

Regional Thermal History Trends From the Idaho Montana Fold - Thrust Belt Using Multiple Low - T Thermochronometers

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RESEARCH ARTICLE

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Fold-and-Thrust Belts: Evolution and Dynamics at All Spatiotemporal Scales

Key Points:

- Apatite and zircon (U-Th)/He, and apatite fission track data records regional thermal histories for the Idaho-Montana fold-thrust belt
- Overlapping cooling signatures from southwestern Montana basement uplifts and the Idaho fold-thrust belt suggest concurrent deformation
- Thermal history models show two phases of cooling at ~125 and ~90–80 Ma, the latter occurring over a wide region of the fold-thrust belt

Supporting Information:

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

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Regional Thermal History Trends From the Idaho-Montana Fold-Thrust Belt Using Multiple Low-T Thermochronometers

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Abstract Low-temperature thermochronometric data can reveal the long-term evolution of erosion, uplift, and thrusting in fold-thrust belts. We present results from central Idaho and southwestern Montana, where the close spatial overlap of the Sevier fold-thrust belt and Laramide style, basement-involved foreland uplifts signify a complex region with an unresolved, long-term tectono-thermal history. Inverse QTQt thermal history modeling of new zircon (U-Th)/He (ZHe, $n = 106$), and apatite (U-Th)/He dates (AHe, $n = 43$) collected from hanging walls of major thrusts systems along a central Idaho to southwestern Montana transect, and apatite fission track results from 6 basement samples, reveal regional thermal and spatial trends related to Sevier and Laramide orogenesis. Inverse modeling of foreland basement uplift samples suggest Phanerozoic exhumation initiated as early as ~80 Ma and continued through the early Paleogene. Inverse modeling of interior Idaho fold-thrust belt ZHe samples documents Early Cretaceous cooling at ~125 Ma in the Lost River Range (western transect), and a younger cooling episode in the Lemhi Arch region (mid-transect) at ~90–80 Ma through the late Paleogene. This cooling in the Lemhi Arch temporally overlaps with cooling in southwestern Montana's basement-cored uplifts, which we interpret as roughly synchronous exhumation related to contractional tectonics and post-orogenic collapse. These data and models, integrated with independent timing constraints from foreland basin strata and previously published thermochronometric results, suggests that middle Cretaceous deformation of southwestern Montana's basement-cored uplifts was low magnitude and preceded tectonism along the classic Arizona-Wyoming Laramide “corridor.” In contrast, Late Cretaceous and Paleogene thrust-related exhumation was more significant and largely complete by the Eocene. The basement-involved deformation was contemporaneous with and younger than along-strike Sevier belt thrusting in central Idaho.

1. Introduction

During the Late Cretaceous and early Cenozoic, the northern U.S. Cordillera encompassed two styles of retroarc deformation: the thin-skinned style Sevier fold-thrust belt where shortening was accommodated by thrusts within sedimentary rocks, and the thick-skinned style Laramide fold-thrust belt typified by reverse faults that carried crystalline basement rocks in their hanging walls (Armstrong, 1968; Perry et al., 1983). Both structural styles were also differentiated by their relative positions in the Cordilleran retroarc: thin-skinned belts most often occur in a retroarc setting and carry sedimentary packages in their hanging walls, whereas Laramide style thick-skinned structures are characterized by foreland thrusts beyond the frontal most thin-skinned structure (Armstrong, 1968). The transition from thin-to thick-skinned thrusting and a coeval inboard sweep of arc magmatism in the U.S. Cordillera is often interpreted as a result of the initiation of shallowly dipping subduction at the plate boundary (Burchfiel & Davis, 1975; Coney & Reynolds, 1977; Copeland et al., 2017; Dickinson et al., 1978; Heller & Liu, 2016; Jordan & Allmendinger, 1986; Lipman et al., 1971; Saleeby, 2003; Yonkee & Weil, 2015). However, more recent studies have advanced debates on the relationship between these two structural styles, including their temporal and spatial overlap, which, in turn, has called into question whether the cause-and-effect relationship between thick-skinned thrusts and shallow subduction has widespread applicability (Garber et al., 2020; Horton et al., 2022; Parker & Pearson, 2021; Schmidt & Perry, 1988; Solum & van der Pluijm, 2007; Yonkee & Weil, 2015). Instead, some regions of thick-skinned thrusting—both within (e.g., Parker & Pearson, 2021) and beyond (e.g., Kley et al., 1999) the North American Cordillera—do not have an apparent spatial or temporal

relationship with subduction of a shallowly dipping plate. Thus, other factors were suggested to exert the primary control on structural style, such as the pre-orogenic crustal architecture and the related influence of changes to the pre-deformation sedimentary basin thickness (Allmendinger et al., 1983; Allmendinger & Gubbels, 1996; Kley et al., 1999; Kulik et al., 1988; Lacombe & Bellahsen, 2016; Marshak et al., 2000; McBride et al., 1992; Parker & Pearson, 2021; Paulsen & Marshak, 1999; Pearson et al., 2013; Tavani et al., 2021).

In most parts of the U.S. Cordillera, thick-skinned thrusting largely occurred after thin-skinned thrusting; some authors use this sequence as indicative and diagnostic of changes in plate boundary dynamics (Anderson et al., 2007; Jordan & Allmendinger, 1986; Jordan et al., 1983; Kay et al., 1987). Therefore, in addition to the spatial relationship between thin- and thick-skinned uplifts, the timing sequence of this transition is important to diagnose the causes of the transition. For example, the effects of upper-crustal architecture on structural style and the influence of pre-existing structural weaknesses on subsequent deformation can create a non-sequential order of thrusting wherein thick-skinned thrusts may not exclusively occur in the foreland, nor do they break only after thin-skinned structures have developed (Marshak et al., 2000; Parker & Pearson, 2021; Pearson et al., 2013; Pfiffner, 2006).

The Idaho-Montana fold-thrust belt and southwestern Montana foreland provide excellent exposures to investigate the timing and potential drivers of the observed change in structural style (Figure 1). Located north of the interpreted influence of the Farallon flat slab, it is unclear if basement-involved thrusts of southwestern Montana can be attributed to flat slab subduction (e.g., Ronemus et al., 2023), as is accepted for other Laramide uplifts farther south (Copeland et al., 2017; DeCelles, 2004; Liu et al., 2010; Peyton et al., 2012). Our study area consists of a transect spanning from the hinterland portion of the Idaho-Montana fold thrust belt northeastward through the basement uplifts of southwestern Montana that offers exposures of both styles of thrusting (e.g., Perry et al., 1983). This portion of the US Cordillera has long been discussed for its overlapping structural styles (Kulik et al., 1988; McDowell, 1997; O'Neill et al., 1990; Parker & Pearson, 2021, 2023; Perry et al., 1988, 1989; Schmidt & Perry, 1988; Skipp, 1988; Tysdal, 1988, 2002). Evidence for temporal overlap between the structures in the Idaho-Montana thrust belt and Montana basement foreland has been documented through sedimentary records of the early onset of basement uplifts (DeCelles, 1986; DeCelles et al., 1987, 1991; Finzel et al., 2023; Garber et al., 2020; Nichols et al., 1985; Perry et al., 1983) and thermochronologic studies of these basement uplifts (Carrapa et al., 2019; Kaempfer et al., 2021; Ronemus et al., 2023). Previously, studies have constrained the timing of exhumation in discrete portions of this transect (Garber et al., 2020; Hansen & Pearson, 2016); however, consideration of data sets from multiple thermochronometers in a well-constrained structural context could constrain thrust exhumation and address the evolution of the transect as a whole.

Here, we present low temperature thermochronometric data from the zircon (U-Th)/He, apatite (U-Th)/He, and apatite fission track (hereafter ZHe, AHe, and AFT) systems, collected along a complete transect across the Idaho-Montana fold-thrust belt. These data are combined with thermal history modeling to examine the timing of cooling. Cooling ages provided by these thermochronometers are interpreted to record exhumation, which provides insight into the timing of hanging wall uplift. Exhumation does not always equate to thrust timing, and could record passive unroofing or post-orogenic erosion (Gavillot et al., 2010; Reiners & Brandon, 2006; Shorten & Fitzgerald, 2019). However, the pattern of rapid cooling displayed in our samples, coupled with contemporary syntectonic sedimentary records of basement exposure (Decelles & Coogan, 2006; Garber et al., 2020), allows us to interpret exhumation as up-ramp movement of hanging wall rocks. Our thermal history results, combined with previous structural insights, constrains the timing of Cretaceous-Paleogene deformation. This approach highlights the importance of placing thermal history information from discrete hanging walls into a broader structural context when striving to understand how thick-skinned thrust systems develop. These thermal histories inform ongoing debates over the controls of thrusting styles in the U.S. Cordillera, specifically the degree of spatial and temporal overlap between thick- and thin-skinned portions of the Cordillera. In our transect, the temporal context in particular reveals whether changing plate boundary conditions, or other factors, such as a pre-existing structural architecture, are most relevant in creating thick- and thin-skinned thrust systems.

2. Geologic Background

2.1. Stratigraphic and Basement Context

Along a northeast-southwest transect across the Idaho-Montana fold-thrust belt, pre-Mesozoic strata in central Idaho consist of an ~7 km thick stack of Neoproterozoic and Paleozoic rift and passive margin sedimentary rocks

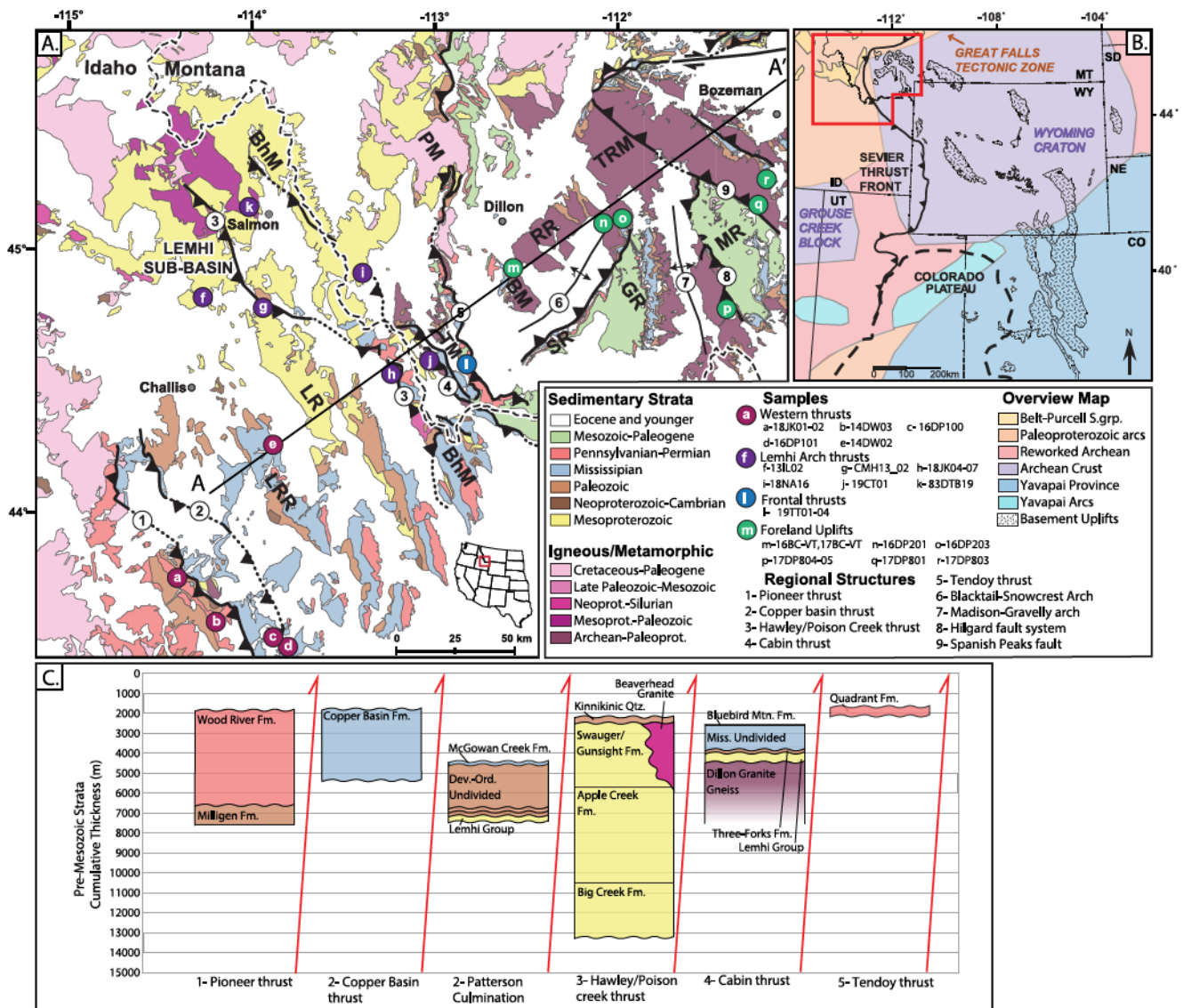


Figure 1. (a) Map of sample area, with a (b) tectonic overview inset map showing region of interest relative to other Laramide uplifts and the Cordilleran thrust front. Line from A to A' is the location and orientation of the transect line used in Figures 2, 8, and 9. Sample locations are as follows: Western thrusts (magenta "a" through "e") are the Pioneer thrust, Copper Basin thrust and the Lost River Range; Lemhi Arch thrusts (purple "f" through "k") include Iron Lake thrust, Poison Creek thrust, and Hawley Creek thrust, the Bloody Dick pluton carried in the Monument thrust sheet, the Cabin thrust; Frontal thrust (blue "m") is the Tendency thrust; Foreland uplifts (teal "l" through "s") including the Blacktail-Snowcrest arch, Madison-Gravelly uplift, and the Northern Madison Tobacco Root uplift. Overview map adapted from Peyton and Carrapa (2013), geologic map adapted from Garber et al. (2020). (c) Schematic stratigraphic columns of Paleozoic stratigraphic overburden carried in thrust hanging walls along our transect. MR: Madison Range; GR: Gravelley Range; PM: Pioneer Mountains; BhM: Beaverhead Mountains; LR: Lemhi Range; LRR: Lost River Range; BM: Blacktail Mountains; TRM: Tobacco Root Mountains.

that thin to the east above a basement high called the Lemhi arch (Figure 1) (Brennan et al., 2023; Grader et al., 2016; Scholten, 1957; Sloss, 1954). To the northeast, above the Lemhi arch in east-central Idaho, ~2 km of Ordovician and younger strata rest unconformably upon tilted Mesoproterozoic quartzites of the Belt Supergroup (Brennan et al., 2023; Parker & Pearson, 2021; Scholten, 1957). These thinner strata host two pre-orogenic unconformities: an Ordovician unconformity placing Kinnikinnick Quartzite over Mesoproterozoic quartzite and rocks of the ~500 Ma Beaverhead plutons, and a middle Devonian unconformity at the base of the Jefferson Formation (Grader et al., 2016; Link et al., 2017; Lonn et al., 2019; Lund et al., 2010; Scholten & Hait, 1962). These unconformities and across-strike stratigraphic changes limited available and continuous weak layers for thrusting during Late Cretaceous to early Paleogene deformation (Parker & Pearson, 2021). East of the Lemhi

arch in southwestern Montana, Archean and Paleoproterozoic crystalline basement rocks are nonconformably overlain by a <2 km thick stack of Cambrian and younger strata (e.g., Perry et al., 1983).

