A Thermochronometric, Microtextural, and Numerical Modeling Approach to Deciphering the Rock Record of Deformation Processes in the Wasatch and Denali Fault Zones

Robert G. McDermott
Utah State University

Follow this and additional works at: https://digitalcommons.usu.edu/etd

Part of the Earth Sciences Commons

Recommended Citation
https://digitalcommons.usu.edu/etd/7957

This Dissertation is brought to you for free and open access by the Graduate Studies at DigitalCommons@USU. It has been accepted for inclusion in All Graduate Theses and Dissertations by an authorized administrator of DigitalCommons@USU. For more information, please contact digitalcommons@usu.edu.
A THERMOCHRONOMETRIC, MICROTEXTURAL, AND NUMERICAL MODELING APPROACH TO DECIPHERING THE ROCK RECORD OF DEFORMATION PROCESSES IN THE WASATCH AND DENALI FAULT ZONES

by

Robert G. McDermott

A dissertation submitted in partial fulfillment of the requirements for the degree of DOCTOR OF PHILOSOPHY in Geosciences

Approved:

Alexis K. Ault, Ph.D. Major Professor

Jonathan Saul Caine, Ph.D. Committee Member

James P. Evans, Ph.D. Committee Member

Peter C. Lippert, Ph.D. Committee Member

Anthony R. Lowry, Ph.D. Committee Member

D. Richard Cutler, Ph.D. Interim Vice Provost for Graduate Studies

UTAH STATE UNIVERSITY Logan, Utah

2020
ABSTRACT

A Thermochronometric, Microtextural, and Numerical Modeling Approach to Deciphering the Rock Record of Deformation Processes in the Wasatch and Denali Fault Zones

by

Robert G. McDermott, Doctor of Philosophy
Utah State University, 2020

Major Professor: Dr. Alexis K. Ault
Department: Geosciences

Linked textural and thermochronometric data from exhumed fault rocks are critical to constraining the timing and style of crustal deformation, from mountain- to slip surface-scale, and elucidating physical mechanisms leading to (a)seismic slip. Hematite, common in faults, can inform fault zone processes via coupled microtextural analysis and (U-Th)/He (He) thermochronometry. I apply fault rock hematite He thermochronometry with field and microtextural observations, conventional thermochronometry of host rocks, and modeling to decipher deformation history and slip mechanisms over different spatiotemporal scales in the Wasatch and eastern Denali fault zones (WFZ, EDFZ).

Thermochronometry and microtextures document fault reactivation, strain localization, and coseismic slip on WFZ small-displacement hematite fault mirrors. Hematite He dates from veins, hematite-cemented breccia, and overlying hematite fault mirrors are ~270–21 Ma, ~69–18 Ma, and ~42–1 Ma. Apatite and zircon He and apatite fission-track (AFT) dates are ~5–3 Ma from adjacent Paleoproterozoic host rocks.
Thermal history models require episodic Paleoproterozoic through Miocene hematite precipitation and coeval ambient reheating (burial) and cooling (exhumation), collectively associated with regional tectonism. Mirrors exhibit microstructures reflecting fluidization, nanoparticle lubrication, and asperity flash heating that promote coseismic dynamic weakening. Mirror-host rock thermochronometry data patterns are consistent with He loss due to friction-generated heat at ~20 to <4.5 Ma, supported by thermomechanical models of coupled asperity flash heating and He loss. Results reveal repeated deformation over 100s of Myr, culminating in $M_w \approx -3$ to 3 earthquakes on hematite fault mirrors.

Hematite He and conventional thermochronometry are used to document fault slip in the EDFZ and the thermotectonic history of the adjacent Kluane Ranges. Apatite He, zircon He, and AFT dates range from ~26–4 Ma, ~94–28 Ma, ~110–12 Ma, and ~137–83 Ma. Thermal history inversions, geological relations, and estimated paleoelevation from palynology reveal a three-stage history linked to Cretaceous terrane accretion, Paleogene enhanced strike-slip on the EDFZ, and Oligocene-present transpression associated with flat-slab subduction. Hematite He dates from foliated microfabrics on hematite slip surfaces in the EDFZ are ~8–4 Ma. Data constrain fault damage development and aseismic slip concomitant with surface uplift and far-field plate boundary processes.

(360 pages)
A Thermochronometric, Microtextural, and Numerical Modeling Approach to Deciphering the Rock Record of Deformation Processes in the Wasatch and Denali Fault Zones

Robert G. McDermott

Fault zones are the primary features that accommodate movement of Earth’s crust, resulting in the formation of mountain belts and damaging earthquakes. Rocks modified by faulting and brought to Earth’s surface by erosion are archives of the mechanical processes involved in earthquakes and(or) aseismic creep. Thermochronometry is a radioisotopic dating system primarily sensitive to temperature and offers a means to constrain dates and rates of thermal processes. Hematite is common in fault zones, amenable to (U-Th)/He (He) thermochronometry, and exhibits distinct microtextures diagnostic of fault zone mechanics. I apply hematite He thermochronometry and microtextural analyses with a suite of other tools to interrogate the evolution of the Wasatch fault zone (WFZ), UT, USA, and the eastern Denali fault zone (EDFZ), Yukon, Canada over different scales in space and time.

Hematite He dates and microtextures from hematite-coated fault surfaces and veins from the WFZ show fault surfaces accommodated ancient seismicity. Although this seismicity is <5 Myr in age, fault surfaces formed within pre-existing hematite that is 100s of Myr older. Microtextures reveal that WFZ earthquakes were facilitated by fluidization of hematite, extreme grain size reduction and rolling between grains, and
breakdown of rough fault surfaces. Small fault surfaces in the WFZ are ultimately the product of deformation processes occurring throughout deep geologic time and at different timescales culminating in earthquakes. Low-temperature thermochronometry is also used to constrain erosion related to surface uplift of mountains adjacent to the EDFZ. Results show growth of topography and deformation along the EDFZ occurred in three stages from ~95–75 Ma, ~75–30 Ma, and ~30 Ma–present, primarily as a response to plate boundary processes >200 km away. Hematite He dates from hematite-coated fault surfaces in the EDFZ constrain hematite precipitation at ~8–4 Ma and reveal a record of faulting that contributes to mountain growth. Hematite microtextures in these samples suggest aseismic fault slip. The collective results of this dissertation highlight the spectrum of deformation of Earth’s crust from the mountain- to fault surface-scales and from 100s of Myr to seconds, as well as the different tectonic and mechanical processes responsible for this evolution.
ACKNOWLEDGMENTS

First and foremost, I need to thank my advisor Alexis Ault. I still remember our first science conversation when I was deciding on graduate schools, and the infectious excitement surrounding this thing no one else was doing—dating hematite on faults. I immediately knew I needed to get to USU and work on this project, and it has gone further than I could have imagined. I would also like to thank my committee: Jonathan Caine, Jim Evans, Pete Lippert, and Tony Lowry. I’m lucky to have had a solid network of mentors over the years, and I cannot express my gratitude to all of you enough.

During my tenure at USU, many family, friends, fellow graduate students, and office mates have been a constant source of support. I would particularly like to thank Mike, Wes, Jace, Jordan, Dave, Amy, Brady, Jerome, Gabriele, Coleman, Heather, Harriet, and all past and present Friends of Noble Gases, Temperature, Time, Textures (and sometimes stromatolites). To my mom, thank you for always encouraging me to chase after my goals and teaching me to think outside the box. To my dad, thank you for showing me what it means to work hard. To my wife, Emily, I could not have done this without you. Thank you for tolerating the long hours, your endless understanding, and the occasional adventure. To our fur children/rabbits Karen and Pip, thank you for dutifully waiting right outside my home office door for me to finish working.

Many colleagues and field assistants were essential to my success. Jordan Jensen trooped through the Kluane Ranges for three weeks, carrying pounds of rocks without a single complaint. Chris Ammon did field work with me in the middle of August, when some of the days were over 100 °F. Kelsey Wetzel prepped, like, 1000 SEM mounts for
me, many of which yielded the observations that form the foundation of chapters 3 and 4. Much of the data in this dissertation benefitted from analytical assistance from, and discussions with, Uttam Chowdhury, Stuart Thomson, and Fen-Ann Shen.

This research was funded by the National Science Foundation (NSF-EAR 1419828), the USGS Mineral Resources Program, a USU Presidential Doctoral Research Fellowship, a USU Dissertation Enhancement Grant, student research grants from the Geological Society of America and Tobacco Root Geological Society, and many USU Geoscience department awards.

Robert G. McDermott
DEDICATION

For Grandpap, who started me on the path to science, and Emily, for her endless support
and love
CONTENTS

<table>
<thead>
<tr>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>ABSTRACT .......................................................... iii</td>
</tr>
<tr>
<td>PUBLIC ABSTRACT ..................................................... v</td>
</tr>
<tr>
<td>ACKNOWLEDGMENTS .................................................... vii</td>
</tr>
<tr>
<td>DEDICATION .......................................................... ix</td>
</tr>
<tr>
<td>LIST OF TABLES .................................................... xiii</td>
</tr>
<tr>
<td>LIST OF FIGURES ................................................... xiv</td>
</tr>
</tbody>
</table>

CHAPTER

I. INTRODUCTION .............................................................................. 1

II. THERMOCHRONOMETRIC AND TEXTURAL EVIDENCE FOR SEISMICITY VIA ASPERITY FLASH HEATING ON EXHUMED HEMATITE FAULT MIRRORS, WASATCH FAULT ZONE, UT, USA ........................................... 21
   Abstract ............................................................................. 21
   Introduction .......................................................................... 22
   Hematite (U-Th)/He thermochronometry of fault rocks ................................. 24
   Multiscale characterization of hematite fault mirrors ..................................... 26
   Hematite and apatite (U-Th)/He thermochronometry results .......................... 29
   Evidence for elevated fault surface temperatures ........................................ 31
   Thermomechanical modeling of flash heating at geometric asperities .......... 34
   Seismicity on hematite fault mirrors ....................................................... 43
   Conclusions ........................................................................... 45
   References ............................................................................ 45

III. HEMATITE (U-TH)/HE THERMOCHRONOMETRY CONSTRAINS EPISODIC PRECIPITATION, FAULT REACTIVATION AND EXHUMATION OVER 100-MYR TIMESCALES IN THE WASATCH FAULT ZONE ................................................................. 50
   Abstract ............................................................................. 50
   Introduction .......................................................................... 51
   Analytical background .................................................................... 54
   Geologic setting ......................................................................... 56
   Methods ............................................................................... 61
   Results ................................................................................. 64
   Preliminary interpretation of hematite (U-Th)/He dates ................................. 82
   Thermal history modeling ................................................................... 84
   Discussion ............................................................................. 95
   Conclusions .......................................................................... 105
References .................................................................................................................. 107

IV. MICROTEXTURAL EVIDENCE FOR MULTIPLE DYNAMIC WEAKENING MECHANISMS DURING COSEISMIC SLIP ON HEMATITE FAULT MIRRORS FROM THE WASATCH FAULT ZONE, UT, USA .......................................................... 117
   Abstract .................................................................................................................. 117
   Introduction ............................................................................................................. 118
   Wasatch fault zone and hematite fault mirrors ..................................................... 120
   Samples and methods ............................................................................................ 123
   Results .................................................................................................................... 129
   Discussion .............................................................................................................. 144
   Conclusions ............................................................................................................ 158
   References ............................................................................................................. 158

V. THERMOTECTONIC HISTORY OF THE KLUANE RANGES AND EVOLUTION OF THE EASTERN DENALI FAULT ZONE IN SOUTHWESTERN YUKON, CANADA ...................................................... 168
   Abstract .................................................................................................................. 168
   Introduction ............................................................................................................. 169
   Geologic setting ..................................................................................................... 173
   Low-temperature thermochronometry and methods ............................................ 181
   Results .................................................................................................................... 185
   Thermal history simulations .................................................................................. 191
   Oligocene to present patterns of Kluane Ranges exhumation and surface uplift 198
   Geological interpretations of thermal history, exhumation, and surface uplift patterns .................................................................................................................. 205
   Conclusions ............................................................................................................ 217
   References ............................................................................................................. 218

VI. DATING FAULT DAMAGE ALONG THE EASTERN DENALI FAULT ZONE WITH HEMATITE (U-TH)/HE THERMOCHEMOMETRY ....... 233
   Abstract .................................................................................................................. 233
   Introduction ............................................................................................................. 234
   Background ............................................................................................................ 236
   Samples and methods ............................................................................................ 241
   Results .................................................................................................................... 242
   Interpretation of (U-Th)/He dates and microtextural observations .................... 248
   Connecting fault-specific deformation with regional strain .................................. 258
   Implications for regional late Miocene tectonics .................................................. 260
   Conclusions ............................................................................................................ 261
   References ............................................................................................................. 263

VII. CONCLUSIONS .................................................................................................... 269
APPENDICES ........................................................................................................274

APPENDIX A ........................................................................................................275
APPENDIX B ........................................................................................................303
APPENDIX C ........................................................................................................313
APPENDIX D (COPYRIGHT RELEASES) .............................................................329
APPENDIX E (VITAE) ........................................................................................338
## LIST OF TABLES

<table>
<thead>
<tr>
<th>Table</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>189</td>
</tr>
<tr>
<td>1</td>
<td>189</td>
</tr>
<tr>
<td>S1</td>
<td>293</td>
</tr>
<tr>
<td>S2</td>
<td>296</td>
</tr>
<tr>
<td>S3</td>
<td>296</td>
</tr>
<tr>
<td>S4</td>
<td>297</td>
</tr>
<tr>
<td>S1</td>
<td>312</td>
</tr>
<tr>
<td>S1</td>
<td>325</td>
</tr>
<tr>
<td>S2</td>
<td>327</td>
</tr>
</tbody>
</table>

### CHAPTER 5

New bedrock (U-Th)/He and fission-track cooling dates

### APPENDIX A

S1 Hematite (U-Th)/He data
S2 Apatite (U-Th-Sm)/He data
S3 Zircon (U-Th)/He data
S4 Apatite fission-track data

### APPENDIX B

S1 Image analysis results

### APPENDIX C

S1 Hematite (U-Th)/He data
S2 Hematite (U-Th)/He weighted mean date, central date, and MSWD results
## LIST OF FIGURES

<table>
<thead>
<tr>
<th>Figure</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>CHAPTER 1</strong></td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>Interpretation of hematite (U-Th)/He dates from fault-related rocks</td>
</tr>
<tr>
<td>2</td>
<td>Location of the Wasatch and eastern Denali fault zones</td>
</tr>
<tr>
<td><strong>CHAPTER 2</strong></td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>Hematite fault surfaces, sample localities, and apatite (U-Th)/He dates for the Wasatch fault zone, UT</td>
</tr>
<tr>
<td>2</td>
<td>Fault surface hematite morphology and microtextures</td>
</tr>
<tr>
<td>3</td>
<td>Hematite and apatite (U-Th)/He thermochronometry data</td>
</tr>
<tr>
<td>4</td>
<td>Schematic hematite fault mirror showing asperities of different diameter that produce spatially and temporally variable heat pulses and textures</td>
</tr>
<tr>
<td>5</td>
<td>Time-temperature histories and He loss for asperities of different diameter at the fault surface</td>
</tr>
<tr>
<td><strong>CHAPTER 3</strong></td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>Geology of northeastern Utah, USA</td>
</tr>
<tr>
<td>2</td>
<td>Geologic maps of the Wasatch fault damage zone</td>
</tr>
<tr>
<td>3</td>
<td>Field photographs of Wasatch fault damage zone and hematite</td>
</tr>
<tr>
<td>4</td>
<td>Structural geology of Wasatch fault zone and hematite</td>
</tr>
<tr>
<td>5</td>
<td>Equal area plots of hematite elements at Tower and Pearsons Canyons</td>
</tr>
<tr>
<td>6</td>
<td>Hematite microtextures</td>
</tr>
<tr>
<td>7</td>
<td>Hematite (U-Th)/He data</td>
</tr>
<tr>
<td>8</td>
<td>Intrasample patterns of hematite (U-Th)/He and conventional bedrock thermochronometry data</td>
</tr>
</tbody>
</table>
CHAPTER 4

1 Structural geology of the Wasatch fault zone .......................................................... 121
2 Microtextures of hematite fault mirrors ................................................................. 131
3 Detail of Domain 1 microtextures in W17-1G and W17-1J ........................................ 133
4 W17-1G polygonal grain size patterns ................................................................. 134
5 Quantitative image analysis results ........................................................................ 136
6 Grain size distribution data .................................................................................... 138
7 Inverse pole figure maps for domains 1 and 2 ....................................................... 140
8 Pole figures for domains 1 and 2 .......................................................................... 141
9 Domain 1 misorientation analysis results ............................................................... 143
10 Model for textural and mechanical evolution of Wasatch fault zone hematite fault mirrors ................................................................. 152

CHAPTER 5

1 Tectonics of the Denali fault system and south-central Alaska-southwest Yukon ................................................................. 171
2 Study area digital elevation model with sample locations .................................... 174
3 Apatite and zircon (U-Th)/He and fission-track data ........................................ 188
4 Thermochronometric data spatial and date-elevation patterns........................191
5 Thermal history models for far-from-fault samples.................................194
6 Thermal history models for near-fault samples .....................................195
7 Integrated thermal history model results.............................................196
8 Phase III net surface uplift, exhumation, and total surface uplift ..........201
9 Simplified Kluane Ranges geology and cooling phases........................207
10 Tectonic and topographic evolution of the Kluane Ranges and environs.....209

CHAPTER 6
1 Geology of the Kluane Ranges and eastern Denali fault zone..................238
2 Microfabrics of low-elevation hematite-coated fault surfaces...............245
3 Microfabrics of high-elevation hematite-coated fault surfaces..............246
4 Hematite (U-Th)/He data .....................................................................248
5 Regional low-temperature thermochronometry and ambient thermal history patterns.................................................................249
6 Pseudo-Arrhenius plot for nonmonotonic cooling histories....................255
7 Schematic of eastern Denali fault zone processes and regional surface uplift and exhumation.................................................................259

APPENDIX A
S1 Photographs of hematite hand samples.............................................281
S2 Library of hematite aliquot reflected light and scanning electron microscope images.................................................................282
S3 Plate width and closure temperature distributions measured from representative SEM mounts.................................................................287
S4 Hematite (U-Th)/He date versus Th/U ratio and effective uranium ........289
S5 Photomicrographs of sample W14-T1 zircon aliquots showing variable preservation of metamictization and radiation damage .............................................. 290

S6 Pilot inverse models of bedrock thermochronometry data .............................................. 291

S7 Predicted zircon (U-Th)/He date-effective uranium curves for forward thermal history simulations .......................................................................................... 292

APPENDIX B

S1 Thin sections of fault mirror samples W17-1G and W17-1J ............................................ 308

S2 Grain tracings for image analyses ..................................................................................... 309

APPENDIX C

S1 Hematite-coated fault surface hand sample photographs .................................................... 315

S2 Additional photomicrographs of select hematite-coated fault surfaces ....................... 316

S3 Image catalogue of (U-Th)/He-dated aliquots and scanning electron microscope images from representative samples ......................................................... 317

S4 Energy dispersive spectroscopy maps and spot analyses of hematite-coated fault surface samples ...................................................................................... 322

S5 Grain size and closure temperature distributions ............................................................... 323

S6 Hematite (U-Th)/He date versus Th/U ratio and effective uranium concentration ........................................................................................................ 255
CHAPTER 1
INTRODUCTION

EXHUMED FAULT ROCKS AS ARCHIVES OF FAULT ZONE EVOLUTION
AND SEISMOGENESIS

Fault zones are fundamental structures that accommodate strain, advect heat, and transport mass through Earth’s upper lithosphere. At the landscape and regional scales, fault zones focus tectonic stresses that influence mountain growth and(or) basin formation and reflect the larger geodynamic setting (Norris and Carter, 1982; Wallace, 1984; Oldow et al., 1990). The rheological structure of a fault zone impacts surface deformation (e.g., Savage and Burford, 1973) and controls the distribution of seismic and aseismic slip in space and time (e.g., Sibson, 1983). Classic views of fault rheology rely on either temporally and(or) spatially (e.g., depth) invariant deformation regimes (Sibson, 1983; Scholz, 1998). However, a growing number of geodetic, seismological, and rock record observations challenge this view and instead indicate rheology is transient in space (m’s to km’s) and time (s to Myr; Heaton, 1990; Peng and Gomberg, 2010; Price et al., 2012; Melosh et al., 2018; Fagereng and Biggs, 2019). Understanding the timescales and mechanisms of different slip behaviors are critical to an integrated view of fault zone evolution in space and time.

Exhumed fault rocks offer an important opportunity to study various fault zone processes and the mechanisms of transient rheology in situ (Niemeijer et al., 2012; Rowe and Griffith, 2015). For example, earthquake nucleation and seismic slip is dependent on weakening of fault materials concomitant with slip (i.e., dynamic weakening) at a rate
faster than the release of tectonic stresses (Brace and Byerlee, 1966; Dieterich, 1992). Laboratory experiments offer insight into a range of dynamic weakening processes, but only studies of natural fault rocks can confirm the mechanisms leading to real earthquakes (Di Toro et al., 2011; Niemeijer et al., 2012; Rowe and Griffith, 2015). Weakening mechanisms can vary in space and time, and over the duration of a single slip event (Kirkpatrick and Shipton, 2009; Brantut and Platt, 2017). Recognizing these mechanisms requires first identifying evidence for friction-generated heat, or lack thereof, and deformation mechanisms from microtextural observations (e.g., Rowe and Griffith, 2015).

A significant advance in our knowledge of fault zone deformation is the development of geo- and thermochronometers that can track fault zone thermal and chemical evolution. Existing radioisotopic methods for dating brittle regime deformation include: K-Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ dating of gouge, cataclasite, and pseudotachylyte (Vrolijk and van der Pluijm, 1999; Sherlock et al., 2009; Klepeis et al., 2019); U-Pb dating of calcite or opal precipitates and U-series dating of calcite slip surfaces and cements (e.g., Williams et al., 2017; Nuriel et al., 2019); and Rb-Sr dating of calcite and illite slickenfibres (Tillberg et al., 2020). Techniques for dating plastic deformation include in situ analyses of synkinematic minerals, such as $^{40}\text{Ar}/^{39}\text{Ar}$ or Rb-Sr in muscovite and U-Pb in monazite (Dunlap et al., 1991; Freeman et al., 1997; Williams and Jercinovic, 2002). An inherent challenge in any of these techniques, but particularly brittle regime methods, is unraveling the competing thermal effects of exhumation, fluid-rock interactions, and friction-generated heat on a given radioisotopic system (Ault et al., 2015; Scheiber et al., 2019; Ault, 2020).
Low-temperature thermochronometry of minerals formed in or deformed by faults also offers the potential to not only directly date deformation, but also inform fault slip temperatures, deformation mechanisms, and fault strength when combined with textural observations. Hematite is common in fault rocks, as it precipitates from synkinematic fluids present in or circulating through fault zones, forming veins and thin coatings on slip surfaces. Hematite incorporates trace amounts of U and Th and negligible initial \textsuperscript{4}He upon crystallization, and retains \textsuperscript{4}He over geological timescales (Bähr et al., 1994).

Additionally, hematite exhibits a suite of microtextures that can reflect the chemical conditions of precipitation and strain rates ranging from seismic to aseismic (Skulan et al., 2002; Siemes et al., 2003; Vallina et al., 2014; Ault et al., 2015; McDermott et al., 2017; Moser et al., 2017; Ault, 2020; Calzolari et al., 2020).

Timing constraints derived from geo- and thermochronometry are critical for linking episodes of exhumation and(or) topographic change to fault networks. In intracontinental settings, making these linkages are particularly important to understanding how or if fault activity responds to far-field plate boundary processes (e.g., Fitzgerald et al., 1995; Benowitz et al., 2011; Klepeis et al., 2019). Strain localization and fault reactivation that influence mechanical properties of fault materials and catalyze seismogenesis may occur over variable spatial and temporal scales, ranging from Myr to s (e.g., Sibson, 1985, 1992; Chester and Logan, 1986; Ikari, 2015; Smith et al., 2015; Fagereng and Biggs, 2019). Exhumed fault rocks will show the net effect of these overprinting processes (e.g., Sibson, 1977). Dates of deformation from hematite He thermochronometry, when combined with microtextural characterization and estimates of exhumation rate derived from conventional low-temperature thermochronometry, can
constrain the depth range over which a certain deformation mechanisms or mineralization/cementation styles may operate. These integrated perspectives are critical to forming a more comprehensive view of (1) how deformation style and fault zone mechanics vary with depth, (2) how deformation style may evolve with changing ambient pressure and temperature as a function of exhumation, and (3) how fault strength evolves through the seismic cycle.

HEMATITE (U-TH)/HE THERMOCHRONOMETRY TRACKS FAULT ZONE THERMAL EVOLUTION

Central to this dissertation is the application of hematite He thermochronometry to constrain fault zone thermal processes. Radiogenic $^4$He is produced by alpha ($\alpha$) decay of $^{238}$U, $^{235}$U, $^{232}$Th, and $^{147}$Sm (Rutherford, 1907). Retention of $^4$He is temperature dependent, such that a mineral will retain or lose $^4$He depending on when it is above or below, respectively, a certain temperature (termed “closure temperature”, or $T_c$ hereafter; Dodson, 1973). The $T_c$ of any thermochronometric system is dependent on the kinetics of $^4$He diffusion (i.e., activation energy and diffusivity at infinite temperature) in a given mineral, diffusion domain lengthscale (commonly grain radius), a geometric factor that describes diffusion domain shape, and cooling rate (Dodson, 1973). In the hematite He system, individual plates are the diffusion domain and their thickness exerts a dominant control on $T_c$ (Farley, 2018; Jensen et al., 2018). The $T_c$ may range from $\sim$250 °C for a 1 mm-thick plate to $\sim$70 °C for a 20 nm-thick plate (Farley, 2018). Aliquots of natural hematite large enough for hematite He dating are typically polycrystalline. Thus, most analyzed aliquots have an effective “bulk” $T_c$ dependent on grain size distribution (Farley
Variability in temperature sensitivity of the hematite He system has led to a wide variety of applications to date, including dating hydrothermal alteration and/or subsequent exhumation (Lippolt et al., 1995; Wernicke and Lippolt, 1997; Farley and Flowers, 2012; Evenson et al., 2014; Farley and McKeon, 2015; Jensen et al., 2018), diagenesis and/or water table migration (Reiners et al., 2014; Cooper et al., 2016; Garcia et al., 2018), and in conjunction with goethite, paleosol formation and paleo-weathering (Shuster et al., 2005, 2012; Danišík et al., 2013).

Hematite is common in fault-related rocks and precipitates and offers a means to fingerprint the thermochronometric signature of fault zone thermal processes such as seismic slip, hydrothermal alteration, fluid flow, and exhumation (Fig. 1; Ault et al., 2015, 2016; McDermott et al., 2017; Moser et al., 2017; Calzolari et al., 2018, 2020; Ault, 2020). Hematite may form over a range of temperatures (~100 to >~300 °C, depending on additional factors such as oxygen fugacity and pH), corresponding to depths of ~4 km to >~10 km (Catling and Moore, 2003; Cornell and Schwertmann, 2003). Thus, hematite can precipitate above or below its effective $T_c$, which depends on initial grain size distribution. If hematite forms at depths less than the $T_c$ isotherm associated with its grain size, a corresponding hematite He date may reflect the timing of precipitation (Fig. 1a). For hematite forming at temperatures greater than the $T_c$, hematite He dates may reflect later cooling through the $T_c$ and thus exhumation (Fig. 1b). Hematite in fault zones that as already cooled below its $T_c$ may be subject to additional thermal processes resulting in partial or full thermal date resetting (Fig. 1c). For example, hematite within fault rocks that experience friction-generated heat during seismic slip may be isotopically reset by ~10-100%, depending on the temperature and duration of
slip and grain size distribution (Ault et al., 2015; McDermott et al., 2017; Calzolari et al., 2020). Similarly, because hematite-precipitating fluids can be ~100 to >500 °C (e.g., Jensen et al., 2018; Walter et al., 2018, 2020), local fluid-rock interactions may also perturb hematite He dates. “Mixed” hematite He dates may result from newly-formed hematite incorporating pre-existing hematite and sampling (possibly inadvertently) multiple generations for analysis, if the gap between successive generations spans timescales greater than the hematite He date uncertainty.

Figure 1. Interpretation of fault rock hematite (U-Th)/He dates. (a) Hematite forming below its closure temperature (T_c) at time t_0 and exhumed over the time period t_1 until surface exposure at time t_2 will potentially preserve formation. (b) Hematite above its T_c at t_0 will record a cooling date reflecting exhumation. (c) Hematite that experiences sufficient temperatures below its T_c may undergo textural changes and yield a He date reflecting thermal resetting. Modified from Ault (2020).

Microtextural analyses are a critical component for interpretation of hematite He thermochronometry data from fault rocks and help connect dates to fault-specific
processes (Ault et al., 2015; McDermott et al., 2017; Ault, 2020). First, well-characterized grain size distributions are essential to constraining sample- and(or) aliquot-specific $T_c$. Hematite He dates can be compared to independent thermochronometers with $T_c$ that overlap or bracket a sample-specific hematite He $T_c$ (e.g., apatite and zircon (U-Th)/He, $T_c = 90$ to $30$ °C and $220$ to $<25$°C, respectively; Flowers et al., 2009; Guenthner et al., 2013) to rule out or confirm an exhumation-related interpretation. Second, microtextural observations may reveal the nature of thermal processes impacting hematite He dates. Temperatures of ~300 to ~1200 °C may result in full or partial date resetting, but also leave a distinct microtextural fingerprint via annealed and(or) recrystallized grains (Ault et al., 2015; McDermott et al., 2017; Calzolari et al., 2020). In contrast, hematite deformed by aseismic or subseismic slip may exhibit foliated textures, allowing thermal resetting by frictional heating to be excluded for these samples (Moser et al., 2017; Calzolari et al., 2018). Third, microtextural observations can inform sampling strategies for hematite He analysis and data interpretation. The minimum aliquot size for hematite He dating is ~120 μm (Evenson et al., 2014). Microtextural observations can aid in determining whether isolation of specific microtextural domains or hematite generations is compatible with this size constraint, or whether aliquots and dates will encapsulate a mixture of various processes. Where evidence for competing thermal processes is observed, microtextures can inform coupled thermal-$^4$He diffusion models that quantify and isolate the thermochronometric signature of each process (McDermott et al., 2017).
SCIENTIFIC GOALS AND STUDY LOCALES

The primary scientific goals of this dissertation are two-fold. First, I aim to develop workflows for the integrated interpretation of hematite He dates and microtextural data. As discussed above, hematite He dates from fault rocks may record a variety of thermal processes in fault zones. Accurate interpretation of hematite He data is critical to linking results to a process of interest. Second, I intend to utilize coupled textural and thermochronometry data to identify potential indicators of seismic or subseismic slip in the rock record. Textural analysis also informs coseismic dynamic weakening mechanisms and the role of other fundamental fault zone processes such as strain localization and fault reactivation in seismogenesis and fault zone evolution.

This dissertation focuses on two study areas: the Wasatch fault zone (WFZ) in northeastern UT, USA, and the eastern Denali fault zone (EDFZ) in southwest Yukon, Canada (Fig. 2). The WFZ and EDFZ are classic, archetypal examples of intracontinental normal and strike-slip fault systems, respectively (St. Amand, 1957; Machette et al., 1991). Each fault zone includes an abundance of hematite reported in prior literature (Evans and Langrock, 1994; Caine et al., 2013, 2014, 2015). These locales are thus ideal natural laboratories for (1) the application of hematite He thermochronometry and (2) reasonable analogues for other fault systems. The WFZ and EDFZ have experienced protracted, multi-phase deformation (Hedge et al., 1983; Naeser et al., 1983; Bryant, 1988; Yonkee, 1992; Israel et al., 2005; Cobbett et al., 2016). Constraints on the timing of brittle deformation and the magnitude and tempo of regional exhumation are essential to linking observations from presently exposed rocks to the broader tectonic and geodynamic context. Prior to the conception of this dissertation, elements of the larger
tectonic history and related exhumation of each of these fault systems were under constrained. Investigations thus require conventional low-temperature (e.g., apatite and zircon He) thermochronometry, intended to refine the thermotectonic history and provide a backdrop for interpretation of hematite He data from WFZ and EDFZ fault rocks.

**Figure 2.** Location of the Wasatch and eastern Denali fault zones

**DISSERTATION OUTLINE**

The primary scientific contributions of this dissertation are presented over five chapters. Each chapter contains relevant figures, tables, and references and is prepared in a format for stand-alone publication in peer-review journals. Supplemental figures, tables of raw analytical data, and detailed analytical methods specific to each chapter are provided in the appendices. Chapters 2, 3, and 4 present results from the WFZ. Chapters
4 and 5 pertain to the EDFZ. A final concluding chapter, Chapter 7, links key outcomes of each chapter with broader research goals outlined here.

Chapter 2 is published in *Earth and Planetary Science Letters* (McDermott et al., 2017). New hematite He data, detailed microtextural observations from pre-screened aliquots, and host rock apatite He data provide evidence for He date resetting by friction-generated heat during seismic slip. Numerical models show the degree of resetting required by hematite He-apatite He data patterns and the spatial distribution of high-temperature microstructures is best explained by the localized and elevated temperatures attained by flash heating at geometric asperities. These results reveal that the WFZ fault mirrors preserve a record of shallow (≤ 2 km) seismicity, and that this seismicity was at least in part facilitated by asperity flash heating as a dynamic weakening mechanism.

Chapter 3 is in preparation for submission to the *Geological Society of America Bulletin* and presents a multi-faceted dataset including field and microtextural observations, geologic mapping, and structural analyses of WFZ hematite alteration; new and previously published (Ault et al., 2015; McDermott et al., 2017) hematite He dates; new zircon He, apatite He, and apatite fission-track (AFT) thermochronometry dates; and a suite of thermal history models that constrain cooling (exhumation) and reheating (burial) of rocks and hematite precipitation ages. Hematite He dates from different microtextural domains including specularite veins, hematite-cemented breccias, and overlying fault mirrors are reported. Hematite He dates from veins and breccia record a combination of initial precipitation and He loss during subsequent ambient reheating and cooling. Although fault mirrors typically accommodated Miocene and younger normal slip, compatible with Basin and Range extension, precursor vein and breccia material pre-
dates this deformation by 10s to 100s of Myr. Integrated results constrain a model for the structural evolution of rocks in the present day WFZ that includes episodic precipitation, fault reactivation, and coseismic slip set against progressive burial and exhumation of host rocks.

Chapter 4 is in preparation for submission to the *Journal of Structural Geology* and expands on the microtextural observations of chapters 2 and 3. The same hematite-cemented breccia and overlying fault mirror samples analyzed in Chapter 3 are targeted for detailed microtextural analyses. Reflected light and scanning electron microscopy, quantitative image analyses, and electron backscatter diffraction mapping expand evidence for different dynamic weakening mechanisms that facilitate WFZ seismicity. In addition to asperity flash heating (Chapter 2), we identify evidence for fluidization and nanoparticle lubrication. Results show dynamic weakening is intimately related to coseismic strain localization, but that textural and rheological development can vary over spatial scales of 100s of μm. When integrated with timing constraints from Chapter 3, these samples demonstrate how precursor damage influences host rock weakening and initial strain localization, favoring reactivation and later coseismic slip.

Chapter 5 is published in *Tectonics* (McDermott et al., 2019) and employs apatite He, zircon He, apatite FT, and zircon FT thermochronometry to interrogate exhumation and surface uplift in the EDFZ-bounded Kluane Ranges over the past 100 Ma. Thermal history inversions constrain three thermotectonic phases corresponding to Cretaceous terrane accretion and EDFZ initiation, Paleogene to Oligocene enhanced strike-slip activity on the EDFZ, and Oligocene to present transpression as a far-field response to flat slab subduction. We leverage previously published palynological data to broadly
bracket surface uplift of the last ~30 Ma. These data refine the structural history of the EDFZ in response to far- and near-field processes and highlight the corresponding landscape response.

Chapter 6 has been submitted to *Earth and Planetary Science Letters* and presents a suite of microtextural observations and new hematite He data from hematite slip surfaces in the EDFZ. Straightforward interpretation of these hematite He data is challenging owing to variable analytical uncertainties, a paucity of apatite He thermochronometry data with complementary $T_c$, and microtextural evidence for multiple hematite generations included in each aliquot. Despite these complexities, statistical tests, interpretation of microtextures and strain rate, and hypothetical He loss calculations suggest hematite He dates record episodes of Miocene-Pliocene hydrothermal alteration. In contrast to WFZ hematite, microtextures are consistent with reactivation of hematite at aseismic to subseismic slip rates at shallow ($\leq 2$ km) depths. Hematite He dates overlap with documented plate-reorganization and surface uplift. The results highlight connections between surface uplift and the fault networks that accommodate topographic change and also provide additional insight to far-field processes that may influence deformation along the EDFZ.

**REFERENCES**


Caine, J.S., Israel, S., Murphy, D.C., and Benowitz, J., 2015, A fault rock record of late Eocene strain partitioning along the Denali fault in southwestern Yukon, in Anchorage, AK.

Caine, J.S., Israel, S.A., Siccard, K.R., and Hults, C., 2013, Denali damage: bedrock deformation, kinematics, and fluid flow in the brittle regime along the Denali fault zone, Kluane Lake, Yukon Territory, Canada, in Denver, CO.

Caine, J.S., Murphy, D.C., and Israel, S.A., 2014, Bedrock geology, internal structure, kinematics, and carbonaceous fault rocks of the Denali fault zone at Telluride and Kimberley Creeks, southwest Yukon, Canada, in Vancouver, BC.


Hedge, C.E., Stacey, J.S., and Bryant, B., 1983, Geochronology of the Farmington Canyon Complex, Wasatch Mountains, Utah, in Geological Society of America


Smith, S.A.F., Nielsen, S., and Di Toro, G., 2015, Strain localization and the onset of

St. Amand, P., 1957, Geological and geophysical synthesis of the tectonics of portions of

Tillberg, M., Drake, H., Zack, T., Kooijman, E., Whitehouse, M.J., and Åström, M.E.,
2020, In situ Rb-Sr dating of slickenfibres in deep crystalline basement faults:

Enhanced magnetic coercivity of α-Fe2O3 obtained from carbonated 2-line
ferrhiydrite: Journal of Nanoparticle Research, v. 16, p. 2322,
doi:10.1007/s11051-014-2322-5.


Wallace, R.E., 1984, Patterns and timing of Late Quaternary faulting in the Great Basin
Province and relation to some regional tectonic features: Journal of Geophysical

Walter, B.F., Gerdes, A., Kleinhanns, I.C., Dunkl, I., von Eynatten, H., Kreissl, S., and
Markl, G., 2018, The connection between hydrothermal fluids, mineralization,
tectonics and magmatism in a continental rift setting: Fluorite Sm-Nd and
hematite and carbonates U-Pb geochronology from the Rhinegraben in SW
Germany: Geochimica et Cosmochimica Acta, v. 240, p. 11–42,

Walter, B.F., Jensen, J.L., Coutinho, P., Laurent, O., Markl, G., and Steele-MacInnis, M.,
2020, Formation of hydrothermal fluorite-hematite veins by mixing of continental
basement brine and redbed-derived fluid: Schwarzwald mining district, SW-
Germany: Journal of Geochemical Exploration, v. 212, p. 106512,

Wernicke, R.S., and Lippolt, H.J., 1997, (U+Th)–He evidence of Jurassic continuous
hydrothermal activity in the Schwarzwald basement, Germany: Chemical

Williams, R.T., Goodwin, L.B., Sharp, W.D., and Mozley, P.S., 2017, Reading a
400,000-year record of earthquake frequency for an intraplate fault: Proceedings
of the National Academy of Sciences, v. 114, p. 4893–4898,
doi:10.1073/pnas.1617945114.

CHAPTER 2

THERMOCHRONOMETRIC AND TEXTURAL EVIDENCE FOR SEISMICITY VIA ASPERITY FLASH HEATING ON EXHUMED HEMATITE FAULT MIRRORS, WASATCH FAULT ZONE, UT, USA

Abstract

Exhumed faults record the temperatures produced by earthquakes. We show that transient elevated fault surface temperatures preserved in the rock record are quantifiable through microtextural analysis, fault-rock thermochronometry, and thermomechanical modeling. We apply this approach to a network of mirrored, minor, hematite-coated fault surfaces in the exhumed, seismogenic Wasatch fault zone, UT, USA. Polygonal and lobate hematite crystal morphologies, coupled with hematite (U-Th)/He data patterns from these surfaces and host rock apatite (U-Th)/He data, are best explained by friction-generated heat at slip interface geometric asperities. These observations inform thermomechanical simulations of flash heating at frictional contacts and resulting fractional He loss over generated fault surface time-temperature histories. Temperatures of > ~700–1200 °C, depending on asperity size, are sufficient to induce 85–100% He loss from hematite within 200 μm of the fault surface. Spatially-isolated, high-temperature microtextures imply spatially-variable heat generation and decay. Our results reveal that flash heating of asperities and associated frictional weakening likely promote small

---


a Utah State University, b University of Arizona
earthquakes (Mw ≈ -3 to 3) on Wasatch hematite fault mirrors. We suggest that similar thermal processes and resultant dynamic weakening may facilitate larger earthquakes.

1. Introduction

Friction-generated heat is a primary by-product of seismicity (Sibson, 1975; Spray, 1992; Rowe and Griffith, 2015 and references therein). Heat is a first-order control on the physical mechanisms behind coseismic fault strength reduction, which is a prerequisite for earthquake nucleation and propagation (Brace and Byerlee, 1966; Scholz, 1998). Fault surface paleothermometers (e.g., Sibson, 1975; Polissar et al., 2011; Rowe and Griffith, 2015) can be used to estimate coseismic shear stress, a key variable in understanding dynamic rupture and slip (Brace and Byerlee, 1966; Zheng and Rice, 1998; Lapusta and Rice, 2003), earthquake energy budgets (Kanamori and Rivera, 2006), and recurrence intervals (Dieterich and Kilgore, 1996). In situ documentation and quantification of fault paleotemperatures are critical to deciphering the rock record of earthquakes and understanding the physics of earthquake processes at all scales.

Mineral coatings on slip surfaces preserve nano- to micro-scale deformation textures that reflect coseismic temperature and strength changes. For example, relict silica colloids on natural, silica-rich slip surfaces suggest gel lubrication as a weakening mechanism (Kirkpatrick et al., 2013). Amorphous Ca-oxide on carbonate faults imply decarbonation and concomitant weakening from elevated coseismic temperatures (Collettini et al., 2013). These observations complement rotary shear experiments illustrating thermally activated lubrication or flash heating-induced low frictional strength (e.g., Hirose and Bystricky, 2007; Goldsby and Tullis, 2011; Di Toro et al., 2011 and
references therein). The diverse lithologic makeup of natural fault zones merits investigation of the signatures of earthquake processes in other mineral phases. Hematite is commonly found on fault surfaces and may record thermomechanical information related to the seismic cycle.

Hematite is amenable to (U-Th)/He (He) thermochronometry, and we use this method to detect elevated paleotemperatures from past earthquakes on mirrored or high gloss, light reflective hematite-coated fault surfaces from Wasatch fault zone (WFZ), northern Utah. The diffusion of He from hematite, and thus hematite He dates, respond to short-duration, high-temperature, localized thermal anomalies that characterize fault slip (Ault et al., 2015). Prior geochemical, microtextural, and thermochronological study of some WFZ hematite slip surfaces suggest they may record elevated fault surface temperatures (Evans et al., 2014; Ault et al., 2015). We present 114 new and published (Ault et al., 2015) individual hematite He dates and apatite He data. A subset of hematite aliquots were prescreened with scanning electron microscopy (SEM) prior to (U-Th)/He analysis to link textural development with thermochronometry data patterns. We then parameterize models of flash heating at asperities, or frictional contacts, with attendant hematite He loss using microtextural and thermochronometric observables to determine if this dynamic weakening mechanism can explain our results. Modeled temperature rise at asperities is consistent with our data patterns, suggesting this process potentially promotes seismicity on these fault mirrors.
2. Hematite (U-Th)/He thermochronometry of fault rocks

Hematite He dates from hematite-coated fault surfaces reflect the thermal and mechanical processes operative within fault systems. Retention of radiogenic He in polycrystalline hematite depends on the distribution of individual diffusion domains, temperature, and cooling rate (Farley and Flowers, 2012; Evenson et al., 2014). The temperature transition from open to closed system behavior, or onset of He retention – the closure temperature, $T_c$ (Dodson, 1973), is between ~25–250 °C in the hematite He system (Farley and Flowers, 2012; Evenson et al., 2014). Individual crystallites are inferred to be the He diffusion domains and $T_c$ increases with grain (domain) size (Evenson et al., 2014). Variations in bulk $T_c$ are controlled by the grain size distribution in each polycrystalline aliquot.

Hematite mineralization occurs over a range of depths within the Earth that correspond to ambient temperatures either below or above its $T_c$, given an initial grain size distribution. If hematite forms and remains below its $T_c$ (i.e., lower T, shallower depth), hematite He data record the timing of hematite formation (Ault et al., 2015; Ault et al., 2016). Alternatively, hematite may form at greater depths with higher ambient temperatures than its $T_c$. Subsequent cooling to below the hematite He $T_c$ via exhumation yields a He date that reflects this cooling history (Farley and Flowers, 2012; Evenson et al., 2014). Hematite on fault surfaces experience heat and/or strain associated with fault slip. Mechanical grain size reduction during cataclasis can progressively lower the $T_c$. If fault surface hematite accumulates He below its $T_c$, thermal pulses from fault slip induce He loss from the crystal lattice by volume diffusion and/or recrystallization (Ault et al., 2015). Circulation of hot fluids along faults may also reset He dates (Ault et al., 2016).
Under these nonmonotonic cooling scenarios, the temperatures required for appreciable He loss are inversely proportional to the duration of heating (Fechtig and Kalbitzer, 1966; Reiners, 2009; Ault et al., 2015). If the timescale of heating is only a few seconds to minutes, requisite temperatures for He loss and resetting of the He date must substantially exceed the $T_c$.

Interpreting fault surface hematite He dates requires information on the hematite mineralization age, microtextural characterization, grain size (and therefore $T_c$) measurements, and independent constraints on the ambient cooling history. Here we acquire hematite He dates from specular veins, a potential precursor to the fault mirror hematite (Evans and Langrock, 1994; Ault et al., 2015). We characterize fault surface microtextures and aliquot grain size distributions with SEM to identify evidence of heat and deformation on these surfaces and bracket aliquot $T_c$. We use apatite (U-Th)/He thermochronometry to track cooling of the WFZ footwall. The apatite He system is sensitive to temperatures of ~30–90 °C, depending on the accumulation and annealing of decay-induced radiation damage, thermal history, and crystal size (Flowers et al., 2009). In rapidly cooled settings, such as the footwall of a normal fault, the apatite He $T_c$ is typically ~50–60 °C (10 °C/Ma cooling rate), corresponding to cooling though ~1–3 km depth. Prior study of WFZ fault surfaces report hematite crystallite radii (and thus diffusion domain length scales) of 0.05 to 5 μm, corresponding to a hematite He $T_c$ of ~75–150 °C (10 °C/Ma cooling rate; Ault et al., 2015). Complementary apatite He dates from nearby unaltered bedrock place a lower temporal bound on hematite He dates that record ambient cooling (Ault et al., 2015).
3. Multi-scale characterization of hematite fault mirrors

The WFZ hematite-coated fault surfaces are exposed in a ~300–400 m-thick footwall damage zone in Paleoproterozoic Farmington Canyon Complex gneiss (Fig. 1; Evans et al., 2014). Faults are mirrored, or high-gloss and light reflective. They are arcuate to planar, locally iridescent, ≤2 mm-thick, and commonly have <1–10 cm of observed offset (Figs. 1A, B; Evans and Langrock, 1994; Evans et al., 2014). Slickenlines and tool marks and grooves from asperity ploughing indicate dominantly normal, down-to-the-west slip compatible with WFZ extension and thus likely developed after initiation of the WFZ at ~12 Ma (Evans and Langrock, 1994; Ehlers et al., 2003). Fault surfaces exhibit one to several sets of slickenlines, indicating multiple slip events (Evans and Langrock, 1994; Evans et al., 2014). Samples analyzed in this study are generally planar and contain one or two visible lineations (Fig. S1). Field and microtextural observations suggest many now mirrored surfaces originated as specular hematite veins later modified by progressive slip and grain comminution (Figs. 1B; S1; Evans and Langrock, 1994; Ault et al., 2015).

We characterize the texture and grain size distribution of hematite aliquots extracted from specularite veins and high-gloss fault surfaces with secondary electron and back-scattered electron SEM imaging. Instrument operating conditions, detailed sample preparation notes, and the grain size measurement approach are described in the Supplementary Material². Hematite aliquots were extracted using a portable rotary tool and fine point tweezers and selected to avoid tool marks. All aliquots are approximately

² The Supplementary Material for this chapter is available at https://doi.org/10.1016/j.epsl.2017.04.020
cubic or rectangular in shape with fault-surface-perpendicular widths of ~120–200 μm (Figs. S2; S3). Hematite polycrystalline aggregates representative of dated aliquots were mounted in epoxy, polished in cross-sectional view, and imaged under high vacuum. SEM-prescreened aliquots (10–11 per sample) were mounted on copper sticky tape and imaged under low vacuum. Four to five of these aliquots were then selected for (U-Th)/He analysis to encapsulate a range of observed fault surface microtextures. Aliquots were removed from copper sticky tape mounts and inserted into Nb tubes without manipulation or breakage to preserve documented microtextures.
Vein samples (W14-HC-20, A13-6) comprise euhedral to subhedral, locally fractured, hematite plates with 0.07–10 μm half-widths (Table S1; Fig. S2). Fault surfaces comprise comminuted, subhedral, subrounded to subangular grains with minimum and maximum grain radii/half-widths of 0.02 and 1.5 μm, respectively, although the majority are 0.1–0.5 μm (Table S1). Some fault surfaces contain crystals with polygonal, triple junction-forming (~120°) grain boundaries (Fig. 2A-C) and/or lobate grain boundaries (Fig. 2D-F). These grain morphologies lack shape-preferred orientation and occur in spatially isolated ≤1–20 μm clusters at the slip surface surrounded by lobate and/or subangular grains (Figs. 2; S3; S4).

4. Hematite and apatite (U-Th)/He thermochronometry results

We report 66 new hematite He dates from seven fault surfaces and two veins combined with previously published results (Table S2; Fig. 3A; Ault et al., 2015). Hematite aliquots were analyzed for He, U, and Th at the University of Arizona using standard apatite lasing temperatures to prevent U and Th volatilization and zircon dissolution procedures. Additional details are provided in the Supplementary Material. Mean hematite He dates from high-gloss fault surfaces range from 12.3 ± 4.2 Ma to 2.4 ±
1.0 Ma (± 1σ standard deviation) with individual aliquot dates spanning 18.4 ± 0.6 Ma to 1.4 ± 0.2 Ma (± 2σ analytical error). Samples exhibit variable intrasample reproducibility with ~50% of samples yielding <10% sample mean standard deviation. Hematite He dates from undeformed vein samples (W14-HC-20, A13-6) are 88.5 ± 15.2 Ma and 10.7 ± 0.8 Ma. For comparison and to track the ambient fault footwall cooling, we acquired an apatite He date from isoelevational and unaltered host rock gneiss. The apatite He date of 4.5 ± 0.6 Ma overlaps previously published apatite He data from ~300 m to the northwest (method details in the Supplementary Material; Table S3; Fig. 1; Armstrong et al., 2004).

**Figure 2** Hematite morphology and microtextures. Scanning electron microscopy (SEM) secondary electron (SE) images at different scales showing hematite with polygonal (A-C) and lobate (D-F) grain boundaries. Black dashed line denotes fault surface.
Figure 3 Hematite and apatite (U-Th)/He thermochronometry data. (A) Hematite He dates for individual aliquots classified by sample with 2σ analytical error. (B) Relative probability (black line) and cumulative frequency (blue line) of hematite He dates. (C) Subset of SEM-prescreened hematite He dates with complementary microtextural data. Aliquots with polygonal (red hexagon, Fig. 2A-C) and lobate (red square, Fig. 2D-F) grain boundaries highlighted. Gray bar in A, B, and C is new apatite He date (mean ± 1σ standard deviation).

Thermochronometry results exhibit four important patterns. First, fault surface aliquots yield hematite He dates younger than those from specular vein samples W14-HC-20 and A13-6, with the exception of outliers from sample A13-3 (Fig. 3A). Second, 48% of fault surface hematite He dates are younger than the ~4.5 Ma apatite He date (Fig. 3B). Third, hematite He dates from two locations on the same fault surface (W15-16A, C) are internally consistent at <10% 1σ sample mean standard deviation, but
distinct (2.3 ± 0.1 Ma and 2.9 ± 0.2 Ma), mirroring previously published data patterns for other fault surface samples (WF94-17A, B, D, E, F; Ault et al., 2015). Finally, in (U-Th)/He-dated prescreened aliquots that purposefully encapsulate the range of crystal morphologies in each sample, aliquots containing clusters of crystals with polygonal or lobate grain boundaries yield hematite He dates that are typically younger than the apatite He data (Fig. 3C). Polygonal and lobate grain morphologies were documented in 18 of 38 pre-screened aliquots. Of these 18 aliquots, 13 (72%) yield dates statistically younger at \(1\sigma\) ± standard deviation of our acquired apatite He date, two (12%) overlap the apatite He date, and three (16%) are older than the apatite He date. Hematite He dates from these samples also generally display larger intrasample scatter than results from fault surfaces that were not prescreened, with the exception of samples W15-16A and W15-16C.

5. Evidence for elevated fault surface temperatures

Hematite crystal morphologies and microtextures reveal evidence for elevated fault surface temperatures. The polygonal hematite grain morphology is similar to textures interpreted as recrystallized hematite in previous study of WFZ surfaces (Ault et al., 2015). These morphologies are analogous to that observed in long-duration hematite torsion (Siemes et al., 2003; Siemes et al., 2011) and dry heating (Vallina et al., 2014) experiments to 1000-1100 °C. Lobate grains are similar to those observed in torsion and dry heating experiments conducted at lower temperatures (300-800 °C; Siemes et al., 2003; Siemes et al., 2011; Vallina et al., 2014). Laboratory and field studies of calcite slip surfaces report similar grain morphologies associated with high coseismic strain rates and
temperatures (Smith et al., 2013; De Paola et al., 2015). These previous studies suggest such textures are the result of annealing (recrystallization) and/or sintering.

We posit that documented polygonal to lobate hematite grain morphologies from the hematite-coated fault surfaces reflect similar processes. Annealing involves dislocation migration and removal, which require high temperatures but can be facilitated by an applied strain rate. The lack of a shape-preferred grain orientation suggests a component of hematite recrystallization occurred at the low differential stress and strain rate of the “static” post-slip period. On-going annealing under static conditions may occur in areas that experienced rapid dislocation migration and high friction-generated temperatures during the coseismic period (e.g., Trepmann et al., 2007). Comparison of our observed microtextures with experimental data indicates temperatures of $\geq 300$–$1000$ °C at the fault surface.

Thermochronometric data patterns suggest complex and spatially variable fault surface thermal histories. We use $T_c$ calculations as an index of fault surface hematite He retentivity and temperature sensitivity. For this thought experiment, we assume a spherical diffusion length-scale equivalent to the majority of our hematite crystal radii observed with SEM (0.1–0.5 $\mu$m), a 10 °C/Ma monotonic cooling rate, and published hematite He diffusion kinetics (Evenson et al., 2014). We compare these data with a calculated host rock apatite He $T_c$ under the same ambient cooling condition and with the approach of Flowers et al. (2009). The apatite He $T_c$ calculation assumes the diffusion kinetics of Flowers et al. (2009), an effective uranium concentration of 30 ppm, and a diffusion domain length-scale of 35 $\mu$m (Table S3). The (U-Th)/He $T_c$ estimates for fault surface hematite and host rock apatite are $\sim 95$–$110$ °C (Table S1) and $\sim 55$ °C,
respectively. Progressive grain size reduction due to fault slip and cataclasis decreases hematite He $T_c$ during footwall exhumation (Ault et al., 2015). Our measured grain size distribution was produced during the most recent slip event and therefore provides minimum $T_c$ estimates. Fault surface hematite He dates younger than apatite He dates are not consistent with hematite He dates recording ambient footwall cooling. Variable dates from different locations on individual fault surfaces (e.g., W15-16A, C and WF94-17A, B, D, E, F), as well as intrasample date variability for aliquots with similar grain size distributions and thus $T_c$, further argue against some hematite He dates recording footwall exhumation.

Hematite fault mirrors with high-temperature microtextures that yield hematite He dates younger than the apatite He date provide evidence of (1) friction-generated heat at the fault surface and (2) attendant thermal resetting (He loss) of the hematite He system. Our $T_c$ calculations and apatite and hematite He date relationships indicate the hematite He system in the WFZ records these processes at depths shallower than the apatite He $T_c$ isotherm ($\sim \leq 2$ km, assuming a geothermal gradient of 30 °C/km; Blackett, 2004). This does not preclude fault slip events and associated shear heating on these surfaces at greater depths and higher ambient temperatures. Prior faulting caused grain size reduction of specularite veins, decrease in the hematite diffusion domain length scale, and lower $T_c$, facilitating resetting by subsequent shear heating (Ault et al., 2015). Some samples from the same fault surface comprise aliquots with hematite He dates older and younger than the apatite He date, where both populations display high-temperature microtextures (Fig. 3C). This may be a result of slip events that occurred at ambient temperatures in excess of the apatite He $T_c$ isotherm prior to $\sim 4.5$ Ma. The presence of clusters of high-
temperature microtextures directly at the slip interface in SEM-prescreened samples, regardless of corresponding hematite He date, argues against this (Figs. 2; S3). Samples with high intrasample hematite He data scatter may be explained by spatially variable He loss. For example, consider hematite with a pre-slip He date of 6.5 Ma (the oldest observed date from surface W15-16) that experienced a slip event at 2.3 Ma (the mean of sample W15-16C). In this hypothetical scenario, ≤65% He loss over the scale of a measured aliquot will yield an apparent hematite He date that is older, or falls within, the 4.5 ± 0.6 Ma range of our acquired apatite He date. Thus, the relationships presented in Figure 3C and microtextural context of observed polygonal and lobate grains collectively suggest spatially-variable shear heating at shallow depths and post-4.5 Ma.

