The formation of ice-marginal embankments into ice-dammed lakes in the eastern Puget Lowland, Washington, U.S.A., during the late Pleistocene

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Large embankments, typically several kilometers in lateral extent and many tens of meters high, choke the mouths of each alpine valley of the central Cascade Range, Washington State, U.S.A., at or near their junctions with the Puget Lowland. They comprise till and bedded gravel, sand, and silt, aggraded into ice-dammed lakes. The embankments lie within the late-Pleistocene Cordilleran ice-sheet limit and so do not mark the location of the ice-maximum terminus. Reconstruction of the subglacial hydraulic potential field indicates that these ice-dammed lakes would have drained subglacially via spillways located near the junction of each alpine valley and the Lowland. Physical processes tended to stabilize the grounding line for each ice tongue close to its respective spillway location. Because sedimentation rates are highest adjacent to the grounding line, subaqueous sedimentation formed a growing embankment there. In some valleys, subsequent subaerial lake drainage or decay of the active-ice dam resulted in late-stage deposition of deltas or valley trains. This analysis of ice-water behavior is based on physical principles that should be generally applicable to any environment where glaciers terminate against ice-dammed bodies of water.

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A striking characteristic of the topography along the western slopes of the Cascade Range of Washington State, U.S.A., is the presence of constrictions in river valleys as they pass from their bedrock-walled alpine reaches into the Puget Lowland (Figs. 1, 2). These constrictive landforms are not bedrock ridges but rather comprise glacial, fluvial, and lacustrine sedimentary material. They are here termed 'embankments' to avoid the common genetic implications of alternative nomenclature such as 'morainal embankments' or 'delta-moraines' that has been applied to these features. All lie within the limits and near the margin of the Puget lobe of the Cordilleran ice sheet that occupied the Puget Lowland most recently about 15,000 years ago, during the Vashon stade of the Fraser glaciation (Armstrong *et al.* 1965). Most of the embankment sediments are virtually unweathered, indicating transport and deposition during this late-glacial time (Booth 1986). A slightly earlier advance of alpine glaciers that originated in the Cascade Range is inferred to have climaxed and waned before the Vashon maximum (Cary & Carlston 1937; Mackin 1941; Williams 1971; Porter 1976). The lower reaches of alpine valleys were thus free of alpine-glacier ice during much or all of the time that the Puget lobe occupied the Lowland during the Fraser glaciation.

Eleven valley-constricting embankments are recognized in this area. Although a few early workers believed that these embankments were deposited by alpine ice flowing downvalley, most studies of embankment morphology and clast provenance have demonstrated that the sediment was derived from the Puget lobe (Cary & Carlston 1937; Mackin 1941; Knoll 1967; Williams 1971; Hirsch 1975). These studies also documented the presence of extensive ice-marginal lakes, impounded by the Puget lobe, that filled the lower reaches of the alpine valleys. A classic study by Mackin (1941) of one of the major embankment systems characterized it as a set of great glaciofluvial deltas that prograded into the adjacent alpine-valley lakes, veneered by till on the ice-proximal faces. He believed that they formed at the maximum stand of the Vashon-age ice sheet and were composed wholly of Vashon-age sediment. His work outlined the fundamental relationship between the ice sheet, alpine glaciers, and sedimentation associated with these ice masses that has guided all subsequent geologic study in this region.

This study addresses several characteristics of

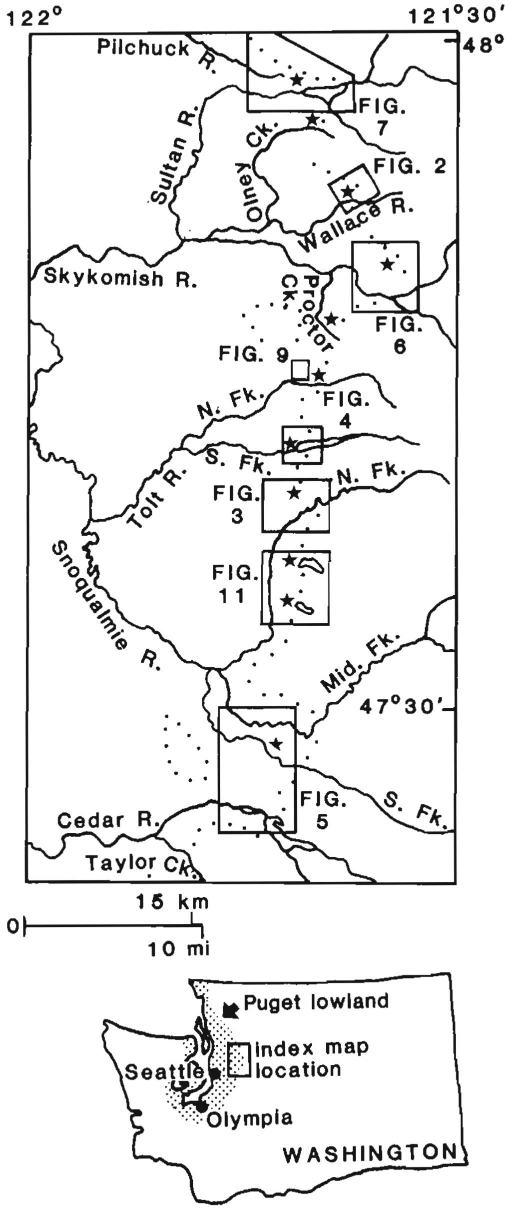


Fig. 1. Index map of geographical localities. Embankments are marked with stars. Vashon-age ice limit is shown with small dots (Booth 1986), which approximates the boundary between the Puget Lowland and the Cascade Range.

these embankments, particularly their common valley-mouth locations and the variety of sediments composing them. The analysis is based in part on available geologic data but stresses the physical behavior of glacier ice near its margin and the attendant depositional processes that op-

Fig. 2. Aerial view of the Wallace embankment (see Fig. 1 for location). View is to the east. The Wallace River has incised through embankment sediments adjacent to the south bedrock wall of the valley in postglacial time.

erate wherever a glacier terminates in an ice-dammed lake. Proposed processes of aqueous deposition adjacent to Pleistocene ice sheets have received considerable recent attention at several localities in eastern and central North America (e.g. Dreimanis 1982; Shaw 1982; Eyles & Eyles 1983; Barnett 1985; Heiny 1985; Kulig 1985). The examples from the Puget Lowland should prove quite relevant to these and other geographically diverse localities. Despite relatively limited exposures of their geologic materials, rendering detailed lithofacies schemes such as proposed by Miall (1977) or Eyles & Eyles (1983) inappropriate, these features provide a unique opportunity to study ice-marginal processes in an area where lake geometry, drainage routes, and configuration of the ice sheet are particularly well constrained.

Description

The eleven embankments in the study area (Fig. 1) have several common characteristics. They occupy alpine valleys, typically at or just above the junction of the valley with the Lowland proper. Their tops are from 100 to several hundred meters above the valley floor and cover 1–10 km^2. All appear to have filled their respective valleys from side to side (Fig. 2). Post-depositional river erosion has incised them along the valley axis and typically has removed much upvalley material. Less commonly, the downvalley faces have been modified as well (Fig. 3).

