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# Sequential Laramide Deformation and Paleocene Depositional Patterns in Deep Gas-Prone Basins of the Rocky Mountain Region

*By* William J. Perry, Jr., *and* R.M. Flores

GEOLOGIC CONTROLS OF DEEP NATURAL GAS RESOURCES IN THE UNITED STATES

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# Sequential Laramide Deformation and Paleocene Depositional Patterns in Deep Gas-Prone Basins of the Rocky Mountain Region

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## ABSTRACT

Successive eastward and northeastward partitioning of the Late Cretaceous Rocky Mountain foreland basin took place in southern Montana and Wyoming during latest Cretaceous and Paleocene time. Economic implications, particularly for deep gas accumulations, are examined in terms of this structural sequence. Calculated basin subsidence rates and associated basin-margin faults and folds are characteristic of transpressional (oblique-contractional) deformation. The sequence of structural events within the Hanna and Wind River Basins is discussed in terms of deep gas occurrences, and the possibility, and possible locations, of undiscovered deep gas are explored.

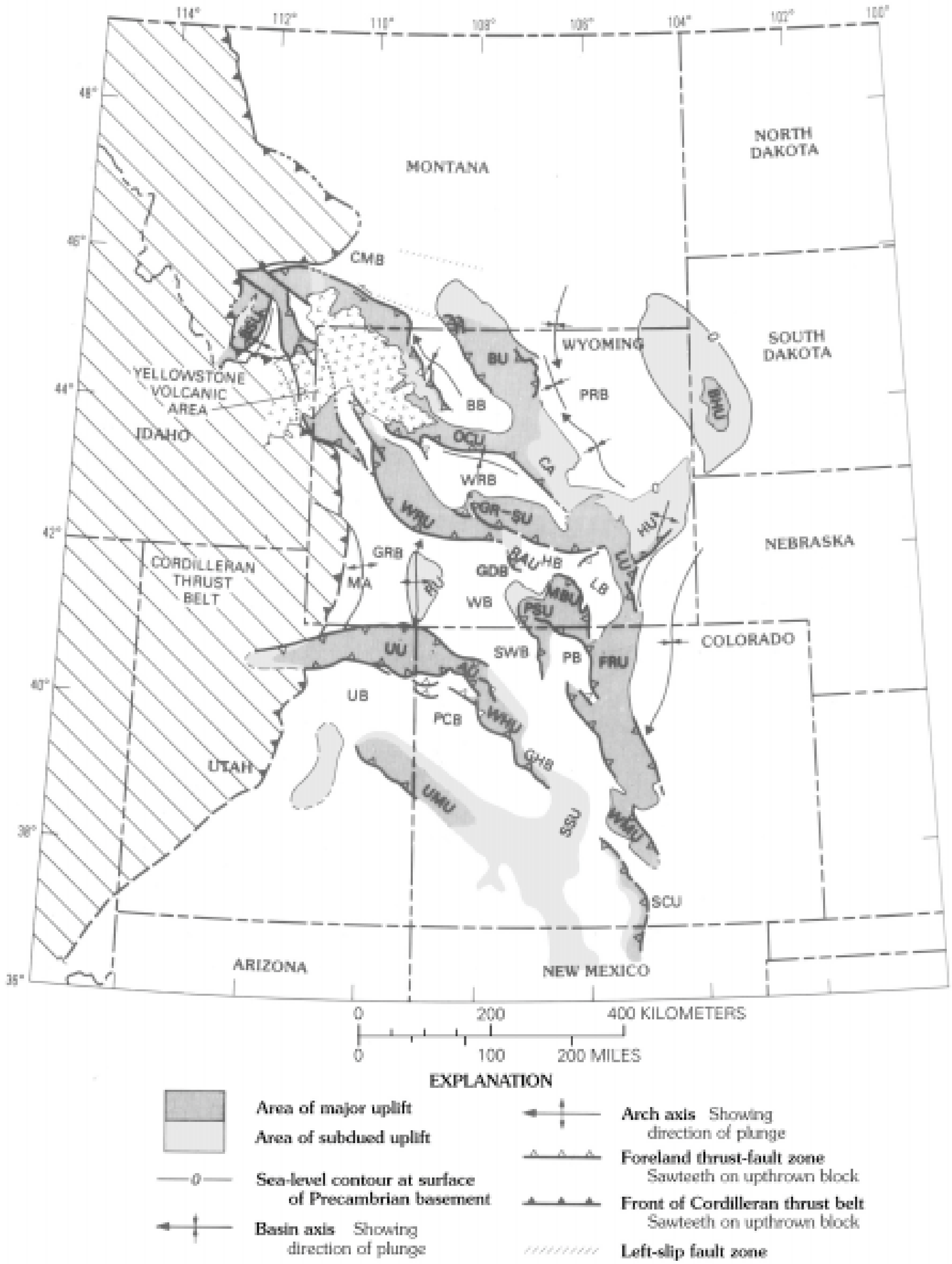
## INTRODUCTION

In this report, we describe the structural history, setting, and trapping mechanism of deep gas accumulations in several Rocky Mountain basins in order to relate these factors to undiscovered natural gas resources. We compare the timing, synorogenic-sediment dispersal patterns, and structural style of two deep Rocky Mountain basins in an attempt to approach this goal.

Perry (1989; this volume) showed that deep natural gas accumulations in the conterminous United States are associated primarily with two structural settings: (1) passive continental margin basins and (2) basins associated with and inland from active continental margins. This latter group of basins (type 2 basins of Perry, 1989) was subdivided by Perry into forearc basins, seaward of the magmatic arc above a continentward-dipping subduction zone; foreland basins, beneath and cratonward of the frontal zone of fold and thrust belts; and extensional or transtensional basins, associated chiefly with transform margins. In this report, we discuss deep gas-prone Late Cretaceous and early Tertiary Laramide basins of the Rocky Mountain foreland, in particular, the Hanna and Wind River basins.

