SYNOROGENIC BASIN DEPOSITS AND ASSOCIATED LARAMIDE UPLIFTS IN THE MONTANA PART OF THE CORDILLERAN FORELAND BASIN SYSTEM

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ABSTRACT

The North American Cordilleran Foreland Basin System extended from the eastern edge of the Cordilleran Sevier fold-thrust belt in western Montana across central and eastern Montana from Jurassic into early Paleogene time. In the central and southern Montana part of the system, isolated Laramide basement-cored uplifts and sedimentary basin pairs characterize the main Laramide Province. The thrust/reverse fault-propagated arches and basins of the province developed primarily during Late Cretaceous through earliest Eocene time, although regionally, initiation of uplift may have occurred during Early to middle Cretaceous time in southwestern Montana. Associated prominent WNW–ESE- and NE–SW-striking linear features reflect Laramide reactivation of basement faults that now extend to the surface, or are entirely in the subsurface. They occur throughout the main Laramide Province and are likely genetically related to development of the uplift-basin pairs.

The asymmetric Laramide Bighorn and Powder River Basins developed in association with uplift along basin-bounding thrust/reverse faults of the Beartooth and Bighorn basement-cored arches, respectively. Thrust loading associated with uplift along the faults propagated asymmetric synclinal basin folds that accommodated the greatest thicknesses of synorogenic deposits adjacent to the range-bounding faults. In Montana, synorogenic deposits of the Bighorn Basin primarily include the Paleocene Fort Union Formation, and those of the Powder River Basin primarily include the Paleocene Fort Union and Eocene Wasatch Formations. The similarly asymmetric Crazy Mountains Basin developed in association with Laramide-style reverse and thrust-fault movement of the ancestral Bridger arch. Synorogenic deposits of the Crazy Mountains Basin primarily include the Upper Cretaceous Livingston Group and the mostly Paleocene Fort Union Formation. An oblique-slip reactivated basement fault that bounds the southern Central Montana Uplift generated the Laramide Bull Mountains Basin primarily during latest Paleocene and early Eocene time. Relative to the other basins, the Bull Mountains Basin did not accommodate much sediment from that uplift. The Black Hills Uplift, which extends into Montana, is the easternmost structure of the Laramide Province. The Black Hills may represent the youngest inception of Laramide uplift, whereas the Laramide uplifts in southwestern Montana may represent the earliest inception.

Other uplifts, basins, and folds associated with reactivated basement structures developed beyond the extent of the main Laramide Province in Montana. These include structures of the western Williston Basin; the Bowdoin Dome; Hoagland, Opheim, Blood Creek, and Circle Basins; and the Bearpaw and Sweetgrass Arches[1].

[1]The original names for *Sweetgrass* Arch, *Bearpaw* Uplift, *Bighorn* Uplift, and *Bighorn* Basin are used in this paper, rather than *Sweet Grass*, *Bears Paw*, and *Big Horn*, applied to certain geographic features. Some reports refer to the *Bull Mountains Basin* as the *Bull Mountain Basin* and to the *Hoagland Basin* as the *Hogeland Basin*.

INTRODUCTION

The main Laramide Province extends from Wyoming northward into the central Montana part of the Cordilleran Foreland Basin System of western North America (Hamilton, 1988; DeCelles, 2004; fig. 1) and is identified as "domains 1 and 2" of the province (Lageson and others, this volume). It is characterized by a series of basement-cored uplifts (Laramide style) and genetically associated, sedimentologically isolated basins that developed primarily during the late Mesozoic to early Cenozoic Laramide orogenic event (Dickinson and others, 1988; fig. 2). The Laramide Province partly overlaps the Sevier fold-thrust belt in southwestern Montana (Kulik and Schmidt, 1988; Schmidt and others, 1988; figs. 1, 2) and in the Bridger Range (Lageson, 1989). Basement-involved uplift may have begun as early as Early Cretaceous in southwestern Montana (DeCelles, 1986; Schwartz and DeCelles, 1988; Carrapa and others, 2019), but probably not until Late Cretaceous (Fan and Carrapa, 2014) to Paleocene time (Lisenbee and DeWitt, 1993) to the east.

The Western Interior Cretaceous seaway that had occupied the foreland (Hendrix, this volume) retreated as orogenesis from contractional tectonism increased in the west. In the western part of the Cordilleran Foreland Basin System, the Upper Cretaceous Livingston Group (fig. 3) records the retreat of the seaway and sedimentation related to Laramide-style uplift. The Upper Cretaceous marginal marine Fox Hills and overlying non-marine Hell Creek/Lance Formations (fig. 3) were deposited across central and eastern Montana in association with the eastward-migrating western margin of the regressing seaway. Deposition of the Hell Creek Formation approximately terminated with the Cretaceous–Paleogene boundary extinction event (fig. 4), including the demise of the dinosaurs (e.g., Fastovsky and Sheehan, 2005; Clemens and Hartman, 2014). Deposition of the Paleocene Fort Union Formation followed this event. In southeastern Montana and isolated areas of the Bears Paw Mountains of north-central Montana (Brown and Pecora, 1949), the latest Paleocene–earliest Eocene Wasatch Formation overlies the Fort Union Formation (figs. 3, 4).

Paleocene and earliest Eocene deposits are sparse and equivocal in the Sevier fold-thrust belt area of western Montana. By contrast, Paleocene deposits are extensive in south-central and eastern Montana, and earliest Eocene deposits occur in a significant part of southeastern Montana (fig. 5).

The Crazy Mountains, Bull Mountains, Bighorn, and Powder River Basins of the Montana part of the main Laramide Province (fig. 2) filled with Late Cretaceous and/or Paleocene—and in some basins lowest Eocene—deposits, which were eroded from associated, coeval, adjacent Laramide uplifts, as well as from uplifts to the west. Synorogenic conglomerate/breccia (Piombino, 1979; DeCelles and others, 1991a); isopach and paleocurrent data (Curry, 1971; Shurr, 1972; Seeland, 1988, 1992; Diemer and Belt, 1991; Whipkey and others, 1991; Belt and others, 1992; Connor, 1992; Hansley and Brown, 1992; Belt and others, 1992; Lisenbee and DeWitt, 1993); sedimentary facies changes (Flores and Ethridge, 1985; Ayers, 1986; DeCelles and others, 1991b; Whipkey and others, 1991); the presence of clasts with distinct lithologies (Merin and Lindholm, 1986; Brown, 1993); unconformities (Belt and others, 1997, 2004; Flores and Bader, 1999); paleolandslides (Garrett, 1963; Belt and others, 2002); and other paleoseismites (Bartholomew and others, 2008; Jackson and others, 2019) record the influence of the Laramide Beartooth, Bighorn, and Bridger Uplifts on sedimentation in their associated basins.

Less pronounced Laramide uplifts and basins occur in northern and easternmost Montana (fig. 1), and are referred to as "domain 3" of the Laramide belt (Lageson and others, 2020). Most synorogenic deposits in this area were eroded or are partly obscured by glacial deposits, so little is known about the relationship between the uplifts and basin sedimentation.

The Laramide Province overlaps the Sevier fold-thrust belt in southwestern Montana (Kulik and Schmidt, 1988; Schmidt and others, 1988; figs. 1, 2). In this area, initiation of Laramide uplift was interpreted from sedimentologic studies of the Early Cretaceous Kootenai and Blackleaf Formations (Schwartz and DeCelles, 1988) and recently based on thermochronological and geochronological data from six basement-cored Laramide uplifts in the overlap zone and the Beartooth Uplift (Carrapa and others, 2019), all within "domain 1" of Lageson and others (2020). The Upper Cretaceous Sphinx Conglomerate (Graham and others, 1986; Ingersoll and others, 1986; DeCelles and others, 1987), and Upper Cretaceous and possibly lower Paleogene conglomerates of the Beaverhead Group (Ryder and Scholten, 1973; Haley and Perry, 1990; fig. 3), were derived from Sevier thrust sheets

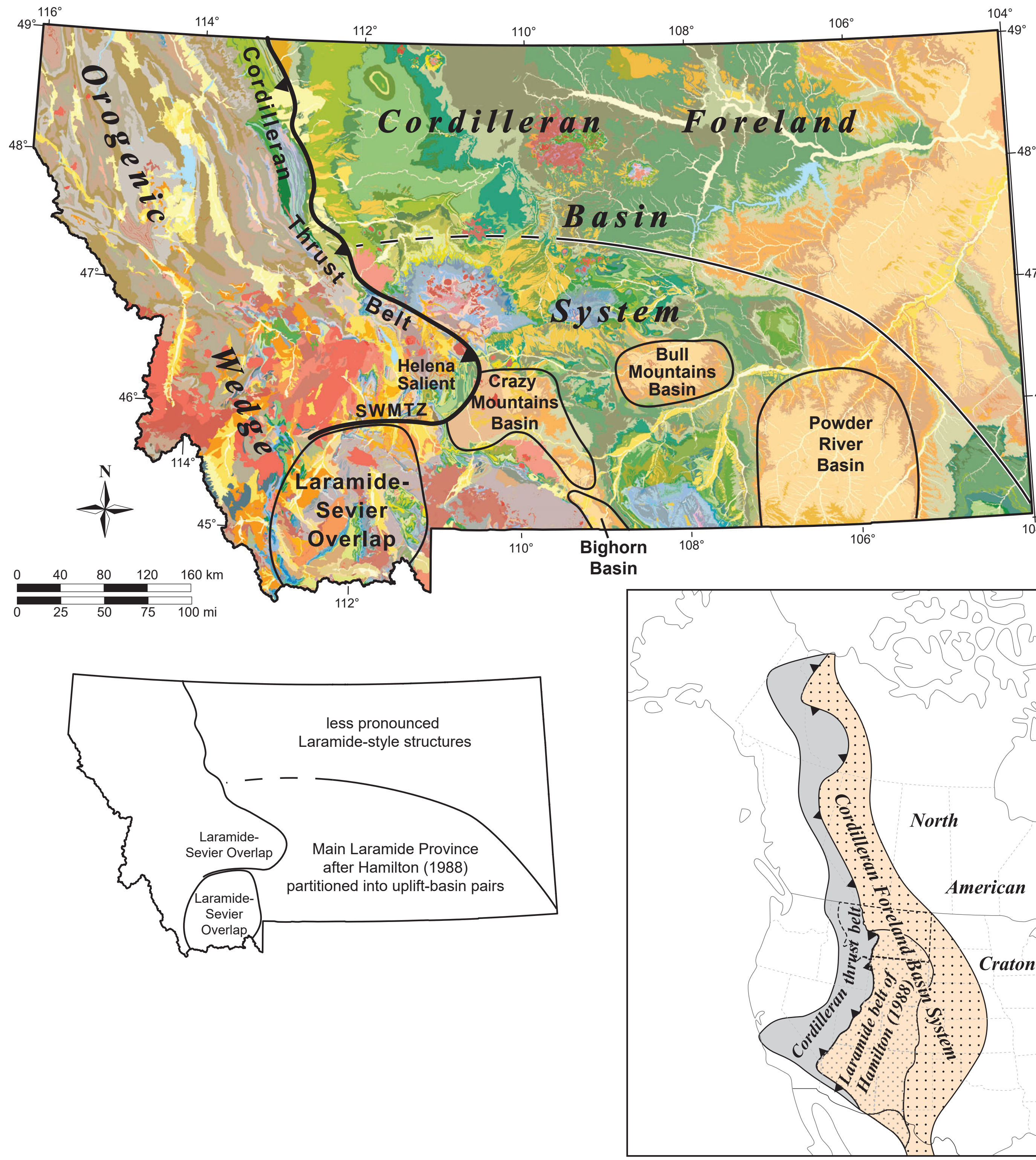


Figure 1. Orogenic provinces and Laramide basins in Montana during Late Cretaceous through earliest Eocene time. Modified from Hamilton (1988), Kulik and Schmidt (1988), and DeCelles (2004). SWMTZ, Southwest Montana Transverse Zone.

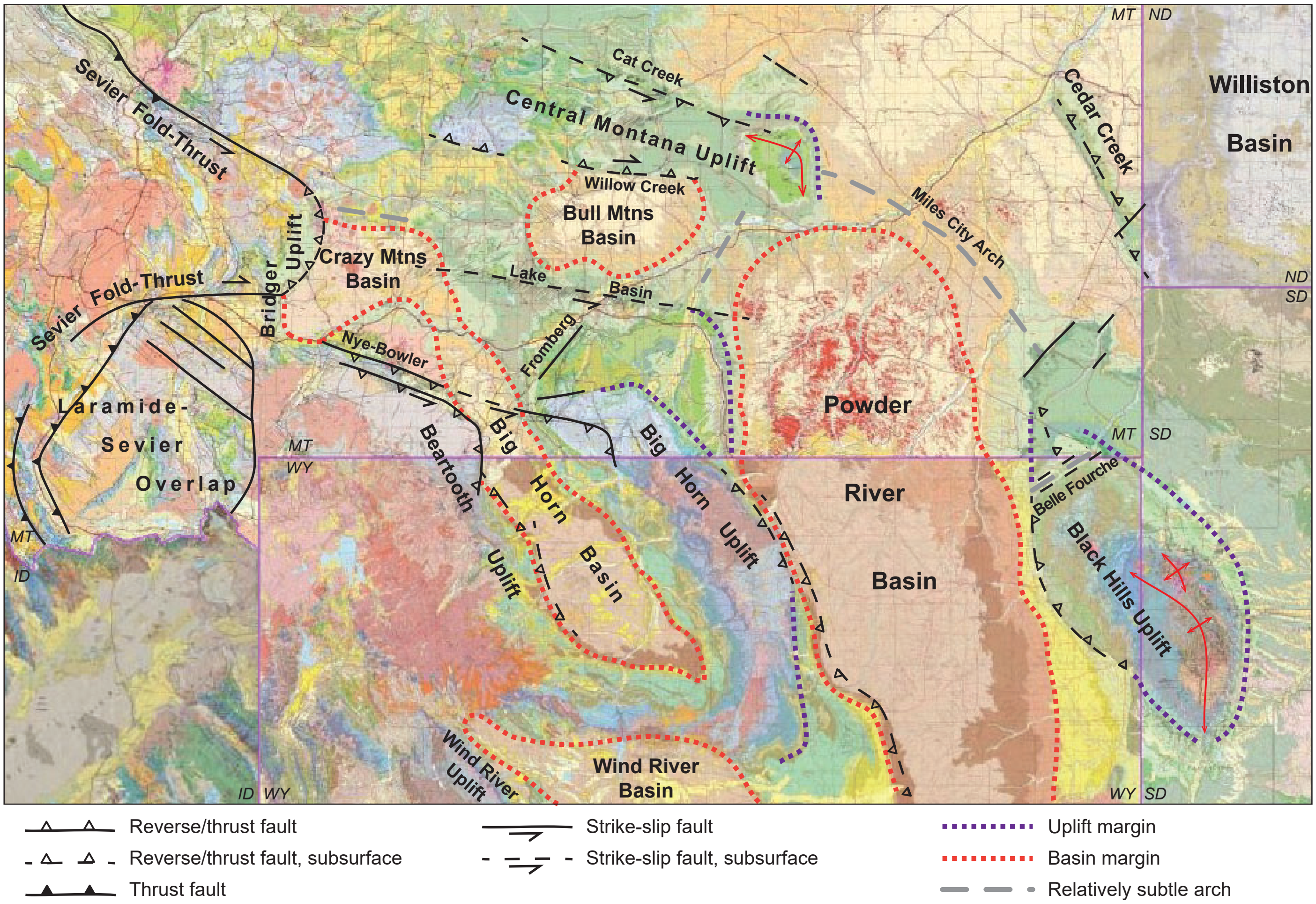


Figure 2. Regional Laramide uplifts, basins, linear features, and other structures. State geologic map base from U.S. Geological Survey National Geologic Map Database.

