Late Cretaceous-early Paleogene extensional ancestry of the Harcuvar and Buckskin-Rawhide metamorphic core complexes, western Arizona

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Key Points:

- Miocene tectonic exhumation at these core complexes was predated by a latest Cretaceous to early Paleocene extensional event.
- This earlier extension was driven by crustal heating and anatexis that triggered gravitational collapse of overthickened crust.
- Recognition of this earlier extension has important implications for models of core complex formation and western North America.

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Abstract

Metamorphic core complexes in the western North American Cordillera are commonly interpreted as the result of a single phase of large-magnitude extension during the middle to late Cenozoic. We present evidence that mylonitic shear zones in the Harcuvar and Buckskin-Rawhide core complexes in west-central Arizona also accommodated an earlier phase of extension during the Late Cretaceous to early Paleocene. Microstructural data indicate substantial top-NE mylonitization occurred at amphibolite-facies, and $^{40}Ar/^{39}Ar$ thermochronology documents post-tectonic footwall cooling to <500°C by the Paleocene to mid-Eocene. Amphibolite-facies mylonites are spatially associated with voluminous and variably deformed footwall leucogranites that were emplaced from ca. 74-64 Ma, and a late kinematic ca. 63 Ma dike indicates this phase of mylonitization had waned by the early Paleocene. Reconstruction of the footwall architecture indicates that this latest Cretaceous – early Paleocene deformation occurred within a NE-dipping extensional shear zone. The leucogranites were likely the result of crustal melting due to orogenic thickening, implying a model whereby crustal heating triggered gravitational collapse of overthickened crust. Other tectonic processes, such as the Laramide underplating of Orocopia Schist or mantle delamination, may have also contributed to this episode of orogenic extension. Miocene large-magnitude extension was superimposed on this older shear zone and had similar kinematics, suggesting that the location and geometry of Miocene extension was strongly influenced by tectonic inheritance. We speculate that other Cordilleran core complexes also experienced a more complex and polyphase extensional history than previously recognized, but in many cases the evidence may be obscured by later Miocene overprinting.

1 Introduction

Metamorphic core complexes in the western North American Cordillera represent sites of large-magnitude extension where mylonitic mid-crustal rocks are juxtaposed against brittlely-deformed upper crustal rocks along a gently-dipping normal (detachment) fault. Core complexes are typically interpreted as a fundamentally distinct mode of crustal extension (e.g. Wernicke, 1985; Davis and Reynolds, 1989; Buck, 1991) due to the high magnitudes and rates of slip, the inferred slip of the detachment fault at low angles, and the exhumation of mylonitic mid-crustal rocks, among other factors. Studies of core complexes have provided important insight into key aspects of continental extension such as the initial geometry of detachment faults (e.g., John and Foster, 1993; Wong and Gans, 2008), the magnitude and rate of detachment fault slip (e.g., Foster and John, 1999; Prior et al., 2016), the mechanics of low-angle normal faults (e.g., Axen, 1992; Selverstone et al., 2012), the structural relationship between detachment faults and mylonites (e.g., Lister and Davis, 1989; Singleton and Mosher, 2012), the rheology of the middle crust (e.g. Hacker et al., 1992; Behr and Platt, 2011), and the role of lower crustal flow in extension (e.g., Gans, 1987; Block and Royden, 1990; McKenzie et al., 2000).

Most models for Cordilleran core complexes interpret them as the product of a single phase of middle to late-Cenozoic extension, where footwall mylonites represent the mid-crustal roots of coeval detachment fault systems (e.g., Wernicke, 1981; Davis et al., 1986; Lister and Davis, 1989; Spencer and Reynolds, 1991). Although it is indisputable that many core complexes in the central and southern Basin and Range experienced a phase of mylonitization during mid-Cenozoic exhumation (e.g., Reynolds et al., 1986; Foster and John, 1999; Wells et al., 2000; Wong and Gans, 2008; Singleton and Mosher, 2012; Zuza et al., 2019; Gottardi et al. 2020), some workers have argued that core complex mylonitization locally predates mid-Cenozoic exhumation and instead records Late Cretaceous extension (John and Musaka, 1990; Applegate and Hodges, 1995; Wong and Gans, 2009; Beyene, 2011) or other crustal flow of unknown tectonic significance (Ducea et al., 2020). If significant footwall mylonitization in some core complexes predated mid-Cenozoic extension, this would raise significant questions about the role of ductile deformation in the formation of core complexes, the tectonic significance of this older deformation, and the potential role of reactivation of these older fabrics in controlling the nature and geometry of mid-Cenozoic extensional shear zones and detachment faults. However, overprinting during Cenozoic deformation at many core complexes has often made it difficult to assess the presence and tectonic significance of older footwall mylonitization.

Here we argue that the Harcuvar and Buckskin-Rawhide core complexes in west-central Arizona (Figure 1) record clear evidence of extensive Late Cretaceous to early Paleogene mylonitization that is a distinctly older event than the large-magnitude Miocene extensional event for which Cordilleran core complexes are most well known. Moreover, we believe this early mylonitization records a substantial Late Cretaceous extensional event that immediately post-dated crustal thickening and partial crustal melting. These conclusions are based on

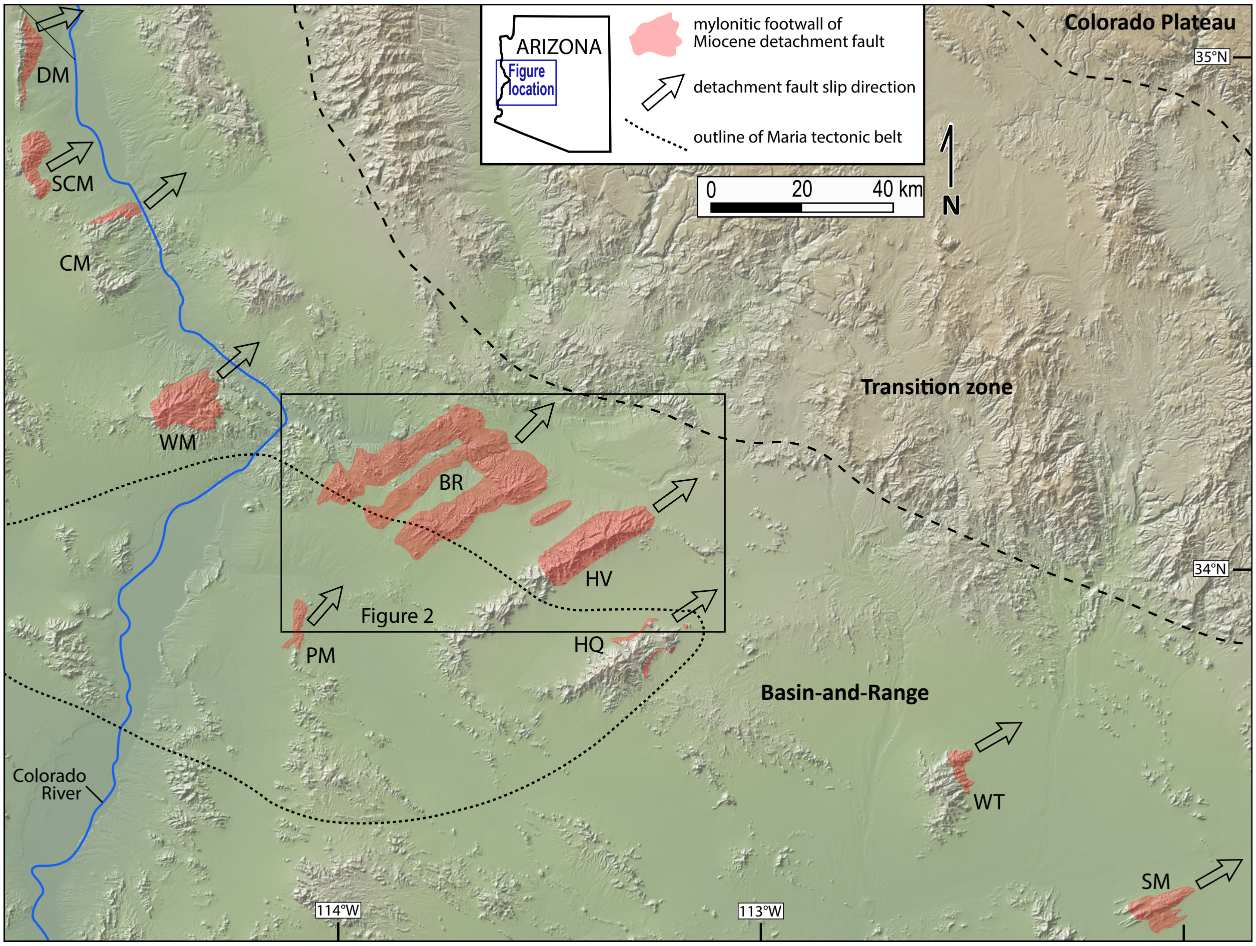

Figure 1. Shaded relief map from Singleton et al. (2018) showing the locations of metamorphic core complexes in southeastern California and western and central Arizona. Mylonitic footwall locations are shown in red and the major detachment fault slip directions are shown by the arrows. The dotted line encompasses the Mesozoic Maria fold-and-thrust belt (after Spencer and Reynolds, 1990). Location names for the mylonite distribution and detachment slip direction: SM—South Mountains, WT—White Tank Mountains, HQ—Harquahala Mountains, HV—Harcuvar Mountains, BR—Buckskin-Rawhide Mountains , PM—Plomosa Mountains, WM—Whipple Mountains, CM—Chemehuevi Mountains, SCM—Sacramento Mountains, DM—Dead Mountains.

geologic mapping, microstructural and electron backscatter diffraction (EBSD) analyses, Ti-in-quartz thermometry, and $^{40}Ar/^{39}Ar$ and U-Pb geochronology. These results demonstrate a composite extensional origin for this belt of core complexes and indicate that single-phase models of core complex formation should be reexamined. Evidence for widespread Late

Cretaceous extension in Cordilleran core complexes also has important implications for understanding the geodynamic evolution of western North America.

2 Geologic background

The Harcuvar and Buckskin-Rawhide Mountains, along with the adjacent Whipple and Harquahala Mountains, form a belt of core complexes within the lower Colorado River extensional corridor (CREC) in eastern California and western Arizona (Figure 2). These core complexes are also located within or adjacent to the Maria fold-and-thrust belt, a zone of Cretaceous basement-involved crustal shortening that was dominantly S- to SW-vergent (e.g., Spencer and Reynolds, 1990). In several ranges across this region (e.g., the Harcuvar, Harquahala, Granite Wash, and Dome Rock Mountains) major Cretaceous thrust faults are cut by Late Cretaceous (ca. 86–70 Ma) granitoids, which in turn are heterogeneously strained (e.g. Rehrig and Reynolds, 1980; Richard, 1988; Laubach et al., 1989; Reynolds and Spencer, 1993; Boettcher et al., 2002). Cawood et al. (2022) interpreted that ductile thrusting in the southernmost part of the Maria fold-and-thrust belt occurred somewhat later at ca. 68–65 Ma. These results place important constraints on the timing of Cretaceous crustal shortening in the region. The total magnitude of crustal shortening across the belt is largely unconstrained, although Chapman et al. (2020) recently applied geochemical proxies to estimate that Late Cretaceous crustal thicknesses may have reached up to 57 ± 12 km in western and southern Arizona. At the end of Cretaceous shortening, accretionary wedge sediments were underplated beneath the Maria fold-and-thrust belt during low-angle subduction of the Farallon plate as the Pelona-Orocopia-Rand Schist (Haxel et al., 2014; Strickland et al. 2018), with schist emplacement occuring by ca. 70 Ma (Jacobson et al., 2017; Seymour et al., 2018).

While most research on the Maria fold-and-thrust belt has focused on contractional structures, growing evidence suggests that Late Cretaceous to early Paleogene NE-directed extension may have also occurred regionally, following the cessation of contraction. Laramide-age mylonitization associated with NE-directed extension has been recognized in several ranges in the CREC and adjacent areas, including the Dome Rock Mountains (Boettcher and Mosher, 1998), Little Maria Mountains (Ballard and Ballard, 1990), Big Maria Mountains (Flansburg et al., 2021), Iron Mountains (Wells et al., 2002), New York Mountains (Wells et al., 2005), and Granite Mountains (Salem, 2009). How widespread this event was and whether it also impacted

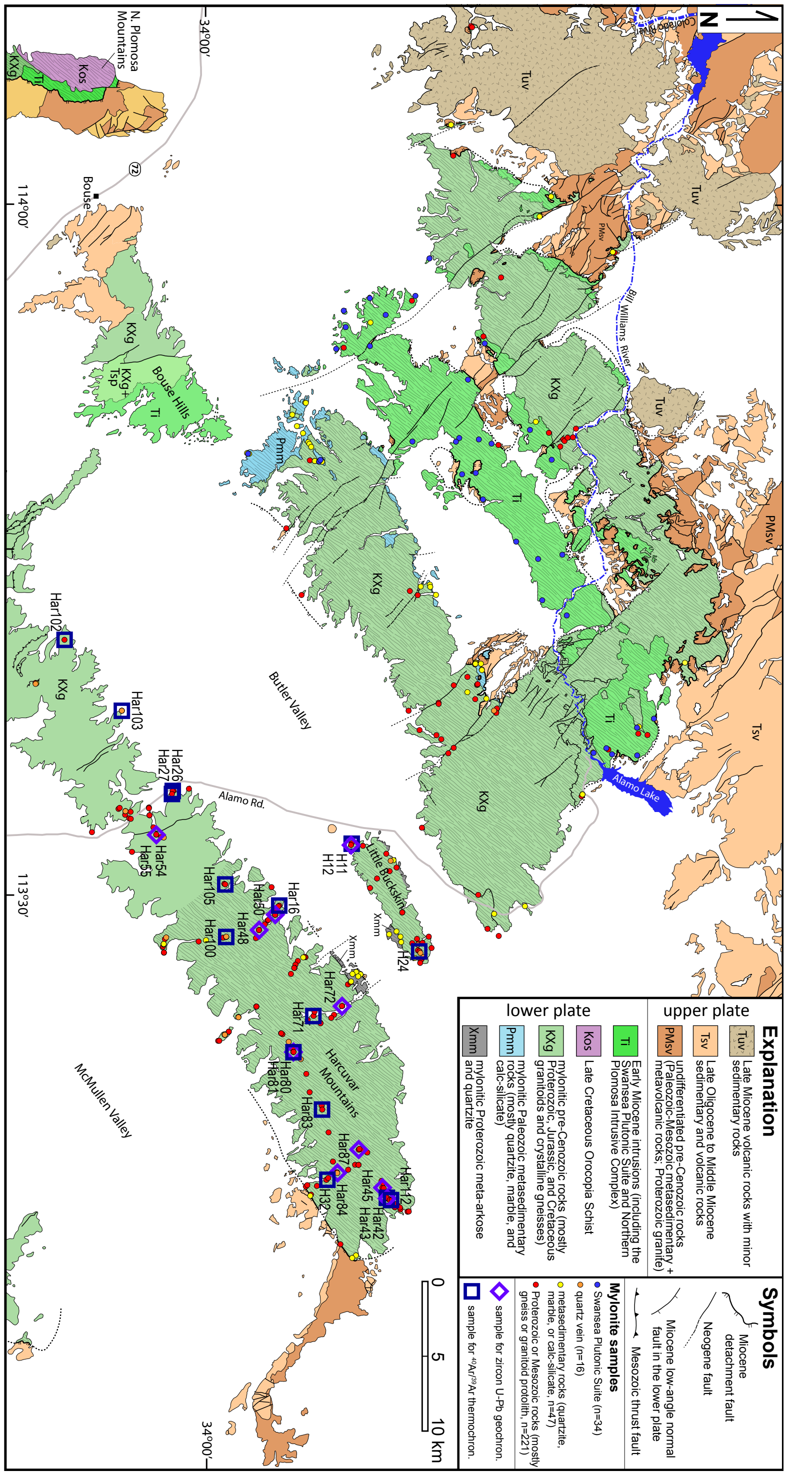

Figure 2. Simplified geologic map of the Buckskin-Rawhide and Harcuvar metamorphic core complexes (after Bryant, 1995; Spencer and Reynolds, 1989). The map also shows the location of samples for petrographic observations (small dots), U-Pb geochronology (diamonds) and $^{40}Ar/^{39}Ar$ thermochronology (squares).

core complexes within the CREC remains unclear, although John (1987) and John and Musaka (1990) suggest that most mylonitization in the footwall of the Chemehuevi detachment fault records top-to-the-NE-directed shearing during the Late Cretaceous (ca. 90–68 Ma).

