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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 (U.S.A. AND SW CANADA) IN THE PALEOGENE

Tracking no: GS2629R

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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 and southern British Columbia in the early-middle Eocene and we summarize a variety of new datasets 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. 60-50 Ma interpreted to reflect low-angle subduction. to 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-Kula or Farallon-Resurrection spreading center. Farther east, peraluminous magmatism occurred in the Omineca belt and Idaho batholith. Accretion of Siletzia and ridge-trench interaction occurred between ca. 53-49 Ma, as indicated by: (i) near-trench magmatism from central Vancouver Island to NW Washington; (ii) disruption and inversion of the Swauk basin during a short-lived contractional event; (iii) 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 (iv) southwestward migration of magmatism across NE Washington. These events suggest that flat slab subduction from ~60-52 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 Ma 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 8-10 kb metamorphic rocks in the North Cascades crystalline core. By ca. 45 Ma, the Farallon - Kula (or Resurrection) - North American triple-junction was likely located in Oregon, 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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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 and southern British Columbia in the early-middle Eocene and we summarize a variety of new datasets 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. 60-50 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-Kula or Farallon-Resurrection spreading center. Farther east, peraluminous magmatism occurred in the Omineca belt and Idaho batholith. Accretion of Siletzia and ridge-trench interaction occurred between ca. 53–49 Ma, as indicated by: (i) near-trench magmatism from central Vancouver Island to NW Washington; (ii) disruption and inversion of the Swauk basin during a short-lived contractional event; (iii) 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 (iv) southwestward migration of magmatism across NE Washington. These events suggest that flat slab subduction from ~60–52 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 Ma 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 8-10 kb metamorphic rocks in the North Cascades crystalline core. By ca. 45 Ma, the Farallon – Kula (or Resurrection) – North American triple-junction was likely located in Oregon, 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.

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 (U.S.A) 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 Cascades 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 about 96–60 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 – 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 – Kula, Farallon – Resurrection, or Farallon – Orcas 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 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 east-west 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, 1984; 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, 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 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, 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; b; Miller et al., 2016, 2022), which enables the construction of a detailed time line 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 ~83 Ma and that the northern boundary of the Farallon plate was a ridge, which intersected the continental margin at a poorly constrained location (e.g., Atwater, 1970; Wood 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 Ma 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 BC 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/Myr) 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 56-49 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, 4). From 52–49 Ma, a triple junction is interpreted to have interacted with the continental margin along central to southern Vancouver Island (Fig. 1), as 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, 2021), and disruption of non-marine basins (Eddy et al., 2016a). The collision of Siletzia, which started by 53 Ma in SW 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 52–48 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).

PRE-PALEOGENE GEOLOGIC SETTING

Prior to 65 Ma, the Pacific Northwest was characterized by a typical convergent margin with a forearc, continental magmatic arc, back-arc 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 oceanic and island arc 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 Ma 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 96–87 Ma, and most biotite Ar/Ar and K/Ar cooling ages are >60 Ma, whereas magmatism in the Chelan block ranges from 92–45 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 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 plutonic and metamorphic rocks of the Omineca belt, including multiple metamorphic core complexes, the Idaho batholith, and Cordilleran passive margin sediments involved in the Rocky Mountain-Sevier fold-and-thrust belt (Fig. 2).

RESTORATION OF STRIKE-SLIP FAULTS

Dextral strike-slip faulting occurred in the northern 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 – 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 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 NE 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 ~115 km of dextral offset (Umhoefer and Miller, 1996).

These known displacements of ~325 km must be considered in tectonic restorations, particularly before 50 Ma. To summarize, after 50 Ma there is approximately 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) 50 km (of total 115 km) of offset between the core and the eastern belt; and 4) a cumulative offset of ~265 km between the western and eastern belts after 50 Ma (Table 1). If we assume that the strike-slip offset from 60 to 50 Ma occurred at rates comparable to those of the ~50–40 Ma interval, the implication is that another approximately 250–300 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 about 450 km south of 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 ~85–55 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 Klamath Mountains northward to the Olympic Peninsula. 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 northeastern-most 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 km or less.

PALEOGENE TECTONIC HISTORY

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

65 – 60 Ma

During this 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 Mountain 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 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, as the upper limit for the mélange is only indicated 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 (up to 12 kbar) Swakane Biotite Gneiss (Valley et al., 2003) and Skagit Gneiss Complex (7 – 10 kbar; Whitney, 1992; Hanson, 2022) (Fig. 3), between ca. 79

– 66 Ma and 74 – 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 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 NE 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 Complex (Fig. 4) initiated at 62 Ma (Carr, 1992; Hinchey and Carr, 2006). In Idaho, 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 NE 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 Complex and in the Rocky Mountain - Sevier fold and thrust belt (e.g., Price, 1981).

60 – 52 Ma

This 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 (southwest Oregon) and by 53.2 Ma outboard of the Northwest Cascades system and mélange belts in western Washington and Vancouver Island in the north, where it continued until at least 48 Ma (Crescent and Metchosin basalts) (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-oceanic-ridge basalt (N-MORB) to shallow water and subaerial flows of enriched mid-oceanic-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 hot spot (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 maximum depositional age (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 Ma 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 – 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 ~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 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, 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, 2021). The alkalinity of magmas increases markedly south of ca. 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., Parrish et al., 1988; Harlan et al., 1988; Brown et al., 2012), led to the formation of metamorphic core complexes and associated extensional basins in NE 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 NE 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 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 Ma 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 northwest Washington by near-trench magmatism from ca. 51 – 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 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 (Clayquot 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 T/low P metamorphism at ~51 Ma (Fairchild and Cowan, 1982; Groome et al., 2003). In NW Washington, local peraluminous magmatism occurred as the ca. 49 Ma Mt. 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 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 SW 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 Ma 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 Myr) 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 SW 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 Gneiss in the crystalline core was probably rapidly exhumed during this interval, in part during distributed ductile shear and top-to-N to –NNE motion on the Dinkelman decollement (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 Schriazza, 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/YbN>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 SW across NE Washington (Fig. 6C) (Tepper, 2016). Metamorphism and deformation continued in the metamorphic core complexes in southern British Columbia, NE 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; Suydan and Gaylord, 1997; Foster et al., 2007; Brown et al., 2012; Rubino et al., 2021). In contrast to NE 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 east-west 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 NW, and the southwestward migration of magmatism in NE Washington is consistent with rollback of the northern Farallon slab (Figs. 5, 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, 2016). 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 this 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 aerially 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 nonmarine sedimentary rocks in the Naches and Barlow Pass units (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 ~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 ~46.5 Ma, and fault reorganization at ~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 ~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 (ca. 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 ~47.9–46.5 Ma, magmatism in the core migrated westward from the Ross Lake fault zone. The ~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., 2021), 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). εNdi 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. 49.3–47 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. 49.5–45 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 to 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 north motion continued on the Dinkelman decollement 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 Gneiss was in part exhumed by the Dinkelman decollement 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 NE Washington (Fig. 6C). In British Columbia, the >200 km^2, granodioritic Needle Peak pluton intruded the Methow basin at ca. 48 Ma (Monger, 1989), and Challis-Kamloops magmatism to the east had largely ended by ca. 47 Ma (Ickert et al., 2009; Dostal and Jutras, 2021).

Extension and sedimentation related to the metamorphic core complexes in NE Washington and British Columbia were on the wane during this interval. Termination of sedimentation at ~48 in NE 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. 50–48 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; Howlet 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

This interval marks the end of slab foundering and the establishment of a new north-south 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, as did Challis-Kamloops magmatism in the eastern belt.

Arc magmatism began at ca. 45 Ma in southwest 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 SW 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 NW Oregon and SW 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 Ma 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. 60–50 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). Note that the Laramide belt in northern Wyoming was directly east of Siletzia at 55 Ma in our reconstruction (Fig. 4).

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 61–48 Ma (Cecil et al., 2018). A projection of the triple junction off central Vancouver Island through the boundary in the Coast Mountains to the NE 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 be kinematically tied to the Tintina fault – Rocky Mountain trench (Price and Carmichael, 1986) and magmatism would occur in a slab window (e.g., Breitsprecher et al., 2003). Seismic tomography and reconstructions of plate motions in the NE 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 plates from ca. 65–50 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 NW edge of the Klamath – Blue Mountains terranes and the flat slab beneath the Pacific Northwest would be 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/Kula ridge with the trench is complicated by the eruption of Siletzia basalts and the construction of an oceanic plateau above a hot spot 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 southwest 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 ~50.8 Ma and 49.3 Ma (Eddy et al., 2016a). The reversal of drainage in the Swauk basin at ~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 Ma 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 ~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 underway in Washington by ca. 52 Ma (Figs. 5, 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 SW.

The collision of Siletzia with the continental margin influenced magmatism much farther eastward than it influenced deformation and sedimentation. We attribute this to the shut off of northeastward flat subduction caused by the collision-related plate reorganization (e.g., Schmandt and Humphreys, 2011). Magmatism migrated to the southwest across NE Washington and reached the Golden Horn batholith at the northeast margin of the Cascades core at ca. 48.3 Ma (Figs. 3, 6C). This migration has been interpreted to result from slab rollback (Tepper, 2016) 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 ca. 70 km into Canada (e.g., Needle Peak pluton), volcanic rocks on the west and south sides of the core, and voluminous dike swarms (Figs. 3, 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 backarc 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 break-off and ridge subduction, and largely involves mantle-derived melts (e.g., Decker et al., 2017; Schwartz et al., 2017; 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 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 occur 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 ~49.5 Ma, the southwest-migrating rollback magmatism had reached the northeast margin of the Cascades core (Tepper, 2016) 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, 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 break-off origin as the crustal component of melt during break-off would be 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 NE Oregon (Figs. 2, 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 about 100 km SE of the North Cascades flare-up and the western breakoff belt west of the Straight Creek fault would have been about 40–50 km closer to the Clarno 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 southeast 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 Fraser River-Straight Creek 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 NE 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 NE Washington range between 51–48 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 Fig 1). These dikes are subparallel to the normal faults that separate the Kettle and Okanogan core complexes from Eocene grabens (Keller, Republic, and Toroda), 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-east-southeast (~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 (ca. 15°–50° counter clockwise) to those of the voluminous N–NE-striking (average of 035°), ~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, 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 W-NW – E-SE (~295-115°) in the Okanogan Complex (Kruckenberg, 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 Gneiss. Thus, ductile extension directions rotate progressively clockwise by ~75° from east to west. The sense of rotation is the same, but the magnitude of rotation is greater, then 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 ~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; Vanderhaege 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 ~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-Kermedec and Izu-Bonin-Mariana system occurring at ca. 53–50 Ma (Sharp and Clague, 2006; Whittaker et al., 2007a; Tarduno et al., 2009). The ~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 Ma 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.

