Oligocene and Miocene arc volcanism in northeastern California: Evidence for post-Eocene segmentation of the subducting Farallon plate

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ABSTRACT

The Warner Range in northeastern California exposes a section of Tertiary rocks over 3 km thick, offering a unique opportunity to study the long-term history of Cascade arc volcanism in an area otherwise covered by younger volcanic rocks. The oldest locally sourced volcanic rocks in the Warner Range are Oligocene (28–24 Ma) and include a sequence of basalt and basaltic andesite lava flows overlain by hornblende and pyroxene andesite pyroclastic flows and minor lava flows. Both sequences vary in thickness (0.2 km) along strike and are inferred to be the erosional remnants of one or more large, partly overlapping composite volcanoes. No volcanic rocks were erupted in the Warner Range between ca. 24 and 16 Ma, although minor distally sourced silicic tuffs were deposited during this time. Arc volcanism resumed ca. 16 Ma with eruption of basalt and basaltic andesite lavas sourced from eruptive centers 5–10 km south of the relict Oligocene centers. Post–16 Ma arc volcanism continued until ca. 8 Ma, forming numerous eroded but well-preserved shield volcanoes to the south of the Warner Range. Oligocene to Late Miocene volcanic rocks in and around the Warner Range are calc-alkaline basalts to andesites (48%–61% SiO2) that display negative Ti, Nb, and Ta anomalies in trace element spider diagrams, consistent with an arc setting. Middle Miocene lavas in the Warner Range are distinctly different in age, composition, and eruptive style from the nearby Steens Basalt, with which they were previously correlated. Middle to Late Miocene shield volcanoes south of the Warner Range consist of homogeneous basaltic andesites (53%–57% SiO2) that are compositionally similar to Oligocene rocks in the Warner Range. They are distinctly different from younger (Late Miocene to Pleistocene) high-Al, low-K olivine tholeiites, which are more mafic (46%–49% SiO2), did not build large edifices, and are thought to be related to backarc extension. The Warner Range is ~100 km east of the axis of the modern arc in northeastern California, suggesting that the Cascade arc south of modern Mount Shasta migrated west during the Late Miocene and Pliocene, while the arc north of Mount Shasta remained in essentially the same position. We interpret these patterns as evidence for an Eocene to Miocene tear in the subducting slab, with a more steeply dipping plate segment to the north, and an initially more gently dipping segment to the south that gradually steepened from the Middle Miocene to the present.

INTRODUCTION

The Cascade volcanic arc in Oregon, Washington, and northernmost California (Fig. 1) has been established close to its present location since the Eocene. Eocene to Late Miocene remnants of the arc consist of a swath of volcanic rocks just west of the modern arc; referred to as the western Cascades (Fig. 1) (e.g., McBirney, 1978; Priest, 1990; Smith, 1993; Sherrod and Smith, 2000; du Bray et al., 2006). A swath of Oligocene to Pliocene volcanic rocks in western Nevada and eastern California (Fig. 1) has been interpreted as a southern continuation of this arc, the ancestral Cascades, active during subduction of the Farallon plate beneath North America prior to northward migration of the Mendocino Triple Junction (Fig. 1) (e.g., Noble, 1972; Christiansen and Yeats, 1992; Putirka and Busby, 2007; Cousins et al., 2008; Busby et al., 2008a, 2008b; Hagan et al., 2009; Busby and Putirka, 2009).

Unlike the western Cascades, however, volcanic rocks of the ancestral Cascades are a subset of a diverse and widespread suite of Cenozoic volcanic rocks erupted across the Basin and Range Province since the Eocene. The ancestral Cascades samples plotted in Figure 1 are those considered by du Bray et al. (2009) to be plausible constituents of the arc based on a variety of criteria, including composition, eruptive style, and location (see du Bray et al., 2009, for a complete list), but they acknowledged that “no clear-cut definition distinguishes constituents of the southern segment of the ancestral Cascades magmatic arc” (p. 3), i.e., from similar-age volcanic rocks related to other tectonic processes, notably the major pulse of mid-Tertiary magmatism thought to result from delamination of the shallow east-dipping Farallon slab (e.g., Armstrong and Ward, 1991; Best et al., 1989; Humphreys, 1995; Henry et al., 2009). Thus, although the existence of an ancestral Cascades arc south of Mount Lassen is a straightforward consequence of Pacific (Farallon)—North American plate interaction, the southern segment of the arc is sufficiently different from the western Cascades that the relationship between the two is not straightforward. Glazner and Farmer (2008), for example, proposed that no Cascade arc ever existed south of Lassen Peak, although recent studies of the ancestral Cascades have concluded otherwise (e.g., Putirka and Busby, 2007; Cousins et al., 2008; Busby et al., 2008a, 2008b; Hagan et al., 2009; Busby and Putirka, 2009).

Volcanic rocks of the inferred ancestral Cascade arc trail off to the north into a poorly mapped area of northern California more than 100 km east of the well-defined western Cascades, which end to the northwest of Mount Shasta (Fig. 1). The region of northeastern California between the inferred ancestral Cascade arc and the undisputed western Cascade arc is mostly covered by Pliocene and younger...
lavas (Fig. 2), and the pre-Pliocene history of the area is largely obscured. Near the Nevada-Oregon-California border, however, the Warner Range (Figs. 2 and 3) exposes nearly 3 km of Tertiary strata as old as Eocene, offering a window into the pre-Miocene history of northeastern California. Although considered part of the Basin and Range Province, the plateau east of the Warner Range in northwestern Nevada is largely unextended (Colgan et al., 2006; Lerch et al., 2008). It is covered by Miocene and younger lava flows and ash-flow tuffs (Middle Miocene bimodal volcanic rocks, Fig. 2); the Middle Miocene tuffs have been inferred to be the earliest eruptions of the Yellowstone hotspot (e.g., Pierce and Morgan, 1992). West of the Warner Range, the Modoc Plateau is mostly covered by flat-lying Pliocene and younger lava flows (Pliocene-Quaternary volcanic rocks, Fig. 2) and has not been broken up by major post-Miocene extension (McKee et al., 1983). Miocene volcanic rocks as old as 14–15 Ma are exposed south of the Warner Range (Middle to Late Miocene volcanic arc rocks, Fig. 2) (Grose, 2000), and similar Miocene volcanic rocks probably extend southeast into northwestern Nevada (Fig. 2), although they have not been mapped in detail.

Pre-Tertiary basement rocks are not exposed in the Warner Range. In the Klamath Mountains (~125 km to the west), basement consists of a complex tectonic assemblage of Paleozoic and Mesozoic oceanic-affinity terranes intruded by Mesozoic plutons (e.g., Snoke and Barnes, 2006, and references therein). In northwestern Nevada (100 km to the east), basement consists of abundant Cretaceous and minor Jurassic granitic plutons intruding Paleozoic and Mesozoic metasedimentary rocks (e.g., Wyld and Wright, 2001). The nearest exposed basement to the south (80–100 km away) consists of Cretaceous granite (e.g., Grose et al., 1992). A seismic refraction profile across northwestern Nevada and northeastern California (including the Warner Range) imaged a low-velocity zone beneath northwestern Nevada (~6.0 km/s) extending to ~16 km depth in the upper crust that Lerch et al. (2007) interpreted as the northern extension of the Mesozoic (Sierra Nevada) batholith beneath Cenozoic cover. Although the western edge of this low-velocity zone is not sharply defined in the seismic data, it is well east of the Warner Range. Pre-Tertiary basement beneath the Warner Range and Modoc Plateau most likely consists of accreted crust similar to that exposed in the Klamath Mountains and northwestern Nevada, rather than the Cretaceous Sierra Nevada batholith, which forms the basement to the ancestral Cascades further south.