An alternative mechanism that could yield hematite He dates younger than apatite He dates is fluid circulation at temperatures below the apatite He $T_c$ isotherm, but this is ruled out based on microtextures and date patterns. Vein hematite He dates are older than fault surface results and suggest multiple generations of vein formation (Fig. 3A). Pervasive comminution in all fault samples and clusters of high-temperature microtextures at the fault surface imply that the most recent events to affect samples were cataclasis followed by localized frictional heating. We observe no textural evidence of neomineralization (i.e., platy hematite growth) overprinting cataclastic and/or recrystallized hematite textures in our aliquots.

6. Thermomechanical modeling of flash heating at geometric asperities

Thermochronometric and microtextural data patterns suggest hematite fault mirrors experienced localized, transient frictional heating from fault slip in the upper two
km of the crust. The primary controls on friction-generated heat on fault surfaces are the coefficient of friction and normal stress (i.e., shear stress), displacement, slip velocity, and shear zone width (Lachenbruch, 1986). At the fast slip rates that characterize earthquakes (i.e., >0.001 m/s; Rowe and Griffith, 2015), heat production outpaces heat dissipation and shear heating at the slip surface may be high (e.g., Lachenbruch, 1986). Laboratory studies of fault friction indicate a correlation between dynamic fault strength and slip velocity (e.g., Dieterich, 1979; Di Toro et al., 2011 and references therein). At slip rates ≥0.1 m/s, thermally-activated weakening mechanisms reduce fault friction and strength (Di Toro et al., 2011). Different dynamic weakening mechanisms inferred from laboratory, theoretical, or field studies reflect lithology, fluid content, and the rate of heat production vs. dissipation controls (e.g., Di Toro et al., 2011 and references therein). These mechanisms include lubrication by the formation of melt (Sibson, 1975; Hirose and Shimamoto, 2005), gels (Kirkpatrick et al., 2013), weak mineral phases at the slip surface (Hirose and Bystricky, 2007), decreases in effective normal stress and/or gouge fluidization by thermal pressurization of pore fluids (Brantut et al., 2008), and weakening of frictional contacts by asperity flash heating (Rice, 2006; Hirose and Bystricky, 2007; Goldsby and Tullis, 2011). Dynamic weakening typically results in measured friction coefficients of 0.1, in contrast to static, or low slip rate, friction coefficients of ~0.4-0.8 (e.g., Byerlee, 1978; Di Toro et al., 2011). Importantly, frictional weakening during slip is a necessary condition for earthquake rupture propagation (Brace and Byerlee, 1966; Scholz, 1998 and references therein).
Figure 4 (A) Schematic of hematite fault mirror showing asperities of different diameters that produce spatially and thermally variable heat pulses and textures. Note different vertical and horizontal scales. (B) Thermomechanical simulation results showing time-temperature paths for hematite at fault surface (z=0 mm) for flash heating at a 20 μm diameter asperity. Dashed portion of x-axis is schematic pre-slip time interval. Dashed and solid yellow bars beneath x-axis indicate pre- and post-slip periods, respectively. Red portion denotes slip event and associated temperature rise. (C) He loss with depth for diffusion domains of different radius r for 20 μm asperity flash heating scenario. Red shaded region indicates bulk He loss for dominant 0.1–0.5 μm grain radii in most fault surface aliquots. Vertical dashed line denotes typical ~200 μm aliquot thickness.
We suggest spatially isolated clusters of polygonal and lobate grains at the surface of hematite fault mirrors are the thermal and mechanical footprints of geometric asperities (Fig. 4A). Micro-scale frictional contacts on the fault surface concentrate stress, resulting in localized thermal anomalies or flash heating during rapid fault slip (Fig. 4A; Dieterich and Kilgore, 1994; Rice, 2006). The correlation between high-temperature microtextures and thermally reset hematite He dates suggests this process operated on exhumed WFZ hematite fault mirrors. We further explore this interpretation with a suite of thermomechanical models. These models calculate the temperature evolution through time at frictional contacts and the fault surface and couple these outputs to a model of He loss from hematite. We first discuss the model framework and parameterization followed by simulation results. An abbreviated description of the model setup is presented here. Additional details of our calculations and a discussion of thermal history sensitivity to parameter choice are described in the Supplementary Material.

6.1. Model framework and parameterization

The temperature at a frictional contact, $T_{fc}$, is (Rice, 2006):

$$T_{fc} = T_{surf} + \frac{\mu_{fc}HV\sqrt{\beta}}{\rho C \sqrt{\alpha}}$$

(1)

where $T_{surf}$ is the macroscopic (i.e., fault surface-averaged) temperature over the slipping patch, $\mu_{fc}$ is the contact coefficient of friction, $H$ is the indentation hardness of the mineral, $V$ is the slip velocity, $\beta$ is the asperity lifetime (diameter/$V$), $\rho$ is density, $C$ is heat capacity, and $\alpha$ is thermal diffusivity. The term $T_{surf}$ in equation (1) is derived from Lachenbruch (1986):

$$T_{surf} = \frac{\mu_{f}Vt^*}{2\rho Ch} + T_{amb} = \frac{rD}{2\rho Ch} + T_{amb}$$

(2)
where \( t^* \) is the duration of slip, \( T_{amb} \) is ambient temperature, \( \mu \) is the coefficient of friction, \( \sigma_n \) is the normal stress, \( \tau \) is shear stress, and \( D \) is displacement.

We calculate the temperature profile beneath and asperity and within the shear zone during and after slip with a 1D thermal model (Lachenbruch, 1986). Temperature, \( T \), is:

\[
T(z, t) = t \left( 1 - 2i^2 erfc \left( \frac{h-z}{\sqrt{4\alpha t}} \right) - 2i^2 erfc \left( \frac{h+z}{\sqrt{4\alpha t}} \right) \right) - (t - t^*) \left( 1 - 
2i^2 erfc \left( \frac{h-z}{\sqrt{4\alpha (t-t^*)}} \right) - 2i^2 erfc \left( \frac{h+z}{\sqrt{4\alpha (t-t^*)}} \right) \right) + T_{amb}
\]

where \( z \) is shear zone depth, \( t \) is time, and \( h \) is half width of the deforming zone. The term \( i^2 erfc(\delta) \) denotes the second integral of the complementary error function of \( \delta \). The time-integrated thermal history as a function of shear zone depth is coupled to a model of He volume diffusion (Fechtig and Kalbitzer, 1966; see Supplementary Material). We posit He loss from hematite crystals during faulting occurs via a combination of recrystallization and thermally activated volume diffusion. Polygonal and lobate hematite crystals interpreted as recrystallization features are isolated in \(<20 \, \mu m\) clusters and rarely extend more than \(~10 \, \mu m\) into the underlying comminuted hematite (Fig. S4). He is completely lost from recrystallized regions, but this only affects a volumetrically small portion of our dated aliquots. For simplicity, we calculate He loss by time-temperature-dependent volume diffusion as a proxy for slip-induced He loss regardless of process. Our use of a 1D temperature model has the implicit assumption that the asperity distribution and spacing across the fault surface was not sparse enough create fault-parallel thermal gradients that exceed the fault-perpendicular gradient (i.e., heat transfer occurs dominantly in one direction). We emphasize that our calculated temperatures
reflect the maximum temperatures at any given depth in the shear zone and that the 1D calculation becomes less realistic with increasing distance from the slip surface.

We discuss our model parameterization below. In our model simulations, we set $V$ at 1 m/s to be consistent with seismic slip rates and evidence for frictional heating on hematite fault mirrors (e.g., Sibson, 1975; Heaton, 1990; Spray, 1992; Rowe and Griffith, 2015 and references therein). We posit that $\mu_{fc}$ evolves from 0.6 to 0.1, but is closer to 0.6, over the contact lifetime and thus assume an average $\mu_{fc}$ of 0.4. We set $H=2.7$ GPa, a typical value for hematite (Chicot et al., 2011). The terms $\rho$, $C$, $\alpha$ are set to 2700 kg/m$^3$, 790 J/kg K, and $1.27\times10^{-6}$ m$^2$/s, respectively, to be consistent with the typical physical properties of granitic host rock (Robertson, 1988). We select asperity contact diameters consistent with the observed spatial extent of high-temperature microtextures, ~1–20 μm (Figs. 2; S4), although we model contact diameters up to 50 μm. In equation (2), $\mu$ is assumed to be 0.1, consistent with the average $\mu$ observed in rotary shear experiments conducted at seismic slip rates (e.g., Di Toro et al., 2011; Goldsby and Tullis, 2011), and displacement $D$ and half-width $h$ are assumed to be 6.5 cm and 1 mm, respectively. These two parameter choices reflect the average displacement and shear zone half-width observed in the field. Observed displacement is likely cumulative and this value represents an upper bound on macroscopic shear heating from equation (2). Hematite and apatite He data patterns suggest that slip events occur at depths $\leq$2 km. Assuming a 30 °C/km geothermal gradient (Blackett, 2004) and lithostatic pressure yields $T_{\text{amb}}$ and $\sigma_n$ of 45 °C and 40 MPa, respectively, at our chosen model depth of 1.5 km. The local coseismic and post-seismic temperature distribution from flash heating at asperities is calculated by equating the total temperature rise from equation (1) to an “equivalent” $\tau$. 
for use in equation (3). He loss is modeled from a range of grain sizes corresponding to our observed grain size distribution and using hematite He diffusion kinetics from Evenson et al. (2014).

6.2. Model results

Thermomechanical simulations indicate flash heating of asperities can generate temperatures required to explain hematite grain morphologies and reset He dates (Fig. 4B). Hematite He dates are most sensitive to the maximum temperature experienced during slip. Peak asperity flash temperature is proportional to contact diameter [equation (1)]. The largest observed diameter of clustered polygonal grains is ~20 μm (Fig. S4) and gives an estimate of the likely peak temperature on these fault mirrors. In this scenario, peak asperity temperature is ~1240 °C (Fig. 4B). He loss occurs from a variety of grain sizes observed in hematite-coated fault surfaces, with ~85-100% fractional loss from domains with radii of 0.1–0.5 μm within 200 μm of the fault surface (typical aliquot thickness; Figs. 4C; S2; S3). Complete resetting occurs for smaller grain radii and plates with 5 and 10 μm half-widths display ~13% and ~7% He loss, respectively (Fig. 4B). Peak slip surface temperatures associated with asperities of 1–50 μm diameters range from 360 to 1900 °C (Fig. 5A) and result in 0-100% He loss from 0.1 μm domains, the lower end of the observed bulk grain size distribution (Fig. 5B). Calculated peak asperity temperatures are not strongly sensitive our choice of $T_{\text{surf}}$, which only contributes a temperature rise above $T_{\text{amb}}$ of ~60 °C (Fig. 5A, inset). Reducing displacement so that macroscopic temperature rise is small [i.e., $T_{\text{surf}} = T_{\text{amb}}$ in equation (1)] results in a negligible change to calculated He loss in all simulations.
Figure 5 (A) Time-temperature histories for asperities of different diameter (d) at the fault surface (z=0 mm). X-axis annotations are those in Figure 4A. Inset shows peak asperity temperature as a function of d with (black line) and without (gray line) contribution from 6.5 cm of displacement. (B) He loss with depth from 0.1 μm domain for different asperity d values. Colors correspond to time-temperature curves in (A). Vertical dashed line denotes typical ~200 μm aliquot thickness.

For comparison to asperity flash heating simulations, we calculate fault surface temperatures and associated He loss from macroscopic, or bulk surface, shear heating [i.e., using equations (2) and (3); Fig. S5]. These simulations vary D (1–10 cm), V (0.1–1 m/s), and average τ (2.7–42.4 MPa). Parameter ranges are further discussed in the Supplementary Material. T_{surf} is ~105 °C for D = 6.5 cm, V = 1 m/s, and τ = 4 MPa and yields negligible He loss. Other simulations indicate that He loss (5-100%) can only be achieved for displacements ≥5 cm and at high τ associated with slip at depths equal to or greater than the apatite He Tc isotherm, and/or high coefficients of friction (μ=0.8).
Observed displacements across hematite fault mirrors are typically <10 cm and average at 6.5 cm. Multiple generations of slickenlines also indicate that displacement potentially accumulated over multiple events, and displacement per event was likely less than even the average estimate. In addition, hematite and apatite He data patterns indicate that many surfaces experience resetting below the apatite He $T_c$ isotherm, where $\sigma_n$ at the macroscopic scale is insufficient to induce measurable He loss. Effective $\sigma_n$ will be lower in the presence of fluid pressure, which we do not consider in these simulations.

Thermomechanical simulations reveal our microtextural and thermochronometry data patterns reflect transient, localized thermal pulses (~700–1200 °C) from asperity flash heating during fault slip. Peak flash temperatures and temperature distribution in space and time was likely heterogeneous on each analyzed fault mirror. Bulk hematite He dates reflect thermal resetting from the time-variant activation of a 2D spatial distribution of asperities across aliquots and slip surfaces (Figs. 3C; 4A). Intrasample scatter in hematite He dates from these samples may also reflect that (1) the size of all analyzed aliquots may exceed regions impacted by high temperatures, and (2) dated, SEM-prescreened hematite aliquots purposefully encapsulate a range of observed microtextures, not solely polygonal grains. In addition, the asperity (flash) temperature distribution will control macroscopic scale heating and fault strength (e.g., Rice, 2006; Beeler et al., 2008). Hematite He dates from samples comprising aliquots with a higher volume fraction of polygonal/lobate grains such as W15-16A and W15-16C are comparatively reproducible versus other samples in Figure 3C. This suggests the asperity distribution, and thus heating, was more uniform across this surface relative to others.
7. Seismicity on hematite fault mirrors

Asperity flash heating and concomitant weakening likely promote earthquakes on these high-gloss fault surfaces. Most rocks experience dramatic strength reduction of micro-contacts at temperatures ≥900–1000 °C (Spray, 1992; Rice, 2006; Goldsby and Tullis, 2011). Inferred paleoasperity dimensions and model simulations parameterized by our data indicate flash-heating temperatures >360–1200 °C (Fig. 5A inset), although measurable He loss by volume diffusion requires temperatures of at least ~700 °C. Hematite asperity contacts that experienced the upper end of this temperature range may have failed, resulting in dynamic weakening. Our results do not preclude other weakening mechanisms operating on these fault surfaces but provide compelling evidence that asperity flash heating does occur during slip. Strength reduction during slip is a requirement for earthquake rupture propagation and seismogenic slip. Evidence for frictional heating and dynamic weakening suggests that WFZ hematite fault mirrors accommodated ancient seismicity.

We calculate potential moment magnitudes (Mw) for documented paleoearthquakes (details in Supplementary Material). Exposed hematite-coated fault mirrors crop out in 0.3–30 m² isolated patches (Evans and Langrock, 1994; Evans et al., 2014). If these are representative of the original dimensions of ruptured slip patches, seismological scaling relationships (e.g., Kwiatek et al., 2011) indicate 0.3 and 30 m² slip patches likely accommodated single-event displacements of ~10–40 μm and 2–4 mm, respectively, yielding Mw = -3.4 to 0.3. The average observed displacement (6.5 cm) likely represents slip accumulated over many events, supported by multiple overprinting slickenline orientations preserved on most surfaces. We calculate an upper bound on Mw
assuming that the average displacement reflects a single event. Theoretical relationships between slip and rupture area (Scholz, 2002; see Supplementary Material) indicate 6.5 cm of displacement would be associated with a slip patch of \( \sim 5340 \text{ m}^2 \), yielding \( M_w = 2.6 \). Reasonable \( M_w \) estimates for earthquakes accommodated on these fault surfaces are thus \( \sim -3.4 \) to 2.6. Following Eaton et al. (2016), seismic events at the lower end of this range of \( M_w \) (i.e., \( M_w = -3.4 \) to 0.3) correspond to nano- to milliseismicity. However, some fault mirrors may have hosted larger seismic events. Hematite and apatite He data patterns constrain the timing and depth of at least some of these earthquakes to post-4.5 Ma and \( \leq 2 \text{ km} \), although earlier slip at greater depths also likely occurred.

We reconstruct the rock record of paleoearthquakes from damage zone slip surfaces that are spatially and temporally correlated with the active, seismogenic Wasatch Fault system. The WFZ has produced \( \geq M_w 7 \) earthquakes every \( \sim 500 \)–\( 2500 \text{ years} \) through the Holocene (DuRoss et al., 2016 and references therein). Small earthquakes on hematite fault mirrors may represent aftershock clouds in response to larger seismic events. Recent studies suggest smaller earthquakes are self-similar with larger earthquakes and that similar physical processes control the genesis of both (e.g., Ide and Beroza, 2001; Kwiatek et al., 2011). Our data argue that flash heating of asperities is a viable mechanism for generating small earthquakes and we suggest this process may also promote larger seismic events. Consideration of dynamic weakening by asperity flash heating in models of earthquake rupture may inform the genesis and behavior of larger earthquakes along the Wasatch Front and in other fault zones.
8. Conclusions

Integrated microtextural observations, fault rock thermochronometry, and thermomechanical modeling quantify fault surface paleotemperatures and reveal a preserved rock record of seismicity preserved on WFZ hematite fault mirrors. Hematite aliquots with clusters of polygonal and lobate grains that yield He dates younger than apatite He data provide evidence of friction-generated temperatures at the fault surface, hematite recrystallization, and attendant He loss. Transient flash temperatures of >~700–1200 °C at frictional contacts on the slipping surface and subsequent weakening likely facilitated seismogenic slip. Hematite and apatite He date patterns constrain some of these events to post-4.5 Ma and at ≤2 km depth. The exhumed damage zone fault mirrors archive thermal and mechanical processes operative along a major normal fault in the western USA during footwall exhumation. Asperity flash heating is hypothesized as a weakening mechanism during earthquake genesis and propagation by laboratory and theoretical studies (e.g., Rice, 2006; Hirose and Bystricky, 2007; Beeler et al., 2008; Goldsby and Tullis, 2011). We provide evidence of this process occurring in a natural fault damage zone. If our documented weakening process scales with larger seismic events, asperity flash heating may be an important process in the propagation of ruptures associated with larger earthquakes.

References


Chicot, D., Mendoza, J., Zaoui, A., Louis, G., Lepingle, V., Roudet, F., Lesage, J., 2011. Mechanical properties of magnetite (Fe₃O₄), hematite (α-Fe₂O₃) and goethite (α-FeOOH) by instrumented indentation and molecular dynamics analysis. Materials Chemistry and Physics 129, 862-870.


CHAPTER 3

HEMATITE (U-TH)/HE THERMOCHRONOMETRY CONSTRAINS EPISODIC PRECIPITATION, FAULT REACTIVATION, AND EXHUMATION OVER 100 MYR-TIMESCALES IN THE WASATCH FAULT ZONE

ABSTRACT

Reactivation of pre-existing structures in brittle fault zones influences fault rheology and seismogenesis. Direct evidence for fault reactivation and the timescales over which it occurs are challenging to identify in polyphase tectonic settings. To address this issue, we present new field, structural, and microtextural observations, with hematite (U-Th)/He (HeHe), zircon (U-Th)/He (ZHe), apatite (U-Th)/He (AHe), and apatite fission-track (AFT) thermochronometry data from rocks now exposed in the extensional Wasatch fault zone (WFZ), UT. Iron-oxide alteration manifests as specularite veins and hematite-coated fault mirrors that cut exhumed Paleoproterozoic host rock and envelop mesoscale normal faults. Most mirrors are normal-sense but exhibit multiple striae orientations, are localized within pre-existing hematite-cemented breccia, and show microstructural evidence for coseismic slip. Texturally-characterized HeHe dates (38 aliquots, N=11 samples) of specularite, breccia matrix, and overlying mirrors are ~270–21 Ma, ~69–20 Ma, and ~42–14 Ma. Specularite and breccia dates are correlated with plate width (HeHe closure temperature). Host rock ZHe (N=1), AHe (N=1), and AFT (N=2) dates are ~5 Ma, ~5 Ma, and ~2.5 Ma. Thermal history models constrain ambient

---

1 Paper formatted for submission to the Geological Society of America Bulletin with author list:
R.G. McDermott and A.K. Ault
Utah State University
reheating (burial) and cooling (exhumation) episodes and suggest HeHe dates retain an isotopic memory of their formation. Hematite precipitation occurred at ~1680 Ma, ~110–75 Ma, and ~≤20–14 Ma, each correlated with regional tectonism. Specularite and breccia formed 100s of Ma prior to earthquakes that created fault mirrors. Integrated data inform a multi-phase history of hematite precipitation and fault reactivation, which ultimately influence the structural evolution and earthquakes in the WFZ.

INTRODUCTION

Brittle fault zones localize stress, strain, fluid flow, and heat transport in the crust. Reactivation of pre-existing structures within fault zones influences these processes; promotes strain localization; and may control the mechanical evolution of faults, crustal strength, and seismogenesis (Sibson, 1985, 1992; Crider and Peacock, 2004; Shipton et al., 2006). Fault reactivation is likely important in the energy budget for various fault zone mechanical processes (Shipton et al., 2006; Kanamori and Rivera, 2013). Structural reconstructions and geometric arguments (Marshak et al., 2000), fault surfaces with multiple striae of opposing slip sense (Marrett and Allmendinger, 1990), and re-brecciated cements in fault rocks (Eichhubl and Boles, 2000) are commonly reported as evidence of fault reactivation. However, the timescales over which this phenomenon occurs are challenging to quantify (Power and Tullis, 1989; Holdsworth et al., 1997, 2001; Eichhubl and Boles, 2000; Caine et al., 2010). Recently, geochronology of fault gouge, fault veins, and slickenfibres suggests fault reactivation is permissible on 10,000 yr through 1 Myr timescales (Ault et al., 2016; Williams et al., 2017; Nuriel et al., 2019; Scheiber et al., 2019; Tillberg et al., 2020). The extent to which strength heterogeneities,
which may serve as loci for fault reactivation, persist over multiple phases of deformation or orogenic events remains unclear.

The Wasatch fault zone (WFZ) in northeastern Utah is an ~400 km-long, seismogenic normal fault trending through northeastern Utah, USA (Fig. 1; Machette et al., 1991). The WFZ demarcates the eastern extent of the Basin and Range physiographic province and associated extension (e.g., Wallace, 1984). Rocks incorporated within the damage zone of this major structure have experienced multiple phases of deformation throughout geologic time (Bryant, 1988; Yonkee, 1992; Evans and Langrock, 1994). Understanding the interaction of pre-existing and new damage development and faulting has implications for the strength, fluid flow properties, and structural history of the WFZ. Hematite is a common secondary mineral in exhumed WFZ footwall rocks, occurring in specularite veins and as networks of high-gloss, reflective fault surfaces (fault mirrors) in the damage zone. This mineral is particularly useful because it can exhibit different microtextures suggestive of different slip rates and deformation style(s) and is amenable to hematite (U-Th)/He (HeHe) thermochronometry (Ault, 2020). HeHe thermochronometry is a useful technique for constraining the timing of ambient cooling and exhumation, (a)seismic slip, or fluid flow in fault zones (Ault et al., 2015, 2016; McDermott et al., 2017; Moser et al., 2017; Calzolari et al., 2018, 2020; Ault, 2020). However, linking HeHe data to fault-surface specific processes (versus simple ambient cooling) requires limited exhumation and(or) that the timing of faulting/fluid flow post-dates significant exhumation (McDermott et al., in prep).

We integrate field observations, structural analyses, and microtextural observations with HeHe thermochronometry from hematite in veins and fault mirrors in
the WFZ, and zircon (U-Th)/He (ZHe), apatite fission-track (AFT), and apatite (U-Th)/He (AHe) thermochronometry of the host rock to constrain the timing of alteration, fault slip, and exhumation in the WFZ. High-spatial resolution hematite sampling and thermal history models are used to isolate textural and thermochronometric signatures of different thermal processes, including hematite precipitation, burial, exhumation, fault reactivation, and deformation preserved in veins and fault mirrors. We show hematite mineralization along the WFZ has spanned 100s of Ma. Hematite veins and hematite-cemented breccia serve as small-scale strength heterogeneities that are repeatedly reactivated through geologic time, localize subsequent fluid flow and strain, and influence development of the Miocene-present WFZ.
Figure 1 (previous page): Geology of northeastern Utah, USA. (a) Simplified geologic map with key tectonic elements denoted. Inset shows location of (b). Modified after Yonkee (2005). (b) Simplified geologic map of our study area. White circles show new and previously published hematite sample locations. Brown circles show location of new and previously published bedrock conventional thermochronometry data. See legend for symbology denoting data type. Subscripts adjacent to sample labels give data source. 1This study, 2Armstrong et al. (2004), 3Ault et al. (2015), 4McDermott et al. (2017), 5Yonkee et al. (2019). Labeled white squares in (a) and (b) show location of previously documented hematite mineralization. Geology modified after Crittenden and Sorensen (1985). (c) Cross section from X-X’ in (b). No vertical exaggeration. After Crittenden and Sorensen (1985).

ANALYTICAL BACKGROUND

Hematite (U-Th)/He thermochronometry in fault rocks

Hematite incorporates U and Th, but negligible initial He, into its crystal lattice upon crystallization and retains radiogenic $^4$He in the crystal structure over geologic timescales (e.g., Bähr et al., 1994). The temperature at which radiogenic $^4$He is retained, or closure temperature ($T_c$; Dodson, 1973) ranges from ~25–250 °C (Farley and Flowers, 2012; Evenson et al., 2014; Farley, 2018). The HeHe $T_c$ is largely controlled by the size of individual hematite crystals (Farley, 2018; Jensen et al., 2018). Natural hematite is commonly a polycrystalline aggregate and the bulk $T_c$ is thus a function of grain size distribution (Evenson et al., 2014). The large variability in $T_c$ of the HeHe system enables broad application in tracking upper crustal processes, including dating hydrothermal or diagenetic precipitation (Wernicke and Lippolt, 1993; Farley and Flowers, 2012; Reiners et al., 2014; Ault et al., 2016), regional cooling and exhumation (Evenson et al., 2014; Farley and McKeon, 2015; Jensen et al., 2018), and fault slip (Ault et al., 2015; McDermott et al., 2017).
Hematite in fault-related rocks may experience a range of thermal processes, such as fluid circulation (and associated new mineral precipitation), fault slip, ambient cooling, or their combination (Ault et al., 2015, 2016; McDermott et al., 2017). Hematite precipitation at temperatures greater than a sample’s specific $T_c$ results in a HeHe date that records host-rock ambient thermal history (e.g., burial and/or exhumation; Calzolari et al., 2018). Precipitation at depths shallower than the HeHe $T_c$ may reflect the timing of precipitation (Ault et al., 2016; Moser et al., 2017). At this same depth range, additional He loss caused by deformation or other non-monotonic cooling processes may also be reflected in bulk dates (Ault et al., 2015; McDermott et al., 2017). Grain size (and $T_c$) reduction by cataclasis may further complicate these scenarios (Ault et al., 2015). The presence of fluids in fault zones can result in precipitation of interstitial U-bearing minerals or U addition to hematite that may impact HeHe dates (Evenson et al., 2014; Reiners et al., 2014). Interpretation of HeHe dates from fault rocks requires constraints on host-rock ambient thermal history (such as other thermochronometers with overlapping $T_c$), data informing grain size ($T_c$) distribution, and microtextural observations on deformation fabrics (e.g., Ault et al., 2020).

**Apatite and zircon (U-Th)/He and apatite fission-track thermochronometry**

The AHe and ZHe thermochronometers have $T_c$ of ~30–120 °C and 25–220 °C, respectively, and thus track exhumation of rocks through the upper ~10 km of the crust (assuming a 25 °C/km geothermal gradient; Zeitler et al., 1987; Wolf et al., 1998; Ehlers and Farley, 2003; Reiners et al., 2004; Flowers et al., 2009; Guenthner et al., 2013). Grain size is a secondary control on $T_c$ in the AHe and ZHe systems (Farley, 2000;
Reiners et al., 2004). The accumulation and annealing of intracrystalline radiation damage are a primary control on He diffusion kinetics and effective Tc in these systems (Flowers et al., 2009; Guenthner et al., 2013). Radiation damage is induced by parent isotope actinide decay, but anneals with increasing temperature, resulting in evolving effective Tc over a sample’s thermal history (Flowers et al., 2009; Guenthner et al., 2013). Increasing radiation damage yields greater He retentivity and effective Tc in the apatite He system and low-damage zircon. In highly damaged zircon, He diffusivity is enhanced after accumulated damage becomes interconnected, and effective Tc decreases (Guenthner et al., 2013). In grains that experienced the same thermal history, effective uranium (eU; [U] + 0.235[Th]) concentration serves as a proxy for accumulated radiation damage. AHe data may exhibit positive date-eU trends, whereas ZHe data may exhibit both positive and negative date-eU trends, depending on thermal history (Flowers et al., 2009; Guenthner et al., 2013).

The AFT thermochronometer depends on the natural production and temperature-dependent annealing of tracks created by fission of $^{238}$U (Naeser, 1967). Effective Tc in this system ranges from ~70–120 °C (Ketcham et al., 2007). Annealing kinetics vary with grain chemistry (Carlson et al., 1999; Ketcham et al., 2007). Etched track diameter (Dpar) varies with grain chemistry and is thus also correlated with annealing and Tc. Because tracks anneal by shortening, track length distributions provide additional constraints on thermal history (Gleadow et al., 1986, 1986).

**GEOLOGIC SETTING**

*Geology and tectonics of the Wellsville and Wasatch Mountains*
The Wellsville and Wasatch Mountains reflect the culmination of a protracted geological history. The oldest exposed rocks are gneiss, migmatite, amphibolite, and pegmatite of the Farmington Canyon Complex (Fig. 1; Crittenden and Sorensen, 1985; Bryant, 1988). The Farmington Canyon Complex is interpreted as Archean (~2.45 Ga) passive margin and(or) rift-related sedimentary, plutonic, and volcanic rocks that formed along the edge of the Wyoming Province (Hedge et al., 1983; Bryant, 1988; Mueller et al., 2011). These rocks experienced peak upper amphibolite facies metamorphism (~600–700 °C and 5–8 kbar) and partial melting in the Paleoproterozoic (~1.7 Ga), associated with subduction and accretion of this and other basement complexes to the southwestern margin of Laurentia (Hedge et al., 1983; Barnett et al., 1993; Nelson et al., 2002; Mueller et al., 2011).

Basement rocks of the Farmington Canyon Complex are overlain by Paleozoic and Mesozoic strata, which are in turn complexly folded and faulted (Fig. 1). Farmington Canyon Complex rocks are unconformably overlain by ~2 km of middle Cambrian through Devonian, westward-thickening, strata formed during passive margin sedimentation along the western margin of North America (Fig. 1; Crittenden, 1972; Peterson and Clark, 1974; Crittenden and Sorensen, 1985). An additional ~1.5 to 2 km of Mississippian through Permian strata record subsidence, ocean basin sedimentation, and local syn-sedimentary faulting, potentially in relation to the Antler and Ancestral Rocky Mountain orogenies (Crittenden, 1972; Jordan and Douglass, 1980; Geslin, 1998). Approximately 60 km east of the Wasatch Mountains, terrestrial Triassic and Jurassic strata are preserved in the footwall of the Sevier and Laramide thrust faults, suggesting Mesozoic strata may have also once overlain our study area (Fig. 1a; Yonkee, 2005).
Our study area lies in the footwall of the Willard thrust (Fig. 1b, c; Crittenden and Sorensen, 1985). Displacement on the Willard thrust from ~140 Ma to ~90 Ma transported a ≥10–15 km-thick sequence of Neoproterozoic through Permian miogeoclinal strata ≥50 km eastward, resulting in tectonic burial and reheating of footwall rocks (Fig. 1a; Crittenden, 1972; Crittenden and Sorensen, 1980, 1985; Yonkee et al., 1989, 2019; Yonkee, 1992). Following Willard thrust activity, deformation continued to shift eastward, causing duplexing, rock uplift, and exhumation of previously buried footwall rocks until ~50 Ma (Naeser et al., 1983; Yonkee, 1992; Decelles, 1994). This latter phase of deformation also led to development of the Wasatch Anticlinorium (Fig. 1a; Bryant, 1984; Yonkee, 1992; DeCelles, 1994).

The present-day physiography of the Wasatch mountains is in part the product of the range-bounding WFZ associated with Miocene-present Basin and Range extension (Fig. 1; Crittenden and Sorensen, 1985). Our study area is within the Brigham City segment of the WFZ (Fig. 1b; Crittenden and Sorensen, 1985; Machette et al., 1991). Available estimates for the timing of initiation of extension and exhumation along the WFZ are from the Salt Lake City and Weber segments and range from ~18 Ma to ~10 Ma, based on K-Ar dating of hydrothermal sericite and low-temperature thermochronometry data (Naeser et al., 1983; Parry and Bruhn, 1986; Kowallis et al., 1990; Ehlers et al., 2003). The southernmost portion of the Brigham City segment of the WFZ has accommodated ~2.8 km of stratigraphic separation since its Miocene inception (Personius et al., 2012). Paleoseismic trenching reveals characteristic Mw ~7.0 earthquakes at a recurrence of ~1060-1500 yrs, although partial segment ruptures
initiated on the Weber segment to the south are also documented. Calculated Pleistocene-present slip rates are ~1.3 mm/yr (Personius et al., 2012; DuRoss et al., 2016).

**Previous field study and hematite (U-Th)/He thermochronometry results**

Prior field study of the southernmost Brigham City segment of the WFZ documents an ~300-400 m-thick zone of deformed rock and locally abundant hematite alteration adjacent to the mapped trace of the WFZ (Evans and Langrock, 1994). Fault rocks include ductilely deformed chlorite-rich phyllonites and weakly foliated chlorite breccia, cut by localized bands of Fe-oxide-rich and highly-weathered cataclastic rocks, which are in turn cut by hematite fault mirrors. West-dipping and down-to-the-west chlorite phyllonites are the oldest fault rock assemblage and interpreted as resulting from fluid-assisted plastic deformation during Basin and Range extension, although similar shear zones with the Farmington Canyon Complex have been interpreted as recording reverse faulting during the Sevier Orogeny and subsequent rotation by footwall tilt (Evans and Langrock, 1994; Yonkee, 1992). Cataclastic rocks and hematite fault mirrors preserve cm-scale displacement and dip- and strike-slip striae generally consistent with east-west extension, likely associated with deformation along the WFZ. Fault mirrors are commonly collocated with, or underlain by, specularite veins (Evans and Langrock, 1994; Ault et al., 2015; McDermott et al., 2017). Mirror-forming hematite appears comminuted, suggesting mirrors develops from precursor hematite (Ault et al., 2015; McDermott et al., 2017).

Previous microtextural and geochemical data suggest evidence for transient, localized elevated temperatures during fault mirror development (Evans et al., 2014; Ault
et al., 2015; McDermott et al., 2017). Hematite fault mirrors comprise ~200–1000 μm-thick bands of hematite cataclasite and underlying specularite plates (Ault et al., 2015; McDermott et al., 2017). Many fault mirrors preserve ~5–10 μm-thick zones of hematite with polygonal and triple-junction forming or lobate grain boundaries (Ault et al., 2015; McDermott et al., 2017). These particle morphologies are similar to those observed in dry heating and high-temperature torsion experiments (Siemes et al., 2003; Vallina et al., 2014), suggesting fault mirrors experienced transient elevated temperatures associated with seismic slip. Trace Fe$^{2+}$ on fault mirrors, characteristic of high-temperature reduction, also supports past hot temperatures and seismic slip along fault mirrors (Evans et al., 2014).

HeHe and AHe thermochronometry, combined with microtextural analysis, document evidence for frictional heating and associated He loss during seismic slip on hematite fault mirrors (Ault et al., 2015; McDermott et al., 2017). HeHe dates are generally younger than AHe data from adjacent, unaltered bedrock despite higher hematite $T_c$. Hematite aliquots yielding the youngest HeHe dates have high-temperature crystal morphologies (Ault et al., 2015; McDermott et al., 2017), interpreted as reflecting recrystallization and He loss due to flash heating at geometric asperities during seismic slip. HeHe and AHe data patterns bracket the timing of this He loss, and thus the most recent earthquake events, to <4.5 Ma (McDermott et al., 2017). Initial HeHe dates from two specularite veins are Late Cretaceous to Miocene (McDermott et al., 2017). These dates reflect the minimum ages of hematite precipitation and suggest multiple generations of hematite, which we explore further in this work.
METHODS

Geologic mapping and structural analyses

We present new geologic mapping, field observations, and structural analyses of hematite alteration in the Brigham City segment of the WFZ and its relation to broader damage zone architecture. We specifically focus on two study locales, Tower Canyon and Pearsons Canyon (see Figure 1 for location), where drainage development leads to exceptional exposure and hematite veins and fault mirrors are present in high-density. Mapping along transects was completed on orthorectified aerial image basemaps provided by the National Agriculture Imagery Program and at 1:1,000 scale. At select outcrops along these transects, we measured the distribution and intensity of fracture, specularite vein, and hematite fault mirrors over ~1 m-increments and approximately parallel to the average pole of planar surfaces (Hudson and Priest, 1983).

We measured the attitudes of hematite veins and collected fault slip data from hematite fault mirrors where observed in the field. Detailed structural analyses are presented in Evans and Langrock (1994). Our goal is to obtain detailed, sample-specific field data that can be directly correlated with subsequent HeHe thermochronometry. Structural data from hematite veins were compiled using Stereonet (v. 9.3) software (Cardozo and Allmendinger, 2013). We use 1% area contouring of poles to vein orientations and calculate mean orientations of these clusters with Fisher statistics.

Fault slip data were collected as fault surface and striae attitude, sense-of-slip, and sense-of-slip confidence level. Sense-of-slip was determined and given quality rankings using observable offsets and(or) kinematic indicators outlined in Petit (1987). In our study area, the most common sense-of-slip indicators are offset gneissic foliations, Riedel
shears or other secondary fractures, and asymmetric grooves resulting from ploughing of cm-scale asperities. Multiple shear sense indicators were collected where possible. Clear offsets resulted in an “A” ranking, secondary fractures resulted in a “B” ranking, and those from asymmetric asperity profiles or otherwise more ambiguous indicators were given a “C” ranking.

Fault slip data are also presented as kinematic inversions using FaultKin (v. 8.1) software (Marrett and Allmendinger, 1990). This approach calculates extension (T) axes for each fault-striae pair by bisecting a plane containing the pole to the fault and striae (movement plane; Aleksandrowski, 1985). Compression (P) axes lie along this same plane at 90° to the T axis. Populations of P and T axes are analyzed with Bingham distribution statistics to calculate directional maxima, or average incremental strain axes, and corresponding pseudo-focal-mechanism solutions. We assume fault populations are scale invariant and apply equal weighting to each data point, although sense-of-slip confidence was used to identify potential outliers (e.g., Marrett and Allmendinger, 1991).

**Microtextural analysis**

We analyzed hematite samples with multi-scale microscopy to characterize grain size distribution and microtextures. Aliquots were isolated from samples via Dremel™ tool, mounted in epoxy, and hand-polished. We imaged samples by reflected light and scanning electron microscopy (SEM) at the Utah State University Microscopy Core Facility. Mounts were dual imaged with secondary electron (SE) and backscatter electron (BSE) detectors at magnification ranging from 100,000x to 500x to highlight crystal morphologies and identify atomic/mineralogical contrasts correlated to distinct
microtextural domains. SEM images were taken under high-vacuum (<10⁻⁶ Torr), 20.0 kV accelerating voltage, 8-10 mm working distance, and 1.0 A current.

Grain size distributions for each sample were measured from a combination of reflected light (for coarse-grained samples) and SEM images with ImageJ software (Schneider et al., 2012). A minimum of 70 measurements were made per sample. Because our ultimate goal is to characterize Tc in our samples, we present grain size as plate-width (e.g., Ault et al., 2015; McDermott et al., 2017; Jensen et al., 2018). We converted grain size measurements to Tc following Dodson (1973) and assuming a spherical diffusion domain with a radius corresponding to plate half width, a monotonic cooling rate of 10 °C/Myr, and the diffusion kinetics of Farley (2018), consistent with prior HeHe thermochronometry studies (Ault et al., 2015; McDermott et al., 2017; Jensen et al., 2018; Farley, 2018).

**Hematite, apatite, and zircon (U-Th)/He and apatite fission-track thermochronometry**

Hematite chips for HeHe thermochronometry were extracted from samples via Dremel™ tool at the Utah State University Mineral Microscopy and Separation Laboratory (USU M²SL). Under reflected light, chips were broken into several replicate aliquots. A subset of these aliquots was pre-screened prior to HeHe analysis via reflected light for coarse-grained samples or SEM for finer-grained samples (McDermott et al., 2017). Measurements of U, Th, and He were made at the Arizona Radiogenic Helium Dating Laboratory (ARHDL) at the University of Arizona.
Apatite and zircon were extracted from samples of unaltered Farmington Canyon Complex host rock with standard mineral separation techniques. Inclusion-free and prismatic apatite and zircon encapsulating a range of metamictization were hand-selected under cross-polarized light and loaded into Nb tubes. In selecting zircon, we followed the protocol of Ault et al. (2018) and purposefully selected crystals exhibiting variable opacity and discoloration to maximize the spread in radiation damage, and thus effective $T_c$, encapsulated by our samples. Measurements of U, Th, and He were conducted at the ARHDL. AFT data were acquired by the External Detector Method (Gleadow, 1981) at the University of Arizona Fission Track Laboratory. Detailed HeHe, AHe, ZHe, and AFT analytical methods are provided in the Supplemental Information.

RESULTS
Field mapping and structural analyses

New mapping results and field observations are presented in Figures 2 and 3. In our mapping, we primarily focused on the distribution and slip sense of mesoscale faults and hematite fault mirrors and documenting the extent of the damage zone. At both Tower and Pearsons Canyon localities, host rock units include Paleoproterozoic gneiss and migmatite of the Farmington Canyon Complex (Figs. 2; 3a; see also Crittenden and Sorensen, 1985). The dominant mineralogy of these units is quartz, orthoclase, and plagioclase feldspar with $\sim$10% (Xfcg) –30% (Xfcq) hornblende and biotite, similar to prior description of the Farmington Canyon Complex (Crittenden and Sorensen, 1985; Bryant, 1988). Felsic minerals are segregated into gneissic banding that defines a

---

2 Supplemental Information for this chapter is available in Appendix A
moderately west-dipping foliation. Gneisses and migmatites are cut by ~1 to >5 m thick, tabular bodies of hornblende-plagioclase-phlogopite amphibolite (Xfca) and granite pegmatite. Amphibolite and pegmatite bodies are typically partially to fully transposed into the regional foliation (Figs. 2; 3a).

We define and map a damage zone unit (Dz) as a zone of enhanced fracture density and hydrothermal alteration overprinting Farmington Canyon Complex host rocks. This unit comprises chlorite breccia and protobreccia, punctuated by zones of cataclasite and protocataclasite (Fig. 3a and inset; Evans and Langrock, 1994). Similar to Evans and Langrock (1994), we observe weakly-developed west-dipping foliation at some locales (Fig. 3a, inset). Chlorite phyllonite units documented elsewhere within our broader study area (Evans and Langrock, 1994) are not present in either detailed mapping locales. Quaternary units include Holocene colluvium and alluvial and lacustrine deposits of Pleistocene Lake Bonneville (Fig. 2). More detailed descriptions of the Quaternary geology of our study area are presented by Personius (1990).
Figure 2: Geologic maps of the Wasatch fault damage zone. Maps show distribution of Wasatch fault damage zone elements and samples at Tower (left) and Pearsons Canyon (right) locales. Location of new (white) and previously published (gray; Ault et al., 2015; McDermott et al., 2017) hematite samples shown.
Figure 3: Field photographs of Wasatch fault damage zone and hematite mineralization. (a) Panorama of the wall of Pearsons Canyon showing damage zone (Dz) rocks (Figure 2) and host rock gneiss (Xf cg) and amphibolite (Xf ca; red dashed lines). White dashed lines denote outcrop-scale faults. Foliation trajectories shown by solid colored lines (Dz-green, Xf cg-pink). Inset shows detail of Dz weakly foliated chlorite protobreccia and protocataclasite. (b) Detail of synthetic and antithetic normal faults on ridge of Pearsons Canyon. Red dashed lines outline offset amphibolite bodies. (c) Specularite veins (likely conjugate) associated with coarse-grained pegmatite. Qtz-quartz, fsp-feldspar. (d) Hand sample photograph of fault mirror W17-1G with multiple striae and underlying hematite-cemented breccia. White arrows show striae trend and
hanging wall slip sense. Note strike and dip symbol. (e) Field photograph of fault mirror with underlying specularite and pegmatite and multiple striae. (f) E-W trending specularite vein crosscut and offset by N-S trending breccia and fault mirror. (g) N-S striking specularite vein cutting E-W trending vein. Red arrow denotes intersection.

Damage zone architecture varies between Tower and Pearsons Canyon locales. The width of the damage zone, although partially obscured by Quaternary cover, is ~130 m wide at Tower Canyon and ~250 m wide at Pearsons Canyon (Fig. 2). At both locales, the contact between damage zone and host rock is gradational over <10 m (Fig. 3a). Subsidiary faults are synthetic, antithetic, and subparallel to the main trace of the WFZ. These structures have trace lengths ranging from ~>25 to >50 m and locally involve zones of gouge and(or) cataclasite and breccia, flanked by zones of enhanced fracture density (Fig. 3a). Faults that offset distinct markers (typically amphibolite or pegmatite) bodies show ~5 to ~20 m of normal-sense vertical separation (Figs. 2, 3a, 3b). Most faults documented at map scale are normal faults. Reverse faults are rare and only observed at scales below that of our mapping. Mesoscale faults and collocated damage zone rocks together define an anastomosing geometry that encapsulates pods of less-deformed and altered host-rock (note pods of Xfcg surrounded by damage zone, Dz, in Fig. 2), similar to other fault damage zones (e.g., Caine et al., 2010; Faulkner et al., 2010). Lastly, subsidiary faults with clear normal offset persist outside of what we formally define as the enhanced zone of fracturing and alteration associated with our damage zone unit (Figs. 2; 3a, b).

We observe various forms of Fe-oxide in our study area. Diffuse Fe-oxide alteration is pervasive within the damage zone, and particularly concentrated along mesoscale faults (Fig. 3a and inset). Consistent with prior work, we observe two primary
modes of hematite occurrence: specularite veins and fault mirrors (Fig. 3c-g; Evans and Langrock, 1994; Evans et al., 2014; Ault et al., 2015; McDermott et al., 2017). Specularite veins typically comprise nearly pure, coarse-grained hematite, are ≤1 mm-thick, and form a sharp contact with adjacent Farmington Canyon Complex host rock (Fig. 3c; g). Specularite veins are commonly, but not exclusively, associated with granite pegmatite pods and veins (Fig. 3c, e). Hematite fault mirrors are similar to those described in prior work (Evans et al., 2014; Ault et al., 2015; McDermott et al., 2017), and many are underlain by specularite or hematite-cemented breccia (Figs. 3e, d; S1). Breccias comprise mm-scale clasts of gneiss wall-rock in hematite matrix, are matrix-supported, and range in thickness from ≤1 mm to >1 cm (Figs. 3d; S1).

Fracture, vein, and fault mirror density measurements and structural data are superimposed on cross sections in Figure 4. Density results indicate variable distribution of hematite in the WFZ damage zone (Fig. 5). Specularite veins and fault mirrors (including mirrors with underlying specularite or breccia) range in density from <1/m to >40/m. We observe five spatial patterns in hematite occurrence. First, higher densities of specularite and fault mirrors are collocated. Second, there are no outcrops where only one type of hematite feature is observed. Third, at Pearsons Canyon, hematite is sporadically distributed and shows no spatial relation to the main trace of the WFZ, although peaks in vein and fault mirror density are associated with mesoscale faulting in the damage zone. At Tower Canyon, vein and mirror density is highest near the modern trace of the WFZ, although drops rapidly at ~130 m from the WFZ’s main trace. Fourth, the highest hematite density is observed in unaltered rock at both locales near where mesoscale structures tip out. Fifth, the density of unmineralized fractures systematically decreases
with distance from the WFZ, typical of other damage zones (e.g., Schulz and Evans, 2000; Mitchell and Faulkner, 2009).
Figure 4 (previous page): Structural geology of Wasatch fault zone hematite mineralization. Cross sections showing structural detail of Tower (top) and Pearson’s (bottom) Canyons. No vertical exaggeration. See Figure 2 for location of A-A’ and B-B’ lines. Density of specularite veins (white triangles), fault mirrors (white circles), and unmineralized fractures (black circles) from scan lines shown. White stars show location of field stations projected onto cross sections. Gray bars denote areas of higher hematite mineralization. Equal area plots show fault mirror and vein(breccia)-fault mirror pair attitudes and striae divided by location and slip sense. Black-normal faults, green-reverse faults. Shaded circles show fault slip data for previously published and new samples analyzed by (U-Th)/He (see keys next to each plot). Tension (T) and compression (P) axes, mean kinematic axes, and pseudo-focal mechanism solutions from FaultKin overlain. Focal mechanism solutions shown separately for normal and reverse faults. Shaded areas (gray or green) correspond to extensional domains, white areas are compressional domains.

We collected a total of 84 structural measurements from specularite veins and 71 measurements (planes and striae) from fault mirrors at our two detailed mapping locations (Figs. 4, 5). We identify three dominant specularite vein orientations at Tower Canyon. Two sets of N-S striking veins are subparallel to the local WFZ trace, dip E and W, and yield an interplane angle of ~43°, broadly consistent with the expectation for conjugate veins (Fig. 5a). A third vein set strikes E-W and dips shallowly SSW (Fig. 5a). At Pearsons Canyon, two sets of N-S striking veins dip steeply E and W and have an interplane angle (35°) consistent with conjugate geometry (Fig. 5b). Two sets of E-W striking veins dip steeply N and S and similarly exhibit conjugate geometry (32° interplane angle; Fig. 5b). A third Pearsons vein set strikes NW-SE and moderately dips SW (Fig. 5b), subparallel to the local trace of the WFZ (e.g., Figs. 1; 2). Generally, mean specularite vein orientations are consistent between Tower and Pearsons canyons.
Figure 5: Equal area plots of hematite elements at Tower (top) and Pearsons (bottom) canyons. (a, c) Poles to specularite veins (gray squares), poles to mean orientations (blue squares), and corresponding mean vein (blue lines). Labeled black dashed arrows show angle between select planes. (b, d) Poles to fault mirrors (black circles) and striae trend/plunge (green circles). Best fit girdle (green lines) to striae shown. In all plots, contouring is 1% and color-coded to corresponding dataset. Note local trace of WFZ (ticks with ball-and-bar on downthrown side). Plots created with Stereonet software (Cardozo and Allmendinger, 2013).

Fault mirrors are variably oriented and typically multiply striated (Fig. 3d, e). At Tower Canyon, fault mirrors strike mostly N-S and accommodate almost exclusively normal or oblique normal slip (Fig. 4a; 5c). Striae define a best-fit girdle that parallels the mean trace of the WFZ (Fig. 5c). Pseudo-focal mechanism solutions for these data are consistent between multiple field stations at Tower Canyon and show E-W trending and subhorizontal extension (Fig. 4a). At Pearsons Canyon, fault mirrors are typically NW-
SE or SW-NE trending with variable dip, subparallel and perpendicular (respectively) to the local WFZ trace (Fig. 5d). Striae define NW-SE- and SW-NE-striking girdle distributions, with the maximum striae directions lying near the intersection of the best-fit girdles (Fig. 5d). Pearsons Canyon fault mirrors exhibit normal, reverse, and strike-slip kinematics, occasionally on the same mirror surface (Figs. 3d, e, f; 4b). Pseudo-focal mechanism solutions show fault mirrors nearest the WFZ main trace and eastern margin of the damage zone have predominantly normal or oblique normal slip with subhorizontal SSE and SSW trending extension directions, respectively (Fig. 4b). Solutions from Xf gc outcrops within and outside of the damage zone exhibit both normal (in gray, Fig. 4b) and reverse slip (in green, Figure 4b). Normal-sense solutions show subhorizontal and N-S or E-W trending extension, whereas reverse faults show subhorizontal, SW trending contraction and sub-vertical extension (Fig. 5). We note that, although our dataset is smaller than that presented by Evans and Langrock (1994), our fault attitude and striae measurements are in good agreement with their results.

Key cross-cutting relationships are observed in the field. At both Tower and Pearsons canyons, E-W specularite veins are consistently cut by N-S trending veins (e.g., Fig. 3g). We also observe E-W trending specularite veins cut by N-S trending fault mirrors and(or) fault mirrors with underlying breccia (e.g., Fig. 3f). On fault mirrors with multiple striations, relative striae ages are difficult to establish, but normal slip is typically youngest (e.g., Fig. 3d, e).
Hematite Samples and Microtextural analyses

Samples

We present microtextural observations from seven samples, corresponding to those ultimately targeted for HeHe thermochronometry (see Figure S1 for hand sample photographs). We specifically target specularite veins and fault mirrors with underlying breccia visible in the field such that discrete domains can be isolated for later HeHe dating. We analyzed a total of three samples from specularite veins (W17-1F, W17-1E, W16-21, W16-6), specularite from a dilational jog in a fault surface (W17-2), and three fault mirrors with underlying breccia (W17-1G, W17-1H, W17-1J). Samples were chosen based on hematite purity sufficient for HeHe thermochronometry (discussed below), but samples with robust field context were prioritized. All samples are from Pearsons Canyon (Figs. 1; 2), except for vein sample W16-6, which is from Tower Canyon. These specularite veins are E-W striking, aside from N-S striking vein sample W16-6. Fault mirror samples W17-1G and W17-1H are E-W striking, whereas W17-1J is N-S striking. Fault mirrors samples all exhibit normal or oblique normal slip (annotated on Figure 4b), but W17-1G exhibits both oblique reverse and normal slip (Fig. 3d).

Microtextural observations and closure temperature estimates

Representative microtextures of our analyzed samples are shown in Figure 6. A full catalogue of our SEM images is provided in the Supplemental Information (Fig. S2). Specularite veins exhibit complex boundaries with pegmatite and(or) gneiss host rock at the micron scale. For example, in sample W16-21, hematite and quartz/feldspar grain boundaries are strongly embayed, with hematite inclusions occurring within feldspar
crystals (Fig. 6a). Specularite comprises undeformed and high aspect ratio plates ranging an order of magnitude in size (i.e., ~1 μm x >10 μm compared to ~10 μm x >100 μm; Fig. 6a, b, c). In some samples, we observe smaller plates along vein margins and larger plates in vein centers (Fig. 6b), however most samples consist of intermixed larger and smaller plates with no distinct fabric (Figs. 6c; S2). Specularite occurring within and along embayed margins of K-feldspar (e.g. Fig. 6a) is texturally similar to prior work showing Fe$^{3+}$ exsolution may occur during feldspar crystallization (e.g., Hofmeister and Rossman, 1984; Sun et al., 2013). Specularite with discrete grain size domains (e.g., Fig. 6b) may reflect multiple precipitation episodes, host-rock-fluid temperature gradients, or vein formation by a crack-seal mechanism.

SEM observations of fault mirrors with underlying breccia with a hematite matrix reveal two distinct microtextural domains: (1) breccia with sub-mm angular clasts of gneissic host rock supported by a matrix of high aspect ratio hematite plates (~2 μm x >10 μm), in sharp contact with (2) an ~200 μm-thick zone of relatively homogenous and deformed hematite (Fig. 6d, e). The latter domain has the fault mirror surface at its upper boundary. Although “fault mirror” describes an optical property, it is created and defined by a volume of material and we hereafter refer to this material as “fault mirror volume” in a microtextural context. Fault mirror volume material is dominantly cataclastic and comprises subangular and equant hematite (~0.1–0.5 μm in diameter) interspersed with high-aspect ratio plates (~2 μm x >5 μm; Fig. 6f, g). Clasts of underlying breccia are also observed (Fig. 6e). In sample W17-1H, there is also evidence for multiple generations of post-brecciation cataclasite in the fault mirror volume (Fig. 6e). Microtextures within ~5–10 μm of the fault mirror surface are variable, ranging from ultracataclasite (W17-1H;
Fig. 6i) to triple-junction-forming polygonal grains similar to those previously documented in other WFZ fault mirrors (W17-1G; Fig. 6h; Ault et al., 2015; McDermott et al., 2017). The sharp boundary between fault mirror volume material and underlying breccia, with deformed versus euhedral hematite in these respective domains, and clasts of breccia included in the fault mirror volume collectively indicate hematite-cemented breccia predate overlying fault mirrors.

Calculated $T_c$, informed by sample-specific grain size distributions, shows large intra- and intersample variation (Fig. S3). In specularite veins, we estimate median $T_c$ ranging from $\sim$130–150 °C, with minimum and maximum estimates of $\sim$110 and 170 °C, respectively. Hematite in the breccia matrix has a median $T_c$ of $\sim$115 °C that is consistent across samples, but $T_c$ ranges from $\sim$100 °C to 125 °C. Material in the fault mirror volume from these same samples has median $T_c$ values of $\sim$105–110 °C, and minimum and maximum values of $\sim$90 and 130 °C.

**Thermochronometric data**

*Hematite aliquot selection and sampling*

We acquired a total of 10 HeHe dates (38 individual aliquots) from the seven samples described above. We chose aliquots to minimize non-He-retentive interstitial phases (e.g., quartz) that would result in anomalously young dates not related to any thermal process (Evenson et al., 2014). There was no available $\alpha$-ejection correction factor for hematite when these data were collected (see Huber et al., 2019). We selected aliquots $\geq$120 μm in size and away from non-He-retentive interfaces such that He loss
due to ejection was likely balanced by implantation (i.e., Farley and Flowers, 2012; Evenson et al., 2014).