The Harcuvar and Buckskin-Rawhide Mountains are dominated by variably mylonitized footwall rocks that include Proterozoic and Mesozoic crystalline gneisses, Late Cretaceous granitoids of the Tank Pass Plutonic Suite, Early Miocene granitoids of the Swansea Plutonic Suite, and minor pre-Cenozoic metasedimentary rocks (Bryant, 1995). The footwalls of these two core complexes are bound by one or more regional low-angle detachment faults that experienced tens of kilometers of top NE-directed extensional slip (e.g., Spencer and Reynolds, 1991; Singleton et al., 2014). The total slip across the detachment fault system in the Harcuvar Mountains is estimated to be ~45–50 km based on the correlation of distinct Jurassic clasts in an upper plate megabreccia to their likely footwall source and other lines of evidence (Reynolds and Spencer, 1985; Spencer and Reynolds, 1991; Prior et al., 2016). Extension may have begun in the late Oligocene based on the timing of some basin-fill deposits (e.g. Lucchitta and Suneson, 1993; 1996), but the main phase of detachment fault slip initiated at ca. 21 Ma and continued until ca. 12 Ma (e.g., Carter et al., 2004; Singleton et al., 2014; Prior et al., 2016, 2018).

Footwall fabrics in the Harcuvar and Buckskin-Rawhide Mountains are exposed for up to 35 km in the extension direction (Fig. 2) and are dominated by LS- or L>S-mylonitic tectonites with NE-SW-trending stretching lineations that trend parallel to the detachment fault slip direction. It is clear that significant lower-plate mylonitization occurred during mid-Cenozoic extension based on the presence of mylonitic fabrics in Early Miocene granitoids both locally (Bryant and Wooden, 2008; Singleton and Mosher, 2012) and within nearby core complexes such as the Whipple, Chemehuevi, and northern Plomosa Mountains (e.g., Anderson, 1988; LaForge et al., 2016; Gans and Gentry, 2016; Strickland et al., 2018). The similarity in the geometry and top-to-the-NE kinematics of mylonitic fabrics to the detachment faults further supports models that view these mylonites as the mid-crustal roots of mid-Cenozoic brittle detachment faulting (e.g., Richard et al., 1990; Spencer and Reynolds, 1991; Behr and Platt, 2011; Singleton and Mosher, 2012). This model has often led to the presumption that most or all lower-plate mylonitization of these core complexes is Miocene in age. However, other workers have argued that mylonites with similar fabric geometries and kinematics in the footwall of the Chemehuevi (John, 1987; John and Musaka, 1990), and Riverside (Lyle, 1982) detachment faults are instead Mesozoic in age, which raises the possibility that at least some of the footwall fabrics in the Harcuvar and Buckskin-Rawhide ranges formed prior to mid-Cenozoic extension. Given this uncertainty, assessing the age and significance of footwall fabrics in the Harcuvar and Buckskin-Rawhide ranges is critical to understand the tectonic development of these core complexes.

3 Results

3.1 Footwall fabrics

Footwall rocks in the northeastern ~30–35 km of the Harcuvar and Buckskin-Rawhide Mountains are dominantly well foliated and lineated mylonites (L-S tectonites; Figure 3), although parts of the Early Miocene Swansea Plutonic Suite are characterized by L>S mylonitic fabrics suggestive of constrictional strain (Singleton and Mosher, 2012). Northeast of the mylonitic front, footwall rocks of all ages and lithologies are are variably mylonitic, but locally Late Cretaceous granitoids of the Tank Pass Plutonic Suite are weakly strained, and steeply-dipping gneissic fabrics are exposed beneath metasedimentary mylonites in the southwestern Buckskin Mountains (Singleton et al., 2018). Footwall foliations are generally subhorizontal but dip gently northwest or southeast on the flanks of the ranges, broadly mimicking the corrugations in the bounding detachment fault. These corrugations define fold axes that plunge gently NE or SW, subparallel to the brittle slip direction (Singleton, 2015; Singleton et al., 2019). Lineations in mylonites are typically defined by quartz ribbons, streaks of mica, and/or aligned feldspar porphyroclasts with recrystallized tails. In most areas mylonitic lineations plunge gently NE-SW with a mean trend and plunge of 234°/ 03° in the Harcuvar Mountains, 038°/03° in the Little Buckskin Mountains, 225°/03° in the Ives Peak corrugation in the southern Buckskin Mountains, and 221°/13° in the Clara Peak and Planet Peak corrugations of the Buckskin-Rawhide Mountains (Figure 4).

3.2 Microstructural analysis and deformation conditions

We documented microstructures in >300 oriented petrographic thin sections from across the Buckskin-Rawhide, Little Buckskin, and Harcuvar Mountains (Fig. 2) to evaluate deformation conditions associated with footwall mylonitization (see Supplementary Table S1). Thin sections were cut perpendicular to the mylonitic foliation and parallel to the stretching lineation, and microstructural observations include mineralogy, kinematic indicators, and quartz

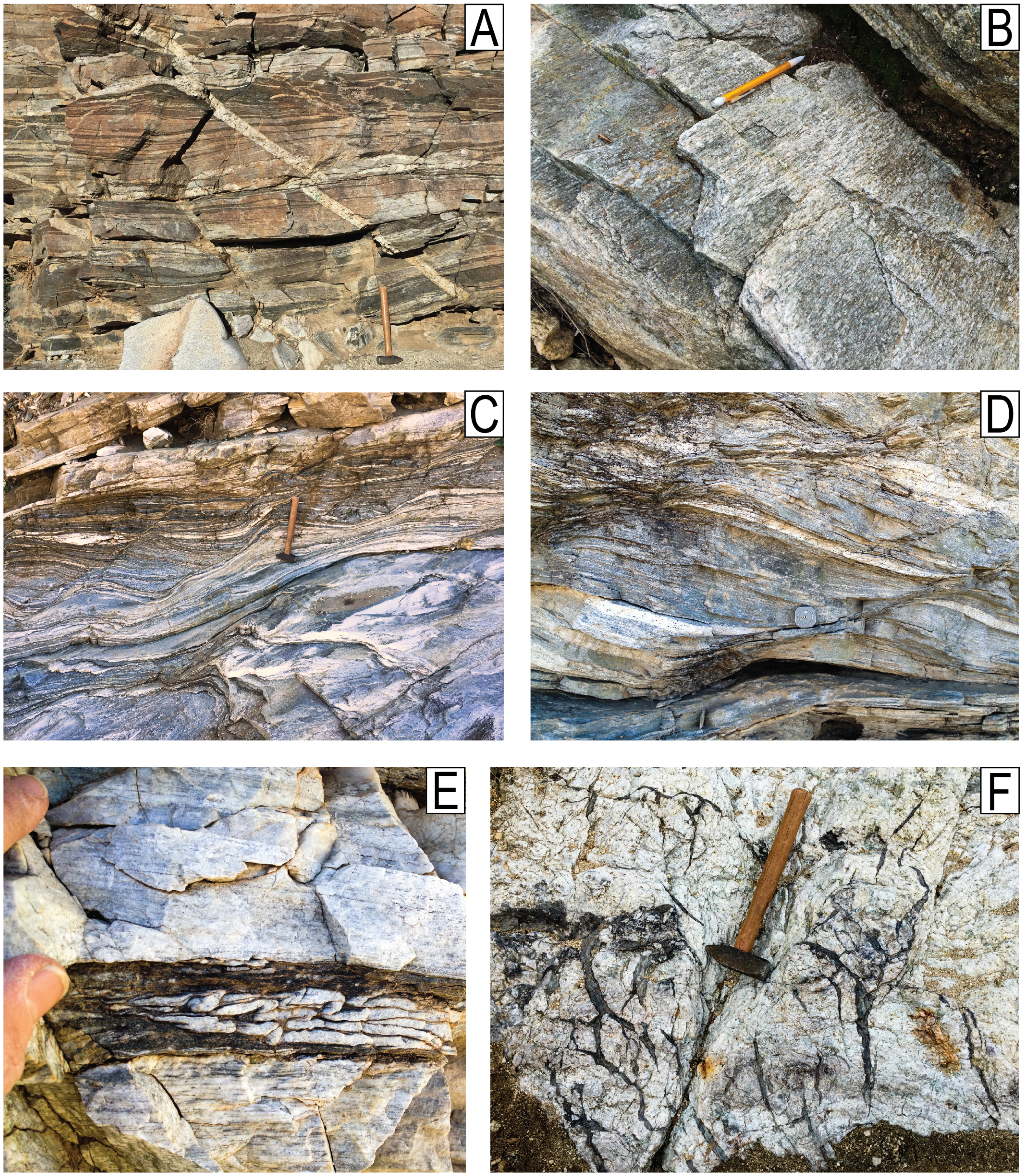

Figure 3. Field photographs from the footwall rocks in the Harcuvar-Buckskin Mountains. A) Gneissic fabric near the mylonitic front east of Cunningham Pass cut by a pegmatite dike, which is displaced by discrete biotite-rich shear zone. B) Leucogranite mylonite along the northwest flank of the Harcuvar Mountains near Burnt Well. Pencil parallels NE-SW trending stretching lineation. C) Top-NE (top-left) shear bands in mylonitic gneiss along the SE flank of the Harcuvar Mountains near Bullard Peak. D) Top-NE (top-left) shear bands in mylonitic gneiss from Miller Wash, eastern Harcuvar Mountains. E) Leucogranite ultramylonite with cm-scale NE-vergent folds (top-right) in the Little Buckskin Mountains. E) Pseudotachylyte (dark material) in leucogranite along a subdetachment fault near Burnt Well in the central Harcuvar Mountains. This fault separates greenschist-facies mylonites in Proterozoic metasedimentary rocks (above) from amphibolite-facies mylonites in Late Cretaceous leucogranite (below).

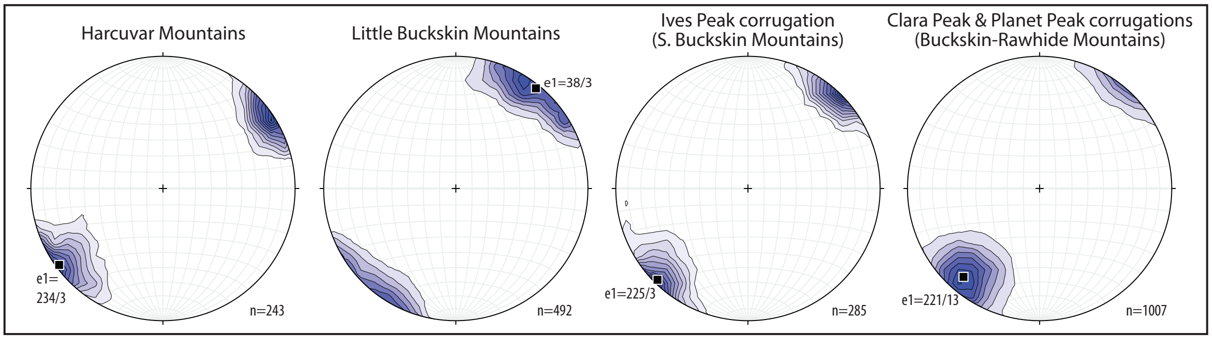

Figure 4. Contoured stereonet plots of mylonitic footwall lineations across different regions within the study area. The shallowly NE-SW plunging lineation direction is highly consistent across the study area.

and feldspar deformation/recrystallization mechanisms. We also estimated dynamically recrystallized quartz grain size for 141 thin sections of pre-Miocene rocks, which we compare to quartz grain sizes of Swansea Plutonic Suite mylonites previously analyzed by Singleton and Mosher (2012). Grain sizes were estimated by tracing ≥50 (average ~86) well-defined recrystallized grains from photomicrographs of relatively pure quartz domains and converting grain areas to an equivalent spherical diameter. For samples with variable grain sizes (typically associated with larger grains produced by grain boundary migration recrystallization) we strived to capture the full range of grain sizes, but mean grain size estimates in these samples have large standard deviations and are used primarily for relative comparison purposes (see Supplementary Figure S1).

Microstructures in footwall mylonites record a wide range of textures and interpreted deformation conditions. In nearly all quartzofeldspathic samples, quartz has undergone dynamic recrystallization and crystal-plastic flow, whereas feldspar records variable degrees of dynamic recrystallization and brittle fracturing. We organized samples into 4 categories based on characteristic quartz and feldspar microstructures that have been correlated with general deformation conditions (e.g., Passchier and Trouw, 2005), with category 1 representing the lowest temperature/highest stress conditions and category 4 representing the highest temperature/lowest stress conditions. Mylonites in all deformation categories are characterized by thinly-spaced foliation, penetrative stretching lineations, and grain-size reduction primarily via dynamic recrystallization.

