Extension of the Anaconda metamorphic core complex: 
$^{40}\text{Ar}/^{39}\text{Ar}$ thermochronology and implications for Eocene tectonics of the northern Rocky Mountains and the Boulder batholith

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ABSTRACT

Thermochronologic data define the extension and exhumation history of the Anaconda metamorphic core complex and have implications for the Eocene tectonic setting of the northern Rocky Mountains. The $^{40}\text{Ar}/^{39}\text{Ar}$ data indicate that relatively rapid extension on the Anaconda detachment started at ca. 53 Ma and continued through ca. 39 Ma. Apatite fission-track data reveal that faulting and exhumation of the footwall continued until ca. 27 Ma. The average displacement rate on the Anaconda detachment was on the order of 1 mm/yr between ca. 50 and 39 Ma based on the lateral gradient in mica $^{40}\text{Ar}/^{39}\text{Ar}$ ages in the direction of fault slip. The total displacement along the Anaconda detachment in Eocene and Oligocene times is estimated to be $\geq$25–28 km based on reconstruction of the Cretaceous Storm Lake Stock with its detached roof, which is now exposed in the Deer Lodge Valley. Extension exhumed crust from ~12 km depth and exposed middle-greenschist-facies mylonites in the easternmost part of the Anaconda complex footwall. On a regional scale, the Anaconda detachment dips east beneath the Cretaceous Boulder batholith, indicating that the batholith and the Butte mineralization were transported east in the hanging wall. The Anaconda metamorphic core complex formed at the transition between the Cordilleran hinterland and the foreland at the same time as extension occurred in the Bitterroot and Priest River metamorphic core complexes but exhumed a shallower part of the Eocene crustal section than the contemporaneous complexes to the west.

INTRODUCTION

Eocene metamorphic core complexes are a significant tectonic element of the northern Rocky Mountains, United States, and southern Canadian Cordillera. Exposure of the core complexes resulted from widespread extension that began between 55 and 53 Ma during the final stages of or immediately after shortening ended in the Cordilleran thrust belt (e.g., Constenius, 1996; Foster et al., 2007). Several hypotheses have been proposed to explain the onset of rapid Eocene extension and associated magmatism in the northern Cordillera including: orogenic collapse, asthenospheric upwelling within a slab window, rapid rollback of the Farallon slab, regional transpression associated with northward motion of the Kula plate, or accretion of the Siletzia terrane (Armstrong et al., 1977; Sevinghaus and Atwater, 1990; Morris et al., 2000; Vanderhaeghe and Teyssier, 2001; Breitsprecher et al., 2003; Haeussler et al., 2003; Foster et al., 2001, 2007; Humphreys, 2009a). The $^{40}\text{Ar}/^{39}\text{Ar}$ thermochronologic data from the Anaconda metamorphic core complex in western Montana (Fig. 1) have implications for the timing, duration, and rate of extension in the easternmost core complex in the Cordillera, as well as the Eocene tectonic setting in the northern U.S. Cordillera.

The Anaconda metamorphic core complex (O’Neill et al., 2004; Foster et al., 2007), along with the Bitterroot, Priest River, Clearwater, Shuswap, Kettle, Okanagan, and Walhalla core complexes, constitutes a belt of hyperextended terrains exposed from western Montana to northeastern Washington and British Columbia (Fig. 1B). Unlike the other core complexes, the Anaconda complex is located at the easternmost margin of the Cordilleran metamorphic-plutonic hinterland. Published geochronology and thermochronology indicate that regional extension and core complex exhumation began in the early to middle Eocene at ca. 54–52 Ma, coincident with onset of Challis-Colville-Kamloops-Absaroka magmatism (Armstrong and Ward, 1991; Morris et al., 2000; Foster et al., 2001, 2007; Breitsprecher et al., 2003; Haeussler et al., 2003).

GEOLOGICAL BACKGROUND

The Anaconda metamorphic core complex is located along the eastern edge of the Cordilleran hinterland in western Montana (O’Neill et al., 2004; Foster et al., 2007). This extensional terrain is south of the Lewis and Clark fault zone, east of the Idaho batholith and Bitterroot metamorphic core complex, west of the Boulder batholith, and within the eastern part of the Helena salient of the Cordilleran thrust belt (Fig. 1) (Foster et al., 2007). The Anaconda complex is composed of three structural-metamorphic domains: (1) a metamorphic-plutonic footwall exposed in the Anaconda and Flint Creek Ranges (Figs. 2 and 3), (2) a low-grade hanging wall exposed along the western edge of and within the Deer Lodge Valley (Fig. 2), and (3) a brittle-plastic detachment fault system exposed along the eastern flanks of the Anaconda and Flint Creek Ranges (Figs. 2 and 3).

Footwall rocks of the Anaconda complex are made up of Late Cretaceous to Eocene granitic plutons intruded into metamorphosed Mesoproterozoic...
Extension of the Anaconda metamorphic core complex

Figure 1. (A) Tectonic map of the northern U.S. Rocky Mountains showing major Phanerozoic structures and tectonic elements in the vicinity of the Anaconda detachment, which bounds the Anaconda metamorphic core complex. The box shows the area of the map in Figure 2. (B) Inset map showing a larger area depicting the location of the Anaconda metamorphic core complex and other Eocene metamorphic core complexes. The core complexes are shaded gray, with names corresponding to the abbreviations listed in the key. The box shows the area of the map in A.
Belt Supergroup and Middle Cambrian to Cretaceous shelf-platform strata (Figs. 2 and 4; Fig. DR1) (Emmons and Calkins, 1913, 1915; Desmarais, 1983; Heise, 1983; Wallace et al., 1992; Lonn et al., 2003; Grice, 2006). In the Flint Creek Range, footwall rocks are composed of granodiorite to granite plutons of the Late Cretaceous Mount Powell batholith, Royal Stock, and Lost Creek Stock. These plutons intruded, deformed, and metamorphosed Middle Cambrian to Cretaceous strata and in a few areas metamorphosed Belt strata (Emmons and Calkins, 1913, 1915; Allen, 1966; Hyndman et al., 1982; Lonn et al., 2003; O’Neill et al., 2004). In the Anaconda Range, the footwall is largely Late Cretaceous diorite to granodiorite and early to middle Eocene granitic plutons, which intruded deformed Belt Supergroup and metamorphosed Middle Cambrian strata (Fig. 4; Fig. DR1 [see footnote 1]) (Desmarais, 1983; Wallace et al., 1992; Lonn et al., 2003; O’Neill et al., 2004; Foster et al., 2007). Upper-amphibolite-facies metamorphism and nappe-style folding (Fig. 3C) of the Belt and Cambrian strata occurred in Late Cretaceous time, with peak metamorphic temperatures (>650–700 °C) accompanying intrusion of quartz diorite–granodiorite plutons at ca. 78–75 Ma, based on U-Pb zircon data (Grice, 2006). Cretaceous metamorphism and deformation took place at pressures of 4.6–6.0 kbar based on metamorphic thermobarometry of garnet-bearing metapelitic rocks (Grice, 2006).

