

## Tectonics®

## RESEARCH ARTICLE

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## Key Points:

- Multi-chronometer time-temperature inversions resolve key aspects of the >2.5 Gyr thermal history of the Beartooth Mountains
- $^{40}\text{Ar}/^{39}\text{Ar}$  and [U-Th(-Sm)]/He data constrain newly recognized Proterozoic and Paleozoic thermal events
- Models suggest complex, spatially variable cooling associated with basement-involved thrusting initiating >90 Ma

## Supporting Information:

Supporting Information may be found in the online version of this article.

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## Orogens of Big Sky Country: Reconstructing the Deep-Time Tectonothermal History of the Beartooth Mountains, Montana and Wyoming, USA

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**Abstract** Archean rocks exposed in the Beartooth Mountains, Montana and Wyoming, have experienced a complex >2.5 Gyr thermal history related to the long-term geodynamic evolution of Laurentia. We constrain this history using “deep-time” thermochronology, reporting zircon U-Pb, biotite  $^{40}\text{Ar}/^{39}\text{Ar}$ , and zircon and apatite [U-Th(-Sm)]/He results from three transects across the basement-core of the range. Our central transect yielded a zircon U-Pb concordia age of  $2,805.6 \pm 6.4$  Ma. Biotite  $^{40}\text{Ar}/^{39}\text{Ar}$  plateau ages from western samples are  $\leq 1,775 \pm 27$  Ma, while those from samples further east are  $\geq 2,263 \pm 76$  Ma. Zircon (U-Th)/He dates span  $686.4 \pm 11.9$  to  $13.5 \pm 0.3$  Ma and show a negative relationship with effective uranium—a proxy for radiation damage. Apatite (U-Th)/He dates are  $109.2 \pm 23.9$  to  $43.6 \pm 1.9$  Ma and correlate with sample elevation. Multi-chronometer Bayesian time-temperature inversions suggest: (a) Cooling between  $\sim 1.90$  and  $\sim 1.80$  Ga, likely related to Big Sky orogeny thermal effects; (b) Reheating between  $\sim 1.80$  Ga and  $\sim 1.35$  Ga consistent with Mesoproterozoic burial; (c) Cooling to  $\leq 100^\circ\text{C}$  between Mesoproterozoic and early Paleozoic time, likely reflecting continental erosion; (d) Variable Paleozoic–Jurassic cooling, possibly related to Paleozoic tectonism and/or low eustatic sea level; (e) Rapid Cretaceous–Paleocene cooling, preceding accepted proxies for flat-slab subduction; (f) Eocene–Miocene reheating consistent with reburial by Cenozoic volcanics and/or sediments; (g) Post-20 Ma cooling consistent with Neogene development of topographic relief. Our results emphasize the utility of multi-chronometer thermochronology in recovering complex, non-monotonic multi-billion-year thermal histories.

**Plain Language Summary** The Beartooth Mountains in southern Montana and northern Wyoming consist of some of the oldest rocks in North America. These rocks have been exhumed to Earth's surface and reburied numerous times. However, sedimentary rocks recording many of these events have been eroded, erasing much of the primary record of the geologic history of this region. To recover this history, we use minerals within existing rocks that act as heat-sensitive clocks. These “clocks” record a cooling date, reflecting the last time the rock was within a specific temperature range. We measure cooling dates for several different types of minerals and model the history of temperature fluctuations they record, corresponding to burial and erosion events. We interpret the results of these models by comparing them with the geologic history recorded in nearby preserved sedimentary rocks. We find that the Beartooth Mountains were involved in mountain-building events not previously recognized and cooling associated with building of the modern Rocky Mountains began earlier, and is more complex, than is generally recognized. The ability to recover such long histories of temperature fluctuations is important, suggesting we may be able to learn about events far in Earth's past even when the rock record is missing.

## 1. Introduction

The Beartooth Mountains of southern Montana and northern Wyoming primarily consist of  $\sim 3.5$ – $2.8$  Ga basement rocks of the Wyoming craton (e.g., Mueller & Frost, 2006; Mueller et al., 2008; Wooden et al., 1988). This region, located  $\sim 100$  km southeast of the northwest craton margin (Figure 1), has been involved in tectonic events spanning Archean to Cenozoic time (e.g., Chamberlain et al., 2003; Dickinson & Snyder, 1978; Mogk et al., 1992). Precambrian basement in the Beartooth Mountains was exhumed to the surface as an intraforeland uplift associated with the development of the North American Cordillera in late Mesozoic–early Cenozoic time

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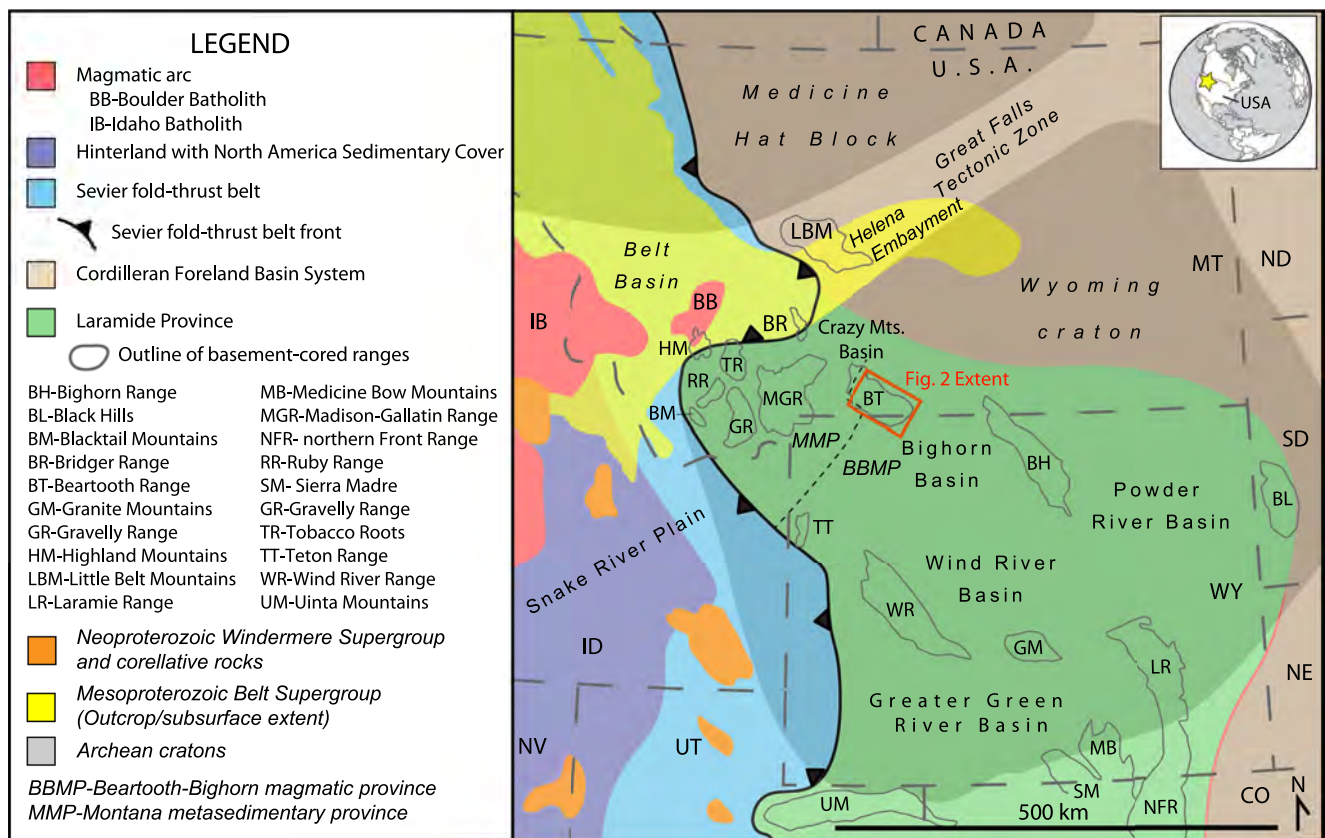
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(e.g., Dickinson & Snyder, 1978). Consequently, the Beartooth Mountains provide not only an opportunity to interrogate the timing and nature of this exhumation event, but also a window into the Archean to recent geodynamic evolution of western Laurentia, where basement rock is commonly inaccessible due to Phanerozoic burial.

Previous application of high- to medium-temperature thermochronometers in the Beartooth Mountains typically yield Archean–early Paleoproterozoic ages (e.g., Gast et al., 1958; Wooden et al., 1988), while conventional low-temperature thermochronology has been used mainly to explore late Mesozoic–early Cenozoic thermal events (e.g., Carrapa et al., 2019; Cervený, 1990; Omar et al., 1994; Peyton et al., 2012). These disparate data sets comprise snapshots into a vastly more complex thermal history which remains largely unresolved. However, recent work has demonstrated the utility of multi-chronometer thermal history inversions—coupling the broad temperature sensitivity of the zircon (U-Th)/He system with additional information from medium-temperature (i.e.,  $^{40}\text{Ar}/^{39}\text{Ar}$  methods) and/or other low-temperature (that is, apatite (U-Th-Sm)/He (AHe) and apatite fission track (AFT)) systems—in constraining nearly continuous thermal histories below  $\sim 400^\circ\text{C}$  (see review in McDannell & Flowers, 2020). This approach enables the recovery of ancient thermal events in regions where other evidence of these events is overprinted or eroded (e.g., Baughman & Flowers, 2018, 2020; DeLucia et al., 2018; Krob et al., 2019; McDannell et al., 2018, 2019, 2022). Here, we apply this approach to investigate the deep-time thermal history of the Beartooth Mountains. Key tectonic/geodynamic questions of open debate are testable using an integrated thermal history of this study area. These include:

1. What was the extent and nature of Paleoproterozoic thermotectonism in southwest Montana? Paleoproterozoic ( $\sim 1.8$ – $1.7$  Ga) thermal overprinting in the region has long been recognized, with early work suggesting a maximum extent  $\sim 100$  km west of the Beartooth Mountains (Giletti, 1966). However, subsequent work revealed partial resetting of medium-temperature thermochronometers east of this transitional boundary (e.g., Montgomery & Lytwyn, 1984) and a complex spatiotemporal pattern of tectonism (Condit et al., 2015).
2. How did Proterozoic rifting of the western Laurentian margin influence the thermal history of cratonic rocks to the east? The western margin of Laurentia experienced rifting events in early–middle Mesoproterozoic and middle–late Neoproterozoic time (e.g., Lonn et al., 2020; Yonkee et al., 2014). Previous studies have suggested the Neoproterozoic event caused erosional exhumation and cooling of cratonic rocks east of the rifting margin (e.g., Kaempfer et al., 2021; Peak et al., 2021). However, other authors have shown late Neoproterozoic cooling was likely not restricted to regions near the Laurentian margin, instead favoring Cryogenian glaciation as the dominant driver of continental exhumation in this interval (McDannell et al., 2022). Additional data from regions near the margin of Laurentia are essential to increase the spatiotemporal resolution of the thermal history of the craton, potentially helping constrain the thermal effects of rifting and driver(s) of Precambrian exhumation of the Laurentian continent.
3. What is the Paleozoic burial and uplift history of the Beartooth Mountains? Sedimentologic and stratigraphic studies in the 1990s suggested the Beartooth Mountains comprised a paleohigh during late Paleozoic time, potentially reflecting uplift associated with the Antler and/or Ancestral Rocky Mountains (ARM) orogenic events (Dorobek et al., 1991; Maughan, 1990, 1993). Despite this sedimentologic support, recent thermochronologic investigation has failed to resolve Paleozoic cooling associated with this hypothesized episode of tectonism (e.g., Carrapa et al., 2019; Peyton et al., 2012).
4. What was the timing of late Mesozoic–early Cenozoic intraforeland basement uplift? Interest in the timing of basement deformation in southwest Montana has recently been renewed following the publication of thermochronologic and sedimentologic evidence for exhumation as early as  $\sim 100$  Ma (Carrapa et al., 2019; Garber et al., 2020; summarized by Orme, 2020). If this is the case, basement exhumation significantly preceded evidence for flat-slab subduction in the region, provoking reconsideration of many existing geodynamic models for intraforeland basement uplift in the North American Cordillera (e.g., Bird, 1984, 1998; Carrapa et al., 2019; Dickinson & Snyder, 1978; Parker & Pearson, 2021). Although the Beartooth Mountains figure centrally in this debate, existing estimates for the timing of cooling initiation vary wildly between studies (that is,  $\geq 110$  Ma from Carrapa et al. (2019) versus  $\leq 65$  Ma from Omar et al. (1994)) and no study has yet been able to reconcile these disparate results.
5. When and why did the Laramide ranges acquire their modern relief? Early thermochronologic work on the Beartooth Mountains suggested the development of significant topographic relief during Miocene–early Pliocene time but was unable to differentiate late-stage erosion of basement versus Oligocene–Miocene reburial and subsequent basin evacuation (Omar et al., 1994). Work in the neighboring Big Horn and Wind River ranges largely support the latter interpretation (e.g., Caylor & Carrapa, 2021; McKenna, 1980; McKenna &



**Figure 1.** Tectonic map of the Northern Rocky Mountains region, showing the distribution of major tectonic features. The study area (orange extent box) in the Beartooth Mountains lies in the northwestern portion of the Archean Wyoming craton and in the northernmost portion of the Laramide Province, as classically defined (e.g., Dickinson & Snyder, 1978). Dashed line divides provinces of the Wyoming craton (e.g., Mogk et al., 1992). Modified from Ronemus et al. (2020) after Whitmeyer and Karlstrom (2007) and Yonkee and Weil (2015). Other tectonic elements adapted from: Basement-cored ranges—Peyton et al. (2012), Belt Supergroup—Lonn et al. (2020), and Vuke et al. (2007), Windermere Supergroup—Brennan et al. (2020).

Love, 1972; Steidtmann & Middleton, 1991; Steidtmann et al., 1989), but this issue has only been cursorily reexamined in the Beartooth Mountains (Carrapa et al., 2019).

We address these questions using a newly reported data set of zircon U-Pb, biotite  $^{40}\text{Ar}/^{39}\text{Ar}$ , ZHe, and AHe geo- and thermochronologic data from samples collected along three transects in the Beartooth Mountains. We input information from these and previously reported data sets in a series of Bayesian QTQt time-temperature ( $t$ - $T$ ) inversions to explore the thermal evolution of the range. In turn, we integrate results of  $t$ - $T$  inversions with structural, sedimentologic, and stratigraphic observations in a kinematic evolution model capable of explaining observed spatiotemporal trends in the Phanerozoic thermal history.

Collectively, our results indicate Paleoproterozoic tectonism affected at least the western Beartooth Mountains, leading to  $\leq 1.8$  Ga biotite  $^{40}\text{Ar}/^{39}\text{Ar}$  ages. Models suggest late Paleoproterozoic to early Mesoproterozoic reheating—possibly reflecting burial by unpreserved Mesoproterozoic sediments—followed by cooling to  $<200^\circ\text{C}$  by the end of Mesoproterozoic time. Model resolution in Neoproterozoic time is generally poor, but our best-resolved models suggest an episode of cooling initiating  $800 \pm 200$  Ma. Models ubiquitously indicate reheating during early Paleozoic time, consistent with burial by passive margin deposits, and suggest at least one previously unresolved later Paleozoic to Mesozoic cooling event. Modeled spatially variable, rapid cooling in middle Cretaceous–Paleogene time reflects the complex evolution of tectonic exhumation likely associated with both foreland basin flexure and compressional stresses linked to retroarc shortening. Several transects suggest Cenozoic reheating coeval with Eocene volcanism and/or Oligocene–early Miocene intermontane basin sedimentation. Final cooling of these samples occurred post-20 Ma, suggesting Neogene basin evacuation and development of intra-foreland topographic relief. This study demonstrates multi-chronometer deep-time thermochronology

can reconstruct broadly continuous Precambrian-to-recent thermal histories, even in regions that have experienced significant Phanerozoic burial and tectonism.