We targeted distinct microtextural elements for HeHe thermochronometry analyses. “Breccia matrix” aliquots contain exclusively hematite from breccia domains (e.g., Fig. 6d, e). In fault mirror samples W17-1H and W17-1J, we isolated the fault mirror volume (W17-1HA, W17-1JA) and underlying breccia matrix (W17-1HB, W17-1JB). In fault mirror sample W17-1G, we were unable to isolate breccia hematite of sufficient purity and only report a date for fault mirror material (W17-1GA). Given α-ejection considerations and the dimensions of different textural domains, “fault mirror” aliquots necessarily predominantly comprise what we describe above as fault mirror volume material above (Fig. 6d-g). Although care was taken to avoid sampling breccia matrix for these aliquots, some may also contain a portion of underlying breccia material. Due to plate size in our specularite vein samples, we were not typically able to isolate individual plates and the majority of aliquots are thus polycrystalline. The exception to this is sample W17-2, where we isolated a single ~40 μm-thick plate from within several hundred μm of surrounding hematite (although other aliquots for this sample are polycrystalline).
**Figure 6 (previous page): Hematite microtextures.** (a) Backscatter electron (BSE) image of specularite vein sample W16-21 showing hematite (bright hues) as inclusions and embayed with pegmatite (gray tones). (b) BSE image of W16-21 showing distinct size domains (black dashed line) potentially indicative of multiple hematite precipitation events. (c) BSE image of W17-1F showing co-existing plates of highly variable grain size. (d, e) BSE images of fault mirrors (W17-1G and W17-1H) with underlying breccia. Contact between breccia and mirror volume shown as black dashed line. Red dashed line denotes fault mirror interface. (f, g) BSE images showing detail of fault mirror volume cataclasite. (h, i) BSE images showing detail of fault mirror interface. Polygonal grains with triple junctions (black arrows) characterize W17-1G (h), hematite cataclasite characterizes W17-1H (i). Abbreviations are hem-hematite, qtz-quartz, fsp-feldspar.

**Hematite (U-Th)/He results**

We present new single aliquot HeHe dates in Figure 7 and Table S1. Fault mirror and specularite vein HeHe dates from Ault et al. (2015) and McDermott et al. (2017) are also shown for comparison. Intrasample scatter in our new data ranges from ~8% to 55% standard deviation (1σ) and we thus report sample results as individual aliquot dates ± 2σ analytical uncertainty. Tower Canyon specularite vein sample W16-6 yields dates ranging from 31.7 ± 0.9 Ma to 14.8 ± 0.4 Ma. Pearsons Canyon specularite vein samples yield dates ranging from 269.5 ± 7.4 Ma (W17-1F) to 17.8 ± 0.6 Ma (W17-2). The single plate isolated from W17-2 has a date of 74.8 ± 4.7 Ma.

Hematite isolated from Pearsons Canyon breccia matrix samples yield individual aliquot dates ranging from 68.8 ± 1.9 Ma (W17-1HB) to 18.4 ± 0.5 Ma (W17-1JB), whereas corresponding aliquots from the fault mirror volume have dates ranging from 42.2 ± 1.2 Ma (W17-1GA) to 13.1 ± 0.4 Ma (W17-1HB). Fault mirror samples W17-1HA and W17-1JA have the most reproducible dates of our dataset, with mean dates of 14.1 ± 1.2 Ma (1σ standard deviation) and 21.0 ± 2.3 Ma, respectively. For reference, previously published HeHe dates from fault mirrors range from 6.4 ± 0.2 Ma to 1.6 ± 0.2
Ma at Tower Canyon and 18.4 ± 0.6 Ma to 3.8 ± 0.3 Ma at Pearsons Canyon (Fig. 7; Ault et al., 2015; McDermott et al., 2017).

**Figure 7: Hematite (U-Th)/He data.** New and previously published hematite (U-Th)/He (HeHe) dates (Ault et al., 2015; McDermott et al., 2017). HeHe data are divided by location (Tower Canyon, left, and Pearson’s Canyon, right) and sample type. Symbol colors match for mirrors with underlying breccia, but with breccia dates shown as squares and fault mirror dates as circles. X-axis is arbitrary. Dates shown are for a single aliquot with 2σ analytical uncertainty.

Intra- and intersample HeHe data patterns are presented in Figure 8a and 8b. We group HeHe dates by sample type (i.e., specularite, breccia matrix, fault mirror), locale, and plate width. We add previously published samples W14-HC-20 and A13-6 to our Tower and Pearsons Canyon groups, respectively (Fig. 7; McDermott et al., 2017). Sample W14-HC-20 is an E-W trending specularite vein and A13-6 is specularite derived for a WNW-ESE trending fault surface with weakly developed striae (but not mirrored). Specularite aliquot dates increase with median plate-width and define two arrays (Fig. 8a; termed “upper array” and “lower array” hereafter). Each array is defined by samples from each canyon. The steeper, upper array comprises HeHe dates ranging from ~270 Ma to ~48 Ma at ~6 μm and ~8 μm median plate width (Fig. 8a). The shallower lower array
includes dates ranging from ~75 Ma to ~14 Ma at ~40 μm and ~8 μm median plate width (Fig. 8a). Breccia matrix and fault mirror HeHe dates do not exhibit as clear a relationship with median plate width, perhaps due to the lack of aliquot-specific grain size measurements for these samples (Fig. 8b). Some samples (i.e., W17-1HB and W17-1GA; Fig. 8a) plot in the HeHe date-plate width space defined by our defined upper array, whereas other samples could plot along either array.

Figure 8: Intrasample patterns of hematite (U-Th)/He (HeHe) and conventional bedrock thermochronometry data. (a) Specularite single aliquot HeHe date (± 2σ) vs. median plate width (± maximum/minimum). Dashed ovals denote upper and lower data arrays (see text for details). (b) Detail of lower plate-widths in (a). (c) New and previously published single grain zircon and apatite (U-Th)/He (AHe, ZHe) dates (± 2σ) vs. effective uranium (eU) concentration. (d) Detail of low-eU portion of (c) showing detail of AHe dates. Orange bar is apatite fission-track (AFT) central dates (± 2σ) for bedrock samples W14-T1 and W14-T5 (this study).
Results of new conventional low-temperature thermochronometric analyses are detailed in Tables S2, S3, and S4 and shown in Figure 8c and d. In reporting results from AHe and ZHe thermochronometers, we follow Flowers and Kelley (2011) and report ranges of single grain dates (± 2σ analytical uncertainty) or mean dates (± 1σ standard deviation) for samples with ≥20% or ≤20% standard deviation, respectively. ZHe dates from W14-T1 form a negative date-eU trend and range from 21.8 ± 0.7 Ma (~350 ppm eU) to 4.6 ± 0.1 Ma (~1050 ppm eU). AHe dates from W14-T1 average 4.1 ± 0.6 Ma. Apatite eU varies from ~13 to 35 ppm eU and shows no relationship with AHe date (Fig. 8d). New AHe dates are consistent with previously published data in our study area (Figs. 1; 8d; Armstrong et al., 2004; McDermott et al., 2017). AFT dates from samples W14-T1 and W14-T5 are 2.4 ± 0.5 Ma (central date ± 1σ) and 3.2 ± 0.7 Ma, respectively (Fig. 8d). All samples pass the χ² test, indicating a single date population (Table S4). However, a large fraction of counted grains does not preserve spontaneous tracks (i.e., single-grain AFT ages are 0 Ma; Table S4). Low U concentrations and recent cooling of these samples through the AFT Tc suggest reported central dates may be anomalously young.

PRELIMINARY INTERPRETATION OF HEMATITE (U-TH)/HE DATES

We suggest specularite HeHe data arrays and the spatial distribution of these HeHe dates in the study area are best explained by at least two episodes of hematite precipitation with subsequent ambient reheating and(or) cooling. Several lines of evidence support this first-order interpretation. First, samples from both locales define the two arrays. Except for dates from aliquots comprising thinner plate widths, most samples
have individual analyses in only one array. Second, at Pearsons Canyon, specularite vein samples separated by <1 m plot within different arrays (e.g., W17-1F and W17-1E). Third, the positive relationships between plate width, and thus $T_c$, and HeHe dates that define both arrays indicate that, following mineralization, hematite experienced variable post-formation He loss. Because deformation-related He loss can be ruled out for undeformed specularite samples, the positive trends reflect the control of the ambient thermal history on He loss. The different slopes of each array suggest samples precipitated at a minimum of two different times and are sensitive to different components of the post-formation ambient thermal history. The oldest dates of ~270 Ma and ~75 Ma provide a minimum formation age for aliquots in the upper and lower arrays, respectively.

An alternative interpretation for specularite date-plate width arrays is that hematite aliquots, particularly those that define our upper array, experienced U volatilization during laboratory degassing of He and thus have anomalously old dates. We consider this unlikely as preferential volatilization of U versus Th in this scenario results in HeHe dates correlated with Th/U, which we do not observe in our data (Fig. S4; Vasconcelos et al., 2013). U volatilization may also result in negative HeHe date-eU trends, which we do not observe in any of the samples defining our upper array (Fig. S4; Hofmann et al., 2020).

Breccia matrix and some fault mirror HeHe data likely reflect the cumulative effects of hematite precipitation and subsequent ambient thermal history, similar to specularite samples, as well as comminution effects. Breccia matrix sample W17-1HB and fault mirror sample W17-1GA overlap with the upper array defined by specularite
data. Other breccia matrix and fault mirror samples are difficult to uniquely assign to one array (Fig. 8a, b). In these samples, we observe microtextural evidence indicative of multiple episodes of hematite precipitation and deformation including grain size reduction, which modifies the diffusion domain length scale (Fig. 6d-i). Inclusion of older breccia material (e.g., Fig. 6e) within younger fault mirror volume material could impact bulk He dates from the latter.

Fault mirror HeHe dates may also reflect a component of He loss due to frictional heating during fault slip. Polygonal grains in sample W17-1GA are identical to those previously identified in WFZ fault mirrors and interpreted as indicating frictional heating to temperatures >1000 °C (McDermott et al., 2017). Prior thermomechanical modeling of fault slip on these fault mirrors suggest the temperatures required to yield these textures may be highly localized and induce ~50 to 100% He loss for grains between 0.1 and 2 μm in diameter (see Figure 4 in McDermott et al., 2017). These models also show He loss of ~10–20% up to 500 μm from the fault mirror for the grain sizes typical of our breccia aliquots (e.g., Fig. 8b). This suggests the potential for frictional heating at the fault mirror interface to impact underlying material comprising fault mirror volume and even underlying breccia.

THERMAL HISTORY MODELING

Approach and model setup

Thermal history modeling of our data allows us to explore the impact of competing thermal processes on HeHe data patterns and test our hypothesis that intrasample data patterns reflect multiple episodes of hematite precipitation with
superimposed reheating and cooling. In addition, models evaluate the consistency of this hypothesis with conventional thermochronometry data and independent geological constraints. Available thermochronometric inverse modeling packages do not permit simultaneous exploration of ambient thermal history and potentially multiple mineralization events, and developing such a modeling package is beyond the scope of this contribution. We thus opt for a forward modeling approach with the goal of testing whether specific combinations of hematite precipitation ages and thermal histories can reproduce first-order specularite, breccia matrix, and fault mirror data patterns.

We implement hematite He diffusion kinetics of Farley (2018) and assume diffusion domain length scale is equivalent to plate half-width (Farley, 2018; Jensen et al., 2018). For zircon, we use the Zircon Radiation Damage Accumulation and Annealing Model (ZRDAAM; Guenthner et al., 2013). Our ZHe dates exhibit a negative date-eU trend, consistent with radiation damage controls on effective $T_c$ in these samples (Guenthner et al., 2013). Analyzed zircon grains exhibit variable degrees of visual metamictization in polarized light (Fig. S5), providing evidence for different amounts of accumulated radiation damage that impact each grain’s $T_c$ (Ault et al., 2018). ZRDAAM implements zircon FT annealing data (Yamada et al., 2007) as a proxy for radiation damage annealing kinetics, and thus damage anneals at $\sim$310–220 °C in ZRDAAM. More recent experimental work suggests annealing temperatures may be as high as $\sim$500 °C (Ginster et al., 2019). For AHe and AFT data, we use the most recently available diffusion and annealing models, respectively (Ketcham et al., 2007; Flowers et al., 2009). We set $D_{par}$ to 1.62, reflecting the average of samples W14-T1 and W14-T5 (Table S3).
Forward models of AHe and ZHe dates take a given time-temperature path and grain size as input and generate a date-eU curve over the range of observed intrasample eU for comparison with thermochronometry results. We assume a single grain radius of 40 μm and 35 μm for ZHe and AHe models, respectively, reflecting the average of our analyzed grains (Table S2). For zircon, we solve the production-diffusion equation using algorithms presented in Ketcham (2005) but implemented in MATLAB to facilitate modeling of a larger number of time-temperature paths using code developed by Guenthner et al. (in revision). AHe and AFT dates are modeled with the HeFTy software (Ketcham, 2005). Time-temperature paths are then used to generate curves of HeHe date versus plate width, solving the production diffusion equation in MATLAB over the approximate range of hematite plate half-widths we observe in our samples (0.1–50 μm).

Time-temperatures paths used in forward models are simple but consistent with independent geological data (Fig. 9a). The initial condition of our input thermal histories is set at 1720 Ma and 650 °C, corresponding to peak amphibolite facies metamorphism and well above radiation damage annealing temperatures in zircon (Hedge et al., 1983; Barnett et al., 1993; Guenthner et al., 2013; Ginster et al., 2019). We require near-surface temperatures at 500 Ma indicated by the Paleoproterozoic-Cambrian unconformity preserved in our study area. Previously reported Paleozoic strata thicknesses are converted to burial temperatures assuming a 25 °C/km geothermal gradient. Because of uncertainty in the presence and associated thickness of Mesozoic strata in our study area, the estimated timing and magnitude of Phanerozoic reheating is broad (see labeled red box in Fig. 9a; Crittenden, 1974; Yonkee, 2005). The Willard thrust was active from ~140 to 90 Ma and may have emplaced ~10-15 km of hanging wall strata on rocks in our
study area (Fig. 1; DeCelles, 1994; Yonkee, 2005; Yonkee et al., 2019), erosion likely accompanied thrust belt evolution, however, and thus net burial of footwall rocks (i.e., our study area) was likely less than this. We thus allow peak temperatures up to 375 °C, corresponding to the full thickness of strata in the thrust sheet (e.g., Yonkee, 2005), but consider lower temperatures as well (see labeled red box in Fig. 9a). We impose a temperature of 0 °C at 0 Ma.

We employ an iterative forward modeling approach to explore thermal histories consistent with all thermochronometric data. We first identify a range of permissible thermal history models that satisfy the ZHe, AHe, and AFT data patterns from Farmington Canyon Complex host rock beginning with high-temperature metamorphism in the Proterozoic. We then consider variations of the best-fit result of these thermal histories, truncated at different intervals corresponding to hypothesized hematite mineralization dates. At each stage, we cross-check the predicted model outcomes (i.e., dates as a function of hematite plate width or zircon eU) on each dataset to evaluate the most likely thermal history to explain all data.

**Initial forward models of zircon (U-Th)/He, apatite (U-Th)/He and apatite fission-track data**

Initial time-temperature paths are aimed at delineating endmember thermal histories for Farmington Canyon Complex rocks in our study area where geological constraints are broad or non-existent. First, we consider two time-temperature paths involving Paleoproterozoic rapid *versus* slow cooling that are informed by preliminary inverse models of our data (dashed vs. solid lines in Fig. 9a; see Supplemental
Information for details of pilot inverse models). Second, we choose a restricted range of peak Mesozoic reheating temperatures informed by previously published fluid inclusion homogenization temperatures from the base of the Willard thrust in our initial thermal history simulations (Fig. 9a; Yonkee et al., 1989). We further explore the influence of the magnitude of reheating on our data in the subsequent section. We incorporate cooling to variable Cenozoic temperatures following peak reheating until 50 Ma (inset in Fig. 9a), corresponding to exhumation associated with uplift of the Wasatch Anticlinorium (DeCelles, 1994). At this stage, we do not allow pre-5 Ma temperatures of <120 °C in order to be consistent with AFT dates and the Tc of this system (Fig. 9a; Ketcham et al., 2007). Lastly, the timing of final exhumation is varied between 12 to 5 Ma, reflecting the onset of modeled exhumation elsewhere along the Wasatch fault and consistent ~5 Ma dates across multiple thermochronometers in our study area (Fig. 8c, d; Ehlers et al., 2003; Armstrong et al., 2004; McDermott et al., 2017).

Results from forward modeling of ZHe, AHe, and AFT data place limits on acceptable thermal histories for Farmington Canyon Complex. We compare predicted time-temperature paths with our observed ZHe and AHe date vs. eU patterns and AFT dates. In general, time-temperature paths characterized by rapid cooling in the Paleoproterozoic followed by radiation damage accumulation in zircon and apatite from residence at low temperatures, require Mesozoic reheating to peak temperatures of ~120 to 150 °C (Fig. 9a, b). This tT history is consistent with preserved visual metamictization in zircon crystals (Fig. S5). However, if our ~21 Ma ZHe date at low eU is an outlier, hotter peak re-heating temperatures (≥200 °C) are preferred by ZHe data (Figs. 9b; S6). AHe and AFT are relatively insensitive to this set of initial forward model paths,
although better fits are provided by paths with lower peak temperatures and cooling at 5 Ma (Fig. 9c).

**Figure 9:** Thermal history modeling of bedrock conventional thermochronometry data. (a) Tested thermal histories (solid and dashed lines) with independent geological information indicated (see text for details). Red boxes show range of permissible time-temperature space from strata thicknesses. (b) Predicted ZHe dates vs. eU (solid lines). Colors correspond to paths in (a). Dashed lines are thermal histories with additional rapid cooling in Paleoproterozoic. (c) Predicted AHe dates vs. eU (solid lines). Predicted AFT dates shown as triangles. Legend for observed data as in Figure 8.

**Forward models of hematite (U-Th)/He-plate-width data patterns**

We next consider thermal histories that best reproduce our HeHe dates and are consistent with conventional thermochronometry dates. We first test the influence of peak Mesozoic reheating temperature on HeHe dates. We build on the first set of forward models: thermal histories are similar in form to the dashed lines in Figure 9a, but include three variants with peak reheating to temperatures of 150 °C, 165 °C, and 180 °C, followed by exhumation to temperatures of 100 °C, 120 °C, and 130 °C, respectively, by 50 Ma (Fig. 10a). For simplicity, we assume a hematite mineralization age of 1720 Ma for these simulations. These models reveal (1) the potential antiquity of hematite mineralization, chiefly in the upper array of HeHe data, (2) the upper array (Fig. 8a) is
best fit by the lowest allowed peak reheating temperatures, and (3) the lower array is best fit by the highest allowed temperatures (Fig. 10b). Lower reheating temperatures preferred by a large portion of our HeHe data yield increasingly poorer fits to ZHe data (Fig. 10c).

This stage of modeling reveals two discrepancies between HeHe and ZHe data. First, HeHe and ZHe data prefer lower and higher Mesozoic reheating temperatures, respectively. Second, upper and lower linear arrays defined by HeHe data prefer lower and higher Mesozoic reheating temperatures, respectively, if the same hematite precipitation age is assumed (Fig. 10). Once formed, hematite must experience the same time-temperature history as host rock thermochronometers. Thus, only one reheating scenario should fit the data. These discrepancies may be due to uncertainties in grain size distribution, HeHe diffusion kinetics, and(or) ZHe diffusion kinetics. Alternatively, a single thermal history, but with variable hematite mineralization ages may explain observed HeHe date-plate width trends and also satisfy ZHe data, as we explore in the next section.
Figure 10: Pilot thermal history models of specularite vein hematite (U-Th)/He data. (a) Tested thermal histories assuming a 1720 age of hematite precipitation. Boxes reflect Mesozoic reheating bounds as in Figure 9a. (b) Predicted HeHe date vs. plate width (solid lines). Colors correspond to paths in (a). Symbology of observed data as in Figure 8a. (c) Predicted ZHe date vs. eU trends for paths shown in (a).

Consideration of variable hematite precipitation ages

We conduct a series of forward models that consider different hematite precipitation ages and systematically vary peak Mesozoic reheating temperatures over the
full space permitted by geological constraints (labeled red boxes in Fig. 9a). Modeled hematite precipitation ages are justified by field observations and regional geology. First, field and microtextural observations in our study area suggest some hematite co-precipitated with pegmatitic pods and veins hosted in the Farmington Canyon Complex (e.g., sample W16-21; Fig. 6a). Pegmatites yield concordant ~1690-1670 Ma zircon U-Pb dates (Mueller et al., 2011) and some hematite may thus be this age. Second, Pb-Zn sulfide, quartz, and hematite mineralization is hosted in Cambrian strata ~2 km east of our samples and at an elevation of ~2200 m.a.s.l. (Fig. 1b, King Solomon Mine; Doelling et al., 1980). The age of these deposits is unknown, but Cambrian host rock requires a maximum ~500 Ma age. Third, published HeHe dates from specularite hosted in Permian strata, ~20 km NNE from our study area, are ~280 Ma across a range of plate widths, similar to the oldest single aliquot HeHe date we report (Figs. 1a; 7b; 8a; Calzolari et al., 2020). We also consider mineralization at 110 Ma and 12 Ma, concomitant with prior constraints on peak Willard thrust emplacement and initiation of Basin and Range extension, respectively (Ehlers et al., 2003; Yonkee et al., 2019).

We generate initial time-temperature paths consistent with best fit model solutions in prior sections. These paths involve (1) Paleoproterozoic cooling to be compatible with our ZHe dates (see dashed line, Fig. 9a), (2) a post-50 Ma temperature of 120 °C, the lowest allowable by conventional thermochronometric data and still broadly consistent with initial HeHe thermal modeling results, and (3) rapid cooling from 5 Ma to present. Because modest changes in peak reheating temperature can produce dramatically different HeHe dates (Fig. 10), we increase reheating temperatures at ~4 and 8 °C increments for reheating due to burial by Paleozoic strata and thrusting, respectively.
From these paths, we model ZHe date-eU trends, AHe data, and AFT data. Each path is then truncated at a hypothesized hematite precipitation age and rerun to predict HeHe date-plate width patterns. In Figures 11 and 12, we show HeHe date-plate width and zircon date-eU patterns, as modeled AHe and AFT dates were identical to those shown for light blue paths in Figure 9.

Modeling results reproduce all thermochronometric data patterns to first order and indicate different HeHe date-plate width arrays require different ages for hematite (Fig. 11). For time-temperature histories with peak reheating temperatures that do not exceed \(~150\) °C, our upper linear array of specularite vein HeHe dates can be reasonably fit by 1680 and 500 Ma mineralization dates (Fig. 11a, b). The minimum assumed mineralization age of \(~280\) Ma for this subset of data yield poorer fits (Fig. 11c). This same suite of time-temperature paths and assumed hematite mineralization at 110 Ma yields an adequate fit to our lower linear array of specularite HeHe data (Fig. 11d). Assuming a 12 Ma precipitation age is not applicable to HeHe dates >12 Ma, but can reasonably explain specularite sample A13-6 which yields uniform \(~11.7\) Ma dates despite a wide plate-width distribution (Fig. 12a). This model set also reasonably reproduces Farmington Canyon Complex host rock ZHe data (Figs. 11; 12a). Although the low-eU \(~21\) Ma ZHe date is not fit by any of the models considered, model simulations still predict the first-order negative zircon He date-eU trend with a “pediment” at \(~5\) Ma. We also considered ZHe forward models of tT paths in Figures 11 and 12a with Ginster et al. (2019) annealing kinetics and no annealing (Fig. S7). These models only slightly decrease the discrepancy between modeled and observed ZHe dates (Fig. S7).
Figure 11: Thermal history models with variable reheating and hematite formation age. Rows correspond model outputs for assumed hematite precipitation at 1680 Ma (a), 500 Ma (b), 280 Ma (c), and 110 Ma (d). Left panels show tested thermal histories, middle panel shows predicted HeHe date-plate width trends for thermal histories with observed specularite vein data only. Right panel shows Predicted ZHe date-eU trends for models. Colors correspond to thermal histories in left panel. Thermal histories generated for ZHe data also include time-temperature points shown in black. Predicted AHe and AFT dates are as in Figure 9c.
Figure 12: Thermal history models shown with breccia and fault mirror data (a)
Thermal histories assuming hematite formation at 12 Ma, predicted HeHe date-plate
width trends, and predicted ZHe-eU trends as in Figure 11. (b) Detail of thermal histories
shown in Figure 11 and against breccia matrix and fault mirror HeHe dates. Symbology
as in Figure 8b. Assumed hematite formation ages in black italics.

DISCUSSION

Key outcomes of thermal history modeling

Multiple hematite precipitation events

Thermal history modeling outcomes support the hypothesis that HeHe dates from
undeformed specularite reflect multiple generations of hematite precipitation. First, HeHe
dates from veins that form the upper array (W16-21, W17-1F, and W14-HC-20) are
consistent with earlier mineralization followed by post-formation He loss from ambient
reheating and cooling. Hematite in this array formed ≥270 Ma but is likely as old as
~1680 Ma based on the quality of model fits for simulations with this precipitation age
(compare Fig. 11a and 11b). Second, HeHe dates defining the lower array (W17-2, W17-
1E, and W16-6) reflect an ~110 to ~75 Ma generation of hematite and He loss from
subsequent ambient cooling (Fig. 11d). Third, models that use a ~12 Ma hematite
formation age predict a date “plateau” across a range of plate widths (Fig. 12a). Specularite sample A13-6 from McDermott et al. (2017) has reproducible dates of ~11.7 Ma, despite a wide plate-width distribution (Fig. 8b), suggesting these data record Miocen hematite precipitation. Assuming an older date for this sample yields a poor fit to observed data (Fig. 12b-d). Each of these samples exhibits intrasample date scatter. We suggest this reflects on-going He loss from different-size plates within the analyzed aggregates during the post precipitation thermal history (Figs. 6a, b, c; 8a, b).

HeHe data patterns from hematite breccia matrix and fault mirrors likely reflect hematite precipitation prior to ~270 Ma, at ~20 Ma, and at ~14 Ma, and subsequent partial He loss by friction-generated heat associated with deformation and mirror formation. Breccia matrix sample W17-1HB overlaps with the upper array defined by specularite, suggesting precipitation >270 Ma and potentially as old as ~1680 Ma (Fig. 8a). Fault mirror dates from this sample (W17-1HA) are reproducible at ~14 Ma, ~46 Ma younger than the average of W17-1HB aliquots (Figs. 7; 8b). Tc estimates from these two textural domains overlap (Fig. S3). This comparison, and the lack of high-temperature microtextures that form at the same temperatures required to induce substantial He loss during deformation (cf. McDermott et al., 2017), suggest W17-1HA captures hematite precipitation at ~14 Ma. Texturally this sample comprises cataclasite, but grain size reduction does not appear to have induced He loss in this sample.

Breccia matrix W17-1JB and corresponding fault mirror W17-1JA yield reproducible ~23–18 Ma dates (Figs. 7; 8b). These samples have overlapping plate width and Tc distributions that are similar to W17-1HB and W17-1HA. Owing to the similar and limited range of grain sizes (i.e., thinner and uniform plate widths), we would
anticipate similar dates between these samples if they recorded ambient cooling. Instead we observe distinct HeHe dates from W17-1JB/W17-1JA and W17-1HA and W17-1HB. This comparison supports our interpretations of W17-1HB and W17-1HA and suggests W17-1JB and W17-1JA also record hematite precipitation at ~20 Ma.

HeHe dates from W17-1GA overlap the upper array (Fig. 8a), suggesting precipitation at >270 Ma for hematite comprising the fault mirror volume in this sample. This sample exhibits microtextures consistent with past friction-generated temperature rise sufficient to at least partially reset HeHe dates (Fig. 6h; Ault et al., 2015; McDermott et al., 2017). We consider it likely younger individual dates for this sample reflect at least some He loss from frictional heating. Some of these younger dates overlap with analyses from W17-1JA and W17-1JB, indicating fault slip on W17-1GA occurred broadly in tandem with hematite precipitation. Sample W17-1GA is distinct from other fault mirrors analyzed in this study and in prior work (Ault et al., 2015; McDermott et al., 2017). Hematite comprising the W17-1GA fault mirror volume either incorporates a component of older hematite (i.e., underlying breccia) during deformation or the fault mirror volume itself is markedly older than other analyzed fault mirrors (Fig. 7).

*Constraints on ambient thermal history, burial, and exhumation*

A key result of our modeling is that hematite of different age and bulk T_c are sensitive to different portions of Farmington Canyon Complex thermal history, which in turn can also refine the burial and exhumation history of these rocks. Geologic constraints on exhumation (e.g., strata thicknesses) are broad and do not delineate key parts of the thermal history (i.e., post-Sevier exhumation). Modeling shows earlier-formed hematite
with generally older HeHe dates retain a portion of its He budget through reheating, effectively limiting likely peak Mesozoic and Cenozoic temperatures to ~120–150 °C (Fig. 11). However, hematite forming the lower array precipitated concomitant with or following peak Mesozoic burial, and is thus only sensitive to Cretaceous through present-day thermal history characterized by cooling due to exhumation. The rate and magnitude of this cooling is also directly dependent on temperature constraints imposed by upper array HeHe dates.

Our HeHe data from different hematite generations and modeling outcomes refine the burial and exhumation history of the Farmington Canyon Complex since the Mesozoic. Models suggest Cretaceous reheating to ~120–150 °C, Cretaceous-Eocene cooling (or lack thereof, see red paths in Fig. 12 that best fit upper array HeHe data) to ~120 °C, and late Miocene or Pliocene cooling from ~120 to ~0 °C. Assuming a geothermal gradient of 25 °C/km and a surface temperature of 0 °C, this suggests ~4-6 km of Mesozoic sedimentary and thrust sheet burial, ~0 to ~1 km of cooling due to erosional exhumation during Wasatch anticlinorium development, and ≤5 km of cooling due to exhumation related to Basin and Range extension.

Broadly speaking, thermal models for HeHe dates are consistent with those for conventional thermochronometry data, but there are some discrepancies. The same time-temperature path can explain HeHe, AHe, and AFT dates, but our models have difficulty honoring both HeHe and ZHe data (Figs. 10; 11). ZHe dates from moderate- and high damage zircons are reasonably fit by time-temperature paths that best fit the HeHe data from the upper array. But modeled ZHe dates from low-damage zircon differ by 100s of Myr for the thermal histories most preferred by HeHe data (see red paths in Figure 11).
HeHe dates of the lower array are better fit by time-temperature paths with hotter Mesozoic reheating temperatures, which also yield better fits to the ZHe data (compare blue and black curves in Figure 10). Prior work utilizing thermokinematic models and ZHe data from the Willard thrust upper plate constrain rapid cooling from ~75-50 Ma as implied by our HeHe data, but also predict hotter peak Mesozoic reheating temperatures (~200–240 °C) for footwall rocks near our study area (Yonkee et al., 2019). Resolving inconsistencies between our thermochronometric datasets and prior work is beyond the scope of this contribution, but we speculate these differences reflect uncertainties in diffusion kinetics of both the HeHe and ZHe systems. Hematite may be more He retentive than published diffusion kinetics imply (Farley, 2018), in effect causing HeHe data to suggest lower post-mineralization ambient temperatures than samples actually experienced. Uncertainties in ZHe diffusion kinetics and corresponding effects on thermal history models are also well-documented (e.g., Mackintosh et al., 2017; Johnson et al., 2017; Ginster et al., 2019). Because hematite in our samples likely formed within a broad partial retention zone (PRZ) defined by variable grain size distributions, and the ZHe PRZ is transient over protracted geologic histories due to evolving radiation damage accumulation, discrepancies between modeled and real diffusion kinetics for each system are also likely exacerbated in our particular samples.

**Evidence for fault reactivation and slip localization**

We document field, microtextural, and thermochronometric evidence for fault reactivation and slip localization along small faults within the WFZ. At the field-scale, fault mirrors with multiple striae provide evidence for repeated fault reactivation during
development of the Wasatch fault damage zone during Basin and Range extension (Fig. 3d, e; Evans and Langrock, 1994; Evans et al., 2014). Structural data suggest the majority of fault mirrors record normal slip and E-W or NW-SE directed extension (Fig. 4a, b; Evans and Langrock, 1994). Striae define best-fit girdles similar to the observed mean strike and expected dip for the main trace of the WFZ (Fig. 5c, d), further suggesting kinematic compatibility between fault mirror deformation and Basin and Range extension (Fig. 4e, f).

Microtextural observations support fault reactivation on fault mirrors. Evidence for reactivation includes multiple generations of hematite precipitation, sharp boundaries between the fault mirror volume and underlying breccia cemented by hematite, and breccia clasts incorporated within fault mirror volume material. These observations indicate slip surfaces developed within pre-existing breccia and that on-going deformation is associated with slip localization (Fig. 6d-i). This is similar to prior observations showing most fault mirrors likely formed by reactivation of pre-existing specularite and breccia (Ault et al., 2015; McDermott et al., 2017).

HeHe data from breccia matrix and fault mirror volume microtextural domains complement microtextural observations of slip localization and fault reactivation and bracket the timescales over which these processes occur. For example, sample W17-1H comprises likely Paleoproterozoic breccia that served as the locus for additional hematite precipitation at ~14 Ma and subsequent normal slip (see pseudo-focal mechanism annotations in Fig. 4b; Fig. 7). In contrast, W17-1J shows microtextural evidence for reactivation and hosts normal slip (Fig. S2; Fig. 4b) but yields similar HeHe dates, implying shorter timescales between hematite precipitation and fault mirror development.
Analyzed hematite from the fault mirror volume of W17-1G fully or partially comprises Paleoproterozoic hematite, but exhibits evidence for deformation enhanced He loss at ~20 Ma and hosts oblique reverse and normal slip (Figs. 3d; 4). A13-6 specularite likely indicates ~12 Ma mineralization and shows field evidence for incipient fault surface development (Figs. 7; 12a).

Collectively, field, microtextural, and thermochronometric data require multiple hematite precipitation events that are preserved in what is the WFZ today and that individual slip surfaces capture not only more than one generation of hematite but also more than one deformation event. Hematite has a coefficient of friction (μ) of ~0.28, whereas gneissic host rock likely has more typical μ of ~0.6-0.7 (Byerlee, 1978; Calzolari et al., 2020). This strength contrast indicates hematite veining created small-scale strength discontinuities within host rock that served to localize slip, in some cases potentially 100s of Ma after original precipitation.

Model for tectonic evolution of the Wasatch Front and Wasatch fault zone

We synthesize geologic map patterns, field data on hematite occurrence, structural analyses, thermochronometric data, and thermal history models to describe the long-term history of hematite precipitation, fault reactivation, and deformation preserved in rocks of the now exhumed WFZ (Fig. 13). We first divide new and previously published HeHe dates into four groups (Fig. 13a, inset). Groups are defined in terms of process (i.e., precipitation and heating/cooling or deformation) and cross-cutting relationships observed in the field (e.g., Fig. 3g, f). HeHe date groups are presented as probability density functions created with IsoplotR (Fig. 13a, inset; Vermeesch, 2018). Group 1 (red)
includes specularite veins and breccia from the upper array, including breccia matrix sample W17-1HB (Fig. 8a, b), which are collectively oriented E-W (Fig. 5a, b). Date-plate width patterns in this group reflect combined hematite precipitation >270 Ma and likely as old 1680 Ma and subsequent He loss through ambient heating and cooling. Group 2 (gray) comprises specularite veins and breccia matrix from our lower array both characterized by E-W trending and potentially N-S trending veins and fault mirrors (e.g., W16-6; Fig. 8a). HeHe dates in this group reflect ~110–75 Ma hematite precipitation and post-formation ambient cooling. Group 3 (blue) includes breccia matrix W17-1JB (N-S trending) and A13-6 specularite (WNW-ESE trending). Group 3 also includes fault mirror dates (blue, dashed) W17-1JA, W17-1HA, and W17-1GA, which have variable orientation and normal slip (Fig. 4b). This date grouping predominantly reflects Miocene hematite precipitation, reactivation of E-W trending specularite and breccia (with group 1 and 2 dates), and normal slip. Lastly, we include previously published data from largely N-S trending fault mirrors with normal slip (e.g., equal area plot annotations in Fig. 4a) in Group 4 (pink, dashed; Ault et al., 2015; McDermott et al., 2017). Individual aliquot dates in this group range from ~18 Ma to ~1 Ma, but McDermott et al. (2017) suggested most of these data reflect either full or partial He loss from friction-generated heat and seismic slip ≤4.5 Ma. Although these groups are largely based on HeHe results, they honor key field observations. For example, older groups 1 and 2 comprise E-W trending veins. N-S trending veins and fault mirrors are part of younger groups 3 and 4 and cross-cut E-W trending veins (Fig. 3g, f).
Figure 13: Thermal and structural evolution of the Wasatch fault zone and associated rocks. (a) Family of preferred thermal histories from Figures 11, 12 (dark gray, see text for details). Probability density functions (PDFs) of hematite groups (see text for details). Inset shows detail of PDFs at <50 Ma. Gray bars denote timing of regional tectonic events. Lower panel gives inferred hematite formation ages. Colors correspond to PDFs above. Italics lettering denotes representative block diagrams below. (b) First stage of hematite mineralization and formation of E-W trending specularite veins. (c) Second stage of hematite mineralization with largely E-W trending veins, and reactivation of prior vein sets. (d) Third stage of hematite mineralization and formation of Wasatch fault damage zone, formation of N-S trending veins, and reactivation of new and previous hematite. Black lines denote unmineralized fractures. Block diagram insets show detail of individual specularite vein or breccia repeatedly reminerlized and reactivated. Colors in block diagrams and insets correspond to generations in (a).

Specularite veins and fault mirrors record at least three phases of deformation corresponding with regional tectonism from the Paleoproterozoic to present (Fig. 13). First, E-W trending specularite veins and hematite-cemented breccias in Group 1 formed
during Paleoproterozoic tectonism within the Farmington Canyon Complex (Fig. 13b). Second, additional E-W trending veins, and lesser N-S trending veins (e.g., W16-6) in Group 2 reflect additional precipitation and associated deformation during the Sevier orogeny (Fig. 13c). These E-W orientations are consistent with a subhorizontal and E-W-directed maximum principal stress associated with transport of the Willard thrust sheet (Yonkee, 1992; DeCelles, 1994). Some E-W trending fault mirrors with reverse and oblique reverse slip (see reverse-sense pseudo-focal mechanism solutions in Fig. 4) may also correspond to this time interval. For example, sample W17-1G is E-W trending and exhibits normal striae that overprint reverse striae (Figs. 3d; 4b). Hematite in this sample may be as old as Paleoproterozoic and reverse striae could instead reasonably reflect pre-Sevier contraction. Third, Miocene Basin and Range extension resulted in the formation of N-S oriented specularite veins, breccia, and fault mirrors subparallel to the WFZ in groups 3 and 4 (Fig. 13d). E-W trending fault mirrors from prior tectonic phases were reactivated in normal slip (e.g., W17-1HA and W17-1GA; Fig. 4b) during this stage. We observe evidence for each precipitation and deformation event at Tower and Pearsons canyon locales, indicating consistency of these processes over at least a few km’s (Fig. 2).

Our data bear on the evolution of the WFZ and reveal multiple stages of faulting during Basin and Range extension. Some HeHe dates suggest initiation of normal faulting by at least ~14 Ma, consistent with previous estimates of the initiation of extension along the WFZ (Naeser et al., 1983; Parry and Bruhn, 1986; Kowallis et al., 1990; Ehlers et al., 2003). Other HeHe dates support normal faulting, at least locally, as early as ~20 Ma (Fig. 7). Evidence of earlier faulting overlaps with the timing of
extension documented elsewhere along the Wasatch Front (Constenius, 1996). Thermal history models do not support rapid exhumation during ~20 Ma WFZ activity (Fig. 13a). In contrast, rapid exhumation from ~5 Ma-present overlaps with previously published Group 1 HeHe dates (Ault et al., 2015; McDermott et al., 2017). HeHe dates from fault mirrors associated with this deformation, albeit only reflecting small strains, may capture a component of tectonic denudation via seismic slip.

High density-corridors of E-W trending and N-S specularite veins and fault mirrors (1) occur with, and reflect the orientation of, outcrop-scale normal faults at both Tower Canyon and Pearsons Canyon (Figs. 2; 4) and (2) parallel the local WFZ trace at Tower and Pearsons canyons (Figs. 1; 2; 5). Pearsons Canyon exhibits a more complex spatial pattern of WFZ-related deformation at map- and outcrop-scales. For example, this locale has a greater proportion of E-W trending veins and fault mirrors and lies near the intersection of NW and N-S striking strands of the WFZ main trace (Figs. 1; 2; 5b, d). Small-scale fault-related hematite mineralization significantly precedes Basin and Range faulting and likely played a role in the localization of larger structures and accommodation of strain at zones of complex faulting between different WFZ fault segments (Fig. 13d). Our collective dataset thus suggests inherited structural complexity in host rock may be reflected in outcrop- and map-scale patterns of faulting.

CONCLUSIONS

We present integrated field, structural, microtextural, and thermochronometric data from specularite, hematite-cemented breccia, and fault mirrors that reveal protracted precipitation, fault reactivation, and deformation that influence the structural evolution
of, and seismogenesis in, the WFZ. HeHe dates from specularite veins and hematite fault mirrors form two arrays in HeHe date-plate width space. Thermal history modeling suggests intra- and intersample data patterns reflect multiple generations of hematite precipitation, followed by ambient reheating and cooling, and additional He loss related to coseismic deformation as fault mirrors develop. Our data interpretations are broadly consistent with AHe, AFT, and ZHe thermochronometric data from the host Farmington Canyon Complex. Some mismatch between the thermal histories preferred by HeHe versus ZHe data is likely due to uncertainties in diffusion kinetics for these respective systems.

We identify at least three generations of hematite precipitation within rocks comprising the present day exhumed WFZ. The earliest generation of hematite documented in our study area significantly predates development of the modern WFZ and is at least ~270 Ma but may be as old as ~1680 Ma. Additional hematite precipitation occurred in the Cretaceous and Miocene, related to the Sevier Orogeny and Basin and Range extension, respectively. Microtextural observations and HeHe data isolated from specific microtextures show specularite veins and breccia, despite forming 100s of Ma prior, focused subsequent synkinematic fluid flow and hematite precipitation, followed by reactivation to form hematite fault mirrors. Fault mirrors accommodate dominantly Miocene coseismic normal faulting associated with Basin and Range extension and WFZ development. Field observations, structural analyses, and thermal models of HeHe data reveal hematite faults were reactivated multiple times over distinct tectonic phases, resulting in complex fault kinematic patterns and thermochronometric signatures of He loss related to coseismic frictional heating and grain size reduction via cataclasis.
Our data indicate fault reactivation and strain localization at the cm- to m-scales along discrete slip surfaces are important processes in the development of the WFZ and potentially other brittle fault zones. Our results suggest relative strength contrasts between cement and host rock influence fault reactivation processes on discrete slip surfaces, potentially over 100s of Myr. Prior work has shown hematite fault mirrors accommodated seismic slip, and our data thus indicate precursor damage may influence earthquake nucleation processes (Ault et al., 2015; McDermott et al., 2017). Zones of hematite precipitation during orogenesis and subsequent fault reactivation may also catalyze development of outcrop- and map-scale faults, thus demonstrating a link between prior tectonic phases and the resulting architecture of the WFZ in our study area.

REFERENCES


Schulz, S.E., and Evans, J.P., 2000, Mesoscopic structure of the Punchbowl Fault, Southern California and the geologic and geophysical structure of active strike-


Wernicke, R.S., and Lippolt, H.J., 1993, Botryoidal hematite from the Schwarzwald (Germany): heterogeneous uranium distributions and their bearing on the helium


CHAPTER 4

MICROTEXTURAL EVIDENCE FOR MULTIPLE DYNAMIC WEAKENING MECHANISMS DURING COSEISMIC SLIP ON HEMATITE FAULT MIRRORS FROM THE WASATCH FAULT ZONE, UT, USA

Abstract

Thin, reflective fault surfaces or fault mirrors (FMs) provide textural archives of weakening mechanisms that spur strain localization and seismogenesis in the shallow crust. We document the microtextural, thermal, and mechanical evolution of three hematite FMs from the Wasatch fault zone (WFZ) with scanning electron microscopy, quantitative image analysis, and electron backscatter diffraction. FMs comprise three textural domains: hematite-cemented breccia, cut by ~200 μm-thick hematite cataclasite, and ~5–20 μm-thick hematite ultracataclasite with local polygonal grains at the FM surface. Grain size distributions, aspect ratios, and orientations suggest cataclasite and ultracataclasite reflect fluidized granular flow at seismic strain rates. Polygonal grain morphologies, crystallographic preferred orientation, and intragrained misorientations show evidence for crystal plastic deformation activated at coseismic elevated temperatures. Each FM preserves textural evidence for spatially-variable (over 100s of μm) temperature, deformation, and dynamic weakening mechanisms operative during seismic slip. Inferred weakening mechanisms include fluidization, nanoparticle lubrication, and asperity flash heating. Integrated microtextural and previously published

---

1 Paper formatted for submission to the Journal of Structural Geology with author list: R.G. McDermott\(^a\), A.K. Ault\(^a\), K.F. Wetzel\(^a\), J.P. Evans\(^a\), Fen-Ann Shen\(^a\)

\(^a\)Utah State University
hematite (U-Th)/He thermochronometry data reveal feedbacks between temperature rise, dynamic weakening, and strain localization during FM development, and also the antiquity of hematite rheological heterogeneities that are exploited during seismogenesis.

1. Introduction

Studies of exhumed fault-related rocks complement geophysical, laboratory, and numerical modeling data to inform a mechanistic understanding of earthquake processes. For example, conceptual and laboratory models of earthquake nucleation and propagation rely on unstable stick-slip (Brace and Byerlee, 1966; Dieterich, 1992; Scholz, 1998 and references therein). Unstable slip arises when shear stress reduction or weakening occurs more rapidly than the dissipation of tectonic stresses (Scholz, 1998). Dynamic weakening processes are a critical component of seismogenesis (e.g., Di Toro et al., 2011). Identifying dynamic weakening mechanisms operative in fault-related rocks is key to microphysically-based models of earthquake rupture and the seismic cycle, understanding conditions favoring earthquake nucleation, and predicting down- and up-dip rupture paths (e.g., Noda et al., 2009; Niemeijer and Vissers, 2014; Thomas et al., 2014; Chen and Spiers, 2016; Brantut and Platt, 2017). Textural and chemical evidence for the occurrence and distribution of these mechanisms from brittle fault rocks underpin these models.

Rotary shear experiments and field studies document various mechanically- and thermally-activated dynamic weakening mechanisms, including melt lubrication, flash heating at geometric asperities, nanoparticle lubrication, lubrication by phase transformations, fluidization, thermal pressurization, and crystal plastic deformation and flow (Rice, 2006; Hirose and Bystricky, 2007; Kirkpatrick and Shipton, 2009; Di Toro et
Weakening mechanisms depend on a range of factors, including lithology and strain rate (Di Toro et al., 2011; Spagnuolo et al., 2016). Progressive localization of slip further promotes temperature rise and activation of additional dynamic weakening mechanisms (Rice, 2006; Platt et al., 2014). Strain localization is a typical component of brittle fault zones that have experienced multiple earthquake cycles and alteration, and likely reflects fault reactivation and strength reduction over longer timescales (Sibson, 1982; Chester and Logan, 1986). The evolution of strain localization and static or dynamic weakening over both geologic time and the duration of an individual earthquake are thus intimately related.

High-gloss fault surfaces, or fault mirrors (FMs), and other thin slip surfaces are common in natural fault zones and represent archives of dynamic weakening, strain localization, and potentially seismic slip (e.g., Power and Tullis, 1989; Kirkpatrick et al., 2013; Siman-Tov et al., 2013, 2015; Evans et al., 2014; Ault et al., 2019). Experimental work connecting textures and friction evolution indicates FMs can develop by various dynamic weakening mechanisms, including the generation of nanoparticles, asperity flash heating, and(or) plastic deformation (Han et al., 2011; Smith et al., 2013, 2015; De Paola et al., 2015; Siman-Tov et al., 2015; Pozzi et al., 2018; Calzolari et al., 2020). FMs may also be produced during subseismic slip and without significant dynamic weakening (Verberne et al., 2013). Temperature rise can vary widely within a slipping zone (Coffey et al., 2019), and thus a range of weakening mechanisms are likely during FM development (cf. Kirkpatrick and Shipton, 2009).
The Brigham City segment of the seismogenic Wasatch fault zone (WFZ) in northeastern, UT, USA, hosts hematite-coated FMs that accommodated paleoearthquakes (Evans et al., 2014; Ault et al., 2015; McDermott et al., 2017). Prior work applied hematite (U-Th)/He thermochronometry and microtextural analysis to document highly localized temperatures of >1200 °C on hematite FMs, strongly suggesting asperity flash heating and associated dynamic weakening (Ault et al., 2015; McDermott et al., 2017). These studies also documented variability in microtextures both across and perpendicular to the FM surface. Here, we investigate three additional WFZ FMs with multi-scale observations from the field to the nanoscale with scanning electron microscopy (SEM), quantitative image and grain size analyses, and electron backscatter diffraction (EBSD) mapping. Each FM preserves textural evidence for different deformation and dynamic weakening mechanisms operative during seismic slip. We combine our microtextural observations and interpretations with prior timing constraints from hematite (U-Th)/He thermochronometry (McDermott and Ault, *in prep*) to capture different stages of FM microtextural and mechanical development.

2. Wasatch fault zone and hematite fault mirrors

The WFZ is an ~370 km-long, seismogenic, and segmented normal fault that demarcates the easternmost extent of Miocene-present Basin and Range extension in northeastern Utah (Machette et al., 1991). The southernmost end of the Brigham City segment of the WFZ juxtaposes footwall Paleoproterozoic hornblende- and biotite-bearing quartz monzonite gneiss and migmatite, amphibolite, and quartz-rich granitoid pegmatite against younger sedimentary rocks in the hanging wall (e.g., Fig. 1a, inset;
Crittenden and Sorensen, 1985). Full- and partial-segment-rupturing earthquakes of \(\sim M_w 7\) have occurred \(~1060-1500\) years during the Quaternary (DuRoss et al., 2016).

Figure 1: Structural geology of the Wasatch fault zone. (a) Geologic map of study area showing outcrop-scale structures and location of breccia-fault mirror pairs. After McDermott and Ault (in prep). Inset shows simplified map of regional tectonic setting. Red star denotes study area. Xfc-Farmington Canyon Complex, Z-Neoproterozoic rocks, Pz-Paleozoic rocks, Cz-Cenozoic rocks and sediments. Modified from Yonkee (2005). (b) Equal area plot of fault mirror attitudes (black) and striae (white circles). Arrows give hanging wall movement. Normal sense-black, reverse sense-blue. (c) Panorama of E-W trending normal fault zone (see (a) map). Red dashed lines outline Fe-oxide patches. White dashed lines denote fault surfaces. Note sample locations. Inset shows view along
principal fault zone with zone of fractures rock. (d) Hand sample photograph of W17-1H showing breccia and crosscutting veins and fault mirror material. (e) Field photograph of W17-1G. White arrow shows striae and hanging wall slip sense.

A ~300–400 m-wide fault damage zone crops out within the Brigham City segment of the WFZ and hosts chlorite breccia and phyllonite, rare gouge, localized Fe-oxide rich cataclasite, disseminated Fe-oxide alteration, and abundant hematite veins and FMs (Fig. 1a, c; Evans and Langrock, 1994). Hematite FMs and veins are principally distributed along mesoscale damage zone normal faults. Prior field study reveals hematite FMs are often localized within mm- to cm- thick hematite-cemented breccia or specular hematite veins (Fig. 1d, e; Evans and Langrock, 1994; Ault et al., 2015; McDermott et al., 2017). FM striae record complex, but dominantly down-to-the-west normal slip, consistent with Basin and Range extension accommodated along the WFZ (Fig. 1b; Evans and Langrock, 1994; McDermott and Ault, in prep).

Prior geochemical, textural, and thermochronometric analyses indicate FMs likely accommodated seismic slip and transient elevated temperatures facilitated by asperity flash-heating. Localized coseismic temperature rise is supported by (1) the presence of surficial Fe$^{2+}$, indicative of high-temperature Fe reduction (Evans et al., 2014), (2) polygonal and(or) lobate grains directly at the FM, suggestive of high-temperature sintering and(or) annealing (Ault et al., 2015; McDermott et al., 2017), and (3) hematite (U-Th)/He data from FMs, which show spatially variable thermal resetting of the (U-Th)/He system (i.e, He loss), in regions characterized by high-temperatures textures. Temperatures >1200 °C are likely due to asperity flash heating (Ault et al., 2015; McDermott et al., 2017).
Hematite (U-Th)/He thermochronometry from a broader suite of specularite veins, FMs, and hematite matrix in mm-scale breccia underlying the FMs suggest slip surface development is multi-staged and involves fault reactivation and slip localization (McDermott and Ault, *in prep*). Thermal modeling of hematite (U-Th)/He thermochronometry data indicates some undeformed specularite veins and hematite in fault breccias are likely as old ~1680 Ma. Others may have mineralized during Cretaceous thrusting (~110–75 Ma) and Basin and Range extension (~20–≤12 Ma) (Fig. 1a, inset; McDermott and Ault, *in prep*; Yonkee et al., 2019). FMs, localized within veins and breccia, most recently accommodated normal slip from ~18 Ma to ≤4.5 Ma and at ambient temperatures of ≤120 °C (≤4–5 km depth, assuming a 25–30°C/km geothermal gradient; Ault et al., 2015; McDermott et al., 2017; McDermott and Ault, *in prep*).

3. Samples and Methods

We analyzed three FM samples (Fig. 1; W17-1G, W17-1H, and W17-1J) with multi-scale microscopy, image analysis, and EBSD mapping to characterize microtextural evolution and grain-scale deformation mechanisms. These samples were previously analyzed with high spatial resolution hematite (U-Th)/He thermochronometry to target distinct textural domains by McDermott and Ault (*in prep*). Samples W17-1H and W17-1G comprise hematite that likely precipitated in the Paleoproterozoic and Miocene (~1680 Ma and ~14 Ma), and W17-1J comprises Miocene (~20 Ma) hematite.
3.1 Microscopy

We prepared aliquots for microtextural analysis by creating epoxy mounts of each of our samples. Samples were cut perpendicular to the FM surface and oriented so that the dominant striae direction were parallel to mount surface, with slip sense towards the left. Mounts were imaged by reflected light at 73x magnification and scanning electron microscopy (SEM) at magnifications ranging from 500x to 80,000x. We additionally mounted unpolished FM chips on double-sided copper sticky tape for map and oblique perspectives, and created thin sections of samples W17-1G and W17-1J to aid in microtextural characterization and image analysis (Fig. S1). SEM microscopy was conducted at the Utah State University (USU) Microscopy Core Facility (MCF) using a FEI Quanta FEG 650 field emission SEM in high-vacuum mode. We dual imaged our samples with secondary electron (SE) and backscatter electron (BSE) detectors to highlight “topographic” and geochemical data, respectively. BSE imaging is particularly useful for distinguishing hematite from other phases and highlighting grains boundaries in areas of uniform polish.

3.2 Quantitative image analysis

We conducted image analysis of select reflected light and SEM images from hematite FM samples to quantify their grain size distribution and fabric characteristics. Grains were traced using an iPad and Adobe Draw from a combination of BSE, SE, and reflected light images from our epoxied mounts and plane-polarized-light images. Gray-

---

2 The Supporting Information for this chapter is provided in Appendix B
scale images were then imported into ImageJ (Abramoff et al., 2004) and analyzed with grain distribution algorithms included in the software. We measured grain area, short- and long-axis lengths, and long-axis orientations from the best-fit ellipses to grain outlines. We quantified grain size distribution from microscopy images following standard image analysis procedures by calculating “equivalent diameter (eqD)”, or the diameter of a circle having the same area as a grain. A library of our analyzed images and tracings is provided in the Supporting Information (Fig. S2).

Grain size distributions in fault rocks and precipitates reflect a variety of factors, including fragmentation mechanisms, initial grain size distribution and associated precipitation kinetics, and(or) chemical modification (Epstein, 1947; Turcotte, 1986; Sammis et al., 1987; Jébrak, 1997; Kile et al., 2000). Grain size distributions in many fault rocks can be described by the power-law relationship and a fractal dimension (D):

$$N(eqD) = c \ast eqD^{-D}$$  \hspace{1cm} (1)

where \(N(eqD)\) is the frequency of grains with size eqD, and \(c\) is an arbitrary constant (Turcotte, 1986; Sammis et al., 1987). Power-law grain size distributions plot as straight lines in log \(N(eqD)\)-log (eqD) space, but may exhibit “rollovers” at higher and lower grain sizes as a function of detection and(or) the grinding limit of deformed grains (Keulen et al., 2007). D-values correspond to the slope of these lines and can be 2- or 3 dimensional \((D_{2D} \text{ and } D_{3D}, \text{ respectively})\), depending on whether grain size is measured from an image or by sieving and weighing, and are related by the relationship \(D_{2D}+1=D_{3D}\) (Mandelbrot, 1982; Heilbronner and Keulen, 2006). Lognormal distributions are also observed in fault-related rocks and are arcuate in log \(N(eqD)\)-log (eqD) space (Epstein, 1947; Blenkinsop, 1991; An and Sammis, 1994; Kile et al., 2000; Melosh, 2019). Here,
we calculate \( N(\text{eqD}) \) as the number of grains \( \geq \) a particular \( \text{eqD} \), normalized by image area, and plot this value against \( \text{eqD} \) in log-log space (Blenkinsop, 1991). For power-law distributions, we also calculate \( D_{2D} \) and associated uncertainty by regression through linear segments. For grain size distributions not characterized by a power-law relation (i.e., Eq. (1)), we do not calculate \( D_{2D} \) and instead formally test for lognormal distribution at 95% confidence via Kolmogorov-Smirnov test.

