Northwest Geological Society

Society Publications in Pacific Northwest Geology

FIELD TRIP GUIDEBOOK #048

OLYMPIC MOUNTAINS GEOLOGIC FIELD TRIP

October 3-4, 2015

Jim Aldrich, Los Alamos National Laboratory (Retired)

NWGS FIELD TRIP GUIDEBOOK SERIES

Of particular note, some stops on these trips may be located on private property. ***Publication of this guide does not imply that public access has been granted to private property.*** If there is a possibility that the site might be on private property, you should assume that this is the case. ***Always ask permission before entering private property.***

NORTHWEST GEOLOGICAL SOCIETY FIELD GUIDEBOOK SERIES
Field Trip Guidebook #048

Published by the Northwest Geological Society, Seattle, WA.
Publication Date: © 2015

***Cover photo**: Deformed Eocene turbidites (broken formation) of the Upper Olympic Structural Complex on Hurricane Ridge Road. J. Aldrich photo.*

TABLE OF CONTENTS

OLYMPIC MOUNTAINS GEOLOGIC FIELD TRIP

Jim Aldrich, Los Alamos National Laboratory (Retired)

I. INTRODUCTION

The Olympic Mountains of northwestern Washington are the result of the Juan de Fuca plate descending beneath the North American plate at the Cascadia subduction zone, accreting oceanic crust onto the Cascadia subduction wedge (**Figure 1**). The accreted oceanic sediments, structurally overlain on the east by Coast Range terrane, are largely below sea level but are exposed on land in the Olympic Mountains (**Figure 2**). The Cascadia forearc, the area between the volcanic arc and subduction zone, has a topographic low (from north to south the Georgia Straits, Puget Sound, and Willamette Valley) and high (Insular Range of Vancouver Island and Coast Range of Washington and Oregon) (**Figure 3**). The forearc high is the crest of the accretionary wedge.

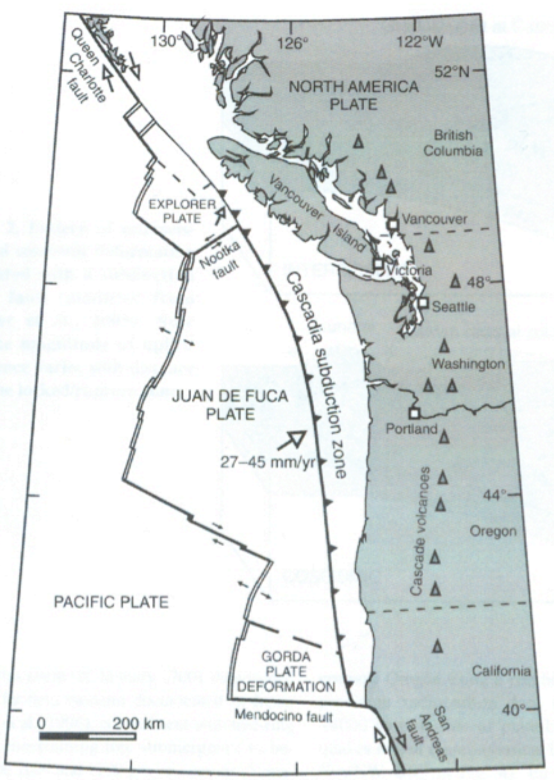

Figure 1. Plate tectonic setting of the Cascadia subduction zone. (From Leonard et al., 2010, Fig. 1)

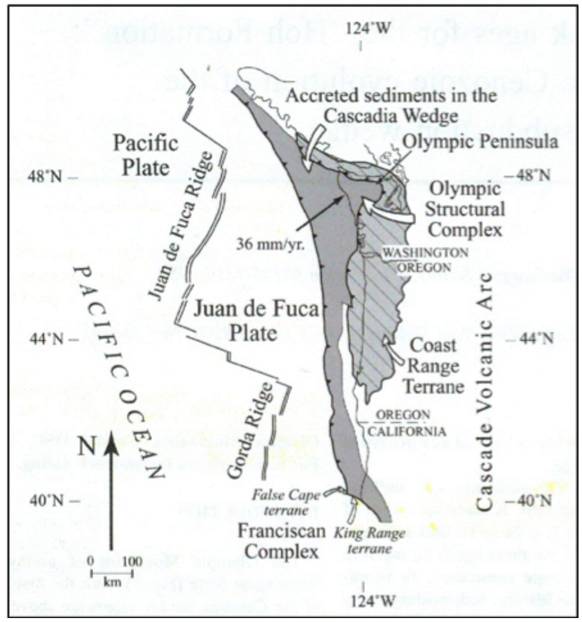

Figure 2. Map showing accreted terranes of the Cascadia sunduction wedge, which includes on-land exposures in the Olympic Mountains (Olympic Structural Complex) and the northern California terranes of the Coastal belt. Convergence velocity of the Juan de Fuca plate relative to North America, at the latitude of the Olympic Mountains, is closer to 43mm/yr than the 36 mm/yr shown. (From Stewart and Brandon, 2004, Fig. 1).

Early investigations of the subduction process envisioned a snowplow model with accreted material being piled up against a rigid backstop in the overriding plate; however, the backstop material was

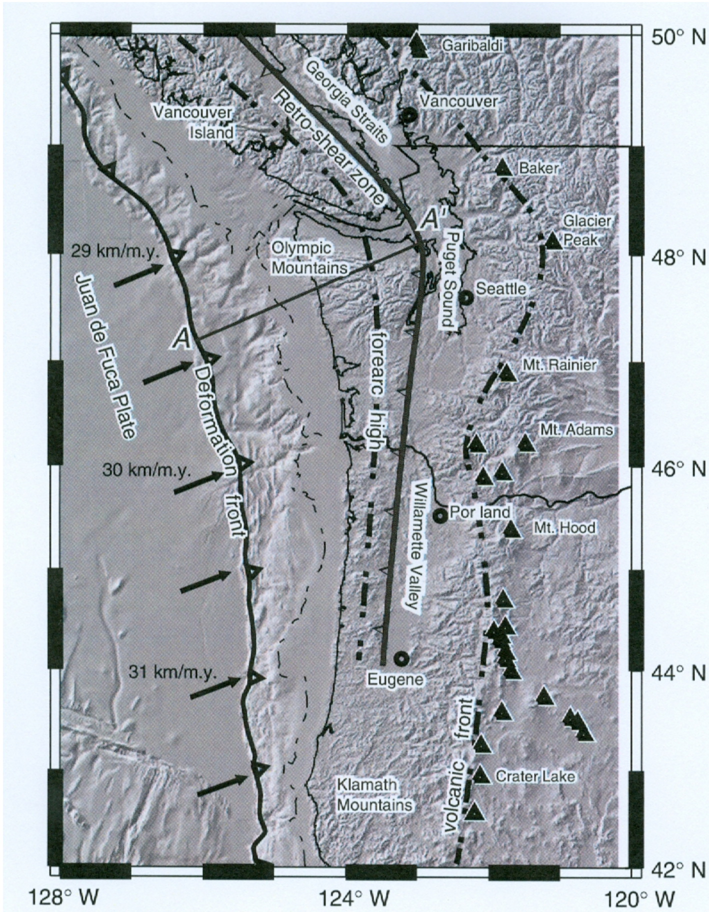

Figure 3. *Tectonic setting of the Cascadia margin showing the plate boundary, volcanic arc, and physiography of the modern margin. (From Brandon, 2004, Fig. 22.2.9)*

deforming along with the accreted sediments in many convergent plate boundaries. Sandbox experiments by Davis et al. (1983) provided a solution to the backstop problem. The experimental box consisted of horizontal layers of sand (upper plate) overlying a sheet of Mylar on a flat rigid base with adjustable dip. As the Mylar sheet was pulled beneath the sand forward- and backward-verging thrusts formed creating a double-sided wedge (a pro-wedge and retro-wedge). At the point S the "subducting" plate (pro-plate), moved below the overriding plate (retro-plate) (**Figure 4**). These experiments served as the basis for an accretionary wedge model in which the crust of the overriding plate drapes over the subduction zone of mantle (at S or retro-shear zone), causing it to deform over a broad area while creating a doubly vergent wedge (Willett et al., 1993).

As the subducting Juan de Fuca plate moves toward the overriding North American plate at about 43 mm/yr, accretion of oceanic sediments occurs mostly at the leading edge (referred to as "frontal accretion") of the North American plate, creating a doubly vergent (i.e. structures are inclined in opposite directions) wedge that progressively increases in size, maintaining a self-similar form as it grows. A far lesser amount of the sediments move down the subduction thrust fault and are added to the bottom of the wedge, a process called "underplating" (**Figure 5**). The doubly

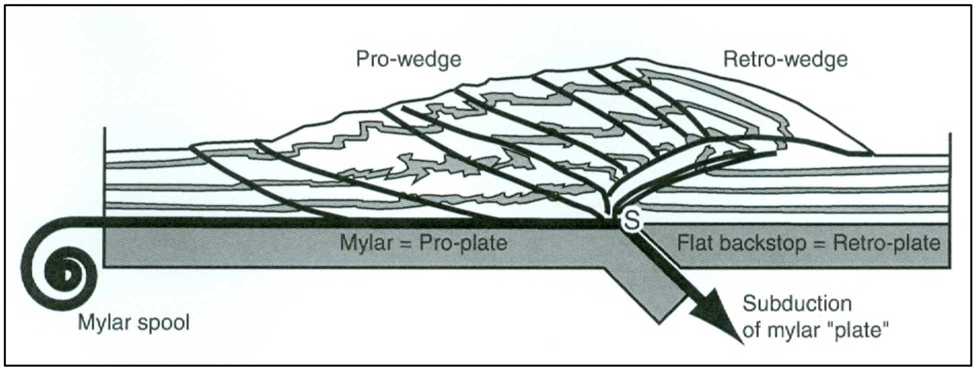

Figure 4. *Sandbox experiment producing a double-sided wedge. (Depiction by Brandon, 2004, Fig. 22.2.2 of experimental results of Davis et al, 1983.)*

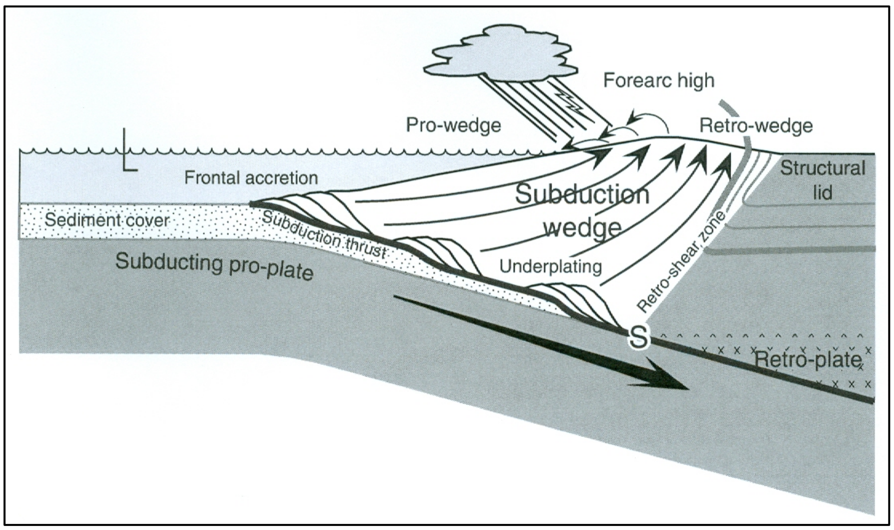

Figure 5. *Schematic cross section of a subduction wedge. "S" refers to the S point, the subduction point, where the pro-plate is subducted beneath the retroplate. (From Brandon, 2004, Fig. 22.2.1.)*

vergent Cascadia wedge has a prowedge (proside) that overrides the subducting Juan de Fuca plate accreting oceanic sediments and a retrowedge (or retroside) on the east side of the Coast Range. In the Cascadia wedge the change in vergence occurs at the crest of the forearc high. The proside of the Cascadia wedge consists of accreted marine sediments while the retroside has both accreted oceanic crust and Coast Range terrane. Mount Olympus is the approximate location of change in vergence in the Olympic Mountains, from west vergence on the proside of the wedge to east vergence on the retroside (**Figure 6**) (Pazzaglia and Brandon, 2001). The retro-shear zone in the Olympic Mountains is a large east vergent fold taking up shear between the forearc low and the eastern/retro component of the double-sided wedge.

II. STRATIGRAPHY

Rocks in the Olympic Mountains have been divided into two major tectonic sequences

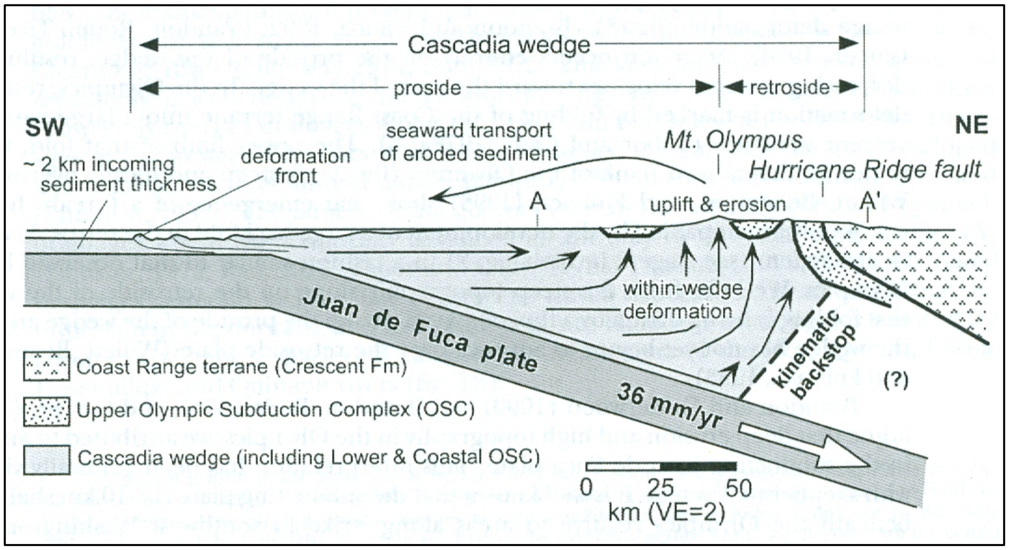

Figure 6. *Schematic cross section showing regional-scale structure of the Cascadia accretionary wedge. (From Pazzaglia and Brandon, 2001, Fig. 3., after Brandon et. al., 1998.)*

(**Figure 7**). The structurally higher sequence, called Crescent "terrane" by Babcock et al., (1994), Coast Range terrane by Brandon et al. (1998) and Stewart and Brandon (2004), crops out in a horseshoe-shape on the north, east, and south sides of the Olympic Mountains (**Figure 8**). It consists of Eocene marine sediments and oceanic basalts disconformably overlain by younger marine clastic sediments. Tabor and Cady (1978a and 1978b) and Babcock et al. (1994) use the term "peripheral rocks" for the basalts and overlying marine sedimentary units while Brandon et al. (1998) called the sedimentary units above the disconformity the "Peripheral sequence". The marine sediments and basalts below the disconformity are part of Siletzia terrane that was sutured to North America during the Eocene (Wells, 2007; Schmandt and Humphreys, 2011).

The structurally lower sequence, separated from the Coast Range terrane by the Hurricane Ridge fault, consists of Eocene turbidite sandstones (graywacke) and siltstones, a considerably lesser amount of oceanic basalts, and a small problematical slice of Mesozoic rocks at Point of the Arches in the northwest corner of the Olympic Peninsula. This sequence was called Olympic core rocks by Tabor and Cady (1978a and 1978b) and Olympic subduction complex by Brandon and Calderwood (1990) and Brandon and Vance (1992). More recently Stewart and Brandon (2004), following the International Stratigraphic Code (Salvador, 1994), named it Olympic Structural Complex (OSC).

