

Age and volcanic stratigraphy of the Eocene Siletzia oceanic plateau in Washington and on Vancouver Island

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ABSTRACT

Geophysical, geochemical, geochronologic, and stratigraphic observations all suggest that the basalts that underlie western Oregon and Washington (USA), and southern Vancouver Island (Canada) form a coherent terrane of Eocene age, named Siletzia. The total volume of basalt within Siletzia is comparable to that observed in large igneous provinces and several lines of evidence point toward the terrane's origin as an accreted oceanic plateau. However, a thick sequence of continentally derived turbidites, named the Blue Mountain unit, has long been considered to floor the northern part of the terrane and its presence has led to alternative hypotheses in which Siletzia was built on the continental margin. We present new high-precision U-Pb zircon dates from silicic tuffs and intrusive rocks throughout the basaltic basement of northern Siletzia, as well as detrital zircon age spectra and maximum depositional ages for the Blue Mountain unit to help clarify the volcanic stratigraphy of this part of the terrane. These dates show that northern Siletzia was emplaced between 53.18 ± 0.17 Ma and 48.364 ± 0.036 Ma, similar to the age and duration of magmatism in the central and southern parts of the terrane. Turbidites in the basal Blue Mountain unit have maximum depositional ages as young as 44.72 ± 0.21 Ma and are distinctly younger than the basaltic basement that forms Siletzia. This age relationship implies that they were thrust under the terrane after 44.72 ± 0.21 Ma along one or more enigmatic faults. The younger age for these sedimentary rocks no longer requires construction of Siletzia on the continental margin, and we consider our revised stratigraphy to provide further support for the origin of the terrane as an accreted oceanic plateau.

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INTRODUCTION

Siletzia is a composite Eocene terrane, composed of the Siletz (Snively et al., 1968) and Crescent (Babcock et al., 1994) terranes, that forms a thick basaltic basement under much of western Washington and Oregon (USA), as well as southern Vancouver Island (Canada; Fig. 1A). The thickness of the terrane ranges from 10 km in the extreme north to 20–30 km throughout the rest of Siletzia, and estimates for the volume of erupted basalt are comparable to those seen in large igneous provinces (Trehu et al., 1994). Geochemical data further suggest that Siletzia includes basalts with isotopic compositions indicative of a plume-like mantle source (Pyle et al., 2009, 2015; Phillips et al., 2017). These

observations have led to the hypothesis that the terrane represents an accreted oceanic plateau or chain of seamounts (Duncan, 1982; Murphy et al., 2003; McCrory and Wilson, 2013; Wells et al., 2014; Murphy, 2016). Such an origin is further supported by evidence for Eocene-aged regional shortening on Vancouver Island (Johnston and Acton, 2003), western and central Washington (Tabor et al., 1984; Johnson, 1985; Eddy et al., 2016; Miller et al., 2016), and southern Oregon (Wells et al., 2000), as well as geophysical studies that show that Siletzia is connected with subducted oceanic crust (Gao et al., 2011; Schmandt and Humphreys, 2011). However, the hypothesis that Siletzia is an accreted oceanic terrane remains controversial, in part because a thick (>1–2 km) section of continentally derived turbidites is considered to underlie and interfinger with the northern part of the terrane (Tabor and Cady, 1978a, 1978b; Einarsen, 1987; Babcock et al., 1992; Brandon et al., 2014). The presence of these sedimentary rocks suggests that the basalts were erupted on the continental margin,

and several studies have proposed alternative tectonic settings for the construction of Siletzia, including a marginal rift (Wells et al., 1984; Babcock et al., 1992; Brandon et al., 2014) or near-trench magmatism related to ridge-trench interaction (Haeussler et al., 2003).

We present new high-precision U-Pb zircon geochronologic data from throughout the northern part of the Siletzia terrane that help us to better constrain the depositional and eruptive history of this area. These dates indicate that the basalts that form much of northern Siletzia were built between 53.18 ± 0.17 Ma and 48.364 ± 0.036 Ma, similar to the age and duration of magmatism within central and southern Siletzia (Wells et al., 2014). Detrital zircon data from the Blue Mountain unit provide maximum depositional ages that range between 47.775 ± 0.057 Ma and 44.72 ± 0.21 Ma, indicating that these sedimentary rocks are distinctly younger than Siletzia and that they were thrust under the terrane after 44.72 ± 0.21 Ma. This revised stratigraphy no longer requires that the terrane was

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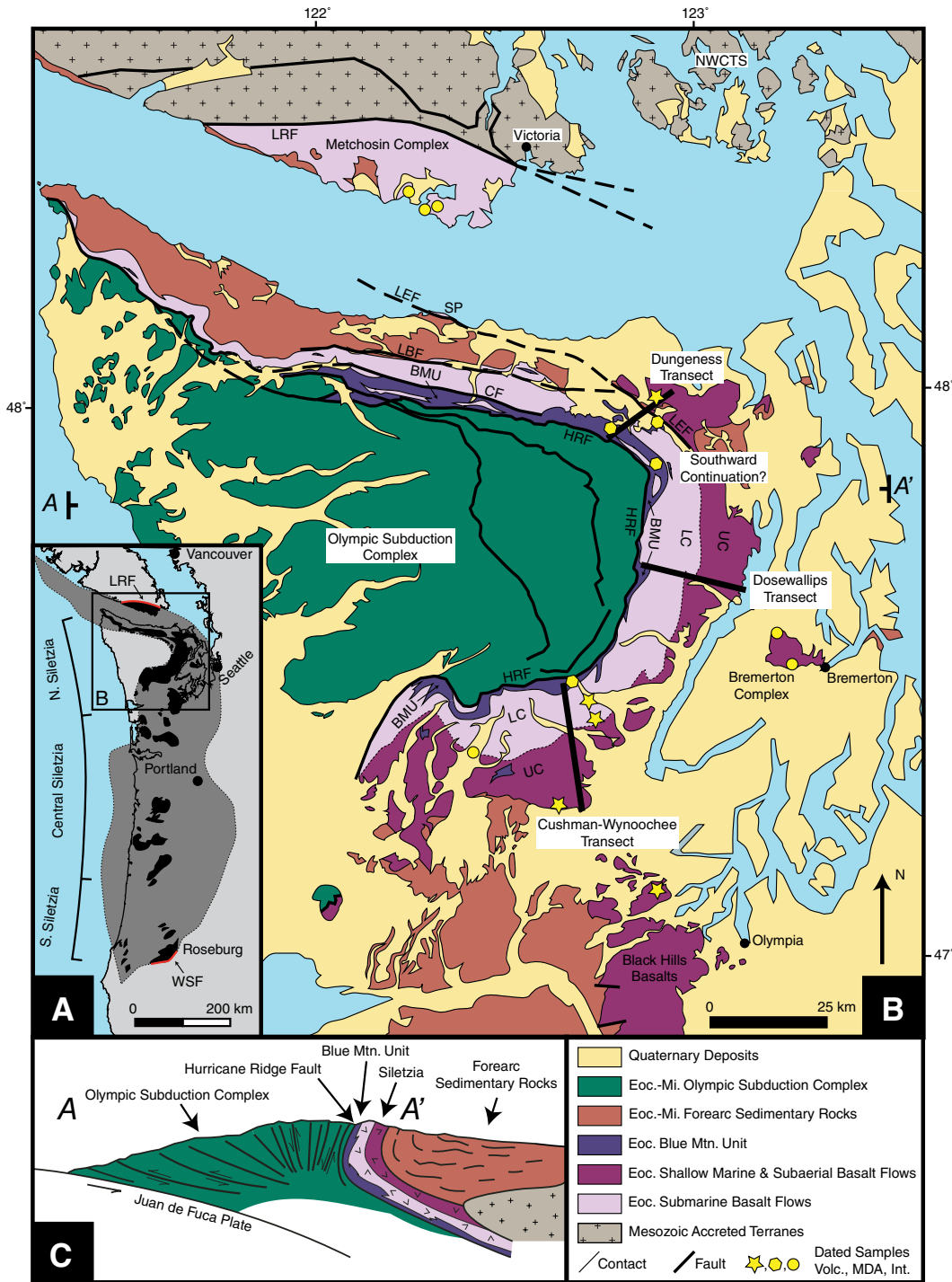


Figure 1. (A) Exposed (black) and subsurface (gray) portions of Siletzia within western Oregon, western Washington, and southern Vancouver Island modified from Wells et al. (2014). The terrane-bounding Wildlife Safari (WSF) and Leech River (LRF) faults are shown in red and the regional distinctions between northern, central, and southern Siletzia are also shown. (B) Simplified geologic map of the Olympic Peninsula and surrounding areas modified from Tabor and Cady (1978a), Walsh et al. (1987), Dragovich et al. (2002), and Massey et al. (2005). Abbreviations: BMU—Blue Mountain unit, CF—Crescent fault, HRF—Hurricane Ridge fault, LBF—Lake Creek–Boundary Creek fault, LRF—Leech River fault, LC—lower Crescent Formation, LEF—Lower Elwha fault, NWCTS—Northwest Cascades thrust system, SP—Striped Peak, UC—upper Crescent Formation. In this figure we follow Tabor and Cady (1978a) and show the upper Crescent Formation as largely subaerial on the southern Olympic Peninsula. However, mapping by one of us (K. Clark) shows that significant portions of the upper Crescent Formation are submarine in this area. The new geochronologic data presented in this study suggest that the submarine basalts on the northern Olympic Peninsula and an unknown thickness of submarine basalt on the southern Olympic Peninsula are unrelated to the Crescent Formation. Abbreviations in the key: Eoc.—Eocene, Int.—intrusive, MDA—maximum depositional age, Mi.—Miocene, Volc.—volcanic. (C) West-east cross section through the Olympic Peninsula modified from Tabor and Cady (1978b) showing the generalized structure of the region.

built on the continental margin and we consider our data to support the hypothesis that Siletzia is an accreted oceanic plateau.

