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Research Article

Late Middle Miocene Emergence of the Olympic Peninsula Shown by Sedimentary Provenance

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The Olympic Peninsula is the uplifted portion of the Cascadia accretionary wedge and forms the core of a 200 km scale oroclinal bend on the west coast of Washington State. The accretionary wedge started forming 45 million years ago following the accretion of the Siletzia igneous province along the Cascadia subduction margin. Low-temperature thermochronology studies have shown that the core of the peninsula has been continuously exhumed for the last 14 million years. The earlier onset of oroclinal bending, uplift, and emergence remains poorly documented. Here, we explore the Cenozoic drainage history of the Cascadia forearc and accretionary wedge to reconstruct the deformation history of the Olympic Peninsula. We use detrital zircon provenance and grain petrography data from modern rivers draining the Cascades, the Cascadia forearc and accretionary wedge, as well as from Eocene to late middle Miocene sedimentary units from the same areas. We first show a clear difference of sedimentary provenance between sedimentary units in the accretionary wedge, with older units reflecting mélange and imbricated strata that began as part of Siletzia, and younger units reflecting trench-fill material sourced from the Cascades and accreted to the wedge. We show that the accretionary wedge was directly fed from the Cascade arc until at least 16.5 ± 0.5 Ma, providing a maximum age for the emergence of the Olympic Peninsula. Fluvial deposits in the Cascadia forearc basin dated at 13.3 ± 1.3 Ma display zircon age spectra and sedimentary grain petrography features typical of recycled accretionary wedge material. Although these deposits may also reflect local input, middle Miocene exhumation rates suggest the Olympic Peninsula was an active sediment source. Our results bracket the timing of emergence of the Olympic Peninsula to a narrow window in the late middle Miocene. We suggest that the initial onset of accretionary wedge deformation and oroclinal bending predates this by at least 10 million years, in the upper Oligocene, and is marked by flexural subsidence and high sedimentation rates recorded in strata of the Seattle Basin. Our results support a composite history for the development of the Cascadia accretionary wedge rather than models predicting a gradual and steady build-up.

1. Introduction

The Olympic Peninsula of Western Washington State is the subaerially exposed portion of the Cascadia accretionary wedge [1, 2], which developed following the Eocene accretion of the Siletzia igneous province along the west coast of North America [3, 4]. Substantial deformation and uplift of the Cascadia accretionary wedge have resulted in a large oroclinal bend with the Olympic Mountain range at its core [5]. These mountains are unique in that they are larger and higher than all other coast mountain ranges formed by accretionary wedges on the Pacific Coast of North America. Low-temperature thermochronology studies have shown that the core of the peninsula has been continuously exhumed for the last 14 million years, though the tempo and evolution of exhumation rates through recent times remain debated [1, 6–10]. Rock uplift has been proposed to be driven by crustal thickening related either to margin-parallel shortening [11], frontal accretion [7, 10], or underplating at the trench [12]. High

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precipitation rates along the coast enhance exhumation, contributing to the shape and rate of uplift [13].

The actual timing of surface uplift of the accretionary wedge remains virtually undocumented. It is unclear when the accretionary wedge started forming a submarine high, creating a topographic barrier between the Cascadia trench and the forearc, and when the peninsula finally emerged. Documenting the chronology of these events is essential to reconstruct sedimentary fluxes into the trench, to understand how fast exhumational steady state was reached in the Olympic Mountains, and to identify the main drivers of rock uplift [12].

Here, we use detrital zircon provenance and sandstone petrography to place age constraints on the uplift and subaerial exposure of the Olympic Mountains. We use U-Pb zircon ages and petrographic data from modern rivers of the Olympic Peninsula and the North Cascades to determine the provenance character of modern drainages. We then use the same methods on Cenozoic sedimentary rocks of the Cascadia forearc and accretionary wedge to reconstruct the regional drainage history and track the emergence of the Olympic Mountains as a topographic high.

2. Geologic Setting

2.1. The Olympic Peninsula. The Cascadia accretionary wedge, Cascadia forearc, and the Cascade Range run parallel to the Cascadia subduction zone, where the Gorda and Juan de Fuca Plates subduct beneath the North American Continental Plate. The Olympic Peninsula is the portion of the Cascadia accretionary wedge that sits on the west coast of Washington State, separating the Puget Lowland from the Pacific Ocean with a high mountain range. The Puget Lowland is the portion of the Cascadia forearc in Washington, which is bounded by the Olympic Peninsula to the West and the Cascade Range to the East, draining sediment from both. In Canada, the forearc forms the Georgia Basin, bounded by Vancouver Island to the West and the Coast Mountains to the East, the northward continuation of the Cascade Range (Figure 1). Reference to "the Cascades" in this manuscript specifically regards the North Cascades that extend from Central Washington to southern British Columbia. The Cascades comprise an exhumed metamorphic core, Mesozoic to Eocene plutons—the Coast Mountains Batholith—and an Andean-type volcanic arc that has been active since the middle Eocene—the Cascade Arc [14].

The seminal study of Tabor and Cady [2] provides detailed structural and geological description of the Olympic Peninsula and has guided all other studies to follow. The peninsula is composed of two primary terrains: peripheral rocks that comprise early Eocene basalts of the Crescent Formation overlain by little-deformed Eocene to Miocene, mostly marine sedimentary rocks, and a core of extensively deformed marine sedimentary rocks called the Olympic subduction complex (OSC) by Brandon and Calderwood [1].

The Crescent Formation is in fault contact with the OSC along the Hurricane Ridge Fault, wrapping around the east of the Olympic Peninsula to form a crescent shape (Figures 1 and 2; [15]). The Crescent Formation comprises primarily thick basal pillow and columnar basalts overlain by felsic volcaniclastic sandstones, lignite, and mudstones [16]. K-Ar hornblende dating, $^{40}Ar/^{39}Ar$ whole rock ages, and U-Pb dating of felsic tuffs constrain the eruption of the Crescent Formation to between 53.2 and 48.4 Ma [4, 17]. Some have proposed slab window volcanism as a result of ridge subduction to be the source of the Crescent Basalt [18, 19]. An alternative hypothesis has linked the Crescent Formation to a flare-up of the long-lived Yellowstone hotspot that resulted in the eruption of a marine large igneous province [4, 20, 21]. The latter model groups together contemporary basalts from Oregon, Washington, and British Columbia, including the Crescent Formation, into the Siletzia terrane; the Siletzia terrane would have erupted as a marine large igneous province, forming a submarine plateau on the Farallon plate before being accreted to North America in the early Eocene [4, 17]. Regardless of which model for the origin of the Crescent Formation is the most accurate, the grouping of the Siletzia terrane and accretion timing are well agreed upon. In this manuscript, "Siletzia" is used to refer to the entire accreted terrane, and "Crescent Formation" is used in reference to rocks of the Siletzia terrane that crop out in Western Washington.

The first marine sediments were accreted to the Cascadia accretionary wedge beneath the Siletzia terrane in the Eocene, following the initiation of the Cascadia subduction zone, which caused significant slip along the Hurricane Ridge Fault (tens of kilometers; [7]). The initiation of the Cascadia subduction zone is dated at ca. 42-45 Ma based on the age of the oldest post-Siletzia accretion magmatic rocks in the Cascade Arc [14, 22]. The accreted sedimentary rocks that have been underthrust beneath the Crescent Formation form the core of the Olympic Peninsula today, and they have been thoroughly mapped [2, 15]. These rocks were initially divided into five informal lithic assemblages: the Hoh, Elwha, Western Olympic, Grand Valley, and Needles-Gray Wolf assemblages (Figure 2; [2]). Although these rocks have been mapped with high precision, the extreme amount of deformation and displacement has made it difficult to discern stratigraphic relationships between the five assemblages; in general, the rocks become younger from East to West [2, 15, 23]. Brandon and Vance [23] grouped rocks of the Cascadia accretionary wedge into the Olympic subduction complex (OSC) and reorganized the five lithic assemblages into three structural units using thermochronological and structural data: the coastal OSC, equating to the Hoh assemblage; the lower OSC, grouping the Western Olympic and Grand Valley assemblages; and the upper OSC, grouping the Needles-Gray Wolf and Elwha assemblages (Figure 2; [23]). Stratigraphic and chronologic relationships between these structural units are not clearly defined; biostratigraphic ages for the OSC range from Eocene to upper Miocene, and generally the oldest ages are found in the upper OSC and the youngest in the coastal OSC [24].

The upper OSC is the structurally highest unit of the OSC, and it is interpreted to be the oldest [23]. The upper OSC locally contains pillow basalts that are compositionally similar to the Crescent Formation and are interbedded with fossiliferous limestones that yield latest Paleocene to

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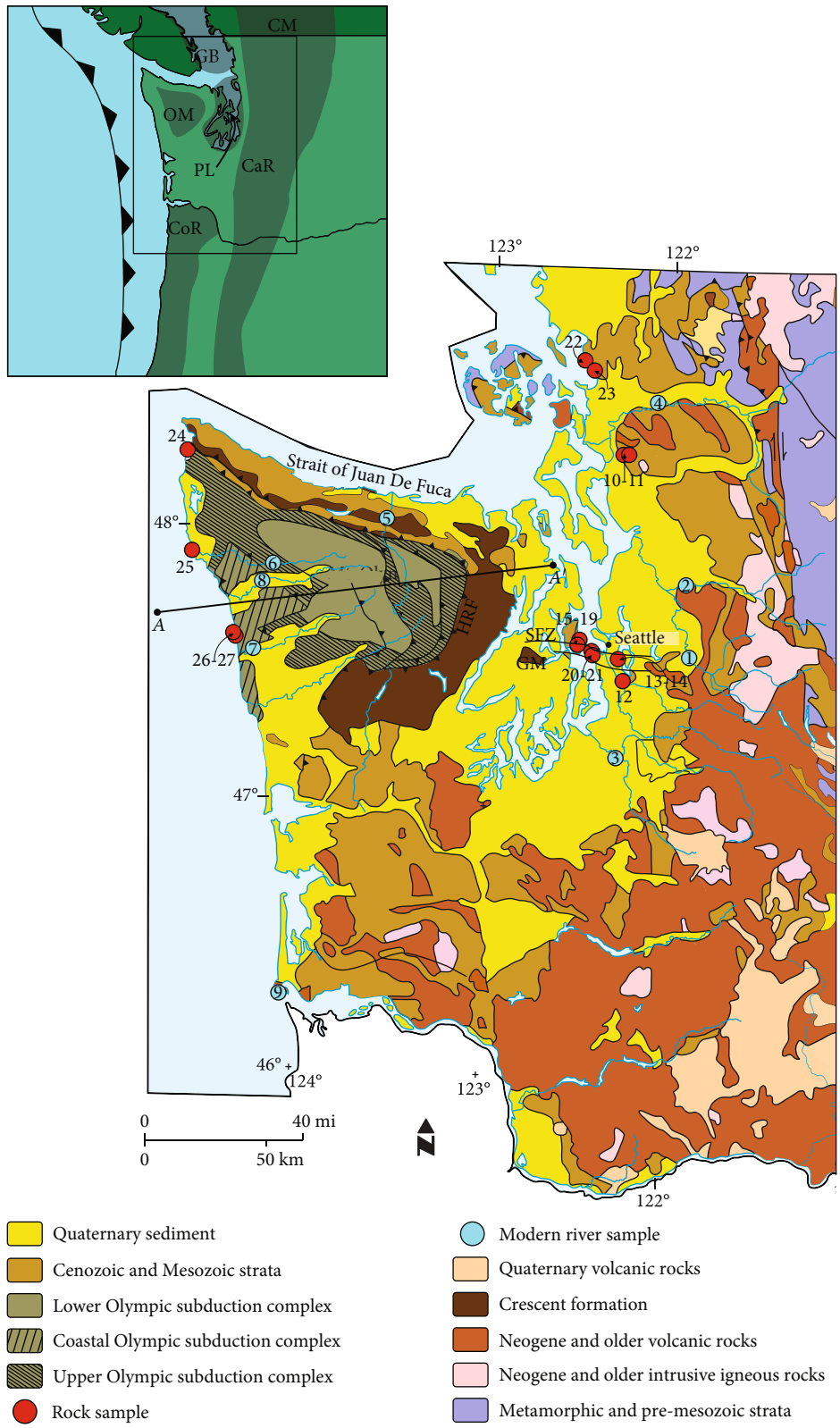

