

Research Article

Post-Laramide Extension and Erosion in the Madison and Gallatin Ranges of Southwest Montana from Apatite (U-Th-Sm)/He Thermochronometry

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Since the end of the Laramide Orogeny (~50 Ma), southwest Montana has experienced complex tectonic, climatic, volcanic, and mantle dynamic processes that have left an imprint on the landscape. Here, we examine the impact of post-orogenic and recent hotspot-related processes on the landscape by quantifying the Cenozoic exhumation history of the Madison and Gallatin Ranges, located on the northern flank of the Yellowstone hotspot (YSH) in southwest Montana. We apply the apatite (U-Th-Sm)/He (AHe) thermochronometer to Cretaceous and Paleogene intrusions from three transects to constrain the Cenozoic cooling history. We also present three new zircon U-Pb crystallization ages. AHe dates from 16 samples produced dates ranging from 67 ± 8.3 Ma to 6.2 ± 0.76 Ma. Most dates are between 45 and 20 Ma and younger than their crystallization age. Samples from the elevation transect with the largest relief display a positive relationship between AHe date and elevation, and thermal history modeling shows a phase of exhumation from ~30–23 Ma. AHe dates in the Madison Range young as they approach the Madison Fault, the range-bounding normal fault, and we ascribe most of the exhumation in the Madison Range to extension and tectonic exhumation due to footwall uplift. We interpret the ~30–23 Ma cooling to represent fault initiation and a phase of Oligocene extension that shows that post-orogenic extensional faulting and collapse propagated into the Laramide domain at that time. Late Miocene AHe dates near the fault represent a renewed phase of motion in the Miocene to recent, though our data lack the resolution to constrain the specific timing. Erosional exhumation due to YSH-driven regional uplift appears to be minimal.

1. INTRODUCTION

The topography and landscape evolution of southwest Montana reflect the complex interplay of tectonics, lithospheric processes, mantle dynamics, climate, and erosion that have affected the region. This part of the North American Cordillera (Figure 1) experienced substantial mountain building during the Sevier (~155–55 Ma) [1] and Laramide (>80–50 Ma) [2–4] Orogenies. This region contains the transition and potential spatial overlap between zones of thin-skinned Sevier fold and thrust belt-style deformation and thick-skinned Laramide basement-cored uplifts (Figure 1) [5–7]. Mountain building was followed by a period of post-orogenic collapse and

extension starting at ~53 Ma that was initially accompanied by substantial magmatism [8, 9]. Eocene to Miocene sedimentary basins throughout southwest Montana have been critical for understanding post-Orogenic extensional collapse and topography within the Sevier belt [8, 10–12], though whether extension propagated into the Laramide foreland at this time is less well documented. Since the Miocene, Basin and Range-style normal faults have dissected the landscape, and recently the arrival of the Yellowstone hotspot (YSH) deformed the region.

Previous studies investigating the tectonothermal history of this region using low-temperature thermochronology largely focused on determining the timing and nature of Sevier and Laramide tectonism or extension in

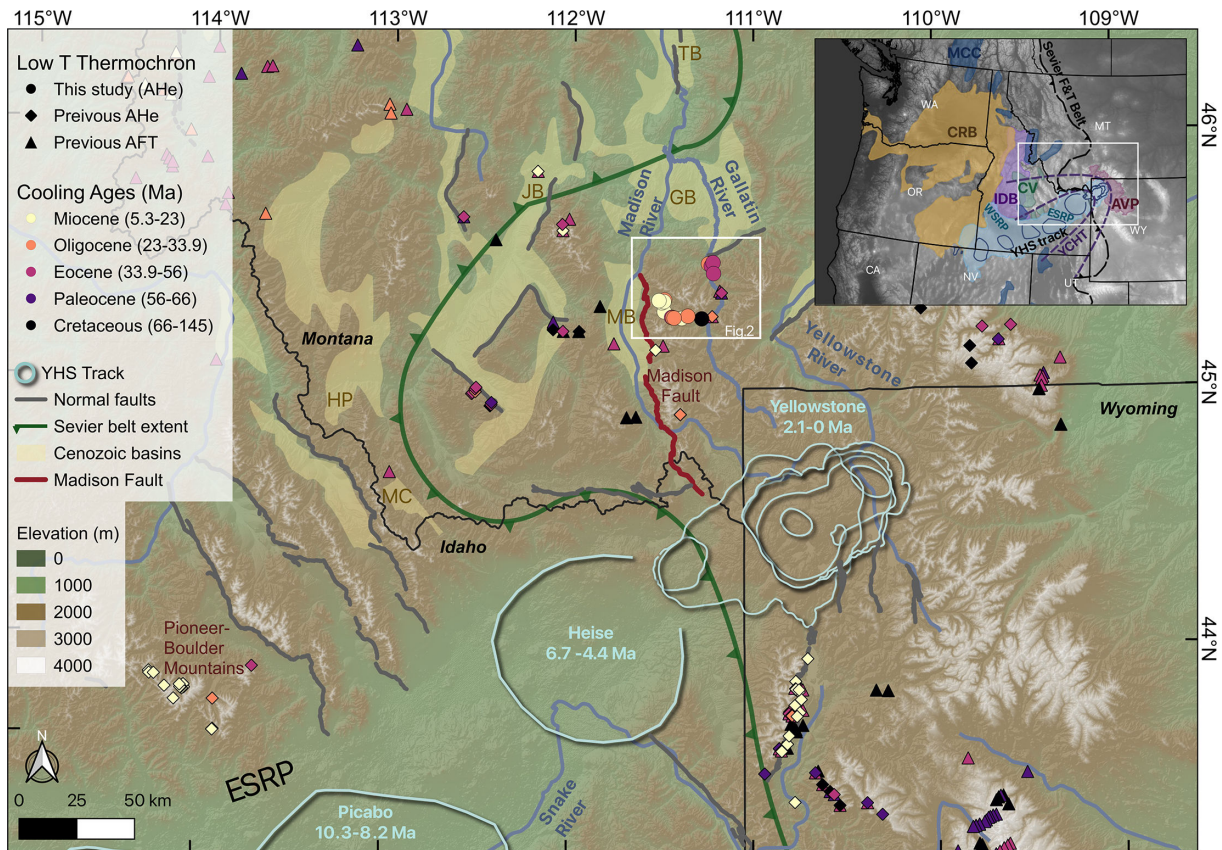


FIGURE 1: Regional terrain map of southwest Montana, eastern Idaho, and northwest Wyoming. The inset map shows the northwest region of the United States with highlighted provinces (AVP—Absaroka Volcanic Province, CRB—Columbia River Basalts, CV—Challis Volcanic Province, ESRP—Eastern Snake River Plain, MCC—Metamorphic Core Complex, IDB—Idaho Batholith, YCHT—Yellowstone Crescent of High Terrain, WSRP—Western Snake River Plain, YHS—Yellowstone hotspot), a generalized boundary of the Sevier fold and thrust belt, and a white box indicating the area of Figure 1. The main map shows the study area (white box, location of Figure 2) which includes the Northern Madison and Gallatin Ranges. Also shown are regional cooling dates from previous low T thermochronology studies (color symbols, [2, 4, 13–16, 19, 20, 63, 110, 111, 130–133]). Also shown are other major features mentioned in the text, including the Eastern Snake River Plain (ESRP), the track of the Yellowstone hotspot (YHS Track), the extent of the Sevier Fold and Thrust Belt, the Madison Fault, and other normal faults in the region [66], the and Cenozoic basins of interest (JB—Jefferson Basin, TB—Townsend Basin, GB—Gallatin Basin, MB—Madison Basin, MC—Muddy Creek Basin, HP—Horse Prairie Basin).

metamorphic core complexes [2, 4, 13–20]. Thermal histories have shown that Laramide-style deformation began in Montana as early as 100 Ma, substantially earlier than the inception of Laramide deformation to the south [2, 15, 18–20]. Studies also support Laramide deformation coeval with the end of the Sevier Orogeny and important structural linkages between Sevier and Laramide deformation [16, 18]. The thermal histories of rocks within metamorphic core complexes in the Sevier belt have been critical for constraining the timing of the transition from compression to post-orogenic extension in the region [14, 17, 21]. However, outside the deeply exhumed core complexes, work on post-orogenic exhumation has been more limited because disentangling the most recent events from the long, complex thermal histories of these rocks has been challenging.

Here, we focus on discerning the post-Laramide erosion and landscape change by applying the apatite (U-Th-Sm)/He (AHe) thermochronometer to Late Cretaceous and Paleogene intrusive rocks in the Gallatin and Madison

Ranges (Figure 1). This area sits just east of the Sevier fold and thrust belt in the Laramide foreland within the Montana Basin and Range and potentially experienced both post-orogenic extension and more recent extension (Figure 1). It also sits on the northern rim of the active manifestation of the YSH and is an ideal location to investigate how mantle processes are impacting the landscape. Small volume Cretaceous and Paleogene-aged intrusions are found throughout the Madison and Gallatin Ranges (Figure 2) [22–25]. The relatively young age of these intrusions (69 ± 1 Ma [23]; 49.0 ± 0.2 Ma [26]) makes them useful recorders of post-Laramide exhumation. By applying the AHe technique with its low-temperature sensitivity to these intrusive lithologies, which record shorter thermal histories than surrounding rocks, we isolate lower magnitude exhumation events after the Laramide. We present new AHe data that documents phases of post-Laramide cooling, exhumation, and extension in the Madison and Gallatin Ranges and interpret these within the broader context of tectonics and landscape evolution in southwest Montana.

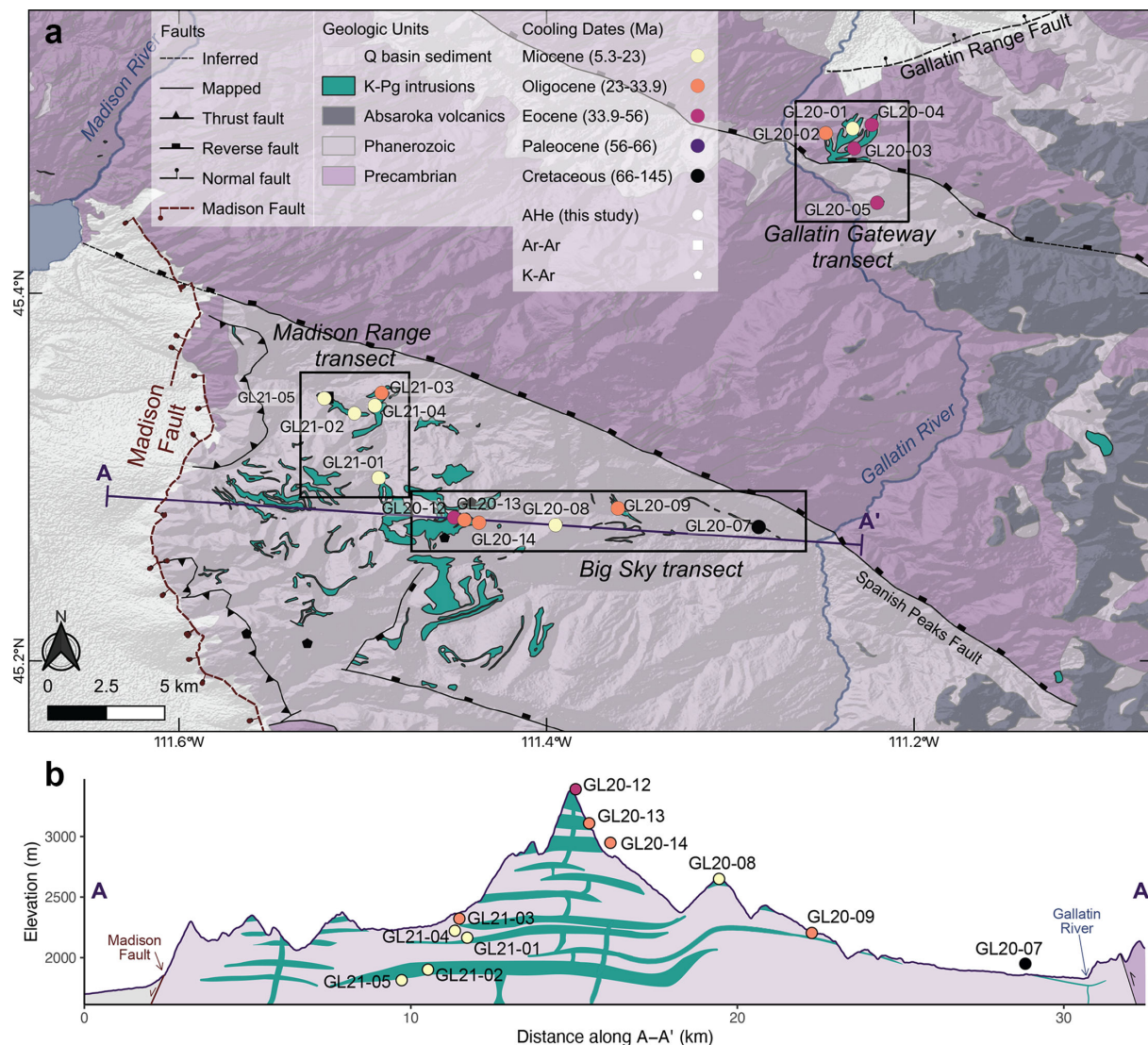


FIGURE 2: Simplified geologic map and topographic profile of our study area. (a) Simplified geologic map after Vuke [82] highlights targeted intrusions. Sample localities are shown as circles colored by their mean AHe date. The Gallatin Gateway transect in the Gallatin Range and the Big Sky and Madison Range transects in the Madison Range are outlined in black boxes. Other features of interest include the Madison and Gallatin Range normal faults, major Laramide structures, and eastern extent of Absaroka Volcanics east of the Gallatin River Valley. The area shown is a white box on Figure 1. Line A-A' shows the location of the profile shown in (b). (b) Topographic profile showing sample locations for Madison and Big Sky Transects with a simplified schematic cross section after Vuke [82].