GEOLOGIC SETTING

Exposures of Siletzia occur throughout the Coast Ranges of Washington and Oregon and along the southern tip of Vancouver Island (Fig. 1A). The terrane's eastern boundary is exposed as the Leech River fault on Vancouver Island (Figs. 1A, 1B; Groome et al., 2003) and the Wildlife Safari fault in southern Oregon (Fig. 1A; Wells et al., 2000), both of which place Mesozoic metasedimentary rocks over the terrane (Wells et al., 2000; Groome et al., 2003). Between these two faults the boundary is covered by middle Eocene and younger sedimentary and volcanic rocks. However, aeromagnetic data (Wells et al., 1998), ambient noise tomography (Gao et al., 2011), and body wave tomography (Schmandt and Humphreys, 2011) suggest that the terrane is continuous beneath this cover and that it is connected to a hanging slab of subducted oceanic crust in the upper mantle beneath eastern Washington and Oregon. To the west, the terrane is underthrust by the accretionary complex for the modern Cascade arc (e.g., Clowes et al., 1987; Trehu et al., 1994; Fleming and Trehu, 1999). This boundary is exposed on the Olympic Peninsula as the Hurricane Ridge fault (Fig. 1B; Cady, 1975; Tabor and Cady, 1978a, 1978b) and is traceable offshore to the north (Clowes et al., 1987) and south (Trehu et al., 1994; Fleming and Trehu, 1999).

Wells et al. (2014) used geochronologic, biostratigraphic, and paleomagnetic data to produce a sedimentary and volcanic stratigraphy of Siletzia; they showed that it consists of a thick basaltic basement that transitions from submarine to subaerial flows upsection. The basalts are predominantly mid-oceanic ridge basalt-like (MORB) with subordinate oceanic island basalt-like (OIB) and alkaline basalts that contain Pb, Sr, Nd, He, Hf, and Os isotopic compositions consistent with their derivation from a plume-like mantle source (Pyle et al., 2009, 2015; Phillips et al., 2017). Eruption of these basalts likely started by 56 Ma in the south and by 53 Ma in the central and northern parts of the terrane, based on available $^{40}\text{Ar}/^{39}\text{Ar}$ dates, while the transition to subaerial eruption is constrained between 52 ± 1 Ma and 49 ± 0.8 Ma by U-Pb zircon dates of silicic tuffs in central Siletzia (Wells et al., 2014). Following the transition to subaerial eruptions, southern Siletzia was deformed into a west-vergent (following restoration of $\sim 80^\circ$ of clockwise rotation) fold-thrust belt that is best documented near Roseburg, Oregon (Fig. 1A; Wells et al., 2000). Biostratigraphic constraints

as well as maximum depositional ages for non-deformed sedimentary rocks that overlie the basaltic basement in this area require that deformation occurred prior to 48 Ma (Dumitru et al., 2013; Wells et al., 2014). A similar fold-thrust belt has not been recognized in central and northern Siletzia. However, Tabor et al. (1984) and Johnson (1985) documented shortening in non-marine Eocene sedimentary sequences throughout central and western Washington that Eddy et al. (2016) constrained to have occurred between 51.309 ± 0.024 Ma and 49.933 ± 0.059 Ma. Similarly, Johnston and Acton (2003) documented a period of shortening on southern Vancouver Island ca. 50 Ma, and Wells et al. (2014) documented folding in central Siletzia between 49.0 ± 0.8 Ma and ca. 48 Ma.

Deposition of sedimentary rocks that belong to a regional forearc sedimentary basin began soon after Siletzia was shortened, and these rocks blanket the terrane. Biostratigraphic data and maximum depositional ages from detrital zircons in these rocks constrain the transition to have occurred by 50–48 Ma in southern Oregon (Dumitru et al., 2013; Wells et al., 2014) and the reestablishment of east-west paleoflow following regional shortening in nonmarine sedimentary rocks of central Washington by 45.910 ± 0.021 Ma may provide a minimum date for this transition in Washington (Eddy et al., 2016). Subsidence of this forearc basin was nearly continuous from the middle Eocene to the Miocene (e.g., Babcock et al., 1994; Ryu et al., 1996), when the basin was inverted during uplift of the modern accretionary prism. Interbedded with the basin's sedimentary rocks are isolated and geochemically diverse basaltic magmatic centers that range between 48 and 34 Ma in age. These centers include the basalt of Hembre Ridge, Tillamook Volcanics, Grays River volcanics, and Yachats Head volcanics (Wells et al., 2014), some of which contain isotopic evidence for derivation from plume-like mantle (e.g., Chan et al., 2012).

The northernmost exposures of Siletzia occur on southern Vancouver Island and in western Washington (Fig. 1B). This area has a stratigraphy similar to that of central and southern Siletzia and consists of a basaltic basement overlain by middle Eocene to Miocene forearc sedimentary rocks (Fig. 1B; Babcock et al., 1994; Wells et al., 2014). Outcrops of the basaltic basement occur as the Crescent Formation, Metchosis complex, Bremerton complex, and Black Hills basalt. Existing geochronology (Duncan, 1982; Clark, 1989; Babcock et al., 1992, 1994; Hirsch and Babcock 2009; Polenz et al., 2012a, 2012b, 2016) shows that the timing of magmatism and initial deposition of the overlying sedimentary rocks is similar to that in central

and southern Siletzia (Wells et al., 2014). However, a >1 – 2 -km-thick section of continentally derived turbidites known as the Blue Mountain unit has been interpreted as the base of Siletzia in Washington (Fig. 1B; Tabor and Cady, 1978a, 1978b; Einarsen, 1987; Babcock et al., 1992, 1994). The presence of these sedimentary rocks is difficult to reconcile with the hypothesis that Siletzia represents an accreted oceanic terrane and it has spurred alternative hypotheses that consider it to have formed on the continental margin as a marginal rift (Wells et al., 1984; Babcock et al., 1992; Brandon et al., 2014), or as near-trench magmatism (Haeussler et al., 2003). Nevertheless, the stratigraphy of northern Siletzia has never been rigorously tested using high-precision geochronology.

U-Pb ZIRCON GEOCHRONOLOGY

Previous geochronology from northern Siletzia largely consists of whole-rock K-Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ dates (Duncan, 1982; Clark, 1989; Babcock et al., 1992, 1994; Hirsch and Babcock 2009; Polenz et al., 2012a, 2012b, 2016). These data have helped establish that this part of the terrane is early Eocene in age, but they are relatively imprecise and largely prevent fine temporal correlations between isolated outcrop areas. Furthermore, variable metamorphism (Timpa et al., 2005; Hirsch and Babcock, 2009) and alteration of basalts and gabbros within northern Siletzia make it difficult to unambiguously interpret whole-rock K-Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ dates as crystallization ages. U-Pb zircon geochronology circumvents these problems, but existing data are limited.

We present 11 new U-Pb zircon chemical abrasion–isotope dilution–thermal ionization mass spectrometry (CA-ID-TIMS) dates from throughout northern Siletzia that help better constrain the regional stratigraphy. The methods for this procedure are slightly modified from Mattinson (2005) and are described in detail in Eddy et al. (2016, Appendix A therein). All isotopic ratios were measured on either the VG Sector 54 or Isotopx X62 thermal ionization mass spectrometers at the Massachusetts Institute of Technology (MIT) and the data are presented in [Table DR1](#)¹. We use the $^{206}\text{Pb}/^{238}\text{U}$ date for all of our interpretations because it offers the most precise date for rocks of this age, and we correct for preferential exclusion of ^{230}Th during zircon crystallization using the calculated $[\text{Th}/\text{U}]_{\text{zircon}}$ and assuming a $[\text{Th}/\text{U}]_{\text{magma}} = 2.8$ for silicic tuffs

¹GSA Data Repository Item 2017232, U-Pb isotopic data and concordia plots, is available at <http://www.geosociety.org/datarepository/2017>, or on request from editing@geosociety.org.