Coast Range Terrane

The Coast Range terrane lies above the Hurricane Ridge fault, and is present in the Tofino basin on the north side of the Olympic Mountains and on the southern end of Vancouver Island (Figure 8). It is approximately 6 km (3.7 miles) thick (Tabor and Cady, 1978b; Snavely et al., 1980; Armentrout, 1987; Garver and Brandon, 1994) and consists of the Eocene Blue Mountain unit and Crescent Formation overlain disconformably by the "Peripheral

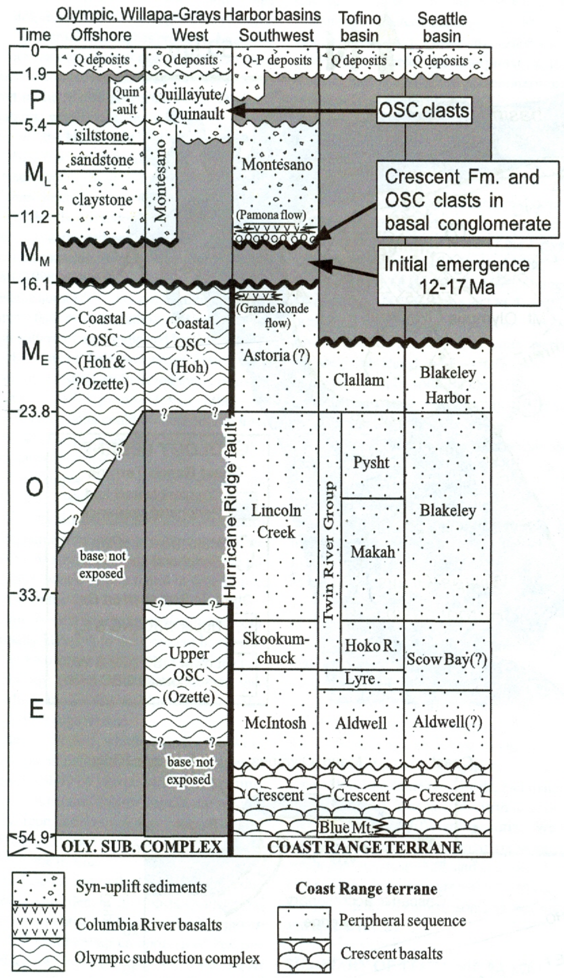

Figure 7. *Stratigraphy of the Olympic Structural Complex and Coast Range terrane. The Tofino basin is on the north side of the Olympic Mountains. E = Eocene, O = Oligocene, M_E = early Miocene, M_M = middle Miocene, M_L = late Miocene, P = Pliocene, Q = Quaternary, Fm = Formation. (From Brandon et al., 1998, Fig. 3.)*

sequence" of Brandon et al. (1998). The Peripheral sequence, in the Tofino basin on the north side of the Olympic Mountains, includes, from oldest to youngest, the Aldwell Formation, Lyre Formation, Twin River Group, and Clallam Formation. Tabor and Cady (1978a) use the term "Peripheral rocks" for the entire stratigraphic section above the Hurricane Ridge fault, not restricting it to the strata above the Crescent Formation.

Blue Mountain unit*.* The Blue Mountain unit is an informal name for marine sediments that underlie and are interbedded with Crescent basalts (Tabor and Cady, 1978a). It is overlain by basalts of the Crescent Formation except in the northeastern part of the "horseshoe" where it is directly overlain by the basal formation (Aldwell) of the upper sedimentary section, apparently filling the area between two volcanic centers (Babcock et al., 1994). Detailed investigations by Einarsen (1987) show the contact is faulted in the eastern part of the Olympic Peninsula as well as in the northwest. Babcock et al. (1994) report baked zones at the contact in two locations, one of which is along the road to Deer Park. The Blue Mountain unit largely consists of thinly-bedded turbidites but also has massive sandstones beds and a lesser amount of massive siltstone beds. The sediments probably reflect various parts of depositional fans in which the thinly-bedded turbidite deposits are middle to outer fan deposits and the massive siltstone beds are distal fan deposits. Einarsen (1987) identified a plagioclase-rich feldspathic arenite facies and chert-rich lithic arenite facies in the unit. He interpreted the sources of the sediments to be from the north to northeast with the feldspathic facies coming from the Coast Plutonic Complex and the chert-rich facies from the San Juan Islands provenance. The few paleocurrent data of Snavely and Wagner (1963) indicate southwest flow for the Blue Mountain unit, consistent with Einarsen's suggested sources for the sediments. There are no radiometric dates or fossil data for the unit. It is assumed to be the same age or older than the Crescent Formation (Rau, 1964).

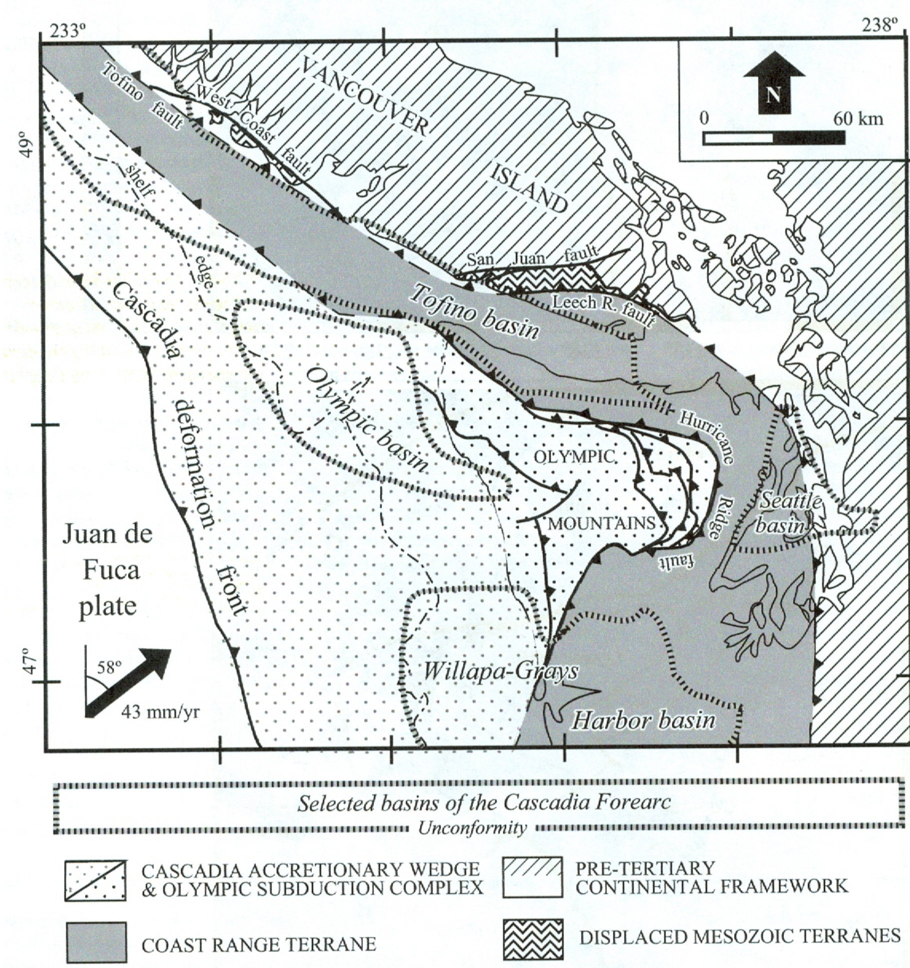

Figure 8. *Generalized geologic map of the Cascadia margin in western Washington and southern Vancouver Island. The ruled lines enclose the modern basins (Miocene and younger) of the Cascadia forearc. (From Brandon et al, 1998, Fig. 2.)*

Crescent Formation. The Crescent formation was named by Arnold (1906) for basalts that crop out at Crescent Bay. Canadian geologists call equivalent volcanic rocks on southern Vancouver Island Metchosin Formation, Metchosin Volcanics, and Metchosin Complex (Clowes, et al., 1987; Massey, 1986) despite the precedence of Arnold's work. While the Metchosin volcanics are stratigraphically continuous with the Crescent Formation mapping by Massey (1986) shows they are a typical ophiolite section with gabbros, sheeted dikes, and submarine volcanics capped by subaerial volcanics. Massey (1986) concluded the Metchosin volcanics formed as an island. Areas of diabase and gabbro sills in the Crescent Formation (Tabor and Cady 1978a) may be partial ophiolite sections. Tabor et al. (1972) and Cady et al. (1972a, 1972b) divided the Crescent Formation into two units, an oceanic lower basalt unit and subaerial upper basalt unit. The lower unit consists of mostly pillowed basalts with diabase dikes and sills and beds of volcanic breccias with minor basaltic sandstone, chert, and red limestone. (e.g. Tabor and Cady, 1978a; Suczek et al.,1994).

The upper unit consists primarily of massive basalt flows with occasional oxidized tops and paleosols.

Foraminifera ages and radiometric dates constrain the age of the Crescent Formation. Rau (1981) identified Penutian (early Eocene) to Ulatisian (middle Eocene) foraminifera in interbedded Crescent sediments consistent with $^{40}Ar/^{39}Ar$ dates. Crescent submarine basalts from the east and north sides the Olympic Mountains have yielded $^{40}Ar/^{39}Ar$ dates ranging 56.0±1.0 Ma to 45.4±0.6 Ma (see Babcock et al., 1994). The foraminifera data and radiometric dates indicate the Crescent Formation is late Paleocene to early Eocene and possible as young as middle Eocene. The age of the Crescent Formation on the Olympic Peninsula and Metchosin volcanics on Vancouver Island are virtually identical. Duncan (1982) reported an $^{40}Ar/^{39}Ar$ date of 57.8±0.8 Ma for the Metchosin volcanics on southern Vancouver Island. Published K-Ar dates for these volcanics range from 53.1±0.7 to 39.2±1.6 Ma (cited by Massey, 1986).

The Crescent basalts are one of several upper Paleocene to middle Eocene basalt units interbedded with continental-margin sediments in the Cascade forearc. Collectively the basalt units, called Siletzia, include the Roseburg and Siletz River formations in Oregon, the Crescent Formation in Washington, and Metchosen Formation/Complex on Vancouver Island. Four tectonic models, all involving hotspot volcanism, have been proposed to explain these continental-margin basaltic volcanics: (1) Yellowstone hotspot volcanism producing seamount chains that were subsequently accreted (Duncan, 1982); (2) oblique rifting of the North American plate margin as it overrode the Yellowstone hotspot (Wells et al., 1984); (3) Yellowstone hotspot volcanism through a northward moving slab window formed by subduction of a spreading ridge (Babcock et al, 1992); (4) formation of an oceanic plateau, by an oceanic hotspot, that was subsequently accreted to the continental margin (Wells, 2007). Pyle et al. (2009) report identical isotopic compositions for the Siletzia lavas and Columbia River Basalt magmas indicating they were derived from the same mantle source, although at different times. They suggest the Siletzia volcanics emanated from the plume-head phase of the Yellowstone hotspot.

Peripheral Sequence. On the north side of the Olympic Mountains post-Crescent shallow and deep marine sediments were deposited in the Tofino basin that extended northwest approximately parallel to the Strait of Juan de Fuca and north to Vancouver Island (Figure 8). From oldest to youngest, the stratigraphic units of the sequence are the Aldwell and Lyre formations, Twin River Group (Hoko River, Makah, and Pysht formations), and Clallam Formation (e.g. Babcock et al., 1994; Brandon et al., 1998). Tabor and Cady (1978a) include the Montesano Formation, found in the southwestern Olympic Peninsula, in the Peripheral sequence.

The basal Aldwell Formation disconformably overlies the Crescent Formation. It consists mostly of thin-bedded siltstone, with interbedded sandstone, and minor conglomerate and pillow basalt (Tabor and Cady, 1978a) deposited in cold, deep water (Rau, 1964). Foraminifera ages indicate the unit is Narizian (middle Eocene) (Rau, 1981); however, mapping by Squires et al., (1992) suggest the lower part of the unit may be Penutian (lower Eocene), at least in the eastern part of the Olympic Peninsula.

Uplift of southern Vancouver Island contributed coarse clastics to the basin, forming the conglomerate-rich upper Eocene Lyre Formation (Brown et al., 1956; Snavely et al, 1986a, 1989; Snavely and Wells, 1996). The age of the unit is based on late Narizian (late Eocene) foraminifera (W. W. Rau cited in Snavely, 1983). Differences in the sediments in the west and east indicate they had separate sources (Babcock et al., 1994).

The Twin River Group intertongues and overlies the Lyre Formation and includes the Hoko River, Makah, and Pysht formations. The middle Eocene Hoko River Formation consists largely of massive to thin-bedded siltstone. The overlying Makah Formation is late Eocene to Oligocene in age and consists of mostly thin-bedded siltstone with interbedded sandstone but also has a water-laid tuff and an olistostrome. The upper Oligocene Pysht Formation has a gradational contact with the Makah Formation and consists of mudstone, sandy siltstone, and a pebble and boulder conglomerate at its base. The Twin River Group is approximately 5,000 m. thick.

The early Miocene Clallam Formation conformably overlies the Pysht Formation, the upper unit of the Twin River Group. It is approximately 800 m. thick. Marine fossils indicate its predominantly feldspathic and lithic sandstone and conglomerate, that contain abundant wood fragments, were deposited in shallow water. Babcock et al, (1994) suggest the sedimentary sources for the Clallam Formation may have not been restricted to Vancouver Island and the Coast Plutonic Complex like those of the Twin River Group but also included contributions from the Coast Range.

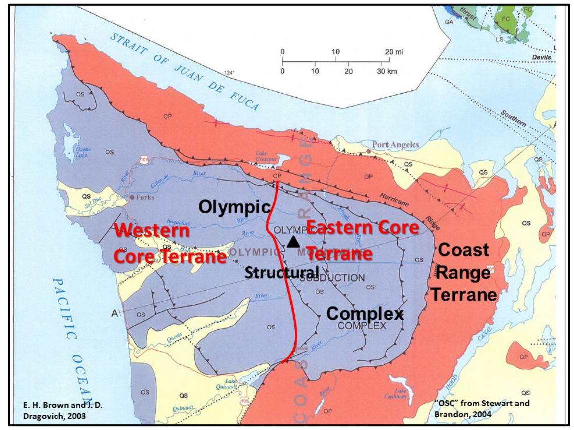

Figure 9. *Olympic Peninsula section of Brown and Dragovich (2003) tectonic map showing eastern and western core terranes of Tabor and Cady (1978b).*

Olympic Structural Complex

The Olympic Structural Complex (OSC) was subdivided by Tabor and Cady (1978b) into western and eastern core rocks (**Figure 9**). The western core, consists of Eocene to Miocene age rocks, and extends from the forearc high to the Pacific Coast. It is nonslaty and complexly folded and faulted, although it has areas that are largely stratigraphically continuous. Locally along the coast it contains mélange. The eastern core consists of Eocene to early Oligocene slaty rocks that are highly sheared and imbricated.

