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Upper-plate response to ridge subduction and oceanic plateau accretion, Washington Cascades and surrounding region: Implications for plate tectonic evolution of the Pacific Northwest (USA and southwestern Canada) in the Paleogene

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ABSTRACT

The interaction between subduction zones and oceanic spreading centers is a common tectonic process, and yet our understanding of how it is manifested in the geologic record is limited to a few well-constrained modern and ancient examples. In the Paleogene, at least one oceanic spreading center interacted with the northwestern margin of North America. Several lines of evidence place this triple junction near Washington (USA) and southern British Columbia (Canada) in the early to middle Eocene, and we summarize a variety of new data sets that permit us to track the plate tectonic setting and geologic evolution of this region from 65 to 40 Ma. The North Cascades segment of the voluminous Coast Mountains continental magmatic arc experienced a magmatic lull between ca. 60 and 50 Ma interpreted to reflect low-angle subduction. During this period of time, the Swauk Basin began to subside inboard of the paleo-trench in Washington, and the Siletzia oceanic plateau began to develop along the Farallon plate–Kula plate or Farallon plate–Resurrection plate spreading center. Farther east, peraluminous magmatism occurred in the Omineca belt and Idaho batholith. Accretion of Siletzia and ridge-trench interaction occurred between ca. 53 and 49 Ma, as indicated by: (1) near-trench magmatism from central Vancouver Island to northwestern Washington, (2) disruption and inversion of the Swauk Basin during a short-lived contractional event, (3) voluminous magmatism in the Kamloops-Challis belt accompanied by major E-W extension east of the North Cascades in metamorphic core complexes and supra-detachment basins and grabens, and (4) southwestward migration of magmatism across northeastern Washington. These events suggest that flat-slab subduction from ca. 60 to 52 Ma was followed by slab rollback and breakoff during accretion of Siletzia. A dramatic magmatic flare-up was associated with rollback and breakoff between ca. 49.4 and 45 Ma and included bimodal volcanism near the eastern edge of Siletzia, intrusion of granodioritic to granitic plutons in the crystalline core of the North Cascades, and extensive dike swarms in the North Cascades. Transtension during and shortly before the flare-up led to >300 km of total offset on dextral strike-slip faults, formation of the Chumstick strike-slip basin, and subhorizontal ductile stretching and rapid exhumation of rocks metamorphosed to 8–10 kbar in the North Cascades crystalline core. By ca. 45 Ma, the Farallon–Kula (or Resurrection)–North American triple junction was likely located in Oregon (USA), subduction of the Kula or Resurrection plate was established outboard of Siletzia, and strike-slip faulting was localized on the north-striking Straight Creek–Fraser River fault. Motion of this structure terminated by 35 Ma. These events culminated in the establishment of the modern Cascadia convergent margin.

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INTRODUCTION

Plate tectonic margins vary from long-lived stable settings to those that change rapidly from one type of boundary to another over only a few million years. The modern Cascadia subduction zone, in the Pacific Northwest (USA) and southwest British Columbia (Canada), has been a convergent plate margin since the mid-Eocene (≤45 Ma) (du Bray and John, 2011). Earlier, the northern Washington (USA) Cascade Range was part of a long-lived continental magmatic arc that is also manifested as the Coast Mountains batholith and parts of the Idaho batholith (e.g., Gehrels et al., 2009). The North Cascades segment of the Coast Mountains arc was active from ca. 96 to 60 Ma and changed from a contractional-convergent to oblique-convergent regime during that time (e.g., Brown and Talbot, 1989; Miller et al., 2009, 2016). Between the older Coast Mountains and Cascadia magmatic arc regimes was an ~25 m.y. period, from ca. 65 to 40 Ma, during which the Washington Cascades and the surrounding region experienced many dynamic changes that can be linked to two major Paleogene tectonic events: spreading ridge–trench interaction and the formation and accretion of an oceanic plateau.

Plate reconstructions suggest that the Farallon plate–Kula plate, Farallon plate–Resurrection plate, or Farallon plate–Orcas plate spreading ridge(s) interacted with North America near the Pacific Northwest during the Paleogene (e.g., Atwater, 1970; Wells et al., 1984; Engebretson et al., 1985; Haeussler et al., 2003; Madsen et al., 2006; Clennett et al., 2020; Fuston and Wu, 2021) (Fig. 1). Based on ca. 51–49 Ma near-trench magmatism from central Vancouver Island (British Columbia) to northwestern Washington, a ridge is assumed to have intersected North America near these locations at that time (e.g., Cowan, 2003; Madsen et al., 2006), although how this triple junction migrated along the margin prior to 52 Ma is poorly understood. The Siletzia terrane, a basaltic oceanic plateau, formed along this oceanic spreading center and was accreted to the Pacific Northwest ca. 50 Ma (e.g., McCrory and Wilson, 2013; Wells et al., 2014). Farther inland, there was a change from a long-lived thrust belt (e.g., Mudge and Earhart, 1980; Price, 1981) to E-W extension and widespread magmatism at ca. 55–53 Ma (e.g., Ewing, 1980; Parrish et al., 1988). These and other changes in the upper plate of the system are the basis for our attempt at a comprehensive model of the 65–40 Ma tectonic evolution of the Washington Cascades and Pacific Northwest.

In this paper, we synthesize data on the ages and types of sedimentary basins (Evans, 1994; Johnson, 1984; Eddy et al., 2016a; Donaghy et al., 2021), age, geochemistry, and spatial patterns of magmatism (e.g., Breitsprecher et al., 2003; Madsen et al., 2006; Miller et al., 2009), and deformation styles and exhumation patterns across Vancouver Island to the Washington Cascades (e.g., Johnston and Acton, 2003; Miller et al., 2016) (Figs. 2 and 3). We present this 25 m.y. geologic history in a series of time slices and place the discussion in the context of the greater region from northern California (USA) to southern British Columbia and inland to the Rocky Mountains (Fig. 2). Integrated within this discussion are a series of new maps that restore slip on the major Paleocene–Eocene strike-slip faults (Figs. 4–7). Boundaries between time slices coincide with transitional periods in at least one of the major processes emphasized in the synthesis (i.e., magmatism, sedimentation, metamorphism, deformation, and/or exhumation). A critical aspect of this work is the incorporation of new high-precision U-Pb zircon age constraints tied to detailed field observations (e.g., Eddy et al., 2016a, 2017a, 2017b; Miller et al., 2016, 2022), which enables the construction of a detailed timeline not previously possible. Moreover, the varied levels of exhumation within the region allow us to study how the changing tectonic setting was manifested at a wide range of Eocene crustal levels. In particular, we explore the upper-plate events in the Washington Cascades and surrounding region in relation to changing plate boundaries, especially the formation and accretion of Siletzia (Wells et al., 2014), and the shifting location of ridge-trench interaction. The study area is described in terms of western, central, and eastern regions, which roughly correspond to the forearc, arc, and backarc regions of the North Cascades segment of the Coast Mountains batholith in the Late Cretaceous (Fig. 2). We utilize these geographic terms because the dynamic tectonic changes described herein make it difficult to define regions typically associated with a stable subduction zone.

PLATE TECTONIC SETTING

There has long been uncertainty about the Late Cretaceous to early Cenozoic plate configuration in the northeast Pacific basin. There is general agreement that the Kula plate originated from rifting of the Pacific plate at ca. 83 Ma and that the northern boundary of the Farallon plate was a ridge that intersected the continental margin at a poorly constrained location (e.g., Atwater, 1970; Woods and Davies, 1982; Engebretson et al., 1985; Stock and Molnar, 1988; Thorkelson and Taylor, 1989). Subsequent models proposed the potential existence of a now-subducted Resurrection plate (e.g., Haeussler et al., 2003; Madsen et al., 2006; Fuston and Wu, 2021) (Fig. 1) or Orcas plate (Clennett et al., 2020). During the interval from ca. 85 to 60 Ma, the northern Cordillera was an oblique, transpressional convergent margin (e.g., Engebretson et al., 1985; Doubrovine and Tarduno, 2008), and northward translation of the Washington Cascades may have been rapid as the southern part of the Insular superterrane (the Baja British Columbia hypothesis; e.g., Cowan et al., 1997; Umhoefer and Blakey, 2006). Relative to North America, motion of the Farallon plate was to the NE to ENE, and motion of the Kula (Resurrection or Orcas?) plate was to the N to NNE and thus more oblique than that of the Farallon plate. Both oceanic plates were moving rapidly (50–150 km/m.y.) during this time (e.g., Engebretson et al., 1985; Doubrovine and Tarduno, 2008; Wright et al., 2015; Fuston and Wu, 2021).

Formation of the Siletzia terrane was a major factor in the Paleogene tectonic evolution of the Pacific Northwest. This terrane represents a large igneous province that developed between 56 and 49 Ma near an oceanic spreading center, and it is probably an early manifestation of the Yellowstone hotspot (e.g., Gao et al., 2011; McCrory and Wilson, 2013; Wells et al., 2014; Camp and Wells, 2021). We support previous work that infers the triple junction between the Farallon–North America–Kula (or Resurrection or Orcas) plates lay along central Vancouver Island by 55–53 Ma (e.g., Madsen et al., 2006) (Figs. 1 and 4). From 52 to 49 Ma, a triple junction is interpreted to have interacted with the continental margin along central to southern Vancouver Island (Fig. 1), given that this interval is marked by near-trench magmatism (Groome et al., 2003; Madsen et al., 2006), geochemically anomalous backarc magmatism (Ewing, 1980; Breitsprecher et al., 2003; Ickert et al., 2009; Dostal and Jutras, 2022), and disruption of non-marine basins (Eddy et al., 2016a). The collision of Siletzia, which started by 53 Ma in southwestern Oregon (Wells et al., 2014) and by 51 Ma in northern Washington and southernmost Vancouver Island, led to a major change in plate geometries and profound changes in the upper plate of the system from 52 to 48 Ma, which we describe in more detail below. The plate boundary later shifted outboard (west) of Siletzia, resulting in the new Cascadia subduction system at ca. 45–40 Ma (e.g., Wells et al., 1984, 2014; Schmandt and Humphreys, 2011; McCrory and Wilson, 2013; Eddy et al., 2017a; Kant et al., 2018).