(e.g., Machlus et al., 2015) or a $[Th/U]_{\text{magma}} = 3.2$ for mafic intrusive rocks (e.g., Gale et al., 2013). An arbitrarily assigned uncertainty of ± 1 (2σ) is assigned to all $[Th/U]_{\text{magma}}$. We assume that zircon does not incorporate Pb during its crystallization and that all measured common Pb (Pb_c) is from laboratory contamination. We correct for this contamination using the measured mass of ^{204}Pb and a laboratory Pb_c isotopic composition of $^{206}Pb/^{204}Pb = 18.145833 \pm 0.475155$ (1σ absolute [abs.]), $^{207}Pb/^{204}Pb = 15.303903 \pm 0.295535$ (1σ abs.), and $^{208}Pb/^{204}Pb = 37.107788 \pm 0.875051$ (1σ abs.), calculated from 149 procedural blanks measured in the MIT isotope geochemistry laboratory. Samples were spiked with the EARTHTIME ^{205}Pb - ^{233}U - ^{235}U isotopic tracer (Condon et al., 2015; McLean et al., 2015) and all data reduction was done on the U-Pb_REDUX software package (Bowring et al., 2011) using the U decay constants presented in Jaffey et al. (1971). Our preferred date for each sample is presented in Table 1 and the data are shown as traditional concordia plots in Figure DR1. With the exception of a single xenocrystic or antecrystic grain in CR-MPE-003, zircons from all igneous samples have age dispersion within the 2σ variability of the mean square of weighted deviates (MSWD; Wendt and Carl, 1991). Therefore, all of our preferred dates for igneous samples represent weighted means. Uncertainties are reported in the format $\pm X/Y/Z$ in Table 1, where X represents the

analytical uncertainty, Y includes uncertainty in the isotopic tracer calibration, and Z represents the full uncertainty and includes uncertainty in the ^{238}U decay constant. However, other references to the dates within this text report the analytical uncertainty only.

Detrital zircons were analyzed from four sandstones collected from the Blue Mountain unit by laser ablation–inductively coupled plasma–mass spectrometry (LA-ICP-MS) on a Thermo-Fisher Element 2 at the Arizona Laser-Chron Center (Tucson, Arizona). The methods for these analyses follow those in Ibañez-Mejía et al. (2015) and all isotopic data are presented in Tables DR2, DR3, DR4, and DR5. To obtain maximum depositional ages for the sandstones, we removed some of the youngest grains identified during LA-ICP-MS analysis and dated them by chemical abrasion–isotope dilution–thermal ionization mass spectrometry. These dates are presented in Table DR1 and the youngest grain dated by CA-ID-TIMS in each sample is reported in Table 1 as a maximum depositional age.

RESULTS

Blue Mountain Unit and Crescent Formation

The thickest exposed section of Siletzia is on the Olympic Peninsula in western Washington (Fig. 1B), where uplift and antiformal doming of

the Olympic subduction complex has tilted the terrane (Figs. 1B, 1C; Tabor and Cady, 1978b; Brandon and Calderwood, 1990; Babcock et al., 1992, 1994; Wells et al., 2014). In this area Siletzia is traditionally considered to consist of, in stratigraphic order, continentally derived turbidites belonging to the Blue Mountain unit, a thick section of submarine basalt flows named the lower Crescent Formation, and a thick section of shallow-marine or subaerial basalt flows named the upper Crescent Formation. All three units can be traced around the perimeter of the Olympic Peninsula (Fig. 1B), but the validity of this inferred stratigraphy has never been confirmed using geochronologic data. Babcock et al. (1992) estimated the thickness of the Blue Mountain unit and Crescent Formation along the Dosewallips River (Fig. 1B), where the section consists of ~ 1 – 2 km of turbidites belonging to the Blue Mountain unit, ~ 8.4 km of submarine basalt flows in the lower Crescent Formation, and ~ 7.8 km of shallow-marine or subaerial basalt flows in the upper Crescent Formation. Elsewhere on the Olympic Peninsula exposed portions of these units are thinner.

The Blue Mountain unit was described in detail by Einarsen (1987) and is composed of turbidites and pebble conglomerates interbedded with minor basalt. The base of the unit is truncated by the Hurricane Ridge fault, which places rocks of the Olympic subduction complex under the Blue Mountain unit and the

TABLE 1. U-PB ZIRCON CA-ID-TIMS GEOCHRONOLOGY RESULTS

Sample	Lat. ($^{\circ}N$)	Long. ($^{\circ}W$)	Lithology	Th-corrected $^{206}Pb/^{238}U$ Date (Ma)*,†	Interpretation
Metchosin complex					
CR-MPE-004A	48.33886	123.71416	Quartz Diorite	$51.115 \pm 0.070/0.077/0.094$ (MSWD=0.88, n=5)	Emplacement age
CR-MPE-007	48.35397	123.68497	Plagiogranite	$50.986 \pm 0.023/0.033/0.064$ (MSWD=1.74, n=6)	Emplacement age
CR-MPE-010A	48.37499	123.75426	Gabbro	$51.176 \pm 0.023/0.033/0.064$ (MSWD=2.01, n=6)	Emplacement age
Bremerton complex					
CR-MPE-003	47.59377	122.79125	Gabbro	$50.075 \pm 0.016/0.027/0.060$ (MSWD=1.74, n=6)	Emplacement age
BH07	47.53390	122.78248	Andesitic Dike	$48.209 \pm 0.057/0.068/0.085$ (MSWD=0.78, n=9)	Emplacement age
Black Hills basalt					
CR-MPE-017B	47.13748	123.14439	Silicic tuff	$49.729 \pm 0.014/0.026/0.059$ (MSWD=1.84, n=8)	Eruption age
Blue Mountain unit					
CR-MPE-015	47.51211	123.34895	Sandstone	$47.775 \pm 0.057/0.064/0.082$	MDA#
CR-MPE-018	47.97032	123.11235	Sandstone	$46.428 \pm 0.048/0.053/0.073$	MDA#
CR-MPE-020	47.89255	123.12283	Sandstone	$44.72 \pm 0.21 /0.22 /0.22$	MDA#
CR-MPE-022	47.94748	123.22704	Sandstone	$45.15 \pm 0.21 /0.22 /0.23$	MDA#
Crescent Formation					
CR-MPE-024	47.38776	123.60406	Gabbro	$48.624 \pm 0.023/0.032/0.061$ (MSWD=1.25, n=6)	Emplacement age
HS-11-13-97-8 [§]	48.00833	123.12056	Rhyolite	$51.424 \pm 0.027/0.040/0.068$ (MSWD=2.12, n=7)	Eruption age
KC-176.1 [§]	47.46661	123.28426	Silicic tuff	$53.18 \pm 0.17 /0.22 /0.23$ (MSWD=0.43, n=7)	Eruption age
KC-226 [§]	47.44855	123.28672	Silicic tuff	$51.059 \pm 0.068/0.084/0.10$ (MSWD=2.10, n=5)	Eruption age
KC-300 [§]	47.29864	123.38645	Silicic tuff	$48.364 \pm 0.036/0.054/0.075$ (MSWD=0.79, n=6)	Eruption age

*Th correction was done using $[Th/U]_{\text{magma}}=2.8 \pm 1$ (2σ) for silicic volcanic rocks and $[Th/U]_{\text{magma}}=3.2 \pm 1$ (2σ) for gabbro, quartz diorite, and plagiogranite.

†Uncertainties are reported in the format $\pm X/Y/Z$ where X is the analytical uncertainty, Y includes uncertainty in the EARTHTIME ^{205}Pb - ^{235}U - ^{238}U isotopic tracer, and Z includes uncertainty in the ^{238}U decay constant.

§Latitude and longitude for these samples is estimated from marked positions on 1:24,000-scale topographic maps. All other samples were collected with a GPS.

#MDA—Maximum depositional age.

structurally overlying Crescent Formation. Around most of the Olympic Peninsula, the Blue Mountain unit is ~1–2 km thick and conformably overlain by submarine basalts (Einarsen, 1987; Babcock et al., 1992, 1994). On the northeastern Olympic Peninsula, the Blue Mountain unit has been mapped as a thick section (~6.8 km) of turbidites and conglomerates that are interbedded with submarine basalts along the upper Dungeness River and its tributaries (Dungeness transect in Fig. 1B). Initial geologic mapping suggested that these rocks included a syncline of younger sedimentary rock that unconformably overlies the Blue Mountain unit (Cady et al., 1972). However, subsequent mapping has consistently considered all of the sedimentary rock in this area to belong to the Blue Mountain unit (Tabor and Cady, 1978a; Einarsen, 1987; Gerstel and Lingley, 2003) and the most detailed study of this area considered it to represent a large channel through which the sediment deposited in the rest of the Blue Mountain unit was fed (Einarsen, 1987). The best existing date for the Blue Mountain unit is a maximum depositional age of ca. 48.7 Ma based on detrital zircons separated from a turbidite collected 10 km to the north of the Dosewallips section (Wells et al., 2014).