The hanging wall of the Anaconda core complex is made up of an array of asymmetric fault-bounded basins containing unmetamorphosed Cenozoic clastic, volcaniclastic, and volcanic strata. These strata are exposed in the Deer Lodge Valley and preserved in a reentrant between

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**Figure 2.** Geologic map and cross section of the central part of the Anaconda metamorphic core complex, Boulder batholith, and adjacent regions. The map was compiled from Emmons and Calkins (1913), Lewis (1998), Lonn et al. (2003), O’Neill et al. (2004), Foster et al. (2007), and Vuke et al. (2007). Locations of samples used for 40Ar/39Ar analyses that are not plotted on Figure 4 are also shown. Cretaceous intrusive rock at location DF02-114, in the hanging wall of the Anaconda detachment, is coarse-grained granodiorite interpreted to be the detached top of the Storm Lake pluton in the footwall. Abbreviations: SP—Sapphire pluton; CJ—Chief Joseph pluton; SLP—Storm Lake pluton; PB—Pioneer batholith; PP—Philipsburg batholith; MPP—Mount Powell pluton; RS—Royal Stock; HL—Hearst Lake plutonic suite; LC—Lost Creek Stock; HLF—Hidden Lake fault. The box shows the area of Figure 4.

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1GSA Data Repository Item 2010208, geologic map of the NE Anaconda Range, tabulated 40Ar/39Ar data, and 40Ar/39Ar age spectra, is available at www.geosociety.org/pubs/ft2010.htm, or on request from editing@geosociety.org, Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301-9140, USA.
Figure 3. (A) Photograph of the footwall and hanging wall of the Anaconda metamorphic core complex looking WNW into the Mill Creek Valley. (B) Photograph showing a listric normal fault cutting greenschist-facies mylonite looking north at the canyon wall of the Mill Creek Valley. (C) Photograph of the north wall of the Mill Creek Valley showing a large Cretaceous nappe defined by metamorphosed Belt Supergroup and Cambrian strata. The ductile strain in the nappe occurred at upper-amphibolite conditions at ca. 75 Ma. (D) Photograph of late-stage semibrittle extensional shear bands cutting greenschist-facies mylonite and ultramylonite in the Mill Creek Valley (view is to the north). (E) Photograph of the south side of the Clear Creek Valley showing listric normal faults that cut through and sole into the greenschist-facies mylonites beneath the Anaconda detachment (view to the south).
Figure 4. (A) Geologic sketch map of the northern Anaconda range (simplified from the map in Fig. DR1 [see text footnote 1]) showing sample locations. The two colors for the Storm Lake Stock are for the granodiorite and quartz diorite compositions. Abbreviations: gd—biotite granodiorite; gr—two-mica granite; sz—shear zone. (B) Map of the same part of the footwall of the Anaconda metamorphic core complex with 40Ar/39Ar cooling ages (errors are 2σ). The thick black lines are “contours” of the cooling ages for muscovite and biotite. The 40Ar/39Ar ages for biotite and muscovite are concordant for each location in the eastern two thirds of the footwall, because of rapid cooling during Eocene time. Abbreviations: ms—muscovite; bt—biotite; kfs (LT)—K-feldspar low-temperature plateau ages; TF—total fusion.
the Anaconda and Flint Creek Ranges (O’Neill et al., 2004; Foster et al., 2007) (Fig. 2). The stratigraphically lowest rocks in these basins are moderately west tilted (~50°–60°), poorly sorted, and poorly consolidated conglomerates, sandstones, breccias, and megabreccias (Kalakay et al., 2003; O’Neill et al., 2004). These strata grade upward into progressively less tilted (~0°–25°) felsic lava flows, tuffs, and volcanioclastic deposits of the Eocene Lowland Creek volcanic field (ca. 53–49 Ma; Dudas et al., 2010). The upward decrease in the tilt of these basin-fill strata indicates deposition synchronous with extension.

Metamorphic and plutonic rocks of the footwall are juxtaposed with the hanging-wall rocks along an east-dipping, low-angle, brittle-plastic detachment system, which shows top-to-the-east-southeast displacement (Emmons and Calkins, 1913; O’Neill and Lageson 2003; Kalakay et al., 2003; O’Neill et al., 2004; Foster et al., 2007). The Anaconda detachment has a mapped strike length of at least 100 km from the northern Flint Creek Range to the southern Anaconda Range (Kalakay et al., 2003; O’Neill et al., 2004; Foster et al., 2007).

The Anaconda detachment is characterized by greenschist-facies mylonite, ultramyolinite, pseudotachylyte, and overprinting brittle normal faults (Figs. 3 and 5). Along the eastern flank of the northeastern Anaconda Range, the detachment is characterized by a 300–500-m-thick lower-to-middle-greenschist-facies mylonitic shear zone of stretched two-mica granite, biotite granite, granodiorite, and mylonitic micaceous quartzite (Emmons and Calkins, 1913; Kalakay et al., 2003; O’Neill et al., 2004, Foster et al., 2007; Grice, 2006). Fractured K-feldspar porphyroclasts in the granitoids are encased by a matrix of plastically deformed quartz. The micaceous quartzite exhibits unannealed quartz ribbons with undulatory extinction and mica fish (Fig. 5). These metamorphic textures are indicative of deformation at temperatures less than ~400–450 °C (Passchier and Trouw, 2005). Greenschist mylonites exhibit shallow-plunging mineral stretching lineations and kinematic indicators, which show top-to-the-east-southeast (102°–110°) sense of motion (Kalakay et al., 2003; O’Neill et al., 2004; Grice, 2006). Strain in the greenschist mylonites is heterogeneous and distributed into 0.1–2-m-thick zones of ultramyolinite alternating with ~5–15-m-thick zones of mylonite and protomylonite (Fig. 5) and some bands of pseudotachylyte (Kalakay et al., 2003; Foster et al., 2007). The mylonites exposed in the northeastern Anaconda Range are cut by an array of closely spaced, east-dipping brittle normal faults. Many brittle faults are listric and become subhorizontal with depth and parallel to ultramylonite zones in the granitoids (Figs. 3B and 3E). Slickenline striations on the brittle fault surfaces show top-to-the-east-southeast (100°–110°) displacement parallel to the stretching direction in the greenschist-facies mylonites (Kalakay et al., 2003).

