Thermochronometric and textural evidence for seismicity via asperity flash heating on exhumed hematite fault mirrors, Wasatch fault zone, UT, USA

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Highlights:

- We present hematite microtextures and (U-Th)/He dates from fault mirrors.
- Results suggest localized > 1200 °C fault surface temperatures.
- Data patterns consistent with thermomechanical asperity flash heating model.
- Asperity flash heating may promote earthquakes along the Wasatch Front.

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 earthquakes ($M_w \approx$

-3 to 3) on Wasatch hematite fault mirrors. We suggest that similar thermal processes and resultant dynamic weakening may facilitate larger earthquakes.

Keywords:

Hematite (U-Th)/He thermochronometry, hematite microtextures, dynamic weakening, earthquakes, fault mirrors, Wasatch fault

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 through ~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 Material. Hematite aliquots were extracted using a portable rotary tool and fine point tweezers and selected to avoid tool marks. All aliquots are approximately 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 ($\sim 120^\circ$) 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 $\leq 1\text{-}20$ 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 laser 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 ($\pm 1\sigma$ standard deviation) with individual aliquot dates spanning 18.4 ± 0.6 Ma to 1.4 ± 0.2 Ma ($\pm 2\sigma$ 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).

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 prescreened aliquots. Of these 18 aliquots, 13 (72%) yield dates statistically younger at 1σ 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 ≥ 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 μm), 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 μm (Table S3). The (U-Th)/He T_c estimates for fault surface hematite and host rock apatite are ~ 95 -110 °C (Table S1) and ~ 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 $30^\circ\text{C}/\text{km}$; 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 ~ 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, $\leq 65\%$ 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 2 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).

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):

$$T_{fc} = T_{surf} + \frac{\mu_{fc}HV\sqrt{\beta}}{\rho C\sqrt{\pi\alpha}} \quad (1)$$

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

$$T_{surf} = \frac{\mu\sigma_n V t^*}{2\rho Ch} + T_{amb} = \frac{\tau D}{2\rho Ch} + T_{amb} \quad (2)$$

where t^* is the duration of slip, T_{amb} is ambient temperature, μ is the coefficient of friction, σ_n is the normal stress, τ 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:

$$T(z, t) = \frac{\tau V}{2\rho Ch} \left[t \left(1 - 2i^2 \operatorname{erfc} \frac{h-z}{\sqrt{4\alpha t}} - 2i^2 \operatorname{erfc} \frac{h+z}{\sqrt{4\alpha t}} \right) - (t - t^*) \left(1 - 2i^2 \operatorname{erfc} \frac{h-z}{\sqrt{4\alpha(t-t^*)}} - 2i^2 \operatorname{erfc} \frac{h+z}{\sqrt{4\alpha(t-t^*)}} \right) \right] + T_{amb} \quad (3)$$

where z is shear zone depth, t is time, and h is half width of the deforming zone. The term $i^2 \operatorname{erfc}(\delta)$ denotes the second integral of the complementary error function of δ . 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 μm clusters and rarely extend more than $\sim 10 \mu\text{m}$ 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 μ_{fc} evolves from 0.6 to 0.1, but is closer to 0.6, over the contact lifetime and thus assume an average μ_{fc} of 0.4. We set $H=2.7$ GPa, a typical value for hematite (Chicot et al., 2011). The terms ρ , C , α are set to 2700 kg/m³, 790 J/kg K, and 1.27×10^{-6} m²/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, $\sim 1\text{-}20$ μm (Figs. 2; S4), although we model contact diameters up to 50 μm . In equation (2), μ is assumed to be 0.1, consistent with the average μ 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 ≤ 2 km. Assuming a 30 °C/km geothermal gradient (Blackett, 2004) and lithostatic pressure yields T_{amb} and σ_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” τ 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 $\sim 13\%$ and $\sim 7\%$ 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_{surf} , which only contributes a temperature rise above T_{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.

