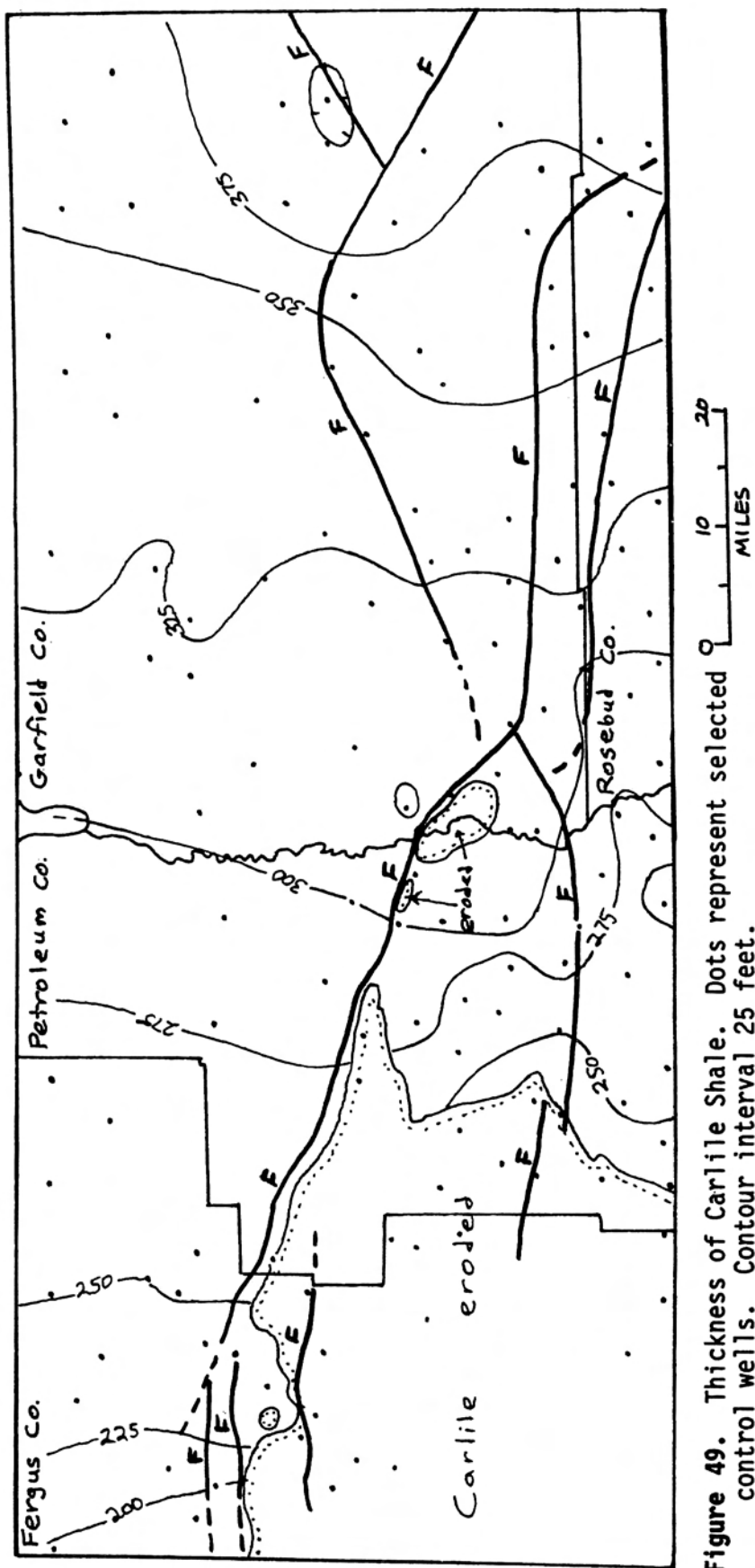


Figure 49. Thickness of Carlile Shale. Dots represent selected control wells. Contour interval 25 feet.

F — Fault.





