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Chapter 4

Timing and processes of deglaciation along the southern margin of the Cordilleran ice sheet

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INTRODUCTION

Approach

The Cordilleran ice sheet covered the northwest part of the North American continent during the last glaciation (Fig. 1). It developed from an initial core of coalescing mountain glaciers on Vancouver Island and the British Columbia mainland, spreading outward over a period of 10,000 to 15,000 yr. South of the ice-sheet limit, isolated alpine glaciers fluctuated in size, leaving a similar but not identical record of glacial advance and retreat. From the behavior of these glaciers, three questions have been posed and, in part, addressed by previous studies (including Crandell, 1965; Porter, 1976; Hicock and others, 1982; Clague, 1981; Waitt and Thorson, 1983). (1) Why did the Cordilleran ice sheet attain its maximum 5,000 yr later than its smaller counterparts to the south? (2) What was the extent of the alpine glaciers during the ice-sheet maximum, and why were most apparently well back from their maximum position? (3) What physical factors determined the rate and character of final ice-sheet retreat?

This chapter approaches these questions by applying current knowledge of glacial mechanics, both theoretical and empirical, to various aspects of the inferred or reconstructed Cordilleran glaciers. The record of advance and retreat should reflect changes in the external environment (regional climate, sea-level changes), the glaciers' physical responses to those changes, and changes that in turn result from ice growth and decay (isostasy, local climate). Existing data on the late-glacial advance-retreat chronology and climate constrain this analysis and provide an independent check on the conclusions suggested by this approach.

Although the primary focus here is on the mechanics of deglaciation, this chapter considers the record of Cordilleran ice advance as well for several reasons. Retreat of most of the other North American ice sheets (Prest, 1969) coincided with the Cordilleran ice's achievement of maximal advance, about 15 ka (Clague and others, 1980). Retreat of Cordilleran glaciers was neither monotonic nor entirely synchronous, in that some glaciers had retreated from their maximum positions 5,000 yr before others had finished advancing. Finally, the full advance-retreat record characterizes the physical behavior of the Cordilleran glaciers more completely, because the same physical principles determine glacier behavior regardless of the direction of motion of the ice terminus.

The most detailed and best studied record of ice advance and retreat is found along the southern margin of the Cordilleran ice sheet, particularly in the Puget and Fraser Lowlands. Because the chronology, physical setting, and environmental changes are particularly well constrained in this region during late-glacial time, the geologic record here offers an excellent opportunity to evaluate the responses of glaciers of various sizes to late Pleistocene environmental changes and interactions among them.

The following analysis of those responses is obviously indebted to nearly a century of geologic investigations of this region, together with recent advances in climatology, palynology, and glaciology. I have also benefited greatly from prior efforts to quantify the physical behavior of Pleistocene glaciers in western North America (Pierce, 1979; Thorson, 1981).

Regional Setting

The southern part of the Cordilleran ice sheet occupied a distinctive physiographic region (Fig. 2). In British Columbia a broad topographic basin extends southeastward as the Georgia Depression and the Fraser Lowland, bounded by Vancouver Island and the mainland coast mountains. The basin continues south into Washington State at the Puget Lowland, bordered by the Olympic Mountains on the west and the Cascade Range on the east. Low hills at about 46°45′ define the southern limit of glacial advance in this basin, but the lowland province itself extends south for at least several hundred kilometers more. The basin has persisted for millions of years, as shown by the great thickness of its Cenozoic fill (e.g., Weaver, 1937). East of the Cascade Range the ice sheet extended from the mountainous highlands in Canada onto the northern edge of the Columbia Basin, primarily along major south-trending valleys.