REGIONAL GEOLOGIC SETTING

The Warner Range marks the western boundary of the Basin and Range Province in northeastern California and was formed by Miocene and younger slip on the Surprise Valley fault (Russell, 1928; Duffield and McKee, 1986; Carmichael et al., 2006; Colgan et al., 2008; Egger et al., 2009). Faulting and uplift of the Warner Range have exposed more than 3 km of Tertiary rocks as old as Eocene, providing a well-exposed but spatially limited window into the pre-Miocene geology of northeastern California. Although considered part of the Basin and Range Province, the plateau east of the Warner Range in northwestern Nevada is largely unextended (Colgan et al., 2006; Lerch et al., 2008). It is covered by Miocene and younger lava flows and ash-flow tuffs (Middle Miocene bimodal volcanic rocks, Fig. 2); the Middle Miocene tuffs have been inferred to be the earliest eruptions of the Yellowstone hotspot (e.g., Pierce and Morgan, 1992). West of the Warner Range, the Modoc Plateau is mostly covered by flat-lying Pliocene and younger lava flows (Pliocene-Quaternary volcanic rocks, Fig. 2) and has not been broken up by major post-Miocene extension (McKee et al., 1983). Miocene volcanic rocks as old as 14–15 Ma are exposed south of the Warner Range (Middle to Late Miocene volcanic arc rocks, Fig. 2) (Grose, 2000), and similar Miocene volcanic rocks probably extend southeast into northwestern Nevada (Fig. 2), although they have not been mapped in detail.

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WARNER RANGE VOLCANIC ROCKS

Eocene to Oligocene Volcanic and Sedimentary Rocks

The oldest rocks exposed in the Warner Range are Eocene to Early Oligocene sedimentary and sparse volcanic rocks that have no formal
designated, but have been divided into various informal subunits by different authors (Russell, 1928; Martz, 1970; Duffield and Weldin, 1976; Duffield and McKee, 1986; Myers, 1998; Egger et al., 2009). We show this unit as undivided Ts in Figure 3, but it can be broadly subdivided into three lithologic packages. (1) The basal unit is minimally exposed but contains several hornblende andesite lava flows, one of which has been dated to 41 Ma (Axelrod, 1966), indicating minor local volcanic activity as early as the Eocene. The base of the Tertiary section is not exposed but presumably rests on Mesozoic and/or Paleozoic basement. (2) The middle unit consists of as much as 1500 m of Late Eocene to Oligocene sandstone, conglomerate, and lacustrine sedimentary rocks, from which Duffield and McKee (1986) reported dates of 34 and 31 Ma on interbedded volcanic ash. Rare clasts of Mesozoic and Paleozoic basement are present in this unit (Colgan et al., 2008; Egger et al., 2009), but the vast majority of conglomerate clasts are volcanic (predominantly andesite with conspicuous phenocrysts of hornblende and plagioclase). In Egger et al. (2009), it was suggested that this unit records erosion and redeposition from a nearby volcanic arc to the south. (3) The upper unit consists of 500–740 m of reddish-weathering volcaniclastic rocks called the Lost Woods formation by Martz (1970). These rocks consist primarily of angular unsorted lava blocks to several meters across supported by a fine-grained matrix. Blocks consist of dark, generally phenocryst poor lava. The unit contains abundant fragments of petrified wood, including whole logs several meters long. These deposits have not been directly dated but they overlie rocks younger than 33.4 Ma and are overlain by rocks as old as 27.5 Ma. They are interpreted as lahars and indicate volcanic activity in the general area of northeastern California and southern Oregon in the Late Oligocene (Duffield and McKee, 1986; Egger et al., 2009).

**Oligocene Volcanic Rocks**

**Lake City Basalts (28–26.5 Ma)**

North of Cedar Pass, Oligocene volcanioclastic rocks are overlain by a sequence of mafic lava flows that we informally refer to as the Lake City basalts (unit Tlb, Fig. 3). The package of flows ranges in thickness from a few tens of meters near Cedar Pass to more than 2 km at Buck Mountain at the north end of the study.
Figure 3. Geologic map of the Warner Range, simplified from Egger and Miller (2011) (legend on following page).
area (Fig. 4). The underlying sedimentary rocks are not exposed at this latitude, but they were penetrated by a geothermal well (LCSH-5, Figs. 3 and 4), putting an upper limit of ~2700 m on the thickness of the Lake City basalts. Individual flows are a few meters thick, dark gray to black when fresh, but often reddish weathering. Flow tops are generally vesicular and flow margins are strongly brecciated, with more massive interiors where the entire thickness of the flow is exposed. Flows vary from aphyric to moderately porphyritic with phenocrysts of plagioclase, olivine, and pyroxene; the olivine is commonly altered to iddingsite and the plagioclase is usually partly altered to clay or white mica. Some flows contain abundant (>50%) large (>1 cm) plagioclase phenocrysts; these have a sieve-like texture in thin section with abundant inclusions of melt and/or opaque oxides. Hornblende is conspicuously absent compared to the slightly younger rocks described in the next section.

Two samples of the base of the Lake City basalts collected ~3.5 km apart yielded 40Ar/39Ar ages of 27.49 ± 0.33 Ma from plagioclase (sample AE05-WR03) and 27.83 ± 0.21 Ma from groundmass (sample JC08-WR405) (Table 1; Fig. 5). Plagioclase from a sample (07-C-6) midway up the sequence of flows yielded a 40Ar/39Ar age of 27.14 ± 0.08 Ma (Table 1; Fig. 5). Plagioclase from a sample of the uppermost flow, just beneath a Late Miocene rhyolite dome at the north end of the study area (WR07-AE40), yielded a disturbed 39Ar release spectrum from which we calculate an isochron age of 25.70 ± 0.94 Ma (Table 1; Fig. 5). This age is consistent with its position at the top of the section, but east of Bald Mountain, the Lake City basalts are overlain by an ash-flow tuff with a sanidine 40Ar/39Ar age of 26.64 ± 0.08 Ma (Fig. 4), which we consider a more precise upper age limit for the Lake City basalts. From these dates we conclude that the Lake City basalts were erupted during a relatively brief period (<1–1.5 m.y.) in the Late Oligocene, from ca. 28 to 26.5 Ma. Where exposed, the basal contact of the Lake City basalt sequence is subparallel to bedding in the underlying volcaniclastic rocks, not cut down into them, indicating that the basalt flows primarily built an edifice on top of the older rocks, rather than being deposited in paleotopography. The map pattern of the basalt flows (Fig. 3) indicates that this edifice was ~2 km high with a southern slope dipping 7°–8°, suggesting a moderate-sized shield volcano.

Cedar Pass Complex (26.6–24.5 Ma)

The name Cedar Pass complex is our informal term for andesitic breccias, lava flows, and hypabyssal intrusive rocks exposed in the vicinity of Cedar Pass in the central Warner Range (Figs. 3, 6A, and 6B). They are exposed as far north as Franklin Creek, where they overlie the Lake City basalts, and grade southward into Duffield and McKee’s (1986) “composite volcanic unit” in the South Warner Wilderness (unit Tcu in Fig. 3), where they overlie older Oligocene volcaniclastic rocks (Fig. 3).

