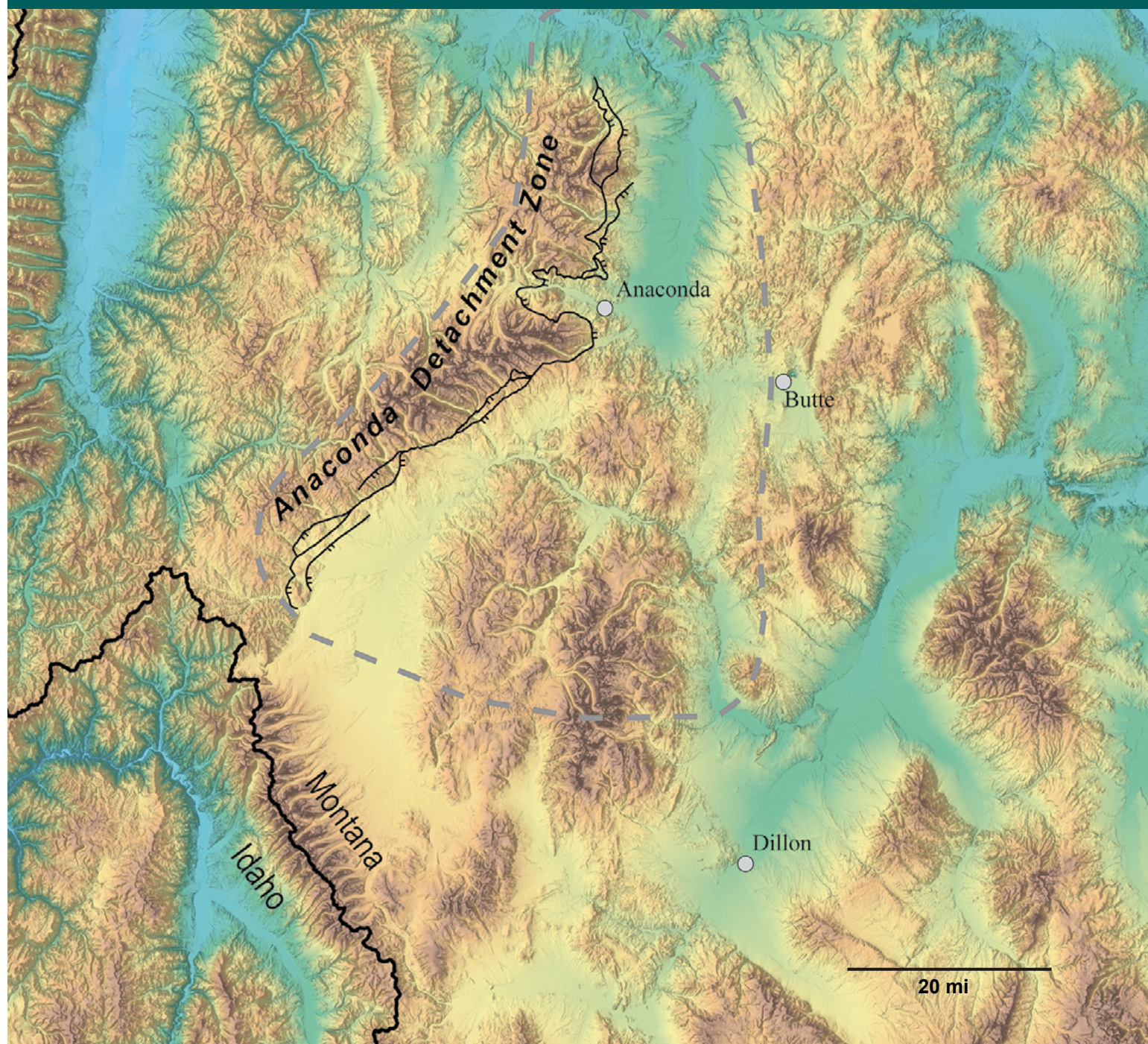


Paleogene Landscape and Basin Evolution in the Missouri River Headwater Region of the Basin and Range Province in Southwestern Montana: Implications for Paleogene Extensional Tectonic Models and Magmatism



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Cover photo: Lidar hillshade map showing study area (dashed line) within the Rocky Mountain Basin and Range Province of southwestern Montana. The Paleogene landscape of southwestern Montana consisted of elongate mountainous highlands, intermontane paleovalleys, and a longitudinal drainage system fundamentally similar to that of today. The paleovalleys were fluvially incised into the Late Cretaceous–Paleocene foreland, similar to that of modern continental thrust belts, and structurally modified and filled during subsequent extension. Late Cretaceous granitic plutons, including two-mica varieties, were exposed in all basin-bounding highland regions of the study area prior to extension.

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ABSTRACT

The intermontane basins that characterize the modern physiography of southwestern Montana (SWMT) were structurally influenced by Eocene–Oligocene extension, including motion on the Anaconda, Bitterroot, Salmon, and Muddy–Grasshopper detachment systems, as well as by middle Miocene Basin and Range-style extension and late Miocene–Pliocene Yellowstone hotspot-related extension. However, interpretations of initial, earlier Cenozoic basin evolution and the relationships to tectonism and surficial processes remain controversial. Key to deciphering basin evolution is the character of the paleolandscape and the geographic distribution of exposed rock types in highland areas, both prior to and during Paleogene extension. To this end, we use a combination of depositional facies, paleocurrent, and compositional data from the Paleogene Renova Formation and equivalents, as well as zircon U–Pb data from both the Renova Formation and granitic plutons exposed in inferred sediment source areas to study the evolution of the Deer Lodge, northern Big Hole, Divide, and Melrose basins.

The Paleogene landscape of SWMT was characterized by elongate mountainous regions separated by intermontane basins occupied by longitudinal drainage systems, both prior to and during Paleogene deposition. Landscape evolution consisted of a pre-Eocene erosional phase during which fluvial systems incised intermontane-scale paleovalleys into the thrust belt infrastructure, similar to modern drainage patterns in continental thrust belts, and provided partial accommodation for future sediments, similar to that of piggy back basins. The locations of intermontane basins and intervening uplifts were similar to those that exist today. Headwaters of the ancestral Missouri River were established during the Paleocene and occupied all basins except the northern Deer Lodge Basin, which contained headwaters of the ancestral Columbia River system. Eocene to Oligocene extensional subsidence, associated with low-angle detachment faulting, greatly added to accommodation and resulted in filling of the intermontane basins.

Late Cretaceous granitic plutons were exposed in all basin-bounding highland regions of the study area. Characteristic of Renova Formation deposits are first-cycle, pluton-derived, detrital zircons that make up Late Cretaceous age groups, with peak ages between 78 and 68 Ma. The age groups closely match those in granitic source areas, an observation corroborated by sample locations, depositional facies, clast compositions, and paleocurrent data. A single Late Cretaceous age group is characteristic of batholiths and stocks in the eastern side of the study area, whereas plutons of the North American Cordilleran Anatectic Belt in the Flint Creek and Anaconda Ranges contain Late Cretaceous igneous-zircon age groups and Paleocene–Eocene igneous-zircon age groups (peak ages 61, 50, and 40 Ma). The younger age groups are not represented in Renova Formation deposits.

Across the basin network, paleo-downstream mixing of zircons derived from an increasing number of plutonic source areas resulted in a unimodal Late Cretaceous age group that exhibits only slight shifts in peak values between ca. 76 and 73 Ma. The identification of specific plutonic sources in the longitudinal fluvial detrital zircon record becomes increasingly speculative in the direction of paleodownstream mixing. Paleocene–Eocene volcanic sources (~56–27 Ma) locally supplied abundant lithic clasts and first-cycle zircons. Thrust belt strata provided Mesoproterozoic quartzite cobbles that typically dominate coarse-clast composition in most longitudinal fluvial deposits. Up to 86% of detrital zircons in the longitudinal fluvial deposits are recycled Precambrian grains from Mesoproterozoic, Paleozoic, and Mesozoic strata.

Sediment derived from two-mica granitic sources is abundant in the Renova Formation in the northern Deer Lodge, Deep Creek, northern Big Hole, and Big Hole Canyon (paleovalley) basins, documenting widespread exposure of two-mica granitic plutons in basin-bounding highland regions (i.e., the Flint Creek Range, Anaconda Range, northern Beaverhead Mountains, and West Pioneer Mountains) during the Paleogene. Detrital zircon U–Pb data characterizing Renova Formation strata in these basins indicate that nearby two-mica plutons were sources of exclusively Late Cretaceous first-cycle plutonic zircons. In contrast, detrital zircon U–Pb data

for Miocene and younger strata in the same basins suggest that two-mica plutons were sources of mixed Late Cretaceous through Eocene zircons. We interpret this trend to reflect progressive magmatism and unroofing of the two-mica plutonic complexes: Late Cretaceous plutons were exposed first, during the Paleogene, whereas Paleocene–Eocene plutons containing Late Cretaceous xenocrysts were exposed beginning in the Miocene.

Our findings indicate that Late Cretaceous two-mica granitic plutons and associated high-grade metamorphic rocks were exposed in the ancestral (pre-Eocene) Flint Creek and Anaconda Ranges prior to Eocene detachment and made up much of the future hanging wall plate. Also, metamorphic–thermobarometric data indicate that >10 km of erosional exhumation and pluton exposure must have occurred prior to detachment in order for Late Cretaceous plutonic detritus to be present in the oldest (~ 54 Ma) Renova Formation deposits. Overall, exhumation history for the Anaconda and Flint Creek Ranges as well as composition of the hanging wall and footwall plates are inconsistent with the current application of a traditional core complex model for the Anaconda Detachment Zone.

The long-term geochronologic record for two-mica plutons in the study area indicates a protracted magmatic history of over 50 Myr, with seemingly cyclical events having an average periodicity of ~11 Myr. This suggests that the two-mica igneous complexes evolved such that Late Cretaceous crystallization ages were preserved at structural levels that, as uplift and exhumation progressed, moved upwards with mixed Late Cretaceous to Eocene age groups developing in what had been deeper structural levels. As plutons were progressively exhumed, the Late Cretaceous to mixed Late Cretaceous–Paleocene–Eocene igneous-age record supplied zircons to younger and younger sedimentary deposits. Overall, magmatism (heating) evidently played an important role in structural evolution, elevation increase, and incisional relief development leading into the Paleocene. Subsequently, semicontinental-scale filling of the greater northwest U.S. intermontane basin system required additional drivers, e.g. subcrustal tectonics and climate change that resulted in regional subsidence, accommodation, and sedimentation.

INTRODUCTION

Continental strata of the late early Eocene to early Miocene (ca. 54–21 Ma) Renova Formation and equivalents (lower Bozeman Group, fig. 1) are the first record of widespread sedimentation and volcanism in southwestern Montana (SWMT), USA, following the end of compression within the Idaho–Montana sector of the North American Cordillera and associated regional unconformity development during the Paleocene and early to middle Eocene (Hanneman and Wideman, 1991; Vuke, 2020a). The strata are key to understanding the Paleogene landscape and tectonic controls on sedimentation, as well as deciphering the tectonic processes and landscape response leading up to the onset of regional extensional deformation, including the exhumation of a proposed metamorphic core complex (the Anaconda Metamorphic Core Complex; O'Neill and others, 2004). With minor exceptions (e.g., Hanneman and Lofgren, 2017), the strata are restricted to modern intermontane basins that exhibit various degrees of Cenozoic extensional tectonic control (Hanneman and Wideman, 1991; Vuke, 2020a).

The tectonic and paleogeographic significance of the Renova Formation and equivalents has long been a subject of much debate. Hypotheses for Paleogene basin origin (see summary by Vuke, 2020a) include development of (1) a network of grabens and half grabens that formed along reactivated thrust faults (Constenius, 1996); (2) grabens and half grabens that formed within a north–south-oriented rift zone that restricted sediment to the rift basin complex, as well as a broad alluvial plain (named the “Renova Basin”) that extended eastward from the rift shoulder and trapped sediment in flexural or erosional valleys (Jänecke, 1994; Thomas, 1995; Stroup and others, 2008); (3) a single, broad alluvial basin (named the “Dillon–Renova Basin”) that existed prior to Basin and Range-style faulting across SWMT (Fritz and others, 2007); and (4) a network of paleovalleys that were incised by fluvial systems during Late Cretaceous(?)–Paleocene compression and uplift and then filled beginning in the Eocene, concurrent with various degrees of extensional faulting and subsidence (Schwartz and Schwartz, 2009a,b, 2013; Rothfuss and others, 2012; Schwartz and others, 2019b). These interpretations of Paleogene

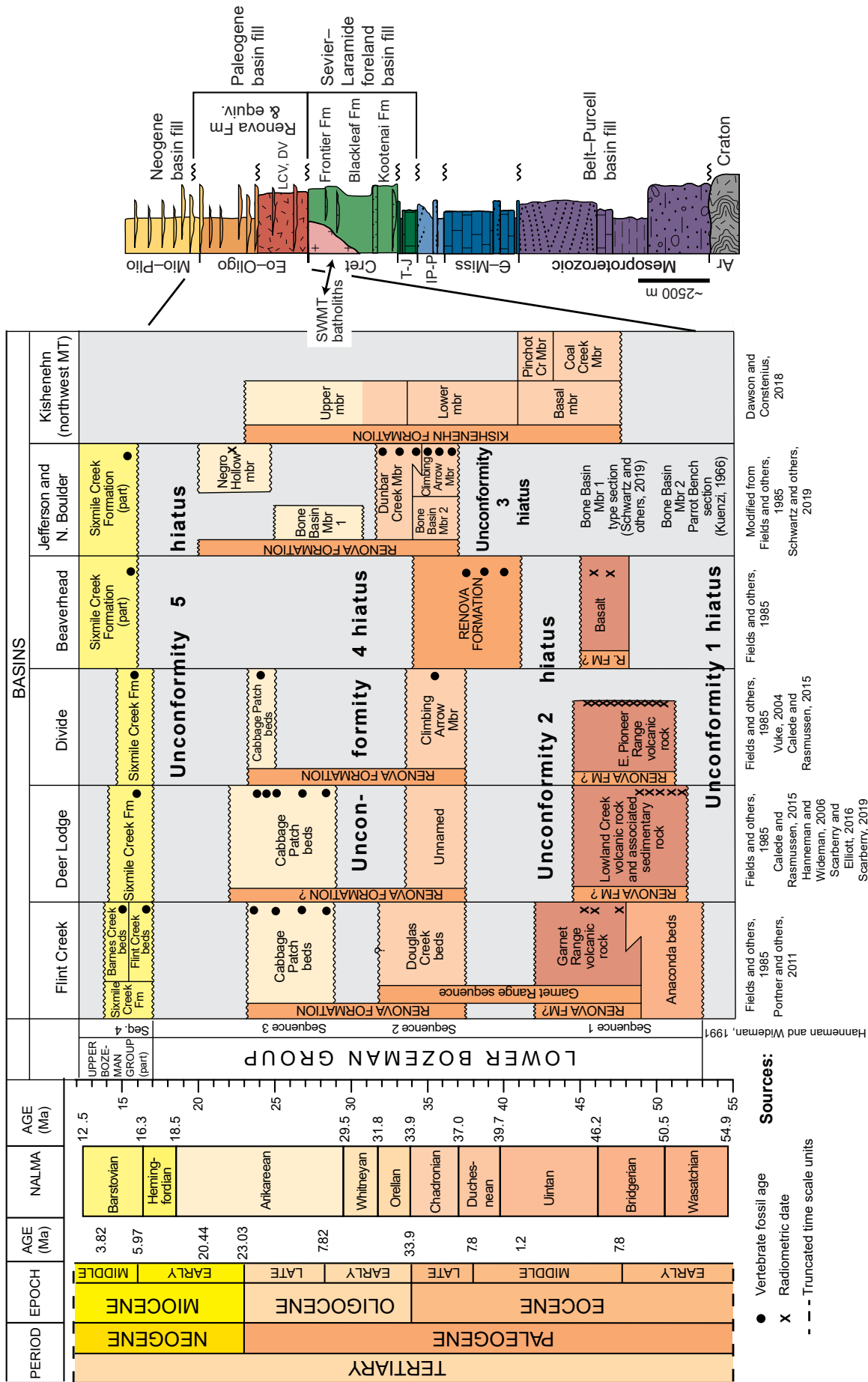


Figure 1. Correlation chart (left) showing Paleogene units in several western Montana intermontane valley basins and corresponding units in northwestern Montana (modified from Vuke, 2020a; North American Land Mammal Ages (NALMA) from Barnosky and others, 2014). Simplified stratigraphic column (right) for southwestern Montana that shows the relative thickness and lithologic properties of Mesoproterozoic through Neogene stratigraphic units exposed in the study area (Ar, Archean; Cret, Cretaceous; _Miss, Cambrian–Mississippian; DV, Dillon volcanics; Eo-Oligo, Eocene–Oligocene; equiv, equivalent; Fm, Formation; LCV, Lowland Creek Volcanics; mbr, Member; Mio-Plio, Miocene–Pliocene; *-P, Pennsylvanian–Permian; SWMT, southwest Montana; T-J, Triassic–Jurassic; modified from Schwartz and others, 2019).

paleotopography and intermontane basin development are highly varied (Vuke, 2020a), in part due to the complicated history of Late Cretaceous through Cenozoic tectonism in the region.

Relevant to determining landscape and basin development during the Paleogene is consideration of the physiography of SWMT prior to Eocene extensional tectonism. Lageson and others (2020) suggested that much of the Late Cretaceous–Paleocene Cordilleran fold-thrust belt and magmatic arc in western Montana and adjacent Idaho made up a tectono-magmatic plateau, the “Montanaplano,” that connected topographically to the south with the Nevadaplano as defined by DeCelles (2004). DeCelles (2004) characterized the Nevadaplano as a broad, high-elevation, low-relief plateau that developed above thickened hinterland crust in the Cordilleran thrust belt of Nevada, Utah, and southern Idaho. Modern analogues for the Nevadaplano include high-elevation plateaus in interior regions of convergent orogens, such as the Andean Puna–Altiplano, Turkish–Iranian, and Tibetan Plateaus (Yildirim and Moores, 1999). Internal drainage is a principal characteristic. As with the Nevadaplano, Lageson and others (2020) describe the Montanaplano as a high-elevation, low-relief, plateau (or “mountain plain”) that was present prior to Paleogene extensional dismemberment in SWMT and, implicitly, to Basin and Range-style extension.

This report focuses on the Paleogene history of intermontane basins and flanking highlands referred to in this study as the “western basins region” (WBR; fig. 2). Key to this effort is the combined consideration of pre- and syndepositional structural setting, paleo-landscape, and sediment provenance, including the compositions and ages of both the Paleogene basin-fills and source rocks exposed in paleo-highland regions. To address these items, we present an integrated analysis of coarse-grained sedimentary facies (i.e., breccia-, conglomerate-, and/or sandstone-dominated facies) and their provenance based on cobble compositions, sandstone modal compositions, and/or detrital zircon U-Pb data. Our findings expand upon those of Schwartz and Schwartz (2009a,b; 2013), providing a detailed, first-time synthesis of observations for Paleogene deposits in the WBR, including adjacent to the bounding normal faults of the Anaconda Metamorphic Core Complex (AMCC). Altogether, our integrated

approach to data interpretation provides evidence for: (1) landscape character and landscape-forming processes prior to deposition of the Renova Formation; (2) individual Paleogene basin geometries and the interconnectedness of adjacent basins; (3) the provenance relationships of basin-fills to basin-bounding paleo-highlands as well as modern highlands; and (4) the relationship between Paleogene basin-fills and the proposed AMCC.

STUDY AREA

Tectonic and Physiographic Setting

SWMT lies within the northern Rocky Mountain region of the North American Cordillera that is marked by the northern extension of the Basin and Range physiographic province (fig. 2). The modern terrain is dominated by north- to northeastward-trending mountain ranges with intermontane hydrographic basins approximately 10–20 km wide and 20–40 km long (fig. 2B). Many of the modern basins are bound by faults of mixed contractional, strike-slip, and/or extensional origin. Extensional structures reflect two major phases of activity, one beginning in the early Eocene (ca. 53 Ma) and the other, Basin and Range-style extension, beginning in the middle Miocene (ca. 15 Ma; Constenius, 1996; Foster and others, 2010; Houston and Dilles, 2013). Most of the basins are currently connected by headwater tributaries of the upper Missouri River, whereas some to the far west are connected by headwaters of the upper Columbia River (figs. 2, 3). Intervening mountain belts contain various assemblages of Archean metamorphic rocks; deformed Proterozoic, Paleozoic, and Mesozoic strata; and Cretaceous and Paleogene plutons and volcanic rocks. Exposures of Paleogene strata largely reside within the bounds of modern basins and unconformably overlie pre-Cenozoic rocks that are of similar character to those exposed in adjacent uplifts (Vuke, 2020a). The maximum thickness of Paleogene strata ranges from 1,200 to 2,700 m in the western part of SWMT (Deer Lodge and Big Hole Basins; Noble and others, 1982; Fields and others, 1985; McLeod, 1987; Hanna and others, 1993) but decreases to <500 m in the east (Smith River Basin; Groff, 1965; Gogas, 1984; Runkel, 1986).

The WBR lies in the southwestern part of the Helena salient of the Cordilleran fold-thrust belt and is one of three, arbitrarily defined regions that con-

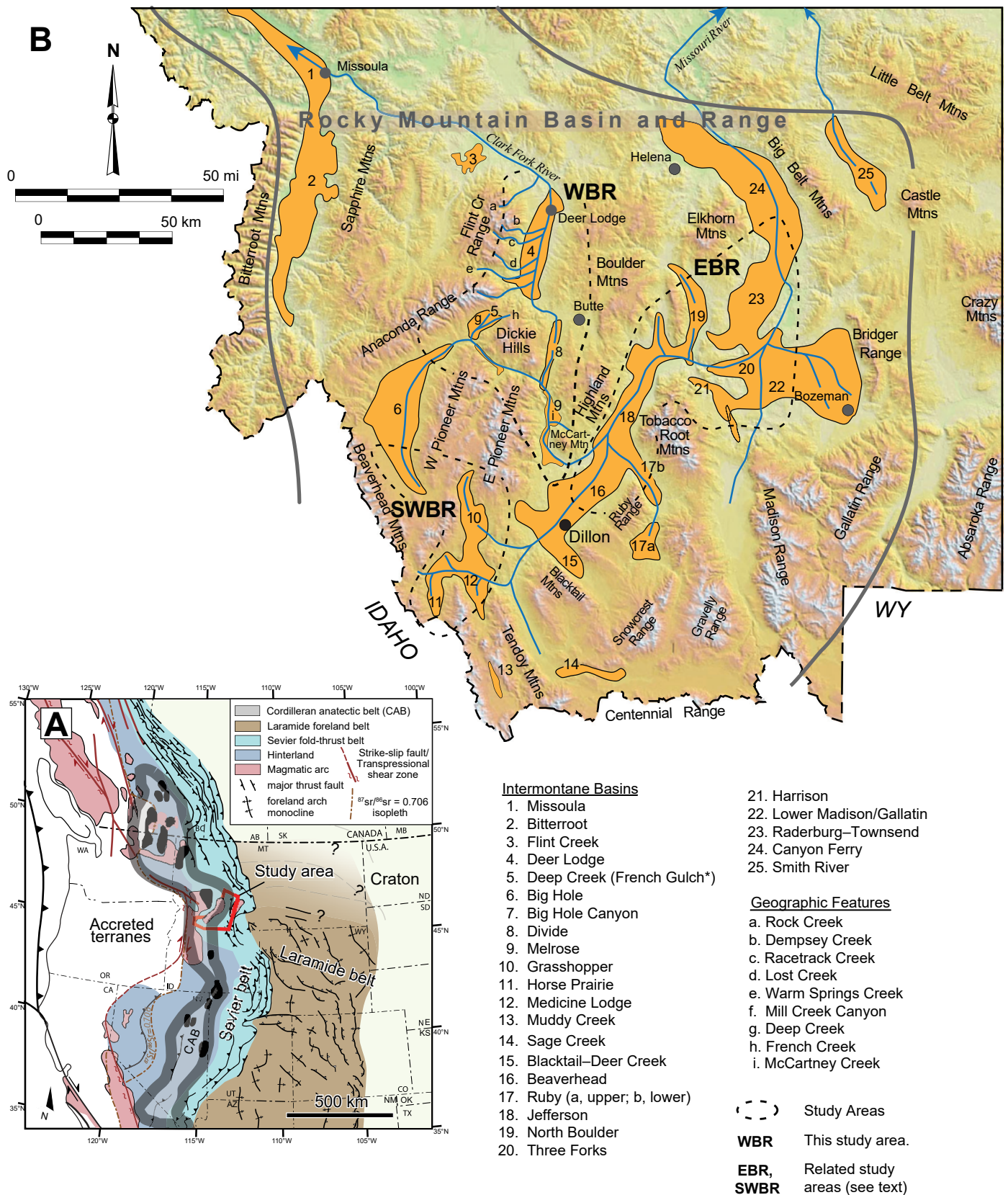


Figure 2. Maps showing (A) tectonic context of the study area (modified from Yankee and Weil, 2015; Parker and Pearson, 2023). (B) Intermontane basins of southwest Montana known to contain lower Bozeman Group (early Eocene to early Miocene) deposits. Blue lines are modern drainages. Dashed lines demarcate the western basins region of this study (WBR) and the southwestern (SWBR) and eastern (EBR) basins regions of related studies.

tain headwaters of the modern Missouri River (fig. 2B). The modern Divide and Melrose Basins make up the eastern half of the WBR and are flanked by the Boulder Mountains–Highland Mountains–McCartney Mountain tract on the east and the East Pioneer Mountains on the west. The modern Deer Lodge, Deep Creek, and Big Hole Basins make up the western half of the WBR and are respectively flanked by the Boulder Mountains and West Pioneer Mountains to the east and by the Flint Creek Range and Anaconda Range–Beaverhead Mountains to the west. We omit the southern Big Hole Basin from this study because of its established Paleogene drainage connection with the “southwestern basins region” (SWBR; fig. 2B; Matoush, 2002; Stroup and others, 2008; Barber and others, 2013). Multiple Late Cretaceous to Eocene plutons associated with emplacement of the Idaho, Boulder, and Pioneer Batholiths (ca. 100–53 Ma; Lund and others, 2002; Gaschnig and others, 2011) are present throughout the region and overprint Early to Late Cretaceous Cordilleran structures (e.g., Vuke and others, 2007). Plutons within the Flint Creek Range, Anaconda Range, and northern Beaverhead Mountains are considered part of the North American Cordilleran Anatectic Belt (CAB), a north–south-oriented chain of Late Cretaceous to Eocene intrusive and anatectic rocks produced by crustal melting (Chapman and others, 2021). In general, magmatic rocks of the CAB include peraluminous, biotite- and muscovite-bearing (i.e., two-mica) granitic intrusions and (meta)igneous rocks associated with metamorphic core complexes (Chapman and others, 2021, their fig. 1). The AMCC, as classified by O’Neill and others (2004), lies along the western margin of the WBR and is the most prominent structural feature preserved by the Anaconda and Flint Creek Ranges. In a classic metamorphic core complex, footwall rocks are tectonically exhumed from the middle and lower crust with upper crustal rocks (any type at the time of detachment) making up the detached hanging wall (Lister and Davis, 1989). In the published AMCC model (O’Neill and others, 2004; Foster and others, 2010; Kalakay and others, 2014), a footwall of Late Cretaceous to Eocene plutons and Late Cretaceous high-grade metamorphic rocks is separated from unmetamorphosed Cenozoic sedimentary and igneous strata by a low-angle detachment zone. Our new results challenge these fundamental aspects of the AMCC model (see Discussion); for this reason, we use the terms “Anaconda Range” and/

or “Flint Creek Range” when referring to rocks composing modern or ancestral highlands coincident with those locations. We also adopt the term “Anaconda Detachment Zone” (ADZ) when referencing the set of low-angle normal faults that mark the topographic expression of the range front. Prior studies of Paleogene sedimentary deposits in the WBR include McLeod (1987), Thomas (1995), O’Neill and others (2004), Stroup and others (2008), Feeney (2009), Roe (2010), Rothfuss and others (2012), and Portner and others (2011).

We provide further context for the WBR by summarizing recent findings for other basins located immediately to the east and southwest. The “eastern basins region” (EBR) is located east of the Boulder and Highland Mountains in the frontal region of the Helena salient (fig. 2B; Camilleri and others, 1997; DeCelles and Graham, 2015). The EBR hosts northward-flowing tributaries of the Missouri River, downstream from headwaters located farther to the west and southwest (fig. 2). Paleogene deposits in the EBR document that Paleogene depocenters were flanked by prominent highlands that were similar in location and lithologic character to modern ranges (Schwartz and Schwartz, 2013). Based upon the difference in elevation of late Eocene–Oligocene paleohillside colluvium and contiguous modern highlands where there are no known intervening syn- or postdepositional faults, paleorelief was a minimum of 2 km. Geologic map relations, sedimentary facies, and compositional data indicate that the depocenters initiated as paleovalleys during latest Cretaceous–Paleocene fluvial incision along zones of stratigraphic and structural weakness associated with pre-existing Cordilleran structures (Schwartz and Schwartz, 2013). Deposition in the EBR was synchronous with the onset of regional extension, the magnitude of which decreased and became more obscure from west to east (Schwartz and Schwartz, 2013). In contrast to the EBR, Paleogene deposits in the SWBR (fig. 2B) have been interpreted as deposits of extensional basins that developed within a narrow, elongate, north–south-trending corridor that extended northward from the Great Basin of Nevada, through eastern Idaho and western Montana, to British Columbia (Jänecke, 1994; Constenius, 1996).

Despite interpretive differences in basin origin, Paleogene strata exposed across the EBR, SWBR, and WBR (as detailed herein) indicate a shared deposi-

tional architecture that defines a regionally connected Paleogene basin network. Paleogene strata consist of variably intercalated deposits of basin margin environments (e.g., alluvial fans and hillslopes) and basin interior environments (e.g., alluvial plains, rivers, floodplains, and lakes; e.g., VanDenburg, 1997, Matoush, 2002; Jänecke, 2007; Roe, 2010; Portner and others, 2011; Rothfuss and others, 2012; Barber and others, 2013; Schwartz and Schwartz, 2013; this study). Paired with sediment provenance and paleocurrent data, the geographic distribution of basin margin versus basin interior deposits defines a rugged Paleogene topography across SWMT with mountainous uplands in similar locations, and composed of similar lithologies, as those that exist today. In other words, the modern landscape of SWMT may be viewed as a coarse analogue for the Paleogene landscape, with subsequent modifications by Eocene extension (particularly in the SWBR and WBR; Jänecke, 1994; Foster and others, 2010), Miocene Basin and Range-style extension (beginning ca. 15 Ma; Jänecke, 2007; Camp and others, 2015), and extension and volcanism associated with passage of the Yellowstone hotspot (beginning ca. 6 Ma; Sears and others, 2009).

Paleogene Stratigraphy

In SWMT, early Eocene to late Oligocene–early Miocene strata of the Renova Formation and equivalents make up the lower Bozeman Group (fig. 1; Robinson, 1963; Hanneman and Wideman, 1991; Rasmussen, 2003; summary by Vuke, 2020a). The lower Bozeman Group is bound by two significant, regionally extensive unconformities (Paleocene–Eocene “Unconformity 1” and early Miocene “Unconformity 5”; fig. 1), and is subdivided into many formal and informal lithostratigraphic members with variable nomenclature (fig. 1; review in Vuke, 2020a). Hanneman and Wideman (1991) further group the lithostratigraphic members into three stratigraphic sequences defined by intraformational unconformities (fig. 1). Given the problematic, inconsistent subdivision of the Renova Formation into many formal and informal stratigraphic units, we adopt the unit nomenclature synthesized in Vuke (2020a) for reporting unit names throughout this manuscript (fig. 1).

Age relations of the Cenozoic strata of SWMT are primarily based upon mammalian biostratigraphy, i.e., North American Land Mammal Ages (NALMAs; e.g.,

Fields and others, 1985; Tabrum and others, 1996; Rasmussen, 2003 and references therein) and similarities between lithologic assemblages (e.g., Hanneman and Wideman, 1991; Schwartz and Schwartz, 2013; Vuke, 2020a). Radiometric age constraints are relatively few, but recent detrital zircon results have suggested Paleogene ages for coarse-grained deposits of previously assumed Neogene age (see review by Vuke, 2020a; this study). In this paper, we broadly use the term “Renova Formation” in reference to all strata within the lower Bozeman Group (fig. 1). This interval includes volcanic and associated volcanoclastic rocks of the Lowland Creek, Dillon, Melrose, and Virginia City volcanic fields (ca. 53–39 Ma; Rasmussen, 2003; Fritz and others, 2007; Dudás and others, 2010; Scarberry and others, 2019, 2020). Although early investigations reported that the Renova Formation is predominantly fine-grained (Kuenzi and Fields, 1971; Fields and others, 1985), more recent studies (e.g., Hanneman, 1989; O’Neill and others, 2004; Schwartz and others, 2009; Schwartz and Schwartz, 2009b, 2013; Rothfuss and others, 2012; Barber and others, 2013; Michalak and others, 2010; Chamberlin and others, 2010) have demonstrated that the Renova Formation is lithologically diverse, ranging from tuffaceous limestone and mudstone to siliciclastic mudstone, sandstone, and boulder–cobble conglomerate and breccia. Depositional facies analyses (Rasmussen, 1977, 1989; Feeney, 2009; Schwartz and Schwartz, 2009a,b, 2013; Schwartz and others, 2009, 2011; Chamberlin and others, 2010; Michalak and others, 2010; Portner and others, 2011; Rothfuss and others, 2012; Barber and others, 2013) document that depositional environments include basin margin settings (e.g., hillslope, alluvial fan, and tributary fluvial environments) and basin interior settings (e.g., alluvial plain, lacustrine, and longitudinal fluvial environments). Although fluvial systems connected the various basins of SWMT (fig. 1), in some areas there were periods of lake development (e.g., Monroe, 1981; Hanneman, 1989; Rasmussen, 1989; Rasmussen and Prothero, 2003; Matoush, 2002; Jänecke, 2007; Portner and others, 2011), possibly reflecting temporary basin closure.

The composition and provenance of Renova Formation sandstone and conglomerate have been addressed to various degrees in a number of studies (Rasmussen, 1977, 1989; Thomas, 1995; Stroup and

others, 2008; Feeney, 2009; Schwartz and Schwartz, 2009a,b, 2013; Schwartz and others, 2009, 2011; Chamberlin and others, 2010; Michalak and others, 2010; Portner and others, 2011; Rothfuss and others, 2012; Barber and others, 2013). Sandstone compositions in longitudinal fluvial deposits are generally feldspar- and lithic-rich, most commonly subarkosic to arkosic (after Folk, 1974), reflecting sediment derivation from Late Cretaceous granitic plutons (e.g., Schwartz and Schwartz, 2013). Other grain types reflect derivation from locally exposed Archean metamorphic rocks and Mesoproterozoic, Paleozoic, and Mesozoic strata, as well as Paleogene volcanic rocks. Overall, sandstone compositions fall within the dissected arc to recycled orogen provenance categories (Schwartz and Schwartz, 2013; after Dickinson and others, 1983). Across the region, pebbles, cobbles, and/or boulders in basin margin deposits reflect derivation from rock types that match those currently exposed in adjacent uplifts. In contrast, pebbles and cobbles in longitudinal fluvial deposits tend to be more polymictic, with individual clast types reflecting local to distal source areas. The most common conglomerate clasts in the fluvial deposits are ultradurable, feldspar-bearing quartzite clasts derived from the Mesoproterozoic Belt Supergroup (“duracasts” of Schwartz and Schwartz, 2013). Muscovite-biotite (“two-mica”) granitic gravel and sand grains, as well as the co-occurrence of detrital muscovite and biotite in trunk-fluvial sandstone of some basins, have received special attention regarding their use as indicators of two-mica granite source rocks (Thomas, 1995; Stroup and others, 2008; Portner and others, 2011; Schwartz and Schwartz, 2013; this study).

METHODS

Report Strategy

Lithostratigraphic members of the Renova Formation (fig. 1) are commonly differentiated on geologic maps and in published reports that document the geology of the WBR. However, many outcrop localities are not formally named because of poor exposure or lack of geochronologic data. We broadly correlate such deposits to the Renova Formation based on the geographic distributions of outcrops, similarities in lithologic properties, and estimated depositional ages from detrital zircon U-Pb data. Rather than grouping and summarizing the characteristics of sedimentary

deposits common to all basins, we selectively present the details of key outcrops to emphasize important observations that provide insight into paleogeography, sediment provenance, and tectonic activity. We augment the results of Schwartz and Schwartz (2013) for the Divide and Melrose Basins and the Big Hole Canyon area, as well as provide the first documentation of outcrops within the Deer Lodge and Deep Creek Basins (fig. 2). We present our results and interpretations in a basin-to-basin manner along interpreted paleodrainage pathways to facilitate paleogeographic understanding. Although data were evaluated on the outcrop scale and sequence scale (after Hanneman, 1989), consistent results through time and across basins allow for synthesis for all of Paleogene time.

Preferred Nomenclature for Modern and Paleogene Physiographic Features

Our results highlight close parallelism between the modern and Paleogene landscapes of SWMT. For clarity in reporting modern versus ancient physiographic features, we herein use the following terms as defined below:

- **Basin (capitalized, e.g., “Deer Lodge Basin”):** A modern hydrographic basin typically named for a local geographic feature, such as a river or town. Such basins are typically elongate, topographically low-lying areas that occur between modern mountainous/highland areas; i.e., they are *intermontane basins* with their scales determined by the modern geographic distribution of topography (e.g., fig. 2). The basins are occupied by longitudinally flowing rivers fed by tributaries draining adjacent highlands.
- **Canyon (capitalized, e.g., “Big Hole Canyon”):** A deep, narrow valley carved into bedrock and occupied by a modern river that may connect adjacent hydrographic basins. Connecting canyons are inferred to be *water gaps*, in which the river occupying the canyon was antecedent to the uplift and/or incision event(s) that resulted in carving of the canyon.
- **Paleovalley:** An ancient river valley incised into older bedrock or sediment as a result of long-term rock uplift, long-term base level fall, and/or other climatic forcings. Paleovalleys may occur at multiple scales across a region. Ideally, lower-order *paleovalley tributaries* occupy higher-elevation catchments and con-

verge downstream to form higher-order paleovalleys. The variably scaled paleovalleys may comprise a *paleovalley network* akin to modern river networks; accordingly, paleovalley width and/or depth may change downstream in a paleovalley network. Paleovalleys may or may not preserve significant sedimentary deposits.

- **Paleovalley basin:** Significant preservation of sedimentary deposits within paleovalleys (e.g., Henry, 2008; Schwartz and Schwartz, 2013; Lundstern and others, 2024; this study) identifies them as an important type of *sedimentary basin*. We systematically use fundamental map unit relations and sedimentary facies distributions of Renova Formation strata to interpret the presence of a *paleovalley basin network* across SWMT, the location of which was at least partly controlled by the structural and stratigraphic framework of the Sevier fold-thrust belt and adjacent Laramide intraforeland region.
- **Paleocanyon:** A paleogeographic analogue to modern canyons. A relatively deep, narrow paleovalley carved into bedrock that connected adjacent, larger-dimension paleovalley tracts. Paleocanyons may or may not preserve sedimentary deposits correlative to those preserved in associated paleovalley basins.
- **Extensional basin:** A sedimentary basin for which accommodation is largely controlled by extension-related faults. Our results indicate that paleovalley basins and extensional basins may not have been mutually exclusive basin types during the Paleogene in SWMT. Rather, there could be geographic and/or temporal variability in basin-forming processes.

Field Methods

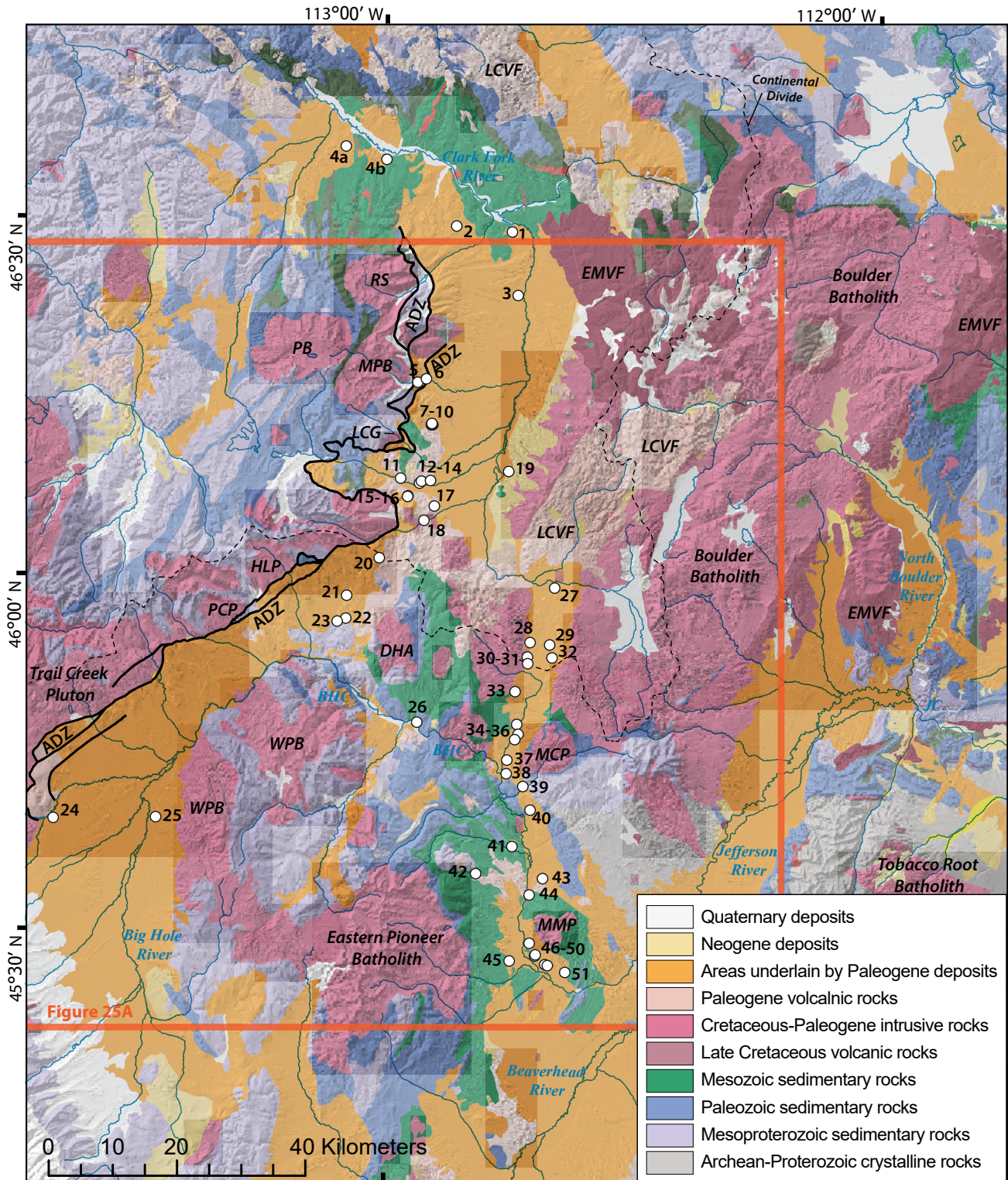
Fifty-one outcrops of Renova Formation-equivalent strata were examined in detail across the WBR (fig. 3; appendix A). Depositional facies were interpreted based upon widely accepted facies models and systematic observations of repeated assemblages of stratigraphic characteristics. We divide coarse-grained deposits into two general categories (following facies definitions by Schwartz and Schwartz, 2013): basin margin deposits and longitudinal fluvial deposits. Basin margin deposits generally document alluvial fan and hillslope environments that exhibit compositional evidence for spatially restricted sediment source re-

gions and/or paleocurrent evidence for paleoflow away from adjacent highlands. In some cases, basin margin deposits may also include the deposits of paleovalley tributaries (i.e., fluvial deposits) that drained upland regions. In contrast, longitudinal fluvial deposits document fluvial channel environments that commonly exhibit compositional evidence for mixing of sediment from multiple sediment source regions and/or paleocurrent evidence for sediment transport parallel to Paleogene basin axes. We treat fluvial style in a general manner, although most exposures of longitudinal fluvial deposits are diagnostic of high-energy (possibly braided) channels. Site-specific aspects of basin margin and longitudinal fluvial facies are presented in the text and used in combination with paleocurrent, sediment composition, and detrital zircon U-Pb data to establish paleolandscape character, sediment provenance, and sediment transport pathways throughout the WBR. Paleocurrent data were collected primarily from imbricated clasts, but also from trough cross-stratification (after DeCelles and others, 1983) and trends of channel axes and current-oriented scour features on the bases of beds. Data collected from tilted beds with a dip angle greater than ten degrees were rotated and restored to horizontal. Pebble- and cobble-composition counts of at least 100 clasts in longitudinal fluvial and basin margin deposits were used to help determine provenance (appendix B), ultimately informing the interpretation of detrital zircon U-Pb data for select outcrops.

Detrital Zircon U-Pb Geochronology

Compilation of Existing Detrital Zircon U-Pb Data

We compiled published detrital zircon U-Pb data for 25 samples from the Paleogene Renova Formation and equivalents exposed in the WBR to enhance sediment provenance interpretations. Samples were originally collected from medium- to coarse-grained alluvial fan and longitudinal fluvial deposits to characterize provenance and constrain depositional age. Location information and references for the compiled samples are presented in appendix C. Prior studies have demonstrated significant lateral variability in sediment provenance trends for the Renova Formation and equivalents in the EBR and SWBR, ultimately related to the variability in sediment sources that were exposed across the rugged Paleogene landscape of SWMT at the time of deposition (Rothfuss and others,



ADZ: Anaconda Detachment Zone
 BHC: Big Hole Canyon
 DHA: Dickie Hills area
 EMVF: Ekhorn Mountains volcanics
 HLP: Hearst Lake plutonic suite

JC: Jefferson Canyon
 LCG: Lost Creek granite
 LCVF: Lowland Creek Volcanics
 MCP: Moose Creek Pluton
 MMP: McCartney Mountain Pluton

MPB: Mount Powell Batholith
 PB: Philipsburg Batholith
 PCP: Pintler Creek Pluton
 RS: Royal Stock
 WPB: Western Pioneer Batholith

Figure 3. Simplified geologic map of southwestern Montana (after Montana Bureau of Mines and Geology, 2025) that displays the distribution of bedrock exposed in the study area. Numbered circles indicate outcrop sites; site coordinates available in appendix A. Solid black line denotes approximate extent of the Anaconda Detachment Zone (ADZ); teeth on hanging wall side. Dashed black line denotes the modern Continental Divide. See figure 2 for map footprint in broader regional context.

2012; Schwartz and Schwartz, 2013; Schwartz and Graham, 2017; Schwartz and others, 2019a). We therefore constrain interpretations of detrital zircon source regions and sediment routing with depositional facies and paleocurrent data, as well as associated cobble/pebble compositions.

In addition, we establish a detrital geochronological fingerprint for Renova Formation samples within the WBR with a composite age distribution ($n = 25$, 1859 grains; fig. 4) to inform provenance analysis on a more regional scale. The composite age distribution for the WBR is similar to that for Paleogene strata across the entire SWMT region (e.g., Schwartz and others, 2019a) and includes grains representing both first-cycle derivation from crystalline sources and recycling from sedimentary sources exposed throughout the region (Stroup and others, 2008; Schwartz and others, 2019a). First-cycle grains were primarily derived from the Paleogene Lowland Creek and Dillon volcanic fields (~56–27 Ma; ~10% of all ages), Late Cretaceous–early Paleogene plutons, and/or the Late Cretaceous Elkhorn Mountains Volcanics (~90–60 Ma;

~50% of all ages). Recycled grains range in age from Archean to early Late Cretaceous (i.e., >90 Ma) and were primarily sourced by Mesoproterozoic to Upper Cretaceous strata that are present throughout the study area (Schwartz and others, 2019a). We note that the recycled Archean (>1.8 Ma; 9%), Proterozoic (1.9–0.61 Ma; 31%), and Paleozoic (543–250 Ma; 2.4%) grains were originally derived from North American crustal provinces and eastern orogens (Schwartz and others, 2019a). Grains between 250 and 120 Ma (~1% of all grains, $n = 13/1384$) correspond to ages for the Triassic–Cretaceous Sierra Nevada arc (250–100 Ma; Schwartz and others, 2019a) and/or the Jurassic–Cretaceous southern Omineca belt (~172–117 Ma; Rosenblume and others, 2021). Grains between 118 and 92 Ma (~6% of all grains, $n = 83/1384$) correspond to early volcanism in the Idaho Batholith (Zartman and others, 1995; Stroup and others, 2008; Gaschnig and others, 2016; Schwartz and others, 2019a).

For new interpretations of detrital zircon U-Pb data presented herein, we focus primarily upon the distributions of grains <120 Ma for several reasons.

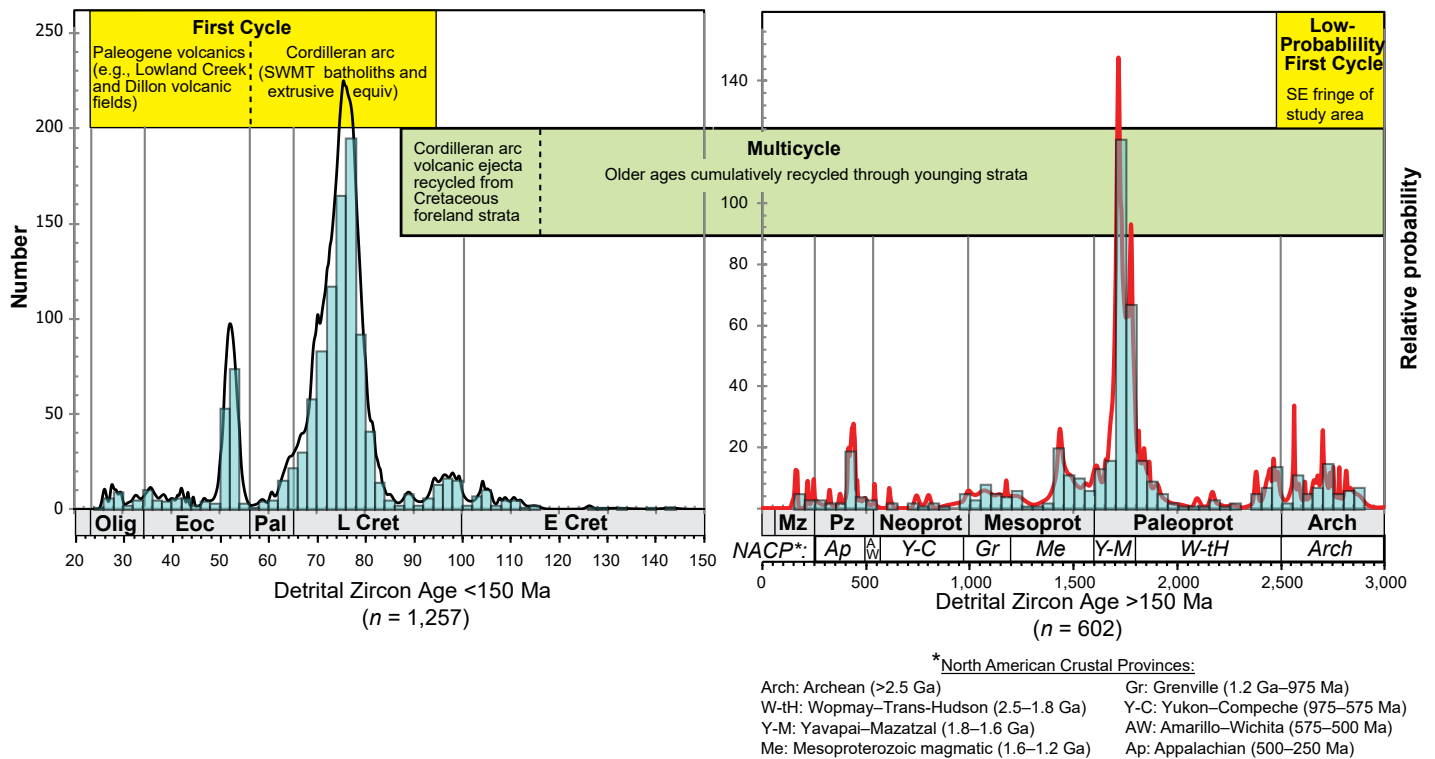


Figure 4. Composite U-Pb age distribution for all detrital zircon samples ($n = 25$, 1,859 grains) within the study area. Interpretation of first-cycle vs. multicycle zircon derivation for this study area modified from Zartman and others (1995), Link and others (2005), Stroup and others (2008), and Schwartz and others (2019). Exposures of Archean crystalline basement near southeast fringe of study area allow for localized primary derivation in addition to recycling. (Arch, Archean; E Cret, Early Cretaceous; Eo, Eocene; L Cret, Late Cretaceous; Mesoprot, Mesoproterozoic; Mz, Mesozoic; Neoprot, Neoproterozoic; Olig, Oligocene; Pal, Paleocene; Paleoprot, Paleoproterozoic; Pz, Paleozoic).