Exposures of the brittle-plastic detachment are not continuous along strike in the Anaconda and Flint Creek Ranges because segments have been removed by erosion, cut out by younger brittle normal faults, and covered by hanging-wall fault slivers or thick talus (Foster et al., 2007). Isolated exposures of the detachment along eastern flanks of the central and southern Anaconda Range are characterized by low-grade mylonitic two-mica granites and granodiorite cut by a series of east-dipping, northwest-trending brittle normal faults similar to those found in the northeastern Anaconda Range (Wallace et al., 1992). Along the eastern flanks of the Flint Creek Range, greenschist-facies mylonite is found locally in the Lost Creek Stock and metasedimentary rocks equivalent to the Belt and Middle Cambrian section (Allen, 1966; Lonn et al., 2003; O’Neill et al., 2004). These mylonites are cut by high-angle normal faults similar to those found in other parts of the detachment (O’Neill et al., 2004).

Along the eastern flanks of the Flint Creek and Anaconda Ranges, the detachment dips gently (~10°–30°) beneath the Deer Lodge Valley. The gentle dip of the detachment is also revealed by industry exploration wells, which intersected greenschist mylonite at the base of the Tertiary basin fill in the western Deer Lodge Valley at depths of ≤5 km (Fig. 2; McLeod, 1987). The downward projection of the low-angle detachment is aligned with subhorizontal seismic reflectors beneath the Boulder batholith (Vejmelek and Smithson, 1995), suggesting that the detachment shallows with depth and continues to the east (Fig. 2; Foster et al., 2007). This consistent shallow dip along with the listric faults soling into the shear zone is consistent with the deeper parts of the Anaconda detachment originating at low angles within the brittle-plastic transition.

The detachment is not well exposed along the western margin of the Anaconda core complex. The trace of the detachment is inferred in several places by the juxtaposition of brittle faulted upper-plate rocks with plastically deformed metamorphic and plutonic rocks. The western part of the detachment probably originated as a series of east-dipping listric normal faults east of a breakaway zone that is inferred to have been located east of the Georgetown thrust and is either no longer exposed or was removed by erosion (O’Neill et al., 2004; Foster et al., 2007). Upper-amphibolite-facies mylonite and extreme attenuation of footwall strata in the western part of the complex footwall (O’Neill et al., 2004) are related to Cretaceous deformation and are not Eocene structures (Grice et al., 2005); the Cretaceous fabrics are locally overprinted by the Eocene brittle-plastic fabrics.

Figure 5. (A) Field photograph of greenschist-facies mylonite and ultramyolinite developed within Eocene granite from the eastern part of the Mill Creek Valley (orientation: SSE to the right side of the photo). (B) Photomicrograph showing mica fish that grew in Cambrian quartzite mylonite beneath the Anaconda detachment (orientation: SSE to the right side of the photo).
40Ar/39Ar THERMOCHRONOLOGY

Samples were collected for 40Ar/39Ar analysis from lower-plate rocks along a transect parallel to the slip direction on the Anaconda detachment in the northeastern Anaconda Range (Fig. 4). All samples were collected as close to the level of detachment as possible. Samples include: (1) high-grade Mesoproterozoic Belt Supergroup and Middle Cambrian metamorphic rocks; (2) granodiorite of the Late Cretaceous Storm Lake Stock; and (3) granite-granodiorite intrusions of the Eocene Hearst Lake suite (Table 1). Two samples were collected from the upper plate in the Deer Lodge Valley: (1) biotite-hornblende granodiorite (DF02-114) that is correlated with the Storm Lake Stock in the footwall (Fig. 2), and (2) crystal-lithic rhyolitic tuff (DF04-113) of the Eocene Lowland Creek volcanic sequence. One sample was collected from greenschist-facies biotite granite mylonite of the Lost Creek Stock in the lower plate on the eastern Flint Creek Range (Fig. 2).

METHODS

Biotite, muscovite, and K-feldspar were separated using standard density and magnetic techniques, followed by handpicking to >99% purity. The mineral separates were irradiated at the Radiation Center at Oregon State University and underwent 40Ar/39Ar analyses at the University of Florida following analytical procedures described by Foster et al. (2009). The 40Ar/39Ar data are summarized in Table 1, age spectra along with inverse isochron plots are shown in Figure DR2 (see footnote 1), complete step-heating data are presented in Table DR1 (see footnote 1), and interpreted 40Ar/39Ar ages are plotted on Figure 4.

RESULTS

The biotite 40Ar/39Ar ages (Table 1; Fig. DR2 [see footnote 1]) define a lateral gradient across the lower plate where ages become younger toward the ESE (Fig. 4). In the westernmost part of the footwall, biotite ages from the Storm Lake Stock range from ca. 79 to 64 Ma (samples SL-38, WG04-114, and WG04-052). East and south of the Storm Lake Stock, biotite 40Ar/39Ar ages become abruptly younger. Biotite from sample WG04-112, a muscovite- and biotite-bearing quartzite gneiss, yielded a 40Ar/39Ar cooling age of 52.8 ± 0.7 Ma. East of the Storm Lake Stock, in the central part of the exposed footwall, biotite 40Ar/39Ar ages from two-mica granite, biotite granite, and a dacite dike (samples WG04-109, WG04-101, ME-6, WG04-092, and WG04-033) range from ca. 52 to 47 Ma. Biotite from a deformed Cretaceous quartz diorite sill (Ug-1) gave a 40Ar/39Ar age of 50.5 ± 0.8 Ma. In the easternmost part of the northeastern Anaconda Range, biotite from two-mica granite (DF02-116a) and biotite granite (DF02-120) mylonites yielded ages of ca. 41–39 Ma.