Most of the Cedar Pass complex consists of andesitic breccias composed of angular lava blocks ranging from a few centimeters to >2 m across in a matrix of more finely broken rock fragments. Bedding is occasionally well developed (e.g., deposits on far end of ridge in Fig. 6A), but breccias are more typically massive and unsorted. Lava blocks range from dark, phenocryst-poor lava to lighter gray-green and abundantly plagioclase-phyric and hornblende-
Figure 4. Oligocene and younger stratigraphy of the Warner Range.
phyric lava, and multiple types of lava blocks are often found within the same outcrop. Some blocks are glassy and some have well-preserved radial fractures (Fig. 6B), indicating that they were hot at the time of emplacement and have been minimally disturbed since. We interpret these deposits as block-and-ash flows that range from minimally reworked (glassy lava blocks with well-preserved radial fracturing) to significantly reworked (e.g., the well-bedded deposits with well-preserved radial fracturing) to significantly reworked (e.g., the well-bedded deposits shown in Fig. 6A).

Porphyritic lava flows containing phenocrysts of plagioclase, hornblende, and lesser pyroxene are interbedded with the andesitic breccias at the summit of Cedar Mountain and on the north edge of the South Warner Wilderness (Fig. 3). Dark, gray-weathering, phenocryst-poor lava flows with a medium-grained groundmass of plagioclase and pyroxene (but no hornblende) overlie andesitic breccias on Payne Peak. At least one small (<1 km²) porphyritic andesite body intrudes the breccias at Cedar Pass (Fig. 3). This intrusion contains abundant phenocrysts of hornblende and plagioclase in a fine-grained matrix.

Duffield and McKee (1986) interpreted one vent area in the Cedar Pass complex centered on Dry Creek Basin (Fig. 3), on the basis of radial dips in the surrounding deposits. This is consistent with our new mapping (Fig. 3; Egger and Miller, 2011), and we suggest that another vent may exist between Cedar Pass and Cedar Mountain, where reworked block-and-ash flow deposits contain lava blocks as much as 2 m across and are intruded by a porphyritic andesite body that is exposed in roadcuts at Cedar Pass (Fig. 3).

East of Bald Mountain (Fig. 3), the Cedar Pass complex overlies a densely weldedphyric ash-flow tuff with a sanidine ⁴⁰Ar/³⁹Ar age of 26.64 ± 0.08 Ma (sample H08–57, Fig. 7) and is capped by a basalt flow with a ⁴⁰Ar/³⁹Ar age of 24.47 ± 0.34 Ma (sample 07–C10, Fig. 5). A porphyritic, plagioclase- and hornblende-bearing lava flow on the summit of Cedar Mountain (Fig. 3) yielded a ⁴⁰Ar/³⁹Ar age of 25.73 ± 0.05 Ma (sample 07–C19, Fig. 5). Duffield and McKee (1986) reported a K-Ar age of 26.6 Ma from one sample to the southwest. Carmichael et al. (2006) reported a K-Ar age of 28.7 ± 2.2 Ma from the same outcrop, and a K-Ar age of 26.6 Ma from another sample to the southwest. Carmichael et al. (2006) reported a ⁴⁰Ar/³⁹Ar age of 26.6 Ma from another sample to the southwest.
Hays Volcano (ca. 24 Ma)

On the east side of Surprise Valley in the southern Hays Canyon Range, Oligocene ash-flow tuffs are overlain by the remnants of an Oligocene spatter volcano called the Hays volcano by Carmichael et al. (2006) (Fig. 3). These deposits now cover an area ~12 × 12 km and are as thick as 550 m, although their base is not exposed and they may be more extensive in the subsurface. Individual outcrops consist of layers of spatter, agglutinated lavas, and scoria blocks as thick as several meters (typically ~1 m). These layers dip as much as 20° radially away from an inferred central vent that is highly altered and more deeply eroded than the surrounding hills. Carmichael et al. (2006) reported two 40Ar/39Ar dates of ca. 24 Ma from the Hays volcano, similar in age to the youngest dated lava flow in the Cedar Pass complex.

Ash-Flow Tuffs (26.6–25.8 Ma)

Several densely welded ash-flow tuffs are interbedded with the Cedar Pass complex and locally separate it from the Lake City basalts.
Ash-flow tuffs also make up the bulk of the Oligocene section south of Parker Creek (Fig. 3), and in the Hays Canyon Range (Fig. 3). East of Bald Mountain (Fig. 3), the Lake City basalts and Cedar Pass complex are separated by a densely welded, pumice-rich, crystal- and lithic-poor tuff containing sparse phenocrysts of biotite and anorthoclase, which yielded a \(^{40}\text{Ar}/^{39}\text{Ar}\) anorthoclase age of 26.64 ± 0.08 Ma (sample H08–57, Fig. 7). A few meters upsection, a single outcrop of a moderately crystal- and pumice-rich, reddish-weathering tuff with phenocrysts of anorthoclase and biotite yielded a \(^{40}\text{Ar}/^{39}\text{Ar}\) anorthoclase age of 26.35 ± 0.11 Ma (sample JC07-WR303, Table 1). On the ridge north of Squaw Peak (Fig. 3), the Oligocene volcanic section includes several ash-flow tuffs in addition to distal material from the Cedar Pass complex (Fig. 4). The base of the section here consists of ~70 m of poorly welded, lithic- and pumice-poor tuff with abundant biotite and lesser phenocrysts of sanidine, plagioclase, and quartz. This tuff yielded a \(^{40}\text{Ar}/^{39}\text{Ar}\) sanidine age of 26.53 ± 0.06 Ma (sample JC08-WR11, Fig. 7). Higher in the section there is a thick (~160 m) lithic- and sanidine- and biotite-bearing tuffaceous sandstone and siltstone containing small (≤2 mm) biotite and feldspar crystals; some feldspars are distinctly iridescent. Plagioclase from a reworked tuff in the upper part of this deposit yielded a \(^{40}\text{Ar}/^{39}\text{Ar}\) age of 19.22 ± 0.27 Ma (Table 1; Fig. 4). Duffield and McKee (1986) reported a poorly defined biotite \(^{40}\text{Ar}/^{39}\text{Ar}\) age of 25.77 ± 0.06 Ma (sample JC08–WR402, Table 1; Fig. 4). The thickness of the tuffs (to 160 m in the Warner Range; >300 m in the Hays Canyon Range) and the limited lateral extent of individual tuff beds are consistent with deposition in paleovalleys or other low spots in the Oligocene paleotopography; ash-flow tuffs elsewhere in the Basin and Range have been documented to have traveled hundreds of kilometers down such channels (e.g., Henry, 2008, Henry and Faulds, 2010). We cannot correlate these tuffs with any known calderas, but caldera-forming eruptions were ongoing during the Oligocene in northwestern and central Nevada (e.g., Noble et al., 1970; Best et al., 1989; Christiansen and Yeats, 1992), and these areas are plausible sources.

