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ABSTRACT

Links Between Eruptive Styles, Magmatic Evolution, and Morphology of Shield Volcanoes: Snake River Plain, Idaho

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In this study, connections between chemical composition, eruption style, and topographic features of two shield volcanoes on the Snake River Plain, Idaho are examined. Despite their similar ages and geographic locations, two young basaltic shield volcanoes—Kimama Butte (87 Ka) and Rocky Butte (95 Ka)—have strikingly different topographic profiles. The Kimama Butte shield has a diameter of 9 km and a height of 210 m. In contrast, Rocky Butte has a broad 36 km topographic shield that rises 140 m with less than 1° slopes. The vent crater at Rocky Butte developed as a large lava blister inflated and then collapsed forming a crater in which a lava lake developed. Little spatter accumulated throughout the eruption. In contrast, high spatter mounds and spatter-fed flows flank the main summit crater at Kimama Butte.

Major- and trace-element compositions of the basaltic lavas are similar at the two shields, but distinct in Ni and Al₂O₃. The lavas range in TiO₂ concentrations from 2.6–4.5 wt.% for Kimama Butte and 2.6–4.3 wt.% for Rocky Butte. These ranges can be related to magma evolution by fractional crystallization involving plagioclase and olivine without clinopyroxene. Compositions of the pre-eruptive phenocrysts are also similar at both shields but show variation with evolution. Olivine cores in the more primitive lavas are more Mg-rich (Fo80-72) than those in the evolved rocks (Fo65-55). Plagioclase cores are similarly more calcic in the more primitive flows (An78-68) than in the evolved ones (An65-52).

Like other olivine-tholeiites on the Snake River Plain, the fO₂ and fH₂O were probably low with fO₂ at -2ΔQFM and 0.1 wt.% H₂O. Pressure of crystallization estimated from MELTS models is less than 3 kbar (~10 km deep). Calculated temperatures and magma viscosities overlap at both Kimama Butte (1226 to 1147°C and 158 to 14 Pa·s) and Rocky Butte (1251 to 1145°C and 75 to 8 Pa·s). However, Kimama Butte magma viscosities extend ~80 Pa·s higher than those for Rocky Butte lavas. The higher magma viscosities are the result of higher phenocryst proportions in spatter and spatter-fed lavas concentrated near the vent.

Because lava temperature, volatile content, and chemical composition overlap at the two volcanoes, they are probably not important controls of shield-volcano morphology. This suggests that steep-capped shields are not created as a simple function of having more silicic lavas. Melt viscosities are also similar, but Rocky Butte lacks the phenocryst-rich (>30 vol %), higher magma viscosity lavas and the high spatter ramparts that form the cap at Kimama Butte. Thus, we conclude that eruption style and phenocryst content play the most important role in developing a low-shield volcano summit. Where eruptions shifted from lava lake overflow and tube development to late fountaining with short spatter-fed phenocryst-rich flows, steeper, higher shields develop.

Keywords: basalt, shield volcano, geochemistry, geomorphology, eruption style, Snake River Plain, Quaternary
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Perhaps I owe the most thanks to Mr. Anthony Phillips who made an offhanded comment in a third period high school Earth and Space Science class that everyone should become a geologist. Before that moment, I had never considered studying geology. But because of that comment, my life most surely was changed forever.
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INTRODUCTION

Volcanic processes are important not only in Earth’s history but in the history of all the terrestrial planets. Volcanic surfaces on other planetary bodies are studied using photographs combined with limited remote sensing data. Color boundaries, albedo, and topography are used to distinguish lava flows and volcanic morphologies (Greeley and Spudis, 1981; Henderson, 2015). In rarer cases, some samples have been delivered to Earth as meteorites or collected by lunar astronauts, but their relations to specific volcanic features are unknown. Because of these limitations, understanding the volcanic systems on other planetary bodies is difficult. Estimations of the chemical composition of lavas is especially challenging and often constrained based on morphology (Guest et al., 1992; Sakimoto et al., 2003; Lena, 2007). To overcome the difficulties presented by our physical distance from these planetary volcanic landscapes, it is beneficial to study similar terrains on Earth where our access is less limited.

The eastern Snake River Plain, Idaho is one such terrain that is analogous to volcanic provinces on the Moon, Mars, and Venus (e.g., Greeley and Spudis, 1981; Head and Gifford, 1980; Guest et al., 1992; Hughes, 2001; Sakimoto et al., 2003; Hughes et al., 2004; Hauber et al., 2009; Henderson, 2015; Hughes et al., 2019). Among the volcanic features on the Moon are over 50 small shield volcanoes with diameters (3–71 km), heights, slopes (less than 5°) and volumes similar to shields on the eastern Snake River Plain (Head and Gifford, 1980). These so-called lunar domes rise to summits that have pit craters with diameters of 1–3 km (Head and Gifford, 1980; Hauber et al., 2009). Other features on the Moon, like regional dark-mantling deposits, are interpreted to be the result of effusive eruptions like Hawaiian fountains that also occur on the eastern Snake River Plain (Head and Wilson, 1992). Similarly, volcanic lowland plains on Venus contain clusters of small volcanic edifices generally less than 10 km in diameter and with low
flank angles. Wide-spread basaltic volcanism on Mars is exhibited in features such as shield volcanoes with small pit craters and volcanic plains with lava tubes and simple and complex flow features (Greeley and Spudis, 1981; Sakimoto et al., 2002; Henderson, 2015). Since volcano morphology is a parameter that can be accurately assessed with the techniques currently used in planetary exploration, it is important to understand the morphology of terrestrial shield volcanoes. Three categories of morphological profiles, capped, low-profile, or dome-shaped, are present on the Snake River Plain (Sakimoto et al., 2003; Hughes et al., 2004). The low-profile shield summit regions are slightly elevated compared to the rest of the shield while the capped variety of shields have markedly steeper flanks near the vent and a distinct elevated cap at the summit (Hughes et al., 2004). Hughes and Sakimoto (2002) and Sakimoto et al. (2003) suggested that lavas making the elevated summits of the capped shield morphology have more evolved chemical compositions than lavas that make up the distal flows. However, in a more recent study, Brady (2005) concluded that shield volcanoes with steeper caps do not necessarily have more evolved lava compositions than shields without steep-summits.

Kimama Butte and Rocky Butte represent the range of morphologies typical of shield volcanoes on the Snake River Plain. Kimama Butte, with its summit cap, is the steepest shield on the eastern Snake River Plain while Rocky Butte is one of the lowest, broadest shields (Fig. 1; Henderson, 2015). These differences in morphology make Kimama Butte and Rocky Butte ideal for better understanding the role eruptive style and chemical composition play in the creation of low-shield volcanoes. The answer to why these shields differ can then be used as a guide for understanding if-and-how eruptive style and composition can be inferred from volcano morphology, especially on other planetary bodies where only aerial imagery is available.
GEOLOGIC SETTING

Kimama Butte and Rocky Butte are two of approximately 300 low-shield volcanoes (Henderson, 2015) scattered along the 400 km long, 100 km wide eastern Snake River Plain that trends east-northeast across southern Idaho to the edge of Yellowstone Plateau volcanic field (Fig. 2). The region is bounded by Basin and Range mountains on the north and south. The area contains an extensive record of bimodal volcanism (basalt and rhyolite) with few intermediate compositions (Hughes et al., 1999; Christiansen and McCurry, 2008; Colón et al., 2018). The time-transgressive Miocene-Quaternary rhyolites are associated with the Yellowstone hot spot track (Hughes et al., 1999). Covering the rhyolitic calderas and volcanic centers are younger basaltic lavas and sediments from eolian, fluvial, and lacustrine processes (Hughes et al., 2002).

The eastern Snake River Plain is a long-lived basin that has been subsiding for millions of years. Thermal contraction following in the wake of the Yellowstone hot spot combined with the emplacement of a dense mafic sill into the middle crust are accepted causes for the subsidence (Anders and Sleep, 1992; McQuarrie and Rodgers, 1998; Rodgers et al., 2002). The presence of a sill is supported by a 9–10 km thick intermediate seismic velocity zone (Mabey, 1982; Sparlin et al., 1982; Wood and Clemens, 2002). Below this zone lies Archean crust (Leeman et al., 1985) and above the zone is a 5 km thick upper crustal layer of deformed Paleozoic sedimentary rock capped with 3–6 km of volcanic rock and interbedded sediment. The total crustal thickness is estimated to be ~42 km, 5–10 km thicker than adjacent Basin and Range region crust (Peng and Humphreys, 1998).