Several sedimentary basins in the central Rocky Mountains contain substantial volumes of sedimentary rock at depths greater than 15,000 ft (4,572 m); the largest of these basins are the Green River and Uinta basins, respectively north and south of the Uinta uplift (fig. 1). These basins initially developed during Cretaceous time as foredeeps in front of the eastward-prograding Wyoming and Utah salients of the Cordilleran thrust belt. A southeastward progression of major uplift and consequent basin development in the Rocky Mountain foreland began in extreme southwestern Montana west of the Neogene Yellowstone volcanic area (fig. 1) during Cenomanian-Turonian time (Perry and others, 1990; Haley and others, 1991). Investigation of the sequence of Laramide deformation and relative timing of Rocky Mountain foreland basin development (Perry and others, 1990, 1992) has begun to revolutionize our understanding of the Late Cretaceous and early Tertiary history of the Rocky Mountain region (Flores and others, 1991; Keighin and others, 1991; Nichols and others, 1991; Roberts and others, 1991). The following comments concerning the history of Laramide deformation are summarized from Perry and others (1992).

No evidence of Campanian or older Cretaceous Laramide-style deformation (other than tectonic welts of low relief) is present in the Rocky Mountain foreland east or southeast of the Blacktail-Snowcrest and Wind River uplifts (fig. 1), based on available palynostratigraphic dating of preorogenic and synorogenic sediments, with the exception of gravels of unknown origin in the Frontier Formation in the northwestern part of the Bighorn Basin. The Front Range uplift began by 69 Ma and culminated in exposure of the


114°
112°
110°
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44°
42°
40°
38°
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MONTANA
NORTH DAKOTA
SOUTH DAKOTA
WYOMING
IDAHO
YELLOWSTONE VOLCANIC AREA
NEBRASKA
COLORADO
UTAH
ARIZONA
NEW MEXICO
CORDILLERAN THRUST BELT
CMB
BU
BB
PRB
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PCB
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GHB
UMU
SSU
SCU
0 200 400 KILOMETERS
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EXPLANATION
Area of major uplift
Area of subdued uplift
Sea-level contour at surface of Precambrian basement
Basin axis Showing direction of plunge
Arch axis Showing direction of plunge
Foreland thrust-fault zone Sawteeth on upthrown block
Front of Cordilleran thrust belt Sawteeth on upthrown block
Left-slip fault zone

crystalline basement by early Paleocene time (Kluth and Nelson, 1988; Wallace, 1988). Subsequent Laramide deformation spread northeastward from the Granite Mountains–Shirley Mountains uplift (fig. 1) in south-central Wyoming. The Laramide deformation front reached the Black Hills by late Paleocene time, creating first the Wind River Basin and then the Powder River Basin, partitioning these basins from an earlier continuous foreland basin with minor welts (Merewether and Cobban, 1986). These broad structural welts of low relief, such as the San Rafael Swell in eastern Utah, had begun to grow in the Rocky Mountain foreland by mid-Cretaceous time (about 90 Ma). A major east-west crustal discontinuity along the Wyoming-Colorado State line separates Archean basement rocks to the north from Proterozoic basement rocks to the south. South of this discontinuity, upper crustal Laramide deformation probably proceeded from east to west, opposite in direction from that in the north, culminating along and defining the eastern boundary of the Colorado Plateau in late Eocene, chiefly Green River time (Perry and others, 1992) (fig. 2).

Economic implications of this newly recognized sequence of deformation of the northern and central Rocky Mountain foreland include progressive opening and subsequent blockage of migration paths for hydrocarbons generated from Paleozoic source rocks in southeastern Idaho, southwestern Montana, Wyoming, Colorado, and eastern Utah. Deep natural gas, generated during the Tertiary, has likely migrated from the deeper parts of these foreland basins into structural traps formed during Laramide deformation.

Within the Rocky Mountain foreland, the Laramide Green River and Uinta Basins are followed in order of size of area deeper than 15,000 ft (4,572 m) by the Wind River Basin, the Great Divide and Washakie Basins, and, perhaps the deepest of all, the Hanna Basin, all in Wyoming (fig. 1). These latter four basins began to subside in Late Cretaceous time as part of the Hanna trough (Thomas, 1949; LeFebre, 1988). This trough extended from the front of the Cordilleran thrust belt, in northeastern Utah and southeastern Idaho, eastward across southern Wyoming. Southward thinning of the upper Maastrichtian siliciclastic sequence along the southern margin of this trough along the eastern flank of the late Laramide Washakie Basin is shown in great detail by Hettinger and others (1991). The region of the Great Divide, Hanna, and Washakie basins was partitioned from the Green River Basin to the west in latest Cretaceous time by growth of the Rock Springs uplift (Kirschbaum and Nelson, 1988; Hettinger and Kirschbaum, 1991) following Late Cretaceous development of the Wind River–ancestral Teton–Granite Mountains uplift (Perry and others, 1990). The Rawlins uplift finally isolated the Hanna Basin from the other basins most likely in latest Paleocene to Eocene time, subsequent to deposition of Paleocene coals of P2 age according to R.D. Hettinger (oral commun., 1992).

