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Eocene initiation of the Cascadia subduction zone: A second example of plume-induced subduction initiation?

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ABSTRACT

The existing paradigm for the major ca. 56–48 Ma subduction zone reorganization in the Pacific Northwest of North America is that: (1) the Siletzia large igneous province erupted offshore to the west of North America, forming an oceanic plateau; (2) Siletzia then collided with North America, clogging the Pacific Northwest segment of the Cordilleran subduction zone; and (3) the oceanic lithosphere west of Siletzia then ruptured to initiate the new Cascadia subduction zone. Oceanic lithosphere is strong and difficult to rupture, so this would represent a rare example of such a rupture initiating a new subduction zone. This paper explores an alternative hypothesis for the reorganization, a plume-induced subduction initiation (PISI) mechanism, which has previously been applied to the initiation of Caribbean plate subduction zones in the Cretaceous. In this PISI hypothesis, a newly formed, ~1200-km-diameter Yellowstone mantle plume head rose at ca. 55 Ma beneath western North America, generating Siletzia in situ on the North American margin, as well as generating the ~1700-km-long Challis-Kamloops volcanic belt ~600 km to the east of Siletzia. This destroyed the existing Cordilleran subduction zone and allowed the new Cascadia subduction zone to form by collapse of thermally weakened oceanic lithosphere over the hot western margin of the plume head. This PISI hypothesis provides an integrated framework for understanding Siletzia, the Challis-Kamloops belt, Eocene core complexes from Idaho (U.S.) to British Columbia (Canada), underplated mafic rocks beneath Oregon and Washington (U.S.), post–17 Ma manifestations of the Yellowstone plume, and geophysical characteristics of the lithosphere beneath the Pacific Northwest.

INTRODUCTION

About one-third of all modern subduction zones formed during Cenozoic time (Gurnis et al., 2004), a time period for which plate motions are well understood and geologic evidence is well preserved. This observation implies that subduction initiation must be a reasonably easily started and frequent plate-tectonic process. This is an extremely exciting constraint, but it does not tell us how new subduction zones form. For that information, we must study individual examples and use these to constrain numerical models. Here, we focus on how the new Cascadia subduction zone in the Pacific Northwest of North America (United States–Canada) formed at ca. 55 Ma.

There are two principal ways that a new subduction zone may form: it may be induced or it may form spontaneously (Fig. 1). Induced nucleation of a subduction zone (INSZ) is expected if plate convergence continues after a collision event arrests convergence in a preexisting subduction zone. There are two varieties of INSZ, polarity reversal and transference. Polarity reversal INSZ (Fig. 1B) occurs when a subduction zone dipping beneath an arc is clogged by buoyant crust, and a new subduction zone dipping in the reverse direction forms on the opposite side of the arc. Transference INSZ (Fig. 1A) occurs when a new subduction zone (with unchanged polarity) forms within oceanic lithosphere outboard of the collision zone. There are Cenozoic examples of polarity reversal INSZ (e.g., accompanying the Miocene collision of the Ontong Java Plateau with the Melanesian arc; Holm et al., 2013); polarity reversal INSZ occurs readily because lithosphere around a hot arc is weak and easy to rupture. In contrast, there are no incontrovertible Cenozoic examples of transference INSZ. We expect that a new subduction zone would have formed by transference INSZ in the Indian Ocean south of the India-Asia collision zone, but this has not happened in the 50 m.y. since this collision began. This indicates that transference subduction initiation does not generally occur because old oceanic lithosphere is too strong to rupture under compression. Support for this interpretation comes from numerical models for the collision of arcs, plateaus, and terranes with continents; these do not reproduce INSZ by transference. Instead, these models predict that only the crustal portion of the arc, plateau, or terrane would accrete to the continental margin, whereas the underlying lithospheric mantle would continue to subduct (e.g., Tetreault and Buiter, 2012; Vogt and Gerya, 2014). Thus, although the trench would jump seaward, the subducting slab would remain intact, and so this does not represent the creation of a new subduction zone.

Spontaneous nucleation of a subduction zone (SNSZ) involves collapse along a lithospheric weakness juxtaposing old, dense oceanic lithosphere with buoyant oceanic or continental lithosphere (Stern and Gerya, 2018). There are three modes of SNSZ: passive margin collapse, transform collapse, and plume head margin collapse. Passive margin collapse (Fig. 1C) is expected as part of the Wilson cycle (Wilson, 1966), but there are no Cenozoic examples, again suggesting that the strength of old oceanic lithosphere prevents it from rupturing, thus impeding subduction initiation (Baes and Sobolev, 2017). Transform collapse (Fig. 1D) occurs when old, dense oceanic lithosphere is

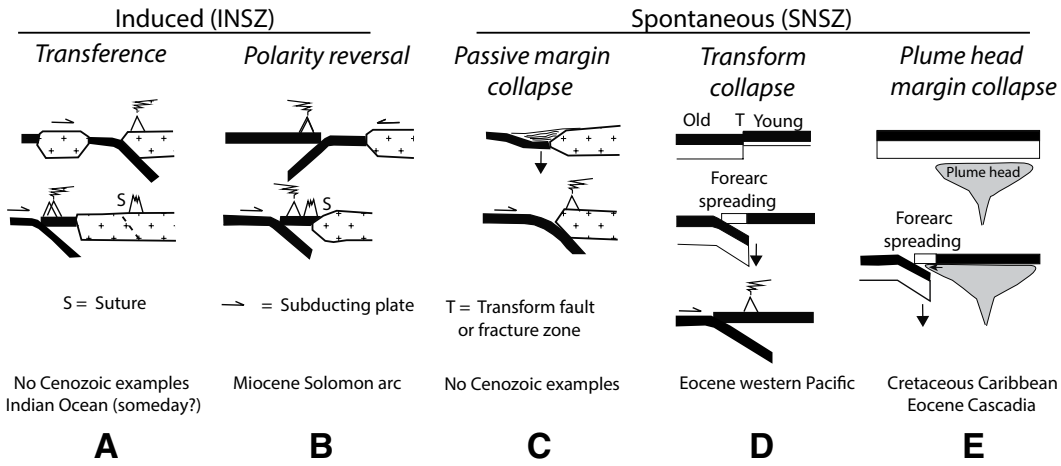

Figure 1. General classes, subclasses, and examples of how subduction zones form, modified after Stern (2004) by addition of "Plume head margin collapse" based on research of Whattam and Stern (2015) and Gerya et al. (2015).

juxtaposed with young buoyant oceanic (or continental) lithosphere across a sufficiently long (at least ~1000 km) weak zone that cuts the entire lithosphere; this commonly occurs along a fracture zone (Gerya et al., 2008). The 2800-km-long Izu-Bonin-Mariana convergent margin in the western Pacific is a good example of this subduction initiation mode (Brandl et al., 2017).

The last SNSZ mode—plume head margin collapse (Fig. 1E)—involves a large, hot mantle plume that weakens oceanic lithosphere over a large enough area that the adjacent older and denser oceanic lithosphere collapses to form a new subduction zone. Plume-induced subduction initiation (PISI) is proposed to explain the formation of Late Cretaceous convergent margins on the southern and western flanks of the Caribbean plate, and it is clearly important to search for other potential unrecognized examples. This motivates us to see if a plume-induced model is viable for the origin of the Cascadia convergent margin in Eocene time, but first we outline the PISI model in greater detail.

The Plume-Induced Subduction Initiation (PISI) Model

There is a great diversity of models for mantle plumes, and here we summarize some basic points from the plume models of Campbell (2007). A new, rising plume consists of a large head ~1000 km in diameter followed by a narrower tail 100–200 km in diameter. When the head encounters the base of the lithosphere, calculations suggest it flattens and spreads out over a diameter of 2000–2500 km. Emplacement of the hot head is expressed at the Earth's surface by ~1000 m of thermally driven uplift, voluminous basaltic eruptions, and lithospheric extension. New plume heads are a popular mechanism invoked to explain the formation of many basaltic large igneous provinces (LIPs). LIPs are defined as having areas >100,000 km^2 and igneous volumes >100,000 km^3, and typically form during intense igneous pulse(s) of notably brief duration (<1–5 m.y.) during which most of the total igneous volume is emplaced (Bryan and Ernst, 2008; Ernst, 2014). This is generally compatible with short emplacement durations calculated in plume models. The initial LIP phase is generally followed by continued eruptions of much smaller volumes, which represent the plume tail and which may persists for 100 m.y. or more, creating hotspot tracks across the overlying plate (Richards et al., 1989).

Since 180 Ma, new basaltic LIPs have formed across the Earth at a rate of about one every 10 m.y. (Ernst, 2014, his section 1.4). Given that roughly 10% of the Earth's surface lies within 1000 km of a subduction zone, we might thus expect a new active plume head to rise up beneath a preexisting active subduction zone roughly once every 100 m.y. (cf. Fletcher and Wyman, 2015).

It is increasingly recognized that a sufficiently hot and large plume head can rupture the lithosphere to cause subduction initiation. The idea was first suggested by Ueda et al. (2008) on the basis of two-dimensional (2-D) numerical experiments. Independent 2-D numerical experiments by Burov and Cloetingh (2010) confirmed that plume-induced subduction initiation (PISI) could result if a sufficiently buoyant plume head was emplaced beneath rheologically stratified crust and lithospheric mantle. These models show that the plume head flexes the lithosphere upward, exerts basal shears on it, and imposes local extensional stresses, while at the same time heating it. The plume head erodes the mantle lithosphere and intrudes between buoyant crust and dense mantle lithosphere, allowing the lithosphere to sink and in some cases form a self-sustaining subduction zone. Burov and Cloetingh (2010) concluded that a large plume head was required; small plumes cannot cause PISI.

These ideas were intriguing, but the absence of plausible natural examples impeded serious consideration. This changed when Whattam and Stern (2015) made a strong argument that PISI happened in Late Cretaceous time to form

subduction zones around the southern and western margins of the Caribbean plate. These formed in response to emplacement of an unusually large, hot plume head which was responsible for the ca. 100 Ma Caribbean LIP. Three key observations led to this conclusion: (1) trace element chemistry of most units 100 Ma and younger interpreted as Caribbean LIP that are exposed along the southern margin of the Caribbean plate and northwestern South America record subduction additions, which increased with time beginning at 100 Ma; (2) there is no known hiatus between Caribbean LIP and younger arc lava sequences, suggesting continuous tectonomagmatic evolution from plume to arc; and (3) generation of the Caribbean LIP and the earliest overlying arc units overlap in time, space, and chemical and isotopic compositions, with both sequences derived from Galapagos plume–like mantle, which became increasingly subduction-modified with time. Whattam and Stern (2015) noted that the scale of the Caribbean PISI event was consistent with the expected large scale of the process, affecting the plume margins along two great arms some 1400 km long from southern Costa Rica–Panama to western Colombia and 1700 km long from Ecuador to the Leeward Antilles.

The PISI model was tested by three-dimensional numerical thermomechanical modeling by Gerya et al. (2015) to demonstrate that three key physical factors combined to trigger self-sustained subduction: (1) strong and negatively buoyant (i.e., sufficiently old) oceanic lithosphere; (2) focused magmatic weakening and thinning of lithosphere above the plume; and (3) lubrication of the slab interface by hydrated crust. The numerical experiments confirmed that arrival of a sufficiently large, hot mantle plume head could weaken strong and dense oceanic lithosphere to produce new self-sustaining subduction zones around the plume head, always dipping beneath the plume head, as seen for the present Caribbean.

Existing Paradigm for Formation of the Modern Cascadia Subduction Zone

Ever since the idea was first proposed by Duncan (1982), the Cascadia subduction zone has generally been thought to have formed by transference INSZ in Eocene time, although it is not usually described this way. The dominant model holds that a large oceanic plateau—now preserved as Siletzia—formed a short distance offshore of North America. Soon afterwards, Siletzia entered the preexisting, east-dipping Cordilleran subduction zone and caused it to fail, accreting to North America and causing the new east-dipping Cascadia subduction zone to form outboard of Siletzia (Duncan, 1982; Wells et al., 2014). In a recent study, Phillips et al. (2017, p. 301) concluded that "…the evidence for the Siletz terrane representing an accreted oceanic plateau linked to a mantle plume is compelling." If these interpretations are correct, then transference INSZ is a more important subduction initiation mechanism than heretofore thought. A variant of this model is that only the crustal portion of Siletzia accreted to the North America, whereas the underlying lithospheric mantle portion of the slab remained intact and continued to subduct, as outlined previously. This seems unlikely, as both a new trench and a new magmatic arc apparently formed several hundred kilometers to the west of the previous trench and arc. We note that the model of Tetreault and Buiter (2012) and Vogt and Gerya (2014), discussed above, could apply to emplacement of Siletzia, but this would not explain the other related phenomena discussed below.

This paper focuses on exploring the plausibility of a third model, a PISI model, in which a large (~1200 km diameter) plume head rose partly under the subducting Farallon oceanic plate and partly under North America itself. This disrupted the old Cordilleran subduction zone and caused lithospheric collapse of the Farallon plate along the western margin of the plume head to form the modern Cascadia subduction zone.