### TABLE 1. 40Ar/39Ar DATA FROM THE ANACONDA METAMORPHIC CORE COMPLEX

<table>
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<tr>
<th>Sample</th>
<th>Rock type*</th>
<th>Latitude (°N)</th>
<th>Longitude (°W)</th>
<th>Elevation (m)</th>
<th>Mineral*</th>
<th>Age (Ma)†§</th>
<th>% 39Ar†</th>
<th>MSWD</th>
<th>Comments</th>
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<td>SL-38</td>
<td>Hb bt granodiorite</td>
<td>46°06’06″</td>
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<td>2311</td>
<td>bt</td>
<td>79</td>
<td>1.2</td>
<td>78</td>
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<td>0.9</td>
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<td>81</td>
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<td>1.0</td>
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<td>Two mica granite</td>
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<td>113°10’15″</td>
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<td>Mylonitic bt granodiorite</td>
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<td>112°59’11″</td>
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<td>112°59’03″</td>
<td>–</td>
<td>bt</td>
<td>53.7</td>
<td>1.4</td>
<td>66</td>
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<td>bt</td>
<td>76.3</td>
<td>1.1</td>
<td>95</td>
<td>0.57 From upper plate – Mill Creek</td>
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Note: MSWD—mean square of weighted deviates for plateau age.
†Weighted plateau age unless noted otherwise in comments.
‡Errors.
§% 39Ar—percent of 39Ar used in weighted plateau age calculation.
Muscovite \(^{40}\text{Ar}/^{39}\text{Ar}\) ages (Table 1; Fig. DR2 [see footnote 1]) also become younger to the ESE across the lower plate. In the west, muscovite from WG04-112 gave a \(^{40}\text{Ar}/^{39}\text{Ar}\) age of 56.9 ± 0.7 Ma. To the east, muscovite from a two-mica granite (samples WG04-109, WG04-101, ME-6), pegmatite dike (WG04-100), garnet-bearing leucogranite dike (WG04-089), and mylonitic muscovite-bearing quartzite (samples WG04-103 and WG04-103) gave ages ranging from ca. 52 to 46 Ma. In the eastern part of the footwall, muscovite from a two-mica granite mylonite (DF02-116a) yielded a cooling age of 40.5 ± 2.0 Ma.

Furnace step heating of K-feldspar from granodiorite of the Storm Lake Stock (SL-38) gave a saddle-shaped age spectrum indicative of excess argon (e.g., Foster et al., 1990). Despite the presence of excess \(^{40}\text{Ar}\) in the K-feldspar, some useful age information is deduced from the low-temperature steps. From ~2% to 20% \(^{39}\text{Ar}\) released, the spectrum is characterized by a saw-tooth pattern, where older apparent ages alternate with slightly younger ages. These steps correspond to isothermal heating steps at temperatures from 550 to 750 °C. Harrison et al. (1994) demonstrated that excess \(^{40}\text{Ar}\) released in the relatively older isothermal steps is commonly derived from fluid inclusions, and that the second isothermal step often provides meaningful age information. Removal of the older isothermal steps over the low temperature portion of the age spectrum gives an error plateau age of 57 ± 2 Ma (Fig. DR2 [see footnote 1]).

Hornblende-biotite granodiorite (DF02-114) from the upper plate east of the northeastern Anacanda Range yielded a biotite age of 76.3 ± 1.1 Ma. Crystal-lithic tuff sample DF02-113 yielded a biotite age of 53.7 ± 1.4 Ma. Mylonitic biotite granite (DF02-116) from the Lost Lake Stock (SL-38) gave a biotite age of 53.7 ± 1.4 Ma. Mylonitic biotite granite (DF02-118) from the Lost Creek Stock in the eastern Flint Creek Range gave a biotite age of 38.8 ± 1.6 Ma.

**FISSION-TRACK ANALYSES**

Apatite fission-track data from five samples are presented in Table 2. The fission-track analyses were performed at the University of Melbourne using methods summarized by Green (1986). All samples yielded sufficient apatite grains and high enough uranium concentrations for statistically sound age populations. The apatite fission-track ages for samples of greenschist-facies mylonite from the footwall are concordant within error at 27 ± 1 Ma and give mean fission-track lengths of ~14.0 µm. One sample of granodiorite from the hanging-wall block, interpreted to be the detached top of the Storm Lake Stock (DF02-114), gave a fission-track age of ca. 44 Ma and mean track length of ~12.8 µm.

**DISCUSSION**

The \(^{40}\text{Ar}/^{39}\text{Ar}\) cooling ages provide constraints on the cooling and exhumation history of the Anacanda metamorphic core complex, and the timing of extension. The \(^{40}\text{Ar}/^{39}\text{Ar}\) ages for coexisting muscovite and biotite from every location in the footwall, with the exception of the western-most part of the footwall, are concordant within error. This indicates that rapid cooling took place through the argon closure temperature interval of muscovite (~450–350 °C) and biotite (380–330 °C) (closure temperatures for rapid cooling; e.g., McDougall and Harrison, 1999) for all of the central and eastern parts of the footwall in Eocene time. For example, samples WG04-101 and WG04-109 cooled through this temperature interval in less than 1 m.y., or at a rate of >100/m.y. (Fig. 6, path 2). Eocene granitic rocks (e.g., DF02-116: U-Pb zircon age of 53 ± 1 Ma; Foster et al., 2007) within the greenschist-facies mylonite in the eastern part of the footwall also cooled through the muscovite and biotite closure intervals at this rapid rate, but nearly 10 m.y. after WG04-101 (Fig. 6, path 3).

Based on the \(^{40}\text{Ar}/^{39}\text{Ar}\) data, the western part of the Anacanda footwall was below 300–250 °C before middle Eocene time (Fig. 6, path 1). Results from this part of the footwall record relatively rapid Late Cretaceous cooling of the 75 ± 1 Ma Storm Lake Stock (U-Pb zircon age; Grice et al., 2005; Grice, 2006), followed by slower cooling until early Eocene time (Fig. 6, path 1). The K-feldspar low-temperature release data from the Storm Lake Stock sample show that the western part of the footwall had cooled through the temperature interval of ~250–200 °C by ca. 55–57 Ma. Collectively, these \(^{40}\text{Ar}/^{39}\text{Ar}\) data from the western part of the footwall require that the amphibolites-facies metamorphism of the footwall rocks was Late Cretaceous in age and not associated with Eocene extension.