Category 1 mylonites are characterized by quartz ribbons with incomplete recrystallization and <25 μm recrystallized grain sizes, commonly associated with small

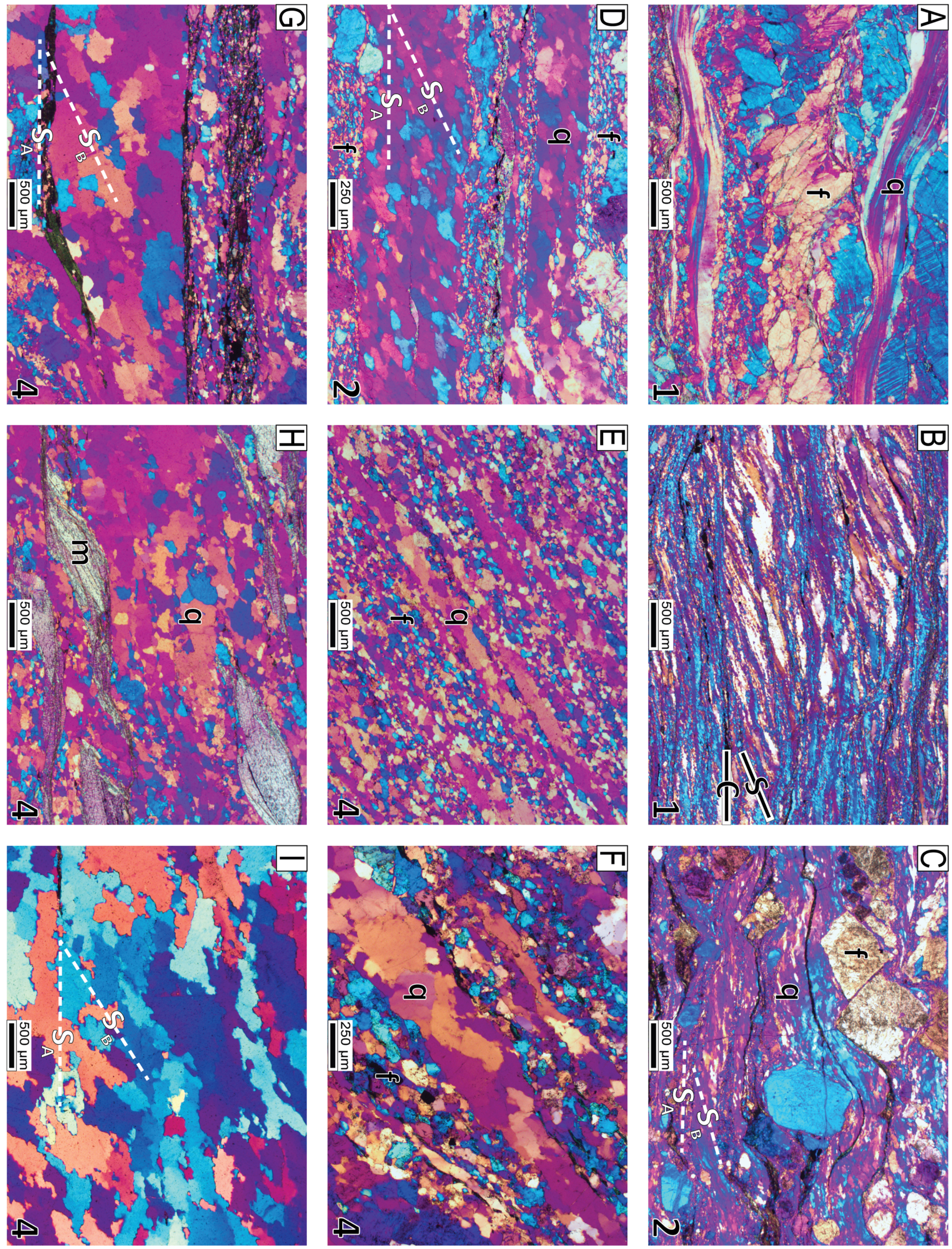

Figure 5. Photomicrographs of mylonites from the Buckskin-Rawhide and Harcuvar metamorphic core complexes. All photomicrographs are from X:Z thins sections in cross polarized light with the gypsum plate inserted, and the northeast side of the macroscopic lineation is on the right. Deformation conditions categories (1-4) are listed in the lower right (see text for details). A) Leucogranite from near the mylonitic front in the central Harcuvar Mountains (H3). Feldspar porphyroclasts (f) record brittle fracturing and cataclasis, whereas quartz ribbons (q) record dislocation creep with very minor BLG. B) Quartzite mylonite from the northwestern flank of the Harcuvar

Mountains near Burnt Well (15-JB3). Quartz records SGR+BLG; S-C fabric records top-NE shear; deformation category 1. C) Swansea Plutonic Suite in the central Buckskin Mountains (5-9). Feldspar (f) records brittle fracturing and minor BLG, whereas quartz records SGR and a grain shape foliation (S_B) that is oblique to the macroscopic foliation (S_A; top-NE shear). D) Leucogranite from the central Buckskin Mountains (4-636); feldspar is deformed by fracturing and BLG (f), whereas quartz is recrystallized primarily via SGR (q) with an oblique grain shape fabric (top-NE shear). E) Leucogranite from the eastern Buckskin Mountains (H20); nearly complete recrystallization of feldspar (f) primarily via SGR and quartz (q) via GBM. F) Leucogranite from the Little Buckskin Mountains (LB-154); feldspar records SGR; quartz records GBM. G) Leucogranite from the Little Buckskin Mountains (LB-190) with quartz GBM and an oblique grain shape fabric (top-NE shear). H) Leucogranite from the Little Buckskin Mountains with quartz GBM (q) and muscovite fish (m; top-NE shear). I) quartz vein from the Little Buckskin Mountains (LB-69); quartz records GBM and a grain shape fabric (S_B) that is oblique to the macroscopic foliation (S_A; top-NE shear).

subgrains and grain boundaries with irregular bulges or sutures (Figure 5), suggesting a combination of subgrain rotation and bulging recrystallization. Feldspar in these samples is dominated by brittle fracturing or cataclasis with minor dynamic recrystallization of <10 μm grains rimming porphyroclasts. Chloritization is common, and fresh biotite is rare. All non-quartzofeldspathic lithologies, predominately calcite-rich metasedimentary mylonites, also fall within this category due to their fine grain size. Category 1 metasedimentary mylonites are common <100 m below the bounding detachment system and were locally sheared through the brittle-plastic transition (Singleton et al., 2018).

In category 2 mylonites, quartz exhibits straight grain boundaries and relatively uniform grain sizes of ~30–70 μm that are similar to subgrains (Fig. 5), suggesting subgrain rotation recrystallization. Dynamic recrystallization of feldspar is more common than in category 1 samples, although porphyroclasts are still typically fractured. Feldspar subgrains and subgrain rotation recrystallization are rare, and overall chloritization is less abundant than in category 1 samples.

Category 4 mylonites are characterized by average quartz grain sizes between 80 and 250 μm with variable size distributions and irregular (amoeboid-like) grain boundaries (Fig. 5), suggesting a dominance of fast grain boundary migration recrystallization. Feldspar in these samples has undergone pervasive dynamic recrystallization into polygonal grains with undulatory extinction and subgrains, suggesting subgrain rotation recrystallization. Feldspar in category 4 granitoid ultramylonites is locally completely recrystallized, and chlorite is rare to absent. Category 3 mylonites have mixed features from category 2 and category 4 mylonites and may either represent an intermediate between these two categories or a lower-temperature overprint of category 4 mylonites.

3.3 Kinematics

The majority of the mylonite samples record a clear microstructural sense of shear. Dynamically recrystallized quartz grain shape fabrics oblique to foliation and C' shear bands are the most abundant shear sense indicators, and S-C fabrics, mica fish, and asymmetric porphyroclasts are also common (Fig. 5). These kinematic indicators consistently indicate a top-NE sense of shear across the study area. Of the 250 oriented thin sections evaluated for shear sense, ~86% record top-NE shear, 13% have symmetric structures or an unclear sense of shear, and 1% record top-SW shear. These microstructural kinematics are consistent with dozens of field observations supporting a dominance of top-NE shear (Fig. 3), which applies to all footwall lithologies and mylonite categories. The clearest top-SW kinematic indicators are from discrete shear zones located near the mylonitic front, which matches observations near the mylonitic front in other Arizona core complexes (e.g., Reynolds and Lister, 1990; Singleton et al., 2019), where antithetic shears have been interpreted to accommodate arching of the footwall during late-stage mylonitization (Reynolds and Lister, 1990).

3.4 Quartz crystallographic preferred orientation analyses

Crystallographic orientation patterns were determined with electron backscatter diffraction (EBSD) at Colgate University using a JEOL JSM6360LV scanning electron microscope with an Oxford Nordlys EBSD detector and processed using the HKL Channel 5 software and the MTEX Matlab toolbox (Bachmann et al., 2010). Step size was variable based on sample grain size but ranged from 5–30 μm. Crystallographic axes in pole figures were reduced to show one point per grain using a misorientation of 10° as a grain boundary threshold.

EBSD analyses of dynamically recrystallized quartz in category 3-4 mylonites reveal a strong CPO with *c*-axis fabrics typically defining a clear Y-axis strain maxima or patterns intermediate between Y-axis maxima and single girdle patterns (Figure 6). These patterns are consistent with inferred deformation temperatures above 500°C (Law, 2014). Based on these patterns, prism <*c*> slip did not play a significant role as a slip system during deformation. *C*-axis and *a*-axis fabric asymmetries, where present, are consistent with the top-NE sense of shear inferred from petrographic observations. Category 1-2 mylonites also have a strongly developed

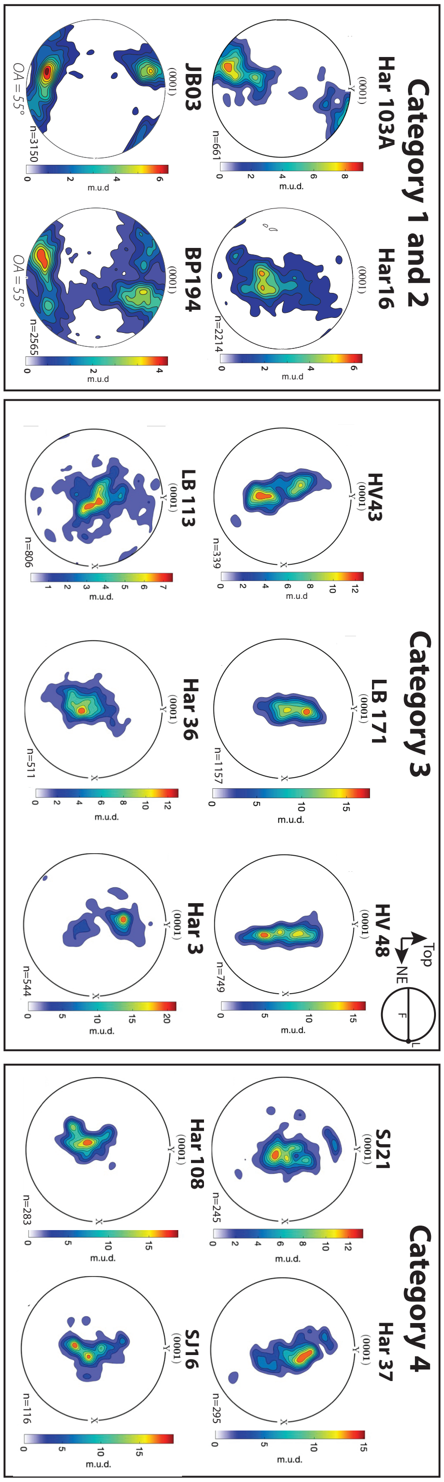

Figure 6. Crystallographic preferred orientation plots of quartz c-axes for representative mylonitic samples within different deformation categories. Sample orientations are shown perpendicular to foliation (F) and parallel to lineation (L) with top as up and NE to the right. Category 1-2 samples typically show cross girdle or single girdle c-axes patterns, while category 3-4 samples have c-axis fabrics typically defining a clear Y-axis strain maxima or patterns intermediate between Y-axis maxima and single girdle patterns. Sense of shear based on pattern asymmetry is top-NE where present.

CPO with distinctive *c*-axes patterns that typically form a cross-girdle or less commonly a single girdle pattern, which is consistent with dynamic recrystallization via BLG and SGR mechanisms at lower temperatures (Stipp et al., 2002b; Faleiros et al., 2010). Two of the category 1-2 mylonites yield *c*-axis cross-girdles with opening angles (OA) of ~55° (Fig. 6), which implies deformation temperatures of 425 ± 50°C (Faleiros et al., 2016).

3.5 Lithologic and spatial patterns of deformation conditions

When viewed across the entire study area, mylonite samples from the Buckskin-Rawhide and Harcuvar core complexes are evenly distributed across the 4 deformation conditions categories, with category 4 samples (~28%) being slightly more common than the other categories (Figure 7). However, the deformation conditions of mylonitization are strongly correlated with lithology. Category 1 samples are dominantly (~80%) metasedimentary mylonites derived from Proterozoic to Paleozoic quartzite and marble, while category 2 samples are typically (~71%) Early Miocene Swansea Plutonic Suite mylonites. The vast majority of category 3 and 4 samples (~85%) are leucogranite mylonites that are derived from Late Cretaceous plutons.

The deformation categories of mylonitization also show strong spatial patterns. Category 1 mylonites are most common in metasedimentary rocks found along the flanks of the footwall corrugations just beneath the detachment fault (Figure 8, Singleton et al., 2018). Some category 1 mylonites are also located near the mylonitic front at the southwestern part of the footwall. Category 2 mylonites are most common in the central part of the Buckskin-Rawhide footwall, where Swansea Plutonic Suite mylonites are prevalent. Category 3 and 4 mylonites are typically located in the interior parts of the Harcuvar, Little Buckskin Mountains, and southern Buckskin Mountains footwall, which also correspond to where Late Cretaceous leucogranite is most common.

To determine how mylonitization conditions vary geometrically and structurally with respect to the detachment fault system, we evaluated structures and samples from several transects that cross the flanks of the footwall corrugations (Figure 9). These transects include the northwest flank of the Ives Peak corrugation near Lincoln Ranch (previously mapped by Singleton et al., 2014), the southeastern flank of the Little Buckskin Mountains corrugation (previously mapped by Singleton, 2011), and the northwest and southeast flanks of the Harcuvar

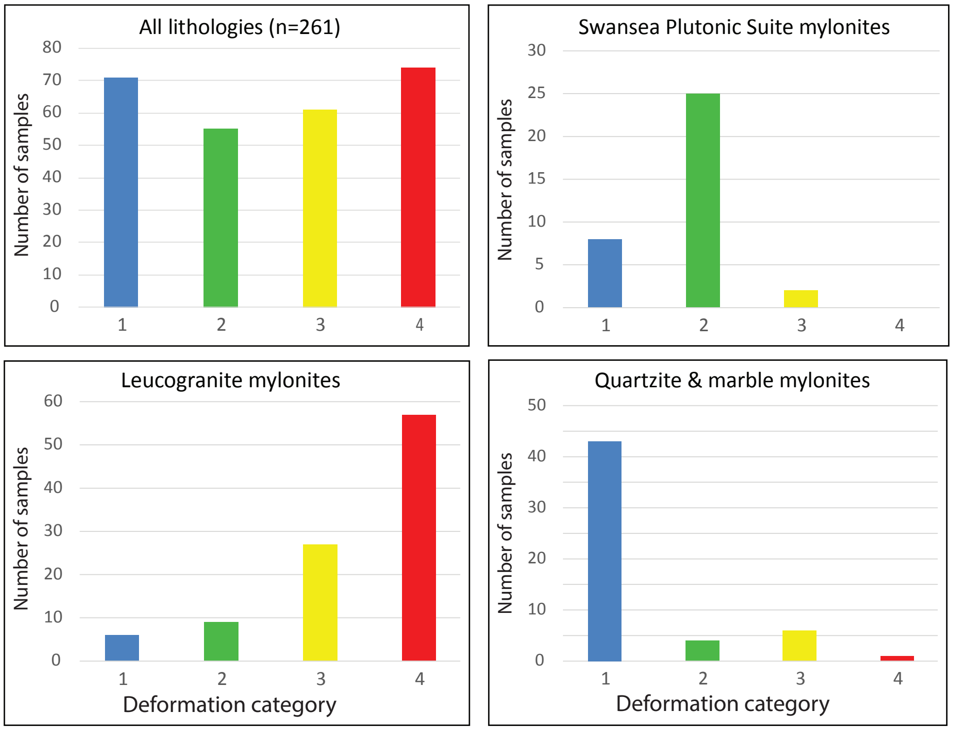

Figure 7. Histograms of the deformation categories (1-4) for mylonites within the study area, organized by lithology. While mylonites within the study area as a whole are equally spread across the deformation categories, there is a strong lithologic influence. Cretaceous leucogranite mylonites dominate the 3-4 deformation categories, Miocene Swansea plutonic suite mylonites are predominantly category 2, and metasedimentary mylonites are predominantly category 1.

