1 STRAIN AND PERMEABILITY GRADIENTS TRACED BY STABLE
2 ISOTOPE EXCHANGE IN THE RAFT RIVER DETACHMENT SHEAR
3 ZONE, UTAH

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21 **ABSTRACT**

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23 Combined geochronological and stable isotope data of quartzite mylonite from the footwall of
24 the Raft River detachment shear zone (NW Utah, USA) reveal that an important phase of ductile
25 deformation and infiltration of meteoric water in the shear zone occurred in Miocene time. $^{40}\text{Ar}/^{39}\text{Ar}$
26 release spectra are complex, and plateau ages decrease systematically from 31.1 ± 0.8 Ma at the top to
27 20.2 ± 0.6 Ma at the bottom of the quartzite mylonite section, capturing a segment of the ~40-15 Ma
28 geochronologic record that has been documented regionally and is likely related to partial to total
29 overprinting of Eocene white mica $^{40}\text{Ar}/^{39}\text{Ar}$ ages in the Miocene. Hydrogen stable isotope values of
30 syn-kinematic muscovite range from -123 ‰ to -88 ‰ and suggest that meteoric water infiltrated the
31 detachment shear zone during mica (re)crystallization and mylonite development. Bulk stable isotope
32 analyses from fluid inclusions in quartz support a meteoric origin for the fluid (low D/H and $^{18}\text{O}/^{16}\text{O}$
33 ratios). Quartz and muscovite oxygen isotope analyses show varying degrees of ^{18}O depletion,
34 suggesting spatially variable time-integrated interaction of meteoric fluids with recrystallizing shear
35 zone minerals. The overall pattern of D/H and $^{18}\text{O}/^{16}\text{O}$ ratios indicates that fluids were channelized
36 along restricted layers or shear zones within the deforming detachment system. The variability in
37 $^{18}\text{O}/^{16}\text{O}$ ratios of both quartz and muscovite and the fluid-rock isotopic exchange results can be
38 explained by variations in the shear zone permeability (confined versus diffuse flow) along with strain
39 variations along the transport direction (from flattening to constriction).

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1. INTRODUCTION

The presence of fluids during active deformation has a profound impact on the thermal state and the rheology of the crust, even at deep levels where rocks deform ductilely (e.g. [Kohlstedt et al., 1995](#); [Ranalli, 1997](#); [Bürgmann and Dresen, 2008](#)). Actively migrating fluids influence heat and mass transport, facilitate melting and mechanical deformation, and are an important control on the thermomechanical evolution of crust, particularly in the detachment shear zones that bound metamorphic core complexes (e.g. [Mulch et al., 2006](#); [Person et al., 2007](#), [Gottardi et al., 2013b](#)). However, our understanding of the hydrology and permeability structure of ductile shear zones is incomplete. Here we combine strain and stable isotope analyses in an extensional detachment shear zone to shed light on how (meteoric) fluids and rock interact at the grain scale and how shear zones transmit fluids during deformation.

2. HYDROLOGY OF DETACHMENT SHEAR ZONES

When thick orogenic crust undergoes extension, the cool brittle upper crust is commonly separated from the hot lower crust by a detachment zone that localizes deformation, fluid flow, and thermal exchange (e.g. [Brun et al., 1994](#); [Rey et al., 2001](#); [Teyssier et al., 2005](#)). Detachment activity eventually leads to the formation of metamorphic core complexes (MCC) in which heat transfer produces strong thermal/metamorphic gradients ([Brun et al., 1994](#); [Rey et al., 2001](#); [Teyssier et al., 2005](#), [Mulch et al. 2006](#); [Tirel et al., 2008, 2009](#); [Gessner et al., 2007](#); [Rey et al., 2009](#); [Huet et al., 2011](#), [Whitney et al., 2013](#)). Extensive brittle faulting in the upper crust results in significant increase in permeability and porosity, with the potential of enhancing the circulation of fluids at the crustal scale. Based on oxygen ($\delta^{18}\text{O}$) and hydrogen (δD) stable isotope analyses, the presence of surface-derived (meteoric) fluids during mylonitization has been documented in many different extensional detachments including those that bound the Thor-Odin dome ([Holk et al., 1997, 2000](#); [Mulch et al., 2004, 2006](#)), the Kettle dome ([Mulch et al. 2007](#)), the Bitterroot MCC ([Kerrick and Hyndman al., 1986](#); [Quilichini, 2012](#)), the Raft River Mountains ([Gottardi et al., 2011](#)), the Ruby Mountains ([Fricke](#)

et al, 1992), the Snake Range (Gébelin et al., 2011; Gébelin et al. in press), the Whipple Mountains (Morrison and Anderson, 1998; Gébelin et al. 2012), the Simpon detachment in the Central European Alps (Campani et al. 2012), the Menderes MCC in Turkey (Hetzl et al. 2013) as well as the South Tibetan Detachment at Mount Everest (Gébelin et al. 2013) even though there exist cases where low δD and $\delta^{18}O$ values may be derived from surface alteration prior to burial (e.g. Raimondo et al. 2013).

Modeling studies of crustal-scale fluid, heat, and oxygen isotope transport and exchange in regions of crustal extension show that fluid migration and fluid-rock exchange at mid- to lower-crustal levels is primarily controlled by the permeability contrast between the different crustal rocks (Person et al., 2007; Gottardi et al., 2013a). High permeability contrast between adjacent rock units leads to focused fluid flow and can result in significant lowering of the $\delta^{18}O$ of the country rock, given that large amounts of fluid with $\delta^{18}O$ values much lower than the country rock is present for a period sufficiently long to equilibrate with the rock.

However, complete understanding of the hydrology of detachment zones based on field observations is commonly limited owing to lack of exposure and a restricted number of analyses. Thus, samples are usually collected along transects across or along detachment zones (Kerrick and Hyndman, 1986; Fricke et al, 1992; Holk et al., 1997, 2000; Morrison and Anderson, 1998; Mulch et al., 2004, 2006, 2007; Gottardi et al., 2011; Gébelin et al., 2011, 2012, 2013, 2014; Quilichini, 2012, Campani et al. 2012). Here, we combine oxygen and hydrogen isotope geochemistry and $^{40}Ar/^{39}Ar$ geochronology from a quartzite mylonite of the Miocene Raft River detachment shear zone (Utah) to evaluate the permeability along and across the shear zone and the extent and duration of fluid flow during extension. The combined data show that an extensive hydrological system draining low D/H and low $^{18}O/^{16}O$ meteoric fluids was active during the evolution of the shear zone in Miocene time. Quartz and muscovite oxygen isotope analyses show different degrees of ^{18}O depletion that are likely related to: (1) different time-integrated interaction of synkinematically (re)crystallized minerals with meteoric fluid, (2) variation in permeability in the basement units (confined versus diffuse flow), and (3) strain variations from flattening (planes) to constriction (pipes) along the transport direction of the detachment shear zone.

3. GEOLOGIC SETTING

The Raft River Mountains form the eastern limb of the Albion–Raft River–Grouse Creek metamorphic core complex in NW Utah and expose amphibolite- to greenschist-facies Archean to Permian rocks (Fig. 1). These rocks experienced alternating crustal thickening and thinning during the Mesozoic to early Cenozoic Sevier orogeny (Wells, 1997; Wells et al., 1998; Hoisch et al., 2002; Harris et al., 2007), followed by episodes of extension on the west-rooted Eocene to early Oligocene Middle Mountain shear zone (Compton, 1983; Saltzer and Hodges, 1988; Wells, 1997; Wells et al., 2000). This Paleogene shearing was accompanied by a long-lived period of magmatism that is manifested by the 32 Ma Middle Mountain gneiss, 30 Ma Almo pluton, and 29–28 Ma Vipont pluton (Strickland et al., 2011), and the 25 Ma Red Butte pluton (Compton et al., 1977; Egger et al., 2003, Wells et al. 2012; Konstantinou et al., 2012, 2013).

