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FIELD TRIP GUIDEBOOK #047

SOME GEOLOGIC HIGHLIGHTS OF WHIDBEY AND GUEMES ISLANDS IN THE PUGET LOWLAND OF WASHINGTON

April 24 – 26, 2015

Eric S. Cheney, Professor Emeritus
University of Washington Earth and Space Sciences Department

Darrel S. Cowan, Professor
University of Washington Earth and Space Sciences Department

Steven R. Grupp, Professor
Everett Community College Geology Department

NWGS Field Trip Guidebook Series

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Field Trip Guidebook #047

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Eric S. Cheney, Professor Emeritus, University of Washington Earth and Space Sciences Depart.
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I. ABSTRACT

by Eric S. Cheney

With the exception of a few outcrops of bedrock north of the Devils Mountain fault on its northern end, Whidbey Island consists entirely of Pleistocene sediments. Deposits of three glaciations occur: Double Bluff (125 to 185 ka), Possession (60 to 80 ka), and Fraser (14 to 18 ka). The Puget lobe of the Fraser Glaciation deposited Vashon till between 18 and 14 ka. At about 14 ka the Puget lobe collapsed, which initiated deposition of glaciomarine drift of the Everson Interstade. The Partridge Gravel near Coupeville is generally believed to be a glaciomarine delta that was built at an elevation of 67 m by a lobe of ice in Penn Cove. The gravel is mantled with Everson glaciomarine drift containing shells dated at 11,500 ^{14}C yrs B.P.

The pre-Fraser stratigraphic units are locally well exposed in the coastal bluffs. Inland portions of the island are mostly blanketed by Vashon till of the Fraser glaciation, the advance outwash that accompanied it, and Everson glaciomarine drift. The coastal bluffs are ideal places to observe sedimentary textures and structures, stratigraphy, and shoreline processes. The bluffs also are sites of prehistoric and present landslides.

In the subsurface, the westerly trending Devils Mountain fault separates pre-Tertiary crystalline rocks of the San Juan Islands and the Cascade Range on the north from Tertiary strata of the Everett structural basin on the south. The southern Whidbey Island fault zone (SWIFZ) strikes northwesterly across the southern part of the island. In the subsurface SWIFZ separates Paleocene basaltic rocks of the Olympic Peninsula and southernmost Vancouver Island on the southwest from Eocene to Miocene clastic strata of the Everett basin on the northeast.

SWIFZ is on the northeastern limb of the Kingston arch, which is the anticline between the synclinal Seattle and Everett basins. This anticline causes the pre-Fraser Pleistocene sediments to crop out on Whidbey Island.

II. ABOUT THIS GUIDEBOOK AND TRIP

by Eric S. Cheney and Steven R. Grupp

The purpose of this two-day fieldtrip is to examine some of the Pleistocene geology of Guemes and Whidbey islands and the Mesozoic bedrock of Fidalgo Island. Because a recent field guide for Fidalgo Island exists (Cheney, 2013), we use it without modification. Thus, this present guide only covers aspects of the

Pleistocene geology of Guemes and Whidbey islands.

This is a multi-authored guidebook. The authorship of individual sections of the guidebook is noted. Authors are neither responsible for, nor necessarily agree with, sections attributed to other authors.

The timing of some stops is constrained by tidal conditions. Thus, on Day One we will do the stops in the Fidalgo Island guide (Cheney, 2013) in the following order: 2, 3, 5, 6, 7, and 11. After Stop 11 we will take the ferry to Guemes Island for the stop in this guidebook. Upon returning from Guemes Island, we will revert to the Fidalgo Island guidebook and finish Day One with Stops 9, 10, and, time permitting Stop 8. On Day Two we will do Stop 4 of the Fidalgo Island guidebook before departing for the five or six stops on Whidbey Island in this guidebook (**Fig 1**).

Time may not permit the long walk involved in the sixth stop, Double Bluff. Therefore, that stop is optional.

The road logs of this guide and the Fidalgo Island guide are, where possible, keyed to mile posts (MP) on the highways.

To return to the mainland at the end of the second day, we will take the Mukilteo ferry from Clinton near the southern end of Whidbey Island.

The itineraries of both days are reasonably ambitious. Therefore, we ask that participants be in the vehicles at 8:02 AM each day, ready to depart. Bring all necessary gear for the entire day, including lunch and drink.

In the interests of time and safety, we will neither scramble up large outcrops nor venture into tidal waters or private property. Some stops are on busy highways; participants must be keen of eye and judgment and, perhaps, fleet of foot.

The usual NWGS policy on the consumption of alcoholic beverages applies during this trip.

III. REGIONAL GEOLOGIC SETTING

by Eric S. Cheney

Regional Folds

The Puget lowland is part of the Neogene egg-crate of the Pacific Northwest. The egg-crate is caused by the intersection of northerly and northwesterly trending folds (Cheney, 2015). The Puget Lowland occupies the Willamette-Puget-Fraser (WPF) syncline between the northerly tending Cascade Range anticline and the northerly trending Coast Ranges anticline. Plate tectonicists prefer to call the synclinal Puget Lowland the fore-arc basin of the Cascade magmatic arc.

The Cascade Range anticline folds strata as young as the Miocene Columbia River Basalt Group (CRBG). More especially, near Ellensburg the uplift of the anticline is marked by the composition and dip of the 4 Ma Thorp Gravel (Cheney, 2014, 2015).

A number of northwesterly trending folds segment the WPF syncline into smaller structural (not depositional) basins, such as the Tacoma, Seattle, Everett, and Whatcom basins (Cheney, 2015). One indication of the youthfulness of this

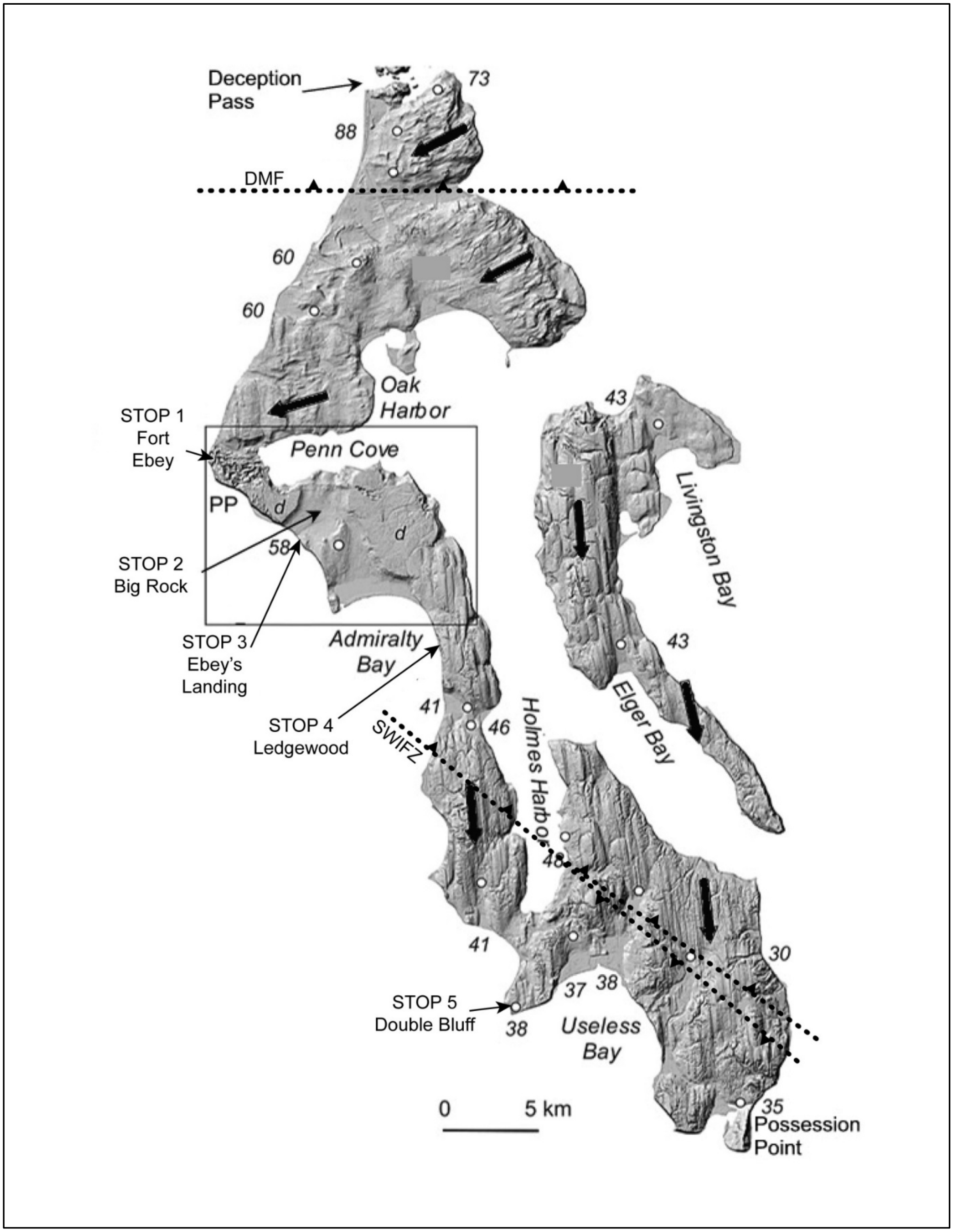
Deception Pass
73
88
DMF
60
60
Oak Harbor
STOP 1 Fort Ebey
Penn Cove
PP
d
d
58
STOP 2 Big Rock
STOP 3 Ebey's Landing
Admiralty Bay
STOP 4 Ledgewood
43
Livingston Bay
43
Elger Bay
41
46
SWIFZ
Holmes Harbor
48
41
30
37
38
STOP 5 Double Bluff
38
Useless Bay
0
5 km
35
Possession Point