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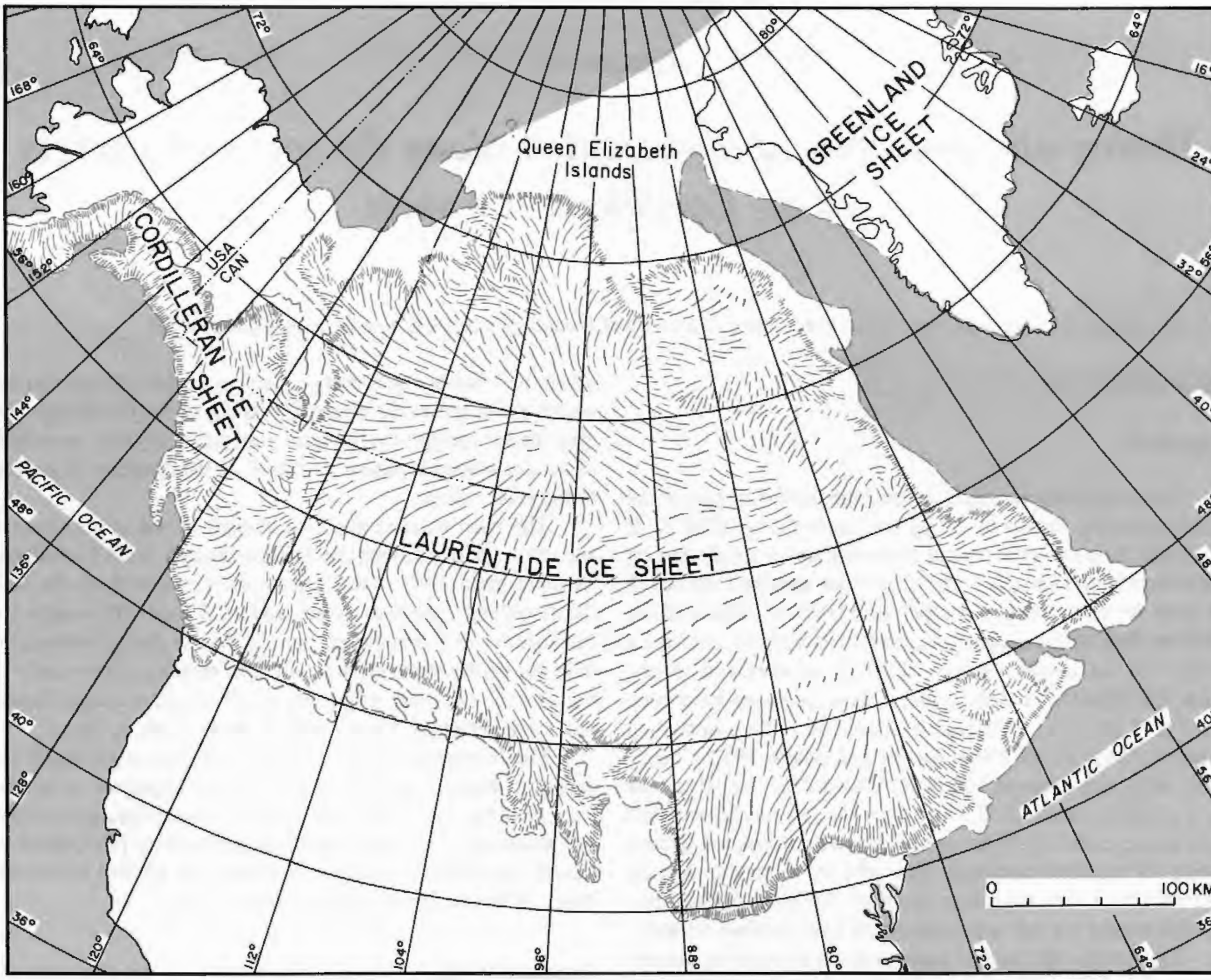

Figure 1. Extent of North American ice sheets during the last glaciation. Dashed lines indicate maximum proposed reconstructions along the northern, eastern, and southern boundaries.

In the west at latitude 48°15′, the Strait of Juan de Fuca separates Vancouver Island from the Olympic Mountains and opens westward to the continental shelf. Bottom depths in the strait, typically 100 to 200 m below modern sea level, are comparable to those throughout the Georgia Depression and in the deeper parts of the Puget Lowland that are occupied by Puget Sound. They lie more than 2,000 m below the summits of the adjacent coastal mountains. In contrast, the continental shelf is a low-relief surface over 50 km wide, gently sloping seaward to a depth of about 200 m before dropping precipitously down the continental slope.

Growth of the Cordilleran ice sheet has occurred several times during the Pleistocene, leaving a discontinuous record of glacial advances throughout the Puget and Fraser Lowlands (Crandell and others, 1958; Armstrong and others, 1965; Easterbrook and others, 1967). By analogy with the well-documented pattern of the last glaciation (the Fraser glaciation of Armstrong and others, 1965), coalescing mountain ice caps advanced into the Georgia Depression and split around the Olympic Mountains into the Juan de Fuca and Puget lobes (Fig. 2). East of the Cascade Range, ice from the eastern slopes of the Canadian Coast Ranges probably met west-flowing ice from the Rocky Mountains of Canada, extending south as a series of lobes and sublobes into eastern Washington, Idaho, and Montana. The Cascade and Olympic mountains south of the ice-sheet limit were locally high enough to support isolated mountain glaciers as well.