SILETZIA AND FORMATION OF THE MODERN CASCADIA SUBDUCTION ZONE

The convergent margin of western North America was established in Triassic time, and the Late Cretaceous magmatic arc can be traced along a path from the Sierra Nevada batholith of California through northwestern Nevada to the Idaho batholith (United States), and then to the Coast Plutonic Complex of British Columbia (Canada) (e.g., Dickinson, 2004; Fig. 2). Much of this margin is associated with an important isotopic boundary—the "0.706 line" (Fig. 2)—separating Mesozoic and Cenozoic igneous rocks with initial $^{87}Sr/^{86}Sr$ <0.706 to the west from those with initial $^{87}Sr/^{86}Sr$ >0.706 to the east. The 0.706 line separates juvenile crust (ensimatic or oceanic) in the west from ancient continental crust of the Wyoming "craton" to the east (craton is in quotes because the ancient rocks of this region have been strongly affected by especially Cenozoic tectonic and magmatic activity). The region between the Olympic-Wallowa lineament (Washington, NE Oregon) and the Klamath Mountains–Blue Mountains (NW California and E Oregon) is sometimes called the Columbia embayment (C.E. in Fig. 2). A major tectonic reorganization occurred in Paleogene time, whereby the older, contorted convergent margin was abandoned and the new, straighter Cascadia convergent margin was established to the west of the Ancestral Cascades volcanic arc (Armstrong and Ward, 1991; Madsen et al., 2006; du Bray and John, 2011). This study addresses why the old convergent margin shut down and how the new one formed.

Formation of the Cascadia subduction zone was heralded by emplacement of Siletzia, a huge mafic volcanic construction exposed in the Coast Ranges of Oregon and Washington (United States) that comprises the modern Cascadia forearc (e.g., Wells et al., 2014; Fig. 2). We note that forearcs preserve evidence for how the associated subduction zone formed (Stern and Gerya, 2018). Siletzia is mostly basaltic, with lower units of tholeiitic basalt and upper units of alkali basalt, a sequence that is consistent with a mantle plume, rift, or both (Wells et al., 2014). Siletzia is dominated by pillow lavas, massive flows, and intrusive sheets. Thin (~10 km thick) crust, sheeted dikes, and transitional chemistry are found in the north; thicker crust with ocean-island basalt (OIB) chemistry is found in the south (Wells et al., 2014). Siletzia has a preserved area of ~100,000

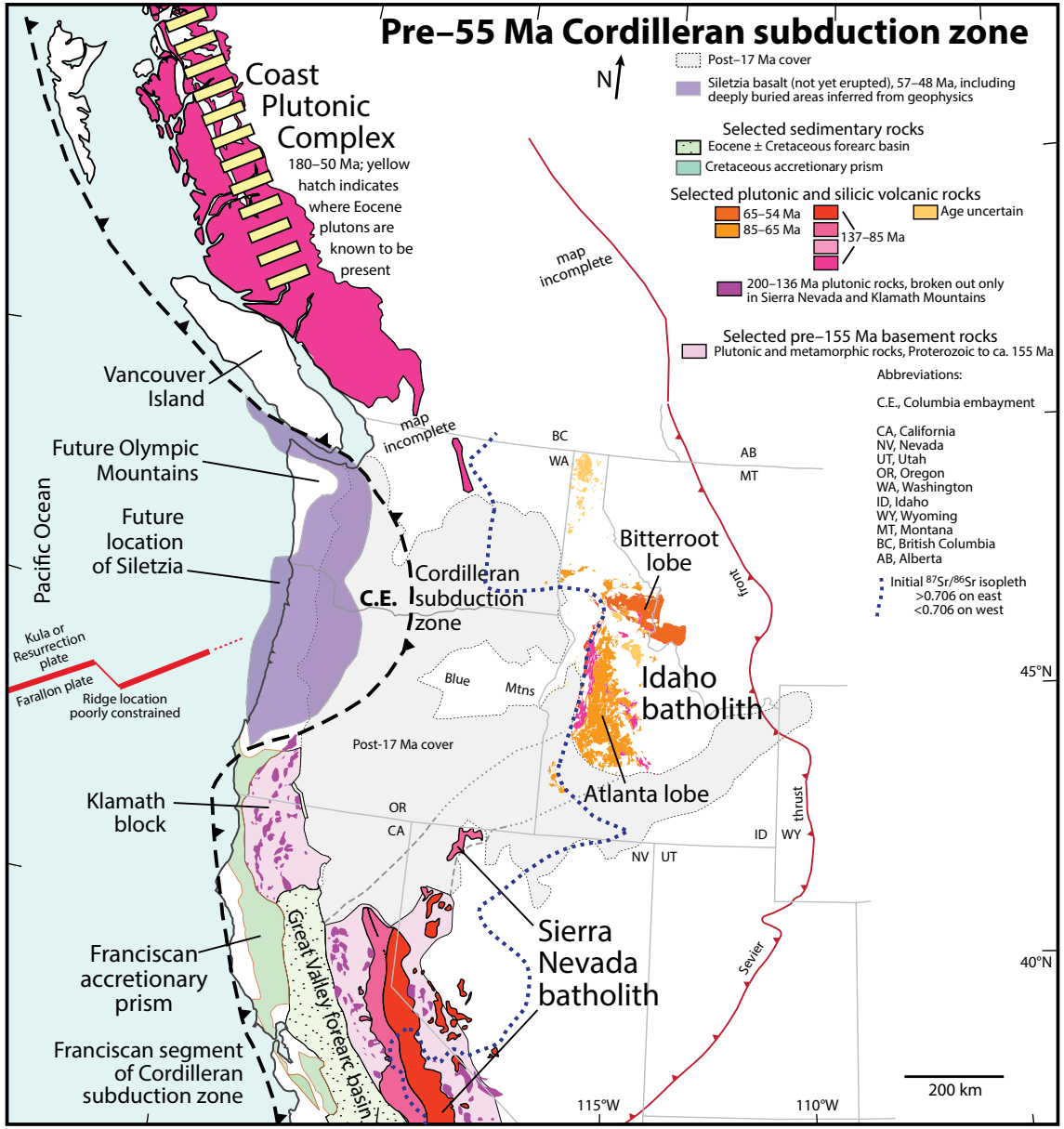

Figure 2. Map of selected tectonic elements of the Cordilleran convergent margin between central California (United States) and British Columbia (Canada) at ca. 55 Ma, shortly before the old Pacific Northwest segment of the Cordilleran subduction zone was destroyed. We have not restored post–55 Ma displacements of various blocks, including vertical-axis rotations (e.g., Wells et al., 2014, p. 707–708), Basin and Range extension, and strike-slip translations. See Table 1 for sources.

TABLE 1. SOURCES OF DATA FOR MAPS IN FIGURES 2, 3, 5, AND 10

Base map, continent-scale geology, map projection	Chiefly Reed (2004); Gehrels et al. (2009) for Coast Plutonic Complex and coastline of British Columbia; Wells et al. (2014, their figure 1) for backarc volcanic rocks in British Columbia
Post–17 Ma cover	Reed (2004)
Coast Plutonic Complex	Gehrels et al. (2009)
Idaho batholith, Challis volcanic complex, plutonic rocks in northern Idaho, Belt Supergroup basin	Gaschnig et al. (2010, 2017); Chetel et al. (2011); Aleinikoff et al. (2012); MacLean and Sears (2016); Digital Geology of Idaho (undated, http://geology.isu.edu/Digital_Geology_Idaho, accessed August 2015)
Klamath block	Irwin and Wooden (1999); Allen and Barnes (2006); Sharman et al. (2015)
Sierra Nevada	Irwin and Wooden (2001); Sharman et al. (2015)
Northwestern Nevada	Lerch et al. (2007)
Franciscan Complex and Great Valley forearc basin	Jennings (1977); Redwine (1984); California Division of Mines and Geology (2000); Dumitru et al. (2010, 2013, 2016, 2018)
Sevier thrust front	Dickinson et al. (1988); Reed (2004)
0.706 isopleth of initial $^{87}Sr/^{86}Sr$ ratio	North to south: Carlson et al. (1991); Bedrosian and Feucht (2014); Gaschnig et al. (2011); Kistler (1990); Tosdal et al. (2000); Chapman et al. (2012)
Siletzia, Yakutat terrane	Wells et al. (2014)
Farallon–Kula/Resurrection spreading ridge	McCrory and Wilson (2013); Wells et al. (2014)
Tyee forearc basin	Heller et al. (1985, 1992); Wells et al. (2000, 2014); Dumitru et al. (2013, 2015).
Tyee River, Princeton River	Heller et al. (1985); Dumitru et al. (2013, 2015, 2016)
Olympic subduction complex	Brandon and Calderwood (1990); Stewart et al. (2003)
Idaho River	Chetel et al. (2011)
Eocene core complexes	Foster et al. (2007); Parrish et al., (1988); in northeastern Washington and southern British Columbia, note that relative locations between core complexes and Eocene volcanic rocks may be inaccurate due to difficulties in registering source maps
Laramide basins (Uinta, etc.)	Dickinson et al. (1988, 2012); Smith and Carroll (2015)
Eocene volcanic rocks in backarc region	Haeussler et al. (2003); Wells et al. (2014); Ootsa: Bordet et al. (2014); Kamloops: Dostal et al. (2003); Absaroka: Feeley et al. (2002), Hiza (1999); Bearpaw: MacDonald et al. (1992).
Cowichan fold-and-thrust belt	England and Calon (1991); Johnston and Acton (2003)
Swauk basin	Eddy et al. (2016b, 2017)
Umpqua fold-and-thrust belt	Wells et al. (2014)
Hornbrook strata	Surpless (2015)
Ancestral Cascades arc	du Bray and John (2011); du Bray et al. (2014); Mullen et al. (2018)
Pacific-Farallon ridge	Wells et al. (2014)
Columbia River Basalt Group	Reidel et al. (2013)
Snake River Plain–Yellowstone	Pierce and Morgan (1992, 2009); Dickinson (2004, 2013)
Eocene volcanic rocks in the forearc region	Breitsprecher et al. (2003); Madsen et al. (2006); Ickert et al. (2009); Kant et al. (2018)

km^2 and an estimated volume of 1,700,000–2,600,000 km^3, qualifying it as a LIP (Wells et al., 2014). Documented volcanism occurred from 53.1 to 48.4 Ma near the northern end of Siletzia (Eddy et al., 2017) and from 56 to 52 Ma near the southern end (Wells et al., 2014, their figure 9). Deeper, unexposed parts of Siletzia are presumably mostly older, although given that LIPs may form very rapidly, they may be only slightly older and may also contain substantial volumes of post–56 Ma sills, intrusions, and/or underplated rocks. Trace elements and radiogenic isotopes implicate a slightly enriched mantle source and show little evidence that these magmas interacted with continental crust (Phillips et al., 2017). Phillips et al. (2017) compared Siletzia magmas to those responsible for producing the Ontong Java and Caribbean LIPs. They further concluded that Siletzia is evidence for a mantle plume, possibly the Yellowstone hotspot.

The exposed Paleogene igneous rocks of Siletzia and their overlying sediments define a reasonably continuous succession. There are local unconformities, especially in the late Eocene, but no major unconformities or hiatuses lasting >1 m.y. are known (Wells et al., 2014, their figure 2). The oldest lavas of the Ancestral Cascades arc are tholeiitic basalt (du Bray and John, 2011). Like the Caribbean PISI sequence (Whattam and Stern, 2015), the record from clearly OIB-like (50–56 Ma) alkali basalt of Siletzia to Ancestral Cascades arc (<40 Ma) tholeiitic basalt shows no obvious breaks that could be ascribed to a collision or other major tectonic event.