Embankment sediments comprise diamicton and bedded gravel, sand, and silt in variable pro-

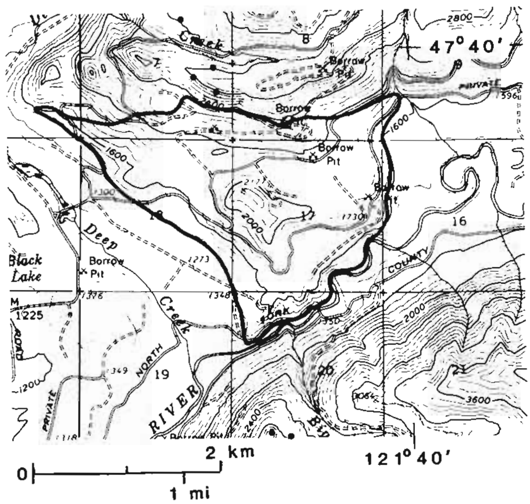


Fig. 3. North Fork Snoqualmie embankment (outlined). Postdepositional erosion is evident both in the transverse incision by the North Fork Snoqualmie River (N 1/2 section 20) and by modification of the ice-proximal face by recessional meltwater flowing south down the valley of Deep Creek (Sections 18–19). Topography from USGS Mount Si 15' quadrangle. Ice-maximum limit shown by black dots; altitude in feet.

portions. Fluvial sediment, deposited as either proglacial outwash or delta topset beds, typically forms much or all of their upper surfaces. The cores of these embankments are more heterogeneous and include significant amounts of subaqueously or subglacially deposited sediment. These nonfluvial sediments are typically weakly stratified and lie in abrupt contact with the overlying material.

Three examples, treated in detail, exemplify both their similarities and differences. The South Fork Tolt embankment is composed of thin, discontinuous fluvial deposits covering a rounded ridge of primarily glacially transported sediments. The Middle-South Forks Snoqualmie embankments, by contrast, are characterized by a planar surface of fluvial sand and gravel that extends both upvalley and laterally along the mountain front for many kilometers. The Pilchuck-Sultan embankment shares important characteristics of both; it is a particularly useful example because of good exposure and extensive subsurface data.

South Fork Tolt embankment

The South Fork Tolt embankment occupies an area of about 2 km^2, forming a broad north-rising ridge with maximum relief of 150 m (500 ft) (Fig. 4A). Truncated sedimentary layers on its steep eastern face indicate postglacial erosion of the embankment by the South Fork Tolt River. Through the southern part of the embankment, recent excavations have exposed a 100-m vertical section of glacial, glaciolacustrine, and glaciofluvial sediments.

This section is subdivided into three units (Fig. 4B). The lowest (unit 1) is an oxidized bouldery

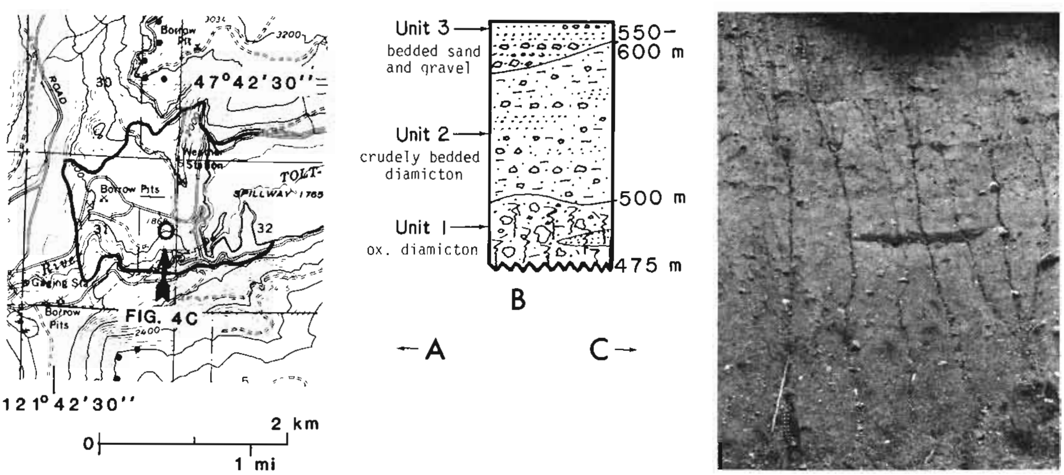
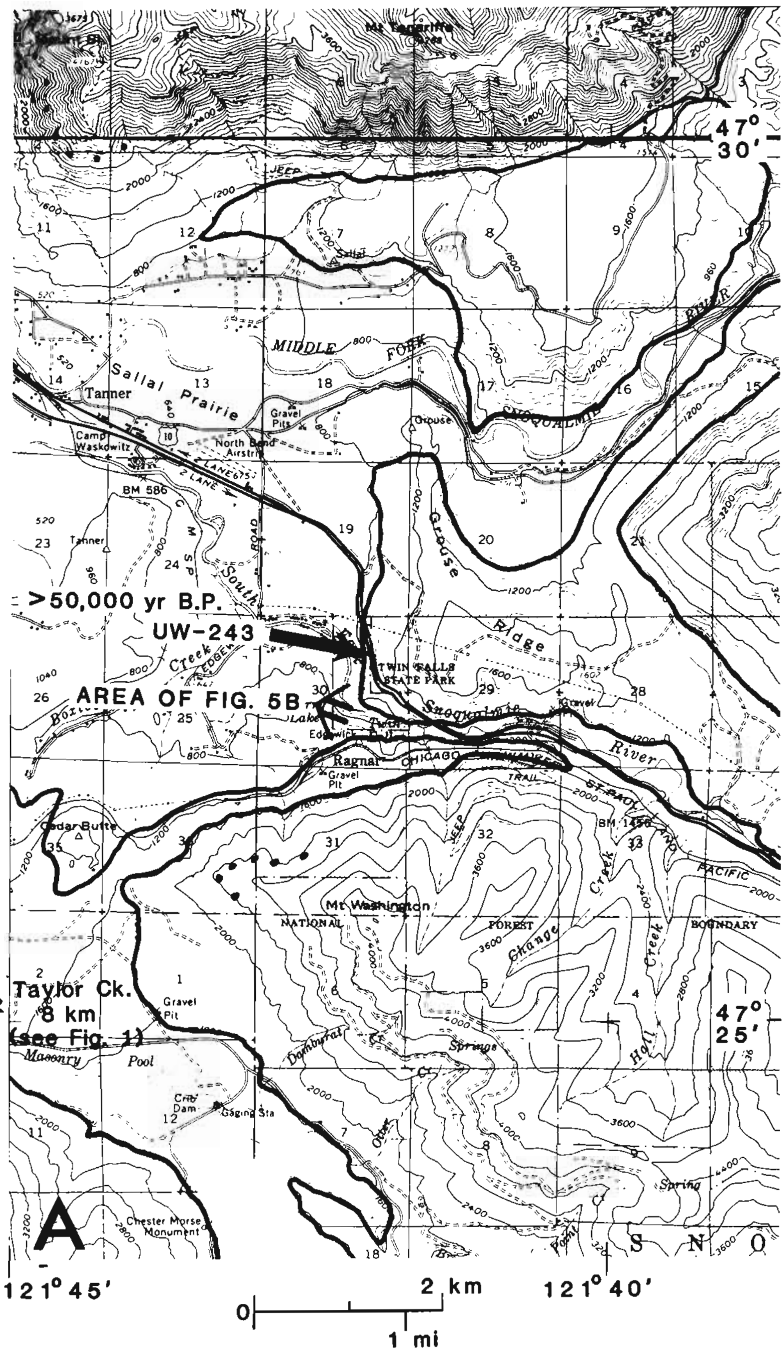


Fig. 4. South Fork Tolt embankment. A: Topography from USGS Mount Si 15' quadrangle (embankment is outlined). Ice-maximum limit shown by black dots; altitude in feet. B: Stratigraphy of embankment sediments exposed downstream of dam. Contacts between units are not necessarily horizontal and so span a range of altitudes within the represented region. C. Pebbly silt diamicton with crude horizontal bedding (location shown on Fig. 4A; altitude 520 m (1,710 ft)).