The east-rooted Raft River detachment studied here is the youngest shear zone in the Raft River Mountains (Fig. 1) (Malavieille, 1987; Wells et al., 2000; Wells, 2001; Wells et al. 2012). In the Ten Mile Canyon area, located in the easternmost part of the Raft River detachment shear zone (see cross-section B-B' in Fig. 1), reported muscovite $^{40}\text{Ar}/^{39}\text{Ar}$ ages from the mylonitic quartzite are about 15 Ma (14.88 ± 0.17 Ma and 15.05 ± 0.18 Ma, Wells et al., 2000). Recently documented muscovite $^{40}\text{Ar}/^{39}\text{Ar}$ ages in the Clear Creek Canyon area (cross section A-A' in Fig. 1) can also be similarly young (13–14 Ma, Cosca et al. 2006). Apatite fission-track ages from the footwall of the Raft River detachment shear zone record an episode of progressive east-directed unroofing along the detachment fault between 13.6 and 7.4 Ma (Wells et al., 2000). This study focuses on this Miocene Raft River detachment zone exposed at Clear Creek (cross section A-A' in Fig. 1, Fig. 2), where a continuous section through the detachment shear zone and along the transport direction is well exposed (Compton, 1975; Wells, 2001; Sullivan, 2008), and where numerous profiles of this shear zone have been collected to address the 2D patterns of fluid-rock interaction.

The Green Creek Complex (Armstrong and Hills, 1967; Armstrong, 1968) is the lowest unit in the Raft River detachment shear zone at Clear Creek (Fig. 1 and 2) and consists of ca. 2.55 Ga gneissic monzogranite (metamorphosed adamellite of Compton, 1972) that intrudes metatrandhjemite, metagabbro, and hornblende-biotite schist (Armstrong, 1968; Compton et al., 1977; Todd, 1980;

124 [Miller, 1980](#)). The Clear Creek canyon cuts through the Green Creek Complex and exposes the
125 transition between relatively undeformed Archean monzogranite that underlies the western part of the
126 study area, and the more intensely deformed schist units that underlie the eastern part of the Raft River
127 shear zone ([Fig. 2](#)) ([Compton, 1980, Sullivan, 2008](#)).

128 These basement units are separated from the overlying Elba Quartzite by a 10 to 50 cm thick
129 conspicuous schist unit that has been interpreted as a metamorphosed paleosol/regolith ([Armstrong,](#)
130 [1968](#)) or a metamorphosed basement-derived arkose ([Compton, 1975](#)). The overlying Elba quartzite
131 and schist ([Wells et al., 1998; Sullivan, 2008](#)) display a distinct stratigraphy that includes, from bottom
132 to top, a basal quartzite-cobble metaconglomerate, an alternating sequence of white quartzite and
133 muscovite-quartzite schist, a distinctive layer of red quartzite, and a sequence of alternating feldspar-
134 rich quartzite, pure quartzite, and pebble-gravel metaconglomerate, which we refer to as
135 microconglomerate. In the first 15 m above the contact between the basement and the Elba quartzite,
136 the quartzite is interlayered with centimeter- to meter-thick muscovite-rich units. Between
137 approximately 15 and 22 m above the contact a white mica schist unit makes a distinctive “notch” in
138 the quartzite cliff. The base of this physiographic notch is characterized by an abrupt transition from
139 typical cliff-forming Elba quartzite to a zone where muscovite makes up 80-100% of the friable rock.
140 The upward remainder of the section consists of relatively pure quartzite, the red quartzite unit, and the
141 pebble-gravel microconglomerate ([Fig. 3](#)). Quartz veins (1 to 8 cm wide) cross cutting the mylonitic
142 foliation show variable degree of ductile deformation (folding or boudinage), and are locally
143 transposed parallel to the foliation. The hanging wall of the Miocene Raft River detachment system is
144 not well exposed above Clear Creek. However, Paleozoic metasedimentary rocks are preserved as a
145 few scattered klippen above the quartzite and schist units ([Fig. 1](#)) ([Compton, 1975, Wells, 1997, 2001,](#)
146 [2009; Wells et al., 1998](#)).

147 Throughout the Clear Creek area foliation and lineation in the Miocene shear zone are strongly
148 developed, display a constant orientation ([Fig. 2](#)), and are defined by flattened and elongated quartz
149 and muscovite grains ([Compton, 1980; Wells, 1997; Sullivan 2008](#)). With the goal of exploring further
150 the extent of fluid rock interaction in the detachment shear zone, we sampled five profiles across the
151 sub-horizontal shear zone, from the Archean basement to the uppermost quartzite and

microconglomerate. The nearly vertical profiles are distributed along the E-W striking Clear Creek canyon, and straddle the transition in basement lithology from meta-monzogranite in the west to schist in the east (Fig. 2).

4. MICROSTRUCTURES

4.1. QUARTZ MICROFABRICS

The Elba quartzite is strongly deformed and shows two populations of quartz grains (Fig. 4). In the first population coarse elongate (>500 μm long) grains define the macroscopic fabric and exhibit typical dislocation creep microstructures, such as strong undulose extinction (Fig. 4a, b), deformation bands, and deformation lamellae (Fig. 4c), indicating that quartz deformed under high flow stress conditions (e.g. Schmid and Handy, 1991; Hirth and Tullis, 1992; Drury et al., 1993). The other population consists of finer recrystallized grains (20 - 100 μm) that occupy the boundaries of coarser grains (Fig. 4b). Recrystallized grains are commonly equant to slightly elongate and have a direct relation to subgrains present in large grains, suggesting that the dominant process of dynamic recrystallization was subgrain rotation, where recovery is accommodated by dislocation climb (Regime II of Hirth and Tullis, 1992). A few grains with lobate grain boundaries indicate that grain boundary migration was also active but limited, and did not play a significant role in producing the observed microstructures.

Recrystallized grains locally form a low angle oblique secondary foliation at 30-35° to the macroscopic foliation, indicating top-to-east sense of shear (Fig 4d). This sense of shear determined microscopically is consistent with macroscopic kinematic criteria such as extensional shear bands, drag folds, S-C fabrics, and the sigmoidal shapes of pebble-size grains, indicating a top-to-east sense of shear irrespective of position within the Miocene Raft River detachment shear zone.

The quartz crystallographic preferred orientation of the Elba quartzite has been analyzed by electron backscattered diffraction (EBSD) analysis on six samples from Section 1 and nine samples from Section 5 (Gottardi and Teyssier, 2013). All analyzed samples display a strong crystallographic fabric that developed by deformation in the dislocation creep regime. Quartz c-axis fabrics display a

Type-I cross-girdle (Lister, 1977), with maxima indicating dominant prism- $\langle a \rangle$, rhomb- a , and basal- a slip systems (Gottardi and Teyssier, 2013). The c-axis cross-girdle is largely symmetrical, suggesting a strong coaxial component of deformation during fabric development, an observation that was also made by Compton (1980), Wells (2001), and Sullivan (2008). Sullivan (2008) also noted an increase in the coaxial component of flow toward the east along the quartzite layer.

4.2. FLUID INCLUSIONS

Fluid inclusions are common throughout the detachment shear zone in quartz of the Elba quartzite and synkinematic quartz veins (Gottardi, 2012). The most common fluid inclusions are located at grain and subgrain boundaries (Fig. 5a) and have sizes up to 10 μm . A few grains have preserved isolated fluid inclusion planes that crosscut subgrain boundaries but not grain boundaries (Fig. 5b). Deformation lamellae are also commonly decorated with fluid inclusions (Fig. 5c) (Gottardi and Teyssier, 2013). The samples located close to the “notch” (Fig. 3) show large amounts of fluid inclusion planes in quartz that are parallel and closely spaced ($\sim 25 \mu\text{m}$) and spatially related to micro-vein zones (Fig. 5d). With a diameter up to 30 μm , they are the largest of all the observed inclusions. In those samples, fluid inclusion planes cross several adjacent relict grains (Fig. 5d). Fluid inclusion planes are interpreted to have been open fluid-filled microfractures that subsequently healed (cf. Roedder, 1984; Samson et al., 2003).

4.3. MUSCOVITE MICROSTRUCTURE

The mylonitic Elba quartzite typically contains $\sim 5\text{-}10\%$ muscovite. Muscovite grains are 50 to 200 μm long and a few tens of microns thick (Fig. 6a), and rarely form mica fish (Gottardi et al. 2011). Instead, they are concentrated in bundles or layers that define the foliation. These muscovite layers or sheets are 10 to 50 μm thick and up to 1 cm long. Depending on bulk mica content these layers are more or less continuous, and are spaced on average 500 μm to 1 mm apart in sheets that define the foliation. The grains have a blocky or platelet-like texture: they have straight boundaries and elongated rhomb shapes with their (001) plane oriented parallel to the mylonitic foliation. Larger muscovite grains have fuzzy rims and display intense recrystallization of small muscovite grains along their rims but are devoid of intragrain deformation features, such as kink bands or microfractures.