Based largely on differences in sandstone petrology and to a lesser extent their age and deformation, Tabor and Cady (1978a and 1978b) defined five informal lithotectonic assemblages in the Olympic core rocks. From west to east, they are the Hoh lithic assemblage, Western Olympic lithic assemblage, Elwha lithic assemblage, Grand Valley lithic assemblage, and Needles-Gray Wolf lithic assemblage (**Figure 10**). They consist primarily of marine turbidite deposits (which include a large amount of graywacke sandstones) and a lesser amount of pillow basalts and are highly deformed by

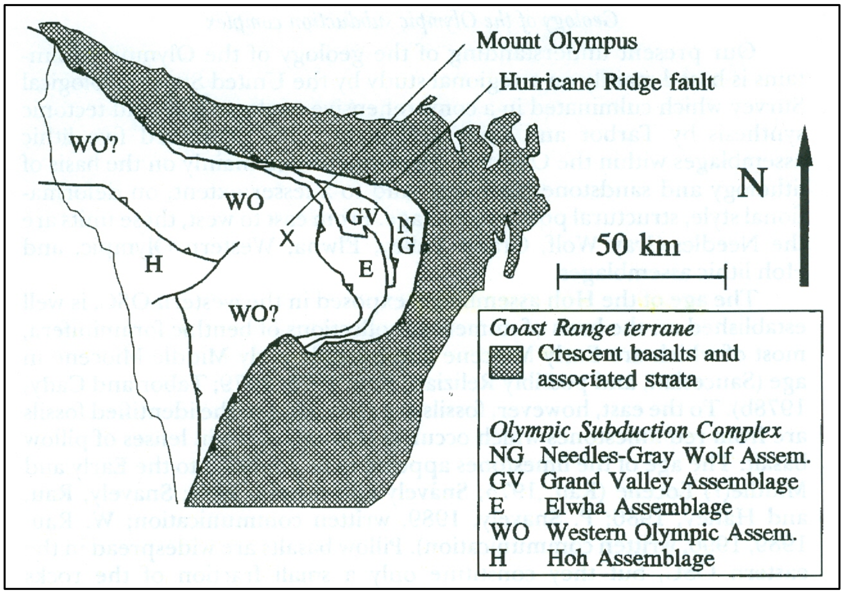

Figure 10. *Geologic map of the Olympic Mountains after Tabor and Cady (1978a). The Hurricane Ridge fault separates the Coast Range terrane from the highly deformed Olympic Structural Complex. (From Brandon and Vance, 1992, Fig. 3.)*

imbricate thrusts and folds. The ages of the assemblages mostly increase from west to east; however, based on a few fossils Tabor and Cady (1978b) suggest a probable late Eocene age for the Needles Gray-Wolf lithic assemblage and an early to middle Eocene age for the Elwah lithic assemblage. They interpret the Western Olympic lithic assemblage to be late Eocene to early Oligocene given its lithologic continuity with other fossiliferous rocks.

Using new zircon fission-track data Brandon and Vance (1992) proposed Tabor and Cady's (1978a) five lithic assemblages be combined into three units named the Upper, Lower, and Coastal Olympic Subduction Complex (OSC) units (**Figure 11**) and suggest there is a "subtle" younging of rocks westward from ~19 Ma at Mount Olympus to ~14 Ma at the Pacific coast. With the formal name change of the core rocks by Stewart and Brandon (2004) the OSC abbreviation now means Olympic Structural Complex.

Upper OSC. The Upper OSC of Brandon and Vance (1989), characterized by turbidite clastic rocks and large blocks of pillow basalts, is the structurally highest unit. Unreset zircon fission-track dates of 48-32 Ma from Upper OSC rocks indicate they are lower Eocene to lower Oligocene (Brandon and Vance, 1992). In the east the Upper OSC includes the Needles-Gray Wolf and Elwha assemblages. In the west it includes the western half of the Western Olympic assemblage (Tabor and Cady, 1978a) and the Cape Flattery area Ozette and Soos terranes (Snavely et al. (1986a). Brandon and Vance (1992) point out that foraminifera fossils in Ozette and Soos strata indicate the terranes are Eocene to Oligocene (Snavely et al. 1986b) and are, therefore, coeval with their fission-track zircon ages of Upper OSC sandstones.

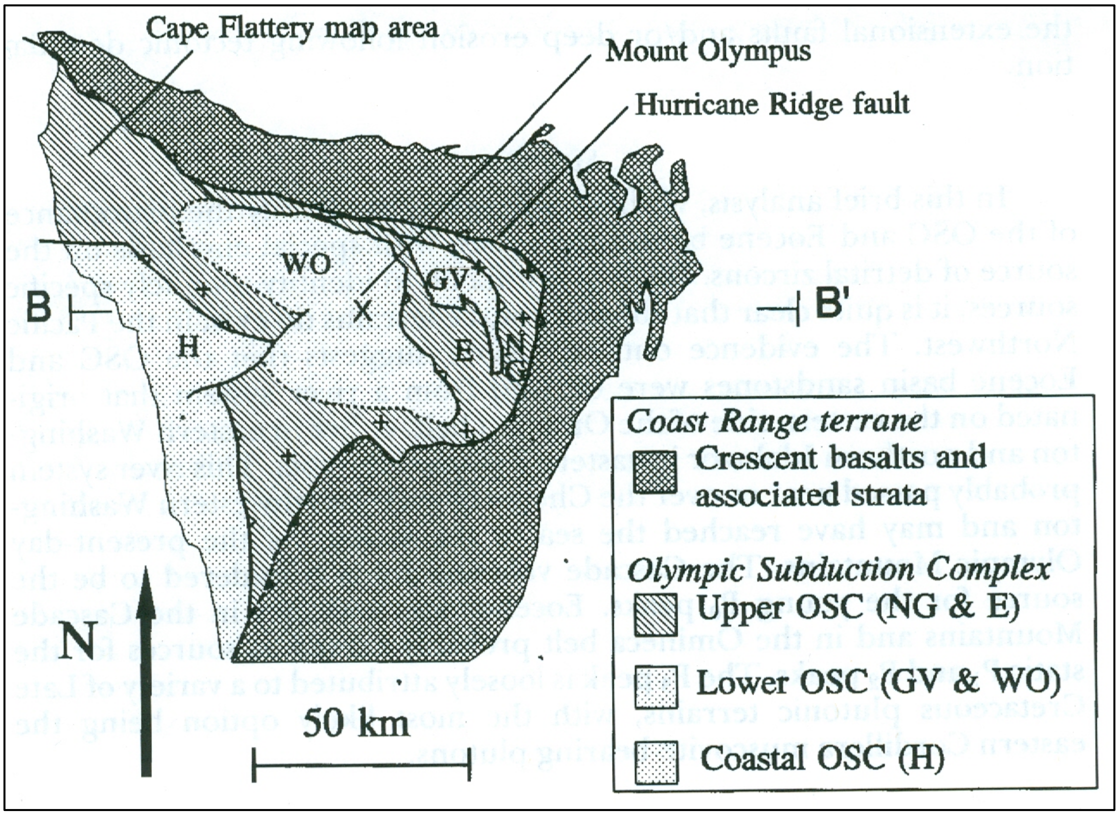

Figure 11. *Geologic map of Brandon and Vance (1992, Fig. 15) showing revised structural interpretation of the Olympic Structural Complex developed by factoring in zircon fission-track ages.*

In the Mount Claywood area in the eastern OSC the pillow basalt blocks are overlain by a "stratigraphic cover" of maroon pelagic limestone, green chert, sandstone, and shale and have an oceanic-island chemistry (Applegate and Brandon, 1989). Fossils in the interbedded limestones indicate the basalt blocks are early Eocene in age (Rau, 1975, 1981; Tabor and Cady, 1978b; and Snavely et al., 1986b). Applegate and Brandon (1989) and Brandon and Vance (1992) concluded the basalt blocks are Crescent Formation to which they are stratigraphically and chemically identical and suggest the Upper OSC clastic rocks were deposited on lower Crescent-like basement. Based on these observations Brandon and Vance (1992) infer the Upper OSC clastic rocks did not form by accretion but rather as part of the Coast Range terrane, possibly by a section of the western Coast Range terrane being broken off and underthrust eastward with the initiation of subduction. They suggest the underthrusting and imbrication of the Upper OSC must have started after ~33 Ma as its youngest zircon ages of 39-33 Ma are from clastic rocks of the Needles-Gray Wolf assemblage.

Lower OSC. The Lower OSC is exposed in the central part of the Olympic Mountains in an area around Mount Olympus with the greatest topographic relief on the Olympic Peninsula. A metamorphic mineral assemblage in this unit (prehnite + pumpellyite) suggests it is the most deeply exhumed part of the Olympic Mountains, having been buried to a depth of 11 km (Brandon and Calderwood, 1990). The Lower OSC is inferred to underlie the Upper unit (Brandon and Vance, 1992). It includes the eastern half of the Western Olympic assemblage and the entire Grand Valley assemblage. Unlike the Upper OSC it has no pillow basalts and consists entirely of clastic sediments that are mostly turbidites. Young zircon dates, which range from 27-19 Ma, indicate the unit was deposited from the late Oligocene to early Miocene. Brandon and Vance (1992) suggest the Lower OSC reflects material formed strictly by accretion through subduction underplating in contrast to the Upper OSC which apparently formed by shortening within the leading edge of the North American plate.

Coastal OSC. The Coastal OSC is equivalent to the Hoh lithic assemblage of Tabor and Cady (1978a) and has also been called Hoh Formation (Weaver, 1916, 1937), and Hoh rock assemblage (Rau 1975, 1979). It was formally named Coastal unit of the Olympic Structural Complex (Coastal OSC) by Stewart and Brandon (2004). The Coastal unit consists of mostly northeast-dipping imbricated slices of turbidites and mélange. The west-verging orientation of the thrusts mimic those described by Davis and Hyndman (1989) in bathymetric data off Vancouver Island and may penetrate down to or close to the underlying oceanic crust. Foraminifera in the siltstones have ages ranging from late Eocene to middle Miocene (Rau 1973, 1975, 1979; Rau and Grocock, 1974). Based on fission-track dates for detrital zircons from 34 sandstone and 2 volcanic ash samples and other criteria Stewart and Brandon (2004) determined the Coastal OSC sediments came from an active volcanic arc and older units, including Cretaceous metamorphic rocks, during the early Miocene (ca. 23–16 Ma).

Rau (1975, 1979) found deep-water (~2000 m.) benthic foraminifera to be common in the Coastal OSC as well as shallow-water foraminifera and concluded the water was >200 m. deep. He also found Miocene megafossils in the Coastal unit that live in very shallow water (10 to 30 m.). Applying the paleobathymetry biofacies of Ingle (1980) to the deepest water taxa, Stewart and Brandon (2004) concluded the unit was deposited in the Cascadia trench in water 2000 to 4000 m. deep and the mixed assemblage of both deep and shallow water fossils to be the result of downslope transport by turbidite flows.

Diapiric and fault (shear)-zone mud-matrix mélanges are present in the Coastal OSC (Orange, 1990). They contain clasts of volcanic and sedimentary rocks. Orange et al. (1993) describe northeast moderately-to-steeply-dipping fault zones in the Coastal unit, ~200 m. to >1000 m. thick, that have highly deformed mud-matrix mélange. They are characterized by pervasive uniform foliation orientation, very intensely-developed scaly foliation, exotic clasts, and a consistent fold vergence that indicate thrusting to the west-southwest. Broad to tight folds in the foot and hanging walls are oriented obliquely to and are truncated by the fault-zone mélange, indicating they formed after the folding. They cite evidence for the presence of considerable fluids in the Coastal OSC including petroliferous sandstone and matrix that have been called "smell muds" (which when freshly broken give off an odor like kerosene) associated with the mélange fault-zones and veins in both the mélange and foot walls and to a lesser extent in the hanging walls.

Diapiric mélange, in contrast to fault-zone mélange, is characterized by an arcuate foliation pattern in map view, increasing matrix scaly foliation from the core toward the margin, opposing fold vergence on opposite margins, and rare exotic clasts (Orange, 1990). Orange and Campbell (1997) suggest the Coastal OSC mélange diapirs formed when the migration of fluids in fault-zone mélange was restricted causing the mélange to become overpressured, resulting in buoyant rise of mélange on faults (**Figure 12**).

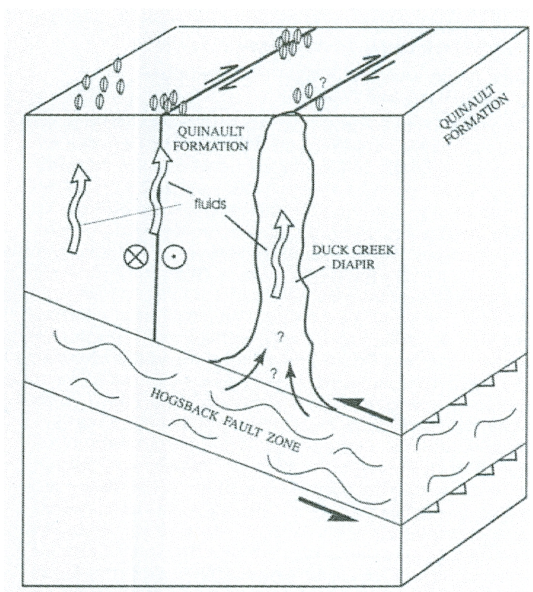

Figure 12. *Schematic diagram illustrating the relation of a fault zone mélange (Hogsback) and diapir mélange (Duck Creek) just north of Taholah, Washington. (From Orange and Campbell, 1997, Fig. 12.)*

Exotic blocks of Eocene pillow basalt and sedimentary rocks are present in the Coastal unit. Tabor and Cady (1978a) mapped these older rocks as Western Olympic assemblage. Brandon and Vance (1992) suggest the basalt blocks, like those in the Upper unit, were derived from overlying Coast Range terrane as they are similar in age and chemistry to Crescent basalts. They suggest downslope mass-wasting moved the volcanic blocks and their connected sedimentary rocks from the overlying Crescent basalt to the trench and subsequently incorporated them into the accretionary wedge. Stewart and Brandon (2004) report zircon fission-track minimum ages for Coast OSC mélange sandstone blocks of 39 to 15 (late Eocene to middle Miocene) Ma. They suggest that the presence of Eocene fossils in the mélange sedimentary rocks (Rau, 1975, 1979) is consistent with the basalt blocks in the mélange being derived from the Eocene Crescent Formation. Some of the basalt clasts in the Coastal OSC mélange might be oceanic basalt caught up in mud-matrix mélange fault-zones that penetrate to the oceanic crust, like the thrusts described by Davis and Hyndman (1989) on the Vancouver Island continental shelf.

III. TECTONIC OVERVIEW

Two major geologic features make up the Olympic Mountains segment of the Cascadia subduction wedge: an accretionary complex, called Olympic Structural Complex (OSC), and the peripheral Coast Range terrane. Rau (1975, 1979) and Tabor and Cady (1978a, 1978b) proposed a steep imbricated structure model for the Olympic Mountains (**Figure 13A**). More recently, Brandon and Vance (1992) and Brandon et al, (1998) proposed a domal imbricated structure model in which the Coast Range terrane is a structural lid on the accreted part of the wedge (**Figure 13B**). The overall structure of the Olympic Mountains is an east-plunging antiform.

The structurally higher section of the Coast Range terrane, directly overlying the OSC, is a fault-bounded basement slab of upper

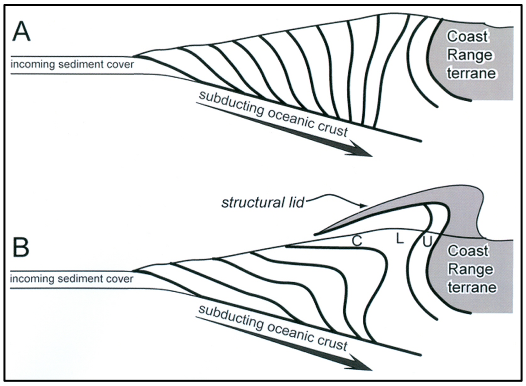

Figure 13. Schematic cross sections of the Cascadia wedge. (A) Steep imbricate structure as proposed by Ray (1975, 1979) and Tabor and Cady (1978a, 1978b) and (B) Domal imbricate structure as proposed by Brandon and Calderwood (1990) and Brandon and Vance (1992), Abbreviations refer to tectonic units exposed in the Olympic Mountains: U = Upper OSC, L = Lower OSC, and C = Coastal OSC. (From Stewart and Brandon, 2004, Fig. 11.)

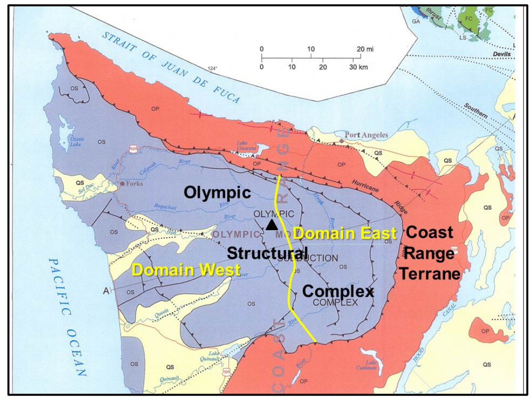

Figure 14. Structural domains of Tabor and Cady (1978b). (Adapted from Tabor and Cady, 1978b, Fig. 24.)