Proterozoic grains are common across the study region and show no significant spatial trends in abundance (appendix D); in addition, they cannot be accurately associated with any one sedimentary source rock because they are ubiquitous across the Mesoproterozoic–Cretaceous stratigraphy of SWMT (Schwartz and others, 2019a). Similarly, recycled grains between ~540 and 120 Ma are sparse and show no significant spatial trends in abundance. In contrast, grains <120 Ma are abundant and can be compared to potential crystalline source rocks that are exposed at multiple locations throughout the study area, thus facilitating detailed comparisons between individual outcrops and source areas.

Compilation of Geochronology Data for Igneous Source Rocks

Felsic to intermediate, Late Cretaceous to Paleogene plutons are currently exposed in all major high-land regions across the WBR (fig. 3; Scarberry and others, 2020). Prior studies have demonstrated that many of the plutons were exposed during the Paleogene and contributed significant volumes of sediment to adjacent basins (Schwartz and Schwartz, 2013; Schwartz and others, 2019a; Scarberry and others, 2020). For this reason, we compile published single-crystal dates for plutons in the region (appendix C) and plot the compiled dates in the same manner as for detrital zircon U-Pb data (i.e., as age distributions; fig. 5) to enable detailed comparisons. Although primary comparative emphasis is based on zircon U-Pb dates, we also compile available single-crystal dates (e.g., from K-Ar, $^{40}\text{Ar}/^{39}\text{Ar}$, and fission track methods) from other minerals (e.g., mica, feldspar, hornblende, apatite) in the same plutons to more fully characterize the plutons with available data. We note that some of the non-zircon dates may record cooling rather than crystallization, and as such, we systematically provide separate plots and assessments of zircon *versus* non-zircon data (e.g., fig. 5A top and bottom).

Boulder Batholith, Eastern and Western Pioneer Batholith, and McCartney Mountain Pluton. Composite age distributions for the zircon and non-zircon data from these four regions are unimodal with a narrow spread. The peak zircon age for the Boulder Batholith is ~76 Ma (fig. 5A) compared to ~71 Ma for the eastern Pioneer Batholith (fig. 5B) and ~73 Ma for the western Pioneer Batholith (fig. 5C). The relatively few

dates available for the McCartney Mountain Pluton show an age distribution nearly identical to that of the Boulder Batholith (fig. 5D). The strong similarity in zircon and non-zircon age distributions in both the Boulder Batholith (fig. 5A) and eastern Pioneer Batholith (fig. 5B) reflect relatively simple emplacement and crystallization histories and suggest that non-zircon dates for those intrusions may be used to corroborate and augment available zircon U-Pb dates for use in comparisons to detrital data. In contrast, the non-zircon dates characterizing the western Pioneer Batholith, which include granitic rocks similar to those in the Anaconda Range (Elliott and Lonn, 2021; Elliott, 2022; Lonn and Scarberry, 2022), reveal a slightly younger peak (~69 Ma) than the zircon dates (~73 Ma; fig. 5C), which may reflect a more complex crystallization and cooling history such as that recorded in the nearby Anaconda Range.

Plutons of the northern Beaverhead Mountains, Anaconda Range, and Flint Creek Range. In contrast to the Boulder Batholith and Pioneer Batholith, zircon and non-zircon age distributions for the Trail Creek Pluton of the northern Beaverhead Mountains (fig. 5E) and plutons within the Anaconda Range (fig. 5F) and Flint Creek Range (fig. 5G) are multimodal, with peaks between ~75 and 40 Ma. The Trail Creek Pluton of the northern Beaverhead Mountains shows a bimodal zircon age distribution with peaks at ~72 and ~64 Ma (fig. 5E). The Anaconda Range plutons have a more complex age distribution with zircon age peaks at ~75 and 61 Ma, as well as non-zircon age peaks at ~50 and 41 Ma (fig. 5F). The coexistence of Late Cretaceous and Paleogene zircons in Anaconda Range plutons has been interpreted to represent Paleocene–Eocene pluton emplacement and crystallization (Foster and others, 2007, 2010; Howlett and others, 2021) wherein Late Cretaceous zircons are xenocrysts (Howlett and others, 2021). The two younger, non-zircon peaks in the Anaconda Range, at ~50 and 41 Ma, have previously been attributed to low-volume, late-stage magmatism and exhumation-related cooling in the Anaconda Range (AMCC of Grice, 2006; Foster and others, 2010; Howlett and others, 2021). Flint Creek Range plutons also have a complex age distribution, with zircon age peaks at ~75, 68, and 40 Ma, as well as non-zircon peaks at ~74 and 66 Ma (fig. 5G), reflecting a complicated, poly-phase history of pluton emplacement and cooling.

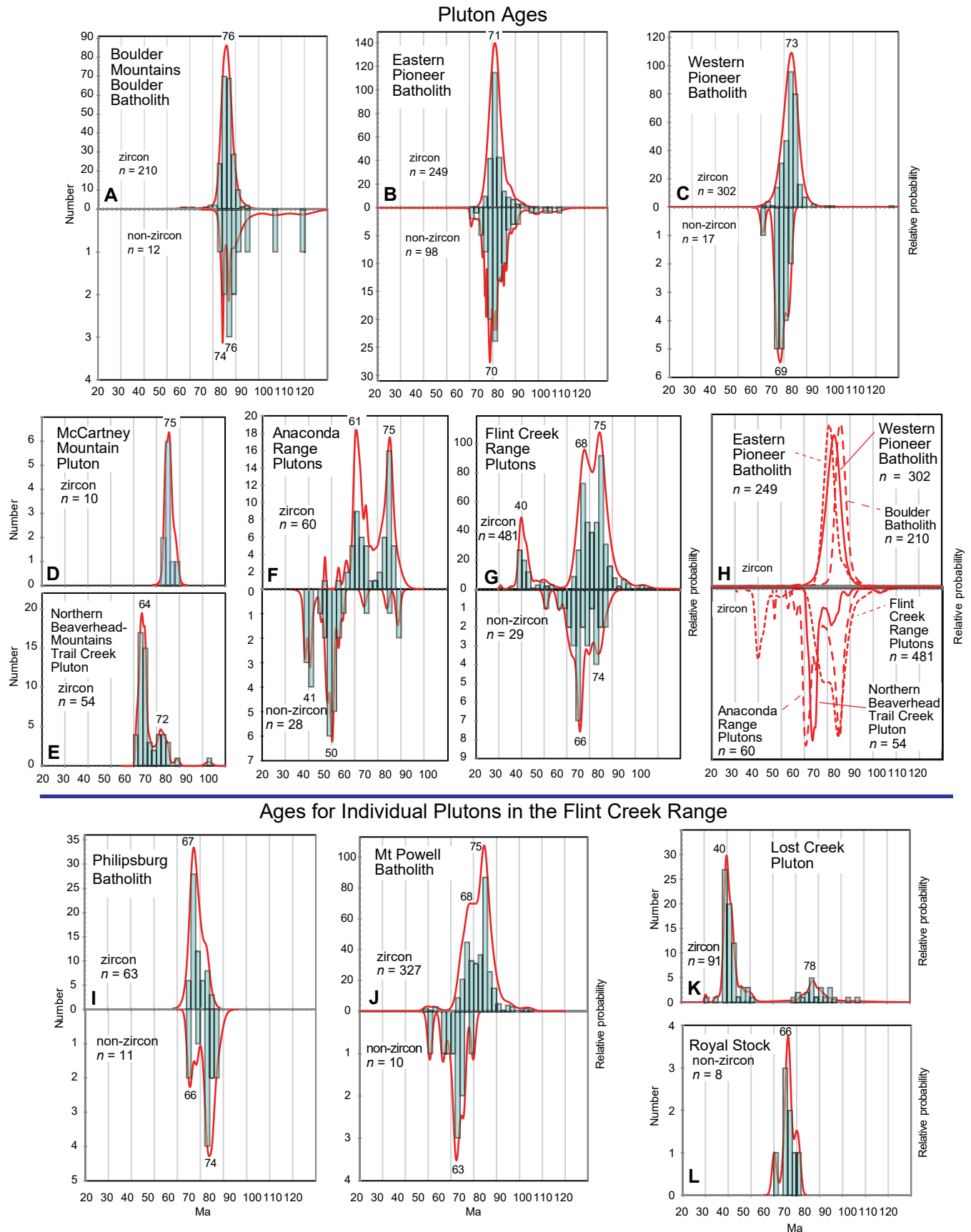


Figure 5. Frequency and relative probability plots of igneous mineral age plutons in the study area. Plots are subdivided into zircon (upper panels) and non-zircon (lower panels) data where both types of data are available. Bin width = 2 Myr. (A–G) Igneous ages grouped for mountain ranges. (H) Comparison of relative probability distributions for zircon ages in groups A–G. (I–L) Igneous ages for individual plutons within the Flint Creek Range.

RESULTS AND INTERPRETATIONS: OUTCROP ANALYSES AND PROVENANCE

In this section, we systematically present observations and interpretations for outcrops, sediment composition, and detrital zircon data in basin tracts of the WBR. For each set of outcrops, we provide a description of the modern basin setting followed by descriptions of longitudinal fluvial and basin margin deposits of Renova Formation age; we note that some basins do not contain good outcrops of both types of deposits, and thus descriptions may be limited to only one deposit type. For outcrops that preserve diverse depositional facies assemblages, we subdivide outcrop descriptions to emphasize key facies components. Conglomerate- and breccia-clast compositional data are presented in appendix B.

Northern Deer Lodge Basin

Basin Setting

The Deer Lodge Basin is an approximately 20-km-wide and 50-km-long, north-trending hydrographic basin in the northwest corner of the study area that contains the headwaters of the modern Columbia River watershed (fig. 2). We treat the northern and southern parts of the Deer Lodge Basin as two separate, but adjoining, Paleogene depocenters based upon observed trends in paleocurrent and compositional data. The northern Deer Lodge Basin, which includes the Gold Creek Basin of Feeney (2009), is flanked on the west by the Flint Creek Range and on the east by the Boulder Mountains. The Flint Creek Range is bound on the east by the early Eocene ADZ (53–39 Ma), whose footwall is primarily composed of Late Cretaceous to Eocene granitic plutons, including the two-mica Mount Powell Batholith, intruded into Mesoproterozoic, Paleozoic, and Cretaceous strata (fig. 3). In contrast, the Boulder Mountains expose the Eocene Lowland Creek Volcanics, Cretaceous Elkhorn Mountains Volcanics, and Late Cretaceous intrusive rocks of the Boulder Batholith (fig. 3). Although Eocene to late Oligocene deposits (sequences 1–3 of Hanneman and Wideman, 1991) are present within the northern Deer Lodge Basin (Portner and others, 2011; Vuke, 2020a), only longitudinal fluvial deposits of the late Oligocene Cabbage Patch beds of the Renova Formation (e.g., Konizeski and Donohoe, 1958; Rasmussen, 1989; Portner and others, 2011; summarized in Vuke, 2020a) are well exposed and described in this study (fig. 1).

Longitudinal Fluvial Deposits

Outcrop descriptions and provenance data (fig. 3, sites 1–3). Ash-bearing mudstone deposits within the Cabbage Patch beds of the Renova Formation contain isolated to laterally coalesced and vertically stacked lenses (~1.5–4.5 m thick) of medium- to coarse-grained sandstone and conglomeratic sandstone (fig. 6). Depending upon location, conglomerate clasts range from scattered pebbles to abundant cobbles and boulders (maximum clast diameter 50 cm). The lenticular sandstone and conglomeratic bodies are characterized by irregular to concave scour bases, contain internal scour surfaces, display upward-fining trends in grain size, and contain trough- and low-angle cross-stratification (~0.5–1.0 m set thickness). Paleocurrent measurements from imbricated clasts and trough cross-strata at individual sites range from west- to northwest- and northeast-directed (fig. 7).

Sandstone is typically coarse- to very coarse-grained and contains abundant angular to subangular quartz, feldspar, and lithic clasts as well as relatively abundant muscovite and biotite. Mica grains are commonly euhedral, with some grain diameters up to 6.5 mm (Mutch, 1960). Granitic lithic grains (some of which contain both muscovite and biotite) account for 11–30% of some sandstones and foliated metamorphic lithic grains account for 4–8% (Feeney, 2009; Schrickler, 2014). Conglomerate clasts at site 1 (fig. 3) consist of 37% Mesoproterozoic Belt Supergroup quartzite and 11% other Belt Supergroup metasedimentary clasts, 23% Elkhorn Mountains Volcanics, 16% two-mica granite and 1% hornblende–biotite granite, 9% Eocene Lowland Creek Volcanics, 2% lithic sandstone, and 1% foliated mylonite. Scattered pebbles and cobbles observed at other locations are consistent with these clast types.

Data from three detrital zircon samples are used to characterize the longitudinal fluvial deposits. Data from site 1 include dates from two-mica granite cobbles as well as dates from their two-mica sandstone matrix. Zircon dates from the cobbles (fig. 8A) consist of two age groups (77 and 68 Ma peaks) that form a bimodal Late Cretaceous distribution (78%, $n = 25/32$, 79–65 Ma), with subordinate Proterozoic dates (22%, $n = 7/32$, 1670–1433 Ma; appendix D). Detrital zircon dates from the sandstone matrix (fig. 8B) show an age distribution dominated by a single Late Cretaceous age

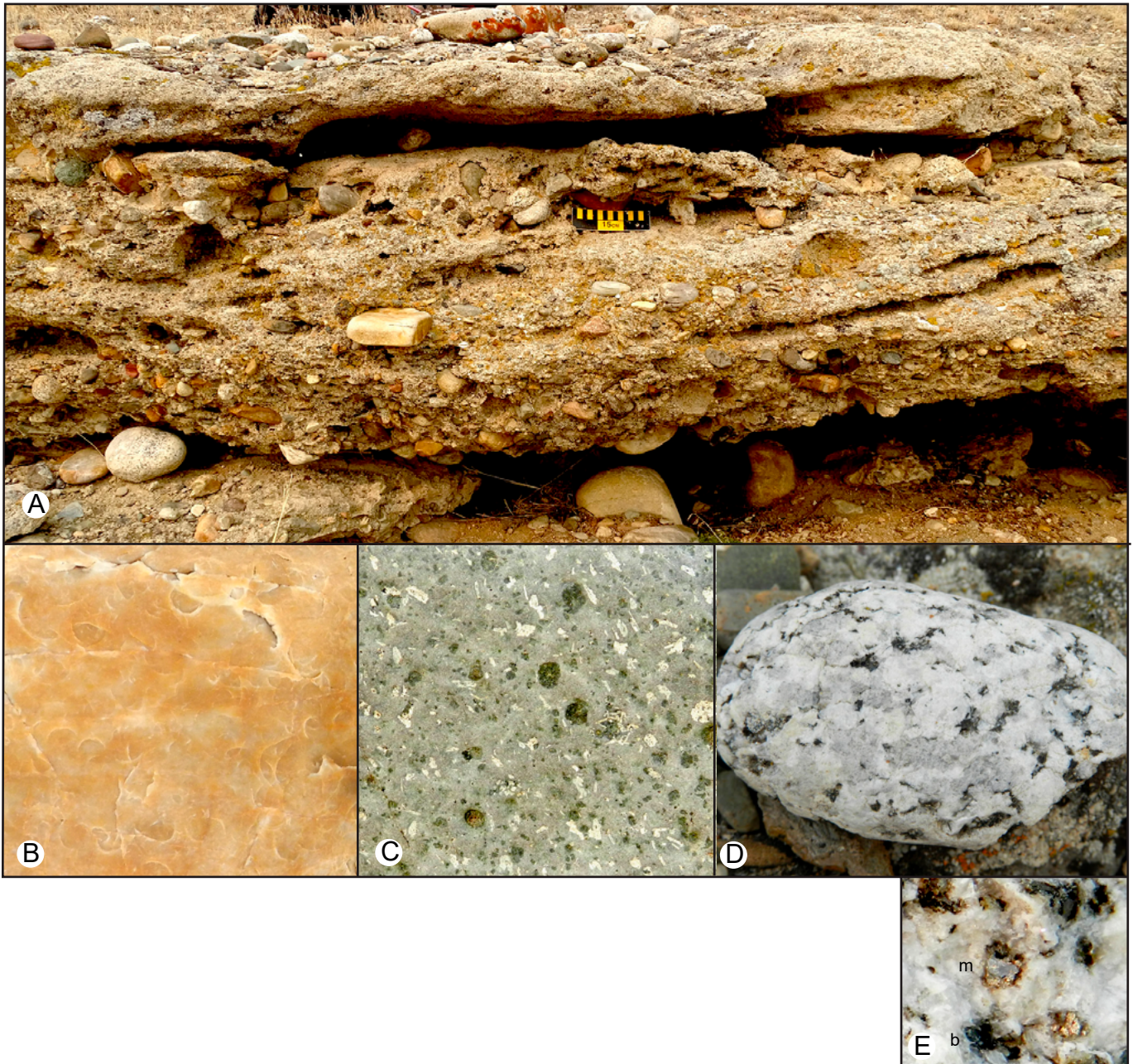


Figure 6. Longitudinal fluvial deposits in the northern Deer Lodge Basin (site 1). (A) Trough cross-stratified conglomeratic sandstone containing boulders and cobbles from source rocks in the ancestral Flint Creek Range and Boulder Mountains. (B) Mesoproterozoic Belt quartzite boulder with percussion marks, source Flint Creek Range. (C) Elkhorn Mountain Volcanics boulder, source Boulder Mountains. (D) Two-mica granitoid cobble, source Mount Powell Batholith, Flint Creek Range. (E) Closeup of two-mica (m, muscovite; b, biotite) granitoid clast. Photographs by R.K. Schwartz.

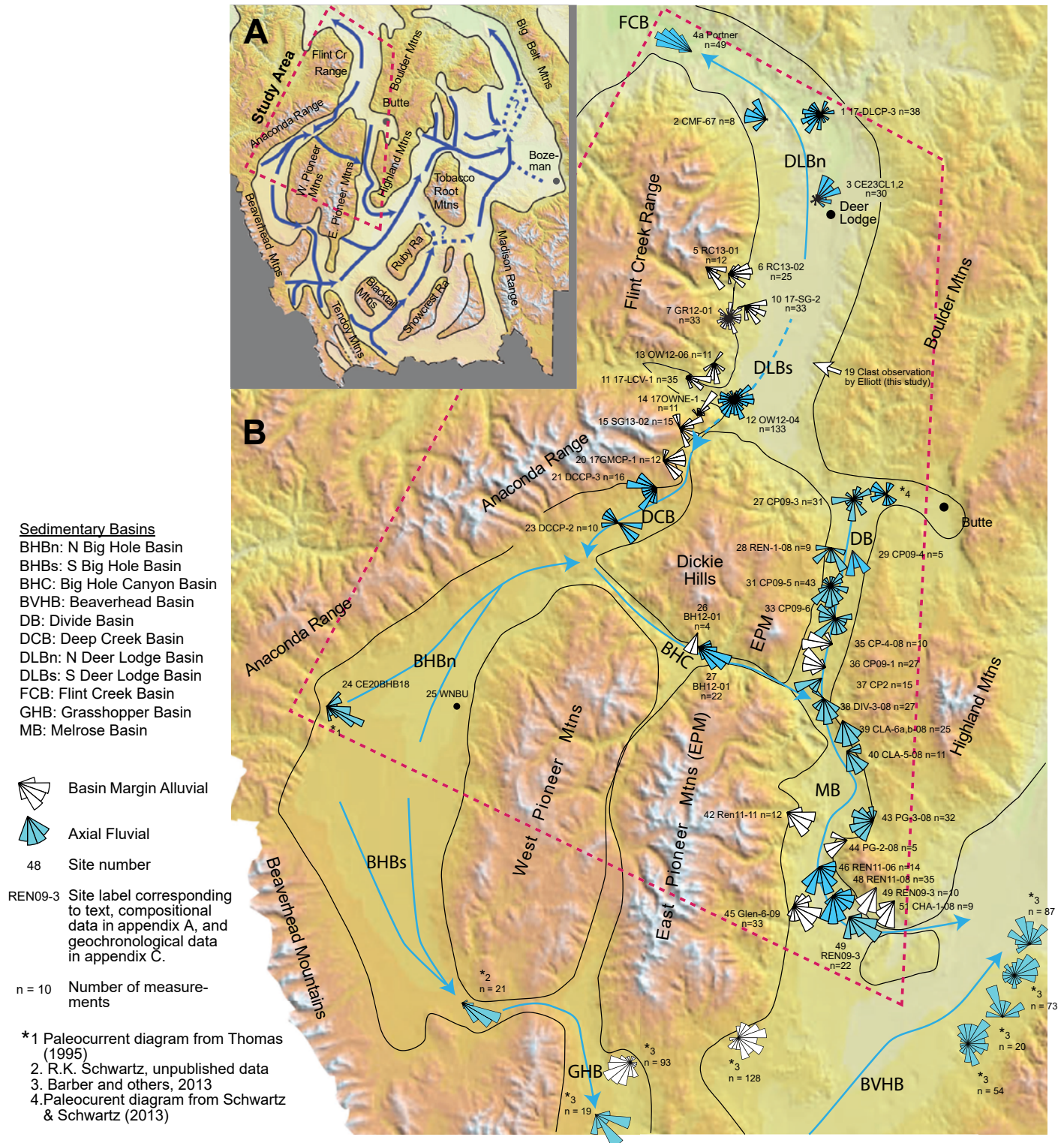


Figure 7. (A) Paleogeographic map summarizing longitudinal fluvial drainage pathways (blue lines) for Renova Formation in the study area (red dashed line) in context of drainage for other intermontane basins in SWMT. Modified from Vuke (2020a). (B) Paleocurrent rose diagrams for basin margin and longitudinal fluvial facies in the study area. Interpreted longitudinal fluvial drainage (blue lines) based upon paleocurrent, compositional, and detrital zircon data.

group (77%, 94/122, 83–70 Ma) with a 77 Ma peak as well as subordinate Proterozoic (12%, $n = 14/122$, 2605–929 Ma), Eocene (7%, $n = 8/122$, 57–46), Early Cretaceous (2%, $n = 2/122$, 111 and 109 Ma), Triassic (1%, $n = 1/222$, 236 Ma), and Cambrian (1%, $n = 1/122$, 474 Ma) dates (appendix D).

Similar to the sandstone data for site 1, the data from site 2 show a single Late Cretaceous age group ($n = 73\%$, 56/76, 82–68 Ma) with a peak age of 78 Ma (fig. 8C). Other subordinate dates are Proterozoic (21%, $n = 16/7$, 1909–1015 Ma), Eocene (3%, $n = 2/76$, 53–52), and Silurian (3%, $n = 2/76$, 441–434 Ma; appendix D).

Interpretations (sites 1–3). The spatial distribution and lithologic properties of the coarse-grained lenticular bodies indicate deposition in a fluvial channel belt characterized by multiple channels encased in overbank mudstone (Rust and Koster, 1984; Bridge, 2006; Miall, 1996). Paleocurrent data indicate a basin-axial orientation of the fluvial system with net northwestward drainage toward the Flint Creek Basin (fig. 7).

Conglomerate clast compositions match rock types that are currently exposed on both sides of the basin. Two-mica granite clasts match the composition of the Mount Powell Batholith, located ~14 km to the southwest of the study sites (fig. 3). The batholith is currently deeply incised by modern valleys (e.g., Dempsey Creek and Rock Creek) that lead northeastward into the Deer Lodge Basin; we suggest that these valleys may mark the locations of Paleogene paleovalley tributaries (this study). Hornblende–biotite granitoid clasts suggest source rocks of similar composition, most likely in the nearby (~10–20 km) Royal stock to the west or, less likely, in the more distant (20–40 km) Boulder Batholith on the eastern side of the basin. Both are implausible sources of two-mica detritus in that muscovite is rare or absent in both (Allen, 1966; Scarberry and others, 2019). Portner and others (2011) also precluded the Boulder Batholith as a source of two-mica detritus. Belt Supergroup strata are currently present in the northern Flint Creek Range, but absent in the western Boulder Mountains, indicating a western source area for quartzite clasts. Low-grade metamorphic and metasedimentary clasts are also consistent with source areas in the Flint Creek Range. Clasts derived from the Elkhorn Mountains Volcanics and Lowland Creek Volcanics support derivation from the eastern and southeastern side of the basin.

The detrital zircon data from the northern Deer Lodge Basin are consistent with provenance interpretations based upon pebble/cobble compositions. A Mount Powell Batholith source for the two-mica detritus is corroborated by the correspondence between zircon dates from the two-mica granite cobbles and those that characterize the Mount Powell Batholith (fig. 8A). Unimodal Late Cretaceous age distributions for both sandstone samples also overlap the age distributions of the Mount Powell and Boulder Batholiths, suggesting that the Boulder Batholith and Elkhorn Mountains Volcanics (86–74 Ma) were likely additional sources. Although zircon U–Pb data are not available for the Royal stock, the presence of Late Cretaceous zircon-age groups in other plutons of the Flint Creek Range (i.e., the Mount Powell Batholith and Lost Creek Pluton; figs. 5J, 5K), as well as the Late Cretaceous non-zircon dates for the Royal Stock (fig. 5L), indicate that it was also a likely contributor of Late Cretaceous zircons. Although it is likely that some Proterozoic detrital zircons were recycled from Mesozoic and Paleozoic strata (Schwartz and others, 2019a) and possibly the batholiths (e.g., Foster and others, 2006; this study), the abundance of Mesoproterozoic and older detrital zircons in association with abundant Belt Supergroup quartzite clasts indicate recycling from Mesoproterozoic strata. Similarly, the rare Cambrian, Silurian, Triassic, and Early Cretaceous grains were likely recycled from Paleozoic and/or Mesozoic strata (e.g., Schwartz and others, 2019a). Early Eocene zircons (e.g., 53–52 Ma) were likely derived from the Lowland Creek Volcanics. Overall, detrital zircon data support the mixing of sediment derived from opposing sides of the basin.

Similar findings are reported for correlative fluvial deposits of the Cabbage Patch beds in the Flint Creek Basin, located about 20 km northwest of the northern Deer Lodge Basin (fig. 3, sites 4a, 4b). A comparative petrographic analysis of sandstone of the Cabbage Patch beds in the northern Deer Lodge and Flint Creek Basins by Feeney (2009) established that the deposits are similar in composition and provenance, and detrital zircon data from Cabbage Patch beds in the Flint Creek Basin (fig. 8D and appendix D; Link and others, 2008; Stroup and others, 2008; Portner and others, 2011) are strikingly similar to those from the northern Deer Lodge Basin. Northwestern paleocurrent data for the Flint Creek fluvial bodies

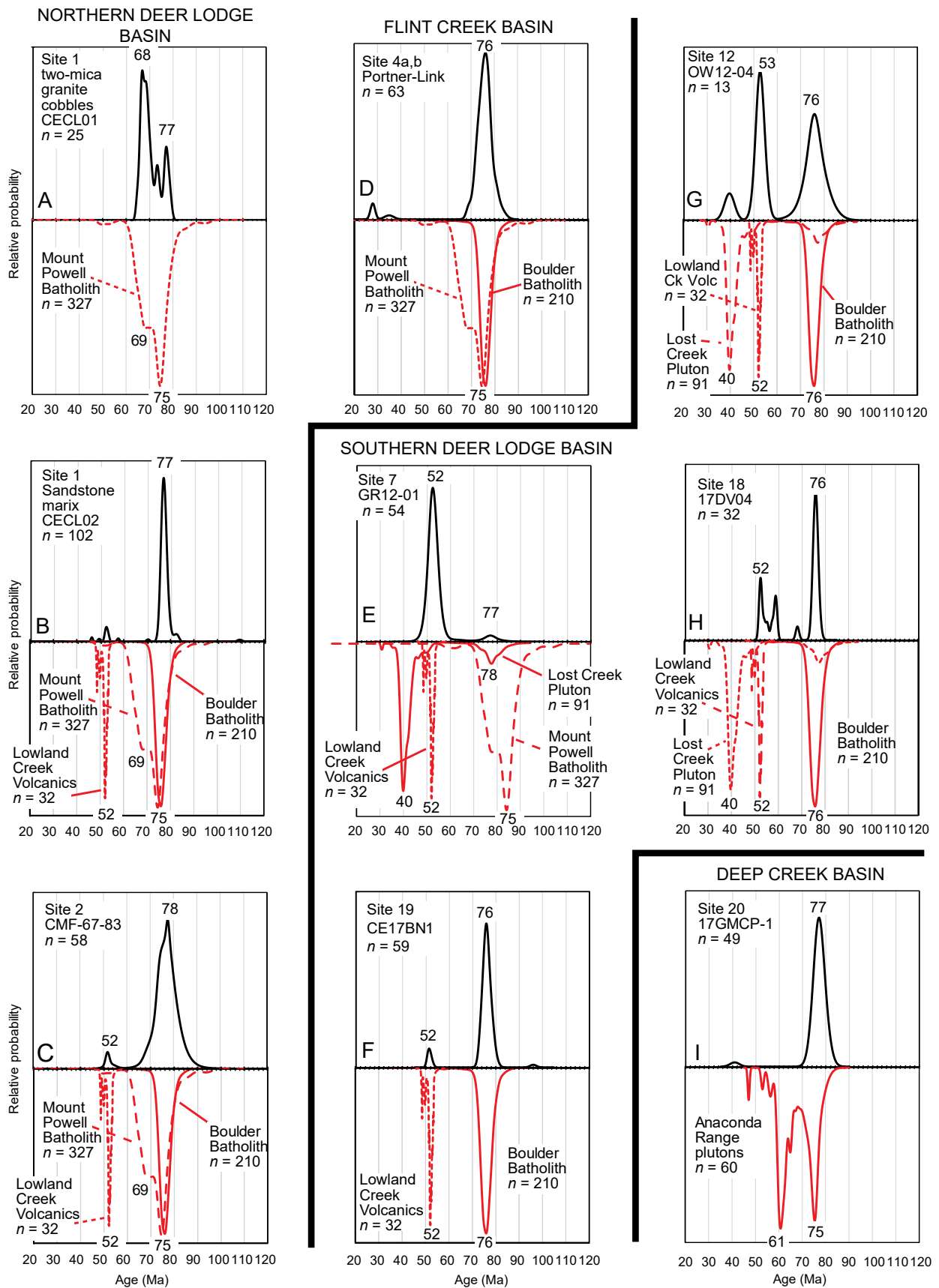


Figure 8. Age frequency and relative probability plots for detrital zircon samples (upper panels) compared with zircon-age probability plots for potential igneous-source rocks (lower panels). Bin width = 2 Myr for all plots. Plots showing the full age range for detrital zircon samples are shown in appendix D. Dashed line in upper part of plot L is for a detrital granite cobble. Bin width = 2 Ma for all plots. Plots showing the full age range for detrital zircon samples are shown in appendix D.

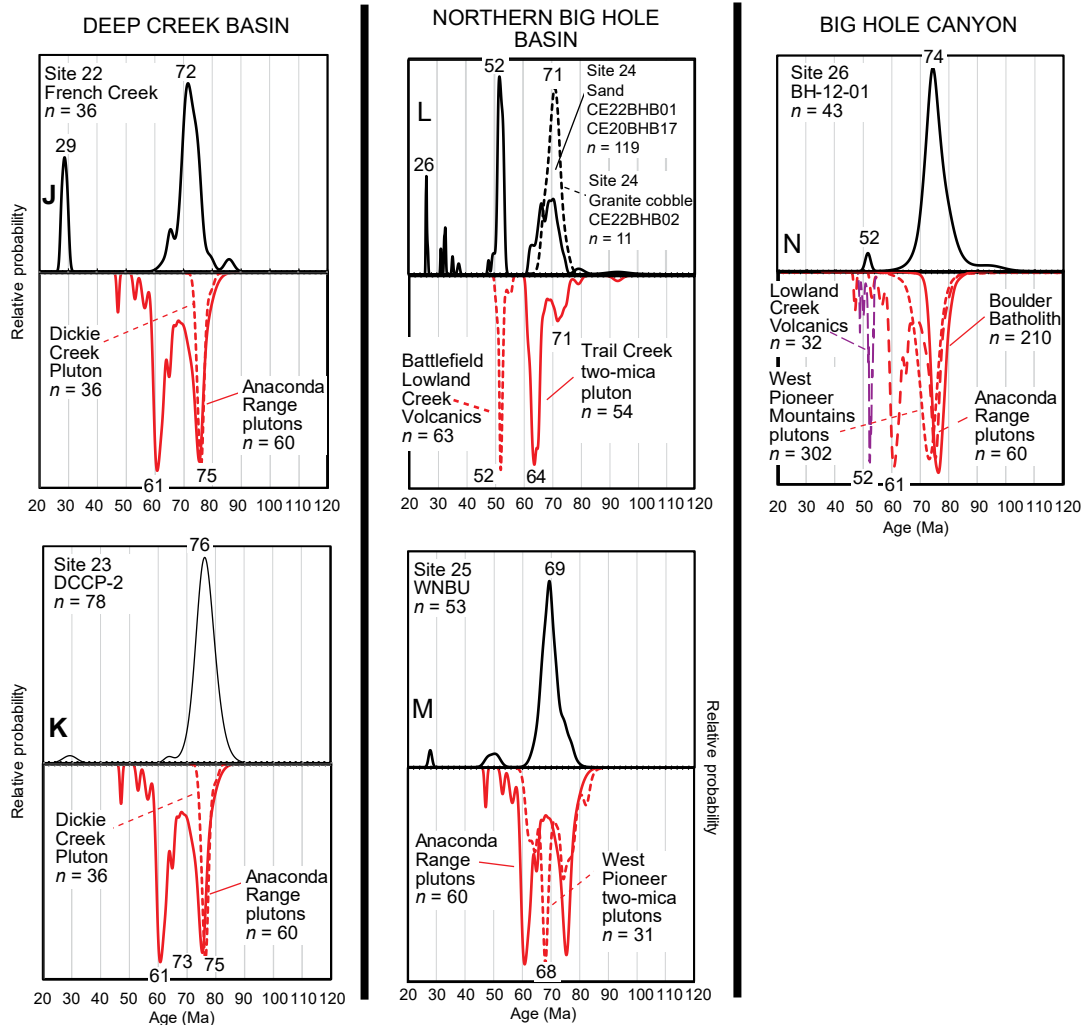


Figure 8—Continued.

(Portner and others, 2011) support northward longitudinal drainage from the northern Deer Lodge Basin into the Flint Creek Basin (fig. 7).

Southern Deer Lodge Basin

Basin Setting

The southern Deer Lodge Basin extends southward from the northern Deer Lodge Basin toward the Pioneer Mountains, where it bifurcates into two narrow zones, one trending southeastward toward the northern end of the Divide Basin and the other trending southwestward toward the Deep Creek Basin (fig. 7). The southern Deer Lodge Basin is bound on the west by the southern Flint Creek Range and on the east by the Boulder Mountains. On the western side of the basin, the east-dipping ADZ bounds the southern Flint Creek Range, whose footwall includes the Late Cretaceous to Eocene Lost Creek Pluton that is intruded into Mesoproterozoic, Paleozoic, and Cretaceous strata

(fig. 3). The Boulder Mountains expose the Eocene Lowland Creek Volcanics, Late Cretaceous Elkhorn Mountains Volcanics, Late Cretaceous intrusive rocks of the Boulder Batholith (fig. 3), and small areas of metamorphosed Paleozoic and Mesozoic strata. Most exposures of Paleogene deposits are present along the western margin of the basin, adjacent to the southern half of the Flint Creek Range and at the northeasternmost end of the Anaconda Range. Paleogene deposits in the southern Deer Lodge Basin are generally older (early to middle Eocene) than the Cabbage Patch beds (late Eocene to Oligocene) of the northern Deer Lodge Basin (Scarberry and others, 2019). The Paleogene deposits are assigned to Sequence 1 of Hanneman and Wideman (1991; fig. 1), which overlies the sub-Cenozoic unconformity and is temporally associated with the early to middle Eocene Lowland Creek Volcanics (Vuke, 2020a). Stratigraphic correlation and nomenclature on an outcrop-by-outcrop basis is problematic due to lithologic complexity and post-depositional

deformation. We consider all but one of the sites in the southern Deer Lodge Basin (site 12, OW12-01-04) as correlative to the Lowland Creek Volcanics, the oldest interval of the Renova Formation (fig. 1).

Basin Margin Deposits (Western Side)

Outcrop descriptions and provenance data: Siliciclastic alluvial deposits and carbonate clast debris flows at Spring Gulch and Racetrack Creek. An ~8 km-long tract of Eocene Lowland Creek Volcanics deposits along the northwestern margin of the basin, primarily exposed between Racetrack Creek and Spring Gulch (fig. 3, between sites 6 and 10), exposes basin margin deposits that we describe in a collective manner due to similar provenance relationships.

Spring Gulch succession (fig. 3, Sites 7–10).

The Lowland Creek Volcanics at Spring Gulch contains a ≥ 200 -m-thick stratigraphic succession that includes a lower, >20 -m-thick, grayish assemblage of siliciclast-bearing conglomerate and sandstone and an upper, ~180-m-thick, reddish assemblage of matrix- and framework-supported limestone clast-rich conglomerate, stacked mudstone units, and thin siliciclastic sandstone and conglomerate lenses. The lower grayish part of the succession contains several widespread, tabular, coarse- to very coarse-grained, lithic sandstone and granule–pebble conglomerate bodies that are interbedded with plant fossil-bearing mudstone (fig. 9A). The sandstone and conglomerate bodies range from ~3 to 10 m thick and are characterized by erosional bases, concave internal scour surfaces overlain by meter-thick lenticular beds, trough and tabular-planar cross-stratification up to 60 cm thick, and strong to subtle upward-fining. Conglomerate clasts are angular to subrounded and commonly show imbrication. Detrital plant fragments are common in the mudstones. Paleocurrent data are highly variable but are dominantly east–southeast-oriented (fig. 7). Sand grains consist of quartz, feldspar, volcanic rock fragments and, commonly, euhedral biotite. No detrital muscovite was observed. Pebble-sized clasts consist of Belt Supergroup quartzite and Lowland Creek Volcanics tuff. Detrital zircon dates from a sample of the lowest fluvial sandstone (site 7 GR12-01) form an age distribution that is dominated by Paleocene to early Eocene dates (86%, $n = 51/59$, 62–50 Ma) as well as lower proportions of Late Cretaceous (6%, $n = 3/59$, 77–76 Ma), Proterozoic (6%, $n = 4/59$, 2702–1433

Ma), and Triassic (2%, $n = 1/59$, 247 Ma) dates (fig. 8E and appendix D).

The overlying reddish succession derives its color from abundant, red calcareous mudstone within the various lithologies. Massive, mudstone-rich, matrix- and framework-supported pebble–cobble conglomerates are thickest (~1.5–3 m) and most abundant in the lower part of the succession (figs. 9B, 9C). Clasts range from angular to rounded. Imbricated clasts suggest southeastward paleoflow (fig. 7). The conglomerate is dominated by Belt Supergroup quartzite (48%), Mississippian limestone (some with healed fractures; 26%), and greenish Belt Supergroup limestone (12%). Minor clast types include marble (5%), lithic sandstone (4%), quartz arenite (2%), and $\leq 1\%$ each of chert, siltstone, tuff, and hornfels.

The lower reddish conglomerate part of the succession rapidly transitions upwards into stacked reddish mudstone beds that are commonly marked by a reworked sandstone cap. Thin ripple-bedded siltstone and lenticular siliciclastic bodies are interbedded with the mudstones and increase in frequency and thickness (up to 2 m) toward the top of the succession. The lenticular bodies consist of distinctly bedded, framework-supported, sandstone, conglomeratic coarse-grained sandstone, and granule–pebble conglomerate with internal scour surfaces, clast imbrication, trough cross-stratification, and normally graded beds, and are overall upward-fining (figs. 9D, 9E). Paleocurrent data indicate flow ranging from northeast to southeast (fig. 7). Conglomerate clasts are dominated by Belt Supergroup quartzite (69%) with minor Mississippian limestone (12%), hornfels (9%), and greenish Belt Supergroup limestone (4%) (fig. 9E). Rarer clasts include marble (3%), quartz arenite (2%), and lithic sandstone (1%).

Racetrack Creek succession (fig. 3, sites 5 and 6). The conglomerate unit within the Lowland Creek Volcanics at Spring Gulch extends ~8 km northward beyond Racetrack Creek and is paralleled along its western side by a zone of reddish limestone breccia that is well exposed at site 5 (Elliott and others, 2013). The reddish limestone breccia is flanked to the west by another elongate, brecciated zone (ranging from sedimentary breccia to cataclasite) of mixed Mississippian limestone and Pennsylvanian Quadrant Formation quartzite (Elliott and others, 2013).



Figure 9. Basin margin deposits at Spring Gulch (sites 7–10). (A) Fluvial conglomeratic sandstone body within the paleo-valley/alluvial plain stratigraphic succession. Staff is 1.5 m. (B, C) Matrix- and framework-supported debris flow deposits in the lower part of the reddish mudstone-rich alluvial fan succession. Limestone clasts (ls) range up to ~38% and indicate proximity of relatively high-relief Precambrian and Mississippian source rocks. (D) Distributary channel fill deposits within upper part of the alluvial fan succession; red lines note scour surfaces. (E) Close-up of channel fill showing well-developed clast imbrication, improved sorting, and decreased abundance of labile carbonate clasts. Photographs by R.K. Schwartz.

At site 5, the succession of reddish limestone breccia is up to 60 m thick, matrix- to framework-supported, and structureless to very crudely bedded (fig. 10A). Compositionally, clasts in the breccia are nearly monomictic, consisting primarily of crinoid-bearing Mississippian limestone. These include abundant angular blocks up to 60 cm in diameter with bed- and joint-surface margins, scattered boulders up to 1.3 m in diameter, and smaller, brittle-fractured limestone clasts. Clast imbrication documents eastward transport (fig. 7). Although Mississippian limestone is dominant, in other locations the breccia may include minor amounts of quartzite, chert, quartz–mica schist, and mylonite. Similar clast types are present in the mapped sedimentary and cataclastic breccia zones located immediately to the west (Elliott and others, 2013).

At site 6, an ~10-m-thick upward coarsening succession of siliciclastic alluvium is present within the conglomerate unit, located approximately 1.4 km east of site 5. Although the stratigraphic position of site 6 relative to site 5 is uncertain owing to structural complexity, site 6 does lie adjacent to patchy exposures of limestone–quartzite breccias similar to those observed at site 5, suggesting that the outcrops are correlative (Elliott and others, 2013). The outcrop hosts a 5 m succession of current ripple-bedded to parallel-laminated, fine-grained sandstone containing scattered, thin lenses of granule to pebble breccia overlain by 5 m of crudely bedded, moderately sorted, cobble breccia containing scattered imbricated clasts (figs. 10D, 10E). Clast imbrication indicates eastward paleoflow (fig. 7). Clasts are a mixture of Eocene mudstone/siltstone intraclasts (37%), Belt Supergroup quartzite (22%), Paleozoic chert (11%) and marble (11%), Mississippian limestone (9%), Eocene tuff (9%), Paleozoic quartz arenite (2%), and lithic sandstone (2%).

Interpretations: Spring Gulch and Racetrack Creek (sites 5–10). At Spring Gulch, the lowest sandstone and conglomerate successions are characteristic of fluvial deposits (Rust and Koster, 1984; Bridge, 2006; Miall, 1996). Their widespread distribution and superposition above plant-bearing mudstone suggest deposition in a fluvial system characterized by flashy discharge and flooding over a vegetated valley floor or low-relief alluvial setting. Although the scattered paleocurrent data alone are not definitive of paleogeographic location within the basin, the association with

overlying coarse-grained debris flow facies within the Lowland Creek Volcanics and other fluvial deposits with basinward paleoflow (fig. 7, site 10) suggest a basin margin setting. Pebbles and granules in the lower fluvial deposits indicate local reworking of Lowland Creek Volcanics deposits and derivation from more distal Mesoproterozoic Belt Supergroup strata. Detrital zircon data support voluminous local input from the Lowland Creek Volcanics, which overwhelmed Proterozoic zircons likely recycled from Belt Supergroup strata and/or Paleozoic–Mesozoic strata (Schwartz and others, 2019a) and possibly the Lost Creek Pluton (F.Ö. Dudás, Massachusetts Institute of Technology Department of Earth, Atmospheric, and Planetary Sciences, written commun., 2022). Similarly, the single Triassic zircon grain was most likely recycled from the Cretaceous Kootenai Formation (e.g., Schwartz and others, 2019a), which is currently exposed along Dempsey Creek and west of the study area. The few Late Cretaceous detrital zircon grains (77–76 Ma), as well as the abundance of quartz and feldspar in sandstones, reflect first-cycle derivation from a plutonic source, most likely the nearby Lost Creek Pluton, located ~5 km to the southwest, which has a similar Late Cretaceous peak at 78 Ma (fig. 8E). Although the Mount Powell Batholith (fig. 8E) is located only ~5 km to the northwest, we preclude it as a source given the absence of two-mica detritus in the deposits.

Debris flow deposits in the overlying reddish succession at Spring Gulch reflect alluvial fan progradation (Blair and McPherson, 1994a,b) with derivation of much coarser and texturally immature clasts of Belt Supergroup quartzite and limestone, as well as Mississippian limestone, from relatively local sources. The abundance of angular limestone in the debris flow deposits indicates tectonized source rocks, which is consistent with rapid development of a high-relief source area. The subsequent upward decrease in debris flow deposits and increase in mudflow deposits and lenticular, siliciclastic channel bodies reflects deposition in more distal alluvial fan settings characterized by distributary channels. The siliciclastic channel bodies near the top of the succession are similar in composition to the debris flow deposits, but are marked by a decrease in Belt Supergroup and Mississippian limestone clasts and an increase in hornfels and Belt Supergroup quartzite clasts.

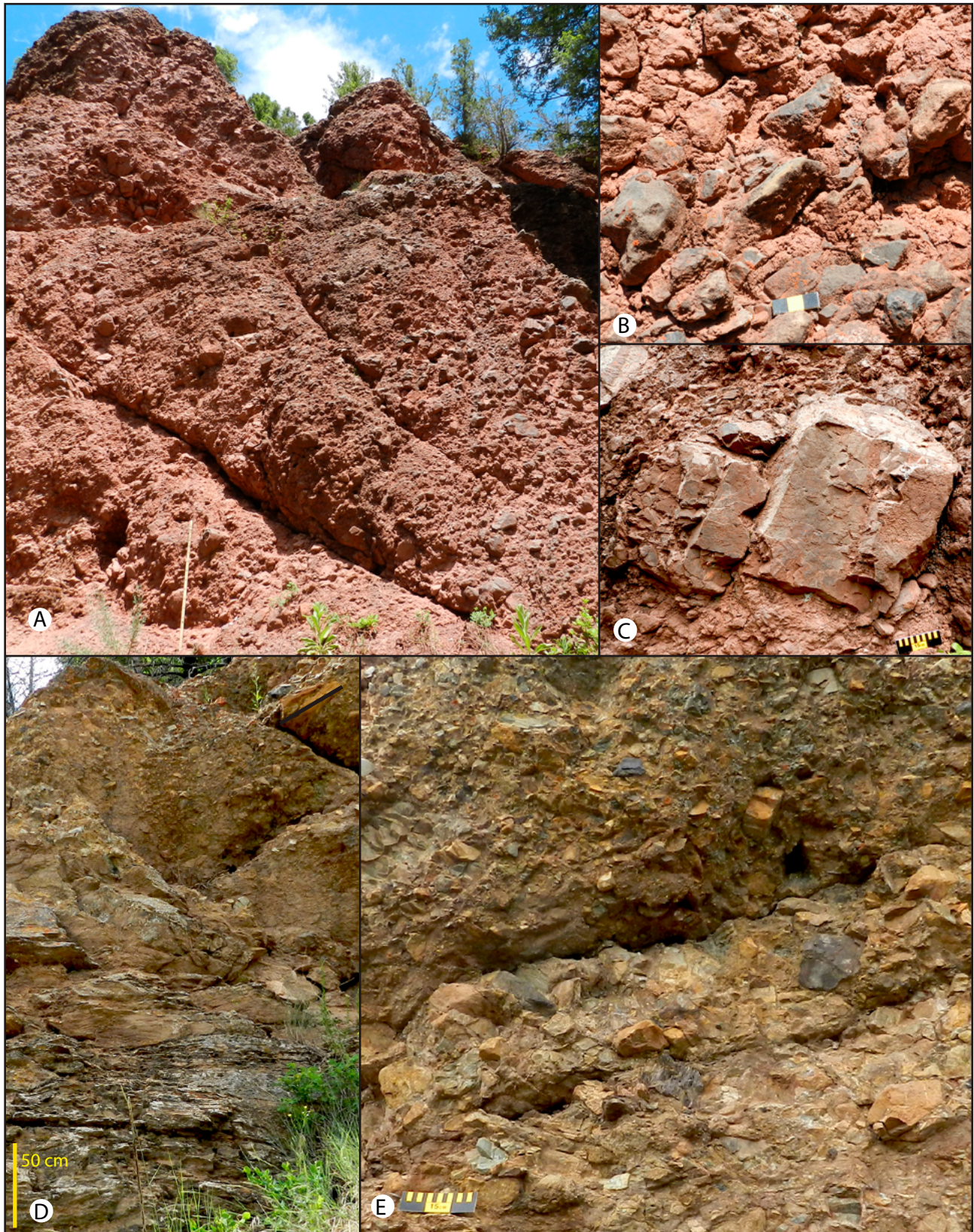


Figure 10. Basin margin deposits near Racetrack Creek (sites 5 and 6). Site 5: (A) Overview of massive to crudely bedded, carbonate-rich debris flow deposits. Staff 1.5 m. (B) Imbricated limestone pebbles in crudely bedded debris flow deposits document basinward (eastward) paleoflow. Scale 3 cm. (C) The size and angularity of carbonate boulders document proximity to a relatively high-relief source area. Scale 15 cm. Site 6: Progradational alluvial fan deposits. (D) Upward coarsening succession from floodplain sandstone and interbedded thin breccia lenses to cobble-sized debris flow breccia. (E) Crude bedding and moderately sorting in breccia. Scale 15 cm. Photographs by R.K. Schwartz.

Overall, fluvial deposits in the lower Spring Gulch succession indicate an initial phase of deposition in a topographically low paleovalley tributary or alluvial plain. Fluvial clast composition and detrital zircon ages document derivation from both range-interior and range-front sources. The abrupt appearance of coarse, proximally derived, alluvial fan detritus followed by a retrogradational fan succession suggests syn-sedimentary faulting and uplift along the range front, presumably along the ADZ, and enhanced localized basin subsidence.