Conceptually, \( D \)-values reflect the relative abundance of smaller versus larger grains, with higher \( D \) values indicating a larger proportion of smaller grains. For example, cataclasis produces a greater number of smaller grains through spalling and attrition of larger grains compared to brecciation. Although \( D \)-values reported in the literature overlap, breccias or cracked grains typically have \( D_{2D} \) values of \( \sim <1 \) to \( \sim 1.4 \) (Storti et al., 2003; Keulen et al., 2007; Mort and Woodcock, 2008; Melosh et al., 2014; note that we convert literature values of \( D_{3D} \) to \( D_{2D} \)). Cataclasite and gouge \( D_{2D} \) values cluster at \( \sim 1.6 \), but range from \( \sim 1 \) to \( \sim 2.5 \), with \( D_{2D} \) increasing proportional to cataclasite maturity (Sammis et al., 1987; Marone and Scholz, 1989; Blenkinsop, 1991; Keulen et al., 2007). Modeling of gouge formation and experimental studies suggest \( D_{2D} \) values of \( \sim 1.5 \) to \( 1.7 \) indicate constrained comminution (fragmentation is dependent on neighboring grains), which also favors velocity weakening and slip localization (Sammis et al., 1987; Biegel et al., 1989; Mair and Abe, 2008). \( D \)-values are sensitive to the thickness of the deforming zone, confining pressure, strain, lithology, and initial variations in the grain size distribution. Thus, fault reactivation and(or) progressive deformation processes, as well as neomineralization, can complicate straightforward interpretation of the physical significance of \( D \)-values (e.g., Keulen et al., 2007). \( D_{2D} \) values can also be modified by
post-deformation processes. For example, processes that preferentially effect smaller grains, such as pressure solution, can influence $D_{2D}$ values (Melosh, 2019). Detailed microtextural context is thus critical to interpretation of $D_{2D}$.

Alternative grain size distributions, such as lognormal distributions, may reflect different fragmentation and/or precipitation processes. Lognormal grain size distributions are predicted by models of unconstrained comminution (grain fragmentation is independent of neighboring grain size; Epstein, 1947; Sammis et al., 1987; Blenkinsop, 1991). Past studies have suggested this may result from deformation contemporaneous with dilation, such as what may occur in the presence of high-fluid pressures (Luther et al., 2013). Additionally, synkinematic precipitation of matrix minerals may also exhibit lognormal grain size distributions for nucleation rates that decay with decreasing clast surface area (Kile et al., 2000).

We evaluated grain shape preferred orientation (SPO) following the workflow outlined in French and Chester (2018). We measured the angle ($\theta$) between the long-axis of clast $i$ and the principal slip surface, where -$\theta$ denotes counterclockwise rotations and +$\theta$ denotes clockwise rotations, and calculated the vector mean orientation ($\hat{V}$) following:

$$\hat{V} = \frac{1}{2} \tan^{-1} \frac{\Sigma_{i=1}^{N} \sin 2\theta_i}{\Sigma_{i=1}^{N} \cos 2\theta_i}$$

We calculated vector strength ($\bar{a}$) to quantify relative SPO development following (Chayes, 1975):

$$\bar{a} = \frac{1}{N} \left[ (\Sigma_{i=1}^{N} \sin 2\theta_i)^2 + (\Sigma_{i=1}^{N} \cos 2\theta_i)^2 \right]^{1/2}$$

Cladouhos (1999) showed randomly oriented fabrics may have $\bar{a} \leq 0.17$, and we thus also employed this value as our cutoff for evaluating whether SPO is present in our samples. All calculations referenced above, including interactive selection of grain size

...
distribution segments for calculation of D-values, were implemented in MATlab. Our 
script is provided in the Supporting Information.

3.3 Electron Backscatter Diffraction

Epoxied mounts previously analyzed by SEM were repolished at 1 μm grit to 
remove carbon-coating and subsequently polished under colloidal silica suspension with 
a Vibro Met 2 vibratory polisher. Hematite crystal orientation and phase data were 
collected at 70° sample tilt with an Oxford NordlysMax detector and analyzed with the 
Inorganic Crystal Structure Database. Data were acquired at low-vacuum (typically ~0.05 
Torr) conditions with 20 kV accelerating voltage and at an ~3.1 Hz collection rate. 
Because of the delicate nature of our samples and large grain size variation between 
analyzed areas, EBSD map areas corresponding exactly to initially SEM imaged areas 
(i.e., section 3.1) were impossible to maintain. Analyzed areas nearest the FM surface 
range from 375 to 600 μm² and data was collected with a 0.1 μm step size. Data collected 
from coarser-grained and more texturally heterogenous areas below the fault mirror range 
in area from ~400 to 2800 μm² and were collected with a 0.5 μm step size.

EBSD data were post-processed with AztecCrystal following best practices 
outlined in Prior et al. (2009). We applied automatic noise reduction algorithms to target 
and remove mis-indexed pixels (i.e., phase or orientation outliers of ≤4 pixels) and zero 
solution pixels. All data are presented in the sample reference frame, where x is 
horizontal (or parallel to slip), y is vertical, and z is perpendicular to the mount face. We 
report our results as maps overlain band contrast (related to indexing quality), inverse 
pole figure orientations in the sample x-direction (IPFx), and grain boundaries at 2° and
10° misorientation. Pole figures were generated by automated grain detection and exported at one-point-per-grain. Only grains comprising ≥10 pixels were included. Additionally, we combined adjacent grains with 85 ± 5° misorientation about the 02-21 plane into single grains, as these likely represent rhombohedral twinning (Ávila et al., 2015).

4. Results

4.1 Field and microtextural observations

At our sampling locale, minor mesoscale normal faults are manifest as zones of highly fractured rock flanked by planar bedrock surfaces with localized ~1 to ~4 m² patches of ruddy hematite (Fig. 1a-c; McDermott and Ault, in prep). The damage zones of these mesoscale faults exhibit little hydrothermal alteration (e.g., Fig. 1e), but contain abundant hematite FMs with underlying hematite breccia (Fig. 1d, e). FMs are subparallel to larger bedrock planar surfaces and exhibit oblique normal and reverse slip (Fig. 1b).

We define three microtextural domains in hematite FM samples on the basis of textural similarity and relative age. Domain 3 is hematite-cemented breccia located farthest from the FM surface and cut by other subsequently described domains. At field-scale, Domain 3 is manifest as mm- (W17-1J, -1G) to cm-scale (e.g., W17-1H) matrix-supported hematite breccia in sharp contact with Farmington Canyon Complex host rock (Figs. 1d, e; 2). Angular breccia clasts are both poly- and monomineralic and largely consist of plagioclase and orthoclase feldspar and quartz derived from adjacent host rock. Breccia clasts are ~100 to 3000 μm wide and ~250 to 4000 μm-long (Fig. 2a, f, k).
Matrix hematite comprises high-aspect-ratio, ~2 μm-thick x ≥10 μm-long plates with little evidence for deformation. Breccia clasts exhibit “jigsaw” geometry (sensu Sibson, 1986) and locally show injection of hematite matrix into veins approximately perpendicular to clast boundaries (Fig. 2c, m). In sample W17-1J, we observe two cross-cutting hematite breccia generations (Fig. 2f). Hand sample and field observations of samples W17-1H and W17-1G show specular hematite veins cross-cutting hematite breccia (Fig. 1d, e).

Domain 2 is hematite cataclasite and crosscuts Domain 3 hematite-cemented breccia. In all samples, Domain 2 ranges between ~50 and ~300 μm-thick and has a cuspatelobate lower contact with Domain 3 (Fig. 2a, f, k). Domain 2 comprises high-aspect ratio (~1 μm-thick x ≥5 μm-long) and relatively undeformed hematite plates interspersed with more equant and angular grains (~0.1-1.5 μm in diameter), indicative of limited cataclasis (Fig. 2). We note W17-1J has the highest proportion of comminuted hematite in Domain 2. Wall-rock quartz or feldspar clasts are rare in this domain (e.g., W17-1H; Fig. 2l). Domain 2 is associated with subsidiary shear fractures which extend into, and offset breccia clasts within Domain 3 (Figs. 2a; S1). On the basis of angular relationships between these fractures and Domain 2 and shear sense relative to the dominant FM striae set, we interpret these features as R, R’, and potentially P’ shears. In W17-1H, we observe two cross-cutting layers of Domain 2, separated by a sharp contact, with the lower layer containing clasts of Domain 3 hematite breccia (Fig. 2l).
Domain 1 includes the FM surface and underlying ~5 to 20 μm of material and exhibits the most textural heterogeneity (Figs. 2o; 3). Common to all samples is finely comminuted (~0.5 μm), subangular, and equant hematite grains termed hematite ultracataclasite (Fig. 2o, inset; 3f, i, k). Domain 1 exhibits a gradational relationship with Domain 2 (e.g., Figs. 2o, 3i), with grain size decreasing towards the FM surface.

Domain 1 in sample W17-1H consists exclusively of hematite ultracataclasite (Fig. 2o, inset), but Domain 1 in W17-1G and W17-1J are microtexturally diverse. The most common W17-1G Domain 1 microtexture is polygonal and triple-junction-forming hematite grains at the FM surface (Figs. 2e; 3a-e). W17-1G polygonal grains occur in a band ~5–20 μm-thick and ~>100 μm-long (parallel to the FM). Polygonal grains show systematically decreasing grain size (eqD), from ~2.5 μm to ≤0.5 μm, with distance from the FM surface (Figs. 3b; 4). At lower magnifications, polygonal grains appear to form a sharp contact with hematite plates in Domain 2 (Fig. 3b). At higher magnifications, however, plates from Domain 2 can are observed as intruding at high-angle into Domain 1 (Fig. 3b-d). Plate tips show polygonal grain formation, but relict plate boundaries are
still recognizable (Fig. 3d). In W17-1J, Domain 1 comprises deformed hematite plates (Fig. 3g), polygonal grains (Fig. 3h, i, j), and ultracataclasite (Fig. 3i, k) directly at or near the FM surface. These different textures and crystal morphologies are commonly laterally adjacent within an ~5-10 μm-thick band (Figs. 2i, j; 3g, h, i). Zones of cataclastic particles and polygonal grains have reduced apparent visible porosity and also mantle larger plates (Fig. 3h-k). The frequency of plates with finely comminuted or polygonized edges decreases with distance from the FM surface, effectively grading to Domain 2 plates and cataclasite (Fig. 3i).

**Figure 3: Detail of Domain 1 microtextures in samples W17-1G and W17-1J.** (a) Polygonal grains (poly. grains) from W17-1G. Black lines outline triple-junction-forming grain boundaries (b) Polygonal grains from W17-1G showing fault mirror-perpendicular grain size gradient and Domain 2 plates. (c, d) Detail of Domain 1 polygonal grains in (b) overprinting Domain 2 hematite plates. Black lines in (d) outline relict plate boundaries.
(e) Map view image of W17-1G polygonal grains. (f) W17-1G ultracataclasite (u-cataclasite). (g, h) Detail of Domain 1 in W17-1J at progressive magnification levels. Black dashed lines in (h) outline triple-junction-forming grain boundaries. (i) Oblique view secondary electron (SE) image of W17-1J Domain 1 and Domain 2. Black dashed lines denote relict plate boundaries. (j) Detail of plate margin with polygonal grains. Black dashed lines outline triple-junction-forming grain boundaries. (k) Detail of W17-1J Domain 1 ultracataclasite. White dashed lines in all images show fault mirror surface.

Figure 4: W17-1G polygonal grain size patterns. Equivalent diameter (EqD) of polygonal grains with distance perpendicular to fault mirror at three locations (circles). Underlying pink histogram shows EqD distribution of W17-1G ultracataclasite (see Figure 5).

4.2 Image analysis results

We targeted representative textures for each microtextural domain for image analysis (Fig. S2). For select samples, we analyzed replicate images to evaluate the reproducibility of our results. In Domain 3, we analyze clasts of wallrock (e.g., quartz and feldspar) and hematite matrix. In Domain 1, we targeted primarily ultracataclastic grains and plates that reflect brittle deformation processes. Although there is abundant microtextural evidence for multiple deformation events encapsulated within domains 1 or
2, we purposefully restricted our analyses to areas where only one event is apparent. Image analysis results are summarized in Table S1 and Figures 5 and 6.

Domain 3 wall-rock clasts exhibit similar distributions of eqD, aspect ratio, and orientation and a statistically significant SPO (except for W17-1H; Table 1; Fig. 5). Jigsaw clast geometries indicate little clast rotation, however, suggesting SPO development is in part a function of approximately consistent hematite vein orientations that compose the breccia matrix (Figs. 2b, c, m; S1).

Hematite analyzed in all domains highlight systematic patterns in grain size/aspect ratio reduction and SPO development. Aspect ratio decreases from an average of ~6, ~4, and ~2 for Domains 3, 2, and 1, respectively (Table S1; Fig. 5). Domain 3 aspect ratios are close to the theoretical limit for euhedral hematite, consistent with these hematite plates experiencing limited deformation relative domains 1 and 2 (e.g., Nesse, 2012). Hematite average eqD also decreases between domains, from ~4 μm for Domain 3, ~2 μm for Domain 2, and ~0.5 μm for Domain 1 (Table S1; Fig. 5). Distributions of aspect ratio and eqD are widest for Domain 3, and narrowest for Domain 1 (Fig. 5). Most analyzed images and samples exhibit statistically significant SPO (a=0.18 to 0.46; Table S1), except for ultracataclasite in W17-1J (Table S1; Fig. 5). SPO strength is weakly correlated with domain. Domains 1 and 2 typically show a greater degree of SPO development than Domain 3 hematite matrix, and Domain 2 has stronger SPO than Domain 1 (Table S1; Fig. 5).
Microtextural domains exhibit variability in grain size distribution (Fig. 6).

Domain 1 ultracataclasite and Domain 3 breccia clasts in all samples exhibit power-law grain size distributions (Fig. 6a, b). These data form a straight line in log(N>eqD)-log(eqD) space, consistent with theoretical expectations for a power law distribution (Fig. 6a, b; Sammis et al., 1987). Power law distributed data also show a “rollover” to shallower slope at approximately similar eqD, perhaps reflecting the detection limit set by the magnification of our images or the grinding limit of deformed grains (e.g., Fig. 6b, Keulen et al., 2007). Data we infer as power-law distributed fails lognormal tests if we only incorporate linear segments, but passes lognormal tests if we include what we subjectively consider to be grinding or detection limit “rollover” (the exception is W17-1J, which does not pass tests for a lognormal distribution regardless of data used; Table S1). Values of D_{2D} calculated only for these data and by regression of steep linear segments (i.e., excluding data past the rollover) range from 1.6 ± 1.0 to 2.5 ± 1.0 for Domain 1 ultracataclasite and 1.3 ± 1.0 to 1.9 ± 1.0 for Domain 3 breccia clasts (Fig. 6a; Table S1). In contrast, hematite grain size distributions from Domain 2 cataclasite and Domain 3 hematite matrix are notably curved in log(N>eqD)-log(eqD) space and pass statistical tests for lognormal distributions (Fig. 6a, b; Table S1).
**Figure 6: Grain size distribution data.** (a) Plot of eqD vs total number of grains (N) greater than eqD. Each sample is offset along Y-axis to enhance visualization. Data are binned by sample (colors) and microtextural domain (shapes). Distribution type denoted (see (b) for details). 2-dimensional fractal dimensions ($D_{2D}$) ± 1σ reported only for power law distributions (Table S1). Values are plotted near respective distribution data and in color corresponding to sample. (b) Schematic showing expected from for different grain size distribution type. Note power law distribution plots as a straight line in log (eqD)-log(N>EqD) space but may show rollover reflecting grinding and(or) detection limits (cf. Keulen et al., 2007). (c-f) Example line drawings of grain boundaries from Domain 1 ultracataclasite (c), Domain 2 catalcasite (d), Domain 3 breccia clasts (e), and Domain 3 hematite matrix (f) from W17-1H. Corresponding distribution data for each panel denoted in (a).

### 4.3 Electron backscatter diffraction results

We acquired EBSD data from a total of seven areas from the three samples. We collected data from both Domain 1 polygonal grains and ultracataclasite from W17-1G, Domain 1 polygonal grains and collocated or underlying plates from W17-1J, and Domain 1 ultracataclasite from W17-1H. In addition, we collected one map of Domain 2
cataclasite from each of our samples. Domain 3 was not analyzed. Indexing rates were typically better than 85%, with the exception of W17-1G ultracataclasite for which ~35% of the map area yielded no solutions due to poor sample quality (i.e., compare black areas to colored areas in Figure 7). We were unable to directly analyze material directly at and below the FM surface due to plucking of grains at the sample-epoxy interface during additional polishing, and thus Domain 1 maps are ~2–5 μm below the FM surface. In some samples, the result is that our mapping includes the gradation between domains 1 and 2. Results from EBSD analyses are presented as IPFx maps and pole figures in Figures 7 and 8. Pole figures are constructed for c- and a-axes (see schematic in Figure 7; Fig. 8).

In sample W17-1G domain 1, we observe polygonal grains and larger, deformed and twinned hematite grains (Fig. 7b). Larger grains host a high intensity of subgrain boundaries (<2° misorientation; Fig. 7b). Polygonal grains mantle larger plates, and some grain boundaries between plates and polygonal grains appear serrated (Fig. 7b). Polygonal grains have weak to moderate, bimodal CPO, with c- and a-axes aligned parallel and perpendicular, respectively, to the FM surface and a-axes aligned perpendicular (Fig. 8a). In sample W17-1J, Domain 1 polygonal grains and associated plates form discrete bands, although some polygonal grains can be observed mantling less deformed plates with intragranular orientation contrasts (Fig. 7c). Polygonal grains and plates both exhibit CPO, with c-axes aligned obliquely to the FM (Fig. 8b, d). Domain 1 ultracataclasite from W17-1G and W17-1H both lack significant CPO (Figs. 7a, d; 8c, e).
Figure 7 (previous page): Inverse pole figure (IPF) maps for domains 1 and 2. Maps are overlain band contrast, IPF, and grain boundaries. Coloring gives crystallographic orientation with respect to sample X-direction (IPFx). IPFx coloration scheme, grain (>10°, black) and subgrain (>2°, gray) boundaries, sample coordinates, and schematic hematite plate with crystallographic axes (gray arrows) shown in upper left. (a) W17-1G ultracataclasite. Black areas show no solution. (b) W17-1G polygonal grains and relict clasts. White arrows show polygonal grain example, asterisked white arrow denotes twinning. (c) W17-1J Domain 1 polygonal grains with plates. White arrow shows polygonal grain examples. Asterisked white arrow denotes serrated grain boundaries. (d) W17-1H ultracataclasite. (e-g) Maps of Domain 2 cataclasite for W17-1G (e), W17-1J (f), and W17-1H (g).

Figure 8: Pole figures for domains 1 and 2. Equal area plots (upper hemisphere projection) of crystallographic axes (c-{0001} and a-{10-10}, see Figure 7). Pole figures correspond to entire map area in Figure 7 unless otherwise indicated. (a) W17-1G polygonal grains and clasts (Fig. 7b). (b) W17-1J polygonal grains only (Fig. 7c). (c) W17-1G ultracataclasite (Fig. 7a). (d) W17-1J polygonal grains plus deformed plates (Fig. 7c). (e) W17-1H ultracataclasite. (f-g) W17-1G (f), W17-1J (g), and W17-1H (h) ultracataclasite (Fig. 7e-g). Contouring is by multiples of uniform density (MUD). One-point-per-grain. Grains per plot denoted (n).
Grain boundary and orientation mapping in Domain 2 further highlight microtextural differences between samples. Sample W17-1G comprises a larger portion of twinned hematite plates (Fig. 7e). Moderate CPO in this sample is defined by c- and a-axes perpendicular and parallel to the shear plane, respectively (Fig. 8f). Domain 2 cataclasite in samples W17-1J and W17-1H show decreasing proportion of subrounded particles (Fig. 7f, g). Domain 2 in W17-1J has a weak CPO (Fig. 8g). Domain 2 in W17-1H has a CPO, with c-axes defining a girdle perpendicular to the FM surface and a-axes clustering parallel to the FM (Fig. 8h). In summary, the weakest CPO in our samples is exhibited by Domain 1 ultracataclasite, whereas Domain 1 polygonal grains and Domain 2 cataclasite have weak to moderate CPO.

We conducted misorientation analyses (Fig. 9) of Domain 1 EBSD-mapped regions of W17-1G (Fig. 7b), W17-1J (Fig. 7c) and ultracataclasite from W17-1H (Fig. 7d). Map coloration is a function of the orientation difference between any individual pixel and the mean orientation of a grain (termed misorientation; Fig. 9a-c). Histograms of misorientation data for these samples are generated by (1) measuring misorientation between two random grains anywhere in the analyzed area and (2) measuring misorientation between two neighboring grains (Fig. 9d-f). Our samples show grains with low (<2°) and high (~15°) intragranular misorientations, although some W17-1G hematite grains have misorientations as high as 30°. Polygonal grains are generally misorientation free but form serrated boundaries with larger and misorientation-riddled grains (Fig. 9a, b). Sample W17-1H ultracataclasite grains have low misorientation, although larger grains exhibit localized areas of higher misorientation near grain tips (Fig. 9c). Misorientation angle histograms also reveal differences in intragranular strain
between samples, with W17-1G, W17-1J, and W17-1H showing progressively less misorientation angles >2° but <10°, indicative of subgrain formation (Fig. 9d-f). All samples show a strong peak at 85° misorientation in neighboring grains, consistent with r-twinning (Avila et al., 2015) in these samples. Mismatch between random pair and theoretical misorientation distributions in W17-1G and W17-1J, versus congruency in W17-1H, is consistent CPO development in the former two samples (Fig. 9a-c).

Figure 9: Domain 1 misorientation analysis results. (a-c) Maps of intragranular misorientation. Coloration shows orientation of any individual pixel relative to grain mean. Grain and subgrain boundaries overlain (legend as in Figure 7). Maps shown are for W17-1G (a) and W17-1J (b) polygonal grains and W17-1H (c) ultracataclasite. White arrows in (a), (b) show high misorientation grains flanked by low misorientation grains with highly serrated boundaries. Asterisked white arrows in (c) show larger grains with intragranular misorientations focused at grain tips. (d-f) Misorientation histograms corresponding to maps in (a-c). Orange is random grains anywhere in map (n=10,000
iterations). Blue shows neighboring grain misorientations. Solid black line is theoretical distribution for a truly random sample with triclinic symmetry (i.e., no crystallographic preferred orientation; Wheeler, 1971). Peaks at 85° likely represent r-twinning (Avila et al., 2015).

5. Discussion

5.1 Interpretation of microtextural observations

Microtextures reveal a multi-stage development, culminating in seismogenesis, of what are now FMs. Domain 3 is (1) farthest from the FM surface, (2) the first to develop, and (3) similar in all analyzed samples. Matrix-supported breccia, jigsaw cast geometry, and hematite injection veins at high angles to clast boundaries (Fig. 2c, h) are consistent with overpressurized fluids contributing to brecciation. A clast $D_{2D}$ value of 1.3 from sample W17-1H is consistent with those reported for other breccias formed by hydraulic fracturing (Jebrak, 1997; Mort and Woodcock, 2008). Higher $D_{2D}$ values of 1.5–1.9 from W17-1G and W17-1J indicate a higher proportion of smaller grains in these samples than is typical for implosion breccias (Fig. 6a; Table S1). This may reflect grain size reduction from overprinting deformation, or other factors that cause variability in $D_{2D}$ that are difficult to constrain with available data (e.g., confining pressure; Keulen et al., 2007).

The grain size distribution of hematite in the matrix of Domain 3 breccia is lognormally distributed (Fig. 6a). For comparison, lognormal grain size distributions in calcite reflect decaying rate- and surface-controlled crystallization from supersaturated solution and may accompany rapid pressure drops associated with fracturing and brecciation (Kile et al., 2000). Prior study of siderite- and calcite-cemented cataclasite also suggested lognormal distributions reflect rapid and potentially coseismic precipitation (Boullier et al., 2004). Assuming hematite growth mechanisms are similar to
carbonate, we suggest hematite precipitation occurred by analogous processes during rapid cementation of breccia fragments. Multiple generations of Domain 3 implosion breccia in sample W17-1J thus indicate repeated fault reactivation, transient fluid pressure, and hematite mineralization (Fig. 2f).

Domain 3 is cut by hematite veins, which we interpret as providing the material for microtextural domains 2 and 1 (Fig. 1d, e). Domain 2 has sparse clasts of Farmington Canyon Complex-derived quartz and feldspar compared to Domain 3, contains rare clasts of Domain 3, and has a distinct boundary with Domain 3 (Fig. 2a, b, f, g, k, l). These relationships indicate domains 3 and 2 reflect discrete hematite precipitation events. Because Domain 2 hematite is comminuted and deformed with little evidence for neomineralization, it must have existed prior to cataclasis within this layer (Fig. 2e, j, o).

We suggest deformation in Domain 2 reflects dilation and unconstrained comminution, characteristic of fluidized granular flow. Three observations support this interpretation. First, Domain 2 forms a cuspatelobate boundary with underlying Domain 3 (Fig. 2b, g), and in some cases shows vein-like geometry (Fig. 2f). Cuspatelobate boundaries in fault rocks are commonly interpreted as reflecting “bulk” fluid-like behavior (Otsuki et al., 2003; Fondriest et al., 2012). Second, hematite plates in Domain 2 preserve SPO and weak CPO (Fig. 5; 8). During fluidization, non-equant grains may align parallel to flow lines, resulting in SPO and(or) CPO (Borradaile, 1981; Ujiie et al., 2011). SPO has been observed in experimentally generated fluidized gouges (French and Chester, 2018). Third, hematite plates are comminuted, but still preserve high-aspect ratios and a lognormal grain size distribution (Figs. 2e, j, o; 6; Table S1). These observations imply low interparticle stresses and dilation accompanying unconstrained
comminution, potentially as a result of high pore fluid pressures (Sammis et al., 1987; Luther et al., 2013). The degree of comminution (e.g., aspect ratio and mean grain size) and SPO development in Domain 2 varies within and between our samples, which may reflect initial grain size distribution effects or variability in dilation over short (100s of μm to mm) spatial scales (Figs. 2; 5; Table S1).

Domain 1 exhibits evidence for concomitant brittle and plastic deformation. Domain 1 ultracataclasite is marked by intense grain size reduction, production of a large volume of nanoparticles (defined as particles with eqD <0.1 μm), and formation of approximately equant hematite grains (Fig. 5). This deformation occurred by predominantly brittle intra- and inter-granular fracturing (Figs. 2o, inset; 3f, k).

Ultracataclasite is also characterized by a weaker SPO and CPO relative to Domain 2, reflecting sliding and rotation between particles and cataclastic granular flow (Borradaile, 1981). The grain size distribution of the ultracataclasite grain size is characterized by a power law with D_{2D} ranging from 1.6 to 2.8 (Fig. 6). The lower bound of these values is identical to the theoretical limit for constrained comminution, whereas higher values are consistent with those reported for more mature cataclasite (Sammis et al., 1987; Storti et al., 2003; Keulen et al., 2007). The gradational contact between Domain 1 ultracataclasite and underlying Domain 2 cataclasite suggests contemporaneous deformation in these microtextural domains (e.g., Fig. 2i, o, inset).

Polygonal grain morphologies, intragranular misorientations, and CPO indicate local plastic deformation in Domain 1, likely indicative of elevated temperatures and thus seismic slip. Domain 1 polygonal grains from W17-1G are similar to those produced in hematite deformation experiments at temperatures of 800 to >1100 °C and lower strain
rates between $10^{-4}$ and $10^{-6}$ s$^{-1}$ (Fig. 3a; Siemes et al., 2003, 2011). EBSD mapping of Domain 1 in this sample, and to a lesser degree, W17-1J, shows the formation of subgrains at the margins of preserved larger hematite plates and serrated, bulging grain boundaries separating plates with high intragranular misorientations (Figs. 7b, c; 9a, b). These textures collectively suggest crystal plastic deformation by grain size insensitive (GSI) deformation mechanisms (i.e., migration of dislocations; Siemes et al., 2003; 2011). Development of CPO within polygonal grains relative to ultracataclasite in W17-1G suggests that this process was not uniformly operative across the FM surface, however (Fig. 8a, c). Because comparative experimental GSI deformation microtextures were formed at much lower strain rates than our samples likely experienced, we suggest 800 to 1100 °C is the minimum temperature required by our observations.

In sample W17-1G, polygonal grain eqD systematically decreases with perpendicular distance from the FM surface, and the majority of polygonal grains are larger than co-occurring cataclastic particles (Figs. 3b; 4). In the absence of EBSD data, this evidence for grain growth implies rapid grain size sensitive (GSS) deformation by diffusion and sintering in Domain 1, which by itself does not typically result in CPO (e.g., Rutter et al., 1994). We observe an overprinting relationship between polygonal grains and Domain 2 plate tips (Fig. 3c, d). This suggests continued growth of at least some polygonal grains post-deformation, most likely by sintering along fault-mirror-perpendicular temperature gradients immediately following seismic slip (e.g., Ault et al., 2019).

Our interpretation of polygonal grains reflecting GSI deformation, overlapping with or followed by GSS deformation is supported by results of prior experimental work.
First, static annealing experiments on quartz deformed by dislocation creep shows growth of recrystallized grains, but only moderate reduction of pre-annealing CPO and misorientation density (Heilbronner and Tullis, 2002). Second, rotary shear experiments in carbonates, stopped at different displacements and thus stages of microtextural evolution, show GSI and GSS mechanisms may operate within the same slip event (Pozzi et al., 2019). GSS mechanisms occur later and persists in the post-slip period (Pozzi et al., 2019). Because sintering is a diffusive and GSS process (Rutter et al., 1994), this would explain our observed CPO in smaller polygonal grains, and that scant larger grains are those with textural evidence for GSI processes in EBSD maps (Fig. 9a).

Samples W17-1J and W17-1H also show some evidence for plastic deformation within Domain 1, but it is more limited than W17-1G. Polygonal grains in Domain 1 of W17-1J, are rare and occur at the margins of larger plates within cataclastic particles (Fig. 3i). We did not observe polygonal grains in W17-1H (Fig. 3h, i, j). Similarly, misorientation angle histograms show progressively low, moderate, and high <10° misorientations and thus variable magnitude of plastic deformation across W17-1H, W17-1J, and W17-1G (Fig. 9d-f). In W17-1J, large and small grains have serrated grain boundaries and CPO, but this texture is not pervasive. It is also difficult to uniquely separate the CPO of W17-1J polygonal grains from that imparted by precursor hematite plates (Figs. 8; 9b). We speculate these textural relations reflect polygonal grain formation by incipient and highly localized GSI creep mechanisms. If GSS creep by sintering occurred in this sample, then we would expect more extensive zones of polygonal grains. In W17-1H, rare intragranular misorientations concentrated at the tips of larger grains suggest limited crystal plastic deformation and GSI creep, but these
processes apparently did not advance far enough to impart a strong textural signature. Thus, all samples exhibit evidence for plastic deformation related to the higher temperatures attained during seismic slip, but the extent of these inferred mechanisms vary within individual samples and between samples. We interpret the combination of these textures in association with ultracataclasite indicating brittle deformation reflects transient and spatially variable temperature rise from coseismic frictional heating.

5.2 Interpretation of dynamic weakening mechanisms

Our microstructural observations are consistent with multiple dynamic weakening mechanisms operative in concert with coseismic temperature rise during progressive development of domains 2 and 1. Evidence for fluidization in Domain 2 suggests reduction in effective normal stress through either acoustic fluidization from normal interface vibrations along the shear zone walls, or thermal pressurization of pore fluids related to ongoing frictional heat (Melosh, 1979; Lachenbruch, 1980; Otsuki et al., 2003; Rice, 2006; Ujiie et al., 2011). Coincident with this deformation, microstructures and grain size distributions suggest development of ultracataclasite via constrained comminution in some regions of Domain 1. Cataclasis is not uniquely weakening and is independent of strain rate (e.g., Keulen et al., 2007). But, progressive constrained comminution may activate interparticle slip, resulting in strain localization, enhanced temperature rise, and weakening (Biegel et al., 1989). The lack of CPO in Domain 1 ultracataclasite supports intergranular slip and bulk flow (Fig. 8; Hutter and Rajagopal, 1994). In our samples, inferred cataclastic flow and interparticle slip of particles <0.1 μm may be analogous to dynamic weakening by nanoparticle lubrication documented in
carbonate rocks (Han et al., 2011). Sample W17-1H domains 2 and 1 are dominated by cataclasite and ultracataclasite, but it does show evidence for limited plastic deformation that requires seismic slip on this surface (Figs. 2o; 9c). In the absence of other dynamic weakening mechanisms suggested by more widespread plasticity (discussed below), fluidization and nanoparticle lubrication likely facilitated seismic slip on some WFZ FMs.

Asperity flash heating to temperatures of \( \sim 300 \, ^\circ \text{C} \) to \( \geq 1200 \, ^\circ \text{C} \) is an additional likely weakening mechanism operative on WFZ FMs (Evans et al., 2014; Ault et al., 2015; McDermott et al., 2017). He loss in hematite is observed in aliquots with patches of polygonal grains, interpreted to reflect the footprints of paleoasperities (McDermott et al., 2017). Non-uniform development of high-temperature polygonal grains on our samples is like those previously studied, suggesting asperity flash heating processes and associated dynamic weakening occurred on these FMs. Evidence for localized activation of GSI (and potential GSS) deformation and viscous flow also implies plastic failure at asperities (e.g., Rice, 2006).

Many experimentally observed dynamic weakening mechanisms that facilitate rupture propagation are thermally activated (e.g., Rice, 2006; Di Toro et al., 2011; Yao et al., 2016). In our samples, we observe evidence for deformation by some temperature-dependent processes, such as GSI and GSS mechanisms. However, other inferred dynamic weakening mechanisms do not necessarily require locally elevated temperatures (i.e., nanoparticle lubrication in Domain 1 and potentially Domain 2, if acoustic vibration as opposed to thermal pressurization is responsible for fluidization). Microtextural heterogeneity at the scale of 100s of \( \mu \text{m} \) to mm on individual FMs implies variable
magnitudes and durations of elevated temperatures. This may reflect thermal gradients at localized asperities, as suggested above and noted on other thin slip surfaces (Goldsby and Tullis, 2011; Evans et al., 2014; Ault et al., 2015; Siman-Tov et al., 2015; DePaola et al., 2015; McDermott et al., 2017). Activation of weakening via asperity flash heating is also dependent on average temperature rise within a slipping zone (e.g., Rice, 2006). Thus, the variable width of the slipping zone, initial shear stress, and(or) total displacement may also play a role in creating observed microtextural variability.

5.3 Model for development of hematite fault mirrors in the Wasatch fault zone

We combine our microtextural interpretations with insights from prior experimental and thermokinematic models, as well as hematite (U-Th)/He thermochronometry, into a conceptual framework for the mechanical and textural evolution of WFZ hematite FMs. Figure 10 presents schematic strength and temperature curves with time and annotations of inferred weakening mechanisms. Because different strain accommodation mechanisms impact fault strength at different spatial and temporal scales, the generic term “strength” in Figure 10 refers specifically to static strength during hydraulic fracturing and dynamic strength (or shear stress) during coseismic slip. The timescales over which various processes occur is also non-linear. For example, individual hydraulic fracturing events and associated hematite precipitation that characterize Domain 3 may occur over minutes, but multiple events may occur at 1000-yr timescales (e.g., Williams et al., 2017). Coseismic slip and evolution of microtextures in domains 1 and 2 likely occurs over seconds. Given that most of our inferred weakening mechanisms
are thermally activated, we consider it more likely that our observations and model captures processes that facilitate rupture propagation rather than lead to its nucleation.

**Figure 10: Model for textural and mechanical evolution of Wasatch fault zone hematite fault mirrors.** Evolution of dynamic strength (solid, black), static strength (dashed, black), and temperature (solid, red) with time. Note time axis is nonlinear and includes deformation over 1000-yr timescales (“~ka”) and second timescales (“~s”). Million-year timing constraints after McDermott and Ault (in prep), denoted. Purple boxes show inferred timing of various deformation and weakening mechanisms (see text for details). GF-granular flow, AFH-asperity flash heating, TP-thermal pressurization, NPL-nanoparticle lubrication, GSI-grain size insensitive, GSS-grain size sensitive. Circled letters along stress/strength curve key to schematic block diagrams below. Scales are schematic for improved visualization.
Microtextural evolution of domains 1 and 2 and interpreted dynamic weakening mechanisms are placed in the context of dynamic shear strength evolution based on prior experimental work. Experimental studies of a range of rock types show initially increasing stress, prior to and at the onset of slip, related to accumulation of elastic deformation, followed by weakening to a new and lower steady state with the onset of high velocity and seismic slip, and post-seismic restrengthening (Dietrich, 1992; Di Toro et al., 2011; Calzolari et al., 2020). Temperature increases with time, but at a rate proportional to shear strength (Fig. 10, red curve). We include schematic block diagrams (a-g) at select stages to illustrate microtextural evolution within the context of our inferred mechanical evolution.

Domain 3 textures reflect hydraulic fracturing and are depicted in terms of static strength (Fig. 10, black dashed line). In this scenario, building fluid pressures within wallrock (Fig. 10, diagram a) result in bulk weakening, loss of strength during fracturing, and rapid strength recovery synchronous with hematite precipitation (Fig. 10, diagram b). At least one of our samples, W17-1J, exhibits evidence for multiple breccia generations, indicating this process may have occurred multiple times (Fig. 10, diagram c). Hematite (U-Th)/He dates measured from hematite isolated from breccia matrix suggests these hydraulic fracturing events are diachronous and occurred from ~1680 Ma (W17-1H, W17-1G) to ~20 Ma (W17-1J; McDermott and Ault, in prep). Hematite has a coefficient friction (μ) of ~0.28, much lower than that of gneissic host-rock (μ=0.6–0.7; Byerlee, 1968; Calzolari et al., 2020). Strength contrasts between hematite breccia and wall rock are thus likely to result in overall weakening of static frictional strength and cohesion at
the macroscopic scale (Fig. 10). Hematite breccia and veins serve as sites for later reactivation and slip localization.

Domains 2 and 1 developed contemporaneously. (U-Th)/He dates isolated from these domains suggest the original hematite precipitated at ~1680 Ma (W17-1G), ~20 Ma (W17-1J), and ~14 Ma (W17-1H; McDermott and Ault, in prep) (e.g., Fig. 1d, e; Fig. 10, diagram c). Coseismic slip in W17-1J and W17-1H must have occurred later than these dates; coseismic slip in W17-1G occurred at ~≥20 Ma (McDermott and Ault, in prep). Thus, coseismic slip in our samples as indicated on Figure 10 is Miocene (Fig. 10). Prior hematite (U-Th)/He thermochronometry data from a broader sample set also indicates some hematite FMs accommodated Plio-Pleistocene earthquakes (Ault et al., 2015; McDermott et al., 2017). In addition, WFZ FMs typically have multiple striations, and sample W17-1H shows some evidence for multiple generations of Domain 2 (Fig. 2l; Evans and Langrock, 1994; McDermott and Ault, in prep). FMs investigated here may thus also reflect multiple coseismic slip events from the Miocene to Pleistocene.

In our model, during initial slip, deformation was distributed over R, R’, and P’ shears overprinting Domain 3 and within Domain 2 (Figs. 2a; S1; Fig. 10, diagram d). Distributed deformation and early cataclasis may corresponded to strain hardening prior to coseismic weakening, consistent with experimental observations (Biegel et al., 1989; Smith et al., 2015; Pozzi et al., 2019). Pervasive r-twins observed suggest strain hardening and may also correspond to this stage (Fig. 9d-f; Avila et al., 2015; Pozzi et al., 2019), although hematite r-twins and hardening may also occur during post-seismic restrengthening (Ault et al., 2019). Initial weakening is related to fluidization and(or) thermal pressurization with Domain 2, forming the observed Domain 2-Domain 3
undulatory boundary and imparting weak CPO and SPO to comminuted hematite plates (Fig. 10, diagram e). Early weakening leads to progressive strain localization (e.g., Platt et al., 2014), cataclastic flow, and further weakening by nanoparticle lubrication (Fig. 10, diagram e). These collective processes ultimately lead to the formation of microtextural Domain 1.

The final width of the deforming zone in our samples is ~5–20 μm. Although early weakening initiated strain localization, numerical and experimental studies show dynamic weakening processes facilitate additional localization until a stable finite width is reached (Platt et al., 2014; Smith et al., 2015). Asperity flash heating is likely to be an important weakening mechanism early in coseismic slip (e.g., Rice, 2006), but may be further activated and aided by enhanced temperature rise along the slip surface (Fig., 10, red curve). This inference is consistent with theoretical considerations (Rice, 2006) and our observation that development of a localized slip zone likely precedes the development of spatially variable high-temperature microtextures indicative of asperity flash heating (e.g., compare W17-1H versus W17-1G). At asperities, elevated temperatures ≥1100 °C activate GSI deformation mechanisms (e.g., W17-1J and W17-1G; Fig. 10, diagram f; Siemes et al., 2003, 2011). Where even higher temperatures are attained, GSS deformation mechanisms dominate (e.g., W17-1G; Fig. 10, diagram f; Siemes et al., 2003; DePaola et al., 2015; Pozzi et al., 2019). These processes occur only locally on a FM surface, and brittle deformation and cataclastic flow maintain low dynamic strength elsewhere.

After slip has ceased, GSS deformation mechanisms continue resulting in local growth of ultracataclastic particles and those recrystallized by GSI processes (Fig. 10,
Grain growth rate scales with temperature, which decreases perpendicular to the FM surface, ultimately causes polygonal grain growth inversely proportional to distance from the FM (Fig. 4; Fig. 10, diagram f). Because temperature gradients persist into Domain 2, post-seismic static annealing affects smaller grains near the Domain 2-Domain 1 contact, resulting in observed hematite plate tips with polygonal grains (e.g., Fig. 3c; Fig. 10, diagram f). Static annealing is strain hardening, and contributes to restrengthening of the FM interface (Fig. 10; Wright, 1976; Ault et al., 2019; Pozzi et al., 2019).

5.4 Implications for strain localization and fault mirror formation

One of the most striking relationships is the degree of final strain localization indicated by the ~5 to 20 μm width of Domain 1, but the 100s of Myr history that lead to seismogenesis and the width of the deforming zone on these slip surfaces. The microtextural fingerprints of inferred dynamic weakening mechanisms are largely restricted to Domain 1, suggesting an inherent relation between temperature rise, weakening processes, and strain localization in WFZ FMs. Experimental studies with cm- to m-scale displacements suggest strain localization is a necessary precursor to the temperatures required to activate dynamic weakening mechanisms (Ikari, 2015; Smith et al., 2015). The vast majority of WFZ FMs have ~mm- to cm-scale displacements or preserve surface areas consistent with that slip range (Evans et al., 2014; McDermott et al., 2017). Even at the depths (~4-5 km) and thus comparatively higher normal stresses (~100-125 MPa, assuming 25 MPa/km) our samples likely slipped at, temperature rise sufficient to induce weakening in the absence of strain localization is improbable. (U-
Th)/He timing constraints suggest prior brecciation and veining on the scale of ~1 cm to ~200 μm and resultant strength contrasts likely set initial width of the slipping zone, even as early as 1680 Ma, for later seismic slip. This yields more dramatic temperature rise and activation of dynamic weakening mechanisms during seismic slip. WFZ FMs may thus be an example of how prior deformation influences the style of future deformation in natural faults.

Studied WFZ FMs exhibit similarities and differences with previously studied FMs and other thin slip surfaces. We document a suite of dynamic weakening mechanisms previously inferred from a wide range of natural and experimental slip surfaces from a range of lithologies, including nanoparticle lubrication, fluidization (or potentially thermal pressurization), asperity flash heating, and GSS/GSI flow (Han et al., 2011; Smith et al., 2013, 2015; De Paola et al., 2015; Siman-Tov et al., 2015; Pozzi et al., 2018; Calzolari et al., 2020). Despite similar macroscopic appearance, our WFZ FMs are microtexturally heterogenous. Textural differences imply different degrees of activation of these mechanisms. Our new data also present a wider range of microtextures and inferred weakening than prior study of hematite FMs, which focused only on asperity flash heating (Evans et al., 2014; Ault et al., 2015; McDermott et al., 2017). Lastly, we note that the combination of a coupled fluidized layer with a micron-scale principal slip zone is similar to recent experimental work on phyllosilicate gouges (French and Chester, 2018), natural fault surfaces in carbonate (Fondreist et al., 2012), and numerical modeling of serpentinite and carbonate faults (Platt et al., 2014). Given lithological differences between these studies and our samples, our results suggest some weakening processes
and resultant microtextural signatures may be applicable to a wider range of fault materials.

6. Conclusions

We document a multi-stage textural and mechanical evolution for hematite FMs within the WFZ. Field and microscopy observations, image analysis, and EBSD mapping reveal three microtextural domains and their associated deformation mechanisms. These domains include hematite-cemented breccia, hematite cataclasite, and hematite ultracataclasite with local, plastically-deformed grains. We interpret these domains as reflecting hydraulic fracturing, possibly in response to early seismogenesis both pre-dating and concomitant with WFZ activity, followed by fault reactivation, strain localization, and seismic slip. Data reveal multiple dynamic weakening mechanisms facilitating paleoearthquakes that operate at variable degrees on individual FMs and between FMs. Weakening mechanisms include nanoparticle lubrication, fluidization and(or) thermal pressurization, asperity flash heating, and GSI/GSS flow. Our observations combined with (U-Th)/He constraints imply WFZ FMs develop over protracted timescales, with varying impact on the static and dynamic strength evolution of natural fault zones.

References


McDermott, R.G., Ault, A.K., in prep. Hematite (u-th)/he thermochronometry constrains episodic precipitation, fault reactivation, and exhumation over 100 Myr-timescales in the Wasatch fault zone.


CHAPTER 5
THERMOTECTONIC HISTORY OF THE KLUANE RANGES AND EVOLUTION OF THE EASTERN DENALI FAULT ZONE IN SOUTHWESTERN YUKON, CANADA

Abstract

Exhumation and landscape evolution along strike-slip fault systems reflect tectonic processes that accommodate and partition deformation in orogenic settings. We present seventeen new apatite (U-Th)/He (He), zircon He, apatite fission-track (FT), and zircon FT dates from the eastern Denali fault zone (EDFZ) that bounds the Kluane Ranges in Yukon, Canada. The dates elucidate patterns of deformation along the EDFZ. Mean apatite He, apatite FT, zircon He, and zircon FT sample dates range from ~26–4 Ma, ~110–12 Ma, ~94–28 Ma, and ~137–83 Ma, respectively. A new zircon U-Pb date of 113.9 ± 1.7 Ma (2σ) complements existing geochronology and aids in interpretation of low-temperature thermochronometry data patterns. Samples ≤2 km southwest of the EDFZ trace yield the youngest thermochronometry dates. Multi-method thermochronometry, zircon He date-effective U patterns, and thermal history modeling reveal rapid cooling ~95–75 Ma, slow cooling ~75–30 Ma, and renewed rapid cooling ~30 Ma–present. The magnitude of net surface uplift constrained by published paleobotanical data, exhumation, and total surface uplift from ~30 Ma–present are ~1 km, ~2–6 km, and ~1–7 km, respectively. Exhumation is highest closest to the EDFZ trace,

aUtah State University, bU.S. Geological Survey, cUniversity of Arizona
but substantially lower than reported for the central Denali fault zone. We infer exhumation and elevation changes associated with ~95–75 Ma terrane accretion and EDFZ activity, relief degradation from ~75–30 Ma, and ~30 Ma–present exhumation and surface uplift as a response to flat-slab subduction and transpressional deformation. Integrated results reveal new constraints on landscape evolution within the Kluane Ranges directly tied to the EDFZ during the last ~100 Myr.

1. Introduction

Intracontinental strike-slip faults are primary geologic structures along which mass, stress, and strain transfer are modulated in the continental interiors of obliquely convergent orogens (e.g., Fitch, 1972; Molnar and Dayem, 2010; Storti et al., 2003). Strike-slip fault systems can accommodate significant vertical strain manifest by exhumation and the development of near-fault mountain ranges (e.g., Dewey et al., 1998; Fossen and Tikoff, 1998; Spotila et al., 2007). Controls on the spatiotemporal partitioning of strain depend on near-field factors such as fault geometry (e.g., restraining or releasing bends and stepovers) and rheological contrasts, combined with far-field factors including plate boundary coupling and obliquity between convergence direction and fault strike (e.g., Buscher and Spotila, 2007; Calzolari et al., 2016; Ellis, 1996; Molnar and Dayem, 2010; Platt, 1993; Spotila et al., 2007; Sylvester, 1988; Tikoff and Teyssier, 1994; Umhoefer et al., 2007). Patterns of elevation gain (or loss) and exhumation along strike-slip faults reflect interaction and transience of these various factors (e.g., Spotila et al., 2007; Umhoefer et al., 2007). Deciphering the relative contribution of any one factor can be complicated by lack of preservation in the geological or exhumational record. Localized near-fault relief may also reflect geomorphic disequilibrium or older
topography maintained by lateral motion, rather than coincident surface uplift (Duvall and Tucker, 2015). These issues collectively highlight an incomplete understanding of how landscape evolution links to kinematics in strike-slip fault settings and indicates the importance of exhumation and surface uplift records that span the entirety of fault zone evolution.

The ~2000-km long, dextral strike-slip Denali fault zone extends from the Bering Sea through to southwestern Yukon (Fig. 1; St. Amand, 1957). The central segment of the Denali fault zone (CDFZ) bounds the Alaska Ranges and the highest peak in North America, Denali (~6.2 km a.s.l). The Kluane Ranges are the easternmost extent of the St. Elias Mountains and high topography adjacent to the present-day eastern Denali fault zone (EDFZ) in southwestern Yukon, Canada, with an average elevation of ~1.7 km a.s.l. (averaged over ~90 x 20 km; Figs. 1a; 2). GPS velocities, modern seismicity, Quaternary fault patterns, and lower elevations indicate the present-day EDFZ is less tectonically active than the CDFZ (Bemis et al., 2015; Haeussler et al., 2008; Marechal et al., 2015, 2018). However, the largest documented dextral separations along the Denali fault are measured across the EDFZ, suggesting localized deformation on this structure through geologic time (Fig. 1a; Eibacher, 1976; Haeussler et al., 2008; Lowey, 1998). The Denali fault system has potentially been active since the Late Cretaceous and has experienced variable kinematic regimes that influence the geologic record of transpressional deformation (Benowitz et al., 2011, 2014; Cole et al., 1999; Fitzgerald et al., 1995; Miller et al., 2002; Wahrhaftig et al., 1975). Much of the Cenozoic record of exhumation is preserved in the Kluane Ranges and environs, and the EDFZ thus presents an opportunity to investigate exhumation and landscape evolution of an archetypal
intracontinental strike-slip fault zone (Enkelmann et al., 2017; Falkowski and
Enkelmann, 2016).

**Figure 1** Tectonics of the Denali fault system and south-central Alaska-southwestern
Yukon. (a) Digital elevation model (DEM) overlain on hillshade. DEM flooded to 1000
m to enhance topography visualization (e.g., Benowitz et al., 2011). Fault abbreviations
(south-north): EDFZ-eastern Denali fault zone, DRF-Duke River fault, TF-Totshunda
fault, CDFZ-central Denali fault zone, MCT-McCallum thrust, BGF-Bagley fault, SGF-
Susitna Glacier fault, BPF-Broad Pass fault, WDF-western Denali fault, HCF-Hines
Creek fault, NFF-Northern fold/thrust. After (Plafker and Berg, 1994). Available
Quaternary slip rates are for dextral slip in mm/yr, with exception of DRF which gives
horizontal shortening assuming range of fault dips (Haeussler et al., 2017; Marechal et
al., 2018; Matmon et al., 2006). Mountain range abbreviations (south-north): STE-St.
Elias Mountains; WM-Wrangell Mountains, KR-Kluane Ranges, EAR-Eastern Alaska
Range, CAR-Central Alaska Range, WAR-Western Alaska Range. Subducted Yakutat
Microplate extent (black, dashed) from Eberhart-Phillips et al. (2006). 2002 Mw 7.9
rupture extent after Eberhart-Phillips et al. (2003). Yakutat (YA) and North America
(NA) relative plate motion direction and velocity after Elliott et al. (2010). (b) Simplified
terrane and fault network of south-central Alaska-southwestern Yukon after Colpron et
al. (2007). Fault abbreviations (south to north): AT-Aleutian megathrust, FF-Fairweather
fault, CSE-Chugach-St. Elias fault; CMF-Castle Mountain fault, BRF-Border Ranges
fault, TF-Totshunda fault, DRF-Duke River fault, TiF-Tintina fault, NF-Nixon Fork fault, KF-Kaltag fault.

Low-temperature thermochronometry offers a tool to constrain spatiotemporal patterns of exhumation that correspond to the distribution of vertical strain in strike-slip fault systems (e.g., Benowitz et al., 2011; Duvall et al., 2013; Fitzgerald et al., 2014; Niemi et al., 2013; Spotila et al., 2007; Tippett and Kamp, 1993). The apatite He, apatite FT, zircon He, and zircon FT thermochronometers record cooling and exhumation through the ~30–120 °C, ~70–120 °C, ~25–220 °C, and 180–250 °C temperature ranges, respectively (Brandon et al., 1998; Flowers et al., 2009; Guenthner et al., 2013; Ketcham et al., 2007). These thermochronometric systems are thus sensitive to thermal perturbations in the upper ~2–10 km of the Earth’s crust (assuming a typical 25–30 °C/km geothermal gradient) and provide constraints on the timing, amount, and rates of exhumation due to tectonic and(or) surface processes (e.g., Ehlers and Farley, 2003; Reiners and Brandon, 2006).

Previous thermochronometric investigations of the EDFZ focused primarily on interior portions of the Kluane Ranges, to the west of our study area, provide evidence for a complex and multiphase cooling history (Enkelmann et al., 2017; Falkowski and Enkelmann, 2016; Spotila and Berger, 2010). We present new data from EDFZ fault-parallel and fault-perpendicular sampling transects in the adjacent Kluane Ranges that include 25 apatite and 28 zircon single-grain He analyses from seven samples, along with five apatite (~40 grains/sample) and three zircon (~20 grains/sample) single-sample FT dates (Fig. 2). Combined multi-method thermochronometry, single grain date-grain chemistry patterns, and thermal history inversions document a ~100 Myr thermal history
of the Kluane Ranges. We provide quantitative constraints on net surface uplift, exhumation, and total surface uplift for the Kluane Ranges from the past 30 Myr. By integrating these constraints with geological information, we define three thermotectonic phases that correspond to distinct stages in EDFZ evolution and their associated surface uplift and exhumation responses. Phases include Late Cretaceous rapid exhumation and likely increased relief during terrane accretion and possible deformation along the EDFZ; Late Cretaceous to Oligocene slow exhumation, relief reduction, and a shift in the locus of regional contraction and(or) dominantly strike-slip motion along the EDFZ; and Oligocene to present rapid exhumation in response to renewed surface uplift in a transpressional tectonic regime. This ~100 Myr history of surface uplift, exhumation, and landscape evolution within the Kluane Ranges is directly tied to the evolution of the EDFZ.

2. Geologic Setting

2.1 Tectonics of southeastern Alaska and southwestern Yukon

The larger tectonic framework of south-central Alaska and southwestern Yukon is characterized by accreted terranes transected by continental-scale strike-slip faults (Fig. 1b; Colpron et al., 2007; Coney et al., 1980). These terranes can be broadly grouped into Proterozoic-Triassic paraautochthonous assemblages associated with ancestral North America; the Proterozoic-Permian pericratonic Intermontane terranes; Paleozoic-Mesozoic allochthonous Insular terranes; and Mesozoic-Paleogene allochthonous Late Accreted terranes (Colpron et al., 2007).
In the general vicinity of our study area, the Intermontane Yukon-Tanana terrane comprises rocks of the former Laurentian continental margin and Paleozoic-Mesozoic allochthonous arc assemblages (Fig. 1b; Dusel-Bacon et al., 2006; Nelson et al., 2006). Paleozoic oceanic arc and back-arc complexes of the Insular terranes are further subdivided into Wrangellia, the Alexander terrane, and the Peninsular terrane (collectively, the Wrangellia Composite terrane of others; Plafker et al., 1989)).
Wrangellia likely formed on Devonian Alexander basement, with accretion of the Peninsular terrane to the Wrangellia-Alexander outboard margin by the Triassic (Israel et al., 2014; Plafker et al., 1989). Accretion of the combined Insular terranes with the Intermontane terranes began by the mid-Jurassic, as recorded by back-arc basin overlap assemblages and accreted suture zone sediment, with final accretion likely occurring in the Late Cretaceous (Dusel-Bacon et al., 1993; Eisbacher, 1976; McClelland et al., 1992; Nokleberg et al., 1985; Ridgway et al., 2002). Accretionary prism-arc complexes of the Chugach terrane record Mesozoic-early Cenozoic subduction and terrane accretion along the outboard margin of the Insular terranes (Plafker et al., 1989). Flat-slab subduction of the Yakutat microplate, a ~15–35 km-thick oceanic plateau, initiated by ~30–35 Ma (Eberhart-Phillips et al., 2006; Ferris et al., 2003; Finzel et al., 2011; Worthington et al., 2012). Subduction of progressively thicker crust resulted in a transition from subduction to collision of the Yakutat microplate at ~15 Ma (Falkowski et al., 2014; Worthington et al., 2012).