2. BACKGROUND

2.1. Geologic Setting

Our study region in southwest Montana (Figure 1) sits within the North American Cordillera which extends through >6000 km of North America from Central Mexico to the Alaskan and Canadian Arctic. Eastward subduction of the oceanic Farallon and Kula plates beneath the North American plate resulted in inboard deformation that initiated during the Jurassic. Subduction has continued to the present off the coast of the Pacific Northwest and along the western coast of North America [1, 5]. Within the US portion of the Cordillera, initiation of crustal shortening began in the Late Jurassic (~155 Ma) with thin-skinned Sevier deformation that continued until ~55 Ma [1]. The

mid-Cretaceous gave rise to the extensive Idaho Batholith, which produced magmatism between ~98 and 54 Ma into ancestral North American crust [27]. Magmatism also extended into western Montana, forming a number of plutons including the Boulder and Tobacco Root Batholiths that were dominantly emplaced from 77 to 75 Ma [28]. Thick-skinned Laramide deformation in Montana initiated by ~81 Ma and potentially as early as 100 Ma [2, 29–31], though basement involved uplifts did not begin exhuming in the southern Laramide province until ~80 Ma [4]. In Montana, the onset of this thick-skinned style of deformation resulted from the influence of preexisting stratigraphy and a decrease in subduction angle from ~80–50 Ma that was coeval with the tail end of the Sevier Orogeny [7, 18, 32, 33]. Despite extensive deformation during this period, the

Laramide produced little magmatism and was considered largely amagmatic [2, 34].

Steepening and rollback of the subducting Farallon slab at ~53–51 Ma, accompanied by a decreased rate of subduction, resulted in a period of orogenic collapse, extension, and magmatism [8, 35–37]. The eruption of the voluminous mantle-derived “ignimbrite flare-up” throughout western North America suggests delamination of the Farallon slab occurred from this time to ~20 Ma [36, 38]. Within Idaho and Montana, the largest eruptions resulted in the Absaroka Volcanic Province (AVP) and Challis Volcanic Province (CVP) (Figure 1). The AVP extends from southeast Wyoming to southwest Montana with the bulk of magmatism occurring between ~55 and 44 Ma, and the CVP is in east-central Idaho and was active from ~50 to 47 Ma [34, 39, 40]. Eocene extension also resulted in the formation of extensional basins and facilitated the exhumation of metamorphic core complexes within the Sevier belt [41]. In Montana, this includes the Anaconda and Bitterroot core complexes that were active ~53–40 Ma [14, 21, 42, 43].

In SW Montana, Eocene extension and a complex surface response to magmatism are documented by basin sedimentation throughout the region (Figure 1), particularly within the post-compressional Renova Formation [10, 11]. Though they have an uncertainty of ~1 km, $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ stable isotope data from southwest Montana suggest elevations >2 km in the Sevier fold and thrust belt at the start of the Paleocene, uplifting to peak elevations of ~4 km around 50 Ma due to thermal uplift of the lithosphere during magmatism [12, 44–48]. Warm, wet conditions in the late Paleocene and Early Eocene lead to increasing fluvial relief, while extension from ~50 to 25 Ma led to lowering of valley floors relative to uplifted regions as the climate cooled and became more arid [12].

Basin and Range-style extension initiated at ~17 Ma likely due to changing plate boundary conditions on the west coast and continues to today [8, 49]. The timing of motion on structures within the Montana Basin and Range is not well constrained but may be recorded by the deposition of the Sixmile Creek Formation which overlies the Renova Formation. Deposition began at ~17 Ma after the formation of a regional angular unconformity from ~20 to 17 Ma [8, 50].

Magmatism was prominent from the Miocene to present with the near-synchronous eruption of the Columbia River Basalts and the inception of the YSH. The voluminous Columbia River Basalts erupted in several pulses throughout Oregon, Idaho, and Washington from ~17 to 6 Ma with the bulk of magmatism erupting over a relatively short span from ~16.6 to 15.9 Ma [51, 52]. The onset of YSH magmatism overlapped temporally with the earliest of the Columbia River Basalt phases at ~17 Ma [51, 53, 54]. The most accepted model for the YSH is that it is a deep-seated mantle plume interacting with a westward migrating overriding North American plate [55–58]. The approach and passage of the YSH from the early Miocene to present facilitated uplift and subsidence along its path. The Eastern Snake River Plain (ESRP) is an area of low topography and

subsidence following in the hotspot's wake (Figure 1). The impingement of the hotspot on the overriding lithosphere creates a topographic swell termed the Yellowstone Crescent of High Terrain (YCHT) and generates extension and active seismicity on each flank [57, 59]. Detrital zircon U-Pb provenance studies suggest that the YSH produced a dynamic continental drainage divide, forcing drainage to flow radially away from the location of the hotspot causing major reorganization of the paleo-Snake River system [60–62]. Increased exhumation along the northern flank of the YCHT contemporaneous with the passage of the YSH is shown in AHe date within the Pioneer Mountains of central Idaho, suggesting substantial landscape response to hotspot-related processes [63].

2.2. Geology of the Madison and Gallatin Ranges

This study focuses on the northern Madison and Gallatin Ranges of southwest Montana on the northern flank of the YCHT (Figure 1). Both ranges consist of metamorphosed Archean bedrock overlain by Paleozoic and Mesozoic sedimentary sequences (Figure 2). Absaroka volcanic rocks overlie much of the Gallatin Range [23, 39, 64]. The Madison Range contains two major Laramide-age fault systems. The northwest-striking Spanish Peaks Reverse Fault that is part of the larger Bismark-Spanish Peaks-Gardiner Pre-Cambrian fault system stretching from the westernmost part of the Tobacco Root Mountains to the Montana-Wyoming border north of Yellowstone [22, 23, 64]. To the southwest of the Spanish Peaks Fault lies the north-striking Hilgard Thrust system [64]. Major Laramide faulting had ceased by the early Eocene as evidenced by undeformed cross-cutting Absaroka volcanic rocks [22, 64].

The region experienced substantial volcanism during the Cretaceous to Eocene. Absaroka volcanic rocks cover the eastern portion of our study area. The Gallatin Range contains some of the northern and westernmost outcrops of the AVP with eruption in this region occurring from ~55 to 52 Ma [39, 40]. Outcrops of AVP volcanic rocks are largely constrained to the west of the Gallatin River (Figure 2). The Madison Range and northern Gallatin Range are intruded by an extensive dacite porphyry sill complex, particularly at the intrusive complex center within Lone Mountain [23, 64, 65]. The age of the Lone Mountain and Fan Mountain intrusions was considered Late Cretaceous based on hornblende $^{40}\text{Ar}/^{39}\text{Ar}$ ages of 69 ± 1 Ma [23] and 69 ± 2.5 Ma [65]. However, a recent, concordant U-Pb zircon age of 49.0 ± 0.2 Ma [26] indicates that at least part of the intrusive complex is Eocene. Similarities in chemical composition suggest that these intrusions may be a part of the AVP [26].

The Madison and Gallatin valleys both contain deposits of the Eocene-Oligocene Renova Formation and the Miocene Sixmile Creek Formation. The Gallatin Range Fault and the Madison Fault (Figure 2) are both Miocene and younger normal faults [66, 67]. The arrival of the YSH to its current location at the Yellowstone plateau occurred at ~2.05 Ma, contemporaneous with the eruption of the Huckleberry Ridge Tuff [55, 68]. Arrival of the hotspot generated seismicity and deformation within the YCHT and

created new northwest-trending grabens that intersected with preexisting Miocene grabens, changing drainage routes and affecting sedimentation [50, 55, 62].

2.3. Apatite (U-Th-Sm)/He Thermochronometry

The apatite (U-Th-Sm)/He (AHe) technique is a low temperature thermochronometer that takes advantage of the thermally activated diffusion of ^4He within apatite mineral grains, which is produced through the radioactive decay of U, Th, and Sm. Higher temperatures facilitate open system behavior with rapid diffusion of ^4He through the grain lattice to the grain boundary, while lower temperatures result in closed system behavior, causing retention of ^4He . The temperature range between the lower and upper limit of fully open and closed system behavior is the partial retention zone (PRZ) in which there is both retention and diffusional loss of ^4He . For the AHe system, the temperature sensitivity is $\sim 40\text{--}90^\circ\text{C}$ making it a useful tool for inquiry involving processes within the upper $\sim 1\text{--}4$ km of the crust, depending on the geothermal gradient [69–71]. The diffusion kinetics of apatite are affected by radiation damage to the crystal, with higher levels of radiation damage increasing the temperature sensitivity. Thus, correlations between dates and effective uranium (eU, $[\text{U}] + 0.234[\text{Th}] + 0.0046[\text{Sm}]$, a proxy for radiation damage) are sometimes expected depending on a sample's thermal history [71–73]. Additional factors that can cause variability of grain dates include grain shape and size [74, 75], the presence of inclusions that act as either He traps and/or contribute excess He from foreign minerals [76], He implantation from surrounding U-Th rich sources [77], grains that are broken or fractured [78], and zonation of parent nuclides within the grain [79]. Helium is emitted energetically, so an α -ejection correction (F_T) is applied based on grain size and shape to account for He ejected from the crystal [80, 81].

3. METHODS

3.1. Sampling Strategy

We sampled Paleogene-Cretaceous igneous intrusions for AHe analysis with a range of dacitic, felsic, and gabbroic compositions. Many of these are a part of the larger Fan and Lone Mountain intrusive center made up of dacite porphyry sills referred to as a Christmas-tree laccolith complex [65, 82]. We targeted these lithologies because they are likely to contain accessory apatite and zircon and their intrusion post-dates Laramide deformation [65]. Additionally, these younger intrusions may exhibit less radiation damage than would have affected surrounding older rocks, making them sensitive to lower temperatures.

Whole rock samples were collected as three separate transects within the northern Madison and Gallatin Ranges. Two transects were collected over as large an elevation range as possible within the Gallatin River Valley (the Big Sky and Gallatin Gateway transects, Figure 2), vegetation and topography permitting. The Big Sky transect is closer to the YSH, while the Gallatin Gateway is more distal.

A third transect was collected just west of the Big Sky transect approaching the Madison Fault (the Madison Range transect, Figure 2). Samples along the Madison Range were collected as a horizontal transect to investigate any differences in exhumation with proximity to the Madison Fault [66, 67]. This region is far from the region of Quaternary glaciation in the Yellowstone region [83] which should limit any influence from glacial erosion and allow us to focus on tectonic processes. Each transect contains 5–6 samples for a total of 16 samples. We analyzed 5–8 individual apatite grains for each sample. Transect relief in descending order is (1) the Big Sky transect has the largest relief with 1440 m between the highest and lowest elevation sample over a horizontal distance of 13 km, (2) the Madison Range transect with a relief of 481 m over a horizontal distance of 5 km, and (3) the Gallatin Gateway transect with a relief of 355 m over a horizontal distance of 4 km. Samples collected along an elevation transect are ~ 200 vertical meters apart to produce a pseudo-vertical profile of the crust using the natural topographic relief of the region. This sampling strategy is useful for observing differences in exhumation rate between lower and higher elevations within the same transect unit and establishing age-elevation relationships [84].

3.2. Zircon U-Pb Geochronology

Standard density and magnetic mineral separation techniques were used to isolate apatite and zircon crystals in each sample. Zircon separates from three samples were sent to GeoSep Services, Moscow, ID, for laser ablation inductively coupled plasma-mass spectrometer (LA-ICP-MS) U-Pb analyzes. A full description of the methods can be found in Hults et al. [85] and Moore et al. [86]. Zircons from each sample were mounted in epoxy, polished to reveal grain interior, and 30–40 grains were selected for analysis under high magnification optical microscopy. LA-ICP-MS analysis was performed using a New Wave YP213 213 nm solid state laser ablation system and an Agilent 7700 $\times$ quadrupole mass spectrometer by GeoSep Services at the Washington State University Radiogenic Isotope and Geochronology Laboratory, Pullman, Washington, USA. Data reduction was performed following the methods of Chang et al. [87]. The two zircon age standards used for calibration purposes were the 1099 ± 0.6 Ma FC zircon [88] as the primary standard and the 61.2 ± 0.1 Ma Tardey Rhyolite zircon [89] as the secondary standard. During this session, the secondary standard yielded a $^{206}\text{Pb}/^{238}\text{U}$ date of 62.8 ± 0.8 (n = 44, MSWD = 1.4). Common Pb was accounted for using the Stacey and Kramers model [90].

3.3. Apatite (U-Th-Sm)/He

Individual apatite crystals were selected for AHe analyzes by hand using a Leica M165C Stereoscope based on grain size, shape, and lack of inclusions. Acceptable grain sizes were those with a b-axis measurement larger than the standard baseline grain size of $\sim 60 \mu\text{m}$ and an α -ejection correction (F_T) value > 0.58 [81, 91]. Measurements and images of each

selected apatite grain were recorded prior to packing into Nb tube packets for analysis. Grains were analyzed for He, U, Th, and Sm concentrations at the University of Colorado (CU) Thermochronology Research and Instrumentation Laboratory (CU TRaIL), Boulder, Colorado, USA. Grain packets were loaded into an ASI Alphachron for He extraction and measurement. Degassed grains were dissolved and spiked and then analyzed for U, Th, and Sm content using an Agilent 7900 Quadrupole inductively coupled plasma-mass spectrometer. Concentrations of He, U, Th, and Sm were used to calculate AHe dates and all associated data using the methods described in Ketchum et al. [81] and Cooperdock et al. [75].

