Cordilleran Ice Sheet glaciation of the Puget Lowland and Columbia Plateau and alpine glaciation of the North Cascade Range, Washington

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THE CORDILLERAN ICE SHEET

The Puget Lowland (Fig. 1) contains widespread Pleistocene sediments, ranging in age from nearly 2 Ma to 10 ka, well exposed in hundreds of km of sea cliffs and roadcuts. During this four–day field excursion, we will examine all of the late Pleistocene stratigraphic units of the past 300 ka.

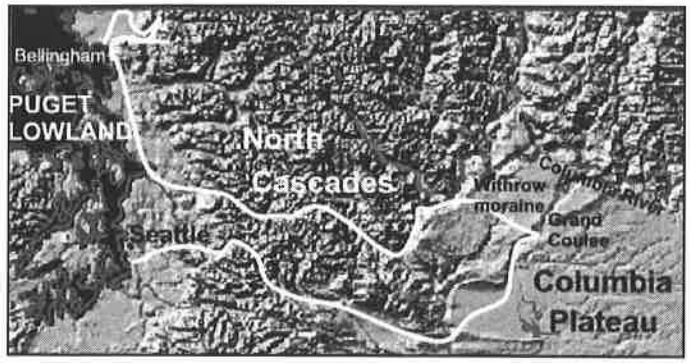

Figure 1. Field trip route map.

DAY 1. DOUBLE BLUFF, POSSESSION, AND VASHON GLACIATIONS; WHIDBEY AND OLYMPIA INTERGLACIATIONS; DISINTEGRATION OF THE CORDILLERAN ICE SHEET

STOP 1-1. DOUBLE BLUFF DRIFT AND WHIDBEY FM. TYPE SECTIONS AT DOUBLE BLUFF

Enroute to STOP 1-1, travel north from Seattle on I–5, take the ferry from Mukilteo to Clinton on southern Whidbey Island, then drive northwest to Freeland. Turn south from Freeland to Double Bluff. Walk the beach from Useless Bay to Double Bluff (about 1.5 km; 1 mi.).

Double Bluff Glaciation. Sea cliff exposures of glacial outwash, till, glaciomarine drift, and glaciolacustrine sediments at Double Bluff (Fig. 2) were designated as the type locality for Double Bluff Drift by Easterbrook et al. (1967). Double Bluff Drift is rarely exposed, but crops out in sea cliffs at or near sea level where it lies stratigraphically beneath the Whidbey Fm. of the last interglaciation (Easterbrook, 1968, 1969, 1986; Easterbrook et al., 1967).

Thermoluminesence (TL) dating of clay beneath Double Bluff Drift has yielded ages of 289 ± 74 ka and 291 ± 86 ka. A TL date of 177 ± 38 ka was obtained from Double Bluff glaciomarine drift, and a date of 320 ± 100 ka was obtained from clay associated with Double Bluff till east of here (Berger and Easterbrook, 1993). Amino–acid ages from mollusk shells in Double Bluff glaciomarine drift range from 111 to 178 ka at the type locality and from 150 to 250 ka elsewhere in the central Puget Lowland (Easterbrook et al., 1982; Blunt et al., 1987).

Whidbey Interglaciation. The type section of the Whidbey Fm. consists of silt and sand, interbedded with, clay, peat, and widely scattered gravel exposed in the ~2–km–long sea cliffs between Double Bluff and Useless Bay (Fig. 2) (Easterbrook et al., 1967). The Whidbey Fm. is underlain by Double Bluff Drift at the west end of the bluffs. Along the sea cliffs in the eastern part of

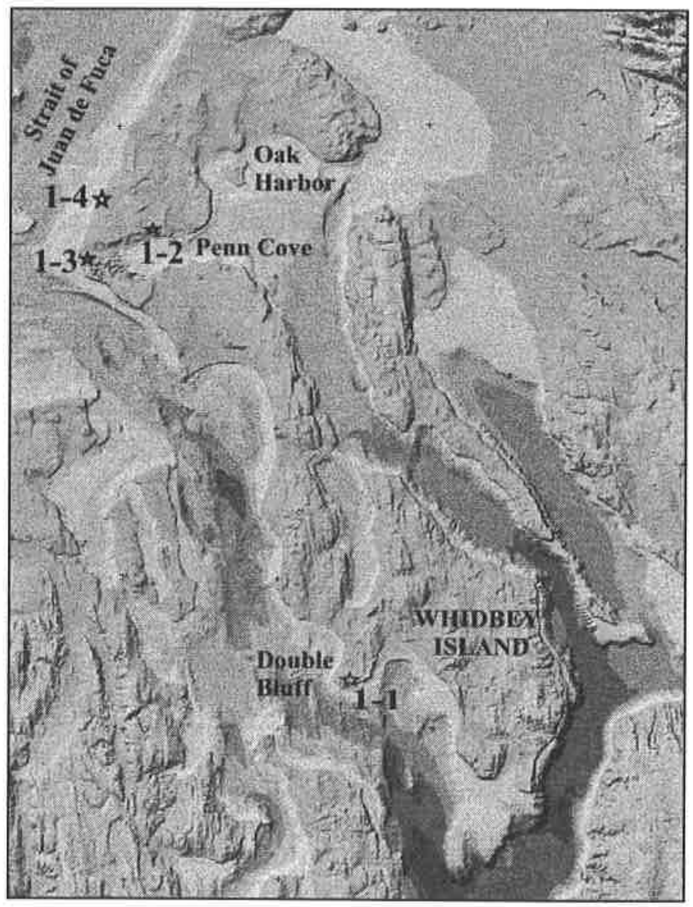

Figure 2. Whidbey Island showing field trip stops. (Modified from shaded digital topographic model by PRISM, University of Washingon.

Easterbrook, D.J., 2003, Cordilleran Ice Sheet glaciation of the Puget Lowland and Columbia Plateau and alpine glaciation of the North Cascade Range, Washington: *in* Easterbrook, D.J., ed., Quaternary Geology of the United States, INQUA 2003 Field Guide Volume, Desert Research Institute, Reno, NV, p. 265–286.

the section, Whidbey sediments are unconformably overlain by Olympia lacustrine silt , ^{14}C–dated at 23,600 ± 275 and 22,210 ± 530 ^{14}C yrs. B.P. (Stoffel, 1980), Vashon advance outwash and till, and Everson glaciomarine drift (Easterbrook, 1968).

Hansen and Mackin (1949) suggested that the Whidbey sediments originated as floodplain deposits formed from "very slow aggradation by meandering streams flanked by floodplain lakes and swamps" in which silty clay and peat were deposited. Pollen studies suggest that the Whidbey Fm. was an interglacial period with a climate somewhat warmer than the present (Hansen and Mackin, 1949; Hansen and Easterbrook, 1974; Heusser and Heusser, 1981).

More than 20 ^{14}C ages beyond the range of ^{14}C dating have been obtained from peat in the Whidbey Fm. Amino-acid ages of 97 ± 35, 96 ± 35 ka and 107 ± 9 ka have been obtained from shells in Whidbey marine sediments to the north (Easterbrook, 1968; Blunt et al., 1987). TL ages measured on clay from Whidbey fluvial sediments are shown below. (Berger and Easterbrook, 1993; Easterbrook, 1994).

TL Age (ka)	Locality
102 ± 38	Lagoon Point, Whidbey Island
106 ± 17	Blowers Bluff, Whidbey Island
142 ± 10	Northern West Beach, Whidbey Island
151 ± 43	Point Wilson, NW of Port Townsend

On route to STOP 1–2, return to Freeland and drive north on Highway 525 (which becomes Highway 20 northward) to the north side of Penn Cove. Turn right off Highway 20 at San de Fuca onto Penn Cove Road and continue to Penn Cove Park. Walk along the beach from here to Blowers Bluff.

STOP 1–2. BLOWERS BLUFF; POSSESSION DRIFT OVER WHIDBEY SILT/SAND; EVERSON GMD

Possession Glaciation. Massive, cliff-making, Possession till and glaciomarine drift overlie peat-bearing sand and silt of the Whidbey Fm. at mid-cliff and are overlain by Vashon advance outwash, till, and fossiliferous Everson gmd. (Easterbrook, 1968).

Amino-acid analyses of shells in massive, poorly sorted, till–like, Possession glaciomarine drift suggest a mean age of 80 ± 22 ka (Blunt and others, 1987; Easterbrook and Rutter, 1981, 1982).

Olympia Interglaciation. Nonglacial floodplain and lacustrine silt, clay, and peat, deposited in the lowland between the Possession and Fraser glaciations, provide the basis for definition of the Olympia Interglaciation (Armstrong et al., 1965) (also referred to as the Olympia nonglacial interval). Radiocarbon dates from Olympia sediments just east of here range from 22-28 ka ^{14}C yrs. B.P. (Hansen and Easterbrook, 1974; Easterbrook, 1976a, 1986; Johnson et al., 1996, Johnson et al., 2002).

Fraser Glaciation. The late Wisconsin Fraser Glaciation consists of several stades; (1) the Evans Creek Stade, an early alpine phase; (2) the Vashon Stade, the maximum advance of the Cordilleran Ice Sheet (CIS) (3) the Everson Interstade, an interval of deposition of glaciomarine drift (gmd) during deglaclation of the lowland; and (4) the Sumas Stade, consisting of several short readvances of the ice terminus before complete deglaciation (Easterbrook, 1963, 1969, 1992, Armstrong et al., 1965; Kovanen and Easterbrook, 2002a; Kovanen, 2002a).

Vashon Stade. The last major climatic episode during which drift was deposited by the CIS is defined as the Vashon Stade (Armstrong et al., 1965). It began with advance of the CIS into the lowland and ended with the beginning of marine and glaciomarine conditions of the Everson Interstade. Ahead of the advancing glacier, meltwater streams deposited an apron of outwash sand and gravel that was later overridden by the ice sheet and covered with till. Timing of the advance is limited by ^{14}C dates from sediments beneath Vashon till. These dates indicate that Vashon ice advanced southward across the Canadian border sometime after 17-18 ka ^{14}C yrs. B.P. and reached the Seattle area soon after 14.5 ka ^{14}C yrs. B.P. (Mullineaux et al., 1965; Deeter, 1979; Easterbrook, 1992; Porter and Swanson, 1998).

Grooves and striations carved on bedrock and elongate drumloidal hills and flutes indicate directions of movement of Vashon ice (Easterbrook, 1979). At its maximum, the Vashon ice extended to approximately 25 km (15 mi.) south of Olympia (Fig. 3). Ice was more than 1,830 m. (6,000 ft) thick near Bellingham and about 915 m. (3,000 ft) thick near Seattle.

Everson Interstade. The CIS underwent sudden, large–scale recession and downwasting about 14.5 ka ^{14}C yrs. B.P. The Puget lobe retreated northward from its southern terminus by backwasting past the Seattle area by about 14 ka ^{14}C yrs. B.P. (Fig. 4) (Rigg and Gould, 1957; Leopold et al., 1982). Rapid thinning of Vashon ice after the terminus had receded north of Seattle allowed marine water from the Strait of Juan de Fuca to flood the lowland (Fig. 4), floating the remaining ice and quickly disintegrating the remaining CIS all the way north to Canada. During this interval, Everson glaciomarine drift (gmd) was deposited over a large area in the central and northern Puget Lowland (Easterbrook, 1963, 1966a,b, 1968, 1969, 1971, 1976a,b, 1979, 1986, 1992, 2003; Dethier et al., 1995; Kovanen and Easterbrook, 2002a; Kovanen, 2002) and British Columbia (Armstrong and Brown, 1954; Armstrong, 1981, 1984; Armstrong and Hicock, 1980a, b).

The gmd consist largely of poorly sorted diamictons deposited when melting of floating ice released rock debris that accumulated on the sea floor. Unbroken, articulated, marine shells, some in growth positions, indicate that the gmd represents *in situ* deposition (Easterbrook, 1963, 1992). The origin of the gmd from floating ice in sea water was first recognized in Washington by Easterbrook (1963, 1966a, 1966b, 1969) and in British Columbia by Armstrong (1957, 1960, 1981, 1984). Since then, more than 150 ^{14}C dates from Washington and British Columbia fix the age of the Everson glaciomarine drift at 11,500 to 13,500 ^{14}C yrs. B.P., (Fig. 5) (Easterbrook, 1992, 2003), making it a valuable stratigraphic marker over the central and northern Puget Lowland.

Return to Highway 20 and drive west. Turn right on the Libby Road to and continue to West Beach.