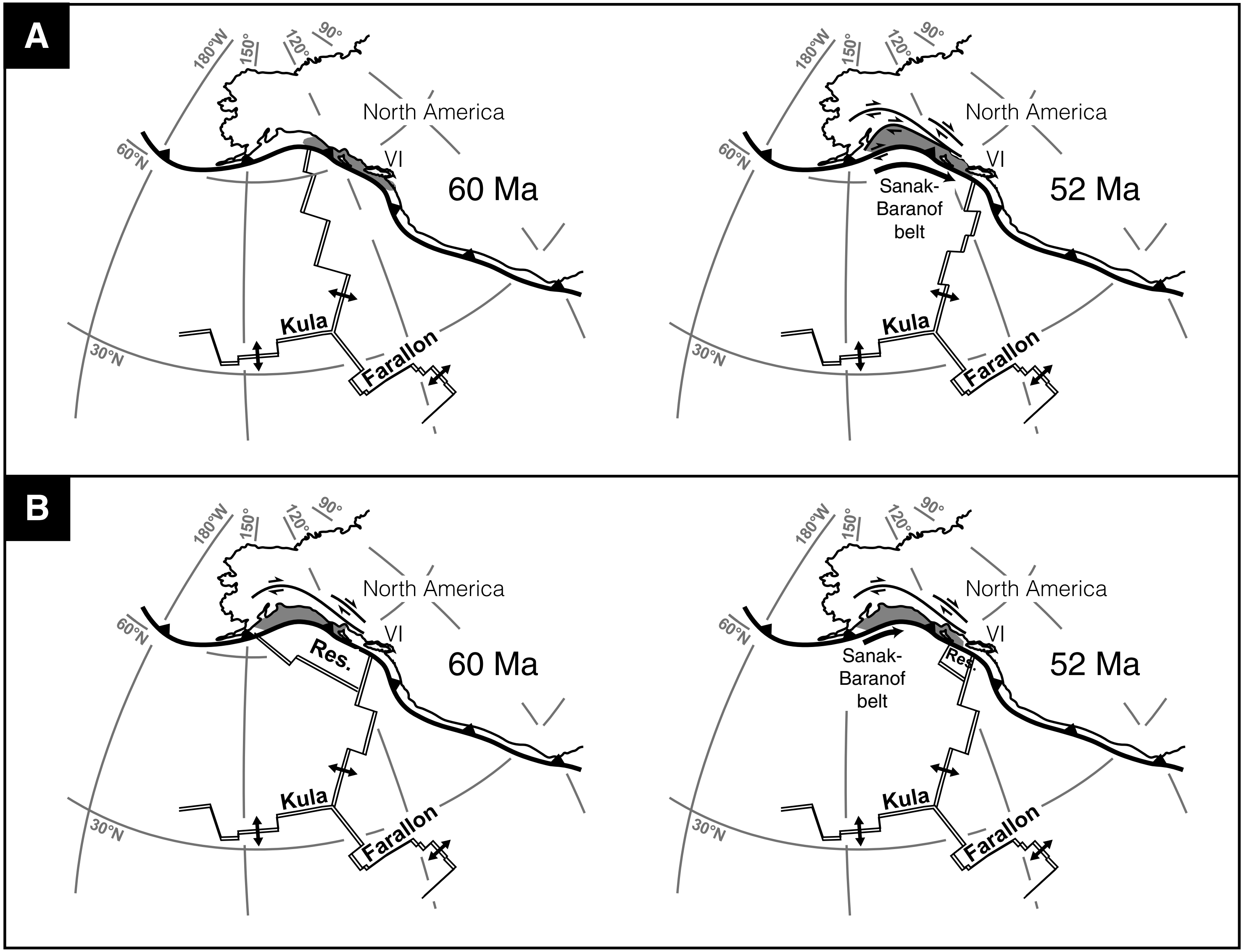

Figure 1. Simple plate reconstruction models of the northeastern Pacific at 60 Ma and 52 Ma showing locations of triple junctions. (A) Kula–Farallon–North America triple junction. Note the southward sweep of the Kula-Farallon ridge from 60 Ma to 52 Ma in this model (e.g., Bradley et al., 2003). (B) Two triple junctions result from the hypothetical Resurrection plate (Res.) (e.g., Haeussler et al., 2003). Note that in either model, there is a triple junction near central to southern Vancouver Island at ca. 52 Ma (e.g., Breitsprecher et al., 2003) and that the Kula-Farallon ridge interacted with the continental margin back to ca. 83 Ma (e.g., Engebretson et al., 1985; Thorkelson and Taylor, 1989). The hypothetical Orcas plate model is on a coarser scale and is not shown; it calls for the final consumption of the plate at ca. 50 Ma (Clennett et al., 2020). Sanak-Baranof (gray shaded area) is a belt of near-trench intrusions, which provide part of the evidence of a ridge interacting with a trench (e.g., Bradley et al., 2003). VI—Vancouver Island.

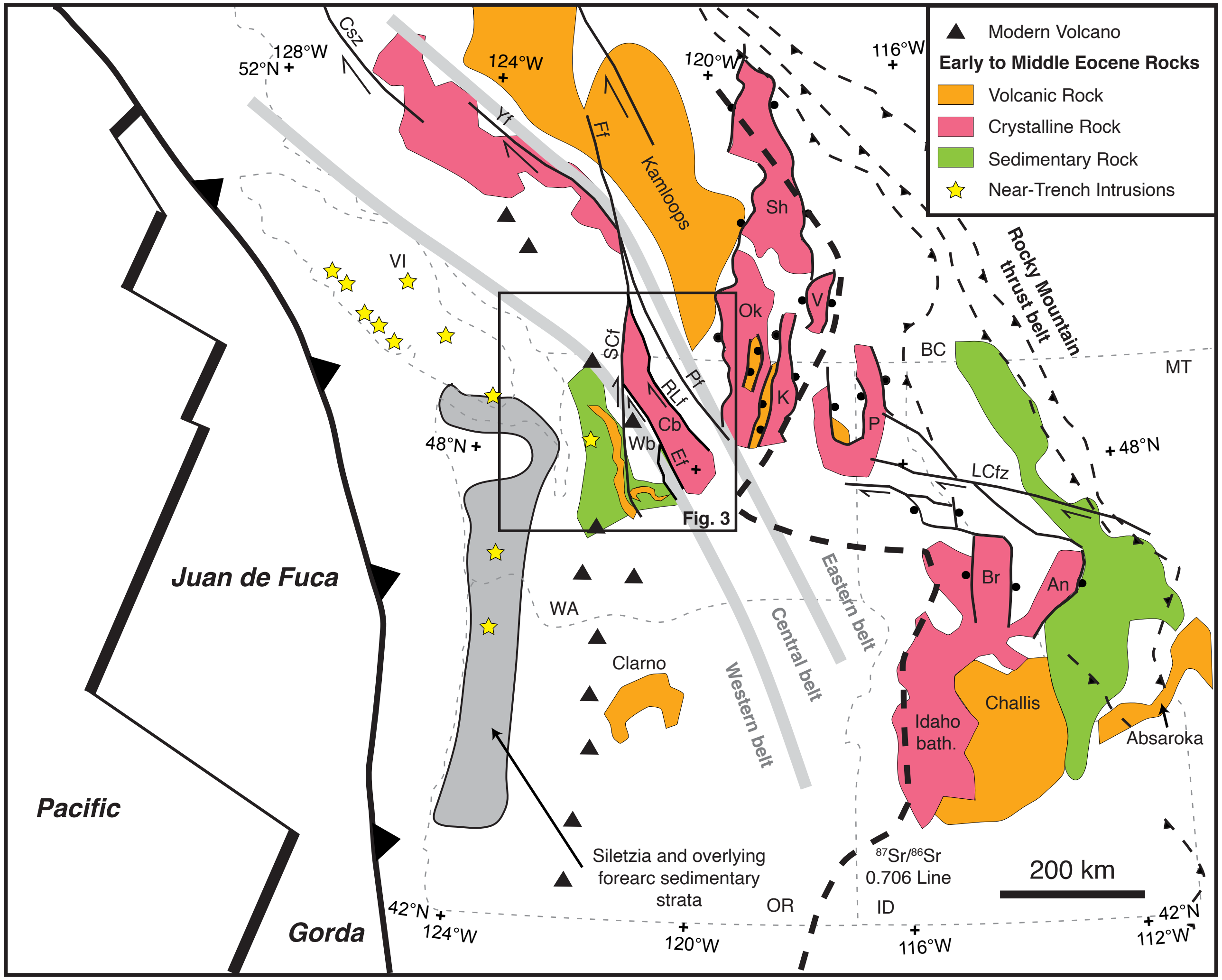

Figure 2. Generalized tectonic map of Paleogene rock types, structures, and tectonics of the greater Pacific Northwest region considered in this study. Note the location of Siletzia (including subsurface), near-trench intrusions, major dextral strike-slip faults, basins, magmatic rocks, and metamorphic core complexes and bounding normal faults. Western, central, and eastern belts are subdivisions used in text. An—Anaconda core complex; Br—Bitterroot lobe of Idaho batholith; Cb—Chelan block of North Cascades crystalline core; Csz—Coast shear zone; Ef—Entiat fault; Ff—Fraser River fault; K—Kettle core complex; LCfz—Lewis and Clark fault zone; Ok—Okanogan core complex; P—Priest River core complex; Pf—Pasayten fault; RLf—Ross Lake fault; SCf—Straight Creek fault; Sh—Shuswap core complex; V—Valhalla complex; VI—Vancouver Island; Wb—Wenatchee block of North Cascades crystalline core; Yf—Yalakom fault. Box shows location of Figure 3. States and provinces: BC—British Columbia; ID—Idaho; MT—Montana; OR—Oregon; WA—Washington.

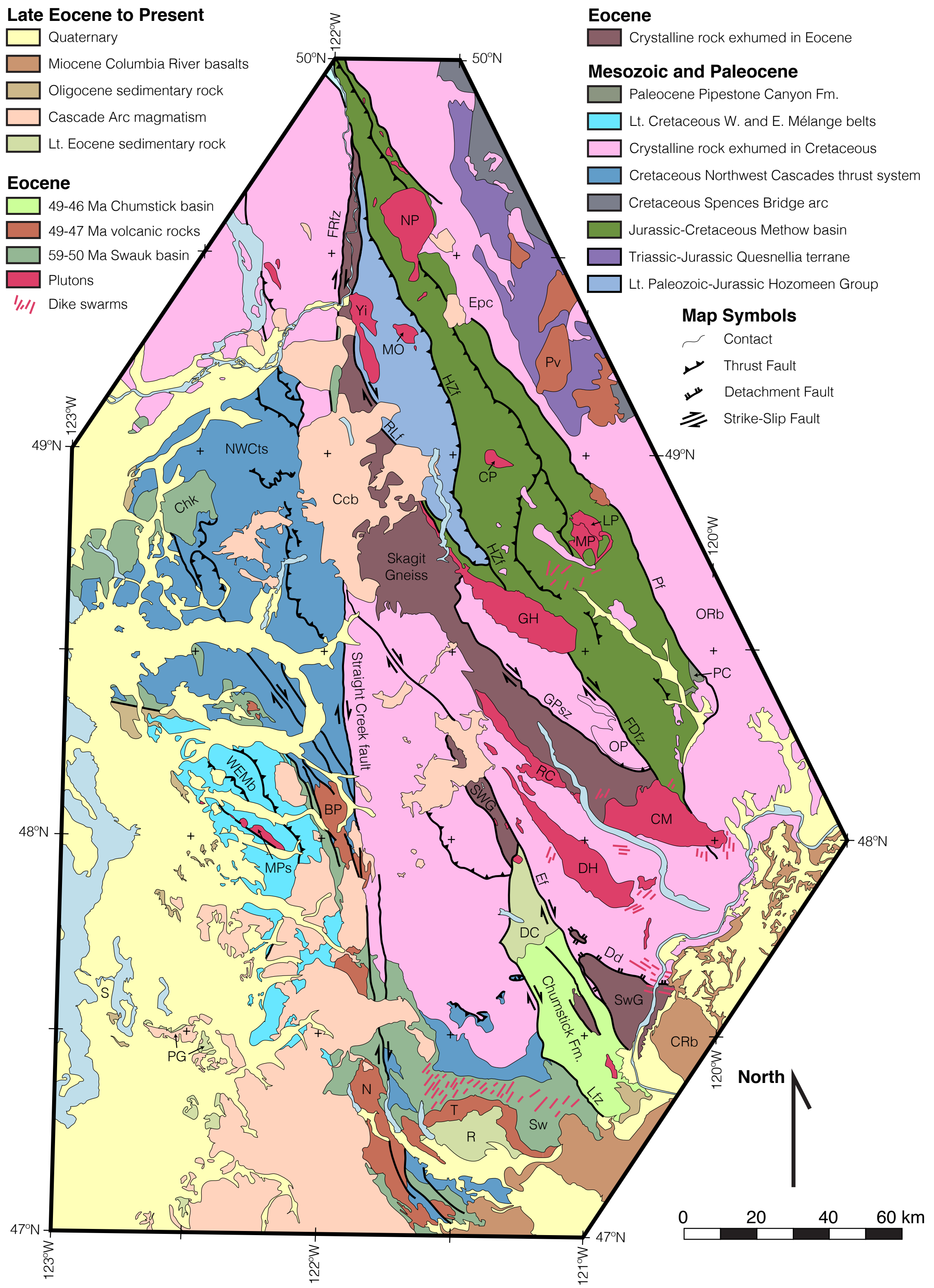

Figure 3. Simplified geologic map of central and northern Washington State and adjacent southern British Columbia (modified from Eddy et al., 2016a). BP—Barlow Pass Volcanics; Ccb—Chilliwack batholith; CHk—Chuckanut Formation; CM—Cooper Mountain pluton; CP—Castle Peak stock; CRb—Columbia River Basalt Group; DC—Deadhorse Canyon unit; Dd—Dinkelman décollement; DH—Duncan Hill pluton; Ef—Entiat fault; Epc—Eagle Plutonic Complex; FDfz—Foggy Dew fault zone; FRfz—Fraser River fault zone; GH—Golden Horn batholith; GPsz—Gabriel Peak shear zone; HZf—Hozameen fault; Lfz—Leavenworth fault zone; LP—Lost Peak stock; MO—Mount Outram pluton; MP—Monument Peak stock; MPs—Mount Pilchuck stock; N—Naches Formation; NP—Needle Peak pluton; NWCts—Northwest Cascades thrust system; OP—Oval Peak pluton; Orb—Okanogan Range batholith; PC—Pipestone Canyon Formation; Pf—Pasayten fault; PG—Puget Group; Pv—Princeton volcanics; R—Roslyn Formation; RC—Railroad Creek pluton; RLf—Ross Lake fault; S—Seattle; SW—Swauk Formation; SWG—Swakane Biotite Gneiss; T—Teanaway Formation; WEMb—western and eastern mélange belts; Yi—Yale intrusions.