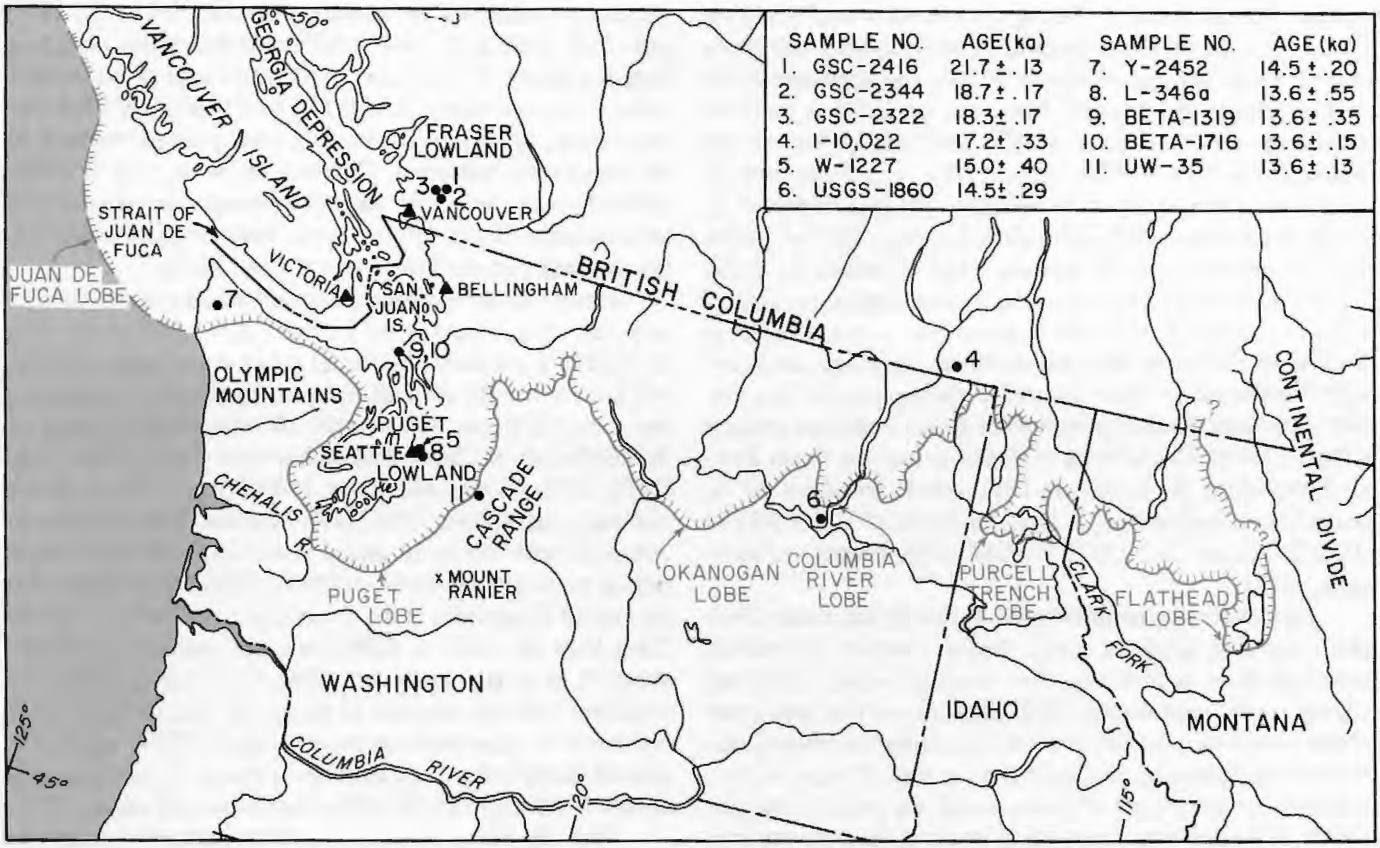

Figure 2. Index map of the southern part of the Cordilleran ice sheet. Location of critical radiocarbon dates discussed in text are indicated by sample number and age in thousands of years (ka).