A deep-mantle plume is the most widely accepted explanation for this mid-plate volcanism (Leeman, 1982a; Pierce et al., 1992; McQuarrie and Rodgers, 1998; Perfit and Davidson, 2000; Hughes et al., 2002; Colón et al., 2018); however, other mechanisms such as an
upper-mantle plume or decompression melting from extension of a rift in the lithosphere are still maintained (Christiansen et al., 2002; Foulger et al., 2015). Evidence supporting the hotspot theory is the decreasing age of the volcanic rocks from 16–15 Ma just west of the Idaho, Nevada, Oregon border to 2–0 Ma on the Yellowstone Plateau (Armstrong et al., 1975; Anders et al., 2014). A two-phase mantle plume model starts 16 Ma when the large plume head initially encountered the base of the lithosphere near the present-day shared Oregon, Idaho, Nevada border. Upon contact, the originally 300 km diameter plume mushroomed to 600 km in diameter. Decompression melting led to initial flood basalts (Columbia River Flood Basalts, OR and the High Rock caldera complex, NV) and then to silicic eruptions along the hot spot track. Around 14–10 Ma, a transition from plume head to plume tail resulted in better aligned rhyolitic volcanic centers and calderas along the axis of the eastern Snake River Plain (Pierce et al., 1992; Coble and Mahood, 2016). Basaltic volcanism began shortly after the demise of rhyolitic volcanism at each center with flow ages ranging from thousands to several million years old (Armstrong et al., 1975).

**Basaltic Volcanism on the Snake River Plain**

Basaltic volcanism on the Snake River Plain has been called “plains style” volcanism as it has distinct flow characteristics, surface morphologies, and geochemistry compared to continental flood basalts and large Hawaiian-type shield volcanoes (Greeley, 1982; Christiansen and McCurry, 2008; Hughes et al., 2018). Clusters of tens to thousands of monogenetic low-shield volcanoes are a characteristic aspect of a basaltic plains region (Walker and Sigurdsson, 2000).

Presumably, development of low-shield volcanoes on the Snake River Plain begins with partial melting of the Yellowstone plume between 80 and 110 km depth (e.g., Leeman and
Vitaliano, 1976; Bradshaw, 2012; Jean et al., 2013; Putirka et al., 2009; Jean et al., 2018). The magma ascends and eventually stagnates in the crust where fractional crystallization and incorporation of local partially melted country rock occurs. Stagnation depth is probably controlled by the strength of the crustal rocks, rather than density, to form the mid-crustal sill, inferred to be a complex network of individual bodies of magma that fractionate independently and represent single pulses of magma that feed individual monogenetic low-shield volcanoes (Potter et al., 2018; Potter et al., 2019). As the basaltic magmas crystallize and evolve, the reservoirs are continually fed by repeated influx of mafic magma before erupting (Leeman and Vitaliano, 1976; Leeman, 1982a; Hanan et al., 2008; Potter et al., 2018).

Small volumes of magma (0.01–15 km$^3$), that have little substantial residence time in the crust, erupt from fissures or central vents over a time span of a few days to years to form large tube- and channel-fed lava fields some tens of kilometers across (Greeley, 1982; Geist et al., 2002; Walker and Sigurdsson, 2000; Hughes et al., 2002; Hughes et al., 2018). Small topographic shields typically less than 200 m high and ~15 km across mark the central vents. The polygenetic lava field grows as volcanoes accumulate on top of each other covering the distal flows but leaving the slightly steeper upper portion of the shields exposed. Eventually, these summits are also buried. The summit regions are typically marked by irregular pit craters filled with small lava lakes, shelly pahoehoe, ropy pahoehoe, and short spatter-fed flows. Lava lake overflow and “underflow” into tubes are important mechanism of shield growth (Greeley, 1982; Hughes et al., 1999; Hughes et al., 2018). The ultimate thickness of the basalt in the eastern Snake River Plain is not well constrained, but the Kimama drill core (Fig. 2) penetrated ~2000 m of basalt from 183 separate eruptive episodes over a 6-million-year time span (Potter et al., 2018).
METHODS

Existing geologic maps (Malde et al., 1963; LaPoint, 1977; Bond et al., 1978; Othberg et al., 2012; Kuntz et al., 2018) were compiled and modified during field exploration and sampling and used to develop eruptive histories. Samples were collected around each shield from the distal flow margins, the flanks of the shield, and near the summit vents. Sample locations and other field information (flow types, emplacement styles, petrographic types, channels, and structural attitudes of deformed flow crusts) were compiled in ArcGIS with a digital topographic map (Fig. 2).

Whole rock major and trace element analyses were performed by x-ray fluorescence spectrometry at Brigham Young University on a Rigaku ZSX Primus II using the methods outlined in Dailey et al. (2016). Neodymium, strontium, and lead isotope ratios were analyzed by Douglas Johnson at the University of Colorado using the methods described in Fritz et al. (2006) and by Garret Hart at Washington State University following the procedure described in Ramos et al. (2004). Inductively coupled plasma- mass spectrometry (ICP-MS) analyses of several samples were conducted commercially by ALS Global.

Polished thin sections were made for several samples and were examined using a research grade Nikon Eclipse E600 petrographic microscope equipped with a digital camera. Abundance of phases were determined using a PELCON automatic point counter. Backscatter electron images of plagioclase and olivine grains were acquired on an Apreo C Low-Vacuum scanning electron microscope (SEM).

Major element compositions of major phases (olivine and plagioclase) and minor phases (Cr-spinel, magnetite, ilmenite) were found using a CAMECA SX-50 electron microprobe (updated to SX-100 standards) at Brigham Young University. Trace-element abundances were
analyzed by laser ablation inductively coupled plasma mass spectrometry (LA-ICP-MS) using an Analyte Excite excimer laser (193 nm) system by Teledyne Photon Machines attached to an Agilent 8900 ICP-MS triple quadrupole mass spectrometer at the University of Utah following techniques outlined in Dailey et al. (2016). Trace element data were processed at Brigham Young University using Iolite (Paton et al., 2011).

GEOLOGY OF ROCKY BUTTE

Rocky Butte has a broad topographic shield 36 km across that rises 100 m to a summit with less than 1° slopes (Fig. 1). The shield consists of approximately 15 km³ of lava that erupted 95 ± 10 ka (40Ar/39Ar weighted mean plateau age; using an age of 27.92 for the Taylor Creek Rhyolite; Tauxe et al., 2004). Lava emplaced through tube- and channel-fed flows filled a former canyon of the Snake River and formed the wide shield. Vesicular pahoehoe lava with multiple tumuli (both as cracked ridges and flat-topped plateaus with inflation pits and fissured margins) formed hummocky flows indicative of moderate effusion rates (e.g. Self, 1998). The flows extend about 40 km from the vent to the southwest as far as the city of Twin Falls where they are found on the north side of the Snake River Canyon (Fig. 2). Some of the outcrops on the canyon wall display distinct pillow textures. Just off Canyon Drive, Twin Falls, ID, a down-section traverse of the canyon wall consists of somewhat altered Rocky Butte compound flows above ~3 m of pillow basalt. The pillow textured layer pinches out and disappears behind a talus pile that is intersected by sub horizontally deposited basalts of a different source on the other side. These textures suggest the distal portions of the Rocky Butte lava flows entered some shallow stagnant body of water near the present Snake River Canyon. Additionally, much of the distal Rocky Butte flow field was scoured by the Bonneville Flood revealing outcrops with only thin loess cover (Othberg et al., 2012). These distal Rocky Butte lavas were originally included in the
informal “Sand Springs basalt” of Malde and Powers (1962) and Malde et al. (1963), but Othberg et al. (2012) showed that the Sand Spring basalt was erupted from multiple mostly older volcanoes and abandoned that name.