Figure 1 (previous page). *Index map of Whidbey Island (on the left) and Camamo Island (on the right), after Easterbrook (2003a, fig. 7). This lidar image shows that the north-south drumlins of the Vashon stade are overprinted in the north by southwesterly trending drumlins. Black arrows show the direction of glacial flow. The numbers along the coast line are the height (in meters) of glaciomarine deposits of the Everson interstade at the sites marked by adjacent circles. The southernmost strand of the Devils Mountain fault (DMF) is from Dragovich et al. (2005); the location of the southern Whidbey Island fault zone (SWIFZ is from Johnson et al. (1996). Other abbreviations are: PP = Point Partridge; d = deltas of Partridge Gravel. The box near Penn Cove is the area of Figure 6.*

northwesterly tending folding is that Miocene strata < 14 Ma are restricted to the synclines; examples are Blakeley Harbor Formation in the Seattle basin, unnamed strata in the Everett basin, and the Boundary Bay Formation in the Whatcom basin.

Another important indication of the youthfulness of the northwesterly trending folds is the regional distribution of Pleistocene strata. Although Fraser-age strata occur throughout the WPF syncline, older Pleistocene strata only occur in the anticlines that intersect the WPF syncline (Booth et al., 2004), such as the Newcastle Hills anticline, which bounds the Seattle basin on the south, and the Kingston arch. The Kingston arch is the anticline between the Seattle and Everett basins, and it includes Whidbey Island.

Booth et al. (2004) noted that although Fraser-age strata are not folded, in places the older Pleistocene sediments do have modest tectonic dips. Johnson et al. (1996, fig. 11) photographed such an example on the west coast of central Whidbey Island.

Major faults

Two major faults underlie Whidbey Island (Fig. 1) but are not particularly obvious in the Pleistocene sediments or their map patterns. The Devils Mountain fault crosses the northern end of the island. The southern Whidbey Island fault zone (SWIFZ) strikes northwesterly across the southern part of the island. Both faults may have originated in the Paleogene as dextral strike-slip faults. Both are considered to be active today.

The Devils Mountain fault separates the pre-Cenozoic crystalline rocks of the Cascade Range, the San Juan Islands, and southern Vancouver Island from the Cenozoic strata of the Everett structural basin (Johnson et al., 1996). The rare outcrops of pre-Cenozoic crystalline rocks on the north end of Whidbey Island are north of this fault (Dragovich et al., 2005). The rocks on Fidalgo Island (Cheney, 2013) are other examples of the pre-Cenozoic crystalline rocks.

The SWIFZ consists of three strands in a 8-km wide zone that extends southeastward from near Victoria, BC, across Whidbey Island to the mainland near Woodville, WA (Kelsey et al., 2004; Sherrod et al., 2008). In general, the fault marks two different types of crystalline basement: the Paleocene Crescent Basalt on the southwest and the diverse pre-Cenozoic rocks of the Cascade Range and Vancouver Island on the northeast. The fault may have originated in the Eocene as part of a major dextral strike-slip fault, the Coast Range Boundary fault of Johnson et al. (1996).

Seismic reflection lines show that northeast of the SWIFZ the base of the Quaternary is 420 m lower and the early Oligocene is 1480 m lower than to the southwest of the fault (Johnson, et al,, 1996). Miocene strata are absent southwest of the fault, but in the synclinal Everett Basin northeast of the fault, coal-bearing Miocene strata are about 300 m thick (Johnson et al., 1996). Trenching reveals that Holocene units are uplifted only 1 to 5.5 meters on the northeastern side of strands of the fault (Kelsey, et al., 2004; Sherrod, et al., 2008). Such evidence for decreasing amounts of offset in progressively younger units suggests that the fault has been active multiple times.

The Partridge Gravel near Coupeville is commonly regarded as post-Vashon in age, but in the alternative presented below it is considered to be slightly older than the Vashon till. All of the known localities of Partridge Gravel and its equivalents on Whidbey Island are in and northeast of the SWIFZ. This suggests that the gravel, which is about 100 m thick, might be offset by the fault (by an amount intermediate between the amounts that the Miocene and Holocene units are offset).

IV. PLEISTOCENE GEOLOGY OF WHIDBEY ISLAND

by Eric S. Cheney

Chronology

The Puget Lowland contains evidence for multiple periods of Pleistocene glaciation in the last 2 million years (Easterbrook, 1994a, fig. 2. However, only the youngest three crop out on Whidbey Island (**Table 1**).

Pre-Vashon Stratigraphy

Figure 2 illustrates the Vashon and older stratigraphy of Whidbey Island. The Double Bluff Drift is overlain by sand, silt, clay, and peat of the Whidbey Formation. The maximum thickness of the Whidbey Formation is about 60 m, at Double Bluff; at Possession Point it is about 30 m thick (**Figs. 2 and 3**). Double Bluff is the type locality for the Double Bluff Drift and for the Whidbey Formation (Easterbrook et al., 1967).

Possession Drift consists of compact till, glaciomarine drift (gmd), and outwash sand and gravel. Easterbrook et al. (1967)

Interval		Lithostratigraphy	Age
FRASIER GLACIATION	Everson Interstade	Glaciomarine drift (gmd)	< 14 ka
	Vashon Stade	Vashon till and advance outwash	14 to 18 ka 18 to 20 ka
OLYMPIA NONGLACIAL INTERVAL		Unnamed	27 to 37 ka?
POSSESSION GLACIATION		Possession Drift	60 to 80 ka
WHIDBEY INTERGLACIAL		Whidbey Formation	80 to 125 ka
DOUBLE BLUFF GLACIATION		Double Bluff Drift	125 to 185 ka

Table 1. *Glacial chronology applicable to Whidbey Island. Ages are from Polenz et al., 2005.*

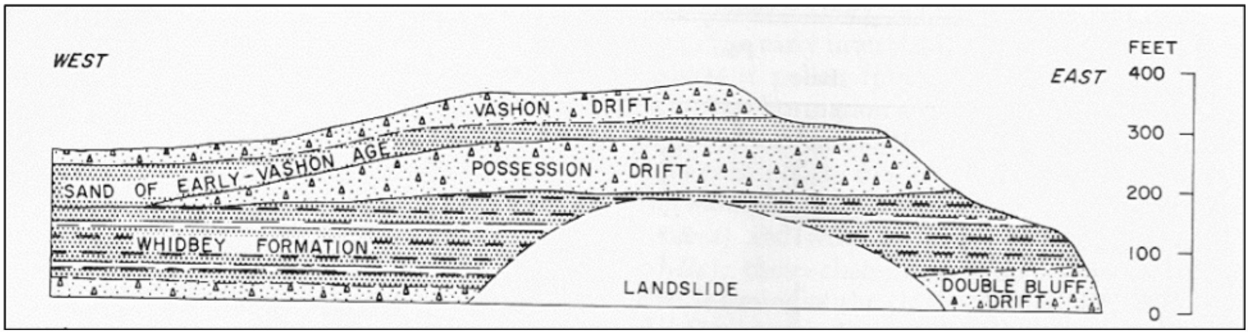

Figure 2. *Cross-section of Possession Point (Easterbrook et al., 1967, Fig. 2*)

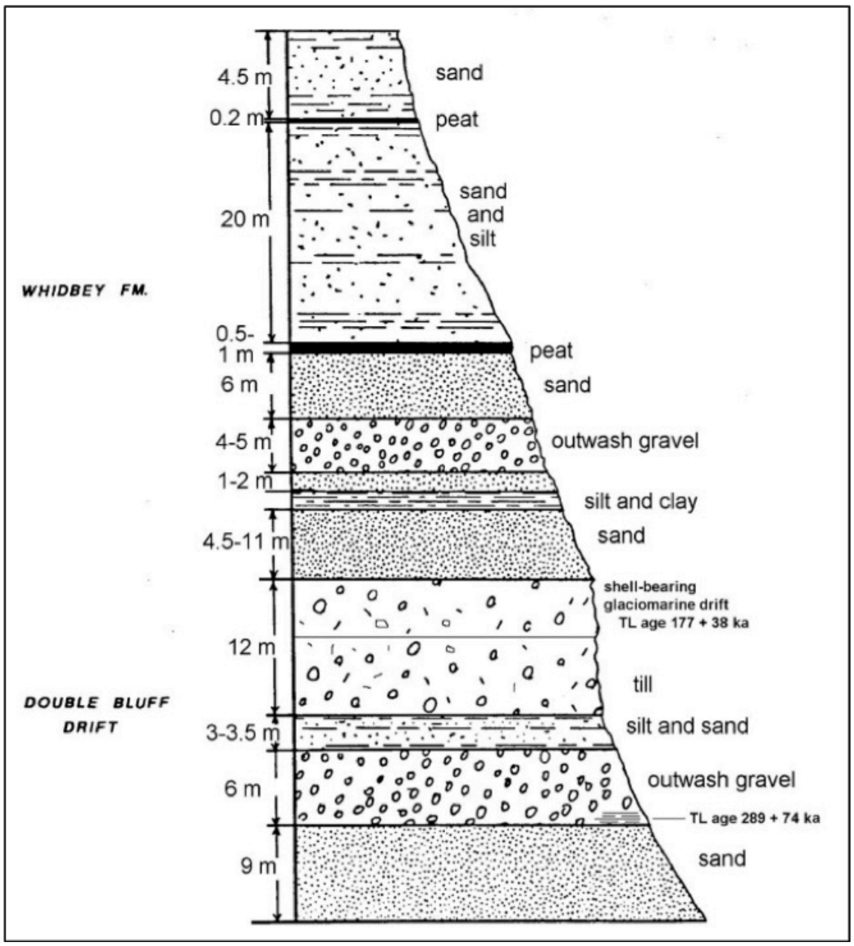

Figure 3. *Composite stratigraphic section at Double Bluff (Easterbrook, 1994a, fig. 9).*

chose Possession Point as the type locality of the Possession Drift.