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 to 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 $\tau = 4$ 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 T_c isotherm, and/or high coefficients of friction ($\mu=0.8$). 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 σ_n at the macroscopic scale is insufficient to induce measurable He loss. Effective σ_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 ($\sim >700$ -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 $\geq 900\text{--}1000\text{ }^{\circ}\text{C}$ (Spray, 1992; Rice, 2006; Goldsby and Tullis, 2011). Inferred paleoasperity dimensions and model simulations parameterized by our data indicate flash-heating temperatures $>360\text{--}1200\text{ }^{\circ}\text{C}$ (Fig. 5A inset), although measureable He loss by volume diffusion requires temperatures of at least $\sim 700\text{ }^{\circ}\text{C}$. 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 (M_w) 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., Kwiitek et al., 2011) indicate 0.3 and 30 m² slip patches likely accommodated single-event displacements of ~10-40 μ m and 2-4 mm, respectively, yielding $M_w = -3.4$ 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 M_w 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 ~5340 m², yielding $M_w = 2.6$. Reasonable M_w estimates for earthquakes accommodated on these fault surfaces are thus ~ -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 ≤ 2 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 ~ 500 -2500 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; Kwiitek 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 $>\sim 700$ -1200 °C at frictional contacts on the slipping surface and subsequent weakening likely enabled 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.

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FIGURE CAPTIONS

Fig. 1. Hematite fault surfaces, sample localities, and apatite (U-Th)/He dates for the Wasatch Fault zone, UT. (A, B) Field photographs of high-gloss, hematite-coated minor faults with local iridescence and slickenlines (A) and specular hematite below fault surface (B). (C) Southern Brigham City segment of WFZ. Red and yellow squares are hematite sample locations in this study and from Ault et al. (2015). Blue and green circles are apatite He sample locations from this study and Armstrong et al. (2004). Dates reported as mean and 1σ standard deviation. Base

map digital elevation model (DEM) is 0.5m LiDAR from Utah Geological Survey. Quaternary fault scarps (black lines) mapped from DEM. Inset shows extent of WFZ. Red square denotes study area.

Fig. 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.

Fig. 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 $\pm 1\sigma$ standard deviation).

Fig. 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 $\sim 200\ \mu\text{m}$ aliquot thickness.

Fig. 5. (A) Time-temperature histories for asperities of different diameter (d) at the fault surface ($z=0\ \text{mm}$). 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 $0.1\ \mu\text{m}$ domain for different asperity d values. Colors correspond to time-temperature curves in (A). Vertical dashed line denotes typical $\sim 200\ \mu\text{m}$ aliquot thickness.

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Figure 1.

Figure 2.

Figure 3.

Figure 4.

Figure 5.

SUPPLEMENTARY MATERIAL-MCDERMOTT ET AL. (2017)

1. Methods

Hematite was extracted from veins and fault surfaces using a dremel tool. Replicate aliquots were obtained by breaking samples into chips using fine-point tweezers and avoiding dremel tool-marked material. Aliquots were selected on the basis of uniform grain size and lack of other mineral phases. Hematite aliquots were dual-imaged using secondary and backscattered electrons on a Quanta 650 FEG scanning electron microscope at Utah State University's Microscopy Core Facility. For samples where aliquots were consumed during U-Th-He analyses, representative aliquots were mounted in 1" epoxy rounds, polished to 1 μm grit, and carbon coated. These were imaged at high-vacuum ($<5 \times 10^{-5}$ Torr). For samples imaged prior to (U-Th)/He dating, ~ 120 - 200 μm thick aliquots were mounted on double-sided copper sticky tape and imaged under low vacuum (0.2-0.9 Torr). A subset of these aliquots were loaded into Nb packets following imaging. General instrument operating conditions included 20-30 kV accelerating voltage, 65 nA current, ~ 8.4 mm working distance, and use of Large Field, Everhart-Thronley, and Concentric Backscatter detectors. Grain size measurements from SEM images use ImageJ software. We identify and categorize hematite as having polygonal, lobate, or subangular grain morphologies from visual inspection of SEM images (Fig. S3).

Aliquots were loaded into Nb packets and heated to temperatures and packet "glow" comparable to apatite for 8 minutes using a diode laser in an ultra-high vacuum gas extraction line at the University of Arizona Radiogenic Helium Dating Laboratory (ARHDL). Extracted He gas was spiked with ^3He , purified using cryogenic and gettering methods, and analyzed on a quadrupole

mass spectrometer. Single gas re-extracts were conducted at higher glow for 10 minutes and until negligible He was released compared to the prior heating step. The majority of samples required only one re-extract to extract all 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. Fish Canyon Tuff zircon was used as a standard to monitor chemistry and instrument performance. Blank-corrected (U-Th)/He dates were calculated with propagated analytical uncertainties from U, Th, and He measurements. No alpha-ejection correction was applied to the hematite He dates because all aliquots are sufficiently thick ($>120\text{ }\mu\text{m}$) such that He ejection is balanced by implantation (Figs. S2; S3; Farley and Flowers, 2012; Evenson et al., 2014).