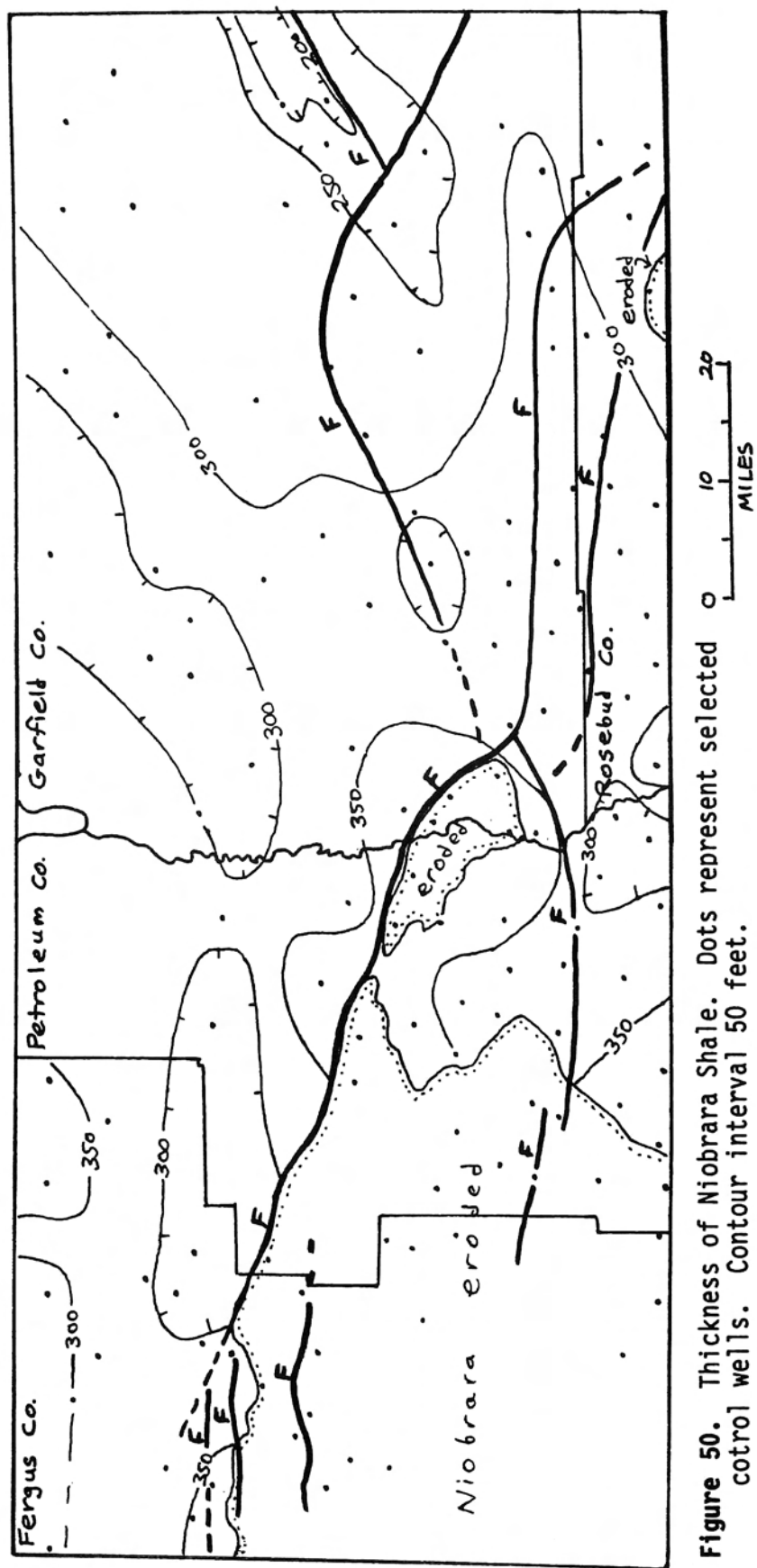