The ages of the Double Bluff Drift and Whidbey Formation are beyond the limit of ^{14}C dating. They are dated by thermoluminescence (TL) and amino-acid techniques (Easterbrook, 1994a; Dragovich et al., 2005; Polenz et al., 2005).

Figure 3 shows the lithologies of the Double Buff Drift at Double Bluff. The uppermost unit of the formation is glaciomarine drift. Note that most of the drift is not till.

Figure 3 shows the thickness and lithologies of the Whidbey Formation in the type area at Double Bluff. Beds of black peat are characteristic of the formation. In the Coupeville area, the Whidbey Formation is < 40 m thick and consists predominantly of two lithofacies (Dragovich et al., 2005; Polenz et al., 2005): planar-bedded silt and buff-weathering, variably cross-bedded sand (with minor gravel); in general, the sandy facies comprises the upper two-thirds of the formation.

Vashon Glaciation

On the upland surfaces of Whidbey Island, sparse meter-scale erratics indicate the extent of Vashon till. Advance outwash is also locally abundant and gradational into till.

The most dramatic indicators of Vashon glaciation are drumlins and flutes (Fig. 1). South of Penn Cove they are north-south (Easterbrook, 2003b, fig. 5; Polenz et al., 2006, fig. 1). North of Penn Cove the southwesterly trending drumlins and flutes overprint the north-south ones (**Fig. 4**).

Post-Vashon history

After the Vashon ice retreated to near the Strait of Juan de Fuca, it collapsed catastrophically (Easterbrook, 2003b). This is documented by **Figure 5**, which shows that ^{14}C dates of Everson glaciomarine drift (gmd) are not progressively younger from Whidbey Island to the Canadian border).

Collapse of the ice sheet in the Strait of Juan de Fuca caused ice still grounded in the northeastern part of the Puget Lowland to be unsupported on the west. The remaining ice then flowed southwesterly (Easterbrook, 2003b).

At the time of collapse, the margin of the ice sheet was along the south shore of Penn Cove (Easterbrook, 2003b, Polenz et al., 2005). The popular belief is that meltwater from this ice built two kame deltas of Partridge Gravel (Easterbrook, 1967, 1994b, 2003b; Carlstad, 1992; Polenz et al., 2005): one with an upper surface at 67 meters east of Coupeville, and one with an upper surface at 61 m west of Coupeville near Fort Ebey State Park (**Fig. 6**). As noted below, closed depressions in these deltas were interpreted as kettles.

Everson gmd unconformably overlies Partridge Gravel at Fort Ebey State Park (Easterbrook, 1994b; Carlstad, 1992; Polenz et al., 2006). The oldest of three dated samples of gmd in the Penn Cove area is 12 ka ^{14}C BP (Easterbrook, 2003b).

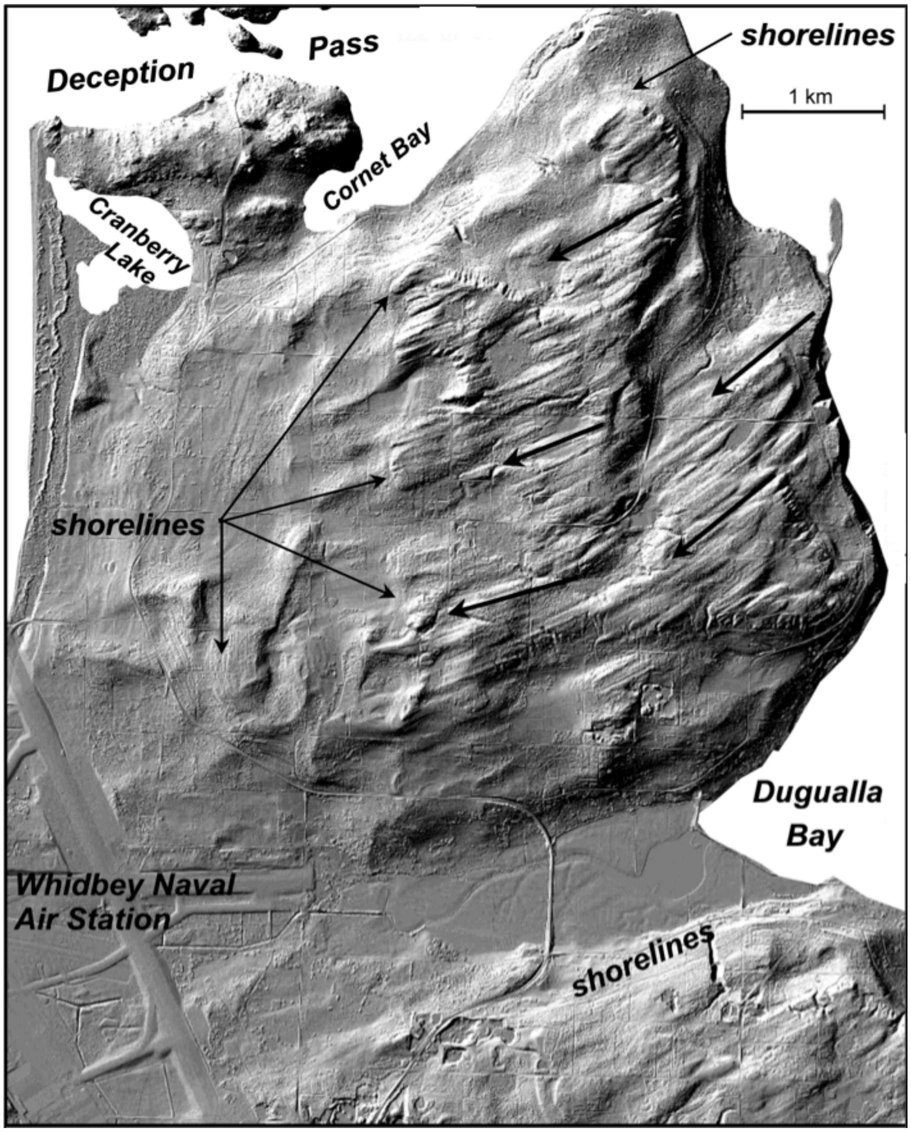

Figure 4. *Southwest-trending drumlins and flutes superimposed on Vashon south-trending drumlins, northern Whidbey Island, Wash (Easterbrook, 2003b, fig. 4). Raised shorelines truncate the drumlins.*

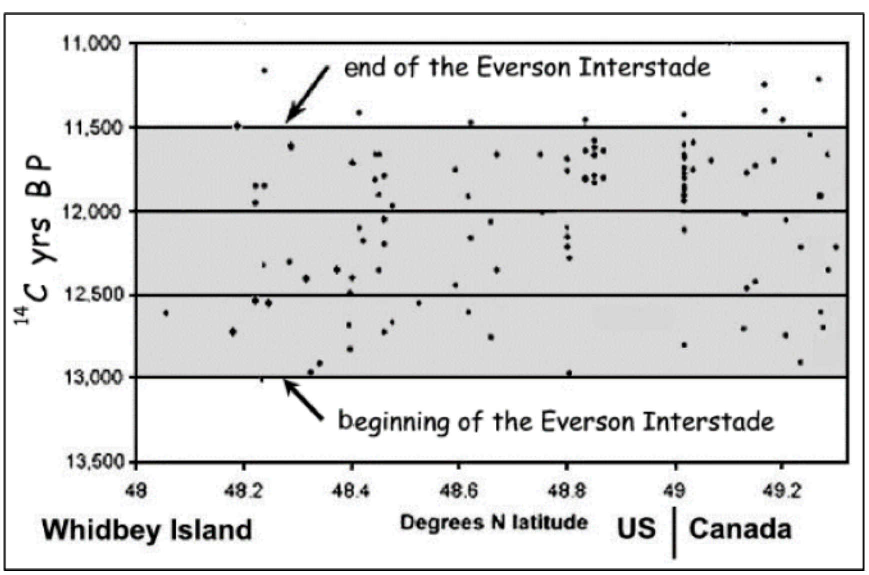

Figure 5. *Geographic distribution of ^{14}C yr BP of Everson glaciomarine drift (Easterbrook, 2014, Fig. 7).*