The summit of Rocky Butte is marked by a shallow crater about 25 m deep and slightly elongated to the north and south (about 970 m by 680 m; Fig. 3). The crater is almost completely surrounded by a narrow ridge (100-140 m wide at ~1375 m elevation) with undulating topography of mounds (1–8 m tall) and intervening lows (Fig. 4). On the southwest portion of the ridge a channel leads away from the summit. The high mounds on the crater rim are discontinuous vesicular blocks that contain sequences of massive lava flows (~1 m thick). The outer blocks dip away from the crater at 30°–50°. At most places along the rim, another set of discontinuous blocks runs parallel to the outermost blocks. These inner sections are a few meters away from the outer sections and dip in the opposite direction, that is into the crater.

Several meters down slope into the crater (~1365 m elevation), there is a bench of randomly oriented blocks (Fig. 3). Large mounds (~4 m tall) and smaller broken up blocks set at different angles are present on the floor of the crater. The lowest section of the crater floor (~1349 m elevation) is in the northwestern part of the crater (Fig. 3). Thin flows and shelly and rope-like pahoehoe surfaces are present in this region. Regardless of their position relative the vent, the basalt at Rocky Butte has 4–29% small phenocrysts of olivine and plagioclase set in a dark, fine-grained vesicular to diktytaxitic matrix (Sup. Tab. 1).

GEOLGY OF KIMAMMA BUTTE

Kimama Butte lavas erupted 87 ± 11 ka ([^40]Ar/[^39]Ar weighted mean plateau age using 1.196 Ma for the age of the Alder Creek Rhyolite standard; Kuntz et al., 2018) and cover an area about 250 km². From the main vent, ~10 km³ of lava distributed in a shield 13 km in diameter
and 240 m high (Fig. 1). Individual hummocky lava flows built on each other to create the broad
shield. Tube-fed flows make up the medial and distal portions of the shield and extend as far as
20 km from the vent. Tumuli developed throughout the flow area. The distribution of lavas with
unique phenocryst percentages shows the general paths of some flows from the summit of the
shield to the distal flow margins (Fig. 2). Thin deposits (< 2 m) of eolian sand and loess cover
local portions of the shield.

About 2.5 km west of Kimama Butte’s summit, the near-vent portion of an older shield
(late middle Pleistocene), Vent 4811, appears as a kipuka. Kimama Butte lavas spill into the
eastern portion of the 30 m deep crater (Kuntz et al., 2018). Similarly, about 8 km to the
northeast of Kimama Butte’s summit, an early middle Pleistocene lava cone, The Crater, rises
100 m above the distal Kimama lava flows that surround it (Kuntz et al., 2018).

Unlike Rocky Butte, Kimama Butte has a cap-like, steep-sided vent region on top of its
typical shield profile (Fig. 1; Christiansen and Hurst, 2004). The main summit vent is marked by
two adjoining sets of north-south trending ramparts instead of a central crater (about 150 m apart
with a combined length of about 500 m; Fig. 3; Fig. 5). The ramparts are as high as 50 m (the
southern ramparts being taller than the northern ramparts) and consist of inclined sheets of thin,
weakly to moderately vesicular flows ranging from 3–10 cm thick. Thicker layers are about 1 m
thick with slightly higher vesicularity but the same phenocryst percentages as the thinner
accumulations. Some sheets of thin flows dip into the crater while others dip away from the
crater. Shelly pahoehoe flows dominate the vent area. The ramparts typically contain greater than
30% phenocrysts with small glomerocrystic/cumulate inclusions about 2–3 cm in diameter
scattered throughout. The inclusions have large, almost radial, laths of plagioclase with smaller
grains of olivine (Fig. 5).
About 350 m north of the main vent area, collapse created a small pit crater 15 m deep and 150 m in diameter. The base of the southern wall of the crater is a 3–4 m thick, phenocryst-poor (3–15 vol% phenocrysts) massive basalt with increasing vesicularity up-section (Fig. 5). Near the top of this unit several smaller 10–20 cm thick flows or flow lobes exist. Above these thinner flows, more massive phenocryst-rich flows (>30 vol% phenocrysts) extend for another 3–4 m. These are similar in composition and phenocryst percentage to the ramparts at the vent to the south but lack the small plagioclase bombs. The top meter of the pit crater wall consists of thinner flows.

MINERALOGY

Mineral grains that are greater than ~0.5 mm across, non-poikilitic, and relatively euhedral were interpreted to be pre-eruptive phenocryst phases (i.e., intratelluric). These pre-eruptive phenocrysts are consistently plagioclase, olivine, and Cr-spinel (most obvious as inclusions in olivine or plagioclase). Groundmass grains of olivine, plagioclase, clinopyroxene, magnetite (subhedral grains), and ilmenite (either rod or semi-euhedral grains) are considered to be post-eruptive phases. Depending on sample location, coarse-grained flow interior or fine-grained flow exteriors, the pre-eruptive phenocrysts can be difficult to distinguish from post-eruptive phases. Often plagioclase and olivine occur in glomerocrysts with the plagioclase arranged in a radial pattern (Fig. 6). Vesicles are common especially in samples from flow exteriors. Total modal percent of phenocrysts based on point counting is between 3 and 40 vol.% on a vesicle-free basis (Fig. 7 and Sup. Tab. 1). Of the total phenocrysts, the range of plagioclase proportions is 64–87 vol. % (57–84 wt.%) and the range of olivine proportions is 13–36 vol.% (16–43 wt.%) for both Kimama Butte and Rocky Butte (Sup. Tab. 1).
In lavas from both shields, pre-eruptive olivine phenocrysts range in size from about 0.5–2.0 mm (Fig. 6). Olivine compositions are summarized in Fig. 8. Most of the phenocrysts have homogeneous centers or “cores” with thin Fe-rich rims. In the most chemically primitive lava from Kimama Butte, pre-eruptive olivine phenocrysts cores are Fo80-72 with narrow 10 μm rims that range to as low as Fo50. In the most evolved lava, olivine cores are generally Fo65-55 while the narrow rims are Fo40-25. Olivine in samples between these two end members have ranges from Fo80-55 with some groundmass grains and rims being even more Fe-rich. Rocky Butte’s most chemically primitive lava contains olivine with F80-66 with the most chemically evolved lava having F65-61. Analysis by LA-ICP-MS shows that trace element concentrations also overlap in olivine from the two volcanoes (Fig. 8). Ni, Cr, and Al decrease with decreasing Fo content (or evolution) while Zn and Ti concentrations increase with decreasing Fo content.

Many of the olivine phenocrysts have Cr-spinel inclusions either completely encased or partially enclosed in the rims of the olivine grains (Fig. 9). When only partially enclosed, the spinel is Fe-rich on the portion not surrounded by olivine. Similar zoning textures and compositions were observed by Bradshaw (2012) for basalts from the nearby Kimama core. Typically, Rocky Butte spinel inclusions are slightly more enriched in Fe3+ than Kimama Butte inclusions (Fig. 9). Most of the inclusions in the chemically evolved lavas have similar compositions to the groundmass oxides.

Typical plagioclase phenocrysts are tabular and euhedral, exhibit faint oscillatory zoning patterns (only seen in high contrast backscatter electron images; Fig. 10), occur as isolated grains or in glomerocrysts with olivine, and range from 0.5 to 5.0 mm across (Fig. 6). Some plagioclase phenocrysts contain Cr-spinel inclusions. A few plagioclase grains have sieve textured cores that
may be the result of decompression, assimilation, or magma mixing (Tsuchiyama and Takahashi, 1983; Nelson and Montana, 1992; Viccaro et al., 2010; Renjith, 2014; Lai et al., 2016).

Plagioclase compositions in the most chemically primitive lava are An\textsubscript{78-68} for Kimama Butte and An\textsubscript{72-68} for Rocky Butte. Compositions in the evolved lava vary from An\textsubscript{65-52} for Kimama Butte and An\textsubscript{63-52} for Rocky Butte (Fig. 10). The plagioclase phenocrysts show a slight variation from calcic cores to sodic rims in sample KB-01-03 (An\textsubscript{78-68} core vs An\textsubscript{74-55} rim) but analyses in the other samples display no significant differences. In general, plagioclase in evolved lavas (KB-02-03 and SS-06-03) are less calcic than the primitive lavas (KB-01-03 and SS-09-03). In all the lavas, Ti, Ba, Fe, and Sr decrease with increasing An in the plagioclase.