ACKNOWLEDGEMENTS

This research was supported by National Science Foundation grants EAR-1119358 and EAR-1945352 to R.B. Miller, EAR-1945260 to M.P. Eddy, EAR-1119252 to J.H. Tepper, EAR-1119063 to P.J. Umhoefer, and EAR-1118883 to S. Bowring. Our late colleague and friend Paul Umhoefer was a driving force for much of this synthesis and particularly the fault reconstructions. The late Sam Bowring also played an important role in the research presented here. M.S. and B.S thesis students at Northern Arizona University, San Jose State University, and the University of Puget Sound contributed to our research project and include Katie Bryant, Erin Donaghy, Melissa Gunderson, Lisa Kant, Jen Pence, and Francesca Senes. We thank Stacia

Gordon, Margi Rusmore, and Noah McLean for helpful discussions. We thank Gene Humphreys and Derek Thorkelson for their thorough and very helpful reviews of the manuscript.

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TABLE

Table 1. STRIKE-SLIP FAULT OFFSETS ACROSS WASHINGTON CASCADES.

FIGURE CAPTIONS

Figure. 1. Simple plate reconstruction models of the NE 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 (e.g., Hauessler 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 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 ~50 Ma (Clennett et al., 2020). Sanak-Baranof 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).

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 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 Fig. 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 Formation; Ccb=Chilliwack batholith; CHK=Chuckanut Formation; CM=Cooper Mountain pluton; CP=Castle Peak stock; CRb=Columbia River Basalt Group; DD=Dinkelman decollement; DH=Duncan Hill pluton; Ef=Entiat fault; Epc=Eagle Plutonic Complex; FDfz = Foggy Dew fault zone; FRfz=Frazer 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; SW=Swauk Formation; SWG=Swakane Gneiss; T=Teanaway Formation; WEMB=Western and eastern mélange belts; Yi=Yale intrusions.

Figure 4. 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 which Siletzia formed on 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. 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.

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. 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.

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 intruded in the central and northern Washington Cascades coincident with a magmatic flare-up. Challis-Kamloops magmatism continued, but is beginning to wane, as is extension associated with core complexes. B. Proposed plate tectonic setting. The approximate location of the idealized slab window assumes that the Farallon plate was moving NE and the Kula-Farallon ridge intersected the continental margin as shown. C. Eocene magmatism across NE to north-central Washington and the pattern of inferred rollback magmatism to the southwest are shown. Filled circles are localities of U-Pb zircon ages 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; Chb = Chumstick basin; Csz =

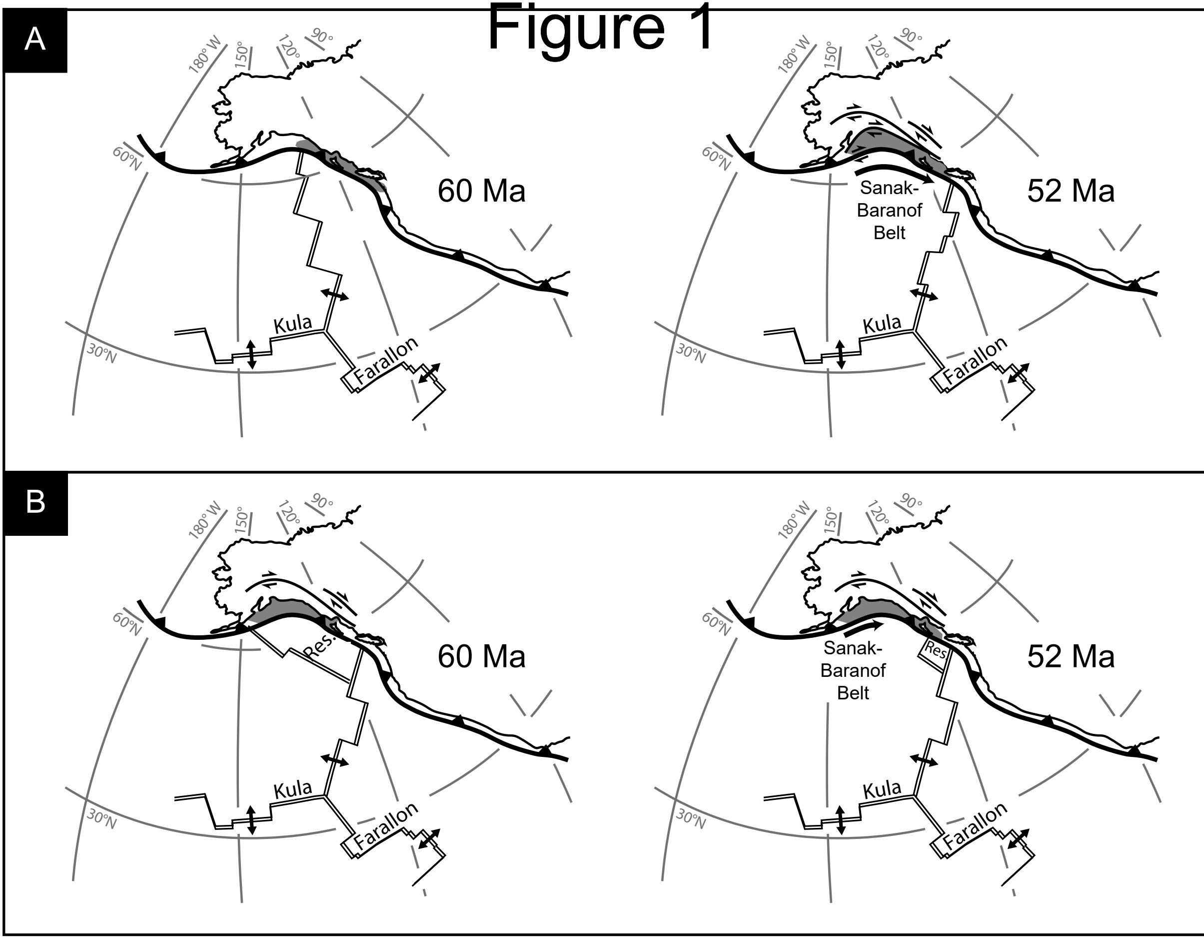
Coast shear zone; Ef = Entiat fault; Fa = Farallon plate; Ff = Fraser fault; K = Kettle core complex; LCfz = Lewis and Clark fault zone; Ok = Okanogan core complex; P = Priest River core complex; Pab = Pasco basin; Pf = Pasayten fault; RLf = Ross Lake fault; SCf = Straight Creek fault; Sh = Shuswap core complex; V = Valhalla complex; VI = Vancouver Island; Wb = Wenatchee block of North Cascades crystalline core. States and Provinces: BC = British Columbia; ID = Idaho; MT = Montana; NV = Nevada; UT = Utah; WA = Washington.

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. VI = Vancouver Island; Yf = Yalakom fault. States and Provinces: BC = British Columbia; ID = Idaho; MT = Montana; OR = Oregon; WA = Washington.

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.

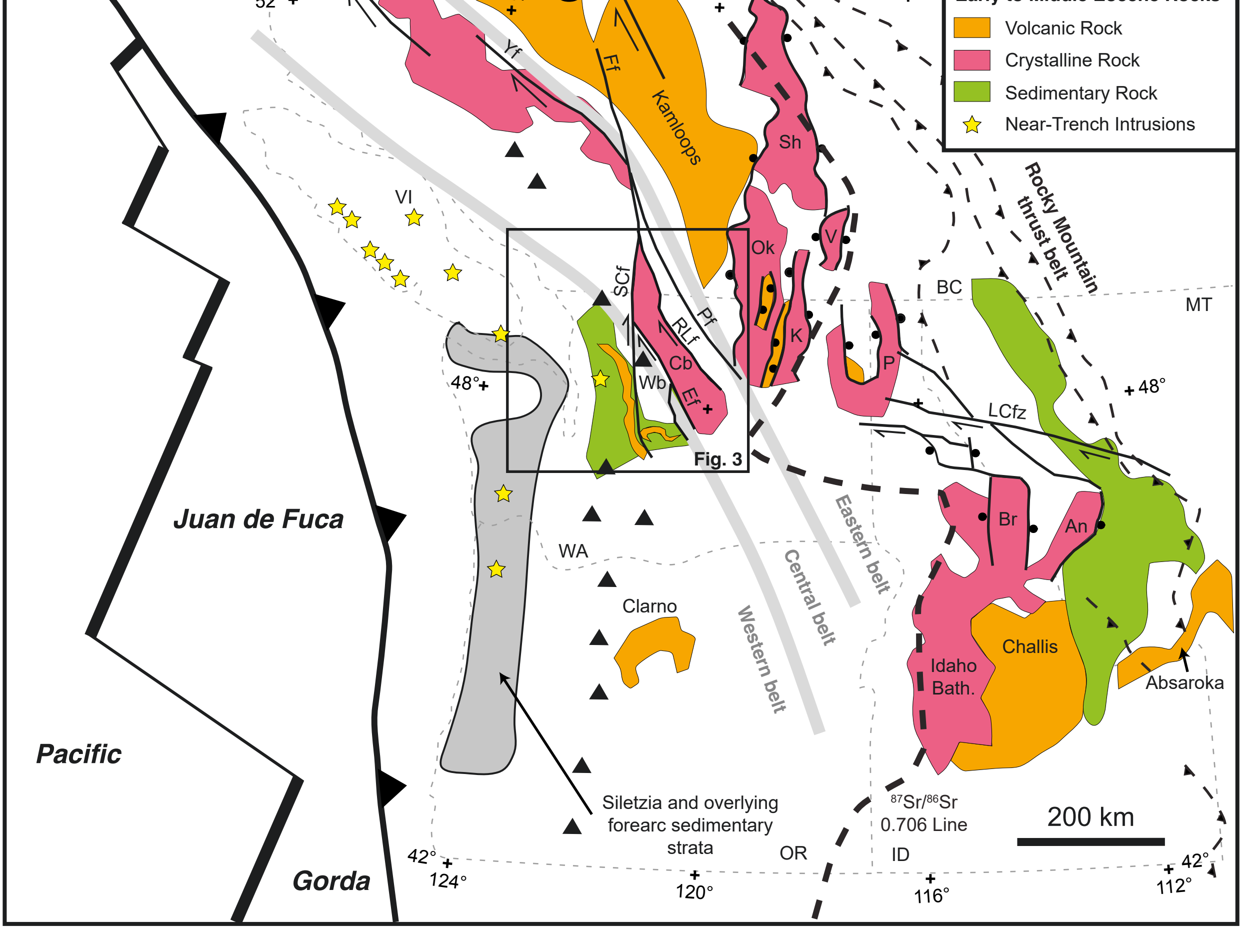
Figure 9. Tectonic model for 49.5–45 Ma magmatic flare-up in the Washington Cascades. The star schematically shows the location of the northern Washington Cascades where the slab has broken, south of the postulated slab window to the north. See text for explanation.