Plate reconstructions and patterns of arc magmatism show multiple modes and configurations of subduction and/or terrane accretion though time. Obliquely convergent, east-dipping, subduction of the Kula plate along the outboard edge of the Insular terranes was likely responsible for Jurassic-Late Cretaceous accretion within a sinistral, followed by dextral, transpressive regime (Doubrovine and Tarduno, 2008; McClelland et al., 1992; Plafker et al., 1989). Magmatic arc belts associated with this subduction are Jurassic to Cretaceous (Chitina arc) and Paleogene (Kluane Arc) and decrease in age from west to east (Dodds and Campbell, 1988; Plafker et al., 1989). Beginning at ~70 Ma, the angle of convergence between North America and the Kula
Plate decreased, while relative velocity increased, until consumption of the Kula Plate at ~40 Ma (Doubrovine and Tarduno, 2008). The demise of a potential additional plate, the Resurrection Plate, has been postulated to have occurred at ~50 Ma (e.g., Haeussler et al., 2003). However, the existence and possible influence of hypothesized Kula-Resurrection ridge subduction is controversial (e.g., Garver and Davidson, 2015). Post-40 Ma relative plate motions between North America and the Pacific/Yakutat plates remained approximately constant and directed northwest, followed by counterclockwise rotation to north-northwest from ~12 Ma-present (Doubrovine and Tarduno, 2008; DeMets and Merkouriev, 2016). Magma sourced from the trailing edge of the subducting Yakutat microplate fed the Oligocene-present Wrangell Arc and similarly record northwestward convergence (e.g., Brueseke et al., 2019; Richter et al., 1990). Flat-slab subduction of the Yakutat microplate is likely responsible for present-day deformation 100s of km inboard from the plate boundary (Benowitz et al., 2011, 2014; Freymueller et al., 2008; Haeussler et al., 2008; Mazzotti and Hyndman, 2002).

2.2 The Denali fault system

The Denali fault system has a complex structural expression and comprises eastern (EDFZ), central (CDFZ), and western (WDFZ) segments, along with several associated thrust splays (Fig. 1a; Haeussler et al., 2008; Plafker and Berg, 1994). The Denali fault occupies the Mesozoic suture zone between the Intermontane and Insular terranes along much of its length. It accommodates ~300–400 km of dextral separation along the EDFZ/CDFZ and ~130 km of dextral separation along the WDFZ (Fig. 1b; Eisbacher, 1976; Lowey, 1998; Miller et al., 2002; Ridgway et al., 2002). Deformed strata along the
EDFZ, CDFZ, and WDFZ suggest dextral activity by ~55 Ma, ~70–60 Ma and ~85–70 Ma, respectively (Cole et al., 1999; Eisbacher, 1976; Miller et al., 2002). Kinematic indicators within mylonites hosted in the southern Intermontane Terranes imply dextral slip between ~85–55 Ma along the former continental margin (Andronicos et al., 1999; Gehrels et al., 2000). More recently, Riccio et al. (2014) suggest most dextral separation occurred post-57 Ma on the basis of offset correlative plutons.

The Denali fault system accommodates ~20–30% of the Quaternary through present-day relative motion between North America and the Yakutat microplate (Fig. 1a; Elliott et al., 2010; Matmon et al., 2006). Present-day activity on the Denali fault is recorded by GPS data and abundant seismicity, including a 2002 Mw 7.9 earthquake that ruptured the CDFZ, Susitna Glacier, and Totshunda faults (e.g., Eberhart-Phillips et al., 2003; Elliott et al., 2010; Fisher et al., 2004; Freymueller et al., 2008;). Quaternary dextral slip rates peak at ~13–14 mm/yr along the CDFZ and decrease both westward and eastward (Haeussler et al., 2017; Marechal et al., 2018; Matmon et al., 2006). The westward decrease in slip rate may reflect partitioning of the slip budget along thrust faults oriented perpendicular to Yakutat microplate convergence, although the exact mechanisms of strain accommodation are debated (e.g., Haeussler et al., 2017). Slip rates along the northern section of the EDFZ and Totshunda fault sum to approximately equal the CDFZ slip rate, suggesting these two faults act in conjunction to accommodate plate boundary strain (Matmon et al., 2006). South of the EDFZ-Totshunda intersection and near our study area, Quaternary dextral slip rates further decrease to ~<1–2 mm/yr on the EDFZ, while the Totshunda Fault accommodates ~15 mm/yr of dextral slip (Fig. 1a; Haeussler et al., 2017; Marechal et al., 2018).
2.3 Geology of the Kluane Ranges

Kluane Ranges bedrock consists primarily of Wrangellian Mississippian to Pennsylvanian volcanic and volcaniclastic rocks of the Skolai Group, unconformably overlain by Triassic-Jurassic flood basalts, volcaniclastics, and marine strata (Dodds and Campbell, 1992; Israel et al., 2005; Read and Monger, 1976; Smith and MacKevett, 1970). Triassic mafic and ultramafic, Late Cretaceous mafic and intermediate, and Oligocene felsic plutonic rocks intrude Paleozoic-Mesozoic sedimentary rocks (Dodds and Campbell, 1988). Latest Eocene to Oligocene terrestrial clastic rocks of the Amphitheatre Formation are found in basins throughout the ranges and are associated with remnants of a pre-depositional, regional, low-relief erosion surface (Eisbacher and Hopkins, 1977). This unit is interpreted to record synorogenic sedimentation in strike-slip basins formed by transcurrent motion along the Denali and other subsidiary strike-slip faults (Eisbacher and Hopkins, 1977; Ridgway et al., 1992). Amphitheatre Formation rocks are conformably overlain by Oligocene to late Miocene mafic and intermediate volcanic, shallow intrusive, and associated volcaniclastic rocks of the Wrangell Volcanic complex (Cole and Ridgway, 1993; Dodds and Campbell, 1988). The Kluane Ranges were sculpted by Pleistocene glaciation and exhibit a variety of glacial landforms and associated sedimentary deposits (Denton, 1974).

Northwest-southeast trending folding and faulting contribute to the regional structural fabric of the Kluane Ranges (Dodds and Campbell, 1992; Israel et al., 2005; Read and Monger, 1976). The principal range-bounding structure of the Kluane Ranges is the EDFZ, which is found adjacent to the mountain front and(or) cuts broad, lowland,
glaciated valleys (Fig. 2). The Duke River fault (DRF), a predominantly reverse-slip fault with minor components of strike-slip, lies ~10 km to the southwest of the EDFZ and approximately follows the contact between Wrangellian rocks and Cambrian-Triassic siliciclastic and volcanic rocks of the Alexander terrane (Figs. 1b; 2; Cobbett et al., 2016). The DRF was active by at least ~100 Ma and accommodates shortening at a rate of ~3.5–6.5 mm/yr over the Quaternary (Fig. 1a; Cobbett et al., 2016; Marechal et al., 2018). The DRF and EDFZ converge in southwest Yukon, with the EDFZ appearing to cross-cut the DRF ~30 km southwest of our study area (Fig. 1a; Cobbett et al., 2016). GPS measurements show ~5–10 mm/yr of distributed horizontal shortening in the adjacent Kluane Ranges in response to ongoing Yakutat microplate collision (Marechal et al., 2015). Focal mechanisms of modern-day earthquakes throughout the larger Kluane Ranges area indicate an active transpressional strain regime (He et al., 2018; Vallage et al., 2014).

2.4 Previous thermochronometry studies of the St. Elias Mountains and Denali fault

Bedrock and detrital thermochronometry in the larger St. Elias Mountains of central-eastern Alaska and southwestern Yukon reveal episodes of terrane-scale rapid cooling in the Jurassic-Cretaceous, Paleocene-Eocene, Oligocene-Miocene, and Pliocene (Enkelmann et al., 2017; Falkowski and Enkelmann, 2016; O'Sullivan and Currie, 1996). These distinct cooling phases have been interpreted as recording the erosional and thermal effects of subduction along the outboard Insular terrane margin and arc magmatism, magmatism from spreading ridge subduction and plate reorganization, Yakutat microplate subduction and collision, and glaciation, respectively (Enkelmann et
Preservation of this protracted cooling record is reflective of limited regional denudation in the Insular terranes during the Cenozoic (Falkowski and Enkelmann, 2016). The southwest margin of the Insular and adjacent Chugach terranes are dominated by Pliocene exhumation resulting from inferred rock and(or) surface uplift along the Fairweather, Malaspina, and Border Ranges fault systems combined with efficient glacial erosion (Fig. 1b; Enkelmann et al., 2009, 2017; Falkowski et al., 2014; Falkowski and Enkelmann, 2016; Spotila and Berger, 2010). Regionally distributed bedrock apatite He data suggest this area has experienced approximately 5x more rock uplift than the Kluane Ranges region (Spotila and Berger, 2010). Late Cretaceous and Miocene cooling in the Kluane Ranges are linked to terrane accretion (Enkelmann et al., 2017; Falkowski and Enkelmann, 2016). Eocene cooling recorded in the Yukon-Tanana terrane and overlap assemblages to northeast of the EDFZ likely reflects rapid post-emplacement magmatic cooling and possible subsequent extensional unroofing (Enkelmann et al., 2017).

The CDFZ has been the focus of numerous thermochronometry studies. In the CDFZ-bounded eastern and central Alaska Range, biotite and k-feldspar $^{40}$Ar/$^{39}$Ar thermochronometry record spatially-variable, rapid cooling and ~6 to ≥11 km of exhumation since ~25 Ma as the far-field response to Yakutat microplate subduction (Fig. 1a; Benowitz et al., 2011, 2014; Lease et al., 2016). In the central Alaska Range, rapid cooling at ~6 Ma is documented by apatite FT data from the Denali massif and potentially indicates a change in Pacific Plate convergence direction (Fitzgerald et al., 1995). Other workers suggest that rapid Neogene cooling of rocks along the CDFZ and other structures is the result of advection of crust through the large restraining bend that
characterizes the CDFZ and attendant large-scale contractional deformation and vertical rock uplift (Fig. 1, inset; Fitzgerald et al., 2014; Lease et al., 2016; Riccio et al., 2014). Bedrock apatite He and detrital apatite FT dates from high-relief, glaciated catchments adjacent to the CDFZ record accelerated Pleistocene cooling and exhumation as the result of glacial erosion (Benowitz et al., 2011; Lease et al., 2016, 2018).

3. Low-Temperature Thermochronometry and Methods

3.1 Apatite and zircon (U-Th)/He systematics

Apatite and zircon He record cooling from temperatures of ~30–120 °C and 25–220 °C, and can track exhumation through the upper ~10 km of the crust, depending on regional geothermal gradient (e.g., Flowers et al., 2009; Guenthner et al., 2013). Closure temperature (T_c) in the He system partially depends on diffusion domain lengthscale (grain size for apatite and zircon) and cooling rate (Dodson, 1973; Farley, 2000; Reiners et al., 2004). A primary control on variations in He diffusion kinetics, and thus effective T_c and dates in these systems, is the accumulation of damage to the crystal lattice by parent isotope actinide decay over the sample’s thermal history (Flowers et al., 2009; Gautheron et al., 2009; Guenthner et al., 2013; Shuster et al., 2006; Shuster and Farley, 2009). Radiation damage anneals with increasing temperature resulting in evolving He diffusivity and effective T_c in an apatite or zircon crystal over a given thermal history. In the apatite He system, increasing amounts of radiation damage from α-recoil create energy traps that impede He diffusion, resulting in higher effective T_c. A similar phenomenon is observed in zircon at low damage. However, because zircon incorporates up to several orders of magnitude more U and Th and requires higher temperatures to
anneal this damage, accumulated damage reaches a percolation threshold resulting in higher diffusivity and lower effective $T_c$ (Guenthner et al., 2013; Ketcham et al., 2013) in highly-damaged zircon grains.

For apatite and zircon grains within a single sample that experience the same thermal history, effective uranium concentration ($eU; U+0.235\times Th$) corresponds to relative radiation damage accumulation. Thermal histories characterized by slow monotonic cooling, residence in the partial retention zone, or reheating yield date-$eU$ trends. Apatite He dates may exhibit a positive relationship with $eU$ and zircon He dates may exhibit positive and(or) negative date-$eU$ trends, depending on damage concentration and thermal history (Flowers et al., 2009; Ketcham et al., 2013). Radiation damage effects expand the effective $T_c$ range of apatite and zircon He systems and can be exploited to better constrain thermal histories (Ault et al., 2009; Flowers, 2009; Guenthner et al., 2017; Orme et al., 2016). Available radiation damage-diffusivity models for apatite and zircon assume damage annealing occurs at the temperatures and rates that fission-tracks anneal in each system, although recent work suggests that $\alpha$-damage annealing may require higher temperatures (Ault et al., 2018; Flowers et al., 2009; Fox and Shuster, 2014; Gautheron et al., 2009; Gerin et al., 2017; Ginster et al., 2019; Guenthner et al., 2013; Willett et al., 2017). In addition to radiation damage effects, possible sources of overdispersion in He dates include strong $eU$ zonation and associated errors in correcting for grain edge $\alpha$ ejection, parentless He implanted by “bad neighbors” or hosted in unrecognized U-Th-rich inclusions, and crystal defects that create fast diffusion pathways (e.g., Farley, 2002; Flowers and Kelley, 2011).
3.2 Apatite and zircon fission-track thermochronometry

FT thermochronometry exploits the natural production of etchable tracks from spontaneous fission of $^{238}\text{U}$ and the time-temperature-dependent annealing of fission-tracks. The $T_c$ for the apatite and zircon FT systems are $\sim70–120^\circ\text{C}$ and $\sim180–250^\circ\text{C}$, respectively, although high Cl apatite and damage-free zircon (not analyzed in this study) may have $T_c$ in excess of $\sim140^\circ\text{C}$ and $300^\circ\text{C}$ (Bernet, 2009; Brandon et al., 1998; Ketcham et al., 2007; Rahn et al., 2004). Annealing of FTs in apatite and zircon is primarily a function of thermal history and grain chemistry (Brandon et al., 1998; Ketcham et al., 1999, 2007; Tagami et al., 1990). FTs anneal yielding shorter track lengths and measurement of the distribution of track lengths in a given sample provides additional constraints on its thermal history (Gleadow et al., 1986; Green et al., 1986). Additional constraints on track annealing can be evaluated using the kinetic parameter $D_{\text{par}}$, or c-axis-parallel etched track width, which although its own stand-alone parameter, is correlated with grain chemistry (Ketcham et al., 2007). Primary sources of systematic analytical uncertainty in FT thermochronometry are analyst bias in track counting and length measurements, and under-/oversaturation of FTs due to low/high U concentrations, respectively (Barbarand et al., 2003; Donelick et al., 2005; Laslett et al., 1982).

3.3 Sampling approach and analytical methods

We collected a total of 15 bedrock samples from an EDFZ-perpendicular elevation transect and an EDFZ-parallel transect. Only seven samples from metabasalt, diorite, monzonite, metagraywacke, and granodiorite yielded material suitable for
thermochronometric analysis (Table 1; Fig. 2). Inclusion-free, whole apatite crystals were selected for He analysis. We followed the zircon selection protocol of Ault et al. (2018) and purposefully selected whole grains exhibiting a range of visual metamictization, a qualitative proxy for intrasample accumulated radiation damage and eU because all crystals within a sample experienced the same time-temperature history. This approach allows us to maximize the likelihood of capturing a broad range of eU in our analyzed crystals. U, Th, Sm, and He concentrations were measured at the Arizona Noble Gas Laboratory at the University of Arizona (UA). Subsets of apatite and zircon crystals were selected for FT analysis from some samples. Apatite and zircon FT track count measurements were conducted at the UA Fission-Track Laboratory. Three samples yielded sufficient apatite track density needed to generate track length distribution data. Dpar was measured on all apatite samples. We also acquired a zircon U-Pb geochronology date by laser ablation inductively coupled plasma mass spectrometry (LA-ICPMS) at the University of Utah to constrain the crystallization age of one sample (61615-5A). This sample’s affinity was previously enigmatic, and this datum provides an upper bound on the time period over which zircon radiation damage could accumulate. Cathodoluminescence (CL) imaging, acquired at the University of Utah, provides textural context for U-Pb dates. Detailed analytical methods for He, FT, U-Pb and CL analyses are provided in the Supporting Information.

The Supporting Information for this chapter is available at https://doi.org/10.1029/2019TC005545
4. Results

Table 1 and Figure 3 summarize new thermochronometry dates and data patterns. To highlight intrasample data patterns, color-coding in Figure 3 is sample-specific and symbols correspond to thermochronometer type as in Figure 2. For apatite and zircon He samples where single-grain dates are within 20% standard deviation of the mean, we report the sample mean and 1σ standard deviation, and for samples with single-grain dates within >20% standard deviation of the mean, we report the range of individual dates with 2σ analytical error (c.f. Flowers and Kelley, 2011). We report FT dates as the central date ± 1σ standard deviation (Galbraith, 1990). Detailed results for U, Th, and 4He measurements in apatite and zircon are provided in Tables S1 and S2 of the Supporting Information. Fission-track count data for apatite and zircon, as well as track length data for apatite, are summarized in Table S3 and raw data are in Supporting Information Datasets 1, 2 and 3. Zircon U-Pb results and CL images are reported in Tables S4 and Figures S1 and S2. To simplify discussion of our dataset, we subdivide our data into “far-from-fault” (FFF; samples 61215-14A, 61215-10A, 61415-1A, 61415-2A) and “near-fault” (NF; samples 61115-2A, 61615-5A, 15-SI-012-1) sample groups. The rationale for sample groupings is partially based on fault-perpendicular distance from the EDFZ trace and elevation above mean sea level (<2 km and 820-1200 m a.s.l. for NF, >2 km and 1696-2215 m a.s.l. for FFF), but each group also exhibits distinct intrasample thermochronometric data patterns discussed below.
4.1 Apatite and zircon He data

We report 25 single-grain apatite He dates from four samples (Fig. 3a, b; 61215-14A; 61415-1A; 61415-2A; 15-SI-012-1). FFF samples 61215-14A and 61415-1A yield mean apatite He dates of 26.2 ± 3.6 Ma (n=5; 14% std. dev.) and 20.1 ± 2.8 Ma (n=7; 14% std. dev.), respectively. Individual aliquot apatite He dates from FFF sample 61415-2A range from 11.6 ± 0.6 Ma to 24.7 ± 0.8 Ma (n=7; 32% std. dev.). NF sample 15-SI-012-1 yields a mean date of 4.3 ± 0.3 Ma (n=6; 7% std. dev.). Individual apatite eU varies from ~10 to 27 ppm across the entire dataset. In any given sample, apatite He dates are broadly uniform as a function of eU, although the eU range is limited (Fig. 3a). We observe no relationship between individual apatite He date and grain size (effective spherical radius, R<sub>es</sub>; Fig. 3c).

We acquired 28 individual zircon He dates from five samples (Fig. 2a; 61115-2A; 61215-14A, 61415-1A, 61415-2A, and 61615-5A). Mean dates from FFF samples 61215-14A (n=6; 19% std. dev.), 61415-2A (n=6; 18% std. dev), and 61415-1A (n=6; 14% std. dev.) are 93.7 ± 17.4 Ma, 84.5 ± 14.8 Ma, and 86.8 ± 12.1 Ma, respectively. FFF samples collectively define a gentle, negative zircon He date-eU trend over ~100-2100 ppm eU values that plateaus at ~80–65 Ma (Fig. 3a). NF sample 61115-2A (n=4; 52% std. dev.) yields a zircon He date range of 14.0 ± 0.4 Ma to 40.7 ± 1.2 Ma. NF sample 61615-5A (n=6; 18% std. dev.) has a mean zircon He date of 60.3 ± 10.8 Ma (Fig. 3a). NF samples yield positive zircon He date-eU relationships from ~190–770 ppm eU and ~14–72 Ma (Fig. 3a). Individual zircon He dates from NF samples are also positively correlated with R<sub>es</sub>, but this trend is not evident in FFF samples (Fig. 3c).
4.2 Apatite and zircon FT data

We report five apatite FT dates, track length distributions on three of these samples, and three zircon FT dates (Fig. 3a, b; Table S3). All apatite and zircon FT results pass the $\chi^2$ test and show low dispersion (<0.01–0.1 %), indicating a single date population for each sample (Table S3). Central apatite FT dates ($\pm 1\sigma$ std. dev.) from FFF samples (61215-10A, 61215-14A, 61415-1A, and 61415-2A) range from 109.5 ± 7.8 Ma to 64.0 ± 15.7 Ma. FFF samples 61215-14A, 61415-1A, and 61415-2A have mean track lengths of 14.22–14.35 µm, and 1σ standard deviations between 0.96 and 1.08 µm. The apatite FT date from NF sample 15-S1-012-1 is 12.2 ± 2.4 Ma. (Table S3). Zircon FT dates from FFF sample 61215-14A and NF samples 61115-2A and 61615-5A are 95.8 ± 4.6 Ma, 136.5 ± 8.5 Ma, and 82.7 ± 4.2 Ma, respectively.

4.3 Zircon cathodoluminescence and U-Pb data

We acquired a zircon U-Pb date from NF sample 61615-5A to determine its crystallization age and thus an appropriate initial condition for thermal modeling discussed below. CL images show zircons exhibit oscillatory zoning typical of magmatic zircon (Fig. S1; Corfu et al., 2003). Concordant zircon $^{206}$Pb-$^{238}$U dates from individual ablation spots range from 125.6 ± 2.5 Ma to 104.4 ± 2.1 Ma (2σ analytical uncertainty; Table S4). Cores consistently yield older dates than rims (Fig. S1). The weighted mean $^{206}$Pb-$^{238}$U date is 112.4 ± 1.4 Ma (2σ; Fig. S2a). A Tera-Wasserburg plot shows that analyses are broadly concordant in $^{207}$Pb/$^{206}$Pb and $^{238}$U/$^{206}$Pb space with little contamination from an additional $^{207}$Pb/$^{206}$Pb source. Correction for common Pb following the two-stage Pb evolution model of Stacey and Kramers (1975) yields an
intercept age of 113.9 ± 1.7 Ma (2σ; Fig. S2b). These data indicate that NF sample 61615-5A is part of the mid-Cretaceous (~117–106 Ma) Kluane Ranges magmatic suite (Dodds and Campbell, 1988).

**Figure 3** Apatite, zircon He and fission-track data. (a) Single-grain apatite He dates (± 2σ analytical uncertainty) and apatite FT central dates (± 2σ standard deviation) plotted as a function of eU for He system and average U concentration for FT system, respectively. (b) Single-grain zircon He and zircon FT dates as a function of eU and U, respectively. (c) Single-grain apatite and zircon He dates as a function of equivalent spherical radius (R_{es}). Color scheme refers to individual samples and thermochronometer type is coded by symbology as in Fig. 2 to highlight intrasample data patterns.
### Table 1: New bedrock (U-Th)/He (He) and fission-track (FT) cooling dates

<table>
<thead>
<tr>
<th>Sample</th>
<th>Lithology</th>
<th>Lat.</th>
<th>Long.</th>
<th>Elev. (m a.s.l.)</th>
<th>Distance from EDFZ (km)</th>
<th>Apatite He&lt;sup&gt;b&lt;/sup&gt;</th>
<th>Apatite FT&lt;sup&gt;d&lt;/sup&gt;</th>
<th>Zircon He</th>
<th>Zircon FT</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Far-from-fault (FFF) samples</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>61215-10A</td>
<td>metabasalt</td>
<td>61.0792</td>
<td>-138.606</td>
<td>1981</td>
<td>3.5</td>
<td>-</td>
<td>64.0 ± 15.7</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>61215-14A</td>
<td>metabasalt</td>
<td>61.0790</td>
<td>-138.585</td>
<td>1696</td>
<td>2.8</td>
<td>26.2 ± 3.6</td>
<td>109.5 ± 7.8</td>
<td>93.7 ± 17.4</td>
<td>95.8 ± 4.6</td>
</tr>
<tr>
<td>61415-1A</td>
<td>diorite</td>
<td>60.9434</td>
<td>-138.378</td>
<td>2215</td>
<td>4.7</td>
<td>20.1 ± 2.8</td>
<td>99.7 ± 7.4</td>
<td>86.8 ± 12.1</td>
<td>-</td>
</tr>
<tr>
<td>61415-2A</td>
<td>diorite</td>
<td>60.9462</td>
<td>-138.385</td>
<td>1947</td>
<td>5.0</td>
<td>17.9 ± 5.7</td>
<td>87.1 ± 6.2</td>
<td>84.5 ± 14.8</td>
<td>-</td>
</tr>
<tr>
<td><strong>Near-fault (NF) samples</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>61615-5A&lt;sup&gt;e&lt;/sup&gt;</td>
<td>monzonite</td>
<td>61.0569</td>
<td>-138.5090</td>
<td>820</td>
<td>1.4</td>
<td>-</td>
<td>60.3 ± 10.8</td>
<td>82.7 ± 4.2</td>
<td></td>
</tr>
<tr>
<td>61115-2A</td>
<td>graywacke</td>
<td>61.3655</td>
<td>-139.1590</td>
<td>941</td>
<td>0.5</td>
<td>-</td>
<td>27.9 ± 14.5</td>
<td>136.5 ± 8.5</td>
<td></td>
</tr>
<tr>
<td>15-SI-012-1</td>
<td>granodiorite</td>
<td>60.8963</td>
<td>-138.1000</td>
<td>1200</td>
<td>0.9</td>
<td>4.3 ± 0.5</td>
<td>12.2 ± 2.4</td>
<td>-</td>
<td>-</td>
</tr>
</tbody>
</table>

<sup>a</sup>measured perpendicular to local strike of EDFZ main trace

<sup>b</sup>He dates reported as mean date ± 1σ std. dev.

<sup>c</sup>n=number of individual dates (He) or crystals counted (FT)

<sup>d</sup>FT dates reported as central date ± 1σ std. dev.

<sup>e</sup>zircon $^{206}$Pb-$^{238}$U date of 113.9 ± 1.7 Ma (2σ)
4.4 Thermochronometry data spatial patterns

Figure 4 integrates new and previously published thermochronometry data from the Kluane Ranges to illustrate spatial data patterns. We add Enkelmann et al. (2017) and Spotila and Berger (2010) samples KLB88 and K6, respectively, to our FFF group and Enkelmann et al. (2017) sample KLB100 to our NF group (Fig. 4). The youngest dates for any given thermochronometric system are in the NF group at fault-perpendicular distances of $\leq 2$ km southwest of the EDFZ (Figs. 4a). Apatite FT, zircon He, and zircon FT dates from FFF samples all approximately overlap within 1σ uncertainty, although some mid- to high-eU individual aliquot zircon He dates are younger than corresponding apatite FT dates (Table 1; Figs. 3; 4a, b). Apatite He dates for FFF samples increase with elevation, with an apparent break in slope at $\sim 30$ Ma (Fig. 4b). FFF apatite FT, zircon He, and zircon FT dates do not vary as a function of elevation (Fig. 4b). Zircon He date-eU/R$_{es}$ trends differ between NF and FFF groups, with FFF samples collectively defining a mild inverse pattern of younger dates with increasing eU and NF samples exhibiting positive date-eU and date-R$_{es}$ trends (Fig. 3c). Zircon FT dates within the NF group are at similar fault-perpendicular distances, but vary substantially parallel to the EDFZ (e.g., compare the $\sim 27$ Ma date of KLB100 to a date of $\sim 136$ Ma 40 km to the northwest and $\sim 83$ Ma $\sim 2$ km to the southeast; Fig. 4a). For comparison, previously published dates northeast of the EDFZ, although not part of the NF and FFF groups defined here, are relatively uniform for any given thermochronometric system (apatite He, apatite FT, zircon He, and zircon FT) and weakly correlated with elevation and perpendicular distance from the fault trace (Fig. 4a; Enkelmann et al., 2017; Falkowski and Enkelmann, 2016).
5. Thermal History Simulations

5.1 Modeling approach

Thermochronometry data spatial patterns and qualitative comparison of zircon He date-eU trends between FFF and NF samples indicate different thermal histories for these two sample groups. To explore and quantify thermal history differences for FFF and NF sample groups, we perform time-temperature (tT) path inversions of new and previously published thermochronometric data using the Bayesian Transdimensional Markov Chain Monte Carlo inversion package QTQt (version 5.6.1d; Gallagher, 2012) and the most recently available diffusion/annealing kinetics for the apatite He, apatite FT, and zircon He systems (Flowers et al., 2009; Guenthner et al., 2013; Ketcham et al., 2007).
Model inputs include apatite and zircon He data adapted for modeling by binning raw (uncorrected) dates by eU and Res (Table S5; Ault et al., 2009; Flowers et al., 2015). We apply the binning approach to account for inherent uncertainties in radiation damage models and 2nd order effects on measured He dates (e.g., parent zonation; Ault and Flowers, 2012; Guenthner et al., 2013). Apatite He data were separated into three bins at 2 ppm eU increments with grain Res averaged for each bin. Zircon He data were binned at 200 ppm eU increments from 0-1000 ppm, and one >1000 ppm bin, with eU bins being further split by Res for samples 61615-5A and 61115-2A (Table S5). We assign ±15% date uncertainty to each bin, or the standard deviation of dates within each bin, whichever is greater (e.g., Johnson et al., 2017). Additional data inputs include single-grain apatite FT dates and Dpar, and apatite FT length distributions. Zircon FT data are incorporated as tT constraints at 220 ± 30 °C, commensurate with the reported Tc of this system in natural radiation-damaged zircon (Bernet, 2009; Brandon et al., 1998). We set zero-date, zero-radiation damage initial conditions for each simulation at temperatures of 25 ± 25 °C above the zircon FT Tc (apatite FT Tc for sample 15-SI-012-1) and 5 Myr prior to the respective date (Figs. 5, 6). For inversions of plutonic samples 61615-6A, 61415-1A, 61415-2A, and KLB88, the age of the initial condition (110 ± 10 Ma) corresponds to our new 113.9 ± 1.7 Ma U-Pb crystallization age (61615-5A) and previously published biotite and hornblende K-Ar data for the Kluane Ranges Suite (Dodds and Campbell, 1988). All inversions were run with 200,000 iterations and 50,000 “burn-in” iterations. Details on data inputs, model settings used, and how we assessed solution convergence are provided in the Supporting Information and Table S6.
We present eight inversions from FFF (61215-14A, 61415-1A, 61415-2A, KLB88) and NF (61115-2A, 61615-5A, 15-SI-012-1, KLB100) samples that have more than one type of thermochronometer date for each sample (Figs. 5; 6). QTQt additionally allows multi-sample inversions where the elevation difference between samples serves as an additional constraint on viable solutions. We thus present an inversion that includes all FFF samples with a span in elevation and at similar EDFZ-perpendicular distances (61215-10A, 61215-14A, 61415-1A, 61415-2A, KLB88; Fig. 7). We include all new apatite He and zircon He data in these simulations because, with the exception of 61415-2A and KLB100, these samples yield either <20% standard deviation between individual grain dates, or apatite/zircon He date-eU trends for samples with >20% standard deviation (Flowers and Kelley, 2011). This indicates that intrasample date variability can be characterized by apatite and zircon radiation damage-diffusivity models. We include 61415-2A and KLB100 in our simulations for completeness but acknowledge that the elevated percent standard deviation for these samples suggests there are factors controlling data dispersion beyond radiation damage accumulation such as those detailed in Section 3.1. We assume eU is uniformly distributed in all apatite and zircon grains, although some analyzed zircon crystals may be zoned (e.g., Fig. S1). This assumption is unlikely to impact modeled apatite He dates but may influence modeled zircon He dates (Ault and Flowers, 2012; Guenthner et al., 2013). Solutions are presented as 1 °C x 1 Ma colormap grids, forming a conditional probability distribution function of tT space, overlain by 95% credible intervals (i.e., the Bayesian form of confidence intervals), and a weighted mean path.
Figure 5 Thermal history model results for far-from-fault (FFF) samples. Left column shows QTQt single-sample thermal history inversion results for samples (a) 61415-1A, (b) 61415-2A, (c) 61215-14A, (d) KLB88 (Enkelmann et al., 2017). Thick black line and envelopes show expected thermal history and 95% credible intervals. Color maps show conditional probability for thermal history simulations (see text). White lines are max-mode tT paths that pass through peak of temperature distribution at each time slice. Red, yellow, and black boxes denote inverse prior, model starting point, and zircon FT constraints, respectively. Insets show slopes corresponding to different cooling rates in °C/Myr. Right column shows corresponding predicted versus observed data for expected time-temperature path. Observed apatite FT length distributions with c-axis projection shown by blue boxes. Predicted length distribution and uncertainty envelopes shown in red and gray, respectively. Insets show observed versus predicted (uncorrected for α-ejection) dates and 1:1 line for reference. Black and gray error bars are observed and model fit uncertainty (respectively).
Figure 6 Thermal history model results for near-fault (NF) samples. QTQt single-sample thermal history inversions (right) and predicted vs. observed date plots (left) for NF samples (a) 61115-2A, (b) 61615-5A, (c) KLB100 (Enkelmann et al., 2017), (d) 15-SI-012-1. See Fig. 5 for description of additional plot elements. Time axis in (a), (b) versus (c), (d) is variable to visualize simulation outputs.
Figure 7 Integrated thermal history model results. FFF multi-sample inversion results include blue, red, and black lines for the expected paths and 95% credible intervals (shaded envelopes) for highest (blue) and lowest (red) elevation samples. Thin black lines show expected time-temperature paths for intermediate elevation samples with credible intervals omitted for clarity. Gray envelopes and corresponding colored paths show results of NF single-sample inversion results (Fig. 6).

5.2 Thermal history simulation results

Inverse thermal history models of all FFF samples and NF samples 61115-2A and 61615-5A reveal a three-phase cooling history with similar features. We define Phase I as a period of rapid cooling (~10 °C/Myr) from ~95 Ma to 75 Ma (Figs. 5; 6a, b; 7). FFF apatite FT, low eU zircon He, and zircon FT data are broadly concordant implying rapid cooling through the effective T_c associated with those systems. Zircon FT dates suggest, however, that the magnitude and rates of cooling for Phase I may be spatially variable (e.g., compare the ~136 Ma zircon FT date of sample 61115-2A with ~95 Ma dates for
Phase II is characterized by near-isothermal to slow cooling conditions, or perhaps minor reheating, at ≤1 °C/Myr from ~75–30 Ma (Figs. 5; 6a, b; 7). FFF single- and multi-sample inversions indicate residence at low-temperatures of ~60–20 °C during this 50 Myr interval. The narrow temperature range is controlled by the interplay between apatite He dates and long mean AFT track lengths of ~14 μm, the latter requiring these samples could not have been heated to temperatures significantly above ~70–60 °C. Thermal history simulations for NF samples 61115-2A and 61615-5A indicate an extended ~70–50 Myr period of slow cooling, but at higher temperatures of ~180–160 °C and ~160–100 °C, respectively (Figs. 6a, b). This temperature range and cooling rate are required by the zircon He date-eU relationships exhibited by these samples (Fig. 3b).

Phase III is characterized by rapid cooling at ~5–10 °C/Myr within the Kluane Ranges beginning by ~30 Ma and continuing until present. Differences between pre-Phase III temperatures for FFF and NF groups suggest gradients in cooling perpendicular to the EDFZ (Figs. 5-7). Pre-Phase III temperatures also vary within our NF group (e.g., compare KLB 100 versus 61115-2A, 61615-5A; Fig. 6), suggesting variable cooling parallel to the EDFZ (Figs. 6; 7). Single- and multi-sample inversions for FFF samples suggest initiation of Phase III cooling from ~30 Ma (61415-14A; Fig. 5c) to 18–10 Ma (KLB88 and multi-sample inversion; Figs. 5d; 7; Enkelmann et al., 2017), although the apatite He-elevation relationship shows a prominent break-in-slope at ~30 Ma (Fig. 4). Inversion results for NF samples 61115-2A and 61615-5A overlap and are consistent with the onset of Phase III cooling between 35-5 Ma and 15-5 Ma, respectively, within
95% credible intervals (Fig. 7). However, tT probability distributions for these simulations exhibit different local T peaks that are suggestive of possible variability in the onset of Phase III cooling (see white max-mode tT paths in Figs. 5; 6). These results favor initiation of Phase III cooling at ~5 Ma and ~25 Ma for NF samples 61115-2A and 61615-5A, respectively (Figs. 6a, b). Inversions for samples KLB100 and 15-SI-012-1 only constrain rapid cooling since ~30 and ~12–10 Ma, respectively (Fig. 6c, d). Model results are thus consistent with spatial variability in the onset of Phase III cooling across the study area, but do not explicitly require it.

6. Oligocene to present patterns of Kluane Ranges exhumation and surface uplift

In this section we calculate the rates and magnitudes of net surface uplift, exhumation, and total surface uplift for the Kluane Ranges from thermochronometric data and from previously published geological information available for Phase III. Our goal is to place limits on the spatial distribution and rates of topographic growth and vertical strain to (1) provide additional insight into processes accommodating deformation in the Kluane Ranges, and (2) serve as a point of comparison to similar datasets published for the CDFZ (Benowitz et al., 2011, 2014; Fitzgerald et al., 1995). The displacement of rock (i.e., a point in a volume) relative to the geoid by deformation and(or) isostasy is defined as rock uplift, the displacement of the Earth’s mean surface relative to the geoid is defined as surface uplift, and the displacement of rock relative to the surface by erosion or tectonic thinning is defined as exhumation (England and Molnar, 1990). These terms are related by the expression (England and Molnar, 1990):

\[ SU = RU - E \]  

(1)
where SU is surface uplift, RU is rock uplift, and E is exhumation.

Surface uplift represents work against gravity and is of interest for quantifying tectonic processes over spatial scales on the order of $10^3$ km$^2$ (Abbott et al., 1997; Brown, 1991; England and Molnar, 1990). In the absence of local thermal anomalies, thermochronometric data record cooling associated with exhumation, but calculation of surface uplift requires knowledge of a region’s paleoelevation prior to the onset of tectonism (e.g., Brown, 1991; England and Molnar, 1990; Reiners and Brandon, 2006). Total surface uplift (TSU) equates to the spatial average of rock uplift at individual points and adjusted for potential isostatic effects of erosion, whereas net surface uplift (NSU) is the residual of TSU and erosion (Abbott et al., 1997; Brown, 1991).

Here we first focus on NSU, the difference between present-day mean elevation and mean paleoelevation, and exhumation derived from thermochronometry data. From these values we calculate TSU following the backstacking approach of Brown (1991). For this exercise, we consider NSU over a ~1900 km$^2$ portion of our study area and next calculate a range of total surface uplift values corresponding to spatially-variable exhumation captured by our NF and FFF sample groups.

6.1 Paleoelevation and net surface uplift calculations

Published paleobotanical data place constraints on paleoelevation of the Kluane Ranges prior to Phase III cooling. We estimate paleoelevation, $Z_0$, from isochronous sea level and land surface temperature differences following (Axelrod, 1997):

$$Z_0 = \frac{T_{se} - T_a}{L} + \Delta H$$  \hspace{1cm} (2)
where, $T_{ss}$ is sea surface temperature, $T_a$ is air temperature at the site of interest, $L$ is an assumed lapse rate, and $\Delta H$ is the difference in sea level relative to today. NSU is thus:

$$NSU = Z - Z_0 = Z - \frac{T_{ss} - T_a}{L} + \Delta H$$

(3)

where $Z$ is modern mean elevation (Fig. 8A).

Paleobotanical data from latest Eocene-Oligocene terrestrial clastic rocks of the Amphitheatre Formation (Ridgway and Sweet, 1995) combined with contemporaneous $T_{ss}$ records place limits on paleoelevation at the beginning of Phase III cooling. Mean annual $T_a$ for the late Eocene lower Amphitheatre Formation is $16.7 \pm 0.4 \, ^\circ C$ and is constrained by the co-existence approach, which utilizes comparisons between paleoflora and modern relatives’ climactic tolerances to define a zone of maximum overlap (Pound and Salzmann, 2017). Contemporaneous $T_{ss}$ at paleolatitudes similar to our study area are assumed to be $20 \pm 3 \, ^\circ C$ based on available organic biomarker ($\sim 19 \pm 3 \, ^\circ C$), Ca/Mg thermometry ($\sim 21.5 \pm 3.5 \, ^\circ C$), and paleobotanical datasets ($\sim 20 \pm 1 \, ^\circ C$) (Evans et al., 2018; Liu et al., 2009; Wolfe, 1978). For $L$ and $\Delta H$, we assume the global average lapse rate of $5.9 \pm 1.1 \, ^\circ C/km$ and a eustatic sea level 100 m higher than present day (Haq et al., 1988; Meyer, 1992). Using these values in Equation (2) and propagating the quoted uncertainties for each parameter yields a paleoelevation of $0.7 \pm 0.6 \, km$. We acknowledge this estimate does not consider factors that may complicate paleobotanical paleoaltimetry or uncertainties in the precise paleolatitude of our study area (c.f. Meyer, 2007). However, our estimate provides a first-order approximation, based on geological data and with conservative uncertainties, through which to calculate NSU and TSU.

Amphitheatre Formation basins are distributed throughout the Kluane Ranges and record similar depositional and climactic conditions (Eisbacher and Hopkins, 1977;
Deposition occurred in small, isolated basins that overlap a previously interpreted regional, low-relief erosion surface with modern elevations of ~1700–2000 m a.s.l. (Fig. 8a; Eisbacher and Hopkins, 1977). The lower Amphitheatre Formation likely corresponds to incipient basin formation and subsidence, and therefore represents only a small departure from the mean elevation of the broader landscape. We assume that our paleoelevation estimate is approximately representative of the mean elevation of the Kluane Ranges at the onset of Phase III cooling. Additionally, we emphasize that this does not preclude the presence of small-scale topographic heterogeneity and that the validity of our calculation rests on this assumption.

Considering a present-day mean elevation of 1.65 km yields NSU of ~1.0 ± 0.6 km (Fig. 8a). Assuming a variable onset of Phase III cooling between 30 and 10 Ma indicated by our thermal history simulations, corresponding NSU rates are calculated between ~0.03 ± 0.02 mm/yr and ~0.1 ± 0.06 mm/yr, respectively.
Figure 8 (previous page) Phase III net surface uplift, exhumation, and total surface uplift. **(a)** Schematic diagram showing relations among topography (Fig. 4a) in the Kluane Ranges, inferred paleoelevation (dark gray, arrows give 1σ uncertainty, dashed where extent is uncertain), and net surface uplift (black arrows). Present-day mean elevation of bedrock pediments shown (black, wavy line with amplitude proportional to std. dev. of pediment modern elevation) **(b)** Exhumation (symbols as in Figure 2) and exhumation rates (gray) from low-temperature thermochronometry data shown as mean ± 1σ standard deviation. Red bar denotes results from regression. Vertical and horizontal extent of bar corresponds to 1σ error or regression and samples included in regression, respectively. **(c)** Total surface uplift assuming Airy isostatic compensation (white boxes). Rates calculated over 30 Ma and 10 Ma shown as gray boxes with teal and blue outlines, respectively. **(d)** Total surface uplift rates assuming no isostatic compensation. Symbology as in (c).

6.2 Exhumation magnitude and exhumation rate calculations

The magnitude and rate of exhumation are estimated from thermochronometric data by assuming a geothermal gradient, effective T<sub>c</sub>, and an onset of exhumation. Geothermal gradient data are not available for the Kluane Ranges; however, elsewhere along the Denali fault zone the geothermal gradient is ~30 °C/km (Fisher et al., 2004). The closest available measurements to our study area in Yukon are ~25 °C/km, although values of ~30 °C/km are measured near the Tintina fault in Yukon, another intracontinental strike-slip fault (Fig. 1b; Fraser et al., 2019). Post 30-Ma magmatism may have also resulted in a higher paleogeothermal gradient within the Kluane Ranges. Although the effect of past magmatism on crustal thermal structure is difficult to quantify, a possible analogue is the presently active portion of the Wrangell arc, where contemporaneous geothermal gradients are ~34 ± 6 °C/km (Batir et al., 2013). Because Wrangell volcanism was not as voluminous in the Kluane Ranges, this gradient estimate is likely an upper bound (e.g., Richter et al., 1990). Thus, we ultimately employ a geothermal gradient of 30 ± 5 °C/km to calculate exhumation.
Exhumation varies markedly between FFF and NF samples. Thermal history simulations for FFF samples indicate samples were at temperatures of ~60 °C prior to the onset of Phase III, suggesting ~2.0 ± 0.3 km of exhumation (Figs. 5; 8). Exhumation recorded by NF samples varies along strike of the EDFZ trace. Pre-Phase III temperatures for samples 61115-2A and 61615-5A are ~160 ± 10 °C and 140 ± 20 °C, respectively (Fig. 6a, b), yielding 5.3 ± 0.9 km and 4.6 ± 1.0 km of exhumation (Fig. 8b). NF samples KLB100 and 15-SI-012-1 do not constrain pre-Phase III temperatures but require a minimum of 7.3 ± 1.6 and 3.3 ± 0.9 km of exhumation, respectively (Figs. 6; 8b).

Exhumation rates are calculated by dividing exhumation by the timing of rapid cooling indicated by thermal history simulations (~30–18 Ma for FFF samples, ~30–10 Ma for NF samples; Figs. 5–7; Table S7), or the cooling date for samples that only constrain minimum exhumation. Rates are ~0.07 ± 0.02 mm/yr to 0.1 ± 0.02 mm/yr for FFF samples and ~0.5 ± 0.2 mm/yr to 0.3 ± 0.06 mm/yr for NF samples (Fig. 8b; Table S7).

An alternative means of estimating apparent exhumation rate that does not require a priori knowledge of the geothermal gradient is from regression of date-elevation relationships (Reiners and Brandon, 2006). Apatite He data from our FFF sample group define a positive correlation with elevation and can thus provide an additional estimate for Phase III exhumation rate (Fig. 4b). Apparent exhumation rate estimates from date-elevation transects can be impacted by topographic influences on isotherms (Stüwe et al., 1994). However, we anticipate this effect is minor in our samples because they are distributed over a horizontal distance of ~4–5 km perpendicular to the dominant topographic grain, which is less than the wavelength over which the apatite He $T_c$ isotherm is likely to be affected by topography (Figs. 2; 4a; Braun, 2002;). Weighted
least squares of <30 Ma FFF apatite He data yields an apparent exhumation rate of 0.07 ± 0.04 mm/yr since 30 Ma, in agreement with our estimate derived by assuming a geothermal gradient (Fig. 8b).

6.3 Total surface uplift calculations

We calculate magnitudes and rates of TSU (sometimes referred to as tectonic uplift, e.g., Brown (1991), or tectonic surface uplift, e.g., Abbott et al. (1997)) over two zones corresponding to FFF and NF sample groups using our estimates of NSU and exhumation. Assuming Airy isostasy, TSU is (rearranged after Equation (2) in Brown (1991)):

$$TSU = NSU + E \left(1 - \frac{\rho_c}{\rho_m}\right)$$  \hspace{2cm} (4)

where $\rho_c$ and $\rho_m$ are the density of the crust and mantle, respectively. This analysis assumes that spatial wavelength-dependent flexural effects associated with the lateral strength of the lithosphere are negligible and may overestimate the rock uplift contribution of isostasy (particularly for NF samples; Turcotte and Schubert, 2014). We thus also present an endmember estimate of TSU that excludes the isostasy term in Equation (4).

From Equation (4) and assuming values of 2700 and 3300 kg/m$^3$ for $\rho_c$ and $\rho_m$, TSU is $\sim$1.4 ± 0.7 km over the zone corresponding to FFF samples and $\sim$2.1 ± 1.4 km over the zone corresponding to NF samples. Corresponding rates are 0.1 ± 0.02 mm/yr (30 Ma) to 0.1 ± 0.07 mm/yr (10 Ma) for the FFF zone and 0.07 ± 0.05 mm/yr (30 Ma) to 0.2 ± 0.1 mm/yr (10 Ma) for NF zone (Fig. 8c). If isostasy is excluded, TSU becomes 3.0 ± 0.7 km and 7.0 ± 1.2 km for FFF and NF zones, respectively. Rates then become 0.1 ± 0.02
mm/yr (30 Ma) to 0.3 ± 0.07 mm/yr (10 Ma) and 0.2 ± 0.03 mm/yr (30 Ma) to 0.7 ± 0.1 mm/yr (10 Ma) for FFF and NF zones (Fig. 8d). Lastly, we note that all TSU rate calculations likely reflect the cumulative effect of more short-lived episodes of local extension and contraction over the duration of Phase III cooling (cf. Israel et al., 2005; Ridgway et al., 1992).

7. Geological interpretations of thermal history, exhumation, and surface uplift patterns

Results of thermal history simulations, as well as exhumation and total surface uplift rates for Phase III are integrated with local and regional geologic data to interpret the long-term topographic and tectonic evolution of the Kluane Ranges and EDFZ (Figs. 7; 9). Interpretive models are summarized in Figure 10 with three schematic block diagrams and representative cross-sections of TSU and mean topography (i.e., NSU) corresponding to Phases I (line A-A’), II (line B-B’), and III (line C-C’). Block diagrams include major fault and magmatic systems active during each phase based on previous work (Fig. 10; Andronicos et al., 1999; Dodds and Campbell, 1988; Cobbett et al., 2016; Gehrels et al., 2000; Lowey, 2000; Ingram and Hutton, 1994; Israel et al., 2011; Rubin et al., 1990). We assume a component of sustained right-lateral displacement along the EDFZ such that the Kluane Ranges (southwest of the EDFZ) were juxtaposed against Intermontane terrane (northeast of the EDFZ) features that are presently observed to the southeast of our study area at various times in the past. We make no inferences regarding the magnitude and rates of lateral displacement across the EDFZ during any phase, but
denote ~400 km of restored separation on the EDFZ on Figure 9b for broad comparison (Lowey, 1998).

7.1 Phase I

We interpret Phase I cooling observed in all Kluane Ranges samples as the exhumational response to final accretion of the Insular terranes (Figs. 7; 9; 10). The present-day EDFZ occupies the ancient suture zone between the Insular and Intermontane terranes, which has been the locus of long-lived erosion and tectonism (Nokleberg et al., 1985; Ridgway et al., 2002). In southwestern Yukon, Jurassic- Late Cretaceous back-arc basin flysch and metamorphic rocks of the Dezadeash Formation and Kluane Schist outcrop as a NW-SE trending, ~225 km belt that parallels, and is cut by, the EDFZ (Fig. 7; Eisbacher, 1976; Erdmer and Mortensen, 1993; Israel et al., 2011; Mezger et al., 2001). The timing of peak amphibolite facies metamorphism of the Kluane Schist is ~80–70 Ma and interpreted to record accretion of the Insular and Intermontane terranes (Fig. 9c; Erdmer and Mortensen, 1993; Mezger et al., 2001; Israel et al., 2005, 2011).

Regionally extensive contraction and possible attendant surface uplift within the Kluane Ranges is recorded by displacement along the DRF, imbricate thrusts, reverse faults, and widespread folding from ~110–80 Ma (Fig. 9b, c; Cobbett et al., 2016; Israel et al., 2005). Late Cretaceous fold-and-thrust belts active at ~100–90 Ma are documented within the Intermontane terranes both southeast and northwest of the Kluane Ranges and provide additional evidence for regional deformation related to terrane accretion (Fig. 9b; Israel et al., 2011; Rubin et al., 1990).
Figure 9 Simplified Kluane Ranges geology and cooling phases. (a) Simplified geologic map highlighting key tectonic elements of the Kluane Ranges that correspond to Phases I, II, and III of cooling paths. Locations of samples considered in thermal history inversions are shown (unfilled and filled circles as shown in Fig. 1). (b) Simplified map of regional geologic elements with labels corresponding to cooling phases discussed in main text. (c) Chronology of local and regional geologic events. References for timeline are: ¹Miller et al. (2002); ²Cobbett et al. (2016); ³Ingram and Hutton (1994); Gehrels et al. (2000); Lowey (2000); ⁴Eisbacher (1976), Cole et al., (1999), Miller et al., (2002), Riccio et al., (2014); ⁵Erdmer and Mortenson (1993), Mezger et al., (2001), Israel et al. (2011); ⁶, ¹⁵, ¹⁶Eisbacher and Hopkins (1977), Ridgway et al., (1992), Ridgway and Sweet (1995); ⁷Nokleberg et al., (1985), Rubin et al., (1990), Dusel-Bacon et al. (1993); ⁸Haeussler et al. (2003); ⁹Falkowska et al. (2014); ¹⁰Finzel et al. (2011); ¹¹, ¹², ¹³, ¹⁴Dodds and Campbell (1988); ¹⁷Denton et al. (1974).
Late Jurassic through Late Cretaceous cooling is documented elsewhere within the Insular terranes with thermochronometric data and attributed to terrane accretion and concomitant magmatism (Enkelmann et al., 2017; Falkowski and Enkelmann, 2016). Our thermochronometric data are from ~117–106 Ma pluton samples or localities <5–10 km away from mapped surface exposures of these units (Figs. 7; 9; Dodds and Campbell, 1988). Owing to the ~10–20 Myr time lag between crystallization ages from these rocks and Phase I cooling, we posit that documented ~95–75 Ma cooling in NF and FFF samples in the Kluane Ranges does not likely reflect post-magmatic thermal relaxation. However, because magmatism continued regionally throughout this time period, we cannot completely rule out this possibility (Dodds and Campbell, 1988).

Phase I exhumation associated with terrane accretion potentially coincides with activity on an incipient EDFZ in the mid-Late Cretaceous. Rapid cooling in the Kluane Ranges adjacent to the EDFZ to may reflect components of exhumation associated with surface uplift and relief creation triggered by contraction within a regional dextral transpressive deformation regime (Doubrovine and Tarduno, 2008), although differences in cooling recorded by NF and FFF sample groups imply spatially heterogeneous exhumation and associated retention of thermal history (Figs. 5–7). Active contraction in both the Insular and Intermontane terranes and arc magmatism are consistent with surface uplift and associated relief generation at regional scales in response to terrane accretion (Fig. 10a, A-A’). The Coast shear zone, interpreted as an extension of the EDFZ, exhibits dextral transpressive deformation at ~85-55 Ma (Andronicos et al., 1999). In addition, the CDFZ and WDFZ preserve evidence for components of dextral displacement at ~70–60 Ma and ~85–70 Ma, respectively (Cole et al., 1999; Miller et al., 2002). Timing
constraints on dextral shear for the Denali fault broadly overlap with the timing of Phase I cooling, suggesting the EDFZ may have been active by the Late Cretaceous.

Figure 10 Tectonic and topographic evolution of the Kluane Ranges and environs. Schematic block diagrams of active faults (red), magmatic systems (orange), and topography associated with EDFZ for Phases (a) I, (b) II, and (c) III time slices. Cross-sections show relative mean topography (topo, black), and total surface uplift (TSU, gray) patterns for each time slice. Topography for Phases I and II are inferred (see section 7). Phase III topography is modern DEM draped over terrain. C-C’ cross-section of mean topography and TSU modified from Figure 8. For all diagrams, inferred geological elements denoted with dashed lines. Question-marked movement arrows in block diagram (a) reflect ambiguity of EDFZ kinematics over Phase I cooling.

7.2 Phase II

We interpret Phase II near-isothermal holding or slow cooling to reflect development of a stable, low-relief landscape over the latter half of the Late Cretaceous and much of
the Paleogene in the Kluane Ranges region (Figs. 7; 10b, B-B’). In this scenario, cessation of surface uplift following Phase I terrane accretion results in decaying topography. Slow cooling and associated low exhumation rates ensue as the landscape approaches a new equilibrium (e.g., Spotila, 2005). Low-relief surfaces and bedrock pediments underlying latest Eocene-Oligocene Amphitheatre Formation are distributed throughout our study area (Figs. 8a; 9). These are interpreted as remnants of an early Cenozoic erosion surface, reflecting relative vertical tectonic quiescence and relief degradation (Eisbacher and Hopkins, 1977). Thermal history simulations for both NF and FFF samples suggest <1 km of exhumation over ~70–30 Ma (Fig. 7).

Development of a low-relief landscape, limited exhumation, and activity along the EDFZ during Phase II are not mutually exclusive. First, strike-slip faulting sense stricto does not induce substantial contraction, hence little surface uplift and erosion/exhumation, and can thus go undetected by thermochronometric techniques (Duvall et al., 2013). Phase II kinematics of the EDFZ may have been dominantly strike-slip relative to Phase I contraction-dominated deformation, as has been suggested for the CDFZ (Cole et al., 1999). The bulk of separation measured along the EDFZ is inferred to have occurred ~post-55 Ma, further suggesting continuing dextral slip along the EDFZ over Phase II (Eisbacher, 1976; Lanphere, 1978; Riccio et al., 2014). Second, the locus of contractual strain and attendant exhumation may have been concentrated elsewhere during Phase II slow cooling in the Kluane Ranges. Over at least part of the Phase II cooling period, our study area was possibly adjacent to previously documented Late Cretaceous–Eocene dextral transpressive or contractual shear zones within the Intermontane terranes but to the southeast of our study area (Fig. 9b, c; Gehrels et al.,
Rapid cooling from ~70–45 Ma within the Intermontane terranes and northeast of the EDFZ, both directly adjacent to and ~ 100 km southeast of our study area, is recorded by low-temperature thermochronometry data (Figs. 2; 4a; Enkelmann et al., 2017; Falkowski and Enkelmann, 2016). Exhumation of the Kluane Schist also occurred post 55-Ma (Mezger et al., 2001). Spatial migration of contractional deformation from the Kluane Ranges to the adjacent Intermontane terranes in Phase II may be due to strength contrasts facilitated by arc magmatism in the Intermontane terranes and a lack thereof in our study area (Figs. 9c; 10b; Enkelmann et al., 2017; Ingram and Hutton, 1994). Thus, thermal history simulations coupled with geologic observations imply subdued topography of the Kluane Ranges adjacent to regions of relatively higher elevation separated by a then-active EDFZ (Fig. 10b, B-B’).