## 2. Geologic Setting and Tectonic History

### 2.1. Geographic and Structural Setting

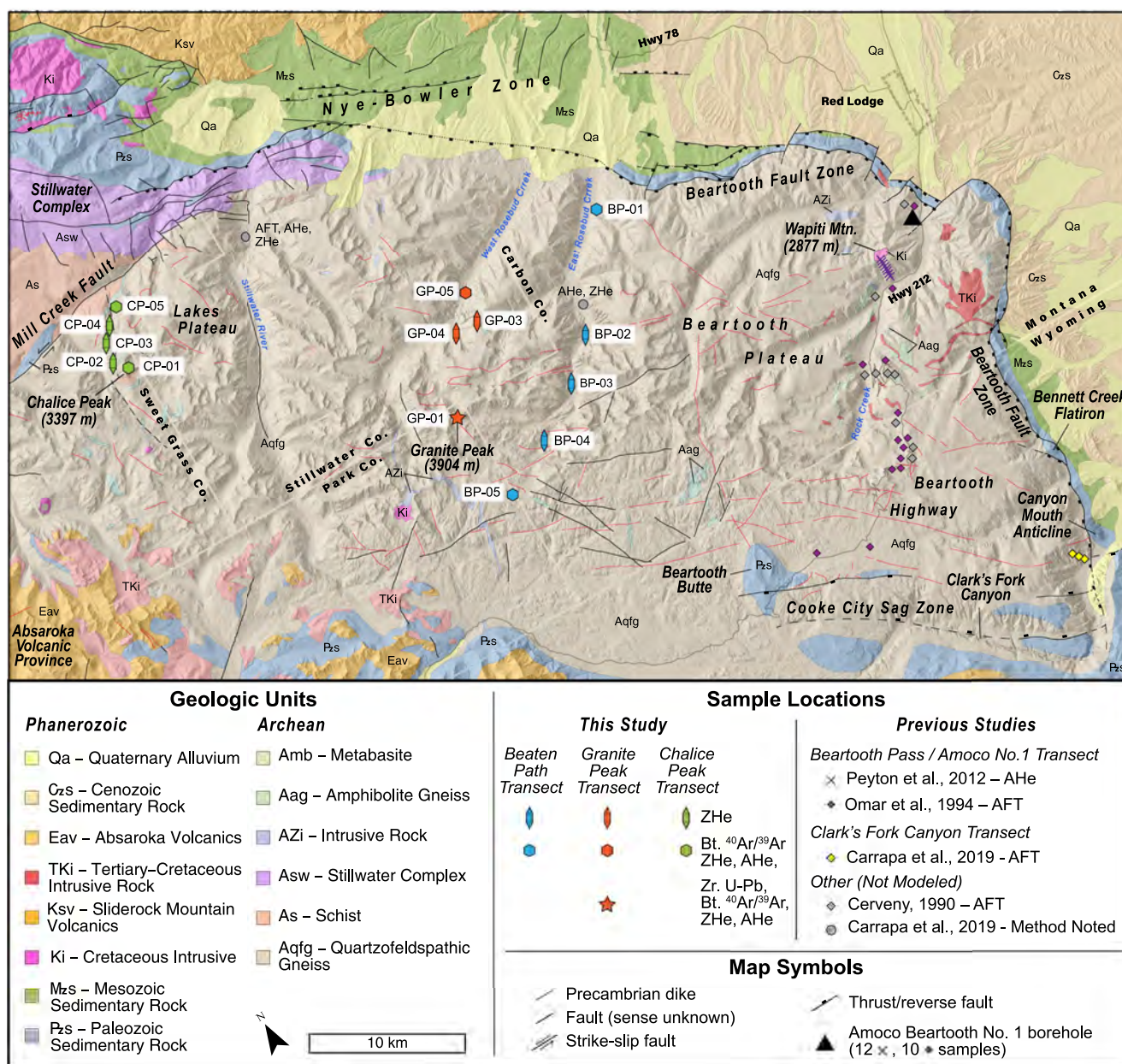
The Beartooth Mountains are a ~8,250 km<sup>2</sup> northwest-southeast trending mountain range in southern Montana and northernmost Wyoming (Figure 1; e.g., Foose et al., 1961). The range stands with  $\geq 2.5$  km of topographical prominence over bordering basins—the Bighorn Basin to the east and southeast, the Crazy Mountain Basin to the north, and the Paradise Valley to the west (Figures 1 and 2). The southern and southwestern regions of the range are buried by the Eocene Absaroka Volcanic Supergroup (Figure 2; Foose et al., 1961). Precambrian rocks in the Beartooth Mountains are divided into four main structural blocks distinguished by differing tectonic histories and lithologies: (a) the main Beartooth block, dominantly orthogneisses comprising the principal structural component of the range, including the study area; (b) the Stillwater mafic layered intrusive complex, exposed along the northeastern margin of the range; and (c) the North Snowy and (d) South Snowy blocks, consisting of rocks of the Montana metasedimentary terrane exposed in northwestern and southwestern portions of the range, respectively (Figure 2; Foose et al., 1961; Mogk et al., 1988, 1992). The eastern Beartooth Mountains form a broad plateau surface (i.e., the Beartooth Plateau)—with an average elevation of  $>3,000$  m (Figure 2).

### 2.2. Precambrian Tectonic History of the Beartooth Mountains and Surrounding Regions

The Beartooth Mountains are volumetrically dominated by latest Mesoarchean–Neoproterozoic (~2.9–2.8 Ga) granitoids and gneisses (Figure 2; Wooden et al., 1988). These and similar rocks comprise the Archean Beartooth-Bighorn magmatic province (BBMP), a major sub-province of the Wyoming craton (Figure 1; Mogk et al., 1992; Mueller et al., 2008). Paleo–Mesoarchean (3.5–3.0 Ga) gneissic rocks of dominantly tonalitic to granodioritic to trondhjemitic composition (TTG suite) are preserved throughout the range as meter-scale xenoliths to km-scale pendants within the 2.9–2.8 Ga suite (Mueller et al., 2008, 2010; Wooden et al., 1988). The Archean metamorphic history experienced by these rocks has been subject to extensive study and is summarized by Wooden et al. (1988) and Mueller et al. (2008).

During Paleoproterozoic time, ~1.86–1.72 Ga thermotectonism associated with the amalgamation of Laurentia impacted basement rocks of the Great Falls tectonic zone and Montana metasedimentary province—north and west of the study area, respectively (Figure 1; e.g., Condit et al., 2015; Gifford et al., 2014, 2020; Giletti, 1966; Harms, Brady, et al., 2004; Mueller et al., 2002). However, existing evidence for this event in the Beartooth Mountains is generally limited to the North and South Snowy blocks on the western margin of the range, where sparse 1.6–1.8 Ga <sup>40</sup>K/<sup>39</sup>Ar and Rb–Sr ages have been linked to Paleoproterozoic reheating (Montgomery & Lytwyn, 1984; Reid et al., 1975). Gast et al. (1958) reported older Rb–Sr (2.53–2.8 Ga) and K–Ar (2.29–2.52 Ga) ages from near the Beartooth Highway, interpreting that late Paleoproterozoic thermotectonism did not significantly affect the main Beartooth block.

During Mesoproterozoic time, the Belt Supergroup was deposited in parts of eastern Idaho, western Montana, southeastern British Columbia, and southwestern Alberta (Figure 1; e.g., Winston, 1986). Lower Belt Supergroup rocks deposited within the eastern arm of the Belt Basin, known as the Helena Embayment, are recognized in the subsurface ~50 km northwest of the study area (Figure 1; Vuke et al., 2007). Sedimentary thicknesses of  $\geq 15$  km are recognized in western portions of the Belt Basin, where the deposition of middle–upper portions of this succession (Ravalli through Missoula groups) is constrained to ~1.49–1.37 Ga (e.g., Anderson & Davis, 1995; Evans et al., 2000; Hirtz et al., 2022; Lonn et al., 2020; Ross & Villeneuve, 2003; Winston, 1986). By contrast, measured thicknesses total  $\leq 3$  km within the Helena Embayment and the depositional timing of these rocks is more poorly constrained, bracketed between ~1.71 and ~1.42 Ga by detrital zircon maximum depositional ages from the LaHood and Helena formations, respectively comprising the lowermost and uppermost units recognized in the Helena Embayment (Hirtz et al., 2022; Mueller et al., 2016). Deposition of the Belt Supergroup slightly preceded the intrusion of a suite of ~1,300 Ma mafic dikes, recognized in the Beartooth Mountains (e.g., Baadsgaard & Mueller, 1973). Both events have been linked to Mesoproterozoic rifting of the western Laurentian margin, possibly related to the breakup of supercontinent Nuna (e.g., Ross & Villeneuve, 2003; Sears & Price, 1978, 2000), also referred to as Columbia (e.g., Rogers & Santosh, 2002). The thermal effects of Mesoproterozoic rifting and sedimentation in the study area are poorly understood.



**Figure 2.** Generalized geology of the southeastern Beartooth Mountains. The range is comprised dominantly of Archean gneiss thrust over Paleozoic–Paleogene sedimentary rocks. Low-temperature thermochronology samples of this study and select previous studies are shown, colored by their corresponding transect used in time-temperature inversions. Geologic units and faults are generalized from Foote et al. (1961), Berg et al. (1999, 2000), Lopez (2001), and Wise (2000), with supplemental original mapping conducted during sample collection. Only Precambrian dikes longer than 2 km are shown. Note that the map is rotated  $\sim 30^\circ$  counterclockwise from north.

During middle–late Neoproterozoic time ( $\sim 780$ – $600$  Ma), the western Montana/Eastern Idaho region was located on the rifting western margin of Laurentia (e.g., Brennan et al., 2020), which developed into the Paleozoic western Laurentian passive margin (Yonkee et al., 2014). Regionally, evidence for the inception of this rifting event includes a suite of  $\sim 780$ – $708$  Ma mafic dikes—preserved in the Beartooth Mountains and elsewhere along the western margin of Laurentia (Figure 2; Baadsgaard & Mueller, 1973; Harlan et al., 2003). Later rifting is recorded by  $\sim 665$ – $650$  Ma alkalic plutonism and  $\sim 667$ – $601$  Ma deposition of syn-rift sedimentary units—likely correlative to the Windermere Supergroup—in east-central Idaho,  $\sim 300$  km SW of the study area (Figure 1; Brennan et al., 2020; Link et al., 2017; Lund et al., 2010). Normal faulting associated with this rifting event has been proposed as a driver of regional erosional exhumation in southwest Montana, based on modeled Neoproterozoic

cooling of basement rock therein (Kaempfer et al., 2021). Notably, Neoproterozoic rifting of western Laurentia temporally overlapped with widespread Cryogenian low-latitude glaciations, including the Sturtian (717–659 Ma) and Marinoan (641–635 Ma) “snowball Earth” events (Hoffman & Schrag, 2002; Hoffman et al., 1998, 2017; Kirschvink, 1992; Prave et al., 2016; Rooney et al., 2015; Walker et al., 1981), and the more enigmatic Gaskiers (~582 Ma; Bowring et al., 2003) and Fauquier (~571 Ma; Hebert et al., 2010) glaciations of the Ediacaran Period. Diamictites associated with Cryogenian glaciation are present in central and eastern Idaho, including within the previously mentioned syn-rift succession, but have not been recognized in Montana (e.g., Brennan et al., 2020; Lund et al., 2003, 2010; Yonkee et al., 2014). Snowball Earth glaciations have similarly been suggested as a driver of km-scale continental erosion based on results of thermochronology (McDannell et al., 2022) and analysis of the global geochemical, sedimentary, and terrestrial bolide impact records (Keller et al., 2019). However, there is no existing geologic or thermochronologic evidence for erosion related to late Neoproterozoic rifting or glaciation in the Beartooth Mountains.

### 2.3. Phanerozoic Burial and Tectonic History

At Beartooth Butte in the southeastern Beartooth Mountains, Phanerozoic sedimentary rocks nonconformably overlie Archean orthogneiss, representing a ~2.3 Gyr “gap” in the rock record (Figure 2; Thomas, 2008). This nonconformity coincides with similar global observations of a major unconformity at or near the base of the Phanerozoic section (e.g., Peters & Gaines, 2012; Walcott, 1914), frequently comprising a nonconformity developed atop Precambrian basement rocks (e.g., Flowers et al., 2020; Karlstrom & Timmons, 2012). The temporal correlation of these unconformities of late Precambrian age is referred to as the “Great Unconformity” (Definition 3 of McDannell et al., 2022; see discussion of variable usage of this term therein). In the region of the study area, the Flathead Sandstone—overlying the Great Unconformity surface—has been variably dated as upper Ediacaran, lower Cambrian, or middle Cambrian via Rb-Sr geochronology, paleopedology, and biostratigraphy, respectively (Chaudhuri & Brookins, 1969; Deiss, 1938; Norris & Price, 1966; Retallack, 2013; Thomas, 2007). Subsequent Paleozoic passive margin-style sedimentation is recorded by onlapping successions of sandstone, limestone, and shale, with intervening unconformities associated with periods of non-deposition and erosion (Mallory, 1972; Maughan, 1993). Regionally, deposition was punctuated by localized uplift during Devonian to Early Mississippian and Late Mississippian to Permian time, likely associated with the Antler and Ancestral Rockies orogenic events, respectively (Dorobek et al., 1991; Maughan, 1990). Along the northeastern Beartooth front, a maximum thickness of ~1 km of Paleozoic stratigraphy is preserved (Lopez, 2001; Maughan, 1993).

During Middle Triassic to Late Jurassic time, the western margin of North America evolved into a tectonically consolidated margin associated with east-dipping subduction of the Farallon oceanic plate beneath the North American continent (e.g., Colpron et al., 2007; Dickinson, 2004). This evolution culminated in the establishment of the North American Cordillera, a retroarc orogenic system extending more than 6,000 km from southern Mexico to Alaska, USA (DeCelles, 2004; Yonkee & Weil, 2015). A complex and extensive foreland basin system developed eastward of the retroarc fold-thrust belt (Figure 1; DeCelles, 2004; Kauffman & Caldwell, 1992). In areas near the Beartooth Mountains, stratigraphy deposited in this foreland basin constitute Jurassic through Paleogene rocks with a maximum thickness of ~4.5 km (Lopez, 2001; Mallory, 1972; Maughan, 1993).

The Beartooth Mountains comprise one of the northernmost features in an elongate series of intraforeland uplifts cored by Precambrian basement rock spanning from New Mexico to southwestern Montana that partitioned the Cordilleran foreland basin during a mountain building event classically referred to as the Laramide orogeny (Figure 1; e.g., Armstrong, 1968; Coney & Reynolds, 1977; Dickinson & Snyder, 1978). The timing of initiation of this event has been subject to recent debate (see Orme, 2020 for a review). Stratigraphic and sedimentologic evidence has been interpreted to suggest segmentation of the foreland basin by incipient intraforeland highs by late Early Cretaceous time (DeCelles, 1986; Parcell & Williams, 2005; Schwartz & Decelles, 1988), while more recent detrital zircon studies suggest partitioning of the southwest Montana foreland basin may have been delayed until ~88 Ma (Garber et al., 2020; Rosenblume et al., 2021). The synorogenic Beartooth Conglomerate member of the Paleogene Fort Union Formation, deposited on the east-northeast flank of the Beartooth Mountains, preserves thrust-related growth strata and an unroofing sequence recording exhumation of the uplift (Ayers, 1986; DeCelles, Gray, Ridgway, Cole, Srivastava, et al., 1991; Gingerich, 1983). Clasts of metamorphic basement occur high in the section, providing unequivocal evidence that basement rocks of the Beartooth Mountains were actively eroding by the deposition of this unit at ~57–55 Ma (DeCelles, Gray, Ridgway, Cole, Srivastava, et al., 1991; Flueckinger, 1970; Jobling, 1974; Koenig, 2015).

### 3. Methodology

#### 3.1. Sample Collection and Mineral Separation

Samples were collected from exposed quartzofeldspathic gneiss along three pseudo-vertical transects in the Beartooth Mountains (Figure 2). Reconnaissance mapping was conducted to ensure samples were collected an adequate distance from intrusions (Figure 2). Five samples were collected along the East Rosebud “Beaten Path” trail; 4 samples were collected on the approach to and summit of Granite Peak; and 5 samples were collected on the approach to and summit of Chalice Peak (Figure 2). Sample elevations ranged from 1,742 to 3,909 m. Biotite, zircon, and apatite grains were extracted by crushing, grinding, and sieving with a 100-mesh screen, followed by separation with water, a Frantz magnetic separator, and heavy liquids.

#### 3.2. Zircon U-Pb Geochronology

U-Pb geochronology was conducted on zircons ( $n = 50$ ) from sample GP-01, collected from the summit of Granite Peak (Figure 2). Though extensive U-Pb geochronology has been conducted on basement rocks from the Beartooth Mountains, these data are concentrated in locations near the Beartooth Highway and relatively few data exist for the central portions of the range (Figure 2; e.g., Carrapa et al., 2019; Mueller et al., 2008, 2010; Wooden et al., 1988). The high  $T_c$  of the U-Pb system means that zircon U-Pb ages generally reflect crystallization of the phase and provide robust starting constraints for thermal history models (e.g., Gehrels et al., 2008). Zircon grains were hand selected from the heavy fraction of mineral separates, prepared for analysis, and imaged with cathodoluminescence (CL) at the Arizona LaserChron Center (ALC; Tucson, Arizona, USA). CL images were used to select analysis spots targeting unzoned cores and rims of zircon grains. U-Pb geochronology of zircons was conducted by laser ablation multicollector inductively coupled plasma mass spectrometry (LA-MC-ICP-MS) at the ALC. Detailed description of methods is provided in Text S1 in Supporting Information S1 and analytical settings are provided in Data Set S1.