There have been many explanations for voluminous Siletzia volcanism, including: (1) an accreted oceanic plateau (Dickinson, 1976; Duncan 1982; Wells et al., 1984; Murphy et al., 2003); (2) continental-margin rifting perhaps due to subduction of the Kula-Farallon spreading ridge (Babcock et al., 1992; Haeussler et al., 2003; Brandon et al., 2014); (3) a tear or slab window in the subducting plate (Madsen et al., 2006; Humphreys, 2008); (4) beginning of the Yellowstone mantle plume (Murphy et al., 2003); and (5) beginning of the Yellowstone plume

rising beneath the Kula-Farallon spreading ridge (Duncan, 1982; McCrory and Wilson, 2013; Wells et al., 2014). Several of these hypotheses can be challenged. For the accreted oceanic plateau model, the existence of a large oceanic plateau in the northeastern Pacific is surprising because such features are unknown from this part of the Pacific Ocean. They are common on Cretaceous seafloor preserved in the western Pacific and were surely abundant on Cretaceous seafloor that was subducted beneath western North America, but oceanic plateaus are rarely constructed on Cenozoic seafloor. Only long-lived hotspots like Hawaii and Tahiti-Marquesas have built significant lava piles on Cenozoic Pacific seafloor, and none of these are oceanic plateaus that approach the scale of Siletzia. It is also surprising that such a uniquely young and large oceanic plateau was fortuitously emplaced so close to the continent that it collided almost immediately after formation. Furthermore, transference would have had to have been the subduction initiation mechanism for this scenario. If such a collision was responsible for forming the modern Cascadia subduction zone, it would be a rare Cenozoic example of transference INSZ (Fig. 1A; Stern and Gerya, 2018). Interpretation as a slab tear is difficult to reconcile with the huge volumes of basalt erupted. Modern examples of igneous activity above slab tears or gaps produce much smaller volumes of melt; for example, no LIPs are associated with slab tears (R.E. Ernst, 2018, personal commun.; see also Ernst, 2014). Also, a separate slab tear is argued by some (e.g., Breitsprecher et al., 2003) to be responsible for the NW-SE–trending belt of Paleogene core complexes that can be traced from central Idaho into British Columbia (Fig. 3), requiring two slab tears at the same time. Continental-margin rifting due to subduction of the Kula-Farallon ridge also seems unlikely because of the huge volumes of magma produced. The huge volume of magma erupted is most consistent with eruption over a plume, as most recently concluded by Phillips et al. (2017). The position of this LIP in a forearc setting is most consistent with formation during subduction initiation. The key question is whether the oceanic plateau produced by this plume was originally offshore and then tectonically accreted to North America, or was magmatically emplaced in about its present position relative to western North America. The former is the existing paradigm, and the latter is the PISI model considered in greater detail below.

ERUPTIVE, ACCRETIONARY, AND SEDIMENTARY HISTORIES NEAR THE SILEZTIA–NORTH AMERICA "SUTURE"

We need to critically assess whether Siletzia was isolated from or attached to North America while it formed. If Siletzia originated as an oceanic plateau that later collided with North America, then there must have been a trench between them. If so, sediment from North America was likely transported along the trench, with little possibility of deposition on top of the elevated plateau. This situation would have persisted until the trench filled with sediment, which would have taken some time to accomplish (a few million years?), even if the collision produced large volumes of sediment. In this case, conglomerate compositions, sandstone compositions, and detrital zircons ages in sedimentary strata interbedded with and deposited on top of Siletzia lavas should be dissimilar from North American sources. In contrast, if Siletzia formed as part of North America, its sedimentary strata should reflect North American sources.

Figure 4 summarizes stratigraphic relationships in southwestern Oregon, which have been extensively documented by Heller and Ryberg (1983), Heller et al. (1985), Ryu et al. (1996), Wells et al. (2000, 2014), and Santra et al. (2013). In this area, Wells et al. (2000, p. 5) reported that the exposed part of the Siletzia volcanic sequence (56–52 Ma) "contains boulder, cobble, and pebble conglomerate interbeds of chert, limestone, greenstone, plutonic rocks, and metagraywacke derived from [Franciscan and]…Klamath terranes [which currently crop out to the south]. Thick conglomerate interbeds are found in the oldest exposed basalt…and in the upper kilometer of section…Turbidite interbeds and pea gravels derived from Klamath source terranes are found throughout the 4 to 6 km thick volcanic section…and are locally more abundant than pillow basalt…Interbedded basalt and conglomerate derived from the Klamath Mountains requires a volcanic source close to the continental margin during eruption of the exposed 4 to 6 km thick flow sequence." Unfortunately, there are no such data from the underlying ~25 km of unexposed Siletzia rocks, and their age is unclear (see section 2).

In southwestern Oregon, Siletzia volcanic rocks are depositionally overlain by strata of the Umpqua Group then the Tyee Formation (Fig. 4). Sandstones of the Umpqua Group (52–49 Ma) are rich in lithic grains, contain little detrital mica, and were apparently sourced from the nearby Klamath Mountains. In contrast, sandstones of the younger Tyee Formation (49.4–46.5 Ma; 1.5–2 km thick) are arkosic, contain abundant detrital muscovite, and were apparently sourced mainly from the Idaho batholith and adjacent regions (Heller and Ryberg, 1983; Heller et al., 1985, 1992; Ryu et al., 1996; Wells et al., 2000, 2014; Dumitru et al., 2013, 2015, 2016; Santra et al., 2013). Figures 4B and 4C shows detrital zircon U-Pb ages from Tyee and Umpqua sandstones that support the inference that these sediments were sourced mainly from North America. The Umpqua data show a prominent, moderately broad age peak at 150 Ma. Examination of individual ages and uncertainties shows that these zircons span a moderate age range rather than clustering around a single age. Plutons in the northern Klamath Mountains cluster at ca. 162–148 Ma and 139 Ma (Fig. 5; Irwin and Wooden, 1999; Allen and Barnes, 2006) and are an excellent match to the broad Umpqua age peak. The small number of zircons older than ca. 175 Ma were probably sourced from the diverse suites of pre–155 Ma metasedimentary and metavolcanic terranes that underlie most of the Klamath province. About 13% of the zircons show ages of 83–117 Ma, younger than the youngest basement rocks (136 Ma) in the Klamath province. These were probably redeposited from sandstones of the Cretaceous Hornbrook Formation, which once covered parts of the Klamath block (Surpless, 2015). Less likely, they could have been redeposited from Franciscan metasandstones, but there are no Franciscan zircon data within 300 km of the southernmost Umpqua outcrops for comparison (Dumitru et al., 2018).

The Tyee Formation shows a very different detrital zircon age spectrum. Tyee ages match those of source rocks in Idaho, including Challis volcanic rocks,

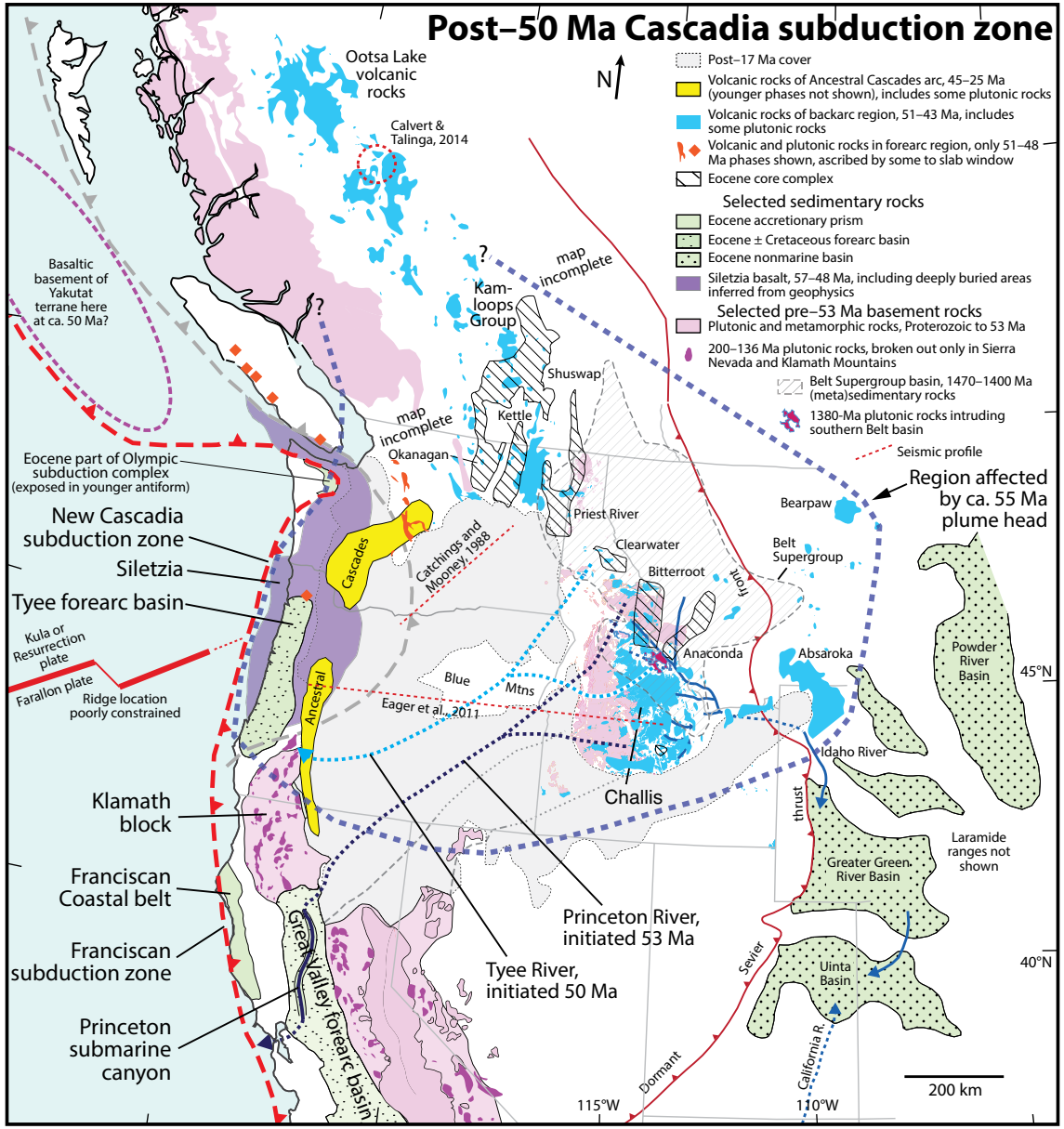

Figure 3. Map of selected tectonic elements shortly after ca. 50 Ma, after the new Cascadia subduction zone formed. We have not restored post–50 Ma displacements. See Table 1 for sources. Blue lines indicate sediment transport pathways in paleorivers and in Princeton submarine canyon; lines are solid where locations are known from paleovalley, submarine canyon, or other data; lines are dashed where specific locations are unknown.

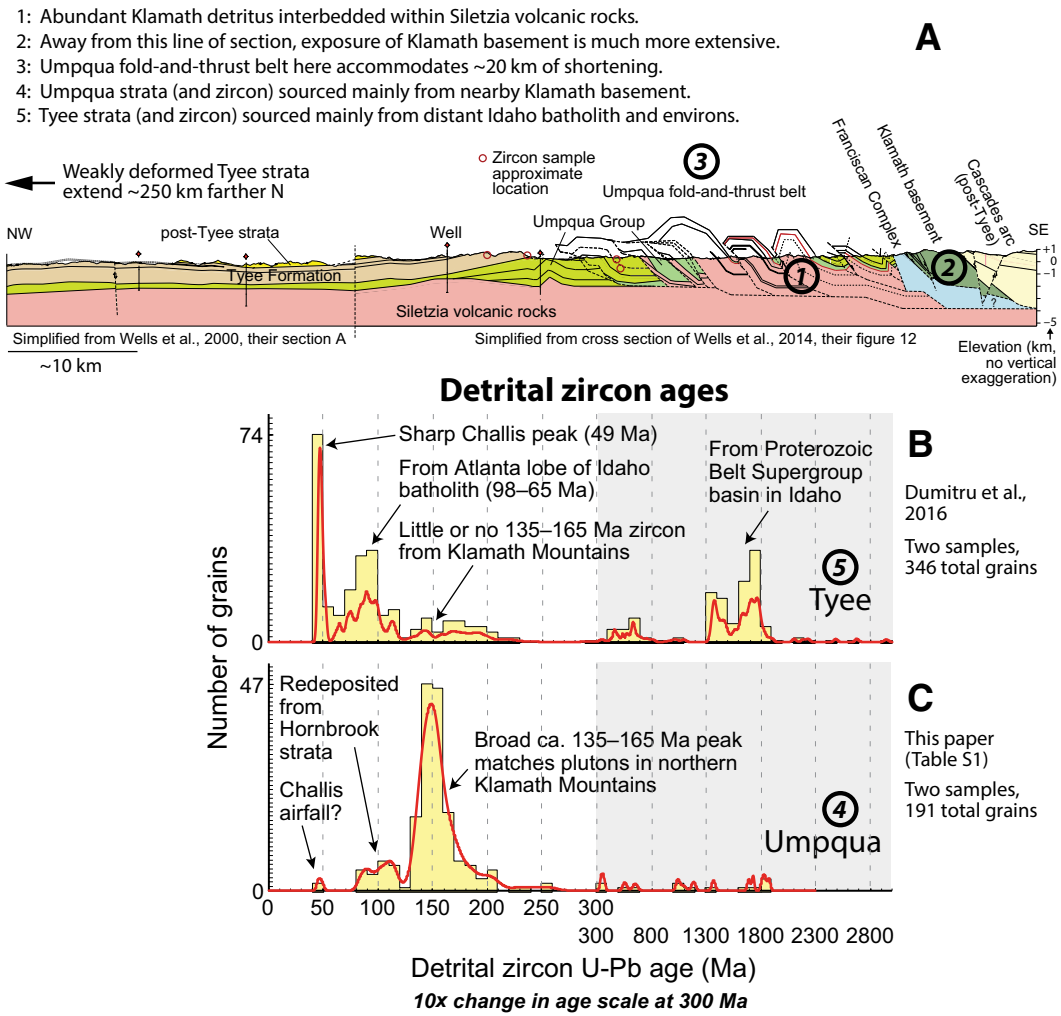

Figure 4. Constraints on the location of Siletzia relative to North America. (A) Cross section of the region of contact between Siletzia and North America near the Klamath Mountains, annotated with structural and stratigraphic constraints. See Figure 5 for location. (B–C) Probability density distributions (red curves) and histograms (yellow fields) of single-grain detrital zircon U-Pb ages from sandstone samples. Note pronounced differences between Umpqua Group and Tyee Formation samples, indicating a major provenance change. See text for discussion. The Umpqua zircon data, additional Franciscan Complex zircon data from Dumitru et al. (2018), and additional interpretive details are available in Table S1 (Supplemental Material [text footnote 1]).

the Atlanta lobe of the Idaho batholith, and detrital zircons and plutonic rocks in the Proterozoic Belt Supergroup basin (Figs. 2 and 3). This is evidence of voluminous sediment transported from highlands in Idaho by the newly established Tyee paleoriver starting at 49.4 Ma (Heller et al., 1985, 1992; Dumitru et al., 2013, 2015, 2016). Detrital zircon data indicate that before this, from ca. 53 to 50 Ma, voluminous sediment was transported from Idaho to northern California (rather than to Siletzia) via the Princeton paleoriver (Fig. 3), which flowed into the northern end of the Great Valley forearc basin (Dumitru et al., 2013, 2015). Preserved volumes of Umpqua strata are roughly 25% of that of Tyee strata (Fig. 4A), so the Umpqua-Tyee contact probably represents a pronounced increase in sediment supply as well as a change in sediment source areas. Additional information on Umpqua and Tyee strata that is mainly of local interest is included in Table S1 in the Supplemental Material[1].