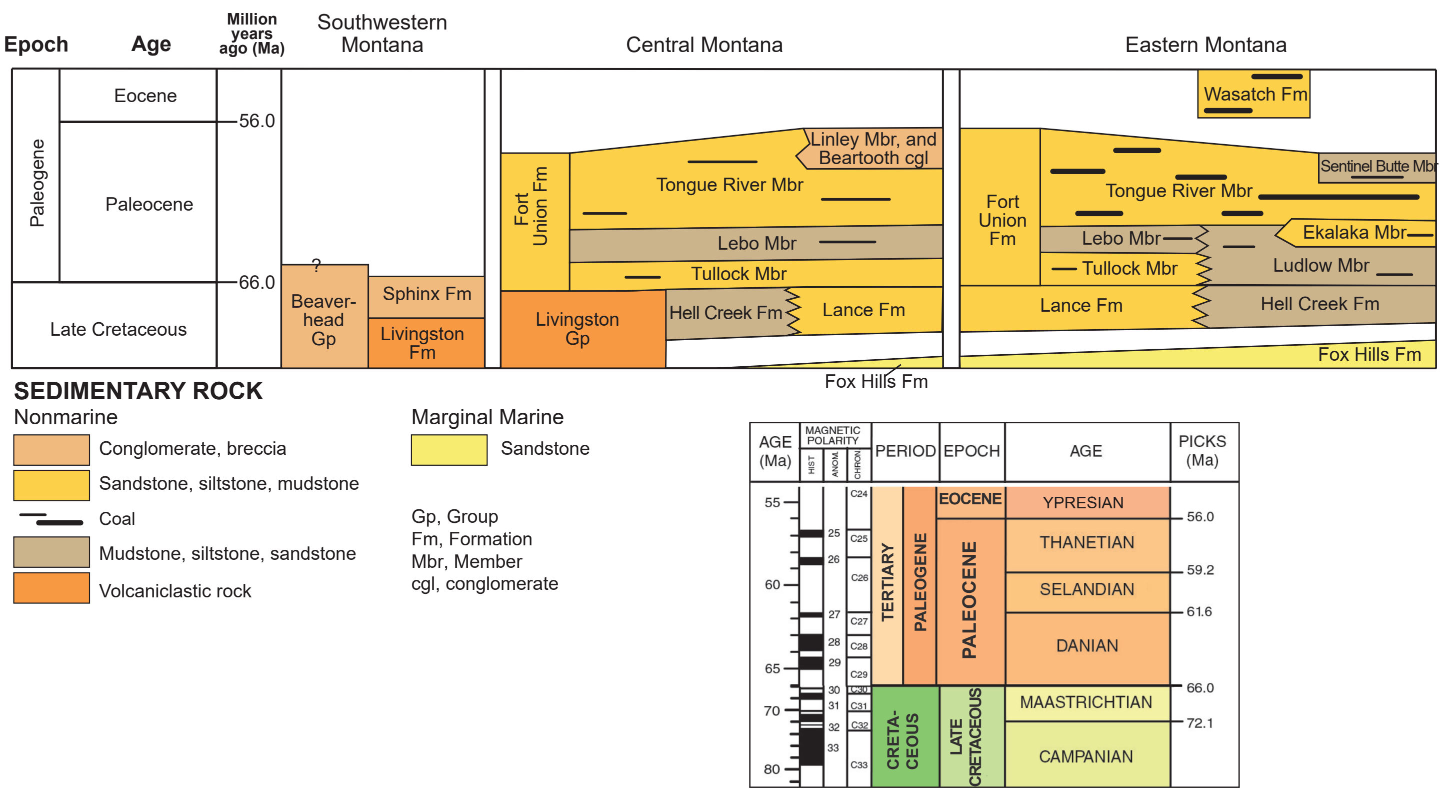


Figure 3. Correlation diagram for synorogenic units in central and eastern Montana, and selected synorogenic units in southwestern Montana. Inset time scale from Walker and others, 2018.

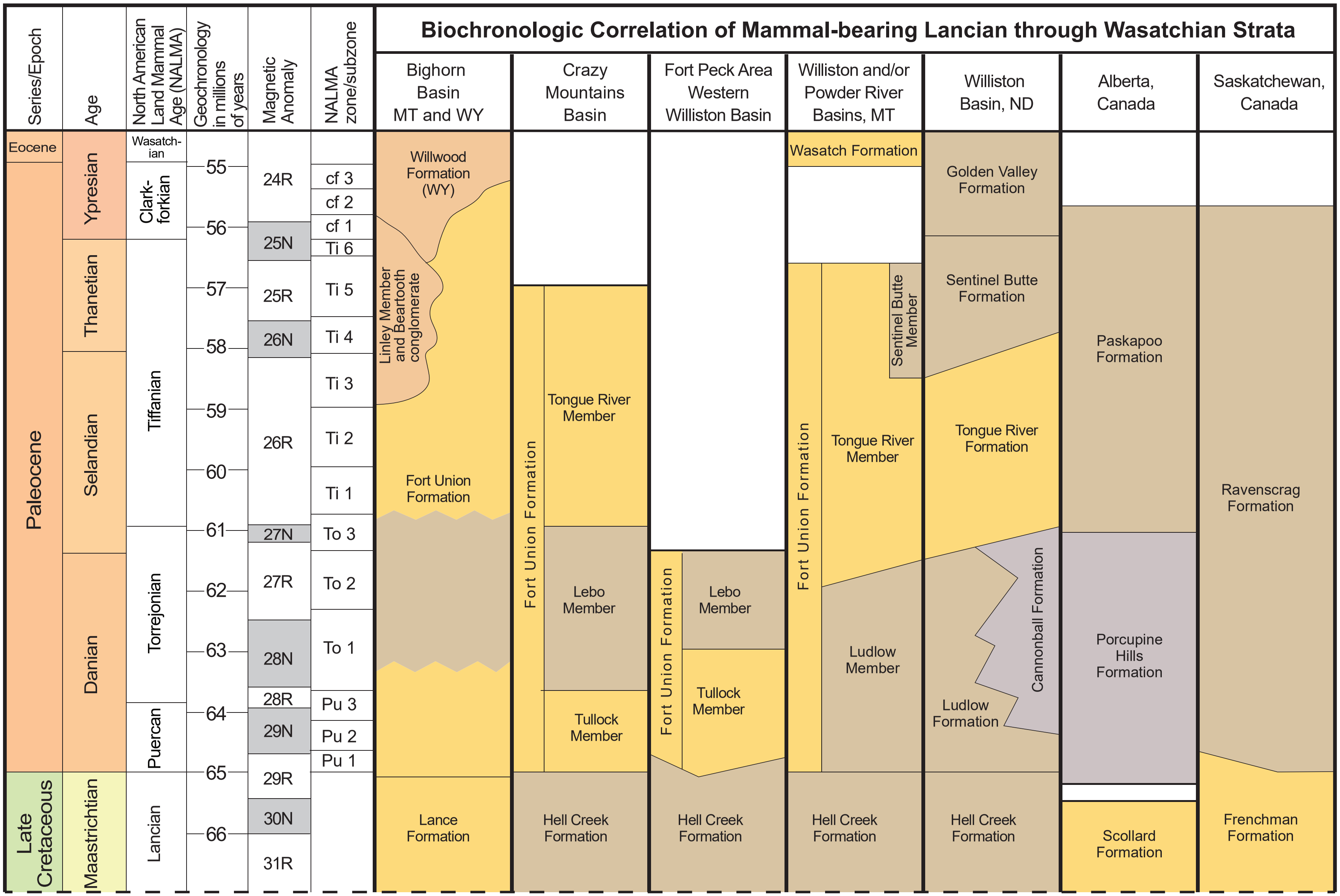


Figure 4. Biostratigraphic (North American Land Mammal Ages), magnetostratigraphic, and chronostratigraphic correlations for upper Hell Creek/Lance, Fort Union, and Wasatch Formations in Montana and correlative units in adjacent areas. Modified from Lofgren and others (2004). Inset time scale in figure 3 shows scale of correlation diagrams in Tertiary papers of this volume (Vuke; Thomas and Sears).

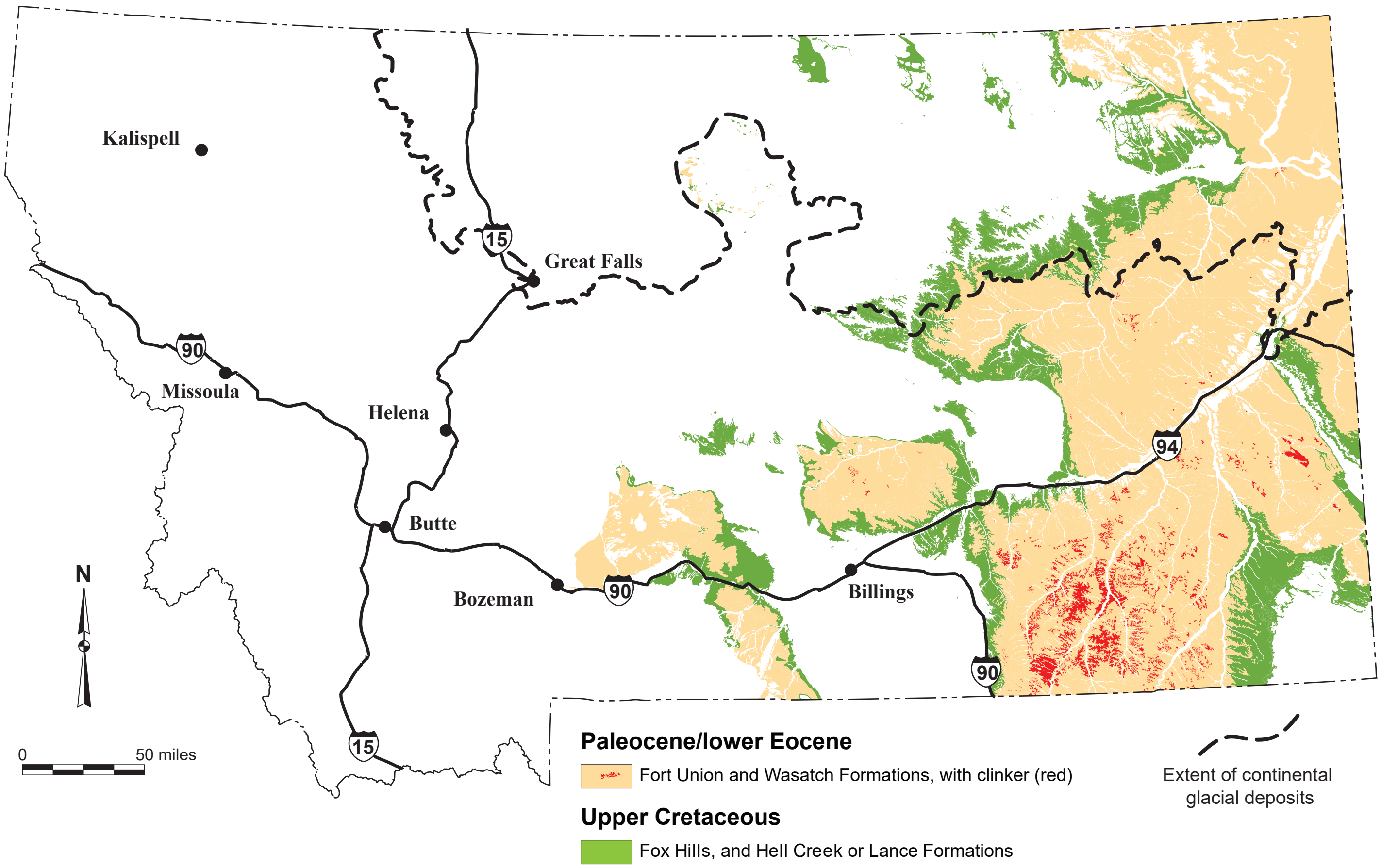


Figure 5. Distribution of Upper Cretaceous Fox Hills, and Hell Creek or Lance Formations; and Paleocene/lower Eocene Fort Union and Wasatch Formations.

and Laramide intra-foreland uplifts in the overlap zone. The conglomerates currently crop out in uplands and on mountaintops, rather than in preserved basins. These and other synorogenic deposits in the Sevier–Laramide overlap area are not discussed in this paper. Rather, the paper focuses on deposits in the part of the Cordilleran Foreland Basin System that partitioned into basement-cored uplifts and associated basins (fig. 1).

UNIT DESCRIPTIONS

Maastrichtian (Upper Cretaceous) formations were deposited during Sevier and Laramide orogenesis in Montana. They are introduced in this section, but the main focus is on the Paleocene Fort Union Formation, which provides the most prominent evidence of sedimentologic response to uplift associated with the Crazy Mountains, Big Horn, Powder River, and Bull Mountains Basins in Montana.

Maastrichtian (Uppermost Cretaceous)

The upper part of the Livingston Group (fig. 3) was deposited as the deep trough of the Crazy Mountains Basin was developing, whereas the generally age-equivalent Fox Hills and Hell Creek/Lance Formations to the east, although broadly influenced by the Laramide event, do not record Laramide basin development related to partitioning of the Cordilleran Foreland Basin System.

Upper Livingston Group

The Maastrichtian Billman Creek and overlying Hoppers Formations are part of the nonmarine upper Livingston Group (Roberts, 1963). The Billman Creek Formation is dominantly massive tuffaceous mudstone interbedded with sandstone and tuffaceous claystone. Sandstone and conglomerate channel deposits make up about one-fourth of the formation and are dominantly filled with volcanic rock fragments derived from the Elkhorn Mountains volcanic field to the west. The overlying Hoppers Formation is dominantly volcanic lithic sandstone interbedded with mudstone and siltstone, also with channel fill sandstones primarily composed of volcanic rock (Roberts, 1972).