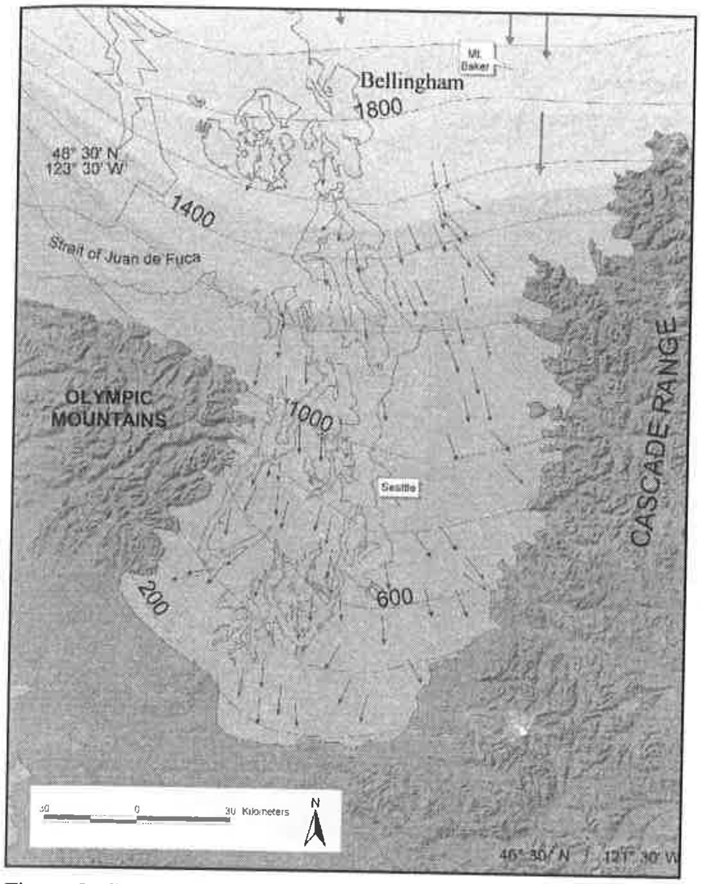

Figure 3. Shaded digital topographic model reconstruction of the Cordilleran Ice Sheet at its late Wisconsin maximum. (D.J. Kovanen and M. Price)

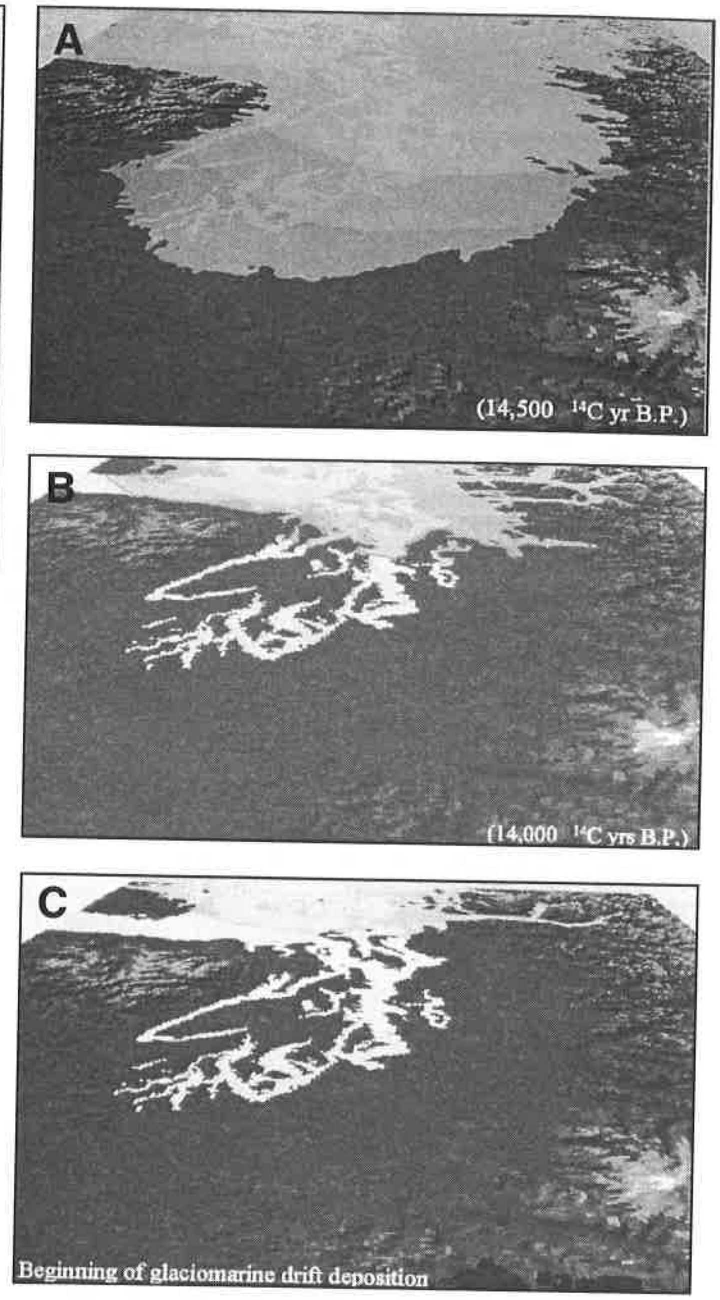

Figure 4. Retreat of the Cordilleran Ice Sheet. (Shaded digital topographic models by D. J. Kovanen and M. Price)

STOP 1–3. DEGLACIATION, EVERSON GLACIOMARINE DRIFT CHRONOLOGY, WEST BEACH

Sea cliffs just north of the park at West Beach consist of fossiliferous Everson gmd, dated at 12,535 ± 300 ^{14}C yrs. B.P. (uncorrected) (Easterbrook, 1968). Applying the new marine reservoir correction for this region (Kovanen and Easterbrook, 2002b) gives a corrected age of 11,400 ^{14}C yrs. B.P.

The first recognition and dating of glaciomarine drift on Whidbey and nearby islands in the 1960s were supported by 11 ^{14}C dates (Easterbrook, 1966a,b, 1968, 1969), and in the intervening years, 140 additional ^{14}C dates were obtained (Fig. 5). The inescapable conclusion from Figure 5, is that the distribution of dates indicates that glaciomarine conditions were essentially contemporaneous over the whole region and that the age of final emergence was about 11,500 ^{14}C yrs. B.P. for the entire area. Thus, the progressive, backwasting, calving terminus concept of Domack (1983) cannot be correct.

Notwithstanding all of these ^{14}C dates, Swanson and Caffee (2001) contend that deposition of the Everson ended 13,500 ^{14}C yrs. B.P., fully 2,000 years earlier. They ignore the 150 dates shown on Figure 5 and based their conclusion on only four ^{14}C dates. The significant implication of this is that they use this time as the basis for calculating a production rate of ^{36}Cl, which they then use for measuring the ages of erratic boulders in the region. Because the base of their production rate calculations is off by 2000 years, their cosmogenic dates are off by 20% (Easterbrook, 2003).

The surface topography in the Partridge Point area is deeply indented with kettles (Fig. 6), many >30 m deep. Just south of West Beach Park, wave erosion has breached a 30–m–deep ket-

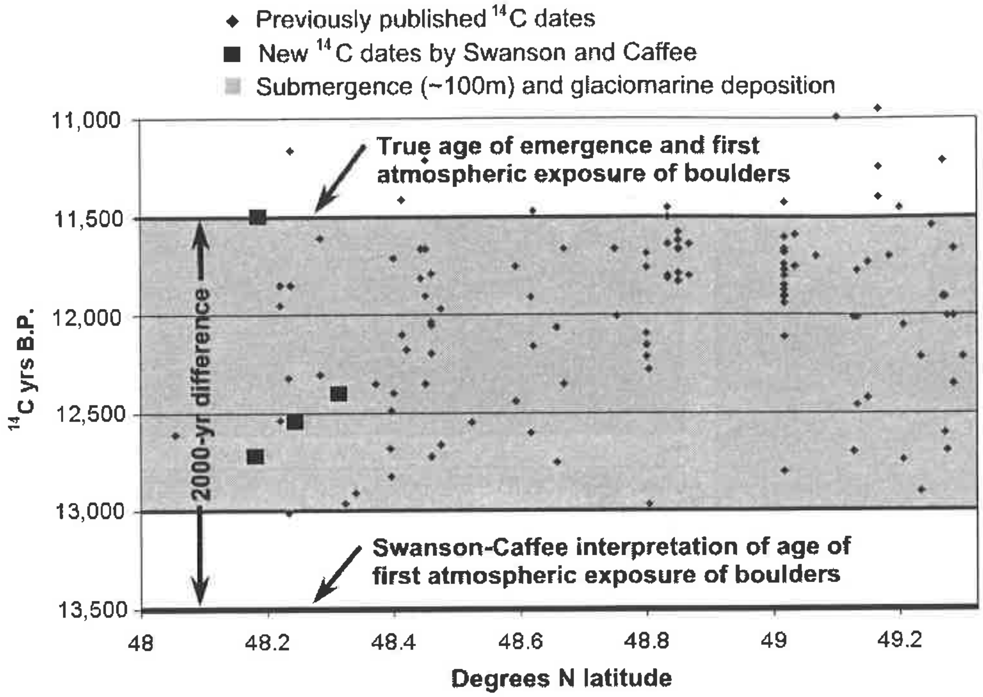

Figure 5. Chronology of Everson glaciomarine drift deposition. (Easterbrook 2003)

tle, exposing the Partridge Point gravel that makes up the kettled topography. Collapse structures are common around the margins of the kettles where sediments slumped into the cavities made by melting of the buried blocks of ice. The gravel typically has strong foreset bedding, with dips up to 20–25°, and contains pumice clasts and many small shell fragments (Easterbrook, 1966a,b, 1968; Carlstad, 1992).

Figure 6. 30m–deep kettles in west–dipping deltaic gravel at Partridge Pt. (Photo by D.A. Rahm, courtesy of EPIC)

The top of the Partridge Point gravel, where not indented by kettles, consists of a flat surface about 55–60 m above present sea level. Relict meltwater channels on this surface are shown on the remarkably detailed LIDAR imagery rendered by Kovanen (Fig. 8). Everson gmd caps the Partridge Point gravel in the bluffs just south of the park (Easterbrook, 1968).

SEA–LEVEL CHANGES, ISOSTATIC MOVEMENTS, AND ICE–FLOW INDICATORS ON WHIDBEY AND CAMANO ISLANDS, PUGET LOWLAND: GEOLOGIC DATA FROM LIDAR–AIDED OBSERVATIONS

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Glacial and marine features

High–resolution, LIDAR (Light Detection and Ranging), digital data of Whidbey and Camano Islands reveal preliminary geologic evidence of former sea–levels and ice–flow directions (Fig. 7A). Raised shorelines in this area are found up to c. 88 m above present sea level and are believed to record glacio–isostatic movements related, in part, to glacial rebound (as well as eustatic and tectonic processes which operate on different spatial and temporal scales) of the Cordilleran Ice Sheet, which inundated the Puget Lowland during the last glaciation.

The timing of ice retreat on central Whidbey Island is constrained between 13,650 ± 350 to 11,850 ± 240 ^{14}C yr B.P. by the

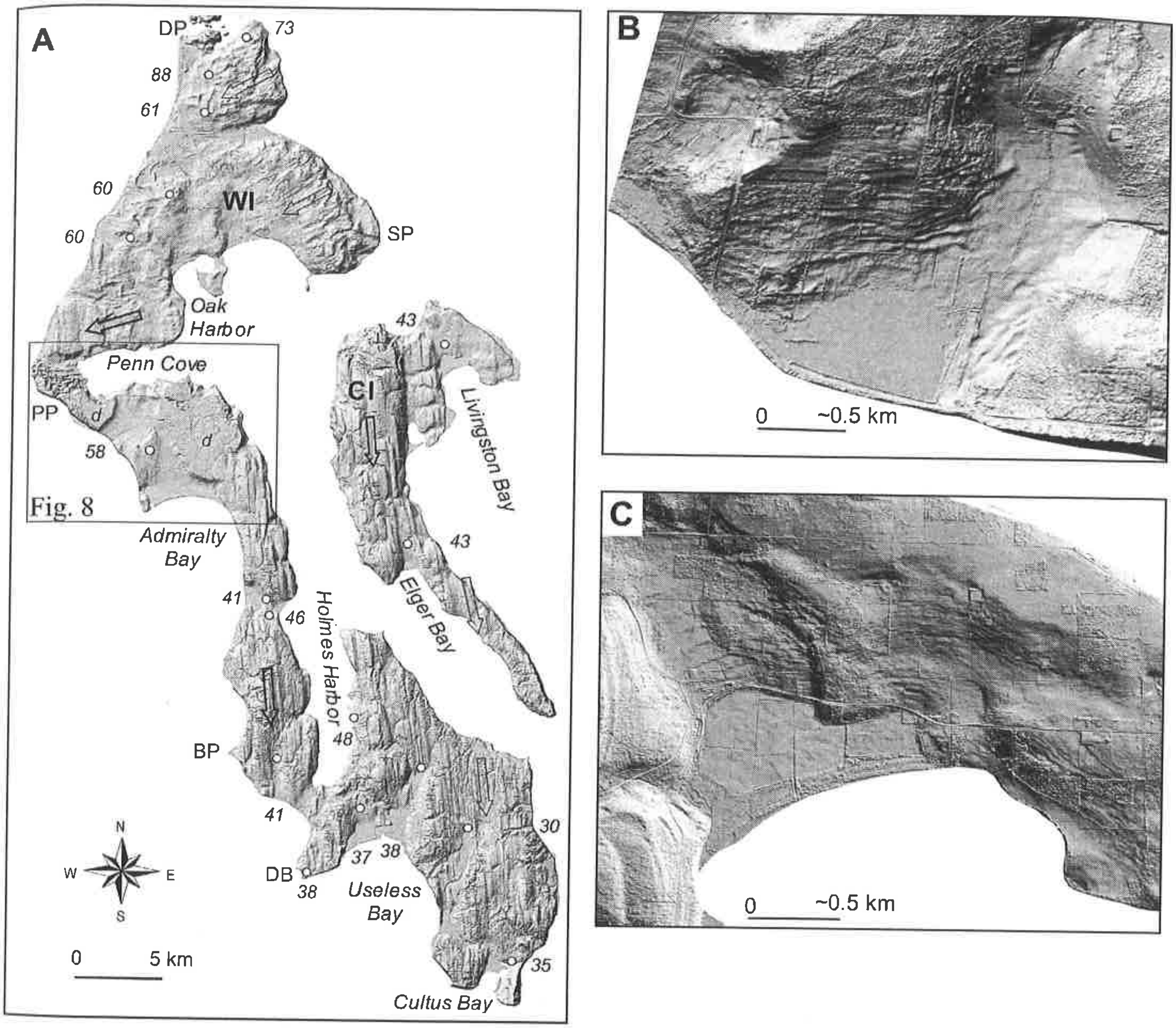

Figure 7. (A) Shaded digital topographic model of Whidbey and Camano Islands from LIDAR data (2 m resolution). White-centered dots indicate the location of raised shoreline sequences and the corresponding number is the approximate upper elevation (m). Arrows indicate inferred ice flow directions and "d"indicates the location of raised deltas. WI–Whidbey Island; CI–Camano Island; DP–Deception Pass; SP–Strawberry Point; PP–Partridge Point; BP–Bush Point; DB–Double Bluff. (B) Raised shoreline sequence northwest of Oak Harbor. (C) Raised shoreline sequence at Livingston Bay.

ages of marine bivalves in glaciomarine sediment (Easterbrook, 1969; Domack, 1983; Dethier et al., 1995; Swanson and Caffee, 2001). Correcting these shell dates for a radiocarbon marine reservoir value of –800 years (Robinson and Thompson, 1981; Southon et al., 1990), yields an age range of ~12,850–11,050 ^{14}C yr B.P. for deglaciation, which may be conservative (see Kovanen and Easterbrook, 2002b). Changes in relative sea level occurred rapidly following deglaciation. The shoreline sequences shown in Figure 7B and 7C have not yet been directly dated and are the subject of future work. Close examination of many of the gullies and raised deltas reveal that they are commonly graded to, and truncated at, an elevation that is equal to the highest, local, raised shoreline elevation.