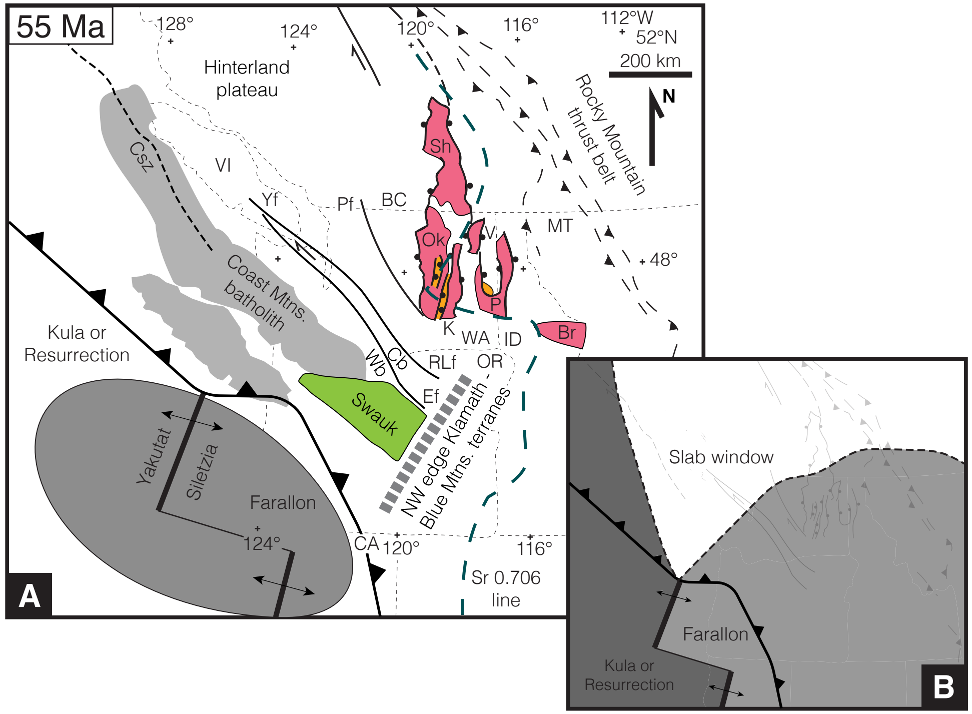

Figure 4. (A) Reconstruction map at ca. 55 Ma based on features in Figures 2 and 3 (see text for details). Note that the western belt has been offset 150 km to the south relative to the central zone and >300 km to the south relative to the eastern belt. The ridge on which Siletzia formed is near central to southern Vancouver Island, and the Swauk Basin has formed inboard of Siletzia in the western belt. There is a lull in magmatism in the southern part of the Coast Mountains and North Cascades arc. Green—Paleocene–Eocene basin; orange—Eocene extensional basin; pink—metamorphic core complex. (B) Proposed plate tectonic setting at ca. 55 Ma based on features in Figures 2 and 3. Note the position of major faults (both active and non-active) in gray for reference. Br—Bitterroot lobe of Idaho batholith; Cb—Chelan block of North Cascades crystalline core; Csz—Coast shear zone; Ef—Entiat fault; K—Kettle core complex; Ok—Okanogan core complex; P—Priest River core complex; Pf—Pasayten fault; RLf—Ross Lake fault; Sh—Shuswap core complex; V—Valhalla complex; VI—Vancouver Island; Wb—Wenatchee block of North Cascades crystalline core; Yf—Yalakom fault. States and provinces: BC—British Columbia; MT—Montana; WA—Washington; OR—Oregon; ID—Idaho; CA—California.

PRE-PALEOGENE GEOLOGIC SETTING

Prior to 65 Ma, the Pacific Northwest was characterized by a typical convergent margin with a forearc, continental magmatic arc, backarc basin, and fold-and-thrust belt that deformed a Paleozoic passive-margin sequence (e.g., Burchfiel et al., 1992). The arc and forearc were originally farther south relative to the inboard rocks by more than 300 km (e.g., Umhoefer and Blakey, 2006; Wyld et al., 2006) and potentially a much greater distance as discussed below.

In the forearc (western belt of Fig. 2) are Paleozoic and Mesozoic island arc and oceanic rocks and overlapping Jura-Cretaceous marine clastic rocks, which were deformed in the mid-Cretaceous Northwest Cascades thrust system (shown as a single Cretaceous unit on Fig. 3) (Misch, 1966; Brown, 1987; Brandon et al., 1988). Structurally above these rocks are mostly Jura-Cretaceous rocks of the western mélange belt (Fig. 3), which is interpreted as an accretionary complex (Tabor, 1994) and contains rocks at least as young as 72 Ma (Dragovich et al., 2014; Sauer et al., 2017a). The Upper Cretaceous to Paleocene Nanaimo Group (e.g., Mustard, 1994), exposed mostly on southern Vancouver Island, is interpreted as a foreland basin to the Northwest Cascades thrust system (Brandon et al., 1988) and has depositional ages extending from at least ca. 84 to 63 Ma (e.g., Matthews et al., 2017; Coutts et al., 2020).

The Cretaceous arc in northern Washington and southern British Columbia is represented by medium- to high-grade metamorphic and plutonic rocks in the crystalline core of the North Cascades and southern British Columbia (central belt of Fig. 2). The crystalline rocks are subdivided into the Wenatchee and Chelan blocks, which are separated by the high-angle Eocene Entiat fault and bounded to the west by the Straight Creek–Fraser River fault (Fig. 3). Magmatism in the Wenatchee block occurred from 96 to 87 Ma and most biotite Ar-Ar and K-Ar cooling ages are >60 Ma, whereas magmatism in the Chelan block ranges from 92 to 45 Ma and Eocene cooling ages are common (e.g., Walker and Brown, 1991; Matzel, 2004; Miller et al., 2009, 2016). The Chelan block also records Paleogene ductile deformation and partial melting in the highest-grade rocks of the Skagit Gneiss Complex (Gordon et al., 2010a).

Pre-Cenozoic rocks directly east of the North Cascades in the eastern belt include: the Mesozoic Methow Basin, ca. 160–105 Ma arc plutonic rocks of the Eagle Plutonic Complex and Okanogan Range batholith, ca. 105 Ma arc volcanic rocks of the Spences Bridge Group, and arc volcanic and sedimentary rocks of the Quesnellia terrane (Fig. 3) (e.g., Greig, 1992; Hurlow and Nelson, 1993). Farther east are the plutonic and metamorphic rocks of the Omineca belt (including metamorphic core complexes), the Idaho batholith, and Cordilleran passive-margin sediments involved in the Rocky Mountain–Sevier fold-and-thrust belt (Fig. 2).

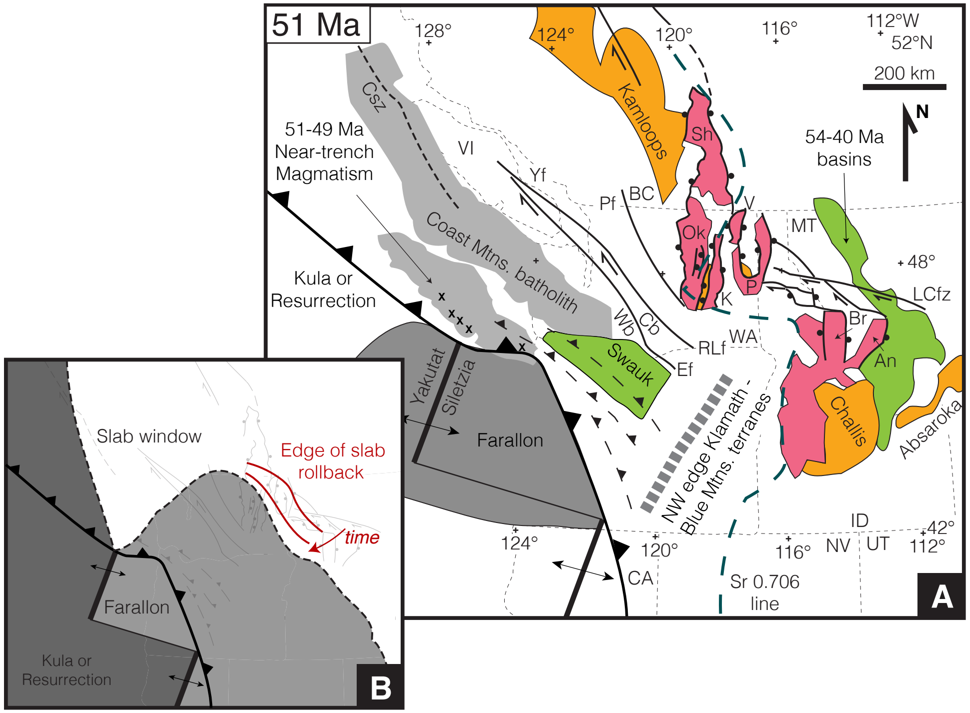

Figure 5. (A) Reconstruction map at ca. 51 Ma based on features in Figures 2 and 3. By 51 Ma, the ridge has reached Vancouver Island, and Siletzia has collided with the continental margin. The Swauk Basin has begun to invert, and Challis-Kamloops magmatism is active, as are core complexes and extensional basins in the eastern zone. (B) Proposed plate tectonic setting. Note the position of major faults (both active and non-active) in gray for reference. Green—Paleocene–Eocene basin; orange—Eocene volcanic rock; pink—metamorphic core complex. An—Anaconda core complex; Br—Bitterroot core complex; Cb—Chelan block of North Cascades crystalline core; Csz—Coast shear zone; Ef—Entiat fault; K—Kettle core complex; LCfz—Lewis and Clark fault zone; Ok—Okanogan core complex; P—Priest River core complex; Pf—Pasayten fault; RLf—Ross Lake fault; Sh—Shuswap core complex; V—Valhalla complex; VI—Vancouver Island; Wb—Wenatchee block of North Cascades crystalline core; Yf—Yalakom fault. States and provinces: BC—British Columbia; CA—California; ID—Idaho; MT—Montana; NV—Nevada; UT—Utah; WA—Washington.

RESTORATION OF STRIKE-SLIP FAULTS

Dextral strike-slip faulting occurred in the North American Cordillera in the Late Cretaceous to Eocene (e.g., Gabrielse, 1985; Wyld et al., 2006), and within our study region, displacements of ~325 km on strike-slip faults active from ca. 60 to 35 Ma are well documented (Table 1). In the west, the N-S–striking Straight Creek–Fraser River fault separates the North Cascades crystalline core of the central belt from the outboard Paleozoic and Mesozoic Northwest Cascades thrust system, mélange belts, and Paleogene rocks of the western belt (Fig. 3). The most recent estimate of dextral offset on this fault is ~150 km (Monger and Brown, 2016). The Leavenworth and Entiat faults (Fig. 3) involve the Cascades core and have a total displacement of ~60 km (Eddy et al., 2017b). The Entiat fault separates the Wenatchee and Chelan blocks within the core (see above), and the northeastern boundary of the Cascades core is the Ross Lake fault system (Ross Lake fault, Gabriel Peak tectonic belt, Hozameen fault, and Foggy Dew fault) (Fig. 3), which probably has ~125 km of dextral offset (Umhoefer and Miller, 1996).