### Miocene Volcanic Rocks

#### Tuffs and Tuffaceous Sedimentary Rocks (ca. 19–13 Ma)

South of Squaw Peak, Miocene mafic lava flows are separated from the Cedar Pass complex by ~100 m of poorly exposed tuffaceous sandstone and siltstone containing small (≤2 mm) biotite and feldspar crystals; some feldspars are distinctly iridescent. Plagioclase from a reworked tuff in the upper part of this deposit yielded a \(^{40}\text{Ar}/^{39}\text{Ar}\) age of 19.22 ± 0.27 Ma (Table 1; Fig. 4). Duffield and McKee (1986) reported a K-Ar sanidine age of 17.3 Ma from a thin (<5 m) densely welded ash-flow tuff at the top of this deposit ~3 km to the north. Biotite- and sanidine-bearing tuffaceous sedimentary rocks are also present south of Parker Creek, between andesitic breccias of the Cedar Pass complex. They were most likely generated by large caldera-forming eruptions and may have traveled significant distances from their sources, as first noted by Duffield and McKee (1986). The thickness of the tuffs (to 160 m in the Warner Range; >300 m in the Hays Canyon Range) and the limited lateral extent of individual tuff beds are consistent with deposition in paleovalleys or other low spots in the Oligocene paleotopography; ash-flow tuffs elsewhere in the Basin and Range have been documented to have traveled hundreds of kilometers down such channels (e.g., Henry, 2008, Henry and Faulds, 2010). We cannot correlate these tuffs with any known calderas, but caldera-forming eruptions were ongoing during the Oligocene in northwestern and central Nevada (e.g., Noble et al., 1970; Best et al., 1989; Christiansen and Yeats, 1992), and these areas are plausible sources.
On the west side of the Warner Range, Carmichael et al. (2006, p. 1198–1199) dated several “rhyolitic ash-flow deposits of various ages”; they did not describe these deposits in detail, but mentioned “notable biotite” (p. 1207) and reported ages of 14.0 and 13.4 Ma (Table 1). These tuffaceous deposits are overlain by Miocene basalt flows at the south end of our Figure 3, and by Pliocene sedimentary rocks and lava flows in the vicinity of Alturas (Carmichael et al., 2006). Because of their proximity, similar position in the section, and similar lithology, we tentatively suggest that these tuffs belong to the same geologic unit as the ones described in the previous paragraph, which would then constitute a package of primarily biotite-bearing, variably reworked tuffs derived from distal sources that accumulated in the Warner Range area from 19 to 13 Ma and possibly more recently.

**Miocene Lavas of the Warner and Hays Canyon Ranges (16–14 Ma)**

Middle Miocene basaltic to andesitic lava flows are exposed in the southern Warner Range as far north as Parker Creek (Fig. 3), and thicken southward to ~500 m at Warren Peak (Fig. 6C) and to >1000 m at Emerson Peak (Figs. 4 and 6D). Duffield and McKee (1986) divided these flows into two map units (Tvm and Tvb) separated by a layer of tuff, but they are indistinguishable from each other in outcrop and are part of the same volcanic edifice or edifices, so we show them as a single unit (Tmb) in Figure 3. Miocene lava flows of similar appearance and composition cap the Hays Canyon Range on the east side of Surprise Valley (Fig. 3), where they unconformably overlie the eroded remnants of Hays volcano (Carmichael et al., 2006). Individual Miocene lava flows are generally 2–5 m
thick, occasionally as thick as 20 m, with scoria­ceous tops and massive interiors. The interiors of some flows exhibit such coarse, gabbro-like textures in hand specimen and thin section that they may actually be sills injected into the stack of flows. Flows often have prominent columnar joints (Fig. 6C), but are locally glassy and flow banded. Phenocrysts are variably abundant and include plagioclase, pyroxene, and olivine.

Miocene lava flows in the Warner Range are locally separated by a section of volcanioclastic rocks as much as 100 m thick (unit Tmt, Fig. 4; equivalent to Duffield and McKee’s Tvt unit (1986). South of Parker Creek, these volcanioclastic rocks are sporadically exposed in roadcuts on the densely forested western slope of the range. They consist of graded, cross-bedded coarse sandstones with abundant plagioclase crystal fragments and dark lava chips, inter­bedded with massive inversely-deposited fragments of angular mafic lava blocks supported by a brown sandy matrix. Below the summit of Warren Peak (Fig. 3), the volcanioclastic section consists of a layer of tuff with a fine-grained gray ash matrix supporting a mixture of small (<1 cm) white pumice lapilli and angular black lava fragments.

Based on published K-Ar and 40Ar/39Ar dates (Table 1), Miocene mafic lava flows in the Warner and Hays Canyon Ranges were erupted during a relatively narrow interval between ca. 16 and 14 Ma; most dates are between 15 and 16 Ma (Duffield and McKee, 1986; Carmichael et al., 2006) (Table 1). Duffield and Weldin (1976) suggested the presence of several vents for the Miocene lava flows at the south end of our Figure 3. Our mapping supports this suggestion, and one of their inferred vents, at Emerson Peak, is pictured in Figure 6D: here, lava flows on Emerson Peak dip west, while flows in the foreground dip to the east, and a prominent mafic dike strikes toward the inferred eruptive center in the low topography below the high peaks. This relationship, and the overall north­ward-thinning of the Middle Miocene lava flows (Fig. 4), is consistent with them being erupted from one or more overlapping shield volcanoes in the southern Warner Range, similar in size and shape to the intact volcanoes of slightly younger age described in the next section.

Miocene Volcanoes South of the Warner Range (14 to as Young as 8 Ma)

Several dozen Miocene volcanoes are exposed south of the Warner Range, extending from near Likely south to Honey Lake Valley (Fig. 2) (Grose, 2000). They average ~8–10 km in diameter and are 500–600 m above the surrounding landscape. Shinn Mountain (Fig. 6E) is typical of one of these volcanic edifices, which exhibit radial drainage patterns, flank slopes of 6°–7°, and summit vent complexes with well-preserved craters. Their flanks are composed of homogeneous basaltic andesite lava flows, with scoria and spatter exposed in the summit vents and/or craters. Extant K-Ar dates from the volcanoes shown in Figure 2 range from 14.5 to 8 Ma; most are in the 10–12 Ma range (Grose, 2000; Grose et al., 1991, 1992). They are thus slightly younger than the Miocene volcanoes in the southern Warner Range, but can be considered part of the same volcanic episode with respect to their composition and eruptive style.

Late Miocene to Pliocene Tholeiitic Basalts (8–3 Ma)

Late Miocene to Pliocene high-alumina, low-potassium olivine tholeiite lava flows crop out extensively on both sides of the Warner Range (Figs. 2 and 3); the lava was referred to as high­alumina olivine tholeiite by Hart et al. (1984), and as low-potassium olivine tholeiite by Carmichael et al. (2006). In outcrop and hand speci­men the flows are usually easy to distinguish from the older lavas by their light to medium gray, nonporphyritic, diktotypic texture. In the area of Figure 3, they range from ca. 8 to 3 Ma, but most are 4–3 Ma (Carmichael et al., 2006). West of the Warner Range, Pliocene lavas (the Devil’s Garden basalt of McKee et al., 1983) are flat lying and overly Middle to Late Miocene lava flows and tuffaceous sedimentary rocks. They crop out partway up the western slope of the range to an altitude of almost 1900 m, ~450 m above the surrounding plain to the west (Fig. 3). In northern Surprise Valley, Pliocene lava flows (the Vya Group of Carmichael et al., 2006) are cut and tilted ~15° west by many small-offset normal faults (Fig. 3).

