

OROGENS OF BIG SKY COUNTRY: RECONSTRUCTING THE DEEP-TIME
TECTONOTHERMAL HISTORY OF THE BEARTOOTH MOUNTAINS,
MONTANA AND WYOMING, USA

by

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DEDICATION

In memory of my friend, Liam “Skinny” Alexander Taylor (1998–2017); his appreciation for the beauty of the natural world lives on in his brother, Koa, mom, Shelly, and the memory of all who loved him. “Love more” and “rally ‘round the Red”.



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ABSTRACT

The southwestern Montana region has experienced a protracted history of orogeny, burial, and erosion recording the development of the western margin of Laurentia, the core of the North American continent. This > 2.5 Gyr record contains clues about the nature of Precambrian tectonism, the development of economic mineral and hydrocarbon reserves, and the long-term geodynamic evolution of Earth. However, aspects of this history remain enigmatic, with events in the geologic record obscured by erosion and thermal overprinting. The manuscript presented herein, bound by introductory and concluding chapters, comprises a deep-time thermochronologic investigation of the Beartooth Mountains. New biotite $^{40}\text{Ar}/^{39}\text{Ar}$, and zircon U-Pb and (U-Th)/He data are presented from 14 samples collected from the Montana part of the range. These data indicate that thermal effects of Paleoproterozoic thermotectonism associated with the Big Sky orogeny (ca. 1.8–1.71 Ga) and/or related mantle metasomatism or mafic underplating penetrated into the core of these mountains. Thermal history model results indicate that this region of the craton experienced multi-phase Proterozoic cooling. The first phase of this cooling is generally coeval with the collapse of the Big Sky orogen. A second phase of Proterozoic cooling culminated in the development of the Great Unconformity surface, across which > 2 Gyr is regionally “missing” from the stratigraphic record. Constraints placing this latter phase between 1.4 Ga and 0.8 Ga preclude mechanisms predicting later Neoproterozoic–Cambrian cooling, such as erosion associated with Snowball Earth glaciation, and support diachronous development of the Great Unconformity surface in Laurentia. Thermal models resolve a Phanerozoic thermal history involving maximum burial temperatures by late Pennsylvanian time and cooling throughout Mesozoic time. This Phanerozoic thermal history, broadly out of sync with nearby basins, underscores the effects of interactions between far-field tectonism and inherited crustal weaknesses in the Beartooth Mountains and reconciles previous interpretations of pre-Late Cretaceous cooling with other evidence only constraining later phases of uplift. Finally, model results suggest Cenozoic reheating—likely due to burial by volcanics—and later cooling to surface temperatures due to erosional removal of these rocks—potentially related to encroachment of the Yellowstone hotspot and/or regional Basin and Range extension.

CHAPTER ONE

INTRODUCTION

Southwestern Montana has experienced one of the most ancient and varied geologic histories in Laurentia, the cratonic core of North America; multiple episodes of deformation, burial, and erosion spanning Archean to recent time have affected the region (e.g., Mueller and Frost, 2006; Mueller et al., 2008). During Phanerozoic orogenesis and the development of the North American Cordillera, Archean–Paleoproterozoic basement rock was uplifted in the hanging-walls of steep reverse faults (e.g., Coney and Reynolds, 1977; Dickinson and Snyder, 1978). These uplifts resulted in the formation of dramatic mountain ranges within the foreland of the Cordilleran system cored by Precambrian basement rock. In southwestern Montana, the Beartooth Mountains—positioned ca. 100 km from the northwest margin of the Wyoming craton, the southernmost major Archean component of Laurentia—comprises one such range.

The rocks exposed in the Beartooth Mountains provide a natural laboratory to investigate the multi-billion-year geologic history of the Laurentian continent and the orogenic, burial, and erosional events recorded therein. Key events with regionally to globally significant implications include: (1) Paleoproterozoic tectonism associated with the amalgamation of Laurentia (e.g., Harms et al., 2004); (2) the development of the Great Unconformity—across which > 2 Gy of time (ca. 2.8–0.54 Ga) is “missing” from the geologic record—and which likely had profound impacts on Earth’s cryosphere, atmosphere, and biosphere (e.g., Karlstrom and Timmons, 2012; Flowers et al., 2020 and references therein); (3) the deposition of Paleozoic rocks containing regionally important hydrocarbon reserves (e.g., Maughan, 1993); (4) basement-involved uplift associated with Phanerozoic orogenesis (e.g., Dickinson and Snyder, 1978); and (5) Cenozoic

exhumation responsible for shaping much of the modern topography of the region, regionally associated with Eocene–recent collapse of the North American Cordillera and, more locally, with the encroachment of the Yellowstone hotspot (e.g., Pardee, 1950; Wegmann et al., 2007). Despite the labor of geologists in this region spanning at least the previous century (e.g., Iddings, 1895; Peale, 1896; Hague, 1899), the specific timing, nature, and tectonic mechanism of many of these events remain largely enigmatic. In this thesis, I use radiometric geo- and thermochronology to elucidate aspects of this record and contribute to an integrated thermal history of southwestern Montana.

The manuscript presented herein (chapter two) comprises a geo- and thermochronologic investigation of the thermal history of the Beartooth Mountains, MT, and the record of geodynamic processes therein. Specifically, I report a large multi-chronometer dataset of zircon U-Pb crystallization ages, medium-temperature biotite $^{40}\text{Ar}/^{39}\text{Ar}$ cooling ages, and low-temperature zircon (U-Th)/He cooling dates, with the latter comprising a pronounced and well-defined negative date-effective uranium (eU; a proxy for radiation damage; e.g., Guenthner et al., 2013) trend. The geo- and thermochronometers utilized here encompass a large range of closure temperatures—the temperature below which a system retains most radiogenic daughter product (e.g., Reiners et al., 2017)—and consequently place constraints on a large volume of the time-temperature history experienced by these rocks. I conduct deep-time thermal history modelling of this dataset and integrate these models with decompacted sedimentary accumulation models and existing thermochronology results (e.g., Omar et al., 1994; Peyton et al., 2012) to explore the multi-billion year thermal history experienced by Archean basement rocks within this range.

This work is timely in that it answers recent calls for integration of multi-chronometer datasets in deep-time thermochronology (e.g., McDannell and Flowers, 2020) and leverages advances of the previous decade in understanding of the effect of radiation damage on helium retentivity within the zircon (U-Th)/He system (e.g., Guenthner et al., 2013; Ginster et al., 2019; Guenthner, 2021). My findings contribute to on-going debates of wide interest concerning the extent and nature of Paleoproterozoic thermotectonism in Montana (e.g., Giletti, 1966, 1971; Harms et al., 2004; Foster et al., 2006; Condit et al., 2015), the geodynamic mechanism for continental erosion associated with development of the Great Unconformity (e.g., DeLucia et al., 2018; Keller et al., 2019; Flowers et al., 2020), and the timing of basement-involved uplift in the foreland of the North American Cordillera (e.g., Carrapa et al., 2019; Garber et al., 2020; Orme, 2020). Broadly, results suggest that thermal events affecting the Beartooth Mountains are linked to both the plate tectonics of Earth's lithosphere as well as, potentially, processes originating within the asthenosphere below (e.g., Wegmann et al., 2007; DeLucia et al., 2018). I suggest that these processes interacted with inherited structural weaknesses—perhaps dating to at least Paleoproterozoic orogenesis—to produce variable cooling and heating out of sync with nearby regions.

Literature Cited

- Carrapa, B., DeCelles, P.G., and Romero, M., 2019, Early Inception of the Laramide Orogeny in Southwestern Montana and Northern Wyoming: Implications for Models of Flat-Slab Subduction: *Journal of Geophysical Research: Solid Earth*, v. 124, p. 2102–2123, doi:10.1029/2018JB016888.
- Condit, C.B., Mahan, K.H., Ault, A.K., and Flowers, R.M., 2015, Foreland-directed propagation of high-grade tectonism in the deep roots of a Paleoproterozoic collisional orogen, SW Montana, USA: *Lithosphere*, p. L460.1, doi:10.1130/L460.1.
- Coney, P.J., and Reynolds, S.J., 1977, Cordilleran Benioff zones: *Nature*, v. 270, p. 403–406, doi:10.1038/270403a0.
- DeLucia, M.S., Guenther, W.R., Marshak, S., Thomson, S.N., and Ault, A.K., 2018, Thermochronology links denudation of the Great Unconformity surface to the supercontinent cycle and snowball Earth: *Geology*, v. 46, p. 167–170, doi:10.1130/G39525.1.
- Dickinson, W.R., and Snyder, W.S., 1978, Plate tectonics of the Laramide orogeny, in Matthews, V. ed., *Laramide folding associated with basement block faulting in the western United States*, *Geological Society of America Memoir*, v. 151, p. 355–366.
- Flowers, R.M., Macdonald, F.A., Siddoway, C.S., and Havranek, R., 2020, Diachronous development of Great Unconformities before Neoproterozoic Snowball Earth: *Proceedings of the National Academy of Sciences*, p. 201913131, doi:10.1073/pnas.1913131117.
- Foster, D.A., Mueller, P.A., Mogk, D.W., Wooden, J.L., and Vogl, J.J., 2006, Proterozoic evolution of the western margin of the Wyoming craton: implications for the tectonic and magmatic evolution of the northern Rocky Mountains, doi:10.1139/e06-052.
- Garber, K.L., Finzel, E.S., and Pearson, D.M., 2020, Provenance of syn-orogenic foreland basin strata in southwestern Montana requires revision of existing models for Laramide tectonism: *North American Cordillera: Tectonics*.
- Giletti, B.J., 1971, Discordant isotopic ages and excess argon in biotites: *Earth and Planetary Science Letters*, v. 10, p. 157–164, doi:10.1016/0012-821X(71)90001-X.
- Giletti, B.J., 1966, Isotopic ages from southwestern Montana: *Journal of Geophysical Research*, v. 71.
- Ginster, U., Reiners, P.W., Nasdala, L., and Chanmuang N., C., 2019, Annealing kinetics of radiation damage in zircon: *Geochimica et Cosmochimica Acta*, v. 249, p. 225–246, doi:10.1016/j.gca.2019.01.033.
- Guenther, W.R., 2021, Implementation of an Alpha Damage Annealing Model for Zircon (U-Th)/He Thermochronology With Comparison to a Zircon Fission Track Annealing Model: *Geochemistry, Geophysics, Geosystems*, v. 22, p. e2019GC008757, doi:https://doi.org/10.1029/2019GC008757.

- Guenthner, W.R., Reiners, P.W., Ketcham, R.A., Nasdala, L., and Giester, G., 2013, Helium diffusion in natural zircon: Radiation damage, anisotropy, and the interpretation of zircon (U-Th)/He thermochronology: *American Journal of Science*, v. 313, p. 145–198, doi:10.2475/03.2013.01.
- Hague, A., 1899, The Early Tertiary Volcanoes of the Absaroka Range: *Science*, v. 9, p. 425–442.
- Harms, T.A. et al., 2004, Character and origin of Precambrian fabrics and structures in the Tobacco Root Mountains, Montana, in *Special Paper 377: Precambrian Geology of the Tobacco Root Mountains, Montana*, Geological Society of America, v. 377, p. 203–226, doi:10.1130/0-8137-2377-9.203.
- Iddings, J.P., 1895, Absarokite-shoshonite-banakitite series: *The Journal of Geology*, v. 3, p. 935–959.
- Karlstrom, K.E., and Timmons, J.M., 2012, Many unconformities make one ‘Great Unconformity’, doi:10.1130/2012.2489(04).
- Keller, C.B., Husson, J.M., Mitchell, R.N., Bottke, W.F., Gernon, T.M., Boehnke, P., Bell, E.A., Swanson-Hysell, N.L., and Peters, S.E., 2019, Neoproterozoic glacial origin of the Great Unconformity: *Proceedings of the National Academy of Sciences*, v. 116, p. 1136–1145, doi:10.1073/pnas.1804350116.
- Maughan, E.K., 1993, Stratigraphic and Structural Summary for Central Montana, in *Field Conference Guidebook: Energy and Mineral Resources of Central Montana*, Montana Geologic Society, p. 3–20.
- McDannell, K.T., and Flowers, R.M., 2020, Vestiges of the Ancient: Deep-Time Noble Gas Thermochronology: *Elements*, v. 16, p. 325–330, doi:10.2138/gselements.16.5.325.
- Mueller, P.A., and Frost, C.D., 2006, The Wyoming Province: a distinctive Archean craton in Laurentian North America: *Canadian Journal of Earth Sciences*, v. 43, p. 1391–1397, doi:10.1139/e06-075.
- Mueller, P.A., Mogk, D.W., Henry, D.J., Wooden, J.L., and Foster, D.A., 2008, Geologic evolution of the Beartooth Mountains: Insights from petrology and geochemistry: *Northwestern Geology*, v. 37, p. 5–20.
- Omar, G.I., Lutz, T.M., and Giegengack, R., 1994, Apatite fission-track evidence for Laramide and post-Laramide uplift and anomalous thermal regime at the Beartooth overthrust, Montana-Wyoming: *Geological Society of America Bulletin*, p. 12.
- Orme, D.A., 2020, New Timing Constraints for the Onset of Laramide Deformation in Southwest Montana Challenge our Understanding of the Development of a Thick-Skinned Structural Style During Flat-Slab Subduction: *Tectonics*, v. 39, p. e2020TC006193, doi:https://doi.org/10.1029/2020TC006193.
- Pardee, J.T., 1950, Late Cenozoic block faulting in western Montana: *Geological Society of America Bulletin*, v. 61, p. 359–406.

- Peale, A., 1896, Description of the Three Forks sheet, Montana: US Geological Survey Geol: Atlas, Folio, v. 24.
- Peyton, S.L., Reiners, P.W., Carrapa, B., and DeCelles, P.G., 2012, Low-temperature thermochronology of the northern Rocky Mountains, western U.S.A.: American Journal of Science, v. 312, p. 145–212, doi:10.2475/02.2012.04.
- Reiners, P.W., Carlson, R.W., Renne, P.R., Cooper, K.M., Granger, D.E., McLean, N.M., and Schoene, B., 2017, Geochronology and thermochronology: John Wiley & Sons.
- Wegmann, K.W. et al., 2007, Position of the Snake River watershed divide as an indicator of geodynamic processes in the greater Yellowstone region, western North America: Geosphere, v. 3, p. 272–281, doi:10.1130/GES00083.1.

CHAPTER TWO

OROGENS OF BIG SKY COUNTRY: RECONSTRUCTING THE DEEP-TIME
TECTONOTHERMAL HISTORY OF THE BEARTOOTH MOUNTAINS,
MONTANA AND WYOMING, USA

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Abstract

Precambrian rocks of the Wyoming craton exposed in the Beartooth Mountains, Montana and Wyoming, have experienced orogenic, burial, and erosional events spanning Archean to recent time. Aspects of this > 2.5 Gyr thermal history—recording important information about the long-term geodynamic evolution of Earth—remain largely enigmatic, with many events thermally overprinted or erased from the geologic record by erosion. We use recent advances in “deep-time” thermochronology to place constraints on this thermal history, reporting a large multi-chronometer dataset of biotite $^{40}\text{Ar}/^{39}\text{Ar}$ and zircon U-Pb and (U-Th)/He results from samples collected therein. Biotite $^{40}\text{Ar}/^{39}\text{Ar}$ analyses from our western samples record dates of 1.62–1.78 Ga; by contrast, biotite $^{40}\text{Ar}/^{39}\text{Ar}$ ages of eastern analyses are 2.26–2.4 Ga. These data indicate the thermal effects of Paleoproterozoic tectonism—likely related to the Big Sky orogeny (ca. 1.8–1.71 Ga)—extended into the core of the Beartooth Mountains, consistent with some previous work documenting a similar thermal overprint farther west in the range. Zircon (U-Th)/He dates show a pronounced negative correlation with effective uranium (eU), a proxy for radiation damage, recording Proterozoic to near-zero dates across a span of > 3400 ppm eU. Results from thermal history modeling indicate multi-phase Proterozoic cooling from ca. $400\text{ }^{\circ}\text{C}$ to $\leq 80\text{ }^{\circ}\text{C}$ between ca. 1600 Ma and 800 Ma. This cooling was likely associated, in part, with post-orogenic collapse of the Big Sky orogen and later continental erosion contributing to the development of the Great Unconformity surface. We resolve maximum Phanerozoic burial temperatures of ca. $145\text{--}155\text{ }^{\circ}\text{C}$ by Late Pennsylvanian time followed by a prolonged phase of Phanerozoic cooling to near-surface temperatures between Late Pennsylvanian and Paleocene time—likely associated with Ancestral Rocky Mountains tectonism and development of the

North America Cordillera. The most rapid cooling occurs at ca. 70–55 Ma, reconciling previous work suggesting ≥ 120 Ma initiation of Laramide uplift with that resolving only latest Cretaceous to Paleocene exhumation. Collectively, this study elucidates a thermal history of a cratonic margin region strongly coupled to diverse geodynamic processes and demonstrates the utility of deep-time thermochronology in resolving the complex thermal histories recorded in Precambrian terranes.

Introduction

The Beartooth Mountains of southern Montana and northern Wyoming are dominantly composed of ca. 3.5–2.8 Ga basement rocks of the Wyoming craton (Wooden et al., 1988; Mueller and Frost, 2006; Mueller et al., 2008 and references therein). This region, located ca. 100 km southeast of the northwest margin of the craton, has been involved in tectonic events spanning more than 2 Gyr—from Archean intracratonic reorganization to Mesozoic–Cenozoic Cordilleran orogenesis (e.g., Dickinson and Snyder, 1978; Mogk et al., 1992; Chamberlain et al., 2003). The Beartooth Mountains have been the subject of geologic investigation since at least the 19th century (e.g., Hague, 1899; Bevan, 1923), including some of the earliest work incorporating radiometric geochronology (Evans et al., 1939). Despite this century of inquiry, many details of the thermal history of the region and the Proterozoic-to-recent geologic events recorded therein remain enigmatic (e.g., Condit et al., 2015; Carrapa et al., 2019). High to medium-temperature geochronometers typically yield Archean–Paleoproterozoic ages (e.g., Gast et al., 1958; Wooden et al., 1988), while conventional low-temperature thermochronology has been leveraged mainly to explore late Mesozoic–early Cenozoic thermal events (e.g., Cervený, 1990; Omar et al., 1994;

Peyton et al., 2012; Carrapa et al., 2019). Consequently, an integrated thermal history spanning this >2.5 Gyr time remains unresolved.

Recent developments in “deep-time” thermochronology have demonstrated the utility of fission track and noble gas methods in characterizing Earth processes occurring on 10^8 to 10^9 -year timescales in the upper regions of the crust (see McDannell and Flowers, 2020 for a review). In regions such as craton margins—which have experienced significant Phanerozoic burial, resetting many low-temperature thermochronometric systems—the broad temperature sensitivity of the zircon (U-Th)/He (ZHe) system is particularly salient (e.g., Orme et al., 2016). The prior decade of advancements in the understanding of the influence of radiation damage and annealing on diffusivity of helium in the zircon crystal show that these factors exert a strong control on the closure temperature (T_c) of the ZHe system (e.g., Guenthner et al., 2013; Johnson et al., 2017; Ginster et al., 2019; Guenthner, 2021). In rocks experiencing protracted low-temperature thermal histories, the damage-diffusivity relationship of zircon commonly manifests as a correlation between single grain date and parent isotope concentration (or effective Uranium; $eU = U + 0.235 \times Th$, a proxy for radiation damage; Flowers et al., 2009). The shape of this date- eU correlation is sensitive to the entirety of a grain’s thermal history during which it accumulated radiation damage and/or radiogenic helium—that spent above the base of the partial retention zone (PRZ). This enables the ZHe system to resolve billion-year low–moderate temperature thermal histories in regions where other evidence of thermal events is overprinted or eroded (e.g., Orme et al., 2016; DeLucia et al., 2018; McDannell et al., 2019; Flowers et al., 2020).

Here, we report a large dataset of zircon U-Pb, biotite $^{40}\text{Ar}/^{39}\text{Ar}$, and ZHe geo- and thermochronologic data from samples collected along three transects in the Beartooth Mountains. We couple this dataset with existing thermochronologic data (e.g., Cervený, 1990; Omar et al., 1994; Bricker, 2016; Carrapa et al., 2019) and decompacted sediment accumulation models to investigate the long-term thermal history of the range. Our biotite $^{40}\text{Ar}/^{39}\text{Ar}$ data suggest that the western portion of the Beartooth Mountains experienced thermal overprinting by Paleoproterozoic thermotectonism. Observed cooling ages are generally coincident with the Big Sky orogenic event—documented to have effected regions to the west in Montana (e.g., Harms et al., 2004; Hames and Harms, 2013; Condit et al., 2015)—suggesting the thermal effects associated with this convergent tectonism extended into the core of the Beartooth Mountains.

Our ZHe data plot along a well-defined negative date-eU correlation, with dates spanning 13.5 Ma (high-eU) to 686 Ma (low-eU). We exploit this observed relationship between damage and diffusivity, as well as integrate thermal information from our biotite $^{40}\text{Ar}/^{39}\text{Ar}$ data, to test major hypotheses for the tectonic and erosional evolution of Laurentia. We specifically tailor our models to test the timing and magnitude of: (1) continental erosion leading to the development of the Great Unconformity surface, where locally more than 2 Gyr (ca. 2.8–0.54 Ga) of stratigraphy are missing across the erosional contact between metamorphic basement rocks and overlying Cambrian sedimentary rocks (e.g., Flowers et al., 2020); (2) burial temperatures experienced during Paleozoic and Mesozoic deposition (e.g., Maughan, 1993; DeCelles, 2004); (3) subsequent cooling associated with Phanerozoic tectonism (e.g., Carrapa et al., 2019); and (4) possible Cenozoic reheating due to Eocene and/or Neogene volcanism and subsequent cooling (e.g., Omar et al., 1994). The former has recently been implicated in

profound changes in Earth's evolution, including the initiation of Cryogenian Snowball Earth glaciation (e.g., Cox et al., 2016), oxygenation of Earth's oceans and atmosphere (Husson and Peters, 2017), and the Cambrian explosion of life (Peters and Gaines, 2012). The latter three events bear directly on the local maturation of hydrocarbon resources and the development of regional topography and ecosystems (e.g., Whitlock et al., 2002; Peyton and Carrapa, 2013). Our results have broad implications for understanding of the geodynamic mechanisms culpable for these major geologic events and contribute to an integrated understanding of the Precambrian to recent thermal history of one of the most ancient regions of Earth's crust.

Geologic Setting

Geographic and Structural Setting

The Beartooth Mountains are a prominent ca. 8250 km² northwest-southeast trending fault-bounded uplift that formed within the North America Cordillera foreland basin system of southern Montana and northernmost Wyoming (Figure 1; Foose et al., 1961; DeCelles, 2004; Yonkee and Weil, 2015). They rise abruptly ca. 1500 vertical meters from the Great Plains to the east to elevations up to 3904 m at Granite Peak, the Montana state highpoint. The range is bounded on the east by the Bighorn Basin and on the northeast and north by the Nye-Bowler lineament (Wilson, 1936) and Crazy Mountain Basin (Figures 1–3A). The south and southwest regions of the range are buried by a thick succession of Cenozoic volcanic and volcanoclastic rocks, which form much of the Absaroka Mountains to the south (Figure 2; Foose et al., 1961).

Precambrian rocks in the Beartooths Mountains comprise four main structural blocks distinguished by differing tectonic histories and lithologies, later described in more detail: (1) the Beartooth block, the main structural component of the range, including the study area in the

southeastern Beartooth Mountains; (2) the Stillwater complex, along the northeastern margin of the range; and (3) the North Snowy and (4) South Snowy blocks, comprising the northwestern and southwestern margins of the range, respectively (Figure 2; e.g., Foose et al., 1961; Reid et al., 1975; Montgomery and Lytwyn, 1984; Mogk et al., 1988, 1992). The eastern Beartooth Mountains form a broad plateau surface (the Beartooth Plateau) with an average elevation of >3000 m; this surface is comprised of smaller plateaus and ridges separated by intervening drainages, including the Stillwater River and Rock and Rosebud creeks (Figure 2).

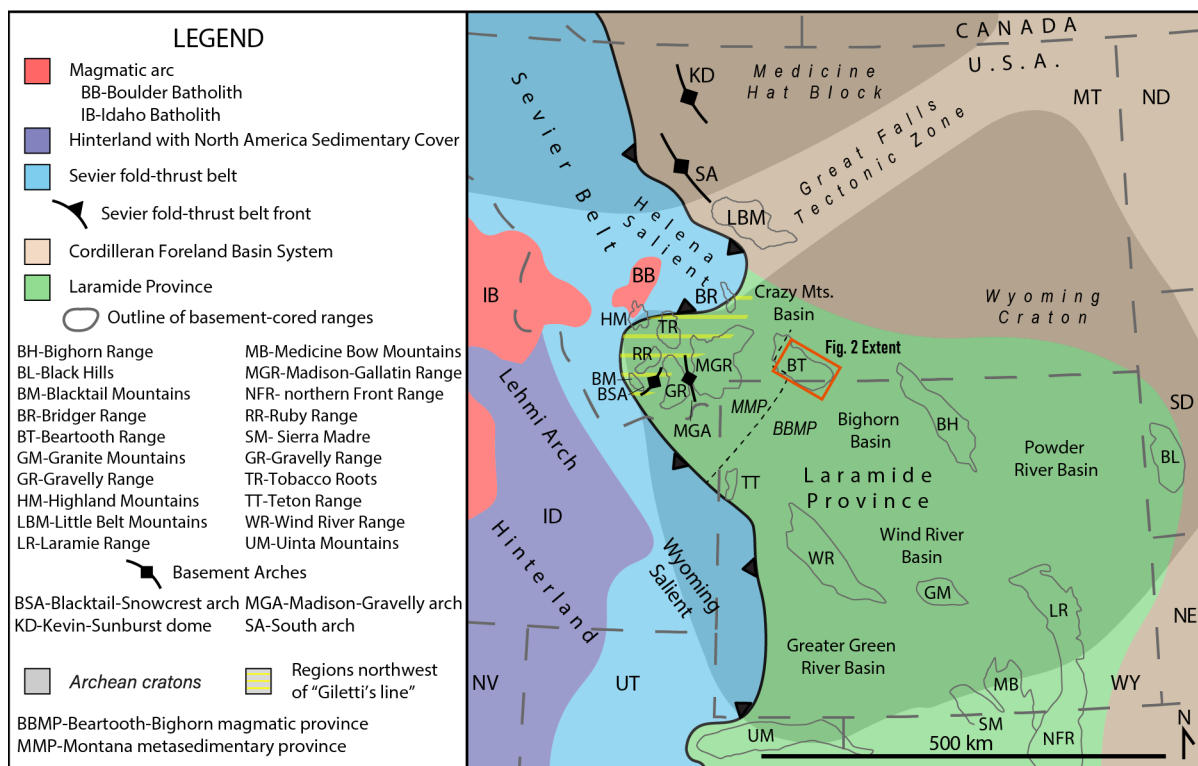


Figure 1: Generalized geologic map of the Northern Rocky Mountains region, showing the distribution of major tectonic features. The study area (orange extent box) in the Beartooth Mountains lies in the northwestern portion of the Archean Wyoming craton and in the northernmost portion of the Laramide Province, as traditionally defined (e.g., Dickinson and Snyder, 1978). Dashed line divides provinces of the Wyoming craton. Yellow stippling defines regions of the Wyoming craton west/northwest of "Giletti's line", where Paleoproterozoic thermal overprinting has been conclusively documented (Giletti, 1966). Abbreviations not in

legend are for US states. Modified from Orme (2020) and Ronemus et al. (2020) after Whitmeyer and Karlstrom (2007) (Precambrian tectonic elements) and Yonkee and Weil (2015) (Phanerozoic tectonic elements).

Precambrian Assembly and Tectonism

The Beartooth block dominantly comprises Mesoarchean igneous and metamorphic rocks, with a component of older Archean rock (Figure 2; Condie, 1976). These and similar rocks comprise the Archean Beartooth-Bighorn magmatic province (BBMP), a major sub-province of the Wyoming craton (Figure 1; Mogk et al., 1992; Mueller et al., 2008). The distribution of this metamorphic basement and the tectonic history it records is summarized here.

The Beartooth Mountains are volumetrically dominated by latest Mesoarchean–Neoarchean (ca. 2.9–2.8 Ga) granitoids and gneisses (Wooden et al., 1988). Paleo–Mesoarchean (3.5–3.0 Ga) gneissic rocks of dominantly tonalitic to granodioritic to trondhjemitic composition (TTG suite) are preserved throughout the range as meter-scale xenoliths to km-scale pendants within the 2.9–2.8 Ga suite (Wooden et al., 1988; Mueller et al., 2008, 2010). The former early Archean suite preserves petrologic evidence of at least one stage of early Archean granulite facies metamorphism and is intercalated with metasupracrustal rocks of varying lithologies in some locations; in contrast, the latter Meso–Neoarchean suite was likely intruded during the waning phases of this protracted thermotectonism and is limited to amphibolite facies metamorphism (e.g., Mueller et al., 2008 and references therein). The Stillwater layered igneous complex is exposed along the northeastern margin of the range, near the Stillwater River (Figure 2). This complex of mafic and ultramafic rocks is a globally important source of platinum-group metals (e.g., Raedeke, 1983). It crystallized in Neoarchean time (ca. 2.709 Ga; e.g., Wall et al., 2018) and was likely tectonically juxtaposed against the main Beartooth block by 2500 Ma, as

suggested by the intrusion of the Mouat quartz monzonite into the Stillwater and main Beartooth blocks (Geissman and Mogk, 1986).

The North Snowy block in the northwestern Beartooth Mountains additionally records tectonism taking place ca. 2.74–2.55 Ga, marked by mylonitic rocks and abrupt changes in metamorphic grade at the contacts between distinct Archean lithologic units (Mogk et al., 1988). This is likely related to tectonic intracratonic reorganization of BBMP and Montana metasedimentary province (MMP)—comprising the suite of dominantly metasupracrustal Archean rocks to the west (Mogk et al., 1992). A similar contact exists within the Snowy shear zone of the western South Snowy block, though a Paleoproterozoic age has been suggested for movement along this structure (Guy and Sinha, 1988; Erslev, 1992; Webber et al., 2019). Rocks of the MMP additionally experienced high-grade metamorphism at ca. 2.45 Ga, recorded in monazite Pb-Pb ages (Roberts et al., 2002; Cheney et al., 2004) that temporally overlap with K-Ar ages reported from the southeastern Beartooth Mountains by Gast et al. (1958).

Existing evidence for later high-grade thermotectonism in the Beartooth Mountains is generally limited to the western margin of the range, which record partially reset $^{40}\text{K}/^{39}\text{Ar}$ and Rb-Sr ages documenting Paleoproterozoic reheating (ca. 1.8–1.7 Ga; Reid et al., 1975; Montgomery and Lytwyn, 1984). Additionally, a Paleoproterozoic metamorphic overprint to greenschist facies of the Stillwater Complex has been suggested (Page, 1977; Page et al., 1985), with ca. 1.7 Ga apatite U-Pb ages reported from therein (Nunes and Tilton, 1971). Significant Paleoproterozoic tectonism is recorded in rocks of the MMP to the west, with upper amphibolite to granulite facies metamorphism in western mountain ranges—including the Highland, Ruby, Tobacco Root, and Madison ranges—grading to greenschist facies near the western margin of

the Beartooth Mountains (Cheney et al., 2004; Harms et al., 2004; Condit et al., 2015 and references therein). Additionally, slightly older (ca. 1.86–1.81 Ga) high-grade tectonism and magmatism is recorded in Little Belt Mountains within the Great Falls tectonic zone to the north (Figure 1; e.g., Boerner et al., 1998; Mueller et al., 2002; Gifford et al., 2014). Gast et al. (1958) suggest that this Paleoproterozoic thermotectonism did not significantly affect the southeastern Beartooth Mountains, reporting older Rb-Sr (2.53–2.8 Ga) and K-Ar (2.29–2.52 Ga) ages from near the Beartooth Highway.

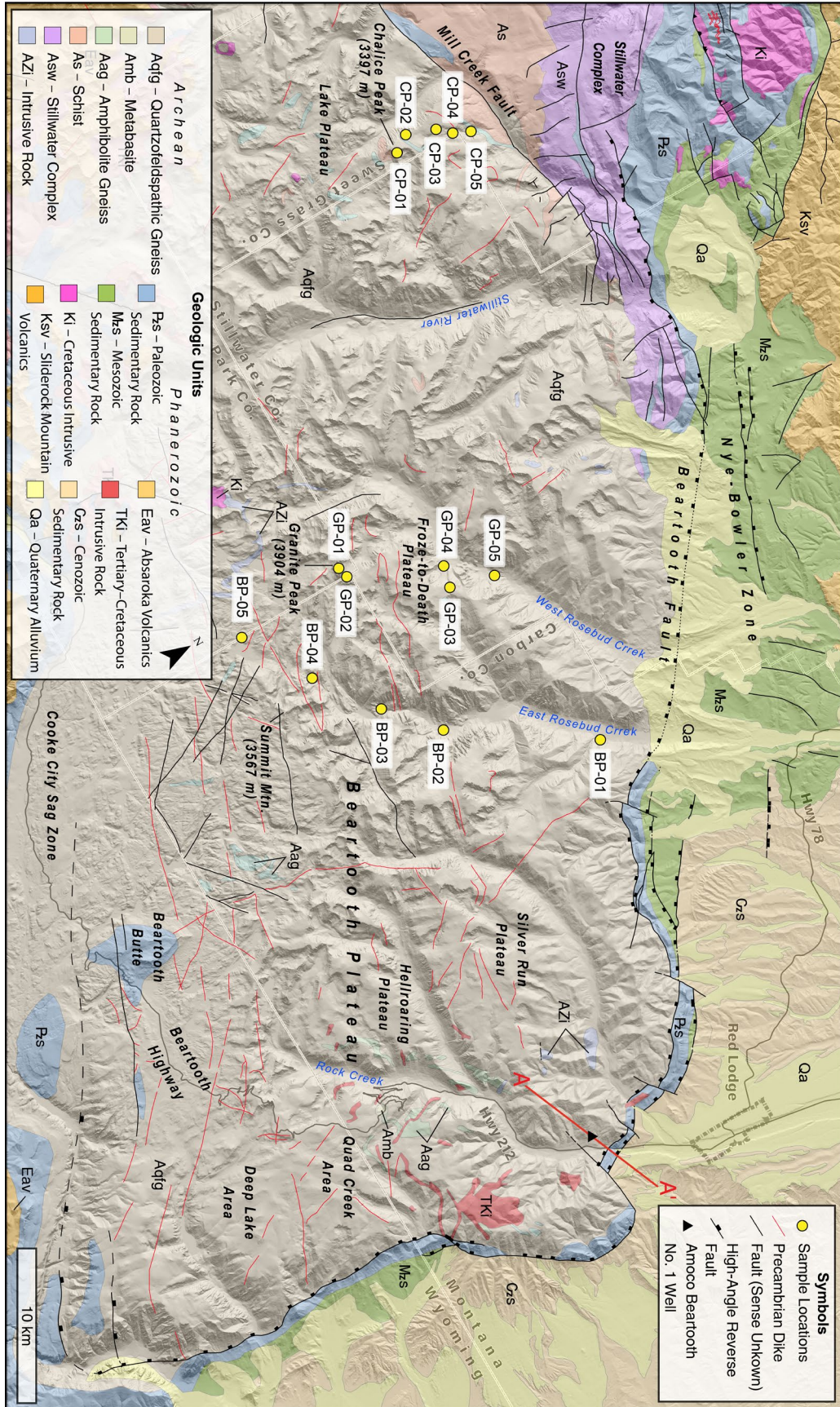


Figure 2: Generalized geology of the southeastern Beartooth Mountains. The range is comprised dominantly of Archean gneiss thrust over Paleozoic–Paleogene sedimentary rocks. Sample locations and names are shown by yellow dots and text boxes, respectively. Geologic units and faults are generalized from Foose et al. (1961; Wyoming geology) and Lopez (2001; Montana geology). Only Precambrian dikes longer than 2 km are shown. A–A’ cross section line corresponds to Figure 3B.

Paleozoic Sedimentation and Tectonism

In western North America, the erosional contact between Archean–Proterozoic basement rocks and overlying Paleozoic strata is known as the “Great Unconformity” (GU). More than 2 Gyr of stratigraphy is missing across this unconformity in the Beartooth Mountains, locally comprising Cambrian Flathead Sandstone (or quartzite) overlying metamorphic and igneous basement rocks of up to ca. 3.5 Ga, described above. In southwestern Montana, deposition of the Flathead Sandstone has been dated at 555 ± 18 Ma using the Rb-Sr whole rock method and 542 Ma by Rb-Sr dating of presumed authigenic glauconite (Chaudhuri and Brookins, 1969). We note that Chaudhuri and Brookins (1969) interpret Middle Cambrian deposition of the Flathead Sandstone, while their reported isotopic results would give an upper Ediacaran age using the modern geologic time scale (Walker et al., 2018). This is likely due to previous assignment of the Ediacaran–Cambrian boundary to ca. 570 Ma (Geyer and Landing, 2017 and references therein). This boundary has been subsequently revised to 541 Ma in the modern geologic time scale (Walker et al., 2018). While the Flathead Sandstone would be assigned Middle Cambrian age in older iterations of the geologic timescale—a designation which continues to pervade modern literature (e.g., Malone et al., 2017; Matthews et al., 2018)—a Fortunian or even upper Ediacaran assignment is plausible considering published isotopic data and the general lack of abundant datable fossils (Chaudhuri and Brookins, 1969; Bell and Middleton, 1978). This is consistent with recent paleopedological evidence suggesting a similar depositional age

(Retallack, 2013). However, we note that some faunal evidence—dominantly from overlying shale units—has instead been interpreted to assign the Flathead Sandstone to the *Albertella* faunal zone (e.g., Deiss, 1939; Norris and Price, 1966; Thomas, 2007), which equates to an early Middle Cambrian age in the modern geologic timescale (Peng et al., 2012). Although a reconsideration of the age of the Flathead Sandstone and the Ediacaran–Cambrian boundary in Montana is beyond the scope of this study, the deposition of this unit nonetheless provides a robust minimum temporal constraint on the time at which basement rocks of the Beartooth Mountains were exposed at the surface. For modeling purposes, we use a 530 ± 10 Ma depositional age for the Flathead Sandstone, which we consider parsimonious in light of the existing evidence for a Cambrian designation and the isotopic results of Chaudhuri and Brookins (1969) described above.

Throughout Paleozoic time, passive margin-style sedimentation regionally continued with the deposition of onlapping sequences of sandstone, limestone, and shale, with intervening periods of non-deposition and erosion (Mallory, 1972; Maughan, 1993). Deposition was punctuated by structural movements during Devonian and Late Mississippian to Permian time, likely associated with the Antler and Ancestral Rockies orogenic events, respectively (Maughan, 1990; Dorobek et al., 1991; Beranek et al., 2016). At present, a maximum thickness of ca. 1040 m of Paleozoic stratigraphy is preserved along the northeastern Beartooth front (Maughan, 1993; Lopez, 2001).

Mesozoic–Cenozoic Sedimentation and Uplift

In Middle Triassic to Late Jurassic time, the western margin of North America evolved from a region of fringing arcs and interarc oceanic basins into a tectonically consolidated margin

associated with east-dipping subduction of the Farallon oceanic plate beneath the North American continent (Dickinson, 2004; Colpron et al., 2007). This evolution culminated in the establishment of the North American Cordillera, a retroarc orogenic system extending more than 6000 km from southern Mexico to Alaska, USA (e.g., DeCelles, 2004; Yonkee and Weil, 2015). A complex and extensive foreland basin system developed predominantly as a consequence of lithospheric flexure from the retroarc fold-thrust belt (Figure 1; Kauffman and Caldwell, 1992; DeCelles, 2004). Near the Beartooth Mountains, stratigraphy deposited in this foreland basin constitute Jurassic through Upper Cretaceous rocks with a maximum thickness of ca. 2140–2410 m (e.g., Mallory, 1972; Maughan, 1993; Lopez, 2001).

The Beartooth Mountains and similar structures comprise an elongate series of intra-foreland uplifts cored by metamorphic basement rock spanning from New Mexico to southwestern Montana (Figure 1; e.g., Coney and Reynolds, 1977; Dickinson and Snyder, 1978; Fan and Carrapa, 2014). These uplifts partitioned the foreland basin of the Western U.S. during a mountain building event broadly referred to as the Laramide Orogeny (e.g., Dickinson and Snyder, 1978). Shortening in each range on the order of several kilometers was accommodated on major range-bounding thrust faults that dip at ca. 30° beneath the ranges and extend to the mid-crust (Figure 3; e.g., Smithson et al., 1978; Sharry et al., 1986). The Beartooth Mountains comprise one of the northernmost basement-cored uplifts (Figure 1; e.g., O'Neill et al., 1996; Burger, 2004).

Subsequent compartmentalized subsidence in a peripheral basin proximally to the east-northeast of the Beartooth Mountains is recorded in the thick nonmarine deposits of the Paleocene Fort Union Formation (Gingerich, 1983; Ayers, 1986). Proximal alluvial to fluvial

facies of the synorogenic Beartooth Conglomerate member of the Fort Union Formation record an unroofing sequence related to exhumation of the Beartooth uplift, in which clasts of younger stratigraphic units generally occur near the base and clasts of older rocks, such as metamorphic basement, occur higher in the section (DeCelles et al., 1991a). Paleontologic and thermochronologic data from several studies on this unit indicate a Paleocene age of ca. 57–55 Ma (Flueckinger, 1970; Jobling, 1974; DeCelles et al., 1991a; Koenig, 2015). This conglomerate overlies an angular basal unconformity on top of Upper Cretaceous rocks, which themselves have been intensely deformed prior to conglomerate deposition (DeCelles et al., 1991b). Therefore, Paleocene deposition of this unit represents only a minimum age for uplift of the range, as significant deformation and uplift may be unrecorded due to sediment bypassing prior to Thanetian time (e.g., DeCelles et al., 1991b; Hoy and Ridgway, 1997).

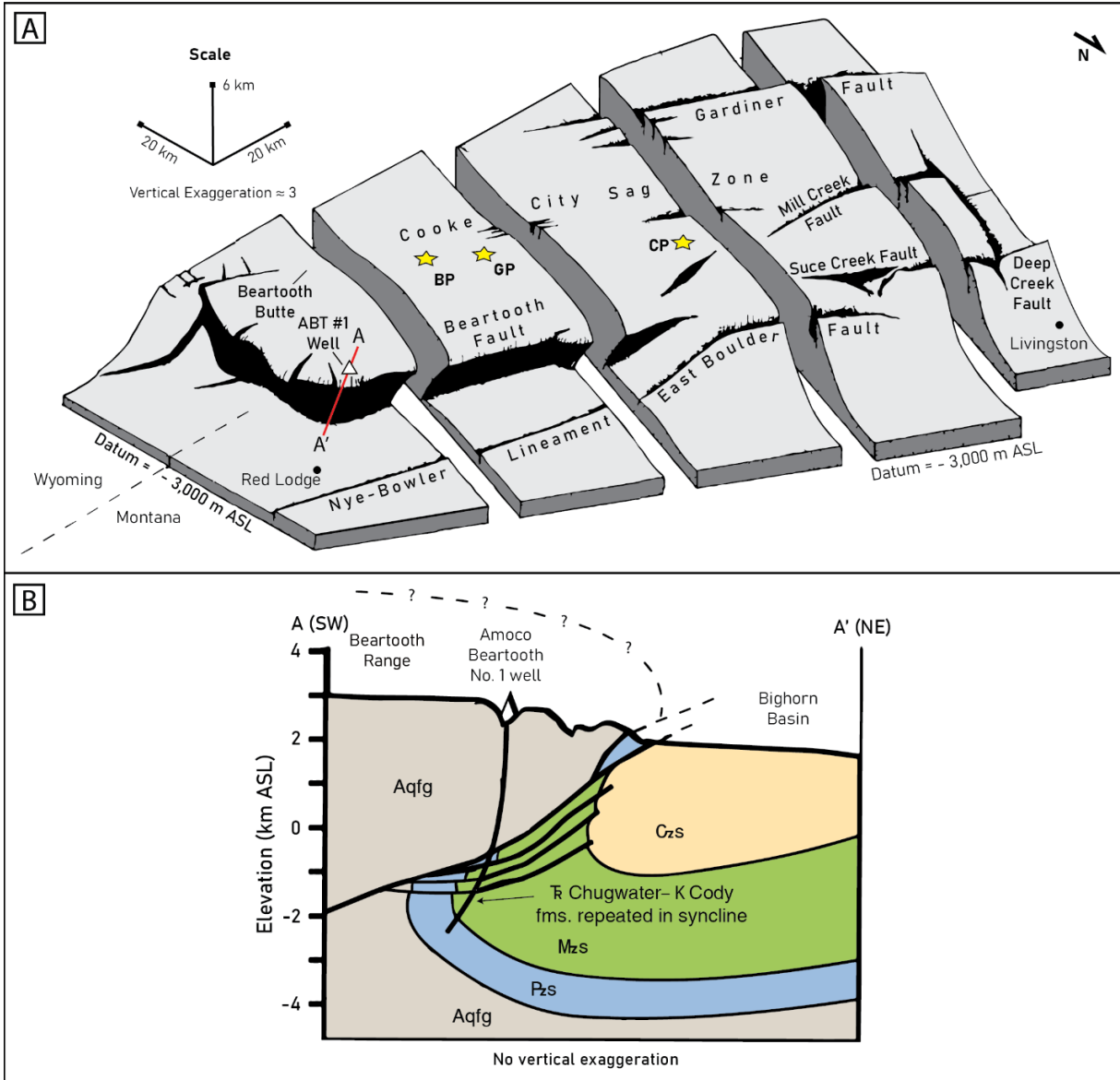


Figure 3: (A) Block diagram showing the structure of the basement surface in the Beartooth Mountains. Major faults are shown in black. Stars represent the approximate location of the highest sample from the indicated transect. Cross section line (A–A') corresponds to panel B. Vertical exaggeration is approximately 3x. ABT #1 well—Amoco Beartooth Number 1 well. Modified from Foote et al. (1961) and Wise (1997). (B) Schematic geologic cross section of the southeastern corner of the Beartooth Mountains. Cross section line (A–A') is shown in Figure 2. The Amoco Beartooth No. 1 well (white triangle) constrains structures and geologic units at depth (Wise, 1997). Cuttings from this well were sampled by previous thermochronologic studies discussed below (Omar et al., 1994; Peyton et al., 2012). Aqfg—Archean quartzofeldspathic gneiss; Pzs—Paleozoic sedimentary rock, undivided; Mzs—Mesozoic sedimentary rock, undivided; Czs—Cenozoic sedimentary rock, undivided. Modified from Peyton et al. (2012).

