



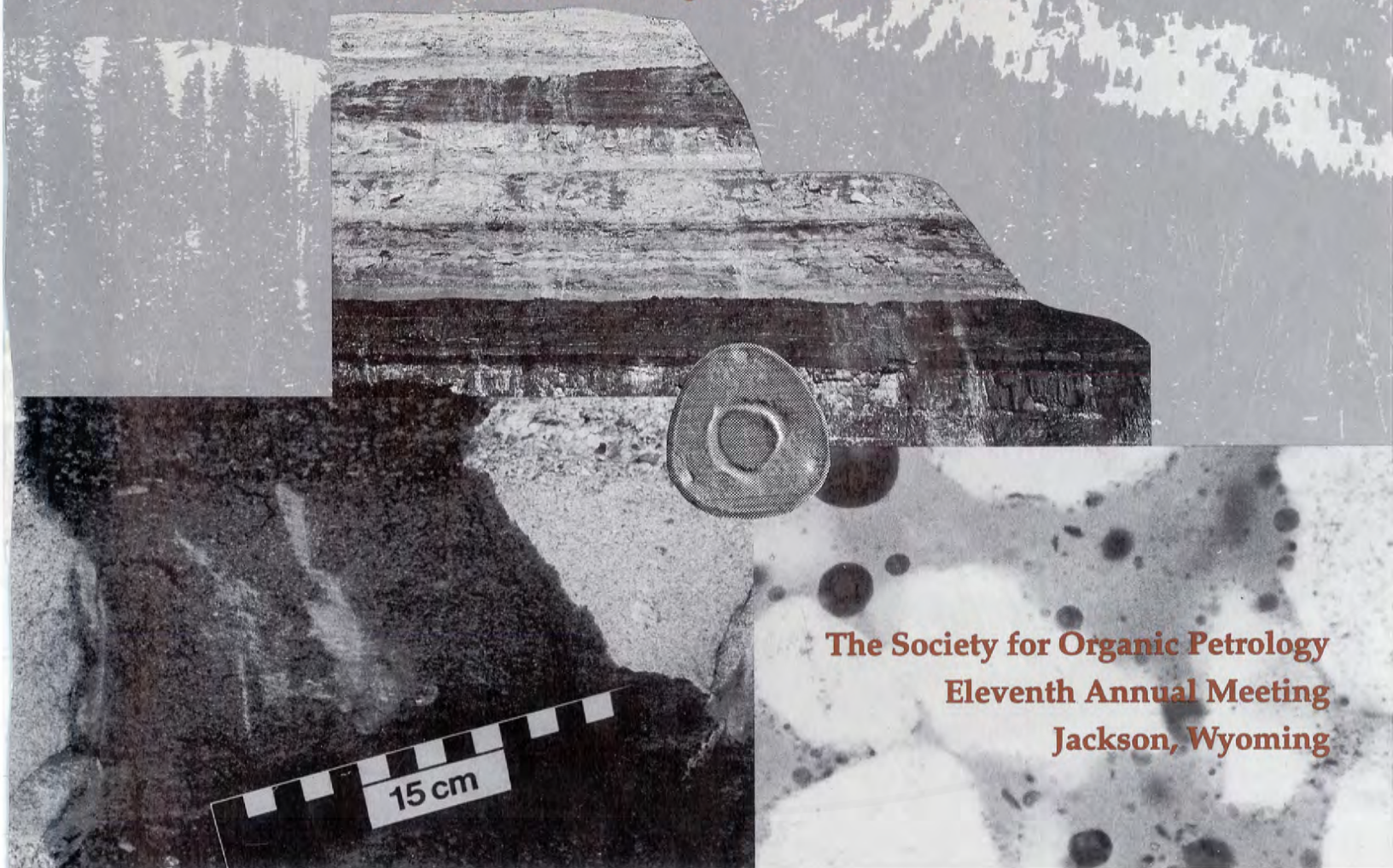
WYOMING STATE GEOLOGICAL SURVEY

Gary B. Glass, State Geologist

ORGANICS AND THE ROCKIES FIELD GUIDE

Edited by

R.M. Flores, K.T. Mehring, R.W. Jones, and T.L. Beck



**The Society for Organic Petrology
Eleventh Annual Meeting
Jackson, Wyoming**

Public Information Circular No. 33

1994

Laramie, Wyoming

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Front Cover: A collage depicting the theme "Organics and the Rockies". The "Rockies" are represented by the Tetons (courtesy of the Wyoming Department of Commerce, Division of Tourism and State Marketing). The "Organics" are represented by the Fort Union Mayfield coal beds in the Grass Creek mine of Northwestern Resources, Inc. in the Bighorn Basin (courtesy of S.B. Roberts), the oil-saturated sandstone and its petrographic thin section showing oil globules of the Fort Union Formation in the Wind River Basin (courtesy of C.W. Keighin), and the Paleocene pollen species *Caryapollenites veripites* (courtesy of D.J. Nichols).

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ORGANICS AND THE ROCKIES

FIELD GUIDE

Edited by

R.M. Flores¹, K.T. Mehring², R.W. Jones², and T.L. Beck²

¹ U.S. Geological Survey

² Wyoming State Geological Survey

*Prepared for The Society for Organic Petrology, Eleventh Annual Meeting, Jackson, Wyoming.
Post-meeting field trip September 25-27, 1994.*



Laramie, Wyoming

1994

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Preface

R.W. Stanton

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Reston, Virginia 22092

The Tertiary basins in the Rocky Mountain area contain vast amounts of coal and petroleum that continue to receive the attention of both researchers and industry. Efforts by the U.S. Geological Survey and the Wyoming State Geological Survey have historically been key to the understanding of the occurrences of gas, petroleum, and coal in these basins. Wyoming has been the leading coal producer in the U.S. since 1988; more than 80 percent of the State's production is from the Wyodak coal bed in the Powder River Basin. Coal is second only to crude oil as the State's most valuable commodity.

This field guide was prepared for a 3-day field trip following the Eleventh Annual Meeting of The Society for Organic Petrology (TSOP) in Jackson, Wyoming, September 25-27, 1994. Members of TSOP engage in microscopic and geochemical characterization of organics, both dispersed in sediments and concentrated in coal. Having the annual meeting in Jackson, Wyoming, afforded TSOP the advantage of a unique opportunity to visit three of the major Tertiary basins in the Rocky Mountain area and view, first-

hand, classic rock types important to petroleum, gas, and coal occurrences in these basins.

Romeo Flores and colleagues assembled a tour of these basins which balances the interests of TSOP members engaged in petroleum exploration as well as those in coal geology. Papers included in this field guide were contributed by members of various research and field teams that have collaborated with Flores. These teams have been able to reconstruct the development of the Tertiary basins as they related to tectonics, sedimentation, and the formation, quality and type of the various organic deposits.

The Society of Organic Petrology is grateful to Romeo Flores of the U.S. Geological Survey; Kim Mehring of the Wyoming State Geological Survey; members of the U.S. Geological Survey; Mr. Gary Glass, the Wyoming State Geologist; Tim Moore, of the Coal Research Association of New Zealand; and Dan Vogler and other staff of the Wyoming State Geological Survey for their support and contribution to this field guide.

Ron Stanton
Chair, TSOP 1994 Annual Meeting

Paleocene paleogeography of the Wind River, Bighorn, and Powder River Basins, Wyoming

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Introduction

The Wind River, Bighorn, and Powder River Basins in Wyoming are Laramide structural basins bounded by a system of interconnected, diverging and converging, anastomosing arches (Erslev, 1993; **Figure 1**). The Wind River Basin is bounded on the west and southwest by the northwest-trending Teton and Wind River arches, and on the south by the west-trending Sweetwater arch (including the Granite Mountains). These interconnected arches join the northwest-trending Casper arch and west-trending Owl Creek arch (including the Washakie Range), which define the eastern and northern margins of the Wind River Basin. The Bighorn Basin is bounded on the south by the Owl Creek arch, which converges with the northwest-trending Beartooth arch along the western margin of the basin. The northwest-trending Bighorn arch, which merges southward with the Casper arch, bounds the Bighorn Basin to the east. The Powder River Basin, in turn, is bounded on the west and southwest by the Bighorn and Casper arches. The Casper arch merges with the northern end of the Laramie and Medicine Bow arches, which bound the Powder River Basin to the south. The north-trending Black Hills arch, which merges with the Laramie arch, bounds the basin on the east.

These arches underwent northeast-southwest compression and shortening, which may have resulted in the development of isolated northwest-trending arches. Subsequent oblique slip resulted in the merging (interconnection) of individual arches (Paylor, 1992). The continuous, anastomosing pattern

of Laramide arches suggests collective deformation by forward and backward thrusting (**Figure 2**). Erslev (1993) suggests that the arches were upper crustal, wedge-shaped popups that developed along bounding back-thrust faults (northeast-dipping) and forward thrust faults (southwest-dipping). He also suggests that these thrust faults were connected to a low-angle master thrust fault in the lower crust. For example, the Wind River and Owl Creek arches (see **Figure 2**) are interpreted to have formed above back-thrust faults extending from a lower-crust detachment.

An interesting test of Erslev's (1993) hypothesis lies in the origin and consequent structural evolution of the intervening intermontane basins. If the Wind River, Sweetwater, Casper, Owl Creek, and Bighorn arches were formed above back-thrust faults, the Wind River, Bighorn, and Powder River Basins could be considered as types of piggy-back basins. The sedimentary history of these basins during the Paleocene should, therefore, shed some light on the evolution and timing of deformation of the intervening arches. This paper will test the hypothesis of Erslev (1993) using the sedimentological record and palynostratigraphic zonation established in the Paleocene Fort Union Formation in the Wind River, Bighorn, and Powder River Basins. In addition, paleogeographic reconstructions are developed for these basins using these palynostratigraphic zonations for correlation.

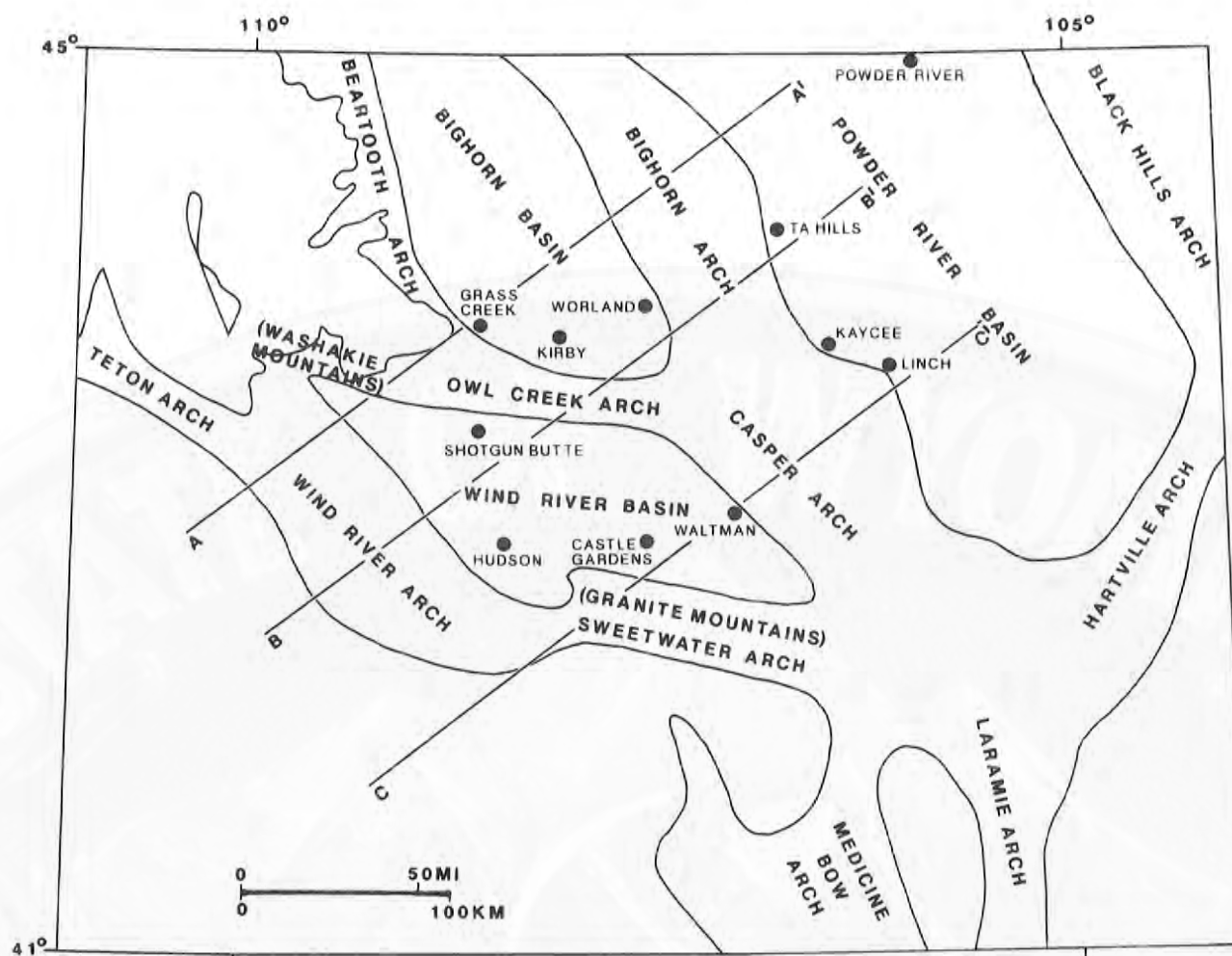


Figure 1. A structural map showing the anastomosed Laramide arches and intervening Tertiary sedimentary basins. Northeast-southwest lines are cross sections shown in Figure 2. Modified from Erslev (1993).

Fort Union correlations

Vertically successive lithofacies intervals in the Paleocene Fort Union Formation in the Wind River, Bighorn, and Powder River Basins were correlated by palynostratigraphic zonation by Nichols (see this volume). This palynostratigraphic zonation provided a mechanism for interbasinal, temporal correlation of rock successions in the Fort Union Formation. The Fort Union Formation in these basins is characterized by palynomorph biozones defined as P1-P2 (lower Paleocene), P3-P4 (middle Paleocene), and P5-P6 (upper Paleocene) (Nichols and Ott, 1978; Pocknall, 1987; Nichols and Flores, 1993).

Lower Paleocene interval

Wind River Basin

The lower Paleocene part of the Fort Union Formation in the Wind River Basin is well represented in strata extending from the western margin of the basin (Shotgun Butte) eastward to the central part of the basin (Waltman area; Figure 3). The Shotgun Butte area is east of the northern Wind River arch and south-east of the Washakie Mountains-western Owl Creek arch. The Hudson area is east of the southern Wind River arch (Figures 1 and 2).

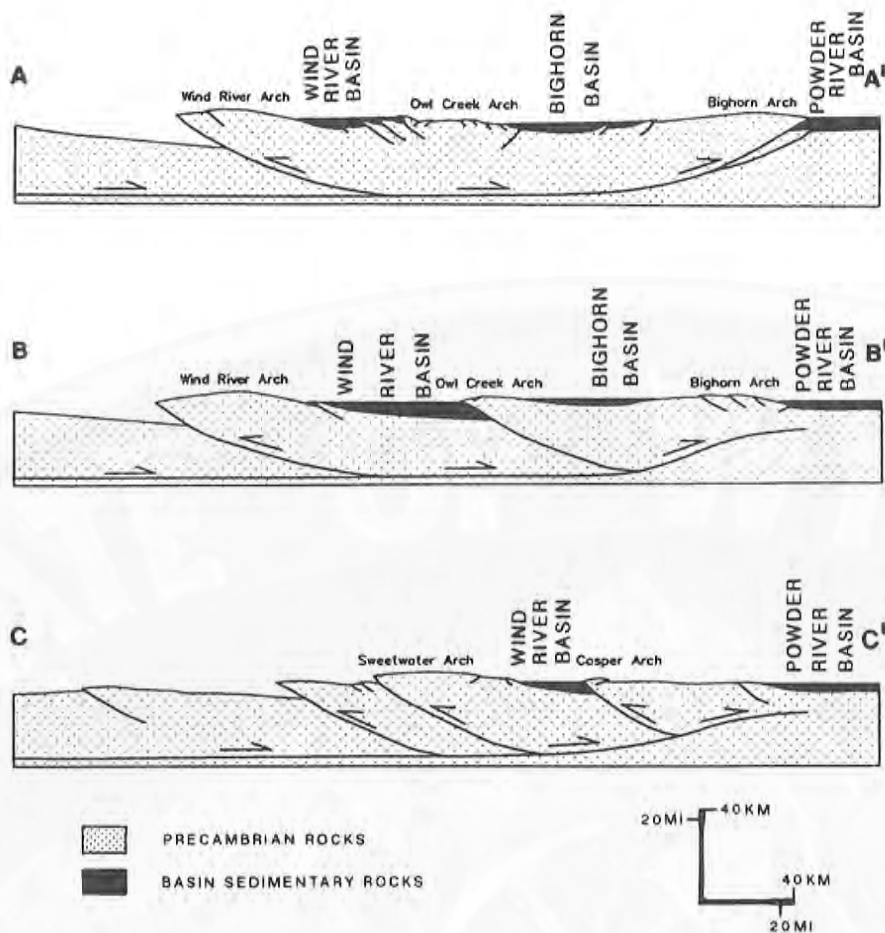


Figure 2. Cross sections of Precambrian basement surface through the Wind River-Owl Creek-Bighorn arches and Sweetwater-Casper arches. Modified from Erslev (1993).

At Shotgun Butte, lower Paleocene strata (see Figures 1 and 3) consist of conglomeratic fluvial channel sandstone and interbedded fluvial channel non-conglomeratic sandstone, floodplain mudstone, siltstone, and silty sandstone, and thin coals and carbonaceous shales representative of short-lived, floodbasin mires (Flores and Keighin, 1992, 1993; Nichols and Flores, 1993). The conglomerates consist of subrounded to rounded black chert, quartzite, porcellanite, and sandstone cobbles and pebbles (Keefer and Troyer, 1964; Keefer, 1965; Flores and Keighin, 1992, 1993). Trough crossbeds in the channel sandstones indicate southeast-oriented paleoflow and clasts in the conglomerates show west-northwest dip imbrication (Flores and Keighin, 1992, 1993).

The Hudson lower Paleocene part (see Figures 1 and 3) of the Fort Union Formation is composed of amalgamated fluvial channel conglomerates and conglomeratic sandstones interbedded with subordinate floodplain mudstones and siltstones (Nichols and Flores, 1993; Flores and others, 1993). The conglomer-

ates are composed of subrounded to rounded chert, limestone, sandstone, and shale boulders, cobbles, and pebbles (Flores and others, 1993). Rare granite cobbles and pebbles are present. Trough crossbeds and clast imbrication in the channel sandstones and conglomerates suggest paleoflow orientations to the east and west, respectively (Flores and Keighin, 1992, 1993; Flores and others, 1993). The Hudson lower Paleocene interval unconformably overlies the Upper Cretaceous Mesaverde Formation, which is palynostratigraphically dated to be Campanian in age (Nichols and Flores, 1993; Flores and others, 1993).

These lithofacies pass basinward (see Figures 1 and 3) in the Castle Gardens area (north of the Granite Mountains-Sweetwater arch) into stacked, amalgamated, fine- to medium-grained fluvial channel sandstones that span the Cretaceous-Tertiary (K-T) boundary and grade up-

ward into fine-grained floodplain mudstones, siltstones, and silty sandstones interbedded with subordinate fine-grained, fluvial channel sandstones and thin coals and carbonaceous shales which formed in topogenous mires. Paleosols (inceptisols) are locally well developed (Flores and others, 1992; Flores and Keighin, 1993; Caudron and others, 1994). Trough crossbeds in the channel sandstones that span the K-T boundary suggest paleoflow orientations to the east-northeast (Flores and others, 1992). The stacked channel sandstones are underlain by a coal bed dated as Maastrichtian in age and overlain by a coal bed dated as early Paleocene (biozone P1) in age.

Farther to the northeast (see Figures 1 and 3), the lower Paleocene interval in the Waltman area (basin center; west of the Casper arch and southeast of the eastern Owl Creek arch) consists of interbedded fine-grained, fluvial channel sandstones, floodplain mudstones, siltstones, silty sandstones, and carbonaceous shales. The fluvial channel sandstones include northeast-oriented trough crossbeds (Nichols and

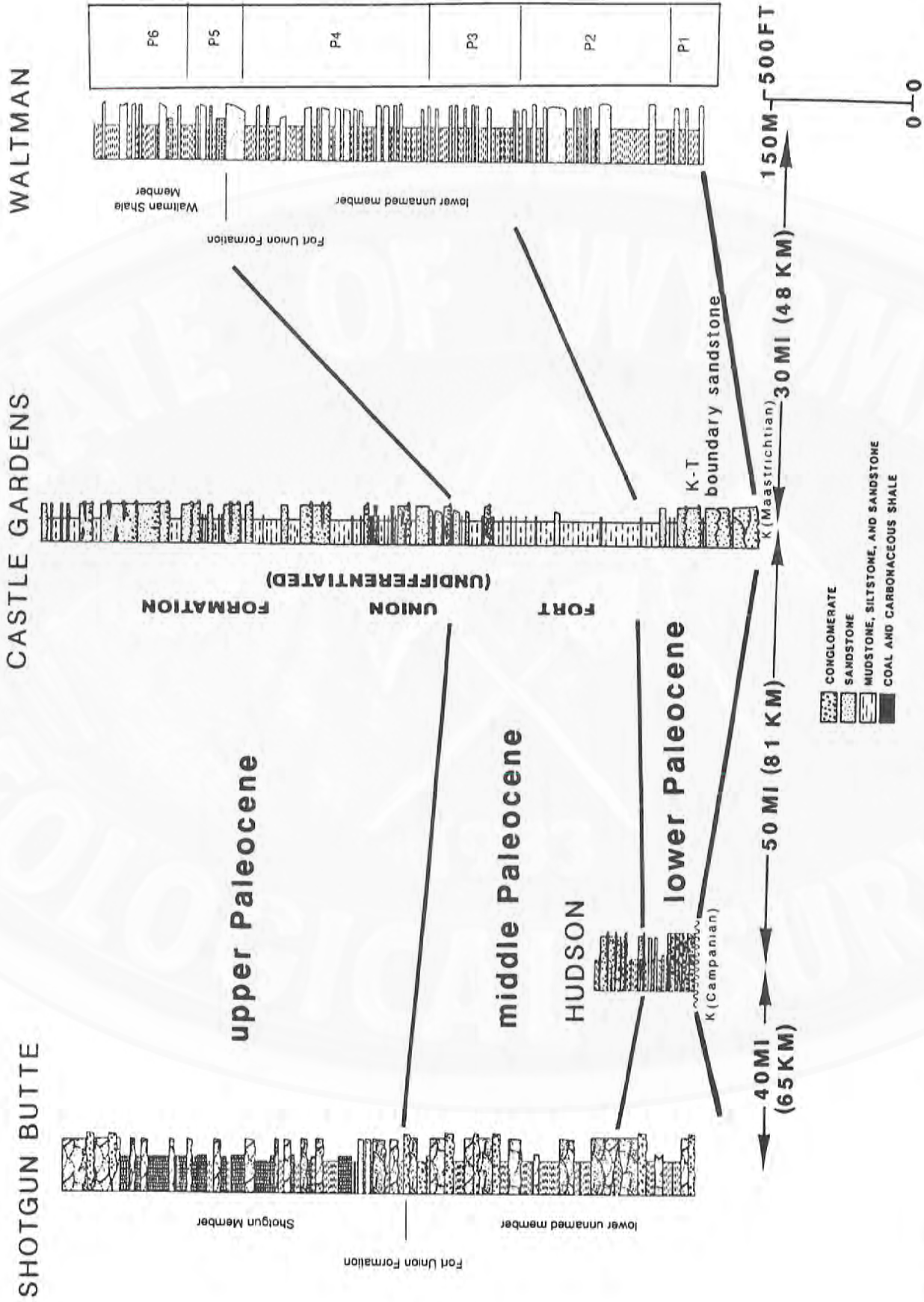


Figure 3. Lithostratigraphic and palynostratigraphic correlation of the Paleocene Fort Union Formation in the Wind River Basin. Palynostratigraphic zones adopted from Nichols (this volume).

Flores, 1993). Subsurface cross sections by Finn (in preparation, a, b) 6 mi (10 km) south of the Waltman area show well-developed stacks of fining-upward sandstone interpreted as amalgamated fluvial channel deposits that are stratigraphically equivalent to the lowermost part of the P1-P2 chrono-succession in the Waltman area. This part of the chrono-succession is overlain by thin, narrow (ribbon) fluvial channel sandstones interbedded by abundant floodplain mudstones. The lowermost amalgamated fluvial channel sandstones are probably a depositional-dip equivalent of the K/T amalgamated fluvial channel sandstones in the Castle Gardens area.

Bighorn Basin

The lower Paleocene part (Figure 4) of the Fort Union Formation in the southern Bighorn Basin is represented by a basin-margin succession in the Grass Creek-Red Ridge area (east and north of the Washakie Mountains-western Owl Creek arch) and a more basin-axial lithofacies in the Kirby-Worland area (north of the Owl Creek arch and southwest of the Bighorn arch; see Figure 1). The basin margin lithofacies includes by quartzite-rich conglomerate, fine- to coarse-grained, single- and multi-storied fluvial channel sandstones and interbedded floodplain mudstone, siltstone, and silty sandstone. Associated thin coals and carbonaceous shale beds are interpreted to have formed in topogenous mires. Floodplain deposits are poorly developed in the Grass Creek area, but become more widespread in basinward facies near Red Ridge. Trough crossbed orientation in the channel sandstones indicates east- and northeast-directed paleoflow in the Grass Creek area (Roberts and Bossiroy, in press), and more easterly directed paleoflow at Red Ridge.

In the Kirby-Worland area, the lower Paleocene interval (see Figures 1 and 4) is represented by fine- to coarse-grained fluvial channel sandstones interbedded with floodplain mudstones, siltstones, and silty sandstones; the latter lithofacies are more abundant than in the Grass Creek area. Sandstones are present as thin to thick, isolated and vertically amalgamated sandstone bodies. Thin coals and carbonaceous shales interbedded with the floodplain sediments are interpreted to have formed in short-lived, topogenous mires. Trough crossbed orientation in the channel sandstones suggests paleoflow directions to the east and northeast.

Powder River Basin

The lower Paleocene part of the Fort Union Formation in the Powder River Basin is best represented

near the southwest margin in the Lynch-Kaycee area (north of the Casper arch and southeast of the Bighorn arch; see Figures 1 and 5). The strata consists of stacked, amalgamated fluvial channel sandstones in the lower part, which are probably the depositional-dip equivalent of the K-T and lower Paleocene fluvial channel sandstones at Castle Gardens and in areas south of Waltman in the Wind River Basin. The upper part of the succession is composed of thin fluvial channel sandstones interbedded with abundant floodplain mudstones and siltstones, paleosols (alfisols and spodosols), and a few thin coals and carbonaceous shales that formed in topogenous mires (Coss and Flores, 1984; Flores and others, 1984; Coss, 1985; Caudron and others, 1994). Trough crossbed foresets in the channel sandstones are oriented to the northeast (Coss, 1985; Flores, 1986; Seeland and others, 1989).

Middle Paleocene interval

Wind River Basin

The middle Paleocene part of the Fort Union Formation near the western margin of the Wind River Basin is shown in Figure 3. At Shotgun Butte, middle Paleocene strata along the northwest margin of the basin (east of the northern Wind River arch and southeast of the Washakie-western Owl Creek arch; see Figure 1) are composed of conglomerate-dominated fluvial channel sandstone interbedded with floodplain mudstone, siltstone, and crevasse-splay silty sandstone and thin coaly carbonaceous shales (Flores and Keighin, 1992, 1993). The conglomerates consist of subrounded to rounded black chert, siliceous shale, quartzite, and sandstone cobbles and pebbles (Keefer and Troyer, 1964; Flores and Keighin, 1992, 1993). Trough crossbeds in the channel sandstones indicate east-southeast paleoflow orientations; clast imbrication in the conglomerates exhibits west-northwest dips (Flores and Keighin, 1992, 1993).

In the Hudson area (east of the southern Wind River arch; see Figure 1), the middle Paleocene interval is represented by conglomeratic, fluvial channel sandstones interbedded with floodplain mudstones and siltstones and some thin, coaly carbonaceous shales (see Figure 3; Flores and others, 1993). The conglomerates are mainly composed of chert, limestone, and sandstone pebbles with minor granitic clasts (Flores and others, 1993). The channel sandstones are feldspathic and arkosic in composition. Trough crossbeds in the channel sandstones indicate east-northeast paleoflow orientations (Flores and others, 1993).

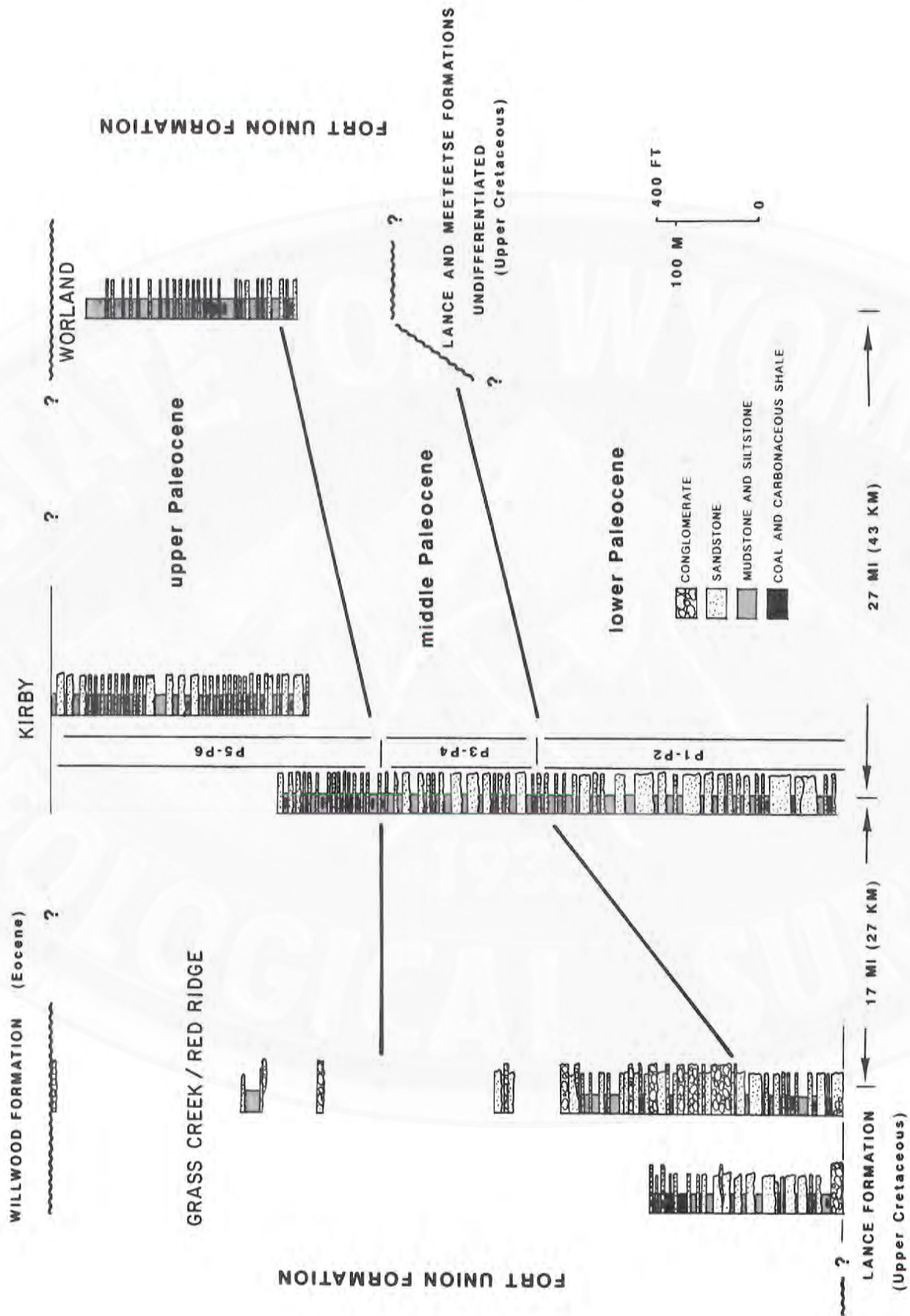


Figure 4. Lithostratigraphic and palynostratigraphic correlation of the Paleocene Fort Union Formation in the Bighorn Basin. Palynostratigraphic zones adopted from Nichols (this volume). Unconformity at the base of the Fort Union Formation (Worland area) inferred from Bown, 1979.

Basinward, middle Paleocene strata in the Castle Gardens area (north of the Granite Mountains and Sweetwater arch; see Figures 1 and 3) consist of floodplain-dominated mudstones and siltstones interbedded with crevasse-splay silty sandstones, fluvial channel sandstones and conglomeratic sandstones (Flores and others, 1992). The mudstones were derived from the Cretaceous marine shales based on montmorillonite clay mineral signatures (Flores and others, 1994; Caudron and others, 1994). These lithofacies are interbedded with thin coal and carbonaceous shales that formed in topogenous mires. Paleosols (inceptisols and vertisols) are well developed (Caudron and others, 1994). The conglomerates consist of rounded black chert, limestone, and siliceous shale pebbles (Keefer, 1965; Flores and others, 1992). Rare, subrounded granitic clasts are present.

At the basin center in the Waltman area (west of Casper arch and south of the eastern Owl Creek arch; see Figure 1), the middle Paleocene section consists of thin fluvial channel sandstones interbedded with abundant floodplain mudstones and siltstones (see Figure 3; Nichols and Flores, 1993). The mudstones dominate the uppermost part of the P3-P4 chronosuccession along the northeastern part of the Wind River Basin. South of the Waltman area, Finn (in preparation, a, b) shows a stratigraphically equivalent succession of fining- and thinning-upward sandstones with interbedded mudstones and siltstones that become increasingly abundant upward. These rocks represent fluvial channel and floodplain deposits; the floodplain deposits are more prominent in the upper part of the section.

Bighorn Basin

The middle part of the Fort Union Formation in the Bighorn Basin is shown in Figure 4. The basin margin lithofacies in the Grass Creek-Red Ridge area (west and north of the Washakie Mountains-western Owl Creek arch; see Figure 1) consists of fluvial channel conglomerates and conglomeratic sandstones interbedded with floodplain mudstones and siltstones, and thick, lenticular coal beds (Roberts and Stanton, this volume). The conglomerates consist of subrounded and rounded quartzite cobbles and pebbles and subordinate fossiliferous chert (Hewett, 1926). Trough crossbeds in the channel sandstones and conglomerates indicate primarily northeast paleoflow orientations.

Toward the basin axis in the Kirby-Worland area (north of the eastern Owl Creek arch and southwest of the Bighorn arch; see Figure 1), the middle Paleocene section (see Figure 4) is composed of medium- to

coarse-grained fluvial channel sandstones, tabular, crevasse-splay sandstones, floodplain mudstones and siltstones, and thin coals and carbonaceous shales (Nichols and others, this volume). Floodplain deposits are more abundant than in the lower Paleocene section. Fluvial channel sandstones occur as single or multi-storied sandstone bodies, and trough crossbeds in the fluvial channel sandstones indicate east-northeasterly paleoflow orientations.

Powder River Basin

The middle Paleocene part of the Fort Union Formation in the Powder River Basin is best represented in the Lynch-Kaycee area near the southwest margin of the basin (north of the Casper arch and southeast of Bighorn arch; see Figures 1 and 5). Here, the section is composed of fluvial channel conglomerates and sandstones interbedded with abundant floodplain mudstones and siltstones, thin coals and carbonaceous shales (topogenous mires), and paleosols (inceptisols and vertisols) (Flores and others, 1984; Coss, 1985; Flores, 1986; Caudron and others, 1994). The conglomerates include fossiliferous chert, quartzite, quartz arenite, mudstone, siltstone, and petrified wood cobbles and pebbles (Coss, 1985). Trough crossbeds indicate paleoflow to the northeast (Coss, 1985; Flores, 1986; Seeland and others, 1989).

The middle Paleocene part of the Fort Union Formation in the central part of the Powder River Basin (Powder River area, west-northwest of the Black Hills arch; Figures 1 and 5) includes fluvial channel sandstones, crevasse-splay sandstones, floodplain mudstones and siltstones, and thin to thick coals and carbonaceous shales that formed in topogenous and raised mires (Flores, 1979a, 1979b, 1981). Trough crossbeds in the channel sandstones indicate northeast paleoflow orientations (Flores, 1986; Seeland and others, 1989).

Upper Paleocene interval

Wind River Basin

The upper Paleocene part of the Fort Union Formation in the Wind River Basin is shown in Figure 3. Near the western basin margin, the section in the Shotgun Butte area (east of the northern Wind River arch and south of the Washakie Mountains-western Owl Creek arch; see Figure 1) is dominated by floodplain mudstones and siltstones interbedded with subordinate fluvial channel sandstones and crevasse-splay sandstones. Interbedded thin to thick coals and carbonaceous shales formed in topogenous mires (Flores and Keighin, 1993). Trough crossbeds in the channel

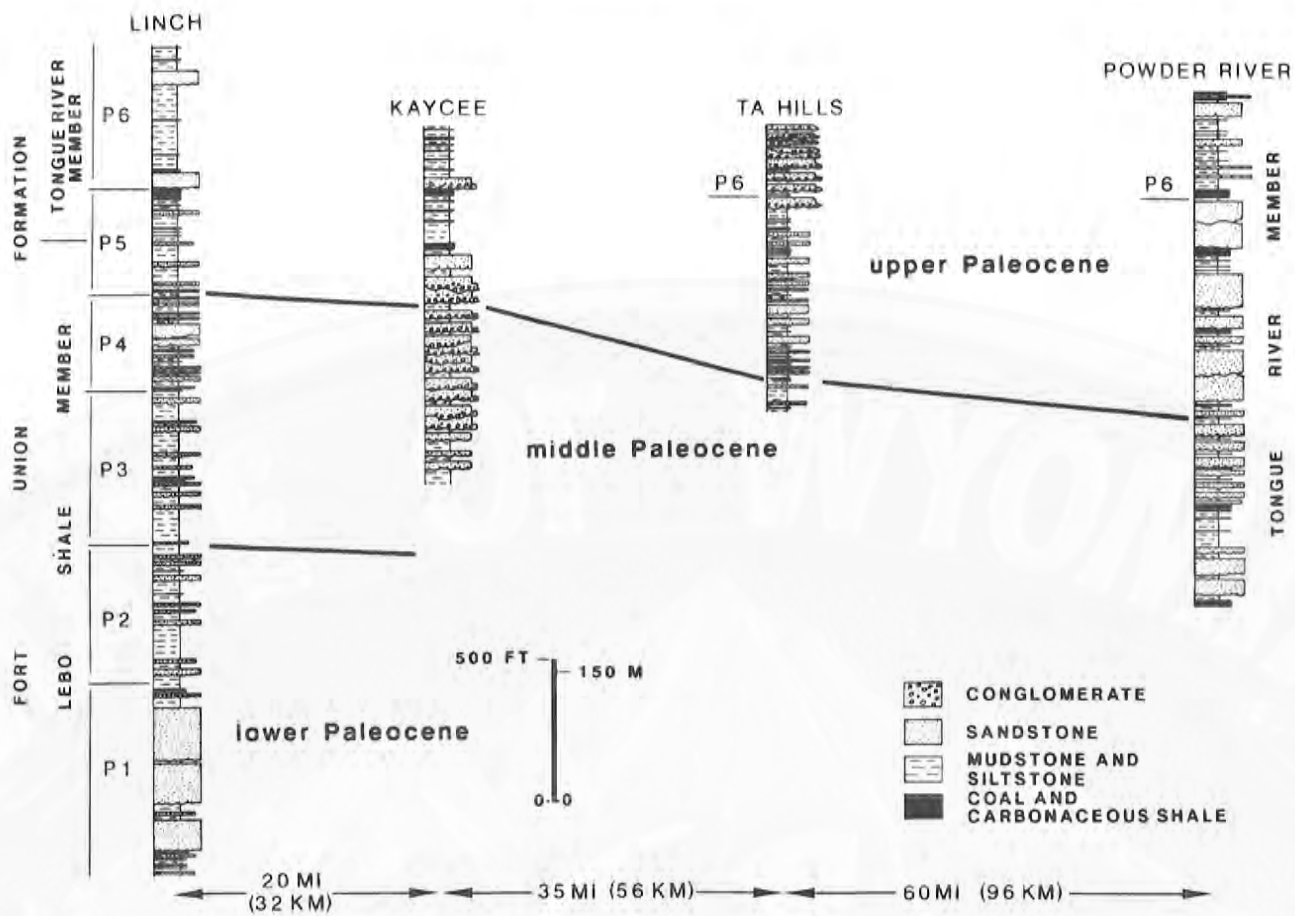


Figure 5. Lithostratigraphic and palynostratigraphic correlation of the Paleocene Fort Union Formation in the Powder River Basin. Palynostratigraphic zones adopted from Nichols (this volume).

sandstones suggest east-northeast paleoflow orientations (Keefer and others, 1993).

Basinward, in the Castle Gardens area (north of the Granite Mountains and Sweetwater arch; see Figure 1), the upper Paleocene section (see Figure 3) consists of conglomeratic fluvial channel sandstones, and crevasse-splay sandstones commonly interbedded with floodplain mudstones, siltstones, and silty sandstones. Thin to thick coals and carbonaceous shales, interbedded with floodplain strata, formed in topogenous mires (Flores and others, 1992). Trough crossbeds indicate a northeast paleoflow orientation (Flores and others, 1992).

Near the basin center in the Waltman area (west of the Casper arch and south of the eastern Owl Creek arch; see Figure 1) the upper Paleocene section (see Figure 3) is composed of mainly of lacustrine mudstones and shales interbedded with delta-front sandstones (Keefer, 1965; Phillips, 1983; Flores and others, 1990). South and west of the Waltman area, this section or interval is a mudstone-rich lake deposit

(Waltman Lake; Keefer, 1965) greater than 3,000 ft (915 m) thick. The lowermost part of the Waltman Lake deposit probably correlates to the biozone P4 (D.J. Nichols, verbal communication, 1994; Nichols, this volume).

Bighorn Basin

The upper Paleocene part of the Fort Union Formation in the Bighorn Basin is shown in Figure 4. In the Kirby-Worland area (north of the eastern Owl Creek arch and southwest of the Bighorn arch; see Figure 1), the section includes recurrent (cyclic) coarsening-upward successions of tabular, fine-grained, splay/levee sandstones capped by fluvial channel sandstones. Interbedded floodplain deposits consist of rooted mudstone, siltstone and silty sandstones, and thin coal and carbonaceous shale beds. Trough crossbeds in thicker channel sandstones indicate north to northeast paleoflow orientations.

East of Worland, the upper Paleocene section (see Figure 4) passes into a fine-grained-dominated inter-

val consisting of thin fluvial channel sandstones encased in abundant floodplain mudstones and siltstones. Interbedded coals and carbonaceous shales are thin and lenticular. Trough crossbeds in thicker fluvial channel sandstones indicate paleoflow orientations to the east-northeast. Eastward, in the Honeycombs area, this part of the Fort Union Formation is represented by an interval of mudrock-dominated strata 100 ft (30 m) thick or less. An unconformity separates upper middle and upper Paleocene strata from underlying Upper Cretaceous deposits in this area of the basin (Bown, 1979).

Powder River Basin

The upper Paleocene part of the Fort Union Formation in the Powder River Basin is shown in Figure 5. The section in the Linch-Kaycee area near the southwest margin of the basin (north of the Casper arch; see Figure 1) consists of conglomeratic fluvial sandstones interbedded with floodplain mudstones and siltstones. Thin to thick coal and carbonaceous shale beds formed in raised mires (Flores and others, 1984; Coss, 1985; Flores, 1986). The conglomerates contain chert, quartzite, limestone, siltstone, and shale cobbles and pebbles (Flores and others, 1984; Coss, 1985). The trough crossbeds in the channel sandstones indicate north-northeast paleoflow orientations (Coss, 1985; Flores, 1986; Seeland and others, 1989).

The upper Paleocene section along the western margin of the Powder River Basin (east of the Bighorn arch) is shown in Figure 5. The section comprises conglomeratic fluvial channel sandstones interbedded with abundant floodplain mudstones and siltstones; thin coal and carbonaceous shale beds are present in this section (Weaver and Flores, 1985, 1987; Flores, 1986). The conglomeratic channel sandstones are within P6 biozone, and the conglomerates consist of chert, limestone, quartz arenite, and silicified siltstone and mudstone cobbles and pebbles (Weaver and Flores, 1985, 1987). Trough crossbeds in the channel sandstones exhibit easterly paleoflow orientations (Flores, 1986).

Near the basin center in the Powder River area (west-northwest of the Black Hills arch; see Figure 1), the upper Paleocene part (see Figure 5) of the Fort Union Formation consists of fluvial channel sandstones and crevasse-splay sandstones commonly interbedded with floodplain lake mudstones and siltstones. Interbedded thick coals formed in raised mires (Flores, 1979a, b; Flores, 1981; Warwick and Stanton, 1988). Trough crossbeds in the channel sandstones indicate a northeasterly paleoflow orientation (Flores, 1986; Seeland and others, 1989).

Paleogeographic interpretations

Correlation of the three Paleocene intervals of the Fort Union from the Wind River Basin through the Bighorn and Powder River Basins permits temporal unfolding of the depositional systems in these basins from early to late Paleocene (Figures 6, 7, and 8). More importantly, the changes in the lithological and petrographical characteristics (e.g., grain size, mineralogy, crossbed orientations, and lithofacies) of Fort Union sediments deposited in these depositional systems provide clues as to the timing of deformation and uplift in the surrounding arches.

Early Paleocene interbasinal depositional systems and tectonic arches

During early Paleocene time (biozones P1-P2), the western margin of the Wind River Basin (Figure 6) was drained by two major fluvial systems; (1) a braided-meandering fluvial system in the Shotgun

Butte area (east of the northern Wind River arch and south of the Washakie Mountains-western Owl Creek arch), and (2) an alluvial fan-braided fluvial system in the Hudson area (east of the southern Wind River arch) (Flores and Keighin, 1992, 1993; Flores and others, 1993). East-southeast sediment transport in the braided-meandering fluvial system, which drained the Shotgun Butte synclinal paleovalley (see Figure 6) near the northwest margin of the basin indicates provenance from the northern Wind River arch to the west-northwest. Quartzite clasts in the conglomerates suggest derivation from Proterozoic and Paleozoic quartzites probably exposed in the Wind River-Gros Ventre Range (Hewett, 1926), and east-central Idaho and southern Montana (Ruppel, 1978). Eastward transport of coarse detritus in the alluvial fan-braided fluvial system near the southwest margin of the basin (Hudson area) indicates provenance from the southern Wind River arch to the west. Here, the presence of granitic clasts suggests exposure of the basement core during early Paleocene time. This alluvial fan-braided fluvial system infilled the Hudson-Castle Gardens

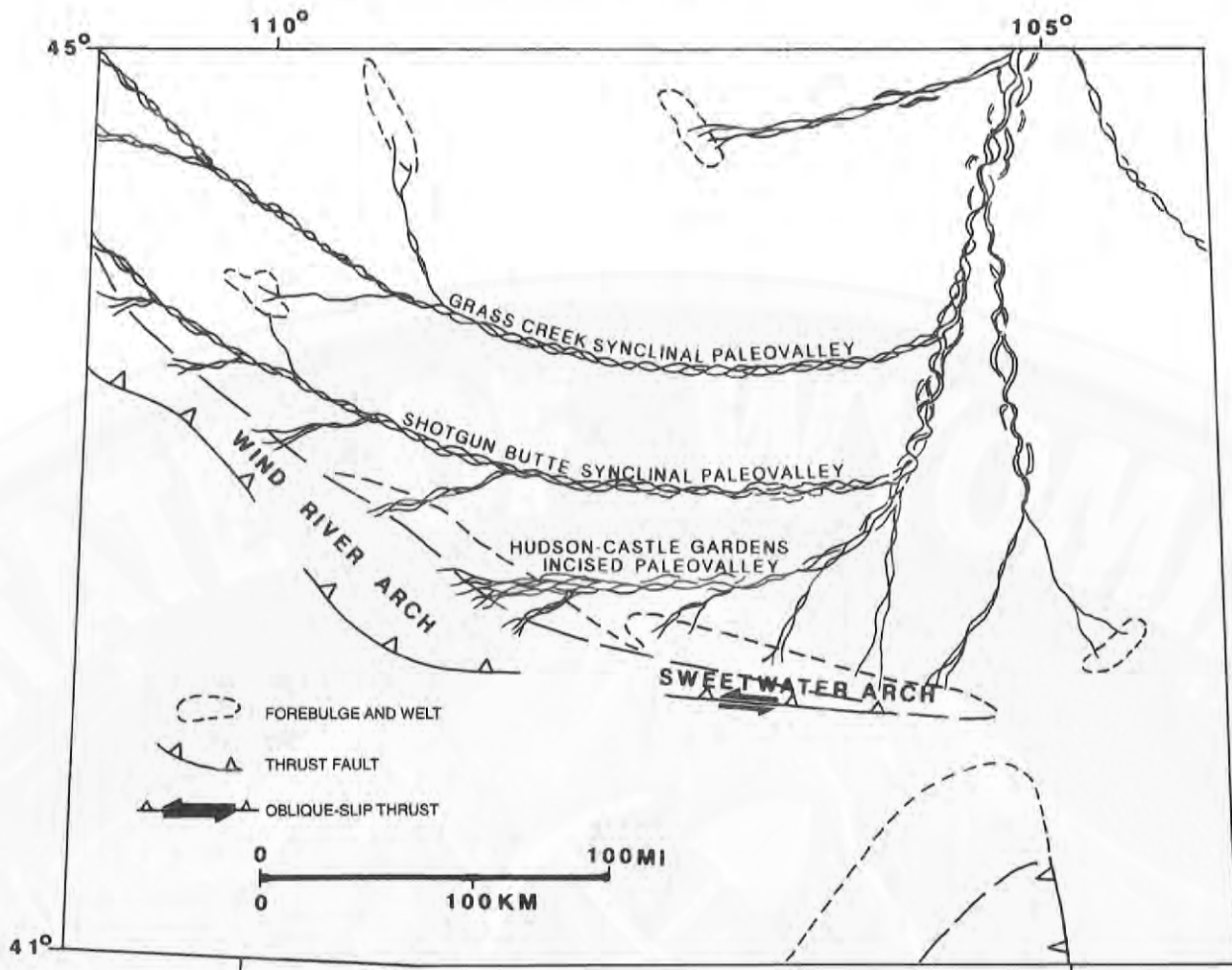


Figure 6. A paleogeographic reconstruction of major paleovalleys during the early Paleocene (biozones P1-P2) showing flow-through fluvial systems in the interconnected Wind River, Bighorn, and Powder River Basins and their paleo-headwaters in the Wind River-Sweetwater arches.

Castle Gardens incised paleovalley (see Figure 6) carved into Upper Cretaceous strata of the Mesaverde Formation (Flores and others, 1993).

The incised paleovalley (see Figure 6) may have extended basinward into the Castle Gardens area where stacked, amalgamated fluvial channel sandstones at the K-T boundary were deposited in a low sinuosity meandering fluvial system (Flores and others, 1992). This fluvial system probably flowed through the Casper arch as indicated by the presence of stacked, amalgamated fluvial channel sandstones in areas south of Waltman and east of Castle Gardens. This paleovalley, which was drained by a meandering, flow-through fluvial system, can be traced to the southwest margin of the Powder River Basin where stacked, amalgamated fluvial channel sandstones are well developed in the Lynch-Kaycee area.

Incision of the Hudson-Castle Gardens paleovalley might be related to uplift (back-thrusting of Erslev, 1993) of the Wind River and Sweetwater arches during early Paleocene time; the latter uplift is indicated by a northeast paleoflow direction observed in fluvial sediments in the Castle Gardens area. Degradation of Upper Cretaceous rocks (Campanian through Maastrichtian) towards the uplifts created an area where net erosion was greater than net deposition (Flores and others, 1993); this erosional hiatus accompanied a regional lowering of base level. At the same time, in basinward settings where the rate of subsidence was less, fluvial systems aggraded within the paleovalley. In the later part of the early Paleocene, increasing subsidence in the basin created accommodation space along the flanks of the uplift and previous areas of degradation were infilled. As a result, paleovalley deposits overlie increasingly

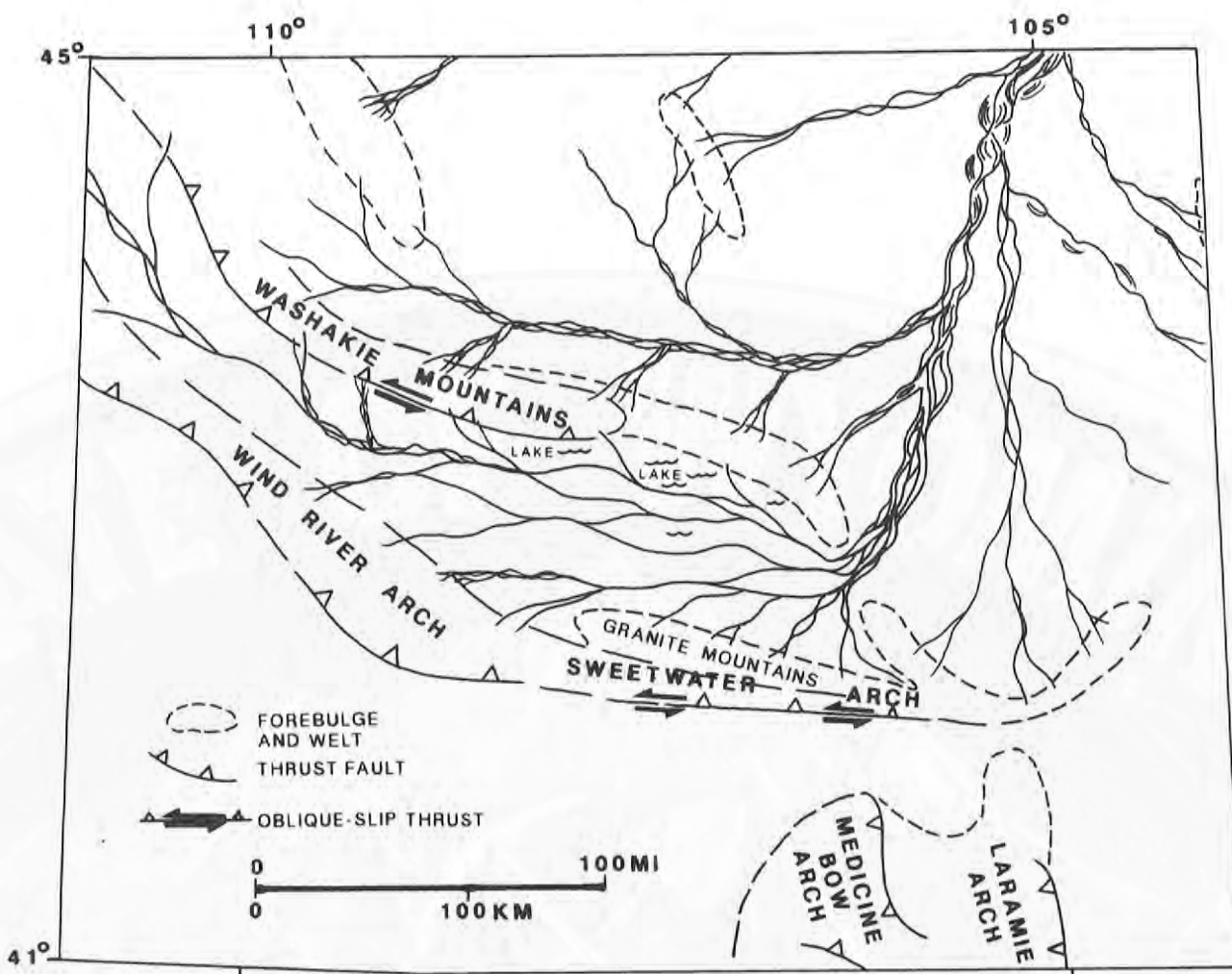


Figure 7. A reconstruction of the middle Paleocene (biozones P3-P4) paleogeography showing an initial structural partitioning of the Wind River, Bighorn, and Powder River Basins by the Washakie-Owl Creek-Casper-Bighorn arches. Flow-through fluvial systems slowly evolved into short-headed fluvial systems and were dammed forming internal basin drainage and lakes.

older rocks towards the paleo-headwaters on the flanks of the ancestral Sweetwater-Wind River arch.

During the latter part of the P1-P2 chronosuccession the Shotgun Butte and Hudson-Castle Gardens paleovalleys were replaced by a series of narrow meandering fluvial channels with associated crevasse-floodplain-mire settings; the latter setting is where coals accumulated (Flores and others, 1992; Johnson and others, this volume). Downstream, these fluvial systems were transformed into less sinuous fluvial channels that were confined by abundant floodplain mudstones derived from initial unroofing of the Cretaceous marine shales. These ribbon, fluvial channels may have been anabranching.

The Shotgun Butte fluvial system and a coeval, east-southeast-flowing braided fluvial system in the

Grass Creek synclinal paleovalley (inherited from a earlier strike-valley fluvial system of Kraus, 1984) near the southwest margin of the Bighorn Basin may have extended eastward and merged with a north-east-flowing trunk fluvial system in the Powder River Basin (see Figure 6). Fluvial aggradation along these synclinal paleovalleys was in response to creation of accommodation space in probably east-west downfolded areas due to crustal shortening. These synclinal paleovalleys are well developed in the Shotgun Butte area where sedimentary rocks of the lower part of the Fort Union Formation have thickened above a syndepositional syncline and thinned over syndepositional anticlines (Keefer and Troyer, 1964; Flores and Keighin, 1993). Thickening and thinning of the fluvial deposits due to downwarping and upwarping suggests confinement of the fluvial system within the downwarp.

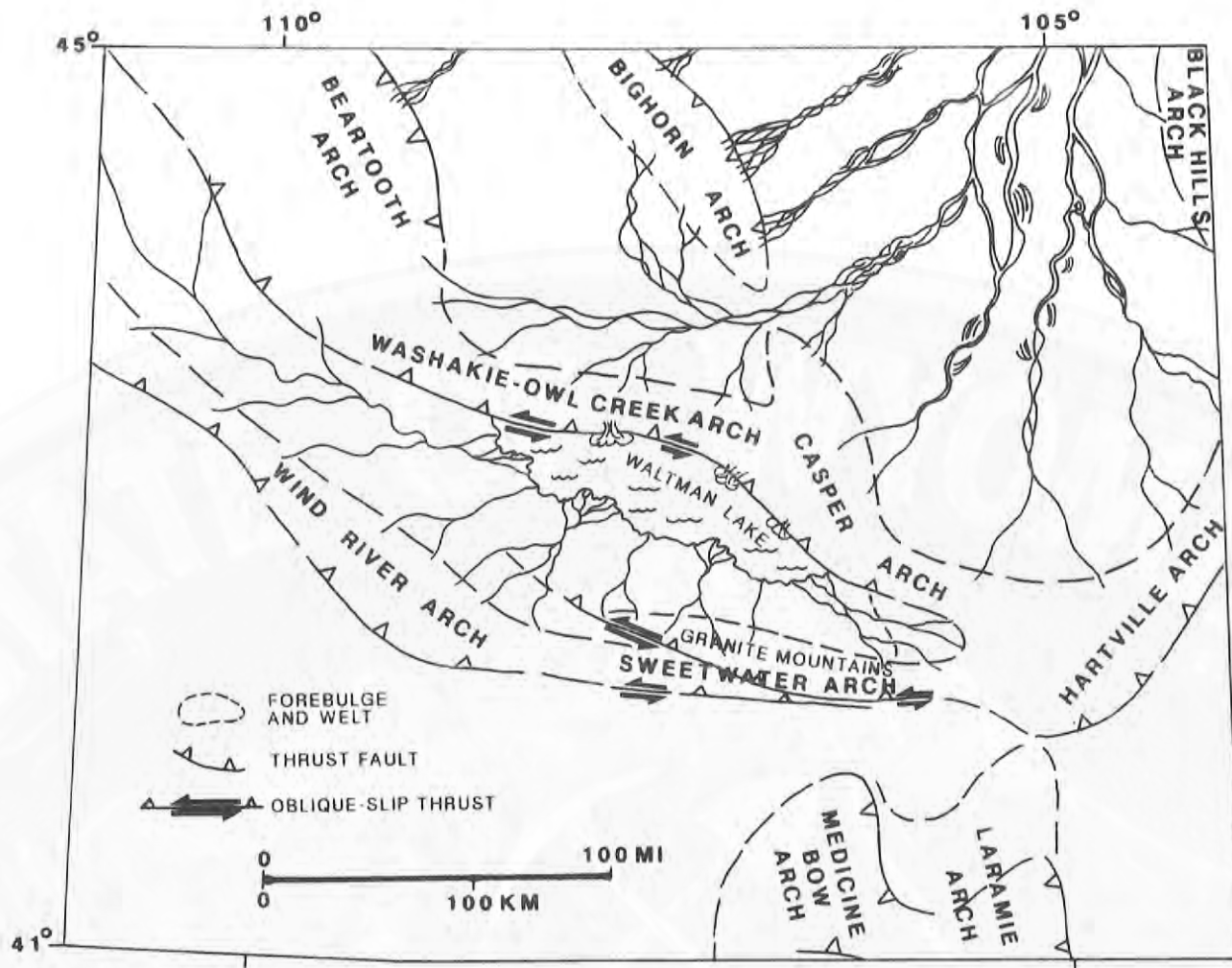


Figure 8. A reconstruction of the late Paleocene (biozones P5-P6) paleogeography showing a complete structural partitioning of the Wind River, Bighorn, and Powder River Basins by the tectonic arches. Fluvial systems completely evolved into internal centripetal drainage and lake systems in structurally closed basins except for the flow-through fluvial system in the Powder River Basin.

Middle Paleocene interbasin depositional systems and tectonic arches

Fluvial systems in the western part of the Wind River Basin (Figure 7) evolved from a braided to an anabranching fluvial system near the northwestern margin of the basin (Shotgun Butte area) and evolved from a braided to meandering-anabranching fluvial system near the southwest basin margin (Hudson area) from early (biozones P1-P2) to middle (biozones P3-P4) Paleocene time (Flores and Keighin, 1992, 1993; Flores and others, 1993). The bedload gravels in the braided streams near the northwest margin of the basin suggest concurrent onset of uplift in the Washakie Mountains-western Owl Creek arch. This uplift may

have been followed by back-thrusting with a oblique-slip component creating a transpressional Wind River Basin (Perry and others, 1994). This interpretation is supported by the presence of abundant gravels and evidence of debris flow in braided stream deposits of early middle Paleocene age (biozone P3) in the southwestern Bighorn Basin (Grass Creek area) north of the arch (Roberts and others, 1994). These braided streams are interpreted to have flowed to the northeast, toward the axis of the basin. Interpreted eastward continuity of the fluvial system through the southern part of the Bighorn Basin suggests that the system may have exited the basin between the Bighorn and Casper arches.

In the Wind River Basin (Figure 7), the basin-margin braided and meandering streams were

gradually dammed as the Casper arch was slowly uplifted; the uplift transformed the flow-through fluvial system into meandering and anabranching fluvial systems, and floodplain lakes due to rapid basin subsidence. Continued stripping of Cretaceous marine shales provided abundant suspended-load sediments of streams, which formed thick floodplain deposits that confined meandering and anabranching, ribbon fluvial channels. These fluvial channels and accompanying floodplain-crevasse-mire settings were common between the margin and central parts of the basin. However, these fluvial channels were transformed into floodplain mudstone-dominated fluvial systems along the central part of the basin, where muds settled in floodplain lakes in response to basin subsidence (Johnson and others, this volume).

Abundant gravel bedload of northeast-flowing braided streams in the Kaycee area (see **Figures 1 and 5**) along the southwest margin of the Powder River Basin indicate initial uplift of the Casper arch in early middle Paleocene; time biozone P3). The chert gravels of these braided streams were probably derived from conglomeratic units of the Cretaceous Frontier and Cloverly Formations (Merewether and others, 1976; May, 1993) as the Casper arch was gradually unroofed. Partial damming of the fluvial systems in the Wind River Basin by the Casper arch due to uplift initiated ponding of fresh water and the development of an anabranching fluvial system and floodbasin lakes ("embryonic" Waltman Lake); the floodplain lakes served as sinks for suspended load of streams derived from Cretaceous marine shales. The granitic cobbles and pebbles in bedload deposits of this fluvial system in the Castle Gardens area during late middle Paleocene time (biozone P4) indicate that the Granite Mountains, north of the Sweetwater arch, were unroofed to the core.

Late Paleocene interbasin depositional systems and tectonic arches

During the late Paleocene (biozones P5-P6) the Wind River Basin was transformed into a closed basin following uplift of the Casper and Owl Creek arches due to back-thrusting (Erslev, 1993). Biozonation of the Waltman Lake deposits as biozones P5-P6 (Nichols and Ott, 1978) and thickening of these deposits to as much as 3,000 ft (915 m) along the northern margin of the Wind River Basin suggests that deposition in the Waltman Lake was caused by the initial back-thrust loading due to dip slip of the Casper arch thrust fault and oblique slip of the Owl Creek thrust fault (Erslev, 1993); this dammed the fluvial systems

during late Paleocene time. Vertical and horizontal movements along these back-thrust faults may have continued into the early Eocene (Keefer, 1965). The mudstones and shales in these lake deposits indicate significant unroofing of Cretaceous marine shales from bounding arches at this time (Flores and others, 1994). Mixed fine- and coarse-grained sediments were eroded, transported, and deposited by short-headed streams, which drained into the lake as fan deltas (Keefer, 1965; Phillips, 1983; Flores and others, 1990). During the late Paleocene, the Granite Mountains were completely unroofed, as indicated by conglomerates containing abundant granitic boulders, cobbles, and pebbles as well as the arkosic composition of conglomeratic channel sandstones in the Castle Gardens area.

A coeval fluvial system in the southern Bighorn Basin flowed to the east-northeast; the northeast component of the crossbeds in channel deposits of this fluvial system suggests provenance to the south and southwest (Washakie Mountains-Owl Creek arch). Closure of the Wind River Basin during the late Paleocene indicates that the Washakie Mountains-Owl Creek arch were continuous with the Casper arch. This fact, coupled with the eastward component of crossbed measurements in channel deposits, suggests that the fluvial system in the southern Bighorn Basin might have flowed into the Powder River Basin through an area between the Casper and Bighorn arches, suggesting a saddle existed in the now-continuous Bighorn arch. This saddle may have been closed during latest Paleocene time.

In the southwestern part of the Powder River Basin, the appearance of conglomeratic channel sandstones containing chert and limestone clasts eroded from Paleozoic rocks and transported by eastward-flowing streams suggests that the Bighorn arch was a source area during late Paleocene time. This interpretation is supported by a petrographic study of late Paleocene age Fort Union channel sandstones in the Powder River Basin (Whipkey and others, 1991), which indicates unroofing of carbonate and chert rock fragments from the Bighorn arch. Late Paleocene age (biozone P6; see **Figure 5**) conglomerates indicate that unroofing resulted from uplift along the eastern margin of the Bighorn arch; this uplift was induced by west-dipping thrust faults (forward thrust faults of Erslev, 1993; Obernayer, 1978; Jenkins, 1986). These synorogenic conglomerates were deposited by alluvial fans along the western margin of the Powder River Basin (Weaver and Flores, 1989). Sequential unroofing of Cretaceous marine shales and sandstones followed by older rocks through Paleozoic

carbonate rocks was documented by Whipkey and others (1991). Unroofing of the granitic core of the Bighorn arch was initiated during early Eocene time (Whipkey and others, 1991). Forward thrusting along the eastern edge of the Bighorn arch may have influenced basin subsidence by thrust loading. This loading and subsidence may have promoted the development of very thick coals in raised mires, which were associated with northeast-flowing anabranching and meandering fluvial systems during late Paleocene time (Flores, 1986; Warwick and Stanton, 1988; Flores and Cross, 1991; Flores, 1993). These axial-trunk fluvial systems were downstream from basin-margin alluvial fans and continuous with flow-through fluvial systems from the southwest margin of the Powder River Basin and from the southern Bighorn Basin. The Bighorn Basin was probably closed at its southeastern margin by merging of the Bighorn and Casper arch in very latest Paleocene or early Eocene time (Bown, 1980). Also, the Powder

River Basin was probably closed along its southern and eastern margins (Laramie and Black Hills arches) during late Paleocene or early Eocene time.

In summary, our study supports the interpretation of anastomosed Laramide arches bounding the Wind River, Bighorn, and Powder River Basins as proposed by Erslev (1993). Timing of uplifts using Nichols' (this volume) palynostratigraphic zonation of the Paleocene Fort Union Formation in these basins suggests a sequential younging and migration of deformation from southwest to northeast through Wyoming consistent with the interpretations of Perry and others (1991). Sequential development of the tectonic arches is also consistent with polyphase deformation of individual arch segments (such as shown by Bergh and Snoke, 1992) of older arches. Finally, the sequential deformation of the tectonic arches was a primary control on the polyphase evolution of depositional systems that developed in these basins.

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Introduction - Part 2

Palynostratigraphic correlation of Paleocene rocks in the Wind River, Bighorn, and Powder River Basins, Wyoming

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Introduction

This report summarizes age determinations and correlations based on fossil pollen and spores (palynostratigraphy) in measured sections of the Paleocene Fort Union Formation in three foreland basins in Wyoming: the Wind River, Bighorn, and Powder River Basins (Figure 1). Measured sections in these basins contain thin to thick deposits of coal, and are the subject of the 1994 field trip of The Society for Organic Petrology (TSOP). This chapter of the TSOP guidebook provides an overview of the age relations among the basins, and provides a temporal framework for interpretations of depositional history, basin evolution, and development of the coal resources (Flores and others, this volume).

Most of the palynostratigraphic age data presented from the measured sections are new. They

were developed by me in conjunction with ongoing studies being conducted in the Wind River and Powder River basins by R.M. Flores and C.W. Keighin and in the Bighorn Basin by R.W. Stanton and S.B. Roberts, all of the U.S. Geological Survey (USGS). Data from the Powder River Basin also incorporate results from my previous studies conducted in collaboration with J.N. Weaver and J.L. Brown of USGS, and studies by D.T. Pocknall, formerly a visiting scientist at USGS and currently with Amoco Production Company. The synthesis of these data and the correlations proposed are new. The correlations utilize the regionally applicable palynostratigraphic zonation of the Paleocene, which divides the series into six zones (P1-P6). Other locally applicable zonations are compared, and some conclusions based on them are incorporated, as appropriate.

Palynostratigraphic zonations, past and present

The earliest published palynostratigraphic zonation of the Paleocene in any Wyoming foreland basin is that of Leffingwell (1970), who conducted a major study of both the Upper Cretaceous Lance Formation and the Paleocene Fort Union Formation in the Powder River Basin. His study areas were the Lance Formation type area near the southeastern edge of the basin; the Gillette coal field near Gillette; and the Spotted Horse coal field near Sheridan. His composite reference section included all members of the Fort Union Formation in the Powder River Basin (Fig-

ure 2). Samples from the Tullock Member came from measured sections in the type Lance area and the Gillette coal field, and samples from the Lebo and Tongue River Members came from the Gillette and Spotted Horse coal fields.

Leffingwell (1970) effectively defined (but did not formally name) three zones in this interval on the basis of three differing assemblages of fossil pollen and spores. Assemblage A of Leffingwell defined a zone of latest Cretaceous (Maastrichtian) age within the

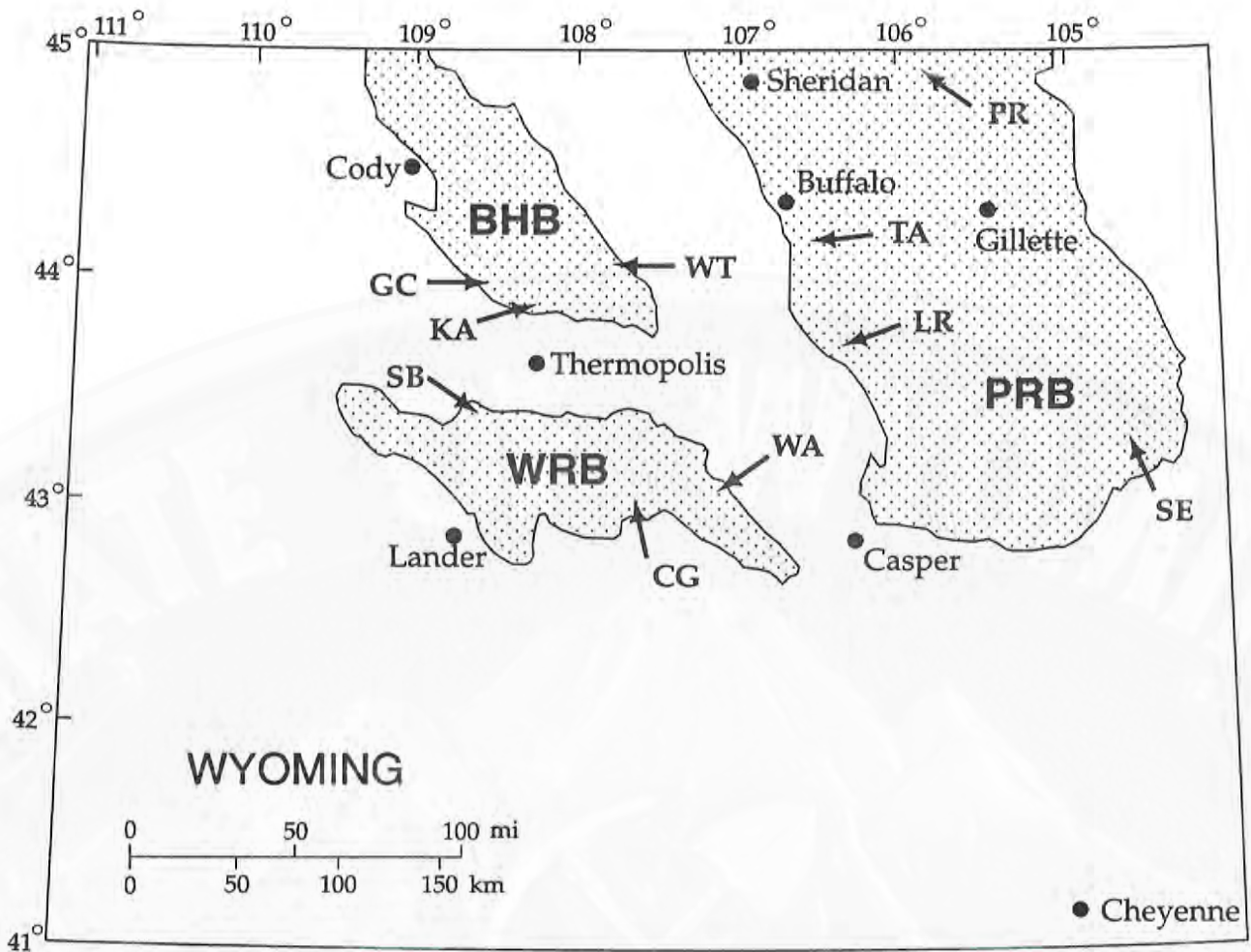


Figure 1. Map of Wyoming showing the basins discussed in this report and indicating approximate locations of measured sections correlated. Basin outlines are defined by Paleocene outcrop. Key: WRB - Wind River Basin (WA - Waltman area, CG - Castle Gardens, SB - Shotgun Butte); BHB - Bighorn Basin (GC - Grass Creek area, KA - Kirby area, WT - Worland-Tensleep highway section); PRB - Powder River Basin (LR - Linch roadcut section, TA - TA Hills, PR - Powder River area, SE - southeastern part of basin).

Lance Formation, and the top of the Assemblage A zone coincides with the top of the Lance. Characteristic species of Assemblage A include many kinds of angiosperm (flowering plant) pollen that are restricted to the Lance Formation. Other species are present in both formations, but among these are several that are common in the Lance but rare in the Fort Union. These strong differences in occurrence serve to clearly distinguish the assemblages. Leffingwell's Assemblage B is present throughout the Tullock Member and in the lower part of the Lebo Member. Assemblage B differs from Assemblage A primarily in lacking those species that are restricted to the Lance Formation. Assemblage C is present in the upper part of the Lebo Member and throughout the Tongue River Member. It is characterized by certain distinctive species of angiosperm pollen (especially fossil species of

the living family Juglandaceae, the walnut family) and a relatively great abundance of pollen probably produced by conifers of the family Taxodiaceae. Leffingwell (1970) noted that his zonation could be traced from Wyoming to Montana, South Dakota, and North Dakota.

Tschudy (1976) studied the palynology of the upper part of the Tongue River Member of the Fort Union Formation and the lower part of the overlying Wasatch Formation. In his study area (Campbell and Sheridan Counties, Wyoming, and near Decker, Montana) both formations are composed of similar continental deposits of sandstone, shale, and coal. Previous investigators had placed the formation contact at various unconformities, at or near the top of the Roland coal bed, at the top of the Smith coal bed, or

	POWDER RIVER BASIN	WIND RIVER BASIN	BIGHORN BASIN
EOCENE (part)	Wasatch Formation	Wind River Formation Indian Meadows Formation	Willwood Formation
PALEOCENE	Tongue River	Shotgun	Fort Union Formation
	Lebo Member	Waltman Shale Member	
	Tullock Member	unnamed lower part	
U. CRET. (part)	Lance Formation	Lance Formation	Lance Formation

Figure 2. Chart showing uppermost Cretaceous and lower Tertiary lithostratigraphy of the basins discussed in this report. Ruled lines indicate a hiatus.

		TONGUE RIVER AREA	POWDER RIVER AREA
EOCENE (part)	WASATCH FORMATION (part)		Scott
			Felix
UPPER PALEOCENE	FORT UNION FORMATION (part)	TONGUE RIVER MEMBER	Arvada
			Roland
			Smith
			Wyodak-Anderson
			Canyon
			Cook
			Wall
			Pawnee
			O' Dell/Cache
			Macklin-Walker
		King	Sawyer (Broadus)

Figure 3. Chart showing coal-bed nomenclature and relative stratigraphic order of selected coal beds of the Tongue River Member in the Tongue River and Powder River areas. The Roland coal bed in the Tongue River area is the Roland coal bed of Baker (1929).

at a change in heavy-mineral suites (coal bed names are shown in Figure 3). Tschudy's objective was to determine if consistently recognizable changes in palynological assemblages occurred within the interval between the Smith and Felix coal beds.

Tschudy (1976) used pollen of the Juglandaceae and certain other angiosperms and conifer pollen of both the Taxodiaceae and Pinaceae to distinguish three zones within the upper part of the Fort Union Formation and the lower part of the Wasatch Formation, from oldest to youngest designated X, Y, and Z. Zone X of Tschudy is present in the interval including the Wall, Canyon, Anderson, Wyodak, and Smith coal beds. Zone Y is present in the interval including the Arvada and Roland beds and rocks to within about 30 m below the Felix bed. Zone Z is present in the interval including the Felix bed and to within about 25 m below the top of the Scott bed.