According to the morphological grouping of mica microstructures, mica grains in the Raft River shear zone belong to Group 6 of [Ten Grotenhuis et al. \(2003\)](#) with high aspect ratio and inverted stair stepping ([Fig. 6b](#)). The rare mica fish that are present are commonly small (30 to 75 μm long, 25 μm thick, [Fig. 6c](#)) and belong to Group 2 of that same classification, with cleavage planes that have sigmoidal shapes converging at one or both tips of the mica grains.

These microstructural observations suggest that mica recrystallized during deformation by migration recrystallization or dissolution-precipitation. Migration recrystallization may explain the geometry of the rare mica fish of Group 2, with rotation of cleavage planes toward the shear plane associated with grain growth parallel to the (001) plane ([Lister and Snoke, 1984](#)). Recrystallization by dissolution-precipitation creep (e.g. [Elliott, 1973](#); [Rutter, 1983](#)) could explain the formation of the smaller platelet-shaped muscovite. This mechanism involving dissolution and mass transport via a fluid phase and precipitation at low pressure sites in the mylonite is supported by the absence of intragrain deformation features in muscovite cores and tips.

Recrystallization by dissolution-precipitation creep is likely to partially reset argon concentrations and equilibrate the oxygen and hydrogen isotopes of muscovite with that of the fluid at elevated temperature (e.g., [Mulch et al., 2005, 2006](#)). Therefore, during dynamic recrystallization of quartz and muscovite in the Miocene Raft River detachment, muscovite likely recorded the isotopic composition of the fluid from which it grew and exchanged isotopically (oxygen and hydrogen) with quartz during deformation ([Mulch et al., 2006, 2007](#); [Wells et al., 2000](#); [Gottardi et al., 2011](#)).

5. $^{40}\text{Ar}/^{39}\text{Ar}$ GEOCHRONOLOGY

We analyzed four muscovite separates from Section C ([Fig. 2](#)), distributed over the ~100 m thick quartzite mylonite by furnace step heating $^{40}\text{Ar}/^{39}\text{Ar}$ geochronology ([Table 1](#)). The four muscovite separates (>125 μm size fraction) show complex age spectra ([Fig. 7](#)), with a decrease in apparent age during the first gas release steps (0 to 5%) followed by a monotonous rise and a “saddle” shaped release spectrum at higher temperatures. Plateau ages for these samples were calculated for the flattest part of multiple contiguous steps including > 50% of total ^{39}Ar released in at least five successive steps.

From bottom to top, the apparent $^{40}\text{Ar}/^{39}\text{Ar}$ ages of muscovite grains are: 20.2 ± 0.6 Ma (RR09-91; at 8 m above the basement-quartzite contact), 24.5 ± 0.4 Ma (RR09-94; 12 m), 26.3 ± 0.4 Ma (RR09-96; 16 m), and 31.1 ± 0.8 Ma (RR09-104; 58 m). Plateau ages correspond to total integrated ages of 20.26 ± 0.06 Ma, 28.01 ± 0.06 Ma, 27.40 ± 0.04 Ma, and 70.01 ± 0.17 Ma, respectively. The $^{40}\text{Ar}/^{39}\text{Ar}$ spectrum from the top of the section is complex, and the “plateau” age is derived from <50% of gas release. The spectra from the three lowest samples (8-16 m of section), which were collected below the “notch”, become progressively flat toward the contact with the basement. Given the youngest plateau ages obtained by Wells et al. (2000) and Cosca et al. (2006) (~14-15 Ma) in the highly sheared section of quartzite at Ten Mile Canyon, we consider the obviously complex ages obtained at Clear Creek as maximum ages for deformation and recrystallization of the quartzite.

6. STABLE ISOTOPE GEOCHEMISTRY

Oxygen isotope ratios ($\delta^{18}\text{O}$) of mineral separates from Sections 1, 3, 4, 5 and C were analyzed at the University of Lausanne (Switzerland) and mineral separates from Section E were analyzed at the University of Wisconsin, Madison. The hydrogen (δD) isotopic ratios were determined at the University of Lausanne (Switzerland) and Leibniz University Hannover (Germany). Analytical facilities and procedures are described in Appendix 2; this section focuses on the analytical results. All $\delta^{18}\text{O}$ and δD values presented here are reported on the VSMOW scale.

6.1. OXYGEN STABLE ISOTOPES

New oxygen stable isotope analyses (Table 2, Fig. 3) are complementary to the data of Gottardi et al. (2011), which are included and integrated here. Quartz $\delta^{18}\text{O}$ values from Gottardi et al. (2011) are variable, ranging from 9.8 ‰ (RR07-57) to 13.1 ‰ (RR07-98) (Fig. 3a). The muscovite $\delta^{18}\text{O}$ values range from 7.1 ‰ (RR07-71) to 10.5 ‰ (RR07-99) (Fig. 3a) (Gottardi et al., 2011). The lowest $\delta^{18}\text{O}$ values for both quartz (9.8 ‰) and muscovite (7.1 ‰) are obtained from a quartzite that lies in the lower half of the shear zone (Gottardi et al., 2011). Sections C and E were collected to the east of the sections reported in Gottardi et al. (2011) and display higher oxygen stable isotope values

259 ranging from 11.7 ‰ to 13.6 ‰ for quartz, and 9.5 ‰ to 12.2 ‰ for muscovite (Table 2, Fig. 3b, c).

260 The largest spread in muscovite $\delta^{18}\text{O}$ values occurs in Section E (9.5 ‰ to 12.1 ‰), which is larger
261 than the spread in Sections 1-5.

262 Values of $\delta^{18}\text{O}$ of the basal metaconglomerate were investigated at a millimeter scale in order
263 to evaluate isotopic heterogeneity on a sample scale. The metaconglomerate is composed of typically
264 pure quartzite clasts (~70 % clasts) in a matrix relatively rich in muscovite (~20%), the remainder
265 being quartz. The quartz microstructures of stretched clasts and matrix are similar and dominated by
266 subgrain rotation recrystallization. One-millimeter thick sections of two samples collected at the
267 bottom of Section C (RR08-15 and -16, see Fig. 3b) were cut into cubes, each ~1 mm³, allowing
268 analysis of both the pure quartz matrix and pure quartz clasts (Fig. 8). Forty-four micro-samples were
269 selected for oxygen stable isotope analysis. Values of $\delta^{18}\text{O}$ fall within the narrow range of 11.0 to 11.2
270 ‰ (Fig. 8) within the clasts (38 samples) and 10.8 to 11.2 ‰ in the matrix (6 samples), revealing
271 complete oxygen isotope homogeneity on the centimeter scale within the basal metaconglomerate
272 across both quartzite clasts and matrix.

273 6. 2. HYDROGEN STABLE ISOTOPES

274 We present two different sets of hydrogen isotope measurements: (a) δD values of muscovite
275 ($\delta\text{D}_{\text{Ms}}$) (Table 3, Fig. 9), and (b) δD values of fluid inclusions in quartz ($\delta\text{D}_{\text{FI}}$) from 5 quartz veins and
276 5 mylonitic quartzite samples distributed from the basement/cover contact to the top of the shear zone
277 (Table 3, Fig. 9). Muscovite δD values were analyzed in 52 samples of mylonitic Elba quartzite
278 collected from 5 different subvertical profiles, distributed from W to E (Figs 2 and 9) along the Raft
279 River detachment shear zone. Muscovite δD values range from -123 to -88 ‰ (± 4 ‰) (Table 3 and
280 Fig. 9). The same range of δD values is observed within single profiles, with no particular vertical
281 trend. However, the spread in δD values is larger in the first 20 m above the contact between the
282 basement and the Elba quartzite where both the lowest and highest δD values are attained (-124 ‰ to -
283 88 ‰). Higher in the quartzite mylonite section (20-70 m), δD values are less variable and average
284 around ~-110 ‰ (Table 3 and Fig. 9).

The δD values of quartz fluid inclusions (δD_{FI}) analyzed in quartzite and synkinematic veins. ,
 Analytical procedures are detailed in Appendix. The fluid-extraction technique precludes separating
 the different generations of fluid inclusions in a sample, and consequently the results represent the
 average δD of the fluids trapped in the sample. Four quartzite samples collected over the entire section
 display a narrow δD range between -94 ‰ and -84 ‰ (Gottardi et al., 2011) and do not show any
 systematic variation with respect to vertical position in the section (Table 3 and Fig. 9). Analyses of
 fluid inclusions from six foliation-parallel quartz veins yield δD values ranging from -83 ‰ to -104
 ‰, with two low- δD samples (RR07-67 at $\delta D_{FI} = -103$ ‰ at 80 m; and RR07-42 at $\delta D_{FI} = -104$ ‰ at
 35 m above basement). Fluid inclusions in both quartz veins and quartzite mylonite have overlapping
 δD values.