Previous Low-Temperature Thermochronology

Previous investigations in the Beartooth Mountains incorporating low-temperature thermochronology have focused almost exclusively on the Phanerozoic thermal history (Figure 4; Giegengack et al., 1986; Cervený, 1990; Cervený and Steidtmann, 1993; Omar et al., 1994; Mueller et al., 2008; Peyton et al., 2012; Bricker, 2016; Carrapa et al., 2019). These previous results serve as constraints on the Phanerozoic portions of our thermal history models and are summarized here.

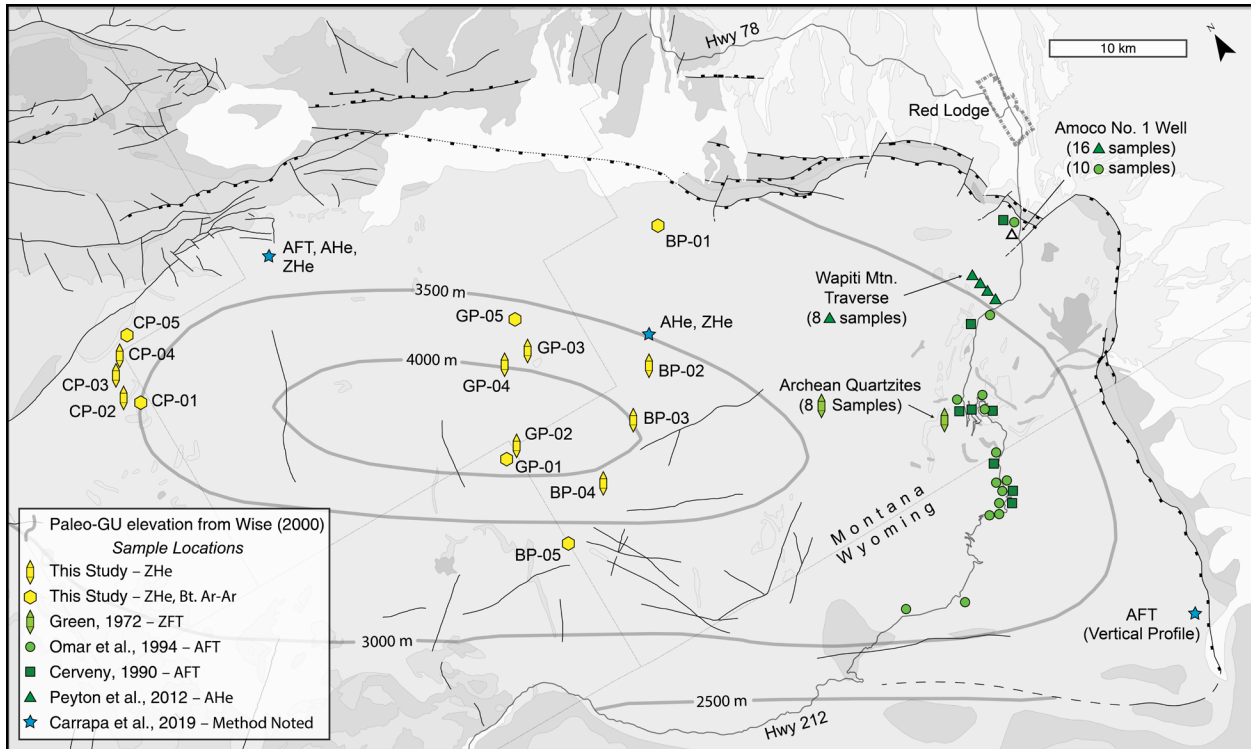


Figure 4: Location of low-temperature thermochronology samples from the Beartooth Mountains with known coordinates. Contours correspond to the interpolated elevation of the paleo-GU surface, modified from isopach maps of Wise (2000); contour interval is 500 m. Greyscale generalized geologic map is shown as base map for reference; refer to Figure 2 for interpretation of this symbology and map data sources. Samples locations of Cervený (1990) are approximated from reported elevation and general spatial description. Samples of Mueller et al. (2008) and

Bricker (2016) are not shown but generally parallel the Beartooth Highway. AHe—Apatite (U-Th-Sm)/He; ZFT—zircon fission track; other method abbreviations described in text.

The earliest thermochronology data in the Beartooth Range include ZFT dates from detrital zircons in Archean quartzites on the Beartooth Plateau, ranging from 1339 Ma to 2346 Ma (Figure 4; Green, 1972). Subsequent fission track efforts, aided by an intervening improvements in calibration of the fission-track dating method (e.g., Hurford and Green, 1983), have focused largely on the apatite fission track (AFT) system. Cervený (1990) and Omar et al. (1994) report AFT dates from sample transects near the Beartooth Highway (Figure 4). These authors interpret rapid Paleocene cooling associated with uplift of the range. Additionally, Cervený (1990) and Omar et al. (1994) document a fossil AFT partial annealing zone (PAZ) near the elevation of the Beartooth Plateau (Figure 5). The authors interpret this as evidence that these highest elevation samples experienced prolonged residence in the AFT PAZ prior to Phanerozoic exhumation, while lower samples experienced hotter temperatures (more than ca. 120 °C) during this time.

Omar et al. (1994) additionally report AFT dates from cuttings of the Amoco Beartooth No. 1 well (Figure 4). These samples suggest Cenozoic residence of lower borehole samples (near sea level and below) within the AFT PAZ (ca. 60–125 °C, depending on hold time and chemistry of the apatite crystal; Reiners and Brandon, 2006; Peyton and Carrapa, 2013 and references therein). Omar et al. (1994) interpret these results to suggest 7–12 km of uplift of the Beartooth Mountains since Late Cretaceous–Paleocene time, with uplift occurring over two stages with an intervening mid-Tertiary period of tectonic quiescence or burial. This interpretation is broadly consistent with AFT data of Giegengack et al. (1986) from the Bighorn basin and flanks of the Beartooth Mountains suggesting similar Cenozoic burial.

Peyton et al. (2012) report AHe cooling dates from 24 samples (89 aliquots) from the Amoco Beartooth No. 1 well and a pseudo-vertical transect on Wapiti Mountain, near the well (Figure 4). The authors do not interpret the preservation of a fossil AHe PRZ at high elevation in the Beartooth Mountains, with some date dispersion in the highest samples attributed to He implantation (Figure 5; Murray et al., 2014). This suggests that Phanerozoic burial temperatures must have been $> 70^{\circ}\text{C}$ to reset the AHe thermochronometer (Peyton et al., 2012). Inverse thermal modeling by the authors suggest burial to ca. $80\text{--}140^{\circ}\text{C}$ by ca. 110 Ma followed by rapid exhumation initiating between ca. 100–70 Ma and slower cooling from ca. 40 Ma to the present. Additionally, the authors calculate a modern geothermal gradient of 27°C for the Beartooth Range, adjusted from borehole measurements from the Amoco Beartooth No. 1 well using the methods of Barker (1996) and Waples et al. (2004). Mueller et al., 2008 also report AHe dates from samples collected along the Beartooth Highway, with dates ranging from ca. 120–35 Ma (Figure 5). Bricker (2016) reports additional data from a similar transect as well as samples from the North and South Snowy blocks, with AHe dates from ca. 190 Ma to near zero attributed to radiation damage, zonation, or implantation effects (Figure 5; e.g., Flowers et al., 2009; Peyton et al., 2012; Murray et al., 2014).

Carrapa et al. (2019) report AHe and ZHe results from basement samples collected at low elevations in the East Rosebud Creek and Stillwater River drainages (Figure 4). Additionally, they report AFT data from a vertical transect in the Clark's Fork River basin (Figure 4), a single basement sample from the Stillwater River drainage, and modern river sand samples from the Stillwater River and Rosebud Creek (Figure 5). Thermal modeling by the authors suggests initiation of Phanerozoic uplift in the Beartooth Mountains between ca. 120 Ma and 80 Ma.

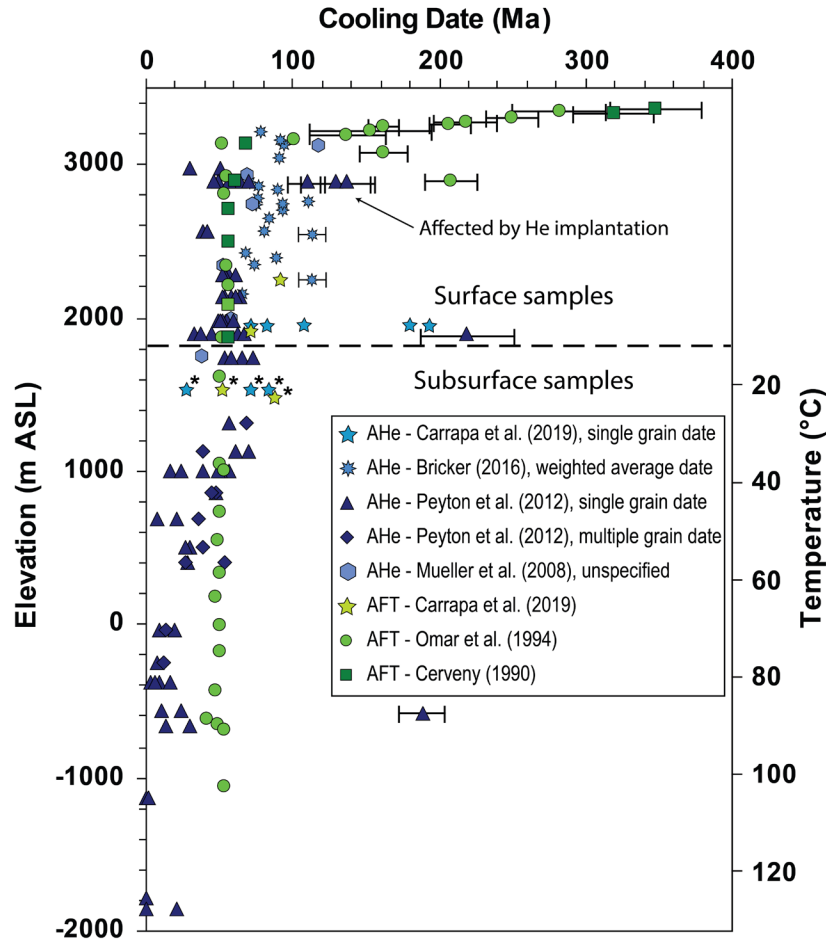


Figure 5: Existing AHe and AFT dates from the Beartooth Mountains as a function of elevation. Error bars are shown at 2σ ; for clarity, error bars are only shown for samples with error greater than ca. 20 Myr. Temperature scale from adjusted borehole measurements at the Amoco Beartooth No. 1 well (Peyton et al., 2012). Sample collection locations are shown/described in Figure 4. Aliquots interpreted to be affected by He implantation by Peyton et al. (2012) are indicated by text in figure. Surface vs. subsurface sample line reflects the elevation of samples from the Beartooth Highway transect and Amoco Beartooth No. 1 well, collared at 1904 m ASL; (*) denotes surface samples collected at lower elevation from the East Rosebud Creek drainage. Modified from Cerveny (1990), Omar et al. (1994), Mueller et al. (2008), Peyton et al. (2012) and Bricker (2016).

Methodology

Sample collection

Samples were collected from exposed quartzofeldspathic gneiss along three pseudo-vertical transects in the Beartooth Mountains at the maximum vertical offset allowable by the topography and exposure (Figure 2 and 4). Reconnaissance mapping was conducted to ensure samples were collected an adequate distance from intrusions (Figure 2). Samples were collected from below the surface of the exposed rock to minimize resetting of thermochronometers by wildfire effects (e.g., Mitchell and Reiners, 2003; Reiners et al., 2007). Five samples were collected along the East Rosebud trail (USFS trail no. 15), colloquially known as the “Beaten Path” (Figure 2; BP- prefix); 4 samples were collected along the approach to and summit of Granite Peak, the Montana state highpoint (Figure 2; GP- prefix); and 5 samples were collected along the approach to and summit of Chalice Peak (Figure 2; CP- prefix). Samples elevations ranged from 1742 m, near East Rosebud Lake, to 3909 m, the summit of Granite Peak. Detailed methodology for all analytic techniques conducted on these samples is reported in Appendix A with detailed isotopic results reported in Appendix B.

Zircon U-Pb Geochronology

U-Pb geochronology was conducted on zircons from sample GP-01, sampled at the summit of Granite Peak (Figure 2). Though extensive U-Pb geochronology has been conducted on basement rocks from the Beartooth Mountains (e.g., Wooden et al., 1988; Mueller et al., 2008, 2010 and references therein), no data exist for this location. The high T_c of the U-Pb system means that zircon U-Pb ages generally reflect crystallization of the zircon crystal (e.g., Gehrels et al., 2008). These data provide robust starting constraints for our ZHe thermal models.

Zircon grains were prepared for analysis and imaged with cathodoluminescence (CL) at the Arizona LaserChron Center (ALC; Tucson, Arizona, USA). CL images were used to select analysis spots targeting unzoned cores or rims of zircon grains. U-Pb geochronology of zircons was conducted by laser ablation multicollector inductively coupled plasma mass spectrometry (LA-MC-ICP-MS) at ALC. Errors are reported at 1σ . See Appendix A for a detailed description of analytical techniques, data reduction methods, and concordance filters.

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

$^{40}\text{Ar}/^{39}\text{Ar}$ thermochronology was conducted on biotite grains from the highest and lowest elevation samples from each transect to provide medium-temperature constraints on thermal history. Existing argon dating in the Beartooth Mountains dominantly utilizes the older K-Ar method and is concentrated on dating of intrusive dikes, which don't provide direct information on the thermal history of the gneissic basement rocks which we sampled (e.g., Condie et al., 1969; Baadsgaard and Mueller, 1973). The T_c of the biotite $^{40}\text{Ar}/^{39}\text{Ar}$ system varies substantially with grain size and composition—we use a conservative estimate of 300 °C to 425 °C to constrain our thermal models (e.g., Grove and Harrison, 1996). Unaltered biotite grains were selected from the light fraction of separates from the highest and lowest elevation samples of each transect. Single-grain aliquots were analyzed using single- and multi-step heating diffusion at the Argon Geochronology for the Earth Sciences Laboratory at Columbia University (New York, New York, USA). Errors are reported at 1σ . See Appendix A for detailed methodology.

Zircon (U-Th)/He Low-Temperature Thermochronology

Zircon (U-Th)/He thermochronology leverages the time-dependent production and temperature-dependent diffusion of radiogenic ^4He through the zircon crystal. Trace U and Th

decay to ^4He at a rate controlled by the half-life of these radioactive parent isotopes and their intermediate daughters; radiogenic ^4He escapes from the crystal lattice at higher temperatures, while it is retained at lower temperatures (e.g., Reiners et al., 2004). We use this thermochronometer to provide constraints on the low-temperature ($<300\text{ }^{\circ}\text{C}$) thermal history of the Beartooth Mountains.

Between 1 and 7 zircons grains were hand-picked from the heavy fraction of mineral separates for each sample for a total of 84 single zircon grain aliquots. ZHe aliquots were prepared at the Tectonic Sedimentology and Thermochronology (TeST) laboratory at Montana State University (Bozeman, Montana, USA). Grains were selected for maximum intrasample eU variability—which is essential to the deep-time ZHe thermochronology approach (e.g., Orme et al., 2016). We used the visual metamictization of grains as a proxy for accumulated radiation damage (and eU concentration), following the methods of Ault et al. (2018), such that clear grains reflect low levels of radiation damage and fully opaque grains represent high levels of radiation damage. Selected grains were photographed, measured to determine surface area to volume ratios for the alpha ejection correction, and packed into Nb foil tubes. Isotopic analysis was conducted at the Helium Analysis Laboratory (HAL) at the University of Illinois, Urbana–Champaign (Urbana, Illinois, USA). Uncertainties on individual ZHe dates—based on propagated analytical uncertainties of U, Th, and He measurements—are reported at the 2σ level. See Appendix A for detailed description of methodology.

Deep-Time Thermochronology from ZHe Date-eU Correlations

ZHe data comprise the core constraints on our thermal history modeling efforts. The temperature range of He retention within the zircon crystal is highly variable and dominantly

dependent on grain-specific properties, such as grain size (Reiners et al., 2002) and accumulated radiation damage to the crystal lattice (Guenther et al., 2013; Ginster et al., 2019). Of these factors, radiation damage exerts the strongest control on He retentivity in samples with very long low to medium-temperature thermal histories (Guenther et al., 2013). Heating of the zircon crystal can anneal accumulated radiation damage, which consequently affects He retentivity (Guenther et al., 2013; Ginster et al., 2019). Owing to this complex interplay between parent isotope concentrations, radiation damage accumulation and annealing, and resulting effects on He retentivity, each zircon crystal of a given level of accumulated radiation damage develops a unique “effective” PRZ and closure temperature and consequently records a unique cooling date (Guenther et al. 2013). Even in zircon crystals from the same rock sample—where all grains must share a common thermal history—an array of dispersed ZHe dates is commonly obtained from grains of different parent isotope concentrations (e.g., Orme et al., 2016).

We note that this date dispersion was considered a debilitating limitation by early practitioners of this technique, who sought to obtain a single “true” age of geologic events from this chronometer (e.g., Strutt, 1906). However, advances in the understanding of the source of this dispersion reveals that this represents ‘a feature, not a bug’ of the system, allowing the extraction of nuanced thermal histories (e.g., Guenther et al., 2013). This underlines that the notion of a single “true” age from which observed ages are dispersed and toward which we would like to correct is largely misguided. When dispersion is considered in the context of radiation damage effects, all ZHe dates are equally valid and contain important information about the thermal history. Consequently, a mean age of a dataset—or a “true” ZHe age of a sample—has little geologic significance. It is partially for this reason we use “date” rather than

“age” herein to refer to that point of time in the past calculated from isotopic decay to ^4He in zircon.

In datasets where cooling date is strongly influenced by radiation damage, this date dispersion manifests as a positive or negative correlation between single-grain ZHe date and eU (Guenther et al., 2013). This relationship develops when the time-temperature (t - T) history is characterized by long residence of the sample at temperatures permitting accumulation of radiation damage—and consequent development of variable He retentivities—followed by moderate reheating and/or slow cooling through the ZHe PRZ; the specific t - T history of a sample controls the shape of the resultant date-eU correlation (Guenther et al., 2013; Ginster et al., 2019; Guenther, 2021). Consequently, only a relatively small number of potential t - T histories are capable of producing an observed date-eU correlation. Even some information about the t - T history of the sample prior to the oldest observed cooling date is contained within ZHe datasets with well-defined date-eU correlations; ancient portions of the t - T history have an influence on ultimate radiation damage accumulation and thus the resultant relationship between date and eU (e.g., Powell et al., 2016; Orme et al., 2016; Boone et al., 2018). For these reasons, “deep-time” thermal modeling of date-eU correlations in ZHe datasets is particularly salient in reconstructing the complex long-term thermal histories of Precambrian basement rocks where other geologic evidence of thermal history has been overprinted or eroded (e.g., Orme et al., 2016; DeLucia et al., 2018; McDannell et al., 2019; Flowers et al., 2020; Reade et al., 2020).

Modeling Methods

Phanerozoic Decompressed Sediment Accumulation Modeling To better understand the timing and magnitude of Paleozoic burial and consequent reheating of basement rocks, we

construct a set of decompacted sediment accumulation curves. Decompacted thickness estimates can aid in assessment of the geologic plausibility of thermal history solutions (e.g., Thacker et al., *in press*). We use the minimum and maximum preserved sedimentary thicknesses on the flanks of the Beartooth Mountains as input (Maughan, 1989, 1990; Lopez, 2001). Additionally, we input initial porosity values and coefficients of porosity-depth relation from Sclater and Christie (1980) for the bulk lithology of various units (averaged when multiple modal lithologies are present). We use the sediment decompaction approach of Allen and Allen (1990) as implemented in MATLAB by Cardozo (OSXBackstrip v.1; e.g., Gómez et al., 2005).

We model maximum decompacted burial depths for our samples by adding estimated paleo-basement overburden to our maximum decompacted sediment accumulation curve. Estimated overburden was calculated by removing the estimated thickness of Cambrian–Devonian rocks from modeled base-Mississippian isopach contours of Wise (2000) to calculate the paleo-GU elevation. We then subtracted the average elevation of each sample elevation group (as defined in results) from the calculated paleo-GU elevation. Isopachs of Wise (2000) were interpolated from modern exposures of Paleozoic rocks and assume no major faults between these exposures and the core of the Beartooth Range and a relatively symmetrical fault-propagation fold shape of the uplift—assumptions consistent with mapping data (Figure 2; e.g., Foote et al., 1961; Lopez, 2001). This estimate predicts a paleo-GU surface nearly coincident with the modern Beartooth Plateau, with <1 km of basement overburden estimated to have been removed during Phanerozoic uplift and erosion. This is supported by interpretations of Simons and Armbrustmacher (1976) based on mapping and topographic features of the Beartooth Plateau surface. Restored cross sections of DeCelles et al. (1991b) likewise predict minimal difference in

elevation between the modern Beartooth Plateau and the paleo-GU surface in the southeastern portion of the Beartooth Mountains.

Contrary to the above, Omar et al. (1994) estimate substantially greater depth (ca. 3 km) below the paleo-GU surface for the Deep Lakes area of the Beartooth Mountains from thermal modeling of AFT samples (locations in Figure 4). We additionally model a second set of maximum decompacted burial depths using their estimates. In all scenarios, we assume negligible compaction of the Precambrian basement. These decompacted sample depth curves inform the Phanerozoic portion of our forward thermal modeling.

Thermal History Modeling We use thermal history modeling to assess the ability of potential t - T paths (thermal history hypotheses) to reproduce observed ZHe date-eU correlations within our dataset. This approach allows us to investigate the reproducibility of observed data from predictions of hypotheses for major thermal events affecting the Beartooth Mountains and constrain aspects of the thermal history of this region not resolved by previous thermochronologic studies. Our forward models utilize the approach of Orme et al. (2016) and Guenthner (2021). This approach implements a MATLAB script incorporating a portion of the relevant code from the program HeFTy (i.e., He diffusion modeling and alpha ejection correction; Ketcham, 2005b) optimized for testing predicted date-eU correlations against a large number of single-grain inputs (publicly available at <https://github.com/wrguenthner/DAAM>). This facilitates visual inspection of the date-eU correlation predicted by a modeled t - T history. Visual inspection is essential when modeling ZHe datasets with high date dispersion; this allows for the assessment of the effect of well-constrained sources of date variability, which are considered by the model (e.g., grain size and radiation damage effects) versus those of less well-

constrained sources. These other sources of date variation—such as zonation of parent isotopes (Ketcham, 2005a; Danišik et al., 2017), implantation of He from neighboring phases (Peyton et al., 2012; Murray et al., 2014), and, to some extent, fragmentation of grains (Brown et al., 2013)—are less well quantitatively characterized and are not well accounted for by most models. Our forward modeling implementation uses the zircon radiation damage and annealing model (ZRDAAM) of Guenthner et al. (2013). To model predicted ZHe date-eU correlations, ZRDAAM combines the radiation damage-diffusivity relationship (Guenthner et al., 2013) with a damage annealing model, defined by a fanning curvilinear fit to the ZFT annealing data of Yamada et al. (2007). These annealing kinetics give a ZFT PAZ of 310–223 °C (0.4–0.8 mean length reduction ratio; 10 Ma isothermal hold-time; Guenthner et al., 2013).

A different radiation damage and annealing model has been developed by Ginster et al. (2019). This annealing model was developed using Raman spectroscopy data—rather than etchable zircon fission tracks (as by Guenthner et al. 2013)—as a proxy for bulk radiation damage. Guenthner (2021) demonstrates that the differing kinetics of Guenthner et al. (2013) and Ginster et al. (2019) manifest as significant differences in predicted date-eU correlations in grains of certain eU for some t - T histories. Guenthner (2021) found that the kinetics of Ginster et al. (2019) do not capture the behavior of grains at very high damage levels in certain data sets and, in some cases, produce worse fits to the observed data. Due to the large number of grains with relatively high-eU concentration in our dataset (i.e., > 800 ppm eU), the bulk of our modeling incorporates the annealing kinetics of Guenthner et al. (2013). We additionally model our refined thermal history solutions using the kinetics of Ginster et al. (2019) as implemented in

MATLAB by Guenthner (2021) to illustrate some of these differences. Results of these models are reported in Appendix D.

Our forward model takes grain size, eU, and t - T paths as input (provided in Appendix B and C). To account for grain size effects on ZHe date, we model three separate date-eU curves for each input t - T path using the mean grain size $\pm 2\sigma$ from measurements of our dataset. The resulting model output is a predicted date-eU envelope with a central curve that shows the extent to which the combined effects of grain size and radiation damage explain the date-eU correlation and variation of our observed dataset. We first model “end-member” Proterozoic scenarios, using the extreme limits of t - T paths permissible by geologic constraints. We then modulate the Phanerozoic portion of the thermal history, testing hypotheses for the timing and magnitude of burial and exhumation. This process allows us to iteratively assess the effect of modifying the timing, rate, and temperature of various thermal events on the predicted ZHe date-eU correlation. This facilitates understanding of the resolution and sensitivity of our models and avoids overinterpretation of the data, which can lead to overly complex models unsupported by—or at odds with—other geologic data.

Following forward model exploration, we utilize inverse modeling to further refine possible thermal histories (e.g., DeLucia et al., 2018; Reade et al., 2020). Inverse modeling facilitates the testing of tens of thousands of potential thermal histories very quickly. We use HeFTy v. 1.9.3 (Ketcham, 2005a), which implements an inverse Monte Carlo search method—testing random thermal histories that intersect defined constraint boxes and assessing the statistical fit of predicted to observed ZHe dates. The model takes eU concentrations, grain radii, measured ages, and age uncertainties as input. HeFTy v. 1.9.3 takes a maximum of 7 grains as

input; due to the large size of our dataset, we group grains defining similar portions of our observed date-eU trend and average input parameters within each group, defining a “synthetic grain”. The date uncertainty of our synthetic grains is equal to the standard deviation of the input dates, except at highest eU (>1900 eU), where uncertainty is extended to include a ZHe date of zero. This modification is necessary because an age uncertainty not inclusive of zero date at high-eU causes HeFTy to fail to find acceptable fits, as we experienced with large (100,000 path) model runs (reported in Appendix D). This is likely due to the poor present understanding of diffusion and annealing kinetics in very damaged zircon crystals, and consequently the inability to reliably predict observed dates at high-eU in thermal models (e.g., Guenthner et al., 2013).

Inverse models explore 50,000 paths each. Output “acceptable” and “good” paths have a goodness-of-fit criteria of >0.05 and >0.05 , respectively. We model upper and lower elevation sample groups separately, as defined in our results section. We then input the best-fit output thermal history into our MATLAB forward model to visually assess the fit of the predicted date-eU envelope to our observed dataset. Complete model constraints and justification, following the methods of Flowers et al. (2015), and parameters of synthetic grains are available in Appendix C. Finally, we forward model a “refined” thermal history. This model incorporates insights from our inverse models with information gleaned from our previous forward models about the sensitivity of various portions of the predicted date-eU trend to thermal events. Here, we also incorporate additional AFT results from Omar et al. (1994) to arrive at a thermal history most compatible with ZHe data from both elevation groups in our dataset as well as existing thermochronologic and geologic data.

While it is relatively common to forward model a dataset prior to inverse modeling (e.g., Peyton et al., 2012; DeLucia et al., 2018; Reade et al., 2020), few studies return back to full exploration of the t - T space defined by good-fit inverse results using forward modeling. In deep-time thermal modeling of ZHe datasets, non-unique solutions are often abundant and model results from input “synthetic grains” may fail to capture important facets of the observed date-eU correlation. This final forward modeling step removes some of the “black box” aspects of inverse modeling and allows a qualitative assessment of the non-unique model solutions represented by good-fit paths and the agreement of their predicted date-eU correlation with observed data. Additionally, geologic data and frameworks that can’t (or shouldn’t, for risk of over-constraining; e.g., Vermeesch and Tian, 2014) be directly incorporated as constraints on inverse models—such as preserved stratigraphy and existing tectonic frameworks—can help narrow this space. This final forward model exploration can facilitate identification of thermal history solutions most parsimonious in consideration of these data.

Analytic Results

Zircon U-Pb Results

Zircon U-Pb ages were determined for 50 grains from the summit of Granite Peak (Sample GP-01). We obtained 11 analyses meeting concordance cutoffs (<20% discordance, <5% reverse discordance). Ages meeting concordance cutoffs (retained analyses) were Neo–Mesoarchean—ranging from 2688.97 ± 12.87 Ma to 3124.01 ± 38.38 Ma. Most retained analyses were centered around ca. 2800 Ma, and only analyses of this age intersect concordia (Appendix B). Most analyses targeted either unzoned grains or cores of zoned grains. A retained analysis targeting a low-U rim yielded an age of 2934 ± 32.92 Ma, suggesting minimal

Neoproterozoic or younger metamorphic growth of zircon. Th/U ratios range from 0.04–0.56 with a weak positive correlation between Th/U and U-Pb age. The probability distribution of ages is defined by a peak at ca. 2800 Ma with a second broader peak centered at ca. 2950 Ma and a tail at lower probability to ca. 3200 Ma (Appendix B). These results are consistent with previous zircon U-Pb geochronology results from the Beartooth Mountains (e.g., Wooden et al., 1988; Mueller et al., 2008). The bulk of ages coincide with the volumetrically dominant ca. 2.8–2.9 Ga magmatic suite in the range, while older ages may reflect inherited zircons from the Mesoproterozoic TTG suite (Mueller et al., 2008 and references therein). See Appendix B for detailed isotopic results and figures.

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

Biotite $^{40}\text{Ar}/^{39}\text{Ar}$ ages were determined for the highest and lowest sample of each transect, for a total of 6 samples. We report both single-step diffusion bulk ages and plateau ages from step-heating diffusion experiments for separate aliquots of each sample (Table 1). We report single-step diffusion bulk ages to demonstrate general agreement with the plateau ages, but dominantly consider plateau ages for our model constraints and interpretation. See Appendix A for details on plateau age assessment and uncertainty calculation and Appendix B for detailed isotopic results and $^{40}\text{Ar}/^{39}\text{Ar}$ gas release spectra.

Biotite $^{40}\text{Ar}/^{39}\text{Ar}$ samples yielded plateau ages ranging from 1622 ± 121.22 Ma to 2402 ± 2.11 Ma (Table 1). Sample ages comprise two distinct groups that do not overlap with one another at 1σ uncertainty: (1) samples from the westernmost transect (Chalice Peak) yielded ages between 1662 ± 121.22 and 1775 ± 27.08 Ma; (2) samples from the eastern two transects

(Granite Peak and East Rosebud trail) yielded ages between 2263 ± 75.44 Ma and 2402 ± 2.11 Ma (Table 1).

Sample	Plateau Obtained?	Plateau Age (Ma)	SEM	MSWD	RCS	Bulk Age (Ma)	Uncertainty $\pm 1\sigma$ (Ma)
			Uncertainty $\pm 1\sigma$ (Ma)		Uncertainty $\pm 1\sigma$ (Ma)		
BP-05	Yes	2402	2.11	4.63	N/A	2354.83	3.22187
BP-01	Forced	2263	5.08	222.47	75.77	2277.97	2.82425
GP-05	Forced	2276	2.85	108.30	29.66	2472.72	3.05836
GP-01	Yes	2369	0.26	0.57	N/A	2387.34	3.02078
CP-05	Forced	1622	10.34	137.43	121.22	1796.73	3.22795
CP-01	Forced	1775	3.14	74.38	27.08	1663.34	2.41543

Table 1: Biotite $^{40}\text{Ar}/^{39}\text{Ar}$ ages and associated error from the Beartooth Mountains. Forced plateau age error was calculated using the reduced chi-squared statistic (RCS).

ZHe Results

ZHe dates were determined for 14 samples from 3 range-perpendicular transects in the Beartooth Mountains (Figure 1). We dated between 1 and 7 single-grain aliquots per sample for a total of 84 individual ZHe dates (see Appendix B for detailed analytic data). ZHe dates range from 686.43 ± 11.88 Ma to 13.54 ± 0.25 Ma while eU concentrations span from 113–3565 ppm. ZHe dates correlate strongly with eU, a proxy for radiation damage (Figure 6A). The correlation is strongly negative for eU values less than ca. 800 ppm. Grains of higher eU form a pediment of relatively restricted ages of ca. 10 Ma to 110 Ma before reaching near-zero dates at ca. 1300–2000 ppm eU (Figure 6A). This is consistent with moderate to high levels of accumulated damage ($>\text{ca. } \alpha/\text{g} \times 10^{17}$), which produce a negative date-eU correlation (Guenther et al., 2013).

ZHe dates in our dataset show only a weak positive correlation with grain size (Figure 6B). Additionally, a weak correlation between ZHe date and structural position is discernable, except for sample BP-05, which lies off this trend (Figure 6C). There is little correlation between $^{40}\text{Ar}/^{39}\text{Ar}$ age and ZHe date (Figure 6D). This is also reflected in the similar date-eU correlation defined by samples from our western transect (Chalice Peak; green triangles in Figure 6A), where $< 1.8 \text{ Ga}$ $^{40}\text{Ar}/^{39}\text{Ar}$ ages are observed, and samples from the eastern transects (East Rosebud trail and Granite Peak; red circles and blue squares in Figure 6A), where $> 2.27 \text{ Ga}$ $^{40}\text{Ar}/^{39}\text{Ar}$ ages are observed.

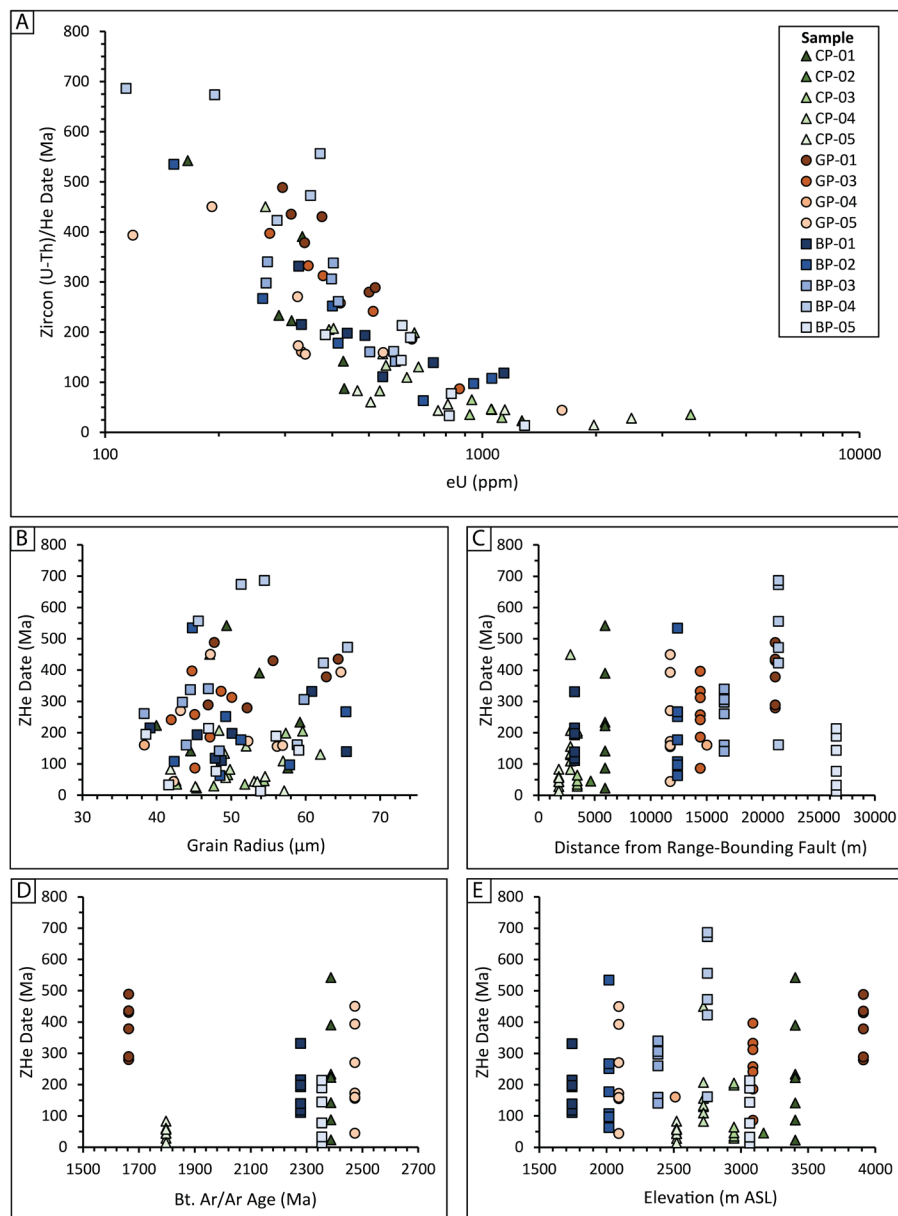


Figure 6: Zircon (U-Th)/He data from three transects in the Beartooth Range: Chalice Peak (green triangles); Granite Peak (orange circles); and the East Rosebud trail (blue squares). ZHe dates are plotted against variables potentially responsible for date dispersion: (A) eU concentrations (ppm); (B) grain size (μm ; equivalent spherical radius); (C) distance from the main range-bounding thrust fault (m); (D) biotite $^{40}\text{Ar}/^{39}\text{Ar}$ cooling age (Ma); and (E) elevation (m) above sea level.

Elevation initially does not appear to have a strong first-order correlation with ZHe date (Figure 6E). However, a second-order control by elevation on ZHe date is evident by plotting this as a secondary independent variable with eU (Figure 7A). The shape of the date-eU curve differs between samples from higher elevations versus from those of lower elevations, with a break in the slope of the date-eU-elevation trend occurring at ca. 2300 m ASL (Figure 7A). At eU values < 800 ppm, samples from upper elevations (> 2300 m) define a steeper date-eU curve than those from lower elevations (< 2300 m; Figure 7B). The upper elevation group of samples additionally define a younger pediment (ca. 30–90 Ma) than the lower elevation group (ca. 100–120 Ma) between ca. 800–1000 ppm eU (Figure 7B). The upper elevation group reaches youngest dates (13.54 Ma) at lower eU (ca. 1290 ppm) than the low elevation group (ca. 44 Ma at 1625 ppm; Figure 7B).

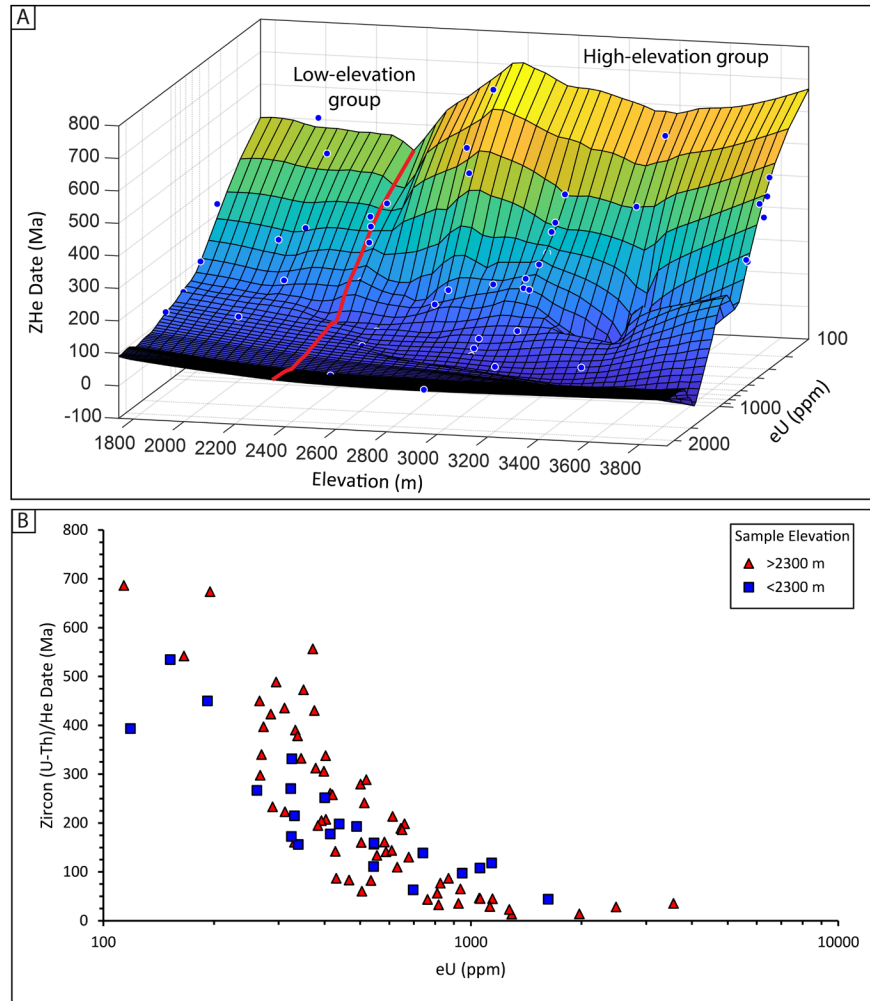


Figure 7: Sample elevation exerts a second-order control on ZHe date, defining two sample elevation groups. (A) 3-D plot showing ZHe date as a function of eU and elevation. Blue dots are measured ZHe data, to which a surface was fit using polynomial locally weighted scatter plot smoothing in MATLAB. Additional ZHe data points lie below the surface. The 2300 m contour on this surface—which we identify as the break in slope between high-elevation sample group and low-elevation sample group date-eU curves—is highlighted in red. Color gradient shows ZHe date and is meant as a visual aid. (B) ZHe date-eU correlations shown for two sample groups at > 2300 m and < 2300 m elevation, respectively. Low elevation samples (blue squares) define a more gradual curve than high elevation samples (red triangles).

Modeling Results

Decompacted Sediment Accumulation Results

Our decompacted sediment accumulation analysis shows that the bulk of preserved stratigraphy on the flanks of the Beartooth Mountains record maximum Phanerozoic burial during latest Cretaceous to earliest Paleocene time (Figure 8). Only ca. 1.5 km decompacted thickness of Paleozoic stratigraphy is preserved. However, a large unconformity separates Lower–Middle Pennsylvanian rocks from overlying sparsely preserved Triassic rocks of the Chugwater Formation or Jurassic rocks of the Ellis Group and stratigraphy may be missing across this unconformity (Figure 8; e.g., Maughan, 1990). Maximum decompacted sediment thickness totals ca. 5.5 km by 57 Ma, following deposition of thick successions of mostly Upper Cretaceous rocks (Figure 8). It is important to note that the entirety of this sedimentary thickness likely never accumulated at the location of the modern Beartooth Mountains—uplift of the range initiated by at least Paleocene time, as evidenced by deposition of the Thanetian Beartooth Conglomerate along the northeastern margin of the range (e.g., DeCelles et al., 1991a)—but is plotted here for completeness.

We estimate a maximum decompacted Paleozoic burial depth of ca. 2 km for our upper group (0.5 km depth below paleo-GU surface) and ca. 3 km for our lower sample group (1.5 km depth below paleo-GU surface) using paleo-basement overburden values calculated from isopachs of Wise (2000; Figure 8). Assuming 20 °C surface temperatures and a 30 °C geothermal gradient, our model predicts maximum Paleozoic burial temperatures of ca. 80 °C and 110 °C for our upper and lower sample groups, respectively. Assuming no intervening uplift of the basement, we model a maximum burial depth of ca. 4.5 km and ca. 5.5 km by the end of

Montana Group sedimentation for our upper and lower sample groups, respectively, by which time uplift of the range had initiated (DeCelles et al., 1991b; Omar et al., 1994). This translates into maximum possible burial temperatures of ca. 150 °C and 180 °C using the assumptions outlined above (Figure 8). We also model a maximum decompacted burial depth curve using estimates of 3–4 km of basement overburden (green lines in Figure 8) derived from AFT modeling of samples collected near the Beartooth Highway by Omar et al. (1994).

We note that very few Phanerozoic sedimentary rocks are preserved in the Beartooth Mountains. Consequently, this sedimentary accumulation model assumes that the Beartooth Mountains accumulated a similar amount of sediment to that preserved in nearby regions. This assumption may not hold for some or most periods of deposition. For instance, previous work suggests the Beartooth region may have formed an intermittent trough or high during portions of Paleozoic and Mesozoic time (e.g., DeCelles, 1986; Maughan, 1993). Despite this, our decompacted sedimentary thickness model provides a framework to assess the compatibility between modeled thermal histories and burial recorded by preserved stratigraphy.