**Olivine-Liquid Thermometry**

Olivine compositions were paired with bulk rock analyses (taken to represent the melt composition) for each of the samples to be used in olivine-liquid thermometers (equation 22 of Putirka (2008) and Beattie (1993)). In order to produce accurate temperatures, the olivine used should be in equilibrium with the liquid. Equilibrium was tested by plotting olivine analyses on a Rhodes diagram (Sup. Fig. 1; Rhodes et al., 1979). According to the diagram, samples KB-01-03 and KB-09-03 contain Mg-rich olivine in equilibrium with the whole rock (that is the olivine crystals have an Fe-Mg exchange coefficient of 0.3 ± 0.03). KB-02-03, KB-52-04, KB-62-03, SS-02-03, SS-06-03 are not far from equilibrium especially if the Fe\textsuperscript{3+} was lower than the value assumed to calculate the Mg\# for the whole rock (Fe\textsuperscript{3+} = 5\%). An average of the olivine compositions in or close to equilibrium were used in the thermometers.

Both of the thermometers used are pressure sensitive, so pressure derived from MELTS modeling (3 kbar) was used as an input for the equations (a pressure difference of 2 kbar created an ~1°C temperature difference with the higher pressure yielding the higher temperature).
Although Putirka (2008) asserts that the Beattie (1993) model with the Herzberg and O’Hara (2002) pressure correction is best in anhydrous conditions (like those anticipated for the Snake River Plain basalts), we will consider the temperature estimates from both models.

Olivine crystallization temperatures at Kimama Butte ranged from 1238°C to 1180°C (Putirka (2008) equation 22) and 1227°C to 1166°C (Beattie, 1993; Table 1). Olivine temperatures at Rocky Butte were similar and ranged from 1202°C to 1199°C (Putirka, 2008) and 1194°C to 1179°C (Beattie, 1993). At both volcanoes and for both models, the least chemically evolved lava has higher olivine temperatures than the most evolved lava. The range of temperatures is similar for both volcanoes aside from Rocky Butte temperatures covering a smaller range than Kimama Butte (perhaps since just two samples are used that cover only the more chemically evolved portion of the whole evolutionary range).

These temperatures are slightly lower than the calculated olivine-liquid temperatures for the Kimama Drill Core lavas that range from 1255 ºC to 1200ºC (Bradshaw, 2012). They are similar to the range of olivine + plagioclase crystallization temperatures (1200°C to 1180°C at 2.8 kbar) of a compositionally similar Snake River Plain olivine-tholeiite established experimentally by Whitaker et al. (2007).

**Plagioclase-Liquid Thermometry**

Similar to the olivine equilibrium test, an Ab-An exchange coefficient of 0.27±0.05 (Putirka, 2008) was used to identify plagioclase in equilibrium with basaltic melt for temperatures greater than 1050°C. Only the primitive samples (KB-01-03 and SS-09-03) and KB-09-03 have coefficients within that range, but due to the lack of additional available data, the other samples were casually considered.
Equation 24a of Putirka (2008) was used for these basalts (Table 1). Plagioclase crystallization temperatures calculated with this model ranged from 1184°C to 1154°C at Kimama Butte. Rocky Butte temperatures were similar and ranged from 1176°C to 1152°C. If only the more primitive samples in apparent equilibrium are considered, the temperatures are 1184°C to 1173°C at Kimama Butte and 1176°C at Rocky Butte. Again, the chemically primitive lavas have higher crystallization temperatures than the evolved, but there is not a significant difference between the two volcanoes.

Overall, the plagioclase temperatures are lower than those calculated with the olivine-liquid thermometer. Bradshaw (2012) also found lower temperatures in the Kimama Drill Core with the use of the plagioclase-liquid thermometer (1200°C to 1160°C) vs the olivine-liquid thermometer (1255°C to 1200°C). The calculated temperatures here overlap with the lower end of his range. The plagioclase-liquid thermometer also matches the temperatures estimated by MELTS experiments (discussed later in the paper) that show at pressures less than 3 kbar olivine and plagioclase begin to co-crystallize between 1196°C and 1180°C.

WHOLE ROCK GEOCHEMISTRY

Major element composition of the basalts from Kimama and Rocky Butte form a single tight cluster within the range of other Snake River Plain olivine-tholeiites in the basalt field (Fig. 11) (Hughes et al., 2002; Christiansen and McCurry, 2008). Concentrations of the major and trace elements in representative samples are presented in Table 2.

The composition of lavas from Kimama Butte and Rocky Butte follow similar trends with more samples falling in the more chemically primitive end of the spectrum and fewer samples extending to the evolved side. MgO, Al2O3, and Fe2O3 show the most difference between the two volcanoes with Rocky Butte basalts having higher MgO and lower Al2O3 at a
given TiO$_2$ content (Fig. 12). Trace element concentrations are likewise similar for the two volcanoes (Fig. 13). Ni and Cr, however, stand out like MgO and Al$_2$O$_3$ in that they are systematically higher in Rocky Butte lavas than in those from Kimama Butte.

Because it is incompatible, immobile, and well-analyzed, TiO$_2$ is used as a measure of magmatic evolution with low concentrations representing more chemically primitive magmas and high concentrations representing evolved magmas. TiO$_2$ almost doubles from 2.5 to 4.4 wt.% for Kimama Butte lavas and from 2.5 to 4.2 wt.% for Rocky Butte. Thus defined, the basaltic lavas follow classic tholeiitic differentiation trends, with strong Fe-enrichment, high Fe/Mg ratios, and slight decreases in silica with differentiation (Sup. Fig. 2). SiO$_2$ decreases only over a small range from 47.5 to 45.3 wt.% at Kimama Butte, and from 47.5 to 46.6 wt.% at Rocky Butte. Other elements that decrease with evolution are Al$_2$O$_3$ (KB-16.3 to 12.7 wt.%; RB-14.9 to 13.1 wt.%) and MgO (KB-8.2 to 4.4 wt.%; RB-9.7 to 6.2 wt.%). Incompatible element concentrations such as total iron as Fe$_2$O$_3$ (KB-13.8 to 18.2 wt.%; RB-14.4 to 17.0 wt.%) and P$_2$O$_5$ (0.4 to 0.9 wt.% for both shields) increase significantly with evolution. A few elements, CaO and Na$_2$O, do not vary systematically.

Major element compositions of the lavas were used to calculate melt viscosities (Shaw, 1972). The viscosities of Kimama Butte lavas range from 7.6 to 19.2 Pa·s and viscosities of Rocky Butte lavas range from 7.0 to 14.3 Pa·s. More chemically primitive melts tend to have lower viscosities (12.0 Kimama Butte; 7.0 Rocky Butte) than the evolved melts (14.2 Kimama Butte; 14.3 Rocky Butte) because of Fe-enrichment and slight Si-depletion characteristic of tholeiitic fractionation trends.

Variations in trace element concentrations with evolution are also similar at the two shield volcanoes (Fig. 13). Like CaO, Sr varies little with evolution (KB-271 to 347 ppm; RB-
La shows the strongest enrichment varying ~2.5-fold even as TiO₂ varies by ~2-fold. Other incompatible elements increase in concentration with chemical evolution. Listed in order of greatest enrichment in Kimama Butte lavas: REE (e.g., Ce (KB- 44 to 112 ppm; RB- 41 to 85 ppm) and Eu (KB- 1.9 to 3.9 ppm; RB- 2.2 to 4.2 ppm)), Zr (KB- 219 to 429 ppm; RB- 199 to 378 ppm), Nb (KB- 20 to 40 ppm; RB- 16 to 35 ppm), Ta (KB- 1.4 to 2.4 ppm; RB- 1.0 to 1.1 ppm), Zn (KB- 104 to 167 ppm; RB- 104 to 170 ppm), Ba (KB- 262 to 639 ppm; RB- 328 to 555 ppm), V (KB- 260 to 402 ppm; RB- 220 to 316 ppm), Hf (KB- 5.3 to 9.2 ppm; RB- 4.8 to 9.1 ppm), and Sc (KB- 25 to 34 ppm; RB 26 to 34 ppm). On the other hand, concentrations of Cr (KB- 225 to 41 ppm; RB- 218 to 111 ppm) and Ni (KB- 112 to 28 ppm; RB- 147 to 68 ppm) decrease with evolution (Fig. 13). Cr decreases the most with a ~3-fold depletion in Kimama Butte lavas; MgO declines by a factor of 1.5 in the same rocks.