3.4. Thermal History Modeling

To investigate the range of possible thermal histories compatible with our AHe data, we used the inverse modeling capabilities of the thermal modeling software HeFTy (v.2.1.5.94) [92] to evaluate time-temperature (t-T) paths and explore cooling histories for each sample in each transect. HeFTy uses a Monte Carlo approach to produce t-T paths that satisfy any geologic constraints set by the user and compares the predicted thermochronometric date with the observed data. Paths that produce good or acceptable fits (defined statistically, [92]) are saved for interpretation. Thus, the resulting paths represent the range of thermal histories that fit the data well [92, 93]. The general constraints imposed on the models were that the temperature was $>120^{\circ}\text{C}$ at the time of emplacement of the intrusions. Samples were hot during emplacement, 120°C is the upper limit of He retention and radiation damage accumulation, and thus our samples are not sensitive to temperatures above that limit [71]. The end constraint for the models was that samples arrive at a present-day surface temperature of 5.9°C , consistent with current surface temperatures in the region [94]. For samples GL20-05, GL20-07, and GL20-09 with emplacement ages that predate the Absaroka volcanic event, we ran an additional set of models to test for the effects of possible reheating at this time. These models have an additional constraint that allows for, but does not require, reheating at 52 ± 2 Ma contemporaneous with Absaroka volcanism.

Grains were combined into bins as HeFTy often struggles to find good fits when many high precision data points are included because of its statistical framework and because the analytical uncertainty does not adequately account for uncertainties in the kinetic models [93]. Grains from each sample were binned based on their eU values and the mean of each bin was modeled to capture any date-eU relationship. The parameters of these bins are consistent across all samples and defined as $20 \text{ ppm} < \text{eU} < 20 \text{ ppm}$. Two samples had high eU values for young dates and were modeled as a sample mean without eU bins due to difficulties finding good and acceptable paths with 2 bins (Table 1). Bin means are entered into HeFTy as “synthetic grains” for each sample with mean values for each grain bin for estimated spherical radius (ESR), uncorrected date, and concentrations of U, Th, and Sm. Uncertainty for each grain is the standard

deviation of each bin or 10% of the bin mean, whichever is larger. The details of the thermal history modeling in HeFTy can be found in Table 1.

To take advantage of the elevation relationships within the Big Sky transect to refine the likely thermal histories, we also modeled several samples together as an elevation transect using the software QTQt [95]. QTQt uses a Bayesian transdimensional Markov Chain Monte Carlo inverse method and results in an ensemble of models that represent the most likely Tt paths given the data and the priors [95]. We modeled four samples, GL20-09, GL20-12, GL20-13, and GL20-14, as having a shared thermal history with a temperature offset applied to the samples to reflect their present-day elevation difference. For QTQt thermal history simulations, we are able to predict individual AHe dates from each sample, as opposed to binning the data as for HeFTy above. We applied both a general prior to search a wide range of time-temperature combinations and a specific constraint using the U-Pb crystallization age of sample GL20-12 from the top of the transect. The past geothermal gradient (and therefore past temperature offset between samples) as well as the rate of temperature change were also allowed to vary quite broadly. We ran initial models both allowing and not allowing reheating during the thermal history. However, we found that models allowing reheating predicted an increase in the temperature of GL20-12 from approximately 20 Ma to the present day, which seems implausible considering that this sample is from the top of the elevation transect. Therefore, we focus on model results where reheating is not permitted; the model parameters for our preferred simulations are reported in Table 2, and input files and output run file can be found in the online supplementary material 1 to 5.

4. RESULTS

4.1. Zircon U-Pb Results

We acquired U-Pb dates for zircons from three samples to constrain the crystallization ages; 30–40 grains from each sample were analyzed and all yielded dominant populations with dates <100 Ma and a smaller population (2–14 grains) with dates >1500 Ma. The IsoplotR software [96] was used to calculate weighted mean and concordia ages from the younger populations using the $^{238}\text{U}/^{206}\text{Pb}$ and $^{207}\text{Pb}/^{206}\text{Pb}$ ratios following the method of Ludwig (1998). Weighted mean and concordia plots are shown in Figure 3. We also calculated the largest young statistical population (LYSP), the youngest population of three or more grains with a mean squared weighted deviation of ~ 1 [97, 98]. Ages from the methods are very similar and overlap within uncertainty, and we use the concordia ages in the text for consistency. Results are summarized in Table 3, and data are reported in full in the Supplemental Materials online supplementary Table S1.

Sample GL20-12 is from near the top of Lone Mountain and represents the dominant lithology of the Lone Mountain intrusive complex, mapped as a dacite porphyry.

TABLE 1: HeFTy thermal history model inputs for southwest Montana samples^{bc}.

1. Thermochronologic data			
Samples and data used in simulations			
Sample	Mapped rock type	eU bins for AHe data ^a	Emplacement age used in model
Gallatin Gateway			
GL20-01	Felsic intrusion	<20 ppm (3 grains), >20 ppm (2 grains)	51 ± 2 Ma
GL20-02	Felsic intrusion	<20 ppm (4 grains), >20 ppm (1 grain)	51 ± 2 Ma
GL20-03	Felsic intrusion	<20 ppm (4 grains), >20 ppm (1 grain)	51 ± 2 Ma
GL20-04	Felsic intrusion	<20 ppm (4 grains), >20 ppm (2 grains)	51 ± 2 Ma (U-Pb age, this study)
GL20-05	Absaroka intrusion	<20 ppm (4 grains)	74 ± 2 Ma (U-Pb age, this study)
Big Sky			
GL20-07	Gabbro dike	<20 ppm (3 grains)	69 ± 1 Ma [23]
GL20-09	Gabbro	<20 ppm (1 grain), >20 ppm (3 grains)	69 ± 1 Ma
GL20-12	Dacite porphyry	4 grains averaged, no bin	49 ± 2 Ma (U-Pb age, this study)
GL20-13	Dacite porphyry	<20 ppm (4 grains)	49 ± 2 Ma
GL20-14	Dacite porphyry	<20 ppm (4 grains)	49 ± 2 Ma
Madison Range			
GL21-01	Dacite porphyry	<20 ppm (4 grains)	49 ± 2 Ma
GL21-02	Dacite porphyry	5 grains averaged, no bin	49 ± 2 Ma
GL21-03	Dacite porphyry	<20 ppm (4 grains)	49 ± 2 Ma
GL21-05	Dacite porphyry	<20 ppm (4 grains)	49 ± 2 Ma
2. Additional geologic information			
Constraint		Explanation and data source	
All rocks			
1	>120°C at time of emplacement	Intrusions hot at crystallization	
2	5.9°C at present day	Surface temperature	

AHe data treatment: Grains from each sample were binned into groups with similar eU values and the mean uncorrected He date of each bin was modeled. α -ejection corrected in HeFTy using [81]. Error (Ma) applied in modeling: The 1 σ sample standard deviation of each bin was applied if $\geq 10\%$. If $< 10\%$, then 10% was applied. r (μm): Mean equivalent spherical radius of each bin. eU (ppm): Mean U and Th for each bin. eU zonation: none assumed.

Emplacement ages from this study used in modeling are rounded concordia ages with $\sim 2\sigma$ analytical uncertainty (Table 3). The U-Pb date for GL20-04 is used for all samples from the same intrusion in the Gallatin Gateway transect. The U-Pb date for GL20-12 is used for all other samples from the Fan and Lone Mountain intrusions.

System and model specific parameters: He kinetic model: RDAAM (radiation damage accumulation and annealing model; [71]). Statistical fitting criteria: GOF values > 0.5 for "good" fits, > 0.05 for "acceptable" fits. The good-fits also must have a minimum GOF of $1/(N+1)$ where N is number of statistics used [92].; Modeling Code: AHe uses HeFTy v1.9.3.; Number of t-T paths attempted: 500 good-fit paths for all models.

GL20-12 yielded an age of 49.3 ± 1.01 Ma. Sample GL20-04 is from a felsic intrusion that composed most of the Gallatin Gateway transect samples. It resulted in an age of 51.2 ± 0.83 Ma. The last sample analyzed, GL20-05, is from an adjacent but separate intrusion from the rest of the Gallatin Gateway transect that was mapped as a Tertiary intrusion related to the Absarokas (Tia; [82]). This sample resulted in an age of 74.5 ± 0.97 Ma.

4.2. Apatite (U-Th-Sm)/He Results

We report AHe results from 82 individual apatite grains from 16 samples (5–8 grains per sample, Table 4). Individual AHe dates and 2σ analytical uncertainties are reported for each grain alongside mean dates for each sample with associated 1 σ standard deviation. Only grains with alpha-ejection correction (F_T) values > 0.58 are reported. A total

TABLE 2: QTQt thermal history model inputs for southwest Montana samples.

1. Thermochronologic data				
Samples and data used in simulations				
Sample	Mapped rock type	Number of (U-Th)/He dates included		Emplacement age used in model
Big Sky				
GL20-09	Gabbro	5		49.5 ± 3.4 Ma
GL20-12	Dacite porphyry	5		49.5 ± 3.4 Ma
GL20-13	Dacite porphyry	5		49.5 ± 3.4 Ma
GL20-14	Dacite porphyry	5		49.5 ± 3.4 Ma
2. Additional model constraints				
Constraint	Time (Ma)	± (Ma)	Temperature (°C)	± (°C)
(1) General Prior	26.4	26.4	100	100
(2) Emplacement	49.5	3.4	700	200
(3) Geothermal gradient (per km)			30	30
(4) Present-day temperature offset			6	6
(5) Maximum temperature change (per Ma)			1000	

AHe data treatment: Individual apatite (U-Th)/He ages from each sample were modeled, using grain specific He, U, Th, and Sm concentrations, as well as grain-specific geometry. Uncertainty applied for each grain was 8%.

Emplacement age and uncertainty used in modeling is mean and standard deviation of all Cenozoic U-Pb dates from GL20-12 (online supplementary Table S1).

He kinetic model: RDAAM (radiation damage accumulation and annealing model; [71]).

System and model specific parameters: QTQt version: 5.6.0; Number of burn-in iterations: 15000; Number of post-burn-in iterations: 20000.

of 5 grains were rejected as outliers as determined by the Grubb's statistical test for outliers [99] and are reported in *italics* in Table 4 but omitted from date means and final interpretations. Whole samples that result in a coefficient of variation >0.2 (sample standard deviation/sample mean, >20% dispersion) are considered dispersed, and the sample mean is not reported or interpreted. Fourteen samples resulted in 18% or less dispersion with reported mean dates between 67 ± 8.3 Ma to 6.2 ± 0.76 Ma (Table 4). Two samples resulted in dispersion >20% (GL20-08, 57% dispersion; GL21-04, 29% dispersion). Dates within these samples are not correlated with eU or grain size online supplementary Figure S1. As dispersion cannot be explained by a predictable mechanism, we do not feel that the mean dates for these samples are geologically meaningful, and they have been omitted from inverse modeling and interpretations. Most samples yield mean dates that are younger than the crystallization ages for the intrusion, though three samples have dates that overlap with their crystallization age.

The Gallatin Gateway transect had a wide range of AHe cooling dates from 43.0 ± 5.0 Ma to 14.5 ± 2.4 Ma (Figures 4 and 5 and Table 4). This transect is the farthest of the three transects from the YSH and has older mean date values for similar elevations than the other transects. This transect only spanned 355 m of elevation difference, but samples

from the highest and lowest elevation values have similar mean date values, and there is no observable relationship between date and elevation for this transect (Figure 4). Plots for eU versus mean AHe dates for samples in this transect show a range of dates across similar eU values and do not show a trend (Figure 5).

AHe cooling dates for the Big Sky transect range from 66.7 ± 8.3 Ma to 23.9 ± 3.8 Ma (Figures 4 and 6 and Table 4). This transect shows a positive trend among date-elevation values over 1440 m of relief (Figure 3) with older dates at higher elevations and younger dates at lower elevations. The oldest AHe date within this transect (66.7 ± 8.3 Ma, GL20-07) does not follow this trend as it has the lowest elevation and oldest mean date. Within samples, similar dates are observed across a range of eU values, showing there is no observable trend between eU and AHe date for these samples (Figure 6).