GEOLOGIC AND CLIMATOLOGIC DATA

Chronology of the Fraser glaciation

Introduction. Although the history of glaciations and interglaciations for the Cordilleran ice sheet is poorly constrained for most of the Pleistocene, the record of the latest glaciation is well documented and extensively dated. Armstrong and others (1965) named and subdivided the Fraser glaciation into three stades and one named interstade (named glaciations, interglaciations, stades, and interstades are used informally in this report, although their original use had presumed them to be formal geologic-climate units). Although the relative ranking of these intervals continues to be deliberated, their overall stratigraphy provides a useful and widely accepted framework for this period. Other recent summaries of the late Pleistocene glaciation are found in Armstrong (1981), Clague (1981), Waitt and Thorson (1983), and Pessl and others (1987).

Pre-Fraser interval. Preceding the Fraser glaciation, nonglacial conditions prevailed in the lowlands of the Pacific Northwest. Named the Olympia interglaciation by Armstrong and others (1965), deposits associated with this period are as old as 40.5 ±1.7 ka and perhaps 58.8 +2.9/−2.1 ka (^{14}C sample numbers GSC-2167 and QL-195; Clague, 1981, p. 5). The end of the Olympia interval is marked stratigraphically by glacial deposits derived from either the expanding Cordilleran ice sheet or from local mountain glaciers in the Cascade Range and Olympic Mountains. Because of the great distance between these glacial-sediment sources, local dates indicating the end of the Olympia nonglacial interval span nearly 10,000 yr.

Growth of interior ice cap. In response to climatic deterioration at the close of the Olympic nonglacial interval, mountain ice caps on Vancouver Island and the British Columbia mainland grew and coalesced. Proglacial outwash, particularly along the Georgia Depression, records the slow southward expansion of this ice. Minimum dates for the outwash include 28.8 ±0.74 ka (GSC-95) in the northern Georgia Depression, 18.3 ±0.17 and 18.7 ±0.17 ka (GSC-2322 and GSC-2344) near Vancouver, and 15.1 ±0.4 ka (W-1227) near Seattle (Dyck and Fyles, 1963, p. 49–50; Armstrong and Clague, 1977, Fig. 1; Mullineaux and others, 1965, p. 7).

Early Fraser maximum stage. During the buildup of ice in the British Columbia mountains, a more short-lived expansion of alpine ice occurred in the mountains of Washington and southern British Columbia. During this interval, named the Evans Creek stade (of the Fraser glaciation) by Armstrong and others (1965), valley glaciers on Mount Rainier and elsewhere in the Cascade Range and Olympic Mountains expanded to their late Pleistocene maximum positions (Crandell, 1963; Crandell and Miller, 1974; Carson, 1970; Porter, 1976). No absolute ages directly date this advance in Washington, but it is correlated by Porter and others (1983) with a glacial advance into the eastern Fraser Lowland from the adjacent coast mountains in British Columbia. Although this correlation implies a regional episode of climatic cooling or increased precipitation, nowhere north of Vancouver or east of the Cascade Range have any conclusive signs been found of other correlative glacial episodes that similarly punctuate the slow growth of the main Cordilleran ice sheet (Clague, 1981). The advance of alpine ice into the Fraser Lowland, depositing the Coquitlam Drift, culminated about 20 ka, according to limiting dates in fluvial sediment of 21.7 ± 0.13 ka (GSC-2416) and 18.7 ± 0.17 ka (GSC-2344; Hicock and Armstrong, 1981).

Lowland nonglacial interval. Following the Evans Creek and Coquitlam advances, alpine glaciers retreated to undetermined positions in their respective mountain valleys. Cary and Carlston (1937) and Mackin (1941) first inferred that these lower alpine valleys were ice-free at some time during the (subsequent) lowland occupation by the Cordilleran ice sheet. Direct evidence, however, for the retreat of Evans Creek ice prior to the *first* arrival of the southward-expanding ice sheet has been reported only in the Fraser Lowland (dated fluvial sediments overlying Coquitlam Drift; Hicock and Armstrong, 1981, sample GSC-2344) and just east of the central Puget Lowland (undated fluvial sediments between alpine and ice-sheet tills; Booth, 1987a).