These known displacements of ~325 km must be considered in tectonic restorations, particularly for before 50 Ma. To summarize, after 50 Ma there has been (1) ~150 km of offset between the western belt and Cascades core of the central belt, (2) 60 km of displacement within the core, (3) 55 km (of total 125 km since 70 Ma) of offset between the core and the eastern belt, and (4) a cumulative offset of ~265 km between the western and eastern belts (Table 1). If we assume that the strike-slip offset from 60 to 50 Ma occurred at rates comparable to those of the ca. 50–40 Ma interval, the implication is that another ~200–250 km of offset occurred across Washington from 60 to 50 Ma. About 60 km of this slip has been documented on the Ross Lake fault system (Miller and Bowring, 1990) and Yalakom fault during that time (Umhoefer and Schiarizza, 1996); precise timing and offset of faults are difficult to document. From this reasoning, at 55 Ma we show Vancouver Island and the western belt as offset ~450 km south relative to the eastern belt (Fig. 4). We note that this is likely a conservative estimate and does not include any distributed dextral ductile displacement or movement on minor cryptic structures. Paleomagnetic data indicate much larger cumulative dextral displacements between ca. 85 and 55 Ma of 2000 km or more between the easternmost part of the eastern belt and the central and western belts and ~1000 km between the western part of the eastern belt and rocks to the west (e.g., Enkin, 2006; Tikoff et al., 2023). From the paleomagnetic data, major displacements of the outboard rocks ended by 55 Ma (e.g., Cowan et al., 1997; Tikoff et al., 2023). Thus, uncertainties are much lower for the positions of units in the region in the 55 Ma and younger reconstructions (Figs. 4–7).

Another potential complication is the rotation in the Oregon Coast Ranges and Cascades, which is probably related to distributed dextral strike slip and Basin and Range extension (e.g., Beck, 1984; Wells and Heller, 1988; Colgan and Henry, 2009; Wells and McCaffrey, 2013; Wells et al., 2014). Rotation increases westward and decreases from the

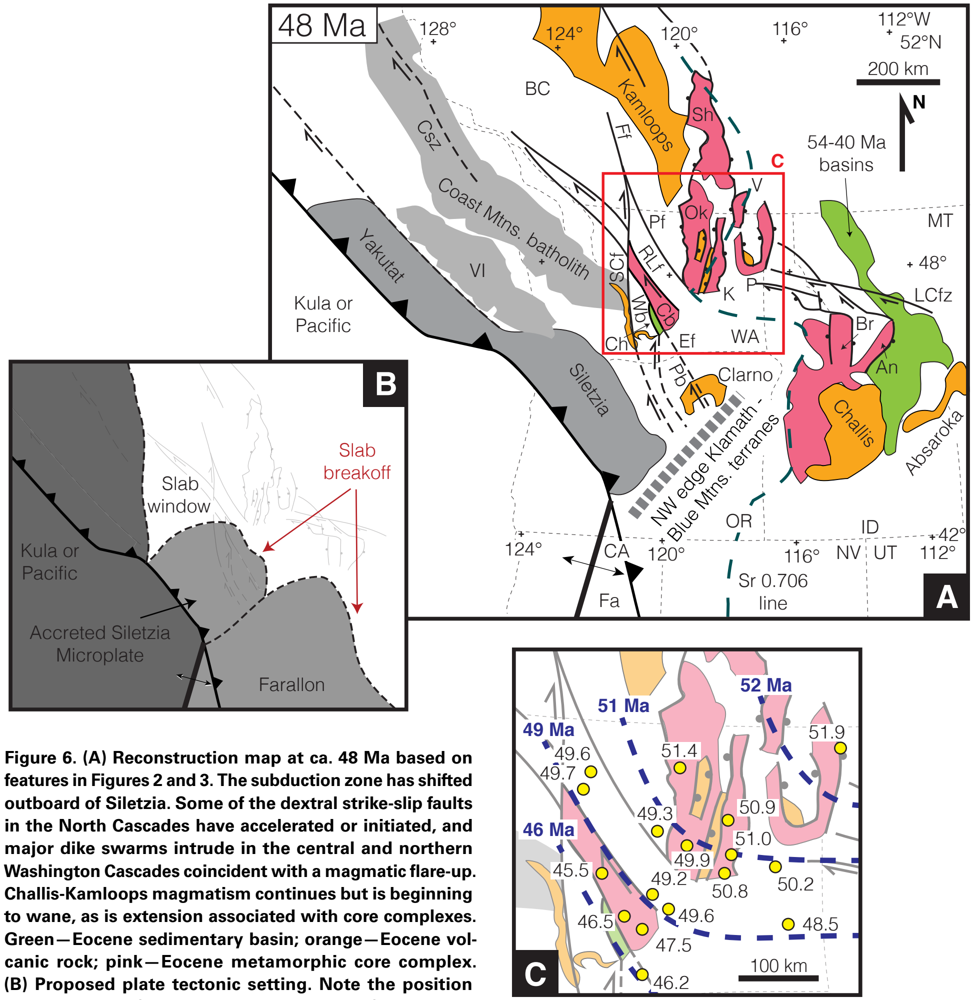

Figure 6. (A) Reconstruction map at ca. 48 Ma based on features in Figures 2 and 3. The subduction zone has shifted outboard of Siletzia. Some of the dextral strike-slip faults in the North Cascades have accelerated or initiated, and major dike swarms intrude in the central and northern Washington Cascades coincident with a magmatic flare-up. Challis-Kamloops magmatism continues but is beginning to wane, as is extension associated with core complexes. Green—Eocene sedimentary basin; orange—Eocene volcanic rock; pink—Eocene metamorphic core complex. (B) Proposed plate tectonic setting. Note the position of major faults (both active and non-active) in gray for reference. The approximate location of the idealized slab window assumes that the Farallon plate was moving northeast and the Kula-Farallon ridge intersected the continental margin as shown. (C) Eocene magmatism across northeastern to north-central Washington and the pattern of inferred rollback magmatism to the southwest. Filled circles are localities of U-Pb zircon ages (in Ma) of plutonic rocks. Heavy dashed lines are contours of the U-Pb zircon ages (references cited in text). An—Anaconda core complex; Br—Bitterroot core complex; Cb—Chelan block of North Cascades crystalline core; Ch—Chumstick Basin; Csz—Coast shear zone; Ef—Entiat fault; Fa—Farallon plate; Ff—Fraser River fault; K—Kettle core complex; LCfz—Lewis and Clark fault zone; Ok—Okanogan core complex; P—Priest River core complex; Pb—Pasco Basin; Pf—Pasayten fault; RLf—Ross Lake fault; SCf—Straight Creek fault; Sh—Shuswap core complex; V—Valhalla complex; VI—Vancouver Island; Wb—Wenatchee block of North Cascades crystalline core. States and provinces: CA—California; BC—British Columbia; ID—Idaho; MT—Montana; NV—Nevada; OR—Oregon; UT—Utah; WA—Washington.

Klamath Mountains (northwestern California and southwestern Oregon) northward to the Olympic Peninsula (Washington). Statistically significant vertical-axis rotation has not occurred after ca. 50 Ma in the Washington Cascades, at least as far south as the present latitude of Seattle (e.g., Beske et al., 1973; Beck et al., 1982; Fawcett et al., 2003). In our reconstructions, we utilize the present trends of structures in the north and restore the Klamath Mountains to northeasternmost California to account for Basin and Range extension (e.g., Colgan and Henry, 2009) and rotation. The resulting trend and position of Siletzia (Washington and Oregon Coast Ranges) in our reconstructions (Figs. 4–7) after accretion is more northerly than in Wells et al. (2014), which suggests that a portion of the rotation in Siletzia was taken up on more local blocks at a scale of a few tens of kilometers or less.

PALEOGENE TECTONIC HISTORY

In this section, we synthesize the Paleogene tectonic evolution across the Pacific Northwest (Fig. 3) and divide this ~25 m.y. history into five intervals. The time slices are generally considered from west to east. The major events from 60 to 40 Ma are summarized in Figure 8.

65–60 Ma

During the 65–60 Ma interval, the plate boundary was one of oblique convergence. This interpretation is based on the arc-type tonalitic intrusions (Miller and Bowring, 1990; Miller et al., 2009), transpressional deformation in the North Cascades and southern Coast Mountains batholith arc (e.g., Brown and Talbot, 1989; Miller and Bowring, 1990), and contractional deformation (e.g., Brown et al., 1986; Simony and Carr, 2011) in the hinterland (eastern belt).

The forearc (western belt) record is sparse, and the timing of deformation in this belt is poorly known (Tabor, 1994; Sauer et al., 2017a). The only known forearc rocks of this age are the uppermost clastic strata of the Nanaimo Group on Vancouver

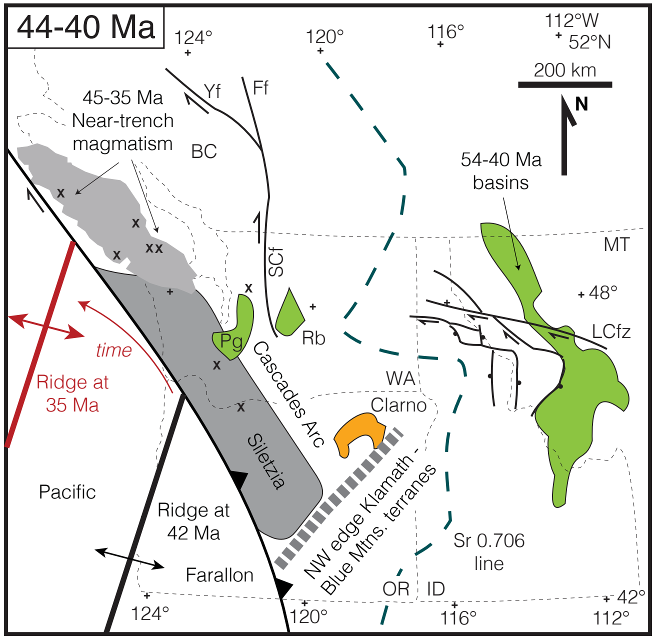

Figure 7. Reconstruction map and proposed plate tectonic setting at ca. 44–40 Ma based on features in Figures 2 and 3. The ancestral Cascades arc ("Cascadia") has initiated, scattered basins extend from the western to eastern belts, and magmatism has ended in the North Cascades and almost all of the Challis-Kamloops belt. The Farallon-Pacific ridge is migrating to the northwest relative to North America. Ff—Fraser River fault; LCfz—Lewis and Clark fault zone; Pg—Puget Group; Rb—Roslyn Basin; SCf—Straight Creek fault; Yf—Yalakom fault. States and provinces: BC—British Columbia; ID—Idaho; MT—Montana; OR—Oregon; WA—Washington.

TABLE 1. STRIKE-SLIP FAULT OFFSETS ACROSS THE WASHINGTON CASCADES

Fault	Fault offsets (km)					
	70–60 Ma	60–50 Ma	50–45 Ma	45–40 Ma	40–35 Ma	Totals
Straight Creek fault	–	–	50	65	35	150
Leavenworth fault	–	–	30	–	–	30
Entiat fault	?	?	10	20	–	30
Ross Lake–Hozameen–Foggy Dew faults	10	60	45	10	–	125
Total offset	10	60	135	95	35	335

Notes: Dash—no slip; question mark—no known slip.

Island, which have maximum depositional ages (MDAs) as young as ca. 63 Ma (Coutts et al., 2020). The youngest dated (MDA) sandstone in the western mélange belt is ca. 72 Ma (Sauer et al., 2017a), and younger rocks may be present in this belt given that the upper limit for the mélange is indicated only by an angular unconformity with Eocene strata.

The 65–60 Ma interval includes the final stage of a magmatic flare-up in the North Cascades core (Chelan block) that began ca. 78 Ma (Miller et al., 2009) and was directly preceded by rapid burial and metamorphism of Cretaceous (protolith age) metasedimentary rocks that comprise the deep-crustal (metamorphic pressure as high as 12 kbar) Swakane Biotite Gneiss (Valley et al., 2003) and Skagit Gneiss Complex (7–10 kbar; Whitney, 1992; Hanson et al., 2022) (Fig. 3) between ca. 79 and 66 Ma and between 74 and 65 Ma, respectively (Sauer et al., 2017b, 2018). Tonalitic magmatism is recorded by the 65 Ma Oval Peak pluton (Fig. 3), which crystallized at 5–6 kbar (Miller and Bowring, 1990), and sheets (now orthogneisses) in the Skagit Gneiss Complex (Miller et al., 2016). Leucosomes of this age also are recognized in the Skagit Gneiss Complex (Gordon et al., 2010a). K-Ar and Ar-Ar biotite cooling ages are sparse, but there is no evidence for major rapid cooling or exhumation of the Cascades core during this interval (Paterson et al., 2004), and no sedimentary or volcanic rocks of this age have been recognized in the arc. Dated deformation during this time interval is limited in the arc region where dextral and reverse shear in the Gabriel Peak tectonic belt of the Ross Lake fault system (Fig. 3) was inferred to be coeval with emplacement of the Oval Peak pluton (Miller and Bowring, 1990).