Mean sample dates from the Madison Range transect have the youngest dates of all three transects ranging from 26.9 ± 5.6 Ma to 6.2 ± 0.76 Ma over 481 m of relief (Figures 2 and 4(f) and Figure 7). There is no observable trend between sample means and elevation (Figure 4) or eU (Figure 7), but mean AHe dates generally increase with distance from the Madison Fault in this transect in the footwall (Figure 8). One sample (6.2 ± 0.76 Ma, GL21-01) diverges from this trend and has an unexpectedly young

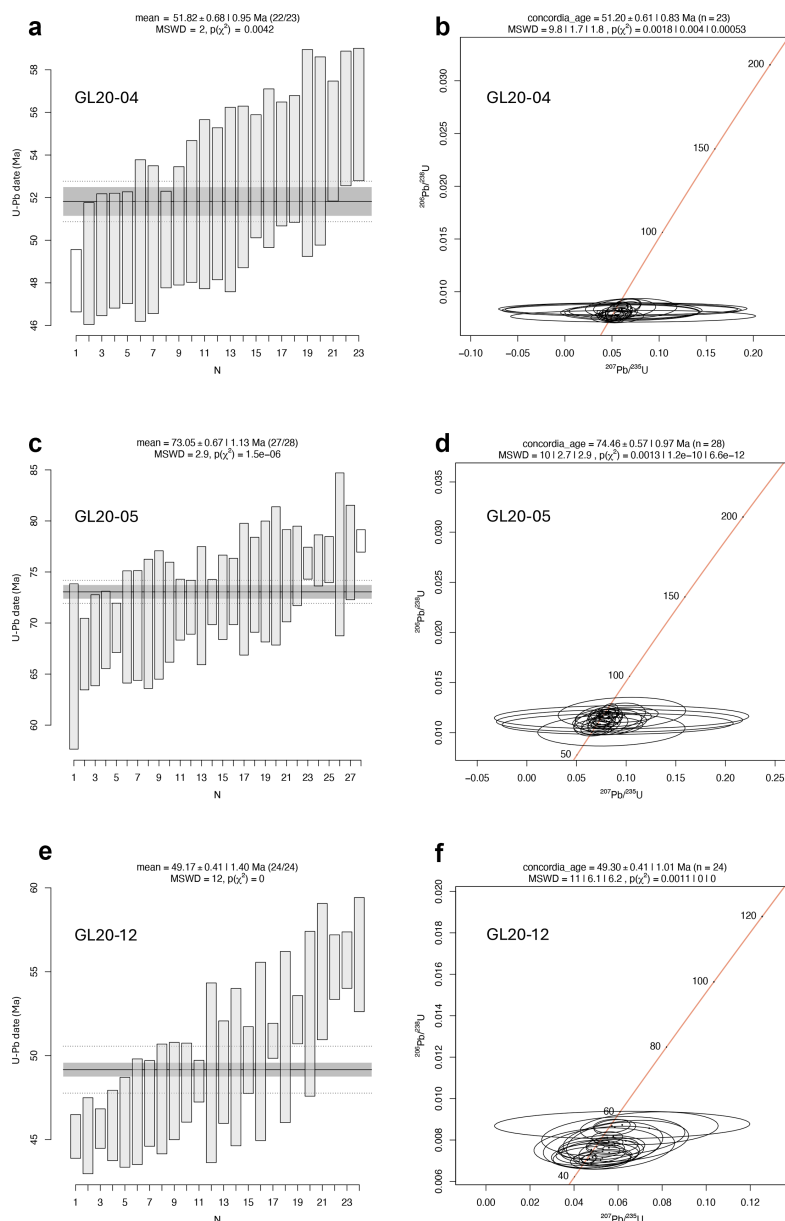


FIGURE 3: Zircon U-Pb dates. Plots of the weighted mean (a, c, e) of individual $^{206}\text{Pb}/^{238}\text{U}$ dates and concordia diagrams (b, d, f) for samples GL20-04 (a, b), GL20-05 (c, d), and GL20-12 (e, f). Vertical bars on weighted mean plots represent individual analyzes and analytical uncertainty, with bars with white fill indicating outliers that were not included in the weighted mean calculation. Horizontal line is weighted mean with gray shading indicating the 95% CI and dashed lines the dispersion. Concordia plots show individual analyzes with 95% confidence error ellipses. Plotted using IsoplotR [96].

date. The increase in AHe date with distance from the Madison Fault continues into the neighboring Big Sky transect, also located in the footwall (Figure 8).

4.3. Inverse Modeling Results

HeFTy inverse modeling plots for each individual sample are shown in Figures 5–7. For discussing thermal history modeling results, we will focus on the times that good-fit paths from samples are passing through the apatite PRZ, ~ 90 – 40°C [69, 71, 100]. This temperature window is where the AHe system is sensitive and thus where the models are constrained by the data. HeFTy modeling results show

variable cooling trends across all samples. Most samples do not cool to surface temperatures ($<40^\circ\text{C}$) until 10–40 Myr after emplacement, except for GL20-04, GL20-07, and GL20-12, which have AHe dates similar to their emplacement ages. These samples show simple t-T paths which cool rapidly to near-surface conditions almost immediately after emplacement (Figures 5(d), 6(a), and 6(c)).

For the majority of samples that do not cool immediately following emplacement, samples cannot cool below $\sim 40^\circ\text{C}$ until after emplacement at a time that roughly corresponds with their mean AHe date and are required to cool below $\sim 50^\circ\text{C}$ 5–10 Myr after that. They show a

TABLE 3: Summary of zircon U-Pb data.

Samples	Latitude	Longitude	Mapped lithology	Concordia			LYSP			Weighted mean		
				n	Age (Ma)	2σ (Ma) ^a	n	Age (Ma)	2σ (Ma) ^b	n	Age (Ma)	2σ (Ma) ^a
GL20-04	45.49176797	-111.2221895	TKfi: Fine-grained intrusion with phenocrysts	23	51.2	0.83	18	51.47	1.34	22	51.8	0.95
GL20-05	45.44927541	-111.2193144	Tia: Medium-grained intrusion (plg-pyx-amp)	28	74.5	0.97	18	71.5	1.71	27	73.1	1.13
GL20-12	45.27798663	-111.4495743	Kdap: Dacite (plg)	24	49.3	1.01	10	48.5	1.43	24	49.2	1.4

^aσ uncertainty for concordia and weighted mean ages includes analytical and decay constant uncertainty.

^bσ uncertainty for LYSP ages includes analytical, systematic, and decay constant uncertainty.

wide range of possible cooling histories with rapid cooling permissible and are not required to spend time in the PRZ. The exception to this is GL20-09 where all good fit paths spend at least 5 Myr within the PRZ. All of these samples allow for, but do not require, 18–46°C of post-Miocene cooling. Results for the additional models that included the possibility of reheating during Absaroka volcanism are very similar to those without the reheating event, and reheating is not required for any of the samples (online supplementary Figure S2).

For the Big Sky transect, we present the QTQt expected model results. QTQt simulations produce a suite of model outputs that summarize the simulation statistics in different ways. The expected model is a weighted mean model of the post-burn-in thermal histories, where the weight of each model is its posterior probability and is commonly used as the summary model result for QTQt simulations [101]. QTQt models suggest that all samples experienced rapid cooling after emplacement at ~50 Ma followed by isothermal holding until ~30 Ma when a phase of accelerated cooling began (Figure 9(a)). This cooling phase lasts until ~23 Ma and samples cool by ~40°C during that time, with most samples cooling to near surface temperatures except the lowest elevation sample GL20-09. Temperatures remain relatively constant until the lower samples see another phase of cooling after 10 Ma, with the lowest sample experiencing ~35°C of cooling (Figure 9(a)). Model predictions match the observations reasonably well, though many of the predicted dates for the highest elevation sample are slightly younger than the observed values (Figure 9(b)).

5. DISCUSSION

5.1. Age of Magmatism in the Northern Madison and Gallatin Ranges

Zircon U-Pb dates from samples analyzed yielded a dominant age population <100 Ma which we interpret as the crystallization age for that sample. A much smaller population (2–14 grains) yielded ages >1500 Ma which we interpret as xenocrysts and omit from our discussion. Sample GL20-12 is from the top of Lone Mountain and

the crystallization age of 49.3 ± 1.01 Ma is in line with a recently published high precision U-Pb zircon date of 49.0 ± 0.2 Ma, also from the Lone Mountain intrusion [26]. Sample GL20-04 is from a different intrusion to the north near Gallatin Gateway (Figure 2) and yields a similar age of 51.2 ± 0.83 Ma. This suggests that the two intrusions are contemporaneous and also overlap with timing of Absaroka volcanism in the region. Dudás et al. [26] used the crystallization age and the adakitic composition of the Lone Mountain intrusion to argue that it was affiliated with the AVP and extended Absaroka volcanism further west. The similar age and lithology for GL20-04 suggest that this intrusion may also be associated with the AVP, though geochemical data would be needed to verify this association.

Sample GL20-05 is from a small intrusion near the intrusion containing GL20-04 but yields an older crystallization age of 74.5 ± 0.97 Ma. This is similar to a hornblende $^{40}\text{Ar}/^{39}\text{Ar}$ age of 69 ± 1 Ma that had been published for the Lone Mountain –Fan Mountain complex [23]. Dudás et al. [26] prefer their younger date for the Lone Mountain intrusion and argue that the $^{40}\text{Ar}/^{39}\text{Ar}$ age may be plagued by excess Ar. They acknowledge that it is possible that there were two separate pulses of magmatism of similar composition in the region separated by 20 Myr but argue that this seems unlikely [26]. However, our age for GL20-05 also suggests that there was a Cretaceous period of magmatism in this region. Additionally, we did not acquire U-Pb dates for GL20-07, but the mean AHe date of 66.7 ± 8.3 Ma shows that the host dike is at least ~67 Ma. GL20-07 is from a gabbro dike that is compositionally distinct from the Lone Mountain intrusion, while GL20-05 is compositionally and texturally more similar to the dacite at Lone Mountain. Crystallization ages suggest there were two pulses of magmatic intrusions in the region, at ~70 Ma and at ~50 Ma, though it appears that the ~50 Ma event may have been more voluminous.

5.2. Eocene Post-Magmatic Cooling

While the majority of the 16 samples, we dated yielded AHe cooling dates that significantly postdate their

TABLE 4: Apatite (U-Th)/He data from southwest Montana.

Sample ^a	Mass(μ g)	R (μ m) ^b	Term. ^c	⁴ He (nmol/g)	U (ppm)	Th (ppm)	Sm (ppm)	eU ^d	Th/U	Raw age (Ma)	F _T ^e	Corrected age (Ma) ^f	2 σ^f (Ma)
Gallatin Gateway transect													
GL20-01: Tertiary-Cretaceous felsic intrusion (TKf): 45.4897°N, -111.2325°W, 2189 m elevation													
a1	3.68	56.8	1T	0.62	7.51	23.45	106.4	13.22	3.12	8.67	0.74	11.74	0.54
a2	2.84	56.5	2T	1.44	10.61	50.87	295.7	23.07	4.79	11.57	0.73	15.74	0.60
a3	6.52	68.6	2T	1.36	10.69	29.89	129.9	17.96	2.80	14.08	0.78	17.96	0.71
a4	5.52	66.6	1T	0.82	7.65	28.88	128.6	14.68	3.77	10.40	0.77	13.41	0.76
a5	1.62	47.3	2T	1.11	10.65	46.79	207.8	22.04	4.39	9.34	0.68	13.62	0.88
Mean: 14.49 $\pm$ 2.40 Ma, 17% dispersion													
GL20-02: Tertiary-Cretaceous felsic intrusion (TKf): 45.4871°N, -111.2472°W, 2006 m elevation													
a1	1.16	36.3	2T	0.66	6.55	5.64	151.8	8.07	0.86	15.24	0.61	24.66	2.85
a2	2.96	55.0	2T	1.01	7.03	11.52	135.3	9.94	1.64	18.88	0.73	25.58	2.50
a3	2.48	52.3	2T	2.22	18.42	6.70	186.6	20.24	0.36	20.30	0.73	27.76	1.68
a4	1.00	37.9	2T	0.61	3.83	5.30	106.8	5.22	1.39	21.75	0.62	34.55	5.56
a5	4.03	59.1	1T	0.70	3.99	7.08	110.1	5.81	1.77	22.29	0.75	29.48	1.81
Mean: 28.41 $\pm$ 3.92 Ma, 14% dispersion													
GL20-03: Tertiary-Cretaceous felsic intrusion (TKf): 45.4788°N, -111.42317°W, 2120 m elevation													
a1	0.76	34.3	2T	0.57	4.94	5.23	113.6	6.33	1.06	16.71	0.59	28.06	5.12
a2	4.06	62.1	2T	0.97	4.10	5.92	105.8	5.63	1.45	31.98	0.76	41.58	2.15
a3	5.82	69.6	2T	7.08	42.84	0.68	39.9	43.05	0.02	30.45	0.80	38.15	1.94
a4	2.02	50.3	2T	1.32	8.77	6.91	166.2	10.62	0.79	23.05	0.72	32.04	3.08
a5	2.97	56.7	2T	1.67	6.50	14.91	127.0	10.20	2.30	30.32	0.74	40.83	1.91
Mean: 36.13 $\pm$ 5.87 Ma, 16% dispersion													
GL20-04: Tertiary-Cretaceous felsic intrusion (TKf): 45.4917°N, -111.2221°W, 2339 m elevation													
a1	4.09	63.9	2T	29.93	133.47	62.95	57.2	148.52	0.47	37.31	0.78	48.01	1.04
a5	1.25	42.7	2T	1.24	4.40	25.91	221.3	10.83	5.89	21.18	0.65	32.39	4.70
a6	1.43	44.4	2T	2.20	6.01	32.56	201.9	14.00	5.42	29.10	0.66	43.74	2.72
a7	1.16	42.5	2T	3.92	22.98	15.45	244.3	26.95	0.67	26.93	0.67	40.21	4.50
a8	1.05	40.9	2T	2.18	8.81	26.78	226.2	15.46	3.04	26.19	0.64	40.59	2.34
a9	0.96	39.1	2T	0.95	3.75	10.64	171.0	6.48	2.84	27.10	0.63	42.68	3.34

(Continued)