Apatite was extracted from bedrock samples using standard magnetic and density separation techniques. Whole grains from mineral separates were examined with a Leica M165 C cross-polarized stereoscope at the Mineral Microscopy and Separation Laboratory at Utah State University. Target crystals were selected on the basis of morphology, clarity, and lack of inclusions. Grains were imaged, their dimensions measured, and subsequently loaded into Nb packets. U-Th-He analyses were conducted at the ARHDL. Apatite grains were laser-heated to $\sim 1065\text{ }^{\circ}\text{C}$ for three minutes with a diode laser without a gas re-extract. Extracted He gas was spiked with ^3He , purified using cryogenic and gettering methods, and analyzed on a quadrupole mass spectrometer. Degassed apatites were retrieved, spiked with a ^{233}U - ^{229}Th - ^{147}Nd - ^{42}Ca tracer, dissolved in HNO_3 , and analyzed on an Element 2 ICP-MS. Apatite dimensional mass was calculated from Ca measurements and stoichiometry and used to calculate parent U-Th-Sm concentrations (Guenther et al., 2016). Durango apatite was used as a standard. Blank-corrected

(U-Th-Sm)/He dates were calculated with propagated analytical uncertainties from U, Th, Sm, and He measurements. An alpha-ejection correction was applied using grain measurements and assuming apatite are unzoned with respect to U, Th, and Sm (Farley et al., 1996).

2. Thermomechanical and He diffusion model

2.1. He diffusion calculations

Shear zone temperature rise and subsequent heat decay results in thermally activated He diffusion from minerals within the shear zone. He diffusion can be described by an Arrhenius relationship as

$$\frac{D}{r^2} = \frac{D_0}{r^2} e^{\frac{-E_a}{RT}} \quad (\text{S.1})$$

where D is the diffusion coefficient, D₀ is a frequency factor, r is the diffusion length scale, E_a is the activation energy, R is the gas constant, and T is temperature. Any square pulse heating event of duration t yields a unique value of (D/r²)* Δt. The integrated thermal history can be represented as “reduced time” t_r (Fechtig and Kalbitzer, 1966):

$$t_r(T, t) = \frac{D_0}{r^2} \int_0^t e^{\frac{-E_a}{RT(t')}} dt' \quad (\text{S.2})$$

where the integrand is integrated from 0 to t for each time step and t' = Δt for the numerical integration. A spherical diffusion domain for each crystal is assumed, and fractional loss (F) for each time-temperature step can thus be approximated as (Fechtig and Kalbitzer, 1966):

$$F \approx \frac{6}{\pi^{3/2}} \sqrt{\pi^2 t_r} - \left(\frac{3}{\pi^2} \right) (\pi^2 t_r), \quad F \leq 0.85$$

$$F \approx 1 - \frac{6}{\pi^2} e^{-\pi^2 t_r}, \quad F \geq 0.85 \quad (\text{S.3})$$

For a series of time-temperature steps, equations (S.2)-(S.3) yield the cumulative fractional loss of He.

Equation (3) in the main text calculates shear zone thermal history as a function of depth, z . For the z - t calculation grid, Δz is set at equal increments and time steps from $t=0$ to $t=1000$ seconds are irregularly spaced to increase computational efficiency (model resolution is discussed in more detail in the following section). The time-temperature path unique to each z -coordinate within the shear zone are treated as a series of square pulses of temperature T and duration t . Following equation (S.2), t_r is calculated for each step. These values are used in conjunction with equation (S.3) to populate a z - F calculation grid. The cumulative F is then extracted for each z -coordinate to produce a composite depth-fractional He loss curve. Asperity temperatures, shear zone thermal history, and fraction He loss calculations are implemented in a code written for Octave, an open-source version of MATLAB. The full software package is available from the authors upon request.

2.2. Discussion of model sensitivity

Model-generated results are sensitive to the chosen variables, material constants, and calculation scheme and thus merit a brief discussion on how the choice of each of these influences the final model output. Because the result of interest to this study is fractional He loss, this discussion will focus on the impact of various parameter choices to those data.

Model runs at various spatial and temporal resolutions were conducted to evaluate the sensitivity of results to the number of steps over which a fractional He loss curve is calculated. Model results are insensitive to the spatial resolution used, but highly sensitive to the temporal resolution used.

For example, calculated fractional loss during the first 20 seconds of post-slip heat decay over 20 steps vs. 500, 1000, and 5000 steps, revealed that a coarse temperature resolution underestimates fractional loss by as much as 30%. Changes in fractional loss between simulations with 500, 1000, and 5000 time steps are modest at <1-3%. Model results are not dependent on the temporal resolution of the model past 20 s. All results presented herein are generated using a model with 1000 time steps during the first 20 s, 20 time steps from 20-100 s, and 30 time steps from 100-1000 s. The spatial resolution is set at 20 steps of $\Delta z=50 \mu\text{m}$.