Trace element concentrations, shown on a primitive mantle normalized diagram (Fig. 14, Fig. 15), are comparable to other basalts from the eastern Snake River Plain (e.g., Leeman, 1982b; Jean et al., 2018). REE, along with Nb, Ta, Pb, and Zr are more enriched in chemically evolved lavas compared to the chemically primitive lavas. LREE are enriched with La/Yb ratios increasing from about 4.5 to 6.0 over the evolutionary range. There is no obvious Eu anomaly in any of the lavas. Rb/Ba ratios decrease slightly with chemical evolution (from 0.036 to 0.021) which may indicate assimilation; however, the correlation is not strong ($R^2 = 0.3883$). Zn/Fe ratios for Kimama Butte and Rocky Butte are high which indicates garnet and clinopyroxene are the dominant residual source phases (OIB with eclogite source signature e.g., Le Roux et al., 2010; Reid et al., 2017). However, Jean et al. (2018) show through low Tb/Yb ratios that the Sugar City and other Snake River Plain basalts were generated from a spinel peridotite mantle.
Leeman (1982b) concurs that Snake River Plain basaltic magmas are produced by partial melting of a spinel peridotite source.

**Sr, Nd, and Pb Isotopes**

The Sr, Nd, and Pb isotopic ratios of the lavas at these two volcanoes coincide with those for other basalts of the eastern Snake River Plain (Christiansen and McCurry, 2008 and references therein; Hanan et al., 2008; Jean et al., 2013; Jean et al., 2018; Fig. 16; Table 3). There is a small increase in $^{87}\text{Sr}/^{86}\text{Sr}$ ratio from the more chemically primitive lavas to the evolved lavas ($0.70641$ to $0.70689$ at Kimama Butte; $0.70634$ to an intermediate lava with $0.70684$ at Rocky Butte) but the range is similar at both shield volcanoes. The range in $\varepsilon\text{Nd}$ is also small with the more primitive lavas having $\varepsilon\text{Nd}$ -4.8 (Kimama Butte) and -4.4 (Rocky Butte) compared to more evolved lavas with -5.2 (Kimama Butte) and -4.4 (Rocky Butte). $^{206}\text{Pb}/^{204}\text{Pb}$ ratios are slightly higher in the primitive lavas ($18.21$ Kimama Butte; $18.14$ Rocky Butte) than in more chemically evolved lavas ($18.01$ Kimama Butte; $18.13$ Rocky Butte). On the other hand, $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$ show small but consistent differences between the shields with lower values at Kimama Butte ($^{207}\text{Pb}/^{204}\text{Pb} = 15.60$; $^{208}\text{Pb}/^{204}\text{Pb} = 38.63$-$38.51$) than at Rocky Butte ($^{207}\text{Pb}/^{204}\text{Pb} = 15.63$; $^{208}\text{Pb}/^{204}\text{Pb} = 38.72$-$38.69$) (Fig. 16).

Overall, the isotopic compositions of these lavas form a tight cluster within the more varied values of other eastern Snake River Plain basalts (Fig. 16). Like other eastern Snake River Plain basalts, those from Kimama and Rocky buttes have high $^{206}\text{Pb}/^{204}\text{Pb}$ ratios compared to those from the Yellowstone Plateau, probably as a result of interaction with younger lithosphere than the Archean lithosphere beneath Yellowstone (e.g. Hildreth et al., 1991, Jean et al., 2018). Assuming $r = 0.2$, the small differences in isotopic ratios between more chemically primitive
and evolved lavas can be explained by 50–60% fractionation and assimilation of continental crust (bulk continental crust from Taylor and McLennan (1995)).

**Magma Evolution**

Previous workers postulated fractional crystallization, assimilation, and magma recharge as the main processes controlling the evolution of the Snake River Plain olivine tholeiites (Leeman, 1982a; Kuntz et al., 1992; Hughes et al., 2002; Geist et al., 2002; Shervais et al., 2006; Putirka et al., 2009; Potter et al., 2018). To test these processes in regard to the production of Kimama Butte and Rocky Butte lavas, major element and trace element modeling techniques were used.

For both the major element and trace element models, parental compositions were chosen from the “primitive” end of the projected trend of whole rock compositions rather than from a single sample. The parental compositions are close to the most chemically primitive sample at each volcano but more accurately account for scatter in the compositional trends.

For the basalt series at Kimama Butte, a major element mass balance model based on a least-square minimization calculation in Microsoft Excel shows that removing ~38 wt.% phenocrysts \( F = 0.62 \) approximated the composition of the daughter lava with a sum of the residuals squared of 0.9. Of the total phenocrysts removed, 26 wt.% was plagioclase (An\(_{58}\)) and 12 wt.% was olivine (Fo\(_{80}\)). From the phenocryst compositions presented, the model selected the more Mg-rich olivine and the more Na-rich plagioclase analyzed in the sampled lavas. The plagioclase/olivine ratio for this model is 2.2. This mass balance model underpredicts concentrations of P\(_2\)O\(_5\) in the daughter by ~25% and K\(_2\)O and Na\(_2\)O by ~18%. Results for Rocky Butte are similar with ~39 wt.% phenocryst removal \( F = 0.61 \), a sum of the residuals squared of 0.9, and 13 wt.% olivine (Fo\(_{74}\)) removed. The model chose two plagioclase compositions to
crystallize, 12 wt.% plagioclase (An\textsubscript{70}) and 14 wt.% plagioclase (An\textsubscript{59}), suggesting that an intermediate composition may fit the model the best.

Rhyolite-MELTS (Gualda et al., 2012; Gualda and Ghiorsor, 2015) fractional crystallization trends were found by using the most Mg-rich, Ti-poor end-member and by varying oxygen fugacity, water content, and pressure. Based on the similarity of the compositional trends, these characteristics appear to be about the same at both shield volcanoes. Estimates for the $f_{O_2}$ of Snake River Plain olivine tholeiites range from -2.3 to 0 $\Delta$QFM (Leeman and Vitaliano, 1976; Whitaker et al., 2007; Brady, 2005; Bradshaw, 2012). In this study, oxygen fugacity was set at -2 $\Delta$QFM, because calculations at -1 and 0 $\Delta$QFM produced olivine compositions that were Mg-rich (as high as Fo\textsubscript{83}) compared to measured compositions (Fo\textsubscript{80-72}). Water concentrations of 0.4 to 0.05 wt.% were tested, but 0.1 wt.% produced trends that best fit the observed compositions. A plagioclase-melt hygrometer shows that the water content is very low, yielding negative water concentrations using temperature and pressure constraints as outlined now (Waters and Lange, 2015). This matches other studies that establish basalts on the Snake River Plain are essentially anhydrous (e.g. Whitaker et al., 2007; Bradshaw, 2012; Ascanio-Pellon et al., 2019) and contradicts the claims of Stefano et al. (2011) that Snake River Plain basalts are water rich. The MELTS models started at the liquidus temperature, about 1240°C, and cooled until magnetite (considered a post-eruptive phase) started crystallizing, typically about 1130°C.

Pressure was initially estimated at 4 kbar (~14 km depth) to coincide with the depth of the mid-crustal sill (Whitaker et al., 2007) where significant differentiation may have occurred when basaltic magma stalled. However, the calculated liquid lines of descent for 3 kbar provided the most satisfactory fit (~10 km depth or upper crust) to the major element compositions. Hurst
and Christiansen (2004) and Brady (2005) corroborate these pressure estimates. The experiments of Whitaker et al. (2007) on a similar Snake River Plain basalt produced comparable fractionation trends, but the starting composition was different than what was involved at Kimama Butte and Rocky Butte. Si, Al, and Ca/Al show sensitivity to pressure with the best fit resulting from the 4.3 and 2.8 kbar experiments (Whitaker et al., 2007).

Crystallization of olivine and plagioclase in the MELTS models results in $F = 0.55$ for the most evolved Fe-Ti basalts. Olivine compositions range from Fo$_{80-63}$ which is similar to the range of measured compositions; however, plagioclase range from An$_{72-60}$ which does not extend to cover the full range of measured compositions (An$_{85-50}$). MELTS models also produced 2–3 times as much plagioclase as olivine at eruption temperatures. These estimates closely match the plagioclase/olivine ratio from the simple mass balance calculations but are slightly lower than the point counting ratios.