The abundant conglomerate interbedded within the marine Siletzia volcanic sequence requires that connection with North America was well established by 53.5 Ma (Wells et al., 2014). Folding and thrusting continued locally through ca. 50 Ma, but ended before deposition of the basal Tyee Formation at 49.4 Ma. The voluminous sediment input from Idaho into the Tyee depocenter strongly suggests that there was not a significant collisional mountain range along the now-covered Siletzia–North America suture, as the Tyee River was able to flow unobstructed to the Tyee basin. Further, the Umpqua and Tyee zircon data do not evidence some missing eroding sediment source, as would be expected if a large mountain range existed. Instead, zircon data from 49.4 to 46.5 Ma Tyee sandstones, as well as from 53 to 50 Ma northern California sandstones, indicate that the main sediment sources were in Idaho. Central Idaho at this time was undergoing intense extension, core complex development, and Challis extensional magmatism, rather than the shortening expected if there had been a strong collisional event (Foster et al., 2007; Gaschnig et al., 2010, 2011; Chetel et al., 2011; Dumitru et al., 2013, 2015). Detritus from central Idaho was apparently transported mainly to northern California by the Princeton River from 53 Ma to 49 Ma and mainly to the Tyee basin by the Tyee River afterwards (Fig. 3). This switch probably does not indicate some implausibly rapid breaching of a major collisional mountain range. Rather, it probably reflects extensive reshaping of landscape and drainage patterns in Idaho during the intense extension, core

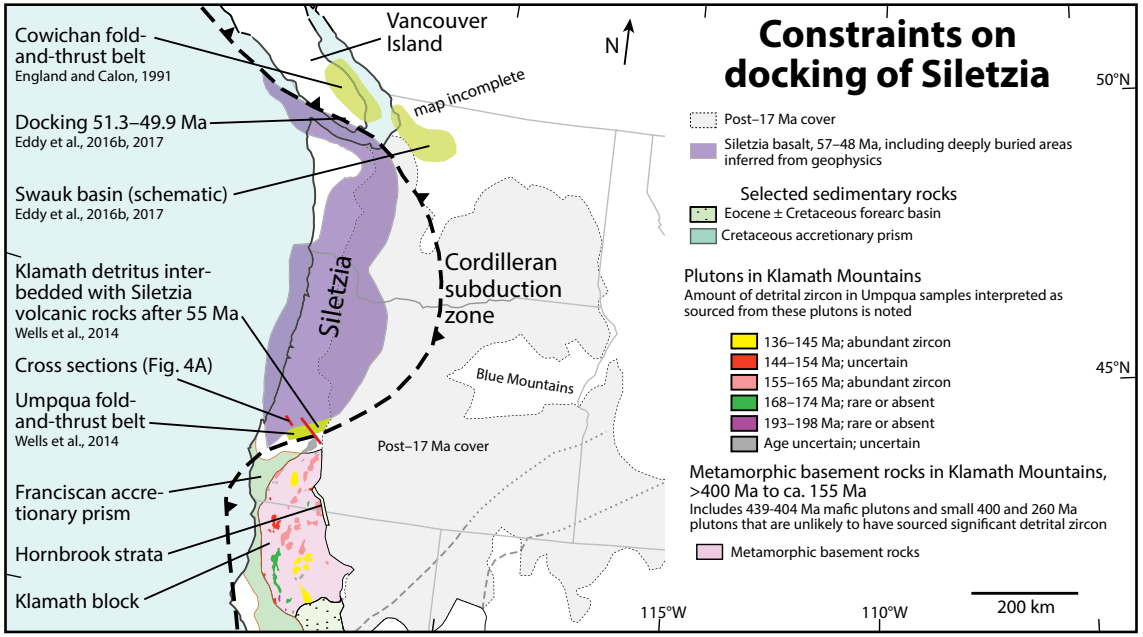

Figure 5. Summary of data from the northern and southern exposed sutures between Siletzia and North America. At the southern suture, Siletzia rotated clockwise ~12° between ca. 55 Ma and ca. 48 Ma, plus an additional ~67° clockwise between 48 Ma and 0 Ma, based on paleomagnetic data (Wells et al., 2014, p. 707–708). At the northern suture, Mesozoic trends on southernmost Vancouver Island were rotated counterclockwise ~20° during suturing (Johnston and Acton, 2003; Wells et al., 2014). See Table 1 for sources.

Data Repository for:
Stern, R.J., and Dumitru, T.A., 2019, Eocene initiation of the Cascadia subduction zone: A second example of plume-induced subduction initiation?: Geosphere manuscript G2050.

Table S1. Umpqua, Tyee, and Franciscan detrital zircon U-Pb data tables and discussion

Umpqua Group, Oregon
Two samples from Umpqua Group (Eocene); Roseburg 1:100,000 Quadrangle, Oregon (Wells et al., 2000).

Sample TD2F-209; mount UCSC3; 43°10.765'N, 123°33.046'W, 605 foot elevation, WGS84 datum.
Twe: Umpqua Group, White Tail Ridge Formation, Coquille River Member
Well bedded medium sandstone; cross bedded; lithic; very little or no detrital white mica visible in hand specimen (in contrast, Tyee sandstones have abundant, coarse detrital white mica).

Sample TD-2F-208; mount UCSC3; 43°10.537'N, 123°31.813'W, 657 foot elevation, WGS84 datum.
Twe: Umpqua Group, White Tail Ridge Formation, Remote Member
Very weathered sample; lithic sandstone; very little or no detrital white mica visible in hand specimen.

Both sample were analyzed by laser ablation–inductively coupled plasma mass spectrometry at the University of California, Santa Cruz, using methods identical to those described in data repository Appendix DR1 of Dumitru et al. (2015). Those methods are available on PDF pages 2-4 and 7 at:
http://www.tandfonline.com/doi/suppl/10.1080/00206814.2015.1008060
See that PDF for abbreviations, etc., used in this table. Plots in Figure 4 were constructed using Isoplot (Ludwig, 2008).

Minor correction to Dumitru et al. (2018): sample TD-2F-418, Yolla Bolly terrane, Franciscan Complex, California
Dumitru et al. (2018) reported U-Pb detrital zircon data from 31 samples, including TD-2F-418 (also called M5). This sample was run in two sessions; the data from one session was inadvertently left out of the 2018 data repository and are included here. The data plots and interpretations in Dumitru et al. (2018) incorporated the data from both sessions, so no changes are needed. In their Fig. 4B, the total number of zircon grains dated should be corrected from n=46 to n=92. In their Table DR5, the total grains accepted should be corrected from 46 to 96 and the total grains rejected from 4 to 8.

Additional interpretative details mainly of local interest for Oregon, Washington, Idaho, Vancouver Island, and Alaska geology

Tyee Formation and Umpqua Group detrital zircon data sets
Dumitru et al. (2013, *Geology*) reported detrital zircon (DZ) U-Pb ages from 7 samples from the Tyee Formation, but those data are not up to the current state of the art. Some of the data were collected about 1999 and others were collected using a quadrupole mass spectrometer that did not allow common-Pb corrections for grains older than 1000 Ma. Number of grains dated per sample was 24 to 53. These data were quite adequate to document abundant Challis-age and Idaho batholith-age detrital zircon in Tyee sandstone, one focus of Dumitru et al. (2013). But they are not adequate for detailed interpretation of the ~25% of grains with Proterozoic ages, which were reported only in data repository tables. We therefore recollected at two of the sample sites and redated ~200 grains per sample using a high-resolution sector mass spectrometer. These data were discussed very briefly in the text of Dumitru et al. (2016, *Geology*) and more extensively in the data repository. Going forward, we recommend that the two samples in the 2016 paper be considered representative of the Tyee Formation and the 7 samples in the 2013 paper be considered useful supplements to the samples in the 2016 paper, but not be included in composite plots.

There is interesting information in the Tyee data that could not be discussed in the short 2013 and 2016 papers and we outline some of it here. This outline also includes more details about the two new Umpqua samples that are not central to the current paper.

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[1]Supplemental Material. Includes Umpqua, Tyee, and Franciscan detrital zircon data tables and detailed discussion. Please visit https://doi.org/10.1130/GES02050.S1 or access the full-text article on www.gsapubs.org to view the Supplemental Material.

complex deformation, and volcanism. This probably created a new Tyee River to Oregon, rather than diverting the existing Princeton River from a mouth in California to a new mouth in Oregon (see also Table S1 [footnote 1]).

The northern Siletzia suture is exposed on the southern tip of Vancouver Island in British Columbia (Fig. 5). Siletzia volcanic rocks on the island and on the Olympic Peninsula of northwestern Washington were emplaced between 53.18 ± 0.17 Ma and 48.364 ± 0.036 Ma, based on eleven new high-resolution zircon U-Pb chemical abrasion–thermal ionization mass spectrometry ages (Eddy et al., 2017). On the peninsula, the base of Siletzia structurally overlies younger subducted sedimentary and volcanic rocks of the Olympic subduction complex, tilted up by Miocene uplift and exposed by erosion of ~11 km of overlying deposits (Brandon and Calderwood, 1990; Stewart et al., 2003). The oldest Siletzia rocks erupted in this area are thus apparently exposed and are ca. 53.2 Ma in age, unless older rocks have been removed by subduction erosion.

On the Olympic Peninsula, the Blue Mountain unit (not to be confused with the Blue Mountains province in Oregon) is a >1–2-km-thick sequence of continent-derived turbidites that was long thought to underlie and to form interbeds within lower Siletzia volcanic rocks. This interlayering relationship was taken as evidence that Siletzia erupted very near the continent (e.g., Cady, 1975; Brandon et al., 2014). However, new high-resolution ages of detrital zircons from four Blue Mountain sandstone samples indicate maximum depositional ages from 47.77 ± 0.57 to 44.72 ± 0.21 Ma, indicating that Blue Mountain strata are younger than Siletzia volcanic rocks (Eddy et al., 2017; see also Wells et al., 2014, their sample JV440). The volcanic rocks that are interlayered with Blue Mountain strata thus postdate the main episode of Siletzia construction; along with the Blue Mountain strata, they probably compose part of the younger Olympic subduction complex that was thrust beneath Siletzia after the latter was emplaced (Eddy et al., 2017). Deformation associated with Siletzia emplacement ("docking") is bracketed between 51.309 ± 0.024 and 49.933 ± 0.059 Ma, as inferred from the timing of shortening in the nonmarine Swauk forearc basin to the east (Fig. 5; Eddy et al., 2016b, 2017).

The new, younger ages for the Blue Mountain unit remove the direct sedimentological evidence that the northern end of Siletzia erupted near the North American continent. The lack of North American sediments interbedded with volcanic rocks in the northern part of Siletzia could mean that this part was farther offshore than the Oregon end. However, there are other good explanations, including that most of northern Siletzia is known to have erupted subaerially, and so may have been an eroding highland rather than a site of sediment deposition (Wells et al., 2014, their figure 2; Eddy et al., 2017, especially their figure 1), or the possibility that rivers and submarine bathymetry funneled North America detritus to other depocenters. Our interpretation of these relationships is summarized in Figure 5. These results show that the currently exposed rocks at the southern end of Siletzia erupted near the Klamath Mountains, whereas the distance between the northern end of Siletzia and Vancouver Island at the time of Siletzia construction is poorly constrained.