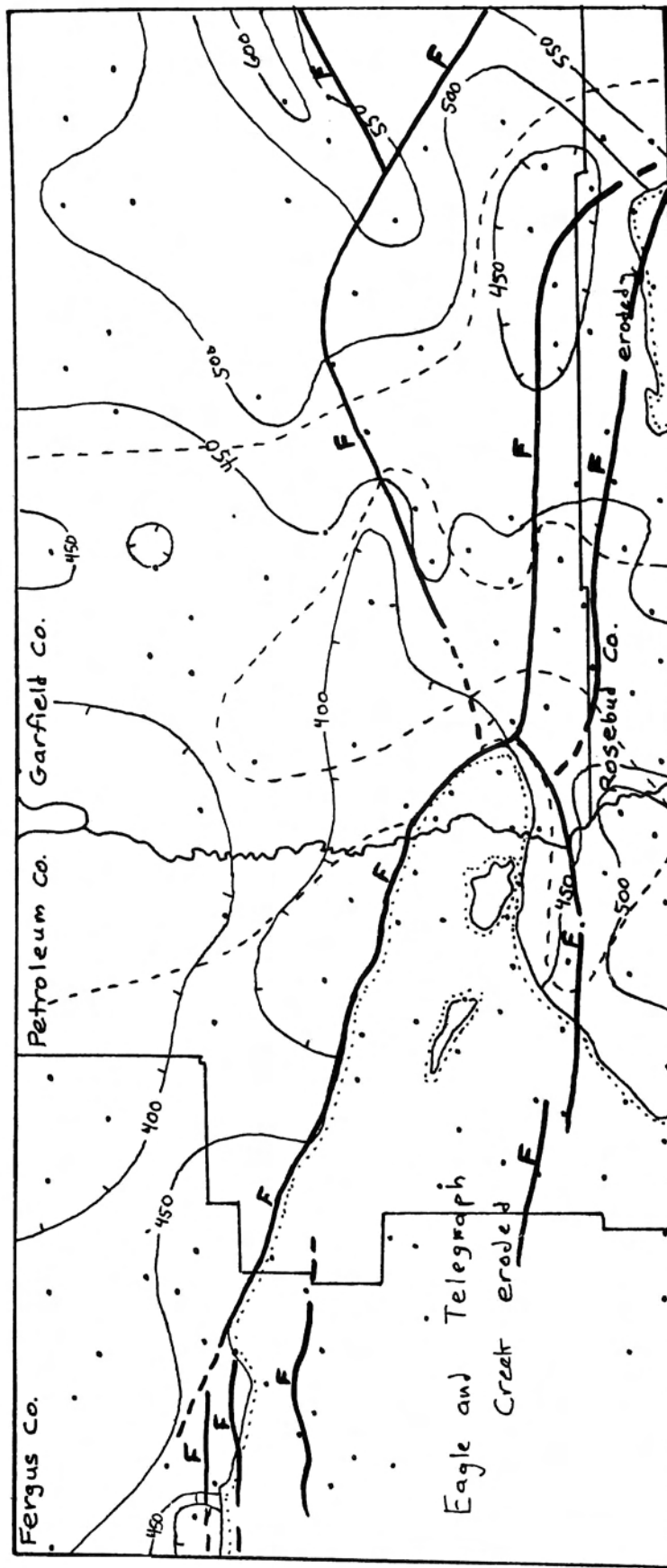