TABLE 4: Continued

Sample ^a	Mass(μ g)	R (μ m) ^b	Term. ^c	⁴ He (nmol/g)	U (ppm)	Th (ppm)	Sm (ppm)	eU ^d	Th/U	Raw age (Ma)	F _T ^e	Corrected age (Ma) ^f	2 σ^f (Ma)
Mean: 41.27 $\pm$ 5.18 Ma, 13% dispersion													
GL20-05: Tertiary intrusion Absaroka (Tia): 45.4492°N, -111.2193°W, 1984 m elevation													
a7	1.63	42.3	1T	1.81	7.89	22.49	126.0	13.40	2.85	25.08	0.65	38.24	1.53
a8	1.96	47.9	1T	2.97	11.28	37.91	95.1	20.41	3.36	26.99	0.69	39.03	1.92
a9	0.91	35.6	2T	2.53	8.87	31.81	129.3	16.59	3.59	28.26	0.59	47.75	2.19
a10	2.81	51.7	1T	2.02	5.80	22.32	96.5	11.23	3.85	33.39	0.71	46.82	2.87
Mean: 42.96 $\pm$ 5.02 Ma, 12% dispersion													
Big Sky transect													
GL20-07: Kgb gabbro dike: 45.2728°N, -111.2837°W, 1950 m elevation													
a6	0.75	35.1	2T	1.24	2.69	12.03	105.9	5.68	4.47	40.32	0.58	68.64	6.94
a7	1.25	41.9	2T	1.48	2.68	11.67	97.44	5.58	4.35	49.03	0.64	75.32	6.78
a8	0.78	34.5	1T	0.99	3.20	10.00	90.73	5.69	3.13	32.26	0.58	55.22	6.51
a9	1.02	37.6	1T	1.96	1.93	9.20	77.79	4.21	4.77	86.17	0.61	140.5	11.85
a10	1.37	43.3	1T	0.99	1.88	9.08	71.34	4.12	4.84	44.66	0.65	67.65	10.05
Mean: 66.71 $\pm$ 8.38 Ma, 12% dispersion													
GL20-08: Cretaceous dacite porphyry intrusion (Kdap): 45.2739°N, -111.3943°W, 2650 m elevation													
a1	1.06	38.6	2T	0.31	5.11	11.37	114.8	7.96	2.22	7.16	0.62	11.39	1.70
a2	1.48	38.2	1T	0.17	3.38	11.42	104.4	6.22	3.38	5.00	0.62	8.04	1.69
a3	1.55	35.6	0T	0.25	3.18	10.99	123.3	5.95	3.45	7.66	0.59	12.83	1.47
a4	1.59	45.3	2T	0.65	4.13	14.42	146.7	7.73	3.50	15.54	0.67	22.95	2.65
a5	1.36	38.7	1T	0.43	4.69	13.80	130.5	8.13	2.95	9.72	0.62	15.49	1.91
a6	2.65	51.0	0T	0.78	2.48	10.56	100.1	5.11	4.27	28.37	0.71	39.90	2.50
a7	1.72	44.1	0T	0.22	4.02	14.35	131.8	7.60	3.56	5.34	0.66	7.99	0.67
a8	2.20	45.8	0T	0.55	2.99	9.69	105.0	5.42	3.25	18.85	0.68	27.63	1.76
57% dispersion of AHe dates													
GL20-09: Tertiary-Cretaceous gabbro (TKga): 45.2830°N, -1113603°W, 2204 m elevation													
a1	2.30	41.31	0T	1.00	7.50	30.64	200.7	15.0	4.09	12.36	0.64	19.18	1.42
a2	1.03	36.66	1T	1.90	9.84	51.80	281.6	22.5	5.27	15.70	0.59	26.21	2.38
a3	2.06	38.46	1T	3.88	19.11	96.26	291.7	42.4	5.04	17.02	0.61	27.66	1.80

(Continued)

TABLE 4: Continued

Sample ^a	Mass(μ g)	R (μ m) ^b	Term. ^c	⁴ He (nmol/g)	U (ppm)	Th (ppm)	Sm (ppm)	eU ^d	Th/U	Raw age (Ma)	F _T ^e	Corrected age (Ma) ^f	2 σ^f (Ma)
a4	0.87	31.16	1T	0.73	7.22	38.53	264.0	16.7	5.34	8.14	0.53	15.20	2.16
a5	4.57	52.81	1T	2.12	9.25	64.26	168.6	24.7	6.94	15.94	0.71	22.38	1.13
Mean: 23.86 $\pm$ 3.83 Ma, 16% dispersion													
GL20-12: Cretaceous dacite intrusion (Kda): 45.2779°N, -111.4495°W, 3390 m elevation													
a1	7.58	71.31	1T	5.06	22.24	25.94	95.59	28.53	1.17	32.86	0.80	41.24	1.80
a2	1.33	39.32	2T	6.03	36.98	4.61	116.88	38.22	0.12	29.24	0.65	45.12	4.07
a3	0.96	39.33	2T	5.85	27.17	38.06	210.00	36.48	1.40	29.73	0.64	46.61	2.80
a4	1.96	41.66	2T	1.92	5.91	19.00	128.81	10.59	3.21	33.70	0.65	51.82	2.07
a5	0.73	31.09	2T	1.13	4.75	15.27	115.00	8.52	3.22	24.53	0.54	45.28	5.60
Mean: 46.01 $\pm$ 3.81 Ma, 8% dispersion													
GL20-13: Cretaceous dacite (Kda): 45.2767°N, -111.4439°W, 3109 m elevation													
a2	2.47	54.02	2T	2.01	12.79	20.31	155.9	17.81	1.59	20.91	0.73	28.55	1.788
a3	1.48	42.87	2T	0.73	4.09	18.35	118.9	8.60	4.49	15.66	0.652	23.88	3.693
a4	0.95	35.76	2T	0.89	3.89	17.00	142.6	8.11	4.37	20.29	0.587	34.15	8.135
a5	0.90	35.30	2T	0.68	3.27	12.82	116.9	6.46	3.92	19.43	0.584	32.88	10.39
Mean: 29.87 $\pm$ 4.66 Ma, 15% dispersion													
GL20-14: Cretaceous dacite porphyry (Kdap): 45.2752°N, -111.4359°W, 2948 m elevation													
a1	2.47	51.09	2T	2.25	9.07	27.15	159.0	15.73	2.99	26.51	0.709	37.22	3.08
a2	3.81	58.19	2T	2.17	10.93	18.74	116.5	15.53	1.71	25.89	0.748	34.51	1.99
a3	10.1	78.08	2T	1.72	8.27	23.14	89.98	13.88	2.80	23.05	0.808	28.46	1.31
a4	2.23	50.36	2T	1.57	8.32	26.65	156.2	14.85	3.20	19.58	0.705	27.68	2.26
a5	3.01	55.26	2T	3.90	6.98	23.02	133.6	12.62	3.30	57.16	0.73	77.96	4.81
Mean: 31.97 $\pm$ 4.65 Ma, 14% dispersion													
Madison Range transect													
GL21-01: Cretaceous dacite porphyry (Kdap): 45.2996°N, -111.4906°W, 2165 m elevation													
a1	1.5	41.7	2T	0.29	11.3	21	179.7	16.5	1.865	3.2	0.65	4.9	0.4
a2	1.7	46.9	2T	0.31	6.6	23.9	150.6	12.5	3.597	4.7	0.68	6.8	0.5
a3	4.8	62.6	2T	0.28	5.4	23.7	105	11.2	4.368	4.7	0.76	6.1	0.4
a4	1.3	43.2	2T	0.39	7.9	34.6	196.2	16.4	4.354	4.4	0.65	6.7	0.5
a5	5.1	68.3	2T	0.24	4.9	16.5	111.7	9	3.363	4.9	0.78	6.3	0.3

(Continued)

TABLE 4: Continued

Sample ^a	Mass(μ g)	R (μ m) ^b	Term. ^c	⁴ He (nmol/g)	U (ppm)	Th (ppm)	Sm (ppm)	eU ^d	Th/U	Raw age (Ma)	F _T ^e	Corrected age (Ma) ^f	2 σ^f (Ma)
Mean: 6.16 $\pm$ 0.76 Ma, 11% dispersion													
GL21-02: Cretaceous dacite porphyry (Kdap): 45.3347°N, -111.5038°W, 1901 m elevation													
a1	1.64	44.1	2T	1.64	8.47	35.52	203.0	17.17	4.19	17.76	0.66	26.71	1.92
a2	2.49	54.4	2T	1.71	10.03	34.00	178.1	18.34	3.39	17.28	0.73	23.74	1.61
a3	2.51	53.2	2T	1.66	10.30	33.51	173.2	18.48	3.25	16.68	0.72	23.10	1.12
a4	4.58	66.3	2T	2.35	13.64	55.03	172.6	26.95	4.03	16.18	0.77	20.91	1.11
a5	1.66	46.7	2T	1.81	15.19	61.06	244.0	30.01	4.02	11.21	0.68	16.43	0.95
Mean: 22.18 $\pm$ 3.8 Ma, 17% dispersion													
GL21-03: Cretaceous dacite porphyry (Kdap): 45.3458°N, -111.4890°N, 2322 m elevation													
a1	6.21	73.0	2T	1.11	4.75	19.68	94.33	9.55	4.14	21.47	0.79	27.03	1.09
a2	3.61	54.3	2T	0.85	4.55	16.26	95.23	8.54	3.57	18.51	0.72	25.45	1.99
a3	5.67	71.4	2T	4.00	13.74	16.54	184.4	17.90	1.20	41.31	0.80	51.76	3.33
a4	2.67	51.7	2T	0.88	5.69	21.79	130.1	11.03	3.83	14.78	0.71	20.72	1.35
a5	1.87	44.4	2T	1.67	6.94	27.13	152.3	13.58	3.91	22.87	0.66	34.23	3.25
Mean: 26.86 $\pm$ 5.6 Ma, 20% dispersion													
GL21-04: Cretaceous dacite porphyry (Kdap): 45.3389°N, -111.4926°W, 2222 m elevation													
a1	8.62	81.34	2T	0.61	9.35	10.61	87.15	11.97	1.13	9.40	0.82	11.43	0.51
a2	4.99	60.15	1T	0.76	4.77	22.20	101.2	10.17	4.65	13.81	0.75	18.39	1.39
a3	9.43	79.92	2T	0.74	5.28	23.97	86.47	11.08	4.54	12.43	0.81	15.32	0.62
a4	8.19	76.95	2T	1.27	5.47	24.86	97.34	11.50	4.55	20.50	0.80	25.50	0.77
a5	4.91	64.92	2T	0.67	5.60	24.80	118.5	11.64	4.43	10.66	0.77	13.87	0.73

(Continued)

TABLE 4: Continued

Sample ^a	Mass(μ g)	R (μ m) ^b	Term. ^c	⁴ He (nmol/g)	U (ppm)	Th (ppm)	Sm (ppm)	eU ^d	Th/U	Raw age (Ma)	F _T ^e	Corrected age (Ma) ^f	2 σ^f (Ma)
<i>29% dispersion of AHe dates</i>													
GL21-05: Cretaceous dacite porphyry (Kdap): 45.3429°N, -111.5202°W, 1841 m elevation													
<i>a1</i>	1.50	42.34	2T	2.76	35.68	13.23	229.4	39.10	0.37	13.11	0.67	19.57	1.01
<i>a2</i>	3.66	60.64	2T	0.50	8.04	28.73	160.8	15.07	3.58	6.14	0.75	8.13	0.65
<i>a3</i>	6.19	71.39	2T	0.46	6.44	23.67	117.2	12.22	3.68	7.01	0.79	8.86	0.50
<i>a4</i>	4.13	59.72	2T	0.46	5.99	24.21	133.8	11.91	4.04	7.18	0.75	9.57	0.43
<i>a5</i>	4.63	65.12	2T	0.59	10.82	24.60	131.1	16.83	2.27	6.46	0.77	8.35	0.39
Mean: 8.73 $\pm$ 0.63 Ma, 7% dispersion													

Note: Rows in italics indicate grains that are outliers from the main population and not included in averages.

^aMean and 1 σ standard deviation of corrected dates reported for samples with <20% dispersion. No mean reported for samples with >20% dispersion. Dispersion is sample standard deviation divided by the mean.

^bEquivalent spherical radius (r), the radius of a sphere with the same surface area to volume ratio.

^cType of grain terminations: 2T—whole grain; 1T—one tip broken off; 0T—both tips broken off.

^deU: Effective uranium concentration, weights U and Th for their alpha productivity, compiled as [U] + 0.235 * [Th].

^eF_T is alpha-ejection correction of Ketcham et al. [81].

^fAnalytical uncertainty based on U, Th, He, and grain-length measurements.

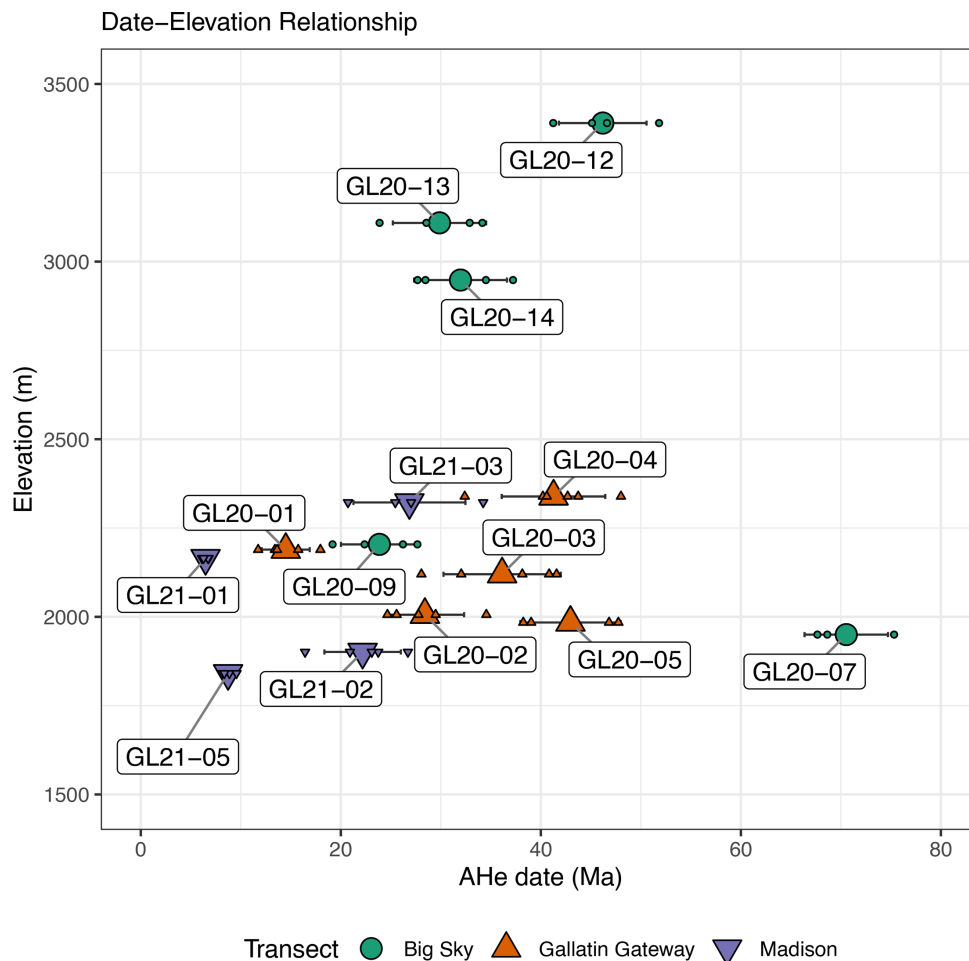


FIGURE 4: Relationship between AHe date and elevation. Mean AHe cooling dates plotted against elevation for all three transects. Large symbols denote sample means and small symbols are dates for individual grains. 1σ standard deviation for each sample mean is plotted as error bars and may be obscured by symbols.