Trace element variation was modeled using the fractional crystallization (FC) and assimilation and fractional crystallization (AFC) models in the Ersoy and Helvacı (2010) FC–AFC–FCA and mixing modeler. Partition coefficients were acquired from the GERM database (clinopyroxene and Cr-spinel) or calculated using our LA-ICP-MS analyses and whole rock compositions (olivine and plagioclase). Most trace element variation trends can be explained by 45% crystallization of olivine and plagioclase ($F = 0.55$). A mineral assemblage including 60–70% plagioclase and 30–40% olivine best explains the variation from parent to daughter. The proportion of magnetite in the fractionating assemblage is estimated to be about 2% based on the amount of Cr depletion (220 to 50 ppm) across the Kimama trend. This percent would be lower if an option for Cr enriched spinel was available.
Some element variation, like Sc, Sr and P, cannot be recreated by low pressure fractional crystallization alone. The models predicted that Sr and P would be lower than what is found in natural rocks while the Sc should be higher. Others have postulated that assimilation of a phosphorous-rich ferrogabbro causes a rapid increase in P in Snake River Plain tholeiites (Geist et al., 2002; Potter et al., 2018); however, this can only explain why the modeled P₂O₅ trend diverges from the natural rocks. Sc concentrations are under-predicted by differentiation models in several other studies of eastern Snake River Plain olivine tholeiites (e.g. Geist et al., 2002; Brady, 2005; Miller and Hughes, 2009). Based on trace element modeling, the fractionating mineral assemblage must include as much as 20% clinopyroxene to create the observed Zr-Sc trends. The role of clinopyroxene is difficult to validate because it is not a pre-eruptive phenocryst and it is not likely to crystallize in Snake River Plain olivine tholeiites at pressures less than 3 kbar as shown by experiment of Whitaker et al. (2007), Whitaker et al. (2008), our MELTS models, and as concluded by Geist et al. (2002) and White (2007). Additionally, Pearce element ratios (Fig. 17; Hiebert et al., 2008) show that there is no major-element evidence supporting clinopyroxene fractionation. Instead, they show that plagioclase-olivine fractionation explains much of the chemical variation.

In short, we see no significant differences in the chemical evolution of these two volcanoes. While there is no correlation between chemical evolution and distance from the vent area at Kimama Butte, evolved lavas are only found near the summit at Rocky Butte (Fig. 18). The near vent lavas at Kimama Butte are not geochemically evolved compared to distal lavas or to the lavas at Rocky Butte. This contradicts the conclusion of Hughes and Sakimoto (2002) and Sakimoto et al. (2003) that shield volcanoes with steep caps, like Kimama Butte, have more chemical variation than volcanoes without steep caps and that the lavas making up the steep summit region are more
evolved than distal lavas. The slight differences in trace element concentrations (e.g. Ni) between the two shields are not the result of different degrees of differentiation but instead may be the result of different degrees of partial melting or slightly different source compositions.

**MAIN CONTROLS ON SHIELD MORPHOLOGY**

Volcanic landforms, whether on Earth or elsewhere, are the result of complex interactions of many factors (e.g., Wilson, 2009; Wilson and Head, 2018). On the Snake River Plain, variations in these aspects produced two adjacent morphologically different shield volcanoes that erupted within a relatively short time of each other. As shown in the previous sections, whole rock and mineral compositions, eruptive temperatures, and melt viscosity overlap at Kimama Butte and Rocky Butte and thus are likely not main controls on shield morphology. Similarly, magma evolution is alike at the two shields.

In this section, we examine additional factors that may differ at the two shields such as magma viscosity, phenocryst abundance, and eruption style in an effort to determine which characteristics, if any, contribute the most to shield growth and the development of the distinctive morphologies of the two volcanoes. In particular, we will test the conclusions of previous studies that have linked the steeper portions of shields with changes in eruptive style (Hughes and Sakimoto, 2002; Brady, 2005) and lavas with higher crystallinity (Brady, 2005).

**Phenocryst Content**

At Kimama Butte, phenocryst content ranges from 3–39 vol.%. Three main varieties of basalt are defined by phenocryst abundance as shown in figure 7 and Sup. Tab. 1. The first type, phenocryst-poor basalt, contains less than 15 vol.% olivine and plagioclase with an average of 7 vol.% total phenocrysts. The second type of lavas at Kimama Butte are intermediate (15-30 vol.%) with plagioclase and olivine comprising an average of 19 vol.%. The plagioclase
phenocrysts are typically smaller in this type than in the phenocryst-rich variety. The third variety, phenocryst-rich basalts (greater than 30 vol.% phenocrysts), have an average of 38 vol.% large plagioclase and olivine phenocrysts. The plagioclase/olivine ratio across all these varieties is 3.8±1.5.

The phenocryst-poor basalts are the most widespread variety found in distal flows, on the flanks of the shield, and at the base of the pit crater walls (Fig. 2). Intermediate type lavas occur only on the flanks of the shield between 5–9 km from the summit. Phenocryst-rich basalts are found within 4 km of the summit. This distribution of phenocryst percentages suggests that phenocryst-poor lavas erupted first from Kimama Butte, as found at other Snake River Plain shields (Casper, 1999; Hughes et al., 2002; Brady, 2005).

Variation in crystal content at Rocky Butte is less pronounced than at Kimama Butte (4–29 vol.%). Rocky Butte lavas, like those at Kimama Butte, have about 3–4 times (by volume) more plagioclase phenocrysts than olivine (Fig. 7). The average percent of phenocrysts at Rocky Butte matches most closely with that of the intermediate lavas of Kimama Butte. Phenocryst-rich lavas, like those found near the vent at Kimama Butte, are not present at Rocky Butte. Additionally, no correlation has been found between position on the shield and phenocryst amount at Rocky Butte. At either volcano, there is no clear relationship between phenocryst content and lava composition (Sup. Fig. 3).

**Magma Viscosity**

The steep summit areas found on some shield volcanoes in the Snake River Plain may be due to an increase in lava crystallinity and hence viscosity near the vent (Sakimoto et al., 2003). Lavas with a higher phenocryst content should have higher viscosities and form shorter thicker flows given similar eruption rates, compositions, and temperatures. Magma viscosities (Fig. 18)
calculated with the Einstein-Roscoe equation (Marsh, 1981) and phenocryst proportions (Sup. Tab. 1) overlap at Kimama Butte (14 to 158 Pa·s) and Rocky Butte (8 to 75 log Pa·s). Phenocryst proportions are a much stronger control on magma viscosity than are the varied melt compositions or inferred eruptive temperatures.

Although the range is similar, the magma viscosities extend to higher values at Kimama Butte (~80 Pa·s) because phenocryst proportions are higher. The distribution of calculated magma viscosities around the vent areas and down to the flanks of the volcanoes may be significant. Crystallinity, and thus viscosity, is significantly higher at the summit of Kimama Butte (up to 158 Pa·s with 39 vol% phenocrysts) than on the flanks (as low as 14 Pa·s with 3 vol% phenocrysts). The crystallinity and viscosities of near-vent lavas at Rocky Butte (43–75 Pa·s with 26–29 vol% phenocrysts) are much less than at the summit of Kimama Butte (Fig. 18). In contrast, the distal flows from Rocky Butte and Kimama Butte have similar phenocryst proportions and low viscosities (<20 Pa·s). Thus, it appears that lava viscosity and phenocryst content may play a role in development of the morphology of these two volcanoes.

Eruption Style

Another factor to consider in the creation of different shield morphologies is differences in eruptive style both between shields and with time at a single shield volcano. Brady (2005) suggested that a viscosity related change in eruption style at Quaking Aspen Butte, another shield volcano on the eastern Snake River Plain, may have focused the fissure to a pipe-like vent. Centralized fountaining from the vent then formed a steep cap. An interpretation of field observations at these two shields suggest a similar event could have happened at Kimama Butte but not Rocky Butte.
The shallow vent crater at Rocky Butte likely developed as a large lava blister that inflated, collapsed and was then filled by a lava lake (Fig. 19). The variably tilted discontinuous vesicular blocks found around the rim and within the crater are the main evidence for a swelling or bulging blister as the tilting occurred after lava emplacement. Inflation would cause extensional fractures along the edges of the uplift. As swelling subsided and the feature deflated or collapsed, the inward dipping ramparts formed.