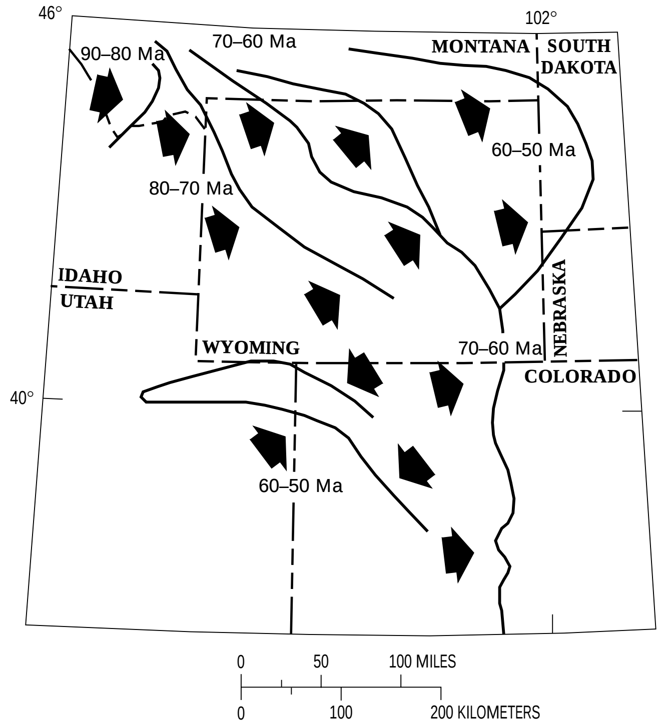


**Figure 2.** Map showing sequence of inception of Laramide deformation in Rocky Mountain region. Arrows indicate direction of migration of Laramide deformation front.

**Figure 1 (previous page).** Map of Rocky Mountain foreland province showing principal Laramide basins and uplifts. Medium shade, major uplift; light shade, broad positive area. Sawteeth on thrust faults point into upper plates. AU, Axial uplift; BB, Bighorn Basin; BHU, Black Hills uplift; BSU, Blacktail-Snowcrest uplift; BU, Bighorn uplift; CA, Casper arch; CMB, Crazy Mountains Basin; FRU, Front Range uplift; GDB, Great Divide Basin; GHB, Grand Hogback uplift; GRB, Green River Basin; GR-SU, Granite Mountains–Shirley Mountains (Sweetwater) uplift; HB, Hanna Basin; HU, Hartville uplift; LB, Laramie Basin; LU, Laramie uplift; MA, Moxa arch; MBU, Medicine Bow uplift; OCU, Owl Creek uplift; PB, North and Middle Parks Basin; PCB, Piceance Creek Basin; PRB, Powder River Basin; PSU, Park–Sierra Madre uplift; RAU, Rawlins uplift; RU, Rock Springs uplift; SCU, Sangre de Cristo uplift; SSU, Sawatch–San Luis uplift; SWB, Sand Wash Basin; UB, Uinta Basin; UMU, Uncompahgre uplift; UU, Uinta Mountains uplift; WB, Washakie Basin; WHU, White River uplift; WMU, Wet Mountains uplift; WRB, Wind River Basin, and WRU, Wind River uplift. Modified from Bayley and Muehlberger (1968).

## HANNA BASIN

The Hanna Basin (fig. 3) contains more than 30,000 ft (9,144 m) of Phanerozoic sedimentary rocks, of which more

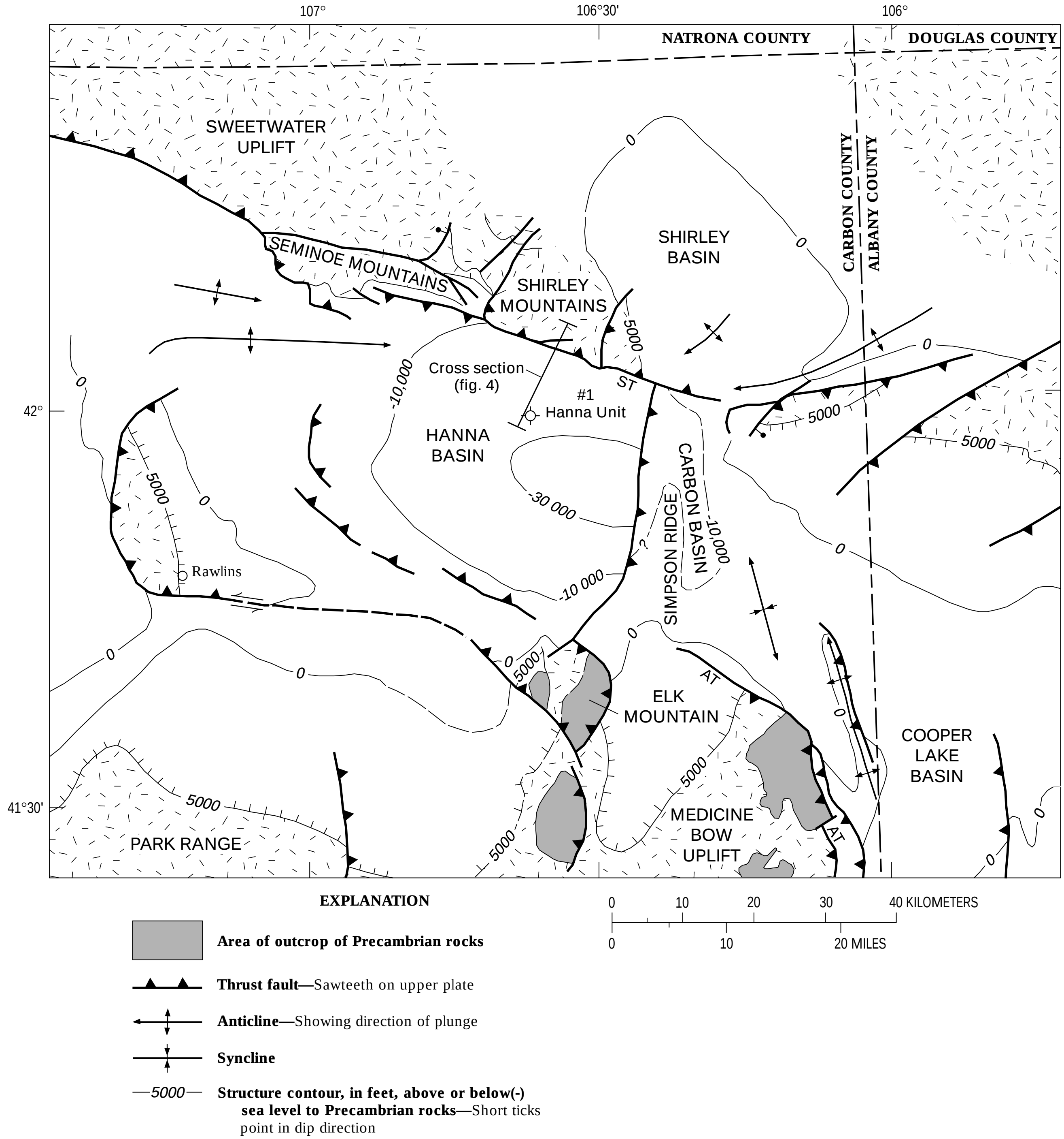


**Figure 3.** Tectonic map of Hanna Basin region, Wyoming, showing the location of No. 1 Hanna Unit well and line of cross section of figure 4. AT, Arlington thrust fault; ST, Shirley thrust fault. Modified from Blackstone (1990; written commun., 1991).