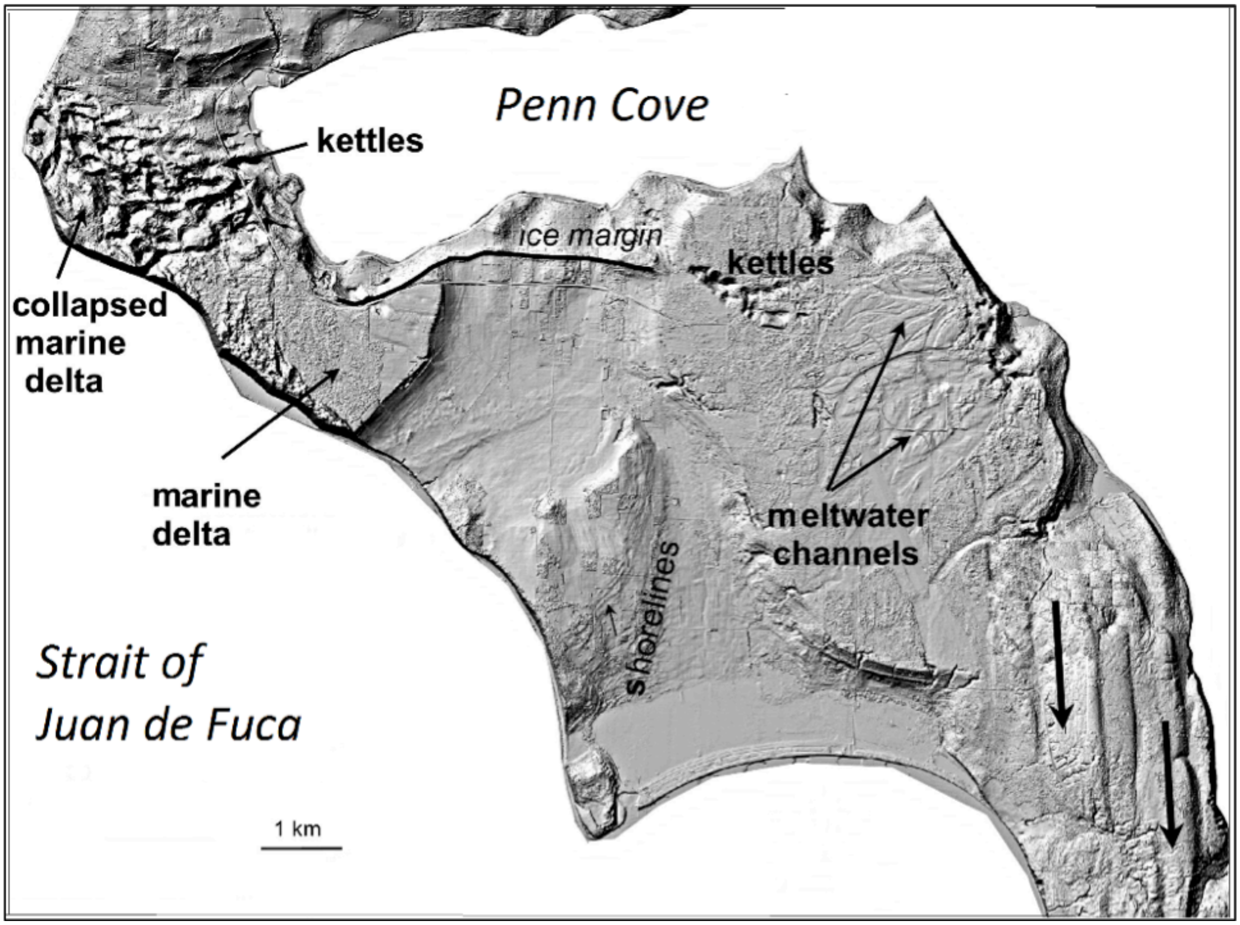

Figure 6. *Lidar image showing landforms in the Penn Cove area (after Easterbrook, 2003a, fig 8 and Polenz et al., 2005, fig. 1). Vashon drumlins and flutes are the north-south features south of Penn Cove. The outwash and marine deltas are composed of the Partridge Gravel. Still younger marine shorelines are slightly younger than Everson glaciomarine drift (11,700 ^{14}C yr BP).*

Numerous marine shorelines are superimposed on the Partridge Gravel (Carlstad, 1992) and on the Vashon and older deposits (Fig. 1). The northward increase in the altitude of these shorelines is due to greater isostatic uplift northward (Easterbrook, 2003b).

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V. AN ALTERNATIVE INTERPRETATION OF THE AGE AND ORIGIN OF THE PARTRIDGE GRAVEL

by Eric S. Cheney

Previous authors (Easterbrook 1968, 1994a, b, 2003a, b; Carlstad, 1992; Polenz et al., 2005), regarded the closed depressions in the Partridge Gravel near Partridge Point as kettles; therefore, they included the gravel as part of the post-Vashon Everson glaciomarine drift. The alternatives proposed here are that the Partridge Gravel is pre-Vashon and that the closed depressions are potholes.

These alternative are based on (1) the distribution of clasts of red chert (CRC) on the beaches of the Puget Lowland, (2) the closed depressions in Partridge Gravel at Fort Ebey (Stop 2-2) are erosional features, and (3) map patterns show that Partridge Gravel underlies Vashon till. In this senario, the Partridge gravel was derived from the Cascade Range in pre-Vashon time. Subsequently, the Puget lobe (and its meltwaters) overrode the gravel and transported CRC at least as far south as Tacoma; the Juan de Fuca lobe transported CRC at least as far west as Port Angeles.

Distribution of clasts of red chert

Since mid-2013 I have been noting the maximum size of clasts of red chert (CRC) on the beaches of northern Puget Sound, Whidbey Island, and the Olympic Peninsula. CRC are an extremely minor but geologically conspicuous component of the beaches. Each data point in **Figure 7** represents a 0.5- to 1.5-km traverse of a beach at low to medium tide.

CRC characteristically are rounded and aphanitic. The most distinctive feature of clasts larger than a few centimeters is one or more generations of white quartz veinlets. Other than the quartz veinlets, most CRC are massive. Some CRC have cm-scale banding or bedding. In thin section some CRC contain radiolarians (D.J. Beal, personal communication, 27 March 2015).

Not all red clasts on the beaches were measured. The following were ingored: (1) red porphyritic and amygdaloidal rocks, (2) reddish, fine-grained clastic rocks, some of which contain CRC up to 2 cm, (3) aphanitic purple, black, gray, and brown rocks, some of which are chert, and (4) bricks.

The two major patterns in Figure 7 are a general southward decrease in size and five radial zones from central Whidbey Island. Significantly, the area of >12 cm CRC south of Penn Cove is coincident with the area of outcrop of the Partridge Gravel, as mapped by Polenz et al. (2005). The Partridge Gravel also underlies prominent coastal bluffs along the shorelines of this part of Whidbey Island.

An area of > 16 cm CRC occurs along the western coast of Whidbey Island. This is most likely due to the vigorous erosion of the coastal bluffs of Partridge Gravel,