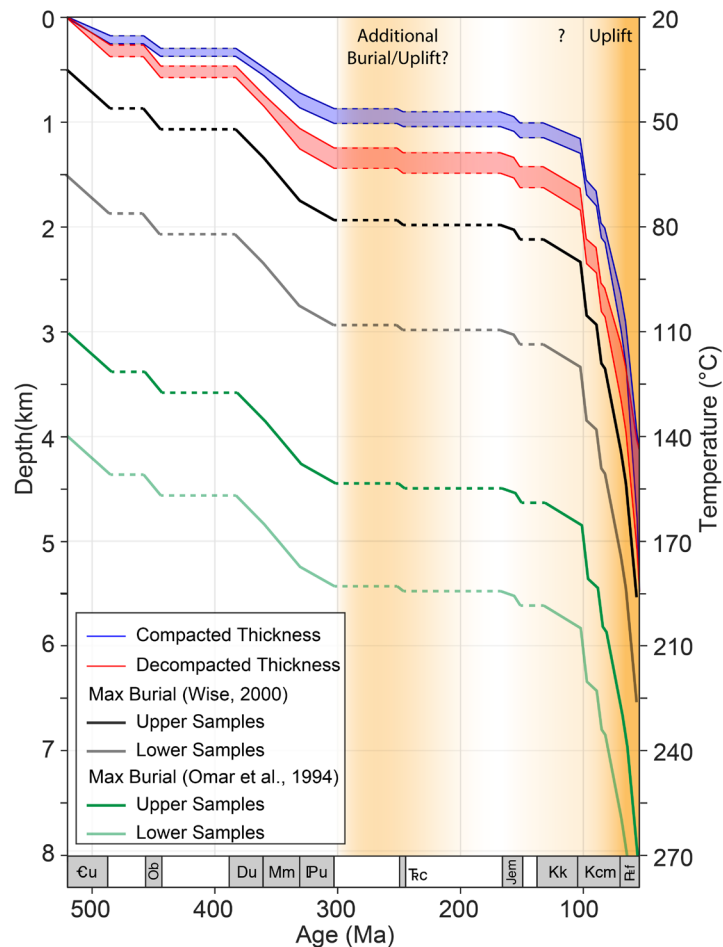


Figure 8: Decompacted sediment accumulation envelopes for stratigraphy near the Beartooth Mountains. The upper line of each envelope reflects the compacted (blue) or decompacted (red) sedimentary thicknesses derived from the minimum thickness of preserved stratigraphy in nearby areas, while the lower line reflects that derived from the maximum; the region between these estimates is shaded. Dashed lines indicate time for which no stratigraphy is preserved. Estimated maximum depth curves for our sample elevation groups reflect basement overburden estimated by Wise (2000; grey lines) and Omar et al. (1994; green lines). Sedimentary thicknesses were tabulated from Maughan (1993) and Lopez (2001) and references therein. Orange shading reflects timing of certain (dark shading) or possible (light shading) uplift of the Beartooth Mountains. Temperature scale assumes 20 °C surface temperature and a consistent 30 °C/km geothermal gradient. Shaded regions at bottom represent periods of deposition recorded by: Cu—Cambrian undivided, Flathead through Snowy Range formations; Ob—Middle Ordovician Bighorn Dolomite; Du—Middle–Upper Devonian undivided, Maywood through Three Forks formations; Mm—Mississippian Madison and Big Snowy groups; IPu—Lower–Middle Pennsylvanian undivided; Amsden Group through Quadrant Sandstone; Trc—Triassic Chugwater Formation; Jem—Jurassic Ellis Group and Morrison Formation; Kk—Lower Cretaceous Kootenai Formation; Kcm—Upper Cretaceous Colorado and Montana groups; Pef—

Paleocene Fort Union Formation. Stratigraphic nomenclature reflects that used in central Montana; see Maughan (1993) for equivalent formations and nomenclature of the Bighorn Basin.

Thermal History Modeling

Result-informed Model Setup We use forward modeling of date-eU correlations to explore the t - T space constrained by our ZHe data. For effective thermal modeling, samples likely sharing a broadly similar thermal history are commonly grouped; this can be determined by assessing the relative shape of date-eU correlation between samples and the influence (or lack thereof) of other variables that may affect ZHe date dispersion (Figure 6–7; e.g., Powell et al., 2016; Orme et al., 2016; Boone et al., 2018; Reade et al., 2020).

Of the variables potentially influencing ZHe date examined above, only sample elevation has a systematic influence in our dataset (Figure 6–7); samples above and below ca. 2300 m ASL define two sample elevation groups with different date-eU correlations, respectively (Figure 7). These differences in date-eU trend likely manifest due to different depths—and thus temperatures—experienced by samples at each elevation over the course of their thermal history (Figure 8). The relatively distinct grouping may result from differential residence with respect to the ZFT PAZ and/or ZHe PRZ. For instance, if upper elevation samples resided above the ZFT PAZ for long duration prior to pre-GU erosion while lower elevation samples resided below or within, differential radiation damage annealing—and thus effective T_c and He retention—would be expected between sample groups (e.g., Guenthner et al., 2013).

A similar negative date-eU correlation exists for analyses of all transects, defining different trends only when elevation is considered (Figure 7B). The fact that the shape of this curve does not strongly differ between samples within each elevation group—and that these samples are not separated by major mappable faults—suggests that all samples within an

elevation group share a broadly similar thermal history and can be grouped in model runs. For these reasons, our models combine single-grain aliquots from individual samples into two sample groups based on elevation—as defined by the break in slopes in our date-eU-elevation plot (Figure 7; ca. 2300 m ASL). Our end-member models fit a single date-eU envelope to both elevation groups to test the general fit of modeled date-eU trends predicted from very different thermal histories. Our subsequent forward and inverse models fit each elevation group separately to test the finer-scale effects of more subtle differences in thermal histories and facilitate testing of paleo-geothermal gradient hypotheses. To account for grain size effects, our forward thermal models incorporate a 2σ grain size envelope.

We use zircon U-Pb and biotite $^{40}\text{Ar}/^{39}\text{Ar}$ cooling ages to constrain the Archean–Paleoproterozoic portions of our thermal models. Zircon U-Pb results collected from our Granite Peak transect resolve similar ages to those collected near other transects (e.g., Mueller et al., 2008; Carrapa et al., 2019). For this reason, we start all models at 600–900 °C at ca. 2.9–2.8 Ga, consistent with crystallization of the most voluminous suite of rocks in the Beartooth Mountains. Our biotite $^{40}\text{Ar}/^{39}\text{Ar}$ results constrain our samples to ca. 300–425 °C (a conservative closure temperature window for biotite $^{40}\text{Ar}/^{39}\text{Ar}$; e.g., Grove and Harrison, 1996) between ca. 1620–2400 Ma—the range of our observed biotite $^{40}\text{Ar}/^{39}\text{Ar}$ ages. While biotite $^{40}\text{Ar}/^{39}\text{Ar}$ ages differ by up to ca. 800 Myr between our western and eastern transects, the similarity in the shape of date-eU correlation between samples of our western transects (< 1.8 Ga $^{40}\text{Ar}/^{39}\text{Ar}$ ages; red circles and blue squares in Figure 6A) and eastern transects (> 1.8 Ga $^{40}\text{Ar}/^{39}\text{Ar}$ ages; green triangles in Figure 6A) suggests that differential Paleoproterozoic thermal histories between these samples have little effect on the thermal history recorded by the ZHe system. Although the

rocks of the western transect almost certainly experienced different Paleoproterozoic temperatures than those of the eastern transect, this similarity in date-eU correlation implies that this history has little effect on ZHe date and these samples can be modelled as having experienced a common Paleoproterozoic thermal history. This additionally implies that ZHe data place few constraints on this portion of the thermal history (>ca. 1.6 Ga; see demonstration in Appendix D); however, we start our thermal models with Archean crystallization for consistency with observed U-Pb ages.

Other aspects of the t - T history of our models are constrained by new and existing data indicated in Figure 9A and discussed in the geologic setting section above. We focus our preliminary forward models on exploration of various end-member Proterozoic cooling histories within the bounds of these constraints. Our subsequent forward models explore various specific hypotheses for Phanerozoic cooling. Together, these models demonstrate the sensitivity (and lack thereof) of aspects of the predicted date-eU trend to differing timings and magnitudes of thermal events. These events are explored in detail in subsequent inverse and “refined” thermal models, on which the same t - T constraints are imposed.

Proterozoic Scenarios and Results Our Proterozoic models test whether earliest or latest Proterozoic cooling can best reproduce the general shape of the date-eU trend of our measured data (Figure 9). These end-member scenarios test the approximate minimum and maximum amount of time that our samples could have spent at near-surface temperatures in the Proterozoic—as constrained by existing geologic and geo/thermochronologic data (red boxes in Figure 9 and references in figure caption)—as well as a middle case with Proterozoic residence in the ZHe PRZ. Our forward model scenarios here are intentionally simplistic, designed to test

the sensitivity of ZHe date/eU correlation to end-member Proterozoic thermal histories. More detailed Proterozoic thermal histories are explored in greater depth in the inverse model results.

Scenario 1 cools from 300 °C to 40 °C from 1622–1520 Ma, the earliest allowable by our $^{40}\text{Ar}/^{39}\text{Ar}$ results assuming moderate cooling rates (ca. 3 °C/Myr). By contrast, scenario 3 explores latest Proterozoic cooling, from 300 °C to 40 °C at 640–540 Ma, the latest allowable by Cambrian deposition of the Flathead Sandstone (Figure 9A). Scenario 2 explores an intermediate hypothesis of cooling from 300 °C to 180 °C at 1622–1400 Ma. We use a Proterozoic upper mid-crustal residence temperature of 180 °C to simulate long-term holding in the conventional ZHe PRZ, which allows the ZHe system to not fully reset prior to more rapid Proterozoic exhumation. Higher residence temperatures would more fully reset ZHe ages, partially defeating the purpose of modeling mid-crustal temperatures (e.g., Guenthner et al., 2013). Following long-duration holding at intermediate temperatures, this scenario cools to surface temperatures between 800 Ma and 715 Ma. This cooling timing explores an intermediate option to the other two scenarios.

We note that our modeled Paleoproterozoic thermal history is likely an oversimplification, as regional studies correlate <1.8 Ga $^{40}\text{Ar}/^{39}\text{Ar}$ and $^{40}\text{K}/^{39}\text{Ar}$ ages with a reheating—rather than cooling—event (e.g., Giletti, 1966; Condit et al., 2015). However, this model simplification is justified due to the lack of correlation between biotite $^{40}\text{Ar}/^{39}\text{Ar}$ age and ZHe date in our dataset (Figure 6D). The Phanerozoic portion of these end-member models is held constant (Figure 9A). We use a maximum Phanerozoic burial temperature of 140 °C at 70 Ma as a compromise between variable previous interpretations (e.g., Omar et al., 1994; Peyton et al., 2012; Carrapa et al., 2019) and for consistency with results of our decompacted sedimentary

thickness models (Figure 8). We explore alternative Phanerozoic burial and cooling histories in our Phanerozoic models below.

Date-eU envelopes predicted using these input t - T histories are shown in Figure 9B. These tests show that the low-moderate portion of the date-eU trend is most sensitive to timing of Proterozoic cooling. In scenario 1, rapid cooling causes grains at lowest eU to record Mesoproterozoic dates, while the long duration spent at surface conditions does not allow annealing of Proterozoic radiation damage. Grains of moderate–high-eU consequently accumulate high amounts of damage—thus attaining lower effective closure temperatures—and are subsequently partially or fully reset by Phanerozoic burial, giving younger dates (envelope I in Figure 9B). In scenario 3, Proterozoic residence at temperatures above the ZFT PAZ facilitates annealing of nearly all Proterozoic radiation damage. Consequently, low-moderate eU grains record only rapid latest Proterozoic cooling and only very high-eU grains are reset by Phanerozoic burial (envelope II in Figure 9B).

Scenario 2 is the best fit to our data (envelope II in Figure 9B). Long-duration holding in the ZHe PRZ allows moderate accumulation of ^4He and radiation damage. Consequently, only the low-eU portion of the curve predicts dates coincident with major Neoproterozoic cooling, while higher eU grains are either fully or partially reset by Phanerozoic burial. This best reproduces the general shape our observed data, suggesting that long-term residence at intermediate temperatures is a viable hypothesis for our dataset. However, this thermal history is a poor fit to our observed data at moderate eU (ca. 400–600 ppm), predicting dates too young for a significant cluster of data in this range.

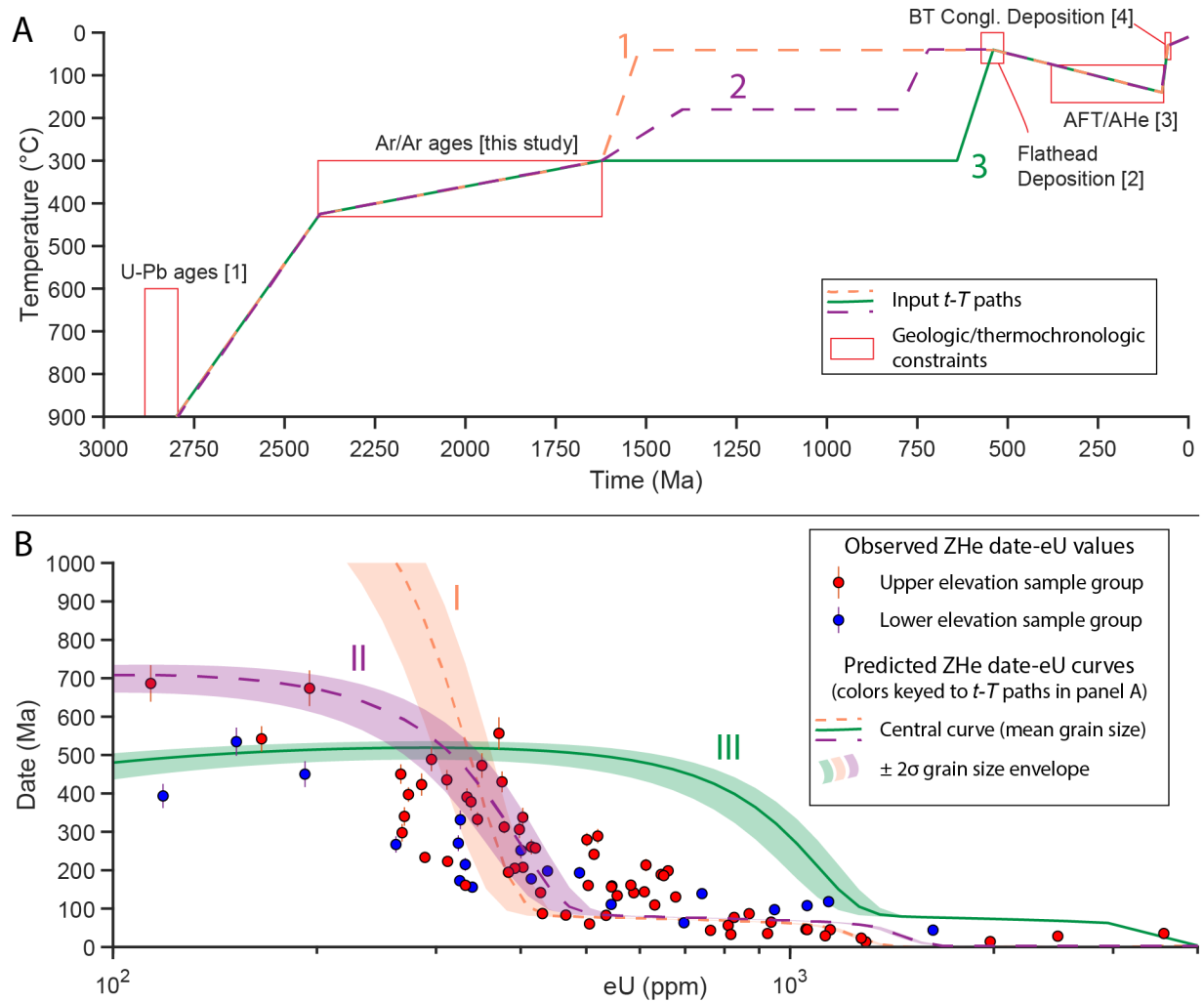


Figure 9: End-member thermal history models. (A) Input t - T histories are shown as orange (Scenario 1), purple (Scenario 2) and green (Scenario 3) lines and are described in the text. Major constraints on t - T history are shown as red boxes. Constraint references: (1) Wooden et al. (1988); Mueller et al. (2008); Carrapa et al. (2019); this study; (2) Chaudhuri and Brookins (1969); Thomas (2007); Retallack (2013); (3) Cervený (1990); Omar et al. (1994); Peyton et al. (2012); (4) Decelles et al. (1991). (B) Resulting date-eU envelopes illustrate the extent to which radiation damage and grain size effects can reproduce the observed ZHe data given the corresponding input t - T path. Observed data from upper elevation samples (>2300 m) are shown as red dots while observed data from lower elevation samples (<2300 m) are shown as blue dots. Error bars for observed ZHe dates are shown at 2σ . The output consists of a central curve (mean grain size = 51 μ m) encompassed by a shaded envelope ($\pm 2\sigma$ grain size = 14 μ m).

Phanerozoic Scenarios and Results In our Phanerozoic thermal models, we explore the effects of varying Paleozoic–Mesozoic burial temperatures, timing of Phanerozoic cooling, and

Cenozoic reheating. We hold the Proterozoic portion of our models constant, using the most acceptable fit scenario from our Proterozoic end-member models (scenario 2 in Figure 9A). We return to exploring alternative timings of Proterozoic cooling in the inverse model results.

Here, we attempt to match the gentler date-eU curve defined by our low elevation sample group by modeling a second t - T path with a temperature difference through time reflecting the paleo-temperature offset between the two elevation sample groups. We test the magnitude of this offset by modulating the temperatures of our lower sample group; we explore a range of temperature offsets from 10 °C/km to 50 °C/km, consistent with commonly observed terrestrial geothermal gradients (e.g., Einsele, 2013; Figure 10A). We note that the broadly constant temperature offsets implemented here are almost certainly an oversimplification and are intended to highlight the effects of varying this parameter on major features of the predicted date-eU correlation. Additionally, we test the effect of varying maximum Phanerozoic burial temperatures on the predicted date-eU relationship. We explore Phanerozoic burial temperatures from 120–160 °C for our upper sample group and 135–210 °C for our lower sample group.

We find the most acceptable fit results from a 45 °C Proterozoic temperature offset and 20 °C Phanerozoic offset between elevation sample groups (Figure 10A). Because the difference in mean average elevations of our sample groups is ca. 1 km, these values translate roughly into predicted geothermal gradient of 45 °C/km and 20 °C/km, respectively. Additionally, we find that maximum Phanerozoic burial temperatures of 140 °C and 160 °C for our upper and lower sample groups, respectively, best fit the shape of our observed date-eU trend (scenario 1 in Figure 10A). By contrast, hotter maximum Phanerozoic burial temperatures (scenario 2 in Figure 10A) cause significant He loss, almost completely resetting all grains regardless of eU in our

lower elevation sample, thus predicting dates too young at low–moderate eU (envelopes IIa and IIb in Figure 10B). Cooler Phanerozoic burial temperatures do not enter the grain-specific PRZ for grains of low–moderate eU, which thus retain most helium accumulated during Proterozoic residence at cool temperatures and predict dates too old for the same range of eU (envelopes IIIa and IIIb in Figure 10B). This illustrates the extreme sensitivity of the ZHe system to late-stage reheating to temperatures within or above the grain-specific effective PRZ, as controlled by eU and modeled t - T history. This effect is especially pronounced in grains of low-moderate eU, which have effective PRZs most sensitive to the range of reheating tested here. A change of 40 °C (180 °C vs. 140 °C) causes a difference of more than 550 Myr (ca. 100 Ma vs. 680 Ma) in dates predicted for low-eU grains (envelopes IIb vs. IIIb in Figure 10B). Comparison with Figure 9 shows that both Proterozoic cooling timing and Phanerozoic burial temperatures control a similar portion of the predicted date-eU trend.

Together, the output date-eU envelopes for scenario 1 encompass ca. 60% of our dataset and are a good fit to the general trend of the data at low–moderate eU (envelopes Ia and Ib in Figure 10). However, both curves are a poor fit to data points at ca. 400–600 eU, predicting corresponding dates ca. 100–200 Myr too young. Additionally, our upper elevation envelope reaches near-zero dates at higher eU than observed in our dataset.

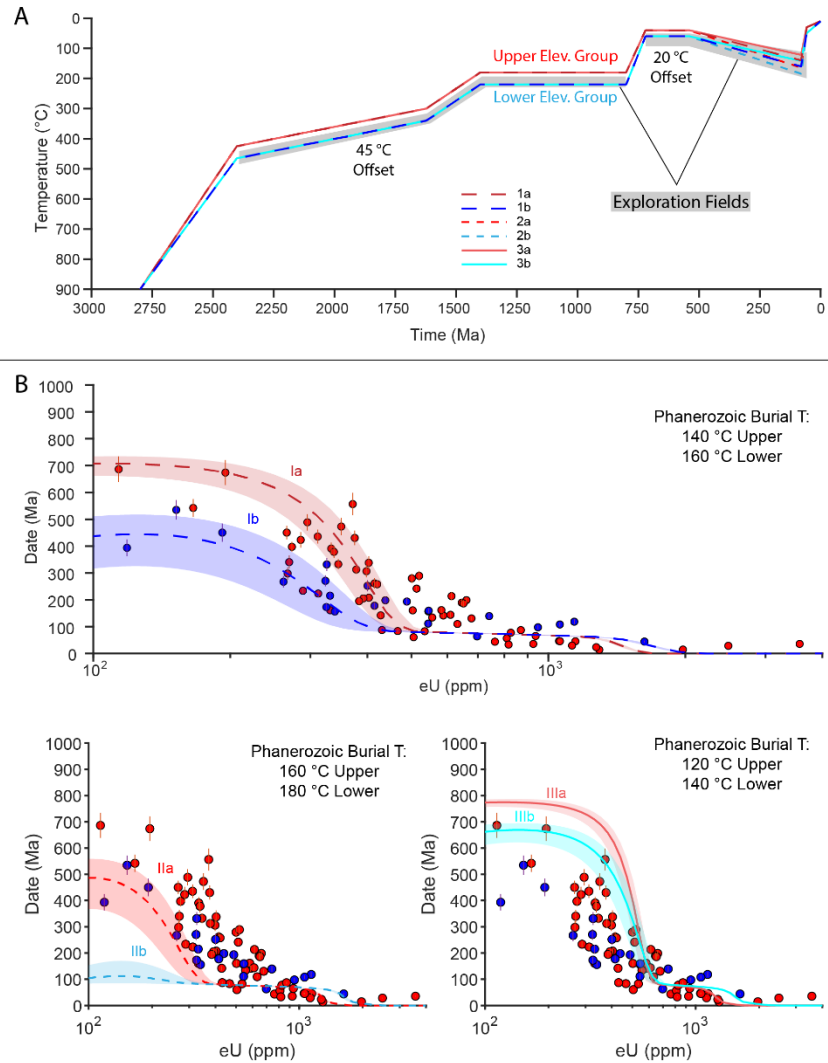


Figure 10: Phanerozoic thermal history models for upper and lower sample elevation groups, testing temperature offset between the sample groups and maximum Phanerozoic burial temperatures. (A) The input t - T histories for the upper elevation sample group (red lines) and lower elevation sample group (blue lines) have temperatures of 180 °C and 225 °C at 1300–800 Ma and maximum Phanerozoic burial temperatures of 120–160 °C and 140–180 °C, respectively. Shaded regions reflect exploration fields—ranges of temperatures tested for this model. (B) Predicted date-eU envelopes for the t - T paths in panel A. Approximately 60% of the observed data are collectively captured by the predicted envelopes for path 1a and 1b.

In the next set of thermal history models, we explore various hypotheses for Phanerozoic burial and exhumation, attempting to improve the model fit to observed data at moderate–high-eU (Figure 11). We use scenario 1 from Figure 10A as an initial input. We test the effects of

Cenozoic reburial, as suggested by AFT models of Giegengack et al. (1986), Omar et al. (1994), and Carrapa et al. (2019). We explore Cenozoic temperatures from 20 °C (no reburial) to 100°C (Figure 11A). Additionally, we modulate the timing of initial Phanerozoic uplift, exploring uplift initiation from 340 Ma to 70 Ma (Figure 11). Tested timings roughly correspond to several tectonic hypotheses for Phanerozoic uplift: (1) Paleozoic Ancestral Rocky Mountains (ARM) tectonism (e.g., Maughan, 1990); (2) Early Cretaceous uplift associated with early flat-slab subduction or far-field effects from the Sevier fold-thrust belt (e.g., Carrapa et al., 2019; Pearson et al., 2019); and (3) Late-Cretaceous–Paleocene flat-slab subduction (e.g., Bird, 1984).

Date-eU envelopes predicted by t - T histories corresponding to these tectonic hypotheses are shown in Figure 11B. We additionally plot the ZHe data of Carrapa et al. (2019) for comparison (green circles in Figure 11B), collected from near the low elevation samples of our East Rosebud trail and Chalice Peak transects. Cenozoic reburial is found to improve the fit to the observed data at high-eU and is implemented in all scenarios shown in Figure 11A. Reburial to 50 °C best matches the upper elevation data, predicting younger ages at high-eU, while the same reburial temperature also retains the fit to the high-eU tail of the lower elevation sample group (Figure 11B). This reduced temperature offset between sample elevation groups is reasonable if much of the incision of the modern topographic surface of the Beartooth Mountains took place prior to Cenozoic reburial, decreasing the basement overburden of the low elevation sample group. Comparing the high-eU portions of the date-eU curves of Figure 10B (no Cenozoic reburial) to Figure 11B (Cenozoic reburial to 50 °C) illustrates the sensitivity of high-eU grains to this relatively modest reheating. In this t - T history involving long durations spent above the ZFT PAZ, high-eU grains accumulate significant amounts of damage owing to their

high concentrations of radioactive isotopes and temperatures not conducive to annealing. These highly damaged grains have low effective T_c and are consequently reset by reheating to only modest temperatures, well below the conventional nominal T_c of the ZHe system (e.g., Reiners et al., 2004). Greater annealing of radiation damage in samples of our lower elevation group due to higher temperatures experienced—particularly during Mesoproterozoic residence in the ZFT PAZ—makes these samples less susceptible to He loss during this late reheating event. Consequently, the t - T history for our lower elevation samples predicts older dates at similar eU despite experiencing the same Cenozoic burial temperatures as the upper elevation samples, resulting in “cross-over” of the predicted date-eU curves, which is consistent with our observed data (e.g., Guenthner et al., 2013; Figure 11B).

Modulating the timing of initial Phanerozoic uplift influences the moderate eU portion of the date-eU trend, which eluded previous models. Envelopes produced by all scenarios shown in Figure 11A capture nearly the same amount of observed data points—ca. 60% of the dataset. Those involving earlier uplift initiation (scenario 1 and 2 in Figure 11A) better match the general trend of the observed date-eU correlation, especially at moderate eU. However, scenarios 1 and 2 require Paleozoic burial temperatures far in excess of that predicted by decompacted sediment accumulation modeling using depth estimates from Wise (2000; Figure 8). Scenario 3 predicts dates substantially too young at moderate eU (envelopes IIIa and IIIb in Figure 11) but is more compatible with the former sediment accumulation model results (Figure 8).

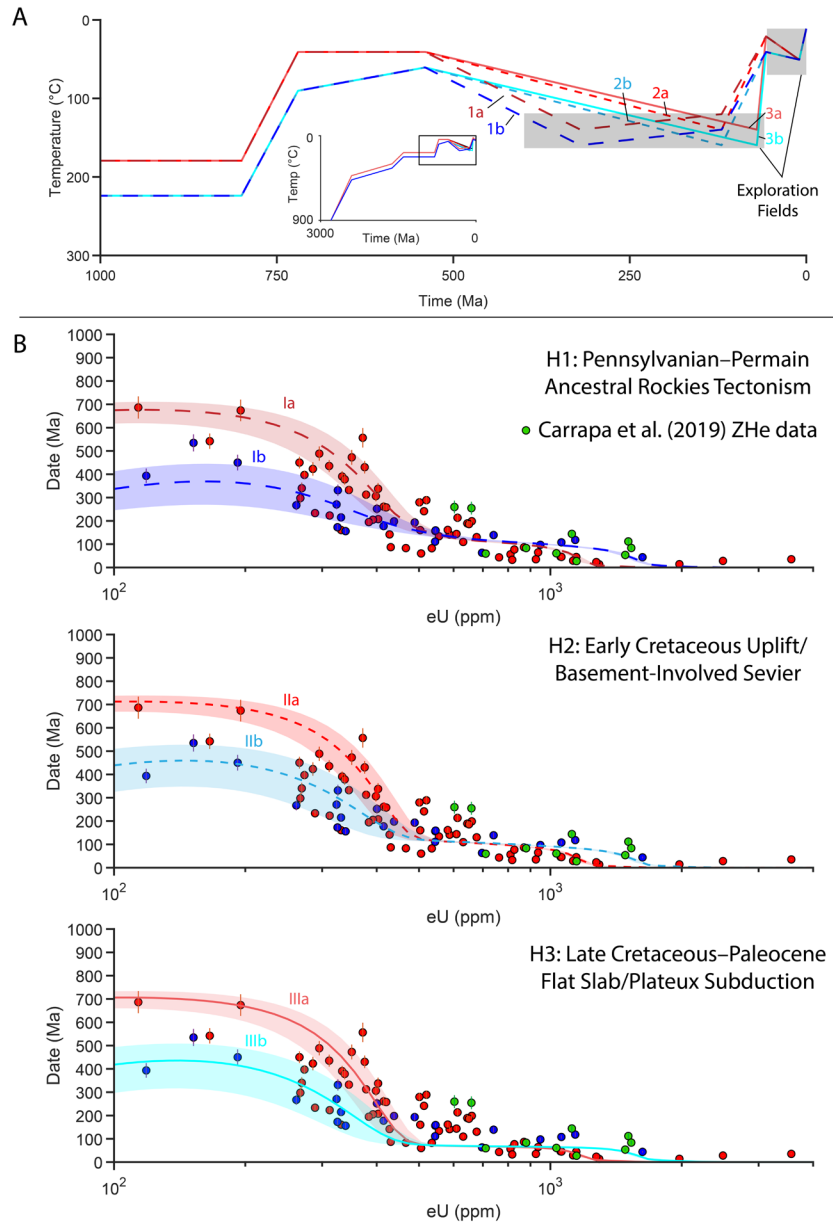


Figure 11: Thermal history models for upper and lower sample elevation groups testing the timing of initial Phanerozoic cooling and Cenozoic reburial/cooling. (A) Input t - T histories for the upper elevation sample group (red lines) and lower elevation sample group (blue lines). For clarity, only the portion of the thermal history from 1 Ga to present is shown. The full modeled t - T space is shown in the inset. Shaded exploration fields show ranges of Phanerozoic uplift timings and Cenozoic reheating temperatures tested. Of these, we show three tectonically plausible uplift histories, described in the text. (B) Predicted date-eU envelopes for the t - T paths in panel A. We additionally plot the ZHe data of Carrapa et al. (2019) for comparison (green circles). Approximately 60% of our data are captured by the envelopes predicted by each thermal history.

Inverse Model Results

Inverse modeling of upper and lower elevation sample groups yielded 11 good-fit t - T paths, respectively, out of a cumulative 100,000 attempted. This low rate of good-fits reflects both the vastness of the t - T space searched and the large volume of date-eU space defined by our dataset. Model outputs are shown in Figure 12, with good-fit paths shown using various colors rather than the default pink from HeFTy to facilitate visually following a path along its duration. This helps illustrate that aspects of all t - T paths within the model space defined by good-fit paths (good-fit envelope) are not simultaneously compatible; paths involving earliest Proterozoic cooling also generally require earliest and hottest Phanerozoic burial (i.e., paths 1a and 1b in Figure 12A), while those involving later Proterozoic cooling are typically associated with later and slightly cooler Phanerozoic burial (i.e., paths 2a and 2b in Figure 12A). This is consistent with our forward model results, showing that earliest Proterozoic cooling predicts old dates at low-eU (Figure 9), which can be brought down to dates more consistent with our observed data by hotter and longer duration residence at Phanerozoic burial temperatures within the effective PRZ (Figure 10). We note that all good-fit t - T paths for our upper elevation samples predict maximum Phanerozoic temperatures of ≥ 150 °C by at least 300 Ma. Such high temperatures in the early Paleozoic are generally incompatible with our decompacted sedimentary accumulation models assuming modest amounts of basement overburden (Figure 8; Wise, 2000). Inverse thermal models with constraints enforcing a closer approximation of burial temperatures indicated by these sedimentary accumulation models found no good fits; these are reported in Appendix D.

Best-fits for upper and lower elevation sample groups achieved similar goodness-of-fit metrics (0.71 and 0.72, respectively). The date-eU correlation predicted by best-fit paths (insets

in Figure 12) are an excellent fit to the synthetic data and capture much of our observed data within a 2σ grain size envelope. Additionally, predicted date-eU correlations are a very good fit to the general date-eU trend of each elevation group. However, best-fit paths for respective elevation groups incorporate different timings and magnitude of Proterozoic cooling, subsequent reburial, and Phanerozoic exhumation. While some offset in magnitude (i.e., temperature) is expected due to the differing mean elevations of each sample group, the timings of major thermal events are mutually exclusive—considering the close spatial proximity of our samples—and therefore solutions making differing predictions are incompatible.

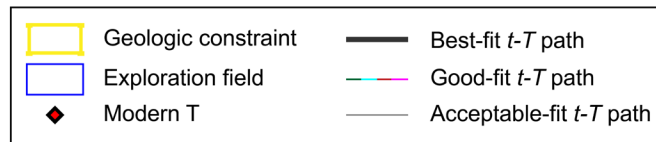
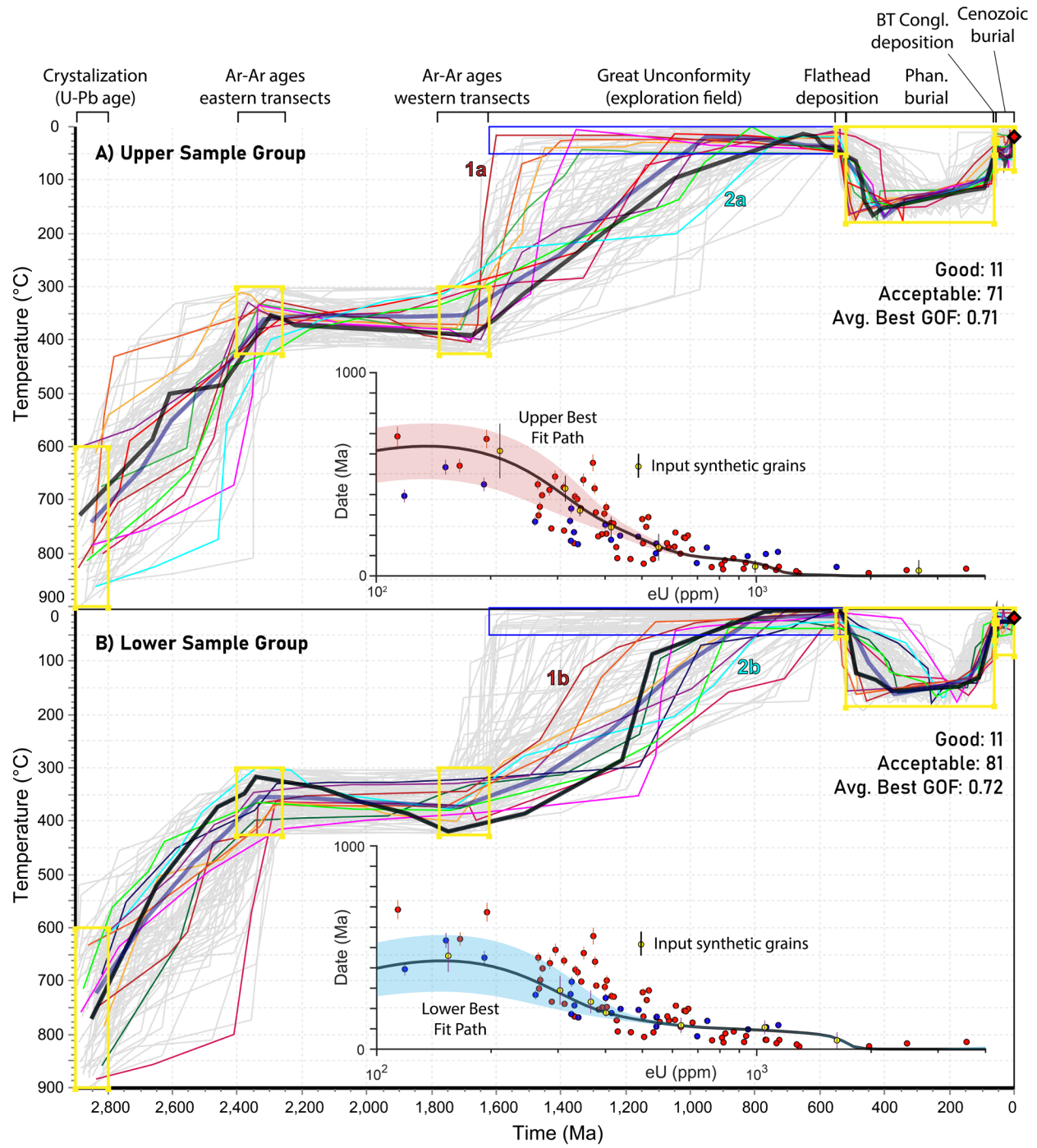


Figure 12: Inverse model results for upper (A) and lower (B) elevation sample groups. Labeled t - T paths are referenced in the text. Geologic and thermochronologic constraints are shown as yellow boxes while the blue Great Unconformity “exploration field” encourages the model to investigate, but does not require, exhumation prior to deposition of the Flathead Sandstone (see Flowers et al., 2020 for an example of this technique). Note that our Phanerozoic constraints allow but do not require reburial, honoring the paucity of preserved sedimentary rocks in the Beartooth Mountains. Date-eU correlations predicted by the best-fit t - T paths are shown in the insets. The fit of each predicted date-eU trend matches that of our observed data well. However, the best-fit paths of the upper and lower elevation groups respectively predict incompatible timings of Proterozoic cooling and Paleozoic maximum burial.

To identify the t - T solution most compatible with both sample groups in our dataset, we overlay the good-fit envelopes of each sample group in Figure 13A (see Sousa and Farley, 2020 for a similar use of this method). Good-fit envelopes have considerable overlap, with only Jurassic time having non-overlapping envelopes; even then, envelopes are separated by a small temperature offset explainable by a modest geothermal gradient (ca. 10 °C; Figure 13A). This suggests the existence of thermal history solutions compatible with both sample groups and allow us to narrow the poorly constrained space defined by each envelope alone (i.e., Figure 12).

We test “refined” forward models by iteratively forward modeling the t - T space defined by overlapping envelopes, or just outside within reasonable temperature offset (≤ 30 °C; Figure 13B), and testing the timing, magnitude, and rate of key thermal events within these bounds. We additionally constrain these refined models by assessing the fit of AFT dates and length distributions predicted by our refined t - T paths to AFT data of Omar et al. (1994), collected near the Beartooth Highway (Figure 4–5). Two groups of three AFT samples—respectively proximally above and below the lower boundary of a fossil AFT PAZ—are grouped and plotted against predicted AFT track length distributions in Figure 13C.

Our best-fit refined thermal history solution for the upper elevation sample group shows early cooling following crystallization to temperatures of ca. 340 °C by ca. 2.4 Ga, followed by

minor reheating to ca. 360 °C by ca. 1.62 Ga, as constrained by our zircon U-Pb and biotite $\text{Ar}^{40}/\text{Ar}^{39}$ results. ZHe model results then predict cooling to 170 °C by 1.3 Ga. Following 250 Myr of residence at upper mid-crustal temperatures, rocks then cool to upper crustal temperatures between 1050 Ma and 800 Ma. Rocks reside at 60 °C from 800 Ma to 540 Ma—consistent with ca. 1.5–3 kilometers of basement overburden. This amount of basement overburden achieves greater compatibility between the Paleozoic temperatures favored by our inverse model results (Figure 12) and our decompacted sedimentary thickness models suggesting only moderate early Paleozoic burial by sedimentary rocks (Figure 8). Phanerozoic reheating occurs relatively rapidly, reaching a maximum temperature of ca. 145 °C by Late Pennsylvanian time. Rocks then cool to 90 °C by 100 Ma, followed by a brief period of Early Cretaceous reheating of 10 °C. Rapid cooling commences at 70 Ma, bringing rocks to near-surface temperatures by 57 Ma. This is followed by modest Cenozoic reheating—reaching maximum temperatures of 40 °C by 25 Ma—after which samples arrive at surface temperatures at the present (Figure 13A). The t - T history for our lower elevation sample group follows the same trajectory, offset by a variable geothermal gradient of: (1) ca. 30–50 °C in the Mesoproterozoic; (2) ca. 15–20 °C in the Paleozoic and Mesozoic; and (3) ca. 0–10 °C following the final pulse of Late Cretaceous–Paleocene cooling (Figure 13A).

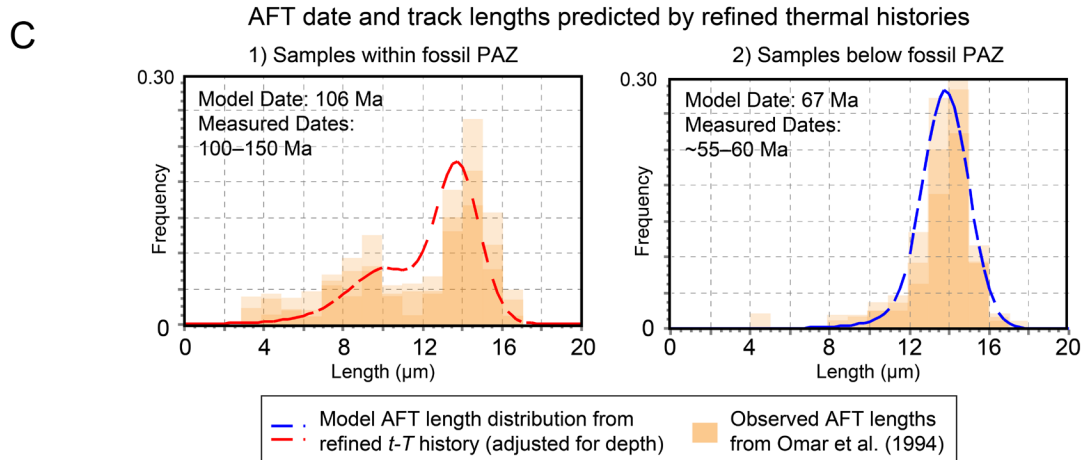
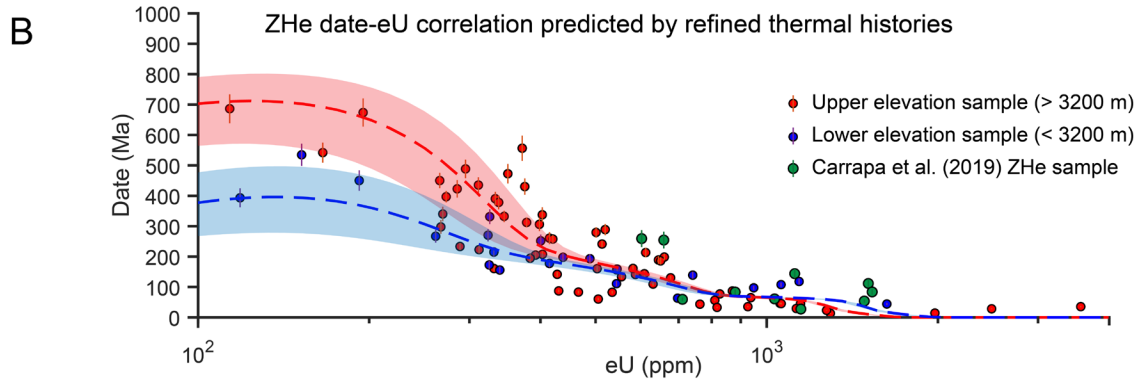
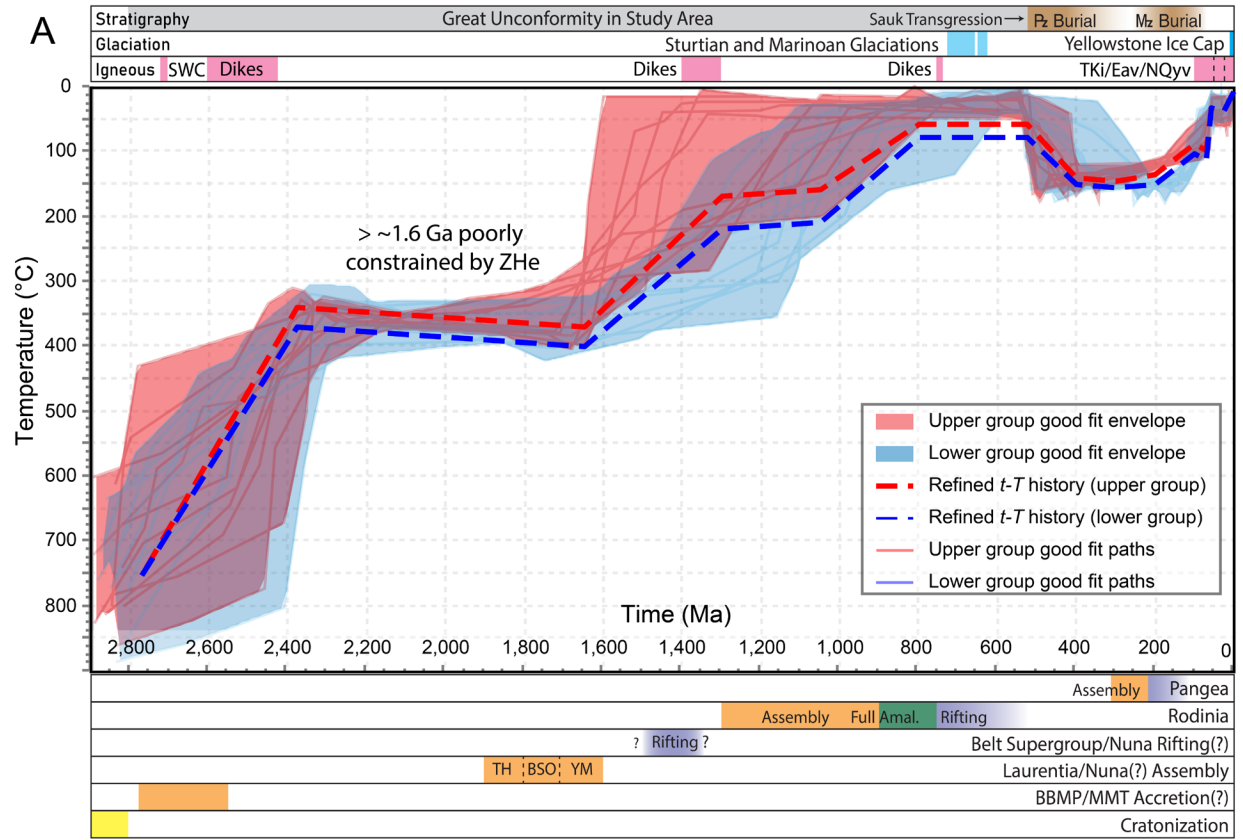


Figure 13: (A) Good-fit envelopes for upper and lower elevation sample groups defined by good-fit t - T paths in Figure 12. Areas of overlap—representing model space with common possible t - T histories—are a purple hue. We forward model a refined thermal history shown by bold dashed red (upper group) and blue (lower group) lines. The timing of major tectonic events affecting the Wyoming craton from 3 Ga to the present is shown as horizontal bars at bottom, from Peterman (1979), Hoffman (1988), Whitmeyer and Karlstrom (2007) and Li et al. (2008). The timing of other important geologic events is shown as horizontal bars at top (Baadsgaard and Mueller, 1973; Maughan, 1990; Godd  ris et al., 2003; Bonnichsen et al., 2008; Dud  s et al., 2010). Gradational shading on events indicates uncertainty in timing. BSO—Big Sky orogeny; Eav—Eocene Absaroka volcanics; Full Amal.—Full amalgamation; Ig. Events—Major igneous events; NQyv—Neogene–Quaternary Yellowstone Volcanics; SWC—Stillwater Complex; TH—Trans-Hudson Orogeny; TKi—Tertiary–Cretaceous intrusives; YM—Yavapai/Mazatzal orogenic events. (B) Date-eU envelopes predicted by refined thermal history in panel A. (C) AFT ages and length distributions predicted by our refined thermal history. Measured track lengths reported by Omar et al. (1994) are shown by transparent orange histograms. Each histogram is constructed from three samples such that darker orange regions contain more overlapping length measurements. Phanerozoic input temperatures were adjusted to match hypothesized depths below the Great Unconformity of AFT sample locations.