Superimposed bedforms (i.e., drumlins, flutes, till lineations) record two flow lines with different orientation, suggesting more than one ice–flow event (Fig. 7A): (1) An "older" event with a north–south, ice–flow direction (Vashon advance?) and (2) a relatively younger event on northern Whidbey with a northeast–

southwest ice–flow direction (during deglaciation). These overlap the north–south ice–flow indicators and therefore did not form contemporaneously. A well–defined, marine, ice–margin exists along the south and west side of Penn Cove (Fig. 8) (see also Hewitt and Mosher, 2001 for paleoenvironments of the eastern Juan de Fuca Strait during the last deglaciation).

Significance and conclusions

The use of high–resolution LIDAR data allow the first recognition of raised shoreline sequences on Whidbey and Camano Islands. The preliminary marine–limit data are generally consistent with the maximum elevation of shell–bearing stony muds interpreted as glaciomarine drift (Easterbrook, 1968, 1992) and postglacial isostatic rebound (i.e., Thorson, 1989) that occurred late in the regional deglacial history. Based on the many closely–spaced shoreline sequences, the argument is made that the land response to transient ice loading was dominantly elastic and rebound is accomplished immediately upon unloading. In this active tectonic regime (Johnson et al., 1996, 2002), the relations between isostatic depression, postglacial rebound, and postglacial eustatic sea level rise are complex. These sites (Fig. 7A) reflect important aspects of uplift and the geologic controls, geomorphic processes, landscape evolution, and environmental change during the late Pleistocene and early Holocene. Future work includes establishing a detailed geochronology of these environmental changes so that they can be compared with local and regional variations in sea level (i.e., Easterbrook, 1963, 1992; Dethier et al., 1995) and with curves of glaciated North America (Dyke and Peltier, 2000) to better understand the geographic patterns of sea level changes during deglaciation.

ACKNOWLEDGEMENTS

The Puget Sound LIDAR Consortium (especially Jerry Harless and Diana Martinez) is thanked for providing the LIDAR data, as well as timely and kind assistance.

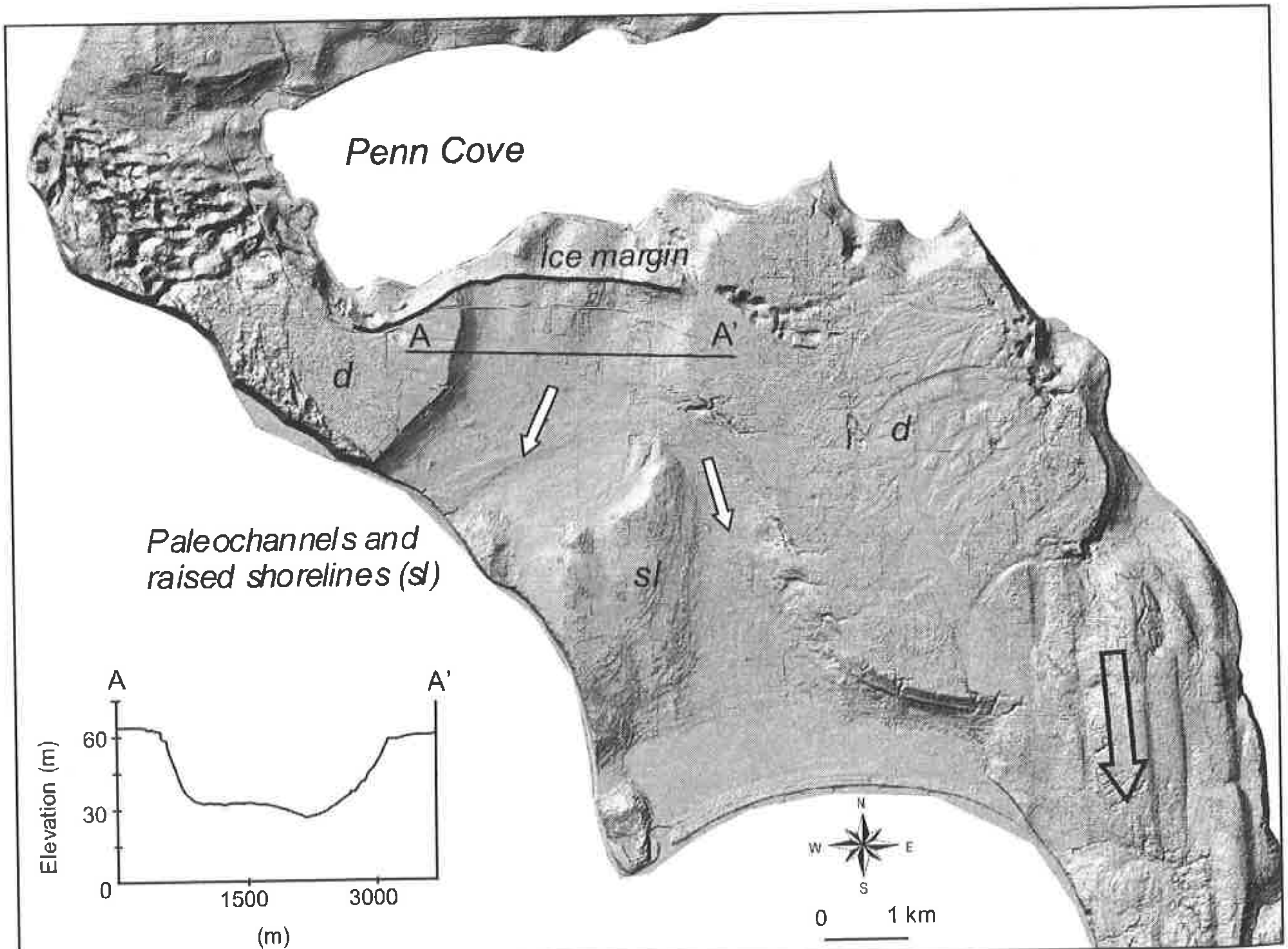

Figure 8. Shaded digital topographic model of the Penn Cove area showing the major geomorphic features; paleochannels (white arrows), raised deltas (d), late glacial ice margin, and raised shoreline sequences (SL). The inferred ice–flow direction in the early stages of Vashon glaciation is north–south.

STOP 1–4. LATE PLEISTOCENE RAISED SHORELINES

The remarkable detail shown on the LIDAR imagery derived by Kovanen (previous section) has opened a new chapter in the understanding of late Pleistocene glacial events and relative sea level changes. Except for the delta surfaces at Penn Cove, the raised shorelines (Fig. 7B,C) that she has discovered on the LIDAR imagery have never been observed before. Also striking is the dramatic shift in ice-flow direction northeast of Penn Cove (Fig. 7A).

That Penn Cove was occupied by an ice-mass that shed sediment into a marine delta and onto stagnant buried ice at Partridge Point has long been recognized (Easterbrook 1966, 1968; Carlstad, 1992), but the details of the ice margin and raised delta surfaces (Fig. 8) are now nicely shown on the LIDAR imagery. The change in ice-flow direction from N-S to NE-SW, shown so vividly on the LIDAR images, means that as the Vashon glacier thinned rapidly during deglaciation and began to disintegrate in the Strait of Juan de Fuca and northward, a remnant of ice was grounded on the eastern part of the lowland long enough to build the features seen near Penn Cove. Because it was no longer directly fed by ice flow from the north, the grounded remnant ice assumed a southwesterly flow direction transverse to the Vashon direction (Fig. 7A), terminating in the sea at the Penn Cove deltas.

Close inspection of the raised shorelines on the LIDAR images shows a fairly regular step-like progression from the highest down to present sea level. Thus, isostatic rebound in this region seems not to have been constant, but rather as a series of short upward jogs, leaving a shoreline at each one.

Drive to the north end the island. Turn right and drive east to I–5. Turn left and drive north on I–5 toward Bellingham.

STOP 1–5. SUMAS OUTWASH CHANNEL INCISED INTO EVERSON GMD ACROSS SHORELINES

Sumas Stade. Emergence of the northern lowland from the sea and readvance of the CIS over its own outwash defines the beginning of the Sumas Stade (Armstrong et al., 1965). Four Sumas phases, spanning the Younger Dryas and part of the Allerød, have now been recognized (Easterbrook and Kovanen, 1998; Kovanen and Easterbrook, 2002a; Kovanen, 2002). We will see the field evidence and chronology of these on DAY 2.

At the Samish River, gmd extends to elevations of 65 m, about the same as the highest shorelines in the area, and is deeply incised by the Samish River, which is badly underfit (Fig. 9). It flows with present–day meander radii of about 0.15 km (0.1 mi.) on a valley floor having meander radii >1.7 km (1 mi.). A much larger stream than the present Samish River obviously made the Samish valley. The meanders are incised into Everson gmd and cut across beach ridges at the marine limit of 65 m (215 ft.) before terminating at a marine delta about 30 m (100 ft.) above present sea level (Easterbrook, 1979, 1994). The outwash was graded to a sea level ~35 m lower than that of the gmd. The outwash channel and terraces belong to the Sumas Stade, C^{14}–dated elsewhere at about 10–11,400 ^{14}C yrs. B.P. A limiting age for similar strandlines nearby is given by a date of 11,700 ± 110 ^{14}C yrs. B.P. from basal peat at 47 m (Siegfried, 1978).

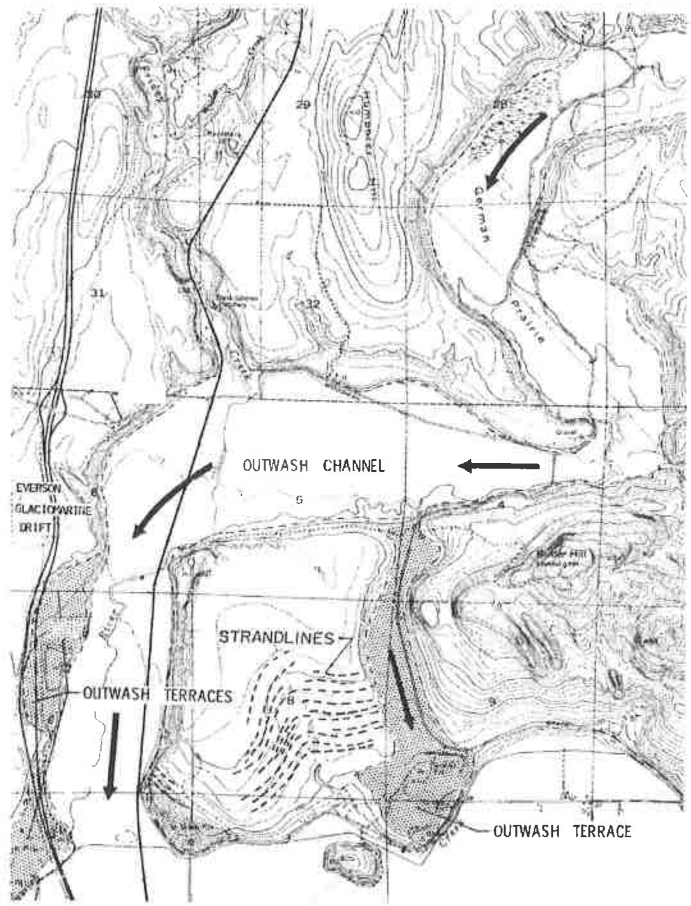

Figure 9. Sumas outwash channel cut into Everson glaciomarine drift and across late Everson shorelines.

Arrive in Bellingham. End of DAY 1.

DAY 2 EVERSON RELATIVE SEA LEVEL CHANGES AND SUMAS READVANCES

EVERSON INTERSTADE

Everson glaciomarine drift (gmd) was deposited from floating ice in the Fraser Lowland (Fig. 10) to elevations up to 180 m. (600 ft) above present sea level and extended more than 10 km (6 mi.) up the Nooksack valley into the North Cascades (Fig. 11) (Easterbrook, 1963, 1992, 1994; Kovanen and Easterbrook, 2001).