Figure 50. Thickness of Niobrara Shale. Dots represent selected control wells. Contour interval 50 feet.

**F** Fault.







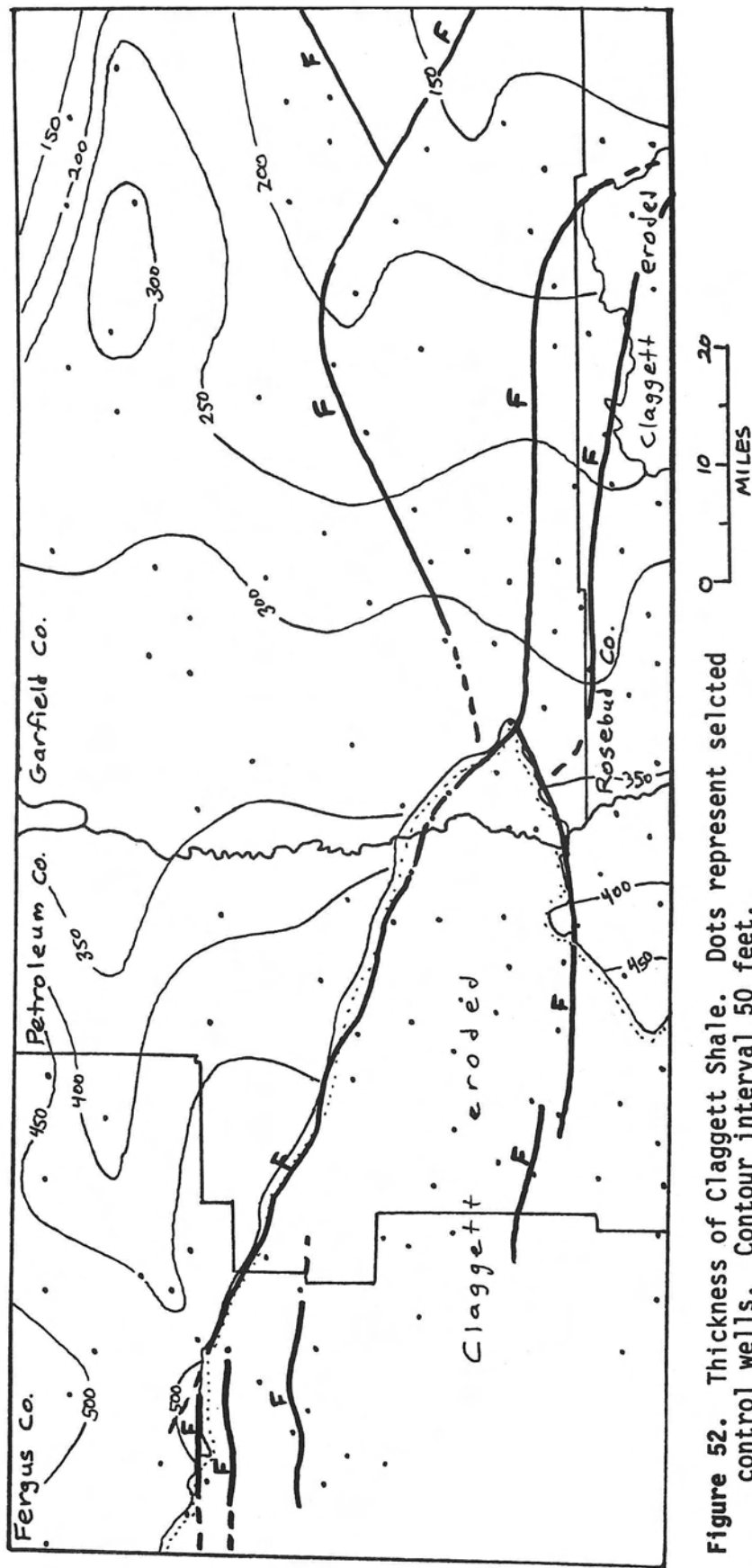


Figure 52. Thickness of Claggett Shale. Dots represent selected control wells. Contour interval 50 feet.

F Fault.



