

Tectonic Implications of Mantle Source Changes recorded by Eocene-Oligocene Basalts from
The Dillon Volcanic Field Southwest Montana (USA)
by

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Abstract

Basalts of the Eocene to Oligo-Miocene Dillon volcanic field Montana (U.S.A.) record the change from dominantly compression to extensional tectonics, within the northern Rocky Mountains and accordingly, also present across the North American Cordillera. The Dillon volcanic field is composed of lavas and intrusives that crop out across southwestern Montana, but most prominently in the Gravelly Range, where they define multiple eruptive centers (e.g., Black Butte, Lion Mountain, etc.). Prior work suggested lower Dillon volcanism-initiated ca. ~50-39 Ma, which occurred adjacent to coevally with the 51-40 Ma central Idaho Challis and the 54-44 Ma Absaroka volcanism. Multiple tectonic models have been proposed to explain the ~33-16 Ma middle and upper Dillon volcanism (e.g., Farallon slab rollback, slab tearing/foundering, lithospheric drip), which erupted during the later period of regional extension and basin sedimentation.

Dillon volcanic rocks have mainly high-K compositions, are primarily subalkaline, but include eruptions of small volume alkali basalts. The lower Dillon rocks tend to range from basaltic trachyandesites to basaltic andesites. Middle/upper Dillon rocks include basanite, trachybasalt, basalt, basaltic andesites, and basaltic trachyandesites. Basalts of the Dillon volcanics are often olivine rich with groundmass of clinopyroxene and plagioclase; however, middle/upper samples tend to contain titanite, reflecting their alkali compositions. Lavas, especially dikes and sills in the Gravelly Range, often contain small ~1-2 cm-wide crustal xenoliths, such as Eocene Renova formation sediment. Dikes and sills cut Renova strata, and some Dillon lavas overlying Renova sediments. These xenoliths, along with microscopic quartz xenocrysts, are evidence of crustal interaction, though assimilation-fractional crystallization modelling shows that this interaction had a small effect on the basalt bulk rock chemistry and

their Sr-Nd isotope-compositions. Trace element and Sr-Nd isotope ratios of the lower Dillon volcanics display arc-like signatures, which we suggest reflect subduction of the Farallon plate (e.g., Ta/Th <0.2, $^{87}\text{Sr}/^{86}\text{Sr}_i$ 0.7072 - 0.7078; $^{143}\text{Nd}/^{144}\text{Nd}_i$ 0.51204 - 0.51229). Middle and upper Dillon rocks have intermediate Ta/Th (0.2-0.6) coupled with Sr-Nd isotopes ($^{87}\text{Sr}/^{86}\text{Sr}$ 0.7059 - 0.7043; $^{143}\text{Nd}/^{144}\text{Nd}_i$ 0.51204 - 0.51236) suggesting magma sources from metasomatized lithosphere mantle. Some middle/upper Dillon rocks have high Ta/Th (>0.6) with low $^{87}\text{Sr}/^{86}\text{Sr}$ (0.7043-0.7049) which might suggest an asthenosphere source, however, these rocks have lower, and thus more enriched, $^{143}\text{Nd}/^{144}\text{Nd}_i$ (0.51203-0.51232), unlike any other documented regional Cenozoic volcanics and we suggest this likely reflects lower crustal interaction.

We propose that the middle Dillon volcanics are sourced from metasomatized subduction-affected lithospheric mantle that melted after the initiation of slab tearing/rollback and upwelling asthenosphere. Eruptions of middle Dillon basalts occurred contemporaneously with the onset of regional extensional tectonics and Renova basinal strata deposition. Continued extension resulted in eruptions of the upper Dillon alkali basalts, which we show were sourced from asthenosphere that interacted with low crust (or possibly lithospheric mantle melts) and resulted in whole rock $^{87}\text{Sr}/^{86}\text{Sr}_i$ and $^{143}\text{Nd}/^{144}\text{Nd}_i$ ratios that resemble enriched mantle (e.g., EM1-like).

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Chapter 1 - Introduction

Basalt geochemistry provides insights into the petrogenesis of mafic magmatism and in combination with modeling and geological constraints, it can elucidate aspects of the regional tectonic setting associated with the generation of mafic mantle melts.

Mafic rocks (basalts) are either sourced from lithospheric mantle or deeper asthenosphere. Basaltic magmas are relatively reliable for constraining the tectonic settings they form within due to their distinctive geochemical characteristics (Xia and Li 2019). Different tectonic settings (e.g., subduction, extension) and processes (e.g., metasomatism of continental lithospheric mantle from subducting oceanic lithosphere, decompression melting of upwelling asthenosphere mantle in extensional settings, flux-melting of asthenosphere in active subduction) impose distinctive geochemical and isotope signatures on the basaltic magmas produced. (Xia and Li 2019; Farmer et al., 2020; Farmer et al., 2024).

Moreover, basalt whole rock incompatible element ratios and isotopic compositions can reveal whether they were sourced from lithospheric mantle and/or the asthenosphere and indicate crustal interaction, as well as record tectonomagmatic changes through time (Fitton et al., 1991; Xia and Li 2019; Farmer et al., 2020; Farmer et al., 2024). Trace element ratios and isotope analyses (e.g., Sr-Nd) are also useful in determining the effects of crustal assimilation, because contamination of melts may occur when magma ascends through thick continental crust.

For example, in north-central Colorado (USA), chemical and isotopic analyses of Late Cretaceous and post-40 Ma volcanic rocks demonstrate compositional variations over geologic time that are interpreted to reflect changing mantle sources (Farmer et al., 2024). During the Cretaceous, low (<0.2) Ta/Th magmas were sourced from subduction-affected asthenosphere associated with flux melting (e.g., “typical” continental arc volcanism) above a shallow dipping

subducting Farallon slab (Farmer et al., 2024). Regional foundering/roll-back of the Farallon slab marked the ending of compressional tectonics and the transition to post-40 Ma extensional tectonics (Crow et al., 2011; Abbot et al., 2022; Farmer et al., 2024). This resulted in asthenosphere upwelling and melting of lithospheric mantle, which is indicated by intermediate (0.2-0.6) Ta/Th mafic melts sourced from metasomatized lithospheric mantle (Farmer et al., 2020; Farmer et al., 2024). In locations where extension is higher and continental lithosphere has undergone great degree of thinning, as in the Basin and Range of the western USA, decompression melting of upwelling asthenosphere occurs resulting in high (>0.6 , ~ 1) Ta/Th basalt magmatism (Farmer et al., 2024).

A similar sequence of changes is recorded in western Canada, where subduction was the dominant tectonic phenomena in the Eocene until the Northern Cordilleran slab window developed in the early Miocene, when asthenospheric mantle displaced the hydrated mantle wedge (Thorkelson et al., 2011). As a result, Eocene magmatism in western Canada, which is the product of subduction tectonics and associated mantle metasomatism, resulted in arc-like incompatible trace element ratios ($\text{La/Nb} > 2:1$, $\text{Nb/Zr} < 0.15$) (Thorkelson et al., 2011). Tectonic conditions changed, however, resulting in a hiatus in magmatism during the Oligocene until intraplate magmatism occurred in the late Oligocene to early Miocene in response to the opening of a slab window (Thorkelson et al., 2011). Miocene mafic igneous rocks here show a compositional change from arc-like geochemistry to OIB-like incompatible trace element compositions ($\text{La/Nb} < 1:1$, $\text{Nb/Zr} > 0.15$) as magma sources changed from slab-metasomatized mantle to upwelling asthenosphere (Thorkelson et al., 2011).

Understanding the petrogenesis of Eocene, Oligocene, and Miocene magmatism in the northern Rocky Mountains (USA and Canada) is vital for the creation of tectonic models that

accurately link the generation of mantle melts and location of volcanism, with their tectonic setting, especially during the period of time just after Laramide compression. Methods used to determine the tectonomagmatic setting of these regions can, thus, be used to determine tectonic models and evolution paths of less known or debated magmatism and volcanism.

Southwestern Montana and the northern Rockies/Pacific Northwest (USA) (Fig. 1) were affected by the subduction of Farallon plate, ~51-43 Ma Challis and Absaroka magmatism (e.g., onset of the ignimbrite flare-up and coincident upper-plate extension) and the subsequent collision of the Siletzia oceanic plateau ca. ~50-45 Ma (Constenius, 1996; Dostal et al., 2021; Donaghy et al., 2024). Between the Challis and Absaroka volcanic fields in southwestern Montana (USA) lies the Eocene to Miocene Dillon volcanic field (DVF) and associated Dillon volcanics (Fig. 1). We use the nomenclature of Fritz et al. (2007) who grouped all the ~51-16 Ma volcanic rocks in southwestern Montana that were not previously mapped as either Absaroka or Challis volcanics, as the Dillon volcanics. Fritz et al. (2007) also defined the overall areal extent of these rocks as the “Dillon volcanic field” (Fig.1), whereas Chadwick (1985) divided DVF rocks into geographically restricted volcanic fields under the broad title “Miscellaneous Montana Fields” (e.g., Virginia City volcanic field; Gravelly Range volcanic field, etc.).

The Dillon volcanics consist of dispersed lavas, intrusions, and volcanic centers (Fig. 1; Fritz et al., 2007). The oldest of the Dillon volcanics, the lower member (54-39 Ma), erupted contemporaneously with the 50-45 Ma Challis volcanic field to the west in central Idaho, and the 53-43 Ma Absaroka volcanic field to the east and southeast in Montana, Wyoming, and eastern Idaho (Fritz et al., 2007). Unlike the Challis and Absaroka fields, the DVF remained active following an ~6 m.y. lull between the lower and middle Dillon members (Fritz et al., 2007;

Mosolf, 2021). Dillon volcanism resumed with eruptions of the middle Dillon (33-27 Ma) and upper (27-16 Ma) Dillon members (Fritz et al., 2007; Broley et al., 2023).

While magmatism of the lower Dillon (54-39 Ma) is thought to be associated with subduction of the Farallon plate (Fritz et al., 2007), and may be an equivalent to central Idaho Challis magmatism, the tectonomagmatic origins of the middle and upper Dillon magmatic episodes are not as clear. Tectonic models have been proposed and are debated to explain the overall prolonged period of DVF magmatism/volcanism i.e., extensional magmatism for the middle and upper DVF, lithospheric drip middle DVF (Fritz et al., 2007; Li et al., 2021).

Previous studies of the DVF rely on a limited suite of bulk rock major and trace element data. This includes a regional study by Fritz et al. (2007), which included both whole rock geochemistry and K-Ar dating of DVF rocks, mapping and whole rock major and trace element analyses of 12 mafic lavas near Virginia City, MT, by Mosolf (2021), as well as a few M.S thesis from the 1980-90's reporting limited bulk rock major element analyses and a few K-Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ ages (Gutmann et al., 1985; Luikart, 1997; and references within). In all the published works, there are few samples from the youngest upper Dillon member and, notably, none include isotope data for any DVF rocks. While major and trace-element data are important for classification and petrologic/tectonic modeling, they have their limitations when distinguishing between lithospheric and asthenospheric melt origins. Basalt melts from both the lithospheric and asthenospheric mantle may result in overlapping incompatible trace element ratios, making it difficult to differentiate between sources. Fractionation of magma and crustal contamination can further alter the original geochemical signature of DVF rocks.

In this study, we combine DVF petrography, whole rock major and trace-element geochemistry and $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ isotope ratios to more accurately assess the origins of

magmas in the volcanic province, where the DVF rocks likely record the tectonic transition from compression to extension. For example, Ta/Th values between 0.2 and 0.6 in DVF basalts, coupled with high $^{87}\text{Sr}/^{86}\text{Sr}$ and low $^{143}\text{Nd}/^{144}\text{Nd}$ ($^{87}\text{Sr}/^{86}\text{Sr} > 0.706$ and $^{143}\text{Nd}/^{144}\text{Nd} < 0.5126$), would be typical of ancient mantle previously affected by subduction (e.g., WY craton lithosphere), while higher Ta/Th (> 0.6), coupled with lower $^{87}\text{Sr}/^{86}\text{Sr}$ values (< 0.7045 and $^{143}\text{Nd}/^{144}\text{Nd} > 0.5126$), suggest an asthenospheric source (Farmer et al., 2020). We can then determine if the upper, middle and lower DVF basalts originated in the lithospheric mantle or asthenosphere mantle and if the sources of these magmas changed with time reflecting different tectonic conditions between ~54 and 16 Ma (e.g., subduction changing to lithosphere drip vs. subduction changing to extension and upwelling of depleted OIB-MORB like asthenosphere). Whole rock $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ isotope ratios in basalts are less affected by fractional crystallization and partial melting, allowing greater clarity in identifying melts originating from the deeper asthenosphere or the more enriched continental lithospheric mantle. Sr-Nd isotope values can be affected by crustal contamination but still are useful in evaluating if DVF melts assimilated mantle and/or crustal materials by recognizing physical evidence (e.g., xenoliths and/or xenocrysts) and non-primary incompatible trace element ratios. These data enhance interpretations of mantle sources in context of the proposed tectonic models (e.g., lithospheric drip or slab rollback/slab window).

This study refines our understanding of how North America, more specifically the northern Rocky Mountains region (USA), responded to the cessation of early Cenozoic subduction-related tectonics and a transition into inboard extension during the Eocene to Oligo-Miocene (where inboard refers to inboard from the Cascades arc).

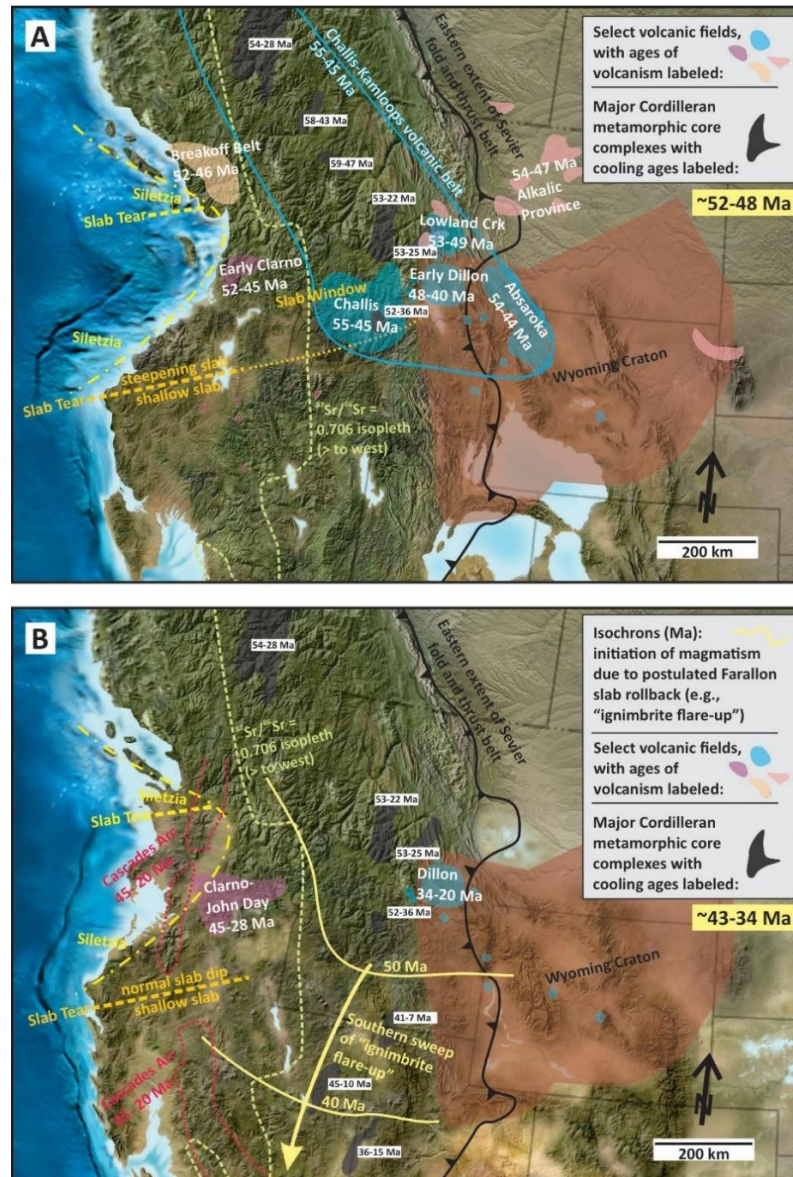


Figure 1: Paleogeographic map of the northern Rockies and Pacific northwest (U.S.A.) and southern Canada at ~52-34 Ma. Also depicted are the Absaroka, Challis, and Lowland Creek volcanic provinces and ages when active (Norman & Mertzman, 1991; McKerverey, 1998; Hiza, 1999; Feeley et al., 2002; Lindsay & Feeley, 2003; Dudás et al., 2010), early volcanism in the Dillon volcanic field (Fritz et al., 2007); Siletzia collision induced “Breakoff-belt” magmatism (Tepper et al., 2023), Challis-Kamloops volcanic belt (Dostal and Jutras, 2021), Montana/Great Plains alkalic province (Dudás and Harlan, 1999; Duke, 2009), early Clarno volcanism (Appel, 2002; Bestland et al., 2002), Cascades and diffuse back-arc volcanism (John et al., 2012; Camp and Wells, 2021), Wyoming craton extent (Bedrosian and Frost, 2023), Cordilleran core complexes and their cooling ages (Howlett et al., 2021), the inferred extent of Siletzia (Schmandt and Humphreys, 2010), northern slab tear in Siletzia (Anderson et al., 2024), and slab tear along southern margin of Siletzia, separating steep-dipping, from shallow-dipping slab segments (Humphreys, 2009). Basemap © 2013 Colorado Plateau Geosystems Inc.

Background

The oldest rocks in the study area are the Archean basement of the Wyoming Craton. Dating from 3.3 to 2.5 Ga, they underlie Wyoming, Montana, and southeastern Idaho (Mogk et al., 2020; Bedrosian and Frost, 2023) (Fig. 2). The DVF is located within and near the western edge of the Wyoming Craton boundary in Montana. West of the craton, near the Tobacco Root Mountains, lies the Selway terrane (~2.5-1.6 Ga) and the northern boundary of the study area is the Medicine Hat Block (~2.5-3.3 Ga) near Butte, Montana (Mogk et al., 2023). The Wyoming Craton is surrounded by collisional Paleoproterozoic orogenic belts, such as the Great Falls Tectonic Zone to the northwest just beyond the boundary of the Wyoming Craton and Medicine Hat Block, and the Farmington Zone to the west at the boundary of the Selway and Wyoming Craton (Mogk et al., 2020). The Wyoming Craton is partially overlain by Mesoproterozoic Belt Basin deposits and is mostly covered by Phanerozoic sedimentary strata (Mogk et al., 2020), but exposed Precambrian craton crops out in the Tobacco Root Mtns, Ruby Range, Greenhorn Mtns, and Blacktail Mtns, as well as the Gravelly Range, which consist mainly of quartzofeldspathic gneisses (Harms and Baldwin, 2022; Mogk et al., 2020).

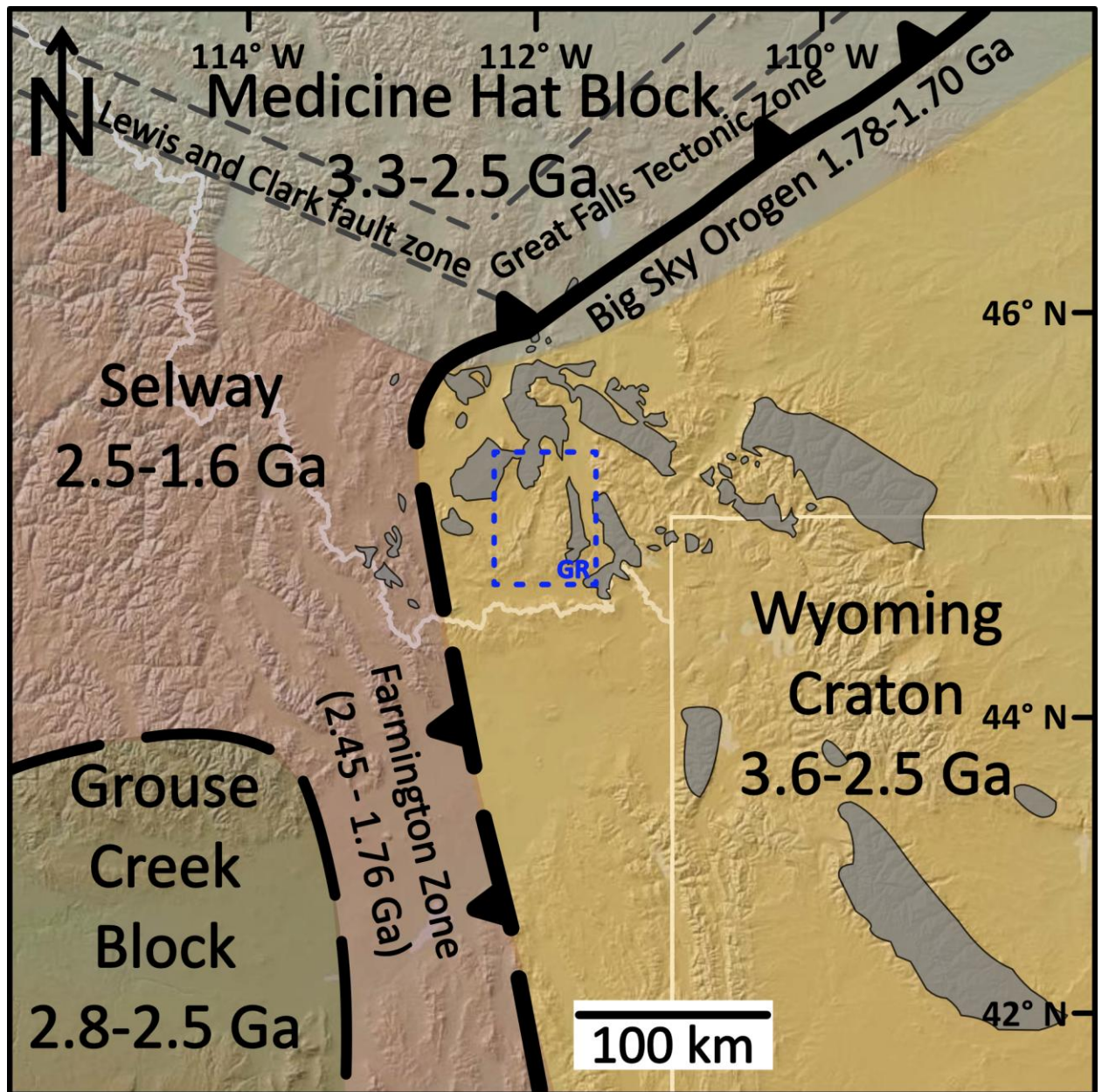


Figure 2: Regional tectonic index map of the Archean Wyoming Province (Wyoming Craton light orange; dark grey areas are exposures of Archean crystalline rocks). Two Paleoproterozoic orogenic belts are shown which surround the Wyoming Craton, The Great Falls Tectonic Zone and the Farmington Zone. Dash blue denotes Gravelly Range study area (GR). Figure based on (Mogk et al., 2020)

During the Late Cretaceous into Eocene, the Wyoming Craton experienced significant uplift due to the collision and likely shallow subduction of the Farallon plate, as well as possibly

one to two oceanic plateaus, and/or preceded by the Insular Terrane ca. ~100 to 85 Ma (e.g., conjugate Shatsky and Hess rises) (Humphreys, 2009; Tikoff et al., 2023). These orogenic events resulted in Sevier and Laramide crustal thickening (Humphreys 2009; Scarberry et al., 2022), respectively. The Laramide orogeny is mostly attributed to a decrease in the subduction angle of the Farallon slab, which enhanced mechanical coupling of the Farallon plate and the overriding North American plate (Aper et al., 2024). The Laramide orogeny affected a large region of the western North American plate from New Mexico to Canada and exhumed some Archean basement, including within the Wyoming Craton (Lageson et al., 2022). Laramide exhumation began at ~100 Ma in southwestern Montana and was well established by ca. 80 Ma, which is earlier in comparison to other parts of the Laramide foreland in southern regions (Carrapa et al., 2019).

Farallon plate subduction-initiated arc magmatism in Montana occurred from 120 to 66 Ma, resulting in the emplacement of batholiths, volcanic fields, and mixed volcanic sedimentary deposits (Scarberry et al., 2022). Magmatism continued, along with the additional collision of the Siletzia oceanic plateau onto the western North American margin, coupled with widespread extension, resulting in ~51-43 Ma Challis and Absaroka magmatism with coeval lower Dillon volcanism (~54-39 Ma) (Constenius, 1996; Fritz et al., 2007; Schmandt and Humphreys, 2010; Dostal et al., 2021; Tepper et al., 2023). Following the collision and accretion of Siletzia terrane at ca. 50 Ma, the breakoff and rollback of the Farallon slab initiated (Tepper et al., 2023). This triggered southwest younging magmatism, the Mid-Tertiary ignimbrite flare-up, which swept southward from Challis region, through the northern and then into the central Great Basin (Humphreys, 2009; Best et al., 2016). Currently proposed regional Eocene-Oligocene tectonic models for the Montana, Idaho and Wyoming tristate region are centered on rollback of the

Farallon slab after shallow subduction resulting in extension (Fritz et al., 2007; Smith et al., 2014; Cassel et al., 2018; Li et al., 2021).

Eruptions of the 54-39 Ma lower Dillon volcanics are almost exclusively located in southwestern Montana (Fritz et al., 2007). The Absaroka Volcanics (53-43 Ma) are found in a northwest-trending belt consisting mainly of stratovolcanoes and primarily located in northwestern Wyoming but also extending into southern Montana and small portions of eastern Idaho (Lindsay and Feeley, 2003; Dudás et al., 2023). The central Idaho Challis volcanic field erupted primarily from fissures and calderas ca. ~50-45 Ma (Norman and Mertzman, 1991). Lower DVF magmatism was proposed by Fritz et al. (2007) to reflect continental-arc origins based on their potassic, alkaline to subalkaline magma affinities and their enrichment in light rare earth elements (LREEs) and large-ion lithophile elements (LILEs), with notable Ta and Nb depletion.

Compressional forces transitioned into extension, marking the end of the Laramide Orogeny, at ~54-51 Ma south of Montana in the Idaho-Utah-Wyoming region and at ~55-50 Ma north of 42°N (Wells et al., 2012). This tectonic event resulted in the Eocene-early Miocene (ca. 49-20 Ma) extensional gravitational collapse of the over-thickened Cordilleran foreland fold-thrust belt, removal of the Farallon slab, formation of Paleogene extensional basins, development of metamorphic core complexes (for example, the ~53 Ma Anaconda metamorphic core complex; Fig. 3), and regional intraplate magmatism resulting from lithospheric thinning and asthenosphere upwelling (Constenius, 1996; Wells et al., 2012; Vuke, 2022). The Renova Formation, a sedimentary unit accumulated locally during extensional basin development in southwestern Montana, is the first record of widespread sedimentation after the end of Sevier and Laramide orogenies in southwest Montana. Renova strata consist of volcanoclastic sediments,

epiclastic sediments and interbedded lavas (Fritz et al., 2007; Schwartz and Graham, 2017; Vuke 2022). Renova Formation deposition records initial subsidence followed by the initiation of regional extension ca. ~49 Ma to 19 Ma (Thoresen et al., 2024). In the Gravelly Range, the Renova sedimentary package is found primarily in a north-south belt along the mountain range crest. It is ~200-250 m-thick and either directly underlies DVF lavas or is cut by DVF dikes and intrusive bodies (Luikart, 1997; this study). The locally exposed Renova formation also hosts an ~3-5 Ma disconformity, where the ~37-33 Ma sediment package is missing (Luikart, 1997).

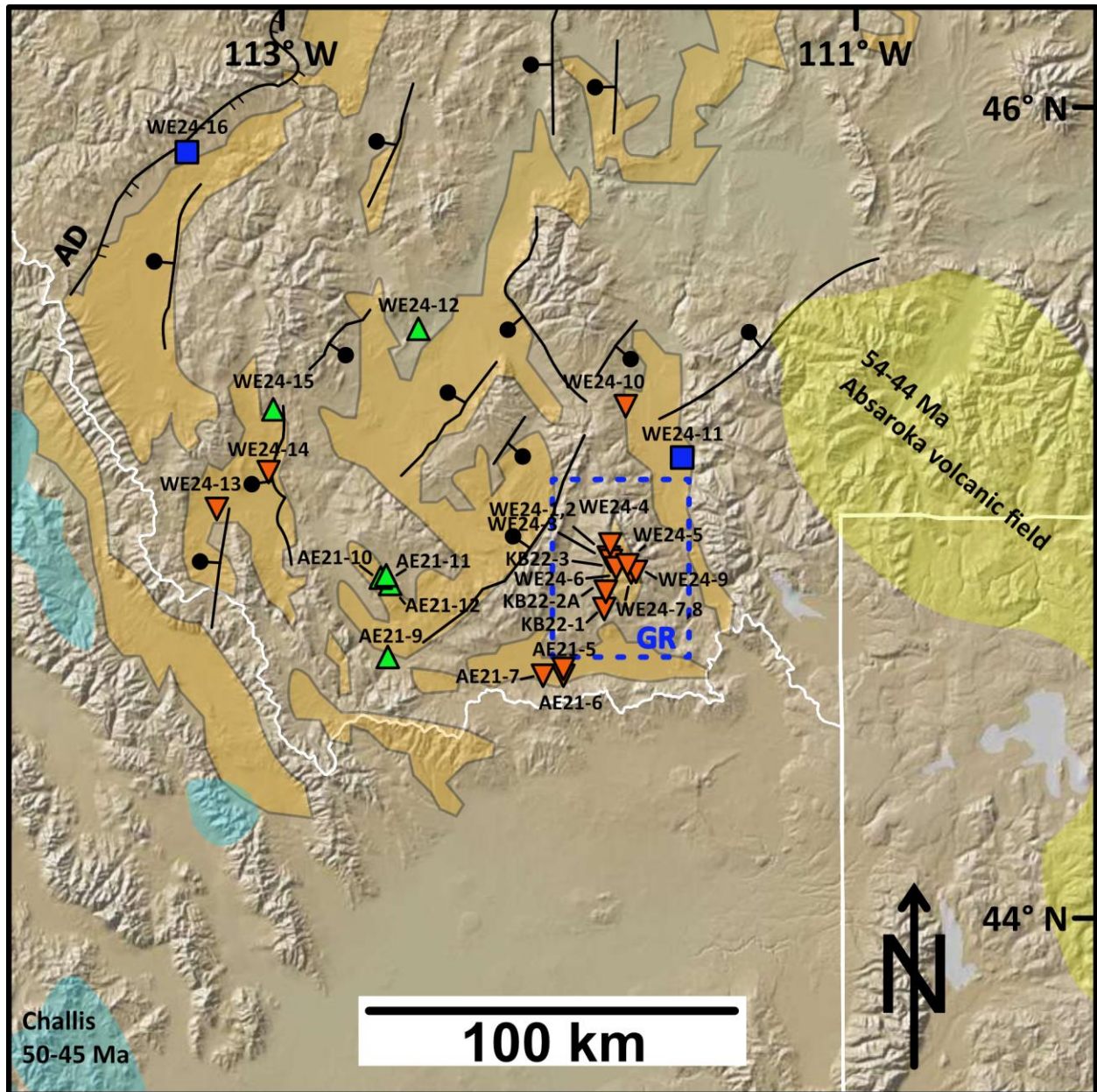


Figure 3: Map showing location of Miocene normal faults, the Anaconda Detachment fault (AD), modern valleys with Paleogene/Neogene deposits older than Sixmile Creek Formation (e.g., Renova Formation brown polygon), Challis volcanics (blue shading), Absaroka volcanic field (yellow shading) and DVF sample locations. Map modified from Stroup (2008), and Vuke (2022).

Continued regional extension in southwest Montana resulted in middle/upper DVF magmatism at ~33-16 Ma (Broley et al., 2023; Brueseke unpublished $^{40}\text{Ar}/^{39}\text{Ar}$ ages). The middle DVF occurred coevally with Renova sedimentation and is volumetrically smaller than the

lower DVF (Fritz et al., 2007; Schwartz and Schwartz, 2013; this study). The middle DVF is predominantly basalt lavas and shallow intrusives; Fritz et al. (2007) speculated that this magmatism occurred due to steepening slab angle and associated asthenosphere melting, coupled with extension. The upper Dillon volcanics contain basalts and at least one basanite lava and are also interpreted to represent derivation from the asthenosphere (Fritz et al., 2007). The upper DVF timing also demarcates the initiation of Basin and Range-style faulting, with eruptions occurring during the hiatus between deposition of Renova and the younger (~17-2 Ma) Sixmile Creek Formation strata (Fritz et al., 2007; Sears et al., 2009; Schwartz and Graham, 2017).

Other proposed tectonic models include lithospheric drip for the middle DVF, in which the continental lithospheric mantle drips into the deeper mantle resulting in uplift directly above the drip (Li et al., 2021) or back-arc extension due to Farallon slab steepening, or the collapse of overthickened continental lithosphere in response to asthenosphere upwelling for middle and upper DVF (Constenius, 1996; Fritz et al., 2007; Smith et al., 2014). The ‘drip’ hypothesis is not widely accepted (nor has it been tested on DVF magmatic products) and could indicate that lithosphere delamination is a much more important process in locations where transitions from compression to extensional tectonics occur.

Extension From Steepening Slab/Rollback (Middle/Upper Dillon)

Late Paleogene extensional basins developed in the Cordilleran foreland fold and thrust belt, resulting from Farallon slab steepening/rollback in which the middle and upper Dillon volcanics coincide spatially with areas of local extensional basin formation (Fritz et al., 2007; Schwartz and Schwartz, 2013; Li et al., 2021). One such extensional feature, the Anaconda metamorphic core complex, formed at ~53 Ma in Montana along a reactivated fault that coincides with the onset of ignimbrite flare-up volcanism (Howlett et al., 2021). The DVF

overprints this transition from compressional to extensional tectonics marked with renewed magmatism with the middle and upper ~33-16 Ma DVF, coeval with continued sedimentation and deposition of the Renova Formation (Schwartz and Schwartz, 2013).

Lithospheric Drip (Middle/Upper Dillon)

Lithospheric drip and accompanying delamination are an alternative model that has been proposed for the entire DVF (Howlett et al., 2021; Li et al., 2021). A model proposed by Li et al. (2021) depicts lower lithospheric removal (lithospheric drip) and concurrent upwelling of the asthenosphere along the margins of the drip contributing to Oligocene Dillon volcanism. Li et al. (2021) favor lithosphere removal as an explanation for ~32-30 Ma regional uplift occurring post-orogenic collapse >17 Ma after the start of extension in southwestern Montana. They interpret this as contradicting previous assumptions that the Cordillera orogen gradually lost elevation during extension. Instead of melting of the upwelling asthenosphere along the margins of the dripping lithosphere, as documented for drip-associated basalt magmatism along the margins of the Colorado Plateau by Crow et al. (2011), the Li et al. (2021) model depicts no asthenosphere melting at the drip margins. Instead, melting is suggested to have occurred within the lithospheric drip center and represents the source melts for the Dillon volcanics in the early Oligocene. Prolonged Cretaceous and early Paleogene contraction and lithosphere thickening is proposed to have set the stage for lithosphere removal, with the subsequent heating and weakening of the lithosphere commencing with shallow-slab rollback (Li et al., 2021).

Howlett et al. (2021) proposes that the foundering of the Farallon slab along a slab tear beneath Idaho and western Montana began at ~55 Ma following the accretion of Siletzia in the northern Cordillera. South of the slab tear, the Farallon slab became steeply dipping resulting in

tearing or necking separation, which enabled southward buckling and dripping of the slab (Howlett et al., (2021).

Chapter 2 - Methods

Samples of mafic lavas/intrusives were collected based on published geochemical datasets and maps of DVF lavas and intrusives from Fritz et al. (2007), Mosolf (2021), and the Centennial. Range/Gravelly map from O'Neill and Christiansen (2004). The aim was to obtain a representative group of samples from the lower, middle, and upper Dillon volcanics (Fig. 4). Emphasis was put into collecting the youngest of the Dillon volcanics (upper member) since published geochemical data from upper DVF samples are lacking. We also focused on locations on and around the Gravelly Range (and Centennial Valley, MT; Figs. 4 and 5), since prior mapping suggested this region hosts the most voluminous and diverse exposures of DVF lavas and intrusives and was likely a major DVF volcanic center.

In total, we collected 26 samples of mafic rocks in southwestern Montana from accessible exposures and only the freshest samples were collected from the field. For geochemistry, fist-sized rock samples were prepared at Kansas State University using a RockLabs Hydraulic Press with WC splitting jaws. All weathered surfaces were removed via diamond-tipped rock saw and a 60-grit sandpaper grinding wheel. All samples were then cleaned in deionized water with a scrub brush and allowed to dry overnight. When dry, all samples were crushed to pea-size gravel and smaller fragments using tungsten crushing plates of a RockLabs Hydraulic Press and randomized using a cone-and-quarter method on a glass plate. Crushed samples (~40 mL) were collected after cone and quartered and powdered for eight minutes using a Spex Industries shatterbox with an alumina puck assemblage.

All rock sample powders were sent to Hamilton Analytical Laboratory for whole rock geochemical analyses. The samples were analyzed on a Thermo ARL Perform'X x-ray fluorescence spectrometer, measuring 44 elements (major and trace). Only major elements are reported here and are as weight percent oxide (SiO_2 , TiO_2 , Al_2O_3 , Fe_2O_3 , MnO , MgO , CaO , Na_2O , K_2O , and P_2O). Total volatiles, as loss on ignition (LOI), were determined for each sample by heating overnight in silica crucibles at 900°C . For LA-ICP-MS trace element analysis, low dilution lithium-borate-fused glass pellets were analyzed to minimize nugget effects. Moreover, an in-house drift monitor (HAL-2A) was used, as it has similar sensitivities at all atomic numbers when compared to the standard NIST 160, which is not consistent across a wide range of atomic numbers. Hamilton Analytical Labs also uses a newer code, which was written to calculate weight means for multiple internal standards. A photon Machines Analyte 193 (G1) ablation station (UV excimer laser, 193 nm, with laminar flow frame cell) attached to a Varian 820 single quadrupole mass spectrometer was used to ablate samples and standard at Colorado School of Mines. The laser spot size is $150\ \mu\text{m}^2$ and has a 7 hz repetition rate. Helium carrier gas carried the ablated material from the laser cell toward the Varian 820 ICPMS, mixing with Ar to 20 ml before entering the plasma torch. The ICP-MS is run in “peak-hopping, time-resolved mode” for 53 major and trace element analyses, with dwell times around 10 ms per mass (Conrey et al., 2019; Conrey et al., 2023). Trace element concentrations are presented as parts per million (ppm). All major, trace, and rare- earth element (REE) data are presented in APPENDIX A. Fe was split according to Le Maitre (1976) and all major element data used in diagrams and the discussion are reported as anhydrous using the split Fe data. Though we collected the freshest possible samples in the field of mafic rock, LOI values and petrographic evidence suggest that some samples have been altered by secondary fluid alteration and/or

weathering. This is based on: (1) petrographic observations of secondary phyllosilicate and/or carbonate phases; (2) eight samples with LOI > 3.5 wt%; and/or (3) anomalous (i.e., non-igneous) relationships between Mg, Fe, Ni, Cr, V, and Sc. Our 26 new bulk rock analyses are combined with 53 major and trace element analyses of DVF mafic rocks previously reported by Fritz et al. (2007), some of which lack modern trace element analyses, Luikart (1997), and Mosolf (2021 – map); in total, 52 mafic DVF rocks are included in our study.