Figure 1



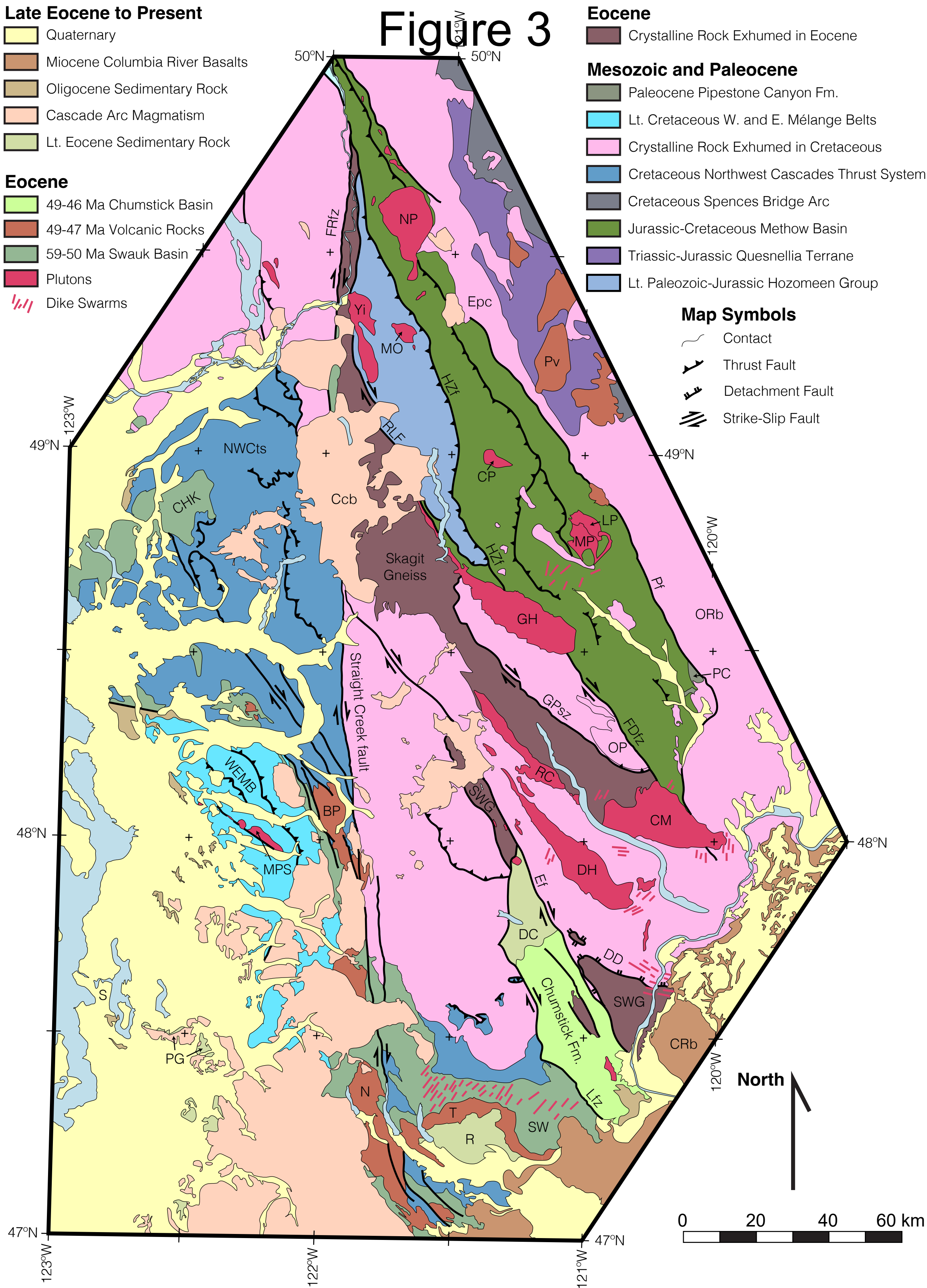
A
180° W
150°
120°
90°
60°N
30°N
60 Ma
52 Ma
Kula
Farallon
Sanak-Baranof Belt
B
Res.
Res

Figure 2



Modern Volcano
Early to Middle Eocene Rocks
Volcanic Rock
Crystalline Rock
Sedimentary Rock
Near-Trench Intrusions
Csz
Yf
Ff
Kamloops
Sh
V
Ok
K
P
SCf
RLf
Pf
Cb
Wb
Ef
Fig. 3
VI
BC
MT
LCfz
Rocky Mountain thrust belt
Juan de Fuca
Pacific
Gorda
WA
Clarno
Western belt
Central belt
Eastern belt
Br
An
Challis
Idaho Bath.
Absaroka
Siletzia and overlying forearc sedimentary strata
$^{87}Sr/^{86}Sr$ 0.706 Line
OR
ID
200 km
128°
52°
124°
120°
116°
48°
42°
112°

Figure 3


Late Eocene to Present
Quaternary
Miocene Columbia River Basalts
Oligocene Sedimentary Rock
Cascade Arc Magmatism
Lt. Eocene Sedimentary Rock
Eocene
49-46 Ma Chumstick Basin
49-47 Ma Volcanic Rocks
59-50 Ma Swauk Basin
Plutons
Dike Swarms
Eocene
Crystalline Rock Exhumed in Eocene
Mesozoic and Paleocene
Paleocene Pipestone Canyon Fm.
Lt. Cretaceous W. and E. Mélange Belts
Crystalline Rock Exhumed in Cretaceous
Cretaceous Northwest Cascades Thrust System
Cretaceous Spences Bridge Arc
Jurassic-Cretaceous Methow Basin
Triassic-Jurassic Quesnellia Terrane
Lt. Paleozoic-Jurassic Hozomeen Group
Map Symbols
Contact
Thrust Fault
Detachment Fault
Strike-Slip Fault
50°N
121°W
49°N
48°N
47°N
123°W
122°W
120°W
FRfz
NP
Yi
MO
Epc
Pv
HZf
RLF
NWCts
CP
CHK
Ccb
LP
MP
Skagit Gneiss
GH
Pf
ORb
PC
GPsz
FDfz
OP
RC
WEMB
BP
Straight Creek fault
SWG
CM
MPS
DH
Ef
DC
DD
Chumstick Fm.
SWG
S
PG
CRb
N
T
R
SW
Lfz
North
0
20
40
60 km

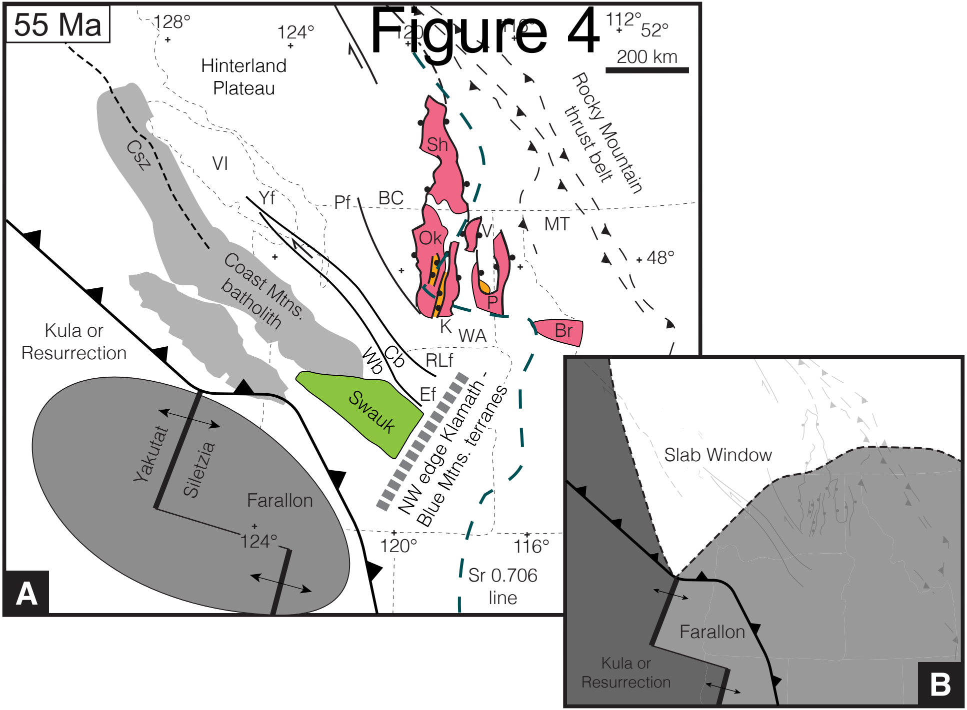

55 Ma
128°
124°
120°
116°
112°
52°
200 km
Hinterland
Plateau
Rocky Mountain
thrust belt
Csz
VI
Yf
Pf
BC
Sh
Ok
V
MT
48°
Coast Mtns.
batholith
P
K
WA
Br
Kula or
Resurrection
Wb
Cb
RLf
Ef
Swauk
NW edge Klamath -
Blue Mtns. terranes
Yakutat
Siletzia
Farallon
124°
120°
116°
Sr 0.706
line
A
Slab Window
Farallon
Kula or
Resurrection
B