Tschudy dated Zone Z as early Eocene by identification of comparable palynological assemblages in samples from vertebrate localities of known early Eocene age in Wyoming and North Dakota. By inference, Zone Y is latest Paleocene in age. An important conclusion follows from this: in parts of the Powder River Basin the lowermost part of the Wasatch Formation is latest Paleocene in age. Tschudy (1976) was also the first to propose that the first appearance of pollen of the juglandaceous genus *Platycarya* marks the Paleocene-Eocene boundary in the Rocky Mountain region.

Nichols and Ott (1978) defined six zones encompassing the entire Fort Union Formation in the Wind River Basin. The Fort Union in the Wind River Basin comprises lower and upper parts; the upper part is divided into the laterally equivalent Shotgun Member and Waltman Shale Member (Figure 2). The palynostratigraphic zonation is based largely on the distribution of species of the juglandaceous genus *Momipites* and of its descendant genus *Caryapollenites*, although some other species of angiosperm pollen supplement the definitions of the zones. The zones, which from oldest to youngest are designated P1 through P6, were defined in a reference section near Waltman, Wyoming, but as noted by Nichols and Ott (1978), they can be identified also in the Powder River, Bighorn, and Hanna Basins in Wyoming and in other deposits of Paleocene age in adjacent states. Jacobson and Nichols (1982) applied the zonation for age deter-

minations throughout the thrust belt of western Wyoming and adjacent Utah, and most recently, Nichols and Flores (1993) demonstrated its utility in correlation within the Wind River Basin.

In the reference section Zone P1 is recognized by the presence of a single species of juglandaceous pollen, *Momipites leffingwellii*, and the absence of Cretaceous pollen and spores. Zone P2 contains *Momipites waltmanensis* and *M. wyomingensis* in addition to *M. leffingwellii*. The lower part of Zone P3 contains *M. actinus* and *M. anellus*, and the upper part contains *M. triorbicularis* and *M. ventifluminis*. Zone P3 also contains *Aquilapollenites spinulosus* (a Paleocene species of a genus usually thought of as Cretaceous; this species does not occur in either the Cretaceous or the lower part of the Paleocene) and two species attributable to either *Intratropipollenites* or *Tilia*. Zone P4 is characterized by presence of species of the genus *Caryapollenites* (*C. wodehousei* is especially common) and reductions in the abundance of species of the genus *Momipites* (*M. leffingwellii* is absent from this zone and *M. actinus* ranges no higher). Zone P5 contains *Caryapollenites veripites*, *Pistillipollenites mcgregorii*, and *Insulapollenites rugulatus*. In Zone P6 *C. veripites* is the dominant juglandaceous species. Beyond these basic patterns of occurrence are more subtle but important

characteristics of the zones. As noted, some of the species of *Momipites* are restricted in range, although others persist in occurrence after their first appearance; additional species of *Caryapollenites* are present in Zones P4 through P6; and the relative abundance of species of *Momipites* and *Caryapollenites* changes through the interval, with *Caryapollenites* tending to become more abundant. Some of these species are illustrated in Figure 4.

The Wind River Basin "P" zones defined by Nichols and Ott (1978) can be compared with the previously described assemblages and zones of the Powder River Basin, although the Wind River zones include more and somewhat different species of juglandaceous pollen, especially those referred to the genus *Momipites*. Zones P1 and P2 are generally equivalent to Assemblage B of Leffingwell (1970), and Zones P3 through P6 are equivalent to Leffingwell's Assemblage C. Zone P5 is in part equivalent to Zone X of Tschudy (1976), although Zone P5 also extends below the stratigraphic level examined by Tschudy. Zone P6 is equivalent to Tschudy's Zone Y, based on the abundance of pollen of *Caryapollenites* and the presence of other species of the juglandaceous complex of fossil pollen, although the zones are defined differently. By superposition, Zones P1 and P2

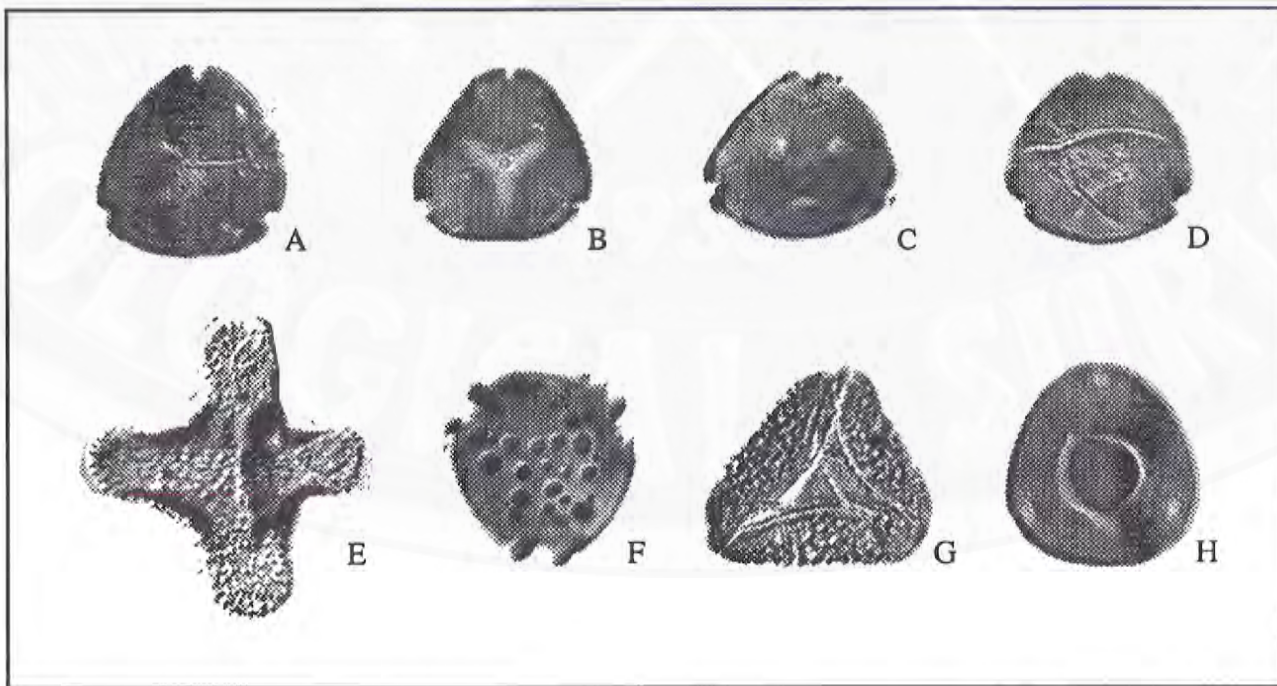


Figure 4. Some of the biostratigraphically important pollen species mentioned in the text. Magnification 1000X. Key: A - *Momipites wyomingensis*, B - *M. actinus*, C - *M. ventifluminis*, D - *Platycarya platycaryoides*, E - *Aquilapollenites spinulosus*, F - *Pistillipollenites mcgregorii*, G - *Insulapollenites rugulatus*, H - *Caryapollenites veripites*.

together represent an informally recognized lower Paleocene, Zones P3 and P4 the middle Paleocene, and Zones P5 and P6 the upper Paleocene. The zones represent undetermined intervals of time, but are unlikely to subdivide equally the approximately 10.5-million-year duration of the Paleocene. Nichols and Ott (1978) also recognized a basal Eocene zone, Zone E, in the Wind River Basin, which is equivalent to Zone Z of Tschudy (1976). In both basins the basal Eocene zone is characterized by the presence of pollen of *Platycarya*.

Wing (1984) formally proposed that the first appearance (lowest stratigraphic occurrence) of *Platycarya platycaryoides* should be used to mark the Paleocene-Eocene boundary in the Western Interior of North America. Wing's reference section is in the southern Bighorn Basin and includes the upper part of the Fort Union Formation and the overlying Willwood Formation. He initially placed the Paleocene-Eocene boundary 100 m above the Fort Union-Willwood contact, based on the first appearance of *P. platycaryoides*. Subsequently, *Platycarya* pollen was identified 35 m below the contact, within the uppermost part of the Fort Union (Wing and others, 1991). No biostratigraphically important mammalian fossils currently are associated with the first appearance of *Platycarya*, but lowermost Eocene (basal Wasatchian) mammals are present just above, in the basal Willwood Formation. Wing and others also noted that the palynological datum is approximately isochronous with the marine nannofossil zone boundary NP9/NP10, which defines the Paleocene-Eocene boundary in marine sections. The geochronologic age of this boundary is $55.07 \text{ Ma} \pm 0.16 \text{ m.y.}$, based on data presented by J.D. Obradovich in Berggren and others (1992).

Pocknall (1987) conducted a detailed palynologic study of the Tongue River Member of the Fort Union Formation and the lowermost part of the Wasatch Formation in the Powder River Basin, on either side of the Wyoming-Montana State line. The interval he studied included coal beds from the Sawyer through the Felix (Figure 3). His reference section was composited from outcrop sections near Broadus, Montana, where the interval from the Sawyer through O'Dell/Cache coal beds is exposed; four drill cores collected along the Powder River north of Arvada, Wyoming, which together sampled the interval from the Pawnee through the Arvada coal beds; and an outcrop section near Arvada, Wyoming, at which the Felix coal bed is exposed. The part of the interval from the Sawyer coal bed up to the Arvada coal bed is the Tongue River Member, and the part of the interval including the Arvada and Felix coal beds and noncoal

rocks above the Felix is the lower part of the Wasatch. Pocknall defined three formal palynostratigraphic zones within this total interval, one of which has two subzones. From oldest to youngest these are the *Aquilapollenites spinulosus* Zone, the *Caryapollenites* Assemblage Zone (including the *Pistillipollenites mcgregorii* Subzone and the *Insulapollenites rugulatus* Subzone), and the *Platycarya platycaryoides* Zone. The base of the lowest zone and the top of the highest zone cannot be defined in the interval investigated because sample coverage is limited to only part of the Fort Union-Wasatch succession.

The *Aquilapollenites spinulosus* Zone is named for the previously mentioned characteristic Paleocene species of an otherwise Cretaceous genus. Other species that define the zone belong to the juglandaceous genera *Momipites* and *Caryapollenites* and other genera of angiosperm pollen, notably *Ulmipollenites*. The zone extends from an undetermined level below the Sawyer coal bed to just below the O'Dell/Cache coal bed. The overlying *Pistillipollenites mcgregorii* Subzone of the *Caryapollenites* Assemblage Zone is defined by the first appearances of *P. mcgregorii* and *C. veripites*. It encompasses the interval from the O'Dell/Cache coal bed to just above the Canyon coal bed. The *Insulapollenites rugulatus* Subzone of the *Caryapollenites* Assemblage Zone is defined by a sharp increase in the relative abundance of *Caryapollenites* pollen relative to *Momipites* pollen; *I. rugulatus*, which is present in the underlying subzone as well, ranges no higher than this subzone. This zone encompasses the interval including the Anderson, Smith, and Arvada coal beds. The base of Pocknall's *Platycarya platycaryoides* Zone, which overlies his *Caryapollenites* Zone, is defined by the first appearances of *P. platycaryoides* and *Intratropopollenites* sp. (cf. *Tilia*; this species differs from both of those previously mentioned as occurring in the Paleocene of the Wind River Basin). The base of this zone is between the Arvada and Felix coal beds; the top of the zone is undefined.

Pocknall's (1987) zones in the Powder River Basin can be compared with others previously proposed for that area and with the regionally applicable zones defined in the Wind River Basin. His *Platycarya platycaryoides* Zone is clearly equivalent to Zone Z of Tschudy (1976) and Zone E of Nichols and Ott (1978). The equivalence of the Paleocene zones is less evident. Pocknall (1987, Figure 4) proposed correlations, but some of them are revised here. The *Aquilapollenites spinulosus* and *Caryapollenites* Zones together are equivalent to Assemblage C of Leffingwell (1970), except that Assemblage C extends below Pocknall's lowest zone and the *Caryapollenites* Zone extends above

the stratigraphic level studied by Leffingwell (as noted by Pocknall, 1987). Based on the stratigraphic intervals studied, Zone Y of Tschudy (1976) must be equivalent to the upper part of the *Caryapollenites* Zone, and Zone X of Tschudy must be equivalent to the middle part of that zone. Some palynological data published by Tschudy (1976) and Pocknall (1987) can be interpreted to indicate that Tschudy's Zone Y is equivalent to the *Insulapollenites rugulatus* Subzone, but this correlation evidently is correct only in part, because Pocknall included the Anderson and Smith coal beds in his *I. rugulatus* Subzone, whereas Tschudy excluded these beds from his Zone Y (and included them in his underlying Zone X). I infer that the *I. rugulatus* Subzone is largely to fully equivalent to Zone P6 of Nichols and Ott (1978), and that the *Pistillipollenites mcgregorii* Subzone similarly corresponds to Zone P5 (Pocknall's correlation of his subzones with P5-P6 differs). Consequently, Zones P5 and P6 together are fully equivalent to Pocknall's *Caryapollenites* Assemblage Zone, and these biostratigraphic units represent the upper part of the Paleocene. Accordingly, the *Aquilapollenites spinulosus* Zone equates with the upper part of Zone P4 (not with the middle part of Zone P5 as indicated by Pocknall's diagram [1987, Figure 4]). These correlations are based on my reevaluation of palynomorph occurrence data reported from the respective reference sections. These and other comparisons of the Paleocene zones are summarized in Figure 5.

The only other palynostratigraphic zone that bears mention here is the *Wodehouseia spinata* Assemblage Zone. This zone, most recently discussed in detail by Nichols and Sweet (1993),

characterizes the uppermost Cretaceous throughout the Western Interior of North America. Recognition of the uppermost Cretaceous *Wodehouseia spinata* Assemblage Zone facilitates identification of the overlying lowermost Paleocene zone, P1, in Wyoming basins. Leffingwell (1970) was the first to note that certain species of pollen and spores (his Assemblage A) are present in the Lance Formation but not in the overlying Fort Union Formation in Wyoming. Nichols and Ott (1978) defined Zone P1 on the basis of occurrence of a species of the pollen genus *Momipites*, but they also noted that species typical of the Cretaceous are absent from P1 assemblages. These species are part of the *Wodehouseia spinata* assemblage and they characterize the upper Maastrichtian rocks of the Wind River, Bighorn, and Powder River Basins.

		Leffingwell 1970	Tschudy 1976	Nichols & Ott 1978	Pocknall 1987	This report
Eocene (part)		(not studied)	Zone Z	E	<i>Platycarya platycaryoides</i>	<i>Platycarya platycaryoides</i>
			Zone Y	P6	<i>Insulapollenites rugulatus</i>	P6
Paleocene	upper		Zone X	P5	<i>Pistillipollenites mcgregorii</i>	P5
		Assemblage C		P4	<i>Aquilapollenites spinulosus</i>	P4
				P3		P3
				P2		P2
		Assemblage B	(not studied)	P1		P1
					(not studied)	
Upper Cret. (part)		Assemblage A		(not studied)		<i>Wodehouseia spinata</i>

Figure 5. Chart comparing palynostratigraphic zonations of the Wyoming Paleocene.

Wind River Basin correlations

At and near the reference section for the Nichols and Ott (1978) palynostratigraphic zonation, near Waltman, Wyoming (WA in Figure 1), the Fort Union Formation is divided into an unnamed lower part and the overlying Waltman Shale Member (Keefer, 1965). As originally described, the lower part of the formation is of fluvial origin; the Waltman Shale Member is now considered to be of freshwater lacustrine origin (Phillips, 1983; Nichols and Flores, 1993). In outcrop, Zones P1 through P4 and the basal part of Zone P5 are within the unnamed lower part, and most of Zone P5 and all of Zone P6 are within the Waltman Shale Member (Nichols and Ott, 1978). Nichols and Flores (1993) suggested that in the subsurface the lower part of the Waltman Shale Member, which thickens into the subsurface and toward the basin axis, may be in the upper part of Zone P4. Near Waltman the Lance Formation, which underlies the Fort Union Formation, presumably contains the upper Maastrichtian *Wodehouseia spinata* Assemblage Zone, although the Lance was not included in the original study. Near Waltman the Wind River Formation, which overlies the Fort Union, contains basal Eocene Zone E of Nichols and Ott (1978), which is equivalent to the *Platycarya platycaryoides* Zone of Pocknall (1987).

The upper part of the Lance Formation, the entire Fort Union Formation, and the lower part of the Wind River Formation are well exposed in the Castle Gardens area along the southern edge of the basin (CG in Figure 1). About 700 m of Fort Union was measured and studied by Flores and others (1992); in this area the formation is not subdivided into members. The composite measured section is west of Castle Gardens park, where thick sandstone units of the Lance Formation are well exposed. Palynomorphs of the *Wodehouseia spinata* Assemblage Zone were recovered from a sample collected below the sandstone in the park. Fine-grained carbonaceous rocks of the Fort Union Formation overlying the thick sandstone in the park yielded no Cretaceous species and are within Zone P1. In the composite measured section the lower 60 m or more is in Zone P1. Above that is a thin (less than 10 m) interval in Zone P2. In succession, Zone P3 is about 100 m thick, and Zone P4 is about 80 m thick. Zone P5 is quite thick in the Castle Gardens area, encompassing more than 290 m. Recovery of palynomorphs is generally poor in the upper part of the Castle Gardens section, but Zone P6 evidently is present in the uppermost 150 m. Poor recovery has so far precluded identification of the *Platycarya platycaryoides* Zone, but the section is still under study.

In the Shotgun Butte area in the western part of the basin (SB in Figure 1), the Fort Union Formation is divided into the unnamed lower part and the overlying, coal-bearing Shotgun Member, both of fluvial origin (Keefer, 1965). The coarse-grained lowermost beds of the unnamed lower part of the Fort Union are difficult to distinguish from those of the underlying Lance Formation, and the upper part of the Shotgun Member is gradational with the overlying red beds of the Eocene Indian Meadows Formation. Nichols and Flores (1993) zoned a composite measured section 825 m thick in this area, beginning near (but not at) the base of the Fort Union. Zones P2 through P3 were identified in the unnamed lower part of the formation (Zone P1 presumably is present in the unsampled interval of the lower part that lies below the sampled interval). Recovery from the Shotgun Member was poor, but samples about 300 m above the base of the Shotgun are in Zone P5. Correlations in the Wind River Basin are summarized diagrammatically in Figure 6.

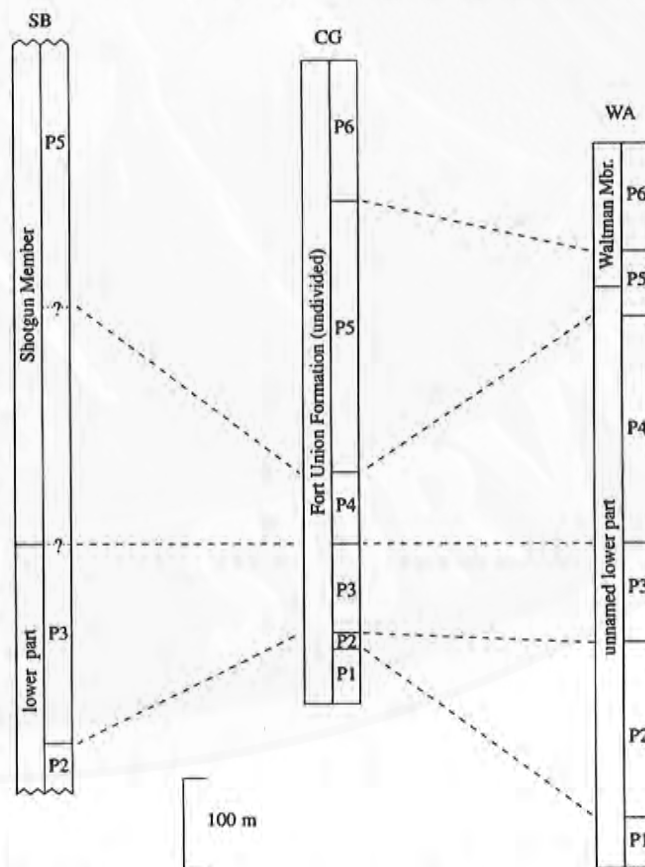


Figure 6. Zonation and correlation of the Fort Union Formation in the Wind River Basin. Key: SB - Shotgun Butte area, CG - Castle Gardens area, WA - Waltman area. Only zones that have been positively identified in measured sections are shown; positions of some zone boundaries are inferred.

Bighorn Basin correlations

In the southwestern part of Bighorn Basin, near Grass Creek (GC in Figure 1), the lower part of the Fort Union Formation contains coal beds as thick as 11 m (Hewett, 1926). The Mayfield coal bed, which currently is being mined at the Grass Creek mine, is about 175 m above the basal bed of the Fort Union Formation (Roberts and others, 1994; Roberts and Stanton, this volume). Palynology samples from the base and the top of the Mayfield coal bed show that it lies in the upper part of Zone P3, based on species of *Momipites* and *Intratropipollenites* present. *Insulapollenites rugulatus* is present in the coal assemblages also, indicating that this species ranges lower in the Bighorn Basin than in either the Wind River Basin (P5) or the Powder River Basin (*Caryapollenites* Assemblage Zone, equivalent to P5 and P6).

On Ilo Ridge, to the east of the Grass Creek mine, a sample collected by Steve Roberts stratigraphically below the Mayfield coal bed (110 m above the basal bed of the Fort Union) yielded an assemblage characteristic of the lower part of Zone P3. A sample about 15 m above the basal bed could not be dated. Roberts collected four samples in his Red Ridge measured section northeast of the Grass Creek mine, three of which provide additional palynostratigraphic control for the lower part of the Fort Union in the southwestern part of the Bighorn Basin. The sample from just above the basal bed of the Fort Union also failed to yield datable material, but another about 60 m above is from Zone P2. A sample about 135 m above the basal bed of the Fort Union is in lower Zone P3, and one about 220 m above the basal bed (and thus stratigraphically higher than the Mayfield coal bed) is from the upper part of Zone

P3. These age determinations are composited in the column for the Grass Creek area in Figure 7.

Roberts collected palynological samples from a composite measured section near Kirby (KA in Figure 1), which as mapped (Love and others, 1979) includes a complete section of the Fort Union Formation from the underlying Lance Formation (Maastrichtian) to the overlying Willwood Formation (Eocene). The measured section includes about 790 m of the Fort Union; see also Flores and others (this volume). A sample from the part of the interval mapped as Lance yielded a Paleocene assemblage indicative of Zone P1.

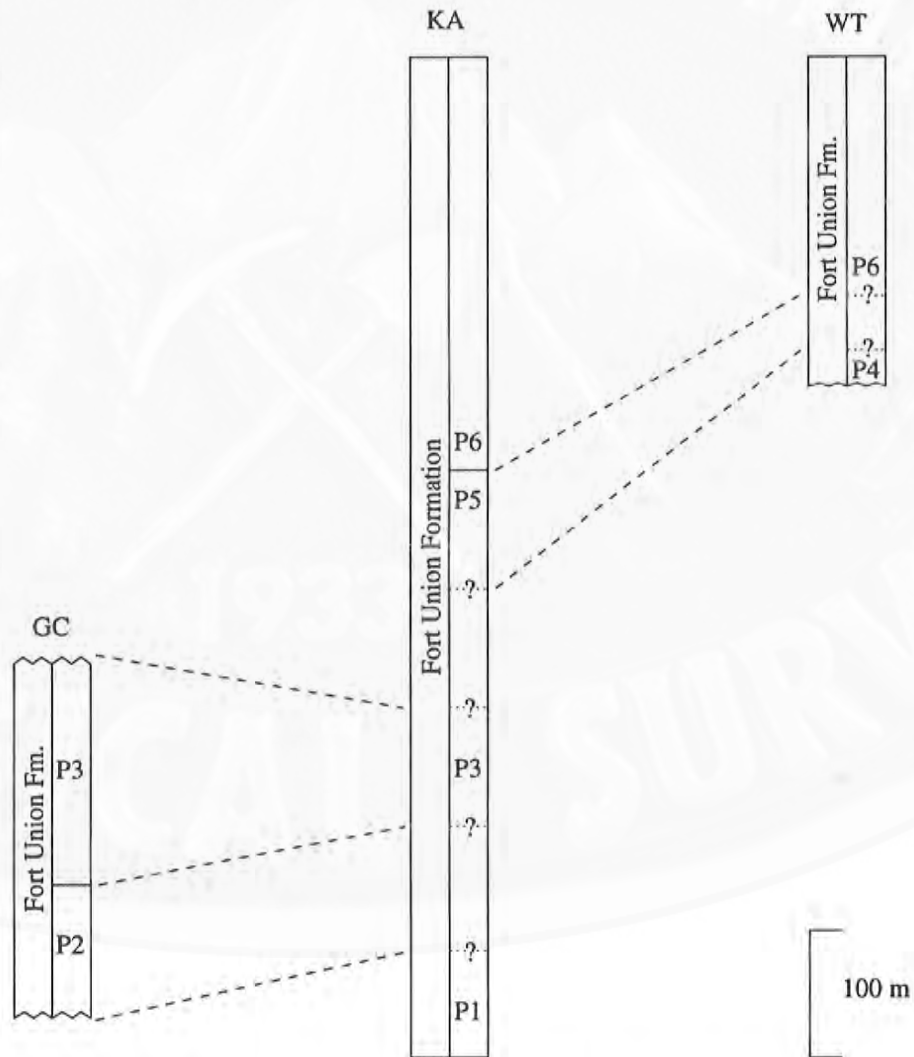


Figure 7. Zonation and correlation of the Fort Union Formation in the Bighorn Basin. Key: GC - Grass Creek area (includes Ilo Ridge and Red Ridge), KA - Kirby area, WT - Worland-Tensleep highway section. Only zones that have been positively identified in measured sections are shown; positions of some zone boundaries are inferred.

A sample about 220 m above is in the lower part of Zone P3. Several samples from a thick interval centered about 250 m above that sample could not be dated, but they contained several species of algal cysts indicative of freshwater lacustrine depositional environments. Two samples from the upper part of the section yielded sparse assemblages of palynomorphs indicative of the upper Paleocene zones P5 and P6, respectively. The uppermost part of the composite section near Kirby has not been dated, but the interval is still under study.

The Fort Union Formation is well-exposed along the highway between Worland and Tensleep, Wyoming, in the southeastern part of the Bighorn Basin (WT in Figure 1; see also Flores and others, this volume). A 260-m-thick section measured here by Steve Roberts is estimated to begin 60-70 m above the Lance Formation as mapped in this area (Love and others,

1978) and end about 45 m below the Willwood Formation. Even if the unmeasured intervals totaling 105-115 m were to be added to the measured section, it is evident that the Fort Union Formation is considerably thinner here than in the Kirby area. Previous mapping in the vicinity (Bown, 1979) suggests that there is an unconformity at the base of the Fort Union in this part of the basin. A palynological sample collected only about 18 m above the base of the measured section is in Zone P4, and another collected about 90 m above the base is in Zone P6; thus there is evidence that much of the Fort Union Formation is indeed missing in this area, and that only the upper part is preserved. Further work, including additional palynologic dates, may further clarify the stratigraphic and structural relationships. Existing palynostratigraphic correlations within the Paleocene of the southern Bighorn Basin are summarized diagrammatically in Figure 7.

Powder River Basin correlations

The Cretaceous-Tertiary (K-T) boundary and the lower part of the Paleocene in the Powder River Basin are well constrained by palynological dating at several places. Janet Brown conducted a basin-wide study of the Tullock Member of the Fort Union Formation (Brown, 1993), and Nichols and Brown (1992) detailed the palynology and age of the Tullock in its type area in the northwestern part of the basin (Montana), and at several localities in the southeastern part of the basin in Wyoming (SE in Figure 1). Near Sussex, Wyoming, in the southwestern part of the basin, a K-T boundary locality was discovered in the course of these studies (Nichols and others, 1992). Palynostratigraphic age determinations obtained for the Tullock study (Nichols and Brown, 1992) show that, at some localities in the basin, lowermost Paleocene P1 assemblages are present also in the uppermost parts of underlying, predominantly Upper Cretaceous units (Lance Formation in Wyoming and Hell Creek Formation in Montana).

Lowermost Paleocene Zone P1 in the Powder River Basin includes two species of *Momipites* not known from the Wind River Basin reference section, *M. inaequalis* and *M. tenuipolus* (the latter had been reported in the basin previously by Leffingwell, 1970). In the Powder River Basin these species evidently replace the Wind River P1 indicator species *M. leffingwellii*. Zone P2 includes the *Momipites* species that characterize Zone P2 in the Wind River and Bighorn basins. Zone P3 assemblages in the Powder

River Basin include the characteristic species of *Momipites*, *M. anellus*, *M. triorbicularis*, and *M. ventifluminis*, as well as *Aquilapollenites spinulosus*.

In work still in progress, Romeo Flores and Bill Keighin measured an apparently complete section of the Fort Union Formation on the western side of the Powder River Basin, along the road near Linch, Wyoming (LR in Figure 1; see also Flores and others, this volume). The Linch roadcut section is about 885 m thick. It begins in the lower part of the Tullock Member and ends in Tongue River Member equivalent, coal-bearing rocks, although existing mapping of the area (Love and Christiansen, 1985) shows only the Lebo Member above the Tullock near Linch. Preliminary palynostratigraphic results indicate that all six zones of the Paleocene are present in the Linch roadcut section.

Recovery is poor from samples collected from the lowermost part of the Linch section, but they presumably represent Zone P1 because they lack Cretaceous species (the *Wodehouseia spinata* assemblage is well represented in the uppermost Lance Formation at the Sussex K-T boundary locality, not far from the Linch section). Zone P1 is more clearly identifiable from better samples about 200 m above the base of the measured section. Zone P2 can be inferred to be within some part of the interval between 200 and 440 m in the measured section, although no palynological samples were collected in that interval because no suitable lithologies

are present. Zone P3 is present from about 440 to 530 m above the base of the section; assemblages here include *Aquilapollenites spinulosus* as well as the characteristic species of *Momipites*. Zone P4 has been identified at about 560 m, and Zone P5 at about 650 m. The thick coal beds of the Sussex coal zone are exposed in the Linch section beginning at about 700 m and continuing to about 750 m. The coal zone brackets the P5-P6 zone boundary; Zone P6 is present at about 745 m. The Sussex coal zone is therefore the palynostratigraphic age-equivalent of the interval including the Anderson and Canyon coal beds in the Powder River area (middle part of the *Caryapollenites* Zone of Pocknall, 1987). Romeo Flores (verbal communication, 1994) stated that the Sussex coal zone in outcrop near Linch is lithostratigraphically correlative with the Big George coal bed in the subsurface of the Powder River Basin, implying that the Big George coal bed is also P5-P6. Zonation of the Linch section is shown diagrammatically in Figure 8.

To the north and also along the western side of the Powder River Basin, Jean Weaver measured sections in conglomeratic and coal-bearing facies in the upper part of the Fort Union Formation in the TA Hills area, near Buffalo, Wyoming (TA in Figure 1). Interpretations of stratigraphy and depositional environment of this interval were published by Weaver and Flores (1985). Palynological

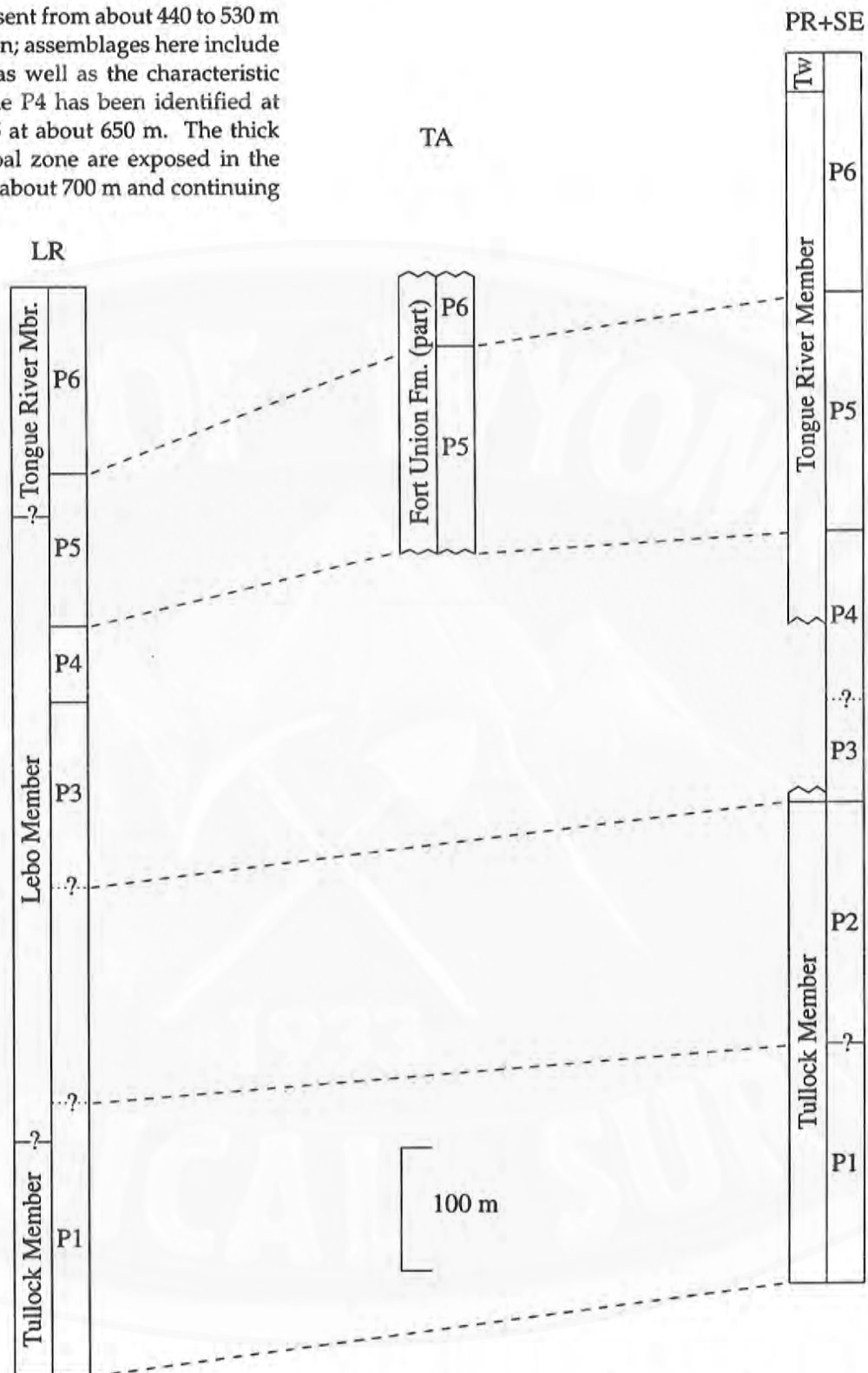


Figure 8. Zonation and correlation of the Fort Union Formation in the Powder River Basin. Key: LR - Linch roadcut section, TA - TA Hills, PR+SE - section composited from Powder River area and localities in southeastern part of basin (the gap in section includes most of the Lebo Member and the lowermost part of the Tongue River Member not sampled in this study; the symbol Tw represents part of the Wasatch Formation in the Powder River area that is of latest Paleocene age). Only zones that have been positively identified in measured sections are shown; positions of some zone boundaries are inferred.

samples collected from Weaver's sections provide age control and permit correlation of this interval with the upper part of the Fort Union in the Linch and Powder River areas. The coal-bearing facies in the TA Hills is in Zone P5, and the conglomeratic facies is in the lower part of Zone P6. The entire succession of Fort Union rocks measured in the TA Hills is correlative with the interval including the Pawnee through Anderson coal beds in the Powder River area, which lies within the *Caryapollenites* Zone of Pocknall (1987); see also Figure 2 in Flores and others (1989).

Weaver and Flores (1987) interpreted the depositional environment of the upper part of the Fort Union Formation in additional measured sections south of the TA Hills. The measured sections are within the upper 300 m of the Fort Union Formation and the succession is presumed to be at least partially equivalent to the succession in the TA Hills area. Palynological samples from section RH-1 in the Reno Hill area show that this interval is in Zone P5. It is evident that much of the upper part of the Fort Union Formation on the western side of the Powder River Basin, which includes both conglomerate and coal beds, is late Paleocene in age and correlative with the Tongue River Member.

The stratigraphy, sedimentology, and depositional environments of the Tongue River Member of the Fort Union Formation have been exhaustively studied by Romeo Flores (e.g., Flores, 1981, 1983, 1986). As previously mentioned, the upper part of Zone P4, Zone P5, and Zone P6 encompass most of the Tongue River Member in the central part of the Powder River Basin, along the Powder River (PR in Figure

1). These zones are equivalent to the *Aquilapollenites spinulosus* and *Caryapollenites* Zones of Pocknall (1987). Little more needs to be said of correlations in the Tongue River Member that is not already covered in this report in the discussion of Pocknall's (1987) zonation and its equivalents. However, a few comments can be made to complete the picture in adjacent areas.

The Wyodak-Anderson coal bed in the Gillette coal field is palynostratigraphically correlative with the Anderson coal bed in the Powder River area. The Wyodak-Anderson coal bed is in the *Caryapollenites* Zone, near the Zone P5-P6 boundary, based on my unpublished data from the Black Thunder mine. It is probable that the Anderson-Dietz coal bed at the Decker mine in the Tongue River area of Montana (a stop on the third day of the TSOP field trip) is also palynostratigraphically correlative with the P5-P6 boundary interval, although no samples from the Decker mine have been analyzed to date. A composite zonation for the Fort Union Formation in the Powder River area and the eastern side of the Wyoming part of the basin is shown in Figure 8. It lacks control in an unsampled interval including the upper part of Zone P3 and the lower part of Zone P4 (the Lebo and basal part of the Tongue River Members), but the correlations implied are unlikely to change significantly with further work. However, future palynostratigraphic studies in local areas can contribute to refinement of interpretations of the depositional history and basin evolution of the Powder River Basin and other foreland basins of Wyoming and to better understanding of the development of coal resources in these basins.

Acknowledgments

This summary of palynostratigraphic relations within the Paleocene of Wyoming basins would not have been possible without the collaboration in the field of Janet Brown, Romeo Flores, Steve Roberts,

Ron Stanton, and Jean Weaver. I thank Romeo Flores for information on coal-bed correlations in the Tongue River Member and Pat Holroyd for discussions about the Paleocene-Eocene boundary.

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Introduction - Part 3

Pre-field trip route from Jackson to Riverton

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Introduction

Jackson Hole, a valley nearly 50 miles (81 km) in length, is bounded by the Grand Tetons to the west, the Wind River Range, and the Gros Ventre Range to the southeast. ("Hole" was a trappers term for a protected valley). The valley is said to have been first explored (by white men) in 1807 by John Colter; it was subsequently entered by various mountain men in the 1820s. David E. Jackson, with William Sublette and Jedediah Smith, bought a fur traders company in 1826; the region is named for Jackson. Settlers had arrived by the early 1880s and started large cattle ranches. Parts of the remote valley were also good hideouts for cattle rustlers.

More detailed information about the geology of the region may be found in Mears and others (1986), deVries and Crockett (1990), and Love and others, (1992). A brief discussion of the region, as well as an introduction to the geology of Wyoming is found in Blackstone (1988).

The Teton Range is the youngest range in the Rocky Mountain range, and rises dramatically above

Jackson Hole, the valley in which the town of Jackson lies. The highest peak, Grand Teton, reaches 13,766 ft (4,196 m) above sea level, while the floor of Jackson Hole is at an elevation of 6,750 ft (2,057 m). The contrast is accentuated by the absence of foothills.

This region experienced two major episodes of tectonism. The Teton/Gros Ventre uplift formed in Laramide time between 80 and 65 million years ago, and produced a southwest-trending arch. This arch was broken by the north-trending Teton normal fault zone about 9 million years ago, when the existing Teton Range began to form. Maximum displacement along this zone of high-angle normal faults is between 25,000 and 30,000 ft (7,620 - 9,144 m). Continued movement is indicated by fault scarps that offset Pleistocene and Holocene deposits along the mountain front.

The Teton fault block has been affected by at least four episodes of glaciation, ranging from Pliocene(?) to late Pleistocene in age, which extensively altered the landscape.

Road log

Mileage:

Incre- ment	Total
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0.0	0.0
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Description:

Begin road log at north edge of Jackson at Flat Creek bridge. The route of this trip is shown on Figure 1. Go north on U.S. Highways 26/89/191.

To the east is Jackson National Elk Refuge where several thousand elk are fed during the winter. Miller Butte, a mile to the northeast, is composed of west-dipping Madison Limestone and Amsden Formation (Mississippian and Upper Mississippian to Middle Pennsylvanian, respectively) in a down-dropped fault block. At the base of Miller Butte is a series of large, warm springs that do not freeze in winter, and provide a year-round haven for trumpeter swans. The timbered highlands east of Miller Butte are the western end of the Gros Ventre Range and comprise west-dipping Madison, Amsden, and Tensleep (Middle and Late Pennsylvanian in

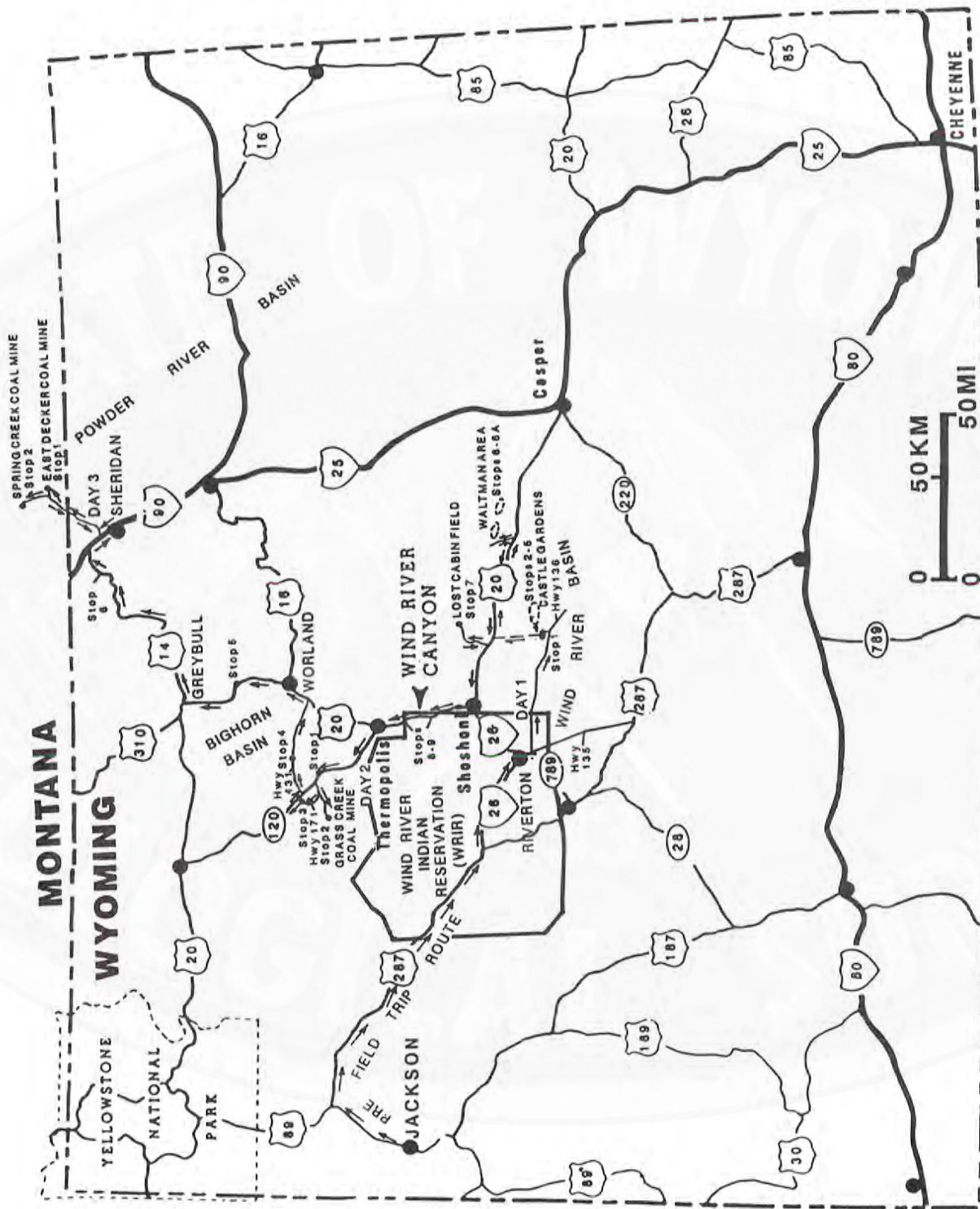


Figure 1. Map showing pre-field trip route and daily routes for the field trips in the Wind River Basin (Day one), Bighorn Basin (Day two), and northwestern Powder River Basin (Day three).

Mileage:		Description:
Increment	Total	
		age) Formations. To the west/northwest is East Gros Ventre Butte, also composed of west-dipping Paleozoic rocks. This butte, and others out of sight farther west, are fault-block remnants of the foundered ancestral Gros Ventre-Teton-Targhee uplift. The uplift foundered along north-trending faults that cut across the older, underlying Cache Creek thrust fault. The thrust is of Laramide age and involves the foreland block that moved southward and southwestward across Mesozoic rocks that dip vertically under the town of Jackson.
		The Paleozoic rocks at the south end of East Gros Ventre Butte are folded down because of drag along the Cache Creek thrust fault. They include strata ranging from the Bighorn Dolomite (Middle and Upper Ordovician), just above town level, to the Phosphoria Formation (Lower Permian) near the top of the Butte. The top of the butte is capped by red and black scoriaceous basalt, probably Quaternary or Tertiary in age.
1.4	1.4	U-shaped saddle cut by pre-Wisconsin glaciers moving through the butte. On the north side of the saddle is brown andesite of late Tertiary or Quaternary age. This is one of eight distinct young volcanic sequences in the East and West Gros Ventre Buttes.
1.3	2.7	Ragged cliffs on butte at right are andesite, which are underlain by a black obsidian which overlies white tuffs, probably of Tertiary age.
0.7	3.4	Sheep Mountain (the "Sleeping Indian") on the skyline to the east is the highest peak in the western part of the Gros Ventre Range. The nearly flat-lying strata on the southwest side of the peak range in age from Cambrian to Mississippian.
2.7	6.1	Bridge across Gros Ventre River.
5.0	11.1	Blacktail Butte to the east is a complex fault block that has been scoured, faceted, and overridden by southward-moving pre-Wisconsin glaciers.
2.7	13.8	Turnoff to Grand Teton National Park Headquarters at Moose.
1.1	14.9	Cross Ditch Creek.
0.2	15.1	To the east, the Lower Gros Ventre landslide, eight miles (13 km) away, is visible. The slide occurred in 1925, and is the largest historic landslide in the US, with a volume of about 50 million cubic yards (38 million cubic meters) of debris. The slide/debris created a dam about 225 feet (69 m) high, impounding a lake more than four miles (6 km) long. The top 50 feet (15 m) of the dam washed out in 1927; the ensuing flood destroyed the town of Kelly, many ranches and highway bridges downstream, and drowned several people.
7.2	22.3	Snake River overlook.
5.5	27.8	Cross Spread Creek. For about 700 feet (213 m), and extending eastward about 3/4 mile (1.2 km) a fault scarp exposes gold-bearing quartzite conglomerate and coal beds intertonguing with marine and brackish water sandstones and shales of the Bacon Ridge Sandstone (Upper Cretaceous). This fault may be related to the Spread Creek thrust fault which is approximately 5 miles to the east, trending nw-se (Love and others, 1992).
3.8	31.6	Moran Junction. Intersection of U.S. Highways 26 and 287. Follow U.S. Highway 26 east toward Dubois, Wyoming. Original roadlog by Love and Love (1980); Flanagan (1964); and Mears and others (1986). Revised by K.A. Sundell and G.F. Winterfeld.
0.8	32.4	Steeply dipping, hard, thin, gray sandstones, shales, and coal beds are part of the Sohare Formation on the west flank of the Spread Creek anticline.
1.5	33.9	East boundary of Grand Teton National Park.
0.6	34.5	Cross Lava Creek bridge. Cliffs visible upstream are Bacon Ridge Sandstone (Upper Cretaceous) at crest of Spread Creek anticline; coal was mined between 1910 and 1916. This is the largest anticline with Upper Cretaceous rocks in its core in Jackson Hole; it is approximately 20 mi (32

Mileage:		Description:
Increment	Total	
		km) long, and has up to 6,000 ft (1,829 m) of closure. It is pre-Tertiary in age, and has several large flammable gas seeps along its crest; it has been tested unsuccessfully for oil and gas in the past 45 years.
0.3	34.8	On the left are old tipples and dumps of abandoned coal mines. Coal occurs in the Upper Cretaceous Bacon Ridge Sandstone.
0.6	35.4	Bridge across Buffalo Fork River.
0.9	36.3	On the skyline to the south, Mount Leidy is composed of Pinyon Conglomerate, a gold-bearing quartzite conglomerate of Late Cretaceous and Paleocene age. It rests in places with a 90° unconformity on vertical Cretaceous rocks along the west flank of the Spread Creek Anticline.
2.1	38.4	Burro Hill is on the north side of the Buffalo Fork River. Drab sandstones and shales are near the base of a 5,000-ft (1,524-m) section of the Harebell Formation of Late Cretaceous age.
2.0	40.4	Ledges on the left are part of the sandy facies of the Harebell Formation (the approximate equivalent of the Upper Cretaceous Lance Formation in central and eastern Wyoming). For the next 5 miles (8 km) [to the east], the roadcuts and landslides on steep slopes are unstable strata of the Harebell Formation; much of the highway for the next few miles is prone to sliding each year.
3.9	44.3	Roadcut on south side of highway is site of a dinosaur quarry.
3.2	47.5	Cross Quaternary fault scarp which trends north across highway and offsets glacial debris. The east block is dropped at least 50 ft (15 m) and has ponded drainage.
0.8	48.3	Togwotee Lodge on the left. Straight ahead is Angle Mountain, part of the Precambrian and Paleozoic core of the Washakie Range of Laramide age. The range was completely capped by Eocene volcanoclastic rocks, but is now being exhumed.
1.1	49.4	Ledges and dip slopes on the left are steeply dipping Mississippian Madison Limestone on the south flank of the Washakie Range. In this area, Paleozoic rocks are thrust southward onto Cretaceous rocks along the Buffalo Fork thrust fault which extends north-northwestward for 40 mi (64 km).
0.9	50.3	At the mouth of the canyon on the north side of the highway are outcrops of steeply dipping to overturned Triassic and Jurassic rocks.
0.6	50.9	On the tree-covered slope at left are red beds of Triassic and Jurassic age rocks which are overturned southward on the south flank of the Washakie Range.
0.9	51.8	Cliffs at stream level to the south are gently dipping sandstones in the Upper Cretaceous Frontier Formation.
0.9	52.7	On the south side of Blackrock Creek, extending upstream for approximately one-fourth mile (0.4 km), are white porcelanites which are the lower marker zone in the Upper Cretaceous Frontier Formation. The strata are folded into a sharp east-trending anticline.
0.7	53.4	On the right, halfway up the south bank of Blackrock Creek is the contact between black Upper Cretaceous Mowry Shale [below] and gray massive basal sandstone of the Frontier Formation. Within the rolling bare hills extending south and southeast from Blackrock Creek are dipping strata in the Mowry Shale and Frontier Formation. These Cretaceous rocks are folded into the Blackrock Anticline.
3.5	56.9	Togwotee Pass, elevation 9,544 ft (2,910 m), on the Continental Divide.
1.1	58.0	Entering the Wind River Basin. Lava Mountain, at 2 o'clock, is composed of nearly 1,000 ft (305 m) of Quaternary lava flows.
0.4	58.4	Blue green volcanic siltstone and sandstone (middle Eocene Aycross Formation) in roadcut.
1.4	59.8	Straight ahead are Pinnacle Buttes comprising the middle and upper Eocene Wiggins Formation on the east side of Brooks Lake.

Mileage:		Description:
Increment	Total	
0.5	60.3	Straight ahead is the high flat summit of the Wind River Range in the vicinity of Union Peak. The summit surface, which was probably the upper limit of Tertiary fill, is variously called the Summit (or Subsummit) surface, Wind River peneplane, or Fremont surface.
3.5	63.8	Pinnacle Buttes on left skyline.
1.8	65.6	Variegated beds on left side of road are considered to be in the lower Eocene Wind River Formation (Rohrer, 1966). Green volcanic sandstone at top of roadcut is probably the base of the Aycross Formation.
3.9	69.5	The lowest clearcut at 3 o'clock is approximately 49 years old. It has been hand planted several times and regeneration remains slow. Other clearcuts, chiefly on landslide debris of Eocene rocks, are younger, and trees are likewise slow to regenerate. The nearly horizontal beds on the north side of the highway for the next five miles (8 km) are greenish-gray and drab tuffaceous facies of the lower Eocene Wind River Formation (Keefer, 1957).
2.0	71.5	Lumpy hills on south side of Wind River are a mixture of landslide and glacial debris.
3.5	75.0	Bridge over Long Creek. Hat Butte, on the right (due south) is composed of Eocene-age tuffs overlying Paleozoic rocks which dip northeastward on the flank of the Wind River Mountains. Lumpy topography on south side of Wind River is largely debris of the Pinedale glaciation (late Pleistocene).
1.4	76.4	Pinedale Till in roadcuts marks confluence of ice that flowed down Wind River with the ice that flowed down the Du Noir River Valley.
0.6	77.0	At left is the broad flat-bottomed valley of Du Noir River. Fluted sides at several levels were developed by different episodes of glaciation that scoured the soft variegated strata of the Wind River Formation. The Ramshorn (at 9 o'clock) is composed of nearly horizontal stratified volcanic conglomerates of middle to late Eocene age.
1.9	78.9	At 1 o'clock, across the Wind River and up the north-trending ravine, are two skid roads and several inconspicuous cabins that mark the site of a coal mine in the Wind River Formation. The coal occurrence is unusual because it accumulated in a narrow steep-sided valley cut in the northeast-dipping Tensleep Sandstone (Middle Pennsylvanian to Lower Permian) and Phosphoria (Lower Permian) Formation. Straight ahead across Wind River are bare ledges of the Phosphoria Formation.
0.9	79.8	Stoney Point. The contact between the Tensleep Sandstone (Pennsylvanian here) and the Phosphoria Formation is approximately one-third of the way up the cliff, at the base of a pebbly sandstone.
0.2	80.0	At 9 o'clock, the Wind River Formation laps across the northeast-dipping Phosphoria Formation. In the next one-half mi (0.8 km), from west to east, is a dramatic facies change from drab, non-variegated claystones and silty sandstones to brightly variegated badlands. The badlands have yielded early Eocene fossil mammals typical of the Lost Cabin (upper member of the Wind River Formation) (Keefer, 1957).
1.4	81.4	From 3 to 4 o'clock, Warm Spring Canyon cuts through the Madison Limestone on the east flank of the Wind River Range. Several terrace levels on the southeast side of the canyon are held up by travertine deposits with a maximum thickness of more than 500 ft (152 m).
0.3	81.7	Looking to the right, down the road to Warm Spring Ranch, one can see northeasterly-dipping red Chugwater Formation of Triassic age, unconformably overlain by nearly horizontal travertine deposits at the level of the prominent ridge.
3.9	85.6	Gray cliffs at 9 o'clock are bleached Lower Jurassic Nugget Sandstone.
0.9	86.5	Dubois. Dinner stop.
		Variegated Wind River Formation at 9 o'clock unconformably overlies red Chugwater Formation.

Mileage:		Description:
Increment	Total	
		Dubois started a post office in 1886, and applied for the name Never Sweat, a more or less humorous reference to some men of the settlement who were models of the early couch potato. Postal officials were not enthusiastic, and suggested the name of an Idaho Senator who served on the Senate's committee on post offices. Dubois later grew into a cow town and center for cutting railroad ties. It is now also a tourist center.
0.6	87.1	Nugget Sandstone (Lower Jurassic) on left at Junction of Horse Creek Road with Main Street. The cave carved into the Nugget has been used at various times as a bar and as a jail.
0.2	87.3	Turn right (south).
0.2	87.5	At 2 o'clock, across the river, are bare, red, dip slopes of Chugwater Formation. The orange-colored badlands are in the Popo Agie Member of the Chugwater, and contain fragments of labyrinthodont amphibians.
2.9	90.4	Roadcuts are in Pinedale terminal moraines.
2.1	92.5	Roadcuts are in Sundance and Morrison Formations (Jurassic age) dipping steeply eastward.
0.3	92.8	At 9 o'clock, the highest red cliffs in the badlands are Indian Meadows Formation (lower Eocene) thrust southwestward onto the Lost Cabin Member of the Wind River Formation (lower Eocene). At the base of the red cliffs is a wedge of gray sandstone in the Frontier Formation on the overriding thrust plate.
2.6	95.4	Bridge across Wind River.
3.5	98.9	For the next mile (1.6 km) on the east side of the Wind River are gray ledges of the Sundance Formation, dull variegated Morrison Formation, and brightly variegated and lilac-colored beds in the Lower Cretaceous Cloverly Formation.
3.9	102.8	Red Grade section of Love and others (1947) at 9 o'clock.
0.9	103.7	Roadcut on right (southwest) shows a northwest-plunging syncline in the Thermopolis Shale which is overlain by thin lower Eocene deposits of the Wind River Formation and Quaternary glacio-fluvial gravels.
0.4	104.1	Roadcut through Pinedale moraine.
0.8	104.9	Cross Dinwoody Creek.
0.8	105.7	Highway crosses onto terrace of Bull Lake age glacio-fluvial outwash plain. For the next 3 mi (4.8 km), the highway follows approximate edge of Pinedale-age moraines visible to south, and Bull Lake outwash terrace to north
8.3	114.0	Highway crosses Cottonwood Draw. Roadcuts are in Quaternary terrace gravels.
7.1	121.1	Historical marker for Crowheart Butte. "In March, 1866, a battle was fought in this vicinity between Shoshone and Bannock Indians on one side and Crow Indians on the other. The contest was waged for the supremacy of hunting grounds in the Wind River Basin. Crowheart Butte was so named because the victorious Washakie, Chief of the Shoshones, displayed a Crow Indian's heart on his lance at the war dance after the battle..."
5.4	126.5	Intersection of U.S. Highways 26-287 and Bull Lake Road to the south. See Mears and others (1986) for a discussion of the local glacial geology.
3.5	130.0	Junction of U.S. Highways 26 and 287; stay left on U.S. Highway 26 toward Riverton.
6.5	136.5	Intersection of U.S. Highway 26 and Diversion Dam Road.
4.1	140.6	Morton Reservoir to the east.
2.1	142.7	Wind River High School.
2.6	145.3	Junction Wyoming Highways 133 to Pavillion to the north (left), and 132 to Ethete to the south (right). Road to Pavillion is also access to Pavillion and Muddy Ridge gas fields. Pavillion gas

Mileage:		Description:
Increment	Total	
		field lies several miles northeast of the town and has produced 130.4 Bcf (billion cubic feet) of gas (to August, 1993) from sandstones in the basal part of the Wind River Formation and from both the lower member and Shotgun Member of the Fort Union Formation (Paleocene). Depths of production range from 1,500 to 5,000 ft (427 to 1,524 m). Muddy Ridge field is adjacent to, and north of, Pavillion field, and has produced more than 26.3 Bcf of gas and 80,000 bbl (barrels) of oil (to August, 1993) from the Fort Union and Upper Cretaceous Lance, Meeteetse, and Mesaverde Formations (Flores and Keighin, 1992; Keefer and others, 1993).
0.9	146.2	Kinnear, Wyoming.
14.4	160.6	Road to Riverton airport.
0.5	161.1	City limits of Riverton. A treaty in 1905 removed the future site of Riverton from the Wind River Indian Reservation, prompting a minor land rush. White settlers filed on plots in town, but claim jumpers from Lander arrived to tear up claim stakes. This prompted some armed retaliation, and the arrival of troops from Fort Washakie. Completion of a federally funded irrigation system enhanced the value, and green potential, of much of the land around Riverton. Riverton is the present home of the Brunton Manufacturing Company, maker of the Brunton compass.
0.4	161.5	Central Wyoming College to north (left).
2.3	163.8	Intersection of U.S. Highway 26 and Wyoming Highway 789; turn left toward motel.

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Wind River Basin - Day one, Part 1

Road log and general geology of the Wind River Basin, Wyoming

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Introduction

Riverton, Wyoming, has a population of 9,600 and its industries include agriculture, oil and gas, uranium, light manufacturing (e.g., home base of internationally known manufacturer of the Brunton compass), tourism, and lumber. Riverton is on the southeastern part of the Wind River Indian Reservation (WRIR), which is jointly governed by the Shoshone and Arapaho Tribes. The Eastern Shoshone Tribe of the WRIR moved out of the Nevada-Utah area into the plains during the 1600s, while the Northern Arapaho Tribe migrated from Minnesota into the plains at about the same time (Ortiz, 1993). The Shoshone Tribe, led by Chief Washakie, was friendly to the American immigrants and a Shoshone woman, Sacajawea, led the Lewis and Clark Expedition through the Pacific Northwest. The Arapaho Tribe was considered more aggressive and, with allies from the Sioux and Cheyenne tribes, the Arapaho fought with immigrants seeking gold and land. Boundaries of the Shoshone Reservation, now referred to as Wind River Reservation, were first set by the Fort Bridger treaty of 1863. In this treaty, the reservation included

about 44,672,000 acres (18,078,510 square hectometers) of land covering parts of Colorado, Idaho, Montana, Utah, and Wyoming (Ortiz, 1993). In 1868, the size of the Reservation was reduced to 2,774,450 acres (1,122,804 square hectometers) as the Shoshone Tribe was forced to accept the second Fort Bridger treaty. Finally, in 1874, a federal government agent talked the Shoshone Tribe into selling 500,000 acres (202,347 square hectometers) for the development of gold mining properties, thus reducing the Reservation to 2,274,450 acres (920,457 square hectometers).

Currently, the WRIR is an area of about 3,500 mi² (9,067 km²) bounded by the Owl Creek Mountains on the north and Sand Draw on the south. The reservation begins just west of the town of Shoshoni on the east and extends westward to the town of Dubois. There are about 2,500 Shoshone and 5,000 Arapaho tribal members on the Reservation. Each tribe is governed by its own General Council, which delegates certain powers to a Joint Business Council responsible for day-to-day activities.

Road log

Mileage:		Description:
Incre- ment	Total	
0.0	0.0	Begin Day 1. Junction of State Highway 135 (Sand Draw road) and State Highway 789 (Lander-Riverton highway). See Figure 1 for the field trip route in the Wind River Basin.
0.6	0.6	Bridge over Popo Agie River.
0.4	1.0	Junction of Sand Draw road (State Highway 135) and Gas Hills-Riverton highway (State Highway 136). Trip continues east on State Highway 136.

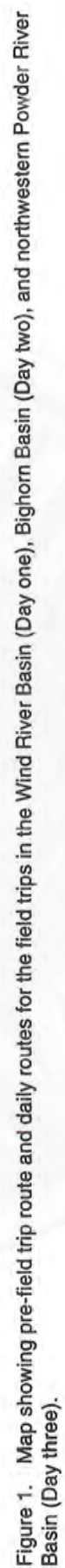


Figure 1. Map showing pre-field trip route and daily routes for the field trips in the Wind River Basin (Day one), Bighorn Basin (Day two), and northwestern Powder River Basin (Day three).

Mileage:		Description:
Incre- ment	Total	
7.1	8.1	To the south is the Conoco East Riverton field. Oil is produced from the Frontier, Muddy, and Cloverly Formations (Cretaceous), the Nugget Sandstone (Lower Jurassic), and the Phosphoria Formation (Lower Permian). Cumulative production from all producing zones (to 1989) is about 68.5 BCFG (Keefer and others, 1993).
3.9	12.0	Eastern boundary of the WRIR. Alkali Butte, which is underlain by Cretaceous and Tertiary rocks, is visible on the skyline to the south.
4.2	16.2	Croft Ranch Road to the north.
1.5	17.7	Poston Ranch Road to the south.
2.3	20.0	Campbell Ridge, which is underlain by the Eocene Wind River Formation, is visible to the north (left). The Fuller Reservoir oil field lies about 8 mi (13 km) north of this point. Oil and gas production in this field are from the Shotgun Member (upper part) of the Paleocene Fort Union Formation at depths ranging from 3,000 to 4,000 ft (915-1,220 m). Cumulative production (to 1989) was 1,390,928 BO (barrels of oil) and 8.5 BCFG (billion cubic feet of gas) (Keefer and others, 1993). The Haybarn field is 6 mi (10 km) east of the Fuller Reservoir field; production is also from the upper part of the Fort Union Formation. Cumulative production in this field (to 1989) was 232,500 BO and 583 MCFG (million cubic feet of gas). The Waltman Shale Member (middle part) of the Fort Union Formation is the source for these hydrocarbons (see Palacas and others, this volume).
2.8	22.8	Outcrops of the Eocene Wind River Formation are on the south side (right) of the highway.
4.5	27.3	Road crossing near milepost 27. Two and one-half miles south is Signor Ridge, which is underlain by rocks of the upper part of the Fort Union Formation. The fluvial channel sandstones and conglomerates of the Fort Union at this locality are oil stained, much like those we will visit at Castle Gardens. These rocks were palynologically dated by D.J. Nichols (written communication, 1993) as late Paleocene (P5-P6 biozone). The reservoir characteristics of these rocks are described by Keighin and Flores (this volume) and the geochemistry of the oil in these reservoirs is discussed in Palacas and others (this volume).
8.5	35.8	Top of a hill. Coal-bearing rocks of the Paleocene Fort Union and Upper Cretaceous Lance Formations crop out north (left) of the highway.
1.7	37.5	Junction of the Castle Gardens road and Gas Hills-Riverton highway (State Highway 136). Continue north on the Castle Gardens road.
0.6	38.1	Stop 1- Orientation. This hill provides a fine view of the southern boundary of the Wind River Basin (the Beaver Divide), which is prominent along the skyline to the south. This divide is an erosional-remnant of upper Eocene and Oligocene rocks that once infilled the Wind River Basin. Arkosic rocks of the Eocene Wind River Formation are major host rocks of uranium deposits in the Gas Hills mining district (see reclaimed areas to the south; Love, 1992). Estimates on the amount of uranium available from the arkoses on the Beaver Divide north to the roll fronts in the Gas Hills suggest that the Wind River Formation alone could have provided enough uranium to account for the total reserves in the Gas Hills district. Restriction of uranium mineralization to the arkosic sandstones indicates that the arkoses served as both a source and host rock. In the middle distance to the southeast, the Dutton Basin anticline is partially exhumed, exposing Mesozoic rocks underlying the Wind River Formation. On the skyline to the north is Castle Gardens, marked by a northward-dipping monocline in strata of the Paleocene Fort Union Formation and Upper Cretaceous Lance Formation.
2.0	40.1	To the west (left), across Muskrat Creek, are rusty orange, fluvial channel sandstones and interbedded coal beds of the Fort Union Formation.
3.7	43.8	Bureau of Land Management (BLM) road to the east leads to the Castle Gardens area. Castle Gardens, a BLM managed park, is about 5.8 mi (9.3 km) east on this road; the park is well known for its Indian petroglyphs and pictographs (described at Stop 5). The road traverses outcrops of

Mileage:
Incre-
ment Total

Description:

the Paleocene Fort Union Formation. The contact with the underlying Upper Cretaceous Lance Formation is marked by K/T sandstone beds composed of white to rusty, continuous sandstone units visible as hogbacks southeast (left) of the road and across the valley. About 0.7 to 1.2 mi (1.1-2.1 km) along the road and immediately to the north is a prominent bleached white, fluvial channel sandstone that contains abundant crayfish burrows (Figure 2; Flores and others, 1992; Hasiotis and others, 1994).

- 3.0 46.8 **Stop 2-Stratigraphic, sedimentologic, and palynologic orientations, and "hands-on" experience.** The Paleocene Fort Union Formation in the Castle Gardens area is at least 2,325 ft (709 m) thick and is underlain by the Upper Cretaceous (Maastrichtian) Lance Formation and overlain by the Eocene Wind River Formation. Excellent outcrops of these formations are present in a belt about 3 mi (5 km) wide and 6 mi (10 km) long. The Fort Union Formation consists of interbedded conglomeratic sandstones, sandstones, siltstones, mudstones, carbonaceous shales, and coals. A measured stratigraphic section showing lithology, depositional facies characteristics, and palynologic (spore and pollen) biozones is shown in Figure 3. Here, the Fort Union formed mainly in fluvial environments consisting of low-sinuosity and braided fluvial channels, floodplain-crevasse splays, mires, and paleosols (histosols, inceptisols, alfisols, and spodosols). The formation is subdivided into palynomorph biozones from P1 (oldest) to P6 (youngest) (Nichols and Ott, 1978; Nichols, this volume) based on 65 samples from coal and carbonaceous shale beds. These biozones permit the correlation of undifferentiated Fort Union Formation rocks in the Castle Gardens area to those rocks included in the lower unnamed member (lower part), Waltman Shale Member (middle part), and the Shotgun Member (upper part) of the Fort Union Formation through the basin axis and to the Waltman area (Stop 6) as shown in Figure 4. These members were deposited in fluvial, deltaic, and lacustrine environments. Vitritine reflectance values of surface samples of the Fort Union coals in the Castle Gardens area range

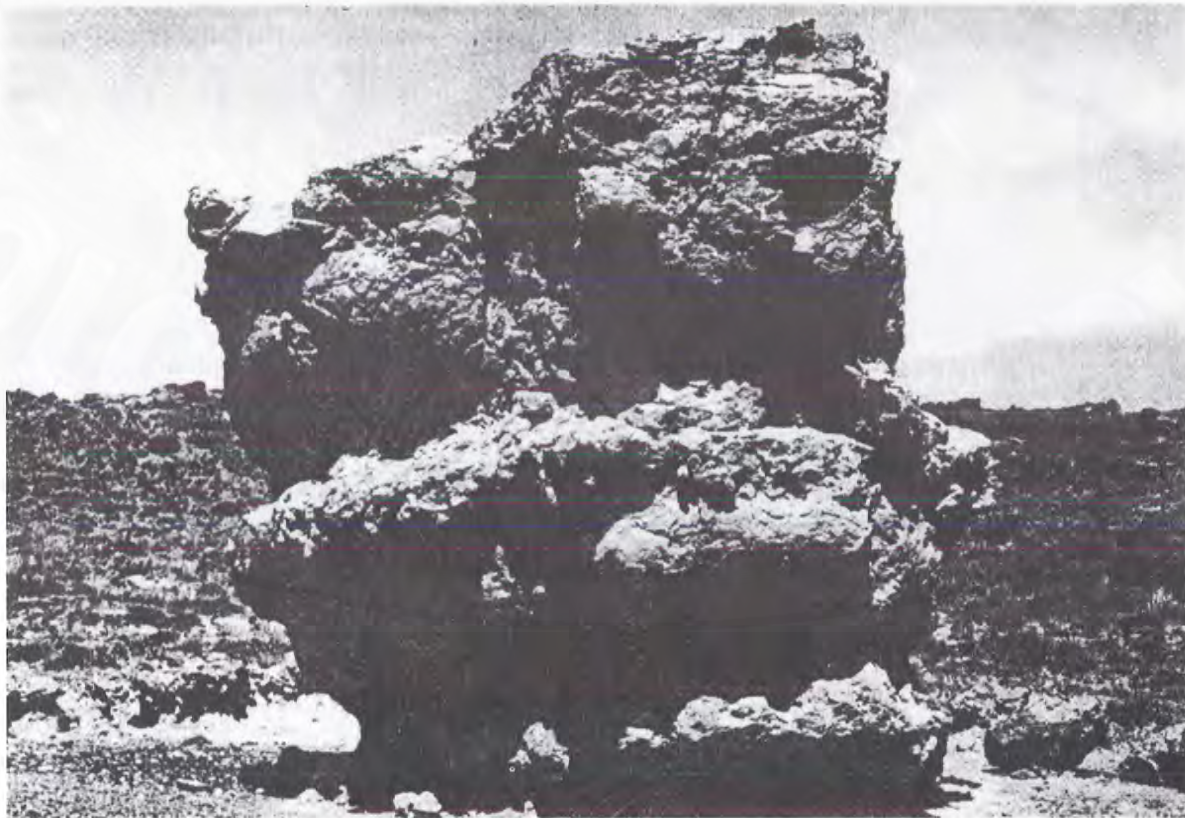


Figure 2. Photo showing zones of crayfish burrows in fluvial channel sandstones of the Fort Union Formation.

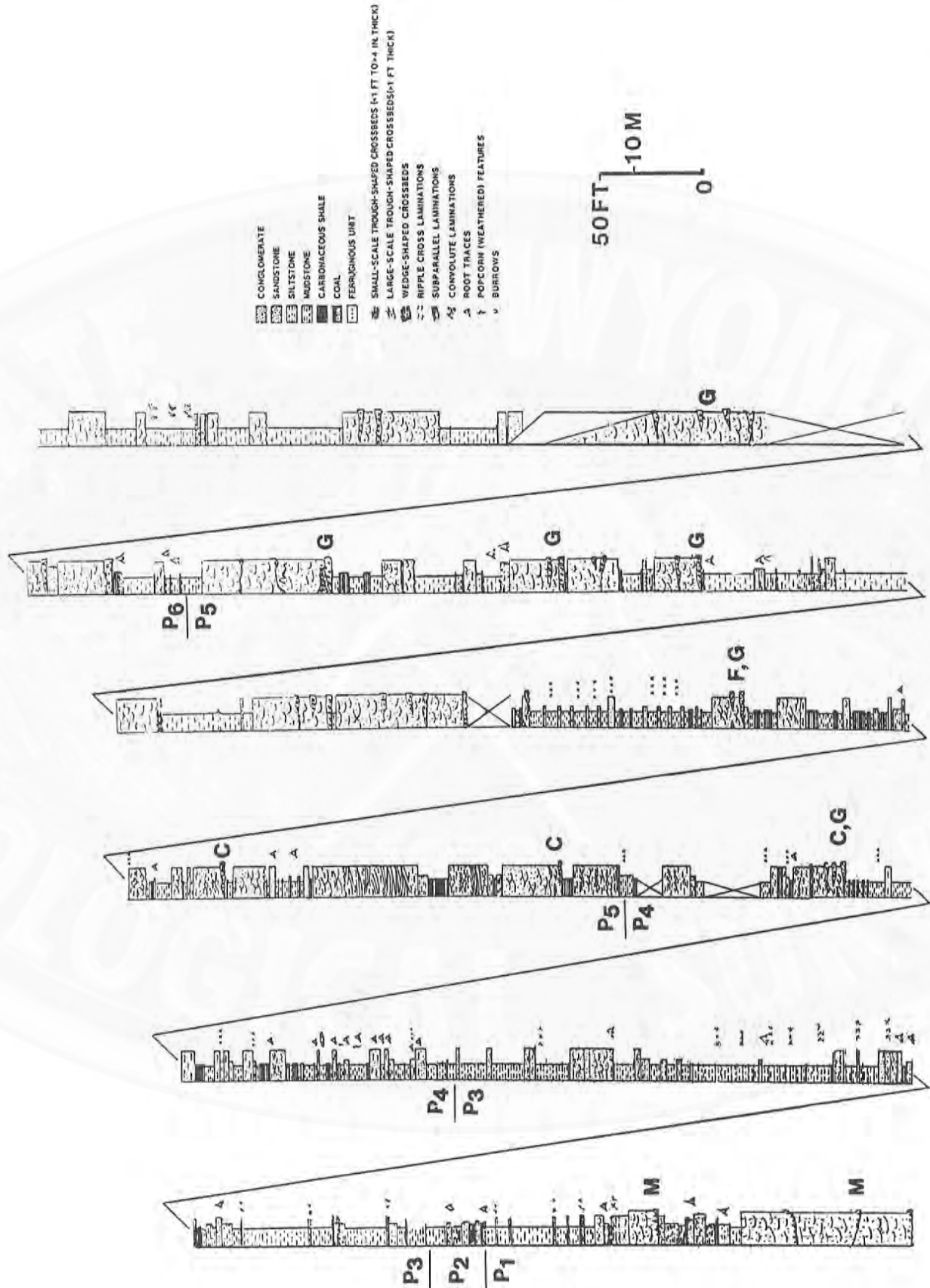


Figure 3. Composite stratigraphic section of the Fort Union Formation in the Castle Gardens area, showing lithology, sedimentary structures and facies types. (M = mudstone, C = chert, F = feldspar, G = granite fragments).

Mileage:
Increment Total

Description:

from 0.36 to 0.40% (Nuccio and others, 1993). In contrast, reflectance values of subsurface coals are higher (0.6 to 0.9%) (Nuccio and Finn, this volume). Fort Union coals in the Wind River Basin attain cumulative thicknesses of as much as 232 ft (71 m), with individual beds as much as 23 ft (7 m) thick and merging to as much as 39 ft (12 m) thick (Johnson and Flores, 1993; Johnson and others, this volume). The majority of the coals are in the lower unnamed member and the Shotgun Member.

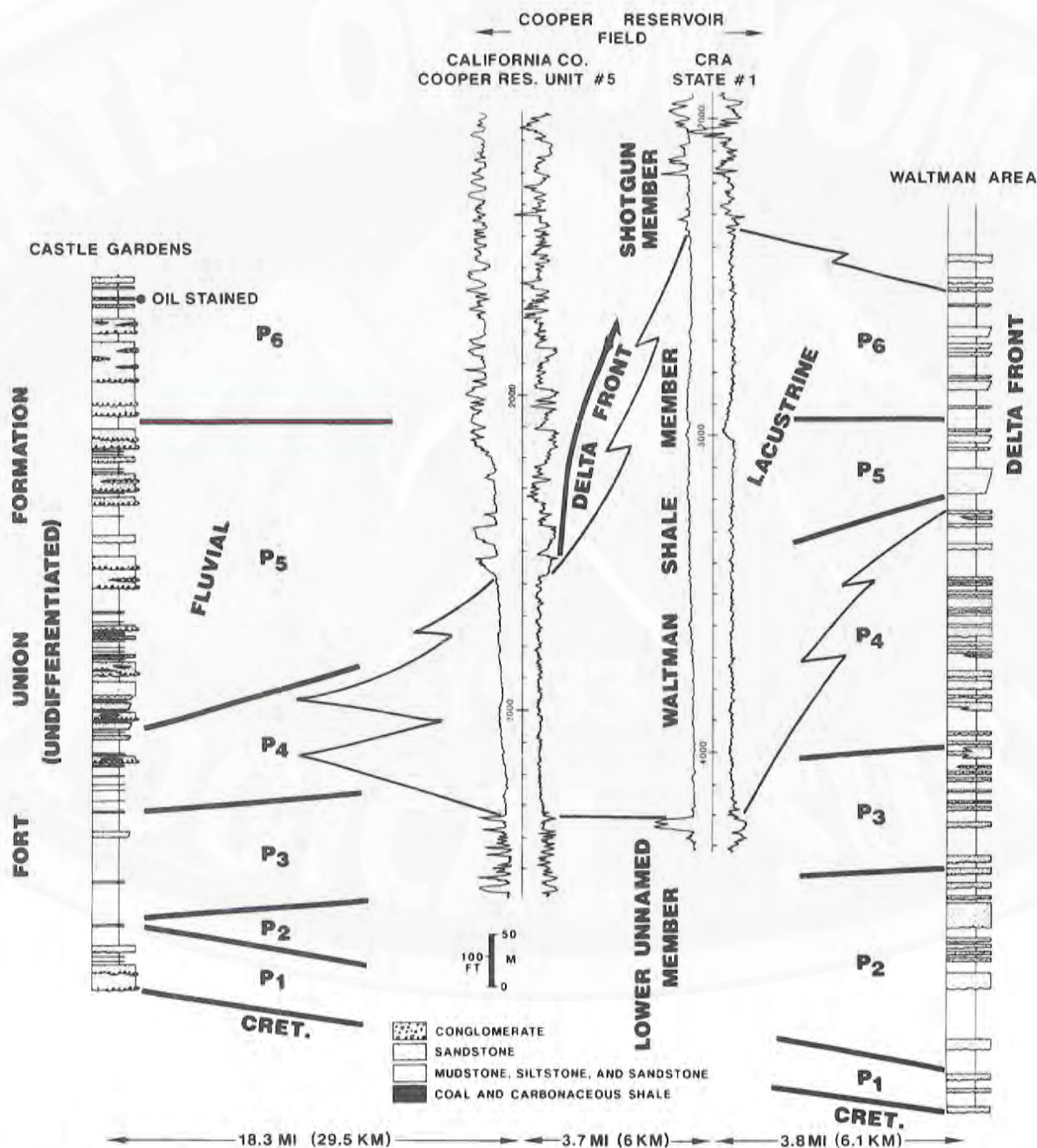


Figure 4. Cross section of the Fort Union Formation from the Castle Gardens area through the Cooper Reservoir field and to the Waltman area.