At Racetrack Creek, the debris flow breccia is also interpreted to represent syntectonic deposition proximal to the ADZ (Elliott and others, 2013), perhaps synchronous with debris flow deposition at the Spring Gulch locale. Greater proximity to an uplifted source area than the deposits at Spring Gulch is clearly supported by clast size, deposit fabric, and the presence of brittle-fractured, mylonitic, and rare schistose clasts. Although clasts were primarily derived from Mississippian Madison Group limestone, the presence of Quadrant Formation quartzite and tectonized clasts establishes a compositional and genetic similarity to the mixed limestone and quartzite tectonic breccia zones located a short distance to the west and southwest of site 5. The upward-coarsening siliciclastic succession at site 6 most likely represents alternating deposition of sheet flow, distributary channel, and fluid-rich debris flow deposits in a distal, progradational alluvial fan setting with eastward transport away from breccia-dominated environments. Clasts of quartzite, quartz arenite, limestone, and marble in deposits adjacent to deformed limestone and quartzite breccias support a genetic linkage between the two. Overall, clast compositions, detrital zircon data, and paleocurrent data for the Spring Gulch and Racetrack Creek deposits document derivation from rock types similar to those presently exposed in the Flint Creek Range.

Outcrop descriptions and provenance data: Granite-clast debris flow (fig. 3, site 13) and alluvial fan distributary channels (site 14). Two closely spaced basin margin deposits of distinctly different character are located on the northern side of West Valley along the southeastern margin of the Flint Creek Range. Site 13 is situated near pre-Cenozoic rocks of the range interior, whereas site 14 lies a short distance (1.4 km) closer to the basin interior.

Granite-clast debris flow (site 13). Several massive to crudely bedded, upward-coarsening units of tuffaceous, cobble to boulder conglomerate make up an ~8-m-thick succession within tuffaceous strata of the Lowland Creek Volcanics. The conglomerates are very poorly sorted and matrix- to framework-supported. Cobbles and boulders consist of ~91% angular to sub-angular Lowland Creek Volcanics (10–50 cm diameter) and ~9% subrounded to rounded, blocky shaped, biotite granite clasts (up to ~1 m diameter), some of which exhibit preserved joint surfaces (fig. 11A). One unit is capped by a horizon of concentrated 0.5- to 1-m granite boulders exhibiting south–southeastward-oriented imbrication as well as infilled fluid-scour pits in front of the northern edges of the boulders (figs. 7, 11B).

Alluvial fan distributary channels (site 14). An ~6-m-thick succession of interbedded cobble–boulder conglomerate, planar-laminated lithic-rich sandstone, and reddish sandy mudstone lenses is present at site 14 (fig. 11C). The conglomerate lenses have erosive bases, are typically <7 m wide, and range from thin stingers up to well-defined, ~1.8-m-thick channel-shaped bodies. A framework-supported fabric and crude upward-fining are typical. Although well sorted, conglomerate clast shapes are extremely variable, ranging from angular with preserved joint-surface margins to well-rounded. Clast imbrication documents northeastward paleoflow (fig. 7). Clast proportions are approximately 64% Belt Supergroup quartzite, 16% granite (mostly two-mica granite), and 16% Lowland Creek Volcanics.

Interpretations: Granite-clast debris flow (site 13) and alluvial fan distributary channels (site 14). The sedimentary fabric of the granite-bearing conglomerate beds at site 13 is diagnostic of debris flow deposits (Costa, 1988; Blair and McPherson, 1994a,b) with clast sizes and shapes indicating a relatively proximal and high-relief source area, including derivation from jointed granitic bedrock. The capping granite boulder-concentrated layer at this site reflects lag development across a southward-sloping paleopediment surface. The tuffaceous matrix and texturally immature volcanic clasts were locally derived from Lowland Creek Volcanics strata similar to those stratigraphically bounding the conglomerates. The biotite granite clasts are similar to biotite granite making up the Lost Creek



Figure 11. Basin margin deposits near the northern side of West Valley (sites 13 and 14). Site 13: (A) Granitic debris flow in Lowland Creek Volcanics (Tlc). Oversized boulders, including imbricated granite (gr), are present in the upper part of the deposit, whereas the lower part primarily consists of Lowland Creek lava and tuffaceous cobbles and smaller boulders. Dashed box area marks closeup in B. Staff 1 m. (B) Closeup of imbricated boulder in A showing Eocene erosional surface and obstacle scour pit (dashed white line) along paleo-upslope margin of the boulder. Downward jet-like scour resulted in stepwise plucking of bed-layer segments along the lower right margin of the tuffaceous mudstone boulder (tm) prior to filling by Lowland Creek hillslope breccia (db). Site 14: (C) Floodplain deposits. The alternation of lenticular gravel bodies and widespread sand reflects high- and low-energy flood events across an alluvial surface near the terminus of paleo-West Valley. Clast imbrication and composition document source rocks in the northeastern Anaconda Range. Photographs by R.K. Schwartz.

Pluton (Winegar, 1971; Scarberry and others, 2019), currently exposed ~5–6 km to the northwest. Together, the compositional and south–southeastward-oriented paleocurrent data suggest a nearby source area at the southeastern end of the Flint Creek Range.

In contrast, at site 14 the combination of paleocurrent data, depositional facies, and clast composition indicate a source area, including two-mica granite, on the opposite side of paleo-West Valley at the northeastern end of the Anaconda Range. Abrupt vertical changes in grain size, lenticular bed geometries, relatively good sorting within beds, lamination within sandstone, and clast imbrication are characteristic of intermittent, flood-generated, water-rich deposits (Costa, 1988; Blair and McPherson, 1994a,b). Clast sizes and shapes indicate mixing from proximal source areas, including first-cycle derivation from jointed bedrock and reworking of previously rounded multicycle quartzite. Conglomerate clast compositions match potential source rocks currently exposed in the northeastern Anaconda Range, <5–6 km from the depositional site. Overall, the stratigraphic succession indicates high- and low-energy flood events that incorporated clasts from the northeastern corner of the Anaconda Range, transporting them northeastward across the mouth of paleo-West Valley.

Outcrop descriptions and provenance data: Paleo-valley tributary and volcanoclastic fan-channel succession (fig. 3, sites 15–16 and site 11). West Valley is a prominent fault-associated valley that separates much of the Flint Creek and Anaconda Range (fig. 2). Located a short distance (~3.0–4.7 km) west–south-

west of sites 13 and 14 in West Valley, sites 15 and 16 host several channelized sandstone bodies within a succession of alluvial siliciclastic to tuffaceous strata in the Lowland Creek Volcanics (fig. 12). Detrital zircon dates from associated tuffaceous sediments in the Lowland Creek Volcanics suggest an age of 53–50 Ma (Scarberry and others, 2019). Stratigraphically lowest are two relatively broad (exceeding 60 m), pebbly to granular, coarse- to very coarse-grained siliciclastic channel fills (site 16, fig. 12A). The channel bodies range from 8 to 14 m thick and are each characterized by an erosional base, internal channel-shaped lenses with basal and internal undulating scour surfaces marked by scour pits, abundant trough cross-stratification, planar bedding, and ripple bedding. Pebble-sized clasts include rounded gray quartzite, angular gray aphanitic volcanics, and black hornfels. Sandstone composition consists of abundant monocrystalline quartz, potassium and plagioclase feldspar, and tuff grains, as well as minor quartzofeldspathic plutonic grains, polycrystalline quartz, schist grains, biotite, and muscovite (Schricker, 2014). In nearby locales, the same interval is dominated by an unsorted coarse gravel to boulder conglomerate (maximum clast diameter >3 m) containing Mesoproterozoic and Paleozoic quartzite, quartzite breccia, basalt, siliceous volcanic rocks, and metasiltstone. Available paleocurrent orientations are generally eastward, toward the southernmost end of the Deer Lodge Basin (fig. 7).

Overlying the siliciclastic channel-shaped units, volcanoclastic channel bodies (1–3 m thick) are present within a thick (>30 m) quartz- and feldspar-bearing

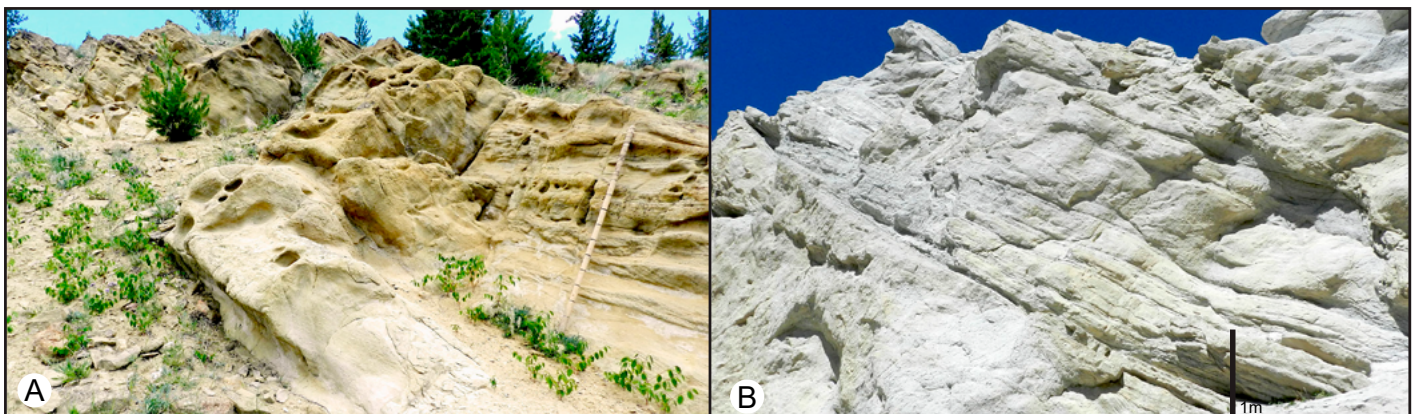


Figure 12. Early Eocene fluvial deposits along the southern margin of West Valley. (A) Terrigenous channel body (site 16) containing clasts derived from Mesoproterozoic through Mesozoic source rocks in the northern end of the Anaconda Range. Staff is 1.5 m. (B) Tuff unit (site 15) containing massive, planar-bedded and channel-shaped deposits documenting fluvial reworking of locally generated pyroclastic debris. Scale is 1 m. Photographs by R.K. Schwartz.

tuff unit (site 15, fig. 12B). Undulating concave scour surfaces, commonly mantled by patches of conglomerate, mark the base and interior parts of the predominantly sandy unit. Sedimentary structures include laminated planar bedding and trough and planar-tabular cross-stratification. Conglomerate clasts consist of whitish gray tuff and rhyolite, and sand-sized grains are compositionally similar to the encasing tuff.

Along the northern margin of West Valley (site 11), an ~10-m-thick, crudely bedded, upward-fining, matrix-rich conglomerate body with an undulating scour base overlies volcanoclastic rocks of the Lowland Creek Volcanics. Conglomerate clasts are commonly ≥ 10 cm diameter, generally range from angular to well-rounded, and include Belt Supergroup quartzite and green metasiltstone, Lowland Creek Volcanics tuff and dacite, Kootenai Formation lithic sandstone, and granite. Some interbeds contain concentrated horizons of small, very angular, Belt Supergroup quartzite clasts. Well-developed clast imbrication indicates paleoflow toward the east-southeast (fig. 7).

Interpretations: Paleovalley tributary-associated fluvial and volcanoclastic fan-channel succession (sites 15–16 and site 11). Sedimentologic properties of the channel-shaped bodies along both margins of West Valley indicate deposition by fluvial systems (Rust and Koster, 1984; Bridge, 2006; Miall, 1996). Along the southern margin of West Valley (sites 15 and 16), the stratigraphic succession most likely represents an alluvial fan distributary setting periodically blanketed by locally derived pyroclastic and epiclastic debris. Net transport of all detritus was toward the southernmost Deer Lodge Basin. The tuff documents a local source of contemporaneous Lowland Creek Volcanic pyroclastic rocks, whereas the siliciclastic fluvial detritus was derived from Mesoproterozoic through Mesozoic metamorphic, sedimentary, and plutonic rocks such as those currently exposed in the northern end of the Anacoda Range (Lonn and others, 2003). The presence of muscovite, biotite, and plutonic grains is consistent with derivation from nearby (~3–4 km) exposures of muscovite–biotite granite, in the same plutonic source area as proposed for site 13.

Along the northern West Valley margin (site 11), the physical properties of the deposits reflect inefficient sorting in an episodic, high-energy fluvial system (Costa, 1988) that also flowed eastward toward the

Deer Lodge Basin. The sediment load included a mixture of previously abraded/reworked (rounded) clasts and locally derived (very angular) clasts. Clast compositions match source rock types currently exposed in headward parts of the West Valley catchment near the southern end of the Flint Creek Range (Lewis, 1998). The most likely source for the small, angular, Belt Supergroup quartzite clasts may have been westward-located paleotalus deposits that mantled paleo-exposures of Belt Supergroup rocks, such as those exposed along the northern side of West Valley (O'Neill and Tysdal, 2005; Schricker and others, 2013).

Although ages of the channel deposits on both sides of the West Valley are most likely diachronous, the combination of paleoflow directions parallel to West Valley, the documentation of source rock types making up headward margins of the valley, and the continuation of Lowland Creek Volcanics and other Tertiary deposits up the valley (Lonn and others, 2003; Elliott and Lonn, 2013) collectively indicate that the modern West Valley marks the location of a paleovalley tributary during the Paleogene. The presence of alluvial fan deposits at the northeastern margin of the valley, having a source terrain on the southwestern side of the valley, documents periodic fan progradation across and beyond the valley mouth (site 14).

Basin Margin Deposits (Eastern Side)

Outcrop description and provenance data: Debris flow sandstone (fig. 3, site 19). A solitary patch of Renova Formation-equivalent strata is present at site 19 along the southeastern margin of the Deer Lodge Basin, adjacent to foothills of the Boulder Mountains (fig. 3). Although we lack detailed sedimentologic data from site 19, a detrital zircon sample was obtained from what has been identified as debris flow deposits within an alluvial fan complex (fig. 8F; Scarberry and others, 2019; Scarberry and Elliott, oral commun., 2020). Detrital zircon analysis of the sandstone yielded five concordant U–Pb dates between 51 and 53 Ma, representing 6% ($n = 5/86$) of the full age distribution (fig. 8F). Most abundant are Late Cretaceous grains (85%, $n = 73/86$, 96–73 Ma). The remainder of the sample includes 1% Jurassic ($n = 1/86$, 166 Ma) and 8% Proterozoic grains ($n = 7/86$; 2614–1262 Ma; appendix D).

Interpretation: Debris flow sandstone (site 19).

The detrital zircon data reflect a suite of source rock types similar to those that are currently exposed along or near the eastern margin of the basin. In order of decreasing abundance, these include Late Cretaceous granite of the Boulder Batholith, Eocene Lowland Creek Volcanics, Upper Cretaceous Elkhorn Mountains Volcanics, and localized patches of Lower Cretaceous Kootenai Formation, Blackleaf Formation, and late Paleozoic carbonate strata.

The Eocene grains provide a maximum depositional age of ~52 Ma. Scarberry and others (2019) reported that these deposits are among the oldest dated siliciclastic Paleogene deposits in western Montana. The abundance of Late Cretaceous detrital zircons reflects proximity and plentiful supply from the Boulder Batholith and possibly the Elkhorn Mountains Volcanics (fig. 8F). The most likely source for the recycled Jurassic grain was Lower Cretaceous Kootenai Formation and/or Blackleaf Formation strata, as suggested by the presence of remnant outcrops. Given the current absence of Proterozoic strata in the adjacent Boulder Mountains, Lower Cretaceous strata and late Paleozoic strata, also indicated by the presence of remnant outcrops, were the most likely sources of recycled Proterozoic grains. Overall, the data document exposure of the Boulder Batholith adjacent to the Deer Lodge Basin as early as ~52 Ma.

Longitudinal Fluvial Deposits

Outcrop description and provenance data: Old works (fig. 3, site 12). A thick (~10 m) cobble–boulder conglomerate unit is well exposed in an ~0.5 km outcrop swath along the southeastern margin of the Flint Creek Range at the juncture of the Deer Lodge Basin and West Valley (figs. 13A, 13B). The conglomerate, included in the informally named “Modesty Creek beds” of Csejtey (1962) and “Anaconda beds” of O’Neill and Tysdal (2005), is in erosional contact with what was designated as “sediments related to Lowland Creek Volcanics (Tlcs)” of Scarberry and others (2019). Although O’Neill and Tysdal (2005) present evidence for interfingering with Lowland Creek Volcanics (53–48 Ma), a detrital zircon maximum depositional age of 39.4 Ma (this study) indicates younger deposition (late middle Eocene or younger), consistent with recent palynological data (Scarberry and others, 2019) from finer grained sedi-

ment within the conglomerate unit that also indicate a late Eocene to early Oligocene age.

The unit consists of multiple laterally and vertically stacked conglomerate bodies interlayered with lithic sandstone, thin coal-bearing mudstone horizons, and mudstones that contain vertical rootlets and gypsum crystals (figs. 13E, 13F). Individual cobble- and boulder-rich (~30 cm maximum diameter) conglomerate bodies are ~4–9 m thick and generally pinch out over tens of meters but may also be tabular over greater distances (≥ 100 m). Basal and internal erosional surfaces, crude bedding, and upward-fining are typical and, in some cases, conglomerate is capped by well-bedded coarse-grained sandstone. The conglomerate is framework-supported to gravel-size matrix-supported and clasts are subrounded to rounded. Well-developed clast imbrication indicates variable paleoflow directions, with a dominant direction toward the south (fig. 7).

Cobbles and boulders are dominated by Belt Supergroup quartzite (51%) and Belt Supergroup limestone (47%), notably containing molar tooth structure, with minor amounts of tuff (1%) and metasiltstone (1%) (figs. 13C, 13D). Although O’Neill and Tysdal (2005) reported rare granitic pebbles, none were identified in this study. Sand-size grains include quartz (monocrystalline 46%, polycrystalline 1%), orthoclase and plagioclase (6%), volcanic rock fragments (37%), metasiltstone/metamudstone (6%), and rare ($\leq 1\%$) schist, mylonite, sedimentary rock fragments, and biotite (Schricker, 2014). Some quartz grains in the dataset (Gazzi–Dickinson method) were components of quartzofeldspathic plutonic clasts (traditional counting method). Detrital zircon dates are dominated by Proterozoic grains (85%, $n = 78/91$, 2815–1420 Ma) with minor Eocene (7%, $n = 6/91$, 54–52, 39 Ma) and Late Cretaceous (8%, $n = 7/91$, 79–74 Ma) grains (fig. 8G; appendix D).

Interpretation: Old Works (site 12). The physical properties and location of the deposits reflect deposition in a high-energy, multi-channel fluvial belt near the western margin of the southern Deer Lodge Basin, whereas the bounding fine-grained deposits indicate overbank deposition with marsh and soil development (Rust and Koster, 1984; Bridge, 2006; Miall, 1996). Imbricated clasts reveal broad variability in paleoflow, with the highest proportion of paleoflow indicators pointing toward the south, approximately parallel to

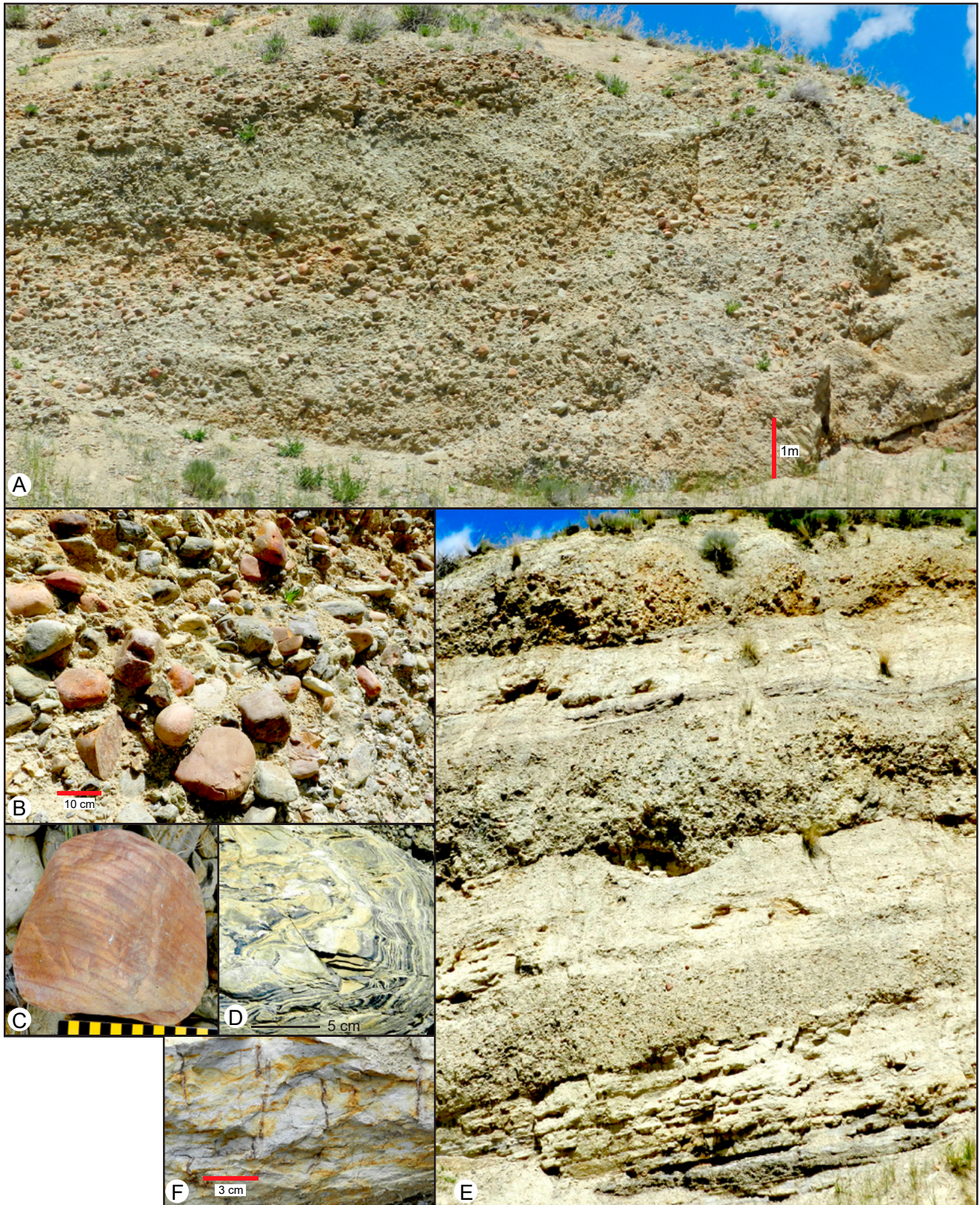


Figure 13. Axial fluvial deposits at the Anaconda Old Works site (site 12). (A) Main channel body consisting of crudely bedded boulder and cobble conglomerate. (B) Closeup of boulder–cobble conglomerate in A. Clast composition is approximately 51% Mesoproterozoic Belt Supergroup quartzite and 47% Belt Supergroup limestone. (C) Mesoproterozoic Belt Supergroup quartzite boulder. Colors range from reddish to various shades of pink, gray, and tan. (D) Mesoproterozoic Belt Supergroup limestone boulder containing tidal laminations. (E) Vertically stacked cobble conglomerate bodies inter-layered with lithic sandstone, very thin coal, and mudstone that contains vertical rootlets and gypsum crystals. Outcrop thickness ~9 m. (F) Sandy mudstone with rootlets in growth position. Scale 3 cm. Photographs by R.K. Schwartz.

the axis of the southern Deer Lodge Basin (fig. 7). The compositional and detrital zircon data are consistent with derivation from rocks currently exposed to the north and northwest in the Flint Creek Range, including exposures of Belt Supergroup carbonate rocks that provided abundant labile conglomerate clasts.

Abundant Proterozoic detrital zircon grains and quartzite clasts similarly indicate recycling from Belt Supergroup strata. Additional, but minor, potential sources for recycled Proterozoic detrital zircons include Paleozoic and Mesozoic strata (Schwartz and others, 2019a) that are relatively prevalent to the northwest, as well as the granitic Lost Creek Pluton that contains xenocrystic Proterozoic zircons (F.Ö. Dudás, written commun., 2022). The association of abundant volcanic lithic clasts and Eocene detrital zircons (ca. 53 Ma) also reflect reworking of Lowland Creek Volcanics that were present along the western side of the southern Deer Lodge Basin (fig. 8G). The source for the single 39 Ma detrital zircon grain is unclear. We posit that deposition by ash-fall processes is most likely, with possible eruptive centers in the Avon and/or Crater Mountain Volcanic Fields (ca. 49–30 Ma), located approximately 50–90 km north–northeast of the study region (McDonald and others, 2020). This is supported by the presence of voluminous ashy detritus preserved in correlative members of the Renova Formation in basins to the east (e.g., the Climbing Arrow Member of the Renova Formation; Vuke and others, 2014).

First-cycle derivation of detrital zircons from granitic source rocks is supported by the Late Cretaceous U-Pb dates (79–74 Ma) in combination with the quartz–orthoclase–plagioclase mineral association observed in sandstones. However, provenance assessment is complicated due to the exposure of similar age source rocks on both sides of the southern Deer Lodge Basin (i.e., in the Flint Creek Range to the west and Boulder Mountains to the east) and the propensity for sediment mixing within the intervening longitudinal fluvial system. Given southward paleoflow indicators at site 12, potential sources include the Mount Powell Batholith and Lost Creek Pluton in the Flint Creek Range and the Boulder Batholith in the Boulder Mountains. Although the 76 Ma age peak in the detrital zircon data is similar to the 75 Ma peak that characterizes the Mount Powell Batholith (fig. 8J), the lack of two-mica detritus at site 12 does not support the

Mount Powell Batholith being a primary source. However, the similarity of the 76 Ma detrital zircon peak to Late Cretaceous peaks that characterize the Lost Creek Pluton (78 Ma) and Boulder Batholith (77–74 Ma; fig. 8G) is compatible with derivation from both sources. Again we note, and later discuss, the disparity between abundant ~50–30 Ma zircon grains in modern exposures of the Lost Creek Pluton (fig. 8K) and the paucity of similar age zircons in the detrital zircon record.

Outcrop descriptions and provenance data: ARCO and Mill Creek sites (fig. 3, sites 17 and 18). In the southwestern corner of the Deer Lodge Basin, pebble–cobble conglomerate and coarse-grained sandstone make up the base of the Lowland Creek Volcanics, which nonconformably overlie a Late Cretaceous biotite- and hornblende-bearing granitic pluton (Scarberry and others, 2019). The sites within the base of the volcanics (Scarberry and others, 2019) lie along the axis of the southwestern part of the basin. Because of restricted access, descriptions of ARCO site 17 are summarized from Scarberry and others (2019) and Scarberry (oral commun.).

Sandstone at site 18 is relatively thick (≤ 5 m) and well sorted. Although locally cross-stratified, other physical structures are indistinct. At nearby site 17 (2.8 km away), a fluvial pebble–cobble gravel contains subrounded to well-rounded quartzite, lithic sandstone, shale/mudstone, granite, and porphyritic volcanic (possibly Elkhorn Mountains Volcanics) clasts. The sandstone at both sites contains abundant detrital plagioclase and potassium feldspar, as well as volcanic, quartzitic, hornfels, and sedimentary lithic grains. No detrital muscovite was observed.

Detrital zircon dates for the sandstone at site 18 include abundant Late Cretaceous grains (55%, $n = 21/38$, 78–68 Ma) as well as Proterozoic (13%, $n = 5/38$, 2614–1243 Ma), Paleocene (13%, $n = 5/38$, 59–55 Ma), and Eocene (16%, $n = 6/38$, 54–52 Ma) grains, plus one Cambrian grain (3%, 494 Ma; fig. 8H and appendix D). The Eocene age group includes five concordant dates ca. 52 Ma (weighted mean 52.19 ± 0.54 Ma).

Interpretations: ARCO and Mill Creek sites (sites 17 and 18). Although the descriptive details of the sandstone bodies are sparse, we provisionally interpret the deposits at both sites as longitudinal fluvial depos-

its based on their proximity and similar stratigraphic positions, as well as their apparent fluvial character. The detrital zircon grains are of mixed recycled and first-cycle origin. The Proterozoic zircons primarily represent recycling from Mesoproterozoic Belt Supergroup strata, supported by the presence of Belt Supergroup clasts, whereas the Cambrian zircon was most likely recycled from Cambrian strata (Link and others, 2017; Schwartz and others, 2019a), both of which currently exist proximal to the sample sites in basin-bounding ranges.

The source for many of the first-cycle Late Cretaceous zircons (78–68 Ma) was likely the underlying and/or adjacent granite, reported to be similar to the Butte Pluton (~76–75 Ma; Lund and others, 2002), or possibly the Elkhorn Mountains Volcanics. Both sources are supported by the presence of granitic and volcanic cobbles. Similar to paleoupstream longitudinal fluvial deposits at site 12 (fig. 8G), the fluvial system would have also allowed for mixing of Late Cretaceous zircons from the Boulder Batholith and the Lost Creek Pluton in the Flint Creek Range (fig. 8H). We preclude the Mount Powell Batholith to the north as a source owing to the lack of muscovite–biotite granitic clasts and detrital muscovite. By extension, we also exclude other two-mica plutons in the northernmost Anaconda Range as potential sources.

The Eocene detrital zircon dates (54–52 Ma) partially overlap known ages for the Lowland Creek Volcanics (53–48 Ma; Dudás and others, 2010). The stratigraphic position of the sandstone below the volcanics at this site helps constrain its depositional age to older than 53 Ma. In addition to site 19 along the eastern basin margin, these deposits are the oldest in the entire Deer Lodge Basin and represent initial deposition associated with pluton exposure in the surrounding ancestral highlands. The source for Paleocene detrital zircon dates (~59–55 Ma) is unknown.

Overall, igneous rocks of Late Cretaceous, Paleocene, and early Eocene age directly contributed detrital zircons to the paleovalley tract. By 54 Ma, Late Cretaceous batholith rocks were exposed adjacent to the study site, as well as in the northern Boulder Mountains. The detrital zircon data are also consistent with exposure of other plutons in the Flint Creek Range (i.e., the Lost Creek Pluton), such as was documented for our stratigraphically younger basin margin sites.

Deep Creek Basin

Basin Setting

A narrow, relatively low-lying basin tract (~11 x 26 km) occupied by the modern Deep Creek and French Creek drainages extends southwestward from the southern Deer Lodge Basin toward the northeast corner of the Big Hole Basin (fig. 2; Elliott, 2015, 2017; Elliott and others, 2025). The modern Continental Divide separates the Deer Lodge Basin from the Deep Creek and French Creek drainages, wherein Deep Creek and French Creek flow southward into the Big Hole River of the Missouri River watershed. We herein refer to this area as the Deep Creek Basin based upon the name of the highest order southward-flowing stream in the basin. The Deep Creek Basin is bound on the west by the northern Anaconda Range and on the east by the Dickie Hills and highlands to the northeast.

The Anaconda Range contains rocks associated with the ADZ footwall, which make up most of the region west of the study area. The footwall rocks include Paleocene to Eocene granitic plutons (Foster and others 2010; Howlett, 2020) and dikes that make up a 5- to 10-km wide zone adjacent to the depositional site, beyond which metamorphosed Mesoproterozoic Belt Supergroup strata, metamorphosed middle Cambrian to Cretaceous strata, and a narrow belt (~0.5–3 km) of Late Cretaceous granodiorite and diorite plutons are present (Wallace and others, 1992; Lonn and others, 2003; Grice, 2006; Foster and others, 2010; Kalakay and others, 2014; Elliott and others, 2025). Farther to the west (~10–20 km), remnants of hangingwall lithologies in the ADZ consist of slightly to unmetamorphosed Mesoproterozoic, Paleozoic, Mesozoic, and Tertiary strata (Kalakay and others, 2014).

The Dickie Hills area includes Late Cretaceous intrusive rocks, Mesoproterozoic Belt Supergroup strata, and minor Paleozoic–Mesozoic strata east of the Deep Creek Basin, as well as Eocene Lowland Creek Volcanics along the northeastern margin of the basin.

Our studied outcrops lie south of the Continental Divide, along the northwestern margin of the Deep Creek Basin where Mill Creek Canyon, a large incisional valley in the Anaconda Range, enters the Deep Creek Basin (fig. 3, site 20). The outcrops continue southward along modern drainages (site 21 and site 22). Site 22 is the French Creek site of Roe (2010).

Basin Margin Deposits

Outcrop descriptions and provenance data: Grassy Mountain (fig. 3, site 20). An ~20-m-thick cobble–boulder conglomerate containing several interbedded, meter-scale units of muddy sandstone and muddy pebble conglomerate is exposed at the northwestern margin of the basin near the mouth of Mill Creek Canyon (Elliott, 2015). Based upon its contact with underlying Lowland Creek Volcanics (53–48 Ma; Dudás and others, 2010; Scarberry and others, 2019) and lateral correlation with likely Oligocene deposits (~31 Ma; Elliott and others, 2025), the deposit is most likely correlative to the Cabbage Patch beds of the Renova Formation. This is weakly supported by the presence of one discordant detrital zircon date that provides an MDA of ~41 Ma.

Physical properties of the conglomerate include crude bedding, subangular to rounded cobble- to boulder-sized clasts (including Late Paleozoic limestone), well-developed clast imbrication, and framework support with a coarse sand matrix. Clast imbrication documents southeastward flow (fig. 7). Approximately 55–65% of conglomerate clasts are metasedimentary and primarily include Mesopro-

terozoic Belt Supergroup quartzite (34%) and Belt Supergroup metacarbonate (including calcsilicate, tremolite- and diopside-bearing metacarbonate, 9%), Mississippian metacarbonate with relict fossils (<5%), and coarsely crystalline white marble (16%; fig. 14). Belt Supergroup limestone clasts displaying algal and molar-tooth structures (fig. 14A) are relatively abundant (15%) and range from unmetamorphosed to greenschist-grade. Other clasts include late Paleozoic limestone (24%) that ranges from undeformed to deformed with shear and vein-filled fractures, Lowland Creek Volcanics tuff (2%), and rare amounts (<0.1% each) of Cretaceous Kootenai Formation conglomerate, hornblende granite, quartz–biotite schist, and two-mica granitic gneiss. Although biotite and muscovite are present in two-mica gneiss clasts, only biotite was recognized in associated sandstone deposits.

Detrital zircon data show abundant and nearly equal proportions of Proterozoic and Archean (46%, $n = 40/89$; 3036–950 Ma) and Late Cretaceous (50%, $n = 45/89$; 81–74 Ma) grains (fig. 8I and appendix D). Rare occurrences include two Cambrian grains (2%, 539 and 501 Ma), a single Silurian grain (424 Ma), and one discordant Eocene grain (41 Ma).

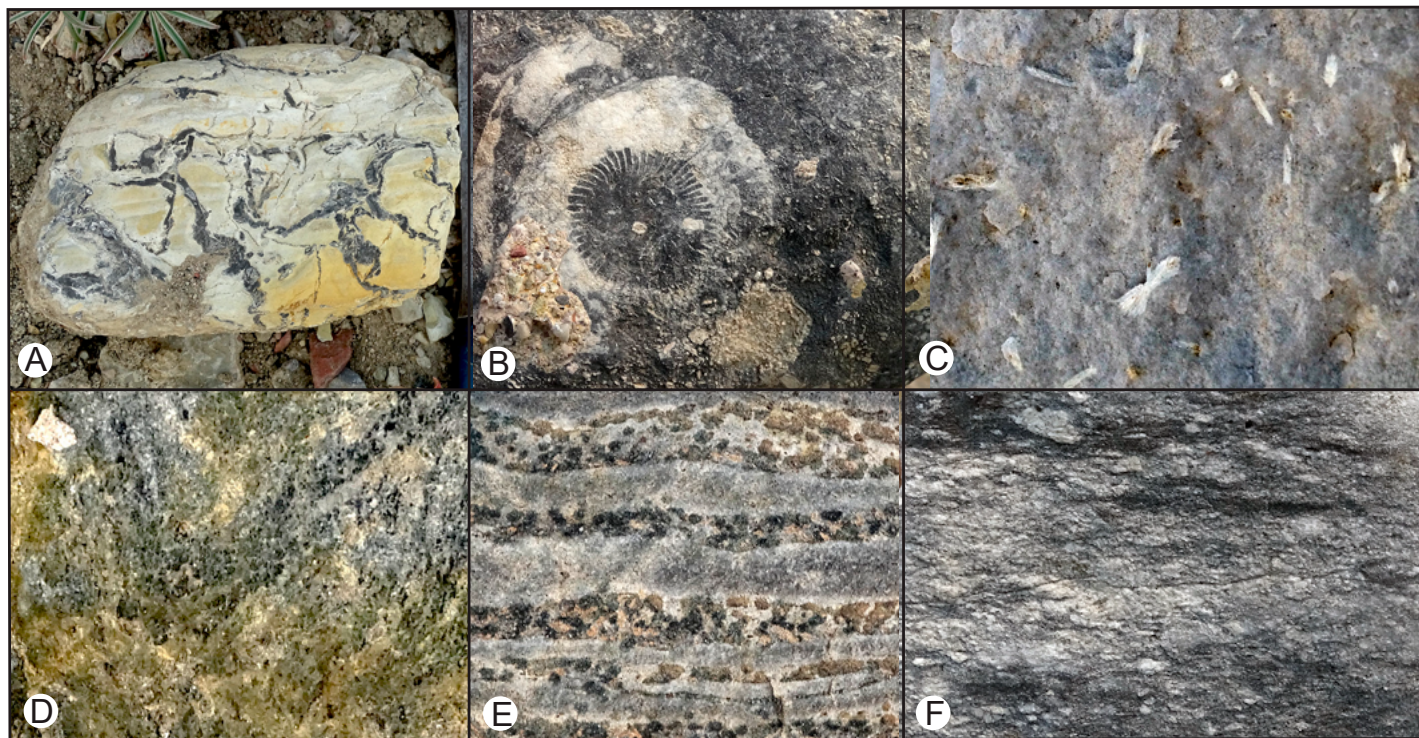


Figure 14. Carbonate and metamorphic hand specimens of basin margin fluvial conglomerate clasts at Grassy Mountain (site 20). (A) Mesoproterozoic Belt Supergroup limestone containing molar tooth structure. (B) Mississippian Madison Limestone. (C) Tremolite-bearing metacarbonate of Mississippian limestone affinity. (D) Muscovite–quartz–unidentified mineral schist. (E) Diopside-bearing metacarbonate of Mesoproterozoic Belt Supergroup limestone affinity. (F) Calcsilicate of Mesoproterozoic Belt Supergroup affinity. Photographs by R.K. Schwartz.

Interpretations: Grassy Mountain (site 20). Physical properties of the conglomerate indicate deposition in a high-energy fluvial system (Rust and Koster, 1984; Bridge, 2006; Miall, 1996) with clast imbrication documenting southeastward flow, away from the mouth of Mill Creek Canyon into the western margin of the Deep Creek Basin (fig. 7). The alignment of paleoflow data with the axis of Mill Creek Canyon and clast provenance indicates the presence of a paleovalley tributary with headwaters extending west into the Anaconda Range, fundamentally similar to the orientation of the modern valley.

Conglomerate clast types match lithologies currently exposed in both the footwall and hanging wall of the ADZ. Although Belt Supergroup quartzite and limestone clasts may have been derived from either side of the ADZ, the quartz–biotite schist, and tremolite–, diopside–, and calcsilicate–metacarbonate clasts were unequivocally derived from Belt Supergroup rocks that underwent deformation and high-grade metamorphism at relatively deep structural levels during the Late Cretaceous (~75–74 Ma) (Wallace and others, 1992; Grice, 2006). Similarly, labile marble clasts were derived from nearby marble units with middle Cambrian limestone protoliths that also underwent Late Cretaceous metamorphism (Grice, 2006). The abundance, large size, subangular to subrounded shape, and deformational fabric of the limestone clasts attest to the presence of brittle-deformed source rocks close to the depositional site. Both the Paleozoic limestone and Cretaceous Kootenai conglomerate clasts match remnant hanging wall exposures located more distally in the up-paleocurrent direction, approximately 10 km to the northwest of the outcrop site (Kalakay and others, 2014).

The detrital zircon data are partially consistent with sandstone/conglomerate compositional data. The abundance of recycled Proterozoic zircons corresponds to the abundance of Mesoproterozoic Belt Supergroup clasts, although some were likely recycled from Paleozoic and Cretaceous Kootenai strata (Schwartz and others, 2019a). Similarly, recycling from Cambrian strata accounts for the Cambrian zircons (Link and others, 2017), whereas recycling of Jurassic–Cretaceous strata in the ADZ hanging wall most likely accounts for the single Silurian grain (Schwartz and others, 2019a).

The remaining detrital zircon dates, Late Cretaceous (50%) and Eocene (1%), are first-cycle grains from igneous source rocks, but do not fully correspond with published ages of plutons currently exposed in the source area, most of which are considered to be Paleocene or Eocene (fig. 8I; Foster, 2010; Howlett, 2020; Howlett and others, 2021). The Late Cretaceous (81–74 Ma) detrital zircon dates do overlap U–Pb ages from quartz–diorite (77–71 Ma; Grice, 2006) and $^{40}\text{Ar}/^{39}\text{Ar}$ and K–Ar ages from granodiorite (79–74 Ma; Wallace and others, 1992; Foster and others, 2010) of the Storm Lake stock, which lies northwest of the Paleocene–Eocene plutonic zone in the ADZ. Although recently acquired zircon U–Pb dates for intervening and more abundant granitoid rocks within the Paleocene–Eocene, two-mica Pintler Creek Pluton show that modern exposures of the batholith dominantly yield Paleocene (65–56 Ma) crystallization ages, Late Cretaceous (76–67 Ma) crystallization ages are also common, and older Jurassic to Early Cretaceous ages are also possible (Howlett, 2020; Howlett and others, 2021). Although visual evidence for muscovite detritus in the sandstone deposits was rare, we interpret that two-mica granitoid rocks of the Anaconda Range were primary sources, and later discuss this discrepancy. We suggest a Paleogene volcanic origin, rather than plutonic, for the single 41 Ma date.

Longitudinal Fluvial Deposits

Outcrop descriptions and provenance data: Deep Creek (fig. 3, sites 21 and 23) and French Creek (site 22). Channel-shaped sandstone and conglomerate bodies are present within the Cabbage Patch beds of the Renova Formation in the central to southern part of the Deep Creek Basin (Elliott, 2017). We present new data for exposures along Deep Creek as well as describe and reinterpret existing compositional and detrital zircon data for deposits along French Creek (Roe, 2010). Deep Creek and French Creek are southward-flowing rivers with headwaters along the western and eastern sides of the basin, respectively; they join downstream before entering the Big Hole River. The “French Creek” sample site of Roe (2010; site 22 herein), only ~1.2 km east of site 23, is located within the French Creek drainage.

At both sites, isolated to relatively widespread, lenticular bodies (<1–10 m thick) of sandstone, pebbly

sandstone, and conglomerate with erosional concave bases are present within clay, silt, and tuffaceous successions in the Cabbage Patch beds (Roe, 2010; Elliott, 2017; this study). Lenticular bodies along the Deep Creek drainage are larger and commonly conglomeratic, whereas those along the French Creek drainage are thin (<1 m), finer grained, and pinch out rapidly. Sandstones are typically feldspathic, coarse-grained, angular to sub-angular, and lithic-rich. Where present, pebbles and cobbles are commonly imbricated. Upward-fining is typical of channel bodies, and although internal structures are seemingly absent for French Creek deposits (Roe, 2010), trough cross-bedding is common at Deep Creek site 23. Paleocurrent data from Deep Creek site 23 document southwestward flow, whereas data from site 21 are inconclusive (fig. 7). Paleocurrent data were not obtainable at French Creek site 22.

Sandstones at the Deep Creek Basin sites contain abundant muscovite and biotite, including large euhedral flakes of both varieties (Elliott, 2017). In addition, black lithic grains, most likely hornfels, and transported wood fragments are common. The conglomerate at site 21 contains Belt Supergroup quartzite (41%), biotite- and muscovite-bearing schist (33%), quartzofeldspathic crystalline clasts (16%), and a minor hornfels (2%). The quartzofeldspathic clasts vary from granitoid plutonic rocks to granitic gneiss that contain either both muscovite and biotite (11%) or a single mica species (5%). Scattered pebbles and cobbles in the pebbly sandstone (site 23) are similar in composition to those in the conglomerate and include Belt Supergroup quartzite and rare two-mica granite clasts or granite clasts with large (>2 cm) muscovite flakes. Detrital zircon distributions for site 23 include abundant Late Cretaceous dates (84%, $n = 75/89$; 82–70 Ma) with one Late Cretaceous outlier (~64 Ma), minor Proterozoic dates (13%, $n = 11/89$; 1778–776 Ma), and rare Oligocene dates (2%, $n = 2/89$; 30–29 Ma; fig. 8K and appendix D).

Roe (2010) describes the sandstone at French Creek (site 22) as including occasional books of biotite but exceptionally rare muscovite (i.e., one grain observed). Roe (2010) also presents 37 detrital zircon dates dominated by Late Cretaceous (81%, $n = 30/37$; 86–64 Ma, peak 72 Ma) and Oligocene (16%, $n = 6/37$; 30–28 Ma, peak 28.9 Ma) grains, as well as a

single Paleoproterozoic grain (3%, $n = 1/37$; 1801 Ma; fig. 8J and appendix D). The Oligocene grains closely correspond to the 29.6 Ma age of an ash-fall tuff that underlies the sandstone (Roe, 2010).

Interpretations: Deep Creek (sites 21 and 23) and French Creek (site 22). Lithologic and paleocurrent properties of the Deep Creek sandstone and conglomerate bodies indicate deposition in a fluvial system (Rust and Koster, 1984; Bridge, 2006; Miall, 1996) with southward paleoflow parallel to the modern basin axis. Clast compositions indicate dominant source areas in the ancestral Anaconda Range with abundant two-mica detritus diagnostic of plutons making up the eastern flank of the modern Anaconda Range (Grice, 2006; Foster and others, 2010; Kalakay and others, 2014; Howlett, 2020). The source of large, euhedral mica flakes is most likely a two-mica granite pegmatite (Grice, 2006), located ~8.5 km to the northwest of site 21, in the headward region of the modern and most likely the corresponding ancestral basin margin drainage. Although Belt Supergroup quartzite and metasedimentary rocks (e.g., the biotite- and muscovite-bearing schist) adjacent to the two-mica source rocks (Grice, 2006) contributed to the axial system, mixing of more durable Belt Supergroup clasts (e.g., quartzite) from distal locales along the axial pathway is also likely. The abundance of two-mica detritus at the Deep Creek sites verifies that two-mica granite was exposed in headwater areas.

The Late Cretaceous detrital zircon age distribution (ca. 76 Ma peak) at site 23, coupled with the presence of two-mica detritus, unequivocally reflects derivation from two-mica granitoid rocks in the adjacent Anaconda Range (ca. 75 Ma peak; fig. 8K). However, there is a paucity of grains younger than ~70 Ma at site 23, despite the abundance of grains <70 Ma in the proposed plutonic source rocks (fig. 8K). Even so, we interpret that two-mica granitoid rocks of the Anaconda Range were primary sources, and later discuss this discrepancy. The Proterozoic detrital zircon distribution is consistent with the presence of Belt Supergroup clasts and recycling from late Paleozoic and Mesozoic strata (Schwartz and others, 2019a). The Oligocene (30–29 Ma) detrital zircon grains at site 23 most likely represent ash-fall input from Oligocene volcanic fields in the Great Basin or ancestral Cascade Mountains, which has also been suggested by Roe (2010) for

similarly aged detrital zircons (30–28 Ma) and the underlying tuff (ca. 29.6 Ma) at the French Creek site (site 22). Moreover, the similarity of the tuff age (~29.6 Ma; Roe, 2010) and maximum depositional ages at site 23 (~28.9 Ma) and site 22 (~29.6 Ma) supports an Arikareean depositional age for all three sites.

Based on a paucity of muscovite at site 22, Roe (2010) also concluded that the Late Cretaceous and Proterozoic detrital zircon dates could reflect source rocks exposed within the Dickie Hills area. Those rocks include muscovite-deficient Late Cretaceous granite and Mesoproterozoic Belt Supergroup quartzites. Zircon age data for a single sample in the Dickie Creek Pluton (80–74 Ma, $n = 36$) partially overlap those of Late Cretaceous detrital zircons at site 22 (fig. 8J; Brennan and others, 2025). We conclude that the ancestral French Creek drainage was similar in location to the modern drainage (i.e., was a paleovalley tributary) and concur with Roe (2010) that nearby uplifts to north and east, within and near the ancestral Dickie Hills, supplied detritus to the eastern side of the Paleogene Deep Creek depocenter. Marked compositional and detrital zircon differences between the Deep Creek and French Creek sites, such as the presence or absence of two-mica detritus, reflect separate western and eastern provenances.

Northern Big Hole Basin

Basin Setting

The Big Hole Basin is a large, elongate (~68 km max by 25 km), north-trending basin in the westernmost part of SWMT that is bound by the Beaverhead Mountains and Anaconda Range along the north to west and the West Pioneer Mountains on the east (fig. 2). Herein, we subdivide the Big Hole Basin into northern and southern sectors; only the northern sector lies within the arbitrarily defined WBR. The northern sector of the Big Hole Basin narrows northeastward from ~22 km wide to ~8 km wide where it meets the Deep Creek Basin and enters the Big Hole River Canyon (<1.5 km wide).

The Anaconda Range and Beaverhead Mountains contain footwall rocks of the ADZ that primarily consist of Paleocene to Eocene two-mica granitic plutons and minor amounts of Mesoproterozoic Belt Supergroup strata (Foster and others 2010; Howlett, 2020; Elliott and others, 2025). The West Pioneer

Mountains contain Late Cretaceous granitic plutons, Late Cretaceous–Paleocene two-mica granitic plutons, Mesoproterozoic Belt Supergroup strata, high-grade Late Cretaceous metamorphic rocks (including granitoid gneiss), and Paleozoic and Cretaceous sedimentary rocks that were metamorphosed in the Late Cretaceous (Elliott and others, 2025).

Longitudinal Fluvial Deposits

Outcrop descriptions and provenance data: *Northern Big Hole Basin* (fig. 3, sites 24 and 25). Two outcrops lie near the western and eastern margins of the northern Big Hole Basin and are located close to branches of the modern Big Hole River (fig. 3, sites 24 and 25). Fossil ages for western site 24 deposits are lacking, whereas Roe (2010) suggests an early Oligocene–early Miocene age (Arikareean; 29–20 Ma) for deposits lying beneath an incisional surface at site 25. Our observations of physical properties of the conglomeratic sandstone and sandstone bodies concur with those of Thomas (1995; site 24) and Roe (2010; site 25), who interpreted the deposits at both sites as fluvial facies. In both cases, multiple meter-scale, upward-fining, lenticular, gravelly coarse- and medium-grained sandstone bodies with erosional bases are interbedded with clay deposits that contain root fossils and casts. Trough (up to 2 m width) and planar cross-stratification are characteristic, although obscure at site 25. Conglomeratic lenses with clasts up to cobble-size are common at site 24, whereas scattered pebbles and granules are present in sandstone at site 25. Paleocurrent data for site 24 document northeastward to southeastward paleoflow, sub-parallel to a major branch of the modern Big Hole River. Paleocurrent data for site 25 were unobtainable (Roe, 2010; this study); however, we note that fluvial sandstone lens elongation is subparallel to the axis of the Big Hole Basin and the adjacent north-flowing trunk channel of the modern Big Hole River.

Sandstone at both locales contains abundant angular feldspar and quartz grains. Abundant, large muscovite and biotite flakes are present in site 24 sandstone; cobbles include abundant two-mica granitoid clasts. In contrast, Roe (2010) reports abundant muscovite at site 25 and an apparent absence of biotite. Although the composition of granules and pebbles in the sandstone deposits was not reported, overlying middle Miocene (Barstovian NALMA, 16.5–11.6 Ma) sand-

stone lenses contain granitic and Mesoproterozoic Belt Supergroup quartzite clasts that were interpreted by Roe (2010) to have a similar provenance as the lower sands.