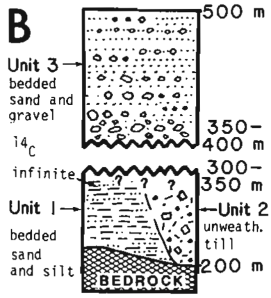


Fig. 5. Middle-South Forks Snoqualmie embankment complex. A: Topography from USGS Mount Si and Bandera 15' quadrangles. Ice-maximum limit shown by black dots; altitude in feet (outlined area includes embankment and its SW continuation past Cedar Butte). B: Stratigraphy of embankment sediments in the vicinity of the Holocene channel of the South Fork Snoqualmie River.

diamicton containing lenses 5–30 cm thick of compact medium sand. Clast lithology and degree of weathering indicate probable deposition by an ice sheet of pre-Fraser age (Booth 1986).

A massive to crudely bedded diamicton (unit 2, Fig. 4B; Fig. 4C), 50 to 100 m thick, overlies the lowest unit. This material constitutes the bulk of the embankment and defines the shape of its upper surface. Gravel-sized clasts are round to subangular, commonly striated, and of mixed lithology. The clasts have no apparent preferred orientation. The diamicton differs from typical Puget-Lowland lodgment till in that it has a far lower degree of compaction and a slightly higher clay content in the matrix (*c.* 10%; cf. Crandell 1963; Mullineaux 1970; Dethier *et al.* 1981). Subhorizontal pebble-poor zones, 5–20 cm thick that extend laterally for several meters, are interspersed throughout the diamicton. Larger discontinuous subhorizontal sand layers, extending 1–20 m, are clustered in, but not limited to, a zone 20–30 m thick in the middle of the unit. Horizontal layering and planar contacts throughout the exposure show that no significant flowing, slumping, or glacitectonic deformation of sediment occurred during or following deposition. In one locality west of the south abutment of the dam, silt and clay laminae drape over, and are depressed locally several centimeters beneath, clasts (dropstones) as large as 15 cm in diameter.

A surface unit (unit 3, Fig. 4B) of oxidized glaciofluvial recessional outwash, generally less than 2 m thick, overlies the diamicton. It drapes most of the embankment and thickens to form a constructional surface only on the western (ice-proximal) slopes below about 500 m (1,600 ft) altitude, where the sediments merge with fluvial deposits of the late ice recession (Booth 1986).

Middle and South Forks Snoqualmie embankments

A nearly continuous, flat-surfaced embankment complex that chokes the valley mouths of the Middle Fork and South Fork of the Snoqualmie River (Fig. 5A) can be traced south for nearly 15 km along the mountain front. It blocks the Cedar River valley and terminates against ice-contact sediment at the northeast edge of the Taylor Creek valley (Frizzell *et al.* 1984). Its surface altitude descends from about 510 m (1,670 ft) on the proximal side of the Middle Fork embankment to 500 m (1,640 ft) at the South Fork embankment to 475 m (1,560 ft) northeast of Taylor Creek. Approximately one-third of this gradient is probably due to isostatic rebound since deglaciation (Thorson 1981).

Most of the sediment exposed in roadcuts and streambanks is gravel and sand, subrounded to well rounded, with better sorted sand layers and lenses throughout. Both median and maximum grain sizes gradually decrease up the Middle and South Fork valleys, with pebble-poor lacustrine silt deposits dominating beyond about 4 km from the ice-proximal face of the embankments. At the Middle Fork embankment, this face is characterized by ice-contact topography and moderately to poorly sorted gravel, sand, and silt and an absence of recognizable lodgment till. A more complex record is revealed at the South Fork embankment (Fig. 5B), perhaps owing to more extensive exposure. Low on its ice-proximal face, an area of fine-grained, horizontally bedded sand and silt is exposed (unit 1). Wood in this sediment, in part converted to lignite, yielded a ^{14}C date of >50,000 years B.P. (Porter 1976; sample number UW-243). Drag folds and thrust faults, at an altitude of about 300 m (1,000 ft) near the top of the deposit, indicate shearing of beds to the east, presumably by overriding ice. Unweathered (i.e. Vashon) till (unit 2, Fig. 5B), which rests directly on striated bedrock, is exposed only along the river topographically below the older sedimentary material. This till must have been deposited in a valley incised through pre-Vashon sediment.

The uppermost sediment of this embankment (unit 3, Fig. 5B) is part of an extensive recessional outwash deposit over 100 m thick in this area. As Mackin (1941) first inferred, much of this fluvial sediment may have been deposited as deltas into ice-marginal lakes. Deltaic progradation, coupled with ice-margin retreat, resulted in deposition of deltaic and lacustrine sediments up the Middle and South Fork valleys that merge with a valley-train fluvial deposit southwest of Cedar Butte (Fig. 5A).

An analogous history is inferred for at least one other embankment in this area. The morphology of the Skykomish embankment (Fig. 6) is controlled by an extensive deposit of late-stage fluvial sediment that extends many kilometers upvalley. Poorly exposed beneath this material, however, is both a basal pebble-poor diamicton and a recessional moraine of Vashon age (Booth 1986).

Pilchuck-Sultan embankment

Deposits constituting the embankment at the head of the Pilchuck River (Figs. 7A and 7B) are exposed in excavations associated with the construction and subsequent raising of adjacent Culmback Dam. They are augmented by extensive subsurface exploration (Converse Ward Davis Dixon 1979; Shaffer 1983). The basal embankment sediments are encountered in two drill

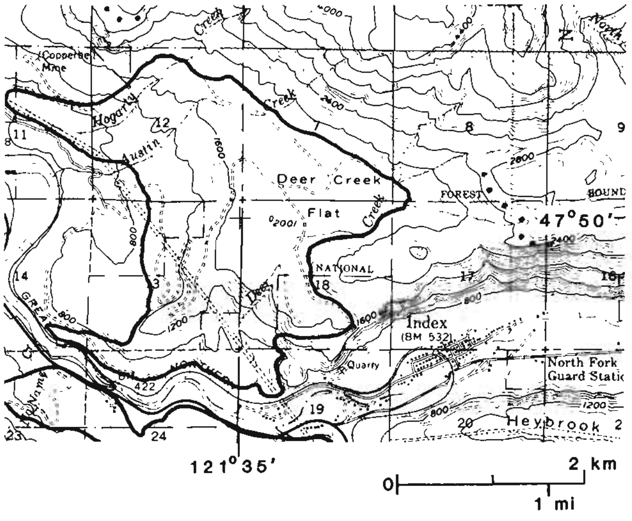


Fig. 6. Skykomish embankment (outlined). Pebble-poor diamicton is exposed in section 24 just south of the highway; the ridge in section 18 (altitude 610 m (2,001 ft)) is a bouldery moraine. Topography from USGS Index 15′ quadrangle. Ice-maximum limit shown by black dots; altitude in feet.

holes near the downstream toe of the embankment. These sediments consist of more than 100 m of fluvial sand and gravel that fill part of the bedrock valley of a pre-Vashon Sultan River beneath the present embankment (unit 1, Fig. 7B). The upper surface of the sand and gravel, at an altitude of about 410 m (1,350 ft), is 20–30 m above the present altitude of the bedrock lip through which the Sultan River now drains into the Sultan gorge. This upper surface may represent a pre-Vashon-age floodplain of the Sultan River, an interpretation that also implies drainage through the Sultan gorge during pre-Vashon time. Alternatively, the surface may represent the level of proglacial fluvial deposition during the Vashon advance only (Shaffer 1983).

Bedded silt and pebbly silty diamicton (unit 2, Fig. 7B) overlie the fluvial sediment. On the south dam abutment the unit is only 5–10 m thick and varies from rather typical massive and stony Vashon till (clast content about 25%) to a poorly sorted, pebble-poor deposit containing discontinuous silt layers 1–2 cm thick and a few gravel-rich lenses. On the north abutment it thickens to nearly 100 m. In this area, clasts constitute less than 5% of this unit (Fig. 7C); those present are subangular to well rounded, 0.5–4 cm in diameter, and commonly striated. The matrix primarily consists of fine silt. Faint color changes and slightly coarser texture define subhorizontal layers in this silt up to 10 cm thick and several meters in lateral extent. Lenses of fine sand as much as 1 m thick are also present. Poor exposures farther east suggest that erosional remnants of this unit extend at least 3 km upvalley of the main embankment crest.