Mileage:		Description:
Incre- ment	Total	
0.4	47.2	Take jeep trail to the north (left) from the BLM road.
1.9	49.1	Stop 3- Oil-stained fluvial channel sandstone and geochemistry of the oil. The oil-stained fluvial channel sandstone is in the uppermost part of the Fort Union Formation (P6 biozone; see Figure 3). The bleached appearance of the sandstone is due to weathering of the oil stain. The sandstone is as much as 15 ft (4.6 m) thick and greater than 0.5 mi (1.2 km) wide, and is composed of multiscoured reservoir bodies (see Keighin and Flores, this volume). Gas chromatography analyses indicate that the oil in the sandstone has sustained biodegradation water washing, and evaporation (see Palacas and others, this volume). Based on biomarker signatures, the oil is mature and similar to the oil hosted by the Fort Union Formation in the Fuller and Haybarn fields; that is, rich in higher plant and coaly-type organic matter (see Palacas and others, this volume). These analyses also suggest that these oils were sourced from organic-rich shales and mudstones of the Waltman Shale Member (lacustrine).
1.9	51.0	Return to the BLM road and turn to the east (left).
0.4	51.4	Stop 4- Overview of the lower Fort Union badlands. The badlands are carved out of the weathering-resistant fluvial channel and crevasse splay sandstones and less-resistant floodplain-mire-paleosol siltstones, mudstones, carbonaceous shales, and coals. The K/T sandstones are the most prominent beds visible to the south and will be seen in more detail in the next stop at the BLM park.
2.0	53.4	Stop 5- BLM park. Overview K/T sandstones and Indian petroglyphs and pictographs. The K/T sandstones in the park include a zone of at least three stacked, thick, amalgamated sandstone beds interbedded with thin coals, carbonaceous shales, mudstones, and siltstones (Figure 5). This zone of K/T sandstone beds is bounded by an overlying coal bed in the P1 biozone, and an underlying coal bed dated as Maastrichtian. The amalgamated sandstone beds were deposited in braided to meandering fluvial channels that may have developed in a major drainage system of an incised paleovalley that extended from the western margin of the Wind River Basin to the southwest margin of the Powder River Basin (see Flores and others, this volume). The pictographs and petroglyphs represent Indian painted symbols and carved, pecked, or incised figures or symbols, respectively (Randall, 1961). Petroglyphs are better preserved and more abundant than the pictographs, which need shelter from weathering for preservation.
5.8	59.2	Junction of Castle Gardens-BLM road; trip continues to the north (right).
2.5	61.7	Quaternary windblown sand dune deposits are visible in the road cuts.
1.7	63.4	Outcrops of variegated units of the Eocene Wind River Formation on the left side of the road.
10.9	74.3	Junction with U.S. Highway 20/26. Turn east (right).
8.6	82.9	Road on left goes to Frenchie Draw field. Cumulative production (to date) in this field from the Fort Union Formation has been 40 BCFG and 1.2 million BO (Keefer and others, 1993).
1.5	84.4	Natrona-Fremont county line.
3.2	87.6	Road cuts in Quaternary sand dune deposits.
5.7	93.3	Hiland Store.
6.3	99.6	Rest Area. Lunch stop. The north end of the Rattlesnake Hills is visible on the skyline to the south. The Rattlesnake Hills are underlain by Cretaceous rocks and the coal-bearing Paleocene Fort Union Formation. Here, we cross the Hiland divide which separates the Powder River and Wind River drainage systems.
2.5	102.1	Waltman Store. The road to the south leads to Rattlesnake Hills, Gas Hills district, and Cooper Reservoir field. The trip continues to the north (left) on the road (State Highway 104) to the town of Armino.



Figure 5. Photo showing the K/T boundary fluvial sandstones in the BLM park.

Mileage:		Description:
Increment	Total	
2.3	104.4	Road junction to gas wells. Take jeep trail to the east (right).
1.9	106.3	Stop 6- Waltman Shale Member type section. This locality is the type section of the Waltman Shale Member of the Fort Union Formation (Keefer, 1961). Here, the Waltman Shale Member (Figure 6) is 645 ft (197 m) thick and is composed of chocolate brown and gray silty mudstones, and fissile shales interbedded with thin beds of bleached, ledge-forming sandstones. The sandstones, which are lenticular, fine-grained, sharp-based, and trough crossbedded and ripple laminated, are interpreted as delta-front deposits (better developed in Stop 6A). The Waltman Shale Member is interpreted as deposits of a freshwater lake (Waltman Lake). The member is unconformably (note the angular discordance) overlain by the Eocene Wind River Formation



Figure 6. Photo of the Waltman Shale Member of the Fort Union Formation in the reference locality. Coarsening-upward interval is delta-front siltstone and sandstone.



Figure 5 continued.

Mileage:		Description:
Incre-	Total	
ment		
		and its base is in contact locally with Wind River strata emplaced by normal faults of small displacement (Keefer, 1961, 1965). These outcrops comprise vertical and overturned beds along the west flank of the Casper Arch. A geochemical study of the Waltman Shale Member by Katz and Liro (1993) indicates that it is kerogenous; much of the organic matter displays a mixed character intermediate between type II and type III kerogens, probably in part, due to presence of recycled organic matter. Pyrolysis-gas chromatography indicates the generated products are a high wax crude oil. Additional geochemical investigation of the Waltman Shale Member by Palacas and others (this guidebook) elaborates on the origin and typing of this oil as well as the recognition of pathways and timing of the oil migration. Rock-eval analysis of surface shale and mudstone samples (Nuccio, personal communication, 1994) from this locality and from Stop 6A indicates that they are highly oxidized due to weathering. However, limited vitrinite reflectance analyses of the Waltman Shale Member by Katz and Liro (1993) from three wells indicated profiles and variations of vitrinite reflectance from less than 0.5% to greater than 0.7% (Figure 7). Palynomorph biozones of the Waltman Shale Member by Nichols (this guidebook) and Nichols and Ott (1978) were established at this locality. Stratigraphic correlation of the undifferentiated Fort Union Formation in the Castle Gardens area and the type locality is shown in Figure 4.
1.9	108.2	Return to the Arminto road (State Highway 104) junction and turn south (left) to the Waltman Store.
2.3	110.5	Road junction with U.S. Highway 20/26. Turn east (left).
3.0	113.5	Road to the north (left) goes to a ranch house. Turn left here.
2.5	116.0	Alternate Stop 6A for Waltman Shale Member at North Hells Half Acre. Exposure of the Waltman Shale Member at this locality is 3.5 mi (5.6 km) southeast of the type section. Here, the Waltman Shale Member is about 600 ft (182 m) thick and consists of numerous, well developed delta-front sandstones interbedded with lacustrine shales, mudstones, and siltstones.
2.5	118.5	Return to road junction of U.S. Highway 20/26; trip continues to the west (right).
31.5	150.0	Moneta. Turn north to oil and gas fields near Lysite and Lost Cabin, including the Madden Unit. Up to 1989, the Madden Unit produced 839,837 BO and 211.7 BCFG from the Paleocene Fort Union Formation and Upper Cretaceous Lance and Mesaverde Formations and Cody Shale (Sundell and Anderson, 1992). Several wells have been drilled to the Mississippian Madison Limestone at depths of 23,000-24,000 ft (7,012-7,317 m); the estimated recoverable gas potential from these wells alone is 20,000-30,000 MCFG (one thousand cubic feet of gas). The gas from the Madison Limestone contains 19% CO ₂ and 11% H ₂ S. This deep structural trap with more than

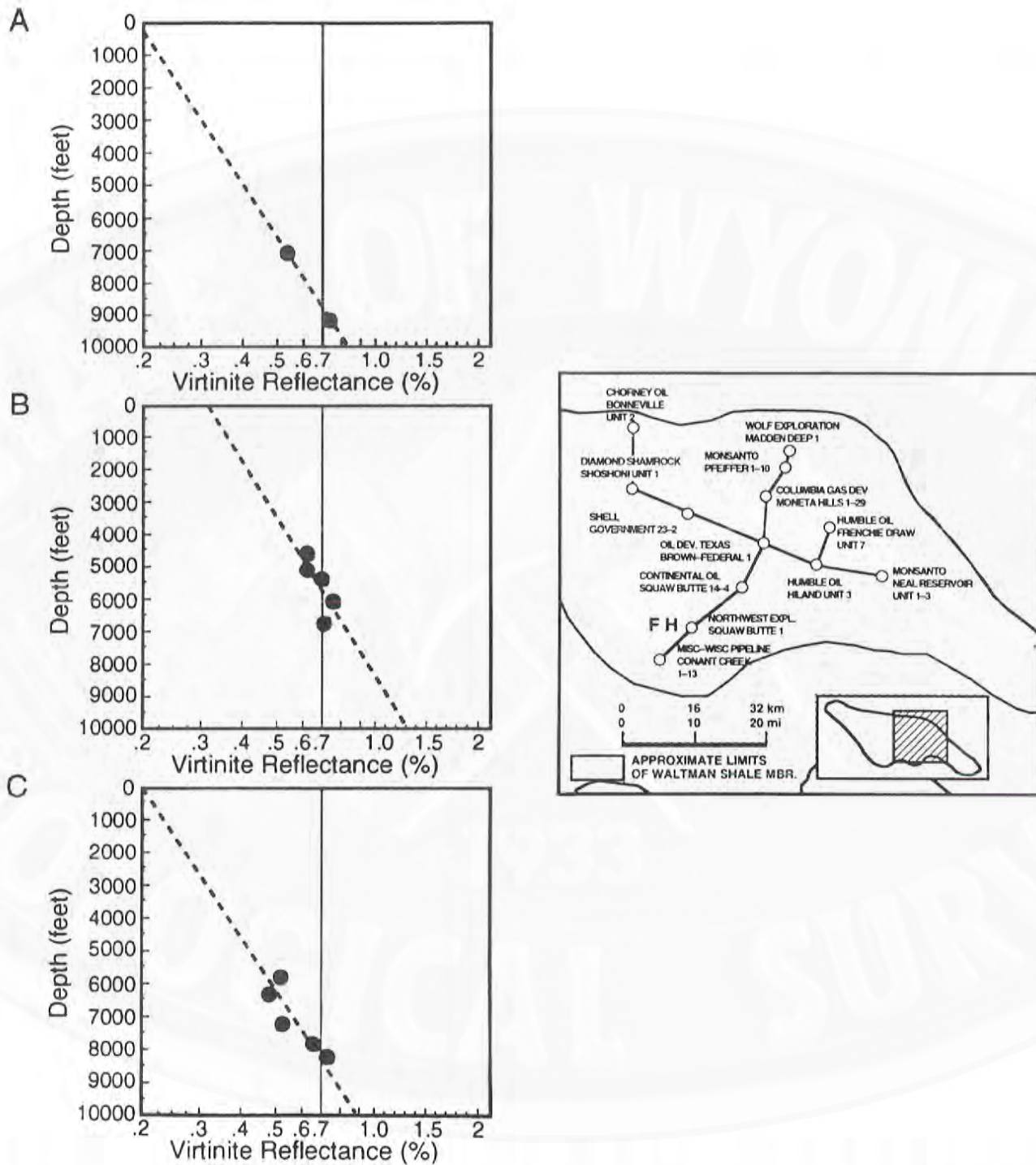


Figure 7. Vitrinite reflectance profiles from the Shoshoni Unit No. 1 (A), Brown Federal No. 1 (B), and Bonneville Unit No. 2 (C). Adapted from Katz and Liro (1993).

Mileage:		Description:
Incre- ment	Total	
		4,500 ft (1,372 m) of closure covers about 40,000 acres (16,188 square hectometers), and could potentially produce several trillion cubic feet of gas. The outcrops between Moneta and Lysite consist of the Eocene Wind River Formation, which includes interbedded fluvial channel sandstones, floodplain mudstones, and variegated paleosols.
8.3	158.3	Lysite, the hub of the Greater Madden field. Turn east (right) at the stop sign 1 block past the grocery store/gas station.
4.2	162.5	Stop 7- Lost Cabin field (Comanche Oil, operator). The Lost Cabin field encompasses about 1.5 mi ² (2.5 km ²) area, and produces from the Shotgun Member of the Paleocene Fort Union Formation. Reservoirs in fluvial channel sandstone are both continuous and discontinuous. Cumulative production is 472,340 BO and 11,569,915 MCFG (to 1989) (Wyoming Geological Association, 1989). Specific gravity of the oil is 36° API. The estimated ultimate recovery is 600,000 BO and 12 BCFG. The geochemical characteristics of this oil are reported in Palacas and others (this volume). Biomarker signatures of the oil suggest that the source is the Waltman Shale Member of the Fort Union Formation.
11.5	174.0	Return to Moneta. At junction of U.S. Highway 20/26, turn to the west (right).
8.9	182.9	Sand dune deposits on the right side of the road.
4.0	186.9	To the right are outcrops of the Eocene Wind River Formation consisting of fluvial channel and crevasse splay sandstones and floodplain mudstones and siltstones.
7.9	194.8	Town of Shoshoni just east of the WRIR. Junction of U.S. Highway 20/26 and State Highway 789. Turn north (right) on U.S. Highway 20 and continue toward the Wind River Canyon and the Owl Creek Mountains. Outcrops north of the town consist of the Eocene Wind River Formation.
8.0	202.8	On the skyline to the west (left) is the Wind River Range. To the east are beds of the Eocene Wind River Formation, which dip north toward the Owl Creek Mountains.
3.0	205.8	At the top of a hill. The Upper Mississippian to Middle Pennsylvanian Amsden Formation, Mississippian Madison Limestone, and Middle and Upper Ordovician Bighorn Dolomite may be seen to the northeast (right). These formations underlie the Middle and Upper Pennsylvanian Tensleep Sandstone.
0.8	206.6	Eocene Wind River Formation onlaps onto Mesozoic rocks.
1.3	207.9	Access road to Boysen Reservoir (Lake) on left and ahead, the south entrance to the Wind River Canyon.
0.4	208.3	A south-dipping outcrop of the Ordovician Bighorn Dolomite can be seen across the reservoir. Immediately south of this outcrop is a wedge of Mississippian Madison Limestone bounded on the north by a normal fault. South of the Madison Limestone wedge, a thrust fault with a maximum displacement of 2,800 ft (854 m) dips 50° north (part of the Owl Creek Mountains fault system). An excellent road cut exposure of the Upper Cambrian Gallatin Limestone is on the east (right) side of the highway. An aerial view looking north into the canyon along the course carved out by the Wind River through the central Owl Creek Mountains is shown in Figure 8. The superimposed Wind River eroded into the soft Tertiary sedimentary rocks in the Wind River Basin and the Bighorn Basin (north of the Owl Creek Mountains) and cut this spectacular gorge into resistant Paleozoic sedimentary and Precambrian rocks of the exhumed Owl Creek uplift. The stratigraphic column of the Paleozoic rocks and underlying Archean rocks in the Wind River Canyon is shown in Figure 9.
2.3	210.6	First of three tunnels in the Precambrian rocks.
1.1	211.7	Stop 8- Overview of the Precambrian rocks. These Archean rocks, which may be part of a larger intrusive body, consist of light-colored, pinkish-gray to white intrusive dikes and pegmatites of quartz monzonite (Gwynne, 1938; Condie, 1967; Maughn, 1987). The darker Archean rocks consist of dark gray to greenish-black, hornblende amphibolite and biotite-quartz schist (Condie,



Figure 8. Aerial photo looking north into the Wind River Canyon in central Owl Creek Mountains. Boysen dam and reservoir are at the bottom of the photo. PIPu = Park City Formation, Tensleep Sandstone, and Amsden Formation; MOu, Madison Limestone and Bighorn Dolomite; Cu = Cambrian rocks, undivided; pCr = Precambrian igneous and metamorphic rocks; f-f = Boysen normal fault. Photograph courtesy of P.T. Jenkins and L.P. House. Adapted from Keefer (1965).

Mileage:

Increment Total

Description:

1967). The protolith (parent rock) of these metamorphic rocks may have been fine- to medium-grained clastic sedimentary rocks (Condie, 1967), or tholeiitic basalts and interlayered high silica basalt and basaltic andesite (Mueller and others, 1985). This rock complex contains the oldest rocks in Wyoming, which were formed about 2,900 Ma and were metamorphosed about 2,750 Ma (Maughn, 1987). These Archean rocks were beveled and are unconformably overlain by the Middle Cambrian Flathead Sandstone. This unconformable contact is best displayed at milepost 117.6, where a layer of granite, 3 ft (1 m) thick or less is found along the contact.

- 10.0 221.7 **Stop 9- Overview of Carboniferous sedimentary rocks (near milepost 127).** Here the Middle and Upper Pennsylvanian Tensleep Sandstone comprises an upper sandy, cliff-forming member and a lower sandy dolomite member that forms the underlying slope. The Tensleep Sandstone is 565 ft (172 m) thick and unconformably overlies the Mississippian and Pennsylvanian Amsden Formation and, locally, the Mississippian Madison Limestone. The dolomitic member consists of silty to sandy dolomite, dolomitic sandstone, and mudstone; sand content increases upward. The overlying sandy member is commonly crossbedded with low- to high-angle, wedge planar sets. The dolomite member was primarily deposited in a shallow marine environment, whereas the upper sandy member was deposited in eolian/interdune pond environments as sand, transported from the north to south, prograded across a shallow submerged shelf (Maughn, 1987).

Mileage:
Increment Total

Description:

0.8 222.5 The highway crosses the Bighorn River in a valley at the southern edge of the Bighorn Basin. Here the Wind River flows from the canyon into the Bighorn Basin and becomes the Bighorn River. Mesozoic rocks are well exposed along the valley walls. Triassic rocks include the Dinwoody Formation, Red Peak Formation, Crow Mountain Sandstone, and Popo Agie Formation; the latter three formations also compose the Chugwater Group. Jurassic rocks (not readily seen from the highway) include the Gypsum Spring, Sundance, and Morrison Formations. These strata dip northward into an east-west fault on the northern side of Thermopolis; this fault localizes groundwater that issues forth from the Bighorn Hot Springs in Thermopolis.

5.0 227.5 End of Day 1- Stop at Holiday Inn, Thermopolis.

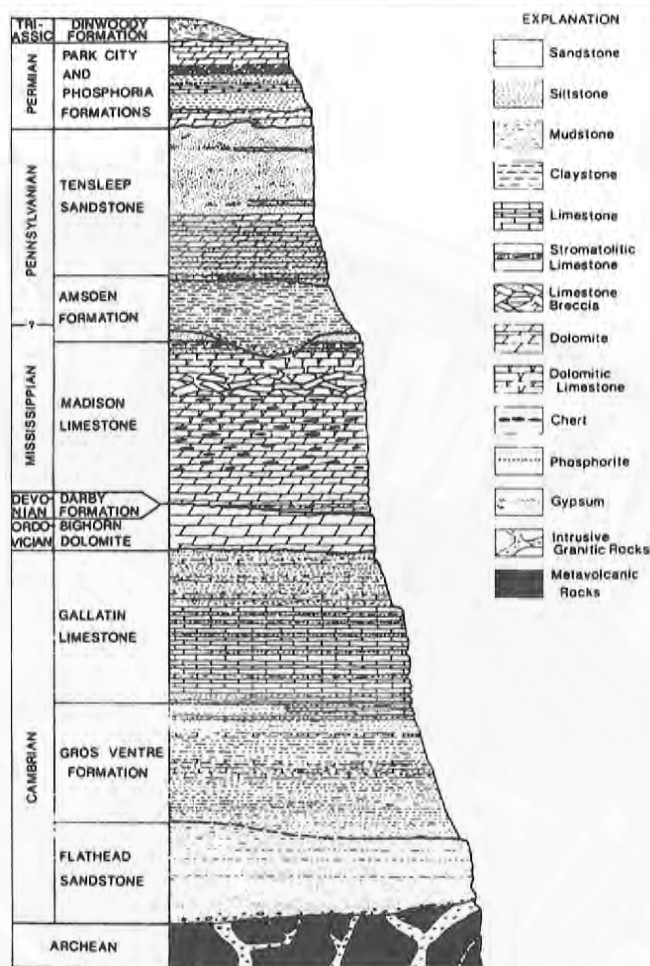


Figure 9. Composite stratigraphic column of the Paleozoic rocks in the Wind River Canyon. Adapted from Maughn (1987).

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Structural and thermal history of the Paleocene Fort Union Formation, central and eastern Wind River Basin, with emphasis on petroleum potential of the Waltman Shale Member

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Introduction

Recent discoveries of oil in the Paleocene Fort Union Formation of the Wind River Basin, Wyoming (Figure 1), have sparked an interest in the organic geochemistry, and burial and thermal history of this formation. The organic geochemistry of the thickest and richest source rock in the Fort Union Formation, the Waltman Shale Member, has recently been characterized by Katz and Liro (1993), and Palacas and others (1993; 1994, this volume). Organic geochemistry is important because it characterizes the type of organic matter contained in the source rock and the type of petroleum that could be generated. The structural and thermal history of the formation are equally important because they detail the burial, temperature,

and thermal maturation of the source rock, and the timing of petroleum generation. The purpose of this report is to present structural and thermal data on the Fort Union Formation in the central and eastern Wind River Basin using cross sections, burial history curves, and vitrinite reflectance data. A kinetic model of petroleum generation is presented for the Waltman Shale Member in the deep part of the basin, and illustrates the timing of oil and gas generation. By combining organic geochemistry data with the structural and thermal history, one can characterize the petroleum potential of the Fort Union Formation and develop a strategy for petroleum exploration in the Wind River Basin.

Geologic setting

The Wind River Basin of central Wyoming is a sedimentary and structural basin that formed during the Laramide orogeny. The basin is bounded by the Owl Creek Mountains on the north, the Casper arch on the east, the Granite Mountains on the south, and the Wind River Range on the west (Figure 1). Strata of early Eocene age occur at the surface throughout most of the central part of the basin (Keefer, 1970).

The pre-Late Cretaceous geologic setting of the Wind River Basin comprised a foreland or stable shelf region that was situated east of the main miogeosynclinal trough. Strata of all systems, except possibly the

Silurian, were deposited during repeated transgressions and regressions of the epicontinental seas across central Wyoming (Keefer and Johnson, 1993).

Beginning in Late Cretaceous time, sediment accumulation shifted eastward in response to tectonic activity in the Overthrust Belt near the present Idaho-Wyoming state line. The Cody Shale represents the last major episode of marine deposition. Much of the overlying Mesaverde Formation is characterized by a series of clastic sediments of coastal floodplain, and swamps, deltas and lagoons (Keefer and Johnson, 1993). Similar conditions persisted during deposition

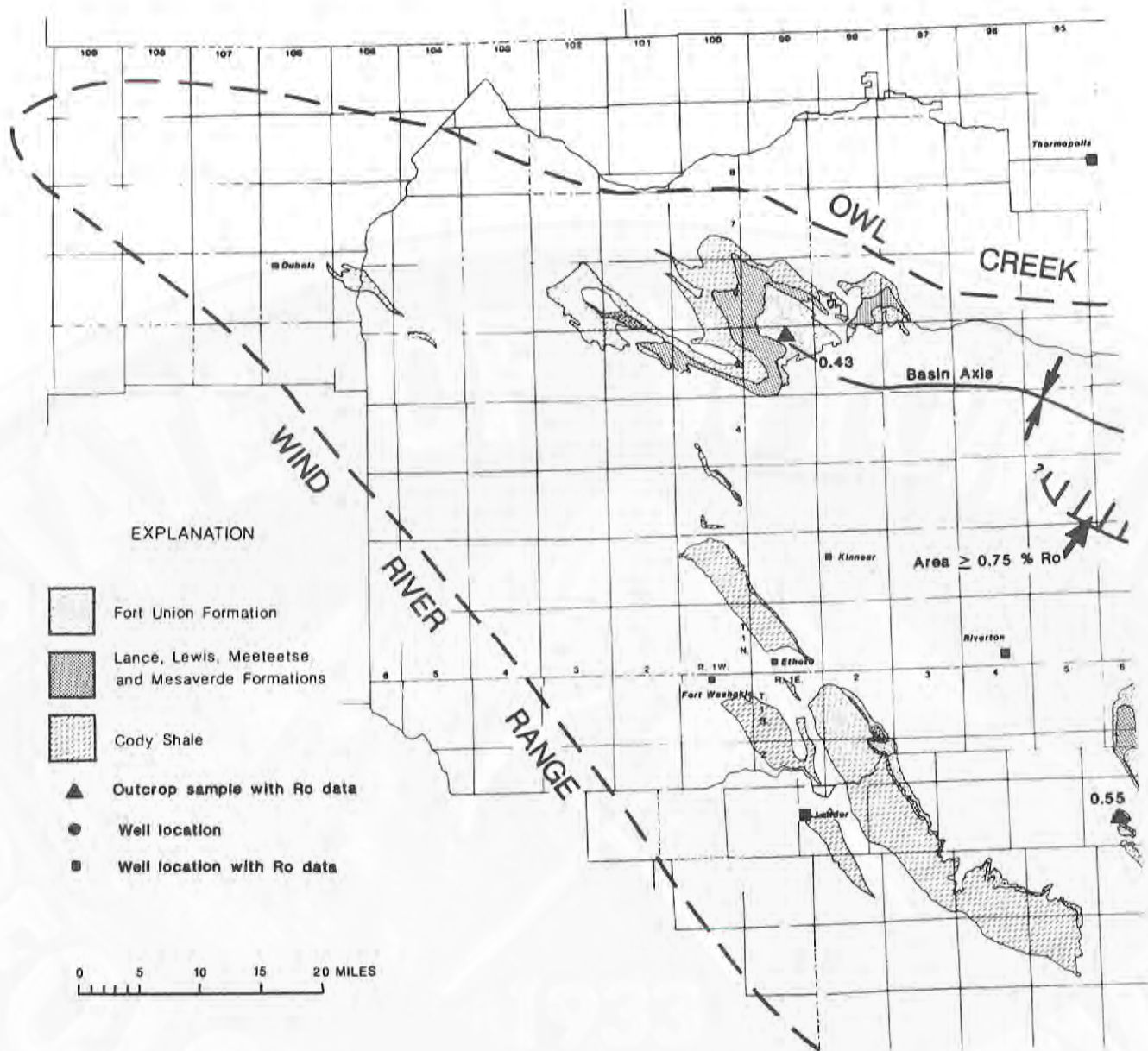


Figure 1. Index map of the Wind River Basin, Wyoming, showing major physiographic features, outcrops of Upper Cretaceous and Tertiary strata, and location of wells (solid circles) used to construct lines of cross sections A-A' and B-B'. Location of wells with vitrinite reflectance data (small squares), and outcrop samples with vitrinite reflectance values (triangles) also are shown. MA = Madden anticline; HRA = Howard Ranch anticline; FR = Fuller Reservoir; HB = Haybarn field. Hachures indicate where the Waltman Shale Member has an R_o of 0.75 percent or greater.

of the next younger Meeteetse Formation/Lewis Shale, but during deposition of the Lance Formation in latest Cretaceous time, downwarping along the axis of the Wind River Basin resulted in deposition of coarse clastics. This initial phase of Laramide deformation was followed by a period of pronounced mountain-building and basin subsidence in Paleocene and Eocene time (see Flores and others, this volume).

In Paleocene time, thick sequences of fine-grained sands, silts, clays, and carbonaceous sediments of the Fort Union Formation were deposited in a subsiding trough in the central and eastern part of the Wind River Basin. In late Paleocene time, the main Wind River Basin trough was flooded by an extensive lake, "Waltman Lake", in which a thick sequence of lacustrine sediments accumulated. This lacustrine

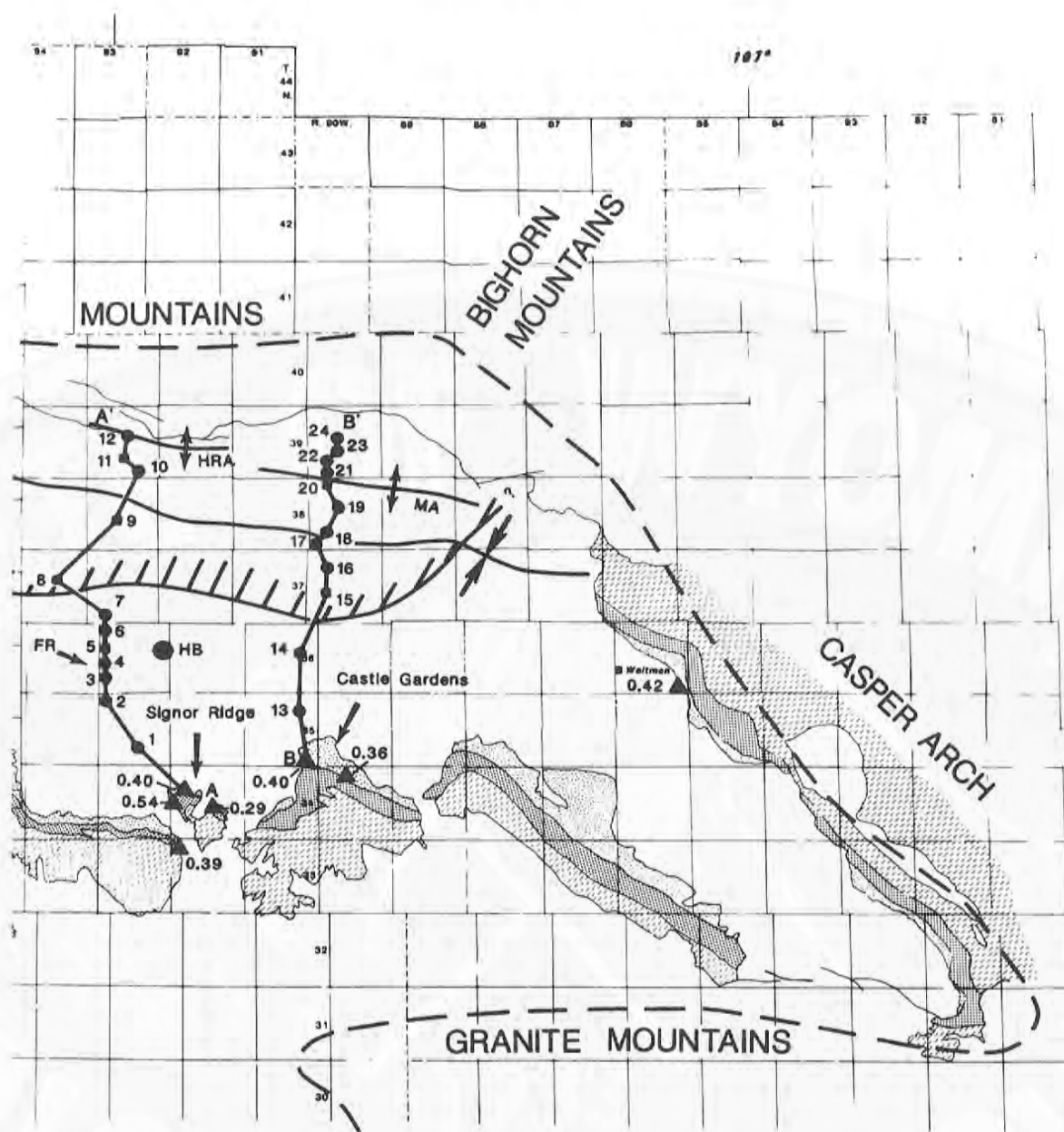


Figure 1 continued.

unit, the Waltman Shale Member of the Fort Union Formation, is of interest to explorationists and is the principal subject of this report.

Fort Union Formation stratigraphy

The Fort Union Formation (Figures 2A and B) overlies the Upper Cretaceous and locally Paleocene Lance Formation with erosional or angular unconformity in nearly all exposures around the margins of the Wind River Basin (Keefer, 1965). The formation crops out along the east, south, and west margins of the

basin and extensively in the northwestern part, in the vicinity of Shotgun Butte (a conspicuous landmark in the southeastern part; T.6N., R.1E.; Figure 1). The thickness of the Fort Union varies along the margins of the basin owing to the erosional unconformity at the base of the overlying lower Eocene rocks. Along the west and south margins, the thickness ranges from 200 ft to approximately 2,300 ft, (61 to 700 m) whereas to the north and east, toward the basin axis, it may be as thick as 8,000 ft (2,438 m) (Keefer, 1965).

The Fort Union Formation is generally divided into three members; the lower unnamed member, the

System	Series	Stratigraphic Unit or event	Thickness (in feet)	Ages (Ma)
Tertiary	mid. Mio.-present	uplift and erosion	-3,380	15 - 0
	Miocene	Split Rock Formation	900	24 - 15
		White River Formation	500	38 - 24
	Eocene	Wagon Bed and Wind River Formations	1,980	55 - 38
		Wind River and Indian Meadows Formations undifferentiated	4,625	
	Paleocene	Fort Union Formation	Shotgun Member	59 - 55
			Waltman Shale Member	60 - 59
			Lower member	66 - 60

reconstructed thicknesses

Figure 2A. Tertiary stratigraphy in the deep part of the eastern Wind River Basin, Wyoming. Thickness and age data used to construct the burial history of the area are also shown.

Waltman Shale Member, and the Shotgun Member (Figure 2A). The Waltman Shale Member grades laterally into the Shotgun Member in the western and southern parts of the Wind River Basin, and, where the Waltman Shale Member is absent, the contact between the lower unnamed member and Shotgun Member is difficult to define and the formation is often defined as undifferentiated. The lower unnamed member of the Fort Union consists of white to gray, fine- to coarse-grained, massive to crossbedded sandstone interbedded with dark-gray to black shale, claystone, siltstone, brown carbonaceous shale, and thin to thick coal beds. The Waltman Shale Member is characterized by chocolate-brown and gray, silty and shaley claystone with a few thin sandstone beds (Keefer, 1965). The Waltman Shale Member contains appreciable amounts of organic matter, including thin, black, shiny coal laminae near the top and bottom. Organic content analyses and re-torting procedures in the early 1960s (Keefer, 1965; p. A28) and recent studies by Katz and Liro (1993), Palacas and others (1993; 1994, this volume), and unpublished organic geochemistry data by J.G. Palacas (U.S. Geological Survey, personal communication, 1994), indicate that the Waltman Shale Member is a good potential source rock for both oil and gas in the Wind River Basin. The Shotgun Member is a series of even-bedded soft claystone, siltstone, shale, and sandstone. Some thin, brown carbonaceous shales and thin coal beds crop out as dark bands on the light-colored slopes.

System	Series	Stratigraphic Unit or event	Thickness (in feet)	Ages (Ma)
Tertiary	mid. Mio.-present	uplift and erosion	-2,280	15 - 0
	Miocene	Split Rock Formation	900	24 - 15
		White River Formation	500	38 - 24
	Eocene	Wagon Bed Formation	480	52 - 38
		Wind River Formation	400	
		uplift and erosion	-1,800	55-52
	Paleocene	Fort Union Formation (part) undifferentiated	1,800	60 - 55

reconstructed thicknesses

Figure 2B. Tertiary stratigraphy of the Castle Gardens area, southern margin of the Wind River Basin, Wyoming. Thickness and age data used to construct the burial history of the area are also shown.

Methods

Two cross sections depict the eastern part of the basin (Table 1, Figure 3). The lines of section illustrate changes in facies, thickness, and thermal maturity of the Fort Union Formation from the structurally deep part of the basin southward to the outcrop areas. Wells containing vitrinite reflectance data were incorporated where possible.

For this study, vitrinite reflectance (R_o) is used to define levels of thermal maturity for selected wells in the central and eastern parts of the Wind River Basin. Vitrinite, a maceral derived from woody plant material, is common in coal and organic-rich shale. R_o is a measurement of light reflected from a polished vitrinite grain. It is related to the degree of metamorphism of the vitrinite grain and can be related to other thermal maturity indicators. Coal samples are generally prepared by crushing, mounting in epoxy on a microscope slide, planing off when hardened, and polishing. To obtain vitrinite from shales, the sample is generally macerated, and the kerogen concentrated before mounting and polishing. The mean random R_o (from randomly oriented indigenous vitrinite grains) is determined using plane-polarized incident white light and a 546 μm monochromatic filter, in immersion oil, on a reflected light microscope with a nonrotating stage (Bostick, 1979; Bustin, 1986).

Burial curves were constructed for two geologically distinctive areas using a variety of information. The thicknesses of the Fort Union Formation and Lower Eocene rocks were measured directly from geophysical logs. Ages and thicknesses of the Fort Union Formation were adapted from Nichols and Flores (1993), and Flores and others (1992). Thicknesses and ages of post-lower Eocene, Oligocene and

Miocene strata were estimated using data from Love (1988), Keefer (1970), and Van Houten (1964). Because

Table 1. Name and location of wells used to construct cross sections A-A' and B-B' (Figure 3).

Well no. (on cross sections)	Well name	Location (Section-Township-Range)
A-A'		
1	Shell Oil, 22-27 Unit	SE1/4, NW1/4, 27-35N-93W
2	Harvey E. Yates Co., Oil Spring Federal no. 1	NW1/4, NE1/4, 6-35N-93W
3	Husky Oil, 12-30 Federal	NW1/4, SW1/4, 30-36N-93W
4	InterNorth Inc., 12-19 Federal	NW1/4, SW1/4, 19-36N-93W
5	InterNorth Inc., 7-18 Federal 886	SE1/4, NE1/4, 18-36N-93W
6	J.M. Huber Corp., No. 6-1 Davison	SW1/4, 6-36N-93W
7	Humble Oil Co., 1 Poison Creek	NE1/4, NE1/4, 32-37N-93W
8	Larry Barnes Petroleum Inc., Carvner Federal no. 22-15	SE1/4, NW1/4, 15-37N-94W
9	Diamond Shamrock, 1 Shoshoni Unit	NW1/4, SE1/4, 21-38N-93W
10	William G. Helis Estate, No. 3 Unit	SW1/4, NE1/4, 35-39N-93W
11	William G. Helis Estate, No. 2 Bonneville Unit	SW1/4, NE1/4, 27-39N-93W
12	Shell Oil Co., 23-15 Howard Ranch	NE1/4, SW1/4, 15-39N-93W
B-B'		
13	Continental Oil Co., Love Ranch 1	NE1/4, SE1/4, 11-35N-91W
14	Continental Oil Co., No. 14-4	NW1/4, NW1/4, 14-36N-91W
15	Oil Development Co. of Texas, Brown Federal no. 1	20-37N-90W
16	Texas Pacific Oil Co., No. 18 Frenchie Draw	NE1/4, 8-37N-90W
17	Columbia Oil and Gas Co., No. 1-31	SE1/4, NW1/4, 31-38N-90W
18	Chaparral Oil Co., No. 1-29 Morreta	NW1/4, SE1/4, 29-38N-90W
19	Monsanto Oil Co., 1-16 Lost Cabin	SE1/4, NW1/4, 16-38N-90W
20	Monsanto Oil Co., Bighorn no. 1-5	SW1/4, NW1/4, 5-38N-90W
21	Inexco Oil Co., No. 1 Federal	NW1/4, 32-39N-90W
22	Monsanto Oil Co., No. 1-29	NW1/4, SE1/4, 29-39N-90W
23	Monsanto Oil Co., No. 1-21	SW1/4, SE1/4, 21-39N-90W
24	Monsanto Oil Co., No. 1-16	SW1/4, NE1/4, 16-39N-90W

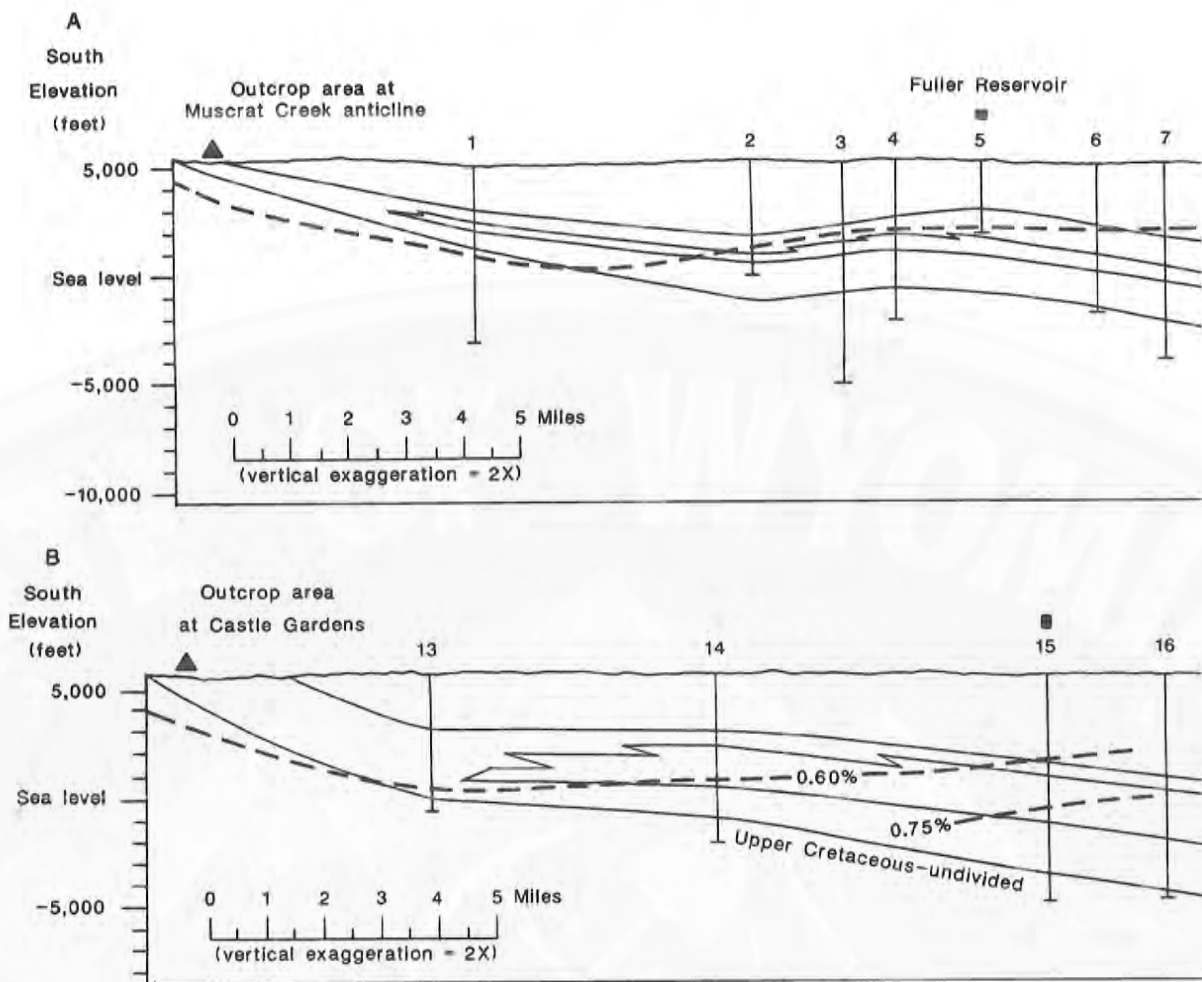


Figure 3. Cross sections A-A' and B-B' extending from the deep part of the Wind River Basin to the southern margin outcrop illustrating variations in facies and thickness of the Fort Union Formation, and overlying Eocene strata. Datum is mean sea level. Tfu = Fort Union Formation, undivided; Tfl = Lower unnamed member of the Fort Union Formation; Tfw = Waltman Shale Member of the Fort Union Formation; Tfs = Shotgun Member of the Fort Union Formation; Tle = lower Eocene rocks undifferentiated. Numbers for wells correspond to Table 1. Squares over numbered drill holes are wells with vitrinite reflectance data and correspond to Table 2. Triangles are locations of surface vitrinite reflectance samples. Generalized vitrinite reflectance lines are superimposed through the cross sections and show the 0.60 percent horizon (beginning of oil generation), the 0.75 percent horizon (beginning of thermogenic gas generation from humic organic matter), and the 1.10 percent horizon (significant gas generation from humic organic matter).

of the uncertainties associated with determining amounts of erosion, unconformities between formations were treated as hiatuses with no erosion. Figures 2A and 2B list the ages, thicknesses, and amounts of erosion used in construction of the burial histories for the two different areas.

Present-day heat flow values for the Wind River Basin range from 36 to 48 mWm⁻² (milliwatt/meter²) (Barker and Crysdale, 1993). In their models, Barker and Crysdale (1993) used heat flow values of 35 to 40 mWm⁻² for Late Cretaceous to Eocene, and 45 to 65 mWm⁻² for Eocene to Miocene to reach pre-

dicted temperatures in the Wind River Basin. Because present-day heat flow values are similar to those in the Tertiary, a constant heat flow model through time (66 Ma to present) is used for this report. Vitrinite reflectance data are used to calibrate the thermal history models of the Fort Union, and, assuming that our burial curves are fairly representative of the two different areas, heat flow is adjusted either upward or downward until the modeled R_o trends match the measured R_o data. The modeled vitrinite profile is based on Lawrence Livermore National Laboratory (LLNL) vitrinite kinetics used in the BasinMod program (Platte River

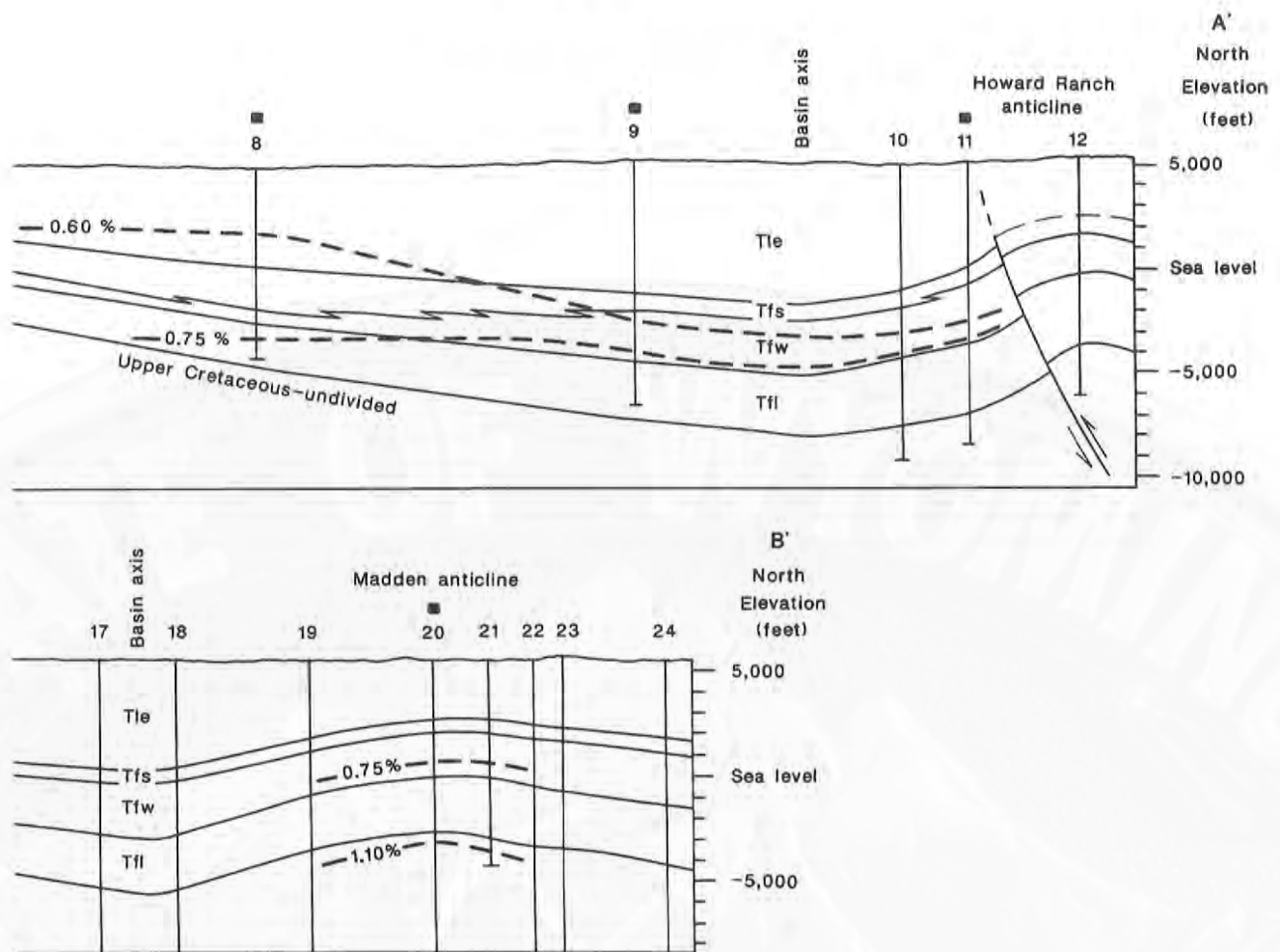


Figure 3A and B continued.

Associates¹). Heat flow values required in our models fit closely with those of Barker and Crysdale (1993).

After matching modeled R_0 values with the measured R_0 , a hydrocarbon-generation kinetic model was constructed to define the timing of petroleum generation for the Waltman Shale Member of the Fort Union Formation. The computer program BasinMod was used for the kinetic modeling. Kinetic modeling is useful in estimating the time when the Waltman Shale Member was capable of generating oil. It is the best method for predicting the petroleum generation history because it is based on the kinetic reactions of

organic matter during the burial and thermal maturation process. All reactions are treated as kinetically first order; that is, the rate is proportional to the amount of reactant. The Arrhenius equation is used to describe the temperature dependence of the rate constant. The model that results for a particular kerogen or mix of kerogens relates the amounts of oil and gas to time and temperature. For a complete explanation of kinetic modeling, see discussions in Tissot and Espitalie (1975), Ungerer (1983), Yukler and Kokesh (1984), Sweeney and others (1987), and Tissot and others (1987).

Kerogen types and related hydrocarbons

Three general types of kerogen have the potential, under optimum conditions, to generate hydrocarbons: type I, alginite (sapropelic or lipid-rich); type II,

exinite (phytoplankton, zooplankton, and other microorganisms); and type III, vitrinite and huminite (terrestrial plant debris). There is no absolute point

¹ Any use of trade, product, or firm names in this publication is for descriptive purposes only and does not imply endorsement by the U.S. Government.

at which hydrocarbon begins to be generated, and it probably begins over a range of maturity values (and temperatures) depending on the specific type of organic matter. In addition, several models have been developed relating the generation of hydrocarbons to types of kerogen and thermal maturity (Tissot and others, 1974; Dow, 1977; Waples, 1980, 1985). It should be noted that there is another kerogen, type IV, which comprises inert organic matter (oxidized and biologically altered organic matter, charcoal, and recycled organic matter); it has low hydrogen content and virtually no hydrocarbon-generation potential.

Type I kerogen is hydrogen-rich, occurs primarily in marine and lacustrine rocks, and generates mainly oil during catagenesis. The R_o value for the onset of oil generation from type I organic matter varies depending on the model one chooses. Dow (1977) used 0.50 percent R_o as the onset of oil generation for type I kerogen, whereas Anders and Gerrild (1984) and Tissot and Welte (1984) used 0.70 percent R_o . The end of oil generation also occurs over a range of R_o values, but 1.35 percent R_o is commonly accepted as the value at which oil begins to break down into shorter-chain hydrocarbons.

Type II kerogen occurs mainly in marine rocks, but can occur in lacustrine rocks, and generates oil and gas during catagenesis. Waples (1985) stated that oil generation begins over a range of R_o values of about 0.45 to 0.50 percent for high-sulfur kerogen to 0.60 percent for "typical" type II kerogen.

Huminite and vitrinite, or type III kerogen, is oxygen-rich and hydrogen-poor, occurs mainly in terrestrial, marginal-lacustrine, or marginal-marine rocks, and generates mostly gas (methane) during catagenesis. For type III kerogen, R_o is the best and most widely used measure of thermal maturity. Two important R_o thresholds—0.75 percent and 1.10 per-

cent—are used to define regions of gas generation from type III kerogen. An R_o of about 0.75 percent represents the maturity required for the onset of significant gas generation (Juntgen and Karweil, 1966; Juntgen and Klein, 1975). Gas accumulations found in rocks with an R_o less than 0.75 percent contain either early biogenic gas or gas migrated from more mature source rocks. An R_o of 1.10 percent represents the level of maximum gas expulsion from type III kerogen (Meissner, 1984). The upper limit of maturity for gas preservation is still unknown but could be as high as 3.5 percent R_o (Dow, 1977) or 4.0 percent R_o (Waples, 1980). Wet gas is generated from mixed lacustrine-marine-terrestrial organic matter and from the breakdown of oil between R_o values of 0.80 percent and 2.0 percent. Dry gas, or thermogenic methane, is generated from humic organic matter and from the breakdown of wet gas between R_o values of about 1.0 percent and 3.0 percent. Biogenic gas can be generated at levels of maturity as low as those for peat (0.20 percent R_o). Figure 4 illustrates, generally, the hydrocarbon generation thresholds represented by R_o values used in this report.

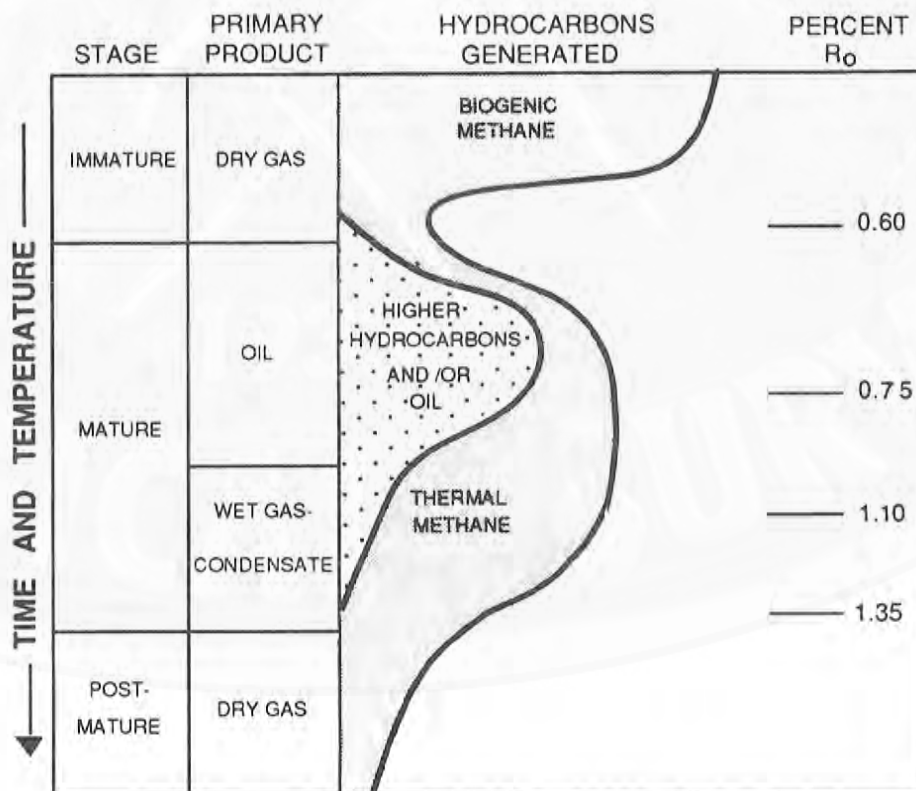


Figure 4. Generalized scheme for oil and gas generation as a function of thermal maturity (expressed as R_o). 0.60 percent R_o = onset of oil generation; 0.75 percent R_o = onset of gas generation from humic organic matter; 1.10 percent R_o = significant gas generation from humic organic matter; 1.35 percent R_o = end of oil generation.

Source rock potential of the Waltman Shale Member

Total organic carbon (TOC) content is a useful parameter for evaluating the quantity of organic material and the petroleum generation potential of a source rock. It is often accepted that shales having TOC values ≥ 0.50 percent are fair source rocks, and those with TOC values ≥ 1.0 percent are good source rocks. Katz and Liro (1993) analyzed 226 samples of the Waltman Shale Member from depositional facies representing the margin, as well as the deepest part of "Waltman Lake". TOC values for the Waltman Shale Member ranged from 0.93 to 6.21 weight percent, with a mean value of 2.71 weight percent. These high values demonstrate that there are very organic-rich zones in the Waltman Shale which have excellent petroleum-generation potential.

Another parameter used to evaluate the hydrocarbon-generating capacity of a source rock is the genetic potential or (S1 + S2). Tissot and Welte (1984) established the genetic potential classification which is as follows:

(S1+S2)

<2.0 mg/g	poor or non source rock
2.0-6.0 mg/g	moderate source rock
>6.0 mg/g	good to excellent source rock

Katz and Liro (1993) state that the genetic potential of 218 samples ranged from 0.30 to 24.04 mg/g. Of these, all but 14 had a genetic potential >2.5 mg/g. One hundred thirty-two samples had values >6.0 mg/g. It is possible that the genetic potential of some of the samples is artificially high due to the presence of unexpelled hydrocarbons (see Figure 7 in Katz and Liro, 1993). However, even after extraction, the genetic potential of the majority of samples is in the moderate-good to excellent source rock range.

Organic matter type determines the quality and kind of hydrocarbons that will be generated. The most widely used parameter for defining the type of organic matter contained in a source rock is

the modified van Krevelen diagram (van Krevelen, 1961; Tissot and Welte, 1984). These diagrams are a plot of the hydrogen index (HI) and oxygen index (OI) as determined from Rock-Eval analysis, and illustrate the type of organic matter and its relative position on the maturation evolutionary path. Figure 5 is a modified van Krevelen diagram (Figure 10 from Katz and Liro, 1993) for Waltman Shale Member samples. The diagram shows mainly types II and III organic matter, sources for both oil and gas. Katz and Liro (1993) believe that the organic input was both algal and higher plant material, or a mixture of algal and recycled (or reworked) organic material. It should be noted that the recycled organic material in the samples is depleted in hydrogen and plots closer to the type III kerogen reference line. This recycled organic material contributes to the carbon content of the rock but not to the hydrocarbon generation potential. In the case of the Waltman Shale Member, the relative proportion of type III, gas-generative kerogen may be exaggerated by the relatively low hydrogen index of the recycled material (Katz and Liro, 1993).

The wide distribution of samples in Figure 5 is probably not only the result of differences in organic matter type and source, but in the degree of thermal maturation as well. In general, plotted samples migrate toward the origin of the diagram (HI of 0.0, and OI of 0.0) with increasing thermal maturity. Caution should be taken when interpreting the origin and

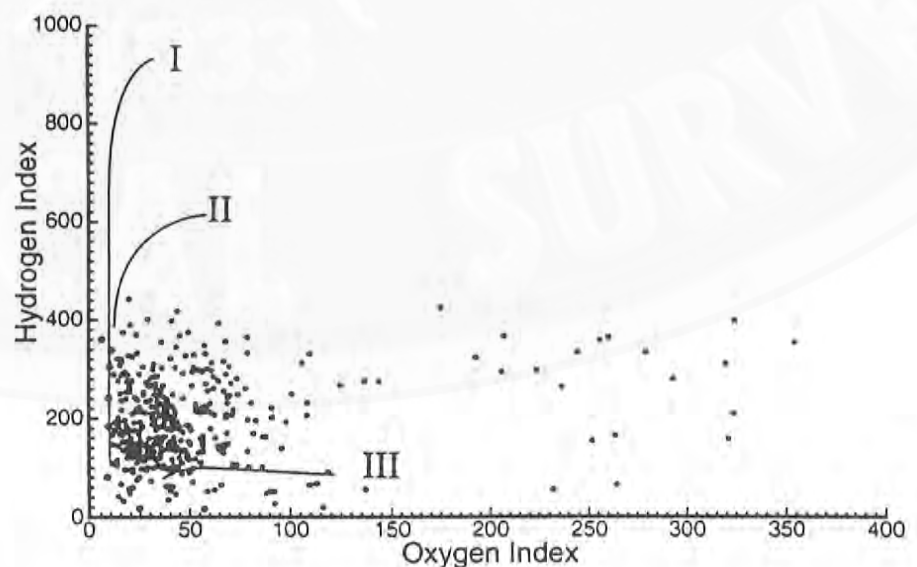


Figure 5. Modified van Krevelen diagram showing type of organic matter in Waltman Shale Member samples. From Katz and Liro (1993).

source of organic matter in samples that have a high level of thermal maturation. For example, a rock that originally had an HI of 400 and an OI of 75 (type II kerogen of marine origin), will decrease in HI and OI with increasing thermal maturation (migrate toward

the origin of the diagram), and will ultimately plot as a type III kerogen. The wide range illustrated in Figure 5 is possibly due, in part, to the differences in the thermal history (some are more mature than others) of the samples throughout the basin.

Thermal maturity of the Fort Union Formation, central and eastern Wind River Basin

The thermal maturity of the Fort Union Formation in the central and eastern part of the Wind River Basin is fairly well established based on studies of Katz and Liro (1993), Nuccio and others (1993), and Pawlewicz (1993). Katz and Liro (1993) provided vitrinite reflectance profiles for the Waltman Shale Member, and Pawlewicz (1993) presented R_o profiles for the Fort Union Formation and other units, for drill holes throughout the basin. Nuccio and others (1993) provided vitrinite reflectance data and described the thermal maturity of the Fort Union at several outcrop localities around the periphery of the basin.

The level of thermal maturity of the Fort Union Formation ranges from very immature (0.29 percent R_o) in outcrops in the southern part of the basin, to mature (0.99 percent R_o) for the lower unnamed member in the deep part of the basin (see Figure 3 and Table 2). Surface R_o values range from 0.29 to 0.55 percent but generally are in the 0.40-0.50 percent range. Subsurface R_o values increase toward the deeper part of the basin, and each individual well profile shows an increasing-with-depth trend (see Figure 3 and Table 2). The extreme differences in

thermal maturity between the shallow and deep parts of the basin are probably due largely to differences in burial history, as there is no evidence for significant heat flow variations in the eastern part of the basin. This will be discussed further in the next section.

Based on the vitrinite reflectance data, the Fort Union Formation in the deep part of the basin has reached the proper level of thermal maturity (>0.60 percent R_o) for both oil and thermogenic gas generation (see Figure 4). The Waltman Shale Member in the deep part of the basin has achieved R_o values of around 0.75-0.80 percent (we have not found any areas in the basin where the Waltman is greater than around 0.80 percent). Approaching the outcrop, levels of thermal maturity decrease, and, where R_o values are <0.60 percent, the Fort Union is considered to be immature for significant oil or thermogenic gas generation (see Figure 3) for purposes of this study. These areas of low thermal maturity do not preclude the possibility of biogenic gas generation from some of the coals and carbonaceous shales in the Fort Union Formation.

Burial, thermal, and petroleum generation models

Deep basin

The burial, thermal and petroleum generation history of the Fort Union Formation in the deep part of the Wind River Basin is illustrated using data from the Larry Barnes Petroleum, Carvner Federal 22-15 well located in sec. 15, T. 37N., R. 94W. (Figure 6). During the Paleocene and Eocene, this part of the basin was continuously subsiding, and 4,365 ft (1,330 m) of Fort Union Formation strata, and 6,605 ft (2,013 m) of strata comprising the Indian Meadows, Wind River, and Wagon Bed Formations were deposited. Following

this rapid deposition of a thick sequence of sediments, and with a period of relative quiescence just prior to late Tertiary uplift, a relatively modest 1,400 ft (427 m) of Oligocene and Miocene strata were deposited (see Figure 2A and Figure 6).

A vitrinite reflectance profile was constructed for the previously mentioned well and is used to calibrate the thermal history (Figure 7). R_o values range from 0.61 percent in the lower part of the Wind River Formation to 0.90 percent in the lower unnamed member of the Fort Union Formation (Figure 7). R_o for the top

Table 2. Surface (outcrop) and subsurface vitrinite reflectance (R_o) data from the Fort Union Formation and other rock units in the Wind River Basin. Outcrop data are indicated by triangles and subsurface data by squares on Figure 1.

on Figure 1.

Outcrop data				
Location (Section-Township-Range)		R _o (percent)		Fort Union Formation
NE1/4, NE1/4, 3-5N-1E		0.43		undifferentiated
SE1/4, 36-36N-86W		0.42		undifferentiated
SE1/4, 35-35N-91W		0.40		undifferentiated
SW1/4, 4-34N-90W		0.36		undifferentiated
NW1/4, 6-33N-92W		0.39		undifferentiated
E1/2, 19-34N-92W		0.54		undifferentiated
SW1/4, 17-34N-92W		0.40		Shotgun Member
S1/2, 22-34N-92W		0.29		undifferentiated
SW1/4, 22-33N-95W		0.55		undifferentiated
Outcrop R _o data from Nuccio and others (1993)				
Subsurface data				
Well number (on cross sections)	Well name	R _o (percent)	depth (ft)	Fort Union Formation and other rock units
5	[3] InterNorth Inc., 7-18 Federal 886	0.46	2,358	Shotgun Member
		0.50	2,458	Shotgun Member
8	[1] Larry Barnes Pet. Inc., Carvner Fed. no. 22-15	0.61	3,270	lower Eocene rocks
		0.67	4,525	lower Eocene rocks
		0.67	4,665	Shotgun Member
		0.68	4,795	Shotgun Member
		0.64	4,865	Shotgun Member
		0.66	5,025	Shotgun Member
		0.68	5,330	Shotgun Member
		0.62	5,535	Shotgun Member
		0.63	5,705	Shotgun Member
		0.71	6,065	Shotgun Member
		0.75	7,745	lower member
		0.90	7,905	lower member
		0.90	8,125	lower member
		0.80	8,225	lower member
9	[2] Diamond Shamrock, 1 Shoshoni Unit	0.89	8,585	lower member
		0.89	8,725	lower member
11	[2] William G. Helis Estate, No. 2 Bonneville Unit	0.90	9,055	lower member
		0.54	7,100	Shotgun Member
		0.73	9,100	Waltman Shale Member
		0.52	5,800	Shotgun Member
		0.49	6,400	Waltman Shale Member
15	[2] Oil Develop. Co. of Texas, Brown Federal no. 1	0.52	7,200	Waltman Shale Member
		0.65	7,800	Waltman Shale Member
		0.73	8,200	Waltman Shale Member
		0.62	4,500	Shotgun Member
		0.62	5,000	Waltman Shale Member
20	[3] Monsanto Oil Co., Bighorn no. 1-5	0.70	5,400	Waltman Shale Member
		0.75	6,100	Waltman Shale Member
		0.71	6,700	Waltman Shale Member
		0.97	7,875	lower member
		0.99	8,115	lower member
		1.05	8,305	Lance Formation
		1.07	8,603	Lance Formation
		1.28	8,985	Lance Formation
[1] R _o data from this report [2] R _o data from Katz and Liro (1993) [3] R _o data from Pawlewicz (1993)				

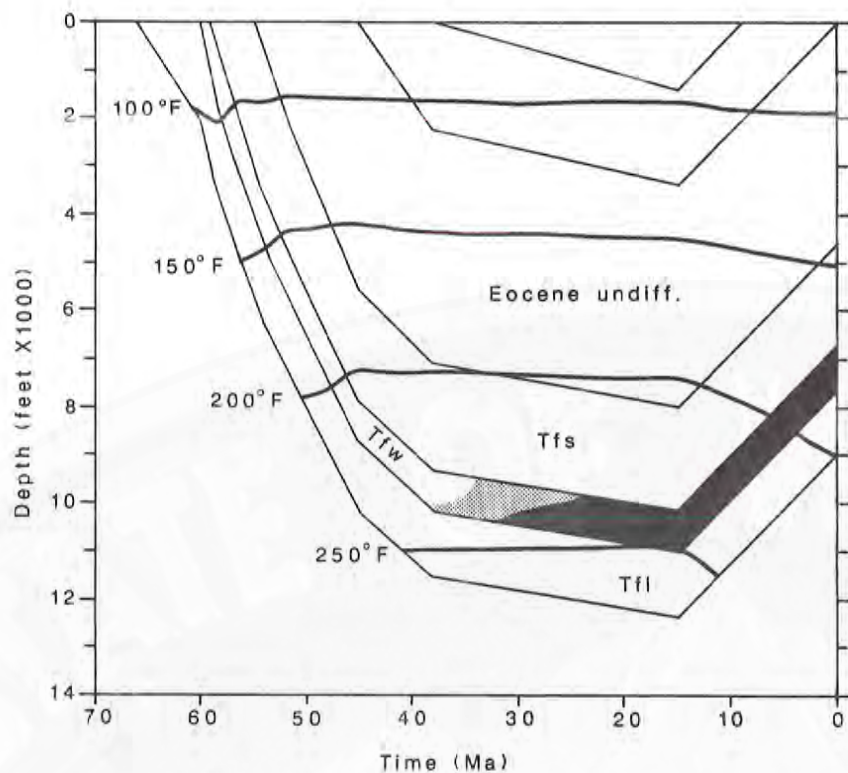
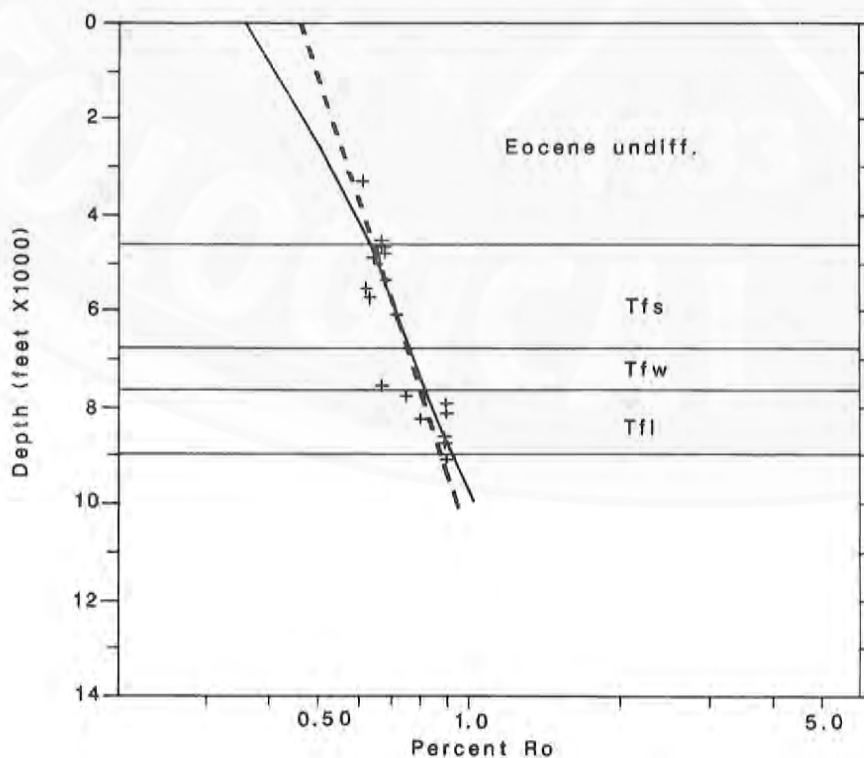


Figure 6. Burial and thermal history of the deep part of the eastern Wind River Basin, Wyoming. Isotherms illustrate temperature through time. Petroleum generation from the Waltman Shale Member is shown. Light pattern represents early oil generation; black indicates significant oil generation. Data used to construct the burial curve is presented in Figure 2A. Unit symbols are as defined for Figure 3.



of the Waltman Shale Member is approximately 0.75 percent, and at the base the R_o is around 0.80 percent. A constant heat flow of 45 mWm^{-2} was required to match the modeled R_o (solid line on Figure 7) to the measured R_o (dashed line on Figure 7). This heat flow is consistent with that determined by Barker and Crysdale (1993).

The base of the Fort Union Formation in this part of the basin was buried to a depth of 12,400 ft (3,780 m) and temperatures greater than 250°F (121°C) were reached at maximum burial 15 Ma (Figure 6). At this time (1) the base of the Waltman Shale Member (the main petroleum source rock in the area) reached a burial depth of 11,000 ft (3,553 m) and a temperature of 250°F (121°C); and (2) the top of the Waltman Shale was buried to 10,150 ft (3,094 m) and reached temperatures of approximately 240°F (116°C). Petroleum generation within the Waltman Shale is based on the burial and thermal model mentioned above, a kerogen mix of 60 percent type II, 10 percent type III, and 30 percent type IV, and an average TOC of 2.0 percent. Employing these parameters, early oil generation began at 38 Ma at the

Figure 7. Measured and modeled vitrinite reflectance profiles for the Larry Barnes Petroleum, Carver Federal 22-15 well in the deep part of the central Wind River Basin. Individual R_o data points are represented by (+), and correspond to Table 2. The dashed line is the measured profile, and the solid line is the modeled profile. A heat flow of 45 mWm^{-2} was required to match the modeled and measured lines. Unit symbols are as defined for Figure 3.

base of the Waltman Shale and 34 Ma for the top (Figure 6). Significant oil generation, amounts necessary for accumulation and possible migration, began around 32 Ma at the base of the Waltman Shale and 25 Ma at the top. The rate of oil generation likely slowed after 15 Ma with regional uplift, erosion, and subsequent cooling; however, in the structurally deeper parts of the basin, generation may have continued after 15 Ma.

Castle Gardens area (southern margin of basin)

The burial and thermal history of the Fort Union Formation on the southern flank of the basin is illustrated in Figure 8. As is immediately apparent, this area is quite different than the deep part of the basin discussed above. The Fort Union Formation is not divided into members in the southern part of the basin, and the Waltman Shale Member was never deposited there (Figure 2B). Surface coal samples having R_o values of 0.36 and 0.40 percent, collected in the lower part of the Fort Union Formation, were used to calibrate the burial and thermal history of this area. The coal beds were buried by approximately 1,800 ft (549 m) of younger Fort Union strata in this marginal area. In early Eocene time (between approximately 55 to 52 Ma), about 1,800 ft (549 m) of rocks were stripped by uplift and erosion, leaving the coal beds lying again near the surface (Figure 8). Following this structural event, 2,280 ft (695 m) of Eocene, Oligocene, and Miocene sediments were deposited over the coal-bearing strata (Figures 2B and 8); these strata were then removed during another period of uplift and erosion beginning

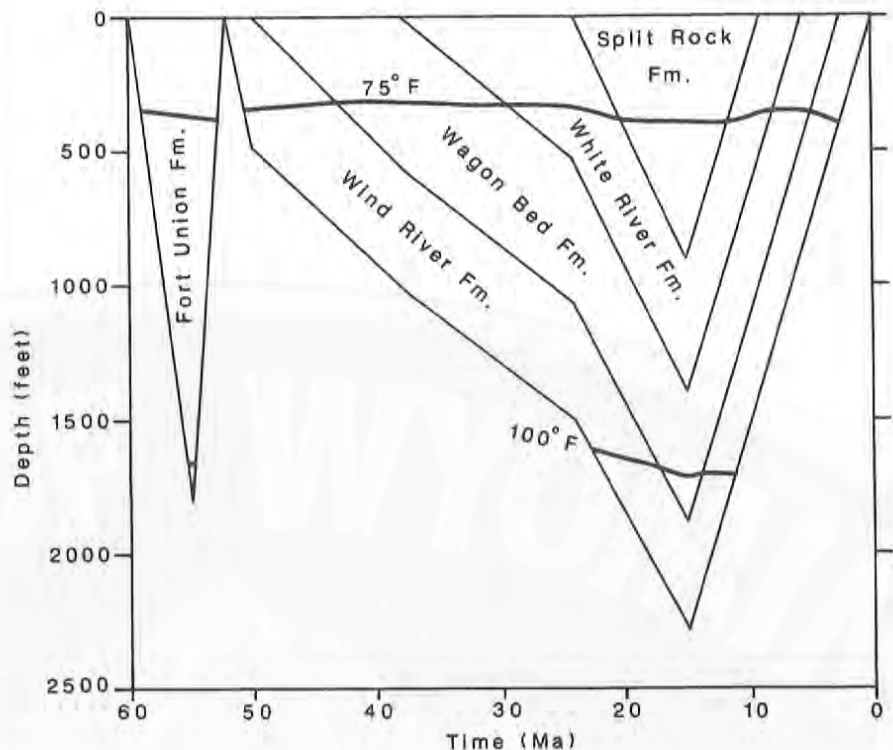


Figure 8. Burial and thermal history of the Castle Gardens area at the southern margin of the Wind River Basin, Wyoming. Isotherms illustrate temperature through time. This area is immature for oil generation, however, coal beds may have the potential for biogenic gas generation. Data used to construct the burial curve is presented in Figure 2B.

around 15 Ma, leaving the coals in their present outcropping position.

A constant heat flow of 45 mWm^{-2} was also used in the Castle Gardens area and temperatures through time are shown in Figure 8. Owing to the complex history of uplift and erosion in the Castle Gardens area, and in other areas around the margin of the basin, the lower part of the Fort Union Formation never reached high temperatures or high levels of thermal maturity. In the Castle Gardens area, the lower Fort Union has not reached the thermal requirements for petroleum generation, nor are significant source rocks present. It is possible, however, that the thin, low maturity coal beds present in the Fort Union Formation in this area, and in adjacent down-dip areas, could be a source for biogenic gas.

Discussion

Data presented in this report indicate that the Fort Union Formation, and more specifically the Waltman Shale Member is mature enough to have generated

significant amounts of petroleum in the deeper, central and eastern parts of the Wind River Basin. Figure 3 illustrates the approximate areas where the

Waltman Shale grades from mature (>0.60 percent R_o) to immature (<0.60 percent R_o). For cross section A-A', the Waltman Shale is mature through most of the deeper part of the basin until near the area of well no. 2, approaching Signor Ridge on the north flank of the Muskrat Creek anticline, where it is <0.60 percent R_o . For cross section B-B', the Waltman again reaches maturity in the deeper part of the basin, but becomes immature in an area between wells no. 14 and 15, and remains immature southward to the Castle Gardens area. As mentioned earlier, one would not expect significant generation of hydrocarbons in these immature areas.

We suggest that in the deeper parts of the basin, oil generation near the base of the Waltman Shale began around 32 Ma and at the top of the unit around 25 Ma. Southward, toward areas of less burial, generation from the Waltman would have begun later. At around 15 Ma, generation of oil from the Waltman Shale throughout most of the basin would have ceased.