Paleocene and lower Eocene marine rocks consisting of the Blue Mountain unit and Crescent Formation basalts. The slab's lower boundary fault, the Hurricane Ridge fault, is exposed in the Olympic Mountains. The suture boundary of the slab is not exposed in the Olympic Mountains but is exposed on the southern end of Vancouver Island as the Leech River fault (Clowes et al., 1987).

The structurally lower OSC consists of highly deformed Eocene to middle Miocene marine sediments with minor igneous rocks. Common structural elements include folds, imbricate faults, cleavage, and pencils. Tabor and Cady (1978b) identified two major structural domains in the OSC based on pencil orientation because they found pencils to be the "most consistent" structural feature. Planar structures dip east and pencils plunge east in Domain West. In Domain East planar structures dip west and pencils plunge west (**Figure 14**). The boundary separating these domains is sub-parallel to the boundary of their eastern and western core terranes and the Coast Range forearc high (**Figure 15**). The west-verging structures of Domain West reflect pro-wedge kinematics and east-verging structures of Domain East reflect retro-wedge kinematics. It is instructive that the overall geometry of the OSC structures mimic those of Davis et al.'s (1983) sandbox model (**Figure 16**).

The active wedge is some 200 – 250 km wide and is bounded by on the east by the forearc low and on the west by the Cascadia trench, located ~140 km west of Olympic Peninsula coastline. Stewart and Brandon (2004) viewed the Coast Range terrane as part of the active wedge because of its folding and uplift associated with development of the doubly vergent wedge. Brandon and Vance (1992) infer Siletzia (Coast Range terrane) must have sutured onto North America along the sinistral Leech River fault on southeast Vancouver Island at ~38 Ma because it places rocks metamorphosed to greenschist and

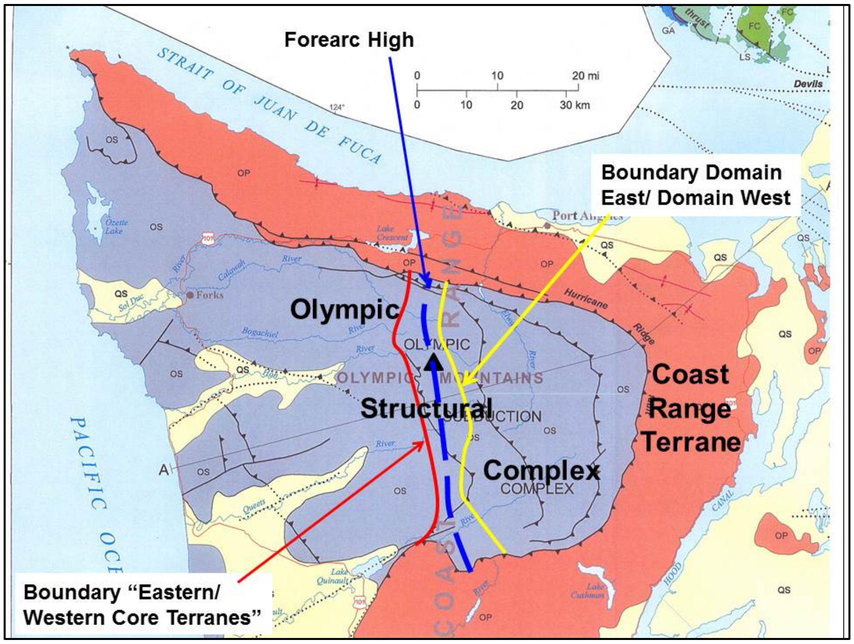

Figure 15. *Olympic Peninsula section of tectonic map by Brown and Dragovich (2003) showing the subparallel orientations and close proximity of Tabor and Cady's (1978b) eastern and western core terrane boundary and Domain East/Domain West boundary to the forearc high.*

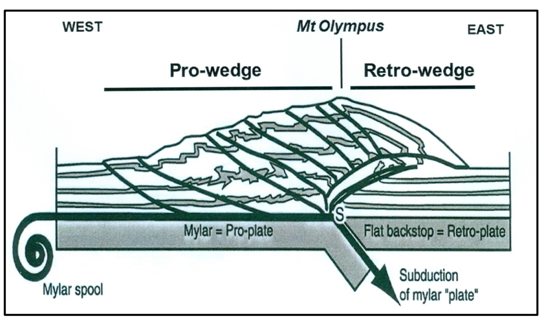

Figure 16. *Sandbox model of Davis et al. (1983) mimics structure of the Olympic Mountains.*

amphibolite facies at ~40 Ma (Fairchild and Cowan, 1982) against low-grade Siletzia Metchosen basalts. Based on the suturing of the Coast Range terrane onto the North American plate at ~38 Ma, the onset of volcanism at ~36 Ma in the Cascade volcanic arc (e.g. Armstrong and Ward, 1991), and slip on the Hurricane Ridge fault after ~33 Ma indicated by the ages of the youngest clastic rocks (late Eocene 39 to 33 Ma Needle-Gray Wolf assemblage) underthrust beneath it, Brandon and Vance (1992) suggest the Cascadia subduction zone developed at ~35 Ma in the late Eocene rather than at 55 to 50 Ma as interpreted by previous investigators (e.g. Wells et al., 1984; Heller et al., 1987; Snavely, 1987).

Several recent investigations favor the earlier ~55 Ma date for the docking of Siletzia terrane. Schmandt and Humphreys (2011) concluded the Cascadia subduction zone developed its current configuration by ~40 Ma, following initial accretion of the Siletzia microplate in the Columbia Embayment at ~55 Ma. Sigloch and Mihalynuk (2013) used subducted slab relics (which reflect ancient plate configurations) to identify accretion events and then determined the time of their collision and reconciled them with plate reconstructions and surface geology. Applying this analysis to Cascadia they inferred a (Siletzia) volcanic arc, over a west-dipping intra-ocean subduction zone, was overridden by westward movement of the continent resulting in accretion of Siletzia to North America at 55±7 Ma and development of an eastward dipping subduction zone. Wright et al.'s (2015) modelling of sea-floor spreading history quantitatively showed the Pacific-Farallon spreading rates increased between ~58 and ~56 Ma. They suggest that this increase in Farallon absolute plate motion may have resulted from the development of east-dipping subduction, which Sigloch and Mihalynuk (2013) link to Siletzia accretion. The Schmandt and Humphrey (2011, Fig. 2) model may offer a way to reconcile evidence for Siletzia accretion at both ~55 Ma and ~38 Ma with the terrane suturing like a zipper starting in the Columbia Embayment in the early Eocene and progressively suturing northward, closing on Vancouver Island in the late Eocene.

Development of the Cascadia subduction zone resulted in extensive imbrication of the western Coast Range terrane, and possibly major westward displacement of Coast Range terrane on the Hurricane Ridge fault as Brandon et al. (1998) suggest. Subsequently the Coast Range terrane was eroded away exposing the Olympic Structural Complex. As the Juan de Fuca and North American plates converge, currently at a rate of 36 mm/yr (Demets and Dixon, 1999), apparently all of the sediment is accreting onto the wedge (e.g. Davis and Hyndman, 1989; Pazzaglia and Brandon, 2001).

Origin of Coast Range Horseshoe

Two hypotheses have been suggested for the origin of the horseshoe arcuate pattern of the Coast Range terrane: 1) Oroclinal bending of the forearc about a vertical axis caused by northeast directed compressive stress of the subducting Juan de Fuca plate (e.g. Tabor and Cady, 1978b; Beck and Engebretson, 1982) or clockwise rotation of the Cordilleran margin caused by extension in the Basin and Range province (Brandon and Calderwood, 1990), and 2) More gentle dip of the Juan de Fuca plate beneath the Olympic Peninsula (Crosson and Owens, 1987), relative to the areas to the south and north, leaving less space to accommodate the wedge increasing in size (Brandon et al, 1998) thousands of feet above sea level (e.g. Davis and Hyndman, 1989, Warnock et al, 1993). Pazzaglia et al. (2003) propose a variation of these models in which northeast movement of the OSC bulldoze the Coast Range rocks into the horseshoe configuration rotating the rocks north of the OSC counterclockwise and clockwise south of the OSC. Paleomagnetic studies show rocks south of the OSC have been rotated clockwise (Wells and Coe, 1985; Wells, 1990; England and Wells, 1991; Prothero and Nesbitt, 2008) and counterclockwise north of the OSC (Bates et al.,1981; Beck and Engebretson, 1982; Prothero et al., 2008; Prothero et al, 2009), supporting the Pazzaglia et al. (2003) model.

Uplift of Olympic Mountains

Using Benioff zone seismicity, Crosson and Owens (1987) were the first to document the Juan de Fuca plate arches upward beneath the Olympic Peninsula. Brandon and Calderwood (1990) superimposed cross sections of geophysical data (seismicity data, refraction and gravity data, reflection data, and teleseismic receiver function analysis) across southern Vancouver Island, the Olympic Mountains, and southwestern Washington. The cross sections show the slab beneath the Olympic Peninsula is an arch ~10 km higher than areas to the north and south. They attribute the anomalous uplift and deep exhumation of the Olympic Mountains to the arch and suggest it began developing at about 15 Ma as a result of extension in the Basin and Range province causing clockwise rotation of the Cordilleran margin which bowed the subducting slab. More recently Pazzaglia and Brandon (2001), using deformed erosional benchmarks, found that the rates of uplift and erosion of the Olympic Mountains are consistent with the convergence accretionary flux. On the basis of this apparent relationship of uplift and erosion to the accretionary flux they suggest that long-term permanent strain is primarily occurring across the orogen over a 140 km northeast-southwest baseline at a rate of 3 km per Myr, parallel to the convergence direction of Juan de Fuca-North American. Batt et al. (2001) use a two-dimensional kinematic and thermal model, constrained by fission track and (U-Th)/He ages for apatite and zircon, that produces distributions to the mineral ages similar to those found in the Olympic Mountains. They conclude the Cascadia wedge has been in steady state flux since about 14 Ma with the accretionary flux balanced by the erosional flux and that the apparent balance of the fluxes is consistent with their argument that uplift of the Olympic Mountains has been primarily driven by accretion of material from the subduction Juan de Fuca plate and not margin-parallel transport.

Based on the movements of subplates or blocks indicated by paleomagnetic and GPS data and structures of deformed Quaternary deposits, other workers (e.g. Wells et al., 1984; McCaffrey and Goldfinger, 1995; McCrory, 1996; Wang, 1996; McCaffrey, et al, 2007) have attributed uplift of the Olympic Mountains to margin-parallel shortening. In this model the Olympic Peninsula is being compressed between the northward moving Oregon block and the comparatively stable rocks of southeastern Canada and northwestern Washington. There is, in fact, considerable evidence that the Olympic Peninsula has been under north-south compression since the middle Miocene. Evidence near the Washington coast for Quaternary northward shortening cited by McCrory (1996) includes steeply dipping and faulted glaciofluvial strata at the crest of an east-northeast trending ridge; thrust and reverse faults farther east on the ridge striking northeast (N65^{0}E to N80^{0}E) and dipping northwest (25^0 to 55^0) offset Pleistocene gravel beds; borehole break-out data of Magee and Zoback (1992) that yield principal compressive stress orientations of N14^{0}E and N5^{0}W; and Pleistocene gravels oriented N80^{0}W, 26^{0}N in an east-trending ridge. This ridge and the other ridge, appear to be anticlinal folds produced by blind thrusts.

Wells et al. (1998) and Wells and Simpson (2001) attribute the north-south compression to the Neogene rotations of the Sierra Nevada block moving N47±5^{0}W at 11±1 mm/yr from Global Positioning System (GPS) data (Dixon et al., 2000) and a semi-ridged Oregon coastal block rotating clockwise at an average Cenozoic rate of 1.19±0.10^0/my. In the revised rotation model of Wells and Simpson (2001) the Olympic Peninsula, which is under uplift and transpression in the Wells et al. (1998) model, is a region of north-south shortening and ~6 mm/y of shortening with respect to North America (**Figure 17**) due to the migrating forearc block abutting against the comparatively stationary Mesozoic and older rocks of southwestern Canada and northwestern Washington (Johnson et al., 1996; Wells et al., 1998; Johnson et al., 2004).

It can be argued that the convergent accretionary flux/erosional flux and margin-parallel transport models for uplift of the Olympic Mountains are not necessarily mutually exclusive. It is possible the two mechanisms, north-directed compression from the rotating Oregon block and accretionary flux associated with northeast convergence of the Juan de Fuca plate, are acting together to produce the uplift. There is no apparent kinematic requirement for an accretionary/erosional flux balance of the Olympic Mountains to be incompatible with coeval margin-parallel shortening.

Exhumation of the Olympic Mountains

The exhumation of the Olympic Mountains

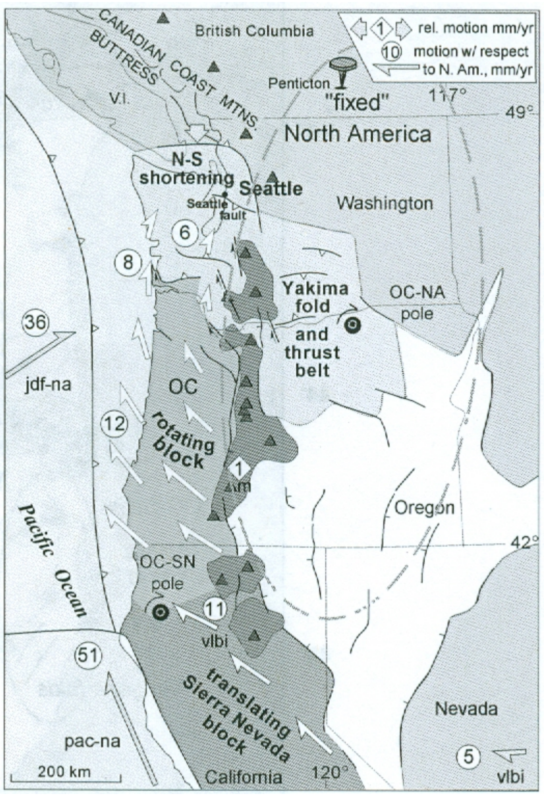

Figure 17. *Velocity field for Cascadia forearc calculated for Oregon Coast (OC) – North America (NA) pole of Wells et al. (1998) revised microplate model. (From Wells and Simpson, 2001, Fig. 4B.)*

has been investigated using the fission-track method. The fission-track method of dating minerals is done by determining the density of tracks (damage zones in the lattice of minerals) left by the spontaneous fission of ^{238}U. Track density is a function of the rate of production of new tracks and the removal of old tracks by thermal annealing that resets the mineral age. Unreset samples of detrital zircons and apatite retain the pre-depositional fission-track ages which reflect the pre-depositional source areas. Reset samples define the cooling age of rocks that are initially subjected to temperatures elevated enough to anneal fission-tracks followed by a reduction in temperature to levels at which fission-tracks are retained. Comparing the minimum age with the depositional age - in unreset mineral grains the minimum age is older than the time of deposition and in reset mineral grains the minimum age is younger than the time of deposition. Brandon et al. (1998) used the zircon fission-track data of Brandon and Vance (1992) and apatite fission track data to determine the exhumation history of the Olympic Mountains. Reset minimum ages of zircon (~14 Ma) and apatite (~15 Ma) define a concentric pattern with the youngest ages in the central massif of the mountains and progressively older ages away from the reset area (**Figure 18**).