The Everson in the Fraser Lowland consists of two fossiliferous glaciomarine drifts, the Kulshan (lower) and Bellingham

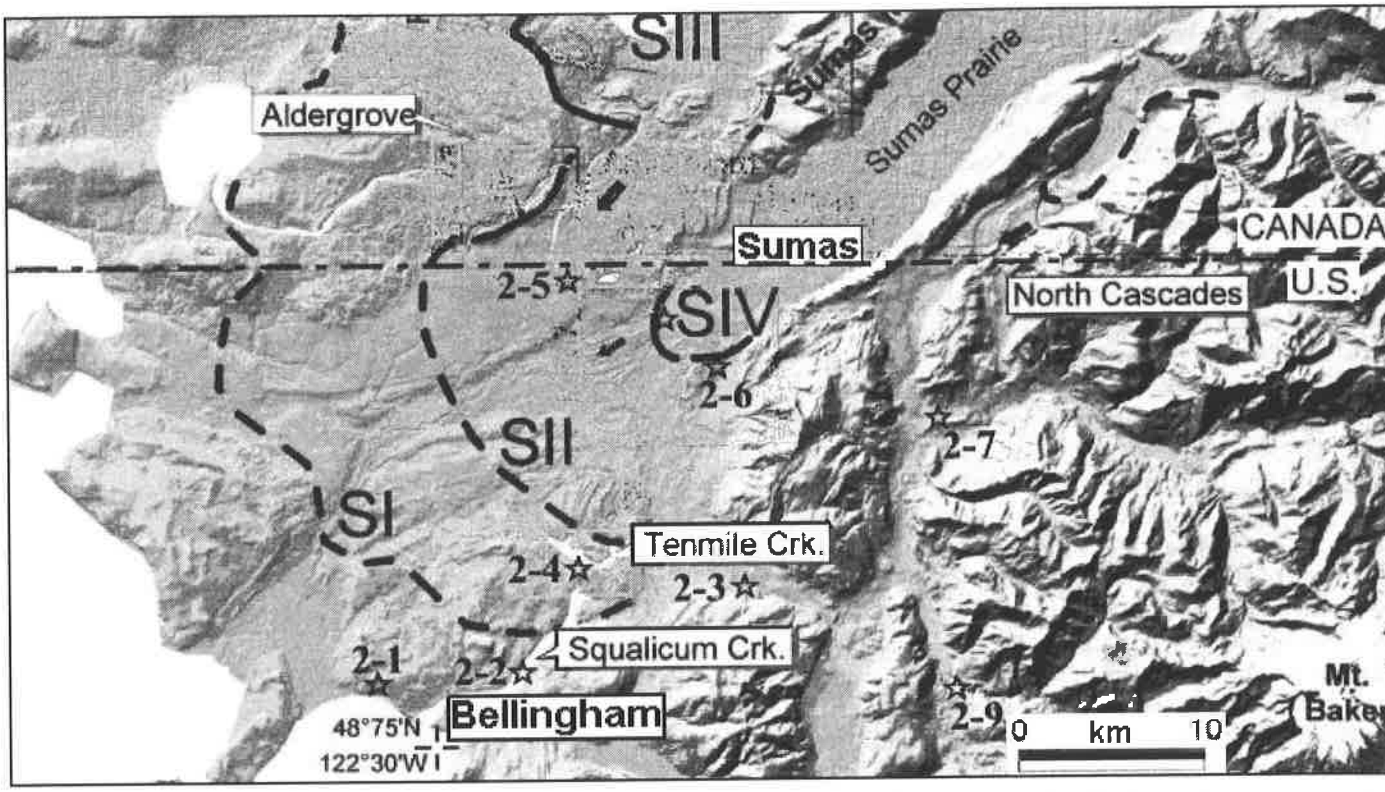

Figure 10. Location of stops in the Fraser Lowland and position of ice margins of the four Sumas phases. (Modified from Kovanen and Easterbrook, 2002a)

(upper), separated by subaerial, fluvial, Deming sand with paleochannels, crossbedding, *in situ* tree stumps (Fig. 12), and basal peat lacking pollen from any salt-tolerant plants. The sandwiching of the non–marine Deming sand between the two glaciomarine drifts demonstrates extraordinary fluctuations of relative sea level in this area over short periods of time. Relative sea level stood at ~90 m (300 ft) above present level during deposition of the Kulshan gmd, dropped to within 8-12 m (25-40 ft.) of present sea level, then rose again to more than 180 m (600 ft.), and finally dropped to present sea level by 11,400 ^{14}C yrs. B.P. during the Sumas Stade (Fig. 11) (Easterbrook, 1963, 1992; Easterbrook and Kovanen, 1996; Kovanen and Easterbrook, 2002a). Sumas outwash channels that cut deeply into Bellingham gmd were graded to a sea level close to that of the present, indicating that the 180 m (600 ft.) of post-Bellingham-gmd emergence took place quickly.

Clearly, a complicated combination of isostatic, eustatic, and tectonic events is needed to explain the ups and downs of relative sea level spanning about 2000 years in the Everson Interstade (Easterbrook 1963, 1992). They cannot be explained by any combination of only isostatic and eustatic changes because of submergences that occurred during a times of isostatic uplift. A strong local tectonic influence is required to adequately explain these late Pleistocene relative sea level changes. Interestingly, analysis of recent seismic records in this area show a remarkable pattern (more than 400 earthquakes since 1969).

SUMAS STADE

Following deposition of the Everson drift, the Cordilleran Ice Sheet readvanced several times during the Sumas Stade, depositing moraines, till, ice-contact sediments, and outwash (Easterbrook, 1963, 1969, 1971, 1976a, 1986, 1992). This readvance reversed the wholesale deterioration of the ice sheet that began shortly before ~14,500 ^{14}C–yrs B.P. The Sumas Stade began with emergence of the area from the Everson sea and readvance of the Cordilleran Ice Sheet (Armstrong et al., 1965). The multiple advances of Sumas ice have been documented by Easterbrook (1994), Easterbrook and Kovanen, (1998), Kovanen and Easterbrook, 2002a, Kovanen (2002), and Mark and Ojamma (1980).

Multiple moraines are associated with oscillations of the Cordilleran Ice Sheet in the Fraser Lowland. The paleogeography of the ice margin and timing of ice retreat during the Sumas Stade is reconstructed and bracketed by 70 radiocarbon dates, which are secured by morphologic and stratigraphic evidence. Four topographically distinct phases of the Sumas readvances have been documented (Fig. 13) (Easterbrook and Kovenan, 1998; Kovanen and Easterbrook, 2002a; Kovanen, 2002).

Leave WWU campus, drive northeast on Garden St., and turn left (northwest) on Holly St. Holly St. becomes Marine Drive to the north. Turn left on an unnamed secondary road north of

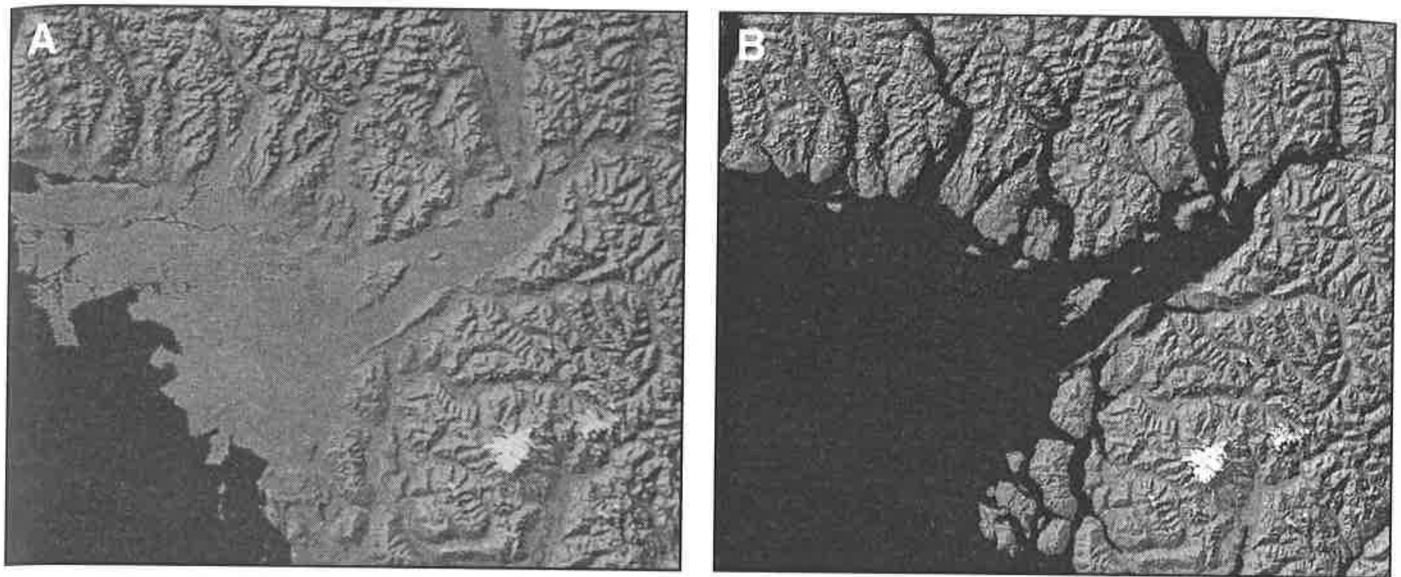

Figure 11. A. Present day shoreline. B. Everson shoreline at 180 m. (Shaded digital topographic model by D.J. Kovanen)

the old cement plant. Walk down the trail to the beach and proceed northward up the beach to sea cliff exposures.

STOP 2-1. BELLINGHAM GLACIOMARINE DRIFT OVERLYING BEACH DEPOSITS OF DEMING SAND, BELLINGHAM BAY

The purpose of this stop is to examine the Kulshan-Deming-Bellingham sequence and establish sea level during deposition of the Deming beach deposits. The sea cliff exposes Bellingham glaciomarine drift (gmd) lying on beach deposits of the Deming sand (Fig. 14) (Easterbrook, 1963). Kulshan gmd is exposed beneath Deming sand near beach level to the south.

The Bellingham gmd consists of stony silt/clay containing marine fossils and makes up the surface topography in the upland to the east where it extends to elevations up to 180 m (600 ft.) along the Cascade foothills. Six ^{14}C dates from wood and marine shells in the gmd range from 11,580 to 11,830 ^{14}C yrs. B.P. and marine fossils in beach deposits of the Deming sand here were dated at 11,685 ± 85 and 11,760 ± 85 (Kovanen and Easterbrook, 2002a). The fossiliferous Deming beach deposits range in elevation from 7.5 m (25 ft.) to 20 m (65 ft.) and establish sea level limits for the Deming sand. Evidence for the beach origin of these deposits is documented in Easterbrook (1963) and Weber (2001). Fossiliferous Kulshan gmd, exposed in bluffs just above the modern beach and in the wave platform seaward from the bluffs, underlies the Deming sand. Shells from the Kulshan have been dated at 12,210 ± 82 ^{14}C yrs. B.P.

The significance of the Kulshan-Deming-Bellingham sequence is that it replicates the same sequence exposed at the Everson type section (STOP 2–2) and establishes the position of sea level for the Deming sand between the two gmds.

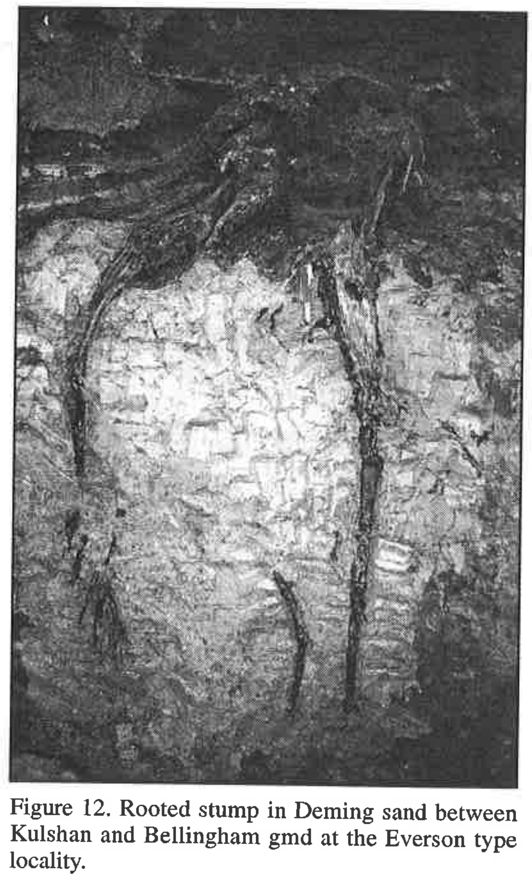

Figure 12. Rooted stump in Deming sand between Kulshan and Bellingham gmd at the Everson type locality.

Drive east on Sunset Drive (Mt. Baker Highway). Turn left and drive north on the Hannegan Road. Across the Squalicum outwash channel (Sumas I)

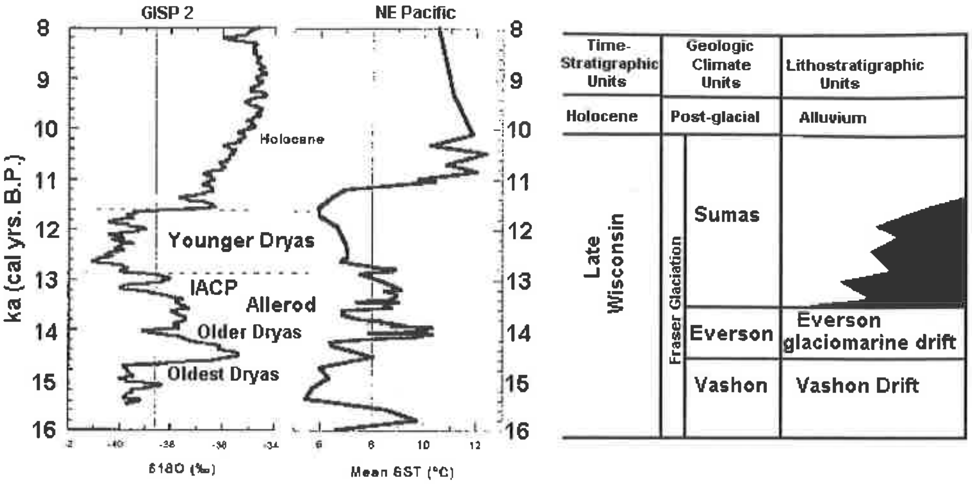

Figure 13. Correlation of the chronology of Sumas phases with the GISP2 ice core and sea surface paleo–temperatures in the northeast Pacific. (From Kovanen and Easterbrook, 2002a)

STOP 2-2

This outwash channel (Fig. 10) originated just to the east at a brief stand of the earliest phase of the Sumas Stade (Sumas I). The channel incises Bellingham gmd and is thus post–11,600 ^{14}C yrs. B.P. and is older than Sumas II (11,400 ^{14}C yrs. B.P.). No well–defined moraine occurs at the head of the channel, it just pinches out at the position of the glacier terminus.

Turn right on Van Wyck Road, cross Squalicum outwash channel, and continue east on the Mt. Baker Highway.

STOP 2-3. EVERSON INTERSTADE TYPE LOCALITY

The type locality of the Everson Interstade (Armstrong and others, 1965) includes fluvial Deming sand sandwiched between Kulshan and Bellingham gmd (Easterbrook, 1963, 1992) (Fig. 15), demonstrating at least two submergences and two emergences of the region in only 1500–2000 years (the "*yo–yo*" effect). The base of the section consists of 5 m (15 ft.) of stratified, crossbedded, pebbly sand interbedded with thin, silt beds (Fig. 15), probably fluvial overbank deposits. A twig and a thin, *in situ*, organic mat a few mm thick interbedded with silty-clay were dated at 12,070 ± 80 and 12, 185 ± 80 ^{14}C yrs. B.P., providing a lower limiting age for the Kulshan gmd and indicating that this area was above sea level before it was submerged to deposit Kulshan gmd. The significance of this is that it demonstrates that neither the Kulshan nor the Bellingham gmds could have resulted from a retreating, calving Vashon ice sheet.