Sr-Nd Isotopes

Ten representative DVF samples that encompass all three DVF members (e.g., lower, middle, upper) were analyzed for Sr-Nd isotope analyses at Washington State University via a Thermo-Finnigan Neptune 28 multicollector (MC)-ICP-MS. Methods for Sr and Nd analysis follow those found in Vervoort and Blichert-Toft (1999) and Vervoort (2004), with important steps highlighted here for information. For Sr and Nd whole-rock isotopic analysis, ~0.25 g aliquots of whole-rock powders were dissolved at high pressure in sealed, steel-jacketed Teflon bombs with a 10:1 mixture of concentrated HF and HNO₃ acids at 150°C for 5-7 days. After conversion from fluorides to chlorides, samples were spiked with mixed ¹⁴⁹Sm-¹⁵⁰Nd. Sr aliquots were purified using micro-columns (0.18 ml resin volume) with Sr-spec resin and HNO₃. Nd and Sm were separated from the LREE aliquot on columns with HDEHP-coated Teflon powder and HCl (Vervoort and Blichert-Toft, 1999). Sr analyses were corrected for mass fractionation using ⁸⁶Sr/⁸⁸Sr=0.1194 and normalized to a value of 0.71024 for the ⁸⁷Sr/⁸⁶Sr ratio of the NBS-987 standard. Nd analyses were corrected for mass fractionation using ¹⁴⁶Nd/¹⁴⁴Nd=0.7219 and normalized to a value of 0.512138 for the ¹⁴³Nd/¹⁴⁴Nd ratio of the Ames Nd standard and 0.511858 for the La Jolla standard. Sm analyses were corrected for fractionation using ¹⁴⁷Sm/¹⁵²Sm=0.56081. All mass fractionation corrections were made using the exponential

law. Recorded $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ values for the DVF samples were corrected to initial Sr-Nd isotope values ($^{87}\text{Sr}/^{86}\text{Sr}_i$, $^{143}\text{Nd}/^{144}\text{Nd}_i$) using published age dates or to 30 Ma with no reported geochronology.

Chapter 3 - Results

Field Relations and Physical Characteristics

Fieldwork in 2021, 2023 and 2024 targeted intrusive and extrusive mafic rocks of the DVF (Fig. 4). Fieldwork within the Gravelly Range (Fig. 5) targeted primarily volcanic centers, as well as possible related lavas to the south in the Centennial Valley that were originally mapped by O'Neil and Christiansen (2004) as Tmv. The Gravelly Range, especially in the vicinity of Black Butte and Lion Mountain, were targeted because they are eroded volcanoes with vent facies, plugs, and dikes in which we collected samples for modern geochemical and isotope analysis. A lava north of Black Butte (WE24-4) and lavas south of the vent facies of Lion Mountain (WE24-7, WE24-8) were also sampled. Samples of the DVF were also collected outside of the Gravelly Range from lavas in the adjacent Madison valley, as well as the Big Hole River Valley, and scattered outcrops to the west around Dillon, Bannack, and Virginia City Montana (Fig. 4).

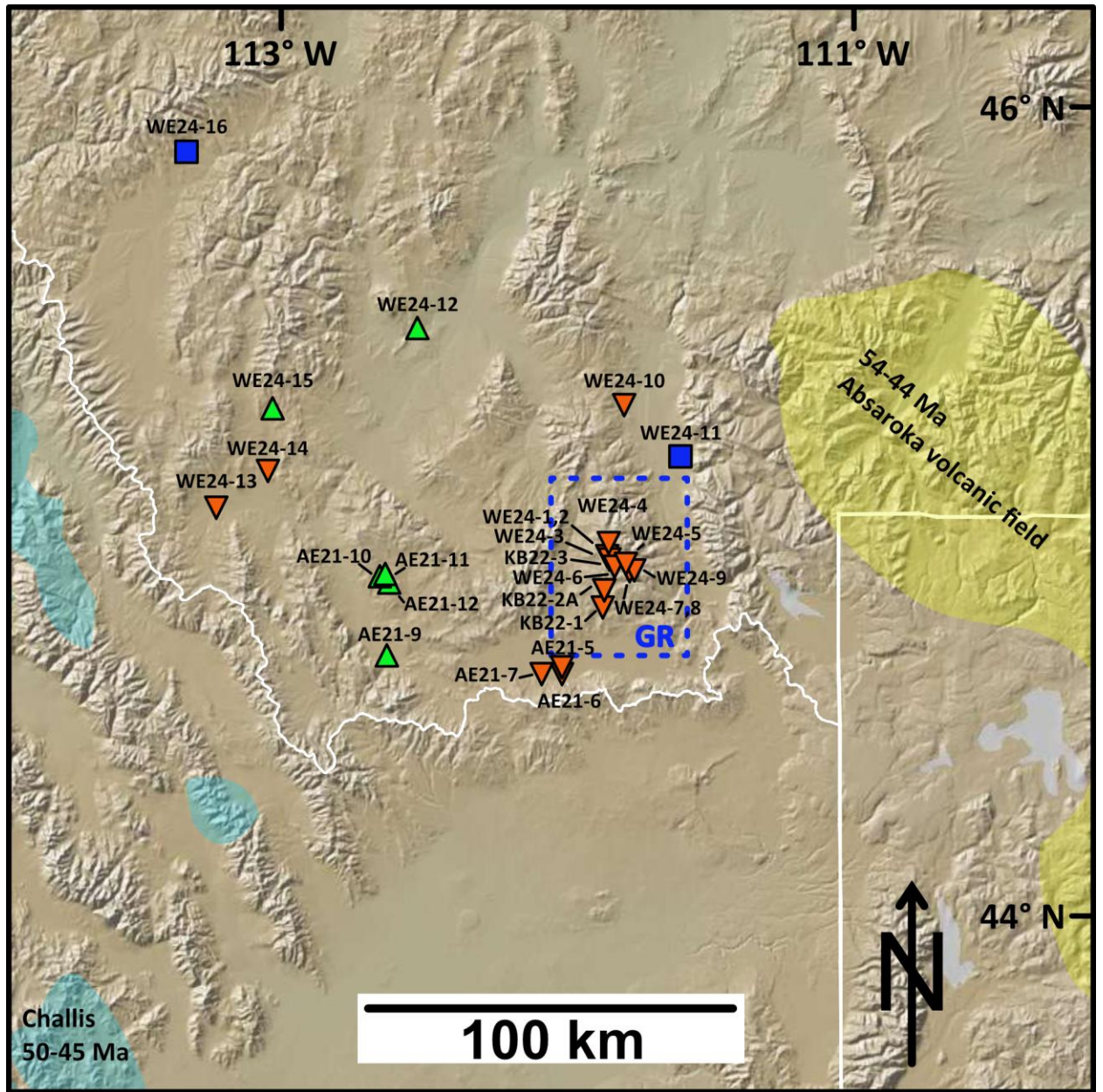


Figure 4: Map showing mafic rock sample locations during 2021, 2023, and 2024 field work in the DVF. Blue dashed box denotes the Gravelly Range (GR), yellow region is the Absaroka volcanic field, and the blue region is the Challis volcanic field outcrops. Sample symbols as in Fig.3, where blue square = upper DVF, orange triangle = middle DVF, and green triangle = lower DVF.

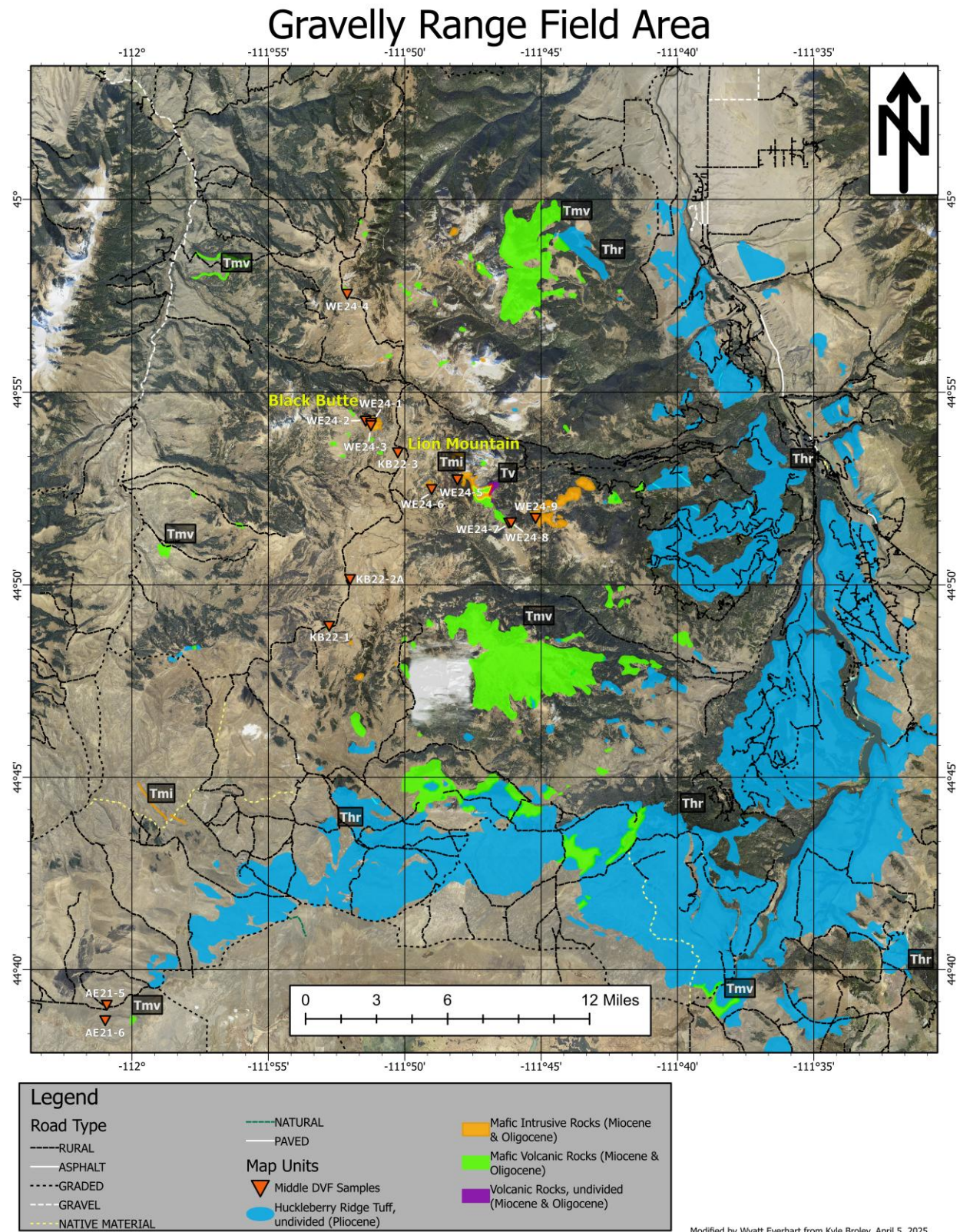


Figure 5: Map of the Gravelly Range in southwestern Montana showing multiple map units including DVF lava units and sample locations.

Black Butte

The ~28.6 Ma Black Butte (Gutmann et al., 1989; K-Ar date by Fritz et al., 2007) is an eroded volcanic plug and eruptive center within the Gravelly Range at an elevation of 3,215 m (Fig. 6). Columnar jointing is observable along most of its sides with piles of talus at its base. Pyroclastic vent facies remain on its south slope, including a ~5-10 m thick mantle of scoria and bombs and near the summit on the south slope (Fig. 6). The summit is composed of lava, cut by an intrusive apatite magnetite olivine gabbro. Hyaloclastic pyroclastic deposits around the base of the plug dip away from its core at ~20 degrees and overlie Renova strata; these and are interpreted to reflect initial phreatomagmatic eruption activity (Marvin et al., 1974; Gutmann et al., 1989; this study).

Lion Mountain

Lion Mountain is an eroded volcanic center that was active at ~30.8 Ma (K-Ar date by Gutmann et al., 1989; Lofgren et al., 2020). It is characterized by pyroclastic vent facies at its summit, as well as dikes and sills and a ~50 m thick package of at least eight ridge-forming lavas to its south (Figs. 7, 8). Multiple dikes cut through the ~240 m-thick Renova formation, which is now exposed, unconformably overlying Mississippian Madison limestone, and forms slopes and cliffs that make up Lion Mountain and the immediate vicinity. A ~50 m-thick sill was observed cutting horizontally through Renova formation with only a few meters of Renova sediments at its top with a ~ 50-m-thick lava package overlying the Renova sediments above. The top lava on the observed and sampled south ridge lava package, approximately 3 km south of the summit of Lion Mountain, is vesicular, porphyritic and more silicious in appearance (52.9 wt% SiO₂) while the lower lava in the package is vesicular and slightly more mafic (50.1 wt% SiO₂). Rock glaciers occur on the eastern slope of Lion Mountain. A separate columnar intrusive body over

0.1 km² in areal extent was observed and sampled ~2 km west of Lion Mountain, along with an additional intrusive body sampled ~ 1km from the ridge lava, ~3.5 km southeast of Lion Mountain (Fig. 5).

Scattered Lava Outcrops

Numerous mafic samples were collected within and outside the Gravelly Range and Centennial Valley. The outcrops ranged from small, poorly exposed lavas within modern valleys, to thick ~80 – 100 m, often columnar jointed lavas, such as those sampled from the Block Mountain and Bannack Bench Road columnar lavas (Figs. 7, 9, 10). These scattered outcrops of lava throughout southwestern Montana outside of the Gravelly Range region may represent a more monogenetic type of volcanism in which lava erupted and then flowed down existing paleovalleys rather than forming larger volcanic centers as observed in the Gravelly Range. For example, a columnar lava along Bannack RD in Montana is ~88-m-thick and is similar in appearance to Columbia River Flood Basalts outcrops suggesting this lava flowed down a paleovalley in which it then inflated and thickened. Lavas in Fig. 10A and 10C overlie Renova sediments which is evidence of interstitial Dillon volcanism during basin sedimentation. These lavas are both faulted, and one is tilted, which shows continued extension in the region.

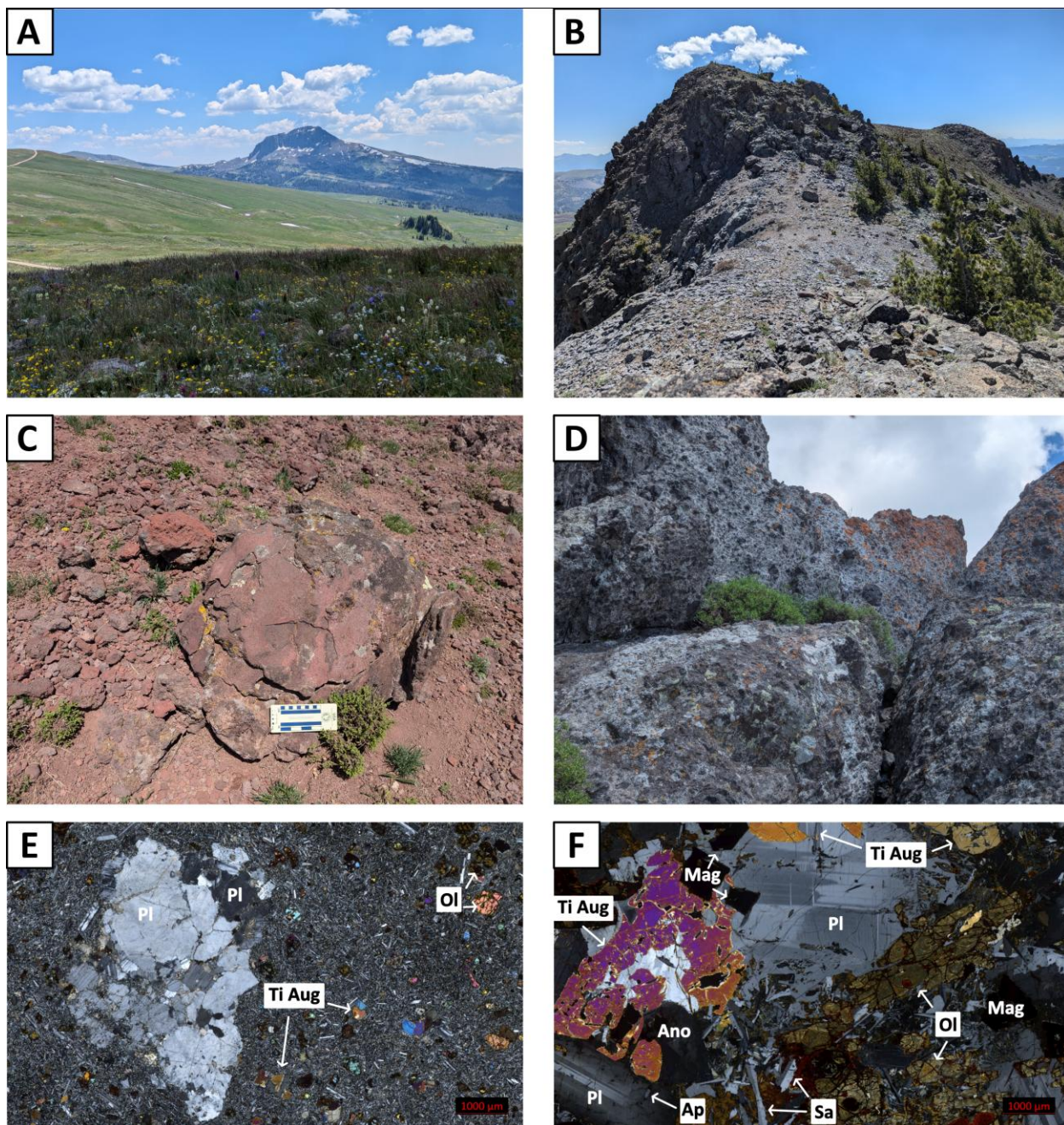


Figure 6: Field photos and photomicrographs of Black Butte. (A) Looking south to Black Butte from lava outcrop where sample WE24-4 was collected. (B) Summit of Black Butte looking to the east. (C) Volcanic bomb on the south slope of Black Butte. (D) Hyaloclastic pyroclastic deposits along the southern slope of Black Butte. (E) Photomicrograph of WE24-1 showing large plagioclase clots with phenocrysts of olivine and titanite; the sample was taken from the summit of Black Butte. (F) Photomicrograph of WE24-2, a more evolved, apatite, magnetite bearing gabbro taken from the summit of Black Butte.

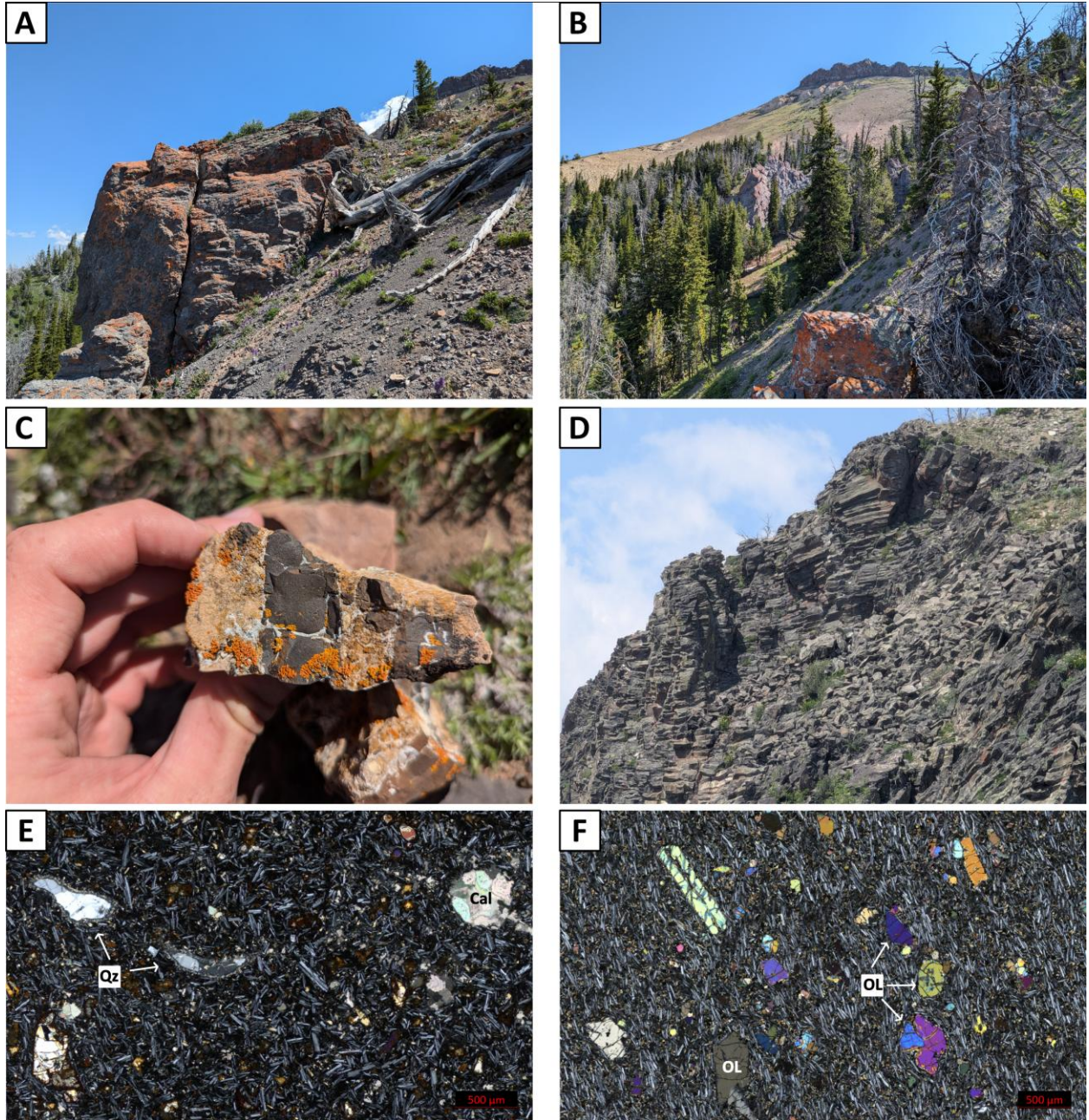


Figure 7: Field photos and photomicrographs of DVF in the Gravelly Range. (A) Dike cutting through Renova Formation along the western slope of Lion Mountain (B). (C) Hand sample of dike margin and Renova sediments. (D) Looking west at large columnar intrusive body located ~2 km west of Lion Mountain. (E) Photomicrograph of intrusive dike (WE24-5) showing calcite amygdaloids and quartz xenocrysts; the samples were collected from location in photo (A). (F) Photomicrograph of sample WE24-6, showing olivine phenocrysts and groundmass plagioclase that displays trachytic flow textures; the sample was collected from the location shown in photo (D).

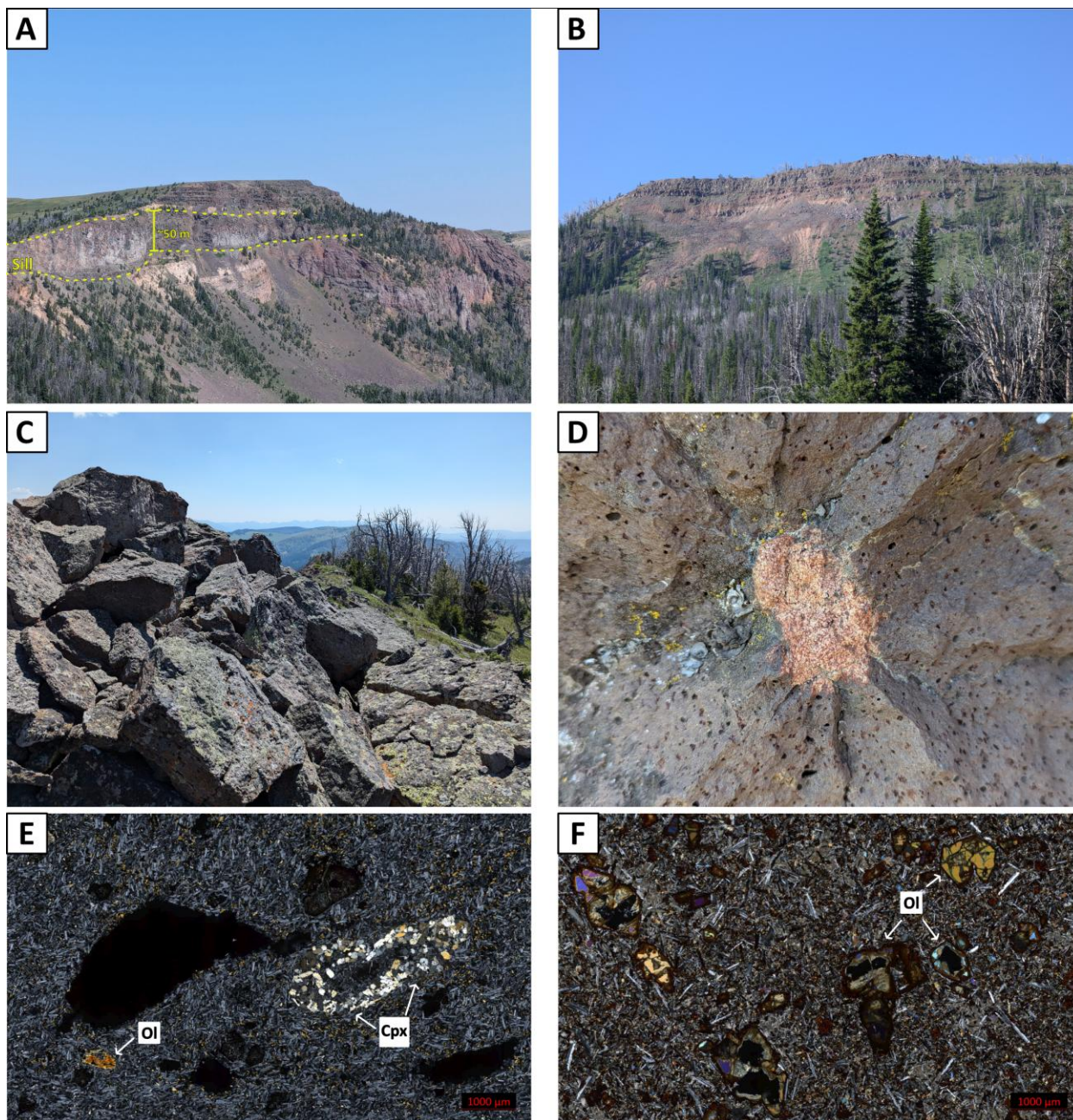


Figure 8: Field photos and photomicrographs of Lion Mountain. (A) Looking north to eroded exposures of lava and a sill south of Lion Mountain (yellow dashed lines outline sill ~50m thick). Yellow to orange outcrop is Renova sediments that lie below the sill and are found as a thin ~9 m outcrop above it with a package of lava capping Renova. (B) Looking southwest up at a package of multiple lavas south of Lion Mountain where photo (A) was taken. (C) Top of lava package that sample WE24-7 was collected. (D) Lava near the base of lava package which contains a sedimentary xenolith (E) Photomicrograph of sample WE24-7 showing large oval shaped vesicles and a crystal clot consisting of mostly clinopyroxene. (F) Photomicrograph of sample WE24-8, collected from basal lava in photos (B) and (D), containing altered (iddingsitized) olivine phenocrysts.

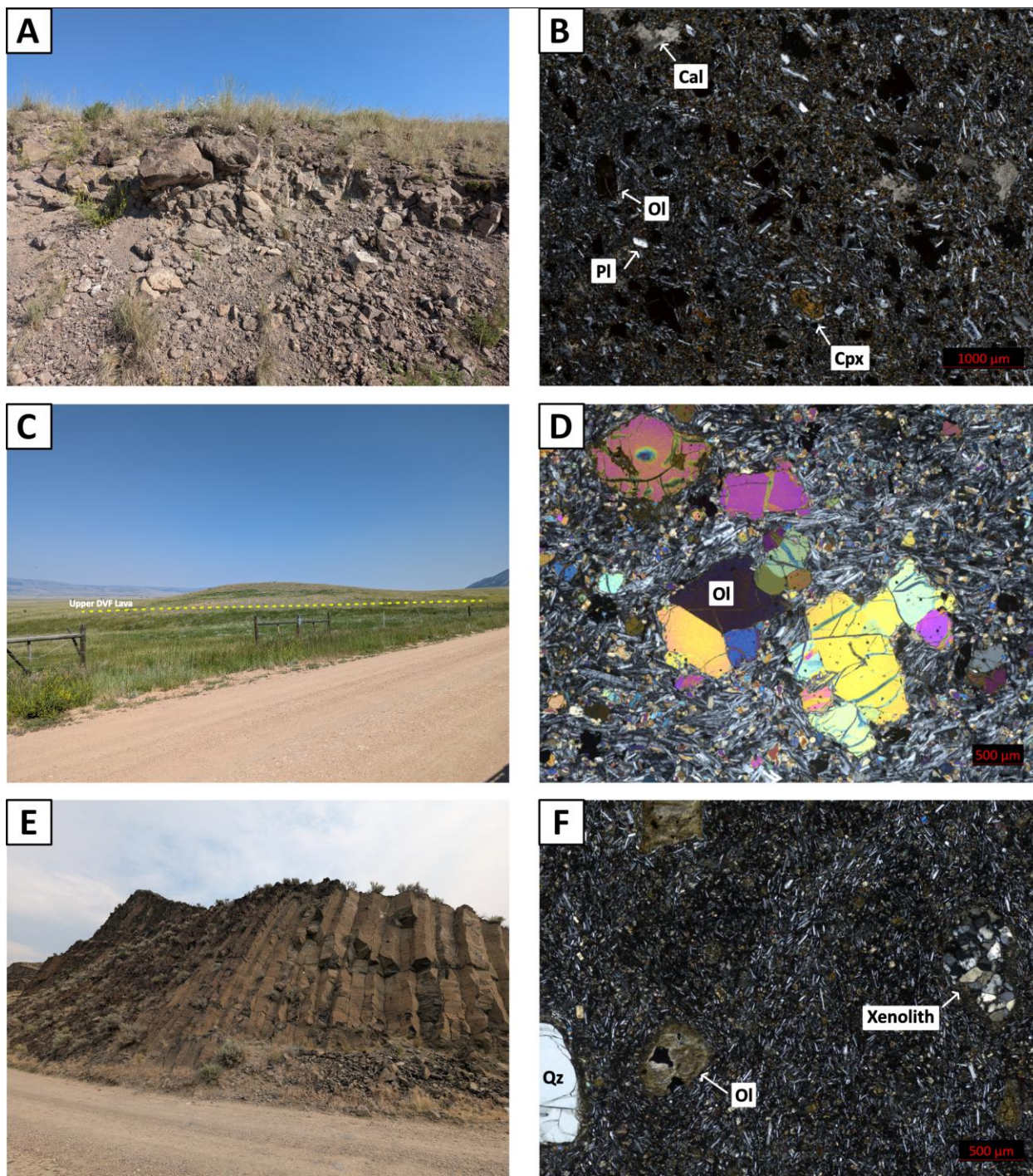


Figure 9: Field photos and photomicrographs of regional DVF exposures. (A) Small outcrop of lava along Antler Trail in Varney Montana, along the north flanks of the Gravelly Range (sample WE24-10). (B) Photomicrograph of lava (A), which is highly vesicular and contains quartz xenoliths. (C) Small hill composed of lava where sample WE24-11 was collected from a cobble near the road, in the eastern Madison Valley. (D) Photomicrograph of sample WE24-11 showing multiple olivine phenocrysts in a relatively unaltered sample. (E) Columnar lava south of Block Mountain located along Burma Rd Montana. (F) Photomicrograph of Block Mountain lava

(WE24-12) containing iddingsitized olivine phenocrysts, quartz xenoliths and pilotaxitic groundmass.

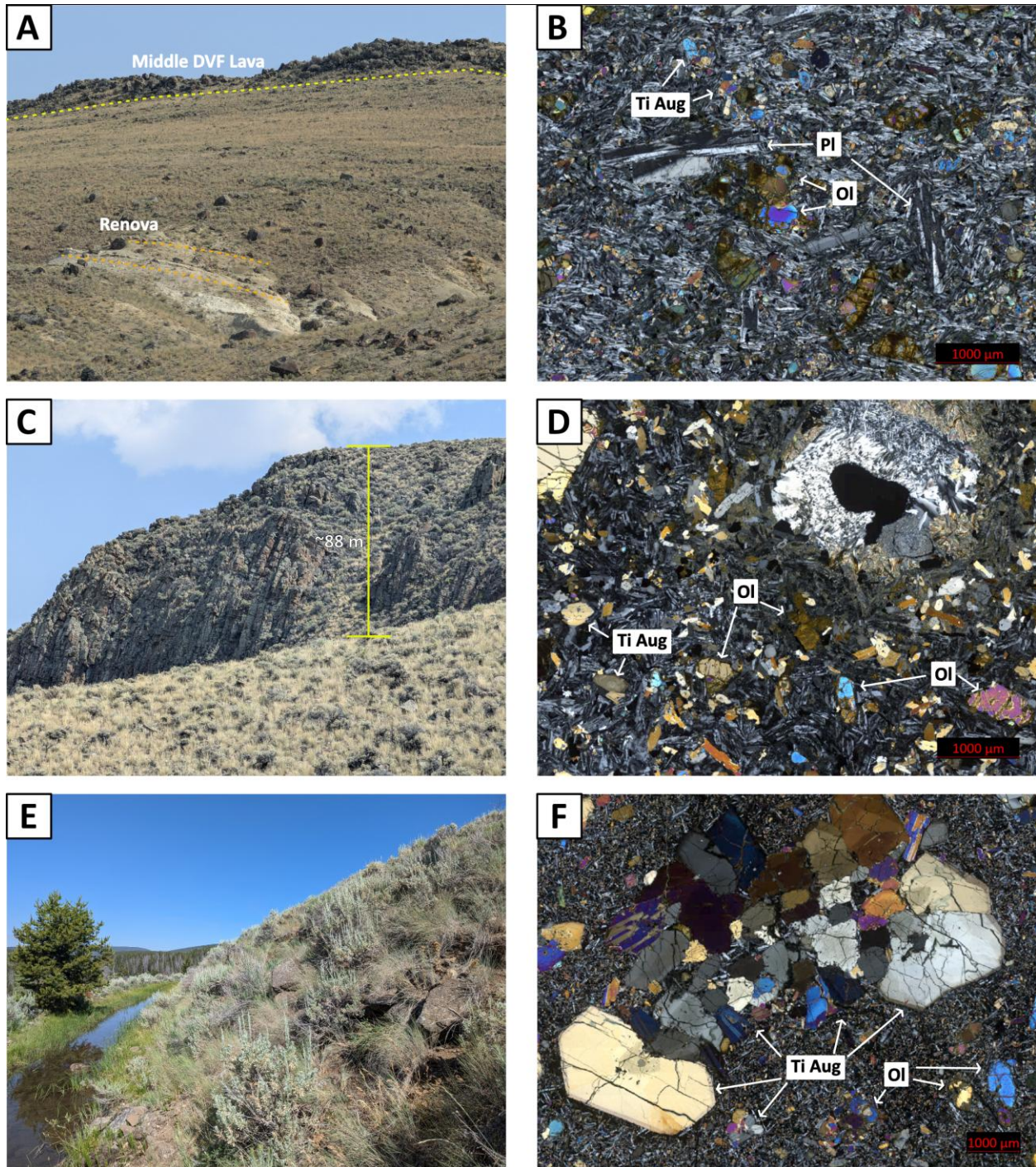


Figure 10: Field photos and photomicrographs of regional DVF exposures. (A) Outcrop of lava above Renova Formation ~700 m north of Red Butte Montana. Entire outcrop is tilted. (B) Photomicrograph of WE24-13 collected from lava in (A) showing olivine and plagioclase phenocrysts with iddingsite alteration in some olivine. (C) Columnar outcrop on the east side of Bannack Bench Rd Montana (WE24-14). (D) Photomicrograph of WE24-14 with amygdaloidal

textures. (E) Small outcrop of lava in northern Big Hole River Valley (WE24-16). (F) Photomicrograph of WE24-16 showing zoned and twinned titanaugite and olivine phenocrysts, as well as a crystal clot of titanaugite.

Bulk Rock Geochemistry of the Dillon Volcanic Field

While the entire DVF includes rock that ranges from basalt to rhyolite (Fritz et al., 2007; Mosolf, 2021), our study focused on mafic samples (defined for this study as rocks with <57 wt% SiO_2). All our newly sampled DVF rocks are basalts to basaltic andesites, except for one andesite (e.g., lower Dillon Block Mountain lava). All data plotted below includes samples prepared at K-State, along with samples from Luikart (1997), Fritz et al. (2007), and Mosolf (2021) (e.g., $n = 52$ total).

All rocks plot within sub-alkaline and transitional arrays on a total alkali vs. silica diagram (Le Bas et al., 1986) (Fig. 11A). The lower Dillon members are more evolved and all display > 52 wt% SiO_2 . The middle Dillon samples plot as sub-alkaline and transitional, ranging from mafic compositions (< 52 wt% SiO_2) to slightly more evolved (>52 wt% SiO_2). The upper Dillon rocks are all mafic and plot as sub-alkaline basalts to alkali basalts, basanites, and trachy-basalts. Sub alkaline DVF rocks plot as calc-alkaline on an alkali-iron-magnesium (AFM) diagram (Irvine and Barager, 1971) (Fig. 11B). The majority of DVF rocks plot as high-K and have absarokite and shoshonite affinities (Fig. 12). On a Harker diagram (Fig. 13), upper DVF rocks tend to have lower wt% K_2O in comparison to the middle and lower DVF, and Cr concentrations (>600) are higher in the lower DVF than the middle/upper DVF. All collected DVF rocks are strongly potassic.