Detrital zircon data for site 24 (fig. 8L and appendix D) include abundant Paleogene (28%, $n = 63/221$, 64–26 Ma), Late Cretaceous (25%, $n = 55/221$, 95–65 Ma) and Proterozoic–Archean (44%, $n = 97/221$, 3410–550 Ma) dates. The Paleogene grains include Paleocene (3%), Eocene (21%), and Oligocene (4%) dates. Youngest ages in the Oligocene group ($n = 6$, 26.7–26.0 Ma) yield a maximum depositional age of ~26.3 Ma. Rare dates include single Early Cretaceous (107 Ma), Jurassic (193 Ma), Triassic (226 Ma), Permian (265 Ma), Pennsylvanian (312 Ma), and Mississippian (338 Ma) dates, as well as two Devonian (393–391 Ma) dates.

Detrital zircon data for site 25 include abundant Late Cretaceous (53%, $n = 49/93$, 77–64 Ma) and Proterozoic–Archean (43%, $n = 40/93$, 2892–1432) dates (fig. 8M and appendix D). All other grains are Paleogene, including 3% early Eocene ($n = 3/93$, 51–48 Ma) and Oligocene ($n = 1/93$, 27.7 Ma) dates.

Interpretations: Northern Big Hole Basin (sites 24 and 25). The deposits at both sites represent fluvial channels in an alluvial plain, or possibly a distal fan (site 24), setting (Rust and Koster, 1984; Bridge, 2006; Miall, 1996). Maximum depositional ages at both sites (~28–26 Ma), combined with the presence of Arikarean faunas (Roe, 2010) at site 24, collectively indicate an Oligocene and possibly early Miocene depositional age. Coarse-grained deposition at site 24 was proximal to the juncture of the ancestral Beaverhead Mountains and Anaconda Range with paleotransport eastward into the north-central interior of the Big Hole Basin (fig. 7). Finer-grained deposition at site 25 took place in a lower-energy setting, with sand body elongation and sediment provenance evidence (described below) indicating northward paleoflow. The combined evidence from both outcrops suggests convergent transport toward the northeastern corner of the basin and entrance to the ancestral Big Hole Canyon (fig. 7). In contrast, paleoflow, compositional, and detrital zircon data for Renova Formation deposits in the southern Big Hole and Grasshopper Basins (fig. 7; Stroup and others, 2008; Roe, 2010; Barber and others, 2013), indicate fluvial transport from the southern Big Hole

Basin into the Grasshopper Basin, thus serving as evidence for the presence of an ancestral drainage divide within the central part of the Big Hole paleovalley (Portner and others, 2011; Vuke, 2020a; this study).

Site 24 provenance. We interpret the westward-adjacent terrain at the juncture of the ancestral Beaverhead Mountains and Anaconda Range, including the Bitterroot lobe of the Idaho Batholith, as the source region for site 24. Based on regional compilations of detrital zircon data (e.g., Schwartz and others, 2019a), Proterozoic and Archean zircon grains were primarily recycled from Mesoproterozoic strata in the west-adjacent thrust belt. Similarly, some Proterozoic–Archean grains as well as the rare Devonian through Early Cretaceous grains were most likely recycled from Paleozoic to early Mesozoic strata that used to be exposed in the same area. Adjacent exposures of Lowland Creek Volcanics (Elliott and others, 2025) served as source rocks for Eocene grains (fig. 8L).

The youngest Oligocene detrital zircons (27–26 Ma) are of volcanic origin and likely reflect ash-fall processes, such as was interpreted for site 23 in the Deep Creek Basin, with volcanic source areas located in the western Idaho–eastern Oregon region (Norman and others, 1986; Walker and Robinson, 1990; Kauffman and others, 2006). The older Oligocene detrital zircons (38–31 Ma) possibly reflect volcanic ash-fall events emanating from the same source region, but were most likely recycled from older, laterally adjacent Renova Formation deposits. The Late Cretaceous detrital zircon dates (93–61 Ma), in combination with paleocurrent data and abundant detrital muscovite, biotite, and muscovite–biotite-bearing crystalline clasts, document first-cycle derivation from the westward-adjacent two-mica Trail Creek Pluton of the Bitterroot lobe of the Idaho Batholith (95–65 Ma; fig. 8L).

Site 25 provenance. Based upon the proximity of site 25 to the West Pioneer Mountains and detrital zircon data, we interpret the West Pioneer Mountains as the primary source region for the Renova Formation in the northeastern part of the Big Hole Basin. Zircon ages for two-mica plutons in the West Pioneer Mountains closely match Late Cretaceous detrital zircon dates (fig. 8M). In contrast, the age distribution for plutons in the Anaconda Range (fig. 5F) is a poor match with the detrital zircon data, supporting our interpretation of northward fluvial transport toward the

Anaconda Range, rather than southward toward this site and into the southern half of the basin. Although detrital biotite was not reported by Roe (2010) at this locality, the most likely sources for abundant detrital muscovite were two-mica plutons and migmatitic metasediments in the West Pioneer Mountains (e.g., Elliott and Lonn, 2021; Elliott, 2022). The single 52 Ma zircon grain was most likely derived from Lowland Creek Volcanics, whereas the 27.7 Ma grain reflects western Idaho–eastern Oregon volcanism. The Proterozoic–Archean detrital zircon distribution reflects recycling from late Paleozoic and Mesozoic strata (Schwartz and others, 2019a).

Big Hole Canyon

Basin Setting

The Big Hole River crosses the West and East Pioneer Mountains through the Big Hole Canyon, after which it turns south into the Divide Basin. Exposed southwest of the canyon are hanging wall rocks of the ADZ, including Mesoproterozoic Belt and possible Cambrian quartzite, Mississippian–Cretaceous metasedimentary rocks, Late Cretaceous gneiss, and Paleogene granitoids (Elliott and others, 2025). Northwest of the canyon (includes Dickie Hills area) are Mesoproterozoic Belt and possible Cambrian quartzite, Late Cretaceous granitoids, and a minor amount of late Paleozoic–Mesozoic strata (Elliott and others, 2025; McDonald and others, 2012). Northeast and southeast of the canyon are Late Cretaceous granitoids of the Pioneer Batholith and late Paleozoic–Mesozoic strata in the East Pioneer Mountains (McDonald and others, 2012).

Longitudinal Fluvial Deposits

Outcrop descriptions and provenance data: *Big Hole Canyon (site 26).* Boulder–cobble conglomerates along the northern flank of the Big Hole Canyon (fig. 15) overlie a steep erosional surface that truncates Mesoproterozoic Belt Supergroup quartzite and Pennsylvanian and Mesozoic siliciclastic strata. The bright red conglomerate succession is ~10 m thick and marked by an erosional base. It is boulder-dominated (diameters 30–70 cm) at its base but fines upward, with beds characterized by crude stratification, well-developed clast imbrication, and rare trough cross-stratified coarse-sand lenses (fig. 15). Conglomerate clast shapes are varied, ranging from subangular to very

well rounded, but also include angular blocks of Belt Supergroup quartzite up to 115 cm in length, similar in composition to adjacent Belt Supergroup outcrops. Clast imbrication documents southeastward paleoflow, parallel to the modern Big Hole Canyon axis (fig. 7).

The conglomeratic succession is overlain by thin reddish mudstones and matrix-supported breccia units containing angular chert and other siliciclastic rocks that match the composition of Pennsylvanian and Permian strata currently exposed upslope of the outcrop. Boulders and cobbles are dominated by Belt Supergroup clasts, including quartzite (~67%) and metasiltstone (~5%). Other constituents include Pennsylvanian Quadrant Formation quartz arenite (~11%), quartzofeldspathic schist and gneiss (~8%), granite (~4%), tuff (~3 %), and rare lithic sandstone. Most of the granite, gneiss, and schist clasts are two-mica bearing. Interstitial sand within the conglomerate body contains abundant large flakes of muscovite and biotite. Detrital zircon data from a sandstone lens within the conglomerate (fig. 8N and appendix D) include abundant Late Cretaceous grains (55%, $n = 43/78$, 95–70 Ma, peak 75 Ma) and Proterozoic–Archean grains (42%, $n = 33/78$, ~2760–1430 Ma). A single Eocene grain (52 Ma) and two Pennsylvanian–Permian grains (321 and 262 Ma) are also present.

Interpretations: Big Hole Canyon (site 26). Physical properties of the conglomerate indicate deposition in a high-energy fluvial system (Schwartz and Schwartz, 2013). Paleoflow was eastward, through a narrow paleovalley coincident with the modern canyon, with deposition at this site located along the northern paleohillslope margin where fluvial deposits were overlain by debris flow deposits. The reddish color of the conglomerate is because of staining from the overlying red alluvial units, which are analogous to reddish Renova Formation hillslope deposits that unconformably overlie upper Paleozoic bedrock paleovalley margins in the EBR (Schwartz and Schwartz, 2013; Vuke, 2020a).

The abundance of recycled Proterozoic–Archean zircons (42%) is consistent with the presence of abundant Belt Supergroup conglomerate clasts (~72%), the size, shape, and compositional characteristics of which indicate nearby source areas in the West Pioneer Mountains as well as potentially more distant sources. A small proportion of Proterozoic zircons may have

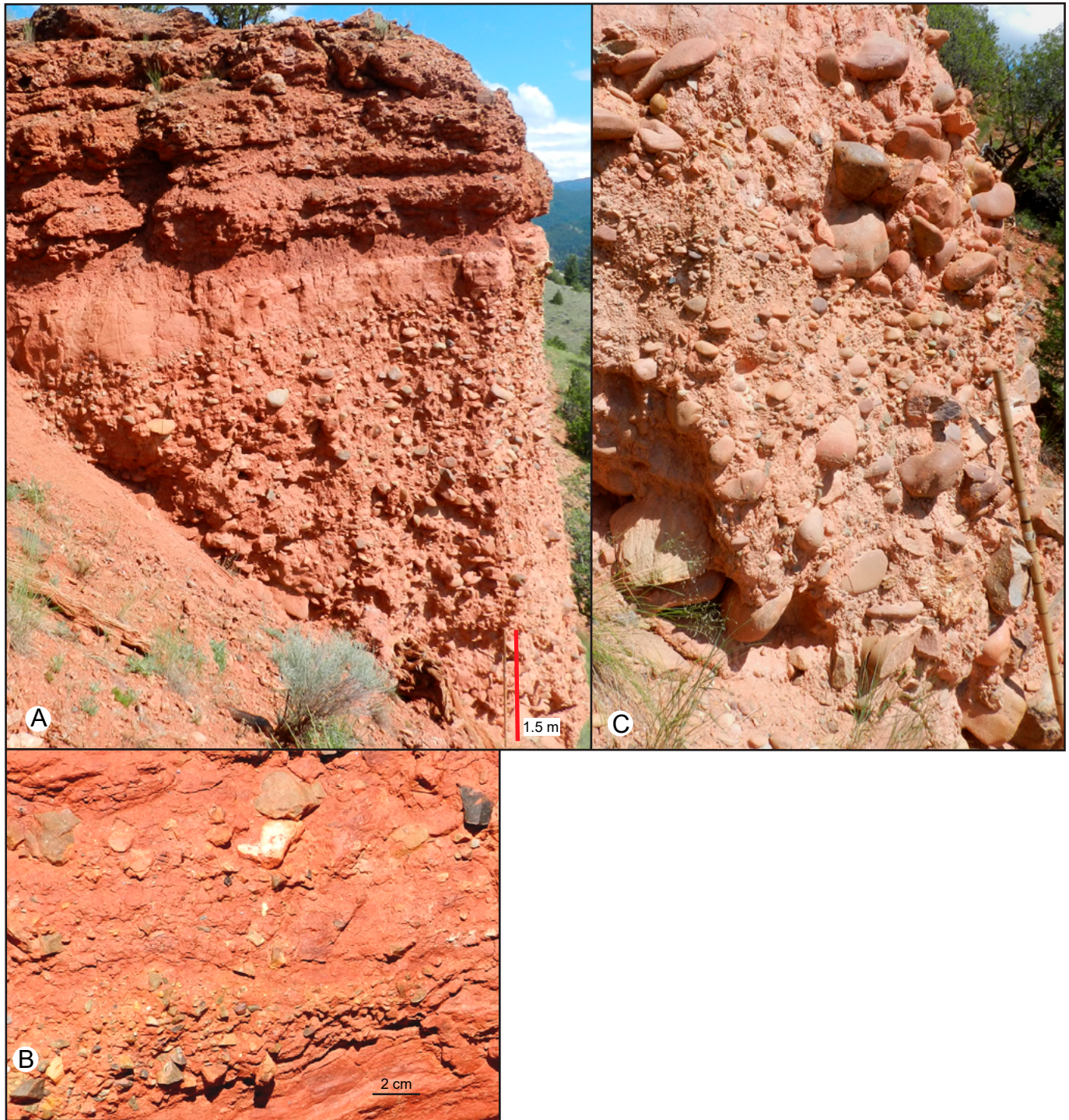


Figure 15. Axial fluvial and basin margin alluvial deposits along the Big Hole paleocanyon (site 26). (A) Upward-fining longitudinal fluvial boulder–cobble conglomerate capped by debris flow and paleosol deposits. Fluvial paleoflow was eastward (into image) compared to debris flow transport from adjacent paleohillside into the basin (left to right). (B) Closeup of matrix-supported debris flow bed in upper part of A containing angular chert and other siliciclastic rocks derived from upslope upper Paleozoic strata. Red coloration of the fluvial conglomerate below is due to staining from the debris flow unit. (C) Boulder conglomerate near base of channel body contains a mix of more distally derived well-rounded Mesoproterozoic Belt Supergroup quartzite boulders and locally derived angular/subangular blocks of Mesoproterozoic Belt Supergroup quartzite and Pennsylvanian Quadrant Formation sandstone. Shown staff length 1.2 m. Photographs by R.K. Schwartz.

been recycled from Cambrian quartzites as well as Pennsylvanian, Permian, and Mesozoic thrust belt strata (Schwartz and others, 2019a) that are relatively abundant along the modern drainage system. Similarly, late Paleozoic detrital zircons were likely recycled from Mesozoic strata. The single 52 Ma zircon grain was most likely derived from the Lowland Creek Volcanics in the southeastern Deep Creek Basin (Elliott, 2017) or westward areas along the Big Hole River (Elliott and others, 2025). A possible Eocene age for the deposit is supported by a detrital zircon MDA of ~52 Ma (this study, fig. 8N and appendix A), which would preclude the addition of ~26–27 Ma zircons such as occur in paleoupstream Renova, albeit Oligocene, deposits in the Deep Creek and northern Big Hole Basins (sites 22–25).

Although quartzofeldspathic igneous and meta-igneous clasts make up a minor proportion (~12%) of conglomerate clasts, the abundance of Late Cretaceous detrital zircon grains (55%) reflects intergranular breakdown of crystalline source rocks into voluminous sand-sized sediment. Sources for the Late Cretaceous detrital zircon group include plutons in the surrounding West Pioneer Mountains, and more remotely, in the Anaconda Range, and uplifts flanking the southern Deer Lodge Basin (fig. 8N). The abundance of muscovite and biotite grains and muscovite–biotite crystalline clasts further document supply from two-mica plutonic and metaplutonic sources in these regions. In addition to two-mica plutons in the north-central Anaconda Range (Grice, 2006; Foster and others, 2010; Kalakay and others, 2014), which are interpreted to have shed abundant detritus into the Deep Creek Basin region, other potential two-mica granite sources within reach of the Big Hole paleodrainage system include the Pintler Creek Pluton (central Anaconda Range; Howlett, 2020; Howlett and others, 2021), Trail Creek Pluton (Bitterroot lobe of the Idaho Batholith in the southernmost Anaconda Range; Elliott and others, 2025), and plutons in the West Pioneer Mountains (Snee, 1982; Marvin and others, 1983; Zen, 1988; Roe, 2010; Elliott and Lonn, 2021; Elliott, 2022).

Similar to the two-mica longitudinal fluvial deposits in the Deep Creek Basin (site 23), pluton-derived grains less than ~70 Ma are absent despite an abundance of younger grains in the proposed plutonic source rocks (fig. 8N). Again, we interpret that two-mica granitoid rocks of the Anaconda Range and West

Pioneer Mountains were primary sources, and later discuss this discrepancy.

Divide Basin

Basin Setting

The Divide Basin is a narrow, north–south-trending basin (~4.8–6.5 x 29 km) flanked by the Highland Mountains along its eastern margin and by the northern part of the East Pioneer Mountains along its western margin. The Continental Divide crosses the central part of the basin, causing divergent drainage northward into the Columbia River watershed through the Deer Lodge Basin, and southward, along Divide Creek, into the Big Hole River of the Missouri River watershed. The Cabbage Patch beds of the Renova Formation are exposed at scattered locations throughout the Divide Basin and lie in depositional contact with the Late Cretaceous Boulder Batholith along its western to northern margin (Vuke, 2004; McDonald and others, 2012). The underlying Climbing Arrow Member of the Renova Formation is locally exposed at the southernmost end of the Divide Basin where it transitions into the Melrose Basin; for this reason, we discuss Climbing Arrow Member deposits in the context of the Melrose Basin in the next section. Based upon conglomerate clast composition and source-rock types in basin-flanking highlands, we discuss the Divide Basin relative to arbitrarily defined northern and southern sectors: the northern sector includes about two-thirds of the basin, bordered by granitoid rocks of the Boulder Batholith as well as by the Lowland Creek volcanic field at its northern end (McDonald and others, 2012; Scarberry and others, 2019). The southern sector is flanked by uplifts containing a mixture of Boulder Batholith rocks and upper Paleozoic to Mesozoic strata of the fold-thrust belt (McDonald and others, 2012).

Basin Margin Deposits

Outcrop descriptions and provenance data: Alluvial fan deposits (sites 34–36). Basin margin deposits of the Cabbage Patch Beds of the Renova Formation are well exposed along the southeastern flank of the Divide Basin and were interpreted by Schwartz and Schwartz (2013) as a mix of debris-, mud-, sheet-flow, and channel deposits making up alluvial fans (fig. 16). Clast imbrication data document westward transport away from the Highland Mountains (fig. 7).

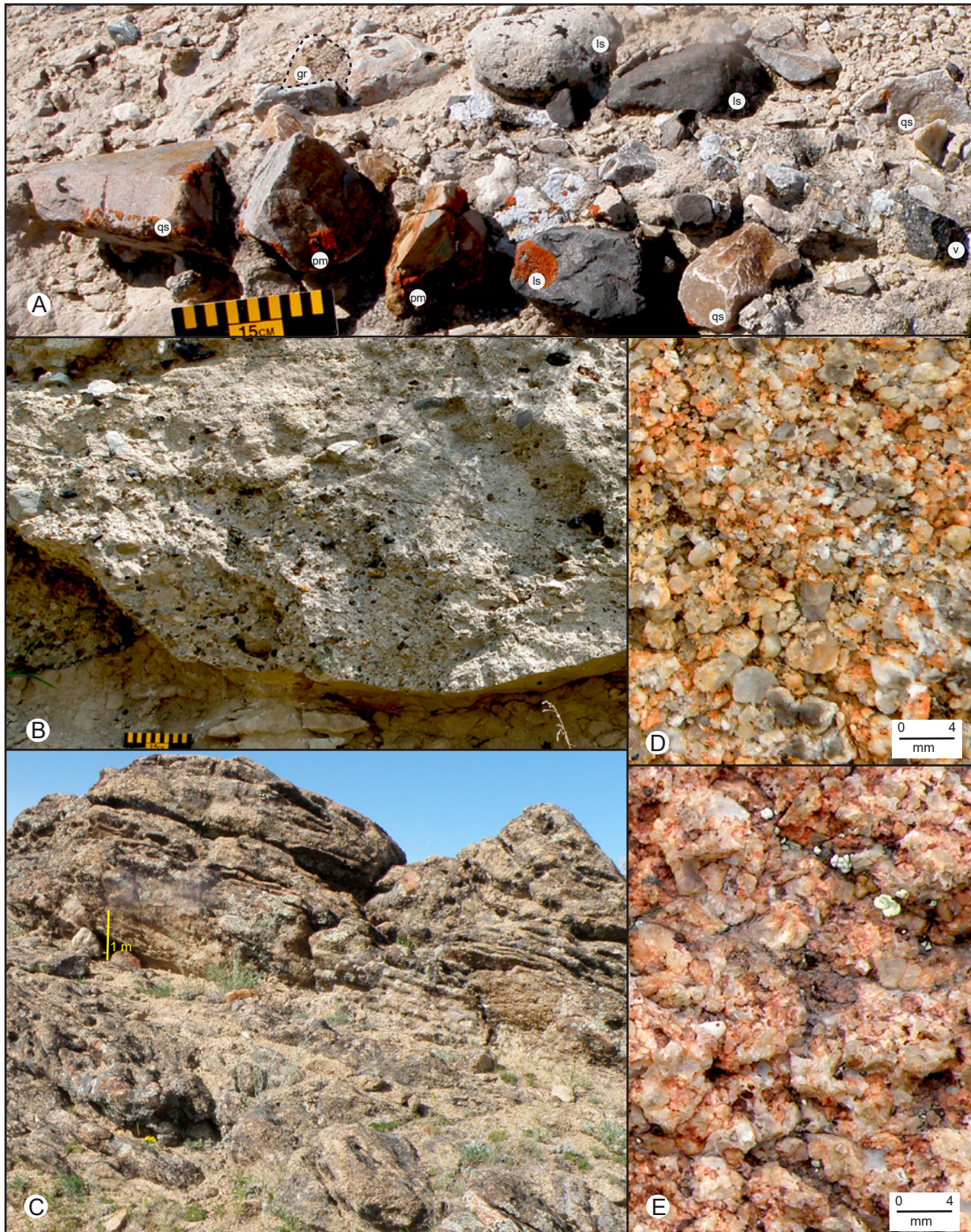


Figure 16. Basin margin (sites 33 and 35) and longitudinal fluvial (site 31) deposits of the Cabbage Patch beds of the Renova Formation in the Divide Basin: (A) Thin lenticular debris flow (site 33) documenting Paleozoic and Mesozoic sedimentary and Cretaceous igneous source rocks in the Boulder Mountains along the eastern side of the basin. Cobbles in A include Paleozoic limestone (ls) and quartz arenite (qs), Paleozoic–Mesozoic siliciclastic rocks (pm), and Cretaceous (Boulder batholith) granite (gr) and Elkhorn Mountains Volcanics (v). (B) Channelized debris flow deposit (site 35). (C) Axial fluvial sandstone body at site 31. Scale 1m. (D) Closeup of fabric of longitudinal fluvial deposits at site 31 showing immature, very coarse-grained to granular, quartz, feldspar, and quartzofeldspathic clasts derived from the nearby Boulder batholith. (E) Grus-like quartzofeldspathic granules and pebbles making up longitudinal fluvial deposits along the western margin of the fluvial system shown in C, at its contact with the Boulder Batholith in the East Pioneer Mountains. Photographs by R.K. Schwartz.

Angular to subrounded cobbles and boulders in debris flow and fan-channel conglomerates typically include Late Cretaceous granite (10–21%), Late Cretaceous and/or Eocene volcanics ($\leq 3\%$), Mesoproterozoic LaHood Formation sedimentary clasts (22–40%), Mississippian limestone (20–40%), and Paleozoic quartz-rich sandstone (9–30%). Clasts of Lower Cretaceous Blackleaf Formation lithic sandstone, siltstone, and porcellanite (vitric tuff, commonly welded) are rare. Granite and limestone clast sizes range up to 2 m.

Interpretations: Alluvial fan deposits (sites 34–36). Paleocurrent data and clast compositions document source areas in the western half of the Highland Mountains. The meter-scale clasts of Late Cretaceous granite and Mississippian carbonate reflect nearby high-relief paleotopography, as do the labile LaHood Formation clasts. Primary source rocks included the Moose Creek Pluton of the Boulder Batholith (fig. 3; McDonald and others, 2012) and its Mesoproterozoic, Paleozoic, and Mesozoic host strata.

Longitudinal Fluvial Deposits

Outcrop descriptions and provenance data: Fluvial deposits (sites 27–31, 33, 37, 38). Eight outcrops of the Cabbage Patch Beds of the Renova Formation along the northern Divide Basin were interpreted by Schwartz and Schwartz (2013) to represent a multi-channel, longitudinal fluvial system that flowed southward along a Divide paleovalley (figs. 7, 16C–16E). The channel-shaped sandstone bodies are laterally and vertically stacked and commonly reach a composite thickness of about 20 m. Here, we describe compositional properties and present detrital zircon data for a northern site (site 28) and a southern site (site 37).

Contrary to most longitudinal fluvial deposits in other basins of the study area, coarse-grained, granule-rich deposits within the Divide Basin typically contain insufficient pebble-sized or larger clasts for quantitative provenance assessment. Thus, we instead pair petrographic data (Schwartz and Schwartz, 2013) with a qualitative assessment of scattered pebbles, cobbles, and boulders. Results of the Gazzi–Dickinson point counting technique (Dickinson and Suczek, 1979) demonstrate that fluvial bodies throughout the basin are arkosic with average QFL proportions of 38% quartz, 55% feldspar, and 7% lithic grains (primarily volcanic and sedimentary; Schwartz and Schwartz,

2013). However, the traditional point counting technique (Folk, 1974; Ingersoll and others, 1984), in which polyminerally, coarsely crystalline grains are counted as lithic fragments, illustrates significant differences between northern and southern locales. These results demonstrate that about 18% of grains in the north are lithic, 62% of which are biotite granite, 25% are volcanic, and 13% are sedimentary (Schwartz and Schwartz, 2013). In contrast, about 28% of grains in the south are granitic, 21% of which contain biotite, 19% contain muscovite, and 2% contain both muscovite and biotite (Schwartz and Schwartz, 2013). Detrital mica content in the northern sandstone is $<0.05\%$ muscovite and about 9% biotite, whereas in the southernmost sandstones (sites 37 and 38), mica content ranges up to about 4% muscovite and 8% biotite.

Gravel-size clasts, although scattered and relatively rare, provide key information. At the northernmost site (site 27), subangular to subrounded felsic–volcanic pebbles and cobbles (up to 7 cm) are relatively abundant compared to smaller amounts of similar sized biotite–granite clasts. Granitic clasts become dominant over volcanic clasts toward the south. At some sites along the western margin of the basin (e.g., sites 30 and 31), the sandstone bodies are in depositional contact with the Boulder Batholith and include high concentrations of grus-like granitic granules and pebbles. Rare, angular to subangular, Paleozoic and Mesozoic sedimentary clasts are present about mid-basin, near sites 30 and 31. Pebbles and cobbles become more abundant, larger, and compositionally diverse at the southern end of the basin with two-mica granitic cobbles and Mesoproterozoic quartzite cobbles and boulders present at site 38.

Detrital zircon data from the northern and southern ends of the Divide basin are also markedly different. In the north (site 28), Late Cretaceous detrital zircon dates between 85 and 65 Ma make up 96% ($n = 86/90$) of the age distribution with a peak value at 78 Ma (fig. 17A). Other dates include one 95 Ma grain, one 139 Ma grain, and two Eocene grains (53 and 49 Ma). Detrital zircons of Proterozoic age are absent. In the south (site 37), data from a two-mica-bearing sandstone (sample CP2 of Link and others, 2008) are characterized by 73% ($n = 43/59$) Late Cretaceous dates with a younger peak value (74.6 Ma) and a much wider range of Cretaceous dates (96–66 Ma) than in

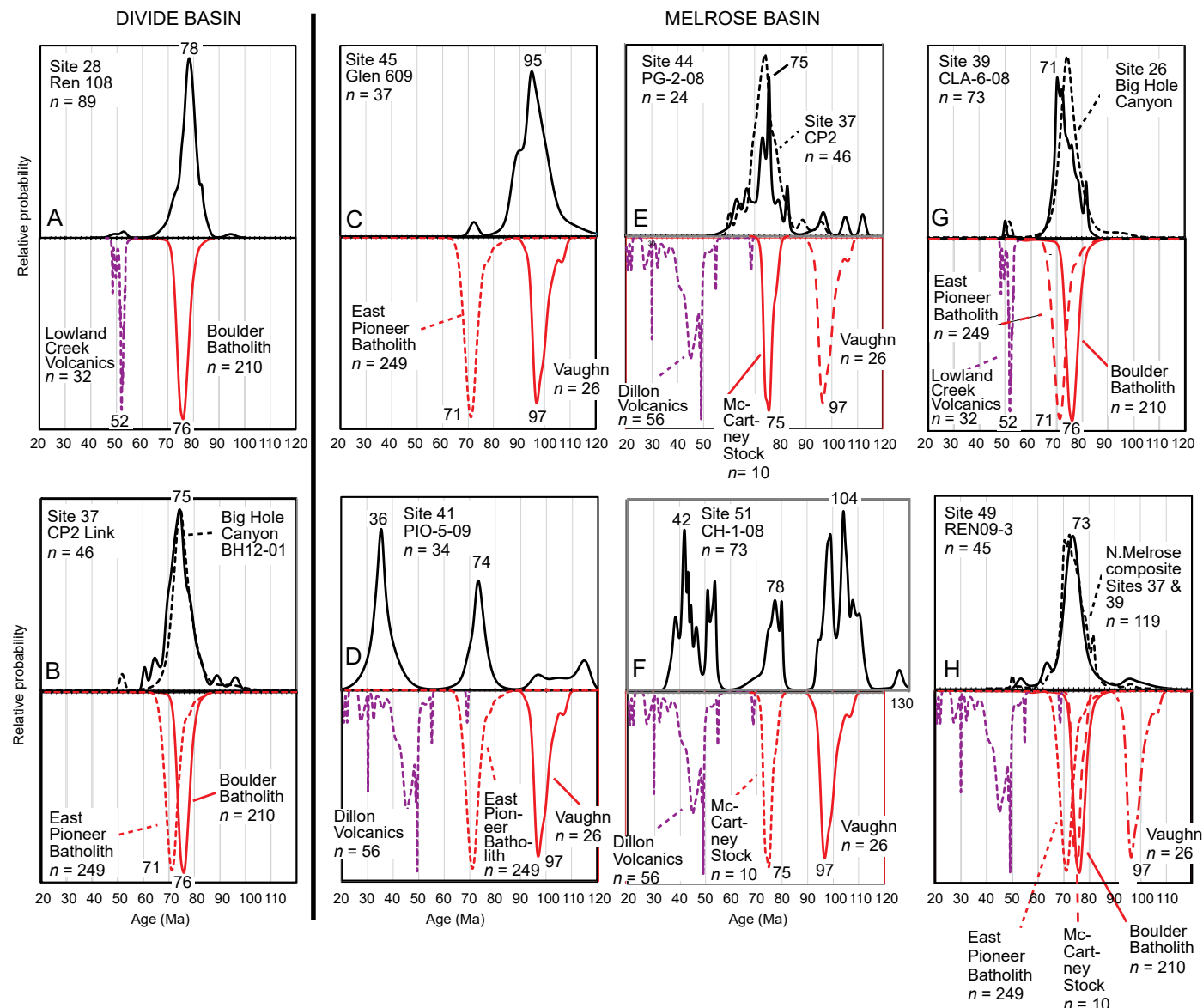


Figure 17. Age frequency and relative probability plots for detrital zircon samples (upper half of diagrams) compared with zircon-age probability plots for potential igneous-source rocks (lower half of diagrams). Dashed black probability plots in upper half of some diagrams show comparisons with other detrital zircon sites, as labelled. Bin width = 2 Myr for all plots. Plots showing the full age range for detrital zircon samples are shown in appendix D. In C–F and H, “Vaughn” refers to the volcanoclastic Vaughn Member of the Blackleaf Formation and “McCartney Stock” refers to the primary pluton making up the core of McCartney Mountain.

the north (85–65 Ma; fig. 17B). Other notable differences include the presence of several grains between 65 and 60 Ma, one Triassic grain (221 Ma), and a minor but significant number of Proterozoic–Archean grains (15%, $n = 9/59$, 2854–745 Ma), several of which are Neoproterozoic ($n = 3$; 993–745 Ma; appendix D).

Interpretations: Fluvial deposits (sites 27–31, 33, 37, 38). The combined detrital zircon, petrographic, and clast composition data for the northern sector of the Divide Basin are consistent with derivation from

rock types currently exposed in upland areas encircling the northern basin. The textural immaturity, relative abundance, and size of volcanic clasts indicate proximity to volcanic source rocks which, supported by the Eocene detrital zircon ages, was most likely rocks of the Lowland Creek volcanic field that were exposed in the northernmost part of the paleodrainage system.

All detrital zircon data from the northern part of the basin support first-cycle derivation from igneous source rocks (Schwartz and others, 2019a). As shown

in figure 17A, the detrital zircon distribution and peak values nearly match those of the Boulder Batholith. The few detrital zircon dates between 65 and 60 Ma are consistent with ages for dikes and sills (66–60 Ma) within the batholith (Lund and others, 2018). The single 139 Ma grain may have been a plutonic xenocryst given that Mesozoic sedimentary rocks, a potential source for recycled zircons (Schwartz and others, 2019a, fig. 3), are not currently present in the source area (McDonald and others, 2012).

A continuous source of granitic detritus from the basin margin along the northern tract, supported by the abundance of *grus*-like grains in Paleogene deposits resting unconformably on batholith rocks (figs. 16D, 16E), resulted in the progressive dilution of volcanic clasts toward the south. The appearance of Paleozoic and Mesozoic cobbles in the middle part of the basin (site 33) is significant in that modern exposures of corresponding fold-thrust belt strata along basin margin drainages are present in the flanking uplifts (McDonald and others, 2012).

Changes in sediment composition and detrital zircon age distributions in the southern basin reflect local mixing of detritus from the northern part of the East Pioneer Mountains and the Big Hole paleocanyon, the confluence of which coincides with sample sites 37 and 38. Notably, sites 37 and 38 contain two-mica granitic debris and clasts of Belt Supergroup rocks. Since two-mica and Belt Supergroup source rocks are absent along the northern Divide Basin, we interpret that those sediments were delivered to the southern Divide Basin along the ancestral Big Hole paleocanyon, a geomorphic feature that we interpret to be persistent through the Cenozoic. The Late Cretaceous detrital zircon grains in the southern Divide Basin are best explained as sourced from the eastward-adjacent Boulder Batholith, westward-adjacent Pioneer Batholith, and igneous rocks that contributed to the Big Hole Basin depositional system. Triassic through Cretaceous strata are the most likely source rocks for the Triassic grain, whereas Neoproterozoic strata or Pennsylvanian through Cretaceous strata are the most likely source rocks for the recycled Neoproterozoic grains (Schwartz and others 2019a,b, fig. 3), all of which are exposed in areas flanking the modern Big Hole Canyon, a short distance to the west (McDonald and others, 2012).

Melrose Basin

Basin Setting

The Melrose Basin extends southward from the Divide Basin as an elongate (~10 km wide x 32 km long) intermontane basin flanked by the East Pioneer Mountains along its western margin and the Highland Mountains and McCartney Mountain along its eastern margin (fig. 2). The modern Big Hole River flows southward through the Melrose Basin until it wraps around the southwestern margin of McCartney Mountain and turns eastward, passing through a water gap before entering the Jefferson Basin (figs. 2, 3). An ancestral version of the Big Hole River is interpreted to have flowed along the same general trackway (Schwartz and Schwartz, 2013). Lithologic assemblages in the bounding highlands of the Melrose Basin are considerably more complex than those bounding other basins in the study area, with units including Paleogene volcanic rocks, Late Cretaceous plutons, Proterozoic through Mesozoic fold-thrust belt strata, as well as Archean metamorphic basement rocks exposed in the Highland Mountains. Outcrops described herein include deposits of the Climbing Arrow Member of the Renova Formation, as well as others without NALMA designations (fig. 1) that are mapped as Renova Formation-equivalent. Outcrops of the Renova Formation and equivalents in the Melrose Basin unconformably overlie Mesozoic and Paleozoic rocks (Vuke, 2004). We integrate new and published data for basin margin deposits exposed along the western and eastern margins of the basin, as well as for longitudinal fluvial deposits.

Basin Margin—Western Side

Outcrop descriptions and provenance data: *Alluvial fan deposits (site 45).* Coarse-grained deposits at this site are interpreted by Schwartz and Schwartz (2013, site BH-1) to represent alluvial fan deposits consisting of eastward-directed debris flow, sheet flow, and fan channel facies (figs. 18A–18C). The clasts are almost entirely sedimentary and consist of 90% Cretaceous siliciclastic rocks, 3% upper Paleozoic quartzite, and 2% upper Paleozoic limestone. The Cretaceous siliciclastic rocks include abundant platy, very angular to subrounded cobbles of greenish gray and brown, lithic and tuffaceous sandstone and siltstone, and light gray, indurated, crystal-vitric tuff and porcellanite. Boulders are particularly abundant in the debris flow deposits and commonly have very angular shapes with

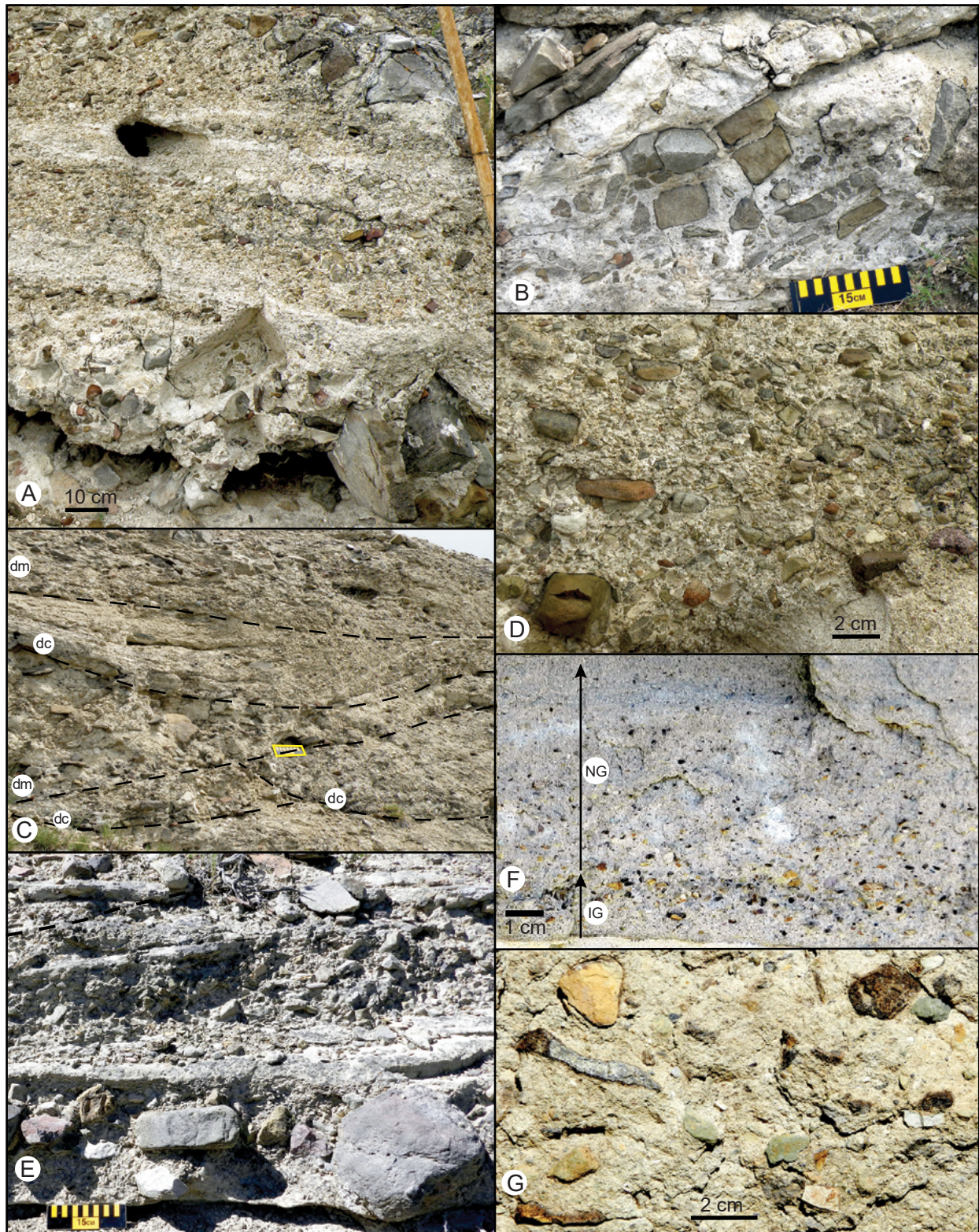


Figure 18. Basin margin deposits along the western side of the Melrose Basin. Site 45: (A) Massive debris flow unit overlain by bedded sheet flow deposits containing imbricated clasts and local inverse grading. (B) Closeup of the basal debris flow unit at an adjacent location showing matrix-supported and imbricated Cretaceous sedimentary clasts. (C) Bedded debris flow-filled channel (dc) and wider spread massive debris flow (dm) deposits. Scale 15 cm. (D) Closeup of channel fill showing better sorting and rounding of sedimentary debris. Site 42: (E) Upward-fining channel-fill succession with abundant Paleogene aphanite boulders in basal unit and mixed amounts of aphanite and Cretaceous sedimentary clasts in overlying units. (F) Inverse (IG) and normal grading (NG) in tabular sheet-flow deposit. (G) Matrix-supported debris flow deposits containing locally derived Cretaceous sedimentary clasts. Photographs by R.K. Schwartz.

well-preserved bedding plane and joint-surface margins. Lithic sandstone blocks up to 35 cm in length are present. Other clast types include about 3% subrounded to rounded basalt cobbles and 2% vein quartz.

Detrital zircon data show almost entirely Cretaceous (59%, $n = 37/63$, 112–72 Ma) and Proterozoic–Archean (36%, $n = 23/63$, 2736–1093) dates, two Jurassic (163, 165 Ma) dates, and a single Devonian (377 Ma) date. The Cretaceous distribution consists of a single Late Cretaceous date (72 Ma) and a transitional Early–Late Cretaceous distribution (57%, $n = 36/63$, 112–85 Ma; fig. 17C and appendix D).

Interpretations: Alluvial fan deposits on paleopediment (site 45). Conglomerate composition, clast size and shape, and extremely poor sorting, in combination with paleocurrent data, document local derivation and eastward transport away from the East Pioneer Mountains toward the Melrose Basin. Diagnostic clasts include lithic sandstone from the Lower Cretaceous Kootenai, Blackleaf, and Frontier Formations, quartzite from the Pennsylvanian Quadrant Formation, and limestone from the Mississippian Madison Group. In particular, the abundance of Lower Cretaceous sedimentary clasts, especially tuff, tuffaceous sandstone, porcellanite, and varicolored (siliceous) fine-grained clasts, are diagnostic of derivation from the Vaughn Member of the Blackleaf Formation and the Frontier Formation. Source rocks for the basaltic clasts are not currently present in the interpreted source area, suggesting post-Paleogene erosion of that source.

Although the eastern Pioneer Batholith (~6 km away) was exposed during deposition of the Renova Formation (Rothfuss and others, 2012; Schwartz and Schwartz, 2013) and alluvial transport was eastward, away from the batholith, the Cretaceous detrital zircon age distribution is almost entirely discrepant with that of the batholith (fig. 17C). Only one Late Cretaceous detrital zircon date (72 Ma) falls within the age range of the Uphill Creek Pluton ($n = 21$; 76–68 Ma; Marvin and others, 1983; Foster and others, 2012) that is exposed immediately to the west. Five other igneous ages (99–88 Ma; Snee, 1982; Marvin and others, 1983) for the eastern Pioneer Batholith overlap the detrital zircon age distribution, but they characterize plutonic rocks in the northern part of the East Pioneer Mountains and represent an implausible distant source. Thus, local paleotopography and paleodrainage must

have precluded significant contribution of sediment from the interior of the ancestral Pioneer highlands.

Instead, the transitional Early–Late Cretaceous group (~112–85 Ma) of detrital zircons represents recycling from tuffaceous strata in the adjacent Blackleaf and Frontier Formations. Dates for a small set of volcanogenic zircons in porcellanite beds of the Vaughn Member of the Blackleaf Formation in the eastern side of the East Pioneer Mountains (127–93 Ma, $n = 19$; Zartman and others, 1995) largely match the Cretaceous age distribution in figure 17C. Similar detrital zircon dates (117–94 Ma, $n = 58$) are reported for tuffaceous beds in the Vaughn Member near Drummond, MT (Link and others, 2008; Zartman and others, 1995), which likely emanated from eruptive centers associated with the Sierra Nevada or Idaho Batholith (e.g., Schwartz and others, 2019a). The Proterozoic–Archean detrital zircon dates, as well as the Jurassic and Devonian dates, are also due to recycling from Kootenai, Blackleaf, and Frontier Formation strata (Zartman and others, 1995; Link and others, 2008; Schwartz and others, 2019a; Rosenblume and others, 2021, 2022). Overall, the data for this site indicate an eastward paleodispersal pattern with local sediment sources largely restricted to Paleozoic and Mesozoic thrust belt strata. Although the core of the East Pioneer Mountains (including Cretaceous batholith rocks and Mesoproterozoic strata) was nearby and exposed (Rothfuss and others, 2012; Schwartz and Schwartz, 2013; this study), deposition at this site was atop a paleopediment with Paleozoic and Mesozoic strata upgradient from the depositional site, similar to that of the modern, wherein pediments with local, upgradient source areas are present between incised valleys that reach farther into the range.

Outcrop descriptions and provenance data: Paludal–lacustrine, debris flow, and sheet flow deposits (sites 41 and 42). Two closely spaced sites on the northwestern side of the Melrose Basin lie within a stratigraphic setting similar to site 45 and provide additional provenance and paleogeographic information. At one site (site 41), near the confluence of Trapper Creek and the Big Hole River, sandy, tuffaceous mudstone deposits of the Chadronian (37–34 Ma) Climbing Arrow Member of the Renova Formation (Vuke, 2004) were interpreted by Rothfuss and others (2012) to represent a paludal to lacustrine setting. The other site (site 42), located 7 km southwest of site 41 and

more proximal to the East Pioneer Mountains, consists of coarse-grained volcanic-rich deposits that are interbedded with the Dillon volcanics (Lowland Creek Volcanics of Zen, 1988; Fritz and others, 2007). Fritz and others (2007) suggest dates of 49 to 42 Ma for the volcanic rocks in this region, whereas Zen (1988) provides dates of 27 Ma for nearby volcanic rock and an ash conglomerate. Overall, the ages for volcanic lithologies associated with site 42 indicate correlation to the Renova Formation. We present new sedimentologic data for site 42 and reinterpret detrital zircon data for site 41.

Site 42 deposits include pebble- to boulder-rich breccia and conglomerate and lithic-rich sandstone (figs. 18D–18G). The breccia contains angular to subrounded sedimentary clasts, many with bedding plane and joint-surface margins, and up to 90% angular volcanic rocks, some with columnar-shaped joint margins. The sedimentary clasts include gray, brown, and greenish siltstone and mudstone; lithic and volcanic-rich sandstone; and light green to tan volcanoclastic rocks. The volcanic clasts consist of mafic to intermediate aphanites and scoria. The coarse deposits range from meter-scale beds of massive, matrix- and framework-supported mixtures of cobble–boulder breccia and sandstone to units with better-defined stratigraphic architecture. Some tabular (<50 cm thick), very coarse to granular sandstone units are characterized by amalgamated, well-sorted, thin beds (5 cm thick) showing inverse and normal grading. Other units are channel-shaped (~1–4 m thick) with a scour base, filled by upward-thinning and -fining bedsets, including: (1) massive, matrix-supported, pebble breccia (0.5 m thick) overlain by normally graded, granular breccia-to-sandstone beds (10 to 5 cm thick) with faint internal laminations, and (2) massive to inversely graded boulder and cobble breccia (~20 up to 5 cm thick) overlain by well-bedded deposits with imbricated clasts indicating southeast-directed paleoflow.

The deposits at site 41 include bioturbated mudstone containing gastropods, land-mammal fossils, and isolated lenses of siltstone and very fine-grained volcanic-rich sandstone (Rothfuss and others, 2012). Detrital zircon data from Rothfuss and others (2012) exhibit four significant age distributions including Eocene–early Oligocene dates (31%, $n = 17/55$, 42–32 Ma), Late Cretaceous dates (20%, $n = 11/55$; 75–70 Ma), transitional Early–Late Cretaceous dates

(13%, $n = 7/55$, 132–96 Ma), and Proterozoic–Archean dates (34%, $n = 19/55$, 1069–2811 Ma; fig. 17D and appendix D). A single Permian (281 Ma) date is also present.

Interpretations: Paludal–lacustrine, debris flow, and sheet flow deposits—paleotributary terminus and adjacent paleopediment (sites 41 and 42). Although paleocurrent data are not available for site 41, its location alongside the terminus of Trapper Creek, a modern stream occupying an interpreted paleovalley, and its proximity to the basin center suggests a regional eastward-directed paleogradient with ponding near local base level. In contrast, site 42 deposits, located on the margin of a dissected pediment and closer to the East Pioneer Mountains, document various sediment and fluid gravity flow processes typically associated with locally steep watersheds and flood events. The tabular, disorganized breccia and well-sorted, graded, granular sandstone beds reflect debris flow and sheet-flood processes (Costa, 1988; Blair and McPherson, 1994a,b), respectively. Similarly, vertical successions in the lenticular bodies reflect aggradation within erosional furrows or ephemeral channels on an alluvial surface by debris, hyperconcentrated, and possibly water-rich flows associated with flood events (Costa, 1988; Blair and McPherson, 1994a,b). Upward-thinning and -fining successions indicate a progressive decrease in flow energy associated with channel filling. Overall, site 42 facies, coupled with site location, unconformity relationship, and paleocurrent data, document an upgradient alluvial setting with generally eastward transport across a topographically irregular terrain, most likely a paleopediment.

The site 42 breccias establish nearby source rock compositions. The sedimentary and volcanoclastic clasts match lithologies within Lower through Upper Cretaceous strata of the Kootenai, Blackleaf, and Frontier Formations, all of which are exposed along the eastern flank of the Pioneer Mountains adjacent to sites 41 and 42. The volcanic clasts match lithologies within the Eocene Dillon volcanics (Zen, 1988; Fritz and others, 2007), which are local to sites 41 and 42 and associated with a nearby eruptive center (Zen, 1988).

The detrital zircon grains at site 41 were derived from the same-age source rocks as interpreted for site 42 as well as additional, westward-located lithologies.

Eocene–early Oligocene dates (42–32 Ma) represent detritus from the Dillon volcanics (fig. 17D). Although there are few interpreted ages for volcanic rocks in the immediate area ($n = 7$; 50–46 and 27 Ma; Marvin and others, 1983; Zen, 1988; Fritz and others 2007), the range of detrital zircon dates generally matches for the age of the lower and middle Dillon volcanics in SWMT, with the middle Dillon volcanics (34–27 Ma) being contemporaneous with Renova Formation deposition (Fritz and others, 2007). Similar to the detrital zircon data that characterize site 45, detrital zircon dates >90 Ma represent recycling from fold-thrust belt sedimentary strata, in this case exposures westward along the Trapper Creek watershed. Also similar to site 45, the Early–Late Cretaceous transitional dates (132–96 Ma) are consistent with derivation from Cretaceous strata, including the Vaughn Member of the Blackleaf Formation (127–93 Ma). The Permian grain was most likely recycled from Cretaceous strata or exposures of the Triassic Dinwoody Formation (Schwartz and others, 2019a). Proterozoic and Archean detrital zircons are recycled from Mesoproterozoic Belt Supergroup, Paleozoic, and Mesozoic strata (Schwartz and others, 2019a). In contrast to the detrital zircon record for site 45, the abundance of Late Cretaceous zircons (75–70 Ma) at site 41 document derivation from the Pioneer Batholith (fig. 17D). The narrow range of Late Cretaceous dates matches that of granitoid rocks in the headwater region of Trapper Creek (80–70 Ma, avg. 74 Ma, $n = 10$; Marvin and others, 1982, 1983; Snee, 1982; Zen, 1988; Fritz and others, 2007), further indicating that ancestral Trapper Creek extended into the interior of the ancestral East Pioneer Mountains.