Mountains corrugation (new mapping in this study). In each of these areas, category 1 mylonites are present just beneath the detachment system, while category 3-4 mylonites are present several hundred meters below the detachment system.

Along the northwest flank of the Ives Peak corrugation, a 50–100 m-thick section of marble, calc-silicate, and quartzite parallel the gently-NW-dipping detachment fault (Fig. 9). These metasedimentary rocks consistently record top-NE sense of shear and category 1 deformation conditions, and the marble maintains coherent mylonitic fabrics up to ~0.4 m below the Buckskin detachment fault principal slip plane (Singleton et al., 2018). The metasedimentary mylonite zone overlies crystalline mylonites along a sheared contact. The crystalline mylonites consist primarily of orthogneiss with abundant Late Cretaceous leucogranite layers that also record top-NE-directed shear and with category 3-4 deformation characteristics. The contact corresponds to an abrupt change in quartz deformation from bulging and subgrain rotation recrystallization and ~14–25 µm mean grain sizes in the metasedimentary mylonites to grain

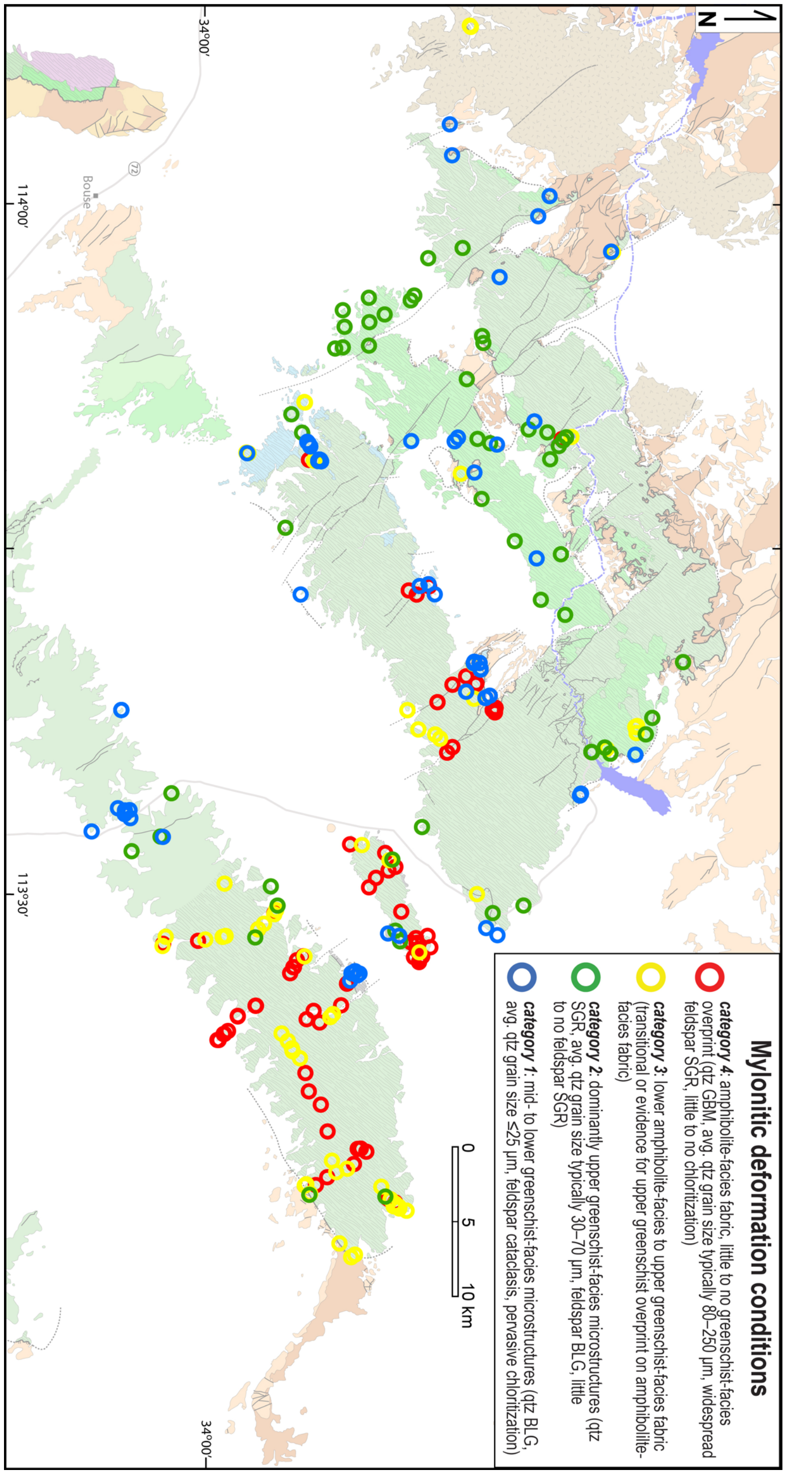

Figure 8. Map showing the spatial distribution of deformation categories across the study area. Category 1 mylonites are most common in metasedimentary rocks found along the flanks of the footwall corrugations just beneath the detachment fault and near the mylonitic front at the southwestern part of the footwall. Category 2 mylonites are common in the central part of the Buckskin-Rawhide footwall, where Swansea Plutonic Suite mylonites are prevalent. Category 3 and 4 mylonites are typically located in the interior parts of the Harcuvar, Little Buckskin Mountains, and southern Buckskin Mountains footwall where Late Cretaceous leucogranite is most common. See Fig. 2 for an explanation of geologic units.

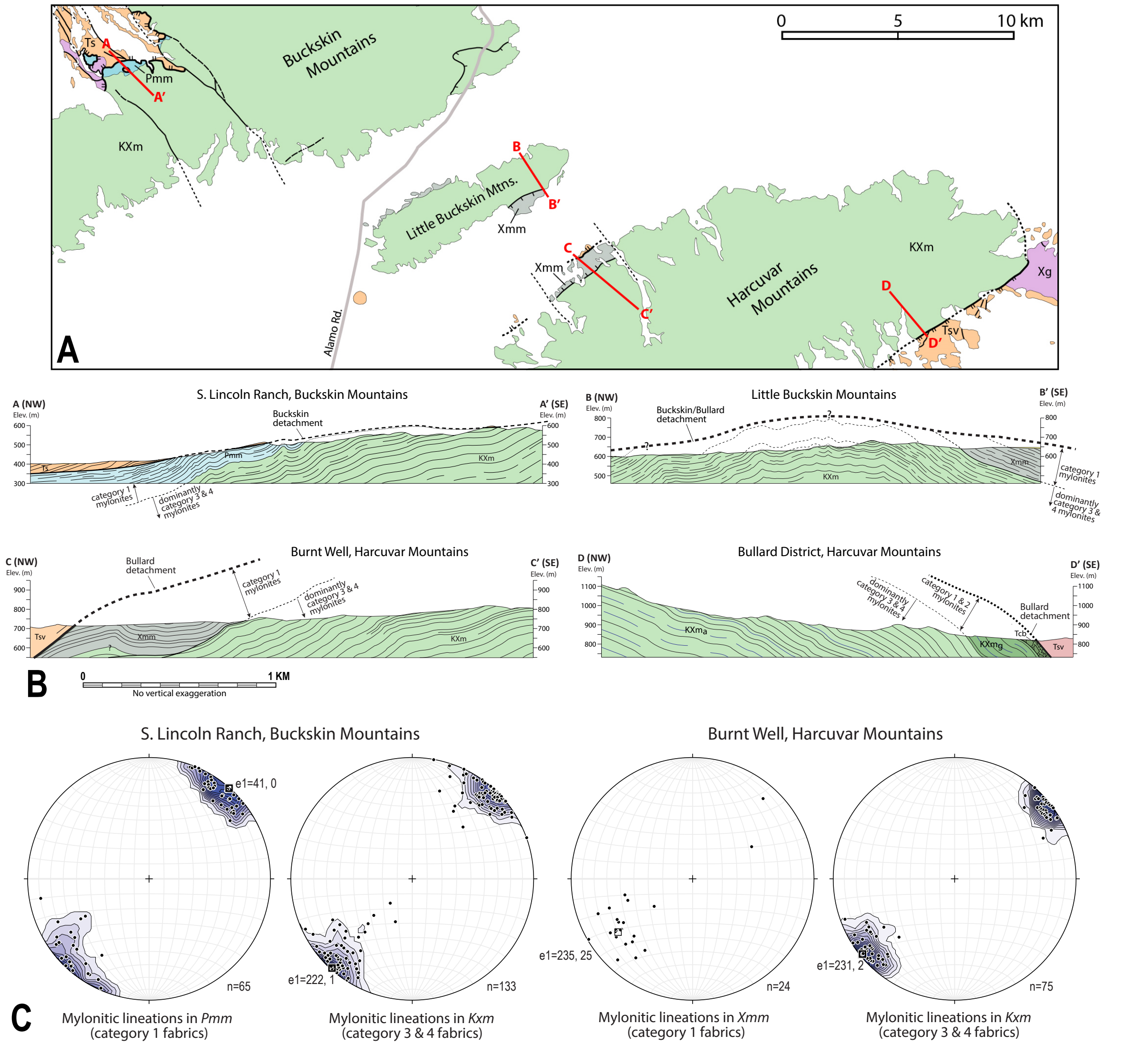

Figure 9. Geologic map (A) and detailed cross sections (B) in the South Lincoln Ranch, Buckskin Mountains (A-A'), Little Buckskin Mountains (B-B'), the Burnt Well locality (C-C') and the Bullard district (D-D') in the Harcuvar Mountains. The cross sections highlight that category 1-2 mylonites are typically located within several hundred meters below the detachment fault and commonly within metasedimentary units. Category 3-4 mylonites are located below this zone and are commonly developed in Cretaceous leucogranites and other crystalline basement rocks. Stereonet plots of lineations (C) highlight that lineation directions are indistinguishable within these different categories of mylonites.

boundary migration and subgrain rotation recrystallization and~60–100 µm mean grain sizes in the crystalline mylonites.

Along the southeast flank of the Little Buckskin Mountains corrugation and northwest flank of the Harcuvar Mountains corrugation near Burnt Well, similar spatial patterns in mylonitization occur (Fig. 9). In these areas, pervasively chloritized mylonites derived from

Mesoproterozoic meta-arkose and quartzite are present at the top of the footwall. These category 1-2 mylonites consistently record top-NE shear subparallel to the slip direction of the bounding detachment fault. This zone of metasedimentary mylonites is up to ~250 m thick and is juxtaposed against crystalline mylonite along a brittle fault that parallels the overlying detachment fault. Along the northwest flank of the Harcuvar Mountains this brittle footwall fault preserves pseudotachylyte veins (Fig. 3), demonstrating that the fault slipped seismically. Crystalline mylonites below this fault are dominantly Late Cretaceous leucogranite that consistently record category 3-4 deformation. Quartz in the metasedimentary mylonites has undergone subgrain rotation and bulging recrystallization with mean grain sizes of 10–50 μm, whereas quartz in the structurally lower leucogranites primarily records grain boundary migration recrystallization and mean grain sizes of 75–300 μm. As with the Ives Peak corrugation, this transition is abrupt and corresponds to the lithologic change from metasedimentary to crystalline lithology. Similar patterns are also observed along the southeast flank of the Harcuvar Mountains corrugation near the Bullard mineral district, where SE-dipping mylonites primarily derived from leucogranite parallel the detachment fault (Fig. 9). Near the top of the lower plate these mylonites are chloritically altered and record quartz subgrain rotation recrystallization with ~30–50 μm mean grain sizes. Chlorite alteration decreases and quartz grain size increases towards deeper structural levels in the footwall, and ≥200–250 m below the detachment fault fabrics are dominated by quartz grain boundary migration recrystallization (mean grain size >100 μm) and feldspar subgrain rotation recrystallization.

Based on these transects, it is clear that category 1-2 mylonites are concentrated within a ≤250 meter-thick carapace at the top of the footwall, whereas category 3-4 mylonites dominate at deeper structural levels. We did not observe a structural base of the category 3-4 mylonites, and based on cross sections, these mylonitic fabrics are likely >1 km thick (Fig. 9). Despite the notable differences in mylonitic deformation conditions preserved in the footwall, fabric orientations within these different zones are remarkably consistent. Where a clear boundary is present between the different category fabrics near Lincoln Ranch and Burnt Well, there is no statistical difference in lineation trend between the different category fabrics (Fig. 9), and a top-NE sense of shear is consistent throughout.

3.6 Ti-in-quartz thermometry

While deformation mechanisms and CPO patterns can provide important constraints on deformation temperatures, these estimates can be influenced by strain rate and other factors (e.g., Law, 2014). TitaniQ thermobarometry, which relies on the temperature-dependent substitution of Ti^{4+} for Si^{4+} in the quartz unit cell, can provide more direct constraints on deformation conditions (Wark and Watson, 2006; Thomas et al., 2010). This approach has been applied to estimate the *P–T* conditions of deformation in quartz-rich rocks (e.g., Kohn and Northrup, 2009; Behr and Platt, 2011; Grujic et al., 2011; Kidder et al., 2013, 2018; Nachlas et al., 2014; Bestmann and Pennacchioni, 2015; Cross et al., 2015), and experiments by Nachlas et al. (2018) demonstrated that dynamic recrystallization of quartz re-equilibrated Ti concentrations to reflect pressure-temperature conditions of deformation.