In the eastern belt, igneous activity was sparse during this interval, and volcanic rocks are absent. In northeastern Washington, magmatism was limited to a few ca. 64–56 Ma plutons (e.g., Stoffel et al., 1991). North of the international border, intrusion of the quartz monzonitic to granitic, peraluminous Ladybird granite suite into high-grade Shuswap Metamorphic Complex (Shuswap core complex) (Fig. 4) initiated at 62 Ma (Carr, 1992; Hinchey and Carr, 2006). In Idaho (USA), peraluminous magmatism in the Bitterroot lobe (Fig. 4) of the Idaho batholith began at ca. 66 Ma and peaked at ca. 60 Ma (Gaschnig et al., 2010). These peraluminous rocks are part of the "Cordilleran anatectic belt" of Chapman et al. (2021a), and the magmatism is ascribed to partial melting of crustal rocks (Mueller et al., 1996; Hinchey and Carr, 2006; Gaschnig et al., 2011).

Sedimentary rocks of this age are also very rare in northeastern Washington. Aside from a <30 km^2 body of Paleocene conglomerate (Pipestone Canyon Formation) directly west of the Pasayten fault (Fig. 3) (Kriens et al., 1995), no other strata have been recognized between central Washington and the Sevier foreland basins. The scarcity of sedimentary rocks and the evidence of crustal melting are compatible with the existence of a high-standing orogenic plateau in the hinterland during this interval (Whitney et al., 2004; Bao et al., 2014). Thrusting also occurred in the eastern belt in the Shuswap

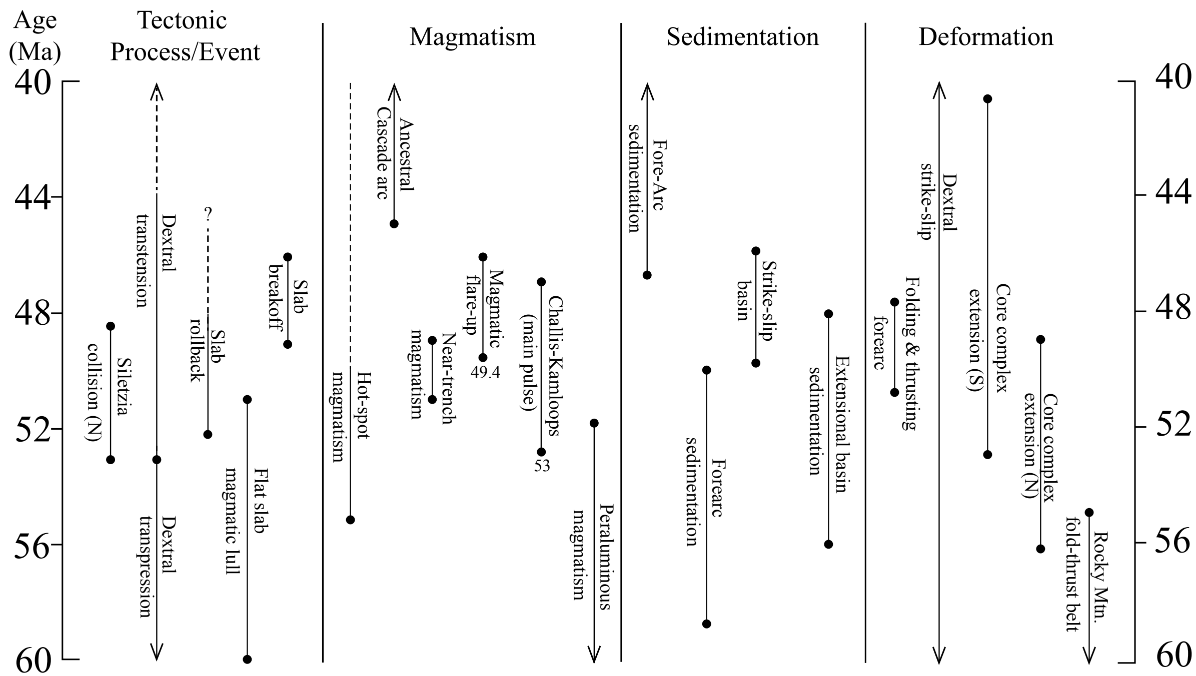

Figure 8. Summary of timing of major events described in the text. Within the magmatism, sedimentation, and deformation panels, features are generally arranged from west (left) to east (right). Arrows designate where processes began before 60 Ma or ended after 40 Ma.

Complex and in the Rocky Mountain–Sevier fold-and-thrust belt (e.g., Price, 1981).

60–52 Ma

The 60–52 Ma interval is marked by major changes in magmatism and sedimentation throughout the region. Near-trench intrusions strongly suggest that an oceanic spreading center lay off central to southern Vancouver Island by 52–51 Ma (Fig. 5) (Groome et al., 2003; Madsen et al., 2006). Magmatism and sedimentation occurred in the western belt near the spreading ridge, but igneous activity was nearly absent in the Cascades core and eastern belt until the onset of Challis-Kamloops magmatism at ca. 53 Ma (e.g., Ickert et al., 2009). The formation of metamorphic core complexes and associated basins in the eastern region also started at ca. 56 Ma (e.g., Brown et al., 2012).

Basaltic magmatism began in the Siletzia terrane by ca. 55 Ma in the south (southwestern Oregon) and by 53.2 Ma outboard of the Northwest Cascades thrust system and mélange belts in western Washington and Vancouver Island in the north, where it continued until at least 48 Ma (Crescent Formation and Metchosin igneous complex) (Fig. 2) (Wells et al., 2014; Eddy et al., 2017a). Siletzia consists of thick sequences of basalt that transition from deep-water lava flows of normal mid-ocean-ridge basalt (N-MORB) to shallow-water and subaerial flows of enriched mid-ocean-ridge basalt (E-MORB) and oceanic-island basalt (OIB) (e.g., Wells et al., 2014). Siletzia is comparable in volume to other large igneous provinces (Trehu et al., 1994; Wells et al., 2014), and this, combined with isotopic evidence, supports its formation over a "plume-like" mantle source, thought to be the Yellowstone hotspot (e.g., Pyle et al., 2015; Phillips et al., 2017; Stern and Dumitru, 2019; Camp and Wells, 2021). In southern Oregon, the submarine basalts were overlain by deep-water sediments (Umpqua Group) in this time interval (Wells et al., 2014), while in Washington, sedimentation was initiated in the non-marine Chuckanut and Swauk Formations of the greater Swauk Basin (Fig. 3) (Eddy et al., 2016a). This basin developed on accreted Paleozoic and Mesozoic rocks of the Northwest Cascades thrust system and the southern end of the Cascades core. A 56.8 Ma tuff from the lower part of the Chuckanut Formation and a 59.9 Ma MDA near the base of the Swauk Formation are compatible with sedimentation in the greater Swauk Basin starting at 60–57 Ma (Eddy et al., 2016a). The 56.8 Ma tuff, a 53.7 Ma tuff higher in the Chuckanut section (Breedlovestrout et al., 2013), and a 53.7 Ma tuff with arc affinities (Summit Creek section; Kant et al., 2018) in the southern Washington Cascades are the only record of volcanism inboard of Siletzia in the western belt. There is also no well-documented deformation between 60 and 52 Ma, although a local angular unconformity in the middle to lower part of the Swauk Formation may be a link to the early collision of Siletzia (Doran, 2009).

In the North Cascades core, a magmatic lull began at ca. 60 Ma (Miller et al., 2009), and that lull extended into the southern Coast Mountains to the northwest (Cecil et al., 2018). The transpressional Gabriel Peak belt (Fig. 3) of the Ross Lake fault system continued to be active between at least 60 and 55(?) Ma and was cut by the transtensional Foggy Dew fault zone of the Ross Lake system at ca. 55–53 Ma (Miller and Bowring, 1990). Ductile deformation probably occurred in domains in the Skagit Gneiss Complex, but otherwise, deformation is not well documented.

In northeastern Washington, magmatism is represented only by scattered, small-volume intrusions until ca. 53 Ma, while small mafic bodies began intruding the Idaho batholith region at ca. 58 Ma (Foster and Fanning, 1997; Gaschnig et al., 2010). Peraluminous magmatism (Ladybird granite suite), metamorphism, and migmatization continued during the 60–52 Ma time interval in the Shuswap and Okanogan complexes (e.g., Crowley et al., 2001; Kruckenberg et al., 2008; Gervais et al., 2010; Brown et al., 2012), and peraluminous magmatism persisted in the Bitterroot lobe of the Idaho batholith until ca. 53 Ma (Gaschnig et al., 2010) and the Anaconda core complex of Montana (USA) until ca. 56 Ma (e.g., Howlett et al., 2021). This magmatism in Idaho was directly followed by the Challis magmatic event (ca. 53–43 Ma; e.g., Janecke and Snee, 1993; Ickert et al., 2009; Gaschnig et al., 2010), which extended from Oregon to South Dakota and Washington and into central British Columbia as the Kamloops belt (Figs. 5 and 6) (e.g., Ewing, 1980; Breitsprecher et al., 2003). Shallow plutons, dikes, and volcanic rocks characterize this magmatic event, with geochemical affinities ranging from arc to within plate, and some rocks being almost entirely crustal melts and others only weakly contaminated melts of the lithospheric mantle (Ewing, 1980; Thorkelson and Taylor, 1989; Lewis and Kiilsgaard, 1991; Morris et al., 2000; Breitsprecher et al., 2003; Ickert et al., 2009; Dostal and Jutras, 2022). The alkalinity of magmas increases markedly south of ~51.5°N, and the width of the belt widens south of the international border (e.g., Breitsprecher et al., 2003).

In the eastern belt, ductile deformation and thrusting continued in the hinterland of the Rocky Mountain fold-and-thrust belt for the early part of this interval (e.g., Simony and Carr, 2011). A major transition from contraction to extension, which was time transgressive (e.g., Harlan et al., 1988; Parrish et al., 1988; Brown et al., 2012), led to the formation of metamorphic core complexes and associated extensional basins in northeastern Washington, British Columbia, Idaho, and Montana (Fig. 4). Core complexes (e.g., Priest River, Okanogan) and associated basins initiated earlier north of the WNW-striking Lewis and Clark fault zone than to the south (Anaconda, Bitterroot) (Foster et al., 2007). Sedimentary basin formation initiated from ca. 56 Ma next to the Okanogan core complex directly east of the North Cascades to ca. 53 Ma adjacent to the Bitterroot and Anaconda core complexes (e.g., Foster et al., 2007; Howlett et al., 2021) and in northeastern Washington, continued to 48 Ma (Pearson and Obradovich, 1977; Suydam and Gaylord, 1997). The absence of sedimentary deposits between the Swauk Basin in the west and the foreland basin east of the thrust belt until extension began and basins formed suggests that the hinterland region continued to be a high orogenic plateau until ca. 55 Ma (Whitney et al., 2004; Bao et al., 2014).

We postulate that the near-complete termination of arc-type magmatism in the North Cascades core and southern Coast Mountains and the paucity of magmatism east of there records a change to low-angle subduction of the Farallon plate at ca. 60 Ma. The peraluminous magmatism in the east probably resulted mainly from concentrated crustal thickening (e.g., Gaschnig et al., 2010).

In the eastern belt, the shift to shallow, widespread, and diverse magmatism at ca. 53 Ma accompanied by extension points to a major change from the earlier peraluminous magmatism. This shift marks the onset of Challis activity and is discussed in more detail in the next section.

52–49.5 Ma

A fundamental change in plate boundary stresses occurred between 52 and 49.5 Ma, as Siletzia encountered the subduction zone in southern Oregon. Collision progressed northward during this time interval from Oregon to Washington and southern Vancouver Island (Wells et al., 2014). This collision was coincident with major changes in magmatism, sedimentation, and the strain field in the upper plate. The Siletzia collision also ultimately led to a westward shift in the location of the plate boundary (e.g., Schmandt and Humphreys, 2011).