Using the calibration of Suzuoki and Epstein (1976) and exchange temperatures between
 350°C to 450°C (Gottardi et al., 2011), the calculated hydrogen isotope fractionation between fluid
 and muscovite ($\Delta_{Ms-Fluid}$) ranges from 38 ‰ to 23 ‰, respectively. At such temperatures, mica with
 $\delta D_{Ms} = -120$ ‰ to -95 ‰ equilibrated with a fluid composition ranging from $\delta D_W = -97$ ‰ (at 350°C)
 to -57 ‰ (at 450°C), respectively. Assuming that the measured δD_{FI} values are representative of δD_W
 present during deformation and recrystallization, these values are in good agreement with those
 calculated from the most negative δD_{Ms} values (Gottardi et al., 2011).

7. DISCUSSION

The partitioning of oxygen and hydrogen isotopes in minerals and waters has been used for
 several decades to elucidate fluid-rock interaction (e.g. Valley, 1986) at various timescales and
 different geological settings such as intrusions and fault zones, including thrust faults (Burkhard and
 Kerrich, 1988; Rye and Bradbury, 1988; Banks et al., 1991; Moore and Vrolijk, 1992; Sibson, 1994,
 Sibson and Scott, 1998; Crespo-Blanc et al., 1995; Goddard and Evans, 1995; McCaig et al., 1995;
 Ghisetti et al., 2001; Kirschner and Kennedy, 2001), strike-slip faults (O'Neil and Hanks, 1980; Unruh
 et al., 1992; Barton et al., 1995; Sibson, 1998; Kharaka et al., 1999; Pili et al., 2002), and extensional
 detachment systems (e.g. Bowman and Willett, 1991; Bowman et al., 1994; Gerdes et al., 1998; Cook
 et al., 1997; Mulch et al., 2004, 2007; Gébelin et al., 2011, 2012, 2013; Campani et al. 2012;

Siebenaller et al., 2013, Hetzel et al. 2013). However, the mechanisms of fluid transport and mineral-water isotope exchange at the transition between brittle and ductile deformation regimes in the crust are still poorly understood, because the critical conditions under which oxygen isotope exchange occurs between a mineral and an intergrain fluid is controlled by a number of factors including fault geometry, permeability fields, oxygen diffusivity, temperature and cooling rate, grain size, and deformation mechanisms. Several stable isotope studies have shown that in normal fault systems, meteoric water can circulate to significant depth, down to the ductile crust (e.g. Wickham and Taylor, 1987; Fricke et al., 1992; Morrison, 1994; Morrison and Anderson, 1998; Bebout et al., 2001; Mulch et al., 2004, 2006, 2007; Person et al., 2007; Gébelin et al., 2011; 2013; Gottardi et al., 2011, 2012; Siebenaller et al., 2013; Catriona et al., 2014).

One-, two-, and three-dimensional numerical modeling of fluid flow, heat and oxygen isotope transport and exchange has helped to identify the parameters that control the rate and extent of isotopic exchange and transport (Cole et al., 1983; Lassey and Blattner, 1988; Bowman and Willett, 1991; Bowman et al, 2004; Gerdes et al., 1995; Person et al., 2007; Savard et al, 2007; Gottardi et al., 2012). Bowman and Willett (1991) and Bowman et al. (1994) show that the time-space distribution of oxygen isotope composition of rocks is controlled by: (1) the rate of fluid infiltration, (2) the diffusive and dispersive properties of the rock, (3) the rate of isotopic exchange, and (4) the oxygen isotope ratio. In particular, solutions and scaling relationships of the one-dimensional isotopic transport and exchange during fluid flow presented by Bowman and Willett (1991) and Bowman et al. (1994) show the relative importance of advective versus diffusive transport in controlling the isotopic exchange front that develops during infiltration in rock. These competing processes can be accurately described by two dimensionless parameters, the Peclet number and the Damköhler number (Bowman and Willett, 1991; Bowman et al., 1994).

Although forward extrapolation of controlling parameters from transport theory helps to understand the geometry of isotopic fronts, fitting theory to natural measurement, such as the data from the Raft River detachment shear zone presented here, is challenging. The first restriction comes from the limited number of measurements. The data presented here were collected from west to east, along the transport direction, over some thickness of the detachment shear zone, which limits our

understanding of the geometry of the isotopic front in this nearly 2D sampling space. Second, fluid infiltration and fluid-rock exchange occurred during deformation and exhumation of the shear zone, and the processes of isotopic exchange during dynamic recrystallization are poorly understood in such evolving geometry. Third, although the detachment shear zone is localized in a ~100 m thick quartzite mylonite, the subjacent basement lithology changes along the transport direction. Such change in lithology may introduce permeability heterogeneities that induce flow variations along the shear zone. In the following we use the geochemical results from the Raft River detachment shear zone at Clear Creek to assess the extent of fluid rock exchange and to investigate the effect of permeability and strain heterogeneities on fluid flow during extensional tectonics.

7.1. MIOCENE FLUID-ROCK INTERACTION

The muscovite $^{40}\text{Ar}/^{39}\text{Ar}$ spectra obtained at Clear Creek yield complex spectra with Oligo-Miocene plateau segments that show, from the top to the bottom of the section, a progressive younging from 31.1 Ma to 20.2 Ma. The admittedly complex spectra flatten toward the contact with the basement. Given the young plateau ages obtained at Ten Mile Canyon (14.88 ± 0.17 Ma and 15.05 ± 0.18 Ma, Wells et al., 2000; and 13-14 Ma, Cosca et al. 2006), we consider the ages obtained at Clear Creek as maximum ages for deformation and fluid-rock interaction.

Throughout the Raft River detachment at Clear Creek, the textural relation of muscovite grains to recrystallized quartz grains suggests that muscovite grew during recrystallization of quartz. The $^{40}\text{Ar}/^{39}\text{Ar}$ ages from muscovite, therefore, record the combined effects of cooling and exhumation as well as this (Oligo-)Miocene deformation and recrystallization event. If intragrain argon concentration gradients were affected during mylonitic deformation and mica recrystallization, it would be unlikely for the D/H and $^{18}\text{O}/^{16}\text{O}$ systematics to remain unaffected and to preserve inherited compositions if there was an out-of-equilibrium fluid in contact with the minerals.

In addition, pristine fluid inclusions in quartz grains form planes at high angle to the foliation, and fluid inclusion trails frequently crosscut grain and subgrain boundaries of the mylonitic Miocene fabric. Moreover, the hydrogen isotope composition of fluid inclusions (δDFI) is consistent with the δDMs of synkinematic muscovites. These two observations are consistent with the percolation of a

Miocene meteoric fluid with $\delta D \sim -90$ ‰ that exchanged with muscovite and was trapped in fluid inclusions.

The possibility that the host rocks were altered in the near surface prior to reburial (cf. [Raimondo et al., 2013](#)) can be discounted for two reasons: (1) $\delta^{18}O$ and δD values in the Raft River metamorphic core complex away from the detachment have “typical” metamorphic values and the low- δD excursions are clearly defined by the detachment shear zones, and (2) the quartzite studied here underwent high-temperature metamorphism prior to exhumation which likely homogenized δD values in muscovite.

7.2. OPEN SYSTEM STABLE ISOTOPE EXCHANGE

Prior to deformation and rapid exhumation in the Raft River detachment, the Elba quartzite likely resided at amphibolite-facies conditions, which favored isotopic homogenization (e.g. [Valley, 2001](#)). Therefore, the relatively large variations (3 ‰) in $\delta^{18}O_{Qtz}$ values in the Raft River detachment are probably not related to a primary sedimentary signal; such variations would have been largely erased during high-grade metamorphism. Furthermore, in other detachments in which quartzite underwent a similar thermal evolution, quartz $\delta^{18}O$ values tend to be higher, around $\sim +12$ to $+13$ ‰ ([Mulch et al., 2006, 2007](#); [Gébelin et al., 2011](#)). Even during intense quartz deformation and recrystallization in detachment mylonites, these values change significantly only as a result of high time-integrated water-rock ratios. Although detachments are, in general, rock-buffered systems with respect to oxygen, the $\delta^{18}O$ variations in muscovite and quartz in the Raft River detachment likely indicate variable degrees of fluid-rock interaction during deformation rather than protolith inheritance ([Gottardi et al., 2011](#)). Strong evidence of open system behavior in the Raft River detachment comes from a single quartzite layer, where $\delta^{18}O$ varies from 9.8 - 12.7‰ in the west to 11.7 - 13.6‰ in the east, a result inconsistent with pre-mylonitic inheritance.