Southwestern Montana and central Idaho basement rocks consist of the northwestern Wyoming Craton and Montana metasedimentary province, with the oldest rocks recording zircon U-Pb ages between 3.5 and 3.2 Ga (Jones, 2008; Mogk et al., 1992, 2020; Mueller et al., 2004, 2011). High-grade metamorphism between 2.7 and 2.45 Ga is observed in exposed basement rocks in the Pioneer Mountains of central Idaho (Link et al., 2017) as well as the Ruby Range, Tendoy Mountains, and Beaverhead Mountains of southwestern Montana (Alcock & Muller, 2012; Anderson, 2017; Cramer, 2015; James, 1990; Jones, 2008; Kellogg et al., 2003; Mogk et al., 2020; Mueller et al., 1993, 2016; Pearson et al., 2017). The Montana metasedimentary province experienced >700°C metamorphism during the Great Falls and Big Sky orogenies at ~1.86–1.74 Ga (Condit et al., 2015; Gifford et al., 2020; Harms & Baldwin, 2020, 2023; Harms et al., 2004; Mueller et al., 2002, 2016; O'Neill & Lopez, 1985). Basement rocks were likely exhumed during late Neoproterozoic to early Cambrian time (Kaempfer et al., 2021; Ronemus et al., 2023), coincident with early Rodinia rifting (e.g., Brennan et al., 2023), and exposed at the surface prior to unconformable deposition of the middle Cambrian Flathead Sandstone.

2.2. Structural Context

Within the study area, Parker and Pearson's (2021) "double-decker" model for the fold-thrust belt consists of thin-skinned thrusts that deformed a relatively thin stack of sedimentary cover rocks, underlain by a system of thick-skinned thrusts that deformed Mesoproterozoic quartzites and crystalline basement rocks beneath and east of the Lemhi arch unconformity. Thrusts developed within the thin-skinned central Idaho portion of the Idaho-Montana fold-thrust belt include the westernmost Pioneer and Copper Basin thrusts, which were detached in lower Paleozoic sedimentary rocks (Figure 1; Dover, 1983; Link et al., 1996; Rodgers & Janecke, 1992; Rodgers et al., 1995). Farther east, above and on the flanks of the Lemhi arch, thin-skinned thrusts were detached in Devonian and Mississippian sedimentary rocks, including an unnamed thrust system in the Lost River Range (Anastasio et al., 1997), and the Thompson Gulch, Fritz Creek, Medicine Lodge, McKenzie, and Tendoy thrusts in the Lemhi Range, Beaverhead Mountains, and Tendoy Mountains of east-central Idaho and southwestern Montana (Figure 1; e.g., Parker & Pearson, 2021, 2023).

Within the structurally deeper deck of the double-decker thrust system, thick-skinned structures include thrusts within the Patterson culmination (Janecke et al., 2000), the Iron Lake fault (Burmester et al., 2023; Lund et al., 2011), Poison Creek/Hawley Creek fault (interpreted as along-strike equivalents or imbricates; Evans & Green, 2003; Hansen & Pearson, 2016; Lund, 2018; Skipp, 1988; Tysdal, 2002), Diamond Creek fault (Burmester et al., 2020, 2023; Lund et al., 2003) and Cabin/Monument/Freeman thrust (interpreted as along-strike equivalents; Lonn et al., 2016; Skipp, 1988) (Figures 1 and 2). These thrusts carried Mesoproterozoic quartzites or Archean and Paleoproterozoic basement rocks, and are interpreted to have been detached in the middle crust (Parker & Pearson, 2021, 2023). The Poison Creek and Hawley Creek faults were interpreted by prior workers to be reactivated Paleozoic or older normal faults (Hansen & Pearson, 2016; Parker & Pearson, 2021).

In southwestern Montana, the more deeply detached, basement-involved thrusts of the lower thrust system dominate the fold-thrust belt (Figure 2; Parker & Pearson, 2021). These include the Snowcrest-Greenhorn thrust system, which carried the Blacktail-Snowcrest arch in its hanging wall (Schmidt & Garihan, 1983) and was interpreted to reactivate an older Paleozoic normal fault (McBride et al., 1989; Perry et al., 1983); the Scarface-Hilgard thrust system, which carried the southern Madison-Gravelly arch in its hanging wall (DeCelles et al., 1987; Perry et al., 1983); and the Bismark-Spanish Peaks-Gardiner fault system, which carried the Tobacco Root-northern Madison arch in its hanging wall (e.g., Garihan et al., 1983).

2.3. Existing Timing Estimates for Thrusting

Existing timing constraints for thrusting in the Idaho-Montana fold-thrust belt come from depositional ages of foreland basin strata and suggest that shortening primarily occurred during Late Cretaceous-Paleogene time (DeCelles et al., 1987; Finzel et al., 2023; Garber et al., 2020; Gardner et al., 2022; Laskowski et al., 2013; McDowell, 1997; Nichols et al., 1985; Perry et al., 1983, 1988; Perry & Sando, 1982; Rosenblume et al., 2021, 2022; Ryder & Scholten, 1973) or more rarely, from the ages of cross-cutting igneous rocks (Kellogg & Harlan, 2007; Rodgers et al., 1995). Thus, additional timing constraints from thrust hanging walls have the potential to better constrain the progression of thrusting in the Idaho-Montana fold-thrust belt.

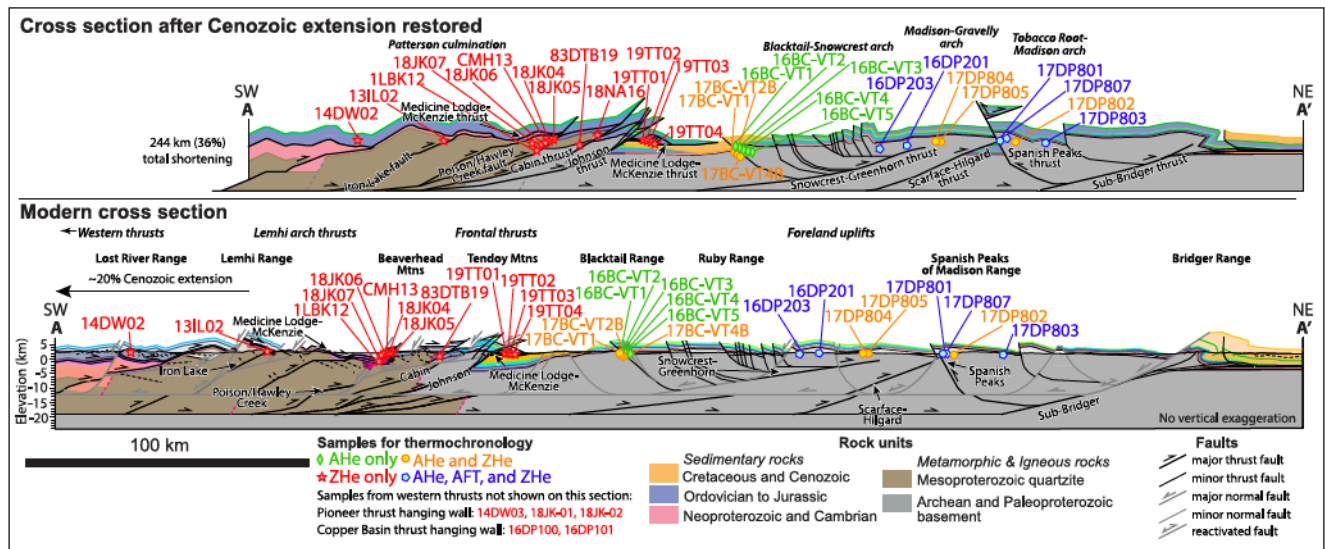


Figure 2. Balanced cross-section (modified from Parker & Pearson, 2023) along line A to A' in Figure 1 with samples projected into their corresponding structural positions. The top panel shows Cenozoic extension restored, whereas the bottom panel shows the modern positions of faults and projected samples. Names in bold above the section show the names of major structures and mountain ranges. Various faults are also labeled on the cross sections.

Within the thin-skinned, central Idaho portion of the Idaho-Montana fold-thrust belt, lower Paleozoic sedimentary rocks are interpreted to have been exhumed by the Pioneer and subsidiary thrusts and recycled into foreland basin strata of the Kootenai and lower Blackleaf formations in southwestern Montana, suggesting thrusting between 135 and 104 Ma (Finzel & Rosenblume, 2021; Rosenblume et al., 2021, 2022). At ~106 Ma, the source for foreland basin sediments in the Blackleaf Formation shifted to recycled sediment interpreted to come from Triassic to Permian strata exhumed by thin-skinned thrusts in east-central Idaho as the fold-thrust belt propagated in-sequence into middle Paleozoic and younger strata above the Lemhi arch (Gardner et al., 2022). Thin-skinned thrusting in the upper deck of the double-decker thrust system is interpreted to have continued throughout deposition of the 93–85 Ma Frontier Formation through deposition of Beaverhead Group conglomerates in southwestern Montana (Finzel et al., 2023; Garber et al., 2020). Early exhumation along thick-skinned thrusts of the deeper thrust system as well as the basement-involved Blacktail-Snowcrest arch in southwestern Montana, is interpreted to have begun at or prior to ~87 Ma during deposition of the Frontier Formation and Beaverhead Group (Finzel et al., 2023; Garber et al., 2020). One ZHe sample collected from Mesoproterozoic quartzites from the structurally deeper, thick-skinned Patterson culmination beneath the Lemhi arch records an average date of 87.7 ± 4.6 Ma (Fayon et al., 2017).

Collectively, these data suggest that an early thin-skinned thrust system developed within sedimentary rocks in central Idaho to southwestern Montana, followed by activation of structurally deeper thrusts that carried Mesoproterozoic quartzites and Archean and Paleoproterozoic basement rocks. The best available “absolute” timing constraints suggest ~135a to 69 Ma thin-skinned thrusting (Gardner et al., 2022; Rosenblume et al., 2021, 2022), with initiation of thick-skinned thrusting at ~87 Ma (Finzel et al., 2023; Garber et al., 2020; Nichols et al., 1985) and cessation after ~66 Ma and before 57 Ma (DeCelles et al., 1987; Garber et al., 2020; Kellogg & Harlan, 2007). Additional reconnaissance detrital and bedrock low-temperature thermochronologic data from southwestern Montana suggest that basement-involved exhumation in southwestern Montana may have occurred as early as 100 Ma (Carrapa et al., 2019; Garber et al., 2020); however, these data have alternatively been interpreted to reflect development of a broken flexural forebulge that predates basement-involved thrusting in the region (DeCelles, 2004). A multi-thermochronometric investigation by Ronemus et al. (2023) records a ~90 Ma onset of exhumation in the western Beartooth Mountains and ~60 Ma in the eastern Beartooth Mountains. Considering these inconsistent interpretations, in particular for the timing of onset of thick-skinned thrusting in the region, more data are needed—in a structural context—to help evaluate the timing and spatial patterns of thrust-related exhumation within thin- and thick-skinned thrusts of the Idaho-Montana fold-thrust belt.

2.4. Post-Thrusting Tectonism

In addition to rock cooling that is interpreted to have resulted from exhumation during active Cordilleran orogenesis, central Idaho and southwestern Montana were affected by multiple magmatic and post-Sevier and Laramide deformation events that may have influenced the rock thermal record. Idaho batholith emplacement began to the west and north of our study area between 98 and 87 Ma, and magmatism continued through ~54 Ma in the Bitterroot batholith (Gaschnig et al., 2010, 2011; Porter, 2021). The Challis Volcanic Group, overlying the western half of our transect, was deposited between 51 and 43 Ma (Gaschnig et al., 2011). At the western extent of our study area, the Pioneer metamorphic core complex and associated Pioneer Intrusive Suite (~50–46 Ma) were exhumed during ca. 50–33 Ma, northwest-southeast extension, including along the northwest dipping Wildhorse detachment fault (Vogl et al., 2012). Development of this metamorphic core complex defines the initiation of orogenic collapse in our study area, which was followed by ~16 Ma initiation of northeast-southwest directed extension associated with Basin and Range faulting, which continues to the present (Bennett, 1986; Rodgers et al., 2002).

3. Methods

3.1. Sampling Strategy

We collected samples from the Idaho-Montana fold-thrust belt, spanning from central Idaho to southwestern Montana, with a focus on the hanging walls of major thrusts (Figure 1). Samples of Mesozoic or older igneous or metamorphic rocks—rather than clastic sedimentary rocks—were collected where available (e.g., Hawley Creek pluton, Deep Creek pluton, crystalline basement rocks in basement uplifts in southwestern Montana) to ensure that sample aliquots (mineral grains) share the same pre-thrusting thermal histories. Samples of sedimentary or metasedimentary rocks were collected where crystalline rocks were not available, including from the hanging walls of the Pioneer thrust, Copper Basin thrust, an unnamed thrust system detached in Mississippian rocks in the Lost River Range, the Poison Creek thrust, and the Tendoy thrust. Given the fine-grained nature of these sedimentary samples, apatite grains were either completely absent or of poor quality, precluding AHe dating. Based on a regional sample transect to constrain maximum burial temperatures using Raman spectroscopy of carbonaceous material (Parker et al., 2022), all samples of sedimentary rock that we analyzed for ZHe thermochronometry were heated above 215°C prior to exhumation, with one exception: the Pennsylvanian Quadrant Formation in the hanging wall of the Tendoy thrust likely was not subjected to burial temperatures in excess of ~115°C. A minimum of two samples were collected per thrust sheet, as proximal to the fault surface as possible, but away from any later stage (i.e., Basin and Range) basin-bounding normal faults. For the Hawley Creek, Tendoy, and Blacktail-Snowcrest thrust hanging walls, sufficient relief, exposure, and access was available to collect samples in sub-vertical transects.

3.2. Thermochronology Analytical Methods

Whole rock samples were separated for apatite and zircon crystals at Idaho State University and the University of Iowa using standard mineral separation processes. Samples were pulverized with a jaw crusher and disc mill before undergoing gravity separation with a Wilfley water table, magnetic separation with a Frantz separator, and density separation with heavy liquids. For (U-Th)/He analysis, a minimum of four single grain zircon and apatite aliquots were selected from each sample. In areas where crystalline basement is exposed in the hanging walls of thick-skinned thrusts, zircon grains were picked for a representative range of apparent radiation damage following the methods of Ault et al. (2018).

We conducted AHe and ZHe dating at the University of Illinois at Urbana-Champaign and the University of Arizona Radiogenic Helium Dating Laboratory; one sample (83DTB19) was analyzed at the University of Colorado Boulder Thermochronology Research and Instrumentation Laboratory. Aliquot degassing followed standard procedures described by Guenther et al. (2016): single aliquot-bearing Nb tubes underwent laser heating with a diode or CO₂ laser, followed by cryogenic purification, and measurement of ⁴He concentrations with isotope dilution techniques (³He spike) using a quadrupole mass-spectrometer. After dissolution, U, Th, and Sm were measured via isotope dilution, inductively coupled plasma mass-spectrometry on an Element2 at the University of Arizona or University of Colorado Boulder, or an iCAP Q at the University of Illinois. Zircon and apatite (U-Th)/He analytical data are reported in Tables 1 and 2, respectively.