The apatite fission-track data from the eastern parts of the footwall in both the Anacanda and the Flint Creek Ranges define the cooling history to temperatures lower than those recorded by the \(^{40}\text{Ar}/^{39}\text{Ar}\) data. The mean track lengths of ~14 µm for the samples that give fission-track ages of ca. 27 Ma (Table 2) indicate that these samples record relatively rapid cooling through the apatite partial annealing zone (110–60 °C) and no subsequent thermal annealing. The apparent average cooling rate between ca. 40 Ma and 27 Ma was therefore ~20 °C/m.y. (Fig. 6, path 3). It is possible, however, that cooling and exhumation were not continuous from 40 to 27 Ma at this average rate, because the rate of cooling may have decreased after ca. 40–39 Ma, and then increased again at ca. 27 Ma. The ca. 27 Ma apatite fission-track ages, therefore, either record continuous cooling and exhumation at ~14 °C/m.y. during ~40–39 Ma or are representative of exhumation from a ~40 °C/m.y. cooling rate at ~39–30 Ma.

**TABLE 2. APATITE FISSION-TRACK DATA**

<table>
<thead>
<tr>
<th>Sample number</th>
<th>Number of grains</th>
<th>Standard track density (×10^6 cm^-2)</th>
<th>Fossil track density (×10^6 cm^-2)</th>
<th>Induced track density (×10^6 cm^-2)</th>
<th>Uranium content (ppm)</th>
<th>Chi square probability (%)</th>
<th>Fission-track age*† (Ma)</th>
<th>Mean track length (µm)</th>
<th>Std. dev. (µm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>DF02-114</td>
<td>Mill Creek</td>
<td>7</td>
<td>1.081 (3495)</td>
<td>3.475 (149)</td>
<td>1.623 (696)</td>
<td>18</td>
<td>85</td>
<td>44 ± 4</td>
<td>12.8 ± 0.2 (88)</td>
</tr>
<tr>
<td></td>
<td>(upper plate)</td>
<td>Granodiorite</td>
<td></td>
<td></td>
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<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>DF02-116A</td>
<td>Clear Creek</td>
<td>30</td>
<td>1.092 (3495)</td>
<td>0.365 (82)</td>
<td>0.308 (692)</td>
<td>4</td>
<td>7</td>
<td>25 ± 3</td>
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<td></td>
<td></td>
<td>Granite</td>
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<td></td>
<td></td>
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<td></td>
<td></td>
</tr>
<tr>
<td>DF02-117A</td>
<td>Clear Creek</td>
<td>25</td>
<td>1.104 (3495)</td>
<td>1.257 (118)</td>
<td>0.97 (1397)</td>
<td>11</td>
<td>58</td>
<td>27 ± 2</td>
<td>14.0 ± 0.3 (47)</td>
</tr>
<tr>
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<td>Granodiorite</td>
<td></td>
<td></td>
<td></td>
<td></td>
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</tr>
<tr>
<td>DF02-119B</td>
<td>Race Track</td>
<td>20</td>
<td>1.127 (3685)</td>
<td>7.06 (671)</td>
<td>5.408 (5140)</td>
<td>60</td>
<td>64</td>
<td>28 ± 1</td>
<td>14.0 ± 0.1 (100)</td>
</tr>
<tr>
<td></td>
<td>Creek</td>
<td>Monzodiorite</td>
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<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>DF02-120A</td>
<td>Mill Creek</td>
<td>30</td>
<td>1.138 (3685)</td>
<td>1.083 (127)</td>
<td>0.887 (1040)</td>
<td>10</td>
<td>30</td>
<td>27 ± 3</td>
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<td>Granite</td>
<td></td>
<td></td>
<td></td>
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</tbody>
</table>

*All ages are central ages.
†1σ error used.
The cooling history of Cretaceous granodiorite (DF02-114) within the hanging-wall block east of the NE Anaconda Range is remarkably similar to that of the Storm Lake Stock in the footwall (Fig. 6). This displaced block cooled below biotite argon closure in Late Cretaceous time, and more slowly through the apatite partial annealing zone in Eocene time based on the ~12.8 µm mean fission-track length.

Approximate contours of biotite and muscovite 40Ar/39Ar ages from the Anaconda core complex are shown in Figure 4. The “contours” represent the time when lower-plated rocks in the respective locations cooled through ~450–330 °C (closure temperatures for rapidly cooled micas; e.g., Lister and Baldwin, 1996; McDougall and Harrison, 1999). These data reveal a preserved Paleocene–early Eocene thermal gradient across the lower plate and an eastward progression in the timing of rapid cooling. The mica 40Ar/39Ar ages highlight that the western part of the lower plate cooled through ~350–300 °C (muscovite and biotite closure temperatures for slow cooling; McDougall and Harrison, 1999) in Late Cretaceous time to early Paleocene time (Fig. 6), whereas the easternmost lower plate remained above ~450–350 °C until the late Eocene, as indicated by the mica ages of ca. 41–39 Ma. Cooling of the western part of the footwall through mica closure predated the development of the Anaconda detachment. In fact, cooling of the western area due to tectonic exhumation beneath the Anaconda detachment was restricted to less than 200 °C, based on the K-feldspar data. In contrast, the total amount of cooling of the rocks exposed in the eastern part of the complex due to slip on the detachment was greater than 350–400 °C. This temperature gradient implies that the Anaconda detachment originally dipped east at an angle less than 15° for any reasonable range of paleogeothermal gradients (20–50 °C/km) (e.g., Foster and John, 1999).

Published K-Ar and 40Ar/39Ar data further define the thermal history of the Anaconda core complex for areas outside the northeastern Anaconda Range. Granitic rocks of the Royal Stock and Mount Powell batholiths, from the footwall in the Flint Creek Range, gave mica cooling ages of ca. 65–62 Ma (Marvin et al., 1989), indicating cooling through ~300 °C by Paleocene time. The Philipsburg batholith, a large granodiorite intrusion in the western Flint Creek Range, yielded hornblende and biotite K-Ar ages of ca. 77–72 Ma (Hyndman et al., 1972), indicating rapid cooling to below 300 °C in Late Cretaceous time. Foster and Raza (2002) reported an apatite fission-track age of ca. 57 Ma for the Philipsburg batholith, indicating the intrusion had cooled below 110 ± 10 °C by Paleocene time. These thermochronologic data are consistent with the interpretation that the Philipsburg batholith lies west of the original breakaway zone of the Anaconda detachment and was intruded at very shallow depths (O’Neill et al., 2004; Foster et al., 2007).