This refined thermal history predicts date-eU curves that are a good fit to both our upper and lower elevation sample groups, respectively. Predicted envelopes capture ca. 70% of our observed data—the most of any forward model—and fit the general trend of the dataset well (Figure 13B). As an additional test, we plot ZHe data of Carrapa et al. (2019) for comparison. Few of the data points of Carrapa et al. (2019) are captured by our envelopes, but there is considerable scatter in their dataset and our model predictions match the general date-eU trend of their data well (Figure 13B).

This t - T history also accurately predicts AFT track length distributions of Omar et al. (1994; Figure 13C). The bimodal distribution of samples interpreted by Omar et al. (1994) to reside within in a fossil AFT PAZ is matched well by our refined t - T history (inset 1 in Figure 13), after adjusting burial temperatures to 15   C cooler than our upper elevation sample t - T path (red dashed line in Figure 13A) to account for the hypothesized shallower depth of these samples with respect to the paleo-Great Unconformity due to folding of the plateau surface (Figures 3 and

4). Likewise, we achieve a good fit to the unimodal distribution of samples interpreted to reside below the fossil AFT PAZ (inset 2 in Figure 13C) using the refined t - T path of our upper elevation samples, which likely resided at similar depths (Figure 13A).

Discussion

Missing Argon: The Nature and Extent of Paleoproterozoic Thermotectonism in Southwest Montana

Biotite $^{40}\text{Ar}/^{39}\text{Ar}$ results from our western samples comprise the easternmost instance of <1.8 Ga medium-temperature thermochronometer ages in southwestern Montana and the first from the BBMP (Table 1; Figure 14 and references in figure caption). Though evidence of Paleoproterozoic thermal overprint has not been extensively documented in much of the Beartooth Mountains, exposed basement elsewhere in Montana records major convergent tectonism associated with the Proterozoic amalgamation of Laurentia (Figure 1; Harms et al., 2004; Mueller et al., 2005; Condit et al., 2015). Magmatism and associated metamorphism in the Little Belt Mountains at ca. 1.86–1.81 Ga has been suggested to record closure of the “Little Belt Ocean” (e.g., Gifford et al., 2014) and collision of the Archean Medicine Hat and Wyoming cratons, comprising the Paleoproterozoic Great Falls tectonic zone (Figure 14; Mueller et al., 2005; Foster et al., 2006). Additionally, slightly younger thermotectonism is recorded in rocks of the MMP in basement-cored ranges to the west in Montana. This region, known as the Big Sky orogen, documents peak metamorphic conditions of ca. 0.8–1.2 Gpa and >700 °C between ca. 1.78–1.72 Ga (e.g., Harms et al., 2004; Ault et al., 2012; Condit et al., 2015). Similarly, mantle metasomatism related to Paleoproterozoic orogenesis has been documented in mantle xenoliths

in southwestern Montana (e.g., Carlson and Irving, 1994; Dahl et al., 1998; Rudnick et al., 1998; Downes et al., 2004; Facer et al., 2009).

Documented conclusive evidence for thermal overprinting from the Big Sky orogenic event in southwestern Montana is dominantly limited to basement exposures west/northwest of a northeast-trending transitional boundary approximately coinciding with the modern Gallatin canyon (Figure 14; Giletti, 1966; Mueller and Cordua, 1976; James and Hedge, 1980; O'Neill et al., 1988; Weyand, 1989; Mueller et al., 1993). Original interpretation of this feature, commonly referred to as “Giletti’s line” (e.g., Harms et al., 2004; Condit et al., 2015), proposed that this boundary marked a Paleoproterozoic thermal transition between hotter temperatures to the northwest—inducing resetting of the medium-temperature $^{40}\text{K}/^{39}\text{Ar}$ and Rb-Sr systems—and cooler temperatures to the southeast—where un-reset ($>\text{ca. } 1.8 \text{ Ga}$) $^{40}\text{K}/^{39}\text{Ar}$ and Rb-Sr ages have been observed (Giletti, 1966). However, subsequent work has revealed excess Ar in biotite in key $^{40}\text{K}/^{39}\text{Ar}$ analyses originally defining Giletti’s line and called into question the precise position and nature of this thermal boundary (Giletti, 1966, 1971).

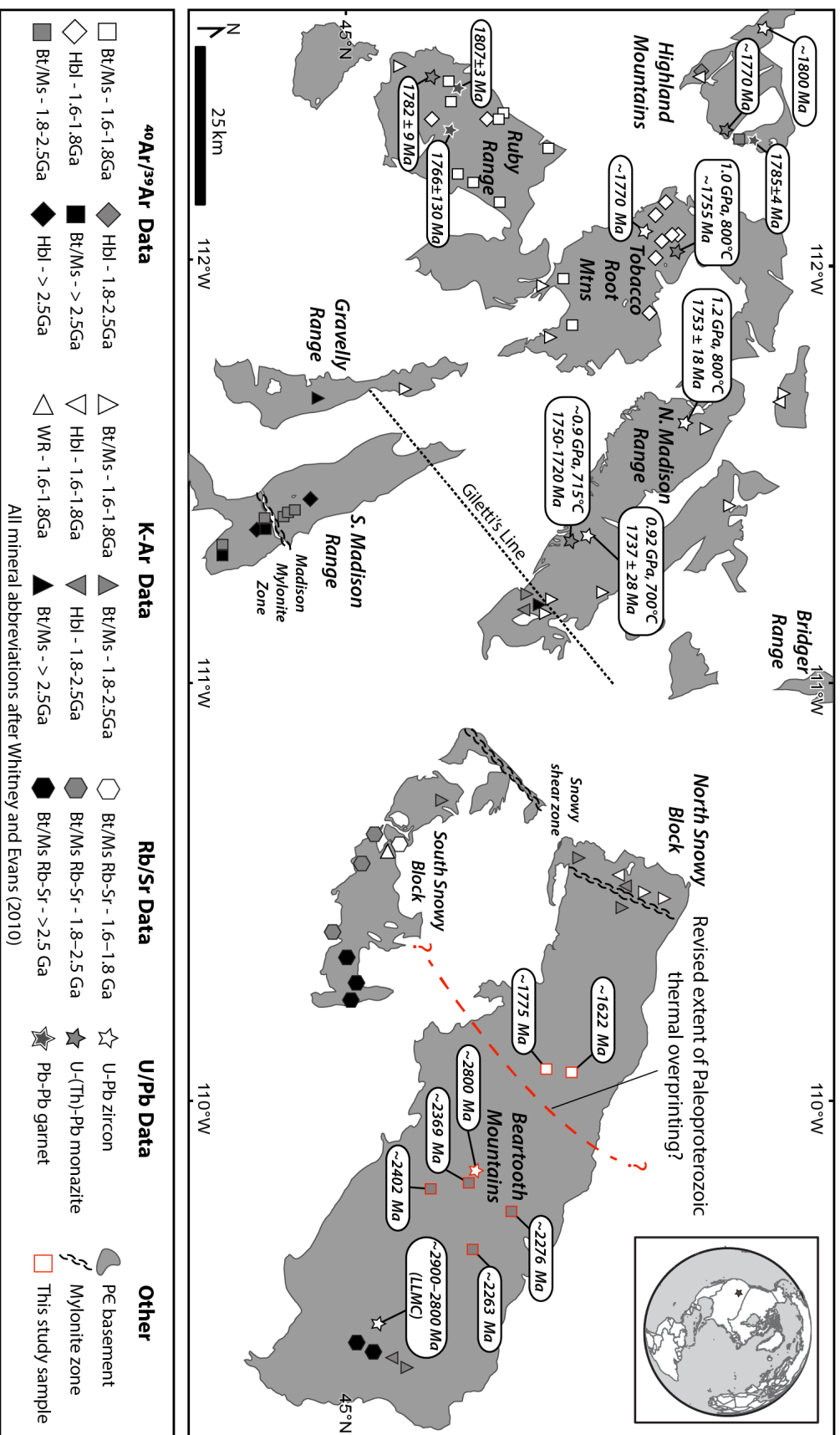


Figure 14: New and existing geochronologic and thermochronologic data constrain the extent of Big Sky Orogeny-related (1.8–1.71 Ga) thermotectonism. Data from this study are outlined in red. Red dashed line denotes the potential revised extent of Paleoproterozoic thermal overprinting suggested by our data. Data citations: $^{40}\text{K}/^{39}\text{Ar}$ —Giletti (1966, 1971); Reid et al., (1975); $^{40}\text{Ar}/^{39}\text{Ar}$ —Erslev and Sutter (1990); Harlan et al. (1996); Roberts et al. (2002); Brady (2004); Hames and Harms (2013); this study; U-Pb zircon geochronology—Mueller et al. (2004, 2008); Ault et al. (2012); Condit et al. (2015); this study; U-Th-Pb monazite geochronology—Cheney et al. (2004); U-Th-Pb monazite electron microprobe geochronology—Alcock et al. (2013); Condit et al. (2015); Pb-Pb garnet geochronology—Roberts et al. (2002); Pb-Pb garnet thermobarometry—Cheney et al. (2004); Ault et al. (2012); Condit et al. (2015); Rb-Sr—Reid et al. (1975); Montgomery and Lytwyn (1984). WR—Whole Rock; LLMC—Long Lake Magmatic Complex (Mueller et al., 2008).

The general overlap of biotite $^{40}\text{Ar}/^{39}\text{Ar}$ ages of our western samples with the timing of Big Sky thermotectonism to the west suggests that the thermal effects of this orogenic event, or perhaps associated mantle metasomatism (e.g., Carlson and Irving, 1994; Rudnick et al., 1998; Downes et al., 2004; Facer et al., 2009), likely reached temperatures above the biotite $^{40}\text{Ar}/^{39}\text{Ar}$ closure temperature ($>300\text{ }^{\circ}\text{C}$) at least as far east as the Lakes Plateau region of the Beartooth Mountains, ca. 100 km east of Giletti's line (Figure 14). This interpretation is supported to some degree previous results documenting partial resetting of $^{40}\text{K}/^{39}\text{Ar}$ and Rb-Sr ages at ca. 1.8 Ga in the North and South Snowy blocks (Figure 14; Reid et al., 1975; Montgomery and Lytwyn, 1984) and previous suggestions of Proterozoic movement on the Snowy shear zone near Yankee Jim canyon (Erslev, 1992; Webber et al., 2019). Paleoproterozoic thermal overprinting of portions of the Beartooth Mountains is perhaps likewise consistent with documented greenschist facies metamorphism within the Stillwater Complex, where apatite U-Pb ages of ca. 1.7 Ga have been documented (Nunes and Tilton, 1971; Page, 1977; Page et al., 1985). However, Thacker et al. (2017) attribute a Cretaceous age to similar greenschist metamorphism documented in Stillwater Complex rocks near the South Prairie Fault. We note that our ZHe models are generally agnostic to this portion of the thermal history, with little to no information from this

period preserved by the ZHe system due to the likely residence of samples below the ZHe PRZ and/or ZFT PAZ (Figure 6; Appendix D).

In contrast to our western transects, unreset biotite $^{40}\text{Ar}/^{39}\text{Ar}$ ages from our eastern samples—along with existing $^{40}\text{K}/^{39}\text{Ar}$ and Rb-Sr results from near the Beartooth Highway (Gast et al., 1958)—suggest that the eastern portion of the Beartooth Mountains did not reach temperatures of $> \text{ca. } 300\text{--}425\text{ }^{\circ}\text{C}$ after at least $\text{ca. } 2.2\text{ Ga}$ (Figure 14). Alternatively, samples collected from the eastern transects may reflect a structurally higher portion of the Beartooth block—positioned at higher, and thus cooler, crustal levels during Paleoproterozoic time and subsequently tilted during Precambrian tectonism. This is largely consistent with previous geobarometry results suggesting $\geq 0.8\text{ Gpa}$ pressures near the Lakes Plateau region of the Beartooth Mountains and $\leq 0.6\text{ Gpa}$ pressures further east (i.e., near the Beartooth Highway) in rocks of similar Archean age now residing at similar elevation to rocks to the west (e.g., Henry et al., 1982; Mueller et al., 2008; Mogk et al., 2012; Will, 2013 and references therein). Large-scale tilting of the Beartooth block, if present, must have occurred prior to Cambrian time, as Cambrian rocks preserved at Beartooth Butte in the southeastern portion of the range dip only shallowly ($\text{ca. } 8^{\circ}$) to the southwest (e.g., Foote et al., 1961; Pierce and Nelson, 1971). Additionally, Cambrian rocks are deposited on a tilted section of layered intrusives of the Stillwater Complex (e.g., Lopez, 2001), providing further evidence for some degree of tilting of Archean rocks of the Beartooth Mountains prior to Cambrian sedimentation. However, the specific timing of this tilting and its potential effect on observed biotite $^{40}\text{Ar}/^{39}\text{Ar}$ age distributions in the Beartooth Mountains are poorly constrained.

These new and previously published medium-temperature thermochronology results suggest that reinterpretation of the position and nature of Giletti's line may be warranted. Our results indicate that, in the Beartooth Mountains, a refined maximum extent of Paleoproterozoic thermal overprinting preserved in present surface exposures of Archean rock may exist between the Lake Plateau and Froze-to-Death Plateau regions of the Beartooth Mountains, roughly coincident with the main fork of the Stillwater River (Figures 2 and 14). Significant uncertainty remains regarding the extent and nature of Paleoproterozoic orogenesis in Montana. Specifically, the documented significant spatial variability of reset (< 1.8 Ga) thermochronometer ages east/southeast of Giletti's line (e.g., Gast et al., 1958; Reid et al., 1975; Montgomery and Lytwyn, 1984; this study) is difficult to reconcile. Additionally, the mechanism of heat transfer to the central Beartooth Mountains—located > 150 km east of regions such as the Tobacco Root and Highland mountains where high-grade Paleoproterozoic tectonism is documented (e.g., Harms et al., 2006; Condit et al., 2015)—is unresolved. Spatially variable heating of the Wyoming craton lithosphere associated with mantle metasomatism or mafic underplating related to Paleoproterozoic tectonism may provide a mechanism for partial resetting of medium-temperature thermochronometers east/southeast of Giletti's line; evidence of such processes active in Paleoproterozoic time has been interpreted from geochronology and trace element results from xenoliths of the sub-GFTZ mantle lithosphere in Montana (e.g., Carlson and Irving, 1994; Dahl et al., 1998; Rudnick et al., 1998; Downes et al., 2004; Facer et al., 2009; Gifford et al., 2014). However, the role of mantle fluids in Paleoproterozoic heat transfer to the Beartooth Mountains remains speculative at present, with most evidence for these processes documented in exposures of the Little Belt and Bearpaw mountains to the north of the present study area (e.g.,

Downes et al., 2004; Facer et al., 2009; Gifford et al., 2014). More detailed thermochronology work in the region is clearly justified, given the broad utility of this comparatively well-exposed Paleoproterozoic zone of convergent tectonism in the study of the evolution of Precambrian orogenic crust (e.g., Harms et al., 2004; Mueller et al., 2004; Condit et al., 2015).

Missing Time: Timing(s) and Mechanism(s) of Proterozoic Cooling

Mesoproterozoic Cooling from Post-Orogenic Collapse

Our refined thermal models suggest that Proterozoic cooling of rocks in the Beartooth Range likely occurred in multiple phases, with an intervening period of residence at intermediate temperatures (Figure 13A). All good fits to our dataset predict some degree of cooling soon after 1.6 Ga (Figure 12), with overlapping good-fit envelopes of our upper and lower sample groups predicting this initial stage of cooling lasting until at least ca. 1.3 Ga (Figure 13A). The onset of this cooling occurs just after reheating from the Paleoproterozoic Big Sky orogenic event—as recorded in our biotite $^{40}\text{Ar}/^{39}\text{Ar}$ results. This may indicate that this episode of cooling was initially triggered by post-orogenic collapse of the Big Sky orogen. Such collapse has been suggested by Condit et al. (2015) to explain observed spatial and temporal trends of Big Sky tectonism in the region and by Jamieson and Beaumont (2011) to explain similar trends in metamorphic grade of the Grenville orogen.

Additionally, this phase of cooling is roughly coincident with the deposition of the Belt Supergroup, with the onset of deposition within the Helena Embayment northwest of the Beartooth Mountains constrained to post-1.78 Ga by the age of the youngest detrital zircons observed in the LaHood Formation (Mueller et al., 2016) and pre-1.45 Ga by the intrusion of the Moyie sills into the western Lower Belt stratigraphy (Prichard and Aldridge formations;

(Symons and Timmins, 1992; Anderson and Davis, 1995; Evans et al., 2000). The termination of Mesoproterozoic cooling of the Beartooth Mountains intermediate temperature—required to produce the “cross-over” of predicted date-eU correlations of our upper and lower elevation sample groups (Figure 13A)—is consistent with the paucity of ca. 2.8–2.9 Ga detrital zircons, the dominant ages of the BBMP, in rocks of the Helena Embayment (Guerrero et al., 2016; Mueller et al., 2016; Anderson, 2017; Fox, 2017; Ronemus et al., 2020). Extensional collapse of the topography associated with the Big Sky orogen was interpreted by Mueller et al. (2016) to have resulted in the development of the “Proterozoic equivalent of the modern Basin and Range” topography. These authors interpret that this topography isolated the Belt Basin from 2.8–2.9 Ga sources, with the location of the modern Beartooth Range perhaps comprising a down-dropped area of low topography below regional base level. This model is consistent with our interpretation of rocks having experienced some Mesoproterozoic cooling, perhaps associated with the collapse of the Big Sky orogen and/or subsequent extension, but not having been fully exhumed to the surface during this time (Figure 13).

Neoproterozoic Cooling and the Formation of the Great Unconformity Surface At the location of the Beartooth Range, the Great Unconformity comprises ca. 540 Ma sedimentary rocks overlying ca. 2.9–2.8 Ga gneisses of the main Beartooth block and ca. 2.71 Ga rocks of the Stillwater Complex (e.g., Lopez, 2001), leaving an unconstrained period of >2 Gyr during which the development of this erosional surface may have taken place. The vast and widespread period of continental erosion contributing to the formation of this surface almost certainly played an important role in the evolution of Earth’s atmosphere and biosphere (e.g., Peters and Gaines, 2012; Husson and Peters, 2017; Keller et al., 2019). However, the precise nature of its influence

is poorly constrained. The cause of this widespread erosion is widely debated, with hypotheses (modified from Flowers et al., 2020) including: (H1) Stenian to Tonian exhumation associated with the assembly, full amalgamation, or early diachronous breakup of supercontinent Rodinia (Godd  ris et al., 2003; Zhong et al., 2007; Halverson et al., 2007; Li and Zhong, 2009; Macdonald et al., 2010; Horton, 2015; Cox et al., 2016; DeLucia et al., 2018); (H2) Cryogenian Snowball Earth glaciation (Planavsky et al., 2010; Sahoo et al., 2012; Pogge von Strandmann et al., 2015; Lau et al., 2017; Brocks et al., 2017; Keller et al., 2019); and (H3) Ediacaran to Early Cambrian erosional transgression, rifting, or assembly of supercontinent Pannotia (Brasier, 1980; Squire et al., 2006; Peters and Gaines, 2012; Husson and Peters, 2017; Karlstrom et al., 2018; Nance and Murphy, 2019). Alternatively, Flowers et al. (2020) propose that this erosion occurred diachronously in different regions as a result of multiple tectonic causes, rather than representing a globally synchronous event. Many of these hypotheses make specific predictions for the timing of pre-Great Unconformity continental erosion which can be tested with deep-time thermochronology.

Most of our inverse thermal model good-fit paths indicate a second episode of Proterozoic cooling in late Mesoproterozoic or early Neoproterozoic time (Figures 12 and 13). Overlapping good-fit envelopes between our upper and lower elevation sample groups suggest the bulk of this cooling initiated between ca. 1.4 Ga and 1 Ga, with samples arriving at upper-crustal temperatures near the GU surface between ca. 1.1 Ga and 0.8 Ga (Figure 13). This large-magnitude Proterozoic cooling is broadly consistent with previous model results from several other datasets from western/central Laurentia, although these previous studies generally predict cooling consistent with only the youngest portions of the overlapping good-fit envelope defined

by our model results (Figure 13; Orme et al., 2016; DeLucia et al., 2018; McDannell et al., 2018; Flowers et al., 2020). Several of these studies (DeLucia et al., 2018; McDannell et al., 2018) investigated regions > 1000 km from the Beartooth Mountains and all implemented modeling techniques somewhat different from those in the present study, inhibiting direct comparison of model results. However, we discuss these findings to place our results in the broader context of pre-GU erosion of Laurentia defined by this previous work.

(1) Multi-chronometer thermochronology results—including K-feldspar $^{40}\text{Ar}/^{39}\text{Ar}$ and ZHe—from the interior of the Superior and Sask cratons of central Canada suggest exposed basement rock in these regions was ca. 200 °C by early Neoproterozoic time; large-magnitude cooling is interpreted to have initiated at ca. 1000 Ma, reaching near-surface temperatures by ca. 400 Ma (McDannell et al., 2018). This cooling was attributed by the authors largely to exhumation associated with magmatic underplating during Midcontinent Rift extension and Grenville orogenesis. We note that most samples of this study were collected > 1000 km from the location of the Beartooth Mountains and used different modelling techniques—utilizing the *Arvert* software package (www.ees.lehigh.edu/EESdocs/geochron/software.html) to model multi-diffusion domain data from the K-feldspar $^{40}\text{Ar}/^{39}\text{Ar}$ system. Additionally, this study sampled interior regions of Laurentia which have largely not been overprinted by significant Phanerozoic burial and orogenesis (McDannell et al., 2018).

(2) DeLucia et al. (2018) interpret Neoproterozoic cooling in the Ozark Plateau region of Missouri, where inverse modeling of ZHe, AHe, and AFT data suggest initiation of pre-GU cooling at ca. 900–800 Ma and termination at ca. 700–550 Ma (DeLucia et al., 2018). Additionally, modeling of AHe and AFT results of this dataset resolves an additional ca. 225–

150 Ma pulse of exhumation (DeLucia et al., 2018). The authors note that these modeled cooling phases coincide with the timing of Rodinia and Pangea supercontinent break up, respectively, and suggest this cooling was likely related to broad mantle upwelling and perhaps lithospheric delamination associated with this phase of the supercontinent cycle. Subsequent modeling of this dataset by Flowers et al. (2020) demonstrated that a somewhat older timing of significant cooling is also permissible by these data, finding good-fit t - T paths with major cooling to < 100 °C as early as 1037 Ma. As in the above case, we note that the Ozark Plateau region is situated in the interior of Laurentia, >1500 km from the location of the Beartooth Range, and almost certainly experienced a different thermal history than samples of the present study.

(3) Deep-time ZHe modeling of cratonic basement exposed in the Bighorn and Wind River ranges, Wyoming, predicts similar largely Neoproterozoic cooling (Orme et al., 2016). In contrast to results from the interior of Laurentia, the tectonic settings of the Big Horn and Wind River mountain ranges are similar in many regards to that of the Beartooth Mountains—comprising Archean rock buried by Paleozoic sedimentation and subsequently uplifted in the hanging-walls of Laramide reverse faults. The favored forward model t - T paths of this study involve an initial phase of Paleoproterozoic–Mesoproterozoic cooling similar to our model results. Additionally, the authors interpret a second phase of cooling initiating at ca. 900 Ma—post-dating that defined by our overlapping good-fit model results by ca. 100 Myr (Figure 13)—and continuing until the inception of Paleozoic sedimentation at ca. 520 Ma—terminating significantly later (> 280 Myr) than that favored by our inverse model results (Figure 13). In this instance we note that Orme et al. (2016) implemented only forward modeling of their dataset, with their results intended primarily to broadly demonstrate the ability of deep-time ZHe thermal

modeling to resolve billion-year thermal histories and not to pinpoint the precise timing of pre-GU cooling.

(4) In the Front Range of Colorado, the ca. 1.07 Ga Pikes Peak batholith is unconformably overlain by the ca. 510 Ma Sawatch Sandstone across the Great Unconformity (Flowers et al., 2020). ZHe inverse thermal model results from this batholith indicate large-magnitude cooling between ca. 1 Ga and 717 Ma (Flowers et al., 2020). These models were additionally constrained by the presence of the Cryogenian Tavakaiv sandstone injectites, interpreted to constrain the maximum age of the Great Unconformity surface in this region to pre-Sturtian Snowball Earth glaciation based on the presence of highly weathered angular fragments of basement rock entrained therein and Cryogenian (U-Th)/He dates from associated specular hematite (Murray et al., 2018; Jensen et al., 2018; Siddoway et al., 2019; Flowers et al., 2020). As the Pikes Peak batholith crystalized in latest Mesoproterozoic time and Tavakaiv injectites described above constrain the bulk of cooling to pre-717 Ma, the geology of this region alone essentially limits the bulk of pre-GU erosion to Tonian time. Given the Archean age of the basement and the notable lack of Neoproterozoic injectites in the Beartooth Mountains, the thermal history of the present study is much more poorly constrained. Despite this, our inverse model results are likewise consistent with the bulk of pre-GU erosion taking place prior to Cryogenian time (Figure 13).

Overlapping good-fit envelopes in the present study encompass a relatively broad swath of Proterozoic t - T space, to some extent inhibiting discrimination of the precise geodynamic mechanism culpable for this erosion. However, all overlapping good-fit t - T paths predict termination of pre-GU cooling prior to ca. 800 Ma. This effectively precludes hypotheses 2 and 3

described above, which predict Cryogenian and Ediacaran to Cambrian cooling, respectively. Additionally, the largest magnitudes of cooling predicted by our inverse model results (Figure 13) overlap more so with the assembly and full amalgamation of supercontinent Rodinia rather than its break up, as suggested to have been associated with exhumation in the Ozarks region (DeLucia et al., 2018).

Together, the deep-time thermal history modeling results of this study and those previous (Orme et al., 2016; DeLucia et al., 2018; McDannell et al., 2018; Flowers et al., 2020) concerning the Proterozoic cooling history of western-central Laurentia suggest somewhat diachronous timings for the initiation and termination of this cooling. Indicated timings for initiation of this cooling range from ≥ 1000 Ma (McDannell et al., 2018; Flowers et al., 2020; this study) to ca. 800 Ma (DeLucia et al., 2018) and that for its termination from ca. 800 Ma (this study) to ca. 400 Ma (McDannell et al., 2018). This underscores the notion that Great Unconformity erosion was likely *not* a globally synchronous phenomenon with a single tectonic cause, as pointed out most recently by Flowers et al. (2020). Rather, various regions of the Earth's surface—even across a single continent—most likely experienced exhumation at different times. The margins and the interiors of continents likely respond differently to processes associated with supercontinent growth and break up (e.g., DeLucia et al., 2018) and more localized effects—such as magmatic underplating (McDannell et al., 2018)—perhaps only add to this diachroneity in erosion timing. Results of our study suggest a pre-800 Ma timing for the bulk of pre-GU erosion in the Beartooth Mountains—one specific region near the western margin of Laurentia. The integration of these results with continued efforts to understand the

timing and magnitude of this erosion will facilitate a greater understanding of the composite Great Unconformity surfaces and the geodynamic mechanisms culpable for their formation.

Missing Temperature: Phanerozoic Burial and Uplift History

Paleozoic Burial Temperature Discrepancies The majority of good-fit t - T paths in our inverse models suggest maximum Phanerozoic burial at the location of the Beartooth Mountains occurred prior to Late Pennsylvanian time (Figures 12 and 13). This is a somewhat surprising result given modest locally preserved sedimentary thicknesses of Paleozoic strata (≤ 1.5 km), far exceeded in thickness by Mesozoic rocks (Figure 8). Using a geothermal gradient of 30 °C, temperatures explainable at the estimated depths of our upper sample locations (e.g., Wise, 2000) from the decompacted maximum thicknesses of preserved strata reach < 80 °C by Late Pennsylvanian time (Figure 8). This is well under the ca. 145 °C burial temperatures required by our thermal models, resulting in a Paleozoic temperature discrepancy of ≥ 65 °C (Figures 11–13). Hypotheses to account for some or all of this missing temperature may include: (1) anomalously high geothermal gradient, perhaps owing to the low temperature conductivity of fine-grained sediment (e.g., Thacker et al., *in press*); (2) thicker accumulations of Paleozoic rocks than that preserved in nearby regions (e.g., Maughan, 1993); and/or (3) greater overburden of metamorphic basement rocks at our sample locations than predicted by previous studies (e.g., Simons and Armbrustmacher, 1976; Wise, 2000), as suggested by AFT thermal model results of Omar et al. (1994; Figure 15).

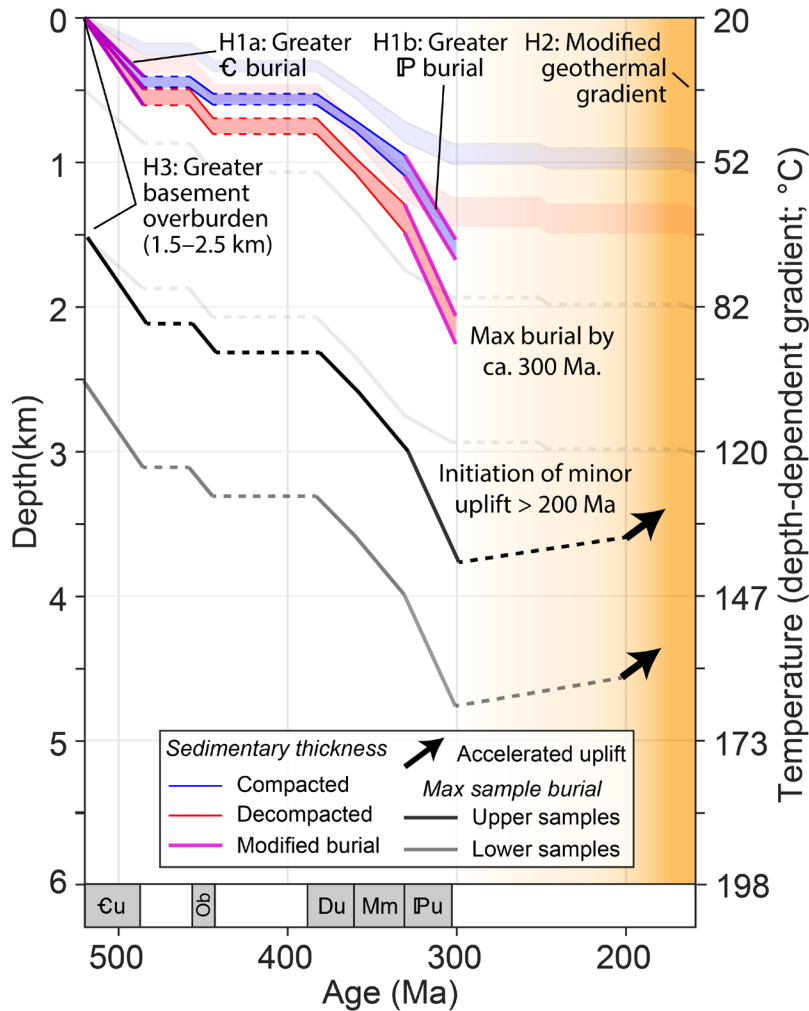


Figure 15: Modified decompacted sedimentary accumulation curves reflecting hypotheses to explain the discrepancy between modeled Phanerozoic burial temperatures: (1) additional Cambrian (1a) or Pennsylvanian (1b) burial; (2) a modified geothermal gradient; (3) greater basement overburden. Original estimates using preserved sedimentary thicknesses and 0.5–1.5 km of basement overburden are shown transparently in background. Shaded orange region represents the timing of gradual (light shading) and more rapid (dark shading) cooling favored by our refined model results. Depth-dependent temperature scale is adapted from the modern southwest African passive margin, a thermally equilibrated passive margin basin (Gholamrezaie et al., 2018). Abbreviations and citations listed in Figure 8 caption.

Abnormally hot geothermal gradients have been recently suggested to explain similar discrepancies between temperatures predicted by decompacted sedimentary accumulation and thermal history model estimates in the Zuni Mountains of New Mexico (Thacker et al., *in press*).

However, we model a temperature offset between our upper and lower elevation sample groups during the Paleozoic of ca. 10–20 °C, suggesting an abnormally cool, rather than hot, geothermal gradient. Strongly depth-dependent geothermal gradients have been recognized in passive margin settings, with the gradient tapering off by > 10 °C/km within the first 6 km of depth (e.g., Gholamrezaie et al., 2018). This may, to some extent, reconcile seemingly incompatible model predictions of moderate temperature offset between sample groups and high burial temperatures (Figure 15). However, thermally equilibrated passive margin basin setting, such as that which existed during early Paleozoic time in Montana, generally have geothermal gradients < 32 °C (e.g., Einsele, 2013; Gholamrezaie et al., 2018). Even the Vøring Basin of Norway, a hyperthermal passive margin basin, reaches a mean geothermal gradient of ca. 40°C only in its upper several kilometers (e.g., Gholamrezaie et al., 2018), which is still incapable of producing our modeled burial temperatures from decompacted Paleozoic sedimentary thicknesses of preserved strata alone (Figure 8).

Burial by additional Paleozoic rocks unaccounted for in preserved stratigraphic thicknesses initially presents a more attractive hypothesis. Isopach maps of Maughan (1993) indicate Cambrian rocks thicken somewhat into the location of the Beartooth Mountains, perhaps suggesting they constituted a local trough during this time. However, the greater estimated thicknesses by these maps are on the order of 10^2 m, not the many kilometers required (Figure 15). Pennsylvanian–Permian time also presents an opportunity for accumulation of unpreserved stratigraphic thickness. The Beartooth Mountains have been implicated in ARM tectonism (e.g., Maughan, 1990). Thick Late Pennsylvanian–Permian successions are common near the periphery of ARM uplifts, such as the Paradox Basin of southeast Utah and southwest Colorado

which reaches thicknesses of ca. 4.5 km in some regions (Nuccio and Condon, 1996). Similar thicknesses would be more than adequate to account for this missing Paleozoic burial temperature assuming even a modest geothermal gradient (Figure 8). However, Maughan (1990) suggests that the location of the Beartooth Mountains comprised an uplifted, rather than down-dropped, region by at least Late Pennsylvanian time (i.e., the “Beartooth Shelf”) based on sedimentologic and stratigraphic observations including thinned or missing Pennsylvanian–Permian stratigraphy near the modern Beartooth Mountains. This interpretation is consistent with Late Pennsylvanian–Permian initiation of uplift in most of our good-fit t - T paths (Figures 12 and 13) but does little to rectify this Paleozoic temperature discrepancy.

It is plausible that the location of the Beartooth Range experienced some—or even considerable—Pennsylvanian sedimentation which has subsequently been removed by pre-Jurassic uplift; this may result if southern regions of the Pennsylvanian–Permian Beartooth shelf experienced uplift prior to northern regions, where our samples were collected. Such uplift patterns would predict flexural subsidence and increased accommodation space at the location of our samples (Figure 15; e.g., Allen and Allen, 1990). However, it seems unlikely that the full several kilometers of sedimentary thickness required to account for this Paleozoic temperature discrepancy were so rapidly deposited and removed, leaving little to no trace in the stratigraphic record.

In consideration of the issues with these former explanations, we examine hypothesis (3)—unaccounted for basement overburden. A 0.5 km and 1.5 km depth of our upper and lower elevation sample groups below the paleo-GU surface, respectively, were used in our modeling for consistency with several sources suggesting the modern Beartooth Plateau surface generally

approximates that of the paleo-GU, with <1 km of basement overburden removed (Simons and Armbrustmacher, 1976; DeCelles et al., 1991b; Wise, 2000; Mueller et al., 2008). However, Omar et al. (1994) reach a different interpretation involving greater (ca. 3 km) basement overburden from analysis of AFT samples collected near the Beartooth Highway. The contrast in these estimates is concerning, as constraining the amount of basement removed from the Beartooth Mountains is necessary before thermal modeling results may be extrapolated to nearby basins and the maturation of hydrocarbons therein (e.g., Peyton and Carrapa, 2013).

In our refined model, we explore hotter temperatures during residence at the Great Unconformity surface more consistent with this latter scenario (Figure 13). Due to the sensitivity of low–moderate eU grains in our dataset to both temperatures of Proterozoic crustal residence and those experienced during Phanerozoic reheating (Figures 9–12), attempting to rectify this Paleozoic temperature discrepancy solely through removed basement overburden is somewhat self-defeating. Additional basement overburden implies higher temperatures during residence at the GU surface, which causes the ZRDAAM model to predict less radiation damage accumulation. Consequently, even higher Paleozoic temperatures are required for sufficient resetting of low–moderate eU grains necessary to match observed data (which in this exercise must be explained by even more basement overburden, leading to recursion).

We find that 60 °C (upper samples) and 80 °C (lower samples) temperatures during Neoproterozoic upper-crustal residence most effectively reproduce our observed date-eU trend without requiring unrealistically high Paleozoic temperatures inconsistent with other data, such as a preserved fossil AFT PAZ (Omar et al., 1994). These upper-crustal residence temperatures are consistent with ca. 1.5–2.5 km of removed basement overburden from the Beartooth

Plateau—an estimate intermediary between the two described above (Figure 15). The discrepancy between this and lower estimates (i.e., 0.5 km of Wise, 2000) may be explainable by a steeper fault-propagation fold geometry than that assumed by the former authors, perhaps accommodated by folding along the Beartooth sag zone (e.g., Foose et al., 1961) or significant offset on an unrecognized fault separating locations of preserved Paleozoic strata and/or samples collected near the Beartooth Highway from our sample locations further west.

In summary, the observed discrepancy between Paleozoic burial temperatures predicted by our decompacted sedimentary accumulation models (Figure 8) and thermal models (Figures 12 and 13) is most likely attributable to some combination of the above hypotheses. In Figure 15, we demonstrate a hypothetical scenario broadly consistent with these hypotheses and our model results. However, some modification of geothermal gradient with depth and with time is required to reproduce modeled temperature offset and upper-crustal residence temperatures, respectively.

Nature and Timing of Phanerozoic Uplift Our refined thermal model results suggest that the later Phanerozoic history of the Beartooth Range comprised subdued Paleozoic–Triassic cooling (here a proxy for uplift) which increased in rate and magnitude during Jurassic time, followed by rapid cooling in the Late Cretaceous–Paleocene (Figure 13). This is somewhat at odds with previous models resolving only late Mesozoic–early Cenozoic cooling (e.g., Cervený, 1990; Omar et al., 1994; Peyton et al., 2012). However, our model results draw support from several prior observations:

(1) Maughan (1990) interprets that the Beartooth Range comprised an elevated lowland region by at least Late Pennsylvanian into Permian time based on interpreted provenance and sedimentary thickness patterns of late Paleozoic rocks in Montana. Pennsylvanian–Permian

initiation of cooling may have been associated with far-field effects of ARM tectonism, producing uplift responsible for these stratigraphic and sedimentologic observations, including erosionally thinned or missing strata of this age on the flanks of the range (e.g., Maughan, 1990).

(2) Cervený (1990) and Omar et al. (1994) document the preservation of a fossil AFT PAZ at high elevations on the Beartooth Plateau (Figure 5). Samples from this PAZ record AFT dates of up to ca. 340 ± 40 Ma (Cervený, 1990), which requires residence of these samples above the base of the AFT PAZ (< 125 °C; Reiners and Brandon, 2006 and references therein) since at least late Pennsylvanian time.

(3) Fuentes et al. (2011) interpret a “broken back-bulge”, segmented by pre-existing basement highs during Late Oxfordian–Kimmeridgian deposition in the Montana foreland basin. The quickening pace of predicted cooling in Early Jurassic time may be attributable to local uplift accommodating tectonic stress associated with lithospheric flexure of the foreland basin (e.g., Fuentes et al., 2011). Such localized uplift is supported by intraformational unconformities in Lower Jurassic strata above similar basement features (e.g., Parcell and Williams, 2005). This high topography and/or uplift may have persisted into Late Jurassic time, perhaps associated with migration of the forebulge into the area, producing convex lithospheric flexure defined by eroded or thinned sedimentary rocks; evidence for passage of the forebulge during this time interval is supported by the angular unconformity between Jurassic and Cretaceous strata in the Gravelly Range to the southwest of the Beartooth Mountains (DeCelles and Burden, 1992).

(4) DeCelles (1986) interprets that intra-foreland uplifts partitioned the foredeep depozone by Early Cretaceous time. He reports paleocurrents in the Kootenai Formation strongly deflected around the location of the Beartooth Mountains, suggesting that they formed a regional

high during this time. Early Cretaceous uplift may also draw support from detrital zircon provenance interpretations; the well-integrated provenance signature of Jurassic foreland basin strata in Montana become more spatially variable along strike by Early Cretaceous time, consistent with tectonic partitioning of the foreland basin (e.g., Ronemus et al., 2020). Cooling predicted by Early Cretaceous uplift is consistent with AFT models of Carrapa et al. (2019) suggesting substantial cooling of the Beartooth Mountains as early as 120 Ma.

Our refined forward models additionally incorporate a brief period of reheating during Late Cretaceous time (Figure 13A), required to reproduce the bimodal age distributions of AFT samples from high elevations of Omar et al. (1994; Figure 13C). This may suggest the ongoing modest cooling—and implied exhumation—of the Beartooth Mountains could not keep pace with voluminous Late Cretaceous sedimentation and were consequently minimally buried in this time. Alternatively, dynamic subsidence from mantle activity implicated in the rapid subsidence of the Western Interior Seaway (e.g., Liu et al., 2014) may have caused brief inversion of the Beartooth Mountains. Either interpretation is consistent with recently reported isopach data from Cretaceous rocks of central Montana (Eichler et al., 2020). These isopach maps suggest the Beartooth Mountains formed a local high during Coniacian–early Campanian time, with sedimentary thicknesses rapidly decreasing toward the location of the range. However, this trend breaks down in Late Campanian time, when models suggest regions proximal to the Beartooth Mountains were buried by up to 800 m of sediment (Eichler et al., 2020).

Our inverse thermal model results predict this brief period of late Campanian burial was followed by rapid cooling by Maastrichtian time, with our refined models incorporating additional AFT data of Omar et al. (1994) suggesting initiation of cooling at ca. 70 Ma and

termination at 57 Ma (Figure 13A). The timing of this rapid cooling is broadly consistent with existing AFT and AHe thermochronology data, which position the initiation of rapid uplift between ca. 120 Ma and 60 Ma (e.g., Cervený, 1990; Omar et al., 1994; Peyton et al., 2012). Additionally, magnitudes of cooling predicted by our models are generally compatible with balanced cross sections of DeCelles et al. (1991b) which estimate a total of ca. 5.5 km of Late Cretaceous–Paleogene uplift. Our refined models cool from 100 °C (upper samples) and 120 °C (lower samples) to near-surface conditions during this episode and predict a temperature offset between our upper and lower elevation samples of ca. 15 °C. Using this as a rough approximate for the paleo-geothermal gradient, we estimate a similar ca. 5.6 km of uplift during Maastrichtian–Thanetian time.

Eocene–Miocene burial is suggested by most of our good-fit inverse and refined forward models (Figures 12 and 13) and predicted by previous AFT data (Giegengack et al., 1986; Omar et al., 1994; Carrapa et al., 2019). This is consistent with burial by Eocene volcanic rocks—such as the Absaroka volcanics—and/or volcanism associated with early caldera-forming eruptions of the Snake River Plain (i.e., McDermitt caldera; e.g., Dudás et al., 2010; Henry et al., 2017). This burial is geologically supported by the preservation of Eocene Absaroka volcanics in the southern portion of the Beartooth Mountains (e.g., Foote et al., 1961), which reach thicknesses of several kilometers in the Absaroka Mountains to the south (e.g., Feeley, 2003).

Subsequent cooling to surface temperatures could feasibly be attributable to glaciation associated with the Yellowstone Ice Cap, which has been implicated in incision of major valleys in the range (e.g., Graf, 1970). However, model results suggest that the thermal offset between our sample elevation groups reached low values prior to this burial—required to fit the high-eU

portion of our dataset (Figure 11). This suggests much of the incision defining the modern topography of the range took place prior to Cenozoic reheating. In consideration of this, a less focused erosional mechanism leading to late Cenozoic cooling may be more parsimonious. The encroachment of the Yellowstone hot spot into the region of the Beartooth Mountains has been geodetically demonstrated to produce uplift in nearby regions (e.g., Dzurisin et al., 2012). Additionally, watershed analysis suggests this feature resulted in increased prominence of nearby basement-involved uplifts, including the Beartooth Mountains (Wegmann et al., 2007). This indicates that the modern elevations of the range and cooling likely associated with erosion of Cenozoic volcanic cover may have responded to this mantle feature.