7.3 Phase III

Spatially and temporally variable rapid Phase III cooling is interpreted as exhumation triggered by surface uplift and concomitant erosion from dominantly transpressive deformation within and west of the EDFZ over the last ~30 Myr. Deposition of latest Eocene/Oligocene-Miocene polymictic fluvial sediments and volcanics of the Amphitheatre Formation and Wrangell Volcanics in local, diachronous, transtensional and transpressional strike-slip basins is coeval with the onset of Phase III cooling (Fig. 9c; Eisbacher and Hopkins, 1977; Ridgway et al., 1992; Ridgway and Decelles, 1993; Ridgway and Sweet, 1995). Individual basins are distributed along the DRF and EDFZ, although most are located to the west of our thermochronometry sample locations (Fig. 9a). Paleocurrents in these basins are variable, although the mean direction
and facies distributions of basins within our study area indicate generally southwestward flow. Clasts are dominantly derived from the Insular terranes, although some basins contain clasts from Intermontane terranes (Fig. 9a; Eisbacher and Hopkins, 1977; Ridgway et al., 1992; Ridgway and Decelles, 1993). Individual basins with a component of Intermontane terrane provenance, despite no eastward flowing rivers presently connecting these basins to the Intermontane terranes, are possibly indicative of drainage reorganization in response to topographic growth along the EDFZ. Geologic mapping reveals thrusting of Paleozoic basement on basin sediments in some locales, recording post-depositional basin inversion and contraction (Israel et al., 2005). Cobbett et al. (2016) also document thrusting on the DRF over this same time interval (Figs. 9c; 10c).

Broad spatial and temporal trends in exhumation and surface uplift are consistent with geological evidence for regional and local transpression, as well as elevation gain over Phase III. Cooling and exhumation recorded by our thermochronometry data, combined with paleocurrent and provenance data from synorogenic basins to the west of our sample locations, suggest emergent topography and near-field relief generation along the Denali fault over Phase III. Phase III NSU is indicative of vertical strain distributed over our study area, with an enhanced zone of exhumation and TSU <2 km from the EDFZ, consistent with geological evidence for contractional deformation (Cobbett et al., 2016; Israel et al., 2005). Previously published apatite He dates within the Intermontane terranes and at ~800 m a.s.l. average ~10–20 Myr older than FFF dates at similar EDFZ-perpendicular distances and higher elevations, indicating more limited exhumation northeast of the modern EDFZ (Fig. 4a; Enkelmann et al., 2017; Falkowski and Enkelmann, 2016). Present-day seismicity is also more subdued northeast of the EDFZ.
relative to the Kluane Ranges, similarly indicating spatially asymmetric Phase III deformation (He et al., 2018).

Patterns of cooling, exhumation, and TSU may be complicated by additional factors. First, apparent elevated exhumation rates near the EDFZ may be influenced by heat advection by fluid flow, either through local compression of the geothermal gradient in a mountain aquifer discharge zone or resetting of thermochronometric dates by sufficiently hot fluids after passage through the $T_c$ for any of the thermochronometric systems used here (Ehlers and Chapman, 1999; Forster and Smith, 1989; Reiners, 2009). Second, enhanced exhumation within NF samples may be a function of rock damage proximity to the EDFZ trace, rather than true deformation gradients (e.g., Molnar et al., 2007). Lastly, we assume cooling is entirely related to exhumation rather than localized re-heating and cooling by Miocene volcanism of the Wrangell suite (Fig. 9a, c). However, presently-exposed shallow intrusives of the Wrangell suite and sedimentation patterns also independently indicate Phase III exhumation (Dodds and Campbell, 1988; Eibshach and Hopkins, 1977; Ridgway et al., 1992).

We suggest Kluane Ranges Phase III transpression is the far-field response to Yakutat microplate subduction and collision. Flat-slab subduction of that Yakutat terrane began by ~35–30 Ma, with transition to microplate collision at ~15–12 Ma (Fig. 9c; Falkowski et al., 2014; Finzel et al., 2011). Adakatic and calc-alkaline lavas of the Wrangell Volcanic belt record magmatism associated with the slab edge by at least ~30–25 Ma in the Kluane Ranges (Fig. 9a; Brueseke et al., 2019; Richter et al., 1990). From ~18–11 Ma, lavas erupted along “leaky” strike-slip and normal faults with MORB-like signatures, suggesting a change in magma reservoir and possible asthenospheric
upwelling following northwestward migration of the slab edge (Fig. 9a; Richter et al., 1990; Skulski et al., 1992). Surface uplift across our study area is consistent with regional crustal thickening associated with transpression and(or) possible dynamic surface uplift related to post-11 Ma asthenospheric upwelling.

Rapid cooling and exhumation at ~30 and ~15 Ma are recorded throughout the broader St. Elias Mountains and similarly attributed to Yakutat subduction and microplate collision dynamics, respectively (Enkelmann et al., 2017; Falkowski et al., 2014; Falkowski and Enkelmann, 2016; O’Sullivan and Currie, 1996; Spotila and Berger, 2010). For example, Kluane Ranges samples that record Phase III cooling and exhumation at ~15–10 Ma (61115-2A and 15-SI-012; Fig. 6) may reflect a response to microplate collision, which is also inferred in the St. Elias Mountains (Falkowski et al., 2014; O’Sullivan and Currie, 1996; Spotila and Berger, 2010). Alternatively, exhumation in the Kluane Ranges may reflect other factors such as changing plate convergence angle with respect to EDFZ orientation (DeMets and Merkouriev, 2016). Oligocene-Miocene cooling and(or) deformation is documented along the broader Denali fault system, although variable in timing, and also attributed to flat-slab subduction of the Yakutat microplate (Fig. 9b; Benowitz et al., 2011, 2014; Lease et al., 2016; Trop et al., 2004; Waldien et al., 2018). Our data implies EDFZ activity triggered by far-field plate boundary processes over timescales beyond the resolution of our thermochronometric data.
7.4 Oligocene to present patterns of regional exhumation and surface uplift along the Denali fault zone

Spatial patterns of Phase III exhumation and surface uplift along the EDFZ in comparison with the CDFZ are of markedly different magnitude, but have similar spatial patterns at the local scale (i.e., adjacent to the fault trace). Persistent but spatially variable Phase III-equivalent exhumation and surface uplift is documented since ~6 and ~25 Ma within the CDFZ-bounded Denali massif and eastern Alaska Range, respectively (Benowitz et al., 2011, 2014). Total exhumation in the Alaska Range is ~5–6 km and locally >11 km (rates of ~0.5–1 mm/yr; Benowitz et al., 2011, 2014; Fitzgerald et al., 1995). At Denali, estimates of NSU and TSU are ~3 km (~0.5 mm/yr) and ~9 km (~1.5 mm/yr; ~4 km and ~0.7 mm/yr accounting for isostasy). Exhumation, NS, and TSU in the Kluane Ranges are ~2-6 km, ~1 km, and ~1-7 km (~1-2 km accounting for isostasy; Fig. 8). We document an ~2 km-wide zone of near-fault enhanced exhumation and total surface uplift along the EDFZ in the Kluane Ranges, qualitatively similar to the CDFZ although the latter exhibits fault-perpendicular exhumation and TSU gradients over 10s of km (Fig. 8; Benowitz et al., 2011, Fitzgerald et al., 1995). We suggest fault-perpendicular gradients in deformation reflect near-field partitioning of contractional strain adjacent to the EDFZ, similar to the CDFZ and other strike-slip fault systems, although over more localized spatial scales (e.g., Fitzgerald et al., 2014; Spotila et al., 2007; Tikoff and Teyssier, 1994; Tippett and Kamp, 1993).

Different magnitudes of exhumation and surface uplift over 100’s of km along-strike between the CDFZ and EDFZ are likely the result of constructive interactions between fault geometry, regional faulting patterns, and geodynamic factors. First, the Denali fault
exhibits ~70° of along-strike curvature (St. Amand, 1957). The Insular terranes block is thought to move along the Denali fault system through counterclockwise rotation, although fault-convergence obliquity is greater for the CDFZ than EDFZ (Figs. 1; 9b; Freymueller et al., 2008). Pronounced differences in position along the fault relative to curvature between the EDFZ and CDFZ may thus explain along-strike heterogeneity in exhumation and surface uplift, as suggested by prior work (Fitzgerald et al., 2014; Lease et al., 2016; Riccio et al., 2014). Second, our study area also lies to the southeast of the Totshunda fault-EDFZ intersection. The Totshunda fault presently accommodates part of the total Denali fault slip budget and was active by at least 30 Ma (Brueseke et al., 2019; Haeussler et al., 2017; Marechal et al., 2018; Matmon et al., 2006). It is thus permissible that Phase III deformation was not exclusive to the EDFZ, but rather shared between the EDFZ and Totshunda faults. Lastly, the distance between the subducted extent of the Yakutat Microplate and Denali fault varies along-strike. The Alaska Range currently overlies part of the subducted plate and is only ~100–200 km inboard of the northern edge of the flat-slab region (Fig. 9b; Eberhart-Phillips et al., 2006). The Kluane Ranges do not presently overlie the Yakutat microplate, and may not have for approximately one third of Phase III if inferences regarding northwestward slab migration at ~11 Ma are correct (Figs.1; 9b; Richter et al., 1990). Plate boundary stresses are more efficiently transferred to continental interiors by flat-slab subduction relative to normal subduction, indicating that spatiotemporal patterns of subducting-upper plate coupling may play a role in the spatial expression of deformation (e.g., Gutscher et al., 2000).
8. Conclusions

We present a multi-thermochronometer dataset for the EDFZ-bounded Kluane Ranges that reveals linkages between the evolution of the EDFZ and the thermotectonic history of its environs. Our dataset demonstrates the utility of exploiting multiple thermochronometers and radiation damage accumulation/annealing effects to reveal regionally-consistent thermal histories in lieu of poor mineral yields across a larger sample set. For all thermochronometric systems, the youngest dates and greater magnitude of exhumation are found near the modern trace of the EDFZ. Data patterns also indicate the occurrence of EDFZ-parallel gradients in differential exhumation. Time-temperature simulations reveal a three-phase cooling history shared by all samples characterized by ~95–75 Ma rapid cooling (Phase I), ~75–30 Ma slow cooling (Phase II), and spatially and temporally variable relatively rapid cooling from ~30 Ma to present (Phase III). Phase III TSU rates are calculated from exhumation rate data and by assuming paleoelevation constraints from previously published palynological data. Our results are consistent with vertical strain focused immediately adjacent to the EDFZ.

Time-temperature histories from inversion of thermochronometric data and cooling date spatial patterns are integrated with local and regional geology to document exhumation and surface uplift over ~100 Ma in a strike-slip fault-bounded mountain range. We interpret Phase I cooling as the erosional response to surface uplift caused by contraction associated with accretion of the Insular terranes to the former continental margin, formation of a broad region of elevated topography, and deformation along the incipient EDFZ. Slow cooling during Phase II is inferred to reflect decreasing relief, exhumation, and surface uplift within the Kluane Ranges. This can be reconciled with
continued motion along the EDFZ if strike-slip motion was dominant along the EDFZ during Phase II and (or) contractional deformation was focused elsewhere as a possible consequence of shifting arc magmatism patterns. Phase III rapid cooling is spatially and temporally variable and interpreted to reflect surface uplift, relief generation, and exhumation triggered by transpressive deformation as a far-field response to flat-slab subduction of the Yakutat microplate.

Quantitative constraints on paleoelevation over Phase III reveal that ~60% of the Kluane Ranges modern mean elevation was attained in the past 30 Myr. Phase III patterns of total surface uplift imply strain partitioning at distances of ≤2 km perpendicular to the EDFZ. Phase III exhumation, NSU, and TSU rates for the EDFZ are markedly lower than those documented for the CDFZ. We speculate comparatively lower EDFZ rates are due to combined geometric complexity along the full extent of Denali fault system, interactions with other regional-scale faults, and variations in the degree of subducting-upper plate coupling that transmits plate boundary stresses into the continental interior. Our integrated results and interpretations indicate spatial and temporal variability in deformation and topographic evolutionary patterns along intracontinental strike-slip fault systems can be transient and reflect the interplay between multiple processes.

References


Buscher, J. T., and J. A. Spotila (2007), Near-field response to transpression along the southern San Andreas fault, based on exhumation of the northern San Gabriel Mountains, southern California, Tectonics, 26(5).


Dodds, C., and R. Campbell (1992), Geology of SW Kluane Lake Map Area (115G and F (E1/2)), Yukon Territory, Geological Survey of Canada, Open File, 2188(1), 250,000.

Dodson, M. H. (1973), Closure temperature in cooling geochronological and petrological systems, Contributions to Mineralogy and Petrology, 40(3), 259-274.

Donelick, R. A., P. B. O’Sullivan, and R. A. Ketcham (2005), Apatite fission-track analysis, Reviews in Mineralogy and Geochemistry, 58(1), 49-94.


England, P., and P. Molnar (1990), Surface uplift, uplift of rocks, and exhumation of rocks, Geology, 18(12), 1173-1177.


Falkowski, S., E. Enkelmann, and T. A. Ehlers (2014), Constraining the area of rapid and deep-seated exhumation at the St. Elias syntaxis, Southeast Alaska, with detrital zircon fission-track analysis, Tectonics, 33(5), 597-616.


Farley, K. A. (2002), (U-Th)/He dating: Techniques, calibrations, and applications, Reviews in Mineralogy and Geochemistry, 47(1), 819-844.


Fitch, T. J. (1972), Plate convergence, transcurrent faults, and internal deformation adjacent to southeast Asia and the western Pacific, Journal of Geophysical research, 77(23), 4432-4460.


Gleadow, A., I. Duddy, P. F. Green, and J. Lovering (1986), Confined fission track lengths in apatite: a diagnostic tool for thermal history analysis, Contributions to Mineralogy and Petrology, 94(4), 405-415.


Israel, S., A. Tizzard, J. Major, and D. Emond (2005), Bedrock geology of the Duke River area, parts of NTS 115G/2, 3, 4, 6 and 7, southwestern Yukon, Yukon exploration and geology, 139-154.


Israel, S., D. Murphy, V. Bennett, J. Mortensen, and J. Crowley (2011), New insights into the geology and mineral potential of the Coast Belt in southwestern Yukon, Yukon exploration and geology, 101-123.


Lowey, G. W. (2000), The Tatshenshini shear zone (new) in southwestern Yukon, Canada: Comparison with the Coast shear zone in British Columbia and southeastern Alaska and implications regarding the Shakwak suture, Tectonics, 19(3), 512-528.


Mazzotti, S., and R. D. Hyndman (2002), Yakutat collision and strain transfer across the northern Canadian Cordillera, Geology, 30(6), 495-498.


Pound, M. J., and U. Salzmann (2017), Heterogeneity in global vegetation and terrestrial climate change during the late Eocene to early Oligocene transition, Scientific Reports, 7, 43386.


Read, P. B., and J. Monger (1976), Pre-Cenozoic volcanic assemblages of the Kluane and Alsek Ranges, southwestern Yukon Territory.


Ridgway, K., P. Decelles, A. Cameron, and A. Sweet (1992), Cenozoic syntectonic sedimentation and strike-slip basin development along the Denali fault system, Yukon Territory, Yukon Geology, 3, 1-26.

Ridgway, K. D., and P. G. Decelles (1993), Stream-dominated alluvial fan and lacustrine depositional systems in Cenozoic strike-slip basins, Denali fault system, Yukon Territory, Canada, Sedimentology, 40(4), 645-666.


Spotila, J. A. (2005), Applications of low-temperature thermochronometry to quantification of recent exhumation in mountain belts, Reviews in Mineralogy and Geochemistry, 58(1), 449-466.


St. Amand, P. (1957), Geological and geophysical synthesis of the tectonics of portions of British Columbia, the Yukon Territory, and Alaska, Geological Society of America Bulletin, 68(10), 1343-1370.

Stacey, J. t., and J. Kramers (1975), Approximation of terrestrial lead isotope evolution by a two-stage model, Earth and planetary science letters, 26(2), 207-221.


Tagami, T., H. Ito, and S. Nishimura (1990), Thermal annealing characteristics of spontaneous fission tracks in zircon, Chemical Geology: Isotope Geoscience Section, 80(2), 159-169.


Wolfe, J. A. (1978), A paleobotanical interpretation of Tertiary climates in the Northern Hemisphere: Data from fossil plants make it possible to reconstruct Tertiary climatic changes, which may be correlated with changes in the inclination of the earth's rotational axis, American Scientist, 66(6), 694-703.

CHAPTER 6

DATING FAULT DAMAGE ALONG THE EASTERN DENALI FAULT ZONE WITH HEMATITE (U-TH)/HE THERMOCHRONOMETRY¹

Abstract

Unraveling complex slip histories in fault damage zones and links between deformation, hydrothermal alteration, and topographic growth remains a challenge. The dextral eastern Denali fault zone (EDFZ; southwest Yukon, Canada) bounds the Kluane Ranges and hosts a variety of fault rocks, including hematite fault surfaces, that have been exhumed through the brittle regime over a protracted period of geologic time. Scanning electron microscopy-based microtextural observations and hematite (U-Th)/He (hematite He) thermochronometry from these surfaces indicate multiple generations of foliated, high-aspect ratio hematite plates. Single-aliquot hematite He dates (n=38) from 11 samples range from 11.5 ± 3.2 Ma (2σ) to 3.4 ± 2.1 Ma and exhibit moderate inter- and intrasample dispersion. A subset of dates is 15-20 Myr younger than previously published apatite (U-Th)/He dates from collocated host rocks, despite similar closure temperatures, which precludes a simple ambient cooling interpretation for our hematite He data. Statistical tests define hematite He date populations at ~8 Ma, ~6 Ma, and ~4 Ma, and when combined with microtextural observations, support episodes of hydrothermal alteration and reactivation at aseismic to subseismic slip rates. Additionally, there is no evidence that hematite experienced deformation- or

®Utah State University, ®U.S. Geological Survey
hydrothermal fluid-related He loss. Hematite He dates spatially and temporally overlap with previously documented Kluane Ranges surface uplift, highlighting a connection between deformation processes and the fault networks that accommodate topographic change. Hematite He dates appear consistent with a far-field response to Yakutat microplate dynamics, providing a link between the evolution of the EDFZ and plate boundary processes.

1. Introduction

Radiometric dating of fault zone materials provides critical constraints on deformation histories of crustal-scale fault zones and context for adjacent topographic growth. Commonly utilized methods for dating brittle regime faulting include $^{40}$Ar/$^{39}$Ar or K-Ar dating of clay-rich fault gouge and U-series or U-Pb geochronology of carbonate fault rocks (e.g., Mottram et al., 2020; Nuriel et al., 2012; Roberts and Walker, 2016; Vrolijk and van der Pluijm, 1999). Brittle fault zones focus deformation, alteration, and heat transport over micron to kilometer scales and at durations of seconds to millions of years (Louis et al., 2019; Sibson, 1994). Radiometric dates from fault rocks may reflect integrated and overprinting exhumation, thermal resetting by deformation, and(or) fluid-rock interaction, as well as multiple generations of each process to varying degrees (e.g., Ault, 2020). Deconvolving these various factors embedded within geo- or thermochronometry dates is essential to linking fault-specific processes to regional strain. Not all fault zones yield material amenable to existing geochronologic techniques, motivating the development of additional approaches for documenting the timing of fault zone thermal processes.
Hematite is common in fault rocks, is amenable to (U-Th)/He (hematite He) thermochronometry, and provides an opportunity to infer deformation processes by combining thermochronometry with microtextural analysis (Ault, 2020; Ault et al., 2016, 2015; McDermott et al., 2017; Moser et al., 2017). Hematite precipitates in faults from hydrothermal fluids or oxidation of Fe-bearing phases over a broad range of depths and temperatures (e.g., Cornell and Schwertmann, 2003), but radiogenic He is retained in hematite at temperatures of ~50-250 °C, depending on grain size (Evenson et al., 2014; Farley, 2018). Hematite He data constraints on fault processes beyond simple cooling due to exhumation depend on hematite formation depth (i.e., ambient temperature) versus temperature-sensitive He retentivity. In addition to exhumation, hematite He dates from fault surfaces can record a variety of temperature- and grain-size sensitive phenomena in fault zones. These include fluid flow and hematite precipitation (Ault et al., 2016; Moser et al., 2017), He loss from frictional heating (Calzolari et al., 2020), or temperature-induced recrystallization (Ault et al., 2015; McDermott et al., 2017). Only hematite formed below its exhumation-related closure temperature (T_c) can potentially reflect the timing of brittle deformation.

Hematite-coated fault surfaces adjacent to in the eastern Denali fault zone (EDFZ; Caine et al., 2015) offer an opportunity to directly date fault slip, and link these timing constraints to the surface uplift history and broader geodynamic processes. The dextral Denali fault system is an archetypal, segmented, intracontinental strike-slip fault system influenced by far-field plate boundary effects (Benowitz et al., 2011; Fitzgerald et al., 1995). Limited exhumation throughout the EDFZ-bounded Kluane Ranges in southwestern Yukon yields country rock that preserves a ~100 Ma to present
thermotectonic and surface uplift record accessible through conventional thermochronometry and paleoelevation data (Enkelmann et al., 2017; Falkowski and Enkelmann, 2016; McDermott et al., 2019). This limited exhumation masks the conventional thermochronometric record of recent EDFZ faulting and how it may (or may not) mirror the full extent of local and regional deformation patterns. However, limited exhumation also favors preservation of thermochronometric signatures of fault-related deformation, as opposed to ambient cooling, from the fault rocks themselves.

Here, we exploit the limited exhumation adjacent the EDFZ to explore the timing of deformation and hydrothermal alteration with hematite He thermochronometry and complementary microtextural observations from EDFZ hematite-coated fault surfaces. These new data are combined with prior constraints on the thermotectonic and paleogeomorphic history of the Kluane Ranges to examine relations between formative fault zone processes, topographic change, and geodynamics. We illustrate how hematite He thermochronometry from fault surfaces, together with field and microtextural observations, provide direct connections between regional surface uplift and intracontinental fault networks that accommodate topographic change.

2. Background

2.1 Tectonic framework and Denali fault zone

The Northern Cordillera comprises Phanerozoic accreted terranes transected by Mesozoic and Cenozoic intracontinental strike-slip faults such as the Denali fault in southwestern Yukon (Fig. 1a; Colpron et al., 2007). Translation of terranes along these dextral intracontinental strike-slip faults initiated in the Late Cretaceous or Eocene and
continues to the present (Plafker et al., 1989). The Denali fault system separates the Insular terranes, or Paleozoic-Mesozoic allochthonous assemblages of arc and associated tectonic affinities (Wrangellia, Alexander Terrane, and Peninsular Terrane), and parautochthonous assemblages of the ancestral North American terranes (Fig. 1a; Colpron et al., 2007). The present-day Denali fault system consists of western, central, and eastern segments (Haeussler et al., 2017). The eastern segment comprises the EDFZ and Totshunda faults (e.g., Haussler et al., 2017). Sparse piercing points suggest ~400 km of post-Eocene, dextral strike-slip separation along the eastern and central segments Denali fault and possibly as much as ~5 km fault-perpendicular shortening (Eisbacher, 1976; Waldien et al., 2018). The EDFZ segment hosts the largest recorded right-lateral separation, but presently hosts the lowest slip rates anywhere along the Denali fault system, suggesting variable slip magnitude, distribution, rate, and duration of the EDFZ through geologic time (Eisbacher, 1976; Haeussler et al., 2017). Oligocene-present transpression and rapid exhumation is widely documented along the Denali fault and inferred to be a far-field response to flat-slab subduction and subsequent collision of the Yakutat microplate (Benowitz et al., 2011; Enkelmann et al., 2017; McDermott et al., 2019).
Figure 1: Geology of the Kluane Ranges and eastern Denali fault zone. (a) Simplified terrane map of southwestern Yukon and south-central Alaska (after Colpron et al., 2007). Thick black lines are dextral intracontinental strike-slip faults. Red star denotes study area. Terrane abbreviations (south to north): Ya-Yakutat, Ch-Chugach, In-Insular, ulnt-undifferentiated Intermontane and overlap, uNA-undifferentiated North America. Dashed-gray line is subducted extent of Ya microplate (b) Simplified geologic map of the Kluane Ranges (Colpron et al., 2016) with location of hematite-coated fault surfaces and previously published conventional bedrock thermochronometry (Spotila and Berger, 2010; Enkelmann et al., 2017; McDermott et al., 2019). He-(U-Th)/He, FT-fission-track. (e) Outcrop photograph (looking NW) of hematite-rich, subsidiary fault zone. Red arrows denote hematite fault surfaces. (d)-(e) Field and hand sample photographs of hematite-coated fault surface (looking SW in d). White dashed lines show striae.

2.2 Geology and thermotectonics of the Kluane Ranges

The Kluane Ranges are part of the broader St. Elias mountains and the easternmost extent of high topography associated with the Denali fault system. Bedrock of the Kluane Ranges consists of Paleozoic and Mesozoic volcanics, volcaniclastics, and carbonate
strata of Wrangellia and Alexander terranes juxtaposed against Triassic-Eocene sediments and plutonic rocks of the Intermontane Terranes (Fig. 1b; Colpron et al., 2016). Jurassic-Late Cretaceous metasedimentary and flysch overlap assemblages, plutons of the Late Cretaceous Chisana arc record, and Late Cretaceous thrusting along the Duke River fault (DRF) document accretion of the Insular and Intermontane terranes (Cobbett et al., 2016; Eisbacher, 1976). Wrangellia bedrock is unconformably overlain by Eocene-Oligocene clastic rocks deposited in isolated, kilometer-scale strike-slip basins and Neogene volcanic rocks of the Wrangell Arc (Fig. 1b; Ridgway, et al., 1992). Wrangellian bedrock, documented Neogene-present contraction on the DRF, and both dip-slip and strike-slip small faults associated with the EDFZ collectively record transpression along the EDFZ since ~30 Ma (Caine et al., 2015; Cobbett et al., 2016; Ridgway et al., 1992).

Previous apatite (U-Th)/He (apatite He), apatite fission track (AFT), zircon (U-Th)/He (zircon He), and zircon fission track (ZFT) thermochronometry data, combined with thermal modeling and palynology-based paleoaltimetry data, document a three-stage thermotectonic and surface uplift history for the Kluane Ranges (Enkelmann et al., 2017; McDermott et al., 2019). This includes rapid cooling, exhumation, and relative surface uplift from ~95-75 Ma; slow cooling and limited exhumation from ~75-30 Ma; and renewed rapid cooling and ~1-3 km of net surface uplift (i.e., total uplift minus denudation) from ~30 Ma to present (Enkelmann et al., 2017; McDermott et al., 2019; Spotila and Berger, 2010). These phases are interpreted as corresponding to Late Cretaceous final accretion of Wrangellia to the former continental margin and possible activity along the EDFZ, relief degradation, and transpression associated with Oligocene
Yakutat flat-slab subduction (Enkelmann et al., 2017; McDermott et al., 2019). Oligocene to present surface uplift was likely a response to crustal thickening and(or) asthenospheric upwelling associated with Yakutat flat-slab subduction dynamics (McDermott et al., 2019).

2.3 Hematite (U-Th)/He thermochronometry in fault rocks

Hematite is amenable to (U-Th)/He thermochronometry because it incorporates trace amounts of U and Th into its crystal lattice, has negligible initial He upon crystallization, and retains radiogenic He over geologic timescales (e.g., Farley and Flowers, 2012). This technique has been used to date precipitation and exhumation of hydrothermal hematite in a variety of settings (Ault et al., 2016; e.g., Evenson et al., 2014; Farley and Flowers, 2012). In the hematite He system, $T_c$ ranges from ~25-250 °C, assuming a 10 °C/Ma cooling rate, depending on diffusion domain length scale (plate width; Farley and Flowers, 2012; Evenson et al., 2014; Farley, 2018). Natural hematite is commonly a polycrystalline aggregate, particularly in fault rocks, and bulk $T_c$ is thus a function of the grain size distribution (Evenson et al., 2014).

Hematite He dates from fault surfaces reflect the thermal signature of various fault-related processes such as fluid circulation (with or without new mineral precipitation), fault slip, exhumational cooling, or a combination thereof (Ault et al., 2015, 2020; McDermott et al., 2017, Moser et al., 2017). Because hematite can form and(or) recrystallize above or below its $T_c$, deciphering which of these processes a hematite He date records depends on ambient temperature conditions during hematite formation, post-formation thermal history, and effective $T_c$ of the analyzed hematite aliquot. If
precipitation occurs at depths deeper than the hematite \( T_c \) isotherm, hematite He dates will record some post-formation process such as ambient cooling due to exhumation (Ault, 2020). Alternatively, He dates record the timing of precipitation if this process occurs at depths shallower than the hematite \( T_c \) isotherm (Moser et al., 2017). Ambient cooling may be punctuated by thermal pulses associated with frictional heating during fault slip or transient heat advection by fluid flow (Ault et al., 2015, 2016; Calzolari et al., 2020; McDermott et al., 2017). These scenarios are further complicated by the potential for variable grain size distribution, and thus \( T_c \), through time associated with grain size reduction by cataclasis (Ault et al., 2015), as well as superimposed thermal and geochemical signatures from frictional or fluid related processes that lead to He loss or U gain/loss (Ault, 2020). Meaningful interpretation of hematite He data thus requires, (1) measurements of grain size distribution and calculations of the associated \( T_c \) range assuming monotonic cooling, (2) microtextural constraints on deformation mechanisms that may impact He loss, and (3) independent constraints on the ambient thermal history of hematite-bearing rocks from more conventional thermochronometers with overlapping \( T_c \) ranges (e.g., apatite and(or) zircon He data).

### 3. Samples and Methods

Oriented samples of hematite-coated fault surfaces were collected from outcrops of the EDFZ along transects parallel and perpendicular to fault strike (Figs. 1; S1\(^2\)). Hematite from fault surfaces was isolated by Dremel™ tool and fine point tweezers for microtextural analyses and hematite He thermochronometry. Microtextural analyses were

---

\(^2\) The Supplementary Material for this chapter is provided in Appendix C.
conducted on samples representative of dated material as well as surrounding wall rock using reflected light microscopy, and secondary electron (SE) and back-scatter electron (BSE) scanning electron microscopy (SEM). Energy dispersive spectroscopy (EDS) was conducted on a subset of samples to characterize sample major element chemistry and mineralogy. Grain size distributions were measured from SE images with the National Institute of Health’s ImageJ software. Sample preparation and SEM instrument operating conditions are detailed in the Supplementary Material.

The U, Th, and He concentrations of selected aliquots were measured at the Arizona Radiogenic Helium Dating Laboratory at the University of Arizona. Given the concentration of interstitial phases in these samples that may impact He retentivity (cf. Evensen et al., 2014), we pre-screened several candidate aliquots with BSE microscopy. This approach was used to identify a subset of aliquots that are pure hematite and to purposefully identify and target foliated hematite textures, as described below. Additional analytical details are provided in the Supplementary Material.

4. Results

We divide our hematite-coated fault surfaces into three groups, based on location, to facilitate discussion of microtextural observations and hematite He data (Fig. 1b). Sample groups are, from NW to SE, Williscroft (samples 61114-4A, 61714-7A, 61815-6A, 61215-2A, and 61215-5A), Alaska Highway south of Williscroft Creek (samples 61014-6A and 61914-17A), and Telluride Creek (samples 61414-2C and 61315-10E). The Williscroft group consists of relatively low (~900-940 m a.s.l.) and high elevation (~2200-2300 m a.s.l.) samples collected along Williscroft Creek and an adjacent ridge,
respectively (Fig. 1a). Williscroft low-elevation samples were collected at distances of ~0.8-1.3 km and high-elevation samples are ~3.8-4.2 km perpendicular to the mapped trace of the EDFZ. Alaska Highway samples range from ~790 to ~900 m a.s.l. in elevation and are ~0.4 to ~0.7 km from the EDFZ. Telluride Creek samples were collected at ~960 m a.s.l. and ~50 m from the fault trace.

4.1 Field and microtextural observations

Hematite-coated fault surfaces outcrop in a ≥4 km-wide zone adjacent to the mapped trace of EDFZ and are locally concentrated along Fe-rich fault zones (Fig. 1b, c; Caine et al., 2015). Low-elevation samples in all sample groups have a metallic, “burnished,” and striated appearance. High-elevation samples from Williscroft are similar, but exhibit localized high-gloss, mirror-like patches that overprint burnished hematite (Fig. S1). Hematite striae show strike-slip and dip-slip motion (Figs. 1d, e; S1; Caine et al., 2015).

Foliated and non-foliated hematite microfabrics occur in low-elevation samples (Fig. 2; S3). Regardless of fabric type, hematite grains are euhedral, high-aspect ratio plates, 28 ± 10 nm-thick (1σ std. dev; n=604) by >300 nm in length (Figs. 2c; S2; S3; S4). Foliated hematite (± clay) occurs as ~10-50 μm-thick sublayers that are collectively ~10-200 μm-wide (Fig. 2a, b, c) or as veinlets (Fig. 2e). In most of our samples, hematite exhibits SC-like geometry (Fig. 2d; Fig. S3). Packages of sigmoidally foliated hematite plates are bound by clay-rich seams, which we interpret as shear bands (Fig. 2d).

Elemental mapping and long-count spot analyses via EDS of clays with SC-like fabrics reveal elevated Al concentrations, but no K. We thus infer clays are smectite and(or) kaolin (Fig. S5). Non-foliated hematite plates are disseminated throughout ~50-300 μm-
thick quartz ± calcite ± clay cataclasite and veins or mantle comminuted grains in the cataclasite (Fig. 2a, b, e). Non-foliated, hematite-bearing cataclasite and veins are commonly cut by multiple seams of foliated hematite. However, we locally observe mutual cross-cutting relationships between these fabric types, indicating multiple episodes of hematite precipitation and slip (Figs. 2a, b, e; S3).

Microfabrics in high-elevation Williscroft samples 61215-2A and 61215-5A are similar to our low-elevation sample set, but with some exceptions (Fig. 3). Foliated hematite comprises plates (± clay) that are morphologically similar and also exhibit SC-like geometry, but with a higher abundance of subangular quartz, feldspar, and chlorite clasts (Figs. 3a, b; S3; S5). High-gloss patches visible on the fault surface in hand sample are composed of packages of tightly packed, spherical nanoparticles (~36 ± 16 nm in diameter; n=91; Fig. 3d), that overlie and cut foliated hematite. Locally the nanoparticles are observed indenting their neighbors (Fig. 3e). Elemental analyses via EDS indicate the nanoparticles are Fe-oxide (Fig. S5). In high-elevation Williscroft samples, foliated hematite seams cut chlorite + quartz ± feldspar cataclasite with a non- to weakly-foliated hematite matrix (Figs. 3a, b; S2; S3).
Figure 2: Microfabrics of low-elevation hematite-coated fault surfaces. (a) Photomicrograph of 61714-7A showing gross sample fabrics and locations of scanning electron microscopy (SEM) images. (b) SEM secondary electron (SE) image of foliated hematite plates (bounded by white-dashed lines) cutting non-foliated hematite with quartz clasts and clots of hematite + clay. White arrow denotes plucked clast impression with hematite mantle. (c) SE blow-up of foliated hematite plates in (b). (d) SE image of hematite plates with SC-like fabric. (e) SE and overlain backscatter electron (BSE) image showing foliated hematite veinlets cut by younger calcite ± quartz veins with disseminated and locally foliated hematite (white dashed lines). White box denotes location of inset showing foliated hematite detail. Bright gray tones correspond to hematite. Hem-hematite, cl-clay, qtz-quartz, cal-calcite, fol.-foliated.
Figure 3: Microfabrics of high-elevation hematite-coated fault surfaces. (a) Photomicrograph of 61215-5A. (b) SE image of foliated hematite plates. Dashed white lines denote shear bands similar to those seen in Fig. 2d. (c) SE blow-up of Domain 1 foliated hematite plates in (b). (d) SE image of tightly-packed hematite nanoparticles. (e) Blow-up SE image of hematite nanoparticles in (d). White arrows denote grains with intruding boundaries. Mineral abbreviations as in Figure 2, chl-chlorite, fsp-feldspar.
4.2 Hematite (U-Th)/He data

We acquired 38 single-aliquot hematite He dates from 11 locations on eight, individual fault surfaces (Fig. 1; Table S1). From high-elevation Williscroft sample 61215-2A and Telluride Creek sample 61315-10E, data come from distinct strike-slip (e.g., 61215-2A_1) and dip-slip (e.g., 61215-2A_2) striae sets on the same fault surface. All aliquots were isolated from foliated hematite layers only. Low U, Th, and He yields from small aliquots reflect our effort to analyze only material from such foliated hematite. This resulted in elevated analytical uncertainties on single-aliquot dates ranging from ~2% to ~31% and 1σ standard deviations of mean sample dates from ~3% to ~38%. We thus report the range of mean sample dates (± 1σ std. dev.) and single-aliquot dates (± 2σ analytical uncertainty) from each group.

Mean sample dates from Williscroft range from 3.8 ± 0.3 Ma to 6.9 ± 2.6 Ma and single aliquot dates from this sample group range from 3.4 ± 2.1 Ma to 9.8 ± 2.4 Ma. Alaska Highway mean dates span 7.0 ± 1.8 Ma to 10.5 ± 0.8 Ma, and single aliquot dates are 5.7 ± 2.6 Ma to 11.5 ± 3.2 Ma. Telluride Creek mean dates range from 5.3 ± 1.0 Ma to 6.4 ± 0.2 Ma, whereas single aliquot dates are 4.0 ± 0.5 Ma to 6.6 ± 0.2 Ma. Mean dates derived from different striae sets are indistinguishable within 1σ error for samples 61215-2A and 61315-10E. Hematite He dates are not correlated with Th/U or eU, indicating parent isotope volatilization during He degassing likely did not impact these data (Fig. S6; Hofmann et al., 2020).
5. Interpretation of hematite (U-Th)/He dates and microtextural observations

5.1 Do hematite (U-Th)/He data record ambient cooling and exhumation?

Hematite He dates potentially record different EDFZ-related processes: ambient cooling and exhumation, deformation and(or) fluid-enhanced He loss, or precipitation. To discriminate between these possibilities, we first evaluate if hematite He dates record ambient cooling. This requires estimates of the hematite He $T_c$ in our samples and comparison to conventional thermochronometry data from host rocks. We calculate the bulk $T_c$ of our samples following Dodson (1973) and assume the diffusion kinetics of Farley (2018). We apply a monotonic cooling rate of 10 °C/Myr and a spherical diffusion
domain length scale equal to the half-width of individual hematite plates (n=70 for each sample). This calculation yields a $T_c$ of $\sim 65 \pm 4 ^\circ{C}$ (mean $\pm 1\sigma$ std. dev.) across the entire dataset (Fig. 5a). $T_c$ estimates for individual samples are shown in Figure S5 and range from $\sim 62 \pm 3 ^\circ{C} (61914-17A)$ and $67 \pm 3 ^\circ{C} (61014-6A)$.

Figure 5: Regional low-temperature thermochronometry and ambient thermal history patterns. (a) Histogram of foliated hematite plate widths (top) and corresponding closure temperature estimates (bottom) for all samples. See text for calculation details. Mean values reported with $1\sigma$ std. dev. (b) Hematite, apatite, and zircon (U-Th)/He (He) and apatite, zircon fission-track (FT) mean ($\pm 1\sigma$ std. dev.) date versus elevation (after Spotila and Berger, 2010; Enkelmann et al., 2017; McDermott et al., 2019). (c) Hematite, apatite, zircon He and apatite, zircon FT date versus EDFZ-perpendicular distance. Note axis break. (d)-(f) Ambient thermal history of sample groups with hematite He dates and individual sample mean $T_c$ (see Fig. S3). Solid black lines are weighted mean paths, gray envelopes are 95% credible intervals (i.e., Bayesian confidence intervals). Red outlines in (d) denote Williscroft high-elevation samples. Thermal history inversions after McDermott et al. (2019).
Foliated hematite targeted for He dating have a $T_c$ on par with the apatite He system (Fig. 5a; Flowers et al., 2009). Previously published apatite He dates provide an independent constraint on the ambient cooling and exhumation history of bedrock overprinted by hematite fault surfaces (cf. Enkelmann et al., 2017; McDermott et al., 2019). Low-elevation hematite and apatite He dates are indistinguishable, although there is a paucity of collocated apatite He data for comparison (Figs. 1b; 5b, c). Hematite He dates from Williscroft high-elevation samples are ~77-88% younger than collocated apatite He dates (Figs. 1b; 5b). These data patterns suggest that at least high-elevation hematite He dates are too young to record ambient cooling and exhumation. Similarly, superposition of sample mean hematite He dates and $T_c$ on thermal history inversions (McDermott et al., 2019) from conventional thermochronometry data show Williscroft high-elevation samples have dates that are inconsistent with the ambient cooling history, whereas all other samples yield hematite He dates that overlap host rock time-temperature paths (Fig. 5d-f).

Statistical approaches offer an alternative means to test an ambient cooling interpretation for hematite He dates, particularly for low-elevation samples without complementary apatite He data. All hematite aliquots have uniform $T_c$, including low elevation samples with only modest elevation differences (Fig. 5b; S5). We propose a null hypothesis that if hematite He dates record ambient cooling only, then they should belong to a single statistically significant date population. Analytical uncertainties vary within our data, but qualitative inspection of aliquots with analytical uncertainties on the lower end of what we observe suggests real inter- and intra-sample dispersion, potentially due to post-formation He loss or the presence of distinct date populations (Fig. 4a). We first
calculate the mean squared weighted deviate (MSWD) or the goodness of fit of observed data to the *weighted mean date*. For MSWD=1, data variance can be explained by analytical uncertainty. MSWD >1 implies underestimated analytical uncertainties *or* data dispersion due to factors beyond measurement error; MSWD <1 suggests overestimated analytical uncertainties. We also use radial plots as an alternative means to evaluate dispersion. This approach is specifically designed for data with variable precision, and does not require assuming data and errors are parametric (Galbraith, 1990). Radial plot software (IsoplotR; Vermeesch, 2018) also determines the MSWD and degree of dispersion from the *central date* and observed dispersion.

MSWDs and dispersion across the hematite He dataset and within individual sample groups indicate overdispersed data (Fig. 4a, Table S2). The MSWD for the weighted mean and central dates across the entire dataset are 33.8 and 28.0, respectively, with dispersion calculated from the radial plot at 26.2% (Fig. 4). For sample groups, MSWDs from the weighted mean dates and central dates, as well as dispersion (%), respectively, are 13.6, 37.0, and 26.8% for Williscroft high-elevation samples; 4.2, 3.5, and 27.0% for Williscroft low-elevation samples; 5.7, 5.4, and 6.5% for Alaska Highway samples; and 20.5, 17.0, and 12.8% for Telluride creek samples (Table S2).

We apply finite mixture modeling using IsoplotR (Vermeesch, 2018) to define potential discrete date populations within our hematite He data (solid lines in Figure 4b). This approach only requires specifying how many populations are present, with the actual date of each population determined by the algorithm. Allowing three populations fit the data substantially better than two, but addition of a fourth population negligibly impacts misfit. Resultant hematite He date populations are 4.2 ± 0.1 Ma, 6.2 ± 0.1 Ma, and 7.7 ±
0.1 Ma (Fig. 4b). Date peaks at ~6 and 4 Ma are present in all low-elevation sample groups and also observed dates in the Williscroft high-elevation sample group, where hematite He dates do not record ambient cooling (Fig. 4).

We suggest overdispersion is best explained by the presence of multiple date populations within and among samples, rather than miscalculated analytical uncertainties for our data. High MSWDs and overdispersion persist in sample groups where all aliquots have relatively high precision (e.g., Williscroft high-elevation and Telluride Creek). We thus infer ambient cooling and exhumation cannot solely explain hematite He dates, even for low-elevation samples. Hematite He dates populations may record hydrothermal alteration and/or fault slip-enhanced He loss at depths shallower than the hematite He $T_c$ isotherm for these samples (~2 km, assuming a 30 °C/km geothermal gradient; McDermott et al., 2019).

5.2 Textural observations inform small fault deformation

Hematite He dates are isolated from layers of foliated hematite, which we typically observe cutting, and thus post-dating, non-foliated hematite (Figs. 2a, b, e; 3a, b). Multiple seams of foliated hematite in a single sample (Figs. 2a; 3a) and cross-cutting packages of hematite (Figs. 2e; 3b) imply multiple precipitation events. Earlier hematite plates may also be incorporated into younger cross-cutting hematite veinlets.

Foliated hematite plates show evidence for post-precipitation alignment with slip likely distributed between individual plates and along anastomosing clay-rich slip surfaces (Figs. 2c, d; 3c; S3). We also do not observe evidence for progressive strain localization, such as gradients in plate rotation towards a single slip surface within
foliated hematite layers. Rather, slip is distributed along multiple presumably simultaneously active slip surfaces and along plate boundaries (Figs. 2b, c; S3). These observations are suggestive of hematite deformation by distributed granular flow (Borradaile, 1981).

Granular flow mechanisms are operative at strain rates ranging from subseismic to seismic (e.g., Rowe and Griffith, 2015). To the best of our knowledge, hematite deformation experiments conducted at relevant slip rates, ambient temperatures, and comparable starting material do not exist (cf. Calzolari et al., 2020). SC-like and foliated fabrics in crystallographically similar phyllosilicate gouges are commonly interpreted as representing deformation at subseismic slip rates (e.g., Rowe and Griffith, 2015 and references therein). Recent experimental work on clay-rich gouges have documented nano-scale foliations can also develop seismic slip rates when slip zones are ≤150 μm (Aretusini et al., 2019). These experimental foliations are reminiscent of foliated hematite and clay textures exhibited in our samples. Other experimental work has demonstrated that, although deformation in <200 μm-thick zones is a prerequisite for seismic slip, it is not uniquely linked to fast slip rates (Ikari, 2015). We observe little evidence for intraplate deformation and comminution in our samples, which typically accompanies granular flow mechanisms at seismic slip rates (Figs. 2c, d, 3c; S3; Aretusini et al., 2019; Calzolari et al., 2020). We thus infer that foliated hematite experienced deformation at aseismic and(or) subseismic slip rates with likely negligible friction-generated heat.

Fe-oxide spherical nanoparticles in Williscroft high-elevation samples resemble those documented on other hematite fault surfaces inferred to have formed by sintering during seismic slip (Fig. 3e-g; McDermott et al., 2017; Calzolari et al., 2020). Sintered
nanoparticles formed during frictional heating can exhibit decreasing grain size with distance from the slip surface, reflecting high thermal gradients perpendicular to the fault surfaces characteristic of seismic slip (Ault et al., 2019). Nanoparticles do not exhibit this trend and form a sharp boundary with underlying foliated hematite (Figs. 3d-e; S3). Similar hematite crystal morphologies also form by precipitation under a range of chemical and thermal conditions, however, and are thus not uniquely linked to frictional heating (e.g., Sinha et al., 2015). We cannot rule out the potential that nanoparticles reflect sintering associated with seismic slip but suggest this texture more likely reflects new hematite precipitation under different chemical and thermal conditions.

We suggest our hematite He dates are most likely linked to microfabrics that experienced aseismic or subseismic slip, given the targeted sampling of exclusively foliated hematite (and not cataclasite or high-gloss zones comprising nanoparticles). Multiple generations of foliated hematite may be incorporated within a single hematite aliquot for (U-Th)/He analysis. Microtextural observations also reveal potentially heterogenous deformation processes over small spatial scales in our samples. Below we consider the potential impact of variable slip rates and the associated thermal evolution associated with multiple precipitation events on our hematite He data.

5.3 Assessment of He loss due to deformation and fluid interactions

EDFZ hematite fault surface microtextures provide evidence for episodic hematite precipitation and potentially variable strain rates within small volumes of fault rock. Here we evaluate the potential influence of non-monotonic thermal histories on our hematite He dates via processes such as frictional heating or fluid-mediated heat advection
associated with hematite precipitation. To evaluate the impact of these processes on our hematite He dates, we consider a range of non-monotonic time-temperature conditions that could induce (partial) resetting of hematite He dates and compare these thermal histories to microtextural observations and thermochronometric data patterns. Figure 6 shows contoured fractional He loss in hematite for square pulses of constant temperature over some duration (i.e., in Pseudo-Arrhenius space; Reiners, 2009). We assume a hematite plate half-width of 14 nm, consistent with our microtextural observations (Fig. 5a), and diffusion kinetics of Farley (2018). For comparison, we plot contours of 1% and 99% fractional loss for apatite He, assuming an average apatite equivalent spherical radius of 40 μm (see McDermott et al., 2019) and Farley (2000) diffusion kinetics.

Figure 6: Pseudo-Arrhenius plot for nonmonotonic cooling histories. Contours of 1%, 10%, 50%, and 99% fractional loss calculated for hematite (U-Th)/He (He; black, solid) and apatite He (gray, dashed) corresponding to square pulses of constant temperature and duration. Assumed diffusion kinetics are Farley (2018) for hematite He and Farley (2000) for apatite He. Diffusion domain lengthscale (a) is 14 nm for hematite He (Fig. 5a) and 40 μm for apatite He (McDermott et al., 2019). Vertical blue bar denotes assumed fluid temperatures. Horizontal blue and purple bars are durations of heating for hematite.
precipitation and sustained regional-scale geothermal perturbation, respectively. Horizontal red bar is inferred duration of frictional heating.

He loss induced by hydrothermal fluids may occur over multiple spatiotemporal scales, such as through localized thermal pulses associated with new hematite precipitation or regional-scale protracted geothermal anomalies. Fluid temperatures experienced by our samples are difficult to constrain. However, smectite and(or) kaolin is observed with hematite, and thus we infer fluid temperatures on the order of ~50-200 °C (see vertical blue bar in Figure 6; Figs. 2c, d; 3c; S5; Beaufort et al., 1995). Laboratory experiments suggest hematite precipitates at rates of 0.015-1.7 μg Fe/ml/day for fluid [Fe] ranging from 1-1000 ppm (Skulan et al., 2002). Assuming a fluid-saturated crack and typical hematite density of 5.15 g/cm³, precipitation of a 20 to 50 μm-thick hematite vein would take between ~5 s to 9 min and ~12 s to 23 min, respectively, after which the thermal pulse would quickly decay. For these durations and our inferred fluid temperatures, resetting of pre-existing hematite is ≤10% (see horizontal and vertical blue bar intersection in Figure 6). In contrast, sustained geothermal gradient anomalies over 100s of m and 10³-10⁶ Myr (e.g., Ault et al., 2016; Louis et al., 2019) that result from fluid-driven heat advection suggest resetting of hematite He dates is possible (see horizontal purple bar in Figure 6). But, these conditions also predict partial to complete resetting of collocated apatite He thermochronometry data, which we do not observe (e.g., compare Williscroft high-elevation sample hematite He data with nearby apatite He dates; Figs. 1a; 5b, c; see apatite He contours in Fig. 6).

Frictional heating is related to the fast slip rates characteristic of earthquakes and occurs over second-long timescales (Rowe and Griffith, 2015). Although we suspect our
samples deformed largely by aseismic or subseismic slip, we evaluate the temperatures required to induce He loss in our samples if frictional heating affected our samples. For an assumed characteristic timescale of 1 s for a seismic slip thermal pulse, temperatures of ~350, ~430, and ~490 °C are required for 10, 50 and 99% fractional He loss from hematite, respectively (red bar in Figure 6). Dry heating and deformation experiments on hematite conducted at seismic slip rates suggest temperatures ≥300 °C induce sintering and annealing, which produce rounded crystal morphologies (Calzolari et al., 2020; McDermott et al., 2017). We do not observe these textures in our dated aliquots, but rather high-aspect ratio platelets (Figs. 2c; 3c). If these slip surfaces did accommodate seismic slip, temperatures <300 °C and negligible He loss occurred (Figs. 2; 3a-d).

5.4 Summary of hematite slip surface (U-Th)/He date and microtextural interpretations

Hematite He dates most likely record hematite precipitation rather than exhumation or other post-precipitation thermal processes. Microtextural observations reveal multiple episodes of hematite precipitation and IsoplotR mixture modeling of thermochronometry data suggests date populations of ~8, ~6, and ~4 Ma. Hematite foliations and SC-like geometries likely reflect reactivation and deformation of original hematite at aseismic and(or) subseismic slip rates (Figs. 2f, g; 3e-g). We infer that any subsequent episodes of hydrothermal precipitation and(or) deformation were not associated with temperatures high enough to reset hematite He dates of earlier hematite generations. Because the dimensions of dated materials exceed the spatial scale over which discrete slip surface sets occur, bulk hematite He dates and associated scatter may thus reasonably reflect mixtures between multiple age populations present within a single slip zone or “mixed
ages”. We suggest microtextural and hematite He data patterns collectively capture the timing of hydrothermal alteration and repeated reactivation by coupled hydrothermal-strain events at ~1 Ma timescales. Additionally, samples with overprinting striae yielding similar dates (e.g., 61215-2A and 61315-10E; Fig. 4) suggest discrete precipitation events beyond the resolution of hematite He data.

6. Connecting fault-specific deformation with regional strain

Hematite He dates capture a window of Miocene to Pliocene hydrothermal alteration and deformation within the EDFZ (Figs. 4; 7). Date populations of ~8, ~6, and ~4 Ma are approximately consistent over km’s along-strike and up to 4 km’s perpendicular to the main trace of the EDFZ. The ~8 Ma hematite He date population is observed in Williscroft low-elevation and Alaska Highway groups. The ~6 Ma date population is observed in all sample groups, and the ~4 Ma date population is observed in Telluride Creek and both Williscroft sample groups (Fig. 4).

McDermott et al. (2019) calculated 30 Ma–present exhumation rates from conventional low-temperature thermochronometry of 0.3 ± 0.06, 0.07 ± 0.04, 0.2 ± 0.04, and 0.3 ± 0.09 km/Myr for locations corresponding to our Williscroft low- and high-elevation, Alaska Highway, and Telluride Creek sample groups, respectively. Assuming the above exhumation rates, hematite precipitation depths are estimated at 2.3 ± 0.5 km to 1.3 ± 0.3 km (errors propagated in quadrature) for Williscroft low-elevation samples, 0.4 ± 0.2 km to 0.3 ± 0.2 km for Williscroft high-elevation samples, 1.5 ± 0.3 km to 1.2 ± 0.2 km for Alaska Highway samples, and 1.9 ± 0.6 to 1.3 ± 0.4 km for Telluride Creek samples. Hematite yielding ~6 and ~ 4 Ma, versus ~8 Ma, dates was likely at paleodepths
shallower than, or approaching, the hematite He \( T_c \), respectively. This provides further support for most hematite He dates recording hematite precipitation, although \(~8\) Ma may in fact record ambient cooling.

**Figure 7: Schematic of eastern Denali fault zone processes and regional surface uplift.** Volume of hematite fault surface-bearing, altered, and deformed crust shown (transparent purple box). Black arrows denote crustal thickening, net surface uplift and differential exhumation after McDermott et al. (2019). Arrow lengths are schematic but proportional to magnitude. Numbered insets show progressive evolution of hematite fault surfaces informed by microtextural observations. Numbers correspond to those in block diagram. Colors correspond to microfabric types (foliated-brown, nonfoliated-green). Additional fabric elements shown schematically. Solid brown lines-hematite plates, gray lines-foliation and shear surfaces, white circles-comminuted wall-rock clasts.

Our data likely track regionally consistent episodes of hydrothermal alteration and deformation at depths of \(~\geq 2\) to 0.5 km. Enhanced differential exhumation \(\leq 2\) km perpendicular to the EDFZ and exposed paleodepths result in preserving different snapshots of a crustal volume with similar alteration and deformation characteristics (Fig.
Alteration and deformation concomitant with exhumation cause overprinting deformation in this volume, observed both at the intrasample and regional scales (Fig. 7).

Hydrothermal alteration and deformation within the EDFZ are broadly coincident with regional net surface uplift of the Kluane Ranges (Fig. 7; Enkelmann et al., 2017; McDermott et al., 2019). Exhumation calculations from conventional low-temperature thermochronometry and paleoaltitude data reveal regional post-30 Ma crustal thickening and net surface uplift throughout the Kluane Ranges (McDermott et al., 2019). Inferred hematite precipitation-deformation at ~8 Ma, ~6 Ma, and ~4 Ma occurs in the latter third of this ~30 Myr time interval (Figs. 4; 5d-f). Although the sampled hematite fault surfaces individually represent small strains, some fault scaling studies suggest the cumulative contribution of small faults may account for a large fraction of total strain (e.g., Marrett and Allmendinger, 1991), particularly given the substantive volume of bedrock where we have observed pervasive small faults coated by hematite (cf. Caine et al., 2015). We suggest our data reveal coupling among regional surface uplift and hydrothermal alteration, in part accommodated by variably aseismic to subseismic small strains (Fig. 7).

7. Implications for regional late Miocene tectonics

Oligocene to present exhumation and net surface uplift in the Kluane Ranges is interpreted as a far-field response to Yakutat flat slab subduction and collision at ~30 and 15 Ma, respectively (Enkelmann et al., 2017; McDermott et al., 2019). Hematite He date populations of ~8 Ma, ~6 Ma, and ~4 Ma may thus reflect a late-stage component of EDFZ activity as a far-field response to Yakutat microplate dynamics. Approximately 6
Ma and ~4 Ma populations documented within our study area temporally overlap with regional-scale, rapid exhumation and deformation recorded elsewhere within the Denali fault system, St. Elias Mountains, and Alaska Range. This includes ~6 Ma exhumation in the Denali Massif and environs (Fitzgerald et al., 1995), enhanced exhumation in the eastern Alaska Range (Benowitz et al., 2011; Waldien et al., 2018), a post-6 Ma increase in slip rate along the Totshunda fault (Waldien et al., 2018), southward migration of the St. Elias syntaxis at 6-4 Ma (Enkelmann et al., 2017, 2008), and post ~5 and ~2 Ma rapid exhumation along the Fairweather fault system (McAleer et al., 2009). Past work has attributed synchronous regional exhumation and deformation as related to a change in Pacific-North America plate convergence to more oblique angles at ~5 Ma (Fitzgerald et al., 1995) and(or) ~5 Ma migration of an unstable Pacific-Yakutat-North America triple junction (Enkelmann et al., 2008; Gulick et al., 2013). Although we cannot discriminate between these potential mechanisms using our data, the temporal coincidence is consistent with hematite-coated fault surfaces exposed along the EDFZ recording fault zone deformation as a response to far-field plate boundary processes (cf. Caine et al., 2015; McDermott et al., 2019). Because recent exhumation has been insufficient to reset more conventional thermochronometers, hematite He data yield additional insight to the regional extent and style of late Miocene deformation.