Figure 1: Regional geologic map of the Olympic Peninsula, Puget Lowland, and the eastern extent of the North Cascades (modified from [7, 94]). Sample locations are shown by red dots (rock outcrop) and blue dots (river sands) with sample numbers labelled (see Table 1 for sample numbers). HRF = Hurricane Ridge Fault; SFZ = Seattle Fault Zone; GM = Green Mountain; CaR = Cascades Range; CM = Coast Mountains; CoR = Coastal Ranges; OM = Olympic Mountains; PL = Puget Lowland; GB = Georgia Basin. Paths of major rivers are shown by blue lines. Cross section of A-A' shown in Figure 2.

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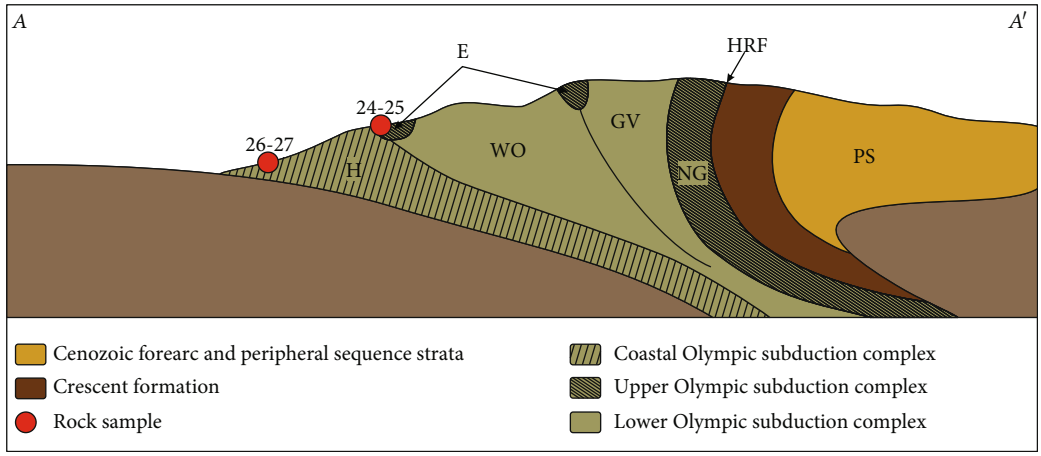

Figure 2: Schematic cross section showing generalized structure of Olympic Peninsula (A-A' line on Figure 1), informal lithic assemblages, and corresponding members of the OSC. H = Hoh; WO = Western Olympic; GV = Grand Valley; E = Elwha; NG = Needles-Gray Wolf; PS = Peripheral Sequence; HRF = Hurricane Ridge Fault.

middle Eocene fossil ages [15, 24]. Detrital zircons from this unit have yielded fission-track maximum depositional ages ranging from early Eocene to early Oligocene (48-32 Ma; [23]). Some localities attributed to the upper OSC yield upper Eocene to lower Miocene microfossils [25, 26].

The lower OSC is distinguished from the upper OSC by the absence of pillow basalts, and only comprises marine clastic sedimentary rocks including turbidite sequences and highly deformed mudstone-rich mélanges. It structurally underlies the upper OSC. Detrital zircons from this unit have yielded fission-track maximum depositional ages ranging from 27 to 19 Ma [23, 27].

The coastal OSC is the structurally lowest unit of the OSC and is lithologically similar to the lower OSC [23]; the two units are differentiated by the higher degree of metamorphism observed in the lower OSC [2, 23]. Detrital zircons from turbidites of the coastal OSC have yielded fission-track maximum depositional ages ranging from 26 to 11 Ma; mélange blocks have yielded maximum depositional ages ranging from 39 to 15 Ma [27]. The youngest fission-track ages, however, display high uncertainties (2 s ± 3 to 4 Ma; [27]).

The three units of the OSC have been attributed different sediment sources and accretionary histories, though little work has been done to quantify these differences [23]. It has been interpreted that the lower OSC represents sediment drained from the Cascades into the subduction trench, while the coastal OSC represents mass wasting of material from the continental slope, with these two units constituting the true accretionary wedge [23, 27]. Brandon and Vance [23] propose that early Eocene deposits of the upper OSC originated as the westernmost clastic strata of the Siletzia terrane. These strata were imbricated and underthrust beneath the Crescent Formation following the initiation of the Cascadia subduction zone. Brandon and Vance [23] also propose that this imbrication could not have occurred until after deposition of the youngest strata of the upper OSC (~33 Ma), which is several million years later than the proposed slip event along the Hurricane Ridge Fault and initiation of subduction (~45 Ma; [7, 14]).

Detrital apatite and zircon fission-track and (U-Th)/He ages from the core of the peninsula show continuous exhumation since at least 14 Ma [7–9, 23, 28]. The exhumation of the OSC has created a wide oroclinal bend, resulting in the arcuate shape of the Crescent Formation [5]. It has been proposed, based on regional-scale plate motion and GPS data, that this oroclinal bend and the unusual elevation of the OSC is caused by margin-parallel motion, which contributes to uplift, with the Olympic Peninsula caught between the northward motion of the Oregon Coast Range and the relatively stable Vancouver Island [11, 29–31]. This margin-parallel motion is evidenced by the numerous southeast- and northeast-striking oblique-slip faults that accommodate trench-parallel shortening and westward block extrusion within the Olympic Peninsula [32–34].

Alternatively, it has been proposed that uplift and oroclinal bending are generated by margin-normal deformation resulting from accretion along the wedge [10, 28]. It has been estimated that 80 to 100% of the sediment from the subducting Juan de Fuca Plate is accreted to the wedge front, and that this accretionary flux is in equilibrium with exhumation rates, with accreted material horizontally traversing most of the wedge before being exhumed [1, 28]. The presence of low-grade metamorphic rocks at high elevations in the Olympic Mountains shows that accretion is likely complemented by underplating of accreted sediment driving material flow upwards within the wedge [1, 7]; however, the contribution of this underplating component remains debated [10, 12]. For these scenarios, the greater elevation of the Olympic Peninsula compared to southern parts of the Cascadia wedge is explained by the concavity of the North American margin where the Juan de Fuca plate subducts beneath Washington; the curvature of this contact forces the underriding slab into an antiformal configuration, shallowing the slab dip beneath the Olympic Peninsula and driving margin-normal shortening and underplating [1, 35, 36]. High sedimentation rates at the trench in the Pacific

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Northwest could also have amplified accretion rates and underplating [7].

Margin-normal and margin-parallel drivers on oroclinal bending are nonexclusive and both could have contributed to shaping the Olympic Peninsula, either at the same time or at different periods. For both types of drivers, the onset of deformation is commonly associated with the beginning of basin and range extension in the middle Miocene; this extension would have changed block motion along the North American margin and enhanced its curvature [1]. Early low-temperature thermochronology studies have suggested that exhumation in the core of the peninsula has been fairly constant at a rate of 0.75 to 1 mm/yr since 14 Ma and is at accretionary steady state [7, 23, 28]. More recent studies have highlighted a more complex exhumation story, with initially high (>2 mm/yr) exhumation rates followed by a sharp decrease to values < 0.3mm/yr at ~5-7 Ma, coeval to a lowering in convergence rates [9]. Exhumation rates were later increased with the onset of Plio-Pleistocene glaciation [8, 9].

While it is established that the core of the peninsula has been steadily exhumed since at least 14 Ma, the timing of the initiation of uplift remains poorly understood. It is unknown when the accretionary wedge started forming a submarine topographic high, which is critical to understand sediment transfer from the arc to the trench and the distribution of deformation along the subduction margin [37]. It is also unclear when the Olympic Peninsula became subaerially exposed, confining the Puget Lowland and allowing fluvial erosion to significantly increase exhumation rates.

Brecciation and quartz veining in low-grade metamorphic rocks in the Olympic core dated at 17 Ma may be associated with early exhumation [38]. Using an assumption of a depth of accretion at 14.5 km, a constant exhumation rate of 0.75 km/Myr, and a geothermal gradient of 4.4°C/km, Brandon et al. [7] combined zircon and apatite fission-track ages to calculate that exhumation of the core of the Olympic Mountains should have begun at ca. 18 Ma. Uplift of the Olympic Peninsula resulted in the deformation and further partitioning of the Cascadia forearc basin into smaller subbasins, a process that had begun after collision with the Siletzia terrane [17]. This uplift could be reflected in the angular unconformity between 22 and ~13 Ma found in the Seattle Basin [39, 40], or the younger angular unconformity between 7.5 and 6 Ma in the Astoria Basin [41]. Bigelow [42] used sedimentary grain petrography to characterize the provenance of the upper Miocene Montesano Formation of Southwest Washington and Northwest Oregon to support the claim that the Olympic core and Crescent Formation had been uplifted and emerged by 10 Ma. There are no other, more precise constraints for the age of uplift and emergence of the Olympic Mountains, particularly in Northwest Washington near the highest peaks of the range.