Fox Hills and Hell Creek/Lance Formations

The Upper Cretaceous Fox Hills Formation (figs. 3, 4, 6) contains brackish-water trace fossils (Flores and Lepp, 1983; Wilde, 1985; Flight, 2004) and represents marginal marine deposits associated with the last Cretaceous cycle of the retreating Western Interior seaway. The interfingering terrestrial Hell Creek and Lance Formations overlie the Fox Hills Formation in south-central Montana (figs. 3, 4). As mapped in Montana, sandstone beds in the Lance Formation are much thicker and more continuous than sandstone beds in the Hell Creek Formation (Lopez, 2001), and the Lance Formation apparently lacks smectitic beds, which are prevalent in the Hell Creek Formation (Vuke and Wilde, 2004). The Hell Creek Formation (figs.6,7)hasyieldednumerousdinosaurfossils (Horner and Hanson, 2020). A detailed lectostratotype was established for the entire Hell Creek Formation section in Garfield County (Hartman and others, 2014) in the reference area (type area) originally designated by Brown (1962).

Paleocene and Lowest Eocene

Fort Union and Wasatch Formations

Brown (1962) designated the Hell Creek Formation contact with the overlying Fort Union Formation (figs. 3, 4, 7B) as "the base of the lowest coal zone above the latest remains of dinosaurs," restricting the upper contact of the Hell Creek Formation to approximately the Cretaceous–Tertiary (K-T) or Cretaceous–Paleogene (K-Pg) boundary. Prior to that designation, the upper contact of the Hell Creek Formation included what is now the lower member of the Fort Union Formation. Brown's definition of the formation boundary has both synchronous (i.e., latest dinosaur remains) and diachronous (i.e., lowest persistent coalbed) components. Therefore, the Hell Creek/Lance contact with the Fort Union Formation may be, but is not necessarily, coincident with the K-Pg boundary. The Hell Creek–Fort Union contact crosses magnetic polarity zones from west to east (fig. 4), but faunal and floral changes parallel the polarity zones (Archibald and others, 1982).

In the western Crazy Mountains Basin in central Montana, Roberts (1972) defined the contact between the Livingston Group and Fort Union Formation (fig. 3) using sedimentary provenance. Conglomerate in the upper Livingston Group was designated as containing almost entirely volcanic clasts, and conglomerate in the overlying basal Fort Union Formation as containing igneous (intrusive), metamorphic, and sedimentary clasts derived from Precambrian, Paleozoic, and Mesozoic units. Using this contact designation, the lowest part of the Fort Union Formation in the Crazy

Figure 6. Upper Fox Hills Formation showing Colgate Member, and lower Hell Creek Formation near Glendive, Montana.

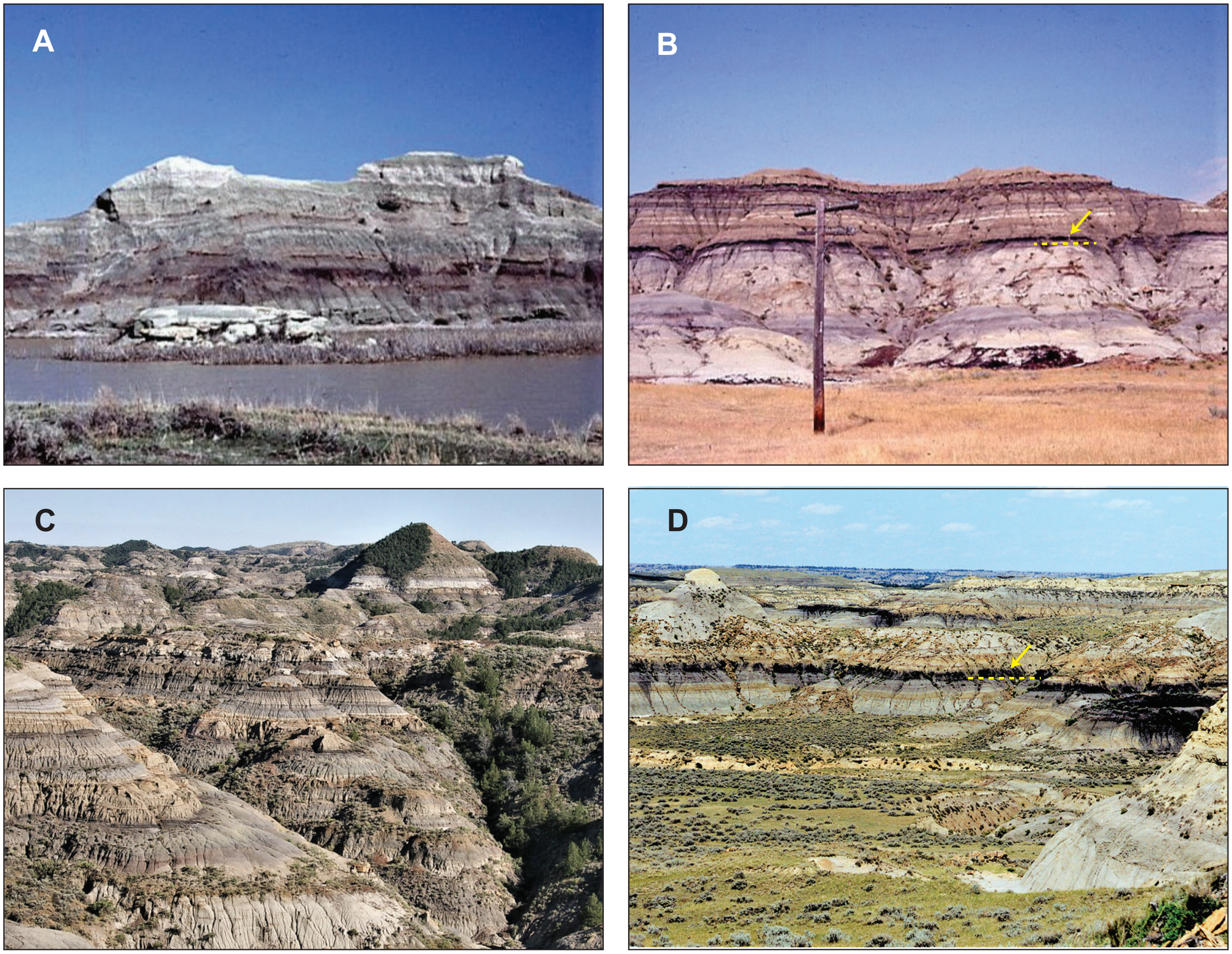


Figure 7. (A) Hell Creek Formation near Webster, Montana. (B) Hell Creek Formation–Ludlow Member contact (~K-Pg boundary) north of Baker, Montana. (C) Ludlow Formation, Makoshika State Park, S.M. Roberts photo. (D) Hell Creek Fm–Tullock Member contact (~K-Pg boundary), Garfield County, Montana, J.H. Hartman photo. Arrows indicate contact positions in (B) and (D).

Mountains Basin is latest Cretaceous (Roberts, 1972), whereas in the Powder River Basin, the uppermost part of the underlying Hell Creek Formation is very earliest Paleocene (Brown, 1993). However, the distinction between clast composition at the Livingston Group–Fort Union Formation contact in the Crazy Mountains Basin was not necessarily discernable by other workers (Dripps, 1992).

The Fort Union Formation ranges from 6,600 ft (2,010 m) thick in the Crazy Mountains Basin (Roberts, 1972) to more than 3,000 ft (915 m) thick in the Powder River Basin. It is only approximately 300 ft (90 m) thick in the Montana part of the Williston Basin (Anna, 1986).

In easternmost Montana, the Paleocene Fort Union Formation is composed of the Ludlow Member and the overlying Tongue River Member (fig. 3), distinguished primarily by color and lithology. The Ludlow Member (fig. 7C) is dominantly gray and brown, greater than 50 percent mudstone, and has a clay fraction dominated by smectite. The Tongue River Member (fig. 8A) is dominantly orange and tan, generally has a higher percentage of sandstone, and a clay fraction dominated by illite and kaolinite (Belt and others, 1992). To the west, the Tullock (figs. 4, 7D) and overlying Lebo Members replace the Ludlow Member (fig. 3). In southeastern-most Montana, the Ekalaka Member (fig. 8B) is in the stratigraphic position of the upper Ludlow Member (fig. 3), but unlike the Ludlow Member, it contains greater than 50 percent sandstone (Belt and others, 2002). All members of the Fort Union Formation consist dominantly of sandstone, mudstone, shale, and coalbeds, but the Tullock, Ekalaka, and Tongue River Members contain a higher percentage of sandstone than the Lebo and Ludlow Members. The Tongue River Member has the

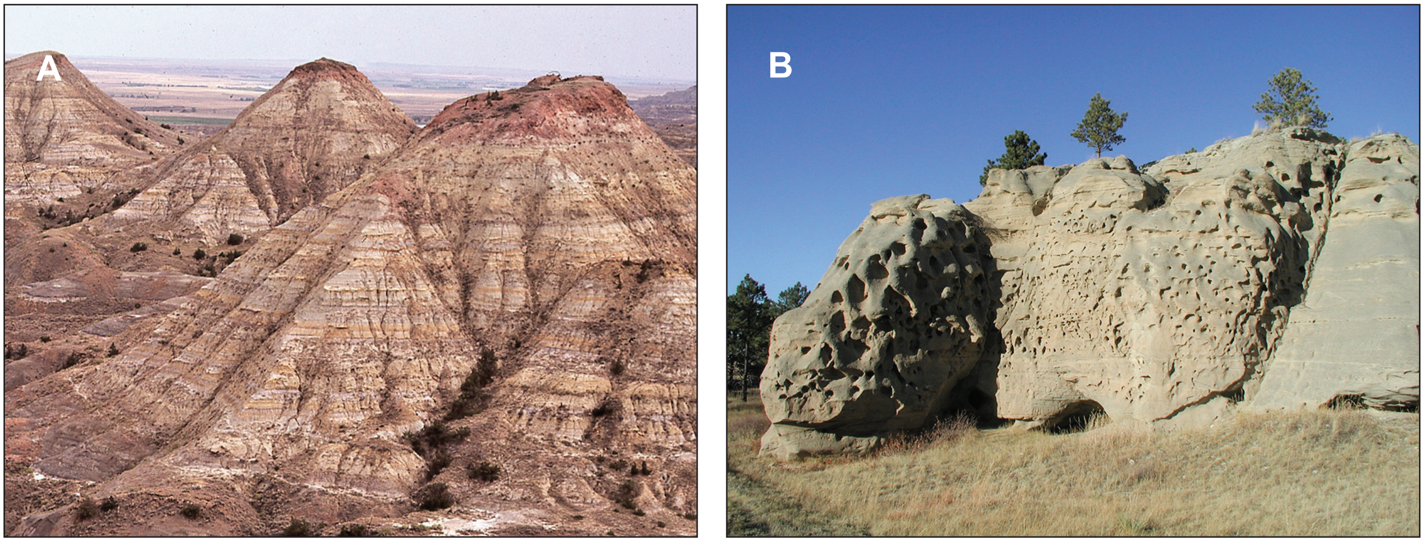


Figure 8. (A) Tongue River Member, near Savage, Montana. (B) Ekalaka Member, near Ekalaka, Montana.

most numerous and thickest coalbeds (fi g. 9A), which have served as a signifi cant energy resource (Gunderson and Wheaton, 2020).

Along the western margins of the western Laramide basins in Montana (fig. 2), synorogenic alluvial fan and braided stream deposits of the Fort Union Formation are dominantly coarse-grained sandstone and conglomerate, and contain a higher percentage of volcanic clasts than deposits located to the east (Piombino, 1979; DeCelles and others, 1991a,b). Clasts of crystalline Precambrian basement rock are present in Fort Union Formation deposits in the Crazy Mountains, Bighorn, and Powder River Basins. The average grain size of fluvial deposits fines toward the basin center, where lacustrine environments in the Lebo Member may be present, with clastic input from both sides of the Bighorn and Powder River Basins (Yuretich and others, 1984; Ayers, 1986). Alternatively, poorly resistant Mesozoic rock sources may have provided the fine-grained sediment in those basins (DeCelles and others, 1991b; Flores and Bader, 1999).

Although Fort Union alluvial fan and braided stream deposits may also be present along basin margins in eastern Montana (Flores and Bader, 1999), sedimentary environments were dominated by coastal plain swamps (Belt and others, 2005) and delta plain swamps (Flores and Lepp, 1983; Belt and others, 1984). Inland swamps were associated with low-energy, anastomosing (Flores and Hanley, 1984; Brown, 1993) and meandering (Diemer and Belt, 1991; Flores, 1983) fluvial systems that produced lignite and subbituminous coal. Coal swamp development was most widespread during times when tectonic uplift caused paleodrainage reorganization (Diemer and Belt, 1991; Belt and others, 1992; Belt, 1993), when related subsidence increased accommodation, or during base-level changes (Flores and Bader, 1999). Isolated, remnant patches of the Sentinel Butte Member of the Fort Union Formation extend a short distance into Montana (Vuke and Colton, 1998; Vuke and others, 2003c; figs. 3, 4) from widespread deposits in North Dakota. Brown and gray (from its smectite clay fraction), more friable deposits of the Sentinel Butte Member overlie the yellow, tan, and light gray Tongue River Member (Jacob, 1975).

The Wasatch Formation (figs. 3, 4) is extensive in the Powder River Basin, primarily in Wyoming. In Montana, it is present only in the northern Powder River Basin, in several fault blocks of limited extent in the Bears Paw Mountains (Brown and Pecora, 1949; Hearn and others, 1964), and along the northern Bighorn Basin in south-central Montana (Raines and Johnson, 1995; Hart, 2012). In the Powder River Basin, the Wasatch Formation overlies the Fort Union Formation both conformably (central basin in Wyoming) and unconformably (at basin margins; Flores and Bader, 1999). In Montana, the Wasatch Formation is dominantly fluvial sandstone with subordinate fine-grained clasts and significant coalbeds (Culbertson and Mapel, 1976), whereas farther south it includes synorogenic conglomerate where it is adjacent to resistant Precambrian and Paleozoic source rock (Love and Christiansen, 2014).

At least four stratigraphic criteria have been used to define the contact between the Fort Union Formation and overlying Wasatch Formation (Seeland, 1993; Flores and Bader, 1999). Depending on the criteria used, the Wasatch Formation is considered entirely earliest Eocene (e.g., Denson and others, 1990) or partly upper Paleocene, at least locally (e.g., Nichols and others, 1988; Seeland, 1992). In Montana, the contact was placed at a regional unconformity approximately 250 ft (75 m) below a 6- to 8-in-thick (15- to 20-cm-thick), brown-weathering, calcareous coquina (Denson and others, 1990; Vuke and others, 2001b).