Table 1
Zircon (U-Th)/He Data

Sample name	Lat.	Long.	Grain size (μm , equiv. Sphere rad.)	U (ng)	Analyt. $\pm (2\sigma)$	Th (ng)	Analyt. $\pm (2\sigma)$	^4He (pmol)	Analyt. $\pm (2\sigma)$	Ft	U (ppm) ^a	Th (ppm) ^a	Effective U (ppm) ^a	Corr. Age (Ma)	Analyt. $\pm (2\sigma)$
Pioneer Thrust															
18JK-01-2	N43°51.635'	W114°16.960'	29	0.1107	0.0026	0.0678	0.0035	0.0126	0.0006	0.60	107	65	122	30.5	1.6
18JK-01-3			33	0.1707	0.0041	0.0426	0.0026	0.0187	0.0011	0.65	124	31	131	29.7	1.8
18JK-02-1	N43°51.207'	W114°16.298'	56	0.5073	0.0125	0.2010	0.0059	0.0810	0.0024	0.78	76	30	83	34.7	1.3
18JK-02-2			56	0.6218	0.0172	0.4359	0.0200	0.1165	0.0031	0.78	88	62	102	38.3	1.4
18JK-02-3			67	1.7902	0.0636	1.1436	0.0334	0.3101	0.0080	0.81	149	95	172	34.2	1.4
18JK-02-4			73	1.2609	0.0292	1.1808	0.0274	0.2788	0.0073	0.83	78	73	95	40.5	1.3
14DW03Z_1	N43°39.835'	W114°11.297'	65	1.2824	0.0371	0.5729	0.0165	0.2782	0.0039	0.81	141	63	156	44.9	1.3
14DW03Z_2			54	1.7239	0.0497	0.5125	0.0151	0.3734	0.0052	0.77	305	91	327	48.5	1.4
14DW03Z_3			78	2.1785	0.0629	0.7910	0.0228	0.5742	0.0196	0.84	112	41	121	53.5	2.3
14DW03Z_4			71	0.9430	0.0273	0.7147	0.0205	0.1956	0.0067	0.83	66	50	77	39.6	1.7
14DW03Z_5			72	1.0688	0.0310	1.2534	0.0361	0.2968	0.0103	0.83	72	84	92	48.9	2.0
14DW03Z_6			75	2.8310	0.0827	1.0041	0.0294	0.5672	0.0194	0.84	216	76	234	41.1	1.8
Copper Basin Thrust															
16DP101_1	N43°34.999'	W113°40.451'	63	0.6884	0.0177	0.4322	0.0177	0.1483	0.0013	0.80	58	37	67	43.3	1.0
16DP101_2			73	1.1279	0.0273	0.7775	0.0273	0.4241	0.0029	0.83	68	47	79	72.2	1.6
16DP101_3			56	0.2444	0.0063	0.2203	0.0063	0.0461	0.0007	0.78	36	32	43	37.1	1.0
16DP101_4			54	0.5104	0.0139	0.2082	0.0139	0.0874	0.0014	0.77	82	33	89	37.5	1.1
16DP100_1	N43°38.844'	W113°46.209'	79	1.0598	0.0284	0.7355	0.0284	0.2323	0.0021	0.84	42	29	49	41.5	1.0
16DP100_2			66	1.3055	0.0320	1.2516	0.0320	0.2752	0.0033	0.81	112	108	138	39.4	0.9
16DP100_3			62	0.3008	0.0074	0.2142	0.0074	0.0788	0.0012	0.80	28	20	33	52.0	1.3
16DP100_4			54	0.9339	0.0226	0.6426	0.0226	0.2929	0.0041	0.77	133	92	155	64.6	1.6
16DP100_5			52	0.5826	0.0148	0.6261	0.0148	0.1093	0.0016	0.76	93	100	116	36.4	0.9
Lost River Range															
14DW02Z_1	N44°14.712'	W113°48.399'	45	0.6309	0.0184	0.3512	0.0101	0.3364	0.0124	0.73	165	92	187	119	5
14DW02Z_2			45	0.3647	0.0106	0.2025	0.0058	0.3276	0.0122	0.74	116	64	131	198	9
14DW02Z_3			59	1.8399	0.0531	0.8803	0.0253	0.9437	0.0366	0.79	212	101	236	107	5
14DW02Z_4			44	0.2937	0.0086	0.3869	0.0111	0.1509	0.0060	0.73	82	107	107	100	5
14DW02Z_5			59	0.9036	0.0263	0.5518	0.0159	0.6258	0.0245	0.80	109	67	125	140	7
14DW02Z_6			36	0.3551	0.0103	0.2400	0.0071	0.1929	0.0027	0.68	256	173	297	129	4

Table 1
Continued

Sample name	Lat.	Long.	Grain size (μm , equiv. Sphere rad.)	U (ng)	Analyt. $\pm (2\sigma)$	Th (ng)	Analyt. $\pm (2\sigma)$	^4He (pmol)	Analyt. $\pm (2\sigma)$	Ft	U (ppm) ^a	Th (ppm) ^a	Effective U (ppm) ^a	Corr. Age (Ma)	Analyt. $\pm (2\sigma)$
Iron Lake Thrust															
13IL02_Z1	N44°54.442'	W114°11.756'	37	0.2898	0.0084	0.0686	0.0021	0.0821	0.0022	0.68	193	46	203	72.9	2.8
13IL02_Z2			57	6.6606	0.1926	0.4363	0.0131	1.9745	0.0477	0.79	1,010	66	1,025	68.5	2.5
13IL02_Z3			71	4.6643	0.1338	1.7915	0.0509	1.8785	0.0455	0.82	432	166	471	82.7	2.9
13IL02_Z4			44	1.3533	0.0393	0.4434	0.0128	0.5600	0.0085	0.73	387	127	417	97.4	3.0
13IL02_Z5			62	1.5897	0.0457	0.6091	0.0174	0.8112	0.0072	0.80	291	111	317	108	3
Hawley Creek Thrust															
18JK04_1	N44°38.982'	W113°10.104'	56	1.7626	0.0536	1.8465	0.0589	0.7333	0.0082	0.78	264	277	329	79.3	2.1
18JK04_2			44	0.8363	0.0248	0.4898	0.0148	0.3210	0.0049	0.72	252	147	286	86.1	2.6
18JK04_3			36	0.8308	0.0259	1.3453	0.0414	0.3187	0.0042	0.66	413	669	571	77.5	2.1
18JK04_4			61	4.3488	0.1303	2.9463	0.0881	1.8431	0.0295	0.79	430	292	499	84.9	2.6
18JK04_5			52	1.2019	0.0366	1.0120	0.0313	0.4528	0.0092	0.76	218	184	261	75.9	2.5
elevation: 2,867 m															
18JK05_1	N44°38.804'	W113°10.343'	38	1.5361	0.0471	1.1962	0.0358	0.3801	0.0066	0.69	586	456	693	56.4	1.8
18JK05_2			47	2.0098	0.0580	1.4082	0.0630	0.9064	0.0088	0.74	387	271	451	96.5	2.6
18JK05_3			52	12.5960	0.3635	11.2415	0.3905	3.5969	0.0352	0.76	1,804	1,610	2,182	57.3	1.5
18JK05_4			47	5.8685	0.1697	6.8554	0.1967	1.9049	0.0171	0.74	1,373	1,604	1,750	63.6	1.6
elevation: 2,609 m															
18JK06_1	N44°38.944'	W113°10.853'	41	1.9594	0.0569	0.9281	0.0267	0.7071	0.0079	0.71	530	251	589	84.6	2.4
18JK06_2			39	0.8052	0.0234	2.1143	0.0606	0.4303	0.0051	0.68	277	727	448	89.4	2.1
18JK06_3			34	0.5008	0.0147	0.5770	0.0176	0.1925	0.0022	0.65	321	370	408	86.5	2.3
18JK06_4			66	2.4145	0.0697	4.6067	0.1685	1.2312	0.0141	0.81	200	381	290	80.4	2.0
elevation: 2,515 m															
18JK07_1	N44°38.626'	W113°10.886'	63	3.4199	0.0987	4.4021	0.1264	0.3074	0.0047	0.80	359	462	467	16.0	0.4
18JK07_2			32	0.3229	0.0096	0.4464	0.0132	0.1144	0.0026	0.63	252	348	334	78.5	2.6
18JK07_4			38	0.5652	0.0164	0.6877	0.0206	0.5809	0.0068	0.69	252	306	324	214	6
18JK07_5			38	0.5370	0.0155	1.0906	0.0321	0.4022	0.0048	0.68	210	427	310	137	3

Table 1
Continued

Sample name	Lat.	Long.	Grain size (μm , equiv. Sphere rad.)	U (ng)	Analyt. $\pm (2\sigma)$	Th (ng)	Analyt. $\pm (2\sigma)$	^4He (pmol)	Analyt. $\pm (2\sigma)$	Ft	U (ppm) ^a	Th (ppm) ^a	Effective U (ppm) ^a	Corr. Age (Ma)	Analyt. $\pm (2\sigma)$
elevation: 2,356 m															
Poison Creek Thrust															
CMH13_02_Z1	N44°53.617'	W113°51.751'	44	1.8943	0.0545	0.3498	0.0103	0.5456	0.0104	0.73	542	100	566	70.1	2.3
CMH13_02_Z2			45	2.5070	0.0723	0.8979	0.0263	0.7909	0.0184	0.73	789	283	855	73.6	2.6
CMH13_02_Z3			39	1.4825	0.0432	0.3199	0.0093	0.4549	0.0104	0.70	974	210	1,024	77.5	2.8
CMH13_02_Z4			37	0.4088	0.0118	0.1246	0.0037	0.1087	0.0030	0.68	249	76	266	67.7	2.6
CMH13_02_Z5			63	0.8560	0.0247	0.3220	0.0094	0.2577	0.0040	0.80	112	42	122	63.6	1.9
CMH13_02_Z6			39	0.7620	0.0221	0.7497	0.0214	0.2297	0.0034	0.69	311	306	383	65.5	1.8
ILBK12_Z1	45°7.570'N	114°12.852'W	93	2.8899	0.0830	0.8631	0.0246	0.8474	0.0034	0.86	77	23	82	58.6	1.6
ILBK12_Z2			84	2.5267	0.0731	0.7085	0.0205	0.6542	0.0039	0.85	99	28	105	52.8	1.5
ILBK12_Z3			65	0.5574	0.0160	0.2322	0.0067	0.1671	0.0012	0.81	56	23	61	62.4	1.8
ILBK12_Z4			88	2.1092	0.0608	0.8838	0.0253	0.6687	0.0048	0.86	72	30	79	62.2	1.8
ILBK12_Z5			59	0.9138	0.0263	0.4368	0.0124	0.2284	0.0010	0.79	77	37	85	52.5	1.4
Diamond Creek/Brushy Gulch															
83DTB19_z01	N45°23.577'	W114°07.963'	40	2.2486	0.1503	0.1444	0.0043	0.2381	0.0032	0.72	994	64	1,009	60.3	4.0
83DTB19_z02			41	3.6908	0.1599	0.4417	0.1841	0.2085	0.0023	0.73	933	112	960	55.1	2.4
83DTB19_z03			43	1.4191	0.0577	0.1875	0.0101	0.1123	0.0011	0.74	491	65	507	55.4	2.2
83DTB19_z04			45	1.8061	0.0676	0.3204	0.0153	0.1412	0.0015	0.75	571	101	595	58.3	2.1
83DTB19_z05			56	17.2585	0.5451	1.4664	0.9562	0.6257	0.0045	0.80	2,351	200	2,398	60.3	2.0
83DTB19_z06			47	1.2291	0.0364	0.1169	0.0050	0.0871	0.0008	0.76	393	37	402	52.6	1.6
83DTB19_z07			48	1.5630	0.0430	0.2998	0.0809	0.1035	0.0008	0.76	413	79	432	57.9	1.7
83DTB19_z08			40	3.4805	0.2331	0.1318	0.0046	0.3560	0.0035	0.72	1,466	56	1,479	61.3	4.0
Cabin Thrust															
19CT02_1	N44°39.671'	W112°58.293'	54	6.8687	0.1691	1.0365	0.3382	1.8555	0.0150	0.77	882	133	914	62.3	1.5
19CT02_2			59	4.2945	0.1030	1.1326	0.2061	1.3028	0.0127	0.79	463	122	492	66.6	1.7
19CT02_3			70	18.7883	0.4749	2.1681	0.9498	5.7460	0.0499	0.82	1,264	146	1,298	66.8	1.7
19CT02_4			78	7.9530	0.1960	1.6473	0.3919	2.3142	0.0242	0.84	369	76	387	61.1	1.5
19CT02_6			52	5.5707	0.1362	0.9909	0.2723	1.4968	0.0266	0.77	865	154	901	62.3	1.8
19CT02_5			67	12.6055	0.3109	2.1951	0.6217	3.6514	0.0266	0.82	716	125	745	63.0	1.5

Table 1
Continued

Sample name	Lat.	Long.	Grain size (μm , equiv. Sphere rad.)	U (ng)	Analylt. $\pm$ (2 σ)	Th (ng)	Analylt. $\pm$ (2 σ)	^4He (pmol)	Analylt. $\pm$ (2 σ)	Ft	U (ppm) ^a	Th (ppm) ^a	Effective U (ppm) ^a	Corr. Age (Ma)	Analylt. $\pm$ (2 σ)
Monument Fault															
18NA16_z2	N45°0.600'	W113°20.802'	69	16.4501	0.4736	2.2943	0.0659	5.9366	0.1688	0.82	1,236	172	1,277	78.5	3.1
18NA16_z3			76	14.7469	0.4232	2.5082	0.0716	5.5415	0.3425	0.84	829	141	863	79.6	5.4
18NA16_z4			58	11.7750	0.3382	2.2177	0.0635	3.9932	0.2467	0.79	1,235	233	1,289	76.0	5.1
18NA16_z6			71	10.2055	0.2949	2.2165	0.0632	3.2970	0.0230	0.83	693	150	728	68.7	1.9
Tendoy Thrust															
19TT01_1	N44°39.859'	W112°46.818'	49	2.3434	0.0629	0.7733	0.0204	3.1495	0.1034	0.76	398	131	429	300	13
19TT01_2			36	0.3058	0.0075	0.1296	0.0052	0.3306	0.0110	0.68	142	60	156	265	11
19TT01_3			50	2.0358	0.0478	0.6992	0.0182	0.5570	0.0049	0.76	340	117	367	61.9	1.4
19TT01_4			50	1.4442	0.0354	0.8250	0.0212	2.4182	0.0161	0.76	281	161	319	352	8
19TT01_5			60	1.7966	0.0457	0.9768	0.0256	3.1519	0.0557	0.80	189	102	213	353	10
19TT02_1	N44°39.948'	W112°46.754'	47	0.5549	0.0130	0.1844	0.0063	0.7040	0.0060	0.74	130	43	141	288	7
19TT02_2			51	0.7814	0.0191	0.3228	0.0089	1.1023	0.0088	0.77	135	56	148	305	7
19TT02_3			50	0.6408	0.0155	0.3740	0.0107	1.0493	0.0085	0.76	113	66	128	343	8
19TT02_4			44	0.4346	0.0108	0.3409	0.0318	0.6943	0.0052	0.73	110	86	131	337	9
19TT02_5			55	1.1547	0.0271	0.6567	0.0177	2.1679	0.0206	0.78	167	95	189	384	9
19TT03_1	N44°40.040'	W112°46.474'	40	1.0293	0.0265	0.5245	0.0144	1.2583	0.0096	0.71	382	195	427	282	7
19TT03_2			42	0.7349	0.0179	0.3715	0.0109	1.1305	0.0067	0.72	223	113	250	348	8
19TT03_4			57	0.1523	0.0040	0.1730	0.0058	0.5724	0.0042	0.79	18	20	23	667	16
19TT03_5			56	1.4044	0.0348	0.4489	0.0124	2.0274	0.0219	0.78	179	57	192	311	8
19TT03_6			41	0.3037	0.0075	0.1182	0.0231	0.3795	0.0036	0.71	92	36	100	294	9
19TT04_1	N44°39.529'	W112°46.277'	54	1.1035	0.0267	0.5473	0.0154	1.6953	0.0100	0.77	171	85	191	322	7
19TT04_2			62	1.6094	0.0413	1.2811	0.0326	3.4399	0.0290	0.80	147	117	174	405	10
19TT04_3			56	1.7932	0.0456	0.4130	0.0111	3.1011	0.0232	0.79	239	55	252	376	10
19TT04_4			47	0.6051	0.0156	0.3010	0.0087	0.7372	0.0038	0.75	126	63	141	267	6
19TT04_5			44	1.0972	0.0261	0.2552	0.0082	1.3706	0.0139	0.73	274	64	289	294	7

Table 1
Continued

Sample name	Lat.	Long.	Grain size (μm , equiv. Sphere rad.)	U (ng)	Analyt. $\pm (2\sigma)$	Th (ng)	Analyt. $\pm (2\sigma)$	^4He (pmol)	Analyt. $\pm (2\sigma)$	Ft	U (ppm) ^a	Th (ppm) ^a	Effective U (ppm) ^a	Corr. Age (Ma)	Analyt. $\pm (2\sigma)$
Blacktail-Snowcrest															
17BC- VT1-ZR1	N44°55.495'	W112°26.673'	45	2.3899	0.0416	0.3389	0.0317	0.6448	0.0089	0.73	618	88	639	66.0	1.4
17BC- VT1-ZR2			41	0.6250	0.0124	0.2336	0.0441	0.5729	0.0082	0.71	218	81	237	218	6
17BC- VT1-ZR3			61	2.4025	0.0543	0.4921	0.1829	0.9593	0.0138	0.80	253	52	265	87.9	2.7
17BC- VT1-ZR4			41	1.8912	0.0423	0.1780	0.0087	0.5685	0.0084	0.71	573	54	585	76.3	2.0
17BC- VT1-ZR5			51	1.9121	0.0463	0.5398	0.0687	0.6685	0.0043	0.76	310	88	331	79.7	1.9
elevation: 2,680 m															
17BC- VT2B-ZR1	N44°55.810'	W112°26.342'	47	2.2733	0.0470	3.5051	0.0640	0.7144	0.0070	0.74	499	770	680	58.0	1.1
17BC- VT2B-ZR2			39	2.1047	0.0394	0.2043	0.0158	0.1914	0.0015	0.70	697	68	713	23.6	0.5
17BC- VT2B-ZR3			55	3.8806	0.0783	6.0826	0.6813	1.2195	0.0681	0.78	513	804	702	55.1	3.6
17BC- VT2B-ZR4			81	13.7784	0.3688	165.6616	21.2011	4.4202	0.2445	0.85	600	7,219	2,297	18.6	2.0
17BC- VT2B-ZR5			62	6.8726	0.1630	0.9953	0.0266	1.6152	0.0895	0.80	680	99	703	52.5	3.1
elevation: 2,450 m															
17BC- VT4B-ZR3	N44°56.295'	W112°26.246'	61	1.1378	0.0460	0.3030	0.0304	1.5308	0.0208	0.80	120	32	127	287	12
17BC- VT4B-ZR4			41	2.4606	0.0998	0.7376	0.0503	3.3453	0.0464	0.71	745	223	798	323	13
elevation: 2,098 m															

^aEffective U calculated from grain volume as determined through morphology measurements.