Quartz-muscovite oxygen isotope fractionation is quite variable, from ~ 1 ‰ to ~ 4 ‰, with 22 out of 32 samples ranging between 2 ‰ and 3 ‰ ([Table 2, Fig. 10](#)). However, we observe that although some samples located in a similar structural position in the section have drastically different absolute $\delta^{18}O$ values, they have similar $\Delta_{(Qtz-Ms)}$ (also noted in [Gottardi et al., 2011](#)). For example,

RR07-57 has much lower $\delta^{18}\text{O}$ values ($\delta^{18}\text{O}_{\text{Qtz}} = 9.8 \text{ ‰}$, $\delta^{18}\text{O}_{\text{Ms}} = 7.4 \text{ ‰}$) than sample RR09-123 ($\delta^{18}\text{O}_{\text{Qtz}} = 13.4 \text{ ‰}$, $\delta^{18}\text{O}_{\text{Ms}} = 11.1 \text{ ‰}$) despite similar $\Delta_{(\text{Qtz-Ms})}$ of $\sim 2.4 \text{ ‰}$ (Fig. 10). If the minerals were in isotopic equilibrium, such results can only be explained by variable degrees of fluid-rock interaction at different time intervals of deformation.

7.3. PERMEABILITY OF THE DETACHMENT SHEAR ZONE

Changes in permeability of the different rock types might be a possible explanation for the variations in stable isotope data observed at Clear Creek. The basement in the western part of the study area (meta-monzogranite) is essentially undeformed, while the eastern part is underlain by a high permeability, anisotropic schist unit (Green Creek Complex). Those two rock units deformed differentially and likely had very different permeabilities during deformation: the nearly undeformed meta-monzogranite was likely less permeable than the schistose units of the Green Creek Complex. In addition, the fluid-rock exchange depends on the ratio between transport and exchange (Bowman et al., 1994): if the transport is too fast, then exchange is limited; whereas if the transport is slow, the exchange is more likely to reach equilibrium.

In the western part of the study area, as the fluids descended through the quartzite section and reached the meta-monzogranite basement, the basement-cover interface may have acted as a bottom boundary of focused flow (Fig. 11). Therefore, the overlying quartzite column is likely to be in equilibrium with a focused fluid in a confined shear-zone aquifer, resulting in isotopic equilibrium between quartz, muscovite, and fluid.

On the other hand, toward the east (Fig. 11), the more permeable basement may have generated a less focused and more diffuse flow, where the fluid likely flowed through the quartzite column and into the schistose units below; hence, the mineral phases are less likely to have equilibrated their isotope composition owing to high transport/exchange ratio. The metaconglomerate ($\delta^{18}\text{O}_{\text{Qtz}} = 11.0 \pm 0.2 \text{ ‰}$) at the interface between quartzite and basement likely acted as a conduit connecting the two styles of aquifer, and localizing fluid flow and fluid-rock exchange, hence recording more isotopic homogenization and greater depletion in ^{18}O than the overlying quartzite ($\delta^{18}\text{O}_{\text{Qtz}} = 11 \text{ ‰}$). In this model, the contrast in deformation may have influenced the permeability of

the different lithological units and therefore controlled fluid flow as represented by the isotopic variations, leading to both vertical and horizontal oxygen isotope gradients.

7.4. RELATION BETWEEN STRAIN AND STABLE ISOTOPE PATTERNS

Previous structural and microstructural analyses show that the Raft River detachment shear zone at Clear Creek records, from west to east along the tectonic transport direction, an extreme gradient in finite-strain intensity associated with the necking of the shear zone (e.g. Compton, 1980; Wells, 1997; Sullivan, 2008). Three-dimensional finite-strain and strain intensity analysis of the basal metaconglomerate (Sullivan 2008) shows that there is a transition from flattening strain in the western part to constrictional strain in the eastern part, associated with an increase in deformation intensity, that Sullivan (2008) relates to the difference in rheology of the basement units. The western part of the shear zone is developed above the rheologically strong meta-monzogranite, whereas the eastern part of the shear zone, characterized by a more intense deformation and representing the necking of the shear zone, is localized above the biotite and hornblende schist that is rheologically weaker (Fig. 2, Sullivan, 2008).

This contrast in strain and rheology controlled strain localization, which then controlled the fluid flow at the time of mylonitization. In a deforming material, the porosity and permeability tensors that control groundwater flow are sensitive to stress, resulting in fluid flow anisotropy (Morgan, 1987; Sibson, 1996; Zhu and Wong, 1997; Zhang et al., 1999). A material deforming under flattening strain develops permeability anisotropy in the direction perpendicular to the maximum shortening direction, (horizontal planes) if the maximum shortening direction is vertical. On the other hand, a material deformed in constriction will develop a linear anisotropy in the direction of maximum elongation. Therefore, flow in the flattening field will be two-dimensional (planes perpendicular to the main shortening direction), whereas it will be unidimensional in the constrictional field (lines parallel to the stretching direction).

In addition, solid-fluid stable isotope exchange results from three mechanisms: solution-precipitation, fluid-assisted chemical reaction, and diffusion (Giletti, 1985; Cole et al., 1983, 1992; Valley, 2001). These mechanisms are strongly dependent on mineral surface areas between two phases

relative to the volume of solid, which in turn depends on the shape and size of individual particles and packing arrangement of the particles in the aggregate, and stress. In general, small spherical grains have higher surface to volume ratios and will exchange faster with interstitial fluids than larger ellipsoidal grains with low surface to volume ratio (Kirschner, 1992). Therefore the change in particle shape owing to deformation and recrystallization (e.g. from spherical to ellipsoidal) must be considered when interpreting stable isotope data owing to the effects the particle size and shape have on the total surface area-to-volume ratio of the entire aggregate (Kirschner, 1992).

Based on those considerations, it is plausible that fluid flow during development of the shear zone may have followed permeability anisotropies in the deforming material. In the western region of the studied Raft River shear zone, above the rheologically strong meta-monzogranite basement, where strain was partitioned into components of both strike-parallel and strike-perpendicular elongation (Sullivan, 2008), groundwater flow would be distributed on subhorizontal planes. In contrast in the eastern region, where rheologically weak schist basement units accommodated large amounts of transport-parallel elongation, characterized by more constrictional strain (Sullivan, 2008), fluid flow would follow subhorizontal lines, parallel to the lineation (Fig. 11). This difference in flow pattern could explain the degree of fluid-rock isotope exchange. In orogenic systems, fluid flow is driven by convective circulation (hydrothermal system), tectonic compression (Oliver, 1986; Ge and Garven, 1994; McPherson and Grant, 1999), or topography, owing to rock and surface uplift (Garven and Freeze, 1984b; Woodbury and Smith, 1988; Person et al. 2007). When a constant volume of fluid circulates in the system with homogeneous permeability and porosity and flow is distributed in two-dimensions (planes), the flux of fluids will be slower than that of a one-dimensional flow (pipes). Since the fluid-rock exchange depends on the ratio between transport and exchange (Bowman et al., 1994), if the flux of fluid is too rapid, then exchange is limited; whereas if the flux is slow, the exchange is more likely to reach equilibrium.

When integrating those considerations in our model, we observe a change in permeability of basement units along the transport direction, which is also associated with a change in the finite strain ellipsoid and an increase in strain intensity. In the west, the low permeability basement acts as a bottom boundary for fluid flow, and the flow is distributed along planar fabrics, leading to isotopic

equilibrium in a confined aquifer. Toward the east, higher permeability schistose units lead to diffuse flow, distributed along linear fabrics in more a diffuse aquifer, yielding non-equilibrium oxygen isotope values. The permeability contrast between the different lithological units and strain variations along the transport direction lead to both vertical and horizontal oxygen isotope gradients.