We conducted Ti-in-qtz thermometry on a subset of samples (n=9) focused on category 4 mylonitic Late Cretaceous leucogranites and quartz veins in order to provide additional constraints on the mylonitization temperatures. We focused on category 4 mylonites because Ti re-equilibration in quartz is mostly likely to occur in mylonites that experienced grain boundary migration recrystallization (Gruijic et al., 2011). Ti concentrations were determined using a Cameca 6f secondary ion mass spectrometer (SIMS) at Arizona State University. Unknowns were calibrated against three Ti-doped synthetic silica glass samples with known Ti concentrations of 0 ppm, 100 ppm, and 500 ppm (Gallagher and Bromiley, 2013). The use of pure silica standards avoids the ~30% bias introduced by calibrating against the common non-matrix matched standards such as the NIST 610, 612, and 614 glasses (Behr et al., 2010). During this analytical session, NIST 612 was analyzed and a NIST soda-lime vs. silica glass standards bias consistent with values reported by Behr et al. (2011) was confirmed. Reported error values only account for uncertainty in the Ti concentration measurements. Individual Ti spot analyses were averaged to calculate a single temperature estimate for each sample. We used the geothermobarometric calibration of Thomas et al. (2010) to convert Ti concentrations into temperatures.

Important unknowns in this calibration are the pressure and TiO_2 activity (a_{TiO2}). The minimum pressure to reach sufficient temperatures to produce extensive mylonitization of quartzofeldspathic rocks, assuming a reasonable 30°C/km geothermal gradient, is 3 kbar. However, the extensive grain boundary migration recrystallization in quartz (Stipp et al., 2002a;

Faleiros et al., 2010) and subgrain rotation recrystallization of feldspar (e.g., Fitz Gerald and Stunitz, 1993 and references therein) recorded by category 4 quartzofeldspathic mylonites require amphibolite-facies conditions (>500°C), suggesting a pressure range of 4–6 kbar is more appropriate for calculations on category 4 samples. This pressure range is reasonable given preliminary thermobarometry from the Harcuvar Mountains, which indicates that Late Cretaceous metamorphism occurred at 6–10 kbar pressures (Walsh et al., 2016), suggesting that the footwall had been deeply buried. We calculate temperatures using a pressure range of 4–6 kbar (Table 1), with calculated temperatures shifting by ~15°C per kbar under these conditions.

The a_{TiO2} can be difficult to estimate, with most studies assuming values of 0.5-1.0, the general range given for most igneous to metapelitic rocks (e.g., Ghent and Stout, 1984). However, Gruijic et al. (2011) demonstrated that these values systematically underestimated the temperatures of quartz deformation in the well–constrained Tonale shear zone and instead suggested a_{TiO2} of 0.2–0.3 is more appropriate for quartz-dominated samples. Furthermore, Thomas and Watson (2012) used melt inclusion compositions of the Bishop Tuff (Wallace et al., 1999) to calculate a_{TiO2} of ~0.23 using the MELTS program and a_{TiO2} of 0.15 using the Rhyolite–MELTS program, suggesting lower values may be more appropriate for felsic samples such as the leucogranite mylonites in our study. The absence of rutile as an oxide mineral in these samples also supports the use of a lower a_{TiO2} value. As such, we used $a_{TiO2} = 0.3$ in these calculations for our granite and quartz vein samples. Assuming a higher Ti activity of 0.7 would lower calculated temperatures by ~60°C.

For all analyzed category 3 and 4 samples, average Ti concentrations range from 0.5 – 13.5 ppm (mean = 6.7 ppm, Table 1) with a corresponding calculated temperature range of 377–588°C (mean = 512°C) at 6 kbar or 345–546°C (mean =474°C) at 4 kbar (Figure 10). Notably, 7 of the 8 samples in these categories yielded calculated temperatures of >500°C within error at 6 kbar, with one anomalously low temperature of 377 ± 13°C (from a quartz vein sample). The range of calculated temperatures for category 3 and 4 samples did not appear to be controlled by lithology. Although one quartz vein yielded the lowest calculated temperature, two other quartz veins yielded relatively high temperatures. All of the category 4 granitic samples yielded temperatures >500°C within error at 6 kbar. The single category 2 granite sample yielded an average Ti concentration of 5.9 ± 1.9 ppm with a calculated temperature of 520 ± 20°C at 6 kbar (482 ± 10°C at 4 kbar), similar to the category 3 and 4 mylonite samples.

Table 1. Ti-In-Quartz Analyses

Sample name	Description	Deformation category	# of analyzed spots	Average Ti (ppm) ± 1σ	Calculated temp. (°C) at 4 kbar*	Calculated temp. (°C) at 6 kbar*
JSH-36	Granitoid	2	7	5.9 ± 1.9	482 ± 10	520 ± 20
HV-48	Quartz vein	3	8	0.5 ± 0.2	345 ± 12	377 ± 13
HV-69	Granitoid	4	9	13.5 ± 2.3	546 ± 12	588 ± 13
15-JB-96	Granitoid	4	7	10.1 ± 1.4	523 ± 10	564 ± 11
LB-12X	Granitoid	4	7	6.6 ± 0.9	491 ± 9	531 ± 10
LB-197	Granitoid	4	7	2.6 ± 2.5	411 ± 56	447 ± 59
HAR38	Granitoid	4	6	4.8 ± 2.3	463 ± 35	501 ± 36
HV-34	Quartz vein	4	7	9.5 ± 0.4	518 ± 3	559 ± 4
EM-2-27	Quartz vein	4	7	5.9 ± 0.2	484 ± 3	522 ± 3

*Calculated temperatures assume $a_{TiO2} = 0.3$

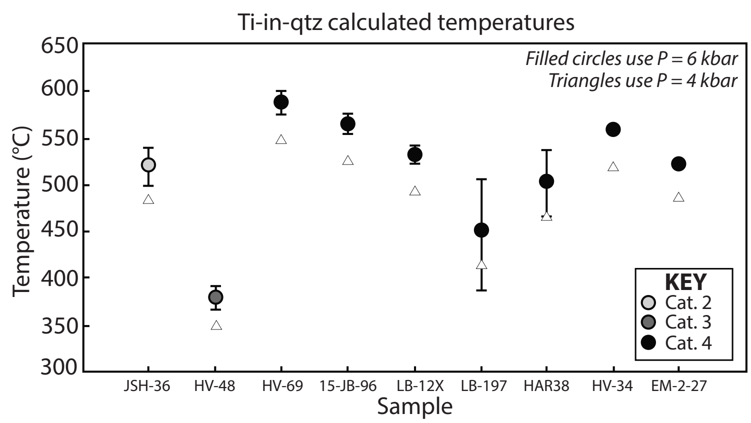

Figure 10. Calculated Ti-in-qtz temperatures for analyzed mylonites organized by deformation category. Filled circles assume pressure = 6 kbar and triangles = 4 kbar ($a_{TiO2} = 0.3$ in both cases). These results show that a significant majority of the samples yield calculated temperatures >500°C within error at 6 kbar pressure and >475°C at 4 kbar. Where no error bars are visible, the error is smaller than the symbol size.

3.7 $^{40}Ar/^{39}Ar$ thermochronology

The microstructural, EBSD, and Ti-in-quartz results all suggest that significant portions of the Buckskin-Rawhide and Harcuvar core complexes footwall underwent mylonitization at relatively high temperatures (amphibolite grade). This raises the question of whether the high temperature mylonitization was part of Miocene extension or a distinctly earlier event. We conducted new $^{40}Ar/^{39}Ar$ analyses on hornblende and biotite (Figure 11) from across the study area to provide new insights into the high temperature thermal history of the footwall and constrain the timing of the high temperature mylonitization. $^{40}Ar/^{39}Ar$ analyses were conducted

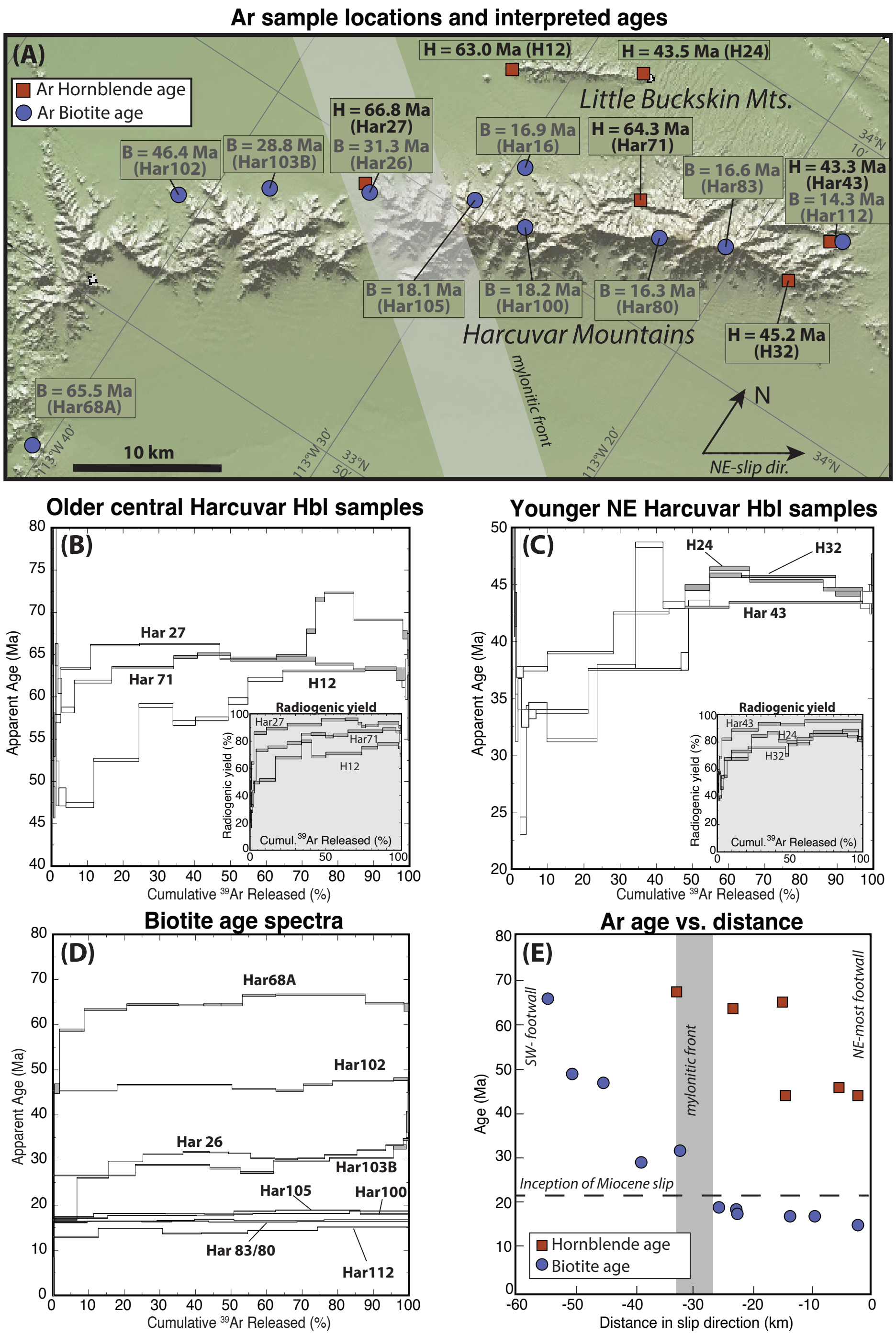

Figure 11. (A) Map of $^{40}Ar/^{39}Ar$ analyses and preferred ages of hornblende and biotite samples from the Harcuvar and Little Buckskin Mountains. (B) $^{40}Ar/^{39}Ar$ age spectra and radiogenic yields for older hornblende samples from the central footwall. Most age spectra show a climbing pattern with a flatter segment at moderate to high temperature steps. See text for additional details. (C) Age spectra for the younger hornblende samples from the northeastern footwall. These spectra also show a climbing pattern but flatten for the last half of the analyzed gas at high temperature steps. (D) $^{40}Ar/^{39}Ar$ age spectra for biotite samples from across the footwall. These spectra are flatter than the hornblende samples and yield readily interpretable ages, although show small variations in step ages. (E) Plot of interpreted hornblend and biotite $^{40}Ar/^{39}Ar$ age versus distance in the slip direction. Ages generally young in the hanging wall slip direction (NE). Hornblende ages are all significantly older than the age of Miocene

detachment faulting (ca. 21 Ma). Biotite ages to the southwest of the mylonitic front are also older than the age of detachment faulting but are younger northeast of the front. The flattening of biotite ages northeast of the front reflects rapid cooling and exhumation during Miocene detachment faulting.

at the $^{40}Ar/^{39}Ar$ geochronology laboratory at UC Santa Barbara. Specific information and more detailed data on the $^{40}Ar/^{39}Ar$ analyses can be found in Supplementary Table S2.

$^{40}Ar/^{39}Ar$ analyses conducted on metamorphic hornblende from the central and NE-central footwall yield the oldest interpreted $^{40}Ar/^{39}Ar$ hornblende ages (Fig. 11b). Har-27 from the central Harcuvar Mountains yields an age spectrum that is relatively flat for 60% of the spectrum (weighted mean age of 66.7 ± 0.1 Ma but no well-defined isochron) with slightly older apparent ages for the highest temperatures steps. Har-71 from the NE-central Harcuvar Range yields a slightly hump-shaped age spectrum with most step ages ranging from 58.5–65 Ma and a weighted mean age of 64.3 ± 0.1 Ma for the flattest part of the age spectrum (80% of final gas release). Alternatively, the sample yields a four-point isochron age of 60.7 ± 0.4 Ma (MSWD = 0.8) that comprises 57% of the gas release ($^{40}Ar/^{36}Ar_{init}$ = 339 ± 4). Sample H-12 from the southwest end of the Little Buckskin range yields an age spectrum with a more pronounced age gradient that climbs from ca. 47–63 Ma with a flat segment at the highest temperature steps (comprising 35% of gas release with the highest radiogenic yields) with a weighted mean age of 63.1 ± 0.2 Ma. A four–point isochron yields a similar age of 66.4 ± 0.8 Ma with no well defined isochron.

Results from the northeastern-most footwall yield distinctly younger $^{40}Ar/^{39}Ar$ hornblende ages (Fig. 11c). Age spectra from these samples are similar in that they tend to climb from minimum ages of 24–37 Ma up to ages of 43–46 Ma for the first ~50% of gas release at low temperature and then flatten significantly for the second half of gas release at high temperature steps. Radiogenic yields follow a similar pattern for these samples. The flatter portion of the age spectrum for Har-43 yields a weighted mean age of 43.3 ± 0.1 Ma and a three-point isochron that comprises 40% of the gas yields a similar age of 43.4 ± 0.14 Ma (MSWD = 0.8, $^{40}Ar/^{36}Ar$ = 300.5 ± 10). The flatter portion of the H-32 age spectrum comprises 51% of the gas release and yields a weighted mean age of 45.2 ± 0.1 Ma. The flatter portion of the age spectra of H-24, from the northeasternmost Little Buckskin Mountains, produces a weighted mean age of 45.3 ± 0.1 Ma for the flat portion. No isochron could be reasonably fit to the H-32 or H-24 sample data. All of these spectra are best interpreted as reflecting varying degrees of argon loss or possibly very slow cooling of older (pre-45 Ma) hornblende.