Analysis of seismic data indicates sediments 2-3 km thick are transported into the subduction zone and increase in thickness to 20 km at the Peninsula's coast and ~35 km below the core of the Olympic Mountains. Zircon fission-track data indicate the 27 to 19 Ma lower Olympic Structural Complex was underplated below the wedge at a depth of 12.1 to 14.5 km and temperatures of ~242º-289ºC, consistent with the occurrence of prehnite-pumpellyite assemblages and pressure-solution cleavage (Brandon and Vance, 1992). A lower metamorphic grade

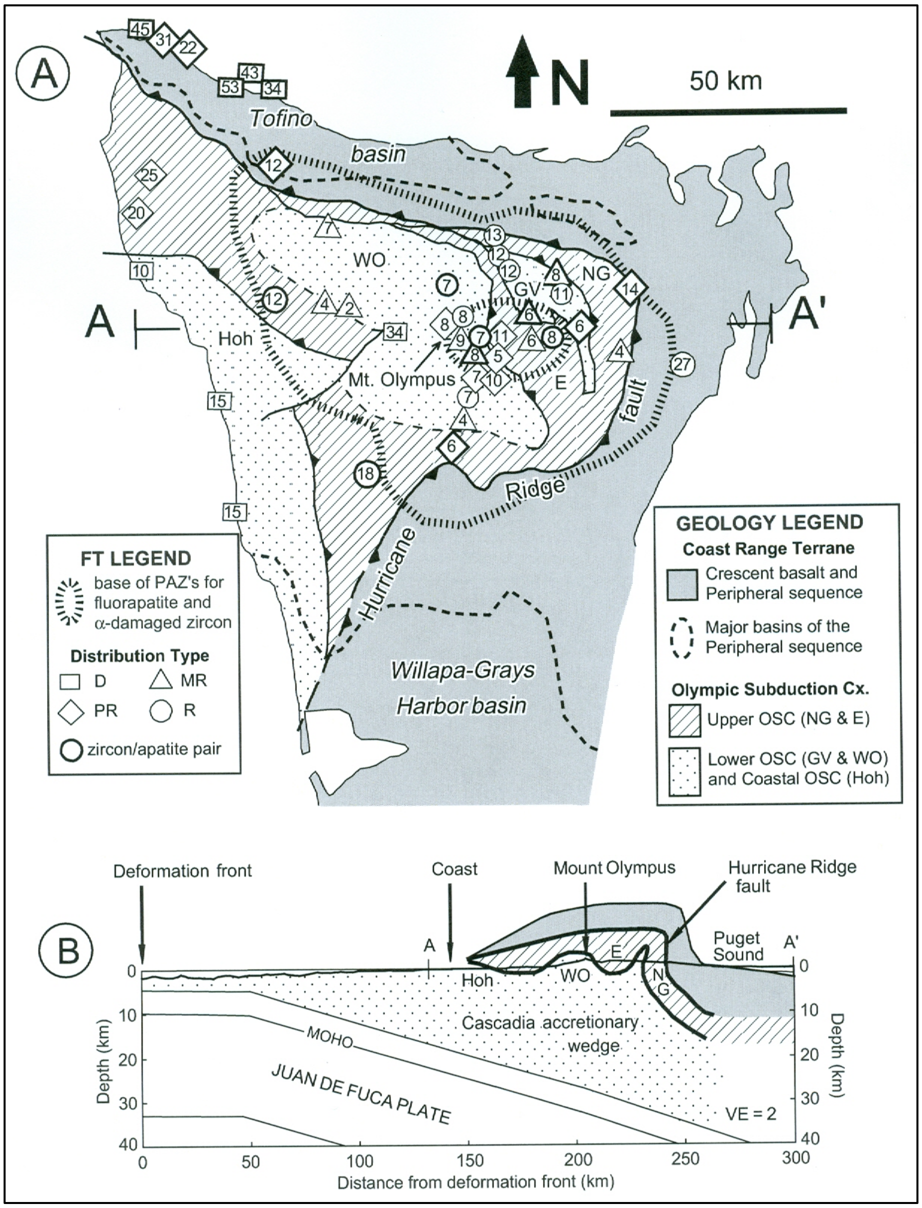

Figure 18. *Generalized (A) geologic map and (B) cross section for the Olympic Structural Complex (OSC). Numbers indicate minimum ages for apatite fission-track (FT) samples and are considered to be reset ages for apatite sample types MR (triangle), PR (diamond), and R (circle). PAZ = partial annealing zone. Bold lines indicate localities with published zircon fission-track ages. The outer and inner ruled lines mark the base of the partial annealing zone for fluorapatite and α- damaged zircon respectively. Tectonostratigraphic units of Tabor and Cady (1978a) are shown as NG = Needles-Gray Wolf, E = Elwha, GV = Grand Valley, WO = Western Olympic, and Hoh. (From Brandon et al., 1998, Fig. 4.)*

and absence of cleavage in the Coastal OSC suggest it was accreted by off-scraping at the front of the wedge (Orange et al., 1993). Apatite and zircon fission-track data indicate the central core of the Olympic Mountains first began emerging above sea level in the early Miocene at ~18 Ma and migrated outward with time, reaching the northwest corner of the Peninsula at ~12 Ma while sedimentary rocks were continually accreted to the wedge (Brandon et al., 1998) (Figure 18). Using three thermochronometer ages (apatite U-Th/He and apatite and zircon fission-track) from the Olympic Mountains that show a nested pattern of reset-age zones (**Figure 19**), Willett and Brandon (2002) infer the exhumation rate of the Olympic mountains did not reached a steady state at ~14 Ma as proposed by Batt et al. (2001) since the reset ages are limited to the central massif. Willett and Brandon (2002) suggest the Olympic Mountains will be in an exhumation steady state when eastward kinematic transport brings the reset-age zones to the northeast edge of the orogen.

Megathrusts

Slip on the basal/subduction thrust, also called the Cascadia fault, of the Juan de Fuca plate during mega-earthquakes (≥ 9.0) apparently accommodates most of the convergence with the North American plate and a lesser amount of convergence by shortening in the subduction wedge. Subduction zone earthquakes on the Cascadia fault are thought to have a recurrence interval of about 500 years in southern Washington. Tsunami deposits in Discovery Bay indicate a shorter earthquake recurrence interval of about 300 years, which might reflect movements on other faults or a subduction zone earthquake on the north end of the Cascadia fault (Brian Atwater, personal communication 2015). The linkage of geologic data from investigations in Washington and historic records of tsunamis arriving on the coast of Japan show that the most recent of these giant earthquakes occurred on January 26, 1700 (Atwater, B. F. et al., 2005).

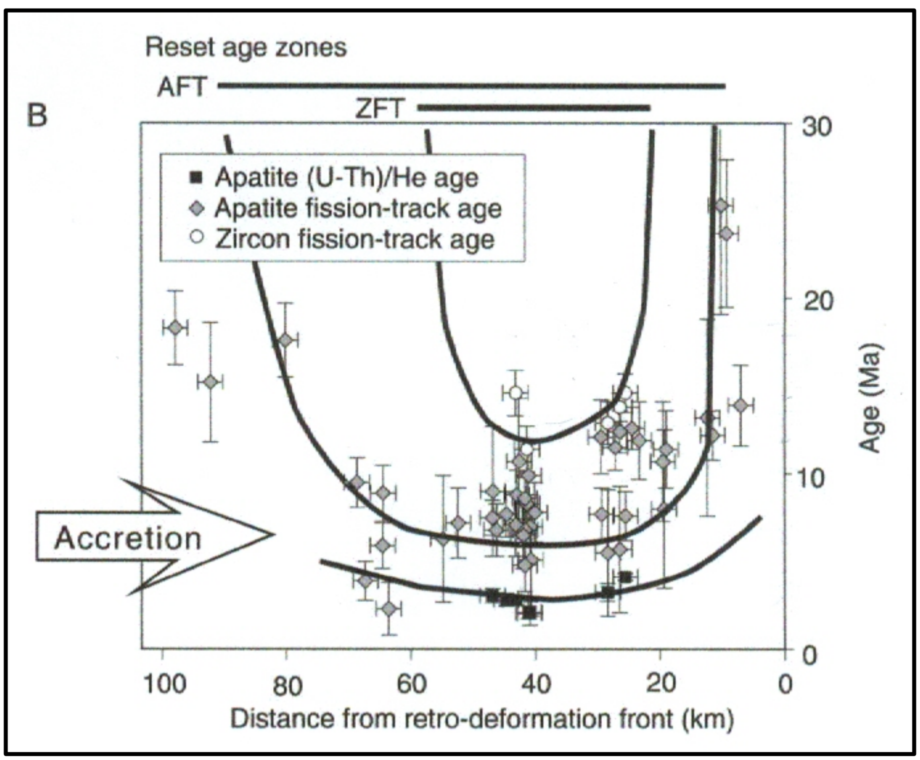

Figure 19. *Ages from three thermochronometers from the Olympic Mountains showing nested pattern of reset zones. (From Willett and Brandon, 2002, Fig. 3B. Adapted from Batt et al., 2001, Fig. 5 with unreset ages older than 30 Ma not plotted.*

ROAD LOG

Planning for the Field Trip

Mileage for each day is measured from the parking lot of the field trip headquarters, the Red Lion Hotel in Port Angeles, Washington. The geology described on each route is based on the geologic map of the Olympic Peninsula by Tabor and Cady (1978a).

When planning this trip set the schedule so as to arrive at Beach 4 approximately three hours before low tide. This will allow sufficient time at Beach 4 and Ruby Beach before the incoming tide becomes a problem. The stop at Lake Crescent, to see and discuss the Crescent Formation and the lake, may be made either before or after Beach 4 and Ruby beach stops depending on the best time with respect to the tide. Non-stop driving time from the Red Lion Hotel to Beach 4 is one and one-half (1½) hours.

On this field trip the Lake Crescent overlook stop is Stop 3 (not Stop 1 as on a field trip in 2013), in order to arrive at Beach 4 at 9:30 am. Low tide is at 11:00 am.

DAY 1: SATURDAY, OCTOBER 3, 2015

Meet in Red Lion Hotel parking lot at 7:45 am for prompt departure at 8:00 am.

Mile

0.0 Turn left out of the parking lot onto North Lincoln Street and then first right onto East Front Street. You'll be driving on continental glacial deposits (Qc) for several miles.

0.7 Left turn onto Tumwater Road (truck route).

2.2 Junction of Tumwater Road and US 101. Keep right to get on US 101 west.

3.2 The low hills on the south side of US 101 are sandstones and siltstones of the lower member (Ttrl) of the Twin River unit. Tabor and Cady (1978a) mapped these as Twin River Formation (Ttr). Snavely et al. (1978) raised the unit from formation to group rank and subdivided it into three new formations. The basal Aldwell Formation is overlain by the Hoko River Formation, a deep-water marine sequence of sandstones and siltstones with lenses of conglomerate and lithic sandstone with basalt and metamorphic rock clasts. Calcareous concretions in the Hoko River contain mega-fossils (e.g. crabs) and carbonized wood. In some locations the Lyre Formation, which consists of conglomerate (Tlc), sandstone and minor siltstone (Tls), andesitic volcanics (Tlv), and breccias and conglomerate (Tlb), is overlain by the Hoko River Formation and in other places interfingers with it. The Makah Formation overlies the Hoko River Formation on a major unconformity and consists of deep-water turbidite deposits. The lower half of the Makah contains late Eocene (Refugian) benthic foraminifera and upper half Oligocene (Zemorrian) foraminifera.

5.8 Intersection of US 101 and WA 112. As the road descends north along the Elwah River we will be passing through the Peripheral rocks of the Coast Range terrane which in descending order are sandstone and siltstone (Ttrl) of the Twin River Group, conglomerate of the Lyre Formation (Tlc), and conglomerate, sandstone and siltstone of the Aldwell Formation (Tal). Just before we cross the Elwah River we will be passing basalts of the Crescent Formation (Tbc) consisting of flows, pillows, and breccia.

7.3 Bridge over the Elwha River. Two dams, one north of US 101 and one south of the highway, were constructed in the early 1900s. They were removed over a three year period starting in September 2011 with the last section of the remaining dam being dismantled in August 2014. Scientists conducted studies on the Elwha River while the dams were in place to develop a baseline against which to assess changes to the geology, fauna, and flora caused by removal of the dams. One encouraging change is the presence of salmon observed above the location of the lower dam, a section of the river the fish had not been able to access for a century. On the west side of the bridge the road is back on continental glacial deposits (Qc).

8.5 Outcrops of Eocene Crescent Formation basalt (Tbc) are along the road for about the next half mile. Baldy Ridge on the south side of the highway is held up by Crescent Formation basalts. The Crescent Formation crops out along the south side of the ridge on the north side of US 101. Its contact with the overlying Aldwell Formation approximately parallels the top of the ridge to about Mile 15.0

13.0 Granny's Café on left.

13.9 Cross Indian Creek. Lake Sutherland, to the west, drains into this creek.

14.5 Weathered exposures of Crescent basalt becoming less weathered between Mile 15.0 and 15.2. Lake Sutherland on left.

15.5 From this point to Mile 16.9 the road crosses over landslide debris that came from the ridge on the right (north). The geologic history of this landslide, that separated a single large lake into lakes Crescent and Sutherland, will be discussed at the Lake Crescent overlook stop.

17.0 Crescent basalt roadcuts are present along the highway from this point to Mile 21.4.

17.1 Boundary sign of Olympic National Park.

17.3 First view of Lake Crescent on right.

18.9 Lake Crescent overlook, Sledge Hammer Point, on right. This is **Stop 1-3** which we will make on the way back to Port Angeles.

20.6 Turnoff for Lake Crescent Lodge and Storm King Ranger Station.

21.4 Contact of Crescent Formation and Blue Mountain unit (Tbm). Roadcuts of Blue Mountain unit, which consists of sandstone and argillite grading to siltstone, are present along the highway from this point to Mile 27.4.

24.8 Turnoff for La Poel picnic site. Sign at entrance stating road is not suitable for bues, trailers, and RVs.

25.5 Pulloff on the right is the best location to stop on this narrow stretch of highway to look at a good exposure of Blue Mountain unit.

27.4 Fairholme Store on right. At this location the road is back on glacial deposits. As you drive up the long grade ahead Blue Mountain unit rocks crop out along the low ridge to the south. On the north side of this ridge micaceous sandstone and slate undifferentiated (Tnmu) of the Needles Gray-Wolf lithic assemblage are in fault contact (Hurricane Ridge fault) with the Blue Mountain unit. The highest terrain that can be seen farther south is held up by sandstones and conglomerates (Two) of the Western Olympic lithic assemblage which are in fault contact with the Needles Gray-Wolf rocks. To the north, basalts of the Crescent Formation (Tcb) extend up the entire slope to the ridge line.

29.3 Turn off on left to Sol Duc Hot Springs.

29.8 Crescent Formation basalts form hills on north side of the road. Hills on the south side are composed of Needles Gray-Wolf lithic assemblage.

34.0 Prominent ridge on right is held up by Crescent basalts.

35.5 Intersection with East Snider Road.

36.1 Bridge over Sol Duc River.

36.7 Turn off to on right to Olympic National Forest Klahowya campground.

37.0 Bridge over Sol Duc River.

37.7 From this point to Mile 39.0 outcrops of Crescent basalt are present along the highway.

39.2 Intersection on right with Wisen Creek Road.

39.5 Sign stating you are leaving Olympic National Forest.

40.5 Hills on right are held up by the Crescent Formation.

42.5 Bear Café on left.

44.4 Enter Sappho

44.8 Intersection of U.S. 101 and Washington 113 and turn off for Cape Flattery. Hill straight ahead (view to west) is composed of Western Olympic lithic assemblage (Two) sandstone and conglomerate.

45.5 Bridge over Sol Duc River.

46.2 Old schoolhouse on right. Built in 1916 for $6,000, it had the name Beaver School on a tin sign over the door. It functioned as schoolhouse from 1916 to 1968. The building then was remodeled for other uses. In 1989 the structure was restored to its original design. In November 1992 it was placed on the National Registry of Historic Places.

48.4 Beaver, Washington

52.5 Clallam County Sheriff Fork's Office on right. To left side of the road are hills of Western Olympic lithic assemblage (Two).