The effect of the Paleocene/Eocene Thermal Maximum (PETM) on deposition of the Wasatch Formation has not been studied in Montana. Research of the age-equivalent Willwood Formation in the Bighorn Basin of Wyoming suggests that the PETM resulted in more seasonal or episodic precipitation because of the temperature increase (Baczynski and others, 2017), which facilitated changes in flora and fauna (Gingerich, 2003; Wing and others, 2009). The flora and fauna of the Wasatch Formation in the Powder River Basin were also affected by the PETM (Wing and others, 2003; Wagner, 2013).

Clinker

Red clinker, prominent in the Fort Union and Wasatch Formations (figs. 5, 9), formed as rocks were baked or melted during burning of underlying coal intervals (Heffern and Coates, 1997; Heffern and others, 2013). Clinker deposits cap many ridges and escarpments in eastern Montana, because they are more resistant than the underlying rock, thus playing a role in modern topographic development (Heffern and Coates, 1997; fig. 9C). Clinker also forms lines along bedding within the Fort Union and Wasatch Formations where overburden was not removed by erosion (fig. 9D).

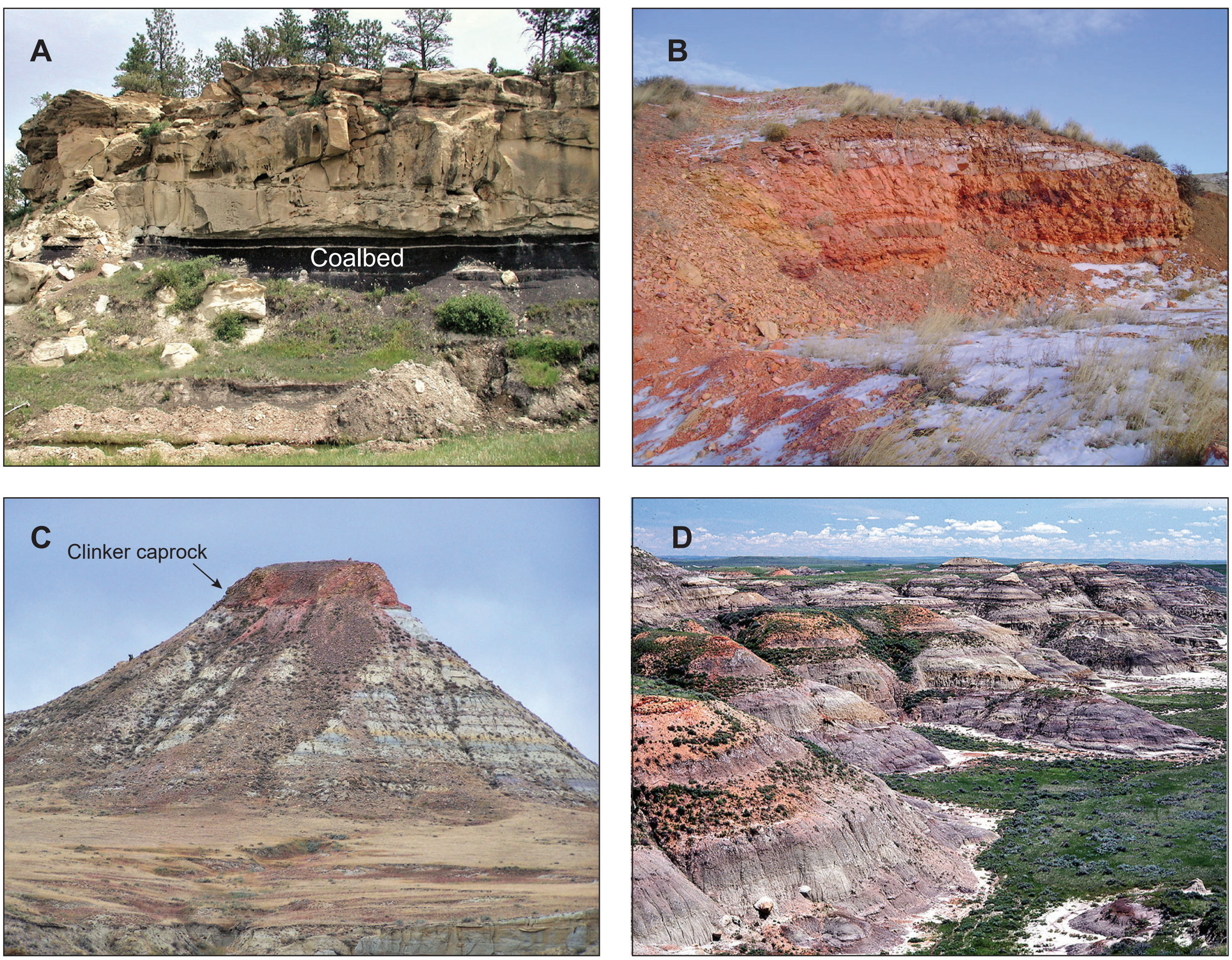


Figure 9. (A) Mammoth coalbed, Bull Mountains Basin near Roundup, Montana, K.B. Waren photo. (B) Clinker bed, C.W. Schwartz photo. (C) Tongue River Member with clinker caprock near Terry, Montana, G.N. Abdo photo. (D) Clinker along Hell Creek–Ludlow contact, near Baker, Montana.

Biostratigraphic, Magnetostratigraphic, and Chronostratigraphic Age Constraints

Biostratigraphic dating of the Hell Creek and Fort Union Formations has relied primarily on the presence or absence of dinosaur fossils, North American Land Mammal ages (NALMA; Krause, 1980; Gingerich and others, 1983; Archibald and others, 1987; Robinson and Honey, 1987; Lofgren and others, 2004; Gingerich, 2016), molluscan faunal changes (Hartman and Roth, 1998), and palynomorph zones (Nichols and Brown, 1992; Pocknall and Nichols, 1996; Hotton, 2002; Nichols, 2003). NALMA designations for the synorogenic deposits include Lancian (upper Hell Creek and Lance Formations); Puercan, Torrejonian, Tiffanian, and Clarkforkian (Fort Union Formation); and Wasatchian (Wasatch Formation; fig. 4). Mammal fossils have been studied extensively in the Crazy Mountains Basin (e.g., Krause and Gingerich, 1983; Hartman and others, 1989; Boyer and Bloch, 2003; Boyer and others, 2004; Bloch and others, 2006), and in eastern Montana and western North Dakota (e.g., Holtzman, 1978; Krause, 1987; Strait and Krause, 1988; Kihm and others, 1993; Hunter and Pearson, 1996; Hunter and others, 1997; Hunter, 1999; Hunter and Archibald, 2002; see Horner and Hanson, 2020).

Magnetostratigraphic and chronostratigraphic studies of the Hell Creek and Fort Union Formations provide age constraints in specific areas (Archibald and others, 1982; Butler and others, 1987; Belt and others, 1997; Lund and others, 2002; Peppe and others, 2009, 2011; Hicks and others, 2002; Swisher and others, 1993; Warwick and others, 2004; Hart, 2012; Buckley, 2018; fig. 4).

Influence of the Cannonball Sea on Synorogenic Sedimentation

The Cretaceous epicontinental seaway that had occupied the Cordilleran Foreland Basin System retreated from Montana during latest Cretaceous time, but did not completely leave the Western Interior region (Hartman and Kirkland, 2002; Hoganson and Murphy, 2002). During Paleocene time, the seaway (known as the "Cannonball Sea" at this time) primarily occupied the craton area of western North and South Dakota (fig. 10), where the marine deposits are represented by the Cannonball Formation (Cvancara, 1976). Mammal fossils, foraminifera, coccoliths, and ostracodes constrain the ages of bounding strata and tongues of the Cannonball Formation, and therefore the incursions of the Cannonball Sea. The lowest tongue of the Cannonball Formation is constrained to upper Puercan, and part of the lower Tongue River Member that intercalates with Cannonball tongues is constrained to upper Torrejonian through lower Tiffanian (Hartman and others, 1998, 1999; fig. 4). In eastern Montana, the lower Tongue River Member contains beds with brackish-water trace fossil associations and marine diatoms that closely alternate with freshwater facies (Belt and others, 2005). Torrejonian deposits of the Ekalaka Member in southeastern Montana (figs. 4, 8) display complex intercalations of marine, estuarine, and nonmarine deposits (Belt and others, 2002) that reflect high-frequency alternations between freshwater and marginal-marine deposits in Montana (Belt and others, 2005).

Paleovalleys and mature paleosols, including silcrete beds (siliceous paleosols) on the interfluves between channels, indicate unconformities related to marine regressions (Belt and others, 2004). Paleovalleys cut into the upper Lebo, Ludlow, or Ekalaka Members (fig. 3) as base level dropped, and were then backfilled with Tongue River sediments during subsequent transgressions (Belt and others, 2002, 2004).

Regionally, the zone that contains the paleosols spans the upper Lebo or Ludlow Members and the lower Tongue River Member in much of eastern Montana (Vuke and Colton, 1998, 2003; Vuke and others, 2001a–f; 2003a,b; 2011). The most significant unconformity may occur at the Lebo–, Ludlow–, or Ekalaka–Tongue River contact (fig. 3) where a thick silcrete bed and incised channels occur locally (Wehrfritz, 1978; Belt and others, 2002, 2004). Plants preserved in the silcrete beds are dominantly non-branching *Equisetum* ("scouring rush"; Christensen, 1984), a marsh plant with diminutive leaves and roots that deposits silica in its outer epidermal walls. The main source of silica in the silcrete beds may have been windblown sand (Christensen, 1984), perhaps derived from marginal marine quartz-rich sands associated with the Cannonball Sea to the east (fig. 10).

Magnetostratigraphic studies indicate a time-correlative unconformity associated with the Lebo–Tongue River contact near Miles City in eastern Montana and with the Ludlow–Tongue River contact in the Little Missouri River area of western North Dakota. The unconformity represents a 0.26 to 0.62 million-year long depositional hiatus (Peppe and others, 2009, 2011).

Figure 10. Paleogeographic map showing a waning stage of the Cannonball Sea. Sea-level fluctuations brought marginal marine environments into eastern Montana intermittently during Paleocene time at least to the extent of the blue line. Figure from Blakey (2016).

In the Little Missouri River area of westernmost North Dakota, a single silcrete bed (the Rhame bed) occurs locally at the contact between the Ludlow Member equivalent and the Tongue River Member equivalent (Wehrfritz, 1978). Mature paleosols including well-developed silcrete beds also occur at the contact between the Ekalaka Member and the overlying Tongue River Member (Belt and others, 2002; fig. 3). A well-developed silcrete bed is also present in the Cave Hills of South Dakota at the contact between the Ekalaka Member equivalent and the Tongue River Member (Belt and others, 2002). Ages of mammal teeth (Krause, 1987; Strait and Krause, 1988) above and below the unconformity at the top of the Ekalaka Member in southeastern Montana document a significant time gap related to a regressive episode of the Cannonball Sea (Belt and others, 2002), which is correlative with unconformities located elsewhere in eastern Montana and western North Dakota at the top of the Ludlow or Lebo Members. The Cannonball Sea—the final epicontinental sea to inundate the Western Interior of North America—persisted into late Paleocene time east of Montana (Lund and others, 2002), but its influence in eastern Montana diminished during later Tongue River deposition (fig. 10).

MAIN LARAMIDE PROVINCE

The main Laramide Province of the North American Cordilleran Foreland Basin System (fig. 1) extends northward into Montana from the Central Rockies of Wyoming, although the placement of its northern boundary varies (e.g., Hamilton, 1988; DeCelles, 2004; Yonkee and Weil, 2014). The northern placement of Hamilton (1988) incorporates the main Laramide uplifts and associated basins discussed in this section (figs. 1, 2).

Regional Late Cretaceous Eastward-Propagating Uplift

Magnetostratigraphic, biostratigraphic, and isotopic age constraints have identified a pre-Lancian (fig. 4) unconformity at the base of the Hell Creek and Lance Formations in southern Montana and northern Wyoming, and have determined regional trends in its magnitude and duration (Belt and others, 1997). The period of erosion represented by the unconformity is greatest in the west and becomes younger to the east by 3.5 million years (fig. 3). Belt and others (1997) concluded that the unconformity was the result of a wave-like migration of tectonic uplift, rather than related to the regressing Western Interior Sea to the east. The eastward-propagating phase of uplift (and associated erosion) moved 600 km (373 mi) in 3.5 Ma, equal to a rate of 17.14 cm (6.8 in)/yr—greatly exceeding the rate of flexural response to Sevier thrust loading, and beyond the range of forebulge migration (Belt and others, 1997). This eastward propagation of uplift and erosion is broadly consistent with recent geodynamic modeling of the topographic development in the Cordilleran System (Copeland and others, 2017), and preceded compartmentalization into uplift arches and associated basins (e.g., Carrapa and others, 2019).

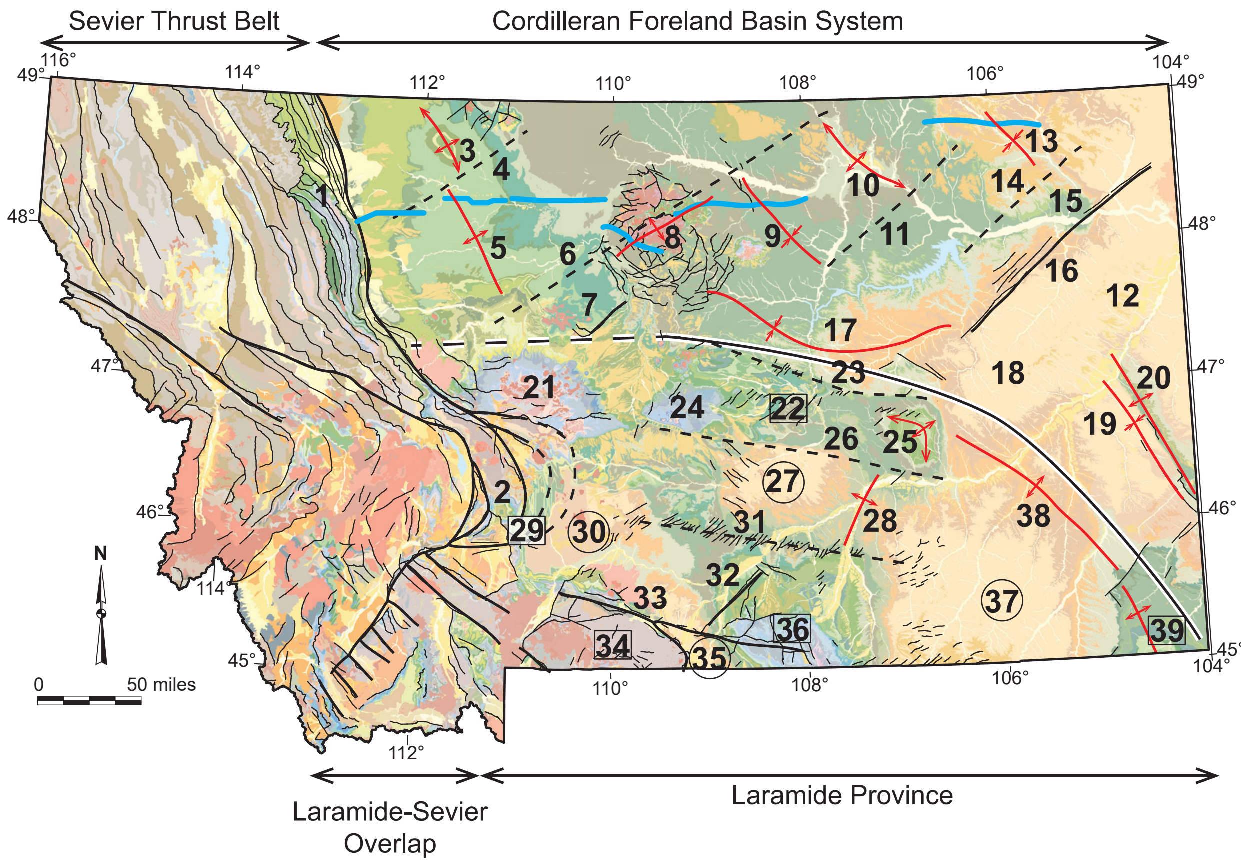
The unconformity at the base of the Hell Creek Formation may occur regionally in eastern Montana and western North Dakota (Jensen and Varnes, 1964), with scour features and rip-up clast breccia or lithic clast lags at the base (Jensen and Varnes, 1964; Vuke and Colton, 1998; Flight, 2004). However, Flight (2004) related the unconformity at the base of the Hell Creek Formation in northeastern Montana to sea level fluctuation, rather than tectonically driven surface uplift. Elsewhere in northeastern Montana an unconformity is not recognized at the base of the Hell Creek Formation (Hartman and others, 2014).