While most of the $^{40}Ar/^{39}Ar$ hornblende data do not yield age spectra with simple plateaus, we interpret the preferred ages reported here as geologically meaningful and reflecting the time of footwall cooling through hornblende $^{40}Ar/^{39}Ar$ closure temperature of ~480–550°C (Harrison, 1981; McDougal and Harrison, 1988). Excess argon contamination can be a common issue with $^{40}Ar/^{39}Ar$ analyses of metamorphic hornblende, but the standard hallmarks of significant excess argon issues, such as strongly U-shaped spectra or highly irregular step ages (e.g. McDougall and Harrison, 1999) are not evident in the data. The isochron plots also do not suggest significant excess argon issues where good fits were possible. The spatial trend of hornblende ages that young towards the structurally deeper northeastern footwall also support that these ages are geologically meaningful. Given the correlation between the shape of the age spectra and radiogenic yields, the oldest high temperature age steps likely record cooling through hornblende closure temperature during the Late Cretaceous to Paleocene with the younger low temperature steps documenting moderate argon loss during alteration, perhaps during Miocene deformation. Taken together, these results suggest that most of the footwall had cooled through ~480–550°C by ca. 65 Ma and the entire footwall by ca. 43 Ma.

New $^{40}Ar/^{39}Ar$ biotite age results also shed light on the thermal history of the Harcuvar footwall. Overall, the $^{40}Ar/^{39}Ar$ biotite age spectra are relatively flat and show little complexity, making them readily interpretable (Fig. 11d). The biotite ages show clear spatial trends that consistently young towards the northeast, with a few minor anomalies. Biotite ages range from a maximum of 65.5 Ma in the southwesternmost footwall and young significantly to ca. 31–28 Ma ages in the central footwall near the mylonitic front. Further northeast, there is a distinct break in the age vs. distance slope (Fig. 11e) and $^{40}Ar/^{39}Ar$ biotite ages decrease from ca. 18 Ma to a minimum of 14.3 Ma in the northeasternmost footwall. These results indicate that the footwall southwest of the mylonitic front had cooled below 325 ± 30°C (biotite closure temperature, McDougall and Harrison, 1999) by the start of Miocene extension at ca. 21 Ma but the footwall northeast of the mylonitic front was above this temperature at that time.

3.8 U-Pb geochronology

Significant portions of the mylonitic zone in the Harcuvar footwall occur in leucogranite. Given the variability of deformation in this unit and its extensive presence within the footwall, we conducted LA–ICP–MS U-Pb zircon geochronology on leucogranite units, leucogranite/pegmatite dikes and sills, and other granite units, in order to constrain the timing of footwall magmatism and its relationship to deformation. U-Pb geochronology was conducted at the Laser Ablation Split Stream facility at the University of California, Santa Barbara. Analysis spots 20 µm in diameter were picked based on cathodoluminescence images of polished zircon mounts to avoid inherited cores and metamorphic overgrowths. For each sample, we report weighted mean U^{238}/Pb^{206} ages of analyses with <5% discordance. Data analysis was conducted using the IsoplotR program (Vermeesch, 2018) and we allowed the program to reject outlier ages for weighted mean age calculations.

Geochronologic results indicate that a major pulse of magmatism occurred in the latest Cretaceous to early Paleocene, with most leucogranite units yielding ages from ca. 74–64 Ma (Figure 12, Table 2). These results are similar to but somewhat younger than the 80–78 Ma U-Pb date of the type locality of the Tank Pass Granite in the western Harcuvar Mountains (DeWitt and Reynolds, 1990). We infer these leucogranitic intrusions to all be part of the Late Cretaceous Tank Pass granite suite and these results document a more protracted phase of magmatism during this time period than previously recognized. The preponderance of latest Cretaceous U-Pb ages in our results still undersells the volumetric significance of the leucogranites, which make up substantial portions of the Harcuvar Mountains footwall as stocks and sills intruded into other footwall units. These Late Cretaceous leucogranites are variably deformed, typically as intensely strained category 4 mylonites. Leucogranite units as young as ca. 64 Ma (Har-81, Table 2) record significant mylonitization in northeastern footwall.

Minor magmatism continued into the Paleocene and early Eocene, primarily in the form of rare pegmatite dikes. Although minor volumetrically, these units provide important constraints on the timing of footwall fabrics. In several localities in the Harcuvar and Little Buckskin mountains, rare pegmatite and leucogranite dikes cut the mylonitic foliation at high angles (Figure 13). These dikes are weakly deformed, with the surrounding mylonitic foliation feathering weakly into their margins. Given these characteristics, we interpret these dikes as synkinematic intrusions that were emplaced towards the end of mylonitization. One such dike (Har-50, strike and dip of 220/88 NW) cuts category 3-4 granitic mylonites at a high angle along the northwest flank of the Harcuvar footwall. Zircon U-Pb results on this dike yielded nearly concordant dates with a weighted mean age of 63.0 ± 0.5 Ma (MSWD = 1.2, n=7, Fig. 13). In the Cunningham Pass region in the central Harcuvar footwall, completely undeformed pegmatitic dikes crosscut foliation within a non-mylonitic biotite-feldspar gneiss. U-Pb dates from one of

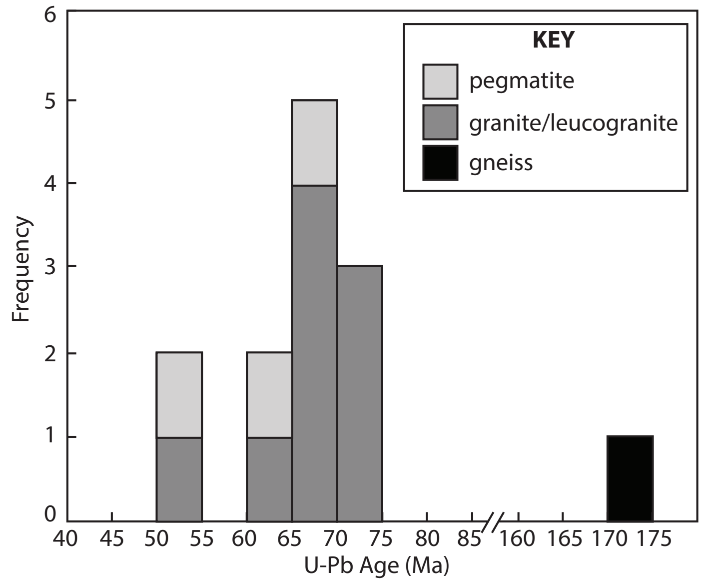

Figure 12. Summary histogram of U-Pb geochronology from the Harcuvar footwall. These analyses focused on the geochronology of the Tank Pass leucogranites, although some other granitic units and pegmatite dikes were also dated. The main pulse of magmatism in the Harcuvar footwall occurred from ca. 74-64 Ma, with minor magmatism (mainly pegmatite dikes) occurring until ca. 53 Ma. The Cretaceous leucogranites are a substantial volumetric unit and make up large portions of the Harcuvar footwall. Note the break in the age scale on the x-axis.

these dikes (Har-55) yield younger concordant ages that typically form zoned rims around inherited Jurassic cores, with the rim ages yielding a weighted mean age of 54.4 ± 0.7 Ma (MSWD = 1.2, n = 10).

4 Discussion

4.1 Temperature conditions of footwall mylonitization

The range of deformation styles evident from the microstructural data strongly suggest that footwall mylonites of the Harcuvar and Buckskin-Rawhide core complexes formed under a wide range of temperature conditions. At strain rates of ~10^{-12} to $10^{-13}/s^{-1}$, quartz bulging recrystallization dominates in the lower greenschist-facies (~280–400°C; chlorite zone), whereas quartz subgrain rotation recrystallization dominates in the upper greenschist-facies (400–500°C), and grain boundary migration dominates in the amphibolite-facies (>500°C) (Stipp et al., 2002a; Faleiros et al., 2010). At slower strain rates of ~$10^{-14}/s^{-1}$, the transition from quartz bulging recrystallization to subgrain rotation recrystallization likely occurs in the lower greenschist-facies (~350°C), whereas the transition from subgrain rotation to grain boundary migration

Table 2. U-Pb Geochronologic Results

Sample name	Description	Latitude*	Longitude	# of analyzed spots†	U^{238}/Pb^{206} weighted mean age ± 2σ	MSWD
HAR-68A	unstrained leucogranite	33.7413	-113.6735	16	74.1 ± 0.7	4.0
HAR-72	mylonitic leucogranite	34.0822	-113.4194	3	70.2 ± 3.9	5.0
HAR-87	mylonitic leucogranite	34.0924	-113.3160	13	70.0 ± 1.2	12
HAR-42	protomylonitic leucogranite sill	34.1085	-113.2807	21	69.9 ± 0.6	4.0
LB-H-11	ultramylonitic leucogranite	34.0878	-113.5360	3	69.8 ± 1.9	1.8
HAR-48	mylonitic leucogranite sill	34.0322	-113.4746	10	68.9 ± 0.7	2.4
HAR-45	mylonitic leucogranite	34.1068	-113.2874	14	68.3 ± 0.5	2.1
HAR-84	mylonitic pegmatite dike	34.0789	-113.2990	21	67.3 ± 0.6	1.9
HAR-81	mylonitic leucogranite	34.0529	-113.3863	22	63.7 ± 0.4	0.8
HAR-50	protomylonitic pegmatite dike	34.0420	-113.4856	6	63.0 ± 0.5	1.2
HAR-55	unstrained pegmatite dike	33.9705	-113.5438	10	54.4 ± 0.7	1.2
HAR-54	fine grain granite sill	33.9705	-113.5436	7	54.2 ± 0.8	0.9

*Latitude and Longitude locations are given in the NAD27 datum.
† Spot analyses that yielded ages with >5% discordance are not reported here.

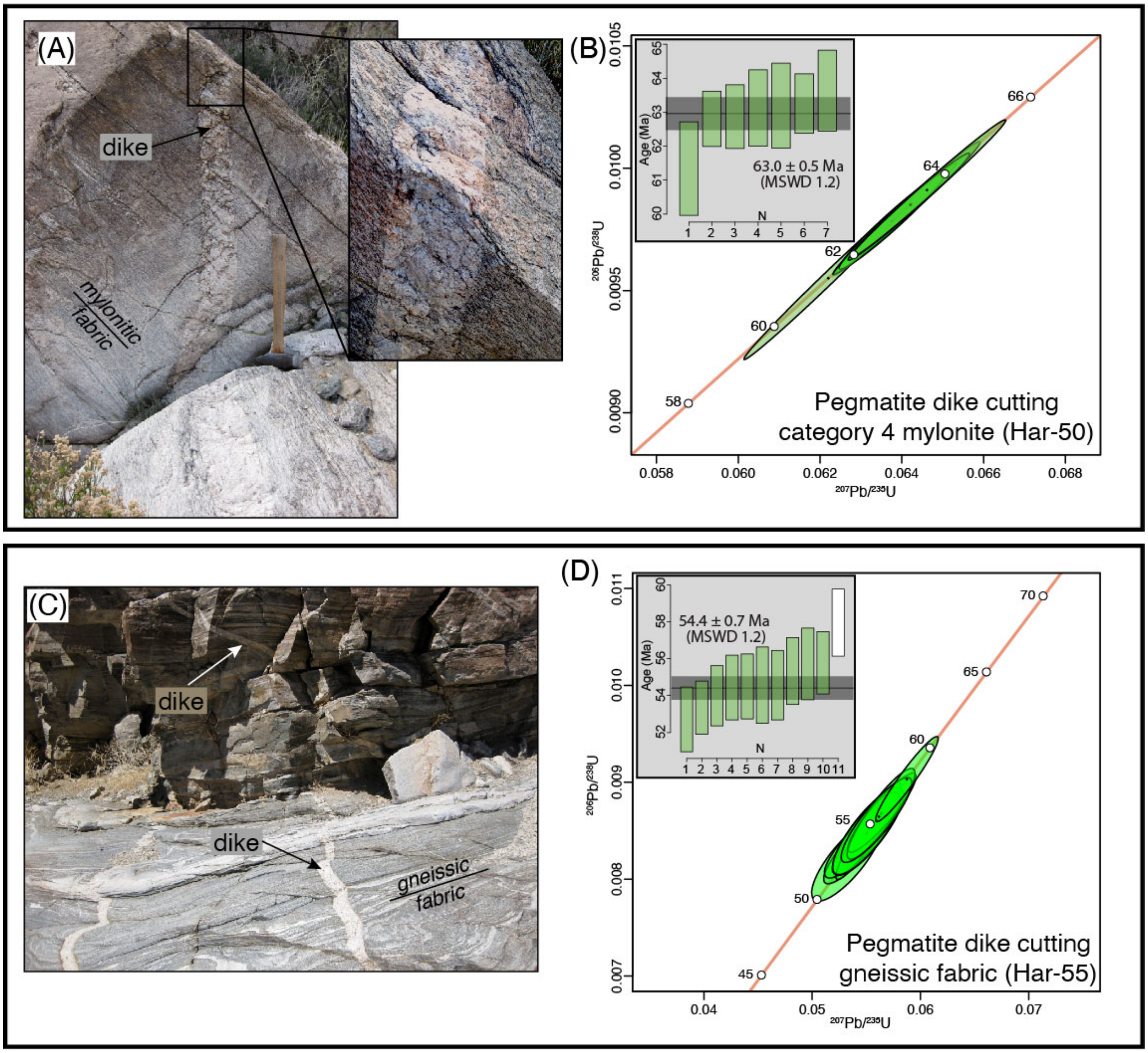

Figure 13. Images of cross-cutting pegmatite dikes and related U-Pb zircon geochronology. (A) Pegmatite dike (Har50) cross-cutting highly deformed category 4 mylonites at a high angle. Closeup view shows the straight but slightly feathered dike margins and a very weak fabric in the dike. (B) Concordia plot and weighted mean U-Pb ages (63.0 ± 0.5 Ma) of Har50 illustrating the concordant nature of the ages. (C) Pegmatite dikes (Har54) cutting gneissic fabrics in the central footwall. (D) Concordia plot and weighted mean U-Pb ages (54.2 ± 0.8 Ma) of the Har54 sample.

recrystallization likely occurs in the middle greenschist facies (~400°C) (Stipp et al., 2002b; Law, 2014).

The correlation between feldspar deformation mechanisms and temperature is complicated by several processes that may occur at a range of conditions, including chemically-driven alteration and breakdown, fracturing, diffusion creep, and solution-precipitation (e.g., Tullis and Yund, 1991; Fitz Gerald and Stunitz, 1993; Fukuda and Okudaira, 2013). Dislocation creep and subgrain rotation recrystallization typically become important in the amphibolite facies (Simpson, 1985; Gapais, 1989; Pryer, 1993; Fitz Gerald and Stunitz, 1993 and references therein; Kruse et al., 2001).