We have found that R_o values for the Waltman Shale Member, even in the deeper areas of the basin, are not greater than around 0.80 percent. Palacas and others (this volume) demonstrate that oils produced at the Haybarn field in the moderately-deep part of the

basin have a nearly identical geochemical signature to oil analyzed from oil-saturated sandstones at Signor Ridge and Castle Gardens. Palacas and others (this volume; J.G. Palacas, personal communication, 1994) further believe that these oils were generated from the Waltman Shale Member at thermal maturities of around 0.75-0.85 percent R_o . Referring to Figures 1 and 3, for cross section A-A' the Waltman achieved maturities of 0.75 percent R_o or greater only around wells 8, 9, 10 and 11. For cross section B-B', the Waltman Shale is 0.75 percent R_o or greater near wells 15 through 21 (the hachured area on Figure 1 illustrates the area where the Waltman Shale Member is 0.75 R_o or greater). The thermal maturity of the Waltman Shale at Haybarn field is similar to that at Fuller Reservoir immediately to the west (no. 5 on cross section A-A'), around 0.60 to 0.70 percent R_o . The thermal maturity of rocks at the outcrop at Castle Gardens is around 0.36 percent R_o . The fact that oil generated at maturities of 0.75-0.85 percent R_o is present in both Haybarn field and at Castle Gardens indicates migration distances of at least 12 mi (19 km) and 18 mi (29 km), respectively, from the deeper basin areas (see Figure 1 of Palacas and others, this volume). The time of migration is constrained to between 32 and 15 Ma. Moreover, the relationships indicate that an R_o of 0.75 to 0.85 is necessary before oil migration can occur.

Summary

The thermal maturity of the Fort Union Formation and petroleum potential of the Waltman Shale Member has been characterized in the central and eastern parts of the Wind River Basin, Wyoming. The following statements summarize our investigation:

1. The Waltman Shale Member is a good to excellent source rock with an average TOC value of 2.71 weight percent, and genetic potential values as much as 24.04 mg/g for those samples analyzed.

2. A modified van Krevelen diagram of Waltman Shale samples indicates mainly types II and III organic matter, making it a source for both oil and gas; however, the presence of type IV kerogen may be causing an overestimation the amount of type III organic matter, and the gas potential.

3. The level of thermal maturity of the Fort Union Formation ranges from very immature (0.29 percent R_o) in outcrops in the southern part of the basin, to

mature (0.99 percent R_o) for the lower unnamed member in the deep part of the basin.

4. Surface R_o values range from 0.29 to 0.55 percent but generally are in the 0.40-0.50 percent range. Subsurface R_o values increase toward the deeper part of the basin, and each individual well profile shows an increasing-with-depth trend. The extreme differences in thermal maturity between the shallow and deep parts of the basin are largely due to differences in burial history.

5. The Fort Union Formation in the deep part of the basin has reached the proper level of thermal maturity (>0.60 percent R_o) for both oil and thermogenic gas generation. The Waltman Shale Member in the deep part of the basin has achieved R_o values of around 0.75-0.80 percent.

6. In the structurally shallower areas of the basin, levels of thermal maturity decrease, and, where R_o

values are <0.60 percent, the Fort Union Formation is considered, for purposes of this study, to be immature for significant oil or thermogenic gas generation. However, coals in the Fort Union in these areas of low thermal maturity may have the potential for biogenic gas generation.

7. In the deep part of the basin, significant oil generation from the Waltman Shale Member, in amounts necessary for accumulation and possible migration, began around 32 Ma. The rate of oil generation probably slowed or ceased after 15 Ma with regional uplift, erosion, and subsequent cooling.

8. Owing to the complex history of uplift and erosion in the Castle Gardens area, and in other areas

around the periphery of the basin, the Fort Union Formation has not reached the thermal requirements for petroleum generation, nor are significant source rocks present.

9. Oils produced at the Haybarn field have a nearly identical geochemical signature to oils analyzed from exposed oil-saturated sandstones at Signor Ridge and Castle Gardens. These oils were generated from the Waltman Shale Member in the deep part of the basin at thermal maturities of around 0.75-0.85 percent R_o . Migration of as much as 18 mi (29 km) is necessary to get the mature oils updip into these moderately-mature to immature reservoirs.

Acknowledgments

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A preliminary study of coal-forming environments during deposition of the Paleocene Fort Union Formation, Wind River Basin, Wyoming

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Introduction

The coal deposits in the Paleocene Fort Union Formation of the Wind River Basin in central Wyoming (**Figure 1**) have received considerably less attention than the Fort Union coals in other Rocky Mountain Basins primarily because they occur mainly in the subsurface and do not crop out extensively. A recent study by Johnson and Flores (1993) using geophysical logs, however, revealed that the Fort Union Formation contains more than 100 ft of total coal in some parts of the Wind River Basin. Despite these large thicknesses, little is known about the environments of deposition of these coals. Drillholes are relatively sparse in the Wind River Basin, and a detailed analysis of the coal-forming environments is difficult. This paper presents a general analysis of Paleocene coal-forming environments in the Wind River Basin using geophysical logs from most of the available drillholes. Sandstone geometries were studied using resistivity and gamma-ray logs, and sonic, density, and gamma-ray logs were used to determine coal thicknesses.

The Fort Union Formation in central Wyoming was deposited during a time of transition, when the Rocky Mountain foreland basin was gradually being partitioned into smaller Laramide structural and sedimentary basins by rising uplifts. In the Wind River Basin the formation is divided into three members, in ascending order (**Figure 2**): the lower unnamed member, the Waltman Shale Member, and the Shotgun

Member (Keefer, 1961a; 1965; 1969). The lower member is of mainly fluvial origin and is recognized on the surface and in the subsurface throughout most of the basin. The Waltman Shale Member is composed mainly of lacustrine shale and mudstone deposited in offshore areas of Lake Waltman, a large lake that, at its maximum extent, covered most of the central part of the Wind River Basin. The Waltman Shale Member occurs largely in the subsurface and crops out only in the Waltman area in the northeast part of the basin which was designated by Keefer (1961a) as the type section. The Shotgun Member is a complex unit that is thought to be largely time equivalent to the Waltman Shale Member. It consists of fine-to coarse-grained, in part conglomeratic, fluvial and marginal lacustrine rocks deposited around the margins of Lake Waltman. The Shotgun Member was named for Shotgun Butte in the western part of the basin, where it is best exposed.

The contact between the Fort Union Formation and the underlying Lance Formation in the Wind River Basin is generally marked by a change from shale, claystone, and minor sandstone in the uppermost part of the Lance to predominantly sandstone and conglomerate in the lowermost part of the Fort Union. In outcrops along much of the basin margins, Keefer (1965) considered the Fort Union to be entirely Paleocene in age and to rest unconformably (angular and/or erosional) on the Upper Cretaceous Lance For-

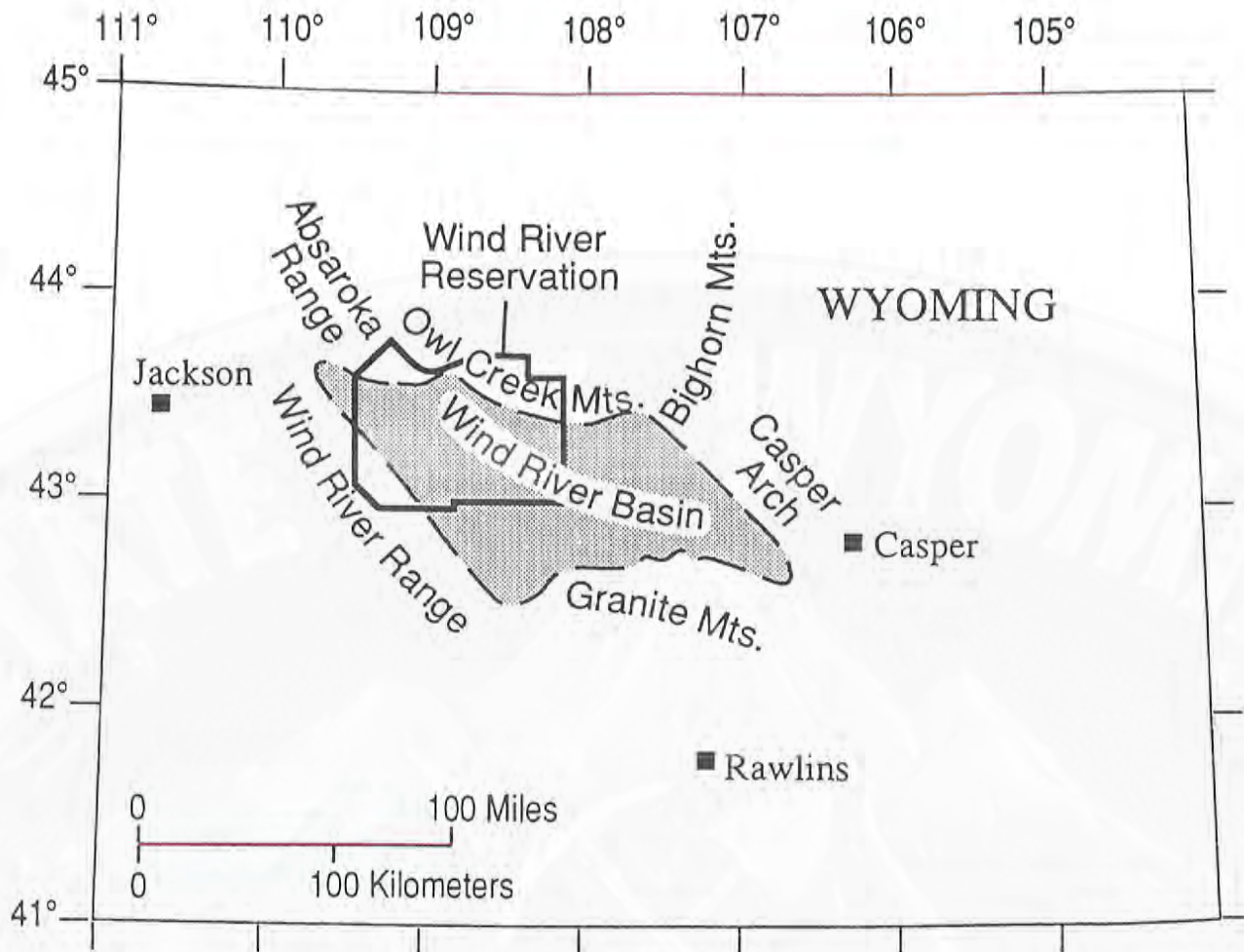


Figure 1. Map showing location of Wind River Basin and surrounding uplifts. Location of Wind River Reservation is also shown.

mation. He believed, however, that the two units were conformable a short distance down-dip from outcrop. Although the contact was picked on the basis of this lithologic change, Keefer (1965, p. A21) cited paleontological evidence which indicated that the contact coincides closely with the Cretaceous-Tertiary boundary; an exception was noted at the Waltman locality, where the uppermost part of the Lance appeared to be

of earliest Paleocene age. Keefer attributed this shift from fine- to coarser-grained sediments to the onset of uplift on the Wind River uplift southwest of the basin. In the southwest part of the basin, adjacent to the Wind River Range, the Fort Union Formation overlies progressively older rocks toward the southwest (Keefer, 1970; Keefer and Johnson, 1983).

Recent outcrop studies of the Fort Union Formation in the Wind River Basin

The Fort Union Formation crops out extensively around the margins of the basin, but these outcrops include relatively minor amounts of coal; the coaly facies occurs almost totally in the subsurface. Detailed outcrop studies, however, were recently published for several areas in the basin (Figure 3), and these studies

are reviewed here to try to better understand the environments of deposition of the coaly facies. The Fort Union has recently been studied in outcrop in the Castle Gardens area in the south-central part of the basin (Flores and others, 1991), near Hudson in the southwestern part of the basin (Flores and others,

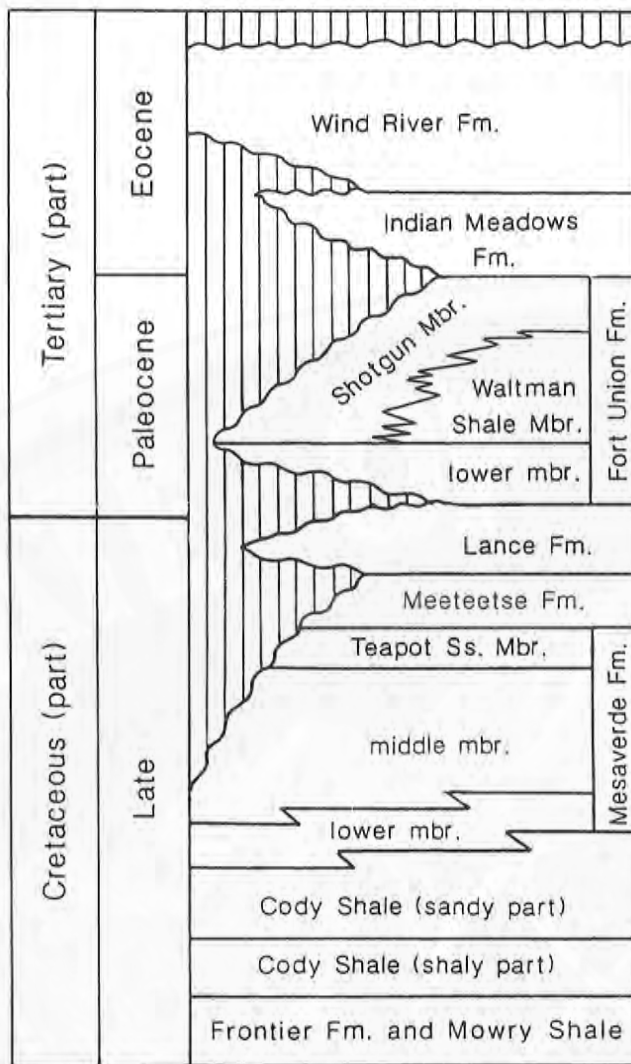


Figure 2. Generalized stratigraphic chart for Upper Cretaceous and lower Tertiary rocks in the Wind River Basin. From Johnson and others (1993).

1992), and in the Shotgun Butte area in the northwestern part (Flores and Keighin, 1993). Detailed palynomorph studies have also recently been completed for these areas (Flores and others, 1992; Nichols and Flores, 1993). These studies applied the six-fold subdivision of the Paleocene, P1 through P6, using a zonation based on palynomorphs that was developed by Nichols and Ott (1978). The P1 and P2 biozones are equivalent to the early Paleocene, the P3 and P4 biozones are equivalent to the middle Paleocene, and the P5 and P6 biozones are equivalent to the late Paleocene. Palynomorphs were also studied in one drillhole in the western part of the basin (Nichols and Flores, 1993) where the coaly facies is present, hence the age of the coaly facies is precisely known at this locality.

The stratigraphic units defined in other areas of the basin, on the basis of lithology, are difficult to trace into Castle Gardens primarily because the entire Upper Cretaceous and lower Tertiary section is very sandy here. Flores and others (1992) used palynomorphs to determine that the Paleocene section is about 2,300 ft (701 m) thick. They placed the base of the Fort Union Formation a few hundred feet lower stratigraphically than previous workers (Keefer, 1970; Gillespie and Fox, 1991). All six Paleocene biozones are present at Castle Gardens. About 1,450 ft (442 m) of this interval was deposited during P5-P6 time when Lake Waltman was present farther north in the basin, and Flores and others (1992) suggest a relationship between this rapid rate of subsidence and the existence of the lake. The Fort Union sandstones were deposited by sinuous bedload streams that varied from fairly deep and wide to shallow and narrow. The lithology of pebbles in the sandstones changes upwards in the section from mudstone in the P1 biozone, to chert and some granite in the upper part of the P4 and lower part of the P5 biozone and finally to abundant granite in the upper part of the P5 and the P6 biozones. These changes seem to record the progressive unroofing of the Granite Mountains to the south. Coal beds occur in the P4 and P5 biozones and are most commonly found above the larger fluvial channel and crevasse-splay complexes. The thickest coal is as much as 12 ft (3.7 m) thick and occurs in the P5 biozone.

Near Hudson, along the southwest flank of the Wind River Basin, the Fort Union Formation is only a few hundred feet thick, is largely conglomeratic, and unconformably overlies the lower part of the Mesaverde Formation. The member subdivisions cannot be recognized in this locality but palynomorphs indicating a P3 biozone age were recovered from the middle part of the interval. The outcrops occur on the northeast flank of a series of northwest-trending, sharply folded and faulted anticlines that parallel the nearby flank of the Wind River Range (Figure 1). Flores and others (1993) believed that the anticlinal folding, as well as uplift on the Wind River Range, resulted in removal of much of the Upper Cretaceous section in this area creating a broad 15-mile (24 km) wide northeast-trending paleovalley. The Upper Cretaceous Meeteetse and Lance Formations and much of the Mesaverde Formation were eroded along this trend before the coarse-grained fluvial rocks of the Fort Union Formation were deposited.

Flores and others (1993) divided the Fort Union Formation near Hudson into three parts. The lower

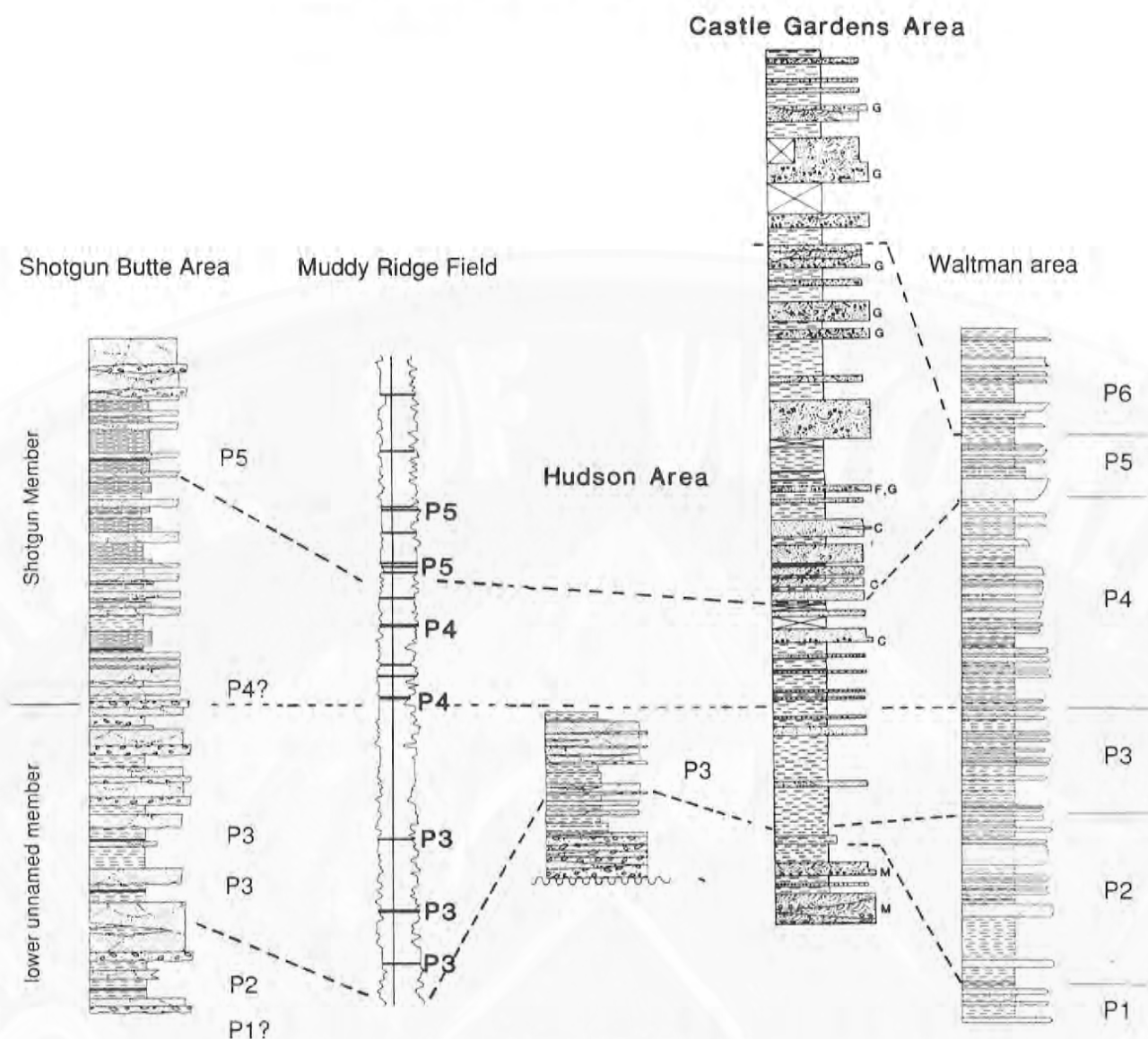


Figure 3. Generalized correlation chart for the Fort Union Formation at Shotgun Butte, Hudson, Castle Gardens, and Waltman showing the P1- through P6 biozones. Modified from Flemming and Nelson (1991), Flores and others (1992), Nichols and Flores (1993).

part is about 115 ft (35 m) thick and consists mainly of conglomerate and conglomeratic sandstone. Individual conglomerate beds are as much as 110 ft (34 m) thick and 1.7 mi (2.7 km) in lateral extent. Conglomeratic sandstones are as much as 45 ft thick and 1.4 mi in lateral extent. The middle part is 170 ft (52 m) thick and consists mostly of siltstone and mudstone with subordinate conglomeratic sandstone and rare carbonaceous shale and coal beds. The upper part is 115 ft (35 m) thick and is composed of conglomeratic sandstone and subordinate siltstone and mudstone and rare carbonaceous shale and coal beds. Conglomerates form lenses in the sandstones that are as much as 10 ft (0.3 m) thick and 0.4 mi (0.6 km) in lateral extent.

The upper and lower parts were deposited by low-sinuosity braided streams while the middle part reflects deposition in poorly drained to well drained floodbasins and low-lying mires.

The Fort Union was studied in exposures in the Shotgun Butte area, and in the subsurface to the east in the Pavillion-Muddy Ridge gas field area by Flores and Keighin (1993). The study area is in a series of southeast plunging synclines that was an area of rapid subsidence during Upper Cretaceous and Paleocene time (Keefer and Troyer, 1964). This rapid subsidence resulted in a conformable contact between the Lance Formation (Upper Cretaceous and lower Paleocene)

and the Fort Union Formation (Paleocene) along the length of the depositional valley. Flores and Keighin (1993) believed that these rapid subsidence rates helped confine the river systems to the area of the synclines. Nichols and Flores (1993) were able to identify biozones P2, P3 and P5 here but poor recovery of spores and pollen prevented the identification of zones P1, P4, and P6. The upper part of the Lance Formation, as mapped by Keefer and Troyer (1964), was in the P2 biozone. Flores and Keighin (1993) and Nichols and Flores (1993) included this Paleocene interval of the Lance in their lower member of the Fort Union Formation. The lower unnamed member, as used by Flores and Keighin (1993) and Nichols and Flores (1993) is within biozones P-2 and P-3 while P5 pollen was collected from the Shotgun Member.

The P2 biozone interval, in the uppermost part of the Lance Formation of Keefer and Troyer (1964), consists mainly of sandstones, siltstones, and mudstones and rarely carbonaceous shales and coal beds. This interval was interpreted by Flores and Keighin (1993) as deposited by a mixed suspended load and bedload braided, meandering, and anastomosed stream sys-

tems. The base of the Fort Union as originally defined by Keefer and Troyer (1964) occurs at the base of an interval of conglomerate and conglomeratic sandstones as much as 100 ft (30 m) thick and subordnately siltstone and mudstone. This interval is interpreted by Flores and Keighin (1993) as deposited in a bedload braided stream system. The Shotgun Member, in contrast, is mud dominated with carbonaceous shales and thin coals. The comparatively thin (less than 39 ft or 12 m thick) sandstones in the Shotgun Member are interpreted to have been deposited by low-to high-sinuosity, meandering and anastomosed streams confined by floodplain sediments that were partially covered by poorly drained, low-lying mires.

Nichols and Flores (1993) were also able to identify the P3 through P5 biozones from cuttings from a well in the Muddy Ridge field about 10 mi (16 km) southeast of outcrop. The contact between the lower unnamed member and the Shotgun Member in the drillhole occurs at about the base of the P4 zone. Coals occur in the upper part of the lower unnamed member and the lower part of the Shotgun Member in the P3 through P5 biozones.

Occurrence of Fort Union Formation coals in the subsurface of the Wind River Basin

A recent study by Johnson and Flores (1993) used subsurface information to define areas of thick Fort Union coal accumulations in the Wind River Reservation, which covers the western part of the structural Wind River Basin. The purpose of the study was to define areas where Fort Union coals could potentially be developed for coalbed gas, but the environments of deposition of the coals were not studied in detail. They found two areas based on drillhole data which contained significant thicknesses of Fort Union coal, one south of Little Dome anticline, where as much as 100 ft (30 m) cumulative thickness of Fort Union coal in beds 2 ft thick or greater is present, and one east of the Pavillion-Muddy Ridge gas field, just south of the trough of the basin, where over 60 ft (18 m) of coal in beds 2 ft (0.6 m) thick or greater occurs. The accumulation south of Little Dome occurs in a syncline and the authors suggested that rapid subsidence rates there may have created an environment favorable to peat accumulation. The other accumulation, just south of the basin trough, occurs near the fluctuating shoreline of Lake Waltman in an area roughly between two mapped river systems, the one at Hudson (Flores and others, 1993) and the one near Shotgun Butte (Flores and Keighin, 1993).

Results of present study

This study extends the subsurface study of Fort Union coals into the central and eastern parts of the Wind River Basin, east of the Wind River Reservation. Some additional work was also done on the Reservation to better define environments of deposition and coal correlation. **Figure 4** is a north-south cross section which shows the typical lithologic facies relationships in the Fort Union Formation across the central part of the Wind River Basin. Five general lithologic facies are recognized: 1) a sandy fluvial facies; 2) a coaly fluvial facies; 3) a mud-dominated facies; 4) a lacustrine shale facies; and 5) a lacustrine deltaic facies.

Datum on the cross section (**Figure 4**) is the base of the Waltman Shale Member of the Fort Union Formation, which is easily recognized on geophysical logs. The Lance-Fort Union contact shown is the base of the sandy interval at the base of the Fort Union as defined by Keefer and Troyer (1964). The uppermost part of the Lance is Paleocene in age in this conformable sequence, as noted earlier. The Paleocene-Cretaceous boundary cannot be located precisely without fossil

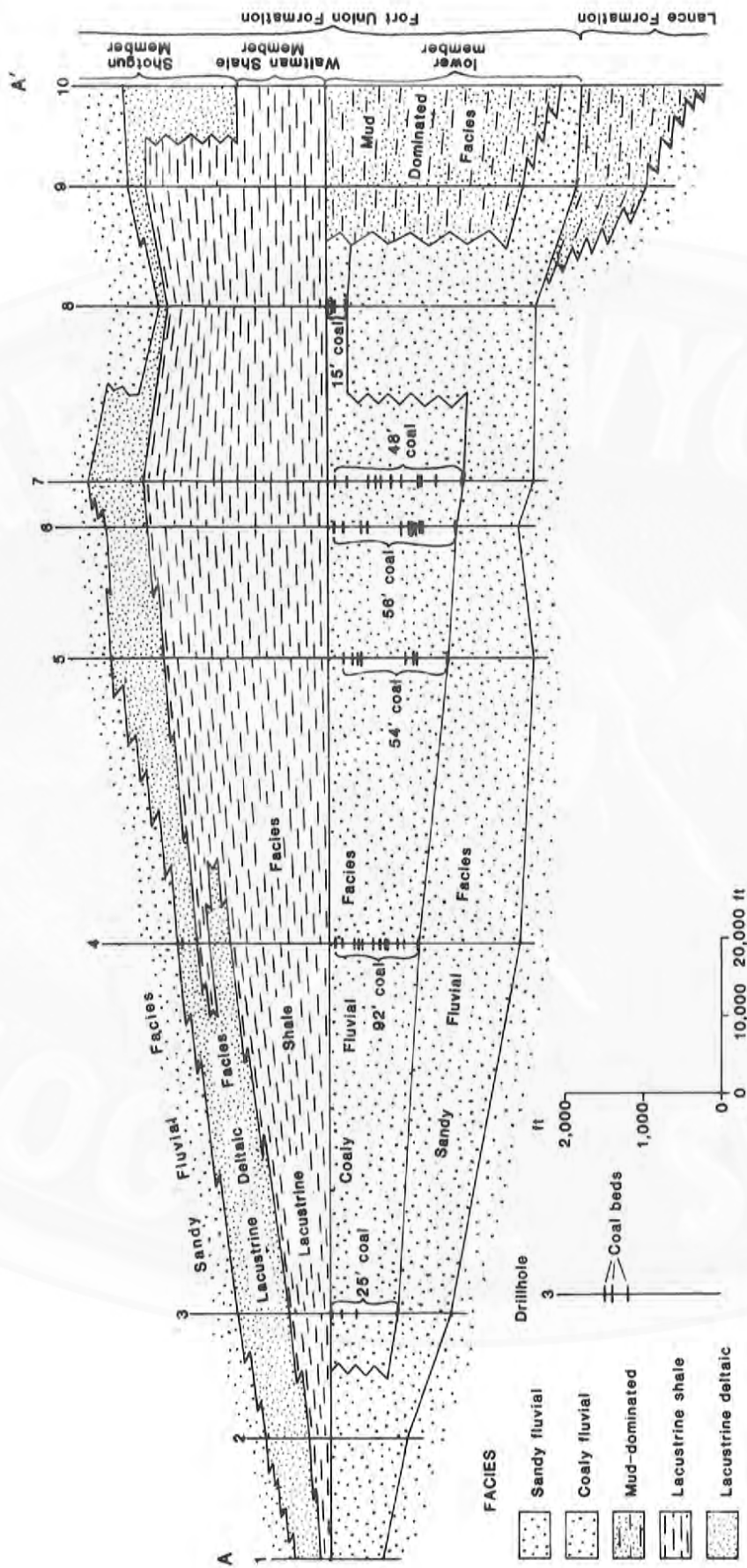


Figure 4. Generalized north-south cross section from the southern margin of the basin to the basin trough showing lithofacies in the Fort Union Formation. Location on Figure 5.

evidence and therefore is not a mappable contact. The base of the Waltman Shale in the basin east of the Wind River Reservation does not appear to be highly time transgressive. Liro and Pardus (1990), in a seismic stratigraphic study of the Waltman Shale Member in the central and eastern part of the basin, found that the basal contact of the Waltman coincides with one specific seismic event, strongly suggesting that Lake Waltman transgressed relatively rapidly across this area. Flemming and Nelson (1991) and Nichols and Flores (1993) found that the base of the Waltman Shale Member at the type section near Waltman corresponded approximately to the base of the P5 biozone.

Toward the western part of the basin, in contrast, a significant stratigraphic rise of the base of the Waltman appears to occur. The Waltman Shale Member has not been identified as extending into this part of the basin, but the Shotgun Member, which is believed to be largely laterally equivalent to the Waltman (Keefer, 1961 a and b; 1965), has been mapped there on both the surface (Keefer and Troyer, 1964) and in the subsurface (Keefer and Johnson, 1993). The lower part of the Shotgun is largely fluvial. Coarsening upward deltaic sequences, which were probably deposited in the nearshore areas of Lake Waltman, occur considerably above the base of Shotgun. Drillhole data is as yet too sparse to fully understand the transgressive nature of the base of the Waltman in the western part of the basin.

The sandy fluvial facies occurs in the lower part of the lower member of the Fort Union Formation and in the Shotgun Member above the lacustrine deltaic facies. In the lower member, it varies from about 650 to 2,400 ft (198 to 732 m) thick along the line of section and from 0 to 2,450 ft (0 to 747 m) thick in the basin. It contains about 30 to 50% fluvial sandstones and little coal. Most individual fluvial sandstones in this facies are from 15 to 35 ft (5 to 11 m) thick, and become finer-grained upwards, but complex multistory bodies from 50 ft to over 300 ft thick are also common.

The coaly fluvial facies occurs above the sandy fluvial facies in the lower member and below the Waltman Shale Member. It is somewhat less sandy than the sandy fluvial facies, with about 25 to 40% fining-upwards sandstone units from 15 to 35 ft (5 to 11 m) thick, and rarely contains complex multistory sandstone units. It varies from about 270 to 1,700 ft (82 to 518 m) thick along the line of section and from 0 to 1,700 ft (0 to 518 m) thick throughout the basin as a whole. It grades to the south into the sandy facies and to the north into the mud-dominated facies. The coaly facies reached its maximum extent immediately prior

to Lake Waltman forming. Maximum total thickness of coal in beds 2 ft (0.6 m) thick or greater found thus far in the basin is 105 ft (32 m). Well no. 4 (Figure 4) contains the most coal along the line of section with an estimated 92 ft (28 m) cumulative thickness. Total coal thickness decreases both north and south of well no. 4. Several intervals with thick coal beds occur in the facies, and some of these coaly intervals appear to be fairly persistent laterally. For instance, thick coal beds occur near the top and near the base of the coaly facies throughout much of the central part of the basin. Some coal beds occur above channel sandstones but most occur in fine grained intervals such as carbonaceous shales. Further study is needed to better understand the distribution of coal beds in the coaly facies.

The mud-dominated facies occurs at the north end of the cross section along the rapidly subsiding trough of the basin. This facies, which attains a thickness of as much as 3,500 ft (1,067 m) along the deep basin trough, does not appear to crop out extensively in the basin. Sandstones in this facies are generally less than 10 ft (3 m) thick. A similar mud-dominated facies occurs in the Lance Formation near the trough of the basin. These rocks crop out near Waltman in the northeastern part of the basin. These outcrops were most recently studied by Gillespie and Fox (1991) who suggested that this facies of the Lance was deposited in waterlogged floodplains, lakes, and swamps that formed along the rapidly subsiding trough of the basin.

The lacustrine shale facies and the lacustrine deltaic facies have been extensively studied in the past (Keefer, 1961; Ray, 1982; Liro and Pardus, 1990; Katz and Liro, 1993); these facies were not studied in detail as part of the present study. The lacustrine shale facies, which crops out only in the vicinity of the designated type section near Waltman, was deposited in offshore areas of Lake Waltman. The lacustrine deltaic facies, which also crops out in the vicinity of Waltman, was deposited in the marginal areas of Lake Waltman. The coarsening upward cycles that characterize this latter facies were interpreted by Phillips (1983) to be fan deltas.

Three generalized lithofacies maps of the Fort Union Formation were constructed for this study (Figures 5-7). Figure 5 shows the distribution of lithofacies in the lower non-coaly part of the lower member. The sandy fluvial facies covers all but a narrow area along the rapidly subsiding trough of the basin and a relatively small area in the eastern part of the basin where mud-dominated facies prevailed. Fluvial sandstones generally appear to be scattered

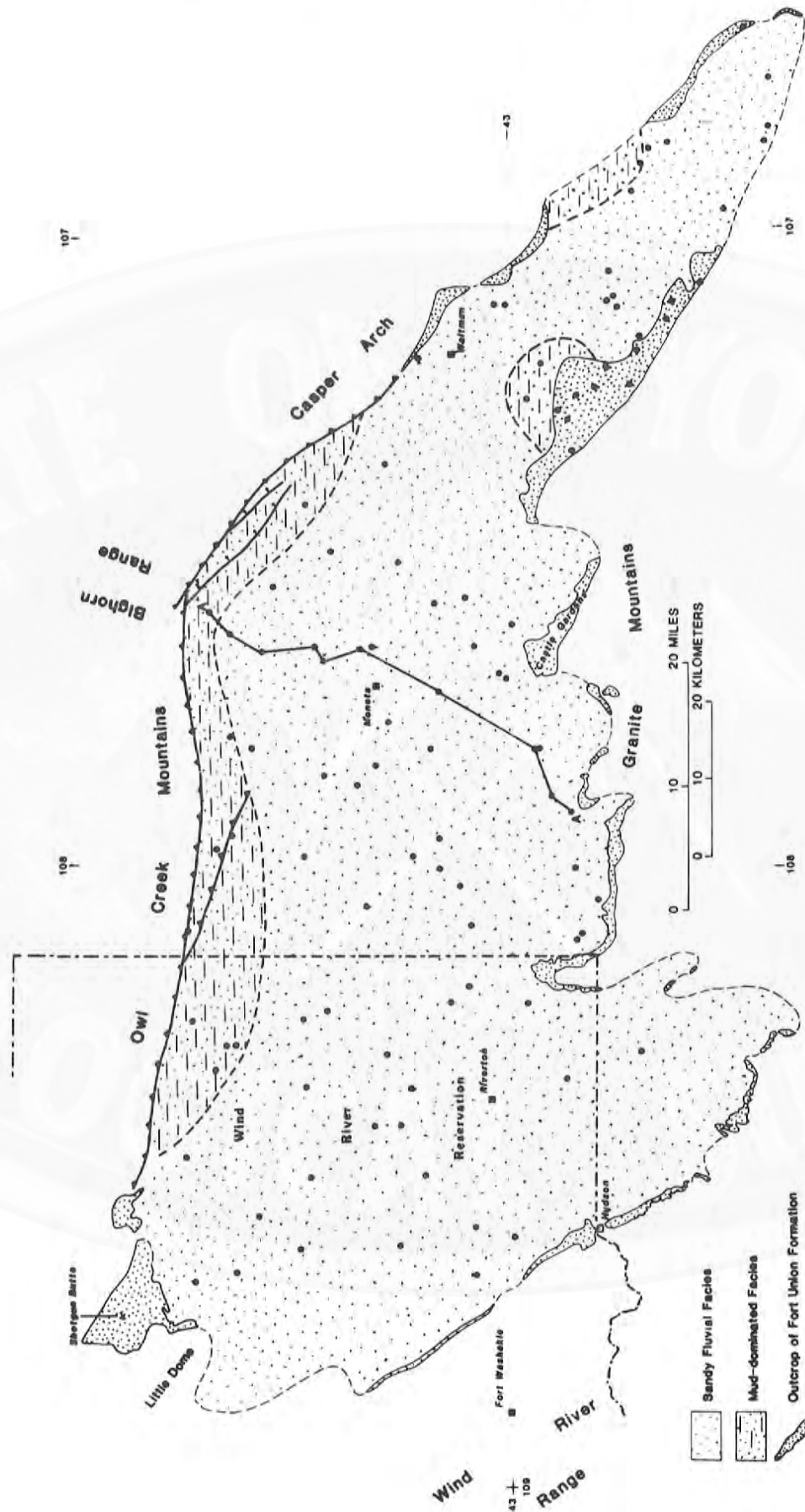


Figure 5. Generalized lithofacies map for the lower part of the lower unnamed member of the Fort Union Formation. Dots are drillholes used for control.

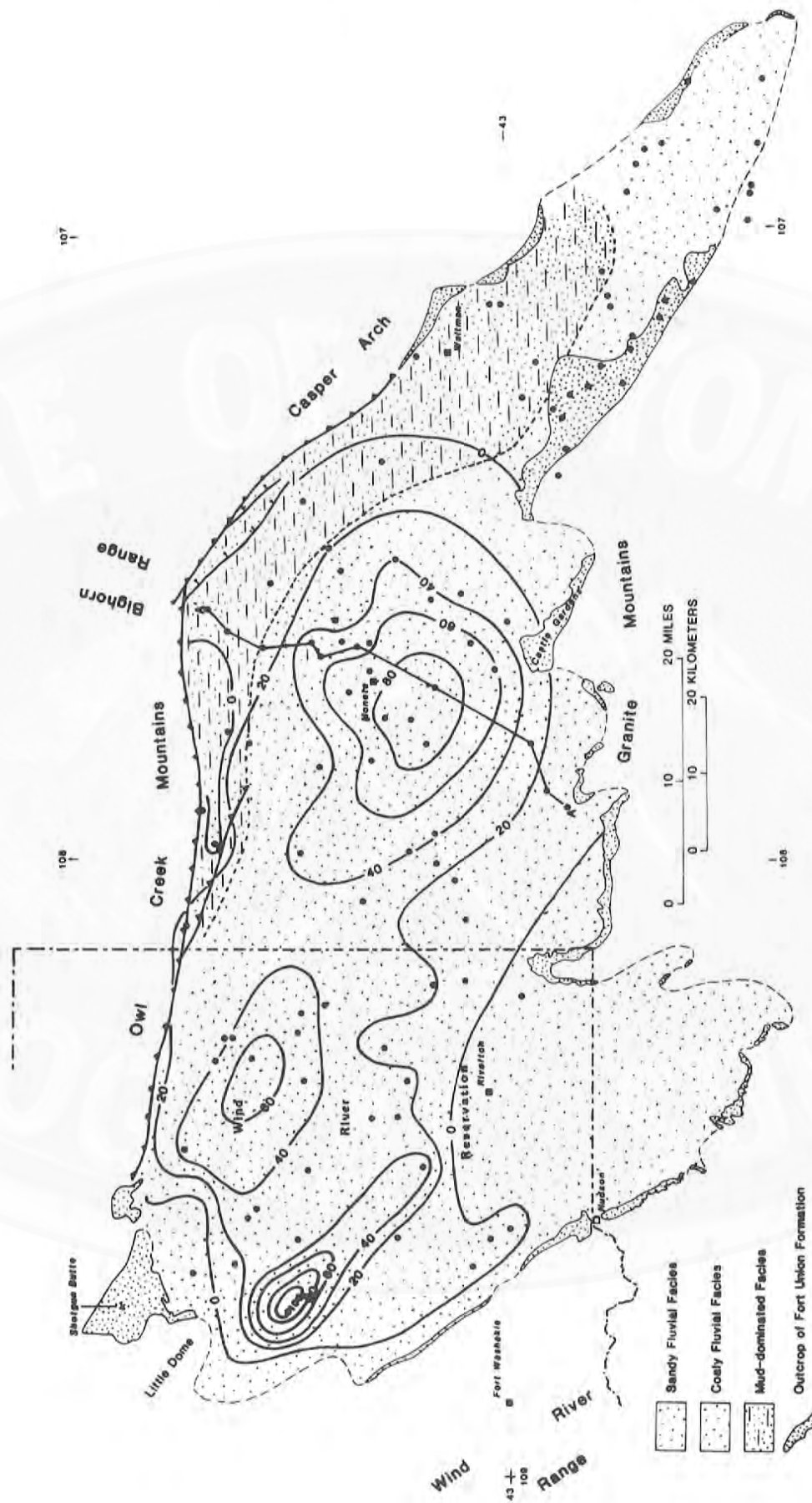


Figure 6. Generalized lithofacies map during maximum extent of the coaly fluvial facies of Fort Union Formation. Dots are drillholes used for control. An isopach map of total coal in the coaly facies in beds 2 ft (0.6 m) thick or greater is shown. The extent of the coaly facies is approximately outlined by the 20 ft (6 m) isopach line.

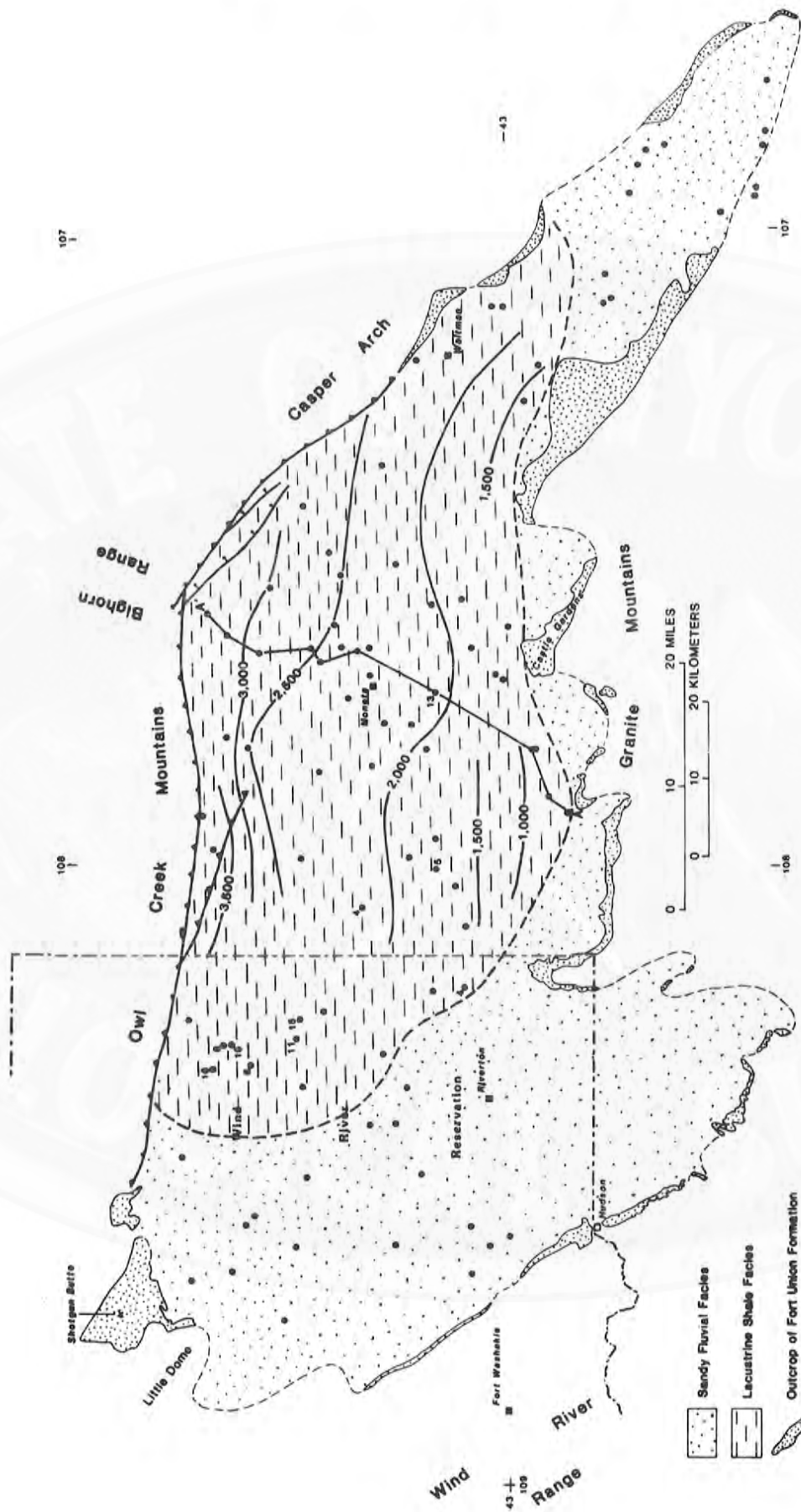


Figure 7. Generalized lithofacies map during maximum extent of Lake Waltham. Dots are drillholes used for control. An isopach map of the lower member of the Fort Union Formation is included. The number next to a drillhole is the total thickness of coal in beds 2 ft (0.6 m) thick or greater in the deltaic facies associated with Lake Waltham.

throughout the 400 to 1,200 ft (122 to 366 m) interval, and more detailed work will be required to define the locations of individual fluvial channel systems during deposition of this interval.

Figure 6 is a lithofacies map for the coaly interval of the Fort Union Formation and an isopach map of total coal in beds 2 ft (0.6 m) thick or greater. The coaly facies occurs along a broad belt in the western two-thirds of the basin. It grades into the non-coaly fluvial facies in the southern and eastern parts of the basin, beyond the zero isopach line, and into the mud-dominated facies to the north along the basin trough. The mud-dominated facies generally extends farther to the south than it did during deposition of the lower part of the lower unnamed member. The top of the coaly interval is placed at the top of the highest significant coal (Figure 4). In the eastern and central parts of the basin, the highest coal bed occurs just below the base of the Waltman Shale. In the western part of the basin, the highest coal generally occurs just below the fan deltas in the middle of the Shotgun Member, indicating that coal-forming environments persisted in the western part of the basin after Lake Waltman had formed farther to the east. The isopach map includes all coal in the coaly facies even though the upper contact of the facies is time transgressive.

The thickest Fort Union coals occur in three areas, two in the western part of the basin and one in the central part. General descriptions of the two thick coal accumulations toward the west are given by Johnson and Flores (1993). The northernmost of these two includes as much as 69 ft (21 m) of coal and is just south of the trough of the basin. Maximum thickness for an individual coal bed in this area is about 17 ft (5

m). The second thick accumulation, farther to the southwest, is in a northwest trending syncline between Little Dome anticline and the southwest flank of the basin. This accumulation contains as much as 101 ft (31 m) of coal with a maximum thickness for an individual coal bed of about 23 ft (7 m). The thick accumulation in the central part of the basin, heretofore unreported, includes as much as 92 ft (28 m) of coal. Maximum thickness for an individual coal bed in this area is about 20 ft (6 m).

Figure 7 shows the maximum extent of lacustrine rocks deposited by Lake Waltman in the Wind River Basin. It is a general map for the entire Lake Waltman interval, and does not distinguish between the lacustrine shale and the deltaic facies. More complete facies analysis of the Waltman Shale have been published by other workers (Ray, 1982; Liro and Pardus, 1990). The map extends Lake Waltman into the western part of the basin where the fan deltas occur in the middle part of the Shotgun Member. A generalized isopach map of the lower member of the Fort Union Formation (Figure 7) shows that it thickens from less than 1,000 ft (305 m) along the south margin of the basin to more than 3,500 ft (1067 m) along the main basin trough.

The small numbers next to the plotted control points are total thicknesses of coal in feet in the deltaic and sandy fluvial facies associated with the infilling of Lake Waltman. A maximum of only 16 ft (5 m) of coal was found in this interval indicating that these environments were generally not favorable for thick peat accumulation. Most of the geophysical logs examined indicated no coal in this interval. Most of the coal in this facies occurs in the western part of the basin, but some coal also occurs in the central part of the basin.

Discussion

The results of this study indicate that environments favorable for significant peat accumulation were widespread in the Wind River Basin during deposition of the Paleocene Fort Union Formation. As much as 100 ft (30 m) or more of total coal in beds 2 ft (0.6 m) thick or greater is present along a broad belt north of the slowly subsiding southern margin of the basin and the rapidly subsiding basin trough to the north. Environments favorable for peat accumulation began to form during P3 biozone time as the river systems, flowing from the nearby Granite Mountains and Wind River Range to the south and southeast, evolved from very sandy and conglomeratic, braided and meandering bedload systems, to finer grained,

more meandering and anastomosing mixed-load systems. These river systems appear to have flowed northeastward. The predominantly fluvial facies grades into a mud-dominated facies along the basin trough. This mud-dominated facies was not favorable for peat accumulation, probably due to suffocation of low-lying mires by mud, and hence has little or no coal.

From the data available, it appears that fluvial channel sandstones are scattered throughout the coaly facies along most if not the entire southern margin of the Wind River Basin. This suggests that several river systems were probably present during deposition,

and that avulsion events were fairly common. It seems highly unlikely that any single coal bed would be persistent laterally for any great distance, however, several coaly zones in the coaly facies appear to be fairly continuous. Available drillholes are too sparse for a detailed analysis of the coal-bed geometries within these zones.

This style of deposition appears to be somewhat different than that found in the coaly Paleocene intervals in the Powder River Basin to the east-northeast where the meandering and anastomosed channel systems remained relatively stationary for extended periods of time producing sinuous areas with little coal known as "want areas" (Flores and Hanley, 1984; Flores, 1986; Warwick and Stanton, 1988). The maxi-

mum thickness of about 20 ft (6 m) for an individual coal bed in the Wind River Basin is also considerably less than in the Powder River Basin where individual coal beds attain thicknesses of as much as 250 ft (76 m).

The maximum development of the coaly fluvial facies occurred immediately prior to the formation of Lake Waltman, and peat accumulation ceased with the initial transgression of Lake Waltman. Lake Waltman transgressed rapidly across the central and eastern parts of the basin at about the beginning of P5 biozone time, whereas the lake may not have reached the western part of the basin until well into P5 biozone time. The deltaic environments associated with Lake Waltman were generally not favorable for significant peat accumulations.

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Origin of oil in oil-saturated sandstones in the upper part of the Fort Union Formation (Paleocene), Castle Gardens and Signor Ridge areas, Wind River Basin, Wyoming

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Introduction

The Paleocene Fort Union Formation in the Wind River Basin produces mainly gas and lesser quantities of oil. However, in the Castle Gardens and Signor Ridge areas, located in the south-central portion of the

basin (Figure 1), there are exhumed oil-saturated fluvial channel sandstones that occur stratigraphically near the top of the Fort Union Formation. Nearest Tertiary Fort Union oil production is at the Haybarn

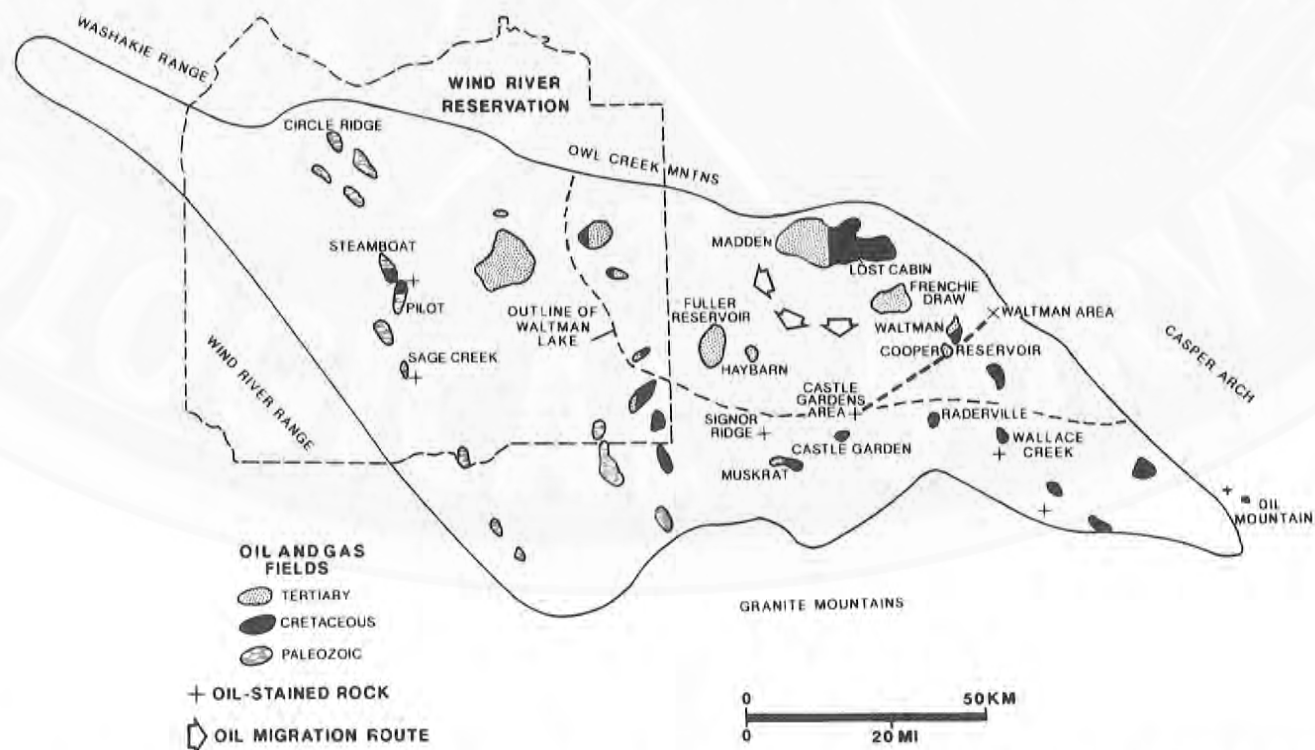


Figure 1. Location map of: 1) principal oil and/or gas fields in Wind River Basin, 2) some of the oil-stained sandstone deposits studied, and 3) former Lake Waltman. Tertiary fields are mainly gas; Tertiary oil production is only at Haybarn, Fuller Reservoir, and Greater Madden (Lost Cabin) fields.

field some 12 to 15 mi to the north-northwest. Preliminary estimates of volumes of oil staining at the Signor Ridge locality indicate that there may be 23-47 million bbls of heavy, mostly immobile, biodegraded oil in-place. At the Castle Gardens area, no total volume estimates were made. However, based on a single analysis of an oil-saturated sandstone sample and assuming that the surface sandstone deposit persists into the subsurface, calculations indicate that there could be as much as 1,100 bbls of oil per acre foot (Palacas and others, 1992).

The question immediately is asked "What is the source of the oil?" This study addresses that question by utilizing standard organic geochemical parameters (gross chemical composition, gas chromatography of saturated hydrocarbons and stable carbon isotopes)

and computerized gas chromatographic-mass spectrometric biomarker analyses to characterize the oils. These parameters were then used as oil-oil correlation tools to try to match the oils to one of the three principal genetic oil families in the Wind River Basin. The three main oil families are: 1) Tertiary-type oils, presumably derived chiefly from the Paleocene Fort Union (Waltman) lacustrine shales, 2) Cretaceous-type oils, attributed to Cretaceous marine shales, and 3) upper Paleozoic (Phosphoria)-type oils, derived from Lower Permian Phosphoria Formation source rocks (Palacas and others, 1993a, 1993b). Discovering the genetic oil type permits identification of source rocks and prediction of migration pathways. Such information can help to target potential areas of exploration in the basin.

Regional setting

The Wind River Basin formed during the Laramide deformation that was initiated during the Late Cretaceous at which time the surrounding major mountain ranges to the south (Granite Mountains), west (Wind River Range), and north (Washakie Range) were uplifted (Figure 1). This initial phase of Laramide deformation was followed by a period of continued uplifts and thrust faulting along the basin margins, accompanied by rapid basin subsidence, and structural closure of the basin at the east-northeast by the Owl Creek Mountains and Casper arch during the Paleocene (see Flores and others, this volume). The Laramide deformation culminated during the Eocene resulting in continued uplift and denudation of the surrounding ancestral mountain ranges. The uplifts surrounding the Wind River Basin consist of folded and faulted rocks of Precambrian, Paleozoic, and Mesozoic ages; the latter age rocks flank the basin. The interior part of the basin consists of Tertiary rocks deposited during the Paleocene and Eocene time. In general, the Paleocene rocks dip basinward and in turn are overlain by near flat-lying Eocene rocks. The Tertiary rocks are deformed by anticlines, synclines, and minor faults and these rocks mask structural complexities of older rocks (Keefer, 1965; Keefer and Johnson, 1993).

The Paleocene strata in the Wind River Basin consist mainly of the coal-bearing Fort Union Formation (Figure 2). The Fort Union Formation is undifferentiated along the southern and western margins of the basin. Basinward to the north-northeast, the Fort Union Formation is divided into two parts, an un-

named lower part, and an upper part divided into two members — Waltman Shale Member below, and Shotgun Member above. A basinward stratigraphic relationship of the undifferentiated Fort Union Formation from the Castle Gardens area on the south-central part of the basin and its members through the basin axis (Cooper Reservoir field) and to the Waltman area on the northeast part of the basin is shown in Figure 3. The Waltman Shale Member and Shotgun Member are

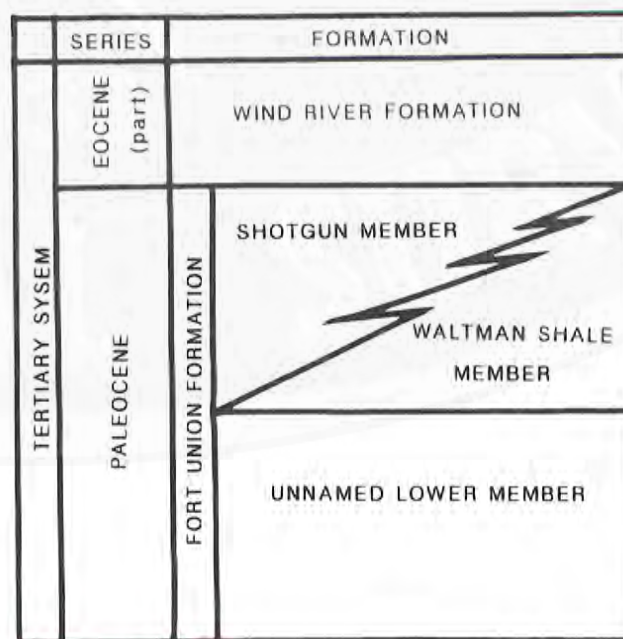


Figure 2. Generalized stratigraphic column of Fort Union Formation.

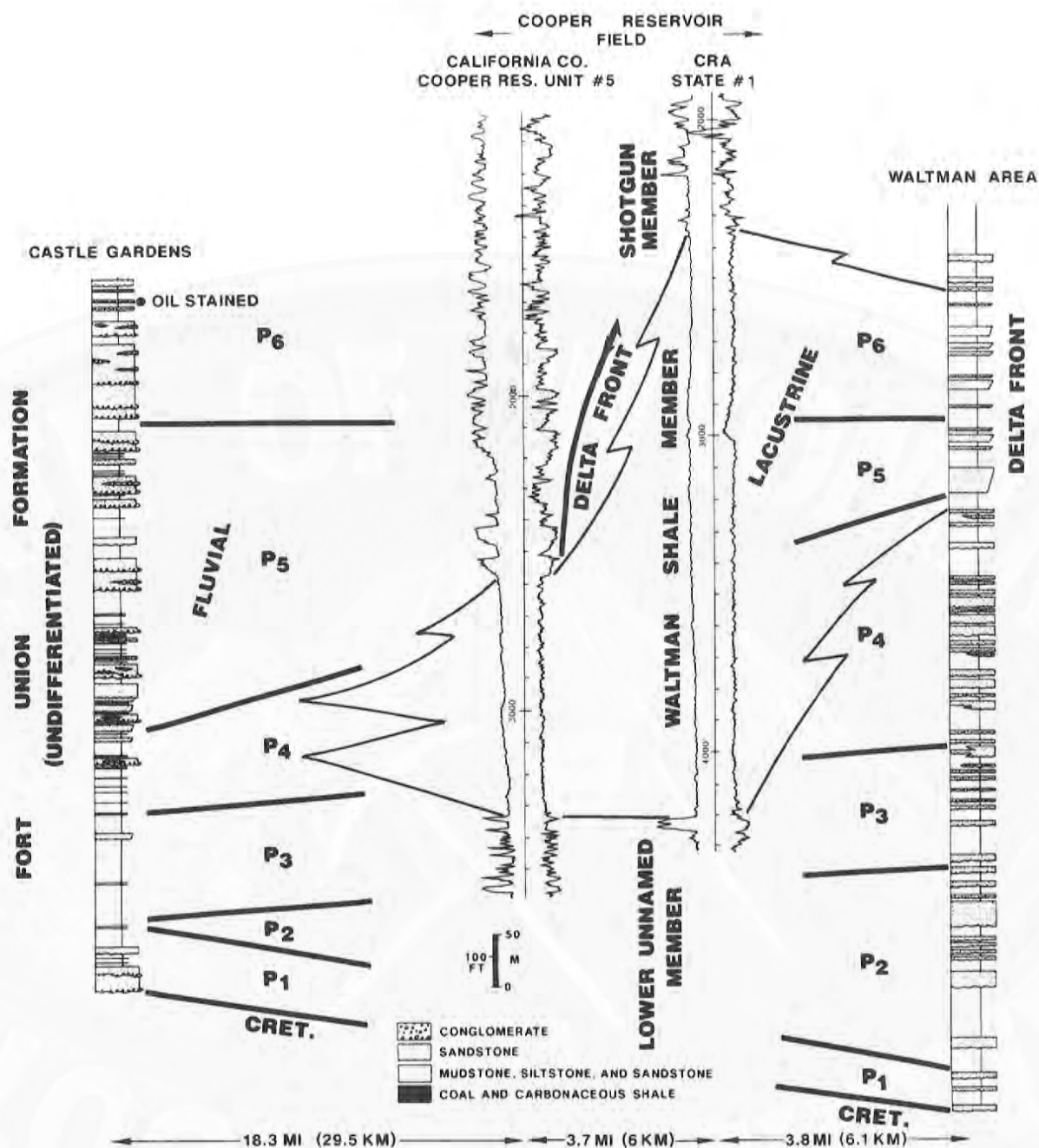


Figure 3. SW-NE cross section of Fort Union Formation. See Figure 1 for line of section.

intertongued with the latter member thinning and pinching out basinward. The Waltman Shale Member also intertongues with the unnamed lower part.

The coal-bearing, Paleocene Fort Union Formation represents deposits of fluvial, deltaic, and lacustrine environments (Keefer, 1965; Phillips, 1982; Robertson, 1984; Flores and others, 1992; Flores and Keighin, 1992; Flores and Keighin, 1993; Flores and others, 1993; Johnson and Flores, 1993; Nichols and Flores, 1993). The basin was initially infilled by detri-

tal sediments, which are represented by the unnamed lower part and stratigraphically equivalent Fort Union Formation, of a flow-through trunk-tributary fluvial system during the early to middle Paleocene. The tributaries of this fluvial system drained the surrounding uplifts and joined the basin-axis trunk stream that flowed to the east-northeast. However, when the east and northeast margins of the basin were closed by uplifts of the Owl Creek Mountains and the Casper arch during the middle to late Paleocene, it was infilled by fluvio-deltaic and lacustrine sedi-

ments. The Waltman Shale Member was mainly deposited in a large freshwater lake that persisted through late Paleocene. This Waltman Lake (see Figure 1) was drained by fluvial and deltaic systems and their deposits are represented by the upper part of the undifferentiated Fort Union Formation, the sandy part of the Waltman Shale Member, and Shotgun Member. The coal beds interbedded with the undifferentiated Fort Union Formation, unnamed lower part, and Shotgun Member were probably formed in low-lying paleomires associated with the fluvial floodplains, abandoned river courses, and deltaic-lobe complexes.

Organic matter content of the organic-rich shales and mudstones of the Waltman Shale Member may have been derived from physical weathering and erosion of peat deposits from these paleomires and/or reworking of older Cretaceous coal beds, in addition to in situ input of freshwater algal material. Matura-

tion of these organic-rich shales and mudstones upon deep burial may have enabled these source rocks to generate and expel the highly paraffinic oil produced in the Fort Union Formation. The reservoir rocks for this oil consist of fluvial channel sandstones as typified in the Castle Gardens and Signor Ridge areas (see paper by Keighin and Flores, this volume). These reservoir rocks are composed of texturally heterogeneous fine- to medium-grained sandstones locally partitioned by lenses of pebble and cobble conglomerates. It is also internally compartmentalized by either an orderly or randomly arranged pattern of sedimentary structures and by persistent shallow to deep internal scours (see paper by Keighin and Flores, this volume). These channel sandstones were probably deposited by braided and meandering fluvial systems, and a single, low-sinuosity stream on deltaic plains. Delta-front sandstones may also have served as reservoir rocks for the waxy oil sourced from the Waltman Shale Member.

Reservoir characteristics

Sandstones collected from outcrop and from cores, are typically quartz-rich, and contain minor to significant amounts of rock fragments, usually chert. The sandstones are generally at least moderately well sorted. Porosity is highly variable, but log porosity is usually less than 20%; modal (thin section) porosity was measured as high as 26%, but this high value may reflect influence of surface weathering. Much of the porosity is secondary and due to leaching. Clay intervals, predominantly kaolinite and illite/smectite, are seen both in thin section and by scanning electron microscopy (SEM). Oil-stained interstitial clays occur

both at Castle Gardens and Signor Ridge, although free oil is also seen in thin sections at Signor Ridge. The presence of clays has caused significant problems during well completions at the Fuller Reservoir field (Pirnor, 1978, p. 281). SEM analysis and capillary pressure/mercury injection test on related reservoir sandstones show that small ($<0.1\mu\text{m}$) pore throats are dominant in controlling fluid flow in these sandstones. Discussion of fields producing hydrocarbons from similar rocks are given by Pirnor (1978; Fuller Reservoir) and Robertson (1984; Haybarn field).

Samples and methods

Samples

Two oil-saturated sandstone samples were collected from exposures of exhumed fluvial channel sandstones present in the top part of the Paleocene Fort Union Formation. One sample was from the Castle Gardens area located some 16 mi southeast of the Haybarn field and about 12 mi west of the Raderville field in the south-central portion of the Wind River Basin (Figure 1, Table 1). The other was from the Signor Ridge area some 12 mi west of the Castle Gardens locality and approximately 4 mi north of the Muskrat field (Figure 1).

Three crude oil samples, each typical of the three principal genetic oil types determined in the Wind River Basin, namely Tertiary, Cretaceous, and upper Paleozoic (Phosphoria) (Palacas and others, 1993a and b), were used for comparative purposes (Table 1). Representative of the Tertiary family of oils is a Fort Union Formation crude oil from the Haybarn field. Two other Tertiary (Fort Union) crude oils (Fuller Reservoir and Lost Cabin fields), which have essentially identical compositional features, could also have been used. For the Cretaceous oil type, a Dakota Formation equivalent crude from the Steamboat Butte field was used. However, for illustrating the biomarker distri-

butions (m/z 191 and m/z 217), crude from the Upper Cretaceous Shannon Sandstone Member of the Cody Shale from the Castle Garden field was selected. Typifying the Phosphoria genetic oil type, an oil from the Tensleep Sandstone (Middle Pennsylvanian to Lower Permian) from the Steamboat Butte field was used.

Methods

The oil in the sandstones was extracted with chloroform in a Soxhlet extractor. Activated copper strips were used to remove free sulfur. The extracted oil was separated by liquid-solid chromatography, using a modified Baker column fractionation technique (Anders and others, 1987). Rather than a cyano-silica gel stationary phase, an alumina-silica gel packing was used to obtain three successive fractions: 1) saturated hydrocarbons (eluted with isoocane), 2) aromatic hydrocarbons (eluted with benzene), and 3) non-hydrocarbons composed of resins (N-S-O compounds) and asphaltenes (eluted with benzene-methanol). It should be pointed out that the Baker column fractionation technique generally yields a higher percentage of saturated hydrocarbons relative to aromatic hydrocarbons, which in turn, yields a higher saturate to aromatic ratio than commonly obtained by other methods. Therefore, caution should be exercised when comparing values obtained by the Baker column technique versus the standard silica-gel alumina glass column chromatographic technique.

Gas chromatography was performed on the saturated hydrocarbon fraction using a HP5880 Gas Chromatograph (GC) equipped with a 50 m by 0.32 mm, 5% phenyl methyl silicone (D)

Table 1. Comparison of bulk geochemical data of two oils from oil-saturated sandstone outcrops and representative crude oils of the three major oil types in the Wind River Basin, Wyoming.

Sample number	Field or locality	Well name	Section, Township, Range	County	Formation	Depth feet	Extract (bitumen) ppm	Sat HC %*	AROM HC %*	Resins plus asphaltenes %*	sat/AROM	Pr/Ph	Pr/n-C17	Ph/n-C18
Oil saturated sandstones at surface														
92 005 022	Castle Gardens	Outcrop	29 35N 90W	Fremont	Fort Union	0	46,680	68.6	7.1	24.3	9.7	-	-	-
92 069 012	Signor Ridge	Outcrop	17 34N 92W	Fremont	Fort Union	0	-	53.7	23.8	22.5	2.3	-	-	-
Tertiary-type crude oil														
92 130 001	Haybarn	Northwest 3-24 Union Federal	24 36N 93W	Fremont	Fort Union	4408-4433	NA	89.7	7.6	2.7	11.9	1.86	.38	.22
Cretaceous-type crude oil														
92 069 010	Steamboat Butte	Chevron 2 Tribal	9 3N 1W	Fremont	Dakota equivalent	4408	NA	84.7	9.5	5.8	8.9	1.79	.55	.34
Upper Paleozoic (Phosphoria)-type crude oil														
92 069 011	Steamboat Butte	Chevron-ES Tribal	5 3N 1W	Fremont	Tensleep	6843	NA	59.9	29.8	10.3	2.0	.63	.28	.47

SAT HC = saturated hydrocarbons; AROM = aromatic; Pr = pristane; Ph = phytane; - = not available, NA = not applicable *Normalized to 100%.

glass capillary column. The GC was programmed at 50°C for 2 minutes then from 50° to 320°C at 4°/min. and then held for 10 minutes using hydrogen as the carrier gas. The detector output was digitized and stored on computer disk using the Nelson chromatography software.

Gas chromatography-mass spectrometry (GC-MS) was performed using an HP5890 GC coupled to a VG7070 MS. GC-MS analyses were performed on combined saturated and aromatic hydrocarbon fractions and distributions of several biomarker compound classes were determined by selective ion monitoring of diagnostic ions (tricyclic, tetracyclic and pentacyclic terpanes m/z 191, and steranes m/z 217).

Compound identifications were based on relative retention times and comparison with published data. Relative abundances of biomarkers were determined from peak areas or peak heights of diagnostic ions.

$^{13}\text{C}/^{12}\text{C}$ ratios of the saturated- and aromatic-hydrocarbon fractions were determined on a Finnigan-MAT 251 mass spectrometer. The hydrocarbon fractions were converted to CO_2 by combustion under oxygen with subsequent purification in a high-vacuum system. $^{13}\text{C}/^{12}\text{C}$ ratios are expressed in the usual δ -notation; $\delta^{13}\text{C}$, per mil = $\{[(^{13}\text{C}/^{12}\text{C})_{\text{sample}} / (^{13}\text{C}/^{12}\text{C})_{\text{PDB}}] - 1\} \times 10^3$ relative to the PDB marine-carbonate standard.

Results and discussions

Bulk composition

Table 1 compares the bulk composition of oils (bitumens) in the oil-impregnated sandstones exposed at the surface at Castle Gardens and Signor Ridge and three crude oils typical of the three major families of oils in the Wind River Basin.

One clear-cut difference observed in comparing bulk compositions is the much higher relative percent of nonhydrocarbons (resins and asphaltenes) in the Castle Gardens and Signor Ridge oils (average ~23.5%) than in the crude oils (2.7-10.3%). This distinct difference is attributed primarily to biodegradation, as will be demonstrated below. The disparity in saturate/aromatic hydrocarbon ratios in the two surface oils, which are presumably derived from the same source, is also attributed to biodegradation. Despite the subjection of both oil accumulations to severe biodegradation, however, the composition of both oils still retains a relatively high percentage of total hydrocarbons (average 77%).

From the standpoint of using gross chemical composition alone as a tool for oil-to-oil correlation, it is difficult, if not impossible, to try to match the surface oils to any of the three prominent families of oils. The severity and variability of biodegradation hampers any viable correlation.

Gas chromatography

In Figure 4, gas chromatograms show the distribution of the $\text{C}_{10}+$ saturated hydrocarbons in Castle

Gardens and Signor Ridge oils (column A) and in typical oils of the three major oil types (column B).

Analyses indicate that the oils in the Castle Gardens and Signor Ridge sandstone outcrops are bacterially degraded and probably water washed. Biodegradation is evidenced by: 1) the complete absence of normal (n)-paraffins (n-alkanes) and acyclic isoprenoid hydrocarbons (such as pristane and phytane), and 2) the conspicuous unresolved complex mixture (UCM) of compounds or "hump" that rises significantly above the baseline. The prominent spikes projecting above the hump, especially in the Castle Gardens sample, are the more bacterially-resistant polycyclic saturated hydrocarbons, referred to as biological markers or simply biomarkers. Of the two oil-impregnated sandstone samples, the Signor Ridge oil is distinctly more severely biodegraded as indicated by the greater intensity of the hump and by the destruction and modification of many of the standard biomarkers used for typing oils, such as the hopanes and steranes (discussed below).

In sharp contrast, the three crude oils, which appear unscathed by bacterial alteration effects, contain the full complement of $\text{C}_{10}+$ saturated hydrocarbons. The prominent peaks are the n-paraffin components which are distributed in a regular homologous series.

It is obvious that, owing to the effects of severe biodegradation, gas chromatographic (GC) analysis as a correlation tool also falls short in attempting to match the two surface oils to any of the three major crude oil types in the Wind River Basin. Fortunately,

GAS CHROMATOGRAMS - SATURATED HYDROCARBONS

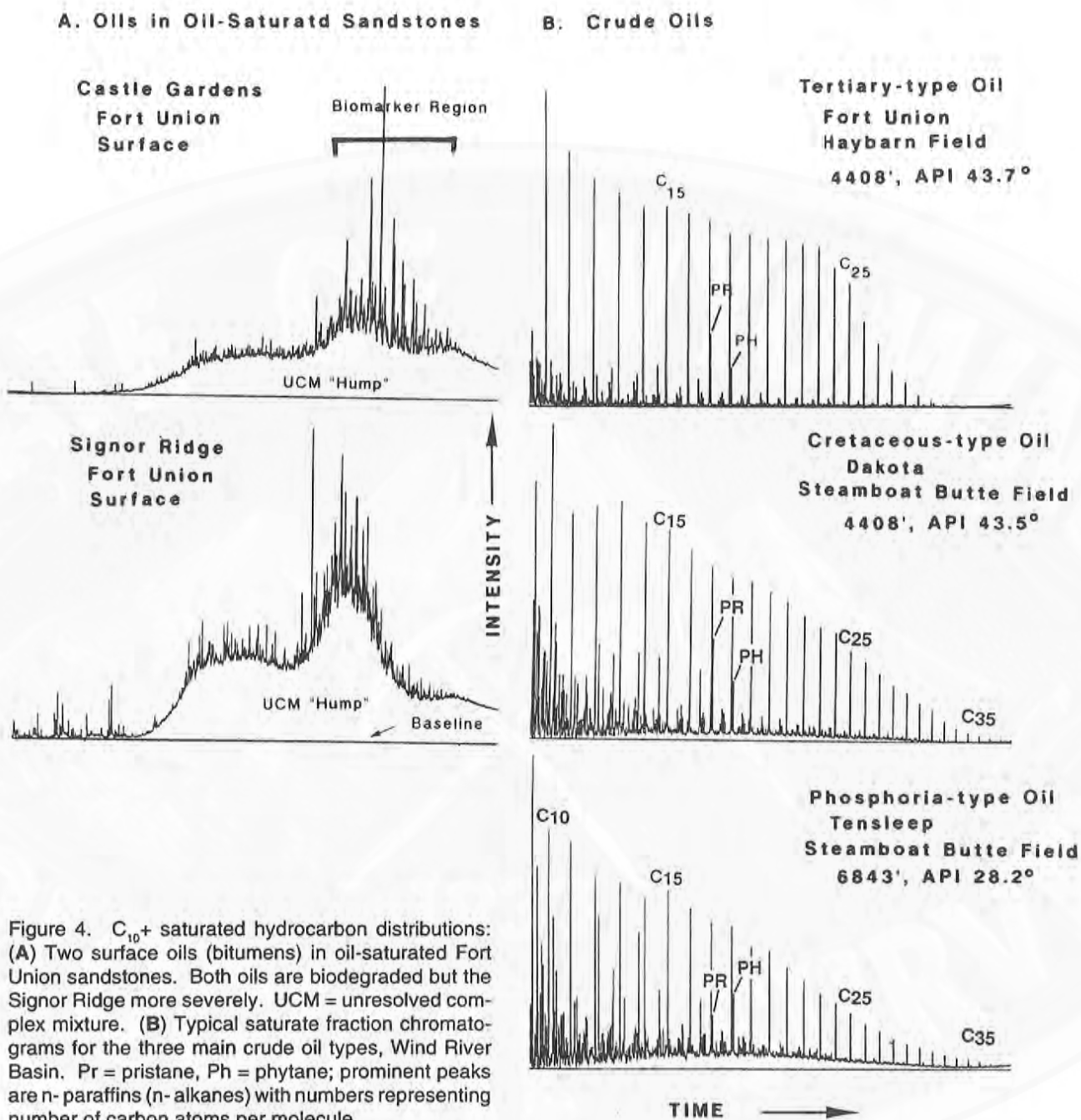


Figure 4. C₁₀+ saturated hydrocarbon distributions: (A) Two surface oils (bitumens) in oil-saturated Fort Union sandstones. Both oils are biodegraded but the Signor Ridge more severely. UCM = unresolved complex mixture. (B) Typical saturate fraction chromatograms for the three main crude oil types, Wind River Basin. Pr = pristane, Ph = phytane; prominent peaks are n-paraffins (n-alkanes) with numbers representing number of carbon atoms per molecule.

however, because of the preservation of key biomarker distributions, primarily in the Castle Gardens sample, the two surface oils can indeed be correlated to one of the major oil families.

