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Strontium, Lead, and Oxygen Isotopic Signatures of Mid-Miocene Silicic Volcanism in Eastern Oregon

Emily Nancy Hess
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Strontium, Lead, and Oxygen Isotopic Signatures of Mid-Miocene Silicic
Volcanism in Eastern Oregon

by

Emily Nancy Hess

A thesis submitted in partial fulfillment of the
requirements for the degree of

Master of Science
in
Geology

Thesis Committee:
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Portland State University
2014

ABSTRACT

Widespread, mid-Miocene rhyolite volcanism of eastern Oregon that are coeval or slightly postdate flood basalts of the Columbia River Basalt Province allows for mapping crustal domains using radiogenic and stable isotopes. Rhyolites are thought to be derived in large part by partial melting of the crust and thus yield direct information on the composition of the crust. Silicic volcanism is expressed in the form of numerous domes and tuffs exposed over a wide area (~300 km in N-S dimension and ~200 km in E-W dimension) west of the presumed craton boundary, which runs parallel but mostly east of the Oregon-Idaho state border as delineated by geophysical characteristics and isotopic transitions, including the $^{87}\text{Sr}/^{86}\text{Sr}_i = 0.7060$ line (MSL) and $^{87}\text{Sr}/^{86}\text{Sr}_i = 0.7040$ line (CSL).

$^{87}\text{Sr}/^{86}\text{Sr}_i$ of twenty-seven silicic units are variable and some are high. Sr isotopic ratios are inconsistent with the location of the traditional MSL and CSL boundaries. A primary control on the $^{87}\text{Sr}/^{86}\text{Sr}_i$ isotope variations may reflect changes in the crustal make-up of Paleozoic accreted terranes of a particular area rather than arising from a westward-dipping decollement that moved cratonic lithosphere below accreted terranes in eastern Oregon. A secondary control on observed isotopic ratios may be related to the amount and composition of basalt involved in the generation of rhyolites. This could lead to higher or lower $^{87}\text{Sr}/^{86}\text{Sr}_i$ relative to the surrounding crust because de facto coeval mafic magmas of the Columbia River Basalt Group have a wide range of Sr isotopic signatures.

While Pb isotope data is incomplete for all samples of this study, the available data indicate a significant range in Pb isotopes. Yet, data of individual regions tend to

plot close to one another relative to the entire data distribution. Comparison of samples from this study in a more regional view indicates the samples generally fall within the previously defined lead isotope boundaries of the main-phase Columbia River Basalt Group lavas.

$\delta^{18}\text{O}$ values range from below 2‰ to above 9‰. In addition, there is a crude trend of rhyolites having lower $\delta^{18}\text{O}$ and more radiogenic $^{87}\text{Sr}/^{86}\text{Sr}_i$ ratios. The lowest oxygen ratios (<2‰) are found in rhyolites ~80 km west of the cratonic margin, potentially reflecting remelting or assimilation of hydrothermally altered crust. Low $\delta^{18}\text{O}$ of selected rhyolite flows cannot be explained by remelting of Cretaceous plutons of the Idaho Batholith and appear irreconcilable with remelting of altered silicic rocks at centers of multiple, confocal caldera cycles— both processes that have been proposed to explain low $\delta^{18}\text{O}$ of rhyolites of the Snake River Plain–Yellowstone area.

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INTRODUCTION

During mid-Miocene time, eruption of voluminous flood basalts and widespread eruptions of rhyolite ash flows and lava flows occurred in eastern Oregon, northern Nevada, and southwestern Idaho (Cummings et al., 2000; Hooper et al., 2002; Shervais and Hanan, 2008; Camp, 2013). At least $\sim 3,900 \text{ km}^3$ of silicic magma erupted from centers covering over $\sim 25,000 \text{ km}^2$ (Coble and Mahood, 2012). While numerous origins, source materials, and contributing components have been proposed for this volcanism, the majority of workers agree that a mantle plume produced the Columbia River Basalt Group (CRBG) and the Snake River Plain – Yellowstone hotspot track to the east (e.g., Wolff et al., 2008; Camp, 2013).

The McDermitt volcanic field, located on the border of Nevada and Oregon, has been proposed as the initiation site of plume head magmatic activity related to the Yellowstone hotspot track with the continental crust (Pierce and Morgan, 2009; Coble and Mahood, 2012; Camp, 2013). $^{40}\text{Ar}/^{39}\text{Ar}$ ages of silicic rocks around McDermitt caldera are $16.54 \pm 0.11 \text{ Ma}$ for a biotite rhyolite and $16.49 \pm 0.04 \text{ Ma}$ for a peralkaline rhyolite (Henry et al., 2006). However, there are numerous silicic centers further north with ages as old as that of the McDermitt center, thus extending the potential crustal interaction area of plume head activity considerably. With a proposed diameter of $\sim 400 \text{ km}$, the initial impingement of the Yellowstone plume head could have underlied much of eastern Oregon, northern Nevada, and southwestern Idaho (Coble and Mahood, 2012).

Flood basalts and associated rhyolitic volcanism occurred in an area of significant compositional changes of the lithosphere. Approximately 116°W is the boundary that

separates Precambrian continental crust underlain by old lithospheric mantle from Mesozoic accreted oceanic terranes to the west that were added to the North American continent during the Cretaceous (Leeman et al., 1992). Armstrong et al. (1977) were the first to propose that the cratonic margin is preserved as a sharp isotopic boundary as defined by the $^{87}\text{Sr}/^{86}\text{Sr}_i = 0.7060$ line (Figure 3). More recent work has further refined the location of the “0.706 line,” which was defined by Armstrong et al. (1977) as the Mesozoic Strontium Line (MSL) (e.g., Kistler and Peterman, 1978; Fleck and Criss, 1985; Leeman et al., 1992; Pierce and Morgan, 1992; Fleck and Criss, 2004; Pierce and Morgan, 2009). This isotopic transition from Precambrian crust to Paleozoic accreted terranes was further delineated by the $^{87}\text{Sr}/^{86}\text{Sr}_i = 0.7040$ (Cenozoic Strontium Line [CSL]; Armstrong et al., 1977). The CSL is thought to parallel the MSL in central Idaho (Figure 1) but diverges west through eastern Oregon to the south (Shervais and Hanan, 2008). Divergence of both boundaries results from the more gradational isotopic transition across this lithospheric boundary in southern Oregon compared to further north. In areas where no appropriate crustal sections are exposed, magmas that formed and/or interacted with the lithosphere on either side of the transition have been used as evidence for the presence of accreted terrane, Precambrian craton, or transitional lithosphere (e.g., Leeman and Whelan, 1983). In this sense, silicic magmas and unfractionated basaltic magmas provide likely complementary evidence. The former characterizes crustal sources while the latter tracks mantle sources (e.g., Savov et al., 2009). In the case of fractionated mafic magmas, crustal, as well as mantle sources, can be mirrored in their isotopic compositions (e.g., Wolff et al., 2008; Wolff and Ramos, 2013).

Widespread rhyolitic volcanism near the lithospheric transitions in eastern Oregon allows for mapping crustal domains using radiogenic isotopes. Rhyolites of Oregon are thought to be derived in large part by partial melting of the crust, and thus, yield direct information on the composition of the crust (e.g., Streck, 2002; Streck and Grunder, 2008; Ford et al., 2013; Steiner and Streck, 2013). Mid-Miocene silicic volcanism is expressed in the form of numerous domes and tuffs exposed over a wide area (~300 km in N-S dimension and ~200 km in E-W dimension, Figure 1) west of the craton boundary (Cummings et al., 2000; Camp et al., 2003; Ferns and McClaughry, 2013).

This study reports $^{87}\text{Sr}/^{86}\text{Sr}_i$ ratios of groundmass and feldspar mineral separates obtained from samples of twenty-seven, mostly mid-Miocene silicic volcanic centers in eastern Oregon. In addition, Pb and O isotopes are included to gain insight into source materials for rhyolites. To date, only sparse isotopic data of a few such silicic centers exist (Leeman et al., 1992; Ford, 2011).

This study also reports the age of select rhyolite centers using $^{40}\text{Ar}/^{39}\text{Ar}$ dating. The silicic geologic units in this area are poorly known and little modern age data exist. If these rhyolites are contemporaneous in age with those found in the McDermitt-Virgin Valley area, these silicic volcanic centers likely mark the area where magmatic reservoirs feeding CRBG flood basalts resided. Most of the volume of CRBG lava (Steens, Imnaha, and Grande Ronde basalt members) erupted between 16.7 and 15.6 Ma (Wolff et al., 2008; Tolan et al., 2009; Barry et al., 2013).

Figure 1. Location map of 17 to ~10 Ma rhyolites in red (after Christiansen et al., 2002; Ma et al., 2009; Steiner and Streck, 2013) and basaltic lava flows of the CRBG in green (after Camp and Ross, 2004). Dark brown dashed lines depict terrane boundaries (after LaMaskin et al., 2009). Included are the MSL (dark blue) and the CSL (light blue) modified from Pierce and Morgan (2009), which demarcates the western edge of North American craton. White box shows the areal coverage of Figure 29.

BACKGROUND

Geologic Setting

Accreted Terranes

Prior to the development of an active continental margin in Cretaceous time, numerous oceanic and craton-fringing crustal blocks were accreted onto the western Laurentian margin of North America as allochthonous and para-autochthonous terranes (DeCelles, 2004; Dickinson, 2004). Several competing tectonic models for the evolution of the western U.S. Cordillera exist and estimates for the age of accretion vary (LaMaskin, 2009). The traditionally defined terranes within the Blue Mountains Province are, from north to south, the Wallowa, Baker, Olds Ferry, and Izee terranes (Figure 2). These late Paleozoic and early Mesozoic terranes were accreted in the late Mesozoic (Ferns and McClaughry, 2013). These blocks were variably faulted, folded, and intruded by middle Mesozoic granitic bodies prior to accretion (Ferns and McClaughry, 2013).

The rocks of the Wallowa, Baker, and Olds Ferry terranes originated from oceanic or island arc settings. (Brooks, 1979; Ferns and McClaughry, 2013). Some terrane material has recently been proposed to originate from being deposited after the arc had been accreted to North America - including the rocks of the Izee terrane (Dorsey and LaMaskin, 2007). The terranes are separated either by faults or unconformities that trend to the east or northeast and tend to converge northward towards the western Idaho shear zone (Brooks and Vallier, 1978). Dorsey and LaMaskin (2007) propose the Wallowa and Olds Ferry terranes were separate arcs that were active at the same time due to opposite dipping subduction of an intervening oceanic plate. The collision of the terranes began in

the Late Triassic with final accretion taking place in the Late Jurassic (Dorsey and LaMaskin, 2007). They suggest the Baker terrane is a composite of the accretionary prisms that formed in front of each arc during subduction. Meanwhile, because the Izee terrane is a regional overlap assemblage that formed after the three tectonic terranes had been united, it should not be referred to as a terrane (Dorsey and LaMaskin, 2007).

Figure 2. Map illustrating distribution of accreted terranes, the Idaho batholith, and Cenozoic cover of the Blue Mountains province from Tumpane (2010) with $^{87}\text{Sr}/^{86}\text{Sr}_i = 0.706$ line modified from Armstrong et al. (1977).

The Wallowa Terrane is a deformed block of Permian and Triassic oceanic island arc composed of volcanic rocks and age-equivalent basinal sedimentary rocks (Ferns and McClaughry, 2013). Its base consists of a complexly deformed mass of lavas, domes, mudflow breccias, and tuffs (Ferns and McClaughry, 2013).

The Baker Terrane separates the Wallowa and Olds Ferry terranes and represents a complexly deformed series of crustal blocks segregated into three components (LaMaskin, 2009). The Bourne subterrane is the northernmost component and consists largely of weakly metamorphosed rocks of oceanic affinity (Ferns and Brooks, 1995). The Greenhorn subterrane is dominated by serpentinite-matrix *mélange* (Ferns and McClaughry, 2013). The more highly metamorphosed greenschist-facies rocks of the Burnt River Schist that compose the southernmost component (Gilluly, 1937; Kays et al., 1987) are oceanic and island arc fragments (Ferns and McClaughry, 2013). In the Baker terrane, metaplutonic rocks have $^{87}\text{Sr}/^{86}\text{Sr}_i = 0.7033 - 0.7034$ while siliceous argillites have $^{87}\text{Sr}/^{86}\text{Sr}_i = 0.7073 - 0.7094$ (Schwartz et al., 2010).

The most inboard of the terranes of the Blue Mountains province, and thus, in the most direct contact with cratonic North America is the Olds Ferry terrane (Tumpane, 2010). It consists of carbonate and clastic sedimentary rocks and associated metavolcanic rocks, as well as plutonic rocks of the Huntington Formation (Tumpane, 2010). The most common volcanic rock type is meta-andesite with a submarine origin, as evidenced by the presence of fossiliferous marine clastic beds (Brooks and Vallier, 1978). The Huntington Formation was deposited unconformably on top of plutonic basement rocks of Middle and Late Triassic age and resulted from volcanic activity in the Late Triassic to Early

Jurassic (Tumpane, 2010). Based on field evidence, geochronology, and geochemistry, Tumpane (2010) split the formation into the lower and upper Huntington Formation. In the Huntington Formation, $^{87}\text{Sr}/^{86}\text{Sr}_i$ ranges from 0.7036 to 0.7057 (Tumpane, 2010).

The Izee terrane does not represent a far-traveled terrane because the rocks were deposited over older rocks and structures of the other terranes (LaMaskin, 2009). According to Dorsey and LaMaskin (2007), the Izee basin formed and subsided deeply on crust from the amalgamated terranes during Early to Late Jurassic time. The growing marine basin provided accommodation space for sediment deposition in a non-volcanic collisional basin. The tectonic closure of a backarc basin, growth of the Cordilleran thrust belt, regional metamorphism, and crustal thickening and mountain building in western Idaho and Nevada occurred in response to terrane-continent collision (Dorsey and LaMaskin, 2007).

In summary, the isotopic signatures of these terranes vary considerably with the Olds Ferry terrane having $^{87}\text{Sr}/^{86}\text{Sr}_i$ ratios between the less radiogenic metaplutonic rocks of the Baker terrane and the more radiogenic siliceous argillites of the Baker terrane.

Distribution of Initial Strontium Isotope Ratios – Historic Perspective

The margin of cratonic North America in Idaho was located at about 116°W longitude at the time of terrane accretion (Fleck and Criss, 2004). Armstrong et al. (1977) were the first to recognize this margin as a sharp isotopic boundary separating Precambrian crust to the east from younger Late Paleozoic and Mesozoic accreted lithosphere of oceanic affinity to the west (Figure 3). Armstrong et al. (1977) examined

granitic rocks emplaced during the Mesozoic and suggested the isotopic change resulted from contamination and assimilation processes involving magmas from the mantle and enclosing crustal rocks.

Figure 3. Map from Armstrong et al. (1977) showing $^{87}\text{Sr}/^{86}\text{Sr}_i$ for Mesozoic and Cenozoic igneous rocks in the northwestern United States (references used are listed in Armstrong et al., 1977).

Many publications have further refined the location of the MSL (Kistler and Peterman, 1978; Fleck and Criss, 1985; Leeman et al., 1992; Fleck and Criss, 2004) and the CSL, which is a similar lithospheric boundary that is coincident with the $^{87}\text{Sr}/^{86}\text{Sr}_i = 0.7060$ line in central Idaho but propagates west through eastern Oregon to the south (Shervais and Hanan, 2008). Basalts that erupt within the transition zone between the MSL and CSL tend to have $^{87}\text{Sr}/^{86}\text{Sr}_i$ ratios ranging from 0.7040 to 0.7060 (Figure 4). The range in $^{87}\text{Sr}/^{86}\text{Sr}_i$ of those rocks is thought to reflect isotopic heterogeneities in

source regions (Leeman and Whelan, 1983). Further west of the CSL, $^{87}\text{Sr}/^{86}\text{Sr}_i$ ratios of most basalts are < 0.7040 (Shervais and Hanan, 2008). Although understudied, sparse prior data on rhyolites in eastern Oregon indicate greater $^{87}\text{Sr}/^{86}\text{Sr}_i$ than basalts in eastern Oregon (Figure 4; Savov et al., 2009). While both the Dinner Creek Tuff and Littlefield Rhyolite lie west of the MSL, they retain $^{87}\text{Sr}/^{86}\text{Sr}_i = 0.7066$ (Leeman et al., 1992). These data call into question the sharpness of the isotopic transitions between accreted terranes and craton.

Figure 4. East-west variations in $^{87}\text{Sr}/^{86}\text{Sr}_i$ of silicic volcanic rocks (Savov et al., 2009). Samples are: Yellowstone intracaldera (Y-IC) and extracaldera (Y-EC) rhyolites, Bruneau-Jarbridge (BJ) rhyolites, eastern Oregon (EOR) rhyolites; miscellaneous (Misc) samples include Craters of the Moon (COM) latite and Big Southern Butte (BSB) rhyolite. Vertical dashed line shows position of western Idaho suture zone (WISZ). Gray shading indicates compositional range for Neogene basalts from the High Lava Plain and Snake River Plain (after Savov et al., 2009).