All DVF rocks are plotted on an expanded CIPW normative basalt tetrahedra in (Fig. 15). Here, lower Dillon rocks range from silica saturated (olivine tholeiites) to silica oversaturated (quartz tholeiites). Lower DVF samples lie mostly along the 1 atm cotectic (Fig. 15). Middle Dillon rocks show variability through the entire range but are mostly silica saturated

with only three samples that are oversaturated (quartz-tholeiite). Upper Dillon rocks tend to be undersaturated alkali basalts, though one sample, WE24-16, plots as an olivine tholeiite which could be due to contamination from observed quartz xenocrysts and alteration; they all lie on or near the 8 to 35 kbar cotectic (Fig. 15).

Chondrite normalized REE patterns of the DVF show a gentle decreasing slope from LREE to heavy rare earth elements (HREE) with LREE enrichments from ~30 to 300 more than chondrite (Fig. 16). Lower DVF rocks have lower LREE and HREE concentrations than the other DVF rocks (Fig. 16). Middle DVF displays greater enrichment than the upper DVF in REE while upper DVF rocks tend to have lower HREE (Fig. 16). The upper DVF rocks vary; one sample has the lowest LREE enrichment but has overlapping MREE and HREE to the rest of the DVF, where the other sample has low HREE but much more enriched LREE than most other DVF rocks (Fig. 16). Generally, HREE patterns for all DVF rocks are flat to slightly concave upwards (Fig. 16). However, sample WE24-16, an upper DVF rock, has a steeper LREE to HREE pattern, different from the rest of the DVF samples (Fig. 16).

Trace and major element concentrations for the DVF show slight variations within each of the DVF members. Although all members of the DVF display variable major element characteristics, some trace elements stand out, such as the multiple samples from the lower DVF that display high Cr (between 500-1200 ppm) and Rb (between 105-150 ppm) (Fig. 14). The lower DVF is characterized by its enrichment in LREEs and large ion lithophile elements (LILE), notably in Rb, relative to primitive mantle (Fig. 17). The lower DVF is depleted in Ta and Nb and shows enrichment in Pb, relative to primitive mantle; these depletions are not present in either of the other two DVF units apart from one middle DVF rock, sample VC-35# from Mosolf (2021). The middle DVF shows similar REE concentrations as the lower DVF with one

difference being its higher enrichment in Cs. Middle DVF rocks also have greater enrichments in Ba, Nb and Ta when compared to the lower DVF (Fig. 18). The upper DVF collectively show the lowest concentrations of Cs, Ba and Pb relative to the other DVF members and relative to the other two members, flatter patterns that are more OIB-like (Fig. 19).

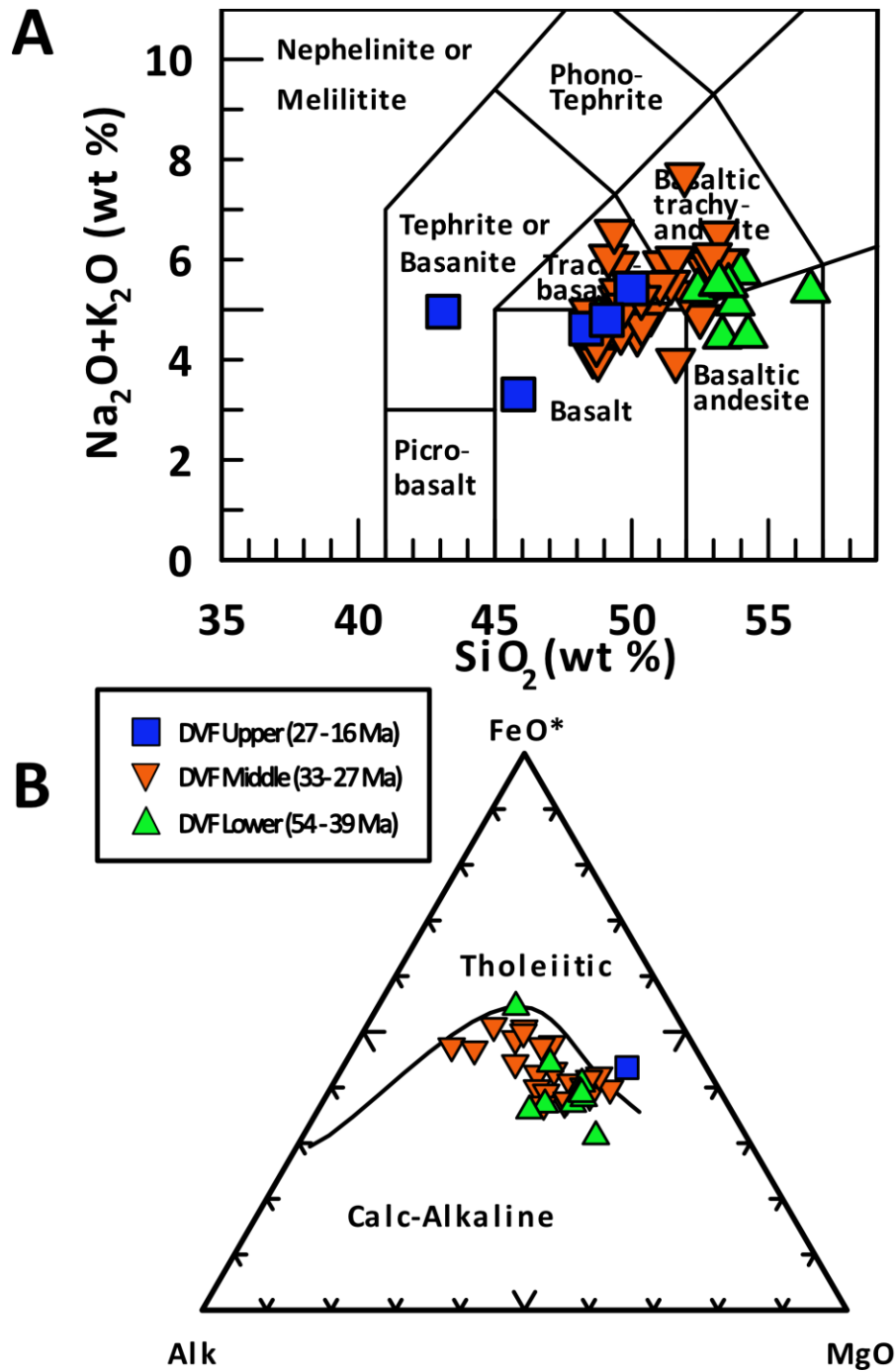


Figure 11: (A) Total alkali versus silica diagram for the Dillon Volcanic Field with classification grid of Le Maitre et al. (1989). (B) AFM diagram after Irvine and Barager (1971) for the subalkaline series of the Dillon Volcanic Field (Luikart, 1997; Fritz et al., 2007; Mosolf 2021; this study).

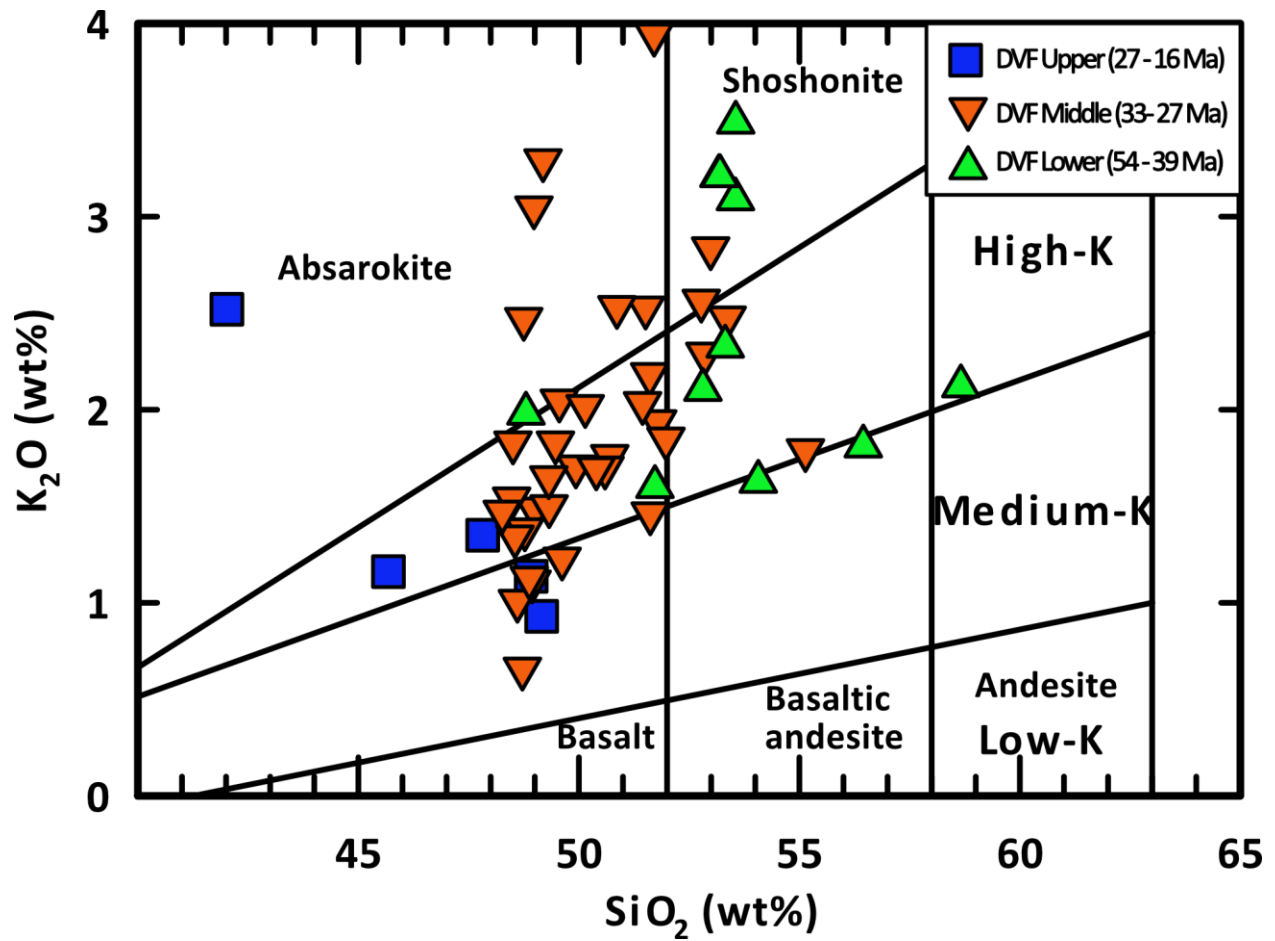


Figure 12: K₂O vs. SiO₂ discrimination diagram after Peccerillo and Taylor (1976). Low-K (tholeiites), Medium-K (calc-alkaline), High-K (K-rich calc-alkaline rocks).

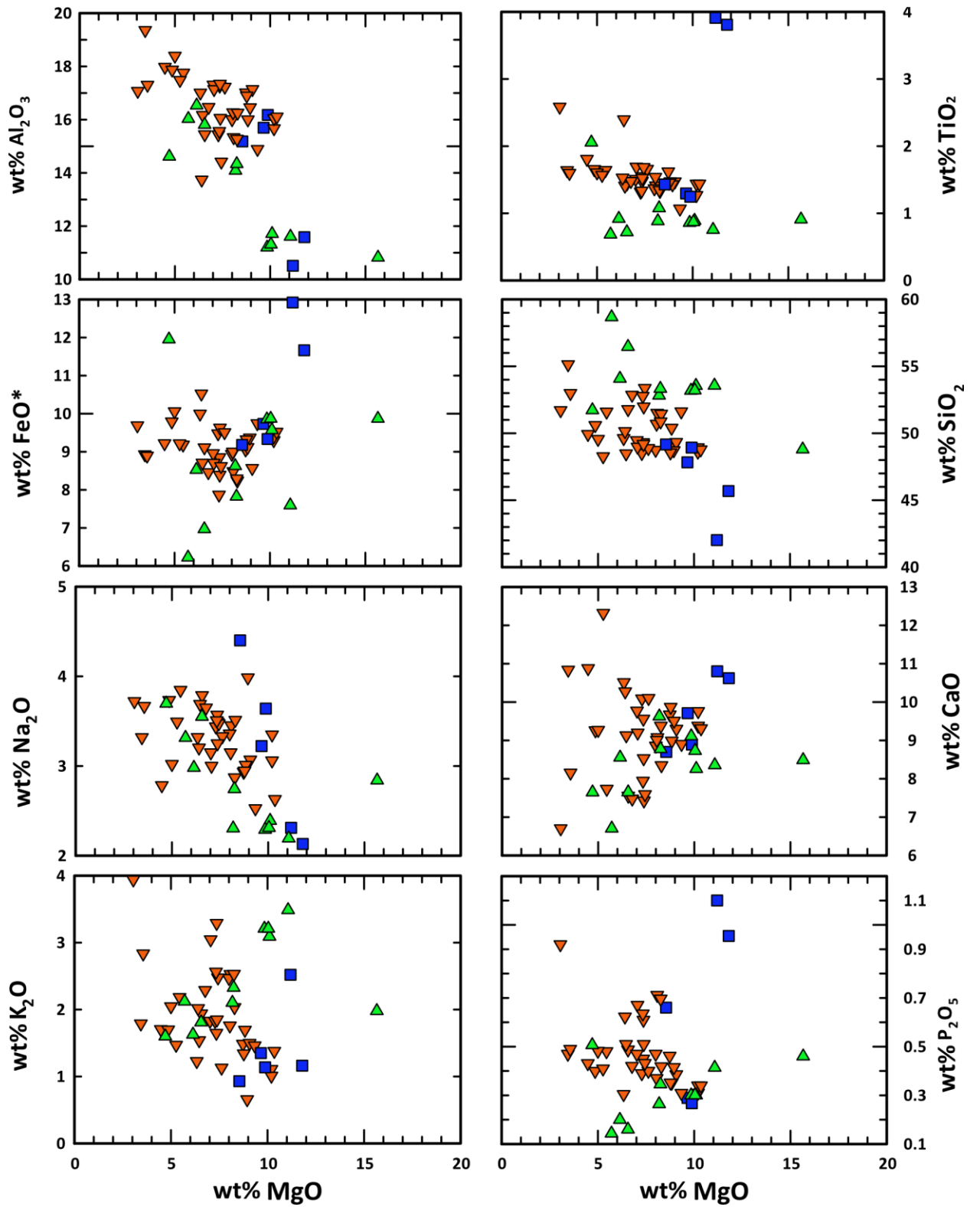


Figure 13: Harker diagrams illustrating major element variations with wt% MgO.

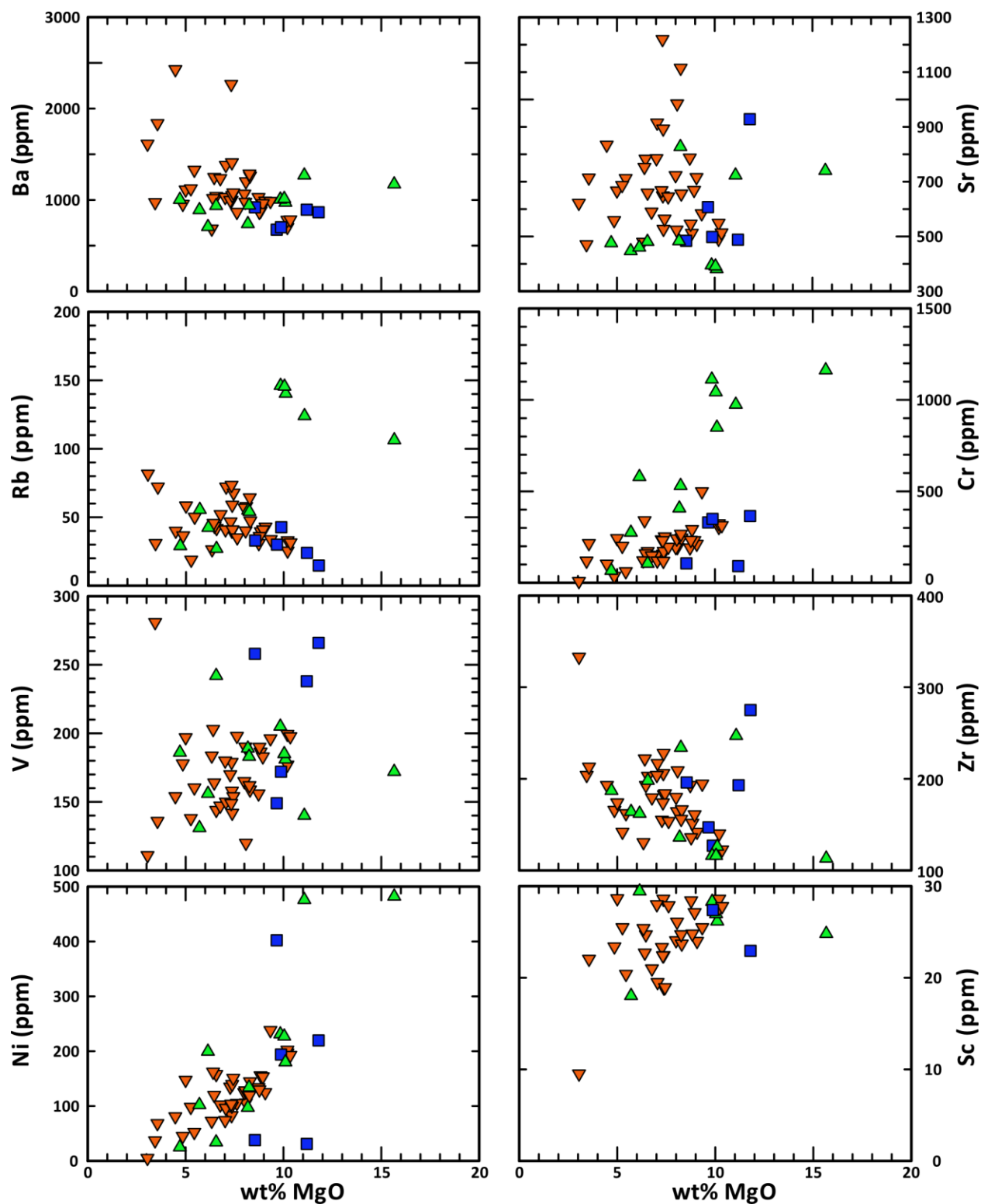


Figure 14: Harker diagrams illustrating trace element variations with wt% MgO.

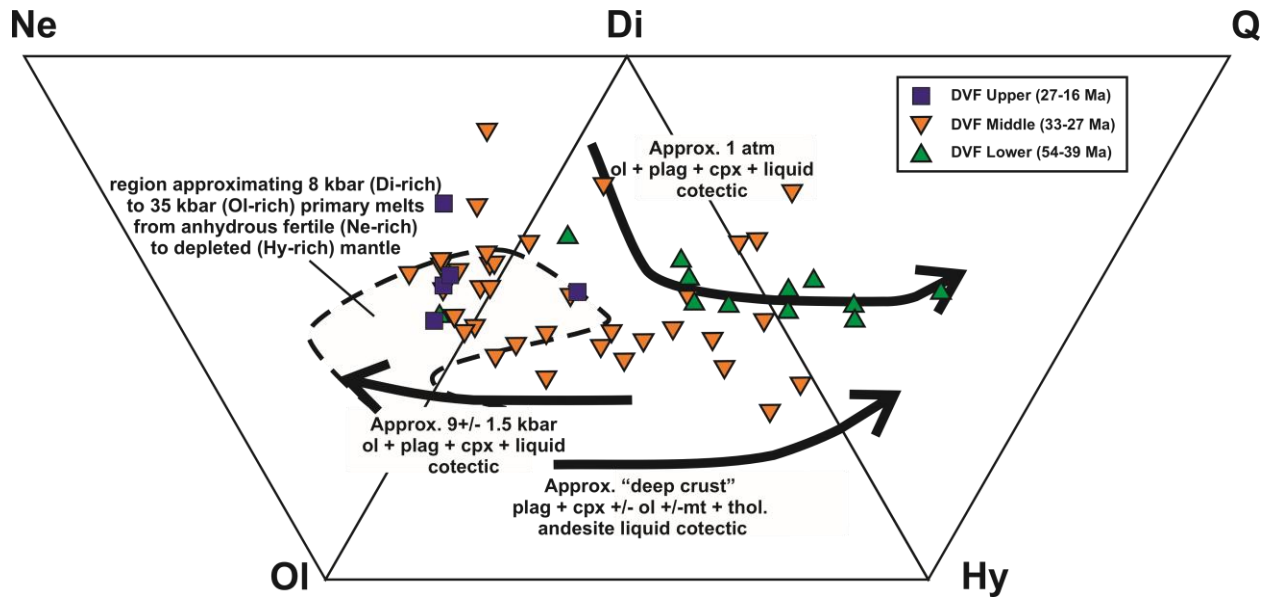


Figure 15: Classification diagram for DVF rocks based on their CIPW normative compositions projected in Ne-Ol-Di-Hy-Q compositional space after Thompson (1984).

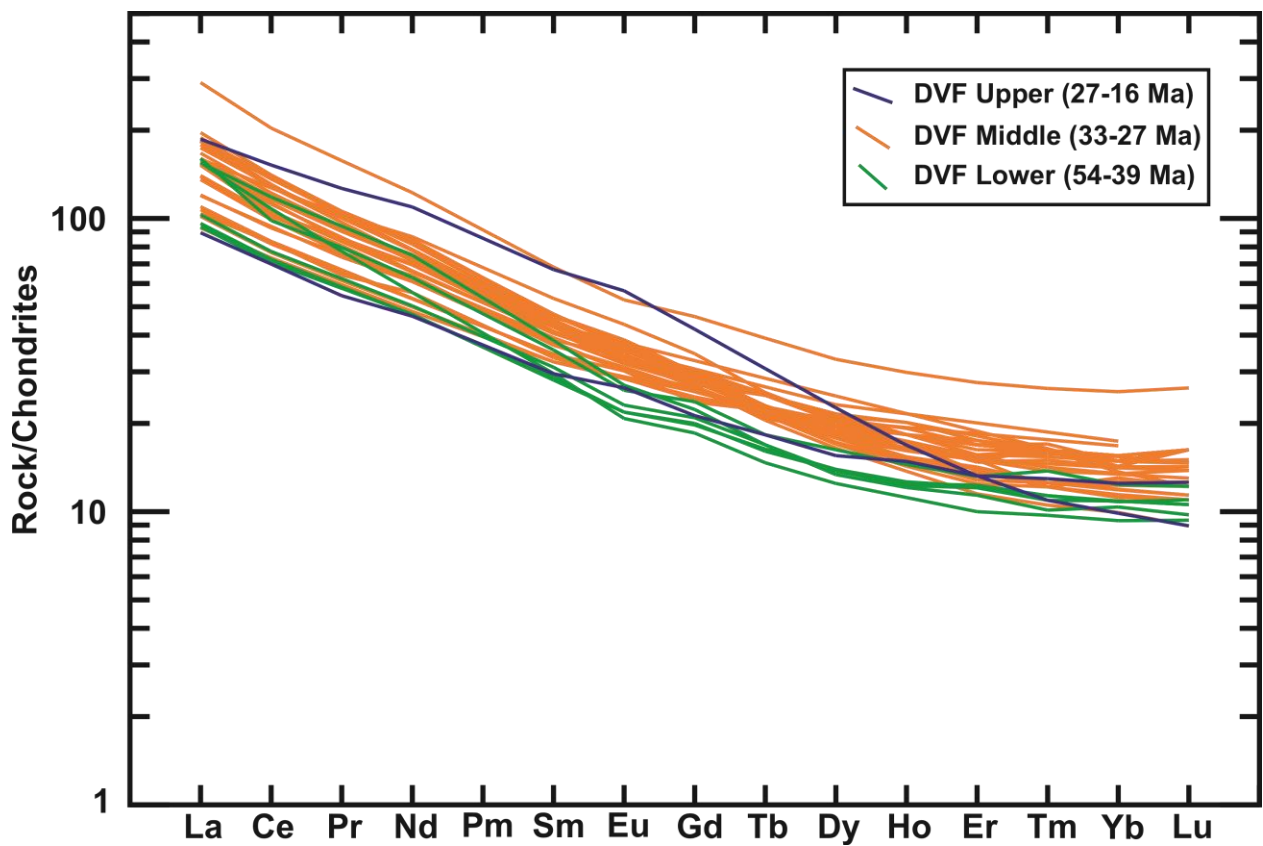


Figure 16: Chondrite normalized multi-element diagram for DVF rocks (Sun and McDonough, 1989).

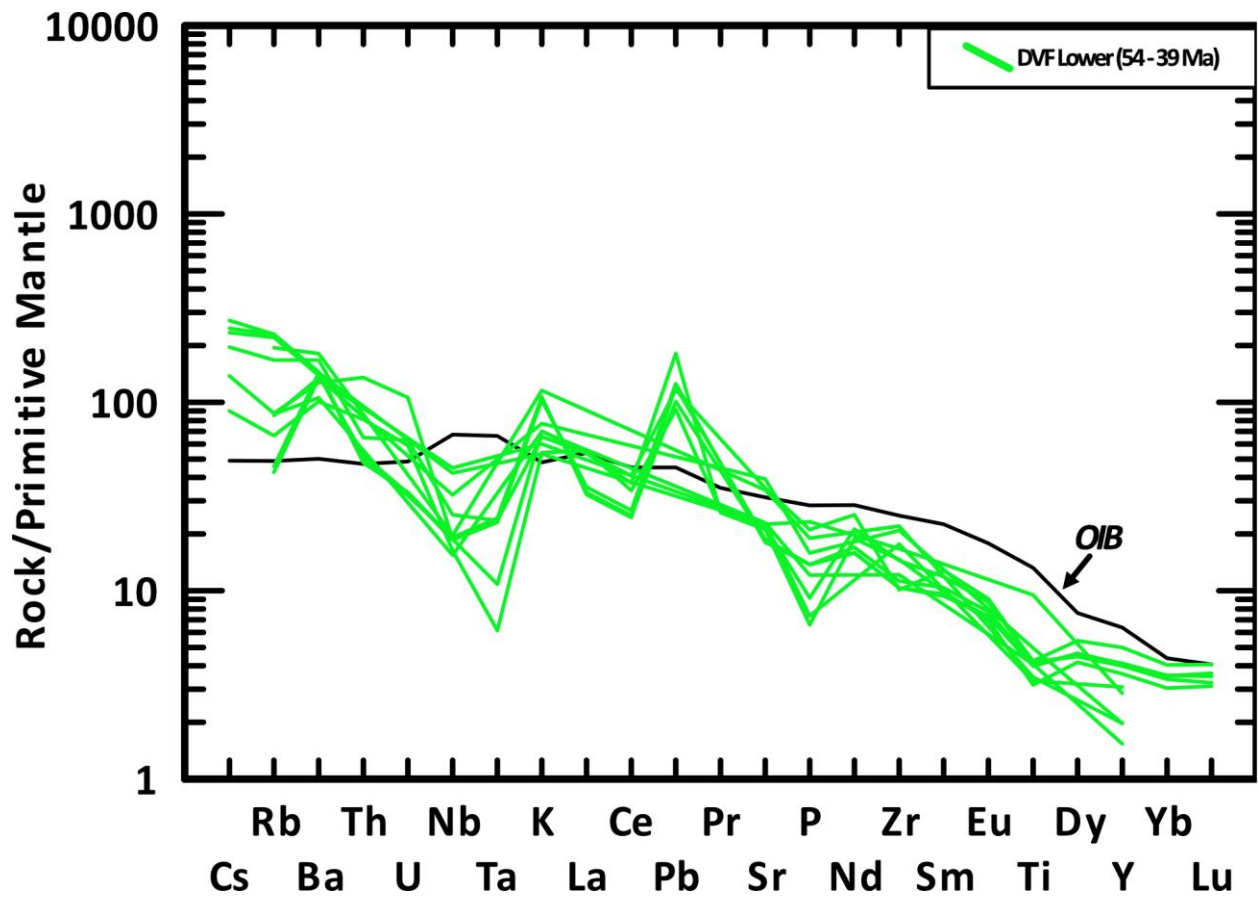


Figure 17: Primitive mantle-normalized (Sun and McDonough, 1989) trace element variations of the lower DVF compared to OIB (Sun and McDonough, 1989).

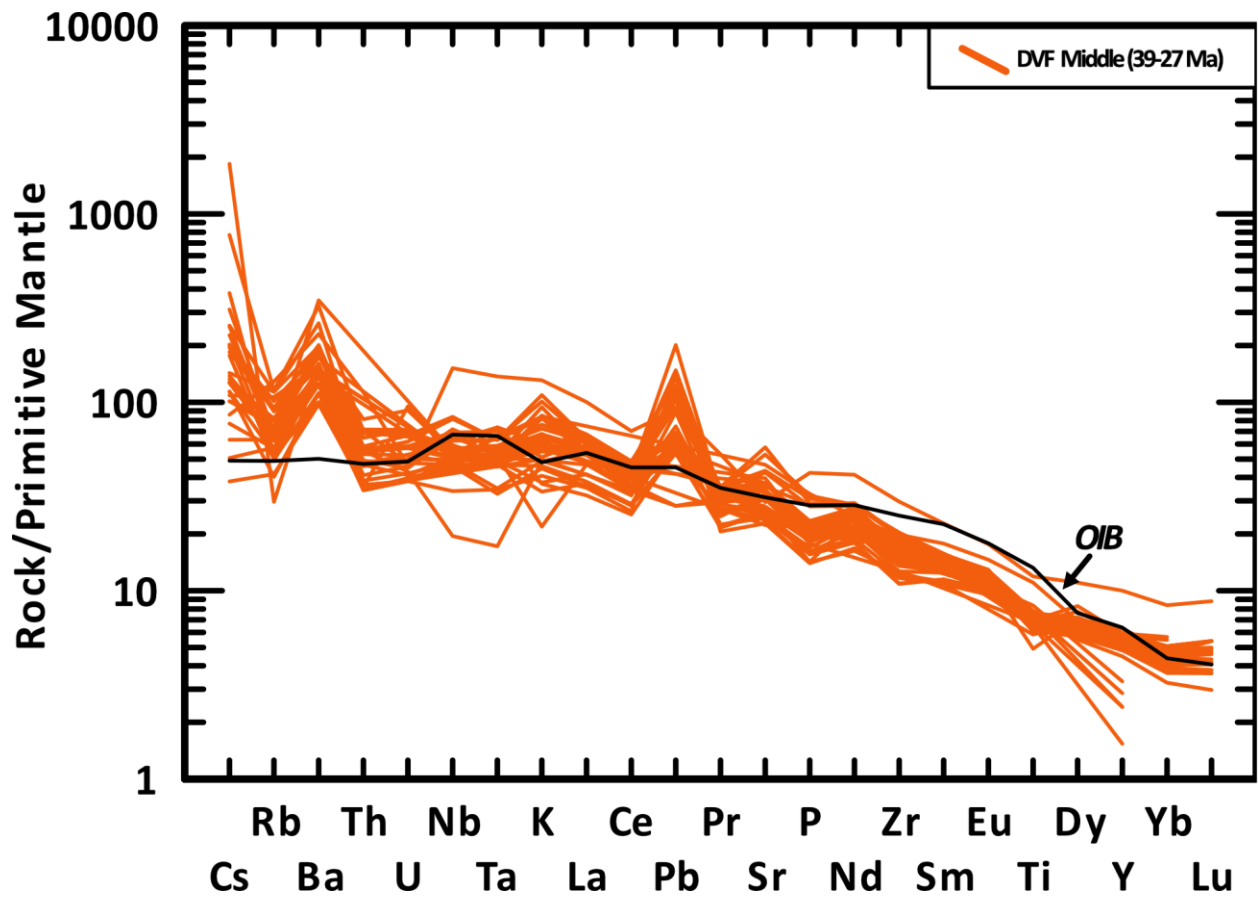


Figure 18: Primitive mantle-normalized (Sun and McDonough, 1989) trace element variations of the middle DVF compared to OIB (Sun and McDonough, 1989).

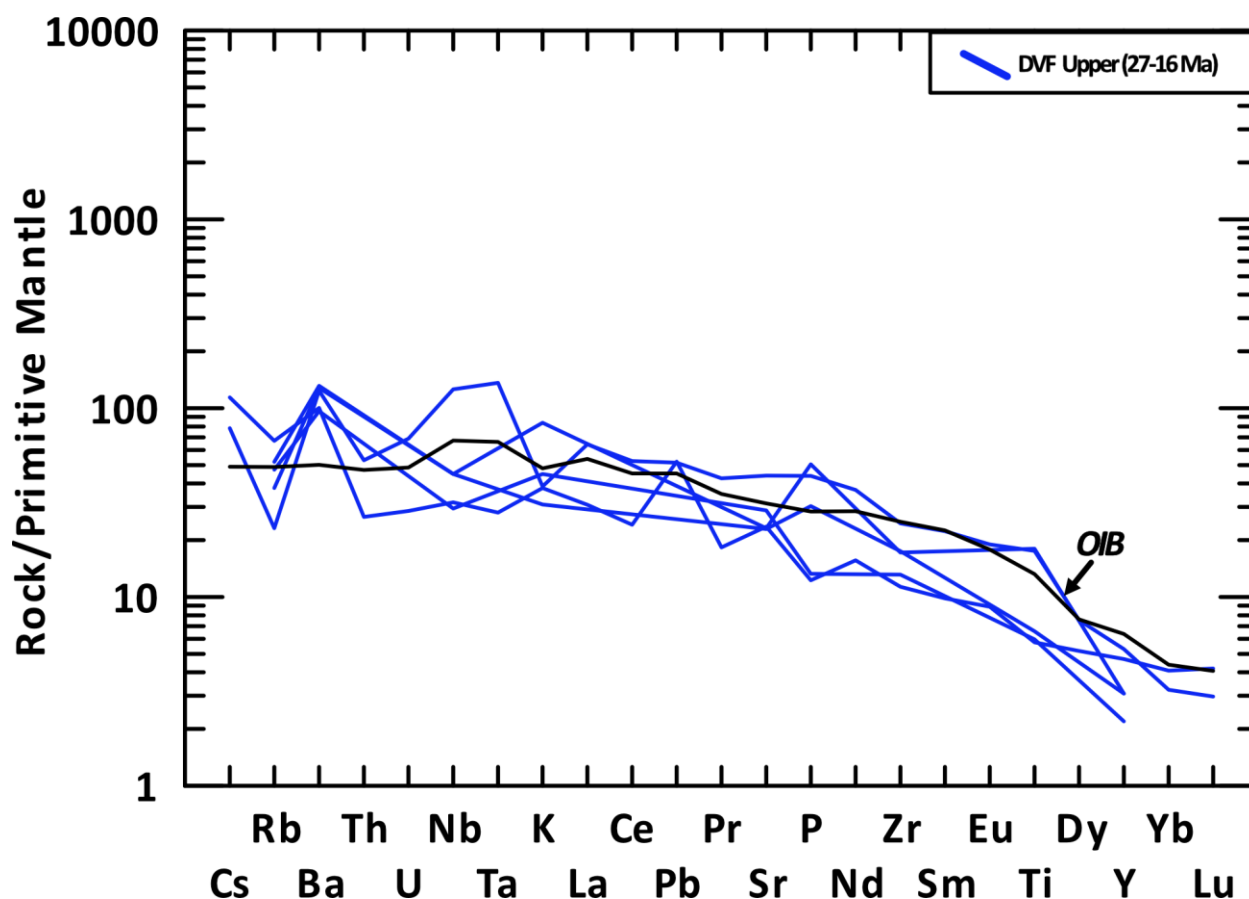


Figure 19: Primitive mantle-normalized (Sun and McDonough, 1989) trace element variations of the upper DVF compared to OIB (Sun and McDonough, 1989).

Sr-Nd Isotopes

Fig. 20 illustrates the $^{87}\text{Sr}/^{86}\text{Sr}_i$ and $^{143}\text{Nd}/^{144}\text{Nd}_i$ variations observed in DVF samples. Measured and age-corrected DVF isotope ratios are provided in Appendix A. Plotted DVF Sr and Nd ratios are age-corrected to known eruptive ages, or if not known, to 30 Ma and are considered initial ratios.

Overall, $^{87}\text{Sr}/^{86}\text{Sr}_i$ values range from 0.704286 to 0.707805 and $^{143}\text{Nd}/^{144}\text{Nd}_i$ values range from 0.512039 to 0.512362. Lower DVF rocks ($^{87}\text{Sr}/^{86}\text{Sr}_i$ 0.707157-0.707793 and $^{143}\text{Nd}/^{144}\text{Nd}_i$ 0.512039-0.512290) plot within the field exhibited by Eocene Challis volcanic rocks of central Idaho (McKervey, 1998) and partially overlap the compositions of Quaternary Snake River plain/Yellowstone basalts (Fig. 20). Middle and upper DVF rocks generally have less radiogenic

$^{87}\text{Sr}/^{86}\text{Sr}_i$ (e.g., <0.7065) but overlapping $^{143}\text{Nd}/^{144}\text{Nd}_i$ values relative to the lower DVF. On exception is middle DVF sample WE24-13, which has $^{87}\text{Sr}/^{86}\text{Sr}_i$ of 0.707805 and $^{143}\text{Nd}/^{144}\text{Nd}_i$ of 0.512166. All DVF rocks have much higher $^{143}\text{Nd}/^{144}\text{Nd}_i$ than Eocene Absaroka volcanics, as well as other Eocene-Oligocene igneous rocks from the central Montana alkalic province (Fig. 20). They also generally have dissimilar Sr and Nd isotope ratios than regional ultramafic xenoliths hosted in central Montana alkalic province volcanics (Fig. 20). Most middle and upper DVF samples have Sr and Nd isotope ratios that resemble the EM1 mantle component (Fig. 20).