Basin Margin Deposits (Eastern Side)

Two outcrops of basin margin deposits (sites 44 and 51) and a combined longitudinal fluvial to basin margin succession (site 43) are located on the south-eastern side of the Melrose Basin near McCartney Mountain (fig. 19).

Outcrop descriptions and provenance data: Debris flow deposits at McCartney Mountain (sites 44 and 51). The deposits at sites 44 and 51 are presented in a collective manner due to their similar settings along the perimeter of McCartney Mountain, similar facies types, and similar provenance relationships (fig. 19). Both are considered to be Renova Formation-equivalent as indicated by a Paleocene maximum deposition-

al age of 62 Ma for site 44 and a late Eocene maximum depositional age of 37 Ma for site 51, consistent with the age of Chadronian strata underlying site 44 (ca. 37–36 Ma; Tabrum and others, 1996; McHugh, 2003). Lenses of cobble- to boulder-sized conglomerate at both sites have been interpreted as debris flow deposits with clast imbrication, suggesting westward (site 44) and south–southwestward (site 51) transport, away from McCartney Mountain (fig. 7; Schwartz and Schwartz, 2013). The strata at site 51 are laterally confined by an older Paleogene andesite sill and deformed Mesozoic strata.

At site 44 (fig. 20A), sedimentary lithic clasts are most abundant and consist of 25–30% angular to subrounded quartz arenite, 20–25% rounded fossiliferous Mississippian limestone, and ~15% angular to subrounded, indurated, very fine-grained sandstone, siltstone, mudstone, and <2% angular porcellanite. Igneous clasts include ≤10% subangular to subrounded dark aphanite and 10–15% subrounded to rounded granite. Detrital zircon data for site 44 show two major age distributions, including Late Cretaceous to Paleocene (24%, $n = 20/85$, ~82–62 Ma) dates and Proterozoic (60%, $n = 50/85$, 2872–614 Ma) dates, as well as several Early–Late Cretaceous (5%, $n = 4/85$, 112–94 Ma), Silurian (5%, $n = 4/85$, 443–423 Ma), and Ordovician (5%, $n = 4/85$, 471–448 Ma) dates (fig. 17E; appendix D).

At site 51, pebble- to boulder-sized clasts are very angular to subangular, commonly with preserved joint surfaces, and occur within matrix-supported to muddy framework-supported conglomerates and breccias (fig. 20B). Most abundant are sedimentary, metasedimentary, and volcanoclastic clasts (75%) consisting of gray, brown, and greenish siliceous mudstone and siltstone, tuffaceous siltstone and sandstone, porcellanite (fig. 20D), and chert-rich sandstone. Igneous clasts include granodiorite–diorite (5%) and dark aphanite (20%; fig. 20E), some of which are boulders with columnar shapes (fig. 20B). Detrital zircon data that characterize site 51 show four age groups including Eocene (27%, $n = 25/93$, 54–37 Ma), Late Cretaceous (13%, $n = 12/93$, 80–69 Ma), transitional Late–Early Cretaceous (39%, 36/93, 129–94 Ma), and Proterozoic–Archean (18%, $n = 17/93$, 2722–543 Ma) dates, as well as single Jurassic (171 Ma), Triassic (250 Ma), and Silurian (421 Ma) dates (fig. 17F and appendix D).

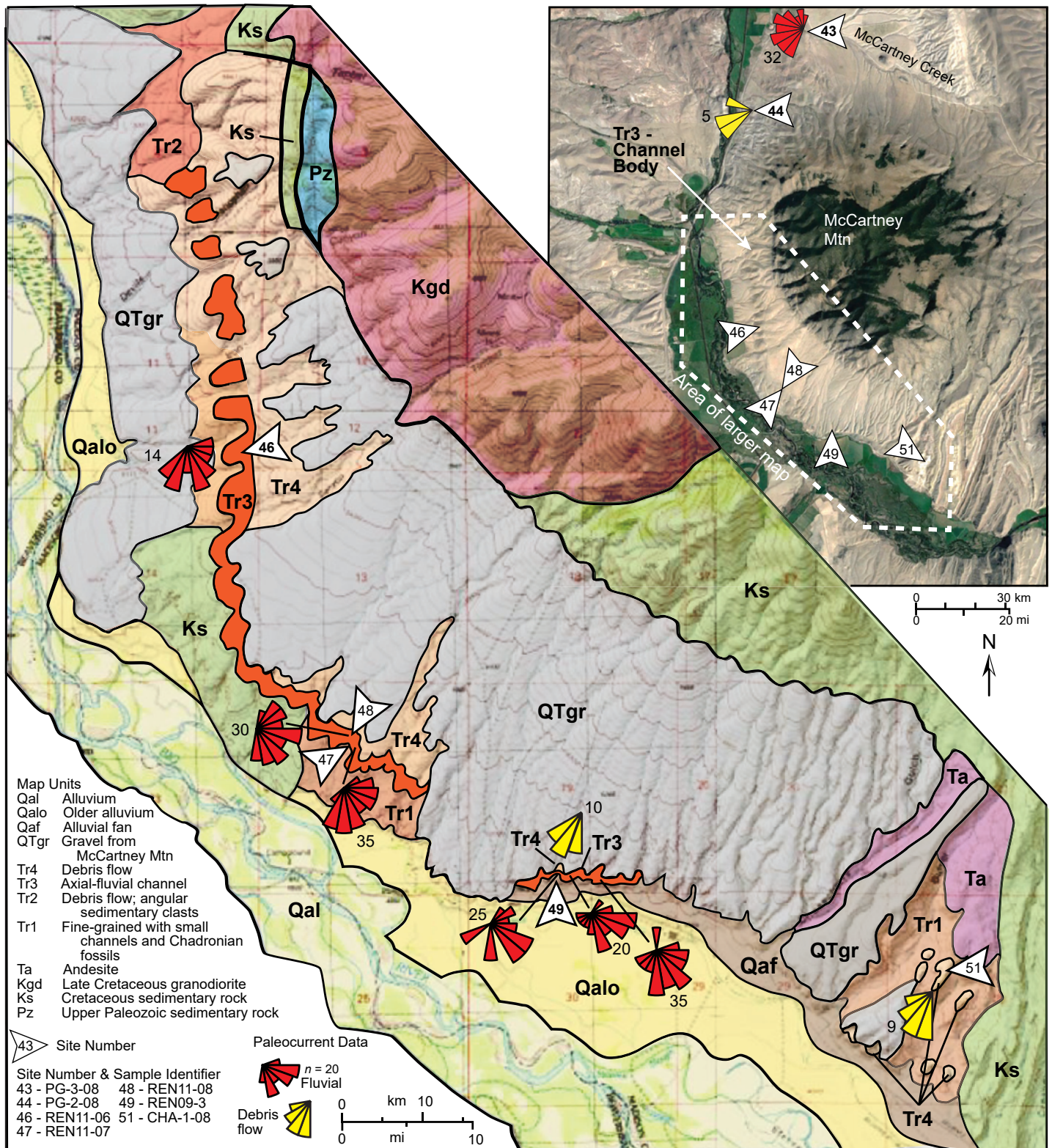


Figure 19. Satellite imagery of the McCartney Mountain area (upper right) and geologic map of the study area along the southeastern margin of the Melrose Basin (satellite imagery from Google Earth, acquired 2024). Designated in the map are informal Renova Formation facies units, Tr1-4. Renova Formation fluvial deposits (Tr3) mark the remnant eastern margin of an axial channel system that abutted the ancestral McCartney uplift as represented by the dark southward zone in the image.



Figure 20. Basin margin deposits along the southeastern side of the Melrose Basin. Site 44: northwestern flank of McCartney Mountain: (A) Debris flow. Boulders and cobbles include Cretaceous lithic sandstone (ss), Mississippian limestone (ls), Cretaceous granite (gr), and Cretaceous mudstone (md). Scale 3 cm. Site 51: southeastern flank of McCartney Mountain: (B) Debris flow dominated by angular and columnar-shaped Paleogene volcanic cobbles and boulders (reddish brown) with minor amounts of Cretaceous granite and Cretaceous sedimentary clasts (gray). (C) Cretaceous granodiorite clast derived from the McCartney Pluton. (D) Porcellanite clast derived from Cretaceous strata. (E) Weathered vesicular aphanite similar to nearby Paleogene volcanic rocks. Photographs by R.K. Schwartz.

Interpretations: Debris flow deposits at McCartney Mountain (sites 44 and 51). The debris flow deposits at sites 44 and 51 document divergent transport away from sediment sources similar to those currently located on the northwestern and southeastern margins of McCartney Mountain. Lateral confinement of site 51 deposits by older strata reflect deposition in a paleovalley tributary setting. Clast types at both sites document exposure of the granitoid core of McCartney Mountain, surrounding siliciclastic and metasiliciclastic rocks of the Cretaceous Blackleaf and Frontier Formations, as well as local Paleogene volcanic rocks. Porcellanite clasts are diagnostic of the Vaughn Member of the Blackleaf Formation and the Frontier Formation. In the northwest (site 44), paleocurrent data and quartzite clasts support derivation from the Pennsylvanian Quadrant Formation, which is currently exposed ~3.5–4 km away, along the western flank of McCartney Mountain. The abundant limestone clasts at site 44 were most likely derived from Mississippian strata, typically found in association with Pennsylvanian strata, although not currently exposed. Similarly, the abundance of Paleogene volcanic clasts and absence of corresponding source rocks at site 44 suggest post-depositional erosion of those sources. In contrast, the abundant volcanic clasts at site 51 are similar to a columnar-jointed aphanitic sill located a short distance away (<0.3 km) in the up-paleocurrent direction.

The detrital zircon data at both sites are compatible with source rocks indicated by conglomerate clast types and paleocurrent data. Both sites contain Late Cretaceous–Paleocene dates, which reflect first-cycle derivation from the McCartney Mountain Pluton, as well as transitional Late–Early Cretaceous dates, which reflect recycling from tuffaceous lithologies in the adjacent Blackleaf Formation (Vaughn Member) and Frontier Formation. Although relatively few dates are available to characterize the McCartney Mountain Pluton ($n = 10$, 78–74 Ma; Snee, 1982; Murphy and others, 2002; Foster and others, 2012), the peak pluton age of 75 Ma matches those of other Late Cretaceous plutons (figs. 17E, 17F). The minor Ordovician and Silurian detrital zircon dates at site 44 reflect recycling from Pennsylvanian–Permian strata (Schwartz and others, 2019a), as supported by the presence of Pennsylvanian Quadrant Formation clasts. Given the absence of Permian–Pennsylvanian strata in the site 51 source area, the single Silurian, Triassic, and Jurassic

dates at that location were most likely derived from Cretaceous strata (Blackleaf and Frontier Formations; Schwartz and others, 2019a) that are locally abundant.

Markedly different at the two sites are the occurrences of Paleocene vs. Eocene detrital zircon age groups. The Paleocene dates at site 44 likely reflect first-cycle derivation from early Paleocene volcanic rocks, supported by the presence of aphanitic volcanic clasts. The absence of Paleocene volcanics in the interpreted source area suggests post-depositional erosion of those source rocks. The Eocene dates (54–37 Ma) at site 51 reflect recycling from adjacent Eocene exposures of tuffaceous deposits similar to the underlying Chadronian unit (37–36 Ma; Tabrum and others, 1996) and possibly first-cycle derivation from the adjacent sill (considered equivalent to 50–46 Ma volcanics in the East Pioneer Mountains; Ruppel and others, 1993). Proterozoic strata are not currently present in the source area and are therefore disregarded as a source for the Proterozoic–Archean detrital zircons. However, Archean to Neoproterozoic zircons are common in late Paleozoic and Cretaceous (Blackleaf and Frontier Formations) strata (Schwartz and others, 2019a; Rosenblume and others, 2022), all of which are present in the area.

Longitudinal Fluvial Deposits

Schwartz and Schwartz (2009 a,b; 2013) and Rothfuss and others (2012) describe a series of outcrops in the Melrose Basin that they interpret as trunk fluvial deposits of the ancestral Big Hole River. Here, we add to existing data and modify provenance and paleogeography interpretations. Two of the sites (39 and 40) are located near the northern end of the Melrose Basin where it abuts the Divide Basin and Big Hole Canyon, and the others (sites 43, 46–49) are in the southeastern part of the Melrose Basin where the Big Hole River turns eastward and exits the basin. In most cases, the outcrops record interbedding of longitudinal fluvial and basin margin deposits.

The fluvial outcrops are relatively widespread, up to 20 m thick, and range from polymictic, framework-supported conglomerate to medium-grained sandstone. Sedimentary structures include planar to low-angle bedding, large-scale trough cross-stratification, and well-developed clast imbrication. Other properties pertinent to our treatment of specific sites are presented below.

Outcrop descriptions and provenance data: Northern Melrose Basin (sites 39 and 40). Sites 39 and 40 are respectively located 3.4 and 6.8 km southeast of longitudinal fluvial site 38 at the mouth of the Big Hole Canyon. They contain fluvial bodies bound by alluvial mudstone and channelized debris flow deposits (fig. 21; Schwartz and Schwartz, 2013). Fluvial paleocurrent data are oriented south–southeast, subparallel to the Melrose Basin axis, whereas paleocurrent data for the debris flow deposits are oriented westward, away from the Boulder Mountains (fig. 7; Schwartz and Schwartz, 2013). Site 39 is a cobble–pebble conglomerate (fig. 21A) and site 40 is primarily a pebbly to conglomeratic, coarse-grained sandstone. In both cases, rare sand-sized lithic grains include two-mica granite. Biotite mica flakes are common, whereas muscovite mica flakes were not observed. Although sandstone at site 40 contains gravel-size clasts of similar composition to site 39, we focus on site 39 because it contains abundant conglomerate clasts and has associated detrital zircon data.

Conglomerate clasts include abundant Mesoproterozoic sedimentary/metasedimentary clasts (68%) and minor Paleozoic sedimentary (14%) and Paleogene igneous (18%) clasts. Most of the Mesoproterozoic clasts are rounded to well-rounded, feldspathic, Belt Supergroup quartzite and metasiltstone cobbles and pebbles (64%), but subangular to subrounded LaHood Formation siliceous mudstone clasts (4%) are also present (figs. 21B, 21C). Recognizable Paleozoic clasts include subangular to subrounded quartz arenite (~10%) and rounded limestone (~4%). Igneous clasts are ~11% Cretaceous granite, ~6% Lowland Creek Volcanics andesite porphyry (fig. 21D), and <1% red scoria and other volcanic clasts of possible Eocene age.

The compositions of clasts in the bounding debris flow units are relevant to those in the fluvial body. Clasts range from pebbles to boulders with subangular to subrounded shapes. Although proportions of clast types vary between exposures, the debris flow deposits at site 39 include 40% Mesoproterozoic LaHood Formation siliceous mudstone, 40% Paleozoic limestone, 9% Paleozoic quartz arenite, 10% Cretaceous granite, and 1% Cretaceous volcanic rocks (figs. 21E–21G).

Detrital zircon data for site 39 include abundant Late Cretaceous (79%, $n = 71/90$, 84–67 Ma), minor Proterozoic (17%, $n = 16/90$, 2484–1444 Ma), and

rare Eocene (2%, $n = 2/90$, 52–50 Ma) and Jurassic (1%, $n = 1/90$, 152 Ma) dates (fig. 17G; appendix D).

Interpretations: Northern Melrose Basin (sites 39 and 40). Some of the clast types in the longitudinal fluvial deposits at sites 39 and 40 demonstrate a compositional link to paleoupstream longitudinal fluvial deposits in the Big Hole Canyon (site 26) and those in the southern end of the Divide Basin (sites 37 and 38), which mark the confluence of the Divide and Big Hole paleobasins. Most diagnostic of the linkage are the abundance of rounded Mesoproterozoic Belt Supergroup quartzite clasts and the presence, although rare, of two-mica granitic sand grains, both of which are signatures of upstream source areas. Derivation of Belt Supergroup quartzite clasts and two-mica sands from source rocks local to the Divide and Melrose Basins is precluded by the absence of corresponding source rocks in bounding uplands. Causes for greatly diminished two-mica clast and muscovite content compared to that of the Big Hole and Deep Creek sites include dilution by sediments from the Divide basin as well as two-mica clast disaggregation.

Sediment contributions from basin margin sources is well supported by clast types within adjacent debris flow deposits. Most diagnostic are the texturally immature Mesoproterozoic LaHood Formation siliceous mudstone and siltstone, Paleozoic quartz arenite, Cretaceous Elkhorn Mountains Volcanics, and Paleozoic limestone clasts, all of which match nearby source rocks in the Boulder Mountains. The source of granitic clasts is more difficult to assess because of mixing from upstream source areas. However, subrounded to subangular granite in the debris flow deposits clearly documents contribution from the adjacent (3–6 km away) Moose Creek Pluton in the western Boulder Mountains.

The detrital zircon data are deceptively simple regarding the diversity of source-rock types and the number of source areas that have contributed to the entire basin network. On a fundamental level, the detrital zircon distribution of site 39 (fig. 17G) could be assumed as consistent with the detrital zircon data for paleoupstream longitudinal fluvial deposits (e.g., sites 26 and 37). However, the Late Cretaceous detrital zircon dates (84–67 Ma) are also similar to ages of plutonic rocks in the Boulder and Pioneer Batholiths, with peaks between 77 and 71 Ma (fig. 17G). The

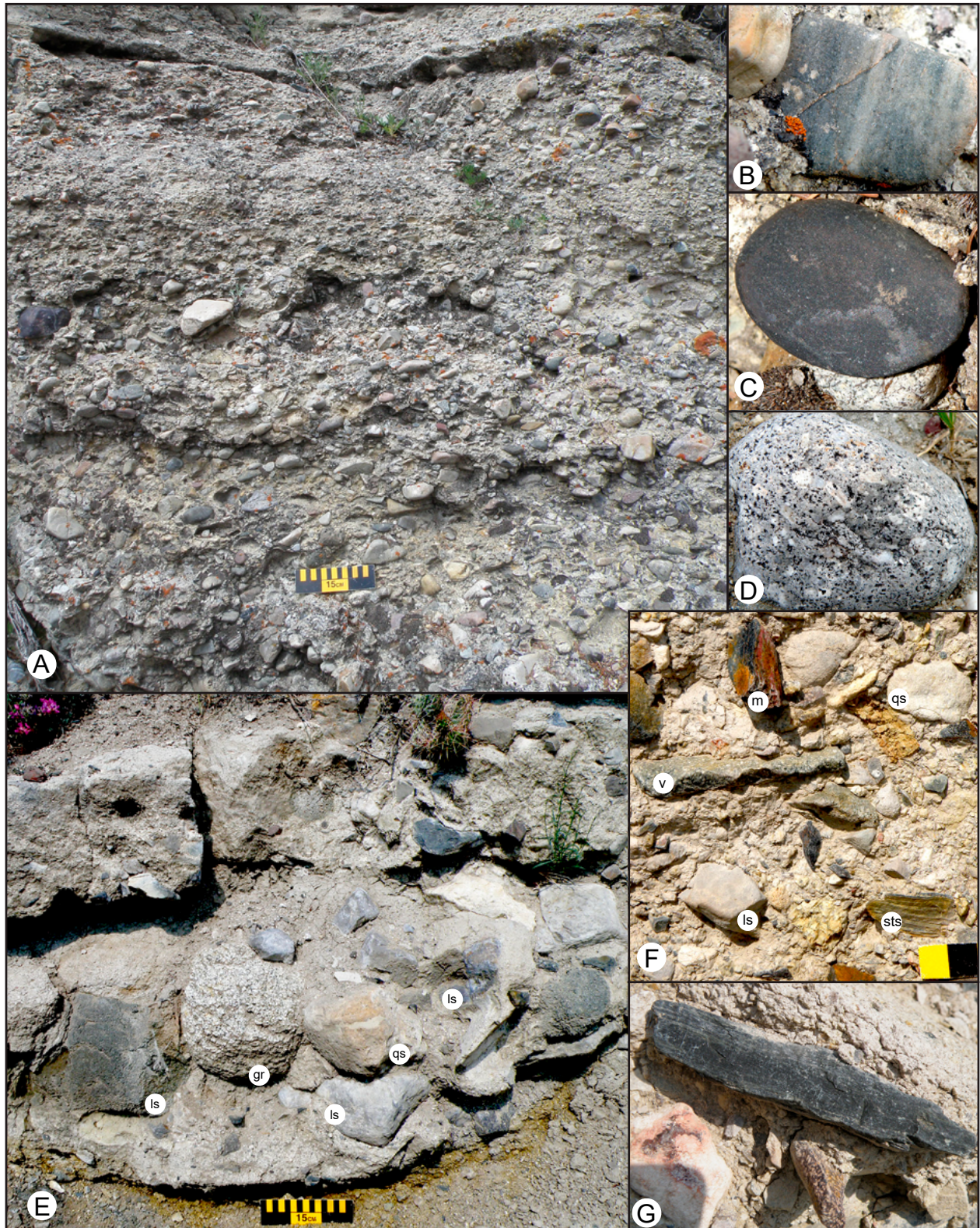


Figure 21. Axial fluvial and basin margin deposits in the northern Melrose Basin. Site 39: (A) Upward-fining cobble to pebble fluvial conglomerate. Diagnostic clasts include Belt Supergroup quartzite (B) as occurs in paleoupstream bodies of the Big Hole River system and Proterozoic LaHood Formation siliceous siltstone (C) and Cretaceous Elkhorn Mountains Volcanics (D) as occurs in adjacent debris flow deposits. (E) Channelized debris flow conglomerate documenting a Boulder Mountains source. Diagnostic clasts include Mississippian limestone (ls), Paleozoic quartz arenite (qs), and Cretaceous granite (gr). (F) Debris flow breccia containing Proterozoic LaHood Formation mudstone (m) and siltstone (sts), Paleozoic limestone (ls) and quartz sandstone (qs), and Cretaceous Elkhorn Mountains Volcanics (v). Scale 2 cm. (G) Closeup of LaHood Formation breccia clast similar in composition to fluvial cobble shown in C. Photographs by R.K. Schwartz.

adjacent Moose Creek Pluton of the Boulder Batholith has been dated at 72 Ma (Wooden and others, 2008; du Bray and others, 2012), also matching the 72 Ma detrital zircon peak. Dates of ca. 83–76 Ma in the detrital zircon sample also likely represent partial derivation from the Elkhorn Mountains Volcanics (Tilling, 1974; Schwartz and others, 2019a, table S1) as supported by the presence of Elkhorn Mountains Volcanics clasts in the conglomerate. The Eocene dates represent first-cycle derivation from igneous rocks, including the Lowland Creek Volcanics that fed both the Divide and Big Hole longitudinal fluvial systems. Recycling from Mesoproterozoic, Paleozoic, and Mesozoic strata in numerous locations along the entire paleodrainage system account for the Proterozoic detrital zircon age group, including recycling from the adjacent Mesoproterozoic LaHood Formation, whereas recycling from Cretaceous strata accounts for the single Jurassic grain (Schwartz and others, 2019a).

Outcrop descriptions and provenance data: Southern Melrose Basin, McCartney Creek (site 43). Schwartz and Schwartz (2013) designated the boulder–cobble conglomerate body at site 43 as Renova Formation-equivalent based on its physical similarity to Chadronian exposures closer to McCartney Mountain (fig. 19). In this study, we present additional details for the conglomerate body and new data for an overlying breccia unit to define the depositional systems and provenance relationships more specifically.

Physical characteristics of the 8- to 10-m-thick conglomerate body include upward-fining, tabular and wedge-shaped bedding, planar-tabular and large-scale trough cross-stratification, and well-developed clast imbrication indicating west–southwest paleoflow (figs. 7, 22A). Lithologies include coarse-grained conglomeratic sandstone, framework-supported conglomerate, and matrix-supported conglomerate with the matrix consisting of mixed mud, sand, and pebbles.

Conglomerate clasts fundamentally occur in two clast-shape groups, angular to subrounded and subrounded to well rounded (fig. 22). The angular–subrounded group includes Mississippian crinoidal limestone (~45%), medium to dark gray fine-grained sandstone and siltstone, some of which appears to be volcanic lithic grain-rich (15–20%), brown sandstone and siltstone (5–10%), pinkish, light gray, and yellowish gray quartz arenite (~5%), and mafic aphanite

(2–5%). In some cases, the limestone clasts make up well-sorted, nearly oligomictic layers with well-developed imbrication. Also present are rare occurrences (1–3%) of angular chert, indurated tuff, and porcellanite. The subrounded to well-rounded group includes feldspathic Mesoproterozoic Belt Supergroup quartzite (up to 15%; 45 cm maximum clast size) and granite (1–3%).

The conglomerate body is transitionally overlain by a several meter thick, pebble to boulder (maximum 20 cm) breccia unit made up of laterally continuous, well-defined, decimeter-scale tabular beds that encase localized, ≤ 1.5 -m-thick lenticular bodies, some of which are distinctly channel shaped (figs. 22B, 22C). The lenticular bodies contain concave scour surfaces that are locally mantled with cobbles and include crude trough cross-stratification and tabular beds. The tabular beds consist of mudstone-matrix-supported breccia to sheets of well-sorted gravel and cobble clasts, some of which exhibit vertically alternating planar-bedded couplets of medium pebble gravel and very coarse pebble and cobble gravel as well as inverse grading. Clast imbrication is common and indicates westward paleoflow (fig. 7).

Depending on location, 85 to >95% of the breccia clasts consist of angular to subangular, commonly platy, medium to dark gray and grayish brown, indurated mudstone, siltstone, and very fine-grained sandstone, some of which appear to be volcanic-rich (fig. 22C). Minor to rare angular and subangular components include greenish porcellanite and indurated tuff, mafic aphanite, lithic sandstone, light gray and light yellowish brown quartzose sandstone, light gray, fine-grained quartzite, and quartz–amphibole–biotite schist (10 cm length). Also rare are subrounded to rounded Cretaceous granodiorite, subrounded crinoidal limestone, and mixed round and rare angular clasts of feldspathic quartzite.

Interpretations: Southern Melrose Basin, McCartney Creek (site 43)—Ancestral confluence of Big Hole River and McCartney Creek. Schwartz and Schwartz (2013) interpreted the conglomerate body to represent a high-energy, braided fluvial, basin margin tributary system (ancestral McCartney Creek) in an east–west-oriented paleovalley located between McCartney Mountain and the southern Boulder Mountains. We herein augment the interpretation of the site to repre-



Figure 22. Axial fluvial and tributary alluvial deposits at confluence of ancestral Big Hole River and McCartney Creek, southeastern Melrose Basin (site 43). Axial fluvial body (A) capped by alluvium (B and C) that was channeled along ancestral McCartney Creek tributary. A scale 15 cm, B staff 1.5 m. (C) Abundant locally derived Cretaceous sedimentary clasts (gray angular) within the debris-filled channel. (D) Fluvial body fabric and composition where angular Cretaceous debris flow detritus (includes dark gray carbonaceous mudstone) from ancestral McCartney Creek was mixed with axial system. (E) Mesoproterozoic Belt Supergroup quartzite boulder within the longitudinal fluvial body. (F, G) Locally derived Mississippian Madison Limestone cobbles and granitic clasts from McCartney Pluton are present in both the longitudinal fluvial body and tributary debris flows. Photographs by R.K. Schwartz.

sent the confluence of ancestral McCartney Creek and the Big Hole River.

Although physical properties of the conglomerate body substantiate a relatively large and energetic fluvial system (Schwartz and Schwartz, 2013), the interbedding of very poorly sorted, matrix-supported conglomerates containing texturally immature clasts similar to those in the overlying breccia unit indicate periodic sediment delivery by hyperconcentrated and mass flows (Costa, 1988). In the overlying breccia unit, the well-sorted gravel couplets in tabular beds and inverse grading are characteristic of alluvial sheet flow deposits (Costa, 1988; Blair and McPherson, 1994a,b). The lenticular bodies reflect small-scale channel incision and filling across a planar alluvial surface. The entire conglomerate-to-breccia facies succession represents mixing between a southwestward-flowing longitudinal fluvial system and a westward-directed flashy basin margin system.

The Belt Supergroup quartzite clasts, up to boulder-sized, are similar to Belt Supergroup quartzite clasts that characterize longitudinal fluvial bodies to the north. The texturally immature clast types in the breccia and conglomerate match lithologies that are currently exposed in the region extending from the northern half of McCartney Mountain to the southern end of the Boulder Mountains (McDonald and others, 2012). The granitic clasts and abundant texturally immature gray mudstone, siltstone, very fine-grained sandstone, porcellanite, and tuff in the conglomerate and breccia reflect proximal sources on the northern side of McCartney Mountain. The texturally immature crinoidal limestone clasts document the proximity of a faulted, relatively high-relief Mississippian Mission Canyon Limestone source represented by remnant outcrops to the east. Similarly, the quartz arenite, brown sandstone/siltstone, and chert clasts represent derivation from Cambrian and Ordovician strata and other Paleozoic lithologies within the same region.

The presence of small feldspathic quartzite clasts within the capping alluvial breccia facies, however, is problematic. Although feldspathic Mesoproterozoic quartzite is locally absent, remnant patches of Belt Supergroup quartzite are present ~18 km northeastward, possibly suggesting post-Paleocene erosion from more proximal areas. Alternatively, and similar to clast mixing at site 49, Belt Supergroup clasts from the under-

lying conglomerate may have mixed into overriding flows of the superposed alluvial system.

Overall, the fluvial mixing model accounts for the conglomerate body having physical properties and rounded quartzite cobble and boulder components similar to those of the other longitudinal fluvial bodies. The texturally immature clasts in the conglomerate and mass-flow conglomerates and breccias represent mixing and eventual progradation of debris that was channeled along ancestral McCartney Creek from sources between the southern Boulder Mountains and McCartney Mountain.

Outcrop descriptions and provenance data: Fluvial deposits along perimeter of McCartney Mountain (sites 46–49). A narrow zone (<0.6 km wide x ~12.5 km long) of coarse-grained sandstone, conglomerate, and localized breccia encircles the northwestern to southern perimeter of McCartney Mountain (fig. 19). The deposits are Renova Formation-equivalent, as indicated by the presence of Chadronian fauna directly below the outcrop belt (Riel, 1963; McHugh, 2003) and an early Eocene detrital zircon maximum depositional age of 53 Ma at one of the sites (49; fig. 17H). Schwartz and Schwartz (2013) interpreted the coexistence of longitudinal fluvial and basin margin facies. Fluvial paleocurrent data are, for the most part, parallel to the curved outcrop belt and therefore parallel to the curved axis of the Melrose Basin (fig. 7). Several sites within the fluvial tract are presented below to specify along-strike compositional properties of the deposits, provide evidence for local sediment sources on McCartney Mountain, and demonstrate compositional linkage with fluvial bodies in the northern part of the Melrose Basin.

At site 46, fluvial bodies consist of distinctly bedded, well-sorted granule and pebble conglomerate and very coarse-grained sandstone (fig. 23A). The pebbles are primarily granitic clasts, whereas the finer components consist of granitic lithic grains and texturally immature feldspar, quartz, and rare volcanic lithic grains. Cobbles of rounded granite and subangular volcanic rock are locally concentrated at the bases of beds. Fluvial bodies are stratigraphically bound by massive beds of unsorted granule and pebble conglomerate and scattered thin lenses of better-sorted granule conglomerate. In these beds, most of the granule and pebble debris is granitic with a *grus*-like appearance. Cobbles

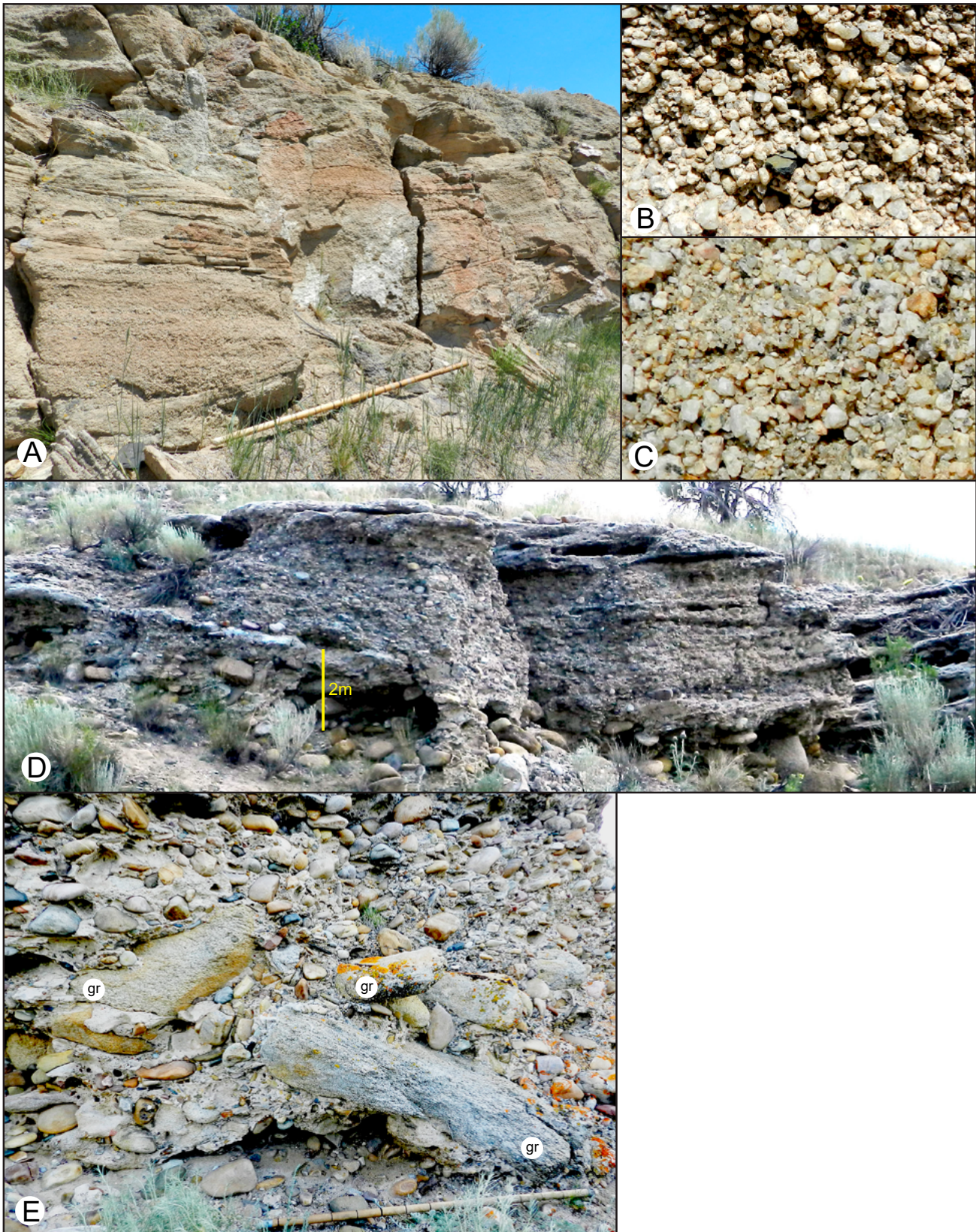


Figure 23. Axial fluvial deposits along perimeter of McCartney Mountain (sites 46 and 47). (A) Granular fluvial body at site 46. Staff 1.5 m. (B) Quartz, feldspar, and quartzofeldspathic granules make up most of the fluvial body. (C) Granular granitic detritus in adjacent debris flow facies (Tr4) that served as a locally dominant source for site 46. (D) Conglomerate body at site 47. Boulder-rich unit at base grades up to cobble and pebble conglomerate. (E) Out-sized granite boulders (gr) in base of D derived from adjacent debris flow deposits. Smaller boulders and cobbles in the conglomerate are predominantly (~70%) Mesoproterozoic Belt Supergroup quartzite. Staff 1.5 m. Photographs by R.K. Schwartz.

and boulders of round granite and subangular volcanic rocks, primarily andesite and quartz porphyry, occur as isolated matrix-supported clasts within the massive beds or in the thin lenses. Overall, grain shapes and the compositions of the massive beds are similar to those of the intervening fluvial bodies (figs. 23B, 23C).

The fluvial deposits at sites 47 and 48 occur about ~2.2 km southeast of site 46 and, although in the same fluvial belt, are remarkably different in stratigraphic architecture and composition. Juxtaposed channel bodies consist of upward-fining, highly polymictic, framework-supported boulder and cobble conglomerates with interspersed thin, medium- to very coarse-grained sandstone lenses (figs. 23D, 23E). Unsorted, massive beds conformably overlie the polymictic fluvial bodies; they contain abundant granite boulders and cobbles within a granular, quartzofeldspathic matrix. The massive deposits, interpreted as debris flows (Schwartz and Schwartz, 2013), commonly contain highly rounded to subangular or subrounded and elongate (0.3 to 1.0 m) boulders with preserved joint surfaces. The fluvial conglomerate contains abundant Mesoproterozoic Belt Supergroup quartzite clasts (70%) with subordinate amounts of Cretaceous siliciclastic rocks (17%), Cretaceous granite (6%) and volcanic rocks (~1%), Paleozoic limestone (4%), and basalt of uncertain age (~2%). The Cretaceous siliciclastic rocks consist of lithic sandstone, siltstone, mudstone, and porcellanite, some of which are metamorphosed. Boulder and cobble sphericity and shape vary with composition. The quartzite clasts and limestone clasts are rounded to well-rounded with high sphericity. Granite clasts range from rounded to subangular with preserved joint surfaces, showing high to low sphericity (blocky). Elongate granite boulders up to 1.5 m long are present along the boulder-rich base of the deposit (fig. 23E). Cretaceous sedimentary and metasedimentary clasts are typically platy and subangular to rounded.

Site 49 is the farthest downstream outcrop in the McCartney fluvial tract. Similar to sites 46 and 47, the deposit consists of an upward-fining, framework-supported boulder and cobble conglomerate with medium- to very coarse-grained, subarkosic sandstone matrix. However, the conglomerate is relatively thin (~1 m) and bracketed by breccia-rich debris flow,

mudflow, sheetflow, and fan-channel deposits shed from McCartney Mountain (figs. 24A, 24B; Schwartz and Schwartz, 2013). Breccia clasts are mixed with the conglomerate in the lowermost and uppermost parts of the fluvial body. The fluvial conglomerate clasts consist of Mesoproterozoic Belt Supergroup feldspathic quartzite (72%), Paleozoic quartz arenite (3%), Paleozoic chert (1%), Cretaceous granite (3%), Cretaceous volcanics (2%), and Cretaceous sedimentary and metasedimentary clasts (19%) including lithic sandstone, black mudstone, and green metasiltstone (fig. 24C). The Mesoproterozoic quartzite, Paleozoic quartz arenite, and Cretaceous granite are rounded to well-rounded, whereas the Cretaceous sedimentary and metasedimentary clasts are typically platy and subangular to subrounded. The breccia clasts are similar in fabric and composition to those in the bounding breccia units as described below. The bounding breccia units primarily consist of angular to subangular Cretaceous sedimentary clasts (fig. 24D). These include lithic sandstone (4%), green mudstone (18%), tuffaceous siltstone/sandstone (13%), and volcanic tuff (61%), of which 11% is porcellanite. Rare clast types include angular to subangular clasts of Paleozoic quartzite and Cretaceous aphanitic rocks, as well as scattered, rounded Mesoproterozoic Belt Supergroup clasts within the base of the overlying breccia unit. Detrital zircon data for site 49 show two major age distributions, Late Cretaceous (41%, $n = 38/93$; 86–70 Ma) and Proterozoic–Archean (50%, $n = 46/93$; 2884–1245 Ma; fig. 17H and appendix D). Minor Eocene (4%, $n = 4/93$; 64–53 Ma), transitional Early–Late Cretaceous (3%, $n = 3/93$; 104–95 Ma), and Devonian grains (2%, $n = 2/93$; 408–377 Ma) also occur.

Interpretations: Fluvial deposits along perimeter of McCartney Mountain (sites 46–49). Overall, we interpret this series of outcrops to represent the eastern margin of a braided channel belt that occupied the Melrose Basin. This is supported by the recurring, conformable association of longitudinal fluvial and basin margin deposits and their abutment to the rocks that compose McCartney Mountain, as well as provenance trends observed along the outcrop belt.

At site 46, we interpret the massive beds enclosing the fluvial body to be debris flow deposits derived from upslope exposures of the Late Cretaceous McCartney Pluton and associated volcanic rocks. We

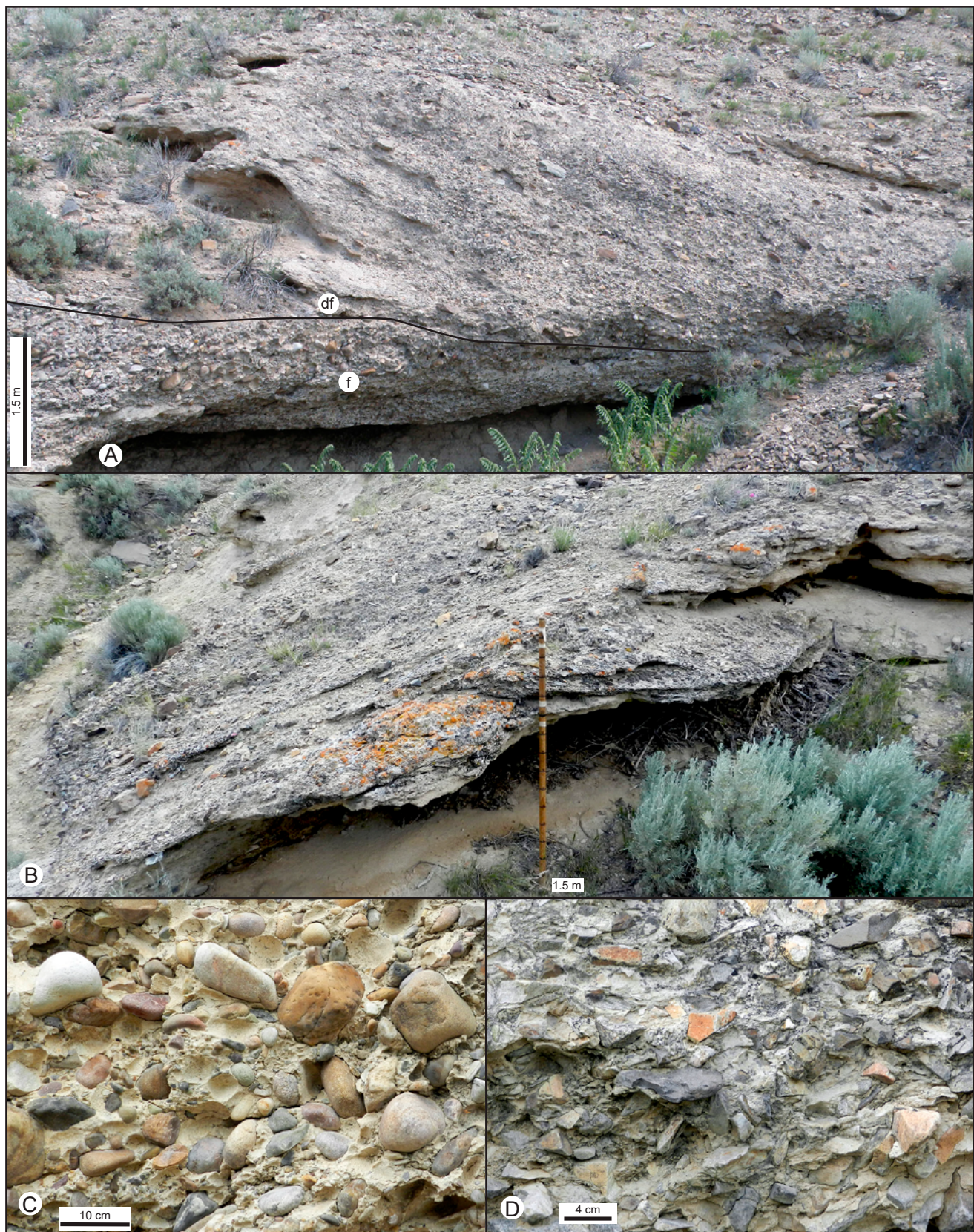


Figure 24. Axial fluvial and debris flow deposits at site 49. (A) Axial fluvial conglomerate (f) overlain by debris flow breccia (df). Breccia of debris flow origin is mixed with fluvial detritus in upper part of the fluvial body. Scale 1.5 m. (B) Clinoform profile of debris flow leading down into fluvial channel body. Staff 1.5 m. (C) Mesoproterozoic Belt Supergroup quartzite makes up ~70% of the fluvial conglomerate at this site. (D) Debris flow breccia primarily includes Cretaceous sedimentary clasts. Photographs by R.K. Schwartz.

interpret the *grus*-like appearance of sand and spheroidal shape of granite cobbles and boulders as linked to chemical–mechanical weathering in the adjacent plutonic source area. The presence of identical clasts in associated fluvial deposits indicates coeval derivation from similar debris flow deposits. Overall, the fluvial bodies at this locale record the influx of locally derived alluvium into an otherwise well-mixed fluvial system characterized by polymictic conglomerates.

The fluvial bodies at sites 47 and 48 also record the local influx of sediment from basin margin sources on McCartney Mountain, at the same time retaining compositional similarity to longitudinal fluvial conglomerates in the Divide Basin. Most diagnostic of linkage with the Divide Basin is the abundance of extremely durable, rounded Belt Supergroup quartzite clasts that do not have any corresponding source areas local to outcrop sites. In contrast, input of locally derived sediment is recorded by clast types that are unique to associated basin margin deposits, which can be matched to bedrock lithologies in the McCartney Mountain area. For example, the labile, well-rounded limestone was likely derived from basin margin sites slightly north of McCartney Mountain, such as those documented for sites 43 and 44. Platy to subangular clasts of Cretaceous sedimentary rocks and porcelanite were also likely derived from basin margin deposits (e.g., sites 43 and 44) that had Blackleaf and Frontier Formation source rocks exposed along the perimeter of McCartney Mountain at the time of deposition. Although some of the Cretaceous granite and volcanic cobbles may have been introduced by the longitudinal fluvial system, the presence of exceptionally large and angular granite boulders supports local derivation from granitic alluvium sourced by the McCartney Pluton.

Similar to adjacent outcrops, longitudinal fluvial deposits at site 49 record fluvial linkage with upstream sites, indicated by the presence of rounded Mesoproterozoic quartzite clasts and Cretaceous granitic clasts. Fluvial transport of more angular Cretaceous sedimentary clasts, similar to those at sites 43 and 44, may account for some observed at site 49, but local input is recorded by the same clast types in bounding debris flow deposits (fig. 23D) that had local source rocks, including the Cretaceous Kootenai, Blackleaf, and Frontier Formations. Detrital zircon data for site

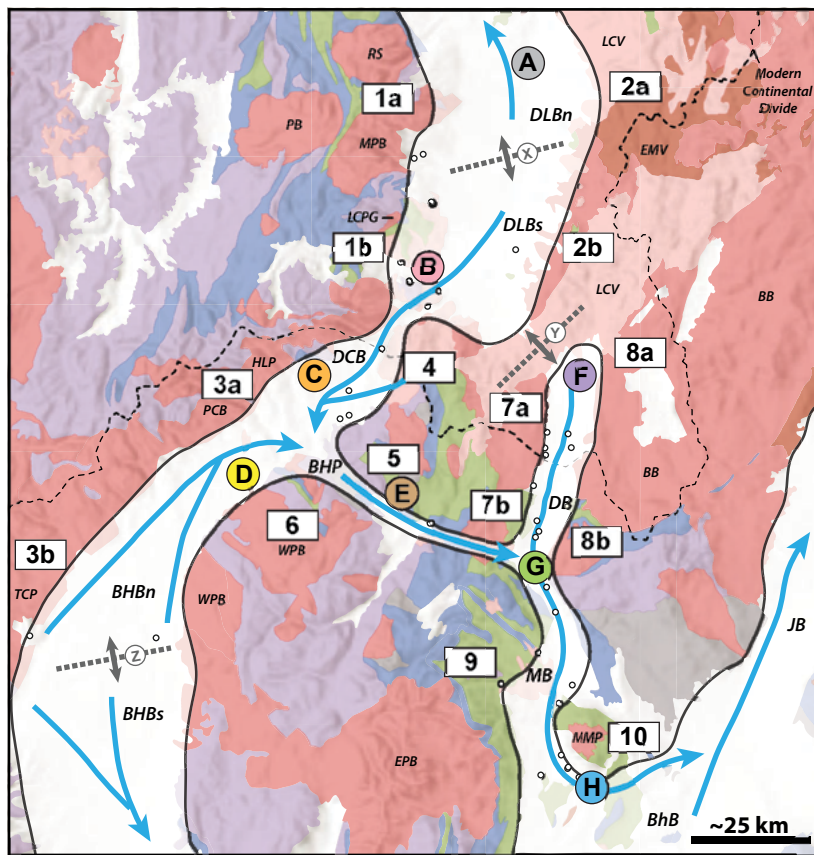
49 help to elucidate specific sources. Eocene (64–53 Ma) and transitional Early–Late Cretaceous (104–95 Ma) grains record derivation from the Dillon volcanics (e.g., in fig. 17F) and the Vaughn Member of the Blackleaf Formation (fig. 17H). Abundant Late Cretaceous grains (86–70 Ma) were derived first-cycle from Late Cretaceous plutons, but identification of exact source(s), other than the McCartney Pluton, is difficult due to overlap between the detrital age distribution and age distributions for alternative igneous sources. Similarity between the detrital zircon age distributions of site 49 and those for upstream deposits (sites 37 and 39; fig. 17H, upper half) suggests plutonic sources located upstream of McCartney Mountain, such as the Boulder and/or eastern Pioneer Batholith (fig. 17H, lower half). However, the presence of locally derived granitic debris along the outcrop belt and overlap in the site 49 age distribution and dates for the McCartney Mountain Pluton (fig. 17H, lower half; Snee, 1982; Murphy and others, 2002; Foster and others, 2012) strongly indicate input by local granitic sources. Abundant Proterozoic–Archean grains correspond to the abundant Mesoproterozoic quartzite clasts observed in the longitudinal fluvial bodies, but may in large part represent recycling from any Mesoproterozoic and younger sedimentary rocks in upstream areas.

DISCUSSION

Paleogene Paleogeography of Southwest Montana

Our findings indicate that the Paleogene landscape in the WBR consisted of elongate, mountainous highland regions separated by paleovalley basins similar in location to the modern intermontane basin network of SWMT (fig. 25) and analogous to coeval intermontane landscapes interpreted for the EBR and SWBR (fig. 7A; Schwartz and Schwartz, 2009a,b, 2013; Rothfuss and others, 2012; Barber and others, 2013). Rugged paleotopography in the study region is consistent with paleoaltimetry studies from both northwestern Montana (Fan and others, 2017, 2021) and SWMT (Kent-Corson and others, 2006; Schwartz and others, 2019b) that indicate the presence of high-elevation (i.e., ≥ 4 km), high-relief (i.e., ≥ 2 km) terrain in westernmost Montana in latest Cretaceous to Eocene time.