Figure 4

Figure 5

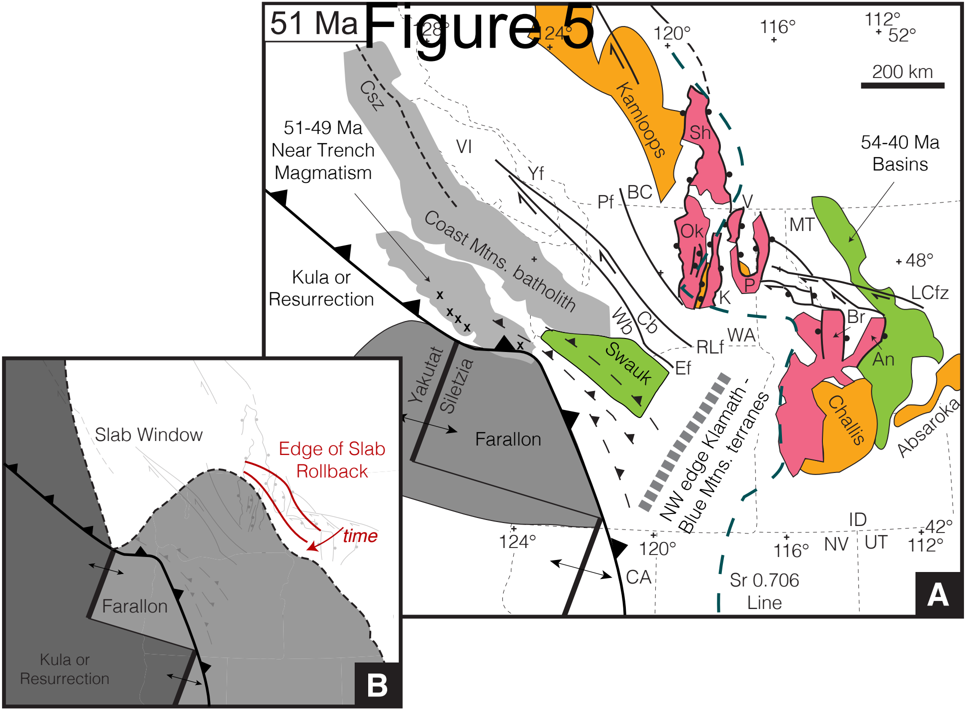

51 Ma
128°
124°
120°
116°
112°
52°
200 km
Csz
51-49 Ma
Near Trench
Magmatism
VI
Yf
Kamloops
Sh
54-40 Ma
Basins
Pf
BC
V
Ok
MT
48°
Coast Mtns. batholith
P
K
LCfz
Kula or
Resurrection
Wb
Cb
Br
WA
RLf
An
Swauk
Ef
Yakutat
Siletzia
Farallon
NW edge Klamath -
Blue Mtns. terranes
Challis
Absaroka
ID
42°
124°
120°
116°
NV
UT
112°
CA
Sr 0.706
Line
A
Slab Window
Edge of Slab
Rollback
time
Farallon
Kula or
Resurrection
B

Figure 6

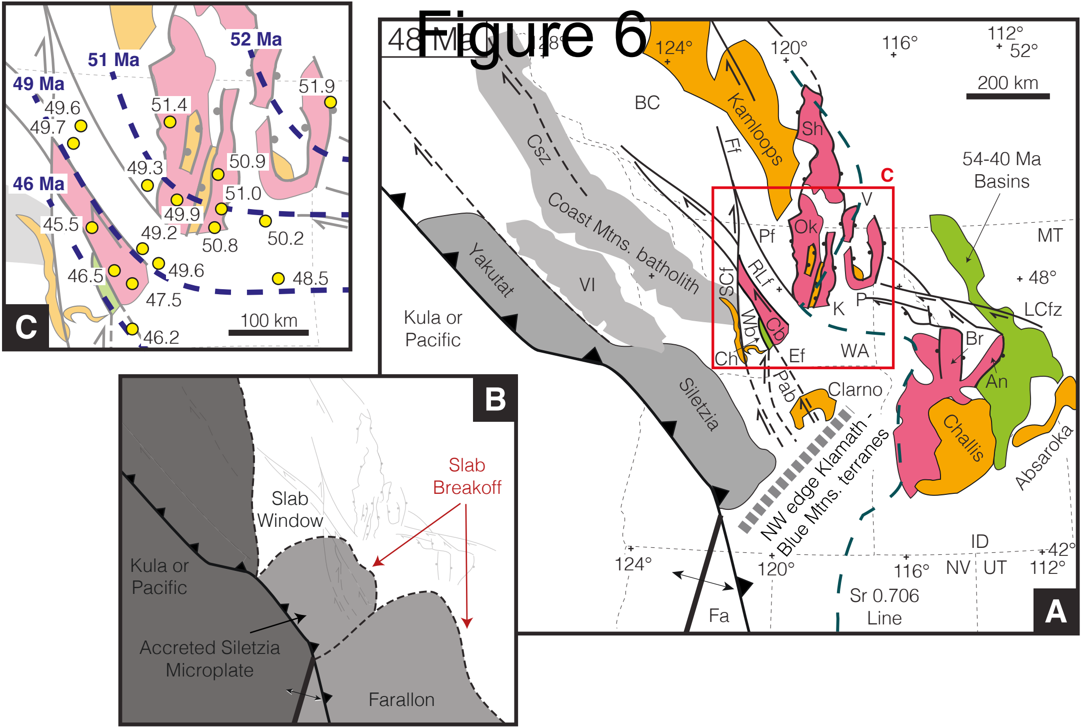
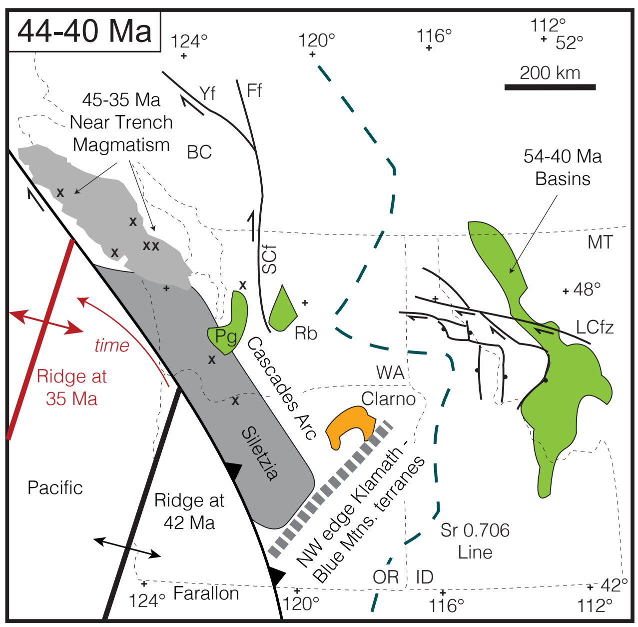
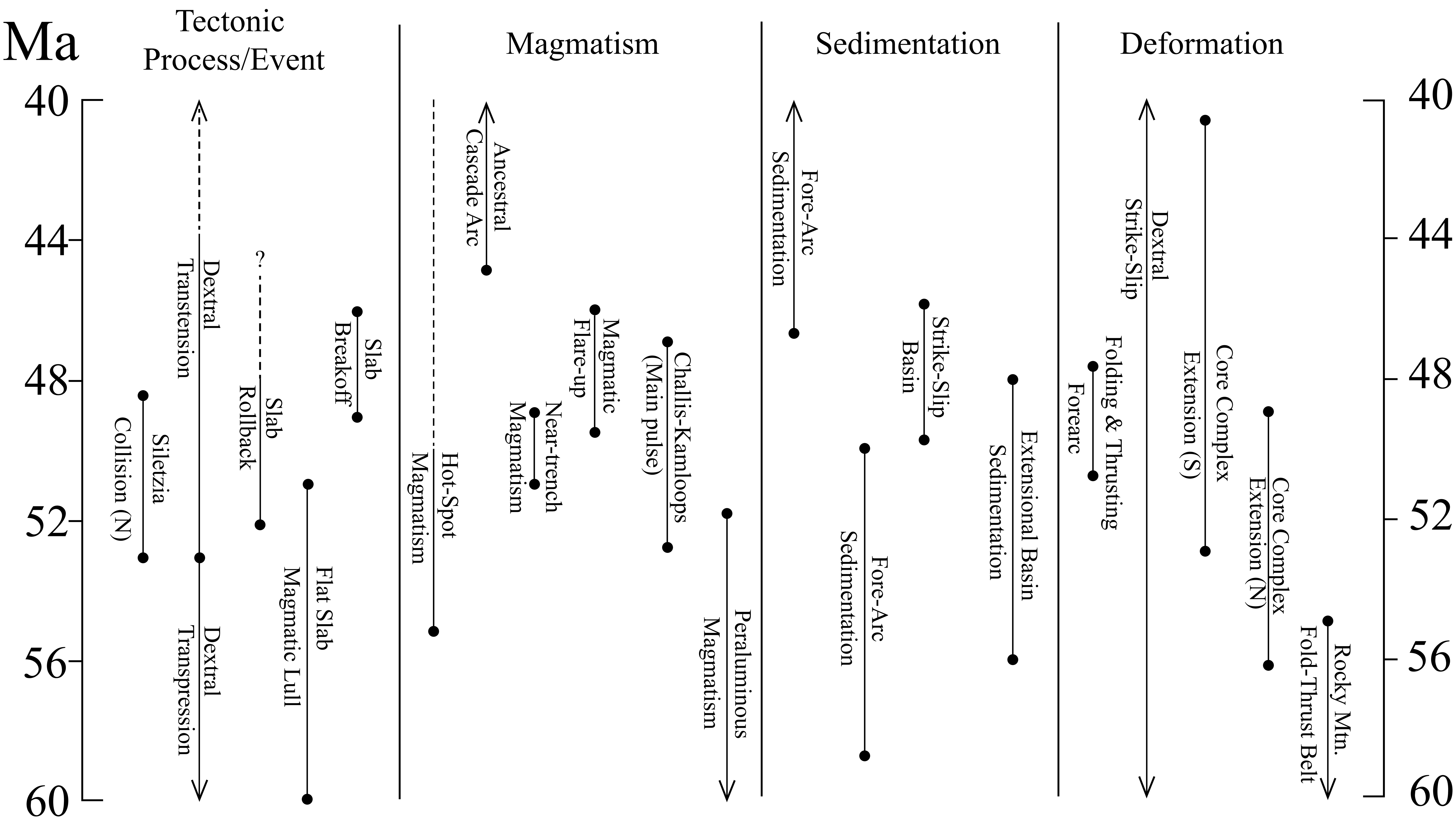