8. Conclusions

Combined microtextural observations and hematite He thermochronometry of hematite-coated fault surfaces constrain the timing of hydrothermal alteration and deformation within the EDFZ. Multiple generations of hematite precipitation occurred ~8
Ma, ~6 Ma, and ~4 Ma, followed by reactivation, deformation, and development of foliation that we interpret to have occurred at aseismic or subseismic slip rates. Straightforward interpretation of hematite He thermochronometry data is challenging in this setting due to synchronous, potentially temperature-sensitive processes including slip, alteration, and exhumation, as well as large analytical uncertainties associated with small aliquots from targeting specific textures. Despite these complexities, our workflow reasonably isolates the timing of fault-specific processes from ambient exhumation. We show Miocene and Pliocene hematite He dates are inconsistent with recording ambient cooling during exhumation; fault surfaces do not exhibit microtextural evidence of the temperatures required for substantial He loss from these samples by frictional heating; and that likely fluid temperatures present during hematite precipitation and deformation did not induce He loss from earlier hematite precipitation events. Because aliquots used for hematite He analysis are larger than the spatial dimensions over which discrete hematite generations occur, some dates are likely “mixed ages” and this is reflected in intrasample dispersion.

Hematite-coated fault surfaces record aseismic and(or) subseismic strain at depths of ~≥2 km to 0.5 km, providing new constraints on deformation mechanisms within the shallow EDFZ. Our data capture hydrothermal alteration and deformation broadly concomitant with previously documented post-30 Ma regional exhumation and surface uplift (Enkelmann et al., 2017; McDermott et al., 2019). Documented Miocene and Pliocene date populations also overlap with phases of deformation previously documented within the Denali fault system and Northern Cordillera (Fitzgerald et al., 1995; Enkelmann et al., 2008; Benowitz et al., 2011; Waldien et al., 2018) and
temporally coincident with changes in plate convergence angle and(or) plate reorganization (Fitzgerald et al., 1995; Gulick et al., 2013). This phase of deformation is not recorded by conventional thermochronometers in the Kluane Ranges due to limited exhumation. New hematite He data thus refine the structural history of the EDFZ, with implications for the spatial extent and timescales over which shallow plate boundary and deformation processes are transmitted to continental interiors.

REFERENCES


continental-scale strike-slip fault: The Denali fault of the eastern Alaska Range. Geosphere 7, 455–467. https://doi.org/10.1130/GES00589.1


Caine, J.S., Israel, S., Murphy, D.C., Benowitz, J., 2015. A fault rock record of late Eocene strain partitioning along the Denali fault in southwestern Yukon. Presented at the Geological Society of America Cordilleran Section Meeting, Anchorage, AK.


CHAPTER 7
CONCLUSIONS

In this chapter, I tie the principal contributions of the chapters comprising this dissertation to larger themes in tectonics and structural geology and research goals outlined in Chapter 1. Case studies centered on hematite (U-Th)/He (He) thermochronometry offer broad insight into the application of this method and other thermochronometers to fault-related rocks. Complementary microtextural observations, in conjunction with depth and timing constraints from thermochronometry and numerical models, yield additional insight to the mechanical structure of fault zones. The contributions herein offer perspectives on the different spatiotemporal scales, from individual crystals on discrete slip surfaces to mountain ranges and from seconds to millions of years, over which fault zones evolve.

IMPLICATIONS FOR THERMOCHRONOMETRY OF FAULT-RELATED ROCKS

In three of the presented chapters, I develop and apply a multi-pronged workflow for isolating the thermochronometric signature of different fault zone thermal processes. Chapter 2 advocates a novel approach of pre-screening hematite aliquots via microtextural and grain size characterization prior to He analysis. This approach proves important for interpretation of (1) hematite He data with intrasample dispersion, (2) disparate dates from locations on the same slip surface, and (3) microtextural observations supporting $^4$He loss by friction-generated heat. Numerical modeling
supports the hypothesis that some hematite He dates from Wasatch fault zone (WFZ) fault mirrors record deformation-enhanced $^4$He loss. In Chapter 3, an expanded hematite He and microtextural dataset from different types of hematite occurrence in the WFZ reflect mineral precipitation, exhumation (and burial), and(or) deformation-enhanced He loss from fault slip. To address these complexities, I develop and apply a novel iterative numerical modeling approach that varies mineralization age and ambient thermal history to identify episodic hematite mineralization and characterize host rock thermal history. Models reveal that some hematite may be >1 Ga older than the oldest hematite He date for the sample.

In Chapter 6, straightforward interpretation of hematite He dates from the eastern Denali fault zone (EDFZ) is impeded by intrasample dispersion, large (~40 %) analytical uncertainties in some samples, a lack of independent thermochronometry data with complementary closure temperature, and ambiguous microtextures. Despite these complexities, statistical tools and hypothetical He loss calculations effectively rule out exhumation-, deformation-, or fluid-related He loss scenarios for these samples. Although this dissertation exclusively focuses on hematite He thermochronometry, many of these interpretive techniques could be applied to any fault rock thermochronometers.

IMPLICATIONS FOR THE SPECTRUM OF SLIP BEHAVIORS IN THE SHALLOW CRUST

Hematite He data and combined microtextural observations reveal variable deformation mechanisms in the shallow crust. A significant outcome of Chapter 2 is the use of hematite He thermochronometry as a paleotemperature proxy. Numerical models,
microtextures, and He loss implied by hematite He-apatite He date patterns suggest
coseismic slip at temperatures \( \geq 1200 \, ^\circ\text{C} \). For the range of observed displacements on
these fault surfaces, these temperatures can only be achieved by asperity flash heating
and associated weakening. Hematite He thermochronometry thus directly constrains a
dynamic weakening process leading to seismic slip and shows this process is active at \(<2
\) km depth. In Chapter 4, I identify a broader range of dynamic weakening mechanisms
operative on WFZ fault mirrors that accompany coseismic strain localization on
individual slip surfaces. These include, in addition to asperity flash heating, fluidization
(by either thermal pressurization or acoustic fluidization) and nanoparticle lubrication.
Importantly, microtextures, and thus inferred temperatures and dynamic weakening, vary
over 100s of \( \mu\text{m} \) during a single slip event. Thermochronometric constraints (Chapter 3)
suggest this suite of fault mirrors potentially slipped at greater depths (\(<\sim 4 \) to 5 \( \) km)
relative to slip events documented with thermochronometry in Chapter 2. Thin hematite-
coated fault surfaces from the EDFZ (Chapter 6) similarly reflect strain localization and
were operative at comparable depths (\(<\sim 2 \) km), but instead have microtextural evidence
for aseismic or subseismic slip. The reason for the difference in slip rates and associated
deformation mechanisms, despite similar depths, are presently unclear and could be
caused by a range of factors, including initial grain size, elastic properties of the host
rock, and(or) fluid content. Regardless of cause, these comparisons demonstrate shallow
crustal deformation is heterogenous, even in similar fault rock lithologies.
IMPLICATIONS FOR FAULT ZONE EVOLUTION ACROSS DIFFERENT SPATIOTEMPORAL SCALES

The combination of hematite He thermochronometry with more conventional low-temperature thermochronometers provides perspective on fault zone evolution over different timescales and within the context of regional tectonics. In The WFZ, fault mirrors capture Miocene-present coseismic slip, temperature rise, and dynamic weakening and are an expression of the modern extensional tectonic regime (chapters 2 and 3). However, hematite precipitation was episodic over 100s of Myr, with events correlated to exhumation following Proterozoic metamorphism, regional thrusting during the Cretaceous Sevier orogeny, and Miocene-present extension. Specularite veins and breccia served to localize future deformation and facilitate fault mirror development. The polyphase deformation history of rocks in the present day WFZ influenced younger deformation via strain localization and fault reactivation in networks of discrete slip surfaces (chapters 3 and 4).

Thermochronometric results from the Kluane Ranges, together with regional geologic constraints, inform exhumation and surface uplift linked to the evolution of the EDFZ (Chapter 5). A three-phase history over 100 Ma is linked to both near- and far-field plate boundary processes (i.e., terrane accretion and outboard flat-slab subduction). The present-day elevation was attained over the last ~30 Ma, reflecting crustal thickening and transpression. Hematite He and microtextural results from the EDFZ directly date fault damage associated with the later part of this phase of surface uplift and capture a component of aseismic to subseismic deformation in the EDFZ. These data provide direct connections between surface uplift, the fault networks that accommodate components of
topographic change, and regional geodynamics. The combination of conventional and novel thermochronometers provide an unprecedented window into related shallow deformation on individual slip surfaces and tectonism at the scale of the mountain range.
Appendices
Supporting Information for Chapter 3

ANALYTICAL METHODS

Hematite, apatite, and zircon He, U, Th, and Sm measurement procedures

Polycrystalline chips of hematite were extracted from specularite veins and fault mirrors via Dremel™ tool, broken into replicate aliquots, and loaded in Nb packets. Apatite and zircon were isolated from bedrock via standard magnetic and density separation techniques. Separates of apatite and zircon were examined under microscope and single-grain aliquots chosen based on morphology, clarity, and lack of inclusions. Grains had their dimensions measured and were subsequently loaded into Nb packets. Aliquots were laser-heated to ~900 °C for three and 10 minutes for apatite and hematite, respectively, and ~900-1300 °C for 15 minutes for zircon, with a diode laser. For hematite and zircon only, we conducted re-extracts at higher temperature and longer duration until ⁴He yield changed by <1%. Liberated He gas was spiked with ³He, purified with cryogenic and gettering methods, and analyzed via quadruple mass spectrometer. Degassed aliquots were dissolved in HNO₃ (apatite) and HF (hematite, zircon) acid and spiked with a ²³³U-²²⁹Th-¹⁴⁷Nd-⁴²Ca tracer. Extracts were analyzed for ²³⁸U, ²³²Th, and ¹⁴⁷Sm (for apatite) on an Element 2 Inductively Coupled Plasma Mass Spectrometer (ICP-MS). We used Durango apatite, Fish Canyon tuff zircon, and WF94-17 hematite (Ault et al., 2015) as standards to monitor machine performance. For hematite aliquots, we estimated aliquot mass by weighing Nb tubes before and after loading. Mass
uncertainties were calculated by propagating the uncertainties associated with repeat measurements of each Nb packet and the manufacturer-reported microbalance sensitivity. In apatite and zircon aliquots, aliquot mass was calculated by [Ca] and [Zr], respectively, and stoichiometry (Guenthner et al., 2016). Blank-corrected (U-Th ±Sm)/He ages were calculated with propagated analytical uncertainties. For apatite and zircon, we applied the $\alpha$-ejection correction of Farley (1996) and Hourigan et al. (2005), respectively, with grain measurements and assuming grains were unzoned with respect to U, Th, and Sm.

**Apatite fission-track analyses**

Apatite grains were mounted in epoxy, polished for fission-track counting. Spontaneous fission tracks were revealed by etching with 5.5M HNO$_3$ at 20-21 °C for 20 seconds (Donelick et al., 2005). $[^{238}\text{U}]$ was calculated by the external detector method of Gleadow (1981) with low-U annealed muscovite sheets. Samples were irradiated at the Oregon State University Triga Reactor with neutron fluence monitored by European Institute for Reference Materials and Measurements (IRMM) uranium-dosed glass IRMM 540R. Muscovite sheets were etched with 48% HF for 18 minutes to reveal induced fission tracks. Densities of spontaneous and induced tracks were measured on an Olympus BX61 microscope equipped with a Kinetek Stage system and at 1250x magnification. Dpar values were measured via attached drawing tube and digitizing tablet calibrated against a stage micrometer and using FTStage software written by Trevor Dumitru of Stanford University (Donelick, 1991; 1993). Central ages and 1σ uncertainties were calculated using the zeta-calibration approach (Hurford and Green, 1983; Galbraith and Laslett, 1993). An apatite IRMM 540R zeta calibration factor of
368.1±14.9 was obtained by repeat calibration against international age standards of Durango and Fish Canyon apatite.

PILOT INVERSE MODELING OF APATITE AND ZIRCON (U-TH)/HE AND APATITE FISSION-TRACK DATA

We conducted a series of inverse models with AHe, ZHe, and AFT data to inform forward models presented in the main text. Inverse models were conducted with the HeFTy software (v1.9.3; Ketcham, 2005). We used the most recently available diffusion models, namely the zircon and apatite radiation damage accumulation and annealing models (RDAAM and ZRDAAM, respectively; Flowers et al., 2009; Guenthner et al., 2013) for (U-Th)/He data. We used the fission-track annealing kinetics of Ketcham et al. (2007) for AFT data.

In our models, we considered data from sample W14-T1 (see Figs. 1;8, main text). AHe data for this sample are available in the Supplemental Material of McDermott et al. (2017) and ZHe and AFT data are presented in Tables S2 and S3. We employed the “binning” approach to create synthetic grains at specific effective uranium (eU) concentrations suitable for modeling (e.g., Ault et al., 2013; Flowers et al., 2015). The rationale for this approach includes practical reasons, as HeFTy limits models to seven inputs, and binning offers an informal way to treat uncertainty in diffusion models by averaging dates that may be affected by additional factors outside of radiation damage effects, but not included a priori (e.g. zoning; Flowers et al., 2015). AHe data have a restricted range of eU, and were thus collapsed to a single bin with date, eU, and equivalent spherical radius (R_s) that reflects the average of all grains (see Table S3 in
McDermott et al., 2017). ZHe data were binned at 200 eU increments, which for our data translated to a total of three synthetic grains at low, middle, and high eU. We used the average date, and Rs, and eU of all grains within a particular bin. For both AHe and ZHe data, we applied either the standard deviation of all grains within that bin, or 20% uncertainty (following recommendations in Ketcham et al., 2018), whichever was greater. For AFT data, the software does not allow grains with zero spontaneous tracks, common in our samples, to be modeled, and these were thus grains were thus not included in AFT input data. We assumed a single kinetic population, reflecting the presence of only one date population in our data (see main text and $\chi^2$ in Table S3), and a Dpar kinetic parameter that averaged all data with spontaneous tracks (Table S3; Ketcham et al., 2007).

Independent geological constraints were included in our inverse models to optimize the model search of time temperature space. Because radiation damage accumulation and annealing are highly dependent on thermal history (Flowers et al., 2009; Guenthner et al., 2013), we considered the entire ~1800 Ma time-temperature path of our rocks. Model initial conditions were set at temperatures of 700-600 °C, far above the temperatures at which damage in either apatite or zircon anneals, and 1800-1620 Ma, reflecting previously published garnet-biotite thermometry and zircon and monazite U-Pb data (Hedge et al., 1983; Barnett et al., 1993; Flowers et al., 2009; Guenthner et al., 2013; Ginster et al., 2019). The Paleoproterozoic-Cambrian unconformity preserved in our study area requires near surface conditions by ~520-510 Ma (see Figure 1b in main text; Petersen, 1974). Our samples lie ~2 km below the projected trace of this unconformity (see Figure 1c cross section in main text), yet this depth may also reasonably be the result
of unquantified normal faulting in the Farmington Canyon Complex (Crittenden and Soresnsen, 1985). We thus allow Cambrian temperatures to vary between 0 and 50 °C, the latter reflecting the potential for our samples to reside 2 km below the surface and assuming a 25 °C/km geothermal gradient. We apply Mesozoic time-temperature constraints derived from strata thicknesses discussed in the main text. Lastly, we assume a temperatures of 0-10 °C and 0 Ma.

We conducted a total of three thermal history inversions (Fig. S6). Each inversion tested 100,000 paths and allowed the number of time-temperature inflection points (and resultant length of each thermal history segment) between each geological constraint to be random. “Good” paths, those supported by the data have a goodness-of-fit (GOF) between modeled and observed data ≥0.5, whereas “acceptable paths”, or those that cannot be ruled out by the data are those with 0.01 ≤ GOF ≤ 0.49 (Ketcham, 2005). In our first inversion, we modeled all data with the exception of our ~21 Ma ZHe date at low eU, considering this one datapoint may be an outlier. A second inversion includes this ~21 Ma, but returns a low number of solutions due to the tighter restrictions this datum imposes on viable thermal histories, but the large solution space explored by the MC algorithm. Thus, we also ran a third inversion that includes all available data, but with the addition of a wide “optimizing” constraint from 1800-1000 Ma and 0-400 °C, reflecting the region time-temperature space preferred by our data in the second inversion (see Ketcham and Gallagher, 2018 for a discussion of this approach). This last inversion yields similar results to the second, but returns a more viable solutions (Fig. S6). Because time-temperature paths do not fill the entirety of the optimizing constraint applied, our
models were still able to explore a regional portion of time-temperature space such that
the inclusion of this box is not solely responsible for the final model outcome.
Figure S1: Photographs of hematite hand samples.

○ (U-Th)/He date sampling location

Specularite samples

W16-6

W16-21

W17-1E

W17-1F

W17-2

Fault mirrors with underlying breccia

W17-1G

W17-1H

W17-1J
**Figure S2:** Library of hematite aliquot reflected light (top row) and scanning electron microscope (SEM; bottom row) images.

**W16-6**

![W16-6](image)

**W16-21**

![W16-21](image)
Figure S2 (continued): Library of hematite aliquot reflected light (top row) and scanning electron microscope (SEM; bottom row) images.
Figure S2 (continued): Library of hematite aliquot reflected light (top row) and scanning electron microscope (SEM; bottom row) images.

W17-2

Fault mirror volume aliquots

W17-1G

Fault mirror volume aliquots
Figure S2 (continued): Library of hematite aliquot reflected light (top row) and scanning electron microscope (SEM; bottom row) images.

**W17-1H**

*Fault mirror volume aliquots*

*Breccia matrix aliquots*
**Figure S2 (continued):** Library of hematite aliquot reflected light (top row) and scanning electron microscope (SEM; bottom row) images.

**W17-1J**

*Fault mirror volume aliquots*

*Breccia matrix aliquots*
Figure S3: Plate width (left) and closure temperature (right) distributions measured from representative SEM mounts. Closure temperature calculations assume Farley (2018) diffusion kinetics, a spherical diffusion domain lengthscale equivalent to plate half-width, and a 10 °C/Myr cooling rate.

Specularite

W16-6

Plate width (μm)

W16-21

Plate width (μm)

W17-1E

Plate width (μm)

W17-1F

Plate width (μm)

W17-12

Plate width (μm)

Closure temperature (°C)

median (+/- range): 5.0 +8.9/-3.4

median (+/- range): 7.3 +17.9/-4.7

median (+/- range): 4.7 +11.4/-4.3

median (+/- range): 7.6 +15.4/-18.3

median (+/- range): 15.6 +24.7/-12.5

median (+/- range): 131.8 +16.3/-16.8

median (+/- range): 137.5 +20.5/-15.9

median (+/- range): 130.9 +19.6/-33.8

median (+/- range): 138.2 +18.3/-14.7

median (+/- range): 149.9 +16.5/-25.6
Figure S3 (continued): Plate width (left) and closure temperature (right) distributions measured from representative SEM mounts. Closure temperature calculations assume Farley (2018) diffusion kinetics, a spherical diffusion domain lengthscale equivalent to plate half-width, and a 10 °C/Myr cooling rate.

Fault mirror volume ("A" suffix) with underlying breccia ("B" suffix)
Figure S4: (U-Th)/He date versus Th/U ratio (a) and effective uranium (eU) concentration (b).
Figure S5: Photomicrographs of sample W14-T1 zircon aliquots showing variable preservation of metamictization and radiation damage.
Figure S6: Pilot inverse models of bedrock thermochronometry data.
Figure S7: Predicted zircon (U-Th)/He date-effective uranium (eU) curves for forward thermal history simulations (left). Ginster et al. (2019) radiation damage annealing kinetics (center) or Guenthner et al. (2013) kinetics, but with no damage annealing (right) considered. Tested paths are those of Figures 11 and 12a of the main text (see annotations in bold italics). Colors in time-temperature paths correspond to those in date-eU plots.
### Table S1. Hematite (U-Th)/He data

<table>
<thead>
<tr>
<th>Specularite samples</th>
<th>Mass (μg)</th>
<th>± 1σ</th>
<th>U (ng)</th>
<th>± 1σ</th>
<th>U-Th/He (ng)</th>
<th>± 1σ</th>
<th>± 1σ</th>
<th>± 1σ</th>
<th>± 1σ</th>
<th>Med. PW (μm)</th>
<th>±/b</th>
<th>Date (Ma)</th>
<th>Error (Ma)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>W16-6, 41.3951, -112.0235</strong>; 1458 masl</td>
<td>H1 68 1.4</td>
<td>0.1531</td>
<td>0.0022</td>
<td>0.2218</td>
<td>0.0033</td>
<td>27.13</td>
<td>0.29</td>
<td>3.02</td>
<td>0.10</td>
<td>0.40</td>
<td>0.01</td>
<td>7.6</td>
<td>3.4/1.8</td>
</tr>
<tr>
<td></td>
<td>H2 78 1.4</td>
<td>0.2720</td>
<td>0.00387</td>
<td>0.4334</td>
<td>0.0062</td>
<td>29.99</td>
<td>0.31</td>
<td>4.79</td>
<td>0.15</td>
<td>0.38</td>
<td>0.01</td>
<td>8.1</td>
<td>2.8/1.0</td>
</tr>
<tr>
<td></td>
<td>H3 113 1.4</td>
<td>0.2933</td>
<td>0.0042</td>
<td>0.4789</td>
<td>0.0070</td>
<td>69.81</td>
<td>0.67</td>
<td>3.59</td>
<td>0.10</td>
<td>0.62</td>
<td>0.01</td>
<td>6.9</td>
<td>6.9/1</td>
</tr>
<tr>
<td>**W16-21, 41.3786, -112.0160; 1501 masl</td>
<td>H1 87 1.4</td>
<td>0.1635</td>
<td>0.0023</td>
<td>0.0020</td>
<td>0.0001</td>
<td>114.20</td>
<td>1.06</td>
<td>1.88</td>
<td>0.04</td>
<td>1.31</td>
<td>0.02</td>
<td>6.1</td>
<td>11.5/2.2</td>
</tr>
<tr>
<td></td>
<td>H2 131 1.4</td>
<td>0.4502</td>
<td>0.0064</td>
<td>0.0042</td>
<td>0.0001</td>
<td>118.01</td>
<td>1.13</td>
<td>3.44</td>
<td>0.06</td>
<td>0.90</td>
<td>0.01</td>
<td>6.0</td>
<td>4.7/2.4</td>
</tr>
<tr>
<td></td>
<td>H3 90 1.4</td>
<td>0.2292</td>
<td>0.0033</td>
<td>0.0038</td>
<td>0.0001</td>
<td>182.31</td>
<td>1.69</td>
<td>2.56</td>
<td>0.05</td>
<td>2.03</td>
<td>0.04</td>
<td>6.0</td>
<td>20.6/1.8</td>
</tr>
<tr>
<td></td>
<td>H4 123 1.4</td>
<td>0.2648</td>
<td>0.0038</td>
<td>0.0010</td>
<td>0.0001</td>
<td>257.61</td>
<td>2.38</td>
<td>2.15</td>
<td>0.04</td>
<td>2.09</td>
<td>0.03</td>
<td>8.1</td>
<td>7.9/3.9</td>
</tr>
<tr>
<td>**W17-1E, 41.3809, -112.0204; 1554 masl</td>
<td>H1 117 1.4</td>
<td>0.4894</td>
<td>0.0071</td>
<td>0.1418</td>
<td>0.0020</td>
<td>134.18</td>
<td>0.58</td>
<td>4.47</td>
<td>0.08</td>
<td>1.15</td>
<td>0.01</td>
<td>23.5</td>
<td>10.2/8.1</td>
</tr>
<tr>
<td></td>
<td>H2 177 1.4</td>
<td>0.6786</td>
<td>0.0097</td>
<td>0.0319</td>
<td>0.0006</td>
<td>109.47</td>
<td>0.44</td>
<td>3.88</td>
<td>0.06</td>
<td>0.62</td>
<td>0.01</td>
<td>14.0</td>
<td>2.5/3.5</td>
</tr>
<tr>
<td></td>
<td>H3 179 1.4</td>
<td>0.7410</td>
<td>0.0107</td>
<td>0.0214</td>
<td>0.0004</td>
<td>85.56</td>
<td>0.35</td>
<td>4.17</td>
<td>0.07</td>
<td>0.48</td>
<td>0.00</td>
<td>15.4</td>
<td>19.2/2.6</td>
</tr>
<tr>
<td></td>
<td>H4 46 1.4</td>
<td>0.1863</td>
<td>0.0026</td>
<td>0.0137</td>
<td>0.0022</td>
<td>37.29</td>
<td>0.18</td>
<td>4.12</td>
<td>0.14</td>
<td>0.81</td>
<td>0.03</td>
<td>13.0</td>
<td>10.5/4.9</td>
</tr>
<tr>
<td>**W17-1F, 41.3809, -112.0204; 1554 masl</td>
<td>H1 131 1.4</td>
<td>0.3547</td>
<td>0.0051</td>
<td>0.048</td>
<td>0.0041</td>
<td>229.95</td>
<td>2.07</td>
<td>2.79</td>
<td>0.05</td>
<td>1.76</td>
<td>0.02</td>
<td>13.7</td>
<td>21.1/4.8</td>
</tr>
<tr>
<td></td>
<td>H2 95 1.4</td>
<td>0.2541</td>
<td>0.0037</td>
<td>0.0655</td>
<td>0.0010</td>
<td>158.06</td>
<td>1.44</td>
<td>2.84</td>
<td>0.06</td>
<td>1.66</td>
<td>0.03</td>
<td>17.9</td>
<td>9.9/4.4</td>
</tr>
<tr>
<td></td>
<td>H3 80 1.4</td>
<td>0.2496</td>
<td>0.0036</td>
<td>0.0194</td>
<td>0.0003</td>
<td>162.45</td>
<td>0.67</td>
<td>3.18</td>
<td>0.07</td>
<td>2.03</td>
<td>0.04</td>
<td>10.8</td>
<td>6.4/6.2</td>
</tr>
<tr>
<td></td>
<td>H4 111 1.4</td>
<td>0.3704</td>
<td>0.0053</td>
<td>0.0588</td>
<td>0.0009</td>
<td>571.78</td>
<td>2.16</td>
<td>3.46</td>
<td>0.06</td>
<td>5.15</td>
<td>0.07</td>
<td>11.1</td>
<td>18.1/5.7</td>
</tr>
<tr>
<td>**W17-2, 41.3808, -112.0206; 1548 masl</td>
<td>H1 20 1.4</td>
<td>0.0563</td>
<td>0.0009</td>
<td>0.0096</td>
<td>0.0002</td>
<td>5.63</td>
<td>0.05</td>
<td>2.93</td>
<td>0.21</td>
<td>0.28</td>
<td>0.02</td>
<td>10.8</td>
<td>16.2/8.5</td>
</tr>
<tr>
<td></td>
<td>H2 29 1.4</td>
<td>0.0356</td>
<td>0.0005</td>
<td>0.0114</td>
<td>0.0002</td>
<td>7.81</td>
<td>0.06</td>
<td>1.32</td>
<td>0.07</td>
<td>0.27</td>
<td>0.01</td>
<td>14.3</td>
<td>3.6/3.0</td>
</tr>
<tr>
<td></td>
<td>H3 23 1.4</td>
<td>0.0233</td>
<td>0.0003</td>
<td>0.0100</td>
<td>0.0002</td>
<td>5.65</td>
<td>0.06</td>
<td>1.11</td>
<td>0.07</td>
<td>0.25</td>
<td>0.02</td>
<td>11.4</td>
<td>9.5/1.3</td>
</tr>
<tr>
<td></td>
<td>H4 10 1.4</td>
<td>0.0051</td>
<td>0.0002</td>
<td>0.0016</td>
<td>0.0001</td>
<td>2.20</td>
<td>0.03</td>
<td>0.54</td>
<td>0.08</td>
<td>0.22</td>
<td>0.02</td>
<td>39.2</td>
<td>NA</td>
</tr>
<tr>
<td>Mass (μg) ± 1σ</td>
<td>U (ng) ± 1σ</td>
<td>Th (ng) ± 1σ</td>
<td>(^{4}\text{He} (\text{fmol}) / u \pm 1σ</td>
<td>eU (ppm) ± 1σ</td>
<td>4He (nmol/g) ± 1σ</td>
<td>Med. PW (μm)</td>
<td>+/− Date (Ma)</td>
<td>Error (Ma)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>--------------</td>
<td>-------------</td>
<td>-------------</td>
<td>---------------------------------</td>
<td>--------------</td>
<td>-----------------</td>
<td>---------------</td>
<td>-------------</td>
<td>-----------</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Fault mirror and underlying breccia samples</strong>&lt;sup&gt;a&lt;/sup&gt;</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>W17-1GA</strong>, 41.3808, -112.0204; 1546 masl</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>H1</td>
<td>129</td>
<td>1.4</td>
<td>1.6319</td>
<td>0.0235</td>
<td>0.1207</td>
<td>0.0017</td>
<td>211.86</td>
<td>0.92</td>
<td>12.87</td>
<td>0.23</td>
<td>1.64</td>
<td>0.02</td>
<td>0.8</td>
</tr>
<tr>
<td>H2</td>
<td>55</td>
<td>1.4</td>
<td>0.7327</td>
<td>0.0106</td>
<td>0.0243</td>
<td>0.0004</td>
<td>77.44</td>
<td>0.42</td>
<td>13.42</td>
<td>0.39</td>
<td>1.41</td>
<td>0.04</td>
<td>0.8</td>
</tr>
<tr>
<td>H3</td>
<td>78</td>
<td>1.4</td>
<td>0.6612</td>
<td>0.0095</td>
<td>0.0435</td>
<td>0.0006</td>
<td>153.32</td>
<td>0.65</td>
<td>8.61</td>
<td>0.20</td>
<td>1.97</td>
<td>0.04</td>
<td>0.8</td>
</tr>
<tr>
<td>H4</td>
<td>37</td>
<td>1.4</td>
<td>0.3831</td>
<td>0.0056</td>
<td>0.0248</td>
<td>0.0004</td>
<td>70.27</td>
<td>0.41</td>
<td>10.51</td>
<td>0.42</td>
<td>1.90</td>
<td>0.07</td>
<td>0.8</td>
</tr>
<tr>
<td><strong>W17-1HA</strong>, 41.3807, -112.0205; 1542 masl</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>H1</td>
<td>104</td>
<td>1.4</td>
<td>1.5115</td>
<td>0.0216</td>
<td>0.4857</td>
<td>0.0071</td>
<td>138.65</td>
<td>0.62</td>
<td>15.63</td>
<td>0.30</td>
<td>1.33</td>
<td>0.02</td>
<td>0.90</td>
</tr>
<tr>
<td>H2</td>
<td>59</td>
<td>1.4</td>
<td>0.6574</td>
<td>0.0095</td>
<td>0.2212</td>
<td>0.0032</td>
<td>51.94</td>
<td>0.28</td>
<td>12.02</td>
<td>0.33</td>
<td>0.88</td>
<td>0.02</td>
<td>0.90</td>
</tr>
<tr>
<td>H3</td>
<td>43</td>
<td>1.4</td>
<td>0.7031</td>
<td>0.0100</td>
<td>0.2311</td>
<td>0.0033</td>
<td>53.60</td>
<td>0.29</td>
<td>17.61</td>
<td>0.62</td>
<td>1.25</td>
<td>0.04</td>
<td>0.90</td>
</tr>
<tr>
<td>H4</td>
<td>37</td>
<td>1.4</td>
<td>0.6610</td>
<td>0.0094</td>
<td>0.3093</td>
<td>0.0044</td>
<td>55.67</td>
<td>0.31</td>
<td>19.83</td>
<td>0.80</td>
<td>1.50</td>
<td>0.06</td>
<td>0.90</td>
</tr>
<tr>
<td><strong>W17-1HB</strong>, 41.3807, -112.0205; 1542 masl</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>H1</td>
<td>115</td>
<td>1.4</td>
<td>0.9825</td>
<td>0.0140</td>
<td>0.0584</td>
<td>0.0009</td>
<td>371.46</td>
<td>1.25</td>
<td>8.663</td>
<td>0.16</td>
<td>3.23005</td>
<td>0.0412</td>
<td>2.0</td>
</tr>
<tr>
<td>H2</td>
<td>80</td>
<td>1.4</td>
<td>0.5552</td>
<td>0.0080</td>
<td>0.0612</td>
<td>0.0010</td>
<td>177.06</td>
<td>0.64</td>
<td>7.119</td>
<td>0.16</td>
<td>2.21321</td>
<td>0.0399</td>
<td>2.0</td>
</tr>
<tr>
<td>H3</td>
<td>98</td>
<td>1.4</td>
<td>0.8543</td>
<td>0.0121</td>
<td>0.0572</td>
<td>0.0009</td>
<td>323.09</td>
<td>2.99</td>
<td>8.855</td>
<td>0.18</td>
<td>3.29684</td>
<td>0.0565</td>
<td>2.0</td>
</tr>
<tr>
<td>H4</td>
<td>126</td>
<td>1.4</td>
<td>1.2297</td>
<td>0.0175</td>
<td>0.0981</td>
<td>0.0014</td>
<td>331.29</td>
<td>3.1</td>
<td>9.942</td>
<td>0.18</td>
<td>2.62925</td>
<td>0.0384</td>
<td>2.0</td>
</tr>
<tr>
<td><strong>W17-1JA</strong>, 41.3807, -112.0207; 1540 masl</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>H1</td>
<td>120</td>
<td>1.4</td>
<td>1.8994</td>
<td>0.02738</td>
<td>0.5043</td>
<td>0.0073</td>
<td>257.35</td>
<td>1.01</td>
<td>16.82</td>
<td>0.3</td>
<td>2.14457</td>
<td>0.0266</td>
<td>0.9</td>
</tr>
<tr>
<td>H2</td>
<td>138</td>
<td>1.4</td>
<td>1.9914</td>
<td>0.02867</td>
<td>0.6104</td>
<td>0.0088</td>
<td>189.08</td>
<td>0.9</td>
<td>15.47</td>
<td>0.27</td>
<td>1.37013</td>
<td>0.0155</td>
<td>0.9</td>
</tr>
<tr>
<td>H3</td>
<td>95</td>
<td>1.4</td>
<td>1.1258</td>
<td>0.01632</td>
<td>0.2817</td>
<td>0.0045</td>
<td>137.79</td>
<td>0.63</td>
<td>12.55</td>
<td>0.25</td>
<td>1.45046</td>
<td>0.0226</td>
<td>0.9</td>
</tr>
<tr>
<td>H4</td>
<td>84</td>
<td>1.4</td>
<td>1.1451</td>
<td>0.01629</td>
<td>0.2459</td>
<td>0.0036</td>
<td>142.14</td>
<td>0.61</td>
<td>14.32</td>
<td>0.31</td>
<td>1.69219</td>
<td>0.0294</td>
<td>0.9</td>
</tr>
<tr>
<td><strong>W17-1JB</strong>, 41.3807, -112.0207; 1540 masl</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>H1</td>
<td>71</td>
<td>1.4</td>
<td>0.7830</td>
<td>0.0112</td>
<td>0.1631</td>
<td>0.0024</td>
<td>102.02</td>
<td>0.42</td>
<td>11.57</td>
<td>0.28</td>
<td>1.44</td>
<td>0.03</td>
<td>1.2</td>
</tr>
<tr>
<td>H2</td>
<td>32</td>
<td>1.4</td>
<td>0.2912</td>
<td>0.0042</td>
<td>0.0837</td>
<td>0.0012</td>
<td>30.93</td>
<td>0.16</td>
<td>9.72</td>
<td>0.44</td>
<td>0.97</td>
<td>0.04</td>
<td>1.2</td>
</tr>
<tr>
<td>H3</td>
<td>66</td>
<td>1.4</td>
<td>0.5525</td>
<td>0.0087</td>
<td>0.2729</td>
<td>0.0040</td>
<td>72.07</td>
<td>0.22</td>
<td>9.34</td>
<td>0.25</td>
<td>1.09</td>
<td>0.02</td>
<td>1.2</td>
</tr>
</tbody>
</table>

<sup>a</sup>Median plate width. Aliquot-specific for specularite vein samples, measured from representative SEM mounts for fault mirror/breccia matrix samples

<sup>b</sup>Plate width error reported as + maximum plate width, -minimum plate width
^1 is propagated error from analytical uncertainties on U, Th, and He analyses.
^2 Latitude and longitude reported in decimal degrees; WGS84
^3 Fault mirror samples denoted "A" at end of sample ID; Breccia matrix samples denoted "B"
^4 Breccia matrix for this sample of insufficient quality for (U-Th)/He
Table S2: Apatite (U-Th-Sm)/He data

<table>
<thead>
<tr>
<th>Mass (mg)</th>
<th>Res (μm)</th>
<th>U (ppm)</th>
<th>Th (ppm)</th>
<th>eU (ppm)</th>
<th>Sm (ppm)</th>
<th>^4He (nmol/g)</th>
<th>F_t c</th>
<th>Raw Date (Ma)</th>
<th>Corr Date (Ma)</th>
<th>Error (Ma)</th>
</tr>
</thead>
<tbody>
<tr>
<td>W14-T5, 41.3741, -112.0092; 1742 masl</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>A1 0.0008</td>
<td>31.9</td>
<td>30.44</td>
<td>9.61</td>
<td>32.70</td>
<td>189.30</td>
<td>0.46</td>
<td>0.57</td>
<td>2.6</td>
<td>4.5</td>
<td>0.5</td>
</tr>
<tr>
<td>A2 e 0.0008</td>
<td>31.3</td>
<td>12.88</td>
<td>4.48</td>
<td>13.94</td>
<td>96.94</td>
<td>1.66</td>
<td>0.57</td>
<td>21.9</td>
<td>38.5</td>
<td>1.4</td>
</tr>
<tr>
<td>A3 e 0.0000</td>
<td>40.0</td>
<td>15.79</td>
<td>25.31</td>
<td>21.74</td>
<td>9.10</td>
<td>0.26</td>
<td>0.64</td>
<td>2.2</td>
<td>3.4</td>
<td>8.8</td>
</tr>
<tr>
<td>A4 0.0007</td>
<td>29.8</td>
<td>10.81</td>
<td>3.76</td>
<td>11.69</td>
<td>179.89</td>
<td>0.17</td>
<td>0.55</td>
<td>2.6</td>
<td>4.8</td>
<td>1.3</td>
</tr>
<tr>
<td>A5 0.0015</td>
<td>39.9</td>
<td>19.46</td>
<td>5.12</td>
<td>20.67</td>
<td>209.81</td>
<td>0.26</td>
<td>0.65</td>
<td>2.3</td>
<td>3.6</td>
<td>0.3</td>
</tr>
<tr>
<td>A6 0.0011</td>
<td>39.1</td>
<td>19.86</td>
<td>6.84</td>
<td>21.46</td>
<td>176.01</td>
<td>0.28</td>
<td>0.64</td>
<td>2.4</td>
<td>3.7</td>
<td>0.4</td>
</tr>
</tbody>
</table>

Mean ± std. dev.: 4.1 ± 0.6 Ma

Table S3: Zircon (U-Th)/He data

<table>
<thead>
<tr>
<th>Mass (mg)</th>
<th>Res (μm)</th>
<th>U (ppm)</th>
<th>Th (ppm)</th>
<th>eU (ppm)</th>
<th>^4He (nmol/g)</th>
<th>F_t c</th>
<th>Raw Date (Ma)</th>
<th>Corr Date (Ma)</th>
<th>Error (Ma)</th>
</tr>
</thead>
<tbody>
<tr>
<td>W14-T1, 41.3726, -112.0137; 1511 masl</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Z1 0.0027</td>
<td>39.6</td>
<td>333.54</td>
<td>81.84</td>
<td>352.77</td>
<td>29.00</td>
<td>0.70</td>
<td>15.2</td>
<td>21.8</td>
<td>0.3</td>
</tr>
<tr>
<td>Z2 0.0024</td>
<td>42.0</td>
<td>510.15</td>
<td>80.74</td>
<td>529.12</td>
<td>11.77</td>
<td>0.71</td>
<td>4.1</td>
<td>5.8</td>
<td>0.1</td>
</tr>
<tr>
<td>Z3 0.0016</td>
<td>40.8</td>
<td>920.50</td>
<td>41.65</td>
<td>930.29</td>
<td>20.36</td>
<td>0.71</td>
<td>4.1</td>
<td>5.7</td>
<td>0.1</td>
</tr>
<tr>
<td>Z4 0.0032</td>
<td>44.4</td>
<td>1015.2</td>
<td>262.14</td>
<td>1076.8</td>
<td>22.38</td>
<td>0.73</td>
<td>3.9</td>
<td>5.3</td>
<td>0.1</td>
</tr>
<tr>
<td>Z5 0.0022</td>
<td>44.8</td>
<td>988.25</td>
<td>259.37</td>
<td>1049.2</td>
<td>18.95</td>
<td>0.73</td>
<td>3.4</td>
<td>4.6</td>
<td>0.1</td>
</tr>
</tbody>
</table>

a mass calculated from Ca (apatite) and Zr (zircon) and stoichiometry [Guenthner et al., 2016]
bRes-equivalent spherical radius
cF_t-alpha ejection correction of Farley et al. [1996]
d1σ propagated error from analytical uncertainties on U, Th, Sm and He analyses
e discarded (A2=outlier; A3=high analytical uncertainty)
## Table S4: Apatite fission-track data

<table>
<thead>
<tr>
<th>Sample Number</th>
<th>W14-T1</th>
</tr>
</thead>
<tbody>
<tr>
<td>No. of Crystals</td>
<td>40</td>
</tr>
<tr>
<td>Zeta Factor ± Error</td>
<td>368.1 ± 14.9</td>
</tr>
<tr>
<td>Rho d (% Relative Error)</td>
<td>1.145E+06 ± 1.65 ± 18913</td>
</tr>
<tr>
<td>N d</td>
<td>3665</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>N s</th>
<th>N i</th>
<th>N g</th>
<th>Dpar</th>
<th>Dper</th>
<th>Rmr0</th>
<th>ρ s</th>
<th>ρ i</th>
<th>ρ s / ρ i</th>
<th>U ppm</th>
<th>Age (Ma)</th>
<th>Age error</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>54</td>
<td>49</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>0.000E+00</td>
<td>1.722E+06</td>
<td>0.0000</td>
<td>22.6</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>1</td>
<td>118</td>
<td>70</td>
<td>1.83</td>
<td>0.61</td>
<td>0.826</td>
<td>2.232E+04</td>
<td>2.634E+06</td>
<td>0.0085</td>
<td>34.5</td>
<td>1.79</td>
<td>1.79</td>
</tr>
<tr>
<td>1</td>
<td>52</td>
<td>40</td>
<td>1.54</td>
<td>0.48</td>
<td>0.844</td>
<td>3.906E+04</td>
<td>2.031E+06</td>
<td>0.0192</td>
<td>26.6</td>
<td>4.05</td>
<td>4.09</td>
</tr>
<tr>
<td>0</td>
<td>43</td>
<td>36</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>0.000E+00</td>
<td>1.866E+06</td>
<td>0.0000</td>
<td>24.4</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>0</td>
<td>29</td>
<td>40</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>0.000E+00</td>
<td>1.133E+06</td>
<td>0.0000</td>
<td>14.8</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>0</td>
<td>20</td>
<td>20</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>0.000E+00</td>
<td>1.563E+06</td>
<td>0.0000</td>
<td>20.5</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>0</td>
<td>17</td>
<td>30</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>0.000E+00</td>
<td>8.854E+05</td>
<td>0.0000</td>
<td>11.6</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>0</td>
<td>38</td>
<td>50</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>0.000E+00</td>
<td>1.188E+06</td>
<td>0.0000</td>
<td>15.6</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>0</td>
<td>22</td>
<td>28</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>0.000E+00</td>
<td>1.228E+06</td>
<td>0.0000</td>
<td>16.1</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>1</td>
<td>88</td>
<td>70</td>
<td>1.54</td>
<td>0.44</td>
<td>0.844</td>
<td>2.232E+04</td>
<td>1.964E+06</td>
<td>0.0114</td>
<td>25.7</td>
<td>2.39</td>
<td>2.41</td>
</tr>
<tr>
<td>0</td>
<td>29</td>
<td>25</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>0.000E+00</td>
<td>1.813E+06</td>
<td>0.0000</td>
<td>23.7</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>0</td>
<td>76</td>
<td>50</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>0.000E+00</td>
<td>2.375E+06</td>
<td>0.0000</td>
<td>31.1</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>1</td>
<td>54</td>
<td>36</td>
<td>1.78</td>
<td>0.40</td>
<td>0.829</td>
<td>4.340E+04</td>
<td>2.344E+06</td>
<td>0.0185</td>
<td>30.7</td>
<td>3.90</td>
<td>3.94</td>
</tr>
<tr>
<td>0</td>
<td>25</td>
<td>36</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>0.000E+00</td>
<td>1.085E+06</td>
<td>0.0000</td>
<td>14.2</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>0</td>
<td>67</td>
<td>35</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>0.000E+00</td>
<td>2.991E+06</td>
<td>0.0000</td>
<td>39.2</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>2</td>
<td>58</td>
<td>28</td>
<td>1.45</td>
<td>0.32</td>
<td>0.849</td>
<td>1.116E+05</td>
<td>3.237E+06</td>
<td>0.0345</td>
<td>42.4</td>
<td>7.26</td>
<td>5.23</td>
</tr>
<tr>
<td>2</td>
<td>47</td>
<td>25</td>
<td>1.55</td>
<td>0.31</td>
<td>0.843</td>
<td>1.250E+05</td>
<td>2.938E+06</td>
<td>0.0426</td>
<td>38.5</td>
<td>8.96</td>
<td>6.48</td>
</tr>
<tr>
<td>0</td>
<td>14</td>
<td>28</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>0.000E+00</td>
<td>7.813E+05</td>
<td>0.0000</td>
<td>10.2</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>0</td>
<td>31</td>
<td>28</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>0.000E+00</td>
<td>1.730E+06</td>
<td>0.0000</td>
<td>22.7</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>3</td>
<td>104</td>
<td>80</td>
<td>1.58</td>
<td>0.39</td>
<td>0.841</td>
<td>5.859E+04</td>
<td>2.031E+06</td>
<td>0.0288</td>
<td>26.6</td>
<td>6.08</td>
<td>3.57</td>
</tr>
<tr>
<td>1</td>
<td>59</td>
<td>49</td>
<td>1.39</td>
<td>0.45</td>
<td>0.852</td>
<td>3.189E+04</td>
<td>1.881E+06</td>
<td>0.0169</td>
<td>24.6</td>
<td>3.57</td>
<td>3.60</td>
</tr>
<tr>
<td>2</td>
<td>69</td>
<td>50</td>
<td>1.64</td>
<td>0.39</td>
<td>0.838</td>
<td>6.250E+04</td>
<td>2.156E+06</td>
<td>0.0290</td>
<td>28.2</td>
<td>6.11</td>
<td>4.39</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.00E+00</td>
<td>1.500E+06</td>
<td>0.0000</td>
</tr>
<tr>
<td>---</td>
<td>---</td>
<td>---</td>
<td>---</td>
<td>---</td>
<td>---</td>
<td>---</td>
<td>---</td>
<td>---</td>
<td>----------</td>
<td>----------</td>
<td>---------</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.000E+00</td>
<td>1.328E+06</td>
<td>0.0000</td>
</tr>
<tr>
<td>1</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.00E+00</td>
<td>2.344E+06</td>
<td>0.0000</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.000E+00</td>
<td>1.992E+06</td>
<td>0.0196</td>
</tr>
<tr>
<td>1</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.000E+00</td>
<td>1.505E+06</td>
<td>0.0000</td>
</tr>
<tr>
<td>0</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.00E+00</td>
<td>1.116E+06</td>
<td>0.0000</td>
</tr>
<tr>
<td>0</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.00E+00</td>
<td>1.190E+06</td>
<td>0.0000</td>
</tr>
<tr>
<td>1</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.000E+00</td>
<td>1.611E+06</td>
<td>0.0303</td>
</tr>
<tr>
<td>0</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.00E+00</td>
<td>2.604E+06</td>
<td>0.0000</td>
</tr>
<tr>
<td>1</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.00E+00</td>
<td>2.404E+06</td>
<td>0.0213</td>
</tr>
<tr>
<td>0</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.00E+00</td>
<td>1.250E+06</td>
<td>0.0000</td>
</tr>
<tr>
<td>0</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.00E+00</td>
<td>1.563E+06</td>
<td>0.0000</td>
</tr>
<tr>
<td>2</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.00E+00</td>
<td>1.432E+06</td>
<td>0.0364</td>
</tr>
<tr>
<td>0</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.00E+00</td>
<td>1.875E+06</td>
<td>0.0000</td>
</tr>
<tr>
<td>0</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.00E+00</td>
<td>2.051E+06</td>
<td>0.0000</td>
</tr>
<tr>
<td>0</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.00E+00</td>
<td>2.246E+06</td>
<td>0.0000</td>
</tr>
</tbody>
</table>

|   |   |   |   |   |   |   |   |   |  2.127E+04 |  1.848E+06 |   0.0115 |   24.2 |  2.42 |  0.54 |       |     |      |

| Pooled Ratio | 0.0115 | ± | 0.0026 |
| Mean Ratio | 0.0093 | ± | 0.0021 |
| Pooled Age | 2.42 | ± | 0.54 |
| Mean Crystal Age | 1.96 | ± | 0.45 |
| Binomial Age | 2.50 | + | 1.21 |
|  | "+95%"
|  | "-95%"

Central Age | 2.42 | ± | 0.54 |

Age Dispersion | 0.86 % |
Chi-squared | 28.165 | with | 39 | degrees of freedom | MSWD | 0.21 |

P (Chi-Sq) | 90.08 % |
Table S4 (continued): Apatite fission-track data

<table>
<thead>
<tr>
<th>Sample Number</th>
<th>W14-T5</th>
</tr>
</thead>
<tbody>
<tr>
<td>No. of Crystals</td>
<td>40</td>
</tr>
<tr>
<td>Zeta Factor ± Error</td>
<td>368.1 ± 14.9</td>
</tr>
<tr>
<td>Rho d (% Relative Error)</td>
<td>1.132E+06 ± 1.66</td>
</tr>
<tr>
<td>N d</td>
<td>3621</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>N s</th>
<th>N i</th>
<th>N g</th>
<th>Dpar</th>
<th>Dper</th>
<th>Rmr0</th>
<th>ρ s</th>
<th>ρ i</th>
<th>ρ s / ρ i</th>
<th>U ppm</th>
<th>Age (Ma)</th>
<th>Age error</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>12</td>
<td>50</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>0.000E+00</td>
<td>3.750E+05</td>
<td>0.0000</td>
</tr>
<tr>
<td>3</td>
<td>80</td>
<td>70</td>
<td>1.51</td>
<td>0.44</td>
<td>0.845</td>
<td>6.696E+04</td>
<td>1.786E+06</td>
<td>0.0375</td>
<td>23.7</td>
<td>7.81</td>
<td>4.60</td>
</tr>
<tr>
<td>3</td>
<td>114</td>
<td>100</td>
<td>1.75</td>
<td>0.48</td>
<td>0.831</td>
<td>4.688E+04</td>
<td>1.781E+06</td>
<td>0.0263</td>
<td>23.6</td>
<td>5.48</td>
<td>3.21</td>
</tr>
<tr>
<td>0</td>
<td>18</td>
<td>50</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>0.00E+00</td>
<td>5.625E+05</td>
<td>0.0000</td>
<td>7.5</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>1</td>
<td>46</td>
<td>70</td>
<td>1.66</td>
<td>0.46</td>
<td>0.836</td>
<td>2.232E+04</td>
<td>1.027E+06</td>
<td>0.0217</td>
<td>13.6</td>
<td>4.53</td>
<td>4.58</td>
</tr>
<tr>
<td>0</td>
<td>22</td>
<td>35</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>0.00E+00</td>
<td>9.821E+05</td>
<td>0.0000</td>
<td>13.0</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>0</td>
<td>22</td>
<td>36</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>0.00E+00</td>
<td>9.549E+05</td>
<td>0.0000</td>
<td>12.7</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>1</td>
<td>45</td>
<td>60</td>
<td>1.64</td>
<td>0.35</td>
<td>0.838</td>
<td>2.604E+04</td>
<td>1.172E+06</td>
<td>0.0222</td>
<td>15.5</td>
<td>4.63</td>
<td>4.68</td>
</tr>
<tr>
<td>0</td>
<td>31</td>
<td>50</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>0.00E+00</td>
<td>9.688E+05</td>
<td>0.0000</td>
<td>12.8</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>1</td>
<td>59</td>
<td>36</td>
<td>1.42</td>
<td>0.43</td>
<td>0.850</td>
<td>4.340E+04</td>
<td>2.561E+06</td>
<td>0.0169</td>
<td>33.9</td>
<td>3.53</td>
<td>3.56</td>
</tr>
<tr>
<td>0</td>
<td>19</td>
<td>50</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>0.00E+00</td>
<td>5.938E+05</td>
<td>0.0000</td>
<td>7.9</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>0</td>
<td>43</td>
<td>50</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>0.00E+00</td>
<td>1.344E+06</td>
<td>0.0000</td>
<td>17.8</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>1</td>
<td>37</td>
<td>50</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>3.152E+04</td>
<td>1.156E+06</td>
<td>0.0270</td>
<td>15.3</td>
<td>5.63</td>
<td>5.71</td>
</tr>
<tr>
<td>0</td>
<td>31</td>
<td>40</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>0.00E+00</td>
<td>1.211E+06</td>
<td>0.0000</td>
<td>16.0</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>0</td>
<td>26</td>
<td>49</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>0.00E+00</td>
<td>8.291E+05</td>
<td>0.0000</td>
<td>11.0</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>2</td>
<td>57</td>
<td>50</td>
<td>1.63</td>
<td>0.38</td>
<td>0.838</td>
<td>6.250E+04</td>
<td>1.781E+06</td>
<td>0.0351</td>
<td>23.6</td>
<td>7.31</td>
<td>5.27</td>
</tr>
<tr>
<td>0</td>
<td>11</td>
<td>49</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>0.00E+00</td>
<td>3.508E+05</td>
<td>0.0000</td>
<td>4.6</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>0</td>
<td>58</td>
<td>70</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>0.00E+00</td>
<td>1.295E+06</td>
<td>0.0000</td>
<td>17.2</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>2</td>
<td>58</td>
<td>40</td>
<td>1.64</td>
<td>0.50</td>
<td>0.838</td>
<td>7.813E+04</td>
<td>2.266E+06</td>
<td>0.0345</td>
<td>30.0</td>
<td>7.18</td>
<td>5.17</td>
</tr>
<tr>
<td>0</td>
<td>35</td>
<td>49</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>0.00E+00</td>
<td>1.116E+06</td>
<td>0.0000</td>
<td>14.8</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>0</td>
<td>44</td>
<td>36</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>0.00E+00</td>
<td>1.910E+06</td>
<td>0.0000</td>
<td>25.3</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>0</td>
<td>41</td>
<td>40</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>0.00E+00</td>
<td>1.602E+06</td>
<td>0.0000</td>
<td>21.2</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td></td>
<td>3</td>
<td>51</td>
<td>70</td>
<td>1.49</td>
<td>0.54</td>
<td>0.846</td>
<td>6.696E+04</td>
<td>1.138E+06</td>
<td>0.0588</td>
<td>15.1</td>
<td>12.24</td>
</tr>
<tr>
<td>---</td>
<td>----</td>
<td>----</td>
<td>------</td>
<td>------</td>
<td>------</td>
<td>-------</td>
<td>-----------</td>
<td>----------</td>
<td>--------</td>
<td>------</td>
<td>-------</td>
</tr>
<tr>
<td>0</td>
<td>78</td>
<td>60</td>
<td></td>
<td>-</td>
<td>-</td>
<td>0.000E+00</td>
<td>2.031E+06</td>
<td>0.0000</td>
<td>26.9</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>0</td>
<td>55</td>
<td>36</td>
<td></td>
<td>-</td>
<td>-</td>
<td>0.000E+00</td>
<td>2.387E+06</td>
<td>0.0000</td>
<td>31.6</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>0</td>
<td>18</td>
<td>70</td>
<td></td>
<td>-</td>
<td>-</td>
<td>0.000E+00</td>
<td>4.018E+05</td>
<td>0.0000</td>
<td>5.3</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>1</td>
<td>30</td>
<td>40</td>
<td>1.43</td>
<td>0.47</td>
<td>0.850</td>
<td>3.906E+04</td>
<td>1.172E+06</td>
<td>0.0333</td>
<td>15.5</td>
<td>6.94</td>
<td>7.06</td>
</tr>
<tr>
<td>0</td>
<td>25</td>
<td>40</td>
<td></td>
<td>-</td>
<td>-</td>
<td>0.000E+00</td>
<td>9.766E+05</td>
<td>0.0000</td>
<td>12.9</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>0</td>
<td>18</td>
<td>5</td>
<td></td>
<td>-</td>
<td>-</td>
<td>0.000E+00</td>
<td>5.625E+06</td>
<td>0.0000</td>
<td>74.5</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>1</td>
<td>54</td>
<td>49</td>
<td>1.43</td>
<td>0.47</td>
<td>0.850</td>
<td>3.189E+04</td>
<td>1.722E+06</td>
<td>0.0185</td>
<td>22.8</td>
<td>3.86</td>
<td>3.90</td>
</tr>
<tr>
<td>1</td>
<td>43</td>
<td>36</td>
<td>1.69</td>
<td>0.34</td>
<td>0.835</td>
<td>4.340E+04</td>
<td>1.866E+06</td>
<td>0.0233</td>
<td>24.7</td>
<td>4.84</td>
<td>4.90</td>
</tr>
<tr>
<td>0</td>
<td>23</td>
<td>36</td>
<td></td>
<td>-</td>
<td>-</td>
<td>0.000E+00</td>
<td>9.983E+05</td>
<td>0.0000</td>
<td>13.2</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>0</td>
<td>28</td>
<td>42</td>
<td></td>
<td>-</td>
<td>-</td>
<td>0.000E+00</td>
<td>1.042E+06</td>
<td>0.0000</td>
<td>13.8</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>0</td>
<td>27</td>
<td>48</td>
<td></td>
<td>-</td>
<td>-</td>
<td>0.000E+00</td>
<td>8.789E+05</td>
<td>0.0000</td>
<td>11.6</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>0</td>
<td>54</td>
<td>50</td>
<td></td>
<td>-</td>
<td>-</td>
<td>0.000E+00</td>
<td>1.688E+06</td>
<td>0.0000</td>
<td>22.4</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>1</td>
<td>63</td>
<td>70</td>
<td>1.47</td>
<td>0.39</td>
<td>0.848</td>
<td>2.232E+04</td>
<td>1.406E+06</td>
<td>0.0159</td>
<td>18.6</td>
<td>3.31</td>
<td>3.34</td>
</tr>
<tr>
<td>3</td>
<td>65</td>
<td>50</td>
<td>1.67</td>
<td>0.45</td>
<td>0.836</td>
<td>9.375E+04</td>
<td>2.031E+06</td>
<td>0.0462</td>
<td>26.9</td>
<td>9.61</td>
<td>5.69</td>
</tr>
<tr>
<td>0</td>
<td>48</td>
<td>50</td>
<td></td>
<td>-</td>
<td>-</td>
<td>0.000E+00</td>
<td>1.500E+06</td>
<td>0.0000</td>
<td>19.9</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>0</td>
<td>23</td>
<td>50</td>
<td></td>
<td>-</td>
<td>-</td>
<td>0.000E+00</td>
<td>7.188E+05</td>
<td>0.0000</td>
<td>9.5</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>1</td>
<td>33</td>
<td>50</td>
<td>1.75</td>
<td>0.44</td>
<td>0.831</td>
<td>3.125E+04</td>
<td>1.031E+06</td>
<td>0.0303</td>
<td>13.7</td>
<td>6.31</td>
<td>6.41</td>
</tr>
</tbody>
</table>

Pooled Ratio 0.0152 ± 0.0031
Mean Ratio 0.0112 ± 0.0026
Pooled Age 3.17 ± 0.65
Mean Crystal Age 2.33 ± 0.53
Binomial Age 3.25 + 1.44 "+95%"
- 1.21 "-95%"

Central Age 3.16 ± 0.65

Chi-squared 30.435 with 39 degrees of freedom
P (Chi-Sq) 83.51 %
MSWD 0.25

Age Dispersion 9.87 %
REFERENCES


Farley, K., R. Wolf, and L. Silver (1996), The effects of long alpha-stopping distances on (U-Th)/He ages, Geochimica et cosmochimica acta, 60(21), 4223-4229.