The consistent bimodal dip distribution of the rim ramparts at the Rocky Butte crater also developed around the smaller (about 420 m by 180 m) Lava Cascades, a perched lava lake in Craters of the Moon National Monument, Idaho (Fig. 4). The voids created by extension of the domed crust partially filled with basalt rubble and eventually with loess and vegetation. They are more obvious at the younger Craters of the Moon feature which developed in the 2.2 ka Blue Dragon flow than they are in the 95 ka Rocky Butte shield.

After the crater collapsed, a lava lake developed at Rocky Butte. The benches stepping down into the crater are interpreted as margins of a sequentially draining lava lake, and the lowest section of the crater floor is taken as the last area of lava lake retreat. Small fountains along the lake margins formed little spatter deposits seen as large mounds on the crater floor (Fig. 3). Moderate to high accumulation rates resulted in welded spatter that flowed short distances (Head and Wilson, 1989). On the southwest portion of the ridge, a channel leads away from the summit suggesting a draining event late in the shield’s development.

In contrast, a topographic break occurs at Kimama Butte ~7 km from the vent where the broad gently sloping shield steepens. The phenocryst-rich basalt variety is only present at the cap within 4 km of the vent. Short flows, interpreted as spatter-fed because of vesicle shapes and trains, make up the high ramparts near the vent region. The calculated viscosity of these spatter
fed flows are as much as 8 times higher than the earlier tube and overflow lavas (14 vs. 158 Pa·s).

The high ramparts at Kimama Butte likely formed by moderate to high rates of spatter accumulation during prolonged fountaining (Fig. 19; Head and Wilson, 1989). The southern ramparts are taller than the northern ramparts suggesting two separate fountains formed with one shorter-lived than the other. Bimodal dip distribution of sheets near the vent is likely due to where spatter landed and if it flowed back into the crater or away, although some inward dipping sheets are merely slumped blocks of what was outward dipping spatter. The pit crater north of the spatter ramparts formed after fountaining ceased as it exhibits the succession of lava found around Kimama Butte (Fig. 5), yet no new lava flowed over the rim into the crater.

The evidence of fountaining late in Kimama Butte’s history could explain the formation of the steep cap that is present there but missing from Rocky Butte. One modern example, the well-studied eruption of Mauna Ulu, Kilauea Volcano, Hawaii, illustrates how this late fountaining or changing eruption style could affect shield growth (Wright et al., 1975; Swanson et al., 1979; Wright and Tilling, 1980; Hofmann et al., 1984; Tilling et al., 1987; Vinet and Higgins, 2010).

Mauna Ulu is an appropriate analog for the eruptive activity on the Snake River Plain because eruptions occur from a vent for only a short period of time and shield edifices and lava fields grow through tube and channel-fed flows in both areas (Greeley, 1982; Swanson et al., 1979). The compositions of Mauna Ulu lavas are also similar to those on the Snake River Plain except they have lower total alkali content and higher MgO (Wright et al., 1975; Wright and Tilling, 1980; Fig. 11; Fig. 20). Average melt viscosities are alike at Mauna Ulu (12.6 Pa·s), Kimama Butte (12.6 Pa·s) and Rocky Butte (13 Pa·s). Magma viscosities are difficult to
constrain given the absence of specific phenocryst contents recorded for lavas; however, since phenocryst abundance is lower at Mauna Ulu (3–12% Byrnes et al., 2004 vs. 5–39% for Kimama Butte), we can assume that magma viscosities overlap but did not extend as high as those estimated here.

Most of the major shield growth occurred at Mauna Ulu during the first several months of the eruption (Fig. 20). There was also some growth in the last few months of the eruption but not as drastic as the beginning. Consequently, characteristics that are variable with time are not likely controls on shield development, e.g. chemical composition or evolution (MgO 9–12 wt.%) and melt and magma viscosities. Additionally, the volume of lava erupted was similar in the first part of the eruption from May 1969 to October 1971 (0.18 km$^3$; Swanson et al., 1979) when most of the growth occurred and in the second part of the eruption from February 1972 to July 1974 (0.16 km$^3$; Tilling et al., 1987).

Mauna Ulu’s changing eruption style (fountaining vs lava lake overflow) does seem to be a controlling characteristic of shield development. At the beginning of the eruption in 1969, lava fountains (50–300 m high) built a 40 m high shield over the period of ~7 months (Fig. 20). Weak spatter and lava lake overflow from the summit crater over the next 6 months caused the shield height to double to 80 m (Swanson et al., 1979). After that, shield height increased (only ~20 to 40 m) from occasional lava lake overflow until fountaining resumed in the last 7 months of the eruption. From December 1973 to January 1974 intermittent fountaining produced 15 m of new basalt on top of the summit. The final fountaining event in May 1974 increased the shield height another 4 m in about 36 hours (Tilling et al., 1987).

The broad topographic shields at Rocky and Kimama Butte were formed similarly from initial fountaining followed by lava lake overflow combined with the buildup of tube- and
channel-fed flows. After initial shield development occurred, the eruptive activity diverged at the two shields when late fountaining at Kimama Butte, like seen at Mauna Ulu, added several meters of height to the shield that is missing from Rocky Butte where no late fountaining occurred (Fig. 19).

**CONCLUSION**

Understanding Kimama Butte and Rocky Butte, two shield volcanoes on the eastern Snake River Plain of Idaho, enlarges the arsenal of methods that can be used to constrain the characteristics of volcanoes throughout the solar system that we are not able to sample or study in situ. Rocky Butte is a broad low-shield (36 km across and 100 m tall with ~1° slopes) with a summit crater; Kimama Butte is taller and has a steep-summit cap that lacks a central pit crater (18 km across and 210 m high). The vent at Rocky Butte developed late in its eruptive history as a large lava blister inflated and then collapsed, forming a crater in which a lava lake developed. Little spatter accumulated there. In contrast, high spatter mounds and spatter-fed flows flank the main summit crater at Kimama Butte. Chemical composition and evolution, eruption temperatures, magma viscosity, and eruption styles were assessed to determine their relationship to the distinctive topography and summit characteristics of these two Quaternary low-shield volcanoes.

Major and trace-element variation as well as mineral assemblages and compositions are similar at the two shields. Major and trace element models (including MELTS) suggest that lavas from both shield have evolved by fractional crystallization at 3 kbar (~10 m depth). Pre-eruptive phenocrysts show variation with evolution with more primitive olivine being more Mg-rich and more primitive plagioclase being more calcic (Fo_{80-72} vs Fo_{65-55}; An_{78-68} vs An_{65-52}). Eruptive temperatures overlap at Kimama Butte (1238–1154°C) and Rocky Butte (1202–1152°C). The
similarity of these factors at the two volcanoes suggests that they are probably not important controls on shield volcano morphology. Thus, it would be difficult to use shield profiles to speculate about magma composition differences or differences in the degree of magma evolution for similar shield volcanoes on other planetary bodies as suggested by Hughes and Sakimoto (2002) and Sakimoto et al. (2002).

Shield morphology does seem to be related to phenocryst content (and therefore magma viscosity) and eruption style. Kimama Butte lavas contain 3–39 vol.% phenocrysts while Rocky Butte lavas contain 4–29 vol.%. Magma viscosities overlap at Kimama Butte (158–14 Pa·s) and Rocky Butte (75–43 Pa·s); however, Kimama Butte magma viscosities extend to double the value of the maximum Rocky Butte viscosities. Additionally, the high magma viscosities (up to 158 Pa·s with 39 vol% phenocrysts) at Kimama Butte are confined to the steep summit cap of the shield.

A reasonable history of shield growth at Kimama Butte and Rocky Butte involves a similar initial period of fountaining, lava lake overflow, and tube- and channel-fed flows. Late fountaining at Kimama Butte deposited the phenocryst-rich (>30 vol.%), higher viscosity lavas and high spatter ramparts that form the cap. Rocky Butte did not experience an episode of late fountaining, so no steep cap developed there. Thus, eruption style, viscosity, and phenocryst content may be reasonably estimated from the morphology of volcanic features on other planetary bodies, but composition and eruption temperature cannot.