48 Ma
128°
124°
120°
116°
112°
52°
200 km
BC
Kamloops
Sh
Ff
Csz
Coast Mtns. batholith
54-40 Ma
Basins
C
V
Pf
Ok
MT
Yakutat
VI
SCf
RLf
K
P
48°
LCfz
Kula or
Pacific
Wb
Cb
Ch
Ef
WA
Br
An
Siletzia
Pab
Clarno
Challis
Absaroka
NW edge Klamath -
Blue Mtns. terranes
ID
124°
120°
116°
NV
UT
42°
112°
Sr 0.706
Line
Fa
A
52 Ma
51 Ma
49 Ma
46 Ma
49.6
49.7
51.4
51.9
49.3
50.9
51.0
49.9
45.5
49.2
50.8
50.2
49.6
46.5
48.5
47.5
100 km
46.2
C
B
Slab
Breakoff
Slab
Window
Kula or
Pacific
Accreted Siletzia
Microplate
Farallon

Figure 7



44-40 Ma
124°
120°
116°
112°
52°
200 km
Yf
Ff
45-35 Ma
Near Trench
Magmatism
BC
54-40 Ma
Basins
MT
48°
SCf
Pg
Rb
LCfz
time
Ridge at
35 Ma
Cascades Arc
WA
Clarno
Siletzia
Pacific
Ridge at
42 Ma
NW edge Klamath -
Blue Mtns. terranes
Sr 0.706
Line
OR
ID
42°
124°
Farallon
120°
116°
112°

Figure 8



Ma
Tectonic Process/Event
Magmatism
Sedimentation
Deformation
40
44
48
52
56
60
Dextral Transtension
Dextral Transpression
Siletzia Collision (N)
?
Slab Rollback
Flat Slab Magmatic Lull
Slab Breakoff
Ancestral Cascade Arc
Hot-Spot Magmatism
Near-trench Magmatism
Magmatic Flare-up
Challis-Kamloops (Main pulse)
Peraluminous Magmatism
Fore-Arc Sedimentation
Strike-Slip Basin
Fore-Arc Sedimentation
Extensional Basin Sedimentation
Dextral Strike-Slip
Folding & Thrusting Forearc
Core Complex Extension (S)
Core Complex Extension (N)
Rocky Mtn. Fold-Thrust Belt

Figure 9

TABLE 1. STRIKE-SLIP FAULT OFFSETS ACROSS WASHINGTON CASCADES

		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		115
TOTAL OFFSET	**10**	**60**	**135**	**95**	**35**	**325**

We appreciate the thoughtful and constructive comments (in italics below) by the reviewers. Our responses directly follows each of the comments.

Response to Comments by Gene Humphreys (Reviewer 1)

Reviewer #1 (Comments to the Author):

This was a well written and organized paper that was easy to read. More important, it is a thoughtful and comprehensive synthesis of a complex area. it represents an enormous amount of previous research that has been brought together by people who understand the geology and the geological relations better than anyone, and who also have a geographically broad perspective within which to place the North Cascades. The area is important to the making of the PNW, and to observations bearing on ridge-continent interaction and terrane accretion.

Practically speaking, the compilation of observational support for a PNW history will be useful to many, and their synthesis of the tectonic history through the use of tectonic reconstructions provides a good reference model for the area.

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Detailed comments for the authors to consider-

53. end of line needs a word or to two to make sense.

“Coast Mountains” added to clarify wording.

58-61 & 98-100. Another possibility that has been suggested by the Muller and Sigloch groups is what they call the Orcas plate (Clennett et al., GGG, 2020). I don't know what to make of their models; while there seem to be clear errors when looking at a local level of their model, the arguments for a different plate tectonic setting in the NE Pacific basin (based on mantle tomography images of ocean slab) demand a different plate configuration and history.

We don’t explore the implications of this recent model in detail, but now refer to the Orcas plate in the introduction (line 59) and Clennett et al. (2020) is cited. The potential Orcas plate is also referred to on lines 104, 110, and 119 of the text and in the caption to Fig. 1.

95. First paragraph of Plate Tectonic Setting. What's the point? And missing is anything about the transport of Baja/BC, which is an important point for the North Cascades. Also, since strongly oblique subduction is required, the subducting plate is more oblique that the Farallon plate. The plate subducting beneath the North Cascades region must be the Kula, Resurrection, Orcas, or whatever plate is north of Farallon. (I think a lot is not known about what was happening west of North America north of the Farallon plate, but the important thing is that a spreading center defines the northern limit of the Farallon plate.).

Reviewer 2 also commented on the transport of Baja BC in his second and third general comments. In response to these comments, we have made our most extensive edits in the sections on “Plate Tectonic Setting” (lines 97-105) and “Restoration of Strike-Slip Faults” (lines 192-201). For the plate tectonic setting we now state that the northern boundary of the Farallon plate was a ridge, regardless of whether the northern plate was the Kula, Resurrection, or Orcas plates, and added a phrase that emphasized the motion of the northern plate was more oblique than that of the Farallon plate relative to N.A.

"…northward translation of the Washington Cascades may have been rapid as the southern part of the Insular superterrane (the Baja BC 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/Myr) during this time (e.g., Engebretson et al., 1985; Doubrovine and Tarduno, 2008; Wright et al., 2015; Fuston and Wu, 2021)."

The paleomagnetic data indicating major dextral translation is addressed more extensively in the section on strike-slip offsets (lines 192 to 201) – see response to Reviewer #2, general comment #3 for more detail

111. To state "One explanation" is too weak, considering the abundance of different observations that support Well's basic interpretation.

Replaced "one explanation for this feature is that" with "probably is."

135. Which rocks in Fig. 3 are being referred to?

The Northwest Cascades system is labelled on Fig. 3., but to clarify we have slightly modified the text so that it reads "Paleozoic and Mesozoic oceanic and island arc rocks and overlapping Jura-Cretaceous marine clastic rocks, which were deformed in the mid-Cretaceous Northwest Cascades thrust system (<u>shown as a single Cretaceous unit on Fig. 3</u>).

137. It should be "which are."

With the comma after "which", we think that the use of the singular for "mélange belt" is proper?

147, 156, 167, 172. Which rocks in Fig. 3 are being referred to? It's hard to go from the text to Fig. 3.

All of the features on these lines are on Fig. 3 and we are not quite certain on how to clarify this figure. We made the following additions to help clarify locations. We did add some words to help guide the reader and added an abbreviation to Fig. 3.

Line 142, … Northwest Cascades thrust system "(shown as a single Cretaceous unit on Fig. 3)"

For line 147, we added "of the central belt".

For line 167, we added "of the western belt" to help the reader find these features on Fig. 3.

For line 172,” Hzf” was added along the southern part of the Hozameen fault in Fig. 3, which compliments the label already present on the northern part.

200-567 'PALEOGENE TECTONIC HISTORY.' A thorough presentation of the relevant observations. I learned a lot.

582-83. I don't know why the approach of Siletzia would help flatten the Farallon slab. I could believe that some of Siletzia that subducted north of the suture on southern Vancouver Island or east of the Crescent formation could provide some local buoyancy. Or that the flat Farallon beneath Wyoming helped hold up the adjacent Farallon.

This material has been deleted. We have also shortened the subsequent text and added a reference to the Laramide. “The shallowing of the slab may be attributable to the rapid subduction of young buoyant lithosphere. Strong suction in the mantle wedge may have played a role, as proposed for the Laramide belt (Humphreys, 2009; O’Driscoll et al., 2009). Note that the Laramide belt in northern Wyoming was directly east of Siletzia at 55 Ma in our reconstruction (Fig. 4).”

585-89. I thought Schellart's idea is that a wide slab acts like a parachute. But since about 85 Ma, the Farallon's northern end (beneath western U.S.) is defined by a ridge, and would not be parachute like. Nonetheless, I think the important thing is that the geological observations support the presence of a flat-slab beneath the PNW at this time.

Our interpretation of the wide slab and shallowing has been deleted in response to Humphreys’ comment.

649. Why does the approach of Siletzia influence magmatism? I can imagine this if Siletzia extended beyond the boundaries of what we know today, and that this area was being subducted. I suspect this was likely. It would be remarkable if the entirety of Siletzia accreted, as though subduction stopped at the first encounter of Siletzia with the subduction zone.

We deleted “approach of” following the reviewer’s comment.