2.2. Stratigraphy of the Cascadia Forearc. Before the accretion of Siletzia, Western Washington hosted the Swauk Basin, a large nonmarine sedimentary basin into which the several kilometer-thick Chuckanut Formation was deposited between 60 and 51 Ma (Figure 3; [17]). Well exposed along

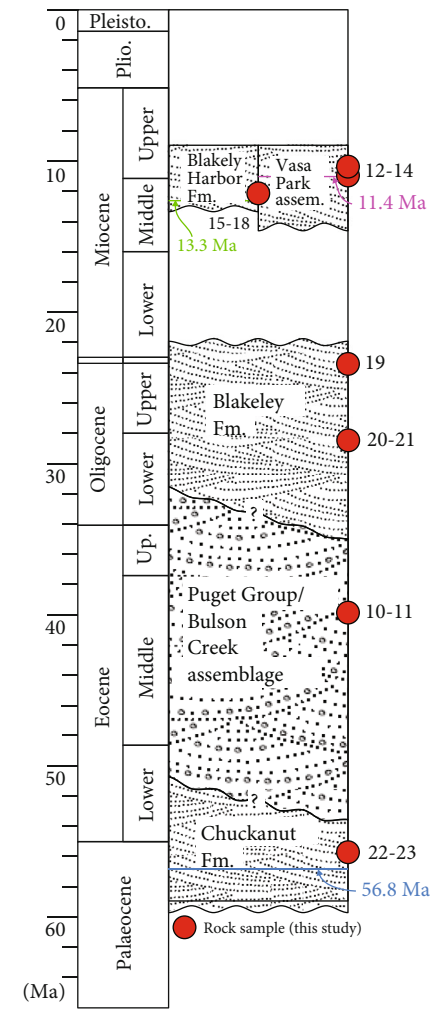

FIGURE 3: Composite stratigraphic column displaying age relationships of units from the Cascadia forearc sampled for this study. Stratigraphic locations of samples are shown and sample numbers are labeled (see Table 1 for sample numbers). Important age constraints are labeled (see main text for references).

the coast near Bellingham, Washington, the Chuckanut Formation consists of 6 members: the Bellingham Bay, Slide, Governer's Point, Padden, Warnick, and Maple Falls Members, in stratigraphic order [43]. The Chuckanut Formation consists of coarse sandstones and conglomerates, mudstones and siltstones, and abundant coal [43–45]. The sedimentary provenance of the Chuckanut Formation has been thoroughly studied and shows a composite history for the unit. The Bellingham Bay and Slide Members have been proposed to be sourced by sediment from the metamorphic core of Eastern Washington, transported by large, competent fluvial systems, which decreased in size through time [43]. The Padden Member contains abundant chert pebbles and is likely derived from the Western mélange belt (WMB)—a large belt of Jurassic to Cretaceous mélange that was accreted to North America in the Late Cretaceous and today is exposed in the foothills of the Cascades [46]. The Warnick and Maple Falls Members are more likely sourced from uplift at the northern portion of the Swauk Basin, and they

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comprise interfingered alluvial fan deposits [17, 44]. Strike-slip faulting in the Swauk Basin began at 51 Ma resulting from collision of Siletzia with North America [4, 17]. Subsequent faulting caused large-scale deformation of the Chuckanut Formation and partitioning of the Swauk Basin throughout the Eocene [43]. The faulting and deformation of the Chuckanut Formation have obscured the top of the section, and stratigraphic relationships with adjacent strata are difficult to discern.

Middle Eocene to early Miocene forearc deposits that postdate deposition of the Chuckanut Formation and accretion of Siletzia have been given various local names; they are grouped under the Puget Group within the Puget Lowland [47], and under the Northern and Southern Peripheral sequences north and south of the Olympic Peninsula, respectively [48]. The Puget Group commonly comprises lignite-bearing fluviodeltaic and shallow marine deposits, interfingered with volcaniclastic sandstones [49]. In the Lake McMurray area near Mt. Vernon, Washington, 1500 m of these deposits have been given the informal name of the Bulson Creek assemblage (Figure 3; [45]). The northern and southern Peripheral Sequences are dominated by deeper marine facies [48].

Further South in Puget Lowland, in the Seattle Basin, Oligocene to Miocene deposits are particularly thick (>7 km). Thickening in this region is explained by local flexural loading along the Seattle Fault Zone (SFZ; [50, 51]). The SFZ consists of multiple east-trending, north-verging thrust faults, which accommodate northward shortening inboard of the Olympic massif [52]; it merges further west with the northeast-striking oblique-slip fault system that accommodates trench-parallel shortening within the Olympic Peninsula [33, 34]. The timing of onset of SFZ activity and associated flexural loading is debated. Johnson et al. ([50], 1998) propose a late Eocene to Oligocene age for the initiation of the Seattle Fault; this interpretation is supported by a single apatite fission-track age of 32 ± 5 Ma obtained from the Green Mountain area in the Seattle Fault hanging wall [53]. By contrast, ten Brink et al. [51] date the onset of the Seattle Fault to the early to middle Miocene based on seismic reflection data.

The upper Oligocene-lower Miocene Blakeley Formation in the Seattle Basin is commonly distinguished from the Puget Group as it is dominated by deeper marine facies [54]. Deposited between 32 and 22 Ma, the Blakeley Formation is divided into two members: the lower Orchard Point and the upper Restoration Point Members [39], both representing submarine fan deposition [54]. The Orchard Point Member, which is well exposed along the Sinclair Inlet and along Alki Beach in Seattle, is a coarse clastic sandstone with local siltstones and fine sandstones, and one layer of tuffaceous clay-shale [39]. The Restoration Point Member is well exposed on the south of Bainbridge Island and is finer grained, with fine sandstones, abundant siltstones and shales, and rare pebbly sandstone layers [39]. The top of the Blakeley Formation is marked by an angular unconformity [55].

Sitting atop the unconformity that caps the Blakeley Formation is the Blakely Harbor Formation, which is well exposed along the Southern beaches of Bainbridge Island [39, 55]. While this contact is not expressed at the surface, it has been well documented with subsurface imaging [51]. A zircon fission-track age at 13.3 ± 1.3 Ma from a tephra layer near the base of the Blakely Harbor Formation and a middle Miocene pollen assemblage give a late middle Miocene age for the base of the unit [40]. The Blakely Harbor Formation is made of coarse fluvial and overbank deposits, composed of clastic, basalt-rich sandstones and conglomerates interbedded with claystone and siltstone, and is locally rich in organic matter [39]. The numerous basalt clasts found in the Blakely Harbor Formation have been attributed to erosion of the Crescent Formation as a sediment source [39].

Further east within the Seattle Basin, a sequence of sedimentary rocks, contemporary with the Blakely Harbor Formation, crops out in ravines within Vasa Park [56]. Some have considered these strata part of the Blakely Harbor Formation [40]; however, we refer to this section informally as the Vasa Park assemblage. The Vasa Park assemblage contains two lithofacies: pebbly conglomerate containing andesitic and felsic volcanic cobbles, and tuffaceous sandy siltstone and silty sandstone rich in organic matter [56]. K-Ar ages of tuffs constrain deposition of the assemblage to between 14.7 and 9.3 Ma, though the extremely low radiogenic argon content of these samples (8%) leaves high uncertainty [57]. A more recently acquired $^{40}Ar/^{39}Ar$ age dates the middle of the assemblage to 11.40 ± 0.61 Ma [58].

3. Methods

Sediment samples were collected from several modern rivers in the Washington Cascades and Olympic Peninsula as well as sedimentary rock samples from turbidites and fluvial sandstones in the Puget Lowland and on the Olympic Peninsula. Sample locations, determined with a handheld GPS, are shown in Figure 1 and labeled in Table 1. We collected samples of 1 to 5 kg of medium to coarse sand from sand bars of the following modern rivers: the Skykomish, Puyallup, Skagit, and Snoqualmie rivers, which drain the Cascades into Puget Sound; the Hoh, Elwha, Bogachiel, and Queets rivers, which drain the Olympic Peninsula into the Pacific Ocean and the Strait Juan de Fuca; and the Columbia River, which drains Eastern Washington State into the Pacific Ocean.

Sedimentary rock samples were collected from medium to coarse fluvial sandstones and turbidites, and samples were cleaned to avoid contamination from nearby quaternary alluvium. In the Olympic Peninsula, we collected two samples of the coastal OSC from Kalaloch Beach, mapped as lower to middle Miocene [25]. We collected one sandstone from the upper OSC at Shi Shi Beach, just South of a block of Eocene pillow basalt that correlates to the Crescent Formation, in turbidites attributed to the upper Eocene-Oligocene [26]. We collected a second sandstone of the upper OSC at Second Beach just south of La Push, Washington, from a local mélange attributed to diapirism and considered as being upper Eocene to lower Miocene in age [25].

In the Puget Lowland, we collected two samples in the Bellingham Bay Member of the Chuckanut Formation. We collected two samples of the Bulson Creek assemblage from the upper lithofacies described by Marcus [45]; one

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Table 1: Maximum depositional ages (MDA) calculated for every sample. *n*: number of zircon ages; MSWD: mean square weighted deviation.