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Figure 1. Rocky Butte is a low-profile shield while Kimama Butte has a steep ‘cap’ near the summit. 15:1 vertical exaggeration.
Figure 2. A. Simplified geologic map showing the location of volcanic rocks in the northwestern United States. Ages of the rhyolitic volcanic centers and calderas show the progression of the Yellowstone hotspot from underneath the Nevada, Oregon, Idaho border to present-day Yellowstone National Park. Map modified from Schruben et al., 1994. B. Bedrock geologic map of the Kimama Butte and Rocky Butte lava fields. Sample locations are labeled and phenocryst content is shown for Kimama Butte lavas. The Kimama drill core site is shown by the black star (Potter et al., 2018). Modified from Malde et al. (1936), Kuntz et al. (2018), and Othburg et al. (2012).
Figure # “Summits.” A. Colored shaded relief map showing the Rocky Butte and Kimama Butte lava fields. B. Summit of Rocky Butte. Strike and dip of tilted lava blocks are shown. Spatter accumulations are shown by yellow triangles. The last location of the lava lake is outlined in black. C. Summit of Kimama Butte. Lava types are designated by colored circles.
Figure 4. The summit of Rocky Butte is broad and shallow (25 m deep). A. The summit pit crater at Rocky Butte. B. The lava lake rim down to the lava lake floor. Several distinct benches can be seen at different elevations on the rim and are attributed to the lava blister swelling and then collapsing. C. A steeply dipping spatter rampart on the lava lake floor. D. A section of the rim of the caldera created by the collapse of the blister thus the differing slope directions. E. The rim of the Lava Cascades lava lake, Craters of the Moon, Idaho.
Figure 5. A. The summit crater showing two sets of high ramparts, one in the foreground and one in the background. The small pit crater is in the distance. B. Large diktytaxitic glomerocryst consisting of large laths of plagioclase with smaller olivine grains found in the spatter rampart near the summit. C. Thin spatter-fed lava flows make up the ramparts at the summit. D. The wall of the pit crater. Lavas at the bottom of the wall are phenocryst-poor (3-4 m) while lavas in the top half are intermediate or phenocryst-rich.
Figure 6. Photomicrographs of representative Kimama Butte and Rocky Butte lavas (3 mm across in cross polarized light). Plagioclase and olivine phenocrysts set in a groundmass of varying grain size. Intermediate type lavas are shown by KB-02-03 (19 vol% phenocrysts). Phenocryst-poor type lavas are represented by KB-01-03 (3 vol% phenocrysts). Sample SS-06-03 has 26 vol% phenocrysts and SS-09-03 has 4 vol% phenocrysts. Some vesicles are marked by a “V.”
Figure 7. Kimama Butte lavas are classified as phenocryst-rich, intermediate, or phenocryst-poor depending on the total phenocryst amount. Rocky Butte lavas coincide with the intermediate and phenocryst-poor Kimama Butte percentages.
Figure 8. Olivine compositions overlap in Kimama Butte (A) and Rocky Butte (B) lavas. The gray field represents all the data collected while the colored bars show the end member analyses. C. Gray field shows density of 90% of electron microprobe data. Dots are for LA-ICP-MS analyses. D. BSE image of SS-06-03_o11. Contrast is high to show the bright, Fe-rich rim more clearly.
Figure 9. A. Spinel inclusions are visible in this BSE image of a typical olivine grain (light gray, subhedral). Note the bright Fe-rich rims where the grains are in contact with the matrix. Also pictured are tabular ilmenite grains and euhedral magnetite grains in the groundmass. B. A ternary plot of Cr, Fe\(^{3+}\), and Al for opaque oxides in Kimama Butte (red symbols) and Rocky Butte (blue symbols). Open symbols indicate groundmass grains while closed symbols are inclusions in olivine. The arrows show change in composition from core to rim for two inclusions that are only partially enclosed in the olivine grain.
Figure 10. Plagioclase composition also overlap for Kimama Butte (A) and Rocky Butte (B) lavas. The gray field represents all the data collected while the colored bars show the end member analyses. C and D. BSE images of two oscillatory zoned plagioclase grains.
Figure 11. TAS diagram comparing Kimama Butte and Rocky Butte compositions with other basaltic lavas from the Snake River Plain (shown by the blue field). Data from the 1969-1974 Mauna Ulu eruption are also included from Wright et al. (1975) and Wright & Tilling, (1980).
Figure 12. Major element concentrations are similar in lavas at Kimama Butte (red) and Rocky Butte (blue).
Figure 13. Trace element concentrations are similar in Kimama Butte (red) and Rocky Butte (blue lavas).
Figure 14. Normalized trace element diagrams for lavas from Kimama Butte and Rocky Butte. Gray fields signify distribution of data while the most (KB-02-03; SS-06-03) and least (KB-01-03; SS-09-03) chemically evolved samples from each volcano are shown specifically.
Figure 15. Rare earth element patterns for lavas from Kimama Butte and Rocky Butte. Gray fields signify distribution of data while the most (KB-02-03, SS-06-03 with elevated REE) and least (KB-01-03, SS-09-03 with low REE) chemically evolved samples are shown specifically.
Figure 16. Kimama Butte and Rocky Butte Sr-Nd-Pb isotopic compositions compared against other basalts from the Snake River Plain and Yellowstone shown by the blue field. Data from Christiansen and McCurry (2008) and Jean et al. (2018).
Figure 17. Pearce element ratio diagrams show that major element variation is controlled by the fractionation of olivine and plagioclase (A) and not by clinopyroxene (B).
Figure 18. TiO$_2$, MgO, and calculated viscosity as a function of distance from the summit at the two volcanoes. Magma viscosities were determined using the Einstein and Roscoe equation as found in Marsh (1981). Phenocryst percentages were determined by point counting. Magma viscosities at the summit of Kimama Butte are higher and more variable than at Rocky Butte. Some of the low viscosities near the vent at Kimama Butte are from the base of the pit crater and thus represent earlier flows.
Figure 19. A schematic cross section of Kimama Butte and Rocky Butte shows how the different morphologies were developed. The vent crater at Rocky Butte developed as a large lava blister that inflated and then collapsed forming a crater in which a lava lake developed. In contrast, high spatter mounds and spatter-fed phenocryst-rich flows flank the main summit crater at Kimama Butte.
Figure 20. A. Mauna Ulu shield profile with the profiles of Kimama Butte and Rocky Butte. B. Comparison of the growth of Mauna Ulu shield and chemical evolution as the eruption progressed from May 1969 to July 1974. The eruption occurred in two periods with a hiatus of activity from October 1971 to February 1972. The shaded background denotes the main eruptive activity that was occurring during that time interval.
Table 1. Summary of temperature for Kimama Butte and Rocky Butte lavas

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**Note:** Thermometers from Putirka (2008)

Bolded values indicate samples with phenocrysts in or near equilibrium

Plagioclase are considered in equilibrium if the following condition is met:
for T>1050°C Kd(Ab-An) = 0.27±0.05

Olivine are considered in equilibrium if Kd(Fe-Mg) is within 0.3±0.03
Table 2. Major and trace elements compositions of Kimama Butte and Rocky Butte basalts

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Table 2 continued. Major and trace elements compositions of Kimama Butte and Rocky Butte basalts

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Note: xx Sample KB-52-04 missing by 2019

*Indicates trace element analyses by LA-ICP-MS as opposed to XRF
Table 3. Isotopic compositions of representative Kimama Butte and Rocky Butte lavas

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<th>(\varepsilon\text{Nd})</th>
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| Lab                           | Reference  | Sample | \(^{206}\text{Pb}/^{204}\text{Pb}_m\) | \(^{207}\text{Pb}/^{204}\text{Pb}_m\) | \(^{208}\text{Pb}/^{204}\text{Pb}_m\) |
|-------------------------------|------------|--------|---------------------------------|----------------|----------------|----------------|
| University of Colorado        | Doug       | KB-01-03 | 18.210                        | 15.600          | 38.630          |
| University of Colorado        | Doug       | KB-02-03 | 18.010                        | 15.600          | 38.510          |
| Washington State Univ         | Garret     | KB-09-03 | 18.007                        | 15.613          | 38.566          |
| Washington State Univ         | Garret     | KB-14-03 | 18.006                        | 15.614          | 38.559          |
| Washington State Univ         | Garret     | SS-02-03 | 18.126                        | 15.632          | 38.720          |
| Washington State Univ         | Garret     | SS-09-03 | 18.135                        | 15.627          | 38.639          |

Note: subscript "m" indicates a measured value
      subscript "i" indicates a calculated initial value