677-699. As I understand it, the North Cascades basins trend into the Pasco basin area, which was very volcanically active at this time (Catchings and Mooney, 1988). This supports the idea that the flareup was not simply arc related, but related to the removal of the Farallon flat slab in some fashion.

We have added the Pasco basin area to our interpretation (lines 707-712) and consulted with the reviewer (Gene Humphreys) to make certain we understood his comment.

“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 NE Oregon (Figs. 2, 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..”

Reviewer #2 (Comments to the Author):

Review by D.J. Thorkelson, January 19, 2023

The submission by Miller et al., on the Paleogene plate tectonic and geologic history of the Pacific Northwest and related areas is very good and should be published with few revisions.

The paper stands out as being well written, with complex ideas expressed simply and effectively. It is a pleasure to read and provides the reader with a near-encyclopedic source of information linked to larger ideas and tectonic models. The paper does not purport to solve all of the interesting local and regional geological problems, but neither does it set it to do so. It is a comprehensive statement on the current state of knowledge and provides a balanced and fair depiction of geological features, history and processes.

Although the paper is good, it could use a little improvement in a few areas, and I hope the authors will take the time to consider my remarks and to act on them.

I have annotated the manuscript using the sticky-notes function. I expect the authors to scroll through the annotated file and read my remarks.

Here, I will summarize four issues that need a little attention.

First, there are a couple of key references missing. I know, not all references can be cited, but please try to include the 2013 paper by McCrory and Wilson (Tectonics) and the 1980 paper by Tom Ewing (Journal of Geology). The Ewing paper is remarkable and is the first paper to try to pull a story of Paleogene evolution together. The McCrory and Wilson paper is a modern view of the region that attempts to do much of what the current submission does. The reader should be alerted to both.

We now cite the papers omitted from the introduction. Ewing (1980) is cited on line 69 of the introduction. This paper was cited in 5 other places in the original manuscript. The paper by McCrory and Wilson (2013) is now cited on line 67 of the introduction and line 117 elsewhere in the text.

Second, the plate tectonic setting is reasonably covered for the Paleogene, but that information could be set within a broader context, more clearly than it is. Specifically, the interaction between the Kula-Farallon ridge and North America in the Paleogene is quite an easy sell, and has been well utilized from Thorkelson and Taylor (1989) onwards. However, the work by Engebretson and colleagues in the mid-1980s and Woods and Davies in 1982 (and subsequently Thorkelson and Taylor) show that the Kula-Farallon ridge began as an oceanic rift in the late Cretaceous, circa 1983, and that the Kula-Farallon ridge would have intersected the western North American margin from that point on - until plate consumption and reorganization in the Eocene. However, one would never know that from the current submission. Instead, the notion of Paleogene ridge subduction is moved to the front of the discussion without being placed in a broader, longer tectonic history. I sincerely hope that the authors do their readers a favour by making the earlier history of K-F-NA interactions a little better known, even if all the *answers regarding exact locations and consequences are imperfectly known. The Paleogene ridge system was connected both physically and thematically to the late Cretaceous one. A possible diachronous scenario was provided by Thorkelson and Taylor and, although it need not be taken as unique, it does provide a foundation that could and should be built upon.*

We have added material on the Late Cretaceous history in the plate tectonic setting (lines 97-102) and in the caption to Fig. 1 (figure on tectonic models). "There is general agreement that the Kula plate originated from rifting of the Pacific plate at ~83 Ma and that the northern boundary of the Farallon plate was a ridge, which intersected the continental margin at a poorly constrained location (e.g., Atwater, 1970; Wood and Davies, 1982; Engebretson et al., 1985; Stock and Molnar, 1988; Thorkelson and Taylor, 1989)." The caption now states "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 a ridge in either scenario interacted with the continental margin back to ca. 83 Ma (e.g., Engebretson et al., 1985; Thorkelson and Taylor (1989)."

Third, the thorny notion of dextral translation is a subject that the authors take on, and kudos to them for including it in the geological history and restorations. However, I think they could be a little more forthcoming by including a brief statement on the broader aspects of the dextral translation debate, specifically the much larger magnitudes of displacement indicated in paleomagnetic studies (and I believe in some detrital zircon studies). The authors are basing their estimates of translation on studies of specific fault zones, and that is a adequate approach to take. However, distributed strain along much small features is also worth considering, and the 1000-3000 km (or more) magnitudes are still, in some people's minds, not out of the question. Adding up displacement on specific faults will, most likely, provide only a minimum. Look at the GPS data for the region and marvel and how much displacement is currently taking place along distributed minor or cryptic structures. I am not asking the authors to re-do their estimates, but they should be more open on this subject and write a few sentences about the fact that the actual amounts of overall displacement, and how they vary from east to west, and temporally from mid-K to Eocene, are not perfectly understood, and may be much larger than fault-only based restorations.

The following sentences were added to the section on lines 193-201 to satisfy this comment. "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 ~85–55 Ma of 2000 km or more between the easternmost part of the eastern belt and the central and western belts, and perhaps 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)." An additional phrase was also added to the introduction to the pre-Paleogene geologic setting on lines 138-39. "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), <u>and potentially a much greater distance</u> as discussed below."

Fourth, the shape of the slab window as shown in their Figure 6 is adequate but is not specifically mentioned in the caption. In fact, the three frames, A, B and C, are not individually addressed in the caption, and must be. For the slab window shape, the authors, I will bet, did not do any modeling of their own, and instead have chosen to show a stylized one. That is OK, but tell the reader exactly that - not modeled, but schematic, based on ... give references.

Each frame is now addressed, as was the caption for Fig. 5. Added to the caption – “The approximate location of the idealized slab window assumes that the Farallon plate was moving NE and the Kula-Farallon ridge intersected the continental margin near the Oregon-Washington state boundary.”

Response to Comments on pdf by Derek Thorkelson (Reviewer 2)

LINES

20. *Is this a new finding or have others made the call, and you are supporting it?*

This lull has been pointed out by two of the authors (Miller et al., 2016; 2021) in two papers published in Geosphere. A flat slab has been proposed by others for the Pacific Northwest, but relating it to a flat slab is new in this manuscript. To clarify without adding details in an abstract, the text now reads “interpreted to reflect” rather than “we attribute”.

71 and 94. Compliments – no response needed (though appreciated).

94. – *The 1980 Journal of Geology article by Tom Ewing should be cited in this introduction. It was the first paper to cover the Paleogene story on both sides of the international border and was the first, I believe, to illustrate core complex formation at that time*. Ewing (1980) is now cited in the introduction on line 69. The paper was cited four times in the text in the original manuscript and these citations remain.

94 – *Surely the paper by McCrory and Wilson, Tectonics, 2013, should be cited as a previous attempt to make sense of the Paleogene history in the region.*

McCrory and Wilson (2013) is now cited on lines 67 and 117.

108. *Yes, but it would appear that the K-F triple junction first formed when the K plate broke from the Farallon in the 85-80 Ma range, and so the Paleogene history of ridge subduction and slab window formation is likely to be part of a continuum of interactions extending back to the mid-Cretaceous. An entirely plausible model was put forward by Thorkelson and Taylor in 1989. It accounts for the younging of the adjacent oceanic slabs and the subsequent flat-slab configuration. I bet readers would benefit from knowing that the Paleogene slab window history belongs to a longer history. Just inform the reader that the Paleogene situation is part of a broader history that goes beyond "transpression."*

We have added the following about the Cretaceous history and the intersection of a ridge bounding the Farallon plate with North America. “There is general agreement that the Kula plate originated from rifting of the Pacific plate at ~83 Ma and that the northern boundary of the Farallon plate was a ridge, which intersected the continental margin at a poorly constrained location (e.g., Atwater, 1970; Wood and Davies, 1982; Engebretson et al., 1985; Stock and Molnar, 1988; Thorkelson and Taylor, 1989).”

109. Better: *"Formation of the Siletzia Terrane was a major...".* We now start the sentence with “Formation of the Siletzia terrane” as suggested.

113. *Infer? Rather, "we support previous work that...."*

Now reads: “We support previous work that infers the triple junction”

185. *It would be good to tell the reader that your geologically based estimates are lower than those identified by paleomagnetism. The Spences Bridge Group may have been 1000 km farther south when it formed, and units farther west have been shown to have greater displacements from the south. I don't think you should feel obliged to make a comprehensive review of the information (which includes detrital mineral studies), but you should alert the reader that the estimates of translation from the south are on the low end, given all the information available. Displacement estimates on specific faults and fault sets will likely be less than the sum of those displacements plus distributed strain from less prominent.*

This comment amplifies the third general comment by Reviewer Thorkelson. The potential for much larger offsets of outboard rocks is now addressed on lines 193 to 201. See the response to the third general comment above.

206*. Again, since the Kula broke from the Farallon at about 83 Ma, it stands to reason that there would have been a K-F-NA triple junction and slab window somewhere along the coastline. No, we don't know exactly where it was, but do we know it wasn't near, or within, the study area?*

This paragraph at the beginning of the section focuses on the geologic evidence for oblique convergence. We hope that the material on lines 97-102, which mentions the earlier history, including the uncertain location of the triple junction, and the addition to the caption for Fig. 1 are sufficient. “There is general agreement that the Kula plate originated from rifting of the Pacific plate at ~83 Ma and that the northern boundary of the Farallon plate was a ridge, which intersected the continental margin at a poorly constrained location (e.g., Atwater, 1970; Wood and Davies, 1982; Engebretson et al., 1985; Stock and Molnar, 1988; Thorkelson and Taylor, 1989).” The caption now states “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 a ridge in either scenario interacted with the continental margin back to ca. 83 Ma (e.g., Engebretson et al., 1985; Thorkelson and Taylor (1989).”

241. *But no slab window despite the lack of proper arc magmatism?*

We are not certain if this is a parenthetical statement. This section describes the basic geologic features and reports on work by others on petrogenesis without focusing on tectonic setting.