In the eastern half of the WBR, the ancestral Boulder, McCartney, and Pioneer highland regions flanked the Divide, Melrose, and Big Hole Canyon

A**Bedrock Features**

BB: Boulder Batholith
EMV: Elkhorn Mountains Volcanics
EPMB: East Pioneer Batholith
HLP: Hearst Lake plutonic suite
LCP: Lost Creek Pluton
LCV: Lowland Creek Volcanics
MMP: McCartney Mountain Pluton
MPB: Mount Powell Batholith
PB: Philipsburg Batholith
PCB: Pintler Creek Pluton
RS: Royal Stock
TCP: Trail Creek Pluton
WPB: West Pioneer Batholith

Paleogene paleovalleys/basins

BhB: Beaverhead Basin
BHBn: Big Hole Basin, north
BHBs: Big Hole Basin, south
BHP: Big Hole paleocanyon
DB: Divide Basin
DCB: Deep Creek Basin
DLBn: Deer Lodge Basin, north
DLBs: Deer Lodge Basin, south
JB: Jefferson Basin
MB: Melrose Basin

1–10 Provenance areas

F

Representative locations in basins for corresponding data sets shown in panels B & C.



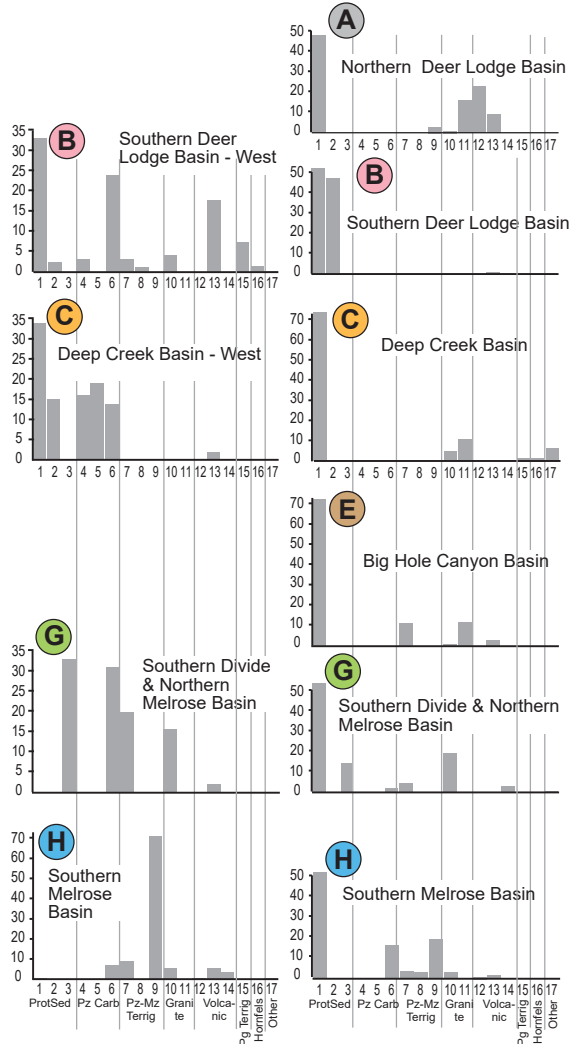
Drainage divide

(X) (Y) (Z)

Figure 25. (A) Paleogeographic map showing Paleogene paleovalleys, paleodrainage networks, and provenance areas superposed on modern bedrock geology, which serves as a close proxy for pre-Paleogene rocks exposed in Paleogene uplands. All paleodrainages other than in the northern Deer Lodge Basin contribute to the ancestral Missouri River system. Provenance areas are defined on the basis of paleocurrent data, clast composition, detrital zircon age, and igneous zircon ages in potential source rocks. (B) Summary of composition of gravel-size clasts in basins along the paleodrainage network. Basin margin deposits (left column) exemplify the variety of clast types provided by adjacent highland areas. Longitudinal fluvial deposits are characterized by a dominance of Proterozoic Belt quartzite clasts throughout the paleodrainage network. (C) Zircon age distributions for plutonic source rocks and longitudinal fluvial deposits along drainage pathway. Circled letters on data plots correspond to circled letter locations on map. (ProtSed, Proterozoic sedimentary rocks; Pz Carb, Paleozoic carbonate rocks; Pz-Mz Terrig, Paleozoic to Mesozoic terrigenous rocks; Pg Terrig, Paleogene terrigenous rocks). See figure 3 for map footprint in regional context.

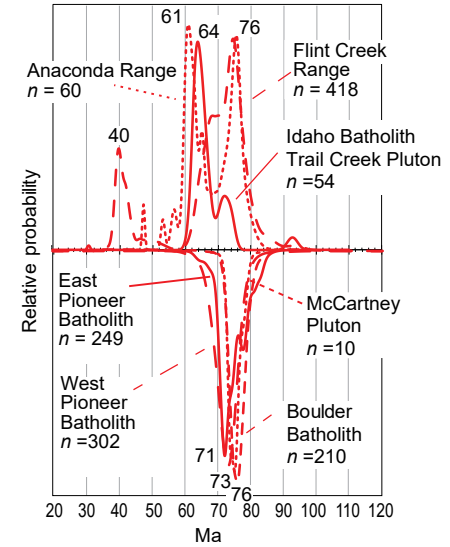
B

Basin Margin Deposits Longitudinal Fluvial Deposits

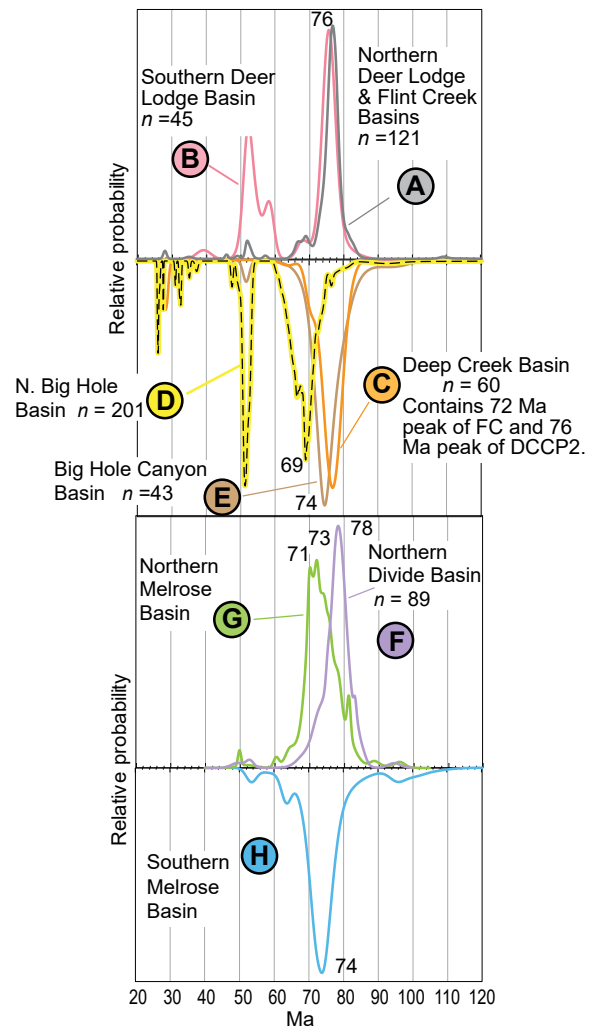


C

Plutonic-Zircon Ages Modern Exposures



Detrital Zircon Ages Renova Axial-Fluvial Deposits



paleovalleys, as initially described by Schwartz and Schwartz (2009a,b; 2013). New results for the western half of the WBR demonstrate a fundamentally similar scenario. Basin margin alluvial facies, including mass flow deposits and channelized distributary bodies, along the western side of the Deer Lodge Basin document the presence of ancestral highlands coincident with the Flint Creek Range that shed sediment into a Deer Lodge paleovalley. Although Paleogene deposits are largely unmapped along the eastern margin of the Deer Lodge Basin, one basin margin site (site 19) and compositional data for some longitudinal fluvial sites (sites 1 and 3) clearly reflect the presence of ancestral highlands coincident with the Boulder Mountains flanking the eastern side of the Deer Lodge paleovalley. The presence of the Boulder Mountains as a highland region throughout the Paleogene was also established by Schwartz and Schwartz (2009a,b; 2013) and Rothfuss and others (2012) in their studies of Paleogene outcrops in the Divide, Melrose, and Jefferson Basins. Although we only report details for a single basin margin site along the western side of the Deep Creek Basin (site 20), combined facies and compositional data (sites 20, 21, and 23) support the presence of a flanking highland region coincident with the Anaconda Range during Renova Formation deposition. Longitudinal fluvial compositional and detrital zircon data from the eastern side of the Deep Creek Basin (site 22) similarly support the presence of highlands coincident with the Dickie Hills region, part of the overall Pioneer highland region, along the eastern margin of the basin. Existence of northern East Pioneer Mountains highlands was also documented by Schwartz and Schwartz (2009a,b; 2013) and Rothfuss and others (2012) in their studies of Paleogene deposits in the Divide and Melrose Basins.

Paleocurrent data from fluvial deposits in central to marginal parts of the Deer Lodge and Deep Creek Basins that are axis-parallel to the modern physiographic basins document the presence of elongate paleovalleys containing longitudinal rivers. These, combined with the presence of thick, fine-grained deposits, including overbank, lacustrine, and swamp (coal) facies, in central parts of the basins (e.g., McLeod, 1987; Rasmussen, 2003; Roe, 2010; Portner and others, 2011) represent basin-center lowlands, similar to those observed in the EBR (Schwartz and Schwartz, 2013). The presence of fluvial deposits associated with basin

margin facies, coupled with basin-directed paleocurrent orientations and detritus matching uplift-interior lithologies (e.g., sites 7, 11, 15, 16, and 20), document the presence of paleovalley tributaries with headwaters in uplands of the ancestral Flint Creek and Anaconda Ranges.

Headwaters of the Ancestral Missouri and Columbia Rivers

Parallelism between Paleogene paleovalleys and headwaters of the modern Missouri River indicates that an ancestral version of the Missouri River drained SWMT as early as the Paleocene (Schwartz and Schwartz, 2009a,b, 2013; summary in Vuke, 2020a). Results from this study allow us to further delineate the paleodrainage system for the Deer Lodge, Flint Creek, and Deep Creek Basin areas relative to that already established for the Big Hole Canyon, Divide, and Melrose Basins (fig. 25; Schwartz and Schwartz, 2013). We acknowledge that discontinuity in exposures and a lack of age control prohibit robust correlations between outcrops and basins, and that all deposits discussed herein are not precisely correlative on the epoch-scale. Further, it is most likely that the interpreted longitudinal fluvial system was not consistently throughgoing during the Paleogene and intermittently contained lentic water bodies and possibly temporary endorheic systems along the basin network. In spite of these limitations, our evidence for persistent geologic and paleogeographic trends throughout the Paleogene supports a time-averaged paleodrainage interpretation.

We interpret longitudinal fluvial paleocurrent data for the Deer Lodge paleovalley to document two divergent drainages in the basin. One system flowed north–northwestward through the northern Deer Lodge paleovalley to paleovalleys coincident with the Flint Creek and Missoula Basins (fig. 25), possibly contributing to the ancestral Columbia River watershed. Available data suggest that the other flowed southward through the southern Deer Lodge paleovalley, a headward part of the ancestral Missouri River watershed. The two drainages were separated by a local, east–west-oriented drainage divide that connected ancestral Flint Creek and Boulder highland regions (fig. 25, location X). The paleodivide was most likely associated with a cross-basin (ENE–WSW oriented) basement-pluton high as indicated by gravity and magnetic data (Hanna and others, 1994; Langhans and others, 2023).

The modern physiography associated with the southern end of the Deer Lodge Basin is characterized by low hills leading into the northern Divide Basin and slightly larger hills leading into the Deep Creek Basin (fig. 25, locations C and F). Because paleoflow interpretations are southward in both the Deep Creek and Divide paleovalleys, we evaluated both as potential continuations for longitudinal drainage from the southern Deer Lodge paleovalley. Eocene boulder conglomerates near the southwest corner of the Deer Lodge Basin (site 12) document a high-energy fluvial system that flowed south–southwest, directly toward the junction with the Deep Creek paleovalley. The composition of Oligocene deposits in the Deep Creek Basin (sites 21 and 23) reflects two-mica granite and Belt Supergroup quartzite sources from the westward-adjacent Anaconda Range and is also consistent with detrital influx from longitudinal fluvial and basin margin deposits of the southern Deer Lodge paleovalley. Outcrops for assessing fluvial paleoflow directions in the southeast corner of the Deer Lodge Basin are seemingly absent. However, there are strong compositional differences between longitudinal fluvial deposits in the Deer Lodge and Divide Basins. For example, Cabbage Patch beds in the Deer Lodge Basin commonly contain Mesoproterozoic Belt Supergroup clasts, Proterozoic detrital zircon grains, and two-mica granite detritus, but similar sediments are absent in Cabbage Patch beds of the northern Divide Basin. These differences support a lack of fluvial connectivity between the southern Deer Lodge and northern Divide paleovalleys. Consequently, we propose a local paleo-drainage divide between the southern Deer Lodge and northern Divide paleovalleys in the area of the Eocene Lowland Creek volcanic field (fig. 25, location Y).

The combination of paleocurrent, compositional, and detrital zircon data indicates that longitudinal fluvial drainage continued southward from the southwestern Deer Lodge paleovalley through a Deep Creek paleovalley, ultimately merging with a northeastward-directed drainage in the northern part of a Big Hole paleovalley (fig. 25, locations C and D). We interpret that the river network continued eastward through a transverse paleovalley or paleocanyon (coincident with the Big Hole Canyon) that cut across the ancestral West and East Pioneer Mountains, becoming confluent with southward-directed drainage in the Divide paleovalley. Continued drainage was

southward through the Melrose paleovalley and then eastward into the ancestral Beaverhead–Jefferson drainage system (Schwartz and Schwartz, 2013). Our interpretation of northeastern flow out of the northern Big Hole paleovalley is consistent with interpretations by Portner and others (2011, their fig. 13B), who suggested a local Paleogene drainage divide within the Big Hole Basin (fig. 25, location Z). Opposing longitudinal fluvial drainage was southward into paleovalleys of the SWBR (Stroup and others, 2008; Barber and others, 2013).

Overall, the regional drainage was a trellis pattern consisting of longitudinal fluvial systems in paleovalleys that were connected by river segments (paleocanyons) that cut across intervening ridges (fig. 7A). The development of trellis drainage systems is typical of fold-thrust belts, in this case similar to the sub-Andean foreland region of northwestern Argentina (Streckler and others, 2011, their fig. 25.5), and most likely reflects initial incision into the Cordilleran fold-thrust belt during Paleocene and possibly latest Cretaceous times.

Sediment Provenance

Geologic Controls on Sediment Composition

Basin margin deposits. The composition of coarse-grained clasts in basin margin deposits varies significantly throughout the study area (e.g., fig. 25B). Compositional diversity resulted from differences in source-rock assemblages for each basin compounded by local paleogeography and related sediment dispersal controls that led to selective, and commonly disproportionate, representation of potential source-rock assemblages. Deposits at different locales along basin margins consist of alluvium from mountain belt perimeters, fluvial sediment from mountain belt interiors, or mixtures of both, such as where locally sourced debris flows are mixed with basin margin paleovalley tributary deposits. Topographically prominent areas may have been dominated by granitic plutons (e.g., adjacent to the northern Divide paleovalley), Mesoproterozoic Belt Supergroup strata and minor amounts of Paleozoic–Mesozoic strata intruded by plutonic rocks (e.g., the western margins of the Deer Lodge and Deep Creek paleovalleys), and/or plutons flanked by abundant Paleozoic to Mesozoic strata (e.g., the northern margin of the Big Hole paleocanyon, western Melrose paleovalley, and southern Divide paleovalley). Other

sources causing strong local diversity in clast types included the Lowland Creek and Dillon volcanic fields.

Axial fluvial deposits. The quartzofeldspathic composition of longitudinal fluvial sandstone is generally consistent throughout the basin network due to the mixing and dilution of heterogeneous basin margin-derived detritus with abundant pluton-derived quartz and feldspar in the longitudinal river systems. In addition to the relative difficulty of determining provenance by means of sandstone composition (Schwartz and Schwartz, 2013), the geographic repetition of many source-rock types, and the similarity of pluton ages along the drainage system compounds the difficulty of provenance analysis for both sand- and gravel-size constituents. However, variability in the composition of basin margin deposits adjacent to longitudinal fluvial pathways provides key insight into provenance relationships. Most apparent are settings where paleogeography exerted strong local control upon longitudinal fluvial composition. Noteworthy examples include locales where the longitudinal fluvial system flowed along, and incorporated sediment from, basin margin bedrock and hillslope colluvium (e.g., the Big Hole paleocanyon, fig. 15, and western margin of the Divide paleovalley, figs. 16C–16E) or reworked the margins of alluvial fan deposits (e.g., the McCartney Mountain area in the eastern Melrose paleovalley; figs. 19, 23, 24). Strong compositional modification also occurred at the confluences of longitudinal fluvial systems (e.g., the juncture of the Big Hole paleocanyon–southern Divide–northern Melrose paleovalleys; fig. 21) or at the confluence of longitudinal fluvial and basin margin tributary systems (e.g., the eastern Melrose paleovalley; fig. 22). In spite of the cumulative increase in number of source areas and the changing proportion of source-rock types along the longitudinal fluvial pathway, the most persistent and abundant gravel constituent observed in longitudinal fluvial deposits was ultradurable Mesoproterozoic Belt Supergroup quartzite (fig. 25B, right side).

Plutons and Detrital Zircon Signature

Depositional facies, paleocurrent, clast composition, and detrital zircon data document that each basin in the study area received first-cycle detritus from Late Cretaceous plutons that were exposed in all of the basin-bounding highlands. At least seven plutonic complexes were exposed along the interconnected

basin network, including in ancestral highlands coincident with the Flint Creek Range, Anaconda Range, Boulder Mountains, northern Beaverhead Mountains, East and West Pioneer Mountains, and McCartney Mountain. Comparisons of zircon-age distributions for each plutonic source area and its corresponding detrital zircon product demonstrates the nature of both igneous and detrital age records and relative accuracy of the detrital zircon signal (figs. 26A–26H).

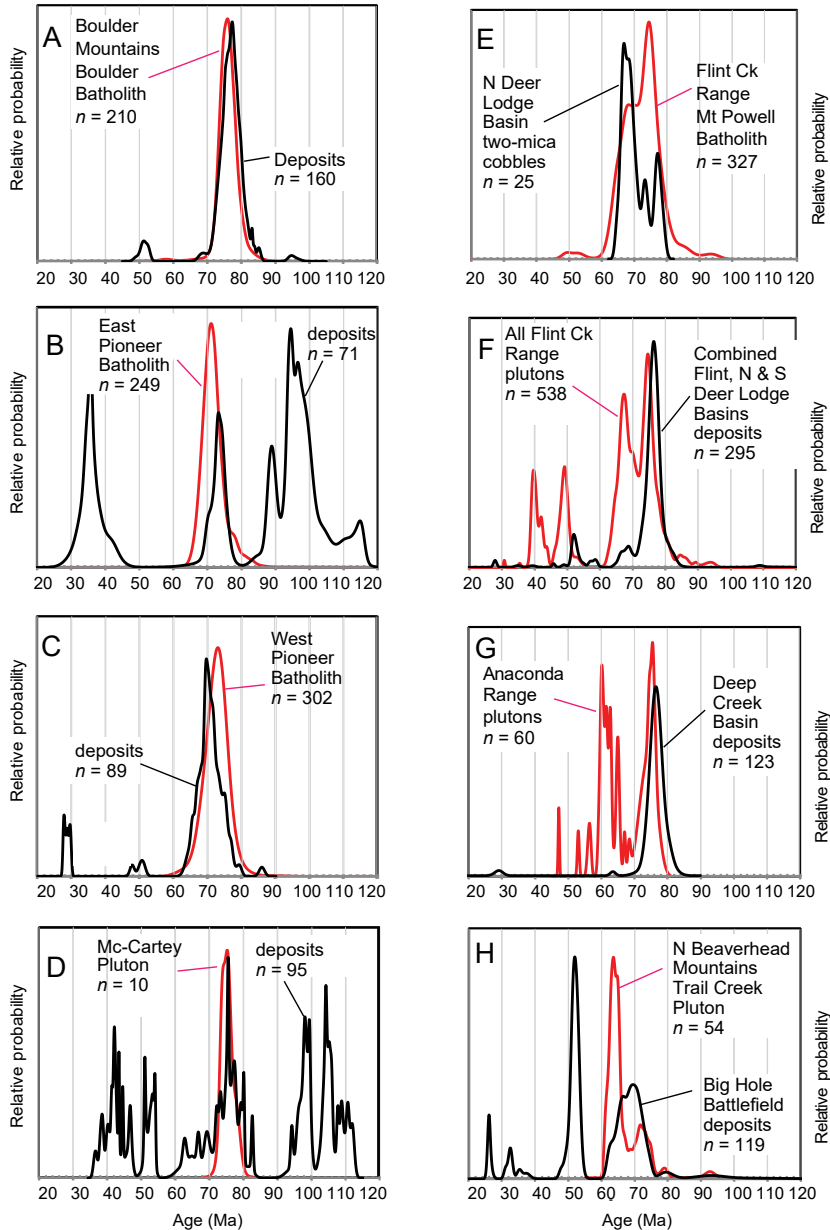
Detrital zircon age distributions for sediments derived from the Boulder Batholith, eastern Pioneer Batholith, and McCartney Mountain Pluton are characterized by unimodal Late Cretaceous peaks that closely match unimodal Late Cretaceous peaks of corresponding igneous source rocks currently exposed in those areas (figs. 26A–26D). In contrast, most igneous zircon age distributions that characterize plutons currently exposed in the Flint Creek Range, Anaconda Range, and northern Beaverhead Mountains are multimodal, with dates ranging from Late Cretaceous to Eocene (figs. 5, 26E–26H). Corresponding detrital zircon age distributions from Paleogene deposits derived from the ancestral Anaconda Range and northern Beaverhead Mountains (figs. 26G, 26H), however, are characterized by unimodal Late Cretaceous peaks. Although the detrital zircon age distribution for plutonic grains derived from the Mount Powell Batholith in the Flint Creek Range is bimodal, both modes are Late Cretaceous and similar to those in the pluton-age distribution (figs. 26E, 26F).

We interpret younger grains (e.g., <60 Ma) in the detrital zircon distributions shown in figure 26 to reflect derivation from local volcanic rocks, which is supported by the presence of volcanic cobbles in many cases. Most grains older than early Late Cretaceous through Archean (i.e., >90 Ma; appendix D) were recycled from Mesoproterozoic to Upper Cretaceous strata that are present throughout the study area (Schwartz and others, 2019a).

Synopsis of Provenance Change Along the Paleodrainage Network

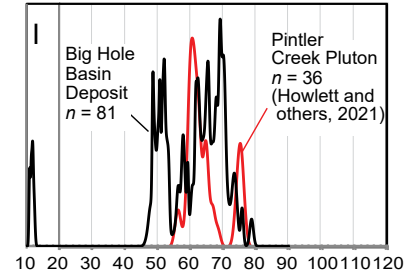
Deer Lodge paleovalley. A local paleodrainage divide between the ancestral Missouri and Columbia River watersheds was located within the Deer Lodge Basin (fig. 25A). Sediment was transported from highlands coincident with the Flint Creek Range,

Plutonic Source Rocks and Renova Age Deposits

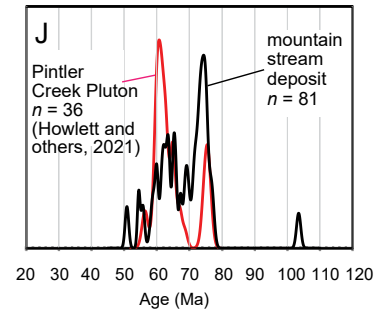


Plutonic Source Rocks and Post-Renova Deposits

Middle-Late Miocene



Modern



Plutonic Source and Linked Detrital Zircon Deposits

- A.** Source: Boulder Batholith
S Deer Lodge Basin—site 19, CE17BN1
N Divide Basin—site 28, Ren-1-08
- B.** Source: East Pioneer Batholith
Melrose Basin—site 41, PIO-5-09
Melrose Basin—site 45, Glen-6-09
- C.** Source: West Pioneer Batholith
Deep Creek Basin—site 22, FC Roe Big Hole Basin—site 25, WNBURoe
- D.** Source: McCartney Mountain Pluton
Melrose Basin—site 44, PG-2-08
Melrose Basin—site 51, CH-1-08

- E.** Source: Flint Creek Range Mt Powell Batholith
N Divide Basin—site 1, CE23CL-01
- F.** Source: Composite Flint Creek Range plutons
N Deer Lodge Basin—site 1, CE23CL-01 & 02
N Deer Lodge Basin—site 2, CMF-67
Flint Creek Basin—sites 4a,b, Portner-Link
S Deer Lodge Basin—site 12, OW12-04
S Deer Lodge Basin—site 18, 17DV04
- G.** Source: Anaconda Range plutons
Deep Creek Basin—site 20, 17GM01
Deep Creek Basin—site 23, DCCP-2

- H.** Source: N. Beaverhead Mtns. Trail Creek Pluton
Big Hole Basin—site 24, CE22BB01 & CE20BHB17
- I.** Source: Pintler Creek Pluton in Anaconda Range
Wisdom West sample (Roe, 2010)
- J.** Source: Pintler Pluton
N Divide Basin—Mussigbrod Creek (Link and others, 2008)

Figure 26. Comparison of zircon ages from plutons (red probability lines) with detrital zircon ages (black lines) for deposits linked to the same plutonic source area. (A–G) Comparisons for Renova deposits. All detrital zircon grains younger than 60 Ma and older than 90 Ma are of volcanic or recycled origin, respectively. (I–J) Comparison for post-Renova Formation deposits (Ck, Creek; N, S, North, South).

northern Anaconda Range, and Boulder Mountains (fig. 25A, areas 1 and 2). Quartzite, carbonate, granite, volcanic, hornfels, siliciclastic sedimentary, and rare metamorphic clasts in basin margin deposits (e.g., fig. 25B, histogram B left) document source rocks to the west that included Mesoproterozoic, Paleozoic, Mesozoic, and Eocene strata, Late Cretaceous granitic plutons, and locally, Paleocene–Eocene volcanic rocks (fig. 25A, areas 1a and 1b). Two-mica plutonic source rocks (Mount Powell Batholith) were present in area 1a and shed abundant detritus into the northern Deer Lodge paleovalley of the ancestral Columbia River watershed. A two-mica pluton in the northern Anaconda Range (southernmost area of 1b) supplied two-mica detritus to the southern Deer Lodge paleovalley of the ancestral Missouri River watershed. Source rocks along the eastern side of the paleovalley (areas 2a and 2b) were less diverse and primarily consisted of Late Cretaceous Boulder Batholith granitoid rocks, Late Cretaceous Elkhorn Mountains Volcanics, and Eocene Lowland Creek Volcanics. Pluton-associated detrital zircon age groups in longitudinal fluvial deposits of the northern and southern Deer Lodge and Flint Creek paleovalleys (peaks ~76 Ma; fig. 25C, probability lines A and B) reflect derivation from plutons exposed in the ancestral Flint Creek Range, northern Anaconda Range, and Boulder Mountains.

Deep Creek paleovalley. The scenario for the Deep Creek paleovalley is fundamentally similar to that of the southern Deer Lodge paleovalley except that source rocks along its eastern margin (fig. 25A, area 4) were those of the ancestral Dickie Hills area, consisting primarily of muscovite-deficient, Late Cretaceous granitic plutons and Belt Supergroup quartzites, as well as Lowland Creek Volcanics. Many lithologies currently exposed in the Anaconda Range (area 3a) served as source rocks during the Paleogene, primarily Late Cretaceous granitic plutons and Mesoproterozoic, Paleozoic, and Mesozoic strata and their metasedimentary equivalents. Plutonic source rocks in the ancestral Anaconda Range included two-mica granitoids (e.g., Hearst Lake and Pintler Creek Plutons) that provided abundant two-mica detritus to the Deep Creek longitudinal fluvial system.

Detrital zircon data from confluent, southward-flowing, longitudinal fluvial systems document different plutonic provenances. An eastern river system

drained the ancestral Dickie Hills area (area 4) and contains Late Cretaceous detrital zircons (72 Ma peak), reflecting derivation from plutons in the Dickie Hills area (73 Ma peak; fig. 25C, probability line C). A centralized river system containing two-mica detritus and a different age group of detrital zircons (76 Ma peak) reflects derivation from plutons in the ancestral Anaconda Range (76 Ma peak) as well as longitudinal fluvial input from the southern Deer Lodge paleovalley (76 Ma peak; fig. 25C, probability lines B and C).

Northern Big Hole paleovalley. The northern Big Hole Basin contains Paleogene deposits that define a paleovalley at the headward reaches of the ancestral Missouri River network. Detritus was derived from highlands coincident with the northern Beaverhead Mountains (area 3b), the southwestern Anaconda Range (area 3a), and the West Pioneer Mountains (area 6). Late Cretaceous granitic plutons, including two-mica granite and Belt Supergroup strata, were present in all of the source areas and fed abundant two-mica detritus and quartzite clasts into the basin.

Although detrital zircons of the 75 Ma age group in the Pintler Creek Pluton (fig. 3, Anaconda Range) were delivered to the northern margin of the basin, plutons in the northern Beaverhead Mountains (Idaho Batholith) and West Pioneer Mountains resulted in a younger composite age (69 Ma peak) for pluton-derived zircons in mixed longitudinal fluvial deposits (fig. 25C, probability line D). Lowland Creek Volcanics along the northwestern margin of the Big Hole Basin (Elliott and others, 2025) were likely a significant source of 55–48 Ma detritus, whereas younger grains reflected minor late Eocene to Oligocene (37–26 Ma) volcanic contributions.

Big Hole paleocanyon. The highlands flanking the northern and southern margins of the Big Hole Canyon (areas 5 and 6), as well as longitudinal fluvial detritus from the Deep Creek and Big Hole Basins, fed the paleovalley that was coincident with the modern Big Hole Canyon (fig. 25A, location E). Source rocks to the north primarily included Late Cretaceous granitoid rocks, Mesoproterozoic Belt Supergroup strata, and Paleozoic–Mesozoic strata. Pennsylvanian quartz arenite boulders and angular blocks of Belt Supergroup quartzite were derived from adjacent basin margin hillsides. Source rocks near the southern margin most likely consisted of Late Cretaceous granite and

Belt Supergroup strata, i.e., the same group of rocks that provided detritus to the northeastern Big Hole Basin. Abundant two-mica detritus was supplied by the northern Big Hole and Deep Creek paleovalleys. Phanerozoic thrust belt strata became more abundant in source areas adjacent to the Big Hole paleocanyon, similar to the modern, as evidenced by a significant increase in Phanerozoic clasts. The Late Cretaceous detrital zircon distribution that characterizes fluvial deposits in the Big Hole paleocanyon (74 Ma peak; fig. 25C, probability line E) reflects mixing of plutonic detritus from adjacent parts of the Pioneer Batholith and longitudinal fluvial plutonic detritus from the Big Hole (69 Ma peak) and Deep Creek (72 and 76 Ma peaks) Basins.

Divide and northernmost Melrose paleovalleys.

The Divide paleovalley was a southward-oriented headwater drainage basin for the ancestral Missouri River. Sediment in the northern part of the paleovalley (fig. 25A, location F) was primarily derived from surrounding Late Cretaceous granitoid rocks (Boulder Batholith) and Lowland Creek Volcanics (areas 7a and 8a). We propose the presence of a drainage divide (fig. 25A, location Y) between location F and the southern Deer Lodge paleovalley based upon strong compositional differences between longitudinal fluvial deposits in the basins, including the absence of Mesoproterozoic Belt Supergroup clasts, Proterozoic detrital zircon grains, and two-mica granite detritus that should otherwise be present if southward fluvial connectivity existed between the basins. Compared to the northern Divide paleovalley, the southern part of the paleovalley, near the confluence with the Big Hole paleocanyon, contains a significant increase in the variety of sedimentary clasts (Mesoproterozoic, Paleozoic, and Mesozoic sources; fig. 25B, histogram plot G left and right), reflecting basin margin supply from the ancestral Pioneer and Boulder Mountains (areas 7b and 8b) and mixing with longitudinal fluvial sediment from the Big Hole paleocanyon. Two-mica plutonic clasts delivered via the Big Hole paleocanyon become much less abundant as a result of dilution and intergranular breakdown of larger clasts. In contrast, ultradurable Belt Supergroup quartzite clasts remained abundant, making up the largest component in conglomerates (fig. 25B, plot G right), also revealing primary source areas to the west. The age distribution of the Late Cretaceous detrital zircon group in the paleo-confluence

region of the Big Hole and Divide paleovalleys, as indicated by the double-peak values (73 and 71 Ma) and age distribution (fig. 25C, probability line G), is younger than that in the Big Hole paleocanyon (74 Ma peak, probability line E) and northern Divide paleovalley (78 Ma peak, probability line F). Despite mixing with older granitic detritus, the younger distribution reflects major contribution from nearby plutonic rocks of the northeastern part of the eastern Pioneer Batholith (fig. 25C left, 71 Ma peak).

Melrose paleovalley. South of the confluence of the Big Hole and Divide paleovalleys, source rocks within the east-central Pioneer, southern Boulder, and McCartney Mountains, as well as localized Paleocene–Eocene volcanic rocks, fed detritus into the Melrose paleovalley from both margins (fig. 25A, source areas 8b, 9 and 10). In contrast to all of the paleoupstream basins, basin margin deposits in the Melrose paleovalley document that Paleozoic–Mesozoic strata flanked interior, primarily batholith-cored, regions of the ancestral highlands; Mesoproterozoic strata were much less abundant than in westward source areas (fig. 25B, histogram G left). Longitudinal fluvial conglomerates maintain a dominance of Belt Supergroup quartzite clasts (fig. 25B, histogram G right), primarily reflecting derivation from the westward-located basins. However, mixing from basin margin sources resulted in locally abundant Paleozoic and Mesozoic carbonate and siliciclastic clasts in addition to granitic clasts. Overall, the composition of deposits in the Melrose paleovalley became more complex than for other basins in the study area, consistent with greater complexity in makeup of the bounding highlands.

In the southern Melrose paleovalley, near the longitudinal fluvial outlet for the entire study area, detritus from all Late Cretaceous plutons in the study area were included in the longitudinal fluvial mix. The pluton-associated detrital zircon group (fig. 25C, probability line H) is slightly older (74 Ma peak) than in the confluence region, reflecting strong localized input from the McCartney Pluton (75 Ma peak). Increased abundance of zircon grains <65 Ma and >90 Ma reflect increased recycling from nearby Paleocene–Eocene Dillon volcanics source areas and regionally abundant Late Cretaceous strata, respectively.

Paleogene Evolution of the Deer Lodge and Deep Creek Basins

Rocks exposed in the Anaconda–Flint Creek structural culmination record a two-phase tectonic history of: (1) Late Cretaceous–Paleocene contraction, fold-thrust belt deformation, and plutonism, followed by (2) Eocene–Miocene extension including crustal detachment, localized rifting, and continued plutonism (e.g., Wallace and others, 1992; Foster and others, 2010; Kalakay and others, 2014; Howlett and others, 2021). Prior to this study, interpretations of Paleogene basin evolution adjacent to the ranges have primarily focused upon extensional tectonics, specifically related to development of the AMCC, i.e., the ADZ of this study. Our data offer additional perspectives regarding inception of the Deer Lodge and Deep Creek Basins and attendant exhumation of the flanking uplifts.

Basin Evolution

The age range of Paleogene deposits exposed in the Deer Lodge and Deep Creek Basins (~53 Ma to 22 Ma; Vuke, 2020a) temporally corresponds to estimates for extension on the ADZ (~53 Ma to ca. 25 Ma; Foster and others, 2010). Moreover, early Eocene maximum depositional ages (54–51 Ma) for deposits near the southwestern and eastern margins of the Deer Lodge Basin (sites 17 and 19), as well as the stratigraphic position of sites 17 and 18 below the Lowland Creek Volcanics (53–48 Ma), establish near synchronicity between deposition at these locations and extension on the ADZ (55–53 Ma; Foster and others, 2010). These age relations, combined with the local presence of normal fault-bound deposits and progressively tilted basin-fills along the western basin margin (Kalakay and others, 2003; O'Neill and others, 2004; Grice, 2006; Foster and others, 2007, 2010) and the projection of the detachment beneath the Deer Lodge Basin (McLeod, 1987; Foster and others, 2010), have been used as evidence for a supradetachment fault-related origin and extensional segmentation of the basin (Jänecke, 1994; Stroup and others, 2008).

In contrast, our results indicate an older, pre-extensional phase of basin development. For all basins in the WBR (fig. 2), the mere presence of coarse alluvial basin margin deposits indicates the presence of a rugged paleotopography with sediment transport into intervening basins (Schwartz and Schwartz, 2013; this study). In the Deer Lodge Basin, the oldest observed

Eocene deposits (~54 Ma), located on opposite sides of the basin (sites 18 and 19), record basin margin alluvial deposition upon the pre-Cenozoic unconformity (erosion surface), near the estimated initiation of extension at ~53 Ma (Foster and others, 2010), indicating that an intermontane basin coincident with the Deer Lodge Basin existed prior to extension. We argue that a pre-extensional paleovalley was formed during a phase of Late Cretaceous–Paleocene fluvial erosion, with incision focused along zones of stratigraphic and/or structural weakness. A rugged paleotopography and the occurrence of a Late Cretaceous–Paleocene incisional phase followed by Paleogene deposition is consistent with findings for the EBR (fig. 2; Schwartz and Schwartz, 2013). Findings for the SWBR (fig. 2) also document a rugged Paleogene landscape with coarse alluvial basin margin deposits and finer grained alluvial, lacustrine, and longitudinal fluvial basin interior deposits (e.g., M'Gonigle and others, 1991; VanDenberg, 1998; Matoush, 2002; Jänecke and others, 2005). Although basin origin in the SWBR has largely been attributed to extensional tectonics (summary by Jänecke, 2007 and references therein), pre-extensional lower Paleogene deposits exist within paleovalleys beneath middle Eocene Challis Volcanics (VanDenberg, 1998; Jänecke and others, 1999, 2000; Thoresen and others, 2024), suggesting a pre-extensional phase of intermontane paleovalley incision into the Cordilleran infrastructure. Other examples of pre-extensional paleovalley incision into contractionally deformed Mesozoic rocks comes from the Basin and Range Province of Nevada, where paleovalleys with estimated relief of up to ~1.5 km were incised into the contractional framework of the Sevier hinterland prior to post-Eocene extension (Henry, 2008; Henry and others, 2012; Zuza and others, 2021; Lundstern and others, 2024).

Fluvial incision of longitudinal intermontane paleovalleys and paleocanyons (water gaps) within fold-thrust terrains is well documented for active to recently active orogenic belts, such as in the Kuqa fold-thrust belt of western China (Delcaillau and others 2022), High Atlas of Morocco (Babault and others, 2012), Qilan Shan fold-thrust belt along the northeastern Tibetan Plateau (Ma and others, 2022), and Eastern Cordillera of northwest Argentina (Pingel and others, 2013). Numerical modeling (Koons, 1995) demonstrates that longitudinal drainage patterns are

inherited from earlier stages of fluvial organization that developed during upper crustal folding and thrusting, whereas transverse drainage segments across major ridges are formed by antecedence or stream capture resulting from headward erosion. Field studies document that the incisional basins intermittently aggrade, become endorheic, and are excavated in association with spatial and temporal changes in deformation, climate, and the erodibility of exposed bedrock (e.g., Hilley and Streckler, 2005). Similar to the deposits characterized in this study, those within active thrust belts consist of variably distributed slope, alluvial fan, alluvial plain, fluvial, and lacustrine deposits (Pingel and others, 2013; Piacentini and Miccadei, 2014; Delcaillau and others, 2022). A fold-thrust intermontane basin system may become fluvially excavated of sediment during widespread unconformity development, such as during the Paleocene in SWMT, resulting in a linked network of basins providing significant amounts of sediment accommodation, e.g., up to 1000 m (Hilley and Streckler, 2005; Streckler and others, 2009; Piacentini and Miccadei, 2014).

Following the Paleocene, the Deer Lodge, Deep Creek, and Big Hole Basins were variably influenced by extension that caused localized rifting, basin margin uplift, basin widening, and overall basin subsidence. Given the vast basin-fill thicknesses in the Deer Lodge and Big Hole Basins, extension locally played a major role in sediment accommodation. However, basin locations and orientations remained similar to their pre-extensional paleovalleys and were at least intermittently connected. The entire set of SWMT intermontane basins was filled throughout the remainder of the Paleogene. The widespread phenomenon of Paleogene basin filling was largely independent of local extension, and was instead associated with broader-scale drivers such as subcrustal dynamics (e.g., Cassel and others, 2018; Howlett and others, 2021) and/or climatic change (Schwartz and Schwartz, 2013).

Exhumation of Plutons in Basin-Bounding Highlands Prior to Extension

The early Eocene deposits of the southern Deer Lodge Basin provide a record of plutonic source rocks that were exposed prior to, and synchronous with, initial Paleogene deposition. At sites 17 and 18 (fig. 3), early Eocene deposits nonconformably overlie a Late Cretaceous granitic pluton, documenting pluton

exposure within the basin. Compositional and detrital zircon data from the same sites document sources that included Late Cretaceous plutonic rocks as well as Mesoproterozoic metasedimentary and Cambrian sedimentary rocks in the northern Anaconda and southern Flint Creek Ranges. Detrital zircon data from the eastern side of the basin (site 19) document a Boulder Mountains source area containing Late Cretaceous plutonic (Boulder Batholith) and volcanic rocks (Elkhorn Mountains Volcanics) as well as Paleocene–early Eocene volcanic rocks. Thus, Cretaceous plutons were exposed in basin-bounding uplifts across the region prior to extension. Other results from slightly younger deposits within the Lowland Creek Volcanics (e.g., 53–48 Ma) in the Deer Lodge Basin (sites 7, 13, 14, and 15) also support widespread exposure of Late Cretaceous plutons in the northern Anaconda and southern Flint Creek Ranges. Because of their depositional age, they also reflect continued exhumation during extension.

The exposure of Late Cretaceous plutons in highland regions prior to the onset of Paleogene deposition is consistent with provenance studies for other intermontane basins in SWMT, including those beyond the immediate influence of Eocene extension in the EBR (Schwartz and Schwartz, 2013; Rothfuss and others, 2012). Pre-extensional exposure of plutons is also supported by structural and thermobarometry data that suggest about 7 km of crustal removal and exposure of the Boulder Batholith by 53 Ma (Houston and Dilles, 2013).

Plutons and Related Detrital Zircon Records

That plutons may contain zircons of multiple ages, including those that represent magmatic pulses as well as those containing xenocrystic grains incorporated from host rocks (e.g., Proterozoic group in this study), is now well established (e.g., Miller and others, 2007; Claiborne and others, 2010; Barboni and others, 2013; Gaschnig and others, 2016). Further, the tendency for plutons within the CAB to contain mixed crystal-age populations is well documented and signifies a protracted history of repeated thermal events including magma emplacement, anatexis, and melt reworking (Chapman and others, 2021).

In similar fashion, composite age distributions for ADZ plutons show multiple age groups (fig. 27). The composites reveal up to seven zircon-age groups (peak

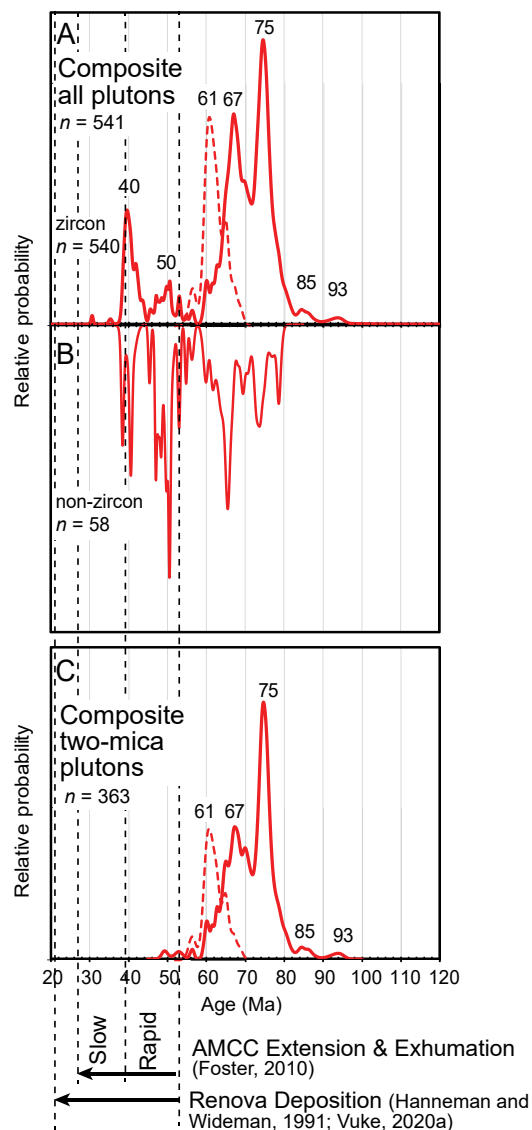
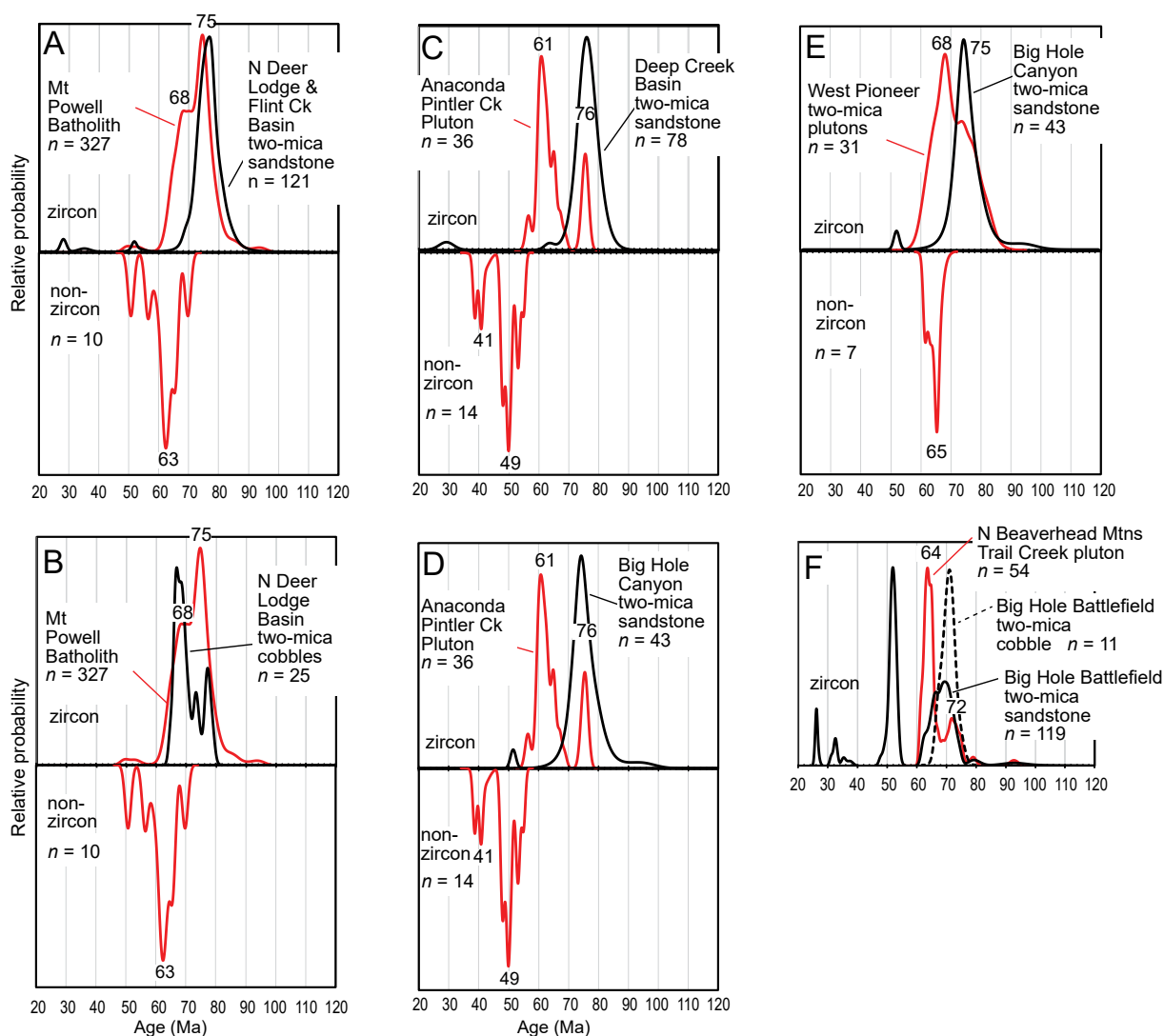


Figure 27. Composite zircon-age distributions for granitic plutons in the ADZ (Flint Creek and Anaconda Ranges) compared to timing of extension-related exhumation and Renova deposition. (A) Combined data for two-mica and non-two-mica plutons. (B) Data for two-mica plutons. Dashed red line, not to scale, illustrates the 61 Ma age group in Anaconda Range plutons that is included in the composite plot but suppressed due its proportionately small sample size (see figs. 5F and 28C). Bin width = 2 Myr (AMCC, Anaconda Metamorphic Core Complex).

ages ~93, 85, 75, 67, 61, 50, and 40 Ma; fig. 27A) with matching non-zircon-age groups (fig. 27B) for all but the ~93 and ~85 Ma zircon-age groups. The two-mica pluton dataset is characterized by just the five oldest zircon-age groups (93, 85, 75, 67, and 61 Ma; fig. 27C). Although the ~61 Ma peak is indistinguishable from the ~67 Ma peak in both datasets because of its proportionately small sample size, we schematically indicate its presence as documented in an individual pluton plot (fig. 5F).

Two-Mica Pluton Ages and Detrital Zircon Record

In contrast to multimodal composite zircon-age distributions that characterize plutons currently exposed in the Flint Creek and Anaconda Ranges, detrital zircon distributions that characterize Renova Formation deposits derived from plutons in those mountain ranges only contain Late Cretaceous grains (figs. 26E–26G). In order to address this discrepancy, we investigate the datasets that characterize two-mica plutons. Facies and paleocurrent data from the Renova Formation allow the identification of discrete source-to-sink connections with two-mica plutons in the Flint Creek Range (Mount Powell Batholith) and Anaconda Range (Pintler Creek Pluton) as well as in other ADZ-proximal locations (Trail Creek Pluton in the northern Beaverhead Mountains and Clifford Creek, Foolhen Mountain, and Doolittle Creek Plutons in the West Pioneer Mountains). The single-crystal ages for two-mica plutons are shown in figure 28 (red lines) and, where available, are subdivided into zircon U-Pb ages (upper half of diagrams) vs. non-zircon ages (i.e., K-Ar, $^{40}\text{Ar}/^{39}\text{Ar}$, and fission track ages; lower half of diagrams) to demonstrate relationships among mineralogy, chronometer, and interpreted age. Considering all chronometers together, age distributions for the Mount Powell, Pintler Creek, and West Pioneer two-mica plutons are multimodal. Zircon U-Pb ages form prominent Late Cretaceous age groups (75–76 Ma peaks), as well as transitional latest Cretaceous–Paleocene and Paleocene age groups (68 Ma and 61 Ma peaks; figs. 28A–28E). Non-zircon ages partially overlap the transitional latest Cretaceous–Paleocene zircon U-Pb distributions (figs. 28A, 28E), but also form age distributions that are dominantly younger owing to cooling rather than crystallization. Similar to the above two-mica plutons, the zircon U-Pb distribution characterizing the Trail Creek Pluton in the



Two-Mica Pluton Source Rocks and Linked Detrital Zircon Deposits

- A. Source: Mt Powell Batholith (two-mica), Flint Creek Range
N. Divide Basin sandstone - Site 1, CECL02
N. Divide Basin sandstone - Site 2, CMF-67-83
Flint Creek Basin sandstone - Site 4a & b, Portner-Link
- B. Source: Mt Powell Batholith (two-mica), Flint Creek Range
N. Divide Basin cobbles - Site 1, CECL01
- C. Source: Anaconda Range Pintler Creek Pluton (two-mica)
Deep Creek Basin sandstone - Site 23, DCCP-2
- D. Source: Anaconda Range Pintler Creek Pluton (two-mica)
Big Hole Canyon sandstone - Site 26, BH12-01
- E. Source: Western Pioneer Batholith (two-mica)
Big Hole Canyon sandstone - Site 26, BH12-01
- F. Source: N. Beaverhead Mountains Trail Creek Pluton (two-mica)
Big Hole Battlefield sandstone - Site 24, CE22BHB01 & CE20BHB17
Big Hole Battlefield two-mica granite cobble - Site 24, CE22BHB02

Figure 28. Comparison of two-mica pluton ages (red lines) with ages of linked two-mica Renova Formation deposits (black lines). Partitioned diagrams show zircon ages in upper half and non-zircon ages in lower half. Zircon U-Pb ages for all of the two-mica plutons are bimodal and include a Late Cretaceous age group as well as a transitional latest Cretaceous–Paleocene age group (Ck, Creek; Mt, Mount; N, North).

northern Beaverhead Mountains is bimodal, marked by age groups with 72 and 64 Ma peaks (fig. 28F). Younger Paleogene and older Mesozoic zircon ages in the plutons are rare to absent. It is noteworthy that xenocrystic Proterozoic grains also occur in all of the plutons.