While the map positions of the MSL and CSL are well-defined in the region between the Columbia River Plateau and Mesozoic Idaho Batholith in Idaho, the southern

extension of the lines has diverse locations because of incomplete coverage of reference data (Hughes and McCurry, 2002). The original map by Armstrong et al. (1977) has a narrow CSL-MSL transition extending southward about 46°N into the Owyhee Plateau and then splitting near 43°N, with the CSL continuing west-southwest to a widening CSL-MSL transition in northern Nevada (Figure 3). While the selection of samples is large throughout much of northern Oregon, there is sparse data between the 44.5°N to 42°N latitudes constraining the proposed contours. In contrast, Kistler and Peterman (1978) examined CSL and MSL strontium isotopic compositions of Mesozoic rocks in California, but their map varies slightly in that the separation of the CSL from the MSL is less gradual in southern Oregon and northern Nevada (Figure 5). Pierce and Morgan (1992) show a separation of the CSL-MSL lines at about 43°N, with the CSL extending southwestward and the MSL extending southeastward based on geologic evidence (Figure 6). The Santa Rosa-Calico volcanic field lies between the two isopleths in the southern region, deflecting the MSL eastward approximately 200 km to the Nevada-Utah-Idaho border (Pierce and Morgan, 1992).

Figure 5. Location of CSL and MSL in western United States, from Kistler and Peterman (1978).

Figure 6. CSL (*), MSL (x), and other major geologic features in the western United States associated with the late Cenozoic track of the Yellowstone hot spot (Pierce and Morgan, 1992).

Leeman et al. (1992) describe the MSL coinciding with the western Idaho suture zone (WISZ), which is at the contact between western terrane rocks and Idaho Batholith plutonic rocks (Figure 7). They suggest the regional isotopic signatures define a narrow accretionary N-S suture zone about 116.5°W to 119°W that occurred along a westward-

dipping decollement at lithospheric mantle depths during Rocky Mountain thrusting from 100-55 Ma (Leeman et al., 1992). The two overlapping sources that generated the transitional isotopic compositions in magmatism within the ~150-km-wide zone were accreted oceanic lithosphere and cratonic subcontinental mantle. Volcanic systems west of this zone were derived from accreted lithosphere with oceanic affinities while isotopic signatures east of this zone represent cratonic lower crust and subcontinental lithosphere (Leeman et al., 1992).

Figure 7. Map of northwestern United States showing the location of MSL and features associated with Cenozoic volcanism (left). Schematic lithospheric cross sections and $^{87}\text{Sr}/^{86}\text{Sr}$ isochrones from different events (right) (Leeman et al., 1992).

Fleck and Criss (2004) studied plutonic rocks and wall rock hosts in western Idaho and southeast Washington to establish a detailed characterization of isotopic variations across the terrane boundary between Precambrian crust and Late Paleozoic-Mesozoic island arc rocks. They delineate the WISZ by its juxtaposition of distinct rock

types and its degree of deformation, as well as geochemical and isotopic variations in plutons emplaced within and into the two different crustal types (Figure 8). The WISZ coincides with the change in Sr and Nd initial-ratios across a narrow 5- to 15-km-wide zone (Fleck and Criss, 2004). An updated map by Pierce and Morgan (2009) has the separation between the CSL and MSL occurring further north at around 44°N with much less of an eastward trace of the MSL around the Nevada-Oregon-Idaho border (Figure 9).

Figure 8. Map of the WISZ outlined in red. Red circles correspond to samples with Sr isotope data and black lines correspond to age traverses (Fleck and Criss, 2004).

Figure 9. Map showing location of the Snake River-Yellowstone hot spot track and associated volcanism (Pierce and Morgan, 2009). The locations of the CSL (.-) and MSL (...) lines are also shown.

Columbia River Basalt Province

Figure 10. Map of digital topography of eastern Oregon, northern Nevada, and southern Idaho depicting eruptive complexes associated with the Snake River-Yellowstone hot spot track (Shervais and Hanan, 2008). The locations of the $^{87}\text{Sr}/^{86}\text{Sr}_i = 0.7040$ (CSL) and 0.7060 (MSL) lines are displayed.

Eruptive activity of all rhyolitic centers in this study is related to the continental Large Igneous Province (LIP) of the Columbia River Basalt-Yellowstone hotspot system. Most discussions regarding the origin of volcanism of the Columbia River Province have centered around a deep-seated mantle plume (Camp, 2013). At ~17 Ma, a mantle plume incepted southeastern Oregon and generated Steens, Innaha, and Grande Ronde lavas. These lavas erupted from vents associated with the Steens, Chief Joseph, and Monument dike swarms located in eastern Oregon and Washington from 16.8 to 15.5 Ma (Barry et al., 2013). Silicic volcanism contemporaneous (~16.5 Ma to ~14.5 Ma) with this activity

is widespread in Oregon, Nevada, and Idaho as voluminous flood basalt magmas induced partial melting in the crust and/or fractionated to produce rhyolites (Christiansen and Yeats, 1992; Cummings et al., 2000; Brueseke et al., 2008; Shervais and Hanan, 2008). A plume head-to-tail transition occurred from 14-10 Ma, as the plume head was decapitated by thick cratonic boundary due to the southwest-directed North American plate motion. The plume tail was fixed in the relatively stationary mantle (Pierce and Morgan, 2009). This tail, which is currently located beneath Yellowstone, left a track of eruptive complexes in the direction opposite to the overriding plate movement along the Snake River Plain (Figure 10; Shervais and Hanan, 2008).

Silicic Volcanic Centers of this Study

In total, twenty-seven silicic units were examined in this study. Prior knowledge regarding each of these units varies widely from just being described in an unpublished geologic map to being well characterized including chemical analyses, age dates, and isotope data. Rhyolites of this study were collected over a wide area in eastern Oregon, 200 km in the east-west dimension and 150 km in the north-south dimension. These rhyolites reflect volcanism of the northern to central sector of the rhyolite flare up of the Columbia River Basalt province (Figure 10). All samples are Miocene in age with the majority between 16 to 14 Ma. Published age determinations detailed in Table 1 fall in the range of 9.48 Ma (Circle Bar Rhyolite) to 22.8 Ma (Dam Rhyolite, which has a more recently obtained age of 13.5 Ma).

Table 1. Published radiometric age determinations for silicic centers associated with this study. Some units have several age determinations. Only the four most recent age dates are shown.

Rock unit	Method	Age (Ma)	Reference
Beulah Reservoir Trd	-	-	-
Buchanan Dome Complex	K/Ar	15.1 ± 0.5	Fiebelkorn et al. (1983)
	K/Ar	13.9 ± 0.9	Evans and Brown (1981)
Circle Bar Rhyolite	K/Ar	9.48 ± 0.3	Fiebelkorn et al. (1983)
Dacite, Ironside Mountain	-	-	-
Dam Rhyolite	K/Ar	22.8 ± 2.6	Brown and Petros (1985)
	K/Ar	13.5 ± 3.4	Ferns and Cummings (1992)
Devine Canyon Tuff	⁴⁰ Ar/ ³⁹ Ar	9.63 ± 0.05	Ford (2011)
	⁴⁰ Ar/ ³⁹ Ar	9.74 ± 0.02	Jordan et al. (2004)
	⁴⁰ Ar/ ³⁹ Ar	9.68 ± 0.03	Streck et al. (1999)
Dinner Creek Ash-Flow Tuff	⁴⁰ Ar/ ³⁹ Ar	15.1 ± 2.0	Hooper et al. (2002)
	⁴⁰ Ar/ ³⁹ Ar	15.4 ± 0.6	Hooper et al. (2002)
	⁴⁰ Ar/ ³⁹ Ar	15.38 ± 0.17	Streck et al. (2011)
	⁴⁰ Ar/ ³⁹ Ar	15.9 ± 0.13	Streck et al. (2011)
Donnelly Butte	-	-	-
Dooley Mountain Rhyolite	K/Ar	14.7 ± 0.4	Evans (1992)
Jump Creek Rhyolite	K/Ar	10.6 ± 0.3	Rytuba et al. (1991)
	K/Ar	11.1 ± 0.2	Armstrong et al. (1980)
Littlefield Rhyolite	⁴⁰ Ar/ ³⁹ Ar	16.6 ± 0.1	Hooper et al. (2002)
	⁴⁰ Ar/ ³⁹ Ar	14.7 ± 0.96	Lees (1994)
	⁴⁰ Ar/ ³⁹ Ar	16.8 ± 0.4	Lees (1994)
	⁴⁰ Ar/ ³⁹ Ar	15.24 ± 0.31	Lees (1994)
McIntyre Ridge Rhyolite	K/Ar	15.8 ± 0.6	Ekren et al. (1984)
Rhyolite at Stockade Mountain	-	-	-
Rhyolite of Bully Creek	-	-	-
Rhyolite of Cottonwood Mountain	⁴⁰ Ar/ ³⁹ Ar	15.5 ± 0.7	Hooper et al. (2002)
	⁴⁰ Ar/ ³⁹ Ar	15.7 ± 0.2	Hooper et al. (2002)
	⁴⁰ Ar/ ³⁹ Ar	14.6 ± 1.0	Hooper et al. (2002)
Spring Creek Tuff	K/Ar	15.4	Rytuba and VanderMeulen (1991)
	⁴⁰ Ar/ ³⁹ Ar	15.95 ± 0.06	Benson et al. (2013)
	⁴⁰ Ar/ ³⁹ Ar	15.4 ± 0.2	Hooper et al. (2002)
	⁴⁰ Ar/ ³⁹ Ar	15.64 ± 0.09	Marcy et al. (2013)
Strawberry Mountain Volcanics	K/Ar	12.5 ± 0.4	Walker et al. (1974)
	K/Ar	12.2 ± 0.4	Walker et al. (1974)
Swamp Creek Rhyolite	K/Ar	17.3 ± 0.36	Fiebelkorn et al. (1983)
	K/Ar	15.7 ± 0.3	Fiebelkorn et al. (1983)
	K/Ar	16.1 ± 0.3	Fiebelkorn et al. (1983)
Three Fingers Intra-caldera Rhyolite	⁴⁰ Ar/ ³⁹ Ar	15.64 ± 0.08	Marcy et al. (2013)
Trd North of Drewsey	-	-	-
Trd South of Drewsey	-	-	-
Tuff, North of Unity	⁴⁰ Ar/ ³⁹ Ar	16.45 ± 0.14	Streck et al. (2011)
	K/Ar	19.6 ± 0.8	Fiebelkorn et al. (1983)
	K/Ar	19.5 ± 0.6	Fiebelkorn et al. (1983)

Many samples are rhyolites belonging to the Lake Owyhee Volcanic Field (LOVF) (Cummings et al., 2000; Ferns and McClaughry, 2013). These are nearly co-spatial with the presumed topographic uplift associated with the thermal anomaly of the Columbia River province (Pierce and Morgan, 2009), as well as with the convergence zone of CRBG dikes, aeromagnetic anomalies and other large-scale features.

^{87}Rb - ^{87}Sr Decay, the $^{87}\text{Sr}/^{86}\text{Sr}$ Isotope System, and the MSL and CSL

^{87}Rb naturally decays to ^{87}Sr by beta emission with a half-life of 4.88×10^{10} years (Faure and Mensing, 2005). Because rubidium substitutes for K^+ in potassium-bearing minerals, the principal carriers of rubidium are biotite, muscovite, lepidolite, orthoclase, and microcline (Faure and Powell, 1972). Igneous processes generally concentrate rubidium in the residual magma relative to potassium. Strontium, which substitutes for Ca^{2+} in calcium-bearing minerals, is mostly incorporated into plagioclase feldspar, apatite, clinopyroxene, and hornblende. The strontium concentration decreases with increasing degree of fractionation of the magma because it is removed from the liquid phase and is concentrated primarily in early-formed calcic plagioclase (Faure and Powell, 1972). The $^{87}\text{Sr}/^{86}\text{Sr}_i$, and how much ^{87}Rb decays to ^{87}Sr , depends on the Rb/Sr ratio of the sample and on the amount of time it has had to decay (Faure and Powell, 1972). The $^{87}\text{Sr}/^{86}\text{Sr}_i$ ratio in the crust increases more rapidly than the mantle because the Rb/Sr ratio of the continental crust is five to ten times greater than the Rb/Sr ratio in the upper mantle (McDougall, 1976). Magmas generated in the continental crust are expected to have a higher $^{87}\text{Sr}/^{86}\text{Sr}_i$ ratio than those generated in the asthenospheric mantle.

The MSL that separates cratonic crust to the east and the accreted lithosphere of oceanic affinity to the west displays this relationship with igneous rocks east of the line generally having higher ratios (Shervais and Hanan, 2008). The CSL that runs southwest from Idaho through Oregon was placed using basalts and intrusive rocks to determine its location. Data are lacking to establish a $^{87}\text{Sr}/^{86}\text{Sr}_i$ relationship using rhyolites. Furthermore, the issue of what $^{87}\text{Sr}/^{86}\text{Sr}_i$ of basalts versus rhyolites means has not been sufficiently addressed. If basalts do not interact with crust, they should record a mantle value. Yet, if partial melting of the crust produces the rhyolites, they record a crustal value. Often there is coherence between the results of basalt and rhyolite. This correlation may result from the mantle under accreted terrane being depleted with time giving rise to lower $^{87}\text{Sr}/^{86}\text{Sr}_i$ in the basalt or the mantle under the craton is enriched giving rise to higher $^{87}\text{Sr}/^{86}\text{Sr}_i$. A separate signal may be the rhyolites recording a depleted accreted terrane crustal melt because it has an overall lower $^{87}\text{Sr}/^{86}\text{Sr}_i$ ratio.

METHODS

Sample Collection

Samples were collected from outcrops in eastern Oregon using published geologic maps, field guides, and unpublished location information. Coordinates of sample locations were recorded by GPS. In total, thirty-three samples from twenty-two silicic units were collected for this study (Figure 11, Table 2). Detailed sample descriptions can be found in Appendix A. Sampled lithologies are ash-flow tuffs, rhyolite dome rocks, and silicic lava flows.

Figure 11. Location map for all samples collected for this study.

Table 2. Unit assignments and locations for each sample in this study. Latitude and longitude are listed in decimal degrees. The inferred source location for Dinner Creek Tuff is 44.04°N, 118.17°W and the inferred source location for Devine Canyon Tuff is 43.55°N, 118.78°W.

Sample ID	Unit ID	Unit	Latitude	Longitude
EJ-12-01	DNB	Donnelly Butte	43.8678	-119.5415
EJ-12-02	DNB	Donnelly Butte	43.8595	-119.5371
EJ-12-03	BDC	Buchanan Dome Complex	43.6453	-118.6258
EJ-12-04	BDC	Buchanan Dome Complex	43.5922	-118.6231
EJ-12-05A	CBR	Circle Bar Rhyolite	43.3389	-118.4225
EJ-12-05B	CBR	Circle Bar Rhyolite	43.3389	-118.4225
EJ-12-06A	SCR	Swamp Creek Rhyolite	43.3585	-118.2938
EJ-12-06B	SCR	Swamp Creek Rhyolite	43.3585	-118.2957
EJ-12-07	BER	Beulah Reservoir Trd	43.9074	-118.1704
EJ-12-08	DIT	Dinner Creek Tuff	43.7905	-118.1408
EJ-12-09	LIR	Littlefield Rhyolite (upper)	43.7817	-117.6906
EJ-12-10	LIR	Littlefield Rhyolite (upper)	43.7897	-117.7023
EJ-12-11	OMR	Old McIntyre Ridge Rhyolite	43.5177	-117.1515
EJ-12-12	OMR	Old McIntyre Ridge Rhyolite	43.5213	-117.1518
EJ-12-13A	JCR	Jump Creek Rhyolite	43.4178	-117.1162
EJ-12-13B	JCR	Jump Creek Rhyolite	43.4178	-117.1162
EJ-12-14	YMR	Young McIntyre Ridge Rhyolite	43.4708	-117.1242
EJ-12-15	DAM	Dam Rhyolite	43.6428	-117.2417
EJ-12-16	CMR	Rhyolite of Cottonwood Mountain	44.0786	-117.5433
EJ-12-17	RBC	Rhyolite of Bully Creek	43.9837	-117.5720
EJ-12-18	WET	Westfall Trv	43.9814	-117.6396
EJ-12-19	TSD	Trd South of Drewsey	43.7652	-118.3884
EJ-12-20	TND	Trd North of Drewsey	43.8067	-118.3923
EJ-12-21A	GRC	Griffen Creek Rhyolite	43.8562	-118.5354
EJ-12-21B	GRC	Griffen Creek Rhyolite	43.8562	-118.5354
EJ-12-22A	IRM	Dacite, NE Flank Ironside Mountain	44.3122	-118.1513
EJ-12-22B	IRM	Dacite, NE Flank Ironside Mountain	44.3122	-118.1513
EJ-12-23A	DMR	Dooley Mountain Rhyolite	44.5260	-117.8430
EJ-12-23B	DMR	Dooley Mountain Rhyolite	44.5260	-117.8430
CR-U3A	UNI	Rhyolite lava flow, North of Unity	44.5404	-118.2218
CR-U4B	UNI	Tuff, North of Unity	44.5397	-118.2065
MS-12-14b	UNI	Rhyolite lava flow, West of Unity	44.4666	-118.2330
CR-U35a	DIT	Dinner Creek Tuff	44.3567	-118.1595
MS-PCIT 1-A	DIT	Dinner Creek Tuff	44.4709	-118.7787
MS-12-29.1	DIT	Dinner Creek Tuff	44.2665	-117.6339
MS-11-20	DIT	Dinner Creek Tuff	44.6650	-118.3051
AS-SV-151	STR	Strawberry Mountain Rhyolite	44.0591	-118.7539
AS-SV-190	STR	Strawberry Mountain Rhyolite	44.2071	-118.7689
AS-SV-243	DCT	Devine Canyon Tuff	43.7978	-118.8447
MS-08-DR2	DMR	Dooley Mountain Rhyolite	44.5893	-117.8466
MS-LFR1	LIR	Littlefield Rhyolite (lower)	43.7816	-117.7298
MS-11-45	STK	Rhyolite at Stockade Mountain	43.2313	-117.9533
MS-11-15	SCT	Spring Creek Tuff	43.5120	-117.1235
TF-88A	TFI	Three Fingers Intra-caldera Rhyolite	43.4259	-117.1888
MS-13-12	BLB	Black Butte Rhyolite	43.7558	-118.2097

MS-13-09	MCB	McEwen Butte Rhyolite	43.4394	-118.1481
MS-13-10	VCR	Visher Creek Rhyolite	43.4367	-118.1455
MS-13-05	IDL	Rhyolite West of Idlewild	43.9267	-119.0739
AL-13-015b	BDC	Obsidian near Buchanan Dome Complex	43.6328	-118.6051

Analytical Methods

Thin Section and Petrographic Analyses

Petrographic characteristics of samples were determined using plane- and crossed-polarized transmitted light. Features from thin section analyses include mineral occurrences, sizes, and proportions, as well as mineral and rock textures (Appendix A).