The Crescent Formation structurally overlies the Blue Mountain unit and is informally divided into lower and upper members based on a change from submarine to shallow-marine and subaerial basalt flows upsection (e.g., Cady, 1975; Tabor and Cady, 1978a; Babcock et al., 1992, 1994). The lower Crescent Formation is dominantly of normal MORB composition, while the upper Crescent is more geochemically variable and contains basalts with enriched MORB and OIB-like compositions (Babcock et al., 1992). On the basis of pervasively fractured basalt near the contact between these units, Glassley (1974) interpreted the two units to be tectonically juxtaposed along a major fault. However, more recent mapping has considered the contact between the two informal units to be conformable and to represent a change in eruptive environment from deep to shallow water (e.g., Cady, 1975; Tabor and Cady, 1978a; Babcock et al., 1992). Sedimentary rocks belonging to a regional Eocene to Miocene forearc basin overlie the upper Crescent Formation. This contact is locally unconformable. However, in many areas the sedimentary rocks are described as interbedded with the uppermost Crescent Formation (Babcock et al., 1994).

We sampled two composite sections through the Blue Mountain unit and the Crescent Formation for high-precision U-Pb zircon geochronology. The first sampling transect is along the upper Dungeness River and adjacent areas in the

northeastern Olympic Peninsula (Fig. 1B). This is the area that Einarsen (1987) considered to represent a channel and consists of ~6.8 km of sedimentary rock interbedded with submarine basalt flows. These rocks are separated from the upper Crescent Formation and unconformably overlying Eocene to Oligocene sedimentary rocks by the Lower Elwha fault (Figs. 1B and 2; Schasse, 2003). Three sandstone layers were collected from turbidites within the Blue Mountain unit in order to generate maximum depositional ages. All three samples are from areas that have consistently been mapped as the Blue Mountain unit and are outside of the area that Cady et al. (1972) proposed to include a syncline of younger strata. Two of the samples from the base of the Blue Mountain unit (CR-MPE-020 and CR-MPE-022) have maximum depositional ages of 44.72 ± 0.21 Ma and 45.15 ± 0.21 Ma (Fig. 2), while a sandstone from the top of the Blue Mountain unit (CR-MPE-018) has a maximum depositional age of 46.428 ± 0.048 Ma (Fig. 2). The progression from younger to older maximum depositional ages upsection may indicate that the Blue Mountain unit is not stratigraphically continuous below the Lower Elwha fault. However, because these are maximum depositional ages, this interpretation is speculative. Above the Lower Elwha fault, we dated a rhyolite within the upper Crescent Formation (HS-11–13–97–8). This sample yielded an eruption or deposition age of 51.424 ± 0.027 Ma (Fig. 2), which is older than the structurally underlying Blue Mountain unit and indicates that the Lower Elwha fault represents a significant stratigraphic discontinuity in this area.

The second sampling transect was through the area between Lake Cushman and Lake Wynoochee (Fig. 1B), where the combined thickness of the Blue Mountain unit and Crescent Formation is ~12 km (Fig. 2). Mapping by K. Clark (personal data) shows that the boundary between the lower and upper Crescent Formation in this area is slightly higher in the section than the boundary mapped by Tabor and Cady (1978a) and is marked by a continuous (~25 km along strike) section of turbidites and other marine sedimentary rocks that are locally unconformable on the underlying pillow basalts. We dated five samples from this transect. A sandstone layer from a turbidite sequence in the Blue Mountain unit (CR-MPE-015) has a maximum depositional age of 47.775 ± 0.057 Ma (Fig. 2), while two silicic tuffs within the lower Crescent Formation (KC-176.1 and KC-226) yielded eruption or deposition dates of 53.18 ± 0.17 Ma and 51.059 ± 0.068 Ma (Fig. 2). A gabbro that intrudes the lower Crescent Formation has an emplacement date of 48.624 ± 0.023 Ma, and a silicic tuff from within the upper Crescent

Formation (KC-300) has an eruption and deposition date of 48.364 ± 0.036 Ma (Fig. 2). The maximum depositional age for the Blue Mountain unit is distinctly younger than overlying samples, indicating that the section is not stratigraphically continuous and that a major fault is between CR-MPE-015 and KC-176.1 (Fig. 2).

A probability density function of the 956 detrital zircons from the Blue Mountain unit that were analyzed during this study is shown in Figure 3. Age peaks align with major periods of pluton construction in the adjacent Coast Mountain batholith (Gehrels et al., 2009) and its southern extension in the North Cascades (Miller et al., 2009). This observation is in good agreement with Einarsen's (1987) interpretation that the Blue Mountain unit was locally derived from the Coast Mountain batholith and metasedimentary rocks within the Northwest Cascades thrust system (Fig. 1B; e.g., Brown, 1987), which also contain detrital zircon peaks corresponding to major periods of magmatism in the Coast Mountain batholith (Brown and Gehrels, 2007). Two Proterozoic age peaks ca. 1380 and between 1800 and 1600 Ma are also present and are common in Late Cretaceous to Eocene sedimentary rocks along the northern Cordilleran margin (e.g., Garver and Davidson, 2015; Dumitru et al., 2016). These peaks have been attributed to deposition adjacent to the Yavapai-Mazatzal and Mojave Provinces (1800–1600 Ma) and Granite-Rhyolite Province (1400–1300 Ma) in the American southwest and subsequent margin-parallel, northward translation (Garver and Davidson, 2015), or recycling of detrital zircons (1800–1600 Ma) and erosion of magmatic rocks (1380 Ma) from the Belt Basin after its uplift during the Laramide orogeny (Dumitru et al., 2016). No paleomagnetic data exist for the Blue Mountain unit. However, paleomagnetic data from pillow basalts adjacent to the thickest part of the Blue Mountain unit on the northeastern part of the Peninsula suggest little to no poleward motion since the middle Eocene (Warnock et al., 1993). We therefore consider the presence of the 1380 and 1800–1600 Ma detrital zircon peaks to indicate either sediment transport from the uplifted Belt Basin to the North American margin during the Eocene (e.g., Dumitru et al., 2016), or recycling of detrital zircons from uplifted Late Cretaceous to early Paleogene sedimentary rocks along the North American margin.

Metchosin Complex

The Metchosin complex is exposed on the southern tip of Vancouver Island (Fig. 1B) and has a partial ophiolite stratigraphy composed of a basal section of gabbro, quartz diorite, and plagiogranite overlain by a thin sheeted