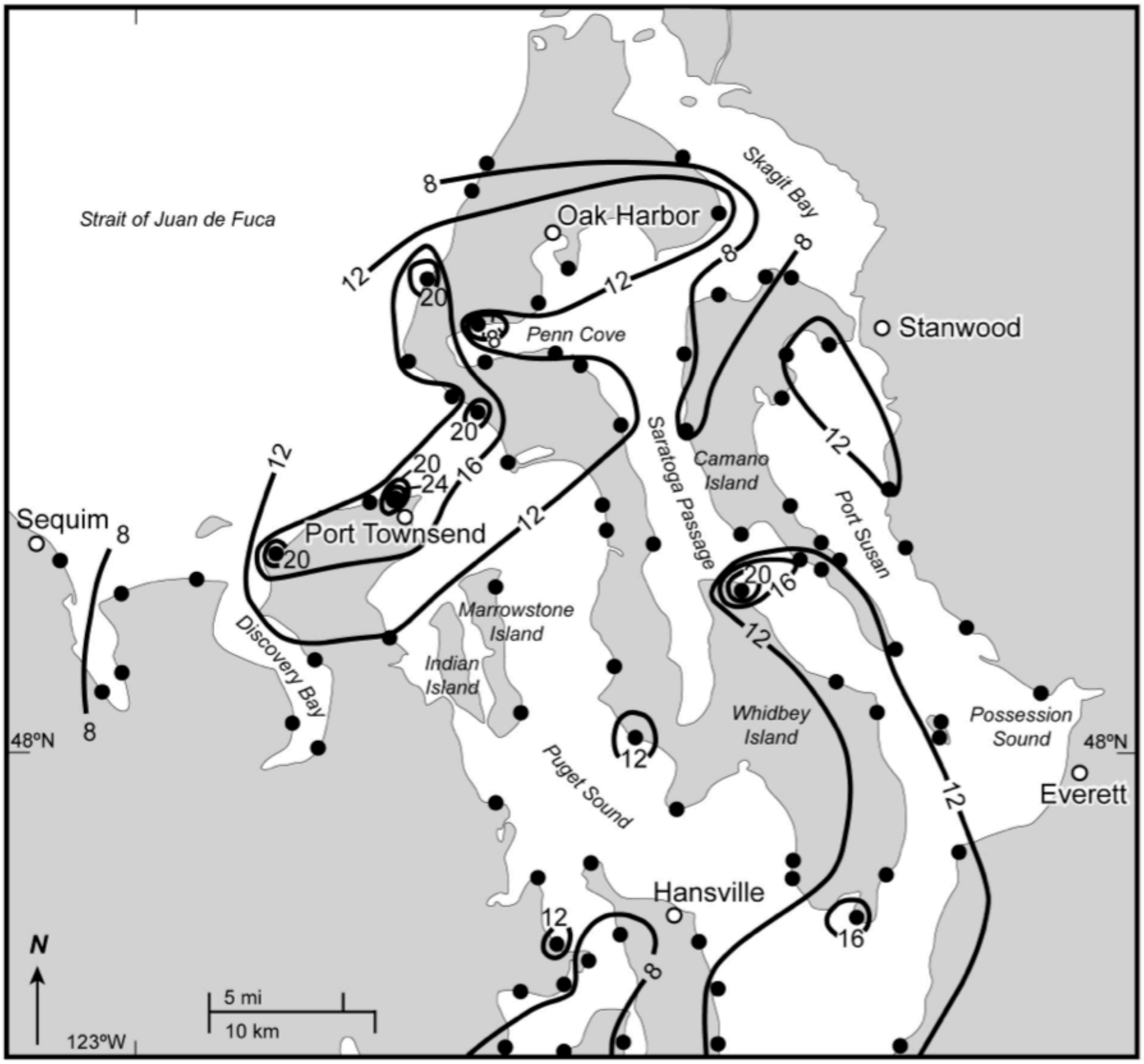

Figure 7. *The maximum size of cobbles of red chert on beaches is contoured in 4-cm intervals. Dots locate 0.5 to 1.5 km traverses of beaches. See text for explanation.*

thereby residually increasing the abundance of larger clasts on the beaches.

CRC > 16 cm do not occur south of Whidbey Island. Figure 7 and the presence of smaller CRC on beaches at least as far south as Tacoma implies that the Puget lobe abraded and smeared CRC southward.

The smallest of the four radial zones of large clasts is north of Fort Ebey. This area of ≥ 16 cm clasts extends north of the area of the Partridge Gravel mapped by Polenz et al. (2005). This extension probably is due to longshore currents.

A > 12 cm area of CRC jogs southwestward from Whidbey Island to west of Port Townsend. The cause of this “dog leg” pattern was either transport by the Juan de Fuca lobe (as it diverged westward from the Puget lobe) or by presently unknown areas of Partridge Gravel southwest of Whidbey Island. The area within the l6-cm isopleth in the Port

Townsend area is underlain by Vashon till (Schasse and Slaughter, 2005). The topography in Schasse and Slaughter (2005) and lidar imagery show that this area has a weak southwesterly trending topographic grain, which favors the glacial alternative.

Two Zones of >12 cm CRC occur in Figure 7 on either side of Camano Island. Significantly, Polenz et al. (2009) mapped the equivalent of Partridge Gravel on Camano Island. Therefore, erosion of Partridge Gravel in areas now occupied by Saratoga Passage and Port Susan on either side of Camano Island may be the source of these two zones. The larger zone in the southeastern part of Figure 7 extends another 14 km southward to Seattle.

An easterly trending zone of > 12 CRC occurs north of Penn Cove. The probable source of CRC for this zone is a NNE trending band of gravel 3 to 4 km wide that passes through the town of Oak Harbor to the northern border of Figure 7. The units in this band ($Qgog_{e,}$ $Ogof_e$ and $Ogod_e$ of Dragovich et al., 2005) seem to be similar to Partridge Gravel.

Closed depressions at Fort Ebey State Park

Easterbrook (1968, 1994a. b, 2003a, b), Carlstad (1992), and Polenz et al. (2005) noted that numerous closed depressions occur in the Partridge Gravel in the isthmus between Point Partridge and Penn Cove. The seemingly obvious conclusion was that these are kettles. Because kettles are syndepositional features in glaciated areas, the Partridge Gravel was, therefore, interpreted to be an Everson (post-Vashon) kame-delta complex.

The scrub forest on the isthmus is so thick that ground-based investigations and aerial photography are virtually useless. The 7.5-minute Coupeville and Port Townsend North topographic quadrangles are revealing, but lidar imagery is even better (**Fig. 8).** Figure 8 is similar to (but at a larger scale than) the relevant portions of Figure 6 and of Figure 1 of Polenz et al. (2005).

Two characteristics in Figure 8 suggest that the closed depressions are not kettles. Firstly, overall, the depressions have a rectilinear reticulated, not a random, pattern. Six or seven westerly trending ridges cause this reticulated pattern. Secondly, remnants of the 61-m high terrace of Partridge Gravel occur as buttes between some of the depressions; the most obvious of these, Fisher Ridge, is, in essence, an erosional fin. These characteristics indicate that the depressions are erosional features incised into the 61 m terrace. Thus, they are giant potholes, not kettles.

The point is that whereas kettles are syndepositional features, potholes and channels are generally post-depositional ones. The potholes and channels may well have been carved in Everson time, but the Partridge Gravel, in which they are carved, is older than Everson.

The suggestion here is that these features were caused by jökulhlaups issuing from the body of glacial ice that occupied the isthmus and, especially, Penn Cove on strike to the east. If at the time the depressions were formed, they were overlain glacial ice, the flow of water could have had considerable hydrostatic head, which would have caused pronounced erosion.

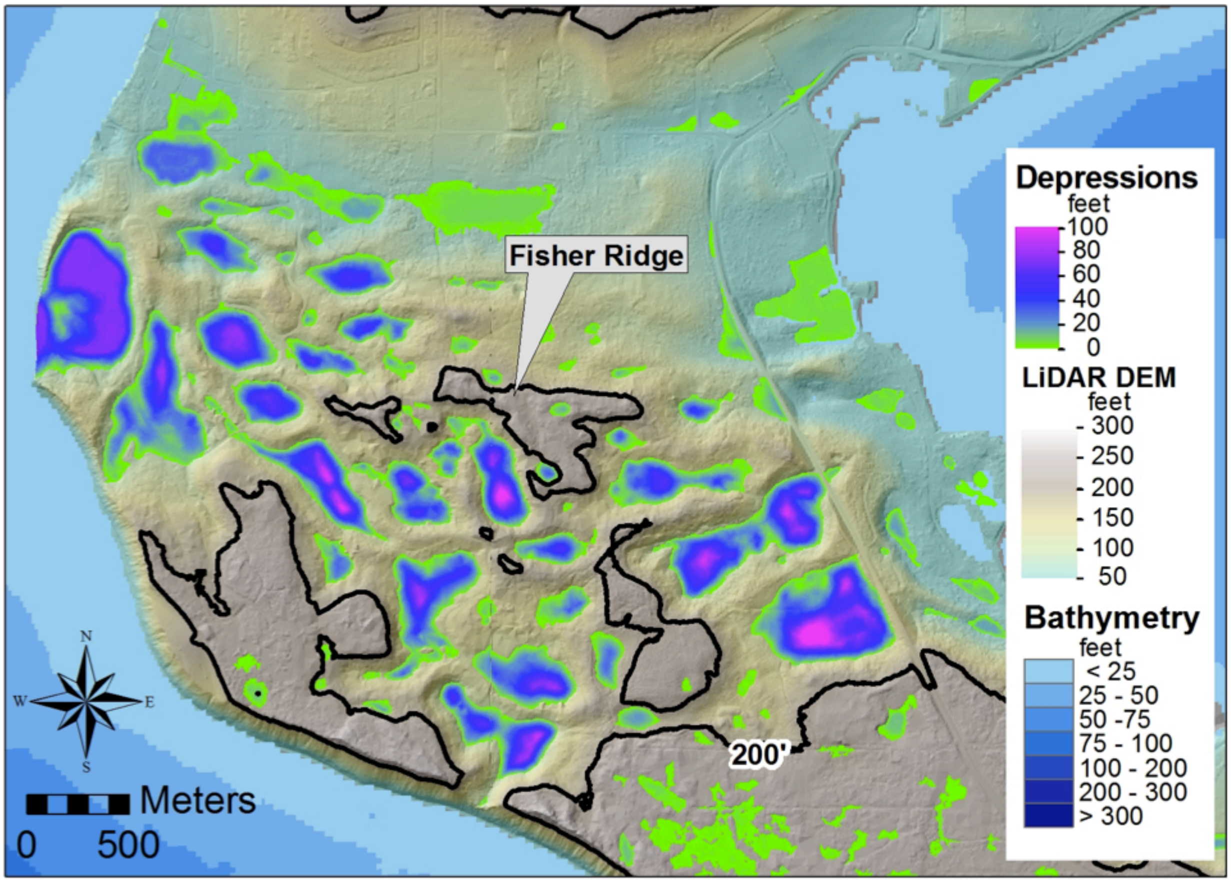

Figure 8. *Closed depressions in Partridge Gravel near Point Partridge. This lidar map was drafted by H.M. Greenberg (ESS, UW) and has illumination form the northwest at 45 degrees. See text for explanation.*

Ebey's Prairie (Stops 2-3 and 2-4) is south of Coupeville and between the terraces of Partridge Gravel (Carlstad, 1992; Polenz et al., 2005). It is the lowland east of the feature labeled "marine delta" in Figure 6. This lowland might also have been caused by jökulhlaups from ice in Penn Cove. However, the surface is now obscured by unconformably overlying Everson gmd (Polenz et al., 2005), so potholes (if any) are obscured.