GEOCHEMISTRY

We obtained major, trace, and rare earth element (REE) data for samples of Oligocene and younger volcanic rocks (Supplemental Table 1), and Sr and Nd isotopic analyses of a few representative samples (Table A3). Additional analyses of Warner Range samples from Carmichael et al. (2006) and V.E. Camp and M.E. Ross (2007, personal commun.) are compiled in Supplemental Table 1 (see footnote 1) and included in the following discussion (symbols are the same in Figs. 8–13). Oligocene samples are plotted in three groups: Lake City basalt flows, Cedar Pass complex (analyses of lava flows, intrusive rocks, and lava blocks), and analyses of the Hays volcano southeast of the Warner Range (from Carmichael et al., 2006). Miocene samples are divided into two groups: Miocene lavas in the Warner and Hays Canyon Ranges, which range from ca. 16 to 14 Ma, and slightly younger Miocene lava flows, including the basaltic andesite shield volcanoes south of the Warner Range, which range from ca. 14 to 8 Ma. Pliocene low-K olivine tholeiites are plotted separately.

Oligocene and Miocene Warner Range lavas range in composition from basalt to andesite, the Miocene lavas on average being slightly more alkalic (Fig. 8). (Duffield and McKee also mapped rhyolites in the Warner Range (Fig. 3) that we did not analyze in this study.) Oligocene volcanism began with the more mafic Lake City basalts and progressed to the more andesitic Cedar Pass complex. The youngest Oligocene lavas, i.e., the ca. 24 Ma Hays volcano (Fig. 2), are similar to the andesite parts of the Cedar Pass complex but slightly more alkalic. Miocene lavas in the Warner Range consist primarily of basalt, whereas slightly younger lavas to the east, west, and particularly south (Fig. 2) consist mostly of basaltic andesite and andesite (Fig. 8). Compared to rocks in the western Cascades, Warner Range lavas are overall relatively enriched in incompatible elements (K, Ba, Sr, La), the Miocene lavas being somewhat more enriched than the Oligocene (Fig. 9). In this respect the Warner Range lavas are more similar to rocks of the ancestral Cascades in California and Nevada, possibly because both suites were built on thicker, less mafic crust than the western Cascades.

Both Oligocene and Miocene Warner lavas are relatively enriched in the light REEs; the Miocene lavas are somewhat more enriched than the Oligocene ones (Fig. 10). They are distinctly different from the Pliocene low-K, high-Al basalts, which have much flatter REE patterns (Fig. 10). Oligocene and Miocene andesites and basaltic andesites have the prominent depletions in high-field-strength elements (Nb, Ta, Ti, Zr) (Fig. 10) characteristic of magmas generated in a subduction setting. Lavas from the Warner Range and nearby areas are relatively nonradio­genetic, ranging from 87Sr/86Sr = 0.7035 to 0.7045 (Fig. 11), consistent with their position well to the west of inferred North American continental crust. The Miocene lavas are distinctly more radiogenic than the Oligocene lavas, however, with 87Sr/86Sr > 0.7039, whereas the Oligocene lavas plot in a more restricted range at 87Sr/86Sr < 0.7039 (Fig. 11). The Pliocene low-K basalts have distinctly lower 87Sr/86Sr than either the Miocene or Oligocene lavas (Fig. 11). With the exception of the younger Miocene lavas,
Warner Range lavas are isotopically more similar to rocks of the Washington and Oregon Cascades than they are to the ancestral Cascades in California and Nevada.

DISCUSSION

Oligocene Volcanism

Andesitic volcanism in northeastern California and southern Oregon was ongoing by the Oligocene, as indicated by the thick sequences of lahars in the lower part of the Warner Range section. Local volcanic activity began with eruption of the Lake City basalts, which built a broad basaltic shield volcano between ca. 28 and 26.5 Ma. Subsequent eruptions in the Warner Range were more silicic and built one or more overlapping volcanic edifices mostly composed of andesitic block-and-ash flows and lesser lava flows (Fig. 14B). The youngest lava flow in the Cedar Pass complex is ca. 24 Ma, although peak eruptive activity may have ceased earlier. The full duration of magmatism at Hays Volcano on the east side of Surprise Valley is less clear, but by 24 Ma it had built a spatter cone (or several overlapping cones) at least 550 m high (Fig. 3). Local Oligocene volcanism thus took place between 28 and 24 Ma, although the life span of any individual eruptive center was probably much less. Volcanism ceased in the Warner Range between ca. 24 and 16 Ma, but was probably active during this time elsewhere in the region (see following discussion).

Major, trace element, and isotopic data are most consistent with a subduction source for the Oligocene volcanic rocks, and we interpret them to be part of the arc formed by subduction of the Farallon plate beneath northern California, which was ongoing in the Oligocene (e.g., Atwater and Stock, 1998). Relative to the Miocene lavas, Oligocene Warner Range lavas have lower \( {\text{Sr}}^{87}/\text{Sr}^{86} \) (Fig. 12A) and Ce/Pb (Fig. 12B) ratios, higher primitive-mantle-normalized Sr/P (Fig. 12C), and are more depleted in Nb (Fig. 12D). Since there is little correlation between Sr/P ratio and SiO2 content (Fig. 12E), we interpret these patterns to indicate that the Oligocene lavas were derived from mantle with a relatively large (compared to the Miocene) fluid flux from the subducting slab. These fluids had relatively low \( {\text{Sr}}^{87}/\text{Sr}^{86} \) ratios, similar to \( {\text{Sr}}^{87}/\text{Sr}^{86} \) ratios in modern southern Cascade arc mafic lavas (Borg et al., 1997). The overall low \( {\text{Sr}}^{87}/\text{Sr}^{86} \) ratios from the Oligocene samples are consistent with variable fluid flux from subducted basaltic crust and minor (if any) contribution from subducted sediments or continental crust.

Miocene Volcanism

No volcanic rocks were erupted in the Warner Range between ca. 24 and 16 Ma, although minor distally sourced tuffs accumulated in at least one locality. However, volcanic rocks between 24 and 16 Ma are exposed in northwestern Nevada and southeastern Oregon, suggesting that active volcanism may have shifted to nearby areas in the Early Miocene rather than ceasing in the region altogether. The nearest examples to the Warner Range are volumetrically minor mafic lava flows erupted between 30 and 21 Ma in the Pine Forest and Black Rock Ranges in Nevada, ~100–120 km to the east (Colgan et al., 2006; Lerch et al., 2007), and a 23–21 Ma complex of rhyolite domes and andesitic volcanoes in Oregon, ~100 km to the north (Scarberry et al., 2010). Abundant, dominantly andesitic rocks with lesser basaltic andesite and rhyolite also erupted between 24 and 18 Ma in Nevada north of Reno (Garside et al., 2000; 2003). Volcanic activity in the Warner Range resumed ca. 16 Ma and formed one or more overlapping basalt to basaltic andesite shield volcanoes. Subsequent eruptions formed more than 12 basaltic andesite shield volcanoes to the south and west of the Warner Range (Fig. 2). Published K-Ar dates from these volcanoes are somewhat younger than the lavas in the Warner Range (mostly 12–11 Ma); this pattern may indicate an age progression in volcanic eruptions from the Warner Range to the south and/or west, but more systematic and precise dating is necessary to confirm this.