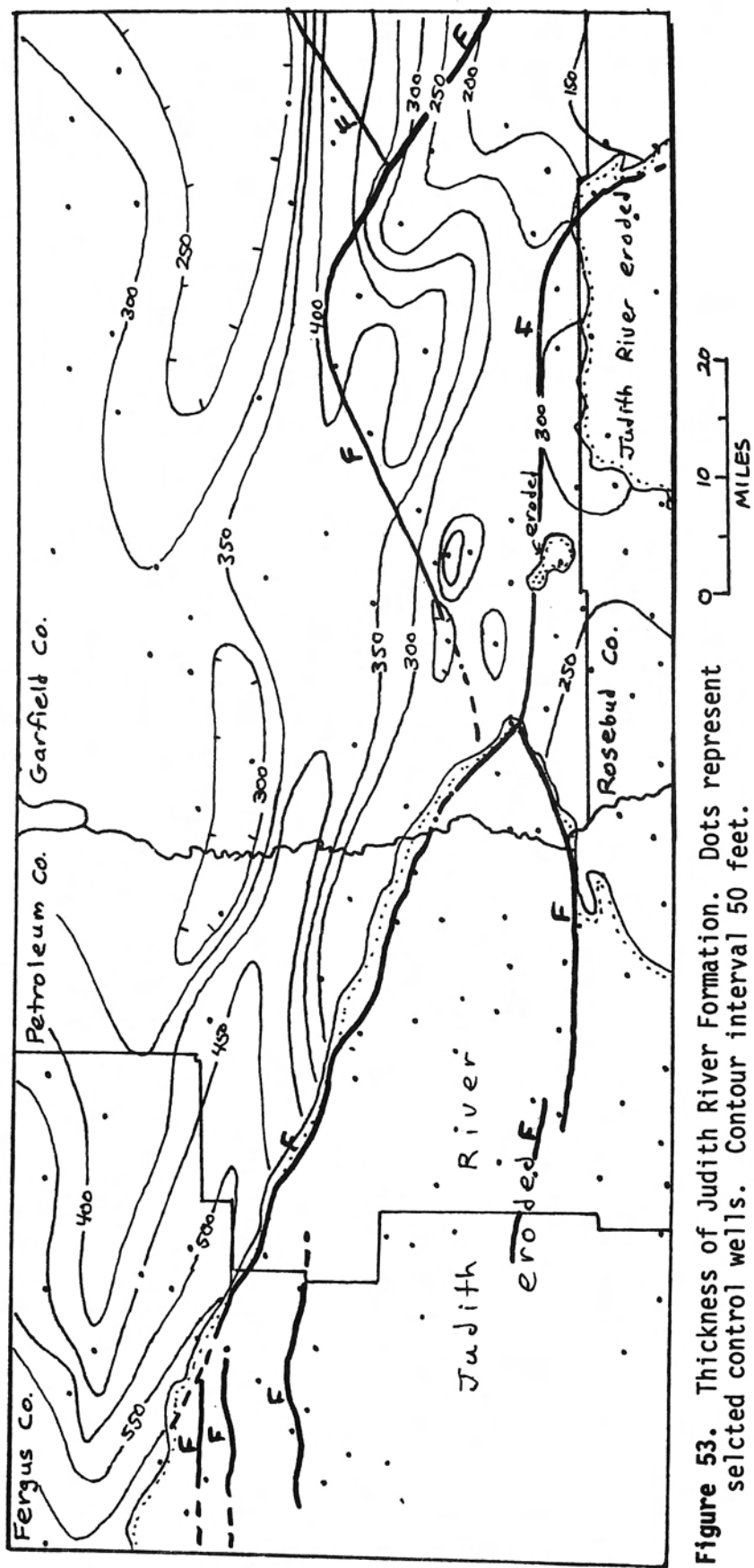


Figure 53. Thickness of Judith River Formation. Dots represent selected control wells. Contour interval 50 feet.

F Fault.