Dry Falls in central Washington in the Miocene Columbia River Basalt Group (**Fig. 9**) is an interesting analog. The falls and the channels upstream and downstream from them were caused by one or more of the Missoula floods. The altitude of the base of the falls is 1207 feet; the lip of the falls is 1500 feet, and a flood-caused trimline on the slope to the northwest of the falls is at about 1700 feet. Thus, at times the water over the falls was at least 200 feet deep. Note that Umatilla Rock is an erosional fin.

The important point of this analogy is that upstream from the falls, deep, high-velocity water caused ridges and a reticulated pattern of closed depressions (potholes). Some closed depressions downstream from Dry Falls may also be potholes. Of course, on Whidbey Island, no waterfalls occurred west of Penn Cove.

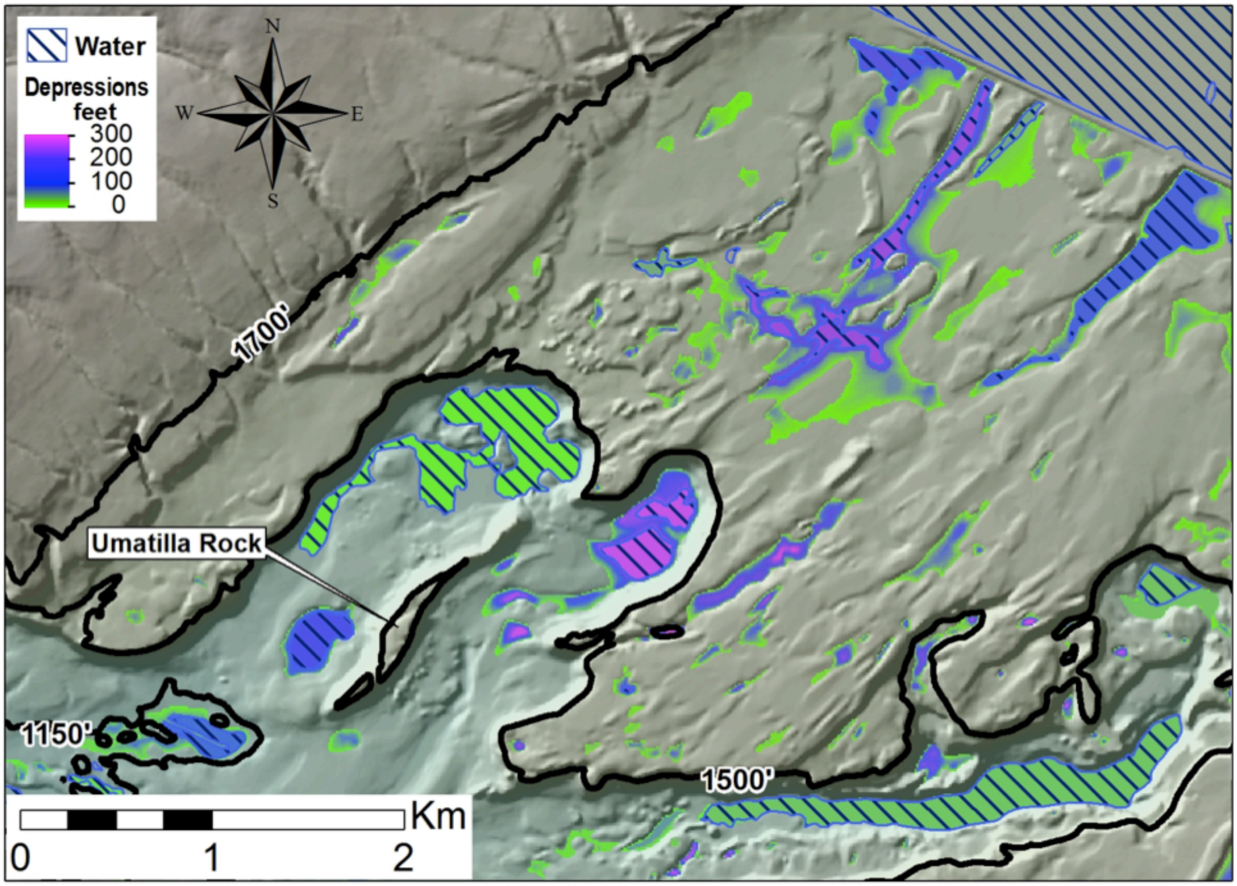

Figure 9. *Dry Falls of the Grand Coulee, WA. This 10 m - Digital Elevation Model was drafted by H.M. Greenberg (ESS, UW) from 7.5-minute topographic quadrangles; it has illumination from the northwest at 45 degrees. Contour lines shown are for 1150, 1500, and 1700 feet. See text for explanation.*

Map patterns

Additional evidence for the age and origin of the Partridge Gravel is the mesa-like topography and geologic map patterns of central Whidbey Island. Although some relief occurs on their basal contacts, the Partridge Gravel and Vashon till in central Whidbey Island are regionally subhorizontal units.

The map pattern of these two subhorizontal units consistently shows that Vashon till is at higher altitudes than adjacent Partridge Gravel. Thus, the gravel is overlain by the Vashon till. The map of Polenz et al. (2005) indicates this relationship northeast of Point Partridge. The map of Polenz et al. (2009) shows that topographically higher Vashon till adjacent to Partridge Gravel ($Qgom_e$) both at Smith Prairie southeast of Coupeville and on the western side of Camano Island. The map of Polenz et al. (2006) has the same pattern north and south of Lake Hancock (about 14 km southeast of Coupeville).

Partridge Gravel may prove to occupy almost the same stratigraphic position as Vashon advance outwash. Therefore, vertical outcrops, such as sea cliffs, also should be evaluated. One example is at Hastie Lake Road (Stop 2-1). There. Dragovich et al. (2005) lumped gravel below the Vashon till with the till. The units are too thin to show separately, and the authors probably assumed that the gravel is Vashon advance outwash, which it may be. However, maybe it is Partridge

Gravel. Perhaps, mineralogical, lithological, and other criteria can be developed to determine whether such gravel is correlative with (or distinct from) the Partridge Gravel.

Other criteria

Two other criteria that might be used to determine the relative age of the Partridge Gravel are its degree of compaction and the presence or absence of erratics. So far, these criteria appear to be indeterminate.

Deposits that were overridden by glacial ice generally are well compacted. Although some of the Partridge Gravel weathers to the angle of repose, suggesting that it is not well compacted, cuts in the sand and gravel pit west of Coupeville and the lower part of the bluffs at Point Partridge (Stop 2-2) are compacted enough to form cliffs.

Erratics resting on Partridge Gravel would indicate the relative age of Vashon till and the gravel. For erratics to be determinant, they would have to be larger than the largest boulders in the gravel. At Point Partridge (Stop 2-2) the largest boulders in the gravel are 1.3 m. Boulders larger than 1.3 m are not evident from roads on the terraces east and west of Coupeville. Two l.6 m boulders on the eastern scarp of the western 61 m terrace could be landscape rocks (emplaced by humans) or dropstones associated with Everson gmd.

VI. FIELD STOP ON GUEMES ISLAND

by Darrel S. Cowan

Directions

In downtown Anacortes continue north on Commercial Avenue, the main thoroughfare, to 6th Street. Turn left (west), and drive about 0.75 km to the intersection of I Avenue. The ferry terminal for the Anacortes-Guemes ferry is at the north end of I Avenue. Northwest of the intersection is a small paved parking area for patrons of the ferry. The ferry schedule and information are at: http://www.skagitcounty.net/Departments/PublicWorksFerry. Walk-on passengers must purchase tickets before boarding the ferry. A ticket is good for round-trip travel.

After disembarking from the ferry, walk east along paved South Shore Road a few meters and drop down to the sandy beach. The westernmost sea cliffs are a couple of hundred meters from the ferry terminal. From there, continue for about 600 or 700 meters east along the beach.

Field Stop

The goal is to examine the poorly consolidated sediments in the cliffs and to discuss their environments of deposition and the genesis of the structures in them. I thank Ralph Haugerud of the USGS for telling me about these exposures.

A consensus would be that deposition was related to glacial activity in the Puget Lowland. Debate occurs about whether the structures are *glaciotectonic*—related to the movement of ice—or tectonic—related to post-depositional deformation unrelated to ice. I strongly favor a glaciotectonic origin.

Along the cliffs are folds that we typically expect to see in metamorphic rocks, not in sediments or strata folded at and near the earth's surface. We characterize the latter as "buckle" or "flexure" folds, wherein the layers have been folded by layer-parallel compression and the thickness of the layers is constant.

However, many of the folds here show evidence of flow of sand from the limbs into the hinges, but this flow, typical of "similar" folds in metamorphic rocks, was accomplished when the sand was partly consolidated and featured high fluid pressures. Note that many folds have gently dipping or horizontal axial planes and gently plunging or horizontal hinges. Some remarkable folds, which also are only rarely observed in high-strain zones in metamorphic terranes, look like eyes. **Figure 10** schematically shows how early buckle folds can be distorted as the hinges are drawn out to resemble socks or sheaths. Cross sections of the sheaths are eye-shaped.

Note the southerly trending lineations on some of the bedding planes. Lineations are common features in sheath folds (Fig. 10). In material undergoing increasing shear strain with the sense of shear shown by arrows in Figure 10, early folds are distorted as the curved hinges are drawn out into sheaths.

Also note a couple of gently dipping or nearly horizontal planes in the cliff a few meters above the beach. The planes are defined by sediment that is slightly more resistant to weathering. Their origin is a good topic for discussion: slip planes, hydraulic barriers, or narrow conduits for groundwater?