than 15,000 ft (4,572 m) are Upper Cretaceous in age and predominantly marine in origin. Less than 2,500 ft (762 m) of pre-Cretaceous Phanerozoic sedimentary rocks are present (from sections by Blackstone, 1983). Uppermost Cretaceous and Paleocene nonmarine rocks are more than 14,000 ft (4,270 m) thick. The nonmarine formations penetrated are gas prone, and these more shallowly buried rocks are being exploited for coal-bed methane.

The major compressional structural framework along the southern margin of the basin was defined by Beckwith (1941). Dobbin and others (1929) named and mapped the Tertiary rocks of the basin. Gill and others (1970) discussed the stratigraphy and nomenclature of Upper Cretaceous and lower Tertiary rocks in the area, and they indicated that a major unconformity is present between the Paleocene

Hanna Formation and Cretaceous rocks in the northern Hanna Basin. A deep drill hole to the south of the surface expression of this unconformity shows, however, that at least 6,500 ft (1,980 m) of intervening rocks is present within the basin (fig. 3); most of these rocks are now known to be early to middle Paleocene in age (Cavaroc and others, 1992). The basin contains numerous Upper Cretaceous to Paleocene coals (Glass, 1975; Glass and Roberts, 1984) for which precise palynologic dates have not been previously reported. Eight carefully selected samples from these formations, provided by Dr. G.B. Glass, State Geologist of Wyoming, were processed for pollen by D.J. Nichols of the U.S Geological Survey. The results indicate that virtually the entire coal-bearing section of the Hanna Basin, a more than 8,150-foot-thick (2,480 m) sequence primarily composed of nonmarine siliciclastic rocks, is Paleocene in age (table 1).

Recent work by Cavaroc and others (1992) shows detailed palynostratigraphic biozones of the Ferris and Hanna Formations in the Hanna, Carbon, and Cooper Lake basins. The Ferris Formation in the Hanna Basin is as thick as 5,500 ft (1,676 m) and is dated as $P_1$–$P_3$ in age (early to early middle Paleocene) above the basal conglomeratic sandstone of the Ferris. The overlying Hanna Formation is as thick as 6,500 ft (1,981 m) and is dated as $P_3$–$P_6$ in age (early-middle to late Paleocene)(Cavaroc and others, 1992). Zone $P_6$ is thinner than $P_5$ and is directly overlain by a thick carbonaceous shale (gyttja formed in a lake) dated as Eocene in age. In the Carbon Basin east of the Hanna Basin (fig. 3), the Ferris Formation is not present, and the Hanna Formation is as thick as 1,100 ft (335 m) and represents $P_4$–$P_5$ time (late-middle Paleocene–late Paleocene). Farther to the east-southeast in the Cooper Lake Basin, the Hanna Formation is as thick as 680 ft (207 m) and is dated as $P_5$–$P_6$ in age (late Paleocene).

Palynomorph dates and crossbed measurements (Ryan, 1977; Cavaroc and others, 1992) of the Ferris and Hanna Formations in the Hanna Basin area suggest that a northeast-flowing fluvial system drained through the rapidly subsiding Hanna Basin from latest Cretaceous through $P_3$ time (early-middle Paleocene). The outlet may have been in the area of the present Shirley Basin. Provenance to the south may have been the western flank of the broad latest Cretaceous Front Range uplift. This fluvial system initially was in the form of high-gradient braided streams and evolved into low-gradient meandering-anastomosed streams. The thick Ferris coals, as thick as 30 ft (9 m), accumulated in low-lying mires of these low-gradient streams. The boundary ($P_3$) between the Ferris and Hanna Formations is marked by conglomeratic sandstone found in east-flowing high-gradient braided streams. These high-gradient streams evolved into a southeast-through-flowing low-gradient, meandering and anastomosed stream during $P_3$–$P_6$ time. The thick Hanna coals (as thick as 30 ft [9 m]) accumulated in associated low-lying mires. Provenance shifted to the Granite, Seminoe, and Shirley Mountains in late Paleocene time.

The fluvial paleodrainage system, consisting of braided to meandering streams, flowed to the southeast from the Hanna Basin to the Carbon Basin during $P_4$–$P_5$ time (late middle Paleocene to early late Paleocene). The Carbon Basin was either a nondepositional or an erosional area in $P_1$–$P_3$ time, and it began to subside during $P_4$ time. Thick Hanna coals (as much as 20 ft [6 m]) in the Carbon Basin accumulated in low-lying mires of the braided and meandering fluvial systems.

The fluvial system drained to the southeast into and through the Cooper Lake Basin during $P_5$–$P_6$ time. A shift in dispersal of fluvial sediments from south to northeast in the Cooper Lake Basin developed during late $P_5$–$P_6$ time. Conglomerate and conglomeratic sandstone dispersed by alluvial fans from the Medicine Bow Range into the Cooper Lake Basin suggests that uplift of the hanging wall of the Arlington thrust (and thus thrusting) was occurring at this time.