Sample #	Sample name (and river)	MDA ± 2 s uncertainty	Sample location
Cascade rivers			
1	19-SNOR-01 (Snoqualmie)	18.5 ± 0.5 Ma (n = 19, MSWD = 0.8)	Three Forks Natural Area
2	18-CASCR-01 (Skykomish)	20.8 ± 0.7 Ma (n = 1)	Sultan, WA
3	18-CASCR-02 (Puyallup)	0.27 ± 0.21 Ma (n = 3, MSWD = 3.9)	Puyallup, WA
4	18-CASCR-03 (Skagit)	0.15 ± 0.04 Ma (n = 12, MSWD = 1.8)	Minkler, WA
Olympic rivers			
5	18-ELWA-01 (Elwha)	23.9 ± 1.6 Ma (n = 1)	Maddison Falls Trailhead
6	19-BOGR-01 (Bogachiel)	23.5 ± 1.8 Ma (n = 1)	Bogachiel State Park
7	19-QUER-01 (Queets)	20.3 ± 1.1 Ma (n = 1)	Near Clearwater Rd.
8	18-HOHR-01 (Hoh)	24.3 ± 2.4 Ma (n = 3, MSWD = 3.9)	Near Allens Bar Campground
Columbia River			
9	18-COLR-01 (Columbia)	0.072 ± 0.09 Ma (n = 4, MSWD = 2.8)	Cape Disappointment State Park
Bulson Creek assemblage			
10	18-BLC-01	48.2 ± 3.4 Ma (n = 3, MSWD = 1.7)	Lake McMurray, WA
11	18-BLC-02	29 ± 0.8 Ma (n = 41, MSWD = 1.2)	Lake McMurray, WA
Vasa Park assemblage			
12	18-RNT-01	12.3 ± 0.3 Ma (n = 40, MSWD = 1.3)	Lakemont Park
13	18-VAP-01	11.8 ± 0.3 Ma (n = 30, MSWD = 1.2)	Vasa Park
14	18-VAP-03	11.1 ± 0.3 Ma (n = 50, MSWD = 1.2)	Vasa Park
Blakely Harbor Formation			
15	17-BLH-02	29.8 ± 1.5 Ma (n = 4, MSWD = 1.5)	Bainbridge Island
16	18-BLH-01	22 ± 0.9 Ma (n = 1)	Bainbridge Island
17	18-BLH-02	28.8 ± 1.5 Ma (n = 1)	Bainbridge Island
18	18-BLK-02	30 ± 0.8 Ma (n = 32, MSWD = 1)	Bainbridge Island
Blakeley Formation			
19	18-BLK-01	26.6 ± 0.9 Ma (n = 3, MSWD = 0.1)	Bainbridge Island
20	19-BLK-01	26.1 ± 0.7 Ma (n = 4, MSWD = 0.7)	Alki Beach
21	19-BLK-02	22.8 ± 0.7 Ma (n = 4, MSWD = 0.7)	Alki Beach
Chuckanut Formation			
22	18-CHK-01	65.1 ± 2 Ma (n = 11, MSWD = 1)	Larrabee State Park
23	18-CHK-02	59.3 ± 2.4 Ma (n = 1)	Larrabee State Park
Upper OSC			
24	19-OSC-01	82.1 ± 2.9 Ma (n = 1)	Second Beach
25	19-OSC-02	41.3 ± 1.4 Ma (n = 1)	Shi Shi Beach
Coastal OSC			
26	18-HOH-01	16.7 ± 0.5 Ma (n = 6, MSWD = 1)	Kalaloch Beach
27	18-HOH-02	16.5 ± 0.5 Ma (n = 8, MSWD = 1)	Kalaloch Beach

sample of the Blakeley Formation from the Restoration Point Member on Bainbridge Island; two samples from the Orchard Point Member where it crops out on Alki Beach in Seattle [39, 55]; four samples from the Blakely Harbor Formation on Bainbridge Island near the 13.3 ± 1.3 Ma tephra layer reported by Sherrod [40]; and three samples from the Vasa Park assemblage close to or at the locality described by Dillhof et al. [56], where a $^{40}Ar/^{39}Ar$ age of 11.4 ± 0.6 Ma is reported [58].

In total, we collected samples from 9 modern rivers and 18 rock outcrops. All samples but two were analyzed for grain petrography. For each modern river sample, a small fraction of collected sediment was mounted in synthetic resin. For each hard-rock sample, a small (~ 1.5 × 3 cm) fragment was cut and prepared as a thin section. Petrographic results were acquired on these thin sections using the Gazzi-Dickinson method to discern the relative abundance of quartz, feldspar, and lithic grains in each sample [59]. All petrographic data and GPS locations of the samples are available in Supplementary Table 1.

All samples were analyzed for detrital zircon provenance; analytical set-up is presented by Licht et al. [60], and the complete procedure is detailed in Supplementary File 1. Zircons were extracted by traditional methods of heavy

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mineral separation, including concentration with a Holman-Wilfley™ gravity table, density separation with methylene iodide, and magnetic separation with a Frantz™ Magnetic Barrier Separator. U-Pb ages were generated using laser-ablation inductively coupled-plasma mass-spectrometry (LA-ICP-MS), using an iCAP-RQ Quadrupole ICP-MS coupled to an Analyte G2 excimer laser at the University of Washington, using a spot diameter of 25 microns and Plešovice zircons as calibration reference material [61]. Data reduction was conducted with *Iolite* (Version 3.5), using their *U_Pb_Geochron4* Data Reduction Scheme to calculate U-Pb dates uncorrected for common lead [62]. In addition, date uncertainties for all samples were calculated using a modified version of the method of Matthew and Guest [63], implemented in MATLAB, that takes into account the impact of ^{207}Pb beam intensity on date uncertainties [64]. The dates used for plotting and in the discussion are ^{206}Pb/^{238}U for dates < 1400 Ma and ^{207}Pb/^{206}Pb for dates > 1400 Ma. Dates > 300 Ma were screened for concordance using a discordance filter at >20% discordance (<80% concordance) and >5% reverse discordance (>105% concordance); we used the ^{206}Pb/^{238}U vs. ^{207}Pb/^{235}U ratio to calculate discordance for dates < 1300 Ma, and the ^{206}Pb/^{238}U vs. ^{207}Pb/^{206}Pb ratio for older dates. These parameters are detailed and justified in Supplementary File 1.

The ten zircon validation reference materials used during these sessions yielded offset around TIMS ages < 1% in most cases, <2% otherwise. In total, our new dataset includes 1478 zircon ages from modern rivers and 2713 zircon ages from sedimentary units; detailed data are available in Supplementary Table 2.

The maximum depositional age for detrital samples is the weighted average of the youngest zircon dates when the youngest three or more dates overlap [65], calculated with *TuffZirc* [66]; if there is no overlap between youngest zircons, we used the youngest zircon date as maximum depositional age. The final age uncertainty around maximum depositional ages is the quadratic sum of the uncertainty of *TuffZirc* age calculation or youngest zircon date and of the systematic uncertainty (~2.67% for the ^{238}U/^{206}Pb ratios). Age distributions are given in the form of kernel density estimate (KDE) diagrams and age histograms obtained with MATLAB.

4. Results

Figure 4 displays detrital zircon age distributions for each of the samples processed over the 0-300 Myr interval; complete age distributions over the 0-3000 Myr interval are displayed in Supplementary Figure 1. Few samples yield zircons older than Mesozoic in age; these older zircons are present in small proportions (commonly < 20%) and display the same age populations: 1.05-1.2 Ga, 1.3-1.4 Ga, 1.5-1.8 Ga, 1.9-2.1 Ga, 2.3-2.4 Ga, and 2.5-2.8 Ga. Older populations are significant (>5%) in the sands of the Elwha and Columbia rivers and in samples of the OSC (sample #27), the Chuckanut Formation (samples #22 and 23), and the Blakely Harbor Formation (Samples #15-18). Maximum depositional ages are displayed in Table 1.

4.1. Modern Rivers. Samples #1-4 from Cascade rivers are rich in quartz and volcanic lithic fragments, plotting in the "recycled orogen" domain on the QFL plot. Age distributions from modern Cascade rivers contain >97% zircons younger than 250 Ma, with >90% of the zircons younger than 110 Ma. Two main age populations can be distinguished, with their contribution varying between samples: 15-40 Ma and 85-100 Ma. Samples #3 (Puyallup River, draining Mount Rainier) and #4 (Skagit River, draining two active volcanoes, Mount Baker and Glacier Peak) yield recent (<5 Ma) zircons.

Samples #5-8 from Olympic rivers are rich in quartz and sedimentary lithic fragments, plotting in the "recycled orogen" domain on the QFL plot. Age distributions from modern Olympic rivers contain >85% zircons younger than 250 Ma, but the contribution of older Cretaceous and Jurassic ages (110-250 Ma) is higher (~23% of zircons), with two prominent age populations at 140-170 Ma (all samples) and 180-220 Ma (Hoh and Elwha rivers). All Olympic rivers display the 85-100 Ma age peak found in modern Cascade rivers; samples #5 and #7, from the Elwha and Queets rivers, display a well-defined population centered at 50 Ma. The 15-40 Ma population found in Cascade river sands is barely expressed in Olympic rivers (<4% of zircons). The youngest grains for all Olympic rivers span from 20 to 24 Ma (Table 1).

Sample #9 from the Columbia River is rich in quartz and volcanic lithic fragments, plotting in the "recycled orogen" domain on the QFL plot. Sample #9 yields ages spanning the last 180 Ma, with few (~5% of zircons) pre-Mesozoic ages, despite its wide drainage basin covering Precambrian strata near its sources. Two age populations are particularly well marked: 45-55 Ma and <5 Ma.

4.2. OSC. Samples #26 and 27 of the Miocene coastal OSC display petrographic assemblages similar to modern Olympic rivers, dominated by quartz and lithic fragments, plotting in the "recycled orogen" field. The two samples also display age distributions similar to those of samples from modern Olympic rivers. They contain a major population at 85-100 Ma and subordinate populations at 45-55 Ma and 140-170 Ma. Samples #26 and #27 both display a significant population of early Miocene ages with maximum depositional ages at 16.5 and 16.7 ± 0.5 Ma (2 s; Table 1).

Samples #24 and 25 from the upper OSC (late Eocene to Oligocene/early Miocene) display the same petrographic results as coastal OSC samples, but significantly different age distributions. Both age distributions are dominated by one single 170-230 Ma peak, with only few younger and older grains. The characteristic 0-110 Ma age range of the modern Cascade rivers is notably absent, with most of the zircons significantly older than the depositional ages of the samples.

4.3. Puget Lowland. Samples #22 and 23 from the Chuckanut Formation are distinct from other samples as they are particularly rich in quartz and metamorphic lithic fragments, with sample #22 plotting in the "continental block domain" on the QFL plot (Figure 5). Their age distributions are rich in pre-Mesozoic grains (30 to 40% of zircons); younger grains cover the 55-200 Ma range, with a significant Late Cretaceous-Paleocene population centered around 70 Ma (Figure 4).