Laramide Uplifts and Associated Structural-Sedimentary Basins

Bridger Uplift/Crazy Mountains Basin (figs. 2, 11, 12)

The reactivated Mesoproterozoic Pass fault of the Southwest Montana Transverse Zone transects the central Bridger Range and separates the Helena Salient of the Sevier fold-thrust belt north of the fault zone, from the main Laramide Province to the south (fig. 1). Basement-involved (Laramide-style) uplift associated with the sub-Bridger fault zone deformed and folded the thin-skinned (Sevier-style) thrust sheets at the south margin of the Helena Salient in the present-day northern Bridger Range, and produced a composite, northward-plunging, ancestral Bridger Range arch during Paleogene time (Lageson, 1989; Skipp and others, 1999). Subsequent Neogene extension down-dropped the crest and western limb of the ancestral arch beneath the Gallatin Valley (Craiglow, 1986). The asymmetric Crazy Mountains Basin developed from thrust loading along its western margin, which produced a deep basin trough adjacent to the ancestral Bridger Arch. Based on structure contour maps, the trough probably formed largely during Maatrichtian time (Johnson and Finn, 2004; fig. 4). Seismic data interpretations indicate that Upper Cretaceous rocks are thicker in the basin trough than near the basin margins along the extent of the basin, whereas lower Mesozoic rocks appear to have uniform thickness, suggesting that Laramide-related subsidence began during Late Cretaceous time (Taylor, 2004). Uplift to the west caused rapid erosion of the Elkhorn Mountains volcanic field, and basin subsidence kept pace with and accommodated the influx of immature sediment dominantly from that source (Roberts, 1972; Taylor, 2004). A change from east-directed to south-directed paleocurrents in Upper Cretaceous deposits in the northeastern part of the basin also suggests that the basin began to develop during Late Cretaceous time (Borrell, 2000). As uplift continued, a series of coalescing alluvial fans of the dominantly Paleocene Fort Union Formation developed along the western basin margin (Piombino, 1979). Members of the Fort Union Formation were mappable in the eastern part of the basin, but were not distinguishable in the western part (Berg and others, 2000; McDonald and others, 2005).

The Crazy Mountains Basin is the deepest Laramide basin in Montana and is one of the deepest in the Western Interior (Dickinson and others, 1988), having accommodated 10,300 ft (3,140 m) of Upper Cretaceous and more than 6,600 ft (5,000 m) of Paleocene sedimentary deposits along its axis (Roberts, 1972). Although the Helena Salient, which borders much of the western Crazy Mountains Basin, is referred to as a salient of the fold-thrust belt (e.g., Harlan and others, 1988) it also represents a deep crustal segment that translated eastward along high-angle reactivated basement faults (Reynolds, 2004) and functioned as a Laramide-style uplift (Lageson, 1989; Skipp and others, 1999). In that sense it is part of the Laramide–Sevier overlap zone (Kulik and Schmidt, 1988; fig. 1).



Cordilleran Thrust Belt
1. "Disturbed belt"
2. Helena Salient

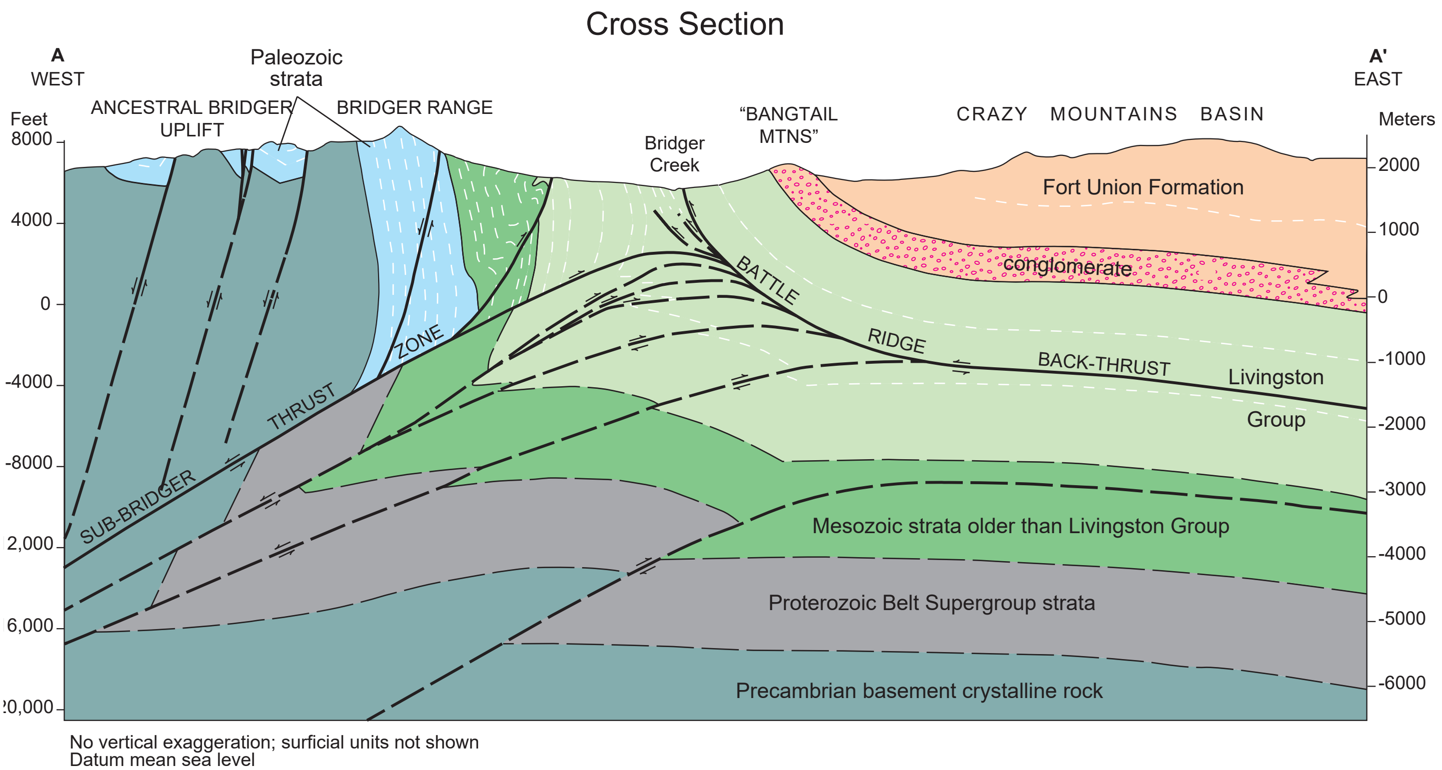
Cordilleran Foreland Basin System
North of Main Laramide Province
3. Kevin-Sunburst Dome of Sweetgrass Arch
4. Pendroy Fault and Scapegoat-Bannatyne Trend
5. South Arch of Sweetgrass Arch
6. Great Falls Tectonic Zone
7. Arrow Creek Fault
8. Bearpaw Uplift
9. Coburn Syncline
10. Bowdoin Dome
11. Hinsdale Fault
12. Western Williston Basin
   13. Opheim Syncline or Basin
   14. Poplar Fault
   15. Poplar Dome
   16. Weldon-Brockton/Brockton-Froid Fault Zone
   17. Blood Creek Syncline or Basin (included by some in Williston Basin)
   18. Circle Basin
   19. Sheep Mountain Syncline
   20. Cedar Creek Anticline/subsurface fault

Cordilleran Foreland Basin System
Main Laramide Province (after Hamilton, 1988)
21. Little Belt Uplift
22. Central Montana Uplift
   23. Cat Creek Anticline/subsurface fault
   24. Big Snowy Uplift
   25. Porcupine Dome
   26. Willow Creek Fault
27. Bull Mountains Basin
28. unnamed arch
29. Bridger Uplift
30. Crazy Mountains Basin
31. Lake Basin Fault Zone
32. Fromberg Fault
33. Nye-Bowler Fault Zone
34. Beartooth Uplift
35. Bighorn Basin
36. Bighorn Uplift
37. Powder River Basin
38. Miles City Arch
39. Black Hills Uplift

Fold or arch
Surface fault (bold = generalized)
Subsurface fault
COCORP seismic line
Northern limit, Main Laramide Province
31 Laramide basin
36 Laramide uplift

Laramide Uplift - Basin Pairs
22. Central Montana Uplift - 27. Bull Mountains Basin
29. Bridger Uplift - 30. Crazy Mountains Basin
34. Beartooth Uplift - 35. Bighorn Basin
36. Bighorn Uplift - 37. Powder River Basin

Figure 11. Cordilleran Foreland Basin system uplifts, basins, faults, and other structures in central and eastern Montana. COCORP seismic lines from Latham and others (1988).



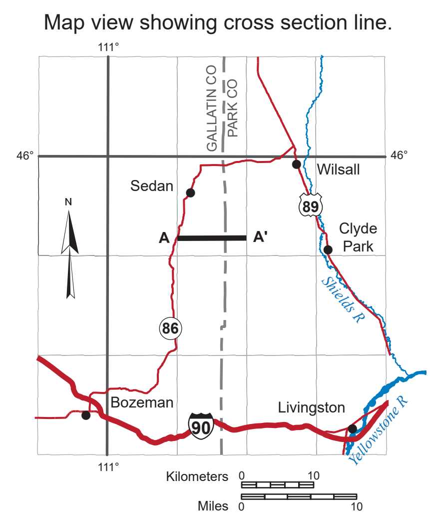


Figure 12. Cross section from Bridger Uplift into Crazy Mountains Basin. Modified from Skipp and others (1999).

Battle Ridge and the "Bangtail Mountains" in the western Crazy Mountains Basin are surface expressions of a west-directed back-thrust zone that was genetically related to the sub-Bridger thrust zone (Skipp and others, 1999; fig. 12). The back-thrusts steeply folded and locally overturned strata of the upper Livingston Group and those of the Fort Union Formation as young as late Paleocene near their leading edge, producing folds within the basin (Skipp and others, 1999). The most significant Laramide deformation occurred during late Paleocene and early Eocene time (Craiglow, 1986; Harlan and others, 1988).

Beartooth Uplift/Bighorn Basin (figs. 2, 13)

A WNW–ESE-striking, basement-involved reverse fault bounds part of the north side of the Beartooth Uplift, but the fault trace deflects southward at the Red Lodge corner (fig. 13). New thermochronological data suggest that exhumation of the Beartooth Uplift may have begun as early as Early Cretaceous (Carrapa and others, 2019). Initially, fine-grained sediment derived from unroofing of 3.6 km of poorly resistant Mesozoic rock bypassed the uplifted footwall of the fault and was deposited in the distal basin (DeCelles and others, 1991b). When Paleozoic, and later, Precambrian rocks in the eroding hanging wall were breached, Paleocene synorogenic alluvial fan conglomerate (DeCelles and

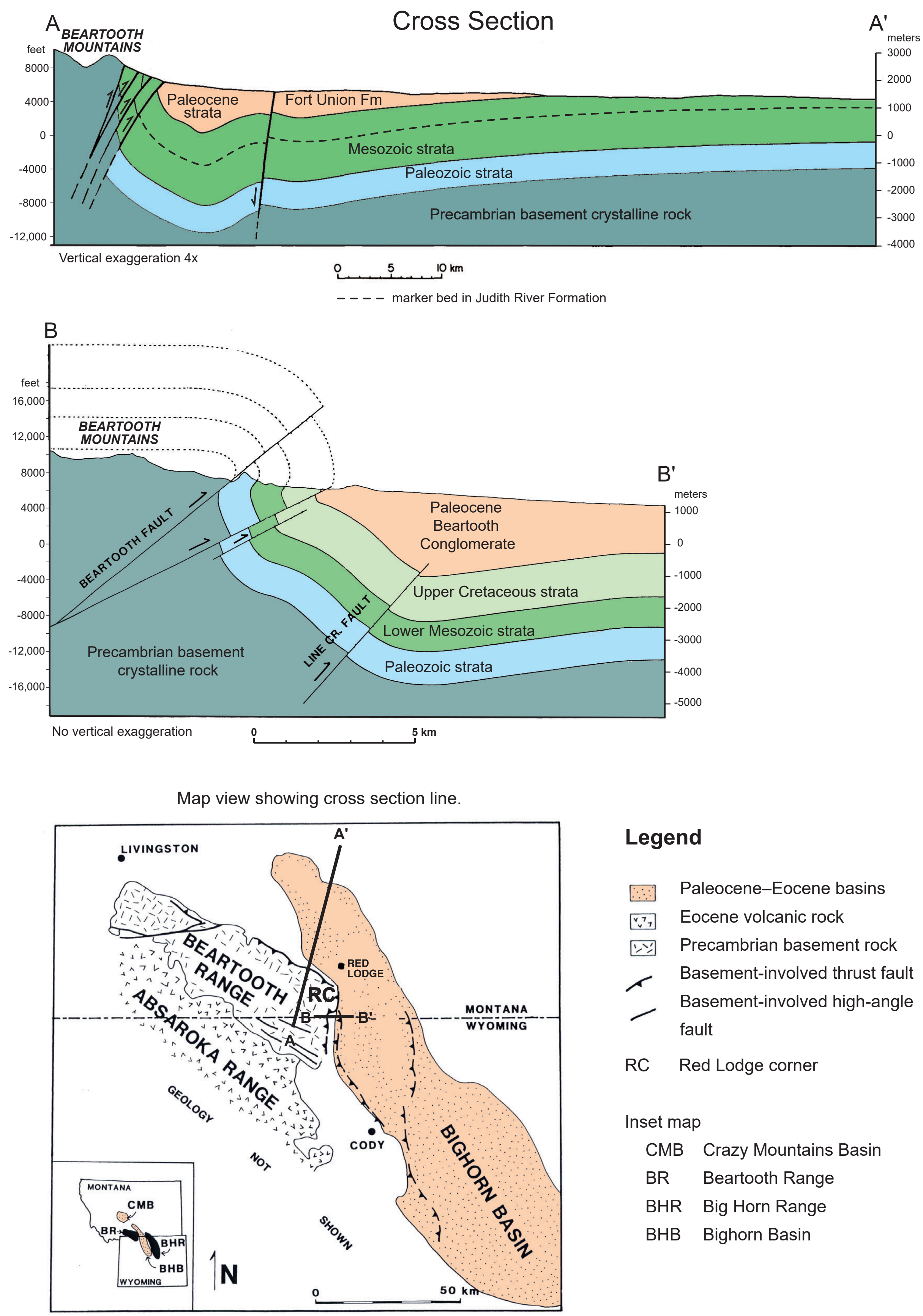


Figure 13. Cross sections across Beartooth Uplift into the Bighorn Basin. A–A' modified from Hagen and others (1985); B–B' and index map modified from DeCelles and others (1991b).

others, 1991a) was shed into an asymmetric, east-vergent growth syncline along the western margin of the Bighorn Basin (Hoy and Ridgway, 1997). As much as 7,000 m (23,000 ft) of sediment was deposited in the syncline (Foose and others, 1961; Blackstone, 1986).