In the quartzofeldspathic category 4 mylonites in this study, the dominance of quartz grain boundary migration recrystallization and feldspar subgrain rotation recrystallization strongly suggest that deformation occurred in the amphibolite facies, likely >500°C. This temperature is consistent with the quartz CPO patterns, where the strong Y-axis strain maxima evident in most samples typically correlates with dominance of prism <a> slip and deformation temperatures >500°C (Law, 2014). The lack of evidence for prism <c> slip suggests that deformation temperatures were largely below ~600°C (Lister and Dornseipen, 1982; Mainprice et al., 1986; Okudaira et al., 1995; Stipp et al., 2002a). Although deformation mechanisms may be influenced by factors other than temperature, including water content and strain rate, the pervasive subgrain rotation recrystallization in feldspar and fast grain boundary migration in quartz observed in many of the category 3 and 4 samples likely requires temperatures >500° C, regardless of other variables. Moreover, estimates of strain rates during mylonitization along this belt of core complexes are relatively high (~10^{-11} to 10^{-14} s^{-1}; Behr and Platt, 2011; Campbell-Stone and John, 2002), so it is unlikely that low strain rates under greenschist-facies conditions account for development of these microstructures.

The Ti-in-quartz results provide independent support that deformation of category 3 and 4 mylonites occurred in the amphibolite facies, with the vast majority of samples in these categories yielding calculated temperatures >500°C within the error of the analyses, assuming 6 kbar pressure (Fig. 10). At 4 kbar, 7 of the 9 samples still yield calculated temperatures above 475°C within error. An important question is whether the Ti concentrations were fully reset during mylonitization. Nachlas et al. (2014) demonstrated that Ti concentrations in quartz were re-equilibrated during experimental dynamic recrystallization. In addition, Grujic et al. (2011) and Behr and Platt (2012) have applied these methods to well-constrained shear zones, demonstrating the utility of this approach in constraining mylonitization temperatures. A few of our results are somewhat more complex. For example, two of the category 3–4 samples yield calculated temperatures <500 °C and one category 2 sample yields a temperature above 500 °C (Fig. 10). These results are likely the product of uneven or partial re-equilibration during lower-temperature deformation. Grujic et al. (2011) noted that Ti diffusion at the greenschist facies may be too slow to fully reset the Ti system in quartz, and thus these temperatures may reflect significant Ti inheritance from the earlier high temperature mylonitization. This may especially be an issue when deformation is not accompanied by significant dynamic recrystallization (Nachlas et al., 2014). Aside from these minor complexities, the Ti-in-quartz results, when combined with the petrographic and EBSD data, strongly suggest that substantial portions of the mylonite zone were deformed under amphibolite-facies conditions (>500°C), especially in the Harcuvar, Little Buckskin, and southern Buckskin mountains (Fig. 8).

Our results also demonstrate that some mylonitization occurred at lower temperatures in the greenschist facies. The presence of quartz bulging recrystallization, the dominance of feldspar fracturing/cataclasis, significant chloritization and distinct CPO patterns of category 1 mylonites strongly suggest deformation in the middle to lower greenschist facies (e.g. Passchier and Trouw, 2005). Category 2 and 3 mylonites likely represent deformation in upper greenschist to lower amphibolite-facies conditions. In many areas, these samples may record an incomplete upper greenschist-facies overprint of amphibolite-facies (category 4) mylonitic fabrics. Greenschist-facies mylonites appear to be more widespread in the Buckskin-Rawhide footwall and within the metasedimentary carapace located <250 m below the detachment fault across the entire footwall but are rare within the crystalline core of the Harcuvar footwall (Figs. 8 and 9). Taken together, our results suggest that the mylonitic shear zone formed during both amphibolite and greenschist-facies conditions, and that there is a strong spatial control on the location of these distinct mylonite categories.

4.2 Timing of amphibolite vs. greenschist-facies mylonites

The presence of both greenschist and amphibolite-facies mylonites within the footwall raises important questions about the timing of footwall mylonitization and whether all of the deformation was coeval with Miocene core complex development. Given that the ca. 22–21 Ma Swansea Plutonic Suite in the Buckskin-Rawhide core complex was deformed at middle to upper greenschist-facies conditions (Singleton and Mosher, 2012), it is clear that significant mylonitization occurred in parts of the footwall in concert with the Miocene initiation of detachment faulting. The highest footwall deformation temperatures during Miocene extension were likely where Swansea Plutonic Suite magmatism was concentrated in the central part of the Buckskin-Rawhide core complex (Fig. 2). Accordingly, we interpret the prevalence of (category 2) upper greenschist-facies fabrics in the Swansea Plutonic Suite in the Buckskin-Rawhide footwall (Fig. 8) to record peak mylonitization conditions during Miocene core complex development.

Further away from the Swansea intrusions, the footwall was likely cooler in the Miocene, although our new $^{40}Ar/^{39}Ar$ biotite results, combined with those of Scott et al. (1998), demonstrate that the mylonitic footwall was hotter than biotite Ar closure temperature (325 ± 30°C) at the inception of Miocene extension. As a result, it is likely that all greenschist-facies mylonites across the study area are Miocene in age. The dominance of category 1 (middle- to lower greenschist-facies) fabrics in metasedimentary mylonites (Fig. 8) suggests that these relatively weak rocks preferentially absorbed Miocene strain directly below the detachment system as the footwall was sheared through the brittle-plastic transition. Although metasedimentary mylonites only form a thin footwall carapace (<1 to 100 m-thick) and are volumetrically minor, they were likely present along 25–35% of the detachment system in the Buckskin-Rawhide core complex (Singleton et al., 2018), and thus played an important role in absorbing Miocene detachment-related strain. Category 1 fabrics that were not developed in metasedimentary mylonites are largely found near the southwestern end of the mylonitic footwall and reflect shallower Miocene structural levels near the mylonitic front.

It is less clear whether the amphibolite-facies mylonites identified in this study also formed during Miocene extension or during an earlier and distinct tectonic event. Given that the geometries (Fig. 9) and top-NE kinematics (Fig. 5) of the two types of mylonites are indistinguishable, the simplest interpretation would be that both fabrics formed as part of the same Miocene extensional shear zone that evolved from amphibolite to greenschist-facies temperatures. However, this simple model is not supported by our results. First, the ~65–43 Ma $^{40}Ar/^{39}Ar$ hornblende ages from the Harcuvar footwall suggest that all of the mylonitic footwall had cooled below the amphibolite facies well before the Miocene (assuming a hornblende closure temperature of 525 ± 40°C; McDougall and Harrison, 1999). Although not all of the age spectra yield simple plateaus, the results are consistent with cooling through the closure temperature in the Paleocene-Eocene followed by variable Ar loss due to minor retrograde (Miocene?) alteration. The preservation of a hornblende age gradient in the slip direction (similar to the biotite data), with younger $^{40}Ar/^{39}Ar$ hornblende ages in the northeasternmost (structurally deepest) footwall (Fig. 11) also suggest that these data are geologically meaningful.

Previous work in the Buckskin-Rawhide footwall yielded similar 70–45 Ma $^{40}Ar/^{39}Ar$ hornblende ages (Richard et al., 1990; Scott, 1995; Fryxell in Bryant, 1995; Scott et al., 1998). Although Richard et al. (1990) reported a few ca. 29–26 Ma $^{40}Ar/^{39}Ar$ hornblende ages, these samples were located near the Early Miocene Swansea Plutonic Suite (Bryant, 1995) and likely experienced substantial argon loss during reheating. Thus, all available thermochronology suggests that, outside of local zones of Miocene reheating in the Buckskin-Rawhide footwall, the footwall had cooled below amphibolite-facies conditions no later than ca. 43 Ma and largely before ca. 65–60 Ma. These results strongly suggest that amphibolite-facies mylonites predate Miocene deformation and instead formed during a discrete event that predated the ca. 65–60 Ma $^{40}Ar/^{39}Ar$ hornblende cooling ages.

Our new U-Pb geochronologic results further refine the timing of this earlier phase of mylonitic deformation. The development of amphibolite-facies mylonitic fabrics in the Late Cretaceous Tank Pass leucogranite indicates that at least some of this deformation postdated the 74–64 Ma emplacement of this suite. The variably deformed nature of the Tank Pass leucogranite also suggests that this unit was intruded syn-tectonically and that various phases captured different degrees of deformation as it intruded the footwall. Finally, the weakly deformed ca. 63 Ma pegmatitic dike that cuts amphibolite-facies mylonites in the central Harcuvar footwall (Fig. 13) provides direct evidence that this phase of mylonitization was waning by the early Paleogene in that part of the footwall. The fact that some leucogranite units as young as ca. 64 Ma in the northeasternmost footwall record amphibolite-facies mylonitization suggests that some strain probably continued after ca. 64 Ma in the structurally deepest part of the footwall, likely aided by heating from intrusion of the leucogranites themselves. Taken together, the microstructural, thermochronologic and geochronologic results provide strong evidence that amphibolite-facies mylonitization occurred during a discrete event in the latest Cretaceous to early Paleogene.

4.3 Evidence for Late Cretaceous extension

The presence of pre-Miocene mylonitic fabrics within the footwalls of these core complexes raises important questions about the kinematics and tectonic significance of this older phase of mylonitization. It is perhaps not surprising that mylonitic deformation occurred in mid-crustal rocks in this region during the Late Cretaceous. During this time, the entire western margin of North America was experiencing subduction, and the Maria fold-and-thrust belt southwest of the study area experienced significant Cretaceous shortening (e.g. Hamilton 1982, 1987; Laubach et al. 1989; Spencer and Reynolds, 1990; Boettcher et al., 2002; Cawood et al., 2022). Within the study area, Cretaceous thrust faulting has been described in the Granite Wash Mountains (e.g. Laubach et al., 1989) and in the adjacent Harquahala Mountains (e.g., Richard, 1988). However, thrusting in these areas is interpreted as top-SW directed, whereas the kinematics of the amphibolite-facies mylonites in the Harcuvars are top-NE, which is the same as Miocene extension. In addition, shortening in the Maria fold-and-thrust belt was largely complete by ca. 80 Ma (e.g. Martin et al., 1982; Isachsen et al., 1999; Flansburg et al., 2021), whereas we interpret top-NE shearing to have continued until ca. 63 Ma. This suggests that amphibolite-facies mylonites record an earlier episode of top-NE extensional deformation during the latest Cretaceous. Other workers have also reported evidence for Late Cretaceous top-NE extension in the Maria fold-and-thrust belt and adjacent areas, including the Dome Rock Mountains (Boettcher and Mosher, 1998), Little Maria Mountains (Ballard and Ballard, 1990), Iron Mountains (Wells et al., 2002), and Granite Mountains (Salem, 2009), suggesting a significant and regional tectonic event. Taken together, these data suggest that amphibolite-facies mylonites in the Harcuvar and Rawhide-Buckskin core complexes accommodated top-NE extension during the latest Cretaceous.

Fundamental to this interpretation is the assumption that the footwall had a similar structural orientation during the latest Cretaceous as it did during top-NE directed Miocene extension. Thermochronologic data confirms that the exposed footwall dipped northeast in the Late Cretaceous. The 65.5 Ma $^{40}Ar/^{39}Ar$ biotite age from the southwestern Harcuvar footwall (Har68A, Fig. 11) suggests this part of the footwall cooled below 325 ± 30°C (biotite closure temperature, McDougall and Harrison, 1999) by the latest Cretaceous. A similar 64.5 ± 5 Ma zircon (U-Th)/He age from a Proterozoic granitoid in the southwesternmost footwall of the Buckskin detachment fault (Singleton et al., 2014) suggests that this part of footwall had cooled below ~180–200°C by that time. In contrast, the central and NE-central footwall must have been much hotter in the latest Cretaceous, as ca. 67–63 Ma $^{40}Ar/^{39}Ar$ hornblende ages from this area (Fig. 11) indicate cooling through hornblende closure temperatures of 525 ± 40°C at that time. Finally, the northeasternmost footwall must have been even hotter than ~525°C in the latest Cretaceous, given the ca. 43–45 Ma $^{40}Ar/^{39}Ar$ hornblende ages in this area. Taken together, these data strongly indicate that the northeastern footwall was >200–350°C hotter and therefore structurally deeper than the southwest footwall during the latest Cretaceous. As a result, we conclude that latest Cretaceous top-NE-directed mylonitization in the study area records normal-sense shear and an episode of extensional deformation with the same kinematics as that of Miocene extension.

The recognition of two discrete extensional fabrics raises questions about the prevalence of latest Cretaceous versus Miocene mylonitization. In the Harcuvar footwall, only the greenschist-facies mylonites can plausibly have formed in the Miocene. While upper greenschist-facies mylonities locally overprint structurally deeper portions of the Harcuvar footwall (producing category 2 and 3 mylonites), pervasive greenschist-facies mylonitization is limited a relatively narrow zone of mylonites ≤250 m below the detachment fault (Figs. 8 and 9). In the Buckskin-Rawhide footwall, significant Miocene mylonitation is evident from deformation of the Miocene Swansea Plutonic Suite (Singleton et al., 2012) and widespread upper greenschist-facies fabrics (Fig. 8), which almost certainly reflects hotter footwall temperatures associated with plutonism. Amphibolite-facies fabrics are locally preserved in the southern Buckskin-Rawhide footwall, where Late Cretaceous leucogranite is common and Swansea Plutonic Suite intrusions are rare. Thus, it is likely that the Buckskin-Rawhide footwall also experienced significant Late Cretaceous mylonitization, although it is more difficult to evaluate the extent of this phase of deformation given significant Miocene overprinting.

Significant Late Cretaceous extension should have produced thermal, metamorphic, and other geologic signals that could provide additional evidence for these interpretations. Previous $^{40}Ar/^{39}Ar$ thermochronology from the Colorado River Extensional Corridor indicate widespread cooling in the Late Cretaceous to early Cenozoic (e.g., Knapp and Heizler, 1990; Foster et al., 1990; Richard et al., 1990). While this cooling has been interpreted to record erosion (e.g., Knapp and Heizler, 1990) or refrigeration from flat slab subduction (e.g., Dumitru et al., 1991), we suggest it may instead record regional extensional exhumation during the latest Cretaceous to Paleogene. In addition, there is tantalizing thermobarometric evidence for significant tectonic exhumation during this time period. For example, Anderson et al. (1988) interpreted that Late Cretaceous granitoids in the lower plate of the Whipple Mountains core complex were emplaced at ~30 km and were exhumed to <20 km depths prior to Oligo-Miocene detachment faulting. This exhumation may record Laramide extensional unroofing. Similarly, Walsh et al. (2016) reported preliminary thermobarometry and monazite geochronology on rare pelitic garnet ± kyanite schists from the Harcuvar footwall that suggests Late Cretaceous (76–70 Ma) garnet growth during decompression from pressures as high at ~10 kbar to as low as ~4 kbar during that time. These data suggest a major exhumational event of ~4–6 kbar occurred during the Late Cretaceous. Finally, in the nearby northern Plomosa Mountains, the Late Cretaceous (~73 Ma) Orocopia Schist was underplated as a subduction complex and subsequently exhumed to ~3–5 km depths prior to ~21 Ma (Strickland et al. 2018; Spencer et al., 2018), requiring major exhumation before the inception of Miocene detachment faulting. Paleogene exhumation of the Orocopia Schist in southern Arizona has been associated with extension (Jacobson et al., 2002, 2007; Oyarzabal et al., 1997).