Similar to the Oligocene lavas, major, trace element, and isotopic data are most consistent
Pliocene high-Al olivine tholeiite
• Miocene lava flows; northeastern California (14–7 Ma)
• Miocene lava flows; Warner Range (16–14 Ma)
• Oligocene Hays volcano
• Oligocene Cedar Pass complex
• Oligocene Lake City basalts
• Oregon and Washington Cascades 45–4 Ma (n = 2130)
• Inferred California and Nevada arc 35–4 Ma (n = 558)

Figure 9. Variation diagrams for Warner Range volcanic rocks. Oregon and Washington Cascades from du Bray et al. (2006); inferred California and Nevada arc from du Bray et al. (2009). Warner Range volcanic rocks are consistently more similar to the inferred California and Nevada arc than they are to the Oregon and Washington Cascades.
with a subduction source for the Miocene volcanic rocks, and we also interpret them to be part of the arc formed by subduction of the Farallon plate beneath northern California, which was ongoing in the Miocene (e.g., Atwater and Stock, 1998). Neither the Oligocene nor the Miocene lavas show obvious evidence of crustal contamination based on their similar \(^{87}\text{Sr}/^{86}\text{Sr}\) over the entire range of \(\text{SiO}_2\) (Fig. 12A). This does not rule out some component of crustal contamination, however, since the isotopic signature of the accreted crust that underlies the region would be minor and hard to trace. The Miocene lavas have more radiogenic \(^{87}\text{Sr}/^{86}\text{Sr}\) (Fig. 12A), smaller negative Nb anomalies (Fig. 12D; lower Sr/P, Fig. 12C), and slightly higher Ce/Pb (Fig. 12B) than the Oligocene lavas. Sr isotope ratios are not negatively correlated with Ce/Pb (Fig. 12B), and Ce/Pb does not correlate at all with \(\text{SiO}_2\) content (Fig. 12F), so we attribute these patterns to the Miocene lavas having a different mantle source than the Oligocene suite. The Miocene source may include a smaller fluid component derived from the subducting slab, but one with more radiogenic Sr. The more radiogenic Sr is accompanied by less radiogenic Nd, so this component cannot simply be seawater. The shift to higher Ce/Pb from the Oligocene to the Miocene mantle sources is also inconsistent with a larger sedimentary component in the Miocene source.

Miocene lavas in the Warner Range and nearby to the south have a distinctly different composition and eruptive style from volcanic rocks of overlapping age erupted in the Basin and Range Province to the east, notably the Steens Basalt, with which they have sometimes been included on regional maps (e.g., Camp and Ross, 2004; Brueseke et al., 2007; Camp and Hanan, 2008). Miocene lavas in northeastern California erupted to form a chain of shield volcanoes, while the Steens Basalt...
Northeastern California arc

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was erupted from roughly north- to northeast-trending dikes in eastern Oregon (e.g., Camp and Ross, 2004). Middle Miocene lavas in the Warner Range are younger (mostly 16–15 Ma, Table 1) than the ca. 16.6 Ma (Jarboe et al., 2008) Steens Basalt, although they do overlap in time with the most voluminous (Grande Ronde) phase of the Columbia River Basalts (e.g., Hooper et al., 2002). Origin of the Steens Basalt and related Columbia River Basalts is much debated, particularly their relationship to a deep mantle plume, but they are clearly not ordinary subduction-related lavas (e.g., Chesley and Ruiz, 1998; Hooper et al., 2002; Camp and Ross, 2004; Camp and Hanan, 2008). Compared to the Steens Basalt, Middle Miocene mafic lava flows in the Warner Range are distinctly depleted in Nb, Ta, and Ti, for a given MgO content (Fig. 13).

Figure 11. Plot of Sr and Nd isotopic data for Warner Range volcanic rocks. Pliocene and younger Oregon and Washington Cascades from GEOROC (2007) database; inferred California and Nevada arc from du Bray et al. (2009). Rocks of the inferred California and Nevada arc are overall distinctly more radiogenic than the Oregon and Washington Cascades; Warner Range volcanic rocks plot in the middle and overlap the more radiogenic end of the Oregon and Washington Cascades. NE CA—northeastern California.

was erupted from roughly north- to northeast-trending dikes in eastern Oregon (e.g., Camp and Ross, 2004). Middle Miocene lavas in the Warner Range are younger (mostly 16–15 Ma, Table 1) than the ca. 16.6 Ma (Jarboe et al., 2008) Steens Basalt, although they do overlap in time with the most voluminous (Grande Ronde) phase of the Columbia River Basalts (e.g., Hooper et al., 2002). Origin of the Steens Basalt and related Columbia River Basalts is much debated, particularly their relationship to a deep mantle plume, but they are clearly not ordinary subduction-related lavas (e.g., Chesley and Ruiz, 1998; Hooper et al., 2002; Camp and Ross, 2004; Camp and Hanan, 2008). Compared to the Steens Basalt, Middle Miocene mafic lava flows in the Warner Range are distinctly depleted in Nb, Ta, and Ti, for a given MgO content (Fig. 13).

Implications for Evolution of the Cascade Arc

We interpret Oligocene (28–24 Ma) and Middle Miocene (16–8 Ma) volcanism in the Warner Range and nearby to be a direct consequence of subduction of the Farallon plate beneath northern California. In map view, these volcanic centers are continuous with the northern extent of the ancestral Cascades (Fig. 1), but are more than 100 km east of the Eocene to Miocene western Cascades, which extend as far south as Mount Shasta (Figs. 2 and 14). Does this offset in arc segments reflect the geometry of the subducting slab, post-Eocene tectonic deformation, or a combination of the two? If it is related to the subducting slab, why are the arc segments apparently offset in such a manner?

All or part of the apparent offset of the two arc segments could be the result of significant Eocene or younger westward translation of the Klamath Mountains relative to the northern Sierra Nevada (Fig. 14), accommodated by east-west extension in the region shown in Figure 2. Geologic studies of northern California indicate that such translation took place, but consistently conclude that it happened during the Cretaceous (Jones and Irwin, 1971; Constienius et al., 2000; Wyld et al., 2006). Although some reconstructions have suggested that such motion was Eocene or younger (e.g., Dickinson, 2002;
Humphreys, 2009), geologic evidence for this is absent. Late Miocene and Pliocene rocks in northeastern California are locally cut by many small normal faults, but these have only minor offset and indicate very little post-Miocene extension. Older rocks that would record pre-Miocene extension are mostly covered between the Warner Range and Klamath Mountains (Fig. 2), but where they are exposed they are conformable with overlying units. In the Warner Range, no angular unconformity separates rocks ranging from older than 34 Ma to 14 Ma, consistent with no significant extension at the time (Duffield and McKee, 1986; Colgan et al., 2008; Egger et al., 2009; Egger and Miller, 2011; this study). Helley and Harwood (1985) mapped no angular unconformity between the Pliocene Tuscan Formation and the Eocene Montgomery Creek Formation on the east edge of the Klamath Mountains (Fig. 2). No evidence was found (Colgan et al., 2006) for significant extension in northwestern Nevada prior to ca. 12 Ma, and Scarberry et al. (2010) documented only minor extension (<5% strain) since the Early Miocene north of the Warner Range in southern Oregon. The western Cascades (Fig. 14) have undergone significant clockwise rotation since the Eocene, but this motion was accommodated relative to the unrotated Sierra Nevada by a pivot in northern California, not by substantial westward motion of the Klamath block relative to the northern Sierra Nevada (Wells et al., 1998; Wells and Simpson, 2001). Overall, because there is geologic evidence for Cretaceous separation of the Klamath Mountains and northern Sierra Nevada, and no geologic evidence for such motion in the Cenozoic, we conclude that the position of the arc segments in Figure 14 is a function of where they formed.