54.5 Bridge over the Sol Duc River.

55.7 Junction La Push Road, Washington Route 110.

56.4 Bridge over Calawah River.

56.4 City limits of Forks, Washington. Forks was named for the intersections of the Quillayute, Bogachiel, Calawah, and Sol Duc rivers. The town was incorporated in 1945. As you drive south through the town the hills on the left (east) are strata of the Western Olympic lithic assemblage (Two). This is the town author Stephenie Meyer chose for the location of her *Twilight* series of books dealing with the love life of a high school student and a vampire but throwing in some werewolves too. Forks is still sponsoring its annual Twilight Festival to attract tourists.

59.2 Southern city limits of Forks.

62.9 Turn off for Bogachiel State Park. Western Olympic lithic assemblage (Two) makes up hills in the distance, straight ahead.

63.3 Bridge over the Bogachiel River.

63.9 Hills straight ahead are sandstones, siltstones, and conglomerates of the Hoh lithic assemblage (Thsr, Thc, Thr, Thts, and Thsp).

70.0 Upper Hoh Road turn off on left to Olympic National Park Hoh Rain Forest.

71.6 Intersection of Oil City Road and turn off for Cottonwood Recreation Area. Oil City is on the north side of the Hoh River's mouth, about a third of a mile from the coast. In December 1911 the Olympic Oil Company filed a plat for Oil City, due to the presence of oil seeps. This area is one of two areas of highest concentrations of hydrocarbons on the coast (Snavely and Kvenvolden, 1989). Many plats were bought in the early 1900s on the hope for becoming rich; some plats were used as vacation sites. Oil City had a store and a cannery. The road from US 101 was built to the town in 1931-1932. The *Oil City News* published at least three issues that mostly promoted oil. Apparently the first oil wells were drilled in the area in the 1920s. There was lots of hype over the potential for oil. The August 3, 1931 issue of the Port Townsend Leader reported "Lots of oil signs. At the spudding in of the big 3,000 pound drill Saturday, it is reported that many signs of oil were evident." The Leslie Petroleum Company drilled Sims No. 1 which, according to the Leader, was producing 20 barrels of oil a day in 1931. Drilling continued for a few more years in the 1930s before the search for a good producing well was abandoned. The Jefferson oil seep near Oil City has a compound thought to be a marker for terrigenous plant input. (Kvenvolden et al., 1989). At **Stop 1-2** we will be discussing the occurrence of hydrocarbons on the Washington coast.

72.4 Narrow bridge (built 1931) over Hoh River.

73.1 Turn off on left onto Hoh Mainline for Hoh Clearwater State Forest.

73.3 Good view of Hoh River on right.

78.6 Turn off on left to Duncan Memorial Big Cedar on Nolan Creek Road. The tree is 178 feet tall and 19.4 feet in diameter. It was left standing at the request of loggers when its magnitude was recognized.

81.5 Junction with Lower Hoh Road to Hoh Tribal Center.

82.3 Entering Olympic National Park.

84.5 Turn off for Ruby Beach, **Stop 1-2.**

85.7 Overlook on right provides good view of Pacific Ocean.

86.8 Turn off for "Big Cedar Tree" on left. This tree is 130 feet high and estimated to be about 1,000 years old.

88.8 STOP 1-1. Beach 4. Turn right and drive into Beach 4 parking lot. There are restrooms here. Take the short trail down to the beach. Cross over wood bridge onto the Hoh Formation also called Hoh rock assemblage (Rau, 1975), Hoh lithic assemblage (Tabor and Cady, 1978a), and Coastal Olympic Structural Complex (Coastal OSC) by Brandon and Vance (1992).

The Hoh Formation outcrops at this stop consist primarily of lower Miocene (ca. 24 to 16 Ma) thick- to thin-bedded turbidite sandstone - also referred to as graywacke (an informal term for coarse-grained sandstone with poorly sorted subangular to angular quartz, feldspar, and rock fragments all mixed together in a clayey matrix), thin to medium bedded siltstone, and shale. At the end of the trail are steep, southeast-dipping sandstones. Graded beds show bedding in this outcrop is overturned. Ripple drift cross-laminations, groove, flute, and flame casts are present in thicker beds as well as crossbedding and channels. The numerous holes in the rock are thought to be pholad clam borings.

Work your way around the outcrop to the north side where you will see an angular unconformity (2.7 m altitude) of essentially horizontal Pleistocene and Holocene sediments on steeply-dipping Hoh turbidites. Thackery and Pazzaglia (1994) and Pazzaglia et al. (2003) interpret the unconformity, which can be traced 80 km along the coast, as a wave-cut surface that formed during the last major interglacial sea level highstand at 122 ka (oxygen isotope substage 5e). The gray sand and pebbly beach deposit directly above the unconformity is radiocarbon dead. The layered silt, clay, and peat with minor gravel overlying the beach sediments are the Brown's Point Formation (Thackery, 1998, Fig. 3, stratigraphic section b) which are thought to reflect fluvial, glaciofluvial, and marsh sedimentation (Pazzaglia et al., 2003). Radiocarbon dates from the Brown's Point Formation yield progressively younger ages from >49,000 at the bottom to 16,700±160 radiocarbon yr B.P. near the top (Heusser, 1972; Thackery, 1998). The sequence is capped by loess.

A short distance farther north the beds have been deformed into a series of folds (fold train). Structural relationships indicate boudinage developed early in thin sandstone beds followed by shortening and duplication of bedding along splay faults (Aldrich, 2014). The folds have northeast-striking axial planes, and hinge lines with an average orientation of 53^0 N. 54^0 E., are truncated at the base of the outcrop by a low angle thrust fault. Structural analysis of the fold train shows the strata have shortened ~40%. The geometry of the folds and other structural criteria indicate they formed by a combination of flexural slip and flexural flow (Aldrich, 2014). Northward directed compressive stress across the Olympic Peninsula (cf. Wells and Simpson, 2001) may have driven development of these structures. The youngest structures at this stop are a set of right-lateral eastnortheast trending strike-slip faults and a set of left-lateral northwest trending strike-slip faults. Several geologic criteria collectively indicate the two fault sets are coeval.

Go south on the beach to the large dome-shaped knob ~3/10 mile beyond the outcrop at the end of the trail. At this location the

122 ka wave-cut terrace lies below the surface and the Pleistocene beach deposits are considerably thicker than those above the unconformity north of the trail. Several northeast-striking and south dipping thrust faults have displaced the beach deposits and appear to have also displaced the overlying Brown's Point Formation.. Apparent offsets on the beach deposits are ~10 cm.

Return to the trail and take it back to the parking lot. Turn left out of the parking lot onto US 101 and drive north to **Stop 1-2**.

92.9 STOP 1-2. Ruby Beach. Take the trail from the parking lot to the beach. You should expect to have to cross a stream to get to the mélange outcrops. I recommend bringing knee-high rubber boots to ford the stream. Once you are on the beach go north.

At this stop the Coastal OSC, Hoh rock assemblage of Rau (1975), includes graywacke sandstone and mélange rocks of intensely sheared claystone and siltstone containing blocks of indurated siltstone and graywacke sandstone and altered volcanic rocks. The outcrops were mapped by Tabor and Cady (1978a) as the same Hoh lithic assemblage unit as at Beach 4, except for a small offshore outcrop of the Lyre Formation. The identification of this outcrop as Lyre Formation is at odds with the map by Rau (1975) that Tabor and Cady (1978a) list as their source for the geology of this area depicted on their map. Rau's map (1975) shows all of the rocks at Ruby Beach as Hoh Formation and mélange. The rocks composing this outcrop are, most likely, part of the Coastal OSC. The rocks at Ruby Beach don't look anything like those at Beach 4 and were more properly mapped by Rau, at a scale of 1:62,500, as mélange. The Tabor and Cady (1978a) map is half the scale (1:125,000) of Rau's map, justifying using the same unit designation as at Beach 4. The description of the unit, however, could have been improved by mentioning the presence of mélange. Snavely and Kvenvolden (1989) correlated the mélange here with mélange at Ozette Lake and, therefore, use the name Ozette mélange for the Ruby Beach unit.

Walk north on the beach past the first sea stack (of Hoh Formation) to the sea stacks and cliff of mélange. Several different origins for Hoh mélange have been proposed including gravity tectonism, shear zones between large structural blocks, and diapirism. Earlier studies by Rau (1973, 1975, 1979) and Rau and Grocock (1974) identified both shear-zone and diapiric mélanges in the Coastal OSC. Orange (1990) and Orange et al. (1993) have done detailed investigations of Hoh mélanges that demonstrate the complexity of geologic conditions under which they form. The mélange at Ruby Beach is a shear-zone mélange with numerous exotic clasts. The clasts range widely in size from small pebbles to blocks more than a meter across. The mélange is referred to as smell muds (Snavely and Kvenvolden, 1989) because of the petroliferous odor given off by fresh pieces of the rock (remember we are in Olympic National Park so we are not allowed to break open rock to create a fresh surface). Large eye-shaped, fault-bounded structural blocks and through-going high angle faults, well-exposed on the cliff face, accommodated tectonic shortening driven by the subducting Juan de Fuca plate.

Take the trail back to the parking lot. Turn left onto US 101 and retracing our route back. It is approximately 65 miles to **Stop 1-3**.

158.4 STOP 1-3. Lake Crescent Overlook, Sledge Hammer Point. There are several overlooks on this stretch of highway so be careful to not go past this stop. The Stop 1-3 overlook has several large trees and a couple of inclined, waist-high informative signs. Roadcut in Crescent Formation, Tcb unit of Tabor and Cady (1978a). Flows of black pillow basalt striking approximately east-west and dipping steeply (~85^0) north; dense to highly vesicular; contains microphenocrysts of clinopyroxene and calcic to soda plagioclase. A submarine flow at Crescent Lake, just below the contact with the overlying Aldwell Formation yielded an $^{40}Ar/^{39}Ar$ date of 52.9±4.6 Ma while the base of the submarine Crescent Formation flows on Hurricane Ridge Road yielded an $^{40}Ar/^{39}Ar$ date of 45.4±0.6 Ma. These two dates suggest the Crescent Formation, while mapped as a single unit between these two locations, had more than one eruptive center (Babcock et al., 1994). There is disagreement among investigators as to whether the chemistry of the basalt justifies separating the formation into lower and upper members. Glassley (1974) and Muller (1980) maintain that the chemistry points to two members – a lower mid-ocean ridge basalt (MORB) and upper oceanic island basalt (OIB) member. Cady (1975) and Babcock et al, 1994) argue there is no clear difference in chemistry between the upper and lower members. More work needs to be done to resolve this issue.

Lake Crescent and Lake Sutherland to the east were one large lake after the last of the Cordilleran glacial ice that occupied the valley melted, leaving a basin that drained into Indian Creek. A massive landslide, from the north side of the valley, occurred after the ice retreated splitting the basin into two lakes. The landslide raised the level of Lake Crescent (about 25 m (83 ft.) higher than Lake Sutherland) causing it to drain north out the Lyre River. Lake Sutherland still drains east through Indian Creek. Tabor (1975) speculates the landslide occurred shortly after the ice was gone in the early Holocene or late Pleistocene.

Interestingly the Quileute and Klallam tribes share a common legend on the origin of Lake Crescent. The legend tells of a battle between the Quileute and Klallam peoples that lasted several days. Mount Storm King became angry and took a great piece of rock from his crest, hurled it into the valley killing all who were fighting and separating the big lake (called Tsulh-mut by Klallam tribal members) into two lakes. While Mount Storm King on the south side of the valley is the source of the rocks in the legend and the actual source was the north side of the valley, the legend is consistent with the geologic process that divided the lake in two.

Wegmann et al. (2014) collected Chirp seismic imagery from the lake across the east-trending, oblique left-lateral Lake Creek-Boundary Creek fault (LCBCF) and two subaqueous mass wasting deposits. One of their seismic stratigraphic sections near the LCBCF reveals evidence for five post-glacial (≤ 14 ka) earthquakes on the fault. Beneath the youngest mass wasting deposit five distinct late Pleistocene-Holocene deformation events were identified with the total stratal offset decreasing up-section. Pringle et al. (2010) obtained ^{14}C ages of 4,450-4,350 calibrated ybp on submerged trees in the lake and suggested the earthquake that triggered the landslide that now separates Lake Crescent and Lake Sutherland may have occurred during that hundred year interval; however, they did not know if the trees are rooted. Wegmann et al. (2014) reported a rooted tree in the lake, at a depth of 18 m, yielded a ^{14}C age of 4,350±50 calibrated ybp and suggested the tree might have been submerged during the last earthquake-triggered landslides on the

Lake Crescent segment of the Lake Creek-Boundary Creek fault. The left- lateral displacement on the Lake Creek-Boundary Creek fault is consistent with the Pazzaglia et al. (2003) model for development of the Coast Range terrane horseshoe in which eastward movement of the OSC, bulldozing the rocks in its path, rotates rocks to the south clockwise and rocks to the north counterclockwise.

End Day 1. Return to Red Lion Hotel in Port Angeles.

DAY 2: SUNDAY, OCTOBER 4, 2015

Mile

0.0 Red Lion Hotel parking lot. Turn left from the parking lot onto North Lincoln Street. Drive south across East Front Street. For about the next 3.6 miles you will be driving over continental glacial deposits identified as Qc on Tabor and Cady's geologic map of the Olympic Peninsula (Tabor and Cady, 1978a).

0.1 Bear left into turn lane for left onto East First Street.

0.8 Turn right onto Race Street.

1.8 Enter Olympic National Park. Turn off on right for the main visitor center of the park.

1.9 Bear right onto Hurricane Ridge Road.

3.6 It is approximately here that you begin passing the first vegetation covered roadcuts of the Twin River Group (Ttrl).

5.7 This is where you first begin to see roadcuts not heavily covered with vegetation.

6.2 Overlook on left. Poor exposure of Crescent Formation (Tcb) basalts in the roadcut. Basalts are better exposed over next 0.2 miles.

6.9 Intersection with Lake Dawn Road.

7.1 Ranger station.

7.2 Turn off on left for Heart of the Hills Campground.

9.4 Large roadcut of Crescent basalt (Tcb).

10.6 Start of continuous roadcuts of Crescent Formation (Tcb) just south of the entrance of the first of three tunnels cut through the basalts.

10.9 Overlook on left. Crescent basalts (Tcb) in roadcut.

11.0 Entrance to first tunnel.

11.3 Exit third tunnel.

12.6 Pullout on left. This is **Stop 2-3** that we will make on the return trip as it is safer to enter and exit the road on the adjacent lane.

12.9 The Crescent basalts here have some of the better developed pillows that we will see.

13.6 Overlook on left.

15.7 Contact of Crescent Formation (Tbc) and Blue Mountain unit (Tbm).

16.8 Switchback Trail head.

17.9 Hurricane Ridge fault. This is **Stop 2-2** on the return trip.

19.2 Sharp curve in road. Roadcut of Needles Gray-Wolf lithic assemblage, Upper OSC, on right.

19.4 Roadcut of Upper OSC.

19.8 Hurricane Ridge visitors center. Continue straight ahead to Hurricane Hill road. NOTE: this is a narrow windy road with a 15 mph speed limit.

20.2 U curve! Keep as far right as possible when navigating this very sharp curve in the event you meet someone coming the other way.

21.1 STOP 2-1. Hurricane Hill Trail. Parking lot of Hurricane Hill trail Park and take the (paved) trail toward Hurricane Hill. This trail crosses three stratigraphic units of Tabor and Cady (1978a). Approximately the first 0.7 km (2000 ft) of the trail traverses the Tnm unit of the Needles Gray-Wolf lithic assemblage, which they describe as a micaceous sandstone, with less than 60% siltstone and slate. The angular, medium-grained, lithic to feldspathic sandstone is poorly sorted. Calcite and slate chips are common. It is thin to very thick bedded with small crossbeds, and rare graded beds, ripple marks, and load casts. Slate is micaceous and highly fissile; it grades to siltstone.