Guenthner, W. R., P. W. Reiners, and U. Chowdhury (2016), Isotope dilution analysis of Ca and Zr in apatite and zircon (U-Th)/He chronometry, Geochemistry, Geophysics, Geosystems.


MATlab code for processing image analysis results

%ImgAn
%for plotting results of image analysis from Image J
%input is tab-delimited text file of columns in following order:
%grain area, long-axis length, short-axis length, orientation (0-180);
%creates 3 histograms of grain size, orientation, and aspect ratio. Uses
%interactive plot to select portion of grain-size distribution to calculate
%D value from power-law fit.
clear;
clc;

%load input file. Preserve .txt extension and quotes
DAT=load('W17-1HB-3c.txt');

%give title for output figure
ID={'W17-1JA Domain 1b'};
%set plotting options
%xmin=-1.5;
%xmax=2;
%give title for output figure
%extrapolation limits for particle size distribution plot
%x1min=0;
%x1max=5;
y1max=inf;
w1=0.16;
%orientation histogram, set plotting limits for y-axis, x is fixed
y3max=inf;

%%%%%%NO EDITS BELOW THIS
LINE%%%%%%NO EDITS BELOW THIS
%plotting stuff for later
edgesAR=[w2:w2:xmax];
edgesGS=[w1:w1:xmax];

%partition out vectors with data
A=DAT(:,1);
L=DAT(:,2);
l=DAT(:,3);
O=DAT(:,4);

%cumulative area
Ac=sum(A);

%create equivalent diameter and frequency vector
eqD=sqrt(A.*(1/3.14))*2;
[eqDs,ln]=sort(eqD);
N=numel(eqD);
freq=N-1:-1:1;
freq=freq*(1/Ac);
leqD=log(eqDs);
leqD=leqD(1:N-1,1);
lfreq=log(freq);

%do test for lognormal distribution using K-S test, 95% confidence
pd=fitdist(eqD,'lognormal'); %fit logn params
h1=kstest(eqD,'CDF',pd); %perform ks test
%set answer
if h1==0
    lognanswer='yes';
else
    lognanswer='no';
end

%aspect ratio
AR=L./l;

%orientation, revert to -90 to 90 scale
for i=1:N
    if O(i)<90
        O(i)=O(i);
    elseif O(i)>90
        O(i)=(180-O(i))*(-1);
    end
end
%call up interactive plotting
figure(1);
hold on;
%set(gca,'XScale','log');
plot(leqD,lfreq,'ko');
axis square;
title('right-click upper/lower points (in that order). Hit enter twice when done');
xlabel('log(clast diameter)');
ylabel('log(# N>D/Ac)');

%eqDs(1:N-1,1)
[x,y]=ginput;
plot(x,y,'or','MarkerSize',6,'MarkerFaceColor','r');

waitforbuttonpress;
R1=x(1,1);
diff1=abs(leqD-R1);
[M,I]=min(diff1);

R2=x(2,1);
diff2=abs(leqD-R2);
[~,Q]=min(diff2);

leqD_sub=leqD(I:Q);
lfreq_sub=lfreq(I:Q);
plot(leqD_sub,lfreq_sub,'ro');
title('selected points. Hit enter to close plot and continue script');

waitforbuttonpress;
close(gcf);

%calculate D and error by regression through selected points
n=numel(leqD_sub);
G=[ones(n,1),leqD_sub];
mest=(G'*G)
\theta(G'*lfreq_sub');
lfreqpre=G*mest;
D2d=mest(2,1)*(-1);
%error estimated from misfit between regression and observed
e=lfreq_sub'-lfreqpre;
Emin=e'*e;
Var=Emin/(n-2);
Cm=Var*(inv(G'*G));
errD2d=diag(Cm);

%refline
X=xmin:1:xmax;
G = [ones(numel(X),1),X'];
Y = G*mea;

%plot figure
figure (1);
clf;
hold on;
set(gca,'units','inches','Position',[0 0 8.5 11]);
subplot(2,2,1);
hold on;
plot(leqD,lfreq,'ko');
plot(leqD_sub,lfreq_sub,'ro');
plot(X,Y,'r-');
set(gca,'LineWidth',1,'TickLength',[0.015 0.02],'FontName','Arial','XMinorTick','on','YMinorTick','on','XColor','k','FontSize',8);
xlabel('log[Eq. Diameter (\mum)]','FontWeight','Bold','FontSize',10);
ylabel('log[n>Eq.Diameter]','FontWeight','Bold','FontSize',10);
axis square;
box on;
title(sprintf('%s',ID{1,1}),'FontName','Arial','FontWeight','Bold','FontSize',10);

subplot(2,2,2);
histogram(eqD,edgesGS,'FaceColor',[0.5 0.5 0.5],'LineWidth',1);
set(gca,'LineWidth',1,'TickLength',[0.015 0.02],'FontName','Arial','XMinorTick','on','YMinorTick','on','XColor','k','FontSize',8);
xlabel('Eq. Diameter (\mum)','FontWeight','Bold','FontSize',10);
ylabel('Count','FontWeight','Bold','FontSize',10);
if U==2
    xlim([x1min x1max]);
ylim([y1min y1max]);
end
axis square;

subplot(2,2,3);
histogram(AR,edgesAR,'FaceColor',[0.5 0.5 0.5],'LineWidth',1);
set(gca,'LineWidth',1,'TickLength',[0.015 0.02],'FontName','Arial','XMinorTick','on','YMinorTick','on','XColor','k','FontSize',8);
xlabel('Aspect Ratio','FontWeight','Bold','FontSize',10);
ylabel('Count','FontWeight','Bold','FontSize',10);
if U==2
    xlim([x2min x2max]);
ylim([y2min y2max]);
end
axis square;

subplot(2,2,4);
histogram(O,18,'FaceColor',[0.5 0.5 0.5],'LineWidth',1);
set(gca,'LineWidth',1,'TickLength',[0.015 0.02],'FontName', 'Arial','XMinorTick','on','YMinorTick','on','XColor','k','FontSize',8); xlabel('Orientation (°)'); ylabel('Count'); xlim([-90 90]); if U==2 ylim([0 y3max]); end axis square; %calculate stats muD=mean(eqD); stdD=std(eqD); muAR=mean(AR); stdAR=std(AR); %for orientation, take vector mean and vector strength sin=sind(2*O); cos=cosd(2*O); sin=sum(sin); cos=sum(cos); muO=atand(sin/cos)*(1/2); %mean vector a=sqrt((sin^2+cos^2))*(1/N); %vector strength

figure(2) %visually inspect cdf's to confirm lognormal fit hold on; cdfplot(eqD); x_values=linspace(min(eqD),max(eqD),100); plot(x_values,cdf(pd,x_values),'r-'); title('visual comparison of lognormal dist fit'); xlabel('Equivalent diameter (um)'); ylabel('Cumulative probability'); 

%write results to screen fprintf('D2d: %f +/- %f (1s)', D2d, 10^(errD2d(2,1))); fprintf('Mean eqD: %f +/- %f (1s)', muD, stdD); fprintf('Mean Asp. Ratio: %f +/- %f (1s)', muAR, stdAR); fprintf('Mean orientation: %f, vector strength: %f, muO, a); fprintf('Cumulative Area: %f, Ac); fprintf('Lognormal?: %s', lognanswer); 

%output fractal stuff for later plotting dlmwrite('D2d_all.txt',[leqD,lfreq],'delimiter','t'); dlmwrite('D2d_reg.txt',[leqD_sub,lfreq_sub],'delimiter','t');
Figure S1: Thin sections of fault mirror samples W17-1G and W17-1J. Plane polarized light. Note R and R' shears in W17-1G. Notable shear fractures in W17-1J are likely from an additional deformation event (see main text for details).
Figure S2: Grain tracings for image analyses. All are of hematite, with the exception of Domain 3 breccia clasts.
Figure S2 (continued): Grain tracings for image analyses. All are of hematite, with the exception of Domain 3 breccia clasts.
Figure S2 (continued): Grain tracings for image analyses. All are of hematite, with the exception of Domain 3 breccia clasts.
<table>
<thead>
<tr>
<th>Domain</th>
<th>n</th>
<th>$A_c$</th>
<th>$D_{2D}$</th>
<th>± LogN?</th>
<th>eqD ($\mu m^2$)</th>
<th>± AR</th>
<th>± $\hat{V}$</th>
<th>$\bar{a}$</th>
<th>SPO?</th>
</tr>
</thead>
<tbody>
<tr>
<td>Domain 1</td>
<td>145</td>
<td>41.0</td>
<td>2.3</td>
<td>1.0</td>
<td>Y</td>
<td>0.5</td>
<td>0.4</td>
<td>2.0</td>
<td>1.2</td>
</tr>
<tr>
<td>Domain 2</td>
<td>219</td>
<td>869.6</td>
<td>N/A</td>
<td>1.0</td>
<td>Y</td>
<td>1.9</td>
<td>1.2</td>
<td>4.4</td>
<td>2.8</td>
</tr>
<tr>
<td>Domain 2b</td>
<td>214</td>
<td>1539.1</td>
<td>N/A</td>
<td>1.1</td>
<td>Y</td>
<td>2.6</td>
<td>1.5</td>
<td>4.1</td>
<td>2.3</td>
</tr>
<tr>
<td>Domain 3 (matrix)</td>
<td>179</td>
<td>3025.2</td>
<td>N/A</td>
<td>1.0</td>
<td>N</td>
<td>4.2</td>
<td>2.2</td>
<td>6.2</td>
<td>3.6</td>
</tr>
<tr>
<td>Domain 3 (clast)</td>
<td>230</td>
<td>8.9E+06</td>
<td>1.5</td>
<td>1.0</td>
<td>N</td>
<td>156.0</td>
<td>150.0</td>
<td>2.1</td>
<td>0.8</td>
</tr>
<tr>
<td>Domain 1</td>
<td>223</td>
<td>518.5</td>
<td>2.5</td>
<td>1.0</td>
<td>Y</td>
<td>1.4</td>
<td>1.0</td>
<td>2.9</td>
<td>2.1</td>
</tr>
<tr>
<td>Domain 1b</td>
<td>119</td>
<td>9.2</td>
<td>1.6</td>
<td>1.0</td>
<td>Y</td>
<td>0.3</td>
<td>0.2</td>
<td>2.2</td>
<td>1.4</td>
</tr>
<tr>
<td>Domain 2</td>
<td>135</td>
<td>405.2</td>
<td>N/A</td>
<td>1.0</td>
<td>Y</td>
<td>1.7</td>
<td>1.0</td>
<td>2.8</td>
<td>1.9</td>
</tr>
<tr>
<td>Domain 2b</td>
<td>179</td>
<td>1853.2</td>
<td>N/A</td>
<td>1.0</td>
<td>Y</td>
<td>3.3</td>
<td>1.6</td>
<td>2.7</td>
<td>1.4</td>
</tr>
<tr>
<td>Domain 3 (matrix)</td>
<td>171</td>
<td>1204.0</td>
<td>N/A</td>
<td>2.1</td>
<td>Y</td>
<td>2.7</td>
<td>1.3</td>
<td>4.1</td>
<td>2.2</td>
</tr>
<tr>
<td>Domain 3 (clast)</td>
<td>343</td>
<td>1.8E+07</td>
<td>1.9</td>
<td>1.0</td>
<td>N</td>
<td>168.0</td>
<td>197.0</td>
<td>2.0</td>
<td>0.9</td>
</tr>
<tr>
<td>Domain 1</td>
<td>163</td>
<td>72.1</td>
<td>2.8</td>
<td>1.0</td>
<td>N</td>
<td>0.6</td>
<td>0.4</td>
<td>2.5</td>
<td>1.7</td>
</tr>
<tr>
<td>Domain 2</td>
<td>184</td>
<td>732.1</td>
<td>N/A</td>
<td>1.1</td>
<td>Y</td>
<td>1.8</td>
<td>1.3</td>
<td>3.0</td>
<td>2.0</td>
</tr>
<tr>
<td>Domain 3 (matrix)</td>
<td>161</td>
<td>3720.0</td>
<td>N/A</td>
<td>1.1</td>
<td>Y</td>
<td>4.9</td>
<td>2.3</td>
<td>5.4</td>
<td>3.1</td>
</tr>
<tr>
<td>Domain 3 (clast)</td>
<td>251</td>
<td>2.6E+07</td>
<td>1.3</td>
<td>1.0</td>
<td>N</td>
<td>170.0</td>
<td>316.8</td>
<td>2.1</td>
<td>1.3</td>
</tr>
</tbody>
</table>

Notes:
(i) abbreviations are: $n=$number of particles, $A_c=$cumulative area, $D_{2D}$=2D fractal dimension, LogN=log-normal, eqD=equivalent diameter, AR=aspect ratio, $\hat{V}$=mean orientation, $\bar{a}$=vector strength, SPO=shape preferred orientation
(ii) $D_{2D}$ uncertainty calculated from linear regression
(iii) eqD and AR reported as mean with standard deviation
(iv) Column “LogN?” gives results of Kolmogorov-Smirnov test for statistically significant lognormal distribution at 95% confidence.
Some distributions are more qualitatively consistent with power-law distribution with rollover at small grain sizes. See text for details.
Analytical methods

Hematite was extracted from fault surfaces using a Dremel™ tool (see Figure S1 for sampling locations). Isolated chips were the broken into 10-12 replicate aliquots per sample with fine-point tweezers and attached to a 1-inch metal stub with double-sided adhesive copper tape. Hematite aliquots were imaged with backscattered electron, sensitive to atomic contrast, on a Quanta 650 FEG scanning electron microscope (SEM) at Utah State University’s Microscopy Core Facility (USU MCF) to rapidly identify and exclude aliquots with a high proportion of interstitial (i.e., non-hematite) phases. SEM imaging was conducted at low-vacuum and 10.0 kV accelerating voltage. From these pre-screened samples, ~3 to 4 aliquots were selected and loaded into Nb packets for U, Th, and He analysis.

We acquired hematite (U-Th)/He data over three separate analytical sessions in 2014, 2017, and 2018 at the Arizona Radiogenic Helium Dating Laboratory (ARHDL) at the University of Arizona. Aliquots were heated to temperatures and packet “glow” comparable to apatite for 5 minutes using a diode laser in an ultra-high vacuum gas extraction line. Liberated $^4$He gas was spiked with $^3$He, purified using standard cryogenic and gettering methods, and measured on a quadrupole mass spectrometer. We conducted gas re-extracts at higher glow and for 6 minutes and until <1% He was released relative to initial degassing. Most samples required one re-extract to extract all $^4$He. Degassed
samples were dissolved in HF acid in pressure digestion vessels. Isotopes of U and Th were measured on an Element 2 ICP-MS following addition of a $^{233}$U-$^{229}$Th spike, equilibration, and dissolution. For the 2014 and 2016 analyses, Fish Canyon Tuff zircon was used as a standard to monitor chemistry and instrument performance. For 2018 analyses, we used WF94-17 hematite (Ault et al., 2015) as an internal standard. Blank-corrected (U-Th)/He dates were calculated with propagated analytical uncertainties from U, Th, and He measurements. For the 2016 and 2018 analytical sessions, aliquot mass was determined prior to degassing by differencing loading and empty Nb packet weights. We thus calculating elemental concentrations from absolute abundance data for a subset of our samples. No alpha-ejection correction was applied to the hematite He dates because all aliquots are sufficiently thick (>120 μm) such that He ejection is balanced by implantation (Fig. S2; S3 Farley and Flowers, 2012; Evenson et al., 2014).

Hematite chips representative of our dated samples were mounted in 1” epoxy rounds, polished, and carbon-coated for SEM imaging, Energy Dispersive Spectroscopy (EDS), and grain size measurements at the USU MCF. Reflected light images of our mounts are provided in Figure S2. Samples were dual-imaged with secondary and backscatter electrons and at high-vacuum and 20-30 kV accelerating voltage. Semiquantitative elemental abundances were measured by SEM-EDS with an Oxford X-Max detector (Fig. S4). We conducted mapping and spot analyses, allowing ~1-3 minutes of measurement time per analysis. We measured hematite plate half-widths from SEM images using ImageJ software (Fig. S5; Scheider et al., 2012).
Figure S1: Hematite-coated fault surface hand sample photographs. White strike/dip symbols give fault surface orientation. White dashed lines highlight striae directions. Location of hematite (U-Th)/He dates shown as white hexagons.
Figure S2: Additional photomicrographs of select hematite fault surfaces. Hem-hematite, Cal-calcite, Qtz-quartz, Chl-clorite.
Figure S3: Image catalogue of (U-Th)/He-dated aliquots (reflected light, top row) and scanning electron microscope images from representative samples (secondary electron, bottom row). White and blue dashed lines denote slip surface and foliation, respectively.

61014-6A

61414-2C
Figure S3 (continued): Image catalogue of (U-Th)/He-dated aliquots (reflected light, top row) and scanning electron microscope images from representative samples (secondary electron, bottom row). White and blue dashed lines denote slip surface and foliation, respectively.

61714-7A

61315-10E

61315-10E_1

61315-10E_2
**Figure S3 (continued):** Image catalogue of (U-Th)/He-dated aliquots (reflected light, top row) and scanning electron microscope images from representative samples (secondary electron, bottom row). White and blue dashed lines denote slip surface and foliation, respectively.
Figure S3 (continued): Image catalogue of (U-Th)/He-dated aliquots (reflected light, top row) and scanning electron microscope images from representative samples (secondary electron, bottom row). White and blue dashed lines denote slip surface and foliation, respectively.
**Figure S3 (continued):** Image catalogue of (U-Th)/He-dated aliquots (reflected light, top row) and scanning electron microscope images from representative samples (secondary electron, bottom row). White and blue dashed lines denote slip surface and foliation, respectively.
**Figure S4:** Energy dispersive spectroscopy maps and spot analyses of hematite fault surface samples. Phase identifications are based on relative elemental abundances. Hem-hematite, Cal-calcite, Cl-clay (kaolin and/or smectite), Qtz-quartz, Chl-chlorite, Fspar-feldspar.

**61714-7A**

*Foliated hematite*

![Image of Foliated hematite](image1)

**61215-5A**

*Foliated hematite + cataclasite*

![Image of Foliated hematite + cataclasite](image2)

**Nanoparticles**

![Image of Nanoparticles](image3)
Figure S5: Grain size (left) and closure temperature ($T_c$, right) distributions. $T_c$ calculations assume Farley (2018) diffusion kinetics, a cooling rate of 10 °C/Myr, and diffusion domain lengthscale equal to plate half-width.

61014-6A (n=70)

![Histogram showing grain size distribution with mean ± st. dev: 35.3 ± 11.4 nm and closure temperature distribution with mean ± st. dev: 67.1 ± 3.4 °C.]

61414-2C (n=70)

![Histogram showing grain size distribution with mean ± st. dev: 26.2 ± 7.4 nm and closure temperature distribution with mean ± st. dev: 64.0 ± 2.8 °C.]

61914-17A (n=70)

![Histogram showing grain size distribution with mean ± st. dev: 22.3 ± 7.5 nm and closure temperature distribution with mean ± st. dev: 62.2 ± 3.0 °C.]

61114-4A (n=70)

![Histogram showing grain size distribution with mean ± st. dev: 37.5 ± 11.2 nm and closure temperature distribution with mean ± st. dev: 67.8 ± 3.0 °C.]

61714-7A (n=70)

![Histogram showing grain size distribution with mean ± st. dev: 29.3 ± 8.6 nm and closure temperature distribution with mean ± st. dev: 65.2 ± 3.4 °C.]

Plate width (nm) | Closure temperature (°C)
Figure S5 (continued): Grain size (left) and closure temperature (T_c, right) distributions. T_c calculations assume Farley (2018) diffusion kinetics, a cooling rate of 10 °C/Myr, and diffusion domain lengthscale equal to plate half-width.

61315-10E (n=70)

61215-2A (n=70)

61215-5A (n=70)

61815-6A (n=70)
<table>
<thead>
<tr>
<th>Mass (μg)</th>
<th>± 1σ</th>
<th>U (ng)</th>
<th>± 1σ</th>
<th>Th (ng)</th>
<th>± 1σ</th>
<th>4^1^He (fmol)</th>
<th>± 1σ</th>
<th>eU (ppm)</th>
<th>± 1σ</th>
<th>4^4^He (nmol/g)</th>
<th>± 1σ</th>
<th>Date (Ma)</th>
<th>Error (Ma)</th>
</tr>
</thead>
<tbody>
<tr>
<td>61014-6A, 61.079, -138.5329; 792 masl</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>H1 --</td>
<td>0.015</td>
<td>0.00025</td>
<td>0.0182</td>
<td>0.0003</td>
<td>0.84</td>
<td>0.09</td>
<td>--</td>
<td>--</td>
<td>8.21</td>
<td>0.92</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>H2 --</td>
<td>0.011</td>
<td>0.00019</td>
<td>0.0052</td>
<td>0.0002</td>
<td>0.39</td>
<td>0.09</td>
<td>--</td>
<td>--</td>
<td>5.74</td>
<td>1.31</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>H3 --</td>
<td>0.0039</td>
<td>0.0001</td>
<td>0.0036</td>
<td>0.0001</td>
<td>0.85</td>
<td>0.09</td>
<td>--</td>
<td>--</td>
<td>33.09</td>
<td>3.72</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>H4 --</td>
<td>0.004</td>
<td>0.00011</td>
<td>0.0048</td>
<td>0.0001</td>
<td>0.69</td>
<td>0.09</td>
<td>--</td>
<td>--</td>
<td>24.50</td>
<td>3.25</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>61414-2C, 60.8927, -138.122; 969 masl</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>H1 --</td>
<td>0.0356</td>
<td>0.0006</td>
<td>0.0052</td>
<td>0.0002</td>
<td>1.06</td>
<td>0.05</td>
<td>--</td>
<td>--</td>
<td>5.37</td>
<td>0.28</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>H2 --</td>
<td>0.0547</td>
<td>0.0008</td>
<td>0.0049</td>
<td>0.0001</td>
<td>1.61</td>
<td>0.05</td>
<td>--</td>
<td>--</td>
<td>5.34</td>
<td>0.19</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>H3 --</td>
<td>0.0475</td>
<td>0.0008</td>
<td>0.0038</td>
<td>0.0001</td>
<td>1.34</td>
<td>0.05</td>
<td>--</td>
<td>--</td>
<td>5.15</td>
<td>0.20</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>H4 --</td>
<td>0.0525</td>
<td>0.0008</td>
<td>0.0095</td>
<td>0.0002</td>
<td>1.69</td>
<td>0.06</td>
<td>--</td>
<td>--</td>
<td>5.74</td>
<td>0.23</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>61914-1A, 61.0783, -138.5314; 792 masl</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>H1 --</td>
<td>0.0379</td>
<td>0.0006</td>
<td>0.0156</td>
<td>0.0002</td>
<td>1.81</td>
<td>0.05</td>
<td>--</td>
<td>--</td>
<td>8.07</td>
<td>0.25</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>H2 --</td>
<td>0.0455</td>
<td>0.0007</td>
<td>0.0075</td>
<td>0.0001</td>
<td>1.91</td>
<td>0.05</td>
<td>--</td>
<td>--</td>
<td>7.48</td>
<td>0.22</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>H3 --</td>
<td>0.0570</td>
<td>0.0009</td>
<td>0.0051</td>
<td>0.0001</td>
<td>2.08</td>
<td>0.05</td>
<td>--</td>
<td>--</td>
<td>6.63</td>
<td>0.19</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>H4 --</td>
<td>0.0575</td>
<td>0.0008</td>
<td>0.0500</td>
<td>0.0007</td>
<td>2.87</td>
<td>0.07</td>
<td>--</td>
<td>--</td>
<td>7.69</td>
<td>0.21</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>61111-4A, 61.0843, -138.5512; 897 masl</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>H2 57</td>
<td>1.414</td>
<td>0.0068</td>
<td>0.0001</td>
<td>0.0042</td>
<td>0.0001</td>
<td>0.26</td>
<td>0.05</td>
<td>0.14</td>
<td>0.00</td>
<td>0.00</td>
<td>6.14</td>
<td>1.18</td>
<td></td>
</tr>
<tr>
<td>H3 65</td>
<td>1.414</td>
<td>0.0072</td>
<td>0.0001</td>
<td>0.0012</td>
<td>0.0001</td>
<td>0.40</td>
<td>0.05</td>
<td>0.11</td>
<td>0.00</td>
<td>0.01</td>
<td>9.84</td>
<td>1.19</td>
<td></td>
</tr>
<tr>
<td>H4 28</td>
<td>1.414</td>
<td>0.0070</td>
<td>0.0004</td>
<td>0.0075</td>
<td>0.0003</td>
<td>0.22</td>
<td>0.05</td>
<td>0.31</td>
<td>0.03</td>
<td>0.01</td>
<td>4.71</td>
<td>1.00</td>
<td></td>
</tr>
<tr>
<td>61714-7A, 61.0844, -138.5517; 895 masl</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>H1 96</td>
<td>1.414</td>
<td>0.0057</td>
<td>0.0001</td>
<td>0.0018</td>
<td>0.0001</td>
<td>0.38</td>
<td>0.05</td>
<td>0.06</td>
<td>0.00</td>
<td>0.00</td>
<td>11.46</td>
<td>1.57</td>
<td></td>
</tr>
<tr>
<td>H2 23</td>
<td>1.414</td>
<td>0.0050</td>
<td>0.0001</td>
<td>0.0040</td>
<td>0.0001</td>
<td>0.32</td>
<td>0.06</td>
<td>0.26</td>
<td>0.02</td>
<td>0.01</td>
<td>9.96</td>
<td>1.77</td>
<td></td>
</tr>
<tr>
<td>H3 49</td>
<td>1.414</td>
<td>0.0051</td>
<td>0.0001</td>
<td>0.0027</td>
<td>0.0001</td>
<td>0.31</td>
<td>0.05</td>
<td>0.12</td>
<td>0.00</td>
<td>0.01</td>
<td>10.06</td>
<td>1.73</td>
<td></td>
</tr>
<tr>
<td>61315-10E_1, 60.8928, -138.1226; 976 masl</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>H2 33</td>
<td>1.414</td>
<td>0.1281</td>
<td>0.0020</td>
<td>0.0024</td>
<td>0.0001</td>
<td>4.60</td>
<td>0.04</td>
<td>3.90</td>
<td>0.18</td>
<td>0.14</td>
<td>6.63</td>
<td>0.11</td>
<td></td>
</tr>
<tr>
<td>H3 54</td>
<td>1.414</td>
<td>0.1354</td>
<td>0.0019</td>
<td>0.0046</td>
<td>0.0001</td>
<td>4.71</td>
<td>0.04</td>
<td>2.53</td>
<td>0.07</td>
<td>0.09</td>
<td>6.40</td>
<td>0.10</td>
<td></td>
</tr>
<tr>
<td>H4 13</td>
<td>1.414</td>
<td>0.0367</td>
<td>0.0005</td>
<td>0.0013</td>
<td>0.0001</td>
<td>1.26</td>
<td>0.03</td>
<td>2.85</td>
<td>0.31</td>
<td>0.10</td>
<td>6.31</td>
<td>0.17</td>
<td></td>
</tr>
<tr>
<td>Mass</td>
<td>U</td>
<td>Th</td>
<td>(^4)He</td>
<td>eU</td>
<td>Date</td>
<td>Error</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>-------</td>
<td>------</td>
<td>------</td>
<td>----------</td>
<td>------</td>
<td>------</td>
<td>-------</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(µg)</td>
<td>(ng)</td>
<td>(ng)</td>
<td>(fmol)</td>
<td>(ppm)</td>
<td>(Ma)</td>
<td>(Ma)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(\pm 1\sigma)</td>
<td>(\pm 1\sigma)</td>
<td>(\pm 1\sigma)</td>
<td>(\pm 1\sigma)</td>
<td>(\pm 1\sigma)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>61315-10E_2</td>
<td>60.8928, -138.1226; 976 masl</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>H1</td>
<td>34</td>
<td>1.414 0.1131</td>
<td>0.0016</td>
<td>0.0025</td>
<td>0.0001</td>
<td>3.83</td>
<td>0.04</td>
<td>0.00</td>
<td>6.24</td>
<td>0.11</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>H2</td>
<td>40</td>
<td>1.414 0.1322</td>
<td>0.0019</td>
<td>0.0031</td>
<td>0.0001</td>
<td>4.36</td>
<td>0.04</td>
<td>0.13</td>
<td>0.11</td>
<td>0.00</td>
<td>6.09</td>
<td>0.10</td>
<td></td>
</tr>
<tr>
<td>H3</td>
<td>22</td>
<td>1.414 0.0329</td>
<td>0.0005</td>
<td>0.0036</td>
<td>0.0001</td>
<td>0.90</td>
<td>0.02</td>
<td>0.10</td>
<td>0.04</td>
<td>0.00</td>
<td>4.94</td>
<td>0.14</td>
<td></td>
</tr>
<tr>
<td>H4</td>
<td>9</td>
<td>1.414 0.0150</td>
<td>0.0003</td>
<td>0.0021</td>
<td>0.0001</td>
<td>0.34</td>
<td>0.02</td>
<td>0.27</td>
<td>0.04</td>
<td>0.01</td>
<td>4.03</td>
<td>0.26</td>
<td></td>
</tr>
<tr>
<td>61215-2A_1, 61.0802, -138.6251; 2330 masl</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>H1</td>
<td>25</td>
<td>1.4   0.0089</td>
<td>0.0002</td>
<td>0.0066</td>
<td>0.0001</td>
<td>0.3501</td>
<td>0.03</td>
<td>0.014</td>
<td>0.0013</td>
<td>6.23</td>
<td>0.483</td>
<td></td>
<td></td>
</tr>
<tr>
<td>H2</td>
<td>16</td>
<td>1.4   0.0042</td>
<td>0.0001</td>
<td>0.0036</td>
<td>0.0001</td>
<td>0.1601</td>
<td>0.02</td>
<td>0.315</td>
<td>0.03</td>
<td>0.01</td>
<td>0.0017</td>
<td>5.89</td>
<td>0.893</td>
</tr>
<tr>
<td>H3</td>
<td>17</td>
<td>1.4   0.0038</td>
<td>0.0001</td>
<td>0.0025</td>
<td>0.0001</td>
<td>0.12</td>
<td>0.03</td>
<td>0.257</td>
<td>0.02</td>
<td>0.0068</td>
<td>0.0017</td>
<td>4.90</td>
<td>1.131</td>
</tr>
<tr>
<td>H4</td>
<td>22</td>
<td>1.4   0.0046</td>
<td>0.0001</td>
<td>0.0028</td>
<td>0.0001</td>
<td>0.22</td>
<td>0.03</td>
<td>0.239</td>
<td>0.02</td>
<td>0.01016</td>
<td>0.0017</td>
<td>7.89</td>
<td>0.908</td>
</tr>
<tr>
<td>61215-2A_2, 61.0802, -138.6251; 2330 masl</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>H1</td>
<td>29</td>
<td>1.4   0.027</td>
<td>0.00042</td>
<td>0.0112</td>
<td>0.0002</td>
<td>1.15</td>
<td>0.03</td>
<td>0.1026</td>
<td>0.05</td>
<td>0.0398</td>
<td>0.0022</td>
<td>7.20</td>
<td>0.228</td>
</tr>
<tr>
<td>H2</td>
<td>36</td>
<td>1.4   0.01</td>
<td>0.00018</td>
<td>0.0083</td>
<td>0.0001</td>
<td>0.77</td>
<td>0.03</td>
<td>0.323</td>
<td>0.02</td>
<td>0.02152</td>
<td>0.0012</td>
<td>12.34</td>
<td>0.56</td>
</tr>
<tr>
<td>H3</td>
<td>25</td>
<td>1.4   0.013</td>
<td>0.00024</td>
<td>0.0067</td>
<td>0.0001</td>
<td>0.42</td>
<td>0.03</td>
<td>0.593</td>
<td>0.03</td>
<td>0.01698</td>
<td>0.0015</td>
<td>5.31</td>
<td>0.374</td>
</tr>
<tr>
<td>H4</td>
<td>24</td>
<td>1.4   0.015</td>
<td>0.00024</td>
<td>0.0076</td>
<td>0.0002</td>
<td>0.58</td>
<td>0.03</td>
<td>0.685</td>
<td>0.04</td>
<td>0.02401</td>
<td>0.0019</td>
<td>6.50</td>
<td>0.346</td>
</tr>
<tr>
<td>61215-5A, 61.0811, -138.6194; 2199 masl</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>H1</td>
<td>38</td>
<td>1.4   0.0237</td>
<td>0.0004</td>
<td>0.0105</td>
<td>0.0002</td>
<td>0.58</td>
<td>0.03</td>
<td>0.69</td>
<td>0.03</td>
<td>0.02</td>
<td>0.00</td>
<td>4.13</td>
<td>0.21</td>
</tr>
<tr>
<td>H2</td>
<td>36</td>
<td>1.4   0.0273</td>
<td>0.0004</td>
<td>0.0135</td>
<td>0.0003</td>
<td>0.59</td>
<td>0.03</td>
<td>0.85</td>
<td>0.04</td>
<td>0.02</td>
<td>0.00</td>
<td>3.56</td>
<td>0.17</td>
</tr>
<tr>
<td>H3</td>
<td>54</td>
<td>1.4   0.0557</td>
<td>0.0008</td>
<td>0.0217</td>
<td>0.0003</td>
<td>1.27</td>
<td>0.03</td>
<td>1.13</td>
<td>0.03</td>
<td>0.02</td>
<td>0.00</td>
<td>3.86</td>
<td>0.11</td>
</tr>
<tr>
<td>H4</td>
<td>39</td>
<td>1.4   0.0324</td>
<td>0.0005</td>
<td>0.0135</td>
<td>0.0002</td>
<td>0.69</td>
<td>0.03</td>
<td>0.91</td>
<td>0.04</td>
<td>0.02</td>
<td>0.00</td>
<td>3.60</td>
<td>0.17</td>
</tr>
<tr>
<td>61815-6A, 61.0837, -138.5565; 935 masl</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>H1</td>
<td>19</td>
<td>1.4   0.0031</td>
<td>0.0001</td>
<td>0.0021</td>
<td>0.0001</td>
<td>0.07</td>
<td>0.02</td>
<td>0.19</td>
<td>0.02</td>
<td>0.0035</td>
<td>0.0011</td>
<td>3.41</td>
<td>1.07</td>
</tr>
<tr>
<td>H2</td>
<td>12</td>
<td>1.4   0.0033</td>
<td>0.0001</td>
<td>0.0006</td>
<td>0.0001</td>
<td>0.12</td>
<td>0.02</td>
<td>0.29</td>
<td>0.03</td>
<td>0.0098</td>
<td>0.0021</td>
<td>6.34</td>
<td>1.17</td>
</tr>
<tr>
<td>H3</td>
<td>4</td>
<td>1.4   0.0024</td>
<td>0.0001</td>
<td>0.0007</td>
<td>0.0001</td>
<td>0.08</td>
<td>0.02</td>
<td>0.64</td>
<td>0.22</td>
<td>0.0202</td>
<td>0.0089</td>
<td>5.84</td>
<td>1.58</td>
</tr>
<tr>
<td>H4</td>
<td>25</td>
<td>1.4   0.0033</td>
<td>0.0001</td>
<td>0.0006</td>
<td>0.0001</td>
<td>0.14</td>
<td>0.02</td>
<td>0.14</td>
<td>0.01</td>
<td>0.0056</td>
<td>0.0009</td>
<td>7.58</td>
<td>1.11</td>
</tr>
</tbody>
</table>

\(^a\) is propagated error from analytical uncertainties on U, Th, and He analyses.

\(^b\) Latitude and longitude reported in decimal degrees; WGS84

\(^c\) discarded due to high analytical error
<table>
<thead>
<tr>
<th>Sample Group</th>
<th>Weighted Mean Date (Ma)</th>
<th>1σ (Ma)</th>
<th>MSWD a</th>
<th>Central Date (Ma)b</th>
<th>1σ (Ma)</th>
<th>MSWDc</th>
<th>Disp. (%)d</th>
</tr>
</thead>
<tbody>
<tr>
<td>All samples (n=38)</td>
<td>5.7</td>
<td>0.1</td>
<td>33.8</td>
<td>6.1</td>
<td>0.3</td>
<td>28.0</td>
<td>26.2</td>
</tr>
<tr>
<td>AK Highway (n=6)</td>
<td>7.4</td>
<td>0.1</td>
<td>5.7</td>
<td>7.5</td>
<td>0.3</td>
<td>5.4</td>
<td>6.5</td>
</tr>
<tr>
<td>Williscroft Creek (all; n=21)</td>
<td>4.1</td>
<td>0.1</td>
<td>11.0</td>
<td>6.0</td>
<td>0.5</td>
<td>24.0</td>
<td>33.2</td>
</tr>
<tr>
<td>Williscroft Creek (low el.; n=10)</td>
<td>6.9</td>
<td>0.4</td>
<td>4.2</td>
<td>7.5</td>
<td>0.8</td>
<td>3.5</td>
<td>27.0</td>
</tr>
<tr>
<td>Williscroft Creek (high el.; n=11)</td>
<td>4.1</td>
<td>0.6</td>
<td>13.6</td>
<td>5.1</td>
<td>0.4</td>
<td>37.0</td>
<td>26.8</td>
</tr>
<tr>
<td>Telluride Creek (n=11)</td>
<td>6.0</td>
<td>0.4</td>
<td>20.5</td>
<td>5.7</td>
<td>0.23</td>
<td>17.0</td>
<td>12.8</td>
</tr>
</tbody>
</table>

aMean square of the weighted deviates, with respect to weighted mean date
bCentral date of Galbraith (1990)
cMSWD with respect to central date
dDispersion wth respect to central date
References


APPENDIX D

Copyright release for Chapter 5

JOHN WILEY AND SONS LICENSE
TERMS AND CONDITIONS

Oct 05, 2020

This Agreement between Utah State University -- Robert McDermott ("You") and John Wiley and Sons ("John Wiley and Sons") consists of your license details and the terms and conditions provided by John Wiley and Sons and Copyright Clearance Center.

License Number
4916250278514

License date
Sep 25, 2020

Licensed Content Publisher
John Wiley and Sons

Licensed Content Publication
Tectonics

Licensed Content Title
Thermotectonic History of the Kluane Ranges and Evolution of the Eastern Denali Fault Zone in Southwestern Yukon, Canada

Licensed Content Author

Licensed Content Date
Aug 15, 2019

Licensed Content Volume
38

Licensed Content Issue
8

Licensed Content Pages
28

Type of use
Dissertation/Thesis

Requestor type
Author of this Wiley article

Format
Electronic

Portion
Full article

Will you be translating?
No

Title
Thermotectonic History of the Kluane Ranges and Evolution of the Eastern Denali Fault Zone in Southwestern Yukon, Canada

Institution name
Utah State University

Expected presentation date
Oct 2020
Order reference number

1

Requestor Location

Utah State University
4505 Old Main Hill
Department of Geosciences
Utah State University
LOGAN, UT 84322
United States
Attn: Utah State University

Publisher Tax ID

EU826007151

Total

0.00 USD

Terms and Conditions

TERMS AND CONDITIONS

This copyrighted material is owned by or exclusively licensed to John Wiley & Sons, Inc. or one of its group companies (each a"Wiley Company") or handled on behalf of a society with which a Wiley Company has exclusive publishing rights in relation to a particular work (collectively "WILEY"). By clicking "accept" in connection with completing this licensing transaction, you agree that the following terms and conditions apply to this transaction (along with the billing and payment terms and conditions established by the Copyright Clearance Center Inc., ("CCC's Billing and Payment terms and conditions"), at the time that you opened your RightsLink account (these are available at any time at http://myaccount.copyright.com).

Terms and Conditions

- The materials you have requested permission to reproduce or reuse (the "Wiley Materials") are protected by copyright.

- You are hereby granted a personal, non-exclusive, non-sub licensable (on a stand-alone basis), non-transferable, worldwide, limited license to reproduce the Wiley Materials for the purpose specified in the licensing process. This license, and any CONTENT (PDF or image file) purchased as part of your order, is for a one-
time use only and limited to any maximum distribution number specified in the license. The first instance of republication or reuse granted by this license must be completed within two years of the date of the grant of this license (although copies prepared before the end date may be distributed thereafter). The Wiley Materials shall not be used in any other manner or for any other purpose, beyond what is granted in the license. Permission is granted subject to an appropriate acknowledgement given to the author, title of the material/book/journal and the publisher. You shall also duplicate the copyright notice that appears in the Wiley publication in your use of the Wiley Material. Permission is also granted on the understanding that nowhere in the text is a previously published source acknowledged for all or part of this Wiley Material. Any third party content is expressly excluded from this permission.

- With respect to the Wiley Materials, all rights are reserved. Except as expressly granted by the terms of the license, no part of the Wiley Materials may be copied, modified, adapted (except for minor reformatting required by the new publication), translated, reproduced, transferred or distributed, in any form or by any means, and no derivative works may be made based on the Wiley Materials without the prior permission of the respective copyright owner. For STM Signatory Publishers clearing permission under the terms of the STM Permissions Guidelines only, the terms of the license are extended to include subsequent editions and for editions in other languages, provided such editions are for the work as a whole in situ and does not involve the separate exploitation of the permitted figures or extracts. You may not alter, remove or suppress in any manner any copyright, trademark or other notices displayed by the Wiley Materials. You may not license, rent, sell, loan, lease, pledge, offer as security, transfer or assign the Wiley Materials on a stand-alone basis, or any of the rights granted to you hereunder to any other person.

- The Wiley Materials and all of the intellectual property rights therein shall at all times remain the exclusive property of John Wiley & Sons Inc, the Wiley Companies, or their respective licensors, and your interest therein is only that of having possession of and the right to reproduce the Wiley Materials pursuant to Section 2 herein during the continuance of this Agreement. You agree that you own no right, title or interest in or to the Wiley Materials or any of the intellectual property rights therein. You shall have no rights hereunder other than the license as provided for above in Section 2. No right, license or interest to any trademark, trade name, service mark or other branding ("Marks") of WILEY or its licensors is granted hereunder, and you agree that you shall not assert any such right, license or interest with respect thereto.

- NEITHER WILEY NOR ITS LICENSORS MAKES ANY WARRANTY OR REPRESENTATION OF ANY KIND TO YOU OR ANY THIRD PARTY, EXPRESS, IMPLIED OR STATUTORY, WITH RESPECT TO THE MATERIALS OR THE ACCURACY OF ANY INFORMATION CONTAINED IN THE MATERIALS, INCLUDING, WITHOUT LIMITATION, ANY IMPLIED WARRANTY OF MERCHANTABILITY, ACCURACY,
SATISFACTORY QUALITY, FITNESS FOR A PARTICULAR PURPOSE, USABILITY, INTEGRATION OR NON-INFRINGEMENT AND ALL SUCH WARRANTIES ARE HEREBY EXCLUDED BY WILEY AND ITS LICENSORS AND WAIVED BY YOU.

- WILEY shall have the right to terminate this Agreement immediately upon breach of this Agreement by you.

- You shall indemnify, defend and hold harmless WILEY, its Licensors and their respective directors, officers, agents and employees, from and against any actual or threatened claims, demands, causes of action or proceedings arising from any breach of this Agreement by you.

- IN NO EVENT SHALL WILEY OR ITS LICENSORS BE LIABLE TO YOU OR ANY OTHER PARTY OR ANY OTHER PERSON OR ENTITY FOR ANY SPECIAL, CONSEQUENTIAL, INCIDENTAL, INDIRECT, EXEMPLARY OR PUNITIVE DAMAGES, HOWEVER CAUSED, ARISING OUT OF OR IN CONNECTION WITH THE DOWNLOADING, PROVISIONING, VIEWING OR USE OF THE MATERIALS REGARDLESS OF THE FORM OF ACTION, WHETHER FOR BREACH OF CONTRACT, BREACH OF WARRANTY, TORT, NEGLIGENCE, INFRINGEMENT OR OTHERWISE (INCLUDING, WITHOUT LIMITATION, DAMAGES BASED ON LOSS OF PROFITS, DATA, FILES, USE, BUSINESS OPPORTUNITY OR CLAIMS OF THIRD PARTIES), AND WHETHER OR NOT THE PARTY HAS BEEN ADVISED OF THE POSSIBILITY OF SUCH DAMAGES. THIS LIMITATION SHALL APPLY NOTWITHSTANDING ANY FAILURE OF ESSENTIAL PURPOSE PROVIDED HEREIN.

- Should any provision of this Agreement be held by a court of competent jurisdiction to be illegal, invalid, or unenforceable, that provision shall be deemed amended to achieve as nearly as possible the same economic effect as the original provision, and the legality, validity and enforceability of the remaining provisions of this Agreement shall not be affected or impaired thereby.

- The failure of either party to enforce any term or condition of this Agreement shall not constitute a waiver of either party's right to enforce each and every term and condition of this Agreement. No breach under this agreement shall be deemed waived or excused by either party unless such waiver or consent is in writing signed by the party granting such waiver or consent. The waiver by or consent of a party to a breach of any provision of this Agreement shall not operate or be construed as a waiver of or consent to any other or subsequent breach by such other party.

- This Agreement may not be assigned (including by operation of law or otherwise) by you without WILEY's prior written consent.
• Any fee required for this permission shall be non-refundable after thirty (30) days from receipt by the CCC.

• These terms and conditions together with CCC's Billing and Payment terms and conditions (which are incorporated herein) form the entire agreement between you and WILEY concerning this licensing transaction and (in the absence of fraud) supersedes all prior agreements and representations of the parties, oral or written. This Agreement may not be amended except in writing signed by both parties. This Agreement shall be binding upon and inure to the benefit of the parties' successors, legal representatives, and authorized assigns.

• In the event of any conflict between your obligations established by these terms and conditions and those established by CCC's Billing and Payment terms and conditions, these terms and conditions shall prevail.

• WILEY expressly reserves all rights not specifically granted in the combination of (i) the license details provided by you and accepted in the course of this licensing transaction, (ii) these terms and conditions and (iii) CCC's Billing and Payment terms and conditions.

• This Agreement will be void if the Type of Use, Format, Circulation, or Requestor Type was misrepresented during the licensing process.

• This Agreement shall be governed by and construed in accordance with the laws of the State of New York, USA, without regards to such state's conflict of law rules. Any legal action, suit or proceeding arising out of or relating to these Terms and Conditions or the breach thereof shall be instituted in a court of competent jurisdiction in New York County in the State of New York in the United States of America and each party hereby consents and submits to the personal jurisdiction of such court, waives any objection to venue in such court and consents to service of process by registered or certified mail, return receipt requested, at the last known address of such party.

WILEY OPEN ACCESS TERMS AND CONDITIONS

Wiley Publishes Open Access Articles in fully Open Access Journals and in Subscription journals offering Online Open. Although most of the fully Open Access journals publish open access articles under the terms of the Creative Commons Attribution (CC BY) License only, the subscription journals and a few of the Open Access Journals offer a choice of Creative Commons Licenses. The license type is clearly identified on the article.

The Creative Commons Attribution License

The Creative Commons Attribution License (CC-BY) allows users to copy, distribute and transmit an article, adapt the article and make commercial use of the article. The CC-BY license permits commercial and non-
Creative Commons Attribution Non-Commercial License

The Creative Commons Attribution Non-Commercial (CC-BY-NC) License permits use, distribution and reproduction in any medium, provided the original work is properly cited and is not used for commercial purposes. (see below)

Creative Commons Attribution-Non-Commercial-NoDerivs License

The Creative Commons Attribution Non-Commercial-NoDerivs License (CC-BY-NC-ND) permits use, distribution and reproduction in any medium, provided the original work is properly cited, is not used for commercial purposes and no modifications or adaptations are made. (see below)

Use by commercial "for-profit" organizations

Use of Wiley Open Access articles for commercial, promotional, or marketing purposes requires further explicit permission from Wiley and will be subject to a fee.

Further details can be found on Wiley Online Library http://olabout.wiley.com/WileyCDA/Section/id-410895.html

Other Terms and Conditions:

v1.10 Last updated September 2015

Questions? customercare@copyright.com or +1-855-239-3415 (toll free in the US) or +1-978-646-2777.
October 5th, 2020

Dear co-author,

I am in the process of preparing my written dissertation for my Doctoral degree at Utah State University. I am seeking permission to include the following article as a chapter in my dissertation:


The full author list, respective institutional affiliations, publication year, and journal of publication appear on the first page of the chapter. Please advise me of any changes you require in your acknowledgement. Please indicate your approval of this request by signing the letter where indicated below and returning a copy of this letter via email to robert.mcdermott@usu.edu.

Thank you for your time.

Sincerely,

Robert G. McDermott  
Department of Geosciences  
Utah State University  
4505 Old Main Hill  
Logan, UT 84322  
robert.mcdermott@usu.edu

**PERMISSION GRANTED FOR THE USE REQUESTED ABOVE:**

Signed:

Title: Dr. Peter Reiners

Date: 6 Oct 2020
October 6th, 2020

Dear co-author,

I am in the process of preparing my written dissertation for my Doctoral degree at Utah State University. I am seeking permission to include the following article as a chapter in my dissertation:


The full author list, respective institutional affiliations, publication year, and journal of publication appear on the first page of the chapter. Please advise me of any changes you require in your acknowledgement. Please indicate your approval of this request by signing the letter where indicated below and returning a copy of this letter via email to robert.mcdermott@usu.edu.

Thank you for your time.

Sincerely,

Robert G. McDermott
Department of Geosciences
Utah State University
4505 Old Main Hill
Logan, UT 84322
robert.mcdermott@usu.edu

PERMISSION GRANTED FOR THE USE REQUESTED ABOVE:

Signed:

Title: Dr. Stuart Thomson

Date: 10/7/2020
APPENDIX E

Curriculum Vitae

Robert G. McDermott
Department of Geosciences
Utah State University
4505 Old Main Hill
Logan, UT 84322
robert.mcdermott@usu.edu

Education

PhD Utah State University, Geosciences August 2014-present
Dissertation (in progress): A thermochronometric and numerical modeling approach to deciphering the rock record of deformation processes in the Wasatch and Denali Fault zones: implications for seismogenesis and fault zone evolution
Advisor: Dr. Alexis Ault

BS University of Pittsburgh, Geology and Planetary Sciences August 2010-April 2014
Undergraduate Honors Thesis: An 11,000-year record of natural drought variability from Rock Lake, Montana using lake sediment stable isotope geochemistry
Advisor: Dr. Mark Abbott
Concentration in mathematics, departmental honors, magna cum laude

Professional Preparation

Research Assistant, USU, Logan, UT January 2019-present
Instructor, USU, Logan, UT August 2018-December 2018
Presidential Doctoral Research Fellow, USU, Logan, UT August 2014-May 2018
Undergraduate Research Assistant, U. Pitt., Pittsburgh, PA April 2012-April 2014

Awards and Fellowships

USU College of Science Claude E. Zobell Graduate Scholarship (May 2018)
USU Dept. of Geology Outstanding PhD Student Award (April 2017)
USUSA Graduate Enhancement Award (April 2016)
American Geophysical Union annual meeting Outstanding Student Paper Award (Dec. 2015)
USU Presidential Doctoral Research Fellowship (August 2014-May 2018)
National Association of Geoscience Teachers/US Geological Survey Cooperative Summer Geosciences internship recipient (May 2014)
Research

PUBLICATIONS

*undergraduate student mentee


CONFERENCE ABSTRACTS

*undergraduate student mentee


**GRANT FUNDING**

- USU Dissertation Enhancement Award (July 2017; $10,000)
- USUSA Graduate Enhancement Award (April 2016; $4,000)
- USU Dept. of Geology J. Stewart Williams Field Scholarship (April 2016; $1,000)
- USU Dept. of Geology Undergraduate Field Assistant Award (April 2016; $1,000)
- Tobacco Root Geological Society Field Scholarship (April 2015; $500)
- Geological Society of America Graduate Student Research Grant (April 2015; $2,500)
- U. Pitt Undergraduate Research Award (March 2013; $1,000)
- Geological Society of America, Northeastern Section Undergraduate Research Award (March 2013; $1,000)

**FIELD, ANALYTICAL, AND SOFTWARE EXPERIENCE**

(U-Th)/He thermochronometry (hematite, apatite, zircon); zircon U-Pb, $^{14}$C, and $^{210}$Pb geochronology; optical and scanning electron microscopy; $\delta^{18}$O and $\delta^{13}$C stable isotope measurements; MATLAB programming; ArcGIS software; lake core collection; paleoseismic trenching

**Teaching**

**CLASSES TAUGHT**

- **Fall 2018**
  - GEO 4700, *Geologic Field Methods*, 3 credits, 8 students, USU

**TEACHING ASSISTANTSHIPS AND GUEST LECTURES**

- **Spring 2020**
  - Teaching Assistant, *Applied Geophysics*, 4 credits, 4 students, USU

- **Spring 2018**
  - Guest Lecturer (2x), *Low-Temperature Thermochronology Methods and Applications*, USU

**Service and Mentoring**

**UNDERGRADUATE MENTEES**

Kelsey Wetzel, USU, Laboratory research assistant August 2017-present

*Funded by a USU Dissertation Enhancement Award to R.G. McDermott*

Christopher Ammon, USU, Field and research assistant August 2016-Dec.2016

Undergraduate Research Project: “Thermotectonics of the Basin and Range Province, USA”

Jordan Jensen, USU, Field assistant in Kluane Lake National Park, YT, Canada June 2015

**OUTREACH**

Utah Public Radio, Logan, UT (Fall 2019)


Promontory School of Expeditionary Learning, Perry, UT (Spring 2016, Fall 2017, Fall 2019)

-earthquake and landslide hazards workshop, visit to USU scanning electron microscope facility, and field trip to the Wasatch Fault for grades 5-6

Geology of Ogden Canyon field trip (Spring 2018)

-designed and led field trip to introduce local church youth groups to geologic history of Utah, local hydrogeology, and landslide and earthquake hazards

USU-Dept. of Geology (Spring 2017, Spring 2018)

-participated in *Rock and Fossil Days* to teach citizens of Cache Valley about the rock record of earthquake processes

USU-Eastern Blanding Campus (Spring 2016)

-taught geochronology laboratory techniques to students from underrepresented populations in STEM fields as part of the *Native American Student Mentoring Program*

Century Elementary School, Bear River City, UT (Spring 2016)

-science fair judge

Western Pennsylvania School for the Deaf, Edgewood, PA (Fall 2013)

-designed and built educational rock display

**PROFESSIONAL SERVICE**

Graduate student representative to the USU Search Committee for Professor of Sedimentary Processes

USU Undergraduate Research Cooperative Opportunity review panel, July 2017 and 2019

Speaker and group leader, USU Undergraduate Research Fellow Mentoring Relationships Workshop, November 2019

EXTRACURRICULAR INVOLVEMENT
Graduate Student Representative to the Faculty, USU (2015-2016 academic year)
USU Photogrammetry Workstation set-up and preparation (Spring 2016)
U. Pitt. Geology Club Secretary (2013-2014 academic term)
Sigma Gamma Epsilon Earth Science honor society (2012-2014)

PROFESSIONAL ORGANIZATIONS
Geological Society of America
American Geophysical Union