Eocene rocks in the Hanna and Cooper Lake Basins are marked by thick carbonaceous shale and mudstone and a few coarse-grained, conglomeratic sandstone beds. Carbonaceous shale in the northern part of the Hanna Basin indicates gyttja or shallow lake and paludal deposition, whereas mudstone in the Cooper Lake Basin suggests a deeper lake deposit. The gyttja is overlain by burrowed

**Table 1.** Summary of results of palynologic studies of coal samples from the Ferris and Hanna Formations, Wyoming.

[Palynologic analyses by D.J. Nichols; samples provided by Dr. Gary Glass, State Geologist, Wyoming Geological Survey. Bed designations are given, sampled intervals described, and relative positions in the sequence shown in Glass (1975) and Glass and Roberts (1984)]

| Stratigraphic unit | Bed | Sample number | Geologic age and zone |
|---|---|---|---|
| Hanna Formation | 80 | 74–24 | Late Paleocene, probably zone P5.[1] |
| Hanna Formation | 76 | 75–14 | Late Paleocene, probably zone P5.[1] |
| Hanna Formation | Brooks Rider[2] | 75–14 | Middle Paleocene, zone P3. |
| Ferris Formation | 65 | 75–16 | Early Paleocene, probably zone P2. |
| Ferris Formation | 60 | 77–6 | Early Paleocene, probably zone P2. |
| Ferris Formation | 25 | 77–14 | Early Paleocene, possibly zone P1. |
| Ferris Formation | 24 | 74–24 | Early Paleocene, possibly zone P1. |

[1]Late but not latest Paleocene (D.J. Nichols, written commun., 1991).

[2]Near base of Hanna Formation.

coarsening-upward mudstone, siltstone, and rippled sandstone beds. Scouring of the rippled sandstone by a channel sandstone indicates a delta front–distributary channel complex. This sequence is overlain by carbonaceous shale that contains fish remains indicating expansion of the lake. The lake deposit, in turn, is overlain by a coarsening-upward delta-front deposit. Thus, during the early Eocene, the Hanna trough was transformed into a closed lacustrine basin, probably resulting in rapid subsidence and (or) rising of groundwater table above a broad alluvial floodplain.

Thus, infilling of the Hanna and associated basins and direction of dispersal of fluvial sediments were controlled by uplift of a south-southwestern source area (west flank of early Front Range uplift) during $P_1$–$P_3$ time followed by uplift of an active northern source area (Granite-Seminoe-Shirley Mountains) during $P_3$–$P_5$ time and succeeded by a southern source area (Medicine Bow Mountains) during $P_5$–$P_6$ time. The high ash and sulfur content in Hanna Basin coals resulted from erosion of Cretaceous marine fine-grained sediments from nearby uplifts, from which detrital and soluton loads entered restricted fluvial pathways and were ponded, interacting with low-lying mires in the rapidly subsiding basin (Ellis and others, 1992). Ponding of these fine-grained sediments continued into the Eocene lake that extended from the Hanna to the Cooper Lake Basin, and during this time rapid subsidence also was accompanied by closure of the Hanna trough.

Twenty-five vitrinite samples from the No. 1 Hanna Unit well (fig. 3) have been analyzed by Ben Law of U.S. Geological Survey. This dry hole was drilled to a depth of 12,496 ft (3,809 m) but did not reach the base of the Ferris Formation. To a depth of almost 10,000 ft (3,048 m), vitrinite reflectance values are less than 0.7 $R_o$ percent. Below 10,000 ft, vitrinite values increase rapidly to a value of 1.23 $R_o$ percent near the bottom of the hole at a depth of 12,485 ft (3,805 m). The high reflectance values near the base of the drill hole suggest that the more deeply buried marine Cretaceous rocks in the basin should yield thermogenic natural gas; however, only one small gas field, on the northwest flank of the basin, has been developed.

The Hanna Basin is surrounded by Laramide thrust faults that are imprecisely dated. The coal-bearing nonmarine sequence represented by the Hanna and Ferris Formations may represent the time of maximum thrust-related subsidence. This more than 12,000-foot (3,657 m)-thick Paleocene sequence (Cavaroc and others, 1992) may be corrected for compaction (Angevine and others, 1990). If an original mean porosity of 45 percent and a present mean porosity of 25 percent (both very rough estimates) are assumed, then a simple decompaction coefficient of 1.36 results. Using this coefficient to expand the presently known conservative thickness of at least 12,000 ft (3,657 m), more than 16,000 ft (4,974 m) of subsidence may have occurred during the Paleocene in the northern part of the Hanna Basin during a period of about 8.6 m.y., or roughly 1.9 ft (0.57 m) per $10^3$ years decompacted or 1.4 ft (0.425 m) per $10^3$ years uncorrected for compaction. These values compare to Cenozoic subsidence rates in southern California in small pull-apart basins of 2.3 ft/$10^3$ years (0.7 m/$10^3$ years) in the Eocene-Miocene and 3.3 ft/$10^3$ years (1.0 m/$10^3$ years) in the post-Miocene (Yeats, 1978), in which the extreme subsidence rates are driven by major strike-slip faulting. Representative tectonic subsidence histories are given for various types of basins by Angevine and others (1990, fig. 6.1); maximum subsidence rates for foreland basins range from 0.085 to 0.57 ft/$10^3$ years (0.02–0.17 m/$10^3$ years), whereas rates for strike-slip basins range from 0.5 to 2.18 ft/$10^3$ years (0.15–0.66 m/$10^3$ years). Clearly, anomalously high subsidence rates occurred in the Hanna Basin, well outside the average range for foreland basins but well within the range of rates typical of strike-slip related basins. The northern margin of the Hanna Basin is interpreted to represent the locus of a significant zone of latest Cretaceous to Paleocene accommodation (strike-slip) faulting at the northern margin of the zone of east-west Laramide shortening represented by the present Colorado Front Range and Laramie Range.