The synorogenic conglomerate adjacent to the northern part of the WNW–ESE-striking fault was designated the Linley Conglomerate (Calvert, 1916), and was included as a member of the Paleocene Fort Union Formation (Lopez, 2000a). South of Red Lodge, a similar unit has been called the Beartooth Conglomerate (DeCelles and others, 1991a).

The conglomerate proximal to the fault contains a depositionally inverted section of approximately 3 km (1.86 mi) of Phanerozoic and Precambrian crystalline rock, reflecting the progressive erosion and unroofing of the Beartooth Arch (Dutcher and others, 1986). Progressive angular unconformities, with dips as high as 50° along the west side of the basin and more subdued farther out into the basin, further document the relationship between progressive uplift and basinward rotation of proximal synorogenic conglomerates (Dutcher and others, 1986; Koenig, 2015).

According to one model, fault-propagation folding of the Beartooth Conglomerate alluvial fan/alluvial plain deposits produced considerable uplift before breakout of the Beartooth thrust fault along the western side of the Bighorn Basin, south of Red Lodge (DeCelles and others, 1991a). Another interpretation suggests that this period of uplift involved only minor deformation (Stewart and others, 2008).

The uplift-proximal Beartooth Conglomerate interfingers with the more distal, finer-grained Tongue River Member of the Fort Union Formation in the Bighorn Basin (Nielsen, 2009; fig. 4). Late Paleocene floral dates from within the Beartooth Conglomerate along the eastern front of the Beartooth Range in Montana (Hickey, 1980) correlate with late Paleocene floral and faunal fossil dates in the Fort Union Formation in the Bighorn Basin to the east (Flueckinger, 1970). However, the floral age of the uppermost conglomerate may extend into Eocene time (Hickey, 1980). Upper conglomerate beds along the northern front of the Beartooth Range may also be earliest Eocene (Raines and Johnson, 1995; Hart, 2012). A NE-flowing trunk drainage system is interpreted to have occupied the Bighorn Basin during earliest Eocene time (Seeland, 1992).

The oldest deposits of the Beartooth Conglomerate were carried in the hanging wall of the Beartooth fault (Nielsen, 2009), which overrode younger Beartooth Conglomerate deposits (DeCelles and others, 1991a). At the Red Lodge corner of the basin where the trace of the Beartooth thrust fault changes orientation (fig. 13), the hanging wall locally completely overrode the Beartooth Conglomerate and currently rests on more distal facies of the Fort Union Formation (Bartholomew and others, 2008).

Fission track dating indicates that the Beartooth Uplift was actively rising from early Paleocene to early Eocene time (Omar and others, 1994; fig. 4). Preserved paleoseismites are abundant in the late Paleocene Tongue River Member of the Fort Union Formation (clastic dikes, dewatering structures, contorted laminae), and in the Eocene Willwood Formation in Wyoming (liquefaction structures) in the Bighorn Basin (Bartholomew and others, 2008; Jackson and others, 2019). Their presence suggests that the main phase of deformation of the Beartooth Uplift extended from late Paleocene into early Eocene time (Bartholomew and others, 2008; Stewart and others, 2008). This agrees with paleoelevation, thermochronology, and provenance data that indicate the main phase of uplift was late Paleocene into early Eocene time (Fan and Carrapa, 2014), although initiation of uplift may have begun during Late Cretaceous time in Wyoming (Fan and Carrapa, 2014; Jackson and others, 2019) or as early as Early Cretaceous time in Montana (Carrapa and others, 2019).

Big Horn Uplift/Powder River Basin (figs. 2, 14, 15)

Four structural domains make up the northern part of the greater Big Horn Uplift or Arch in Montana (fig. 14): (1) the Big Horn Uplift proper is a back-thrusted domain with 3,000 m (9,800 ft) of structural relief that exposes the basement core of the arch. This domain includes a graben on its western side. (2) The Pryor Mountain domain lies to the immediate west of the graben, forms the eastern boundary of the Bighorn Basin, and has 3,400 m (11,150 ft) of structural relief. (3) The Billings Arch domain is an arched platform composed of Mesozoic strata located north of the Pryor Uplift, and is bisected by the NE–SW-striking Fromberg fault zone. The domain is bounded on the north by the Lake Basin fault zone. (4) The Hardin Platform is located east of the Billings Arch, is dominantly composed of Cretaceous rock at the surface,

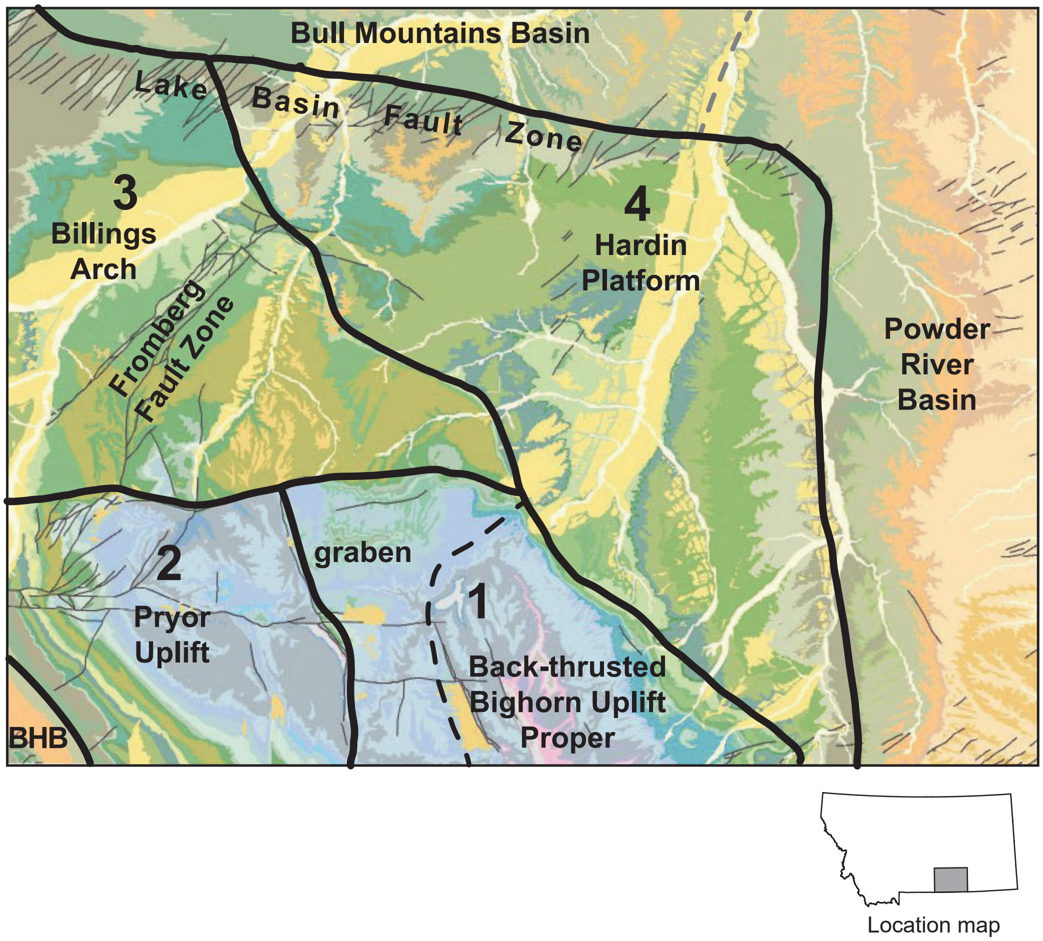


Figure 14. Greater Bighorn Uplift domains and adjacent basins. BHB, Bighorn Basin. Modified from Naus (2000).

and is also bounded on the north by the Lake Basin fault zone, which extends into the Powder River Basin (Robinson and Barnum, 1986). The Hardin Platform domain bounds the Powder River Basin to the east and north (Naus, 2000). Geophysical data suggest that the subsurface structure of the northwestern Powder River Basin extends along the western side of the Hardin Platform (Robbins, 1994). Seismic profiles across the east flank of the Big Horn Mountains indicate that two major west-dipping thrust faults are present close to the surface in northern Wyoming (Robbins and Grow, 1992; Robbins, 1994; fig. 15).

The Powder River Basin is asymmetrical with strata dipping as many as 20–25 degrees along its western margin and only 2–5 degrees along its eastern margin (Flores and Bader, 1999). Gravity "lows" are lowest on the west side of the basin, reflecting the basin's asymmetry (Robbins, 1994). The Fort Union Formation is over 5,200 ft (1,585 m) thick along the basin axis in the western part of the basin in Wyoming (Curry, 1971). Abundant mudstone in the central part of the basin led to interpretations of an internally drained lacustrine environment for the Lebo Member of the Fort Union Formation (Ayers, 1986). However, glauconite in the Lebo Member (Whipkey and others, 1991) suggests that unroofing of Cretaceous marine shale from the Bighorn Uplift contributed the fine-grained sediment to the central part of the basin, having bypassed coarser sediment along the basin margin (Flores and Bader, 1999).

Isopach data show little change in thickness of the Upper Cretaceous Fox Hills and Hell Creek Formations across the northern Powder River Basin (Connor, 1992). Abundant first-cycle carbonate clasts in the lowermost part of the Paleocene Tullock Member of the Fort Union Formation in the northwestern part of the Powder River Basin suggest doming and erosion of Paleozoic and Mesozoic rocks in the Big Horn Uplift area during earliest Paleocene time (Hansley and Brown, 1992; Brown, 1993). However, Tullock Member streams flowed east–northeast across a gently sloping alluvial plain toward the Cannonball Sea (Hansley and Brown, 1992), suggesting that basin formation was not developed enough at the surface to deflect the flow direction. Paleocurrent and isopach data indicate that more significant uplift began during

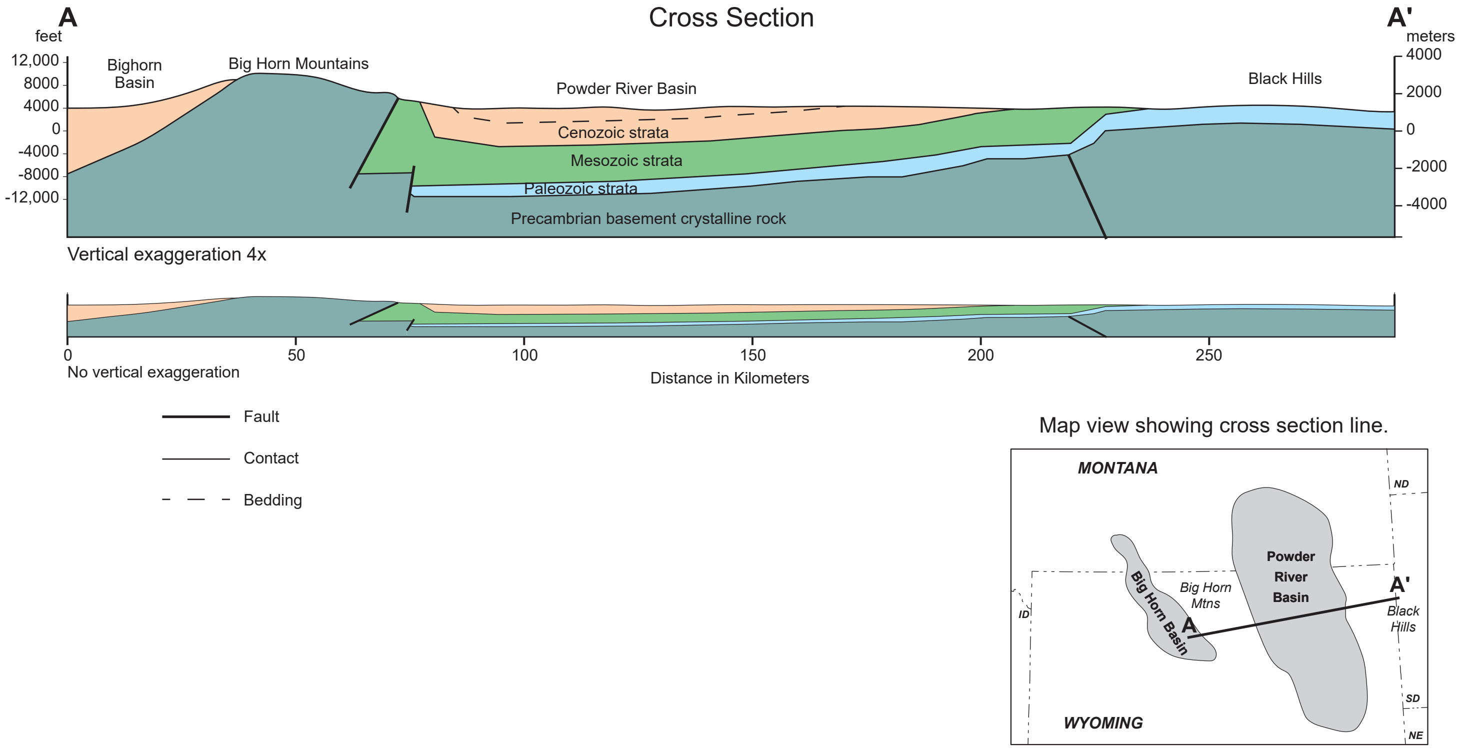


Figure 15. Cross section from eastern Bighorn Basin across Big Horn Mountains and Powder River Basin to Black Hills. Modified from Robbins (1994).

early Puercan (fig. 4) deposition of the Lebo Member of the Fort Union Formation (Curry, 1971; Whipkey and others, 1991; Belt and others, 1992). Based on sedimentary facies studies, the Bighorn Uplift became a major source of sediment for the Powder River Basin beginning in late Paleocene time (Flores and Ethridge, 1985). Additionally, according to paleoelevation and thermochronology data, the highest rate and magnitude of uplift associated with the Big Horn Uplift and thrust-loading of the Powder River Basin occurred during late Paleocene–early Eocene time (Fan and Carrapa, 2014).