It is also noteworthy that in comparing the GC signatures of the three crude oils, the Phosphoria-type crude can be easily distinguished from the Creta-

ceous- and Tertiary-type oils. The former crude oil has a Pr/Ph ratio less than one, whereas, the latter two oils are characterized by Pr/Ph ratios greater than one. Furthermore, one characteristic that may be used to differentiate between Tertiary- and Cretaceous-type oils, is that the former crude has a higher proportion of higher molecular-weight n-paraffins in the n-C₁₉ to n-C₂₅ region. This enrichment of higher

molecular-weight n-paraffins may account in large part for the characteristically high pour point of these Tertiary oils (~55-80°F) in contrast to pour points of ~5°F for the Cretaceous oils. In addition, the Tertiary oils, in contrast to the Cretaceous oils, commonly display a much higher ratio of odd to even numbered n-paraffins in the region of C_{25} - C_{31} , although the odd predominance is not so evident in the chromatogram of the Haybarn oil (Figure 4). This odd predominance is well illustrated in other Tertiary oils, such as the Fort Union Formation oil in the Fuller Reservoir field and the Waltman Shale Member oil in the Lost Cabin field (Palacas, unpublished data). The above distinguishing features, of course, apply to crude oils of comparable maturity and to oils not subjected to high levels of thermal maturation.

Carbon isotope correlations

Stable isotope ratios were used to attempt to correlate the biodegraded oils to their respective genetic oil types. The $^{13}C/^{12}C$ ratio of $C_{15}+$ saturated and aromatic hydrocarbon fractions of oils are plotted against each other in Figure 5. The encircled areas, based on about four data points each, are representative of the cluster of values obtained for the three principal oil families (Palacas, unpublished data). The Castle Gardens and Signor Ridge biodegraded oils have values that are significantly depleted in ^{13}C , especially for the saturated hydrocarbon fraction

(~ -30.5 to -32.0 ‰). From the plot it can be readily seen that the two biodegraded oils cannot be correlated to any of the oil families. In short, carbon isotope ratios, as an oil-oil or oil-source correlation tool proved ineffective for the above two biodegraded oils. However, lest any erroneous conclusions be drawn, oil-oil correlation is not precluded for all biodegraded oils. For example, in the Wind River Basin, three other heavily biodegraded surface oils—one oil seep, hosted in the Triassic Chugwater Formation near Sage Creek field, and two "tar sand" oils, hosted in Lower Cretaceous Cloverly Formation/Lakota Formation rocks near Oil Mountain area and Wallace Creek field—were successfully correlated to their genetically-related oil families, i.e., a Phosphoria-type oil and a Cretaceous-type oil, respectively, (Palacas, unpublished data).

Biomarker characteristics and correlations

Gas chromatographic-mass spectrometric (GC-MS) analysis (Figures 6, 7, and 8) revealed that key biomarker distributions in the Castle Gardens surface oil sample had escaped the effects of bacterial degradation and could be used for oil-oil correlation. On the other hand, the Signor Ridge sample sustained extensive alteration in the pentacyclic terpane (hopane) region (Figure 6) and in the regular sterane region (Figure 7), precluding reliable correlations with the three major oil types. However, based on overall distribution patterns of some of the more bacterial resistant biomarkers, the diasteranes (rearranged steranes; Figure 7) and the tricyclic terpanes (Figure 8), the Signor Ridge oil seems to closely resemble the Castle Gardens oil. The geographic proximity (Figure 1) and the comparable stratigraphic position of the oil-charged sandstones at Signor Ridge and Castle Gardens give further support for the similarity and genetic linkage of the two surface oils. Figure 3 shows the stratigraphic position of the oil-stained sandstones at the Castle Gardens section. The oil-saturated sandstones at Signor Ridge are also believed to occur at nearly the same position as those at Castle Gardens (i.e., either in the palynostratigraphic zone of P6 or in the uppermost part of the underlying zone P5). In the ensuing discussion, the Castle Gardens oil sample, by virtue of its intact biomarker distributions, will be primarily used for comparative purposes.

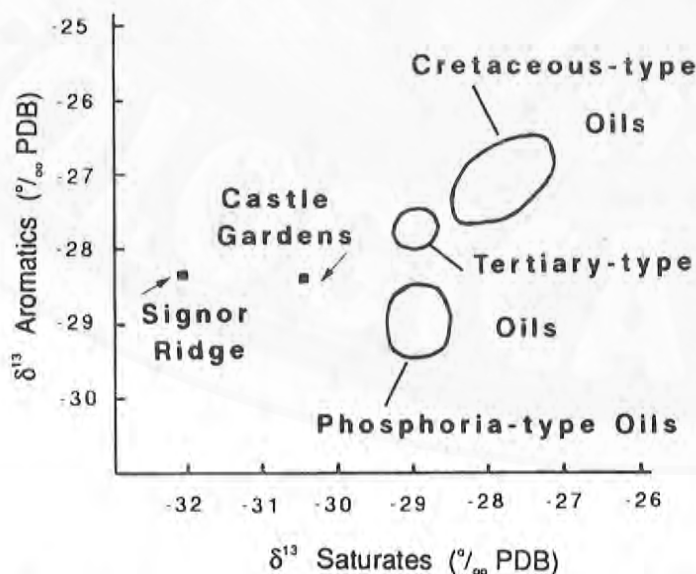
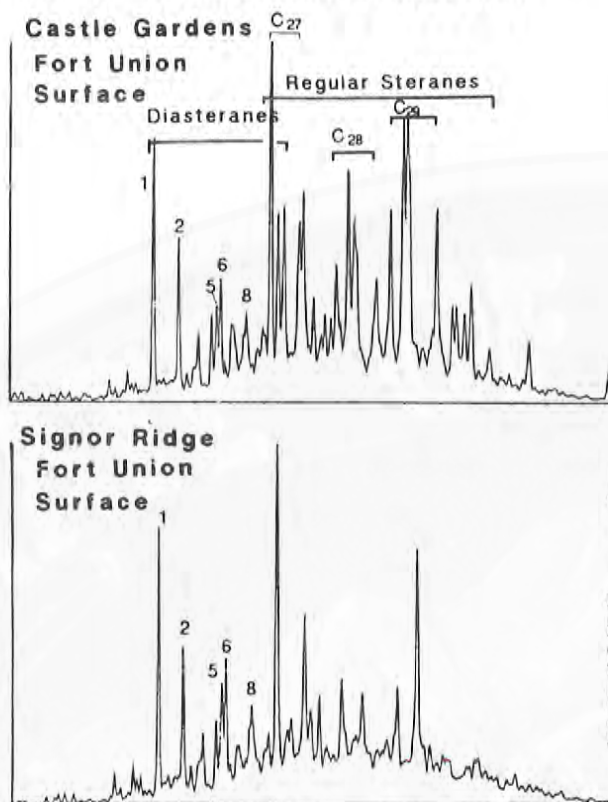


Figure 5. Cross plot of $\delta^{13}C$ of saturated and aromatic hydrocarbons. Oils in oil-saturated Fort Union sandstones at Castle Gardens and Signor Ridge are compared to oils of the three major genetic oil types, Wind River Basin.

STERANE (m/z 217) FRAGMENTOGRAMS

A. Oils in Oil-Saturated Sandstones



B. Crude Oils

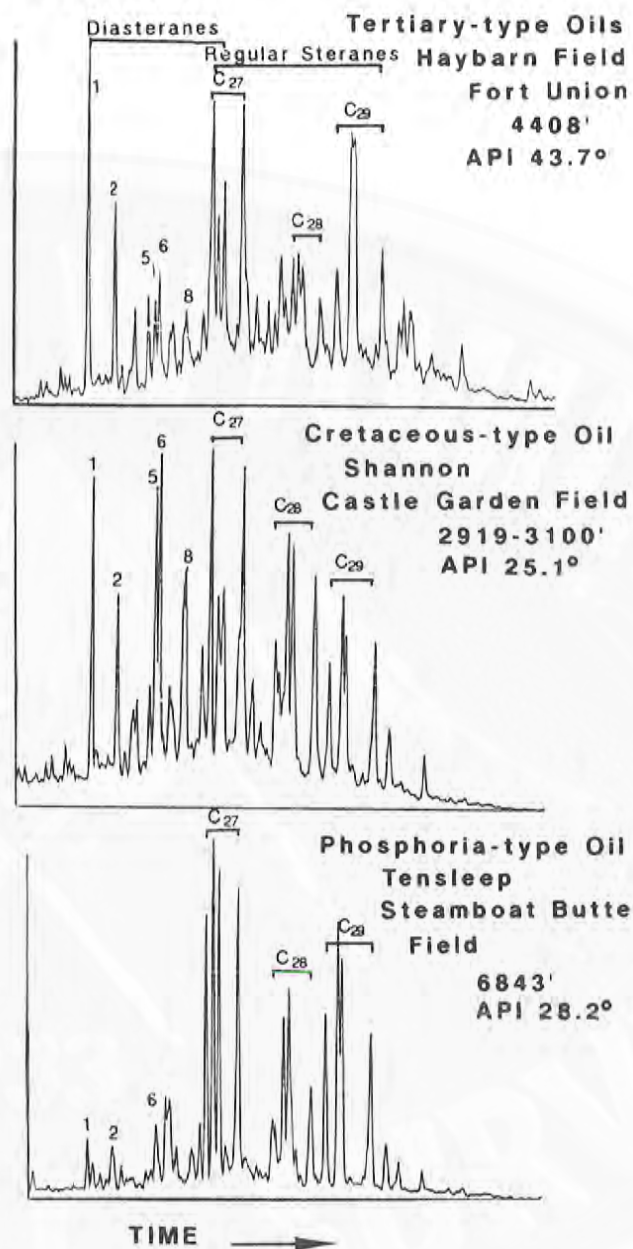


Figure 7. Sterane distributions as shown by m/z 217 mass fragmentograms. (A) Fort Union oil-saturated sandstones at surface. Sterane distributions at Castle Gardens sample appears essentially intact. At Signor Ridge, the regular steranes are severely biodegraded but the diasteranes (rearranged steranes) appear mostly intact. (B) Typical fragmentogram patterns for the three main oil types, Wind River Basin. Peaks 1 and 2 = C_{27} , $13\beta(H)17\alpha(H)20S$ and $20R$ diasteranes, respectively; 5, 6, and 8 are C_{28} diasteranes; C_{27} , C_{28} and C_{29} are regular steranes.

matched in essentially all biomarker characteristics. For example, the Phosphoria oil is characterized by: 1) a high proportion of tricyclic terpanes relative to the pentacyclic terpanes (with a C_{23}/C_{30} terpane ratio = 2.1 versus 0.04 for Castle Gardens oil, Figure 6), 2) a dominance of the C_{27} re-regular steranes over the C_{28} and C_{29} regular steranes and 3) a very low proportion of diasterane peaks 1 and 2 (Figure 7).

The Cretaceous oil type, however, is not so easily discounted because of similar biomarker features - such as the overall terpane pattern as shown in Figure 6. Yet, upon close examination, distinct differences can be identified to establish incompatible genetic linkage.

TRICYCLIC TERPANE (m/z 191) FRAGMENTOGRAMS

A. Oils in Oil-Saturated Sandstones

B. Crude Oils

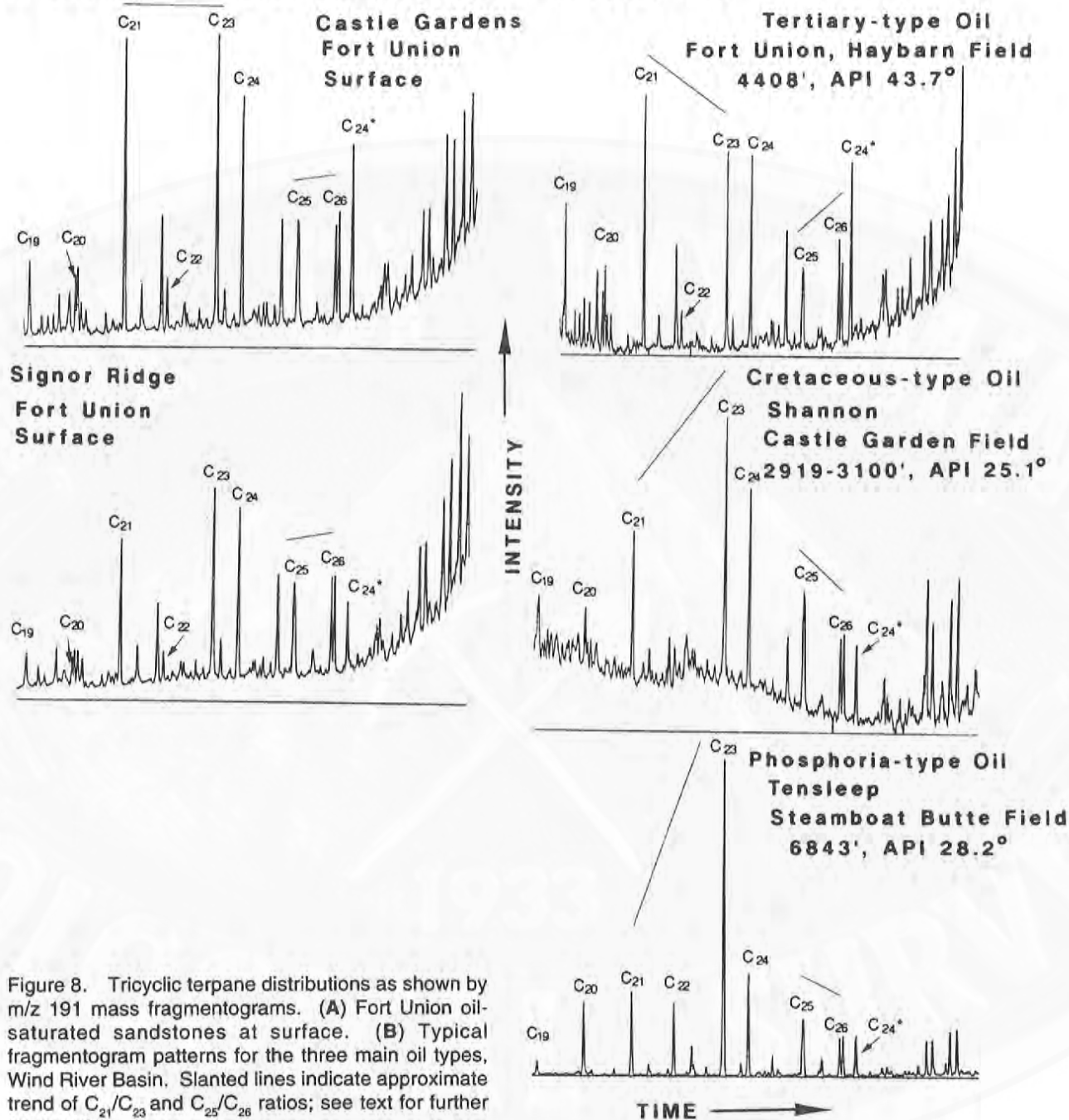


Figure 8. Tricyclic terpene distributions as shown by m/z 191 mass fragmentograms. (A) Fort Union oil-saturated sandstones at surface. (B) Typical fragmentogram patterns for the three main oil types, Wind River Basin. Slanted lines indicate approximate trend of C_{21}/C_{23} and C_{25}/C_{26} ratios; see text for further comments.

One of the most striking biomarker differences observed is in the sterane distribution (Figure 7). The Cretaceous oils are characterized by an unusually high abundance of C_{28} diasteranes labeled as 5, 6, and 8. In sharp contrast, the Castle Gardens oil exhibits the commonly observed low abundance of the 5, 6, and 8 diasteranes. The increase in relative amounts of these

diasteranes is not a function of increased thermal maturation but is attributed to the original source facies from which these Cretaceous oils were derived.

Other differences inherent in the Cretaceous oils that set them apart from the Castle Gardens oil are as follows: 1) $C_{21} < C_{23}$ tricyclic terpanes and $C_{25} > C_{26}$ tricy-

clic terpanes, as indicated by the slanted lines over the specific peaks (Figure 8) and 2) the C_{29} steranes are the least abundant of the regular steranes (Figure 7). From the combination of all the above biomarker traits, we can say with confidence that the Castle Gardens oil does not belong to the Cretaceous oil family and, therefore, is not derived from Cretaceous source rocks.

The Tertiary type oils, on the other hand, as exemplified by the Fort Union oil in the Haybarn field, have biomarker features that are essentially identical to those of the Castle Gardens biodegraded oil (Figures 6, 7, and 8). Any minor variations in peak abundances in the biodegraded oil are attributed mainly to mild biodegradation and/or migration effects. A biomarker crossplot (one of many that could be given) that illustrates the very positive correlation between the Castle Garden oil and the typical Tertiary oil is shown in Figure 9. Consequently, both the Castle Gardens oil and its genetically related counterpart, the Signor Ridge oil, are strongly believed to belong to the Tertiary family of oils. The Tertiary oils, in turn, are believed to have been derived principally from source rocks of the lacustrine Waltman Shale Member of the Fort Union Formation (see below).

Organic-rich shales of the Waltman Shale Member are designated as the prime source rocks based on both geological and organic geochemical evidence.

Geologically, Waltman shales, as much as 3,000 ft (914 m) thick, grade or interfinger with fluvial and delta-front sandstones of the Shotgun Member or equivalent undifferentiated units of the upper part of the Fort Union Formation (Figure 3, this report; Keefer, 1961, 1965; Robertson, 1984; and Katz and Liro, 1993). The juxtaposition of source and reservoir rock should provide ideal conduits for migrating generated hydrocarbons into structural and stratigraphic traps.

Geochemically, Waltman shales are rich enough in organic matter to generate and expel commercial amounts of petroleum. Total organic carbon (TOC) contents range up to 6.2 wt. % and average 2.7 wt. % and Rock-Eval genetic potentials (S_1+S_2) range from 0.30 to 24.04 mg HC/g rock (Katz and Liro, 1993).

Qualitatively, because certain compositional traits unique to a non-marine (lacustrine) depositional environment are present only in the Tertiary oils of the Wind River Basin, the lacustrine Waltman Shale Member appears to be the only viable source rock that could yield these compositional characteristics. One of these traits is the generation of high paraffin wax, high pour point oils (Tissot and Welte, 1984; Katz and Liro, 1993). Katz and Liro (1993) showed that a pyrolysis-gas chromatographic analysis of a Waltman shale sample yielded a hydrocarbon distribution characterized by an abundance of waxy components ($>n-C_{20}$) and low amounts of naphthenes, comparable to the distribution shown for the typical Tertiary oil in Figure 4. Relatively low abundances of naphthenes are also typical of "nonmarine organic matter derived from lacustrine algae or leaf cuticle, spores and pollen" (Katz and Liro, 1993, p. 168). In contrast, oils derived from marine organic matter are generally enriched in naphthenes relative to the adjoining n-paraffins as illustrated in the typical marine-derived Phosphoria- and Cretaceous-type oils in Figure 4.

Another trait common to lacustrine-derived oils is the relatively higher proportion of hopanes (bacterially derived) to steranes (algae derived) (Moldowan and

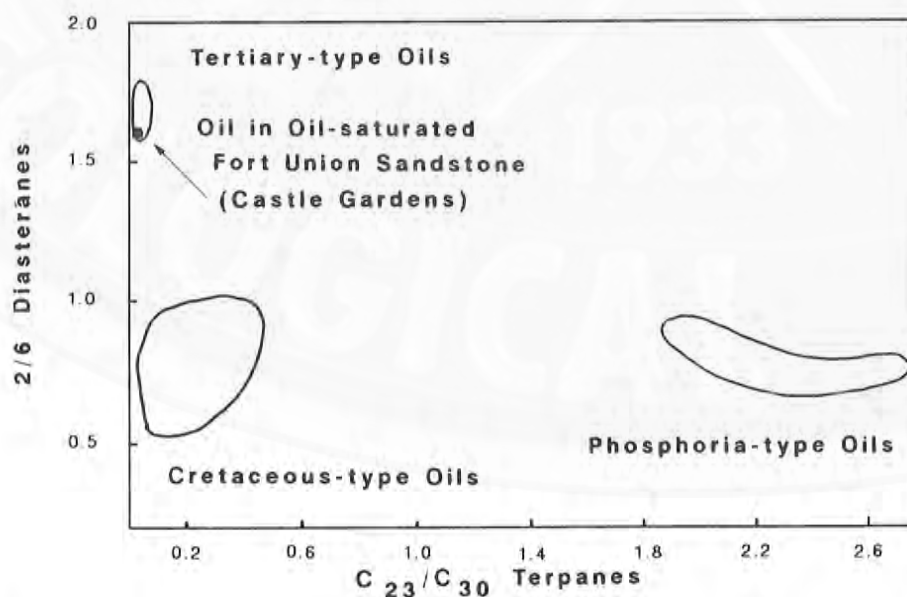


Figure 9. Cross plot of 2/6 diasteranes (C_{27} 13 β (H) 17 α (H) 20R diasterane/ C_{28} diasterane) versus C_{23}/C_{30} terpanes (C_{23} tricyclic terpene/hopane). Peaks 2 and 6 are indicated in Figure 7 and C_{23} and C_{30} in Figure 6. Note that oil impregnating Fort Union sandstones at Castle Gardens unequivocally belongs to the Tertiary family of oils.

others, 1985; Sofer, in press). Depicting this relationship, the Tertiary oils exhibited C_{30} hopane/ C_{29} (20R) sterane ratios of ~26.0-56.0, whereas, the predominantly marine Cretaceous oils manifested ratios of ~5.0-8.5 (Palacas, unpublished data).

Two other features that are quite diagnostic of lacustrine-derived oils are the relationships of the

tricyclic components C_{25} to C_{26} and C_{21} to C_{23} . Most lacustrine oils are characterized by C_{25}/C_{26} ratios <1 and C_{21}/C_{23} ratios >0.7 (Sofer, in press). Such ratios are clearly evident in the Tertiary-type oils (Figure 8). Conversely, most marine oils exhibit C_{25}/C_{26} ratios >1 and C_{21}/C_{23} ratios <0.7. As expected, the Cretaceous oils exhibit the diagnostic marine-related tricyclic ratios.

Summary and conclusions

Standard organic geochemical parameters (bulk chemical composition, gas chromatography of saturated hydrocarbon fractions, and carbon isotope ratios) and computerized gas chromatographic-mass spectrometric biomarker analysis were used to characterize and type the two oils in the oil-saturated Fort Union Formation sandstones at Signor Ridge and Castle Gardens. Because of severe bacterial degradation, the standard parameters proved ineffective in typing the oils. Biomarker distribution patterns, however, showed very positive correlation to the Tertiary family of oils. Cretaceous-type oils and upper Paleozoic (Phosphoria)-type oils were rejected as having any possible genetic linkage to the biodegraded oils based on clear-cut biomarker mismatches.

Geological and organic geochemical evidences also indicate that the Paleocene Waltman Shale Member, deposited in a lacustrine environment, is the primary source rock for the lacustrine-type Tertiary oils and consequently for the Castle Gardens and Signor Ridge oils.

If indeed the Castle Gardens/Signor Ridge oils were derived from these lacustrine shales, it marks the farthest extent of migration of oils derived from mature Waltman shales which are located to the north close to or in the depocenter of the Wind River Basin. The extent of lateral migration is of the order of 20 mi (32 km) or more—a distance consistent with similar findings of Katz and Liro (1993). The direction of probable migration pathways is indicated in Figure 1. The subsurface depth of the organically rich, thermally mature source shales range from 6,900-9,000 ft (2100-2740 m)—a depth interval determined to be the main phase of hydrocarbon generation and expulsion, commonly referred to as the "hydrocarbon kitchen" (Demaison, 1984; Katz and Liro, 1993; Nuccio and Finn, this volume).

Of major significance is the fact that the oil staining in the sandstones, particularly at Signor Ridge, is not merely a localized, pinpoint phenomenon but rather pervasive throughout an appreciable areal extent. According to R.D. Robertson (personal communication), based on areal surveying and subsurface control, these oil-saturated sandstones, located on the northern end of the northwest plunging Muskrat anticline, occupy an area of about 3,000 acres and are characterized by an average porosity of ~20%, an average water saturation of ~50% and a composite thickness of 50 ft (15 m) of oil-saturated sandstones. In our limited studies, however, we found it difficult to determine with confidence the total combined thickness of oil-saturated sandstones because of insufficient surface and subsurface data. Yet, from the available data, we estimate that there are between 10-20 ft (3-6 m) of oil-saturated sandstones (assuming 100 percent oil saturation). Using 10 ft (3 m) as the thickness value multiplied by the other factors listed above and by 7,758 (number of barrels/acre foot), the resulting volume of oil saturation would be about 23 million barrels of oil. For a 20-ft (6 m) thickness, the volume would be about 47 million bbls of oil. Whatever is the more realistic value, certainly in the past before biodegradation, water washing, and flushing of the sands by groundwater movement, there could easily have been double that amount.

Recognizing that oil generated and expelled from the Waltman Shale Member can migrate laterally long distances even to the periphery of former Lake Waltman (see Figure 1) and can charge, in significant volumes, fluvial and perhaps delta-front and shoreline sandstones, should open up new potential areas for oil exploration.

Acknowledgments

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Characteristics of oil-saturated fluvial channel sandstone reservoirs, Paleocene upper Fort Union Formation, south-central Wind River Basin, Wyoming

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Introduction

Natural gas and limited quantities of oil have been produced from Paleocene Fort Union Formation in the Wind River Basin since 1959 (i.e., Cooper Reservoir). Although Upper Cretaceous and Tertiary rocks in the basin are typically gas prone, Fuller Reservoir and

Haybarn fields yield significant quantities of oil from Tertiary reservoir rocks (Robertson, 1984). Current production is concentrated in the Fuller, Haybarn, Greater Madden, Pavillion, and Muddy Ridge fields (Figure 1, and Wyoming Oil and Gas Conservation

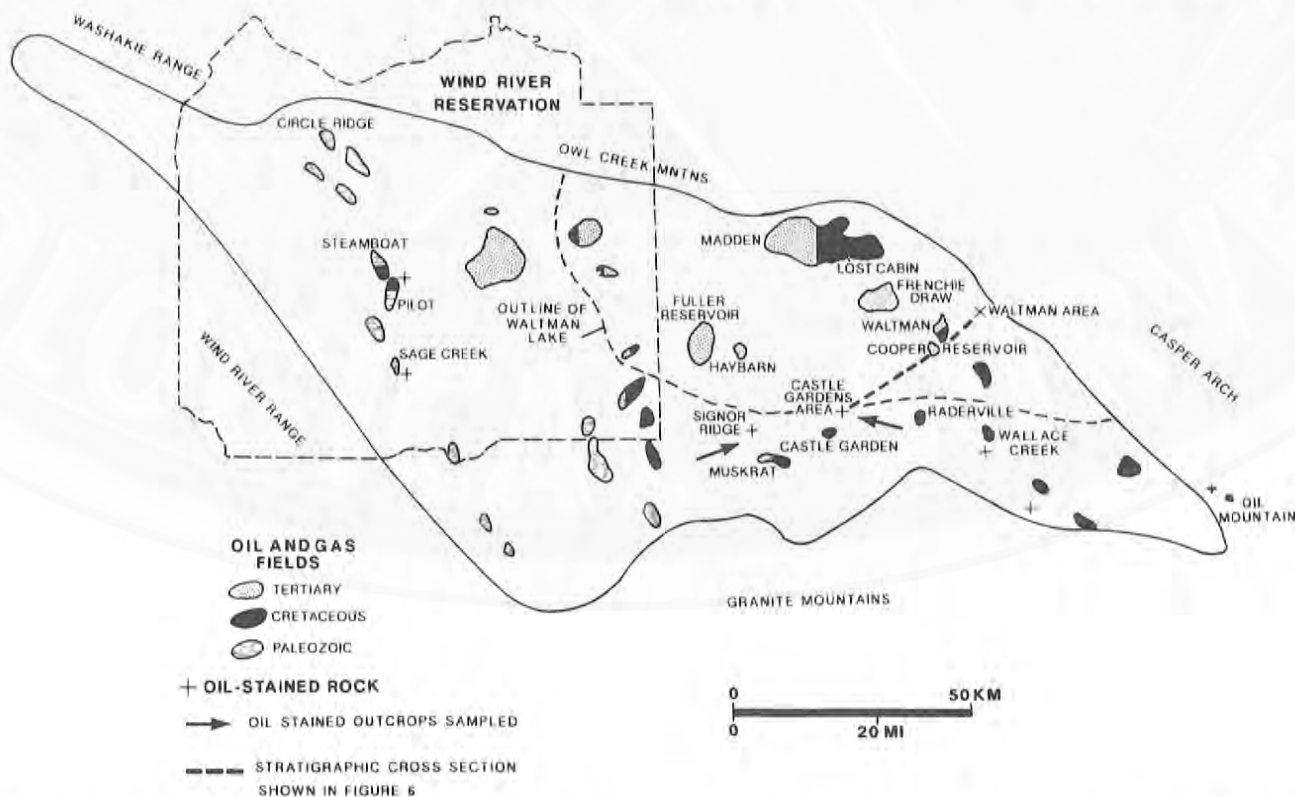


Figure 1. Index map showing general location of oil-saturated fluvial channel sandstones (arrows) discussed. For a detailed route map, see Flores and others, this volume.

Commission, 1992). Fort Union reservoirs are typically reported to be lenticular sandstones and produce from depths between approximately 3,000 and 9,600 ft (915 to 2,900 m). Depositional environments and stratigraphic relationships of the Paleocene Fort Union Formation are described in detail by Keefer (1965), Robertson (1984), Flores and Keighin (1992), Flores and others (1993), and Nichols and Flores (1993); a simplified stratigraphic section is shown in Figure 2.

Porosity observed in thin sections of the Fort Union oil-stained sandstones is high (20-25%), and due primarily to secondary leaching. Porosity and permeability of other Fort Union reservoir sandstones collected from producing fields varies widely due to variations in grain size and sorting, and highly irregular distribution of carbonate cements (Keighin and Flores, 1993).

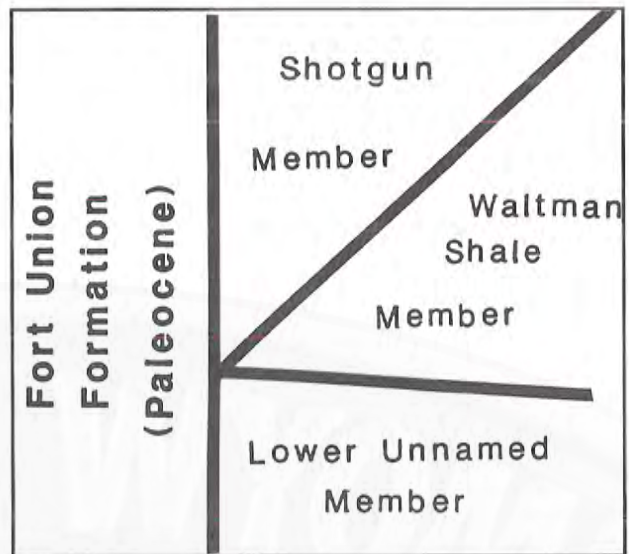


Figure 2. General stratigraphic section for strata in Signor Ridge and Castle Gardens areas, Wind River Basin.

Reservoir sandstones

During detailed investigations of outcrops of Paleocene upper Fort Union strata in the Wind River Basin, two heavily oil-stained sandstones (Figure 1)

were examined; views of these outcrops are shown in Figures 3-5. Architecture of the reservoir rocks is complex (Figures 4 and 6), and varies from multiple



Figure 3. Outcrop of Paleocene Fort Union Formation along Signor Ridge (se nw sec. 17, T. 34 N., R. 92 W., Fremont County, Wyoming). Outcrop extends nw-se for approximately 2,000 ft (610 m), and illustrates complex reservoir architecture resulting from multiple depositional environments. Arrow marks approximate location of a largely covered coaly interval dated as late Paleocene (P5-P6 palynostratigraphic zones of Nichols and Ott, 1978; verbal communication, 1993). Details of some outcrops in this sequence are shown in Figure 4A-C.

stacked, multiscoured conglomeratic sandstones representing deposits of meandering and braided streams to multilateral, multiscoured sandstones formed by meandering streams. Oil staining occurs in very fine to coarse sand-size and gravel-size material deposited in fluvial environments; oil is now trapped in interstitial clays. The source of the oil saturating the outcrops at Castle Gardens area and at Signor Ridge is geochemically similar to other Paleocene oils in the Wind River Basin (Palacas and others, this volume). Seals are formed by clay drapes within the sandstone reservoirs and interbedded shales and mudstones between the sandstone reservoirs. Thickness and lateral continuity of potential Fort Union sandstone reservoir rocks is highly variable. Individual sandstone units within the sandstone reservoir complex range in thickness from <3.3 ft (<1 m) to 6.6-9.8 ft (2-3 m). Although sandstone packages may be thicker, the sandstones are interrupted by mudstones and shales. In addition, texture, internal scours, and types and arrangement of sedimentary structures often interrupt the continuity of the individual units. Lateral extent of the individual sandstone units varies from a few hundred feet to as much as 1.5 mi (2.5 km). At the Castle Gardens area, Fort Union strata are exposed within a zone about 2 mi (3.2 km) wide and 6 mi (10 km) long.

Detailed measurement of many closely-spaced stratigraphic sections in the Castle Gardens area clearly shows the complex and usually discontinuous nature of sandstone in the middle part of the Fort Union Formation (Flores and others, 1992, Figure 12). Tabular crevasse splay sandstones may extend laterally for over 1 mi (1.6 km), but are rather thin (<15 ft or <5m). Lenticular fluvial-channel and (or) crevasse-splay sandstones may be up to 50 ft (15 m) thick, but are rarely traceable laterally for more than 0.3 mi (0.5 km). Oil-saturated sandstones were observed only in the uppermost part of the Fort Union Formation at Signor Ridge and in the Castle Gardens area (Figures 1 and 6).

Examination of thin sections from both outcrops shows the sandstones are very similar in petrographic characteristics. Both are dominantly composed of fine grained, subrounded to subangular detrital quartz grains, and abundant intergranular, oil-saturated clay minerals (Figure 7). No feldspar grains were observed in the samples from Castle Gardens, and less than one volume percent was seen in the Signor Ridge samples. Some of the potassium feldspar in these samples is authigenic (Figure 8). It is possible that some (or most) detrital feldspar was removed by leaching. De-



Figure 3 continued.



Figure 4. Series of outcrops observed stratigraphically above point marked by arrow in **Figure 3**; (A) is stratigraphically highest in this sequence. Coarse sandstone and conglomerate (arrows) deposited in a braided stream. (B). Conglomeratic lens (arrows) intertonguing with crossbedded, fine- to medium-grained sandstone; conglomeratic sandstones are irregularly distributed throughout this upper part in the Fort Union Formation. (C). Coarsening-upward crevasse splay (S) overlain by multistory, stacked sandstones separated by abundant mudstone drapes. Arrow indicates where a light gray surface has been removed, showing a dark, oil-stained sandstone; oil staining pervades this 2 to 4-ft (0.6 to 1.2-m)-thick section. **Figure 7A**, a thin section photomicrograph of an oil-stained sandstone in this outcrop, shows abundant porosity, in addition to oil.

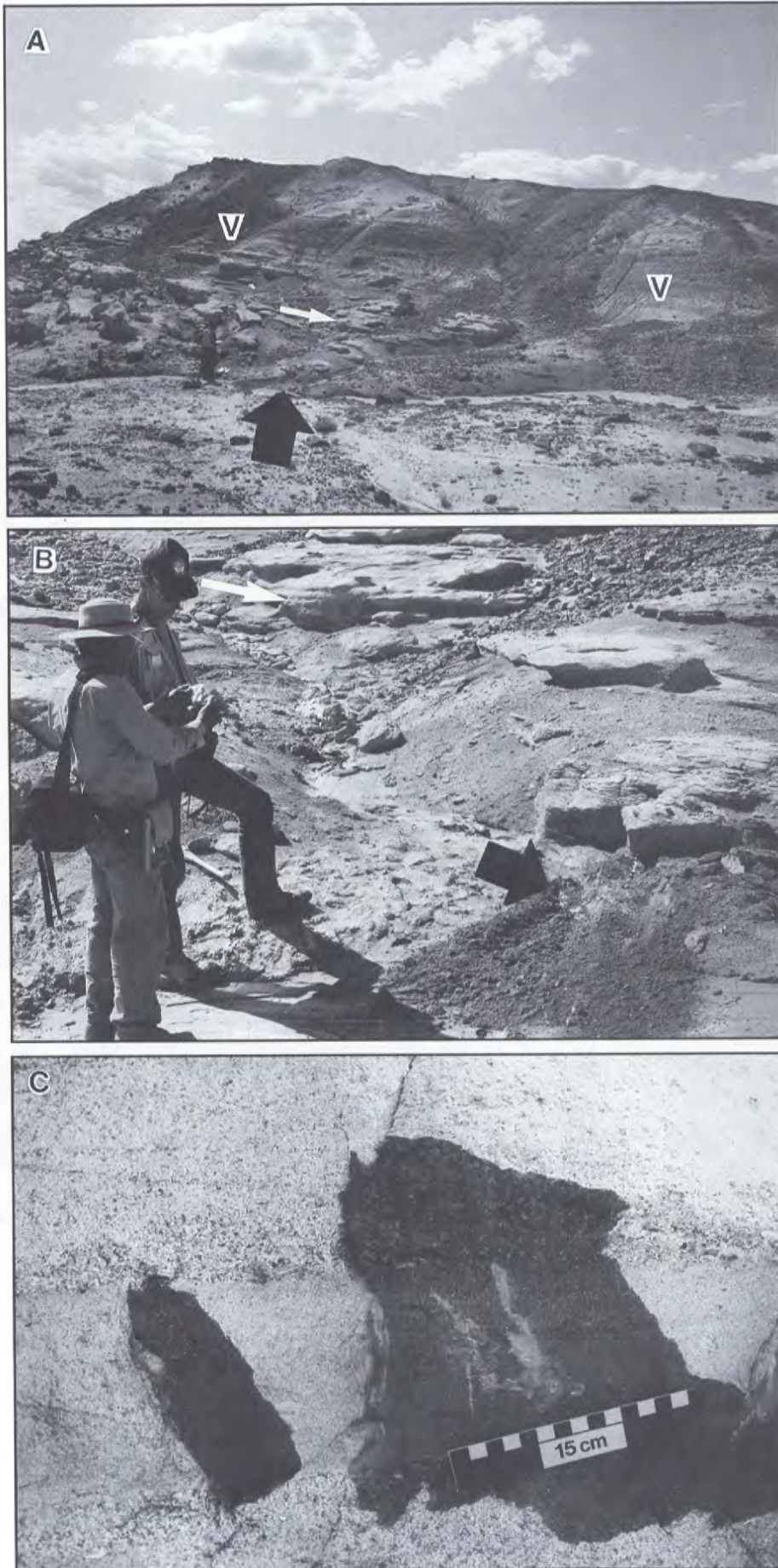


Figure 5. (A). Outcrop of channel sandstones and overlying mudstones in the uppermost part of the Fort Union Formation, Castle Gardens area (ne sw, sec. 17, T. 34 N., R. 90 W.). White arrow marks approximate sample location of oil-stained sandstone shown below (Figure 5C), and in thin section in Figure 7B; black arrow marks site sampled for identification of clay minerals (Thorez and others, in press). Variegated (V) clay-rich paleosol complex (Thorez and others, in press) acts as an upper seal for this reservoir complex. Approximately 50 ft (15 m) stratigraphically below the black arrow, a sample was determined to be late Paleocene age (P6 palynostratigraphic zone, Nichols, verbal communication, 1993). The Fort Union Formation produced natural gas from the now-abandoned Squaw Butte field, secs. 1, 26, 31, T. 36 N., R. 90 W. (B). Detail of photograph above. Oil-stained sandstones (white arrow) are typically <3 to 7 ft (<1 to 2 m) thick; a thin section photomicrograph of the sandstone is shown in Figure 7B. (C). Portion of oil-saturated sandstone illustrating pervasive nature of oil-saturation, and wide variation of grain size (fine to medium in center and coarse above and below).

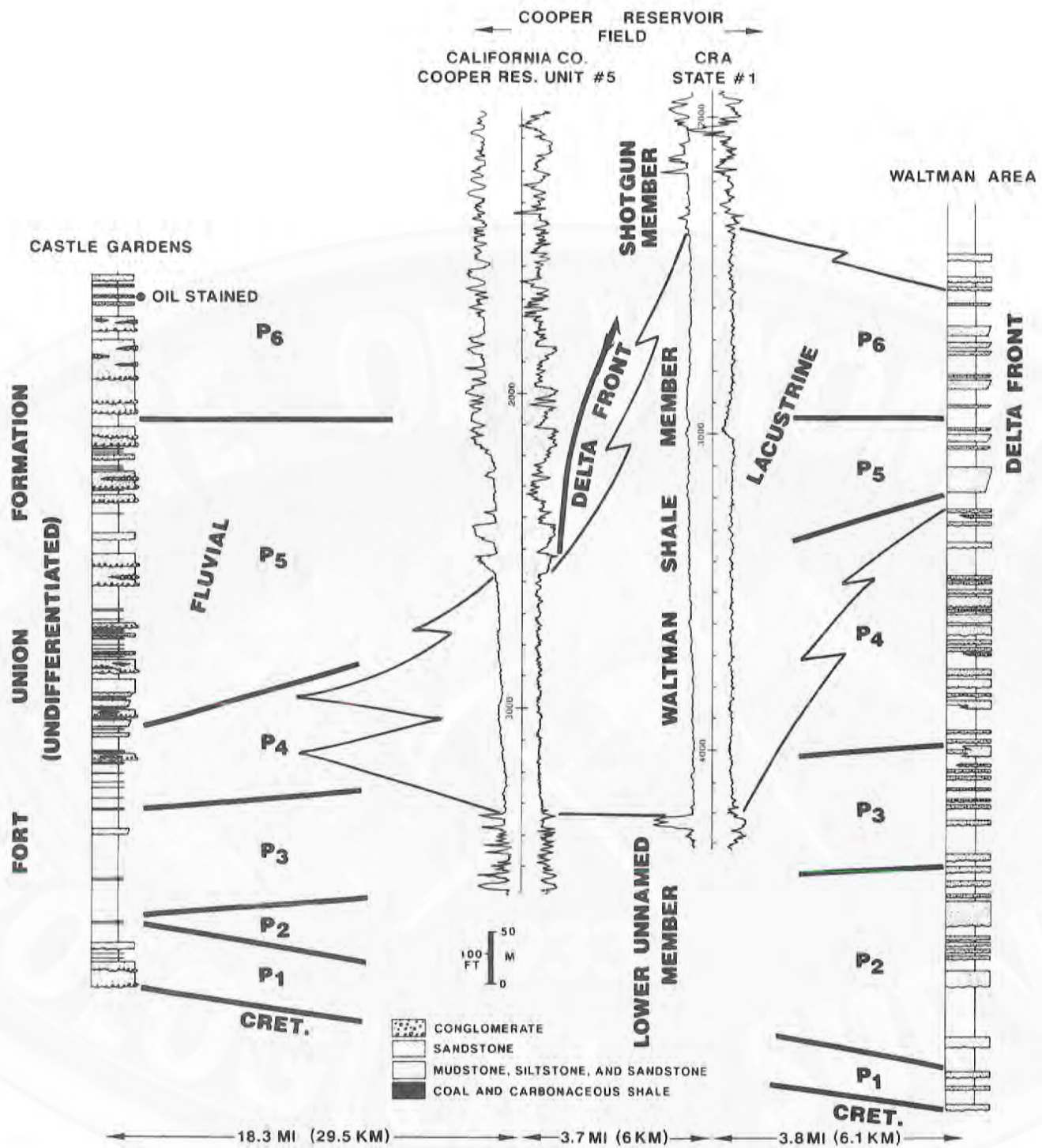


Figure 6. Stratigraphic cross section from Castle Gardens area to Waltman (see Figure 1) illustrating complex geometry and age relationships (P1 - P6 palynostratigraphic zones; Nichols and Ott, 1978) of potential sandstone reservoirs. The oil-stained upper portion of the Castle Gardens measured section is shown in Figure 6A; outcrops near Waltman will be visited on the field trip.

trital chert is seen in both sandstone reservoirs, and some grains contain micro porosity formed by secondary leaching. Compaction of framework grains is very minor, although most grains are embedded in the

abundant intergranular largely detrital, intergranular clay minerals. No carbonate cements are present in these oil-saturated samples; the abundant intergranular clays act as a cementing agent.

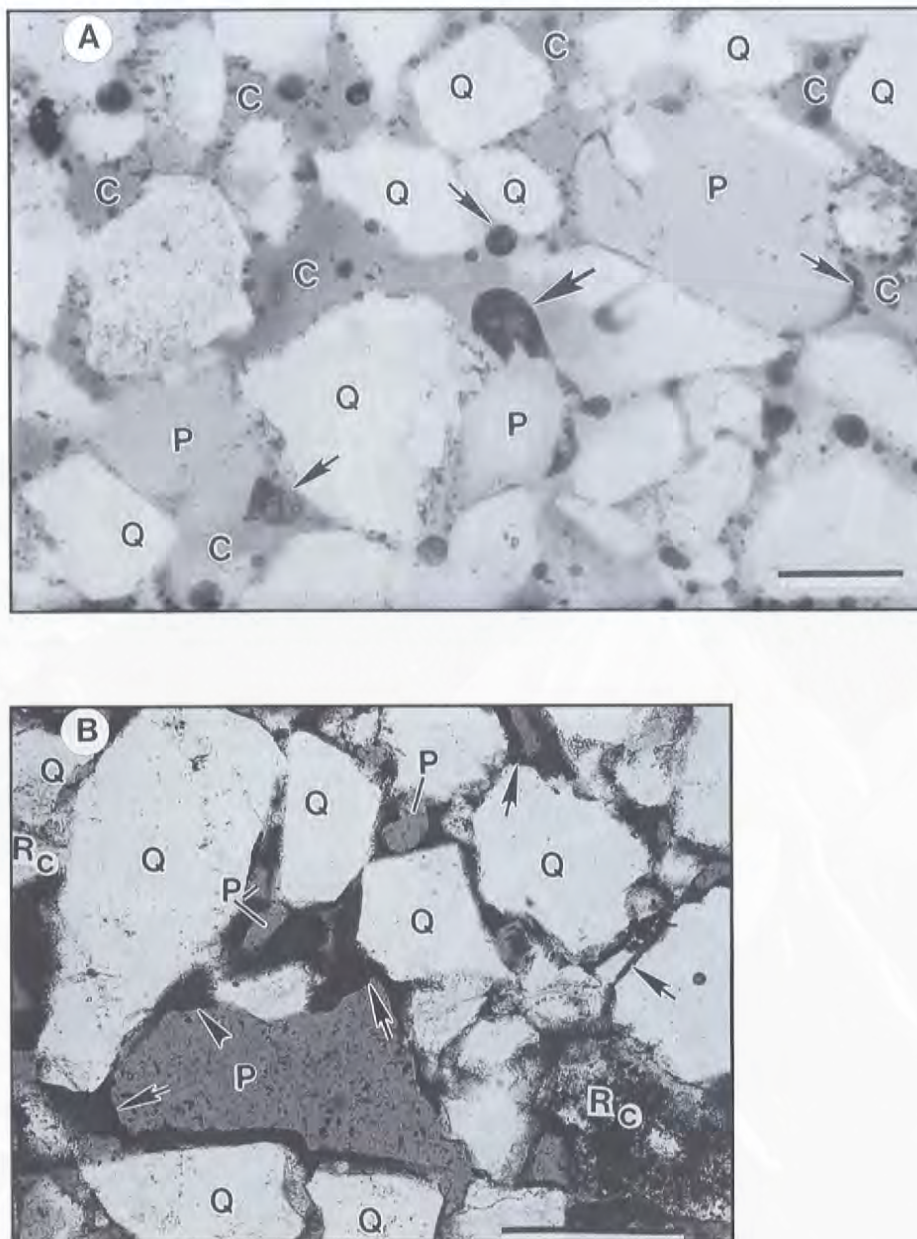


Figure 7. Thin section photomicrographs of outcrop samples of oil-stained sandstones from (A) Signor Ridge (sw sw ne sec. 17, T. 34 N., R. 92 W.), and (B) Castle Gardens area (ne sw sw sec. 28, T. 35 N., R. 90 W.). Both photomicrographs were taken with plain polarized light. (A). Sandstone is composed almost entirely of fine-grained angular to subangular detrital quartz grains (Q), oil-stained intergranular clay(s) (C), and open pores (P). Rock fragments, dominantly chert, occur, but none are visible in this view. Oil (arrows) migrated into the sample after formation or emplacement of intergranular clay(s), but before secondary leaching which formed now-open pore space (P). Globular nature of some oil may be due to mobilization by epoxy used in preparation of the thin section. Scale bar represents 100 mm. (B). Framework composition, grain size, and angularity of the sandstone from the Castle Gardens area are essentially the same as at Signor Ridge, although partially altered chert grains (Rc) are visible here. Oil staining (arrows) is limited to intergranular and grain-coating clays, but is pervasive throughout the thin section, as well as in the outcrop. Open pores (P) vary widely in size, and are primarily due to secondary leaching; most were formed after emplacement of oil which stains the clay. Scale bar represents 250 mm.

Case studies

Characteristics of reservoir sandstone in the Signor Ridge and Castle Gardens area

Because the depositional history of the potential reservoir sandstones in the Paleocene Fort Union Formation in the Wind River Basin is highly variable over short distances, numerous closely-spaced and detailed measured sections are necessary to clearly delineate reservoir geometry and characteristics, and to

correlate reservoir sandstones. Examination of outcrops at Signor Ridge indicates a complex of four reservoir types, including a coarsening-upward crevasse splay, multistory stacked sandstone deposited by meandering streams, separated by mudstone drapes, and multiscoured conglomeratic sandstones consisting of sand-dominated or gravel-dominated braid bars deposited by braided streams. Abundant mudstone drapes and interbedded mudstones act as local seals. Relatively thin (5-10 ft, or 1.5-3 m) fine grained, multiscoured channel-form sandstones deposited by

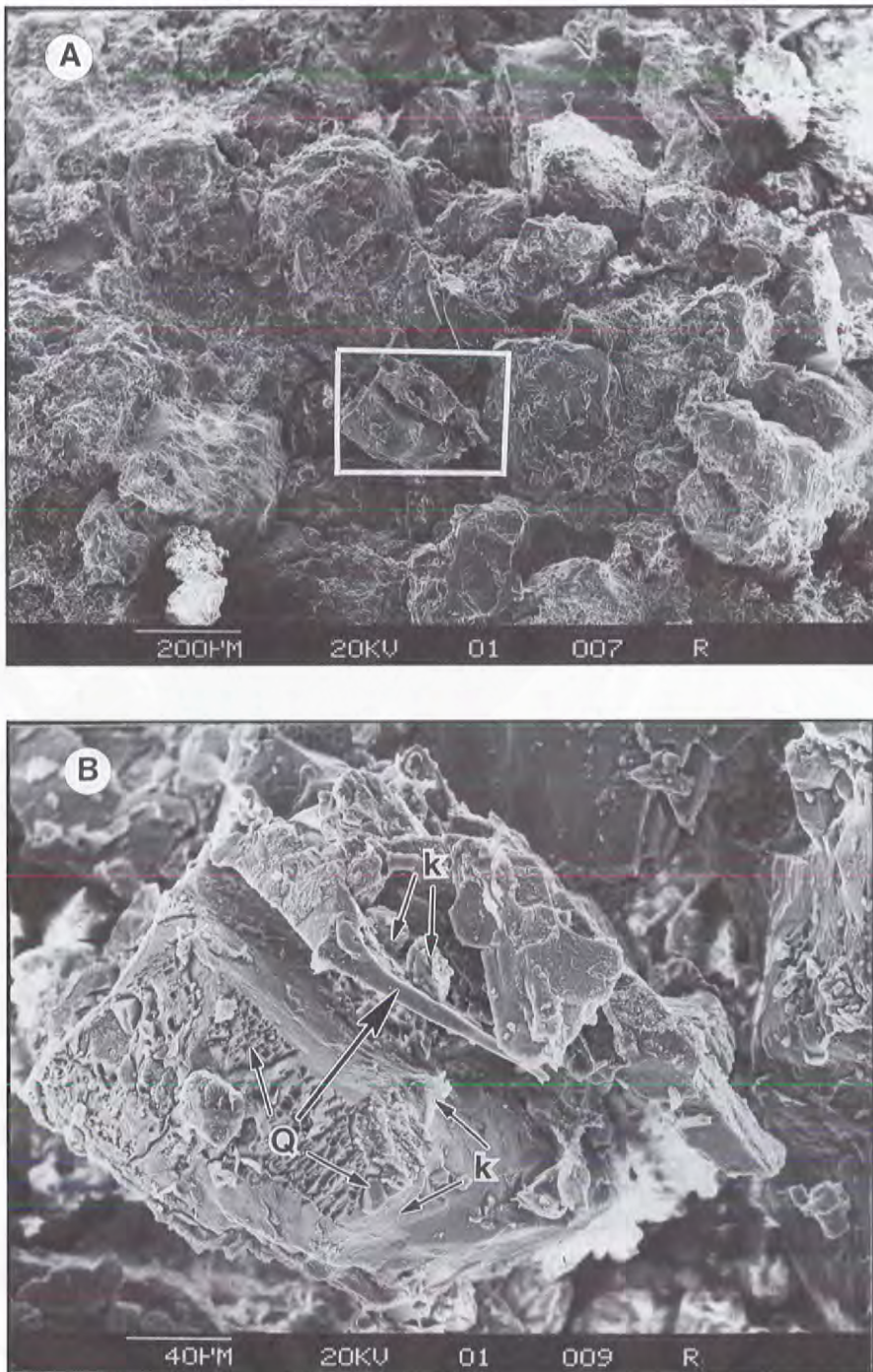


Figure 8. Scanning electron micrographs of outcrop samples of oil-stained sandstones from the Castle Gardens area (ne sw sw sec. 28, T. 35 N., R. 90 W.). (A). An upper fine- to medium-grained sandstone illustrating variable nature of intergranular porosity, due primarily to relative abundance of clay minerals. Some visible microporosity is due to leaching of rock fragments. Area outlined is enlarged in next figure. (B). Detrital quartz grain (Q) illustrating episodes of diagenesis which influence the reservoir properties of the sandstone. Quartz (Q), occurs as 1) detrital quartz grain (under the Q) which displays etch pits, 2) authigenic, euhedral crystal which formed on the surface of the detrital grain, and 3) as a partial overgrowth (larger arrow above Q) on a mineral grain which has been removed by leaching. Partially coating the detrital quartz grain is a thin (5-10 μm) coating of clay; EDS (Energy Dispersive Spectroscopy) suggests the clay is primarily kaolinite (k). A portion of the clay coat has started to peel away from the detrital grain (see arrow). Kaolinite also occurs in clusters of well-crystallized platelets within the pore formed by leaching of a detrital grain. (C). All mineral grains in this micrograph are coated with oil-stained (?) clay.

low sinuosity streams in the uppermost zone of the Fort Union Formation are oil-saturated in the Castle Gardens area. At this location, overlying variegated paleosol mudstones act as a seal. Oil from a Paleocene source impregnates intergranular clay(s) at both sandstone-reservoir sandstones. Secondary porosity, formed by leaching of framework grains is common. These open pores are, however, not filled by oil, indicating that oil impregnation of intergranular clays occurred before leaching selectively removed some

framework grains (possibly feldspar). Much of the porosity visible in thin sections is probably due to removal of intergranular clay, possibly during preparation of thin sections. The high content of intergranular clays could cause significant problems during drilling, completion, or production. Clays have caused well completion problems at Fuller Reservoir field, which produces from the Fort Union Formation (Pirner, 1978).

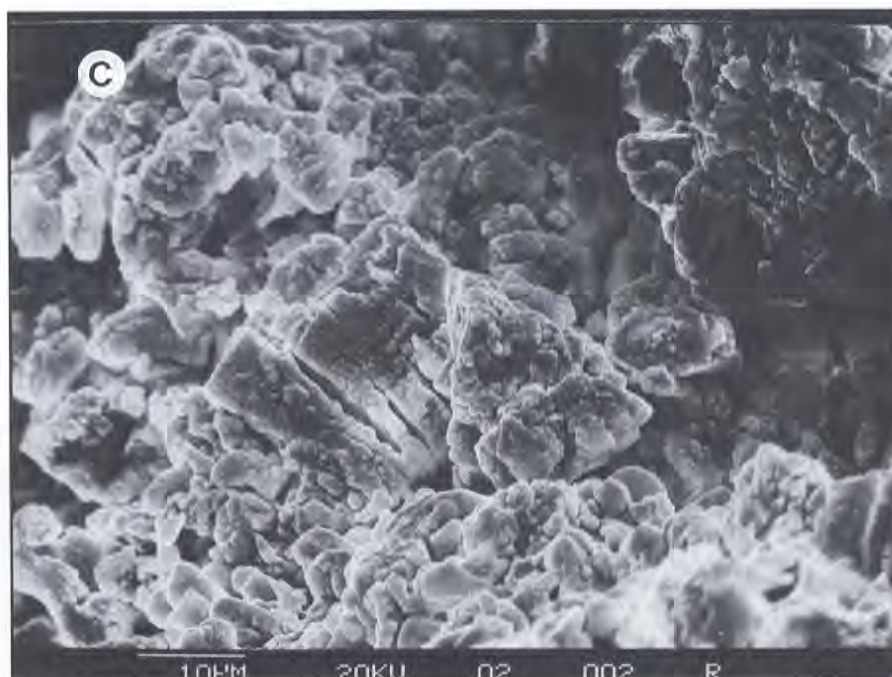


Figure 8 continued.

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Road log and general geology of the Bighorn Basin, Wyoming

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Introduction

Thermopolis has a population of 3,287 and the town relies heavily on tourism as a source of income. The town's other industries include agriculture (ranching), oil and gas, and timber. The town is strategically situated north of the Wind River Canyon and numerous hot springs are developed within Thermopolis; the therapeutic value of the thermal springs is the mainstay of tourism in this area. The Big Spring in Hot Springs State Park discharges water at rates to 2,900 gallons per minute and at temperatures of 135° F (43° C). The water is derived from recharge areas on the north flank of the Owl Creek Mountains, from where it percolates downslope, passes underground, is heated, and reemerges on the surface via fractures along the flank of the Thermopolis anticline (Figure 1; Blackstone, 1988).

Thermopolis is on the southernmost end of the Bighorn Basin, which is rimmed on all but the north end by mountain ranges. The basin is bounded by the Owl Creek Mountains to the south, the Absaroka Range to the west-southwest, the Beartooth Mountains on the northwest, and the Bighorn-Pryor Mountains on the east. The Absaroka Range is composed of Eocene volcanic rocks including lava flows (trachyandesite), intrusives (granodiorite, monzonite, and syenite), and volcanic-derived breccias, agglomerates, sandstones, siltstones, and ashes (Bergendahl, 1975; Pruss, 1975; Wil-

son, 1975). Volcanics of the Absaroka Range now bury remnants of the Washakie Range, a Paleocene highland that developed along the southwestern margin of the basin during the Laramide orogeny. The Beartooth Mountains are composed of a Precambrian basement core, which consists of quartzites, amphibolites, migmatites, schists, granitic gneisses, granodiorites, monzonites, and granitoids (Mueller and others, 1982; Lafrenz and others, 1986). The granites were dated to be 2,752 Ma (Lafrenz and others, 1986). The Beartooth Mountains are flanked by Paleozoic and Mesozoic sedimentary rocks. The Bighorn Mountains contain a Precambrian core of Early Archean, layered granitic gneiss, amphibolite, and felsic gneiss in the southern part. These rocks are intercalated with metagraywacke, quartzite, and other metasedimentary rocks. The northern part of the Bighorn Mountains core consists of Middle Archean diorites and monzonites, which have been intruded by amphibolites (Love and Christiansen, 1985). The mountains are flanked by Paleozoic and Mesozoic sedimentary rocks.

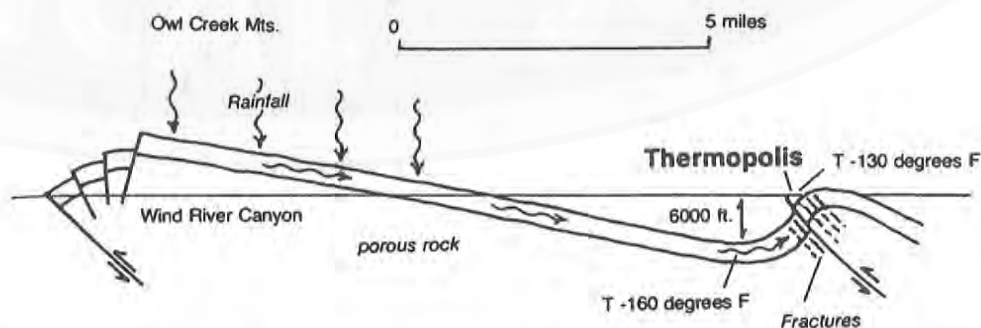


Figure 1. Groundwater flow and heating system of the hot springs in Thermopolis. Adapted from Lageson and Spearing (1988).

Road log

Mileage:		Description:
Increment	Total	
0.0	0.0	Begin Day 2. Road log begins at the junction of U.S. Highway 20 and State Highway 120 in central Thermopolis. See Figure 2 for field trip route in the Bighorn Basin. Proceed west on Highway 120 (the Thermopolis-Meeteetse road).
2.0	2.0	The highway follows the southwest limb of the Thermopolis anticline and passes steeply-dipping outcrops of the Upper Cretaceous Cody Shale and Frontier Formation, the Lower Cretaceous Cloverly Formation, the Upper Jurassic Morrison Formation and Middle and Upper Sundance Formation, and the bright-red, Triassic Chugwater Group on the northeast side of the highway (Lageson and Spearing, 1988). Resistant sandstone beds form the pinyon-covered ridges or hogbacks. A generalized stratigraphic column of the Bighorn Basin is shown in Figure 3.
4.6	6.6	Highway crosses Owl Creek.
9.0	15.6	Large concretions in sandstone in the Frontier Formation are visible on the north or right side of the highway.
1.2	16.8	Cody Shale on both sides of the highway.
4.3	21.1	Stop 1- Orientation and overview near junction with Hamilton Dome road. To the east-southeast side of the highway are low-lying cliffs overlying the Upper Cretaceous Cody Shale (marine). The cliffs are formed in strata of the Upper Cretaceous Mesaverde Formation, and include massive sandstones and coal-bearing rocks deposited in shoreface, coastal plain, and fluvial environments. The Mesaverde Formation is overlain by fluvial and coal-bearing rocks of the Upper Cretaceous Meeteetse and Lance Formations. The Paleocene Fort Union Formation, which was deposited in fluvial environments, overlies the Lance Formation (Hewett, 1926; Lageson and Spearing, 1988). To the southwest, the Hamilton Dome oil field produces from the Lower Cretaceous Muddy Sandstone, the Lower Permian Phosphoria Formation, the Middle and Upper Pennsylvanian Tensleep Sandstone and Upper Mississippian to Middle Pennsylvanian Amsden Formation, the Mississippian Madison Limestone, and the Upper Ordovician Bighorn Dolomite. Cumulative production (to 1988) is about 227,329,824 BO (Wyoming Geological Association, 1989). This structure is a doubly-plunging anticline in which the crest is composed of the Lower Cretaceous Thermopolis Shale flanked by the Lower Cretaceous Mowry Shale, and Upper Cretaceous Frontier Formation and Cody Shale. To the northwest is Ilo Ridge, a syncline underlain by the Lance Formation and capped by the Fort Union Formation. The contact between these formations is marked by a prominent basal conglomerate unit. The Grass Creek coal mine is about 8.5 mi (13 km) northwest of Ilo Ridge.
0.2	21.3	Crossing Cottonwood Creek.
2.8	24.1	Crossing the estimated contact between the Upper Cretaceous Lance and Paleocene Fort Union Formations.
2.1	26.2	Road junction of State Highways 120 and 171. Turn west on State Highway 171.
0.5	26.7	Outcrops of the Cretaceous Lance Formation are visible on the south (left) side of the road.
0.9	27.6	Crossing the estimated contact between the Cretaceous Lance and Meeteetse Formations.
0.5	28.1	Coal-bearing rocks of the Meeteetse Formation are on the left. At this point, we are crossing the approximate west-northwest trending axis of the Grass Creek dome (Hewett, 1926) to the right. Structurally, the basin is a dome that is part of a doubly plunging anticline cored by Cretaceous Cody Shale and flanked by the Upper Cretaceous Mesaverde, Meeteetse, and Lance Formations (Figure 4). The Grass Creek oil field is localized within the dome and produces from the Upper Cretaceous Frontier Formation, Lower Cretaceous Muddy Sandstone, the Upper Triassic Crow Mountain Formation (Chugwater Group), the Lower Permian Phosphoria Formation, and the Pennsylvanian Tensleep Sandstone. Cumulative production up to 1989 was about 238,746,000 BO (Wyoming Geological Association, 1989).

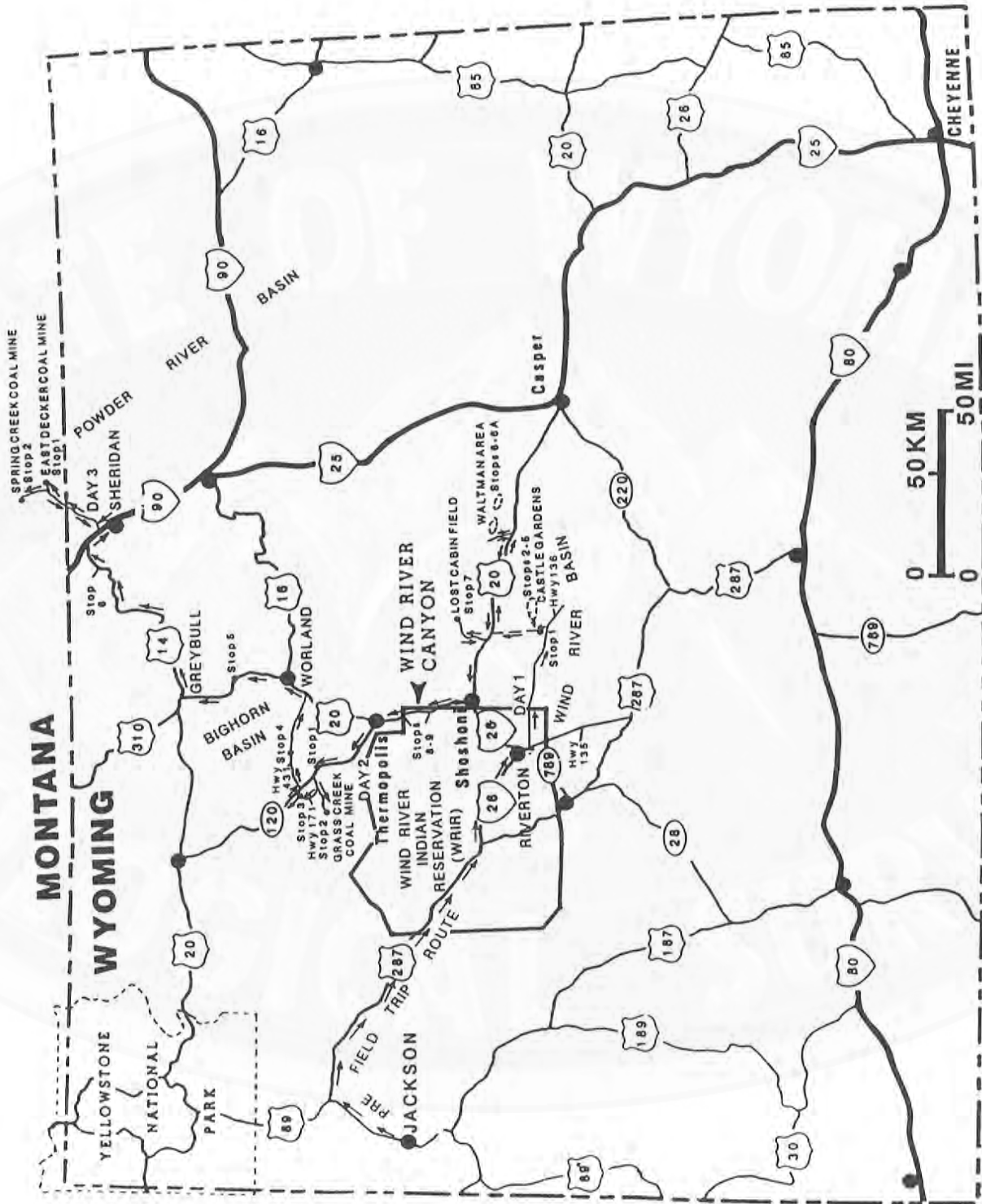


Figure 2. Map showing pre-field trip route and daily routes for the field trips in the Wind River Basin (Day one), Bighorn Basin (Day two), and northwestern Powder River Basin (Day three).

		QUAT	Recent & Pleistocene	Gravel & Terraces	Approx. thickness	General outcrop description	Main outcrop locality	
CENOZOIC	Tertiary	Eocene	Absaroka igneous flows and sediments			Igneous flow rocks, volcanic-derived sediments, dikes, sills, plugs.	Absaroka Plateau, west of State highway 120.	
			Wiggins	0-4000'	Gray, white, volcanic-derived shale, sandstone, conglomerate, breccia. Plugs, dikes, and sills prevalent.			
			Tatman	0-600'	Drab sandstone and shale. May be part volcanic-derived. Remnants form highest topography in central part of the basin.	East of State 120 near Gooseberry Cutoff. South of U.S. 20-14 (Tatman Mt.).		
			Willwood	0-8000'±	Brightly colored (mostly reds) shale, claystone, sandstone (conglomeratic in part) on west flank, grading to pastel to drab colored on east flank. Forms badlands.			
		Paleocene	Fort Union	0-8000'±	Drab sandstone, mudstone, and coal beds. Some fossil leaves. General tan appearance on outcrop. Coal mined at Bear Creek and Red Lodge, Montana.	Central basin area and on flanks of basin near most major highways.		
			Cretaceous	Upper	Lance	700-1200'	Sandstone, shale, and few coals. Forms resistant ridges.	Outcrop parallels many basin roads.
					Meeteetse-Lewis-Bearpaw	700-1500'	Massive white sandstones, and many thin shales and coals. Generally less resistant than formations above or below.	State 120, near Meeteetse, Wyoming (type section).
					Mesaverde Group	1500'±	Sandstone, shale, coal. Forms resistant ridges. Outlines many anticlines in the basin.	Most major highways are near outcrops in less resistant shale valleys.
					Cody	1400-2100'	Gray, black shale. Few thin sandstones. Type section in Shoshone River cut west of Cody, Wyo. 2120' measured. Microfossils and megafossils abundant. Forms valley and flat areas.	Near many highways. Good exposure at type section locality.
Niobrara?								
MESOZOIC	Lower	Frontier	500-1000'	Forms an alternating series of sandstone ridges and shale valleys.	Most major highways rimming the basin.			
		Mowry	370'	Forms rounded steep slopes with some resistant, siliceous bands. Silver-gray color on outcrop.				
		Thermopolis	700'	Dark gray to black shale. Muddy Sandstone member, near center of section, generally poorly resistant.	Type section near Thermopolis, Wyo.			
		Cloverly Group	300'	Thin sandstones and shales, tan to brown, sideritic. Some conglomerate lenses in lower part.	Near most major highways around the basin.			
		Morrison	200'	Shales, sub-waxy, and some sandstone and conglomerate. White and light red appearance on outcrop.				
		Sundance	200'	Glaucconitic sandstone, thin limes and shales. Greenish appearance on outcrop.				
		Gypsum Springs	275'	Gypsum, shale and thin limestone. Light red, pink, and white on outcrop.				
		Triassic	Chugwater	500-1200'	Red, maroon sandstone and shale. Pale red Dinwoody at base.	U.S. 20-State 789. Good section in road cut south of Thermopolis.		
			Dinwoody	50'	Gray. Forms resistant dipslopes.			
		PALEOZOIC	Permian	Phosphoria	50-300'	Cliff-forming, cross-bedded sandstone.	U.S. 20-State 789 in the Wind River Canyon, south of Thermopolis, Wyo.	
Tensleep	50-300'			Dolomite, red shale, sandstone. Darwin Sandstone at base.				
Pennsylvanian	Amsden		300'	Cliff-forming. Light gray limestone group.				
	Madison Group		330-900'	Shale and dolomite.	U.S. 20-14 West of Cody, Wyo.			
Mississippian	Three Forks? Jefferson?		0-400'	Massive, tan, cliff-forming dolomite.				
	Bighorn		0-500'	Poorly resistant limestones, sandstones, and shales. Generally held up by the overlying resistant Bighorn Dolomite.	U.S. 20-789 Wind River Canyon.			
Devonian	Gallatin		1000-1200'					
	Gros Ventre							
Cambrian	Flathead							
	PRECAM.							

Figure 3. A composite stratigraphic column from Precambrian to Cenozoic rocks in the Bighorn Basin. Modified and adapted from deVries and Crockett (1990).



Figure 4. Geologic map showing the structural geology of the Grass Creek anticline and outlying areas. Adapted from Hewett (1926).

Mileage:		Description:
Incre- ment	Total	
1.0	29.1	Outcrops of the coal-bearing Cretaceous Mesaverde Formation are visible on the left.
1.2	30.3	Crossing the approximate contact of the Mesaverde Formation and Cody Shale.
1.5	31.8	Road junction. Turn left. On the skyline to the southeast is Ilo Ridge, which is an east-west syncline. The basal conglomerate and lower part of the Paleocene Fort Union Formation are well exposed along the flanks (Figure 4). Ilo Ridge is capped by an upper conglomeratic succession about 394 ft (120 m) thick (Roberts and others, 1994; Roberts and Bossiroy, in press).
1.0	32.8	Outcrops of the Mesaverde Formation on the ridge to the south (left).
1.5	34.3	Ridge on the north (right) is underlain by the Cretaceous Mesaverde and Meeteetse Formations. The ridge on the south (left) is underlain by the Cretaceous Lance Formation and Paleocene Fort Union Formation.
1.9	36.2	Crossing the approximate contact of the Cretaceous Lance Formation and Paleocene Fort Union Formation. The pit on the right is for extracting gravel from the basal conglomerate of the Fort Union Formation; the gravel is used locally for road ballast. The conglomerate and overlying deposits are on the northeast limb of the Grass Creek syncline (Hewett, 1926; Stewart, 1975), which trends northwest-southeast, and roughly parallels the Grass Creek anticline (see Figure 4). The road follows this limb (southwest side) for about 1 mi (1.6 km) into the axis of the syncline in the vicinity of the Grass Creek coal mine.
0.5	36.7	Stop 2- Grass Creek Coal Mine (Western Energy Company). The Grass Creek coal mine, which is the only remaining active coal mine in the Bighorn Basin, consists of small, L-shaped strip pits with long dimensions oriented north-south and east-west (Roberts and others, 1994). The coal beds in the mine include, in ascending order, the Mayfield, unnamed, and rider 1 and 2 coal beds; these beds are present in a zone about 98 ft (30 m) thick (Figure 5; see Roberts and Stanton, this volume). This coal zone, informally designat-ed as the Mayfield coal zone, is about 575 ft (175 m) above the top of the basal conglomerate of the Fort Union Formation. D.J. Nichols (written communication, 1993) places the Mayfield coal bed within the P3 biozone (Nichols and Ott, 1978), or lower middle Paleocene. Based on analyses of ash yields and maceral data, the Mayfield and related coal beds are interpreted to have developed in low-lying, topogenous mires associated with fluvial environments (Roberts and others, 1994). The Mayfield



Figure 5. Composite stratigraphic column of the Mayfield coal beds and related beds in the Grass Creek coal mine. Adapted from Roberts and others (1994).

Mileage:		Description:
Incre- ment	Total	
		coal bed is as much as 36 ft (10.9 m) thick and is high volatile C bituminous in apparent rank. Ash yields average about 17% (dry basis), total sulfur content (dry, ash-free basis) is generally less than 0.5%, and volatile matter content (dry, ash-free basis) averages about 45%. The depositional environments and coal petrology of the Mayfield and associated coal beds are discussed in Roberts and Stanton (this volume).
4.9	41.6	Return to road junction on State Highway 171. Turn west (left). Between the road junction and the next stop (6.2 mi or 10 km), we will be traversing Cody Shale and the central part of the Grass Creek oil field; note the closely spaced wells, which are typically within 0.1 mi (0.2 km) of one another (Hewett, 1926). Wells were unitized within the extent of the oil pool in the sandstone reservoirs of the Cretaceous Frontier Formation; the pool encompasses about 1,600 acres (648 hectares).
3.1	44.7	Crossing Grass Creek and outcrops of the Cody Shale.
1.5	46.2	Cliff-forming sandstones on the right are the lowermost part of the Cretaceous Mesaverde Formation overlying Cody Shale.
1.5	47.7	Stop 3- Overview escarpments near Sand Draw gap. Stop locality is on the Cody Shale. To the north and east is the northeast flank of the Grass Creek anticline; this flank dips 15° to 22° to the north-northeast. Here, the Cretaceous Cody Shale grades upward into the Mesaverde Formation. The sandstones in the lower part of the Mesaverde Formation cap coarsening-upward units of mudstone, siltstone, and silty sandstone, and these stacked, coarsening-upward units represent parasequences. Marley and others (1984; 1985) interpreted the Mesaverde Formation in the western Bighorn Basin as progradational successions of shoreface, barrier, and coastal plain deposits. The coarsening-upward units in the lower part of the Mesaverde Formation are overlain by a coal-bearing interval. Along the escarpments of the Grass Creek anticline Hewett (1926) described the Mesaverde coal beds to be lenticular and as much as 14 in (36 cm) thick; the lateral extent of most coal beds was estimated to range from 3,000 to 7,000 ft (915 to 2134 m). However, some beds may extend from 15,000 to 18,000 ft (4,573 to 5,488 m). The coals average 10% ash (dry basis), 1.0% sulfur (dry, ash-free basis), and 46% volatile matter content (dry, ash-free basis). Heating value averages about 13,000 Btu (dry, ash-free basis). To the south of this vantage point, the southwest flank of the Grass Creek anticline, which dips 20° to 58° to the southwest, is visible on the skyline. On the uppermost parts of the ridge are rocks of the Paleocene Fort Union Formation, which form the northeast flank of the Grass Creek syncline. The basal conglomerate of the Fort Union Formation may be seen to approximate the contact of the Tertiary and underlying Cretaceous formations.
1.4	49.1	Road junction of State Highways 171 and 120. Turn left (northwest) on Highway 120. Coal-bearing outcrops of the Upper Cretaceous Meeteetse Formation visible on the road cuts.
2.5	51.6	Rest Area. Lunch stop. The thick succession of sandstone at the rest area is included in the Cretaceous Lance Formation. The uppermost sandstone capping the succession in this area is interpreted as the basal bed of the Fort Union Formation (Hewett, 1926). After lunch, the trip turns around and proceeds southeast from the rest area on Highway 120.
4.8	56.4	Junction of State Highways 120 and 431. Turn left (east) on Highway 431.
0.5	56.9	Crossing approximate contact of the basal bed of the Fort Union Formation (ridges on both sides of the highway) and underlying Cretaceous Lance Formation. Here the basal bed of the Fort Union Formation is essentially non-conglomeratic. Beds in the Fort Union Formation dip 20° to 24° to the northeast.
0.6	57.5	Crossing Red Ridge. Outcrops on Red Ridge include conglomeratic fluvial channel sandstones which may be stratigraphically equivalent to conglomeratic sandstones on the upper part of Ilo Ridge.