I was unable to find a published date for the sediments exposed in these cliffs. However, Easterbrook (1969, p. 2278) published a radiocarbon date of “>40,000 yr B.P.” from “wood in peat” collected at Yellow Bluff on the west side of the island, about 2.8 km west of these exposures. Kathy Troost of UW contacted Sue Kahle (USGS), who provided a memo dated 10/26/08, in which Charles Lindsay reports a “conventional radiocarbon age,” also from Yellow Bluff, of 38640 +/- 2030 B.P. If the sediments at Yellow Bluff and those here are correlative, the “ferry section” is likely ca. 38,000 years old and, therefore, of Possession age.

VII. FIELD STOPS ON WHIDBEY ISLAND.

by Eric S. Cheney and Seven R. Grupp

This road log begins at MP 42.1 of SR 20, which is the island between the bridges at Deception Pass. This is also the site of

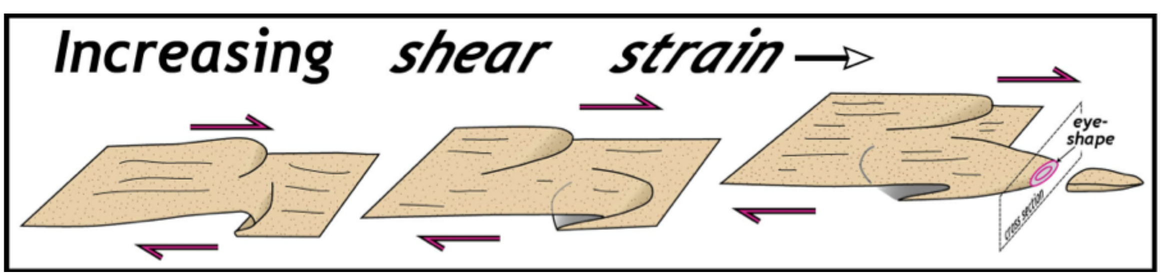

Figure 10. *Schematic development of sheath folds (wmblogs.wm.edu). In a material undergoing increasing shear strain with the sense of shear shown by arrows, early folds are distorted as the curved hinges are drawn out into sheaths. The thin black lines represent lineations, which are parallel to the direction of tectonic transport.*

Stop 3 in the guidebook for Fidalgo Island (Cheney, 2013). If necessary, we will stop at the toilets at the south end of the southern bridge. Otherwise, we will proceed southward about 12.7 miles to the turn-off to Stop 2-1. Before that, about 6 miles from here, SR 20 crosses Clover Valley, which here marks the covered trace of the southern strand of the Devils Mountain fault (Fig.1).

Drive to Stop 2-1

South of Oak Harbor at MP 28.4, turn west on Hastie Lake Road. Follow this road 2.4 miles to its end, which is a parking lot and boat ramp adjacent to the beach.

Stop 2-1, Hastie Lake Road

The goals of this stop are to examine the Whidbey Formation and its contact with the overlying gravel and Vashon till. Walk about 0.6 km southwest along the beach to beyond the highest portion of the sea cliffs.

The Whidbey Formation consists of two lithologies in beds 0.5 to 2 m thick, Planar bedded and laminated, micaceous silt contains black peat ≤ 14 cm thick. In contrast, most of the cliff, especially its upper part, is planar bedded but variably cross-bedded, buff-weathering sand.

In this area the Whidbey Formation contains clasts of dacite and grains of hypersthene typical of Glacier Peak. These characteristics suggest that the formation was derived from the drainage of the Skagit River (Dragovich et al., 2005; Polenz et al., 2005).

The cliff is capped by 5 to 8 m of Vashon till. Below the till weakly bedded cobblely gravel is 0 to 10 m thick. However, at the first ravine in the cliffs, gravel with foreset beds dipping southward, truncates the entire interval of Whidbey Formation. Thus, the relief on the upper contact of the Whidbey Formation is about 30 m.

An important issue is whether this gravel is equivalent to the Partridge Gravel of Stop 2-2. What mineralogical, lithological, or other criteria could be used to correlate (or to distinguish between) the two?

Note also the abundance of erratics ≥ 1 m on the beach. Some of these have photogenic lithologies. More importantly, they attest to the nearby presence (or former presence) of Vashon till, which is an issue at Stops 2-2 and 2-4.

Drive to Stop 2-2

Return to SR 20 and turn right (southbound**).** At MP 25.3 on SR 20 opposite Penn Cove, turn west on Libbey Road and begin following signs to Fort Ebey State Park. In 0.9 miles turn left on Hill Valley Road. At 0.1miles from Libbey Road the Hill Valley Road crosses the main drainage of the isthmus and then begins to skirt closed depressions (Figure 8). Thick vegetation obscures most observations of the depressions, but gravel does mantle a few roadcuts. Proceed another 0.8 miles to Park Headquarters. At Park Headquarters, turn right, following signage to "beach"; proceed 0.5 miles to the end of the road. At the end of the road are toilets and the trailhead to Lake Pondilla.

Stop 2-2: Fort Ebey State Park

Hike five minutes on the trail to Lake Pondilla; the lake is in one of the larger depressions. Hike around the southwestern side of the lake and ascend the ridge adjacent to the beach. East of the crest of the ridge the trail is incised into fine-grained dune sand. Subtle discontinuous patches of dune sand also occur along the trail on the west side of the ridge.

Where the trail begins to descend the ridge to the beach is a good place to guesstimate the rate of retreat of the costal bluffs due to wave action. Because this part of the coastline of Whidbey Island is exposed to the full fetch of the Strait of Juan de Fuca, erosion of the bluffs probably is maximal here. Note that kelp is growing about 100 m offshore (kelp usually grows in less than 10 m of water). NOAA Chart 18423 (1:80,000) shows that the 5-fathom isopleth is about 320 m offshore. Assuming that the 320 m-wide swath is a mostly a wave-cut bench (which it may not be) and that sea level attained it present height about 10,000 years ago, you can guesstimate the rate of erosion of the bluffs.

As the trail descends the ridge, note the classic example of wave refraction around Point Partridge to the south.

Once on the beach proceed 100 m north to observe outcrops of the Partridge Gravel. Foreset beds dip northward. The beds generally are 0.5 to 2 m thick and consist of alternating matrix- and clast-supported gravel. Clasts up to 1.3 m are weathering out of the upper part of the bluff, and similar sized clasts litter the beach. In 2013, a 16 cm CRC occurred on this beach (Fig. 7).

At the base of the bluff the gravel is not at the angle of repose; rather, it is compacted enough to produce steep faces. Could this compaction be due to formerly overriding glacial ice? Note that a 1.9 m boulder of metabasalt on the beach might support this hypothesis.

Proceed another 100 m to the north to the last outcrop of Partridge Gravel. Here the gravel is very poorly bedded but the orientation of cobbles indicates north-directed foreset beds. Note that this gravel is well compacted. More importantly, at the top of the cliff is buff-weathering, well bedded Everson gmd unconformable on the gravel.

If time permits, continue another 250 m northward to the parking area at the end of Libbey Road. Just north of the parking area is fine-grained sediment with isolated cobbles (dropstones) of the Everson gmd. This is part of the unit that overlies the Partridge Gravel to the south.

Return southward along the beach. About 75 m south of the access trail from Lake Padilla is a major re-entrant in the bluff, and logs in it are at about sea level. Contemplate the origin of this re-entrant.

Continue another 30m to the south to the next access trail. Take this trail to the paved road, turn left, and return to the vehicles.

Drive to Stop 2-3

Return to SR 20 and turn south toward Coupeville. At 1.1 miles to the south (MP 24.2) the road is on top of the mesa (terrace) that is underlain by Partridge Gravel. At MP 22.7 the road begins to descend the eastern edge of the mesa onto the younger glaciomarine drift that underlies Ebey's Prairie. At the traffic light at MP 21.8 turn right (south) on South Main Street and proceed 0.1 mile. Park behind the building on the right that hosts Coupeville Coffee and Bistro. This is only a brief photo opportunity.

Stop 2-3: Big Rock, the Coupeville erratic

Big Rock is an erratic of greenstone about 8 m high. Its bedrock source most likely is Mount Erie, which is about 40 km north of here (and north of the Devils Mountain fault). Big Rock lies at an altitude of about 30 m and is just south of the former margin of the Continental Ice Sheet that occupied Penn Cove (Fig. 6). The north-south orientation of Big Rock and similar trending striations on its top imply that the erratic was deposited subglacially (T. W. Swanson, written communication, 15 April 2015).

This area at Big Rock is underlain by 5 to 8 m of younger Everson gmd containing shells dated at 11,535 ± 300 ^{14}C yrs BP at West Beach. This gmd is, in turn, underlain here by Partridge Gravel (Polenz et al., 2005). Thus, maybe half of Big Rock is embedded in gmd. Whether Big Rock rests on Partridge Gravel is unknown.

Drive to Stop 2-4

Proceed southward on South Main street for 0.2 miles and turn right (west) on Terry Road. Continue on Terry Road for 0.3miles and turn left (south) on Ebey Road. Follow Ebey Road and Ebey's Landing Road 1.4 miles southwestward along Ebey Prairie to Ebey's Landing State Park at the beach.