We explore the potential for parameter variations in macroscopic shear heating calculations to induce measurable He loss from fault surface hematite (Fig. S5). These simulations vary velocity (V), displacement (D), and shear stress (τ). V of 0.1, 0.5, and 1 m/s and D of 1, 5, 6, 7, 8, 9, and 10 cm are considered. Values of τ are set at 2.65, 21.2 and 42.4 MPa, corresponding to depths of 1, 1.5, and 2 km (assuming lithostatic pressure) and friction coefficients (μ) of 0.1, 0.5, and 0.8, respectively. All other parameters are as previously described. All simulations yield negligible He loss from a hematite crystallite with a $0.1 \mu\text{m}$ radius, with the exception of those with $\tau=42.4 \text{ MPa}$ and $D > 5 \text{ cm}$ (Fig. S5). Peak temperature in these simulations is $\sim 600\text{-}1000^\circ\text{C}$ and results in 10-100% He loss from hematite (Fig. S5, inset). We note that these peak temperature estimates are highly sensitive to shear zone half-width (1 mm). We consider these mechanical scenarios an unlikely explanation for our data. Shear stresses of $\sim 40 \text{ MPa}$ were likely not attained on these surfaces at the depths implied by apatite and hematite He data patterns. Values of $\mu > 0.6$ could also be argued as being unlikely at seismic slip rates based on laboratory experiments (e.g., Di Toro et al., 2011). These simulations are also not consistent with observed data patterns. Fault-surface-average heating of this magnitude would result in spatially uniform hematite He dates and

high-temperature microtexture development across the fault surface. We observe the exact opposite of this in our samples.

He diffusion from hematite is dependent on time and temperature with longer durations of equal temperature pulse resulting in progressively higher degrees of fractional He loss. Model results will thus be sensitive to the temporal resolution of the model (e.g., Reiners, 2009) during both the co- and post-slip periods. This predicts that the use of equivalent τ to define an isothermal square pulse for a duration of t^* [e.g., equation (3), main text] to describe coseismic temperature rise, when the syn-slip temperature will increase more or less linearly, will overestimate fractional He loss at $z=0$. For example, the fractional loss from a diffusion domain with $0.1\ \mu\text{m}$ radius subjected to an isothermal pulse of $T=1300\ ^\circ\text{C}$ for a duration of 0.065 seconds is 1.0, while the fractional loss achieved by summing square pulses of 0.001 second duration and steadily increasing temperature to $1300\ ^\circ\text{C}$ is 0.99. However, this effect becomes more prominent at decreasing temperature. Temperatures of $900\ ^\circ\text{C}$ and $700\ ^\circ\text{C}$ result in an overestimate of 10-30% fractional loss, respectively, when assuming an isothermal square pulse of 0.065 s as opposed to a linearly increasing temperature over the same time interval. Fractional loss curves associated with asperity temperatures may thus be overestimates, with the severity of this overestimate corresponding to asperity size (i.e., temperature). Fractional loss curves for macroscopic, or average, fault surface heating may therefor also be overestimates. However, this does not change the conclusions drawn in the main text.

We also explore the effect of varying host rock thermal parameters, such as density (ρ), heat capacity (C), and thermal conductivity (and thus thermal diffusivity), within the envelope of other values reported for granitic host rock (Robertson, 1988). For example, varying ρ between 2700 and 2900 kg/m³, C between 2.6 and 2.7 J/kg K, and thermal conductivity between 2.6-2.7 W/m K results in only modest changes of a few percent to fractional loss curves in all simulations. Our model simulations and final interpretations are therefore not strongly sensitive to host rock conductive heat transport parameters.

We also tested the choice of diffusion kinetics on model outputs. Farley and Flowers (2012) report an E_a of 157 kJ/mol for He diffusion from hematite. Using this value of E_a and letting $D_0=2.2 \times 10^{-4}$ cm²/s (Evenson et al., 2014) results in ~20% less fractional loss than when $E_a=147.5$ kJ/mol, although this affect is only manifested in certain grain sizes. For example, flash heating at a 20 μ m asperity results in no change in the fractional loss curve for a grain of 0.1 μ m radius, but He loss from a grain with a 0.5 μ m radius decreases from ~90% (see Fig. 4, main text) to ~70%. For flash heating at a 10 μ m asperity, fractional loss within 200 μ m of the slip surface decreases from ~80% (see Fig. 5, main text) to ~60% for a 0.1 μ m diffusion domain. For macroscopic shear heating, the negligible He loss calculated for low- τ slip events (Fig. S5) is even less for these scenarios with higher E_a and negligibly different in simulations with lower E_a (e.g., 110 kJ/mol; Lippolt et al., 1993; Bähr et al., 1994). This suggests that the major conclusions and interpretations drawn in the main text are not sensitive to the choice of diffusion kinetics.