#### 3.3. Biotite $^{40}\text{Ar}/^{39}\text{Ar}$ Medium-Temperature Thermochronology

$^{40}\text{Ar}/^{39}\text{Ar}$  thermochronology was conducted on biotite from the highest and lowest elevation samples from each transect ( $N = 6$ ). This thermochronometer provides information about the medium-temperature ( $\sim 300^\circ\text{C}$ ) thermal history of our samples (Grove & Harrison, 1996). Existing Ar dating in the Beartooth Mountains predominantly utilizes the K-Ar method and is concentrated on dating of intrusive dikes, which don't provide direct information about the thermal history of the gneissic basement rocks we sampled (e.g., Baadsgaard & Mueller, 1973; Condie et al., 1969). Visually unaltered, inclusion-free flakes of biotite were selected from the light fraction of separates and co-irradiated at the Oregon State TRIGA Reactor. Single-grain aliquots were analyzed using single- and multi-step heating experiments on a VG5400 and an Isotopx NGX mass spectrometer at the Argon Geochronology for the Earth Sciences Laboratory at the Lamont-Doherty Earth Observatory, Columbia University (Palisades, New York, USA). We conducted bulk age analyses (the samples were fused and analyzed in a single step) and two sets of step-heating experiments—the first with 9 heating steps to determine the Ar release characteristic of the samples, and the second with 12 heating steps. We calculated plateau biotite  $^{40}\text{Ar}/^{39}\text{Ar}$  ages from the results of the second step-heating experiment. A natural plateau was defined following accepted criteria ( $2\sigma$  overlap of each step,  $\geq 50\%$   $^{39}\text{Ar}$ ,  $\geq$  three subsequent steps). Where natural plateaus were not achieved, we calculated forced plateau ages and then weighed the apparent standard error of the mean (SEM) uncertainties by the square root of the mean square weighted deviation ( $\sqrt{\text{MSWD}}$ ) to account for the additional spread beyond analytical uncertainties in the heating steps forced onto the plateau (Powell et al., 2020). Analytical uncertainties for each heating step are reported as  $1\sigma$  SEM (see Data Set S2). See Text S2 in Supporting Information S1 for a detailed description of biotite  $^{40}\text{Ar}/^{39}\text{Ar}$  analytical methods.

#### 3.4. Apatite and Zircon [U-Th(-Sm)]/He Low-Temperature Thermochronology

Apatite and zircon [U-Th(-Sm)]/He thermochronology leverage the time-dependent production and temperature-dependent diffusion of radiogenic  $^4\text{He}$  through the crystal lattice. Trace U, Th, and Sm decay to  $^4\text{He}$  at a rate controlled by the half-life of these radioactive parent isotopes; radiogenic  $^4\text{He}$  escapes from the crystal lattice at higher temperatures, while it is retained at lower temperatures (e.g., Reiners et al., 2004). We use these thermochronometers to provide constraints on the  $< 200^\circ\text{C}$  thermal history of the Beartooth Mountains.

Recent advancements in the understanding of the influence of radiation damage and annealing on helium diffusivity in apatite and zircon show that these factors exert a strong control on the closure temperature of the systems (e.g., Flowers et al., 2009; Ginster et al., 2019; Guenther, 2021; Guenther et al., 2013; Johnson et al., 2017). Damage to the crystal lattice accumulates as uranium undergoes alpha decay, while heating of the crystal can anneal accumulated radiation damage; the level of net radiation damage consequently affects He retentivity (Flowers et al., 2009; Guenther et al., 2013). Owing to these effects, each crystal of a given level of accumulated radiation damage develops a unique “effective” partial retention zone (PRZ) temperature and consequently commonly records a unique cooling date (Flowers et al., 2009; Guenther et al., 2013). For samples with a  $t$ - $T$  history characterized by long duration at temperatures permitting accumulation of radiation damage followed by moderate reheating and/or slow cooling, date dispersion commonly manifests as a relationship (also referred to as correlation or trend; e.g., Guenther et al., 2013; Orme et al., 2016) between single-grain date and parent isotope concentration (or effective uranium; eU; a proxy for radiation damage, formally derived by Cooperdock et al., 2019;  $eU = [U] + 0.28[Th] + 0.0012[Sm]$ ). In apatite, radiation damage generally decreases the diffusivity of the crystal, resulting in a positive relationship between date and eU (Flowers et al., 2009). Zircon exhibits similar behavior at low levels of radiation damage, while moderate to high levels increase diffusivity, resulting in a negative relationship between date and eU (Guenther et al., 2013). The specific  $t$ - $T$  history of a sample controls the shape of the resultant date-eU relationship, such that key aspects of the former can be recovered from the latter (Flowers et al., 2009; Guenther, 2021; Guenther et al., 2013). For these reasons, thermal modeling efforts incorporating a radiation damage and annealing model are particularly salient in reconstructing complex deep-time thermal histories (e.g., DeLucia et al., 2018; Flowers et al., 2020; McDannell et al., 2019; Orme et al., 2016; Reade et al., 2020).

Between 1 and 7 zircon grains were hand-picked from the heavy fraction of mineral separates from each sample for a total of 84 single zircon grain aliquots ( $N = 14$ ;  $n = 84$ ). Grains were selected for maximum intrasample eU variability. We used the visual metamictization of grains as a proxy for accumulated radiation damage (and eU concentration) such that clear grains reflect low levels of radiation damage and opaque grains reflect high levels of radiation damage (Ault et al., 2018). Euhedral grains of similar half-widths ( $\sim 60 \mu\text{m}$ ) free from fluid or mineral inclusions were preferentially selected, when possible. The presence of broken crystal terminations or internal imperfections was noted when observed (see Data Set S3). Grain photos are available in Data Set S4, <https://doi.org/10.5281/zenodo.7443913>. Selected grains were photographed, measured to determine surface area to volume ratios for alpha ejection correction (Farley et al., 1996; Hourigan et al., 2005; Ketcham et al., 2011), and packed into Nb foil tubes to ensure even heating of the grain and prevent volatilization of parent nuclides during He extraction. Isotopic analysis was conducted at the Helium Analysis Laboratory (HAL) at the University of Illinois, Urbana-Champaign (Urbana, Illinois, USA). Uncertainties on individual ZHe dates, reflecting the combination of measurement and systematic error, are reported at the  $2\sigma$  level. Detailed analytical methods for ZHe thermochronology are provided in Text S3 in Supporting Information S1.

AHe thermochronology was conducted on apatites from the highest and lowest elevation samples from each transect. Three apatite grains were hand-picked from the medium-density fraction of mineral separate for each selected sample for a total of 18 single grain aliquots ( $N = 6$ ,  $n = 18$ ; see Data Set S5). Grains were photographed, measured, and packed into Nb foil tubes in the same manner as described above for ZHe analyses. He extraction and measurement, as well as isotopic dissolution for U-Th-Sm content, was conducted at the University of Colorado (U-Th)/He thermochronology laboratory (Boulder, CO, USA). Analytical uncertainties reflecting the combination of measurement and systematic error are reported at the  $2\sigma$  level. Detailed analytical methods for AHe thermochronology are provided in Text S4 in Supporting Information S1.

### 3.5. Bayesian Time-Temperature Inversions

To explore plausible time-temperature histories of the Beartooth Mountains, we performed Bayesian inverse thermal history modeling of newly reported biotite  $^{40}\text{Ar}/^{39}\text{Ar}$ , ZHe, and AHe data and previously published AFT and AHe data from the Beartooth Mountains using the QTQt v. 5.8.0 software package (Gallagher, 2012). This approach facilitates the testing of many ( $10^6$ ) potential thermal histories using a transdimensional reversible jump Markov chain Monte Carlo (rjMCMC) approach. QTQt infers the complexity of the thermal history solution from the data directly and will generally favor simpler thermal histories unless the model fit is significantly improved by additional complexity (i.e., additional  $t$ - $T$  points).

The helium model in QTQt takes parent isotope concentrations, grain radii, measured dates, and date uncertainties as input. The radiation damage and annealing models of Flowers et al. (2009) and Guenther et al. (2013) were implemented for modeling of AHe and ZHe data, respectively. To expand uncertainty accounting, we implemented Hierarchical Bayesian resampling of [U-Th(-Sm)]/He date uncertainty. Analytical uncertainties were input into QTQt and, during inverse modeling, were randomly rescaled by a value of 1–5 (Gallagher, 2012; Malinverno & Briggs, 2004). Thereby, the variance of the date uncertainties is estimated from their most probable value, given the data. This approach helps account for the discrepancy between typically small calculated analytical uncertainties (typically  $\leq 5\%$ ) and the much larger uncertainty indicated by studies analyzing large numbers of replicate analyses of natural aliquots and continuous ramped heating experiments ( $\geq 15\%$ ; e.g., McDannell et al., 2018, 2022).

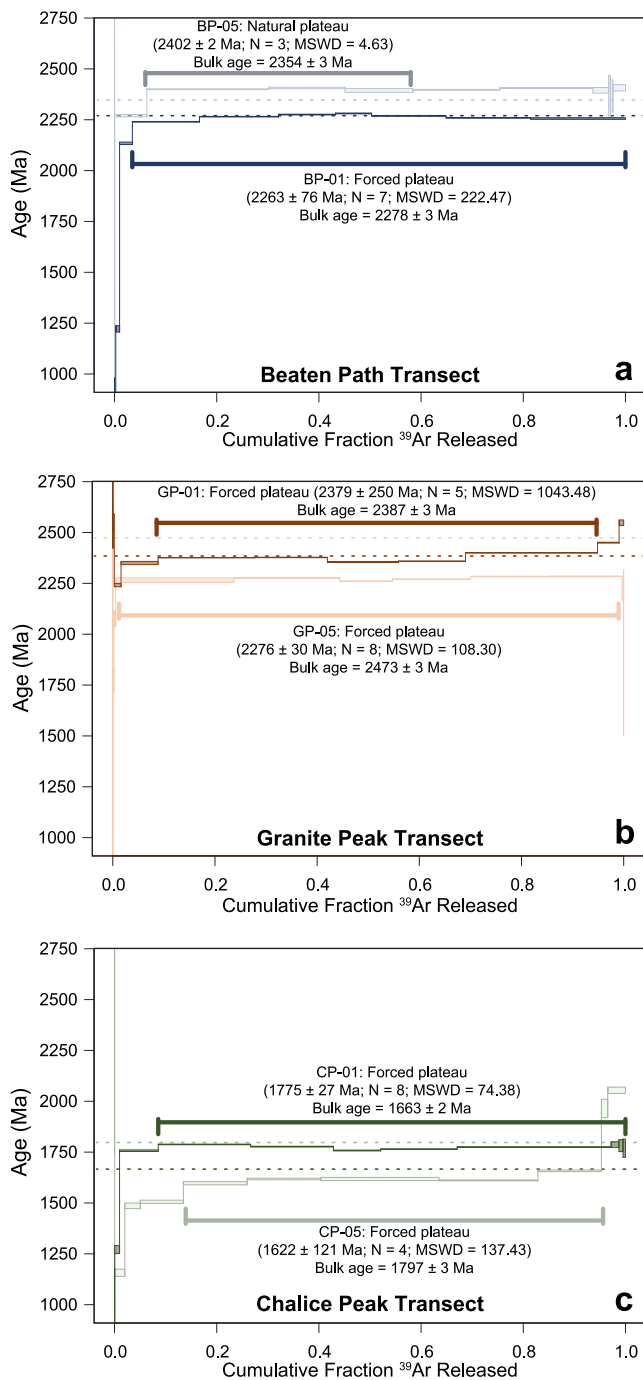
To invert biotite  $^{40}\text{Ar}/^{39}\text{Ar}$  data, we used the helium model in QTQt modified by inputting diffusion parameters for  $^{40}\text{Ar}^*$  diffusion in biotite calculated by Grove and Harrison (1996) for an infinite cylinder geometry. We assumed a 150  $\mu\text{m}$  grain size. Although we did not measure individual biotite grains, this assumption is reasonable because we selected the largest biotite flakes which passed through a 100-mesh screen ( $\sim 150 \mu\text{m}$ ), and previous studies have suggested a  $\sim 150 \mu\text{m}$  effective radius of diffusion (Harrison et al., 1985). Biotite  $^{40}\text{Ar}/^{39}\text{Ar}$  plateau ages and corresponding uncertainties, described in Section 3.3, were input into QTQt. This allowed us to screen spurious heating steps, potentially affected by processes not accounted for by the model (e.g., McDougall & Harrison, 1999). Additional details on our approach to modeling biotite  $^{40}\text{Ar}/^{39}\text{Ar}$  data in QTQt are available in Text S2 in Supporting Information S1.

The AFT multi-kinetic model of Ketchum et al. (2007) was implemented for modeling of AFT data previously published by Carrapa et al. (2019) and Omar et al. (1994). All relevant AFT count and length data from Carrapa et al. (2019) were input into the AFT model directly. The full data set of Omar et al. (1994) was not available, so the values of  $p_d$  and  $N_d$  were resampled using the reported date, date uncertainty, and  $N_s/N_i$  for each sample; the  $D_{\text{par}}$  value was resampled from a distribution of  $2.0 \pm 2.0$ , and the QTQt default values of 340 and 5 were accepted for  $\zeta$  and  $\partial\zeta$ , respectively. Although portions of these data sets were modeled by their original authors, we remodel them herein using the same software and parameters as our newly reported models to facilitate direct comparison of results. Additionally, our approach allows for the integration of multiple samples in a vertical profile, which was not used in previous studies.

All thermal history inversions employed a general prior of 2,812–0 Ma and 0–400°C. This broad prior allows models to explore time space from the present to the older limit of observed concordant zircon U-Pb ages and temperature space spanning surface temperatures to those above the closure temperature of the highest-temperature thermochronometer modeled. We note that modeled thermal history complexity is generally inversely proportional to the size of the general prior in QTQt because more data are required to resolve the model across a greater range of time and temperature space. We justify our broad general prior because we input a large and diverse suite of thermochronology data recording dates spanning  $\sim 2.5$  Gyr and sensitive to a broad range of temperatures. In the Bayesian context of QTQt, our unconstrained models therefore comprise a lower limit on the complexity of the thermal history required to fit the observed data (Gallagher, 2021).

A subset of models included three constraints, using robust geologic and geochronologic information to guide the inversions: (a) 2,812–2,799 Ma and 390°C–400°C—the crystallization age of the main orthogneiss suite in the Beartooth Mountains (e.g., Carrapa et al., 2019; Wooden et al., 1988; Mueller et al., 2008; this study); (b) 550–520 Ma and 0°C–100°C—the timing of deposition of the Cambrian Flathead Sandstone atop the Great Unconformity surface (e.g., Chaudhuri & Brookins, 1969; Retallack, 2013; Thomas, 2008), constraining basement rocks near the unconformity to near-surface temperatures; and (c) 60–50 Ma and 0°C–100°C—the depositional timing of the syn-orogenic Beartooth Conglomerate, which includes clasts of basement rock eroded at the surface (DeCelles, Gray, Ridgway, Cole, Srivastava, et al., 1991). The relatively broad temperature constraints ( $\leq 100^\circ\text{C}$ ) for (b) and (c) reflect the uncertainty regarding the thickness of basement rock eroded from the Beartooth Mountains (e.g., Omar et al., 1994; Simons & Armbrustmacher, 1976).

To promote thorough exploration of modeled time-temperature space, more complex models were accepted for equivalent likelihood and proposal jumps were resampled if the proposal fell outside of the general prior. Models were run for 1,500,000 total iterations, consisting of 750,000 burn-in iterations that were discarded and an additional 750,000 post-burn-in iterations retained to estimate the posterior probability of the model parameters. The sampling distribution reached stationary—with no structure in the likelihood or posterior chains—by the end of



**Figure 3.** Biotite  $^{40}\text{Ar}/^{39}\text{Ar}$  gas release spectra for samples from (a) Beaten Path, (b) Granite Peak, and (c) Chalice Peak transects in the Beartooth Mountains. Steps used for plateau determination are highlighted by bold brackets. Bulk ages are indicated by dotted lines. For samples not meeting the accepted criteria for a plateau, a plateau calculation using steps comprising a visual plateau was forced, and uncertainty was calculated using the product of the SEM uncertainty and the square root of the mean square weighted deviation (as defined in the text).

the burn-in period, signifying model convergence. Model input and output files, detailed information on model setup, and justification for constraints are available in Data Set S6, <https://doi.org/10.5281/zenodo.7443913>, Tables S1 and S2 in Supporting Information S1, respectively.