This unit is covered by bedded gravel and sand (unit 3, Fig. 7B) that is at least 100 m thick on the northern part of the embankment. This fluvial sediment forms a nearly continuous kame terrace that begins at the 700-m (2,300-ft) altitude at a bedrock notch on the north wall of the Pilchuck valley, continues upvalley as a feature only a few tens of meters wide, and broadens abruptly at the embankment itself. The gradient of the constructional surface is about 12 m/km (65 ft/mi), which projects south to the altitude of a spillway at Olney Pass (Fig. 7A).

Reconstruction of the ice-marginal environment

Ice-dammed lakes

The occupation of the Puget Lowland by the Cordilleran ice sheet required major readjustments in the prevailing westward and northwestward drainage out of the Cascade Range. Rivers were first diverted south along the advancing ice margin. Eventually, the ice reached the eastern edge of the Lowland, covering all ice-free routes. Glacial lakes then formed in the lower reaches of the west-draining alpine valleys of the Cascades (Cary & Carlston 1937; Mackin 1941; Williams 1971; Porter 1976; Booth 1985, 1986).

Ice-marginal lakes of similar size and geometry are common in modern glacial environments

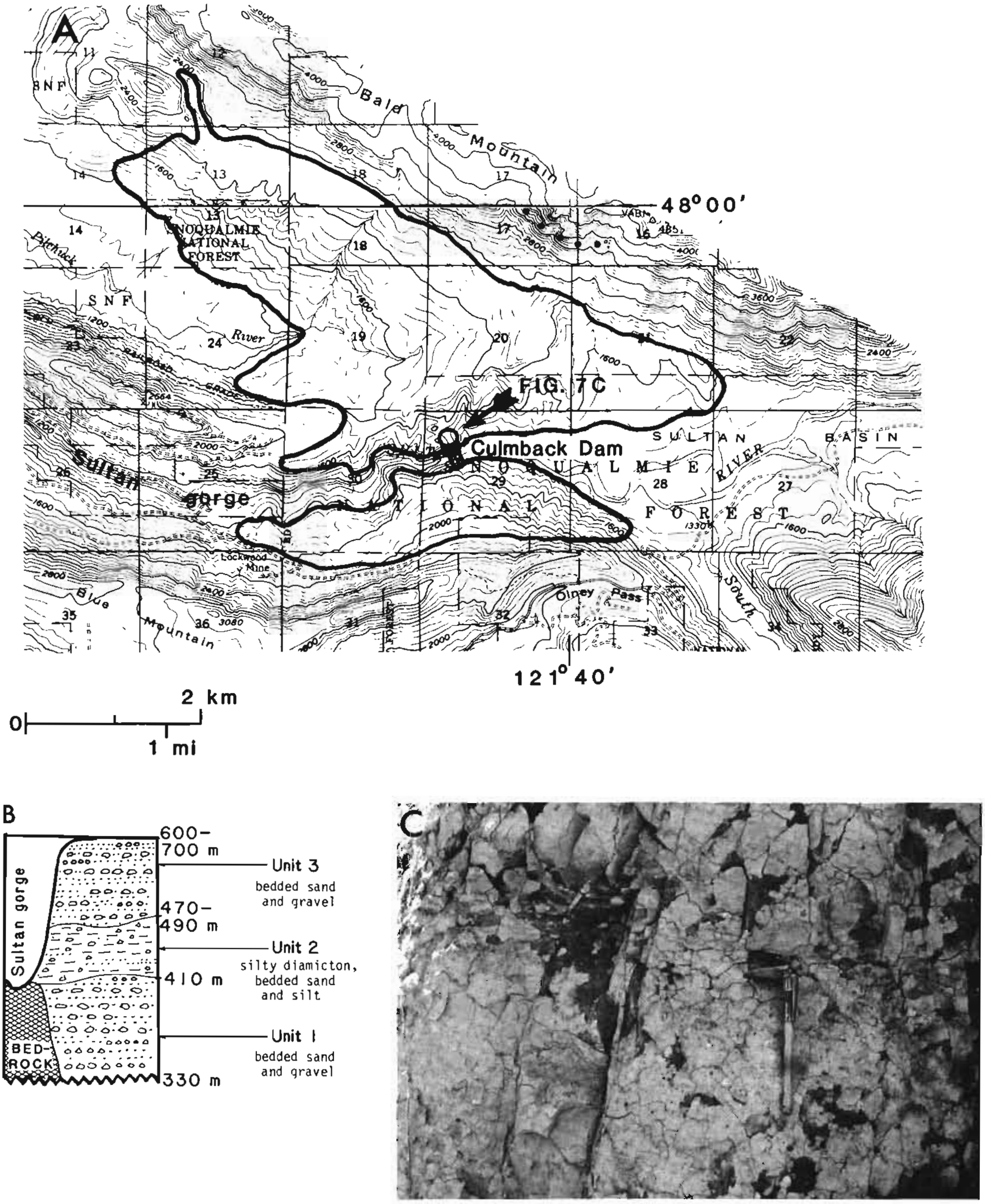


Fig. 7. Pilchuck-Sultan embankment. A: Topography from USGS Index and Silverton 15′ quadrangles (embankment is outlined). Ice-maximum limit shown by black dots; altitude in feet. B: Stratigraphy of embankment sediments north of Culmback Dam (adapted from Shaffer 1983). C: Pebble-poor silty diamicton; zones of horizontally bedded silt are present above and below shovel (location shown on Fig. 7A; altitude 440 m (1,440 ft)).

(Stone 1963; Post & Mayo 1971). These lakes typically drain either through subaerial spillways upvalley of the ice or beneath the damming ice itself (reports of long-term lake drainage *over* the ice surface are notably rare).

In valleys adjacent to the Puget Lowland, bedrock divides upvalley of the maximum ice margin stand well above the altitude of the damming ice. Thus, drainage of each lake could only be over, under, or around the ice dam (see below). Even at lower ice stands, only the Sultan River valley had exposed a spillway over the bedrock divide (at Olney Pass; see Fig. 7A) that permitted subaerial lake drainage while ice still blocked the valley mouths. All other valleys drained via the ice dam for as long as active glacier ice remained at their mouths.

Sedimentological processes in ice-marginal environments

The origin of deposits similar to the stratified, pebbly silt and diamict units in both the South Fork Tolt and Pilchuck-Sultan embankments (unit 2, Figs. 4B and 7B) has been discussed by many authors (e.g. Dreimanis 1979; Evenson *et al.* 1977; May 1977; Gibbard 1980; Eyles & Eyles 1983; McCabe *et al.* 1984; Barnett 1985; Heiny 1985). Common to all hypotheses is an inferred environment of subaqueous deposition, indicated in particular by laterally continuous, nondeformed stratification that presumably precludes deposition (lodgment) by ice sliding over its bed.

All of the characteristics of the diamict units in these Puget-Lowland embankments are compatible with, though not demanding of, sedimentation processes characterized by fallout of debris from the base of melting ice combined with sediment influx from subglacial streams and the settling of suspended lacustrine sediments. The variable sediment load of ice and subglacial water yields irregularly distributed pebble-poor layers and lenses within a more uniform lithology of fine-grained suspended material. Traction currents associated with water flow into or out of the lake locally produce stratified and sorted coarse-grained layers. The absence of sediment-flow features rules out significant deposition by subaqueous flowtill (Evenson *et al.* 1977). The lack of overconsolidation or glaciodynamic structures (Dreimanis 1976) near the top of most of these deposits implies that the ice may have only grounded rarely against these sediments (Gibbard 1980; Eyles *et al.* 1982; Shaw 1982; Eyles & Eyles 1984).