North of Mount Shasta, Oligocene volcanic rocks of the western Cascades are west of the Quaternary arc (Figs. 1 and 2). South of Mount Shasta, Oligocene and Miocene rocks of any kind are absent; Pliocene volcaniclastic rocks of the Tuscan Formation (ca. 3.4–1.8 Ma) are deposited directly on pre-Tertiary basement and Eocene sedimentary rocks of the Montgomery Creek Formation (Helley and Harwood, 1985; Irwin, 1997). The westernmost volcanic arc rocks of possible Late Miocene age are undated rocks along a ridge south of Medicine Lake (Fig. 2), and the westernmost dated Miocene volcanic centers are between Alturas and Susanville (Fig. 2). Fundamentally arc-related volcanic activity was ongoing in the Warner Range in the Oligocene, and from 16 to ca. 8 Ma in the Warner Range and the belt of Miocene volcanoes extending another 100 km south (Fig. 2). With a few minor exceptions, post-Miocene volcanism in the area of these older volcanic rocks has been limited to HAOT lavas with a distinctly different mantle source than the subducting slab (e.g., Guffanti et al., 1990; Carmichael et al., 2006). No Quaternary vents have been mapped in this area (east of ~120°30′W, Fig. 2), and the axis of the Quaternary Cascade arc is 100–150 km to the west. The eastern edge of the arc must therefore have migrated west between the Miocene and the present (e.g., Guffanti and Weaver, 1988; Guffanti et al., 1990; Muffler et al., 2009). The absence of pre-Pliocene volcanic rocks beneath the Pliocene–Quaternary arc south of Mount Shasta indicates that volcanism migrated west into this area after the
Late Miocene. Together, these patterns suggest that the segment of the Cascade arc south of Mount Shasta migrated ~100 km west between the Middle Miocene and the Quaternary, while the Cascade arc north of Mount Shasta remained in place or migrated east.

If the Oligocene and Miocene arc in northern California was active ~100 km east of its present location and not elsewhere, why and how did it move to where it is now? A simple explanation is that the subducting slab south of present-day Mount Shasta dipped more shallowly in the Oligocene and Miocene, then steepened either gradually or abruptly sometime after the Middle Miocene. North of Mount Shasta, however, the arc either remained in place (albeit more diffuse, e.g., Taylor, 1990) or actually migrated east since the Eocene (e.g., Verplanck and Duncan, 1987; Wells, 1990). We reconcile these observations by proposing that the Eocene to Middle Miocene subducting slab along the western North American plate margin was broken by a northeast-trending tear located south of Mount Shasta and north of the Warner Range (red dotted line in Fig. 14). Humphreys (2009) proposed that Early Eocene accretion of the Siletzia block in central Oregon (Fig. 14) caused a new subduction zone to form outboard of the accreted block, with a tear along its southern margin separating it from the still gently dipping slab to the south (Fig. 14). Based on the pattern of offset arc segments described in this study, we suggest a more southern location for the slab tear (Fig. 14) that is more consistent with a major change in slab dip across it; however, we agree with the mechanism proposed by Humphreys (2009) for its formation. Once formed, the northern segment of the slab subducted at the same angle, or perhaps shallowed slightly to the present, while the southern segment initially remained attached to the gently dipping Laramide slab beneath the western United States interior. The transition from Basin and Range ignimbrite volcanism to the ancestral Cascades remains poorly understood, but recent work indicates that normal subduction was established beneath eastern California and western Nevada by the Miocene, if not earlier (Putirka and Busby, 2007; Cousens et al., 2008; Busby et al., 2008a, 2008b; Busby and Putirka, 2009; Hagan et al., 2009; Vikre and Henry, 2010). The southern segment of the slab continued subducting at a shallower dip than the northern portion until the Middle Miocene, before steepening and moving the arc westward to its present location by the late Pliocene.

CONCLUSIONS

Arc volcanism in the Warner Range began in the Oligocene, and volcanic centers were active locally from 28 to 24 Ma and from 16 to 14 Ma. Oligocene and Middle Miocene volcanic rocks both have clear subduction sources, but involved different degrees of fluid in the melting. From 14 Ma, arc volcanism migrated south and west, and Pliocene eruptions around the Warner Range were limited to HAOT lavas with a different mantle source than the subducting slab. The Oligocene and Middle Miocene arc south of Mount Shasta was continuous with the inferred ancestral Cascades in western Nevada and eastern California, but was 100 km east of both the modern arc and the Eocene to Pliocene western Cascade arc north of Mount Shasta. We conclude that separation of these arc segments is the result of a northeast-trending tear in the subducting slab that formed in the Eocene, at which time the western Cascades began to form. The ancestral Cascades began to form by the Oligocene, as the southern segment of the slab progressively steepened and the arc migrated west, reaching its present position by the late Pliocene.
APPENDIX 1. \(^{40}\text{Ar}/^{39}\text{Ar}\) ANALYTICAL METHODS

Samples were coarsely crushed and sieved to ~20–40 mesh, then run through a magnetic separa-
tor to concentrate sanidine, quartz, and plagioclase. These were leached with 5% HF for ~1 h to remove any matrix. Sanidine was then handpicked under a

binocular microscope. Samples were irradiated at Texas A&M University for 7 h and analyzed at the New Mexico Geochronology Research Laboratory (New Mexico Institute of Mining and Technology), using procedures described in McIntosh et al. (2003). Neutron flux was monitored using interlaboratory standard Fish Canyon Tuff sanidine FC-1 with an assigned age of 28.02 Ma (Renne et al., 1998). Individual sanidine grains were fused using a CO\(_2\) laser operating at 10 W for 5 s. Extracted gases were purified with SAES GP-50 getters. Argon was analyzed with a Mass Analyzer Products model 215–50 mass spectrometer operated in static mode. Variance-weighted mean ages of the 9–23 grains reported in Table A1 were calculated by the method of San-
son and Alexander (1987), using decay constants of

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TABLE A1. \(^{40}\text{Ar}/^{39}\text{Ar}\) ANALYTICAL DATA (NEW MEXICO TECH UNIVERSITY)

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Note: Data are from the New Mexico Tech University (New Mexico Institute of Mining and Technology, Socorro, New Mexico). J is the irradiation parameter (e.g., McDougall and Harrison, 1999) determined by analyses of Fish Canyon Tuff sanidine with an assumed age of 28.02 Ma (Renne et al., 1998).
APPENDIX 2. \(^{40}\text{Ar}/^{39}\text{Ar}\) ANALYTICAL METHODS