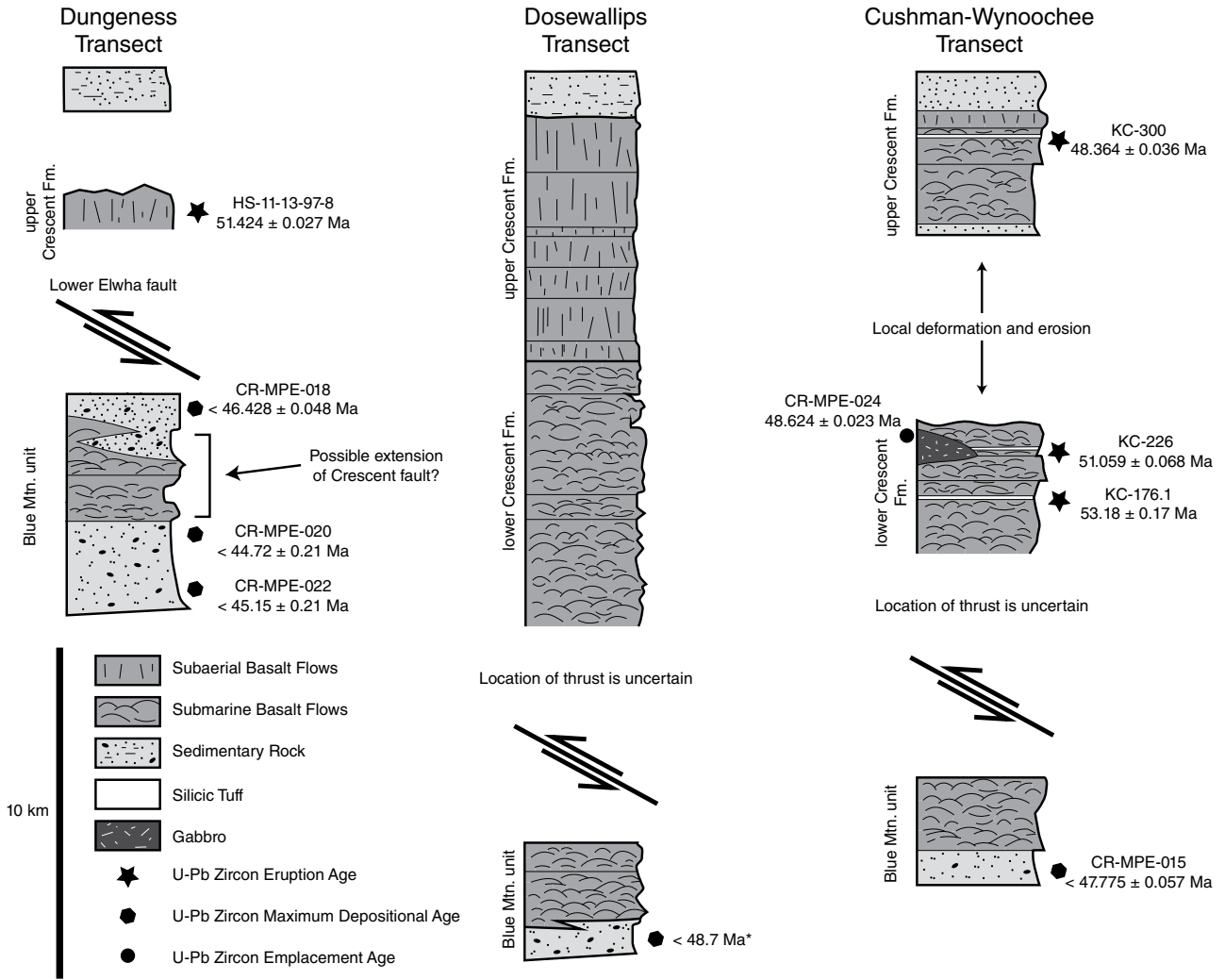


Figure 2. Stratigraphic columns showing the locations of dated samples for the Dungeness, Dosewallips, and Cushman-Wynoochee transects through the Blue Mountain (Mtn.) unit and Crescent Formation (Fm.). Thicknesses are estimated from Cady et al. (1972) for the Dungeness transect and from our mapping for the Cushman-Wynoochee transect (K. Clark). The column for the Dosewallips section is modified from Babcock et al. (1992). All dates are from this study with the exception of a U-Pb detrital zircon date from Wells et al. (2014; denoted by an asterisk).

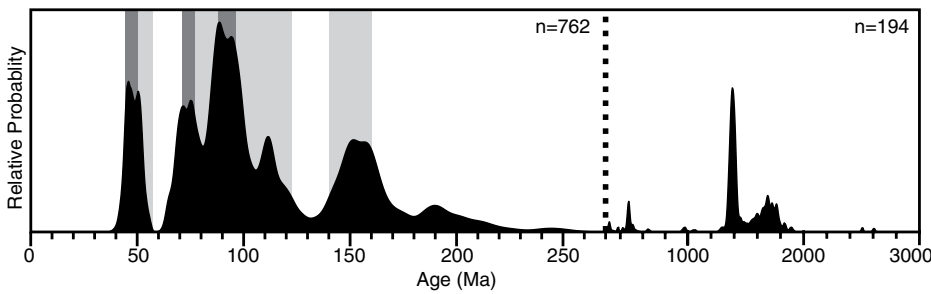


Figure 3. Probability density function for the 956 detrital zircons from the Blue Mountain unit that were dated as part of this study. Major periods of Mesozoic and Paleogene magmatism in the Coast Mountain batholith (Gehrels et al., 2009) and its southern extension in the North Cascades (Miller et al., 2009) are shown as light and dark shaded bars, respectively. See text for discussion about the origin of the Proterozoic age peaks. Note the change in scale at 270 Ma.

dike complex and ~2.5 km of basalt flows that transition from submarine to subaerial upsection (Fig. 4; Massey, 1986). The lower part of the complex is not exposed. However, seismic reflection data suggest that it is truncated by sedimentary rocks belonging to the accretionary prism of the modern Cascade arc along a major thrust fault (Clowes et al., 1987). The Metchosin complex was thrust under Mesozoic metasedimentary rocks during the Eocene along the Leech River fault (Fig. 1B; Groome et al., 2003) and records metamorphic grades between prehnite-actinolite and amphibolite facies that are likely related to this event (Timpa et al., 2005). Exhumation to the surface was complete by the late Eocene and is marked by the unconformable deposition of sedimentary rocks belonging to the Carmanah group above the complex (Muller, 1977). The only previous U-Pb zircon date for the Metchosin complex is a poorly documented 52 ± 2 Ma date from a gabbro (personal commun. from R. Zartman to M.T. Brandon, cited in Massey, 1986). We dated three samples from the intrusive part of

the Metchosin complex (Fig. 4): a quartz diorite (CR-MPE-004B), a plagiogranite (CR-MPE-007), and a gabbro (CR-MPE-010A). All three samples were emplaced between 51.176 ± 0.023 Ma and 50.986 ± 0.023 Ma and are in good agreement with the previously published U-Pb date for the complex.

Bremerton Complex

The Bremerton complex is exposed in a structural uplift near Bremerton, Washington (Fig. 1B). It has a partial ophiolite stratigraphy, similar to the Metchosin complex, and includes a basal section of gabbro and plagiogranite overlain by a thin section of sheeted dikes and <1.4 km of basalt flows that transition from submarine to subaerial upsection (Clark, 1989). Gravity and magnetic surveys suggest that ultramafic rocks may exist immediately below these exposures (Haeussler and Clark, 2000; Tabor et al., 2011), indicating that a nearly complete ophiolite stratigraphy exists in this area. The upper contact of the complex is

not exposed. However, hornblende dacite dikes with an adakite-like composition cut all of the exposed units and provide a constraint on the end of basaltic magmatism (Tepper et al., 2004). A date of 50.4 ± 0.6 Ma from a leucogabbro is the only previously published U-Pb zircon date for the Bremerton complex (Haeussler and Clark, 2000; Tabor et al., 2011). We dated two samples from the area (Fig. 1B). A sample of pegmatitic gabbro within the intrusive part of the complex yielded an emplacement date of 50.075 ± 0.016 Ma and overlaps within uncertainty of the previously published U-Pb zircon date. A hornblende dacite dike gave an emplacement date of 48.209 ± 0.057 Ma and provides a minimum age constraint on the construction of the complex (Fig. 4).

Black Hills Basalts

The Black Hills basalts are exposed west and south of Olympia, Washington (Fig. 1B). The total thickness of this section is >600 m and it consists of mixed submarine and subaerial