Interaction of ice and water at the ice margin

Although the sedimentological evidence supports the existence of lacustrine environments adjacent to and beneath the ice sheet, these data alone cannot conclusively delineate this environment (cf. Karrow *et al.* 1984; Christe-Blick 1985). Furthermore, they offer little insight into the configuration of the ice sheet, the controls on lake-surface altitude, or the determination of where and under what conditions sedimentation will occur. These deficiencies can be corrected by analyzing the mechanics of ice and water along a glacier margin. This analysis not only independently confirms the presence of this inferred environment, but also shows why embankments formed almost exclusively at valley mouths, why they are ubiquitous in this region, and whether they represent climatically induced stillstands of the ice margin.

Ice limits. – Evidence for maximum extent of ice into alpine valleys that enter the eastern Puget Lowland is not well expressed on the steep valley sidewalls. The distribution of glacial erosion and erratics on the adjacent interfluves (Booth 1986), however, indicates the vertical extent of these ice tongues. These data show that ice, at maximum, stood from 100 to over 600 m above the altitude of the present embankment surfaces and in places nearly 700 m above the inferred Vashon-age valley floor (Table 1). Embankment altitudes bear no systematic relationship to adjacent ice-maximum altitudes, and thus embankments represent neither the position of maximum ice advance (Knoll 1967) (because 600-m-thick ice would have continued advancing beyond them) nor a synchronous halt of the whole ice front during retreat.

Marginal lakes and subglacial hydrology. – Once a marginal lake is impounded by ice, the lake's maximum surface altitude equals the maximum hydraulic potential that its drainage route must cross (Nye 1976). For subaerial drainage, this potential is simply the altitude of the spillway. For subglacial drainage paths, the hydraulic potential has two components: the bed altitude of the channel and the pressure of the overlying ice (Shreve 1972). These are analogous to the eleva-

Table 1. Embankment altitudes, inferred glacial-age valley-floor altitudes, and reconstructed ice-surface altitudes in the Puget Lowland.

	Embankment altitude (m)[1]	Valley-floor altitude (m)[2]			Embankment height[3] (m)	Ice altitude[4] (m)	Ice thickness[5] (m)
		min	*max*	*best*			
Pilchuck-Sultan	680/480	390	410	390	290/90	1060	380/580
Olney	510	370	510	490	20	1050	640
Wallace	800	590	610	590	210	980	180
Skykomish	540/160	120	130	130	410/30	860	320/700
Proctor	760	630	650	630	130	910	150
North Fork Tolt	550	430	460	450	100	880	330
South Fork Tolt	610/600	500	540	500	110/100	860	250/260
North Fork Snoqualmie	600/600	270	490	390	210/210	850	250/250
Calligan Lake	710	550	680	590	120	840	130
Lake Hancock	700	550	660	610	90	830	130
Middle-South Forks Snoqualmie	490	230	360	270	220	780	290

[1] First number is the present embankment altitude. Second number is available data on the altitude of the highest non-fluvial sediment deposited subglacially and/or subaqueously.
[2] Evidence for valley-floor altitude includes the top of bedrock or pre-Fraser material (minimum altitude) and the base of recessional or postglacial deposits (maximum altitude). The best estimate assumes a concave-upvalley profile and is probably accurate in all cases to within 50 m.
[3] Equals the difference between the embankment altitude and the best valley altitude.
[4] Booth (1986)
[5] Equals the difference between ice and embankment altitudes.

tion head and pressure head, respectively, common in groundwater terminology.

The hydraulic potential (or total head) at any point on the glacier bed can be expressed as an altitude, ignoring any dynamic component due to sliding of the ice (Booth 1984a):

$$H = z_b + (\varrho_i/\varrho_w)h, \qquad (1)$$

where H = the hydraulic potential, z_b = bed altitude, ϱ_i = ice density, ϱ_w = water density, and h = thickness of ice.

In these terms, H equals the altitude to which water would rise in a borehole that penetrates the glacier to its bed. The hydraulic potential increases with increasing ice thickness and decreases with decreasing bed altitude. The gradient of H specifies the direction of water flow (i.e. from regions of high H to low H). Note from equation 1, however, that any change in H is primarily determined by changes in the altitude of the ice surface; bed altitude exerts an influence only one-eleventh as great (Shreve 1972, 1985; Nye 1976). The hydraulic potential at the base of the ice can therefore be reconstructed (Fig. 8) from the reconstructed topography of the ice surface and the glacier bed (approximately the present ground topography).

The hydraulic potential beneath ice tongues that extended into alpine valleys in the study area thus increased upglacier (to the west and northwest, towards the center of the ice sheet). This was the direction of greater ice-surface altitudes, more than compensating (via eqn. 1) for the descent of the valley floors in this (upglacier) direction. Any potential route of subglacial drainage down these valleys thus faced steadily *rising* hydraulic potentials. At the edge of the Lowland, however, potential subglacial drainage routes were not constrained by valley walls and so could turn southward, in a direction of decreasing ice-surface altitudes, to follow paths where the hydraulic potential decreased monotonically (cf. Fig. 8). A maximum value was therefore defined for the potential along each complete drainage route, between their 'confined' west-flowing and 'unconfined' south-flowing legs, which must have been exceeded before the water impounded in a valley could drain. The location of these maximum potentials, near to where each valley entered the Lowland proper, defined the position of an effective spillway for each lake.

Where subglacial lake drainage flowed over such a spillway upglacier from the snout of each local ice tongue, some ice between the spillway and the glacier margin probably began to float as

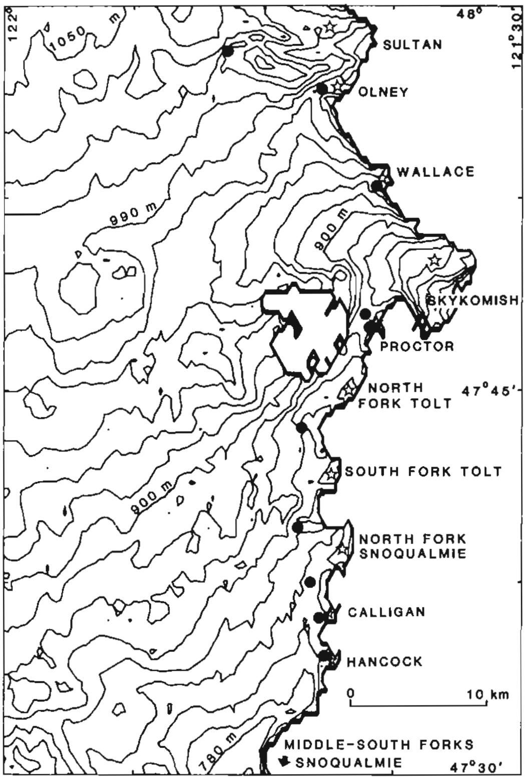


Fig. 8. Contour map of subglacial hydraulic equipotentials calculated using equation 1, contour interval 15 m. Ground surface digitized from 7.5′ and 15′ topographic maps; ice-surface reconstruction from Booth (1984a: fig. 3.8). Embankment locations are marked with open stars, subglacial spillways by heavy dots.

the altitude of the impounded lake approached the value of the spillway's hydraulic potential. Such lacustrine ice shelves have been observed in some areas (Nye 1976; Stone 1963; Post & Mayo 1971, lakes no. 6 and 27; Sturm & Benson 1985) though are absent or ill-defined in others (e.g. Holdsworth 1973; Bindschadler & Rasmussen 1983). An ice shelf will always tend to advance into standing water because its unsupported portion above water level exerts a longitudinal stress greater than the water pressure acting on the submerged ice face (Weertman 1957; Thomas 1973). The shelf thickness necessary for appreciable strain rates, observed in Antarctica to be 150–300 m (Robin 1979; Thomas 1979), was exceeded during the Vashon stade in most of the major alpine valleys entering the Puget Lowland (Table 1). Therefore, shelf ice adjacent to the Puget lobe probably advanced beyond the grounding line into these ice-marginal lakes.