Cenozoic Basin and Range extension may have also played a role in this late-stage cooling to surface conditions predicted by our refined thermal history model results. The western margin of the Beartooth Mountains is bordered by a series of SW-striking normal faults down-dropping the Paradise Valley, > 40 km west of our sample locations (e.g., Pardee, 1950; Berg et al., 2000). Similar structures of the Basin and Range province regionally accommodated at least 10^2 km-scale crustal extension, dominantly during Miocene–recent time (e.g., Long, 2018 and references therein). Extension in southwestern Montana resulted in widespread drainage reorganization and incision of orogenic topography (e.g., Janecke, 2007). Although our samples were not collected in the proximal footwall of a major documented extensional structure, regional patterns of extension—as well as perhaps more local effects from the down-dropping of the Paradise Valley—may have contributed to the erosional removal of the Cenozoic volcanic carapace from the Beartooth Mountains and cooling at the location of our samples. Although a high geothermal gradient is associated both with the Yellowstone hotspot (e.g., Anders and

Sleep, 1992) and Basin and Range extension (e.g., Lachenbruch, 1979), cooling from erosional exhumation associated with one of both of these phenomena may have trumped the heating signal related to a higher geothermal gradient. The Beartooth Mountains likely comprised a zone of uplifted high-relief topography by Paleocene time and would have been relatively insulated from transient modifications to the regional geothermal gradient (e.g., Braun, 2002).

Collectively these interpretations for the Phanerozoic history of the Beartooth Range illustrate a history somewhat at odds with those involving sustained, progressive burial in a relatively tectonically quiescent setting prior to sudden and rapid initiation of uplift in latest Mesozoic or Paleocene time (e.g., Cervený, 1990; Omar et al., 1994). While Cambrian–Ordovician sedimentation may reflect passive margin deposition *sensu stricto*, Pennsylvanian–Permian cooling resolved by our thermal models suggests the Beartooth Mountains comprised a tectonically active feature by at least late Paleozoic time. The location of the Beartooth Mountains likely experienced a different thermal and erosional history than surrounding basins for much of the remainder of Phanerozoic time, perhaps remaining elevated during much of Mesozoic time as surrounding areas rapidly subsided. This may indicate that the faults which accommodated the large magnitudes of uplift lending the Beartooths their impressive modern prominence are not merely Mesozoic–Cenozoic features; rather, inherited crustal weaknesses—perhaps dating back to Paleoproterozoic tectonism suggested by our $^{40}\text{Ar}/^{39}\text{Ar}$ results and/or Mesoproterozoic extension (e.g., Mueller et al., 2016)—reactivated by subsequent tectonism more readily explains this pattern of subsidence and uplift out of sync with nearby regions (e.g., O’Neill et al., 1986). Lithospheric stresses associated with geodynamic events ranging from at least Pennsylvanian Ancestral Rocky Mountains tectonism to perhaps modern uplift associated

with Yellowstone mantle activity and/or Basin and Range extension likely found accommodation in these ancestral features.

Conclusions

We demonstrate the utility of a multi-chronometer approach involving diverse modeling techniques in reconstructing complex and temporally extensive thermal histories of regions near the margins of cratons. A dataset of dispersed but systematically correlated ZHe dates collected across a range of elevations facilitates the exploration of deep-time thermal histories that are unresolvable by lower temperature techniques (AHe and AFT) alone. Our results show evidence of previously unrecognized thermotectonism and cooling experienced by rocks of the northwestern Wyoming craton and constrain the timing, and perhaps geodynamic mechanism, of documented erosion, burial, and uplift events. Constraints on these events from this work include the following:

(1) Western regions of Beartooth Mountains experienced Paleoproterozoic thermal overprinting to temperatures above the biotite $^{40}\text{Ar}/^{39}\text{Ar}$ closure window ($> 300\text{ }^{\circ}\text{C}$). We postulate that this reheating was associated with thermotectonism of the Big Sky orogenic event (e.g., Harms et al., 2004, 2006; Condit et al., 2015) and/or related mantle metasomatism or mafic underplating (e.g., Rudnick et al., 1998; Downes et al., 2004; Facer et al., 2009; Gifford et al., 2014).

(2) The location of the Beartooth Mountains experienced a multi-phase Proterozoic cooling history. We suggest a plausible mechanism for the first phase of Proterozoic cooling may include post-orogenic collapse following Paleoproterozoic orogenesis. Later Neoproterozoic cooling is constrained to ca. 1.4–0.8 Ga, precluding mechanisms such as Snowball Earth glacial

erosion which predict later Neoproterozoic cooling (e.g., Keller et al., 2019) and supports previous interpretations of some degree of diachroneity in the timing of large-magnitude pre-Great Unconformity continental erosion across Laurentia (e.g., DeLucia et al., 2018; Flowers et al., 2020).

(3) The location of the Beartooth Mountains likely experienced maximum Phanerozoic burial temperatures by the end of Pennsylvanian time. This estimate is far earlier (> 200 Myr) than may be demonstrated for nearby regions (e.g., Maughan, 1993). Subsequent Paleozoic cooling indicated by our model results may be associated with Ancestral Rocky Mountains tectonism. Generally continuous cooling at a moderate pace during Triassic–Early Cretaceous time—with later phases perhaps associated with lithospheric stresses in the foreland basin—was followed by rapid cooling during Cretaceous time, likely punctuated by a brief period of largely Campanian reburial. This result reconciles previous observations indicating pre-Late Cretaceous exhumation (DeCelles, 1986; Carrapa et al., 2019) with those only resolving later phases of exhumation and uplift (Cerveny, 1990; Omar et al., 1994).

(4) The Cenozoic history of the Beartooth Mountains likely involves some degree of minor reheating (≤ 70 °C), perhaps attributable to burial by Cenozoic rocks such as Eocene Absaroka volcanics preserved in the southern Beartooth Mountains. Subsequent Neogene–Quaternary cooling was likely associated with the removal of these rocks. We suggest the latter thermal event may owe to dynamic uplift associated with the encroachment of the Yellowstone hotspot—consistent with previous geodetic and geomorphic results (e.g., Wegmann et al., 2007; Dzurisin et al., 2012)—and/or increased erosion related to regional Basin and Range extension (e.g., Janecke, 2007).

Collectively, the addition of these constraints contributes to the development of an integrated thermal history for this region of Laurentia, spanning at least Paleoproterozoic to recent time. This illustrates a thermal history of a region near the western Laurentian margin strongly coupled to geodynamic processes associated with both the tectonics of Earth's lithosphere as well as, perhaps, mantle activity originating from deeper within the planet.

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Literature Cited

- Alcock, J., Muller, P.D., and Jercinovic, M.J., 2013, Monazite ages and pressure–temperature–time paths from anatectites in the southern Ruby Range, Montana, USA: evidence for delamination, ultramafic magmatism, and rapid uplift at ca. 1780 Ma: *Canadian Journal of Earth Sciences*, doi:10.1139/cjes-2013-0035.
- Allen, P., and Allen, J., 1990, *Basin analysis: principles and applications*, Blackwell Scientific Publications: Oxford, London.
- Anders, M.H., and Sleep, N.H., 1992, Magmatism and extension: The thermal and mechanical effects of the Yellowstone Hotspot: *Journal of Geophysical Research: Solid Earth*, v. 97, p. 15379–15393, doi:<https://doi.org/10.1029/92JB01376>.
- Anderson, C.J., 2017, Stratigraphic framework and provenance of the lower belt (Newland Formation), Belt Supergroup, Helena Embayment, Central Montana [Master of Science]: Montana State University - Bozeman, College of Letters & Science.
- Anderson, H.E., and Davis, D.W., 1995, U–Pb geochronology of the Moyie sills, Purcell Supergroup, southeastern British Columbia: implications for the Mesoproterozoic geological history of the Purcell (Belt) basin: *Canadian Journal of Earth Sciences*, v. 32, p. 1180–1193, doi:10.1139/e95-097.
- Ault, A.K., Flowers, R.M., and Mahan, K.H., 2012, Quartz shielding of sub-10 μ m zircons from radiation damage-enhanced Pb loss: An example from a metamorphosed mafic dike, northwestern Wyoming craton: *Earth and Planetary Science Letters*, v. 339–340, p. 57–66, doi:10.1016/j.epsl.2012.04.025.
- Ayers, W.B., Jr., 1986, Lacustrine and Fluvial-Deltaic Depositional Systems, Fort Union Formation (Paleocene), Powder River Basin, Wyoming and Montana: *AAPG Bulletin*, v. 70, p. 1651–1673, doi:10.1306/94886C90-1704-11D7-8645000102C1865D.
- Baadsgaard, H., and Mueller, P.A., 1973, K-Ar and Rb-Sr Ages of Intrusive Precambrian Mafic Rocks, Southern Beartooth Mountains, Montana and Wyoming: *GSA Bulletin*, v. 84, p. 3635–3644, doi:10.1130/0016-7606(1973)84<3635:KARAOI>2.0.CO;2.
- Barker, C., 1996, Thermal modeling of petroleum generation. Theory and applications, <https://www.osti.gov/etdweb/biblio/557106> (accessed March 2021).
- Bell, L.H., and Middleton, L.T., 1978, An Introduction to the Cambrian Flathead Sandstone, Wind River Basin, Wyoming, in *Resources of the Wind River Basin*, 30th Annual Field Conference Guidebook, Wyoming Geological Association, p. 79–88.
- Beranek, L.P., Link, P.K., and Fanning, C.M., 2016, Detrital zircon record of mid-Paleozoic convergent margin activity in the northern U.S. Rocky Mountains: Implications for the Antler orogeny and early evolution of the North American Cordillera: *Lithosphere*, v. 8, p. 533–550, doi:10.1130/L557.1.

- Berg, R.B., Lopez, D.A., and Lonn, J.D., 2000, Geologic map of the Livingston 30' x 60' quadrangle, south-central Montana: Montana Bureau of Mines and Geology, Open File Report MBMG, v. 406, p. 21.
- Bevan, A., 1923, Summary of the geology of the Beartooth Mountains, Montana: *The Journal of Geology*, v. 31, p. 441–465.
- Bird, P., 1984, Laramide crustal thickening event in the Rocky Mountain Foreland and Great Plains: *Tectonics*, v. 3, p. 741–758, doi:<https://doi.org/10.1029/TC003i007p00741>.
- Boerner, D.E., Craven, J.A., Kurtz, R.D., Ross, G.M., and Jones, F.W., 1998, The Great Falls Tectonic Zone: suture or intracontinental shear zone? *Canadian Journal of Earth Sciences*, v. 35, p. 175–183, doi:10.1139/e97-104.
- Bonnichsen, B., Leeman, W.P., Honjo, N., McIntosh, W.C., and Godchaux, M.M., 2008, Miocene silicic volcanism in southwestern Idaho: geochronology, geochemistry, and evolution of the central Snake River Plain: *Bulletin of Volcanology*, v. 70, p. 315–342, doi:10.1007/s00445-007-0141-6.
- Brady, J., 2004, Precambrian Geology of the Tobacco Root Mountains, Montana: Geological Society of America, 270 p.
- Brasier, M.D., 1980, The Lower Cambrian transgression and glauconite-phosphate facies in western Europe: *Journal of the Geological Society*, v. 137, p. 695–703, doi:10.1144/gsjgs.137.6.0695.
- Braun, J., 2002, Quantifying the effect of recent relief changes on age–elevation relationships: *Earth and Planetary Science Letters*, v. 200, p. 331–343, doi:10.1016/S0012-821X(02)00638-6.
- Bricker, A.L., 2016, (U-Th)/He thermochronology of the Yellowstone Crescent High Terrain: Implications for Miocene to recent topographic impacts associated with the Yellowstone Hotspot [Masters Thesis]: University of Florida.
- Brocks, J.J., Jarrett, A.J.M., Sirantoine, E., Hallmann, C., Hoshino, Y., and Liyanage, T., 2017, The rise of algae in Cryogenian oceans and the emergence of animals: *Nature*, v. 548, p. 578–581, doi:10.1038/nature23457.
- Brown, R.W., Beucher, R., Roper, S., Persano, C., Stuart, F., and Fitzgerald, P., 2013, Natural age dispersion arising from the analysis of broken crystals. Part I: Theoretical basis and implications for the apatite (U–Th)/He thermochronometer: *Geochimica et Cosmochimica Acta*, v. 122, p. 478–497, doi:10.1016/j.gca.2013.05.041.
- Burger, H.R., 2004, General geology and tectonic setting of the Tobacco Root Mountains, in Special Paper 377: Precambrian Geology of the Tobacco Root Mountains, Montana, Geological Society of America, v. 377, p. 1–14, doi:10.1130/0-8137-2377-9.1.
- Carlson, R., and Irving, A., 1994, Depletion and enrichment history of subcontinental lithospheric mantle: an Os, Sr, Nd and Pb isotopic study of ultramafic xenoliths from the northwestern Wyoming Craton: *Earth and Planetary Science Letters*, v. 126, p. 457–472.

- Carrapa, B., DeCelles, P.G., and Romero, M., 2019, Early Inception of the Laramide Orogeny in Southwestern Montana and Northern Wyoming: Implications for Models of Flat-Slab Subduction: *Journal of Geophysical Research: Solid Earth*, v. 124, p. 2102–2123, doi:10.1029/2018JB016888.
- Carrapa, B., Wijbrans, J., and Bertotti, G., 2003, Episodic exhumation in the Western Alps: *Geology*, v. 31, p. 601–604, doi:10.1130/0091-7613(2003)031<0601:EEITWA>2.0.CO;2.
- Cerveny, P.F., 1990, Fission-track thermochronology of the Wind River Range and other basement cored uplifts in the Rocky Mountain foreland [Ph.D. thesis]: University of Wyoming, 180 p.
- Cerveny, P.F., and Steidtmann, J.R., 1993, Fission track thermochronology of the Wind River Range, Wyoming: Evidence for timing and magnitude of Laramide exhumation: *Tectonics*, v. 12, p. 77–91, doi:10.1029/92TC01567.
- Chamberlain, K.R., Frost, C.D., and Frost, B.R., 2003, Early Archean to Mesoproterozoic evolution of the Wyoming Province: Archean origins to modern lithospheric architecture: *Canadian Journal of Earth Sciences*, v. 40, p. 1357–1374, doi:10.1139/e03-054.
- Chaudhuri, S., and Brookins, D.G., 1969, The isotopic age of the Flathead Sandstone (Middle Cambrian), Montana: *Journal of Sedimentary Research*, v. 39, p. 364–366, doi:10.1306/74D71C76-2B21-11D7-8648000102C1865D.
- Cheney, J.T. et al., 2004, Proterozoic metamorphism of the Tobacco Root Mountains, Montana, in *Special Paper 377: Precambrian Geology of the Tobacco Root Mountains, Montana*, Geological Society of America, v. 377, p. 105–129, doi:10.1130/0-8137-2377-9.105.
- Colpron, M., Nelson, J.L., and Murphy, D.C., 2007, Northern Cordilleran terranes and their interactions through time: *GSA Today*, v. 17, p. 4, doi:10.1130/GSAT01704-5A.1.
- Condie, K.C., 1976, Trace-element geochemistry of archaean greenstone belts: *Earth-Science Reviews*, v. 12, p. 393–417, doi:10.1016/0012-8252(76)90012-X.
- Condie, K.C., Leech, A.P., and Baadgaard, H., 1969, Potassium-Argon Ages of Precambrian Mafic Dikes in Wyoming: *Geological Society of America Bulletin*, v. 80, p. 899, doi:10.1130/0016-7606(1969)80[899:PAOPMD]2.0.CO;2.
- Condit, C.B., Mahan, K.H., Ault, A.K., and Flowers, R.M., 2015, Foreland-directed propagation of high-grade tectonism in the deep roots of a Paleoproterozoic collisional orogen, SW Montana, USA: *Lithosphere*, p. L460.1, doi:10.1130/L460.1.
- Cox, G.M., Halverson, G.P., Stevenson, R.K., Vokaty, M., Poirier, A., Kunzmann, M., Li, Z.-X., Denyszyn, S.W., Strauss, J.V., and Macdonald, F.A., 2016, Continental flood basalt weathering as a trigger for Neoproterozoic Snowball Earth: *Earth and Planetary Science Letters*, v. 446, p. 89–99, doi:10.1016/j.epsl.2016.04.016.
- Dahl, P., Frei, R., and Dorais, M., 1998, When did the Wyoming Province collide with Laurentia? New clues from step-leach Pb-Pb dating of garnet independent of its inclusions, in *Geological Society of America Abstracts with Programs*, v. 30, p. 109.

- Danišik, M., McInnes, B.I.A., Kirkland, C.L., McDonald, B.J., Evans, N.J., and Becker, T., 2017, Seeing is believing: Visualization of He distribution in zircon and implications for thermal history reconstruction on single crystals: *Science Advances*, v. 3, p. e1601121, doi:10.1126/sciadv.1601121.
- DeCelles, P.G., 2004, Late Jurassic to Eocene evolution of the Cordilleran thrust belt and foreland basin system, western U.S.A.: *American Journal of Science*, v. 304, p. 105–168, doi:10.2475/ajs.304.2.105.
- DeCelles, P.G., 1986, Sedimentation in a tectonically partitioned, nonmarine foreland basin: The Lower Cretaceous Kootenai Formation, southwestern Montana: *GSA Bulletin*, v. 97, p. 911–931, doi:10.1130/0016-7606(1986)97<911:SIATPN>2.0.CO;2.
- DeCelles, P.G., and Burden, E.T., 1992, Non-marine sedimentation in the overfilled part of the Jurassic-Cretaceous Cordilleran foreland basin: Morrison and Cloverly Formations, central Wyoming, USA: *Basin Research*, v. 4, p. 291–314, doi:10.1111/j.1365-2117.1992.tb00050.x.
- DeCelles, P.G., Gray, M.B., Ridgway, K.D., Cole, R.B., Pivnik, D.A., Pequera, N., and Srivastava, P., 1991a, Controls on synorogenic alluvial-fan architecture, Beartooth Conglomerate (Palaeocene), Wyoming and Montana: *Sedimentology*, v. 38, p. 567–590, doi:10.1111/j.1365-3091.1991.tb01009.x.
- DeCelles, P.G., Gray, M.B., Ridgway, K.D., Cole, R.B., Srivastava, P., Pequera, N., and Pivnik, D.A., 1991b, Kinematic history of a foreland uplift from Paleocene synorogenic conglomerate, Beartooth Range, Wyoming and Montana: *GSA Bulletin*, v. 103, p. 1458–1475, doi:10.1130/0016-7606(1991)103<1458:KHOAFU>2.3.CO;2.
- DeLucia, M.S., Guenther, W.R., Marshak, S., Thomson, S.N., and Ault, A.K., 2018, Thermochronology links denudation of the Great Unconformity surface to the supercontinent cycle and snowball Earth: *Geology*, v. 46, p. 167–170, doi:10.1130/G39525.1.
- Dickinson, W.R., 2004, Evolution of the North American Cordillera: *Annual Review of Earth and Planetary Sciences*, v. 32, p. 13–45, doi:10.1146/annurev.earth.32.101802.120257.
- Dickinson, W.R., and Snyder, W.S., 1978, Plate tectonics of the Laramide orogeny, in Matthews, V. ed., *Laramide folding associated with basement block faulting in the western United States*, *Geological Society of America Memoir*, v. 151, p. 355–366.
- Dorobek, S.L., Reid, S.K., Elrick, M., Bond, G.C., and Kominz, M.A., 1991, Subsidence across the Antler foreland of Montana and Idaho: Tectonic versus eustatic effects: *Kansas Geological Survey Bulletin*, v. 233, p. 232–251.
- Downes, H., Macdonald, R., Upton, B.G., Cox, K.G., Bodinier, J.-L., Mason, P.R., James, D., Hill, P.G., and Hearn Jr, B.C., 2004, Ultramafic xenoliths from the Bearpaw Mountains, Montana, USA: Evidence for multiple metasomatic events in the lithospheric mantle beneath the Wyoming craton: *Journal of Petrology*, v. 45, p. 1631–1662.

- Dudás, F.Ö., Ispolatov, V.O., Harlan, S.S., and Snee, L.W., 2010, 40Ar/39Ar Geochronology and Geochemical Reconnaissance of the Eocene Lowland Creek Volcanic Field, West-Central Montana: *The Journal of Geology*, v. 118, p. 295–304, doi:10.1086/651523.
- Dzurisin, D., Wicks, C.W., and Poland, M.P., 2012, History of surface displacements at the Yellowstone Caldera, Wyoming, from leveling surveys and InSAR observations, 1923–2008: Professional Paper Report 1788, doi:10.3133/pp1788.
- Eichler, C.M., Saylor, J., Julia, W., Rudolph, K., and Davis, E.M., 2020, Along-strike variation in geometry of the late Cretaceous western interior basin tracks multiple geodynamic subsidence mechanisms, in Geological Society of America.
- Einsele, G., 2013, Basin classification and depositional environments (Overview), *in* Sedimentary basins: evolution, facies, and sediment budget, Springer Science & Business Media.
- Erslev, E.A., and Sutter, J.F., 1990, Evidence for Proterozoic mylonitization in the northwestern Wyoming province: *GSA Bulletin*, v. 102, p. 1681–1694, doi:10.1130/0016-7606(1990)102<1681:EFPMIT>2.3.CO;2.
- Evans, K.V., Aleinikoff, J.N., Obradovich, J.D., and Fanning, C.M., 2000, SHRIMP U-Pb geochronology of volcanic rocks, Belt Supergroup, western Montana: evidence for rapid deposition of sedimentary strata: *Canadian Journal of Earth Sciences*, v. 37, p. 1287–1300, doi:10.1139/e00-036.
- Evans, R.D., Goodman, C., Keevil, N.B., and Urry, W.D., 1939, Work at Massachusetts Institute of Technology, Exhibit 3, in Report of the Committee on the Measurement of Geologic Time 1937-1938, Appendix J, in Annual Report of the Division of Geology and Geography: National Research Council.
- Facer, J., Downes, H., and Beard, A., 2009, In situ serpentinization and hydrous fluid metasomatism in spinel dunite xenoliths from the Bearpaw Mountains, Montana, USA: *Journal of Petrology*, v. 50, p. 1443–1475.
- Fan, M., and Carrapa, B., 2014, Late Cretaceous-early Eocene Laramide uplift, exhumation, and basin subsidence in Wyoming: Crustal responses to flat slab subduction: Laramide uplift and exhumation: *Tectonics*, v. 33, p. 509–529, doi:10.1002/2012TC003221.
- Feeley, T.C., 2003, Origin and Tectonic Implications of Across-Strike Geochemical Variations in the Eocene Absaroka Volcanic Province, United States: *The Journal of Geology*, v. 111, p. 329–346, doi:10.1086/373972.
- Flowers, R.M., Farley, K.A., and Ketcham, R.A., 2015, A reporting protocol for thermochronologic modeling illustrated with data from the Grand Canyon: *Earth and Planetary Science Letters*, v. 432, p. 425–435, doi:10.1016/j.epsl.2015.09.053.
- Flowers, R.M., Ketcham, R.A., Shuster, D.L., and Farley, K.A., 2009, Apatite (U–Th)/He thermochronometry using a radiation damage accumulation and annealing model: *Geochimica et Cosmochimica Acta*, v. 73, p. 2347–2365, doi:10.1016/j.gca.2009.01.015.

- Flowers, R.M., Macdonald, F.A., Siddoway, C.S., and Havranek, R., 2020, Diachronous development of Great Unconformities before Neoproterozoic Snowball Earth: *Proceedings of the National Academy of Sciences*, p. 201913131, doi:10.1073/pnas.1913131117.
- Flueckinger, L.A., 1970, Stratigraphy, petrography, and origin of Tertiary sediments off the front of the Beartooth Mountains, Montana-Wyoming [PhD Thesis]: Pennsylvania State University.
- Foose, R.M., Wise, D.U., and Garbarini, G.S., 1961, Structural Geology of the Beartooth Mountains, Montana and Wyoming: *GSA Bulletin*, v. 72, p. 1143–1172, doi:10.1130/0016-7606(1961)72[1143:SGOTBM]2.0.CO;2.
- Foster, D.A., Mueller, P.A., Mogk, D.W., Wooden, J.L., and Vogl, J.J., 2006, Proterozoic evolution of the western margin of the Wyoming craton: implications for the tectonic and magmatic evolution of the northern Rocky Mountains, doi:10.1139/e06-052.
- Fox, N.R., 2017, Sedimentary tectonics of the Mesoproterozoic LaHood Formation, southwest Montana: M.Sc. Thesis, v. Montana State University, p. 164.
- Fuentes, F., DeCelles, P.G., Constenius, K.N., and Gehrels, G.E., 2011, Evolution of the Cordilleran foreland basin system in northwestern Montana, U.S.A.: *GSA Bulletin*, v. 123, p. 507–533, doi:10.1130/B30204.1.
- Gast, P.W., Kulp, J.L., and Long, L.E., 1958, Absolute age of early Precambrian rocks in the Bighorn Basin of Wyoming and Montana, and southeastern Manitoba: *Eos, Transactions American Geophysical Union*, v. 39, p. 322–334, doi:10.1029/TR039i002p00322.
- Gehrels, G.E., Valencia, V.A., and Ruiz, J., 2008, Enhanced precision, accuracy, efficiency, and spatial resolution of U-Pb ages by laser ablation–multicollector–inductively coupled plasma–mass spectrometry: *Geochemistry, Geophysics, Geosystems*, v. 9, doi:10.1029/2007GC001805.
- Geissman, J.G., and Mogk, D., 1986, Late Archean tectonic emplacement of the Stillwater Complex along reactivated basement structures, northern Beartooth Mountains southern Montana, U.S., in Aldrich, J.J.Jr. and Laughlin, A.W. eds., 6th International Conference on Basement Tectonics, p. 25–44.
- Geyer, G., and Landing, E., 2017, The Precambrian–Phanerozoic and Ediacaran–Cambrian boundaries: a historical approach to a dilemma: *Geological Society, London, Special Publications*, v. 448, p. 311–349, doi:10.1144/SP448.10.
- Gholamrezaie, E., Scheck-Wenderoth, M., Sippel, J., and Strecker, M.R., 2018, Variability of the geothermal gradient across two differently aged magma-rich continental rifted margins of the Atlantic Ocean: the Southwest African and the Norwegian margins: *Solid Earth*, v. 9, p. 139–158, doi:10.5194/se-9-139-2018.
- Giegengack, R., Omar, G.I., and Johnson, K.R., 1986, A reconnaissance fission track uplift chronology for the northwest margin of the Bighorn Basin: Montana Geological Society and yellowstone bighorn Research Association Joint Field Conference and Symposium: *Geology of the Beartooth Uplift and Adjacent Basins*, p. 179–184.

- Gifford, J.N., Mueller, P.A., Foster, D.A., and Mogk, D.W., 2014, Precambrian Crustal Evolution in the Great Falls Tectonic Zone: Insights from Xenoliths from the Montana Alkali Province: *The Journal of Geology*, v. 122, p. 531–548, doi:10.1086/677262.
- Giletti, B.J., 1971, Discordant isotopic ages and excess argon in biotites: *Earth and Planetary Science Letters*, v. 10, p. 157–164, doi:10.1016/0012-821X(71)90001-X.
- Giletti, B.J., 1966, Isotopic ages from southwestern Montana: *Journal of Geophysical Research*, v. 71.
- Gingerich, P.D., 1983, Paleocene-Eocene Faunal Zones and a Preliminary Analysis of Laramide Structural Deformation in the Clark's Fork Basin, Wyoming: *Geology Of the Bighorn Basin; 34th Annual Field Conference Guidebook*, p. 185–195.
- Ginster, U., Reiners, P.W., Nasdala, L., and Chanmuang N., C., 2019, Annealing kinetics of radiation damage in zircon: *Geochimica et Cosmochimica Acta*, v. 249, p. 225–246, doi:10.1016/j.gca.2019.01.033.
- Goddéris, Y., Donnadieu, Y., Nédélec, A., Dupré, B., Dessert, C., Grard, A., Ramstein, G., and François, L.M., 2003, The Sturtian 'snowball' glaciation: fire and ice: *Earth and Planetary Science Letters*, v. 211, p. 1–12, doi:10.1016/S0012-821X(03)00197-3.
- Gómez, E., Jordan, T.E., Allmendinger, R.W., and Cardozo, N., 2005, Development of the Colombian foreland-basin system as a consequence of diachronous exhumation of the northern Andes: *GSA Bulletin*, v. 117, p. 1272–1292, doi:10.1130/B25456.1.
- Graf, W.L., 1970, The Geomorphology of the Glacial Valley Cross Section: *Arctic and Alpine Research*, v. 2, p. 303–312, doi:10.1080/00040851.1970.12003589.
- Green, D., 1972, Morphologies and fission track ages of zircons from quartzites of the Hellroaring Lakes area, Pine Creek quarry, and Chrome Mountain area, Beartooth Mountains, Montana [Masters Thesis]: University of Cincinnati.
- Grove, M., and Harrison, T.M., 1996, $^{40}\text{Ar}^*$ diffusion in Fe-rich biotite: *American Mineralogist*, v. 81, p. 940–951.
- Guenther, W.R., 2021, Implementation of an Alpha Damage Annealing Model for Zircon (U-Th)/He Thermochronology With Comparison to a Zircon Fission Track Annealing Model: *Geochemistry, Geophysics, Geosystems*, v. 22, p. e2019GC008757, doi:10.1029/2019GC008757.
- Guenther, W.R., Reiners, P.W., Ketcham, R.A., Nasdala, L., and Giester, G., 2013, Helium diffusion in natural zircon: Radiation damage, anisotropy, and the interpretation of zircon (U-Th)/He thermochronology: *American Journal of Science*, v. 313, p. 145–198, doi:10.2475/03.2013.01.
- Guerrero, J.C., Mueller, P.A., and Mogk, D.W., 2016, Provenance study and geochemical analysis of the LaHood Formation of the Bridger Range, Montana: *Northwest Geology*, v. 45, p. 21–28.

- Guy, R., and Sinha, A., 1988, Petrology and geochemistry of Archean basement lithologies in the Yankee Jim and Lamar Canyon areas, Montana and Wyoming: Precambrian and Mesozoic Plate Margins: Montana Bureau of Mines, Special Publication, v. 96, p. 53–68.
- Hague, A., 1899, The Early Tertiary Volcanoes of the Absaroka Range: *Science*, v. 9, p. 425–442.
- Halverson, G.P., Dudás, F.Ö., Maloof, A.C., and Bowring, S.A., 2007, Evolution of the $^{87}\text{Sr}/^{86}\text{Sr}$ composition of Neoproterozoic seawater: *Palaeogeography, Palaeoclimatology, Palaeoecology*, v. 256, p. 103–129, doi:10.1016/j.palaeo.2007.02.028.
- Hames, W., and Harms, T., 2013, Laser $^{40}\text{Ar}/^{39}\text{Ar}$ thermochronology of single crystals from Paleoproterozoic to Archean rocks of the northern Wyoming Province, Big Sky orogen, and Great Falls tectonic zone: Giletti's line revisited, in *Geological Society of America Abstracts with Programs*, v. 45, p. 223.
- Harlan, S.S., Geissman, J.W., Snee, L.W., and Reynolds, R.L., 1996, Late Cretaceous remagnetization of Proterozoic mafic dikes, southern Highland Mountains, southwestern Montana: A paleomagnetic and $^{40}\text{Ar}/^{39}\text{Ar}$ study: *GSA Bulletin*, v. 108, p. 653–668, doi:10.1130/0016-7606(1996)108<0653:LCROPM>2.3.CO;2.
- Harms, T.A. et al., 2004a, Character and origin of Precambrian fabrics and structures in the Tobacco Root Mountains, Montana, in *Special Paper 377: Precambrian Geology of the Tobacco Root Mountains, Montana*, Geological Society of America, v. 377, p. 203–226, doi:10.1130/0-8137-2377-9.203.
- Harms, T.A., Brady, J.B., Burger, H.R., and Cheney, J.T., 2004b, Advances in the geology of the Tobacco Root Mountains, Montana, and their implications for the history of the northern Wyoming province, in *Special Paper 377: Precambrian Geology of the Tobacco Root Mountains, Montana*, Geological Society of America, v. 377, p. 227–243, doi:10.1130/0-8137-2377-9.227.
- Harms, T.A., Brady, J.B., College, S., and Cheney, J.T., 2006, Exploring the Proterozoic Big Sky Orogeny in southwest Montana: 19th Annual Keck Symposium, p. 6.
- Harrison, J.E., Griggs, A.B., and Wells, J.D., 1974, Tectonic features of the Precambrian Belt Basin and their influence on post-Belt structures, doi:10.3133/pp866.
- Henry, C.D., Castor, S.B., Starkel, W.A., Ellis, B.S., Wolff, J.A., Laravie, J.A., McIntosh, W.C., and Heizler, M.T., 2017, Geology and evolution of the McDermitt caldera, northern Nevada and southeastern Oregon, western USA: *Geosphere*, v. 13, p. 1066–1112, doi:10.1130/GES01454.1.
- Henry, D., Mueller, P.A., Wooden, J.L., Warner, J., and Lee-Berman, R., 1982, Granulite grade supracrustal assemblages of the Quad Creek area, eastern Beartooth Mountains, Montana: Montana Bureau of Mines and Geology, Special Publication, v. 84, p. 147–155.
- Hoffman, P.F., 1988, United Plates of America, The Birth of a Craton: Early Proterozoic Assembly and Growth of Laurentia: *Annual Review of Earth and Planetary Sciences*, v. 16, p. 63.

- Horton, F., 2015, Did phosphorus derived from the weathering of large igneous provinces fertilize the Neoproterozoic ocean? *Geochemistry, Geophysics, Geosystems*, v. 16, p. 1723–1738, doi:<https://doi.org/10.1002/2015GC005792>.
- Hoy, R.G., and Ridgway, K.D., 1997, Structural and sedimentological development of footwall growth synclines along an intraforeland uplift, east-central Bighorn Mountains, Wyoming: *GSA Bulletin*, v. 109, p. 915–935, doi:10.1130/0016-7606(1997)109<0915:SASDOF>2.3.CO;2.
- Hurford, A.J., and Green, P.F., 1983, The zeta age calibration of fission-track dating: *Chemical Geology (Isotope Geoscience Section)*, v. 1, p. 285–317.
- Husson, J.M., and Peters, S.E., 2017, Atmospheric oxygenation driven by unsteady growth of the continental sedimentary reservoir: *Earth and Planetary Science Letters*, v. 460, p. 68–75, doi:10.1016/j.epsl.2016.12.012.
- Iddings, J.P., 1895, Absarokite-shoshonite-banakitite series: *The Journal of Geology*, v. 3, p. 935–959.
- James, H.L., and Hedge, C.E., 1980, Age of the basement rocks of southwest Montana: *GSA Bulletin*, v. 91, p. 11–15, doi:10.1130/0016-7606(1980)91<11:AOTBRO>2.0.CO;2.
- Jamieson, R.A., and Beaumont, C., 2011, Coeval thrusting and extension during lower crustal ductile flow – implications for exhumation of high-grade metamorphic rocks: *Journal of Metamorphic Geology*, v. 29, p. 33–51, doi:10.1111/j.1525-1314.2010.00908.x.
- Janecke, S.U., 2007, Cenozoic extensional processes and tectonics in the northern Rocky Mountains: Southwest Montana and eastern Idaho: *Northwest Geology*, v. 36, p. 111–132.
- Jensen, J.L., Siddoway, C.S., Reiners, P.W., Ault, A.K., Thomson, S.N., and Steele-MacInnis, M., 2018, Single-crystal hematite (U–Th)/He dates and fluid inclusions document widespread Cryogenian sand injection in crystalline basement: *Earth and Planetary Science Letters*, v. 500, p. 145–155, doi:10.1016/j.epsl.2018.08.021.
- Jobling, J., 1974, Stratigraphy, petrology, and structure of the Laramide (Paleocene) sediments marginal to the Beartooth Mountains: Montana [Ph.D. thesis] University Park, Pennsylvania, Pennsylvania State University.
- Johnson, J.E., Flowers, R.M., Baird, G.B., and Mahan, K.H., 2017, “Inverted” zircon and apatite (U–Th)/He dates from the Front Range, Colorado: High-damage zircon as a low-temperature (<50 °C) thermochronometer: *Earth and Planetary Science Letters*, v. 466, p. 80–90, doi:10.1016/j.epsl.2017.03.002.
- Karlstrom, K., Hagadorn, J., Gehrels, G., Matthews, W., Schmitz, M., Madronich, L., Mulder, J., Pecha, M., Giesler, D., and Crossey, L., 2018, Cambrian Sauk transgression in the Grand Canyon region redefined by detrital zircons: *Nature Geoscience*, v. 11, p. 438–443, doi:10.1038/s41561-018-0131-7.
- Kauffman, E.G., and Caldwell, W.G.E., 1992, The Western Interior Basin in space and time: *Geologic Association of Canada Special Paper* 39.

- Keller, C.B., Husson, J.M., Mitchell, R.N., Bottke, W.F., Gernon, T.M., Boehnke, P., Bell, E.A., Swanson-Hysell, N.L., and Peters, S.E., 2019, Neoproterozoic glacial origin of the Great Unconformity: *Proceedings of the National Academy of Sciences*, v. 116, p. 1136–1145, doi:10.1073/pnas.1804350116.
- Ketcham, R.A., 2005a, Forward and Inverse Modeling of Low-Temperature Thermochronometry Data: *Reviews in Mineralogy and Geochemistry*, v. 58, p. 275–314, doi:10.2138/rmg.2005.58.11.
- Ketcham, R., 2005b, HeFTy: Forward and inverse modeling thermochronometer systems: *Computational Tools for Low-Temperature Thermochronometer Interpretation*. Mineralogical Society of America, Chantilly, VA.
- Koenig, D. 1981-, 2015, Low-temperature Thermochronology of the Beartooth Conglomerate [Masters Thesis]: University of Houston, Houston, TX.
- Lachenbruch, A.H., 1979, Heat Flow in the Basin and Range Province and Thermal Effects of Tectonic Extension, in Rybach, L. and Stegena, L. eds., *Geothermics and Geothermal Energy*, Basel, Birkhäuser, Contributions to Current Research in Geophysics (CCRG), p. 34–50, doi:10.1007/978-3-0348-6525-8_3.
- Lau, K.V., Macdonald, F.A., Maher, K., and Payne, J.L., 2017, Uranium isotope evidence for temporary ocean oxygenation in the aftermath of the Sturtian Snowball Earth: *Earth and Planetary Science Letters*, v. 458, p. 282–292, doi:10.1016/j.epsl.2016.10.043.
- Li, Z.X. et al., 2008, Assembly, configuration, and break-up history of Rodinia: A synthesis: *Precambrian Research*, v. 160, p. 179–210, doi:10.1016/j.precamres.2007.04.021.
- Li, Z.-X., and Zhong, S., 2009, Supercontinent–superplume coupling, true polar wander and plume mobility: Plate dominance in whole-mantle tectonics: *Physics of the Earth and Planetary Interiors*, v. 176, p. 143–156, doi:10.1016/j.pepi.2009.05.004.
- Liu, S., Nummedal, D., and Gurnis, M., 2014, Dynamic versus flexural controls of Late Cretaceous Western Interior Basin, USA: *Earth and Planetary Science Letters*, v. 389, p. 221–229, doi:10.1016/j.epsl.2014.01.006.
- Long, S.P., 2018, Geometry and magnitude of extension in the Basin and Range Province (39°N), Utah, Nevada, and California, USA: Constraints from a province-scale cross section: *GSA Bulletin*, v. 131, p. 99–119, doi:10.1130/B31974.1.
- Lopez, D.A., 2001, Preliminary Geologic Map of the Red Lodge 30' x 60' Quadrangle South-Central Montana: Montana Bureau of Mines and Geology Open File Geologic Map 423.
- Macdonald, F.A., Schmitz, M.D., Crowley, J.L., Roots, C.F., Jones, D.S., Maloof, A.C., Strauss, J.V., Cohen, P.A., Johnston, D.T., and Schrag, D.P., 2010, Calibrating the Cryogenian: *Science*, v. 327, p. 1241–1243, doi:10.1126/science.1183325.
- Mallory, W., 1972, *Geologic Atlas of the Rocky Mountain region*: Rocky Mountain Association of Geologists: Denver, Colorado, v. 331.

- Malone, D., Craddock, J., and Kenderes, S., 2017, Detrital zircon geochronology and provenance of the Middle Cambrian Flathead Sandstone, Park County, Wyoming: *The Mountain Geologist*, v. 54, p. 86–103, doi:10.31582/rmag.mg.54.2.86.
- Matthews, W., Guest, B., and Madronich, L., 2018, Latest Neoproterozoic to Cambrian detrital zircon facies of western Laurentia: *Geosphere*, v. 14, p. 243–264, doi:10.1130/GES01544.1.
- Maughan, E.K., 1989, Geology and petroleum potential, central Montana province: U.S. Geologic Survey Open-File Report 88-450 N.
- Maughan, E.K., 1993, Stratigraphic and Structural Summary for Central Montana, in *Field Conference Guidebook: Energy and Mineral Resources of Central Montana*, Montana Geologic Society, p. 3–20.
- Maughan, E., 1990, Summary of the Ancestral Rocky Mountains Epeirogeny in Wyoming and adjacent areas: U.S. Geologic Survey Open-File Report 90-447.
- McDannell, K.T., and Flowers, R.M., 2020, Vestiges of the Ancient: Deep-Time Noble Gas Thermochronology: *Elements*, v. 16, p. 325–330, doi:10.2138/gselements.16.5.325.
- McDannell, K.T., Schneider, D.A., Zeitler, P.K., O’Sullivan, P.B., and Issler, D.R., 2019, Reconstructing deep-time histories from integrated thermochronology: An example from southern Baffin Island, Canada: *Terra Nova*, v. 31, p. 189–204, doi:10.1111/ter.12386.
- McDannell, K.T., Zeitler, P.K., and Schneider, D.A., 2018, Instability of the southern Canadian Shield during the late Proterozoic: *Earth and Planetary Science Letters*, v. 490, p. 100–109, doi:10.1016/j.epsl.2018.03.012.
- Mitchell, S.G., and Reiners, P.W., 2003, Influence of wildfires on apatite and zircon (U-Th)/He ages: *Geology*, v. 31, p. 1025–1028, doi:10.1130/G19758.1.
- Mogk, D., Henry, D., Mueller, P., and Foster, D., 2012, Origins of a continent: *Yellowstone Science*, v. 20, p. 22–32.
- Mogk, D.W., Mueller, P.A., and Wooden, J.L., 1988, Archean Tectonics of the North Snowy Block, Beartooth Mountains, Montana: *The Journal of Geology*, v. 96, p. 125–141, doi:10.1086/629205.
- Mogk, D.W., Mueller, P.A., and Wooden, J.L., 1992, The nature of Archean terrane boundaries: an example from the northern Wyoming Province: *Precambrian Research*, v. 55, p. 155–168, doi:10.1016/0301-9268(92)90020-O.
- Montgomery, C.W., and Lytwyn, J.N., 1984, Rb-Sr Systematics and Ages of Principal Precambrian Lithologies in the South Snowy Block, Beartooth Mountains: *The Journal of Geology*, v. 92, p. 103–112, doi:10.1086/628837.
- Mueller, P.A., Burger, H.R., Wooden, J.L., Brady, J.B., Cheney, J.T., Harms, T.A., Heatherington, A.L., and Mogk, D.W., 2005, Paleoproterozoic Metamorphism in the Northern Wyoming Province: Implications for the Assembly of Laurentia: *The Journal of Geology*, v. 113, p. 169–179, doi:10.1086/427667.

- Mueller, P.A., Burger, H.R., Wooden, J.L., Heatherington, A.L., Mogk, D.W., and D'Arcy, K., 2004, Age and evolution of the Precambrian crust of the Tobacco Root Mountains, Montana, in *Special Paper 377: Precambrian Geology of the Tobacco Root Mountains, Montana*, Geological Society of America, v. 377, p. 181–202, doi:10.1130/0-8137-2377-9.181.
- Mueller, P., and Cordua, W., 1976, Rb-Sr whole rock age of gneisses from the Horse Creek area: Tobacco Root Mountains, Montana: *Isochron/West*, v. 16, p. 33–36.
- Mueller, P.A., Heatherington, A.L., Kelly, D.M., Wooden, J.L., and Mogk, D.W., 2002, Paleoproterozoic crust within the Great Falls tectonic zone: Implications for the assembly of southern Laurentia: *Geology*, v. 30, p. 127–130, doi:10.1130/0091-7613(2002)030<0127:PCWTGF>2.0.CO;2.
- Mueller, P.A., Mogk, D.W., Henry, D.J., Wooden, J.L., and Foster, D.A., 2008, Geologic evolution of the Beartooth Mountains: Insights from petrology and geochemistry: *Northwestern Geology*, v. 37, p. 5–20.
- Mueller, P., Mogk, D., Wooden, J., and Spake, D., 2016, U-Pb ages of zircons from the Lower Belt Supergroup and proximal crystalline basement: Implications for the early evolution of the Belt Basin, in *Geological Society of America Special Papers*, Geological Society of America, v. 522, p. 283–303, doi:10.1130/2016.2522(11).
- Mueller, P.A., Shuster, R.D., Wooden, J.L., Erslev, E.A., and Bowes, D.R., 1993, Age and composition of Archean crystalline rocks from the southern Madison Range, Montana: Implications for crustal evolution in the Wyoming craton: *GSA Bulletin*, v. 105, p. 437–446, doi:10.1130/0016-7606(1993)105<0437:AACOAC>2.3.CO;2.
- Mueller, P.A., Wooden, J.L., Mogk, D.W., Henry, D.J., and Bowes, D.R., 2010, Rapid growth of an Archean continent by arc magmatism: *Precambrian Research*, v. 183, p. 70–88, doi:10.1016/j.precamres.2010.07.013.
- Murray, K.E., Orme, D.A., and Reiners, P.W., 2014, Effects of U–Th-rich grain boundary phases on apatite helium ages: *Chemical Geology*, v. 390, p. 135–151, doi:10.1016/j.chemgeo.2014.09.023.
- Nance, R.D., and Murphy, J.B., 2019, Supercontinents and the case for Pannotia: *Geological Society, London, Special Publications*, v. 470, p. 65–86, doi:10.1144/SP470.5.
- Nuccio, V.F., and Condon, S.M., 1996, Burial and thermal history of the Paradox Basin, Utah and Colorado, and petroleum potential of the middle Pennsylvanian Paradox Formation: *U.S. Geological Survey Bulletin* 2000-O.
- Nunes, P.D., and Tilton, G.R., 1971, Uranium-Lead Ages of Minerals from the Stillwater Igneous Complex and Associated Rocks, Montana: *GSA Bulletin*, v. 82, p. 2231–2250, doi:10.1130/0016-7606(1971)82[2231:UAOMFT]2.0.CO;2.
- Omar, G.I., Lutz, T.M., and Giegengack, R., 1994, Apatite fission-track evidence for Laramide and post-Laramide uplift and anomalous thermal regime at the Beartooth overthrust, Montana-Wyoming: *Geological Society of America Bulletin*, p. 12.