The sequence of tectonic events in the Hanna Basin region are as follows: first, development of a sequence of thick marine Upper Cretaceous rocks that trends east-west across southern Wyoming; second, partial isolation of the Hanna Basin as a subarea of the Greater Green River Basin by early Paleocene growth of the Granite Mountains–Shirley Mountains transpressive zone to the north; third, southward tilting, probably in mid-Paleocene time concurrent with growth of the Sweetwater uplift and initial development of the Shirley thrust fault along the northern margin of the basin. The fourth and final phase of structural growth, uplift of the Medicine Bow Mountains and Rawlins uplift concurrent with development of the Arlington thrust fault, probably began in late Paleocene time. The inferred geometry (fig. 4) of the northern margin of the basin suggests that major gas accumulations may be present in the undrilled northern part of the basin beneath the Shirley thrust fault, provided that gas generation continued during and after thrusting. Seismic data (Kaplan and Skeen, 1985) do not clearly define the structure of the northern margin of the Hanna Basin. Much gas may remain to be found in deep Rocky Mountain foreland basins if the scenario described here is correct and if this type of tilting prior to thrusting has occurred in other areas. The high vitrinite reflectance values at depths greater than 10,000 ft (3,048 m) suggest that the deeper Cretaceous units should also yield natural gas; however, only one small gas field has been developed. Very little deep drilling has been conducted in the Hanna Basin, unlike other basins to the west and north, and substantial amounts of deep gas may yet be found in this basin.

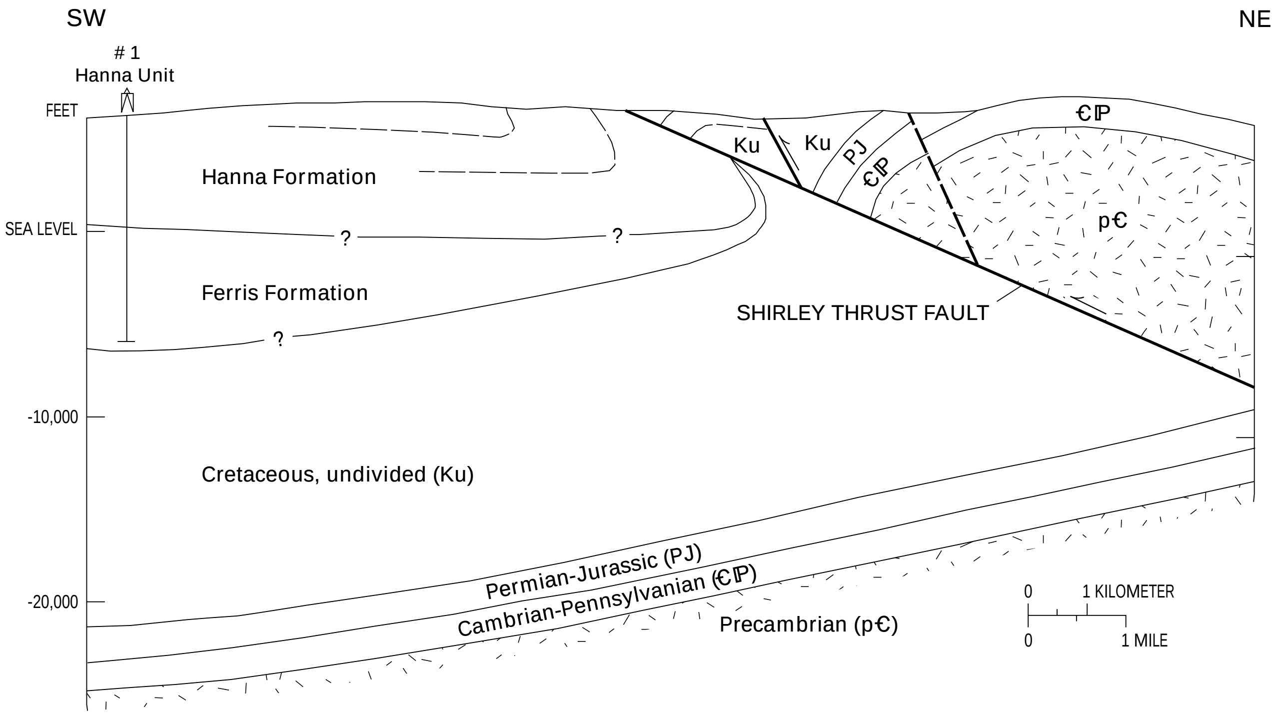


**Figure 4.** Schematic cross section through northern part of Hanna Basin, Wyoming. Line of cross section shown in figure 3. Modified from Blackstone (1983).

## WIND RIVER BASIN

The Wind River Basin, northwest of the Hanna Basin, is separated from the Hanna Basin by the Granite Mountains–Sweetwater uplift (fig. 5), which may have begun to grow in mid-Cretaceous time (Merewether and Cobban, 1986) and was a positive element in Campanian time (Reynolds, 1976). Upper Cretaceous rocks thicken northeastward in the Wind River Basin to more than 18,000 ft (5,486 m). The Walton-Bullfrog field in Upper Cretaceous Frontier Formation sandstone in the northeastern part of the basin (fig. 5) contains the deepest producing Cretaceous gas reservoir in the Rocky Mountain region (more than 18,700 ft [5,700 m] deep). Other significant nearby ultradeep oil and gas fields include West Poison Spider and Tepee Flats; the latter field is beneath the lip of the Casper arch, from which it is separated by a major blind basement-involved thrust system that dips northeastward beneath, and is responsible for, the arch.

The deep Madden gas field in the northern part of the Wind River Basin (in which Madison Limestone and Bighorn Dolostone gas reservoirs are as deep as 23,500–23,900 ft [7,162–7,284 m]) is in front of (south of) the Owl Creek thrust fault that bounds the northern margin of the basin and is likely continuous with the thrust under the lip of the Casper arch. The Madden anticline (fig. 5), the locus of this growing giant gas field, is cored by a thrust wedge, and the north-bounding Owl Creek thrust fault has more than 35,000 ft (10,668 m) of structural relief (Dunleavy and Gilbertson, 1986), comparable to that of the Wichita frontal fault system along the southern margin of the Anadarko Basin (Perry, 1989). The Wind River Basin is thus bounded on two sides by thrust faults, whereas the Hanna Basin is almost surrounded by thrust faults (figs. 3, 5).