3. Earthquake moment magnitude calculation

Theoretical relationships describe the connection between rupture area and displacement during slip. The relative displacement distribution, D_{rel} , along an elastic crack of radius c is given by (Scholz, 2002):

$$D_{rel}(x, y) = \frac{24}{7\pi} \frac{\Delta\sigma}{G} [c^2 - (x^2 + y^2)]^{1/2} \quad (S.4)$$

$\Delta\sigma$ is the stress drop and G is the shear modulus. D_{rel} is half of the total displacement D . Assuming reduction in μ from 0.6 to 0.1 and σ_n of 40 MPa, consistent with our model simulations, yields $\Delta\sigma$ of 20 MPa. A typical value of G for crustal scale faults is 3×10^{10} N/m² (Hanks and Kanamori, 1979). Using these values, equation (S.4) can be solved for maximum displacement ($x, y=0$) assuming a rupture patch of area A , or rearranged to solve for A given an assumed displacement. From Hanks and Kanamori (1979), the moment magnitude (M_w) of an earthquake is given by

$$M_w = \frac{2}{3} \log(GDA) - 6.03 \quad (S.5)$$

where D and A are estimated from field observations and equation (S.4).

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Table S1. Hematite grain size measurements and closure temperature calculations

Sample	Count	r min (μm) ^a	r max (μm) ^a	T _c (°C) ^b
W14-HC-20 ^c	15	0.3	9.6	102-163
A13-6 ^c	25	0.06	2.5	79-137
W14-HC-18A	25	0.04	0.4	74-106
W14-HC-18C	19	0.04	0.3	74-102
W14-HC-23A	26	0.04	0.6	74-113
W14-HC-23B	25	0.02	0.3	65-102
W14-HC-23C	21	0.05	0.3	77-102
W15-8A	43	0.02	1.3	65-126
W15-8C	30	0.02	0.6	65-113
W15-11A	37	0.02	0.6	65-113
W15-11-2A	73	0.02	1.7	65-130
W15-11-2B	64	0.03	1.7	70-130
W15-16A	29	0.02	0.6	65-113
W15-16B	39	0.03	0.9	70-120
W15-16C	62	0.04	0.5	74-110

^agrain size. All measurements reported as subangular grain radius or plate half-width

^bestimated closure temperature assuming $E_a=147.5$ kJ/mol and $D_0=2.2\text{E-}4$ cm²/s