- O'Neill, J.M., Duncan, M.S., and Zartman, R.E., 1988, Proterozoic gneiss dome in the Highland Mountains, southwestern Montana, in Lewis, S.E. and Berg, R.B. eds., *Precambrian and Mesozoic plate margins: Montana, Idaho, and Wyoming*, with field guides for the 8th International conference on Basement Tectonics, Montana Bureau of Mine and Geology Special Publication 96, p. 81–88.
- O'Neill, J.M., Ferris, D.C., Schmidt, C.J., and Hanneman, D.L., 1986, Recurrent movement along northwest-trending faults at the southern margin of the Belt Basin, Highland Mountains, southwestern Montana: Belt Supergroup: Montana Bureau of Mines and Geology Special Publication, v. 94, p. 208–216.
- O'Neill, J.M., Klepper, M.R., Smedes, H.W., Hanneman, D.L., Fraser, G.D., and Mehnert, H.H., 1996, Geologic map and cross sections of the central and southern Highland Mountains, southwestern Montana: U.S. Geologic Survey Miscellaneous Field Studies Map I–2525.
- Orme, D.A., 2020, New Timing Constraints for the Onset of Laramide Deformation in Southwest Montana Challenge our Understanding of the Development of a Thick-Skinned Structural Style During Flat-Slab Subduction: *Tectonics*, v. 39, doi:10.1029/2020TC006193.
- Orme, D.A., Guenther, W.R., Laskowski, A.K., and Reiners, P.W., 2016, Long-term tectonothermal history of Laramide basement from zircon–He age–eU correlations: *Earth and Planetary Science Letters*, v. 453, p. 119–130, doi:10.1016/j.epsl.2016.07.046.
- Page, N.J., 1977, Stillwater Complex, Montana; rock succession, metamorphism and structure of the complex and adjacent rocks: US Govt. Print. Off.
- Page, N., Zientek, M., and Czamanske, G., 1985, Geologic and structural setting of the Stillwater Complex: The Stillwater Complex, Montana: *Geology and Guide*: Montana Bureau of Mines and Geology Special Publication, v. 92, p. 1–8.
- Parcell, W.C., and Williams, M.K., 2005, Mixed sediment deposition in a retro-arc foreland basin: Lower Ellis Group (M. Jurassic), Wyoming and Montana, U.S.A.: *Sedimentary Geology*, v. 177, p. 175–194, doi:10.1016/j.sedgeo.2005.02.007.
- Pardee, J.T., 1950, Late Cenozoic block faulting in western Montana: *Geological Society of America Bulletin*, v. 61, p. 359–406.
- Peale, A., 1896, Description of the Three Forks sheet, Montana: US Geological Survey Geol: Atlas, Folio, v. 24.
- Pearson, D.M., Parker, S.D., Montoya, L.M., and Brennan, D.T., 2019, No flat-slab needed: the transition from thin- to thick-skinned thrusting in the Idaho-Montana fold-thrust belt was controlled by the pre-orogenic stratigraphy, in p. 341061, doi:10.1130/abs/2019AM-341061.
- Peterman, Z.E., 1979, Geochronology and the Archean of the United States: *Economic Geology*, v. 74, p. 1544–1562.
- Peters, S.E., and Gaines, R.R., 2012, Formation of the 'Great Unconformity' as a trigger for the Cambrian explosion: *Nature*, v. 484, p. 363–366, doi:10.1038/nature10969.

- Peyton, S.L., and Carrapa, B., 2013, An Overview of Low-temperature Thermochronology in the Rocky Mountains and Its Application to Petroleum System Analysis: , p. 37–70, doi:10.1306/13381689St653578.
- Peyton, S.L., Reiners, P.W., Carrapa, B., and DeCelles, P.G., 2012, Low-temperature thermochronology of the northern Rocky Mountains, western U.S.A.: *American Journal of Science*, v. 312, p. 145–212, doi:10.2475/02.2012.04.
- Pierce, W.G., and Nelson, W.H., 1971, Geologic map of the Beartooth Butte quadrangle, Park County, Wyoming: *Geologic Quadrangle*, doi:10.3133/gq935.
- Planavsky, N.J., Rouxel, O.J., Bekker, A., Lalonde, S.V., Konhauser, K.O., Reinhard, C.T., and Lyons, T.W., 2010, The evolution of the marine phosphate reservoir: *Nature*, v. 467, p. 1088–1090, doi:10.1038/nature09485.
- Pogge von Strandmann, P.A.E., Stüeken, E.E., Elliott, T., Poulton, S.W., Dehler, C.M., Canfield, D.E., and Catling, D.C., 2015, Selenium isotope evidence for progressive oxidation of the Neoproterozoic biosphere: *Nature Communications*, v. 6, p. 10157, doi:10.1038/ncomms10157.
- Raedeke, L., 1983, Platinum group metals in the Stillwater Complex, in 1983. Symposium on genesis of Rocky Mountain ore deposits: Denver, Colorado, Exploration Geologists Society Denver Region Proceedings, p. 31–37.
- Reade, N.Z., Biddle, J.M., Ricketts, J.W., and Amato, J.M., 2020, Zircon (U-Th)/He Thermochronologic Constraints on the Long-Term Thermal Evolution of Southern New Mexico and Western Texas: *Lithosphere*, v. 2020, doi:10.2113/2020/8881315.
- Reid, R.R., McMannis, W.J., and Palmquist, J.C., 1975, Precambrian Geology of North Snowy Block, Beartooth Mountains, Montana: *Geological Society of America Special Papers*, v. 157, 152 p.
- Reiners, P.W., and Brandon, M.T., 2006, Using Thermochronology to Understand Orogenic Erosion: *Annual Review of Earth and Planetary Sciences*, v. 34, p. 419–466, doi:10.1146/annurev.earth.34.031405.125202.
- Reiners, P.W., Carlson, R.W., Renne, P.R., Cooper, K.M., Granger, D.E., McLean, N.M., and Schoene, B., 2017, *Geochronology and thermochronology*: John Wiley & Sons.
- Reiners, P.W., Farley, K.A., and Hickey, H.J., 2002, He diffusion and (U–Th)/He thermochronometry of zircon: initial results from Fish Canyon Tuff and Gold Butte: *Tectonophysics*, v. 349, p. 297–308, doi:10.1016/S0040-1951(02)00058-6.
- Reiners, P.W., Spell, T.L., Nicolescu, S., and Zanetti, K.A., 2004, Zircon (U-Th)/He thermochronometry: He diffusion and comparisons with $^{40}\text{Ar}/^{39}\text{Ar}$ dating: *Geochimica et Cosmochimica Acta*, v. 68, p. 1857–1887, doi:10.1016/j.gca.2003.10.021.
- Reiners, P.W., Thomson, S.N., McPhillips, D., Donelick, R.A., and Roering, J.J., 2007, Wildfire thermochronology and the fate and transport of apatite in hillslope and fluvial environments: *Journal of Geophysical Research: Earth Surface*, v. 112, doi:10.1029/2007JF000759.

- Retallack, G.J., 2013, Early Cambrian Humid, Tropical, Coastal Paleosols from Montana, USA, in Driese, S.G. ed., *New Frontiers in Paleopedology and Terrestrial Paleoclimatology: Paleosols and Soil Surface Analog Systems*, SEPM (Society for Sedimentary Geology), p. 257–272, doi:10.2110/sepmsp.104.09.
- Roberts, H., Dahl, P., Kelley, S., and Frei, R., 2002, New ^{207}Pb – ^{206}Pb and ^{40}Ar – ^{39}Ar ages from SW Montana, USA: constraints on the Proterozoic and Archæan tectonic and depositional history of the Wyoming Province: *Precambrian Research*, p. 25.
- Ronemus, C.B., Orme, D.A., Campbell, S., Black, S.R., and Cook, J., 2020, Mesoproterozoic–Early Cretaceous provenance and paleogeographic evolution of the Northern Rocky Mountains: Insights from the detrital zircon record of the Bridger Range, Montana, USA: *GSA Bulletin*, doi:10.1130/B35628.1.
- Rudnick, R., Ireland, T., Gehrels, G., Irving, A., Chesley, J., and Hanchar, J., 1998, Dating mantle metasomatism: U-Pb geochronology of zircons in cratonic mantle xenoliths from Montana and Tanzania, in *International Kimberlite Conference: Extended Abstracts*, v. 7, p. 754–756.
- Sahoo, S.K., Planavsky, N.J., Kendall, B., Wang, X., Shi, X., Scott, C., Anbar, A.D., Lyons, T.W., and Jiang, G., 2012, Ocean oxygenation in the wake of the Marinoan glaciation: *Nature*, v. 489, p. 546–549, doi:10.1038/nature11445.
- Sclater, J.G., and Christie, P. a. F., 1980, Continental stretching: An explanation of the Post-Mid-Cretaceous subsidence of the central North Sea Basin: *Journal of Geophysical Research: Solid Earth*, v. 85, p. 3711–3739, doi:10.1029/JB085iB07p03711.
- Sears, J.W., and Price, R.A., 1978, The Siberian Connection: A case for Precambrian separation of the North American and Siberian cratons: *Geology*, v. 6, p. 267–270, doi:10.1130/0091-7613(1978)6<267:TSCACF>2.0.CO;2.
- Sharry, J., Langan, R.T., Jovanovich, D.B., Jones, G.M., Hill, N.R., and Guidish, T.M., 1986, Enhanced Imaging of the COCORP Seismic Line, Wind River Mountains, in *Reflection Seismology: A Global Perspective*, American Geophysical Union (AGU), p. 223–236, doi:10.1029/GD013p0223.
- Siddoway, C., Palladino, G., Prosser, G., Freedman, D., and Duckworth, W.C., 2019, Basement-hosted sand injectites: Use of field examples to advance understanding of hydrocarbon reservoirs in fractured crystalline basement rocks: *Geological Society, London, Special Publications*, v. 493.
- Simons, F.S., and Armbrustmacher, T.J., 1976, High-level plateaus of the southeastern Beartooth Mountains, Montana and Wyoming—remnants of an exhumed sub-Cambrian marine plain: *Jour. Res. US Geol. Surv*, v. 4, p. 387–396.
- Sloss, L.L., 1963, Sequences in the Cratonic Interior of North America: *Geological Society of America Bulletin*, v. 74, p. 93, doi:10.1130/0016-7606(1963)74[93:SITCIO]2.0.CO;2.
- Smithson, S.B., Brewer, J., Kaufman, S., Oliver, J., and Hurich, C., 1978, Nature of the Wind River thrust, Wyoming, from COCORP deep-reflection data and from gravity data: *Geology*, v. 6, p. 648–652, doi:10.1130/0091-7613(1978)6<648:NOTWRT>2.0.CO;2.

- Sobolev, S.V., and Brown, M., 2019, Surface erosion events controlled the evolution of plate tectonics on Earth: *Nature*, v. 570, p. 52–57, doi:10.1038/s41586-019-1258-4.
- Sousa, F.J., and Farley, K.A., 2020, A Framework for Evaluating Variation in (U-Th)/He Datasets: *Minerals*, v. 10, p. 1111, doi:10.3390/min10121111.
- Squire, R.J., Campbell, I.H., Allen, C.M., and Wilson, C.J.L., 2006, Did the Transgondwanan Supermountain trigger the explosive radiation of animals on Earth? *Earth and Planetary Science Letters*, v. 250, p. 116–133, doi:10.1016/j.epsl.2006.07.032.
- Symons, D.T.A., and Timmins, E.A., 1992, Geotectonics of the cratonic margin from paleomagnetism of the Middle Proterozoic Aldridge (Prichard) Formation and Moyie sills of British Columbia and Montana, *in* Bartholomew, M.J., Hyndman, D.W., Mogk, D.W., and Mason, R. eds., *Basement Tectonics 8: Characterization and Comparison of Ancient and Mesozoic Continental Margins*, Dordrecht, Springer Netherlands, *Proceedings of the International Conferences on Basement Tectonics*, p. 373–384, doi:10.1007/978-94-011-1614-5_25.
- Thacker, J.O., Kelley, S.A., and Karlstrom, K.E. Late Cretaceous–Recent low-temperature cooling history and tectonic analysis of the Zuni Mountains, west-central New Mexico: *Tectonics*, v. n/a, p. e2020TC006643, doi:10.1029/2020TC006643.
- Thacker, J.O., Lageson, D.R., and Mogk, D.W., 2017, Subsurface structural and mineralogical characterization of the Laramide South Prairie fault in the Stillwater Complex, Beartooth Mountains, Montana: *Lithosphere*, v. 9, p. 100–116, doi:10.1130/L562.1.
- Thomas, R., 2007, A field guide to the Cambrian section at camp creek, southwest Montana: *Northwest Geology*, v. 36, p. 231–244.
- Thompson, A.B., 1984, Geothermal Gradients Through Time, in Holland, H.D. and Trendall, A.F. eds., *Patterns of Change in Earth Evolution*, Berlin, Heidelberg, Springer, Dahlem Workshop Reports Physical, Chemical, and Earth Sciences Research Reports, p. 345–355, doi:10.1007/978-3-642-69317-5_20.
- Walker, J., Geissman, J., Bowring, S., and Babcock, L., 2018, GSA geological time scale v. 5.0: Geological Society of America.
- Wall, C.J., Scoates, J.S., Weis, D., Friedman, R.M., Amini, M., and Meurer, W.P., 2018, The Stillwater Complex: integrating zircon geochronological and geochemical constraints on the age, emplacement history and crystallization of a large, open-system layered intrusion: *Journal of Petrology*, v. 59, p. 153–190.
- Waples, D.W., Pacheco, J., and Vera, A., 2004, A method for correcting log-derived temperatures in deep wells, calibrated in the Gulf of Mexico: *Petroleum Geoscience*, v. 10, p. 239–245, doi:10.1144/1354-079302-542.
- Webber, J.R., Kalakay, T.J., Peterman, E.M., McAleer, R., and Mogk, D.W., 2019, Deciphering the tectonometamorphic evolution and localization of deformation within the Wyoming Province of southwest Montana, in AGU Fall Meeting Abstracts, v. 2019, p. T13E-0237.

- Wegmann, K.W. et al., 2007, Position of the Snake River watershed divide as an indicator of geodynamic processes in the greater Yellowstone region, western North America: *Geosphere*, v. 3, p. 272–281, doi:10.1130/GES00083.1.
- Weyand, E.L., 1989, U-Pb zircon geochronology of Archean rocks from the Spanish Peaks area, Madison Range, Montana [PhD Thesis]: University of Florida, Gainesville, FL.
- Whitlock, C., Reasoner, M.A., and Key, C.H., 2002, Paleoenvironmental history of the Rocky Mountain region during the past 20,000 years: *Rocky Mountain futures: an ecological perspective*. Island Press, Washington, DC, p. 41–57.
- Whitmeyer, S.J., and Karlstrom, K.E., 2007, Tectonic model for the Proterozoic growth of North America: *Geosphere*, p. 40.
- Whitney, D.L., and Evans, B.W., 2010, Abbreviations for names of rock-forming minerals: *American Mineralogist*, v. 95, p. 185–187, doi:10.2138/am.2010.3371.
- Will, C., 2013, Temperature and pressure conditions of Archean amphibolite-granulite facies metamorphic xenoliths from the eastern Beartooth Mountains, Montana and Wyoming, USA [Master of Science]: Louisiana State University, Baton Rouge, LA.
- Wilson, C.W., Jr., 1936, Geology of Nye-Bowler Lineament, Stillwater and Carbon Counties, Montana: *AAPG Bulletin*, v. 20, p. 1161–1188, doi:10.1306/3D932E20-16B1-11D7-8645000102C1865D.
- Winston, D., 1986, Sedimentation and Tectonics of the Middle Proterozoic Belt Basin and Their Influence on Phanerozoic Compression and Extension in Western Montana and Northern Idaho: Part II. Northern Rocky Mountains: v. 155, p. 87–118.
- Wise, D.U., 2000, Laramide Structures in Basement and Cover of the Beartooth Uplift Near Red Lodge, Montana: *AAPG Bulletin*, v. 84, p. 360–375, doi:10.1306/C9EBCDF1-1735-11D7-8645000102C1865D.
- Wise, D.U., 1997, Pseudo Tear Faults at Red Lodge, MT: Clues to Late-stage Basin Understuffing and Tectonics of the Greater Beartooth Block, Bighorn Basin Symposium Guidebook, p. 71–99.
- Wooden, J.L., Mueller, P.A., Mogk, D.A., Bowes, D.R., Lewis, S., and Berg, R., 1988, A review of the geochemistry and geochronology of Archean rocks of the Beartooth Mountains, Montana and Wyoming: Precambrian and Mesozoic plate margins: Montana, Idaho and Wyoming. Edited by SE Lewis and RB Berg. Montana Bureau of Mines and Geology, Special Publication, v. 96, p. 23–42.
- Yamada, R., Murakami, M., and Tagami, T., 2007, Statistical modelling of annealing kinetics of fission tracks in zircon; Reassessment of laboratory experiments: *Chemical Geology*, v. 236, p. 75–91, doi:10.1016/j.chemgeo.2006.09.002.
- Yonkee, W.A., and Weil, A.B., 2015, Tectonic evolution of the Sevier and Laramide belts within the North American Cordillera orogenic system: *Earth-Science Reviews*, v. 150, p. 531–593, doi:10.1016/j.earscirev.2015.08.001.

Zhong, S., Zhang, N., Li, Z.-X., and Roberts, J.H., 2007, Supercontinent cycles, true polar wander, and very long-wavelength mantle convection: *Earth and Planetary Science Letters*, v. 261, p. 551–564, doi:10.1016/j.epsl.2007.07.049.

CHAPTER THREE

CONCLUSIONS

The manuscript presented in this thesis (chapter 2) reports a multi-chronometer dataset from samples of Archean basement rock collected across large vertical and horizontal distances in the Beartooth Mountains. Integration of thermal modeling results with those of decompacted sediment accumulation models and previously reported low-temperatures thermochronology (Omar et al., 1994; Peyton et al., 2012; Carrapa et al., 2019 and references therein) places new constraints on the Paleoproterozoic–recent thermal history experienced by Archean rocks of the Beartooth Mountains. Collectively, results of this study help constrain an integrated long-term thermal history for this region of southwestern Montana. An abbreviated summary of this Archean–recent geologic history, as informed by new constraints from this work in the Beartooth Mountains, is as follows:

(1) Archean—Cratonization of rocks with the Wyoming province and subsequent intracratonic reorganization and metamorphism to granulite facies led to the juxtaposition of two lithologically distinct Archean terranes comprising the cratonic lithosphere of southwestern Montana, one dominated by orthogneiss and one dominated by metasedimentary rock (e.g., Mogk et al., 1992; Mueller and Frost, 2006; Mueller et al., 2008). Rocks of the former terrane comprise the core of the Beartooth Mountains.

(2) Paleoproterozoic—Thermotectonism and/or mantle metasomatism or mafic underplating associated with the Paleoproterozoic amalgamation of Laurentia (e.g., Harms et al., 2004; Mueller et al., 2005; Gifford et al., 2014; Condit et al., 2015) resulted in heating, across

southwest Montana as far east as the central Beartooth Mountains, as recorded by biotite $^{40}\text{Ar}/^{39}\text{Ar}$ results.

(3) Early Mesoproterozoic—The gravitational collapse of the Big Sky orogen caused widespread cooling—as predicted by our thermal model results from the Beartooth Mountains. Mesoproterozoic extension also accommodated regional deposition of the Belt Supergroup, though the Beartooth Mountains were not a significant source of detritus to this basin (e.g., Guerrero et al., 2016; Mueller et al., 2016; Fox, 2017; Ronemus et al., 2020) and our samples remained at middle-upper crustal temperatures during this time.

(4) Late Mesoproterozoic–Neoproterozoic—Widespread continental erosion resulted in the development of the Great Unconformity surface. This erosion occurred dominantly between ca. 1.4–0.8 Ga, as suggested by our thermal modeling results, and was likely diachronous across the Laurentian continent (e.g., Orme et al., 2016; DeLucia et al., 2018; McDannell et al., 2018; Flowers et al., 2020). The Flathead Sandstone was deposited atop this erosional surface, preserved in regions on the periphery of the Beartooth Mountains.

(5) Paleozoic—Widespread deposition occurred in southwest Montana, likely reaching maximum burial temperatures at the location of the Beartooth Mountains by Late Pennsylvanian time, as suggested by our thermal model results. Here, relatively thick Pennsylvanian burial and subsequent Late Paleozoic uplift out of sync with nearby regions was perhaps associated with the reactivation of inherited crustal weaknesses in the Wyoming craton by far-field stresses of the Ancestral Rocky Mountains orogenic event (e.g., Maughan, 1990).

(6) Mesozoic—Development of the North American Cordillera resulted in the formation of a broad orogenic system, with the Beartooth Mountains located in its foreland basin region

(e.g., DeCelles, 2004; Yonkee and Weil, 2015). This range was likely uplifted with respect to subsiding nearby regions, with continued cooling throughout much of this time suggested by our thermal modeling results. Between Late Cretaceous and Thanetian time, the Beartooth Mountains experienced rapid uplift, as recorded by AFT, ZHe and AHe thermochronometers (e.g., Omar et al., 1994; Peyton et al., 2012; Carrapa et al., 2019).

(7) Cenozoic—The Beartooth Mountains likely experienced reheating associated with burial by Eocene–Neogene volcanics (e.g., Giegengack et al., 1986; Omar et al., 1994). Finally, the Beartooth Mountains experienced a final pulse of cooling, as suggested by thermal model results, likely owing to impingement of the Yellowstone hot spot (e.g., Wegmann et al., 2007) and/or increased erosion related to regional Basin and Range extension (e.g., Janecke, 2007).

The events described here underscore the complex and varied geologic history recorded in this region of western Laurentia. The intimate linkage between tectonism of each time period and the structure, geometry, and perhaps timing of subsequent geologic events cannot be understated. An apt analogy may be found in the zircon crystal, in which continually accumulating damage to the lattice influences helium retentivity during subsequent thermal events. Collectively, the thermal history interpreted herein underscores the notion that, in regions near cratonic margins with ancient records of lithospheric deformation, geologic events ought not be considered in isolation. To bastardize the great naturalist writer John McPhee (1981): “If by some fiat I had to restrict all this writing to one sentence, this is the one I would choose: The summit of [Granite Peak] is [Archean Gneiss].”

Literature Cited

- Carrapa, B., DeCelles, P.G., and Romero, M., 2019, Early Inception of the Laramide Orogeny in Southwestern Montana and Northern Wyoming: Implications for Models of Flat-Slab Subduction: *Journal of Geophysical Research: Solid Earth*, v. 124, p. 2102–2123, doi:10.1029/2018JB016888.
- Condit, C.B., Mahan, K.H., Ault, A.K., and Flowers, R.M., 2015, Foreland-directed propagation of high-grade tectonism in the deep roots of a Paleoproterozoic collisional orogen, SW Montana, USA: *Lithosphere*, p. L460.1, doi:10.1130/L460.1.
- DeCelles, P.G., 2004, Late Jurassic to Eocene evolution of the Cordilleran thrust belt and foreland basin system, western U.S.A.: *American Journal of Science*, v. 304, p. 105–168, doi:10.2475/ajs.304.2.105.
- DeLucia, M.S., Guenther, W.R., Marshak, S., Thomson, S.N., and Ault, A.K., 2018, Thermochronology links denudation of the Great Unconformity surface to the supercontinent cycle and snowball Earth: *Geology*, v. 46, p. 167–170, doi:10.1130/G39525.1.
- Flowers, R.M., Macdonald, F.A., Siddoway, C.S., and Havranek, R., 2020, Diachronous development of Great Unconformities before Neoproterozoic Snowball Earth: *Proceedings of the National Academy of Sciences*, p. 201913131, doi:10.1073/pnas.1913131117.
- Fox, N.R., 2017, Sedimentary tectonics of the Mesoproterozoic LaHood Formation, southwest Montana: M.Sc. Thesis, v. Montana State University, p. 164.
- Giegengack, R., Omar, G.I., and Johnson, K.R., 1986, A reconnaissance fission track uplift chronology for the northwest margin of the Bighorn Basin: Montana Geological Society and yellowstone bighorn Research Association Joint Field Conference and Symposium: *Geology of the Beartooth Uplift and Adjacent Basins*, p. 179–184.
- Gifford, J.N., Mueller, P.A., Foster, D.A., and Mogk, D.W., 2014, Precambrian Crustal Evolution in the Great Falls Tectonic Zone: Insights from Xenoliths from the Montana Alkali Province: *The Journal of Geology*, v. 122, p. 531–548, doi:10.1086/677262.
- Guerrero, J.C., Mueller, P.A., and Mogk, D.W., 2016, Provenance study and geochemical analysis of the LaHood Formation of the Bridger Range, Montana: *Northwest Geology*, v. 45, p. 21–28.
- Harms, T.A. et al., 2004, Character and origin of Precambrian fabrics and structures in the Tobacco Root Mountains, Montana, *in* Special Paper 377: Precambrian Geology of the Tobacco Root Mountains, Montana, Geological Society of America, v. 377, p. 203–226, doi:10.1130/0-8137-2377-9.203.

- Janecke, S.U., 2007, Cenozoic extensional processes and tectonics in the northern Rocky Mountains: Southwest Montana and eastern Idaho: *Northwest Geology*, v. 36, p. 111–132.
- Maughan, E., 1990, Summary of the Ancestral Rocky Mountains Epeirogeny in Wyoming and adjacent areas: U.S. Geologic Survey Open-File Report 90-447.
- McDannell, K.T., Zeitler, P.K., and Schneider, D.A., 2018, Instability of the southern Canadian Shield during the late Proterozoic: *Earth and Planetary Science Letters*, v. 490, p. 100–109, doi:10.1016/j.epsl.2018.03.012.
- McPhee, J., 1981, *Basin and range*: Macmillan.
- Mogk, D.W., 1992, Ductile shearing and migmatization at mid-crustal levels in an Archaean high-grade gneiss belt, northern Gallatin Range, Montana, USA: *Journal of Metamorphic Geology*, v. 10, p. 427–438, doi:10.1111/j.1525-1314.1992.tb00094.x.
- Mueller, P.A., Burger, H.R., Wooden, J.L., Brady, J.B., Cheney, J.T., Harms, T.A., Heatherington, A.L., and Mogk, D.W., 2005, Paleoproterozoic Metamorphism in the Northern Wyoming Province: Implications for the Assembly of Laurentia: *The Journal of Geology*, v. 113, p. 169–179, doi:10.1086/427667.
- Mueller, P.A., and Frost, C.D., 2006, The Wyoming Province: a distinctive Archean craton in Laurentian North America: *Canadian Journal of Earth Sciences*, v. 43, p. 1391–1397, doi:10.1139/e06-075.
- Mueller, P.A., Mogk, D.W., Henry, D.J., Wooden, J.L., and Foster, D.A., 2008, Geologic evolution of the Beartooth Mountains: Insights from petrology and geochemistry: *Northwestern Geology*, v. 37, p. 5–20.
- Mueller, P., Mogk, D., Wooden, J., and Spake, D., 2016, U-Pb ages of zircons from the Lower Belt Supergroup and proximal crystalline basement: Implications for the early evolution of the Belt Basin, *in* Geological Society of America Special Papers, Geological Society of America, v. 522, p. 283–303, doi:10.1130/2016.2522(11).
- Omar, G.I., Lutz, T.M., and Giegengack, R., 1994, Apatite fission-track evidence for Laramide and post-Laramide uplift and anomalous thermal regime at the Beartooth overthrust, Montana-Wyoming: *Geological Society of America Bulletin*, p. 12.
- Orme, D.A., Guenther, W.R., Laskowski, A.K., and Reiners, P.W., 2016, Long-term tectonothermal history of Laramide basement from zircon–He age–eU correlations: *Earth and Planetary Science Letters*, v. 453, p. 119–130, doi:10.1016/j.epsl.2016.07.046.
- Peyton, S.L., Reiners, P.W., Carrapa, B., and DeCelles, P.G., 2012, Low-temperature thermochronology of the northern Rocky Mountains, western U.S.A.: *American Journal of Science*, v. 312, p. 145–212, doi:10.2475/02.2012.04.

- Ronemus, C.B., Orme, D.A., Campbell, S., Black, S.R., and Cook, J., 2020, Mesoproterozoic–Early Cretaceous provenance and paleogeographic evolution of the Northern Rocky Mountains: Insights from the detrital zircon record of the Bridger Range, Montana, USA: *GSA Bulletin*, doi:10.1130/B35628.1.
- Wegmann, K.W. et al., 2007, Position of the Snake River watershed divide as an indicator of geodynamic processes in the greater Yellowstone region, western North America: *Geosphere*, v. 3, p. 272–281, doi:10.1130/GES00083.1.
- Yonkee, W.A., and Weil, A.B., 2015, Tectonic evolution of the Sevier and Laramide belts within the North American Cordillera orogenic system: *Earth-Science Reviews*, v. 150, p. 531–593, doi:10.1016/j.earscirev.2015.08.001.

CUMULATIVE LITERATURE CITED

- Alcock, J., Muller, P.D., and Jercinovic, M.J., 2013, Monazite ages and pressure–temperature–time paths from anatectites in the southern Ruby Range, Montana, USA: evidence for delamination, ultramafic magmatism, and rapid uplift at ca. 1780 Ma: *Canadian Journal of Earth Sciences*, doi:10.1139/cjes-2013-0035.
- Allen, P., and Allen, J., 1990, *Basin analysis: principles and applications*, Blackwell Scientific Publications: Oxford, London.
- Anders, M.H., and Sleep, N.H., 1992, Magmatism and extension: The thermal and mechanical effects of the Yellowstone Hotspot: *Journal of Geophysical Research: Solid Earth*, v. 97, p. 15379–15393, doi:<https://doi.org/10.1029/92JB01376>.
- Anderson, C.J., 2017, Stratigraphic framework and provenance of the lower belt (Newland Formation), Belt Supergroup, Helena Embayment, Central Montana [Master of Science]: Montana State University - Bozeman, College of Letters & Science.
- Anderson, H.E., and Davis, D.W., 1995, U–Pb geochronology of the Moyie sills, Purcell Supergroup, southeastern British Columbia: implications for the Mesoproterozoic geological history of the Purcell (Belt) basin: *Canadian Journal of Earth Sciences*, v. 32, p. 1180–1193, doi:10.1139/e95-097.
- Ault, A.K., Flowers, R.M., and Mahan, K.H., 2012, Quartz shielding of sub-10 μ m zircons from radiation damage-enhanced Pb loss: An example from a metamorphosed mafic dike, northwestern Wyoming craton: *Earth and Planetary Science Letters*, v. 339–340, p. 57–66, doi:10.1016/j.epsl.2012.04.025.
- Ayers, W.B., Jr., 1986, Lacustrine and Fluvial-Deltaic Depositional Systems, Fort Union Formation (Paleocene), Powder River Basin, Wyoming and Montana: *AAPG Bulletin*, v. 70, p. 1651–1673, doi:10.1306/94886C90-1704-11D7-8645000102C1865D.
- Baadsgaard, H., and Mueller, P.A., 1973, K-Ar and Rb-Sr Ages of Intrusive Precambrian Mafic Rocks, Southern Beartooth Mountains, Montana and Wyoming: *GSA Bulletin*, v. 84, p. 3635–3644, doi:10.1130/0016-7606(1973)84<3635:KARAOI>2.0.CO;2.
- Barker, C., 1996, Thermal modeling of petroleum generation. Theory and applications, <https://www.osti.gov/etdeweb/biblio/557106> (accessed March 2021).
- Bell, L.H., and Middleton, L.T., 1978, An Introduction to the Cambrian Flathead Sandstone, Wind River Basin, Wyoming, in *Resources of the Wind River Basin*, 30th Annual Field Conference Guidebook, Wyoming Geological Association, p. 79–88.
- Beranek, L.P., Link, P.K., and Fanning, C.M., 2016, Detrital zircon record of mid-Paleozoic convergent margin activity in the northern U.S. Rocky Mountains: Implications for the Antler orogeny and early evolution of the North American Cordillera: *Lithosphere*, v. 8, p. 533–550, doi:10.1130/L557.1.
- Berg, R.B., Lopez, D.A., and Lonn, J.D., 2000, Geologic map of the Livingston 30' x 60' quadrangle, south-central Montana: Montana Bureau of Mines and Geology, Open File Report MBMG, v. 406, p. 21.

- Bevan, A., 1923, Summary of the geology of the Beartooth Mountains, Montana: *The Journal of Geology*, v. 31, p. 441–465.
- Bird, P., 1984, Laramide crustal thickening event in the Rocky Mountain Foreland and Great Plains: *Tectonics*, v. 3, p. 741–758, doi:<https://doi.org/10.1029/TC003i007p00741>.
- Boerner, D.E., Craven, J.A., Kurtz, R.D., Ross, G.M., and Jones, F.W., 1998, The Great Falls Tectonic Zone: suture or intracontinental shear zone? *Canadian Journal of Earth Sciences*, v. 35, p. 175–183, doi:10.1139/e97-104.
- Bonnichsen, B., Leeman, W.P., Honjo, N., McIntosh, W.C., and Godchaux, M.M., 2008, Miocene silicic volcanism in southwestern Idaho: geochronology, geochemistry, and evolution of the central Snake River Plain: *Bulletin of Volcanology*, v. 70, p. 315–342, doi:10.1007/s00445-007-0141-6.
- Brady, J., 2004, Precambrian Geology of the Tobacco Root Mountains, Montana: Geological Society of America, 270 p.
- Brasier, M.D., 1980, The Lower Cambrian transgression and glauconite-phosphate facies in western Europe: *Journal of the Geological Society*, v. 137, p. 695–703, doi:10.1144/gsjgs.137.6.0695.
- Braun, J., 2002, Quantifying the effect of recent relief changes on age–elevation relationships: *Earth and Planetary Science Letters*, v. 200, p. 331–343, doi:10.1016/S0012-821X(02)00638-6.
- Bricker, A.L., 2016, (U-Th)/He thermochronology of the Yellowstone Crescent High Terrain: Implications for Miocene to recent topographic impacts associated with the Yellowstone Hotspot [Masters Thesis]: University of Florida.
- Brocks, J.J., Jarrett, A.J.M., Sirantoine, E., Hallmann, C., Hoshino, Y., and Liyanage, T., 2017, The rise of algae in Cryogenian oceans and the emergence of animals: *Nature*, v. 548, p. 578–581, doi:10.1038/nature23457.
- Brown, R.W., Beucher, R., Roper, S., Persano, C., Stuart, F., and Fitzgerald, P., 2013, Natural age dispersion arising from the analysis of broken crystals. Part I: Theoretical basis and implications for the apatite (U–Th)/He thermochronometer: *Geochimica et Cosmochimica Acta*, v. 122, p. 478–497, doi:10.1016/j.gca.2013.05.041.
- Burger, H.R., 2004, General geology and tectonic setting of the Tobacco Root Mountains, in Special Paper 377: Precambrian Geology of the Tobacco Root Mountains, Montana, Geological Society of America, v. 377, p. 1–14, doi:10.1130/0-8137-2377-9.1.
- Carlson, R., and Irving, A., 1994, Depletion and enrichment history of subcontinental lithospheric mantle: an Os, Sr, Nd and Pb isotopic study of ultramafic xenoliths from the northwestern Wyoming Craton: *Earth and Planetary Science Letters*, v. 126, p. 457–472.
- Carrapa, B., DeCelles, P.G., and Romero, M., 2019, Early Inception of the Laramide Orogeny in Southwestern Montana and Northern Wyoming: Implications for Models of Flat-Slab Subduction: *Journal of Geophysical Research: Solid Earth*, v. 124, p. 2102–2123, doi:10.1029/2018JB016888.

- Carrapa, B., Wijbrans, J., and Bertotti, G., 2003, Episodic exhumation in the Western Alps: *Geology*, v. 31, p. 601–604, doi:10.1130/0091-7613(2003)031<0601:EEITWA>2.0.CO;2.
- Cervený, P.F., 1990, Fission-track thermochronology of the Wind River Range and other basement cored uplifts in the Rocky Mountain foreland [Ph.D. thesis]: University of Wyoming, 180 p.
- Cervený, P.F., and Steidtmann, J.R., 1993, Fission track thermochronology of the Wind River Range, Wyoming: Evidence for timing and magnitude of Laramide exhumation: *Tectonics*, v. 12, p. 77–91, doi:10.1029/92TC01567.
- Chamberlain, K.R., Frost, C.D., and Frost, B.R., 2003, Early Archean to Mesoproterozoic evolution of the Wyoming Province: Archean origins to modern lithospheric architecture: *Canadian Journal of Earth Sciences*, v. 40, p. 1357–1374, doi:10.1139/e03-054.
- Chaudhuri, S., and Brookins, D.G., 1969, The isotopic age of the Flathead Sandstone (Middle Cambrian), Montana: *Journal of Sedimentary Research*, v. 39, p. 364–366, doi:10.1306/74D71C76-2B21-11D7-8648000102C1865D.
- Cheney, J.T. et al., 2004, Proterozoic metamorphism of the Tobacco Root Mountains, Montana, in *Special Paper 377: Precambrian Geology of the Tobacco Root Mountains, Montana*, Geological Society of America, v. 377, p. 105–129, doi:10.1130/0-8137-2377-9.105.
- Colpron, M., Nelson, J.L., and Murphy, D.C., 2007, Northern Cordilleran terranes and their interactions through time: *GSA Today*, v. 17, p. 4, doi:10.1130/GSAT01704-5A.1.
- Condie, K.C., 1976, Trace-element geochemistry of archaean greenstone belts: *Earth-Science Reviews*, v. 12, p. 393–417, doi:10.1016/0012-8252(76)90012-X.
- Condie, K.C., Leech, A.P., and Baadgaard, H., 1969, Potassium-Argon Ages of Precambrian Mafic Dikes in Wyoming: *Geological Society of America Bulletin*, v. 80, p. 899, doi:10.1130/0016-7606(1969)80[899:PAOPMD]2.0.CO;2.
- Condit, C.B., Mahan, K.H., Ault, A.K., and Flowers, R.M., 2015, Foreland-directed propagation of high-grade tectonism in the deep roots of a Paleoproterozoic collisional orogen, SW Montana, USA: *Lithosphere*, p. L460.1, doi:10.1130/L460.1.
- Cox, G.M., Halverson, G.P., Stevenson, R.K., Vokaty, M., Poirier, A., Kunzmann, M., Li, Z.-X., Denyszyn, S.W., Strauss, J.V., and Macdonald, F.A., 2016, Continental flood basalt weathering as a trigger for Neoproterozoic Snowball Earth: *Earth and Planetary Science Letters*, v. 446, p. 89–99, doi:10.1016/j.epsl.2016.04.016.
- Dahl, P., Frei, R., and Dorais, M., 1998, When did the Wyoming Province collide with Laurentia? New clues from step-leach Pb-Pb dating of garnet independent of its inclusions, in *Geological Society of America Abstracts with Programs*, v. 30, p. 109.
- Danišík, M., McInnes, B.I.A., Kirkland, C.L., McDonald, B.J., Evans, N.J., and Becker, T., 2017, Seeing is believing: Visualization of He distribution in zircon and implications for thermal history reconstruction on single crystals: *Science Advances*, v. 3, p. e1601121, doi:10.1126/sciadv.1601121.

- DeCelles, P.G., 2004, Late Jurassic to Eocene evolution of the Cordilleran thrust belt and foreland basin system, western U.S.A.: *American Journal of Science*, v. 304, p. 105–168, doi:10.2475/ajs.304.2.105.
- DeCelles, P.G., 1986, Sedimentation in a tectonically partitioned, nonmarine foreland basin: The Lower Cretaceous Kootenai Formation, southwestern Montana: *GSA Bulletin*, v. 97, p. 911–931, doi:10.1130/0016-7606(1986)97<911:SIATPN>2.0.CO;2.
- DeCelles, P.G., and Burden, E.T., 1992, Non-marine sedimentation in the overfilled part of the Jurassic-Cretaceous Cordilleran foreland basin: Morrison and Cloverly Formations, central Wyoming, USA: *Basin Research*, v. 4, p. 291–314, doi:10.1111/j.1365-2117.1992.tb00050.x.
- DeCelles, P.G., Gray, M.B., Ridgway, K.D., Cole, R.B., Pivnik, D.A., Pequera, N., and Srivastava, P., 1991a, Controls on synorogenic alluvial-fan architecture, Beartooth Conglomerate (Palaeocene), Wyoming and Montana: *Sedimentology*, v. 38, p. 567–590, doi:10.1111/j.1365-3091.1991.tb01009.x.
- DeCelles, P.G., Gray, M.B., Ridgway, K.D., Cole, R.B., Srivastava, P., Pequera, N., and Pivnik, D.A., 1991b, Kinematic history of a foreland uplift from Paleocene synorogenic conglomerate, Beartooth Range, Wyoming and Montana: *GSA Bulletin*, v. 103, p. 1458–1475, doi:10.1130/0016-7606(1991)103<1458:KHOAFU>2.3.CO;2.
- DeLucia, M.S., Guenther, W.R., Marshak, S., Thomson, S.N., and Ault, A.K., 2018, Thermochronology links denudation of the Great Unconformity surface to the supercontinent cycle and snowball Earth: *Geology*, v. 46, p. 167–170, doi:10.1130/G39525.1.
- Dickinson, W.R., 2004, Evolution of the North American Cordillera: *Annual Review of Earth and Planetary Sciences*, v. 32, p. 13–45, doi:10.1146/annurev.earth.32.101802.120257.
- Dickinson, W.R., and Snyder, W.S., 1978, Plate tectonics of the Laramide orogeny, in Matthews, V. ed., *Laramide folding associated with basement block faulting in the western United States*, *Geological Society of America Memoir*, v. 151, p. 355–366.
- Dorobek, S.L., Reid, S.K., Elrick, M., Bond, G.C., and Kominz, M.A., 1991, Subsidence across the Antler foreland of Montana and Idaho: Tectonic versus eustatic effects: *Kansas Geological Survey Bulletin*, v. 233, p. 232–251.
- Downes, H., Macdonald, R., Upton, B.G., Cox, K.G., Bodinier, J.-L., Mason, P.R., James, D., Hill, P.G., and Hearn Jr, B.C., 2004, Ultramafic xenoliths from the Bearpaw Mountains, Montana, USA: Evidence for multiple metasomatic events in the lithospheric mantle beneath the Wyoming craton: *Journal of Petrology*, v. 45, p. 1631–1662.
- Dudás, F.Ö., Ispolatov, V.O., Harlan, S.S., and Snee, L.W., 2010, ⁴⁰Ar/³⁹Ar Geochronology and Geochemical Reconnaissance of the Eocene Lowland Creek Volcanic Field, West-Central Montana: *The Journal of Geology*, v. 118, p. 295–304, doi:10.1086/651523.
- Dzurisin, D., Wicks, C.W., and Poland, M.P., 2012, History of surface displacements at the Yellowstone Caldera, Wyoming, from leveling surveys and InSAR observations, 1923–2008: *Professional Paper Report 1788*, doi:10.3133/pp1788.

- Eichler, C.M., Saylor, J., Julia, W., Rudolph, K., and Davis, E.M., 2020, Along-strike variation in geometry of the late Cretaceous western interior basin tracks multiple geodynamic subsidence mechanisms, in Geological Society of America.
- Einsele, G., 2013, Basin classification and depositional environments (Overview), *in* Sedimentary basins: evolution, facies, and sediment budget, Springer Science & Business Media.
- Erslev, E.A., and Sutter, J.F., 1990, Evidence for Proterozoic mylonitization in the northwestern Wyoming province: GSA Bulletin, v. 102, p. 1681–1694, doi:10.1130/0016-7606(1990)102<1681:EFPMIT>2.3.CO;2.
- Evans, K.V., Aleinikoff, J.N., Obradovich, J.D., and Fanning, C.M., 2000, SHRIMP U-Pb geochronology of volcanic rocks, Belt Supergroup, western Montana: evidence for rapid deposition of sedimentary strata: Canadian Journal of Earth Sciences, v. 37, p. 1287–1300, doi:10.1139/e00-036.
- Evans, R.D., Goodman, C., Keevil, N.B., and Urry, W.D., 1939, Work at Massachusetts Institute of Technology, Exhibit 3, in Report of the Committee on the Measurement of Geologic Time 1937-1938, Appendix J, in Annual Report of the Division of Geology and Geography: National Research Council.
- Facer, J., Downes, H., and Beard, A., 2009, In situ serpentinization and hydrous fluid metasomatism in spinel dunite xenoliths from the Bearpaw Mountains, Montana, USA: Journal of Petrology, v. 50, p. 1443–1475.
- Fan, M., and Carrapa, B., 2014, Late Cretaceous-early Eocene Laramide uplift, exhumation, and basin subsidence in Wyoming: Crustal responses to flat slab subduction: Laramide uplift and exhumation: Tectonics, v. 33, p. 509–529, doi:10.1002/2012TC003221.
- Feeley, T.C., 2003, Origin and Tectonic Implications of Across-Strike Geochemical Variations in the Eocene Absaroka Volcanic Province, United States: The Journal of Geology, v. 111, p. 329–346, doi:10.1086/373972.
- Flowers, R.M., Farley, K.A., and Ketcham, R.A., 2015, A reporting protocol for thermochronologic modeling illustrated with data from the Grand Canyon: Earth and Planetary Science Letters, v. 432, p. 425–435, doi:10.1016/j.epsl.2015.09.053.
- Flowers, R.M., Ketcham, R.A., Shuster, D.L., and Farley, K.A., 2009, Apatite (U–Th)/He thermochronometry using a radiation damage accumulation and annealing model: Geochimica et Cosmochimica Acta, v. 73, p. 2347–2365, doi:10.1016/j.gca.2009.01.015.
- Flowers, R.M., Macdonald, F.A., Siddoway, C.S., and Havranek, R., 2020, Diachronous development of Great Unconformities before Neoproterozoic Snowball Earth: Proceedings of the National Academy of Sciences, p. 201913131, doi:10.1073/pnas.1913131117.
- Flueckinger, L.A., 1970, Stratigraphy, petrography, and origin of Tertiary sediments off the front of the Beartooth Mountains, Montana-Wyoming [PhD Thesis]: Pennsylvania State University.