Table 2
Apatite (U-Th)/He Data

Sample name	Lat.	Long.	Grain size (μg , equiv. Sphere rad.)	U (ng)	Analylt. $\pm$ (2 σ)l	Th (ng)	Analylt. $\pm$ (2 σ)l	⁴ He (pmol)	Analylt. $\pm$ (2 σ)l	Ft	U (ppm) ^a	Th (ppm) ^a	Effective U (ppm) ^a	Corr. Age (Ma)	Analylt. $\pm$ (2 σ)l
Blacktail-Snowcrest															
17BC-VT1-ap1	N44°55.495'	W112°26.673'	80	0.3303	0.0163	0.0609	0.0080	0.088	0.0016	0.82	22	4	23	56.9	2.8
17BC-VT1-ap2			56	0.5245	0.0275	0.0508	0.0091	0.133	0.0029	0.74	102	10	105	61.2	3.3
17BC-VT1-ap3			54	0.2153	0.0140	0.0646	0.0082	0.058	0.0015	0.73	50	15	54	63.1	4.1
17BC-VT1-ap4			41	0.0337	0.0017	0.0747	0.0144	0.013	0.0004	0.65	17	37	25	70.1	5.4
17BC-VT1-ap5			47	0.0943	0.0075	0.0619	0.0122	0.045	0.0011	0.70	30	20	34	106	8
elevation: 2,680 m															
17BC-VT2B-ap1	N44°55.810'	W112°26.342'	52	0.6580	0.0194	0.3493	0.0231	0.129	0.0025	0.72	174	92	196	44.6	1.5
17BC-VT2B-ap2			88	1.9932	0.0748	0.3934	0.0187	0.634	0.0115	0.83	111	22	116	67.3	2.6
17BC-VT2B-ap3			43	0.1935	0.0063	0.0535	0.0064	0.016	0.0004	0.68	81	23	87	21.2	0.8
17BC-VT2B-ap4			72	0.6813	0.0213	0.1709	0.0112	0.043	0.0006	0.80	65	16	69	13.8	0.4
17BC-VT2B-ap5			58	0.5918	0.0174	0.0542	0.0072	0.092	0.0012	0.75	114	10	117	37.2	1.1
elevation: 2,450 m															
17BC-VT4B-ap1	N44°56.295'	W112°26.246'	66	0.1583	0.0052	0.2130	0.0118	0.062	0.0014	0.78	21	28	28	69.1	2.4
17BC-VT4B-ap2			71	0.4150	0.0140	0.1864	0.0096	0.147	0.0018	0.79	45	20	50	72.9	2.3
17BC-VT4B-ap3			60	0.3095	0.0136	0.1793	0.0081	0.099	0.0016	0.76	53	31	60	67.7	2.8
17BC-VT4B-ap4			58	0.3808	0.0225	0.2973	0.0109	0.105	0.0015	0.75	70	54	82	57.3	2.9
17BC-VT4B-ap5			93	1.6084	0.0718	3.3946	0.1276	0.685	0.0071	0.84	76	161	114	61.5	2.5
elevation: 2,098 m															
16BC-VT1-ap1	N44°58.280'	W112°32.801'	59	0.0623	0.0018	0.0816	0.0024	0.052	0.0016	0.75	12	16	16	153	6
16BC-VT1-ap2			42	0.0597	0.0017	0.0174	0.0005	0.009	0.0003	0.67	28	8	30	40.3	1.7
16BC-VT1-ap3			54	0.2730	0.0079	0.0129	0.0005	0.035	0.0008	0.73	58	3	58	31.4	1.2
16BC-VT1-ap4			41	0.0835	0.0024	0.0116	0.0003	0.008	0.0002	0.66	44	6	46	25.3	1.0
16BC-VT1-ap5			77	0.5217	0.0150	0.1390	0.0040	0.137	0.0032	0.81	43	12	46	56.2	2.0
16BC-VT1-ap6			79	0.6244	0.0179	0.0238	0.0007	0.074	0.0017	0.81	45	2	46	26.5	1.0
16BC-VT2-ap1	N44°58.515'	W112°32.538'	60	0.1942	0.0056	0.0016	0.0001	0.034	0.0008	0.76	31	0	31	42.1	1.5
16BC-VT2-ap2			60	0.0413	0.0012	0.0011	0.0001	0.005	0.0002	0.76	9	0	9	30.7	1.5
16BC-VT2-ap4			67	0.0480	0.0014	0.0116	0.0004	0.012	0.0004	0.79	6	1	6	54.5	2.3
16BC-VT2-ap5			44	0.0144	0.0004	0.0022	0.0001	0.002	0.0001	0.69	5	1	5	30.1	2.2
16BC-VT2-ap6			67	0.0662	0.0019	0.0090	0.0003	0.010	0.0003	0.79	7	1	8	32.8	1.3

Table 2
Continued

Sample name	Lat	Long.	Grain size (μg , equiv. Sphere rad.)	U (ng)	Analyt. $\pm$ (2 σ)1	Th (ng)	Analyt. $\pm$ (2 σ)1	^4He (pmol)	Analyt. $\pm$ (2 σ)1	Ft	U (ppm) ^a	Th (ppm) ^a	Effective U (ppm) ^a	Corr. Age (Ma)	Analyt. $\pm$ (2 σ)1
16BC-VT3-ap1	N44°58.775'	W112°31.622'	45	0.0677	0.0020	0.0635	0.0019	0.005	0.0002	0.68	23	22	28	16.9	0.7
16BC-VT3-ap2			54	0.1457	0.0042	0.0820	0.0023	0.017	0.0004	0.73	36	20	41	25.6	0.9
16BC-VT3-ap3			45	0.0391	0.0011	0.0190	0.0006	0.002	0.0001	0.69	15	7	16	14.8	0.7
16BC-VT3-ap4			44	0.0224	0.0006	0.0125	0.0004	0.003	0.0001	0.68	11	6	12	32.0	1.5
16BC-VT3-ap5			48	0.1493	0.0043	0.0394	0.0011	0.028	0.0006	0.70	42	11	45	46.4	1.5
16BC-VT3-ap6			65	0.1526	0.0044	0.0210	0.0006	0.019	0.0004	0.78	21	3	21	28.8	1.0
16BC-VT4-ap1	N44°59.385'	W112°31.706'	80	1.1977	0.0343	0.4840	0.0138	0.399	0.0068	0.81	84	34	92	68.8	2.1
16BC-VT4-ap3			94	1.4346	0.0411	0.5529	0.0158	0.318	0.0037	0.84	63	24	69	44.5	1.2
16BC-VT4-ap4			61	0.6118	0.0176	0.1330	0.0038	0.153	0.0018	0.76	88	19	92	57.7	1.7
16BC-VT4-ap5			64	1.2819	0.0367	0.1398	0.0040	0.256	0.0029	0.77	148	16	152	46.5	1.4
16BC-VT4-ap6			63	0.5144	0.0148	0.1061	0.0031	0.248	0.0028	0.77	71	15	75	110	3
16BC-VT5-ap1	N44°59.788'	W112°31.393'	59	0.5576	0.0159	0.2229	0.0064	0.162	0.0019	0.75	101	40	111	65.0	1.8
16BC-VT5-ap2			46	0.0448	0.0013	0.0150	0.0005	0.008	0.0002	0.69	16	5	17	42.5	1.5
16BC-VT5-ap3			50	0.2404	0.0069	0.0561	0.0016	0.043	0.0007	0.71	62	14	65	44.5	1.4
16BC-VT5-ap4			53	0.4907	0.0141	0.0160	0.0005	0.136	0.0017	0.73	98	3	99	69.5	2.1

^aEffective U calculated from grain volume as determined through morphology measurements.

The external detector method (Hurford & Green, 1983) was used to obtain AFT data from six samples collected from five crystalline basement rocks in the foreland uplift portion of our transect, and one crystalline basement rock in the hanging wall of the Cabin Thrust. Sample separates were mounted in epoxy and irradiated at Oregon State University. Samples were etched in 5.5 M nitric acid at 21°C for 20 s following the protocols of Donelick et al. (2005). Following irradiation, the mica external detectors were etched in 40% hydrofluoric acid for 45 min at 21°C (Donelick et al., 2005). Samples were analyzed with a Leica DM2700P microscope with a drawing tube located above a digitizing tablet and a Ludl computer-controlled stage driven by the FTStage Program (Dumitru, 1993). Analyses were performed with reflected and transmitted light at 1,250X magnification at the Montana State University Tectonic Sedimentology and Thermochronology (TeST) Laboratory. Twenty individual apatite crystals per sample were analyzed. Track density in many samples precluded accurate length measurements. In the few samples where track lengths could be measured, fewer than 10 tracks permitted accurate length measurements, which is too few to be statistically robust. Summary AFT analytical data are reported in Table 3, and individual counting data from each sample are reported in Tables S1–S6 in Supporting Information S1.

4. Results

4.1. Dating Results

We present 106 new ZHe dates (Table 1), 43 new AHe dates (Table 2), and 6 new AFT data sets (Table 3, Tables S1–S6 in Supporting Information S1). Sample locations, indicated in Figure 1, are divided into the Western Thrust group, the Lemhi Arch group, the Frontal Thrust group, and the Foreland uplifts group, according to their position along a SW–NE transect (see Figure 1). Thus, these groupings follow the predominant structural style outlined in the introduction: thin-skinned thrusting involving sedimentary units (Western Thrust group and Frontal Thrust group), thick-skinned thrusting involving crystalline basement or Mesoproterozoic quartzite units (Lemhi Arch group and Foreland Uplifts group), and a “transitional” grouping where these two styles show close spatial overlap (Frontal Thrust group and structurally deeper Lemhi Arch group). At the western extent of our study area, we collected samples from the Pioneer thrust sheet, the Copper Basin thrust sheet, and from the hanging wall of an unnamed thrust in the Lost River Range (Figure 3). In this figure, and subsequent figures, we plot most of the data against effective uranium (eU) concentration, which is proportional to the degree of radiation damage in a crystal. Radiation damage plays a fundamental role in He diffusivity and can potentially explain date dispersion in a data set, provided the He date and eU are correlated. We use eU concentration here as a metric to assess the amount and potential sources of date dispersion in each data set. For samples collected from sub-vertical transects, we show dates plotted against elevation. Figure 3a shows 2 zircon crystals from one sample of Ordovician Trail Creek Formation (sample 18JK01) and 4 zircon crystals from one sample from the Ordovician–Silurian Phi Kappa Formation (sample 18JK02); both samples crop out in the hanging wall of the Pioneer thrust sheet. Individual crystal dates range from 29.7 ± 1.8 Ma (2 σ analytical uncertainties, here and henceforth) to 40.5 ± 1.3 Ma. Another sample (sample 14DW03) was collected from the Devonian Milligen Formation inferred to be in the hanging wall of the Pioneer thrust on the western side of the Pioneer metamorphic core complex (Dover, 1983). Six zircon crystals from this sample had dates ranging from 39.6 ± 1.7 to 53.5 ± 2.3 Ma. Two samples (samples 16DP100 and 16DP101) were also collected from clastic rocks of the Mississippian Copper Basin Group carried in the hanging wall of the Copper Basin thrust; nine zircon crystals were dated from these samples. Copper Basin thrust sheet grains had dates ranging from 36.4 ± 0.9 to 72.2 ± 1.6 Ma. An additional sample (sample 14DW02) was collected from sandstone of the Mississippian Bluebird Mountain Formation in the hanging wall of an unnamed thrust system detached in Mississippian rocks (Anastasio et al., 1997) in the Lost River Range. Six zircon crystals from this sample had dates ranging from 100 ± 4.6 to 198 ± 9.1 Ma. The Pioneer and Copper Basin samples have relatively flat date–eU trends, and all of the grains from these two thrusts are significantly younger than the Lost River Range samples (Figure 3a).

Multiple samples were collected from the five thrust hanging walls that deformed mechanical basement rocks below the basal Lemhi arch unconformity. Figure 3b shows data from the Iron Lake thrust, Poison Creek thrust, the Diamond Creek/Brushy Gulch fault system, the Cabin thrust, and the Monument thrust. Figure 3c shows the sub-vertical transect data from the Hawley Creek thrust. A total of five zircon crystals were dated from one sample (13IL02) of Mesoproterozoic Swauger Formation (Burmester et al., 2023) in the hanging wall of the Iron Lake thrust. These crystals yielded single grain dates ranging between 68.5 ± 2.5 to 108 ± 2.9 Ma. Two samples were collected from the hanging wall of the Poison Creek thrust. The first (CMH13-02) was sampled from the Mesoproterozoic Apple Creek Formation (Hansen & Pearson, 2016) and produced six ZHe dates ranging from

Table 3
Data Summary of Apatite Fission Track Results

Sample ID	Latitude	Longitude	Lithology	No. Crystals ^a	Rho-S (track/cm ²) ^b	NS ^c	Rho-I (track/cm ²) ^b	NI ^c	P(c) ^d	Rho-D ^e	ND ^f	Central age (Ma)	±1 sigma	Average D _{par} (μm)	U (ppm)
16DP201	N45° 13.166'	W111° 57.733'	Quartzofeldspathic gneiss	20	1.750E+05	350	9.900E+05	1,980	99	2.500E+06	1,250	77.16	5.38	1.92	5.9
16DP203	N45° 13.165'	W112° 2.881'	Quartzofeldspathic gneiss	20	1.990E+05	398	9.145E+05	1,829	99	2.500E+06	1,250	94.85	6.41	1.70	5.5
17DP801	N45° 16.888'	W111° 13.598'	Biotite-plag-k-spar-qtz gneiss	20	7.200E+04	144	6.125E+05	1,225	27	2.500E+06	1,250	52.11	5.32	1.88	3.7
17DP803	N45° 22.419'	W111° 10.522'	Quartzofeldspathic gneiss	20	2.005E+05	401	1.368E+06	2,736	12	2.500E+06	1,250	64.35	4.72	1.86	8.2
19CT01	N44° 39.671'	W112° 58.293'	Quartzofeldspathic gneiss	20	1.015E+05	203	9.210E+05	1,842	87	2.500E+06	1,250	48.21	4.03	1.9	5.5
19BS01	N45° 50.481'	W112° 11.940'	Chloritized K-spar rich gneiss	20	1.185E+05	237	1.137E+06	2,274	82	2.500E+06	1,250	45.60	3.58	1.87	6.8

Note. Refer to Supplementary Material X for complete data tables. DAO zeta 351.33 ± 9.29. ^aNumber of individual crystals dated. ^bRho-S and Rho-I are the spontaneous and induced track density measured, respectively (tracks/cm²). ^cNS and NI are the number of spontaneous and induced tracks counted, respectively. ^d χ^2 (%) is the chi-square probability (Galbraith & Green, 1990; Green 1981). Values greater than 5% are considered to pass this test and represent a single population of ages. ^eRho-D is the induced track density in the external detector adjacent to the CN2 dosimetry glass (tracks/cm²). ^fND is the number of tracks counted in determining Rho-D.