Chemical Data

Major and trace element analyses were performed at the GeoAnalytical Lab at Washington State University using X-ray fluorescence (XRF) and inductively coupled plasma-mass spectrometry (ICP-MS). Detailed methods and data can be found in Appendix B.

Radiogenic Isotopes

To prepare for isotopic analyses, single crystals were etched in 10% HF to remove surface contaminants. Both crystals and whole rock samples were digested using a series of acids (HF, HCl, and HNO₃) over a three-day period. Cation and anion exchange chromatography was used to purify Sr and Pb from dissolved samples. Sr isotopic analyses were performed on whole rock and feldspar crystals at New Mexico State University (NMSU) using thermal ionization mass spectrometry (TIMS). For samples

with Sr concentrations less than 15 ppm as measured by ICP-MS, Sr and Rb concentrations for feldspars and groundmass were determined by isotope dilution using a high Rb/Sr spike. Pb isotopic analyses were performed on whole rock samples at University of Wyoming using multicollector inductively coupled plasma mass spectrometry (MC-ICP-MS). Detailed radiogenic isotopic methods can be found in Appendix C.

Stable Isotopes

Oxygen isotope analyses were performed on bulk samples of quartz, plagioclase, alkali feldspar, and glass at University of Oregon. Oxygen was separated from the crystals using laser fluorination and a 35W Newwave CO₂ IR laser. A Finnigan MAT 253, large radius 10kv gas source mass spectrometer was used for the final oxygen gas analysis. This method yielded particularly precise values with an error of better than $\pm 0.1\%$, 1 standard deviation. Results from the bulk crystal analyses can be used to calculate a bulk rock $\delta^{18}\text{O}$ value (Bindeman and Valley, 2002, 2003; Bindeman et al., 2004). Detailed oxygen isotopic methods and further calculations can be found in Appendix D.

Age Determinations

Feldspar phenocrysts or groundmass were selected from fresh samples for $^{40}\text{Ar}/^{39}\text{Ar}$ dating at the Oregon State University (OSU) $^{40}\text{Ar}/^{39}\text{Ar}$ Geochronology

Laboratory. Detailed $^{40}\text{Ar}/^{39}\text{Ar}$ incremental heating methods and age plateaus can be found in Appendix E.

RESULTS

Petrography

Rhyolitic ignimbrites and lava flows examined in thin section fall evenly into three categories of aphyric, phenocryst poor (<6%), and moderately porphyritic (7% - 25% phenocrysts). Textures include flow banded, vitroclastic, and hyalopilitic. Three samples from the Owyhee region exhibited concentric fractures indicating perlitic texture, which is common to rhyolites after the glass rehydrates. The most prevalent mineral phase among the samples was plagioclase, which ranges from 0-30% by volume. Also present to a lesser degree was sanidine, quartz, pyroxenes, Fe-Ti oxides, and sparse biotite and amphibole instead of pyroxene. A detailed petrographic examination of the samples is displayed in Table 7.

Whole Rock Chemistry

Of the total number of rocks collected, thirty-seven samples were analyzed for major and trace elements (Table 8). Weight percent SiO₂ ranges from 64% to 78% and total alkali contents range from 5.7% to 9.2%. With the exception of EJ-12-22B (Ironside Mountain dacite) and MS-11-20 (Dinner Creek Tuff), all samples are categorized as rhyolite using total alkali versus silica (Figure 12). MS-11-20 is a mingled bulk tuff sample of the Dinner Creek Tuff consisting of high-silica rhyolite and less silicic components. Rhyolites are compositionally diverse ranging from low to high-silica rhyolites (Figure 12). With the exception of AS-SV-243 (Devine Canyon Tuff), which is classified as peralkaline, samples are metaluminous to slightly peraluminous, indicating

alumina saturation (ie. $\text{Al}_2\text{O}_3/(\text{Na}_2\text{O}+\text{K}_2\text{O})$ and $\text{Al}_2\text{O}_3/(\text{CaO}+\text{Na}_2\text{O}+\text{K}_2\text{O})$ are around 1, Figure 13). Both calcium and iron follow the expected trend of a systematic depletion as silica increases (Figure 14 and Figure 15).

Figure 12. Total alkali-silica diagram for samples of this study (Le Bas, 1986). All but two samples are classified as rhyolite.

Figure 13. Molar A/NK versus A/CNK (Maniar and Piccoli, 1989) for samples; one sample falls in the peralkaline field while the remainder are in the metaluminous and the peraluminous fields.

Figure 14. Diagram illustrating CaO versus SiO₂. Overall calcium contents decline with increasing silica contents. Samples from the western and southern transect have less calcium than Littlefield Rhyolite and samples from the northern transect, likely because they tend to have a higher silica content.

Figure 15. Diagram illustrating FeO* versus SiO₂. Overall FeO* content declines with increasing silica, following the same trend as calcium (Figure 14).

According to the discrimination system of Whalen et al. (1987), the majority of the rhyolites (21) in this study show major and trace element chemistry similar to A-type granites and rhyolites, while 16 are similar to I-type granites and rhyolites (Figure 16). Although compositions of A-type rhyolites are diverse in regards to degree of

fractionation and contents of incompatible trace element concentrations, they can usually be distinguished from I-type rhyolites by higher Ba/Sr, lower La/Yb, and generally higher high field strength elements (HFSE) and heavy rare earth elements (HREE) abundances. The typical rock association for A-type rhyolites in a continental setting are bimodal provinces or flood basalt provinces while I-type rhyolites are typically associated with calc-alkaline volcanic suites in volcanic arc and intra-continental settings (Streck and Grunder, 2008).

Figure 16. $(K_2O + Na_2O)/CaO$ versus $Zr + Nb + Ce + Y$ diagram to differentiate A-type granites and rhyolites from other types of granites and rhyolites (after Whalen et al., 1987).

The europium anomaly describes the measured value of Eu compared to the expected value when interpolating between the neighboring REE of Sm and Gd.

Europium anomaly values (Eu/Eu^*) were calculated using ICP-MS data normalized to C1 chondrite composition (Sun and McDonough, 1989) for Eu, Sm, Gd:

$$\frac{Eu}{Eu^*} = \frac{Eu}{\sqrt{Sm \times Gd}} \quad \text{Eq. 1}$$

A negative europium anomaly is generated by removal of Eu enriched plagioclase where Eu^{2+} has substituted for Ca^{2+} . Larger negative europium anomalies are more common in the rhyolites from the southern area and the presumed younger units from the Owyhee region (Figure 17). The Littlefield Rhyolite, Dinner Creek Tuff, and presumed older units of the Owyhee retain smaller europium anomalies.

Figure 17. Diagram illustrating europium anomalies of rhyolites (calculated using Eq. 1).

The rhyolites of the study area show a variety of trends for incompatible trace elements in an average crust-normalized multi-element diagram using values of Rudnick and Gao (2003). Rhyolites were separated into six groups: northern transect, southern transect, central/western transect, Dinner Creek Tuff, Littlefield Rhyolite, and Owyhee region rhyolites. Northern, central/western transect, and southern transect have significant variations in trace element concentrations likely because they cover a wider region than the other three subsets or because the source rocks reflect different petrogenetic processes. The majority of rhyolites in the northern transect are enriched in Ba while the majority of rhyolites in the southern transect have Ba anomalies (Figure 18 and Figure

19). Trace elements of the Littlefield Rhyolite clearly separate two distinct magma types; EJ-12-10 has significant variations in numerous elements, including Ba, Nb, P, and Ti, relative to EJ-12-16, EJ-12-17, and MS-LFR1 (Figure 22). Rhyolite of Bully Creek has previously been recognized as retaining similar chemical characteristics as rhyolite of Cottonwood Mountain (Evans and Binger, 1999), but the lower unit of the Littlefield Rhyolite is chemically similar as well. Kittleman et al. (1967) mapped rhyolite of Bully Creek as part of their Littlefield Rhyolite but they described the Littlefield as stratigraphically above the Hunter Creek Basalt. Ferns et al. (1993) mapped the rhyolite of Cottonwood Mountain as part of the petrographically similar Littlefield Rhyolite.

Figure 18. Average crust-normalized multi-element diagram of rhyolites from the northern transect. EJ-12-22B (dacite of Ironside Mountain) has the greatest Nb anomaly.

Figure 19. Average crust-normalized multi-element diagram of rhyolites from the southern transect. There is greater variation in the trace element concentrations in the southern transect relative to the northern transect (Figure 18).

Figure 20. Average crust-normalized multi-element diagram of rhyolites from the western and central transect. MS-13-12 (Black Butte Rhyolite) creates the greatest variation in trace element concentrations among the samples from these transects.

Figure 21. Average crust-normalized multi-element diagram of Dinner Creek samples. MS-11-20, which is classified as a dacite, has the greatest variation in trace element concentrations among the Dinner Creek samples.

Figure 22. Average crust-normalized multi-element diagram of the Cottonwood Mountain, Bully Creek, and Littlefield Rhyolite. EJ-12-10 (upper Littlefield Rhyolite) has significant variation relative to the other rhyolites.

Figure 23. Average crust-normalized multi-element diagram of rhyolites from the Owyhee region. TF-88A (Three Fingers Intra-caldera Rhyolite) has the greatest Ba, Sr, and P anomaly.

Geochronology

Ar-Ar ages of select rhyolites from this study are displayed in

Table 3. New ages of some rhyolites indicate the initiation of rhyolitic volcanism in the study region was essentially coeval with rhyolitic volcanism in the McDermitt area. Particularly old ages were obtained for EJ-12-10 and EJ-12-17 using the MAP instrument. These samples were reanalyzed using the Argus instrument and the age spectrums produced progressively younger ages (EJ-12-10) and older ages (EJ-12-17) at higher temperatures (Figure 32). This pattern of age varying with incremental heating steps is a result of ^{39}Ar recoil.

The dates obtained by Marcy et al. (2013) for Spring Creek Tuff (SCT) and intra-caldera Three Fingers rhyolites (TFI) of LOVF show little time elapsed between the eruptions of the outflow tuff and intra-caldera rhyolite of the Owyhee region.

Table 3. Ar-Ar age data for select rhyolites. MS-11-15 and TF-88A ages are from Marcy et al. (2013). AS-SV-151 and AS-SV-190 ages are from Steiner per. comm., while all other data were obtained in this study. Ages were calculated with FCT age of 28.201 Ma (Kuiper et al., 2008).

Sample ID	Unit ID	Unit	Age (Ma)	Material	Instrument
EJ-12-03	BDC	Buchanan Dome Complex	16.13 ± 0.11	sanidine	MAP
EJ-12-06A	SCR	Swamp Creek Rhyolite	16.01 ± 0.08	glass	MAP
EJ-12-07	BER	Beulah Reservoir Trd	16.29 ± 0.21	groundmass	MAP
EJ-12-10	LIR	Littlefield Rhyolite (upper)	16.75 ± 0.45	groundmass	MAP
EJ-12-12	OMR	Old McIntyre Ridge Rhyolite	15.94 ± 0.16	sanidine	MAP
EJ-12-17	RBC	Rhyolite of Bully Creek	17.01 ± 0.60	plagioclase	MAP
EJ-12-23A	DMR	Dooley Mountain Rhyolite	15.59 ± 0.04	plagioclase	ARGUS
MS-11-20	DIT	Dinner Creek Tuff	14.84 ± 0.11	anorthoclase	MAP
AS-SV-151	STR	Strawberry Mountain Rhyolite	16.16 ± 0.17	groundmass	MAP
AS-SV-190	STR	Strawberry Mountain Rhyolite	14.79 ± 0.12	groundmass	MAP
MS-11-15	SCT	Spring Creek Tuff	15.74 ± 0.09	sanidine	MAP
TF-88A	TFI	Three Fingers Intra-caldera Rhyolite	15.74 ± 0.08	sanidine	MAP

Isotope Analysis

Oxygen Isotopes

Table 4 contains oxygen isotope results acquired for this study. Complete oxygen isotopic results can be found in Appendix D. Values are highly variable between different silicic centers with $\delta^{18}\text{O}$ ranging from 1.6 to 9.9‰. After applying a correction to determine the approximate bulk rock $\delta^{18}\text{O}$ values (Eqns. D-1, Bindeman et al., 2004; D-2 Bindeman and Valley, 2002; D-3, Bindeman and Valley, 2003), the calculated magmatic values ranged from 2.1 to 10.5‰. Rhyolites from Dooley Mountain in the northern region have the highest $\delta^{18}\text{O}$ values while the Littlefield/Cottonwood Mountain rhyolites have the lowest $\delta^{18}\text{O}$ values.

Table 4. Oxygen isotopic compositions of samples acquired from this study. The magmatic $\delta^{18}\text{O}$ values are calculated using equations D-1, D-2, and D-3 (Appendix D), depending on the mineral analyzed.

Sample		Material	$\delta^{18}\text{O}$ (‰)	Calculated $\delta^{18}\text{O}$ magma (‰)
Owyhee locality				
OMR	EJ-12-12	sanidine	4.001	4.291
JCR	EJ-12-13B	sanidine	8.399	8.689
YMR	EJ-12-14	plagioclase	5.944	6.576
DAM	EJ-12-15	plagioclase	6.674	7.189
SCT	MS-11-15	sanidine	3.959	4.249
TFI	TF-88A	sanidine	5.705	5.995
Littlefield/Cottonwood Mountain				
LIR	EJ-12-10	whole rock	3.881	3.881
LIR	EJ-12-10	plagioclase	2.436	2.905
CMR	EJ-12-16	plagioclase	2.350	2.872
RBC	EJ-12-17	plagioclase	1.635	2.139
LIR	MS-LFR1	plagioclase	1.806	2.308
Dinner Creek				
DIT	EJ-12-08	plagioclase	6.849	7.503
DIT	CR-U35a	plagioclase	6.430	6.981
DIT	MS-PCIT 1-A	sanidine	7.489	7.779
DIT	MS-12-29.1	plagioclase	6.990	7.600
DIT	MS-11-20	plagioclase	6.339	6.724
Northern locality				
UNI	CR-U3A	plagioclase	7.226	7.830
UNI	CR-U4B	quartz	7.230	6.780
UNI	MS-12-14b	plagioclase	7.754	8.353
DMR	EJ-12-23A	plagioclase	9.529	10.150
DMR	EJ-12-23B	-	-	-
DMR	MS-08-DR2	plagioclase	9.905	10.450
IRM	EJ-12-22B	whole rock	8.078	8.078
IRM	EJ-12-22B	plagioclase	5.836	6.132
Strawberry Mountain				
STR	AS-SV-151	plagioclase	7.336	7.862
STR	AS-SV-190	glass	6.702	6.702
Drewsey locality				
TND	EJ-12-20	-	-	-
TSD	EJ-12-19	-	-	-
BER	EJ-12-07	-	-	-
BLB	MS-13-12	plagioclase	6.339	6.968
Southern locality				
CBR	EJ-12-05B	quartz	8.431	7.981
SCR	EJ-12-06A	glass	6.685	6.685
STK	MS-11-45	sanidine	6.487	6.777
MCB	MS-13-09	sanidine	6.010	6.300
VCR	MS-13-10	plagioclase	6.603	7.228

DCT	AS-SV-243	sanidine	5.097	5.387
Western locality				
DNB	EJ-12-01	-	-	-
IDL	MS-13-05	-	-	-
Buchanan				
BDC	EJ-12-03	sanidine	5.303	5.593
BDC	AL-13-015b	glass	6.396	6.396

Radiogenic Isotopes

Sr and Pb isotopic compositions are reported in Table 5 and Table 6, respectively.