Kulshan gmd consists of 9–12 m (30-40 ft.) of poorly sorted pebbly silt and clay containing widely scattered marine shells that overlies the basal sand. No ^{14}C dates have been obtained form the Kulshan here, but shells in the same unit 6 km (3.7 mi.) to the east were dated at 11,870 ± 380 years. The top of the Kulshan here is about 75 m (250 ft.) above sea level. Adding another 30 m for probable water depth gives a sea level about 100 m (350 ft.) higher than present sea level for the Kulshan.

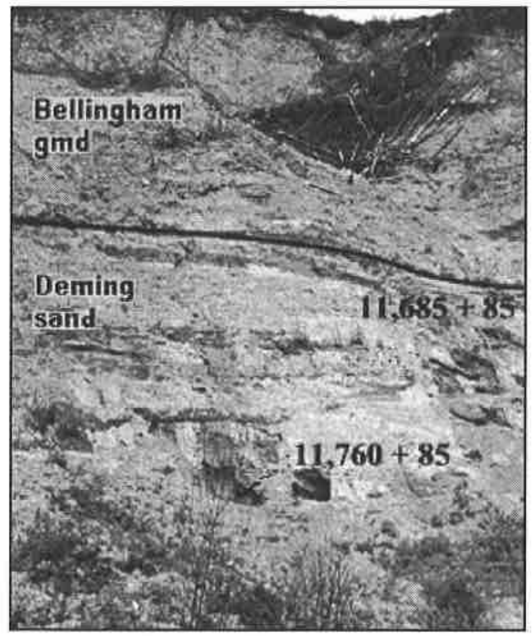

Figure 14. Bellingham glaciomarine drift overlying Deming sand at Bellingham Bay.

About 10 m (33 ft.) of Deming sand overlies the Kulshan gmd (Figs. 15, 16). It consists mainly of crossbedded sand with some interbedded silt Prominent channeling may be seen in bluff exposures. A peat bed about 20-30 cm (8-12 in.) thick at the contact between the Kulshan and Deming units contains *in situ* rooted stumps (Fig. 12). Additional evidence of the nonmarine origin of the Deming sand has been documented in Easterbrook (1963, 1992).

Beach deposits in Deming sand at Bellingham Bay (STOP 2–1) indicate that sea level at this time was about 25–65 ft. above present sea level. ^{14}C ages from the Deming sand at

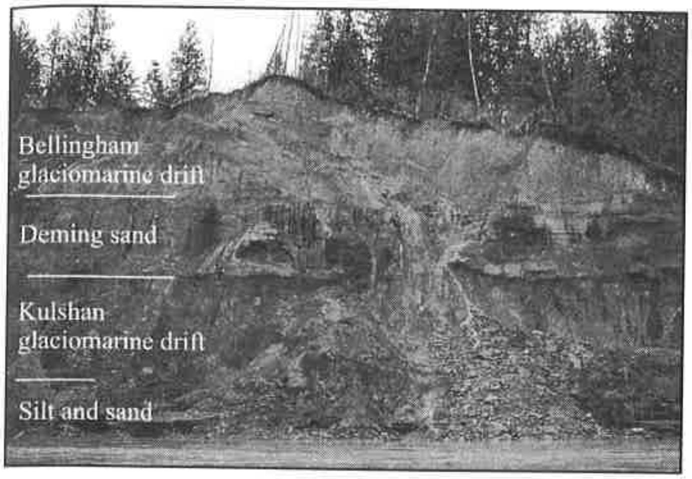

Figure 15. Type section of the Everson glaciomarine drift.

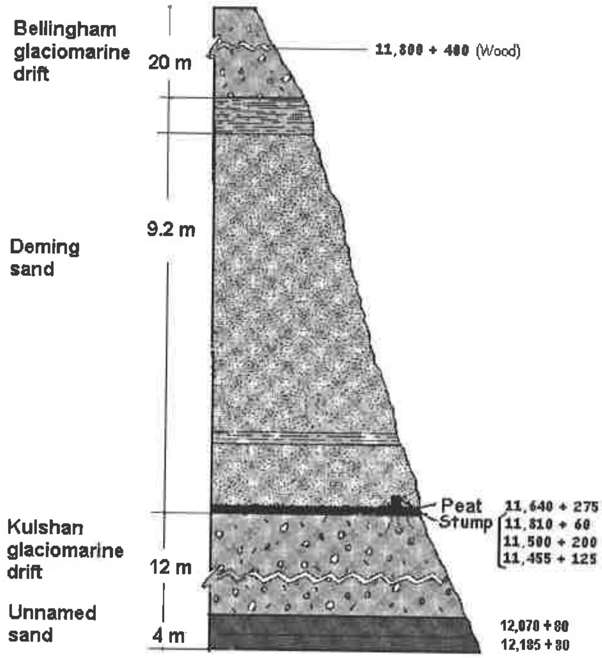

Figure 16. Stratigraphic section at the type locality of the Everson glaciomarine drift.

Bellingham Bay are 11,685 ± 85 and 11, 760 ± 85 ^{14}C yrs. B.P., and the rooted stump at the Everson type locality is 11,810 ±60 ^{14}C yrs. B.P. This means that the Deming sand at the Everson type locality must have been deposited ~60 m (200 ft,) above sea level at that time.

Deming sand is overlain by about 20 m of massive, poorly sorted, Bellingham gmd. Marine fossils, mostly *Nuculana* and *Clinocardium*, are more abundant than in the Kulshan gmd, but are widely scattered. An allocthanous piece of wood in the middle portion of the Bellingham gmd was dated at 11,800 ± 400 years (Easterbrook, 1963). Outcrops of Bellingham gmd extend to elevations of 180–210 m (600–690 ft.) above present sea level, and adding another 30 m for probable water depth gives a relative sea level of 210-240 m (700-800 ft.) above present sea level, a relative submergence of >200 m (>660 ft.) from Deming to Bellingham time. A similar 'sandwich' of fluvial sediment between two gmds occurs in British Columbia (Fig. 17), demonstrating that this is a regional feature. Ten ^{14}C dates from the Bradner pit in B.C. range from 11,600 to 11,800 ^{14}C yrs. B.P., almost identical to the ages from the Everson type locality and Bellingham Bay.

Post-Bellingham gmd relative sea level dropped abruptly about 11,500 ^{14}C yrs. B.P. Sumas outwash, graded to a sea level close to that of the present, was then deposited on Bellingham gmd and outwash channels were incised into the gmd, indicating yet another abrupt relative sea level change of about 200 m. Remarkable as they may seem, these ups and downs of relative sea level appear to be irrefutable–the physical evidence for them can be seen in the single riverbank exposure at the Everson type locality. No correlation of strata or extrapolation of data is required.

Return to Mt. Baker Highway and drive west to the Smith Road intersection. Turn right and continue to the Starry Road. Turn right and drive 1.7 km (1 mi.) to the Kulshan Golf Club on the right. Pull into the parking lot.

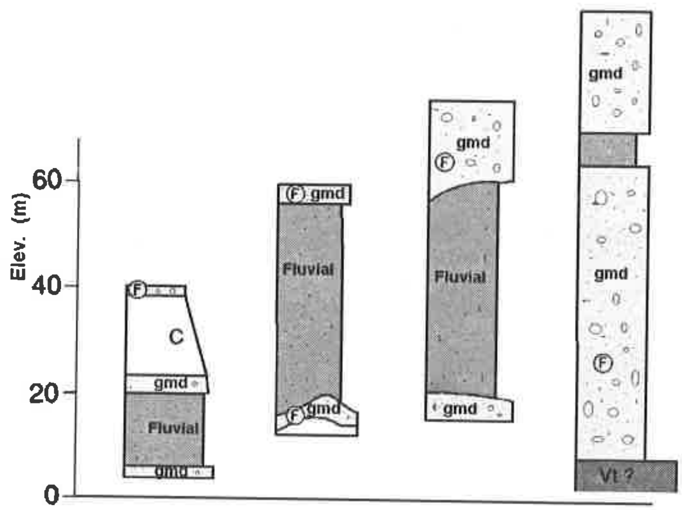

Figure 17. Fluvial sand between two glaciomarine drifts in southwestern B.C. The fluvial sand in B.C. contains rooted stumps that have the same ^{14}C age as the Deming at the type locality and at Bellingham Bay. (Modified from Armstrong, 1984)

STOP 2-4. OVERLOOK OF TENMILE CREEK SUMAS II OUTWASH CHANNEL.

The purpose of this stop is to verify that deep, Sumas, meltwater channels incise the Bellingham gmd, indicating that emergence closely followed deposition of the Bellingham gmd and providing a lower limiting age for the Sumas II phase. The Sumas outwash channel is incised 33m (100 ft.), exposing Bellingham gmd overlying Deming sand in the valley walls. Basal peat from a 6-9 m-(20-30 ft.)-thick bog that floors the channel was dated at 11,080 ± 100 and 11,113 ± 88 ^{14}C yrs. B.P. (Fig. 18). These dates correlate well with the age of a prominent Sumas II moraine in southwestern British Columbia. A basal bog date of 10,400 ± 85 ^{14}C yrs. B.P. was obtained from Fazon Lake, a large kettle just to the north.

Drive west on Axton Road, turn right on Hannegan Road. Cross peat-filled Sumas IV outwash channels, continue through Lynden to near the Canadian border.

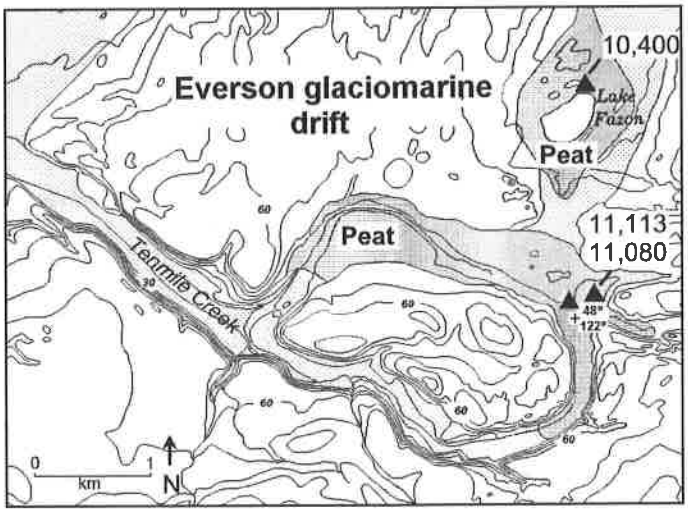

Figure 18. Sumas II outwash channel incised ~30 m in Bellingham gmd and filled with peat. (Kovanen and Easterbrook, 2002a)

Figure 19. Sumas II moraine at the Canadian border (road on the left side). (Photo by D.A. Rahm courtesy of EPIC)

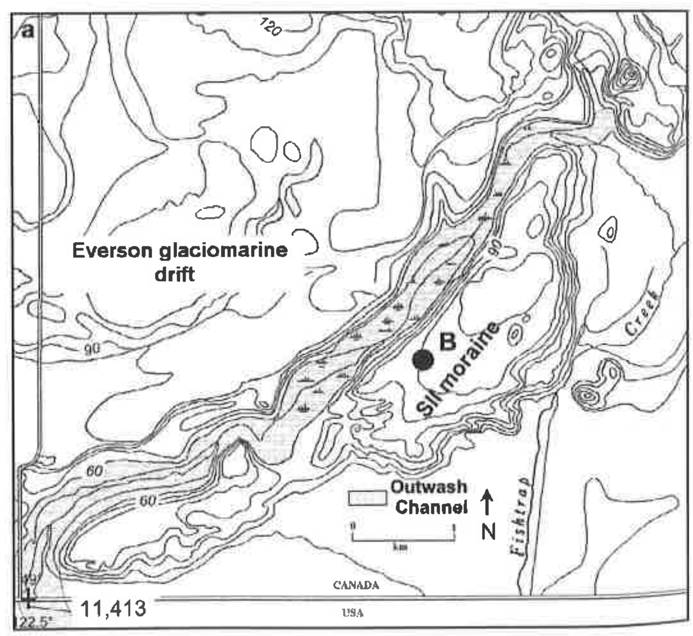

Figure 20. Sumas II moraine at the Canadian border. (Kovanen and Easterbrook, 2002a)

STOP 2-5. SUMAS II MORAINE, SUMAS III OUTWASH PLAIN, ^{14}C LIMITING DATES

Just across the border is an elongate, tapering, Sumas II moraine (Figs. 19, 20) (Armstrong, 1981; Kovanen and Easterbrook, 2002a, Kovanen, 2002). The moraine rests on Everson gmd from which many ^{14}C dates on wood and shells have been obtained, clearly limiting the Sumas II advance at younger than 11,600 . ^{14}C yrs. B.P.

A ^{14}C date of 11,413 ± 75 ^{14}C yrs. B.P. from basal peat in an outwash channel from the moraine (Fig. 20) indicates that the glacier had retreated back from the end moraine by 11.4 ka, cutting off meltwater flow in this channel. A ^{14}C date of 10,980 ± 250 ^{14}C yrs. B.P. from basal peat in a channel *behind* the moraine indicates that the glacier had receded to the Sumas III moraines to the north by this time.

We are presently on the broad Abbotsford outwash plain that heads at Sumas III moraines farther north (Fig. 21). The significance of the position of the outwash plain is that lies behind the Sumas II moraine (Fig. 21b) and must therefore have retreated 20 km to the Sumas III moraines. ^{14}C dates of basal bogs on the outwash plain indicate that the Sumas III phase must be Younger Dryas.