(Evenson et al., 2014), spherical diffusion geometry, diffusion lengthscale

equal to grain size (Bahr et al., 1994; Evenson et al., 2014),

and 10 °C/Myr cooling rate

^cspecularite vein samples

Table S2. Hematite (U-Th)/He data

	U (ng)	± 1s	Th (ng)	± 1s	Th/U	⁴ He (fmol)	± 1s	Date (Ma)	Error (Ma) ^a
SPECULARITE VEIN SAMPLES									
W14-HC-20 , 41.39190, -112.02316 ^b ; 1460 masl									
H1	0.38924	0.0056	0.025	0.0004	0.066	232.46	1.22	108.16	1.59
H2	0.20244	0.0029	0.018	0.0003	0.092	94.16	0.60	83.93	1.28
H3	0.48635	0.007	0.025	0.0004	0.053	240.74	1.19	90.05	1.32
H4	0.1515	0.0022	0.009	0.0002	0.059	59.77	0.31	71.80	1.06
A13-6 , 41.38026, -112.02020; 1500 masl									
H1	0.612	0.009	0.058	0.001	0.097	34.71	0.22	10.28	0.15
H2	1.253	0.018	0.066	0.001	0.054	73.61	0.37	10.77	0.16
H3	0.885	0.013	0.073	0.001	0.084	56.33	0.37	11.59	0.18
H4	1.136	0.016	0.085	0.001	0.077	59.77	0.40	9.59	0.15
H5	0.747	0.011	0.064	0.001	0.088	46.61	0.31	11.35	0.17
HIGH-GLOSS FAULT SURFACE SAMPLES									
W14-HC-23A , 41.39190, -112.02316 ; 1460 masl									
H1 ^c	0.003	0.000	0.011	0.000	3.325	0.29	0.12	9.19	3.80
H2	0.433	0.006	0.873	0.012	2.066	11.00	0.13	3.19	0.05
H3	0.157	0.002	0.578	0.008	3.775	4.90	0.09	3.09	0.06
H4	0.061	0.001	0.191	0.003	3.225	1.90	0.09	3.33	0.17
W14-HC-18A , 41.39190, -112.02316 ; 1460 masl									
H1	1.870	0.027	3.900	0.056	2.140	48.14	0.36	3.20	0.04
H2	3.360	0.048	8.338	0.119	2.546	82.90	0.53	2.88	0.03
H3	2.257	0.033	7.814	0.111	3.551	45.49	0.32	2.05	0.03
H4	0.769	0.011	1.758	0.025	2.345	16.41	0.16	2.57	0.04
W14-HC-18C , 41.39190, -112.02316 ; 1460 masl									
H1	0.080	0.001	0.111	0.002	1.423	0.90	0.06	1.56	0.11
H2	0.136	0.002	0.226	0.003	1.702	3.32	0.09	3.24	0.09
H3	0.085	0.001	0.093	0.001	1.124	0.83	0.06	1.44	0.10
H4	0.630	0.009	0.957	0.014	1.559	14.59	0.15	3.16	0.05
Samples with aliquots extracted from multiple isolated localities on surface W14-HC-23									
W14-HC-23B , 41.39190, -112.02316 ; 1460 masl									
H1	0.534	0.008	0.553	0.008	1.062	11.12	0.13	3.10	0.05
H2	0.312	0.005	0.264	0.004	0.866	5.12	0.09	2.53	0.06
H3	0.945	0.014	1.001	0.014	1.086	17.96	0.19	2.82	0.04
H4	1.054	0.015	0.959	0.014	0.933	20.20	0.19	2.93	0.04
W14-HC-23C , 41.39190, -112.02316 ; 1460 masl									
H1	0.658	0.009	0.421	0.006	0.656	15.45	0.19	3.79	0.07
H2	0.265	0.004	0.172	0.002	0.666	6.15	0.11	3.73	0.08
H3	2.200	0.032	1.570	0.022	0.732	42.07	0.30	3.04	0.04
H4	0.572	0.008	0.391	0.006	0.701	10.40	0.14	2.90	0.05