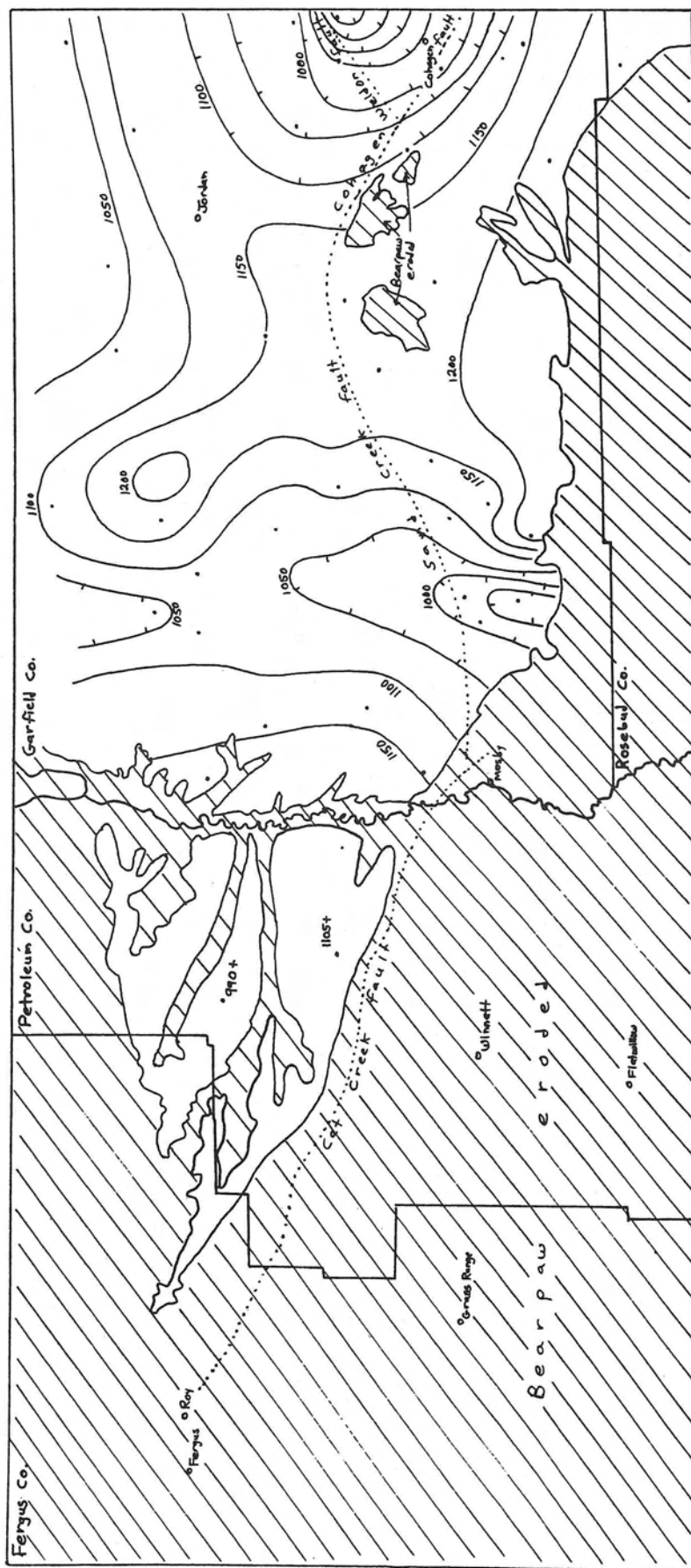


Figure 54. Thickness of Bearpaw Shale. Dots represent selected control wells. Contour interval 50 feet.  
 .... Fault.