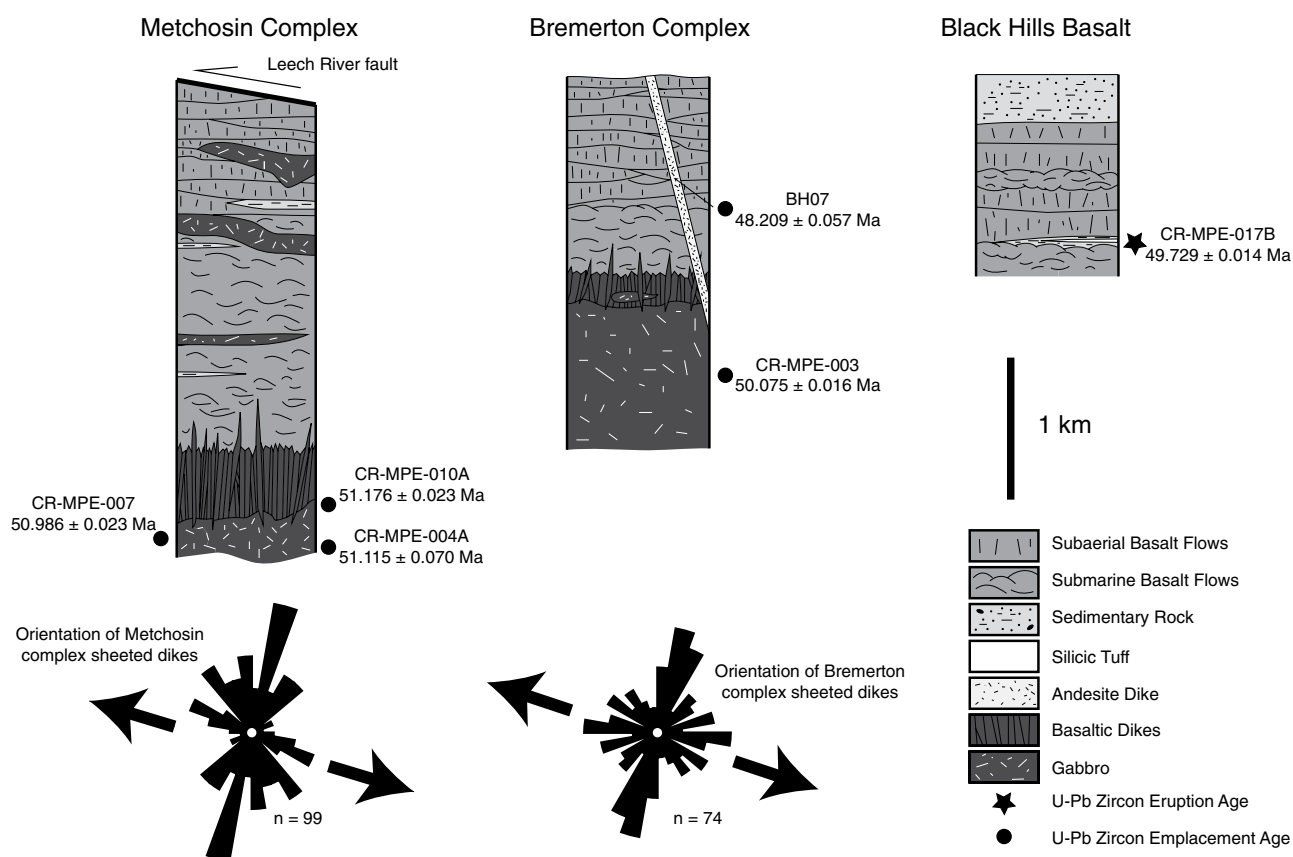


Figure 4. Generalized stratigraphic columns for the Metchosin complex, Bremerton complex, and Black Hills basalts that show unit thicknesses and the locations of dated samples. The columns for the Metchosin and Bremerton complexes are modified from Massey (1986) and Clark (1989), respectively. Rose diagrams of dike orientations within the Metchosin and Bremerton complexes are shown below their respective stratigraphic columns and use data from Massey (1986) and Clark (1989). The heavy arrows show the implied extensional direction for both areas. All dates are from this study.

basalt flows that are overlain by Eocene to Oligocene sedimentary rocks (Globerman et al., 1982). The basalts have traditionally been considered part of the upper Crescent Formation and new whole-rock geochemical data support this hypothesis (Polenz et al., 2016). We dated a thin (~1 cm) bentonite (CR-MPE-017B) within a 1–2-m-thick sedimentary section interbedded with the basalts (Fig. 4) that provided an eruption date of 49.729 ± 0.014 Ma. This is the first U-Pb zircon date from this area.

DISCUSSION

Volcanic Stratigraphy of Northern Siletzia

Our new data provide important constraints on the volcanic stratigraphy of northern Siletzia. They show that the Blue Mountain unit is distinctly younger than the structurally overlying Crescent Formation and must have been thrust under Siletzia after 44.72 ± 0.21 Ma. This relationship has been documented on both the northern and southern part of the Olympic Peninsula and is consistent with the U-Pb zircon maximum depositional age of ca. 48.7 Ma previously reported for the Blue Mountain unit near the base of the Dosewallips section (Fig. 2; Wells et al., 2014). It is interesting that a conformable contact between the upper Blue Mountain unit and the overlying submarine basalt flows suggests that an unknown thickness of basalt is also younger than 44.72 ± 0.21 Ma in age. On the northeastern part of the Olympic Peninsula, where the Blue Mountain unit is interbedded with submarine basalts along the margin of a large sedimentary channel (Einarsen, 1987), the thickness of these basalts may be as much as 6–7 km, and include most of the basalts previously assigned to the lower Crescent Formation in this area. However, the inverted progression in maximum depositional ages within the Blue Mountain unit on the northeastern Olympic Peninsula provides weak evidence that the sedimentary and volcanic rocks in this area may also be structurally thickened.

The fault or faults along which the Blue Mountain unit was thrust under the Crescent Formation remain enigmatic. However, detailed mapping on the northern Olympic Peninsula has identified three high-angle thrust faults that are possible candidates (Fig. 1B): the Lower Elwha, the Lake Creek–Boundary Creek, and the Crescent faults (Brown et al., 1960; MacLeod et al., 1977; Tabor and Cady, 1978a; Atkins et al., 2003; Schasse, 2003). These faults all have north-side-up displacements, can be traced over tens of kilometers, and imbricate the Blue Mountain unit, submarine basalts, and the forearc sedimentary rocks that overlie northern Siletzia. We consider it likely that one or more of these

structures placed the Blue Mountain unit and associated basalts under the Crescent Formation as a west-vergent, low-angle thrust during early construction of the Olympic subduction complex. Subsequent uplift and antiformal doming of the complex (e.g., Tabor and Cady, 1978a, 1978b; Brandon and Calderwood, 1990) may explain their current steep orientations. The possibility for large displacements on the Lower Elwha fault is particularly intriguing because it has placed basalts of the Crescent Formation on top of late Eocene to Oligocene sedimentary rocks near Striped Peak, is traceable over ≥ 50 km, truncates the Lake Creek–Boundary Creek fault near the Dungeness transect, and is mapped as continuous around part of the antiformal dome that defines the geologic structure of the Olympic Peninsula (Fig. 1B; Dragovich et al., 2002). Nevertheless, it is likely that the Lake Creek–Boundary Creek and/or the Crescent faults also have large displacements, because our geochronologic data suggest that the Blue Mountain unit is the same age or younger than the oldest structurally overlying forearc sedimentary rocks on the northern Olympic Peninsula. This relationship precludes a continuous stratigraphic sequence between the Hurricane Ridge and Lower Elwha faults and is also consistent with the inverted progression of maximum depositional ages within the Blue Mountain unit on the northeastern Olympic Peninsula (Fig. 2). No thrust faults equivalent to the Lower Elwha, Lake Creek–Boundary Creek, or Crescent faults have been mapped on the eastern or southern parts of the Olympic Peninsula. However, rough terrain, heavy vegetation, and a lack of clear stratigraphic markers within the Crescent Formation make it difficult to determine the structure of this area, and we emphasize that our geochronologic data require that such structures exist.

Our data show that widespread mafic magmatism occurred throughout northern Siletzia from 53.18 ± 0.17 Ma until slightly after 48.364 ± 0.036 Ma. During that time, the lower and upper Crescent Formation, the Bremerton complex, Metchosin complex, and Black Hills basalts were emplaced. The basement onto which these basalts were erupted is not exposed in the Crescent Formation or in the Black Hills basalts. However, both the Metchosin and Bremerton complexes contain an intrusive basement consistent with their formation at an oceanic spreading center coeval with volcanism throughout the rest of the terrane (e.g., Massey, 1986; Clark, 1989). Therefore, we conclude that northern Siletzia was likely constructed on young oceanic crust. All outcrop areas in northern Siletzia show a progressive change in eruptive environment, from deep-marine pillow lavas to shallow-marine and subaerial basalt flows during construction of the

terrane, suggesting regional emergence between 51.424 ± 0.027 Ma and 48.364 ± 0.036 Ma (Figs. 2 and 4). Shortly after 48.364 ± 0.036 Ma the terrane subsided below sea level and initial deposition of marine sedimentary rocks in a regional forearc basin began. This eruptive and depositional history closely mirrors that of central and southern Siletzia (Fig. 5; Wells et al., 2014), and provides further support that Siletzia represents a coherent terrane.

Tectonic Setting

The tectonic setting of Siletzia has been a longstanding problem in Cordilleran geology, with debate centering on whether it represents an accreted oceanic terrane (Duncan, 1982; Murphy et al., 2003; McCrory and Wilson, 2013; Wells et al., 2014; Murphy, 2016) or magmatism along the continental margin related to either rifting (Babcock et al., 1992, 1994; Brandon et al., 2014) or ridge-trench interaction (Haeussler et al., 2003). One of the most compelling arguments for eruption on the continental margin was the inferred stratigraphic position of the Blue Mountain unit at the base of the terrane in Washington (e.g., Cady, 1975; Tabor and Cady, 1978a, 1978b; Einarsen, 1987; Babcock et al., 1992). However, our new U-Pb geochronology demonstrates that these rocks were thrust under Siletzia after 44.72 ± 0.21 Ma (Fig. 2). Therefore, their presence no longer necessitates that the basalts that compose Siletzia were erupted on preexisting continental crust.