- Foose, R.M., Wise, D.U., and Garbarini, G.S., 1961, Structural Geology of the Beartooth Mountains, Montana and Wyoming: *GSA Bulletin*, v. 72, p. 1143–1172, doi:10.1130/0016-7606(1961)72[1143:SGOTBM]2.0.CO;2.
- Foster, D.A., Mueller, P.A., Mogk, D.W., Wooden, J.L., and Vogl, J.J., 2006, Proterozoic evolution of the western margin of the Wyoming craton: implications for the tectonic and magmatic evolution of the northern Rocky Mountains, doi:10.1139/e06-052.
- Fox, N.R., 2017, Sedimentary tectonics of the Mesoproterozoic LaHood Formation, southwest Montana: M.Sc. Thesis, v. Montana State University, p. 164.
- Fuentes, F., DeCelles, P.G., Constenius, K.N., and Gehrels, G.E., 2011, Evolution of the Cordilleran foreland basin system in northwestern Montana, U.S.A.: *GSA Bulletin*, v. 123, p. 507–533, doi:10.1130/B30204.1.
- Gast, P.W., Kulp, J.L., and Long, L.E., 1958, Absolute age of early Precambrian rocks in the Bighorn Basin of Wyoming and Montana, and southeastern Manitoba: *Eos, Transactions American Geophysical Union*, v. 39, p. 322–334, doi:10.1029/TR039i002p00322.
- Gehrels, G.E., Valencia, V.A., and Ruiz, J., 2008, Enhanced precision, accuracy, efficiency, and spatial resolution of U-Pb ages by laser ablation–multicollector–inductively coupled plasma–mass spectrometry: *Geochemistry, Geophysics, Geosystems*, v. 9, doi:10.1029/2007GC001805.
- Geissman, J.G., and Mogk, D., 1986, Late Archean tectonic emplacement of the Stillwater Complex along reactivated basement structures, northern Beartooth Mountains southern Montana, U.S., in Aldrich, J.J.Jr. and Laughlin, A.W. eds., 6th International Conference on Basement Tectonics, p. 25–44.
- Geyer, G., and Landing, E., 2017, The Precambrian–Phanerozoic and Ediacaran–Cambrian boundaries: a historical approach to a dilemma: *Geological Society, London, Special Publications*, v. 448, p. 311–349, doi:10.1144/SP448.10.
- Gholamrezaie, E., Scheck-Wenderoth, M., Sippel, J., and Strecker, M.R., 2018, Variability of the geothermal gradient across two differently aged magma-rich continental rifted margins of the Atlantic Ocean: the Southwest African and the Norwegian margins: *Solid Earth*, v. 9, p. 139–158, doi:10.5194/se-9-139-2018.
- Giegengack, R., Omar, G.I., and Johnson, K.R., 1986, A reconnaissance fission track uplift chronology for the northwest margin of the Bighorn Basin: Montana Geological Society and yellowstone bighorn Research Association Joint Field Conference and Symposium: *Geology of the Beartooth Uplift and Adjacent Basins*, p. 179–184.
- Gifford, J.N., Mueller, P.A., Foster, D.A., and Mogk, D.W., 2014, Precambrian Crustal Evolution in the Great Falls Tectonic Zone: Insights from Xenoliths from the Montana Alkali Province: *The Journal of Geology*, v. 122, p. 531–548, doi:10.1086/677262.
- Giletti, B.J., 1971, Discordant isotopic ages and excess argon in biotites: *Earth and Planetary Science Letters*, v. 10, p. 157–164, doi:10.1016/0012-821X(71)90001-X.

- Giletti, B.J., 1966, Isotopic ages from southwestern Montana: *Journal of Geophysical Research*, v. 71.
- Gingerich, P.D., 1983, Paleocene-Eocene Faunal Zones and a Preliminary Analysis of Laramide Structural Deformation in the Clark's Fork Basin, Wyoming: *Geology Of the Bighorn Basin; 34th Annual Field Conference Guidebook*, p. 185–195.
- Ginster, U., Reiners, P.W., Nasdala, L., and Chanmuang N., C., 2019, Annealing kinetics of radiation damage in zircon: *Geochimica et Cosmochimica Acta*, v. 249, p. 225–246, doi:10.1016/j.gca.2019.01.033.
- Godd  ris, Y., Donnadieu, Y., N  d  lec, A., Dupr  , B., Dessert, C., Grard, A., Ramstein, G., and Fran  ois, L.M., 2003, The Sturtian 'snowball' glaciation: fire and ice: *Earth and Planetary Science Letters*, v. 211, p. 1–12, doi:10.1016/S0012-821X(03)00197-3.
- G  mez, E., Jordan, T.E., Allmendinger, R.W., and Cardozo, N., 2005, Development of the Colombian foreland-basin system as a consequence of diachronous exhumation of the northern Andes: *GSA Bulletin*, v. 117, p. 1272–1292, doi:10.1130/B25456.1.
- Graf, W.L., 1970, The Geomorphology of the Glacial Valley Cross Section: *Arctic and Alpine Research*, v. 2, p. 303–312, doi:10.1080/00040851.1970.12003589.
- Green, D., 1972, Morphologies and fission track ages of zircons from quartzites of the Hellroaring Lakes area, Pine Creek quarry, and Chrome Mountain area, Beartooth Mountains, Montana [Masters Thesis]: University of Cincinnati.
- Grove, M., and Harrison, T.M., 1996, $^{40}\text{Ar}^*$ diffusion in Fe-rich biotite: *American Mineralogist*, v. 81, p. 940–951.
- Guenthner, W.R., 2021, Implementation of an Alpha Damage Annealing Model for Zircon (U-Th)/He Thermochronology With Comparison to a Zircon Fission Track Annealing Model: *Geochemistry, Geophysics, Geosystems*, v. 22, p. e2019GC008757, doi:10.1029/2019GC008757.
- Guenthner, W.R., Reiners, P.W., Ketcham, R.A., Nasdala, L., and Giester, G., 2013, Helium diffusion in natural zircon: Radiation damage, anisotropy, and the interpretation of zircon (U-Th)/He thermochronology: *American Journal of Science*, v. 313, p. 145–198, doi:10.2475/03.2013.01.
- Guerrero, J.C., Mueller, P.A., and Mogk, D.W., 2016, Provenance study and geochemical analysis of the LaHood Formation of the Bridger Range, Montana: *Northwest Geology*, v. 45, p. 21–28.
- Guy, R., and Sinha, A., 1988, Petrology and geochemistry of Archean basement lithologies in the Yankee Jim and Lamar Canyon areas, Montana and Wyoming: *Precambrian and Mesozoic Plate Margins: Montana Bureau of Mines, Special Publication*, v. 96, p. 53–68.
- Hague, A., 1899, The Early Tertiary Volcanoes of the Absaroka Range: *Science*, v. 9, p. 425–442.

- Halverson, G.P., Dudás, F.Ö., Maloof, A.C., and Bowring, S.A., 2007, Evolution of the $^{87}\text{Sr}/^{86}\text{Sr}$ composition of Neoproterozoic seawater: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 256, p. 103–129, doi:10.1016/j.palaeo.2007.02.028.
- Hames, W., and Harms, T., 2013, Laser $^{40}\text{Ar}/^{39}\text{Ar}$ thermochronology of single crystals from Paleoproterozoic to Archean rocks of the northern Wyoming Province, Big Sky orogen, and Great Falls tectonic zone: Giletti's line revisited, in Geological Society of America Abstracts with Programs, v. 45, p. 223.
- Harlan, S.S., Geissman, J.W., Snee, L.W., and Reynolds, R.L., 1996, Late Cretaceous remagnetization of Proterozoic mafic dikes, southern Highland Mountains, southwestern Montana: A paleomagnetic and $^{40}\text{Ar}/^{39}\text{Ar}$ study: GSA Bulletin, v. 108, p. 653–668, doi:10.1130/0016-7606(1996)108<0653:LCROPM>2.3.CO;2.
- Harms, T.A. et al., 2004a, Character and origin of Precambrian fabrics and structures in the Tobacco Root Mountains, Montana, in Special Paper 377: Precambrian Geology of the Tobacco Root Mountains, Montana, Geological Society of America, v. 377, p. 203–226, doi:10.1130/0-8137-2377-9.203.
- Harms, T.A., Brady, J.B., Burger, H.R., and Cheney, J.T., 2004b, Advances in the geology of the Tobacco Root Mountains, Montana, and their implications for the history of the northern Wyoming province, in Special Paper 377: Precambrian Geology of the Tobacco Root Mountains, Montana, Geological Society of America, v. 377, p. 227–243, doi:10.1130/0-8137-2377-9.227.
- Harms, T.A., Brady, J.B., College, S., and Cheney, J.T., 2006, Exploring the Proterozoic Big Sky Orogeny in southwest Montana: 19th Annual Keck Symposium, p. 6.
- Harrison, J.E., Griggs, A.B., and Wells, J.D., 1974, Tectonic features of the Precambrian Belt Basin and their influence on post-Belt structures, doi:10.3133/pp866.
- Henry, C.D., Castor, S.B., Starkel, W.A., Ellis, B.S., Wolff, J.A., Laravie, J.A., McIntosh, W.C., and Heizler, M.T., 2017, Geology and evolution of the McDermitt caldera, northern Nevada and southeastern Oregon, western USA: Geosphere, v. 13, p. 1066–1112, doi:10.1130/GES01454.1.
- Henry, D., Mueller, P.A., Wooden, J.L., Warner, J., and Lee-Berman, R., 1982, Granulite grade supracrustal assemblages of the Quad Creek area, eastern Beartooth Mountains, Montana: Montana Bureau of Mines and Geology, Special Publication, v. 84, p. 147–155.
- Hoffman, P.F., 1988, United Plates of America, The Birth of a Craton: Early Proterozoic Assembly and Growth of Laurentia: Annual Review of Earth and Planetary Sciences, v. 16, p. 63.
- Horton, F., 2015, Did phosphorus derived from the weathering of large igneous provinces fertilize the Neoproterozoic ocean? Geochemistry, Geophysics, Geosystems, v. 16, p. 1723–1738, doi:https://doi.org/10.1002/2015GC005792.
- Hoy, R.G., and Ridgway, K.D., 1997, Structural and sedimentological development of footwall growth synclines along an intraforeland uplift, east-central Bighorn Mountains,

- Wyoming: GSA Bulletin, v. 109, p. 915–935, doi:10.1130/0016-7606(1997)109<0915:SASDOF>2.3.CO;2.
- Hurford, A.J., and Green, P.F., 1983, The zeta age calibration of fission-track dating: Chemical Geology (Isotope Geoscience Section), v. 1, p. 285–317.
- Husson, J.M., and Peters, S.E., 2017, Atmospheric oxygenation driven by unsteady growth of the continental sedimentary reservoir: Earth and Planetary Science Letters, v. 460, p. 68–75, doi:10.1016/j.epsl.2016.12.012.
- Iddings, J.P., 1895, Absarokite-shoshonite-banakitite series: The Journal of Geology, v. 3, p. 935–959.
- James, H.L., and Hedge, C.E., 1980, Age of the basement rocks of southwest Montana: GSA Bulletin, v. 91, p. 11–15, doi:10.1130/0016-7606(1980)91<11:AOTBRO>2.0.CO;2.
- Jamieson, R.A., and Beaumont, C., 2011, Coeval thrusting and extension during lower crustal ductile flow – implications for exhumation of high-grade metamorphic rocks: Journal of Metamorphic Geology, v. 29, p. 33–51, doi:10.1111/j.1525-1314.2010.00908.x.
- Janecke, S.U., 2007, Cenozoic extensional processes and tectonics in the northern Rocky Mountains: Southwest Montana and eastern Idaho: Northwest Geology, v. 36, p. 111–132.
- Jensen, J.L., Siddoway, C.S., Reiners, P.W., Ault, A.K., Thomson, S.N., and Steele-MacInnis, M., 2018, Single-crystal hematite (U–Th)/He dates and fluid inclusions document widespread Cryogenian sand injection in crystalline basement: Earth and Planetary Science Letters, v. 500, p. 145–155, doi:10.1016/j.epsl.2018.08.021.
- Jobling, J., 1974, Stratigraphy, petrology, and structure of the Laramide (Paleocene) sediments marginal to the Beartooth Mountains: Montana [Ph.D. thesis] University Park, Pennsylvania, Pennsylvania State University.
- Johnson, J.E., Flowers, R.M., Baird, G.B., and Mahan, K.H., 2017, “Inverted” zircon and apatite (U–Th)/He dates from the Front Range, Colorado: High-damage zircon as a low-temperature (<50 °C) thermochronometer: Earth and Planetary Science Letters, v. 466, p. 80–90, doi:10.1016/j.epsl.2017.03.002.
- Karlstrom, K., Hagadorn, J., Gehrels, G., Matthews, W., Schmitz, M., Madronich, L., Mulder, J., Pecha, M., Giesler, D., and Crossey, L., 2018, Cambrian Sauk transgression in the Grand Canyon region redefined by detrital zircons: Nature Geoscience, v. 11, p. 438–443, doi:10.1038/s41561-018-0131-7.
- Kauffman, E.G., and Caldwell, W.G.E., 1992, The Western Interior Basin in space and time: Geologic Association of Canada Special Paper 39.
- Keller, C.B., Husson, J.M., Mitchell, R.N., Bottke, W.F., Gernon, T.M., Boehnke, P., Bell, E.A., Swanson-Hysell, N.L., and Peters, S.E., 2019, Neoproterozoic glacial origin of the Great Unconformity: Proceedings of the National Academy of Sciences, v. 116, p. 1136–1145, doi:10.1073/pnas.1804350116.

- Ketcham, R.A., 2005a, Forward and Inverse Modeling of Low-Temperature Thermochronometry Data: Reviews in Mineralogy and Geochemistry, v. 58, p. 275–314, doi:10.2138/rmg.2005.58.11.
- Ketcham, R., 2005b, HeFTy: Forward and inverse modeling thermochronometer systems: Computational Tools for Low-Temperature Thermochronometer Interpretation. Mineralogical Society of America, Chantilly, VA.
- Koenig, D. 1981-, 2015, Low-temperature Thermochronology of the Beartooth Conglomerate [Masters Thesis]: University of Houston, Houston, TX.
- Lachenbruch, A.H., 1979, Heat Flow in the Basin and Range Province and Thermal Effects of Tectonic Extension, in Rybach, L. and Stegena, L. eds., Geothermics and Geothermal Energy, Basel, Birkhäuser, Contributions to Current Research in Geophysics (CCRG), p. 34–50, doi:10.1007/978-3-0348-6525-8_3.
- Lau, K.V., Macdonald, F.A., Maher, K., and Payne, J.L., 2017, Uranium isotope evidence for temporary ocean oxygenation in the aftermath of the Sturtian Snowball Earth: Earth and Planetary Science Letters, v. 458, p. 282–292, doi:10.1016/j.epsl.2016.10.043.
- Li, Z.X. et al., 2008, Assembly, configuration, and break-up history of Rodinia: A synthesis: Precambrian Research, v. 160, p. 179–210, doi:10.1016/j.precamres.2007.04.021.
- Li, Z.-X., and Zhong, S., 2009, Supercontinent–superplume coupling, true polar wander and plume mobility: Plate dominance in whole-mantle tectonics: Physics of the Earth and Planetary Interiors, v. 176, p. 143–156, doi:10.1016/j.pepi.2009.05.004.
- Liu, S., Nummedal, D., and Gurnis, M., 2014, Dynamic versus flexural controls of Late Cretaceous Western Interior Basin, USA: Earth and Planetary Science Letters, v. 389, p. 221–229, doi:10.1016/j.epsl.2014.01.006.
- Long, S.P., 2018, Geometry and magnitude of extension in the Basin and Range Province (39°N), Utah, Nevada, and California, USA: Constraints from a province-scale cross section: GSA Bulletin, v. 131, p. 99–119, doi:10.1130/B31974.1.
- Lopez, D.A., 2001, Preliminary Geologic Map of the Red Lodge 30' x 60' Quadrangle South-Central Montana: Montana Bureau of Mines and Geology Open File Geologic Map 423.
- Macdonald, F.A., Schmitz, M.D., Crowley, J.L., Roots, C.F., Jones, D.S., Maloof, A.C., Strauss, J.V., Cohen, P.A., Johnston, D.T., and Schrag, D.P., 2010, Calibrating the Cryogenian: Science, v. 327, p. 1241–1243, doi:10.1126/science.1183325.
- Mallory, W., 1972, Geologic Atlas of the Rocky Mountain region: Rocky Mountain Association of Geologists: Denver, Colorado, v. 331.
- Malone, D., Craddock, J., and Kenderes, S., 2017, Detrital zircon geochronology and provenance of the Middle Cambrian Flathead Sandstone, Park County, Wyoming: The Mountain Geologist, v. 54, p. 86–103, doi:10.31582/rmag.mg.54.2.86.
- Matthews, W., Guest, B., and Madronich, L., 2018, Latest Neoproterozoic to Cambrian detrital zircon facies of western Laurentia: Geosphere, v. 14, p. 243–264, doi:10.1130/GES01544.1.

- Maughan, E.K., 1989, Geology and petroleum potential, central Montana province: U.S. Geologic Survey Open-File Report 88-450 N.
- Maughan, E.K., 1993, Stratigraphic and Structural Summary for Central Montana, in Field Conference Guidebook: Energy and Mineral Resources of Central Montana, Montana Geologic Society, p. 3–20.
- Maughan, E., 1990, Summary of the Ancestral Rocky Mountains Epeirogeny in Wyoming and adjacent areas: U.S. Geologic Survey Open-File Report 90-447.
- McDannell, K.T., and Flowers, R.M., 2020, Vestiges of the Ancient: Deep-Time Noble Gas Thermochronology: *Elements*, v. 16, p. 325–330, doi:10.2138/gselements.16.5.325.
- McDannell, K.T., Schneider, D.A., Zeitler, P.K., O’Sullivan, P.B., and Issler, D.R., 2019, Reconstructing deep-time histories from integrated thermochronology: An example from southern Baffin Island, Canada: *Terra Nova*, v. 31, p. 189–204, doi:10.1111/ter.12386.
- McDannell, K.T., Zeitler, P.K., and Schneider, D.A., 2018, Instability of the southern Canadian Shield during the late Proterozoic: *Earth and Planetary Science Letters*, v. 490, p. 100–109, doi:10.1016/j.epsl.2018.03.012.
- McPhee, J., 1981, Basin and range: Macmillan.
- Mitchell, S.G., and Reiners, P.W., 2003, Influence of wildfires on apatite and zircon (U-Th)/He ages: *Geology*, v. 31, p. 1025–1028, doi:10.1130/G19758.1.
- Mogk, D., Henry, D., Mueller, P., and Foster, D., 2012, Origins of a continent: *Yellowstone Science*, v. 20, p. 22–32.
- Mogk, D.W., Mueller, P.A., and Wooden, J.L., 1988, Archean Tectonics of the North Snowy Block, Beartooth Mountains, Montana: *The Journal of Geology*, v. 96, p. 125–141, doi:10.1086/629205.
- Mogk, D.W., Mueller, P.A., and Wooden, J.L., 1992, The nature of Archean terrane boundaries: an example from the northern Wyoming Province: *Precambrian Research*, v. 55, p. 155–168, doi:10.1016/0301-9268(92)90020-O.
- Montgomery, C.W., and Lytwyn, J.N., 1984, Rb-Sr Systematics and Ages of Principal Precambrian Lithologies in the South Snowy Block, Beartooth Mountains: *The Journal of Geology*, v. 92, p. 103–112, doi:10.1086/628837.
- Mueller, P.A., Burger, H.R., Wooden, J.L., Brady, J.B., Cheney, J.T., Harms, T.A., Heatherington, A.L., and Mogk, D.W., 2005, Paleoproterozoic Metamorphism in the Northern Wyoming Province: Implications for the Assembly of Laurentia: *The Journal of Geology*, v. 113, p. 169–179, doi:10.1086/427667.
- Mueller, P.A., Burger, H.R., Wooden, J.L., Heatherington, A.L., Mogk, D.W., and D’Arcy, K., 2004, Age and evolution of the Precambrian crust of the Tobacco Root Mountains, Montana, in Special Paper 377: Precambrian Geology of the Tobacco Root Mountains, Montana, Geological Society of America, v. 377, p. 181–202, doi:10.1130/0-8137-2377-9.181.

- Mueller, P., and Cordua, W., 1976, Rb-Sr whole rock age of gneisses from the Horse Creek area: Tobacco Root Mountains, Montana: *Isochron/West*, v. 16, p. 33–36.
- Mueller, P.A., Heatherington, A.L., Kelly, D.M., Wooden, J.L., and Mogk, D.W., 2002, Paleoproterozoic crust within the Great Falls tectonic zone: Implications for the assembly of southern Laurentia: *Geology*, v. 30, p. 127–130, doi:10.1130/0091-7613(2002)030<0127:PCWTGF>2.0.CO;2.
- Mueller, P.A., Mogk, D.W., Henry, D.J., Wooden, J.L., and Foster, D.A., 2008, Geologic evolution of the Beartooth Mountains: Insights from petrology and geochemistry: *Northwestern Geology*, v. 37, p. 5–20.
- Mueller, P., Mogk, D., Wooden, J., and Spake, D., 2016, U-Pb ages of zircons from the Lower Belt Supergroup and proximal crystalline basement: Implications for the early evolution of the Belt Basin, in *Geological Society of America Special Papers*, Geological Society of America, v. 522, p. 283–303, doi:10.1130/2016.2522(11).
- Mueller, P.A., Shuster, R.D., Wooden, J.L., Erslev, E.A., and Bowes, D.R., 1993, Age and composition of Archean crystalline rocks from the southern Madison Range, Montana: Implications for crustal evolution in the Wyoming craton: *GSA Bulletin*, v. 105, p. 437–446, doi:10.1130/0016-7606(1993)105<0437:AACOAC>2.3.CO;2.
- Mueller, P.A., Wooden, J.L., Mogk, D.W., Henry, D.J., and Bowes, D.R., 2010, Rapid growth of an Archean continent by arc magmatism: *Precambrian Research*, v. 183, p. 70–88, doi:10.1016/j.precamres.2010.07.013.
- Murray, K.E., Orme, D.A., and Reiners, P.W., 2014, Effects of U–Th-rich grain boundary phases on apatite helium ages: *Chemical Geology*, v. 390, p. 135–151, doi:10.1016/j.chemgeo.2014.09.023.
- Nance, R.D., and Murphy, J.B., 2019, Supercontinents and the case for Pannotia: *Geological Society, London, Special Publications*, v. 470, p. 65–86, doi:10.1144/SP470.5.
- Nuccio, V.F., and Condon, S.M., 1996, Burial and thermal history of the Paradox Basin, Utah and Colorado, and petroleum potential of the middle Pennsylvanian Paradox Formation: *U.S. Geological Survey Bulletin* 2000-O.
- Nunes, P.D., and Tilton, G.R., 1971, Uranium-Lead Ages of Minerals from the Stillwater Igneous Complex and Associated Rocks, Montana: *GSA Bulletin*, v. 82, p. 2231–2250, doi:10.1130/0016-7606(1971)82[2231:UAOMFT]2.0.CO;2.
- Omar, G.I., Lutz, T.M., and Giegengack, R., 1994, Apatite fission-track evidence for Laramide and post-Laramide uplift and anomalous thermal regime at the Beartooth overthrust, Montana-Wyoming: *Geological Society of America Bulletin*, p. 12.
- O'Neill, J.M., Duncan, M.S., and Zartman, R.E., 1988, Proterozoic gneiss dome in the Highland Mountains, southwestern Montana, in Lewis, S.E. and Berg, R.B. eds., *Precambrian and Mesozoic plate margins: Montana, Idaho, and Wyoming, with field guides for the 8th International conference on Basement Tectonics*, Montana Bureau of Mine and Geology Special Publication 96, p. 81–88.

- O'Neill, J.M., Ferris, D.C., Schmidt, C.J., and Hanneman, D.L., 1986, Recurrent movement along northwest-trending faults at the southern margin of the Belt Basin, Highland Mountains, southwestern Montana: Belt Supergroup: Montana Bureau of Mines and Geology Special Publication, v. 94, p. 208–216.
- O'Neill, J.M., Klepper, M.R., Smedes, H.W., Hanneman, D.L., Fraser, G.D., and Mehnert, H.H., 1996, Geologic map and cross sections of the central and southern Highland Mountains, southwestern Montana: U.S. Geologic Survey Miscellaneous Field Studies Map I-2525.
- Orme, D.A., 2020, New Timing Constraints for the Onset of Laramide Deformation in Southwest Montana Challenge our Understanding of the Development of a Thick-Skinned Structural Style During Flat-Slab Subduction: *Tectonics*, v. 39, doi:10.1029/2020TC006193.
- Orme, D.A., Guenther, W.R., Laskowski, A.K., and Reiners, P.W., 2016, Long-term tectonothermal history of Laramide basement from zircon–He age–eU correlations: *Earth and Planetary Science Letters*, v. 453, p. 119–130, doi:10.1016/j.epsl.2016.07.046.
- Page, N.J., 1977, Stillwater Complex, Montana; rock succession, metamorphism and structure of the complex and adjacent rocks: US Govt. Print. Off.
- Page, N., Zientek, M., and Czamanske, G., 1985, Geologic and structural setting of the Stillwater Complex: The Stillwater Complex, Montana: Geology and Guide: Montana Bureau of Mines and Geology Special Publication, v. 92, p. 1–8.
- Parcell, W.C., and Williams, M.K., 2005, Mixed sediment deposition in a retro-arc foreland basin: Lower Ellis Group (M. Jurassic), Wyoming and Montana, U.S.A.: *Sedimentary Geology*, v. 177, p. 175–194, doi:10.1016/j.sedgeo.2005.02.007.
- Pardee, J.T., 1950, Late Cenozoic block faulting in western Montana: *Geological Society of America Bulletin*, v. 61, p. 359–406.
- Peale, A., 1896, Description of the Three Forks sheet, Montana: US Geological Survey Geol: Atlas, Folio, v. 24.
- Pearson, D.M., Parker, S.D., Montoya, L.M., and Brennan, D.T., 2019, No flat-slab needed: the transition from thin- to thick-skinned thrusting in the Idaho-Montana fold-thrust belt was controlled by the pre-orogenic stratigraphy, in p. 341061, doi:10.1130/abs/2019AM-341061.
- Peterman, Z.E., 1979, Geochronology and the Archean of the United States: *Economic Geology*, v. 74, p. 1544–1562.
- Peters, S.E., and Gaines, R.R., 2012, Formation of the ‘Great Unconformity’ as a trigger for the Cambrian explosion: *Nature*, v. 484, p. 363–366, doi:10.1038/nature10969.
- Peyton, S.L., and Carrapa, B., 2013, An Overview of Low-temperature Thermochronology in the Rocky Mountains and Its Application to Petroleum System Analysis: , p. 37–70, doi:10.1306/13381689St653578.

- Peyton, S.L., Reiners, P.W., Carrapa, B., and DeCelles, P.G., 2012, Low-temperature thermochronology of the northern Rocky Mountains, western U.S.A.: *American Journal of Science*, v. 312, p. 145–212, doi:10.2475/02.2012.04.
- Pierce, W.G., and Nelson, W.H., 1971, Geologic map of the Beartooth Butte quadrangle, Park County, Wyoming: Geologic Quadrangle, doi:10.3133/gq935.
- Planavsky, N.J., Rouxel, O.J., Bekker, A., Lalonde, S.V., Konhauser, K.O., Reinhard, C.T., and Lyons, T.W., 2010, The evolution of the marine phosphate reservoir: *Nature*, v. 467, p. 1088–1090, doi:10.1038/nature09485.
- Pogge von Strandmann, P.A.E., Stüeken, E.E., Elliott, T., Poulton, S.W., Dehler, C.M., Canfield, D.E., and Catling, D.C., 2015, Selenium isotope evidence for progressive oxidation of the Neoproterozoic biosphere: *Nature Communications*, v. 6, p. 10157, doi:10.1038/ncomms10157.
- Raedeke, L., 1983, Platinum group metals in the Stillwater Complex, in 1983. Symposium on genesis of Rocky Mountain ore deposits: Denver, Colorado, Exploration Geologists Society Denver Region Proceedings, p. 31–37.
- Reade, N.Z., Biddle, J.M., Ricketts, J.W., and Amato, J.M., 2020, Zircon (U-Th)/He Thermochronologic Constraints on the Long-Term Thermal Evolution of Southern New Mexico and Western Texas: *Lithosphere*, v. 2020, doi:10.2113/2020/8881315.
- Reid, R.R., McMannis, W.J., and Palmquist, J.C., 1975, Precambrian Geology of North Snowy Block, Beartooth Mountains, Montana: *Geological Society of America Special Papers*, v. 157, 152 p.
- Reiners, P.W., and Brandon, M.T., 2006, Using Thermochronology to Understand Orogenic Erosion: *Annual Review of Earth and Planetary Sciences*, v. 34, p. 419–466, doi:10.1146/annurev.earth.34.031405.125202.
- Reiners, P.W., Carlson, R.W., Renne, P.R., Cooper, K.M., Granger, D.E., McLean, N.M., and Schoene, B., 2017, *Geochronology and thermochronology*: John Wiley & Sons.
- Reiners, P.W., Farley, K.A., and Hickey, H.J., 2002, He diffusion and (U-Th)/He thermochronometry of zircon: initial results from Fish Canyon Tuff and Gold Butte: *Tectonophysics*, v. 349, p. 297–308, doi:10.1016/S0040-1951(02)00058-6.
- Reiners, P.W., Spell, T.L., Nicolescu, S., and Zanetti, K.A., 2004, Zircon (U-Th)/He thermochronometry: He diffusion and comparisons with $^{40}\text{Ar}/^{39}\text{Ar}$ dating: *Geochimica et Cosmochimica Acta*, v. 68, p. 1857–1887, doi:10.1016/j.gca.2003.10.021.
- Reiners, P.W., Thomson, S.N., McPhillips, D., Donelick, R.A., and Roering, J.J., 2007, Wildfire thermochronology and the fate and transport of apatite in hillslope and fluvial environments: *Journal of Geophysical Research: Earth Surface*, v. 112, doi:10.1029/2007JF000759.
- Retallack, G.J., 2013, Early Cambrian Humid, Tropical, Coastal Paleosols from Montana, USA, in Driese, S.G. ed., *New Frontiers in Paleopedology and Terrestrial Paleoclimatology*:

- Paleosols and Soil Surface Analog Systems, SEPM (Society for Sedimentary Geology), p. 257–272, doi:10.2110/sepmsp.104.09.
- Roberts, H., Dahl, P., Kelley, S., and Frei, R., 2002, New 207Pb – 206Pb and 40Ar – 39Ar ages from SW Montana, USA: constraints on the Proterozoic and Archæan tectonic and depositional history of the Wyoming Province: *Precambrian Research*, p. 25.
- Ronemus, C.B., Orme, D.A., Campbell, S., Black, S.R., and Cook, J., 2020, Mesoproterozoic–Early Cretaceous provenance and paleogeographic evolution of the Northern Rocky Mountains: Insights from the detrital zircon record of the Bridger Range, Montana, USA: *GSA Bulletin*, doi:10.1130/B35628.1.
- Rudnick, R., Ireland, T., Gehrels, G., Irving, A., Chesley, J., and Hanchar, J., 1998, Dating mantle metasomatism: U-Pb geochronology of zircons in cratonic mantle xenoliths from Montana and Tanzania, in *International Kimberlite Conference: Extended Abstracts*, v. 7, p. 754–756.
- Sahoo, S.K., Planavsky, N.J., Kendall, B., Wang, X., Shi, X., Scott, C., Anbar, A.D., Lyons, T.W., and Jiang, G., 2012, Ocean oxygenation in the wake of the Marinoan glaciation: *Nature*, v. 489, p. 546–549, doi:10.1038/nature11445.
- Sclater, J.G., and Christie, P. a. F., 1980, Continental stretching: An explanation of the Post-Mid-Cretaceous subsidence of the central North Sea Basin: *Journal of Geophysical Research: Solid Earth*, v. 85, p. 3711–3739, doi:10.1029/JB085iB07p03711.
- Sears, J.W., and Price, R.A., 1978, The Siberian Connection: A case for Precambrian separation of the North American and Siberian cratons: *Geology*, v. 6, p. 267–270, doi:10.1130/0091-7613(1978)6<267:TSCACF>2.0.CO;2.
- Sharry, J., Langan, R.T., Jovanovich, D.B., Jones, G.M., Hill, N.R., and Guidish, T.M., 1986, Enhanced Imaging of the COCORP Seismic Line, Wind River Mountains, in *Reflection Seismology: A Global Perspective*, American Geophysical Union (AGU), p. 223–236, doi:10.1029/GD013p0223.
- Siddoway, C., Palladino, G., Prosser, G., Freedman, D., and Duckworth, W.C., 2019, Basement-hosted sand injectites: Use of field examples to advance understanding of hydrocarbon reservoirs in fractured crystalline basement rocks: *Geological Society, London, Special Publications*, v. 493.
- Simons, F.S., and Armbrustmacher, T.J., 1976, High-level plateaus of the southeastern Beartooth Mountains, Montana and Wyoming—remnants of an exhumed sub-Cambrian marine plain: *Jour. Res. US Geol. Surv.*, v. 4, p. 387–396.
- Sloss, L.L., 1963, Sequences in the Cratonic Interior of North America: *Geological Society of America Bulletin*, v. 74, p. 93, doi:10.1130/0016-7606(1963)74[93:SITCIO]2.0.CO;2.
- Smithson, S.B., Brewer, J., Kaufman, S., Oliver, J., and Hurich, C., 1978, Nature of the Wind River thrust, Wyoming, from COCORP deep-reflection data and from gravity data: *Geology*, v. 6, p. 648–652, doi:10.1130/0091-7613(1978)6<648:NOTWRT>2.0.CO;2.

- Sobolev, S.V., and Brown, M., 2019, Surface erosion events controlled the evolution of plate tectonics on Earth: *Nature*, v. 570, p. 52–57, doi:10.1038/s41586-019-1258-4.
- Sousa, F.J., and Farley, K.A., 2020, A Framework for Evaluating Variation in (U-Th)/He Datasets: *Minerals*, v. 10, p. 1111, doi:10.3390/min10121111.
- Squire, R.J., Campbell, I.H., Allen, C.M., and Wilson, C.J.L., 2006, Did the Transgondwanan Supermountain trigger the explosive radiation of animals on Earth? *Earth and Planetary Science Letters*, v. 250, p. 116–133, doi:10.1016/j.epsl.2006.07.032.
- Symons, D.T.A., and Timmins, E.A., 1992, Geotectonics of the cratonic margin from paleomagnetism of the Middle Proterozoic Aldridge (Prichard) Formation and Moyie sills of British Columbia and Montana, *in* Bartholomew, M.J., Hyndman, D.W., Mogk, D.W., and Mason, R. eds., *Basement Tectonics 8: Characterization and Comparison of Ancient and Mesozoic Continental Margins*, Dordrecht, Springer Netherlands, *Proceedings of the International Conferences on Basement Tectonics*, p. 373–384, doi:10.1007/978-94-011-1614-5_25.
- Thacker, J.O., Kelley, S.A., and Karlstrom, K.E. Late Cretaceous–Recent low-temperature cooling history and tectonic analysis of the Zuni Mountains, west-central New Mexico: *Tectonics*, v. n/a, p. e2020TC006643, doi:10.1029/2020TC006643.
- Thacker, J.O., Lageson, D.R., and Mogk, D.W., 2017, Subsurface structural and mineralogical characterization of the Laramide South Prairie fault in the Stillwater Complex, Beartooth Mountains, Montana: *Lithosphere*, v. 9, p. 100–116, doi:10.1130/L562.1.
- Thomas, R., 2007, A field guide to the Cambrian section at camp creek, southwest Montana: *Northwest Geology*, v. 36, p. 231–244.
- Thompson, A.B., 1984, Geothermal Gradients Through Time, *in* Holland, H.D. and Trendall, A.F. eds., *Patterns of Change in Earth Evolution*, Berlin, Heidelberg, Springer, Dahlem Workshop Reports Physical, Chemical, and Earth Sciences Research Reports, p. 345–355, doi:10.1007/978-3-642-69317-5_20.
- Walker, J., Geissman, J., Bowring, S., and Babcock, L., 2018, GSA geological time scale v. 5.0: Geological Society of America.
- Wall, C.J., Scoates, J.S., Weis, D., Friedman, R.M., Amini, M., and Meurer, W.P., 2018, The Stillwater Complex: integrating zircon geochronological and geochemical constraints on the age, emplacement history and crystallization of a large, open-system layered intrusion: *Journal of Petrology*, v. 59, p. 153–190.
- Waples, D.W., Pacheco, J., and Vera, A., 2004, A method for correcting log-derived temperatures in deep wells, calibrated in the Gulf of Mexico: *Petroleum Geoscience*, v. 10, p. 239–245, doi:10.1144/1354-079302-542.
- Webber, J.R., Kalakay, T.J., Peterman, E.M., McAleer, R., and Mogk, D.W., 2019, Deciphering the tectonometamorphic evolution and localization of deformation within the Wyoming Province of southwest Montana, *in* AGU Fall Meeting Abstracts, v. 2019, p. T13E-0237.

- Wegmann, K.W. et al., 2007, Position of the Snake River watershed divide as an indicator of geodynamic processes in the greater Yellowstone region, western North America: *Geosphere*, v. 3, p. 272–281, doi:10.1130/GES00083.1.
- Weyand, E.L., 1989, U-Pb zircon geochronology of Archean rocks from the Spanish Peaks area, Madison Range, Montana [PhD Thesis]: University of Florida, Gainesville, FL.
- Whitlock, C., Reasoner, M.A., and Key, C.H., 2002, Paleoenvironmental history of the Rocky Mountain region during the past 20,000 years: *Rocky Mountain futures: an ecological perspective*. Island Press, Washington, DC, p. 41–57.
- Whitmeyer, S.J., and Karlstrom, K.E., 2007, Tectonic model for the Proterozoic growth of North America: *Geosphere*, p. 40.
- Whitney, D.L., and Evans, B.W., 2010, Abbreviations for names of rock-forming minerals: *American Mineralogist*, v. 95, p. 185–187, doi:10.2138/am.2010.3371.
- Will, C., 2013, Temperature and pressure conditions of Archean amphibolite-granulite facies metamorphic xenoliths from the eastern Beartooth Mountains, Montana and Wyoming, USA [Master of Science]: Louisiana State University, Baton Rouge, LA.
- Wilson, C.W., Jr., 1936, Geology of Nye-Bowler Lineament, Stillwater and Carbon Counties, Montana: *AAPG Bulletin*, v. 20, p. 1161–1188, doi:10.1306/3D932E20-16B1-11D7-8645000102C1865D.
- Winston, D., 1986, Sedimentation and Tectonics of the Middle Proterozoic Belt Basin and Their Influence on Phanerozoic Compression and Extension in Western Montana and Northern Idaho: Part II. Northern Rocky Mountains: v. 155, p. 87–118.
- Wise, D.U., 2000, Laramide Structures in Basement and Cover of the Beartooth Uplift Near Red Lodge, Montana: *AAPG Bulletin*, v. 84, p. 360–375, doi:10.1306/C9EBCDF1-1735-11D7-8645000102C1865D.
- Wise, D.U., 1997, Pseudo Tear Faults at Red Lodge, MT: Clues to Late-stage Basin Understuffing and Tectonics of the Greater Beartooth Block, Bighorn Basin Symposium Guidebook, p. 71–99.
- Wooden, J.L., Mueller, P.A., Mogk, D.A., Bowes, D.R., Lewis, S., and Berg, R., 1988, A review of the geochemistry and geochronology of Archean rocks of the Beartooth Mountains, Montana and Wyoming: Precambrian and Mesozoic plate margins: Montana, Idaho and Wyoming. Edited by SE Lewis and RB Berg. Montana Bureau of Mines and Geology, Special Publication, v. 96, p. 23–42.
- Yamada, R., Murakami, M., and Tagami, T., 2007, Statistical modelling of annealing kinetics of fission tracks in zircon; Reassessment of laboratory experiments: *Chemical Geology*, v. 236, p. 75–91, doi:10.1016/j.chemgeo.2006.09.002.
- Yonkee, W.A., and Weil, A.B., 2015, Tectonic evolution of the Sevier and Laramide belts within the North American Cordillera orogenic system: *Earth-Science Reviews*, v. 150, p. 531–593, doi:10.1016/j.earscirev.2015.08.001.

Zhong, S., Zhang, N., Li, Z.-X., and Roberts, J.H., 2007, Supercontinent cycles, true polar wander, and very long-wavelength mantle convection: *Earth and Planetary Science Letters*, v. 261, p. 551–564, doi:10.1016/j.epsl.2007.07.049.

APPENDICES

APPENDIX A

DETAILED METHODOLOGY AND ANALYTICAL PARAMETERS

Appendix A1: Zircon U-Pb

Zircon crystals were extracted from samples by crushing and grinding, followed by separation with a Wilfley table, heavy liquids, and a Frantz magnetic separator. Zircon grains were retained in the final heavy mineral fraction. A split of zircon grains from sample GP-01 was selected and incorporated into a 1" epoxy mount together with fragments of Sri Lanka standard zircon. Approximately 20 μm of epoxy was sanded from the pucks to expose zircon grains. The pucks were then polished. Backscattered electron images of mounts were generated with a Hitachi 3400N scanning electron microscope at the Arizona LaserChron Center at the University of Arizona (Tucson, Arizona, USA) to provide a guide for locating analysis pits in the core of zircon crystals free from excessive zoning or radiation damage.

Zircons were ablated with a Photon Machines Analyte G2 Excimer laser using a spot diameter of 30 microns (Gehrels et al., 2008; Gehrels and Pecha, 2014; Pullen et al., 2018). The ablated material was carried in helium into the plasma source of a Nu HR ICP-MS, which is equipped with a flight tube of sufficient width that U, Th, and Pb isotopes are measured simultaneously. All measurements were made in static mode, using Faraday detectors with 3×10^{11} ohm resistors for ^{238}U , ^{232}Th , ^{208}Pb - ^{206}Pb , and discrete dynode ion counters for ^{204}Pb and ^{202}Hg . Each analysis consists of one 15-second integration on peaks with the laser off (for backgrounds), 15 one-second integrations with the laser firing, and a 30 second delay to purge the previous sample and prepare for the next analysis.

Following analysis, data reduction was performed with a MATLAB-based routine (AgeCalcML; Sundell et al., 2021). ^{204}Hg was subtracted from the 204 signal to yield ^{204}Pb intensity. Common Pb was corrected for using the measured $^{206}\text{Pb}/^{204}\text{Pb}$ and the assumed com-

position of common Pb (Stacey and Kramers, 1975). In-run fractionation was corrected for by comparing measured and known ratios for zircon standards then applying these fractionation factors to unknowns using a sliding-window average. Internal uncertainty was calculated by summing $^{206}\text{Pb}/^{238}\text{U}$, $^{208}\text{Pb}/^{232}\text{Th}$, $^{206}\text{Pb}/^{207}\text{Pb}$, and $^{206}\text{Pb}/^{204}\text{Pb}$ measurement uncertainties with any overdispersion factor. External uncertainty was calculated from the scatter of standard analyses, uncertainty in ages of standards, uncertainties in the composition of common Pb, and uncertainties in the decay constants for ^{235}U and ^{238}U . Ages < 900 Ma reflect the 206/238 ratio while ages > 900 Ma reflect the 206/207 ratio.

Appendix A1.1: Detailed analytical settings for U-Pb geochronology

Laboratory and Sample Preparation

Laboratory name	Arizona LaserChron Center
Sample type/mineral	Zircon
Sample preparation	Conventional mineral separation, 1-inch epoxy mount, polished to 1-micron finish
Imaging	Hitachi 3400N SEM with BSE

Laser ablation system

Make, Model, and type	Photon Machines Analyte G2 Excimer laser
Ablation cell and volume	HelEx ablation cell
Laser wavelength	193 nm
Pulse width	~8 ns
Energy density	~7 J/cm ²
Repetition rate	7 Hz
Ablation duration	10 s
Ablation pit depth/ablation rate	~12 microns & 0.8 microns/sec
Spot diameter nominal/actual	20 microns
Sampling mode/pattern	Spot
Carrier gas	Helium
Cell carrier gas flow	0.11 L/min He in inner cup, 0.29 L/min He in cell

ICP-MS instrument	
Make, Model, and type	Thermo Element2 HR ICP-MS
Sample introduction	Ablation aerosol
RF power	1200 W
Make-up gas flow	0.8 L/min Ar
Detection system	Dual mode Secondary Electron Multiplier
Masses measured	202Hg, 204(Hg+Pb), 206Pb, 207Pb, 208Pb, 232Th, 235U, 238U
Dwell times (ms)	202=5.2, 204=7.8, 206=20.2, 207=28.4, 208=2.6, 232=2.6, 235=15.4, 238=10.4
Total integration time per output data point (sec)	202=1.5, 204=2.3, 206=5.9, 207=8.3, 208=0.7, 232=0.7, 235=4.5, 238=3.0
Sensitivity as useful yield	~5000 cps/ppm
IC dead time	22 ns
Data processing	
Gas blank	8 sec on-peak zero subtracted
Calibration strategy	SLM zircon used as primary standard
Reference material information	Gehrels et al. (2008)
Data processing package used/Correction for LIEF	E2agecalc
Mass discrimination	Normalized to primary standard
Common Pb correction, composition, and uncertainty	Common Pb correction based on measured 206Pb/204 Pb and the assumed composition of common Pb based on Stacey and Kramers (1975)
Uncertainty level and propagation	Uncertainties for individual analyses propagated at 1-sigma. Uncertainty of pooled analyses propagated at 2-sigma.
Quality control/validation	FC-1 and R33 analyzed as secondary standards.
Concordance Cut-offs	< 20% discordance, < 5% reverse discordance
Other information	
Primary and secondary standards mounted together with unknowns.	