## 4. Analytical Results

### 4.1. Zircon U-Pb Results

Eleven zircon U-Pb analyses meeting concordance cutoffs were obtained ( $<20\%$  discordance,  $<5\%$  reverse discordance). Ages meeting concordance cutoffs are Neo-Mesoarchean—ranging from  $2,689.0 \pm 12.9$  to  $3,124.0 \pm 38.4$  Ma (Figure S1 in Supporting Information S1). Four analyses have error ellipses intersecting concordia, giving a concordia age of  $2,805.6 \pm 6.4$  Ma (mean square weighted deviation [MSWD] = 0.79;  $p(\chi^2) = 0.37$ ; calculated using IsoplotR; Vermeesch, 2018; see Figures S2a and S2b in Supporting Information S1). The probability distribution of U-Pb ages is defined by a peak at  $\sim 2,803.5$  Ma with a second, bimodal peak centered at  $\sim 2,957$  Ma and a tail at lower probability to  $\sim 3,200$  Ma (Figure S1c in Supporting Information S1). These results are consistent with previous zircon U-Pb geochronology results from the Beartooth Mountains (Mueller et al., 2008; Wooden et al., 1988). Analyses with ages  $\geq 2,791.4$  Ma, comprising 10/11 concordant analyses, had Th/U values  $> 0.2$  (see Figure S1d in Supporting Information S1), suggesting these ages record magmatic growth of zircon (e.g., Pystina & Pystin, 2019). A lower Th/U value of 0.04 was obtained for a single analysis with an age of  $2,688.97 \pm 12.90$  Ma, consistent with minor Neoproterozoic metamorphic growth of zircon. The bulk of ages coincide with the volumetrically dominant  $\sim 2.8$ – $2.9$  Ga magmatic suite in the range, while older ages may reflect inherited zircons from the Mesoarchean TTG suite (Mueller et al., 2008). Detailed analytical data is available in Data Set S1.

### 4.2. Biotite $^{40}\text{Ar}/^{39}\text{Ar}$ Results

Apart from GP-01, age spectra from biotite  $^{40}\text{Ar}/^{39}\text{Ar}$  step-heating diffusion experiments for all samples rise from younger values to a plateau-like sequence of steps within the first 5%–20% of  $^{39}\text{Ar}$  release (Figure 3). Sample GP-01 exhibits a saddle-like spectrum, with older values exhibited during the first and last  $\sim 5\%$  of  $^{39}\text{Ar}$  release (Figure 3b). Older values within the last several steps are also present in the spectrum of sample CP-05 (Figure 3c). Young step ages during initial increments of  $^{39}\text{Ar}$  release may suggest minor episodic  $^{40}\text{Ar}$  loss, slow cooling through the PRZ, or degassing of (sub) microscopic inclusions within biotites (Onstott & Peacock, 1987; Ross & Sharp, 1988). The saddle-shaped spectrum of sample GP-01 may indicate the presence of minor excess Ar in this sample (Figure 3b; e.g., Lanphere & Brent Dalrymple, 1976). However, all age spectra are generally flat for the central  $\geq 80\%$  of  $^{39}\text{Ar}$  release and yield a clear visual sequence of plateau-like steps, suggesting these potential sources of discordance are of relatively minor importance (Figure 3).

A natural plateau was obtained only for sample BP-05. For other samples, we report forced plateau ages and uncertainties using the methods described in Section 3.3. Using this approach, biotite samples yielded  $^{40}\text{Ar}/^{39}\text{Ar}$  plateau ages ranging from  $1,622 \pm 121$  to  $2,402 \pm 2$  Ma (Figure 3). We additionally report bulk ages from single-step heating experiments for separate aliquots of each sample (Figure 3). These bulk ages range from  $1,663 \pm 2$  to

2,472  $\pm$  3 Ma and demonstrate general agreement with the plateau ages (Figure 3). Biotite  $^{40}\text{Ar}/^{39}\text{Ar}$  plateau ages comprise two distinct groups with internal overlap at  $2\sigma$  uncertainty: (a) samples from the westernmost transect (CP) yield ages between 1,662  $\pm$  121.22 and 1,775  $\pm$  27.08 Ma; (b) samples from the eastern two transects (Granite Peak and Beaten Path) yield ages between 2,263  $\pm$  75 Ma and 2,402  $\pm$  2 Ma (Figure 3). High elevation samples in each transect record older cooling dates than those from low elevation. See Data Set S2 for detailed isotopic results.

### 4.3. Zircon (U-Th)/He Results

Corrected ZHe dates range from 686.4  $\pm$  11.9 to 13.5  $\pm$  0.3 Ma while eU concentrations span 113–3,565 ppm (Figure 4; detailed isotopic results available in Data Set S3). ZHe dates from the Beartooth Mountains are relatively evenly distributed with few obvious date probability peaks (Figure 5b). Correlation between ZHe date and grain size is poor (Figure S2 in Supporting Information S1). By contrast, ZHe dates exhibit a coherent, negative relationship with eU, a proxy for radiation damage, which is expected given that we purposefully selected for grains with a wide range of observable damage (Figure 4). The trend is strongly negative for eU values less than  $\sim$ 500 ppm, consistent with moderate to high levels of accumulated damage ( $>1.0 \alpha/\text{g} \times 10^{17}$ ; Guenther et al., 2013). For all transects, grains of  $>800$  ppm eU form a pediment of ages of  $\sim$ 10 Ma to  $\sim$ 110 Ma before reaching dates  $<50$  Ma at  $>1,200$  ppm eU (Figure 4). Eastern transects (Beaten Path and Granite Peak) define a more gradual transition into this pediment, with an initial inflection to a lower-angle date-eU slope at  $\sim$ 400 ppm and a second inflection to a nearly flat slope at  $\sim$ 800 ppm eU (well-defined for only the Beaten Path transect; Figures 4a and 4b). By contrast, the Chalice Peak transect exhibits a steeper date-eU slope at low-eU concentrations and a more abrupt inflection at  $\sim$ 500 ppm eU (Figure 4c). Elevation has a second-order influence on ZHe date (Figure 4 insets). Although intrasample date dispersion is high—likely primarily owing to radiation damage effects—median ZHe dates of each sample (red Xs in Figure 4 insets) generally trend toward older values at higher elevations. Samples BP-05 and CP-03 noticeably deviate from this trend. However, this is likely due these analyses sampling only moderate–high eU grains (384–3,565 ppm). Consequently, radiation damage effects may mask the date-elevation relationship for these samples.

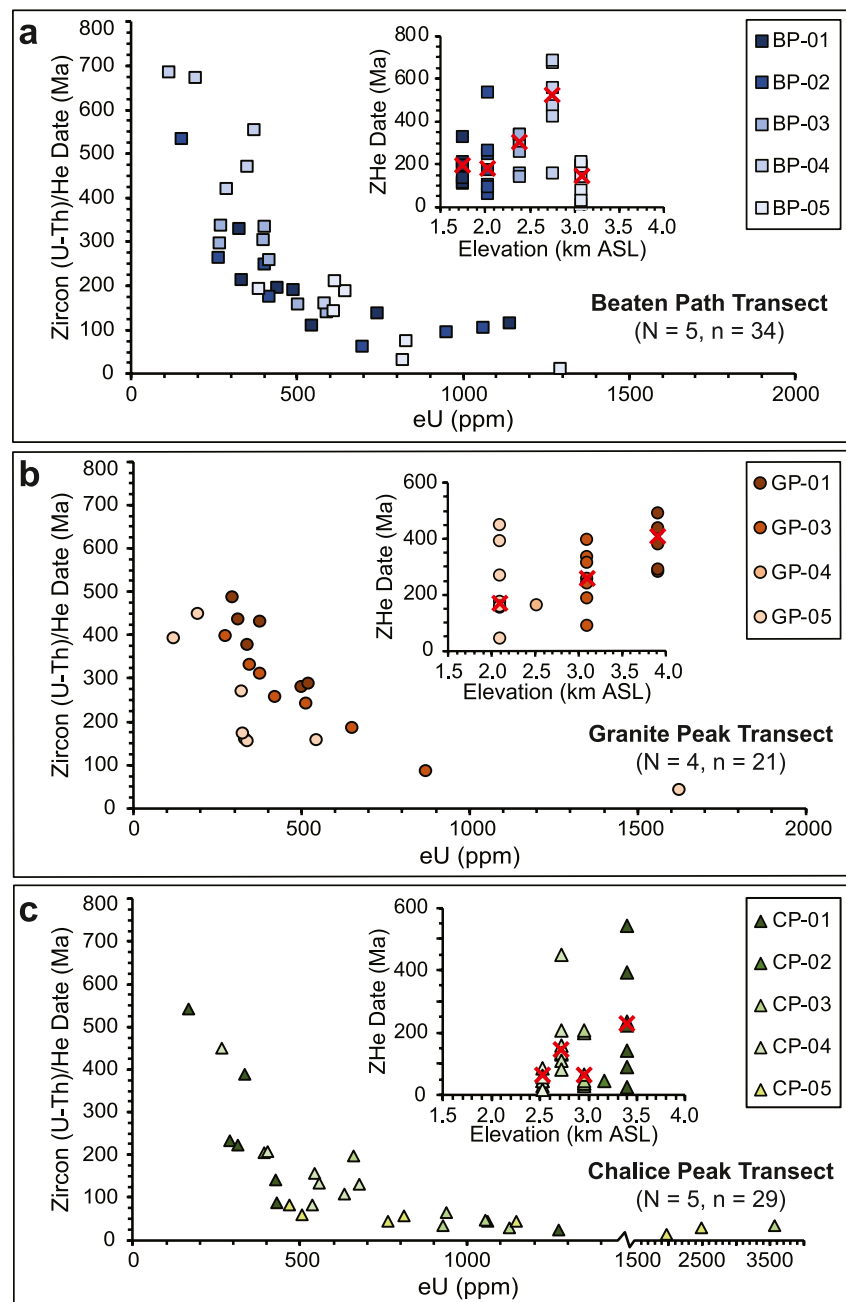
### 4.4. Apatite (U-Th-Sm)/He Results

AHe dates span from 109.2  $\pm$  23.9 to 43.6  $\pm$  1.9 Ma and eU concentrations range from 3.3 to 35.4 ppm (Figure 5; see Data Set S5 <https://doi.org/10.5281/zenodo.7443913> for detailed analytical data). AHe dates exhibit moderate intrasample dispersion (average intrasample  $\sigma$  = 11.87 Myr). Compared to previously reported AHe dates from the Beartooth Mountains, the present distribution of dates is flatter and lacks tails  $<40$  and  $>110$  Ma, but the dominant Paleocene date probability peak is consistent between new and previously reported data sets (Figure 5b; Bricker, 2016; Carrapa et al., 2019; Mueller et al., 2008; Peyton et al., 2012). Newly reported AHe dates are less dispersed than those of Mueller et al. (2008), Bricker (2016), and Carrapa et al. (2019), who reported dates ranging from  $\sim$ 190 to  $\sim$ 32 Ma from samples collected along the Beartooth Highway, East Rosebud Creek, and Stillwater River drainages (Figures 2 and 5).

Correlation between AHe date and grain size is poor (Figure S3a in Supporting Information S1). Aliquots from most newly reported samples span a relatively restricted array of eU concentrations ( $\leq 20$  ppm), and AHe dates are not strongly correlated with eU (Figure S3b in Supporting Information S1). The Beaten Path and Granite Peak transects display a positive date-elevation relationship, while this trend is less coherent in samples from the Chalice Peak transect (Figure 5a). The lack of a clear inflection point toward older AHe dates with increasing elevation suggests that a fossil AHe PRZ is not preserved in the eastern Beartooth Mountains, contrary to suggestions of Bricker (2016) and Mueller et al. (2008). Rather, previously reported, highly dispersed AHe dates at  $\sim$ 1,300–3,200 m ASL may have been affected by other sources of AHe date overdispersion (e.g., Brown et al., 2013; Guo et al., 2021; Murray et al., 2014). For instance, Peyton et al. (2012) observed clear overdispersion in a sample from an intrusive porphyry in the Beartooth Mountains with AHe dates of up to 138  $\pm$  19 Ma and a zircon U-Pb date of 98.3  $\pm$  0.3/–1.0 Ma (gray Xs at 2,868 m ASL in Figure 5a).

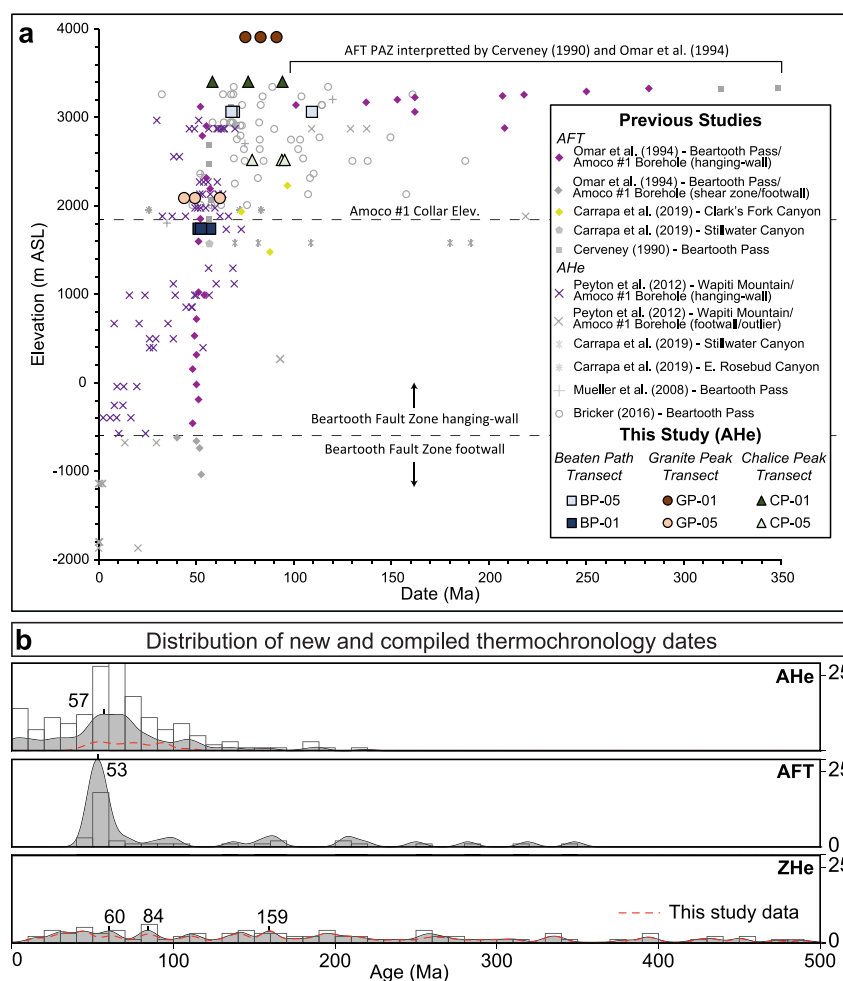
### 4.5. Time-Temperature Inversion Results

Results of Bayesian  $t$ - $T$  inversions implemented in QTQt show the relative probability of thermal history solutions for transects in the Beartooth Mountains (Figures 6 and 7). We report the results of models without geologic



**Figure 4.** Corrected ZHe dates are plotted against eU from three transects in the Beartooth Range: (a) Beaten Path trail; (b) Granite Peak; and (c) Chalice Peak. A negative relationship between ZHe date and eU concentration exists independently all transects. Note the change in x-axis scale at 1,500 ppm eU in panel (c). Insets show corrected ZHe date plotted against sample elevation. Red Xs denote median ZHe date for samples with multiple aliquots.

constraints (“unconstrained models”; Figures 6a, 6c, and 6e) primarily as a sensitivity test, allowing us to better examine the resolution of the thermochronology data themselves and isolate possible artifacts of imposed boundary conditions. These model results represent the simplest thermal history solutions consistent with the observed thermochronology data but do not necessarily honor geologic data. By contrast, models including geologic constraints (“constrained models”; Figures 6b, 6d, and 6f; Figure 7) represent the most probable thermal history solutions considering all available sources of information—both thermochronologic and geologic—and are therefore primarily used for geologic interpretation. The agreement between observed and predicted thermochronometer dates is generally good for all models, with exceptions noted (Figure 8).