The Wind River Basin was partitioned from the remainder of the Rocky Mountain foreland in late Paleocene time by growth of the Casper arch, which led to internal drainage as represented by Lake Walton (fig. 5) (Keefer, 1965). Isolation from long-distance migration of hydrocarbons from previously downdip areas to the west and southwest occurred earlier, during latest Cretaceous to early Palocene time.

The Wind River Basin occupies a critical position with respect to the sequential development of Laramide structure in Wyoming. Conglomerate in the Upper Cretaceous Lance Formation in the northwestern part of the Wind River Basin, nearest the Wind River uplift, contains granule-size fragments and scattered pebbles of chert, siliceous shale, and porcellanite (Keefer, 1965). Here the Lance is about 1,150 ft (351 m) thick (Keefer, 1965, p. A17), and only the lower part is conglomeratic. Keefer found no definite evidence for uplift of the Wind River Range during Cretaceous time, but his control was inadequate along the southwestern margin of the basin (contours dashed, no control points within 30 mi

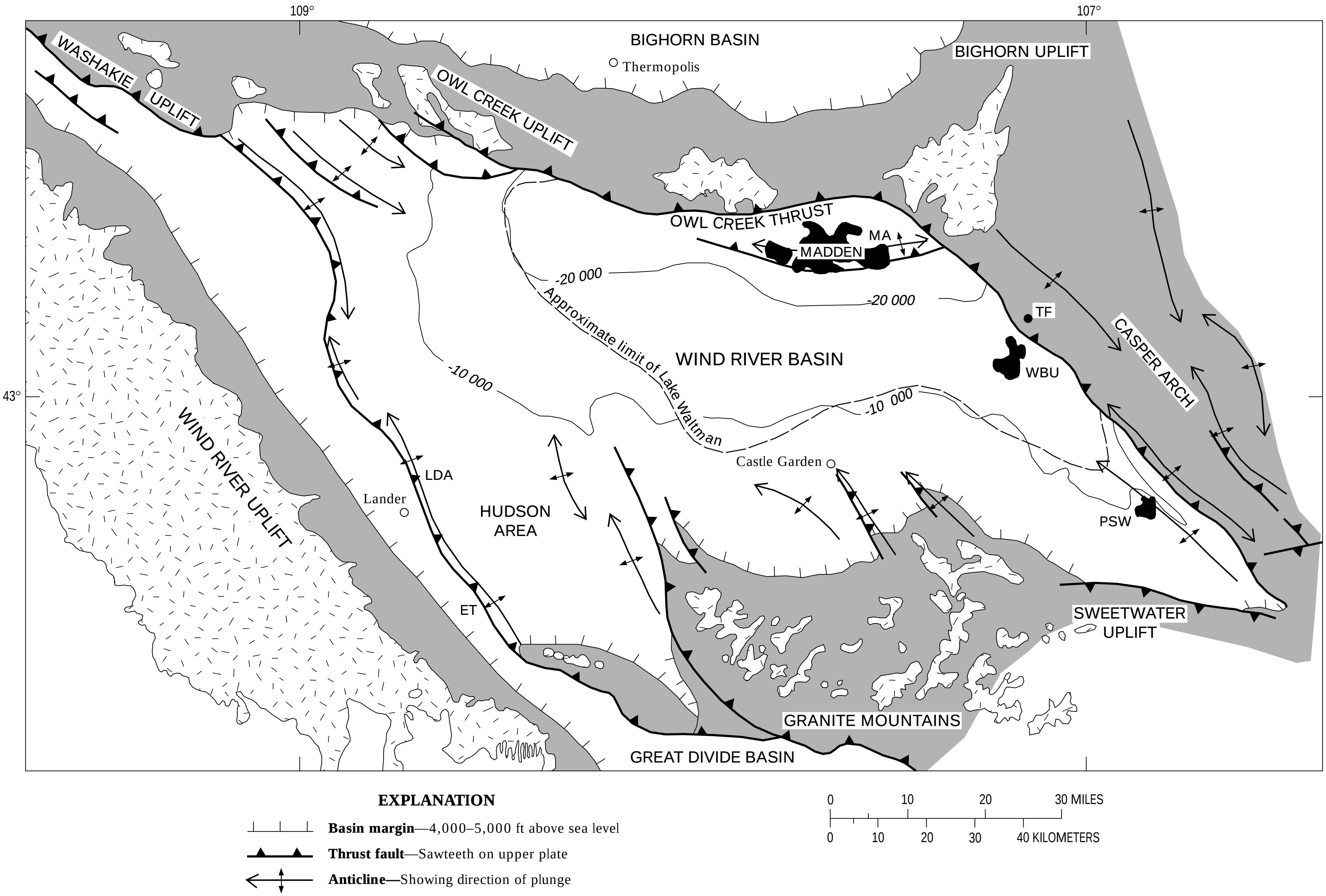


**Figure 5.** Tectonic map of Wind River Basin region, Wyoming. Selected contours (in feet) for top of Pennsylvanian and Permian Minnelusa and Phosphoria Formations. ET, Emmigrant Trail thrust fault; LDA, Little Dome anticline; MA, Madden anticline; PSW, West Poison Spider field; TF, Tepee Flat field; WBU, Walton-Bullfrog field.

[50 km] of the northeastern flank of the Wind River Mountains, his fig. 9). The above described conglomerate in the lower part of the Lance was probably eroded from Frontier and older Mesozoic rocks exposed on the growing Wind River uplift.

Murphy and Love (1958) inferred that a broad, domal uplift occurred in latest Cretaceous time on the southeastern flank of the Wind River Basin in the area of the present Granite Mountains, and Keefer (1965) made similar conclusions. In a summary of the Laramide history of the Granite Mountains area, Love (1971) indicated that uplift of this area did not begin until latest Cretaceous time and culminated in the earliest Eocene time. He suggested that the early phase of this uplift may have been coextensive with that of the south-central part of the Wind River Range.