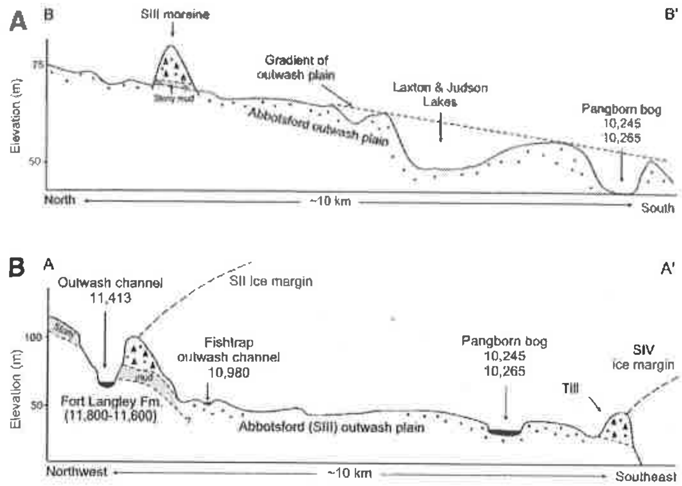

Figure 21. Profiles of the Sumas moraines. (Kovanen and Easterbrook, 2002a) A. N–S profile from Sumas SIII moraines to the Abbotsford outwash plain. B. E–W profile from the Sumas SII moraine to the Sumas SIV moraine.

Drive east on Halverstick Road past Pangborn bog to Clearbrook (Sumas IV moraine across Sumas III outwash channel).
Turn right off Halverstick Road onto road to Clearbrook, then to Van Buren Road. Turn left on Frick Road and take road across the front of the Sumas IV moraine.

STOP 2–6. SUMAS IV MORAINE AND OUTWASH TERRACE

The Sumas IV moraine consists of a hummocky, linear ridge that extends in an east-west direction away from the mountains for about 3.5 km (2.2 mi.) at the head of an outwash plain that slopes southward away from the moraine (Fig. 22).. Erosion has removed it from the central part of the Sumas Valley, but a remnant occurs on the west side of the valley, filling the distal end of a Sumas III meltwater channel (Fig. 22), showing that it represents a readvance of the glacier. This is the final phase of the Sumas advances and represents a late Younger Dryas readvance of the CIS.

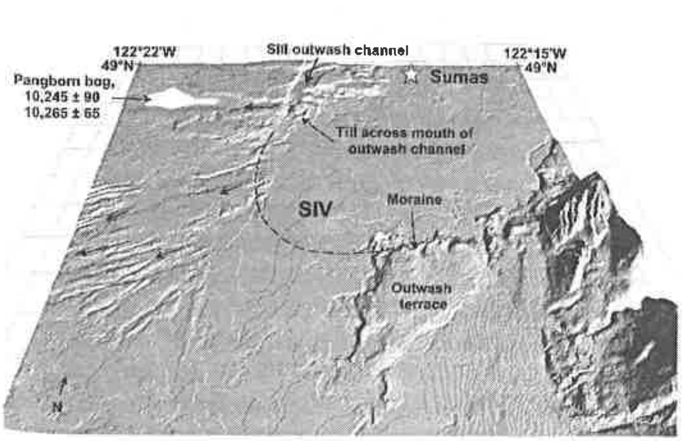

Figure 22. Sumas IV moraine and outwash terrace. (Kovanen and Easterbrook, 2002a).

Continue east on Telegraph Road-Easterbrook Road to Sumas. Turn right on road to Columbia Valley, drive to Kendall.

STOP 2–7. KENDALL MORAINE AND OUTWASH TERRACE

After the CIS had disappeared from the North Cascades, long valley glaciers, heading on Mt. Baker and fed by tributary glaciers, occupied the Nooksack drainage system (Figs. 23, 24) (Kovanen and Easterbrook, 2001). The purpose of this stop is to examine evidence for extension of a post-CIS, Cascade, alpine glacier ~45 km downvalley from its sources on Mt. Shuksan and Mt. Baker. The Kendall moraine forms a distinct ridge several hundred meters long whose axis is transverse to the valley (Fig. 25). It is composed of ~12 m (40 ft.) of till containing abundant, glacially-faceted, striated boulders and cobbles derived from the upper North Fork valley. The upper North Fork has a distinct U–shaped cross-profile at right angles to the direction of flow of the CIS.

The number of well-faceted and striated cobbles and boulders in the till is striking, probably because the valley glacier was so long. The lithology of cobble and boulders in the Kendall moraine is dominated by Chuckanut sandstone (54%), Mt. Baker andesite (12%), and graywacke/greenstone (14%)

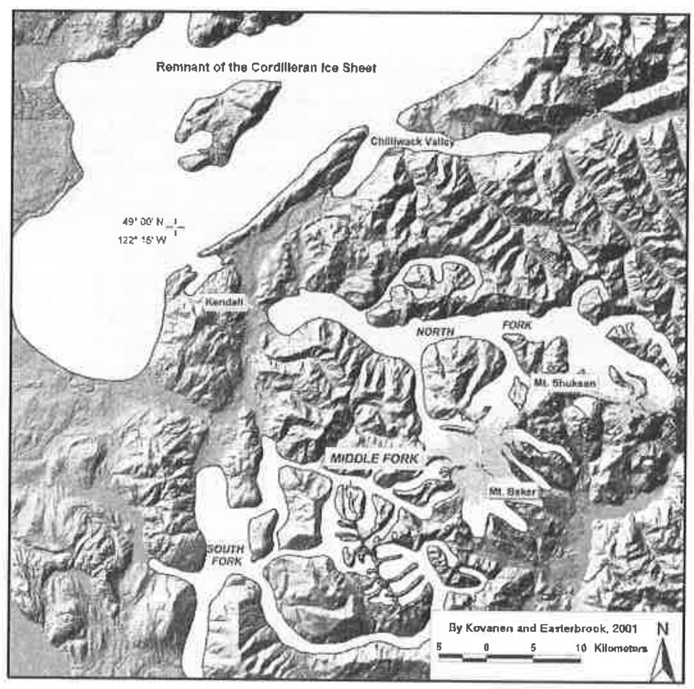

Figure 23. Reconstruction of Nooksack Alpine Glacier System. (Kovanen and Easterbrook, 2001)

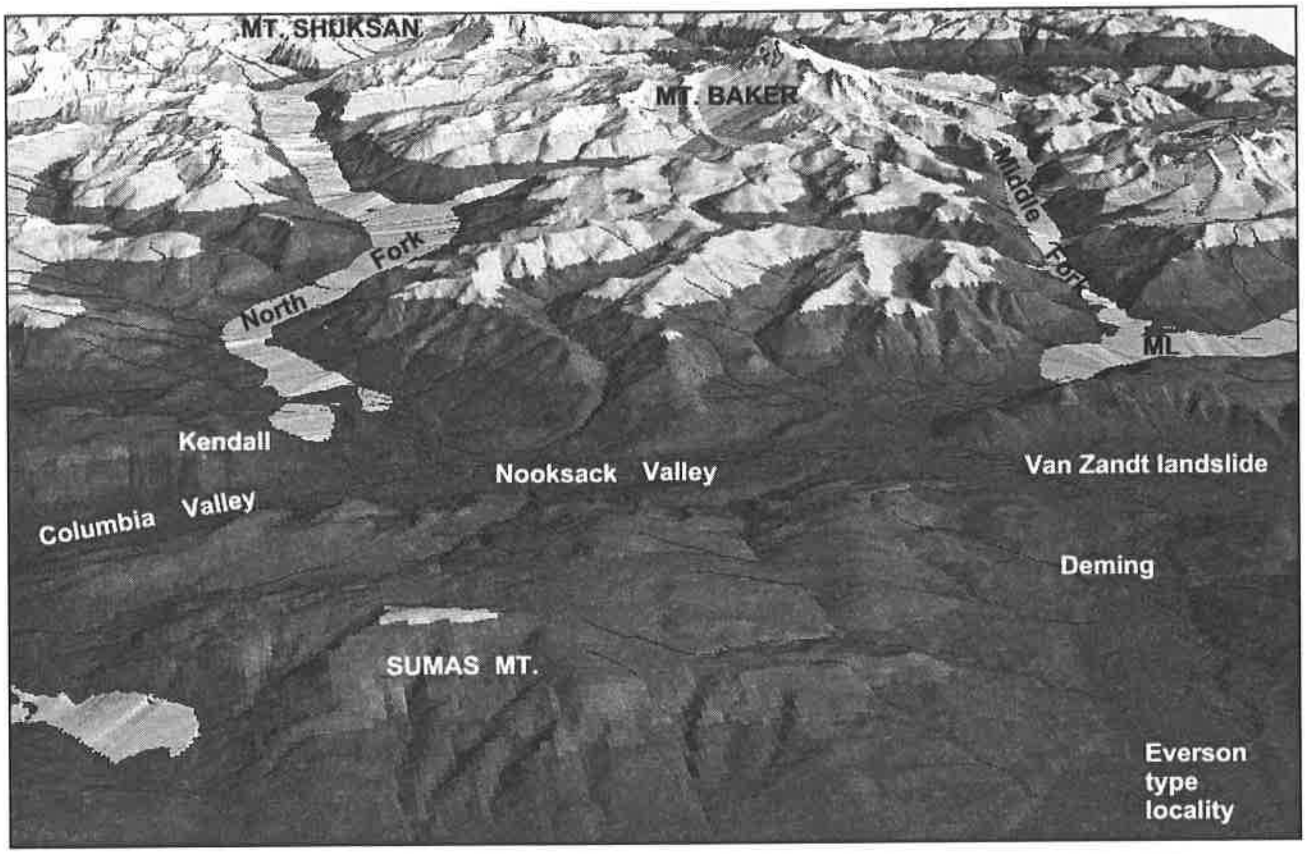

Figure 24. . Reconstruction of post–CIS alpine glaciers in the North Cascades. (DEM image by D.J. Kovanen).

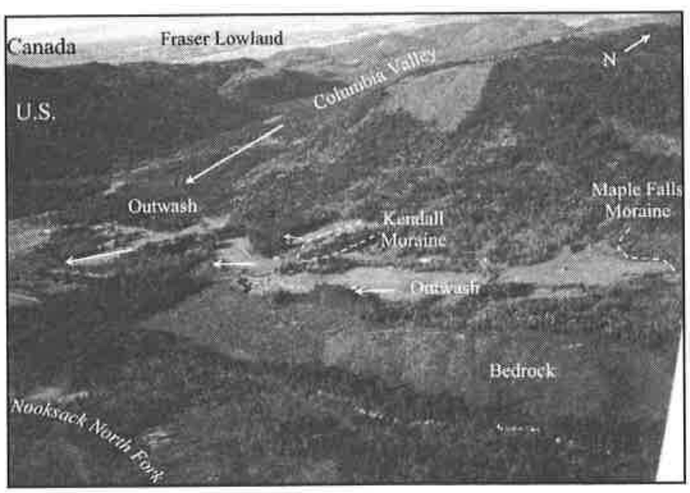

Figure 25. The Kendall moraine and outwash terrace. (From Kovanen and Easterbrook, 2001).

(presumably from the Jurassic Nooksack Group) (Kovanen and Easterbrook, 2001). These lithologies demonstrate that the source of ice was from the North Fork valley to the east, rather than from the CIS to the north.

The Mt. Baker andesite was probably contributed from Mt. Baker via Bagley Creek, Glacier Creek and Wells Creek. Chuckanut sandstone was derived from the valley sides immediately upvalley, and the graywacke/ greenstone came from the valley sides near the town of Glacier.

Not far to the west, outwash from the Kendall moraine and the Maple Falls moraine just upvalley merge with the outwash terrace in the Columbia Valley. A few km downvalley, the outwash terrace rests on gmd dated at 11,910 ± 80 ^{14}C yrs. B.P. Distinct charcoal layers interbedded with sandy outwash terrace gravel gave dates of 10,788 ± 77 and 10,603 ± 69 ^{14}C yrs. B.P. (Fig. 26), indicating a Younger Dryas age of the outwash (Kovanen and Easterbrook, 2001).

Time and weather permitting, continue east on Mt. Baker Highway to Heather Meadows.

STOP 2–8 (OPTIONAL) ARTISTS POINT AND HEATHER MEADOWS, SCENIC VIEWS OF MT. BAKER, MT. SHUKSAN, AND NORTH CASCADES; CORDILLERAN ICE SHEET GROOVES; HOLOCENE CIRQUE GLACIATION; MAZAMA ASH

Table Mt., just west of the Artists Point parking lot, is an andesite flow that has become topographically inverted by erosion. Erratics of Paleozoic greenstone from Mt. Herman, the next peak to the north, indicates that glacial ice rode over Mt. Herman

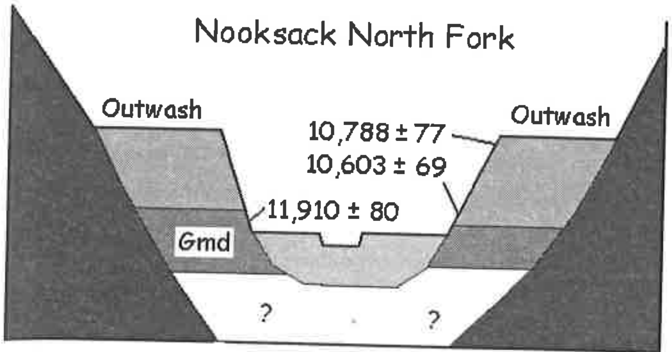

Figure 26. Cross section through the North Fork Sumas outwash

and deposited greenstone erratics on Table Mt. at an elevation of 1725 m (5700 ft.). Adding a few hundred feet for ice thickness means that the surface of the CIS here must have been more than 1800 m (6000 ft.). From the parking lot, walk eastward along the trail to Artists Point and note the beautifully developed grooves and striations carved into the andesite by the CIS. Significantly, all of the grooves are oriented north-south, indicating the direction of ice flow. Thus, the CIS did <u>not</u> flow westward from here down the Nooksack drainage toward the lowland, it flowed almost due south down the axis of the range.
Return to Picture Lake at Heather Meadows.

Picture Lake is contained by a moraine deposited by a cirque glacier. Mazama ash is exposed in a road cut beside the lake.

Return west on Mt. Baker Highway to Welcome. Turn left onto Mosquito Lake Road, cross Nooksack River, stop at overview of Deep Kettle bog.