	U (ng)	± 1s	Th (ng)	± 1s	Th/U	⁴ He (fmol)	± 1s	Date (Ma)	Error (Ma)
HIGH-GLOSS FAULT SURFACE SAMPLES PRESCREENED WITH SEM PRIOR TO DATING									
W15-11A , 41.39190, -112.02316 ; 1460 masl									
H2	0.579	0.008	0.314	0.004	0.556	11.27	0.12	3.20	0.05
H4	0.815	0.012	0.425	0.006	0.534	19.45	0.16	3.94	0.06
H5	1.753	0.025	1.471	0.021	0.861	39.83	0.28	3.52	0.05
H6	0.359	0.005	0.221	0.003	0.632	8.50	0.09	3.84	0.06
Samples with aliquots extracted from multiple isolated localities on surface W15-8									
W15-8A , 41.39190, -112.02316 ; 1460 masl									
H4	0.887	0.013	1.638	0.024	1.896	18.20	0.13	2.65	0.03
H5	0.365	0.005	0.771	0.011	2.165	11.22	0.11	3.80	0.05
H6	0.646	0.009	1.157	0.017	1.836	13.24	0.13	2.67	0.04
H8	0.119	0.002	0.163	0.002	1.408	4.89	0.07	5.77	0.10
H10	0.247	0.004	0.343	0.005	1.426	6.85	0.10	3.88	0.07
W15-8C , 41.39190, -112.02316 ; 1460 masl									
H1	0.691	0.010	3.572	0.051	5.304	17.31	0.20	2.09	0.03
H4	0.462	0.007	1.045	0.015	2.319	7.24	0.12	1.89	0.04
H5	0.727	0.010	1.189	0.017	1.678	18.21	0.21	3.35	0.05
H7	0.355	0.005	1.529	0.022	4.412	11.09	0.14	2.87	0.05
H8	0.068	0.001	0.257	0.004	3.862	1.24	0.06	1.78	0.09
Samples with aliquots extracted from multiple isolated localities on surface W15-11-2									
W15-11-2A , 41.39190, -112.02316 ; 1460 masl									
H3	1.688	0.024	3.445	0.049	2.094	52.64	0.39	3.90	0.05
H5	2.083	0.030	5.912	0.084	2.912	70.26	0.50	3.74	0.05
H6	2.486	0.035	7.764	0.111	3.203	83.48	0.59	3.58	0.04
H7	2.668	0.038	11.186	0.159	4.302	134.97	0.88	4.71	0.06
H10	1.401	0.020	2.031	0.029	1.487	57.12	0.40	5.63	0.07
W15-11-2B , 41.39190, -112.02316 ; 1460 masl									
H4	0.702	0.010	0.917	0.013	1.339	26.82	0.24	5.41	0.08
H5	0.209	0.003	0.230	0.003	1.128	5.24	0.12	3.70	0.10
H6	0.439	0.006	0.313	0.004	0.732	11.30	0.13	4.09	0.07
H7	0.267	0.004	0.343	0.005	1.315	7.72	0.10	4.11	0.07
H8	0.046	0.001	0.116	0.002	2.573	1.02	0.06	2.56	0.15
Samples with aliquots extracted from multiple isolated localities on surface W15-16									
W15-16A , 41.39190, -112.02316 ; 1460 masl									
H4	0.501	0.007	0.690	0.010	1.414	8.61	0.12	2.41	0.04
H7	0.359	0.005	0.550	0.008	1.572	5.72	0.09	2.17	0.04
H8	1.121	0.016	1.451	0.021	1.328	19.12	0.18	2.42	0.03
H10	1.410	0.020	2.047	0.029	1.490	24.85	0.24	2.43	0.04
W15-16B , 41.39190, -112.02316 ; 1460 masl									
H1	0.997	0.014	1.414	0.020	1.455	34.60	0.29	4.82	0.07
H4	0.549	0.008	0.684	0.010	1.278	24.52	0.25	6.40	0.10
H5	0.301	0.004	0.431	0.006	1.466	9.69	0.14	4.46	0.08
H8	0.044	0.001	0.064	0.001	1.502	1.12	0.06	3.53	0.20
H9	0.134	0.002	0.254	0.004	1.936	6.00	0.11	5.72	0.12

W15-16C, 41.39190, -112.02316 ; 1460 masl

H1	0.837	0.012	1.469	0.021	1.801	18.73	0.19	2.93	0.04
H3	0.591	0.008	1.114	0.016	1.935	11.93	0.12	2.59	0.04
H7	0.664	0.009	1.165	0.017	1.799	14.97	0.16	2.95	0.04
H8	1.122	0.016	1.917	0.027	1.752	26.30	0.20	3.10	0.04
H10	0.137	0.002	0.249	0.004	1.867	3.04	0.07	2.88	0.07

^a1s propagated error from analytical uncertainties on U, Th, and He analyses.

^b Latitude and longitude reported in decimal degrees; WGS84

^cdiscarded due to high analytical error

Table S3. Apatite (U-Th)/He data

Ca-based			U	Th	[eU]	Sm	⁴ He			Raw Date	Corr Date	
Mass ^a (mg)	r ^b (μm)	l ^c (μm)	(ppm)	(ppm)	(ppm)	(ppm)	(nmol/g)	Ft ^d		(Ma)	(Ma)	Error ^e (Ma)
W14-T1 , 41.37266, -112.01365; 1512 masl												
A1	0.0015	41.8	121.0	30.063	7.245	31.765	167.343	0.403	0.67	2.34	3.52	0.22
A2	0.0009	28.0	122.3	24.991	7.735	26.808	214.226	0.392	0.56	2.69	4.83	0.53
A3	0.0006	28.1	95.8	34.793	10.470	37.253	182.756	0.469	0.54	2.32	4.31	0.52
A4	0.0013	35.9	138.1	20.135	5.078	21.329	163.480	0.399	0.64	3.44	5.41	0.38
A5	0.0011	35.6	114.4	29.234	8.417	31.212	178.394	0.480	0.62	2.84	4.57	0.30
A6	0.0008	33.7	112.3	29.982	8.064	31.877	246.239	0.442	0.61	2.55	4.21	0.36
										Mean ± st. dev.: 4.5 ± 0.6 Ma		

^amass calculated from [Ca] and stoichiometry (Guenther et al., 2016)^br-radius^cl-length^dFt-alpha ejection correction of Farley et al., 1996^e1σ propagated error from analytical uncertainties on U, Th, and He analyses

SUPPLEMENTAL INFORMATION-MCDERMOTT ET AL.