The Wasatch Formation (figs. 3, 4) conformably overlies the Fort Union Formation in the center of the Powder River Basin in Wyoming, but unconformably along its margins (Flores and Bader, 1999). The interplay between tectonic subsidence and base-level changes promoted development of raised bogs that in turn produced thick, economical, subbituminous and lignite coalbeds in the Tongue River Member of the Fort Union Formation and the Wasatch Formation (Flores and Bader, 1999; Gunderson and Wheaton, 2020). However, along the western margin of the basin in Wyoming, synorogenic conglomerate is present in the late Paleocene upper Fort Union and early Eocene Wasatch Formations (Love and Christiansen, 2014), recording rapid erosion of the Bighorn Uplift (Hoy and Ridgway, 1997). Conglomerate is lacking along the western edge of the Powder River Basin in Montana, where source areas adjacent to the basin primarily consisted of poorly resistant Cretaceous rock, rather than resistant Paleozoic and Precambrian rock as to the south.

During late Paleocene and earliest Eocene time, a trunk fluvial drainage system flowed along the subsiding part of the basin adjacent to the Bighorn Uplift and then diverted to the east along the northern edge of the Black Hills Uplift, flowing toward the Cannonball Sea (Seeland, 1988, 1992; fig. 10).

Central Montana Uplift[1]/Bull Mountains Basin (figs. 2, 16)

The Bull Mountains Basin in central Montana is an asymmetrical syncline (Stricker, 1999) with superimposed anticlines (Luebking and others, 2001). It is bounded on the north by the southern part of the WNW–ESE-striking Central Montana Uplift (fig. 16), and on the south by the parallel Lake Basin fault zone (fig. 2), both of which are associated with blind Laramide reverse and strike-slip faults that were reactivated from Proterozoic basement structures (Nelson,

[2]In this paper *Central Montana Uplift* refers to its original application (e.g., Lageson and others, 2020), not to other applications (e.g., Bader, 2019a).

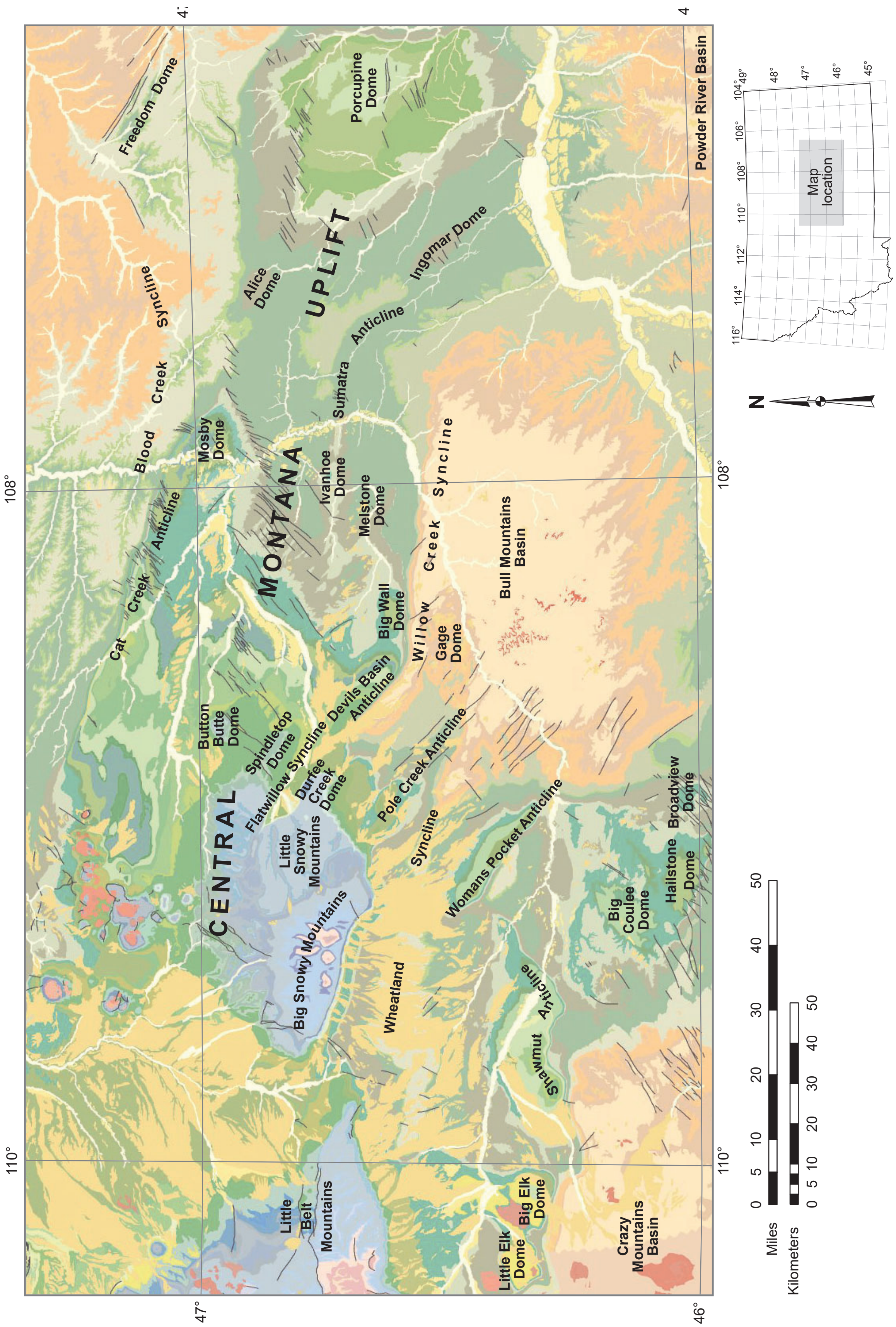


Figure 16. Cordilleran Foreland Basin System structures in central Montana. Modified from Vuke and others (2007).

1992, 1993). The youngest deposits in the basin are those of the Paleocene Fort Union Formation, and the Tongue River Member is the dominant unit exposed (Wilde and Porter, 2000).

The Central Montana Uplift is a foreland structure composed of several structurally positive tectonic elements including the Big Snowy Mountains at its west end, the Porcupine Dome at its east end (e.g., Sonnenberg, 1956; Shurr and Rice, 1986), and numerous smaller domes and folds between (Nelson, 1993; Porter and others, 1996; Porter and Wilde, 1999; Vuke and Wilde, 2004; figs. 2, 11, 16). The Cat Creek Anticline (figs. 16, 17)—a fault-propagation fold (Nelson, 1992, 1993)—and related features bound part of the north end of the Central Montana Uplift, which is underlain by a Laramide-reactivated basement-involved reverse fault that dips to the southwest and was buttressed by stable craton to the north (Nelson, 1992, 1993). The Cat Creek reverse fault displays approximately 1,200 m of vertical separation in addition to an

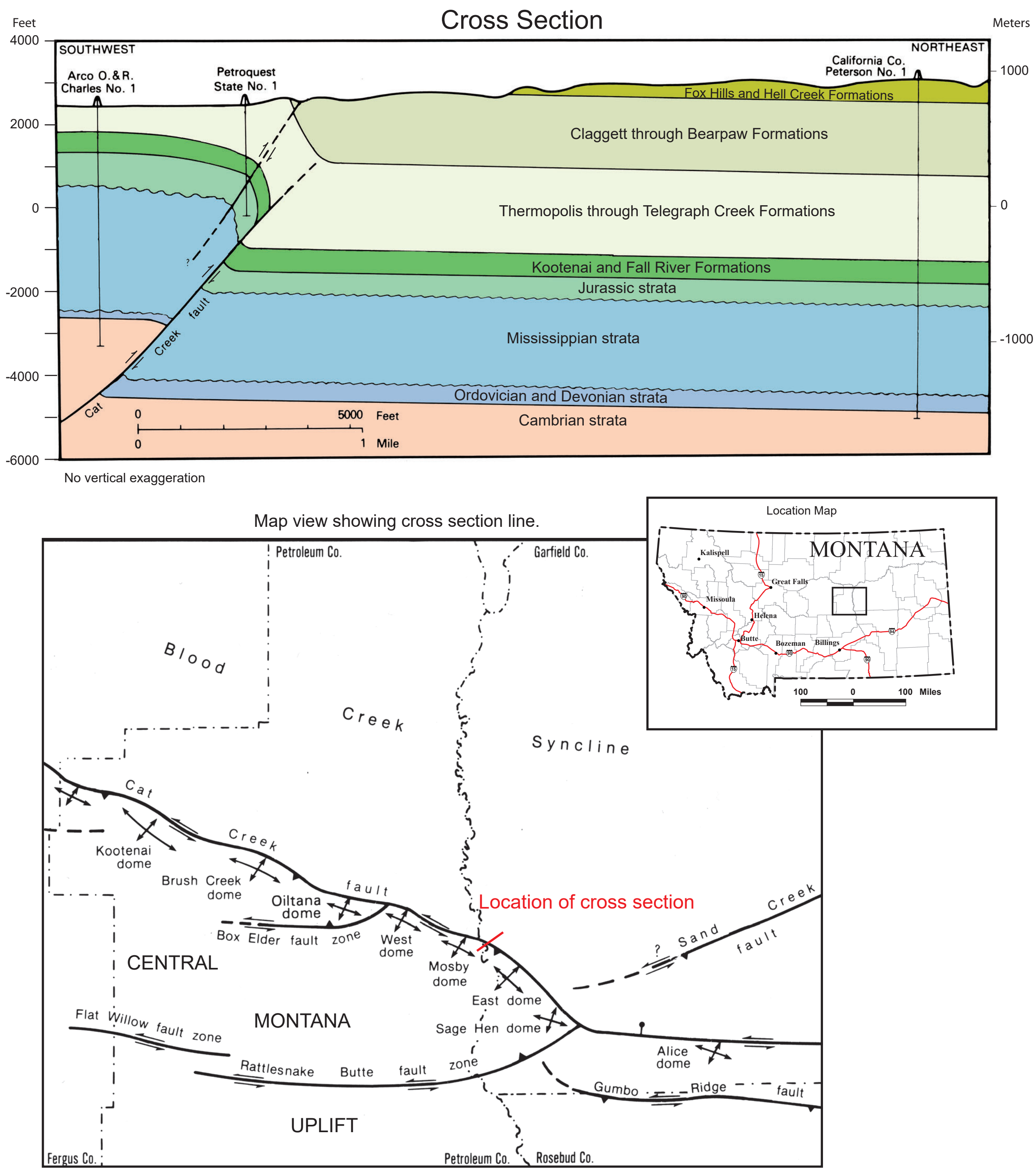


Figure 17. Cross section across Cat Creek Anticline/Fault, northern Central Montana Uplift. Modified from Nelson (1993).

unknown amount of lateral (oblique-slip) movement (Nelson, 1992, 1993). The Willow Creek fault, a parallel subsurface basement fault, bounds the southern margin of the Central Montana Uplift and the northern end of the Bull Mountains Basin (Stone, 1969; fig. 2). This reverse fault extends along the southern Big Snowy Uplift (Baker and Johnson, 2000; figs. 2, 16).

The Willow Creek Syncline (fig. 16), just south of the Central Montana Uplift, represents the trough of the asymmetric Bull Mountains Basin (Woodward and others, 1997). Most of the basin gently dips toward the trough from the Lake Basin fault zone along the southern margin of the basin (Woodward and others, 1997), and the northern margin of the greater Big Horn Uplift (Naus, 2000). Paleoseismites in the Tongue River Member, localized in the Bull Mountains Basin, suggest the occurrence of large-magnitude, late Paleocene earthquakes generated by movement on a nearby fault (Bartholomew and others, 2008).

Clast lithologies and sediment grain-size distributions in the basin indicate that the Central Montana Uplift did not contribute significant amounts of sediment to the Bull Mountains Basin (Shurr, 1972). Although one study demonstrated consistent paleoflow to the south for the Tongue River Member, away from the Central Montana Uplift (Shurr, 1972), another study indicated eastward paleocurrent directions for the Tongue River Member, reflecting a western source and paleodrainage toward the Cannonball Sea (Seeland and others, 1988; fig. 10). The Central Montana Uplift represents the last reversal of a trough in roughly the same position that was episodically present from Proterozoic through Early Cretaceous time (Proterozoic Belt Basin, Paleozoic Big Snowy Trough, and Mesozoic Central Montana Trough; Peterson, 1981). The most significant uplift in this area began during Paleocene time, and reached maximum uplift after Paleocene time (Shurr, 1972; Nelson, 1993). Only a small, unnamed arch (figs. 2, 14), recognized from its gravity signature and to a lesser extent its magnetic signature, separates the Bull Mountains Basin from the Powder River Basin (Robbins, 1994).

Black Hills Uplift and Miles City Arch

The domal, basement-cored Black Hills Uplift is the easternmost of the Laramide basement-cored uplifts that extend into Montana (figs. 2, 15), and had the latest time of initial development (Paleocene time; Lisenbee and DeWitt, 1993). Fault data indicate that subhorizontal shortening was kinematically identical to that of most other Laramide uplifts, although unlike the other Laramide uplifts, basement fabric reactivation was likely insignificant (Singleton and others, 2019). Although the uplift provided sediment to the east side of the Powder River Basin (Merin and Lindholm, 1986; Lisenbee and DeWitt, 1993; fig. 15), it contributed only minimally to basin subsidence relative to the loading influence of the basin-bounding reverse faults associated with the Big Horn Uplift on its western side.