The Siletzia collision was accompanied from central Vancouver Island to northwestern Washington by near-trench magmatism from ca. 51 to 49 Ma (Madsen et al., 2006), which is thought to record the location of a subducting spreading ridge and the Kula–Farallon–North America or Resurrection–Farallon–North America triple junction (Fig. 5) (e.g., Cowan, 2003; Groome et al., 2003; Haeussler et al., 2003; Madsen et al., 2006) that would have been the northern boundary of Siletzia (Wells et al., 2014). This inference is also consistent with the 51 Ma age of the ophiolitic Metchosin Igneous Complex on southern Vancouver Island (Massey, 1986, Eddy et al., 2017a). Near-trench magmatic rocks on Vancouver Island include: 51.2–50.5 Ma bimodal but dominantly dacitic rocks (Flores Volcanics) (Irving and Brandon, 1990); 51.2–48.8 Ma hypabyssal tonalite, trondhjemite, and granodiorite (Clayoquat intrusions) (Madsen et al., 2006); and in the south, peraluminous 50.9–50.7 Ma intrusions (Walker Creek intrusions) (Groome et al., 2003). The Leech River Schist on southern Vancouver Island also records high-temperature, low-pressure metamorphism at ca. 51 Ma (Fairchild and Cowan, 1982; Groome et al., 2003). In northwestern Washington, local peraluminous magmatism occurred as the ca. 49 Ma Mount Pilchuck stock (Fig. 3) and nearby Bald Mountain pluton (Yeats and Engels, 1971).

Farther inboard, but still west of the Cascades core, basaltic to rhyolitic volcanism began with the eruption of 51.4 Ma lavas and tuffs (Silver Pass Volcanic Member) of the upper Swauk Formation (Peterson and Tepper, 2021) and 51.3 Ma dacitic to rhyolitic lavas and pyroclastic rocks (Taneum Formation), which overlie clastic rocks correlative with the Swauk Formation (Fig. 3) (Tabor et al., 1984; Eddy et al., 2016a; Wallenbrock and Tepper, 2017). These units represent the initiation of a magmatic belt that roughly parallels the leading edge of subducted Siletzia in the subsurface (Fig. 2) (Wells et al., 2014) and is attributed to tearing of the Farallon slab (Kant et al., 2018).

The approach and collision of Siletzia is also recorded in folding and changes in paleotopography in the western belt. Sedimentation in the Swauk Basin persisted until at least ca. 50.8 Ma, the youngest MDA from stratigraphically high in the basin (Eddy et al., 2016a; Senes, 2019), but a drainage reversal from SW- to NE-flowing streams occurred at ca. 51 Ma (Eddy et al., 2016a) and may record the initial stages of collision of Siletzia at the latitude of the Swauk Basin. A NW-vergent fold-and-thrust belt developed in southwestern Oregon in response to collision and involved Siletzia basalts, overlying Umpqua Group, and Klamath basement terranes. Unconformably overlying marine strata (Tyee Formation) demonstrate that accretion was completed between 50.5 and 49 Ma at that latitude (Wells et al., 2000, 2014). In the central Washington Cascades, the Swauk Formation is folded and locally faulted under a short-lived (<1.5 m.y.) angular unconformity with the overlying Teanaway Formation (Foster, 1958). The Teanaway Formation includes a 49.3 Ma rhyolite near its base (Eddy et al., 2016a) and is dominated by subaerial basalts, in contrast to the marine strata in southwestern Oregon. Contractional structures also attributed to the accretion of Siletzia are folds in the Chuckanut Formation in the northwestern Swauk Basin (Misch, 1966; Johnson, 1984), some of the upright folds in the Skagit Gneiss Complex of the North Cascades core (Miller et al., 2016), and the Cowichan fold-and-thrust belt on Vancouver Island, which is approximately the same age and has a similar northwesterly trend as the Chuckanut folds (Fig. 5) (Johnston and Acton, 2003).

The magmatic lull continued in the North Cascades core (Miller et al., 2009), although minor partial melting persisted in the Skagit Gneiss Complex (Gordon et al., 2010a). The deep-crustal (9–12 kbar) Swakane Biotite Gneiss in the crystalline core was probably rapidly exhumed during this interval, in part during distributed ductile shear and top-to-the-N to top-to-the-NNE motion on the Dinkelman décollement (Fig. 3) (Paterson et al., 2004). Dextral-normal slip and associated mylonitization continued in the Foggy Dew fault zone, a southern strand of the Ross Lake fault system, and dextral displacement also occurred on the NW-striking Yalakom fault and other faults west of the Straight Creek–Fraser River fault (Fig. 5) (Miller and Bowring, 1990; Umhoefer and Schiarizza, 1996).

East of the Cascades core, magmatism increased with the emplacement of granitoid plutons and dominantly metaluminous tonalites and granodiorites. Although arc-like in mineralogy, many of these plutons have trace element traits compatible with slab-breakoff magmas (e.g., Sr/Y >10, $(La/Yb)_N$ >10; Whalen and Hildebrand, 2019) and Sr-Nd isotopic compositions indicative of significant contributions from older crust (Tepper and Eddy, 2017). The earliest U-Pb date associated with this renewed activity is 52 Ma in central Idaho, and subsequent plutonism appears to have migrated to the southwest across northeastern Washington (Fig. 6C) (Tepper et al., 2023). Metamorphism and deformation continued in the metamorphic core complexes in southern British Columbia, northeastern Washington, Idaho, and Montana, as did Challis-Kamloops magmatism and sedimentation in extensional basins where MDAs of locally derived sediments cluster around 50 Ma in southern British Columbia and northeastern Washington (e.g., Ewing, 1980; Suydam and Gaylord, 1997; Foster et al., 2007; Brown et al., 2012; Rubino et al., 2021). In contrast to northeastern Washington, no pattern of magmatism migration is seen across the Challis to Absaroka area in Idaho and Wyoming (e.g., Feeley and Cosca, 2003). The thermal peak in the Shuswap metamorphism was at ca. 53–49 Ma (Crowley et al., 2001).

Deformation in the eastern belt was dominated by roughly E-W extension, although contraction may have continued at deep levels in the Shuswap metamorphic core complex until ca. 52–49 Ma (Crowley et al., 2001; Gervais et al., 2010; Gervais and Brown, 2011). The peak of extension and exhumation in the Okanogan core complex occurred at 53–50 Ma (Brown et al., 2012). Brittle slip of uncertain sense reactivated the high-angle, ≥250-km-long Pasayten fault (Fig. 3) along the eastern boundary of the Methow Basin and ended in Washington before eruption of ca. 48 Ma volcanic rocks, which overlap the fault (White, 1986).

In summary, the transition from a low-angle, transpressional subduction regime to a dextral transtensional regime was largely complete by the end of this time interval. The collision of Siletzia explains the deformation in the Swauk Basin and along strike to the northwest, and the southwestward migration of magmatism in northeastern Washington is consistent with rollback of the northern Farallon slab (Figs. 5 and 6C). The slab ruptured west of the Cascades core and is marked in part by a belt of magmatism that started at the end of this time period and lasted until ca. 48 Ma (Kant et al., 2018) (Fig. 6). Previous explanations for this Challis-Kamloops magmatism include a decrease in the rate of plate convergence (Constenius, 1996), passage of a slab window (Thorkelson and Taylor, 1989; Breitsprecher, et al., 2003; Ickert et al., 2009), buckling and "sideways" slab rollback (Humphreys, 1995, 2009), and rollback and breakoff of the Farallon slab (Tepper et al., 2023). Slab rollback and breakoff and slab window evolution are the most widely cited scenarios (see review by Humphreys and Grunder, 2022).

49.5–45 Ma

The short-lived deformation episode resulting from the collision of Siletzia was followed by profound changes in the tectonic evolution of the Pacific Northwest. A new subduction zone of the Kula or Resurrection plate beneath North America was established along the west side of Siletzia during the 49.5–45 Ma time interval (Fig. 6) (e.g., Schmandt and Humphreys, 2011). A dextral transtensional regime dominated, and a new non-marine strike-slip basin formed next to the Cascades core (Fig. 6). A magmatic flare-up occurred in the Cascades core and in the adjacent parts of the western belt, and magmatism and extension continued in the eastern belt but were more areally restricted after ca. 48 Ma.

In the west, the effects of the collision of Siletzia were waning by this time as magmatism ended in the southern part of Siletzia at ca. 50–49 Ma (Wells et al., 2014) and in northern Siletzia at ca. 48 Ma (Eddy et al., 2017a). The collision was followed in the Olympic Mountains (northern Siletzia) by deposition of turbidites (Blue Mountain unit) that have maximum depositional ages ranging from 47.8 to 44.7 Ma (Eddy et al., 2017a).

To the east of Siletzia, magmatism attributed to slab rollback, tear, and breakoff continued until ca. 45 Ma, producing compositionally diverse volcanic and plutonic rocks that in part formed parallel to the edge of Siletzia in the subsurface and are commonly near the Straight Creek fault and its splays (Fig. 6) (Trehu et al., 1994; Kant et al., 2018). Distinctive traits of these rocks include their bimodal nature, with OIB affinities of the mafic lavas and crustal signatures of the silicic rocks. On the west side of the Straight Creek fault are basalt and lesser rhyolite flows interbedded with non-marine sedimentary rocks in the Naches Formation and Barlow Pass Volcanics (Fig. 3). East of the Straight Creek fault, the prolific Teanaway dike swarm intruded the deformed rocks of the Swauk Basin (Fig. 3) (Tabor et al., 1984; Miller et al., 2022) and is interpreted to be related to the dominantly basaltic, ca. 49.3 Ma Teanaway Formation. The mafic rocks are medium-K tholeiitic basalts and basaltic andesites (Clayton, 1973; Peters and Tepper, 2006; Roepke et al., 2013), which are derived from mantle that is inferred to have been metasomatized during earlier subduction (Tepper et al., 2008). The NNE (035°) average orientation of the dikes provides the most robust evidence for initiation of right-lateral strike-slip on the Straight Creek fault at ca. 49 Ma (e.g., Miller et al., 2022).

Starting at 49.2 Ma, the Chumstick Basin formed between the right-stepping Leavenworth and Entiat strike-slip faults, directly west of the Chelan block (Evans, 1994; Eddy et al., 2016a) (Fig. 3). Abundant stratigraphic, paleocurrent, and detrital geochronologic data suggest that the basin formed during strike-slip faulting (Eddy et al., 2016a; Donaghy et al., 2021). The main western subbasin formed from 49.2 to ca. 46.5 Ma, and fault reorganization at ca. 46.5–44 Ma started inversion of the western subbasin and the formation of a narrow eastern subbasin next to the Entiat fault (Fig. 3). After this reorganization, strike-slip faulting localized on the Entiat and Straight Creek faults. The youngest (<45.9 Ma) sediments of the Chumstick Formation top the Leavenworth fault and probably correlate with the arkosic Roslyn Formation, which overlies the Teanaway Formation (Evans, 1994; Eddy et al., 2016a) (Fig. 3).

The magmatic lull in the Cascades core ended at ca. 49.4 Ma, close in time to the eruption of Teanaway volcanic rocks south of the Cascades core. The ensuing short-lived (until ca. 45 Ma) flare-up has the highest magmatic addition rate and the shortest duration of the three flare-up events in the North Cascades since the mid-Cretaceous. It began with the ca. 49.6 Ma Lost Peak stock, followed by two large (~300 km^2 each) plutons, the Cooper Mountain and Golden Horn batholiths, which intruded at 49.3–47.9 Ma and 48.5–47.7 Ma (Eddy et al., 2016b; Miller et al., 2016), respectively, across the Ross Lake fault zone and into both the Cascades core and the Methow Basin (Fig. 3). These plutons and coeval variably deformed 49.4–47.2 Ma intrusions (now orthogneisses) in the Skagit Gneiss Complex are commonly granodioritic, in contrast to the mainly Cretaceous tonalitic intrusions of the two older flare-ups (e.g., Misch, 1966; Haugerud et al., 1991; Miller et al., 2009). The ca. 49–48 Ma intrusions also range from gabbro to granite and include alkaline granites. Between ca. 47.9 and 46.5 Ma, magmatism in the core migrated westward from the Ross Lake fault zone. The ca. 46.5 Ma Duncan Hill pluton and 45.5 Ma Railroad Creek pluton (Fig. 3) were the last of the large intrusions in the North Cascades (Miller et al., 2022), and on the basis of their age and location, they appear to be the youngest sizable elements related to slab rollback (Fig. 6C). The youngest magmatic rocks are ca. 44.9 Ma lineated granite sheets (Misch, 1968; Haugerud et al., 1991; Wintzer, 2012; Miller et al., 2016). Initial ε_{Nd} values for some of the 49.3–45 Ma intrusive rocks are the least radiogenic values for North Cascades intrusions and imply a greater crustal component than in earlier flare-ups (Matzel et al., 2008).