Complete isotopic results can be found in Appendix C. Sr isotopic compositions are variable between the different silicic centers. For the majority of rhyolites with feldspar and groundmass data, feldspar phenocrysts are in Sr isotopic equilibrium with groundmass. This is expected given the low proportion of feldspar phenocrysts and petrographic evidence (e.g., euhedral, no significant resorption boundaries) that feldspars crystallized in the host magma.

Table 5. Sr isotope data for feldspar (fsp) and groundmass (GM). $^{87}\text{Sr}/^{86}\text{Sr}$ (initial) calculated using equation C-1 (Appendix C). Concentrations of Sr and Rb are available in Table 9 (Appendix C).

Sample		$^{87}\text{Sr}/^{86}\text{Sr}$ (present)		2σ error		$^{87}\text{Sr}/^{86}\text{Sr}$ (initial)	
		fsp	GM	fsp	GM	fsp	GM
Owyhee locality							
OMR	EJ-12-12	0.704990	0.707632	0.000022	0.000015	0.704644	0.704865
JCR	EJ-12-13B	0.705682	—	0.000010	—	0.705666	—
YMR	EJ-12-14	0.704575	0.708181	0.000015	0.000015	0.704128	0.704603
DAM	EJ-12-15	—	0.704327	—	0.000011	—	0.704144
SCT	MS-11-15	0.704970	0.708270	0.000011	0.000011	0.704606	0.705361
TFI	TF-88A	0.707846	0.731446	0.000015	0.000011	0.704673	0.701976
Littlefield/Cottonwood Mountain							
LIR	EJ-12-10	—	0.705637	—	0.000013	—	0.705142
CMR	EJ-12-16	—	0.707205	—	0.000010	—	0.706712
RBC	EJ-12-17	0.706668	—	0.000022	—	0.706595	—
LIR	MS-LFR1	0.706647	0.707301	0.000021	0.000015	0.706646	0.706677
Dinner Creek							
DIT	EJ-12-08	—	0.705559	—	0.000014	—	0.704551
DIT	CR-U35a	0.704216	0.704921	0.000008	0.000013	0.704207	0.703890
DIT	MS-PCIT 1-A	0.704414	0.705053	0.000010	0.000032	0.704315	0.704261
DIT	MS-12-29.1	0.704164	0.706008	0.000049	0.000015	0.704159	0.704463

DIT	MS-11-20	0.704170	—	0.000014	—	0.704154	—
Northern locality							
UNI	CR-U3A	0.703836	0.704921	0.000011	0.000013	0.703699	0.703828
UNI	CR-U4B	0.703820	0.704271	0.000013	0.000014	0.703819	0.704067
UNI	MS-12-14b	0.703730	—	0.000008	—	0.703730	—
DMR	EJ-12-23A	0.704444	0.705171	0.000010	0.000015	0.704443	0.704575
DMR	EJ-12-23B	—	0.706697	—	0.000013	—	0.704643
DMR	MS-08-DR2	0.704814	0.705088	0.000014	0.000008	0.704783	0.704841
IRM	EJ-12-22B	—	0.703772	—	0.000014	—	0.703727
Strawberry Mountain							
STR	AS-SV-151	—	0.704294	—	0.000013	—	0.704054
STR	AS-SV-190	—	0.704994	—	0.000011	—	0.704303
Drewsey locality							
TND	EJ-12-20	—	0.705807	—	0.000014	—	0.703303
TSD	EJ-12-19	—	0.705554	—	0.000013	—	0.703898
BER	EJ-12-07	—	0.704145	—	0.000015	—	0.703937
BLB	MS-13-12	0.710698	0.714644	0.000046	0.000011	0.703650	0.700718
Southern locality							
CBR	EJ-12-05B	0.704874	—	0.000014	—	0.704800	—
SCR	EJ-12-06A	—	0.704548	—	0.000011	—	0.704078
STK	MS-11-45	0.706244	0.730827	0.000013	0.000014	0.704897	0.704631
MCB	MS-13-09	0.704780	0.706235	0.000010	0.000031	0.704408	0.704128
VCR	MS-13-10	0.706179	0.779998	0.000024	0.000011	0.704912	0.683238
DCT	AS-SV-243	0.705758	—	0.000020	—	0.704958	—
Western locality							
DNB	EJ-12-01	—	0.704237	—	0.000011	—	0.703878
IDL	MS-13-05	—	0.708009	—	0.000011	—	0.703430
Buchanan							
BDC	EJ-12-03	0.709250	0.722852	0.000020	0.000021	0.704851	0.698936
BDC	AL-13-015b	—	0.704802	—	0.000011	—	0.704060

Table 6. Pb isotopic compositions of groundmass and whole rock.

Sample		$^{206}\text{Pb}/^{204}\text{Pb}$	2 σ error	$^{207}\text{Pb}/^{204}\text{Pb}$	2 σ error	$^{208}\text{Pb}/^{204}\text{Pb}$	2 σ error
Owyhee locality							
OMR	EJ-12-12	18.888	0.001	15.610	0.001	38.757	0.002
JCR	EJ-12-13B	18.911	0.001	15.608	0.001	38.673	0.002
YMR	EJ-12-14	-	-	-	-	-	-
DAM	EJ-12-15	-	-	-	-	-	-
SCT	MS-11-15	18.885	0.001	15.607	0.001	38.749	0.002
TFI	TF-88A	18.879	0.001	15.607	0.001	38.723	0.002
Littlefield/Cottonwood Mountain							
LIR	EJ-12-10	-	-	-	-	-	-
CMR	EJ-12-16	-	-	-	-	-	-
RBC	EJ-12-17	-	-	-	-	-	-
LIR	MS-LFR1	18.820	0.001	15.610	0.001	38.800	0.002
Dinner Creek							
DIT	EJ-12-08	-	-	-	-	-	-

DIT	CR-U35a	-	-	-	-	-	-
DIT	MS-PCIT 1-A	18.913	0.001	16.000	0.001	38.493	0.002
DIT	MS-12-29.1	-	-	-	-	-	-
DIT	MS-11-20	-	-	-	-	-	-
Northern locality							
UNI	CR-U3A	18.866	0.001	15.588	0.001	38.495	0.002
UNI	CR-U4B	-	-	-	-	-	-
UNI	MS-12-14b	-	-	-	-	-	-
DMR	EJ-12-23A	18.935	0.001	15.605	0.001	38.550	0.002
DMR	EJ-12-23B	18.934	0.001	15.605	0.001	38.532	0.002
DMR	MS-08-DR2	-	-	-	-	-	-
IRM	EJ-12-22B	-	-	-	-	-	-
Strawberry Mountain							
STR	AS-SV-151	18.868	0.001	15.596	0.001	38.484	0.003
STR	AS-SV-190	18.900	0.001	15.595	0.001	38.506	0.002
Drewsey locality							
TND	EJ-12-20	-	-	-	-	-	-
TSD	EJ-12-19	-	-	-	-	-	-
BER	EJ-12-07	-	-	-	-	-	-
BLB	MS-13-12	-	-	-	-	-	-
Southern locality							
CBR	EJ-12-05B	-	-	-	-	-	-
SCR	EJ-12-06A	-	-	-	-	-	-
STK	MS-11-45	18.799	0.001	15.607	0.001	38.748	0.003
MCB	MS-13-09	-	-	-	-	-	-
VCR	MS-13-10	-	-	-	-	-	-
DCT	AS-SV-243	-	-	-	-	-	-
Western locality							
DNB	EJ-12-01	-	-	-	-	-	-
IDL	MS-13-05	-	-	-	-	-	-
Buchanan							
BDC	EJ-12-03	18.991	0.001	15.603	0.001	38.575	0.002
BDC	AL-13-015b	-	-	-	-	-	-

Overall, the data indicate a significant range in Pb isotopic data, but data of individual regions tend to plot close to one another in comparison to the entire data field (Figures 24, 25, and 26). $^{208}\text{Pb}/^{204}\text{Pb}$ - $^{206}\text{Pb}/^{204}\text{Pb}$ ratios of rhyolites define two trends that converge at high $^{206}\text{Pb}/^{204}\text{Pb}$. Towards lower $^{206}\text{Pb}/^{204}\text{Pb}$ values, one trend develops progressively higher $^{208}\text{Pb}/^{204}\text{Pb}$ values (i.e. originally high Th/U values) while the other trend develops lower $^{208}\text{Pb}/^{204}\text{Pb}$ values (i.e. originally slightly lower Th/U values) compared to $^{208}\text{Pb}/^{204}\text{Pb}$ values with higher $^{206}\text{Pb}/^{204}\text{Pb}$ values.

Figure 24. Variation diagram illustrating $^{87}\text{Sr}/^{86}\text{Sr}_i$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ signatures. $^{87}\text{Sr}/^{86}\text{Sr}_i$ data are from feldspars; if there are no data for feldspar of a particular sample, the $^{87}\text{Sr}/^{86}\text{Sr}_i$ data are from groundmass.

Figure 25. $^{208}\text{Pb}/^{204}\text{Pb}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ variation diagram illustrating rhyolite groundmass signatures.

Figure 27. $^{87}\text{Sr}/^{86}\text{Sr}_i$ versus calculated $\delta^{18}\text{O}$ magma (‰) variation diagram illustrating significant range in $\delta^{18}\text{O}$. $^{87}\text{Sr}/^{86}\text{Sr}_i$ data are from feldspars; if there are no data for feldspar of a particular sample, the $^{87}\text{Sr}/^{86}\text{Sr}_i$ data are from groundmass.

Except for rare examples in the western part of the study area, dispersion of $^{87}\text{Sr}/^{86}\text{Sr}_i$ ratios among both silicic and basaltic rocks occurs eastward of 118.6°W (cf. Ford et al., 2013). For example, rhyolites in the Owyhee region between 117.10°W and 117.25°W have $^{87}\text{Sr}/^{86}\text{Sr}_i$ ratios that range widely from 0.7041 to 0.7057. Highway 20, which runs predominately in an E-W direction in the study region, is the approximate boundary separating higher $^{87}\text{Sr}/^{86}\text{Sr}_i$ rhyolites to the south from lower $^{87}\text{Sr}/^{86}\text{Sr}_i$ rhyolites to the north. Further east of 118°W , this transition trends more NE-SW and incorporates the higher $^{87}\text{Sr}/^{86}\text{Sr}_i$ units of Littlefield Rhyolite and rhyolite of Cottonwood Mountain (Figure 29). In general, higher $^{87}\text{Sr}/^{86}\text{Sr}_i$ ratios (>0.7040) correspond to samples that are A-type rhyolites (Figure 28).

Figure 28. Diagram illustrating Zr+Nb+Ce+Y (ppm) of whole rocks and corresponding $^{87}\text{Sr}/^{86}\text{Sr}_i$. $^{87}\text{Sr}/^{86}\text{Sr}_i$ data are from feldspars; if there are no data for feldspar of a particular sample, the $^{87}\text{Sr}/^{86}\text{Sr}_i$ data are from groundmass. The majority of samples are classified as A-type rhyolites, after Whalen et al. (1987), and A-type rhyolites are in general more radiogenic than others. Samples are coded for whole rock SiO₂ (wt.%) content; black is dacite (<70%), grey is low-silica rhyolite (<75%), unfilled is high-silica rhyolite (>75%).

Figure 29. Map illustrating sample locations with age corrected $^{87}\text{Sr}/^{86}\text{Sr}_i$ ratios from feldspar or groundmass. Shading of icon is defined as: left part of icon reflects $^{87}\text{Sr}/^{86}\text{Sr}_i$ ratio of feldspar while the right part of icon reflects $^{87}\text{Sr}/^{86}\text{Sr}_i$ ratio of groundmass. The traditional $^{87}\text{Sr}/^{86}\text{Sr}_i = 0.7040$ and 0.7060 isochrones are represented by dashed black lines are after Pierce and Morgan (2009). Our proposed 0.7040 (orange) and 0.7060 (blue) $^{87}\text{Sr}/^{86}\text{Sr}_i$ isochrones are dashed. Literature data in the region are included as abbreviations: BB- Burns Butte, IM- Iron Mt, PB- Palomino Butte, DB- Duck Creek Butte, IB- Indian Creek Butte (Streck and Grunder, 2008; Ford, 2011). Devine Canyon Tuff and Dinner Creek Tuff symbols correspond to inferred source locations for these tuffs, although samples were collected from different locations (Table 2).

DISCUSSION

This study presents new Sr, O, Pb isotope data and $^{40}\text{Ar}/^{39}\text{Ar}$ ages of mostly rhyolitic, mid-Miocene volcanic centers in eastern Oregon. These silicic centers, of which many erupted during regional flood basalt activity, represent crustal magmatism west of the Proterozoic cratonic lithosphere.

Sr isotopic data presented here are in large part inconsistent with previously determined locations for the MSL (0.7040) and CSL (0.7060) in eastern Oregon. Silicic centers over 100 km west of the proposed boundaries retain Sr isotope signatures significantly greater than 0.7040, and in a few cases, even greater than 0.7060. Discrepancies with these long established isotopic boundaries call into question how sharp the crustal transitions are or to what extent more radiogenic crustal domains exist west of the cratonic boundary. Leeman et al. (1992) describe these transitional zones as arising from a westward-dipping decollement at lithospheric mantle depths. This resulted in a ~150-km-wide transition zone (119°W – 116.5°W) of mixed accreted oceanic lithosphere and cratonic subcontinental mantle, whereby the cratonic mantle portion progressively decreases westward, which raises a question regarding the likely causes for observed high Sr isotope ratios (>0.7060) in rhyolites from vents located between ~119°W to ~117°W and ~43°N to ~44.5°N.

Position of Terrane/Craton Boundary

The existence of crustal sources retaining Sr isotopic ratios >0.7060 west of the long accepted MSL could signify the following: 1) the position of the 0.7060 line, and

therefore the position of the lithospheric boundary between accreted terranes to the west and cratonic crust to the east on the latitude between 43°N and 45°N, is not where the MSL is currently positioned but rather should be moved westward, or 2) alternatively, crustally derived igneous rocks with Sr isotopic ratios >0.7060 are not as faithfully delineating the position of the accreted terrane/craton boundary as previously thought. In this regard, crustal rocks with $^{87}\text{Sr}/^{86}\text{Sr}_i > 0.7040$ may have even less to do with a transition from crust composed of accreted terranes to crust of North American craton. Geophysical evidence allows evaluation of these two possibilities. An increase in Moho depth occurs at the longitude coinciding with the CSL and MSL in the north and with the CSL in the south, except where there is a shallow Moho embayment trending eastward at $\sim 44^\circ\text{N}$ (Eagar et al., 2011; Gilbert, 2012). Additional geophysical evidence indicates that the MSL does coincide with the lithospheric boundary based on resistivity models (Bedrosian and Feucht, 2013) and that more radiogenic crustal sections exist west of the MSL. This indicates that the location of the MSL does coincide with the transition of cratonic crust to the east and accreted terrane crust to the west.

Higher Sr ratios west of the MSL at latitudes above 43.5°N cannot be explained by a model incorporating a sliver of cratonic lithosphere underlying accreted terrane crust in eastern Oregon (Leeman et al., 1992; Evans et al., 2002). This is only applicable to the Owyhee plateau in southeastern Oregon where the CSL and MSL are separated. Further north, a shallow Moho at ~ 30 km is also inconsistent with accreted terrane crust overlying craton that together would be in excess of 30 km (cf. Evans et al., 2002). The highest Sr ratios in the Littlefield/Cottonwood Mountain area are mirrored by Sr ratios

around 0.704 on either side, requiring an indenter rather than a cratonic keel at 44°N (Figure 29). Lastly, results of this study are too geographically scattered to conform to a gradual E-W transition of $^{87}\text{Sr}/^{86}\text{Sr}_i$ (Figure 29) and a mixed lithospheric mantle could not serve to explain higher Sr isotopic signatures in rhyolites that largely originated from crustal sources.

Variable Accreted Terrane Crust

Isotopic signatures of accreted terranes in eastern Oregon vary considerably depending on rock type and origin. The Olds Ferry terrane has $^{87}\text{Sr}/^{86}\text{Sr}_i$ ratios less radiogenic than metaplutonic rocks but more radiogenic than siliceous argillites of the Baker terrane. Accreted terrane outcrops only occur in the northern part of our study area (Figure 1), but lithologies of xenoliths found in volcanic rocks along the southern part indicate variable meta-sedimentary and igneous lithologies (Evans et al., 2002; Streck, 2002; J. Edwards, pers. comm. 2013) that suggest accreted terranes in the southern part were also composed of variable lithologies.

The northernmost primary terrane is the Wallowa terrane, although none of the units in this study are erupted through this terrane. Schwartz et al. (2010) propose the boundary between the Wallowa island-arc terrane and the Baker accretionary-complex terrane is an example of a broad zone of imbrication composed of slabs and slices of arc crust tectonically mixed within an accretionary complex. The northern margin of the Baker terrane includes the Bourne subterrane with metaplutonic rocks, Elkhorn Ridge Argillite, and Cougar Basin metasedimentary sequence. The metaplutonic rocks have low

$^{87}\text{Sr}/^{86}\text{Sr}_i$ values (0.7033 to 0.7034), the siliceous argillites from the chert-argillite mélange have $^{87}\text{Sr}/^{86}\text{Sr}_i$ values between 0.7073 and 0.7094, while the metasedimentary sequence has $^{87}\text{Sr}/^{86}\text{Sr}_i$ values between 0.7039 and 0.7057 (Schwartz et al., 2010). Elevated $^{87}\text{Sr}/^{86}\text{Sr}_i$ values of the Elkhorn Ridge Argillite indicate that it contains a component of detritus from Proterozoic or older continental crust in addition to volcanogenic material (Schwartz et al., 2010). Among the rhyolites analyzed in this study, both Dooley Mountain and Unity rhyolites have eruption locations within the area delineated as Baker terrane by LaMaskin et al. (2009). $^{87}\text{Sr}/^{86}\text{Sr}_i$ ratios for these units averaged 0.7046 and 0.7038, respectively.