7.5. GEOMETRY OF THE SHEAR ZONE AND TIME-INTEGRATED FLUID FLOW

The geometry of the detachment shear zone at the time of fluid infiltration also has a control on the oxygen isotope distribution. Two- and three dimensional model results (Person et al, 2007; Savard et al., 2007; Gottardi et al., 2012) show that, besides permeability contrasts, the oxygen isotope patterns are also strongly affected by the architecture, spacing, and connectivity of the fault zones that drain fluids in and out of the detachment shear zone. In such models, strong ^{18}O depletion is systematically observed in areas close to the recharge fault zone. However, it is difficult to reconstruct the geometry of exhumed natural hydrologic systems such as the Raft River detachment shear zone. It is likely that Miocene fluid circulation was controlled by a set of major (brittle) fault zones. The tip of such a fault zone may have been located in the western part of the study area, hence leading to more depleted $\delta^{18}\text{O}$ values than that of localities further away from the fault zone. However, no evidence of such major structure(s) is observed in the Clear Creek area.

Finally, the distribution of $\delta^{18}\text{O}_{\text{Qtz}}$ values in the Raft River shear zone could be explained by different time-integrated interaction between fluid and rock. Metamorphic core complexes such as the Raft River Mountains are fast exhuming structures that evolved at the brittle-ductile transition. The detachment shear zone might experience cycles of brittle fracturing followed by dynamic recrystallization (Gottardi & Teyssier, 2013). In such systems, the thermal structure, permeability, and geometry are likely transient (Weis et al., 2012). Fault zones might open and close, therefore affecting the permeability fields and the flux of fluids permeating the shear zone (Sibson, 1990, 1995, 1996). At any given time, different parts of the shear zone may or may not be connected to a meteoric fluid reservoir, and therefore experience different durations of fluid-rock interaction, yielding different oxygen isotope depletion patterns. There is microstructural evidence that the Raft River shear zone

developed close to its peak strength (Gottardi & Teyssier, 2013), however detailed oxygen isotopic work at higher resolution would be necessary to address such issues.

8. CONCLUSIONS

Combined microstructural, geochronological, and stable isotope analyses of the Raft River detachment shear zone at Clear Creek show that an extensive hydrological system was active during the evolution of the shear zone in Miocene time. The most important results are as follow.

1) Hydrogen stable isotope ratios from both muscovite and fluid inclusions show that the fluid that infiltrated the detachment shear zone was meteoric in origin, with a low D/H and low $^{18}\text{O}/^{16}\text{O}$ ratios.

2) Meteoric water infiltrated the detachment shear zone during the Miocene (likely between 20 and 15 Ma) over the time scale of mylonite formation.

3) Quartz and muscovite $\delta^{18}\text{O}$ values show different degrees of ^{18}O depletion, suggesting different time-integrated interaction of the minerals with meteoric water; these fluids would be channelized in preferential layers or shear zones within the deforming system.

4) Variation in permeability in the basement units might affect the flow pattern: confined versus diffuse flow may be responsible for the different degrees of ^{18}O depletion.

5) Strain variations from flattening (planes) to constriction (pipes) along the transport direction might affect fluid pathways resulting in different fluid-rock exchange patterns.

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FIGURES AND TABLES

Table 1: Furnace step heating $^{40}\text{Ar}/^{39}\text{Ar}$ data of muscovite separates from the Raft River detachment shear zone.

Table 2: Oxygen isotope data (Clear Creek section, Raft River) of quartz ($\delta^{18}\text{O}_{\text{Qtz}}$) and muscovite ($\delta^{18}\text{O}_{\text{Ms}}$), and fractionation coefficient $\Delta^{18}\text{O}_{\text{Qtz-Ms}}$.

Table 3: Hydrogen Isotope data from Fluid Inclusions (Clear Creek Canyon, Raft River), $\delta\text{D}_{\text{FI}}$ (‰) vs. VSMOW

Table 4: Hydrogen Isotope data from muscovite (Clear Creek Section, Raft River), $\delta\text{D}_{\text{Ms}}$ (‰)

Figure 1: Simplified geologic map of the Raft River metamorphic core complex showing the location of Fig. 2. Adapted from Wells, (1997, 2000), and Sullivan (2008).

Figure 2: Detailed map of the study area, the Clear Creek canyon, modified after Sullivan (2008).

Figure 3: Oxygen stable isotope data of quartz ($\delta^{18}\text{O}_{\text{Qtz}}$) and muscovite ($\delta^{18}\text{O}_{\text{Ms}}$) from the Elba quartzite at Clear Creek, plotted as a function of elevation (m above the basement). (a) data published by Gottardi et al. (2011); (b) Section C; and (c) Section E. See Fig. 2 for location of sections.

Figure 4: Quartz microstructures. (a) typical quartz microstructures from the Raft River shear zone; (b) close-up of the microstructures, showing that the main recrystallization process is subgrain rotation; (c) deformation lamellae (shown by arrows) are common; (d) shear bands are locally present. Thin sections cut perpendicular to foliation and parallel to lineation, the photomicrographs are taken oriented top to the East.

555

556 **Figure 5:** Fluid inclusion microstructures: (a) intergranular fluid inclusions, located at grain and
557 subgrain boundaries (white arrows). These are the most common type of fluid inclusions observed in
558 quartz from the Elba quartzite; (b) fluid inclusion trails that cross-cut multiple grains (white arrow); (c)
559 fluid inclusions associated with deformation lamellae; (d) fluid inclusions that form subparallel planes
560 as frequently observed close to the “notch”. Thin sections cut perpendicular to foliation and parallel to
561 lineation, the photomicrographs are taken oriented top to the East.

562

563 **Figure 6:** Mica microstructures. (a) the Elba quartzite contains ~5 to 10 % of muscovite; (b) mica with
564 high aspect ratio and inverted stair stepping; (c) rare mica fish. Thin sections cut perpendicular to
565 foliation and parallel to lineation, the photomicrographs are taken oriented top to the East.

566

567 **Figure 7:** $^{40}\text{Ar}/^{39}\text{Ar}$ step-heating spectra of muscovite from four mylonitic quartzites (Section C)
568 distributed over the footwall of the Raft River detachment shear zone.

569

570 **Figure 8:** Spatial pattern of oxygen stable isotope ratios from the basal metaconglomerate. Two
571 samples were cut into 1 mm³ cubes and analyzed. Figure shows $\delta^{18}\text{O}$ of matrix quartz (italic) and
572 deformed quartz pebbles (regular).

573

574 **Figure 9:** δD vs. elevation, from both muscovite and fluid inclusions. Very variable δD values are
575 obtained at and near the “notch”.

576

577 **Figure 10:** $\delta^{18}\text{O}$ of muscovite versus $\delta^{18}\text{O}$ of quartz values from mylonitic quartzite (Raft River
578 detachment shear zone). Temperatures correspond to lines of constant oxygen isotope fractionation.

579

580 **Figure 11:** Conceptual model to explain vertical and horizontal isotopic variability observed in the
581 Raft River detachment shear zone. In the West, flattening fabrics (planes) and low permeability
582 basement (Adamellite) lead to a focused flow in a confined aquifer, enhancing fluid-rock exchange. In

583 the East, constrictional fabrics (pipes) and high permeability basement (schists) lead to a more diffuse
584 flow.
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Figure 2

Figure 3**a. Clear Creek - Section 1-5***(Gottardi et al., 2011 data)* **b. Clear Creek - Section C** **c. Clear Creek - Section E**

Figure 8

Figure 8 shows a grid of 10 rows and 10 columns of squares. The squares are arranged in a grid pattern, with some squares containing numerical values. The values are as follows:

Row	Col 1	Col 2	Col 3	Col 4	Col 5	Col 6	Col 7	Col 8	Col 9	Col 10
1				11.1						
2				11.1						
3								11.0		
4	10.8			11.1						
5						11.2		11.1		
6			11.1		11.0					
7						11.1		11.2	11.1	
8	11.1		11.2							
9						11.1	11.1		11.1	
10			11.0							
11		11.1		11.0	11.7	11.0		11.0		11.0
12		11.1								
13										
14						11.0

Figure 9

Elevation (m above basement)

Figure 10

West

East

Table 1

Table 1. ⁴⁰Ar/³⁹Ar analytical data.