crystallization age, several have dates similar to crystallization age indicating rapid cooling. Samples GL20-12, the top of the Big Sky transect, and GL20-07 have dates that overlap with the zircon U-Pb age (for GL20-12) or the assumed crystallization age (for GL20-07). We interpret cooling in these two samples to be reflecting post-magmatic cooling and not exhumation, suggesting they were intruded shallowly at temperatures below the AHe PRZ. One additional sample, GL20-04, from the top of the Gallatin Gateway Transect has an AHe date younger than the crystallization age, but only slightly (AHe 41.3 ± 5.2 Ma vs zircon U-Pb 51.2 ± 0.83 Ma) and HeFTy models indicate that it can cool to surface temperatures within a few Myr of emplacement (Figure 5). The first phase of cooling in the joint model of the Big Sky transect is rapid Eocene cooling (Figure 9), and individual models for most samples allow rapid Eocene cooling (Figures 5–7). We suggest that this Eocene cooling reflects rapid cooling to ambient temperatures after intrusion of these magmatic bodies into the shallow crust (Figures 10(a) and 10(b)).

Since the top of the Big Sky transect was emplaced shallower than the AHe PRZ but the lower samples in the transect were not, we can approximate the depth of

emplacement. Thermal history models for GL20-12 suggest it cooled to temperatures $<50^\circ\text{C}$ quickly following emplacement (Figure 6(c)), and the joint model shows that it likely cooled to between 40 and 50°C at this time (Figure 9). If we estimate depth using a conservative geothermal gradient of $20^\circ\text{C}/\text{km}$ and a 0°C surface temperature, 50°C would suggest a maximum emplacement depth of 2.5 km for the top of the Lone Mountain intrusion that hosts the Big Sky Transect. Realistically, the emplacement depth was likely shallower as emplacement coincided with the elevated surface temperatures of the Early Eocene Climatic Optimum [102, 103] and regional eruption of the AVP, which likely resulted in an elevated geothermal gradient [40]. Shallow emplacement is consistent with the porphyritic texture of the rocks, and some sills in the western portion of our study area were likely emplaced at <3 km depth based on structural relationships [65]. Thermobarometry on the Lone Mountain intrusion yields equilibrium pressures for amphibole and plagioclase that range from 1.5 to 7.4 kbar, suggesting the shallow sill complex was fed by a partially crystalline magma chamber at 10 – 20 km depth [26].

Sample GL20-07 cooled to near-surface temperatures in the Cretaceous (Figure 6(a)) likely indicating shallow

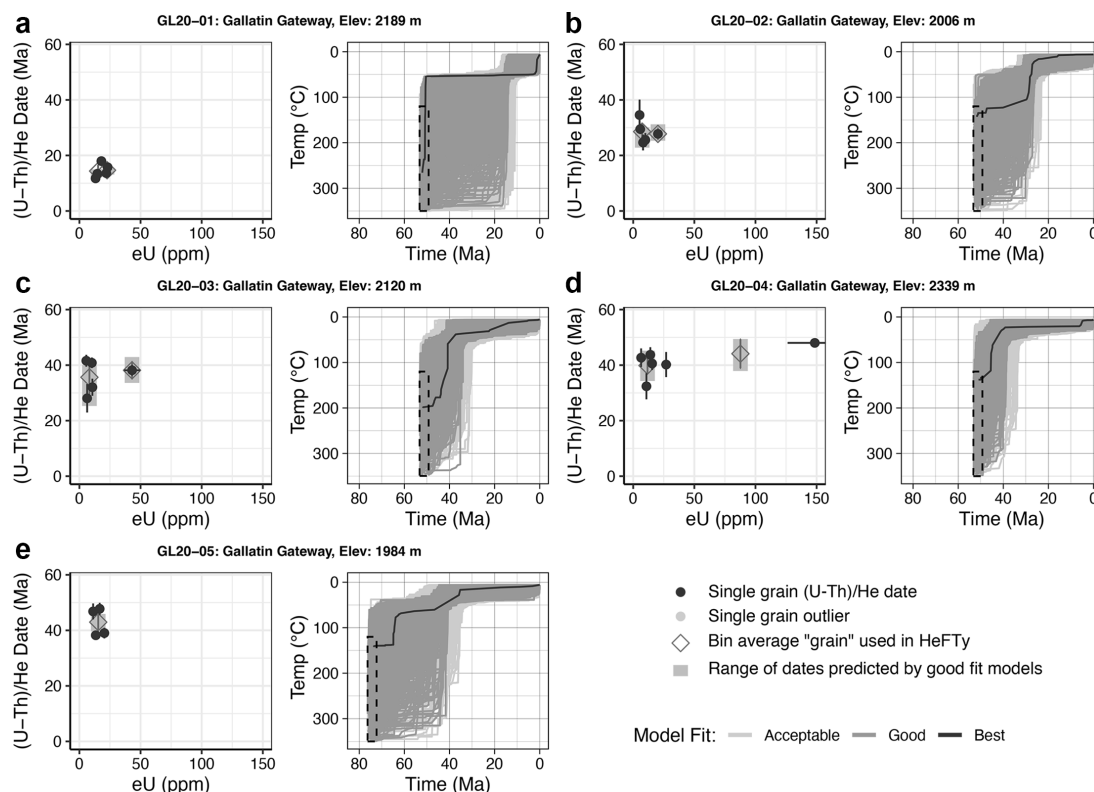


FIGURE 5: Gallatin Gateway transect AHe date and HeFTy thermal history models. (a)-(e) show apatite (U-Th-Sm)/He (AHe) versus effective uranium (eU) and HeFTy thermal history model results (time versus temperature) plotted for individual samples from the Gallatin Gateway transect with <20% dispersion (Table 4). Date-eU plots have outlier grains plotted in light gray, but these dates are not modeled or used in interpretations. Error bars for sample points are the 2σ analytical uncertainties of individual corrected grain dates and may be obscured by points. The “synthetic grains” used for thermal history modeling (the sample mean for each bin) shown as open diamonds with error bars equal to the 1σ standard deviation of the bin or 10% of the mean bin date, whichever is greater. Thermal history plots show modeling constraints as dashed boxes, the weighted mean (black line), best fit path (dark gray), good fit paths (gray), and acceptable fit paths (light gray). The range of dates predicted by good fit models for each synthetic grain is shown on the date-eU plots as a gray rectangle. Thermal history modeling constraints for all modeled samples are listed in Table 2. Abbreviations: Elev—elevation above mean sea level, Temp—temperature.

emplacement and little post-Cretaceous cooling, which is surprising given that this is the lowest elevation sample in the Big Sky transect. However, the rest of the samples in the transect are clustered within 7 km of one another, while GL20-07 is >5 km from the nearest sample and the farthest from the active Madison Fault (Figure 2). It is also lithologically different from the others, taken from a mafic dike that intrudes near the Laramide and older Spanish Peaks Fault [64, 82]. Thus, it is possible that this sample is cooling independently of the others due to the position of faults or structurally controlled topography and/or thermal structure. We therefore did not model it with the remainder of the transect that appears to record a shared exhumation history and suggest its AHe date reflects post-magmatic cooling.

The intrusion that hosts the Gallatin Gateway transect also likely intruded at similar, shallow depths to the Lone Mountain intrusion given the Eocene cooling dates for GL20-04 and GL20-03 (Table 4). Numerical models of the cooling of shallow intrusions suggest that they cool to near ambient temperatures on timescales of 10 ka-1 Ma dependent on the size and depth of

the intrusion [104]. This rapid post-magmatic cooling is followed by a second phase of slower exhumational cooling to bring them to the surface [104]. The AHe dates for these Gallatin Gateway samples are >5 Ma younger than the intrusions crystallization age, suggesting that in addition to the rapid post-magmatic cooling, they likely experienced some low magnitude post-emplacement cooling due to exhumation. GL20-05 from a neighboring intrusion has a Cretaceous crystallization age but an Eocene AHe cooling date similar to GL20-04 (Table 4). This may reflect reheating and cooling due to nearby intrusions or exhumational cooling or a combination.

5.3. Oligocene Extensional Cooling

Thermal history models from the Big Sky elevation transect show a dominant period of cooling in the Oligocene from ~30 to 23 Ma resulting in the positive date-elevation relationship we observe for that transect. This Oligocene phase of cooling was preceded by ~10 Myr of little to no cooling, making it distinct from post-magmatic cooling and

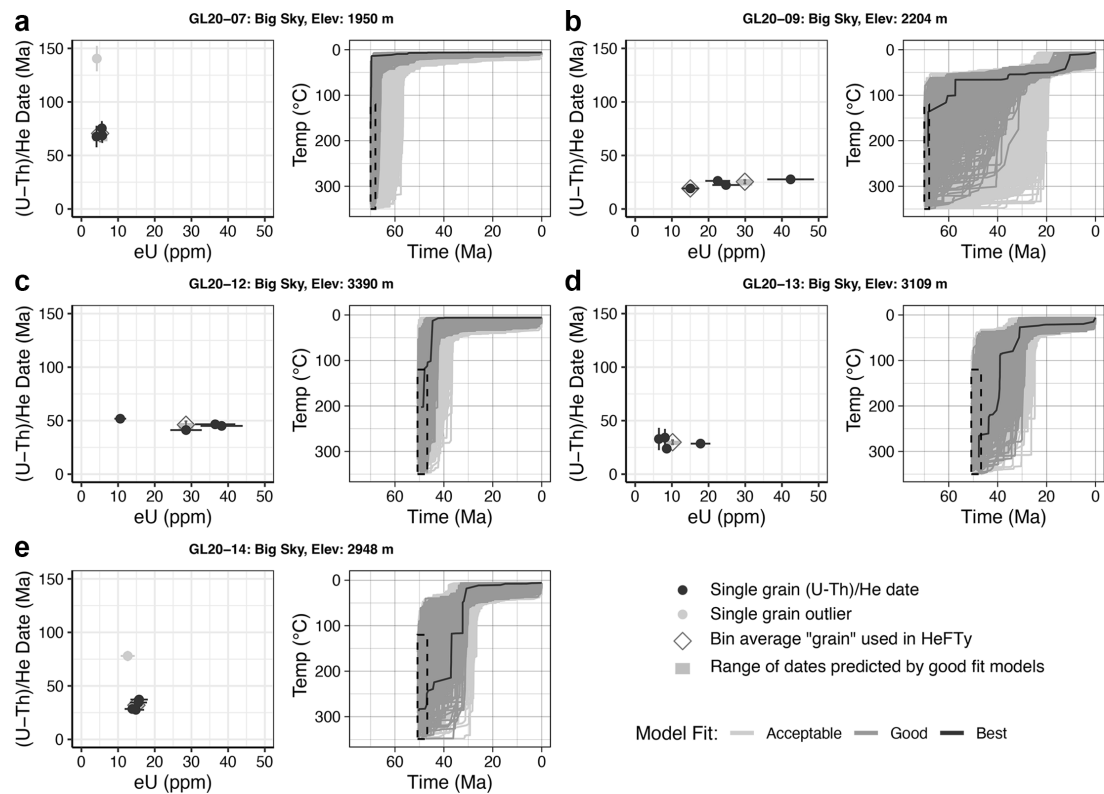


FIGURE 6: Big Sky transect AHe date and HeFTy thermal history models. (a)-(e) show AHe date-eU a HeFTy time-temperature plots for individual samples in the Big Sky transect with dispersion <20% (Table 4). Symbology and abbreviations as in Figure 5.

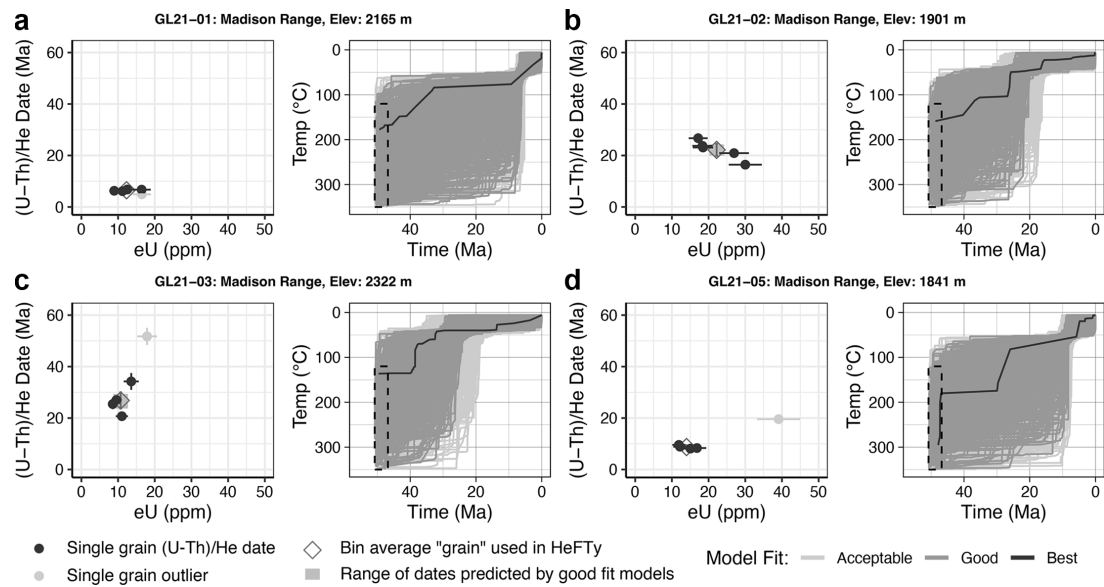


FIGURE 7: Madison Range transect AHe date and HeFTy thermal history models. (a)-(d) show AHe date-eU a HeFTy time-temperature plots for individual samples in the Madison Range transect with dispersion <20% (Table 4). Symbology and abbreviations as in Figure 5.