Another important point can be summarized as the "Siletzia collision conundrum". What is generally interpreted as collision was a remarkably brief and gentle episode that occurred while Siletzia volcanism was still active. There are only two exposed contacts of Siletzia with pre-Cenozoic North America (Fig. 5). One is in southwestern Oregon where Siletzia is thrust against Klamath Mountain basement (Wells et al., 2014); the other is on the southern end of Vancouver Island, where Siletzia is in fault contact with the previously sutured Pacific Rim and Wrangellia terranes (e.g., Johnston and Acton, 2003). Everywhere else, the contact between Siletzia and the continent is concealed under younger volcanic and sedimentary rocks, including the modern Cascades arc.

The total magnitude of shortening on Vancouver Island is not well constrained. The Cowichan fold-and-thrust belt may be the main zone that accommodated shortening. Palinspastic restoration of five balanced cross sections across the belt yield minimum estimates of shortening of 8.6–12 km (England and Calon, 1991). However, alternate restorations assuming thrust faults that cut upsection at shallower angles yield minimum estimates of ~45–75 km (Johnston and Acton, 2003). In southwestern Oregon, collision and suturing involve thrusting of North American rocks over Siletzia and possible less-extensive obduction of some shallow parts of Siletzia over North American rocks. A restorable cross section (Fig. 4A) accommodates ~37% shortening over a distance of ~50 km (Wells et al., 2014, p. 705–707), or roughly 20 km of shortening. This compares with an estimated 350 km of shortening across the Zagros orogeny in Iran (Pirouz et al., 2017) and 1355 ± 250 km for the Himalayas (Guillot et al., 2003). From this, we infer that suturing here was much gentler than expected for collision of a very large, young, warm (and thus buoyant) allochthonous terrane with a convergent margin. The conclusion of modest shortening is supported by the absence of significant crustal thickening and metamorphism associated with the suturing (e.g., Wells et al., 2014). Furthermore, there is no evidence that a highland existed east of Siletzia after 49 Ma, as would have been expected if a buoyant tract of young, thick oceanic lithosphere had collided with North America; instead, the Tyee River seems to have flowed westward from the Idaho batholith to the coast (Dumitru et al., 2013, 2015). It is also noteworthy that the purported collision was accompanied by strong extension and magmatism several hundred kilometers to the east (Fig. 3). The geologic evidence for emplacement of Siletzia indicates that it either (1) experienced a remarkably gentle collision with North America and a subduction step-out (transference INSZ) event, or (1) was emplaced in situ on the margin of North America, where it experienced modest shortening because of changing stress fields, vertical axis rotations (Wells et al., 2014, p. 707–708), or the behavior of the underlying subduction zone.

WAS THERE AN EOCENE PLUME HEAD BENEATH THE PACIFIC NORTHWEST?

Geophysical Evidence

If a large mantle plume head was emplaced in the uppermost mantle beneath the Pacific Northwest in Paleogene time, there should be geophysical evidence. There have been numerous geophysical studies in the region that

examined its crustal, lithospheric, and asthenospheric structure that can be considered. Unfortunately, there is no accepted protocol for identifying fossil plume heads, and there are multiple ways that the data can be interpreted. However, many of the geophysical features of the region highlighted below are best explained as due to a still-warm fossil mantle plume head.

Geophysical profiles through this region image buried crust. These profiles include a 260-km-long NE-SW seismic refraction line through eastern Washington (Catchings and Mooney, 1988), a receiver function study in the High Lava Plains of southeastern Oregon and surrounding areas (Eager et al., 2011), and a network of vibroseis (seismic vibrator) reflection profiles in south-central British Columbia (Calvert and Talinga, 2014). The locations of these lines are shown on Figure 3. There are also results from EarthScope Transportable Array (TA) studies that provide additional insights, as discussed below.

The seismic refraction line through eastern Washington, west of the 0.706 line, shows that the crust beneath the Columbia Plateau is a largely mafic construction, with thinned continental crust (10–15 km thick) above a great mafic pillow, all associated with a ~5-km-thick sedimentary basin above necking crust (Fig. 6). Catchings and Mooney (1988) interpreted this crustal structure as due to rifting. They linked the buried rift basin to the Chiwaukum, Republic, and Methow grabens of Washington, which contain Eocene sediments and trend southward beneath the northernmost extent of the Columbia River Basalt Group flood basalts, implying that the rift formed in Eocene time. Such structure is expected to result when a great mantle plume head is emplaced at the base of continental crust in an extensional setting, as would be expected to accompany PISI (Gerya et al., 2015).

Magmas from the Eocene plume head appear to have affected crust as far north as central British Columbia, where Calvert and Talinga (2014) used results of a network of 330 km of vibroseis reflection profiles to identify strongly reflective lower crust. The lower crustal reflections correlate with P-wave velocities of 6.5–7.0 km/s, indicating mafic composition. Calvert and Talinga (2014) inferred that lower crust reflectors formed by intrusion of mantle-derived basaltic magma into zones of ductile shearing, and that this magma differentiated and caused widespread Paleocene to Eocene volcanism in the region. In the uppermost crust, extension occurred by block faults that soled into shallowly dipping detachments above 10 km depth. Extension in the deeper upper and middle crust was accommodated on a network of anastomosing shear zones that sole out into the top of the reflective lower crust. Reflector dips indicate that extension was approximately E-W, consistent with NNW-trending horsts and grabens filled with Paleocene to Eocene volcanic and volcaniclastic rocks. This is also the trend of Paleocene–Eocene core complexes in Idaho, Washington, and British Columbia (Fig. 3).

Gao et al. (2011) used EarthScope TA results to confirm the presence of magmatic underplate beneath the Columbia embayment. Recent EarthScope TA deployments of seismometers reveal unusually thin lithosphere underlain by electrically conductive and low-velocity asthenosphere (Fig. 7A). This indicates the presence of unusually hot uppermost mantle, as would be expected for a fossil plume head. There is evidence from Rayleigh wave tomography for a layer of partial melt just below the thin lithosphere of the region, also expected for a fossil plume head (Hopper et al., 2014) (Fig. 7B). Profiles of shear wave tomography (Fig. 7C, 7D; Wagner et al., 2010, 2012) image low-velocity regions at 50–100 km depth in the upper mantle beneath the region, as would be expected for a fossil plume head. Finally, a broad region of negative phase velocity mapped beneath the Pacific Northwest by Hopper et al. (2014) is consistent with the presence of a large fossil plume head (Fig. 7E).

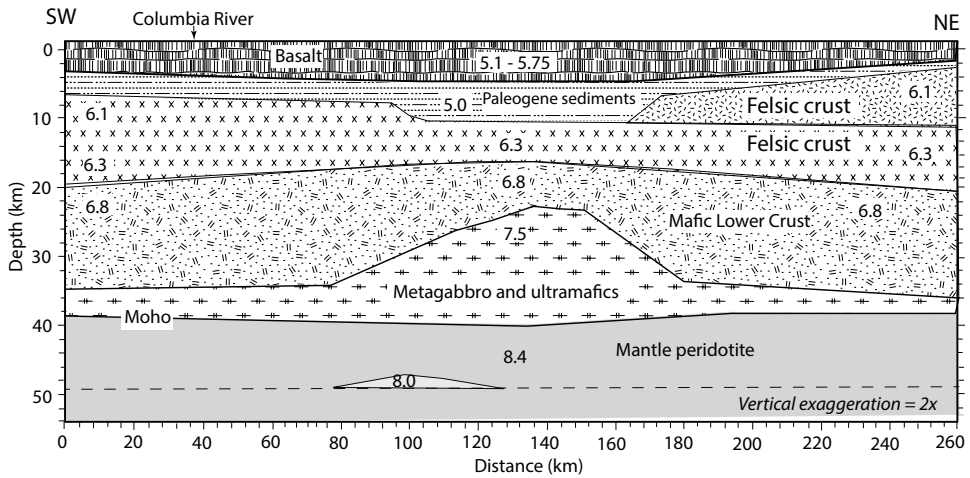

Figure 6. Crustal structure of the Columbia Plateau, modified after Catchings and Mooney (1988). This profile (location shown in Fig. 3) shows stretched crust with a rift-like architecture (including a sedimentary basin). P-wave velocities for various layers are given in kilometers per second; interpretations of lithology are ours. The large volume of lower crust with high P-wave velocities (and densities, not shown) is consistent with a large mafic underplate associated with a mantle plume head. Dashed line at ~50 km depth is a laterally discontinuous reflector. Note that upper (stretched continental) crust is 10–15 km thick; lower underplated crust is 20 km thick.