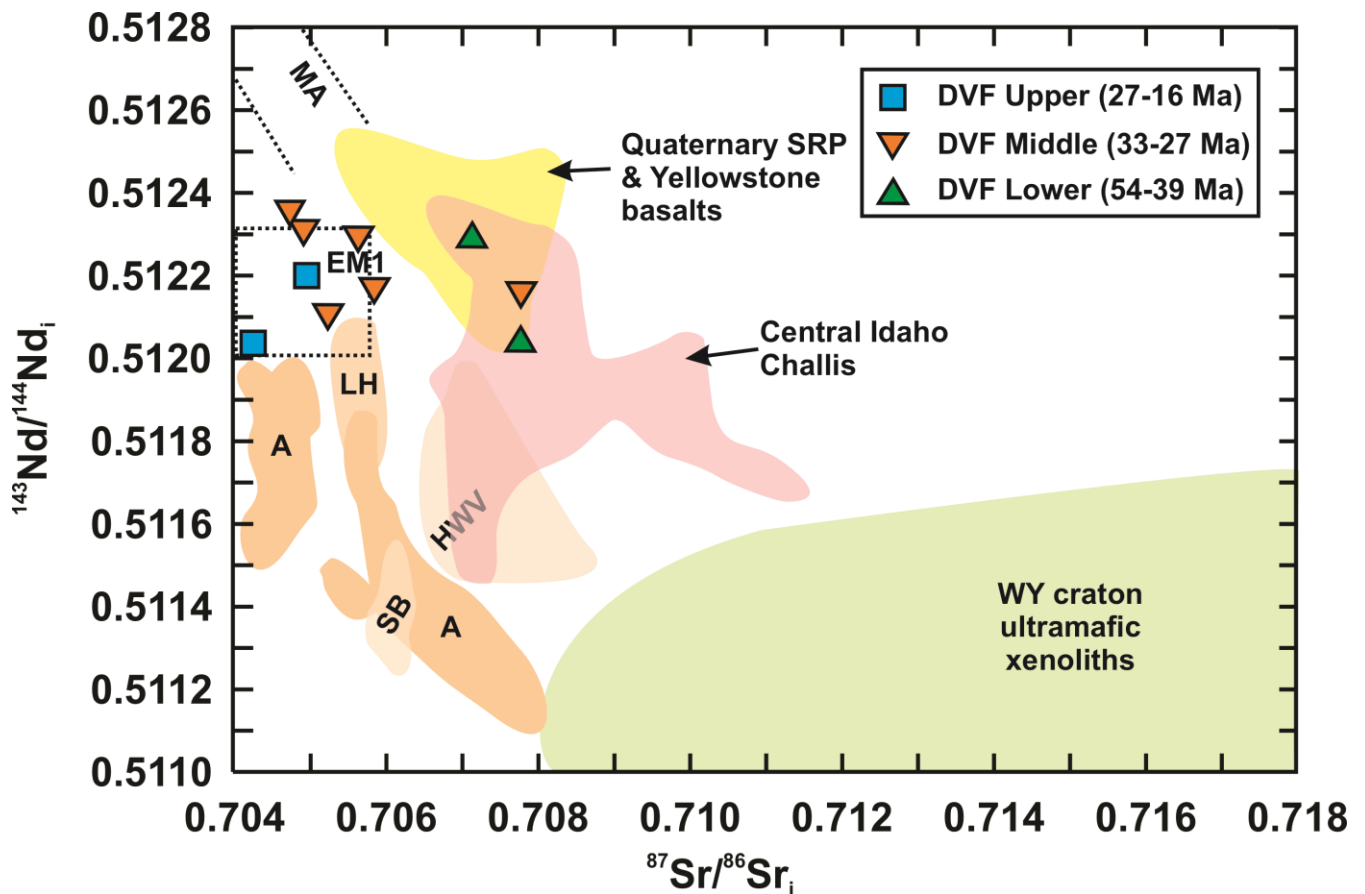


Figure 20: $^{87}\text{Sr}/^{86}\text{Sr}_i$ vs. $^{143}\text{Nd}/^{144}\text{Nd}_i$ for the DVF samples. DVF samples were corrected to published age dates or to 30 Ma with no reported geochronology. Fields depict published data from Yellowstone basalts (McCurry et al., 2009; Swallow et al., 2018), ultramafic xenoliths and Eocene alkalic rocks from the Montana alkalic province (HWV, Highwood Mountains; volcanics; HW/EBP, Highwood Mountains/Eagle Buttes; SB, Smokey Buttes lamproites; Carlson and Irving, 1994; O'Brien et al., 1995); (A), Eocene Absaroka volcanics in northwest Wyoming; (Feely, 2003); Eocene Central Idaho Challis volcanics (McKervey 1998); Pliocene

Leucite Hills lamproites in southwest Wyoming (LH; Mirnejad and Bell, 2006); Wyoming Craton ultramafic xenoliths (Carlson and Irving, 1994) MA, mantle array.

Chapter 4 - Discussion

Dillion volcanic field lavas are spread out across southwestern Montana with major eruptive centers (e.g., Black Butte, Lion Mountain) in the Gravelly Range. Compression during Laramide orogeny and subduction of the Farallon plate generated lower DVF magmatism in southwest Montana. Magmatism continued in southwestern Montana after a transition from compression to extensional tectonics, resulting in concurrent basin sedimentation and eruptions of the middle DVF after an initial ~6 m.y hiatus. The upper DVF represents a renewal of volcanism, coeval with continuation of the extensional tectonic regime. Due to the long eruptive history of the DVF from the Eocene into the Miocene, the DVF magmatism records a shift in tectonomagmatic processes through time. In this study, we investigate whether these postulated tectonic shifts are recorded in the geochemical and isotopic signatures of the lower, middle, and upper DVF.

The lower DVF shows enrichment of LREE and LILE with notable depletions in Ta and Nb, and with enrichment of Pb relative to primitive mantle, all of which are characteristics of continental-arc origins. The middle DVF shows similar LREE and LILE patterns relative to the lower DVF except for them having higher Cs and Ba and shows less depletion of Nb and Ta. The upper DVF shows less enrichment of LREE and LILE and more closely matches average OIB than either of the lower or middle DVF. The upper DVF also has the least amount of Pb enrichment.

Crustal Contamination Effects

Though basaltic magmas may contain geochemical traces of their tectonomagmatic origins, those that form beneath continents may become contaminated by continental crust/continental lithosphere as they ascend to the surface, which can impart subduction-like geochemical signatures (e.g., low Nb, Ta, Ti; Brueseke and Hart, 2009). This can lead to the misidentification if basalts have been contaminated (Xia and Li, 2019). Some DVF samples display visual evidence for crustal interaction, as observed in both the field and at a microscopic scale (e.g., Renova sediment clasts in the Lion Mountain Dike, quartz/crustal xenoliths in some rocks, etc.). Any samples with potential influence of crustal contamination on the geochemical and Sr-Nd isotopic signatures of DVF magmas must be evaluated.

The K/P ratio is useful for determining potential upper crustal contamination since the ratio remains below 3.5 for uncontaminated mantle-derived rocks but exceeds 10 for andesitic compositions and for basalts that have been contaminated by K-rich upper crust (Carlson and Hart, 1987).

All but three of the DVF samples show K/P ratios >3.5 with the lower DVF having the highest K/P ratios (Fig. 21). Only the middle DVF sample WE24-1 and upper DVF samples LM-14 (Fritz et al., 2007) and WE24-16 fall below 3.5, suggesting limited crustal interaction. However, the K/P ratio may be misleading for DVF rocks, due to their high-K concentrations and the alkaline to transitional nature of many DVF rocks.

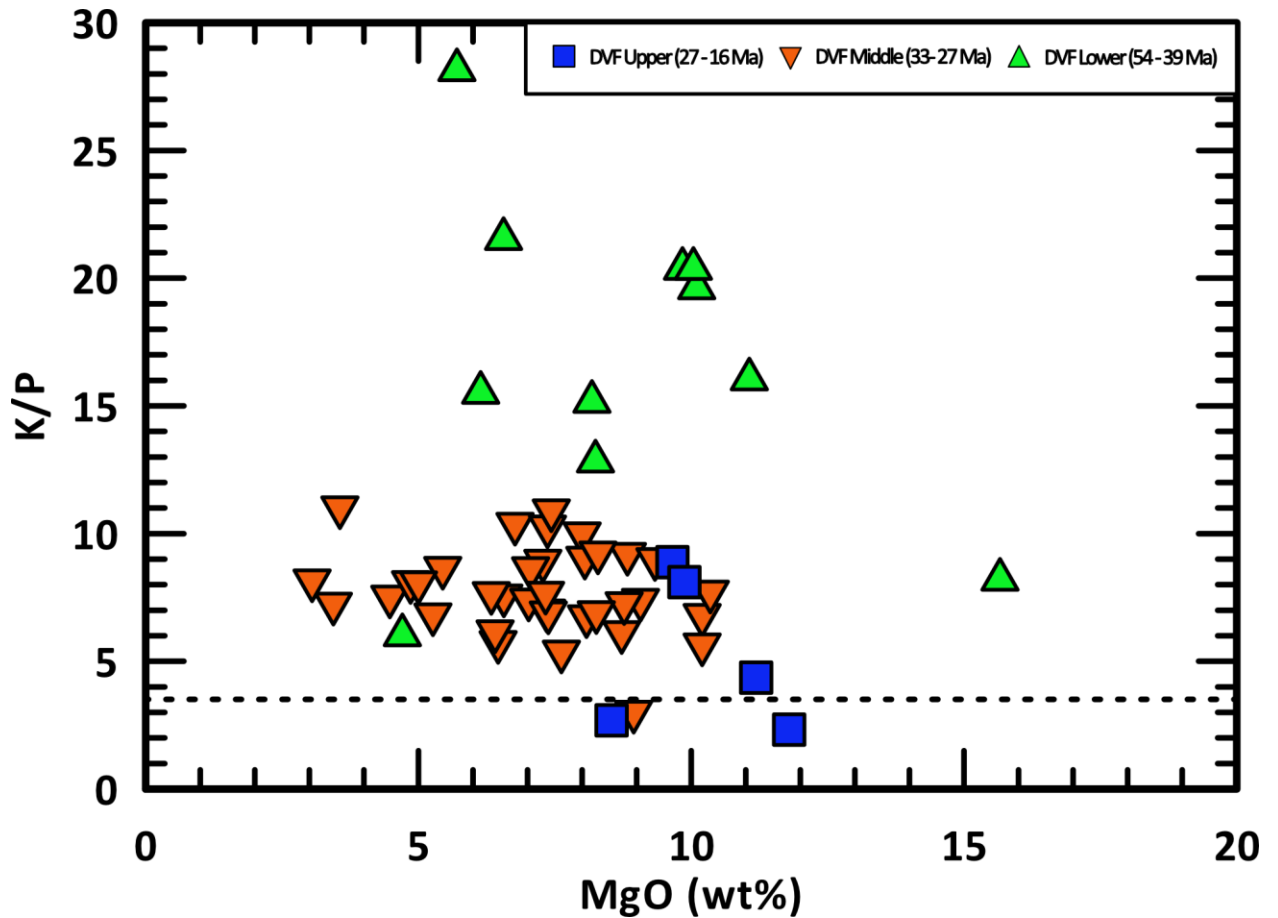


Figure 21: K/P (ppm) over wt% MgO representing samples that may have crustal contamination, ratios >3.5, based on Carlson and Hart (1987).

Thus, other proxies are needed to assess the degree of crustal interaction. Here, we use Ce/Pb and La/Nb values (Fig. 22) to depict crustal contamination from both upper and lower crustal lithologies (Rooney et al., 2007). Lower crustal contamination is associated with La/Nb ratio > 2 and variable Ce/Pb ratios, while samples with Ce/Pb = 25 ± 5 represent both a MORB and OIB mantle array (Hofmann et al., 1986; Rooney et al., 2007). Basalts that are affected by lower crustal contamination have Ce/Pb < 20 while continental crust has a Ce/Pb = 4 (Hofmann et al., 1986; Rooney et al., 2007). All DVF samples but two display evidence for crustal contamination (Ce/Pb < 20 Fig. 22).

The ten selected samples of DVF rocks that have reported $^{87}\text{Sr}/^{86}\text{Sr}_i$, are paired with La/Nb and Ce/Pb to further evaluate crustal interaction (Fig. 23). Higher La/Nb with higher $^{87}\text{Sr}/^{86}\text{Sr}_i$ suggests crustal contamination and lower Ce/Pb with higher $^{87}\text{Sr}/^{86}\text{Sr}_i$ also suggests crustal contamination. One upper DVF sample (WE24-16) lies predominantly within the mantle field (Ce/Pb 25 ± 5) and cannot represent contamination with crustal materials, nor is it likely that the single middle DVF sample with high Ce/Pb (>30) was affected by contamination (Fig. 22).

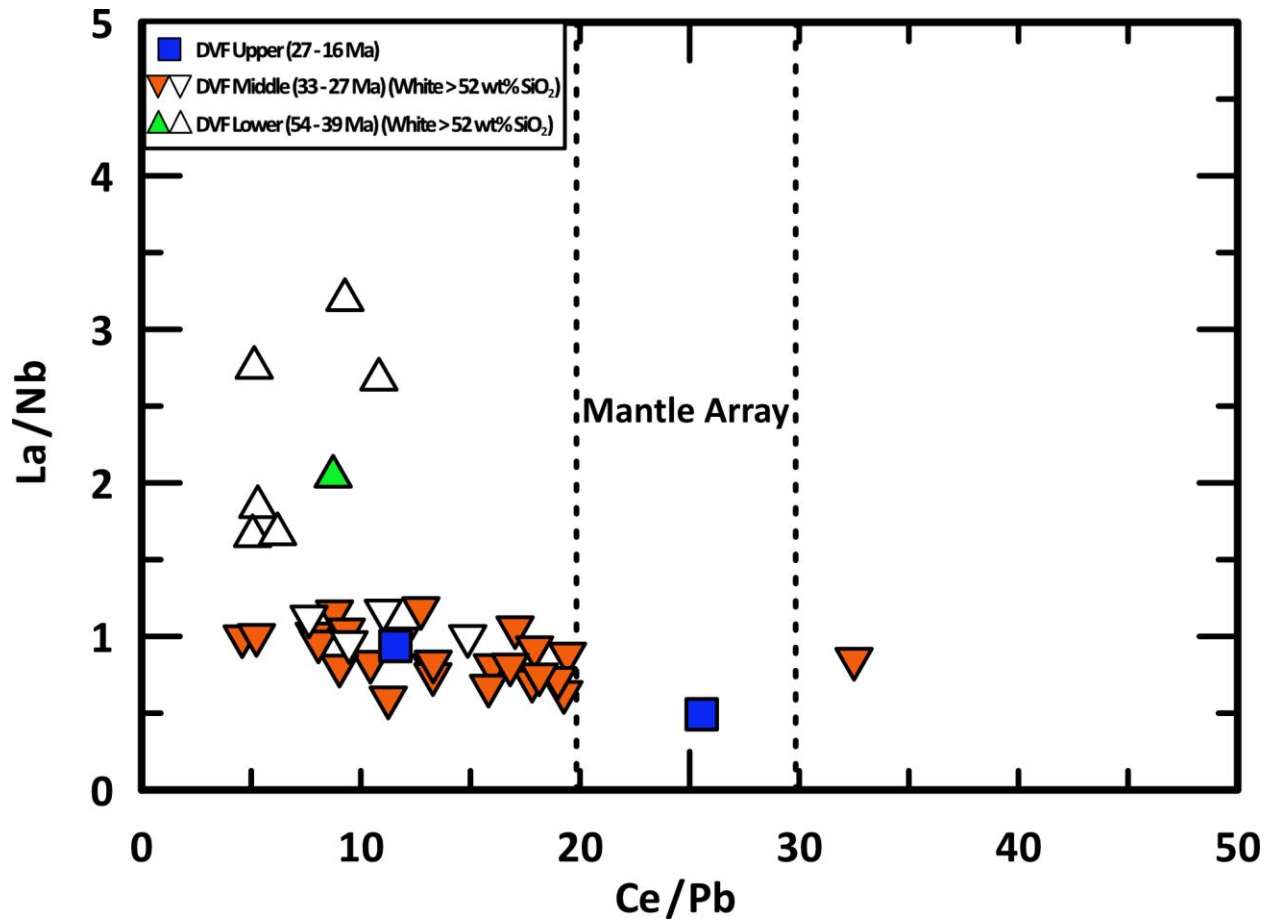


Figure 22: Crustal contamination indicators, Ce/Pb and La/Nb. Figure based on Rooney et al. (2007) plotted for DVF samples.

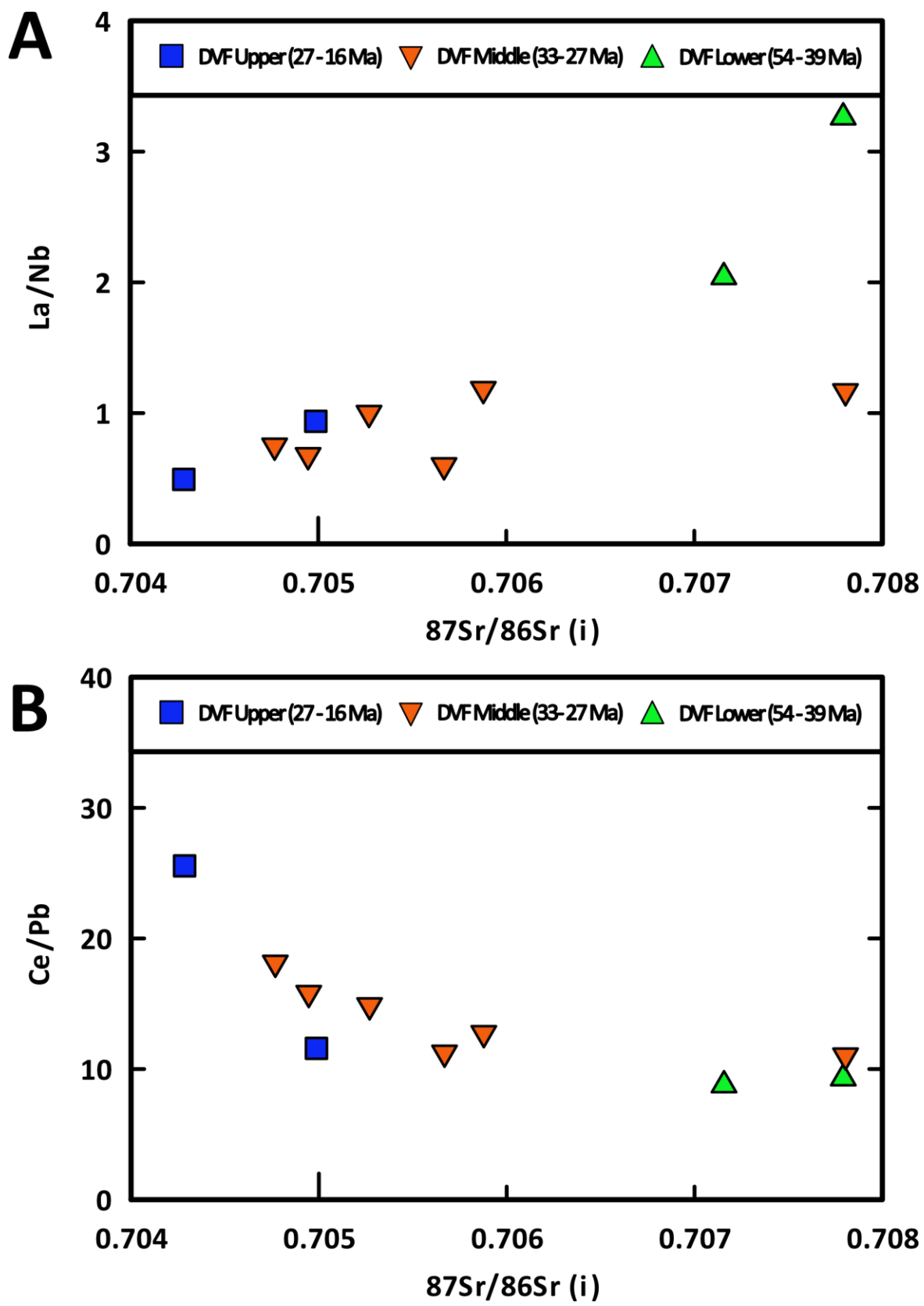


Figure 23: Crustal contamination indicators (A) La/Nb vs. $^{87}\text{Sr}/^{86}\text{Sr}_i$ (B) Ce/Pb vs. $^{87}\text{Sr}/^{86}\text{Sr}_i$

Based off the prior indicators, DVF rocks have some level of crustal contamination. To determine the amount of crustal assimilation into DVF magmas, we have modelled the degree of assimilation using the assimilation and fractional crystallization (AFC) equation of DePaolo (1981). The model results are presented in (Figs. 24-29). Endmembers used in the calculations are sample WE24-1 (Black Butte basalt) and an Archean quartzofeldspathic gneiss from the western Owl Creek Mountains Wyoming (980C2 from Frost et al., 2006; e.g. “upper crust”), a middle crustal xenolith from 49 Ma Lone Mountain Montana (Dudás et al. (2023) and a lower crustal norite (B-1, Leeman et al., 1985). WE24-1 was selected to represent the original magma being contaminated and fractionated since it has low K/P ratios (<3.5) and represents a more primitive sample ($Mg\# = 62.9$) from a volcanic plug. The gneiss was selected as a source of possible upper crustal contamination in DVF rocks, because it represents Wyoming Craton crust exposed at or near the surface via Laramide compressional faulting. To represent Wyoming Craton middle crustal contamination (~ 10 -20 km), a hornblendite xenolith from the Lone Mountain Montana andesite, exposed ~ 40 km east of the Gravelly range (sample 18BS02 from Dudás et al. 2023), was selected. As mentioned, the norite (sample B-1 from the nearby Spencer-Kilgore region, Idaho, Leeman et al. 1985) was selected to represent lower crustal contamination.

For all AFC model figures, the F value (% of liquid magma remaining) is 0.9 and the r values (mass assimilate / mass crystallized) are 0.005, 0.01, and 0.05 (e.g., 0.5 to 5% assimilation). For these calculations, sample WE24-1 (Black Butte) was used to represent primitive basalt with a crystallizing assemblage of olivine (13.4%) + plagioclase (5.84%) + clinopyroxene (2.34%) + orthopyroxene (0.13%) + oxides (0.52%). These mineral proportions

were calculated via thin section point counting of phenocrysts at 1 mm intervals. Samples were selected to act as end members to model with WE24-1 at 0.5% to 5% assimilated based on physical characteristics (e.g., crustal xenoliths), age and locations of the samples. Bulk partition coefficients values (D) are shown in Table 1. AFC modeling using WE24-1 (Black Butte) as a parental magma shows that fractional crystallization and assimilation of upper crustal gneiss, middle crust hornblende, and lower crust norite do not sufficiently change the trace element geochemistry of WE24-1 to replicate values observed in other samples (e.g., WE24-5, AE21-5, WE24-16) and suggest up to ~5% assimilation of crustal materials incorporated. Any more assimilation yields dissimilar modelled geochemical compositions.

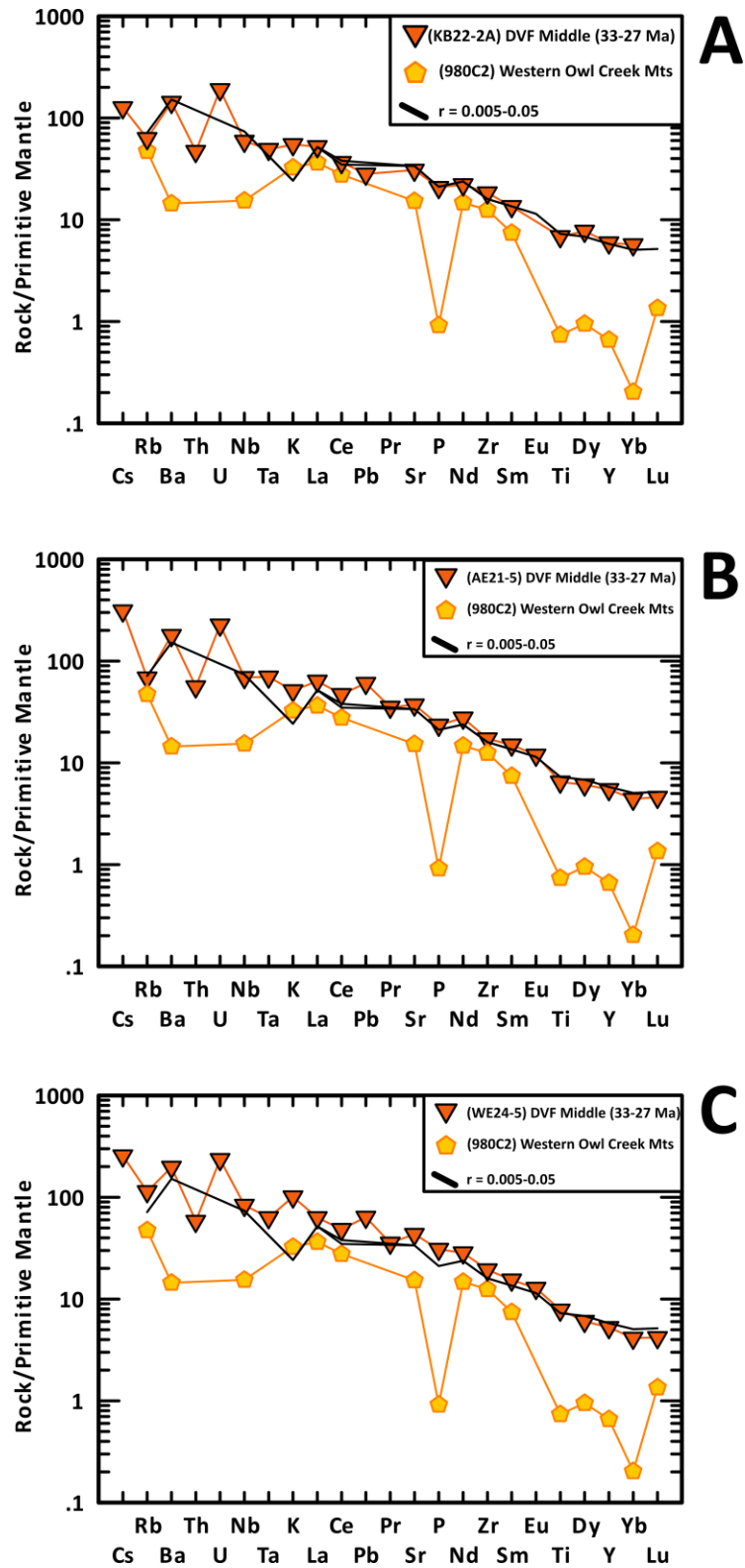


Figure 24: Results from AFC modeling (after DePaolo 1981) between a middle DVF (KB22-2A, AE21-5, WE24-5) and a potential upper crustal contaminant, Gneiss 980C2 from Frost et al. (2006). Primitive mantle after Sun and McDonough, (1989).

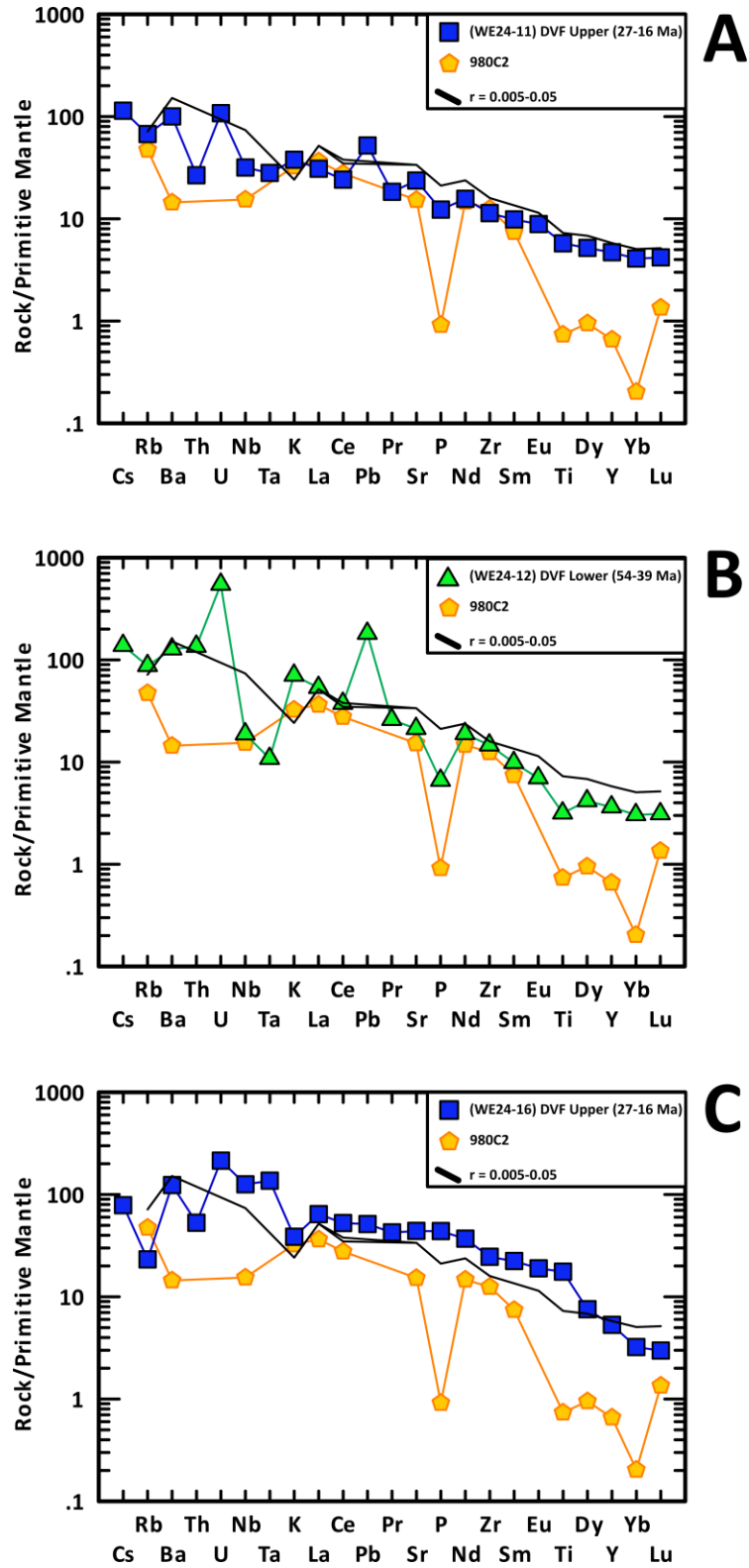


Figure 25: Results from AFC modeling (after DePaolo 1981) between upper DVF (WE24-11, WE24-16), lower DVF (WE24-12) and a potential upper crustal contaminant, Gneiss 980C2 from Frost et al. (2006). Primitive mantle after Sun and McDonough, (1989).

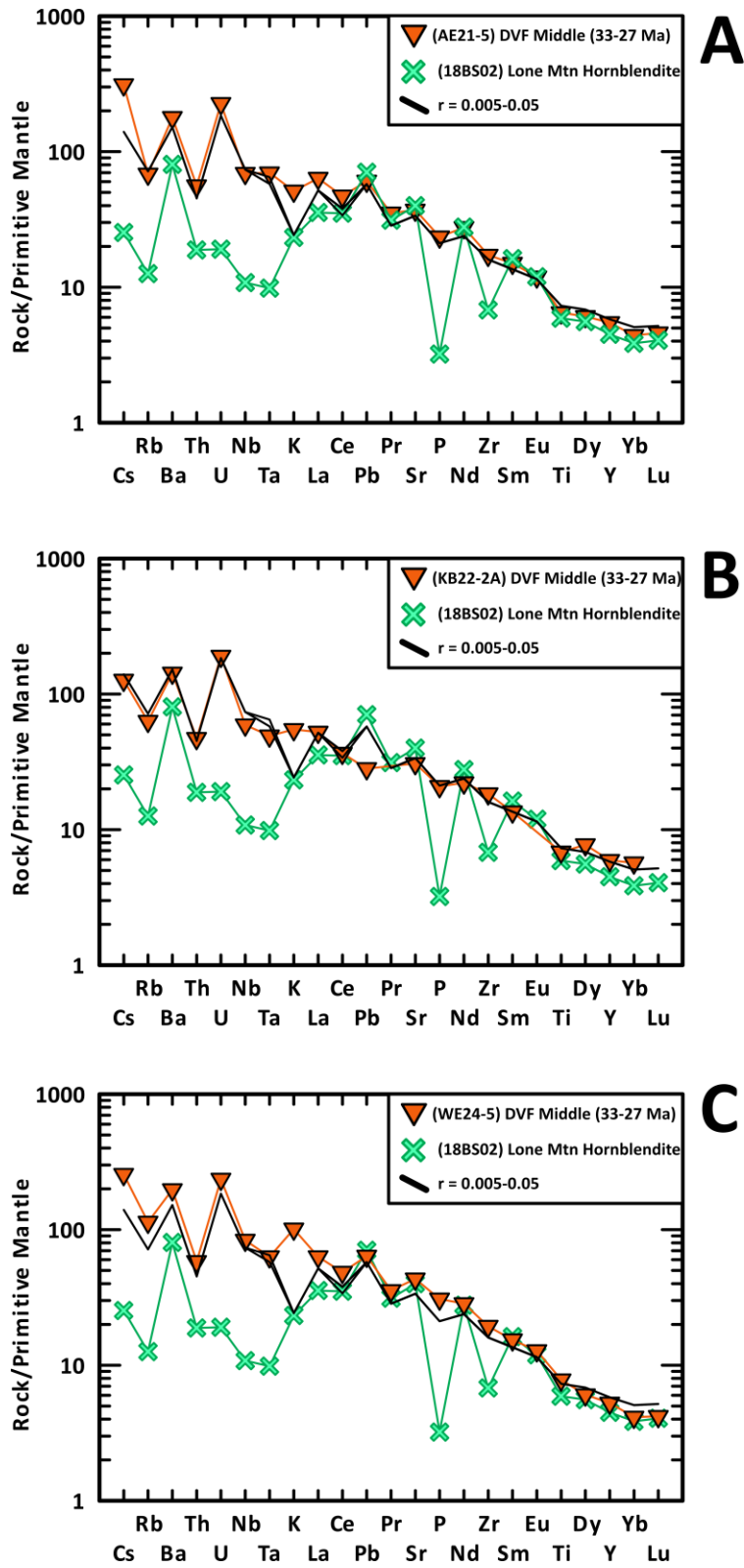


Figure 26: Results from AFC modeling (after DePaolo 1981) between a middle DVF (KB22-2A, AE21-5, WE24-5) and a potential middle crustal contaminant, Hornblende 18BS02 from Dudás et al. (2023). Primitive mantle after Sun and McDonough, (1989).

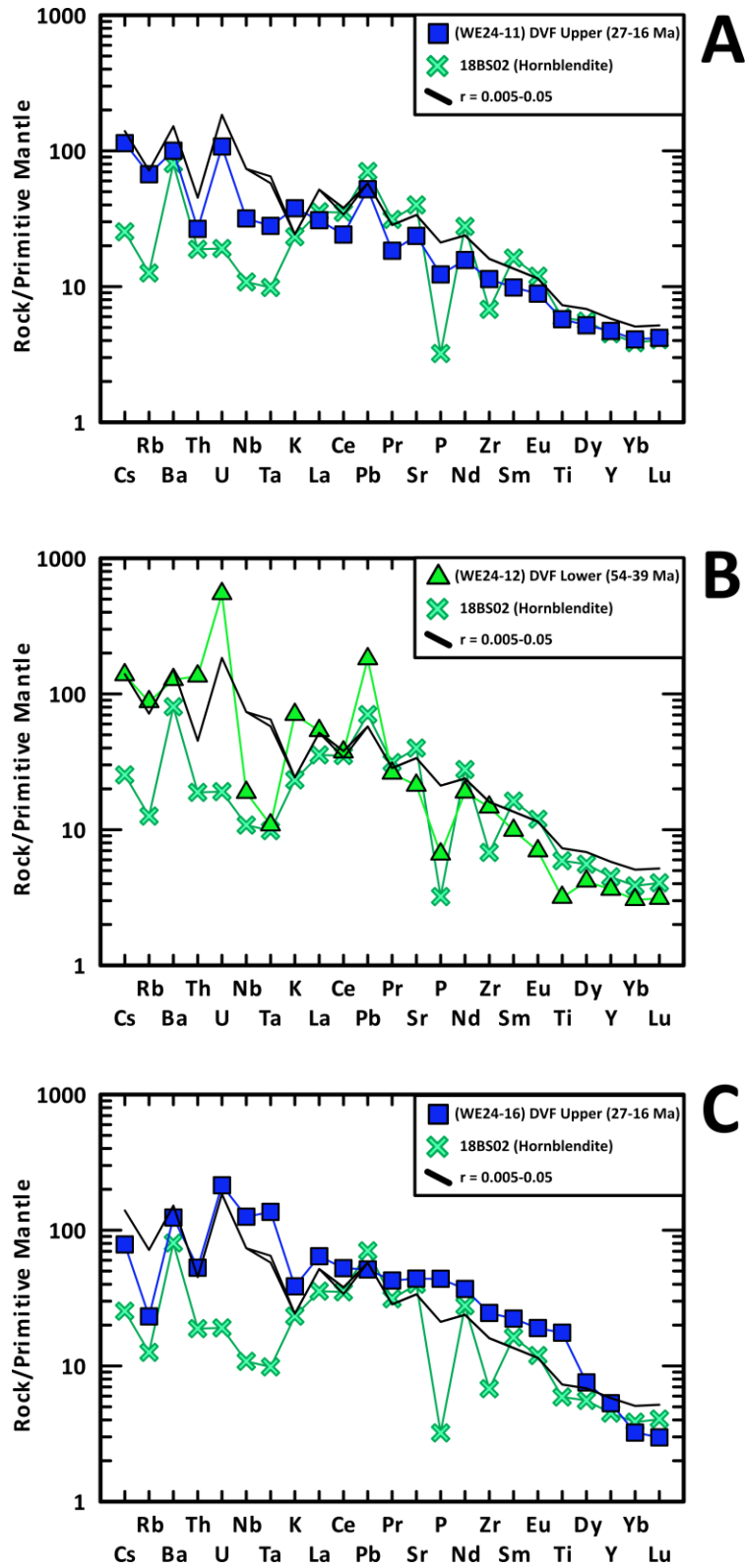


Figure 27: Results from AFC modeling (after DePaolo 1981) between upper DVF (WE24-11, WE24-16), lower DVF (WE24-12) and a potential middle crustal contaminant, Hornblende 18BS02 from Dudás et al. (2023). Primitive mantle after Sun and McDonough, (1989).