The southernmost terrane that lies below many of the sample locations is the Olds Ferry terrane. The Huntington Formation of the Olds Ferry terrane is split into an upper and lower member. Volcanic rocks of the upper Huntington are richer in silica and have elevated $^{87}\text{Sr}/^{86}\text{Sr}_i$ ratios (0.7036 to 0.7057) relative to samples from the lower Huntington Formation (Tumpane, 2010). Two samples from the lower Huntington Formation had $^{87}\text{Sr}/^{86}\text{Sr}_i$ values of 0.7037 and 0.7034 (Tumpane, 2010). More radiogenic $^{87}\text{Sr}/^{86}\text{Sr}_i$ compositions of the upper Huntington and Weatherby Formation likely reflect the increasing influence of continental crust incorporation or lithospheric mantle transitions to more radiogenic values on volcanic rocks through time (Tumpane, 2010).

Figure 30. $^{87}\text{Sr}/^{86}\text{Sr}_i$ - ϵNd diagram showing fields for basement rocks of the Olds Ferry and Wallowa terranes (from Tumpene, 2010). Both sets of basement rocks have $^{87}\text{Sr}/^{86}\text{Sr}_i$ and ϵNd values indicative of an island arc setting and little evidence for elevated $^{87}\text{Sr}/^{86}\text{Sr}_i$ ratios.

The Izee terrane is marine sedimentary rocks that rest in depositional contact on older rocks and structures of the Baker, Wallowa, and Olds Ferry terranes (Dorsey and LaMaskin, 2007). A marine strata section ~12-15 km thick preserves the sequences of rocks deposited in a single large marine basin to form the Izee terrane (Dorsey and LaMaskin, 2007). Sedimentation to the Izee basin was derived in part by material shedding from the craton, which should imprint a $^{87}\text{Sr}/^{86}\text{Sr}_i$ signal of greater than 0.706.

Rather than a progressive decrease of $^{87}\text{Sr}/^{86}\text{Sr}_i$ away from the craton boundary, Sr isotopic ratios show no consistent trend and placing Sr contours in these regions indicates complicated curved paths and a patchy distribution of similar Sr isotopic values (Figure 29). These trends are better explained by the orientation of accreted terrane boundaries

and expected local complications in the distribution of the terrane rocks with high vs. low Sr isotopic values rather than by the progressive thinning of cratonic crust westward.

As a result of this distribution of isotopic values and the highly variable isotopic nature of the accreted terranes, the crustal composition of a specific terrane in a particular area is most likely the primary control on observed $^{87}\text{Sr}/^{86}\text{Sr}_i$. One consequence of this is that not all fractionated basalts with elevated radiogenic ratios would have had to inherit these signatures from cratonic crust. In contrast, some may have inherited such signatures from specific radiogenic accreted terranes or parts of terranes. Based in part on high $^{87}\text{Sr}/^{86}\text{Sr}_i$ of Grande Ronde lavas (0.7041 - 0.7057), Wolff et al. (2008) regarded these lavas as essentially Imnaha basaltic magmas ($^{87}\text{Sr}/^{86}\text{Sr}_i = 0.7038 - 0.7044$) that were more extensively contaminated by cratonic crust. It was also proposed that both Grande Ronde and Imnaha Basalts were derived from a single central crustal magma system that was centered in the area where the western Snake River Plain, Oregon-Idaho graben, and Chief Joseph dike swarm converge (Wolff et al., 2008; Wolff and Ramos, 2013). This area coincides with rhyolites with the highest $^{87}\text{Sr}/^{86}\text{Sr}_i$ as well as surrounding centers (Figure 29). Given this spatial overlap, it is plausible that some of isotopic change observed in Grande Ronde Basalt relative to Imnaha Basalt could possibly also result from the incorporation of highly radiogenic, accreted terrane crust during storage in magma chambers located within specific regions of highly radiogenic accreted terranes.

Variations Induced by Variable Amounts of Basalt Ancestry

Not all rhyolites may be pure crustal melts and the possibility exists that basaltic magma contributed, at least in part, to their compositions. Rhyolites from the Snake River Plain have radiogenic isotopic ratios that lie between those of surrounding crust and of regional basaltic magma (Christiansen and McCurry, 2008; Leeman et al., 2008). This has been interpreted as evidence that these rhyolites obtained isotopic signatures by combined processes such as Assimilation-Fractionation Crystallization (AFC) or other mixing processes. For rhyolites in this study, Sr isotopic ratios could hypothetically become either more or less radiogenic by mixing with mafic magma. This results from the largest pool of mafic magma that is in fact contemporaneous with 16-15 Ma rhyolites is the basaltic-andesitic magmas of the Grande Ronde Basalt, which have isotopic Sr ratios ranging between 0.7041 - 0.7057 and are thought to have been stored at crustal levels over most of the area of this study (Wolff et al., 2008; Wolff and Ramos, 2013; Streck et al., submitted). If mafic magmas with Sr isotopic ratios of >0.7040 interacted with less radiogenic accreted crust (0.7036), the resulting Sr isotopic ratio of the mixture would be higher than a pure crustal melt (~ 0.7036). One possible example of this is the Dinner Creek Tuff. The five Dinner Creek Tuff samples analyzed in this study have an average $^{87}\text{Sr}/^{86}\text{Sr}_i$ of 0.7043 ± 0.0002 , but the likely source area is surrounded by rhyolites with Sr isotopic ratios of 0.7033-0.7039. Additional evidence that this occurred is found in co-eruptive mafic components that are interpreted as *de facto* Grand Ronde Basalt (Streck et al., submitted).

An alternative scenario in lowering Sr isotope ratios of rhyolites could result from interaction of mafic magmas such as Imnaha ($^{87}\text{Sr}/^{86}\text{Sr}_i = 0.7038 - 0.7044$) or Steens Mountain ($^{87}\text{Sr}/^{86}\text{Sr}_i = 0.7033 - 0.7040$) type basalts (cf. Wolff and Ramos, 2013) with more radiogenic crust like the one that is assumed to be tapped for the Littlefield/Cottonwood Mountain rhyolites (i.e. with $^{87}\text{Sr}/^{86}\text{Sr}_i > 0.706$). We speculate that variable amounts of a “basalt component” in rhyolites play a secondary role in contrasting specific Sr isotopic ratios of compositionally variable accreted terrane crust.

Lees (1994) modeled the Dinner Creek Tuff and Upper and Lower Members of Littlefield Rhyolite to determine whether regional basalts are genetically related to the rhyolites. The Malheur Group basalt fractionation model effectively reproduces the composition of the Upper Littlefield Rhyolite by fractional crystallization of clinopyroxene, plagioclase, and ilmenite (Lees, 1994). Beginning with the Hunter Creek Basalt, approximately 50% fractionation is required to obtain the Upper Littlefield Rhyolite composition for most trace elements (Zr, Rb, Ba, Nb, Sr, Y and TiO_2), corresponding to about a 10% rise in silica (Lees, 1994). To achieve an additional 8% increase in silica and reach the SiO_2 composition of the rhyolite, about 70% of fractional crystallization is required. In addition, the similarity of $^{87}\text{Sr}/^{86}\text{Sr}_i$ isotope ratios of the Upper Littlefield Rhyolite ($^{87}\text{Sr}/^{86}\text{Sr}_i = 0.7056$) and the more contaminated Hunter Creek Basalt flows ($^{87}\text{Sr}/^{86}\text{Sr}_i = 0.7044$ to 0.7056) confirm the idea that the Upper Littlefield Rhyolite is related to the Malheur Gorge basalts by fractionation (Lees, 1994). Yet, the Dinner Creek Tuff and the Lower Littlefield Rhyolite were not able to be successfully modeled using crystal fractionation of Malheur Group basalts and therefore are not

genetically related to the Malheur Group basalts. Therefore, they may have an origin from crustal melting (Lees, 1994). The high $^{87}\text{Sr}/^{86}\text{Sr}_i = 0.7069$ of the Lower Littlefield Rhyolite also suggests the Lower Littlefield Rhyolite Member has a different origin than that of the Upper Littlefield Rhyolite Member, such as crustal melting (Lees, 1994).

Lead Isotopes

Because the accreted terranes that exist west of the MSL were extracted from the mantle during Paleozoic time, they are too young to have developed significant Pb isotope contrasts with present-day mantle sources (Wolff and Ramos, 2013). While Pb isotope data is incomplete for all samples of this study, the available data indicate a significant range in Pb isotopes. Yet, data of individual regions tend to plot close to one another in comparison to entire data spread (Figure 25).

Figure 31. Sr-Pb isotope diagrams illustrating relations among CRBG lavas and Pacific mid-ocean-ridge basalts (after Wolff et al., 2008) with rhyolites from this study. (a) $^{207}\text{Pb}/^{206}\text{Pb}$ versus $^{208}\text{Pb}/^{206}\text{Pb}$ and (b) $^{87}\text{Sr}/^{86}\text{Sr}_i$ versus $^{206}\text{Pb}/^{204}\text{Pb}$. Black dotted lines are mixtures between Imnaha-source mantle and depleted mantle (Wolff et al., 2008). Plum-colored arrows labeled “crustal contamination” represent the extent of interaction between Imnaha magmas and crust, which is represented by rhyolite lavas in the vicinity of the CRBG magma storage zone (Wolff et al., 2008).

Samples from this study are compared to a more regional view in Figure 31. For the most part, the samples fall within the boundaries of the main-phase CRBG lavas. MS-LFR1 (Littlefield Rhyolite) and EJ-12-13B (Jump Creek rhyolite) fall outside of the Grande Ronde field in Figure 31b, but they follow the arrow pathway of crustal contamination. Overall, the rhyolites have both craton and accreted terrane trends, but most rhyolites are similar to the Grande Ronde trend.

Oxygen Isotopes

Silicate minerals in rocks of the crust of the Earth are in general enriched in ^{18}O whereas meteoric water and seawater on the surface of the Earth are depleted. As a result, the isotope composition of oxygen in igneous rocks may be altered by isotope-exchange reactions with water at elevated temperatures (Faure, 2001). Two models have been proposed to explain low $\delta^{18}\text{O}$ for rhyolites of the Snake River Plain-Yellowstone hotspot trend. These models differ on the timing of protolith alteration, depth at which alteration occurs, and physical controls on the extent of alteration. The first model involves a single caldera system or multiple, near contemporaneous, overlapping systems. In this scenario, the earliest erupted products become hydrothermally altered and are subsequently melted at shallow depths during cyclical volcanism at the same or overlapping eruptive centers to form low $\delta^{18}\text{O}$ rhyolites (Bindeman and Valley, 2001; Watts et al., 2011). Evidence of small to dramatic ($>5\%$) $\delta^{18}\text{O}$ depletions occur in magmas from Yellowstone that erupted after caldera-forming eruptions (Bindeman and Valley, 2001). In the Yellowstone Volcanic Plateau, the majority of erupted products are normal $\delta^{18}\text{O}$ rhyolites while low-

$\delta^{18}\text{O}$ units are primarily small volume lavas erupted following episodes of caldera collapse.

The second model involves a pre-existing crustal magma that was hydrothermally altered a significant amount of time prior to the onset of Miocene magmatism (Boroughs et al., 2012). Boroughs et al. (2005) determined hydrothermally altered zones of the granitic crust of the Idaho batholith could be a pre-existing low $\delta^{18}\text{O}$ source to generate low $\delta^{18}\text{O}$ rhyolites upon melting in the volcanic fields associated with the magmas of the Snake River Plain.

The low $\delta^{18}\text{O}$ of selected rhyolites of this study cannot be explained by either of these models. The locations of the silicic centers are too far west of the Cretaceous plutons of the Idaho Batholith to be explained by remelting of these pre-existing low $\delta^{18}\text{O}$ Idaho Batholith sources. The lowest oxygen ratios come from EJ-12-17 (Rhyolite of Bully Creek, $\delta^{18}\text{O}_{\text{plag}} = 1.6 \text{ ‰}$) and MS-LFR1 (Littlefield Rhyolite, $\delta^{18}\text{O}_{\text{plag}} = 1.8 \text{ ‰}$), which erupted ~80 km west of the cratonic margin. The model of remelting of altered silicic rocks at centers of multiple, confocal caldera cycles is irreconcilable because there is no evidence of multiple caldera cycles at any of the locations where we observe low $\delta^{18}\text{O}$ values. Hydrothermal alteration of terrane rocks, as opposed to Cretaceous plutons of the Idaho Batholith, prior to melting may explain the low $\delta^{18}\text{O}$ values of this study.

CONCLUSIONS

This study presents new data for Sr, Pb, and O isotopes of mostly mid-Miocene rhyolites erupted in eastern Oregon as well as new $^{40}\text{Ar}/^{39}\text{Ar}$ ages for a number of silicic centers throughout the region. These silicic centers, which formed contemporaneously with regional flood basalt activity, represent mantle and crustal magmatism west of the Proterozoic cratonic lithosphere and the Mesozoic accreted terrains lithospheric transition. Sr isotopic analyses of twenty-seven Miocene rhyolites and dacites in eastern Oregon confirms the presence of high and variable $^{87}\text{Sr}/^{86}\text{Sr}_i$ ratios that are not consistent with the current location of the MSL (0.7040) and CSL (0.7060) boundaries. This suggests that the craton–accreted terrane boundary may not be as well defined isotopically as previously thought and that second order crustal variations westward of the craton are likely responsible for this discrepancy. Instead of elevated $^{87}\text{Sr}/^{86}\text{Sr}_i$ variations arising from a westward-dipping decollement as suggested by Leeman et al. (1992), we suggest that the primary control is the crustal composition of accreted terranes of specific areas of eastern Oregon. Secondary control on observed $^{87}\text{Sr}/^{86}\text{Sr}_i$ ratios is possibly associated with the amount of basalt involved in the generation of these rhyolites. Because $^{87}\text{Sr}/^{86}\text{Sr}_i$ ratios in the contemporaneous mafic magmas of the CRBG vary considerably, changing the type and amount of this “basalt component” in rhyolites in eastern Oregon could generate high or low Sr isotope ratios relative to the surrounding crust.

Pb isotopes also vary, though data of individual rhyolites from the same region are similar to one another relative to entire data spread. In comparison to Sr and Pb

isotope relations among the main-phase CRBG lavas, samples from this study tend to lie along the boundaries of the Grande Ronde, Picture Gorge, and Steens basalts, though samples from the Littlefield Rhyolite and Jump Creek rhyolite lie along trends of crustal contamination (Figure 31). Continued Pb isotope analysis of mid-Miocene silicic volcanic rocks from eastern Oregon would allow for a more robust data set to generate further conclusions.

There is a significant range in oxygen isotopes in samples from this study area, with the approximate bulk rock $\delta^{18}\text{O}$ values ranging from 2.1 to 10.5‰. The lowest $\delta^{18}\text{O}$ samples are from the Littlefield Rhyolite. Low $\delta^{18}\text{O}$ values document high temperature interaction of magmatic protoliths with meteoric waters. Yet, the locations of these units do not conform to the hypothesized models of producing low $\delta^{18}\text{O}$ rhyolites. There must be another cause at play to generate these low $\delta^{18}\text{O}$ units. Hydrothermal alteration of terrane rocks prior to melting may explain the low $\delta^{18}\text{O}$ values of this study.

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APPENDIX A: SAMPLES

Sample Nomenclature

Rhyolite and tuff samples with prefix “EJ” were collected during a one-week fieldwork period in September 2012. Samples with prefix “MS”, “AS”, “TF”, “AL”, and “CR” were collected from various field areas by Martin Streck, Arron Steiner, Phil Marcy, Adam Large, and Chris Ricker, respectively.

Optical Microscopy

40- μm thin sections were prepared on standard rectangular slides to determine the petrography of the samples and investigate mineral textures using a Leica optical microscope (Table 7).

Table 7. Chart describing mineralogical features of samples from thin section. Percent mineral (%Min) corresponds to the approximate percentage of the listed mineral phases, including in the groundmass. %Ph corresponds to the approximate overall phenocrysts percentage.