While transiting Ebey Prairie on Ebey's Landing Road note the scarp to the right (west). This is the edge of the 61 m mesa that is underlain by Partridge Gravel. Ebey Prairie is underlain by younger gmd and is at an altitude of about 37 m. A similar but less obvious scarp occurs in the distance to the east; it is the western side of the body of Partridge Gravel that underlies 67 m Smith Prairie.

Stop 2-4: Ebey's Landing State Park

Weather permitting; this is a scenic place for arm-waving, followed by lunch. We will explore neither the beach nor the bluff.

As for arm-waving, Port Townsend is 8 km to the southwest across Admiralty Inlet. Southwest of Port Townsend are the Olympic Mountains; the portion we see is predominantly Paleocene metabasalt.

Until after WWII this inlet was guarded by coastal batteries at Fort Ebey and at Fort Casey (which is 3 km to the southeast of here), as well as by Fort Worden and Fort Flagler near Port Townsend. Guns at these forts could enfilade any hostile ships trying to enter Puget Sound. .

The swale in the coastal bluff here is underlain by the Everson glaciomarine drift of Ebey's Prairie. The higher bluffs to the northwest and southeast are underlain by Partridge Gravel (Polenz et al., 2005). Atop the grassy bluffs to the northwest are knobs with pine trees on their leeward side; these knobs are inactive sand dunes. Small outcrops at the base of buffs are well-bedded sand (probably Whidbey Formation).

One argument for a post-Vashon age for the Partridge Gravel is that it is so unconsolidated that it occurs at the angle of repose in the sea bluffs. In the bluffs, here, the gravel is mantled by dune sand: the sand is the material at the angle of repose.

Another argument for a post-Vashon age for the Partridge Gravel is that, at least in the Coupeville area, few, if any, glacial erratics larger than the l.3 m clasts in the gravel occur on the upper surface of the gravel. On this beach the largest clasts are < 45 cm. A 12.3 cm CRC occurred on this beach in early 2015 (Fig. 7).

Drive to Stop 2-5A
Return to SR 20 and turn east (southbound) toward Coupeville (and Stop 2-3). From the traffic light at South Main Street in Coupeville, continue southbound on SR 20 for 5.6 miles to the junction leading to the Port Townsend ferry. Beyond this junction the main highway becomes SR 525. Continue southbound and in 2.5 miles at MP 28.0 of SR 525, turn right (west) on Ledgewood Beach Road.

Proceed 0.3 miles on Ledgewood Beach Road and turn right (north) on Fircrest Avenue. In 0.4 miles take the first left, East Seaward Terrace. At the junction in 0.2 miles turn left on Driftwood Way and park in the hairpin curve. Walk to the northern edge of the Ledgewood slide (**Fig. 11)**. We will not venture on to the slide and on adjacent private property.

Stop 2-5A: North side of the Ledgewood Landslide
The Ledgewood landslide (Fig. 11) occurred on 27 March 2013, a year before the more famous Oso landslide in the western Cascade Range. Ledgewood is the biggest historic landslide on Whidbey Island. The Ledgewood area was platted for residential development in the 1960s (before permitting and the evaluation of environmental issues were required).

Like Oso, Ledgewood occurred in glacial sediments and during the end of the annual "monsoon" season. Like Oso it was caused, at least partially, by erosion of the buttress of an older landslide below it. Unlike Oso, no loss of life occurred. Both landslides may yet cause serious new evaluations of landslide hazards in the State (and the issuance of new zoning regulations).

Most of the information reported here is from Cool and Gordon (2013).

The Ledgewood landslide occurred midway along an ancient (prehistoric) landslide complex that is 2.5 km long (Polenz et al. (2009). This complex straddles the trace of the northeastern strand of the SWIFZ (Polenz et al., 2009), but no seismic activity occurred in this area at the time of the landslide.

The head scarp of the 2013 slide is about 180 m long, and its top is at an altitude of about 61 m. The width of the slide ranges from 200 m to 290 m, and it advanced 76 m into the sea. The toe of the slide uplifted beach gravel and driftwood logs about 10 m. Figure 11 shows that by 2014 half of the portion of the slide in the sea had been eroded by wave action (Google Earth, 07/10/2014, visited 03/18/2015). By March 2015, the shoreline was again straight, but much landslide debris still remained in the high tide zone.

Parts of a previous landslide adjacent to the beach were sporadically active in the 15 years prior to the Ledgewood slide. This coastal slide was again active 5 days before the Ledgewood slide; the Island County Public Works Department obliged by repairing the access road prior to the Ledgewood slide.

Wave erosion and monsoon-generated groundwater probably initiated reactivation of the coastal slide. Movement of this older slide likely removed a buttress for the Ledgewood slide. The combined movement of the two slides transported a house on the older slide 41 m horizontally and 12 m

Figure 11. *Google Earth photograph of 7/10/14 of the Ledgewood landslide that occurred on 3/2713. The road shown as Driftwood Way no longer exists in the area of the slide.*

downward. Some of the southern part of the slide acted more like a mudflow than as discrete landslide blocks.

Initial collapse along the 2013 scarp removed 3 to 12 m of the lawns of adjacent houses; a similar amount of lawn was lost thereafter. In the following months, no tension cracks or other indications of movement appeared east (inland) of the scarp.

From our present viewpoint, the 2013 scarp passes both north and south into older (vegetated) scarps of the ancient landslide complex. Note also the vegetated but hummocky topography along the road between us and the parked vehicles.

"Jack-strawed" trees mark the lower part of the slide.

Polenz et al. (2009) mapped the area that became the Ledgewood slide as Vashon till. However, drilling indicates that the area is underlain by very well compacted sand with interbedded layers of silt and clay. On the southern part of the scarp these units dip moderately southward. Silt and clay units in the sand at about sea level may have been the basal planes of movement of the older coastal slide and the distal part of the subsequent

Ledgewood slide (Cool; and Gordon, 2013, figs. 9 and E1 to E9).

Drive to Stop 2-5B
Continue downhill on Driftwood Way. Take the first left and turn around in the parking lot next to the beach. Return uphill. Beyond Driftwood Way, the road ascends the scarp of the prehistoric landslide complex.

On the way to the junction of Fircrest Avenue and Ledgewood Beach Road, notice that some of the houses on the west side of Fircrest Avenue have unimpeded views to the west. At the junction, continue straight 0.1 miles on Fircrest Avenue to where the road curves sharply west. The curve is the top of the emergency road that was built to allow access to the residences south of the 2013 landslide.

Beyond the curve, the road descends the scarp of the prehistoric landslide complex. Before proceeding beyond the curve, be sure that no vehicles are coming up this one-lane road. To help preserve this fragile road, descend one vehicle at a time. When our lead vehicle reaches the bottom of the hill, it should alert other vehicles not to go up the road until all of our vehicles have descended. At the junction in 0.1 miles at the base of hill turn right (north) and proceed 0.15 miles to the end of the road. A short and easy bushwhack leads to the southern edge of the Ledgewood landslide.

Stop 2-5B: South side of Ledgewood landslide
This stop provides a better view than Stop 2-5A of the bedded units in the 2013 scarp. Because these units are so well compacted, have a moderate dip, and seem to be limited to the area of the ancient landslide complex, they may be pre-Vashon in age (D.T. Bedard, personal communication, 3 April 2015).

Drive to Stop 2-6
Return to SR 20 and proceed southbound for 11.0 miles to MP 16.9. Turn right (west) on Double Bluff Road and proceed 1.9 miles to the parking lot. Because the hike along the beach to the western of the two bluffs to see the Double Bluff Drift is more than 2 ½ miles, this is an optional stop, This stop requires several hours; so use of the toilets here may be strategic. Bring water and appropriate protection from the sun. In the interests of safety and time, do not climb the bluffs.

Stop 2-6: Double Bluff (Optional)
In the eastern bluff, bedded deposits extend to the top (97 m).. The basal 37 m consist of alluvial sands that are believed to be derived from the Snohomish River valley shortly before Vashon glaciation (T.W. Swanson, written communication, 15 April 2015). Flame structures (liquefaction features) occur in the sand 12 to 15 m above the beach, and rip-up clasts occur in this and other units. Although the flame structures might have been seismically induced by the nearby SWIFZ, the rip-up clasts imply overloading by floods and sediments derived from them (T. W. Swanson, written communication, 15 April, 2015).

Overlying the alluvial sediments are clays of the Lawton Formation, pebbly sandy outwash (similar to the Esperance sand farther south), and Vashon Till.

In 2013 a 9.7 cm CRC occurred on this beach (Fig. 7).

Midway down the beach blocks of black peat (from the Whidbey Formation) are as large as 10 by 3 m and are about 1/2 m thick. They form “cuestas” in the beach sand.

In the bluffs, an unconformity at the base of the alluvial sediments truncates the Whidbey Formation and its included peat (T. W. Swanson, written communication, 15 April 2015)

Mammoth tusks, bones and teeth (presumably from the Whidbey Formation) once were plentiful on the beach (D.J. Easterbrook, written communication, 29 March 2015).

The Double Bluff Drift is exposed in the western of the two bluffs. Use Figure 3 to interpret the stratigraphy but note that the strata are not horizontal.

Drive to Mukilteo ferry
Return to SR 525 and turn left (west). Proceed l.1 miles and turn left onto Fish Road. After refueling the vehicles here at the Shell Station, return to SR 525 and proceed 9.7 miles to the Mukilteo ferry. Once we land in Mukilteo, we will proceed to Everett Community College, where the trip ends.

VIII. ACKNOWLEDGEMENTS

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