Water flow into a marginal lake will eventually raise the water-surface altitude to equal the spillway potential. If the spillway, and thus the drainage, is subaerial over a bedrock divide, lake level is stable at this point. If drainage is subglacial, a more complex sequence of events modeled by Nye (1976) typically results in total or near-total drainage of the lake (Clarke 1982). Subglacial water flow melts a tunnel, whose rate of expansion due to melting initially exceeds its closure rate due to ice deformation. The hydraulic potential along the tunnel thus loses its component from the ice overburden and drops precipitously to equal the bed altitude. Drainage should now proceed catastrophically as a jökulhlaup until lake level falls to the altitude of the highest *bed* spillway along the drainage route (z_b in eqn. 1) or until ice finally reseals the tunnel during waning flow. Refilling of the lake begins the process anew (e.g. Stone 1963; Post & Mayo 1971; Mottershead & Collins 1976; Sturm & Benson 1985). The lakes along the eastern margin of the Puget lobe probably refilled in a few tens of years

Fig. 9. Boulder gravel deposited by episodic drainage out of glacial Lake Skykomish (see Fig. 1 for location). Note person for scale (circled); boulders in this layer range from 1–3 m in diameter.

(Booth 1984a), and so this process repeated itself numerous times during the Vashon stade. The predicted path of these outburst floods is etched in the landscape of the eastern Puget Lowland as a 40-km-long trough up to 1 km wide and over 100 m deep (Booth 1984b), a portion of which is depicted along the west edge of Fig. 11 (now occupied by the North Fork of the Snoqualmie River). Although largely swept clear of sediment, the trough is choked in one area by a 200-m-thick section of fluvially transported boulder gravel with clasts up to 3 m in diameter (Fig. 9).

Because no subaerial drainage routes out of the upper Cascade valleys existed during the Vashon maximum, each ice-dammed lake in this region would have experienced subglacial drainage during some (usually extensive) period of its existence. Lake drainage *over* or *around* each ice dam is precluded on both theoretical and empirical grounds. Because water is more dense than ice, its hydraulic potential will exceed that of ice at the bed before the water surface ever rises above the ice level (an obvious prerequisite to supraglacial flow). Subaerial flow around an ice dam confined by steep valley walls (Mackin 1941) is equally implausible, requiring the water to flow contrary to the prevailing hydraulic gradient (see below). These physical constraints are borne out by the absence of reported supraglacial lake-drainage systems on modern glaciers. Geologic evidence in the Puget Lowland is equally in accord with these physical expectations; incised channels adjacent to each valley at or near the maximum ice-margin position are notably absent, implying a lack of significant subaerial marginal drainage along the steep interfluves. Even at progressively lower ice stands, only in the Sultan River valley would subaerial paths have been uncovered (the farthest upvalley being Olney Pass; see Fig. 7A). The remaining valleys drained subglacially for as long as active glacier ice remained in their lower reaches.

Sedimentation into ice-dammed marginal lakes

Processes

The glaciolacustrine environment permits a variety of potential depositional processes, generalized in Fig. 10. These include basal meltout of sediment beneath a floating shelf or berg, subaqueous flowtill off the glacier snout, subglacial

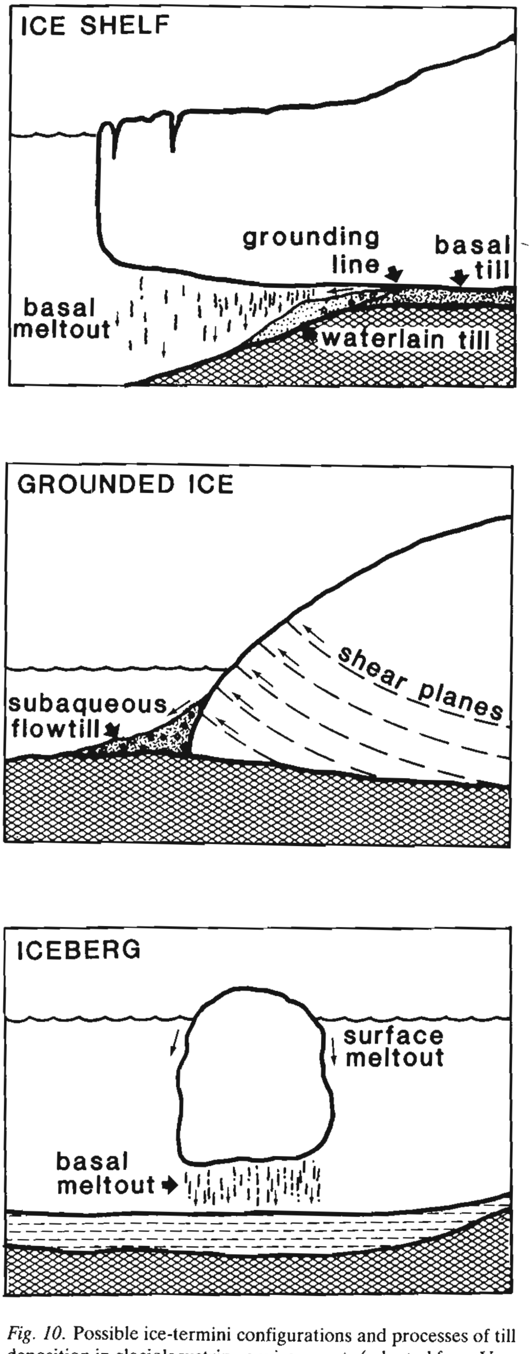


Fig. 10. Possible ice-termini configurations and processes of till deposition in glaciolacustrine environments (adapted from Vorren *et al.* 1983). More detailed discrimination of depositional processes (e.g. Eyles & Eyles 1983: fig. 4) in the Puget-lobe embankments is inhibited by insufficient exposure of the sediments themselves. With a stable grounding line beneath the ice shelf depicted in the upper diagram, waterlain till will accumulate over time to form a ridge regardless of the topography of the cross-hatched substrate.

or supraglacial transport of sediment by running water into the lake, and settlement of suspended lacustrine sediment (variously discussed singly or in combination by Rust & Romanelli 1975; May 1977; Evenson *et al.* 1977; Fecht & Tallman 1978; Dreimanis 1979, 1982; Gibbard 1980; Hicock *et al.* 1978; Hillaire-Marcel *et al.* 1981; Orheim & Elverhøi 1981; Powell 1981; Eyles & Eyles 1983; Vorren *et al.* 1983; McCabe *et al.* 1984). Continued aggradation, lowering of lake level, and/or exposure of a subaerial spillway will permit subsequent deltaic or subaerial fluvial sedimentation as well. These processes are not mutually exclusive; most studies have recognized their close association in even single exposures. The rates and dominant sites of sedimentation, however, may differ for each process.

Sedimentation by most subaqueous processes is concentrated at the grounding line (Orheim & Elverhøi 1981), because the glacier is the primary source of sediment. Except in the presence of localized fast currents near the ice margin, all but the finest suspended sediment should be deposited here (Edwards 1978; Eyles & Eyles 1983). The resulting deposit commonly forms shoals or ridges beneath the ice at the grounding line. These linear ridges have been repeatedly observed or inferred beneath modern glaciers and ice shelves (Holdsworth 1973; Barnett & Holdsworth 1974; Post 1975; Rust & Romanelli 1975; Edwards 1978; Hillaire-Marcel *et al.* 1981; Powell 1981; Sturm & Benson 1985).