Unit	Sample	Mineral	%Min	Size (mm)	%Ph	Texture	Comments
DNB	EJ-12-01	plagioclase	20	0.1	0	flow-banded	Microclites
		oxide	2	0.02		aphyric	Birdseye extinction in biotite
		biotite	1	0.1			
BDC	EJ-12-03	quartz	10	2	15	porphyritic	Groundmass is light cream color
		sanidine	5	1 - 3			
		oxide	1	0.02			
CBR	EJ-12-05B	sanidine	1	0.2 - 1.5	3	hyalopilitic	Basalt xenoliths with plag, biotite, and oxide,
		plagioclase	<1	0.2 - 1.5		porphyritic	0.5 - 2 mm, 5% abundance
		amphibole	<1	0.1 - 0.5		flow-banded	Possibly a tuff because of xenoliths, shard
		biotite	<1	0.1 - 0.4			fragments, flow-banded texture
		quartz	1	0.4			
BER	EJ-12-07	biotite	15	0.3	0	aphyric	Equant pseudomorphs after pyroxene with grain
		plagioclase	10	0.1		hyalopilitic	shape still preserved
DIT	EJ-12-08	plagioclase	2	1.5	2	porphyritic	Relict shard texture
		oxide	1	0.02			Devitrified
LIR	EJ-12-10	plagioclase	3	0.3 - 1.5	3	porphyritic	Some globular regions (0.6 mm diameter) of darker
		Fe-rich clinopyroxene	1	0.3			brown, 5% abundance, is an unknown artifact
OMR	EJ-12-12	sanidine	2	1.5	2	perlitic	There are both resorbed and normal feldspars
		clinopyroxene	<1	0.5			Fracture that has oxidized Fe around it
JCR	EJ-12-13B	plagioclase	20	0.5 - 1.5	20	porphyritic	Pseudomorphs after fayalite
		oxide	1	0.02 - 0.2			Some resorbed plagioclase, including a 15 mm xl
YMR	EJ-12-14	quartz	17	0.7	25	perlitic	Apatite and biotite within plag
		plagioclase	8	1 - 2		vitroclastic	Darker vein 1.5 mm across also shows perlitic GM
		clinopyroxene	1	1.5			with same xl suite on a smaller scale. Appears to be
		zircon	<1	0.5			tuff inclusion because of shard texture. May be
DAM	EJ-12-15	plagioclase	3	0.3 - 1	2	hyalopilitic	insitu fragmentation and opening of shear bands
		Mg-rich orthopyroxene	1	0.1		perlitic	Brown glass better preserved around crystals

CMR	EJ-12-16	plagioclase clinopyroxene oxide	7 3 1	1.5 0.5 0.2	10	porphyritic hyalopilitic	Microspherulites (devitrification feature) 0.1 mm in groundmass, 10% abundance
RBC	EJ-12-17	plagioclase clinopyroxene oxide	7 3 <1	1.5 0.5 0.3	10	porphyritic hyalopilitic	Resorbed feldspars Microlites in glass
TSD	EJ-12-19	plagioclase	30	0.05 - 0.2	0	aphyric	Microlites Incipiently devitrified
TND	EJ-12-20	plagioclase	2	0.02	0	aphyric	Devitrified circles 0.3 mm, 30% abundance
IRM	EJ-12-22B	plagioclase orthopyroxene clinopyroxene	15 3 1	0.2 - 2 0.2 - 0.5 0.2 - 0.5	15	porphyritic hyalopilitic	Dark red/brown groundmass
DMR	EJ-12-23A	plagioclase	<1	0.3 - 0.5	<1	porphyritic	
DMR	EJ-12-23B				0	aphyric	Slightly brecciated, incipiently devitrified

APPENDIX B: XRF AND ICP-MS CHEMICAL DATA

Methods

X-Ray Fluorescence

Bulk analyses of the rock samples were conducted at the GeoAnalytical Lab at Washington State University (WSU) to determine major and trace element abundances. Fresh portions of the samples were crushed into fine gravel-sized fragments using a Chipmunk Rock Crusher. The samples were ground into a powder using a tungsten carbide bowl in a swing mill. The powders were homogeneously mixed with dilithium tetraborate ($\text{Li}_2\text{B}_4\text{O}_7$) at a 2:1 flux to rock ratio in a VSM3 vortex mixer. Following the sample preparation and analytical procedures of Johnson et al. (1999), the mixture was then placed in graphite crucibles, fused into glass beads in a muffle furnace at 1000°C, reground with the swing mill, and fused again prior to analysis to ensure complete homogenization. One duplicate bead was prepared for each set of nineteen unknown samples to verify reproducibility of the analysis. A ThermoARL AdvantXP+ sequential X-ray fluorescence (XRF) spectrometer was used to analyze for ten major elements, Si, Ti, Al, Fe, Mn, Mg, Ca, Na, K, and P, and twenty-one trace elements, Ni, Cr, Sc, V, Ba, Rb, Sr, Zr, Y, Nb, Ga, Cu, Zn, Pb, La, Ce, Th, Nd, U, Cs, and As. The X-ray intensity of each element of the prepared unknown glass bead was compared versus beads prepared from known standards to determine the unknown concentrations.

Inductively Coupled Plasma Mass Spectrometry

An HP4500 inductively coupled plasma mass spectrometer (ICP-MS) was used for further trace element analyses at the GeoAnalytical Lab at WSU. Analyses were

conducted using a 1:1 dilithium tetraborate to rock ratio. The trace elements analyzed were Ba, Rb, Y, Nb, Cs, Hf, Ta, Pb, Th, U, Sr, Zr, and the fourteen naturally occurring rare earth elements (REE), La, Ce, Pr, Nd, Sm, Eu, Gd, Tb, Dy, Ho, Er, Tm, Yb, and Lu. The sample preparation and analytical procedures are after Knaack and Hooper (1994).

Major and Trace Elements

The data for both major and trace elements are shown in Table 8. All samples that were analyzed for Sr isotopes are included in this table with the exception of AL-13-015b.

Table 8. Major and trace element data from XRF and ICP-MS.

Unit	DNB	BDC	CBR	SCR	BER	DIT
Sample	EJ-12-01	EJ-12-03	EJ-12-05B	EJ-12-06A	EJ-12-07	EJ-12-08
XRF analyses, normalized 100% volatile free, weight %						
SiO₂	76.14	78.81	75.34	75.44	71.71	77.92
TiO₂	0.064	0.089	0.198	0.062	0.400	0.159
Al₂O₃	14.23	11.22	13.08	13.75	13.26	11.42
FeO*	0.66	0.80	1.40	0.98	4.51	1.66
MnO	0.072	0.019	0.049	0.087	0.058	0.015
MgO	0.08	0.00	0.23	0.14	1.23	0.23
CaO	1.19	0.05	1.15	1.19	2.93	0.59
Na₂O	3.64	2.36	2.80	3.92	3.17	4.39
K₂O	3.89	6.65	5.65	4.41	2.60	3.58
P₂O₅	0.019	0.009	0.099	0.022	0.123	0.035
XRF analyses, unnormalized ppm						
Ni	0.6	1.6	2.3	1.6	48.6	1.9
Cr	3.7	4.1	5.4	3.3	73.7	3.2
Sc	3.1	4.1	3.4	3.2	11.5	4.6
V	4.1	5.5	14.2	6.3	50.9	12.4
Ba	1120.3	148.9	499.6	871.5	1170.4	1327.7
Rb	73.7	173.7	167.9	98.4	52.2	68.4
Sr	110.8	6.1	103.6	129.3	156.2	39.7
Zr	52.5	271.6	116.5	78.4	85.8	390.6
Y	20.9	36.3	23.0	27.7	38.5	68.6
Nb	11.8	16.1	13.5	12.6	9.5	22.2
Ga	15.8	15.4	14.7	15.0	15.1	19.7
Cu	1.3	6.4	3.6	6.8	45.6	4.7
Zn	27.2	68.8	30.4	42.6	57.3	121.4
Pb	16.7	45.6	23.0	20.8	11.1	10.4
La	20.7	33.7	24.8	15.9	21.8	38.7
Ce	30.8	48.9	47.6	27.3	41.7	85.3
Th	6.6	16.2	16.9	4.4	4.3	7.7
Nd	15.1	33.3	17.3	13.4	22.6	40.6
U	2.3	3.4	8.1	2.9	1.8	3.2

Table 8 continued. Major and trace element data from XRF and ICP-MS.

Unit	DNB	BDC	CBR	SCR	BER	DIT
Sample	EJ-12-01	EJ-12-03	EJ-12-05B	EJ-12-06A	EJ-12-07	EJ-12-08
ICP-MS analyses, ppm						
La	21.04	32.04	25.23	16.37	22.12	37.70
Ce	31.48	47.37	46.83	31.37	40.83	90.30
Pr	4.43	8.93	5.28	3.80	5.46	10.47
Nd	15.41	33.59	18.51	14.52	21.89	41.23
Sm	3.44	7.32	3.83	3.83	5.25	9.81
Eu	0.73	0.43	0.55	0.58	1.14	1.59
Gd	3.07	6.03	3.36	3.83	5.51	9.60
Tb	0.55	1.11	0.58	0.71	0.97	1.83
Dy	3.33	7.14	3.72	4.50	6.21	12.13
Ho	0.66	1.51	0.77	0.91	1.35	2.64
Er	1.83	4.39	2.23	2.55	3.76	7.74
Tm	0.28	0.68	0.35	0.39	0.58	1.23
Yb	1.80	4.62	2.38	2.59	3.69	8.15
Lu	0.28	0.72	0.39	0.41	0.59	1.29
Ba	1151.42	152.84	510.72	894.79	1174.20	1358.46
Th	6.94	13.00	16.89	4.80	3.85	7.36
Nb	8.86	9.28	12.96	11.07	8.46	20.61
Y	18.72	34.19	21.35	25.69	37.05	65.58
Hf	2.42	8.87	3.88	3.31	2.92	10.59
Ta	0.80	0.47	1.67	0.84	0.62	1.33
U	2.66	4.41	7.87	2.54	1.43	3.28
Pb	15.18	43.86	21.65	19.49	10.36	10.39
Rb	75.4	178.5	168.7	98.8	51.0	69.4
Cs	2.02	2.77	8.11	3.02	1.76	1.71
Sr	117	8	108	133	159	44
Sc	2.88	2.62	3.21	3.52	12.08	3.41
Zr	55	282	119	81	85	396

Table 8 continued, Major and trace element data from XRF and ICP-MS.

Unit	LIR	OMR	JCR	YMR	DAM	CMR
Sample	EJ-12-10	EJ-12-12	EJ-12-13B	EJ-12-14	EJ-12-15	EJ-12-16
XRF analyses, normalized 100% volatile free, weight %						
SiO₂	71.08	74.19	73.52	77.14	72.78	73.04
TiO₂	0.434	0.312	0.511	0.153	0.257	0.717
Al₂O₃	12.44	12.92	13.99	12.17	14.44	12.74
FeO*	5.79	3.15	1.81	1.40	1.86	3.51
MnO	0.138	0.071	0.021	0.038	0.054	0.078
MgO	0.23	0.51	0.03	0.14	0.57	0.34
CaO	1.62	1.05	1.46	0.49	2.05	1.64
Na₂O	4.16	4.41	4.11	3.70	3.23	2.81
K₂O	4.04	3.38	4.38	4.74	4.68	4.99
P₂O₅	0.059	0.015	0.154	0.019	0.076	0.145
XRF analyses, unnormalized ppm						
Ni	4.6	2.5	4.0	2.6	5.6	3.9
Cr	2.9	2.0	2.8	1.7	5.3	3.2
Sc	7.5	0.6	7.8	1.3	5.6	11.8
V	3.4	1.0	12.3	4.0	23.1	12.4
Ba	1654.0	1438.5	2000.1	247.1	1520.3	1369.3
Rb	111.1	160.7	82.7	140.4	87.4	147.2
Sr	143.2	33.9	274.2	20.4	244.9	178.8
Zr	534.3	726.7	575.2	296.8	155.7	295.1
Y	87.0	100.0	51.6	88.0	17.1	41.9
Nb	36.3	44.9	44.1	36.8	9.4	19.2
Ga	25.5	25.9	22.0	23.3	14.2	18.1
Cu	7.5	2.6	0.7	4.5	9.0	2.6
Zn	172.6	180.9	102.6	105.3	35.2	91.0
Pb	20.5	24.8	25.3	20.8	15.9	23.9
La	64.3	66.1	82.5	65.3	21.1	47.5
Ce	125.5	147.0	164.7	125.5	41.3	88.8
Th	13.4	15.4	9.0	13.5	6.8	17.0
Nd	65.1	75.1	73.2	56.7	15.7	39.4
U	4.1	4.0	1.5	4.5	1.5	5.7

Table 8 continued. Major and trace element data from XRF and ICP-MS.

Unit	LIR	OMR	JCR	YMR	DAM	CMR
Sample	EJ-12-10	EJ-12-12	EJ-12-13B	EJ-12-14	EJ-12-15	EJ-12-16
ICP-MS analyses, ppm						
La	62.25	71.54	86.46	64.49	24.06	45.99
Ce	128.56	149.47	168.36	129.69	42.55	89.79
Pr	16.26	18.80	20.01	15.60	4.73	10.66
Nd	65.29	75.14	74.07	58.58	16.60	40.45
Sm	15.20	17.35	14.14	13.01	3.16	8.52
Eu	3.74	3.04	2.80	0.50	0.67	1.70
Gd	14.97	17.18	11.33	12.73	2.74	7.71
Tb	2.62	2.99	1.77	2.35	0.44	1.27
Dy	16.50	18.79	10.16	15.16	2.71	7.65
Ho	3.41	3.91	1.96	3.23	0.56	1.53
Er	9.52	10.92	5.16	9.30	1.64	4.11
Tm	1.40	1.63	0.74	1.41	0.25	0.60
Yb	8.81	10.50	4.62	9.02	1.70	3.81
Lu	1.41	1.65	0.69	1.41	0.28	0.58
Ba	1662.12	1456.96	2111.15	251.34	1550.00	1380.32
Th	13.50	15.48	9.52	13.43	6.94	16.66
Nb	34.44	43.94	42.76	36.44	8.19	18.07
Y	86.13	97.97	49.26	85.74	15.83	39.77
Hf	14.06	18.10	13.60	9.62	4.23	8.00
Ta	2.16	2.62	2.40	2.30	0.70	1.36
U	3.87	4.34	2.99	4.08	2.67	5.45
Pb	19.12	22.98	23.76	19.65	14.21	24.29
Rb	109.3	160.0	83.7	142.6	87.2	145.8
Cs	1.89	9.41	0.94	3.48	3.27	5.18
Sr	147	37	285	24	256	185
Sc	8.08	0.86	7.03	1.42	4.34	11.92
Zr	542	744	578	296	158	298

Table 8 continued. Major and trace element data from XRF and ICP-MS.

Unit	RBC	TSD	TND	IRM	DMR	DMR
Sample	EJ-12-17	EJ-12-19	EJ-12-20	EJ-12-22B	EJ-12-23A	EJ-12-23B
XRF analyses, normalized 100% volatile free, weight %						
SiO₂	72.40	77.06	76.83	64.65	76.72	76.90
TiO₂	0.704	0.056	0.059	0.639	0.118	0.051
Al₂O₃	12.65	12.97	13.08	16.70	13.03	13.34
FeO*	3.92	0.55	0.60	4.17	1.38	1.17
MnO	0.089	0.068	0.072	0.069	0.033	0.044
MgO	0.40	0.21	0.08	2.38	0.03	0.03
CaO	1.74	0.87	0.77	5.45	1.05	0.69
Na₂O	2.59	2.75	3.14	3.07	3.12	2.90
K₂O	5.37	5.45	5.34	2.68	4.50	4.86
P₂O₅	0.145	0.023	0.025	0.194	0.013	0.015
XRF analyses, unnormalized ppm						
Ni	3.4	3.6	4.0	22.1	0.1	0.0
Cr	1.1	3.7	2.7	46.7	3.5	3.0
Sc	12.4	3.4	2.8	11.2	3.3	3.3
V	14.1	2.2	1.9	91.8	4.7	0.7
Ba	1286.8	195.4	86.0	870.5	1597.1	579.8
Rb	151.3	107.9	106.6	36.0	92.8	122.3
Sr	168.6	26.7	18.4	469.5	90.6	33.2
Zr	293.3	58.5	62.9	131.8	170.1	72.9
Y	41.4	34.2	34.5	14.6	37.0	35.2
Nb	19.9	21.2	21.7	5.6	10.0	10.7
Ga	17.5	14.3	14.9	18.4	16.9	16.3
Cu	3.5	3.5	1.3	20.1	1.1	0.2
Zn	87.4	23.3	25.4	68.1	60.8	39.6
Pb	24.5	22.6	21.8	9.1	14.2	16.6
La	45.6	8.5	15.4	18.5	32.8	21.3
Ce	93.4	21.4	32.0	35.6	61.8	43.9
Th	17.2	7.4	9.1	3.9	7.8	9.9
Nd	41.0	11.9	13.4	14.8	27.9	19.0
U	3.9	3.4	3.2	1.5	1.5	3.5

Table 8 continued. Major and trace element data from XRF and ICP-MS.