In the southern Anaconda Range, the majority of the lower plate consists of Eocene biotite ± muscovite granitoids and dacite dikes similar to those exposed in the northeastern Anaconda Range. These intrusions have mica cooling ages from 60 to 50 Ma (Desmarais, 1983), which indicate that the lower plate in southern Anaconda Range area had cooled through ~300 °C by the early to middle Eocene. The Chief Joseph batholith, located to the west of this area, has hornblende and biotite cooling ages from ca. 75 to 60 Ma. Foster and Raza (2002) reported apatite fission-track ages of ca. 40–30 Ma for the southern Anaconda Range. Together, these data indicate cooling through 550–350 °C in the Late Cretaceous to late Paleocene and then through 110 ± 10 °C in late Eocene to early Oligocene time. A more comprehensive thermochronological data set will need to be collected from this area for direct comparison with our results from the northeastern Anaconda Range.

**Timing and Rate of Extension**

The 40Ar/39Ar cooling ages obtained from samples collected from extensional fault blocks and exhumed metamorphic core complex footwalls may be used to determine the onset of extension if the base of the partial retention zone for a thermochronologic system is identified within the fault block (John and Foster, 1993; Foster and John, 1999; Stockli, 2005). The partial retention zone corresponds to an interval of crustal depths specific to each thermochronometer where progressively higher temperatures result in only partial retention of radiogenic 40Ar (or radiogenic 4He for U-Th/He thermochronology) (Fig. 7). For biotite, the argon partial retention zone occurs between ~250 and 330 °C. At pre-extension depths deeper than the partial retention zone, temperatures are too hot for radiogenic 40Ar to be retained prior to the onset of extension and cooling (e.g., John and Foster, 1993). At pre-extension depths shallower than the partial retention zone, the crust is cool enough to allow retention of radiogenic 40Ar, and rocks containing K-bearing minerals record 40Ar/39Ar ages related to earlier cooling events. Within the partial retention zone, temperatures progressively increase, giving rise to a progression from near argon closure at shallower depths to almost complete loss of daughter isotopes prior to extension at deeper levels. K-bearing minerals residing within the partial retention zone record “mixed ages” upon cooling (e.g., Foster and John, 1999). At the onset of extension, rapid exhumation and cooling of rocks immediately below the base of the partial retention zone—the depth transition between no argon retention and partial retention—quench...
minerals that previously retained no radiogenic argon, thereby recording the onset of extension (Foster and John, 1999; Stockli, 2005).

A plot of muscovite and biotite 40Ar/39Ar cooling ages against distance along a section in slip direction (105°) of the Anaconda detachment is given in Figure 8A. This diagram was constructed by orthogonally projecting sample locations to the transect line (Fig. 8B) and includes an applied error of ±1 km to account for projection errors and elevation (e.g., Foster and John, 1999; Brichau et al., 2006).

As summarized already, the mica 40Ar/39Ar cooling ages become younger to the ESE. There is a rapid decrease from Late Cretaceous ages (≥74 Ma) to early Eocene ages (ca. 53 Ma) over a distance of ~5 km and then a much more gradual decrease to late Eocene ages farther ESE (ca. 40–39 Ma). The change in the slope of mica cooling ages at ~5 km corresponds to a quenched paleoisotherm marking the base of the biotite partial retention zone, or ~330 °C (Foster and John, 1999; Stockli, 2005). Samples to the ESE of this point were at higher temperatures prior to the onset of exhumation and cooling of the lower plate. The top of the partial retention zone lies to the west of this point, but east of sample WG04-114 from the Storm Lake Stock, which gave a biotite cooling age of ca. 74 Ma.

The mica cooling age at the base of the Eocene partial annealing zone is between ca. 54 and 52 Ma, which we interpret to be the time that tectonic exhumation began to rapidly cool the footwall beneath the Anaconda detachment. Eocene granodiorite of the Hearst Lake suite was intruded into the eastern part of the footwall at 53 ± 1 Ma (Foster et al., 2007) and was subsequently overprinted by greenschist-facies fabrics. Intrusion of the granodiorite was presumably related to the change in tectonic setting that initiated regional extension starting at 54–53 Ma (Foster et al., 2001, 2007). An early Eocene age for the onset of extension is also consistent with the age of oldest Lowland Creek volcanic rocks (ca. 53 Ma; Dudas et al., 2010; this study). The ca. 41–39 Ma mica cooling ages obtained from the greenschist mylonite in the easternmost lower plate record when this part of the lower plate cooled through ~350 °C and was exhumed through the brittle-plastic transition. These cooling ages indicate that extension accommodated by slip on the Anaconda detachment continued at least into late Eocene time.

The inverse of the slope on the age-distance plot for muscovite and biotite 40Ar/39Ar cooling ages ≤51 Ma (Fig. 9) provides the basis to estimate the rate of slip on the Anaconda detachment (Foster et al., 1993; Ketcham, 1996; Foster and John, 1999; Brichau et al., 2006). We calculated inverse slopes of the biotite and muscovite 40Ar/39Ar ages using the least-squares regression option in Isoplot v. 3.09a (Ludwig, 2004). Dacite dike sample WG04-033 was excluded from the calculation because the dike was intruded late in the history of extension. The biotite and muscovite data give extension rates of 0.93 ± 0.33 km/m.y. and 0.87 ± 0.48 km/m.y. (2σ), respectively (Fig. 8). The larger error of the muscovite slip rate is the result of scatter in the muscovite 40Ar/39Ar ages at ~10–11 km and relatively large error in age of sample DF02-116 at ~19 km. The muscovite and biotite data give similar values, suggesting that the average rate of slip on the Anaconda detachment was ~0.9 km/m.y. This average rate does not account for any increase or decrease in slip rate between 51 and 39 Ma. A displacement rate of ~0.9 km/m.y. is 5–10 times slower than Miocene slip rates for detachments in the southern Basin and Range Province (e.g., Harcuvar and Buckskin detachments [Foster et al., 1993; Carter et al., 2004]; Chemehuevi detachment [John and Foster, 1993; Foster and John, 1999; Carter et al., 2006]; Raft River detachment [Wells et al., 2000]), but similar to that estimated for the Ruby detachment in the central Basin and Range (Gifford et al., 2007).