AVP processes. Local climate was cooling and becoming more arid through the Oligocene [12] so climatic conditions are unlikely to have accelerated erosion during that time. Tectonically, the region was in an overthickened state due to extensive contraction during the Sevier and Laramide Orogenies [8]. Extensional collapse started

~53 Ma and persisted throughout SW Montana until ~20 Ma [8, 14, 21], making tectonic exhumation (via normal faulting) and erosional exhumation plausible drivers of this Oligocene cooling phase (Figure 10(c)). Several basins surrounding our study area, including the Jefferson-Three Forks, Clarkston, and Gallatin basins (Figure 1), record

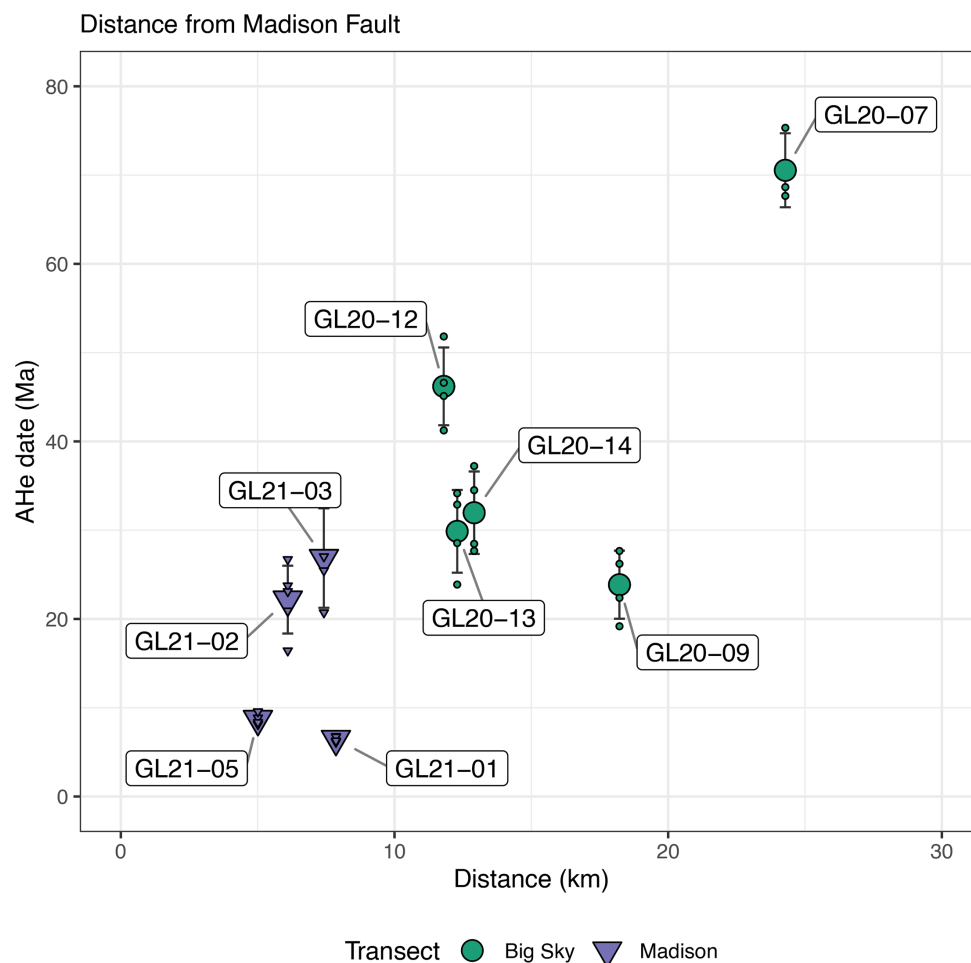


FIGURE 8: Relationship between AHe date and distance from the Madison Fault for the Big Sky and Madison Range transects. Large symbols denote sample means and small symbols are dates for individual grains. 1σ standard deviation for each sample mean is plotted as error bars and may be obscured by symbols.

outcrops of the Renova Formation suggesting potential synextensional deposition occurring at similar longitudes to our study area [10, 11]. Renova deposition initiated in the Jefferson, Townsend, and Clarkston Basins at ~37 Ma based on North American Land Mammal Ages [10], somewhat later than basins further west [105]. This may coincide with the initiation of extension in this region, but this is not necessarily the case as some Renova deposition is attributed to filling of paleovalleys and not extension [106]. The Madison Basin, adjacent to our study area to the west (Figure 1), is thought to contain Renova formation at depth, buried beneath more recent deposits [107]. Interpretation of seismic data suggests that the Renova beds thicken toward the basin bounding Madison Fault, indicating synextensional deposition [107]. Extension along the Madison Fault in the Oligocene driving exhumation and erosional cooling of the footwall block could explain the Oligocene phase of cooling observed in the Big Sky transect.

If the Madison Fault was driving Oligocene exhumation in the Madison Range, this would imply the highest topography and exhumation occurred in the footwall close to the fault, with diminishing exhumation expected moving away. This pattern is consistent with the broad pattern of

thermochronology ages with youngest ages on the west side of the range (Figures 1 and 8), and the Cretaceous date of GL20-07 may indicate that it was too far away from the fault to be affected by exhumation driven by faulting or fault-generated topography. Additionally, volcanic rocks from the Absaroka eruption are widespread in the Gallatin Range and east of the Gallatin River, but outcrops are extremely limited to the west toward the Madison Range and Madison Fault (Figure 2) [24]. The Lone Mountain intrusive complex is part of the AVP [26], so it is likely that AVP volcanic rocks were deposited above in the Madison Range and have since been removed by erosion. This indicates a pattern of increasing erosion moving west toward the Madison Fault as expected if the fault is the dominant control on topography and exhumation. This pattern could also be caused by younger, Miocene to modern motion on the Madison Fault, but is consistent with any fault motion that occurred after the AVP.

The cooling history for the Big Sky transect (Figure 9), particularly the period of rapid cooling from ~30–23 Ma bracketed by periods of slow cooling, suggests that fault-driven exhumation was active within the period of ~30–23 Ma (Figure 10(c)). This implies that fault motion was most

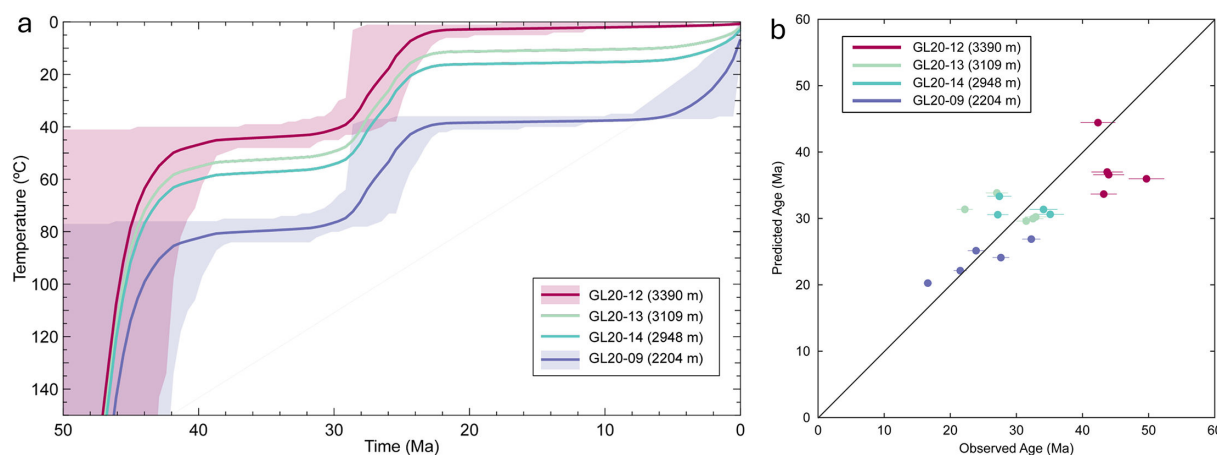


FIGURE 9: Thermal history model for the Big Sky Transect. (a) Expected model results for QTQt inverse modeling of the Big Sky Transect. Lines show the expected time-temperature for each sample, and shaded regions show the 95% credible interval for the top and bottom samples in the transect. For visual clarity, only the expected model is shown for the two intermediate elevation samples without their credible intervals. (b) AHe ages predicted by the expected model compared with the observed AHe ages. Error bars on the observed ages represent 8% uncertainty.

active during that time and likely less active both before and for some time after that period as shown by slow cooling rates (Figure 9). The end of implied fault activity is roughly coincident with the end of the post-orogenic period of regional extension [8, 12, 108]. Fault initiation that we infer at ~30 Ma is later than the initiation of extension and core complex formation that began at ~53 Ma to the west within the Sevier fold and thrust belt [8, 9, 12, 21, 105]. However, even though extension broadly began at ~50 Ma, there are several examples within the Sevier domain of individual fault initiation or renewed motion substantially after the beginning of extensional collapse in the mid-Eocene. For example, in the Muddy Creek Basin, surface breaking faulting does not initiate until ~37 Ma, though the basin began ponding earlier [109], and in the Horse Prairie basin, extension shifts to a new set of N-S oriented faults in the Oligocene that remain active through the early Miocene [108].

The Sevier and Laramide domains demonstrate differing responses to slab rollback from ~50 to 20 Ma. The Sevier domain experienced widespread extension and core complex formation [8, 9, 12–14] while the Laramide uplifts appear to be undergoing burial, heating, and sedimentation [2, 20, 110, 111]. Our study area in the westernmost Laramide domain seems to be transitional between these two responses. We do not see evidence for post-Eocene burial and sedimentation in the Madison Range, and extension on the Madison Fault is dominantly after ~30 Ma. There is evidence for extension on other structures within the westernmost Laramide domain starting at ~34 Ma [112, 113], but on the whole, Eocene-Oligocene extension in this region appears to start later and be less pervasive than to the west in the Sevier province. Paleobotanical proxies suggest that basin floors in the Sevier domain had lowered to elevations of ~1.5 km by 35 Ma, while valley floors in the western Laramide domain retained elevations of up to 2.5 km [114]. Perhaps extension and surface lowering were preferentially hosted in areas that

experienced Sevier-style deformation before propagating into the still elevated western part of Laramide domain while burial and subsidence were beginning farther east.

5.4. Miocene Madison Fault-Driven Exhumation

The Madison transect overall yielded younger AHe dates with two samples with Miocene cooling ages (Table 4). These young dates show cooling that post-dates the main phase of Oligocene cooling documented in the Big Sky transect and indicate a younger phase of cooling close to the Madison Fault. The Madison Fault extends 98 km in length and is sectioned into northern, central, and southern portions with moderately constrained traces and has been active in the Quaternary [67]. It is generally thought to have initiated in the Miocene with Basin and Range extension or due to the YSH [115–117], though seismic interpretations from the Madison Valley [107] and our data suggest a phase of pre-Miocene extension as well. The Madison Fault lies within the bands of active seismicity fanning out from the YSH [117] with proposed peak fault activity from ~2.0 to 0 Ma coeval with hotspot migration [55, 59, 117]. The fault scarp cuts Quaternary and upper Pleistocene deposits [24], and most recently minor offset occurred on some southern segments of the Madison fault during the 1959 Hebgen Lake earthquake [55, 67]. Evidence for movement along the southern portion of the Madison Fault is better constrained and suggests activity with estimated quaternary recurrence interval of ~10–25 ka [67]. Basin depths in the Madison Valley hanging wall may be >4.5 km in the central portion [107, 118], and throw on this fault system from Eocene to present may reach ~8 km [55, 67, 119]. The 2 Ma Huckleberry Ridge tuff is offset by 800–900 m in the central portion of the fault [59].