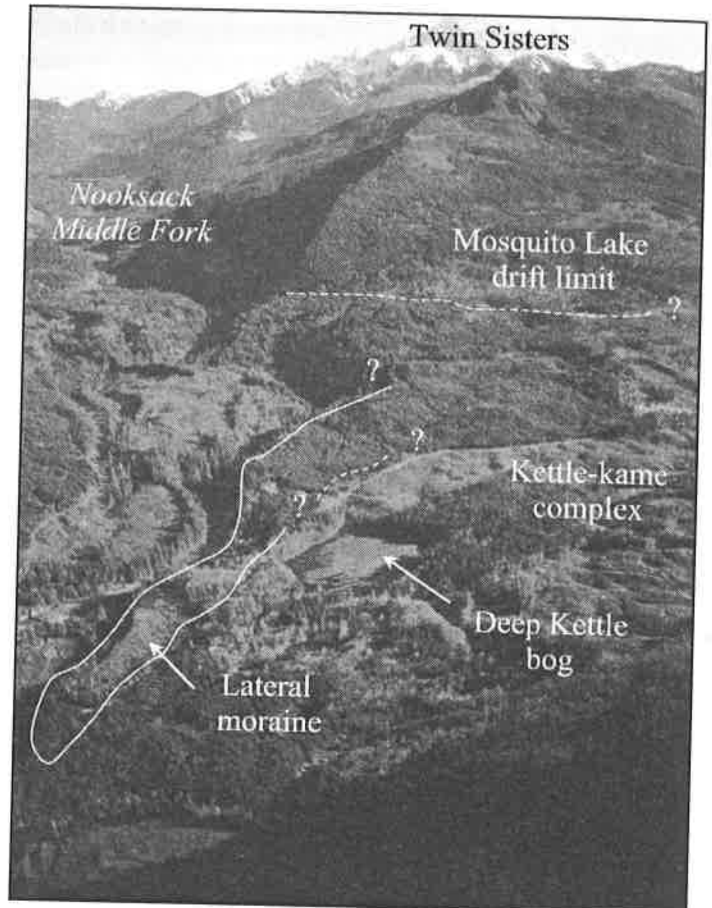

Figure 27. Ice–contact drift, Deep Kettle bog, and lateral moraine, Nooksack Middle Fork. (Kovanen and Easterbrook, 2001)

STOP 2-9. DEEP KETTLE BOG, MOSQUITO LAKE KETTLE-KAME COMPLEX AND LATERAL MORAINE

The purpose of this stop is to examine evidence for the alpine ice origin of a post–CIS lateral moraine and kettle-kame complex. The kettle–kame complex is composed of poorly-to-moderately-sorted sand and gravel containing many glacially faceted and striated boulders and cobles. The lithology of the boulders and cobbles is dominated by Mt. Baker andesite (17.4%), dunite from the Twin Sisters Range (16.6%), and Chuckanut sandstone (45.2%) from upvalley, demonstrating that the glacier that transported them must have originated in the headwaters of the Nooksack Middle Fork (Kovanen and Easterbrook, 1996a, 2001).

The age of the kettle-kame complex is known from a 20 m core from a deep bog in a kettle (Fig. 27). Rootlets in gravel at the base of the bog were ^{14}C–dated at 12,356 ± 115 ^{14}C yrs. B.P. and had to be growing in rock debris on stagnant ice before the ice melted out to make the kettle. A date on wood lying on the basal gravel was 12,165 ± 115 ^{14}C yrs. B.P. (Kovanen and Easterbrook, 2001).

A lateral moraine drapes across the kettle–kame complex (Fig. 27), indicating that ice in the Middle Fork Valley readvanced following the stagnation that produced the kettle–kame complex. The lithology of the moraine is dominated by glacially faceted and striated Mt. Baker andesite and Twin Sisters dunite, proving that it was deposited by an alpine glacier from the Middle Fork. The age of the moraine must be younger than the 12,000 year–old kettle–kame complex because it overlies it, both morphologically and stratigraphically.

The strong dominance of glacially faceted boulders and cobbles from the Middle Fork headwaters 23 km (14.3 mi.) upvalley demonstrates that the kettle–kame complex and the lateral moraines were deposited by true alpine glaciers, not by remnants of the CIS and must therefore postdate the CIS. These alpine glaciers could not have been fed by Cordilleran ice because of the lithology of the drift and, as shown by gmd in the Nooksack drainage, the area to the north was free of Cordilleran ice at that time.

A shallow landslide exposes till of the lateral moraine overlying the ice-contact sediments of the kettle-kame complex, giving stratigraphic, as well as morphologic, evidence of the relationship of the lateral moraine to the kettle-kame complex. The lithology of the till is dominated by glacially faceted and striated Mt. Baker andesite and Twin Sisters dunite. The lateral moraine demonstrates that the Middle Fork alpine glacier readvanced over its previously–deposited kettle–kame complex.

Upvalley, logs in a lateral moraine have been dated at 10,680 ± 70 and 10,500 ± 70 ^{14}C yrs. B.P. (Kovanen and Easterbrook, 2001).

Continue west to I-5, turn right, return to Bellingham.

DAY 3. GLACIATION OF THE NORTH CASCADES

Depart Bellingham, drive 58 mi. (97 km) south on I–5 to Everett. Exit I–5 onto Highway 2, drive to Monroe, then turn right (south) to Snoqualmie falls.

ICE–MARGINAL FEATURES OF THE VASHON GLACIER.

The eastern margin of the Cordilleran Ice Sheet was banked against the Cascade Range as it flowed southward down the Puget Lowland. The glacier extended up major Cascade valleys, blocking drainage and impounding ice–marginal lakes into which large deltas were built (Cary and Carlston, 1937; Mackin, 1941; Booth, 1987). An ice–marginal embankment was constructed in the Snoqualmie valley near North Bend. It consists of a high, flat–topped ridge that extends across the Snoqualmie River valley. Deltaic sediments, consisting mostly of coarse sand and gravel, grading upvalley to sand, then to silt and clay, were deposited in an ice–dammed lake upvalley from the Snoqualmie embankment. Wood from rhythmically–bedded (varved?) lake sediments upvalley was ^{14}C dated at 13,570 ± 130 (UW–35) years. As the ice retreated to the north and west, the newly–opened, ice–free valleys held ice–marginal lakes that became successively lower as new outlets were uncovered (Mackin, 1941).

STOP 3–1. TOKUL CREEK DELTA AND SNOQUALMIE FALLS.

As the Puget lobe retreated northward from north–sloping valleys, an ice–marginal lake formed in the Snoqualmie valley south of the ice margin. Tokul Creek, a tributary of the Snoqualmie River brought large quantities of sand and gravel into the lake, building a delta that extended nearly all the way across the valley. As the ice front retreated northward, new, lower outlets were uncovered and the lake eventually drained. When the Snoqualmie River re–established its course, the delta sediments forced the river to the west side of the valley at the distal edge of delta. As the Snoqualmie cut down through these sediments, it encountered the bedrock of the west valley wall, and, once notched into the rock, couldn't get out. Downstream, the river quickly eroded the valley–filling sediments, but where the river was notched into the bedrock, it could not erode its bed as rapidly, and Snoqualmie falls were born. As the falls retreat upvalley, they will eventually erode through the bedrock spur and into the valley fill upstream. When this happens, the river will quickly cut down through the sediments and bring an end to the falls.

Continue through North Bend, turn left on I–90 and drive 22 mi. (37 km) to Snoqualmie Pass. Exit I–90 to the ski area.

STOP 3–2. LATEST PLEISTOCENE ALPINE MORAINES

Multiple, as-yet-undated, post-LGM moraines occur at Snoqualmie Pass in the North Cascades (Porter, 1978). Half a dozen, closely-spaced moraines at Snoqualmie Pass, well upvalley from the LGM, record several periods of glacial retreat and stillstand. No numerical ages have been published for any of these moraines, although a limiting date of 11,050 ± 50 ^{14}C yrs. B.P. was obtained from wood in gravel on till between two of the moraines (Porter, 1978).

Rejoin I–90 and drive east to Exit 70 to the Easton Campground for Lunch.
Drive east 16 mi. (27 km) on I–90 to Exit 85 to Blewett Pass. Contine 9 mi. (15 km) to Swauk Prairie.

STOP 3–3. SWAUK PRAIRIE MORAINE

Long valley glaciers extended down several tributaries of the Yakima River. At this stop we will examine the oldest moraines of this system.

Drive 37 mi. (62 km) to the junction with Highway 2. Turn left and continue 6 mi. (10 km) to Leavenworth. At the west end of town, turn left up Icicle Creek road and continue 6 mi. (10 km) to Rat Creek. Pull into the campground on the left.

STOP 3–4. YOUNGER DRYAS (?) MORAINE AT RAT CREEK

A double moraine in Icicle Creek (Page, 1939) from an alpine glacier in Rat Creek near Leavenworth has been correlated with the Sumas Stade of the CIS, which has been shown to include the YD (Kovanen and Easterbrook, 2002a). However, no numerical dates have yet been published for these moraines and this correlation is based on the position of the moraines upvalley from LGM moraines.

Return to Leavenworth, turn right on Highway 2 and drive 26 mi. (43 km) to Wenatchee. Exit to the right.

DAY–4. GLACIAL LANDFORMS OF THE OKANOGAN LOBE OF THE CORDILLERAN ICE SHEET; MISSOULA FLOOD FEATURES

The Cordilleran Ice Sheet extended down both sides of the North Cascades, overwhelming the crest of the range and connecting the Okanogan lobe with the Puget Lowland ice across the north Cascades. At the late Wisconsin maximum, the Okanogan lobe reached about 50 km south of the Columbia River onto the Columbia Plateau where it built a massive end moraine and left a virtual museum of glacial landforms beautifully preserved in the arid climate. Drumlins and flutes show the radial movement of ice, and eskers, kames, kame terraces, and recessional moraines were left during ice recession. The Okanogan lobe played an important role in directing giant floods from the bursting of ice dams and draining of glacial Lake Missoula. Glacial landforms are intimately associated with Missoula flood features.

GLACIAL GEOMORPHOLOGY AND ICE–FLOW INDICATORS OF THE OKANOGAN LOBE OF THE CORDILLERAN ICE SHEET: AN ARCHIVE OF GLACIAL FEATURES

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The Okanogan lobe flowed southward through the Okanogan trough from the Interior Plateau of British Columbia and invaded the Waterville Plateau during the late Wisconsin (Fraser Glaciation). This had a profound influence on the landscape and meltwater routing and flooding events in adjacent areas (e.g., Pardee, 1922; Bretz, 1959; Hanson, 1970; Easterbrook, 1979; Baker and Bunker, 1985; Waitt, 1985; Atwater, 1986). To help improve our understanding of the spatial distribution, morphology, processes, and dynamics of the Okanogan lobe, and because vertical stratigraphic sections showing the glacial sediments of the primary landforms are relatively sparse, we extracted glacial bedform information (i.e., drumlin, macro–flutes, till lineations, and eskers) from aerial photographs and 10–m digital elevation models (DEMs) obtained from the Washington State Geospatial Data Archive (University of Washington). More than sixty U.S.G.S. 7.5 minute quadrangles were merged together in the Geographic Information Systems (GIS) environment. The spatial distribution of glacial features and ice–flow pattern is shown on Figures 28 and 29. An advance phase is marked by the streamlining of flutes and drumlins and ended with the deposition of a broad morainal

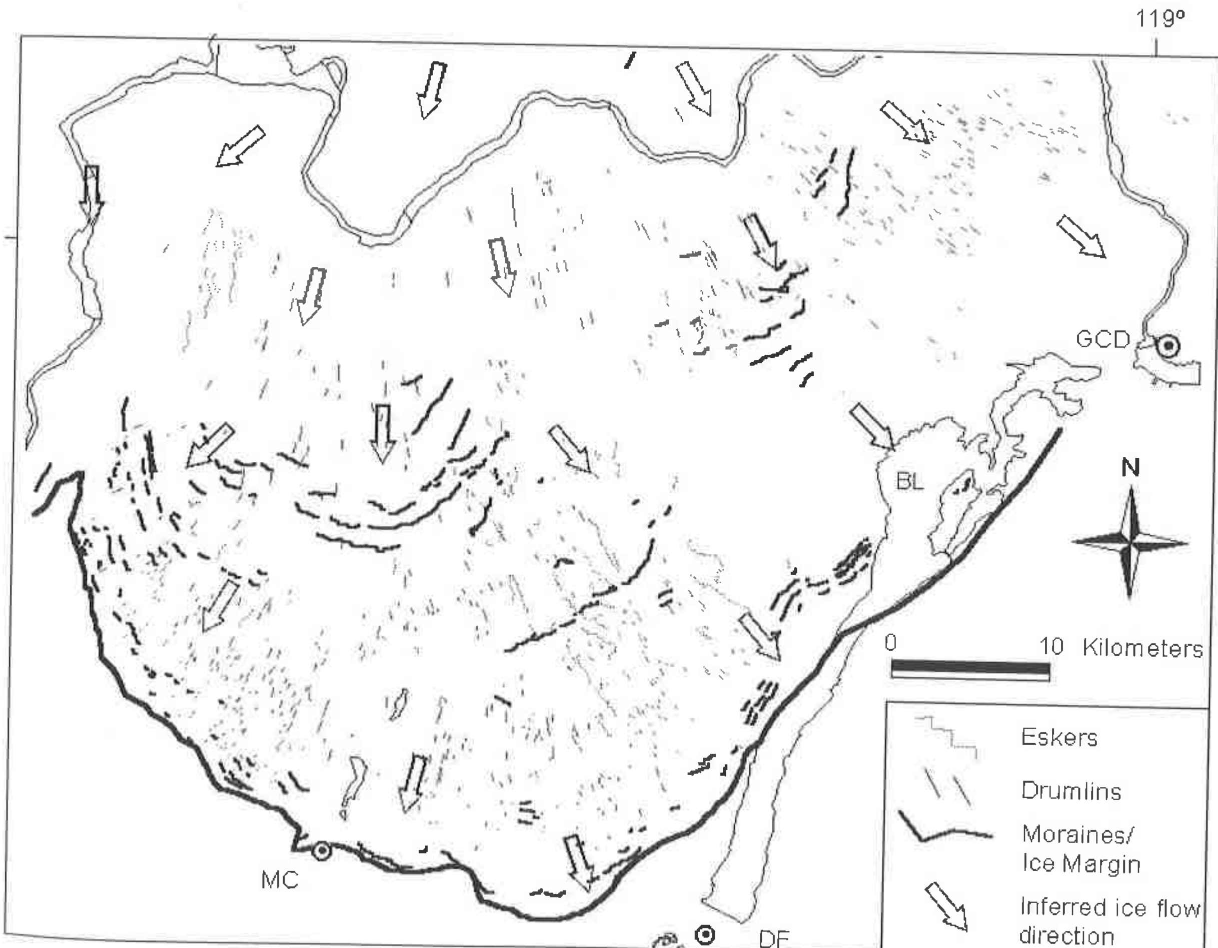

Figure 28. Distribution of glacial landforms and inferred ice–flow directions. Recessional moraines commonly overprint drumlins. DF–Dry Falls; BL–Banks Lake; MC–Moses Coulee; GCD–Grand Coulee dam.