Alternative processes that distribute subaqueous sediment over a wider area are likely to be ineffective in a confined ice-marginal lake. Because the debris content of ice tends to be highest at its base (Pessl & Frederick 1981), basal melting will yield the greatest volume of sediment near the grounding line regardless of how far ice floats or extends into the basin (cf. Anderson *et al.* 1980). Based on study of modern Alaskan glaciers, the volume of supraglacial debris available for more distal transport is negligible by comparison to the basal component (Evenson *et al.* 1985). Furthermore, rafting by icebergs is restricted to those months when the lake surface is unfrozen (Holdsworth 1973).

Sudden lowering of lake level during a period of episodic drainage tends to destabilize an ice shelf and hastens its breakup into bergs (e.g. Post & Mayo 1971: lake no. 5 photo). Slower refilling of the lake would have a similar, though less pronounced effect (Holdsworth 1973; Barnett & Holdsworth 1974). Such changes in ice configuration upvalley of the grounding line, however, should have little effect on the primary pattern of sedimentation, namely a pronounced decrease in sedimentation away from the grounding line. These Puget-Lowland embankments thus reflect anticipated depositional processes focused near the grounding line of a glacier adjacent to a marginal lake.

Location and stability of the grounding line

As the site of maximum subaqueous sedimentation, the position of the grounding line controls the pattern of deposition. Formation of a discrete embankment requires that the glacier margin terminate in or against standing water and maintain a relatively stable grounding line for a considerable time (Rust & Romanelli 1975). A stable grounding line requires, either singly or in combination: (1) a long-term climatically controlled stillstand of the ice margin, coupled with either a fixed water level or one that never rises high enough to float a portion of the ice; (2) presence of a pre-existing topographic rise or shoal; (3) resistance to ice advance by confining topography; or (4) a process that incorporates the interdependence of the thickness of an ice dam with the water depth in an impounded lake.

(1) *Climatically controlled stillstand.* – In the Puget Lowland, evidence does not seem to support this alternative (Thorson 1980; Booth 1986). Limiting radiocarbon dates on the Vashon stade from the central Lowland (Rigg & Gould 1957; Mullineaux *et al.* 1965) require rapid average advance and retreat rates of the terminus of about 100 m/a throughout the glaciation. Commensurate motion of the lateral ice margins would be expected to have followed suit. A single stillstand at or near ice maximum (Mackin 1941) cannot explain the formation of these embankments (Table 1) because the variation in ice thickness above them implies great variability in the length of the ice-maximum tongue beyond them. No geologic evidence in the Lowland suggests the likelihood of eleven such halts, needed to climatically stabilize each of the ice tongues at the present sites of their respective embankments.

(2) *Pre-existing shoals.* – From the record of multiple glaciations in the Lowland (Armstrong *et al.* 1965) and the ubiquitous presence of Vashon-age embankments, pre-existing valley blockades were probably present when the Vashon ice sheet

arrived. Older glacial sediments are indeed observed at the core of several of these embankments (e.g. Porter 1976; Figs. 4B, 5C). This explanation, however, merely displaces back in time the question of why deposition is localized. It is also inapplicable for those embankments were deep exposures or drill cores reveal no such material.

The effect of shoals may nevertheless become important during the formation of a new embankment. Once sediment begins accumulating, the growth of an incipient embankment will tend to stabilize the grounding line (Post 1975) and further localize deposition there.

(3) *Confining topographic effects.* – The position of most of the embankments at valley mouths empirically shows the importance of the valley-wall configuration. Its conceptually most simple consequence is to induce resistance to ice-front advance, thereby halting any advance of the grounding line as well. As an ice tongue advances into a valley, it must thicken and steepen at the valley entrance to compensate for the added lateral drag (Nye 1965). The convergent upvalley geometry typical of most valleys amplifies this effect (Mercer 1961; Funder 1972). If the valley is perched above the Lowland, such as those shown in Fig. 11, this pause will last even longer while ice thickens at the valley mouth. The second consequence of valley-wall configuration is reflected indirectly in its contribution to the location of the subglacial spillways (discussed below).

(4) *Interdependence of ice thickness and water depth.* – Although an ice margin and grounding line may stabilize under the special conditions mentioned above, these conditions cannot account for deposition of all embankments along the Cascade front, because they do not all satisfy one or more of the preceding requirements. The morphological and sedimentological data from these embankments can be understood, however, by considering the mechanics and interaction of ice, subglacial topography, ice-marginal water, and transported sediment.

As thickening ice seals subaerial drainage routes out of a basin, rising lake water, unavoidable in a blocked basin, must eventually float the glacier snout prior to (or simultaneous with) lake drainage. The location of the grounding line then becomes a function of the depth of impounded water and the hydraulic potential at the bed. Wherever the altitude of the lake surface equals

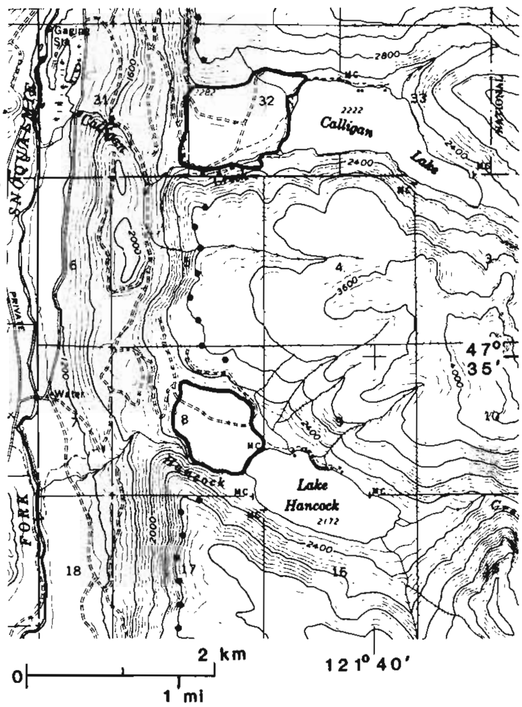


Fig. 11. Calligan and Hancock embankments (outlined). Topography from USGS Mount Si 15′ quadrangle. Ice-maximum limit shown by black dots; altitude in feet.

Wherever the altitude of the lake surface equals or exceeds the hydraulic potential at the bed, the ice floats. The exact extent and configuration of the lake ice is irrelevant, for although the ice front will advance as a shelf if it is sufficiently thick (Weertman 1974; Sanderson 1979), the location of the grounding line does not depend on the position of the ice front.

Rising water can float ever-increasing thicknesses of ice, causing the grounding line to migrate upglacier. As lake level approaches the hydraulic potential of the subglacial spillway and drainage becomes imminent, the grounding line lies across the ice tongue just downglacier of this spillway (Fig. 12). Subsequent drainage may allow the grounding line to migrate upvalley some distance, but repeated cycles of deposition during lake filling will tend to stabilize the ice front. This is partially due to both the effect of shoals (Post 1975) and the shape of natural reservoirs. Water level rises most slowly, and so the grounding line migrates most gradually, when these lakes are filled near to their maximum depths. Thus the grounding line will remain longest at positions