### Magnitude of Extension

Displacement on the Anaconda detachment may be constrained by reconstructing the granodiorite phase of the Storm Lake Stock with collinear Cretaceous granodiorite within a detached block to the east in the Deer Lodge Valley (Fig. 2). The granodiorite in both areas is unfoliated medium- to coarse-grained biotite ± hornblende granodiorite with concordant Late Cretaceous biotite 40Ar/39Ar cooling ages (samples DF02-114 and WG04-114; Table 1), consistent post-Cretaceous cooling histories (Fig. 6), and complementary major- and trace-element concentrations (Grice, 2006). Restoring the displaced top of the Storm Lake Stock gives between 25 and 28 km of heave on the Anaconda detachment. This value of heave is greater than the amount indicated by the slip rate of ~0.9 km/m.y. between 53 and 39 Ma (~13 km), but is consistent with this slip rate if the Anaconda detachment continued to be active until ca. 25–27 Ma as indicated by the apatite fission-track data.

Late Cretaceous metamorphic rocks beneath the detachment were metamorphosed at pressures of 4–6 kbar or depths of ~12–18 km, based on metamorphic thermobarometry data (Grice, 2006). Throw on the Anaconda detachment since the Late Cretaceous, therefore, cannot have been more than 12–18 km, although the amount in Eocene time may have been significantly less due to erosional exhumation of the Cretaceous metamorphic and plutonic complex prior to Eocene extension.

### Relationship to Regional Extension

Extension in the Anaconda metamorphic core complex was coeval with early Eocene extension throughout the northern U.S. Rocky Mountains and Canadian Cordillera, and very similar in timing to the Bitterroot metamorphic core complex (Foster et al., 2007). The extension direction in the Bitterroot complex (100°–110°; Hyndman and Myers, 1988; Foster, 2000; House et al., 2002) is identical to the extension direction in the Anaconda core complex.

Eocene extension within the Bitterroot mylonite initially took place at upper-amphibolite-facies conditions (Foster et al., 2001), at temperatures...
Figure 8. (A) Plot of mica $^{40}$Ar/$^{39}$Ar cooling age against distance for sample locations along a section parallel to the slip direction of the Anaconda detachment (105°). Sample locations were orthogonally projected to section A–A' as shown in part B. The errors on the cooling ages are ±2σ, and errors on the location are fixed at ±1 km to account for uncertainty in projection to section A–A' and elevation. The shaded band on upper plot (A) represents the location, or paleodepth, of the partial retention zone for mica prior to Eocene slip on the detachment and exhumation. The location of the partial retention zone is also shown for reference on the map in B.
considerably higher than the greenschist-facies conditions within the Anaconda mylonite. Overprinting deformation in the Bitterroot mylonite occurred at progressively lower temperatures and produced greenschist-facies mylonite, ultramylonite, and shear-banded chloritic breccia (Hyndman and Myers, 1988; Foster, 2000). The total amount of exhumation is greater for the Bitterroot complex than the Anaconda complex. Peak metamorphic conditions in the Bitterroot complex took place in Cretaceous to Paleocene time and attained pressures of 6–8 kbar and 650–750 °C (House et al., 1997; Foster et al., 2001). Metamorphic pressure-temperature-time (P-T-t) paths in the Bitterroot complex show strong decompression with upper-amphibolite-facies metamorphism and partial melting as young as 53 ± 1 Ma (Foster et al., 2001). The total amount of Eocene exhumation was probably ~20–25 km in the eastern part of the core complex (Foster et al., 2001) and ~10 km in the western part of the complex (House et al., 2002; Foster and Raza, 2002), owing to the asymmetry in the fault system.

U-Pb zircon crystallization ages of pre-extensional and synextensional plutons in the Bitterroot footwall indicate that extension started at ca. 54–53 Ma (Foster and Fanning, 1997; Foster et al., 2001, 2007), which is identical to the estimates from the Anaconda complex based on the U-Pb zircon crystallization age of the synextensional granodiorites of the Hearst Lake suite (Foster et al., 2007) and the 40Ar/39Ar thermochronology. Thermochronologic data (40Ar/39Ar and apatite fission-track data) record rapid exhumation of the Bitterroot complex after ca. 50 Ma (Cris and Fleck, 1987; Foster and Fanning, 1997; House et al., 1997, 2002; Foster et al., 2001; Foster and Raza, 2002), and indicate that the eastern footwall did not cool below 250 °C until ca. 40–38 Ma (Foster et al., 2001). Apatite fission-track data suggest that exhumation of the eastern part of the complex did not occur until 35–25 Ma during high-angle normal faulting and opening of the Bitterroot Valley (Foster and Raza, 2002). The exhumation and cooling history of the Anaconda and Bitterroot complexes in terms of timing and duration of extension is, therefore, identical.

Relationship to the Boulder Batholith

The Boulder batholith (Figs. 1 and 2) is a composite of >15 plutons emplaced at shallow depths (<5 km) into a contemporaneous volcanic carapace (Tilling et al., 1968). Magmatism occurred between ca. 80 and 70 Ma, and the most voluminous plutons, including the Butte pluton (granodiorite), were intruded between ca. 75 and 70 Ma (Tilling et al., 1968; Robinson et al., 1968; Hamilton and Myers, 1974; Kalakay et al., 2001; Lund et al., 2002). The Boulder batholith hosts the Butte porphyry Cu-Mo deposit, which formed between ca. 67 and 62 Ma (Lund et al., 2002).

The Boulder batholith spans the western Helena salient (Fig. 1) with lateral thrust ramps along the northern and southern boundaries. The Lombard thrust, which transported middle Proterozoic Belt rocks over Paleozoic and Mesozoic strata, lies immediately east of the Boulder batholith (Schmidt et al., 1990; Lageson et al., 2001). In most places, the eastern contact is a steeply west-dipping mylonite zone (Rutland et al., 1989; Kalakay et al., 2001). The Boulder batholith, therefore, is bounded by thrust faults, except in the west, where it is in the hanging wall of the Anaconda detachment (Fig. 2). Seismic-reflection studies show a highly reflective and laminated lower crust below the batholith starting at ~12–18 km depth (Vejmelk and Smithsonian, 1995), which could be the downdip projection of the Anaconda mylonite.

The 25–28 km displacement estimate for the Anaconda detachment indicates that the Boulder batholith was translated east at least this amount after emplacement. This also implies that the hydrothermal systems that produced the Butte deposit originated to the west. Late Cretaceous–Paleocene plutons in the Flint Creek and Anaconda ranges, including the Mount Powell batholiths, were intruded at deeper crustal levels than the Boulder batholith (Hyndman et al., 1982), and crystallized within the age range of the Butte deposits (Lund et al., 2002). Equivalent-aged plutons beneath the Deer Lodge Valley could be related to mineralization at shallower levels in the Boulder batholith.