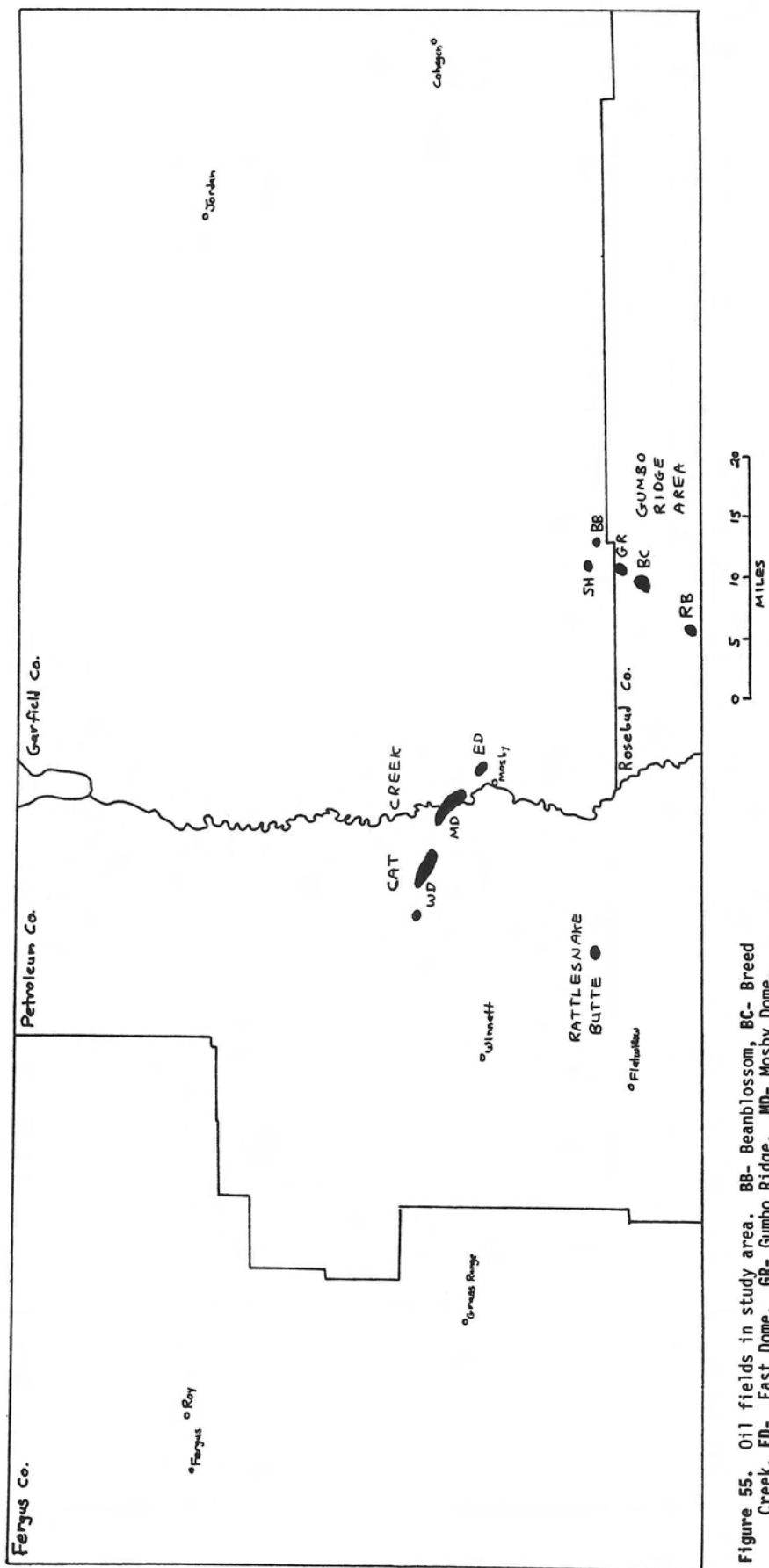
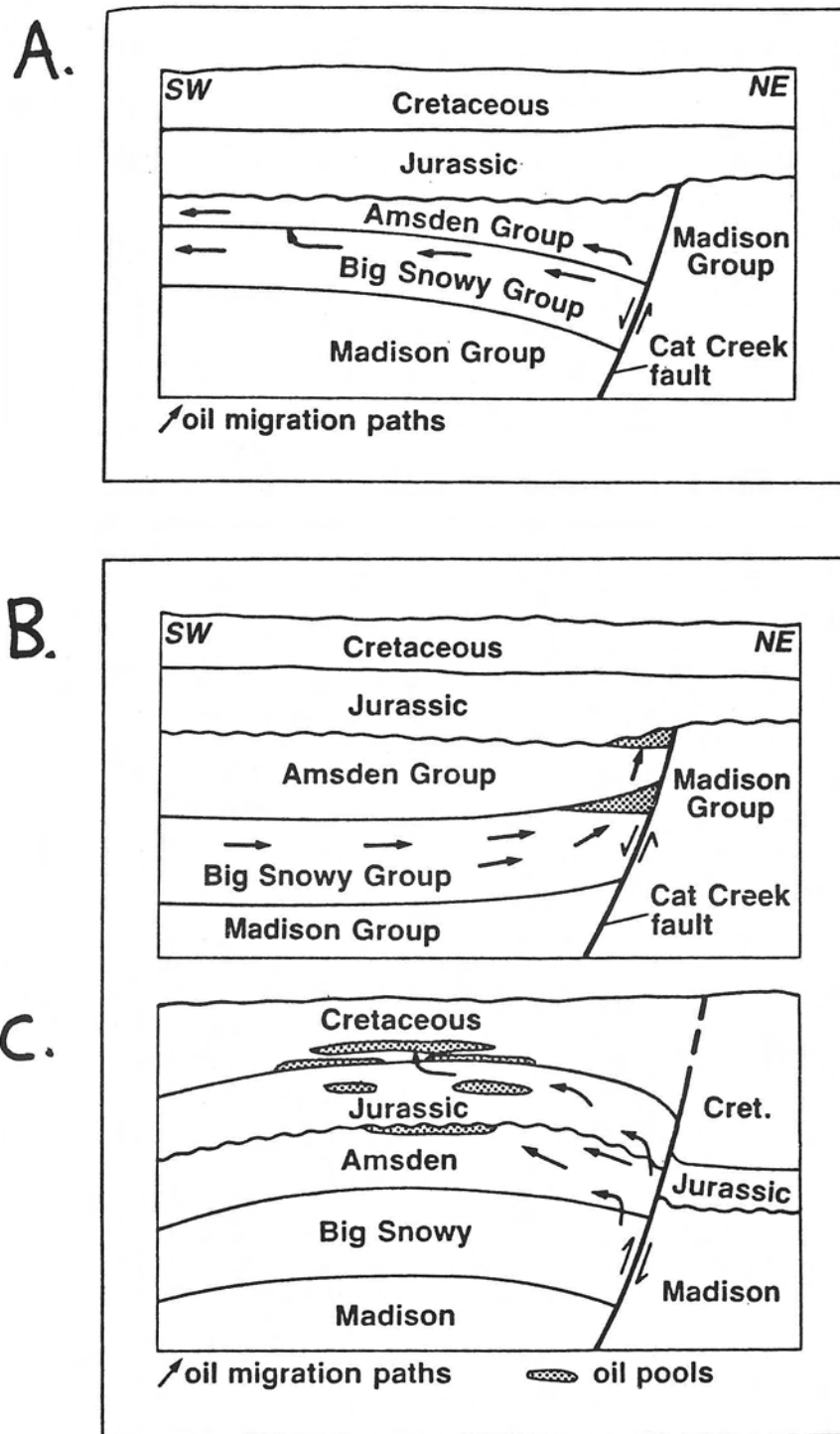


Figure 55. Oil fields in study area. BB- Beanblossom, BC- Breed Creek, ED- East Dome, GR- Gumbo Ridge, MD- Mosby Dome, RB- Rattler Butte, SH- Studhorse, WD- West Dome.





**Figure 56.** Diagram illustrating possible oil-migration pathways in the Cat Creek anticline. **A.** In western part of anticline, oil migrates up-dip and out of the area before the Laramide orogeny. **B.** In eastern part of anticline, oil migrates up-dip and is trapped against the fault before the Laramide orogeny. **C.** In eastern part of anticline, oil re-migrates to the crest of the anticline as it rises during the Laramide orogeny.