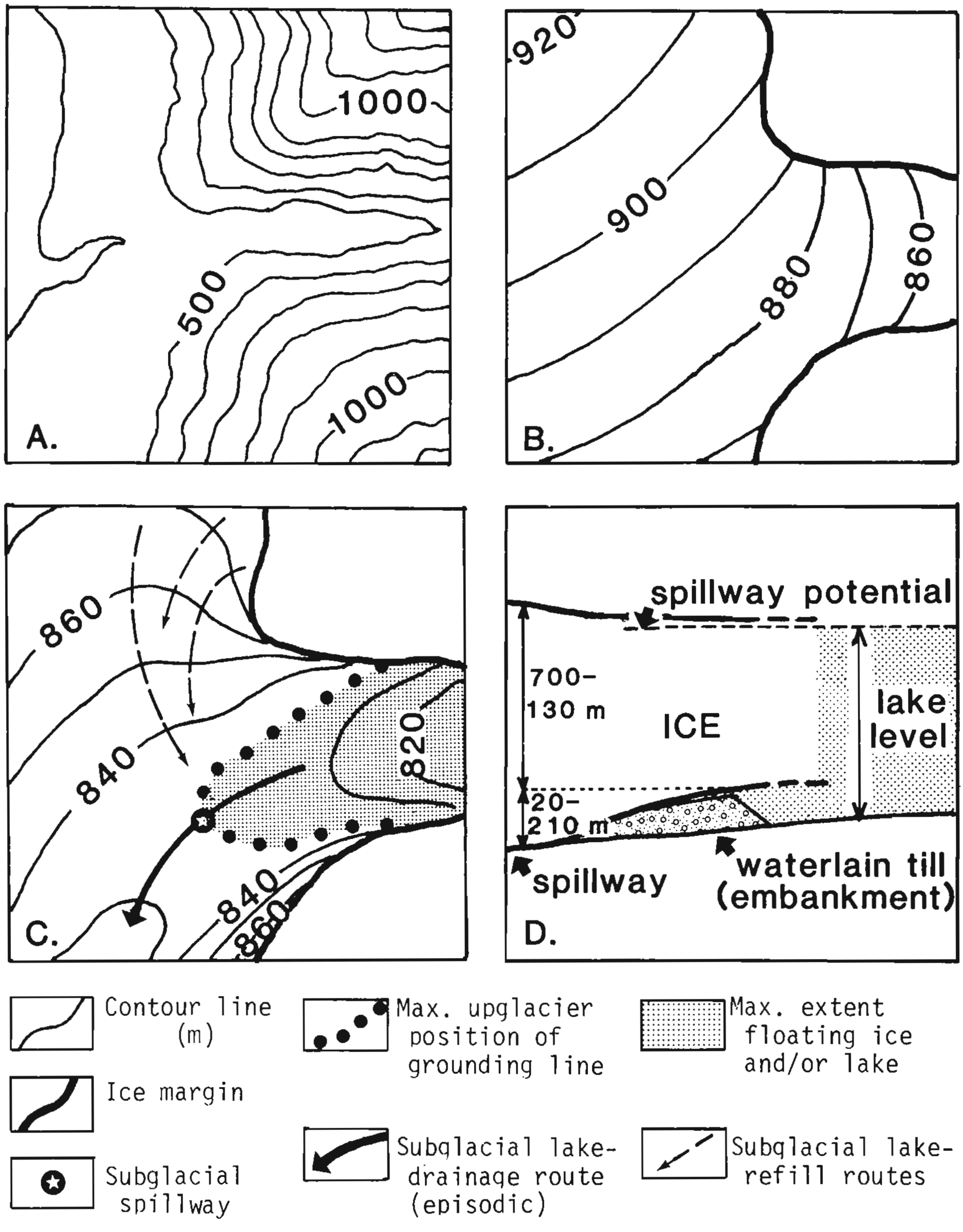


Fig. 12. Reconstruction of subglacial water-flow paths and spillway in a hypothetical valley dammed by active glacial ice. A: Subglacial bed topography; contour interval 100 m. B: Ice-surface topography; contour interval 10 m. C: Subglacial hydraulic equipotentials; contour interval 10 m. Grounding line for a given lake level defined by the position of the equipotential contour whose value equals the lake-surface altitude. The maximum lake altitude will be determined by the potential of the subglacial spillway. Deposition of embankment sediments occurs in the shaded area, at greatest rates near to the maximum upglacier grounding-line position. Note gradient of the hydraulic potential adjacent to ice margin on the steep sideslopes; extensive marginal drainage not physically possible in these areas. D: Cross-sectional view of ice dam, impounded lake, and subaqueous embankment sediment. Observed ranges of embankment heights and maximum ice thicknesses in the eastern Puget Lowland are shown (from Table 1).

closest to that of the spillway. Continued sedimentation there will tend to stabilize that location.

Subglacial spillways are located in the Lowland just south of the mouth of most valleys (Figs. 8 and 12). This geometry focuses subaqueous deposition at the valley mouths and explains the tendency for embankments to form in these positions. These spillway locations, and thus the grounding-line positions, are quite insensitive to changes in the ice thickness (which can be easily if somewhat tediously checked by alternative reconstructions with thinner ice using eqn. 1). Deposition during advance, maximum, and retreat stages therefore will be at virtually the same location.

The Sultan River, by contrast, has a spillway that is not in the Lowland proper (Fig. 7A). The embankment in this valley also occupies an atypical position. As predicted by this analysis, the diamicton (unit 2 of Fig. 7B) of this embankment in the main Pilchuck-Sultan valley was deposited at a point just upvalley of where the subglacial potential (in the main valley) equals the potential of the spillway (located in the Sultan gorge; cf. Fig. 7A).

Other models

The physical processes described here are similar to those invoked by Hillaire-Marcel *et al.* (1981) for formation of 're-equilibration moraines', localized deposits at ice-water margins during uniform retreat of the eastern Laurentide ice sheet. Their model implicitly assumes sedimentation only during periods when the ice is fully grounded in shallow water, with neither a floating nor an actively calving margin. The processes that actively contribute sediments to this environment, however, will be active at the grounding line regardless of the configuration of the ice farther downglacier. This resolves their acknowledged difficulty (Hillaire-Marcel *et al.* 1981:212) in explaining large sediment volumes, hypothesized to accumulate only during the few years that the ice front actually terminates at a specific locality.

Other settings proposed for deposition of fluvial and ice-contact sediment in other localities are clearly not applicable to the initial formation of the Puget embankments. Deltaic sedimentation into a proglacial lake, described by Orombelli & Gnaccolini (1978), lacks a significant basal-meltout component because water depth in their example was controlled by a subaerial spillway at an altitude too low to float the ice. This process, however, was active following the ice-dammed stage of deposition at several of the Puget embankments (e.g. Pilchuck-Sultan, Skykomish, North Fork Tolt, and North Fork Snoqualmie), as reflected by planar topset surfaces and rarely observed upvalley-dipping forsets. Other generalized ice-contact lacustrine environments, such as described by Koteff (1974), all have subaerial spillways distinct from the damming ice. The location of deposition in such cases is therefore determined by the position of the ice front itself during temporary stillstands. This contrasts with the Puget embankments, whose locations were determined by a grounding line fixed by the pattern of subglacial hydraulic potentials and resultant subglacial spillways. Although focused on a particular geographic region, this analysis should apply to any locality where modern or Pleistocene glaciers have terminated against ice-dammed, subglacially draining water bodies.

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Ice age mammals and environments

MATTI ERONEN

Sutcliffe, Antony J. 1985: *On the Track of Ice Age Mammals*. 224 pp. British Museum (Natural History). Price GBP 12.95.

This book is a comprehensive treatment of the biological and physical developments during the Quaternary period. Although the Ice Age mammalian fauna referred to in the title figure prominently throughout the entire book, many different aspects of environmental changes during the Quaternary time are also discussed. Both the title and the appearance of the book are 'selling': on the dust jacket, for example, there is a colourful picture of an Ice Age landscape with a herd of woolly mammoths. Five such colour plates depicting different Pleistocene scenes can be found inside the book, which as a whole is well illustrated with many black-and-white photographs, drawings, maps, and diagrams. The book is aimed primarily at specialists working on different fields related to Quaternary research, but it should also appeal to a wider audience.

Although Antony J. Sutcliffe is an expert in mammalian palaeontology, he skilfully provides an overview of all the main lines of modern Quaternary research in this book. He opens with a discussion of Quaternary climatic variations and their consequences and follows with a fascinating description of the early interpretations of fossil remains. For example, early workers identified animal bones unearthed in the past centuries as remains of such fanciful creatures as dragons and unicorns. The next chapter is concerned with our present knowledge of