Mileage:		Description:
Increment	Total	
1.0	58.5	Dirt road on right goes into Rankine Basin (a topographic low) which is underlain by deposits in the middle and upper part of the Fort Union Formation. This interval of the formation includes fluvial channel sandstones (locally conglomeratic) and variegated floodplain mudstones and siltstones.
0.9	59.4	Crossing the contact of the Paleocene Fort Union Formation and Eocene Willwood Formation. The contact is a slight angular unconformity; the Fort Union Formation dips 15°, whereas the overlying Willwood Formation dips 11° to the northeast. The lowermost part of the Willwood Formation at this locality is composed of two prominent beds of conglomerate.
1.7	61.1	Crossing Gooseberry Creek.
0.2	61.3	Road to the left is BLM access to Squaw Teats, visible on the skyline to the northwest above the valley of the Gooseberry Creek.
2.0	63.3	Stop 4- Overview of the Eocene Willwood Formation badland topography and paleosols. The lower Eocene Willwood Formation in the Bighorn Basin was deposited by fluvial systems including braided to meandering fluvial channels and associated crevasse splay and floodplain environments. Paleosols, marked by variegated units (gray, yellow, orange, red, purple) formed in the overbank-floodplain areas lateral to fluvial channels (Kraus and Bown, 1988). These paleosols can be differentiated on the basis of pedogenetic maturity. That is, the least mature paleosols developed in overbank or channel margin settings where sediments accumulated rapidly during flood stages. The most mature paleosols formed in distal floodplains where sediments accumulated significantly more slowly. Kraus and Bown (1988) described three pedofacies sequences in the Willwood Formation. The first pedofacies is a simple sequence consisting of one or more paleosol horizons bounded by crevasse splays. This sequence was generated by slow and episodic alluviation; pedogenesis was periodically interrupted by crevasse splay progradation during flood stages. A second pedofacies consists of a compound sequence composed of vertically stacked paleosol horizons interbedded with channel sandstones. This sequence underwent alternately increasing and decreasing pedogenesis influenced by channel migration and avulsion. A third pedofacies consists of a megasequence (hundreds of feet or meters thick) superimposed on the first two types of pedofacies; this type of pedofacies shows the overall changes in maturity of the paleosols reflected by allocyclic processes and tectonism. Paleosol profiles and accompanying clay mineralogy of the Willwood Formation from a road cut nearby are discussed in Thorez and others (this volume).
6.2	69.5	Willwood paleosol outcrops along the road studied by Thorez and others (this volume).
7.1	76.6	Approximate contact between the Paleocene Fort Union Formation and Eocene Willwood Formation is across the valley of Gooseberry Creek on the right, on the "backside" (south side) of Blue Ridge. Outcrops on both sides of the road are within the Willwood Formation.
9.9	86.5	Road junction of State Highway 431 and U.S. Highway 20. Outcrops to the south and southeast are in the Fort Union Formation, which extends beyond the skyline. Turn left (northeast) on Highway 20, proceeding toward Worland. The outcrops along the bluffs across the Bighorn River on the right (east) side of U.S. Highway 20 are also part of the Fort Union Formation. Outcrops on the left and right sides of the highway are within the Willwood Formation.
9.1	95.6	City of Worland. Road junction of U.S. Highway 20 and U.S. Highway 16. Turn left (north) on Highway 16. The population of Worland is about 6,000, and the main industries are agriculture, ranching, and manufacturing. The Bighorn River floodplain, on which Worland rests, is farmed for sugar beets, hops, barley, and grains. These agricultural products support related manufacturing companies (e.g., Holly Sugar Corp., Fremont Beverage, PepsiCo., etc.). Drilling mud and other products are manufactured from bentonites mined in beds of Cretaceous age. Oil and gas activity is currently a very minor industry, even though the city is in close proximity to oil fields.
5.0	100.6	Bluffs and badland topography on the right (east) across the Big Horn River valley are composed of the Eocene Willwood Formation. Variegated units are paleosols.

Mileage:		Description:
Incre- ment	Total	
5.5	106.1	Approximate contact of the variegated Willwood Formation and coal-bearing Fort Union Formation may be seen on the bluffs to the right (east) across the Big Horn River valley.
3.4	109.5	Outcrops on the right side of the highway are in the fine-grained dominated, coal-bearing (mainly carbonaceous shale and some thin coal beds) Fort Union Formation.
4.2	113.7	Stop 5- Overview of the Fort Union Formation at the confluence of Big Horn and Nowood Rivers. At this locality, the coal-bearing rocks of the Fort Union Formation dip about 10° west-southwest. The Fort Union Formation is in contact with the underlying sandstones of the upper part of the Lance Formation. On this stop, we can compare the characteristics of the lower part of the Fort Union Formation observed in the western part of the Bighorn Basin (Grass Creek syncline) to these Fort Union outcrops in the eastern part of the basin. Palynologically, the Fort Union coal beds in the Grass Creek area were dated as early-middle Paleocene (P3 biozone) in contrast to these Fort Union beds, which are interpreted to be primarily late Paleocene in age (P5-P6 biozone). Lithologically, the Fort Union rocks in the Grass Creek area are sandstone-dominated (with conglomerate) in contrast to the mudstone-dominated Fort Union rocks at this locality. In addition, the mudstone-dominated Fort Union Formation in the eastern Bighorn Basin is smectite-rich, which is reflected in popcorn-like weathering in mud-rich outcrops. The sandstone-dominated strata in the western part of the basin probably formed in braided (bedload) and meandering (mixed-load) fluvial systems. In contrast, mudstone-dominated Fort Union deposits in the eastern part of the basin were probably deposited in fluvial environments characterized by low sinuosity channels "confined" by mud-rich floodplain deposits.
1.3	115.0	Crossing the Big Horn River. Coal-bearing outcrops on the left side of the road are in the Fort Union Formation.
2.4	117.4	Roadcuts consist of variegated paleosols of the Eocene Willwood Formation.
5.0	122.4	Outcrops of Willwood Formation in roadcuts.
3.5	125.9	Town of Basin. Continue northward on U.S. Highway 20/16 toward Greybull.
7.7	133.6	Town of Greybull. Junction of U.S. Highways 20 and 16 with U.S. Highway 14. Turn right (east) on Highway 14 toward Sheridan.
0.4	134.0	Crossing the Big Horn River.
0.1	134.1	Outcrop of sandstone of the Cretaceous Frontier Formation on the right side of the road.
2.1	136.2	Crossing the approximate contact of the sandstone of the Cretaceous Frontier Formation and the overlying Cody Shale (see left or north side of the road). On the north side of the highway, the Big Sheep Mountain anticline is flanked to the south by Triassic redbeds (Chugwater Group).
2.6	138.8	Frontier sandstones dipping to southwest, away from the Shell Creek anticline.
0.1	138.9	Crossing approximate contact between sandstones of the Upper Cretaceous Frontier Formation and the underlying Lower Cretaceous Mowry Shale (Curry and others, 1947). On the north (left) side of the road is an anticline that exposes the maroon and variegated shales of the Lower Cretaceous Cloverly and Upper Jurassic Morrison Formations. The hills to the northeast are underlain by the Cody Shale.
1.2	140.1	The hills to the south (right) are underlain by the Lower Cretaceous Cloverly and Upper Jurassic Morrison Formations.
0.2	140.3	School house on the left.
0.9	141.2	To the south (right) are the Lower Cretaceous rusty beds dipping away from the east flank of the Shell Creek anticline (Curry and others, 1947).
0.6	141.8	The hills to the south are underlain by the Lower Cretaceous Mowry Shale.
1.3	143.1	Bluffs to the north (left) along the Shell Creek anticline contain Upper Cretaceous Cody Shale.

Mileage:		Description:
Increment	Total	
0.4	143.5	South (right) of the highway, the interval from the Lower Cretaceous Muddy Sandstone to the Upper Jurassic Morrison Formation is well exposed (Curry and others, 1947).
3.4	146.9	On the right side of the road are maroon and variegated beds of the Cloverly and Morrison Formations.
1.0	147.9	The hills to the east are underlain by the Middle and Upper Jurassic Sundance Formation.
0.5	148.4	Town of Shell.
0.8	149.2	Crossing Shell Creek. Rocks of the Sundance Formation are exposed on the north side of the road.
0.8	150.0	Outcrops along the roadside include gypsiferous beds of the Middle Jurassic Gypsum Spring Formation and redbeds of the Triassic Chugwater Group.
0.3	150.3	Alcova Limestone of the Chugwater Group on the left.
2.2	152.5	Permian gypsiferous redbeds (Goose Egg Formation).
0.2	152.7	Crossing Shell Creek. Outcrop of Pennsylvanian Tensleep Formation.
0.4	153.1	Entering Bighorn National Forest.
0.2	153.3	The Mississippian Madison Limestone forms the gateway to the Shell Creek Canyon.
0.5	153.8	Bighorn Dolomite. Crossing approximate contact of the Ordovician Bighorn Dolomite and Cambrian Gallatin Limestone (which includes limestones, green shales, and conglomerates at this locality).
2.9	156.7	Precambrian granite in contact with Cambrian rocks on the left side of the road (gorge overlook).
2.5	159.2	Contact of the Cambrian with Precambrian basement rocks.
3.5	162.7	Granite Creek. Exposures of Precambrian granite on the right side of road.
6.4	169.1	Across the valley (left), the smooth topographic slopes are in Cambrian shales. The caprock is the Ordovician Bighorn Dolomite.
1.2	170.3	Weathered Precambrian granite on both sides of the road.
3.6	173.9	Knobby topography is carved out of the Precambrian granite.
9.6	183.5	Burgess Junction; juncture of U.S. Highways 14 and 14-A. Turn right (east) on Highway 14 toward Sheridan. All bedrock road cuts for the next 13 mi (21 km) are in Precambrian granite.
1.2	184.7	Precambrian outcrop is on the right.
9.1	193.8	Paleozoic rocks dipping to the west are visible to the north (left) across the headwater of the Tongue River.
2.6	196.4	Road cut is composed of Precambrian granite.
0.1	196.5	Steamboat Point to the north-northeast is the tip of a flatiron formed by the lower part of the Ordovician Bighorn Dolomite, which dips 25° to the northeast (Whitehead, 1986). The Bighorn Dolomite cliffs are underlain by the Cambrian Gallatin and Gros Ventre Formations (grassy slopes). To the north is the Box Canyon of the Tongue River, which follows along the Tongue River lineament/fault zone. The Box Canyon section starts, from the bottom to top, in the Cambrian Gallatin Formation and continues through rocks of the Ordovician, Devonian, and Mississippian (Madison Limestone) Systems (see Figures 6 and 7 for the stratigraphic column of the eastern flank of the Bighorn Mountains). Redbeds on top of the hillside are in the Upper Mississippian to Middle Pennsylvanian Amsden Formation. The Precambrian core rocks below the Paleozoic rocks cover an 1,100 mi ² (2,850 km ²) area of the Bighorn Mountains. Areally, the Precambrian rocks may be grouped into granite in the northern part of the Bighorn Mountains and gneiss in the southern part (Heimlich, 1969; Love and Christiansen, 1985). Rb-Sr age dating of these rocks indicate an isochron age of 2.85 Ga (Stueber and Heimlich, 1977).

GENERALIZED PALEOZOIC SECTION along US-14, East flank of BIG HORN MOUNTAINS, WYOMING

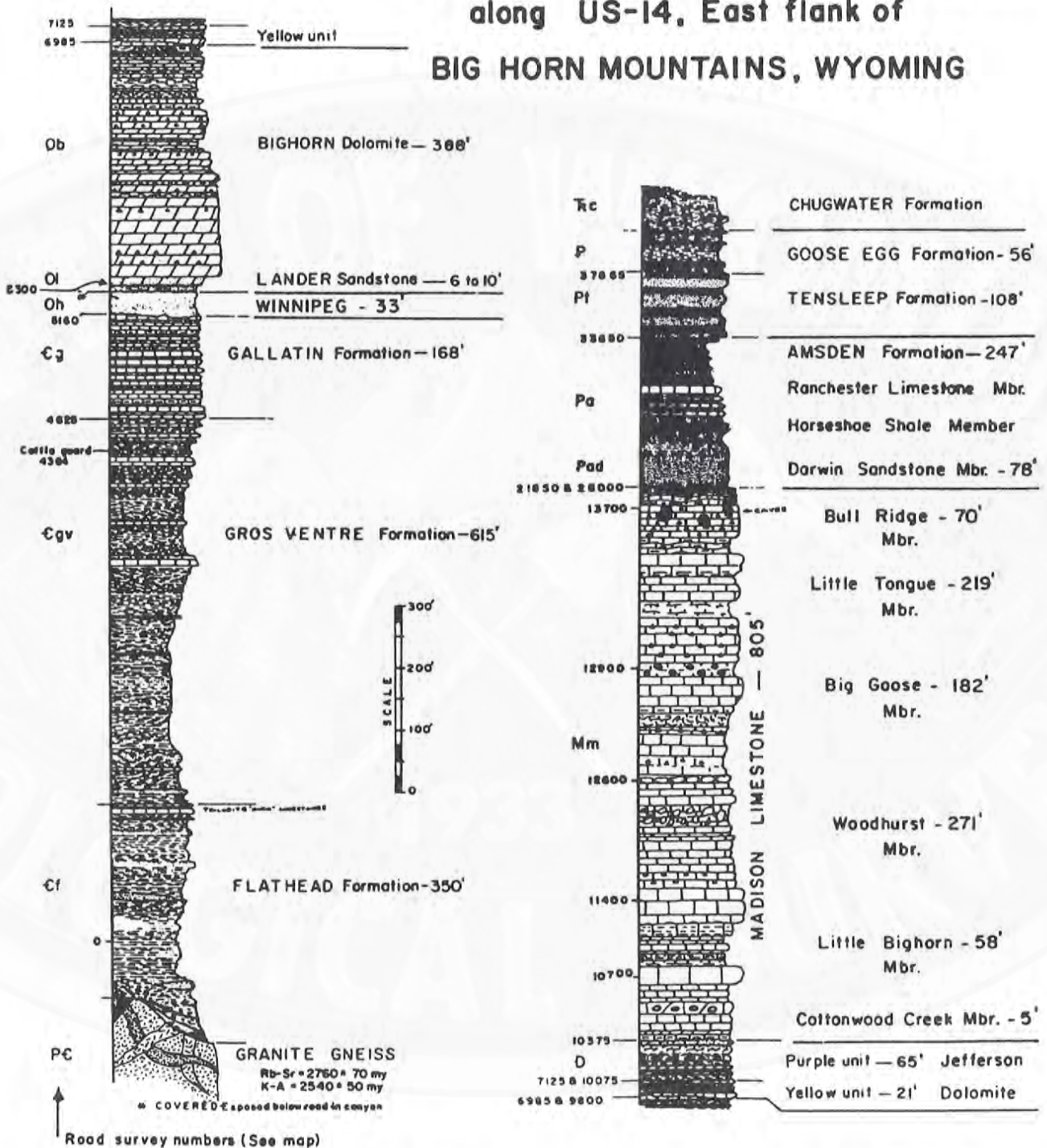


Figure 6. A generalized stratigraphic section along U.S. Highway 14, on the eastern flank of the Bighorn Mountains. Adapted from Kuocky and Cygan (1963).

Mileage:		Description:
Increment	Total	
0.7	197.2	Road cut in green shale, siltstones, and sandstones of the Cambrian Gros Ventre Formation.
2.6	199.8	The turn was named for the Darwin Sandstone Member of the Mississippian and Pennsylvanian Amsden Formation, when it was well exposed here. To the north is the Little Tongue River (Figure 7) and across the valley is the Buffalo Tongue slide. The slide involved the movement of a block of the Ordovician Bighorn Dolomite downslope at or near the top of the Cambrian Gallatin Formation, which in turn, ponded the Little Tongue River. Red and yellow beds of the Upper Devonian Jefferson Formation are exposed on the left of the slide. The Jefferson Formation is overlain by the Mississippian Madison Limestone. The road cut across the highway is the top of the reference section for the Madison Limestone by Sando (1979). Along U.S. Highway 14, Sando (1979) measured about 805 ft (245 m) thick of the Madison Limestone. This limestone is interpreted by Gutschick and others (1980) to be the deposit of a broad cratonic, carbonate platform that lay east of a deep-water clastic basin fed by the Antler orogenic belt to the west. This carbonate platform was transformed into a karst plain by a drop of relative sea level, giving rise to the karst, cave, breccia, and enlarged joint features in the upper part of the Madison Limestone (Sando, 1974; Gutschick and others, 1980). The contact of the Madison Limestone with the Darwin Sandstone Member of the Amsden Formation is on the wooded slope above the road cut at the Sand Turn Overlook.
1.7	201.5	The contact of the Madison Limestone with the Darwin Sandstone Member of the Amsden Formation is on the left.
0.3	201.8	Stop 6- Overview of the northwest Powder River Basin. Straight ahead ("12 o'clock") is the town of Dayton 4.5 mi (7.3 km) in the distance. Just to the right is the town of Ranchester, about 10 mi (16 km) away. At "2 o'clock" (east) is the city of Sheridan approximately 18 mi (29 km) in the distance. On the skyline to the northeast are the Wolf Mountains, which are carved into coal-bearing rocks of the Paleocene Fort Union Formation. Gray to tan beds of the Lebo Member of the Fort Union Formation are present near Ranchester. These beds dip about 1° to the southeast along the plunging axis of the Powder River Basin. To the north and northeast, beyond Ranchester, are light-colored patches. These are spoil piles from the mining of coal beds (Dietz-Anderson coal beds) in the upper part of the Tongue River Member of the Fort Union Formation. A coal silo at the Big Horn Coal Company mine at Acme, Wyoming, is visible to the northeast. On the skyline to the north-northeast are the Badger Hills, just south of the Montana-Wyoming border; they are underlain by the coal-bearing Paleocene and Eocene Wasatch Formation. To the northeast is the Piney Creek salient with the Moncrief Ridge attached to the leading edge. The Moncrief Ridge is underlain by the Moncrief Conglomerate Member of the Wasatch Formation. These conglomerates are interpreted to be mountain front alluvial fans along the western margin of the Powder River Basin (Flores and Warwick, 1984; Flores, 1986). South of Sheridan, the Moncrief alluvial fans drained into a sump, which formed the 250-ft (76 m) thick Lake de Smet coal bed of the Paleocene and Eocene Wasatch Formation (Flores and Warwick, 1984).
1.4	203.2	Contact of Pennsylvanian Tensleep Sandstone and Pennsylvanian Ranchester Member of the Amsden Formation (light-colored carbonates) is on the right.
0.8	204.0	Drive through switchbacks in the Triassic Chugwater-Goose Egg interval.
1.1	205.1	Carbonates and gray-red shales of the Jurassic Gypsum Spring Formation are on the right.
2.4	207.5	On the right is a small hogback probably composed of the Lower Cretaceous Cloverly Formation.
2.8	210.3	Crossing Little Tongue River and entering Dayton (population 701).
0.7	211.0	Crossing Tongue River and leaving Dayton.
1.0	212.0	Outcrops of the Paleocene Fort Union Formation.
6.1	218.1	Entering Ranchester (population 655).
1.6	219.7	Road junction of U.S. Highway 14 and U.S. Interstate 90. Turn right to enter the eastbound Interstate 90.

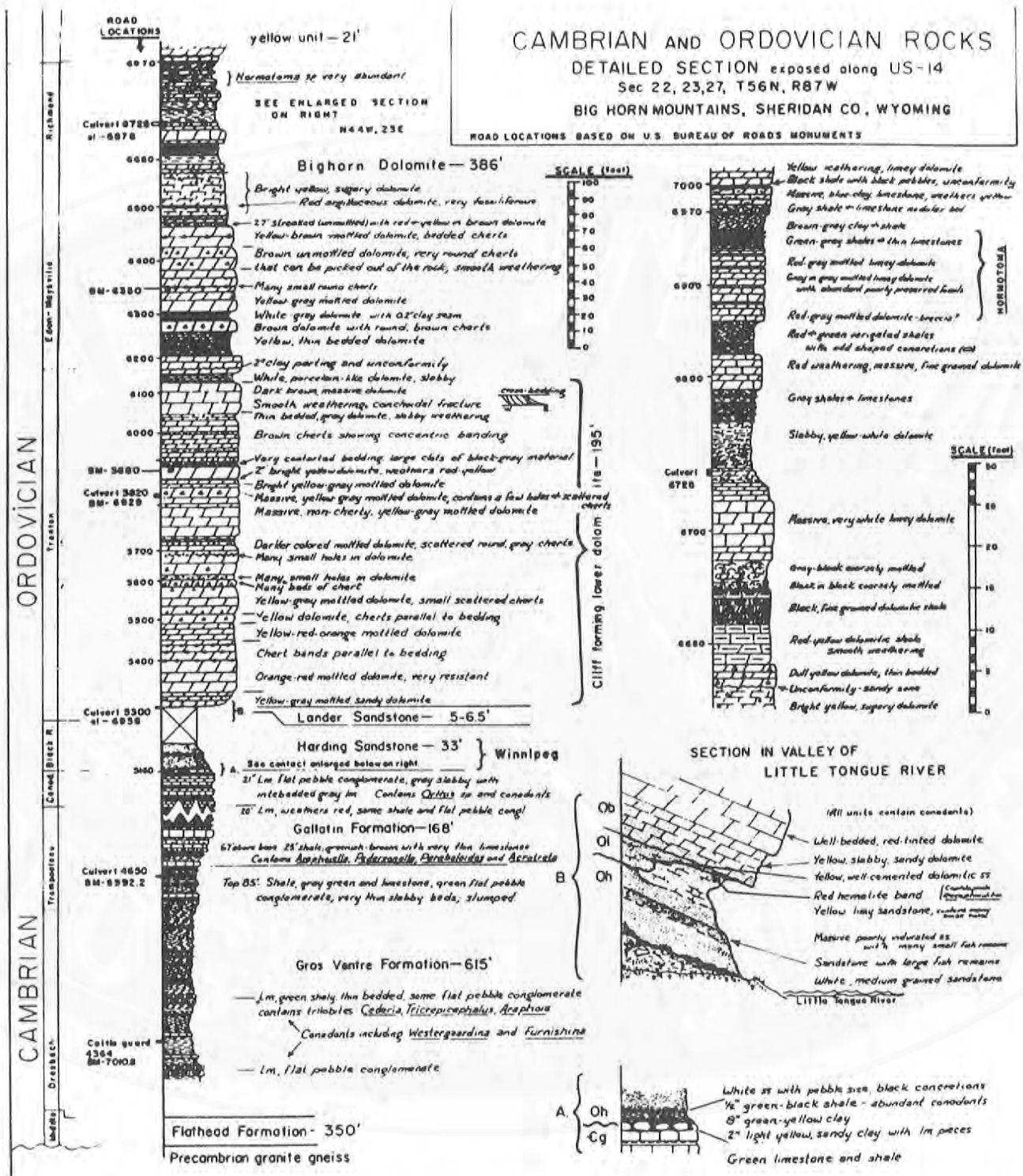


Figure 7. Stratigraphic sections of the Cambrian and Ordovician rocks along U.S. Highway 14, on the eastern flank of the Bighorn Mountains. Adapted from Kuocky and Cygan (1963).

Mileage:		Description:
Incre- ment	Total	
6.7	226.4	Crossing the Tongue River. Silo on the left is in the Bighorn mine. For the next mile the outcrops of sandstone, shale, carbonaceous shale, and coal beds are in the lower part of the Tongue River Member of the Fort Union Formation.
5.8	232.2	Exit on State Highway 338 and Main Street to downtown Sheridan.
1.9	234.1	End of Day 2. Stop at Sheridan Center Motor Inn, Sheridan.

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Bighorn Basin - Day two, Part 2

Stratigraphy and depositional setting of thick coal beds in the Grass Creek coal mine, southwestern Bighorn Basin, Wyoming

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Introduction

Grass Creek basin in the southwestern part of the Bighorn Basin, Wyoming (Figure 1), is a topographic basin coincident with the breached core of the Grass Creek anticline (Hewett, 1926). The basin encompasses an area of roughly 25 mi² (65 km²) and is floored by Upper Cretaceous deposits of the Cody Shale. The Cody Shale is overlain by the Mesaverde Formation, which forms a prominent escarpment surrounding the basin. Strata overlying the Mesaverde Formation include the Upper Cretaceous Meeteetse and Lance Formations and the lower Tertiary Fort Union Formation (Hewett, 1926; Andrews and others, 1947). The Fort Union Formation is overlain by the Eocene Willwood Formation, which crops out sporadically in areas southwest of Grass Creek basin, and forms a prominent ridge (Blue Ridge) along the northern margin of Rankine basin (Figure 2).

The Mesaverde, Meeteetse, and Fort Union Formations are all coal-bearing in the Grass Creek area. Coal production, however, has been restricted to the Mayfield coal bed in the lower part of the Fort Union Formation. This coal bed is the target of exploitation in the Grass Creek coal mine, located approximately 1.5 mi (2.4 km) southwest of Grass Creek basin (Figure 2). The mine, which is the only active coal mine in the Bighorn Basin, is a small surface mine consisting of adjoining strip pits oriented north-south (pit 1) and east-west (pit 2); strip pit lengths are about 280 ft (85

m) and 560 ft (170 m), respectively. The mine has been operated by Northwestern Resources, Inc. since the late 1970s.

We studied coal-bearing strata of the Fort Union Formation in the Grass Creek mine area to gain insight as to the depositional setting associated with thick, Paleocene coal beds. Our study area (Figure 2) encompasses some 30 mi² (78 km²) and includes Fort Union Formation outcrops to the south and southwest of Grass Creek basin (referred to collectively as the coal mine area) as well as outcrops northeast of the basin along Red Ridge. In the coal mine area, the lower part of the Fort Union Formation is exposed along the flanks of the northwest-trending Grass Creek syncline and the east-trending Ilo syncline (Hewett, 1926). The synclines are asymmetrical, with steeply dipping northern flanks (dips to 60°) and more gently dipping southern flanks (average dip of about 20°). Fort Union outcrops in both synclines have undergone substantial erosion, and the maximum preserved thickness of the formation in the coal mine area is 1,030 ft (314 m) (Hewett, 1926). At the mine, about 810 ft (247 m) of the lower part of the Fort Union is preserved (Roberts and Bossiroy, in press). Along Red Ridge and in Rankine basin, the lower and middle parts of the Fort Union Formation crop out as resistant, northwest-trending strike ridges that dip 10° to 20° to the northeast; the upper part of the formation is

intermittently exposed across Rankine basin to the base of Blue Ridge. The total thickness of the formation in the Red Ridge/Rankine basin area is 3,140 ft (957 m) (Hewett, 1926).

Detailed facies analyses of coal beds and associated clastic deposits were completed in the mine

and at locations on Ilo Ridge and Red Ridge. The studied interval incorporates 150 ft (45 m) of strata from about 565 to 715 ft (172 to 218 m) above the top of the basal bed of the Fort Union Formation (Figure 3). The method of study incorporates (1) detailed measured sections and facies analysis of thick coal beds and associated clastic deposits in the mine,

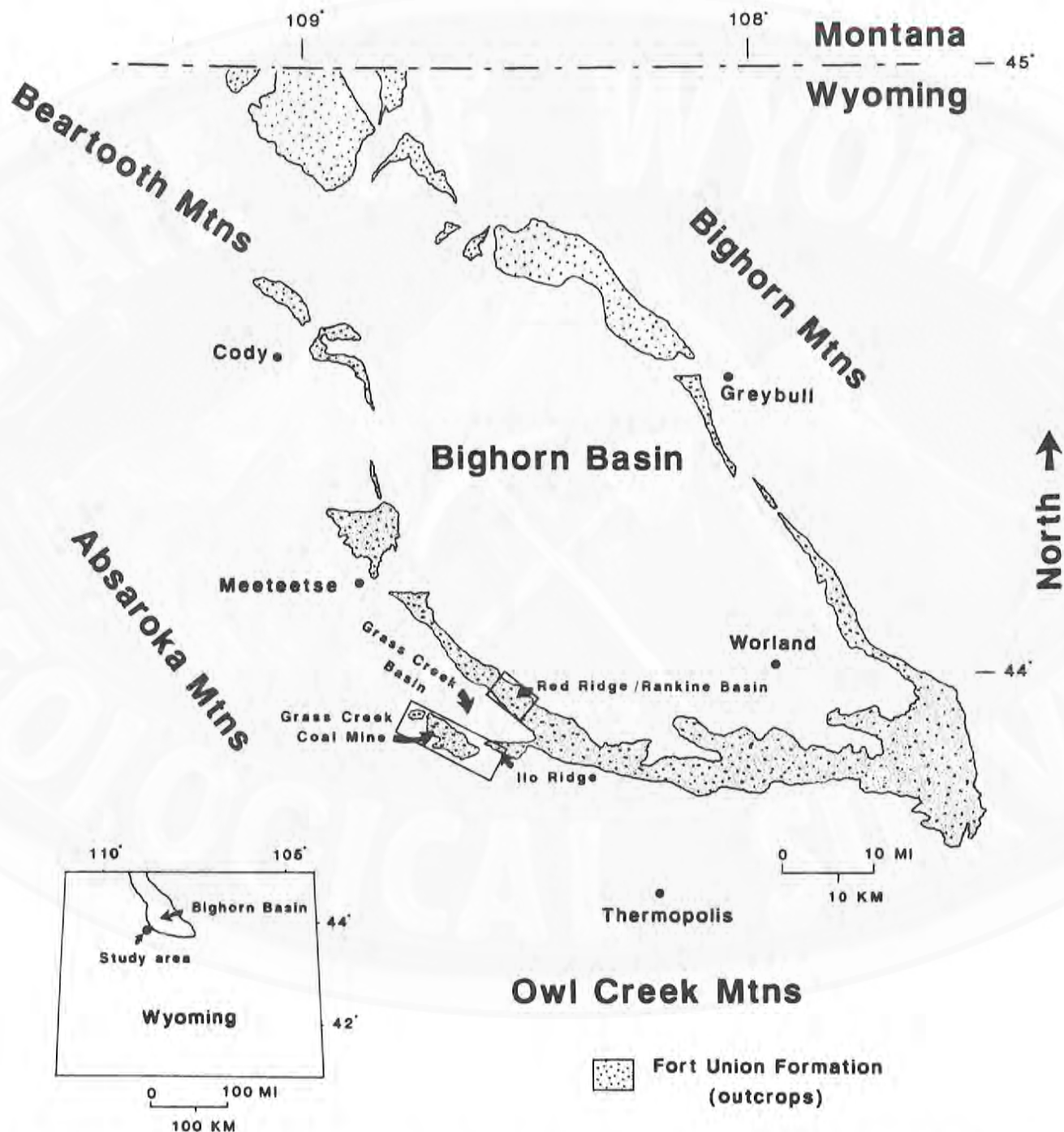


Figure 1. Index maps showing the location of the study area in the southwestern part of the Bighorn Basin, Wyoming. The principal area of study included outcrops in the Grass Creek coal mine and on Ilo Ridge and Red Ridge. Distribution of Fort Union Formation outcrops from Love and Christiansen (1985).

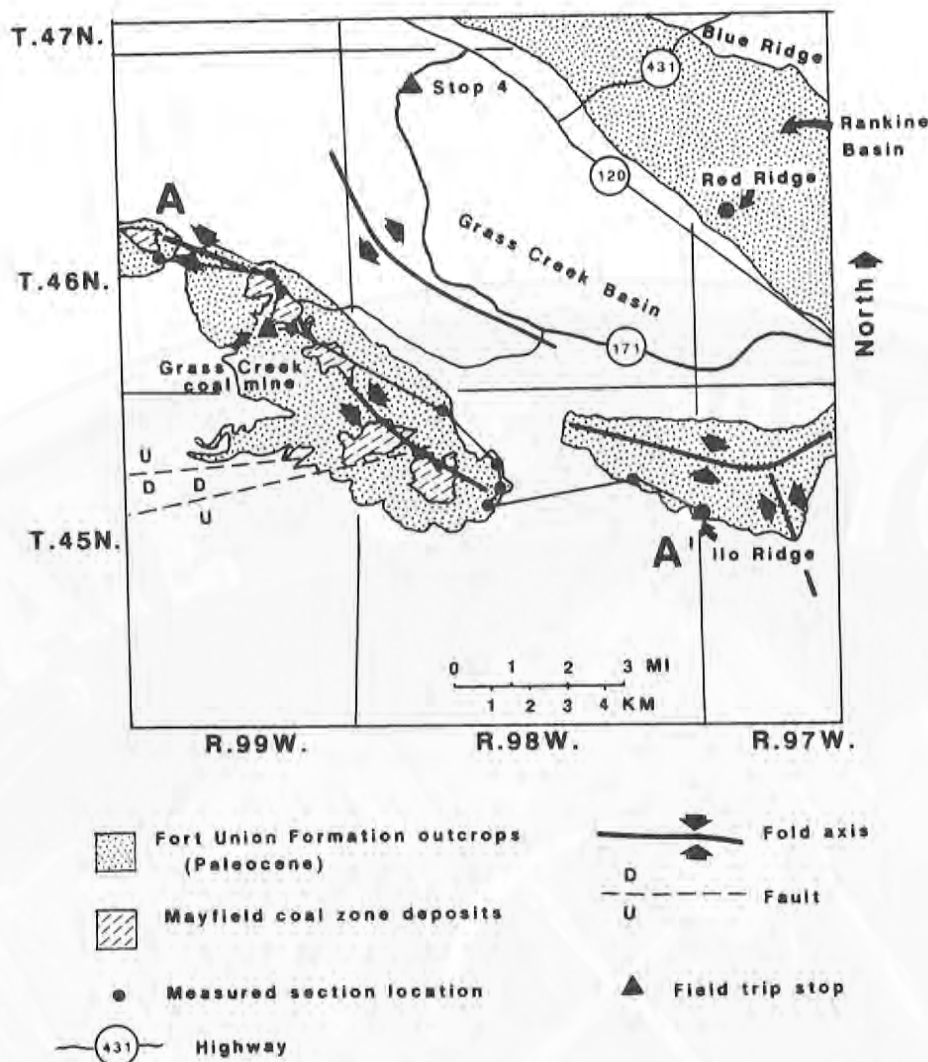


Figure 2. Generalized geologic map showing structural elements and distribution of Fort Union Formation and Mayfield coal zone deposits in the study area. Cross section A-A' is shown in Figure 4. Geology based on Hewett (1926).

Geologic setting

The Bighorn Basin is an asymmetric, intermontane structural basin that developed as the Laramide orogeny reshaped the Rocky Mountain foreland from Late Cretaceous to middle Eocene time. Deposition of the lower part of the Fort Union Formation in the Grass Creek area ensued during early Tertiary structural development of the Bighorn Basin (Love, 1973); the formation is Paleocene in age (Hewett, 1926). Palynomorph data from the Fort Union Formation in the coal mine area indicate deposition spanning the late early (P2 biozone) and early middle (P3 biozone) Paleocene. Fort Union Formation strata in the Red Ridge/Rankine basin area record deposition from

early (P2 biozone) through late (P5 biozone) Paleocene (D.J. Nichols, personal communication, 1994).

The Fort Union Formation overlies Upper Cretaceous fluvial and floodplain deposits of the Lance Formation; the contact has been reported as conformable (cf. Hewett, 1926) and unconformable (Kraus, 1985). Near the coal mine and on the west end of Ilo Ridge, the basal bed of the Fort Union Formation is a deposit of conglomerate that averages some 40 to 50 ft (12 to 15 m) in thickness (Hewett, 1926; Kraus, 1984, 1985); palynomorphs from above and below the basal bed indicate a latest Cretaceous or earliest Paleocene

(2) facies analysis of correlative non-coal-bearing strata on Ilo and Red Ridges, and (3) analyses of coal chemistry and coal petrographic characteristics. A discussion of coal sampling methods is reported in Roberts and others (1994). In addition, outcrop samples from coal and carbonaceous shale beds were collected for recovery of pollen and spores and determination of relative age based on the Paleocene palynostratigraphic zonation of Nichols and Ott (1978).

This paper presents an overview of the stratigraphy and sedimentology of the lower part of the Fort Union Formation in the Grass Creek area, provides descriptions of coal and clastic lithofacies present in the mine and equivalent horizons on Ilo Ridge and Red Ridge, and offers interpretations as to the depositional environment that influenced the development (and termination) of thick coal deposits in the Grass Creek coal mine.

age (Kraus, 1984). Toward the east end of Ilo Ridge, the basal bed grades into a buff sandstone with few or no conglomeratic lenses (Hewett, 1926; Roberts and

Bossiroy, in press) (Figure 3). In the Red Ridge area, the basal bed is a massive, buff-colored sandstone that ranges in thickness from 80 ft (24 m) to 230 ft (70 m),

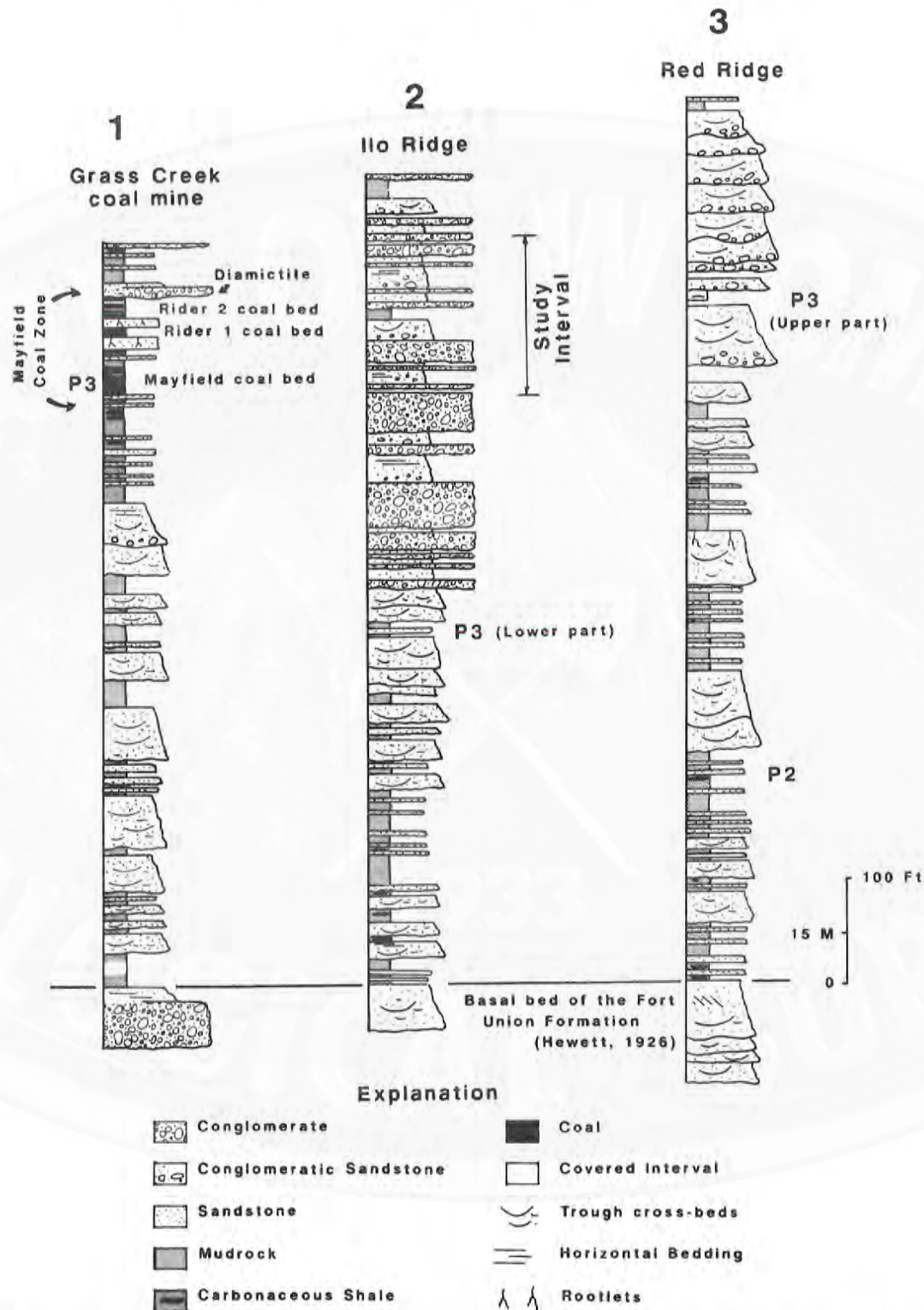


Figure 3. Stratigraphic columns of the lower part of the Fort Union Formation in the Grass Creek basin study area. Palynostratigraphic notation (e.g., P3) is based on Paleocene pollen zonations of Nichols and Ott (1978). The datum is the top of the basal bed of the Fort Union Formation as defined by Hewett (1926).

thickening from southeast to northwest along the ridge (Hewett, 1926); the basal bed is approximately 90 ft (27 m) thick in our Red Ridge measured section.

Strata of the Fort Union Formation overlying the basal bed consist of thin to thick, single and amalgamated (multistoried) sandstone bodies, quartzite conglomerate and conglomeratic sandstone, mudrock (mudstone, siltstone, and interlaminated very fine-grained sandstone), and carbonaceous shale and coal beds. Hewett (1926) identified lower and upper coal zones in the Fort Union Formation in the coal mine area (Figure 4); the focus of our study, the Mayfield coal zone (Roberts and others, 1994), is within the upper coal zone of Hewett (1926). The

Mayfield coal zone is about 100 ft (30 m) thick, and includes the basal Mayfield coal bed, and two overlying coal beds designated as the rider 1 and rider 2 coal beds (Figure 3). Hewett (1926) reported the Mayfield coal bed to be 420 ft (128 m) above the base of the Fort Union Formation in the coal mine area. However, based on unpublished drill hole data and our stratigraphic studies near the mine, we place the coal bed 625 ft (190 m) above the base of the Fort Union Formation in this area.

The Mayfield coal bed is middle Paleocene in age (biozone P3; D.J. Nichols, personal communication, 1991). In the strip pit, the Mayfield coal bed is 26 ft (7.9 m) thick; the coal bed attains a maximum thick-

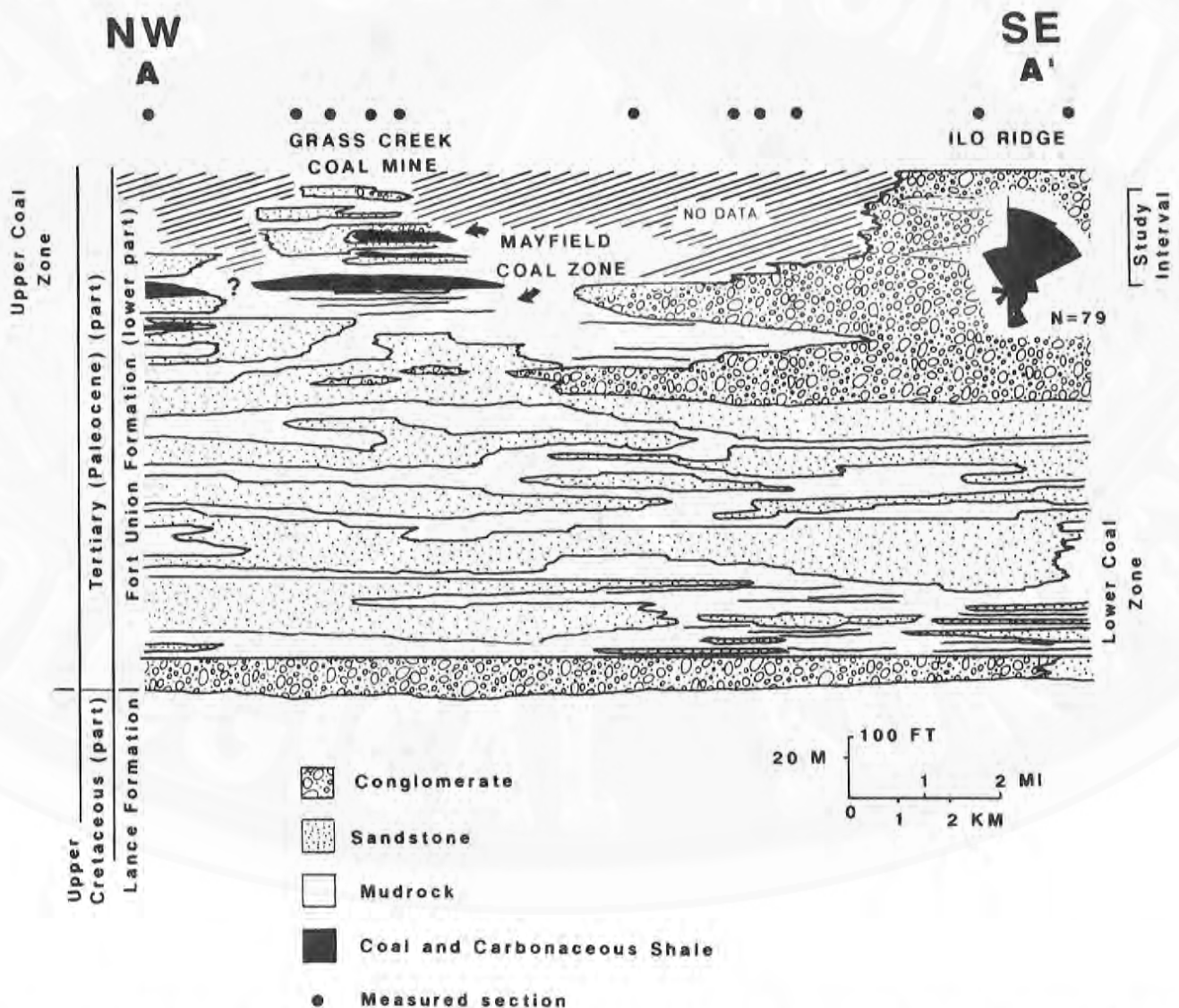


Figure 4. Generalized stratigraphic framework for the lower part of the Fort Union Formation in areas southwest of Grass Creek basin. The cross section is based on 12 measured sections; the location of the cross section is shown in Figure 2. Datum is the top of the basal bed of the Fort Union Formation (Hewett, 1926). The rose plot is based on 79 readings (cumulative) of trough and planar tabular foreset orientations and clast imbrication. Modified from Roberts and others, 1994.

ness of 36.8 ft (11.2 m) about 1 mi (1.6 km) north of the mine (Hewett, 1926). The Mayfield coal bed can be traced to outcrops 1.2 mi (2 km) southeast and 2.5 mi (4 km) northwest of the mine (Stewart, 1975). Within 2 to 3 mi (3 to 5 km) southeast of the mine, outcrops correlative to the Mayfield coal zone (lower part) contain mudrock with minor coal and carbonaceous shale. The rider 1 and rider 2 coal beds are apparently restricted to the immediate mine area, although incomplete exposures preclude accurate delineation of their extent. Maximum thicknesses of the rider 1 and 2 coal beds are 9 ft (2.7 m) and 19 ft (5.8 m), respectively. Both coal beds thin and pinch out to the west along the pit 2 highwall; erosional thinning of the rider 2 coal bed is also evident in both strip-pits.

The Mayfield coal zone is correlative (in part) to conglomerates on Ilo Ridge and conglomerates and sandstones in the Red Ridge area (Figure 3). A car-

bonaceous shale 70 to 80 ft (21 to 24 m) below the base of the conglomerate on Ilo Ridge is of early middle Paleocene age (biozone P3). Conglomerates on Red Ridge are in the lower part of a 400 to 500 ft-thick (120 to 150 m-thick) succession of conglomerate and sandstone that crops out along Red Ridge (upper part) and immediately north in Rankine basin. P3 biozonation of a carbonaceous shale bed in the lower part of the conglomeratic succession on Red Ridge indicates that these coarse clastics were also deposited in early middle Paleocene time, although their initial deposition is interpreted to be later than the onset of deposition of the Ilo Ridge conglomerates (D.J. Nichols, personal communication, 1994). Paleocurrent readings, based on trough and planar tabular crossbed foresets and clast imbrication in conglomerates and sandstones on both Ilo and Red Ridges, indicate northeasterly (basinward) paleoflow (see rose plot for Ilo Ridge, Figure 4).

Lithofacies characteristics

Conglomerate

Conglomerate in the study area is quartzite-rich, consisting primarily of well-rounded clasts of gray, green, and red quartzite. Fragments of silicified wood as well as clasts of chert, sandstone, and volcanic fragments (trace amounts) are additional constituents of the conglomerates (Hewett, 1926). Long axis (A-axis) lengths of larger clasts can exceed 8 to 9 in. (20 to 23 cm), and the outer surface of individual clasts is typically scarred by radial cracks or percussion marks.

On Ilo Ridge, conglomerates comprise an interval that is 380 ft-thick (115 m-thick). The base of the interval is 375 ft (114 m) above the top of the basal bed of the Fort Union Formation, and the conglomerate extends for more than 2 mi (3.2 km) along the upper part of the ridge. The lower 195 ft (59 m) of the conglomeratic interval is characterized by amalgamated beds of framework supported, massive, trough cross-stratified, and planar tabular cross-stratified conglomerate interbedded with matrix supported conglomerate (less common) and medium- to coarse-grained sandstone beds. Normal and inverse graded beds of conglomerate were observed locally. Individual beds of conglomerate are separated by sharp bounding surfaces, and basal contacts typically exhibit erosional relief. Massive and matrix-supported conglomerate is better developed in the lower 60 ft (18 m) of the conglomeratic succession.

The upper 185 ft (56 m) of the Ilo Ridge succession is more sandstone-rich than the lower part, and characterized by interbedded sandstone (medium-grained to conglomeratic) and conglomerate. The overall size of the clasts in conglomeratic interbeds has decreased relative to the underlying, conglomerate-rich part of the succession. The conglomerate is typically trough cross-stratified, and basal contacts of individual beds are sharp; locally, the contacts exhibit erosional relief. A conglomerate lag is common above the basal contacts of interbedded sandstone units.

On Red Ridge, the succession of conglomerate and sandstone begins 580 ft (177 m) above the top of the basal bed of the Fort Union Formation; 240 ft (73 m) of this interval is included in the Red Ridge measured section (Figure 3). These conglomerates are correlative to the upper sandstone-rich part of the conglomerate on Ilo Ridge. Compared to Ilo Ridge, however, sandstone is the more dominant lithology on Red Ridge. The associated conglomerate generally occurs as thin beds of trough cross-stratified and planar tabular cross-stratified conglomerate, as lenses in sandstone, and as lag gravels overlying basal contacts of sandstone beds. North of Red Ridge, in Rankine basin, lenticular beds of cross-stratified conglomerate are present.

Discontinuous exposures of conglomerate extend northwest from Ilo Ridge to the coal mine, along the

axial trend of the Grass Creek syncline. Individual beds within the conglomerates are primarily composed of framework supported, massive and trough cross-stratified conglomerate. Matrix supported conglomerate is less common, although a most striking example of matrix supported conglomerate is a bed of diamictite (conglomeratic mudstone) overlying the rider 2 coal bed in the highwall of the Grass Creek coal mine (Roberts and others, 1994; Figure 3). The diamictite is composed of "floating" quartzite clasts embedded in a sandy mudstone matrix; many of the clasts are oriented with long axes in vertical or near-vertical positions. Larger clasts within the diamictite are up to 10 in (25 cm) in length (A-axis), although more typical lengths are on the order of 4 to 6 in (10 to 15 cm).

In addition to the diamictite, the upper part of the Mayfield coal zone includes framework supported deposits of pebble and cobble conglomerate immediately overlying the diamictite and an additional conglomeratic horizon 45 ft (14 m) above the rider 2 coal bed. Within 3 mi (5 km) southeast of the mine, conglomerate and conglomeratic sandstones are interbedded with mudrock and thin coal beds at horizons equivalent to the lower part of the Mayfield coal zone (Figure 4).

Sandstone

Sandstone in the study interval ranges from very fine- to coarse-grained, and is locally conglomeratic. The mode of occurrence varies from thin, tabular sandstone beds to thick, vertically amalgamated (multistoried) sandstones. In general, the thick sandstones are characterized by well-developed cross-bedding, whereas tabular sandstones are characteristically laminated or cross-laminated.

Tabular sandstones range in thickness from less than 1 ft (0.3 m) to as much as 10 ft (3 m), although most are less than 3 ft (1 m) thick. Upper contacts are sharp; basal contacts are sharp or gradational. There is little or no erosional relief on basal contacts. Most tabular sandstones contain ripple (ripple-drift) or horizontal lamination; thick tabular sandstones may include small to medium-scale trough crossbeds. The sandstones are primarily very fine- to fine-grained, and grain size is generally uniform upward through the beds. Fossil root traces are common, and leaf imprints are preserved on laminae surfaces. Desiccation cracks and small vertical and horizontal burrows are present locally.

Units of cross-stratified sandstone range in thickness from 3 ft (1 m) to as much as 240 ft (73 m). Upper

contacts are sharp and flat; basal contacts generally exhibit erosional relief. Thick cross-stratified sandstones (as on Red Ridge) are composed of vertically and laterally amalgamated sandstone units separated by erosional surfaces. Trough crossbedding is the dominant sedimentary structure, and crossbed heights generally range from 1.5 to 3.0 ft (0.5 to 1.0 m). Planar tabular cross-stratification and horizontal bedding is locally well developed. Sedimentary structures are often highly contorted from soft sediment deformation, and dewatering structures in excess of 1 to 3 ft (0.3 to 1.0 m) in height are not uncommon. The sandstones are medium grained to conglomeratic. Conglomeratic sandstone is particularly abundant on Ilo Ridge and Red Ridge. Individual sandstones fine upward and sedimentary structures generally pass upward from large-scale trough crossbeds to small-scale crossbeds and ripple lamination. Fossil root traces and burrows are present locally.

Mudrock

Mudrock in the study interval includes silty to sandy mudstone, siltstone, and interlaminated very fine grained sandstone. Mudrock units are massive or laminated and weather brown, greenish-gray, and white on outcrop; white mudrock intervals have a granular ("popcorn-like") weathered texture. Locally the mudrock is variegated with reddish or reddish-purple hues. Gradational, upward-coarsening successions of mudstone, siltstone, and thin-bedded very fine-grained sandstone are common in mudrock intervals, and fossil rootlets tend to be well-developed in sandstones capping these successions. Siderite concretions are locally abundant within the mudrock intervals.

Coal

The Mayfield and rider 2 coal beds were described megascopically and sampled for chemical analyses. Ash and sulfur yields and volatile matter content were determined using ASTM standards 3177 and 3175, respectively (American Society for Testing and Materials, 1991). Maceral composition of the rider 2 coal bed was determined by detailed point-count analysis of 21 possible maceral varieties. Coal surfaces were etched for identification of vitrinite maceral varieties; however, for the sake of brevity, we do not use the prefix "crypto" as suggested by the International Committee for Coal Petrology (1971). Results of the coal analyses, previously reported in Roberts and others (1994), are summarized below.

Mayfield coal bed

Where sampled, the Mayfield coal bed is 26 ft (7.9 m) thick; of the total thickness, approximately 4.2 ft (1.3 m) is composed of clay-rich partings. Based on megascopic characteristics, the coal bed can be subdivided into four major units (MA-MD) from the base of the coal bed to the top (Figure 5). Unit MA is characterized by coal with a moderately woody texture, consisting of structured fragments of more than 2 in. (5 cm) in thickness and 30-60% matrix material. The overlying unit MB is composed of coarsely-laminated coal. Unit MC is matrix-dominant and finely laminated coal with clay-rich partings. Unit MD, at the top of the coal bed, consists primarily of woody textured coal with lesser finely-laminated coal. Ash yields (dry basis) for the coal (excluding partings) average about 17%. Medium to high ash coal (>8% ash yield) is closely associated with partings and matrix-dominant coal layers. The total sulfur content (dry, ash-free basis) is generally less than 0.5%. The volatile matter content (dry, ash-free basis) averages about 45%.

Rider 2 coal bed

The 19-ft-thick (5.8-m-thick) rider 2 coal bed, at the top of the Mayfield coal zone, includes six major clay-rich (detrital) partings and consists primarily of finely-laminated coal and coarsely-laminated, woody-dominated coal (Figure 5). Finely laminated coal is composed of thin layers of structured coal and attritus. Coarsely laminated coal consists of fragments of woody structured coal (70%) in attritus matrix. As with the Mayfield coal bed, we subdivided the rider 2 coal bed into four megascopic units (RA-RD) from the bottom to the top of the bed. Unit RA consists of finely-laminated, attritus coal with clay-rich layers; the unit contains three clay-rich (detrital) partings. Unit RB is similar in composition to unit RA, but contains fewer mineral layers. Unit RC consists of a claystone parting and coarsely laminated coal. Unit RD, at the top of the coal bed, includes a thick claystone parting

overlain by coal and interbedded thin, clay-rich partings; there is a distinct fusain layer in the upper part of the unit. Claystone partings decrease in number upward through the coal bed. Ash yields (dry basis) for the rider 2 coal bed average about 26%, and generally decrease upward through the coal bed. Coal with the highest ash yield (48-49%) is associated with partings in the lower part of the coal bed. Sulfur values (dry, ash-free basis) range from 0.8% (base of the coal bed) to 0.3% in the upper part (Table 1); organic sulfur is the dominant form of sulfur. Volatile matter content (dry, ash-free basis) averages 51.3%.

Maceral data (Table 1; Roberts and others, 1994) were organized into five genetically related groups (Stanton and others, 1989; Figure 5). Group I (preserved tissues) consists of telinite and corpocollinite and resinite associated with telinite. Group II (partially degraded tissues) is composed of gray gelocollinite and gray, degraded telinite. Group III (matrix humic gels) consists of black, brown, and micronized gelocollinite. Group IV (liptinite macerals) consists of amorphous resinite and bituminite. Group V is composed of detrital and oxidized fragments including vitrodetrinite, fusinite, semifusinite, macrinite, and inertodetrinite and corpocollinite in attritus or matrix.

Four major petrographic units comprise the rider 2 coal bed: (1) the lower part of the coal bed (samples 5 through 8, Figure 5), which contains high amounts of matrix gels and degradation-resistant liptinites; (2) samples 3 and 4, which contain the highest amounts of preserved cellular vitrinite; (3) sample 2, which consists of high amounts of woody-derived material; and (4) sample 1 (top of the coal bed) which is dominated by matrix humic gels and has the lowest ash yield (8%). Sample 1 also contains an appreciable amount of fusinite and semifusinite related to the megascopically observed fusain layer in the upper part of the coal bed.

Depositional setting

Stratigraphic and palynostratigraphic correlations of coal and clastic facies in the study interval indicate that development of the Mayfield coal zone was coeval with the development of coarse clastic (bed-load) fluvial systems in the Grass Creek basin area. The influx of conglomerates corresponds to interpreted middle Paleocene uplift of the Washakie Range-Owl Creek arch to the west and southwest of Grass Creek basin (Perry and others, 1991; Roberts,

1992; Flores and others, this volume). The dominance of quartzite clasts within the conglomerates may reflect recycling of quartzites in the Upper Cretaceous-Paleocene Pinyon Conglomerate (cf. Kraus, 1985), which was deposited in areas surrounding Jackson Hole to the west of Grass Creek (Lindsay, 1972; Love, 1973). Following uplift in the Washakie Range, the conglomerates were reworked from the flanks of the uplift and transported to the Grass Creek basin area

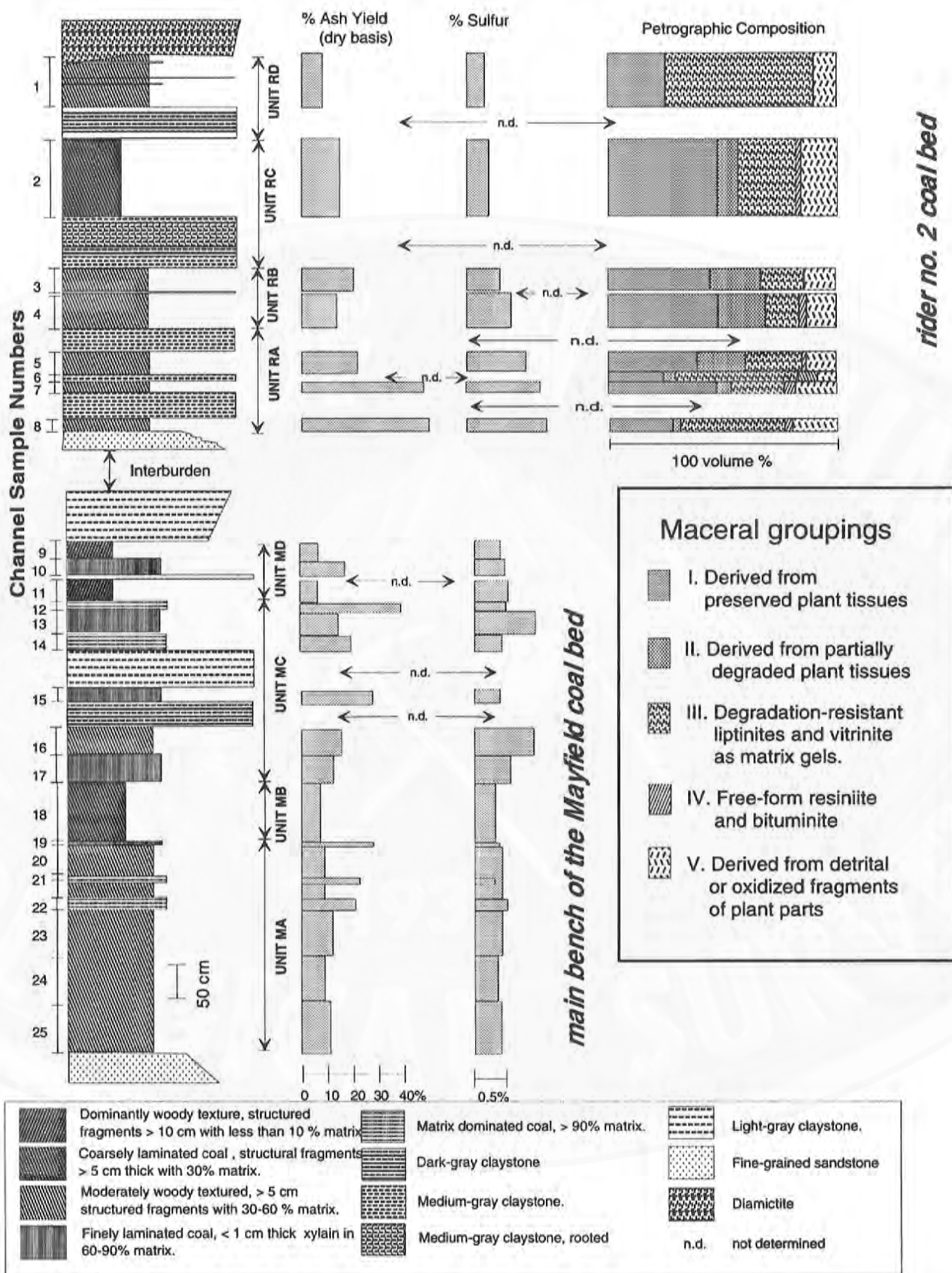


Figure 5. Megascopic coal facies profiles, maceral composition, and variations in ash yield, and sulfur content in the Mayfield and rider 2 coal beds in the Grass Creek coal mine. Ash yields on a dry basis; maceral data on a dry, volume percent, mineral-free basis; all other determinations on a dry, ash-free basis. Modified from Roberts and others, 1994.

Table 1. Maceral composition and quality of 8 samples from the rider 2 coal bed, Grass Creek coal mine in southwestern part of the Bighorn Basin, Wyoming

sam	tel ¹	crp/ tel ¹	deg tiss ²	gy gc ²	mic vit ³	br gc ³	bk gc ³	vd ⁵	fus ⁵	sf ⁵	mac ⁵	id ⁵	sp ³	cut ³	res/ tel ¹	res det ³	am res ⁴	bit ⁴	alg ³	ld ³	crp/ mat ⁵
1	7	3	—	—	6	—	69	2	2	5	T	2	3	T	—	—	—	—	—	T	1
2	32	5	4	6	3	12	7	10	—	4	T	6	6	T	—	1	T	2	1	—	—
3	28	6	12	16	1	13	1	13	—	1	T	1	4	T	—	1	—	1	T	—	2
4	30	7	6	18	4	6	T	10	—	2	T	3	4	1	3	2	—	3	T	—	1
5	20	7	7	19	—	20	3	10	T	1	—	5	5	T	—	2	T	T	—	—	—
6	7	2	3	3	T	1	3	2	T	4	T	5	48	1	1	4	1	8	—	6	—
7	33	3	5	4	2	9	7	8	—	4	—	11	8	1	—	1	—	5	—	—	—
8	13	1	5	—	1	—	28	3	—	7	1	12	18	1	—	7	—	4	—	—	1
Group																					
sam	I	II	III	IV	V	Ash ⁶	OS ⁷	Sulf ⁸	Hyd ⁹	VM ¹⁰											
1	10	0	78	0	12	8.8	0.31	0.33	4.4	46.4											
2	37	11	31	2	19	15.2	0.37	0.40	5.3	46.1											
3	33	27	21	1	18	21.5	0.53	0.61	6.0	49.8											
4	40	24	17	3	16	14.3	0.70	0.74	6.0	51.6											
5	27	25	31	1	16	24.4	0.94	0.97	6.4	53.5											
6	11	6	64	9	11	n.d.	n.d.	n.d.	n.d.	n.d.											
7	36	9	27	5	23	48.0	1.17	1.25	6.4	60.6											
8	14	5	54	4	23	49.6	1.33	1.39	7.0	59.1											
					avg.	25.97	0.76	0.81	5.9	51.3											

[sam=sample number; "Crypto"-macerals: tel=telinite, crp/tc=corpocollinite in telinite; deg tiss=degraded tissues; gy gc=gray gelocollinite; mic vit=micrinite associated vitrinite; br gc=brown gelocollinite; bk gc=black gelocollinite; vd=vitrodetrinite; sf=semifusinite; fus=fusinite; mac=macrinite; id=inertodetrinite; sp=sporinite; cut=cutinite; res/tel=resinite in telinite; resdet=resinite in matrix; am=amorphous (free) resinite; bit=bituminite; alg=alginite; ld=liptodetrinite; cry/mat=crypto-corpocollinite in matrix, T=trace].

¹Group I macerals.

²Group II macerals.

³Group III macerals.

⁴Group IV macerals.

⁵Group V macerals.

⁶Ash on a dry basis.

⁷Organic sulfur on a dry, ash-free basis.

⁸Total sulfur on a dry, ash-free basis.

⁹Hydrogen, on a dry, ash-free basis.

¹⁰Volatile matter on a dry, ash-free basis

via a northeast-flowing braidplain system (Roberts, 1992). Kraus (1985) invokes a similar interpretation for Eocene conglomerates in the basal Willwood Formation on Blue Ridge.

Peat accumulation (Mayfield coal zone) took place during deposition of the upper, sandstone-rich conglomeratic interval on Ilo Ridge and the inception of basinward, coarse clastic deposition on Red Ridge. Our interpretation of the depositional controls on peat accumulation is based primarily on an understanding of (1) coal-bed mire type, as reflected in the megascopic, microscopic (petrographic) and chemical characteristics of the coal beds, and (2) clastic depositional environments associated with coal-bed mires, as determined from facies analysis of sedimentary deposits interbedded with the coal beds. The discussion that follows, which addresses each of these aspects, is summarized from Roberts and others (1994).

The Mayfield and rider 2 coal beds are interpreted as deposits of low-lying, topogenous mires. Ash yields for the Mayfield and rider 2 coal beds (17% and 26%, respectively) are significantly higher than ash yields in coals interpreted to have developed from domed (raised) mires in the Powder River Basin (average ash yield < 5%; Stanton and others, 1989). In addition, the presence of numerous detrital partings and the resulting mineral enrichment of coal (high-ash layers) associated with these partings is also evidence for low-lying conditions, which could have allowed for the mixing of organics and clastics during periodic floods.

The variation in ash yields in both the Mayfield and rider 2 coal beds suggests that the termination of peat accumulation in these mires may have been rapid. A gradual demise, such as drowning of the mire, would probably result in an upward increase in ash yield as incursions of sediment-laden water became more frequent through time. In contrast, the general upward de-

crease in ash yields, particularly in the rider 2 coal bed (Figure 5), suggests a peat accumulating environment in which clastic influx became less frequent during the latter stages of peat accumulation. This factor, coupled with the interpreted low-lying mire conditions, suggests that mire demise was most likely the result of clastic incursion and rapid burial; the mechanisms of mire termination, therefore, are represented in the clastic facies associated with the coal beds (Roberts and others, 1994).

The interpreted depositional environments and processes of mire demise are summarized in Figures 6 and 7, respectively. Coal beds in the lower part of the zone (Mayfield and rider 1 coal beds) are overlain by coarsening-upward successions of mudstone and siltstone (mudrock), capped by tabular, rooted sand-

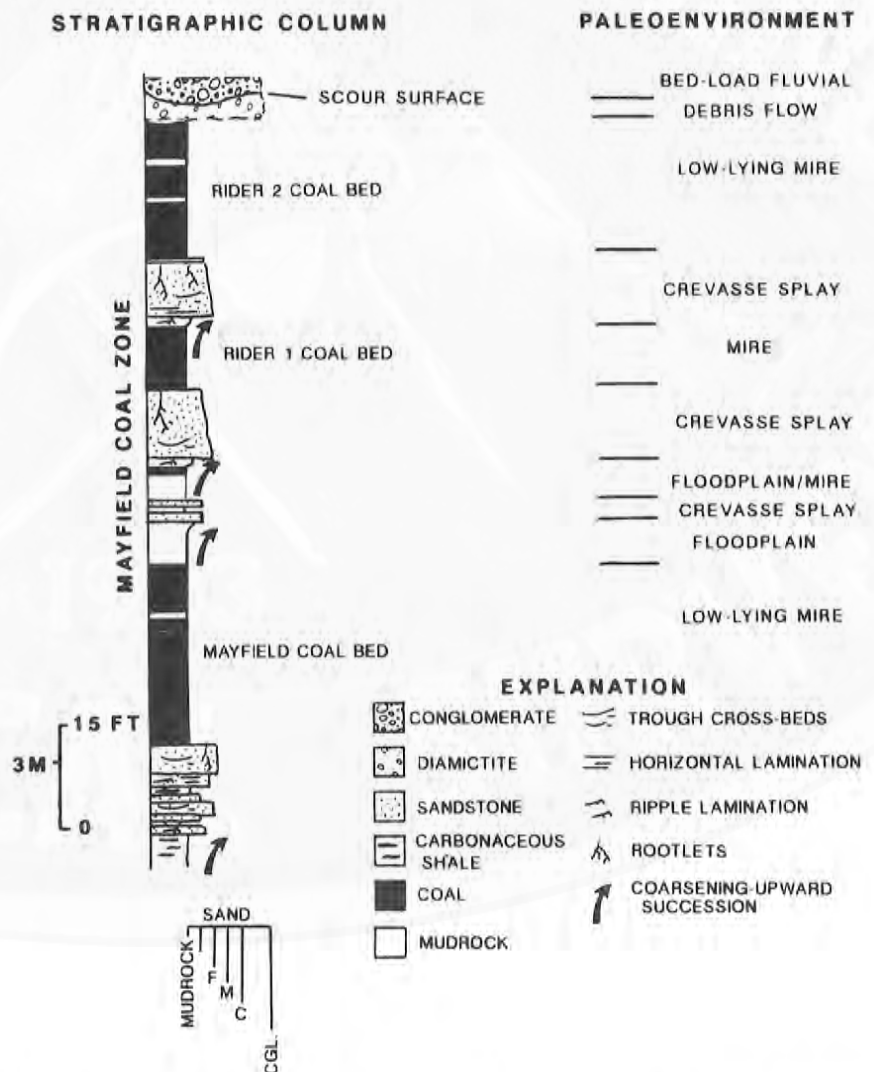


Figure 6. Stratigraphic column of the Mayfield coal zone in the Grass Creek coal mine showing interpreted paleoenvironments. Adapted from Roberts and others, 1994.

stone; these successions are characteristic of prograding crevasse splay systems (e.g., Gersib and McCabe, 1981; Flores, 1985). This association suggests that crevasse splay progradation was the primary cause of mire demise throughout the early history of the coal zone; this process is well documented in modern and ancient alluvial successions (e.g., Smith, 1983; Smith and others, 1989; Flores, 1986; Warwick and Flores, 1987).

The late history of the coal zone is represented by an influx of conglomerate that marks the termination of significant coal development in the Grass Creek basin area, as there are no coal beds in horizons overlying the coal zone. Diamictite, which immediately overlies the rider 2 coal bed, (Figure 3) is interpreted to be the deposit of a debris flow resulting from a short duration (hours to weeks), catastrophic flood, and thereby reflects a virtually instantaneous mechanism of mire demise (Roberts and others, 1994). Following deposition of the diamictite, coarse clastics were transported into the area via a bed-load fluvial system; incursion of the bed-load system

resulted in significant erosion of the diamictite and underlying rider 2 coal bed (Figure 7).

Figure 8 shows the distribution of interpreted depositional environments in the coal mine area during development of the Mayfield coal bed. Peat accumulated in a low-lying, "pod-like" depocenter, resulting in the formation of a thick, areally restricted coal deposit (see Mayfield coal-bed isopach maps of Stewart, 1975). Thick peat accumulation prevailed within 6 to 8 mi (9 to 12 km) of a coarse clastic fluvial belt; gradual deterioration of peat-accumulating environments toward the fluvial belt is evidenced by an increase in mudrock and carbonaceous shale, the lack of thick coal beds, and the interfingering of conglomerates and coal zone deposits in overbank reaches of the fluvial system.

In summary, coal beds in the Grass Creek coal mine are interpreted to be products of peat accumulation in low-lying, topogenous mires. These mires coexisted with a coarse clastic, bed-load fluvial system throughout much of the middle Paleocene; develop-

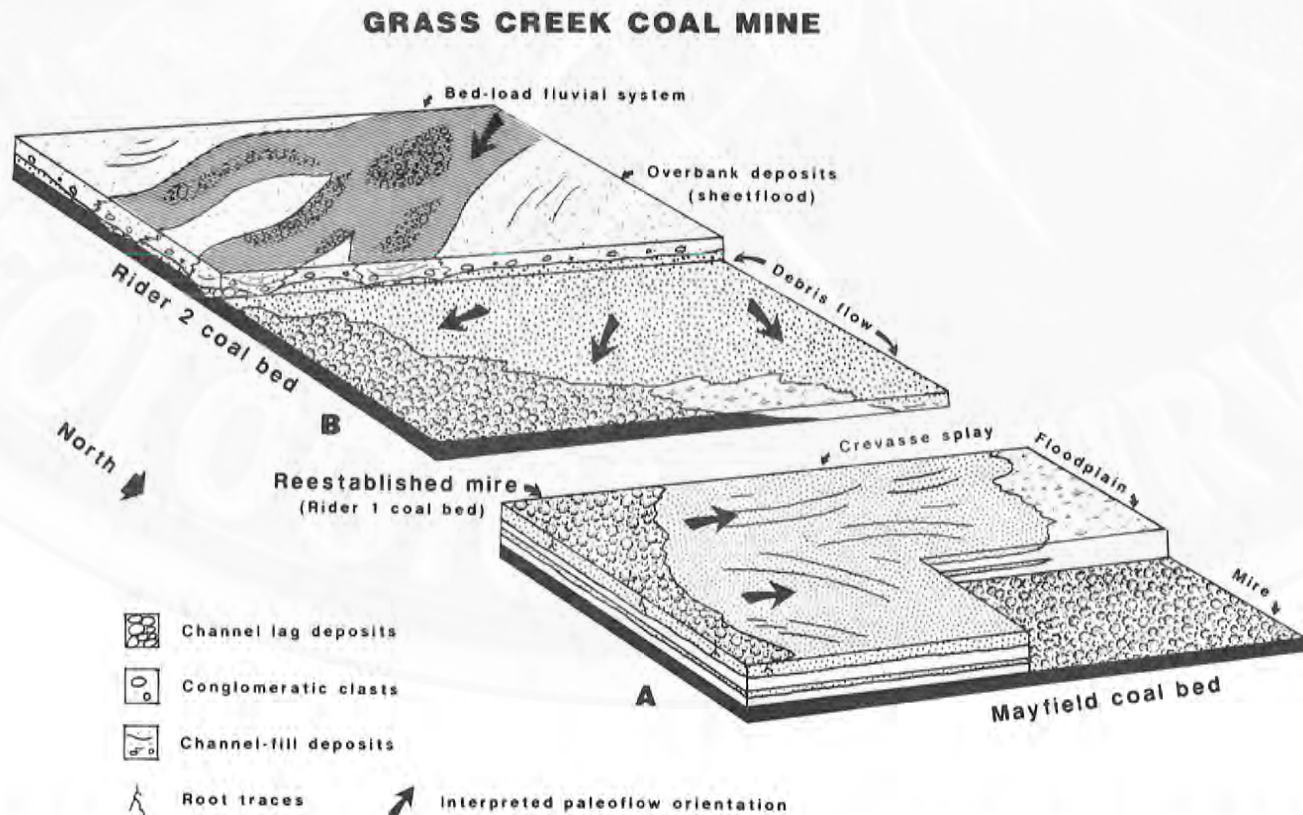


Figure 7. Block diagram illustrating the depositional setting of Mayfield coal zone deposits in the Grass Creek coal mine. Peat accumulation in low-lying mires (Mayfield and rider 2 coal beds) was periodically terminated by clastic incursion associated with crevasse splay progradation (A) and ultimately, catastrophic debris flow (B). Modified from Roberts and others, 1994.