In contrast to the pluton-age distributions, detrital zircon data from adjacent two-mica-bearing deposits (black lines in fig. 28) are predominantly unimodal and only show a strong correspondence with Late Cretaceous pluton dates. That is, they lack any of the younger zircon-age groups that are present in modern exposures of the plutons. The detrital zircon samples contain rare zircon grains younger than 65 Ma that we interpret, based upon independent compositional evidence, as derived from coeval Paleogene volcanic rocks located within the study area. We conclude that the lack of abundant Paleogene detrital zircon grains in Renova Formation deposits indicates that the Paleogene exposures of two-mica plutons were characterized by Late Cretaceous crystallization ages, and that younger phases of the plutons were not exposed until much later.

Temporal Change in the Detrital Zircon Record

For plutons with a mixed mineral-age record, a succession of pluton-derived age groups may be recorded in the stratigraphy depending upon age makeup of the pluton at different times of exposure. For this study, the time lag between exposure and deposition would have been minimized owing to the proximity of source and sink, thus enhancing interpretations of exposure history.

Although detrital zircon data for Neogene and younger sedimentary deposits containing two-mica detritus are sparse, we present two case examples of younger pluton-derived age groups being added to the depositional record. In the northern Big Hole Basin, middle Miocene beds of the Sixmile Creek Formation contain Late Cretaceous, Paleogene, and Eocene age groups that reflect the range of igneous zircon ages in the nearby Pintler Creek Pluton (fig. 26I). Also, nearby modern deposits from a first-order stream draining the Pintler Creek Pluton reflect the range of igneous zircon ages in the batholith (fig. 26J). The absence of ~70 Ma grains in the Pintler Creek Pluton igneous-rock data of Howlett and others (2021) is an artifact of small sample size ($n = 36$), as demonstrated by the

Mussigbrod Creek detrital zircon data (fig. 26J), collected atop the Pintler Creek Pluton (Link and others, 2008), which show a relative low from ~70 to 66 Ma and match the overall range in the detrital zircon ages. Collectively, the data in figures 26 and 28 indicate the exposure of Late Cretaceous plutons prior to the early Miocene (Hemingfordian, fig. 1) unconformity and the subsequent exposure of plutons with mixed Late Cretaceous to Eocene age groups in the same areas during and following unconformity development; that is, in association with Basin and Range extension and uplift that initiated ca. 17–16 Ma ago.

Conceptual Model for Pluton Evolution and Exhumation History of the ADZ

We propose a model (fig. 29) and a summary time chart (fig. 30) for progressive magmatism and pluton exhumation in the ADZ region. The model is based upon the mixed mineral-age record of presently exposed plutons in the Flint Creek and Anaconda Ranges, evidence for sustained sediment accommodation in adjacent basins beginning in the Eocene, and up-section changes in detrital zircon age distributions of the basin-fill.

A Late Cretaceous (~75 Ma) magmatic event occurred within the Flint Creek and Anaconda Ranges in association with Cordilleran deformation, high grade metamorphism, and anatexis (fig. 29A). The presence of ~93 and ~85 Ma zircon U-Pb dates in some plutonic samples (fig. 27) suggest pervasive overprinting by the ~75 Ma event and minimal age retention from the earlier Late Cretaceous magmatic events (e.g., as discussed for the Idaho Batholith by Gaschnig and others, 2016; Schwartz and others, 2021). The ~75 Ma granitoid plutons included both two-mica and biotite–hornblende varieties emplaced at depths of ~10–16 km based on thermobarometric data from associated high-grade metamorphic rocks (Grice, 2006; Haney, 2008).

Earliest Eocene (~54–53 Ma) deposits of the Renova Formation adjacent to the Flint Creek and Anaconda Ranges contain two-mica plutonic detritus and ~75 Ma detrital zircons (fig. 29), which requires that Late Cretaceous (~75 Ma) plutons of the Flint Creek and Anaconda Ranges had to have been exposed at the surface prior to the Eocene. Widespread Late Cretaceous to Paleocene erosion is recorded by a regionally occurring sub-Cenozoic unconformity

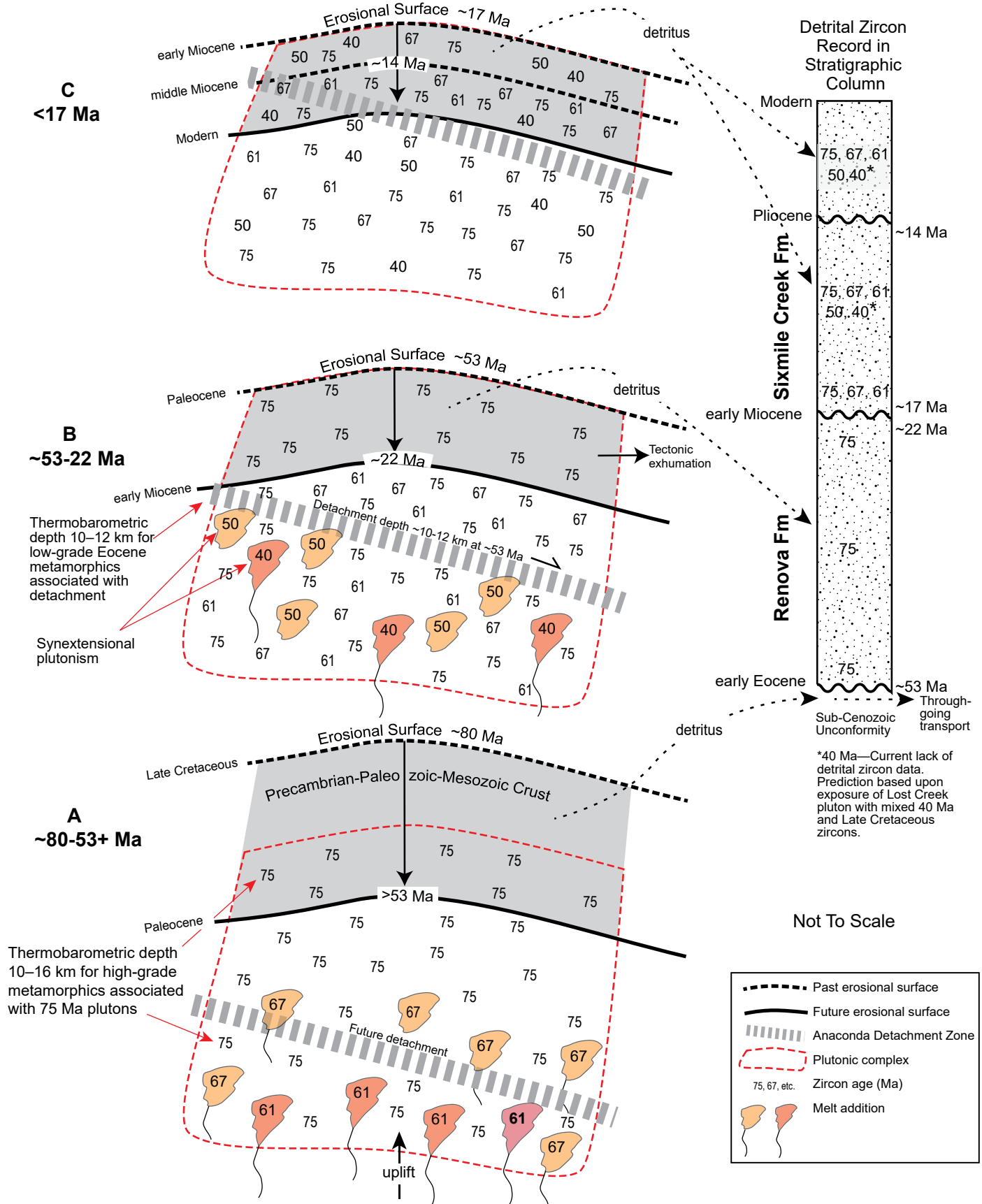


Figure 29. Conceptual model of mixed zircon-age development in Anaconda and Flint Creek Range two-mica plutons (Pintler and Mt Powell), progressive exposure of the igneous age record, and detrital zircon representation in adjacent basin fill. Some other non-two-mica plutons (e.g., Lost Creek, Philipsburg, and Trail Creek) also contain a mixed zircon-age record. Igneous zircon ages are summarized from those shown in figure 27 (Fm, Formation).

(Unconformity 1; fig. 1) that records deep incision of paleovalleys into the fold-thrust belt. Available thermobarometric data for metamorphic rocks associated with the ~75 Ma plutons (Grice, 2006; Haney, 2008) indicate that at least 10 km of crust had to have been eroded by the Eocene to allow for ~75 Ma two-mica plutonic detritus to be present in the basal Renova Formation. Available hornblende barometry data from the adjacent Boulder Batholith parallel those for the Flint Creek–Anaconda Range plutons, suggesting Late Cretaceous (~79–76 Ma) pluton emplacement at depths of ~6–9 km followed by rapid erosion of ~7 km of rock by ~53 Ma (Houston and Dilles, 2013). Such erosional depths and rates are regionally supported by sediment compositional data for Renova Formation deposits surrounding the Boulder Batholith that document widespread batholith exposure by the mid-Eocene (Schwartz and Schwartz, 2013), which would have required pre-Eocene erosion of its ~4.6-km-thick carapace of Elkhorn Mountains Volcanics (e.g., Klepper and Smedes, 1959; Robinson and others, 1968). Two additional magmatic events at ~67 Ma and ~61 Ma overprinted deeper structural levels of the igneous complex in the Flint Creek and Anaconda Ranges as upper ~75 Ma structural levels were being exhumed (fig. 29A).

Renova Formation deposition began with nearly concurrent extension and crustal detachment along the ADZ (~53 Ma; Foster and others, 2010), minor plutonism in the Flint Creek and Anaconda Ranges (~50 Ma; figs. 26F, 26G, 27A), and extrusion of the Lowland Creek Volcanics (53–48 Ma; Dudás and others, 2010). Thermobarometric data from Eocene mylonitic rocks associated with the ADZ indicate detachment depths of ~10–12 km at the onset of extension (Grice, 2006; fig. 29B). Late Cretaceous plutons were progressively exposed and continued to provide Late Cretaceous detrital zircons (~75 Ma) to Renova Formation deposits in adjacent basins; coeval magmatic events (ca. 50 and 40 Ma) at deeper structural levels further overprinted older plutons.

Subsequent to Renova Formation deposition (fig. 29C), plutons characterized by Late Cretaceous to Eocene dates, similar to those in modern exposures of the ADZ footwall, provided detritus to adjacent basins (e.g., figs. 26I, 26J). Although Neogene and younger detrital zircon data are limited, initial exposure of

mixed-age plutons most likely occurred during onset of Miocene Basin and Range extension (~17–16 Ma), correlative with development of the early Miocene unconformity across SWMT (Fields and others, 1985; Barnosky and others, 2007; Harris and others, 2017; Vuke, 2020a). If correct, the thermobarometric depth estimate for the ADZ (Grice, 2006) suggests that another ~10–12 km of crustal exhumation may have occurred by means of extension and erosion throughout the Paleogene and early Miocene, yielding a total of >20 km of crustal removal since Late Cretaceous (~75 Ma) plutonism and metamorphism.

Ramifications of Findings

Inconsistencies of our Data with the AMCC Model

Our findings regarding compositional aspects of the hanging wall and footwall in the ADZ and crustal level of structural detachment bring the AMCC model (O'Neill and others, 2004; Foster and others, 2010; Kalakay and others, 2014) into question, suggesting the ADZ does not fit a traditional metamorphic core complex model (Lister and Davis, 1989). Our data demonstrate that mountainous highlands of Cordilleran ancestry were present in the location of the Anaconda and Flint Creek Ranges during the Late Cretaceous–Paleocene, well before crustal detachment was initiated at the range-bounding ADZ. The data also demonstrate that the pre-detachment highlands included exposures of Late Cretaceous (~75 Ma) two-mica plutons and genetically related Late Cretaceous high-grade metamorphic rocks, as well as Mesoproterozoic, Paleozoic, and Mesozoic strata. Thus, an Eocene low-angle detachment zone would have transected Late Cretaceous plutonic and metamorphic rocks, making them significant components of both the footwall and hanging wall plates. This is supported by other recent findings that document rock types of the described “metamorphic core” (O'Neill and others, 2004; Foster and others, 2010; Kalakay and others, 2014) are exposed in the ADZ hanging wall, including granite near site 18 in the Deer Lodge Basin (Foster and others, 2010; Scarberry and others, 2019) and metamorphosed strata of the Late Cretaceous Kootenai and Blackleaf Formations in the eastern foothills of the Flint Creek Range (Scarberry and others, 2019). Also, the composition of the recently described ADZ hanging wall in the West Pioneer Mountains (Elliott and others, 2025) includes Late Cretaceous–Paleogene two-mica plutons

(fig. 8M) and Late Cretaceous high-grade metamorphic rocks and migmatites (Elliott and Lonn, 2021; Elliott, 2022), which we consider analogous to the rock composition that existed above the ADZ in the Anaconda–Flint Creek Range. Metamorphic–thermobarometric data from footwall rocks in the Anaconda and Flint Creek Ranges (Grice, 2006; Haney, 2008) indicate that high-grade metamorphism was associated with Late Cretaceous plutonism that occurred at about 10–16 km (Grice, 2006; Haney, 2008), which is above mid-crustal depths.

Overall, rather than separating a mid- or lower crustal metamorphic core from an unmetamorphosed hanging wall, the ADZ transected upper crustal rocks that included Late Cretaceous plutonic and metamorphic rocks. Exposure of the plutonic and metamorphic rocks well preceded the initiation of extension and detachment at about 53 Ma (Elliott and others, 2025).

Relevance of AMCC Thermochronologic Data to the ADZ Model and Renova–Sixmile Creek Deposition

Grice (2006) and Foster and others (2010) used thermochronologic data to define extension and exhumation history in their AMCC model. These data also apply to our proposed ADZ model of a preexisting, mountainous topography cut by a low-angle detachment zone under upper crustal conditions. In similar fashion, the $^{40}\text{Ar}/^{39}\text{Ar}$ data indicate relatively rapid exhumation of the ADZ footwall between 53 and 39 Ma, whereas apatite fission-track data indicate continued, but slower, exhumation of footwall rocks until about 27 Ma (Grice, 2006; Foster and others, 2010). Thus, footwall uplift accompanied most of the Renova Formation depositional period (~53–22 Ma), a time when Late Cretaceous plutons in the hanging wall were already supplying detritus to the basins. Although Foster and others (2010) focused upon hanging wall displacement as a tectonic driver for footwall exhumation, we recognize a combined role of tectonism and erosion. Detrital zircons diagnostic of currently exposed footwall rocks (i.e., mixed Late Cretaceous to Eocene zircon dates) were deposited for the first time in middle Miocene Sixmile Creek Formation deposits (~17–14 Ma; fig. 29C), suggesting initial exposure of the footwall at least 5 Myr following the cooling-based estimate for tectonic exhumation, most likely in association with early Miocene unconformity development and Basin and Range tectonism (fig. 30).

Horizontal Displacement Hypothesis for Deer Lodge Basin Origin

Our paleogeographic findings for the Deer Lodge Basin question previous interpretations for the magnitude of horizontal extension along the ADZ and possible relationship to basin origin. The presence of an incisional basin, i.e., large paleovalley, located between the ancestral Flint Creek Range and ancestral Boulder Mountains prior to extensional detachment, not only disfavors a sole detachment origin for the Deer Lodge Basin (Jänecke, 1994; Thomas, 1995; Stroup and others, 2008) but also argues against at least 25–28 km of eastward displacement of the Boulder Batholith away from the Anaconda–Flint Creek Range (Foster and others, 2010; Kalakay and others, 2014) to form a basin of modern scale (Foster and others, 2010; Kalakay and others, 2014). We acknowledge significant extensional modification of the original basin, but preclude tectonic separation as the initial and primary cause of basin location and width.

Core Complex Occurrence, Magmatism, and Extension in the CAB Compared to the Anaconda–Flint Creek Range

In figure 31 we compare our compilation of ages for magmatism, onset of extension, and rapid exhumation with those of core complexes in the CAB (Chapman and others, 2021). Chapman and others' (2021) ages of 53–48 Ma for AMCC magmatism are based upon $^{40}\text{Ar}/^{39}\text{Ar}$ analyses (Wallace and others, 1992; Foster and others, 2010) and are represented by the 50 Ma zircon age group in our ADZ dataset. Similar to our study area, core complexes in the CAB not only contain Cenozoic plutons (e.g., Armstrong and Ward, 1991), but magmatism north of latitude 43° is shown to be concurrent or temporally proximate to extension (fig. 31).

In comparison, zircon U–Pb data from plutons in our study area provide evidence for a long-term magmatic history leading up to, and following, extension (figs. 27, 31). The igneous zircon data indicate that up to five magmatic events (ca. 93, 85, 75, 67, and 61 Ma) preceded ADZ development and that two magmatic events (ca. 50 and 40 Ma) occurred during extension (fig. 27A). The Paleocene ages of Howlett and others (2021) and Eocene ages of Foster and others (2010) are included in the 61 Ma and 50–40 Ma

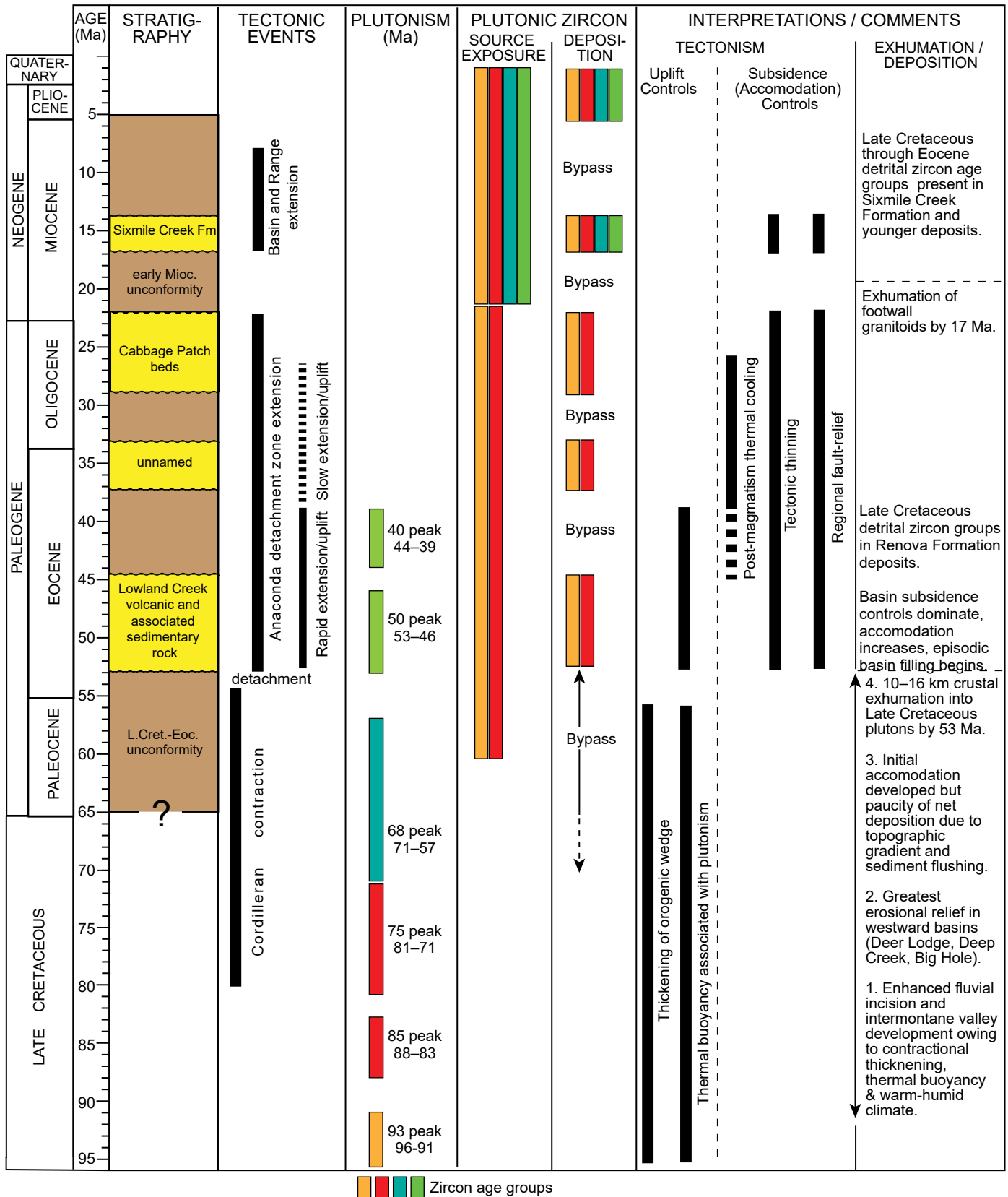


Figure 30. Summary chart of temporal relationships between Paleogene–Neogene deposition (yellow), unconformities (brown), tectonism, and plutonism. Also shown are times of plutonic zircon age-group deposition and inferred times of plutonic age-group exposure as well as interpretations and comments regarding tectonism, pluton exhumation, and deposition. Not included are subcrustal dynamics that may have driven crustal processes (Eoc, Eocene; L Cret, Late Cretaceous; Mioc, Miocene).

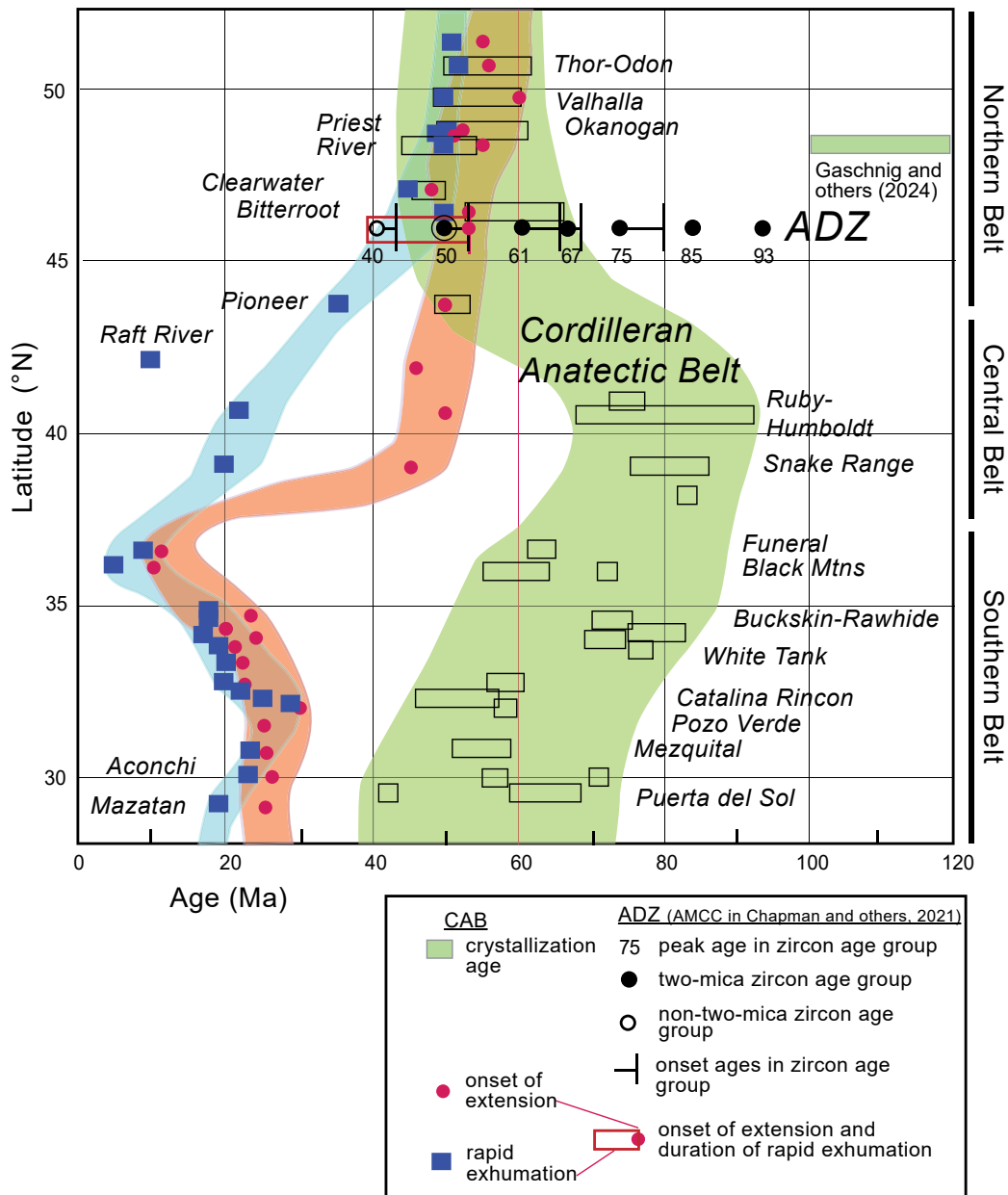


Figure 31. Plot of age vs. latitude for crystallization ages of rocks, rapid exhumation/cooling ages, and timing for onset of extension in the Anaconda Detachment Zone (ADZ) compared to envelopes (colored zones) containing ages for corresponding types of events in metamorphic core complexes of the Cordilleran anatectic belt (CAB) as defined by Chapman and others (2021). Reference locations for most core complexes in the CAB and data for defining the age envelopes are presented in Chapman and others (2021; AMCC, Anaconda Metamorphic Core Complex).

groups, respectively. Figure 27B shows that non-zircon (mica K-Ar and $^{40}\text{Ar}/^{39}\text{Ar}$) age groups match most zircon-age groups within a few Myr. However, data for the 50 and 40 Ma non-zircon-age groups primarily come from the Anaconda Range, where matching zircon-age groups are absent (fig. 5F); data from this region have been primarily attributed to exhumation-related cooling of the ADZ footwall (Grice, 2006; Foster and others, 2010). Even so, the close correspondence between zircon- and non-zircon-age groups in the Flint Creek Range (fig. 5G), as well as overall correspondence in the composite data (figs. 27A, B), suggests possible dependency upon rapid cooling following nearby, but not *in situ*, plutonism (e.g., Lister and Baldwin, 1993).

Overall, the geochronologic record indicates a protracted magmatic history of over 50 Myr with seemingly cyclical events having an average periodicity of ~11 Myr. The cyclical record reflects tectono-thermal high-flux events akin to those reported for the Late Cretaceous–early Paleogene Cordilleran hinterland of the United States, during which upper parts of thickened crust underwent shortening, extension, and partial melting on a 10 Myr time scale (Wells and others, 2012; DeCelles and Graham, 2015).

Spatial and temporal overlap of magmatic activity, crustal extension, and exhumation has led to focus upon the role of magmatism in uplift and core complex development (e.g., Lister and Baldwin, 1993; Hill and others, 1995; Konstantinou and others, 2013; Whitney and others, 2013; Stevens and others, 2016; Chapman and others, 2021). Our data support the notion that Cenozoic tectonic evolution of our study area may be the result of protracted (at least ~16 Ma) heating and partial melting of the deep crust. This is consistent with other evidence for protracted pre-detachment magmatism in core complexes of the CAB (e.g., Konstantinou and others, 2013; Stevens and others, 2016; Gaschnig and others, 2017).

Magmatic History of Two-Mica Granites

Two-mica granitoid rocks are fundamental components of both our study area and core complexes in the CAB (Chapman and others, 2021). The history of magmatism for two-mica granitoids in the Anaconda, Flint Creek, and Pioneer Ranges (fig. 27C) documents that two-mica petrogenesis began in the Late

Cretaceous. Thus, initial two-mica petrogenesis did not occur immediately prior to, or during, low-angle extension and Renova Formation deposition but at least 20–30 Myr earlier. This is fundamentally similar to the prolonged history of pre-extensional magmatism in the central and southern CAB where two-mica granitoid rock crystallization ages range from 92 to 42 Ma with a majority of ages between 80 and 50 Ma (fig. 31). Moreover, new two-mica data from the Priest River metamorphic complex near the U.S.–Canadian border support pre-extensional two-mica magmatism with ages of 120–100 Ma (fig. 31; Gaschnig and others, 2024).

The age trends reproduced in figure 31 led Chapman and others (2021) to suggest that anatectic magmatism, associated with two-mica granite development, started in the central CAB and swept northward and southward. However, Chapman and others (2021) provided the caveat that many plutons in the CAB only have a few dated samples and that where sufficient geochronologic data are available, the duration of crustal melting is typically protracted, lasting 10 Myr or more, e.g., the new data of Gaschnig and others (2024) and the two-mica granite data presented in this study (appendix C).

Two-Mica Magmatism and Uplift-Associated Extension

The above findings further reflect the relationship between the causes of two-mica magmatism and crustal extension within the CAB. Two-mica plutons in the CAB are considered to be partial melting products of thickened continental (metasedimentary) crust (Barbarin, 1999; Chapman and others, 2021). In addition to the study area, the long-term magmatic record for two-mica granitoids in the Idaho Batholith (Gaschnig and others, 2016), the record for two-mica magmatism in the Priest River metamorphic complex (Gaschnig and others, 2024), and the wide temporal spacing (e.g., 15–40 Myr) between magmatic and extensional events in the central to southern CAB (fig. 31) demonstrate a lack of dependency of two-mica magmatism upon core-complex occurrence.

Rather, this suggests that two-mica petrogenesis is dependent upon crustal composition and conditions of partial melting (Barbarin, 1999) in association with subcrustal dynamics that lead to long-term heating

(Ducea and others, 2026), uplift, and ultimately extension with attendant basin subsidence and filling. Such is compatible with Zuza and Cao's (2022) posit that thermally induced buoyant ascent of migmatitic or plutonic cores (e.g., Whitney and others, 2004) may eventually lead to doming and detachment faulting in association with regional extension. The fundamental driver for heating, magmatism, and uplift events, including core complex development, along the northern to central CAB, has been ascribed to delamination or slab rollback of the Farallon slab (e.g., Konstantinou and others, 2013; Cassel and others, 2018; Chapman and others, 2021; Zuza and Cao, 2022). Cogent arguments against slab rollback in this region have also been published (e.g., Tikoff and others, 2023). Consistent with all the data, however, is the need for a long-term crustal scale thermal driver leading to uplift and eventually widespread Paleogene extension.

Montanapiano as a Tectonomorphic Zone in SWMT

Contrary to spatial context of the Nevadapiano (DeCelles, 2004), where an orogenic plateau in the hinterland was flanked by an eastern foreland region and a westward magmatic arc, Lageson and others (2020) propose that the Montanapiano extended eastward from the hinterland in Idaho to include a large part of the Cordilleran fold-thrust belt and its intermixed plutons (fig. 32A), thus placing it adjacent to the Late Cretaceous–Paleocene foreland basin and Western Interior Seaway. Findings from the WBR (this study) and EBR (Schwartz and Schwartz, 2013), mostly located in the fold-thrust belt sector of the proposed Montanapiano, contradict the existence of an orogenic plateau in that region. Instead, evidence indicates that fluvially incised paleovalleys with relief up to 2 km were present during the Paleocene and most likely initiated during regional Late Cretaceous uplift, similar to modeling results for landscape evolution in fold-thrust belt regions of continental mountain belts (Koons, 1995) and piggyback basin development in modern continental fold-thrust belts (e.g., Suriano and others, 2015). Although maximum paleoelevation was high (e.g., ≥ 4 km; Kent-Corson and others, 2006; Schwartz and others, 2019b; Fan and others, 2017, 2021), relief was also high (e.g., up to 2 km). Compared to the internally drained Puna–Altiplano orogenic plateau (e.g., fig. 32B), drainage in the pre-Paleocene foreland of SWMT was necessarily exorheic in order to

allow for intense fluvial incision and focused intermontane valley development. Such drainage would be expected given the paleogeographic location of the eastward thrust belt and plutons, adjacent to the Late Cretaceous–Paleocene foredeep and Western Interior Seaway. Moreover, fluvial incision would have been enhanced because of warm and wet climatic conditions during the Late Cretaceous through early Eocene (Schwartz and Schwartz, 2013; Schwartz and others, 2019b). Overall, the pre-Paleocene foreland landscape in SWMT was similar to modern continental foreland systems and bears a strong resemblance to the “broken foreland” region of northwestern Argentina, where a mixture of thin-skinned and thick-skinned (Laramide-style) structures influence the landscape (fig. 32B; Jordan and Allmendinger, 1986; Hilley and Strecker, 2005; Strecker and others, 2011).

CONCLUSIONS

Paleogeography and Paleodrainage

The Paleogene landscape of SWMT consisted of elongate mountainous highlands (maximum elevations ≥ 4 km) and intermontane paleovalleys (maximum valley margin relief ~ 2 km), similar in location to the intermontane basins that characterize SWMT today. Our findings conflict with the concept of a topographically subdued orogenic plateau making up the region during the Late Cretaceous and Paleocene. Fluvial pathways were dominantly longitudinal, with some transverse connections between paleovalleys via narrow paleocanyons (water gaps) in intervening uplifts. Headwaters of the ancestral Missouri River extended into paleovalleys coincident with the northern Big Hole, southern Deer Lodge, and Divide Basins, all of which contributed to the ancestral Big Hole paleovalley that continued through the Melrose paleovalley and drained eastward, out of the WBR study area, into the ancestral Jefferson River system of the EBR. Paleodrainage in the northern Deer Lodge paleovalley was northwestward into the Flint Creek Basin region as part of the ancestral Columbia River watershed.

Tectonic History and Basin Evolution

Intermontane basin evolution in the study area is associated with several tectonic phases: (1) Late Cretaceous–Paleocene contractional deformation and plutonism (ca. 80 to 53 Ma; Lageson and others, 2020); (2) early Eocene through early Miocene exten-

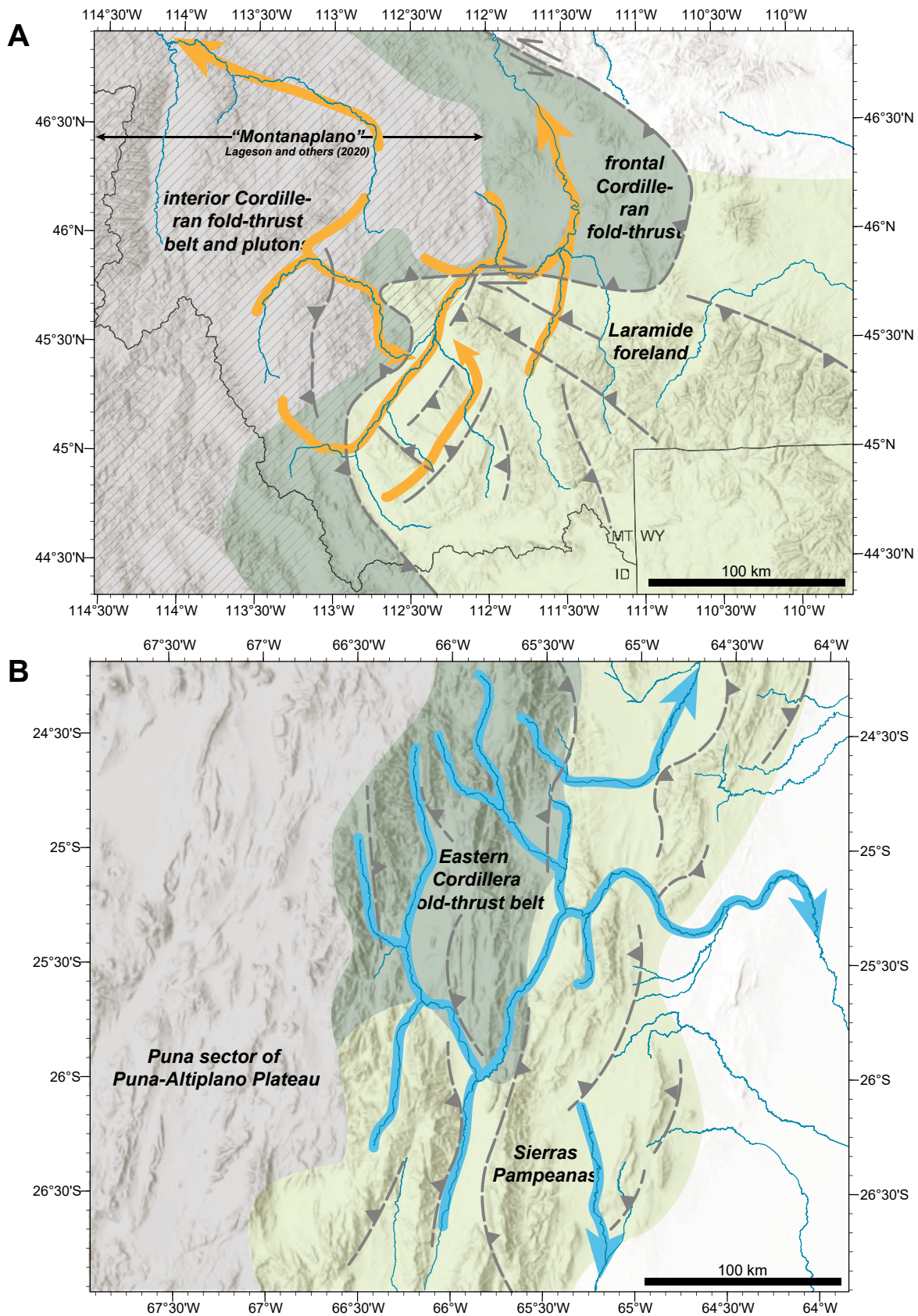


Figure 32. Comparison of structurally controlled drainage patterns interpreted for (A) the Paleogene landscape of Montana (orange lines; adapted from Schwartz and Schwartz, 2013; Schwartz and others 2019b; major faults adapted from Lageson and others, 2020; Vuke and others, 2020b) to (B) modern-day drainages (blue lines) associated with the Puna-Altiplano Plateau and Sierras Pampeanas of Argentina (structural provinces and major faults adapted from Carrapa and others, 2011).

sional deformation (ca. 53–25 Ma), including crustal detachment and magmatism (Foster and others, 2010); and (3) mid-Miocene through present Basin and Range-style extension (ca. 17–modern; Yakovlev and others, 2020; fig. 30). Major crustal thickening and uplift during the Late Cretaceous–Paleocene contractional phase promoted fluvial incision into structurally and lithologically susceptible zones of the Cordilleran tectonostratigraphy, resulting in deep paleovalleys that served as extensionally modified Tertiary intermontane basins. Incision of the paleovalley basins was similar to that documented for intermontane basins of the EBR and those incised in modern continental fold-thrust belts. Subsequent early Eocene–early Miocene basin filling, including intermittent exhumation and unconformity development, was strongly influenced by extension. Shared fundamental properties for the complete set of intermontane basins in the northwestern U.S. reflects a large-scale driver for extension and subsidence, such as subcrustal tectonic events, in addition to that of locally enhanced extensional (e.g., detachment-associated) tectonics.

Zircon Age Distributions in Batholiths and Magmatic History

Frequency distributions of all zircon data from granitic plutons facilitate comparisons with detrital zircon distributions, and provide insight into magmatic history. Plutons in the Boulder Mountains (Boulder Batholith), East Pioneer Mountains (eastern Pioneer Batholith), and parts of the West Pioneer Mountains (western Pioneer Batholith) are characterized by relatively narrow, unimodal, Late Cretaceous zircon-age distributions with peak values between 76 and 71 Ma. In contrast, two-mica plutons in the Flint Creek and Anaconda Ranges and West Pioneer Mountains are characterized by polymodal zircon-age distributions. Most abundant are Late Cretaceous (peak ~75 Ma) and latest Late Cretaceous–Paleocene (peaks ~67–61 Ma) dates with minor occurrences of Eocene dates (peaks ~50 or 40 Ma) and rare older dates ~93 and 85 Ma (figs. 28, 30). Fundamentally, unimodal age distributions in the eastern batholiths indicate simple thermal-event histories whereas polymodal age distributions in the western batholiths indicate multi-phase thermal histories.

Provenance and Initial Exposure of Granitic-Cored Uplifts

Compositional and detrital zircon data demonstrate that Late Cretaceous granitic plutons were exposed in all uplifts flanking the intermontane basins of the WBR. Conglomerate clasts also document the exposure of Proterozoic, Paleozoic, and Mesozoic strata that make up deformed foreland strata. Paleocene–Eocene clasts were incorporated from local sedimentary and volcanic sources.

Detrital zircon distributions that characterize Paleogene deposits typically include abundant first-cycle, Late Cretaceous, pluton-derived zircons and abundant Proterozoic zircons recycled from Proterozoic, Paleozoic, and Mesozoic siliciclastic strata. Volcanic zircons recycled from Late Cretaceous, volcanoclastic foreland strata and Paleogene volcanic rocks are locally abundant.

The sole occurrence of pluton-derived zircons with a Late Cretaceous age (76 Ma peak) in earliest Eocene deposits (ca. 54–51 Ma) documents that granitic plutons were exposed in the Flint Creek Range and Boulder Mountains prior to Anaconda detachment (ca. 53 Ma). Nonconformable relationships between Eocene deposits and Late Cretaceous plutonic rocks in the Deer Lodge Basin also document incision into Cretaceous granitoid rocks and associated metamorphic rocks. The exposure of granitic plutons prior to, and during, Paleogene deposition is consistent with findings for intermontane basins in the EBR, including those beyond the influence of significant Eocene extension. Two-mica granitic clasts were derived from the Mount Powell Batholith in the Flint Creek Range, plutons in the Anaconda Range (e.g., Pintler Creek Pluton and Trail Creek Pluton), and several plutons in the West Pioneer Mountains. Mesoproterozoic Belt Supergroup quartzite clasts, the primary gravel-sized component in longitudinal fluvial deposits of all basins except the Divide Basin, were mostly derived from the western part of the study area, i.e., from upland regions west of the Deer Lodge, Deep Creek, and Big Hole paleovalleys as well as in the vicinity of the Big Hole paleocanyon. Conglomerate clast data also indicate that although Paleozoic and Mesozoic sedimentary strata were prevalent sources in the eastern part of the study area (adjacent to the Divide and Melrose

Basins), gravel-size Mesoproterozoic Belt Supergroup quartzite clasts remained dominant in trunk systems because of their ultra-durability.

Detrital Zircon Record in Longitudinal Fluvial Systems

Longitudinal fluvial detrital zircon data are fundamentally similar throughout the entire drainage system and are characterized by a prominent, first-cycle, pluton-derived Late Cretaceous age group as well as a prominent, recycled, Proterozoic age group. Exceptions include local abundances of Paleocene–Eocene volcanic zircons and local abundances of transitional Early–Late Cretaceous zircons that were recycled from Late Cretaceous volcanoclastic strata. A lack of Proterozoic zircons, as well as Mesoproterozoic Belt Supergroup clasts, in the longitudinal fluvial system of the northern Divide paleovalley indicate a lack of fluvial connectivity with the southern Deer Lodge paleovalley.

At least seven plutonic complexes were exposed across the study area during the Paleogene. Pluton-associated detrital zircon-age groups along the longitudinal fluvial pathway are typically unimodal, similar in age, and show only slight shifts in peak values, typically from ca. 76 to 73 Ma. The most distinct differences occur where plutonic rocks with a markedly different age are the sole or primary source, e.g. in the northern Big Hole Basin (fig. 25C, plot D) and northern Divide Basin (fig. 25C, plot F). Based upon longitudinal fluvial detrital zircon data alone, the contribution of specific plutonic-source areas to the longitudinal fluvial detrital zircons becomes increasingly speculative for deposits in the direction of paleo-downstream mixing.

Two-Mica Plutons, Progressive Unroofing, and Temporal Change in the Source-Rock and Detrital Zircon Record

Two-mica granite clasts and sand within the Renova Formation in the Deer Lodge, Deep Creek, Big Hole, and Big Hole paleovalleys document the exposure of Late Cretaceous two-mica plutons in ancestral highlands coincident with the Flint Creek Range, Anaconda Range, northern Beaverhead Mountains, and West Pioneer Mountains. However, in contrast to the multimodal zircon-age distributions that characterize modern two-mica pluton exposures, the Renova For-

mation two-mica detrital zircon data are unimodal and only show a strong match with the Late Cretaceous igneous zircons. Thus, the detrital zircon data indicate sole exposure of Late Cretaceous plutons prior to, and during, Renova Formation deposition. Available data from younger deposits (e.g., the <17 Ma Sixmile Creek Formation) document that pluton exposures in the same source areas provided first-cycle zircons from Late Cretaceous, latest Cretaceous–Paleocene, and Eocene igneous-age groups after early Miocene unconformity development; that is, during and following initiation of Basin and Range-style extension.

The combined data suggest that the igneous complexes had evolved, by means of melt addition, anatexis, and age overprinting, such that Late Cretaceous crystallization ages were preserved in older upper structural levels of the complexes and that mixed Late Cretaceous to Eocene age groups marked deeper structural levels. As deeper structural levels of the plutons were progressively exhumed, the Late Cretaceous to mixed Late Cretaceous–Paleocene–Eocene igneous-age record supplied zircons to younger and younger sedimentary deposits (fig. 30).

Pre-Extensional Setting and the Core Complex Model

Eocene extension in SWMT overprinted a Cordilleran thrust belt terrain marked by high-elevation, high-relief upland areas separated by fluvially incised intermontane valleys. Our evidence for the presence of a Paleocene intermontane paleovalley, i.e. basin, in the location of the Deer Lodge Basin is inconsistent with sole supradetachment origin of the basin and consistent with evidence for incisional paleovalley inheritance in other intermontane basin locations in the Rocky Mountains Basin and Range Province.

Late Cretaceous granitic plutons, including peraluminous two-mica plutons and high-grade metamorphic rocks, were present and exposed in the ADZ region prior to detachment. Thus, magma was originally emplaced into the Cordilleran infrastructure, into both future footwall and hanging wall locations, during the Late Cretaceous. Tectonic and erosional exposure of Paleocene–Eocene plutons in the ADZ footwall did not occur until late Oligocene–early Miocene unconformity development (fig. 1) and Basin and Range tectonism.

Although two-mica plutons are spatially associated with metamorphic core complexes within the CAB (Chapman and others, 2021), the established presence of Late Cretaceous two-mica granitic exposures in the study area prior to early Eocene extension and the documentation of long-term magmatic records for two-mica granitoids in the Idaho Batholith (Gaschnig and others, 2016) and the Priest River metamorphic complex (Gaschnig and others, 2024) collectively indicate genetic independence from core complex development. Instead, two-mica granitoid occurrence was dependent upon crustal composition, conditions of partial melting, and heating associated with sub-crustal dynamics. Zircon age distributions for plutons in the Anaconda–Flint Creek region indicate a protracted magmatic history of over 50 Myr with seemingly cyclical events having an average periodicity of ~11 Ma, culminating during ADZ development. Contractual tectonic thickening coupled with thermally induced pluton buoyancy between the Late Cretaceous and early Eocene would have enhanced erosion and unconformity development, including sediment flushing from incised valleys (fig. 30). Overall, we question some aspects of the AMCC model and the suitability of a metamorphic core complex model for the Anaconda and Flint Creek Ranges. Fundamentally, rather than separating a mid- or lower crustal metamorphic core from an unmetamorphosed hanging wall (AMCC model), the ADZ transected upper crustal rocks that included Late Cretaceous peraluminous two-mica granites and high-grade metamorphic rocks that were well exposed prior to detachment at ~53 Ma.

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APPENDIX A
SITE AND SAMPLE LOCATIONS

Appendix A. Study sites and sample locations.

Basin	Site #	Outcrop Locale	Sample	Latitude	Longitude	Age or Stratigraphic Unit	Source
Northern Deer Lodge	1	Rock Creek Ranch	CECL01, CEMF02, DLCP-3	46.49356	-112.7772	Late Oligocene–early Miocene	Scarberry and others, 2019
Northern Deer Lodge	2	Deer Lodge west	CMF-67-83	46.49828	-112.88946	Late Oligocene–early Miocene	Scarberry and others, 2019
Northern Deer Lodge	3	Grant-Kohrs Ranch historic site	17-DLCP-2	46.40571	-112.75952	Late Oligocene–early Miocene	Scarberry and others, 2019
Flint Creek	4a	Drummond	Link	46.586309	-113.035805	Late Oligocene–early Miocene	Portner and others, 2011
Flint Creek	4b	Drummond	Portner	46.6022222	-113.1186111	Late Oligocene–early Miocene	Portner and others, 2011
Southern Deer Lodge	5	Racetrack Creek west	RC13-01	46.278496	-112.952706	Renova	Elliott and Lonn, 2013; this study
Southern Deer Lodge	6	Racetrack Creek east	RC13-02	46.28379	-112.935761	Renova	Elliott and Lonn, 2013; this study
Southern Deer Lodge	7	Spring Gulch/Gallick Ranch	GR12-01	46.219826	-112.9215835	Renova–Lowland Creek–Eocene	Elliott and Lonn, 2013; Scarberry and others, 2019
Southern Deer Lodge	8	Spring Gulch/Gallick Ranch	17-SG-1	46.21997	-112.91859	Renova–Lowland Creek–Eocene	Elliott and Lonn, 2013; Scarberry and others, 2019
Southern Deer Lodge	9	Spring Gulch/Gallick Ranch	GR12-02A	46.22032	-112.92065	Renova–Lowland Creek–Eocene	Elliott and Lonn, 2013; Scarberry and others, 2019
Southern Deer Lodge	10	Spring Gulch/Gallick Ranch	17-SG-2	46.22179	-112.92051	Renova–Lowland Creek–Eocene	Elliott and Lonn, 2013; Scarberry and others, 2019
Southern Deer Lodge	11	Anaconda West	17-LCV-1	46.14335	-112.9782	Paleocene–Eocene	Scarberry and others, 2019
Southern Deer Lodge	12	Anaconda Old Works	OW12-04	46.137578	-112.939714	Late Eocene–Early Oligocene	Scarberry and others, 2019; this study
Southern Deer Lodge	13	Anaconda Old Works north	OW12-06	46.14047	-112.93607	Paleocene–Eocene	Scarberry and others, 2019
Southern Deer Lodge	14	Anaconda Old Works northeast	17-OWNE-1	46.14199	-112.91801	Eocene	Elliott and Lonn, 2013
Southern Deer Lodge	15	Sheep Gulch tuff	SG13-02	46.11922	-112.9617	Eocene	Elliott and Scarberry, 2013
Southern Deer Lodge	16	Sheep Gulch sandstone body	SG13-03	46.11826	-112.96245	Eocene	Elliott and Scarberry, 2013
Southern Deer Lodge	17	ARCO site	ARCO	46.1068253	-112.9078684	Paleocene–Eocene	Scarberry and others, 2019
Southern Deer Lodge	18	Mill Creek	17DV04	46.08587	-112.92746	Paleocene–Eocene	Scarberry and others, 2019
Southern Deer Lodge	19	Whitcraft Gulch	CE17BN1	46.159679	-112.76191	Paleocene–Eocene	Scarberry and others, 2019
Deep Creek	20	Grassy Mountain	17-GMCP-1	46.03056	-113.0136	Paleocene–Eocene	Scarberry and others, 2019
Deep Creek	21	Deep Creek	17-DCCP-3	45.97591	-113.07535	Renova	Elliott, 2015
Deep Creek	22	French Creek	FC	45.9437607	-113.0749714	middle Oligocene–early Miocene	Elliott, 2017
Deep Creek	23	Deep Creek	17-DCCP-2	45.93913	-113.09204	middle Oligocene–early Miocene	Elliott, 2017
Northern Big Hole	24	Big Hole Battlefield	CE22BHB01, CE20BHB17	45.6379	-113.6561	middle Oligocene–early Miocene	Elliott, 2017
Northern Big Hole	25	Wisdom north	WNBU	45.6519525	-113.437806	Eocene–Oligocene	Elliott, 2024
Big Hole Canyon	26	Big Hole Canyon	BH-12-01	45.802299	-112.922774	Early Oligocene–Early Miocene/Arikarean	Roe, 2010
Divide	27	Silver Bow south	CP09-3	45.9994806	-112.6587	Renova	this study
Divide	28	Feeley-Buxton north	REN-1-08	45.92149	-112.70287	Cabbage Patch–early Miocene	McDonald and others, 2012
Divide	29	Feeley-Buxton north	CP09-4	45.9194306	-112.6634111	Cabbage Patch–early Miocene	McDonald and others, 2012
Divide	30	Feeley-Buxton north	CP09-5c	45.9009611	-112.7069	Cabbage Patch–early Miocene	McDonald and others, 2012
Divide	31	Feeley-Buxton north	CP09-5b	45.8917389	-112.7055889	Cabbage Patch–early Miocene	McDonald and others, 2012
Divide	32	Feeley-Buxton north	CP09-2	45.9014	-112.6573611	Cabbage Patch–early Miocene	McDonald and others, 2012
Divide	33	Divide north	CP09-6	45.85155	-112.728339	Cabbage Patch–early Miocene	McDonald and others, 2012
Divide	34	Divide north	CP09-7	45.8055694	-112.7220389	Cabbage Patch–early Miocene	McDonald and others, 2012
Divide	35	Divide north	CP4-08	45.79181	-112.71826	Cabbage Patch–early Miocene	McDonald and others, 2012
Divide	36	Divide north	CP09-1	45.7840694	-112.7246194	Cabbage Patch–early Miocene	McDonald and others, 2012
Divide	37	Divide	CP2	45.75478	-112.738372	Cabbage Patch–early Miocene	McDonald and others, 2012
Divide	38	Divide	DIV3-08	45.73557	-112.73853	Cabbage Patch–early Miocene	McDonald and others, 2012

Appendix A—Continued.