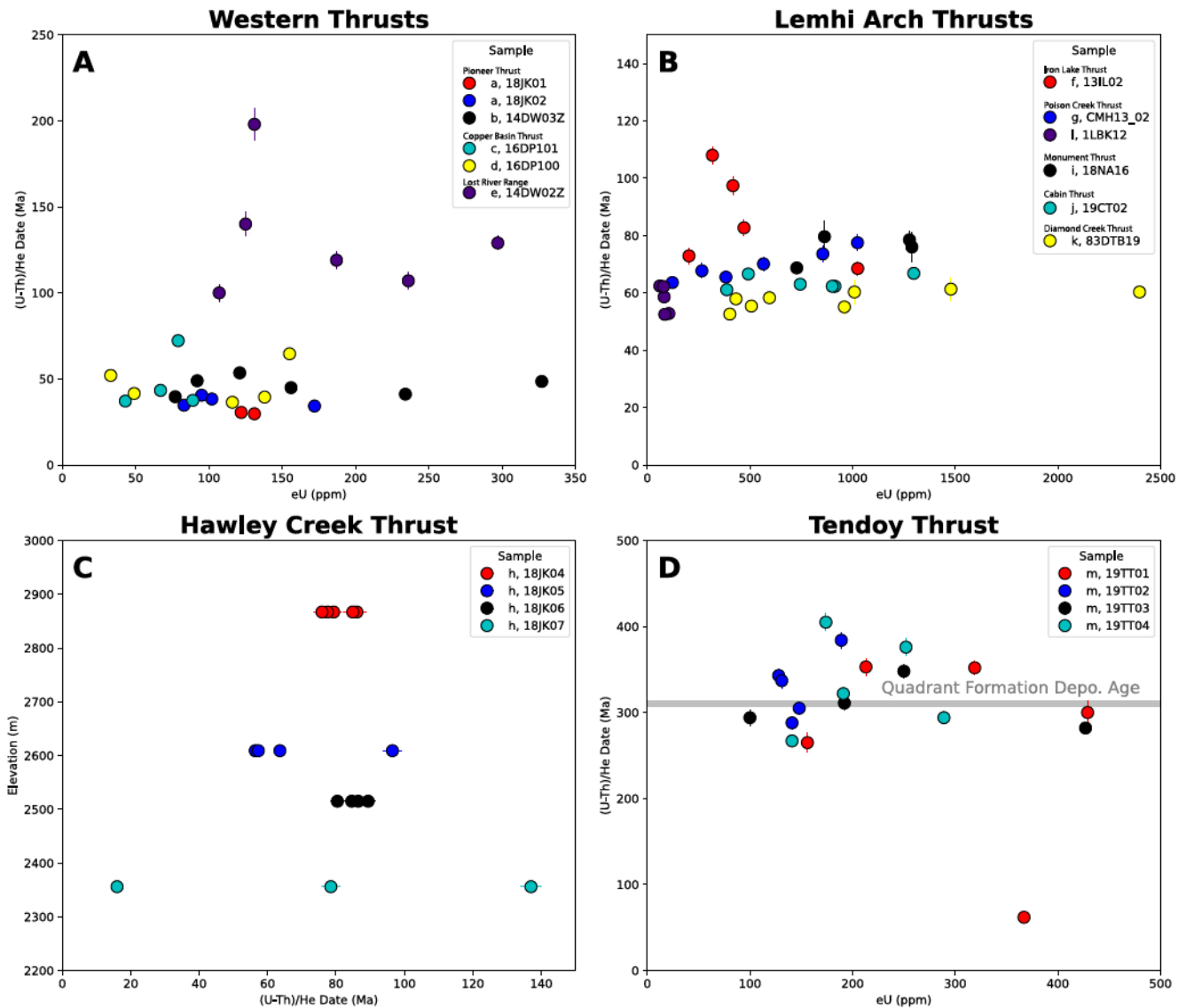


Figure 3. ZHe results for hinterland thrust belt locations. Lower-case letters used in all plot legends match those displayed on the map in Figure 1. (a) ZHe dates plotted against effective uranium (eU) concentration for all samples in the Pioneer thrust (18JK01, 18JK02, and 14DW03Z), Copper Basin thrust (16DP101 and 16DP100), and samples from the Lost River Range (14DW02Z). The Pioneer and Copper Basin data are significantly younger than the Lost River Range data. (b) ZHe dates from the Iron Lake thrust (13IL02), the Poison Creek thrust (CMH13_02 and 1LBK12), the Cabin thrust (19CT02), the Monument thrust (18NA16), and the Diamond Creek/Brushy Gulch thrust (83DTB19), plotted against eU concentration. The mean of all dates from these thrust sheets is ~ 71 Ma, and all possess relatively flat date-eU trends (with the exception of 13IL02). (c) ZHe date-elevation plot for samples from the Hawley Creek sub-vertical transect. The three highest elevation samples cluster around a mean date of 78.3 Ma. The lowest elevation sample has a date spread from 15.9 to 214 Ma. (d) ZHe date plotted against eU concentration for samples from the Tendoy thrust sheet. The data are dispersed, with most dates from this sample clustered around the depositional age of the host unit, the Quadrant Formation. Dates are both older and younger than this depositional age, which suggests only partial resetting of the ZHe system.

63.6 ± 1.9 to 77.5 ± 2.8 Ma. The second sample (1LBK12) was collected ~ 35 km along-strike to the northwest from the Cambrian Deep Creek pluton. This sample produced five ZHe dates ranging from 52.5 ± 1.9 to 62.4 ± 1.8 Ma. Given the relatively long distance between these two samples, we used only one, CMH13-02, in our QTQt inverse modeling. Eight zircon crystals were dated from one sample in the hanging wall of the Diamond Creek fault (Burmester et al., 2020), with dates ranging from 52.6 ± 1.6 to 61.3 ± 4.0 Ma. Four ZHe crystals were dated from one sample in the hanging wall of the Monument fault (Anderson, 2017), with dates ranging from 68.7 ± 1.9 to 79.6 ± 5.4 Ma. Six zircon grains were dated from a sample collected from Archean to Paleoproterozoic crystalline rocks carried in the hanging wall of the Cabin thrust. These dates ranged from 61.0 ± 1.5 to 66.8 ± 1.7 Ma. AFT results from the Cabin thrust hanging wall returned a central age of 48.21 ± 4.0 Ma (Table 3,

sample 19CT01). Collectively, the ZHe dates from our Lemhi Arch group show relatively minimal dispersion (with the slight exception of sample 13IL02), and flat date-eU trends centered at ~65 Ma.

A sub-vertical transect consisting of 4 samples dated with ZHe spaced over ~500 vertical meters was collected from the Ordovician Beaverhead pluton in the hanging wall of the Hawley Creek thrust. The 2,867 m elevation sample has an age range of 75.9 ± 2.5 to 86.1 ± 2.6 Ma, the 2,609 m elevation sample has a date range of 56.4 ± 1.8 to 96.5 ± 2.6 Ma, the 2,515 m elevation sample has a date range of 80.4 ± 2.0 to 89.4 ± 2.1 Ma; these three highest elevation samples have a combined mean age of ~78 Ma. The lowest elevation sample from this transect, collected at an elevation of 2,356 m, has much more dispersed single grain dates that range between 15.9 ± 0.4 and 214 ± 5.6 Ma. When plotted together in date versus elevation space, the three higher elevation samples have a nearly vertical slope (Figure 3c).

Frontal thrust sheets to the east of the Lemhi arch are the toe thrusts of the Idaho-Montana thrust belt. We collected a vertical transect from the Pennsylvanian Quadrant Formation (Saperstone, 1986) in the hanging wall of the Tendoy thrust (Figure 3d). Dates from 20 single grain aliquots across 4 samples, are highly dispersed and range from 61.9 ± 1.4 to 404.7 ± 10.0 Ma, with some dates older than the depositional age of the Quadrant Formation. Given that these grains are not fully reset, we did not include the Tendoy thrust in any of our inverse modeling.

In the Foreland Uplifts group, predominantly in southwestern Montana, samples were collected from the hanging walls of basement-carrying thick-skinned thrusts. Figure 4 shows Blacktail-Snowcrest uplift data in both date-eU (Figures 4a and 4b) and date-elevation (Figures 4c and 4d) space. Samples collected from a sub-vertical transect in the hanging wall of the Snowcrest-Greenhorn fault in the Blacktail-Snowcrest arch yielded 15 single grain AHe dates (samples 17BCVT1, 17BCVT2B, and 17BCVT4B) and 12 single grain ZHe dates (samples 17BCVT1, 17BCVT2B, and 17BCVT4B). An additional five samples were collected from this uplift, and altogether apatite dates range from 13.8 ± 0.4 to 153 ± 5.8 Ma and no correlation between date and eU (Figure 4b). The three sub-vertical transect samples were collected in a vertical elevation ranging from 2,800 to 2,100 m, and display a vertical trend between AHe date and elevation with significant dispersion at the intermediate elevation (Figure 4d). Zircon dates, picked from three samples, range from 18.6 ± 2.0 to 323.1 ± 13.1 Ma, with no trend between date and eU or date and elevation (Figures 4a and 4c). Three samples from the Blacktail-Snowcrest arch returned AFT central ages of 94.9 ± 6.4 , 77.2 ± 5.4 , and 45.6 ± 3.6 Ma (Table 3, samples 16DP203, 16DP201, and 19BS01, respectively). Two samples from the hanging wall of the Spanish Peaks fault in the Northern Madison-Tobacco Root uplift have central AFT ages of 52.1 ± 5.3 and 64.4 ± 4.7 Ma (Table 3, samples 17DP801 and 17DP803, respectively).

5. Thermal History Modeling

5.1. Data Treatment and Thermal History Modeling Methods

Individual thrust sheets and basement uplifts with multiple thermochronometers, or those sampled by vertical transect, were modeled with the program QTQt (Gallagher, 2012). QTQt was chosen because, where independent geologic constraints are limited, as is the case in several of our sample locations (e.g., basement uplifts), QTQt's Bayesian approach allows for an assessment of the relative robustness (or limitation) of the thermochronologic data to constrain timing of cooling. This statistical robustness is conveyed in the form of the posterior distribution of t-T paths, whose color-coded density is directly related to higher t-T path probability, which is crucially based on the thermochronologic data and any known geologic (i.e., stratigraphic constraints, unconformities) constraints. The complexity of the posterior distribution is determined by the model inputs such that fewer inputs yield less resolved solutions, but an overreliance on priors (i.e., too many geologic constraints) dictates the shape of the posterior. Simply put, the extent to which QTQt model output represents a robust solution needs to be assessed within the context of the model inputs used.

The details of the set-up for each individual model are listed in Table S7. In general, we grouped samples from the same thrust sheets into individual models by synthesizing multiple thermochronometers, or samples from vertical transects. In this way we assessed more complex, multi-thermochronometer data sets from single thrust sheets or uplifts in context. Some of the samples used in these groupings were previously reported in Kaempfer et al. (2021). These sample groupings are as follows: (a) Pioneer Thrust: 18JK01, 18JK02; (b) Copper Basin Thrust: 16DP100, 16DP101; (c) Lost River Range: 14DW02Z; (d) Iron Lake Thrust: 13IL02; (e) Hawley Creek

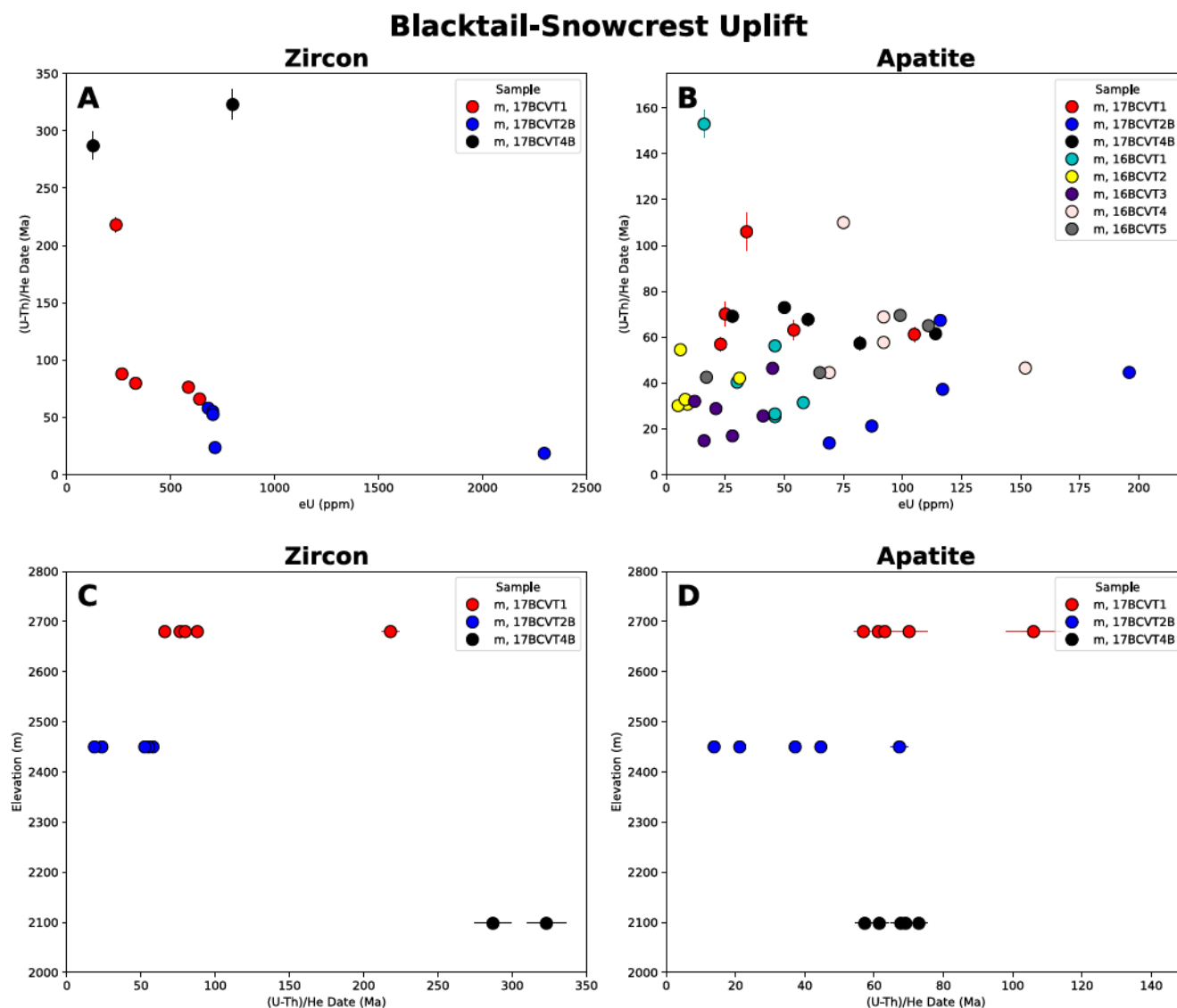


Figure 4. (a) ZHe date and (b) AHe date plotted against effective uranium concentration for all samples from basement rocks in the Blacktail-Snowcrest arch. There is no date-eU correlation in either data set. (c) Date-elevation plot of ZHe and (d) AHe samples from basement rocks in the Blacktail-Snowcrest arch. ZHe dates show significant dispersion, whereas AHe dates are less dispersed and possess a nearly vertical trend. Lower-case letters used in plot legend match those displayed on the map in Figure 1.