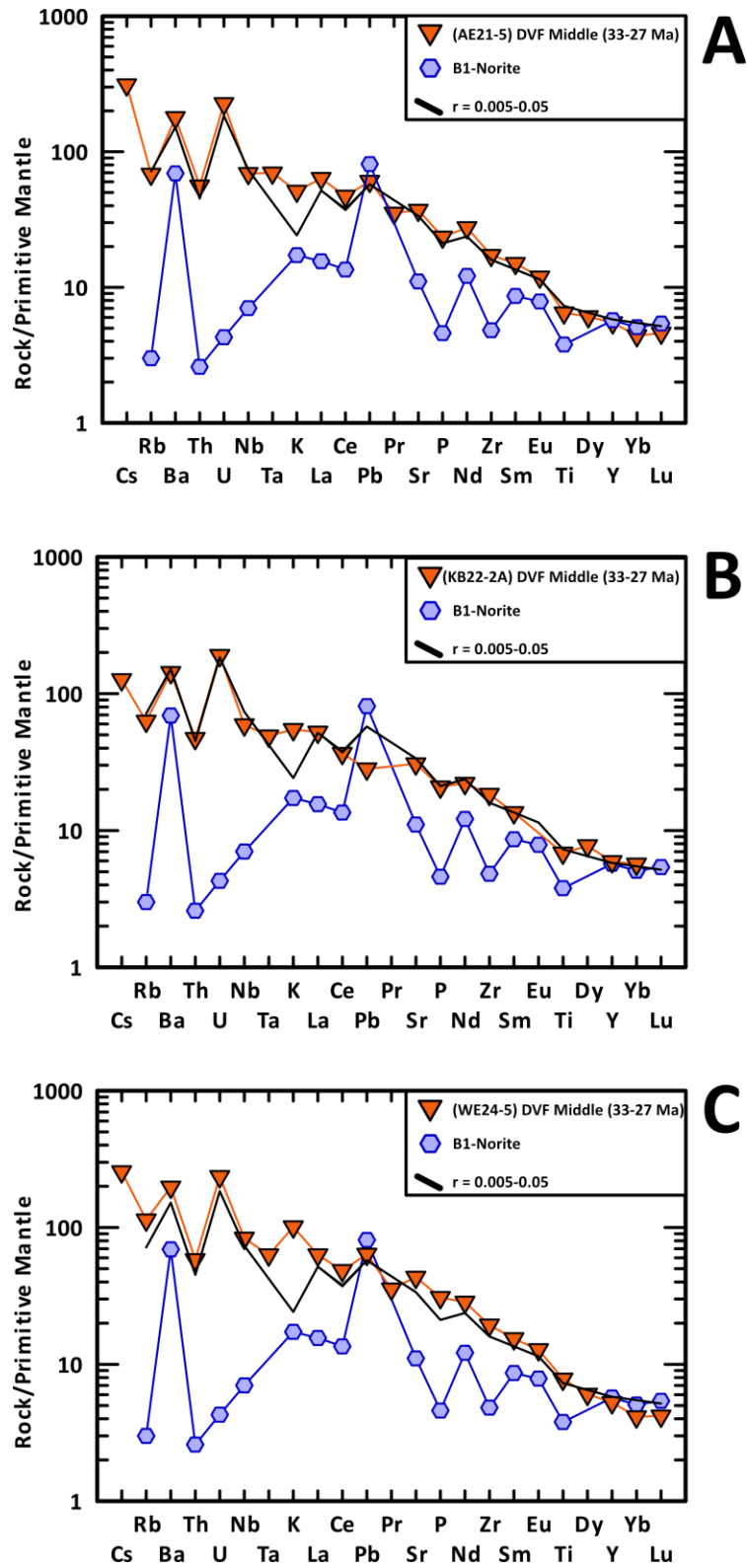


Figure 28: Results from AFC modeling (after DePaolo 1981) between a middle DVF (KB22-2A, AE21-5, WE24-5) and a potential lower crustal contaminant, Norite B-1 from Leeman et al. (1985). Primitive mantle after Sun and McDonough, (1989).

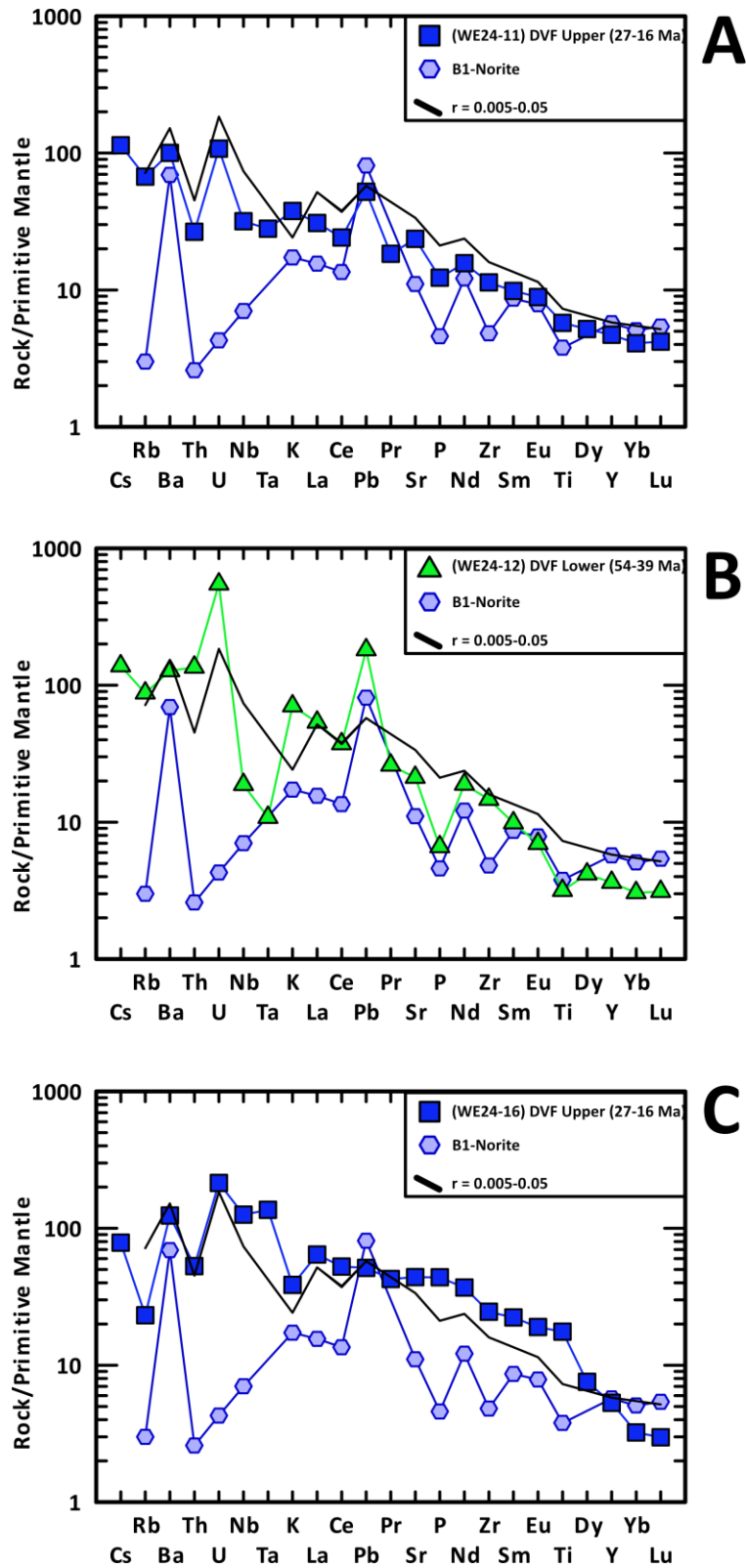


Figure 29: Results from AFC modeling (after DePaolo 1981) between upper DVF (WE24-11, WE24-16), lower DVF (WE24-12) and a potential lower crustal contaminant, Norite B-1 from Leeman et al. (1985). Primitive mantle after Sun and McDonough, (1989).

Name	AE21-5	KB22-2A	WE24-5	980C2	18BS02	B-1	WE24-1	
Location	Centennial Valley	Gravelly Range	Gravelly Range	Owl Creek Mtn Upper Crust	Lone Mtn Middle Crust	Norite Lower Crust	Gravelly Range	Bulk Partition Coefficient
	Contaminant				Contaminant	Contaminant	Source	
Cs	2.46	1	2.02	N.R	0.2	N.R	1.01	0.161
Rb	43.4	40	72.2	30	8	1.9	41	0.032
Ba	1249	998	1381	101	562	484	962	0.066
Th	4.76	4	4.95	N.R	1.6	0.22	3.49	0.085
U	4.76	4	4.95	N.R	0.4	0.09	0.96	0.004
Nb	49.3	42.2	59.63	11	7.7	5	47.45	0.029
Ta	2.84	2	2.57	N.R	0.4	N.R	2.13	0.041
K	12793	13703	25285	8135	5811	4317	5466	0.059
La	43.67	36	43.36	25	24.4	10.7	32.17	0.053
Ce	83.57	65	86.02	49.2	62.3	24	60.78	0.045
Pb	4.3	2	4.53	N.R	5	5.76	3.84	0.430
Pr	9.74	N.R	9.82	N.R	8.6	N.R	7.1	0.067
Sr	783	649	915	322	842	233	669	0.448
P	2232	1978	2927	87	305	436	1812	0.035
Nd	37.31	30	38.58	19.9	37.5	16.4	29.13	0.075
Zr	193	206	217	140	76	54	161	0.017
Sm	6.67	6	6.84	3.3	7.2	3.83	5.47	0.116
Eu	2	N.R	2.15	0.25	2	1.32	1.75	0.136
Ti	8427	8886	10146	959	7674	4916	8571	0.062
Dy	4.48	5.7	4.46	0.7	4.1	N.R	4.6	0.162
Y	24.92	27	23.93	3	20.4	26	25	0.484
Yb	2.17	2.8	2.04	0.1	1.9	2.5	2.29	0.169
Lu	0.34	N.R	0.31	0.1	0.3	0.39	0.35	0.146
Mineral	Olivine	Clinopyroxene	Orthopyroxene	Plagioclase	Oxide			
Modes (%)	13.4	2.34	0.13	5.84	0.52			

Table 1: AFC Modeling data. Lower crust, (B-1 Norite, Leeman et al., 1985), middle crust, (18BS02 Hornblendite, Dudás et al., 2023), upper crust, (980C2, gneiss, Frost et al., 2006). Mineral-melt partition coefficients for trace elements in basalt melts from Rollinson and Pease (2021). N.R. = not reported. Element concentrations in ppm.

Name	WE24-11	WE24-12	WE24-16	980C2	18BS02	B-1	WE24-1	
Location	Madison Valley	Block Mountain lava	Big Hole River Valley	Owl Creek Mtn Upper Crust	Lone Mtn Middle Crust	Norite Lower Crust	Gravelly Range	Bulk Partition Coefficient
	Contaminant				Contaminant	Contaminant	Source	
Cs	0.90	1.09	0.62	N.R	0.2	N.R	1.01	0.161
Rb	42.7	55.5	14.7	30	8	1.9	41	0.032
Ba	701	890	865	101	562	484	962	0.066
Th	2.26	11.49	4.5	N.R	1.6	0.22	3.49	0.085
U	2.26	11.49	4.5	N.R	0.4	0.09	0.96	0.004
Nb	22.62	13.37	89.68	11	7.7	5	47.45	0.029
Ta	1.14	0.44	5.55	N.R	0.4	N.R	2.13	0.041
K	9438	17609	9640	8135	5811	4317	5466	0.059
La	21.19	36.76	44.18	25	24.4	10.7	32.17	0.053
Ce	42.86	66.18	93.27	49.2	62.3	24	60.78	0.045
Pb	3.7	12.86	3.65	N.R	5	5.76	3.84	0.430
Pr	5.06	7.17	11.75	N.R	8.6	N.R	7.1	0.067
Sr	498	447	915	322	842	233	669	0.448
P	1167	626	4165	87	305	436	1812	0.035
Nd	21.21	25.56	50.01	19.9	37.5	16.4	29.13	0.075
Zr	127	164	275	140	76	54	161	0.017
Sm	4.37	4.38	9.91	3.3	7.2	3.83	5.47	0.116
Eu	1.49	1.17	3.19	0.25	2	1.32	1.75	0.136
Ti	7469	4108	22830	959	7674	4916	8571	0.062
Dy	3.82	3.07	5.56	0.7	4.1	N.R	4.6	0.162
Y	21.37	16.49	24.15	3	20.4	26	25	0.484
Yb	2.01	1.50	1.59	0.1	1.9	2.5	2.29	0.169
Lu	0.31	0.23	0.22	0.1	0.3	0.39	0.35	0.146
Mineral	Olivine	Clinopyroxene	Orthopyroxene	Plagioclase	Oxide			
Modes (%)	13.4	2.34	0.13	5.84	0.52			

Table 2: AFC Modeling data. Lower crust, (B-1 Norite, Leeman et al., 1985), middle crust, (18BS02 Hornblendite, Dudás et al., 2023), upper crust, (980C2, gneiss, Frost et al., 2006). Mineral-melt partition coefficients for trace elements in basalt melts from Rollinson and Pease (2021). N.R. = not reported. Element concentrations in ppm.

All DVF samples in the AFC models were calculated with 0.05% to 5% assimilated, with a fractionation of 90%. These results from the AFC modeling indicate that fractional crystallization and assimilation of local-regional upper, middle, and lower crustal contaminants (980C2 western Owl Creek Mts, 18BS02 Lone Mtn hornblendite, and B-1 Norite, respectively) show that at most ~5% crustal contamination can occur and replicate the measured element concentrations of DVF rocks. Although upper crustal xenoliths, such as Renova formation, are

found within samples like WE24-5, they likely had minimal effect on the trace element geochemistry of DVF basalts.

Tectonomagmatic Indicators for The DVF - Lithosphere Mantle and Asthenosphere Sources

Trace element geochemistry and Sr-Nd isotopes ratios are used to determine melt origins of the Dillon mafic rocks and if melt sources changed through time (e.g., upper vs. lower Dillon) in response to evolving tectonomagmatism. To help clarify these possibilities, DVF samples are compared to the Pliocene-Quaternary basalts from the Basin and Range province since these small-volume alkali basalt eruptions have an unequivocally asthenospheric origin and erupted during lithosphere thinning during ongoing Basin and Range extension. Central Nevada ~6-3 Ma Reveille Range basalts (Yogodzinski et al., 1996) and basalts from the 4 to 1.1 Ma Buffalo Valley volcanic field, northern Nevada (Cousens et al., 2013) provide a good proxy for geochemical and isotope ratios that represent North American Cordillera asthenospheric mantle.

The lower, middle and upper DVF rocks are displayed with Reveille Range basalts and the Buffalo Valley volcanic field on a primitive-mantle-normalized diagram (Fig. 30-32). The lower DVF exhibits enrichment in Cs and Rb, and more depletions in Ti than the Basin and Range volcanic fields (Fig. 30). The middle DVF displays higher enrichment in Cs and Ba in comparison to the Basin and Range volcanic fields (Fig. 31). The upper DVF rocks display the most similar trace element signatures of both the Basin and Range rocks plotting within their arrays (Fig. 32).

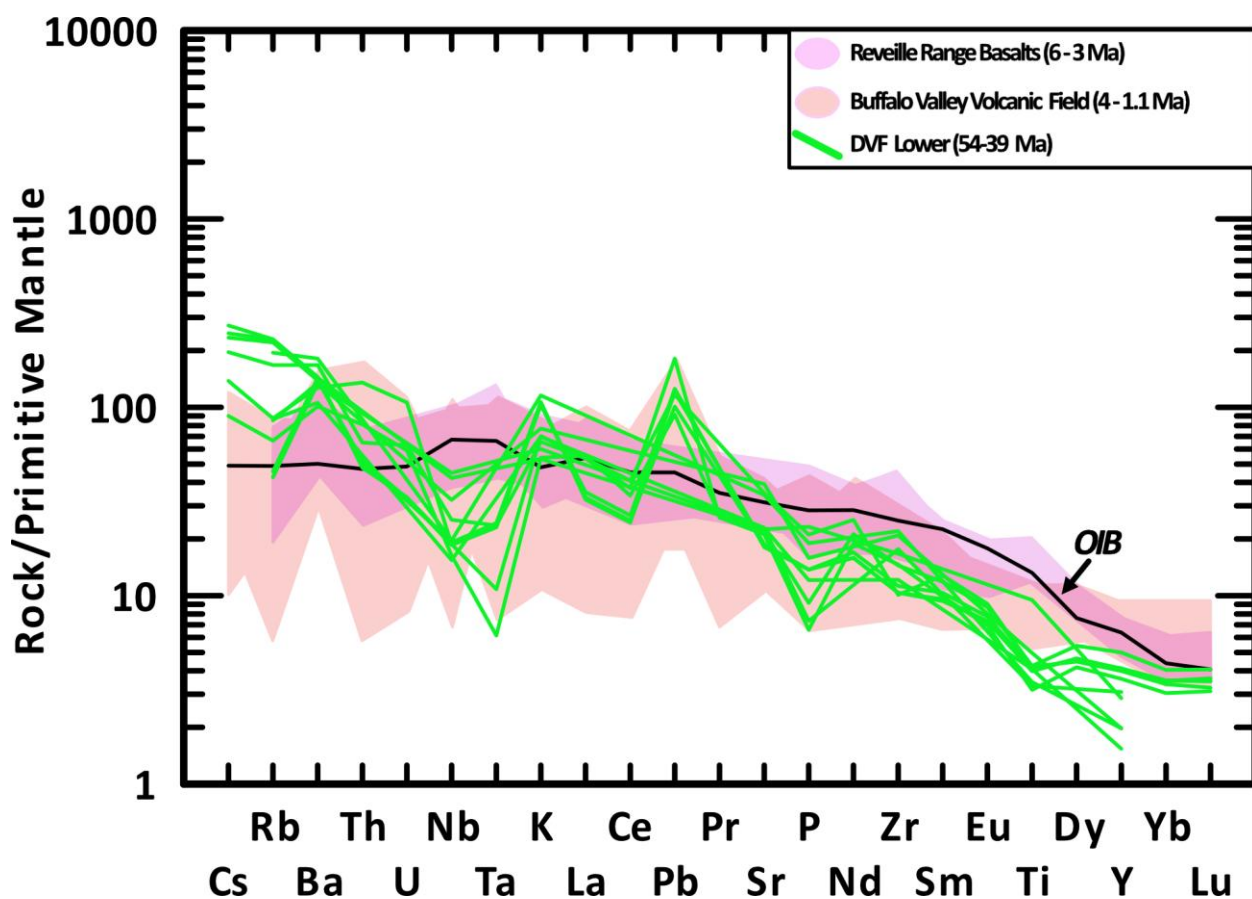


Figure 30: A primitive mantle-normalized multi-element diagram (normalizing values from Sun and McDonough 1989) displaying rocks from the lower DVF in comparison to arrays from the Reveille Range Basalts and the Buffalo Valley Volcanic Field.

Compared to the younger members of the DVF, the lower DVF displays enrichment in LREE and LILE with depletion of Nb, Ta, and Ti (Fig. 30), which we interpret to reflect slab derived components due to subduction/arc-related magmatism. Accessory phases like rutile, which is stable in metasomatized lithospheric mantle and subduction-affected asthenosphere (Farmer et al., 2020), retain Ta during the melting of this mantle, which causes a depletion of Ta and similar high field strength elements in magma.

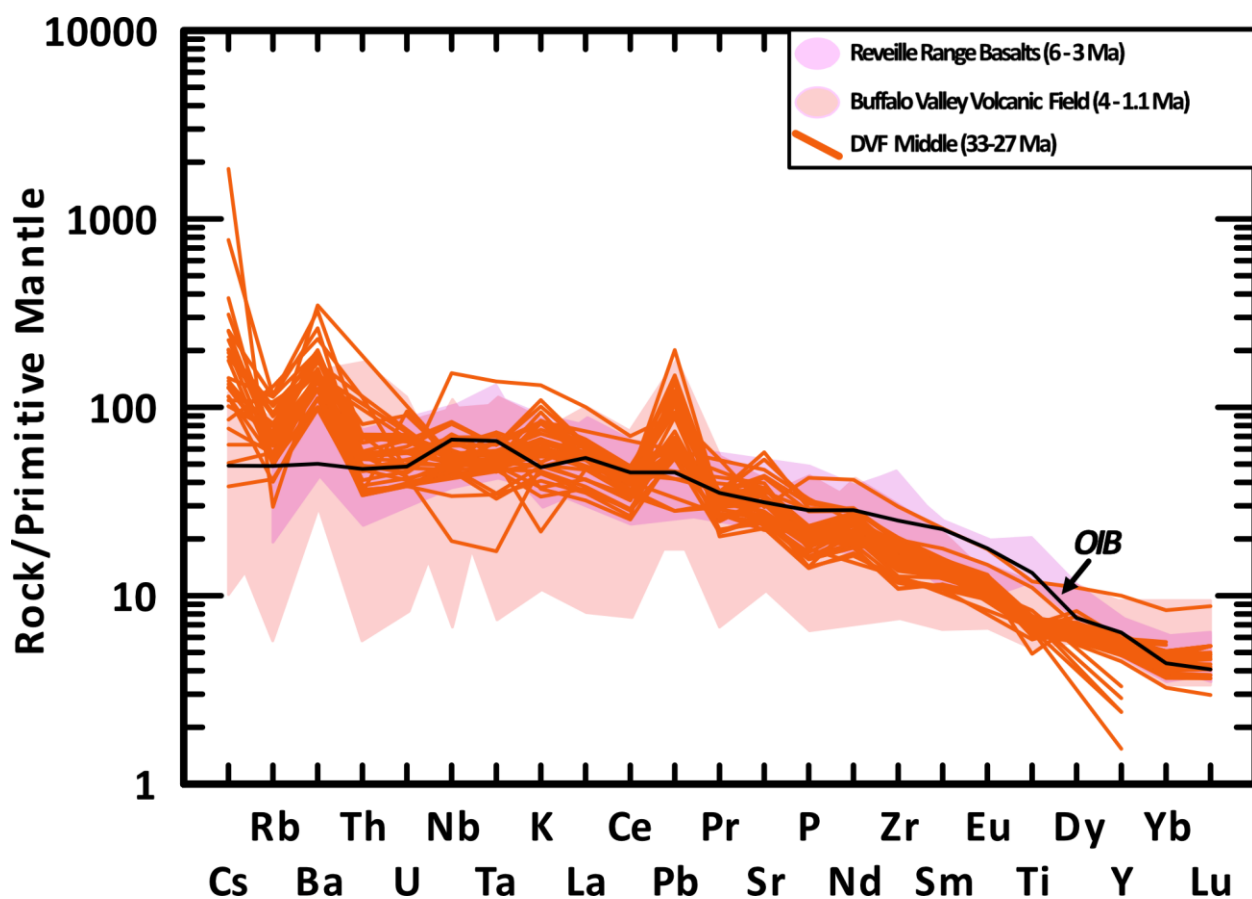


Figure 31: A primitive mantle-normalized multielement diagram (normalizing values from Sun and McDonough 1989) displaying rocks from the middle DVF in comparison to arrays from the Reville Range Basalts and the Buffalo Valley Volcanic Field.

The middle DVF (Fig. 31) shows similar rare earth element patterns to lower DVF rocks which suggest either an enriched source or crustal contamination, with notably less Nb and Ta depletions compared to the lower DVF (Figs. 30-31). Based on AFC modelling (Figs. 23-28), crustal contamination of these rocks is minimal. Some middle DVF rocks have high positive K anomalies, specifically rocks in the Gravelly Range, such as the Lion Mountain dike (WE24-5: Fig. 7). The high positive K anomalies in these rocks could originate due to a component of melts of metasomatized Wyoming Craton lithosphere, as observed in other Cenozoic igneous rocks in the region (Feeley, 2003; Apen et al., 2024). The middle DVF shows less depletion of Nb and Ta

than the lower DVF, which suggests a reduced subduction component influence. Middle DVF samples also show enrichment of Cs however, Cs and other large ion lithophile elements (e.g., Ba, Rb, and Sr) are also readily mobile during surface weathering and alteration and may be an artifact of alteration or potential crustal contamination. (Xia and Li, 2019).

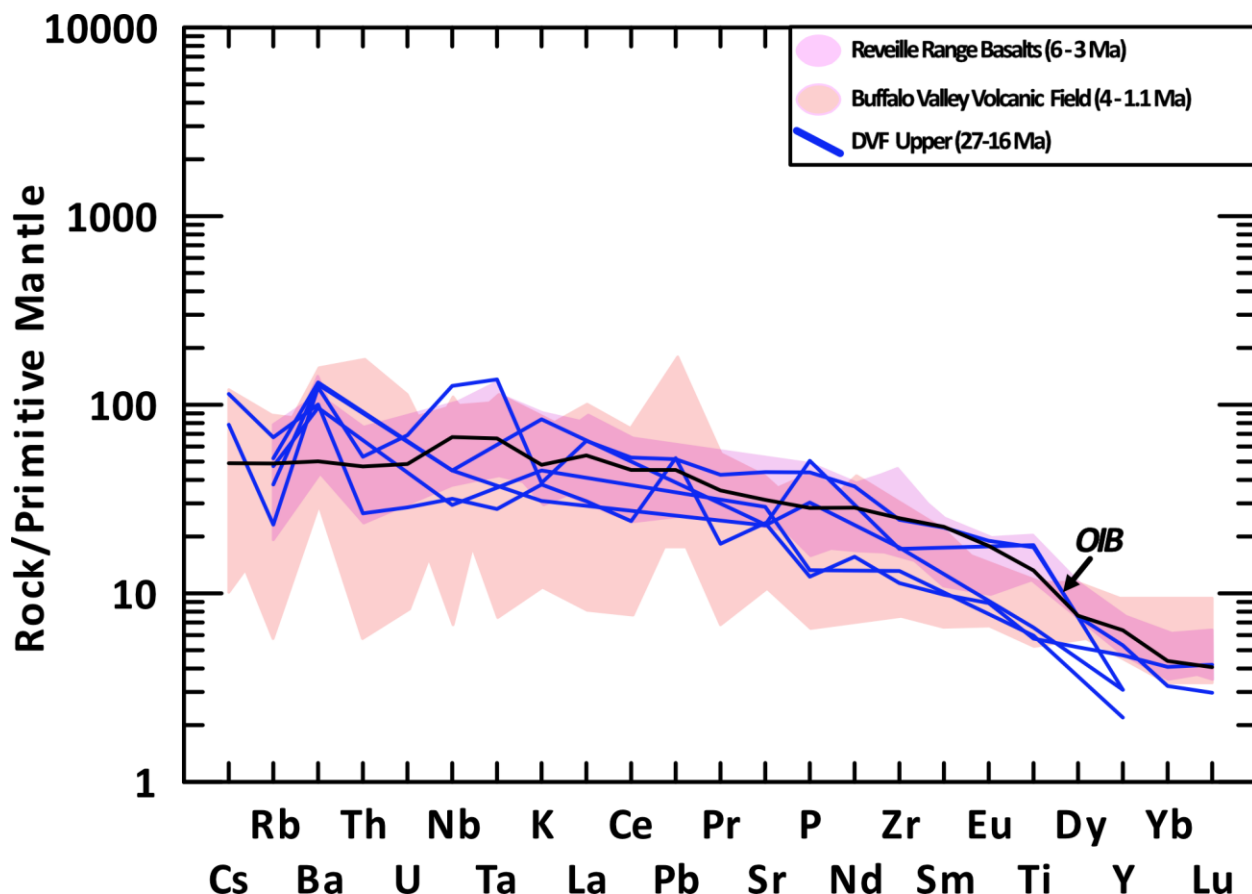


Figure 32: A primitive mantle-normalized multi-element diagram (normalizing values from Sun and McDonough 1989) displaying rocks from the upper DVF in comparison to arrays from the Reville Range Basalts and the Buffalo Valley Volcanic Field.

The upper DVF displays the best fit of trace elements in comparison to Basin and Range basalts, especially those from the Reville Range (e.g., Lunar crater). Overall, these rocks display less enrichment in LREE and LILE (Fig. 32) than the older DVF suites, consistent with

derivation from a mantle less affected by slab-derived components. The upper DVF displays the lowest enrichment in Pb compared to the other DVF members also suggesting less crustal input.

Incompatible elements La, Nb, and Ba are useful for determining if melts are from an OIB-type (asthenospheric) mantle or from more enriched lithospheric mantle. Lithospheric mantle is enriched in Ba and depleted in Nb and Ti (Fitton et al., 1991). Continental basalts that originate from melting of enriched metasomatized lithospheric mantle will deviate from OIB affinities and will be variably enriched in Rb, Ba, and K but depleted in Nb (Fitton et al., 1991). A plot of La/Ba vs La/Nb (Fig. 33) is both insensitive to low-pressure crystal fractionation and partial melting in the source and useful for determining if continental basalts have asthenospheric affinities or a more enriched, lithosphere mantle source. Basin and Range basalts (Fig. 33) tend to plot within the OIB array, as previously documented, and represent melts from the asthenosphere. In general, DVF lavas show a lithospheric influence, i.e., they plot outside the OIB field. The Lower DVF shows the greatest depletion of La/Nb, representing a lithospheric source or slab-derived metasomatized sources or potentially from crustal contamination. Middle DVF rocks plot closer to the OIB field on Fig. 33 and all but one plot just outside the OIB array. We interpret this as evidence for a more depleted source and greater asthenosphere input, as seen in the upper DVF. However, Compositional similarity with OIBs does not mean mafic magmas originated from the asthenosphere (Fitton et al., 1991). The majority of middle DVF rocks plot just outside the OIB array suggesting possible lithosphere input into OIB asthenospheric melts.

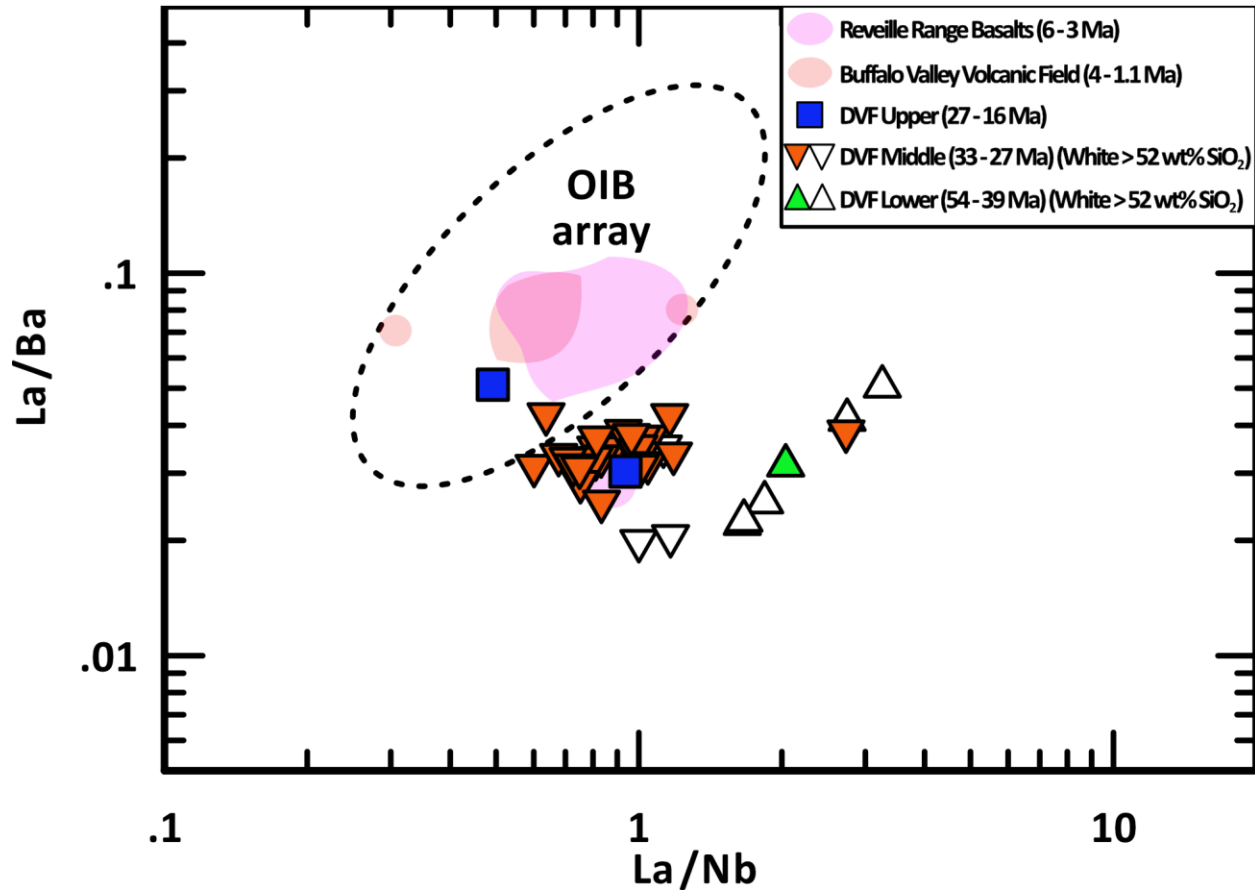


Figure 33: La/Ba versus La/Nb in DVF rocks in southwestern Montana in comparison to Basin and Range basalts from the Reville Range (Yogodzinski et al., 1996) and Buffalo Valley Volcanic Field (Cousens et al., 2013). Diagram is based on (Fitton et al., 1991), where the dashed circle is ocean island basalt (OIB) array. Increases in Ba/Nb (from top left to bottom right) reflects increasing input of slab-derived fluids/metasomatized lithospheric into asthenosphere derived OIB magmas.

Fitton et al. (1991) showed that La vs. Nb can be a useful discriminant for different tectonic affinities for mafic igneous rocks. With respect to the DVF rocks, Fig. 34 shows a transition from arc-type magma for the lower DVF, to OIB-like for middle and upper DVF.

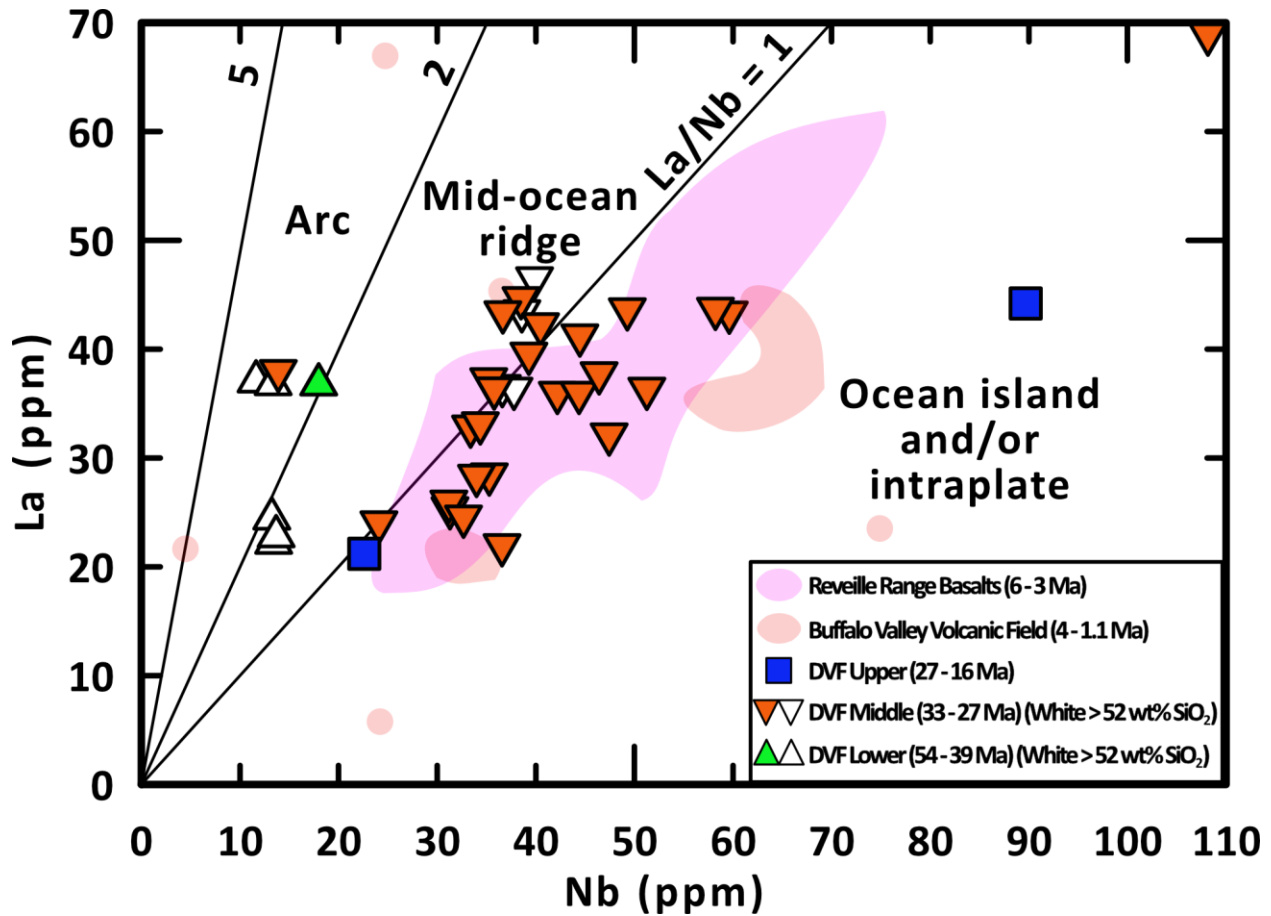


Figure 34: La/Nb tectonic discrimination diagram showing changes in tectonic affinity of DVF rocks in southwestern Montana in comparison to Basin and Range basalts from the Reville Range (Yogodzinski et al., 1996) and Buffalo Valley Volcanic Field (Cousens et al., 2013) based on (Thorkelson et al., 2011).

Melting of mantle wedge can be evaluated using the TiO_2/MnO ratio shown by Thorkelson et al. (2011) to distinguish between MORB or OIB and a volcanic arc-like mantle melts. TiO_2 and Mn are relatively immobile during surface alterations and weathering. TiO_2/MnO ratios show only approximate division between volcanic arc and MORB/OIB compositions with $\text{TiO}_2/\text{MnO} < 10$ representing rocks of possible arc origins (Fig. 35) (Thorkelson et al., 2011). All but one lower DVF falls below the $\text{TiO}_2/\text{MnO} < 10$ in the volcanic arc array while both middle and upper DVF spread across this boundary.

Thorkelson et al. (2011) also showed that Nb/Zr ratio can distinguish among N-MORB, E-MORB and OIB. Low Nb/Zr ratios are typical of volcanic arcs while higher values are common in intraplate fields (Thorkelson et al., 2011). Continental crust will also have very low Nb/Zr ratios (<0.1) (Rudnick and Gao, 2003). Lower DVF has Nb/Zr 0.05 to 0.16, (Fig. 36), indicating potential E-MORB mantle sources of arc-like magmatism. The middle DVF mostly exhibits higher Nb/Zr (0.07 to 0.3, average = 0.23), suggesting deeper, more asthenospheric mantle while the upper DVF has mostly lower Nb/Zr (from 0.14 to 0.33, average = 0.11) overlapping with the E-MORB to OIB fields (Fig. 36).