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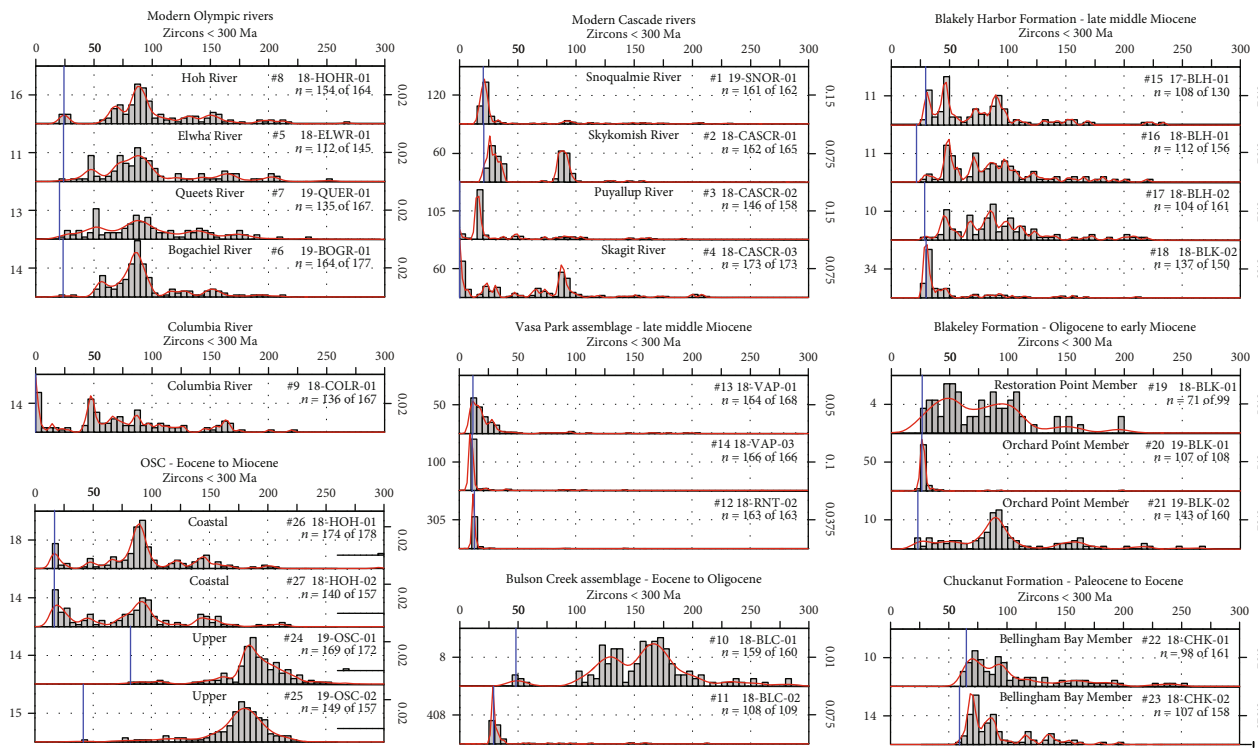

FIGURE 4: Histograms and kernel density estimate diagrams for detrital zircon age distributions of all samples, showing only zircon ages < 300 Ma (for full age distributions see Supplementary Figure 1). River samples are grouped by drainage region, and rock samples are grouped by formation. All ages are displayed in 5 Myr bins. Histogram counts are labeled on the left axis. Probability is labeled on the right axis. Blue line = maximum depositional age.

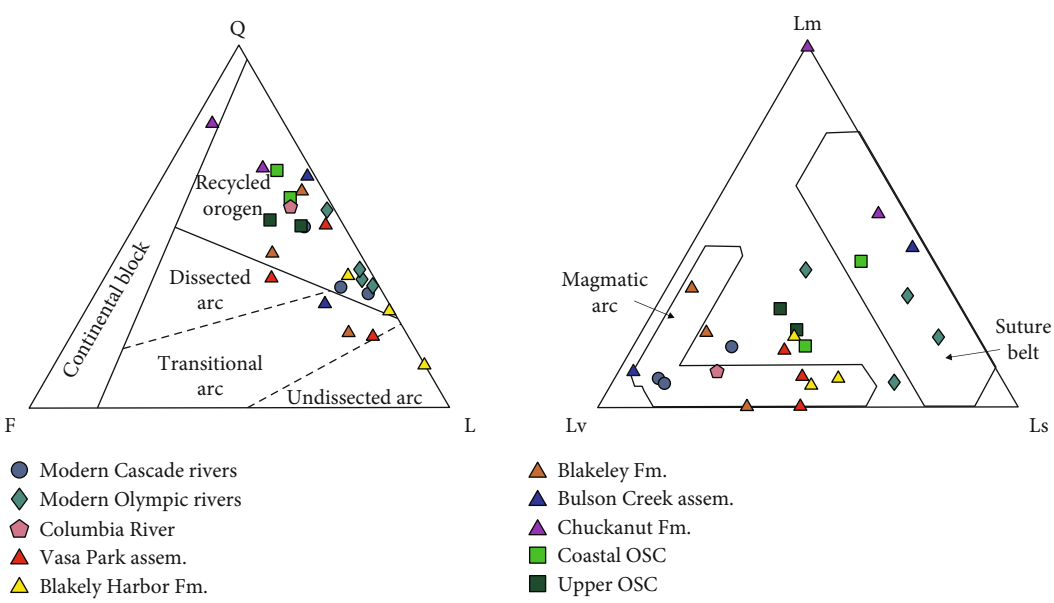

FIGURE 5: Ternary diagrams of sediment modal composition for modern rivers and rock samples from the Puget Lowland and the Olympic Peninsula according to the Gazzi-Dickinson method [59, 95].

Samples #10 and 11 from the Bulson Creek assemblage are notably different from each other. Sample #10 is rich in quartz, metamorphic and sedimentary lithic clasts; it displays one population of younger grains clustered at 50 Ma, with most ages ranging between 115 and 280 Ma (>90% of zircons). Sample #11 consists entirely of volcaniclastic lithic grains and rare quartz grains; its age distribution shows a single prominent population between 28 and 35 Ma, with rare older accessory grains. Sample #11 yields a 29.0 ± 0.8 Ma maximum depositional age (2 s; Table 1).

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Samples of the Blakeley Formation can be divided into two groups. Samples #19 and #21 are rich in quartz and volcanic lithic fragments, and plot in the "recycled orogen" domain; most zircons are in the 20-110 Ma age range with an important Cretaceous population, and a secondary population around 50 Ma for sample #19. Sample #20 is rich in volcanic lithic grains, with rare quartz grains; its age distribution show a characteristic zircon population between 20 and 35 Ma, with a 26.1 ± 0.7 Ma maximum depositional age (2 s; Table 1).

Samples #15-17 of the late middle Miocene Blakely Harbor Formation display similar age distributions to one another, and mirror age distributions of modern Olympic rivers; ages cover the full 20-230 Ma range with a major population around 85-100 Ma and a subordinate population at 45-55 Ma. Sample #18 spans a similar age range, though it has a primary age peak at ~30 Ma and contains only rare accessory grains between ~120 and 230 Ma. In contrast to the Olympic river sands, samples #15-17 are mostly made of basaltic lithic clasts, a feature recognized by Fulmer [39]. Maximum depositional ages are all similar to those of the Blakeley Formation samples (22-30 Ma; Table 1) and much older than the depositional age of the Blakely Harbor Formation (13.3 ± 1.3 Ma; [40]).

Samples #12-14 of the Vasa Park assemblage show distinctive characteristics in their petrography and age distributions. All three samples contain primarily quartz grains and lithic clasts with roughly the same amount of volcanic and sedimentary lithic fragments. Samples #12 and #14 show a single, late middle Miocene population and maximum depositional ages at 12.3 ± 0.3 and 11.1 ± 0.3 Ma (Table 1); this second age is statistically similar to the 11.4 ± 0.6 Ma $^{40}Ar/^{39}Ar$ age that is reported at the same locality [56, 58]. Sample #13 shows a similar maximum depositional age at 11.8 ± 0.3 Ma, but with an age range spanning 10-50 Ma (Table 1; Figure 4).

5. Interpretation

5.1. Provenance of Modern Cascade River Sands. The age distributions of samples #1-4 from modern Cascade river sands are a combination of young zircon populations derived from the Cascade Arc (ages < 45 Ma) and minor, older populations derived from the Coast Mountains Batholith (clusters at 65-75 Ma, 85-105 Ma, 140-165 Ma, and 200-210 Ma; Figure 6). Below, we use these age distributions as a signature for zircons derived from the Cascades; sedimentary rocks that were sourced from the Cascades should contain similar zircon populations and are expected to incorporate young volcanic zircons from the Cascade Arc that are coeval or only slightly delayed in age with the deposition of the unit.

5.2. Provenance of the OSC. Our study highlights two distinct provenances for the OSC. Samples #24 and 25 of the upper OSC (early Eocene to early Miocene in age) are dominated by a single Late Triassic to Early Jurassic population (170-220 Ma) that is absent in every other sample, including modern Cascade rivers (Figure 4). There are Late Triassic-Early Jurassic plutons in the crystalline core of the Cascades [67, 68], but they barely contribute to the zircon load of modern Cascade rivers, as highlighted here. Thermochronological data show that the crystalline core of the Cascades remained mostly unexhumed until the Miocene: rocks from the west flank of the Cascades yield apatite-He cooling ages between 12 and 6 Ma; rocks from the east flank show a broader range of apatite-He cooling ages between 60 and 18 Ma, though they also display far slower exhumation rates throughout the Cenozoic [69]. Mesozoic strata from the Methow Valley, east of the Cascades, contain abundant Late Jurassic and Early Cretaceous zircon populations, but zircons older than 200 Ma are either entirely absent, as seen in the Boston Bar and Twisp Formations [70], or they are extremely rare (>2%), as in the Winthrop, Harts Pass, and Midnight Peak Formations and the Jackass Mountain Group [71].

Another potential source for the 170-220 Ma age peak found in samples #24 and 25 is the Bonanza Arc of the Wrangellia Terrane, which is exposed on Southern and Western Vancouver Island [72]. Rocks of the Bonanza Arc yield U-Pb ages ranging from 168 to 202 Ma; associated volcaniclastic strata indicate that magmatic activity had started earlier in the Triassic, potentially as early as 230 Ma [73]. A similar 170-220 Ma zircon population is found in the age distribution of samples from Cretaceous forearc strata in the Georgia Basin, east of Vancouver Island [74]. This population is particularly prominent in samples from the Comox Formation of the lower Nanaimo Group in the southern part of the Georgia Basin, and is interpreted by Huang et al. [74] as reflecting sediment sourced from the Bonanza Arc (Figure 7). Other age populations found in samples of the Comox Formation include age clusters at 344-364 Ma and 450-455 Ma, and ages younger than 170 Ma; these populations are interpreted as reflecting sediment drained from the Coast Mountains and the metamorphic and igneous basement of Vancouver Island [74]. However, these ages are absent from our samples of the upper OSC. Rocks of the Bonanza Arc crop out on extensively along the southern and western slope of Vancouver Island; this makes it possible for sediment from the Bonanza Arc to drain into the Pacific Ocean without incorporating zircons from other basement rocks, from forearc strata, or from the Coast Mountains [75]. We interpret the 170-220 Ma age population in our samples of the upper OSC as also reflecting sediment drained form the Bonanza Arc on Vancouver Island. Major exhumation of the Wrangellia Terrane is recorded between 50 and 40 Ma, which is contemporary with deposition of the upper OSC and shows that the Bonanza Arc is a viable source of sediment to the upper OSC [76]. We suggest that our samples originated as turbiditic events generated along the Vancouver Island slope during this period of exhumation.