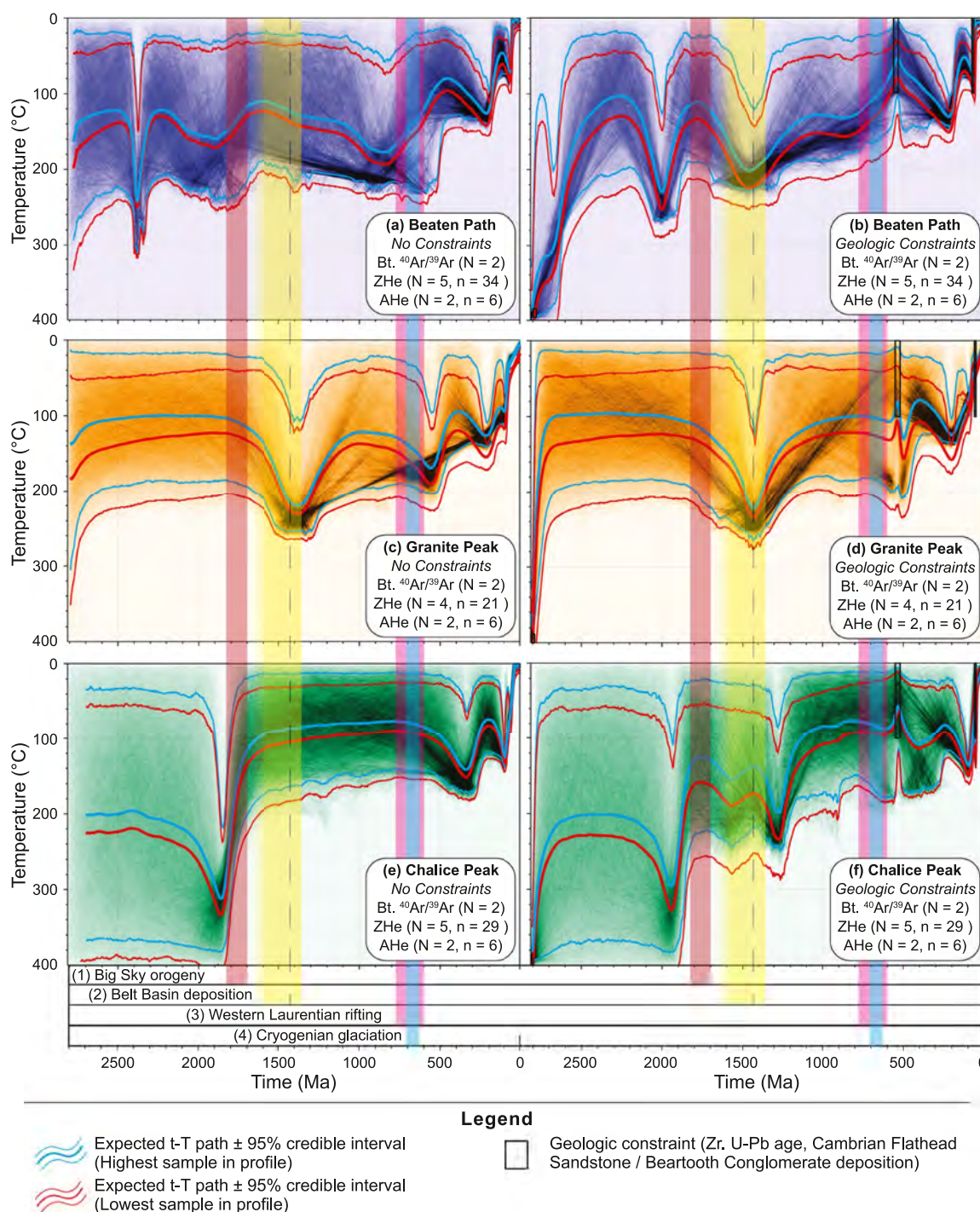


**Figure 5.** (a) Existing and newly reported AHe and apatite fission track (AFT) dates from the Beartooth Mountains as a function of elevation. Sample locations are shown in Figure 2. Data with colored symbols were used in time-temperature inversions, while gray symbols were not. Note that several outlier aliquots from the data set of Peyton et al. (2012), which the authors identified as being overdispersed, were excluded from our modeling. (b) Kernel density estimation (KDE) and histogram plots summarizing new and previously reported AHe, AFT, and ZHe dates from the Beartooth Mountains between 0 and 500 Ma; note that newly reported ZHe dates >500 Ma plot off the scale. KDE and histogram bin widths are 5 and 10 Myr, respectively, and KDEs are normalized such that each plot contains the same area under the curve. Prominent date-probability peaks are indicated by vertical ticks and labeled. Scales at right correspond to the number of analyses in each histogram bin. Data sources: AHe—Bricker (2016), Carrapa et al. (2019), Mueller et al. (2008), Peyton et al. (2012), and this study; AFT—Carrapa et al. (2019), Cervený (1990), and Omar et al. (1994), ZHe—Carrapa et al. (2019) and this study.

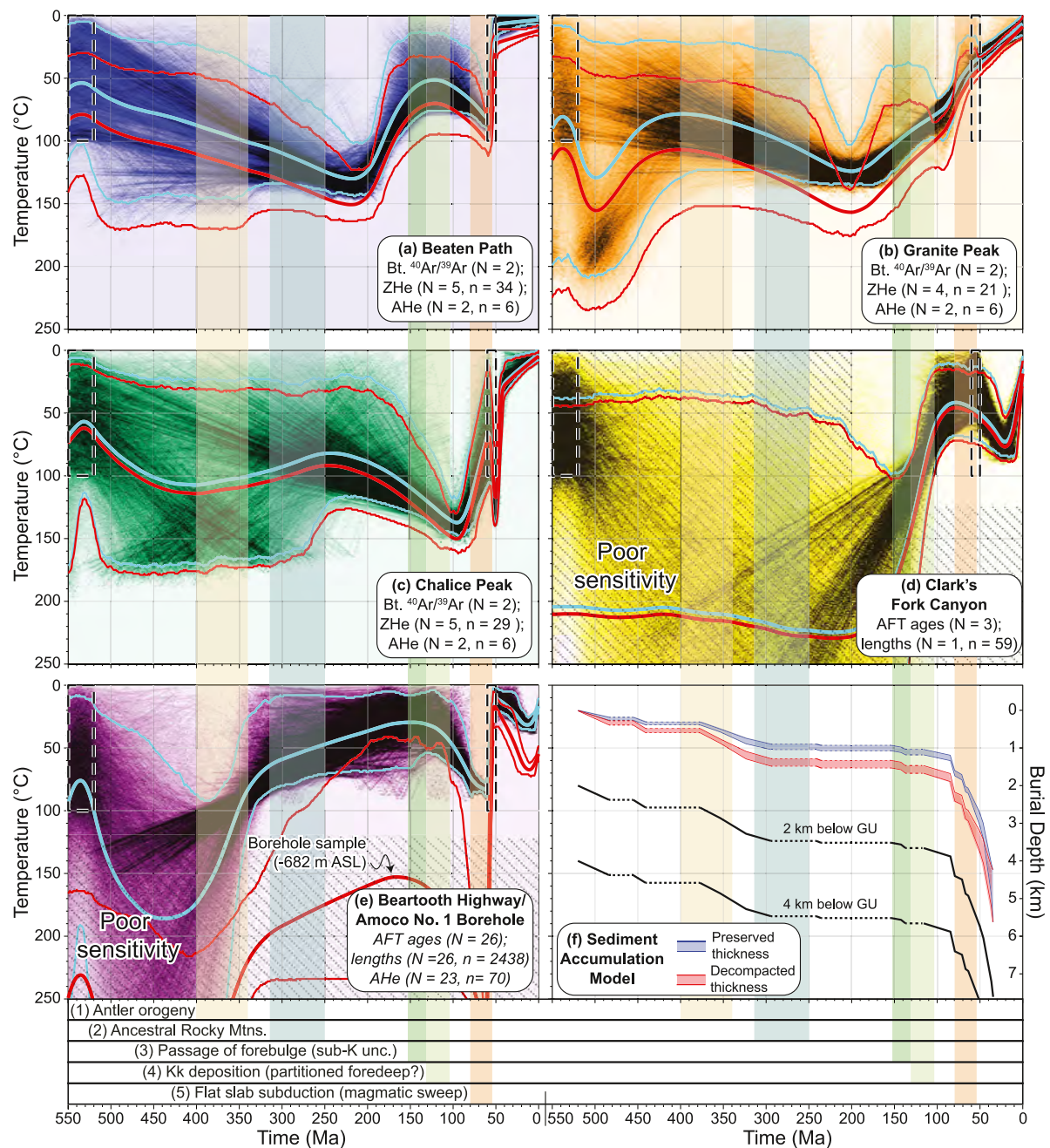
The Precambrian portion of our thermal history models resolves several first-order thermal events previously unrecognized in the Beartooth Mountains, discussed here in detail (Figure 6). Model results are described for the highest elevation sample in each transect. Modeled thermal histories for lower elevation samples follow the same  $t$ - $T$  paths, offset according to a geothermal gradient resampled during modeling. We discuss model results in terms of both relative path density, which is proportional to relative probability, and the expected model—a weighted mean model, where the weighing is provided by the posterior probability for each thermal history solution, and the preferred single model output in QTQt (Gallagher, 2012; Figure 6). The results of supplemental models, testing the effects of modifying the resampling parameters, are available in Figure S4 in Supporting Information S1.

#### 4.5.1. Modeled Precambrian Thermal History

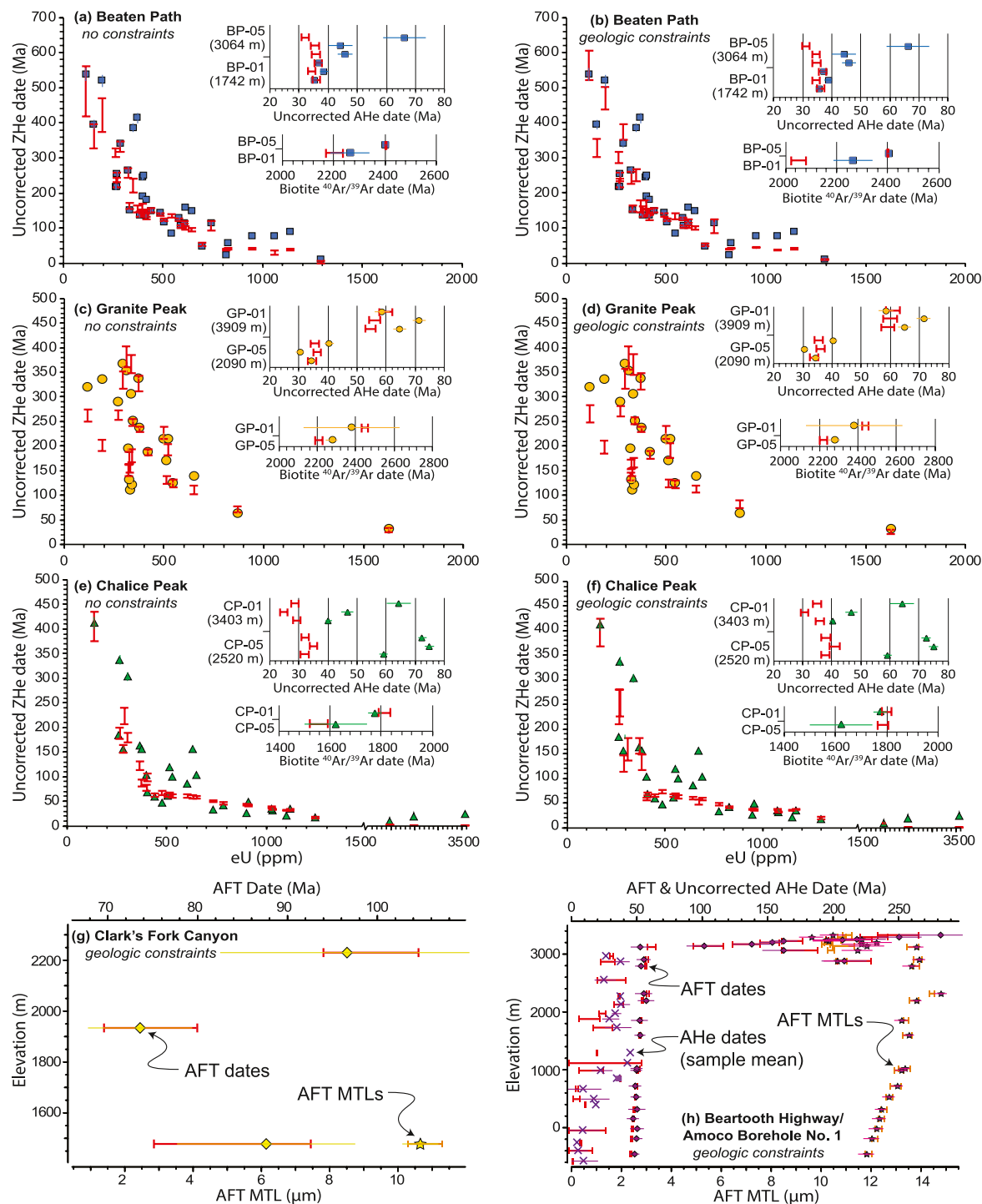
Model results for the Beaten Path transect indicate at least three distinct Proterozoic cooling events (Figures 6a and 6b). The most probable solution for the constrained model, reflected by the highest path density, suggests



**Figure 6.** Results of QTQt  $t$ - $T$  inversions of thermochronology data from (a, b) Beaten Path, (c, d) Granite Peak, and (e, f) Chalice Peak transects. Panels on the left (a, c, and e) show results of unconstrained models while panels on the right (b, d, and f) show results of constrained models. Post-burn-in  $t$ - $T$  path density is shown for the highest elevation sample in each profile by colored gradient (blue, orange, and green). Path density is proportional to relative probability such that higher saturation indicates higher relative probability in  $t$ - $T$  space. Expected model results and corresponding 95% credible intervals are shown by cyan (highest elevation sample in each profile) and red (lowest elevation sample in each profile) envelopes. “N” refers to the number of samples in each transect for the respective thermochronometer, while “n” refers to the total number of aliquots. Constraints imposed on thermal history models are shown as black boxes (described in methods section). Vertical bars show the approximate timing of geologic events potentially responsible for major aspects of the modeled  $t$ - $T$  histories. The dashed line for the Belt Basin deposition event divides poorly-constrained deposition of Lower Belt Supergroup units of the Helena Embayment from better-constrained deposition of the Ravalli through Missoula groups in the main Belt Basin. Geologic event references: (1) Condit et al. (2015) and references therein; (2) Lonn et al. (2020) and references therein; Hirtz et al. (2022); (3) Brennan et al. (2020), Link et al. (2017), Lund et al. (2010), and Yonkee et al. (2014); (4) Hoffman et al. (1998, 2017), Hoffman and Schrag (2002), Kirschvink (1992), Prave et al. (2016), Rooney et al. (2015), and Walker et al. (1981).



**Figure 7.** QTQt  $t$ - $T$  inversions of new and previously reported thermochronology data resolve Phanerozoic cooling and reheating events. The  $\leq 800$  Ma and  $\leq 250^\circ\text{C}$  portion of  $t$ - $T$  inversions are shown for the (a) Beaten Path, (b) Granite Peak, (c) Chalice Peak, (d) Clark's Fork Canyon (Carrapa et al., 2019), and (e) Beartooth Highway/Amoco No. 1 borehole (Omar et al., 1994; Peyton et al., 2012) transects. All models enforced geologic constraints (black boxes). Stippled regions show time-temperature space poorly constrained by AFT/AHe-only models (panels d and e). Path densities and expected model results are symbolized as in Figure 6. (f) Results of decompacted sediment accumulation models: compacted (blue envelopes) and decompacted (red envelopes) sedimentary thicknesses are defined by the minimum (upper envelope bound) and maximum (lower envelope bound) thickness of preserved stratigraphy near the study area. Dashed lines indicate hiatuses in the stratigraphic record. Black lines show modeled reference maximum burial depths for samples originally 2 and 4 km below the sub-Cambrian unconformity surface (i.e., 2 or 4 km of basement overburden). Temperature scale assumes  $20^\circ\text{C}$  surface temperature and a  $30^\circ\text{C}/\text{km}$  geothermal gradient. Shaded vertical bars show the approximate timing of geologic events potentially impacting the thermal history of the region. Geologic event references: (1) Beranek et al. (2016), Blakey and Ranney (2018), and Dorobek et al. (1991); (2) Kluth and Coney (1981) and Maughan (1990); (3) DeCelles (2004), DeCelles and Burden (1992), and Fuentes et al. (2011); (4) DeCelles (1986) and Schwartz and DeCelles (1988); (5) Coney and Reynolds (1977) and Copeland et al. (2017). Kk—Cretaceous Kootenai Formation; unc.—unconformity.



**Figure 8.** Observed and predicted ZHe dates for unconstrained and constrained models for the Beaten Path (a, b), Granite Peak (b, c), and Chalice Peak (e, f) transects (inversion results in Figures 6 and 7a–7c) are shown with respect to eU. Insets show predicted AHe dates and  $^{40}\text{Ar}/^{39}\text{Ar}$  ages, grouped by sample. Observed and predicted AHe and apatite fission track (AFT) dates and AFT mean track lengths (MTL) for constrained models from the Beartooth Pass/Amoco No. 1 Borehole (g) and Clark's Fork Canyon (h) transects (inversion results in Figures 7d and 7e) are plotted with respect to elevation. Colored bars without ticks show reported uncertainty. Red bars with ticks show 95% credible range of dates sampled by post-burn-in models (i.e., predicted dates of the stationary distribution). For samples with AFT length data (g, h), orange bars with ticks show 95% credible range of AFT MTL sampled by post-burn-in models. For readability, AHe dates in (h) are shown as the mean of all sample aliquots. Observed AHe dates for individual aliquots are shown in Figure 5. Note that plotted  $[\text{U-Th}(-\text{Sm})]/\text{He}$  dates are uncorrected for  $\alpha$ -ejection. This is the default input and output of QTQt, which handles corrections during modeling. Observed ZHe and AHe dates corrected for  $\alpha$ -ejection are provided in Figures 4 and 5, respectively.

post-crystallization cooling to  $<150^{\circ}\text{C}$  by  $\sim 2.3$  Ga followed by reheating to  $\sim 250^{\circ}\text{C}$  by  $\sim 2.0$  Ga. Subsequent cooling to  $<200^{\circ}\text{C}$  by  $\sim 1.8$  Ga is followed by reheating to  $\geq 200^{\circ}\text{C}$  by  $\sim 1.4$  Ga (Figure 6b). Model resolution throughout the remainder of the Proterozoic is relatively poor, with subsets of paths indicating either a cooling-reheating-cooling trajectory or monotonic cooling until Cambrian time. The most probable solution suggests relatively slow cooling to  $\sim 150^{\circ}\text{C} \pm 25^{\circ}\text{C}$  between  $1.45 \pm 0.5$  and  $0.8 \pm 0.2$  Ga, followed by more rapid cooling to  $<100^{\circ}\text{C}$  by Cambrian time (Figure 6b). The unconstrained model for the Beaten Path is less well resolved. In contrast to the constrained model, it indicates a reheating event at  $\sim 2.4$  Ga and does not resolve a reheating-cooling inflection point at  $\sim 1.4$  Ga (Figure 6a). However, the unconstrained model similarly resolves Orosirian and Tonian through Cambrian cooling, suggesting these events are directly informed by the thermochronologic data and not an artifact of imposed model boundary conditions.