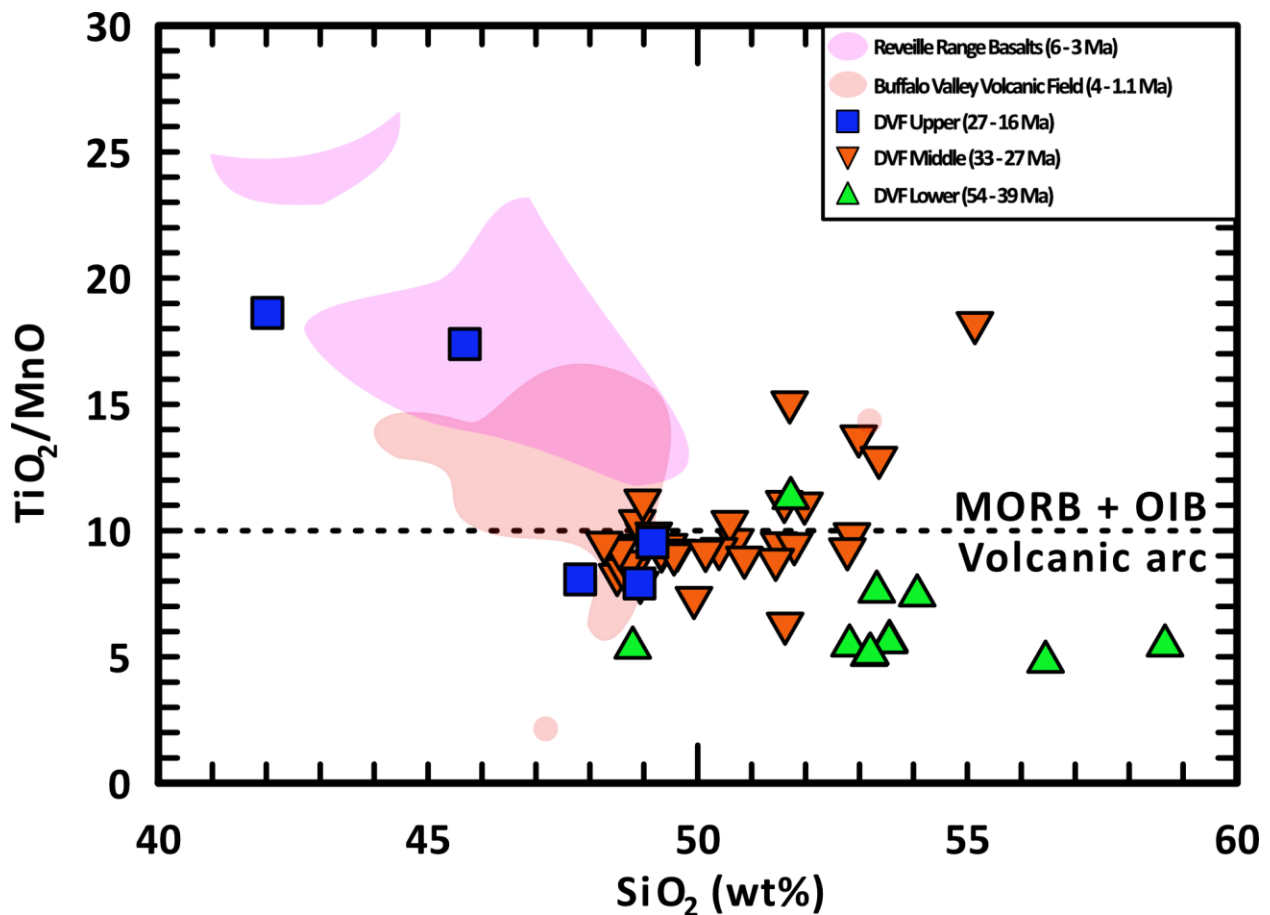


Figure 35: TiO_2/MnO versus wt% SiO_2 showing DVF rock values, in comparison to Basin and Range basalts (Yogodzinski et al., 1996; Cousens et al., 2013). Lower TiO_2/MnO (<10) are associated with volcanic arcs while $\text{TiO}_2/\text{MnO} > 10$ higher ratio values are common in intraplate fields with a larger component of mantle not affected by subduction (Thorkelson et al., 2011).

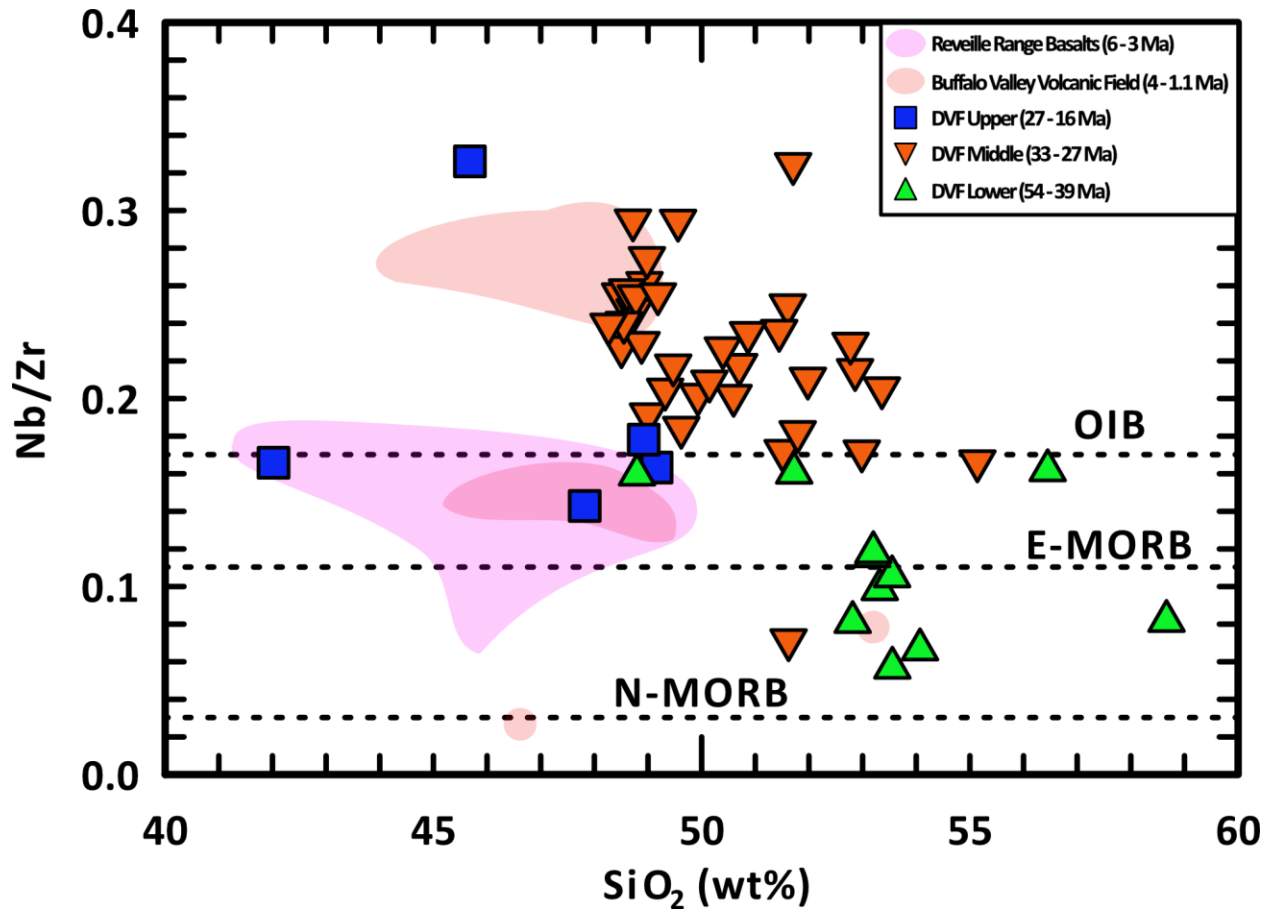


Figure 36: Nb/Zr versus wt% SiO₂ showing DVF rock values for N-MORB (N-Normal) ~0.025, E-MORB (E-enriched) ~0.105, and OIB ~0.165 from Sun and McDonough (1989). Plot is in comparison to Basin and Range basalts (Yogodzinski et al., 1996; Cousens et al., 2013). Lower ratios are associated with volcanic arcs while higher ratio values are common in intraplate fields, possibly associated with more asthenospheric mantle sources; figure base on (Thorkelson et al., 2011)

Ta and Th ratios support these other trace element-based interpretations and are perhaps more sensitive indicators of source because Ta, and Th are relatively immobile during surface alteration and weathering of basalts thus those processes should have little effect on whole-rock Ta/Th (Farmer et al., 2024). Ta and Th are concentrated in accessory minerals with Ta in rutile and Th in apatite which precipitate from Ti and P-rich aqueous fluids ascending and cooling in continental lithospheric mantle undergoing metasomatism (Farmer et al., 2020; Farmer et al., 2024). Rutile is rare and apatite unstable in asthenosphere mantle (Farmer et al., 2020). Aqueous

metasomatism of the deep continental lithosphere below North America is coincident with the long history of prior subduction, including perhaps shallow slab subduction of the Farallon plate (Schmandt and Humphreys, 2010; Apen et al., 2024). Metasomatic hydration of cratonic lithosphere lowers mantle solidus temperature allowing for lower temperature melting of lithospheric mantle either from “arc-like” subduction or later asthenosphere upwelling due to slab-rollback/slab tearing and regional extension. Melting of apatite and rutile bearing mantle will remove rutile, but not apatite from residual mantle at low extents of partial melting ($< \sim 1\%$) (Farmer et al., 2020). Melts from the continental lithosphere which contain rutile (Ta), and apatite (Th) phases can thus be recognized because basalts with Ta/Th between 0.2 and 0.6 (Figs. 37, 38) reflect partial melting of ancient lithospheric mantle previously affected by subduction (e.g., Wyoming craton) (Farmer et al., 2020). Mafic rocks with low Ta/Th (≤ 0.2) are associated with arc settings since Ta is held back by residual rutile in the subducting oceanic crust where coeval metasomatism affected the continental lithospheric mantle (Fig. 39B) (Farmer et al., 2020; Farmer et al., 2024). Mafic rocks with high Ta/Th (> 0.6) are interpreted to have been sourced in asthenospheric mantle.

Phlogopite deposition can also occur during metasomatic processes which distribute it within mantle peridotite or in veins such as glimmerite or similar phlogopite-rich ultramafic rocks and likely occurred within the mantle below our study area due to samples of Wyoming Craton phlogopite-bearing xenoliths (O’Brien et al., 1995). Like rutile, phlogopite also takes in Ti, Nb and Ta (O’Brien et al., 1995). Magmas that interact and assimilate phlogopite can produce wide ranges in Rb concentrations and elevates Ba and K, with partial melting of phlogopite-veined harzburgite lithosphere producing potassic magmas (O’Brien et al., 1995).

Ti, Nb, and Ta, which are usually regarded as immobile elements during slab subduction, can be soluble and mobilized in halogen-bearing or silicate-rich supercritical fluids under subduction P-T conditions (1.5-2.5 GPa and 920-1150 °C) (Hayden and Manning 2008; Manning et al., 2011; Chen et al., 2018). Supercritical fluids released from shallow/flat slab subduction likely allow for Ti, Nb, and Ta to enter the lithospheric mantle. If the subducting slab is at a steeper angle to allow mantle wedge emplaced above it, these Ti, Nb and Ta supercritical fluids are too short-lived to ascend very far into the mantle wedge (Chen et al., 2018).

Overall, trace element geochemistry from the lower DVF (e.g., low Ta/Th, La/Nb, Nb/Zr) indicate that these rocks originated from enriched mantle sources displaying arc-like signatures (Figs. 34, 36, 38). These results suggest that lower DVF magmas originated mantle wedge above a subducting Farallon plate. Middle DVF rocks show more variation in Ta/Th, La/Nb and Nb/Zr which likely reflect variation in mantle source regions and suggest melts originated from enriched source (and/or interacted with crust) but with increasing OIB (asthenosphere) input (Figs. 34, 36, 38, 39). Upper DVF rocks show a continued increase in asthenospheric sources resulting in OIB like geochemistry. However, assimilation of phlogopite bearing lithospheric mantle, interaction with lithospheric mantle melts, or lower crustal materials may offset these ratios (e.g. increasing Ta/Th values) since phlogopite holds Ti, Nb, and Ta and lower crust has an average Ta/Th around 0.5 at 53.3 wt% SiO₂ (Rudnick and Gao, 2003). Thus, melts that assimilated lower crust or lithospheric mantle may alter their trace element ratios (e.g., Ta/Th, La/Nb).

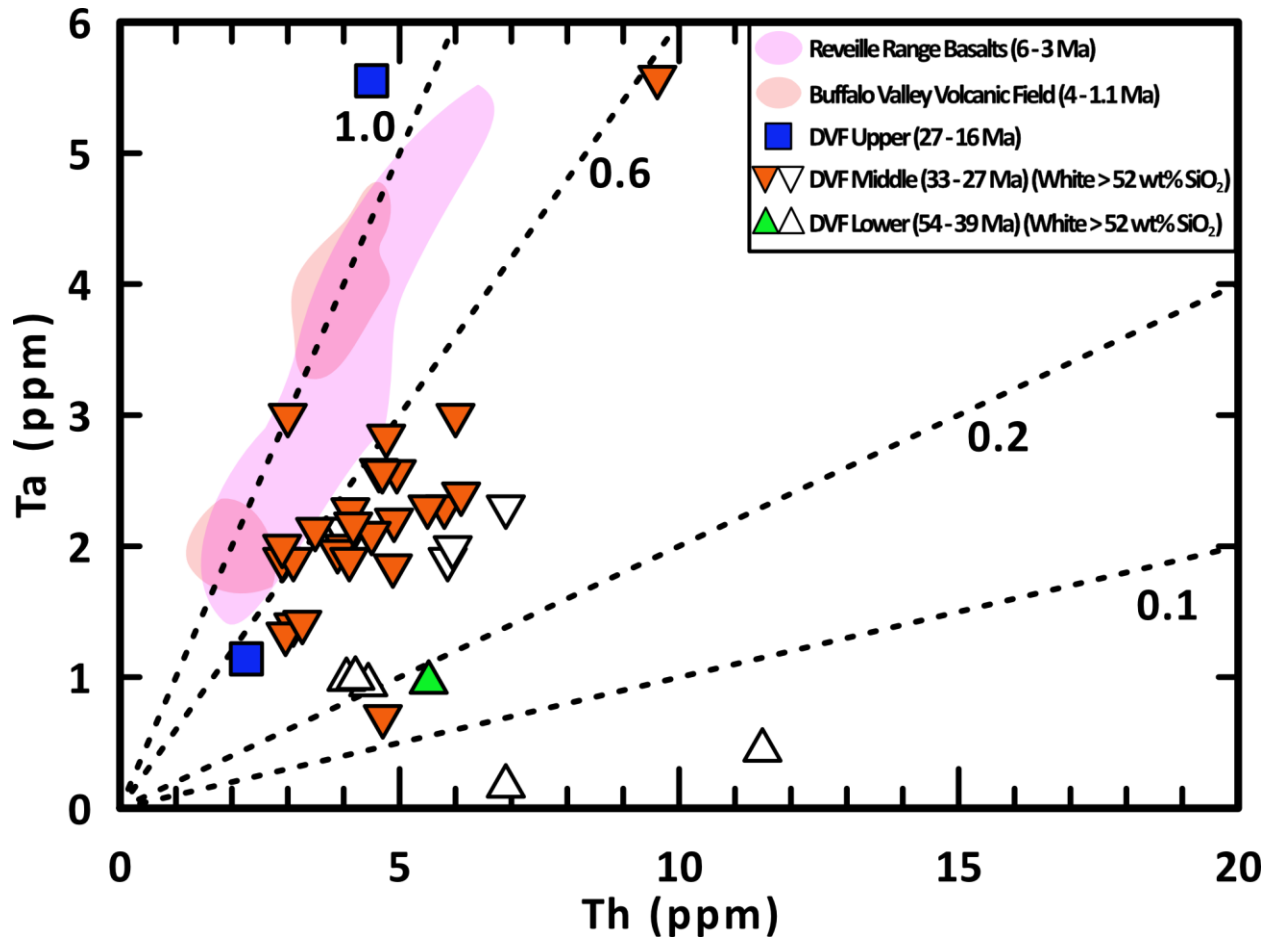


Figure 37: Ta versus Th contents for DVF rocks in southwestern Montana in comparison to Basin and Range basalts (Yogodzinski et al., 1996; Cousens et al., 2013) based on (Farmer et al., 2020; Farmer et al., 2024)

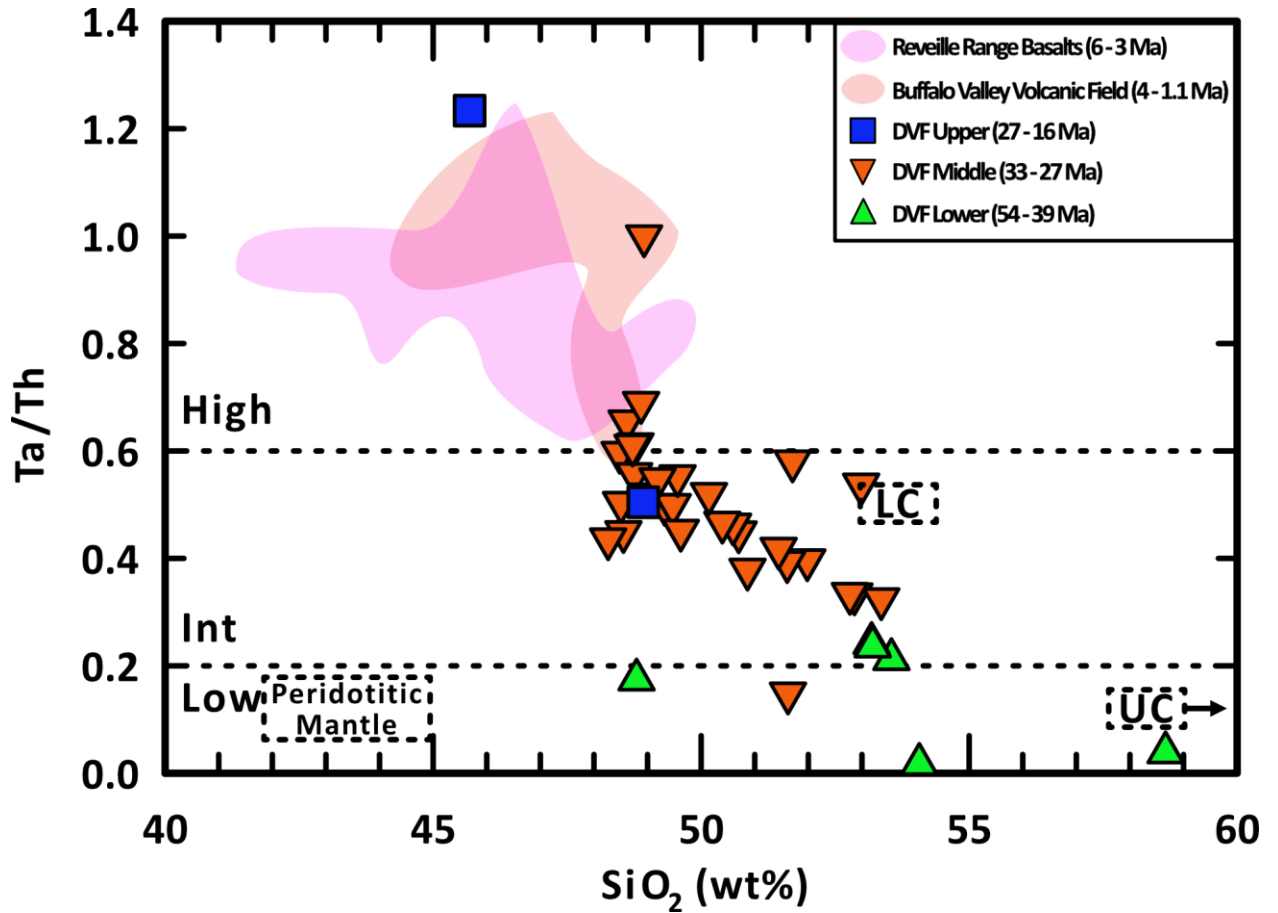


Figure 38: Ta/Th versus wt% SiO₂ for DVF rocks in southwestern Montana in comparison to Basin and Range basalts (Yogodzinski et al., 1996; Cousens et al., 2013) based on (Farmer et al., 2020; Farmer et al., 2024).

The new and first reported $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ isotope ratios for the Dillon volcanic field mafic rocks are also useful for determining melt sources and possible crustal interaction. More radiogenic $^{87}\text{Sr}/^{86}\text{Sr}_i$ and lower $^{143}\text{Nd}/^{144}\text{Nd}_i$ in DVF rocks relative to depleted mantle, could suggest either direct melting of Wyoming Craton lithospheric mantle, assimilation of crust, or mixing of lithospheric mantle-derived melts with DVF melts.

Isotope ratios of Sr-Nd can be additionally coupled with Ta/Th ratios. Ta/Th values between 0.2 and 0.6 in DVF basalts, coupled with high $^{87}\text{Sr}/^{86}\text{Sr}$ and low $^{143}\text{Nd}/^{144}\text{Nd}$ ($^{87}\text{Sr}/^{86}\text{Sr} > 0.706$ and $^{143}\text{Nd}/^{144}\text{Nd} < 0.5126$), are typical of ancient mantle previously affected by subduction

(e.g., Wyoming Craton lithosphere), while higher Ta/Th (>0.6), coupled with lower $^{87}\text{Sr}/^{86}\text{Sr}$ values (< 0.7045 and $^{143}\text{Nd}/^{144}\text{Nd} > 0.5126$), suggest decompression melting of upwelling, rutile and apatite free, asthenospheric mantle due to lithospheric thinning (Farmer et al., 2020).

$^{87}\text{Sr}/^{86}\text{Sr}_i$ values of 10 DVF rocks display decreasing $^{87}\text{Sr}/^{86}\text{Sr}_i$ values with increasing Ta/Th (Fig. 39A). Two lower DVF samples have low Ta/Th (<0.2), coupled with high $^{87}\text{Sr}/^{86}\text{Sr}_i$ (0.7078-0.7072) (Fig. 39A). Middle and upper DVF rocks display both intermediate Ta/Th (0.2-0.6) and high Ta/Th (>0.6) with intermediate and low $^{87}\text{Sr}/^{86}\text{Sr}_i$ (0.7043-0.7059) with one upper DVF sample having the highest Ta/Th (1.23) and lowest $^{87}\text{Sr}/^{86}\text{Sr}_i$ value (0.7043; (Fig. 39A). Basalts from the Basin and Range (Fig. 39A) all display high Ta/Th (>0.6) with $^{87}\text{Sr}/^{86}\text{Sr}_i$ values of 0.7034 - 0.7071, except for one sample displaying low Ta/Th (<0.2) and a higher $^{87}\text{Sr}/^{86}\text{Sr}_i$ value above 0.7071.

$^{143}\text{Nd}/^{144}\text{Nd}_i$ values of 10 DVF rocks display $^{143}\text{Nd}/^{144}\text{Nd}_i$ values between 0.51204 to 0.51236 with lower DVF displaying low Ta/Th (<0.2) and middle and upper DVF displaying both intermediate Ta/Th (0.2-0.6) and high Ta/Th (>0.6) (Fig. 39B). Rocks from the Basin and Range (Fig. 39B) all display high Ta/Th (>0.6) with $^{143}\text{Nd}/^{144}\text{Nd}_i$ values between 0.51214 to 0.51291, higher than all DVF samples, minus one Basin and Range sample displaying low Ta/Th (<0.2) and lower $^{143}\text{Nd}/^{144}\text{Nd}_i$ value of 0.51221 (Fig. 39B). Although the decreasing $^{87}\text{Sr}/^{86}\text{Sr}_i$ values with increasing Ta/Th displayed by DVF rocks suggests greater input of asthenospheric mantle, especially as Ta/Th increases the $^{143}\text{Nd}/^{144}\text{Nd}_i$ values of DVF rocks do not increase with increasing Ta/Th ratios as would be expected for greater asthenospheric input. Instead, $^{143}\text{Nd}/^{144}\text{Nd}_i$ ratios of the DVF display uniform values between 0.51236 - 0.51204. Thus, lower $^{143}\text{Nd}/^{144}\text{Nd}_i$ values (0.51236 – 0.51204) of DVF with decreasing $^{87}\text{Sr}/^{86}\text{Sr}_i$ through time (and

asthenosphere-like trace element ratios) does not directly point to asthenospheric sources for the middle and upper DVF. We will address this discrepancy below.

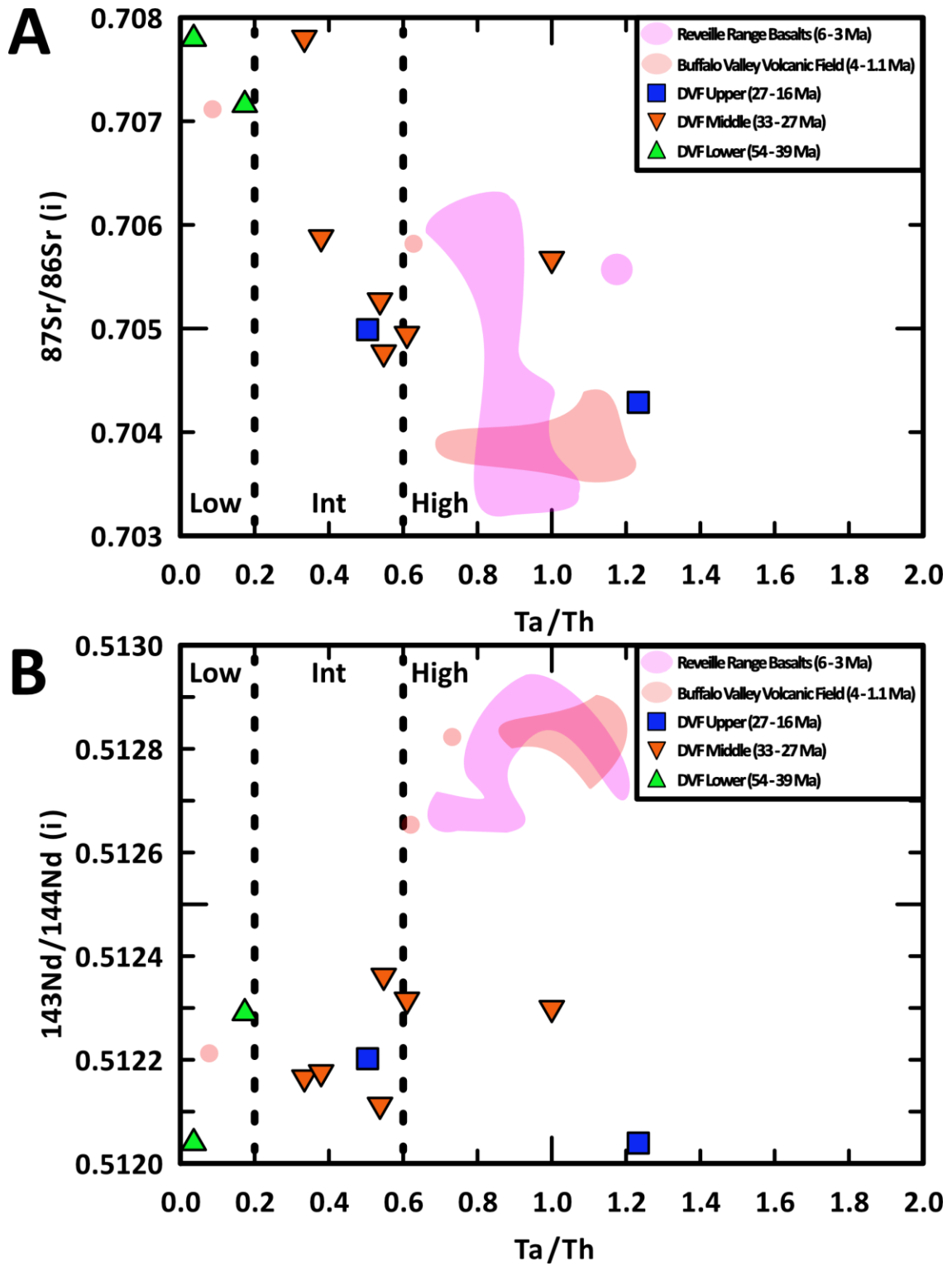


Figure 39: (A) $^{87}\text{Sr}/^{86}\text{Sr}_i$ vs. Ta/Th and (B) $^{143}\text{Nd}/^{144}\text{Nd}_i$ vs. Ta/Th for DVF samples in comparison to Basin and Range basalts (Yogodzinski et al., 1996; Cousens et al., 2013) based on (Farmer et al., 2020; Farmer et al., 2024)

The lower DVF samples range from $^{87}\text{Sr}/^{86}\text{Sr}_i$ (0.7072 - 0.7078) and $^{143}\text{Nd}/^{144}\text{Nd}_i$ (0.51204 - 0.51229). Their compositions overlap the field of Eocene central Idaho Challis volcanic rocks, and they plot above the field for ultramafic xenoliths from HW and EBP (Fig. 40). One middle DVF rock, WE24-13, also over lies the Eocene central Idaho Challis field.

The less radiogenic $^{87}\text{Sr}/^{86}\text{Sr}_i$ and relatively consistent $^{143}\text{Nd}/^{144}\text{Nd}_i$ isotope ratios of the middle/upper DVF rocks plot within the Sr-Nd isotope range for the enriched mantle 1 (EM1) (Fig. 40). EM1-type mantle is believed to reflect asthenosphere mantle affected by recycled crust (e.g., primarily oceanic crust). Ta/Th ratios coupled with the Sr-Nd isotope ratios (Figs. 38, 39) show that middle and upper DVF rocks have either intermediate Ta/Th (0.2-0.6), attributed to melting of metasomatized lithospheric mantle, and high Ta/Th (>0.6) attributed to melting of asthenosphere. For example, the upper DVF sample WE24-16 has the highest Ta/Th (1.23), which suggests an asthenospheric source but also however displays the lowest $^{87}\text{Sr}/^{86}\text{Sr}_i$ (0.7043) which is more asthenosphere-like, but lower $^{143}\text{Nd}/^{144}\text{Nd}_i$ ratio (0.51204), than would be expected of asthenosphere mantle (Fig. 40).

The samples of middle and upper DVF that display high Ta/Th (>0.6) which is associated with asthenosphere sources and have more EM1-like enriched $^{143}\text{Nd}/^{144}\text{Nd}_i$ ratios can be interpreted in a few ways. First, this EM1 mantle signature could represent a yet unidentified mantle source, though no evidence of similar EM1-type Sr and Nd isotope ratios are reported in any other regional Cenozoic basalts. A more likely scenario, is that upper DVF asthenosphere melts, interacted with subducted sediment, crust, or mantle lithosphere causing Nd isotope enrichment (e.g., lower $^{143}\text{Nd}/^{144}\text{Nd}_i$ ratios). To assess these possibilities, binary mixing between depleted MORB mantle and modern Cascadia sediments, lower crustal norite (LC), Lone Mtn middle crust (MC), regional Archean Wyoming craton upper crust (UC), and a representative

Wyoming craton lithosphere melt (LM) was calculated. Cascadia sediment was chosen to reflect the general Sr-Nd isotope character of Pacific Northwest subducting sediment, and the lithospheric mantle melt is the ~4 Ma Pilot Knob lamprophyre that crops out to the southeast of the study area in the upper Wind River basin, Wyoming (Dodd, 2015; Brueseke et al., 2018). Mixing endmembers are reported in Table 2 and the results are depicted in Fig. 40. Of all the four model mixing curves, only the mixing of ~1%, or less LM melt (e.g., Pilot Knob lamprophyre) and/or 4-8% mixing with the lower crustal norite sample replicates the EM1-like $^{87}\text{Sr}/^{86}\text{Sr}_i$ and $^{143}\text{Nd}/^{144}\text{Nd}_i$ values displayed by the middle/upper DVF rocks (Fig. 40). We suggest that just 1% mixing of a lithospheric mantle melt, and/or 4-8% mixing with the lower crust with an asthenosphere melt with trace element characteristics identical to the upper and middle DVF basalts, is enough to pull middle/upper DVF Sr-Nd isotope ratios into the EM1 mantle field. This interaction with the lithospheric mantle and the lower crust likely sways Ta/Th ratios with the average Ta/Th value being ~0.5 in the lower crust (Rudnick and Gao, 2003). This could either be some kind of direct interaction between melts, or perhaps more likely reactive assimilation, as the middle and upper DVF melts were undergoing channelized transport through previously enriched lithospheric mantle as described by Roy et al. (2024) or by stagnation in a MASH zone at the lower crust-mantle boundary which can result in variability in $^{143}\text{Nd}/^{144}\text{Nd}$ (Emo and Kamber, 2022).

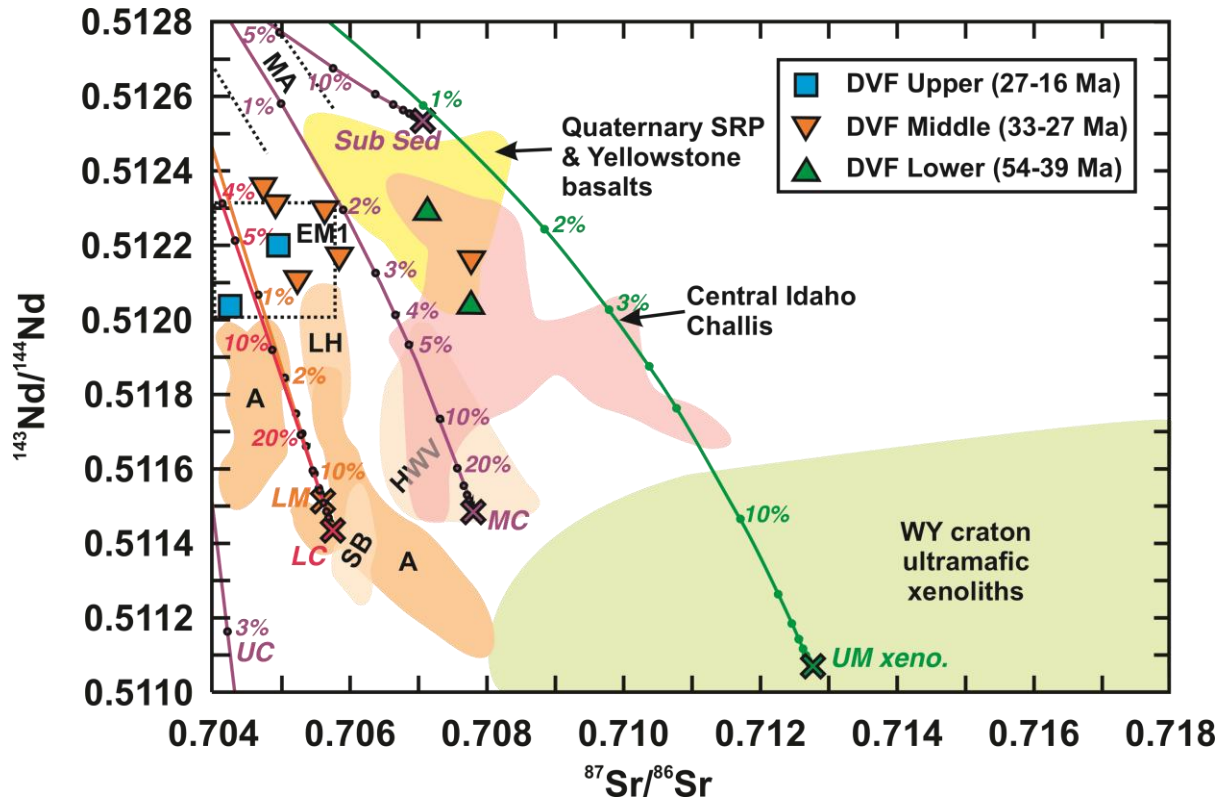


Figure 40: Mixing between DMM (Depleted MORB Mantle), lower crust, LC (B-1) Norite) (Leeman et al., 1985), middle crust, MC (18BS02) (Dudás et al., 2023), upper crust, UC (980C2) (Frost et al., 2006), ultramafic xenolith, (UM) (Carlson and Irving, 1994), Cascadia sediment (Sub Sed) (Plank and Langmuir (1998), and ~4 Ma Pilot Knob lamprophyre (LM) (Brueseke et al., 2018).

Abbreviation	Sub Sed	UC	MC	LC	UM	LM	DM
Endmember	Cascadia Sediment	Upper Crust	Lone Mtn Middle Crust	Norite Lower Crust	Ultra Mafic Xenolith	Pilot Knob	Depleted Mantle
$^{87}\text{Sr}/^{86}\text{Sr}_i$	0.7071	0.70586	0.70783	0.70578	0.71279	0.705612	0.702583
Sr (ppm)	216	322	842	233	778	2147	9.8
$^{143}\text{Nd}/^{144}\text{Nd}_i$	0.512534	0.508857	0.511488	0.511438	0.511065	0.511507	0.512149
Nd (ppm)	21.8	19.9	37.5	16.4	26.9	139.1	0.7

Table 3: Isotope values for endmembers used in binary mixing calculations in Figure 40. Isotope ratios and elemental concentrations of B-1 LC, Leeman et al. (1985), mgab UM, Carlson and Irving (1994), depleted mantle (DM), Chauvel and Blichert-Toft (2001), ZD14-2 Pilot Knob (LM) (Brueseke et al., 2018), 18BS02 MC, Dudás et al. (2023), 980C2 UC, Frost et al. (2006), and Sub Sed, Plank and Langmuir (1998). Initial Sr-Nd isotope ratios of crust and UM xenolith are age corrected to 30 Ma.

In summary, enrichment of lithospheric mantle (Wyoming Craton) occurred during shallow slab subduction of the Farallon plate with slab fluids driving metasomatism which enriched the overlying lithospheric mantle (Fig. 41A). Initiation of slab rollback ca. 54-40 Ma led to arc-like magmatism and eruptions of the lower DVF, with low Ta/Th values (<0.2) that reflect mantle wedge melting (Fig. 41B). Lower and some middle DVF melts underwent 1-3% mixing with ultramafic rocks from the Wyoming Craton which is also consistent with their high Cr and Rb concentrations. This interaction likely occurred at the crust/mantle boundary, where high-Mg olivines and Cr-rich spinels were incorporated in a MASH (melting-assimilation-storage-homogenization; Hildreth and Moorbath, 1988). Middle DVF melts are sourced primarily from melting metasomatized lithospheric mantle suggested by intermediate Ta/Th (0.2-0.6) (Fig. 41C) with one upper DVF sample (WE24-16) clearly being sourced from asthenospheric melts associated with thinning lithosphere and upwelling asthenosphere mantle resulting in high Ta/Th (>0.6) magmatism (Fig. 41D) as well as its steep drop off in HREEs suggesting deeper melting in the presence of garnet (Fig. 16). Both middle and upper DVF interacted with the lower crust and/or the lithospheric mantle with Sr-Nd isotope mixing curves suggesting 4-8% lower crustal mixing with asthenosphere melts and/or $\sim 1\%$ or less of lithospheric mantle melts. These asthenosphere-sourced melts interacted with overlying lithospheric mantle and lower crust leading to the Sr and Nd isotope characteristics of an EM1-like mantle.