The Late Triassic-Early Jurassic population identified in upper OSC samples is barely expressed in samples of the coastal OSC and modern Olympic rivers. The composite age distribution of the (Miocene) samples from the coastal OSC contains the same zircon populations that are found in the composite age distribution of modern Cascade rivers, aside from the complete absence of middle Miocene and younger zircons in the coastal OSC (Figure 6). These features indicate that the coastal OSC was likely derived from Cascade

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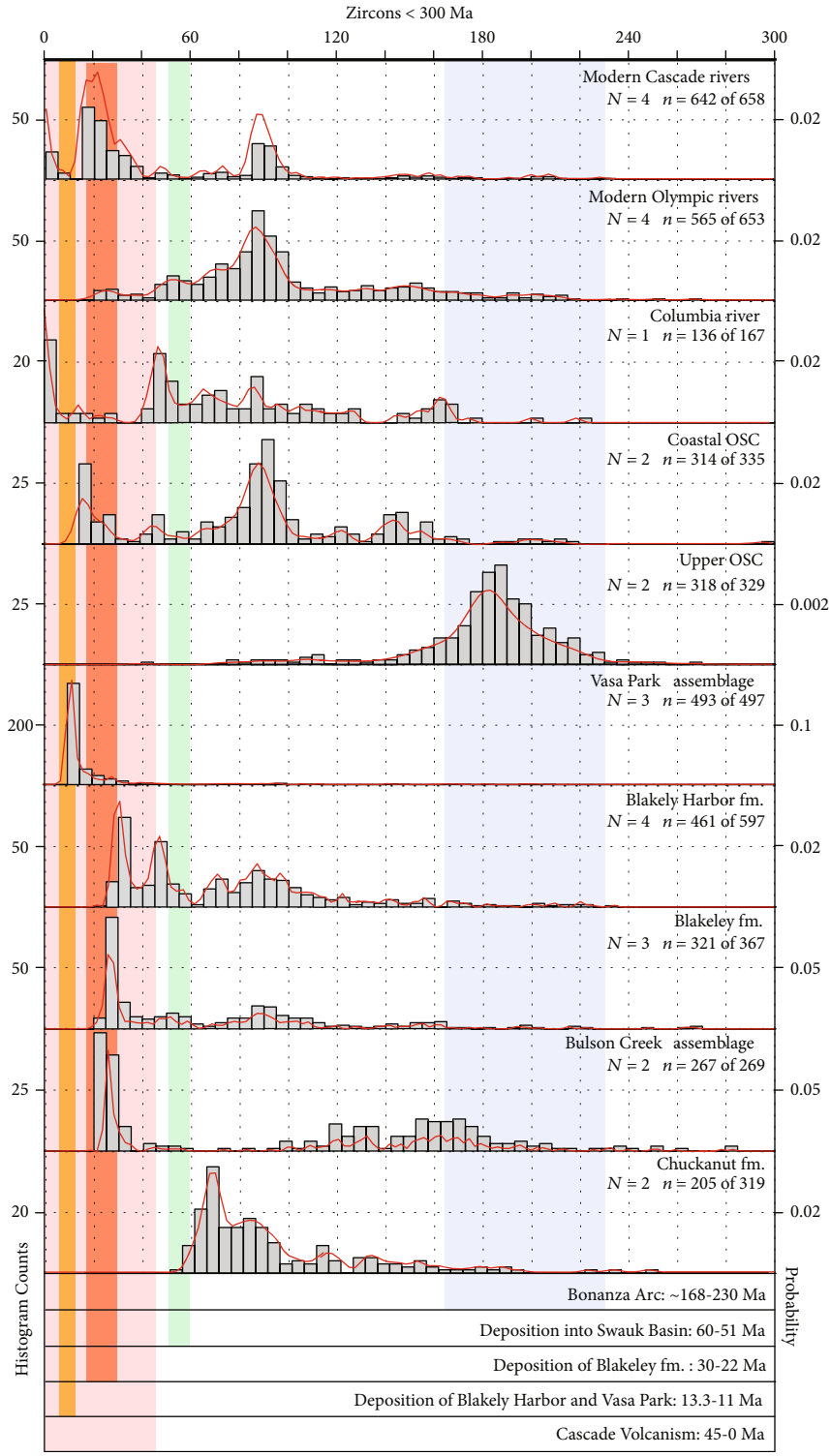

Figure 6: Histograms and kernel density estimate diagrams for sample groups showing only zircon ages < 300 Ma. Samples are grouped by unit/drainage region. Colored bars indicate important depositional ages and age peaks particular to individual source regions. All histograms are in 5 Myr bins.

material. The composite age distribution of modern Olympic rivers is also similar to the composite output of Cascade rivers (Figure 6), with some differences. Olympic rivers contain a relatively low proportion of 0-50 Ma grains (~6.5% of zircons) and display a 110-200 Ma population, which is underrepresented in modern Cascade rivers. The four modern Olympic rivers we sampled all drain the core of the Olympic Peninsula, which is dominated by the oldest (Eocene) OSC rocks; this drainage is consistent with an underrepresentation of 0-50 Ma grains and a higher

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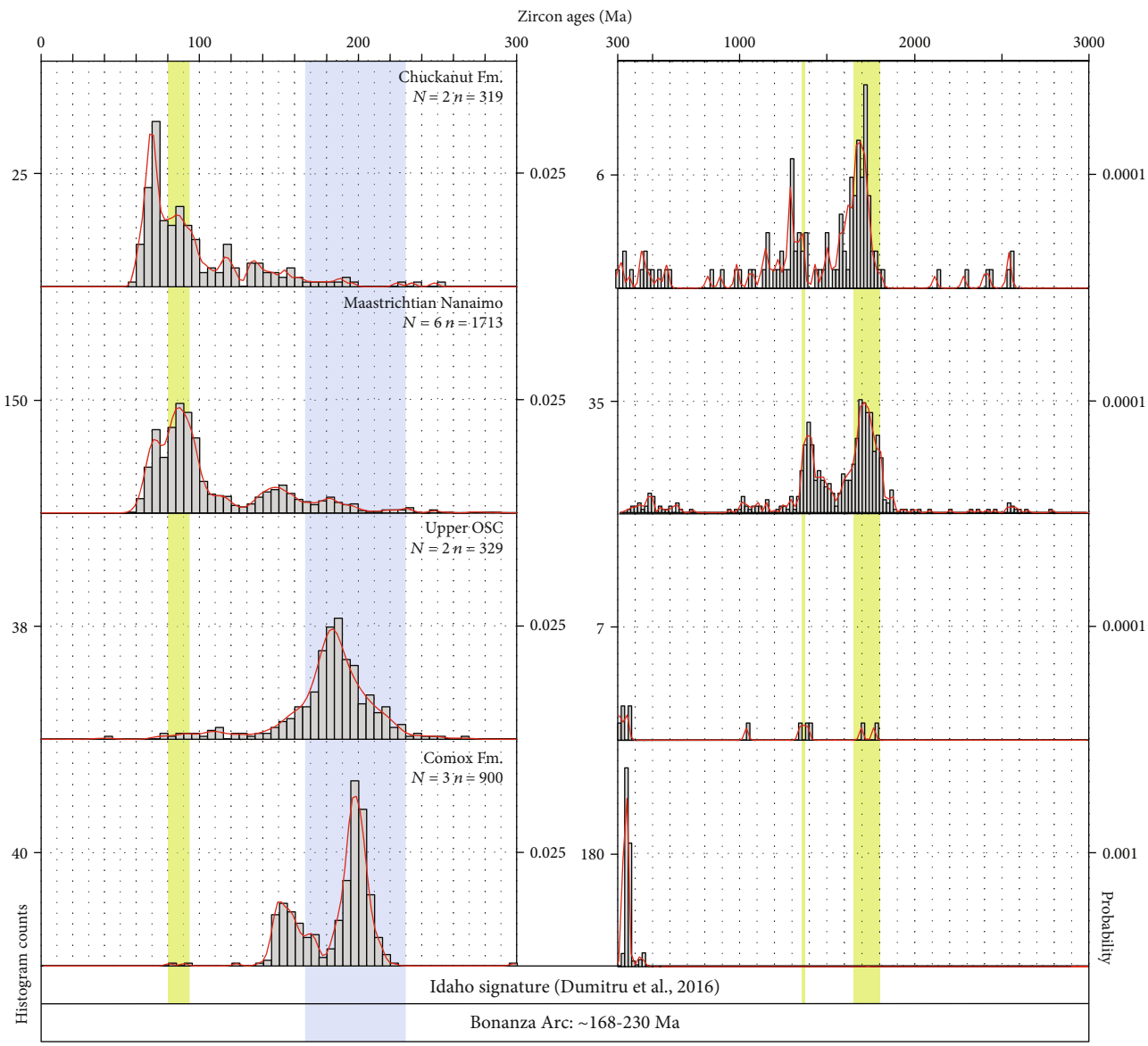

Figure 7: Histograms and kernel density estimate diagrams for the upper OSC, Chuckanut Formation, Comox Formation (samples S19, S20, and S21 from [74]), and Maastrichtian strata of the Nanaimo Group (samples GF1, GF2, GF3, SP, GB3, and GB4 from [78]). Age distributions for each sample are split into ages < 300 Ma with 5 Myr histogram bins, and ages 300-3000 Ma with 20 Myr histogram bins.

representation of older populations. We suggest that despite these differences, the majority of the sedimentary rocks in the core of the Olympic Peninsula were sourced by drainage of the Cascades.

The presence of zircons as young as 20 Ma in Olympic rivers draining the core of the peninsula indicates that the Olympics remained topographically low prior to the early Miocene. The maximum depositional ages of 16.5 and 16.7 ± 0.5 Ma in samples #26 and #27 of the coastal OSC indicate that the OSC was fed by Cascade-derived sediment until at least that time.