The fit between observed and predicted thermochronometer dates is similar for unconstrained and constrained models for the Beaten Path transect, matching the observed date-eU relationship of the ZHe data well (Figures 7a and 7b). Both models achieve good fits to the biotite  $^{40}\text{Ar}/^{39}\text{Ar}$  age of sample BP-05, but the unconstrained model produces a better fit for sample BP-01. This thermochronometer is partially reset by  $\sim 2.0$  Ga reheating in the constrained model, resulting in a predicted age  $\sim 0.2$  Ga younger than that observed (Figures 6b and 7b).

Both unconstrained and constrained models for the Granite Peak transect suggest reheating to  $>200^{\circ}\text{C}$  at  $\sim 1.3$ – $1.4$  Ga (Figures 6c and 6d). This reheating event apparently obscures the earlier Precambrian history. Both models suggest temperatures of  $\leq 200^{\circ}\text{C}$  post- $1.3$  Ga with cooling to  $<150^{\circ}\text{C}$  by early Paleozoic time. The trajectory of this cooling is relatively poorly resolved, with a subset of paths in the unconstrained model suggesting latest Neoproterozoic–early Paleozoic cooling (Figure 6c). However, these paths are largely inconsistent with geologic evidence suggesting near-surface temperatures during deposition of the Cambrian Flathead Sandstone (e.g., Simons & Armbrustmacher, 1976). Enforcing this constraint results in poor resolution of late Precambrian cooling (Figure 6d). Predicted biotite  $^{40}\text{Ar}/^{39}\text{Ar}$  ages overlap at uncertainty with those observed, and predicted ZHe dates generally reproduce the observed date-eU relationship (Figures 7c and 7d). However, predicted dates for the two lowest-eU grains, both from the lowest elevation sample, are  $\sim 50$ – $150$  Myr younger than observed.

Temperatures of  $\sim 300^{\circ}\text{C}$  at  $\sim 1.9 \pm 0.1$  Ga are indicated by both unconstrained and constrained models for the Chalice Peak transect (Figures 6e and 6f). These temperatures are required to reproduce observed Statherian biotite  $^{40}\text{Ar}/^{39}\text{Ar}$  ages (Figures 7e and 7f). Notably, the unconstrained model better reproduces the  $^{40}\text{Ar}/^{39}\text{Ar}$  age of sample CP-05 (Figure 7e). Subsequent cooling to  $<250^{\circ}\text{C}$  by  $\sim 1.75$  Ga is indicated by both models. The unconstrained model is poorly resolved throughout the remainder of Proterozoic time (Figure 6e), while the constrained model indicates reheating to  $>200^{\circ}\text{C}$  by  $1.25 \pm 0.5$  Ga and subsequent, poorly constrained cooling (Figure 6f). While ZHe predictions of both models reproduce the observed steeply negative date-eU relationship at  $<400$  ppm eU, predicted dates are significantly younger than observed for a cluster of 5 zircons at  $\sim 500$ – $800$  ppm eU. Additionally, zero dates predicted for zircons of  $>2,000$  ppm eU conflict with non-zero observed dates (Figures 7e and 7f).

#### 4.5.2. Modeled Phanerozoic Thermal History

Model results suggest that at least two distinct Phanerozoic reheating-cooling events best reproduce the observed thermochronologic data (Figures 7 and 8). Following early Paleozoic reheating, a subset of paths of the constrained Granite Peak and Chalice Peak models suggest an episode of Paleozoic cooling, initiating between 500 and 300 Ma (Figures 7b and 7c). This cooling event is resolved by both unconstrained and constrained models for the Chalice Peak transect (Figures 6e, 6f, and 7c), indicating it is directly informed by the data. By contrast, the unconstrained model for the Granite Peak transect produces a good fit to the data without requiring Paleozoic initiation of cooling, suggesting rapid early Paleozoic reheating followed by cooling indicated by a subset of paths in the Granite Peak constrained model may be an artifact of imposed constraints (Figure 6c). The Beaten Path and Granite Peak transects indicate cooling during Jurassic time, initiating  $200 \pm 20$  Ma (Figures 6a–6d, Figures 7a and 7b). All transects additionally resolve cooling during Late Cretaceous and/or Paleocene time (Figure 7). Following Late Cretaceous–Paleocene cooling, the Beaten Path and Granite Peak transects indicate slow cooling while the Chalice Peak transect suggests an episode of rapid reheating and cooling between  $\sim 60$  and  $\sim 40$  Ma (Figure 7c).

In addition to our newly reported transects, we modeled 3 AFT dates and one set of track lengths from the Clark's Fork Canyon transect of Carrapa et al. (2019; Figure 7d). Model results are similar to those of Carrapa

et al. (2019), who conducted a single-sample HeFTy model of their sample CF4. Their best-fit model solution suggested rapid cooling at  $\sim 110$  Ma followed by reheating to  $\sim 80^\circ\text{C}$  between 40 and 20 Ma followed by monotonic cooling to surface temperatures at the present. Our results demonstrate this thermal history produces a good fit to observed AFT dates of the entire Clark's Fork transect modeled as a vertical profile (Figure 8g).

Finally, we modeled an extensive set of AFT dates and track lengths ( $N = 26$ ,  $n = 2,438$ ) and AHe dates ( $N = 23$ ,  $n = 70$ ) collected from the subsurface in the Amoco No. 1 Beartooth borehole and surface exposures near the Beartooth Highway by Omar et al. (1994) and Peyton et al. (2012; Figure 7e). Omar et al. (1994) modeled the post-80 Ma thermal history of samples from two end-member elevations using the Monte-Carlo approach of Lutz and Omar (1991), which suggested rapid cooling between  $\sim 60$  and 55 Ma followed by either residence at or reburial to  $\geq 50^\circ\text{C}$  until  $\sim 10$  Ma, after which samples cooled to the surface. Peyton et al. (2012) inverted three aliquots from one sample in HeFTy, which indicated  $\geq 30^\circ\text{C}$  of cooling between 74 and 41 Ma and didn't require later reheating. Our inversion approach combines samples from both previous studies into a single vertical profile. Our models indicate very rapid cooling between  $\sim 62$  and  $\sim 50$  Ma (Figure 7e), generally consistent with previous models (Omar et al., 1994; Peyton et al., 2012). However, our models apparently achieve greater sensitivity to the Cenozoic thermal history, favoring reburial to, rather than residence at, temperatures of  $\geq 30^\circ\text{C}$  by 6 Ma. Most significantly, our inversion resolves a major cooling event initiating in Devonian time, not recognized by previous modeling efforts (Figure 7g). Paleozoic cooling is required to reproduce AFT dates of up to  $282 \pm 16$  Ma and short MTLs in samples above 3 km ASL, which define a nearly flat date-elevation relationship (Figure 8h; we note that Cervený (1990) documented dates of up to  $348 \pm 31$  Ma at similar elevation, but we were unable to model his data due to insufficient reporting). Although Omar et al. (1994) reported Paleozoic AFT ages, they began their models at 80 Ma and consequently did not resolve the Paleozoic cooling recorded by their samples. The full model results for these transects (i.e., 2,812–0 Ma) are available in Figure S4 in Supporting Information S1; however, Precambrian thermal events therein are unresolved.

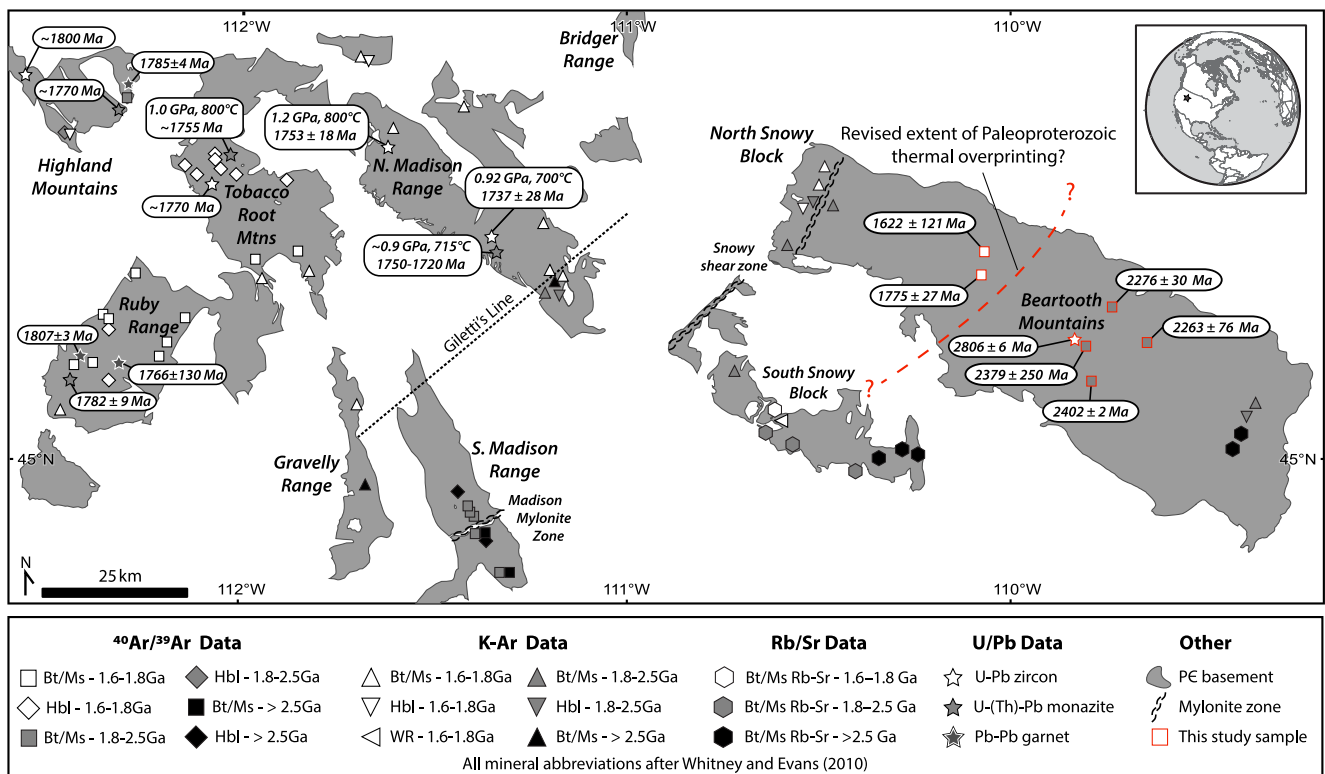
To assess the compatibility of the results of  $t$ - $T$  inversions with the Phanerozoic sedimentary record preserved in nearby areas, we constructed a set of decompacted sediment accumulation models (Figure 7f). Details on the setup of decompacted sediment accumulation models are available in Text S5 and Table S3 in Supporting Information S1. Results show that the total decompacted thickness of all Phanerozoic sedimentary rocks is 4.13–5.47 km. Less than 0.57 km of sediment was deposited prior to Paleozoic cooling resolved by the Chalice Peak and Beartooth Highway/Amoco No. 1 borehole transects, which initiated during an Ordovician–Devonian depositional hiatus ( $\sim 444$ –384 Ma) and persisted throughout the regional deposition of 0.78–0.86 km of Devonian–Pennsylvanian sediment (Figure 7). Assuming a typical geothermal gradient, early Paleozoic reheating temperatures indicated by these models cannot be accounted for based on burial reheating consistent with preserved stratigraphy alone, requiring additional overburden of unpreserved sediments or eroded basement. Initiation of cooling of the Beaten Path and Granite Peak transects at  $\sim 200$  Ma occurred during a Triassic–late Jurassic deposition hiatus. Modeled Cretaceous reheating of the Beaten Path, Chalice Peak, and Beartooth Highway/Amoco No. 1 borehole transects is consistent with reburial by foreland basin deposits, which total 2.83–3.99 km of decompacted thickness. These results inform our interpretations of modeled Phanerozoic thermal events, discussed below.

## 5. Discussion

### 5.1. The Extent of Paleoproterozoic Thermotectonism in Southwest Montana

Biotite  $^{40}\text{Ar}/^{39}\text{Ar}$  results from samples CP-01 and CP-05 comprise the easternmost instance of  $<1.8$  Ga medium-temperature thermochronometer ages recognized in southwestern Montana and the first from the BBMP (Figures 1, 3, and 9; references in Figure 9 caption). These ages partially overlap with Paleoproterozoic ( $\sim 1.78$ –1.72 Ga) thermotectonism observed within the Montana metasedimentary province, west of the Beartooth Mountains, where peak metamorphic conditions of  $\sim 0.8$ –1.2 Gpa and  $>700^\circ\text{C}$  have been documented (Figure 9; e.g., Ault et al., 2012; Condit et al., 2015; Harms, Brady, et al., 2004). This episode of thermotectonism, known as the Big Sky orogeny, has been attributed to collision between the Wyoming and Medicine Hat cratons (i.e., part of the wider Great Falls orogeny; Mueller et al., 2002) and/or between the Wyoming craton and an arc terrane to the west (i.e., Selway terrane of Foster et al., 2006).

Previously documented evidence for thermal effects associated with the Big Sky orogeny is dominantly limited to basement exposures northwest of a northeast-trending transitional boundary, commonly referred to as “Giletti's



**Figure 9.** New and existing geochronologic and thermochronologic data constrain the extent of late Paleoproterozoic thermotectonism affecting basement rocks (gray polygons) in southwest Montana. Data from this study are outlined in red. Biotite  $^{40}\text{Ar}/^{39}\text{Ar}$  data from this study are reported as plateau ages and associated uncertainty. Red dashed line denotes a schematic revised extent of Paleoproterozoic thermal overprinting suggested by our data. Data citations:  $^{40}\text{K}/^{39}\text{Ar}$ —Giletti (1966, 1971) and Reid et al. (1975);  $^{40}\text{Ar}/^{39}\text{Ar}$ —Erslev and Sutter (1990), Harlan et al. (1996), Roberts et al. (2002), Brady et al. (2004), Hames and Harms (2013), and this study; zircon U-Pb—Ault et al. (2012), Carrapa et al. (2019), Condit et al. (2015), and this study; monazite U-Th-Pb—Cheney et al. (2004), Alcock et al. (2013), and Condit et al. (2015); garnet Pb-Pb—Roberts et al. (2002), Cheney et al. (2004), Ault et al. (2012), and Condit et al. (2015); Rb-Sr—Reid et al. (1975) and Montgomery and Lytwyn (1984). WR—Whole Rock. Adapted from Condit et al. (2015).

line” (Figure 9; Condit et al., 2015; Giletti, 1966). However, excess Ar was later identified in samples from key localities originally used to define this boundary (Giletti, 1971). Biotite  $^{40}\text{Ar}/^{39}\text{Ar}$  ages of <1.8 Ga observed in samples CP-01 and CP-05 require late Paleoproterozoic residence of these samples at temperatures above the biotite  $^{40}\text{Ar}/^{39}\text{Ar}$  closure temperature (>300°C; Grove & Harrison, 1996; Figures 2 and 9). Constrained time-temperature inversions of the Beaten Path transect resolve an inflection from cooling to reheating approximately temporally coincident with the Big Sky orogeny (Figure 6b), suggesting this event may have also influenced the thermal history of eastern portions of the Beartooth Mountains. These results—in addition to previously documented partially reset K-Ar and Rb-Sr ages in the North and South Snowy blocks (Montgomery & Lytwyn, 1984; Reid et al., 1975)—suggest that reinterpretation of the position and nature of Giletti’s line is warranted. Significant uncertainty remains regarding the extent and nature of late Paleoproterozoic tectonism in Montana and associated spatial patterns in medium-temperature thermochronometer ages (e.g., Condit et al., 2015; Harms, Burger, et al., 2004; Mueller et al., 2004). Further detailed thermochronologic work in the region may help resolve this period of uncertainty, contributing to improved understanding of the evolution of ancient orogenic systems.