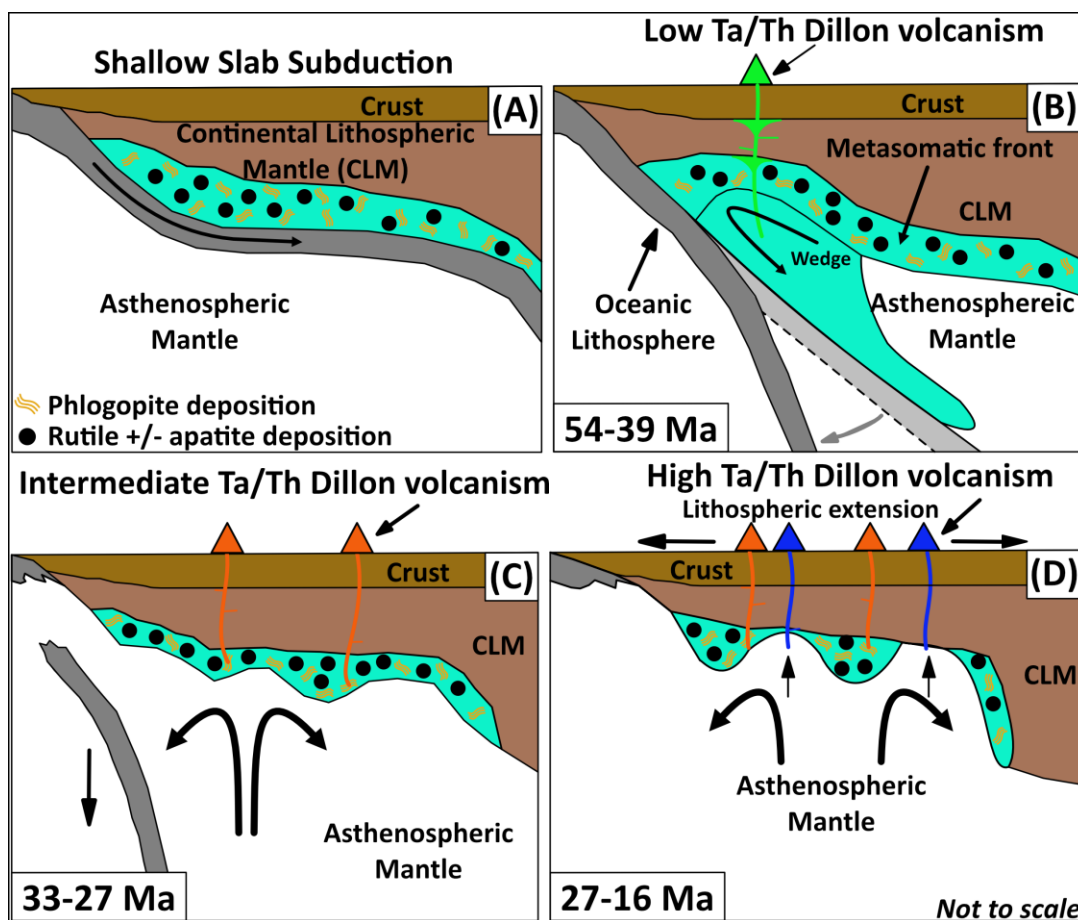


Figure 41: Illustration modified from Farmer et al. (2020) showing (A) lithospheric mantle with metasomatized by low-angle/flat slab subduction with no magmatism, (B) melt generation in continental arc setting with subduction fluids, (C) foundering/slab roll-back triggering asthenosphere upwelling and magmatism, and (D) melt generation from upwelling asthenosphere mantle with a thinning and extending lithosphere.

Tectonomagmatic Interpretations of the Lower DVF- Possible links to Challis magmatism

From this study, we suggest rocks of the lower DVF are associated with arc-like magmatism associated with subduction of the Farallon plate, like regional Challis-Absaroka volcanism. The lower DVF displays enrichment in LREE and LILE with notable depletion in Nb and Ta (Fig.42) along with low Ta/Th ratios (<0.2) (Figs. 37, 38). Other incompatible trace element ratios such as TiO_2/MnO (Fig. 35), La/Nb (Fig. 34) and Nb/Zr (Fig. 36) further support that magmas of the lower DVF are from enriched mantle sources, likely originating from mantle

wedge above the Farallon plate. We interpret lower DVF to represent arc-like magmatism related to the subduction of the Farallon plate equivalent to the coeval Absaroka and Challis volcanic fields.

Due to the proximity of the Challis volcanic field and the timing of Challis and DVF magmatism, we consider the possibility that the lower DVF is an extension of the Challis volcanic group. The eruption of the lower Dillon volcanics (54-39 Ma) was coeval with the Challis volcanic field (51-43 Ma) in central Idaho. We compare published geochemistry from the central Idaho Challis field with our newly collected samples, preliminary samples and previously published data of the “lower Dillon” (Fig. 42).

Lower DVF rocks are highly potassic, with alkaline to subalkaline affinities and arc-like trace element compositions (e.g., LILE, LREE enrichment and HFSE depletion) and are interpreted to represent arc melts generated in a mantle wedge above the subducting Farallon slab (Fritz et al., 2007; this study).

For comparison, only Challis with $\leq$ wt% 57 SiO₂ are included here. The central Idaho Challis rocks from McKervey (1998) show enrichment of LILE and depletion in high field strength elements relative to primitive mantle (Fig. 42). The lower DVF rocks show similar LILE enrichments and HFSE depletions and plot mostly within the Challis array (Fig. 42). The lower DVF is the only member out of the Dillon volcanics that overlaps in age with the Challis volcanic field in central Idaho. Additionally, lower DVF rocks also overlaps within central Idaho Challis volcanic rocks in Sr-Nd isotope space (Fig. 40). Thus, we suggest that the lower DVF is essentially an extension of the Challis volcanic group and should be included as such. This geochemical link defined above, also aligns well with geological constraints, namely the timing of magmatism (esp. relative to younger middle and upper DVF).

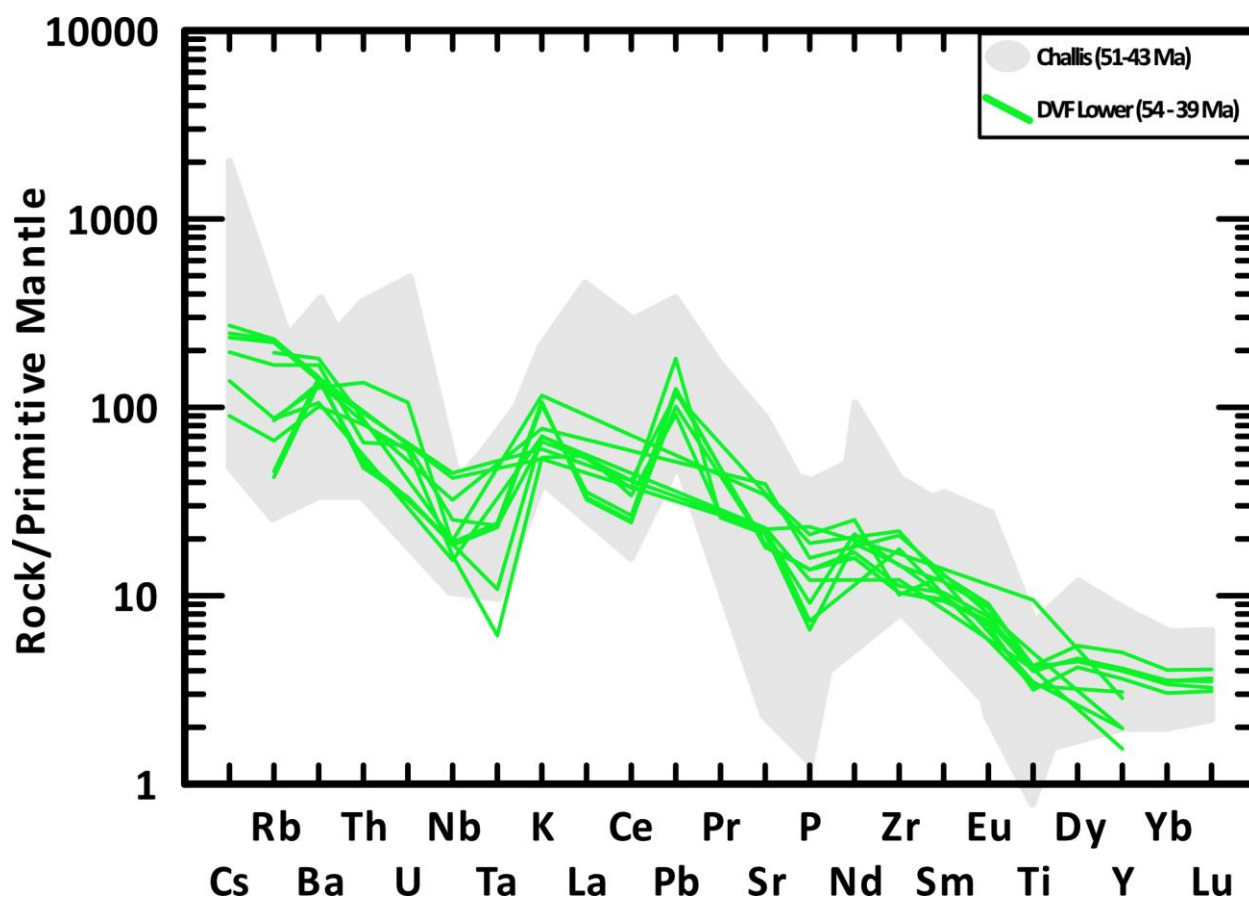


Figure 42: Lower DVF rocks compared with central Idaho Challis rocks with < 57 wt% SiO₂, normalized to Primitive mantle after Sun and McDonough, (1989). Challis data from McKervey (1998).

Tectonomagmatic Interpretations of the Middle DVF – Lithospheric Drip vs. Extension

Early Oligocene uplift, extension, basin development and DVF magmatism in southwestern Montana have been proposed by Li et al. (2021) to result from removal of the lower mantle lithosphere (lithospheric drip) followed by upwelling asthenosphere. This contradicts previous assumptions that the Cordilleran orogen gradually lost elevation during extension associated with rollback of the Farallon flat slab (Li et al., 2021). The lithospheric drip model was proposed by Li et al. (2021) because lithospheric removal is known to be common during orogenesis and can cause rapid uplift. They inferred that prolonged Cretaceous-early

Paleogene contraction and lithosphere thickening set the stage for lithospheric drip. Li et al. (2021), however, note that regional Oligocene magmatism is scarce in southwestern Montana (Fritz et al., 2007). They explain this by proposing that the scale of the lithospheric drip may be small such that the resulting asthenosphere upwelling was not extensive enough for widespread magmatism.

The model by Li et al. (2021) proposes melting of the lithosphere mantle within the drip, with focused Oligocene uplift directly above due to isostatic and thermal effects. If melting of the lithosphere was occurring, the resulting geochemistry of the middle DVF should show incompatible trace element ratios (and isotope data) that coincide with lithosphere melting. However, geochemical trace element and isotope ratios will very likely not be able to show a difference between melting mantle lithosphere due to lithospheric drip vs. decompression melting of lithospheric mantle because of extensional tectonics. Studies outside the scope of this work are instead utilized.

Work by Thoresen et al. (2024) in southwestern Montana attribute extension, subsidence and basin formation in the region to thermal weakening from both contractional features associated with a southwestward Farallon slab rollback and/or lithospheric mantle removal. Uplift is shown to have occurred in the location of modern-day Yellowstone in Wyoming and Idaho between ~46.9 and 35 Ma, whereas extensional valleys and basins developed at this time in southwestern Montana (Thoresen et al., 2024). This aligns with regional studies (e.g., Schwartz and Schwartz, 2013; Vuke 2022) demonstrating that widespread extensional basin development occurred throughout southwest Montana during the Oligocene. A disconformity in local Renova sedimentary strata in the Gravelly Range (e.g., ~37 – 33 Ma) documented by Luikart (1997) as well as the similar Dell-Cook Ranch unconformity noted by Nady and Fan,

(2025), possibly reflects non-deposition preceding localized extension in the Gravelly Range, later associated with the basalt eruptions at Black Butte, Lion Mountain, etc.

Nandy and Fan, (2025) address the results by Li et al. (2021) in which they believe Li's results were likely from variation of Delta D in different depositional environments and from a low sampling resolution in the Cook Ranch member. Volcanic glass delta D values from Nandy and Fan, (2025) do not replicate Oligocene values by Li et al. (2021). Thus, we consider middle DVF magmatism to be a product of extensional thinning of the lithosphere and upwelling asthenosphere due to Farallon slab rollback/foundering and gravitational collapse of Sevier and Laramide style faults rather than from lithospheric drip.

Tectonomagmatic Interpretations of the Upper DVF

Continued extension in southwestern Montana from the Oligocene into the Miocene resulted in continued Dillon volcanism. Though not as numerous in outcrops, upper DVF basalts represent the youngest of the Dillon volcanism and display the least subduction-affected chemical compositions. The upper DVF shows similar rare earth element geochemistry to asthenosphere-derived Basin and Range basalts in Nevada, Buffalo Valley volcanic field (Cousens et al., 2013) and the Reveille Range basalts (Yogodzinski et al., 1996) (Fig. 31). Our geochemical and isotope data show that upper DVF tends to have high Ta/Th (>0.6) (Figs. 36, 37) and geochemical ratios such as Nb/Zr, La/Nb and La/Ba that are typical of OIB (Figs. 32-35). Thus, we suggest upper DVF alkali basalts were formed from low degree partial melting of upwelling asthenospheric mantle, in response to continued extension and lithosphere thinning in southwestern Montana into the Miocene.

Dillon Volcanic Field Petrogenetic Model and Interpretations

Foundering of the Farallon slab is interpreted to have initiated at ~55 Ma due to the arrival and subsequent collision of the Siletzia oceanic plateau in the northern Cordillera (Howlett et al., 2021). Isosurface modeling of North America by Sigloch et al. (2008) indicates a slab gap in the foundered Farallon slab below our study area orientated NE to SW. We interpret this to be the approximate location of tearing and slab-window formation below southwestern Montana beginning ~55 Ma (Fig. 43). The portion of the Farallon slab south of the tear became steeply dipping and resulted in a necking separation that enabled southward slab rollback and foundering and led to upwelling asthenosphere that marks the initiation of the Cordilleran ignimbrite flare-up (Howlett et al., 2021; Tepper et al., 2023). This flare-up swept southward from this region, through the northern Great Basin (Humphreys, 2009; Best et al., 2016). The Farallon slab north of the tear rolled back but did not founder and began hanging steeply, starting at ~40 Ma and continuing to the present beneath eastern Washington to depths of ~230 km and 450-600 km beneath northern Idaho and westernmost Montana (e.g., slab curtain of Schmandt and Humphreys, 2010).

DVF magmatism contains geochemical signatures that evolve with the evolving regional tectonomagmatic regime. The lower DVF is interpreted to derive from the mantle wedge related to subducting Farallon slab. The middle DVF is interpreted to be derived from mixing of upwelling asthenosphere mantle and metasomatized lithosphere mantle, in response to tearing and foundering of the Farallon slab. This is supported by the similar enrichments in LREE and LILE relative to the lower DVF, but also the fact that middle DVF basalts have more OIB-like incompatible trace element signatures (e.g., $\text{Nb/Zr} > 0.17$, $\text{La/Nb} > 1$; intermediate Ta/Th 0.2-0.6). Tearing of the Farallon plate is interpreted to have initiated NW to SE extension with the study area and as shown by approximate NE to SW striking Miocene faults with passive

asthenospheric upwelling through the slab tear (Fig. 43). The upper DVF is interpreted to represent basaltic magmatism coeval with continued regional extension, but with a greater asthenospheric mantle component (e.g., lower enrichments in LREE and LILE, plus OIB-like incompatible trace element ratios such as high Ta/Th > 0.6). The occurrences of alkali basalts in the DVF and similarity to small volume, <5 Ma Basin and Range alkali basalts, supports this interpretation (Fig. 43).

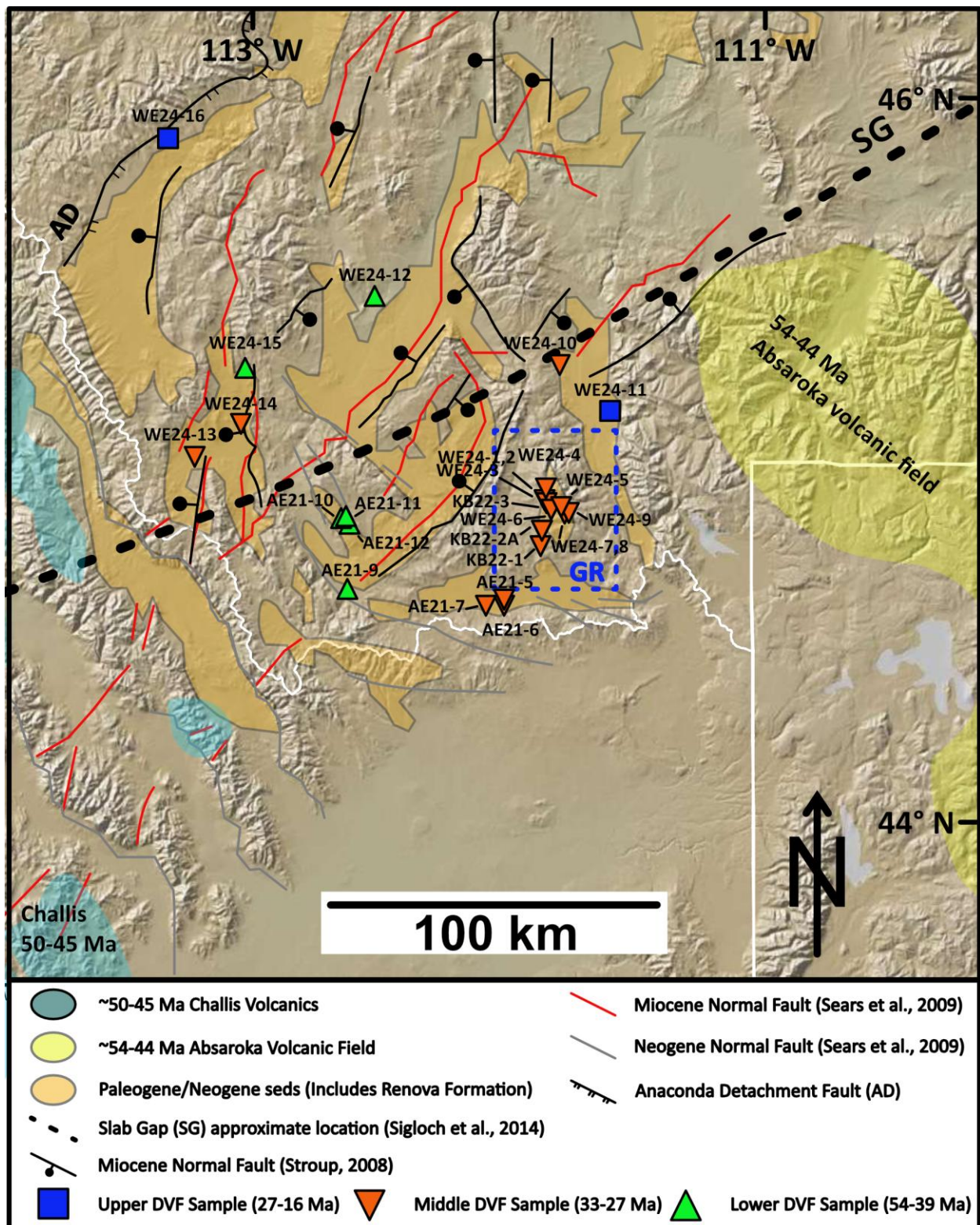


Figure 43: Simplified geologic map of southwestern Montana, Idaho, and Wyoming showing locations of volcanic fields, basin sedimentation, and faulting, plus approximate location of the inferred slab gap (e.g., northern edge of tear; SG) and our DVF sample locations.

Given the tectonic constraints and data, we report here, we suggest DVF magmatism occurred due to Farallon slab foundering and/or rollback, which induced asthenosphere uprising through a slab window and/or gap (Fig. 44A). Farallon slab rollback/tear affecting regional extension has been proposed by prior workers (e.g., Fritz et al. 2007; Smith et al. 2014; Cassel et al. 2018; Howlett et al. 2021; Li et al. 2021). Shallow slab subduction that ended ca. ~55 Ma resulted in arc-like magmatism and the lower DVF. Asthenospheric upwelling through the tear interacted with subduction affected, metasomatized lithospheric mantle, producing magma sources for the middle DVF associated with the switch to an extensional tectonic regime. In this model for DVF magmatism, upwelling asthenosphere undergoes decompression melting creating a mix of both asthenosphere and lithosphere sources primarily for the middle DVF and eventually leading to small degree melting of asthenosphere (e.g., upper DVF alkali basalts) as extension and upwelling continues through time (Fig. 44B, C). The upper DVF in southwestern Montana is associated with continued Basin and Range-like extension and extensional basin development and associated sedimentation.

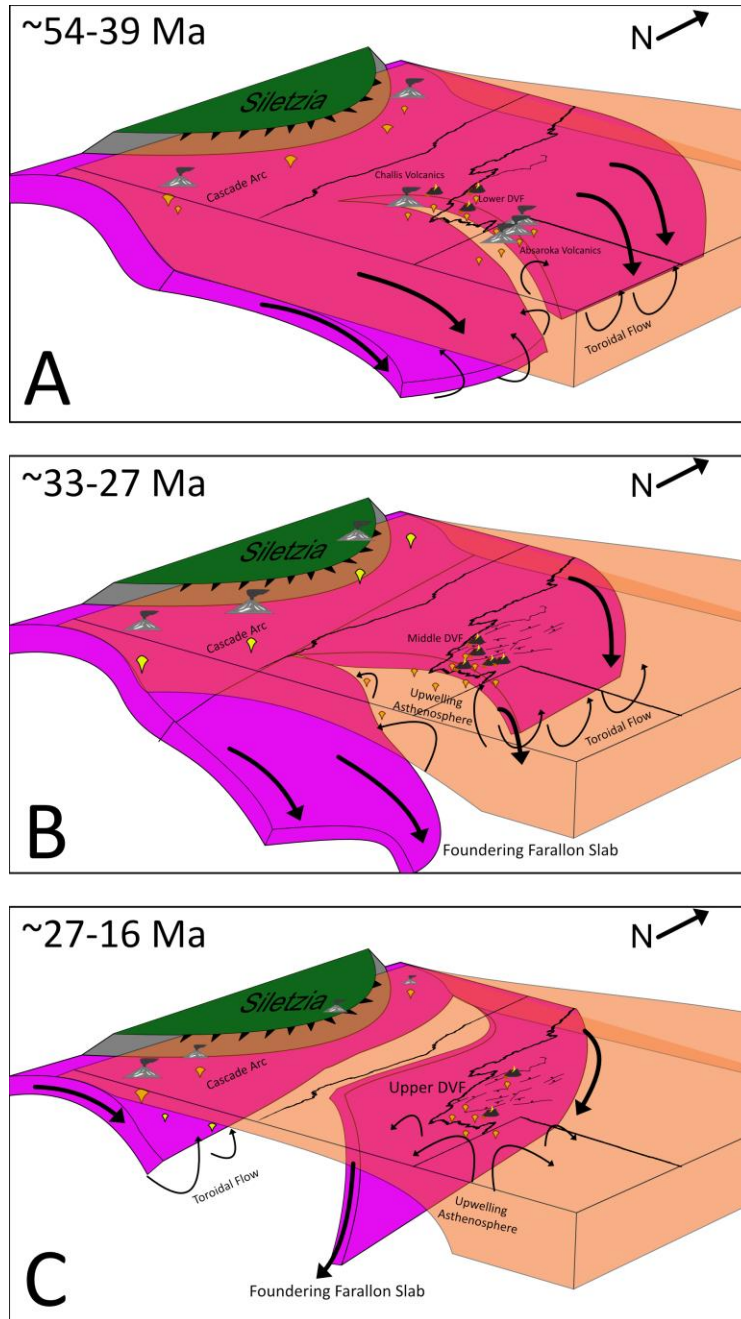


Figure 44: Tectonomagmatic model for DVF magmatism in relation to the collision of the Siletzia oceanic plateau and Farallon slab tear and roll-back. (A) ~54-39 Ma showing the beginning of subduction like processes and end of shallow slab subduction of the Farallon plate triggering magmatism feeding Challis, Lower Dillon, and Absaroka volcanic fields (B) ~33-27 Ma showing foundering of the southern and northern Farallon slabs, upwelling asthenosphere, the initiation of the Cascade Arc to the west (C) Continued asthenosphere upwelling, coupled with upper-plate extension generating upper DVF alkali basalts. Hanging Farallon slab remnant (e.g., "slab curtain") below Washington, Idaho, and Montana, and the Cascade Arc continuing to develop in the west. Figure not to scale.

Chapter 5 - Conclusions

- Lower DVF samples display arc-like subduction-associated geochemical signatures and predate extensional basin development in southwestern Montana. The lower DVF also erupted coevally with the Absaroka and Challis volcanic fields. Lower DVF rocks are calc-alkaline and consist of basalts, basaltic andesites, and andesites. Enrichments in LREE and LILE, plus HFSE depletions, are products of subduction-modification of the mantle source (mantle wedge or metasomatized lithosphere above Farallon slab). The lower DVF geochemically and isotopically (Sr-Nd) overlap with Challis volcanic rocks suggesting “lower Dillon” may represent the easternmost extension of the Challis volcanic field.
- Middle DVF rocks are calc-alkaline to mildly alkalic, consisting of olivine-bearing basalts to basaltic andesites. They display similar LREE and LILE enrichments (and HFSE depletions), to the lower DVF. Key differences include higher $Ta/Th = 0.2-0.6$ and elevated Nb/Zr relative to lower DVF. Middle DVF volcanism is synchronous with extension, and basin formation in southwestern Montana. It is not associated with lithospheric drip, but rather extension and gravitational thinning of the overthickened continental lithosphere, plus slab rollback, ensuing asthenosphere upwelling and melting of asthenosphere mantle, plus interaction with lithospheric mantle and crust. Furthermore, the Gravelly Range hosts a significant middle DVF eruptive complex and related lavas (e.g., ~33-30 Ma Black Butte-Lion Mountain area).
- Upper DVF rocks are mostly olivine-rich alkali basalts and crop out across the study area. These basalts show relative depletions in LREE and LILE and, overall, have more OIB-like trace element compositions relative to the lower/middle DVF (e.g., $Th/Th > 0.6$).

Isotopically, they have less radiogenic $^{87}\text{Sr}/^{86}\text{Sr}_i$, but more enriched $^{143}\text{Nd}/^{144}\text{Nd}_i$, which overlaps with EM1-like mantle. We suggest that upper DVF basalts reflect low-degree partial melting of primarily upwelling asthenosphere, where melting occurred under continued regional extension; mixing/assimilation of ~1% lithospheric mantle and/or 4-8% of the lower crust lowered their $^{143}\text{Nd}/^{144}\text{Nd}_i$ ratios. These rocks are analogous tectonically and to an extent, geochemically, to <5 Ma, small-volume monogenetic basalts of the central Basin and Range province.

- DVF basalts record magmatic source evolution through time associated with the regional tectonomagmatic change from subduction and arc-like magmatism in the Eocene, to extensional tectonomagmatism following slab rollback/foundering of the previously shallow-subducting Farallon plate from the Oligocene into the Miocene, in southwestern Montana.

Future Work

Future work in the DVF should focus on primarily upper DVF. Though we attempted to collect these younger DVF rocks, sampling of upper DVF basalts is still lacking. While known outcrops and volcanic centers are sparse, a few more such as the ~16.9 Ma Lobo Mesa basalt package in the Gravelly Range (Smith, 1992), should be targeted. Accurate age dates on DVF rocks are lacking or need updating, such as the possibly ~16 Ma Loba Mesa basalts, which have only been dated using K-Ar techniques (Smith, 1992) and, thus, may not provide an accurate age for this volcanic center. More and better age dates on DVF rocks will better constrain the timing and rates of tectonomagmatic processes. Other regional volcanic fields of Oligocene-Miocene age should also be evaluated such as the Potlatch, and Kamiah volcanic fields in Idaho, and the Hog Heaven volcanics in northwestern Montana, to

determine whether they are related to the DVF and the overall tectonic-magmatic changes we document.

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Appendix A - Supplementary Geochemistry

Whole rock major and trace element data from volcanic and intrusive rocks of the DVF

Sample	WE24-1	WE24-2	WE24-3	WE24-4	WE24-5
Unit	Middle Dillon	Middle Dillon	Middle Dillon	Middle Dillon	Middle Dillon
SiO ₂	48.72	51.71	49.56	48.55	48.98
TiO ₂	1.43	2.59	1.61	1.45	1.69
Al ₂ O ₃	16.46	17.07	18.40	16.90	17.14
Fe ₂ O ₃	3.57	4.44	3.96	3.48	3.59
FeO	6.14	5.68	6.49	6.18	5.47
MnO	0.16	0.17	0.18	0.16	0.15
MgO	8.95	3.05	5.00	8.77	7.05
CaO	9.51	6.70	9.27	9.87	9.20
Na ₂ O	3.99	3.72	3.02	2.95	3.00
K ₂ O	0.66	3.94	2.05	1.34	3.05
P ₂ O ₅	0.42	0.92	0.48	0.35	0.67
LOI	2.42	3.85	2.67	2.14	3.16
Mg#	63	36	47	62.7	59.1
Ag	0.03	0.07	0.02	0.01	0.04
As	0.60	1.77	1.15	0.69	0.96
V	183	111	197	190	150
Cr	231	8	243	233	120
Ni	154	5	147	130	96
Zn	71	106	64	72	74
Ga	16.70	21.70	17.70	16.80	17.40
Ge	1.69	1.85	1.79	2.61	1.47
Nb	47.45	108.16	51.26	32.68	59.63
Ba	962	1613	1112	867	1381
Bi	0.000	0.01	0.02	0.000	0.02
Cd	0.08	0.08	0.08	0.07	0.08
La	32.17	68.92	36.36	24.62	43.36
Ce	60.78	124.66	71.25	50.79	86.02
Pr	7.10	14.61	8.38	5.95	9.82
Nd	29.13	55.95	33.08	25.53	38.58
Sm	5.47	10.10	6.41	4.96	6.84
Eu	1.75	2.97	1.99	1.61	2.15
Gd	5.17	9.21	6.04	5.23	5.88
Tb	0.77	1.41	0.91	0.75	0.82
Dy	4.60	8.14	5.09	4.62	4.46
Ho	0.94	1.63	1.05	0.93	0.84
Er	2.37	4.41	2.71	2.50	2.27
Tm	0.37	0.65	0.42	0.38	0.31
Yb	2.29	4.13	2.39	2.39	2.04
Lu	0.35	0.65	0.36	0.37	0.31
Rb	41.0	81.7	58.4	36.1	72.2
Sb	0.21	0.28	0.25	0.10	0.17
Sr	669	622	667	546	915
Cs	1.01	0.68	1.46	0.61	2.02
Hf	3.72	6.88	4.12	3.12	4.94
Ta	2.13	5.58	2.28	1.33	2.57
Th	3.49	9.61	4.12	2.96	4.95
Tl	0.01	0.07	0.10	0.04	0.09
Mo	1.58	3.34	1.24	1.16	2.86
U	0.96	0.97	1.04	0.83	1.40
Y	25	45	28	25	24
Zr	161	333	174	136	217
Sc	27.13	9.52	28.66	28.42	19.51
Sn	1.17	2.11	1.24	1.10	1.45
Cu	35	91	28	36	29
Pb	3.84	6.47	4.00	3.82	4.53
⁸⁷ Sr/ ⁸⁶ Sr _m	0.705017				
⁸⁷ Sr/ ⁸⁶ Sr _i	0.704947				
¹⁴³ Nd/ ¹⁴⁴ Nd _m	0.512339				
¹⁴³ Nd/ ¹⁴⁴ Nd _i	0.512316				
εNd _i	-5.4				

Sample	WE24-6	WE24-7	WE24-8	WE24-9	WE24-10
Unit	Middle Dillon	Middle Dillon	Middle Dillon	Middle Dillon	Middle Dillon
SiO ₂	49.19	52.99	50.15	53.36	48.27
TiO ₂	1.52	1.60	2.40	1.68	1.57
Al ₂ O ₃	17.34	17.30	13.74	14.42	17.48
Fe ₂ O ₃	3.61	3.98	4.24	3.76	3.49
FeO	5.13	5.30	6.70	5.23	6.07
MnO	0.15	0.12	0.26	0.13	0.16
MgO	7.37	3.56	6.41	7.44	5.27
CaO	8.53	8.16	10.27	7.59	12.32
Na ₂ O	3.25	3.67	3.20	3.47	3.49
K ₂ O	3.29	2.84	2.02	2.48	1.47
P ₂ O ₅	0.61	0.49	0.62	0.43	0.41
LOI	1.49	1.03	1.93	1.38	4.8
Mg#	61	41.6	52	60.6	50.5
Ag	0.03	0.00	0.03	0.04	0.02
As	2.24	6.86	2.31	0.81	0.63
V	158	136	203	154	138
Cr	119	216	341	252	201
Ni	100	69	162	151	98
Zn	73	85	112	87	63
Ga	17.90	20.30	20.40	19.90	15.6
Ge	1.87	2.03	1.74	1.71	1.53
Nb	58.24	36.61	46.42	37.80	33.99
Ba	1410	1838	1018	1080	1124
Bi	0.02	0.03	0.01	0.02	0.01
Cd	0.07	0.08	0.08	0.08	0.07
La	43.71	36.62	37.80	36.36	28.36
Ce	85.62	73.20	77.89	69.25	57.61
Pr	9.55	8.60	9.54	7.87	6.85
Nd	38.19	34.39	39.59	31.71	27.83
Sm	6.99	6.48	7.90	6.06	5.51
Eu	2.08	2.03	2.45	1.87	1.71
Gd	5.92	5.67	6.88	5.36	5.24
Tb	0.83	0.78	0.92	0.74	0.78
Dy	4.71	4.38	5.08	4.05	4.80
Ho	0.94	0.83	0.93	0.75	0.95
Er	2.36	2.15	2.20	1.84	2.47
Tm	0.36	0.31	0.30	0.26	0.39
Yb	2.27	1.91	1.84	1.60	2.48
Lu	0.34	0.28	0.27	0.22	0.36
Rb	59.00	72.20	45.70	67.80	18.80
Sb	0.20	0.23	0.24	0.19	0.24
Sr	893	714	753	564	687
Cs	1.58	1.13	1.09	0.86	14.55
Hf	4.93	4.92	5.08	4.38	3.46
Ta	2.57	2.13	2.17	1.90	1.42
Th	4.69	3.96	4.17	5.86	3.26
Tl	0.10	0.11	0.47	0.17	0.06
Mo	1.93	0.72	2.30	1.89	0.94
U	1.30	0.86	0.94	1.38	0.88
Y	25	23	24	20	26
Zr	228	213	222	184	142
Sc	22.44	22.04	22.72	18.95	25.49
Sn	1.33	1.42	1.65	1.34	0.97
Cu	26	14	36	26	23
Pb	4.72	4.92	4.63	7.31	4.33
⁸⁷ Sr/ ⁸⁶ Sr _m	0.704799	0.705329			
⁸⁷ Sr/ ⁸⁶ Sr _i	0.704769	0.705271			
¹⁴³ Nd/ ¹⁴⁴ Nd _m	0.512383	0.512135			
¹⁴³ Nd/ ¹⁴⁴ Nd _i	0.512362	0.512113			
εNd _i	-4.5	-9.3			

Sample	WE24-11	WE24-12	WE24-13	WE24-14	WE24-15
Unit	Upper Dillon	Lower Dillon	Middle Dillon	Middle Dillon	Middle Dillon
SiO ₂	48.92	58.67	52.78	50.87	54.07
TiO ₂	1.25	0.69	1.33	1.35	0.92
Al ₂ O ₃	16.18	16.03	15.56	15.28	16.53
Fe ₂ O ₃	3.63	2.80	3.38	3.34	3.39
FeO	6.06	3.70	4.81	5.28	5.46
MnO	0.16	0.13	0.14	0.15	0.12
MgO	9.88	5.71	7.33	8.26	6.14
CaO	8.89	6.70	7.94	9.38	8.55
Na ₂ O	3.64	3.31	3.51	2.87	2.98
K ₂ O	1.14	2.12	2.56	2.53	1.63
P ₂ O ₅	0.27	0.14	0.64	0.70	0.20
LOI	1.58	0.86	2.67	3.93	4.12
Mg#	65.3	62	62.4	64	56.2
Ag	0.04	0.04	0.02	0.02	0.03
As	1.49	1.47	1.18	1.88	1.62
V	172	131	149	162	156
Cr	349	274	232	267	579
Ni	194	102	104	119	199
Zn	75	71	78	77	81
Ga	16.60	19.00	17.60	18.10	18.60
Ge	2.18	1.36	1.67	2.10	1.73
Nb	22.62	13.37	39.88	36.66	11.65
Ba	701	890	2267	1286	706
Bi	0.004	0.041	0.034	0.025	0.016
Cd	0.07	0.07	0.08	0.08	0.07
La	21.19	36.76	46.51	43.40	37.98
Ce	42.86	66.18	86.29	85.01	60.31
Pr	5.06	7.17	9.78	9.60	7.39
Nd	21.21	25.56	38.59	38.63	28.73
Sm	4.37	4.38	6.91	6.92	5.27
Eu	1.49	1.17	2.07	2.17	1.46
Gd	4.23	3.69	5.84	5.94	4.72
Tb	0.66	0.53	0.77	0.77	0.66
Dy	3.82	3.07	4.23	4.18	4.00
Ho	0.81	0.61	0.81	0.78	0.79
Er	2.12	1.60	2.05	1.99	2.11
Tm	0.32	0.24	0.31	0.30	0.34
Yb	2.01	1.50	1.92	1.80	1.99
Lu	0.31	0.23	0.28	0.27	0.30
Rb	42.70	55.50	73.50	64.30	42.30
Sb	0.14	0.18	0.17	0.25	0.27
Sr	498	447	1220	1115	460
Cs	0.90	1.09	6.11	1.59	0.71
Hf	2.99	4.17	4.15	3.77	3.96
Ta	1.14	0.44	1.99	1.85	0.25
Th	2.26	11.49	5.95	4.88	6.89
Tl	0.16	0.22	0.19	0.25	0.07
Mo	0.76	1.11	1.07	1.98	0.82
U	0.60	2.23	1.42	1.23	1.24
Y	21	16	22	22	23
Zr	127	164	174	156	162
Sc	27.39	18.03	22.42	24.72	29.44
Sn	0.96	1.16	1.23	1.14	0.79
Cu	43	22	22	25	28
Pb	3.70	12.86	7.82	6.67	6.50
⁸⁷ Sr/ ⁸⁶ Sr _m	0.705062		0.707834	0.705916	0.707862
⁸⁷ Sr/ ⁸⁶ Sr _i	0.704988		0.707805	0.705879	0.707793
¹⁴³ Nd/ ¹⁴⁴ Nd _m	0.512226		0.512187	0.512197	0.512061
¹⁴³ Nd/ ¹⁴⁴ Nd _i	0.512202		0.512202	0.512039	0.512039
εNd _i	-7.6		-8.3	-8.1	-10.8