Although there is currently a lack of evidence for normal faulting and extensional basin formation in this region during the latest Cretaceous to Paleogene, there has been substantial Miocene denudation, which may have removed much of the surficial geologic evidence of this early phase of extension. It may also be difficult to distinguish older normal faults from Miocene structures, especially given the possibility of reactivation. Alternatively, Hodges and Walker (1992) presented a kinematic model whereby Late Cretaceous mid-crustal extension in the Western U.S. Cordillera was decoupled from upper-crustal deformation. Regardless, existing thermochronology and thermobarometry provide additional support for a significant exhumational event in the latest Cretaceous.

In summary, we conclude that both the footwalls of the Harcuvar and Buckskin-Rawhide core complexes experienced two discrete phases of top-NE extensional mylonitization: an earlier Late Cretaceous to early Paleogene phase followed by overprinting during Miocene extension. The earlier extensional fabrics are well preserved in the Harcuvar footwall where Miocene fabrics are limited, whereas the earlier fabrics are more strongly overprinted in the Buckskin-Rawhide core complex where Miocene magmatism locally reheated the footwall, allowing for more pervasive Miocene mylonitization.

4.4 Implications for core complex formation

Cordilleran metamorphic core complexes have long been interpreted as unique and enigmatic features of crustal extension owing in large part to their low-angle fault geometry, juxtaposition of brittle and ductile features, and the large magnitude and high rate of extension inferred to be integral to their formation (see reviews by Lister and Davis, 1989; Wernicke, 1995; Whitney et al., 2013; Platt et al., 2015). Central to most core complex models is the interpretation that they represent the product of a single phase of Cenozoic extension, where footwall mylonites represent the mid-crustal roots of coeval detachment fault systems (e.g., Wernicke, 1981; Davis et al., 1986; Lister and Davis, 1989; Spencer and Reynolds, 1991).

This study demonstrates that the tectonic evolution of at least some core complexes may be more protracted than these prior models recognized. Our results suggest that the Harcuvar and Rawhide-Buckskin core complexes experienced two discrete phases of extension and footwall mylonitization separated by more than 40 Myr, indicating a protracted and polyphase tectonic evolution. If this conclusion is broadly correct, it raises fundamental questions about the nature of core complexes and the processes by which they form.

One controversial aspect of core complexes has been the low-angle geometry of the mylonitic fabrics and bounding detachment fault. The central issue is whether these normal faults initiated and slipped at their present shallow dips (e.g., Wernicke, 1981, 1985; Scott and Lister, 1992) or formed at steep initial dips and were rotated to shallower dips through time either by rotation on other normal faults (e.g., Proffett, 1977; Gans et al., 1985; Wong and Gans, 2008) or by isostatic rebound (rolling hinge) processes (Buck, 1988; Wernicke and Axen, 1988). The initiation of low-angle normal faults is at odds with classic rock mechanics (Anderson, 1951), and slip on low-angle normal faults appears to be rare in actively extending regions (e.g., Jackson, 1987; Jackson and White, 1989; *cf.* Abers, 1991; Boncio et al., 2000), so this question is fundamental to our understanding of how faults form and slip.

The recognition of a polyphase extensional history at the Harcuvar and Buckskin-Rawhide core complexes adds a new dimension to this long-standing debate. If extension in these core complexes began in the Late Cretaceous, then subsequent Miocene extensional structures were superimposed on earlier crustal fabrics, which may have important mechanical consequences. Our results indicate that the Late Cretaceous shear zone was reactivated during Miocene extension and this pre-existing weakness may have allowed the crust to fail in non-ideal orientations, including at shallower angles than would be predicted for intact crust. Such a reactivation scenario also provides a compelling explanation for the identical geometries of Late Cretaceous and Miocene footwall fabrics in that Miocene structures inherited their orientation from pre-existing weaknesses established by the Late Cretaceous shear zone. In addition, pre-existing shear zones in the middle crust may rotate stress fields to non-Andersonian orientations and allow extensional failure at lower dips (e.g., Wu and Lavier, 2016). Recognition of a polyphase extensional history at these core complexes also raises questions about other fundamental aspects of core complex development including understanding the contribution of earlier phases of extension to the total magnitude and rate of detachment fault slip.

4.5 Regional tectonic implications

The recognition of substantial Late Cretaceous extension within this part of the North American Cordillera raises significant questions about the tectonic evolution of the region. One important question is what drove syn-orogenic extension during the latest Cretaceous. Many workers have evoked gravitational collapse of the Cretaceous Sevier orogen either during or immediately following crustal thickening (e.g., Hodges and Walker, 1992), and our results are consistent with such a model. The spatial and temporal association of voluminous Late Cretaceous Tank Pass leucogranite with coeval extensional fabrics strongly suggests that the leucogranite played an important role in this tectonic event. The leucogranite is typically peraluminous with two micas ± garnet, suggesting it was derived from significant crustal melting (Miller and Bradfish, 1980; Lee et al., 1981; Farmer and DePaolo, 1983; Haxel et al., 1984; Miller and Barton, 1990; Patiño-Douce et al., 1990; Wright and Wooden, 1991, see review by Chapman et al., 2021). This is consistent with a cycle of crustal thickening producing localized crustal melting from heating and dehydration-derived fluids, which triggered crustal collapse of overthickened crust (Figure 14).

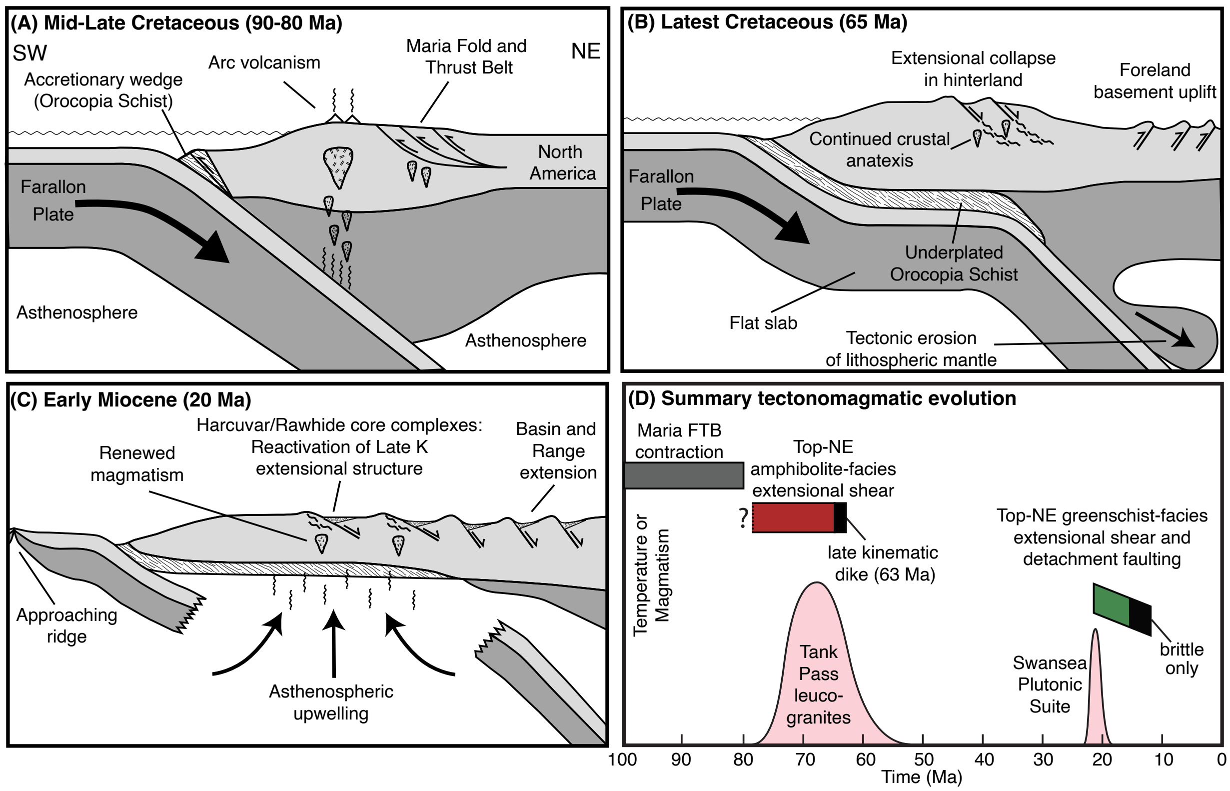

Figure 14. Schematic tectonic evolution of the Buckskin-Rawhide and Harcuvar core complexes in the context of the North American Cordillera. (A) South-vergent crustal shortening and thickening in the Maria fold and thrust belt was occurring during the mid-Late Cretaceous (ca. 90-80) as a result of the subduction of the Farallon plate beneath North America. Thickening generated significant crustal melting. The Pelona-Orocopia schist was forming within the accretionary wedge at that time. (B) During the latest Cretaceous (ca. 65 Ma) the overthickened crust was collapsing due to reheating, resulting in top-NE directed extension. Flat slab subduction of the Farallon underplated the Pelona-Orocopia schist beneath the study area, possibly with the tectonic erosion of some of the North American lithospheric mantle. (C) In the Early Miocene (ca. 21-20 Ma), extension and magmatism resumed, perhaps due to the emergence of a slab window or gap and the resultant influx of heat. In the Rawhide-Buckskin and Harcuvar core complexes, Miocene extension reactivated existing Late Cretaceous extensional structures whereas elsewhere, typical "Basin and Range" extension occurred. (D) Summary chart of the proposed tectonomagmatic evolution of the region.

An important element of this model is the assumption that substantial crustal thickening occurred within the Maria fold-and-thrust belt during mid- to Late Cretaceous contraction. While there are a number of mapped thrust faults and other contractional features documented in the region, the magnitude of crustal shortening and thickening in the region during Sevier/Laramide time remains poorly known. However, Chapman et al. (2020) applied geochemical proxies to suggest that crustal thickness in southern United States Cordillera during the Laramide orogeny (ca. 80–40 Ma) was 57 ± 12 km. In addition, preliminary thermobarometry and monazite geochronology on rare pelitic garnet ± kyanite schists from the Harcuvar Mountains suggests that Late Cretaceous metamorphism occurred at 6–10 kbar pressures at ca. 76–70 Ma (Walsh et al., 2016). Although Spencer et al. (2018) argued that any crustal welt in the region may have been tectonically removed from the base of the crust as a result of Laramide emplacement of the Orocopia Schist (Jacobson et al., 2017; Seymour et al., 2018; Strickland et al., 2018) during low-angle subduction, it is unclear that schist emplacement requires the removal of a Cretaceous crustal welt. Given the strong temporal overlap between Tank Pass leucogranite emplacement and Late Cretaceous extension, combined with the recent evidence for substantial early Laramide-age crustal thickening in the region, our preferred tectonic model is that crustal thickening drove heating and local anatexis which triggered collapse of overthickened crust.

Other tectonic models are also viable and may have acted alone or in concert with an orogenic collapse model. For example, tectonic underplating of the Orocopia Schist during the Laramide may have triggered extension by emplacing rheologically weak schist in the middle to lower crust and/or by hydration weakening due to dehydration reactions (Strickland et al., 2018). Alternatively, synorogenic extension may have been triggered by mantle delamination via density-driven foundering, as has been suggested for the North American Cordillera as a whole (Wells and Hoisch, 2008). Mantle delamination would result in elevated geothermal gradients, partial melting, and rock uplift, which might drive extension. Finally, regional Late Cretaceous extension may have been driven by the subduction of thickened oceanic crust (conjugate of the Shatsky Rise) on the Farallon slab and its passage through this region (Saleeby, 2003; Chapman et al., 2010). This proposed aseismic ridge may have driven low-angle subduction during the Laramide orogeny in the Late Cretaceous to early Paleogene, and the trailing edge of the rise is modeled to have passed through the region at ca. 68 Ma (e.g., Liu et al., 2010).

Looking more broadly at the North American Cordillera as a whole, a number of studies have suggested that the Cordillera experienced significant extensional exhumation during the Late Cretaceous to early Cenozoic (e.g. Hodges and Walker, 1992; Applegate and Hodges, 1995; Wells and Hoisch, 2008), but such studies have often relied on thermobarometric data, and the structures that may have accomplished such exhumation are commonly unclear. This study is significant in that it identifies a structure that accommodated significant extension in the middle crust during the latest Cretaceous. However, this does not resolve the question of why such structures remain poorly identified throughout the Cordillera. We speculate that many other Cordilleran core complexes experienced a similar polyphase tectonic evolution with an older period of extension, but tectonic inheritance and middle Cenozoic overprinting effects may make it difficult to recognize these older footwall fabrics. We anticipate that future studies applying new analytical tools will be able to test this hypothesis.

5 Conclusions

The mylonitic shear zones of the Harcuvar and Rawhide-Buckskin core complexes are the result of two distinct phases of crustal extension. Amphibolite-facies mylonitization had top-NE kinematics and occurred during the Late Cretaceous to early Paleogene. This deformation was spatially associated with voluminous ca. 74–64 Ma footwall leucogranites, which were emplaced syn-kinematically. A late kinematic ca. 63 Ma dike indicates this phase of mylonitization had largely waned by the early Paleogene. The leucogranites were likely the result of crustal melting due to orogenic thickening, implying a model whereby crustal heating from thickening and magmatism triggered gravitational collapse of overthickened crust, although the tectonic underplating of Orocopia Schist and/or mantle delamination may have also played a role in triggering orogenic collapse. Miocene footwall mylonitization occurred in the greenschist facies and is largely restricted to areas within and near the Early Miocene Swansea Plutonic Suite and narrow (<250 m-thick) zones immediately below the detachment fault. Miocene extension was superimposed on the latest Cretaceous to early Paleocene shear zone and had similar kinematics, suggesting that the location and geometry of Miocene extension was strongly influenced by tectonic inheritance. The tectonic development of these core complexes was more protracted and polyphase than previously recognized, suggesting that models of core complex development may need to be reevaluated.

Acknowledgments

This work was supported by the National Science Foundation grant number EAR-1321407 and by the Malcolm ’54 and Sylvia Boyce Fund for Geology Research at Colgate University. We thank Colgate undergraduates Jacky Baughman and Austin Sun for assistance with some of the EBSD measurements and Sam Jackson for work on some of the U-Pb geochronology. Andrew Kylander-Clark provided assistance with the U-Pb geochronology at the

Laser Ablation Split Stream facility at the University of California, Santa Barbara and Rick Hervig at Arizona State University provided assistance with the Ti-in-quartz analyses.

Open Research

Additional figures and data on sample characterization, petrographic analyses, U-Pb and $^{40}Ar/^{39}Ar$ geochronology are provided in the supplementary materials. Topographic maps were generated using the GeoMapApp available at https://www.geomapapp.org and using data from the U.S. Geological Survey's National Elevation Dataset available from https://apps.nationalmap.gov/downloader/. Stereonet plots were generated using Stereonet 10.1.6, which is available at https://www.rickallmendinger.net/. U-Pb geochronology figures were generated using IsoplotR (Vermeesch, 2018) which is available at http://www.isoplotr.com/isoplotr/home/index.html. Analysis of the crystallographic preferred orientation data was conducted using the MTEX toolbox (Bachmann et al., 2010), which is available at https://mtex-toolbox.github.io.

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