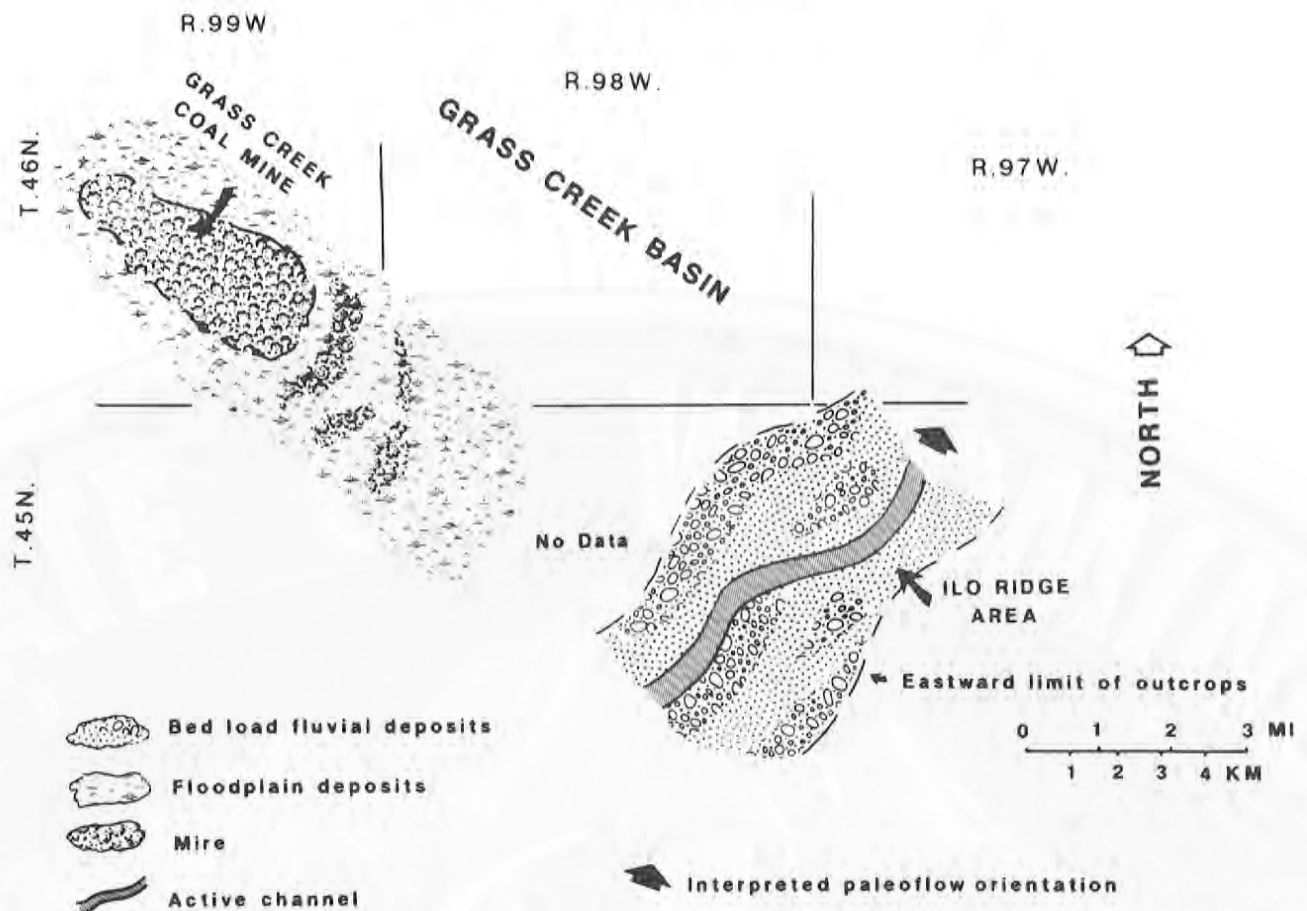


Figure 8. Diagram showing the distribution of interpreted paleoenvironments during deposition of the Mayfield coal bed southwest of Grass Creek basin. The coal bed developed in a low-lying "pod-like" mire lateral to a northeastward-flowing, braided fluvial system in middle Paleocene time. Delineation and areal extent of the Mayfield coal-bed mire based on Hewett (1926).

ment of this fluvial system was concurrent with uplift in the Washakie Range. In the latter part of the middle Paleocene, this fluvial system expanded and overran

areas of previous mire development, resulting in the termination of coal-forming environments in the Grass Creek basin area.

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Clay geology, micromorphology, and geochemistry of an Eocene paleosol in the Willwood Formation, southern Bighorn Basin, Wyoming: an overview

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Introduction

Fluvial deposits in the intermontane Wind River, Bighorn, and Powder River Basins are replete with different types of paleosols. These paleosols were formed mainly in overbank and floodplain environments and display various stages of maturity within these settings. Sedimentation and pedogenesis in the overbank and floodplain environments were controlled by paleoclimate, groundwater level, and the distance from active fluvial channels (Collinson, 1986; Kraus and Bown, 1988; Allen, 1989). Overbank floods resulting from peak channel discharges can inundate a floodplain with suspended sediments every one to two years (Wolman and Leopold, 1957). Deposition of sediments during these floods takes place on the channel margin (overbank and proximal floodplain) or flood-basinward (distal floodplain). Sediments accrete and fine flood-basinward. Vertical accretion and sediment-size differentiation in the flood basin depend on the direction of migration, aggradation (deposition), and degradation (erosion) by the chan-

nel. Coarsening-upward floodplain sediments suggest channel migration accompanied by overbank-crevasse splay aggradation toward that position in the flood basin. Fining-upward floodplain sediments indicate channel migration away from that position in the floodplain resulting in reduced rate of sedimentation and limited erosion (degradation). Between floods and periods of nondeposition, the floodplain was dry and exposed to subaerial weathering and pedogenesis. However, where the floodplain areas are subaqueous, due to high groundwater table, vegetal growth and bioturbation pervaded. In addition, in humid paleoclimates, these submerged floodplains supported permanent vegetation in mires that accumulated peat deposits.

The dynamics of the fluvial floodplain during deposition of the Willwood Formation were influenced by fluvial aggradation and degradation, paleoclimatic changes, and fluctuations of ground-

water level. The dynamics are also reflected by the types and maturity of paleosols. Wright (1989) suggested that paleosols that formed in overbank areas were immature because of well-drained conditions, whereas paleosols that formed in the proximal floodplain varied from immature to mature due to wet and/or dry paleoclimatic changes. Furthermore, Wright (1989) indicated that the moderately mature paleosols formed in the distal floodplains. Allen (1989) summarized the relationship of fluvial regimes and their influence on paleosol development and maturation. He stated that the maturity of paleosols was directly related to non-depositional and non-erosional stages in the overbank-floodplain environments. A long period of pedogenesis during these stages gave rise to mature paleosol development. However, as aggradation recurred toward the floodplain, paleosol development was arrested, giving rise to immature types. In addition, deposition (aggradation) in the floodplain preserved paleosols by burial and erosion (degradation) removed records of paleosol development. Thus, observations by Allen (1989) and Wright (1989) collectively indicate that the formation of mature paleosols is more likely in distal floodplains where non-depositional and non-erosional stages were maintained and pedogenesis was undisturbed.

Retallack (1988) recognized three main characteristics in of paleosols in the field: 1) root traces, 2) soil horizons (e.g., A, B, and C), and 3) structures (such as burrows and glaeboles). In addition, mineral and chemical characterizations are also useful in identification of paleosols and their degree of pedogenesis

(Feakes and Retallack, 1988; Gardner and others, 1988). Similar characterizations have also been utilized by Bown and Kraus (1981a, 1981b) and Bown (1985) for paleosols in the Eocene Willwood Formation of the Bighorn Basin (Figure 1). Paleosols in the Willwood are well exposed along the field trip route (State Highway 430). The purpose of this paper is to summarize the various paleosol types and characteristics based on previous works in the Willwood Formation. Although this summary is just a birdseye view of the paleosols in the Bighorn Basin, the field trip participants are referred to the works of Bown and Kraus (1981a, 1981b) for additional information and a detailed treatment of the subject. In addition, clay mineralogy of selected paleosol types exposed in a roadcut along the field trip route is discussed.

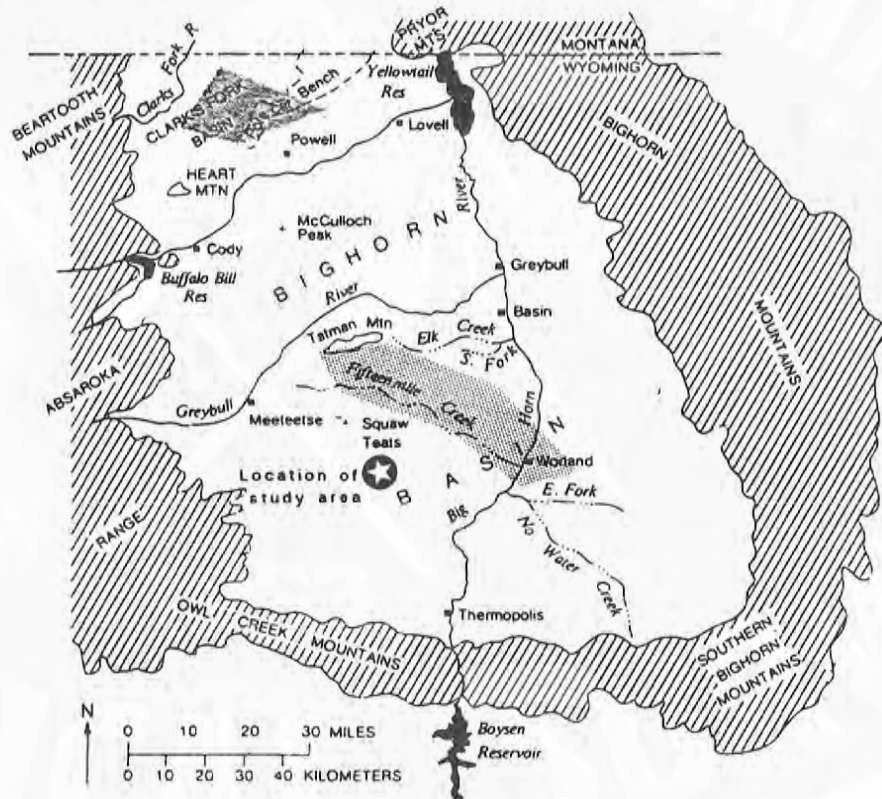


Figure 1. Map of the southern Bighorn Basin, Wyoming showing location of study area. Modified from Kraus and Bown (1988).

Willwood Formation paleosols (Eocene)

Bown (1985) described three principal types of alluvial paleosols in the Eocene Willwood Formation in the Bighorn Basin. These paleosols include spodosols, entisols, and red soils of uncertain classifi-

cation. All are well developed and widespread in overbank-floodplain deposits of the Willwood Formation.

Spodosols described by Bown (1985) as the most mature of the three Willwood paleosols, consist of an A horizon of gray to bluish or greenish-gray color with rhizoliths and vertebrate fossils. The A horizon is organic-rich and depauperate in clay minerals and Fe, Al, and Mn sesquioxides. An underlying B horizon includes an upper part, which is purple, red, and/or orange in color; this part is rich in organic carbon and Fe and Al sesquioxides that occur as interstitial films and plasmic glaebules. The upper part of the B horizon may be deficient in clay minerals and contain some calcrete glaebules. The lower part of the B horizon is rich in hydrated Fe and Al sesquioxides and associated with manganese oxides (interstitial or glaebule forms). Also, this part of the B horizon is locally rich in clay minerals and CaCO_3 glaebules and commonly displays a gray mottling that decreases upward. An underlying C horizon comprises unaltered parent material and is rarely present.

Bown (1985) described the entisols (stage 1) as poorly developed or immature paleosols. He stated that entisols generally have an ill-defined A horizon and an orange or red and orange B horizon with dep-

auperate organic carbon, rhizoliths, and vertebrate fossils. The B horizon is rich in Fe and Al sesquioxides (generally hydrated and rarely in glaebule forms). The B horizon contains some transported clay minerals and abundant CaCO_3 glaebules. Gray mottles are abundant in the lower part of the B horizon. Traces of earthworms (Bown and Kraus, 1983) and unknown vertebrates are locally abundant in this horizon. An underlying C horizon may also be present.

A red paleosol of uncertain classification was described by Bown (1985) as lacking an A horizon and comprising a brick red B horizon that may have a purplish upper part. The red horizon is very thick (as much as 4 m) and commonly mottled with orange in the lower part. The B horizon is rich in Fe and Al sesquioxides (dehydrated), which occur in interstitial and glaebule forms. Clay minerals are not as common as in spodosols and are best developed in association with CaCO_3 glaebules in the lower part of the B horizon. No significant amount of organic carbon or rhizoliths is present. An underlying C horizon is rare to absent.

Willwood paleosol case study

The roadcut to be visited as part of the field trip consists of three vertically superposed paleosols. The two lower paleosols are separated from the upper paleosol by a fluvial channel sandstone (Figure 2). All three paleosols are truncated at the top and lack their original A horizon; only Bw horizons are preserved. The characteristics of these paleosols are summarized in Caudron (1993).

Classification and interpretation of the paleosols

The paleosols below the channel sandstone are vertically stacked (see Figure 3). The lower paleosol is classified as an alfisol-spodosol (alfisol is probably equivalent to the unclassified red paleosols of Bown (1985) using Retallack's (1988) classification), whereas the upper paleosol is an inceptisol (Retallack, 1988). These classifications are macroscopically verified by utilizing the Munsell Color (1975) charts. The superposed colors of the mudstones and silty mudstone horizons, from top to bottom, are: grey with some orange patches; orange with lenticular banding and patches; and purple with patches.

The truncated paleosols below the channel sandstone may be interpreted in light of the work of Kraus and Bown (1988). They observed that the development of a purple B horizon generally occurred near the paleosol surface and thus, above any orange to reddish-colored horizons (Figure 3 in Kraus and Bown, 1988). In the lower paleosol profile, the purple B horizon is overlain and passes upward to an orange Bw horizon, which is in turn capped by a grey A-Bw horizon. This profile may be explained by invoking an erosional (degradational) episode in the floodplain. Degradation was probably caused by floodwaters from an advancing channel, which may have removed the upper part of a former alfisol-spodosol profile preserving only the underlying purple horizon. This degradation was followed by aggradation that deposited younger floodplain sediments, which were subsequently transformed by pedogenesis. Thus, the preserved purple horizon in the lower paleosol served as a C horizon for the upper inceptisol.

The paleosols below the channel sandstone range from paleoferralf-oll (paleo fer alfisol-spodosol), stages 2-3 (Kraus and Bown, 1988) upward to paleofluvept (paleo fluvent inceptisol; stage 2 inceptisol). Micro-



Figure 2. Stacked Willwood paleosols consisting of a inceptisol (I) in the lower part and alfisol-spodosol (II) in the upper part. A and Bw designate paleosol horizons. FC= fluvial channel sandstone.

morphologic analysis indicates mainly a plasma translocation in the purple and the orange Bw-horizons as revealed by Fe-glaebules and quasi-ferranes with scattered Fe-reduced patches generated during pseudo-gleyfication periods (controlled by fluctuation of the water table; see Figures 4A, B, and C).

The upper paleosol profile above the channel sandstone is less developed than the lower, stacked paleosol succession. It contains a Bw horizon, which displays grey color with some reddish patches randomly dispersed in the silty mudstone material. Such patches correspond to Fe-glaebules and are related to the very first step of the pedogenetic processes in the floodplain. This paleosol is also identified as an inceptisol (palefluvept) and corresponds to stage 2 in Kraus and Bown (1988) pedofacies classification.

Geochemistry and clay mineralogy

Geochemical analysis (Figure 3C) indicates weathering from the bottom to the top within each

paleosol. Data expressed as ratios and their interpretation include: 1) $\text{SiO}_2/\text{Al}_2\text{O}_3$ ratio indicates an increase in clay content from base to top in each paleosol; 2) $\text{SiO}_2/\text{Fe}_2\text{O}_3$ ratio supports weathering in the Bw horizon; 3) $\text{Al}_2\text{O}_3/\text{K}_2\text{O}$ ratio reveals no clear variation within each paleosol; 4) MgO/CaO ratio indicates less dolomitization from base to top of the each paleosol; and 5) $\text{Fe}_2\text{O}_3/\text{FeO}$ ratio reveals less oxidation in the lower, stacked alfisol-inceptisol (below the channel sandstone) and a more oxidation in the upper inceptisol (above the channel sandstone).

Clay mineralogy was determined on the <2 m fraction of the paleosols (Figures 3A and B; Figure 5). It is characterized by well-crystallized kaolinite and illite and by smectite as major clay mineral components. Mixed layer illite-chlorite (10-14C) and chlorite-vermiculite (14C-14V) occur as minor mineral components. The smectite component is characterized by a v/p (crystallinity) ratio (Biscaye, 1965; p. 806, Figure 1) between 0.40 and 0.62 referring to a swelling capacity between 65 to 75%.

These values better correspond to an illite-smectite mixed layer than to strictly clay mineral. Li- and K-saturations prior to XRD investigation indicate that the smectitic fraction itself is composed of a predominant neoformed (Wyoming type) montmorillonite associated with a smaller quantity of an illite-beidellite mixed layer 10-(10-14BeI).

The neoformed montmorillonite is interpreted to be the product of halmyrolytic weathering of a volcanic material or ash brought into the floodplain by fluvial process. The transformed illite-beidellite mixed layer is interpreted as the in-situ (para-autochthonous) pedogenetic degradation of a parent illite.

Because of its presence in C-horizons of the paleosols, the illite-beidellite mixed layer may be interpreted as the product of reworked paleosol (B-horizon) material during lateral migration (avulsion) of the fluvial channel. This is indicated by the occurrence of pedorelicts in the C-horizon of the upper inceptisol (above the channel sandstone). Pedorelicts include Fe-stained but consolidated and translocated plasma, and recycled glaebules.

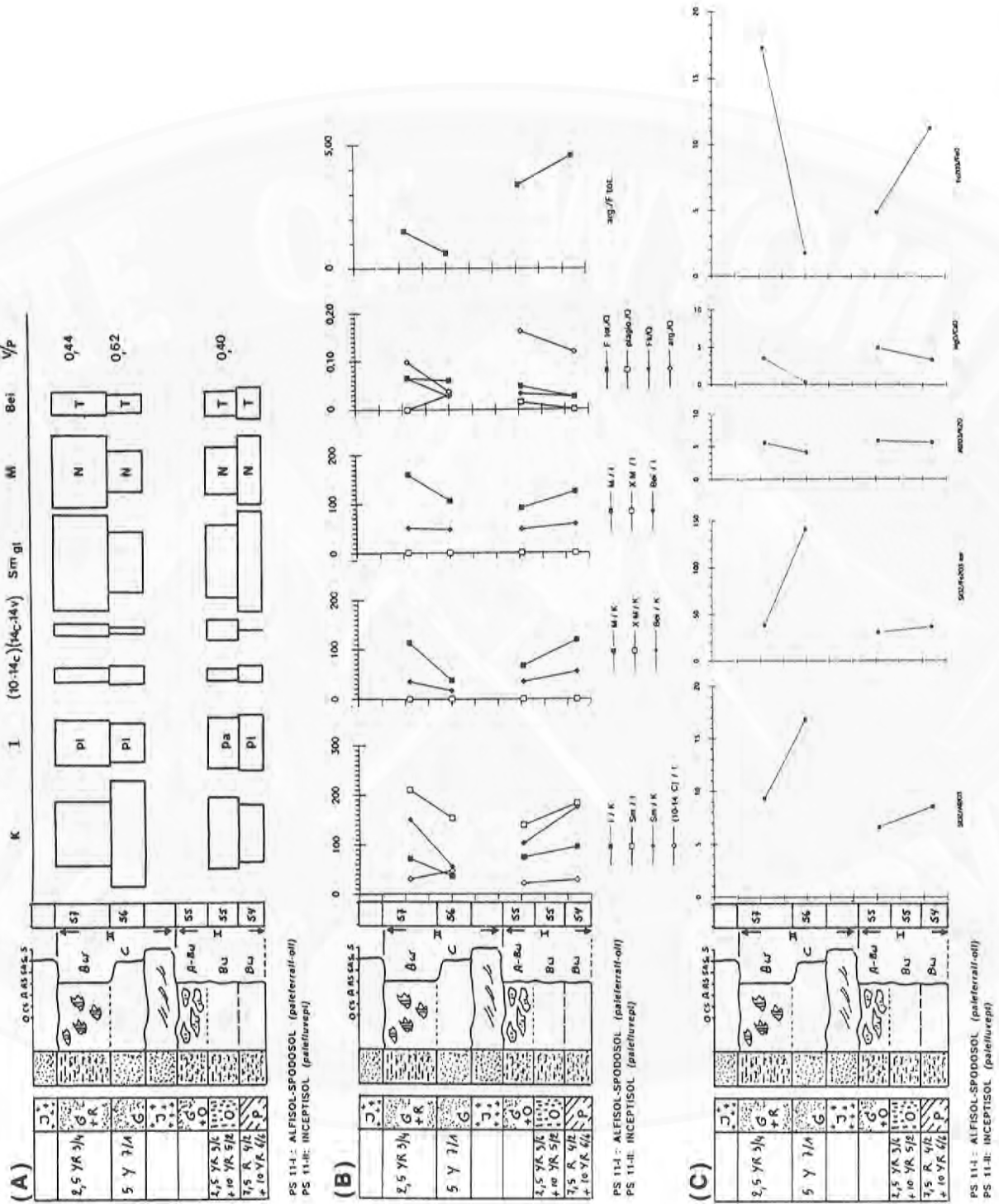


Figure 3. Lithological profiles of the Willwood Formation stacked paleosols (above and below the fluvial channel sandstone) and accompanying clay mineralogy (A and B) and geochemistry (C).

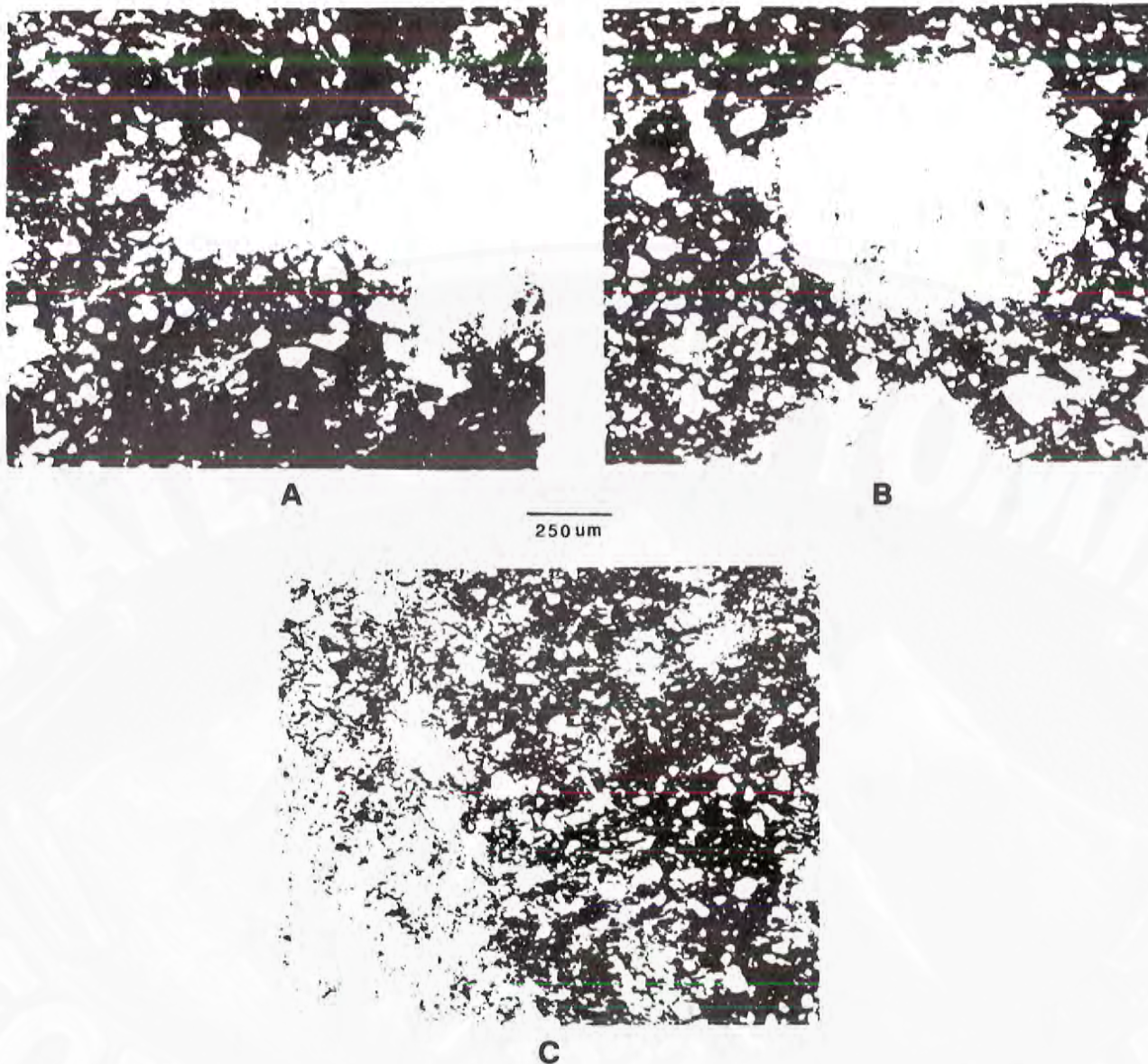


Figure 4. Photomicrographs showing the micromorphology and microstructures of the Willwood paleosols. (A) pseudogleying with decoloration process in the dark ferruginous matrix (plain light, 4x); (B) Fe-reduction spots (pseudogleying; light circular spots) in ferruginous material (plain light, 4x); and (C) diffused mottle (Fe impregnation; left side of photo) in a pedoturbated horizon (plain light, 4x).

Despite the color banding in the paleosols neither geochemical nor clay mineralogical processes indicate an important pedogenesis. This observation is in ac-

cord with the first stages of pedogenesis (stages 2 to 3) exhibited by the investigated paleosols.

Conclusion

The block diagram (Figure 6) shows the interpreted relationships of the studied paleosol types as well as their observed vertical compositional changes with respect to their settings within a fluvial system.

The truncated alfisol-spodosol (with preserved purple upper horizon below the channel sandstone) probably developed in a distal part of the floodplain, where the ground-water table fluctuated because of

Bighorn Basin - Day two, Part 3

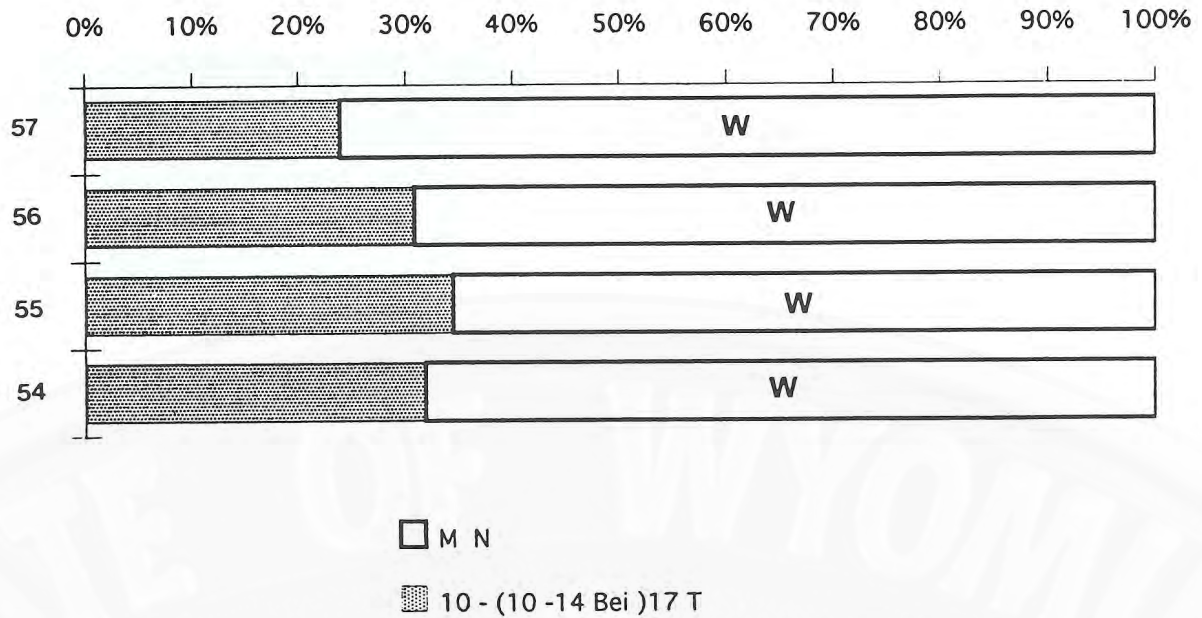


Figure 5. Qualitative combination, percentage, and genetic signature of smectite in the Willwood paleosols. M N= neo-formed smectite (W= Wyoming type). Bei= beidellite. T= transformed smectite.

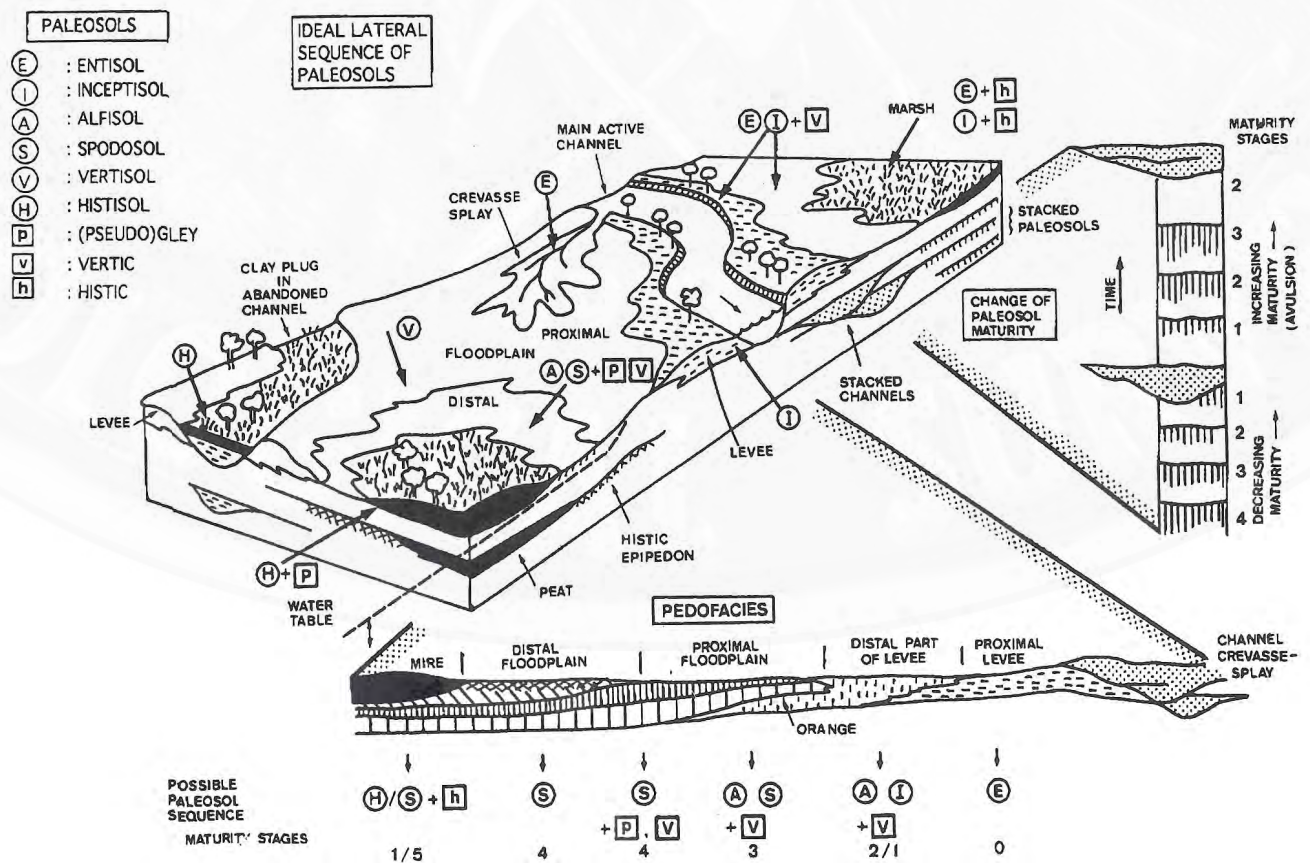


Figure 6. A block diagram showing paleosol types and predicted maturation in various settings in the alluvial plain.

changes in the local and regional paleoclimate. This paleosol was then exposed to erosion by an advancing channel migrating or avulsing. Avulsion is evidenced by channel plug deposits with superposed paleosols shown in Figure 7. Renewed sedimentation in the floodplain from the fluvial channel provided sediments that were subjected to pedogenesis, thus producing the overlying inceptisol. This inceptisol became directly superposed onto the purple horizon belonging to the truncated alfisol, which served as parent material or C horizon for the later inceptisol.



Figure 7. Stacked paleosols in the Willwood Formation above an abandoned fluvial channel deposit. The abandoned fluvial channel deposit (AFC) correlates below to the inceptisol in the studied paleosol profile.

The pedogenetic process was abruptly terminated by channelization of the floodplain as indicated by the overlying channel sandstone. The channelization indicates that the fluvial system was laterally degrading (eroding) towards the floodplain. This process also supports the change of maturity from mature (lower purple B horizon) to immature (upper grey-orange A-B horizon) stages of the two lower, stacked paleosols. Pedogenesis ensued after deposition of the fining-upward channel sandstone and subsequent

abandonment. The silty mudstone above the channel sandstone was pedogenetically transformed into a less well-developed inceptisol.

Thus, the upward-decreasing maturity of the three stacked paleosols indicates that they are directly related to the onset of channel migration towards the floodplain. This observation is in general agreement with the interpretations of Allen (1989) and Wright (1989) and in particular with the observations of Bown and Kraus (1981a), Bown (1985), and Kraus and Bown (1988) in the Willwood Formation.

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Northwestern Powder River Basin - Day three, Part 1

Road log and general geology of the northwestern Powder River Basin, Wyoming and Montana

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Introduction

Sheridan is the sixth largest city in Wyoming; the 1990 population was 13,900. The main industries are ranching and mining. Coal mines include the Big Horn, Decker, and Spring Creek mines north of Sheridan; the coal companies have their headquarters in the city. Tourism is an active industry as the city serves as a "gateway" to several recreational areas,

including the Bighorn Mountains to the west, the Custer National Battleground Monument to the north, and Lake De Smet and Tongue River Reservoirs to the south and north, respectively. The city is also a stop for tourists visiting the Yellowstone/Teton and Glacier National Parks to the west and northwest, respectively.

Road log

Mileage:		Description:
Incre- ment	Total	
0.0	0.0	Begin Day 3. Road log begins at the junction of State Highway 338 and U.S. Interstate 90 north of Sheridan. Turn north on Interstate 90. See Figure 1 for the field trip route in the northwestern Powder River Basin.
0.3	0.3	To the north, across the Goose Creek valley, are outcrops of the Paleocene Fort Union Formation. Red and orange clinker deposits (burned coal and related rocks) of the Roland coal bed are present just below the contact of the Fort Union Formation and overlying Paleocene and Eocene Wasatch Formation. To the north is the silo for the Big Horn Coal Company mine. We will be driving through outcrops of the Fort Union Formation for the next 3 mi (5 km).
2.3	2.6	To the east-northeast ("1-2 o'clock"), across the Goose Creek valley and beyond, is a hilltop pockmarked by subsidence pits (may be difficult to see without binoculars) from collapse of the underground Dietz Mines (Dunrud and Osterwald, 1980).
1.9	4.5	Take Exit 16 to Decker, Montana. Turn right on the exit road. Turn left on State Highway 339.
1.1	5.6	Junction with State Highway 338. Turn left on Highway 338.
0.2	5.8	On the right is an outcrop of a fluvial channel sandstone of the Tongue River Member of the Paleocene Fort Union Formation.
0.2	6.0	Road to the Big Horn coal mine is on the left.

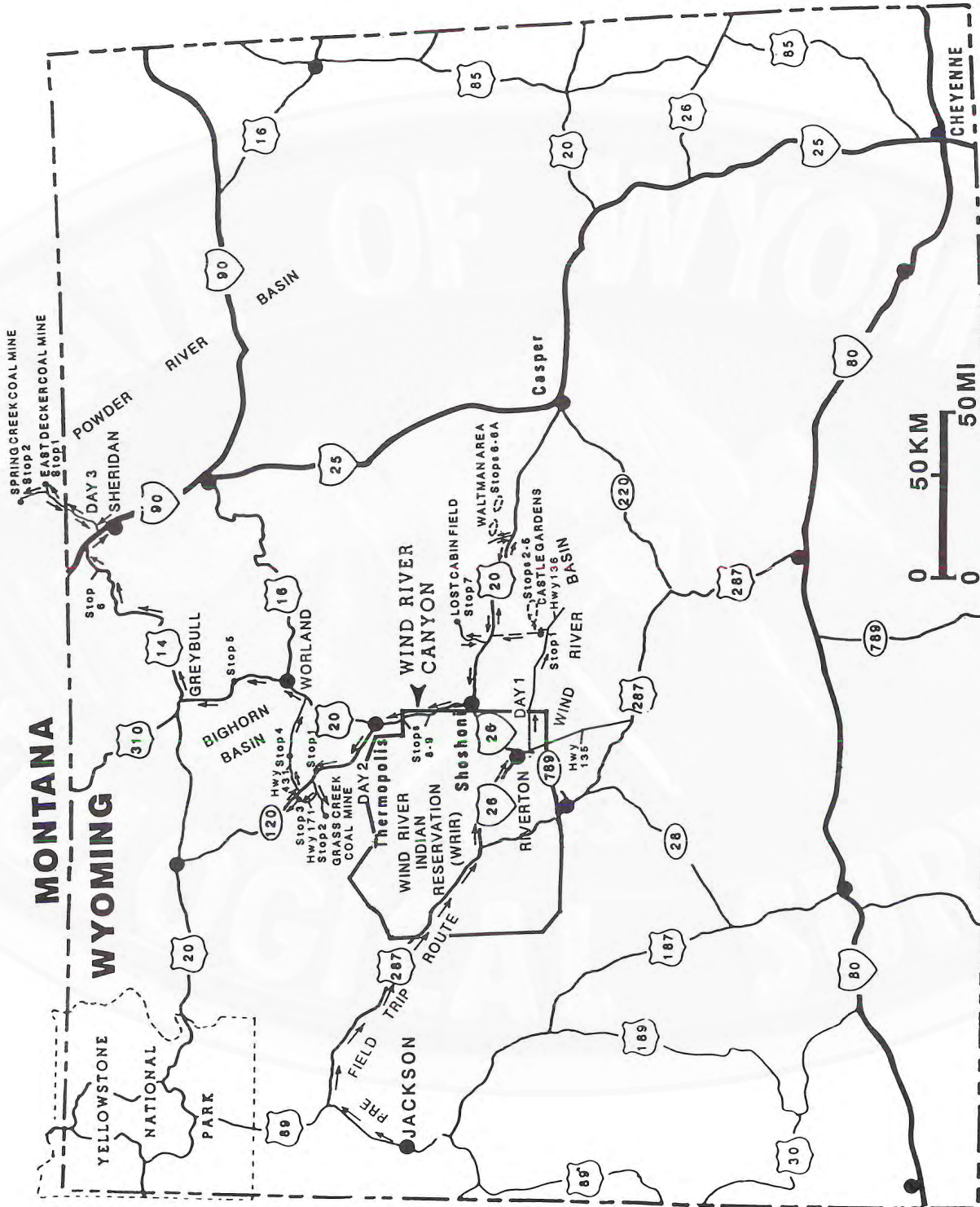


Figure 1. Map showing pre-field trip route and daily routes for the field trips in the Wind River Basin (Day one), Bighorn Basin (Day two), and northwestern Powder River Basin (Day three).

Mileage:		Description:
Incre- ment	Total	
		The Anderson and Dietz coal beds which average 20-25 ft (6.1-7.6 m) in thickness, are being mined here. These coal beds are in the upper part of the Tongue River Member of the Paleocene Fort Union Formation.
6.2	12.2	Crossing the Tongue River. The rocks over the next 6 mi (9.7 km) are in the upper part of the Tongue River Member of the Fort Union Formation (the interval above the Dietz coal bed).
4.3	16.5	State Line of Wyoming and Montana. Proceed into Montana on State Highway 314.
2.2	18.7	Decker Post Office, Montana.
2.9	21.6	Turn off to East Decker mine
5.6	27.2	Stop 1- East Decker coal mine. Operator is the Peter Kiewit Company. Coals mined in the Decker area are included in the uppermost part of the Tongue River Member of the Paleocene Fort Union Formation. The coal-bearing interval contains the merged Anderson-Dietz coal bed, which is as much as 85 ft (25.9 m) thick in the West Decker coal mine. This coal bed splits eastward in the East Decker area into the Anderson, Dietz No. 1, and Dietz No. 2 beds and each averages about 20 ft (6.1 m) in thickness. Here the interburden, which consists of sandstone, siltstone, mudstone, and carbonaceous shale, may reach 150 ft (45.7 m) in thickness. The interburden is interpreted as deposits of fluvial channel, floodplain, crevasse-splay, and lake environments (Sholes and Cole, 1981; Flores and Ethridge, 1985; Flores, 1986; Moore, 1986, 1991; Flores and Moore, this volume. Moore (1991) suggests that the surface of the peat that formed the Anderson and Dietz No. 1 coal beds was domed. The Anderson and Dietz coal beds in the Decker area average about 8% ash (dry basis), 0.5% sulfur (ash-free and moisture-free basis), and the heat of combustion ranges from 12,649 to 13,357 Btu/lb (ash-free and moisture-free basis) (Matson and Blumer, 1973). Environmentally sensitive trace elements (identified in the 1990 Clean Air Act) in the Anderson, Dietz No. 1, and Dietz No. 2 coal beds are discussed by Crowley and others (this volume).
5.6	32.8	Junction of East Decker coal mine road and State Highway 314. Turn right (north).
1.0	33.8	On the left side of the road are the strip pits and reclaimed areas of the West Decker coal mine. Here the Anderson, Dietz No. 1, and Dietz No. 2 coal beds are merged.
8.4	42.2	Junction and entrance to the Spring Creek Coal Mine. Turn left.
2.6	44.8	Stop 2- Spring Creek coal mine. Operator is Kennecott Energy Company. The Anderson, Dietz No. 1, and Dietz No. 2 coal beds in the Spring Creek coal mine are merged into one bed that is about 80 ft (24 m) thick. An isopach map of the merged Anderson-Dietz coal bed shows a north-south direction of elongation subparallel to the southwest-northeast axis of a fluvial channel complex about 5-7 mi (8-11 km) to the east (Flores, 1986; see Flores and Moore, this volume). This fluvial channel complex served as a point source for crevasse splays that aggraded along the margins of the raised mire, resulting in the splitting of the Anderson and Dietz coal beds. The combined Anderson-Dietz coal bed in the Spring Creek coal mine averages 4% ash (as-received), 0.33% sulfur (as-received), 25.5% moisture (as-received), and 9,325 Btu (as-received) (M. Robinson, Spring Creek Coal Mine Co., personal communication, 1994).
2.6	47.4	Mine entrance. Turn right (south) onto Decker highway.
4.5	51.9	Road junction. Turn left (east) to Tongue River Reservoir.
1.0	52.9	Rest Area at the Tongue River Reservoir. Lunch stop. End of field trip. Return to Sheridan via Montana State Highway 314 and Wyoming State Highway 338.
		Optional Stops on the way to Casper, Wyoming. If time permits, stops will be made to view the coals of the Eocene Wasatch Formation between Sheridan and Buffalo. Also a stop will be made to view coals of the upper (P5-P6 biozone) Tongue River Member of the Fort Union Formation north of Linch about 60 mi (97 km) north of Casper.

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Mechanisms of splitting of the Anderson-Dietz coal bed in the Decker area, Montana: a synthesis

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Introduction

Splitting of coal beds has always remained an enigma even though various concepts have been forwarded to explain their mechanisms. These mechanisms vary from the interaction of coal accumulation with depositional environments to its interaction with tectonic settings (Ferm, 1970; Weisenfluh, 1979; Ferm and Staub, 1984; Weisenfluh and Ferm, 1984). Auto-compaction of sediments underlying coal beds was also proposed as a controlling factor (Fielding, 1984). Splitting mechanisms therefore may be grouped into syndepositional (e.g., depositional environments, differential compaction of sediments, autocompaction of peat in mires) and syntectonic (e.g., growths faults, downfolding, upwarping) factors.

Distribution, geometry, continuity, and thickness of mineable coal beds such as the Anderson-Dietz

coal bed and associated coal-bed splits (Anderson, Dietz 1, and Dietz 2) in the Decker area (Figures 1 and 2) are controlled mainly by the nature of splitting. The mechanisms of splitting were not well understood even though early detailed surface and subsurface mapping and correlation have identified the splitting and merging of the Anderson-Dietz coal bed (Taff, 1909; Baker, 1929; Culbertson, 1954; Matson, 1971, Matson and Blumer, 1974; Law and others, 1979; Cole and Sholes, 1980). This paper will summarize the mechanisms of splitting of the Anderson-Dietz coal bed as has been proposed in recent various papers (Sholes and Cole, 1981; Flores, 1986; Moore, 1986, 1991).

Syndepositional mechanisms

The Anderson-Dietz coal bed (Figure 3) was initially recognized by Sholes and Cole (1981) to be split by crevasse-splay and channel-fill deposits. This work, based on subsurface data, indicated that in the vicinity of the Decker mines (Figure 1) the Anderson-Dietz coal bed contains three areas in which two or

more coal beds are merged and that away from these areas, the coal bed is split by variable thicknesses of interburden clastics (see Figure 2). These clastics were interpreted as channel deposits (thick sandstone bodies) and crevasse-splay deposits (thin sandstone bodies). These deposits were principally recognized by

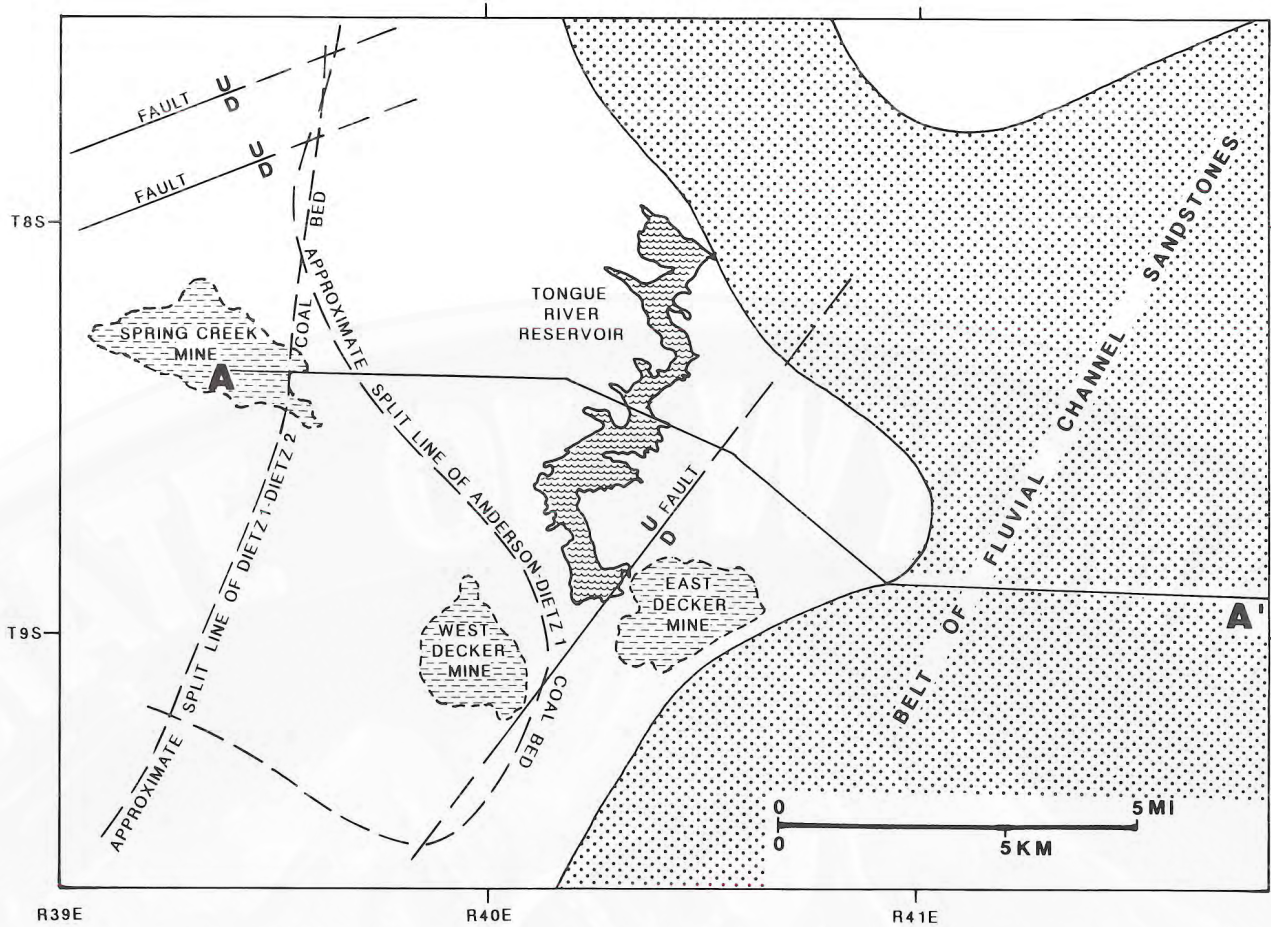


Figure 1. Map showing the locations of the Spring Creek, West Decker, and East Decker coal mines, fault traces, and a belt of fluvial channel sandstones. The line of cross section A-A' is shown in Figure 2.

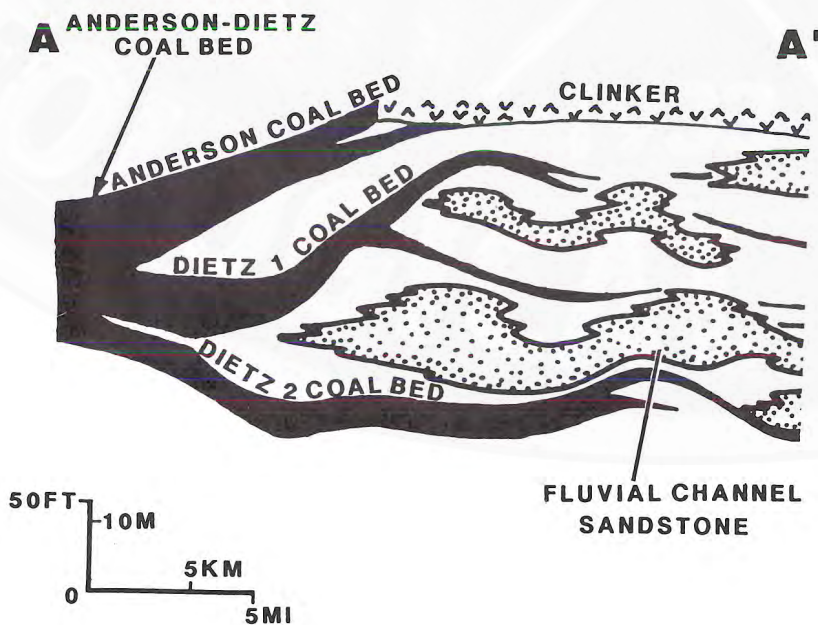


Figure 2. A cross section showing the Anderson-Dietz coal bed splitting into Anderson, Dietz 1, and Dietz 2 coal beds from west to east. Modified from Flores (1986).

their vertical change in grain size; fining upward for the channel sandstones and uniform grain size for the crevasse-splay sandstones which have a sharp base and top. However, the crevasse-splay sandstones, which are normally underlain by mudstones and siltstones in the outcrops, may cap a coarsening-upward succession. In addition, the channel sandstones were identified to be in 30- to 120-ft (9 to 37 m)-thick beds. The crevasse-splay sandstones were recognized as generally less than 10 ft (3 m) thick. These clastic partings represented deposits of a fluvial system, in the absence of which a merged coal bed would have been more than 100 ft (30 m) thick. Two areas of the merged Anderson-Dietz coal bed, which is as much as 80 ft (24 m) thick, is mined in the West Decker



Figure 3. Merged Anderson-Dietz coal bed in the West Decker mine overlain by fluvial channel sandstone.

and Spring Creek coal mines. A cross section of this coal bed and associated coal-bed splits, which are mined in the East Decker area, is shown in Figure 2. The merged coal bed splits into the Anderson, Dietz 1, and Dietz 2 coal beds from the West Decker-Spring Creek mines to the East Decker mine. In addition, the Anderson and Dietz 2 coal beds split farther eastward; both of these coal beds and Dietz 1 also thin eastward. Splitting of the Anderson-Dietz coal bed eastward is towards the direction of an area of stacked fluvial channel sandstones (Flores, 1983, 1986; see Figure 1). The coal-bed splits (e.g., Dietz 1 and Dietz 2 coal beds) thin towards this fluvial channel sandstone complex.

Subsequent investigation by Moore (1986, 1991) of the petrology and mechanisms of splitting of the Anderson-Dietz coal bed has provided additional understanding of the syndepositional factor of splitting. These mechanisms place a major emphasis on the autocompaction of peat and a minor emphasis on the subsidence of the surfaces of peat mires due to differential compaction of the underlying sediments and burning of peat. Autocompaction of peat may have been initiated by detrital influx, which drowned the mire and terminated peat accumulation. Continuing detrital deposition and sediment loading resulted in the compaction of the underlying peat. Moore (1986) suggested that the underlying peat must have compacted 4.5 times greater than the uncompacted peat in order to accommodate unconsolidated sediments such as those between the Anderson and Dietz coal-bed splits.

Compaction of sediments below the peat may have partially controlled splitting of the Anderson-Dietz coal bed (Moore, 1986). This mechanism may be responsible for the subsidence of the mire in which the Dietz 1 peat coal was accumulated. For example, where the coal bed is split, the Dietz 1 coal bed is underlain by compactible mud, and where it is merged, the coal bed is underlain by less compactible sand and silt. According to this model, the peat mire on compactible substrate was transformed into topographic lows that attracted sediment-laden floodwaters in contrast to the peat

mire underlain by less compactible substrate, which developed on a topographic high and shielded from detrital influx. The latter concept, however, does not hold true for the merged Anderson-Dietz 1-Dietz 2 coal bed, which is underlain by compactible mudstones.

Subsidence of the mire's surface may be attained by burning of peat, which creates a topographic low in which sediment-laden floodwaters or fire splays (Staub and Cohen, 1979) are diverted and collected. Staub and Cohen (1979) found that these fire splays were maintained by annual floodwaters until such time as they were abandoned due to a fluvial avulsion. Abandonment permitted reoccupation by vegetal growth on the splay deposit, which was followed by mire development and subsequent peat accumulation. Moore (1986, 1991) recognized a substantial increase in oxidized plant material (possibly charcoal) in the upper 2 ft (<1 m) of the Dietz 1 coal bed (i.e., beneath the splay sediments), suggesting burning of the peat. Recent re-interpretation by Moore and Shearer (1993), Shearer and others (in press) and Moore (this volume) suggests that this interval could also have resulted in oxidation of peat surface through microbial (principally fungi) alteration, although a fire origin can not be fully ruled out. A drop in the water table (i.e., a change in the hydrology of the peat mire) would have resulted in aerobic conditions in the upper surface of the mire, which in turn, would allow intense microbial oxidation to occur. With either fire or fungal alteration, the surface of the peat mire would

deflate (Edil and others, 1986; Staub and Cohen, 1979; Prokopovich, 1985) allowing parts of the mire to be inundated by floodwaters, some of which would deposit sediments. Areas where the Anderson-Dietz 1 is merged and the highly oxidized plant-material interval is still evident represents parts of the mire distal to sedimentation. Initial compaction of the peat in other areas with the first vestiges of sediments probably would have enhanced subsidence of peat surface from the load pressure, thus allowing even more sediments to be deposited. The interval lateral to the splay sediments in the merged Anderson-Dietz 1 coal bed have very low concentrations of oxidized plant material (see Moore, this volume). Therefore, this interval in the merged Anderson-Dietz 1 coal bed probably was flooded with aerated water (explaining its highly degraded nature) but was topographically high enough or distal enough not to be inundated with sediments.

Crevasse-splay sedimentology

A major inference from the studies of Moore (1986, 1991, this volume), Moore and Shearer (1993), and Shearer and others (in press) is that where crevasse-splays occurred, the peat surface must have subsided first, creating a topographic low into which sediments could be deposited. In addition, splay sedimentation was relatively short-lived, affecting only a small portion (aerially and chronostratigraphically) of the mire. These characteristics (i.e., infilling a hole and rapid deposition) are typical of crevasse-splaying. Investigation on the nature of splitting of the Anderson-Dietz coal bed indicates that splitting was effected by an aerially extensive and a short-lived process controlled by subsidence of the surface of the peat mires. In addition, because this mechanism of splitting was a rapid process in which only a part of the mire system was selectively affected while the rest was unaffected, fluvial inundation is indicated. Collectively, these characteristics of the fluvial process are typical of crevasse-splaying. Although Sholes and Cole (1981) applied this syndepositional mechanism

for splitting the Anderson-Dietz coal bed, the sedimentological characteristics of the crevasse-splays were not fully described and therefore will be addressed here.

The sedimentological characteristics of the a crevasse-splay deposit within the interburden clastics between the Anderson and Dietz 1 coal beds are best shown in Figure 4. The vertical succession in this interval is composed of a generally coarsening-upward interval that ranges in thickness from 35 to 50 ft (10 m to 15 m). The interval is composed of two lithofacies associations: 1) fine-grained dominated (Figure 5) and 2) coarse-grained dominated (Figure 6). The fine-grained dominated lithofacies association (see Figure 5) comprises mainly interbedded mudstones and siltstones in the lower part, which locally contains freshwater pelecypods and gastropods. These lithologies grade upward into thin-bedded rippled sandstones, thin, basally erosional, lenticular, trough crossbedded sandstones, and ferruginous limestone lenses.

The coarse-grained dominated lithofacies association (see Figure 6) consists of two sandstone types. One type of sandstone is an erosional-based body as much as 15 ft (7.6 m) thick and 75 ft (23) wide. This sandstone is multiscoured and contains large-scale trough crossbeds and rippled laminations. The other type of sandstone exhibits a sharp to gradational base, about 35 ft (10 m) thick, and is laterally extensive



Figure 4. Crevasse-splay deposit between the Anderson (A) and Dietz 1 (D1) coal beds consisting of coarse-grained (CGLA) and fine-grained FGLA dominated lithofacies associations. Photograph is in the East Decker mine.



Figure 5. The fine-grained dominated lithofacies association consisting of mudstones, siltstones and silty sandstones. The lenticular bodies are small, coeval crevasse channel sandstones (CCSS). A= Anderson coal bed; D1= Dietz 1 coal bed. Photograph is in the East Decker mine.



Figure 6. The coarse-grained dominated lithofacies association comprising flat-bedded, ripple-laminated, crevasse-splay sandstones (CCSS) and erosionally-based, trough crossbedded crevasse channel sandstone (CCSS). Photograph is in the East Decker mine.

(>2000 ft or >610 m). It is mainly ripple laminated but locally trough cross bedded. This sandstone-dominated succession grades laterally into the fine-grained dominated succession (Figure 7). The erosional-based sandstone occurs as contemporaneous, isolated, thin to thick, narrow bodies encased within the ripple-dominated sandstone type (see Figures 8A, B, and C).

siltstone cone-shaped bodies were interpreted by Cavaroc and Saxena (1979) as overbank-floodplain deposits that resulted in overflow from the major channels. In the Decker area, this lithofacies association consists of mudstones, siltstones, silty sandstones, and ferruginous limestones, which locally contain freshwater mollusk fossils. The belt of minor

The characteristics of the lithofacies associations between the Anderson and Dietz 1 coal beds are very similar to those of a large (300 mi² or 777 km² area) crevasse-splay complex between the Middle and Upper Kittanning coal beds in Ohio and Pennsylvania described by Cavaroc and Saxena (1979). This crevasse-splay complex consists of three major lithofacies that are present in the Anderson and Dietz 1 coal-bed interval, which include: 1) major channel sandstones, 2) a belt of minor channel sandstones leading from the major channel sandstones, and 3) interchannel siltstone cone-shaped bodies. Similar characteristics in the Carboniferous coal-bearing rocks of Kentucky were described by Baganz and others (1975), which they compared to recent Mississippi lower delta plain crevasse-splays (Figure 9).

The major channel sandstones were interpreted by Cavaroc and Saxena (1979) as fluvial channels that initially entered the flooded mires. These fluvial channels were infilled by sandy deposits and may be represented by the thick fluvial channel sandstones originally recognized to occur in the Decker area by Sholes and Cole (1981) and verified by Moore (1986; see Figure 8).

The interchannel

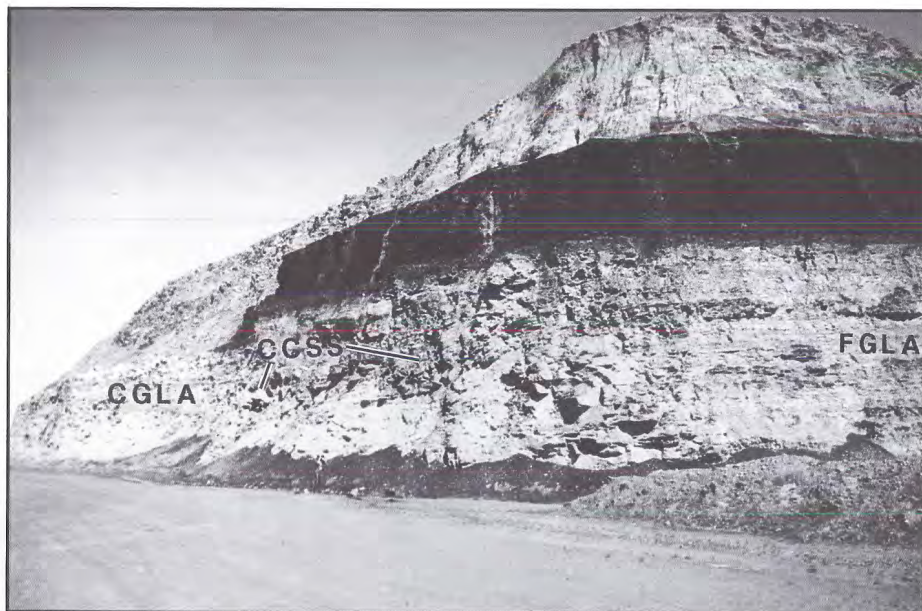


Figure 7. The coarse-grained lithofacies association (CGLA) grades laterally into the fine-grained lithofacies association (FGLA). CCSS= crevasse channel sandstones. Photograph is in the East Decker mine.

channel sandstones, which occupy peripheral areas to the major channels, formed later than the two lithofacies associations by crevassing from the major channels. Thus, this lithofacies association consists of floodplain mudstones and siltstones grading upward into sandstones of the crevasse-splay.

In the Anderson and Dietz 1 coal bed interval the crevasse-splay sandstones comprise an erosional-

based, trough crossbedded, thin to thick, narrow channel deposits enclosed by rippled-laminated silty sandstones. The closely-spaced, coeval, narrow channels are interpreted as diverging and served as pathways of floodwater sediments that drained into the floodplain lakes as evidenced by the freshwater mollusks fossils. The final episode of the crevasse-splay system was subsidence (e.g., due to peat auto-compaction, differential compaction of sediments) and encroachment of outlying mire. The occurrence of mire and subsequent peat accumulation on top of the crevasse-splay system

led to splitting of the coal beds. That is, the peat underlying the crevasse-splay deposit accumulated in the mire that occupied the floodplain prior to formation and progradation of the splay. The peat overlying the crevasse-splay deposits accumulated in the mire after the crevasse-splay had been abandoned and subsided. The two peat deposits merge at the margin of the crevasse-splay; hence the splitting of the coal bed.

Syntectonic mechanisms

Sholes and Cole (1981) suggested that the splitting of the Anderson-Dietz coal bed may have been influenced by yet unrecognized structural control. They recognized that the western and northeastern Decker areas were stable and the southeastern area was unstable. Northeast-striking normal faults were mapped by Law and Grazis (1972) in the Decker quadrangle. Robinson and Barnum (1986) interpreted these faults and associated normal faults in south-central Montana as the surface expression of a left-lateral movement along basement wrench faults. Although the wrench faults are Precambrian features that have undergone reactivation (Lorenz, 1984), no precise movement can be interpreted of the associated normal faults except that some may have formed during or after late Paleocene time (Robinson and Barnum, 1986). Two northeast-striking normal faults are observed in the northwest part of the Decker area where

the Anderson-Dietz coal bed is not split (see Figure 1). A northeast-striking normal fault is also found in the southeast part of the area where the Anderson-Dietz is split. The area between these sets of normal faults is an upthrown block.

A structure contour map by Moore (1986) of the base of the Dietz 1 and Anderson coal beds in the southeast part of the Decker area shows a regional dip to the southeast with coal-bed displacements along the normal fault. This suggests post-depositional faulting. However, Moore (1986) indicated that in most places in the Anderson-Dietz coal bed, the Dietz 2 coal bed shows a rapid elevation drop in the zone of coal-bed splitting, yet the overlying Dietz 1 coal bed displays a lack of displacement along the line of splitting. At some localities the Dietz 1 coal bed also exhibits a rapid elevation drop; however, the overlying

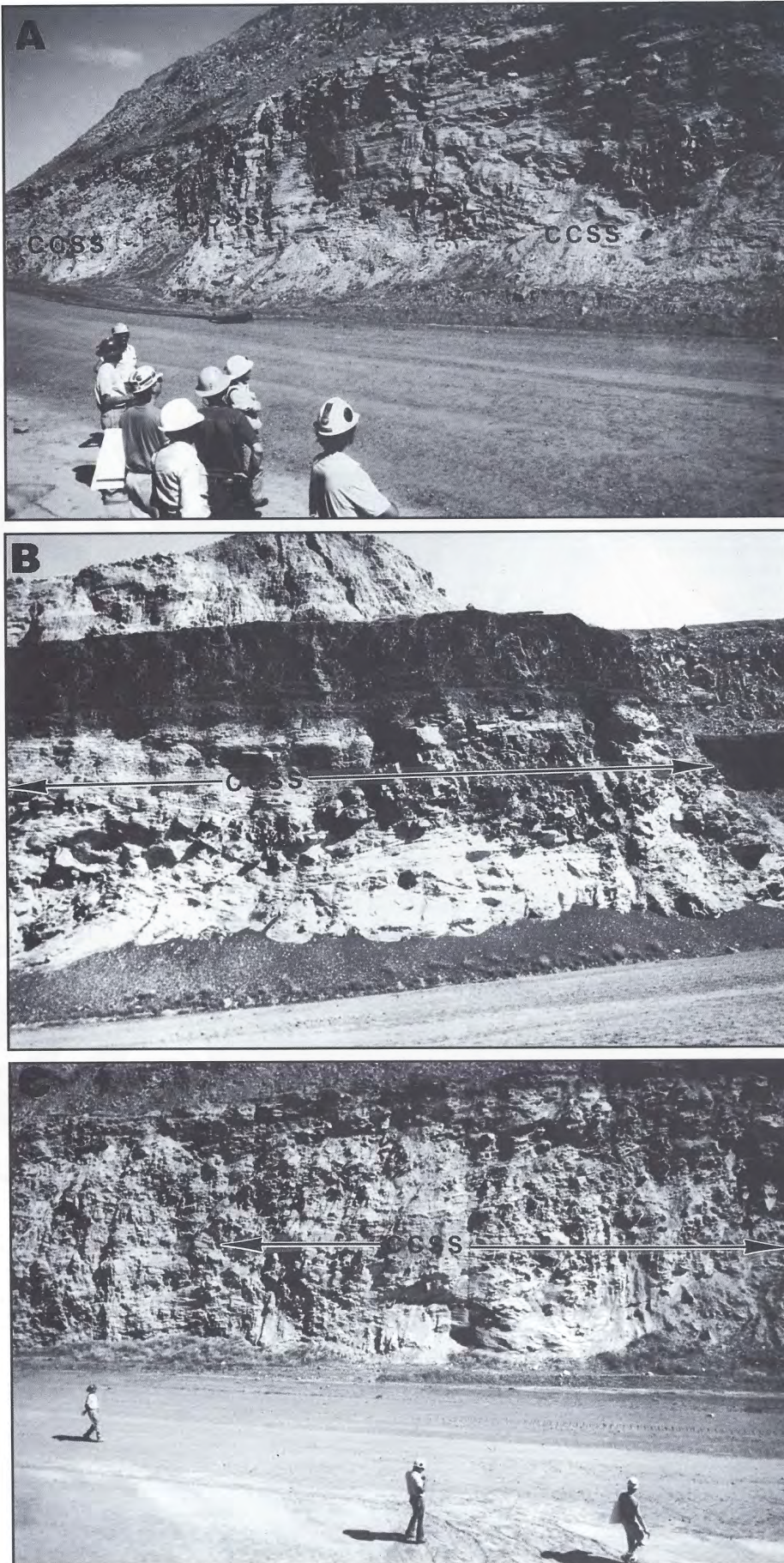


Figure 8. (A) Internal architecture of the coarse-grained lithofacies association consisting of thin to thick, narrow to wide, coeval crevasse channel sandstones (CCSS) protruding like fingers out of the mine highwall. (B) A highwall showing thick, narrow CCSS. (C) A highwall showing a thick, wide CCSS. Photograph is in the East Decker mine.

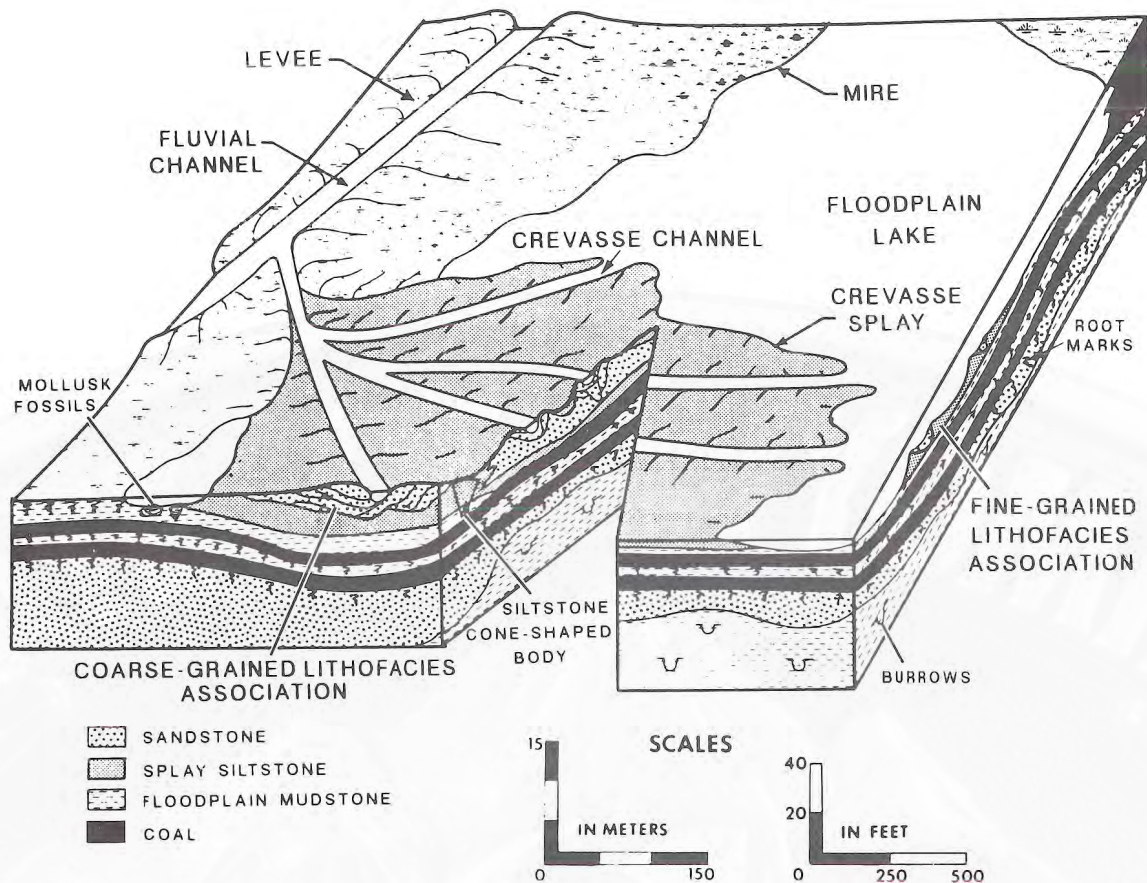


Figure 9. A block diagram of crevasse-splay deposits showing cone-shaped aerial distribution and vertical profile, and diverging and coeval crevasse channels. Modified from Baganz and others (1975).

ing Anderson coal bed lacks displacement along the line of splitting. These observations suggest that if growth faulting occurred during deposition, it was episodic, affecting splitting of the Dietz 1 and Dietz 2 coal beds and Anderson and Dietz 1 coal beds locally.

If subsidence was locally controlled by minor growth faulting, downthrown areas were prone to more rapid inundation by the fluvial system (e.g., crevasse-splays) than on the upthrown blocks or in stable areas.

Summary

In summary, the mechanisms of splitting the Anderson-Dietz coal bed may be grouped into syndepositional and syntectonic factors. The syndepositional factors are mainly related to the depositional environment and subsidence due to coal autocompaction (from fire or microbial alteration),

and/or differential compaction of sediments. The depositional environment appears to be an important controlling factor with the crevasse-splaying process playing a major role of splitting. The syntectonic mechanism mainly by selective local growth faulting was a very minor factor.

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Organic compositional clues to a stacked mire sequence in the Anderson-Dietz 1 coal bed (Paleocene), Montana

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Introduction

A paradigm in coal petrology is that composition is the sum of floral, degradational, climatic, and burial influences. Early work on coal composition stressed the botanical origin of constituents (e.g., Seyler, 1928; Thiessen, 1937) while later work often focused only on industrially relevant parameters such as rank and volatile-matter content (see Stach and others, 1982). More recently however, a number of studies have focused on explaining compositional variation in coal as related to specific parameters, such as geomorphology of the mire, degradational regime (i.e., acidity), climate, clastic-sedimentation patterns, and floral succession (e.g., Teichmüller and Teichmüller, 1968; Cecil and others, 1985; Warwick, 1985; Esterle and Ferm, 1986; Hagemann and Wolf, 1987; Littke, 1987; Warwick and Stanton, 1988; Eble and Grady, 1990; Staub, 1991; Staub and others, 1991; Demchuk and Moore, 1993; Shearer and Moore, in press). Undoubtedly organic compositional variations noted in coal beds are a result of the interplay of some or all of these influential factors and deciphering the relative effects of each factor is a challenging endeavor.

Models proposed by Staub (1991), Moore and Shearer (1993) and Shearer and others (in press) cite

the necessity for recognition of mire sequences in coal beds. They propose that variations in composition become ascribable to specific processes and/or linked processes. For example, one of the most conspicuous and important variations often noted in coal is that of oxidized plant material (e.g., inertinite, using maceral terminology). The model of stacked mire sequences proposed by Shearer and others (in press) suggests that some laterally extensive layers of oxidized plant material may reflect a period of mire demise. In addition, Shearer and others (1994) suggests that layers of highly degraded plant material may represent demise of a mire. There are examples of both these types of layers in the Anderson-Dietz 1 coal bed.

The purpose of this paper is to examine in detail the organic compositional variation seen in the Anderson-Dietz 1 coal bed in the context of a stacked-mire sequence. One important aspect of the Anderson-Dietz 1 coal bed is that there is extensive petrographic and sedimentologic data (see Flores and Moore, this volume) which allow the relative effect of many influences on the peat (e.g., climate, degradation, floral) to be assessed.

Method of analysis

Two cores from the Anderson-Dietz coal zone (Paleocene) were collected from an area in southeastern Montana in the Powder River Basin (Figure 1). The Anderson-Dietz coal zone (Moore, 1986, 1991)

consists of three coal beds (in descending order, the Anderson, Dietz 1 and Dietz 2), which are merged in the west and split by wedges of sediments to the east (Figure 2).

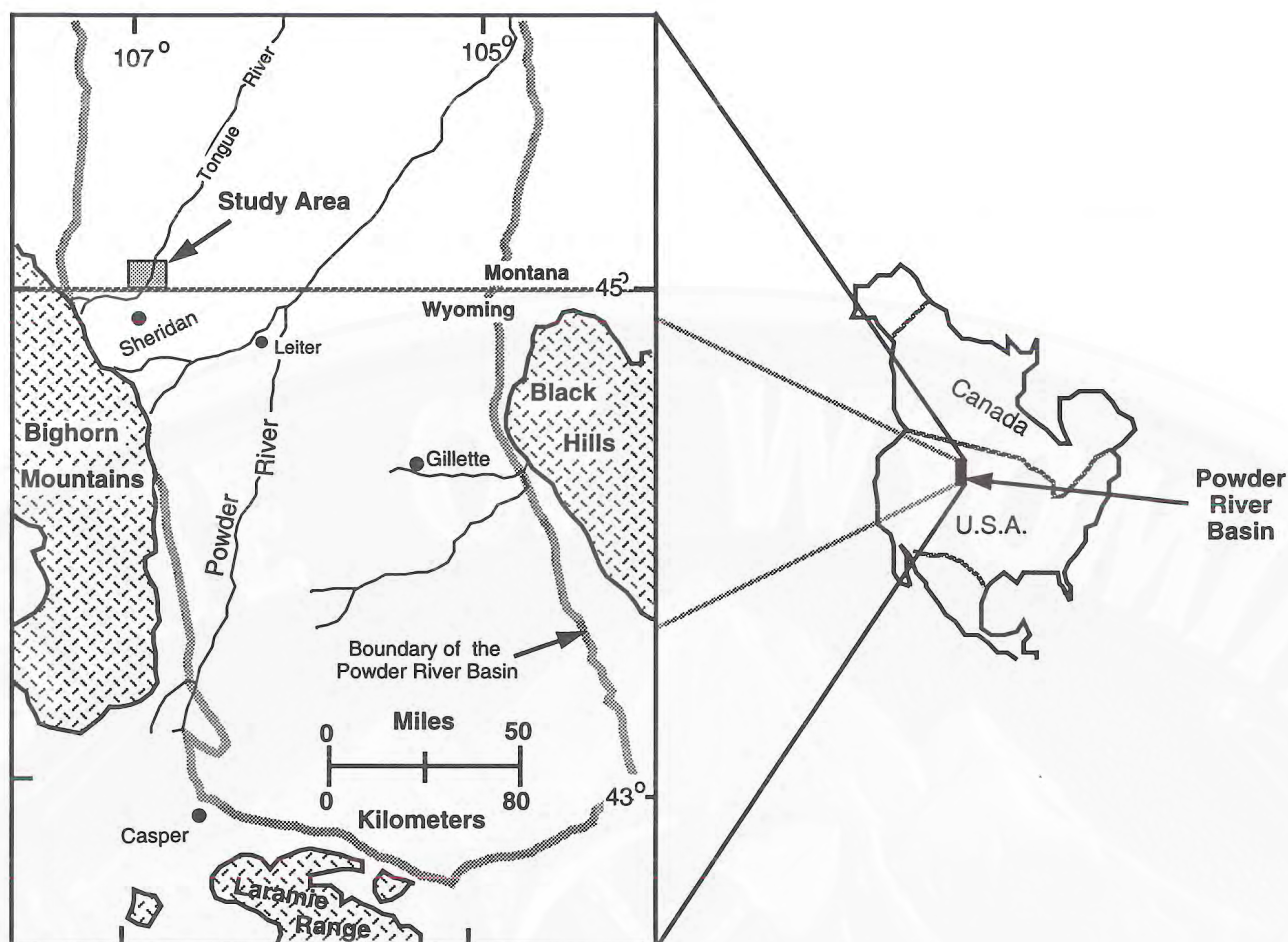


Figure 1. Index map of the Anderson-Dietz 1 coal bed study area.