Sample	WE24-16	KB22-1	KB22-2A	KB22-3	AE21-5
Unit	Upper Dillon	Middle Dillon	Middle Dillon	Middle Dillon	Middle Dillon
SiO ₂	45.68	48.94	49.32	49.47	48.48
TiO ₂	3.81	1.27	1.48	1.51	1.41
Al ₂ O ₃	11.59	15.67	17.27	17.31	16.17
Fe ₂ O ₃	3.89	3.53	3.52	3.49	3.49
FeO	8.15	6.11	5.67	5.81	5.56
MnO	0.22	0.16	0.16	0.16	0.15
MgO	11.80	10.20	7.35	7.02	6.46
CaO	10.62	9.38	9.56	9.77	9.13
Na ₂ O	2.13	3.35	3.57	3.15	3.69
K ₂ O	1.16	1.11	1.65	1.83	1.54
P ₂ O ₅	0.95	0.31	0.45	0.47	0.51
LOI	3.75	1.82	2.29	3.05	3.38
Mg#	64.3	66.1	59.7	58.3	56.9
Ag	0.03	-	-	-	-
As	1.17	-	-	-	-
V	266	177	179	180	164
Cr	365	304	158	152	160
Ni	220	197	83	74	120
Zn	119	197	83	74	120
Ga	20.90	-	-	-	-
Ge	1.85	-	-	-	-
Nb	89.68	36.6	42.2	44.4	49.30
Ba	865	702	998	1030	1249
Bi	0.007	-	-	-	-
Cd	0.09	-	-	-	-
La	44.18	22	36	36	43.67
Ce	93.27	45	65	64	83.57
Pr	11.75	-	-	-	9.74
Nd	50.01	22	30	32	37.31
Sm	9.91	4.8	6.0	6.4	6.67
Eu	3.19	-	-	-	2.00
Gd	8.34	-	-	-	5.64
Tb	1.11	-	-	-	0.82
Dy	5.56	5.0	5.7	6.1	4.48
Ho	0.92	-	-	-	0.89
Er	2.12	-	-	-	2.46
Tm	0.27	-	-	-	0.35
Yb	1.59	2.7	2.8	2.3	2.17
Lu	0.22	-	-	-	0.34
Rb	14.7	33	40	41	43.44
Sb	0.19	-	-	-	-
Sr	928	491	649	785	783
Cs	0.62	3	1	2	2.46
Hf	6.78	3.3	4.7	4.2	4.33
Ta	5.55	3	2	3	2.84
Th	4.50	3	4	6	4.76
Tl	0.05	-	-	-	-
Mo	3.20	-	-	-	-
U	1.45	2	-	1	1.23
Y	24	26	27	27	25
Zr	275	140	206	204	193
Sc	22.93	28.6	28.6	28.0	24.71
Sn	2.00	-	-	-	-
Cu	21	-	-	-	-
Pb	3.65	4	2	4	4.30
⁸⁷ Sr/ ⁸⁶ Sr _m	0.704300	0.705730			
⁸⁷ Sr/ ⁸⁶ Sr _i	0.704286	0.705670			
¹⁴³ Nd/ ¹⁴⁴ Nd _m	0.512061	0.512324			
¹⁴³ Nd/ ¹⁴⁴ Nd _i	0.512039	0.512300			
εNd _i	-10.8	-5.7			

Sample	AE21-6	AE21-7	AE21-9	AE21-10	AE21-11
Unit	Middle Dillon	Middle Dillon	Lower Dillon	Lower Dillon	Lower Dillon
SiO ₂	48.51	48.75	53.55	53.18	53.20
TiO ₂	1.32	1.37	0.89	0.86	0.87
Al ₂ O ₃	15.44	16.01	11.71	11.20	11.31
Fe ₂ O ₃	3.82	3.73	4.05	4.14	4.14
FeO	6.02	5.58	5.91	6.12	6.13
MnO	0.16	0.16	0.16	0.17	0.17
MgO	7.28	8.00	10.10	9.84	10.04
CaO	10.09	8.86	8.26	9.11	8.73
Na ₂ O	3.44	3.36	2.39	2.29	2.31
K ₂ O	1.83	2.47	3.09	3.21	3.21
P ₂ O ₅	0.39	0.47	0.30	0.30	0.30
LOI	1.69	1.24	1.57	2.45	2.53
Mg#	57.8	61.5	65.3	64	64.4
Ag	-	-	-	-	-
As	-	-	-	-	-
V	170	165	181	205	185
Cr	236	194	849	1112	1042
Ni	135	125	180	231	227
Zn	135	125	180	231	227
Ga	-	-	-	-	-
Ge	-	-	-	-	-
Nb	35.23	44.48	13.25	13.44	13.69
Ba	997	1067	971	1006	1013
Bi	-	-	-	-	-
Cd	-	-	-	-	-
La	37.25	41.27	24.41	22.20	22.80
Ce	71.75	79.26	47.32	43.42	44.50
Pr	8.47	9.30	-	-	-
Nd	33.09	35.51	23.00	21.54	21.47
Sm	6.07	6.43	4.60	4.16	4.25
Eu	1.84	1.94	1.30	1.23	1.23
Gd	5.32	5.52	4.16	3.93	3.98
Tb	0.78	0.80	0.61	0.59	0.58
Dy	4.24	4.41	3.38	3.43	3.41
Ho	0.82	0.88	0.68	0.69	0.67
Er	2.24	2.40	1.97	1.95	1.93
Tm	0.31	0.34	0.28	0.28	0.27
Yb	1.98	2.16	1.75	1.74	1.75
Lu	0.30	0.32	0.26	0.27	0.26
Rb	46.70	56.87	140.29	146.07	145.50
Sb	-	-	-	-	-
Sr	668	723	381	395	391
Cs	1.05	1.54	1.85	2.15	1.95
Hf	3.67	4.13	3.35	3.17	3.09
Ta	1.96	2.58	0.94	0.98	0.99
Th	3.89	4.63	4.44	4.05	4.21
Tl	-	-	-	-	-
Mo	-	-	-	-	-
U	1.05	1.20	0.69	0.70	0.67
Y	23	24	19	19	18
Zr	155	180	126	116	116
Sc	23.33	24.07	26.16	28.30	26.96
Sn	-	-	-	-	-
Cu	-	-	-	-	-
Pb	4.21	4.42	8.92	8.56	7.17

Sample	AE21-12	Luikart-1 (MS thesis)
Unit	Lower Dillon	Middle Dillon
SiO ₂	48.80	49.33
TiO ₂	0.91	1.47
Al ₂ O ₃	10.82	17.14
Fe ₂ O ₃	3.81	3.34
FeO	6.43	5.57
MnO	0.17	0.16
MgO	15.66	9.07
CaO	8.49	9.30
Na ₂ O	2.84	3.07
K ₂ O	1.98	1.50
P ₂ O ₅	0.46	0.39
LOI	2.38	-
Mg#	74.0	65.4
Ag	-	-
As	-	-
V	172	-
Cr	1162	212
Ni	482	125
Zn	482	125
Ga	-	-
Ge	-	-
Nb	18.01	-
Ba	1172	990
Bi	-	-
Cd	-	-
La	36.74	-
Ce	72.52	-
Pr	-	-
Nd	34.22	-
Sm	5.68	-
Eu	1.52	-
Gd	4.44	-
Tb	0.61	-
Dy	3.29	-
Ho	0.66	-
Er	1.82	-
Tm	0.25	-
Yb	1.67	-
Lu	0.24	-
Rb	106.31	43
Sb	-	-
Sr	740	716
Cs	1.55	-
Hf	2.96	-
Ta	0.96	-
Th	5.52	3
Tl	-	-
Mo	-	-
U	1.32	-
Y	18	26
Zr	113	142
Sc	24.80	24.0
Sn	-	-
Cu	-	-
Pb	8.29	2
⁸⁷ Sr/ ⁸⁶ Sr _m	0.707225	
⁸⁷ Sr/ ⁸⁶ Sr _i	0.707157	
¹⁴³ Nd/ ¹⁴⁴ Nd _m	0.512310	
¹⁴³ Nd/ ¹⁴⁴ Nd _i	0.512290	
εNd _i	-5.9	

Appendix B - Sample locations and Petrographic descriptions

Black Butte

Sample ID: WE24-01

Latitude: 44.904351

Rock Type: Basalt

Longitude: -111.855063

Thin Section Description: Based on thin section description, the rock is holocrystalline, aphanitic, porphyritic, and contains phenocrysts of olivine (13.4%), plagioclase (5.84%), clinopyroxene (2.34%), orthopyroxene (0.13%), and oxides (0.52%). The sample is approximately (23.4%) phenocrysts and (76.60%) groundmass. Mineral abundances were calculated via point counting at 1 mm intervals. Phenocrysts grainsize range from ~ 0.3 to 1.7 mm. Groundmass grains range from ~0.015 to 0.25 mm with plagioclase, olivine, clinopyroxene and oxides. Olivine phenocrysts range from euhedral to anhedral. Clinopyroxene (titanaugite) phenocrysts are mostly euhedral. Groundmass plagioclase is euhedral and tabular. Alteration within the sample is confined to olivine phenocrysts which display iddingsite alteration along fractures and grain boundaries. Some olivine phenocrysts display resorbed margins. Groundmass plagioclase displays pilotaxitic to felty flow textures and polysynthetic twinning.

Black Butte Intrusive

Sample ID: WE24-02

Latitude: 44.904351

Rock Type: Basaltic trachyandesite

Longitude: -111.855063

Thin Section Description: Based on thin section description, the rock is holocrystalline, aphanitic, porphyritic, and contains phenocrysts of plagioclase (29%), anorthoclase (21.1%), sanidine (8%), magnetite (5%), clinopyroxene (5%), olivine, (4.9%), apatite (1.9%), amphibole (1.3%), oxides (0.6%), and chlorite (0.3%). ~ (20%) of the rock is secondary mineral fill within void space and within cracks/grain boundaries. Mineral abundances were calculated via point counting at 1 mm intervals. Phenocrysts grainsize range from ~ 0.57 to 9 mm. Groundmass grains range from ~ 0.05 to 0.15 mm with plagioclase, clinopyroxene, olivine, and oxides. Plagioclase, and sanidine phenocrysts are euhedral. Clinopyroxene (titanaugite), anorthoclase, magnetite and apatite phenocrysts range from euhedral to anhedral. Apatite is visually present within almost every mineral except olivine. Alteration minerals and secondary mineral fill are confined within void spaces, between grain boundaries and cracks. Chlorite is present in the groundmass. Olivine is completely altered.

Black Butte Bomb

Sample ID: WE24-03

Latitude: 44.90270597

Rock Type: Trachybasalt

Longitude: -111.852777

Thin Section Description: Based on thin section description, the rock is holocrystalline, aphanitic, porphyritic, and contains phenocrysts of plagioclase (2.14%), clinopyroxene (1.74%), olivine (0.93%), and oxides (0.4%). The sample is approximately (13.22%) phenocrysts and (86.78%) groundmass. Mineral abundances were calculated via point counting at 1 mm intervals. Phenocrysts grainsizes range from ~ 0.25 to 0.87 mm. Groundmass grains range from ~ 0.03 to 0.1 mm with plagioclase, and oxides. Plagioclase, clinopyroxene (titanaugite), and olivine phenocrysts are euhedral. Alteration within the sample is confined within the olivine phenocrysts which display iddingsite alteration through most of the mineral. A few titanaugite phenocrysts show sieve textures and show twinning and sector zoning. Groundmass plagioclase displays pilotaxitic flow textures and polysynthetic twinning.

Gravelly Range Lava

Sample ID: WE24-04

Latitude: 44.95947702

Rock Type: Basalt

Longitude: -111.867977

Thin Section Description: Based on thin section description, the rock is holocrystalline, aphanitic, porphyritic, and contains phenocrysts of plagioclase (60.9%), clinopyroxene (13.2%), olivine (12.9%), and oxides (3.6%). The sample is approximately (99.2%) phenocrysts and (0.8%) groundmass. Mineral abundances were calculated via point counting at 1 mm intervals. Phenocrysts grainsize range from ~ 0.05 to 0.25mm. Groundmass grains are < 0.05 mm with plagioclase, clinopyroxene, olivine and oxides. Plagioclase phenocrysts are euhedral. Olivine and clinopyroxene (titanaugite) phenocrysts, are euhedral to subhedral. Slight alteration occurs within the groundmass and individual olivine phenocrysts with chlorite being present in the groundmass and iddingsite occurring along olivine grain boundaries and within cracks. A few titanaugite phenocrysts show sector zoning. Individual olivine phenocrysts contain brown spinel inclusions. Groundmass plagioclase is unoriented and displays polysynthetic twinning.

West Lion Mountain Dike

Sample ID: WE24-05

Latitude: 44.87933404

Rock Type: Trachybasalt

Longitude: -111.800937

Thin Section Description: Based on thin section description, the rock is holocrystalline, aphanitic, porphyritic, and contains phenocrysts of olivine (1.2%). The sample is approximately (1.2%) phenocryst (98%) groundmass, and (0.5%) quartz xenocrysts. Mineral abundances were calculated via point counting at 1 mm intervals. Phenocrysts grainsizes range from ~ 0.05 to 0.20 mm. Groundmass grains are < 0.05 mm with plagioclase, clinopyroxene, olivine, and oxides. Olivine phenocrysts are euhedral to subhedral. Individual olivine phenocrysts display iddingsite

alteration along grain boundaries, cracks and throughout entire individual phenocrysts. Individual olivine phenocrysts show sieve textures and embayment. Secondary mineral fill, calcite, is found filling between grain boundaries in the matrix but primarily filling vesicles (amygdaloidal). Quartz xenocrysts exhibit reaction rims composed of clinopyroxene. Groundmass plagioclase displays some pilotaxitic flow textures and polysynthetic twinning.

Columnar Plug – West of Lion Mountain

Sample ID: WE24-06

Latitude: 44.87630398

Rock Type: Trachybasalt

Longitude: -111.815257

Thin Section Description: Based on thin section description, the rock is holocrystalline, aphanitic, porphyritic, and contains phenocrysts of olivine (7.7%), and oxides (0.1%). The sample is approximately (7.8%) phenocrysts and (92.2%) groundmass. Mineral abundances were calculated via point counting at 1 mm intervals. Phenocrysts grainsizes range from ~ 0.05 to 0.30 mm. Groundmass grains are < 0.05 mm with plagioclase, clinopyroxene, olivine, and oxides. Olivine phenocrysts are euhedral to subhedral. Few olivine phenocrysts display iddingsite alteration within cracks. Some individual olivine phenocrysts show sieve textures and embayment. Slight chlorite alteration is observed in the groundmass. Groundmass plagioclase displays trachytic flow texture and polysynthetic twinning.

Upper Lion Mountain Lava

Sample ID: WE24-07

Latitude: 44.86092503

Rock Type: Basaltic trachyandesite

Longitude: -111.769938

Thin Section Description: Based on thin section description, the rock is holocrystalline, aphanitic, porphyritic, and contains phenocrysts of olivine (5.3%), and clinopyroxene (0.5%). The sample is approximately (10.5%) phenocrysts and (89.5%) groundmass. Mineral abundances were calculated via point counting at 1 mm intervals. Phenocrysts grainsizes range from ~ 0.05 to 0.25 mm. Groundmass grains are < 0.05 mm with plagioclase, clinopyroxene and oxides. Olivine and oxide phenocrysts are euhedral to subhedral. Olivine phenocrysts show high alteration with many showing complete replacement of iddingsite. Little to no alteration in the groundmass. At least one clinopyroxene and plagioclase rich xenolith. The sample is highly vesicular with groundmass plagioclase displaying pilotaxitic flow texture and polysynthetic twinning.

Lower Lion Mountain Lava

Sample ID: WE24-08

Latitude: 44.86001702

Rock Type: Trachybasalt

Longitude: -111.767321

Thin Section Description: Based on thin section description, the rock is holocrystalline, aphanitic, porphyritic, and contains phenocrysts of olivine (10.2%), and plagioclase (1.7%). The sample is approximately (14) phenocrysts, (86%) groundmass, and (0.1%) quartz xenocrysts. Mineral abundances were calculated via point counting at 1 mm intervals. Phenocrysts grainsizes range from ~0.05 to 0.30 mm. Groundmass grains are < 0.05 mm with plagioclase, clinopyroxene, olivine and oxides. Plagioclase, olivine, and clinopyroxene phenocrysts are euhedral to subhedral. Olivine phenocrysts show high iddingsite alteration throughout the grains. Olivine phenocrysts show sieve textures with some containing melt inclusions. Slight alteration is observed in the groundmass. Individual plagioclase phenocrysts contain apatite inclusions. Groundmass plagioclase is unoriented and shows polysynthetic twinning.

Wolverine Basin

Sample ID: WE24-09

Latitude: 44.86339501

Rock Type: Basaltic trachyandesite

Longitude: -111.753341

Thin Section Description: Based on thin section description, the rock is holocrystalline, aphanitic, porphyritic, and contains phenocrysts of olivine (6.7%), and clinopyroxene (0.4%). The sample is approximately (7.1%) phenocrysts and (92.9%) groundmass. Mineral abundances were calculated via point counting at 1 mm intervals. Phenocrysts grainsizes range from ~ 0.05 to 0.50 mm. Groundmass grains range is < 0.05 mm with plagioclase, clinopyroxene, olivine and oxides. Olivine and clinopyroxene phenocrysts are euhedral to subhedral. Olivine phenocrysts show iddingsite alteration along grain boundaries and within cracks. Olivine phenocrysts show embayment. Clinopyroxene phenocrysts show sieve textures. Chlorite alteration is observed in the groundmass along with secondary mineral fill, calcite, filling between mineral grains and within vesicles. Groundmass plagioclase displays pilotaxitic flow textures and polysynthetic twinning.

Verney Lava

Sample ID: WE24-10

Latitude: 45.23430998

Rock Type: Basalt

Longitude: -111.832216

Thin Section Description: Based on thin section description, the rock is holocrystalline, aphanitic, porphyritic, and contains phenocrysts of olivine (11.9%), plagioclase (4.5%), and clinopyroxene (0.1%). The sample is approximately (18.2%) phenocrysts (81.8%) groundmass, and (0.1%) quartz xenocrysts. Mineral abundances were calculated via point counting at 1 mm intervals. Phenocrysts grainsizes range from ~ 0.05 to 0.30 mm. Groundmass grains are < 0.05 mm with plagioclase, clinopyroxene, olivine and oxides. Plagioclase and olivine phenocrysts are euhedral to subhedral. Olivine phenocrysts show complete alteration to iddingsite. Olivine phenocrysts show embayment. The groundmass shows high secondary mineral fill, calcite, filling between grain boundaries and vesicles (amygdaloidal). Xenocrysts of quartz within the

sample with a clinopyroxene reaction rim. Groundmass plagioclase is unoriented and displays polysynthetic twinning.

Madison Valley Lava

Sample ID: WE24-11

Latitude: 45.14437501

Rock Type: Basalt

Longitude: -111.593578

Thin Section Description: Based on thin section description, the rock is holocrystalline, aphanitic, porphyritic, and contains phenocrysts of olivine (15.8%), plagioclase (5.5%), clinopyroxene (0.6%), and oxides (1.6%). The sample is approximately (23.8%) phenocrysts and (76.2%) groundmass. Mineral abundances were calculated via point counting at 1 mm intervals. Phenocrysts grainsizes range from ~ 0.05 to 0.55 mm. Groundmass grains are < 0.05 mm with plagioclase, clinopyroxene, olivine and oxides. Plagioclase phenocrysts are euhedral with olivine phenocrysts being euhedral to subhedral. Olivine phenocrysts show slight iddingsite alteration along grain boundaries and within cracks. Individual olivine phenocrysts show embayment. Portions of groundmass plagioclase show pilotaxitic flow textures and polysynthetic twinning.

Block Mountain Lava

Sample ID: WE24-12

Latitude: 45.44482299

Rock Type: Andesite

Longitude: -112.528776

Thin Section Description: Based on thin section description, the rock is holocrystalline, aphanitic, porphyritic, and contains phenocrysts of olivine (2.9%), clinopyroxene (0.7%), and plagioclase (0.1%). The sample is approximately (5.1%) phenocrysts, (94.9%) groundmass, and (1.5%) quartz xenocrysts. Mineral abundances were calculated via point counting at 1 mm intervals. Phenocrysts grainsizes range from ~ 0.05 to 0.30 mm. Groundmass grains are < 0.025 mm with plagioclase, clinopyroxene, olivine and oxides. Olivine phenocrysts are subhedral to anhedral. Olivine phenocrysts show complete alteration to iddingsite. Groundmass olivines are mostly intact with little to no groundmass alteration. Quartz xenocrysts show reaction rims and embayment. Plagioclase groundmass displays pilotaxitic flow textures and polysynthetic twinning.

North Red Butte Lava

Sample ID: WE24-13

Latitude: 45.02287396

Rock Type: Basaltic trachyandesite

Longitude: -113.228991

Thin Section Description: Based on thin section description, the rock is holocrystalline, aphanitic, porphyritic, and contains phenocrysts of olivine (10.1%), plagioclase (8.6%),

clinopyroxene (5.7%), and orthopyroxene (0.1%). The sample is approximately (24.8%) phenocrysts and (75.2%) groundmass. Mineral abundances were calculated via point counting at 1 mm intervals. Phenocrysts grainsizes range from ~ 0.05 to 0.30 mm. Groundmass grains are < 0.05 mm with plagioclase, clinopyroxene, olivine, and oxides. Plagioclase phenocrysts are euhedral. Clinopyroxene (titanaugite) phenocrysts are euhedral to subhedral. Glomeroporphyritic texture is observed in the sample. Olivine phenocrysts are subhedral to anhedral. Olivine phenocrysts display iddingsite alteration along grain boundaries and within cracks. Groundmass plagioclase groundmass displays pilotaxitic flow textures and polysynthetic twinning.

Bannack Bench Road Columnar Lava

Sample ID: WE24-14

Latitude: 45.10893096

Rock Type: Trachybasalt

Longitude: -113.050473

Thin Section Description: Based on thin section description, the rock is holocrystalline, aphanitic, porphyritic, and contains phenocrysts of clinopyroxene (19.8%), olivine (10.3%), plagioclase (7.2%), and oxide (0.6%). The sample is approximately (39.9%) phenocrysts and (60.1%) groundmass. Mineral abundances were calculated via point counting at 1 mm intervals. Phenocrysts grainsizes range from ~ 0.05 to 0.25 mm. Groundmass grains are < 0.05 mm with plagioclase, clinopyroxene, olivine, and oxides. Plagioclase, Clinopyroxene (titanaugite), and olivine phenocrysts are euhedral to subhedral. Glomeroporphyritic texture is observed in the sample. Olivine phenocrysts display iddingsite alteration along grain boundaries and within cracks. The groundmass shows slight alteration with chlorite present in the groundmass. Sparce vesicles in the sample are filled with secondary minerals (amygdaloidal). Groundmass plagioclase is randomly orientated and displays polysynthetic twinning.

Dyce Creek Road Lava

Sample ID: WE24-15

Latitude: 45.27495501

Rock Type: Basaltic andesite

Longitude: -113.031841

Thin Section Description: Based on thin section description, the rock is holocrystalline, aphanitic, porphyritic, and contains phenocrysts of olivine (16.1%), oxides (0.4%), and clinopyroxene (0.1%). The sample is approximately (16.6%) phenocrysts and (83.4%) groundmass. Mineral abundances were calculated via point counting at 1 mm intervals. Phenocrysts grainsizes range from ~ 0.05 to 0.55 mm. Groundmass grains are < 0.025 mm with plagioclase, clinopyroxene, and olivine. Olivine phenocrysts are euhedral to subhedral. At least one glomerocryst composed of subhedral-anhedral clinopyroxene. All olivine phenocrysts show complete alteration to iddingsite. Some olivine phenocrysts show embayment textures. Olivine in groundmass also displays iddingsite alteration. Groundmass plagioclase groundmass displays pilotaxitic flow textures and polysynthetic twinning.

Big Hole River Valley Lava

Sample ID: WE24-16

Latitude: 45.864928

Rock Type: Basalt

Longitude: -113.347134

Thin Section Description: Based on thin section description, the rock is holocrystalline, aphanitic, porphyritic, and contains phenocrysts of olivine (14.7%), clinopyroxene (14.2%), plagioclase (0.7%), and oxides (0.4%). The sample is approximately (32.4%) phenocrysts, (67.6%) groundmass, and (0.7%) quartz xenocrysts. Mineral abundances were calculated via point counting at 1 mm intervals. Phenocrysts grainsizes range from ~ 0.05 to 0.50 mm. Groundmass grains are < 0.05 mm with plagioclase, clinopyroxene, olivine, and oxides. Olivine and clinopyroxene (titanaugite) phenocrysts are euhedral to subhedral. Olivine phenocrysts show iddingsite and chlorite alteration along grain boundaries and within cracks. Clinopyroxene (titanaugite) phenocrysts show abundant twinning, zoning, and a pinkish rim visible in plain polarized light in phenocrysts. The sample is also glomeroporphyritic with 1 large glomerocryst (~ 1 by 0.50 mm) composed almost entirely of titanaugite. A large quartz xenocryst (~ 0.65 mm) displays a reaction rim composed of clinopyroxene. The groundmass shows moderate alteration between mineral grains. Chlorite is the dominate alteration mineral in the groundmass. Groundmass plagioclase is randomly orientated and displays polysynthetic twinning.

Gravelly Range

Sample ID: KB22-1

Latitude: 44.814818

Rock Type: Basalt

Longitude: - 111.878957

Thin Section Description: Based on thin section description, the rock is holocrystalline, aphanitic, porphyritic, and contains phenocrysts of olivine (13.7%), oxides (2.3%), and plagioclase (0.3%). Mineral abundances were calculated via point counting at 1 mm intervals. Phenocrysts grainsizes range from ~ 0.05 to 0.30 mm. Groundmass grains are < 0.05 mm with plagioclase, clinopyroxene, olivine, and oxides. Phenocrysts are euhedral to subhedral. Olivine phenocrysts show slight iddingsite and chlorite alteration along grain boundaries and within cracks. Some individual olivine phenocrysts show embayment textures and one clinopyroxene phenocrysts display a high sieve texture. Plagioclase phenocrysts show zoning. Groundmass shows slight alteration of olivine. Groundmass plagioclase groundmass displays pilotaxitic flow textures and polysynthetic twinning.

Gravelly Range

Sample ID: KB22-2A

Latitude: 44.836373

Rock Type: Trachybasalt

Longitude: -111.866716

Thin Section Description: Based on thin section description, the rock is holocrystalline, aphanitic, porphyritic, and contains phenocrysts of plagioclase (28%), olivine (8.9%), clinopyroxene (4.9%), and oxides (2%). Mineral abundances were calculated via point counting

at 1 mm intervals. Phenocrysts grainsizes range from ~ 0.05 to 0.55 mm. Groundmass grains are < 0.05 mm with plagioclase, clinopyroxene, olivine, and oxides. Plagioclase, clinopyroxene (titanaugite), and olivine phenocrysts are euhedral to subhedral. Plagioclase phenocrysts display zoning and some have apatite inclusions. Clinopyroxene (titanaugite) phenocrysts display, twinning, zoning, sector zoning and pink rims on larger phenocrysts while smaller phenocrysts are completely pink when observed in plane polarized light. Olivine phenocrysts contain brownish green spinel and display iddingsite alteration along grain boundaries and within cracks. The sample is glomerophyritic and one xenolith is present in the sample and is composed of all phenocryst compositions previously stated. Groundmass shows slight alteration primarily as chlorite. Groundmass plagioclase felty and shows some pilotaxitic flow textures and polysynthetic twinning.

Gravelly Range

Sample ID: KB22-2B

Latitude: 44.836373

Rock Type:

Longitude: -111.866716

Thin Section Description: Based on thin section description, the rock is holocrystalline, aphanitic, porphyritic, and contains phenocrysts of olivine (19%), oxides (1.7%), and plagioclase (0.2%). Mineral abundances were calculated via point counting at 1 mm intervals. Phenocrysts grainsizes range from ~ 0.05 to 0.50 mm. Groundmass grains are < 0.05 mm with plagioclase, clinopyroxene, olivine, and oxides. Plagioclase and olivine phenocrysts are euhedral to subhedral. Plagioclase phenocrysts display zoning. Olivine phenocrysts contain dark brown spinel and display iddingsite alteration along grain boundaries and within cracks. Individual olivine phenocrysts display embayment textures. Alteration in the groundmass is limited to small secondary mineral veins which follow the direction of flow. Sparse vesicles have been filled by secondary minerals (amygdaloidal). Groundmass plagioclase displays pilotaxitic flow textures and polysynthetic twinning.

Gravelly Range

Sample ID: KB22-3

Latitude: 44.891622

Rock Type: Trachybasalt

Longitude: -111.837658

Thin Section Description: Based on thin section description, the rock is holocrystalline, aphanitic, porphyritic, and contains phenocrysts of plagioclase (16.9%), olivine (8.5%), and clinopyroxene (3.5%). Mineral abundances were calculated via point counting at 1 mm intervals. Phenocrysts grainsizes range from ~ 0.05 to 0.55 mm. Groundmass grains are < 0.05 mm with plagioclase, clinopyroxene, olivine, and oxides. Plagioclase, clinopyroxene (titanaugite), and olivine phenocrysts are euhedral to subhedral. Plagioclase phenocrysts display zoning, and individual phenocrysts contain clinopyroxene. Clinopyroxene (titanaugite) phenocrysts display twinning, zoning, and sector zoning. Olivine phenocrysts contain yellow brown spinel and display iddingsite alteration along grain boundaries and within cracks. The sample is glomerophyritic and xenoliths are present in the sample and are composed of all phenocryst

compositions previously stated. The groundmass shows slight alteration with chlorite present. Groundmass plagioclase displays pilotaxitic flow textures and polysynthetic twinning.

Centennial Valley

Sample ID: AE21-5

Latitude: 44.652350

Rock Type: Trachybasalt

Longitude: -112.07269

Thin Section Description: Based on thin section description, the rock is holocrystalline, aphanitic, porphyritic, and contains phenocrysts of olivine (10.9%), plagioclase (0.1%) and oxides (0.1%). Mineral abundances were calculated via point counting at 1 mm intervals. Phenocrysts grainsize range from ~ 0.05 to 0.25 mm. Groundmass grains are < 0.05 mm with plagioclase, clinopyroxene, olivine, and oxides. Olivine phenocrysts are euhedral to subhedral. Individual olivine phenocrysts display iddingsite alteration along grain boundaries, cracks and throughout entire individual phenocrysts. Individual olivine phenocrysts show sieve textures and embayment. Secondary mineral fill, calcite, is found filling between grain boundaries in the matrix but primarily filling vesicles (amygdaloidal). Quartz xenocrysts exhibit reaction rims composed of clinopyroxene. Groundmass plagioclase displays some pilotaxitic flow textures and polysynthetic twinning.

Centennial Valley

Sample ID: AE21-6

Latitude: 44.644610

Rock Type: Trachybasalt

Longitude: -112.017340

Description: Very fine-grained grey matrix, abundant 1-2 mm plagioclase in matrix. Olivine is iddingsite, but one piece has fresh olivine. Pyroxene 2-3 mm. Some relations with gravels above. Lava is ~20 ft thick.

Centennial Valley

Sample ID: AE21-7

Latitude: 44.652350

Rock Type: Trachybasalt

Longitude: -112.072690

Description: Very fine-grained platy outcrop with weathered surface. Fresh 2-3 mm olivine, 1-2 mm black pyroxene, plagioclase in matrix. No vesicles.

Lima Montana

Sample ID: AE21-9

Latitude: 44.633200

Rock Type: Basaltic trachyandesite

Longitude: -112.592500

Description: Very fine grained, dark gray matrix. Plagioclase in matrix is 1 mm. Olivine rich with some displaying iddingsite

Briggs Ranch Quad

Sample ID: AE21-10

Latitude: 44.798361

Rock Type: Basaltic trachyandesite

Longitude: -112.637440

Description: Very fine-grained reddish matrix. Contains clinopyroxene and olivine. Rock contains filled vesicles and granitoid xenoliths.

Briggs Ranch Quad

Sample ID: AE21-11

Latitude: 44.798850

Rock Type: Basaltic trachyandesite

Longitude: -112.633380

Description: Very fine grained dark grey matrix with abundant 1-3 mm orange olivine, euhedral plagioclase matrix and some pyroxenes. Sample collected from upper pile of 4-5 lavas.

Briggs Ranch Quad

Sample ID: AE21-12

Latitude: 44.787961

Rock Type: Basalt

Longitude: -112.633480

Description: Very fine grained dark gray blackish matrix. 1-4 mm oxidized olivine. The plagioclase in matrix is up to 2 mm.

Appendix C - AFC Modeling and Mixing Mass Balance

Assimilation-Fractional Crystallization (AFC) modeling from DePaolo, (1981) to model assimilation of crustal materials (upper, middle, and lower crust) into a parent magma (WE24-1) to then compare best fits to other samples.

$$\frac{C_m}{C_o} = F^{-z} + \left[\frac{r}{r-1} \times \frac{C_a}{zC_o} \times (1 - F^{-z}) \right]$$

$$\text{Where } Z = \frac{r+D-1}{r-1}$$

And r = rate of assimilation (M_a)/rate of fractional crystallization (M_c)

C_o and C_m = initial and final concentration of the trace element

D = the bulk distribution coefficient of the trace element

M_o and M_m = initial and final mass of magmas

F = Mass fraction of magma remaining = M_m/M_o

Bulk distribution coefficient of the trace elements determined using mineral modes (%) (Table 1), calculated via point counting mineral phenocrysts at 1 mm increments of WE24-1 thin section. Mineral modes (%) were then normalized to sum to 100%. Bulk distribution coefficients of the trace elements were then calculated for each phenocrystal mineral phases (e.g., olivine, plagioclase, clinopyroxene, etc) using Rollinson and Pease (2021).

Symbol	Description	Value/Range	Notes
C_o	Initial parent melt concentration	See Table 1	Used WE24-1 for this study
C_a	Assimilate concentration	See Table 1	Experimented with upper, middle, lower crustal materials
r	Assimilation/crystallization ratios	0.005-0.05	Best fit for this study
F	Melt fraction remaining	.90	Variable (1-0.1), .90 best fit for this study

Calculations were done in Microsoft Excel with results (Cl) plotted in Igpet Rock/Primitive Mantle (Sun and McDonough, 1989)

WE24-1	Olivine	Plagioclase	Clinopyroxene	Orthopyroxene	Sum
Original Mineral modes Measurement	0.134	0.0584	0.0234	0.0013	0.2171
Normalized	0.6172271	0.269000461	0.107784431	0.005988024	1

Normalized mineral modes used to calculate bulk D of each trace element

(Example)

Bulk D for Rb = 0.0321939 =

$(\text{olivine} \times 0.007) + (\text{orthopyroxene} \times 0.003) + (\text{clinopyroxene} \times 0.01) + (\text{plagioclase} \times 0.114)$

	Rb		F	F^{-z}	r/(r-1)	Ca/ZCo	1-F^{-z}	Cl
Co	40.99422197		1	1.00	-0.01	0.76	0.00	40.99
r	0.005		0.9	1.11	-0.01	0.76	-0.11	45.41
D	0.03219392		0.8	1.24	-0.01	0.76	-0.24	50.91
Ca	30.00		0.7	1.41	-0.01	0.76	-0.41	57.96
z	0.97		0.6	1.64	-0.01	0.76	-0.64	67.30
			0.5	1.96	-0.01	0.76	-0.96	80.32
			0.4	2.43	-0.01	0.76	-1.43	99.71
			0.3	3.21	-0.01	0.76	-2.21	131.77
			0.2	4.75	-0.01	0.76	-3.75	195.15
			0.1	9.28	-0.01	0.76	-8.28	381.80

$$Z = ((r + D - 1)/(r - 1))$$

Trace elements	Olivine	Plagioclase	Clinopyroxene	Orthopyroxene
Cs	0.0007	0.596	0	0
Rb	0.0007	0.114	0.01	0.003
Ba	0.000023	0.247	0.0002	0.000006
Th	0.0018	0.305	0.013	0.00002
U	0.0013	0.0107	0.006	0.00004
Nb	0.004	0.097	0.0037	0.0013
Ta	0.03	0.075	0.0239	0.0025
K	0.006845	0.203	0.007	0.0003
La	0.000022	0.163	0.086	0.0008
Ce	0.000045	0.096	0.175	0.0016
Pb	0.0013	1.592	0.009	0.0013
Pr	0.000095	0.132	0.289	0.0032
Sr	0.000138	1.629	0.088	0.0012
P	0.019	0.065	0.05	0.03
Nd	0.0002	0.0908	0.47	0.0056
Zr	0.008	0.001	0.1115	0.032
Sm	0.000636	0.106	0.81	0.015
Eu	0.0012	0.1005	1	0.03
Ti	0.0096	0.038	0.38	0.759
Dy	0.004	0.0293	1.4	0.077
Y	0.00719	0.0371	0.67	0.095
Yb	0.0188	0.011	1.42	0.22
Lu	0.028	0.0093	1.16	0.22

Table 4: Mineral-melt partition coefficients for trace elements in basaltic melts (Rollinson and Pease (2021)).