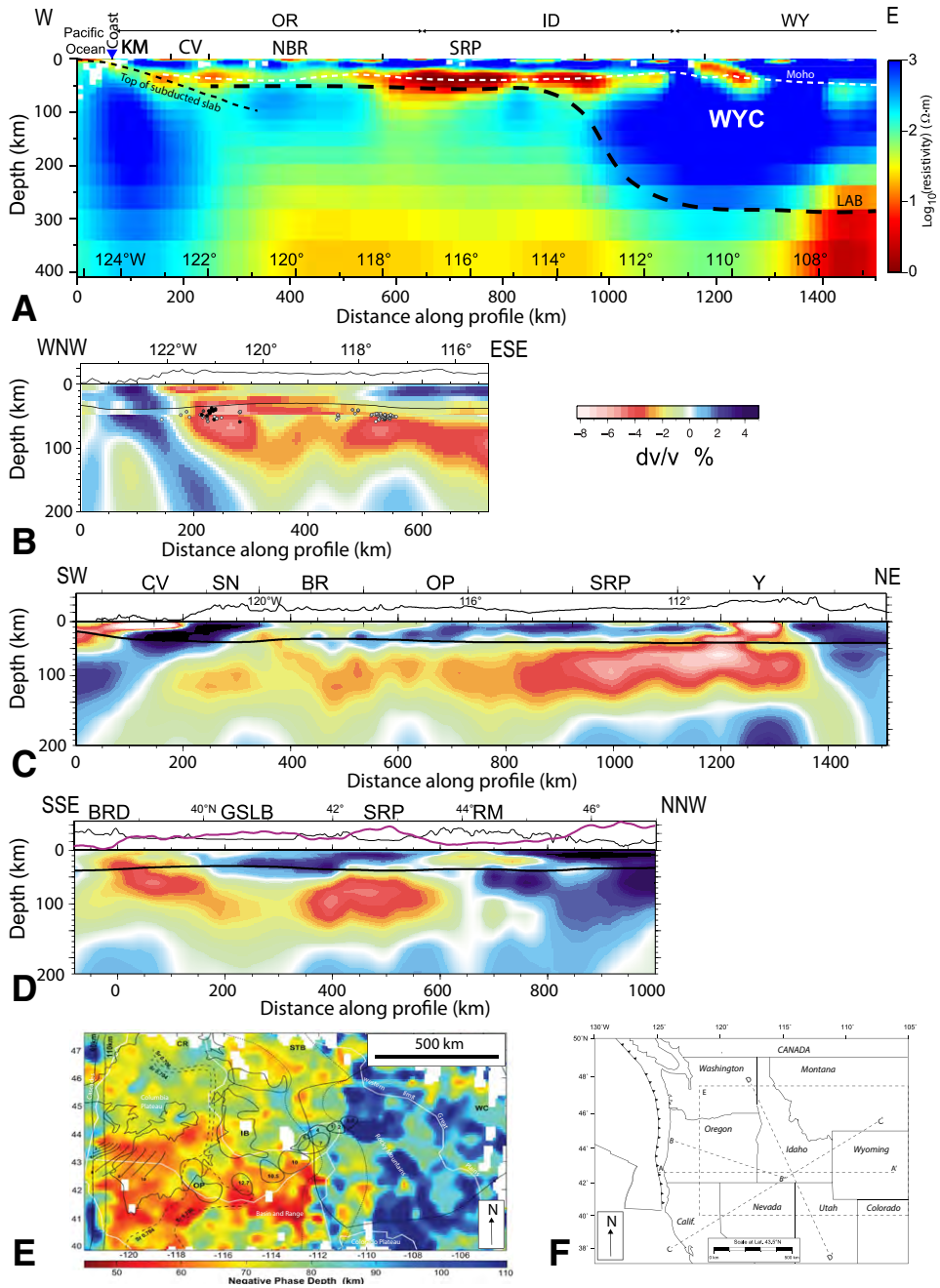

Figure 7. (A) Representative east-west cross section (latitude 42.5°N) illustrating resistivity structure beneath the northwestern United States. Note the thick lithosphere and high-resistivity lithospheric mantle beneath the Wyoming craton of Archean age and the low-resistivity region and thin lithosphere associated with the fossil plume head (Meqbel et al., 2014). (B) ESE-WNW section through southern Oregon and southwestern Idaho (United States), after Long et al. (2012). Background colors show deviations (%dv/v) from a starting one-dimensional velocity model obtained from inverting Rayleigh wave phase velocities derived from ambient noise cross-correlation (shallower than 50 km). (C–D) Southwest-northeast (C) and SSE-NNW (D) shear-wave tomography beneath the northwestern United States (from Wagner et al., 2012). Brown line indicates topography; blue line indicates Moho. Contours are for 1% velocity. (E) Mean depth and amplitude of largest negative Sp (conversion of S waves to P waves at interfaces like the Moho) phase velocity in 50 bootstrapped stacks. White lines are boundaries between tectonic provinces. Black lines are smaller-scale tectonic features, after Hopper et al. (2014). CR—Columbia River. Features noted in section A are U.S. states Oregon (OR), Idaho (ID), and Wyoming (WY). Also shown are Klamath Mountains (KM), Cascade volcanic arc (CV), northern Basin and Range (NBR); Snake River Plain (SRP); and Wyoming craton (WYC). Features noted in section B are subducted Juan da Fuca slab; in section C are Central Valley (CV), Sierra Nevada (SN), Basin and Range (BR), Owyhee plateau (OP), Snake River Plain (SRP), and Yellowstone (Y); in section D are Black Rock desert (BRD), Great Salt Lake Basin (GSLB), Snake River Plain (SRP), and Rocky Mountains (RM). Features noted in section E are depth to subducted Juan da Fuca slab (dotted lines labeled "90km and "110km"); igneous rock $^{87}Sr/^{86}Sr$ = 0.706 and 0.704 (Sr 0.706, Sr 0.704); Sevier thrust belt (STB), Idaho batholith (IB); Owyhee plateau (OP); age of westward migrating felsic volcanism in Ma (thin lines labeled "10", "8", "6", "4", "2"); age in Ma of eastward migrating volcanic centers of the Snake River Plain (irregular ellipses labeled "12.7", "10.5", "10", "7", "6", "4", "2", "1", and "0.6"); and Wyoming craton (WC). Important physiographic features are labeled in white letters. (F) Map showing approximate locations of lines A–D and map E.

A strong, uniform E-W mantle fabric is documented for the region where the low-velocity upper mantle is found (Figs. 7 and 8; Long et al., 2012). The simplest explanation for this fabric is that it reflects upper mantle flow direction, either fossilized in the lithosphere or current asthenospheric flow; less likely is that it reflects fabric of the subducted Gorda plate. The radial-pattern shear wave anisotropy seen in the upper mantle beneath the Columbia embayment may be explained by upper mantle flow away from the plume head.

We conclude that the available geophysical data for the crust and upper mantle beneath the Pacific Northwest is consistent with the presence of a fossil plume head emplaced in Paleogene time.

Evidence from Volcanic Rocks and Core Complexes

There is evidence for a regional magmatic event that could reflect an Eocene plume head centered on the Columbia embayment. About the same time that Siletzia igneous activity was underway, abundant igneous activity occurred in the Challis-Kamloops magmatic belt, which extends for ~1700 km from central British Columbia to northwestern Wyoming (United States) (Fig. 3), the cause of which is controversial. Breitsprecher et al. (2003) ascribed it to a slab window that opened between the subducted Farallon and Kula plates. Others have argued that ridge-trench interaction was responsible (Babcock et al., 1992; Haeussler et al., 2003; Eddy et al., 2016b). In these explanations, there is rarely an attempt to link this activity to what was happening at the same time in Siletzia. It is perhaps not surprising that the two belts of vigorous Eocene igneous and tectonic activity separated by ~1000 km are generally considered to be unrelated. Occam's razor favors considering the possibility that the two belts *are* related; certainly an explanation that treats both as different manifestations of the same cause is worth considering.

We consider the Eocene metamorphic core complexes first. The latest phases of arc magmatism in the Idaho batholith (Bitterroot lobe) occurred in a contractional setting and ended at 53 Ma (Gaschnig et al., 2011), when shortening was abruptly superseded by strong extension and exhumation of the Bitterroot, Anaconda, Clearwater, and Priest River metamorphic core complexes (53–40 Ma; Fig. 3; Foster et al. 2007). Farther north in British Columbia, Vanderhaeghe et al. (1999) proposed that exhumation in the Shuswap core complex (accompanied by detachment faulting) occurred earlier, from ca. 60 to 56 Ma, while the crust was partially molten. However, for the same region, van Rooyen and Carr (2016) recently suggested that crustal-scale orogen-parallel extension commenced ca. 53 Ma, and that pre–53 Ma radiometric ages generally represent events that occurred during the earlier shortening regime. Although this issue remains unresolved, commencement of extension in British Columbia at ca. 53 Ma is more consistent regionally with the age of extension farther south and with the ages of syn-extensional volcanic rocks from central British Columbia to Wyoming.

It is difficult to understand how major regional extension could occur at the same time as Siletzia collision farther west. Extension does accompany collision in orogens such as the Himalaya, but the Eocene core complexes in the Wyoming–British Columbia corridor have a different relationship to Siletzia accretion than those in the Himalaya have to the India-Asia collision. The latter formed in response to ~30 m.y. of crustal thickening and heating, whereas the former formed at the same time as Siletzia emplacement in a region with much less crustal thickening. It is easier to envision how extension could have occurred on the flanks of a mantle plume head, which heated and weakened the crust without greatly thickening it.

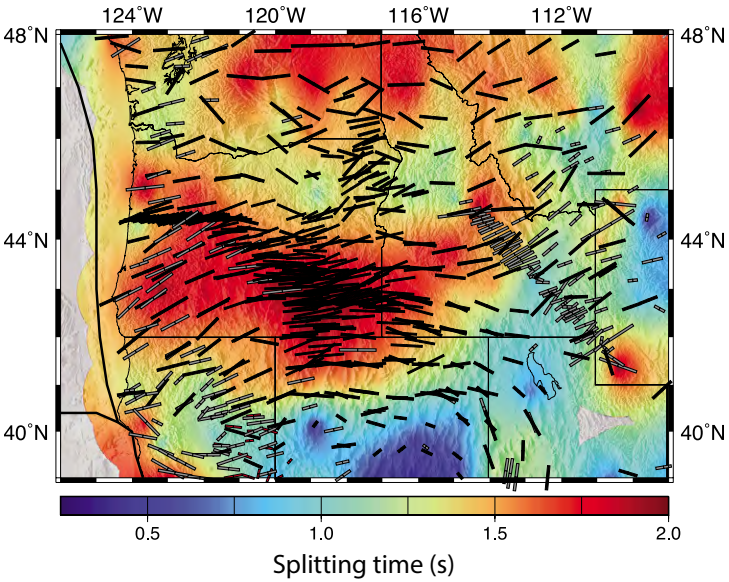

Figure 8. SKS splitting (equated with mantle anisotropy) in the Pacific Northwest of the United States. Orientation and length of bars indicate the fast polarization direction and delay time, respectively. Background colors indicate smoothed, contoured delay times from Long et al. (2012). Note that a strong upper mantle fabric oriented east-west is located at the position of the postulated fossil plume head. Figure from Long et al. (2012, their figure 4).

Now we consider evidence from Eocene volcanism. Figures 3 and 9 highlight the Challis-Kamloops magmatic belt, which erupted in what is now the backarc region (e.g., Haeussler et al., 2003; Wells et al., 2014). These lavas erupted in an extensional setting, as demonstrated by the fact that this region also contains major Eocene extensional metamorphic core complexes. These volcanic rocks and core complexes tend to be slightly younger than the volcanic rocks of Siletzia (Fig. 9). The Challis volcanic-plutonic complex is the most voluminous magmatic complex in the belt. Challis magmatism started at 51 Ma, peaked at 47 Ma, and ended at 43 Ma (Gaschnig et al., 2010, 2011; Chetel et al. 2011). The Absaroka volcanic province near Yellowstone is less well dated, but magmatism probably started ca. 55 Ma and peaked ca. 51.5 Ma to 47 Ma (Feeley et al., 2002; Chetel et al., 2011, their figure 2).

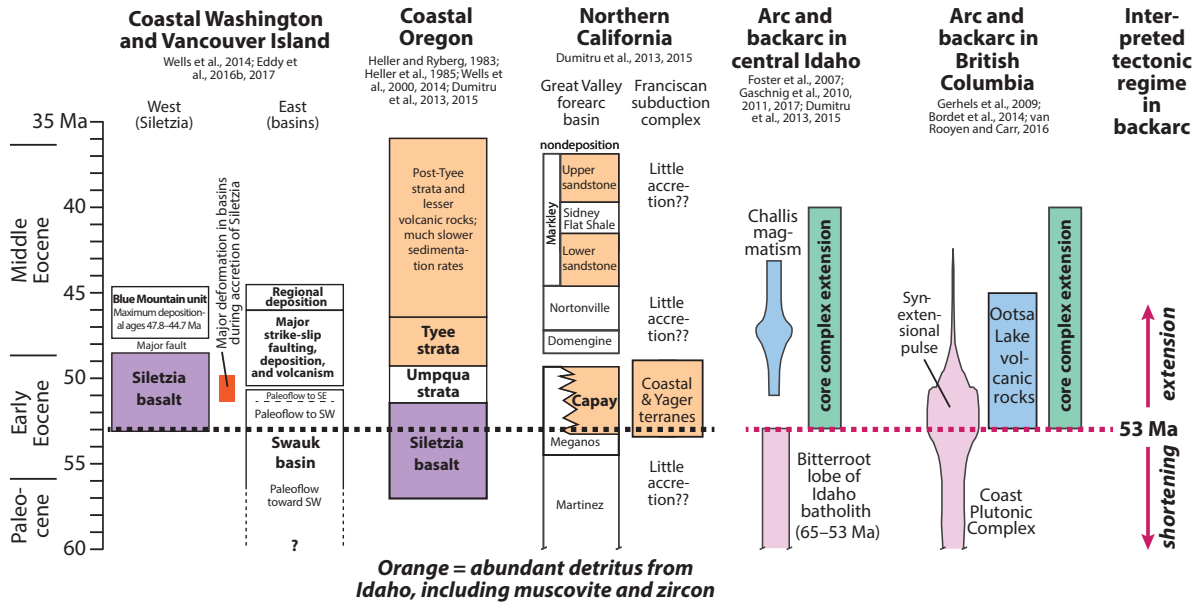

Figure 9. Summary of timing relationships among ca. 60 Ma to ca. 40 Ma rocks in the Pacific Northwest of North America. Note the change in tectonic regime in the backarc at ca. 53 Ma, accompanying the death of the old Cordilleran subduction zone and the birth of the new Cascadia subduction zone.

The Absaroka and Challis volcanics reflect regional lithospheric extension and resultant decompression melting (Feeley et al., 2002; Janecke and Snee, 1993), as would be expected for melting of continental crust above a mantle plume. Rapid extension at 48–44 Ma in the Challis Volcanics of Idaho is shown by the Panther Creek graben, interpreted as an intra-arc rift by Janecke et al. (1997).

In British Columbia, small volumes of Eocene volcanic rocks first erupted at ca. 54 ± 1 Ma, followed by voluminous eruptions from ca. 53.5 Ma to ca. 47 Ma, then relatively minor eruptions until ca. 37 Ma (our inferences mainly from Ickert et al. [2009] and Bordet et al. [2014, their figure 3]). The 53–47 Ma adakitic lavas of the Princeton Group in southern British Columbia (Ickert et al., 2009) may also be related to arrival of the mantle plume head. Deeper levels still are revealed by intrusions like the 48 Ma Golden Horn batholith of Washington (Eddy et al., 2016a), which reflects very high magma accumulation and emplacement rates.

Evidence for the participation of mantle material that can be linked to the arrival of a plume head ca. 55 Ma can be seen in changing Hf and Nd isotopic compositions of intrusive rocks in central Idaho (Gaschnig et al., 2011). Zircon from the 53–65 Ma Bitterroot lobe of the Idaho batholith has $\varepsilon_{\text{Hf-initial}}$ values of –18 to –23 (n = 9), whereas 51–47 Ma Challis intrusive rocks have –2 to –28 (n = 9). $\varepsilon_{\text{Nd-initial}}$ similarly shifted from –14 to –17 in the Bitterroot peraluminous suite (n = 12) to –3 to –17 in Challis igneous rocks (n = 15). This shift to more mantle-like values in many Challis samples is consistent with the arrival of a mantle plume at ca. 53 Ma.