Unit	RBC	TSD	TND	IRM	DMR	DMR
Sample	EJ-12-17	EJ-12-19	EJ-12-20	EJ-12-22B	EJ-12-23A	EJ-12-23B
ICP-MS analyses, ppm						
La	45.44	10.88	13.50	19.36	36.35	22.69
Ce	89.22	22.05	26.90	36.66	69.29	44.39
Pr	10.63	2.81	3.32	4.38	8.06	5.19
Nd	40.70	11.11	12.60	16.42	29.18	18.85
Sm	8.58	3.27	3.47	3.39	5.98	4.53
Eu	1.71	0.32	0.29	0.99	0.79	0.23
Gd	7.76	3.55	3.76	2.90	5.61	4.66
Tb	1.26	0.70	0.74	0.45	0.98	0.87
Dy	7.70	4.79	5.03	2.58	6.14	5.69
Ho	1.52	1.06	1.10	0.50	1.29	1.20
Er	4.11	3.28	3.34	1.32	3.70	3.39
Tm	0.60	0.53	0.55	0.19	0.56	0.52
Yb	3.77	3.71	3.83	1.18	3.61	3.36
Lu	0.59	0.62	0.64	0.19	0.57	0.52
Ba	1302.90	200.24	88.07	879.70	1626.24	588.85
Th	16.69	7.73	8.50	3.62	8.52	9.43
Nb	17.96	20.50	20.80	5.36	9.61	9.49
Y	39.48	31.96	32.36	13.32	35.11	33.87
Hf	8.00	3.09	3.28	3.27	5.14	3.19
Ta	1.36	1.98	2.06	0.41	0.79	1.07
U	5.32	3.32	3.48	1.40	3.11	4.00
Pb	24.14	20.75	21.40	8.20	14.88	16.07
Rb	150.0	108.8	108.3	35.6	93.2	122.5
Cs	5.26	3.84	3.52	1.14	3.87	6.01
Sr	172	31	21	474	96	37
Sc	11.82	2.84	2.97	11.22	3.41	3.14
Zr	294	60	66	128	172	75

Table 8 continued. Major and trace element data from XRF and ICP-MS.

Unit	UNI	UNI	UNI	DIT	DIT	DIT
Sample	CR-U3A	CR-U4B	MS-12-14b	CR-U35a	MS-PCIT 1A	MS-12-29.1
XRF analyses, normalized 100% volatile free, weight %						
SiO₂	76.10	72.07	75.89	74.11	73.82	76.30
TiO₂	0.09	0.27	0.212	0.256	0.32	0.173
Al₂O₃	13.34	15.30	13.83	13.15	13.58	12.30
FeO*	1.19	1.68	0.96	2.61	2.91	2.04
MnO	0.02	0.05	0.017	0.064	0.03	0.054
MgO	0.06	0.48	0.10	0.09	0.31	0.04
CaO	0.72	2.12	1.26	0.48	1.04	0.58
Na₂O	3.25	4.10	3.90	3.16	2.29	3.78
K₂O	5.22	3.84	3.77	6.06	5.64	4.73
P₂O₅	0.01	0.08	0.049	0.021	0.06	0.014
XRF analyses, unnormalized ppm						
Ni	0.0	0.0	3.7	4.0	3.8	2.4
Cr	1.3	5.3	3.8	3.4	4.0	3.4
Sc	3.0	3.8	3.4	1.7	5.1	4.1
V	0.3	21.3	8.5	6.3	10.0	3.1
Ba	1102.1	794.4	1191.3	1815.4	1382.4	1411.0
Rb	106.6	82.3	80.3	84.6	79.1	74.2
Sr	63.8	263.3	173.4	54.4	104.2	27.6
Zr	125.4	118.0	165.8	390.2	384.3	428.8
Y	26.8	16.3	25.4	62.1	76.5	93.0
Nb	17.2	12.3	14.7	24.3	20.1	24.2
Ga	16.9	15.8	17.0	22.0	21.5	21.5
Cu	1.2	0.5	5.0	2.9	5.2	1.4
Zn	50.5	39.2	26.1	149.1	84.5	152.4
Pb	15.2	13.2	14.8	15.1	14.6	18.7
La	31.2	25.4	36.6	44.6	47.6	47.4
Ce	59.1	44.7	69.6	82.8	94.1	91.9
Th	13.0	9.7	9.8	6.6	8.4	7.6
Nd	27.2	17.4	26.1	43.7	46.5	49.5
U	3.8	4.9	3.3	1.8	3.1	4.7

Table 8 continued. Major and trace element data from XRF and ICP-MS.

Unit	UNI	UNI	UNI	DIT	DIT	DIT
Sample	CR-U3A	CR-U4B	MS-12-14b	CR-U35a	MS-PCIT 1A	MS-12-29.1
ICP-MS analyses, ppm						
La	34.51	26.22	40.29	46.34	46.79	46.68
Ce	63.11	47.11	70.95	92.01	94.93	93.30
Pr	7.39	5.24	7.78	11.73	12.29	12.42
Nd	26.00	18.00	26.38	46.24	49.40	49.54
Sm	5.23	3.33	4.59	10.28	11.48	11.60
Eu	0.76	0.79	0.92	1.91	2.24	1.71
Gd	4.57	2.81	3.95	9.91	11.93	12.25
Tb	0.77	0.44	0.62	1.75	2.10	2.30
Dy	4.81	2.68	3.66	11.21	13.51	15.61
Ho	0.94	0.57	0.75	2.37	2.88	3.39
Er	2.62	1.57	2.01	6.74	8.14	10.11
Tm	0.40	0.24	0.30	1.03	1.23	1.56
Yb	2.54	1.64	1.86	6.66	7.77	10.24
Lu	0.40	0.28	0.31	1.06	1.23	1.64
Ba	1119	815	1216	1985	1401	1419
Th	10.94	8.99	10.02	7.36	7.03	7.81
Nb	17.61	12.64	13.55	23.86	19.82	23.28
Y	26.24	15.81	23.96	62.02	74.28	90.28
Hf	4.43	3.39	4.44	10.34	10.29	11.64
Ta	1.64	1.30	1.25	1.33	1.32	1.39
U	3.54	3.45	3.09	3.34	3.03	3.52
Pb	15.84	13.78	13.65	15.25	15.22	17.53
Rb	107.3	83.4	79.7	88.6	78.5	74.6
Cs	3.29	3.35	2.45	2.92	2.78	2.90
Sr	64	267	177	53	101	31
Sc	3.6	3.8	3.3	4.8	5.3	3.8
Zr	132	120	163	424	396	439

Table 8 continued. Major and trace element data from XRF and ICP-MS.

Unit	DIT	STR	STR	DCT	DMR	LIR
Sample	MS-11-20	AS-SV-151	AS-SV-190	AS-SV-243	MS-08-DR2	MS-LFR1
XRF analyses, normalized 100% volatile free, weight %						
SiO₂	67.95	73.19	76.93	76.80	73.91	72.29
TiO₂	1.101	0.35	0.19	0.208	0.33	0.70
Al₂O₃	14.55	14.57	12.66	11.28	14.11	12.69
FeO*	5.69	2.23	1.19	2.63	2.08	3.94
MnO	0.128	0.053	0.037	0.062	0.04	0.09
MgO	1.09	0.23	0.13	0.01	0.03	0.44
CaO	2.97	1.99	0.81	0.14	2.08	1.73
Na₂O	3.12	3.86	4.12	4.19	3.78	2.51
K₂O	3.02	3.422	3.912	4.64	3.62	5.47
P₂O₅	0.375	0.097	0.023	0.026	0.03	0.14
XRF analyses, unnormalized ppm						
Ni	3.6	2.5	3.0	1.8	0.0	0.0
Cr	6.9	5.4	3.2	3.7	2.1	3.1
Sc	16.4			1.4	10.8	11.1
V	68.7	29.8	5.2	8.0	5.7	12.5
Ba	1296.4			63.7	1480.4	1259.3
Rb	54.1			115.5	77.2	150.6
Sr	233.6			7.0	193.0	161.9
Zr	272.4			768.7	362.8	287.7
Y	50.5			56.5	45.8	40.0
Nb	17.7			55.8	14.1	17.7
Ga	21.3	15.7	14.4	28.1	20.0	18.3
Cu	6.6	7.2	2.7	3.3	3.0	1.9
Zn	140.8	48.5	35.2	186.7	84.2	89.1
Pb	10.6			26.5	12.7	22.8
La	33.7			61.3	35.6	43.7
Ce	67.7			149.6	67.6	85.6
Th	5.6			11.0	8.8	15.9
Nd	39.0			63.3	33.3	37.9
U	1.8			2.1	3.5	5.0

Table 8 continued. Major and trace element data from XRF and ICP-MS.

Unit	DIT	STR	STR	DCT	DMR	LIR
Sample	MS-11-20	AS-SV-151	AS-SV-190	AS-SV-243	MS-08-DR2	MS-LFR1
ICP-MS analyses, ppm						
La	33.49	25.41	29.51	62.92	35.07	45.52
Ce	69.91	46.23	59.07	155.99	70.32	88.88
Pr	9.12	5.36	7.04	17.03	8.77	10.72
Nd	37.89	19.62	26.15	63.24	34.54	40.35
Sm	8.81	3.82	5.67	13.96	7.79	8.62
Eu	2.38	0.88	0.80	0.71	1.58	1.67
Gd	8.77	3.35	5.36	11.42	7.51	7.75
Tb	1.46	0.54	0.97	2.08	1.29	1.27
Dy	9.25	3.26	6.22	12.60	8.23	7.63
Ho	1.93	0.67	1.31	2.36	1.71	1.53
Er	5.32	1.88	3.88	6.42	4.91	4.16
Tm	0.78	0.29	0.60	0.98	0.75	0.61
Yb	4.97	1.93	3.96	6.32	4.78	3.80
Lu	0.82	0.32	0.63	0.98	0.75	0.59
Ba	1319	1485.10	1254.51	71	1523	1277
Th	5.32	7.70	8.46	11.30	6.83	16.37
Nb	17.29	8.78	12.29	55.36	13.76	17.94
Y	50.07	17.93	35.96	53.86	45.47	39.45
Hf	7.10	4.59	5.38	17.81	8.80	7.92
Ta	1.07	0.74	1.01	3.37	0.90	1.34
U	2.53	3.67	3.42	2.19	2.73	5.26
Pb	10.65	13.91	13.10	24.61	13.19	23.73
Rb	53.7	86.65	90.49	114.2	77.5	151.1
Cs	2.27	4.20	3.32	1.46	3.46	5.13
Sr	239	231.02	76.82	10	192	162
Sc	16.5	5.83	4.68	0.9	11.3	11.7
Zr	278	171.00	176.24	800	375	296

Table 8 continued. Major and trace element data from XRF and ICP-MS.

Unit	STK	SCT	TFI	IDL	MCB	VCR	BLB
Sample	MS-11-45	MS-11-15	TF-88A	MS-13-05	MS-13-09	MS-13-10	MS-13-12
XRF analyses, normalized 100% volatile free, weight %							
SiO₂	76.73	73.80	77.19	76.31	73.59	76.84	77.00
TiO₂	0.096	0.32	0.16	0.040	0.336	0.082	0.096
Al₂O₃	12.21	12.49	11.58	13.90	13.34	12.53	12.11
FeO*	1.42	3.23	1.88	0.55	2.13	1.04	1.38
MnO	0.035	0.07	0.03	0.097	0.044	0.018	0.049
MgO	0.03	0.05	0.00	0.03	0.35	0.07	0.11
CaO	0.46	0.83	0.32	0.61	1.23	0.44	0.41
Na₂O	3.58	3.40	3.34	3.97	2.72	3.09	3.99
K₂O	5.42	5.79	5.51	4.47	6.17	5.87	4.65
P₂O₅	0.028	0.02	0.01	0.019	0.076	0.015	0.193
XRF analyses, unnormalized ppm							
Ni	0.8	0.4	1.9	0.0	4.4	0.9	13.1
Cr	3.1	2.6	2.1	4.5	4.3	3.0	4.4
Sc	0.4	1.4	0.0	6.1	4.9	1.1	1.0
V	1.6	5.2	2.8	1.0	19.5	3.6	8.3
Ba	20.3	1768.4	130.6	63.7	843.6	87.0	75.2
Rb	152.2	134.9	146.2	130.4	143.3	178.2	144.4
Sr	4.1	27.4	3.0	11.7	90.9	11.0	8.6
Zr	304.6	669.1	670.5	45.5	221.7	163.3	282.8
Y	75.8	93.3	121.6	35.0	36.5	64.5	155.0
Nb	50.0	41.6	47.1	19.9	16.4	40.3	52.2
Ga	21.5	24.3	26.1	17.0	16.3	21.5	23.1
Cu	5.5	5.6	5.2	1.7	6.9	4.6	10.1
Zn	112.8	165.3	193.1	43.4	43.4	79.2	108.6
Pb	26.6	22.6	27.7	20.0	18.1	27.9	24.2
La	59.7	62.4	73.4	7.7	40.3	45.4	136.5
Ce	118.7	139.0	153.0	19.7	73.0	95.7	143.9
Th	14.5	13.5	16.6	4.9	12.5	17.9	15.0
Nd	54.0	68.4	76.1	7.7	30.3	40.9	112.6
U	5.6	5.6	5.0	4.5	6.1	6.7	4.5

Table 8 continued. Major and trace element data from XRF and ICP-MS.

Unit	STK	SCT	TFI	IDL	MCB	VCR	BLB
Sample	MS-11-45	MS-11-15	TF-88A	MS-13-05	MS-13-09	MS-13-10	MS-13-12
ICP-MS analyses, ppm							
La	59.84	67.84	76.27	8.47	40.58	45.44	139.92
Ce	119.95	148.11	164.67	19.69	77.53	96.58	151.34
Pr	14.36	17.63	20.20	2.46	8.99	11.56	29.59
Nd	52.78	71.42	81.25	8.89	32.83	43.01	116.71
Sm	12.28	16.45	19.41	3.06	6.67	10.28	25.11
Eu	0.14	3.36	1.18	0.23	0.90	0.16	2.88
Gd	11.67	16.33	19.95	3.72	5.95	9.76	26.10
Tb	2.12	2.82	3.57	0.80	1.02	1.77	4.21
Dy	13.35	17.70	22.89	5.44	6.45	11.40	25.23
Ho	2.73	3.69	4.81	1.17	1.33	2.35	5.16
Er	7.72	10.39	13.43	3.35	3.87	6.62	13.63
Tm	1.17	1.54	2.01	0.52	0.59	0.99	1.92
Yb	7.34	9.76	12.54	3.34	3.86	6.25	11.25
Lu	1.14	1.62	1.94	0.51	0.63	0.96	1.71
Ba	22	1870	135.18	67	856	92	76
Th	14.54	14.41	17.30	4.90	11.89	16.93	14.66
Nb	50.55	41.26	47.56	19.18	15.94	40.51	52.65
Y	72.96	93.74	122.52	33.82	35.98	62.76	153.14
Hf	10.52	16.85	19.38	2.78	6.48	7.01	10.54
Ta	3.33	2.42	2.98	1.68	1.37	3.00	3.53
U	5.67	4.29	5.12	4.52	5.12	6.53	5.27
Pb	24.96	21.70	26.65	17.87	17.69	25.82	23.70
Rb	153.9	134.9	155.62	131.2	141.1	176.6	144.4
Cs	4.18	3.16	4.05	4.93	3.33	5.00	3.96
Sr	6	29	5.06	13	96	13	10
Sc	0.5	1.4	0.09	4.5	4.7	1.1	0.5
Zr	310	719	714.39	49	226	169	292

APPENDIX C: RADIOGENIC ISOTOPES

Methods

Sample Preparation

Selecting Crystals, Groundmass, and Whole Rock Powders

Whole rock samples were chipped and sieved into three size fractions: 0.991-2 mm, 0.425-0.991 mm, and 0.104-0.425 mm. These chips were then handpicked under a microscope for either fresh groundmass or feldspar crystals, ensuring they were free of inclusions and lacked alteration due to weathering. Approximately 0.5 grams of groundmass or feldspar crystals was collected for each sample. Whole rock powders were prepared for selected samples using a tungsten carbide ring mill at the GeoAnalytical Lab at WSU.

Etching Crystals

Sample etching, digestion, chromatography and analyses were conducted at the Isotope Lab at NMSU. For the samples with feldspar crystals, between 15 and 30 crystals were selected with a LEICA M165C microscope to be etched. The purpose of etching was to remove adhering groundmass and to accentuate cleavage/striations to ensure accurate identification of plagioclase and/or alkali feldspar and not quartz. Crystals were placed in plastic beakers with ~2 mL 10% HF for 30 minutes. The crystals were then rinsed in ultrafiltered H₂O, and placed in a sonicating bath for 15 minutes. Among the crystals etched, approximately ten crystals that did not display altered surfaces (e.g., discoloration, adhering minerals, inclusions) were handpicked for digestion.

Digestion

Groundmass (Sr)

For each sample, ~70 mg were weighed and placed in a 15 mL Teflon™ beaker. Samples were dissolved in the cleanroom at NMSU over the course of three days. First, ~2 mL of concentrated HNO₃ and ~2mL of concentrated HF, both doubly distilled, were added to each beaker. Beakers were sealed and left on a hotplate overnight at 100°C. After ~18 hours, the lids were removed and the acid was evaporated. Next, ~4 mL of 7N HNO₃ was added and the beakers were sealed and left on a hotplate overnight at 100°C. After ~18 hours, the lids were removed and the acid was evaporated. Next, the samples were dissolved in ~4 mL of 6N HCl and the beakers were sealed and left on a hotplate overnight at 100°C. The final step was drying the samples and redissolving them in 1 mL 2.5N HCl. Samples were then transferred to 1.5 mL micro-centrifuge tubes and centrifuged at 5000 rpm for three minutes in preparation for column chromatography. The purpose of centrifuging was to separate any undissolved material, of which there was none. From the 1 mL volume of sample, a 0.5 mL aliquot of 2.5N HCl sample solution was used for column chromatography.