Several researchers have proposed that Siletzia represents an accreted oceanic plateau, or series of oceanic islands, that developed above a hotspot (Duncan, 1982; Murphy et al., 2003; McCrory and Wilson, 2013; Wells et al., 2014; Murphy, 2016; Phillips et al., 2017). Such an origin is consistent with the projected position of a long-lived Yellowstone hotspot near western Oregon during the early Eocene (e.g., Engebretson et al., 1985), and could explain the great volume of basalt erupted within the terrane as well as geochemical evidence for a plume-like mantle source for the basalts (Pyle et al., 2009, 2015; Phillips et al., 2017). Duncan (1982), McCrory and Wilson (2013), and Wells et al. (2014) went further and proposed that along-strike variation in the ages of the basaltic basement in Siletzia may indicate that the terrane was built along an oceanic spreading center, in a setting analogous to present-day Iceland. This proposal is largely based on the observation that initial K-Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ dates become systematically younger toward the center of the terrane (Duncan, 1982), and plate reconstructions that require the intersection of one or more oceanic spreading centers (Kula-Farallon or Kula-Farallon and

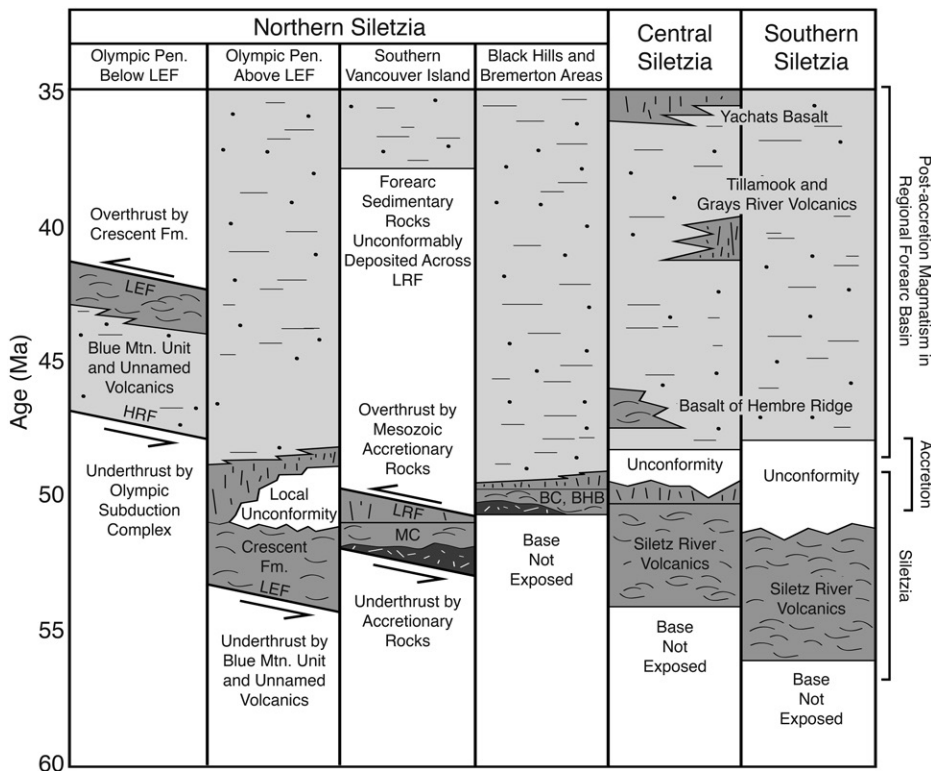


Figure 5. Temporal correlations between magmatism in northern, central, and southern Siletzia. See Figure 1A for a key to the geographic distinctions. Northern Siletzia is divided into four columns; rocks on the Olympic Peninsula between the Hurricane Ridge fault and the Lower Elwha fault (LEF; and its presumed southern continuation), rocks on the Olympic Peninsula above the LEF, southern Vancouver Island, and a generalized section for the Black Hills and Bremerton areas. Note that on the northern Olympic Peninsula the Lake Creek–Boundary Creek and Crescent faults probably imbricate additional thrust sheets of the postaccretion forearc sedimentary basin below the LEF, but this structural complexity is not shown due to the uncertain position of the faults. All of the ages for the units in northern Siletzia are based on the U-Pb geochronology presented in this work. The temporal history of magmatism and sedimentation in central and southern Siletzia is from Wells et al. (2014). Colors and patterns are the same as in Figures 2 and 4. Abbreviations: BHB—Black Hills basalts, BC—Bremerton complex, HRF—Hurricane Ridge fault, MC—Metchosin complex, Mtn.—mountains, Fm—formation.

Resurrection–Farallon) with North America during the Paleogene (e.g., Atwater, 1970; Engebretson et al., 1985; Stock and Molnar, 1988; Haeussler et al., 2003; Madsen et al., 2006; McCrory and Wilson, 2013). While the symmetric age distribution of Duncan (1982) has been revised with new geochronologic data (Wells et al., 2014; this study), there are many reasons to think that Siletzia formed as a ridge-centered oceanic plateau. First, the southern portion of the terrane is older than the central and northern portions, and this age progression remains compatible with the presence of a ridge near present-day Washington (e.g., McCrory and Wilson, 2013; Wells et al., 2014). Second, the Metchosin and Bremerton complexes appear to have been generated at a spreading center between 51 and 50 Ma and are inboard of older basaltic basement in the Crescent Formation (Figs. 2 and 4). This spatial relationship implies that a spreading center lay between the Crescent Formation

and North America prior to Siletzia accretion (Fig. 6). Third, Eddy et al. (2016) documented regional initiation, or acceleration, of dextral strike-slip faulting in Washington after 50 Ma. Such a transition is consistent with a southward jump of the Kula–Farallon–North America, or Resurrection–Farallon–North America triple junction immediately following Siletzia accretion, since plate reconstructions for the Paleogene North Pacific basin consistently predict that the Kula and/or Resurrection plates had a stronger dextral oblique component of plate motion relative to North America than the Farallon plate (Atwater, 1970; Engebretson et al., 1985; Stock and Molnar, 1988; Madsen et al., 2006; McCrory and Wilson, 2013). The presence of near-trench magmatism along Vancouver Island (Groome et al., 2003; Madsen et al., 2006) and in western Washington (Cowan, 2003) from 52 to 49 Ma is also consistent with a triple junction along this part of the margin prior to Siletzia

accretion, since near-trench magmatism is considered one of the most diagnostic features of ridge-trench interaction and is a manifestation of the formation of a slab window (e.g., Thorkelson, 1996). Further inboard, geochemically diverse magmatism in eastern Washington and British Columbia (Breitsprecher et al., 2003), including adakites (Ickert et al., 2009), is compatible with the presence of a slab window in this area ca. 50 Ma. However, tomographic images of a hanging slab of Farallon oceanic crust under eastern Washington suggest that slab breakoff accompanied the accretion of Siletzia (Schmandt and Humphreys, 2011), and this process could have produced similar magma compositions to those documented in Breitsprecher et al. (2003) and Ickert et al. (2009). Nevertheless, there is compelling evidence for the presence of a triple junction near the latitude of southern British Columbia and Washington at the time of Siletzia accretion. One consequence of this ridge position is that Siletzia was likely emplaced on very young oceanic crust, as evidenced by the 51 and 50 Ma dates for the Metchosin and Bremerton complexes. Such a young age for both the plateau and its oceanic basement may explain why Siletzia jammed the subduction zone rather than subducting easily, because only oceanic plateaus that are broad, thick, and young can exert a buoyancy force that equals or exceeds slab pull (e.g., Cloos, 1993; Arrial and Billen, 2013).

Our revised volcanic stratigraphy for northern Siletzia is strikingly similar to the stratigraphy of the rest of the terrane (Fig. 5), and we discuss the origin of Siletzia within the tectonic framework of a ridge-centered oceanic plateau (Fig. 6) as summarized by Wells et al. (2014). Construction of the plateau started by 56 Ma in the south (Wells et al., 2014) and 53.18 ± 0.17 Ma in the north, and was complete shortly after 48.364 ± 0.036 Ma, when deposition of a regional forearc sedimentary basin began. The volume of basaltic magma emplaced during this period is comparable to those seen in large igneous provinces (e.g., Trehu et al., 1994; Wells et al., 2014) and Siletzia may have developed above the Yellowstone hotspot. Pyle et al. (2009) suggested that Siletzia represents initial impingement of the Yellowstone hotspot on the Earth's surface. However, the ~8–5 m.y. duration of volcanism within Siletzia is much longer than the short pulses (<1 m.y.) usually attributed to impingement of a plume head on oceanic or continental lithosphere (e.g., Richards et al., 1989), and this long duration may suggest that Siletzia is a manifestation of a much longer lived Yellowstone hotspot (e.g., Johnston et al., 1996; Murphy et al., 2003; Murphy, 2016).