Thrust (sub-vertical transect): 18JK04, 18JK05, 18JK06, 18JK07; (f) Poison Creek Thrust: CMH_01; (g) Diamond Creek Thrust: 83DTB19; (h) Cabin Thrust: 19CT02; (i) Monument Fault: 18NA16; (j) West Limb of Blacktail Snowcrest (sub-vertical transect): 17BC-VT1, 17BC-VT2B, 17BC-VT4B; (k) East Limb of Blacktail Snowcrest: 16DP201; (l) Southern Madison Range: 17DP804, 17DP805; and (m) Northern Madison Range: 17DP801, 17DP802, 17DP803. Data from the Tendoy Thrust were not modeled as the ZHe dates were not fully reset following deposition (see ZHe results below). The inverse scenarios for the Pioneer, Lost River, Iron Lake, Diamond Creek, Monument, Poison Creek, and Cabin Thrusts were run with 100,000 burn-in and 100,000 post burn-in paths, whereas the Hawley Creek, West and East Limbs of Blacktail Snowcrest, and Southern and Northern Madison Range sample groupings were run with 750,000 burn-in and 750,000 post burn-in paths (burn-in refers here to initial model iterations that are used to fully explore the general prior, but ultimately discarded from the final distribution). More iterations were used where samples needed a much larger general prior (i.e., the frontal basement uplifts), or had a large number of data inputs (i.e., Hawley Creek vertical transect). The models used all of the individual ZHe and AHe dates for a given sample or set of samples as inputs, as well as the track

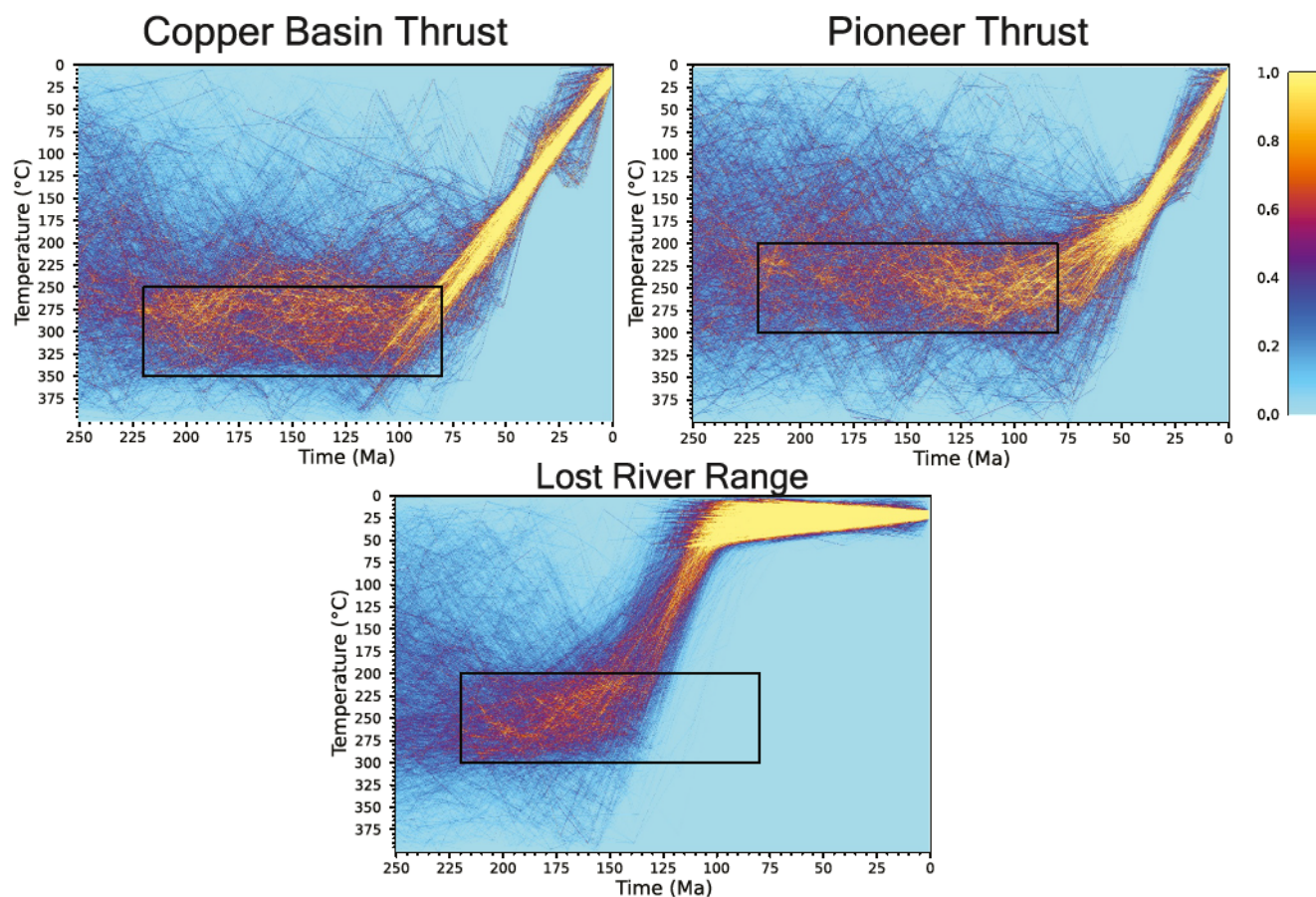


Figure 5. QTQt inverse model output for the western thrust sheets. The posterior distribution is shown for each location in a normalized path density scale. These densities are directly related to probability such that warmer or brighter colors indicate that more paths pass through that region of t-T space. The black outlines indicate constraint boxes, justifications for these constraints can be found in Table S7.

count data for AFT results. We used error resampling on each individual date, and allowed for more complex paths to be accepted given equivalent posterior values.

To the extent that unequivocal geologic priors were available, we included those as well in our modeling, including deposition/emplacement dates and unconformities. The geologic priors for each model varied slightly from location to location, but a complete description of the justification for each one, as well as the relevant citations, are listed in detail in Table S7. For general priors, the foreland basement uplifts were extended back to 1,800 Ma, which is the approximate timing of high-grade metamorphism as measured in regional $^{40}\text{Ar}/^{39}\text{Ar}$ biotite cooling ages (Condit et al., 2015). All other locations used a general prior that extended back to the 550 Ma, with the exception of the Hawley Creek samples where the Ordovician emplacement age of the granitoid was used (Link et al., 2017). Although the non-basement uplift grains are detrital and have an older, Precambrian recycling history, they all experienced temperatures in the Phanerozoic (with the exception of the Tendoy samples) that led to being fully reset. To this point, additional constraint boxes were included in the Phanerozoic to reflect either unconformities observed in the stratigraphic record, or maximum burial temperatures calculated from Raman spectroscopy of carbonaceous material (Parker et al., 2022).

5.2. Thermal History Modeling Results

We present our QTQt inverse model results using the same groupings described in the results section: Western thrust, Lemhi Arch thrusts, and Foreland Uplifts. Summary posterior probability distributions are shown in Figures 5–7. These figures show only the most relevant portions of t-T space for our discussion (250 Ma and younger), although each model was run with a larger general prior (see methods section and Table S7 for full details). Individual model results, which include predicted versus observed date plots and the probability density

plots of the half-maximum isotherms (see below for details), are shown in Figures S1–S13 in Supporting Information S1.

Each of the modeling figures (Figures 5–7) shows variable levels of complexity and resolution on the post-250 Ma thermal history. We focus our results summary on the aspects of the models most relevant to understanding potential cooling related to thrust exhumation; namely (a) when did cooling initiate and (b) what was the duration of cooling through the relevant isotherms that our data are sensitive to? Several features of these models are important to consider when contrasting and comparing results. First, there are variable inputs for each (i.e., ZHe only vs. ZHe and AHe vs. ZHe, AHe, and AFT), which, in turn, might lead to finer or coarser resolution and path complexity on post-250 Ma cooling segments. Second, the path density and therefore warmest colors are directly related to the *conditional* probabilities of paths passing through those portions of t-T space. Although it may be tempting to draw a line connecting only the brightest colored portions of these plots and assume that is the “best” solution, we caution against such overinterpretation as the full path distribution should be considered. For example, our Copper Basin thrust result (Figure 5) shows a nominally straight and monotonic bright region between ~75 and 0 Ma, but closer inspection of the plot shows a large number of paths that cross over this region but display much more complexity and variability in their cooling trajectories. Third, our ability to address the start of cooling, or more accurately stated, the passage of paths through particular isotherms on these plots, is directly dependent on the temperature sensitivity of the thermochronometers used as inputs.

With these considerations in mind, we examine both the qualitative nature of each model result, but also employ a more quantitative measure of timing of cooling that addresses some of the limitations described above. For the western thrusts (Figure 5), path density suggests an earlier timing of youngest initial cooling—as represented by the youngest cooling inflection points drawn from the full posterior distribution (i.e., not just where the highest density portion of the distribution begins)—in the Copper Basin and Pioneer thrusts than the Lost River range: ~50–75 Ma versus ~140 Ma. Plots of predicted versus observed ages for these samples (Figures S1–S3 in Supporting Information S1) show decent agreement, with only one or two grains per sample underpredicted. Lemhi Arch thrusts (Figure 6) suggest cooling trajectories in the more western thrust sheets (Iron Lake, Poison Creek, and Hawley Creek thrusts) that began at ~100 Ma, and slightly younger timing for initial cooling between 90 and 70 Ma in the more eastern thrust sheets (Diamond Creek, Cabin, and Monument thrusts). There is good agreement for nearly all grains in this set of samples between the predicted and observed ages (Figures S4–S9 in Supporting Information S1). Foreland uplifts (Figure 7) in general show timing for initial cooling between ~75 and 50 Ma. The exception here is the model for the Eastern Blacktail-Snowcrest uplift, but we have some skepticism of this model result due to both the structure of the paths in Figure 7 and the modeled single grain output. A comparison of the predicted versus observed age plots for all four of these models (Figures S9–S13 in Supporting Information S1), shows that the model for the Eastern Blacktail-Snowcrest uplift does a poor job of reproducing either the ZHe or the AHe data. The other three foreland uplift models are variable in the degree to which they reproduce observed data, with the Western Blacktail-Snowcrest (Figure S10 in Supporting Information S1) and Northern Madison Range (Figure S13 in Supporting Information S1) uplifts doing a better job than the Southern Madison Range (Figure S12 in Supporting Information S1).

To be more quantitative with our assessments of timing of cooling, we use a method and accompanying Julia code (<https://github.com/OpenThermochronology/CoolingFDHM>) described by McDannell and Keller (2024) that calculates the probability distribution of times when all of the paths for a given model pass through a specific isotherm. These authors refer to this method as the “half-maximum isotherm” approach. This method requires the analyst to pick relevant isotherms based upon the temperature sensitivity of the modeled thermochronometers. Given our inputs, we use the 200°C and 150°C for our sample locations with only ZHe data. We argue that this range fully bounds the partial retention zones for our zircon grains. We show the 100°C isotherm for samples that have AFT, and extend our isotherms to as cool as 50°C for samples with AHe data (note that some locations do not have a 200°C and/or 150°C isotherm test as they were never that hot post-250 Ma). Individual results for each isotherm are shown in the supplementary figures, and the summary result of this exercise is shown in transect form in Figure 8. We have excluded the isotherm results from the Eastern Blacktail-Snowcrest from this summary figure for the reasons described above, as well as the results from the Pioneer and Copper Basin thrusts as those cooling histories are not thrusting related.

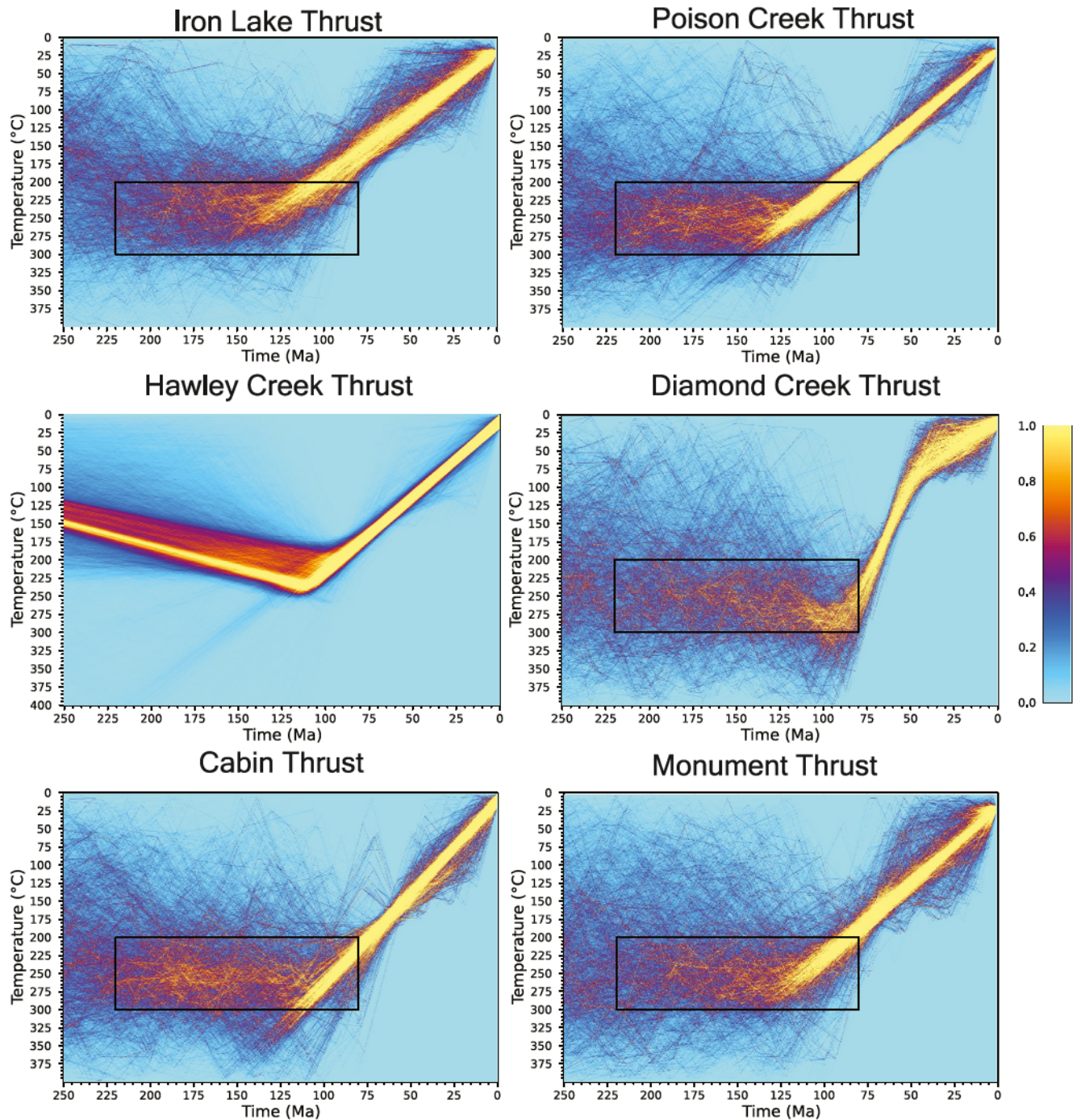


Figure 6. QTQt inverse model output from the Lemhi Arch thrust sheets. The posterior distribution is shown for each location in a normalized path density scale. These densities are directly related to probability such that warmer or brighter colors indicate that more paths pass through that region of t - T space. The black outlines indicate constraint boxes, justifications for these constraints can be found in Table S7.