The Miles City Arch in southeastern Montana (fig. 2) is typically considered the northwestern extension of the basement-cored Black Hills Uplift that is centered in northwestern South Dakota (Lisenbee and DeWitt, 1993; Robbins, 1994), although the Miles City Arch has also been interpreted as a discrete structural entity (Thomas, 1974). Paleocurrent patterns of the Fox Hills and Hell Creek Formations, as well as the lower Ludlow and Tullock Members of the Fort Union Formation, suggest that the Black Hills area did not obstruct eastward drainage toward the Cannonball Sea (fig. 10) during Late Cretaceous and early Paleocene time (Seeland, 1988; Lisenbee and DeWitt, 1993; Belt and others, 1997). Isopach maps also show no stratigraphic thinning across the Miles City Arch during deposition of the Hell Creek and Lance Formations (Connor, 1992).

Initiation of the Black Hills Uplift–Miles City Arch occurred during earliest Torrejonian time based on sedimentologic evidence (Belt, 2004; fig. 4). During late Torrejonian and early Tiffanian time, the Black Hills area was further uplifted and eroded, tilting strata in lower parts of the Tongue River Member to the northwest and producing a regional angular unconformity in southeastern Montana (Vuke and others, 2001a; Belt and others, 2004). The uplift may have caused a shift in depocenters from east of the present Black Hills to the northern Williston Basin during the transition from Puercan to middle Torrejonian time (fig. 4), as reflected by a significant shift in paleodrainage orientation (Diemer and Belt, 1991; Belt and others, 1997) throughout eastern Montana and western North Dakota (Belt, 1993). A locally persistent Puercan unconformity is present at the base of the Ekalaka Member in southeastern Montana, also reflecting this episode of uplift (Belt and others, 2002). In addition, earthquakes associated with uplift may have triggered middle Paleocene mass movement, reflected by large areas of rotated bedding and megabreccias, present along the axis of the Black Hills Uplift (Belt and others, 2002) and Miles City Arch (Garrett, 1963).

Paleozoic clasts in the Tongue River Member in the central part of the Powder River Basin could have originated from the Bighorn Uplift on the west side of the basin or the Black Hills Uplift on the east side. However, the Tongue River Member contains detrital marble and phyllitic rock fragments just south of the Montana border in the central part of the Powder Basin, both of which more likely came from the Black Hills Uplift. This suggests that erosion breached the crystalline core of the uplift during Tongue River deposition and that the sediment source for the central and eastern part of the basin was at least partly from the Black Hills Uplift (Merin and Lindholm, 1986). An unconformity at the base of the Wasatch Formation on the east side of the Powder River Basin suggests a second Black Hills Laramide tectonic pulse during latest Paleocene–early Eocene time (Lisenbee and DeWitt, 1993). During or after uplift of the Black Hills area, a system of NE–SW-striking faults of the Belle Fourche Arch (fig. 2) was active, offsetting the Black Hills monocline, which reflects a blind fault system on the west side of the Black Hills Uplift.

CORDILLERAN FORELAND BASIN SYSTEM NORTH OF THE MAIN LARAMIDE PROVINCE

Outside of the primary extent of the main Laramide Province (Laramide Belt of Hamilton, 1988; “domains 1 and 2” of Lageson and others, 2020; figs. 1, 11), many peripheral basement-cored uplifts and basins are also referred to as Laramide-style structures (e.g., Rice and Shurr, 1978; Thamke and Craigg, 1997). These features are less pronounced than those within the main Laramide Province, and the relationships between uplifts and basins are poorly known. This area of less-pronounced Laramide-style features is “domain 3” of Lageson and others (2020).

Western Williston Basin and Related Structures

The main part of the intracratonic Williston Basin is centered in North Dakota (fig. 2), but its western margin is defined by Laramide-style uplifts in Montana: the Black Hills Uplift, Miles City Arch, Porcupine Dome, and Bowdoin Dome (Shurr and others, 1989a; Anna and others, 2010; fig. 11). The Williston Basin—the largest intracratonic sedimentary basin in North America (Hamke and others, 1966)—may have been dominantly flexural without an associated fault-controlled uplift along its margins, distinguishing it from basins of the Laramide Province (Sloss, 1987). However, a Laramide-style evolutionary phase is interpreted for the Williston Basin based on basement-involved reverse and strike-slip faults that caused most of the mild deformation within the basin (Anna, 1986; Lisenbee and DeWitt, 1993; Herrera, 2013).

The Williston Basin underwent steady, continuous subsidence throughout most of Phanerozoic time (Fowler and Nisbet, 1985), although the sediment accumulation rate increased at the onset of Laramide foreland basin sedimentation (Cherven and Jacob, 1985). Subsequent slow, constant subsidence is indicated by the continuity of Late Cretaceous and early Paleogene rocks in the basin, and the basin received more than 3,000 ft (915 m) of sediment during this time (Anna, 1986). The change from Tongue River Member to overlying Sentinel Butte Member within the Fort Union Formation reflects a shift in source areas that resulted from uplift to the west and northwest (Royce, 1970). The Tongue River and Sentinel Butte Members also thicken toward the center of the basin, indicating significant Paleocene basin subsidence during their deposition (Royce, 1970).

The Cannonball Sea occupied the central part of the Williston Basin during much of early Paleocene time (fig. 10), with only limited incursion into Montana (Belt and others, 1997; Warwick and others, 2004; Belt and others, 2005). The sea dwindled as basin subsidence no longer kept pace with sedimentation rates during deposition of the late Tiffanian Sentinel Butte Member (Royce, 1970; Kihm and others, 1993). The continuity and thickness of lignite beds in the Tongue River and Sentinel Butte Members in the Williston Basin indicate a tectonic setting that was stable enough to allow a consistent water table level, but with dynamic subsidence rates that allowed organic matter to accumulate, conducive to lignite development (Daly and others, 1985).

The asymmetric, NNW–SSE-striking, fault-propagated Cedar Creek Anticline (figs. 11, 18) is a major Laramide structure within the southwestern part of the Williston Basin (Shurr and others, 1989a), hosting significanthydrocarbonyields(Clement1986,1987; Hofmann, 2020). The basement-cored structure may coincide with the fault boundary between the Wyoming and Trans-Hudson Precambrian basement provinces at depth (Sims and others, 1991; Bader, 2019b), although the eastern margin of the Wyoming Province has also been interpreted much farther west (Worth-

Figure 18. Ludlow Member, west limb of Cedar Creek Anticline near Glendive, Montana, looking north.

ington and others, 2015). Subsurface data indicate recurrent movement along the Cedar Creek basement fault, and tectonic stability from Middle Jurassic to post-Paleocene time (Clement, 1986, 1987). However, surface mapping places the basal upper Cretaceous Hell Creek Formation on progressively lower parts of the underlying Fox Hills Formation toward the axis of the anticline, suggesting some Late Cretaceous movement followed by erosional beveling prior to deposition of the Hell Creek Formation (Vuke and Colton, 1998).

Paleocene Ludlow Member fluvial systems flowed across the eroded Cedar Creek Anticline, but during later Paleocene Tongue River deposition, the anticline was again prominent enough to obstruct drainage (Belt and others, 1984). The greatest uplift of the structure was post-Paleocene movement along the high-angle, SE-dipping, reverse fault system that displaced Paleozoic, Jurassic, and Cretaceous rocks up to the east (Clement, 1986, 1987).

The Sheep Mountain Syncline (fig. 11) is adjacent to the steeply dipping west limb of the Cedar Creek Anticline. The Tongue River Member is thicker in the Sheep Mountain Syncline area than on the east flank of the anticline (Vuke and Colton, 1998).

Seismic data demonstrate vertical displacements on basement-rooted faults associated with the Poplar Dome (Shurr and others, 1993; fig. 11). The Fort Union Formation is tilted away from the Poplar Dome in the western part of the Williston Basin, reflecting Laramide-style uplift of the dome that occurred during post-Paleocene time (Orchard, 1987; Thomas, 1974).

Two subbasins are present along the margin of the Williston Basin—the Opheim Basin near Poplar Dome and the Circle Basin near Porcupine Dome (fig. 11). The Blood Creek Basin or Syncline (Thomas, 1974; fig. 11) is arbitrarily included as part of the Williston Basin on some maps (Hamke and others, 1966). Other maps exclude the syncline from the Williston Basin (Shurr and others, 1989a).

Additional Laramide-Style Structures North of the Main Laramide Province

The relationship between foreland structures and sedimentation is also unknown for other structures in the Cordilleran Foreland Basin System that are peripheral to the main Laramide Province. The Consortium for Continental Reflection Profiling (COCORP) deep seismic profiles across northern Montana (fig. 11) show faults interpreted to extend to mid-crustal levels (Baker and Johnson, 2000). The Bowdoin Dome was a long-lived positive tectonic element with a Laramide history (Shurr and others, 1993; Shurr and Monson, 1995). The Bearpaw Arch is a basement-cored Laramide feature (Baker and Johnson, 2000) that was intruded by laccoliths in later Eocene time (Hearn and others, 1964). Similarly, the Little Belt Uplift had a Laramide history, including movement on a basement reverse fault on its northeast side (Baker and Johnson, 2000). The Little Belt area was subsequently uplifted further during the intrusion of later Eocene laccoliths (Marvin and others, 1973), exposing Precambrian crystalline basement rocks in the Little Belt Mountains. The Sweet Grass Hills in northwestern Montana, located along the international border (fig. 11), also had a Precambrian ancestry (Lopez, 2000b).

The Hoagland Basin and associated Coburn Syncline are present between the Bearpaw Uplift and Bowdoin Dome. The Blood Creek Basin (Syncline) (Thomas, 1974) is the most prominent basin in the area and lies along the northern margin of the main Laramide Province (fig. 11).

The Sweetgrass Arch is a basement-involved structure that extends northward into Canada and southeastward to the Little Belt Uplift. The arching of sedimentary rocks is easily discernable on a COCORP deep seismic line (Latham and others, 1988). The structure was strongly influenced by movement on the eastward-migrating Sevier fold-thrust belt to the west (Fuentes and others, 2011; Schwartz and Vuke, 2019) before reaching its present orientation during the final stage of contractile orogenesis (Lorenz, 1982).

LINEAR STRUCTURES

Sets of basement-cored linear structures transect the Cordilleran Foreland Basin System. These include a WNW–ESE-striking set (the Cat Creek, Willow Creek, Lake Basin, and Nye–Bowler left-lateral zones) and a NE–SW-striking set (the Hinsdale, Weldon–Brockton, and Fromberg right-lateral zones), attributed to simple shear (Stone, 1969; Bader, 2019a) from reactivation of Paleoproterozoic conjugate basement fault zones (Bader, 2019a; figs. 2, 11). Subsurface data indicate that many of the faults underwent significant dip-slip reverse movement (Shurr and others, 1989b; Nelson, 1993; Baker and Johnson, 2000). En echelon faults, anticlines, domes, and half domes characterize some of the features (fig. 16), and are manifestations of subsurface basement-involved faults. Others are expressed as zones of linear fault traces. Gravity and aeromagnetic anomalies (Smith, 1970; Woodward and others, 1997) and isostatic residual gravity maps (Bader, 2019a) also help delineate these structures. Reactivation of these basement structures interplayed with basin development in the Laramide Province of the Central and Northern Rocky Mountains (Bader, 2019a), and this framework confines major structural features (Thomas, 1974) throughout the Montana and Wyoming Foreland Basin System.

The Lake Basin and Nye–Bowler WNW–ESE linear structures represent reactivated Precambrian faults (Bader, 2019a) that influenced development of the Laramide Crazy Mountains Basin (Roberts, 1972). The Nye–Bowler structure (Wilson, 1936) is represented by a complex regional anticlinal trend produced by movement on basement faults. The feature separates the Crazy Mountains Basin from the Bighorn Basin (figs. 1, 2), and different structural styles occur on either side of the feature (Johnson and Finn, 2004). The Paleocene Lebo Member of the Fort Union Formation is the youngest preserved unit that is involved in folding associated with the Nye–Bowler linear structure (Johnson and Finn, 2004).

The Cat Creek and Willow Creek linear structures bound parts of the Central Montana Uplift on the north and south, respectively, reflecting reactivated basement faults. A basement-involved thrust fault along the south side of the Big Snowy Uplift aligns with the Willow Creek fault on the south side of the Central Montana Uplift and indicates that it was also involved in Laramide movement (Baker and Johnson, 2000). The Willow Creek and Lake Basin linear features bound the Bull Mountains Basin on the north and south, respectively (Sonnenberg, 1956).

The Bowdoin Dome is geographically associated with the Hinsdale fault, the Poplar Dome is geographically associated with the Weldon–Brockton fault zone (Stone, 1969), and the Blood Creek Syncline is geographically associated with part of the northern Central Montana Uplift. These associations may be genetically related (Thomas, 1974). However, the area north of the Central Montana Uplift may have been stable craton during Laramide tectonism, and movement on the southern part of the Weldon fault postdated Laramide tectonism (Nelson, 1993).

The NE–SW-striking Pendroy fault bisects the Sweetgrass Arch into the northern Kevin–Sunburst Dome and the South Arch, offsetting the fold axes of the two components (Dobbin and Erdmann, 1955; fig. 11). The Pendroy and parallel Scapegoat–Bannatyne subsurface faults have been interpreted as part of a wrench-fault system that was activated by Laramide tectonism (Stone, 1969; Thomas, 1974). A wrench-fault origin of the Williston Basin has also been proposed (Thomas, 1974), specifically involving two NE–SW-striking fault zones that bound the basin (Gerhard and others, 1982), but data from COCORP seismic lines suggest that this hypothesis is unlikely (Latham and others, 1988).

The Belle Fourche paleo-arch is marked by a zone of NE–SW-striking faults in northeastern Wyoming (fig. 2). The faults are related to basement structures, and were active throughout the Phanerozoic. Laramide movement on the faults was sufficient to influence synorogenic basin-fill sediment (Slack, 1981). Some NE–SW-striking faults in southeastern Montana cross-cut the Hell Creek Formation but seem not to persist into the Fort Union Formation; others offset rocks as young as the Tongue River Member of the Fort Union Formation in this area (Vuke and others, 2001a,f). A NE–SW-striking zone of linear features has also been identified in this area, based on satellite images and high-altitude aerial photos (Shurr, 2000).

FUTURE WORK

The distinction between structures of the main Laramide Province and structures identified as "Laramide-style" north of the belt needs clarification, and the relationship between tectonism and sedimentation north of the main Laramide Province needs more research. Some of the structures north of the main Province (figs. 2, 10) extend into Canada, where the term "Laramide" has a different, broader application (Osborn and others, 2006). More research on these structures that transcends the international boundary is needed. Continued thermochronological dating of Laramide uplifts is needed for more refined interpretations of exhumation timing. More work is also needed on the intersecting WNW–ESE and NE–SW faults in central and eastern Montana to update interpretations in light of the deep reverse-fault movement identified on some of them (Nelson, 1992; Shurr and others, 1989b) and their relation to basement structures (Bader, 2019a).

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