The timing of the accretion of Siletzia is constrained throughout the terrane by dates of

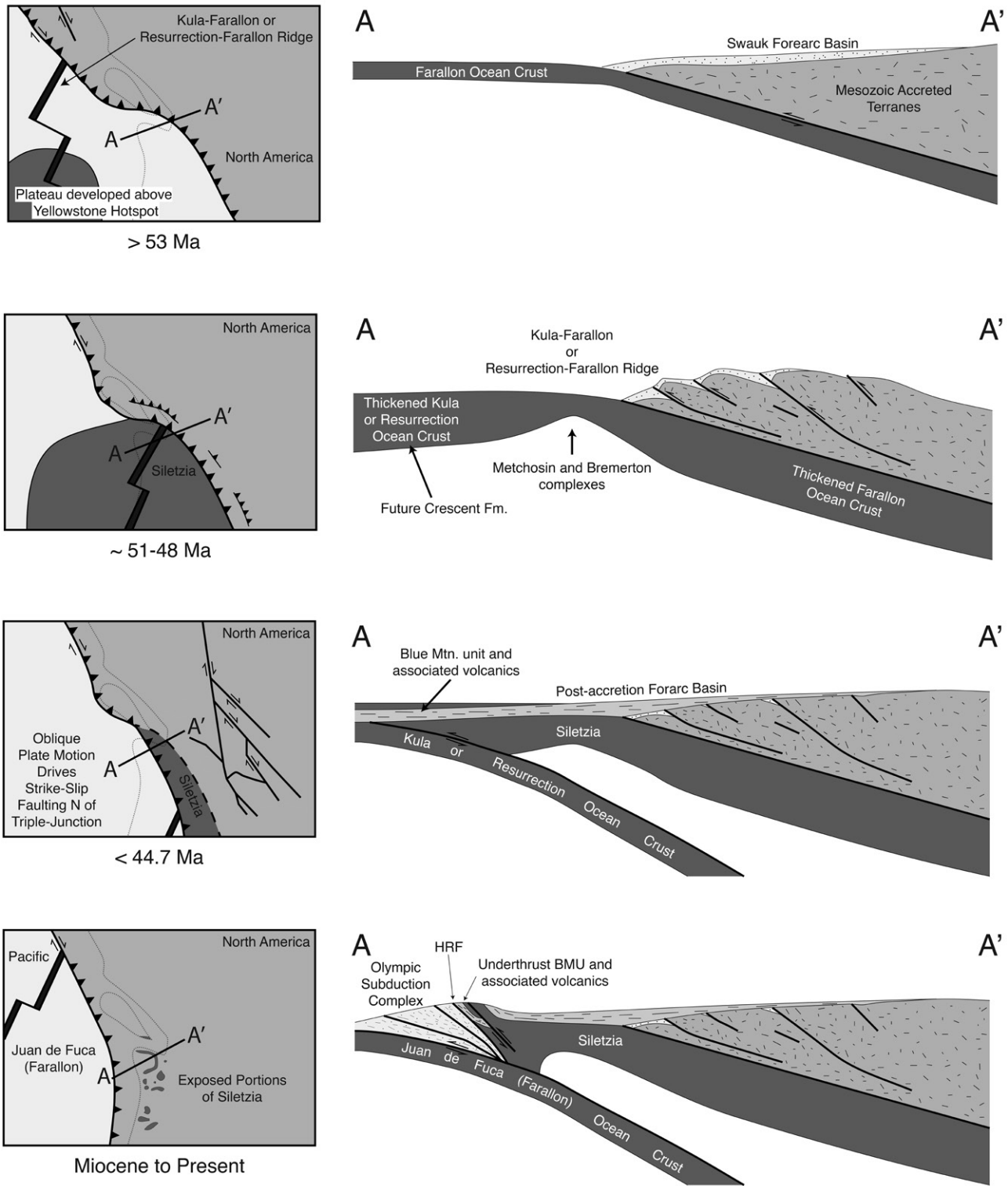


Figure 6. Proposed history of the construction and accretion of Siletzia following Wells et al. (2014). Ocean crust of normal thickness belonging to the Farallon plate was subducting at the latitude of Washington prior to 53 Ma. From 53 to 48 Ma an oceanic plateau developed along the Kula-Farallon, or Resurrection-Farallon, oceanic spreading center. The attempted subduction of this young oceanic plateau (ca. 51–48 Ma) jammed the subduction zone, led to regional shortening, and a jump of the Kula–Farallon–North America, or Resurrection–Farallon–North America triple junction to the south. Oblique motion between the Kula, or Resurrection, and North American plates drove dextral strike-slip faulting along the North American margin during that time. After 44.72 Ma the Blue Mountain unit (BMU) was deposited as a distal part of a regional depositional system. Basalts that are interbedded with and conformably overlie these sediments may represent continued interaction between the Yellowstone hotspot and the North American margin, initiation of subduction following the accretion of Siletzia, or a second period of ridge-trench interaction. After 44.72 Ma, the BMU and associated volcanics were thrust under the rest of Siletzia. Maps are modified from Eddy et al. (2016) and based on the model of Wells et al. (2014). HRF—Hurricane Ridge fault.

deformed and nondeformed rocks. In the north, accretion is bracketed between 51.309 ± 0.024 and 49.933 ± 0.059 Ma by shortening in the nonmarine Swauk forearc sedimentary basin in Washington (Eddy et al., 2016). In central Siletzia, it is constrained between 49.0 ± 0.8 Ma and ca. 48 Ma by ages that bracket deformation within the upper part of the basaltic basement of Siletzia (Wells et al., 2014). In southern Siletzia, it is constrained between ca. 53 Ma and ca. 50–48 Ma by the youngest basalts involved in the fold-thrust belt exposed near Roseburg, Oregon, and the age of overlying nondeformed forearc sedimentary rocks (Dumitru et al., 2013; Wells et al., 2014). The close similarity in the timing of deformation along the length of the terrane suggests that buoyant oceanic crust entered the subduction zone simultaneously over a wide area. The tight brackets on deformation also imply that regional shortening was very brief and that accretion occurred while construction of the plateau was ongoing. The transition to subaerial and shallow-marine eruptions in northern and central Siletzia is coeval with accretion, indicating that it may be a result of tectonic uplift rather than the emergence of a volcanic edifice above sea level due to thermal buoyancy, as suggested by Murphy et al. (2003) and Wells et al. (2014).

Duncan (1982) and McCrory and Wilson (2013) suggested that Siletzia represents a composite terrane composed of a captured piece of the Farallon plate in the south and a captured piece of the Kula or Resurrection plate to the north. These interpretations were based on a systematic younging trend in available age constraints that converged near the Black Hills. New geochronologic data have complicated this age trend by recognizing that magmatism was relatively long lived throughout the terrane (Wells et al., 2014; this study). Thus, the reported age, or age range, of different outcrop areas within Siletzia is likely dependent on the level of exposure and is not necessarily representative of the age of the underlying oceanic crust. Nevertheless, there is some evidence from the Metchosin and Bremerton complexes that the northern part of the terrane represents a captured piece of the Kula or Resurrection plate. Both complexes preserve sheeted dikes with similar north-northeast–south-southwest orientations (Fig. 4) that we consider to record the primary extensional direction, because paleomagnetic data from the Metchosin complex demonstrate that it has undergone little to no postaccretion vertical-axis rotation (Irving and Massey, 1990). The implied spreading direction from these dike orientations is west-northwest–east-southeast, and the difference in age between the ca. 51 Ma Metchosin complex and ca. 50 Ma Bremerton complex would indicate that the spreading center that

generated these complexes was near Bremerton at the time of accretion. If so, then most of northern Siletzia west and north of Bremerton represents part of the oceanic plateau that developed on the Kula or Resurrection plates, while central and southern Siletzia represents parts of the oceanic plateau developed on the Farallon plate. The Yakutat terrane in southern Alaska is of similar age and has a similar stratigraphy to Siletzia, and several researchers have proposed that it represents a displaced fragment of the Siletzia oceanic plateau (e.g., Davis and Plafker, 1986; Wells et al., 2014). If so, it likely represents an additional fragment of the plateau that developed on the Kula or Resurrection plate.

Following Siletzia accretion, the terrane subsided below sea level and a new forearc sedimentary basin was established (Figs. 5 and 6). Given that our maximum depositional ages for the Blue Mountain unit are comparable to the oldest sedimentary rocks in this regional forearc basin, we envision deposition of the unit as a distal portion of this sedimentary system prior to its incorporation in the Olympic subduction complex. The basalts that overlie and are interbedded with the Blue Mountain unit are similar in age to the volcanic centers interbedded with the regional forearc basin in southwestern Washington and western Oregon (Fig. 5) and likely represent a manifestation of the same process. Wells et al. (2014) attributed these volcanic centers to continued interaction between the North American margin and the Yellowstone hotspot, and the basalts associated with the Blue Mountain unit may be the northernmost documented expression of this postaccretion forearc magmatism. Alternatively, they may be related to the formation of a new subduction zone following Siletzia accretion or the northward migration of the Kula–Farallon–North America triple junction, or the Resurrection–Farallon–North America triple junction, after the middle Eocene, as required by our model (Fig. 6).

CONCLUSIONS

Our U-Pb zircon dates from throughout northern Siletzia place new constraints on the terrane's volcanic stratigraphy. They show that the basaltic basement in this area was constructed between 53.18 ± 0.17 Ma and 48.364 ± 0.036 Ma, similar to the rest of the terrane. Furthermore, we show that the continentally derived turbidites of the Blue Mountain unit are younger than the basaltic basement of Siletzia and must have been thrust under the terrane after 44.72 ± 0.21 Ma. These rocks have long been considered to floor the terrane and have been used to argue for emplacement of Siletzia along the continental margin. However, our revised age for

the Blue Mountain unit no longer necessitates such an interpretation. Instead, we interpret the stratigraphy of northern Siletzia to be compatible with its origin as an accreted oceanic plateau possibly developed above a long-lived Yellowstone hotspot.

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