Whole Rock Powder (Pb)

For each sample, between 100-150 mg was weighed and placed in a 60 mL Teflon™ beaker. The digestion steps were the same as those listed above for groundmass, but ~3 mL of concentrated HNO₃ and ~3 mL of concentrated HF were added to each beaker during the first step, ~4 mL of 7N HNO₃ was added during the second step, and ~4 mL of 6N HCl was added during the third step.

Feldspars (Sr)

For each sample, the weight of the ten feldspar crystals was recorded and the sample was placed in a 7 mL or 3 mL TeflonTM beaker. Digestion steps were the same as those listed above for groundmass samples, but ~2 mL of concentrated HNO₃ and ~2mL of concentrated HF were added to each 7 mL beaker and ~1 mL of concentrated HNO₃ and ~2mL of concentrated HF were added to each 3 mL beaker for the first step. For the second step, ~1 mL of 7N HNO₃ was added to each beaker. For the third step, ~2 mL of 6N HCl was added to each beaker.

Chromatography Preparation and Sample Purification

Strontium

Cation exchange resin was used to purify elemental strontium. The columns used were 0.7 cm by 15 cm PyrexTM columns with 22 mL reservoirs and ~4 mL cation exchange resin. After pretreating the columns with 10 mL 2.5N HCl, 0.5 mL of each sample, removed from the top portion of the 1.0 mL centrifuged sample solution, was loaded onto columns via a pipette. Next, two 0.5 mL washes of 2.5N HCl were added by dropping liquid along the reservoir sides to ensure complete sample loading onto the resin bed. Next, 2 mL of 2.5N HCl, followed by 9.5 mL of 2.5N HCl, and then by 5 mL of 2.5N HCl were used to remove elements other than strontium and the REEs. Each step was initiated after all acid had drained from the reservoir into the resin. Strontium was then collected in a 15 mL beaker using 6 mL of 2.5N HCl. Those purified strontium solutions were dried overnight in preparation for thermal ionization mass spectrometry

analysis. The columns were subsequently washed with 25 mL of 6N HCl to remove rare earth elements from the resin and then backwashed with H₂O.

Lead

To obtain pure Pb separates, washes collected from cation exchange chromatography were first dried and dissolved in 1 N HBr. 0.6 cm by 8 cm PyrexTM columns with 6 mL reservoirs were filled with ~2 mL of anion exchange resin. Columns were prepared using 3 mL of 1 N HBr prior to sample loading. Each sample was loaded onto a column in a 0.5 mL aliquot, followed by two washes of 0.5 mL of 1 N HBr to ensure all sample was loaded onto the resin. The columns were then washed four times with 3 mL of 1 N HBr. Each step was initiated after all acid had drained from the reservoir into the resin. Pb was collected using a strip technique with a wash of 1 mL of H₂O followed by a two washes of 3 mL of 1 N HNO₃. Columns were washed with 3 mL of 6 N HCl, 1 mL of H₂O, 3 mL of 7 N HNO₃, and 3 mL of 1 N HNO₃ to clean the resin. Purified Pb solutions were placed on the hot plate and dried.

Mass Spectrometry

Strontium Loading and Analysis

Purified strontium samples were analyzed using a VG 54 Sector Thermal Ionization Mass Spectrometer (TIMS) on center only Cathodeon beads with Re filaments. The filaments were outgassed and left to oxidize for at least 48 hours prior to loading. A dilution factor was determined for each sample to ensure between 10 and 100 ng Sr would be loaded on the filament. In general, ~100 microliters of 5% HNO₃ were added to

the dried samples and left to dissolve for at least one hour. One microliter of both dissolved strontium solution ($\sim 1\%$, $1\ \mu\text{L}$ of $100\ \mu\text{L}$) and tantalum oxide dissolved in $\sim 5\%$ phosphoric acid were loaded onto the center filament and allowed to air dry. Samples were preheated in atmosphere to volatilize the phosphoric gas and to bind strontium and tantalum oxide. Samples were then loaded into the TIMS and the TIMS was pumped down to a vacuum of $<10^{-5}$ atm. Filaments were degassed at ~ 1.5 amps under vacuum for a minute. Approximately 2.4 amps of current were required to obtain $\sim 3\ \text{V}$ ^{88}Sr for analysis. Strontium isotopes were analyzed using five Faraday collectors in dynamic mode with ratios normalized to $^{86}\text{Sr}/^{88}\text{Sr} = 0.1194$ to account for mass fractionation.

Lead Analysis

The purified lead samples were analyzed using a Thermo Scientific multicollector inductively coupled plasma mass spectrometer (MC-ICP-MS) at University of Wyoming High-Precision Isotope Laboratory. Samples were dissolved in $2\% \text{HNO}_3 + 0.01\% \text{HF}$. Thallium was added to each sample to correct for mass fractionation with Pb/Tl ratio target of 3 to 1. Samples were introduced into the plasma using a desolvating nebulizer. Lead and thallium isotopes were analyzed in static mode using six Faraday collectors. Samples were doped with NBS 997 thallium standard and mass discrimination was adjusted to $^{205}\text{Tl}/^{203}\text{Tl} = 2.387075$.

Results

NBS 987 Sr standard results averaged 0.710288 ($n=5$). For samples that had Sr concentration less than 15 ppm as measured by ICP-MS, Sr and Rb concentrations for

feldspars and groundmass were determined by isotope dilution using a high Rb/Sr spike. Concentrations of Rb and Sr determined by isotope dilution are considered accurate to about 1%. There were four groundmass samples that had age-corrected Sr ratios much lower than would be expected for this region: TF-88A, MS-13-12, MS-13-10, and EJ-12-03 (Table 5). While the corresponding feldspar samples had $^{87}\text{Sr}/^{86}\text{Sr}_i$ ranging from 0.7036 to 0.7049, $^{87}\text{Sr}/^{86}\text{Sr}_i$ of the groundmass ranged from 0.6832 to 0.7020. These unrealistic ratios likely reflect the extremely high Rb/Sr ratio of groundmass, which ranged from 13.6 to 31.2. The Rb and Sr concentrations for each of these samples were determined by isotope dilution. A minor change in the Sr concentration when it is small relative to the Rb concentration can greatly influence the Rb/Sr ratio, which is a critical aspect of age corrected initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios:

$$^{87}\text{Sr}/^{86}\text{Sr}_{\text{sample}} = ^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}} + ^{87}\text{Rb}/^{86}\text{Sr} (e^{\lambda t} - 1) \quad \text{Eq. C-1}$$

where t is the age of the sample and λ is the decay constant ($1.42 \times 10^{-11} \text{ yr}^{-1}$). The age, Rb, and Sr concentrations of the samples are available in **Error! Reference source not found.**

We believe that the ratios for TF-88A, MS-13-12, MS-13-10, and EJ-12-03 are incorrect and reflect the challenges of measuring the $^{87}\text{Sr}/^{86}\text{Sr}$ in a sample that has a low Sr concentration relative to Rb. There is likely something happening to the Rb/Sr of the groundmass in these rhyolites post-eruption.

Table 9. Age, Rb, and Sr concentration for calculating $^{87}\text{Sr}/^{86}\text{Sr}$; (Eq. C-1). For Rb and Sr concentrations, WR (whole rock) is from ICP-MS while GM (groundmass) and fsp (feldspar) are from isotope dilution. If isotope dilution was not used to determine Rb and Sr concentration for feldspar samples, estimated partition coefficients were 0.25 for Rb and 2 for Sr.

Sample		Age (Ma) ¹³	Rb (WR, ppm)	Sr	Rb (GM, ppm)	Sr	Rb (fsp, ppm)	Sr
Owyhee locality								
OMR	EJ-12-12	15.94 ²	160	37				
JCR	EJ-12-13B	10.66 ³	84	285				
YMR	EJ-12-14	15.76 ²	143	24				
DAM	EJ-12-15	13.58 ⁴	87	256				
SCT	MS-11-15	15.74 ⁵	135	29				
TFI	TF-88A	15.74 ⁵	156	5	152	3.2	54.6	10.7
Littlefield/Cottonwood Mountain								
LIR	EJ-12-10	16.75 ²	109	147				
CMR	EJ-12-16	15.8 ⁶	146	185				
RBC	EJ-12-17	17.01 ²	150	172				
LIR	MS-LFR1	16.9 ⁷	151	162			1.46	666
Dinner Creek								
DIT	EJ-12-08	~16.0 ¹	69	44				
DIT	CR-U35a	~15.5 ¹	89	53			4.72	339
DIT	MS-PCIT 1-A	15.28 ⁸	101	79				
DIT	MS-12-29.1	16.16 ⁸	75	31			0.74	91.8
DIT	MS-11-20 ¹⁴	14.84 ⁸	54	239			7.86	285
Northern locality								
UNI	CR-U3A	~16.5 ¹	107	64				
UNI	CR-U4B	16.53 ²	83	267			0.55	444
UNI	MS-12-14b	~16.5 ¹	80	177			0.35	885
DMR	EJ-12-23A	15.59 ²	93	96			2.37	1,721
DMR	EJ-12-23B	~15.5 ¹	122	37				
DMR	MS-08-DR2	~15.5 ¹	77	192				
IRM	EJ-12-22B	~15 ¹	36	474				
Strawberry Mountain								
STR	AS-SV-151	16.16 ¹⁰	87	231				
STR	AS-SV-190	14.79 ¹⁰	90	77				
Drewsey locality								
TND	EJ-12-20	~12 ¹	108	21				
TSD	EJ-12-19	~12 ¹	109	31				
BER	EJ-12-07	16.29 ²	51	159				
BLB	MS-13-12	~14 ¹	177	13	161.6	6.4	44.5	3.5
Southern locality								
CBR	EJ-12-05B	9.54 ¹¹	169	108				
SCR	EJ-12-06A	16.01 ²	99	133				
STK	MS-11-45	~10.7 ¹	154	6	160.2	2.6	48.2	15.1
MCB	MS-13-09	~15 ¹	141	96	142.1	40.1	332	531
VCR	MS-13-10	11.1 ⁹	144	10	326.6	1.5	87.8	30.8
DCT	AS-SV-243	9.74 ¹²	116	7				
Western locality								
DNB	EJ-12-01	~14 ¹	75	117				

IDL	MS-13-05	~9 ¹	131	12	134.5	10.7		
Buchanan								
BDC	EJ-12-03	16.13 ²	178	8	196.7	5.3	53.2	7.73
BDC	AL-13-015b	~16 ¹			112.0	95.8		

¹ Ar-Ar or K-Ar age is unknown, but an age estimate was made based on stratigraphic data.

² This study.

³ Armstrong et al. (1980)

⁴ Ferns and Cummings (1992)

⁵ Marcy et al. (2013)

⁶ Hooper et al. (2002)

⁷ Lees (1994)

⁸ Streck et al. (submitted)

⁹ Walker (1979)

¹⁰ Steiner (personal communication, 2013)

¹¹ Fiebelkorn et al. (1983)

¹² Jordan (2001)

¹³ All Ar-Ar and K-Ar ages are recalibrated to FCT = 28.201 Ma.

¹⁴ Feldspars for MS-11-20 crystallized from high-silica rhyolite magma of the tuff.

APPENDIX D: OXYGEN ISOTOPES

Methods

Oxygen isotope analyses were performed at University of Oregon in the Stable Isotope Laboratory. Analyses were performed on quartz, plagioclase, and alkali feldspar crystals, and whole rock powders. Grains were pretreated with 2 Torr BrF₅ overnight to remove surface and water contamination. Oxygen was isolated from the sample using laser fluorination with a 35W Newwave CO₂ IR laser and BrF₅ reagent. Released oxygen gas was analyzed using a Finnigan MAT 253 large radius 10kv gas source mass spectrometer. Ratios for these analyses are shown in Table 4.

Equations (i.e. correction factors) have been developed to convert mineral $\delta^{18}\text{O}$ data to bulk rock $\delta^{18}\text{O}$ (Bindeman et al., 2004). The MELTS program was used to determine how $\delta^{18}\text{O}$ changes with SiO₂ to treat melts (i.e., bulk rock) as mixtures of normative minerals. The correction factor for plagioclase, quartz, and alkali feldspar using bulk rock SiO₂ concentrations is:

$$\text{Plagioclase: } \delta^{18}O_{\text{magma}} = 0.027 (\text{SiO}_2 \text{ wt}\%_{\text{wholerock}}) - 1.45 + \delta^{18}O_{\text{plag}} \quad \text{Eq. D-1}$$

$$\text{Quartz: } \delta^{18}O_{\text{magma}} = \delta^{18}O_{\text{qtz}} - 0.45 \quad \text{Eq. D-2}$$

$$\text{Alkali Feldspar: } \delta^{18}O_{\text{magma}} = \delta^{18}O_{\text{kspar}} + 0.29 \quad \text{Eq. D-3}$$

with equation D-1 based on calculations by Bindeman et al. (2004), equation D-2 based on measurements in the Bishop Tuff (Bindeman and Valley, 2002), and equation D-3 based on the average $\delta^{18}O$, *melt-sanidine* value for seventeen sanidine crystals analyzed in the Timber Mountain/Oasis Valley Caldera Complex in southwest Nevada (Bindeman and Valley, 2003). Approximate bulk rock $\delta^{18}\text{O}$ values can be determined adding this

correction factor to single grain data. This correction factor will not be valid when applied to xenocrysts or crystals involved in magma mixing because the bulk rock SiO_2 content is not the same as the melt the minerals crystallized (Bindeman et al., 2004).

APPENDIX E: AGE DATING

Background

Dating using the Ar-Ar system is based on the formation of ^{39}Ar by the irradiation of potassium bearing samples with thermal and fast neutrons in a nuclear reactor. The desired reaction is $^{39}_{19}\text{K}(n,p)^{39}_{18}\text{Ar}$. ^{39}Ar is unstable and decays to ^{39}K by β -emission with a half-life of 269 years. This short half-life guarantees ^{39}Ar is absent in the sample beforehand and that ^{39}Ar can be treated as though it were stable during the short period involved in analyses (Faure and Mensing, 2005). This dating technique is more powerful than K-Ar because it can distinguish samples that have experienced argon loss or gain from those that have remained in closed systems since formation (Brownlow, 1995). By subjecting a sample to increasing temperature, the $^{40}\text{Ar}/^{39}\text{Ar}$ gas ratio released can be measured and a series of dates can be obtained from this fractional release. The plateau is the margin on the incremental heating plot that is used to calculate the age of the sample. To determine whether an incremental heating experiment yields meaningful crystallization ages, four criteria were proposed by Fleck et al. (1977) and Pringle (1993): high temperature plateaus in the age spectra should include more than three incremental heating steps and at least 50% of the total amount of ^{39}Ar released; the plateau and isochron ages should be concordant at the 95% confidence level; the $^{40}\text{Ar}/^{36}\text{Ar}$ intercepts on the isochron diagrams should be equal to the atmospheric value of 295.5 at the 95% confidence level; and the mean squared weighted deviations for the plateau ages and the isochron ages should be sufficiently small compared to F-statistic distributions.

Methods

Incremental heating experiments were performed at the $^{40}\text{Ar}/^{39}\text{Ar}$ Geochronology Laboratory at OSU on fresh feldspar and groundmass separates that were acid-rinsed prior to irradiation. All samples were loaded into quartz vials with small quantities of mineral monitor FCT-3 biotite and irradiated at the OSU TRIGA research reactor. Samples that were analyzed via the single collector MAP 215-50 mass spectrometer were heated in a double-vacuum, thermocouple-controlled, resistance furnace. Each heating step was followed by gas cleanup and then Ar isotopic composition analysis. Some of the samples were measured on the ARGUS VI, which is a multi-collector mass spectrometer. ArArCALC software provided by Koppers (2002) reduced the isotopic data and calculated age. Resulting age spectra are based on the younger Fish Canyon Tuff (FCT) age of 28.020 Ma or 28.030 Ma (Figure 32). For final age calculations, the samples ages were later recalculated using a FCT age of 28.201 (Kuiper et al., 2008). Further details of the analytical procedures are described in Duncan and Keller (2004). Plateaus generated from the incremental heating experiments are displayed in Figure 32. All analyses were undertaken using the MAP 215-50 mass spectrometer except for EJ-12-10, EJ-12-17, and EJ-12-23A, which were analyzed on the ARGUS VI.

EJ-12-03

EJ-12-06A

EJ-12-07

Figure 32. $^{40}\text{Ar}/^{39}\text{Ar}$ age plateau from incremental heating experiment.

EJ-12-10
(MAP)

EJ-12-10
(Argus)

EJ-12-12

Figure 32 continued. $^{40}\text{Ar}/^{39}\text{Ar}$ age plateau from incremental heating experiment.

EJ-12-17
(MAP)

EJ-12-17
(Argus)

EJ-12-23A
(MAP)

Figure 32 continued, $^{40}\text{Ar}/^{39}\text{Ar}$ age plateau from incremental heating experiment.

EJ-12-23A
(Argus-GM)

EJ-12-23A
(Argus-Plag)

Figure 32 continued. $^{40}\text{Ar}/^{39}\text{Ar}$ age plateau from incremental heating experiment.

MS-11-20

AS-SV-151

AS-SV-190

Figure 32 continued. $^{40}\text{Ar}/^{39}\text{Ar}$ age plateau from incremental heating experiment.