Figure S1: Hand sample and field photographs of hematite samples

W14-HC-20

A13-6

Figure S1 continued: Hand sample and field photographs of hematite samples

W14-HC-18

W14-HC-23 (float)

Figure S1 continued: Hand sample and field photographs of hematite samples
W15-8, -11, 11-2

W15-16

Figure S2: Hematite aliquot photographs and SEM images.

W14-HC-20

In SEM image B, dashed black lines delineate interpreted plate boundaries

Figure S2 continued: Hematite aliquot photographs and SEM images.

A13-6

Figure S2 continued: Hematite aliquot photographs and SEM images. Fault surface at top of photo in SEM images.

W14-HC-18A

Figure S2 continued: Hematite aliquot photographs and SEM images. Fault surface at top of photo in SEM images.

W14-HC-18C

Figure S2 continued: Hematite aliquot photographs and SEM images. Fault surface at top of photo in SEM images.

W14-HC-23A

Figure S2 continued: Hematite aliquot photographs and SEM images. Fault surface at top of photo in SEM images.

W14-HC-23B

Figure S2 continued: Hematite aliquot photographs and SEM images. Fault surface at top of photo in SEM images.

W14-HC-23C

Figure S3: Hematite aliquot photographs and SEM images from samples prescreened prior to (U-Th)/He analysis. Dashed white or black line denotes fault surface.

W15-8A

^alobate grain boundaries, red box highlights example grains.
^bpolygonal grain boundaries, blue box highlights example grains.

Figure S3 continued: Hematite aliquot photographs and SEM images from samples prescreened prior to (U-Th)/He analysis. Dashed white or black line denotes fault surface.

W15-8C

^alobate grain boundaries, red box highlights example grains.
^bpolygonal grain boundaries, blue box highlights example grains.

Figure S3 continued: Hematite aliquot photographs and SEM images from samples prescreened prior to (U-Th)/He analysis. Dashed white or black line denotes fault surface.
W15-11A

^alobate grain boundaries, red box highlights example grains.

Figure S3 continued: Hematite aliquot photographs and SEM images from samples prescreened prior to (U-Th)/He analysis. Dashed white or black line denotes fault surface.

W15-11-2A

^aglobular and lobate grain boundaries; indicated by red bounding box
^bpolygonal grain boundaries, blue box highlights example grains.

Figure S3 continued: Hematite aliquot photographs and SEM images from samples prescreened prior to (U-Th)/He analysis. Dashed white or black line denotes fault surface.
W15-11-2B

^alobate grain boundaries, red box highlights example grains.

Figure S3 continued: Hematite aliquot photographs and SEM images from samples prescreened prior to (U-Th)/He analysis. Dashed white or black line denotes fault surface.

W15-16A

^alobate grain boundaries, red box highlights example grains.

Figure S3 continued: Hematite aliquot photographs and SEM images from samples prescreened prior to (U-Th)/He analysis. Dashed white or black line denotes fault surface.
W15-16B

^alobate grain boundaries, red box highlights example grains.

Figure S3 continued: Hematite aliquot photographs and SEM images from samples prescreened prior to (U-Th)/He analysis. Dashed white or black line denotes fault surface.

W15-16C

^alobate grain boundaries, red box highlights example grains.
^bpolygonal grain boundaries, blue box highlights example grains.

W15-8C_H7

W15-8A_H8

Figure S4: Microtextural evidence of paleoasperity dimensions. Black or white dashed line denotes fault surface. Dashed red lines denote interpreted asperities. Note that these serve as minimum estimates of asperity diameter assuming approximately circular geometry.

W15-16C_H8

W15-11-2A_H3

Figure S4 continued: Microtextural evidence of paleoasperity dimensions. Black or white dashed line denotes fault surface. Red dashed lines denote interpreted asperities. Note that these serve as minimum estimates of asperity diameter assuming approximately circular geometry.

W15-11-2B_H8

W94-17E (Ault et al., 2015)

Figure S4 continued: Microtextural evidence of paleoasperity dimensions. Black or white dashed line denotes fault surface. Red dashed lines denote interpreted asperities. Note that these serve as minimum estimates of asperity diameter assuming approximately circular geometry.

Figure S5: Average He loss from macroscopic shear heating as a function of displacement (D), velocity (V), and shear stress (τ). He loss is average over upper 200 μm of shear zone. Inset shows average He loss for all simulations as a function of peak temperature. Scatter effects are result of different heating durations (controlled by D/V) in simulations with similar temperature rise.