Extensive dike intrusion into a ≥600 km^2 region of the Cascades core and adjacent rocks to the east and south began at ca. 49.3 with the Teanaway dikes and at least one other dike swarm and continued until ca. 45 Ma (Miller et al., 2022). The largest number of dikes intruded between ca. 49.3 and 47 Ma. Many of these rhyolitic to basaltic dikes overlap spatially with the 49–46.5 Ma granodioritic plutons of the core. Some of the dikes have trace element signatures of arc magmas and some are adakites; they are interpreted to be the product of melting of eclogitic lower crust in response to intrusion of mantle-derived basalts (Davidson et al., 2015).

Metamorphism during this time interval is restricted to domains in the Skagit Gneiss Complex of the Cascades core where metamorphic monazite growth continued at least locally until 46 Ma (Gordon et al., 2010a). NW-striking foliation and subhorizontal lineation formed in the complex from ca. 49.5 to 45 Ma (Haugerud et al., 1991; Wintzer, 2012; Miller et al., 2016), and foliation was deformed into upright gentle to open, generally SE- or NW-plunging folds of foliation between ca. 49 Ma and 47 Ma (Miller et al., 2016). Motion of the Ross Lake fault zone ended at ca. 49 Ma, but the Entiat fault was active until at least 46.9 Ma and ended by 44.4 Ma, and the N-S–trending Straight Creek fault experienced dextral slip from ca. 49 Ma and was sealed by 35 Ma (Misch, 1966; Tabor et al., 1984; Miller and Bowring, 1990). Excision and top-to-the-N motion continued on the Dinkelman décollement at least until ca. 49–47 Ma (Matzel, 2004; Paterson et al., 2004). The Eocene dikes also provide information on the strain field. Their average orientation is ~035°, and the resultant extension direction (305°–125°) is oblique to the strike (~320°) of the North Cascades orogen and to the stretching lineation (average trend of 330°–150°) in the Skagit Gneiss Complex (Miller et al., 2022). Overall, these structures are compatible with the regional dextral transtensional tectonic regime.

The 49.5–45 Ma interval was marked by rapid cooling and exhumation of parts of the Cascades core. The 8–12 kbar Swakane Biotite Gneiss was in part exhumed by the Dinkelman décollement and was at the surface in the Chumstick Basin by 48.5 Ma (Tabor et al., 1987; Eddy et al., 2016a). Most of the $^{40}Ar/^{39}Ar$ and K-Ar hornblende, biotite, and muscovite cooling ages in the 7–10 kbar Skagit Gneiss Complex are ca. 50–44 Ma (Engels et al., 1976; Wernicke and Getty, 1997; Tabor et al., 2003; Gordon et al., 2010b), and thermochronology indicates very rapid cooling in some areas, with rates of perhaps 100 °C/m.y. at ca. 47–45 Ma (Wernicke and Getty, 1997).

In the eastern belt, magmatism, sedimentation, and extension all continued during the early part of this interval, and magmatism and extension were largely waning by the end. Igneous activity was still migrating southwestward across northeastern Washington (Fig. 6C). In British Columbia, the >200 km^2 granodioritic Needle Peak pluton intruded the Methow Basin at ca. 48 Ma (Monger, 1989a, 1989b), and Challis-Kamloops magmatism to the east had largely ended by ca. 47 Ma (Ickert et al., 2009; Dostal and Jutras, 2022).

Extension and sedimentation related to the metamorphic core complexes in northeastern Washington and British Columbia were on the wane during this interval. Termination of sedimentation at ca. 48 Ma in northeastern Washington was roughly coeval with the end of volcanism (Suydam and Gaylord, 1997). Mylonitization in the Okanogan Complex ended at ca. 49 Ma with cooling through 47 Ma (Kruckenberg et al., 2008). The Priest River Complex was rapidly exhumed from ca. 50 to 48 Ma (Doughty and Price, 2000; Stevens et al., 2016), but extension and exhumation continued through this interval in Idaho and Montana in the Bitterroot and Anaconda core complexes (Foster et al., 2007, 2010; Howlett et al., 2021). The Lewis and Clark fault zone continued to act as a boundary between the older core complexes to the north and the younger complexes to the south.

45–40 Ma

The 45–40 Ma interval marks the end of slab foundering and the establishment of a new N-S subduction zone and arc that became Cascadia. Subduction was occurring beneath much, if not all, of Oregon and Washington by the end of this period (Fig. 7). Sedimentation occurred in the western belt but ended in the Chumstick Basin, and Challis-Kamloops magmatism in the eastern belt also ended.

Arc magmatism began at ca. 45 Ma in southwestern Washington where local basaltic andesites and andesites erupted (du Bray and John, 2011) and by 40 Ma in southwest Oregon (e.g., Darin et al., 2022). In northwestern Washington, similar volcanic rocks occur in a belt that lies west of the younger part of the Cascades arc and also includes 45–35 Ma granodioritic intrusions, and abundant 45–40 Ma tuffs occur in the Puget Group (Fig. 3) (Vine, 1969; Tabor et al., 1993, 2000; Dragovich et al., 2009, 2011, 2013, 2016; MacDonald et al., 2013). Within this belt, the oldest rocks appear to be at the northern end, but there is a lack of precise dates for units in the south. Local dacite and rhyolite domes (Wenatchee domes) intruded the Chumstick Basin to the east at ca. 44.5 Ma (Gilmour, 2012; Eddy et al., 2017b) and may be the youngest intrusive rocks related to slab rollback and/or breakoff (White et al., 2021). In southwestern Washington and Oregon, the Tillamook magmatic episode occurred from 42 to 34 Ma (Parker et al., 2010; Chan et al., 2012; Wells et al., 2014). This episode included volcanic rocks (Tillamook Volcanics, Yachats Basalt, and Grays River Volcanics) in northwestern Oregon and southwestern Washington, which are interpreted by some workers to be related to the Yellowstone hotspot and were synchronous with margin-parallel extension (e.g., Wells et al., 2014; Camp and Wells, 2021).

Sedimentation in the western belt includes both deep- and shallow-marine deposits on the Olympic Peninsula (Einarsen, 1987; Babcock et al., 1994). Inboard, in the Puget Sound region, the deltaic to shallow-marine middle(?) to late Eocene Puget Group (Fig. 3; Vine, 1969; Buckovic, 1979; Johnson and O'Connor, 1994) was deposited on Siletzia on the west and the older rocks of the western North Cascades on the east. The Puget Sound basin likely formed in the forearc to the early Cascadia arc.

Sedimentation ended in the Chumstick Basin but continued in the overlying, ca. 44–42 Ma arkosic Deadhorse Canyon unit and the Roslyn Formation (Evans, 1994; Eddy et al., 2016a). (Fig. 3). The latter, which rests on the Teanaway Formation south of the Cascades core, may be the easternmost part of the regional depositional system that included the Puget Group.

Magmatism ceased in the Cascades core at ca. 44.9 Ma, and ductile deformation in the Skagit Gneiss Complex had also ended at ca. 45 Ma (Miller et al., 2016). Dextral strike slip ended between 46.9 and 44.5 Ma on the Entiat fault (Evans, 1994; Eddy et al., 2016a) and continued to a later time on the Straight Creek fault, which is intruded by a 34 Ma pluton (e.g., Tabor et al., 2003).

East of the Cascades core, Challis magmatism terminated at ca. 43 Ma (Gaschnig et al., 2010). Extension and cooling of the Bitterroot and Anaconda core complexes continued until ca. 39 Ma, as did sedimentation (Foster et al., 2010; Howlett et al., 2021). Motion on the Lewis and Clark fault zone presumably ended as well.

DISCUSSION

We emphasized in the introduction that the Pacific Northwest in the Paleogene is an excellent place to examine a variety of processes resulting from ridge-trench interaction and oceanic plateau collision. In the following, we explore the upper-plate response shortly before, during, and after the Farallon-Kula or Farallon-Resurrection ridge encountered the trench bordering North America near Vancouver Island and the consequences of the collision of Siletzia.

Relation of the 60–50 Ma Magmatic Lull to Slab Dynamics

It is likely that the end of long-lived arc magmatism in the Cascades core at ca. 60 Ma and the overall low volume of magmatism from ca. 60 to 50 Ma eastward to the Idaho batholith resulted from flat-slab subduction. Moreover, magmatism in the Idaho batholith during this interval probably resulted from crustal thickening and not subduction-related processes (Gaschnig et al., 2010). The shallowing of the slab may be attributable to the rapid subduction of young buoyant lithosphere, as also proposed by others for the greater region (e.g., Thorkelson and Taylor, 1989; Haeussler et al., 2003). Strong suction in the mantle wedge may have played a role, as proposed for the Laramide belt (Humphreys, 2009; O'Driscoll et al., 2009) which at 55 Ma was directly east of Siletzia.

The northern boundary of the flat slab is inferred to be northeast of central Vancouver Island (Fig. 4) where there is a transition in pluton ages within the Coast Mountains batholith. The southern Coast Mountains have a 60–50 Ma magmatic lull much like the Cascades core of this study, whereas to the north, a high-magma-addition event attributed to arc magmatism occurred from 61 to 48 Ma (Cecil et al., 2018). A projection of the triple junction off central Vancouver Island through the boundary in the Coast Mountains to the northeast may run to the northern edge of the Shuswap Complex at this time, which potentially explains the location of the belt of major extension along the eastern edge of the flat slab from British Columbia to southern Idaho and western Montana. Alternatively, the flat slab may have underlain the region of the magmatic lull but just south of most of the Shuswap to Okanogan extensional belt (Fig. 4), in which case the latter would have been kinematically tied to the Tintina fault–Rocky Mountain trench (Price and Carmichael, 1986) and magmatism would have occurred in a slab window (e.g., Breitsprecher et al., 2003). Seismic tomography and reconstructions of plate motions in the northeastern Pacific also suggest a major boundary inboard from Vancouver Island (Fuston and Wu, 2021). Plate motion models indicate rapid northward rates of either the Kula or Resurrection plate from ca. 65 to 50 Ma that were highly oblique to the North American plate boundary (Engebretson et al., 1985; Matthews et al., 2016), and this may have produced a large slab window under western Canada (Fuston and Wu, 2021; cf. Madsen et al., 2006) north of the proposed flat slab.

The magmatic lull and flat slab extended to the south of the crystalline core of the North Cascades, which on the basis of known strike-slip faults (Wyld et al., 2006; this study) was at the latitude of current central Oregon to the Oregon-Washington border at ca. 60–50 Ma. Post–50 Ma volcanic and sedimentary strata obscure relations to the south and east of the Wenatchee block; in our reconstruction at 55 Ma (Fig. 4) and projecting faulting back to 60 Ma, the North Cascades would have lain near the northwestern edge of the Klamath–Blue Mountains terranes and the flat slab beneath the Pacific Northwest would have been continuous with the well-established Laramide flat slab to the south (see Tikoff et al., 2023, for an alternative hypothesis).

Consequences of Collision of Siletzia

The inferred position of the intersection of the Farallon-Resurrection or Farallon-Kula ridge with the trench is complicated by the eruption of Siletzia basalts and the construction of an oceanic plateau above a hotspot mantle plume (e.g., Wells et al., 2014). In the region of the Washington Cascades, major changes occurred in the upper plate of the system due to collision of this oceanic plateau.

Notable aspects of Siletzia collision are the short duration of the associated deformation, its profound inboard influence, and the subsequent change in plate boundary stresses along the newly established North America margin. The most important structural response was the brief shortening that migrated from southwestern Oregon to central Washington and Vancouver Island during the 51–49 Ma interval (Fig. 5) (Wells et al., 2014). In the Swauk Basin, folding and formation of an angular unconformity is tightly bracketed between ca. 50.8 and 49.3 Ma (Eddy et al., 2016a). The reversal of drainage in the Swauk Basin at ca. 51 Ma is probably one of the first signs of Siletzia collision at that latitude (Eddy et al., 2016a). Younger upright folding continued until ca. 48 Ma at deeper crustal levels in the Skagit Gneiss Complex of the Chelan block of the Cascades core ~175 km inboard of Siletzia (Miller et al., 2016). Folding only bracketed between ca. 65 and 48 Ma (Kriens et al., 1995) in the Methow Basin farther to the northeast may have been induced by collision. In contrast, in the eastern belt, ≥235 km inboard of Siletzia, extension in most of the core complexes continued unabated. Peak metamorphism of the voluminous Shuswap Complex and several other core complexes at ca. 53–49 Ma was roughly coincident with the proposed flat slab and Siletzia collision. One explanation for the widespread eastern extension and timing of magmatism and metamorphism may be the rollback of the flat slab, which we propose was under way in Washington by ca. 52 Ma (Figs. 5 and 6C).