Past the trail head there's a very low topographic dip in it, a short distance farther on you will see the first outcrops of the Needles Gray-Wolf lithic assemblage. In this area it is thinly-to thick bedded graywacke sandstone, siltstone and slate. The most pervasive structure in these outcrops is a pencil cleavage, slivers of rock (pencil-like) formed by the intersections of two or more cleavages or, more typically, the intersection of cleavage (a planar fabric created by the rock tendency to split in a particular direction) and bedding. The pencil cleavage may reflect an intermediate stage in the development of slaty cleavage (a foliation defined by elongate domains of quartz or feldspar aggregates separated by anastomosing mica-rich laminae) and, therefore, occur only in weakly metamorphosed rocks, like those on this trail. Tabor and Cady (1978b) found pencil cleavage in the western and northeastern parts (where we are now) of the eastern core lying in bedding. In the central part of the core pencils generally do not lie in bedding but are formed by two cleavages and are perpendicular to fold axes. Because they found pencils to be the most consistent structural element in the core they used pencil orientations to divide it into two large structural domains that they subdivided into 19 subdomains. In the field the boundary between the two main domains, they called Domain East and Domain West, is identified by opposing dips and plunges. In Domain East the planar structures dip west to southwest and pencils plunge west. In Domain West planar structures dip east and northeast and pencils plunge east. The boundary which winds roughly north-northwest across the core, passing about 8 km east of Mount Olympus, separates the west verging structures in the Olympic Mountains from the east verging structures. Hurricane Hill Trail falls within Tabor and Cady's Domain East, Subdomain 1. Their contour diagrams of Subdomain 1 data show bedding mostly striking northwest and dipping steeply southwest, cleavage striking west-northwest and dipping steeply south-southwest, steeply plunging pencils trending south-southeast to south, and steep to moderately steep plunging fold axes plunging moderately to steeply northeast to northwest.

The first sign on the trail is titled "Folded Rock." (This sign is not always mounted. A steel plate, set in concrete, is located at the fold.) The fold has a chevron-like geometry. The axis of this well-developed fold trends northeast. Notice the steeply dipping fault that cuts across the axial plane, creating a small apparent offset of bedding. A short distance farther on the trail very thick graywacke sandstone beds are present. They are devoid of the cleavage that is so prominent in the thinly-bedded layers.

The map by Tabor and Cady (1978a) shows the Hurricane Ridge fault crossing the trail where its grade changes steeply up on the west side of the saddle located about 2,000 feet from the trail head. When you cross the fault the trail is on the Blue Mountain unit. Tabor and Cady describe it as sandstone and argillite (a compact rock derived from claystone, siltstone, or shale that has undergone a somewhat higher degree of induration but is clearly less laminated than shale and without its fissility, and that lacks the cleavage distinctive of slate) - very fine to medium-grained lithic sandstone, volcanic rich; fair to poorly sorted and angular with thin to thick beds.

The rocks in the area of the Hurricane Ridge fault are more highly deformed, probably reflecting a wide zone of deformation associated with displacement along the master fault. The map pattern of the fault (Tabor and Cady, 1978) shows it is nearly vertical here. Bedding orientations change significantly over short distances and there are a significant number of faults with widely varying orientations, many with low dips. The trail sign "Wind the Sculptor" is west of the fault on the Blue Mountain unit. Higher up (about 4,500 feet from the trail head) the trail crosses the contact with the Crescent Formation, which caps Hurricane Hill.

Return to parking lot and drive back to the Hurricane Ridge visitor center. We will have lunch there before going on to the next stop.

24.4 STOP 2-2. Hurricane Ridge Road (Mile Marker 15.9). This is an excellent location to see the Hurricane Ridge fault, the contact of the Blue Mountain unit and the Needles Gray-Wolf lithic assemblage of the core rocks, the same units we saw at Stop 1. The geologic map by Tabor and Cady (1978a) shows the road passes over the fault at about mile marker 16.0. The roadcuts west and east of this mile marker have steeply dipping beds dipping south and north, right side up and overturned, and highly disrupted by imbricate faults. Graded bedding and cross laminations are present in some of the thin greywacke sandstones. On the south side of the mile marker a large elongate (~ 1 meter) block of graywacke is surrounded by thin beds of slate, siltstone and sandstone, that is like the exotic blocks found in tectonic mélange. There are several well-developed faults in these exposures. The fault closest to mile marker 16 may be the master fault of the fault zone or what Tabor and Cady (1978b) call the zone of disruption. Bedding is nearly vertical on both sides of the fault. Drag on the beds flanking the fault and very small drag folds on the fault indicate the north side moved steeply up and west relative to the south side. Look for the tight isoclinal fold about 50 feet south of mile marker 16 and for steeply-dipping splay faults and beds sheared off by well-developed cleavage within the fault zone.

Slate present within the Needles Gray-Wolf lithic assemblage here, and at the previous stop, is the result of shale and mudstone being subducted into the accretionary wedge, subjecting it to increased temperatures and pressures. Tabor and

Cady (1978b) found a general increase in the metamorphic grade from west to east based on the presence of various index minerals in samples they collected in the central and eastern Olympic Mountains and other workers (Stewart, 1974, in the western part of the Olympic Structural Complex and Hawkins, 1967, in the Mount Olympus area). Brandon and Calderwood (1990) concur with Tabor and Cady's metamorphic zonation. Based on fission-track dates for sandstones from the eastern zone they place its temperatures between 100 ± 10^0 and 200 ± 50^0 C, the blocking temperatures for apatite (a mineral consisting of some combination of fluorine, chlorine, hydroxyl or carbonate) and zircon. The slate in the Needles Gray-Wolf rocks was, very likely, formed within this range of temperatures. In the area of Mount Olympus, the topographically highest part of the mountains, Brandon and Calderwood (1990) identified an adjacent zone with an assemblage of minerals that indicate higher temperature (~190 ^{0}C) and pressure (~300 MPa or 3000 kg/cm^2) conditions. Assuming the rocks have an overall density of 2,700 kg/m^{3,} they calculated the rocks in this zone were subducted to a depth of 11 km (6.8 miles) before they began their upward ascent to the surface.

29.7 STOP 2-3. Hurricane Ridge Road (Mile Marker 10.7). This roadcut in the Crescent Formation is dominated by a volcanic breccia and pillow basalt. It provides an excellent display of the faulting experienced by the basalt, juxtaposing different rock types. Note the presence of both moderately dipping and steeply dipping faults and what appears to be large conjugate shears filled with secondary minerals.

There are two basic models for the origin of the Coast Range basalts (Crescent and Siletz formations), a seamount/plume model (a spreading ridge reorganization model is a variation of this) and a marginal basin model. In the seamount/plume model a seamount chain, that formed over a mantle plume, was accreted to the continent (Simpson and Cox, 1977; Duncan, 1982). Its variation, the spreading ridge reorganization model, involves reorganizations of spreading on the Kula-Farallon ridge between 61 and 48 Ma resulted in Coast Range basalts erupting as seamounts and volcanic ridges along leaky transform faults and fractures during changes in spreading directions. The marginal basin rift model involves the outpouring of oceanic basalt during rifting of the continental margin as a result of highly oblique motion of the Kula and Farallon plates relative to the North American plate (Wells et al., 1984; Babcock et al., 1992; Snavely and Wells, 1996). A study by Chan et al. (2012) of Pb isotopes in the 42–37 Ma Grays River volcanics indicates these younger Coast Range basalts at least partly shared a common mantle source with the older (ca. 56 – 45 Ma) Crescent Formation basalts. Tepper (written communication, 2013) suggested the Crescent basalts may reflect a combined seamount/plume and marginal basin model, like the model proposed by Chan et al. (2012) for the Grays River volcanics. In their model the Grays River volcanics (MORB) erupted in a marginal basin, formed in response to oblique subduction of the Kula-Farallon spreading ridge, while oceanic island basalts (OIB) from a mantle plume fed into it.

Schmandt and Humpreys (2011) place the accretion (of the "Siletzia microplate") at ~55 Ma in the early Eocene while Brandon and Vance (1992) favor a younger age of 42-24 Ma for accretion. Brandon and Vance (1992) base the time of accretion primarily on the movement history of the Leech River fault on Vancouver Island, the only place where the continental suture boundary of the Coast Range terrane is exposed.

End Day 2. Return to Red Lion Hotel in Port Angeles.

REFERENCES CITED

Aldrich, M. J., 2014, Structural accommodation of north-directed compression in Cascadia accretionary wedge Miocene turbidites, coastal Olympic Mountains, Washington: Geological Society of America Annual Meeting, Abstracts with Programs, v. 46, no. 6, p. 35.

Applegate, J. D. R. and Brandon, M. T., 1989, An upper plate origin for basalt blocks in the Cenozoic subduction complex of the Olympic Mountains, northwest Washington State: Geological Society of America Abstracts with Programs, v. 21, no. 5, p. 51.

Armentrout, J. M., 1987, Cenozoic stratigraphy, unconformity-bounded sequences, and tectonic history of southwestern Washington: Washington Division of Geology and Earth Resources Bulletin 77, p. 291-320.

Armstrong, R. L. and Ward, P., 1991, Evolving geographic patterns of Cenozoic magmatism in the North America Cordillera: The temporal and spatial association of magmatism and metamorphic core complexes: Journal of Geophysical Research, v. 96, p. 13,201-13,224.

Arnold, R., 1906, A geological reconnaissance of the coast of the Olympic Peninsula, Washington: Geological Society of America Bulletin, v. 17, p. 451-468.

Atwater, B. F., Satoko, M-R., Kenji, S., Yoshinobu, T., Kazue, U., and Yamaguchi, D. K., 2005, *The Orphan Tsunami of 1700*: U. S. Geological Survey and University of Washington Press, 133p.

Babcock, R. S., Burmester, R. F., Engebretson, D. C., Warnock, A, and Clark, K. P., 1992, A rifted margin origin for the Crescent basalts and related rocks, in the northern Coast Range volcanic province, Washington and British Columbia: Journal of Geophysical Research, v. 97, no. B5, p. 6,799-6,821.

Babcock, R. S., Suczek, C. A., and Engebretson, D. C., 1994, The Crescent “terrane”, Olympic Peninsula and southern Vancouver Island: Washington Division of Natural Resources Bulletin 80, p. 141-157.

Bates, R. G., Beck, M. L., and Burmester, R. F., 1981, Tectonic rotations in the Cascade Range of southern Washington: Geology, v. 9, p. 184-189.

Batt, G. E., Brandon, M. T., Farley, K. A., Roden-Tice, M., 2001, Tectonic synthesis of the Olympic Mountains segment of the Cascadia wedge, using two-dimensional thermal and kinematic modeling of thermochronological ages: Journal of Geophysical Research, v. 106, no. B11, p. 26,731-26,746.

Beck, M. E. and Engebretson, D. C., 1982, Paleomagnetism of small basalt exposures in the west Puget Sound area, Washington, and speculations on the accretionary origin of the Olympic Mountains: Journal of Geophysical Research, v. 87, no. B5, p. 3755-3760.

Brandon, M. T., 2004, The Cascadia subduction wedge: the role of accretion, uplift, and erosion, *in* van der Pluijm, B. A. and Marshak, S., *Earth Structure, An Introduction to Structural Geology and Tectonics*, 2nd edition, WCB/McGraw Hill Press, p. 566-574.

Brandon, M. T. and Calderwood, A. R., 1990, High-pressure metamorphism and uplift of the Olympic subduction complex: Geology, v. 18, p. 1252-1255.

Brandon, M. T. and Vance, J. A., 1992, Tectonic evolution of the Cenozoic Olympic Subduction Complex, Washington State, as deduced from fission track ages for detrital zircons : American Journal of Science, v. 292, p. 565-636.

Brandon, M.T., Roden-Tice, M. K., and Garver, J. I., 1998, Late Cenozoic exhumation of the Cascadia accretionary wedge in the Olympic Mountains, northwest Washington State: Geological Society of America Bulletin, v. 110, no. 8, p. 985-1009.

Brown, E. H. and Dragovich, J. D., 2003, Tectonic elements and evolution of northwest Washington: Washington State Department of Natural Resources, Division of Geology and Earth Resources, Geologic map GM-52, scale 1:625,000.

Brown, R. D., Jr., Snavely, P. D., Jr., and Gower, H. D., 1956, Lyre Formation (redefinition), northern part of Olympic Peninsula, Washington: American Association of Petroleum Geologists Bulletin, v. 40, p. 94-107.

Cady, W. M., 1975, Tectonic setting of the Tertiary volcanic rocks of the Olympic Peninsula, Washington: U. S. Geological Survey Journal of Research, v. 3, no. 5, p. 573-582.

Cady, W. M., Sorense, M. L., and MacLoed, N. S., 1972a, Geologic map of the Brothers quadrangle, Jefferson, Mason, and Kitsap counties, Washington: U. S. Geological Survey Geologic Quadrangle Map GQ-969, scale: 1:62,500.

Cady, W, M., Tabor, R. W., MacLoed, N. S., and Sorensen, M. L., 1972b, Geologic map of the Tyler Creek quadrangle, Clallam and Jefferson counties, Washington: U. S. Geological Survey Geologic Quadrangle Map, GQ-970, scale 1:62,500.

Chan, C. F., Tepper, J. H., and Nelson, B. K., 2012, Petrology of the Grays River volvanics, southwest Washington: Plume-influenced slab window magmatism in the Cascadia forearc: Geological Society of America Bulletin, v. 124, n. 7-8, p. 1324-1338.

Clowes, R. M., Brandon, M. T., Green, A. G., Yorath, C. J., Brown, A. Southerland, Kanasewich, E. R., and Spencer, C., 1987, LITHOPROBE – southern Vancouver Island: Cenozoic subduction complex imaged by deep seismic reflections: Canadian Journal of Earth Sciences, v. 24, p. 31-51.

Crosson, R. S. and Owens, T. J., 1987, Slab geometry of the Cascadia subduction zone beneath Washington from earthquake hypocenters and teleseismic converted waves: Recent accretion and deformation: Geophysical Research Letters, v. 14, p. 824-827.

Davis, D., Suppe, J., and Dahlen, F. A., 1983, Mechanics of fold-and-thrust belts and accretionary wedges: Journal of Geophysical Research, v. 88, No. B2, p. 1153-1172.

Davis, E. E. and Hyndman, R. D., 1989, Accretion and recent deformation of sediment along the northern Cascadia subduction zone: Geological. Society of America Bulletin, v. 101, p. 1465-1480.

DeMets, G. and Dixon, T., 1999, New kinematic models for Pacific-North America motion from 3 Ma to present; I, Evidence for steady motion and biases in the NUVEL-1A model: Geophysical Research Letters, v. 26, n. 13, p. 1921-1924.

Dixon, T. H., Miller, M., Farina, F., Wang, H., and Johnson, D., 2000, Continuous GPS motion of the Sierra Nevada block and some tectonic implications for the Basin and Range province, North American Cordillera: Tectonics, v. 19, p. 1-24.

Duncan, R. A., 1982, A captured island chain in the Coast Range of Oregon and Washington: Journal of Geophysical Research, v. 87, no. B13, p. 10,827-10,837.

Einarsen, J. M., 1987, The petrology and tectonic significance of the Blue Mountain unit, Olympic Peninsula, Washington: Western Washington University MS thesis, 59p.

England, P., and Wells, R. E., 1991, Neogene rotations and quasicontinuous deformation of the Pacific Northwest continental margin: Geology, v. 19, no. 10, p. 978-981.

Fairchild, L. H. and Cowan, D. S., 1982, Structure, petrology and trctonic history of the Leech River complex northwest of Victoria, Vancouver Island: Canadian Journal of Earth Sciences, v. 19, p. 1817-1835.

Garver, J. I. and Brandon, M. T., 1994, Erosional denudation of the British Columbia Coast Ranges as determined from fission-track ages of detrital zircons from the Tofino basin, Olympic Peninsula, Washington: Geological Society of America Bulletin, v, 106, p. 1398-1412.

Glassley, W. E., 1974, Geochemistry and tectonics in the Crescent volcanic rocks, Olympic Peninsula, Washington: Geological Society of America Bulletin, v. 85, p.785-794.

Heusser, C. J., 1972, Palynology and paleogeographical evidence of a late Pleistocene refugium near Kalaloch, Washington: Quaternary research, v. 2, p. 180-201.