## 5.2. Potential Triggers of Mesoproterozoic–Neoproterozoic Thermal Events

Constrained models for all newly reported transects resolve reheating to  $\geq 200^\circ\text{C}$  by early Ectasian time followed by cooling to <100°C between late Ectasian and early Paleozoic time (Figures 6b, 6d, and 6f). In the following section, we discuss possible mechanisms driving inflections in this portion of the modeled thermal history. However, we note that unconstrained models do not ubiquitously resolve these inflections, suggesting they are not strictly required by the data. We focus discussion on the best resolved model aspects but emphasize interpretive caution given large model uncertainties.

The Beaten Path and Granite Peak transect suggest peak reheating by  $\sim 1.45$  Ga, while that for the Chalice Peak transect occurs slightly later. The timing of this modeled reheating overlaps with deposition of Belt Supergroup units within the Helena Embayment (Figure 1), bracketed between 1.71 Ga and 1.42 Ga (e.g., Anderson & Davis, 1995; Hirtz et al., 2022; Mueller et al., 2016; Ross & Villeneuve, 2003). Preserved Mesoproterozoic sedimentary rocks in the Helena Embayment are structurally bound by the Perry line, a system of north-dipping Mesoproterozoic normal faults tracing  $\sim 50$  km north of the Beartooth Mountains (e.g., Hatcher et al., 1987; McMannis, 1963; Ross & Villeneuve, 2003; Winston, 1986). Despite this, the occurrence of Mesoproterozoic talc deposits  $\sim 46$ – $80$  km south of the Perry line—defining a WNW–ESE striking “corridor”  $\sim 150$  km southeast of the Beartooth Mountains—has been interpreted as evidence that the eastern Belt Basin originally extended south of this structural boundary (Anderson et al., 1990; Underwood et al., 2014). The detrital zircon record from the eastern Belt Supergroup is consistent with this interpretation. A paucity of  $\sim 2.8$  Ga zircons in eastern Belt Supergroup units suggests the BBMP was not an important sediment source to the basin (Fox, 2017; Mueller et al., 2016; Ronemus et al., 2020), consistent with a lack of erosion of BBMP basement during Belt Supergroup deposition. Coupled with these lines of evidence, modeled Mesoproterozoic reheating of rocks in the Beartooth Mountains may support Mesoproterozoic deposition within the study area (Figure 6). In this context, the preserved extent of Belt Supergroup sediments may not reflect the original extent of the basin, but rather that of later erosional truncation. Later faulting along the Perry line may have contributed to erosional beveling of Mesoproterozoic rocks deposited in the hanging-wall of the fault system. We note that the biotite  $^{40}\text{Ar}/^{39}\text{Ar}$  ages of our samples were not reset in Mesoproterozoic time (Figure 3), and models indicate  $\leq 125^\circ\text{C}$  of Mesoproterozoic reheating (Figure 6). Therefore, any Mesoproterozoic sediments deposited at the location of the study area must have been substantially thinner than the  $\sim 15$  km of Mesoproterozoic stratigraphy preserved in western portions of the Belt Basin.

Following potential Mesoproterozoic reheating, constrained thermal history models indicate samples cooled from  $>200^\circ\text{C}$  to  $<100^\circ\text{C}$  between  $\sim 1.45$  Ga and early Paleozoic time (Figure 6). Meso–Neoproterozoic cooling events are best resolved by models for the Beaten Path transect, which suggest an initial phase of cooling initiating  $\sim 1.45 \pm 0.5$  Ga and a later inflection to more rapid cooling beginning  $\sim 0.8 \pm 0.2$  Ga (Figure 6b). Only the former phase of cooling is well resolved by models for other transects (Figures 6d and 6f). This initial phase approximately coincides with the  $\sim 1.4$  Ga rifting of western Laurentia associated with breakup of supercontinent Nuna, inferred from a reported disappearance of non-Laurentian detrital zircons in the upper Belt Supergroup (e.g., Ross & Villeneuve, 2003; Stewart et al., 2010; see recent debate in Parker & Hendrix, 2022). Similarly, the modeled inflection to more rapid cooling in Neoproterozoic time overlaps with an additional  $\sim 780$  to  $\leq 680$  Ma phase of rifting of the western Laurentian margin associated with the breakup of supercontinent Rodinia (e.g., Brennan et al., 2020; Li et al., 2013; Link et al., 2017; Lund et al., 2010; Merdith et al., 2019).

Collectively, the observation that both modeled Meso–Neoproterozoic cooling events coincide with the timing of hypothesized rifting of the western Laurentian margin is consistent with a causal relationship between marginal rifting and the exhumation of cratonward basement rock. Resolving  $\sim 850$ – $550$  Ma cooling for the Madison and Ruby–Gravelly mountain ranges,  $75$ – $120$  km west of the present study area, Kaempfer et al. (2021) suggested normal faulting associated with the rifting of western Laurentia caused erosional exhumation of horst blocks to the east. They hypothesized southwestern Montana was an eroding topographic high throughout late Neoproterozoic–early Cambrian time. Deep-time thermal history models from the Grand Canyon have similarly been interpreted to suggest a linkage between Rodinia breakup and erosion of the footwall blocks of rift-related normal faults (Peak et al., 2021), although later work showed this phase of cooling may not be required to reproduce observed ZHe dates (Thurston et al., 2022). Similar rifting processes associated with Nuna breakup may have triggered erosional exhumation of southwest Montana basement rock in Mesoproterozoic time as well, although samples apparently experienced reheating during initial phases of Mesoproterozoic normal faulting. We note that inversions of Kaempfer et al. (2021) did not resolve significant Mesoproterozoic reheating or cooling, but this may reflect that their model setup allowed only monotonic cooling during Proterozoic time. Additionally, their data set apparently lacked sensitivity to Mesoproterozoic thermal events, as demonstrated by forward models.

Notably, rifting related to supercontinent breakup is not the only process capable of explaining Neoproterozoic continental exhumation. The modeled cooling timing of basement rock in southwest Montana also temporally overlaps at model resolution with Cryogenian snowball Earth ice-sheet glaciation ( $\sim 720$ – $635$  Ma; Godd  ris et al., 2003; Kaempfer et al., 2021; Li et al., 2013). Recent QTQt  $t$ – $T$  inversions by McDannell et al. (2022; remodeling data from DeLucia et al., 2018, Flowers et al., 2020, and McDannell & Flowers, 2020) highlighted

that late Neoproterozoic cooling was likely not isolated to regions near the western Laurentian margin. Rather, they resolved a similar timing of cooling for basement rock in interior regions of Laurentia, which would have been insulated from the thermal effects of rifting. Therefore, McDannell et al. (2022) argued that erosion associated with snowball Earth ice-sheet glaciation is more consistent with late Neoproterozoic exhumation across both marginal and interior regions of the continent.

In the absence of independent Neoproterozoic geologic constraints in the study area, the resolution of the present models is not sufficient to distinguish between the, notably temporally overlapping, potential cooling mechanisms of Laurentian rifting and snowball Earth glaciations. While continental-scale syntheses may favor ice sheet glaciation as a primary driver in interior portions of Laurentia (i.e., McDannell et al., 2022), rifting and glaciation are not mutually exclusive agents of erosional exhumation. Glaciers most effectively erode regions with preexisting topography, including topography generated by rift-related faulting (e.g., Walsh et al., 2019). Consequently, as most recently noted by McDannell et al. (2022), the “erosional synergy” between rift-related tectonism and ice sheet erosion remains a viable mechanism to explain rapid Neoproterozoic cooling and associated km-scale erosion in regions near the Laurentian margin, such as southwest Montana.

### 5.3. The Paleozoic Beartooth Mountains

Time-temperature inversions for the Chalice Peak and Beartooth Highway/Amoco No. 1 borehole transects indicate a phase of cooling initiating in early to middle Paleozoic time (Figures 7c and 7e). Modeled cooling temporally overlaps with the Devonian–Early Mississippian Antler orogeny and the Pennsylvanian–Permian ARM orogenic event (e.g., Beranek et al., 2016; Blakey & Ranney, 2018; Dorobek et al., 1991; Kluth & Coney, 1981; Maughan, 1990). Although the ARM event is widely credited as triggering intraforeland basement uplift from Oklahoma through Colorado (e.g., Kluth & Coney, 1981), its effects have been more sparsely documented further north (e.g., Maughan, 1990). Existing evidence includes southward thinning of Mississippian–Pennsylvanian units toward the Montana–Wyoming border, evident in isopach patterns (Maughan, 1993). Additionally, Maughan (1990) documented erosional unconformities and lithologic facies indicative of proximal terrigenous sediment sources in Pennsylvanian rocks near the Beartooth Mountains. This led him to suggest the existence of a subaerially exposed Pennsylvanian high approximately spatially coincident with the present study area, comprising one of the northernmost ARM uplifts. Our thermal history results support this hypothesis, consistent with active erosional exhumation of this high (Figures 7c and 7e).

While a Paleozoic cooling event is clearly required to reproduce AFT data of the Beartooth Highway transect (Figures 7e and 8h), the record thereof has gone unrecognized despite previous modeling by Omar et al. (1994). Our construction of a sufficiently deep-time thermal history model incorporating both AFT and AHe samples in a vertical profile and enforcing geologic constraints facilitated resolution of this newly documented cooling event from these previously reported data (Omar et al., 1994; Peyton et al., 2012). Similar modern inversion techniques applied to data sets from other nearby basement-cored ranges (e.g., Cerveny, 1990; Cerveny & Steidtmann, 1993; Peyton et al., 2012) may help further constrain cryptic Paleozoic cooling in the northern US Rocky Mountains.

### 5.4. Mesozoic–Cenozoic Growth of an Intraforeland Basement Uplift

Models for the Beaten Path and Granite Peak transects suggest the initiation of a cooling event near the Triassic–Jurassic transition (Figures 7a and 7b). There is little evidence for the influence of active tectonism in the southern Montana region during this time, with documented Early Jurassic shortening dominantly limited to the Luning–Fencemaker fold-and-thrust belt of northern Nevada and southeastern Idaho (e.g., Wyld, 2002). Instead, this cooling may be related to eustasy rather than tectonics; the Triassic–Jurassic transition coincides with the lowest global sea levels of Paleozoic–Mesozoic time (Marcilly et al., 2022). Low sea levels coupled with a lack of tectonic subsidence facilitated sub-aerial erosion of the continental margin, recorded by a widespread Permian or Triassic to Late Jurassic hiatus in the regional sedimentary record (e.g., Maughan, 1993). Alternatively, modeled Jurassic cooling may comprise a model artifact associated with underestimation of [U-Th(-Sm)]/He date uncertainties; models with larger uncertainty rescaling parameters tend to suggest Paleozoic and/or Cretaceous, rather than Jurassic, cooling events for these transects (Figure S4 in Supporting Information S1).

All models resolve rapid cooling ( $\sim 1.2\text{--}20^\circ\text{C/Myr}$ ) initiating during Early Cretaceous to Paleocene time (Figure 7). The Clark's Fork Canyon transect indicates the earliest cooling, initiating  $\geq 110$  Ma (Figure 7d;

Carrapa et al., 2019). This cooling significantly precedes that indicated by other transects and consequently may reflect localized exhumation which did not necessarily affect other portions of the range (Figure 10a). Structural analysis and mapping by Wise (1983) and DeCelles, Gray, Ridgway, Cole, Srivastava, et al. (1991) demonstrate that NW-SE striking faults—such as those defining the Cooke City sag zone near the southeastern margin of the range—were truncated by later slip along the main range-bounding fault system. The Clark's Fork canyon transect is located in the proximal hanging-wall of both fault systems. By contrast, the nearby Beartooth Highway/Amoco No. 1 borehole transect—recording cooling initiation  $\geq 50$  Myr later—resides in the proximal hanging-wall of only the main range-bounding fault system. Therefore, early cooling recorded by the Clark's Fork Canyon transect may be related to slip on the NW-SE striking system. These faults, oriented sub-parallel to the axis of the Bighorn Basin, may have slipped in response to flexure of the basin during incipient foredeep subsidence and/or during passage of the flexural forebulge, recorded by a regional time-transgressive unconformity at the base of the Kootenai and Cloverly formations (e.g., DeCelles, 2004; DeCelles & Burden, 1992). Incipient Aptian–Albian intra-foreland uplift is consistent with paleocurrent deflections around the periphery of several southwest Montana basement uplifts—including the Beartooth Mountains—observed in the Lower Cretaceous Kootenai Formation (DeCelles, 1986). However, this interpretation was recently disputed by detrital zircon provenance work  $>100$  km west of the study area, which did not find evidence for basement exhumation or significant partitioning of the foreland basin in southwestern-most Montana during Kootenai Formation deposition (Rosenblume et al., 2021).

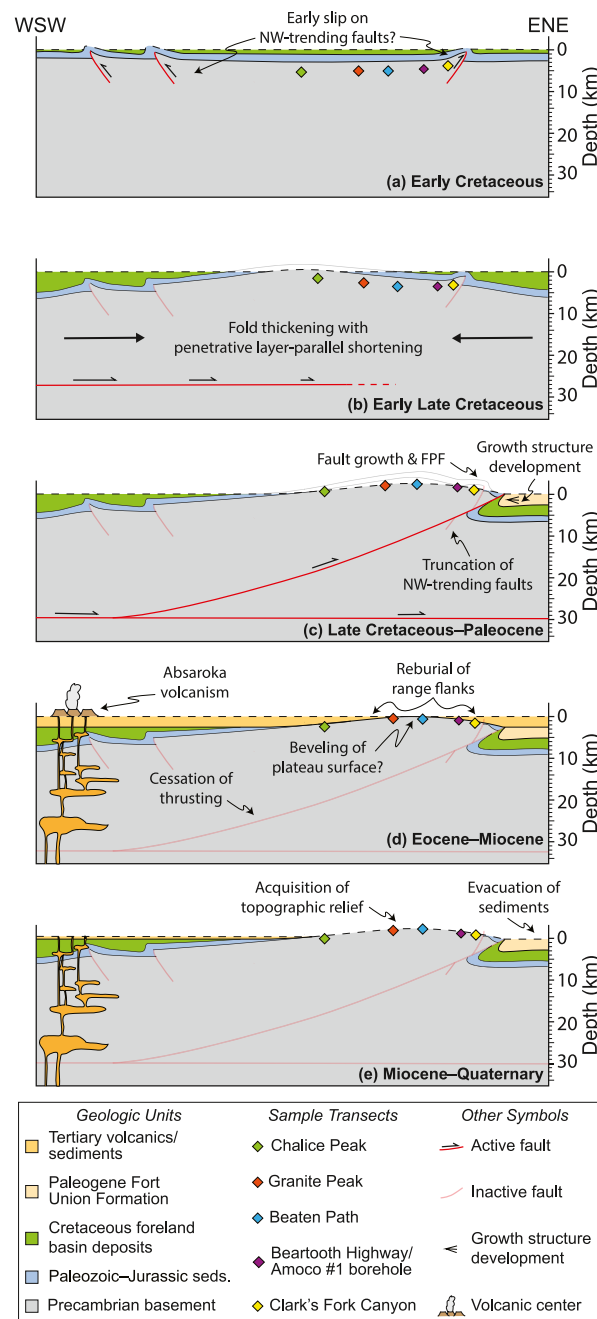
Apart from the Clark's Fork Canyon transect, the timing of modeled cooling initiation generally decreases from west to east; the Chalice Peak and Granite Peak transects indicate an inflection to rapid cooling at  $\sim 90$  Ma, while the Beaten Path and Beartooth Highway transects don't resolve the initiation of this cooling event until  $\sim 60$  Ma (Figure 8). A decreasing age of cooling in the direction of fault growth and transport is compatible with models conceptualizing the growth of intraforeland basement arches through an overlapping sequence of detachment folding and layer-parallel shortening followed by fault-propagation and fault-bend folding (Figures 10b and 10c; e.g., Erslev et al., 2022; Weil & Yonkee, 2012).

Initial growth of the Beartooth uplift by the former processes may have localized initial uplift and exhumation in eastern portions of the range, consistent with  $\sim 95$  Ma cooling resolved by models for the Chalice Peak and Granite Peak transects (Figure 10b). Thinning of Coniacian–Early Campanian foreland basin deposits near the incipient Beartooth uplift, evident from isopach patterns, supports the development of positive relief by at least  $\sim 90$  Ma (Eichler et al., 2020). However, there is no conclusive record of sediment eroding off the Beartooth Mountains until Paleocene time (DeCelles, Gray, Ridgway, Cole, Srivastava, et al., 1991). Earlier uplift of the Beartooth Mountains may have entailed recycling of mainly fine-grained deposits, which were bypassed to distal regions. Proximal accumulation likely did not occur until erosion had exhumed more resistant lithologies and/or the uplift had reached sufficient size to trigger localized flexural subsidence on its periphery (DeCelles, Gray, Ridgway, Cole, Srivastava, et al., 1991; Dickinson et al., 1988).