We highlight several key observations from Figure 8. First, cooling through the 200°C isotherm started as early as ~130 Ma in the Lost River Range, and youngs to ~95 Ma across the Lemhi Arch and ~80 Ma in the Northern Madison Range. And second, cooling through the 150°C and 100°C isotherm occurred between ~60 and 40 Ma in both the Lemhi Arch thrusts and the foreland uplifts.

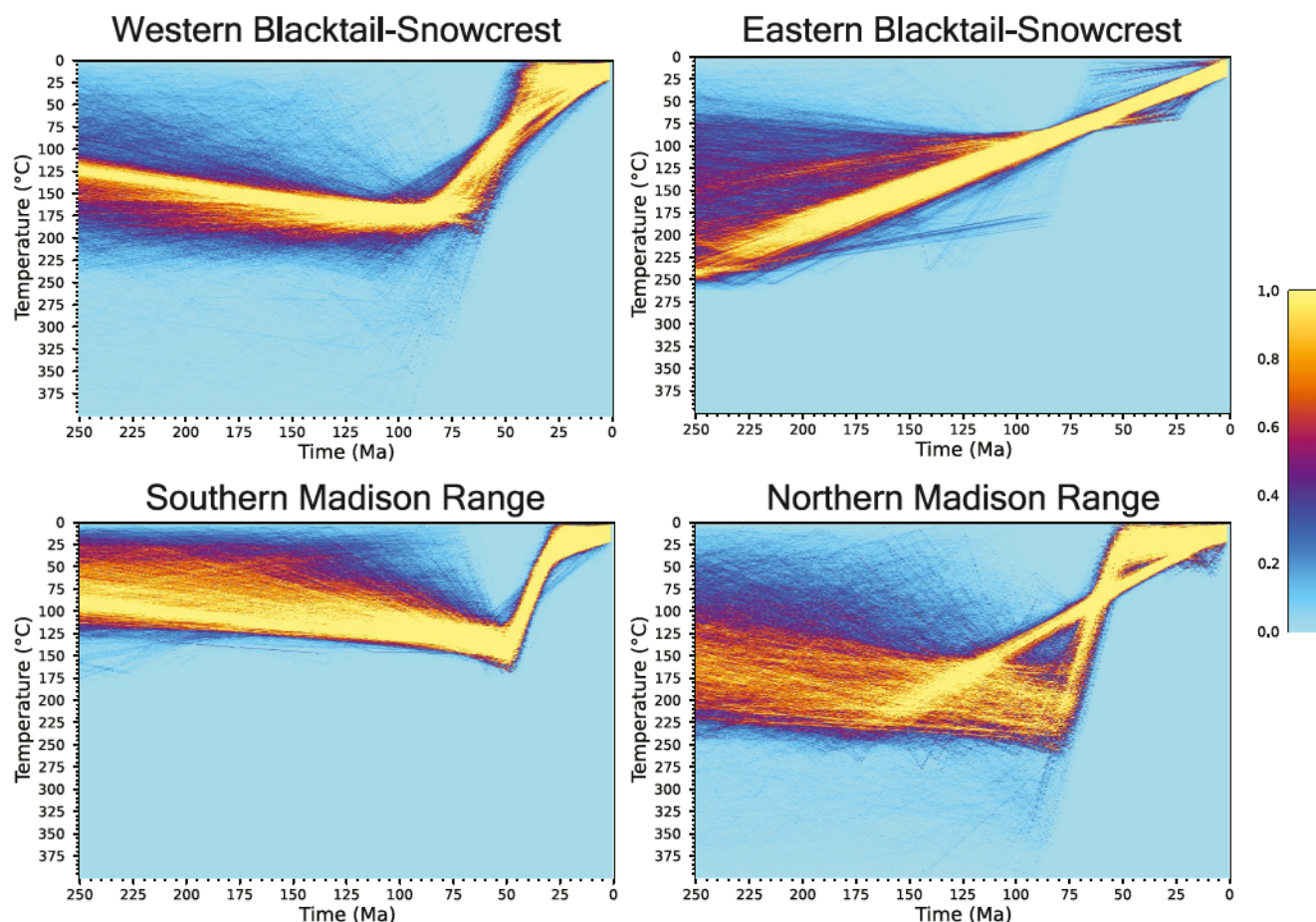


Figure 7. QTQt inverse model output from the foreland uplifts. The posterior distribution is shown for each location in a normalized path density scale. These densities are directly related to probability such that warmer or brighter colors indicate that more paths pass through that region of t - T space.

6. Discussion

6.1. Comparison of Exhumation Timing Across the Transect

The thermochronometric results and thermal history modeling output along our transect demonstrate the timing of deformation-induced exhumation of both thin- and thick-skinned structures in the Idaho-Montana fold-thrust belt. Below, we interpret the relevant data in the context of different drivers of exhumation for different thrust hanging walls. These cooling responses correspond roughly to our regional thrust sheet groupings. Broadly, measured and modeled results show temporal overlap between cooling signatures in the thin-skinned thrusts in Idaho and the thick-skinned uplifts in Montana (Figure 8), likely signaling overlap in timing of exhumation as well. This coincident deformation bridges the end of Sevier tectonism in Idaho and Wyoming (Yonkee & Weil, 2015), and Laramide tectonism toward the southeast (Burtner & Nigrini, 1994; Wiltschko & Dorr, 1983).

6.1.1. Exhumation Timing for Western Thrusts

To the west of the Lemhi arch, in the interior, thin-skinned portion of the Idaho-Montana fold thrust belt, we interpret that post-Cordilleran magmatic events thermally reset our samples, or extension resulted in exhumation of thermally reset ZHe samples, providing little insight into the movement on the Pioneer and Copper Basin thrusts. For these hinterland thrusts, we interpret the ~50–40 Ma ZHe dates (samples 18JK01, 18JK02, 14DW03, 16DP100, 16DP101), as well as the middle Paleogene isothermal cooling signatures in our models, in the context of regional post-orogenic extensional exhumation, which is broadly related to the development of the Pioneer metamorphic core complex. Ages of core complex formation include an early phase of high temperature,

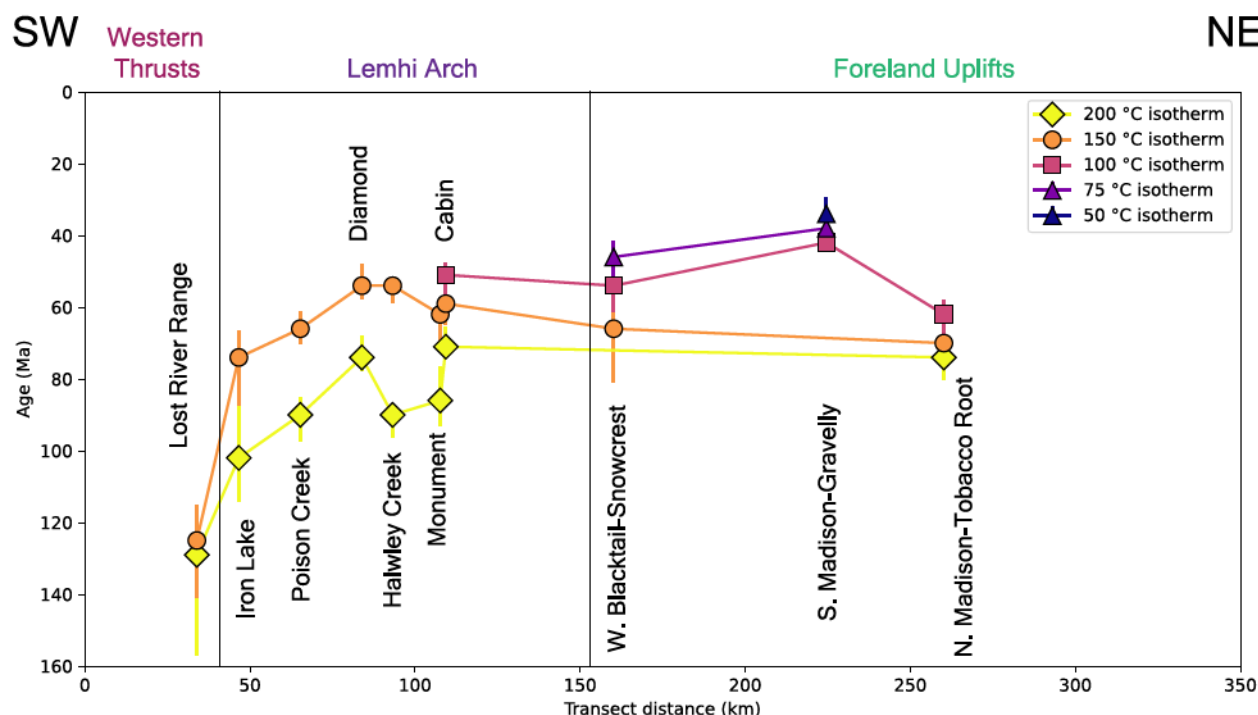


Figure 8. Compiled half-maximum isotherm times across a transect that runs from A to A' in Figure 1 (sample locations are orthogonally projected onto this line). Symbols represent the mean age of all paths that pass through that isotherm, and the error bars represent the 2σ standard deviation about the mean. Lines connecting points are for visual aid. Isotherm ages for the highest temperatures decrease from west to east. Intermediate temperature isotherm ages overlap between the foreland uplifts and the Lemhi Arch thrusts.

extensional strain between 50 and 46 Ma with an enigmatic linkage to upper crustal structures (Vogl et al., 2012), and later tectonic exhumation along the Wildhorse detachment from 38 to 33 Ma (McFadden et al., 2015; Silberberg, 1990; Vogl et al., 2014). Our ZHe dates from both the Pioneer and Copper Basin thrust hanging walls span this full range of core complex extension. Although our samples are located a considerable distance from the immediate footwall of the core complex (i.e., the most deeply exhumed rocks in the area), our results from the hinterland thrust sheets suggest that core complex extension and uplift imparted a more regional exhumation signature that overprinted any earlier thrust belt-related exhumation. Stable isotope data from McFadden et al. (2015) suggest the Pioneer metamorphic core complex underlaid a region of high topography, which may have been actively eroding, imparting a broad signature of exhumation at 38 Ma. Interestingly, the Copper Basin data set, located southeast of the metamorphic core complex, yields model results with isothermal cooling through the 200 and 150°C isotherms at ~58 and ~42 Ma, respectively (Figure S2 in Supporting Information S1). These cooling episodes overlap with the early high temperature extensional phase, and the Copper Basin samples may have been situated in an early break-away zone associated with initial core complex development, as hypothesized by Vogl et al. (2012, 2014). Regardless of the precise mechanism of exhumation, dates from these thrust sheets were influenced by later exhumation unrelated to Cretaceous thrusting, and we will not discuss them further in the context of exhumation directly related to thrust belt evolution.

Farther east, the upper, thin-skinned thrust system was detached predominantly in fine-grained siliciclastic and carbonate rocks, which did not provide the necessary zircon or apatite crystals for low-temperature thermochronometry. One locality where we were able to sample, in the hanging wall of an unnamed thrust system beneath the Lost River Range, gave dates older than 100 Ma, with cooling through the 200 and 150°C isotherms at ~129 and ~125 Ma, respectively (Figure S3 in Supporting Information S1). This sample (14DW03) was collected from the hanging wall of a series of fault-bend and fault-propagation folds detached in weak silty carbonates of the McGowan Creek Formation (Anastasio et al., 1997; Fisher & Anastasio, 1994). This thrust system was interpreted to have been folded and exhumed along the western edge of the Patterson culmination during deformation within the deeper, thick-skinned thrust system developed beneath the Lemhi arch unconformity

(Parker & Pearson, 2023). We thus interpret this Early Cretaceous age—in spite of some dispersion—as a rough age constraint on the older thin-skinned thrust system above the Lemhi arch unconformity.

6.1.2. Exhumation Timing for Lemhi Arch Thrusts

Lemhi Arch thrusts, including the Patterson culmination, Iron Lake thrust, Hawley Creek/Poison Creek thrust, Diamond Creek/Brushy Gulch fault system, and Cabin/Monument/Freeman thrust, contain structural evidence primarily for thick-skinned deformation that initiated following step-down in the structural level of the basal detachment horizon of the fold-thrust belt (Parker & Pearson, 2021). Samples were collected from mechanical basement units beneath the upper thin-skinned thrust system's basal detachment horizon in the hanging walls of all of these thick-skinned thrusts, and ZHe dates from these thrusts are generally Late Cretaceous in age (single aliquot date range from 108 to 52 Ma, Figure 3b). Mean isothermal cooling times through 200 and 150°C are 102 and 74 Ma for the Iron Lake thrust (Figure S4 in Supporting Information S1), respectively; 90 and 66 Ma for the Poison Creek thrust (Figure S5 in Supporting Information S1), respectively; and 90 and 54 Ma for the Hawley Creek thrust (Figure S6 in Supporting Information S1), respectively. The more eastern thrusts in the Lemhi Arch group are younger in terms of their 200 and 150°C isothermal cooling signatures: 74 and 54 Ma for the Diamond Creek/Brushy Gulch fault system (Figure S7 in Supporting Information S1), 71 and 59 Ma for the Cabin thrust (Figure S9 in Supporting Information S1), and 86 and 62 Ma in the Monument thrust (Figure S10 in Supporting Information S1). Integration of measured and modeled results from all of these locations suggests that broader Lemhi Arch exhumation began at ~102–90 Ma in the west and ~86–71 Ma in the east, and continued through subsequent post-orogenic collapse and Basin and Range extension.

These new results from low-temperature thermochronology for the Lemhi Arch area are consistent with detrital zircon provenance studies for synorogenic foreland basin rocks of the Frontier Formation (Finzel et al., 2023), and Beaverhead Group in southwestern Montana (Garber et al., 2020). These studies showed that ~500 Ma zircon grains from the Beaverhead pluton and host rocks of Mesoproterozoic Belt Supergroup quartzites in the hanging wall of the Hawley Creek thrust first appeared in rocks with a maximum depositional age of ~67 Ma. This depositional age is within the range of possible modeled cooling in the Hawley Creek vertical transect, and overlaps with some of the other cooling dates from Lemhi Arch thrusts. The coalescence of Late Cretaceous ages across all faults above the Lemhi arch indicate a pulse of deformation and related exhumation associated with thick-skinned thrusting in the region of overlap between the thin-skinned and thick-skinned fold-thrust belts. Pre-95 Ma thrusting in the upper, thin-skinned portion of the fold-thrust belt, is constrained by models and ZHe ages presented here and 97–94 Ma plutons that crosscut the Pioneer thrust and subsidiary structures (Brennan et al., 2020; Rodgers et al., 1995). Additional field observations indicate that older thin-skinned thrusting was synchronous with and eventually truncated by younger, structurally deeper, thick-skinned thrusts (Brumbaugh & Hendrix, 1981; Lopez & Schmidt, 1985; McDowell, 1997; O'Neill et al., 1990; Parker & Pearson, 2021; Perry et al., 1988; Schmidt & Perry, 1988; Skipp, 1988; Tysdal, 1988). Taken together and in the context of existing results, these data thus suggest that thin-skinned thrusting occurred primarily in Early Cretaceous time (~129 Ma), and may have been synchronous with but was largely exhumed and truncated after ~90 Ma by deeper, thick-skinned thrusts carrying rocks from beneath the Lemhi arch unconformity.