Samples collected for \(^{40}\text{Ar}/^{39}\text{Ar}\) geochronology were crushed and sieved to sizes appropriate for mineral separation of each sample (methods of U.S. Geological Survey, Menlo Park, California). Samples were irradiated in the U.S. Geological Survey TRIGA Reactor Facility (Denver, Colorado), with irradiation times between 10 and 16 h. Plagioclase and groundmass \(^{40}\text{Ar}/^{39}\text{Ar}\) ages (Table 2A) were obtained by incremental-heating analysis; i.e., sequential extraction of the argon from the sample at progressively higher temperatures until the sample was fused. Incremental-heating analyses utilized a low-blank, tantalum and molybdenum, resistance-heated furnace, commonly releasing all of the Ar in 9–13 temperature-controlled heating increments. Ages from incremental-heating experiments are determined by an evaluation of the \(^{40}\text{Ar}/^{39}\text{Ar}\) age spectrum and isochron diagrams of the data. Many of the ages are defined by a \(^{40}\text{Ar}/^{39}\text{Ar}\) plateau, defined as the weighted mean age of contiguous gas fractions representing more than 50% of the \(^{39}\text{Ar}\) released for which no difference can be detected between the ages of any two fractions at the 95% level of confidence (Fleck et al., 1977). Where no plateau is defined by the ages, the \(^{40}\text{Ar}/^{39}\text{Ar}\) isochron age or the integrated age of all increments may be used. The sandine age for sample JC07-WR303 was obtained by laser-fusion analysis in which multiple grains were fused with a CO\(_2\) laser in a single heating step. The reported age for laser-fusion analyses represents the weighted mean of six or more replicate analyses, with the inverse variance of propagated.

### TABLE 2A. \(^{40}\text{Ar}/^{39}\text{Ar}\) ANALYTICAL DATA (USGS SURVEY LABORATORY, MENLO PARK, CA)

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<th>Sample</th>
<th>Lab no.</th>
<th>Step-heating experiment</th>
<th>J =</th>
<th>Age (Ma)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sample: 07-C-10 groundmass</td>
<td>Lab no. IRR266-36</td>
<td>Step-heating experiment</td>
<td>J = 0.002673929</td>
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</tr>
<tr>
<td>Sample: JC08-WR405 groundmass</td>
<td>Lab no. IRR266-48</td>
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<tr>
<td>Sample: AE05WR03 plagioclase</td>
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<td>Step-heating experiment</td>
<td>J = 0.000262108</td>
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<tr>
<td>Sample: 07-C-19 plagioclase</td>
<td>Lab no. IRR266-40</td>
<td>Step-heating experiment</td>
<td>J = 0.002561241</td>
<td></td>
</tr>
</tbody>
</table>

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APPENDIX 3. Sr/Nd ISOPOTE ANALYTICAL METHODS AND DATA

Samples were analyzed for Sr and Nd isotopic ratios at Carleton University (Ontario, Canada), using techniques described in Couzens (1996). Samples were run on a Thermo-Scientific Finnigan Triton T1 thermal ionization mass spectrometer running in static mode. Sr isotope ratios are normalized to \(^{86}\text{Sr}/^{88}\text{Sr} = 0.11940\). Two Sr standards are run at Carleton, NIST SRM987 (\(^{87}\text{Sr}/^{86}\text{Sr} = 0.710245 \pm 16\) and the Eimer and Amend (E&A) SrCO\(_3\) (\(^{87}\text{Sr}/^{86}\text{Sr} = 0.708022 \pm 10\)). Nd isotope ratios are normalized to \(^{143}\text{Nd}/^{144}\text{Nd} = 0.512273 \pm 0.00024\) and the Eimer and Amend (E&A) NdCO\(_3\) (\(^{143}\text{Nd}/^{144}\text{Nd} = 0.511848 \pm 8\)). All quoted uncertainties in Table A3 are 2\(\sigma\) standard deviations of the mean.

APPENDIX 4. GEOCHEMICAL ANALYTICAL METHODS

Major, trace, and rare earth element analyses reported in this study were obtained from two facilities, Washington State Geosynthetic Lab at Washington State University (WSU, Pullman, Washington), and the U.S. Geological Survey (USGS, Denver, Colorado), which at the time of this study contracted analyses out to SGS Laboratories (Toronto, Canada). Samples were analyzed at WSU for 27 major and trace elements by X-ray fluorescence (XRF), using procedures described by Johnson et al. (1999). Samples were analyzed at the USGS for 10 major elements by XRF, and for 55 major, trace, and rare earth elements by inductively coupled plasma–atomic emission spectrometry (ICP-AES) and ICP–mass spectrometry (MS).

Major element data reported in Supplemental Table 1 (see footnote 1) are XRF analyses from either WSU or the USGS. Trace element data reported are USGS ICP-MS and ICP-AES analyses if available, and WSU XRF analyses if not. Rare earth element data are USGS analyses. Specific data sources are noted in Supplemental Table 1 (see footnote 1) for each sample; for example “WSU XRF” means all data reported are XRF analyses from WSU, while “WSU XRF” means all data reported are XRF analyses from WSU and trace and/or rare earth element data are ICP-MS and ICP-AES data from the USGS contract lab.
ACKNOWLEDGMENTS

We thank Wendell Duffield for sharing his notes and field maps from the South Warner Wilderness, Trobridge “Trobe” Grove for sharing his extensive knowledge of northeastern California geology, Vic Camp for sharing unpublished geochemical analyses from the southern Warner Range, and Dick Benoit for access to core samples from the LCSH-5 geothermal well. Robert Christiansen and Keith Putrika reviewed an early version of the manuscript, and Cathy Busby and Anita Grunder provided insightful journal reviews.

REFERENCES CITED


Cathy Busby and Anita Grunder provided insightful review of the manuscript, and Cathy Busby and Anita Grunder provided insightful journal reviews.

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![Image of a geological map and table]

**TABLE A3. Sr AND Nd ISOTOPIC DATA**

<table>
<thead>
<tr>
<th>Sample</th>
<th>Age (Ma)</th>
<th>Rb (ppm)</th>
<th>Sr (ppm)</th>
<th>$^{87}$(Sr)/Sr (present)</th>
<th>$^{143}$(Nd)/Nd (±2σ)</th>
<th>$^{87}$(Sr)/Sr (initial)</th>
<th>$^{143}$(Nd)/Nd (±2σ)</th>
</tr>
</thead>
<tbody>
<tr>
<td>JC08-WR416</td>
<td>30</td>
<td>18.7</td>
<td>453</td>
<td>0.703887</td>
<td>0.000014</td>
<td>0.1066</td>
<td>0.703842</td>
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<tr>
<td>07-C-6</td>
<td>27</td>
<td>24.3</td>
<td>465</td>
<td>0.703800</td>
<td>0.000015</td>
<td>0.1511</td>
<td>0.703889</td>
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<td>07-C-5</td>
<td>26</td>
<td>31.2</td>
<td>506</td>
<td>0.703757</td>
<td>0.000013</td>
<td>0.1783</td>
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<td>07-C-20</td>
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<td>0.703767</td>
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<td>JC08-WR415</td>
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<tr>
<td>08-C-2</td>
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<tr>
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<td>08-C-1</td>
<td>10</td>
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<td>589</td>
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<td>0.000009</td>
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<tr>
<td>08-C-11</td>
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<td>WR08-AE08</td>
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<td>16.2</td>
<td>567</td>
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<td>WR07-AE40</td>
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<td>0.000010</td>
<td>0.1449</td>
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</tbody>
</table>


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