ID	Temperature (°C)	⁴⁰ Ar/ ³⁹ Ar	³⁷ Ar/ ³⁹ Ar	³⁶ Ar/ ³⁹ Ar (x 10 ⁻³)	³⁹ Ar _K (x 10 ⁻¹⁵ mol)	K/Ca	⁴⁰ Ar* (%)	³⁹ Ar (%)	Age (Ma)	±1σ (Ma)
<i>RR09-91, Muscovite, 2.43 mg, J=0.0016011±0.06%, D=1.005±0.001, NM-233K, Lab#=59589-01</i>										
A	600	61.290	0.290	170.863	1.014	1.76	17.7	1.4	31.00	1.66
B	650	9.097	0.061	11.962	1.001	8.33	61.2	2.7	15.99	0.68
C	700	7.027	0.022	5.604	1.897	23.24	76.4	5.3	15.43	0.35
D	750	6.439	0.011	3.710	3.164	45.94	83.0	9.5	15.34	0.19
E	800	6.790	0.013	3.508	4.919	40.75	84.7	16.1	16.52	0.14
F	840	8.534	0.009	3.767	9.596	57.42	86.9	29.1	21.28	0.12
G	880	7.693	0.006	1.662	14.326	83.22	93.6	48.3	20.66	0.06
H	920	7.021	0.007	1.862	9.813	68.42	92.2	61.5	18.57	0.08
I	980	7.297	0.004	2.785	8.774	113.84	88.7	73.4	18.58	0.11
J	1050	8.262	0.005	3.778	6.349	105.81	86.5	81.9	20.49	0.12
K	1110	8.818	0.003	3.230	8.851	199.10	89.2	93.8	22.55	0.11
L	1170	10.183	0.004	2.604	4.599	114.31	92.4	100.0	26.96	0.17
Integrated age ± 1σ			n=12		74.302	44.41	K2O=7.34%		20.26	0.06
Plateau ± 1σ		steps F-K	n=6	MSWD=247.64	57.709		77.7		20.23	0.59

Table 1. $^{40}\text{Ar}/^{39}\text{Ar}$ analytical data.

ID	Temperature (°C)	$^{40}\text{Ar}/^{39}\text{Ar}$	$^{37}\text{Ar}/^{39}\text{Ar}$	$^{36}\text{Ar}/^{39}\text{Ar}$ (x 10 ⁻³)	$^{39}\text{Ar}_\text{K}$ (x 10 ⁻¹⁵ mol)	K/Ca	$^{40}\text{Ar}^*$ (%)	^{39}Ar (%)	Age (Ma)	$\pm 1\sigma$ (Ma)
<i>RR09-94, Muscovite, 9.07 mg, $J=0.0015948\pm0.06\%$, $D=1.005\pm0.001$, NM-233K, Lab#=59588-01</i>										
A	550	154.634	0.176	432.595	2.116	2.90	17.3	0.8	75.54	1.95
B	600	15.241	0.207	24.446	2.306	2.47	52.7	1.6	22.94	0.50
C	650	9.372	0.079	13.709	4.250	6.45	56.8	3.1	15.24	0.26
D	700	7.219	0.021	6.055	8.068	24.79	75.2	6.0	15.53	0.14
E	750	6.903	0.013	4.076	12.873	39.77	82.5	10.5	16.30	0.11
F	800	8.356	0.012	3.324	18.354	41.59	88.2	17.0	21.06	0.08
G	850	9.933	0.007	3.068	49.629	70.18	90.9	34.7	25.76	0.05
H	880	8.791	0.007	1.245	41.220	73.77	95.8	49.3	24.05	0.05
I	900	8.616	0.007	1.125	26.243	77.29	96.1	58.7	23.65	0.07
J	930	8.747	0.007	1.615	21.893	75.10	94.5	66.5	23.61	0.07
K	960	8.874	0.005	1.770	16.742	94.02	94.1	72.4	23.84	0.08
L	990	9.455	0.004	2.291	12.755	124.86	92.8	76.9	25.05	0.10
M	1020	10.060	0.003	2.374	9.793	166.97	93.0	80.4	26.70	0.12
N	1040	10.455	0.003	1.899	7.519	147.05	94.6	83.1	28.21	0.14
O	1060	10.652	0.003	1.674	6.528	200.07	95.4	85.4	28.96	0.18
P	1080	10.837	0.002	1.409	6.271	207.51	96.2	87.6	29.71	0.17
Q	1110	11.523	0.002	3.004	8.219	212.77	92.3	90.6	30.32	0.15
R	1140	10.859	0.002	2.200	9.805	273.67	94.0	94.0	29.11	0.12
S	1170	10.914	0.001	1.514	10.644	424.39	95.9	97.8	29.84	0.12
T	1650	96.589	0.007	121.610	6.109	71.45	62.8	100.0	166.55	0.82
Integrated age $\pm 1\sigma$			n=20		281.340	47.29		K2O=7.47%	28.01	0.06
Plateau $\pm 1\sigma$		steps G-L	n=6	MSWD=210.44	168.484			59.9	24.43	0.39

Table 1. $^{40}\text{Ar}/^{39}\text{Ar}$ analytical data.

ID	Temperature (°C)	$^{40}\text{Ar}/^{39}\text{Ar}$	$^{37}\text{Ar}/^{39}\text{Ar}$	$^{36}\text{Ar}/^{39}\text{Ar}$ (x 10 ⁻³)	$^{39}\text{Ar}_\text{K}$ (x 10 ⁻¹⁵ mol)	K/Ca	$^{40}\text{Ar}^*$ (%)	^{39}Ar (%)	Age (Ma)	$\pm 1\sigma$ (Ma)
<i>RR09-96, Muscovite, 8.79 mg, $J=0.0015921\pm0.07\%$, $D=1.005\pm0.001$, NM-233K, Lab#=59587-01</i>										
A	550	70.598	0.021	68.478	1.694	24.86	71.3	0.7	139.13	0.89
B	600	18.744	0.045	20.811	2.517	11.34	67.2	1.8	35.80	0.41
C	650	8.006	0.009	8.801	4.823	55.27	67.5	3.8	15.43	0.18
D	700	6.713	0.002	4.323	9.012	207.75	80.9	7.6	15.52	0.10
E	750	6.737	0.002	2.815	14.095	232.09	87.6	13.5	16.85	0.08
F	800	8.217	0.002	2.139	19.695	313.72	92.3	21.7	21.63	0.06
G	850	10.025	0.001	1.922	38.325	566.26	94.3	37.8	26.93	0.06
H	880	9.306	0.001	1.013	35.661	780.49	96.8	52.7	25.66	0.05
I	900	9.302	0.001	1.134	23.866	834.35	96.4	62.7	25.55	0.07
J	930	9.603	0.000	1.351	24.397	1766.02	95.8	72.9	26.22	0.06
K	960	10.333	0.000	1.819	12.995	1494.45	94.8	78.4	27.89	0.09
L	990	11.331	0.001	2.327	9.668	843.87	93.9	82.4	30.29	0.12
M	1020	12.433	0.000	2.810	7.823	2270.60	93.3	85.7	32.99	0.13
N	1040	13.278	0.000	3.323	5.940	-	92.6	88.2	34.95	0.17
O	1060	13.845	0.001	3.785	4.987	616.93	91.9	90.3	36.16	0.19
P	1080	14.224	0.000	3.893	4.800	-	91.9	92.3	37.14	0.20
Q	1110	14.321	0.001	3.779	6.743	393.65	92.2	95.1	37.51	0.15
R	1140	13.321	0.001	3.922	6.247	701.12	91.3	97.7	34.57	0.17
S	1170	13.408	0.001	4.692	5.383	644.36	89.7	100.0	34.17	0.19
Integrated age $\pm 1\sigma$			n=19		238.670	314.84		K2O=6.55%	27.40	0.05
Plateau $\pm 1\sigma$			steps G-K	n=5	MSWD=180.90	135.244		56.7	26.27	0.37

Table 1 (continued). $^{40}\text{Ar}/^{39}\text{Ar}$ analytical data.