325. *Surely, the K-F ridge is responsible for the younging and shallowing of the F and possibly the K slabs. So, where was the continental triple junction? Can you show that it wasn't involved? Doesn't seem like a bad fit to me.*

We didn't change this paragraph as we suggested that low-angle subduction accounted for the cessation of arc plutonism. Younging of the subducting slab is proposed for the cause of the low angle in the discussion and references to others whom have stated this for the greater region are cited (lines 597-599). "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)."

418. *See attachments to this review -- Thorkelson 1995 GAC abstract and figures.*

We appreciate getting the abstract and figure. We mention "passage of a slab window" as one of the mechanisms proposed for Challis-Kamloops magmatism.

420. Humphreys now spelled correctly.

438. Are we certain that the triple junction and slab window had nothing to do with these features? OIB -- a common composition of slab window volcanism -- see review in Thorkelson, Encyclopedia of Geology article on slab windows.

Lines 434-5 mentioned a slab window as one of the mechanisms proposed by others in the original text. We emphasize this a little more with the underlined text. "Slab rollback and breakoff, <u>and slab window evolution</u> are the most widely cited scenarios (see review by Humphreys and Grunder, 2022)." We deleted the phrase stating our preference in this paragraph, which just summarizes models in the literature.

516. Changed "but" to "and" as requested.

579. We are not clear what the "and/or" refers to

582. *"ridge close to the trench"... a little vague. Knowing what we do about the K and F Cretaceous history, what is meant by "close to the trench"? What argument do you know of that would position the ridge offshore, and not in contact with the continent? It would seem unavoidable that the ridge would have intersected the continental margin, so, again, what does "ridge close to the trench" even mean?*

Reviewer Humphreys also criticized this idea and the statement has been deleted and a few other lines have been added in response to his comment (see new lines 597-602). Also see the response to Humphreys' comment above.

596. *Cite the papers which dealt with this and proposed a location something like this. This is not something new*.

The following reference is added (underlined) "in which case the latter would be kinematically tied to the Tintina fault – Rocky Mountain trench (Price and Carmichael, 1986) and magmatism would occur in a slab window (e.g., Breitsprecher et al., 2003)." We note that our interpretation is based on our strike-slip restorations and matching up parts of the southern Coast Mountains-North Cascades Cretaceous arc using compilations of recent U-Pb ages, which is new.

598. Made the minor wording change.

679. *I'm going to include an abstract with key figures from a 1995 GAC talk for you to consider.*

We have now stated that the magmatism to the east has been attributed to a slab window and reference Thorkelson and Taylor (1989) and Breitsprecher et al. (2003). We then give our preferred rollback and breakoff model, which takes into account new age data from NE Washington and the Cascades, as well as the tomographic information from Schmandt and Humphreys (2011).

"The Eocene Cascades core plutons have been considered 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 occur 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."

1395. *In this model -- OK -- but say where it comes from*.

Bradley et al. (2003) is now cited for model A and Haeussler et al. (2003) for model B. Madsen et al. (2006) is cited for location of ridge intersection on Vancouver Island.

1396. *from inclusion of the*

Phrase added as requested.

1399. *History from 83-60 not shown, and that's OK, but somewhere in the paper you need to let the reader know that a RTT triple junction did exist prior to 60.*

Please see the response to the second general comment.

1454. *You need to have separate statements about each of the frames, A, B and C.*

Each frame is now described separately. Also, please see response to general comment #4.

UPPER PLATE RESPONSE TO RIDGE SUBDUCTION AND OCEANIC PLATEAU ACCRETION, WASHINGTON CASCADES AND SURROUNDING REGION: IMPLICATIONS FOR PLATE TECTONIC EVOLUTION OF THE PACIFIC NORTHWEST (U.S.A. AND SW 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 and southern British Columbia in the early-middle Eocene and we summarize a variety of new datasets 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. 60-50 Ma ~~that we attribute~~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-Kula or Farallon-Resurrection spreading center. Farther east, peraluminous magmatism occurred in the Omineca belt and Idaho batholith. Accretion of Siletzia and ridge-trench interaction occurred between ca. 53–49 Ma, as indicated by: (i) near-trench magmatism from central Vancouver Island to NW Washington; (ii) disruption and inversion of the Swauk basin during a short-lived contractional event; (iii) 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 (iv) southwestward migration of magmatism across NE Washington. These events suggest that flat slab subduction from ~60–52 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 Ma 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 8-10 kb metamorphic rocks in the North Cascades crystalline core. By ca. 45 Ma, the Farallon – Kula (or Resurrection) – North American triple-junction was likely located in Oregon, 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.

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 (U.S.A) 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 Cascades 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 about 96–60 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 – 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 – Kula, ~~or~~ Farallon – Resurrection, or Farallon – Orcas 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 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 east-west 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, 1984; 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, 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 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, 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; b; Miller et al., 2016, 2022), which enables the construction of a detailed time line 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 ~83 Ma and that the northern boundary of the Farallon plate was a ridge, which intersected the continental margin at a poorly constrained location (~~, including the location of the Kula – Farallon – North America triple junction (~~e.g., Atwater, 1970; Wood and Davies, 1982; Engebretson et al., 1985;~~-~~ Stock and Molnar, 1988; Thorkelson and Taylor, 1989). ~~and~~ 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 ~~Late Cretaceous to earliest Cenozoic,~~ from ca. 85 Ma 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 BC 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 (~~and~~ 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/Myr) 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~~a major factor in the Paleogene tectonic evolution of the Pacific Northwest ~~was the formation of the Siletzia terrane~~. This terrane represents a large igneous province that developed between 56-49 Ma near an oceanic spreading center~~.~~, and it is probably ~~One explanation for this feature is that it represents~~ 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 ~~infer~~ support previous work that infers the triple junction between the~~Kula (or Resurrection) –~~ Farallon – North America – Kula (or Resurrection or Orcas) plates ~~triple junction~~ lay along central Vancouver Island by 55–53 Ma (e.g., Madsen et al., 2006) (Figs. 1, 4). From 52–49 Ma, a triple junction is interpreted to have interacted with the continental margin along central to southern Vancouver Island (Fig. 1), as 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, 2021), and disruption of non-marine basins (Eddy et al., 2016a). The collision of Siletzia, which started by 53 Ma~~a~~ in SW 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 52–48 Ma, which we describe in more detail below. The plate boundary later shifted outboard (west) of Siletzia,

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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).

PRE-PALEOGENE GEOLOGIC SETTING

Prior to 65 Ma, the Pacific Northwest was characterized by a typical convergent margin with a forearc, continental magmatic arc, back-arc 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 ~~at least~~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 oceanic and island arc 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 Ma 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 96–87 Ma, and most biotite Ar/Ar and K/Ar cooling ages are >60 Ma, whereas magmatism in the Chelan block ranges from 92–45 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 d~~D~~irectly east of the North Cascades in the eastern belt~~, pre-Cenozoic rocks~~ include: the Mesozoic Methow basin; ca. 160–105 Ma arc plutonic rocks of the Eagle 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 plutonic and metamorphic rocks of the Omineca belt, including multiple metamorphic core complexes, the Idaho batholith, and Cordilleran passive margin sediments involved in the Rocky Mountain-Sevier fold-and-thrust belt (Fig. 2).

RESTORATION OF STRIKE-SLIP FAULTS

Dextral strike-slip faulting occurred in the northern 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 – 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 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 NE 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 ~115 km of dextral offset (Umhoefer and Miller, 1996).

These known displacements of ~325 km must be considered in tectonic restorations, particularly before 50 Ma. To summarize, after 50 Ma there is approximately 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) 50 km (of total 115 km) of offset between the core and the eastern belt; and 4) a cumulative offset of ~265 km between the western and eastern belts after 50 Ma (Table 1). If we assume that the strike-slip offset from 60 to 50 Ma occurred at rates comparable to those of the ~50–40 Ma interval, the implication is that another approximately 250–300 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~~show Vancouver Island and the western belt about 450 km south of 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 ~85–55 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).

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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 Klamath Mountains northward to the Olympic Peninsula. 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 northeastern-most 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 km or less.

PALEOGENE TECTONIC HISTORY

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

65 – 60 Ma

During this 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 Mountain 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 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, as the upper limit for the mélange is only indicated 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 (up to 12 kbar) Swakane Biotite Gneiss (Valley et al., 2003) and Skagit Gneiss Complex (7 – 10 kbar; Whitney, 1992; Hanson, 2022) (Fig. 3), between ca. 79 – 66 Ma and 74 – 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 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 NE 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 Complex (Fig. 4) initiated at 62 Ma (Carr, 1992; Hinchey and Carr, 2006). In Idaho, 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 NE 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 Complex and in the Rocky Mountain - Sevier fold and thrust belt (e.g., Price, 1981).

60 – 52 Ma

This 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 (southwest Oregon) and by 53.2 Ma outboard of the Northwest Cascades system and mélange belts in western Washington and Vancouver Island in the north, where it continued until at least 48 Ma

(Crescent and Metchosin basalts) (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-oceanic-ridge basalt (N-MORB) to shallow water and subaerial flows of enriched mid-oceanic-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 hot spot (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 maximum depositional age (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 Ma 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 – 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 ~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 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, 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, 2021). The alkalinity of magmas increases markedly south of ca. 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., Parrish et al., 1988; Harlan et al., 1988; Brown et al., 2012), led to the formation of metamorphic core complexes and associated extensional basins in NE 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 NE 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 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 Ma 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 northwest Washington by near-trench magmatism from ca. 51 – 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 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 (Clayquot 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 T/low P metamorphism at ~51 Ma (Fairchild and Cowan, 1982; Groome et al., 2003). In NW Washington, local peraluminous magmatism occurred as the ca. 49 Ma Mt. 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 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 SW 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 Ma 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 Myr) 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 SW 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 Gneiss in the crystalline core was probably rapidly exhumed during this interval, in part during distributed ductile shear and top-to-N to –NNE motion on the Dinkelman decollement (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 Schriazza, 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/YbN>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 SW across NE Washington (Fig. 6C) (Tepper, 2016). Metamorphism and deformation continued in the metamorphic core complexes in southern British Columbia, NE 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; Suydan and Gaylord, 1997; Foster et al., 2007; Brown et al., 2012; Rubino et al., 2021). In contrast to NE 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 east-west 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 NW, and the southwestward migration of magmatism in NE Washington is consistent with rollback of the northern Farallon slab (Figs. 5, 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, 2016). Slab rollback and breakoff, and slab window evolution are the most widely cited scenarios (see review by Humphreys and Grunder, 2022)~~, and we favor this interpretation as discussed below~~.