5.3. Provenance of the Puget Lowland. Samples #22 and 23 of the Chuckanut Formation display a significant population of pre-Mesozoic grains and, in addition to the well-marked 90 Ma population found in all forearc samples, a distinct 75 Ma population, which is not well expressed in the modern Cascade rivers. The age distributions of these samples are remarkably similar to those of Maastrichtian strata of the upper Nanaimo Group in the Georgia Basin, which also display two Cretaceous age peaks centered at 75 and 90 Ma, and older age populations centered at 1380 Ma and 1650-1800 Ma (Figure 7; [77, 78]). The age clusters at 90 Ma, 1380 Ma, and 1650-1800 Ma found in upper Nanaimo samples have been interpreted as derived from Idaho [79] and the younger ages from the Coast Mountains Batholith. Alternatively, Mathews et al. [63] showed that these ages are also found in the Mojave-Sonoran Region (MSR) of the Southwestern United States, advocating for a much lower paleolatitude for Vancouver Island at the time of deposition of the Nanaimo Group followed by a large coastwise translation of the Wrangellia Terrane during the Cretaceous [80].

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The paleolatitude of the Wrangellia Terrane and the distinction between the two possible sediment source regions for the Nanaimo Group are both topics of ongoing debate. The age distributions of our samples suggest that either the Chuckanut Formation and the Maastrichtian strata of the Nanaimo Group share the same sediment source, or that the Chuckanut Formation was sourced from reworking of the Nanaimo Group. A direct provenance from the MSR is unlikely for the Chuckanut Formation as the proposed coastwise translation of accreted terranes would have been achieved several millions of years prior to deposition of the Chuckanut Formation in the early Eocene [80]. At present, our provenance data do not allow us to distinguish between a sediment source for the Chuckanut Formation either in northern Idaho or from reworking of the Nanaimo group. A potential reworking of the Nanaimo group at that time requires further validation as thermochronological data do not show major exhumation of the Nanaimo Group until 50 Ma [76].

The source of the Bulson Creek assemblage is also problematic. The volcaniclastic nature and age distribution of sample #11 indicates a direct sourcing from the Cascade Arc. By contrast, sample #10 displays a dominant, broad age population ranging between 115 and 280 Ma, which is barely expressed in any other sample. This broad population overlaps with the ages of magmatism found in the Eastern Magmatic Belt of the Coast Mountains Batholith in British Columbia, though it is unclear how sediment derived from this region could have been transported far enough southwest to become incorporated into the Bulson Creek assemblage [81]. The Western mélange melt (WMB) crops out in close vicinity to our sampling site; age distributions of samples from the WMB are heterogeneous, with some samples displaying a well-marked 100-200 Ma age population while others do not [82–84]. It is possible that sample #10 is locally derived from the reworking of WMB material.

Samples #19 and 21 of the Oligocene to early Miocene Blakeley Formation are volcaniclastic or quartz-rich, and the composite age distribution of the Blakeley Formation shows prominent age peaks at 30 Ma and 90 Ma, and a small population at 140-160 Ma (Figures 5 and 6). The same peaks are found in the same proportions in the composite age distribution of modern Cascade rivers (Figure 6). The provenance and petrography data of the Blakeley Formation are consistent with drainage of the Cascades as the sediment source.

Both the Vasa Park assemblage and the Blakely Harbor Formation are fluvial, lithic sedimentary units that were deposited into the Cascadia Forearc in the late middle Miocene, with the Vasa Park assemblage deposited in the east of the forearc basin and the Blakely Harbor Formation deposited further west (Figures 1 and 3). Samples #12-14 of the Vasa Park assemblage are entirely volcaniclastic and likely have a source in the Cascade Arc. By contrast, samples #15-18 of the Blakely Harbor Formation show significant (7-22 million years) offset between their maximum depositional ages and their true depositional ages (Figure 8). This offset indicates that the Cascade arc did not supply sediment to the Blakely Harbor Formation. The maximum depositional ages of samples #15-18 (22-30 Ma; Table 1) are the same as those of samples #19-21 of the Blakeley Formation (22.8-30 Ma; Table 1). The composite age distribution of the Blakely Harbor Formation very closely resembles the composite distributions of the modern Olympic rivers and the Blakeley Formation, with the exception of a prominent age peak at 50 Ma that is only well expressed in the Blakely Harbor Formation and the Columbia River (Figure 6). This 50 Ma peak is consistent with the age of the Crescent Formation, and the Blakely Harbor Formation contains abundant basaltic lithic clasts, which have been attributed to the incorporation of felsic and basaltic material from the Crescent Formation [39, 54]. These provenance and petrographic data show that the Blakely Harbor Formation was sourced by sediment that was eroded from the Crescent Formation and from older sedimentary units including the Blakeley Formation. There are two potential source areas where the Crescent Formation, the Blakeley Formation, and associated units of the Puget Group are exposed in close vicinity:

(1) The eastern flank of the Olympic Peninsula, approximately 30 km away from our sample location, where these units are exposed over thousands of square kilometers

(2) The hanging wall of the SFZ in the Green Mountain area, approximately 20 km away from our sample location, where exposure is much more limited (<100 km^2; [15, 53])

Everywhere else in the Puget Lowland, the hanging wall of the SFZ is covered by Puget Group sediment and does not expose the Crescent Formation [50, 51, 85–87]

Our petrographic and detrital zircon data are insufficient to distinguish between these two source areas as they display the same geological units. However, we suggest that a single contribution from the SFZ hanging wall is unlikely. Both the Blakely Harbor Formation and the Vasa Park assemblage are in the footwall of the SFZ [51, 86]; our samples of the Vasa Park assemblage do not show evidence for reworking of Puget Group material, which is immediately adjacent in the SFZ hanging wall.

The only thermochronological data available for the SFZ is an apatite fission-track age from the Green Mountain area dated at 32 ± 5 Ma [53]. This age dates the onset of exhumation of the SFZ hanging wall 10 to 25 Myr earlier than deposition of the Blakely Harbor Formation. Assuming a geothermal gradient of 20°C/km in the forearc [88, 89] and a functional closure temperature of 120°C for the apatite fission-track system [90], this fission-track age limits average exhumation to ~0.2 km/Myr since the Oligocene. Exhumation was to 4-10 times higher in the Olympic Peninsula during the late middle Miocene, with rates estimated between 0.75 and 2 km/Myr [7, 9]. The Olympic Peninsula is a more likely source as it was actively denuded during the period of deposition of the Blakely Harbor Formation. Though further thermochronological work on the SFZ is needed to confirm this interpretation, the plausible presence of material derived from the Olympic Peninsula in the Blakely Harbor Formation suggests that the peninsula had already been subaerially exposed and sufficiently uplifted to form a local topographic

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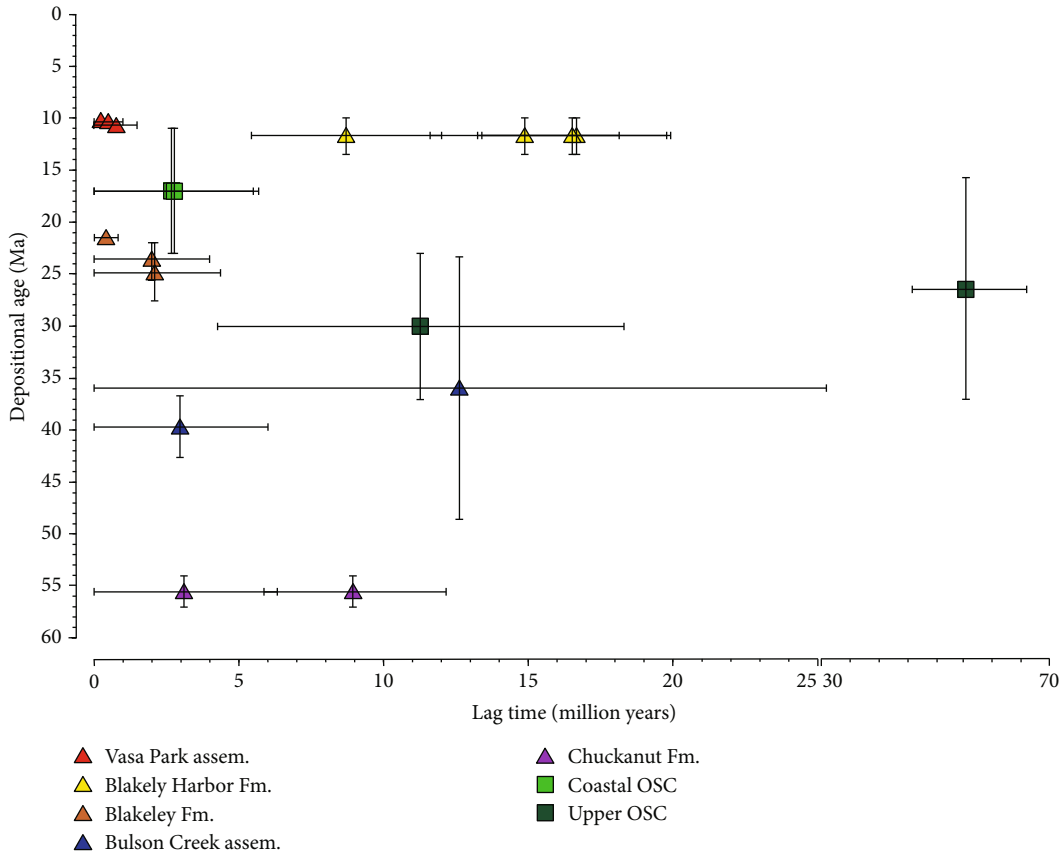

Figure 8: Lag time between the youngest zircon population and depositional age plotted versus depositional age for every sample.

high and supply sediment to eastward draining rivers. The tuff dated at 13.3 ± 1.3 Ma at the base of the unit [40] gives a minimum age for the emergence of the Olympic Peninsula (Figure 9).

All the samples interpreted here as flowing directly from the Cascades display a short lag time between their depositional age and the age of their youngest zircon population, commonly <5 Myr (Figure 8). Samples interpreted as coming from Vancouver Island (upper OSC) or mostly made from the reworking of older material (Chuckanut Formation, Bulson Creek assemblage, and Blakely Harbor Formation) display much longer lag times (9 to 60 Myr), which corroborates their provenance from areas without coeval volcanism.