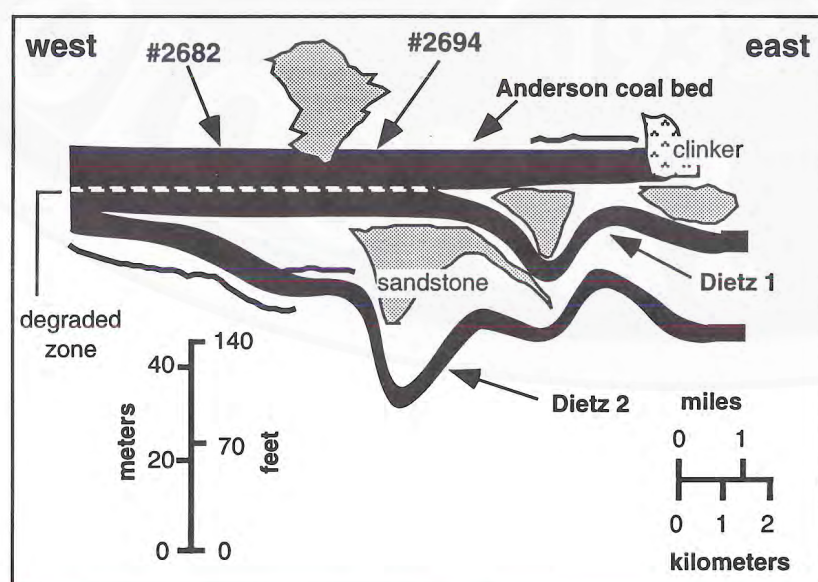


Figure 2. Cross section through the Anderson-Dietz coal zone showing location of cores and degraded zone.

Two cores located in the western part of the study area (cores 2682 and 2694) were sampled at 2.0 ft (0.61 m) intervals to define vertical variations. Samples were crushed, split, and processed for petrographic and chemical analysis. The inorganic constituents in the Anderson-Dietz 1 coal bed were examined to determine the effect of clastic splitting of the coal in the eastern part of the study area. Total sulfur and ash yield were obtained for samples from both cores. For a complete explanation of the sampling and analytical methodologies see Moore (1991).

A total of 54 samples were prepared for petrographic analysis using particulate pellet mounts. These pellets were then etched with a solu-

tion of potassium permanganate, sulfuric acid, and water following procedures outlined in Pontollilo and Stanton (in press). Etching was conducted to reveal differences between tissues, amorphous humic gels and fragments of cell walls and cell fillings. Petrographic analyses were originally performed using maceral nomenclature (Moore, 1986, 1991) but are reported here as botanical components. For the purpose of this paper only 10 components will be discussed as these account for most of the compositional variation seen in the Anderson-Dietz 1 coal bed. These components are given in Table 1 along with their corresponding maceral terminology.

Table 1. Organic components identified in analysis and the corresponding low-rank maceral terminology*.

Organic Component	Maceral Term
Tissue	Humotelinite
Fragments of cell walls	Detrogelinite
Cell fillings in matrix	Corpohuminite
Amorphous bitumen	Bituminite
Crack-filling resin	Exsudatinite
Resin in matrix	Resinite
Fragments of 'resinous' material	Liptodetrinite
Slightly oxidized cell walls	Semifusinite
Highly oxidized cell walls	Fusinite
Fragments of oxidized cell walls	Inertodetrinite

*from Stach and others (1982)

Compositional variation

The major variations noted in the Anderson-Dietz 1 coal bed are in the following inorganic (Figure 3) and organic components (shown in Figure 4 as vertical profiles in cores 2682 and 2694): 1) ash yield, 2) sulfur yield, 3) percentage of cell fillings plus cell wall fragments, 4) percentage of oxidized material, 5) percentage of fluorescing components, and 6) tissue preservation. These variations are described in the paragraphs below.

Total sulfur and ash yields for vertical ply samples in the Anderson-Dietz 1 coal bed increase at the top and bottom of the bed. There is also a peak in the center of the bed, which corresponds to the

top of the Dietz 1 coal bed. Although there is only a slight increase in ash yield in this peak (2 or 3%), sulfur content increases substantially (from < 0.5% to 1%). Laterally, this same interval is represented by clastic detrital sediments deposited by a crevasse-splay event (Moore, 1991; Flores and Moore, this volume). The high ash and sulfur contents of the coal probably indicate that the interval was affected by this crevasse clastic deposition from the east (see Flores and Moore, this volume). Only a small amount of sediments reached the area of peat accumulation (sampled coal), but it appears that the near-neutral pH flood waters allowed pyrite formation (Moore, 1991), increasing the sulfur content of the coal.

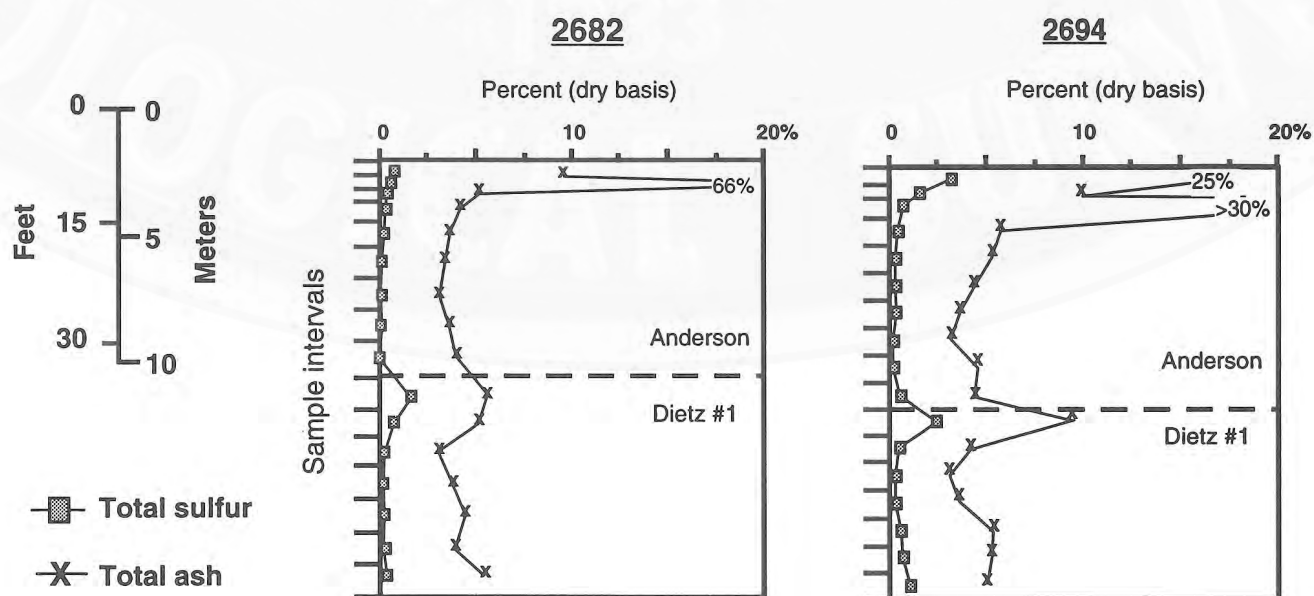


Figure 3. Total ash and sulfur content (dry basis) for two cores in the Anderson-Dietz 1 coal bed (see Figure 2 for location).

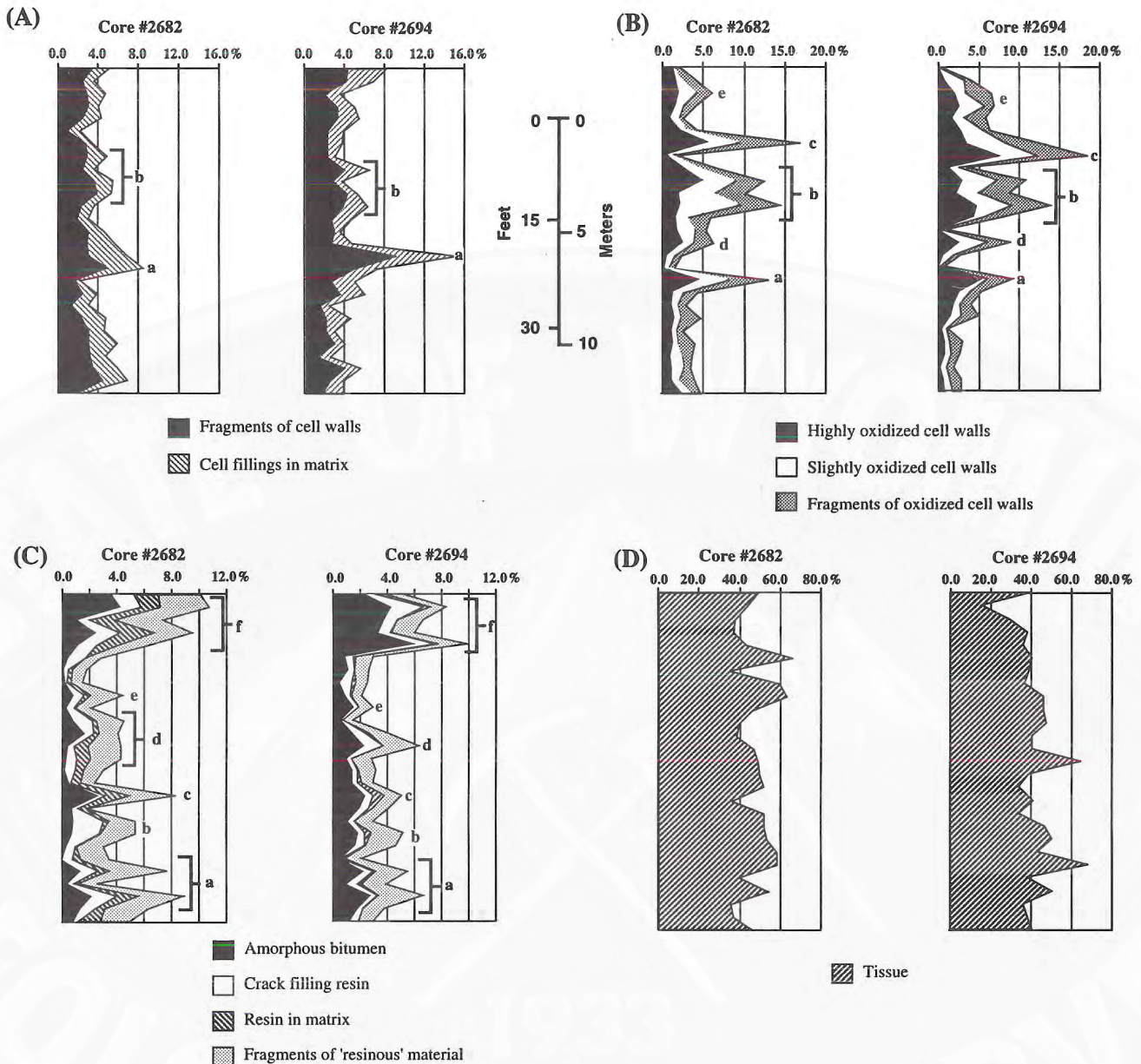


Figure 4. Organic compositional variation in two cores of the Anderson-Dietz 1 coal bed (see Figure 2 for location).

One of the most obvious significant variations of organic components in the Anderson-Dietz 1 coal bed is in the proportions of cell fillings and cell-wall fragments found in the matrix. These two components tend to vary concomitantly, so they are plotted together (Figure 4A). Photomicrographs of typical examples of these components are shown in Figure 5. The sum of these two organic components form a conspicuous peak (labeled 'a' in Figure 4A) approximately in the center of the Anderson-Dietz 1 coal bed. This peak corresponds laterally to a stratigraphic interval where the coal bed is split by detrital sediments (see the cross section in Figure 2) and to the zone of sulfur and ash increases noted above. The proportion

of cell wall fragments and cell fillings is higher in core 2694, which is closer to the sediments, than in core 2682. Two less conspicuous peaks in the proportions of cell fillings and cell wall fragments occur above the peak marked 'a' in Figure 4A. This couplet of peaks, labeled 'b', is laterally correlated between the two cores but no sediments occur at these intervals within the study area.

The proportions of oxidized cell walls and cell wall fragments in the Anderson-Dietz 1 coal bed form three major peaks (labeled 'a' to 'c') and two minor peaks (labeled 'd' and 'e' in Figure 4B). Representative photomicrographs of the types of oxidized cell

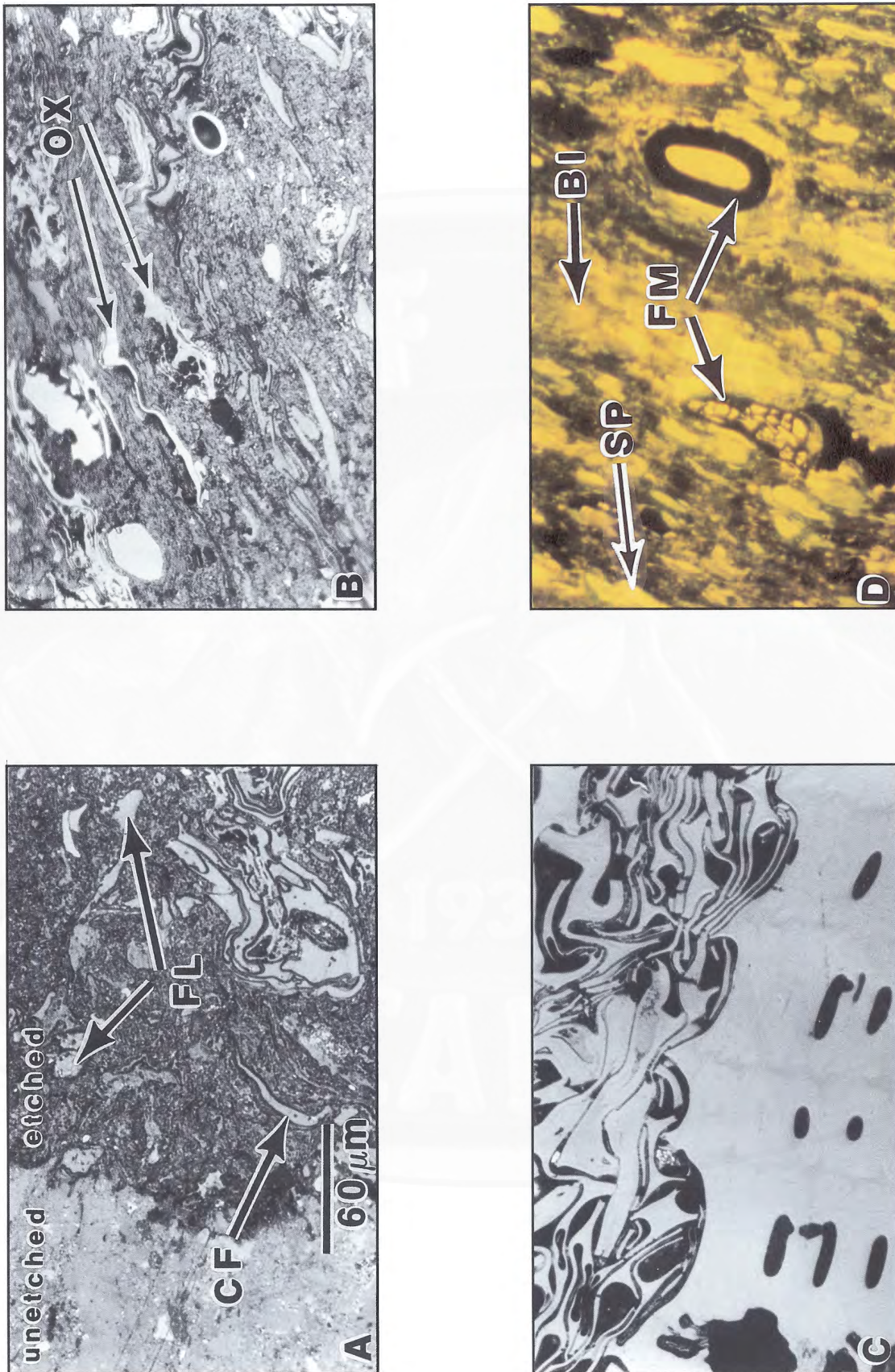


Figure 5. Photomicrographs of organic components in the Anderson-Dietz 1 coal bed. Scale bar applies to all photomicrographs. (A). Unetched and etched surface showing cell wall fragments (CF) and cell fillings (FL). Black ground mass between particles is fossilized colloidal humic gels. (B). Oxidized fragments (OX) of matrix in fossilized amorphous humic gels. (C). Oxidized cell walls. (D). Blue-light illumination showing amorphous bitumen (BI), fungal material (FM) and spores/pollen (SP).

wall components seen in the Anderson-Dietz 1 coal bed is shown in **Figures 5B and C**. Of the major peaks in the oxidized material, the stratigraphically lowest ('a') occurs at what has been inferred to be the top of the Dietz 1 coal bed (Moore, 1991). Peak 'b' is actually a couplet containing from 11 to 14% oxidized plant remains. This interval occurs in the middle portion of the Anderson coal bed and is nearly identical in both cores. The uppermost major peak ('c') is relatively thin (<2 to 4 ft or <0.61 to 1.2 m) in both cores and has the highest proportion of oxidized plant remains of any of the peaks (approximately 17% in both cores). The two minor peaks ('d' and 'e') are generally less than 10% oxidized material and occur at the same intervals in both cores. Finally, it is interesting to note that the proportion of oxidized plant remains is only 1 or 2% in the interval, which corresponds to the high concentration of cell wall fragments and cell fillings previously described.

Fluorescing material, including amorphous bitumen, resin found in matrix, secondary resin filling cracks and unidentifiable fragments, tend to vary in the same way and so are plotted together in **Figure 4C**. Photomicrographs of the major fluorescing components are shown in **Figures 5D**. Although there are a number of peaks (labeled 'a' through 'f'), which appear to be laterally correlative between cores, the ma-

ior variation is that these components are generally high at the top and bottom of the bed. Amorphous bitumen can have concentrations as high as 6% and overall the sum of all the components can reach concentrations greater than 15%, particularly at the top of the cores. The substantial increase in fluorescing material at the top of both cores could have just as much to do with changing floral types as with degree of degradation. Unfortunately there is presently insufficient pollen data to test either of these possible origins.

Tissue preservation is relatively high, averaging approximately 40% (**Figure 4D**), but neither vertical or lateral trends can be noted. The generally high concentration of preserved tissues can be ascribed to abundant vitrain bands noted in the coal face (**Figure 6**). Although no palynological analysis has been conducted in the Anderson-Dietz 1 coal bed, studies in other areas of the Powder River Basin show a great abundance of *Glyptostrobus* pollen (Kremp and others, 1956; Spindel, 1975; Satchell, 1985; Pocknall and Flores, 1987). This indicates that this taxa (which is a gymnosperm) was most likely a significant contributor to the peat matter. Gymnosperms are well known to be resistant to decay preferentially over that of angiosperms and therefore would likely be the source of the abundant vitrain bands (tissue) in the coal (Shearer and Moore, in press).

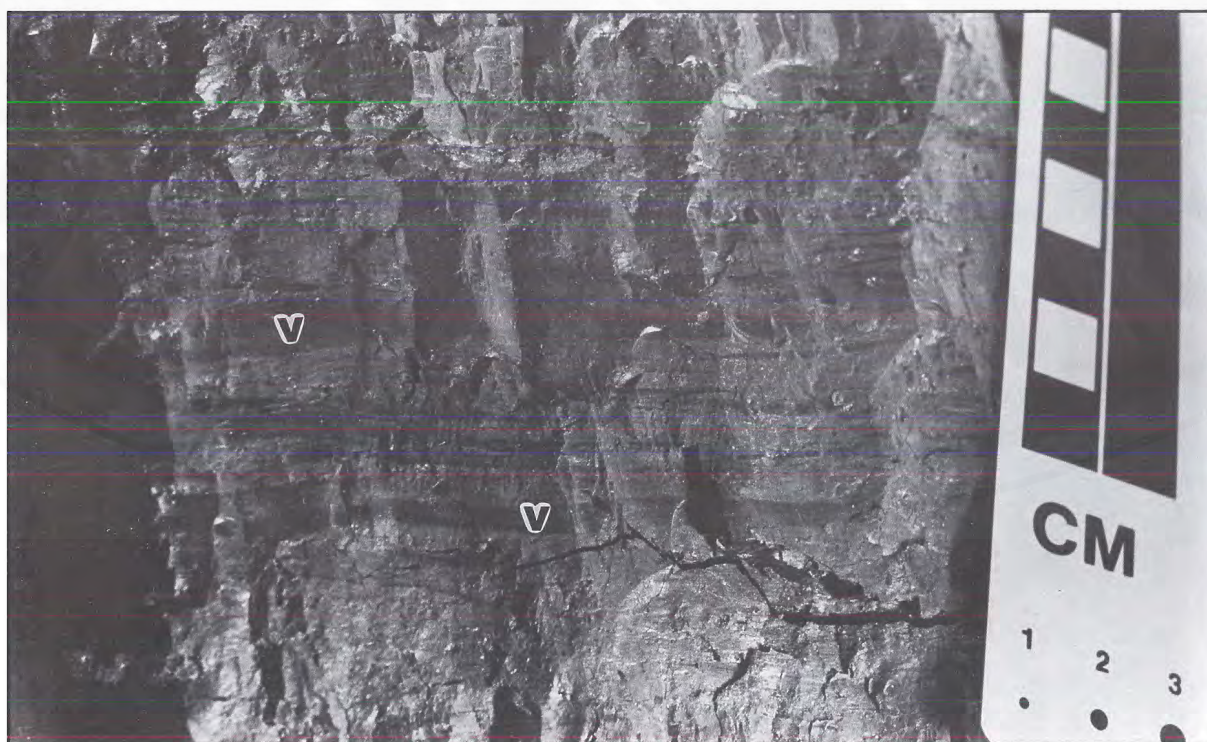


Figure 6. Macroscopic photograph illustrating the banded nature of the Anderson-Dietz 1 coal bed. "V" marks position of vitrain bands which are preserved plant tissue.

Vertical mire stacking

The compositional variation noted in the Anderson-Dietz 1 coal bed can be interpreted in the context of the models proposed by Moore and Shearer (1993) and Shearer and others (in press). These models suggest that most thick coal beds are actually vertically stacked-mire sequences. Laterally extensive "partings" within the coal beds represent demarcations between genetically different mire systems and these "partings" may be either inorganic or organic in nature. Two types of organic partings proposed by Moore and Shearer (1993) and Shearer and others (in press) are laterally extensive layers composed of: 1) oxidized plant material and 2) highly decomposed but non-oxidized remnants of plant material (i.e., fragments of cell walls and cell fillings, bitumen, etc.). The peaks in oxidized plant material and fragments of cell walls and filling in the Anderson-Dietz 1 coal bed can be interpreted as process-linked demarcations which separate different mire systems in a vertically stacked sequence.

The lower part of the Anderson-Dietz 1 coal bed (equivalent to the Dietz 1 coal bed where the coal zone is split) may represent a single mire. It is relatively uniform and low (<5%) in its oxidized plant material concentrations, except near the top where abundances rise to over 10%. This unit of high oxidized material is present as disseminated particles of plant material and not as a fusain layer. Although formation of these oxidized components through combustion can not be discounted (e.g., see Scott, 1989), the gradual rise in their concentration evidenced in core 2694 could indicate a fungal origin. Prokopovich (1985) and Moore and Shearer (1993) argue that a drop in water table in a peat environment will produce aerated conditions conducive to fungal oxidation of peat resulting in deflation of the peat surface. A drop in water table would cause peat to be oxidized most intensely at the top with a decreasing effect of alteration downward to the new position of the water table. During this alteration, peat material would actually leave the system as methane gas and in aqueous solutions due to degrada-

tion, which with the lowered water table, would result in deflation of the peat surface. Consequently, the lowered surface of the peat would become more prone to floods.

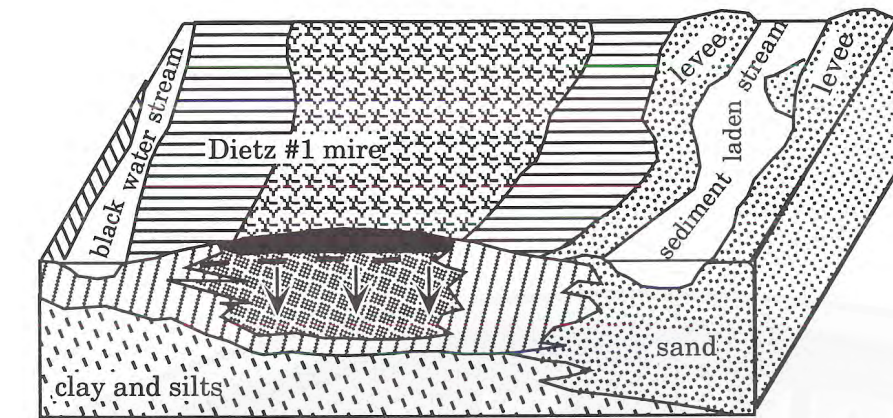
Such a flood event appears to have occurred, as indicated by the concentration of degraded plant material (cell walls and fillings in matrix, amorphous bitumen) and high ash and sulfur, immediately above the oxidized zone in the Anderson-Dietz 1 coal bed. It appears that the peat was degraded for a longer than normal period in a non-oxidizing manner after flooding, possibly by bacteria as fungi do not live subaqueously. The continued degradation suggests that the mire was already dead. To the east (Figure 2), flood waters deposited clastic sediments over the peat, showing without doubt that peat accumulation ceased temporarily. However as suggested by Moore (1991) and Flores and Moore (this volume), the crevasse-splay event responsible for these sediments was short lived. Therefore, peat accumulation resumed, after a time, both over the splay sediments as well as over the highly degraded areas of the Dietz 1 peat. The new mire probably had floristic and hydrologic characteristics quite different from the pre-flood mire and can be considered a genetically different peat system.

Other peaks in oxidized plant material in the upper part of the Anderson-Dietz 1 coal bed (equivalent to the Anderson coal bed in the split coal zone), probably also represent demarcations between genetically different mire systems. If the model proposed by Moore and Shearer (1993) and Shearer and others (in press) is correct then each peak in oxidized plant material could represent a drop in water table with subsequent oxidation and then deflation of the mire's surface. However, unlike the top of the Dietz 1 coal bed, the other peaks in oxidized plant material are not followed by an interval of high concentrations of cell wall fragments and cell fillings. This may indicate that the mire in these places did not deflate to a degree that resulted in floods from surrounding fluvial systems.

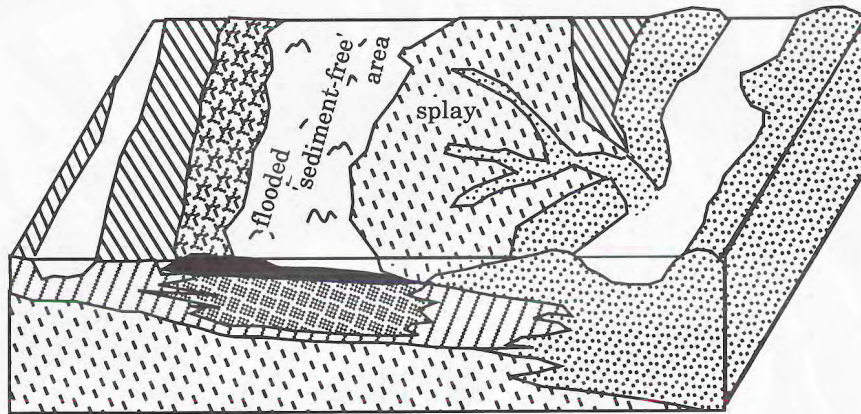
Conclusions

Compositional variations of organic constituents in the Anderson-Dietz 1 coal bed can be related to associated processes, as shown in Figure 7. First, high concentrations of oxidized plant material represent

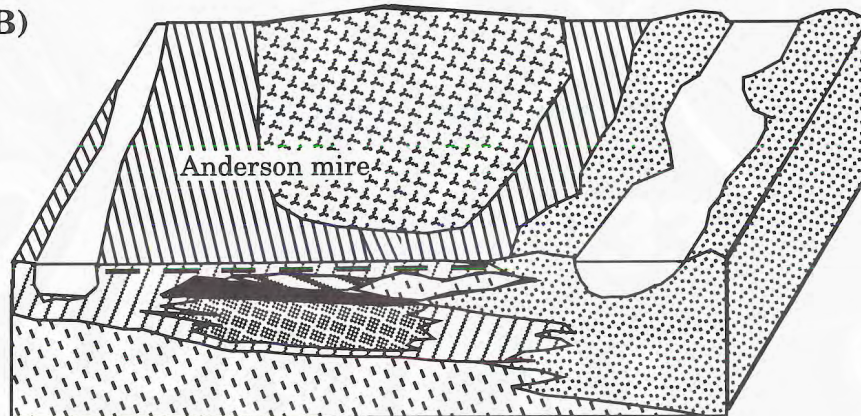
stages of lowered water table in the mire (Figure 7A). A lowered water table resulted in combustion or fungal alteration of the peat, or both. In any event, the surface of the mire was deflated and the topographi-



(A)



(B)



(C)

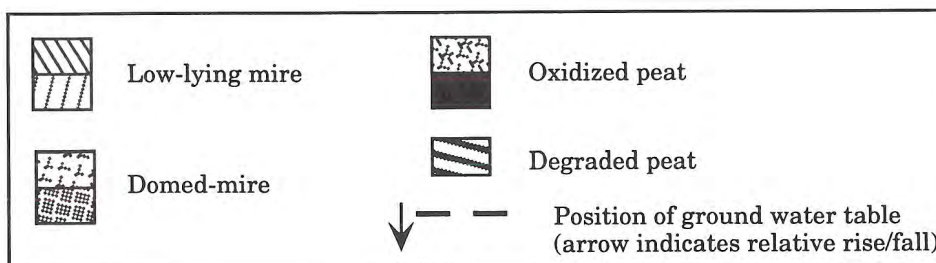


Figure 7. Model for the formation of the Anderson-Dietz coal bed (see text for explanation).

cally low surface became prone to flooding (Figure 7B). It is likely that breaching of levees and accompanying crevasse-splay processes inundated the lowered mire surface. Areas more proximal to the crevasse-splay were covered with sediment while areas more distal were only flooded with sediment-poor waters. During this high pH water regime a peat accumulated, which was generally highly degraded. After sedimentation ceased, subsequent peat accumulation resulted in the formation of a genetically different mire (the Anderson coal bed), which formed over the top of both the crevasse-splay sediments and the highly decomposed top of the Dietz 1 peat mire (Figure 7C). Subsequent oxidizing periods in the Anderson peat mire may have resulted in a deflation of the peat surface but apparently not to a degree which allowed deposition of clastic sediments. Variations seen in fluorescing material in the Anderson-Dietz 1 coal bed may be related to either floral variations or degradational processes affecting the peat; whereas, the high overall tissue content is related to the abundance of decay-resistant gymnosperms in the peat-forming flora.

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Geochemical studies of the Anderson-Dietz 1 coal bed, Powder River Basin: Origin of inorganic elements and environmental implications

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Introduction

The Anderson-Dietz coal bed of the northwestern Powder River Basin (PRB) (Figure 1) is a thick (>65 ft or 20 m), subbituminous coal in the Tongue River Member (Paleocene) of the Fort Union Formation. This coal bed is merged in the West Decker mine area and is split, from bottom to top, into the Dietz 2, Dietz 1, and Anderson coal beds in the East Decker mine area (Figures 2 and 3). Merging of these coal beds occurs progressively westward into Anderson-Dietz 1 and Anderson-Dietz 1-Dietz 2 coal beds. In this study, we examined the inorganic geochemistry of the merged Anderson-Dietz 1 coal bed obtained from three 49-ft (15-m) cores located in the West Decker mine area near Decker, Montana (Figure 2). In this paper, the Anderson-Dietz 1 coal bed refers to this merged part of the coal bed. The objective of the first part of this study was to examine the distribution of 46 inorganic elements in the coal bed, on a coal-ash basis, to identify the factors affecting the origin of these elements. The second phase of this research involved the examination of 12 potentially toxic compounds, as identified in the 1990 Clean Air Act Amendments, that may be released during the combustion of coal. The 12 trace elements are As, Be, Cd, Cr, Co, Hg, Mn, Ni, Pb, Sb, Se, and U (radionuclides). We examined

these elements on a whole-coal basis in the Anderson-Dietz 1 coal bed to determine their impact on mining applications.

Background

The present study complements previous work on the maceral composition of the Anderson-Dietz 1 coal bed (Moore, 1986, 1991). Moore (1991) determined the effects of clastic sedimentation on the organic facies development and distribution of sulfur in the peat swamp. Results of his study indicate that a thin zone (about 4.9 ft or 1.5 m thick) approximately in the middle of the Anderson-Dietz 1 coal bed is stratigraphically equivalent to detrital sediments that split the coal into the Anderson and Dietz 1 benches to the east (Figure 3). This study refers to this zone as the midbed area.

Previous work

Studies on the geochemistry of the Anderson-Dietz coal bed include work by Hildebrand (1988),

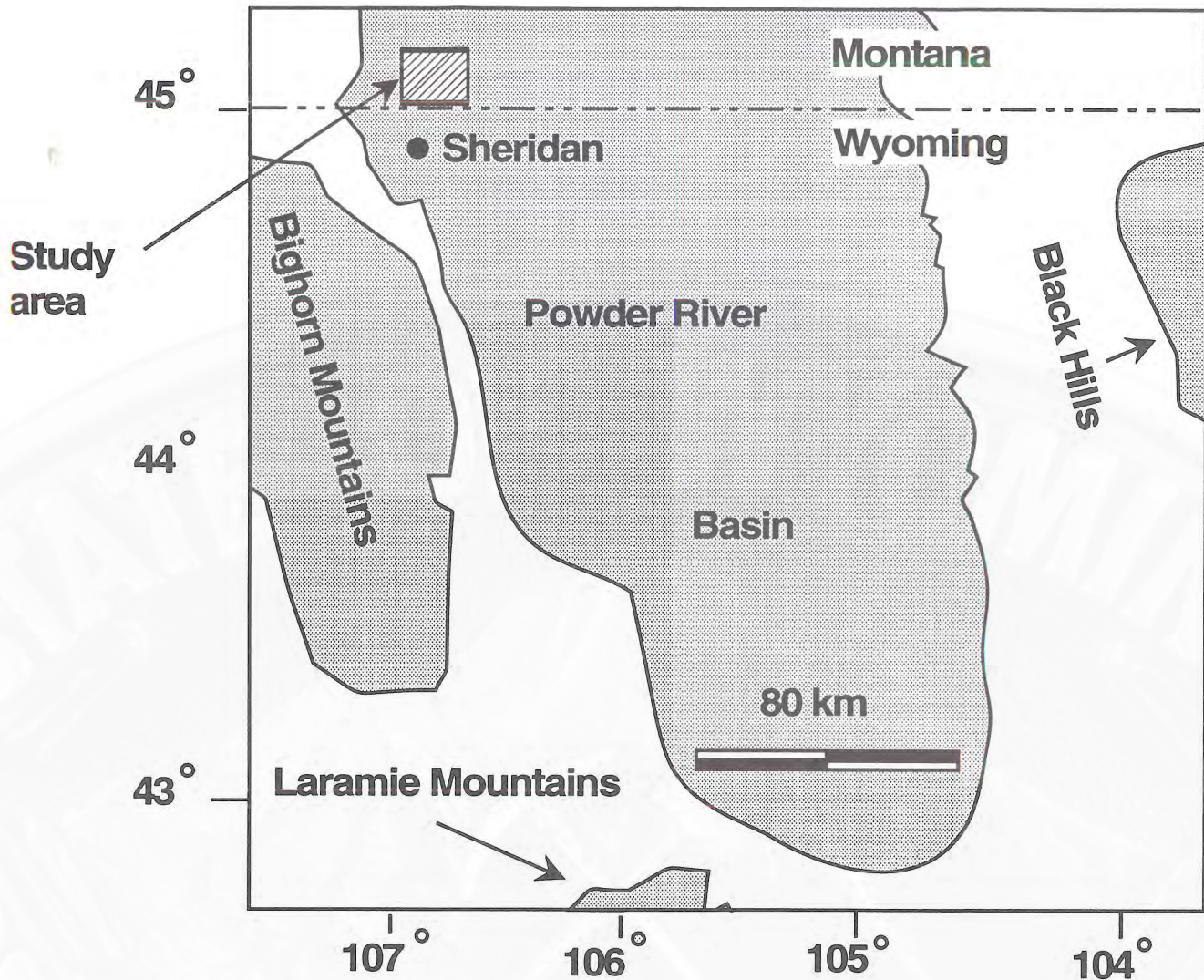


Figure 1. Location of the study area in the Powder River Basin.

who reported that ground-water composition influenced the contents of sodium and associated elements in the Anderson-Dietz coal bed in the Decker, Montana, area. Numerous studies have been made on coal beds correlative with the Anderson-Dietz coal bed, such as the Wyodak-Anderson coal bed (Culbertson and others, 1979). The Wyodak-Anderson coal bed has been described as a thick (72 ft or 22 m), laterally continuous coal body covering 800 mi² (2,000 km²) on the eastern flank of the Powder River Basin (Denson and others, 1973; Denson and Keefer, 1974; Kent, 1986). However, data from closely spaced drill holes indicate that the Wyodak-Anderson coal bed consists of separate, offset, laterally discontinuous bodies of coal (Warwick and Stanton, 1986, 1988a, b; Stanton and others, 1989). An ombrotrophic (rain-fed), raised-peat model has been postulated for the Wyodak-

Anderson coal bed (Flores, 1986; Warwick and Stanton, 1988a, b; Stanton and others, 1989).

Studies have been made on other coal beds correlative with the Anderson-Dietz coal bed. Hobbs and others (1986) provided a detailed stratigraphic and quality analysis of the Anderson coal deposit in the southwestern part of the Powder River Basin. Rich and others (1988) examined the petrography, sedimentology, and geochemistry of the Canyon and Anderson coal beds in the eastern part of the Powder River Basin. Among other conclusions, they reported that the Anderson coal bed was enriched in many inorganic elements in comparison with the Canyon coal bed and suggested that the Anderson coal bed contained more transported material than the Canyon coal bed (Rich and others, 1988).

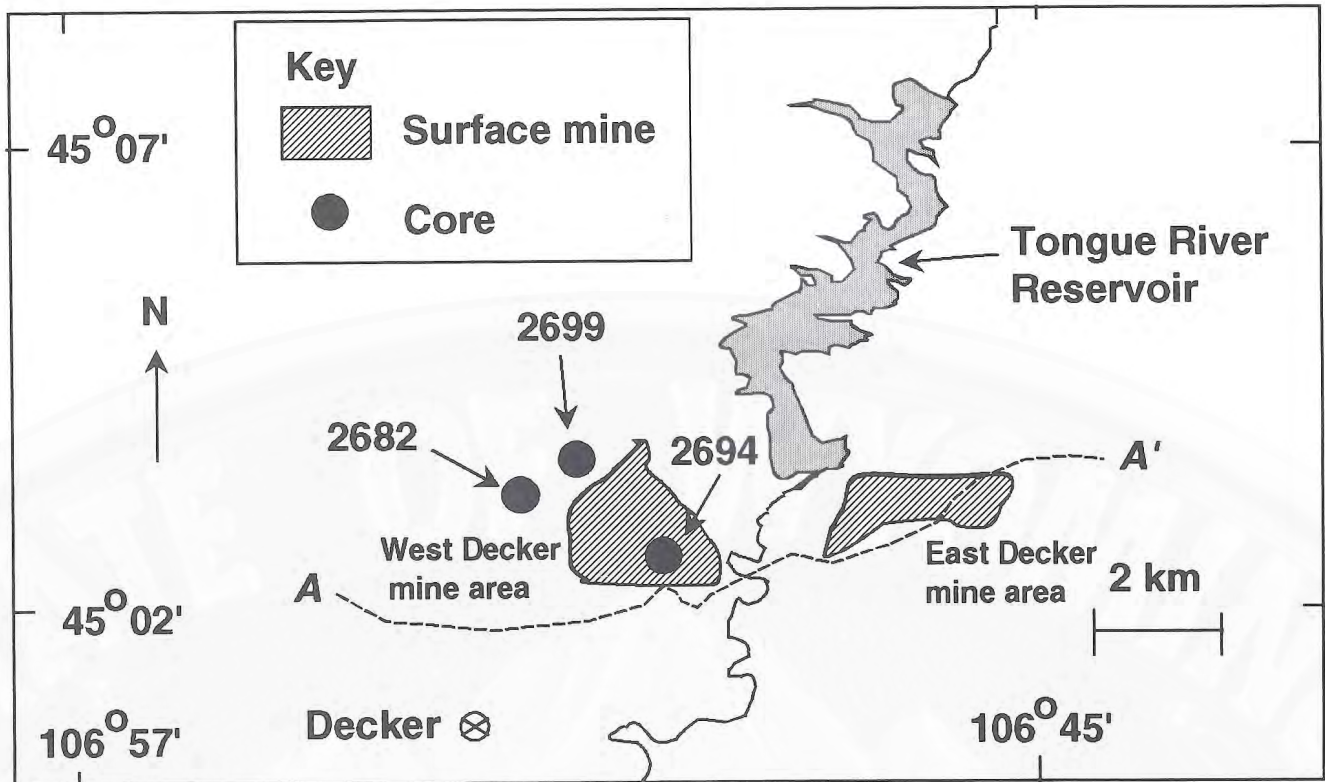


Figure 2. Map of the study area showing the location of the three cores and the line of cross section shown on Figure 3 (modified from Moore, 1991).

Samples and methods

Description of samples

The analysis of inorganic elements was made on 85 samples from three drill cores of the merged Anderson-Dietz 1 coal bed, which is approximately 49.9 ft (15.2 m) thick (Figure 4). Analytical data were not available for one sample in the coal bed (2694-22). Splits of samples for the three drill cores used in this study were obtained from the Decker Coal Company. Eighty-two coal samples and three carbonaceous rock samples (with ash yields greater than 50%) were obtained from the three cores. Seventy-one of the 82 coal samples each represent a 2-ft (0.6-m) interval of coal. Eight coal samples are splits representing from 0.7- to 3.0-ft (0.2- to 0.9-m) intervals of coal; interval thickness data were not available for three coal samples. Two carbonaceous rock samples represent a 0.7 ft (0.2-m) interval of rock; interval thickness data were not available for the third sample (Figure 4). The methods used for chemical analysis have been reported by Crowley and others (1993; in press).

Statistical analysis of 46 inorganic elements

Samples were grouped into compositional clusters by a hierarchical means of comparison (Q-mode cluster analysis). All chemical data used in the analysis were reported on a *coal-ash basis* to evaluate the composition of the inorganic fraction of the coal without the dilution effects of organic matter. Some elements were analyzed directly in the ash. Elements that were analyzed on a whole-coal basis were mathematically converted to a coal-ash basis. To reduce the effect of the relative size of the variables, all chemical data were statistically standardized to the same scale, by producing new variables with a mean of 0 and a standard deviation of 1 (SPSS/PC, Inc., 1986), before use in the cluster analysis. The squared euclidean measure (Aldenderfer and Blashfield, 1984) and average linkage between groups method (Milligan and Isaac, 1980; SPSS/PC, Inc., 1986) were used in

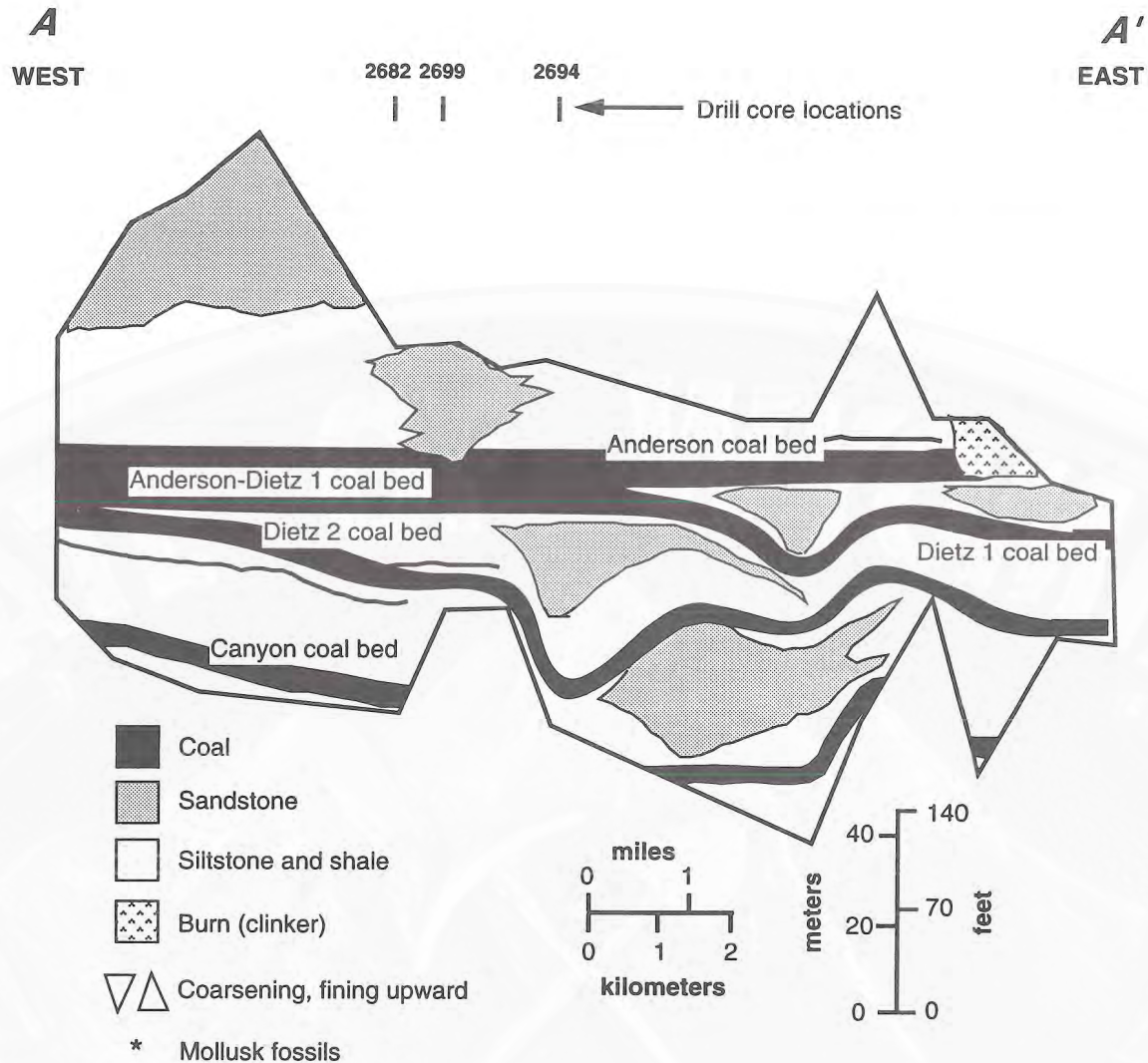


Figure 3. Generalized cross section along A-A' with location of cores studied. Cores are projected to line of cross section (modified from Moore, 1991). Location of cross section is shown in Figure 2.

the cluster analysis. Because the average-linkage-between-groups method uses information about all pairs of distances, not just the nearest or the farthest, it is preferred to the single and complete linkage methods for cluster analysis (SPSS/PC, Inc., 1986). Samples were coded according to their cluster group and plotted on a cross section of the Anderson-Dietz 1 coal bed (Figure 4). Mean values for the major cluster groups were calculated for the chemical data (Table 1).

Analysis of potentially toxic trace elements

Comparisons were made of concentrations of the 12 potentially toxic trace elements defined by the Clean Air Act Amendments of 1990 (As, Be, Cd, Cr,

Co, Hg, Mn, Ni, Pb, Sb, Se, and U (radionuclides)) on a *whole-coal basis* from four parts of the coal bed having different ash yields: (1) the roof area (eight samples having a mean ash yield of 32%), (2) the midbed area (seven samples having a mean ash yield of 7%), (3) the floor rock of core 2682 (two samples having a mean ash yield of 59%), and (4) the rest of the coal bed (63 samples having a mean ash yield of 4%) (Figure 5, Table 2). Although the mean ash yield of the roof area is less than 50%, we refer to this area as "roof rock" in this paper to simplify terminology. In addition, whole-coal-bed analyses of the 12 elements were calculated on the basis of the following mining scenarios: (1) mining that excludes the roof rock and floor rock (in core 2682) of the coal bed and (2) mining that excludes the roof rock, floor rock (in core 2682), and midbed area of the coal bed (Table 3).

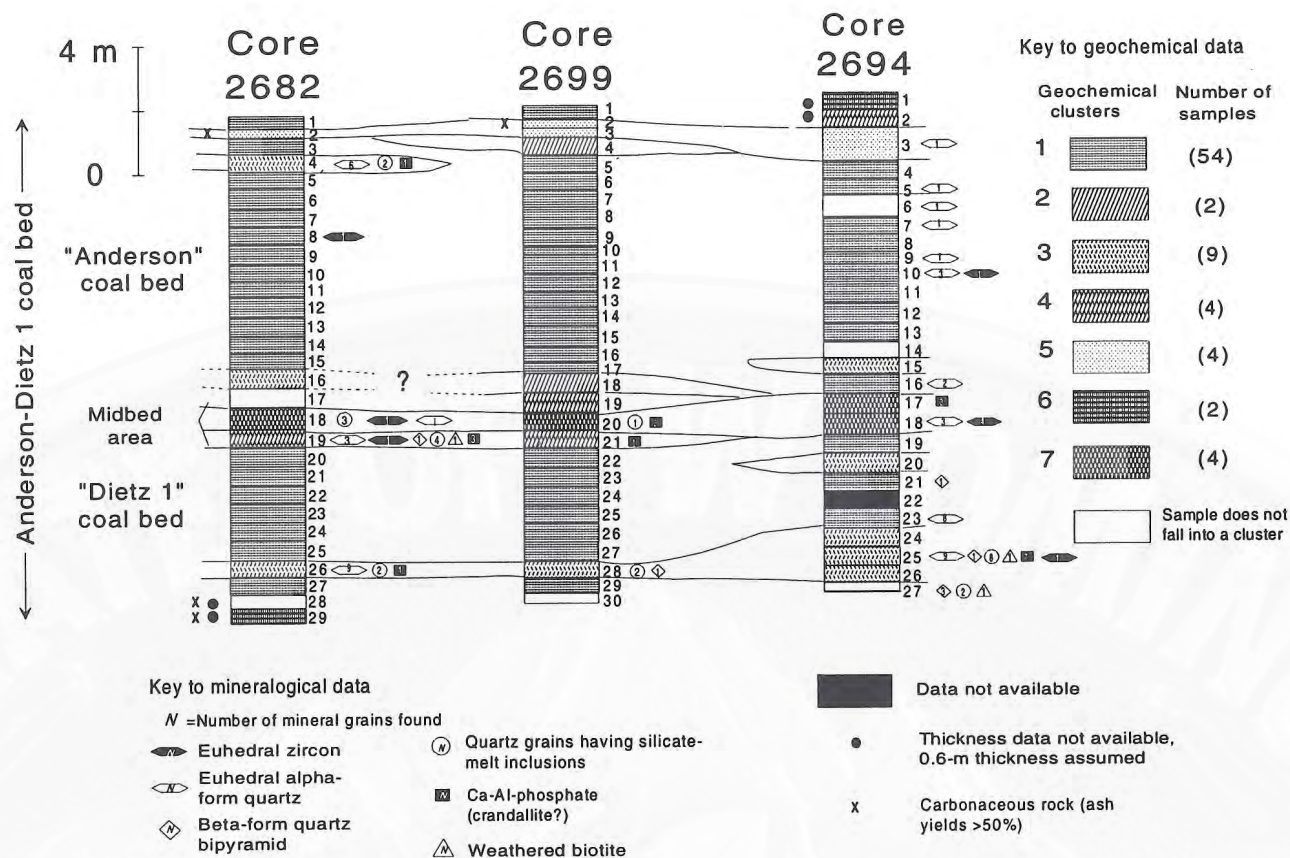


Figure 4. Cross section showing (1) correlation of the geochemical clusters (on a coal-ash basis) and (2) minerals found in the Anderson-Dietz 1 coal bed as reported by Crowley and others (1993). Lines of correlation are dashed where inferred. Numbers adjacent to cores are sample numbers. See Figure 2 for core hole locations.

Results

Statistical analysis of 46 inorganic elements

Seven groups of samples of similar chemical composition were identified by cluster analysis of the chemical data (85 samples and 46 elements on a coal-ash basis). Six samples clustered very weakly with the other samples and lay outside all other sample groupings (Figure 4); these samples are not considered in the following discussion of the clusters. Two of the resulting cluster groups, clusters 5 and 7, correlate within the Anderson-Dietz 1 coal bed throughout the study area (Figure 4). Cluster 5 is composed of two carbonaceous rock samples (having a mean ash yield of 81%) and two coal samples having fairly high ash yields (mean ash yield of 24%) taken from each of the three cores near the top of the coal bed; these samples are characterized by high levels of SiO_2 (Table 1). Cluster 7, which consists of four coal samples taken

from the midbed area, occurs in each of the three cores and contains relatively high levels of 18 of the elements determined (As, Cd, Ce, Co, Cr, Eu, Fe, La, Li, Ni, Sb, Se, Sm, Tb, Th, U, Y, and Zn) in comparison with the other cluster groups (Table 1). Cluster 4 consists of four samples. Because three of the samples are adjacent to cluster 7 samples and one sample is from near the top of core 2694, cluster 4 is limited in its usefulness for correlation. Cluster 4 samples are enriched in many of the same elements that are predominant in cluster 7 samples (As, Cd, Eu, La, Li, Se, Sm, Tb, Th, and U). However, the concentration levels are generally not as high as they are in cluster 7 (Table 1). The nine samples of cluster 3 were predominantly from the lower half of the coal bed. Cluster 3 samples are enriched in P_2O_5 and Zr in comparison with the other clusters (Table 1). Cluster 1 consists of 54 coal samples taken from intervals throughout the three cores. These samples provide the "background" for

Table 1. Mean geochemical values on a coal-ash basis for the seven major clusters formed from the cluster analysis of 85 samples from three cores in the Anderson-Dietz 1 coal bed and geochemical values for six individual samples that clustered very weakly with the other samples.

	Clusters							Non-clustering samples					
	1	2	3	4	5	6	7	2694 -6	2694 -14	2694 •27	2682 -17	2682 -28	2699 -30
Number of samples	(54)	(2)	(9)	(4)	(4)	(2)	(4)						
Oxide content (percent)													
Al ₂ O ₃	13.0	13.7	15.5	16.7	4.0	16.5	15.9	15.4	13.5	14.0	12.9	20.1	17.9
Fe ₂ O ₃	3.4	2.7	3.2	8.5	0.5	4.0	15.5	3.6	7.6	3.7	6.6	3.0	3.4
CaO	17.4	14.0	16.7	10.4	1.7	1.4	7.9	18.6	22.2	16.4	17.5	13.6	14.8
K ₂ O	0.3	0.3	0.3	0.4	0.4	2.5	0.7	0.3	0.6	0.4	0.4	0.3	0.6
MgO	3.8	4.1	2.7	2.4	0.5	2.0	1.8	1.8	1.7	1.6	4.0	0.6	4.0
Na ₂ O	10.5	9.8	7.2	4.5	0.9	4.2	4.5	3.9	3.9	3.8	13.8	0.7	8.9
P ₂ O ₅	0.4	0.3	1.9	1.1	0.1	0.0	0.8	0.3	0.2	1.3	0.3	0.1	1.6
SiO ₂	25	32	24	28	90	67	29	25	19	25	13	30	26.8
SO ₃	10	10	10	14	1	4	11	9	5	1	5	8	9
TiO ₂	0.9	3.0	1.6	0.9	0.5	0.8	1.1	0.6	0.5	1.3	0.6	1.8	1.8
Element content (parts per million)													
Ag	0.3	0.4	0.5	0.5	0.2	0.3	0.5	0.3	0.3	0.3	0.4	0.4	0.6
As	15.2	19.1	8.8	59.7	1.9	11.7	221.3	6.7	7.1	10.5	34.3	1.0	20.4
B	581	405	619	353	112	152	278	610	660	740	610	640	460
Ba	14346	5000	16489	4125	638	1340	4725	22000	16000	14000	19000	5500	4100
Be	1.5	1.0	1.7	4.1	1.0	4.7	3.9	2.1	1.7	49.0	3.0	12.0	14.0
Br	17.7	14.4	16.5	11.1	1.7	1.7	9.2	14.4	10.0	23.3	19.7	1.4	13.7
Cd	0.4	0.3	0.4	0.6	0.2	0.5	0.7	0.2	0.2	0.6	0.8	0.8	0.3
Ce	59	71	73	87	29	79	135	56	45	128	74	6	94
Co	15	14	22	19	1	15	28	16	13	58	24	3	37
Cr	43	59	52	44	29	74	69	44	45	77	57	8	78
Cs	2.0	2.2	2.3	2.3	7	10.1	5.9	1.0	1.2	4.4	1.4	0.3	2.0
Cu	136	295	367	168	31	235	208	77	60	250	160	38	310
Eu	1.3	1.6	1.8	2.0	0.4	1.4	2.8	1.0	1.2	3.7	2.6	0.2	2.9
Hr	6.4	6.6	8.0	5.8	10.5	5.9	6.5	4.4	4.8	15.4	4.0	0.7	7.8
La	38	47	47	57	18	47	84	31	25	80	37	4	61
Li	36	36	41	70	17	61	93	38	42	35	40	38	33
Mn	179	130	157	158	45	180	154	240	1200	180	150	48	170
Nb	20	28	23	16	14	16	23	16	17	62	11	17	33
Ni	38	30	19	74	7	70	109	47	39	100	82	68	88
Pb	44	25	59	47	20	48	53	310	48	30	59	42	29
Rb	1445	2254	587	786	180	104	558	436	341	279	800	35	1224
Sb	3.0	3.1	2.0	10.3	0.7	2.7	17.6	1.9	1.0	3.5	10.0	0.2	3.1
Sc	17	25	28	21	4	15	23	16	15	71	20	6	65
Se	11	14	14	25	3	6	30	9	14	23	12	2	12
Sm	6	7	8	10	2	8	14	6	5	16	10	1	11
Sr	11294	11000	11200	6650	745	575	3425	16000	8600	7200	11000	5700	9500
Ta	1.4	2.8	1.8	1.2	0.7	1.2	1.8	0.7	1.1	1.8	0.9	0.2	1.7
Tb	0.8	0.7	0.9	1.1	0.3	0.9	1.7	0.4	0.5	2.6	1.4	0.2	1.7
Th	10	12	12	15	6	14	19	9	8	14	4	1	10
U	7	6	6	18	3	12	31	6	5	7	13	1	7
V	138	175	230	138	36	150	163	140	150	300	190	410	240
W	27	29	16	17	2	2	11	8	9	11	16	2	35
Y	36	35	46	49	16	36	62	38	40	130	70	83	91
Yb	11	18	6	7	3	4	7	3	3	11	4	1	10
Zn	57	48	64	96	27	100	137	49	66	110	130	110	64
Zr	297	190	402	218	308	220	273	230	320	1100	210	340	280
Ash content (percent) ¹													
Ash (%) ...	3.8 (1.1)	5.0 (1.2)	4.2 (0.8)	6.7 (1.2)	52.4 (33.5)	45.5 (32.0)	7.9 (1.9)	3.9	4.4	4.3	3.5	49.0	4.9

¹Standard deviation in parentheses

the other clusters and contain moderate levels of most of the elements studied, with the exception of Na, which is relatively abundant (Table 1).

Two of the clusters do not correlate with any of the other clusters. Cluster 2 consists of two coal samples from core 2699, one located near the top of the bed and the other near the midbed area. Although cluster 2 samples are similar in composition to cluster 1 samples, they contain high levels of Rb, Ta, and Ti in comparison with the other clusters (Table 1). The two samples composing cluster 6 are from areas at the top and bottom of the coal bed and contain the highest levels of Be, Cr, Cs, and K₂O (Table 1).

Analysis of potentially toxic trace elements

The highest mean values for four of the trace elements (As, 11.4 ppm; Sb, 1.1 ppm; Se, 2.1 ppm; Hg, 0.10 ppm) on a whole-coal basis are obtained from

seven coal samples from the three cores. All seven samples were from the midbed area of the coal bed (Figure 5, Table 2). The distributions of As are shown in a graph of elemental concentration versus depth of the cores in Figure 6. Sb, Se, and Hg show distribution patterns similar to that of As. The mean values for As, Sb, and Se in the midbed area are higher than the mean values for 181 samples having ash yields of 6 to 11% from correlative coal beds in the PRB (As, 2.5 ppm; Sb, 0.3 ppm; Se, 0.9 ppm) (Crowley and others, in press). The concentration of Hg in the midbed area is nearly the same as the mean concentration for Hg (0.09 ppm) in samples having ash yields of 6 to 11% from the PRB (Crowley and others, in press).

The highest levels of U in this coal bed occur in samples from the midbed area. Although the mean level for U in the midbed area is only 1.9 ppm (Table 2), levels of U are as high as 3.5 ppm in core 2694. The pattern of distribution for U is similar to that of As (Figure 6), Sb, Se, and Hg. The mean concentration of U (1.9 ppm) in the midbed area is

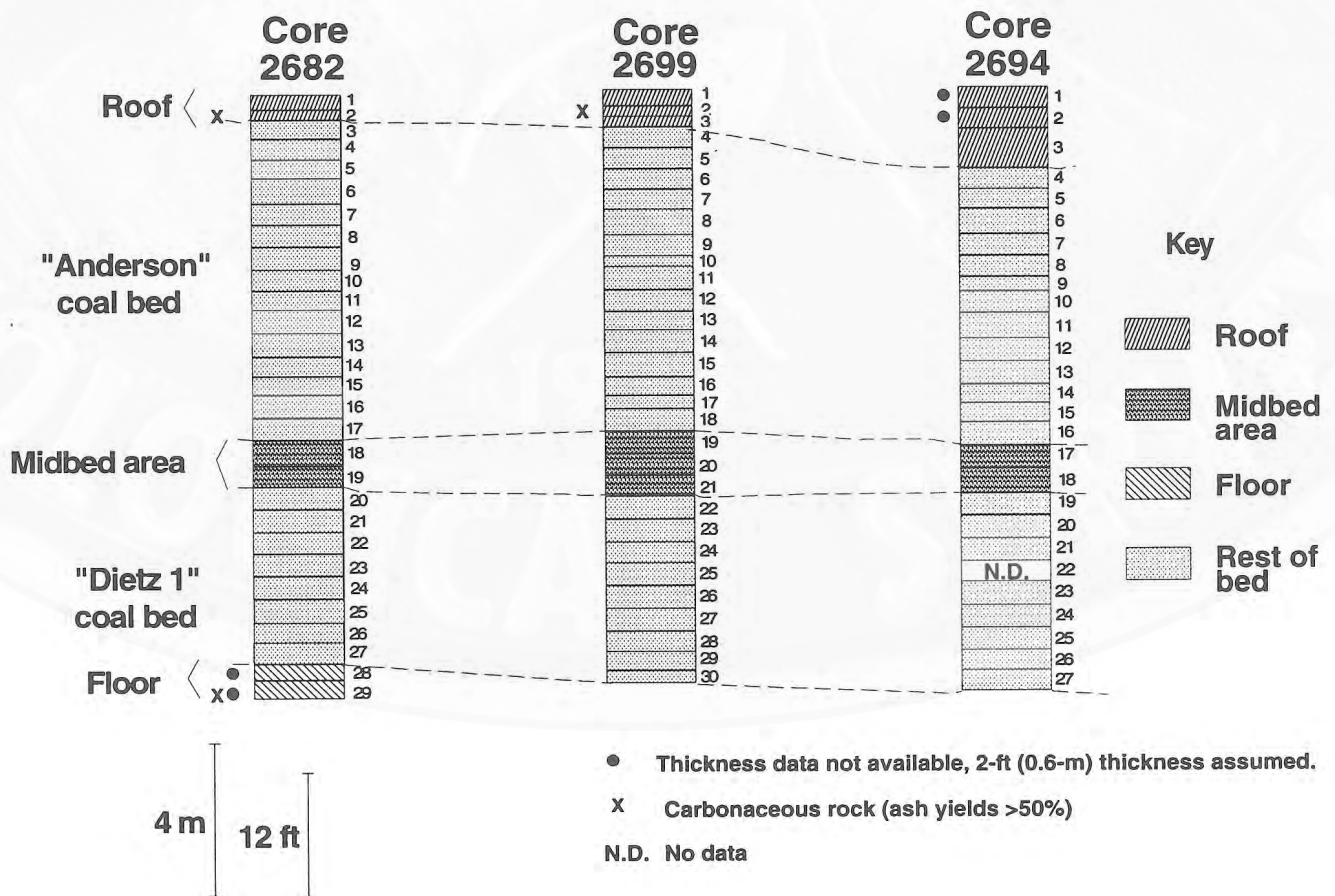


Figure 5. Cross section of three cores of the Anderson-Dietz 1 coal bed showing sample locations and the five areas of the coal bed that were used for comparisons of concentrations of the 12 trace elements studied. See Figure 2 for core hole locations.

higher than the mean concentration for U in samples having ash yields of 6 to 11% from the PRB (0.9 ppm) (Crowley and others, in press). Moderately high levels of U also are found in the roof rock samples (mean concentrations of 1.6 ppm) (Table 2) and floor rock samples (mean concentration of 1.7 ppm) (Table 2).

The highest levels of Be (3.9 ppm), Cd (0.28 ppm), Co (4.8 ppm), Cr (22.5 ppm), Mn (66 ppm), Ni (27 ppm), and Pb (28 ppm) are contained in two samples (a high-ash coal sample and a rock sample) from the floor rock of the coal bed (Figure 5, Table 2).

Table 2. Mean concentrations of the 12 studied elements on a whole-coal basis in the four parts of the Anderson-Dietz 1 coal bed having different ash yields. [All values, except ash, are in parts per million.]

	High ash zones		Low ash zones	
	Floor rock	Roof rock	Midbed	Rest of coal bed
Range of ash yields (percent)	>49	7-87	6-11	3-6
Mean ash yield (percent)	59	32	7	4
Standard deviation of ash	14	31	2	1
Number of samples	2	8	7	63
As	1.5	1.4	11.4	0.5
Be	3.9	0.5	0.3	0.1
Cd	0.28	0.09	0.05	0.02
Co	4.8	1.0	1.7	0.7
Cr	22.5	9.9	4.2	1.8
Hg	0.02	0.01	0.10	0.01
Mn	66	21	10	7
Ni	27	6	7	2
Pb	28	8	3	2
Sb	0.3	0.4	1.1	0.1
Se	1.2	1.0	2.1	0.4
U	1.7	1.6	1.9	0.3

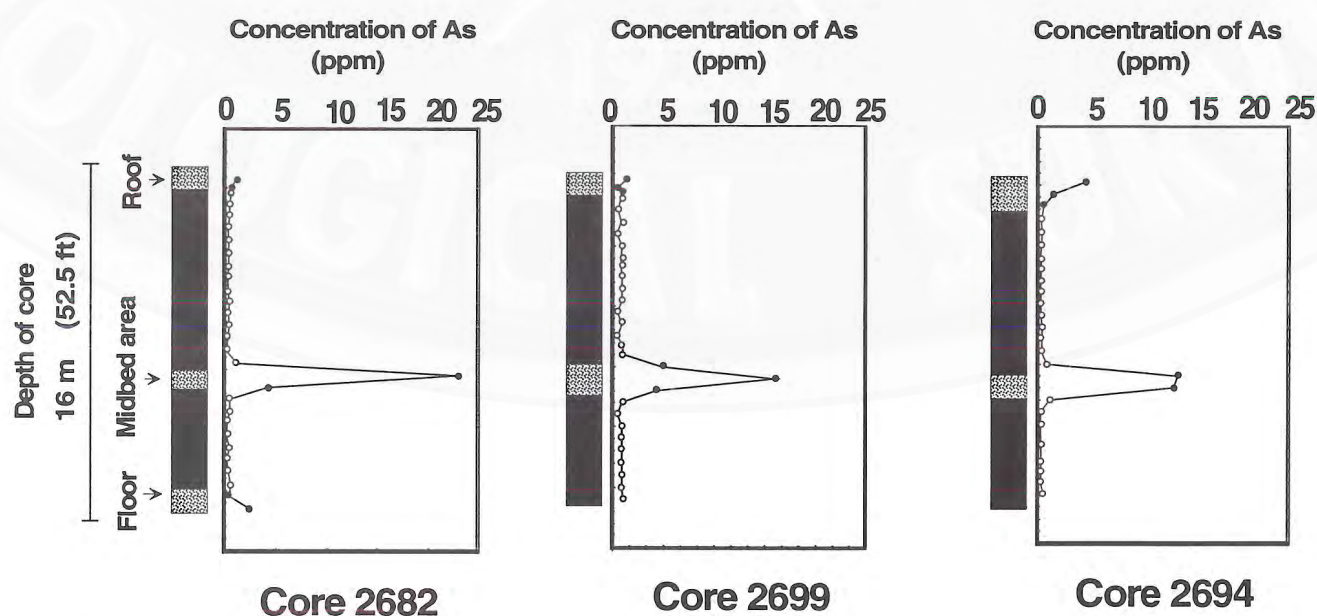


Figure 6. Graph showing concentration of arsenic on a whole-coal basis in roof rock, coal, and floor rock in three cores of the Anderson-Dietz 1 coal bed.

Discussion

Statistical analysis of the 46 inorganic elements correlation using geochemical clusters

Cluster analysis of geochemical data on a coal-ash basis for 82 coal samples and three carbonaceous rock samples from three cores in the Anderson-Dietz 1 coal bed indicates that the sampled intervals in the coal bed can be correlated, to a limited extent, by using their geochemical characteristics (Figure 4). Two of the resulting clusters were found to correlate best: (1) cluster 7, composed of four samples from the midbed area, and (2) cluster 5, composed of four samples taken from each of the three cores near the top of the coal bed. The rest of the clusters are limited in their usefulness for correlation; the samples used in this study (splits that represent a 2-ft (0.6-m-thick) interval of core) probably were not detailed enough to permit correlation of thin layers or lithotypes in the coal bed.

Volcanic ash and its effect on the geochemistry of the coal bed

Previous work on the mineralogy of the Anderson-Dietz 1 coal bed indicates the presence of a minor volcanic ash component in the coal bed (Crowley and others, 1993). Mineralogical data supporting a volcanic-ash component include the presence of a limited number of beta-form quartz bipyramids and euhedral zircons (Figure 4). The presence of quartz grains having silicate-melt inclusions in samples from the midbed area, cluster 3 samples, and sample 2694-27 provides the strongest evidence for a volcanic-ash component in the Anderson-Dietz 1 coal bed.

Several factors indicate that cluster 3 samples (predominantly from the lower part of the coal bed) and cluster 4 and 7 samples (from the midbed area) may be associated with volcanic ash. Ca-Al-phosphate (crandallite?) grains were observed in samples comprising cluster 3, which contains the highest level of P_2O_5 (1.9%) of all the clusters. Ca-Al-phosphate was also found in clusters 4 and 7. Minerals from the crandallite-goyazite group are sometimes found in tonsteins (altered volcanic ashes) associated with coals; these minerals apparently form as alteration products of the original volcanic ash (Triplehorn and Bohor, 1983, 1986; Triplehorn and Finkelman, 1989; Brownfield and others, 1986; Reinink-Smith, 1990). The Ca-Al-phosphate found in the Anderson-

Dietz 1 coal bed may have formed from the alteration of volcanic material that was either transported by fluvial processes to sites of peat deposition or deposited as an air-fall volcanic ash.

Effect of other factors on the geochemistry of the coal bed

Previous mineralogical studies (Crowley and others, 1993) have indicated that most of the mineral grains from the Anderson-Dietz 1 coal bed samples are detrital (water borne) in origin, on the basis of the presence of rounded to subrounded hydrothermal quartz grains and low- to moderate-grade metamorphic quartz grains found throughout the coal bed. However, the total number of discrete grains in the coal is relatively low, as the low ash content of 65 samples in clusters 1, 2, and 3 indicates (mean values of 3.8-5%, Table 1). Cluster 5 samples, located near the top of the coal bed (Figure 4), have the highest mean ash content (52.4%, Table 1) of all of the clusters; 90% of the ash is silica. We infer that these samples received detrital input from the sedimentary processes that formed the roof rock of the coal bed. Samples from clusters 4 and 7 of the midbed area (having mean ash contents of 6.7 and 7.9%, respectively) contain abundant quartz grains, in comparison with the other cluster groups, that we interpret as detrital. As discussed earlier, mineral matter in these samples was probably derived from the fluvial processes that formed the split (see Flores and Moore, this volume) between the Anderson and the Dietz 1 coal beds east of the study area (Figure 3). This hypothesis is supported by the presence of high levels of detrogeinite and corpohuminitite, vitrinite macerals that represent partially decomposed plant fragments, which Moore (1991) observed near the midbed area.

Plant remains probably provided a source for some of the chemical elements in the coal bed. Cluster 1 samples provide the "background" for the other clusters and contain moderate levels of most of the elements studied. If the ombrotrophic (rain-fed), raised-peat model suggested for the Wyodak-Anderson coal bed (Flores, 1986; Warwick and Stanton, 1988a, b; Stanton and others, 1989) is applicable, then detritus from fluvial systems was probably not a primary contributor to the mineral fraction of the peat (Warwick and Stanton, 1988a); for this reason we suggest that many of the elements having moderate concentrations in cluster 1 were derived from plant material.

The movement of ground water may also have influenced the distribution of elements in the coal bed. Sodium is enriched in cluster 1 compared to the other clusters. Because previous work indicates that the occurrence of sodium in coal beds is related to ground-water composition (Hildebrand, 1986; Finkelman, 1987; Fulton, 1989; Demir and others, 1990; Ting, 1990), we cannot rule out the possibility that high concentrations of sodium in the Anderson-Dietz 1 coal bed resulted from the movement of ground water subsequent to peat deposition. Cluster 7 samples in the midbed area contain the highest levels of uranium (31 ppm) of all the cluster groups. The circulation of ground water has been reported to affect the distribution of uranium in other coal beds (Moore, 1954; Breger, 1974; Van der Flier and Fyfe, 1985). The levels of uranium found in cluster 7 samples may be related to the flow of ground water; perhaps the chemical environment of this section in the coal bed was more conducive than that of the other parts of the coal bed to removing uranium ions from the circulating ground water. Alternatively, it is also possible that the levels of uranium found in the midbed area were derived from minerals that were washed into the peat mire.

Analysis of potentially toxic trace elements

The samples from the midbed area of the Anderson-Dietz 1 coal bed have a mean ash yield (7%, on a dry basis) that is slightly higher than that for the rest of

the samples in the coal bed (4%, on a dry basis), excluding samples from the roof or floor rock of the coal bed (Table 2). The samples from the midbed area contain high levels of As, Sb, Se, Hg, and U on a whole-coal basis in comparison with the rest of the coal bed; the high concentration of As in the midbed area is demonstrated in Figure 6. As discussed previously, the elevated levels of these elements probably are associated with the deposition of clastic materials that formed the split between the Anderson and Dietz 1 coal beds to the east of the study area (Figure 3). Higher concentrations of uranium in the midbed area (Figure 4, Table 1) also may be related to the deposition of clastic materials.

The levels of As, Sb, Se, and U in the midbed area are higher than those for samples having comparable ash yields (6 to 11% ash yields) (Crowley and others, in press) from other parts of the PRB. As previously discussed, consistent reductions in the concentrations of As, Sb, Se, and U, on a whole-coal-bed basis, can be achieved by excluding the midbed area of the Anderson-Dietz 1 coal bed from the product (Table 3). Our study indicates that detailed sampling and chemical analysis of other thick coal beds may help to delimit specific low-ash intervals of coal that have elevated levels of the elements listed in the 1990 Clean Air Act Amendments. If analysis of a coal bed shows that toxic elements exceed the limits set by the Environmental Protection Agency (EPA) in a specific horizon of the coal bed, then reduction of the hazardous pollutants can be achieved by selective mining processes.

Table 3. Whole-coal-bed averages for the Anderson-Dietz 1 coal bed based on two mining scenarios: (1) exclusion of roof and floor rocks from product and (2) exclusion of roof rock, floor rock, and midbed area from product. [All units are in parts per million unless noted otherwise. Percentage of reduction is achieved by excluding midbed area. Whole-bed averages were calculated by using thickness of sample interval as a weighting factor.]

	Core 2682			Core 2699			Core 2694		
	Excluding roof, floor	Excluding roof, floor, midbed area	Percentage of reduction	Excluding roof, floor	Excluding roof, floor, midbed area	Percentage of reduction	Excluding roof, floor	Excluding roof, floor, midbed area	Percentage of reduction
As	1.5	0.4	73	1.7	0.8	53	1.5	0.3	80
Be	0.1	0.1	0	0.1	0.1	0	0.2	0.1	50
Cd	0.02	0.02	0	0.01	0.01	0	0.02	0.01	50
Co	0.7	0.6	14	0.7	0.6	14	0.8	0.7	12
Cr	2.0	1.9	5	1.9	1.7	11	2.2	1.8	18
Hg	0.03	0.01	67	0.01	0.01	0	0.01	0.01	0
Mn	7	6	14	6	6	0	10	10	0
Ni	2	2	0	2	1	50	2	2	0
Pb	3	2	33	1	1	0	3	2	33
Sb	0.2	0.1	50	0.2	0.1	50	0.2	0.1	50
Se	0.6	0.4	33	0.5	0.4	20	0.6	0.5	17
U	0.4	0.3	25	0.4	0.2	50	0.5	0.2	60

In the present study, elevated levels of Be, Cd, Co, Cr, Mn, Ni, and Pb are found in high-ash samples (Table 2) from the floor of the coal bed in comparison with the rest of the coal bed. The floor rock of coal beds is sometimes added to the mine product, the

result being a higher ash yield product. Recognition that such high-ash coal or rock may contribute to higher levels of certain toxic elements, as defined by the EPA, may benefit mine planning and mitigation efforts.

Conclusions

The Anderson-Dietz 1 coal bed can be correlated to a limited extent by use of geochemical data alone, as the distribution of 46 inorganic elements (on a coal-ash basis) in three cores of the coal bed indicates. The geochemical data combined with mineralogical analyses from previous work (Crowley and others, 1993) indicate that the geochemistry of the Anderson-Dietz 1 coal bed has probably been influenced by the following factors: (1) detrital input associated with the fluvial processes that formed the split between the Anderson and Dietz 1 coal beds and the roof rock of the coal bed, (2) volcanic materials that were transported by surface water into the peat or deposited as air-fall volcanic ash, (3) plant material and peat-forming processes, and (4) ground-water flow.

The midbed area of the coal bed contains the highest levels of 18 of the elements studied. Several trace elements (As, Sb, Se, and U) that are of environmental interest as defined by the Clean Air Act Amendments of 1990 are present in high concentrations in this midbed area in comparison with the rest of the coal bed, on either a coal-ash or a whole-coal basis. The mean ash content of samples from the midbed area (approximately 7%) is only slightly higher than the ash content for most of the samples from the rest of the coal bed; recognition of such intervals of coal having slightly higher ash yields may be of value in mine planning and coal utilization.

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Common abbreviations used in this publication

Temperature

°F = degrees Fahrenheit

°C = degrees Celsius

Weight

g = gram

mg = milligram

Concentration

ppm = parts per million

ppb = parts per billion

Time

Ma } million years
m.y. }
my }
Ga = billion years

Distance

mi = miles

km = kilometers

ft = feet

m = meters

in. = inches

cm = centimeter

mm = millimeters

µm = micrometers

Volume

bbls = barrels

BO = barrels of oil

MCFG = thousand cubic feet of gas

MMCFG = million cubic feet of gas

BCFG = billion cubic feet of gas

Others

R_o = vitrinite reflectance, percent

TOC = Total organic carbon

XRD = X-ray diffraction

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