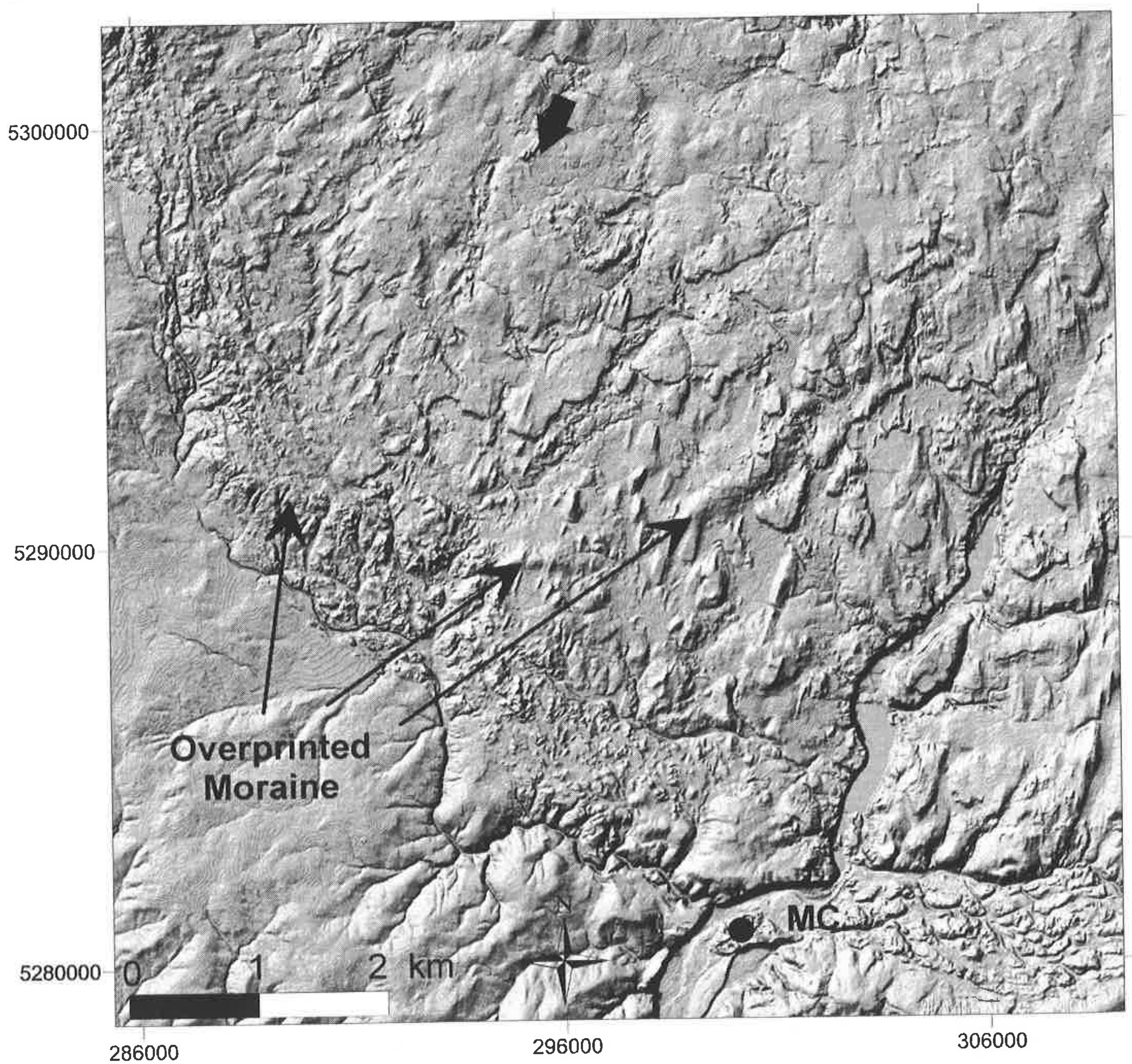

Figure 29. Shaded digital topographic model along the southwestern margin of the terminal area of the Withrow moraine. This image shows a lobate form that has been overprinted by till lineations. Arrows indicate inferred ice–flow direction.

complex which consisted of isolated kames composed of till and occasional, segmented, sharp–crested moraine ridges (the Withrow moraine). The retreat phase is marked by many moraines or morainal segments that are sub–parallel, arcuate in form, and transverse to ice–flow directions (see also Hanson, 1970; Easterbrook, 1979). Some recessional moraines can be traced almost continuously across the plateau and may be up to c. 60 m wide and rise 15–20 m above the local topography. Eskers are distinct, extensive, rise up to 15 m above the local relief, are up to 5 km long (60 m wide), and are typically found in association with back–wasting ice positions. Generally, two types of eskers were noted: (1) those associated with channels incised into bedrock and spatially fixed (i.e., "Nye Channels" or "N channels"; Nye, 1976), and (2) those that were formed when channels were melted upward into the glacier ice (i.e., "Röthlisberger channels" or "R channels"; Röthlisberger, 1972). Some glacial landforms overlap older forms, indicating more than one ice–flow event and possibly a switch in flow and/or thermal regimes.

Timing of ice–flow events and conclusions

The timing of the advance and retreat of the Okanogan lobe is not well constrained and this imposes caveats on the interpretation of the timing of meltwater routing and flooding events which were controlled by topographic features, ice margin configuration, and the opening of drainage routes. The main chronologic controls come from: (1) the presence or absence of volcanic ash in deposits which provide limiting estimates (e.g., Mullineaux *et al.*, 1978; Porter, 1978; Waitt and Thorson, 1983; Carrara *et al.*, 1996), (2) correlation with other major ice lobes, some of which may (Puget lobe) or may not (Columbia River and Purcell Trench lobes) be well–dated (i.e., Richmond, 1965; Easterbrook, 1979); and (3) counting–backwards techniques of dated varve sequences (e.g., Atwater, 1986; Waitt, 1985). The archive of glacial features in this area presents a dynamic picture of the Okanogan lobe. Reconstructions of the ice–surface morphology and estimated driving stresses (17–26 kPa) implied from ice thickness and surface slope in the terminal area suggest fast basal flow.

From Wenatchee, drive north on Highway 2 for 50 miles to Jameson Lake junction. Turn left and drive 2 ½ mi. to Moses Coulee moraine

STOP 4–1. WITHROW MORAINE AND OUTWASH TERRACE DRAPED ACROSS A GIANT MISSOULA FLOOD BAR, MOSES COULEE

Moses Coulee was scoured out by a giant flood from Lake Missoula prior to the last glaciation, leaving a large, dry gorge over 1.7 km (1 mi.) wide and 180 km (600 ft.) deep. During the late Wisconsin glaciation, the Okanogan lobe built the Withrow moraine, which drapes into Moses Coulee and across an earlier giant flood bar (Fig. 30).

Return to Highway 2, turn right and return to Farmer. Turn right and drive 9 mi. to the Withrow moraine.

STOP 4–2. THE WITHROW MORAINE

The Withrow moraine (Fig. 31), is a hummocky moraine, 1–2 km wide and 30–70 m high and extends across the Waterville Plateau. Large basalt erratics commonly dot the surface. The Columbia Plateau here slopes gently northward so meltwater streams could not flow directly away from the ice terminus. They flowed parallel to the ice margin, then down Moses to the Columbia River (Easterbrook, 1979). Dutch Henry Draw, which we crossed just before this stop, is one such channel.

Drive 12 ½ mi. to Mansfield. Cross recessional moraines. Continue for 3 ½ mi. to huge erratic, then 1 mi. to large erratic. Continue for 4.3 mi. to Barnes Butte kame complex. Drive 4.8 mi. past several eskers to Sims Corner.

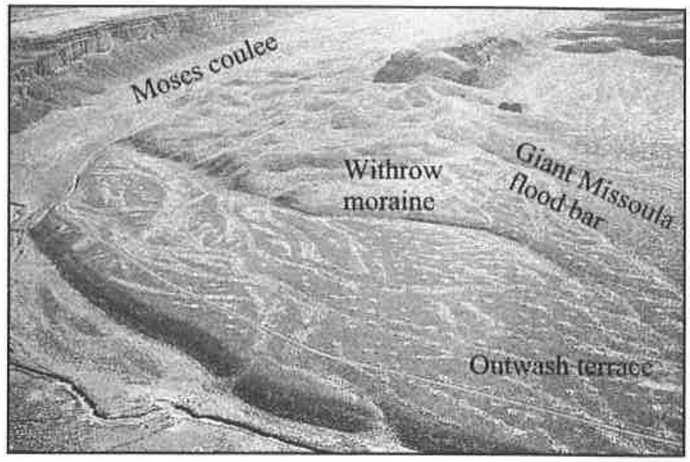

Figure 30. Withrow moraine and outwash terrace draped over a giant Missoula flood bar in Moses Coulee. (Photo by D.A. Rahm, courtesy of EPIC)

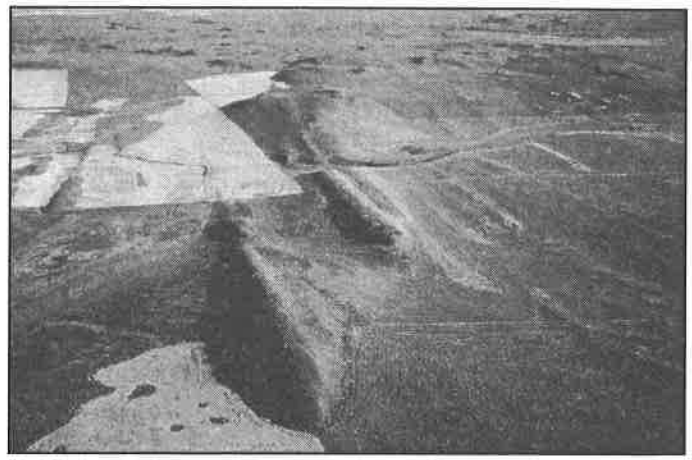

Figure 31. Drumlins behind the Withrow moraine (background). (Photo by D.A. Rahm, courtesy of EPIC)

STOP 4–3. ESKERS AND DISTAL FANS

The Columbia Plateau north of the Withrow moraine contains a profusion of eskers, kames, kettles, kame terraces, and related ice–contact deposits indicative of widespread glacier stagnation during deglaciation of the Okanogan lobe. Eskers which terminate at ice–marginal fans and low morainal ridges mark temporary halts in the recession of the glacier terminus. Three eskers end abruptly at ice–marginal fans which still retain the traces of abandoned braided channels on their surface (Fig. 32). The proximal side of the fans mark the former ice margin.

STOP 4–4. GRAND COULEE, DRY FALLS, MISSOULA FLOOD

Grand Coulee was described by Bretz (1969) as the "deepest, longest, and steepest in gradient. Its system possesses the

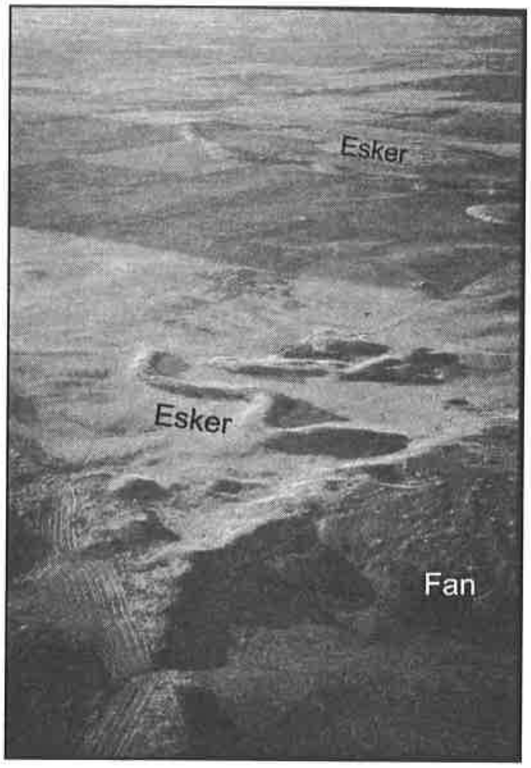

Figure 32. Esker terminating in a distal fan near Sims Corner. (Photo by D.A. Rahm, courtesy of EPIC)

Figure 33. Dry falls. (Photo by D.A. Rahm, courtesy of EPIC).

largest number of abandoned cataracts, the widest one, a record of the highest, the greatest cascade, and the largest number of distributary canyons. It is unique in that, at its initiation, it found no preglacial valley head but crossed 20 miles or so of the northern divide of the plateau before encountering such a valley, and it trenched that divide nearly 1000 feet deep".

Dry Falls (Fig. 33) extends about three miles across and is one of several falls excavated by spectacular quarrying of jointed basalt during catastrophic floods spilling down Grand Coulee when the Okanogan lobe blocked flow down the Columbia River near the site of Grand Coulee Dam and forced it to spill over into Grand Coulee. Much of the large size and depth of Grand Coulee is due to waterfall recession.

Drive 14 mi. to junction with Highway 2 and Dry Falls.
Drive down Grand Coulee to Elphrata and return to Seattle.

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