Basin	Site #	Outcrop Locale	Sample	Latitude	Longitude	Age or Stratigraphic Unit	Source
Divide-Melrose transition	39	Moose Creek	CLA-6-08	45.7186	-112.70382	Orellan-Eocene	McDonald and others, 2012
Divide-Melrose transition	40	Soap Gulch north	CLA-508	45.68576	-112.68713	Orellan-Eocene	McDonald and others, 2012
Melrose	41	Melrose west	PIO-5-09	45.63332	-112.72013	Chadronian-Eocene	McDonald and others, 2012
Melrose	42	Cherry Creek	Ren11-11	45.592654	-112.790253	Renova	McDonald and others, 2012
Melrose	43	McCartney Creek	PG-3-08	45.59	-112.65549	Renova	McDonald and others, 2012
Melrose	44	McCartney northwest	PG-2-08	45.56609	-112.68091	Renova	McDonald and others, 2012
Melrose	45	Glen	GLEN-6-09	45.4715297	-112.714669	Renova	McDonald and others, 2012
Melrose	46	McCartney southwest	Ren11-06	45.49784	-112.6763	Renova	McDonald and others, 2012
Melrose	47	McCartney southwest	Ren11-07	45.48012	-112.66433	Renova	McDonald and others, 2012
Melrose	48	McCartney southwest	Ren11-08	45.48187	-112.66301	Renova	McDonald and others, 2012
Melrose	49	McCartney south	REN09-3	45.4686	-112.6430194	Renova	McDonald and others, 2012
Melrose	50	McCartney south	Ren11-02	45.46741	-112.63783	Renova	McDonald and others, 2012
Melrose	51	McCartney south	CHA-1-08	45.45893	-112.6021	Renova	This study

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APPENDIX B
CLAST COMPOSITION TABLE

Appendix B. Conglomerate/Breccia clast composition in %.

Basin Margin Alluvial	Sample	Lat	Lon	Proterozoic Belt	Proterozoic Ls	Proterozoic Lahood	Marble	Metamorphic-Deformed Pz Carbonate	
S. Deer Lodge Basin	GR12-02 A Gallik Ranch	46.2203	-112.9206	48	12		5		
S. Deer Lodge Basin	GR12-02 H Gallik Ranch	46.2218	-112.9205	69	4		3		
S. Deer Lodge Basin	RC13-01 race trac ck	46.2785	-112.9527						
S. Deer Lodge Basin	RC13-02 race trac ck	46.2838	-112.9358	22			11		
S. Deer Lodge Basin	OW12-06 granite debris flow	46.1405	-112.9361						
S. Deer Lodge Basin	17OWNE-1 basin margin fluvial	46.1420	-112.9180	64					
Deep Creek Basin	17GMCP-1	46.0306	-113.0136	34	15		16	19	
S. Divide Basin	CP-4-08	45.7918	-112.7183			25			
N. Melrose Basin	CLA-6b-08 ALLUV	45.7186	-112.7038			40			
S. Melrose Basin	GLEN-6-09	45.4715	-112.7147						
S. Melrose Basin	PG-2-08	45.5661	-112.6809						
S. Melrose Basin	REN09-3b	45.4686	-112.6430	2					
S. Melrose Basin	CHA-1-08	45.4589	-112.6021						
Axial Fluvial									
N. Deer Lodge Basin	17DLCP-3 axial fluvial	46.4936	-112.7772	48					
S. Deer Lodge Basin	OW12-04 old works	46.1376	-112.9397	52	47				
Deep Creek Basin	17DCCP-3 axial fluvial	45.9391	-113.0920	74					
Big Hole Canyon Basin	BH12-01 big hole gorge axial fluv	45.8023	-112.9228	73					
N. Melrose Basin	CLA-5c-08	45.6858	-112.6871	45		25			
N. Melrose Basin	CLA-6a-08	45.7186	-112.7038	64		4			
S. Melrose Basin	PG-3-08	45.5900	-112.6555	15					
S. Melrose Basin	Ren11-07	45.4801	-112.6643	70					
S. Melrose Basin	REN09-3a	45.4686	-112.6430	72					

	Paleozoic Carb	Paleozoic Terrigenous	Paleozoic- Mesozoic Terrigenous	Cretaceous Terrigenous	Granite	Two- Mica Granite	Cretaceous Volc	Paleocene- Eocene Volc.	Cretaceous- Paleocene- Eocene Volc undiff.	Paleogene Terrigenous	Hornfels	Other
	26		7					1				1
	12	2	1								9	
	100											
	9	13						9		37		
					9			91				
		4			16			8		8		
	14							2				
	21	30			21			3				
	40	9			10		0	1				
	2	3		90	0		0		3			2
	25	30		20	15		0		10			
		1		96			1					
				75	5		0	20				
				3	1	16	23	9				
								1				
					5	11				2	2	6
		11			1	12		3				
					29							1
	4	10			11				7			
	45	6	9	22				3				
	4			17	6		1	2				
		4		19	3		2					

APPENDIX C
GEOCHRONOLOGY DATA

Appendix C. Age, figure, location, and source information for detrital zircon and igneous geochronology data (tables C1–C3).

Table C1. Detrital Zircon: Age, location, and source information for detrital zircon samples utilized in this study.

Site No.	Figure	Sample	Age	Formation	Sedimentary Basin	Lat	Lon	Source
1	8A	CECL01	Late Oligocene–early Miocene	Renova Fm (Cabbage Patch mem.)	Northern Deer Lodge	46.49356	-112.7772	MBMG geochron data archive
1	8B	CECL02	Late Oligocene–early Miocene	Renova Fm (Cabbage Patch mem.)	Northern Deer Lodge	46.49356	-112.7772	MBMG geochron data archive
2	8C	CMF-67 (CMF-67-83)	Paleogene	Renova Fm	Northern Deer Lodge	46.4983	-112.8895	Schwartz and others, 2019
4a	8D	Portner-Link (RAP1127-1, CP1)	Oligocene–Miocene (Arikareean)	Renova Fm (Cabbage Patch mem.)	Flint Creek	46.5863	-113.0358	Link and others, 2008
4b	8D	Portner-Link (II_95_1)	Paleogene	Renova Fm (Cabbage Patch mem.)	Flint Creek	46.5947	-113.1140	Portner and others, 2011
7	8E	GR12-01	Paleogene	Renova Fm	Southern Deer Lodge	46.2200	-112.9187	Schwartz and others, 2019
12	8G	OW12-04	Paleogene	Renova Fm	Southern Deer Lodge	46.1376	-112.9397	Schwartz and others, 2019
18	8H	17DV04	Eocene	Renova Fm	Southern Deer Lodge	46.0859	-112.927	MBMG geochron data archive: Yakovlev and Elliott, Butte North data set
19	8F	CE17BN1	Eocene	Renova Fm	Southern Deer Lodge	46.1597	-112.762	MBMG geochron data archive: Yakovlev and Elliott, Butte North data set
20	8I	17-GMCP-1	Paleogene	Renova Fm	Deep Creek	46.0306	-113.014	MBMG geochron data archive: Yakovlev and Elliott, Butte North data set
22	8J	FC (French Creek)	Oligocene	Renova Fm	Deep Creek	45.9459	-113.0721	Roe, 2010
23	8K	DCCP-2	Oligocene	Renova Fm (Cabbage Patch mem.)	Deep Creek	45.8023	-112.9228	MBMG geochron data archive

¹Sample names that are followed by additional numbers/letters in parentheses are samples whose analytical names differed from their in-text names. Both are included here for clarity.

Table C1—*Continued.*

24	8L	CE22BHB01	Paleogene	Renova Fm	Big Hole Basin	45.6379	-113.6561	Mosolf and others, 2023
24	8L	CE22BHB02	Paleogene	Renova Fm	Big Hole Basin	45.6379	-113.6561	Mosolf and others, 2023
24	8L	CE20BHB17	Paleogene	Renova Fm	Big Hole Basin	45.6379	-113.6561	Mosolf, and Kylander-Clark, 2023a
25	8M	WNBU (Wisdom N below unc)	Eocene–Miocene (Chadronian–Arikareean)	Renova Fm	Big Hole	45.6520	-113.4378	Roe, 2010
26	8N	BH-12-01	Paleogene	Renova Fm	Big Hole Canyon	45.8023	-112.9228	Schwartz and others, 2019
28	17A	REN-1-08	Oligocene–Miocene (Arikareean)	Renova Fm (Cabbage Patch mem.)	Divide	45.9215	-112.7029	Rothfuss and others, 2012
37	17B	CP2 (2PL04)	Oligocene (Arikareean)	Renova Fm (Cabbage Patch mem.)	Divide	45.7548	-112.7384	Link and others, 2008
39	17G	CLA-6-08	Eocene (Chadronian)	Renova Fm (Climbing Arrow mem.)	Melrose	45.7186	-112.7038	Rothfuss and others, 2012
41	17D	PIO-5-09	Eocene	Renova Fm	Melrose	45.6333	-112.7201	Rothfuss and others, 2012
44	17E	PG-2-08	Eocene (Chadronian)	Renova Fm (Climbing Arrow mem.)	Melrose	45.5661	-112.6809	Rothfuss and others, 2012
45	17C	GLEN-6-09	Eocene (Chadronian)	Renova Fm (Climbing Arrow mem.)	Melrose	45.4715	-112.7174	Schwartz and others, 2019
49	17H	REN09-3	Eocene (Chadronian)	Renova Fm (Climbing Arrow mem.)	Melrose	45.4686	-112.6430	Schwartz and others, 2019
51	17F	CHA-1-08	Eocene (Chadronian)	Renova Fm (Climbing Arrow mem.)	Melrose	45.4589	-112.6021	Rothfuss and others, 2012

Table C2. Igneous single ages: Age, location, and source information for previously published igneous single-mineral age data used in this study.

Sample	Figure	Isotopic System	Crystal Type	Single-Grain Age	Error	Pluton	Lat	Lon	Source
Boulder Batholith									
98BL011_s	5A	Ar/Ar	Ser	74.4	0.3	Boulder Batholith	46.2967	-112.2036	Lund and others, 2002
98BL016	5A	Ar/Ar	Muscovite	77	0.4	Boulder Batholith	46.3164	-112.2369	Lund and others, 2002
98BL010	5A	Ar/Ar	Muscovite	74.4	1.2	Boulder Batholith	46.2839	-112.1431	Lund and others, 2002
98BL011_m	5A	Ar/Ar	Muscovite	73.9	0.7	Boulder Batholith	46.2967	-112.2036	Lund and others, 2002
EQ79-1	5A	K/Ar	Biotite	76.6	2.6	Boulder Batholith, Blackfoot City stock	46.6895	-112.465	Schmidt and others, 1994
GBG79-1h	5A	K/Ar	Hornblende	109.1	4.7	Boulder Batholith, Butte granite	46.7941	-112.158	Schmidt and others, 1994
GBG79-1b	5A	K/Ar	Biotite	77.9	2.6	Boulder Batholith, Butte granite	46.7941	-112.158	Schmidt and others, 1994
FQ79-1	5A	K/Ar	Biotite	79.5	2.6	Boulder Batholith, Dalton Mountain stock	46.7771	-112.6912	Schmidt and others, 1994
EQ79-2	5A	K/Ar	Biotite	78.5	2.5	Boulder Batholith, Marysville granodiorite	46.7367	-112.3119	Schmidt and others, 1994
MCQ79-1h	5A	K/Ar	Hornblende	85.4	3.9	Boulder Batholith, Ogden Mountain stock	46.9169	-112.8419	Schmidt and others, 1994
MCQ79-1b	5A	K/Ar	Biotite	81.1	2.8	Boulder Batholith, Ogden Mountain stock	46.9169	-112.8419	Schmidt and others, 1994
HQ79-1	5A	K/Ar	Hornblende	97	4.2	Boulder Batholith, Scratchgravel Hills stock	46.7158	-112.1053	Schmidt and others, 1994
BB_A	5A	U/Pb	Zircon	78.2	0.5	Boulder Batholith, Unionville pluton	46.5591	-112.5082	Wooden and others, 2008
KCS-15-68	5A	U/Pb	Zircon	80.2	0.8	Boulder Batholith	46.226612	-112.0275	Scarberry and others, 2017
BB_R	5A	U/Pb	Zircon	77.6	0.8	Boulder Batholith, Burton Park pluton	45.7911	-112.4754	Wooden and others, 2008

Table C2—Continued.

Sample	Figure	Isotopic System	Crystal Type	Single-Grain Age	Error	Pluton	Lat	Lon	Source
BB_D	5A	U/Pb	Zircon	74.7	0.6	Boulder Batholith, Butte granite	46.1899	-112.3384	Wooden and others, 2008
BB_M	5A	U/Pb	Zircon	76.4	0.7	Boulder Batholith, Butte granite	45.9238	-112.4256	Wooden and others, 2008
BB_F	5A	U/Pb	Zircon	76.5	0.9	Boulder Batholith, Butte granite	46.0919	-112.4822	Wooden and others, 2008
BB_N	5A	U/Pb	Zircon	76.3	0.8	Boulder Batholith, Climax Gulch pluton	45.8387	-112.5169	Wooden and others, 2008
BB_P	5A	U/Pb	Zircon	74.7	0.4	Boulder Batholith, Donald pluton	45.82	-112.3937	Wooden and others, 2008
BB_O	5A	U/Pb	Zircon	76.9	0.6	Boulder Batholith, granophyric phase of Butte granite	45.8306	-112.4228	Wooden and others, 2008
BB_S	5A	U/Pb	Zircon	73.7	0.5	Boulder Batholith, Hell Canyon pluton	45.6294	-112.3774	Wooden and others, 2008
BB_L	5A	U/Pb	Zircon	75.3	0.7	Boulder Batholith, Homestake pluton	45.9238	-112.4256	Wooden and others, 2008
10378-505	5A	U/Pb	Zircon	74.3	1.1	Boulder Batholith, Modoc biotite igneous breccia	46.0211	-112.5042	Lund and others, 2018
BB_T	5A	U/Pb	Zircon	74.2	0.5	Boulder Batholith, Moose Creek pluton	45.7252	-112.6965	Wooden and others, 2008
BB_C	5A	U/Pb	Zircon	76.3	0.5	Boulder Batholith, Pulpit Rock granite	46.1973	-112.1598	Wooden and others, 2008
BB-Q	5A	U/Pb	Zircon	80.7	0.8	Boulder Batholith, Rader Creek granodiorite	45.8485	-112.3003	Wooden and others, 2008

Table C2—*Continued.*

Sample	Figure	Isotopic System	Crystal Type	Single-Grain Age	Error	Pluton	Lat	Lon	Source
East Pioneer Mountains									
DodgsonCreek_s	5B	FT	Sphene	72.4	4.4	Pioneer Batholith	45.8686	-112.9917	Marvin and others, 1983
PH1m	5B	Ar/Ar	Muscovite	62.9	0.8	Pioneer Batholith two mica	45.7525	-113.3647	Snee, 1982
S1m	5B	Ar/Ar	Muscovite	65.3	0.6	Pioneer Batholith	45.6519	-113.0011	Snee, 1982
M313-1h1	5B	K/Ar	Hornblende	79.8	2.2	Pioneer Batholith	45.5933	-112.9528	Marvin and others, 1983
M313-1h2	5B	K/Ar	Hornblende	80.3	2.3	Pioneer Batholith	45.5933	-112.9528	Marvin and others, 1983
M547-1h	5B	K/Ar	Hornblende	71.9	4.3	Pioneer Batholith	45.5939	-112.9	Marvin and others, 1983
IVPh	5B	K/Ar	Hornblende	69.5	2	Pioneer Batholith	45.5208	-112.8389	Marvin and others, 1983
FGh	5B	K/Ar	Hornblende	70.6	32	Pioneer Batholith	45.4006	-112.8325	Marvin and others, 1983
M121-1h	5B	K/Ar	Hornblende	71.3	4.5	Pioneer Batholith	45.5206	-112.9575	Marvin and others, 1983
M1272-2h	5B	K/Ar	Hornblende	68.8	4.2	Pioneer Batholith	45.545	-112.9767	Marvin and others, 1983
BH9850h	5B	K/Ar	Hornblende	70	2	Pioneer Batholith	45.5883	-112.9486	Marvin and others, 1983
DodgsonCreek_h1	5B	K/Ar	Hornblende	95.9	2.8	Pioneer Batholith	45.8686	-112.9917	Marvin and others, 1983
DodgsonCreek_h2	5B	K/Ar	Hornblende	98.8	2.8	Pioneer Batholith	45.8686	-112.9917	Marvin and others, 1983
D1h	5B	Ar/Ar	Hornblende	75.4	1.1	Pioneer Batholith	45.7622	-112.81	Snee, 1982
D2h	5B	Ar/Ar	Hornblende	78.3	0.5	Pioneer Batholith	45.7611	-112.8061	Snee, 1982
D3h	5B	Ar/Ar	Hornblende	80	0.9	Pioneer Batholith	45.7769	-112.8325	Snee, 1982
D4h	5B	Ar/Ar	Hornblende	87.5	1.9	Pioneer Batholith	45.7781	-112.87	Snee, 1982
P2h	5B	Ar/Ar	Hornblende	80.3	4.6	Pioneer Batholith	45.3206	-113.0853	Snee, 1982
P3h	5B	Ar/Ar	Hornblende	79.9	0.8	Pioneer Batholith	45.3167	-113.0867	Snee, 1982
P5h	5B	Ar/Ar	Hornblende	75.7	1.7	Pioneer Batholith	45.3156	-113.0481	Snee, 1982
P6h	5B	Ar/Ar	Hornblende	74.9	0.2	Pioneer Batholith	45.3375	-113.0294	Snee, 1982
P7ah	5B	Ar/Ar	Hornblende	71.6	0.2	Pioneer Batholith	45.3947	-113.0083	Snee, 1982
P7bh	5B	Ar/Ar	Hornblende	71.4	0.7	Pioneer Batholith	45.3947	-113.0083	Snee, 1982
P9h	5B	Ar/Ar	Hornblende	71	0.5	Pioneer Batholith	45.4306	-113.0108	Snee, 1982
TM12h	5B	K/Ar	Hornblende	75.4	0.9	Pioneer Batholith	45.4675	-112.9336	Snee, 1982
TM16h	5B	Ar/Ar	Hornblende	71.9	0.2	Pioneer Batholith	45.3903	-112.9806	Snee, 1982
VP3h	5B	Ar/Ar	Hornblende	76.1	0.3	Pioneer Batholith	45.5736	-112.9122	Snee, 1982
VP6h	5B	Ar/Ar	Hornblende	74.9	1	Pioneer Batholith	45.5933	-112.9528	Snee, 1982

Table C2—Continued.

Sample	Figure	Isotopic System	Crystal Type	Single-Grain Age	Error	Pluton	Lat	Lon	Source
M313-1b	5B	K/Ar	Biotite	72.6	2.5	Pioneer Batholith	45.5933	-112.9528	Marvin and others, 1983
M547-1b	5B	K/Ar	Biotite	74.3	2.7	Pioneer Batholith	45.5939	-112.9	Marvin and others, 1983
IVPb	5B	K/Ar	Biotite	72.6	2.8	Pioneer Batholith	45.5208	-112.8389	Marvin and others, 1983
FGb	5B	K/Ar	Biotite	70.5	1.7	Pioneer Batholith	45.4006	-112.8325	Marvin and others, 1983
M121-1b	5B	K/Ar	Biotite	71.3	2.4	Pioneer Batholith	45.5206	-112.9575	Marvin and others, 1983
M881-1	5B	K/Ar	Biotite	72	2.5	Pioneer Batholith	45.5267	-112.935	Marvin and others, 1983
M1272-2b	5B	K/Ar	Biotite	72.2	2.6	Pioneer Batholith	45.545	-112.9767	Marvin and others, 1983
BC	5B	K/Ar	Biotite	70.4	2.4	Pioneer Batholith	45.4167	-112.8533	Marvin and others, 1983
BH9850b	5B	K/Ar	Biotite	71.5	2.5	Pioneer Batholith	45.5883	-112.9486	Marvin and others, 1983
BHS	5B	K/Ar	Biotite	72.3	2.5	Pioneer Batholith	45.5872	-112.9517	Marvin and others, 1983
M1293-1	5B	K/Ar	Biotite	74.1	2.7	Pioneer Batholith	45.5778	-112.9278	Marvin and others, 1983
M744-2	5B	K/Ar	Biotite	71.3	2.4	Pioneer Batholith	45.5256	-112.8358	Marvin and others, 1983
M1228-1	5B	K/Ar	Biotite	71.7	2.6	Pioneer Batholith	45.5506	-112.9758	Marvin and others, 1983
M1162-1	5B	K/Ar	Biotite	72.3	2.6	Pioneer Batholith	45.7158	-113.0253	Marvin and others, 1983
M516-1	5B	K/Ar	Biotite	66.8	2.4	Pioneer Batholith	45.6644	-113.00	Marvin and others, 1983
M500-1	5B	K/Ar	Biotite	64.6	2.1	Pioneer Batholith	45.65	-112.995	Marvin and others, 1983
M32-1-4	5B	K/Ar	Biotite	64.9	2.2	Pioneer Batholith two mica	45.6447	-112.9772	Marvin and others, 1983
M984-2	5B	K/Ar	Biotite	77.1	2.8	Pioneer Batholith	45.7428	-112.8717	Marvin and others, 1983
DodgsonCreek_b	5B	K/Ar	Biotite	76.4	2.6	Pioneer Batholith	45.8686	-112.9917	Marvin and others, 1983
A1b	5B	Ar/Ar	Biotite	71.5	0.6	Pioneer Batholith	45.2839	-112.8708	Snee, 1982
D1b	5B	Ar/Ar	Biotite	74.7	1.2	Pioneer Batholith	45.7622	-112.81	Snee, 1982
D2b	5B	Ar/Ar	Biotite	73.9	0.3	Pioneer Batholith	45.7611	-112.8061	Snee, 1982
P10b	5B	Ar/Ar	Biotite	68.3	0.5	Pioneer Batholith	45.4447	-113.0808	Snee, 1982
P12b	5B	Ar/Ar	Biotite	68.8	0.3	Pioneer Batholith	45.485	-113.1967	Snee, 1982
P13b	5B	Ar/Ar	Biotite	66.7	0.2	Pioneer Batholith	45.4594	-113.0422	Snee, 1982
P3b	5B	Ar/Ar	Biotite	75.8	1	Pioneer Batholith	45.3167	-113.0867	Snee, 1982
P5b	5B	Ar/Ar	Biotite	72.7	0.7	Pioneer Batholith	45.3156	-113.0481	Snee, 1982
P6b	5B	Ar/Ar	Biotite	73.4	0.6	Pioneer Batholith	45.3375	-113.0294	Snee, 1982
P7ab	5B	Ar/Ar	Biotite	69.9	0.4	Pioneer Batholith	45.3947	-113.0083	Snee, 1982

Table C2—Continued.

Sample	Figure	Isotopic System	Crystal Type	Single-Grain Age	Error	Pluton	Lat	Lon	Source
P7bb	5B	Ar/Ar	Biotite	69.1	0.7	Pioneer Batholith	45.3947	-113.0083	Snee, 1982
P9b	5B	Ar/Ar	Biotite	68.2	0.4	Pioneer Batholith	45.4306	-113.0108	Snee, 1982
TA1b	5B	Ar/Ar	Biotite	71.1	0.6	Pioneer Batholith	45.4108	-112.8481	Snee, 1982
TA2b	5B	Ar/Ar	Biotite	70.3	0.6	Pioneer Batholith	45.4108	-112.8481	Snee, 1982
TM10b	5B	Ar/Ar	Biotite	68.2	1.6	Pioneer Batholith	45.4678	-112.9367	Snee, 1982
TM11b	5B	Ar/Ar	Biotite	68.7	0.5	Pioneer Batholith	45.4561	-112.9731	Snee, 1982
TM12b	5B	K/Ar	Biotite	68.5	0.9	Pioneer Batholith	45.4675	-112.9336	Snee, 1982
TM14bb	5B	Ar/Ar	Biotite	68.2	1	Pioneer Batholith	45.4633	-112.89	Snee, 1982
TM16b	5B	Ar/Ar	Biotite	69.3	0.4	Pioneer Batholith	45.3903	-112.9806	Snee, 1982
TM1b	5B	Ar/Ar	Biotite	68.9	0.6	Pioneer Batholith	45.4242	-112.9703	Snee, 1982
TM2b	5B	Ar/Ar	Biotite	69.4	1.1	Pioneer Batholith	45.4075	-112.9689	Snee, 1982
TM3apliteb	5B	K/Ar	Biotite	68	0.8	Pioneer Batholith	45.4222	-112.9928	Snee, 1982
TM3b	5B	Ar/Ar	Biotite	67.5	0.5	Pioneer Batholith	45.4222	-112.9928	Snee, 1982
VP1b	5B	K/Ar	Biotite	72.5	1.1	Pioneer Batholith	45.5794	-112.9092	Snee, 1982
VP2b	5B	Ar/Ar	Biotite	71.1	1.7	Pioneer Batholith	45.5769	-112.9144	Snee, 1982
DF98-069	5B	U/Pb	Zircon	73.4	1.3	Pioneer Batholith, Anniversary stock	45.1829	-112.9682	Murphy and others, 2002
DF98-067	5B	U/Pb	Zircon	75.1	1.1	Pioneer Batholith, Grasshopper Creek tuff	45.2337	-112.9373	Murphy and others, 2002
DF98-076	5B	U/Pb	Zircon	72.2	1.7	Pioneer Batholith, Grayling Lake granite	45.5281	-112.8635	Murphy and others, 2002
DF98-070	5B	U/Pb	Zircon	77.1	8.1	Pioneer Batholith, Lower Beaverhead andesite	45.1829	-112.9682	Murphy and others, 2002
DF98-074	5B	U/Pb	Zircon	72.2	0.6	Pioneer Batholith, Uphill Creek granodiorite	45.5243	-112.8508	Murphy and others, 2002
DodgsonCreek_z	5B	FT	Zircon	61.8	9.1	Pioneer Batholith	45.8686	-112.9917	Marvin and others, 1983
West Pioneer Mountains									
FR1h	5C	Ar/Ar	Hornblende	71.8	1.4	Pioneer Batholith	45.8461	-113.1506	Snee, 1982

Table C2—Continued.

Sample	Figure	Isotopic System	Crystal Type	Single-Grain Age	Error	Pluton	Lat	Lon	Source
OL1h	5C	Ar/Ar	Hornblende	70.6	0.3	Pioneer Batholith	45.5081	-113.2033	Snee, 1982
PH2h	5C	Ar/Ar	Hornblende	70.2	0.4	Pioneer Batholith	45.8219	-113.2914	Snee, 1982
PH3h	5C	Ar/Ar	Hornblende	70.4	0.6	Pioneer Batholith	45.8206	-113.2717	Snee, 1982
StM2h	5C	Ar/Ar	Hornblende	70.7	0.4	Pioneer Batholith	45.575	-113.2761	Snee, 1982
SM1h	5C	Ar/Ar	Hornblende	71.9	0.8	Pioneer Batholith	45.7425	-113.1331	Snee, 1982
SM2h	5C	Ar/Ar	Hornblende	72.9	0.7	Pioneer Batholith	45.7411	-113.1358	Snee, 1982
FR1b	5C	Ar/Ar	Biotite	68.8	0.4	Pioneer Batholith	45.8461	-113.1506	Snee, 1982
OL1b	5C	Ar/Ar	Biotite	68.4	0.3	Pioneer Batholith	45.5081	-113.2033	Snee, 1982
OL2b	5C	Ar/Ar	Biotite	66.8	0.4	Pioneer Batholith	45.5869	-113.2425	Snee, 1982
PH1b	5C	Ar/Ar	Biotite	61.3	0.6	Pioneer Batholith two mica	45.7525	-113.3647	Snee, 1982
PH2b	5C	Ar/Ar	Biotite	65.7	0.8	Pioneer Batholith	45.8219	-113.2914	Snee, 1982
PH3b	5C	Ar/Ar	Biotite	65.7	0.3	Pioneer Batholith	45.8206	-113.2717	Snee, 1982
PR1b	5C	Ar/Ar	Biotite	60.6	0.4	Pioneer Batholith	45.7114	-113.3294	Snee, 1982
StM1b	5C	Ar/Ar	Biotite	67.2	0.1	Pioneer Batholith	45.5886	-113.3322	Snee, 1982
StM2b	5C	Ar/Ar	Biotite	66.8	0.3	Pioneer Batholith	45.575	-113.2761	Snee, 1982
SM1b	5C	Ar/Ar	Biotite	69.1	0.5	Pioneer Batholith	45.7425	-113.1331	Snee, 1982
SM2b	5C	Ar/Ar	Biotite	69.1	0.3	Pioneer Batholith	45.7411	-113.1358	Snee, 1982
StM4b	5C	Ar/Ar	Biotite	66	0.7	Pioneer Batholith	45.5953	-113.27	Snee, 1982
MM1b	5C	K/Ar	Biotite	67.3	0.9	Pioneer Batholith	45.5247	-113.0894	Snee, 1982
P11b	5C	Ar/Ar	Biotite	69.1	0.4	Pioneer Batholith	45.4469	-113.1197	Snee, 1982
WR1ab	5C	Ar/Ar	Biotite	76.7	1	Pioneer Batholith	45.8594	-112.9906	Snee, 1982
WR1bb	5C	Ar/Ar	Biotite	75.6	0.9	Pioneer Batholith	45.8594	-112.9906	Snee, 1982
MM1h	5C	K/Ar	Hornblende	72.6	1.2	Pioneer Batholith	45.5247	-113.0894	Snee, 1982
WR1ah	5C	Ar/Ar	Hornblende	89.6	1.4	Pioneer Batholith	45.8594	-112.9906	Snee, 1982
WR1bh	5C	Ar/Ar	Hornblende	92.8	4	Pioneer Batholith	45.8594	-112.9906	Snee, 1982
Flint Creek Range									
DKA1893 (AA4)	5G	K/Ar	Hornblende	72	2.5	Philipsburg Batholith	46.328	-113.2706	Hyndman and others, 1972
DKA1888 (AA4)	5G	K/Ar	Biotite	73.4	2.1	Philipsburg Batholith	46.328	-113.2706	Hyndman and others, 1972
DKA1890 (BB6)	5G	K/Ar	Hornblende	76.7	2.5	Philipsburg Batholith	46.3306	-113.2653	Hyndman and others, 1972

Table C2—Continued.

Sample	Figure	Isotopic System	Crystal Type	Single-Grain Age	Error	Pluton	Lat	Lon	Source
DKA1891 (BB6)	5G	K/Ar	Biotite	74	2.1	Philipsburg Batholith	46.3306	-113.2653	Hyndman and others, 1972
FC-271	5G	K/Ar	Biotite	77.2	1.9	Philipsburg Batholith	46.3144	-113.25	Marvin and others, 1989
FC-272	5G	K/Ar	Biotite	73.3	1.8	Philipsburg Batholith	46.2783	-113.1881	Marvin and others, 1989
H29-14	5G	fiss. Track	Apatite	64.8	1*	Philipsburg Batholith	NR ² - Baty map	NR ² -Baty map	Baty, 1973
H29-9	5G	fiss. Track	Apatite	65.9	1*	Philipsburg Batholith	NR ² - Baty map	NR ² -Baty map	Baty, 1973
S30-5	5G	fiss. Track	Apatite	68.2	1*	Philipsburg Batholith	NR ² - Baty map	NR ² -Baty map	Baty, 1973
AA-4	5G	fiss. Track	Apatite	73	1*	Philipsburg Batholith	NR ² - Baty map	NR ² -Baty map	Baty, 1973
BB-7	5G	fiss. Track	Apatite	75	1*	Philipsburg Batholith	NR ² - Baty map	NR ² -Baty map	Baty, 1973
FC-263	5G	K/Ar	Biotite	64.7	1.6	Royal stock	46.4083	-113.0944	Marvin and others, 1989
H23-13	5G	fiss. Track	Apatite	60.1	1*	Royal stock	NR ² - Baty map	NR ² -Baty map	Baty, 1973
H23-8	5G	fiss. Track	Apatite	65.4	1*	Royal stock	NR ² - Baty map	NR ² -Baty map	Baty, 1973
S31-10	5G	fiss. Track	Apatite	65.9	1*	Royal stock	NR ² - Baty map	NR ² -Baty map	Baty, 1973
S1-1	5G	fiss. Track	Apatite	66.7	1*	Royal stock	NR ² - Baty map	NR ² -Baty map	Baty, 1973
H2H-1	5G	fiss. Track	Apatite	66.9	1*	Royal stock	NR ² - Baty map	NR ² -Baty map	Baty, 1973
H31-1	5G	fiss. Track	Apatite	69.3	1*	Royal stock	NR ² - Baty map	NR ² -Baty map	Baty, 1973
S28-4	5G	fiss. Track	Apatite	70.8	1*	Royal stock	NR ² - Baty map	NR ² -Baty map	Baty, 1973
H7-9	5G	fiss. Track	Apatite	50.7	1*	Mt Powell Batholith	NR ² - Baty map	NR ² -Baty map	Baty, 1973
S9-2	5G	fiss. Track	Apatite	56.4	1*	Mt Powell Batholith	NR ² - Baty map	NR ² -Baty map	Baty, 1973
H6-R	5G	fiss. Track	Apatite	62	1*	Mt Powell Batholith	NR ² - Baty map	NR ² -Baty map	Baty, 1973
H10-5	5G	fiss. Track	Apatite	62.9	1*	Mt Powell Batholith	NR ² - Baty map	NR ² -Baty map	Baty, 1973
H10-8	5G	fiss. Track	Apatite	63.6	1*	Mt Powell Batholith	NR ² - Baty map	NR ² -Baty map	Baty, 1973

Table C2—Continued.

Sample	Figure	Isotopic System	Crystal Type	Single-Grain Age	Error	Pluton	Lat	Lon	Source
H10-2	5G	fiss. Track	Apatite	65.4	1*	Mt Powell Batholith	NR ² - Baty map	NR ² -Baty map	Baty, 1973
H14-2	5G	fiss. Track	Apatite	65.7	1*	Mt Powell Batholith	NR ² - Baty map	NR ² -Baty map	Baty, 1973
S16-7a	5G	fiss. Track	Apatite	69.7	1*	Mt Powell Batholith	NR ² - Baty map	NR ² -Baty map	Baty, 1973
FC-273m	5G	K/Ar	Muscovite	61.5	1*	Mt Powell Batholith	46.3064	-113.0044	Marvin and others, 1989
FC-273b	5G	K/Ar	Biotite	59.7	1.4	Mt Powell Batholith	46.3064	-113.0044	Marvin and others, 1989
Anaconda Range									
E1048_b	5F	K/Ar	Biotite	78.7	0.8	Storm Lake Stock	46.0734	-113.2554	Wallace and others, 1992
WG04-114	5F	Ar/Ar	Biotite	74	0.9	Storm Lake Stock	46.0736	-113.2453	Foster and others, 2010
WG04-052	5F	Ar/Ar	Biotite	64.1	1	Storm Lake Stock	46.0858	-113.2294	Foster and others, 2010
SL-38	5F	Ar/Ar	Biotite	79	1.2	Storm Lake Stock	46.1017	-113.2756	Foster and others, 2010
81540_m	5F	K/Ar	Muscovite	54.9	0.6	Pintler Creek Batholith	45.7974	-113.6979	Wallace and others, 1992
81642_m	5F	K/Ar	Muscovite	49.9	0.5	Pintler Creek Batholith	45.8228	-113.6663	Wallace and others, 1992
81540_b	5F	K/Ar	Biotite	53.1	0.8	Pintler Creek Batholith	45.7974	-113.6979	Wallace and others, 1992
81642_b	5F	K/Ar	Biotite	50.7	0.5	Pintler Creek Batholith	45.8228	-113.6663	Wallace and others, 1992
DF02-120	5F	Ar/Ar	Biotite	38.5	0.7	Hearst Lake granite	46.0569	-113.0272	Foster and others, 2010
DF02-118B	5F	Ar/Ar	Biotite	38.6	0.6	Anaconda MCC, intrusives	46.0569	-113.0273	Foster and others, 2012
DF02-116a_b1	5F	Ar/Ar	Biotite	39.6	2.3	Hearst Lake granite	46.0914	-113.0347	Foster and others, 2010
DF02-116a_m	5F	Ar/Ar	Muscovite	40.5	2	Hearst Lake granite	46.0914	-113.0347	Foster and others, 2010
DF02-116Ab	5F	Ar/Ar	Biotite	40.7	0.6	Anaconda MCC, intrusives	46.0913	-113.0348	Foster and others, 2012
DF02-116a_b2	5F	Ar/Ar	Biotite	40.7	0.7	Hearst Lake granite	46.0914	-113.0347	Foster and others, 2010
DF02-116Am	5F	Ar/Ar	Muscovite	41.3	1.9	Anaconda MCC, intrusives	46.0913	-113.0348	Foster and others, 2012
WG04-089	5F	Ar/Ar	Muscovite	45.5	0.6	Hearst Lake granite	46.0489	-113.1383	Foster and others, 2010
WG04-092	5F	Ar/Ar	Biotite	47.1	0.7	Hearst Lake granite	46.0594	-113.1403	Foster and others, 2010
02-120	5F	U/Pb	Zircon	47.1	0.4	Anaconda MCC, intrusives	46.0569	-113.0273	Foster and others, 2012
WG04-109b	5F	Ar/Ar	Biotite	47.6	0.6	Hearst Lake granite	46.0431	-113.1786	Foster and others, 2010
WG04-109m	5F	Ar/Ar	Muscovite	48	0.6	Hearst Lake granite	46.0431	-113.1786	Foster and others, 2010

Table C2—Continued.

Sample	Figure	Isotopic System	Crystal Type	Single-Grain Age	Error	Pluton	Lat	Lon	Source
E0134_b	5F	K/Ar	Biotite	48.5	0.5	Hearst Lake granite	46.0216	-113.2372	Wallace and others, 1992
ME-6b	5F	Ar/Ar	Biotite	48.6	1	Hearst Lake granite	46.0403	-113.1708	Foster and others, 2010
WG04-101b	5F	Ar/Ar	Biotite	49.2	0.9	Hearst Lake granite	46.0608	-113.1669	Foster and others, 2010
WG04-101m	5F	Ar/Ar	Muscovite	49.6	0.6	Hearst Lake granite	46.0608	-113.1669	Foster and others, 2010
WG04-100	5F	Ar/Ar	Muscovite	50.4	1.2	Hearst Lake granite	46.06	-113.1644	Foster and others, 2010
Ug-1	5F	Ar/Ar	Biotite	50.5	0.8	Hearst Lake granite	46.0756	-113.1486	Foster and others, 2010
E0134_m	5F	K/Ar	Muscovite	50.6	0.5	Hearst Lake Batholith	46.0216	-113.2372	Wallace and others, 1992
ME-6m	5F	Ar/Ar	Muscovite	51.5	1.2	Hearst Lake granite	46.0403	-113.1708	Foster and others, 2010
02-117A	5F	U/Pb	Zircon	53	0.6	Anaconda MCC, intrusives	46.0917	-113.0349	Foster and others, 2007
81639	5F	K/Ar	Biotite	53.1	0.6	Elk Park stock	45.8691	-113.5118	Wallace and others, 1992

*NR, not reported.

Table C3. Igneous multiple-age samples: Location and source information for pluton samples with multiple zircon ages.

Sample	Figure	Isotopic System	Crystal Type	Number of Grains ¹	Pluton	Lat	Lon	Source
Boulder Batholith								
Spl 03KL067	5A	U/Pb	Zircon	14	Butte granite Homestake Pass	45.923	-112.417	duBray and others, 2012
spl 03KL068	5A	U/Pb	Zircon	15	Butte granite mining district	46.039	-112.532	duBray and others, 2012
Spl 98BL005	5A	U/Pb	Zircon	16	Unionville granodiorite	46.5125	-112.2597	duBray and others, 2012
Spl 98BL004	5A	U/Pb	Zircon	19	Butte granite, Bison Creek	46.1917	-112.3444	duBray and others, 2012
Spl 04KL057	5A	U/Pb	Zircon	19	Burton Park granodiorite	45.8193	-112.5203	duBray and others, 2012
Spl 03KL072	5A	U/Pb	Zircon	15	Granodiorite Climax Gulch granodiorite	46.5819	-112.2727	duBray and others, 2012
Spl 04KL059	5A	U/Pb	Zircon	12	Moose Creek monzogranite	45.7394	-112.2727	duBray and others, 2012
03KL066	5A	U/Pb	Zircon	17	Homestake monzogranite	45.9211	-112.4209	duBray and others, 2012
03KL069	5A	U/Pb	Zircon	15	Granodiorite of Donald	45.8493	-112.4277	duBray and others, 2012
Spl 03KL071	5A	U/Pb	Zircon	10	Deer Creek granophyre	45.8595	-112.4647	duBray and others, 2012
04KL058	5A	U/Pb	Zircon	15	Hell Canyon monzogranite	46.6353	-112.3435	duBray and others, 2012
98BL002	5A	U/Pb	Zircon	17	Rader Creek Granodiorite	45.8389	-112.31129	duBray and others, 2012
98BL007		U/Pb	Zircon	12	Pulpit Rock monzogranite	46.196	-112.133	duBray and others, 2012
McCartney Mtn								
8-072	5D	U/Pb	Zircon	8	McCartney stock, granodiorite/microdiorite	45.5194	-112.5990	Foster and others 2012
East Pioneer Mtns								
8-1	5B	U/Pb	Zircon	10	Keokirk quartz diorite	45.6703	-112.9526	Foster and others 2012
8-074	5B	U/Pb	Zircon	15	Uphill Creek pluton, granodiorite	45.4843	-112.8851	Foster and others 2012
8-076	5B	U/Pb	Zircon	20	Browns Lake pluton, granite	45.5249	-112.8332	Foster and others 2012
KCS2129	5B	U/Pb	Zircon	28	Diorite	45.5084	-113.0403	MBMG geochron data archive: Molsof and Kylander-Clark, 2022
KCS2136	5B	U/Pb	Zircon	17	Granite	45.5483	-113.3631	MBMG geochron data archive: Molsof and Kylander-Clark, 2022
JM21TM01	5B	U/Pb	Zircon	10	Tonalite	45.4559	-112.9730	MBMG geochron data archive: Molsof and Kylander-Clark, 2022
JM21TM02	5B	U/Pb	Zircon	28	Lucogranite	45.4450	-112.9653	MBMG geochron data archive: Molsof and Kylander-Clark, 2022
JM21TM03	5B	U/Pb	Zircon	29	Granite	45.4805	-112.9655	MBMG geochron data archive: Molsof and Kylander-Clark, 2022
JM21TM04	5B	U/Pb	Zircon	29	Aplite	45.4783	-112.9501	MBMG geochron data archive: Molsof and Kylander-Clark, 2022
JM21TM06	5B	U/Pb	Zircon	19	Granite	45.4823	-112.9454	MBMG geochron data archive: Molsof and Kylander-Clark, 2022
KCS2110	5B	U/Pb	Zircon	29	Quartz porphyry	45.5226	-113.0548	MBMG geochron data archive: Molsof and Kylander-Clark, 2022
KCS2129	5B	U/Pb	Zircon	28	Diorite	45.5084	-113.0403	MBMG geochron data archive: Molsof and Kylander-Clark, 2022
KCS2130	5B	U/Pb	Zircon	30	Granodiorite	45.5095	-113.0386	MBMG geochron data archive: Molsof and Kylander-Clark, 2022
West Pioneer Mtns								
CCG	5C	U/Pb	Zircon	7	Clifford Creek two-mica granite	45.6814	-113.0116	Roe, 2010
NG	5C	U/Pb	Zircon	3	Nichols Granodiorite	45.8219	-113.2902	Roe, 2010
CE18FH13A	5C	U/Pb	Zircon	42	Foolhen granodiorite	45.7900	-113.1500	MBMG geochron data archive: Fool Hen Quad
CE18FH14	5C	U/Pb	Zircon	44	Foolhen granodiorite	45.7900	-113.1900	MBMG geochron data archive: Fool Hen Quad
KCS2111	5C	U/Pb	Zircon	12	Tonalite	45.7239	-113.0363	MBMG geochron data archive: Molsof and Kylander-Clark, 2022
KCS2112	5C	U/Pb	Zircon	65	Granodiorite	45.6494	-113.1800	MBMG geochron data archive: Molsof and Kylander-Clark, 2022
KCS2134	5C	U/Pb	Zircon	28	Granite	45.5400	-113.3565	MBMG geochron data archive: Molsof and Kylander-Clark, 2022
KCS2136	5C	U/Pb	Zircon	17	Granite	45.5483	-113.3631	MBMG geochron data archive: Molsof and Kylander-Clark, 2022
KCS-20-04	5C	U/Pb	Zircon	29	Granodiorite	45.5752	-113.2352	MBMG geochron data archive: O'Dell Lake Quad
KCS-20-12	5C	U/Pb	Zircon	29	Granodiorite	45.5239	-113.2118	MBMG geochron data archive: O'Dell Lake Quad
KCS-20-19	5C	U/Pb	Zircon	30	Granodiorite	45.5119	-113.1799	MBMG geochron data archive: O'Dell Lake Quad
CE19PH-11a	5C	U/Pb	Zircon	15	Two-mica granite	45.7651	-113.2419	MBMG geochron data archive: Pine Hill Quad
CE19PH-7	5C	U/Pb	Zircon	10	Two-mica granite	45.7519	-113.3650	MBMG geochron data archive: Pine Hill Quad

Table C3. Igneous multiple-age samples: Location and source information for pluton samples with multiple zircon ages.

Sample	Figure	Isotopic System	Crystal Type	Number of Grains ¹	Pluton	Lat	Lon	Source
Dickie Hills								
CE23W100k-08_101	8K	U/Pb	Zircon	40	Granodiorite	45.8825	-113.0014	Brennan and others, 2025
Anaconda Range								
UG-1	5F	U/Pb	Zircon	22	Storm Lake quartz diorite	46.0756	-113.1486111	Grice, 2006
CH-1	5F	U/Pb	Zircon	19	Pintler Creek two-mica granite	46.8600	-113.4400	Howlett and others, 2021
CH-8	5F	U/Pb	Zircon	17	Pintler Creek two-mica granite	45.8610	-113.4460	Howlett and others, 2021
5-4	5F	U/Pb	Zircon	22	Dickie Creek pluton, granodiorite	45.8015	-113.3666	Foster and others 2012
Flint Creek Range (FCR)								
PB-BM	5G	U/Pb	Zircon	9	Phillipsburg Bath-Bimetallic pluton, granodiorite	46.3169	-113.2502778	Gray and others, 2022
PB-DT	5G	U/Pb	Zircon	9	Phillipsburg BathDora Thorn pluton, granodiorite	46.3669	-113.1336111	Gray and others, 2022
1620	5G	U/Pb	Zircon	45	Phillipsburg BathStewart Gulch granitic porphyry	NR ²	NR ²	MBMG geochron data archive: Dudas Flint Creek data
18LC01, location 1	5G	U/Pb	Zircon	69	Lost Creek pluton-granite	NR ²	NR ²	MBMG geochron data archive: Dudas Flint Creek data
17LC08, location 3	5G	U/Pb	Zircon	21	Lost Creek pluton-granite	46.2065	112.9909	MBMG geochron data archive: Dudas Flint Creek data
LC10, Location 1	5G	U/Pb	Zircon	45	Mt Powell Bath-two-mica granite	NR ²	NR ²	MBMG geochron data archive: Dudas Flint Creek data
LC08, location 6	5G	U/Pb	Zircon	31	Mt Powell Bath-two-mica granite	NR ²	NR ²	MBMG geochron data archive: Dudas Flint Creek data
FC01, Location 10	5G	U/Pb	Zircon	30	Mt Powell Bath-two-mica granite	46.2196	113.1194	MBMG geochron data archive: Dudas Flint Creek data
17FC01, Location 10	5G	U/Pb	Zircon	97	Mt Powell Bath-two-mica granite	46.2196	113.1194	MBMG geochron data archive: Dudas Flint Creek data
17FC02, Location 10	5G	U/Pb	Zircon	57	Mt Powell Bath-two-mica granite	46.2196	113.1194	MBMG geochron data archive: Dudas Flint Creek data
17LC04, Location 6	5G	U/Pb	Zircon	64	Mt Powell Bath-two-mica granite	NR ²	NR ²	MBMG geochron data archive: Dudas Flint Creek data
FC-RT	5G	U/Pb	Zircon	15	Mt Powell Bath-monozodiorite	46.2669	-112.9669444	Gray and others, 2022
FC-MP	5G	U/Pb	Zircon	19	Mt Powell Bath-two-mica granodiorite	46.2669	-112.9669444	Gray and others, 2022
N. Beaverhead Mountains-Anaconda Range Junctionure								
AH21WW12	5E	U/Pb	Zircon	40	Trail Creek Pluton two-mica granite	45.8358	-113.9855	Mosolf and Kylander-Clark, 2023b
CE20BHB15	5E	U/Pb	Zircon	40	Trail Creek Pluton two-mica granite	45.6519	-113.7086	Mosolf and Kylander-Clark, 2023a

APPENDIX D
DETRITAL ZIRCON PLOTS