Appendix A2: Biotite $^{40}\text{Ar}/^{39}\text{Ar}$

Mica samples were handpicked from purified mineral separated and then co-irradiated at the Oregon State TRIGA Reactor for 14 hours with the Fish Canyon Tuff sanidine monitor

standard (28.201 $\pm$ 0.046 Ma, Kuiper et al., 2008; using decay constants of Min et al., 2000). The samples and standards were loaded together in adjacent pits at the same depth in a 1.9 cm diameter, 0.3 cm depth Al disk (Renne et al., 1998). Stacked disks were wrapped in Al foil and placed in a cadmium-shielded irradiation tube at the TRIGA facility.

The samples and standards were removed from the disk after irradiation and placed in 2 mm diameter pits in a machined Ti sample holder under ultra-high vacuum in a chamber made of stainless steel with a differentially-pumped ZnS viewport. Each aliquot was sequentially heated using an automated Photon Machines CO₂ laser (10.6 μ m wavelength) with a lens positioned to provide a uniform 3 mm diameter beam that completely covers the sample. Heating power was increased to produce gas release spectra until the sample was fully melted and degassed; each heating step was 30 s at full power after 15 s ramp. Gases evolved during heating were exposed to a hot ($\sim$ 400 $^{\circ}$ C) SAES St 101 Zr-Al non-evaporable getter for three minutes.

Purified Ar was released from the getter chamber into the mass spectrometer for 30 seconds, during which the signal was monitored on all detectors. The three stable and two long-lived Ar isotopes (^{36}Ar , ^{37}Ar , ^{38}Ar , ^{39}Ar , ^{40}Ar) were measured on an Isotopx NGX multicollector mass spectrometer with the ion source set to 6000 V accelerating voltage and 200 μ A trap current and tuned for good sensitivity and best isotope ratio linearity. Samples were heated with calibrated air standards and procedural blanks between every three heating steps to account for mass discrimination and instrument backgrounds. The isotopes ^{37}Ar , ^{38}Ar , ^{39}Ar , and ^{40}Ar were measured on ATONA Faraday detectors while the isotope ^{36}Ar was measured on a Hamamatsu ion counting multiplier modified by Isotopx for the NGX. Sample isotope intensities were determined using background correction from the procedural blanks and a correction to the

measured $^{40}\text{Ar}/^{36}\text{Ar}$ ratio using the air standards. The $^{40}\text{Ar}/^{39}\text{Ar}$ ratios for age determination were then calculated using the nuclear interference corrections of (Renne et al., 1998) and a J value calculated using the Kuiper et al. (2008) age of 28.201 $\pm$ 0.046 Ma for the Fish Canyon Tuff sanidine monitor standard.

Plateau ages were defined as three or more contiguous steps corresponding to a minimum of 50% of the ^{39}Ar released and showing no statistically significant slope (Fleck et al., 1977). Using these criteria, plateau ages were obtained for two samples (natural plateau). In some cases, the statistical spread of the individual steps exceeded the individual uncertainties—likely due to the high radiogenic argon concentration ($\geq 99.7\%$) in our Precambrian biotites and resultant high levels of analytic precision—but plateaus meeting the criteria of more than 50% of the ^{39}Ar and no significant slope were apparent (see Appendix A2 for $^{40}\text{Ar}/^{39}\text{Ar}$ gas release spectra). In these cases, we imposed a plateau calculation (forced plateau) and assigned an uncertainty using the reduced chi-squared statistic, equivalent to the standard error multiplied by the square root of the MSWD (mean square weighted deviation; Schoene et al., 2013) in order to conservatively account for the uncertainty of the age determination. Details of the age determinations, and whether the plateaus are natural or forced, are shown in Table 1.

Appendix A3: Zircon (U-Th)/He

Zircon grains were separated from samples using the traditional density and magnetic separation methods described in Appendix 1.1. Aliquots were prepared at the TeST lab at Montana State University, as described in the methods section above. (U-Th)/He analytical methods followed those described in Guenthner et al. (2016). $^4\text{He}/^3\text{He}$ ratios of individual crystal-bearing tubes were measured via in vacuo diode laser heating, cryogenic purification, and

quadrupole mass-spectrometry on a Pfeiffer PrismaPlus to (e.g., Zeitler and Tremblay, 2020 and references therein). Degassed crystals were then retrieved, spiked with calibrated ^{235}U - ^{230}Th - ^{145}Nd tracer, and dissolved in pressurized, high temperature HF-HCl acid-vapor dissolution vessels. U and Th isotopic concentrations were then measured on a Thermo iCAP-Q ICP-MS (e.g., House et al., 2000). Radiogenic ^4He atoms travel up to 20 μm during α -decay, requiring a correction be applied based on crystal size and geometry to account for He lost to ejection outside of the crystal (Farley et al., 1996). The alpha ejection corrections equations of Hourigan et al. (2005)—using measured grain sizes and idealized crystal morphologies—were applied to correct ZHe dates for this effect.

APPENDIX B

DETAILED ANALYTICAL RESULTS

Appendix B1: U-Pb data from sample GP-01.

The ages of analyses not meeting concordance cutoffs (<10% discordance, <5% reverse discordance) are highlighted in orange.

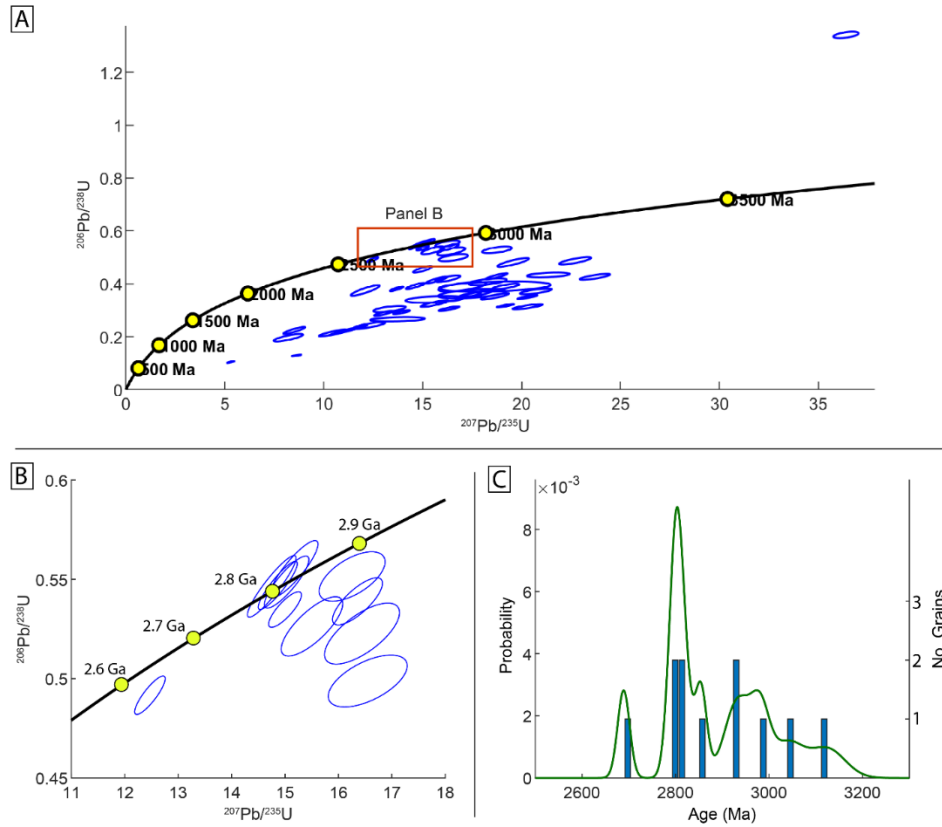
Spot	$^{206}\text{Pb}/^{207}\text{Pb}$	$\pm(\%)$	$^{207}\text{Pb}/^{235}\text{U}$	$\pm(\%)$	$^{206}\text{Pb}/^{238}\text{U}$	$\pm(\%)$	Best Age (Ma)	$\pm$ (Ma)
1	5.07	0.71	14.86	1.64	0.55	1.48	2802.40	11.63
2	5.04	0.96	15.17	1.97	0.56	1.73	2811.07	15.65
3	4.27	1.87	12.10	4.03	0.37	3.57	3080.26	29.85
4	3.39	1.39	19.57	2.72	0.48	2.34	3444.65	21.64
5	3.15	3.19	13.32	4.11	0.30	2.60	3556.59	49.06
6	5.43	0.78	12.47	1.53	0.49	1.31	2688.97	12.87
7	3.30	3.59	8.21	6.02	0.20	4.83	3487.28	55.61
8	5.03	0.94	15.03	1.85	0.55	1.60	2815.40	15.29
9	2.71	2.56	12.29	4.42	0.24	3.61	3784.80	38.74
10	3.02	5.77	15.59	6.27	0.34	2.46	3623.39	88.51
11	4.38	2.07	16.44	2.85	0.52	1.96	3038.98	33.11
12	2.60	1.22	18.66	2.56	0.35	2.25	3846.56	18.42
13	2.47	1.75	20.72	2.59	0.37	1.91	3927.21	26.31
14	2.58	1.09	16.34	2.13	0.31	1.83	3861.55	16.46
15	2.84	1.32	17.40	2.10	0.36	1.63	3715.41	20.04
16	4.15	2.41	16.55	2.98	0.50	1.75	3124.01	38.38
17	2.87	1.49	15.59	2.96	0.32	2.56	3698.62	22.72
18	3.64	1.03	15.70	2.27	0.41	2.02	3332.67	16.12
19	4.18	1.13	14.99	2.12	0.45	1.80	3115.40	17.98
20	5.11	0.80	14.75	2.06	0.55	1.90	2791.36	13.15
21	3.88	2.18	18.75	2.70	0.53	1.58	3231.56	34.45
22	2.82	2.05	10.45	4.48	0.21	3.98	3724.47	31.17
23	4.68	1.60	15.50	2.48	0.53	1.89	2933.46	25.86
24	4.54	1.22	16.36	1.91	0.54	1.47	2981.58	19.60
25	2.38	0.79	20.29	1.66	0.35	1.46	3983.75	11.74
26	2.90	1.11	13.93	2.04	0.29	1.71	3685.93	16.87
27	2.48	1.30	23.70	2.16	0.43	1.73	3920.93	19.55

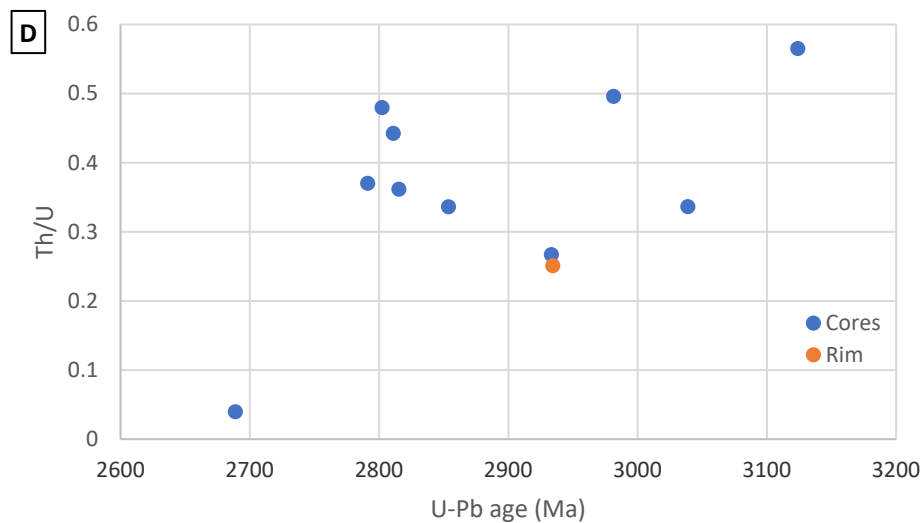
28	2.67	6.41	13.72	6.74	0.27	2.08	3808.04	97.12
29	3.15	3.56	17.00	4.71	0.39	3.09	3555.14	54.82
30	3.08	2.84	18.28	3.53	0.41	2.10	3593.41	43.63
31	2.07	0.91	8.63	1.87	0.13	1.64	4191.79	13.46
32	2.28	0.68	19.18	1.53	0.32	1.37	4047.64	10.12
33	2.86	5.67	17.50	6.53	0.36	3.25	3707.45	86.45
34	4.91	0.78	15.00	1.36	0.53	1.12	2853.85	12.64
35	3.62	1.87	8.51	4.30	0.22	3.88	3339.78	29.18
36	2.70	1.07	5.30	2.35	0.10	2.09	3793.16	16.24
37	3.55	1.66	16.27	2.59	0.42	1.99	3370.65	25.88
38	2.75	0.97	11.24	5.77	0.22	5.69	3763.62	14.79
39	3.05	1.24	13.02	2.45	0.29	2.12	3605.95	18.98
40	2.80	3.07	21.35	3.37	0.43	1.39	3738.13	46.68
41	2.80	1.51	19.20	2.34	0.39	1.79	3738.51	22.98
42	3.69	1.04	14.63	2.19	0.39	1.93	3310.97	16.33
43	5.08	1.00	36.39	1.17	1.34	0.60	2799.59	16.37
44	2.88	9.56	18.65	10.10	0.39	3.28	3694.29	146.10
45	2.13	1.37	20.30	2.49	0.31	2.08	4150.74	20.37
46	2.81	1.47	17.74	3.25	0.36	2.90	3733.24	22.38
47	2.96	1.43	22.70	2.38	0.49	1.90	3653.72	21.91
48	3.84	0.78	13.68	1.69	0.38	1.50	3249.84	12.31
49	4.68	2.04	16.26	2.53	0.55	1.51	2934.57	32.92
50	3.15	0.84	18.72	2.02	0.43	1.84	3555.15	12.95

Appendix B2

U-Pb ages from the summit of Granite Peak yield Neo–Mesoarchean ages. (A) Wetherill concordia diagram (Wetherill, 1956) showing all analyses (including those not meeting concordance cutoffs). (B) Wetherill concordia diagram, showing analyses meeting concordance cutoffs. Analyses intersecting concordia plot at ca. 2800 Ma. (C) Probability density plot and histogram showing distribution of zircon U-Pb ages. The highest probability peak is centered at

ca. 2804 Ma while the distribution has a positive skew toward older ages. Plots were generated using AgeCalcML (Sundell et al., 2021). (D) Zircon U-Pb ages plotted against Th/U ratio, a proxy for metamorphic vs. igneous zircon. No obvious correlation is present.





Appendix B3

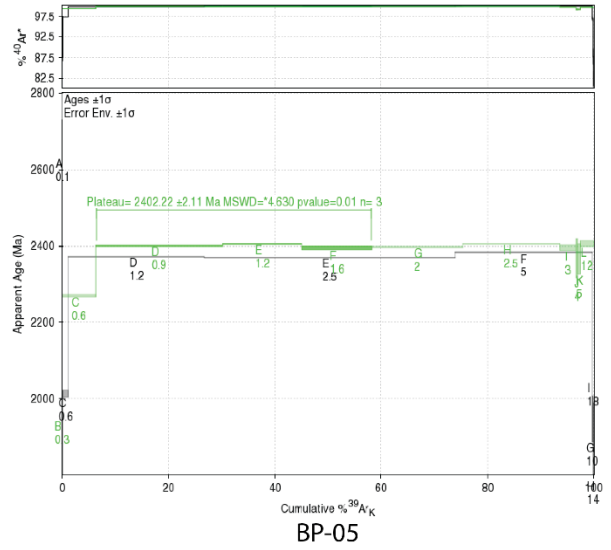
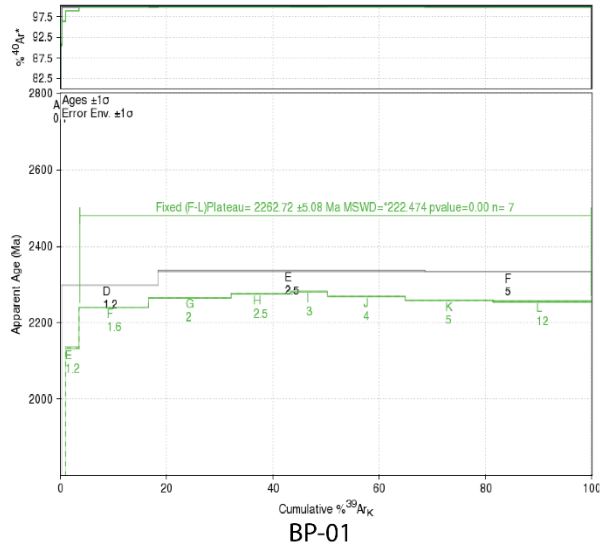
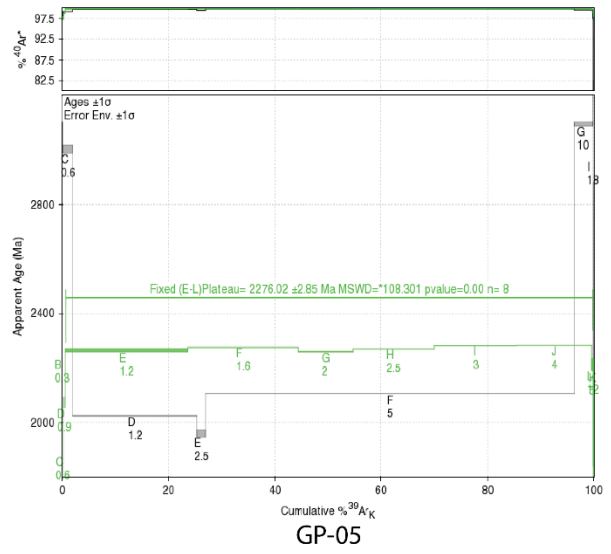
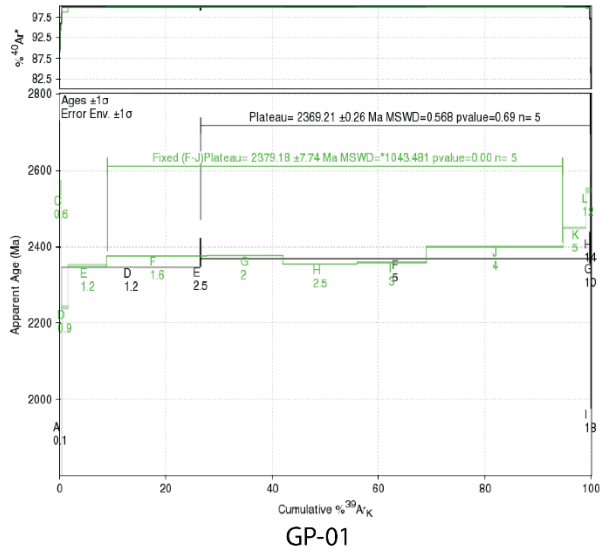
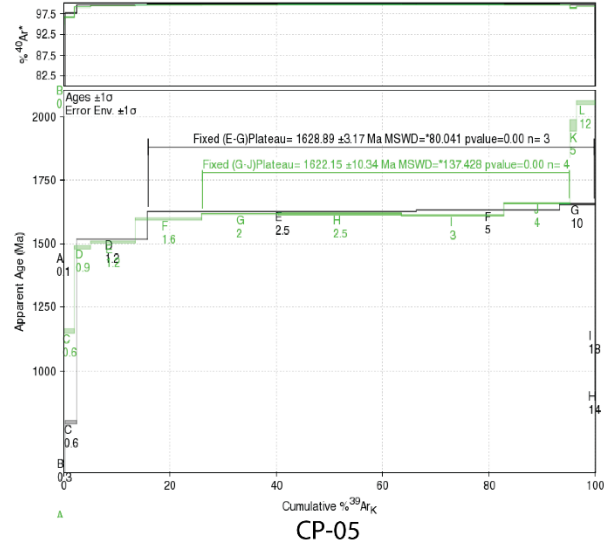
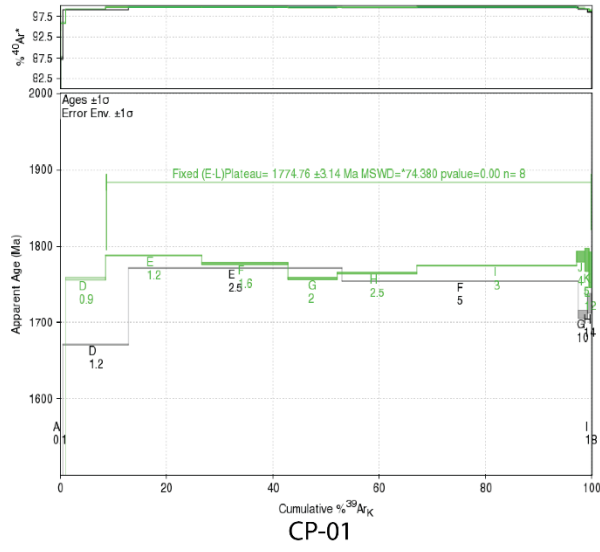
Biotite $^{40}\text{Ar}/^{39}\text{Ar}$ data from single-step diffusion experiments reported in Table 1, included to demonstrate general consistency with multi-step heating results. * denotes radiogenic ^{40}Ar .

Sample	Ca/K	Cl/K	Mol ^{39}Ar	% $^{40}\text{Ar}^*$	Age (Ma)	$\pm$ Age
CP-01	0.01214	-0.00132	0.63	99.9	1663	3
CP-05	0.09962	-0.00044	1.121	99.7	1797	3
GP-01	-0.00113	-0.00223	0.389	100	2387	2
GP-05	0.00271	0.0022	0.972	100	2473	3
BP-01	0.02013	0.00443	0.401	99.9	2278	3
BP-05	0.01651	0.00271	0.849	99.9	2355	3

Appendix B4

Biotite $^{40}\text{Ar}/^{39}\text{Ar}$ gas release spectra from step-heating experiments. Grey lines represent initial analyses conducted to calibrate the NGX mass spectrometer. Green lines represent final gas

release spectra for our samples, with natural or forced plateaus indicated. The $^{40}\text{Ar}/^{39}\text{Ar}$ ratio indicated by the inset at top approaches 1 after initial heating steps, reflecting the large amount of radiogenic argon due to the Paleoproterozoic cooling age of our samples.



Appendix B5: Zircon (U-Th)/He results.

Sample (Coordinates; Datum = WGS84)	Elevation (m)	Aliquot	Rs (μm)	eU (ppm)	Corrected Date (Ma)	$\pm 2\sigma$ (Ma)
CP-01 (45.3125, -110.0751)	3403	CP-01-z1	53.79	333.23	390.51	5.61
		CP-01-z2	49.38	165.72	542.08	8.22
		CP-01-z3	45.34	1274.00	23.24	0.33
		CP-01-z4	59.24	288.84	233.40	2.89
		CP-01-z5	39.94	312.03	222.99	3.18
		CP-01-z6	57.59	430.72	87.14	1.14
		CP-01-z7	44.55	428.06	141.86	2.02
CP-02 (45.3220, -110.0843)	3167	CP-02-z7	42.57	1060.16	45.38	0.63
CP-03 (45.3385, -110.0755)	2948	CP-03-z1	47.68	1127.13	29.26	0.46
		CP-03-z2	42.67	3565.28	35.38	0.50
		CP-03-z3	51.82	926.37	35.25	0.54
		CP-03-z4	54.41	1055.14	46.24	0.62
		CP-03-z5	49.52	937.47	64.88	0.92
		CP-03-z6	57.33	660.50	198.59	2.84
		CP-03-z7	59.60	392.08	205.07	2.92
CP-04 (45.3456, -110.0660)	2721	CP-04-z1	47.12	266.08	450.21	6.31
		CP-04-z2	61.99	677.65	130.41	1.67
		CP-04-z3	52.02	544.65	156.60	2.21
		CP-04-z4	48.38	403.24	207.48	2.90
		CP-04-z5	49.09	555.51	133.82	1.83
		CP-04-z6	56.96	631.13	109.57	1.39
		CP-04-z7	49.82	534.82	82.60	1.17
CP-05 (45.3551, -110.0597)	2520	CP-05-z1	41.86	466.51	83.22	1.23
		CP-05-z2	54.56	505.49	60.31	0.94
		CP-05-z3	53.06	1146.38	45.06	0.64
		CP-05-z4	45.17	2485.68	28.28	0.40
		CP-05-z5	53.50	762.93	43.50	0.59
		CP-05-z6	57.13	1975.38	14.27	0.23
		CP-05-z7	49.26	810.23	56.49	0.73
GP-01 (45.1634, -109.8075)	3909	GP-01-z1	47.71	295.38	488.61	7.65
		GP-01-z2	52.13	501.05	279.56	4.34
		GP-01-z3	55.60	375.39	430.31	6.74
		GP-01-z5	46.90	519.90	288.95	4.61

		GP-01-z6	64.36	311.19	435.41	6.41
		GP-01-z7	62.78	337.84	378.33	5.75
GP-03 (45.2129, -109.7484)	3088	GP-03-z1	44.74	273.12	397.06	4.71
		GP-03-z2	45.09	420.59	257.85	3.33
		GP-03-z3	48.63	345.36	332.41	4.67
		GP-03-z4	50.09	378.21	312.47	4.11
		GP-03-z5	41.96	513.32	241.48	2.60
		GP-03-z6	45.11	870.01	86.82	1.28
		GP-03-z7	47.13	650.70	186.03	2.96
GP-04 (45.2160, -109.7659)	2510	GP-04-z7	38.34	331.52	160.77	2.22
GP-05 (45.2384, -109.7384)	2090	GP-05-z1	43.18	323.58	270.51	5.09
		GP-05-z2	56.17	339.35	155.98	2.97
		GP-05-z3	42.28	1625.97	44.08	0.90
		GP-05-z4	47.19	192.03	450.16	8.44
		GP-05-z5	52.26	325.16	172.69	3.36
		GP-05-z6	64.74	118.56	393.44	7.80
		GP-05-z7	56.91	545.63	159.17	2.12
BP-01 (45.2431, -109.5796)	1742	BP-01-z1	39.10	331.32	215.02	4.17
		BP-01-z2	60.82	325.82	331.42	6.07
		BP-01-z3	45.44	488.66	193.24	3.85
		BP-01-z4	48.67	544.26	110.97	2.24
		BP-01-z5	47.84	1139.99	118.20	2.35
		BP-01-z6	50.10	438.25	197.90	3.43
		BP-01-z7	65.45	741.14	139.10	2.83
BP-02 (45.1684, -109.6509)	2018	BP-02-z1	44.78	152.13	534.77	9.17
		BP-02-z2	48.48	697.10	63.22	1.08
		BP-02-z3	42.34	1059.66	107.91	1.74
		BP-02-z4	57.86	948.59	97.36	1.70
		BP-02-z5	49.26	400.43	251.92	5.19
		BP-02-z6	51.30	414.75	177.64	3.22
		BP-02-z7	65.41	261.73	267.05	5.28
BP-03 (45.1438, -109.6912)	2382	BP-03-z1	43.43	267.30	297.74	5.74
		BP-03-z2	38.31	414.94	260.86	4.87
		BP-03-z3	43.96	503.25	160.39	2.85
		BP-03-z4	48.41	587.59	141.37	2.44
		BP-03-z5	59.77	398.30	306.27	5.58
		BP-03-z6	44.51	402.95	337.72	6.17
		BP-03-z7	46.92	269.37	340.16	6.02
	2750	BP-04-z1	51.32	195.09	673.83	11.61
		BP-04-z2	65.61	350.61	472.68	7.99

BP-04 (45.1185, -109.7412)		BP-04-z3	62.37	285.58	423.00	7.23
		BP-04-z4	54.44	113.69	686.43	11.88
		BP-04-z5	58.85	581.85	161.19	2.88
		BP-04-z7	45.59	371.54	556.47	10.42
BP-05 (45.0954, -109.7987)	3064	BP-05-z1	38.53	383.70	194.75	3.45
		BP-05-z2	53.94	1293.78	13.54	0.25
		BP-05-z3	59.07	609.30	143.83	2.57
		BP-05-z4	47.97	826.86	77.15	1.37
		BP-05-z5	56.02	645.03	189.07	3.21
		BP-05-z6	46.95	612.58	213.43	3.94
		BP-05-z7	41.60	817.61	32.88	0.52

APPENDIX C

DETAILED MODEL CONSTRAINTS AND JUSTIFICATION

Appendix C1: Inverse modelling simulations, samples, and data treatment

Samples: Upper elevation group: BP-03, BP-04, BP-05, CP-01, CP-02, CP-03, CP-04, CP-05, GP-01, GP-03, GP-04 (N = 11; n = 63); Lower elevation group: BP-01, BP-02, GP-05 (N = 3; n = 21).

Data exclusion: No grains excluded.

Treatment: Samples binned into groups defining similar portions of date-eU curve, shown in Appendix 3.2.

He dates (Ma): Mean uncorrected He date of each bin. α -ejection corrected in HeFTy using Ketcham et al. (2011).

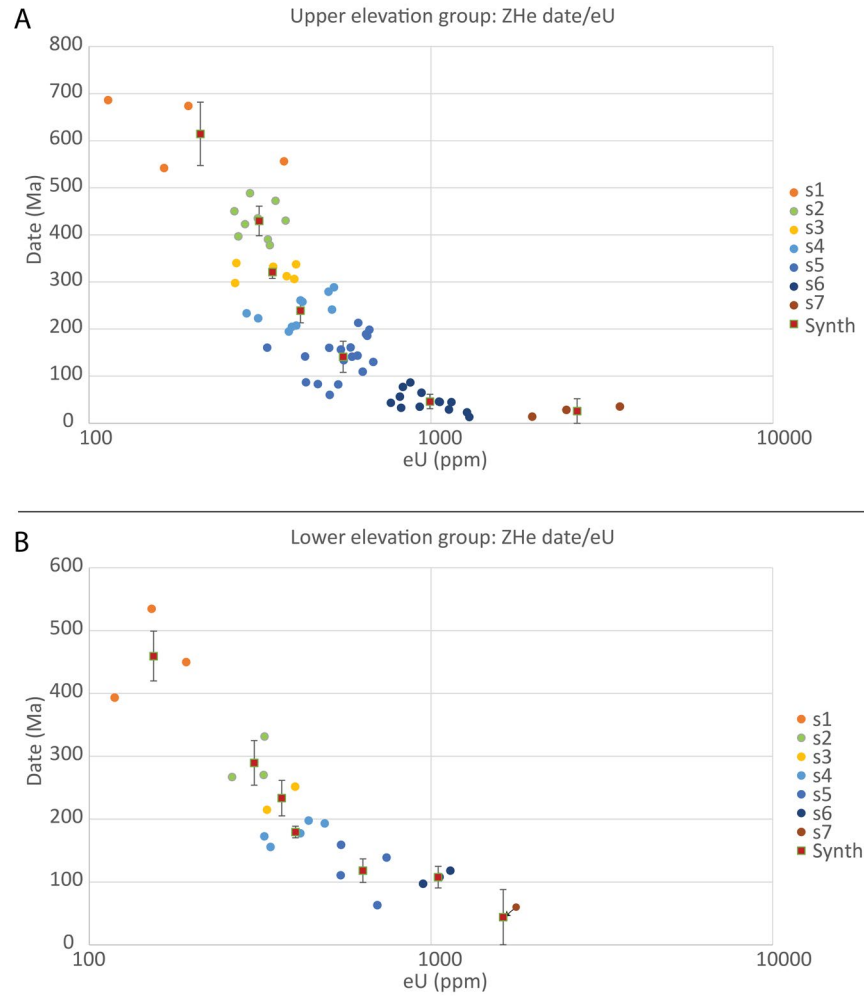
Error (Ma) applied in modeling: The 1-sigma standard deviation of dates of each bin were applied, except for highest eU bins where the error was modified to include a date of 0 Ma (as detailed in text and Appendix 4.3).

R_s (μ m): Mean equivalent spherical radius of each bin.

eU (ppm): Mean eU of each bin.

Appendix C2: Synthetic ZHe grain input

Observed ZHe date-eU defining inverse model input “synthetic grains” for the upper (A) and lower (B) elevation sample groups. Circles represent observed ZHe data, colored according to their bin. Squares represent resultant “synthetic grains” after averaging approach described in Appendix 3.1.



Appendix C3: Additional geologic information used to
impose t - T constraints on inverse thermal models

#	Time (Ma)	Temp (°C)	Reason
1. Crystallization	2900–2800	600–900	Zircon LA-ICPMS U-Pb data for main suite in Beartooth Mountains constrains crystallization to ca. 2900–2900 Ma (Mueller et al., 2008 and references therein; Carrapa et al., 2019; this study).
2. Archean– early	2402–2263	300–425	Biotite $^{40}\text{Ar}/^{39}\text{Ar}$ ages of our eastern transects constrain Archean–early Paleoproterozoic cooling. This is also consistent with K-Ar and Rb-Sr analyses of Gast et al. (1958).

Paleoproterozoic cooling			Temperatures are consistent with a conservative biotite $^{40}\text{Ar}/^{39}\text{Ar}$ closure temperature window (e.g., Grove and Harrison, 1996).
3. Paleoproterozoic cooling	1775–1622	300–425	Biotite $^{40}\text{Ar}/^{39}\text{Ar}$ ages of our western transect constrain later Paleoproterozoic cooling. Temperatures are consistent with a conservative biotite $^{40}\text{Ar}/^{39}\text{Ar}$ closure temperature window (e.g., Grove and Harrison, 1996).
4. Proterozoic cooling exploration field	1622–520	0–50	Encourages the model to explore, but does not require, cooling to near-surface conditions prior to deposition of the Flathead Sandstone.
5. Great Unconformity	550–520	0–50	Deposition of the Flathead Sandstone on the Great Unconformity surface requires Precambrian rocks of the Beartooth Mountains to have been exposed at the surface during this time. This age range is consistent with Rb-Sr dates (Chaudhuri and Brookins, 1969) and paleopedological evidence (Retallack, 2013) as well as younger depositional ages indicated by faunal evidence (Deiss, 1939; Norris and Price, 1966; Thomas, 2007). Temperatures are consistent with residence at near-surface temperatures, as suggested by minimal (< 0.5 km) of basement overburden interpreted by Wise (2000) and others discussed above. Our refined forward models explore hotter temperatures during this time interval.
6. Paleozoic reheating	520–60	Upper samples: 0–180 Lower samples: 0–230	Encourages the model to explore Paleozoic burial consistent with nearby preserved stratal thicknesses (Maughan, 1993; Lopez, 2001 and references therein), as indicated by our decompacted sediment accumulate models (Figure 8). Temperatures inclusive of 0 °C reflect uncertainty around sediment accumulation at the location of the Beartooth Mountains due to the paucity of preserved stratigraphy; however, temperature of > 70 °C are required near the Beartooth Highway where reset AHe dates are documented (Peyton et al., 2012). Temperature offset between samples allows lower elevation samples to be more deeply buried, honoring the difference in paleo-basement overburden.
7. Paleocene cooling	60–56	0–50	The unroofing sequence recorded in the Beartooth conglomerate, including basement clasts, requires rocks of the Beartooth Mountains to be exposed at the surface at this time; this formation has been dated as late Paleocene (DeCelles et al., 1991a and references therein). Temperatures are consistent with residence at near-surface temperatures, as suggested by minimal (< 0.5 km) of basement overburden interpreted by Wise (2000) and others discussed above. Our refined forward models explore hotter temperatures during this time interval.
8. Cenozoic reburial exploration field	56–0.5 Ma	0–80	Encourages the model to explore, but does not require, Cenozoic reburial consistent with previous AFT model results (Giegengack et al., 1986; Omar et al., 1994; Carrapa et al., 2019).
9. Present	0	0–20	Samples collected at surface in 2020.

Appendix C4: System- and model-specific parameters for inverse modelling

He kinetic model: ZRDAAM (Guenthner et al., 2013)

Statistical fitting criteria: GOF values > 0.5 for “good fits; > 0.05 for “acceptable fits”. Good-fits must also have a minimum GOF of $1/(N+1)$ where N is the number of statistics used (Ketcham, 2005a).

Modeling Code: HeFTy v. 1.9.3

Number of t-T paths attempted: 50,000 per sample elevation group (100,000 cumulative).

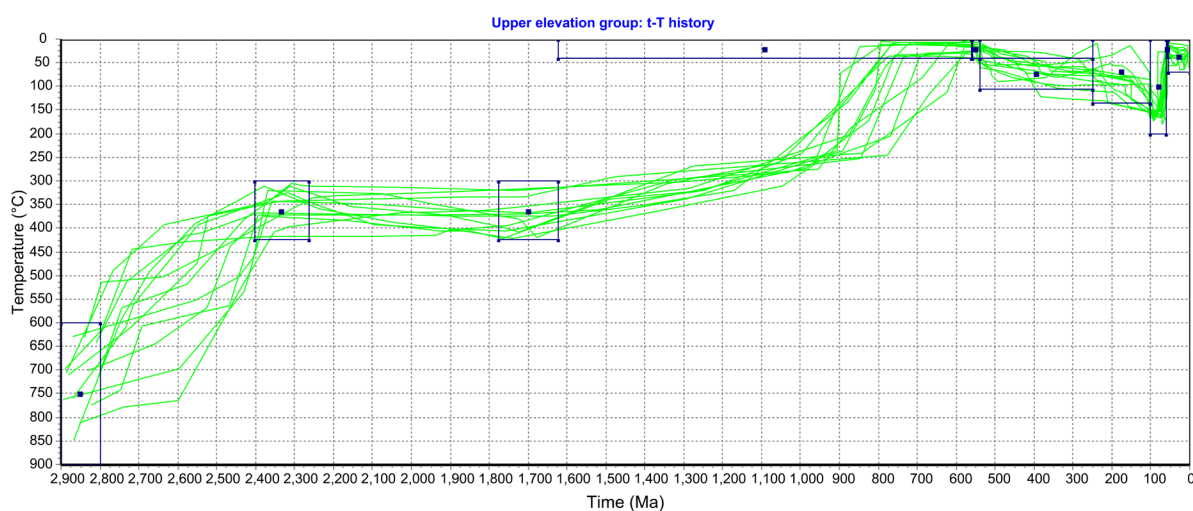
t-T path characteristics: Cooling between constraints 1 to 6 uses monotonic variable intermediate cooling (halve 2 times) to simulate erosional cooling and reheating by burial in a passive margin setting. Cooling between constraints 6 to 9 uses monotonic variable episodic (halve 2 times) to allow more rapid exhumation associated with orogenesis and rapid burial by volcanics.

APPENDIX D

SUPPLEMENTARY THERMAL MODEL RESULTS

Appendix D1: Models Forcing Consistency with Decompacted Sedimentary Thicknesses

Inverse thermal model for our upper elevation sample group with constraints enforcing Phanerozoic burial similar to that predicted by our decompacted sediment accumulation models (Figure 8). This model found 13 acceptable-fits but no good-fits. GOF values are 0.13–0.55.

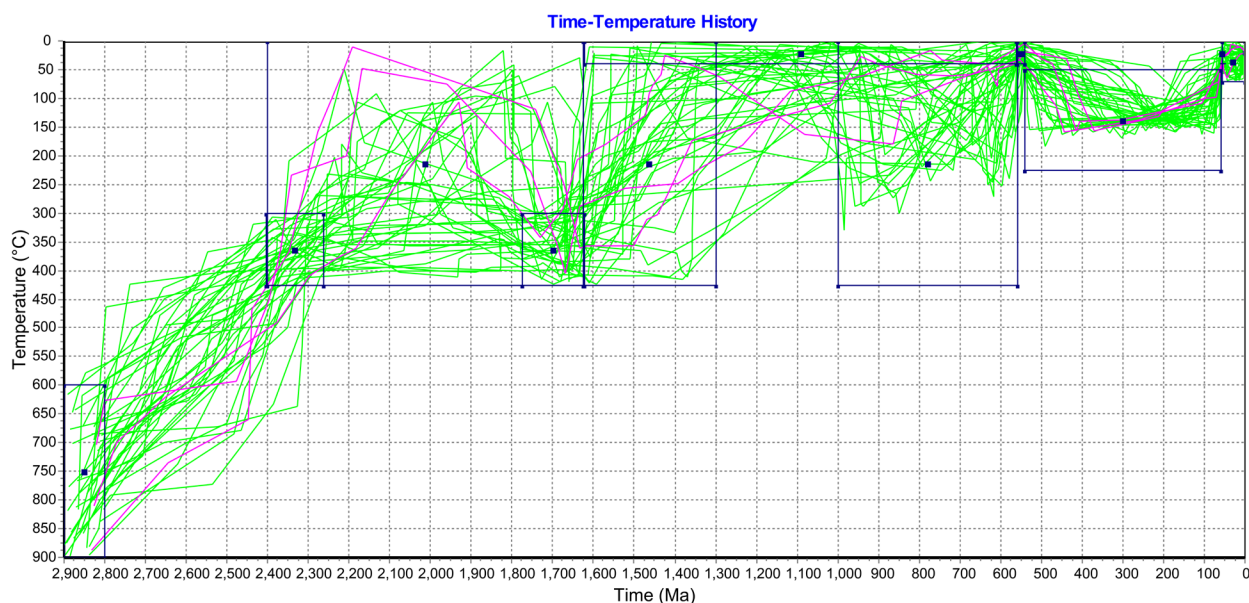


Appendix D2: Models Exploring Proterozoic Reheating

Inverse thermal model for our upper elevation group with additional exploration boxes encouraging model to explore: (1) Paleoproterozoic cooling and reheating; (2) Mesoproterozoic reheating; and (3) Neoproterozoic reheating. This model illustrates the lack of sensitivity to the Paleoproterozoic thermal history. Even with Mesoproterozoic–Neoproterozoic reburial encouraged, the majority of good-fit paths have broadly similar characteristics to those found without these additional constraint boxes (Figure 12); 2/3 good-fit paths cool to near-surface

temperatures by ca. 800 Ma and experiencing minor subsequent Proterozoic reburial.

Additionally, all good-fit paths incorporate Phanerozoic histories similar to our models without these additional exploration boxes (Figure 12). We note that there is no independent geologic evidence for significant Proterozoic reburial at the location of the Beartooth Mountains, so the bulk of our modeling does not encourage exploration of these histories.



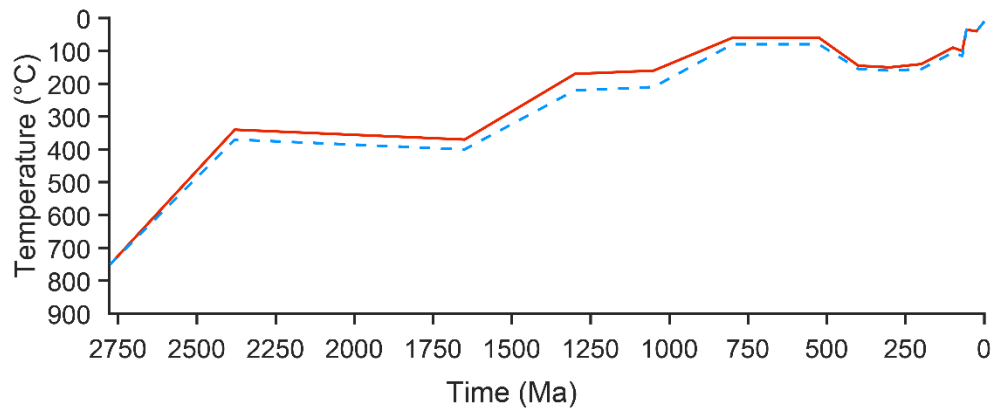
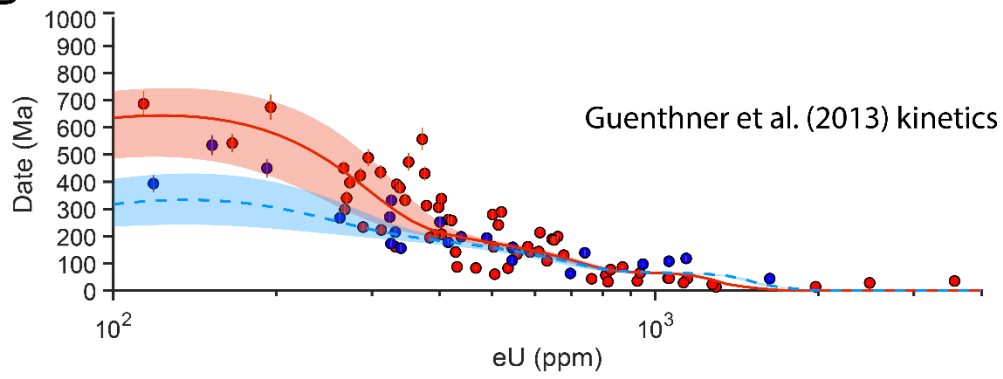
Appendix D3: Unsuccessful Models

We attempted 100,000 path model runs with the uncertainty of our highest eU grains not inclusive of zero, but instead comprising the standard deviation of constituent analyses (as the rest of our synthetic grain inputs; upper sample group = 7.2 Myr) or analytic error (lower sample group, where the highest eU model input grain was comprised of only one analysis with an uncertainty of 0.64 Myr). These models with found no acceptable- or good-fits. This is likely due

to the poor present understanding of helium diffusion in very damaged zircon grains (e.g., Guenthner et al., 2013) and consequently, the inability of thermal models to accurately predict observed ZHe dates in grains with high-eU, as discussed in the text. Consequently, we extended the uncertainty of these highest eU model input grains to include a ZHe date of zero.

Appendix D4: Comparison of Zircon Radiation Damage Annealing Kinetics

Comparison of date-eU correlations predicted from our refined t - T history in Figure 13A using the kinetics of Guenthner et al. (2013) versus those of Ginster et al. (2019), as implemented by Guenthner (2021). The model incorporating kinetics of Ginster et al. (2019) is a poor fit to the data, especially at high-eU, where it predicts zero dates at > 800 ppm eU. This is likely due to the greater temperatures required for annealing of radiation damage with these kinetics and the inability for very damaged grains to fully recover, as illustrated by Guenthner (2021). Because of the large portion of grains in our dataset with relatively high-eU (> 800 ppm), we implement the kinetics of Guenthner et al. (2013)—which have been shown to predict better fits to observed data in some datasets (Guenthner, 2021)—in the bulk of our modeling.

ARefined t - T paths**B****C**