ID	Temperature (°C)	$^{40}\text{Ar}/^{39}\text{Ar}$	$^{37}\text{Ar}/^{39}\text{Ar}$	$^{36}\text{Ar}/^{39}\text{Ar}$ ($\times 10^{-3}$)	$^{39}\text{Ar}_K$ ($\times 10^{-15}$ mol)	K/Ca	$^{40}\text{Ar}^*$ (%)	^{39}Ar (%)	Age (Ma)	$\pm 1\sigma$ (Ma)
<i>RR09-98, Muscovite, .71 mg, $J=0.0016023\pm0.05\%$, $D=1.005\pm0.001$, NM-233K, Lab#=59591-01</i>										
A	650	50.440	0.425	139.487	0.264	1.20	18.4	2.7	26.58	3.16
B	750	7.363	0.030	4.463	0.617	16.77	82.1	9.1	17.37	1.05
C	800	9.761	0.011	4.712	0.716	46.72	85.7	16.5	24.00	0.83
D	850	13.330	0.005	2.432	2.036	113.09	94.6	37.5	36.07	0.35
E	900	9.174	0.006	1.840	2.010	82.93	94.1	58.2	24.75	0.31
F	970	9.200	0.011	1.611	1.371	46.63	94.8	72.4	25.02	0.49
G	1060	12.671	0.011	5.010	0.839	45.53	88.3	81.0	32.04	0.88
H	1150	16.822	0.003	5.849	1.437	156.59	89.8	95.9	43.10	0.54
I	1220	16.905	0.022	18.301	0.355	23.00	68.0	99.5	32.91	2.14
J	1650	1819.047	0.176	5543.521	0.045	2.91	9.9	100.0	459.34	44.55
Integrated age $\pm 1\sigma$			n=10		9.689	24.08		K2O=3.27%	32.61	0.32
Plateau $\pm 1\sigma$		steps E-F	n=2	MSWD=0.21	3.380			34.9	24.83	0.26

Table 1 (continued). $^{40}\text{Ar}/^{39}\text{Ar}$ analytical data.

ID	Temperature (°C)	$^{40}\text{Ar}/^{39}\text{Ar}$	$^{37}\text{Ar}/^{39}\text{Ar}$	$^{36}\text{Ar}/^{39}\text{Ar}$ (x 10 ⁻³)	$^{39}\text{Ar}_\text{K}$ (x 10 ⁻¹⁵ mol)	K/Ca	$^{40}\text{Ar}^*$ (%)	^{39}Ar (%)	Age (Ma)	$\pm 1\sigma$ (Ma)
<i>RR09-104, Muscovite, 1.63 mg, $J=0.0016051\pm0.06\%$, $D=1.005\pm0.001$, NM-233K, Lab#=59590-01</i>										
A	600	55.282	0.181	44.135	1.343	2.83	76.4	4.1	118.38	1.08
B	650	8.345	0.120	10.392	0.880	4.25	63.3	6.8	15.21	0.95
C	700	7.069	0.016	6.453	0.948	32.21	73.0	9.7	14.86	0.93
D	750	7.474	0.004	4.891	1.220	114.78	80.6	13.5	17.35	0.70
E	800	11.856	0.006	5.350	1.462	84.36	86.7	17.9	29.48	0.58
F	840	12.280	0.004	3.648	1.687	144.15	91.2	23.1	32.12	0.49
G	880	13.252	0.003	4.526	2.757	156.24	89.9	31.5	34.15	0.33
H	920	11.128	0.000	2.875	3.766	7267.86	92.4	43.1	29.49	0.23
I	980	11.868	0.000	3.158	4.574	2170.16	92.1	57.1	31.36	0.21
J	1050	20.801	0.000	3.764	4.295	-	94.7	70.2	56.10	0.25
K	1050	28.774	0.000	3.785	1.804	-	96.1	75.8	78.33	0.52
L	1650	65.382	0.002	32.330	7.908	271.67	85.4	100.0	154.80	0.42
Integrated age $\pm 1\sigma$			n=12		32.644	41.08		K2O=4.79%	70.01	0.17
Plateau $\pm 1\sigma$		steps E-I	n=5	MSWD=36.17	14.247			43.6	31.15	0.80

Table 2: Oxygen Isotope data (Clear Creek section, Raft River).

Elevation (m above basement)	Sample	$\delta^{18}\text{O}_{\text{Qtz}}$ (‰)	$\delta^{18}\text{O}_{\text{Ms}}$ (‰)	$\Delta^{18}\text{O}_{\text{Qtz-Ms}}$ (‰)
<i>Gottardi et al (2011) GEOLOGY data (Section 1-5)</i>				
90	RR-07-068	11.3	7.7	3.7
75	RR-07-092	12.7	9.4	3.4
65	RR-07-077	10.6	7.5	3.1
40	RR-07-043	12.2	9.4	2.8
25	RR-07-071	9.8	7.1	2.7
15	RR-07-040	10.2	7.8	2.5
8	RR-07-057	9.8	7.4	2.4
<i>Section C</i>				
58	RR09-103	12.3	11.3	1.0
25	RR09-098	12.4	10.5	1.9
18	RR09-097	12.7	9.9	2.8
16	RR09-096	12.6	10.5	2.1
13	RR09-095	12.5	11.0	1.6
12	RR09-094	12.7	10.6	2.2
11	RR09-093	12.5	10.6	1.9
9	RR09-092	12.0	9.8	2.3
8	RR09-091	12.3	9.9	2.3
4	RR09-090	13.2	12.0	1.2
<i>Section E</i>				
64	RR09-136	11.7	10.5 [†]	1.2
59	RR09-135	12.6	10.6	2.0
55	RR09-134	12.7	11.5 [†]	1.2
49	RR09-133	12.7	9.5	3.3
36	RR09-132	13.0 [†]	12.1 [†]	0.9
29	RR09-131	12.8	11.5 [†]	1.3
23	RR09-130	12.8	10.8	2.0
18	RR09-129	12.8	10.2	2.6
16	RR09-128	12.8	10.5 [†]	2.3
15	RR09-127	12.8	10.6 [†]	2.2
14	RR09-126	12.6	9.6	3.1
12	RR09-125	12.7 [†]	10.7 [†]	2.0
10	RR09-124	13.1 [†]	11.6 [†]	1.4
8	RR09-123	13.4 [†]	11.1	2.3
6	RR09-122	13.6 [†]	11.5 [†]	2.1

[†] single analysis

Table 3: Hydrogen Isotope data from Fluid Inclusions (Clear Creek Section, Raft River)

† Published in Gottardi et al. (2011)

Sample	Lithology	Elevation (m above basement)	H ₂ O (μ L)	$\delta D_{FI}(\text{‰})$ vs. SMOW
<i>Section 1</i>				
RR-07-046 [†]	Quartz vein	46	0.188	-83
RR07-042	Quartz vein	35	0.484	-104
RR07-041 [†]	Quartzite	28	0.134	-89
<i>Section 4</i>				
RR-07-068 [†]	Quartzite	90	0.129	-93
RR07-067	Quartz vein	80	0.234	-103
RR07-059	Quartz vein	25	0.521	-83
RR-07-057	Quartz vein	8	0.247	-88
RR07-057 [†]	Quartzite	8	0.247	-94
<i>Section 5</i>				
RR07-098	Quartzite	90	0.655	-99
RR-07-087 [†]	Quartzite	0	0.229	-84

Table 4: Hydrogen stable isotope data from Muscovite (δD_{Ms})*Bold font denotes samples dated by $^{40}Ar/^{39}Ar$*

Elevation (m above basement)	Sample	δD_{Ms} (‰)
<i>Section A</i>		
51	RR09-158	-89
32	RR09-156	-110
27	RR09-155	-106
21	RR09-154	-111
17	RR09-153	-122
13	RR09-152	-94
8	RR09-151	-93
3	RR09-150	-109
<i>Section B</i>		
61	RR09-86	-112
43	RR09-83	-107
38	RR09-82	-109
30	RR09-80	-106
29	RR09-79	-109
27	RR09-78	-111
19	RR09-77	-109
15	RR09-76	-98
6	RR09-74	-108
2	RR09-73	-114
<i>Section C</i>		
58	RR09-104	-97
25	RR09-98	-116
18	RR09-97	-110
16	RR09-96	-109
13	RR09-95	-108
12	RR09-94	-96
11	RR09-93	-91
9	RR09-92	-88
8	RR09-91	-90
4	RR09-90	-104
<i>Section D</i>		
55	RR09-119	-117
52	RR09-118	-104
49	RR09-117	-107
45	RR09-115	-113
25	RR09-113	-112
22	RR09-112	-111
20	RR09-111	-107
18	RR09-110	-99
15	RR09-109	-100
<i>Section E</i>		
68	RR09-137	-114
64	RR09-136	-105
59	RR09-135	-112
55	RR09-134	-110
49	RR09-133	-123
36	RR09-132	-104
29	RR09-131	-110
23	RR09-130	-112
18	RR09-129	-101
16	RR09-128	-103
15	RR09-127	-107
14	RR09-126	-111
10	RR09-124	-102
8	RR09-123	-106