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 this 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 aerially 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 nonmarine sedimentary rocks in the Naches and Barlow Pass units (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 ~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 ~46.5 Ma, and fault reorganization at ~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 ~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 (ca. 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 ~47.9–46.5 Ma, magmatism in the core migrated westward from the Ross Lake fault zone. The ~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., 2021), 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). εNdi 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. 49.3–47 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. 49.5–45 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 to 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 north motion continued on the Dinkelman decollement 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 Gneiss was in part exhumed by the Dinkelman decollement 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, ~~but~~ and magmatism and extension were largely waning by the end. Igneous activity was still migrating southwestward across NE Washington (Fig. 6C). In British Columbia, the >200 km^2, granodioritic Needle Peak pluton intruded the Methow basin at ca. 48 Ma (Monger, 1989), ~~but~~ and Challis-Kamloops magmatism to the east had largely ended by ca. 47 Ma (Ickert et al., 2009; Dostal and Jutras, 2021).

Extension and sedimentation related to the metamorphic core complexes in NE Washington and British Columbia were on the wane during this interval. Termination of sedimentation at ~48 in NE 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. 50–48 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; Howlet 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

This interval marks the end of slab foundering and the establishment of a new north-south 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, as did Challis-Kamloops magmatism in the eastern belt.

Arc magmatism began at ca. 45 Ma in southwest 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 SW 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 NW Oregon and SW 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 Ma 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. 60–50 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 subduction of~~

~~young buoyant lithosphere~~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). ~~generated at a ridge close to the trench, and by the approach of overthickened oceanic crust of Siletzia.~~ Strong suction in the mantle wedge ~~due to approach of the slab with its decreasing dip toward the craton~~ may have played a role, as proposed for the Laramide belt ~~to the south~~ (Humphreys, 2009; O'Driscoll et al., 2009). Note that the Laramide belt in northern Wyoming was directly east of Siletzia at 55 Ma in our reconstruction (Fig. 4). ~~Another explanation for the postulated flat-slab subduction is that the Farallon subduction zone was old and wide (Schellart, 2020), which is viable if the flat slab before an approximately 60 Ma plate reorganization was a smaller part of the northern Farallon plate subducting beneath western North America since at least the earliest Cretaceous (Engebretson et al., 1985).~~

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 61–48 Ma (Cecil et al., 2018). A projection of the triple junction off central Vancouver Island through the boundary in the Coast Mountains to the NE 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 ~~underlie~~ 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 be kinematically tied to the Tintina fault – Rocky Mountain trench (Price and Carmichael, 1986) and magmatism would occur in a slab window (e.g., Breitsprecher et al., 2003). Seismic tomography and reconstructions of plate motions in the NE Pacific also suggest a major boundary inboard from Vancouver Island (Fuston and Wu, 2021). Plate motion models ~~suggest~~ indicate rapid northward rates of either the Kula or Resurrection plates from ca. 65–50 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 NW edge of the Klamath – Blue Mountains terranes and the flat slab beneath the Pacific Northwest would be continuous with the well-established Laramide flat slab to the south (see Tikoff et al.~~,~~ [2023~~2~~]~~,~~ for an alternative hypothesis~~s~~).

Consequences of Collision of Siletzia

The inferred position of the intersection of the Farallon – Resurrection/Kula ridge with the trench is complicated by the eruption of Siletzia basalts and the construction of an oceanic plateau above a hot spot 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 southwest 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 ~50.8 Ma and 49.3 Ma (Eddy et al., 2016a). The reversal of drainage in the Swauk basin at ~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 Ma 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 ~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 underway in Washington by ca. 52 Ma (Figs. 5, 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 SW.

The ~~approach and~~ collision of Siletzia with the continental margin influenced magmatism much farther eastward than it influenced deformation and sedimentation. We attribute this to the shut off of northeastward flat subduction caused by the collision-related plate reorganization (e.g., Schmandt and Humphreys, 2011). Magmatism migrated to the southwest across NE Washington and reached the Golden Horn batholith at the northeast margin of the Cascades core at ca. 48.3 Ma (Figs. 3, 6C). This migration has been interpreted to result from slab rollback (Tepper, 2016) 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 ca. 70 km into Canada (e.g., Needle Peak pluton), volcanic rocks on the west and south sides of the core, and voluminous dike swarms (Figs. 3, 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 backarc 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 break-off and ridge subduction, and largely involves mantle-derived melts (e.g., Decker et al., 2017; Schwartz et al., 2017; 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 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 occur within a slab window (e.g., Thorkelson and Taylor, 1989; Breitsprecher et al., 2003). ~~but~~ I~~i~~n our view, the flare-up is related to the Farallon slab rollback and breakoff. At ~49.5 Ma, the southwest-migrating rollback magmatism had reached the northeast margin of the Cascades core (Tepper, 2016) 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, 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 break-off origin as the crustal component of melt during break-off would be 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 NE Oregon (Figs. 2, 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. ~~A speculative additional interpretation is that the breakoff-related magmatism continued to the southeast to the Clarno Formation of NE Oregon (Figs. 2, 6).~~ 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 about 100 km SE of the North Cascades flare-up and the western breakoff belt west of the Straight Creek fault would have been about 40–50 km closer to the Clarno 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 southeast 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 Fraser River-Straight Creek 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 NE 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 NE Washington range between 51–48 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 Fig 1). These dikes are subparallel to the normal faults that separate the Kettle and Okanogan core complexes from Eocene grabens (Keller, Republic, and Toroda), 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-east-southeast (~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 (ca. 15°–50°

counter clockwise) to those of the voluminous N–NE-striking (average of 035°), ~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, 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 W-NW – E-SE (~295-115°) in the Okanogan Complex (Kruckenberg, 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 Gneiss. Thus, ductile extension directions rotate progressively clockwise by ~75° from east to west. The sense of rotation is the same, but the magnitude of rotation is greater, then 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 ~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; Vanderhaege 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 ~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-Kermedec and Izu-Bonin-Mariana system occurring at ca. 53–50 Ma (Sharp and Clague, 2006; Whittaker et al., 2007a; Tarduno et al., 2009). The ~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 Ma 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.

ACKNOWLEDGEMENTS

This research was supported by National Science Foundation grants EAR-1119358 and EAR-1945352 to R.B. Miller, EAR-1945260 to M.P. Eddy, EAR-1119252 to J.H. Tepper, EAR-1119063 to P.J. Umhoefer, and EAR-1118883 to S. Bowring. Our late colleague and friend Paul Umhoefer was a driving force for much of this synthesis and particularly the fault reconstructions. The late Sam Bowring also played an important role in the research presented here. M.S. and B.S thesis students at Northern Arizona University, San Jose State University, and the University of Puget Sound contributed to our research project and include Katie Bryant, Erin Donaghy, Melissa Gunderson, Lisa Kant, Jen Pence, and Francesca Senes. We thank Stacia Gordon, Margi Rusmore, and Noah McLean for helpful discussions. We thank Gene Humphreys and Derek Thorkelson for their thorough and very helpful reviews of the manuscript.

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TABLE

Table 1. STRIKE-SLIP FAULT OFFSETS ACROSS WASHINGTON CASCADES.

FIGURE CAPTIONS

Figure. 1. Simple plate reconstruction models of the NE 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 (e.g., Hauessler 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 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 ~50 Ma (Clennett et al., 2020).~~.~~

Sanak-Baranof 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).

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 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 Fig. 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 Formation; Ccb=Chilliwack batholith; CHK=Chuckanut Formation; CM=Cooper Mountain pluton; CP=Castle Peak stock; CRb=Columbia River Basalt Group; DD=Dinkelman decollement; DH=Duncan Hill pluton; Ef=Entiat fault; Epc=Eagle Plutonic Complex; FDfz = Foggy Dew fault zone; FRfz=Frazer

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; SW=Swauk Formation; SWG=Swakane Gneiss; T=Teanaway Formation; WEMB=Western and eastern mélange belts; Yi=Yale intrusions.

Figure 4. 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 which Siletzia formed on 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. 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.

Figure 5. A. Reconstruction map ~~and proposed plate tectonic setting~~ 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. 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.

Figure 6. A. Reconstruction map ~~and proposed plate tectonic setting~~ 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 intruded in the central and northern Washington Cascades coincident with a magmatic flare-up. Challis-Kamloops magmatism continued, but is beginning to wane, as is extension associated with core complexes. B. Proposed plate tectonic setting. The approximate location of the idealized slab window assumes that the Farallon plate was moving NE and the Kula-Farallon ridge intersected the continental margin as shown. ~~In~~ C.~~,~~ Eocene magmatism across NE to

north-central Washington and the pattern of inferred rollback ~~of~~ magmatism to the southwest are shown.~~n~~ Filled circles are localities of U-Pb zircon ages of plutonic rocks.~~, as inferred from~~ 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; Chb = Chumstick basin; Csz = Coast shear zone; Ef = Entiat fault; Fa = Farallon plate; Ff = Fraser fault; K = Kettle core complex; LCfz = Lewis and Clark fault zone; Ok = Okanogan core complex; P = Priest River core complex; Pab = 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: BC = British Columbia; ID = Idaho; MT = Montana; NV = Nevada; UT = Utah; WA = Washington.

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. VI = Vancouver Island; Yf = Yalakom fault. States and Provinces: BC = British Columbia; ID = Idaho; MT = Montana; OR = Oregon; WA = Washington.

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.

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.