There was clearly a huge early Eocene thermal anomaly in the Pacific Northwest, with Siletzia on its western flank and a complementary zone of extension and magmatism on the eastern flank. These two belts are quite different, partly because these two great regions of Eocene igneous and tectonic activity straddle the 0.706 line, which approximates the western edge of the Wyoming craton, which responded very differently to the impact of the mantle plume head. Marginal regions to the east were heated by the plume, but large-scale extension to the point of plate separation was precluded. Nevertheless, the upsurge in igneous activity and formation and uplift of core complexes in this region imply very high thermal gradients in the crust, as would be expected for continental crust above the broad, hot head of a mantle plume (Şengör and Burke, 1978). In this region of thicker lithosphere and crust, it was much more difficult for mantle-derived mafic magmas to reach the surface; instead, magmas interacted with the lithosphere and ponded at the base of the continental crust, resulting in more-evolved magmas and forming core complexes. In contrast, the region to the west was dominated by thinner lithosphere that was easier for mafic magmas to traverse and erupt through. This region also abutted Pacific oceanic lithosphere, where lithospheric collapse resulted in regional extension and large-scale decompression melting of the upper mantle. In addition, the Kula-Farallon spreading ridge may have intersected this area, providing a favorable pathway for the eruption of plume-related lavas (Duncan, 1982; McCrory and Wilson, 2013; Wells et al., 2014).

In the next section, we build on evidence for the existence of a mantle plume head beneath the Pacific Northwest in Eocene time to propose how lithospheric collapse along its western margin created a new subduction zone.

THE CASCADIA PLUME-INDUCED SUBDUCTION INITIATION HYPOTHESIS

The previous section provides evidence that an active mantle plume head reached the base of the lithosphere beneath the Pacific Northwest ca. 55–53 Ma, and that the interaction of the plume head with the overlying lithosphere dominated tectonic and magmatic activity of the Pacific Northwest until ca. 45 Ma. Could such a plume head have caused lithospheric collapse along its western margin to form the modern Cascadia subduction zone? PISI requires a large, hot plume of ~1000 km dimension or more (Gerya et al., 2015). The Pacific Northwest plume head outlined in Figure 3 was ~1200 km across; a plume head of this size could plausibly have been responsible for lithospheric collapse on its western flank to form the modern Cascadia subduction zone. The evidence suggests that such a plume head arrived at the base of the lithosphere of this region in early Eocene time, that it massively affected deformation and magmatism in the region, and that it was responsible for shutting down the old subduction zone and starting a new one.

Lithospheric collapse along a fracture zone explains formation of the Izu-Bonin-Mariana subduction zone ca. 52 Ma, and it is increasingly clear that many forearcs and ophiolites form as a result of similar decompression-related igneous activity accompanying this style of subduction initiation (Stern et al., 2012). There is evidence for a similar episode of lithospheric collapse around the southern and western margins of the Late Cretaceous Caribbean plume head, and that this formed the Central American subduction zone (Whattam and Stern, 2015). An analogous situation appears to have formed the Cascadia subduction zone in the Eocene.

The topographic expression of a large plume head would vary depending on whether it was emplaced beneath a rifting environment—as would be expected during PISI—compared to when it was emplaced beneath a region that was not rifting. In the latter case, flood basalt eruptions would build a high-standing plateau, such as the Ontong-Java Plateau or the Deccan Traps, but formation of a rift basin would more likely if there were strong extension above the plume head, and this appears to have been the situation in the Pacific Northwest. Consequently, much of the region above the purported Eocene plume head is buried beneath Neogene lavas, including the Columbia River flood basalts, Nevada high plains, Snake River Plain, and Yellowstone lava fields (Fig. 10). Direct geologic evidence of the plume head is found only on its margins, in Siletzia and in the Challis-Kamloops magmatic belt (Fig. 3). It is likely that early Eocene flood basalts related to the arrival of the plume head at the base of the lithosphere are buried beneath younger lavas and sediments of this region.

Another key point is that the plume head envisioned here affected two distinct crustal types: younger, ensimatic accreted terranes to the west of the 0.706 line and much older continental crust of the Wyoming "craton" to the east (Fig. 2). The 0.706 line corresponds to a sharp change in crustal thickness, from >35 km in the east to <30 km in the west (Eager et al., 2011; Stanciu et al., 2016).

Figure 11 summarizes our hypothesis for how the Cascadia subduction zone formed in response to arrival of a large plume head. Figure 11A shows the situation at ca. 65 Ma before the plume head arrived. At this time, the Farallon plate subducted eastwards beneath the western margin of North America, which was made up of Archean crust of the Wyoming craton to the east and accreted terranes farther west. Figure 11B shows the arrival of the ~1000-km-diameter plume head at ca. 55 Ma. Thermal buoyancy of the plume and associated igneous activity weakened the Farallon plate, ending subduction to the east and causing widespread extension above the plume head, including forming core complexes near the edge of the craton. Lithospheric weakening on western side of plume head allowed collapse of Farallon plate (Fig. 11C), starting the Cascadia subduction zone.

The simple tectonic scenario shown in Figure 11 is consistent with all large-scale features of the region. It provides a simple and self-consistent explanation for the remarkable Eocene events that affected all of the Pacific Northwest at about the same time. It is similar to the evolutionary sequence shown in Schmandt and Humphreys (2011, their figure 2D) except that the presence of the plume head provides a mechanism for developing a new subduction zone. The Cascadia PISI hypothesis has the added advantage of simplicity: it does not require a host of coincidences to explain the many manifestations of Eocene tectonic and magmatic activity in the region. It does away with the many explanations for different subregions in the affected area: there is no need for an oceanic plateau to have formed and then been immediately accreted to the continent as Siletzia; there is no need to invoke formation of a new subduction zone by transference; there is no need for subduction of a spreading ridge; and there is no need for multiple slab tears. There may not even be a need to invoke the existence of the Resurrection plate, at least offshore Oregon and Washington (Haeussler et al., 2003). The Pacific Northwest PISI hypothesis thus has the advantage of resolving many geoscientific problems in the region.

ISOTOPIC EVIDENCE THAT THE SILETZIA AND YELLOWSTONE PLUMES ARE RELATED

A key question is whether the Siletzia plume and the Yellowstone plume are derived from similar mantle sources, and isotopic data for igneous rocks can illuminate this problem. Because the Yellowstone plume currently rises beneath continental lithosphere, modern lavas are felsic and continental crust has been massively incorporated during magma formation. To see the plume's uncontaminated mantle signal, we must look at older lavas erupted west of the 0.706 line. Wolff et al. (2008) reported such data from the Steens, Imnaha, and Picture Gorge Basalts within the 16.7–15.6 Ma Columbia River Basalt Group (Fig. 10). These show $^{87}Sr/^{86}Sr$ = 0.7030–0.7045, ε_{Nd} = +2 to +10, and $^{206}Pb/^{204}Pb$ = 18.8–19.1. Siletzia basalts show $^{87}Sr/^{86}Sr$ = 0.7030–0.7038, ε_{Nd} = +5 to +8, and $^{206}Pb/^{204}Pb$ = 18.7–19.7 (Phillips et al., 2017). The isotopic similarities do not require that the Siletzia and Yellowstone plumes are the same. However, the similarities are consistent with identifying the Paleogene Siletzia lavas and Neogene Columbia River Basalt Group–Snake River Plain lavas as older and younger melts generated from the same mantle plume.

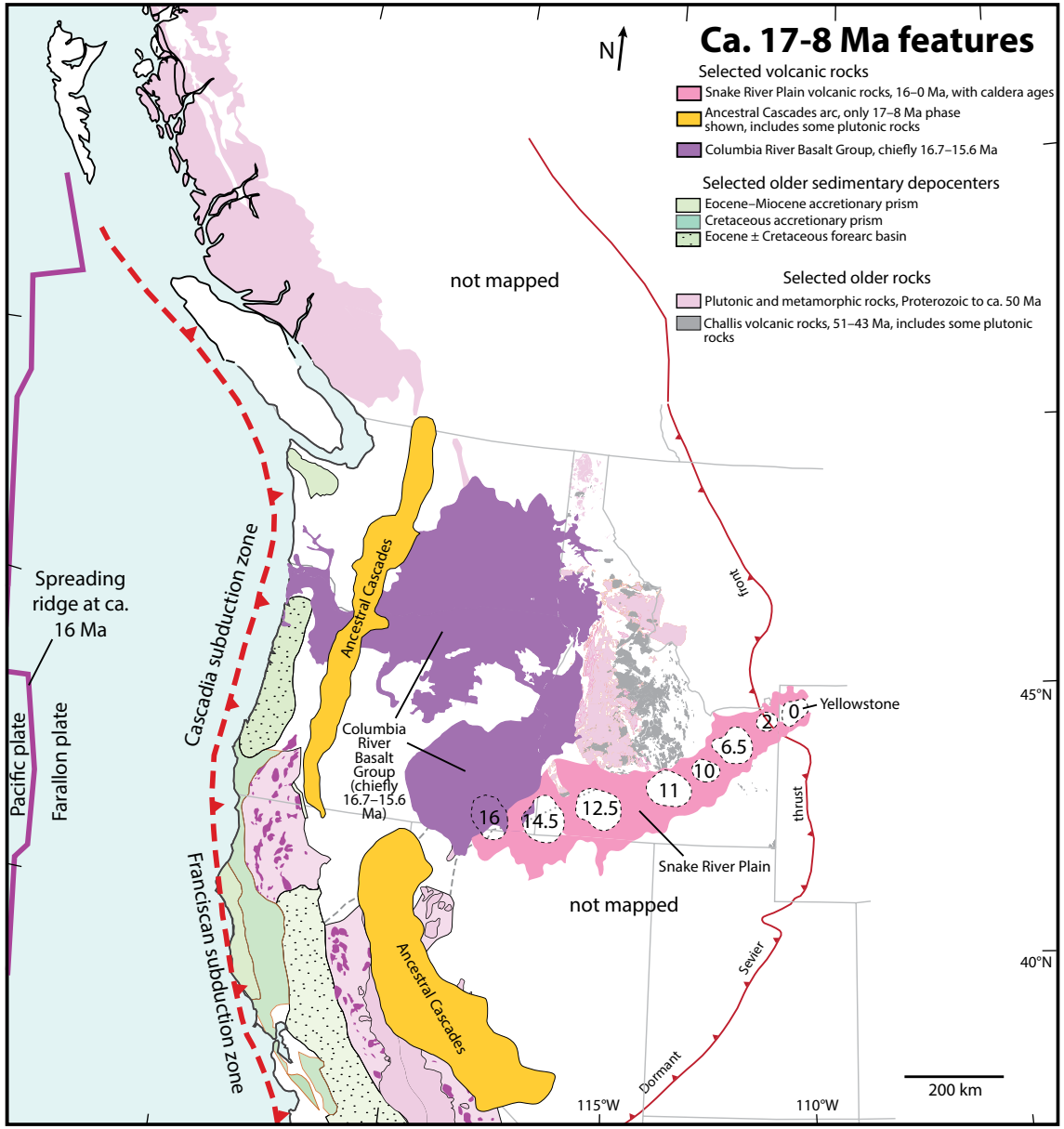

Figure 10. Map of selected tectonic elements shortly after ca. 17 Ma, when the Yellowstone plume began generating the voluminous volcanic rocks of the Columbia River Basalt Group and the Snake River Plain. We have not restored post–17 Ma displacements, such as Basin and Range extension. See Table 1 for sources.

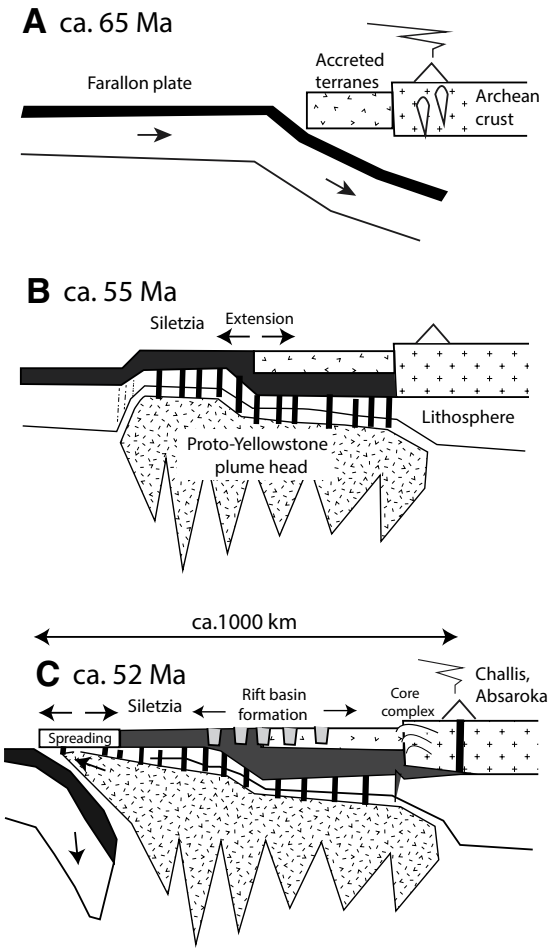

Figure 11. Simplified cross section at latitude ~46°N capturing the most important processes of Cascadia plume-induced subduction initiation. (A) Situation ca. 65 Ma before the plume head arrives. Oceanic lithosphere of the Farallon plate is subducted eastwards beneath the western margin of North America, comprising Archean crust of the Wyoming craton and accreted terranes to the west. Subduction is responsible for the Idaho batholith. (B) Arrival of the ~1000 km diameter Proto-Yellowstone plume head ~55 Ma. Thermal buoyancy of the plume and associated igneous activity weakens the Farallon plate, ending subduction and causing widespread extension above the plume head, including core complexes near the edge of the craton. Lithospheric weakening on western side of plume head allows collapse of Farallon plate (C), starting the Cascadia subduction zone.