During later growth of the main Beartooth Fault, NNE-directed thrusting and associated fault-propagation folding would have focused exhumation on the steep northeastern forelimb (Figure 10c). There, the Beaten Path and Beartooth Highway/Amoco No. 1 borehole transects record rapid cooling between  $\sim 62$  and  $\sim 50$  Ma, coeval with the timing of growth structure development in the Beartooth Conglomerate (DeCelles, Gray, Ridgway, Cole, Srivastava, et al., 1991; Figure 10c). The lack of major surface-breaking faulting until Paleocene time is further supported by cross-cutting of Turonian–Coniacian (Cody Shale;  $\leq 96.5$  Ma; May et al., 2013; Wise, 2000) and likely late Maastrichtian (Lance Formation;  $\leq 67$  Ma; Hicks et al., 2002; DeCelles, Gray, Ridgway, Cole, Srivastava, et al., 1991) rocks by the main range-bounding fault. This thrusting folded Phanerozoic strata into a recumbent and tectonically attenuated footwall syncline and generated complex structures at the northeast corner of the uplift (e.g., Wise, 2000), comprising a crescendo in the uplift of the Beartooth Mountains (Figure 10c).

### 5.5. Implications for Models of Laramide Tectonism

In the previous sections, we suggested that Phanerozoic exhumation and associated cooling in the Beartooth Mountains was a result of protracted tectonism on the margins of Laurentia spanning  $>300$  Myr and culminating in the development of the North American Cordillera. While the previously documented Late Cretaceous–Paleocene phase of rapid cooling was the most significant of these events (e.g., DeCelles, Gray, Ridgway, Cole, Srivastava, et al., 1991; Omar et al., 1994; Peyton et al., 2012), it likely records only an inflection superimposed



**Figure 10.** Schematic kinematic evolution of the Beartooth Mountains consistent with results of thermal history inversions. (a) Early Cretaceous deformation involved low-magnitude slip on thrust faults with a dominantly NW–SE trend. Samples in the Clark's Fork Canyon transect, located in the proximal hanging-wall of one such fault, experienced cooling. (b) Early fault-related folding, likely related to detachment folding and layer-parallel slip, uplifted and exhumed rocks in the southwestern portions of the range, recorded by the Chalice Peak and Granite Peak transects as cooling initiating ~95 Ma. Exhumation was poorly preserved in the sedimentary record due to sediment bypassing (e.g., DeCelles, Gray, Ridgway, Cole, Srivastava, et al., 1991; Hoy & Ridgway, 1997). (c) Growth of the main Beartooth fault triggered fault-propagation and fault-bend folding. Transects in the northeastern portion of the range experienced rapid exhumation and cooling. Deformation was recorded by the development of growth structures in the Paleogene Beartooth Conglomerate (DeCelles, Gray, Ridgway, Cole, Srivastava, et al., 1991). Total structural relief of  $\geq 8$  km was achieved (Blackstone, 1986). (d) Shortening ceased by early Eocene time. Reburial by Eocene Absaroka volcanics and/or by Oligocene–Miocene sediments caused reheating of transects near the flanks of the range. The Beartooth Plateau paleo-surface may reflect beveling of high topography to the regional basin elevation during this time. (e) Evacuation of Cenozoic sediments from the foreland re-exhumed the range flanks, resulting in the acquisition of significant topographic relief. Scales at right reflect depth below regional basin elevation at each time step. The growth of the Beartooth uplift in (b) and (c) is based after two-stage model of Erslev et al. (2022). Depth to detachment is schematic.

on a longer and more complex Phanerozoic cooling history. Here, we discuss potential implications for the tectonic/geodynamic mechanism(s) of intraforeland basement uplift these results entail.

Original consensus on the timing of the “Laramide” intraforeland basement uplift *sensu* Armstrong (1968) and Dickinson and Snyder (1978) favored mid-Campanian (~75 Ma) initiation. In this interpretation, the initiation of basement deformation was approximately coeval with the hypothesized timing of shallowing of the Farallon slab below North America, as recorded in part by shutdown and eastward migration of the Sierra Nevada arc and localized deep foreland subsidence (e.g., Coney & Reynolds, 1977; Cross & Pilger, 1978; Lipman et al., 1971). At the latitude of the study area, apparent northeastward magmatic migration between 80 and 55 Ma has similarly been interpreted to reflect slab flattening (Constenius et al., 2003; Copeland et al., 2017; Lageson et al., 2001; Lund et al., 2002). Due in part to the ostensible similarity in timing between intraforeland basement uplift and proxies for slab flattening, most leading geodynamic models attribute Laramide basement uplift to some variety of interaction between a shallowly subducting Farallon slab and the overriding North American plate, including basal traction from a flat Farallon slab (e.g., Bird, 1984, 1998; Copeland et al., 2017; Dickinson & Snyder, 1978; Heller & Liu, 2016; Lawton, 2019), oceanic plateau(s) thereon (e.g., Liu et al., 2010; Livaccari et al., 1981), or associated hydration and lithospheric weakening of the North American lithosphere (e.g., Humphreys et al., 2003; Saylor et al., 2020).

However, recent timing constraints from southwest Montana and northern Wyoming demonstrate basement uplift in the region likely initiated significantly prior to the inferred timing of regional insertion of a flat slab. Steidtmann and Middleton (1991) presented evidence for uplift in the Wind River Range as early as 110 Ma, which to them “suggest(ed) either a pre-Laramide event of unknown origin or that Laramide crustal shortening began nearly 35 Myr earlier than commonly assumed.” In southwest Montana, sedimentologic and thermochronologic work has produced additional evidence for Aptian–Coniacian exhumation of several basement uplifts—including the Highland, Madison, Blacktail-Snowcrest, and Beartooth mountains (Carrapa et al., 2019; DeCelles, 1986; Garber et al., 2020; Haley, 1985; Ronemus & Orme, 2022; summarized by Orme, 2020). Our results add additional thermochronologic evidence suggesting likely fault-related exhumation of the Beartooth Mountains by at least ~90 Ma. If 80–55 Ma eastward magmatic migration in southwest Montana accurately reflects the timing of regional flat slab subduction, this developing consensus implies initial basement uplift preceded the arrival of a flat slab by  $\geq 10$  Myr (e.g., Carrapa et al., 2019; Garber et al., 2020; this study). This observation presents an obstacle for the former theories.

Other models more consistent with  $\geq 90$  Ma basement uplift include the propagation of basement deformation ahead of the developing flat slab (e.g., Kulik & Schmidt, 1988), potentially as a result of plate margin end loading driving far-field compression (e.g., Axen et al., 2018; Behr & Smith, 2016; Erslev et al., 2022; Jackson et al., 2019; Livaccari & Perry, 1993; Thacker et al., 2022). Alternatively, early basement deformation in southwest Montana may be largely unrelated to flat-slab subduction. In this case, deformation may be linked to injection of ductile lower crust from the fold-thrust belt (e.g., McQuarrie & Chase, 2000), the transmission of hinterland stresses (e.g., Erslev, 1993; Livaccari, 1991), terrane accretion/translation (e.g., Maxson & Tikoff, 1996; Tikoff et al., 2016), and/or the interaction between the retroarc fold-thrust belt and preexisting stratigraphic (Parker & Pearson, 2021) and/or basement architecture (Tavani et al., 2021). Although these models make less specific and/or more compatible predictions for the timing of basement uplift in southwest Montana and northern Wyoming, they have difficulty accounting for observed eastward migrating magmatism, aspects of basement uplift strain distribution, and foreland subsidence patterns (Bird, 1984, 1998; Coney & Reynolds, 1977; Constenius et al., 2003; Erslev et al., 2022). Given the emerging geologic and thermochronologic evidence contesting the dominant Laramide flat-slab paradigm, future regional syntheses examining the timing of basement uplift in southwest Montana are uniquely poised to test these alternative geodynamic models.

### 5.6. Evidence for Eocene–Miocene Reburial and Neogene Development of Relief

Model results indicate the termination of rapid cooling likely associated with basement uplift by ~50 Ma (Figure 7), approximately coincident with a regional shift from compressional to extensional deformation (e.g., Foster & Mark Fanning, 1997; Foster et al., 2001, 2010; Howlett et al., 2021). This transition in tectonic regime is associated with a widespread phase of volcanism (i.e., ignimbrite flare-up; Best et al., 2016). Ignimbrite volcanism is locally represented by the ~53–43 Ma Absaroka Volcanic Supergroup, which is preserved in sections up to several kilometers thick overlying the southern portion of the modern Beartooth Mountains and the Absaroka

Mountains to the south (Figures 2 and 10d; e.g., Feeley, 2003; Feeley & Cosca, 2003; Hiza, 1999). Model results from the Chalice Peak, Clark's Fork Canyon, and Beartooth Highway/Amoco No. 1 borehole transects indicate  $\sim 20^{\circ}\text{C}$ – $45^{\circ}\text{C}$  of reheating in Eocene to Miocene time (Figures 7c–7e). Although Eocene volcanics are not currently preserved outside the southern flanks of the Beartooth Mountains, this may suggest their original extent was more expansive, burying portions of the Beartooth Mountains further north (Figure 10d), as hypothesized by Carrapa et al. (2019). Reheating coeval with eruption of the Absaroka Volcanic Supergroup is best resolved by the Chalice Peak transect, which indicates temperatures of  $110^{\circ}\text{C} \pm 30^{\circ}\text{C}$  at  $50 \pm 5$  Ma (Figure 7c) are required to reproduce a pediment of ZHe dates spanning  $\sim 60$ – $20$  Ma at  $\sim 400$ – $1,200$  ppm eU (Figure 8f).

Model results for the Clark's Fork Canyon and Beartooth Highway/Amoco No. 1 borehole transects indicate Cenozoic reheating continued until at least 20 Ma (Figure 7d) and possibly 6 Ma (Figure 7e), suggesting reheating was longer-lived than Eocene volcanism. Later stages of modeled reheating may be instead associated with burial by Oligocene–Miocene sedimentary rocks, as first proposed in the Beartooth Mountains by Omar et al. (1994). Erosional remnants of rocks of this age (South Pass, White River, and Arikaree formations) have been identified at high elevation ( $\geq 2.2$  km ASL) in the Bighorn and Wind River mountains, WY (McKenna, 1980; McKenna & Love, 1972; Steidtmann & Middleton, 1991; Steidtmann et al., 1989). McKenna and Love (1972) argued that the Bighorn Mountains were buried to the present  $\sim 2.7$  km elevation by these sediments—an interpretation consistent with recent thermochronologic work in the range (Caylor & Carrapa, 2021). Although similar deposits have not been recognized in the Beartooth Mountains, crude extrapolation of McKenna & Love's (1972) estimate of Tertiary basin elevation in the Bighorn Mountains ( $\sim 2.7$  km ASL) to equivalent modern elevations in the present study area yields total burial of the Clark's Fork Canyon transect (1.48–2.23 km ASL) and near-total burial of the Beartooth Highway/Amoco No. 1 borehole transect ( $-1.87$ – $3.33$  km ASL); other transects not resolving Oligocene–Miocene reheating remain largely or wholly above this hypothetical Tertiary basin elevation. Therefore, it is consistent with regional geologic and thermochronologic evidence that at least the eastern flank of the Beartooth Mountains was reburied by Oligocene–Miocene intermontane sedimentary deposits. In this context, the Beartooth Plateau—with an average modern elevation of  $\sim 3$  km—may represent an Oligocene–Miocene paleo-surface, analogous to similar Cenozoic erosional surfaces recognized at high elevation in the Colorado Front Range (Epis & Chapin, 1975; McMillan, 2003 and references therein). In this case, the Oligocene–Miocene relief of the Beartooth Mountains was likely on the order of  $<1$  km, with only peaks of  $>3$  km modern elevation exposed above the basin floor.

The most recent phase of cooling recorded by the Clark's Fork Canyon and Beartooth Highway/Amoco No. 1 borehole transects (Figures 7d and 7e) suggests that the erosion of Cenozoic sedimentary rocks initiated between  $\sim 20$  and 6 Ma. Exhumation likely occurred as these cover rocks were evacuated from the foreland along one or multiple continental-scale paleo-drainage systems (e.g., Corradino et al., 2021; Galloway et al., 2011; Sears, 2013). Considerable debate exists concerning the relative roles of extensional tectonism versus climate change in driving late Cenozoic large-scale drainage reorganization and basin evacuation (e.g., Galloway et al., 2011; McMillan, 2003; McMillan et al., 2006; Sears, 2013). Regardless of the mechanism, our results in addition to other recent thermochronologic data (Caylor & Carrapa, 2021) emphasize that, while the structural relief of ranges in the northern Laramide foreland was likely developed by Paleogene time, their modern topographic relief ( $\geq 2.5$  km in the Beartooth Mountains) is a more recent, Neogene feature.

## 6. Conclusions

We demonstrate the utility of a Bayesian approach constrained by multiple thermochronometers in reconstructing complex deep-time thermal histories involving multiple periods of reheating and cooling. An extensive geo/thermochronologic data set of samples from a range of elevations and positions—including dispersed but systematically correlated ZHe dates—facilitates the exploration of complex thermal events that are unresolvable by conventional geochronologic or thermochronologic techniques alone. Our results show evidence of previously unrecognized thermal events affecting rocks of the Beartooth Mountains and constrain the timing and magnitude of documented burial and exhumation episodes, including:

1. Western regions of Beartooth Mountains experienced late Paleoproterozoic temperatures above the biotite  $^{40}\text{Ar}/^{39}\text{Ar}$  closure window ( $>300^{\circ}\text{C}$ ), likely associated with the Big Sky orogeny (e.g., Condit et al., 2015; Harms, Brady, et al., 2004; Harms et al., 2006). These results suggest the thermal effects of this

- Paleoproterozoic tectonism penetrated significantly (~100 km) further east than previously recognized (e.g., Condit et al., 2015; Giletti, 1966).
2. Early Mesoproterozoic reheating is indicated for all newly reported constrained models. This is consistent with previous hypotheses for deposition of Belt Supergroup sediments south of their preserved extent (Underwood et al., 2014).
  3. Model results suggest late Mesoproterozoic and potentially late Neoproterozoic cooling events. The timing of both events overlaps with separate rifting episodes on the western Laurentian margin, consistent with previous hypotheses for rifting-induced exhumation of cratonic basement in the region (Kaempfer et al., 2021). However, model uncertainty is large and Neoproterozoic cooling overlaps with the timing snowball Earth ice-sheet glaciation, another possible driver of continental erosion (e.g., McDannell et al., 2022).
  4. Models resolve reheating to maximum Phanerozoic temperatures by Devonian time, followed by an episode of later Paleozoic cooling. This cooling is contemporaneous with the Antler and ARM orogenic events, suggesting an episode of tectonic exhumation broadly consistent with sedimentologic and stratigraphic observations (Dorobek et al., 1991; Maughan, 1990). Our data provide the first thermochronologic evidence that the Beartooth Mountains experienced thermal effects associated with Paleozoic tectonism and suggest associated deformation penetrated further into the foreland than formerly appreciated.
  5. A second phase of Phanerozoic cooling records complex, spatially variable Cretaceous–Paleogene exhumation. We suggest Early Cretaceous cooling reported by Carrapa et al. (2019) was localized, possibly due to minor slip of NW–SE striking faults accommodating flexural stresses in the evolving foreland basin. More widespread exhumation of the range likely initiated in Late Cretaceous time and progressed from west to east. Uplift of the Beartooth Mountains reached a crescendo in Paleocene time, when surface-breaking thrusting triggered very rapid exhumation concentrated near the northeastern range front. Fault-related basement uplift likely initiated  $\geq 90$  Ma and preceded the hypothesized arrival of a shallowly subducting slab, adding to a growing number of similar observations requiring reconsideration of canonical flat-slab models for intraforeland basement uplift in the North American Cordillera. Our results help to reconcile previous observations indicating Early Cretaceous uplift of the Beartooth Mountains (e.g., Carrapa et al., 2019; DeCelles, 1986; Schwartz & Decelles, 1988) with those only resolving Late Cretaceous–Paleocene phases of this cooling history (Cerveny, 1990; Omar et al., 1994; Peyton et al., 2012). We suggest the well-documented latter event may reflect only the late stage of a more complex cooling history associated with the development of the orogenic system.
  6. Data from transects near the margins of the main Beartooth block indicate Cenozoic reheating. This reheating was likely due to reburial by Eocene Absaroka volcanics (e.g., Carrapa et al., 2019) and/or Oligocene–Miocene basin fill (e.g., McKenna & Love, 1972; Omar et al., 1994) and suggests much of the modern topographic relief of the Beartooth Mountains was acquired in Neogene time.

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#### Data Availability Statement

Zircon U–Pb, biotite  $^{40}\text{Ar}/^{39}\text{Ar}$ , ZHe, and AHe data sets, alongside supplementary text and figures referenced in the main text, are available in Supporting Information S1 for this paper. Large files supporting this work, comprising QTQt modeling files (Data Set S6) and grain photographs (Data Set S4), are available on Zenodo (<https://doi.org/10.5281/zenodo.7443913>). Analytical data sets may also be accessed using this DOI on Zenodo.

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