Figure 8 shows the Madison Range AHe dates plotted against approximate distance to the Madison Fault. The closest sample to the fault, ~5 km away, is one of the youngest cooling in our sample suite at 8.7 ± 0.6 Ma and

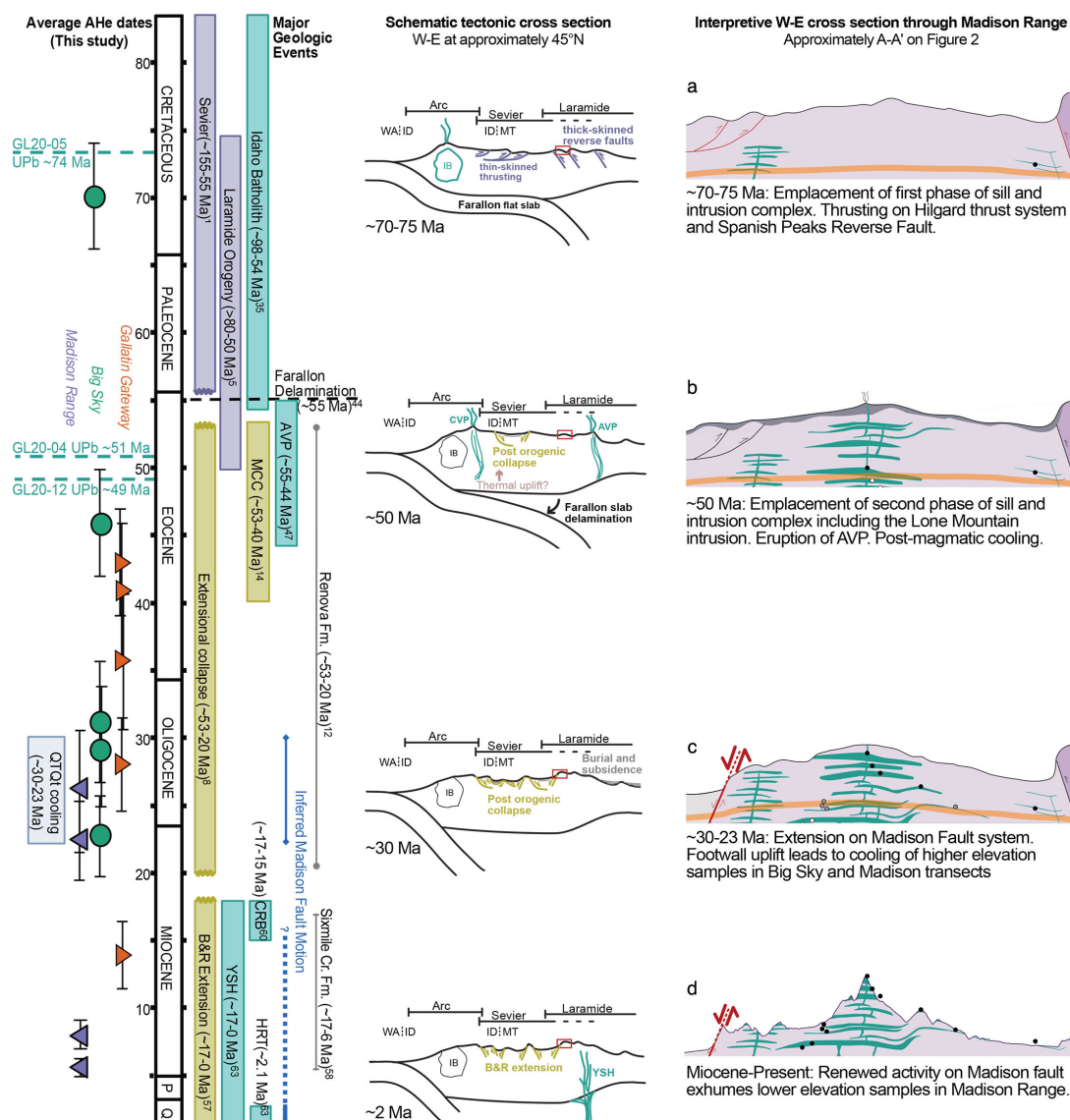


FIGURE 10: Timeline diagram depicting generalized major geologic events within the Cordillera Orogenic system and schematic reconstruction of the Madison Range. References numbers given as superscripts and cited within the text. Mean AHe dates for each transect are presented at the left with error bars representing 1σ standard deviation for each sample. Abbreviations: AVP—Absaroka Volcanic Province, B&R Extension—Basin and Range Extension, CRB—Columbia River Basalts, CVP—Challis Volcanic Province, HRT—Huckleberry Ridge Tuff, MCC—Metamorphic Core Complexes, IB—Idaho Batholith, P—Pliocene, Q—Quaternary, Renova Fm.—Renova Formation, YSH—Yellowstone Hotspot. Dashed teal lines are U-Pb emplacement dates. The middle column shows a schematic cross section of major tectonic events from ~108 to 115°W along ~45°N. Red box indicates approximate location of study area shown in panels A–D. Panels A–D show an interpretive reconstruction of major events recorded in sample thermal histories. Section is approximately A–A' in Figure 2 (same section as Figure 2(b)). Colors of geologic units as in Figure 2. The orange line represents the approximate AHe closure isotherm. Circles represent sample locations with white circles representing a sample above the AHe closure isotherm with open system behavior, gray circles representing samples in the PRZ, and black circles representing samples that have cooled below the closure isotherm.

broadly dates get older away from the fault. Additionally, a Pre-Cambrian basement sample from Sphinx Mountain, ~15 km south of our study area and also in the Madison Fault footwall (Figure 1), yielded a central AFT age of 18 ± 2.4 Ma with individual AHe grains ranging from 12 to 32 Ma, some of the youngest dates from Pre-Cambrian rocks in the region [2]. While some of the pattern shown for the Madison transect in Figure 8 is likely due to Oligocene motion on the fault, the two late Miocene dates close to the

fault and evidence for recent fault motion show a renewed phase of motion beginning in the Miocene (Figure 10(d)).

Our data lack a clear elevation relationship and sufficient resolution to constrain the timing and rate of Miocene motion. The pattern within the Madison transect is also potentially disrupted by secondary faults as the youngest date, GL21-01, is removed from the main fault but near a mapped secondary fault [82]. The 800–900 m offset of the Huckleberry Ridge tuff suggests an average rate of

motion of 0.4–0.45 mm/yr over the last 2 Myr. If the rate of fault motion has been constant since initiation, then the total maximum throw of ~8 km would imply initiation at ~20 Ma. However, we interpret exhumation in our Big Sky transect to indicate a period of fault motion from ~30 to 23 Ma, and the oldest two dates in the Madison Transect coincide with this time window (Figure 10). Seismic interpretation of the subsurface also suggests that deposits of the Renova Formation (Eocene-Oligocene), the Sixmile Creek Formation (Miocene), and recent basin fill all dip and thicken toward the Madison Fault indicating syndepositional fault motion for all of those strata [107]. We therefore suggest that the Madison Fault was actively extending from ~30 to 23 Ma, followed by a reduction or cessation in motion and then renewed activity in the Miocene to present. We cannot determine exactly when the more recent phase of motion initiated, though the final stage of cooling in models of Big Sky transect suggests it may have accelerated after ~8 Ma (Figure 9).

5.5. Minimal YSH Influence on Cooling

All of our AHe dates predate the arrival of the YSH to its current location, marked by the eruption of the Huckleberry Ridge Tuff at ~2.0 Ma (Figure 10) [59]. Our youngest AHe date is 6.5 ± 0.28 Ma (Table 4), which predates the Huckleberry Ridge tuff by 4.5 Myr, though it does overlap with activity in the Heise caldera to the southwest (Figure 1; active 6.7–4.4 Ma [120]). None of our samples require erosion after 2 Ma, meaning the landscape's erosional response to any YSH-induced uplift has been limited to below AHe resolution, but most samples do allow for some cooling during this window (Figures 5–7). The maximum amount of cooling allowed by all samples according to our good-fit HeFTy t-T paths between ~2.0 and 0 Ma is ~18–46°C (Figures 5–7). Using estimates for the geothermal gradient, we can estimate an upper limit for exhumation during this period. Using a typical continental geothermal gradient of ~25°C/km this amounts to a maximum of between ~0.72 and 1.8 km since 2 Ma. Using the present-day estimated geotherm for our study area of ~30°C/km [94, 121], likely increased by the arrival of the hotspot, results in a maximum of ~0.60–1.5 km of exhumation since the arrival of the YSH. Similar maximum magnitudes for most samples are allowed when considering the last 6 Myr, and many samples may have experienced little to no exhumation over this period. The exceptions to this are the two youngest samples from the Madison transect (GL21-01, GL21-05), which we attribute to fault motion. We do not observe a regional km-scale erosion event contemporaneous with the arrival of the hotspot.

Based on established evidence of deformation, topographic evolution, and drainage disruption contributed through dynamic and thermal processes around the YSH and YCHT [50, 57, 59, 61–63, 68, 122, 123], we originally hypothesized that the arrival of the hotspot would have a more significant contribution to exhumation-driven cooling in our study area than we observe. This contrasts with previous work along the hotspot track that attributed

significant exhumation and topographic changes to hotspot activity. In the Pioneer-Boulder Mountains, Idaho (Figure 1), AHe date indicates >2–3 km of exhumation since ~11 Ma that was ascribed to hotspot processes from when the YSH was located at the nearby Twin Falls (12.7–8.5 Ma) and Picabo (10.3–8.2 Ma) volcanic fields [63]. Enhanced river incision in the central Idaho mountains, decreasing in magnitude to the north away from hotspot activity, initiated at this time, likely due to hotspot-lithosphere interactions [124, 125]. The amount of exhumation in the Pioneer-Boulder Mountains exceeds what is possible in our study area even with a “normal” geothermal gradient.

We see two possible explanations for the discrepancies in exhumation magnitudes between the two study areas. First, it is possible that fundamental differences in the lithosphere between central Idaho and SW Montana lead to different amounts of hotspot-driven uplift and exhumation. The Pioneer-Boulder Mountains abut the Cretaceous Idaho Batholith, and the magnitude and extent of river incision in the region at the time were likely driven by density changes in the lithosphere, potentially due to delamination of a dense root beneath the Idaho Batholith [124]. Our study area lies much farther east within the North American craton in Archean lithosphere with a crustal thickness of ~44 km compared to that of the Pioneer-Boulder Mountains in Proterozoic lithosphere with a crustal thickness of ~36–38 km [126]. The interaction of the mantle plume with the older lithosphere in SW Montana may be manifesting differently, resulting in less exhumation than it did in areas with younger lithosphere.

The second explanation for the differences in exhumation magnitude is that the subsidence of the caldera and flexural upwarping of the flanks after the hotspot passes may drive more exhumation than the uplift associated with the approach of the hotspot. Since the hotspot passed near the Pioneer-Boulder Mountains, those calderas have been subject to considerable subsidence and the formation of the ESRP. Some of the exhumation in the Pioneer-Boulder Mountains may have been driven by subsidence, flexure, and lower crustal flow due to loading in the ESRP [127, 128] and increasing topographic relief between the YCHT and the ESRP. Our study area is on the flanks of the active hotspot and has not yet experienced the subsidence phase. The minimal amount of exhumation in our study area may be indicating that subsidence and flexure after the hotspot's passage drives more erosion and landscape change than any thermal or dynamic uplift associated with its arrival. This is consistent with recent work that suggests that the arrival of the YSH has reduced topographic relief in the YCHT [129].

6. CONCLUSIONS

Bedrock AHe thermochronology and zircon U-Pb crystallization ages of Late Cretaceous and Eocene intrusions in the Madison and Gallatin Ranges of southwest Montana provide new insights into post-Laramide cooling and identify Cenozoic drivers of landscape evolution in this region. We report three new crystallization ages for these intrusions, and AHe date and modeling support three main

phases of cooling (Figure 10): (1) Eocene post-magmatic cooling (Figure 10(b)), (2) Oligocene cooling from ~30 to 23 Ma that we attribute to extension on the Madison Fault (Figure 10(c)), and (3) localized Miocene and younger cooling driven by another phase of fault motion (Figure 10(d)).

Zircon U-Pb dates from three intrusions yielded dates of 49.3 ± 1.0 Ma, 51.2 ± 0.83 Ma, and 74.5 ± 0.97 Ma. The Eocene dates are contemporaneous with the AVP, and those intrusions were likely part of the Absaroka magmatic event. The Cretaceous date is similar to a previously published hornblende $^{40}\text{Ar}/^{39}\text{Ar}$ date from a similar intrusive complex in the Madison Range. Together, this suggests a complex history of magmatism in the region with phases of intrusion at ~50 and ~70 Ma (Figures 10(a) and 10(b)).

The cooling dates range from 67 ± 8.3 Ma to 6.2 ± 0.76 Ma, though the majority are between 20 and 45 Ma. Samples within the Madison Range tend to be youngest on the west side of the range near the Madison Fault and get older away from the fault (Figure 8). The Big Sky sample transect has a positive relationship between AHe dates and elevation and, when modeled jointly, thermal history models from this transect show a phase of cooling from ~30 to 23 Ma (Figure 9). We interpret this phase of exhumation to be driven by extensional exhumation and erosion of the footwall of the Madison Fault (Figure 10(c)). This suggests extension initiated on this fault ~30 Ma, showing that post-orogenic extension and collapse propagated into the Laramide domain in the Oligocene.

The thermal history models for the Big Sky transect show a final phase of cooling for the lower samples in the late Miocene (Figure 9), and there are two AHe dates <10 Ma. Samples with these young cooling dates are proximal to the Madison Fault and are indicative of a Miocene and younger phase of motion on the fault (Figure 10(d)). Together, our results suggest an Oligocene phase of extension on the Madison, followed by a period of relative quiescence and an increase in motion from Miocene to present either driven by Basin and Range extension or the YSH or both.

Recent erosion due to regional uplift and doming from the arrival of the YSH is minimal. This contrasts with observations from central Idaho where 2–3 km of hotspot-driven exhumation was inferred from the passage of the YSH through the Twin Falls and Picabo volcanic fields [63]. This either suggests that the impact of the YSH on the lithosphere is different between the two locations or that subsidence after the hotspot's passage drives more erosion than hotspot-driven uplift.

Data Availability

All data are presented in tables in the main text and supplementary materials associated with this article.

Conflicts of Interest

The authors declare that there is no conflict of interest regarding the publication of this article.

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Supplementary Materials

Supplementary materials include AHe date vs. eU and AHe date vs. grain size for dispersed samples GL20-08 and GL21-04 (Figure S1), alternate thermal history models for GL20-05, GL20-07, and GL20-09 (Figure S2), complete data for the zircon U-Pb analyses (Table S1), and input text file and output run file for QTQt models (Appendix 1-5).

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