In the western belt, sedimentation continued in the early stages of collision after the drainage reversal in the Swauk Basin at 51 Ma but presumably ended during folding and certainly before the Swauk-Teanaway unconformity and eruption of Teanaway volcanic rocks at 49.3 Ma. Note that the youngest Swauk Formation strata are in lake and fluvial facies in the far eastern end of the Swauk Basin near the Leavenworth fault (Tabor et al., 1982; Senes, 2019), and their position may be related to an eastward migration of late basin subsidence related to the collision. In the eastern belt, sedimentation continued in the supra-detachment extensional basins and grabens until ca. 48 Ma, just after this slab is inferred to have rolled back to the southwest.

The collision of Siletzia with the continental margin influenced magmatism much farther eastward than it influenced deformation and sedimentation. We attribute this to the shutoff of northeastward flat subduction caused by the collision-related plate reorganization (e.g., Schmandt and Humphreys, 2011). Magmatism migrated to the southwest across northeastern Washington and reached the Golden Horn batholith at the northeastern margin of the Cascades core at ca. 48.3 Ma (Figs. 3 and 6C). This migration has been interpreted to result from slab rollback (Tepper et al., 2023) and breakoff as the Farallon plate detached and formed the subvertical "slab curtain" currently imaged seismically beneath Idaho and eastern Washington (Schmandt and Humphreys, 2011).

What Drove the 49.3 Ma to 45.5 Ma Magmatic Flare-Up?

Plutons in the North Cascades crystalline core and dike swarms across the study area record a major magmatic flare-up at 49.3–45.5 Ma (Miller et al., 2009), shortly after Siletzia collision. This flare-up is concentrated in the Chelan block of the core but also includes plutons that intruded the Methow Basin directly east and northward of the core for ~70 km into Canada (e.g., Needle Peak pluton), volcanic rocks on the western and southern sides of the core, and voluminous dike swarms (Figs. 3 and 6) (e.g., Tabor et al., 1984; Eddy et al., 2016b; Miller et al., 2016, 2022). The Eocene flare-up is marked by the highest magmatic addition rate and shortest duration of any of the magmatic events in the North Cascades.

The factors that control initiation and termination of magmatic flare-ups, such as the Eocene event, are controversial (e.g., Chapman et al., 2021b). Isotopic data from intrusions emplaced during flare-ups in some arcs imply increased crustal melting and have led to the orogenic cycle hypothesis in which flare-ups are driven by melting of fertile back-arc crustal material thrust into the deep levels of an arc or underlying mantle (e.g., Ducea and Barton, 2007; DeCelles et al., 2009). Others have argued that voluminous melting results dominantly from processes external to the arc, including slab breakoff and ridge subduction, and largely involves mantle-derived melts (e.g., Decker et al., 2017; Schwartz et al., 2017; Martínez Ardila et al., 2019), which in turn can drive an increase of partial melting of the crust.

The Eocene Cascades core plutons have been considered to represent the latest pulse of arc magmatism in the North Cascades by earlier workers (e.g., Matzel et al., 2008; Miller et al., 2009), and magmatism to the east in the Challis-Kamloops belt has been interpreted to have occurred within a slab window (e.g., Thorkelson and Taylor, 1989; Breitsprecher et al., 2003). In our view, the flare-up is related to the Farallon slab rollback and breakoff. At ca. 49.5 Ma, the southwest-migrating rollback magmatism had reached the northeastern margin of the Cascades core (Tepper et al., 2023), and the edge of a large slab window may have lain nearby to the north (Fig. 6). The accretion of Siletzia and termination of subduction led the slab to break off, as shown in part by the belt of bimodal volcanic rocks lacking an arc signature near the Straight Creek fault (Figs. 3, 6, and 9) (Kant et al., 2018). The Eocene-age Cascades core plutons have a wider isotopic range than earlier plutons (Matzel et al., 2008), but their geochemistry does not permit distinguishing between an arc or slab breakoff origin because the crustal component of melt during breakoff would have been mafic lower crust of the Late Cretaceous arc. Dextral strike slip, slab rollback, and breakoff were concentrated in and near the Cascades core, and we infer that the slab was ripped apart, leading to upwelling of asthenospheric mantle and decompression melting (Fig. 9).

A speculative additional interpretation is that the breakoff-related magmatism continued to the southeast beneath the Columbia River Basalt Group in the Pasco Basin to the Clarno Formation of northeastern Oregon (Figs. 2 and 6). The Pasco Basin is on strike with the Eocene Chumstick Basin, and seismic velocities suggest that beneath the Miocene basalt is a thick, asymmetric sedimentary basin of probable Eocene age and an associated mafic underplate (Catchings and Mooney, 1988; Gao et al., 2011). These mafic rocks may be similar to the Teanaway Basalt of the flare-up. The Clarno Formation is not well dated, but available ages suggest that the volcanic rocks erupted starting at ca. 53–50 Ma (Bestland et al., 2002). Note that in our reconstruction for 48 Ma, the Clarno area is ~100 km southeast of the North Cascades flare-up, and the western breakoff belt west of the Straight Creek fault would have been ~40–50 km closer to the Clarno area at 50 Ma. If the Siletzia terrane lay on a small microplate within the shrinking northern Farallon plate as we show (Fig. 6), then the southeastern edge of the slab that rolled back and broke off may have been near the Clarno volcanics (cf. Humphreys, 2009).

Upper-Plate Deformation after Siletzia Collision

The ca. 49–45 Ma structural record west of the Straight Creek–Fraser River fault is largely restricted to high-angle NW-striking faults and associated local folds, whereas in the central and eastern belts, a wide array of structures can be used to evaluate deformation. Eocene dikes, dextral strike-slip faults, basins, and ductile structures in the Cascades are broadly coeval with dikes, faults bounding non-marine basins, and ductile fabrics in metamorphic core complexes in northeastern Washington and southern British Columbia (Fig. 6) (e.g., Ewing, 1980; Parrish et al., 1988; Eddy et al., 2016a; Miller et al., 2016). Dikes in the eastern belt are not well dated, but most K-Ar dates from volcanic rocks in northeastern Washington range between 51 and 48 Ma (Pearson and Obradovich, 1977) and thus overlap temporally with the older (49.3–47.5 Ma) dikes in the Cascades and the magmatic flare-up. Dikes intruding the Kettle metamorphic core complex, ~140 km east of the North Cascades, strike ~012°–022° (McCarley Holder et al., 1990, their figure 1). These dikes are subparallel

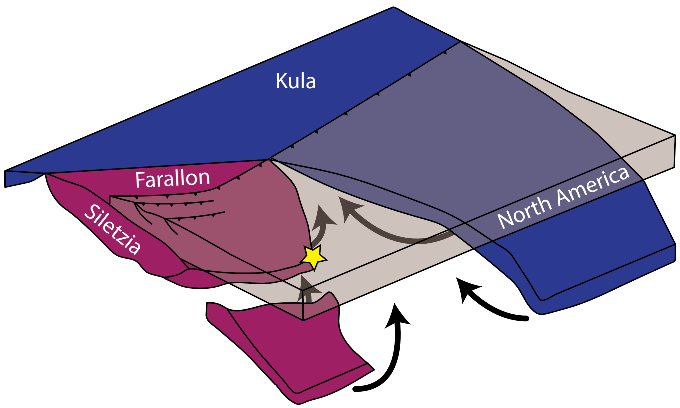

Figure 9. Tectonic model for 49.5–45 Ma magmatic flare-up in the Washington Cascades. The star schematically shows the location of the northern Washington Cascades where the slab has broken, south of the postulated slab window to the north. See text for explanation.

to the normal faults that separate the Kettle and Okanogan core complexes from Eocene grabens (Keller, Republic, and Toroda grabens), which strike 008°–020°. Farther east, ENE-WSW (~075°–255°) brittle slip occurred on the Newport fault, which is the upper boundary of the Priest River Complex (Harms and Price, 1992), and east and south of the Lewis and Clark fault zone, slip on the Bitterroot and Anaconda detachments is top-to-the-ESE (~100°–110°) (Kalakay et al., 2003; Foster et al., 2007). Brittle extension directions from the dikes and faults bounding the grabens suggest that they are oblique (~15°–50° counterclockwise) to those of the voluminous NNE-striking (average of 035°), ca. 49.3–47.5 Ma dikes in the Cascades.

A major difference between faults in the eastern belt and those in the western and central belts is that the eastern faults are apparently purely dip slip, whereas faults (Ross Lake, Entiat, Leavenworth, Straight Creek) in the central and western belts are dextral strike slip, and most have a subordinate component of normal slip. Dextral slip does occur to the east on the Lewis and Clark fault zone (Figs. 2 and 6), but this structure strikes ~E-W and transfers slip between the Anaconda, Bitterroot, and Priest River core complexes (e.g., Foster et al., 2007). The combination of dextral strike-slip faults and dike swarms of the Cascades core region is most compatible with a N-S dextral shear and related WNW-ESE extension.

Eocene ductile stretching in mylonites in core complexes ranges from ~105°–285° in the Bitterroot and Anaconda complexes in Montana (Foster et al., 2007), to 074°–254° in the Priest River Complex (Harms and Price, 1992; Doughty and Price, 1999) near the Washington-Idaho border, to E-W in the Kettle complex (Rhodes and Cheney, 1981), to WNW to ESE (~295°–115°) in the Okanogan Complex (Kruckenberg et al., 2008; Brown et al., 2012) ~40 km east of the Cascades core. Broadly coeval, subhorizontal Eocene ductile stretching in the North Cascades is ~330°-150° in the Skagit Gneiss Complex to close to N-S in the Swakane Biotite Gneiss. Thus, ductile extension directions rotate progressively clockwise by ~75° from east to west. The sense of rotation is the same as, but the magnitude of rotation is greater than that of the upper-crustal structures.

Rotation of extension directions fits with the progressively greater influence of dextral shear closer to the plate margin in response to the plate reorganization at ca. 49.5 Ma after Siletzia collision. Extension and transtension led to orogenic collapse in the core complexes (e.g., Price and Carmichael, 1986; Parrish et al., 1988; Vanderhaeghe and Teyssier, 2001), whereas strike slip occurred to the west on the faults bounding and cutting the North Cascades core.

Eocene Global Plate Reorganization

The dramatic tectonic transitions in the Pacific Northwest region at ca. 52–49 Ma coincide with a fundamental plate reorganization in the Pacific basin and a global change in plate vectors at ca. 53–47 Ma (e.g., Whittaker et al., 2007; O'Connor et al., 2013; Seton et al., 2015). This plate reorganization in the Pacific may have been driven by subduction of the Izanagi-Pacific ridge at ca. 60–46 Ma (Wu and Wu, 2019), with the ensuing initiation of subduction in the Tonga-Kermadec and Izu-Bonin-Mariana system occurring at ca. 53–50 Ma (Sharp and Clague, 2006; Whittaker et al., 2007; Tarduno et al., 2009). The ca. 50 Ma bend in the Hawaiian-Emperor seamount chain also coincides with a change in Pacific plate motion and Australian-Antarctic plate reorganization at that time (Sharp and Clague, 2006; Whittaker et al., 2007). It has been suggested that Pacific-Kula plate spreading also changed at ca. 53.3 to 43.8 Ma (Lonsdale, 1988) and that Kula-North America relative motion became more northerly and faster at 57 Ma (Doubrovine and Tarduno, 2008). Other major global events roughly coeval with the fundamental changes in the Pacific Northwest region include initiation of the Aleutian arc and the dramatic slowing of Greater India at ca. 50 Ma resulting from collision with Asia (e.g., Copley et al., 2010; van Hinsbergen et al., 2011).

It appears that the significant changes in the tectonics of the Pacific Northwest at 52–49 Ma are the consequence of both a global plate reorganization and the regional collision of the ridge-centered Siletzia oceanic plateau. The global plate changes resulted in faster and perhaps more northerly relative plate motion in the Pacific Northwest, which in turn resulted in the formation of the new N-S–striking strike-slip Straight Creek–Fraser River fault. However, most of the complex changes summarized here are the result of the profound changes due to the Siletzia collision, and westward stepping of the subduction zone, and triple-junction migration during the 60–40 Ma interval.

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