Heller, P. L., Tabor, R. W., and Suczek, C. A., 1987, Paleographic evolution of the United States Pacific Northwest during Paleogene time: Canadian Journal of Earth Science, v. 24, p. 1652-1667.

Ingle, J. C., Jr., 1980, Cenozoic paleobathymetry and depositional history of selected sequences within the southern California Continental Borderland: Cushman Foundation for Foraminiferal Research Special Publication 19, p. 163-195.

Johnson, S. Y., Potter, C. J., Armentrout, J. M., Miller, J. J., Finn, C., and Weaver, C. S., 1996, The southern Whidbey island fault, an active structure in the

Puget Lowland, Washington: Geological Society of America Bulletin, v. 108, p. 334-354.

Johnson, S. Y., Blakely, R. J., Stephensen, W. J., Dadisman, S. V., and Fisher, M. A., 2004, Active shortening of the Cascadia forearc and implications for seismic hazards in the Puget Lowland: Tectonics, v. 23, TC1011, 27p.

Leonard, L. J., Currie, C. A., Mazzotti, S., and Hyndman, R., 2010, Rupture area and displacement of past great Cascadia earthquakes from coastal coseismic subsidence: Geological Society of America, v. 122, n0. 11, p. 2,079-2,096.

Magee, M. E. and Zoback, M. L., 1992, Wellbore breakout analysis for determining tectonic stress orientations in Washington State: U. S. Geological Survey Open-File report 92-715, 56p.

Massey, N. W. D., 1986, Metchosin igneous complex, southern Vancouver Island: Ophiolite stratigraphy developed in an emergent island setting: Geology, v. 14, no. 7, p. 602-605.

McCaffrey, R. and Goldfinger, C., 1995, Forearc deformation and great subduction earthquakes: Implications for Cascadia offshore earthquake potential: Science, v. 267, p. 856-859.

McCaffrey, R., Qamar, A. I., King, R. W., Wells, R., Khazaradze, G., Williams, C. A., Stevens, C. W., Vollick, J. J., and Zwick, P. C., 2007, Fault locking, block rotation and crustal deformation in the Pacific Northwest: Geophysical Journal International, v. 169, p. 1,315-1,340.

McCrory, P. A., 1996, Tectonic model explaining divergent contraction directions along the Cascadia subduction margin, Washington: Geology, v. 24, p. 929-932.

Muller, J. E., 1980, Chemistry and origin of the Eocene Metchosin volcanics, Vancouver Island, British Columbia: Canadian Journal of Earth Sciences, v. 17, p. 199-209.

Orange, D. L., 1990, Criteria helpful in recognizing shear-zone and diapiric mélanges: examples from the Hoh accretionary complex, Olympic Peninsula, Washington: Geological Society of America Bulletin, v. 102, p. 935-951.

Orange, D. L., Geddes, D. S., and Moore, J. C., 1993, Structural and fluid evolution of a young accretionary complex: The Hoh rock assemblage of the western Olympic Peninsula, Washington: Geological Society of America Bulletin, v. 105, p. 1053-1075.

Orange, D. L. and Campbell, K. A., 1997, Modern and ancient cold seeps on the Pacific Coast – Monterey Bay, California, and offshore Oregon as modern-day analogs to the Hoh accretionary complex and Quinault Formation, Washington: Washington Geology, v. 25, no 4, p. 3-13.

Pazzaglia, F. J. and Brandon, M. T., 2001, A fluvial record of long-term steady-state uplift and erosion across the Cascadia forearc high, western Washington state: American Journal of Science, v. 301, p. 385-431.

Pazzaglia, F. J., Thackery, G. D., Brandon, M. T., Wegmann, K. W., Gosse, J., McDonald, E., Garcia A. F., Prothero, D., 2003, Tectonic geomorphology and the record of Quaternary plate boundary deformation in the Olympic Mountains, *in* Swanson, T. W., ed., Western Cordillera and adjacent areas: Geological Society of America Field Guide 4, p. 37-67.

Pringle, P., Wegmann, K. Pontbriand, D., Walker, W. W., 2010, Some recent discoveries pertaining to subfossil forests in the Pacific northwest – examples from Lake Crescent and Lower Dry Bed Lakes, Olympic Mountains, Washington: Northwest Scientific Association 82nd Annual Meeting, Cascadia Prairie-Oak Partnership Ecoregional Conference, with Northwest Lichenologists, p. 77-78.

Prothero, D. R. and Nesbitt, E. A., 2008, Paleomagnetism and tectonic rotation of the Restoration Point Member of the Blakely Formation (type Blakely stage), Bainbridge Island, Washington, and the Pacific Coast Oligocene-Miocene boundary: *in* Lucas et al. eds, Neogene Mammals, New Mexico Museum of Natural History and Science Bulletin 44, p. 315-322.

Prothero, D. R., Draus, E., Cockburn, T. C., Nesbitt, E. A., 2008, Paleomagnetism and counterclockwise tectonic rotation of upper Oligocene Sooke Formation, southern Vancouver Island, British Columbia: Canadian Journal of Earth Sciences, v. 45, no. 4, p. 499-507.

Prothero, D. R., Draus, E., and Burns, C., 2009, Magnetostratigraphy and tectonic rotation of the Eocene-Oligocene Makah and Hoko River formations, northwest Washington: International Journal of Geophysics, v. 2009, p. 1-15.

Pyle, D. G., Duncan, R., Wells, R. E., Graham, D. W., Harrison, B., and Hanan, B., 2009, Siletzia: An oceanic large igneous province in the Pacific Northwest: Geological Society of America Annual Meeting, Abstracts with Programs, v. 41, no. 7, p. 369.

Rau, W. W., 1964, Foraminifera from the northern Olympic Peninsula, Washington: U. S. Geological Survey Professional Paper 374-G, 33p. 7 pl.

Rau, W. W., 1973, Geology of the Washington coast between Point Grenville and the Hoh River: Washington Department of Natural Resources, Geology and Earth Resources Division, Bulletin 66, 58p.

Rau, W. W., 1975, Geologic map of the Destruction Island and Toholah quadrangles, Washington: Washington Department of Natural Resources, Division of Geology and Earth Resources, GM-13, scale: 1:62,500.

Rau, W. W., 1979, Geologic map in the vicinity of the lower Bogachiel and Hoh river valleys, and the Washington coast: Washington Department of Natural Resources, Division of Geology and Earth Resources, GM-2A, scale: 1:62,500.

Rau, W. W., 1981, Pacific Northwest Tertiary benthic foraminifera biostratigraphic framework – An overview. *In* Armentrout, J. M., editor, Pacific Northwest Cenozoic biostratigraphy: Geological Society of America Special Paper 184, p. 67-84.

Rau, W. W. and Grocock, G. R., 1974, Piercement structure outcrops along the Washington coast: Washington

Department of Natural Resources, Division of Geology and Earth Resources, Information Circular 51, 7p.

Salvador, A., editor, 1994, International Stratigraphic Guide – A guide to stratigraphic classification, terminology, and procedure: International Subcommission on Stratigraphic Classification, 2nd Edition: The International Union of Geological Sciences and Geological Society of America: 214p.

Schmandt, B. and Humphreys, E., 2011, Seismically imaged relict slab from the 55 Ma Silitizia accretion to the northwestern United States: Geology, v. 39, no. 2, p. 175-178.

Sigloch, K. and Mihalynuk, M. G., 2013, Intra-oceanic subduction shaped the assembly of Cordilleran North America: Nature, v. 496, p. 50-56.

Simpson, R. W. and Cox, A., 1977, Paleomagnetic evidence for rotation of the Oregon Coast Range: Geology, v. 5, p. 585-589.

Snavely, P. D., Jr., 1983, Peripheral rocks – Tertiary geology of the northwestern part of the Earth Science Series, Olympic Peninsula, Washington, *in* Muller, J. E., Snaveley, P. D., Jr., and Tabor, R. W., eds., The Tertiary Olympic terrane, southwest Vancouver Island and northwest Washington: Geological Association of Canada, Mineralogical Association of Canada, Canadian Geophysical Union Field trip Guidebook, Trip 12, p. 6-31.

Snavely, P. D., Jr., 1987, Tertiary geologic framework, neotectonics, and petroleum potential of the Oregon-Washington continental margin, *in* D. W. Scholl, A. Grantz, and J. D. Vedder, eds., *Geology and Resource Potential of the Continental Margin of Western North America and Adjacent Ocean Basins – Beaufort Sea to Baja California*, Circum-Pacific Council on Energy and Mineral Resources, Houston, Texas: Earth Science Series, v. 6, p. 305-335.

Snavely, P. D., Jr. and Kvenvolden, K. A., 1989, Geology and hydrocarbon potential, *in* Preliminary evaluation of the petroleum potential of the Tertiary accretionary terrane, west side of the Olympic Peninsula, Washington: U.S. Geological Survey Bulletin 1892, p. 1-17.

Snavely, P. D., Jr., and Wagner, H. C., 1963, Tertiary geologic history of western Oregon and Washington: Washington Division of Mines and Geological Report of Investigations 22, 25p.

Snavely, P. D., Jr. and Wells, R. E., 1996, Cenozoic evolution of the continental margin of Oregon and Washington, *in* Rogers, A. M., Walsh, T. J., Kockelman, W. J., and Priest, G. R., eds., Assessing earthquake hazards and reducing risk in the Pacific Northwest: U. S. Geological Survey Professional Paper 1560, p. 161-182.

Snavely, P. D., Jr., MacLeod, N. S., Niem, A. R., and Minasian, D. L., 1986a, Geologic map of Cape Flattery, northwestern Olympic Peninsula, Washington: U. S. Geological Survey Open-File Report 86-344B, scale: 1:48,000.

Snavely, P. D., Jr., Niem, A. R., and MacLeod, N. S., 1989, Geology of the coastal area between Cape Flattery and

Cape Alava, northwest Washington: U. S. Geological Survey Open-File Report 89-141, scale 1:24,000.

Snavely, P. D., Jr., Niem, A. R., Pearl, J. E., 1978, Twin River Group (upper Eocene to lower Miocene) defined to include the Hoko River, Makah, and Pysht Formations, Clallam County, Washington: U. S. Geological Survey Bulletin, v. 1457-A, p. 111-119.

Snavely, P. D., Jr., Rau, W. W. and Hafley, J. D., 1986b, Tertiary foraminiferal localities in the Cape Flattery area, northwestern Olympic Peninsula, Washington: U. S. Geological Survey Open-File Report 86-344a, 18p.

Snavely, P. D., Jr., Neim, A. R., MacLeod, N. S., Pearl, J. E., and Rau, W. W., 1980, Makah Formation – a deep marginal basin sequence of late Eocene and Oligocene age in the northwestern Olympic Peninsula, Washington: U. S. Geological Survey Professional Paper 1162, 38p.

Squires, R. L., Goedert, J. L., and Kaler, K. L., 1992, Paleontology and stratigraphy of Eocene rocks at Pulali Point, Jefferson County, eastern Olympic Peninsula, Washington: Washington Division of Geology and Natural Resources Report of Investigations 31, 27p.

Stewart, R. J., 1974, Zeolite facies metamorphism of sandstone in the western Olympic Peninsula, Washington: Geological Society of America Bulletin, v. 85, p. 1139-1142.

Stewart, R. J. and Brandon, M. T., 2004, Detrital-zircon fission-track ages for the "Hoh Formation": Implications for late Cenozoic evolution of the Cascadia subduction wedge: Geological Society of America Bulletin, v. 116, no. 1/2, p. 60-75.

Suczek, C., Babcock, R., and Engebretson, D. C., 1994, Tectonostratigraphy of the Crescent terrane and related rocks, Olympic Peninsula, Washington, *in* Swanson, D., and Haugerud, R., eds., Geological field trips in the Pacific Northwest: 1994 Geological Society of America Annual Meeting, p. 1H1—1H11.

Tabor, R. W., 1975, *Geology of Olympic National Park*: Seattle, University of Washington Press, 144p.

Tabor, R.W. and Cady, W. H., 1978a, Geologic map of the Olympic Peninsula: U. S. Geological Survey Map I-994, scale 1:125,000, 2 sheets.

Tabor, R. W. and Cady, W. H., 1978b, The structure of the Olympic Mountains, Washington – Analysis of a subduction zone: U. S. Geological Survey Professional Paper 1033, 38p.

Tabor, R. W., Yeats, R. S., and Sorensen, M. L., 1972, Geologic Map of the Mount Angeles quadrangle, Clallam and Jefferson counties, Washington: U. SD. Geological Survey Geologic Quadrangle Map GQ-958, scale 1:62,500.

Thackery, G. D., 1998, Convergent –margin deformation of Pleistocene strata on the Olympic coast of Washington, USA, *in* Stewart, I. S. and Vita-Finzi, C., eds., Coastal tectonics: Geological Society [London] Special Publication 146, p. 199-211.

Thackery, G. D. and Pazzaglia, F. J., 1994, Quaternary stratigraphy, tectonic geomorphology, and fluvial evolution of the western Olympic Peninsula, Washington: *in* Swanson, D. A. and Haugerud, R. A., eds., Geologic Field trips in the Pacific Northwest, Geological Society of America Annual Meeting, p. 2A-1-2A-9.

Wang, K., 1996, Simplified analysis of horizontal stresses in a buttressed forearc sliver at an oblique subduction angle: Geophysical Research Letters, v. 23, no. 16, p. 2,021-2,024.

Warnock, A. C., Burmeister, R. E., and Engebretson, D. C., 1993, Paleomagnetism and tectonics of the Crescent Formation, northern Olympic Mountains, Washington: Journal of Geophysical Research, v. 98, no. B7, p. 11,729-11,741.

Weaver, C. E., 1916, The Tertiary formations of western Washington: Washington Geological Survey Bulletin 13, 327p.

Weaver, C. E. 1937, Tertiary stratigraphy of western Washington and northeastern Oregon: Washington University Publications in Geology: v. 4, 266p.

Wegmann, K. W., Bohnenstiehl, D. R., Leithold, E. L., and Pringle, P. T., 2014, Earthquakes, mass wasting, and fish from northern Cascadia: Post-glacial rupture of Lake Creek-Boundary Creek fault revealed by Chirp seismic investigations at Lake Crescent, Washington: Geological Society of America Annual Meeting, Abstracts with Programs, v. 46. no. 6, p. 779.

Wells, R. E., 1990, Paleomagnetic rotations and regional tectonics of Cascade arc, Washington, Oregon, and California: Journal of Geophysical Research, v. 95, p. 19,409-19,418.

Wells, R. E., 2007, Reconsidering the origin and emplacement of Siletzia: Geological Society of America Cordilleran Section Annual Meeting, Abstracts with Programs, v. 39, no. 4, p. 19.

Wells, R. E. and Coe, R. S.,1985, Paleomagnetism and geology of Eocene volcanic rocks of southwest Washington, Implications for mechanisms of tectonic rotations: Journal of Geophysical Research, v. 90, no. B2, p. 1925-1947.

Wells, R. E. and Simpson, R. W., 2001, Northward migration of the Cascadia forearc in the northwestern U. S. and implications for subduction deformation: Earth, Planets, and Space, v. 53, p. 275-283.

Wells, R.E., Weaver, C. S., and Blakely, R. J., 1998, Fore-arc migration in Cascadia and its neotectonic significance: Geology, v. 26, p. 759-762.

Wells, R. E., Engebretson, D. C., Snaveley, P. D., Jr., and Coe, R. S., 1984, Cenozoic plate motions and the volcano-tectonic evolution of western Oregon and Washington: Tectonics, v. 3, p. 275-294.

Willett, S. D. and Brandon, M. T., 2002, On steady states in mountains belts: Geology, v. 30, n. 2, p. 175-178.

Willett, S., Beaumont, C., and Fullsack, P.,1993, Mechanical model for the tectonics of doubly vergent compressional orogens: Geology, V. 21, p. 371-374.

Wright, N. M., Müller, R. D., Seton, M., and Williams, S. E., 2015, Revision of Paleogene plate motions in the Pacific and implications for the Hawaiian-Emperor bend: Geology, v. 43, no. 5, p. 455-458.