Tectonic Setting

The metamorphic core complexes of the northern Cordilleran orogen occupy the hinterland west of the fold-and-thrust belt (Fig. 1). These core complexes developed within either Precambrian basement, the Mesoproterozoic Belt basin, or Phanerozoic accreted terranes. The Bitterroot and Anaconda complexes in western Montana are within the former Belt basin and lie over North American basement. They formed after Jurassic–Cretaceous accretion of oceanic terranes and intense Cretaceous–Paleocene shortening during oblique subduction of the Farallon or Kula plates (Engebretson et al., 1985; Severinghaus and Atwater, 1990; Bird, 2002). The dominant phase of core complex extension started at ca. 54–52 Ma (Foster et al., 2001, 2007), which was within 1–3 m.y. after the end of thrusting in the Cordilleran foreland fold-and-thrust belt at ca. 55 Ma (Harlan et al., 1988; Constenius, 1996; Sears and Hendrix, 2004), and continued to after 40 Ma (Foster and Raza, 2002).

Early Eocene extension throughout the northern Rocky Mountain Basin and Range and Canadian Cordillera was coincident with the widespread Challis–Absaroka–Colville–Kamloops–Bitterroot–Lowland Creek–Montana alkaline province magmatism at 53–45 Ma (e.g., O’Brien et al., 1991; Janecke and Snee, 1993; Foster and Fanning, 1997; Morris et al., 2000; Foster et al., 2001; House et al., 2002; Feeley et al., 2002; Feeley, 2003; Breitsprecher et al., 2003; du Bray et al., 2006; Dudas et al., 2010). This voluminous magmatism has been attributed to regional extension (Morris et al., 2000), subduction (Armstrong et al., 1977), or a slab window between the Farallon and Kula plates (Breitsprecher et al., 2003), but regardless of the generation mechanism, it includes a trace element and
isotopic signature of partial melting of a lithosphere that had been metasomatized by a long history of Mesozoic subduction (e.g., Morris et al., 2000; Feeley, 2003). Any model for the causes of Eocene extension in this region must also explain the voluminous contemporaneous magmatism.

Regional Eocene extension has been attributed to changes in plate-boundary conditions (Haeussler et al., 2003; Foster et al., 2007) or collapse of the orogenic pile due to significant partial melting of the middle crust (e.g., Vanderhaeghe and Teyssier, 2001). There are several alternative models for tectonic plate configurations and dynamics due to uncertainties in the position of the Kula-Farallon spreading center with respect to western North America (e.g., Atwater, 1989). Northward motion of the Kula, or the proposed Resurrection plate, would likely have caused dextral transtension inboard of the plate margin in Eocene time, explaining extension if one of these two plates were adjacent to northwestern North America at the latitude of the U.S.–Canadian border, (e.g., Thorkelson and Taylor, 1989; Haeussler et al., 2003; Breitsprecher et al., 2003). The southern option (Engebretson et al., 1985) for the Farallon-Kula or Farallon-Resurrection spreading center suggests that a slab window opened beneath the northern U.S. and southern Canadian Cordillera in the early to middle Eocene, resulting in the Challis-Colville-Kamloops-Absaroka magmatism and extension (e.g., Breitsprecher et al., 2003).

If the Farallon plate were subducting to the NE under this area in Eocene time (Severinghaus and Atwater, 1990), then the lithosphere north of the flat slab segment, which extended under Colorado in Eocene time, could have been extended (Fig. 10) (Dickinson and Snyder, 1979; Severinghaus and Atwater, 1990; Saleeby, 2003). This is because the angle of subduction was presumably progressively, or sharply, steeper to the north of the flat slab and rolling back to the west. The transition to normal subduction north of the flat slab region could also explain the onset of Challis (and related) arc magmatism at ca. 53–48 Ma, as well as regional extension (Fig. 10). If the region north of the Eocene flat slab had been refrigerated and metasomatized by flat slab subduction in Late Cretaceous–Paleocene times (Dumitrut al., 1991; Humphreys et al., 2003; Humphreys, 2009b) and then subduction steepened as the flat slab segment moved relatively southward, magmatism and extension would be a natural consequence of rising asthenosphere under western Montana, northern Idaho, Washington, and southern British Columbia.

Humphreys (2009a, 2009b) attributed the Challis arc and associated Eocene extension to the collision of the Siletzia terrane (Farallon oceanic lithosphere) into the Columbia embayment. This hypothesis favorably explains the distribution and onset of Challis (and related) magmatism. The locally adakitic signatures of the magmatism (e.g., Breitsprecher et al., 2003) could be attributed to heating and partial melting of the trapped and partially subducted oceanic crust at the leading edge of Siletzia. Extension in this model would be due to the weakening of the orogenic pile during the Eocene magmatic flare-up, asthenospheric upwelling through a tear in the Farallon slab, and/or rollback of the slab at the establishment of Cascadia subduction (Humphreys, 2009b).

Postorogenic extension north of the flat slab segment due to rollback of the Farallon plate, the collision of Siletzia, or a combination of both are consistent with the majority of the data on the timing and distribution of extension and magmatism in the northern Cordillera in Eocene time. The relationship between extension and strike-slip translation of terranes now in the northern Cordillera is also important and was somehow due to oblique subduction beneath North America.

CONCLUSIONS

The 40Ar/39Ar data indicate that extension in the Anaconda metamorphic core complex started at 53 ± 1 Ma and continued until after ca. 39 Ma. The gradient of 40Ar/39Ar muscovite and biotite cooling ages along a cross section in the slip direction of the Anaconda detachment gives an average displacement rate for the fault of ~0.9 mm/yr between 50 and 39 Ma. Additional extension and exhumation of the footwall continued into Oligocene time (27–25 Ma) based on apatite fission-track data. Total displacement of the Anaconda detachment is estimated to be 25–28 km based on reconstruction of the Cretaceous Storm Lake Stock with detached granodiorite in the hanging wall. The timing of extension in the Anaconda complex was coincident with Eocene hyperextension in the Bitterroot, Priest River, and other northern Cordillera core complexes as well as the extensive Challis volcanic arc. Regional extension and magmatism were probably caused by factors related to subduction rollback, slab window formation, or collision of the Siletzia terrane northwest of the shallow subducting segment of the Farallon slab, which did not extend north of Wyoming in Eocene time.

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