Investigations by Flores and Keighin (1993), Flores and others (1993), and Nichols and Flores (1993) suggest that the Wind River Range was an active source area about $P_3$ (early middle Paleocene) time. Flores and Keighin (1993) described conglomerate and conglomeratic sandstone as thick as 500 ft (152 m) in the upper part of the lower member of the Fort Union Formation. The conglomerate is made up predominantly of quartzite probably derived from Precambrian–early Paleozoic metaquartzite. Paleocurrent measurements from clast imbrication and trough crossbeds show a west-northwest provenance. Flores and Keighin (1993) suggested that these rocks were deposited in east-southeast-flowing braided streams along a structural paleovalley that occupied the Shotgun Butte area. They also reported that the overlying Shotgun Member ($P_4$–$P_5$ age) (Nichols and Flores, 1993) in the same area was deposited in shortheaded meandering and anastomosed streams that drained into Lake Waltman. Flores and others (1993) described Fort Union conglomerates dominated by quartzite in the Hudson area at the southwestern part of the Wind River Basin (fig. 5). These Fort Union conglomerates are $P_3$ in age or older (early middle Paleocene). Underlying coals of the Mesaverde Formation are Campanian in age (D.J. Nichols, written commun., 1993). Paleocurrent measurements from clast imbrication of the conglomerate and trough crossbeds of the conglomeratic sandstone show northeastward dispersal of braided streams, a dispersal direction that suggests uplift of the Wind River arch at this time. In the Hudson area a high may have existed on which an incised paleovalley was developed (Flores and others, 1993). This high, which extended eastward and was flanked by the Emmigrant Trail thrust fault, was an area of net erosion and (or) nondeposition during Maestrichtian to early Paleocene time prior to deposition of the Fort Union conglomerates. A northwest line from the Emmigrant Trail thrust fault to southwest of Little Dome anticline (Flores and Keighin, 1993; Flores and others, 1993) represents a hinge line east of which was a rapidly subsiding subbasin of the western Wind River Basin proper. Investigation of the Fort Union conglomerates and conglomeratic sandstone by Flores and others (1992) at Castle Garden (fig. 5) on the south-central flank of the Wind River Basin indicates the appearance of granitic pebbles, cobbles, and boulders by $P_3$ time (early middle Paleocene). Paleocurrent measurements of trough crossbeds show northeastward dispersal associated with a meandering fluvial system. This dispersal direction suggests that the sediments were derived from the Granite Mountains. The age of first appearance of igneous detritus coincides with the $P_3$ biozone determined for dispersal of the Hanna Formation in the Hanna Basin from the Seminoe-Granite Mountains and Shirley Mountains. An increase in amount of conglomeratic boulders during $P_3$–$P_6$ time reflect continued uplift of the Granite Mountains, which provided sediments into Lake Waltman. Nichols and Flores (1993) suggested that the $P_3$–$P_6$ arkosic conglomeratic sandstone at Castle Garden is in part correlative with the $P_5$–$P_6$ Waltman Shale Member of the Fort Union Formation in the northeastern part of the basin. The Waltman Shale Member represents deposition in a lacustrine setting, as a result of rapid subsidence probably influenced by the Owl Creek thrust fault (transpressional or strike-slip) (Molzer, 1992; Paylor, 1992) and closure of the basin by the Casper arch. The Waltman Shale Member is as thick as 3,000 ft (914 m). Lake Waltman was fed by shortheaded meandering and anastomosed streams and associated fan deltas. The Waltman Shale Member served as source rock for high-paraffin oils in the Fort Union reservoir sandstones (Palacas and others, 1992); however, gas and condensate are the most common hydrocarbons in reservoir sandstones in the Fort Union Formation (Wyoming Geologic Association, 1989).

Keefer (1965) concluded that, although the Cretaceous-Tertiary boundary in the Wind River Basin is generally conformable, extensive downwarping occurred at this time along the present-day northern margin of the basin. Along the northeastern margin of the Wind River Basin, Keefer (1965) observed that the oldest conglomerate zones are in the lower Eocene Indian Meadows Formation. The oldest arkosic conglomerate in this part of the basin is at the base of the Lost Cabin Member of the overlying Eocene Wind River Formation. The presence of extensive lacustrine sediments, which first appeared in the Wind River Basin in late Paleocene time (Nichols and Ott, 1978; Phillips, 1983), is indicative of internal drainage that likely reflects initial growth of the Casper arch and the Owl Creek uplift (fig. 1), which closed the outlets of the basin.

Keefer (1965) estimated that more than 8,800 ft (2,682 m) of middle to late Paleocene uplift occurred in the Owl Creek Mountains and that almost 10,500 ft (3,200 m) of subsidence occurred in the adjacent Wind River Basin; these amounts indicate a cumulative vertical separation (uplift+subsidence) rate of slightly more than 4 ft/$10^3$ years (1.2 m/$10^3$ years). Keefer estimated that an additional 8,500 ft (2,591 m) of uplift and an additional 5,600 ft (1,707 m) of subsidence occurred in the early Eocene, yielding a cumulative vertical separation rate of almost 4 ft/$10^3$ years (1.2 m/$10^3$ years). He showed that thrust faults of the Casper arch cut the lower Eocene Indian Meadows Formation and that the rocks deformed by this thrusting are erosionally truncated by the overlying lower Eocene Wind River Formation. The relations date the cessation of major Laramide deformation in the area as early Eocene. These rates are consistent with relatively late Laramide strike-slip-dominated transpressional deformation along the northern margin of the Wind River Basin, similar to the earlier Laramide transpressional boundary along the northern margin of the Hanna Basin to the south. Structurally trapped deep gas may still be discovered north and northwest of the Madden anticline in the northern part of the Wind River Basin.

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