6. Discussion

Our data confirm that the upper OSC has a different origin from other structural units of the Olympic Peninsula, which has important implications for the build-up of the Cascadia accretionary wedge. Age distributions and sedimentary grain petrography of the Miocene coastal OSC and modern Olympic rivers indicate that they are derived from the Cascades; this is consistent with deposition and incorporation in a typical accretionary wedge, with Cascade material transported through the forearc and into the subduction trench [27]. By contrast, our samples of the Eocene upper OSC are derived from Vancouver Island and do not include Cascade-derived zircons; this suggests that parts of the upper OSC were deposited offshore at a significant distance from the Cascadia trench. These data support the interpretation of Brandon et al. [7] who proposed that the upper OSC was deposited as the western continuation of the Siletzia terrane, and was later imbricated and underthrust beneath the eastern strata of Siletzia following the initiation of the Cascadia subduction zone (Figure 9: middle Eocene to middle Miocene). Despite the Cretaceous and middle Eocene maximum depositional ages of our two samples (Table 1), the upper OSC has yielded biostratigraphic ages as young as the lower Miocene, and our two samples are mapped as post-middle Eocene [25, 26]. It is very likely that our samples represent older strata of the upper OSC that have been reprised into a mélange and are part of broken formations, integrating younger foraminifera, as seen in many places in the upper OSC [7, 15].

There is quasicontinuous volcanism within the Cascade Arc since 45 Ma, despite periods of lower flux and varying composition [14]; it is expected that we would observe the presence of continually younger zircons within the OSC until uplift of the Olympic Peninsula precluded Cascade material from entering the accretionary wedge. As such, the presence of the youngest zircon population within the coastal OSC at 16.5 ± 0.5 Ma provides the maximum age at which the Peninsula was not uplifted and had not yet formed a topographic barrier. There are two mechanisms that could bring zircons of postuplift age into the Olympic accretionary wedge. First, younger zircons brought to the sea by the Columbia River, or an older analogue, south of the peninsula

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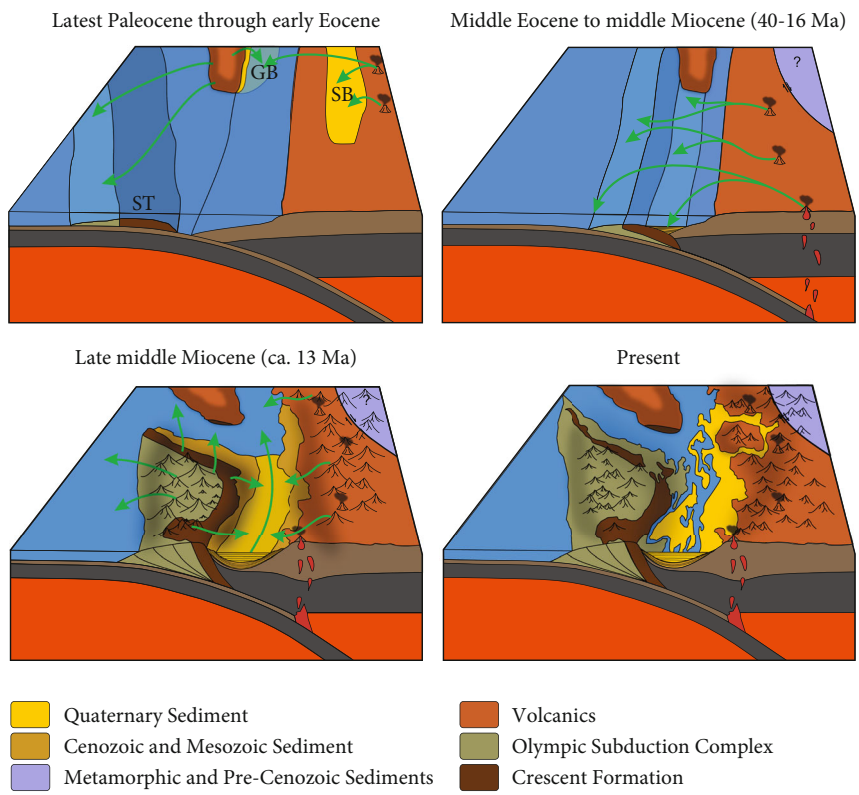

Figure 9: Synthesis of generalized block diagrams showing changes in geology and proposed sediment drainage patterns in Western Washington State during the Cenozoic. ST = Siletzia terrane; GB = Georgia Basin; SB = Swauk Basin.

could have been transported 150 km northward along the Olympic Coast by longshore drift to eventually arrive at the trench in front of the OSC, at the northernmost end of the modern deep-sea Astoria Fan. Today, Columbia River sediment is transported up to 100 km northwards into the Quinault Canyon along the southern part of the Olympic Coast [91]; this has also been observed for Mount Saint Helens ash material [92]. However, most of the modern sediment reaching the Quinault Canyon is very fine grained (silty clay; [93]), well below the grain size of the zircons we analyzed (25-micron minimum laser beam diameter). An alternative way that postuplift zircons could have been transported into the Cascadia trench and subsequently into the accretionary wedge is conveyance via a proto-Strait of Juan de Fuca around the north of the peninsula followed by a 100 km southward transit either by longshore drift or, less likely, deeper trench-parallel turbidity currents. Today, there is almost no southward longshore drift along the Washington coast due to the dominance of southwesterly winds in the winter [92]. Additionally, the age distribution of the coastal OSC is not consistent with either sediment supply from the Columbia River or from Vancouver Island, so it is unlikely that these mechanisms supplied fresh zircons to the OSC after uplift of the Olympic Peninsula cut off drainage from the Cascade Arc (Figure 9).

The late middle Miocene Blakely Harbor Formation in the Washington forearc displays age distributions and grain petrography that are consistent with a direct supply from the Olympic Peninsula. Our data cannot exclude that this formation was sourced by material from the hanging wall of the Seattle Fault, as proposed by ten Brink et al. [51], but we suggest that this source is less likely due to the low exhumation rates of the SFZ at that time and the absence of hanging wall reworking in the Vasa Park assemblage. While thrust loading along the SFZ likely contributed to the subsidence of the Seattle Basin and the deposition of the Blakely Harbor Formation in the late middle Miocene [50, 51], a contribution of the SFZ hanging wall to the sedimentary supply remains to be shown. We show that the Olympic Peninsula very likely contributed sediment to the Blakely Harbor Formation, making the age of the base of the Blakely Harbor Formation a maximum constraint on the emergence of the peninsula and formation of a topographic high.

Our sedimentary provenance data bracket the timing of the emergence of the Olympic Peninsula to a narrow window in the late middle Miocene, after 16.5 ± 0.5 Ma and likely before 13.3 ± 1.3 Ma (2 s), though this latter age remains to be confirmed by a more thorough study of the exhumation of the SFZ. Brandon et al. [7] estimated that sediment accreted to the base of the accretionary wedge would take ~4.4 Myr to rise from the depth of accretion to the α-damaged zircon closure depth of 10 km, assuming a constant exhumation rate of 0.75 km/Myr as observed for the last 14 Ma, and they show that this 10 km depth was reached ca. 13.7 Ma. These estimates date the onset of exhumation to ca. 18.1 Ma, at least two million years before our proposed window for the emergence of the peninsula. Michel et al. [9] proposed a faster rate (2 km/Myr) for the initial phase

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of exhumation, decreasing to bellow modern rates at 5-7 Ma. Using this faster initial exhumation rate, and the same model as Brandon et al. [7], we estimate exhumation to begin at ca. 15.3 Ma, which falls within our proposed time window for the emergence of the Olympic Peninsula. These estimates are based on a simple model for the trajectory and temperature history of material accreted to the wedge and should be assessed with caution. In the framework of this simple model, and assuming that there was little preemergence erosion and exhumation in the accretionary wedge, our timing for the emergence of the peninsula is in agreement with models arguing for an early phase of high exhumation [9] and a composite deformation history for the accretionary wedge.

Our work constrains the initial subaerial exposure of the Olympic Peninsula and the formation of a topographic barrier to the late middle Miocene, but we do not exclude the possibility of earlier deformation and uplift of the accretionary wedge. Deformation predating the emergence of the peninsula may have occurred in the Oligocene to early Miocene, and this deformation may be recorded in the strata of the Blakeley Formation. The Blakeley Formation is dominated by submarine fan deposits, some of which have yielded foraminifera found at bathyal to abyssal depths (deeper than 1 km; [39]). The presence of such depths in a forearc basin requires a mechanism to create high subsidence. Johnson et al. [50] linked this subsidence to an early phase of thrusting along the SFZ, the timing of which is supported by apatite fission-track data [53]. More thermochronological work along the SFZ is needed to confirm this early phase of deformation; if confirmed, this Oligocene episode of thrusting on the SFZ provides evidence for margin-parallel shortening and initiation of the Olympic orocline that predates emergence of the Olympic Peninsula by at least ten million years.

7. Conclusion

Our combined detrital zircon provenance and grain petrography data from the Olympic Peninsula and the Puget Lowland allow us to reconstruct the Cenozoic history of the Cascadia subduction zone. We first show a distinct sedimentary provenance of the upper OSC that differs from other structural units of the Olympic Peninsula. Upper OSC samples comprise only material from Vancouver Island and show no mixing with young (post-40 Ma) zircons. These results indicate that the upper OSC was deposited as part of the Siletzia terrane before its accretion to the Cascadia margin, and was later imbricated and underthrust beneath the Crescent Formation. Other units of the OSC represent accreted Cascadia trench sediment and show input from the Cascades into the accretionary wedge until at least 16.5 $\pm$ 0.5 Ma, providing an older age limit for the emergence of the Olympic Peninsula. The fluvial deposits of the late middle Miocene Blakely Harbor Formation record the first input of recycled Crescent Formation and OSC/forearc material into the Seattle Basin. Though we cannot exclude some contribution from the hanging wall of the Seattle Fault Zone, we suggest that the Blakely Harbor Formation provides a minimum age for input of eroded material from the Olympic Peninsula into the Puget Lowland; this gives a younger age limit of 13.3 $\pm$ 1.3 Ma for the emergence of the peninsula. We finally propose that the initial onset of deformation and oroclinal bending in the Olympic Peninsula predates emergence by at least 10 Myr, marked by flexural subsidence along the Seattle Fault Zone and the deposition of the Blakeley Formation in the Oligocene.

Data Availability

All data used in support of our conclusions are made available in the body of this paper and in the supplementary materials provided with it.

Conflicts of Interest

The authors declare that there is no conflict of interest regarding the publication of this article.

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Supplementary Materials

Supplementary File 1: complete analytical procedure for detrital zircon provenance analysis. Supplementary Table 1: grain-counting data. Supplementary Table 2: U-Pb data. Supplementary Figure 1: histograms and kernel density estimate diagrams for detrital zircon age distributions of all samples. River samples are grouped by drainage region, and rock samples are grouped by formation. All ages are displayed in 20 Myr bins. *(Supplementary Materials)*

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