PETROLOGIC EVIDENCE FOR THE INVOLVEMENT OF PLUME-HEAD MANTLE IN THE SOURCE OF CASCADES BASALTS

If the entire upper mantle beneath the Columbia embayment was the site of a Paleocene–Eocene plume head, evidence for this should be seen in modern Cascades arc magmas. In fact there is strong evidence that the upper mantle beneath the Pacific Northwest has been strongly polluted by plume-like mantle, as would be expected if a large mantle plume head invaded the region. Leeman et al. (2005) examined primitive Neogene Cascades arc lavas and identified two groups: OIB-like group I and arc-like group II. Some arcs have OIB-like magmas erupted in a rear-arc setting (e.g., eastern China; Kimura et al., 2018), but basalts of both groups I and II are widely distributed east-west across the Cascades arc. Leeman et al. (2005) concluded that the mantle responsible for generating group I magmas was deeper than that responsible for group II magmas (50–70 km versus 30–50 km deep), inferring that there was compositional stratification in the mantle wedge. The abundance of OIB-like basalts in the Cascades arc contrasts with typical arc basalts, which generally show strong evidence of modification by slab-derived fluids (Stern, 2002). Leeman et al. (2005, p. 92) further observed that "…Cascades magmatism may differ in important ways from the standard arc paradigm." These observations and inferences, especially for group I Cascades primitive basalts, are of the sort that might be expected from prolonged involvement of the Paleocene–Eocene plume head in the mantle beneath the Cascades arc.

Documentation of OIB-like arc basalts in Cascadia begs the question of whether or not the presence of OIB-like arc magmas should be expected for convergent margins that form by PISI. It is noteworthy that OIB-like arc basalts are also found for the Central American arc, especially in central Costa Rica (e.g., Gazel et al., 2009). This convergent margin is argued to have formed by PISI in Late Cretaceous time (Whattam and Stern, 2015). From these two examples, it seems that the presence of a significant proportion of OIB-like arc lavas may be a distinctive signature of convergent margins formed by PISI.

WHY WAS THE YELLOWSTONE PLUME DORMANT FOR 30 MILLION YEARS?

A potential serious criticism of the Cascadia PISI hypothesis is that there is not a continuous record of plume activity from its first appearance in Paleocene–Eocene time until today. The presumed plume head was apparently very active at 55–50 Ma, and there are quite a few volcanic units erupted between 48 and 34 Ma (Wells et al., 2014), but there is little evidence of voluminous plume activity until the Columbia River Basalt Group began to erupt ca. 16.7 Ma (Reidel et al., 2013). This is a potentially fatal flaw for the Cascadia PISI hypothesis because the tails of mantle plumes typically persist for tens of millions of years after LIPs form and rarely "turn off" until they die. A plausible explanation of the "plume gap"—why the Siletzia-Yellowstone plume disappeared for 30 million years and then reappeared—is needed for the Cascadia PISI hypothesis to be viable.

What happened to the purported plume during the ~30 m.y. interval? It could be that the missing Oligocene–early Miocene igneous rocks are buried beneath Columbia River Basalt Group lavas east of the Cascades. However, evidence for 50–20 Ma plume activity is not observed in outliers from this region. These outliers instead expose Oligocene and early Miocene volcaniclastic rocks of the John Day Formation (Robinson et al., 1984). These reflect activity of the Cascades arc, not a mantle plume.

A plausible explanation for the plume gap is that the newly subducted slab chilled the mantle plume, interrupting upwelling and melting of plume mantle, as shown in Figure 12B. Today, the Yellowstone plume can be traced tomographically to ~660 km depth (Smith et al., 2009; Schmandt et al., 2012), and it is plausible that it has had a similar vertical structure for the past ~55 m.y. If so, the subducted Farallon–Juan da Fuca plate would have intersected the upwelling plume tail soon after subduction began, disrupting mantle upwelling and chilling the upper mantle beneath the region, with the effect of effectively shutting down the plume. If this is the reason that the mantle plume became quiescent, why did it resume activity after a 30 million year hiatus? A possible answer is that, with time, the subduction zone became increasingly

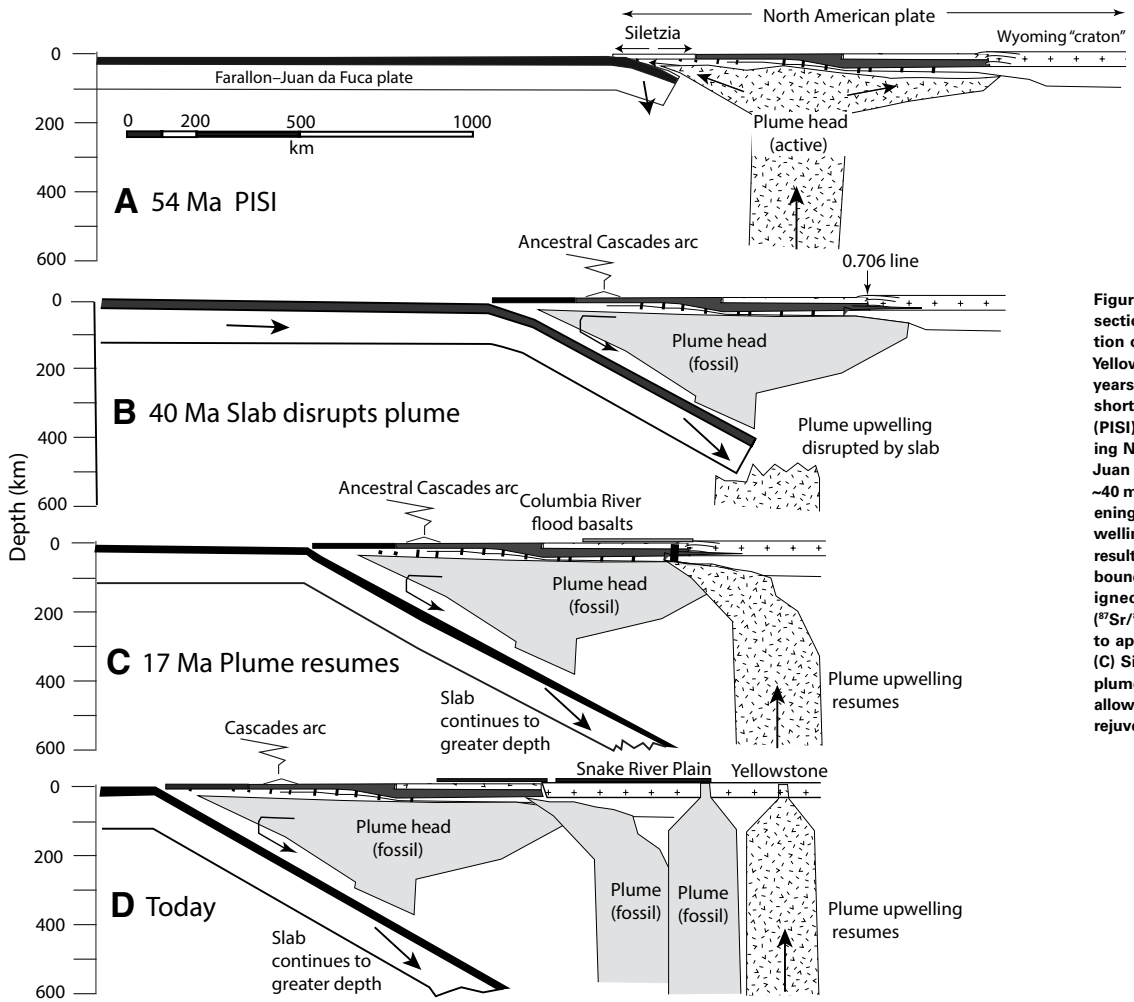

Figure 12. Highly schematic and simplified cross section at latitude ~46°N summarizing the evolution of the Cascadia margin and why the Siletzia-Yellowstone plume was quiescent for ~30 million years between ca. 50 and ca. 20 Ma. (A) Situation shortly after plume-induced subduction initiation (PISI). (B) Margin after 14 million years, assuming North America moves west at ~25 mm/yr and Juan de Fuca–North America plate convergence is ~40 mm/yr (Long, 2016). The lengthening and deepening slab penetrates into the region of mantle upwelling, disrupting upflow and chilling the region, resulting in plume shutdown. The 0.706 line refers to boundary between less radiogenic ($^{87}Sr/^{86}Sr$ <0.706) igneous rocks to the west and more radiogenic ($^{87}Sr/^{86}Sr$ <0.706) igneous rocks to the east, thought to approximate the edge of the Wyoming craton. (C) Situation at 17 Ma when separation between plume and subducted slab increased sufficiently to allow mantle upwelling to result, resulting in plume rejuvenation. (D) Present situation.

separated from the position of the plume. As North America continued to move west-southwest at 20–30 mm/year (Gripp and Gordon, 2002; Smith et al., 2009), the slab moved away from the region of plume upwelling. In 30 million years, the Cascadia subduction zone and the plume would have separated by 600–900 km, enough separation between plume and subducted slab to allow the plume to resume upwelling and partially melt to generate the Columbia River Basalt Group (Fig. 12C). After this time, the distance between the subducted Juan da Fuca plate and the Yellowstone mantle plume has continued to increase, and the plume has operated continuously (Fig. 12D). This scenario warrants detailed geodynamic modeling to further assess and constrain this unusual example of plume–subduction zone interaction.

In should be noted that our model implies that the Columbia River Basalt Group was generated by a reformed plume head, one that combined elements of a plume head and tail. The estimated volume of the Columbia River Basalt Group is 210,000 km^3 (Reidel et al., 2013), roughly 10% of Siletzia's volume of 1,700,000–2,600,000 km^3 (Wells et al., 2014).

CONCLUSIONS

We propose a new hypothesis (below "the hypothesis") for the origin of the Cascadia subduction zone in early Paleogene time. The hypothesis calls on a mantle plume head ~1200 km in diameter to have reached the base of the lithosphere beneath the Columbia embayment. This disrupted an older eastward-dipping subduction zone and caused a wide range of regional phenomena, most importantly lithospheric collapse along its western margin to form the modern Cascadia subduction zone. The hypothesis is motivated by weaknesses of the current paradigm that Siletzia formed as an oceanic plateau offshore and was accreted to North America, causing a new subduction zone to form outboard (below "the paradigm"). The hypothesis is based on a modern understanding of how new subduction zones do and do not form. The paradigm and hypothesis agree that understanding the origin of the Siletzia LIP is key to understanding the early Paleogene tectonic transformation of the Pacific Northwest, but the hypothesis explains other phenomena in the region as well. We stress the following 10 points:

1. The Yellowstone mantle plume is still active in the region and was almost certainly the plume called for in both the paradigm and the hypothesis.
2. The paradigm requires subduction initiation by transference following collision of Siletzia with North America. Transference is difficult to accomplish because of the strength of the oceanic lithosphere and is not documented among Cenozoic examples of subduction initiation.
3. Abundant clastic sediment shed from the Klamath Mountains is interbedded with Siletzia lavas in southwestern Oregon. The paradigm implies that a trench separated Siletzia from North America, which should have diverted North American detritus away from the growing Siletzia plateau prior to accretion. The hypothesis explains that Siletzia grew in situ in the Columbian embayment.
4. Early Paleogene shortening is modest (9–75 km) and is consistent with either the paradigm or the hypothesis.
5. Geophysical studies show evidence of much mafic crust beneath the Columbia Plateau. Regional considerations suggest emplacement in early Paleogene time. The upper mantle beneath the region shows evidence of a fossil plume head. Seismic anisotropy is consistent with east-west mantle flow expected by the hypothesis.
6. The hypothesis helps explain the abundant igneous activity of the ~1700-km-long Challis-Kamloops magmatic belt and the formation of early Paleogene extensional core complexes along the eastern margin of the plume head.
7. Isotopic compositions of Siletzia lavas and uncontaminated Neogene Yellowstone lavas are similar, consistent with derivation from a similar mantle plume source. Invasion of an early Paleogene plume head beneath the Pacific Northwest explains the presence of Cascades arc lavas with strong OIB-like affinities.
8. The plume apparently was quiescent for ~30 m.y. This can be explained by interaction between the newly subducted slab and the plume. The deepening slab intersected the plume, chilling the mantle and disrupting plume upflow. Southwest migration of North America at 20–30 mm/yr increasingly separated slab and plume. After 30 m.y. and 600–900 km of separation, plume upwelling resumed, forming the Columbia River flood basalts.
9. The global conclusion that it is difficult to initiate subduction by transference is supported, and a second example (after the Caribbean in the Late Cretaceous) of plume-induced subduction initiation is identified.
10. Further work is needed to test and refute or refine the hypothesis.

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