Samples from the hanging wall of the Tendoy thrust have substantially older ZHe ages than the Cabin thrust. These data show a large degree of scatter and are older than the depositional age of the Pennsylvanian Quadrant Formation host rock (Nichols et al., 1985). Combined, these observations suggest that zircon grains in the Tendoy sheet were not reset post-deposition, and that the Tendoy hanging wall was exhumed from a shallow structural depth, which is corroborated by maximum burial temperatures constrained by Raman spectroscopy of carbonaceous material to be less than ~115°C (Parker et al., 2022). These dates from the Tendoy thrust therefore demonstrate the decreasing structural depth, and limited exhumation, of the thin-skinned portion of the belt, as well as the dominance of thick-skinned structures in accommodating horizontal shortening of the crust in the absence of a thick sedimentary column. The Tendoy thrust thus marks a key transition in our transect. As we argue below, this transition can be directly interpreted in the context of spatial and temporal transitions from thin-to thick-skinned thrusting.

6.1.3. Exhumation Timing for Foreland Uplifts

The foreland basement uplifts observed in southwestern Montana with close proximity to the Idaho-Montana thrust front show evidence of concurrent deformation with the Idaho-Montana thrust belt (e.g., Kaempfer et al., 2021; Ronemus et al., 2023), which preceded the established timing of Laramide activity in nearby Wyoming uplifts (Peyton et al., 2012; Stevens et al., 2016). Existing data suggest that Blacktail-Snowcrest growth overlapped with thin-skinned deformation in the fold-thrust belt (DeCelles, 1986; Finzel et al., 2023; Garber et al., 2020; Nichols et al., 1985; Perry et al., 1983). The Kootenai Formation, interpreted by DeCelles (1986) to record paleoflow changes in response to the uplifting Madison, Gravelly, Tobacco Root, and Blacktail-Snowcrest arch, has a depositional age of ~135 to ~110 Ma (Finzel & Rosenblume, 2021; Rosenblume et al., 2021), suggesting Early Cretaceous, low-magnitude deformation in these ranges. Significant basement-involved deformation within the Blacktail-Snowcrest arch is recorded by recycling of Cretaceous and underlying Pennsylvanian to Permian strata atop the arch into the Frontier Formation by 87–85 Ma (Finzel et al., 2023). The presence of basement clasts in the Sphinx conglomerate of the Beaverhead Group provides further evidence for exhumation of the Madison and Gravelly ranges along the east-verging Scarface thrust, occurring between 73 and 65 Ma (DeCelles et al., 1987). Finally, the appearance of Blacktail-Snowcrest arch-derived basement clasts (Garber et al., 2020; Schmitt et al., 1995) and first-cycle detrital apatite in Beaverhead Group conglomerates establishes exhumation of these ranges by Campanian time (Finzel et al., 2025).

Our dates from these basement ranges in southwestern Montana provide further constraints on the timing and magnitude of basement-involved thrusting in this portion of the Laramide province. Cooling through the 200 to 100°C isotherms for the Western Blacktail-Snowcrest, Southern Madison, and Northern Madison models is broadly overlapping and spans ~74 to 42 Ma (Figure 9; Figures S10–S12 in Supporting Information S1). We therefore interpret a major phase of cooling in these uplifts to have initiated between 74 and 66 Ma (highest temperature isotherms in Madison Range and Blacktail-Snowcrest uplift, respectively), which overlaps with some of the timing estimates described above, as well as the timing of cooling for thrust sheets in the eastern portion of the Lemhi Arch (Figure 9). In combining our model results from basement-involved structures in both the Lemhi Arch and the foreland uplifts, we suggest that early basement-involved thrusting resulted in mid-Cretaceous (~90 Ma) break-up of the once contiguous southwestern Montana foreland basin, followed by a more significant phase of Late Cretaceous (~75 Ma) basement-involved shortening, arch exhumation, and cooling.

There is also some evidence in our thermal history models of foreland exhumation prior to 100 Ma. Model results suggest cooling paths from the Eastern Blacktail-Snowcrest arch (Figure S11 in Supporting Information S1) began prior to 100 Ma from temperatures in excess of 200°C. These models do a poor job of reproducing the ZHe data and thus lack resolution at higher temperatures. However, AFT central ages of 94.8 and 77.1 Ma from the eastern Blacktail-Snowcrest arch support this concurrent exhumation with the Idaho thrust belt, placing basement rocks near the surface leading up to the Late Cretaceous appearance of basement clasts in the Lima conglomerate (Garber et al., 2020; Schmitt et al., 1995). More convincing evidence for a pre-100 Ma cooling or exhumation signal is seen in results for the Western Blacktail-Snowcrest and Northern Madison models that show paths with cooling from high temperature beginning prior to 100 Ma (Figure 9). This observation is echoed in the half-maximum isotherm histogram plots for these two models (Figures S10 and S13 in Supporting Information S1). We particularly note the long tails that stretch to ages older than 100 Ma in the plots of the 150°C and 200°C isotherms. Although more speculative than some of our other results, this timeline of an earlier (>100 Ma), low magnitude basement-involved uplift event in the foreland is corroborated by DeCelles' (1986) interpretation of the Kootenai Formation in this area, reconnaissance low-temperature thermochrometric data from Carrapa et al. (2019), and growing evidence of similar periods of early uplift from the broader Laramide province (Caylor et al., 2023; Ronemus et al., 2023).

6.1.4. Summary of Exhumation Timing

Collectively and in the context of prior data constraining the timing of horizontal shortening in the region, data from our SW-NE transect suggest: (a) early, <130 to >90 Ma, thin-skinned thrusting west of and structurally above the Lemhi arch; (b) a ~90 Ma age of activation of the structurally deeper, thick-skinned thrust system involving Mesoproterozoic quartzites and crystalline basement rocks beneath the Lemhi arch unconformity that may have been coincident with a decrease in shortening and exhumation magnitude during late, thin-skinned thrusting on frontal thrusts including the Tendoy; and (c) Late Cretaceous (86–64 Ma), high magnitude

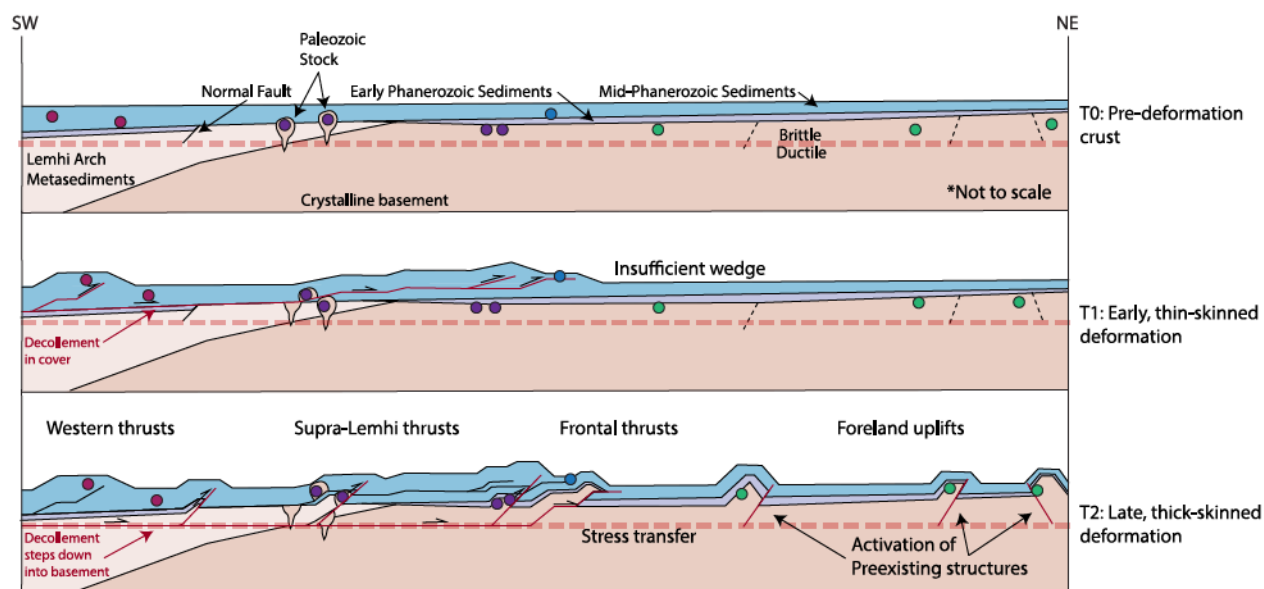


Figure 9. Schematic cross sections depicting sample locations at various phases of deformation. T0 shows pre-thrusting stratigraphy including Lemhi arch metasediments, early Paleozoic strata, and Mesozoic strata. T1 shows thrusting with a décollement in sedimentary rocks. T2 shows Late Cretaceous to early Paleogene thick-skinned thrusting of mechanical basement; thick-skinned thrusts crosscut prior thin-skinned thrusts (Parker & Pearson, 2021). This secondary thick-skinned deformation requires the décollement to down-step to the basement, which can accommodate stress transfer to the foreland.

exhumation northeast of the Tendoy thrust, especially along the Blacktail-Snowcrest arch, where some amount of exhumation may have commenced prior to ~100 Ma. Deformation related-exhumation in southwestern Montana basement-cored ranges therefore temporally overlapped with thick-skinned, Lemhi Arch thrusts (Figure 8), with a region of reduced exhumation, as represented by the unreset dates from the Tendoy thrust, in between.

We argue that this distribution of exhumation ages along the ID-MT transect supports a “double-decker” model whereby an earlier phase of thin-skinned deformation was followed by a later thick-skinned deformation causing coincident exhumation driven by uplift in two distinct regions (Lemhi Arch thrusts and Foreland Uplifts) of the thrust belt (Parker & Pearson, 2021). The temporal and spatial relationships among our thrust sheet groups can be best explained by a thrust belt that shifted, in specific locations, from thin-to thick-skinned (Figure 9). The double-decker system encompasses an initial phase of thin-skinned deformation along an upper-level detachment rooted in sedimentary cover (130–90 Ma uplift west of and structurally above the Lemhi arch), which was then cross-cut by thrusts rooted in the underlying mechanical basement along a lower-level detachment (~90 Ma activation of the structurally deeper, thick-skinned thrusts) (Lacombe & Bellahsen, 2016; Molinaro et al., 2005; Mouthereau et al., 2007; Parker & Pearson, 2021). Although the classic critical taper model (Davis et al., 1983) dictates a single, fixed detachment, an explanation of the driving mechanism behind the two detachment model still draws heavily from critical taper theory. That is, in thrust belts with thin sedimentary cover, secondary thick-skinned thrusting along a lower detachment can be an effective way to increase taper and permit thrust belt propagation (Lacombe & Bellahsen, 2016; Nemcok et al., 2006; Stuart et al., 2011). However, in contrast to the presence of a frictional detachment, which favors in-sequence deformation (Davis et al., 1983; Stockmal et al., 2007), activation of a second, deeper detachment strongly favors out-of-sequence deformation (Ruh et al., 2012). Parker and Pearson (2021) interpreted structural relationships in the Lemhi arch region, covered by a relatively thin package of sedimentary units, as reflective of this process (i.e., Figure 1c).

6.2. Implications for the Broader U.S. Cordillera

The co-exhumation of thin- and thick-skinned thrusts in the Idaho-Montana fold-thrust belt and southwestern Montana basement uplifts has implications for the ongoing debates over the controls of thrusting styles in the U.S. Cordillera and foreland. Spatial and temporal overlap between the thick- and thin-skinned portions of our transect further refutes the traditional understanding that these deformation styles are confined to discrete regions or times in the U.S. Cordillera (Dickinson et al., 1978). Further, the relationship between these two

endmember thrusting styles in our study area suggests that thick-skinned structures may not be intrinsically linked to changes in the plate boundary conditions or flat slab subduction. Our thermal history modeling shows that exhumation of thick-skinned structures in both the foreland and hinterland was ongoing by ~ 90 –80 Ma, which predates thermochronologic estimates of exhumation ≤ 70 Ma in “classic” Laramide structures in Wyoming (Peyton & Carrapa, 2013). The timing of exhumation in the Wyoming thick-skinned uplifts has been directly related to flat-slab passage (Fan & Carrapa, 2014; Peyton & Carrapa, 2013) as the region correlates spatially and temporally with the trajectory of shallow subduction and underplating (Liu et al., 2010; Saleeby, 2003). The substantially older dates of thick-skinned exhumation reported along our transect and by others for southwestern Montana (DeCelles, 1986; Garber et al., 2020; Nichols et al., 1985; Ryder & Scholten, 1973; Schmitt et al., 1995; Schwartz & DeCelles, 1988), suggests that uplift of thick-skinned structures cannot be directly related to the flat slab in this region. Furthermore, at the time the thick-skinned features of southwest Montana were exhuming, they were on the extreme periphery (about 500 km from the axis) of the modeled Shatsky-Hess conjugate plateau under-riding the basement cored uplifts in Colorado and Wyoming (DeCelles, 2004; Liu et al., 2010).

Further implications of our findings along the ID-MT transect contribute to the conversation around the importance of pre-existing structural architecture, or structural inheritance. Authors have posited that increased tectonic stresses may reactivate underlying weak zones in the upper crust (relict faults or shear zones) in order to accommodate deformation (Marshak et al., 2000; Wiltshko & Eastman, 1983). Structural reactivation of pre-thrust normal faults has been observed in the Lemhi Arch faults, as the mechanism by which thrusting style transitioned from thin-skinned to thick-skinned (Parker & Pearson, 2021). This pattern of reactivated faults was also observed in the Montana foreland basement rocks by McBride et al. (1989) who described multiple reactivations along the Snowcrest-Greenhorn lineament during the Paleoproterozoic Great Falls orogeny, Neoproterozoic rifting, Late Cretaceous basement involved thrusting, and finally Miocene extension. The repeated collisional and extensional episodes experienced by the northwestern boundary of the Wyoming craton provided a surfeit of deeply penetrating structures to reactivate during the Late Cretaceous uplift of foreland basement-cored ranges (Kaempfer et al., 2021). Specific to our transect, this pre-thrusting architecture had two influences. First, relatively thin stratigraphy over the Lemhi arch due to intra-Devonian unconformities (Grader et al., 2016) promoted down-stepping in the basal detachment and the development of a large ramp to increase wedge taper (Parker & Pearson, 2021). Second, strain was transferred along this detachment to foreland uplifts that exploited pre-existing structural weaknesses in southwestern Montana that had experienced repeated reactivation in the past (Garihan et al., 1983; Kaempfer et al., 2021; McBride et al., 1989; Perry et al., 1983; Perry Jr, 1986; Schmidt et al., 1993; Schmidt & Garihan, 1983, 1986).

7. Conclusions

The thermochronometric results of this study outline that cooling in thrust sheets overriding and eventually exhuming the Lemhi arch basement high began during the Early Cretaceous (~ 130 Ma) and continued through Late Cretaceous time. A Late Cretaceous cooling pulse from ~ 90 to 80 Ma through the end of the Cretaceous is also observed in the southwestern Montana foreland uplifts. We interpret these coincident cooling pulses as tectonic exhumation: cooling as a result of uplift along thrust faults and associated erosion. Exhumation observed in the southwestern Montana foreland uplifts pre-dates estimations of flat-slab related exhumation in neighboring areas. We consider the basement-cored uplifts in southwestern Montana, therefore, as linked to the concurrent thrust-belt deformation in Idaho, possibly via slip transfer along a mid-crustal detachment. This model explains, to some degree, the region with a limited magnitude of uplift (i.e., Tendoy thrust) between the Idaho-Montana thrust belt and Montana foreland.

The implications our results have on thrusting style and structural inheritance argue for the consideration of regional geology as a whole when interpreting low temperature thermochronologic results. Understanding and accounting for the long term tectono-thermal histories of ZHe, AHe, and AFT results is necessary to produce informed models that reflect the observed geology. Thermochronology, when observed at the orogen scale, can provide important timing constraints if placed in structural context and combined with independent constraints.

Data Availability Statement

The apatite and zircon (U-Th)/He and apatite fission track data presented in this study, as well as the modeling setup tables, are available at zenodo.org via <https://doi.org/10.5281/zenodo.13754918> (Kaempfer et al., 2024) with open access conditions.

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