Ice-sheet advance to maximum. The Vashon stade of the Fraser glaciation of Armstrong and others (1965) encompasses the growth of the Cordilleran ice sheet to its late Pleistocene maximum position (Figs. 3 and 4). In the Fraser and northern Puget Lowlands, this ice occupied the terrain probably covered previously by glaciers of the Evans Creek/Coquitlam advance after 18.3 ka (GSC-2322; Armstrong and Clague, 1977, Fig. 2; see also Clague and others, 1980, Table 1). In this region the ice sheet probably merged with the Vashon-age remnants of these alpine glaciers. At the lateral margins of the central and southern Puget Lowland, tongues of the ice sheet extended up into alpine valleys over areas previously occupied by Evans Creek glaciers (Knoll, 1967; Williams, 1971; Booth, 1986a, 1987a), but at no time did ice from these two sources meet. A readvance of alpine glaciers, of smaller magnitude than during Evans Creek time, has been tentatively correlated with the Vashon advance (Porter, 1976).

Precise dating of the Vashon maximum suffers from a paucity of dates near the terminus. Postglacial sediments on the northwestern Olympic Peninsula, about 50 km behind the maximum ice limit of the Juan de Fuca lobe, specify a minimum retreat date of 14.46 ±0.20 ka (Y-2452; Heusser, 1973). In the Seattle area, 100 km north of the Puget-lobe limit, the Vashon maximum is bracketed between dates of 15.0 ±0.4 ka (W-1227) and 13.65 ±0.55 ka (L-346a) (Mullineaux and others, 1965, p. 7; Rigg and Gould, 1957). East of Seattle at the edge of the lowland, wood in lake sediments dated at 13.57 ±0.13 ka (UW-35; Porter and Carson, 1971) provides the only other possible constrain on the Puget-lobe maximum. The lake, however, was probably debris-dammed long after ice had retreated from the adjacent lowland areas (Booth, 1987b); active Vashon-age ice had probably retreated from the Seattle area before 13.6 ka.

Limits on the Fraser maximum are broader east of the Cascade and Coast ranges. They include a date of 17.24 ±0.33 ka (I-10,022) in proglacial (advance) outwash just north of 49°N, 100 km north of the ice limit, followed by advance to maximum and retreat of at least 80 km in the following 6,000 yr, based on the distribution of Glacier Peak tephra-layer G at 11.18 ±0.15 ka (WSU-2668) (Clague and others, 1980; Porter, 1978, p. 38–39; Mehringer and others, 1984). Lake Missoula flood deposits, interbedded with varves of glacial Lake Columbia that contain detrital wood dated at 14.49 ±0.29 ka (USGS-1860), suggest that the Purcell Trench lobe of the Cordilleran ice sheet lay across the Clark Fork for 2,000 to 3,000 years and reached its terminus about 15 ka or later (Atwater, 1986). These interpretations are consistent with the presence of Mount St. Helens tephra-set S between Lake Missoula flood deposits (Waitt, 1980), which indicates blockage of the Clark Fork by the Purcell Trench lobe until at least 13.08 ±0.30 ka (W-3404; Mullineaux and others, 1978).

Thus the Fraser maximum on both sides of the Cascade Range was at least broadly synchronous (Waitt and Thorson, 1983, p. 67). The Fraser maximum coincides with the maximum advance of parts of the Laurentide ice sheet in central North America but lags 2,000 to 6,000 yr behind the culmination of most of the rest of the Laurentide ice sheet (Prest, 1969; Mickelson and others, 1983).

Ice-sheet retreat and the Everson interval. Rates and timing of the initial retreat of the Puget and Juan de Fuca lobes are relatively well constrained by radiocarbon dates. The Juan de Fuca lobe retreated 50 km by 14.46 ka (Y-2452; see above). The next-oldest sediments are found in the north-central Puget Lowland (48°30′N latitude), 150 km north of the Puget-lobe limit and 250 km east of the Juan de Fuca limit. Dates of 13.60 ±0.15 ka (BETA-1716) and 13.65 ±0.35 ka (BETA-1319) (D. P. Dethier and others, written communication, 1986) from shells in glacial-marine sediment specify the maximum southern limit of grounded ice at this time.

During the preceding 1,000 years, the rate and pattern of ice retreat are known only indirectly from undated landforms and deposits. North of about 47°30′N latitude, the western part of the Puget lobe left little erosional or depositional record of its retreat, suggesting a rapid and chaotic decay (Thorson, 1980). South of this line and near the eastern Puget-lobe margin, abundant marginal channels and sequential recessional deposits imply a systematic and probably slower marginal recession (Newcomb, 1952; Crandell, 1963; Curran, 1965; Anderson, 1965; Knoll, 1967; Thorson, 1980; Booth, 1987a, 1987b). Conversely, irregu-

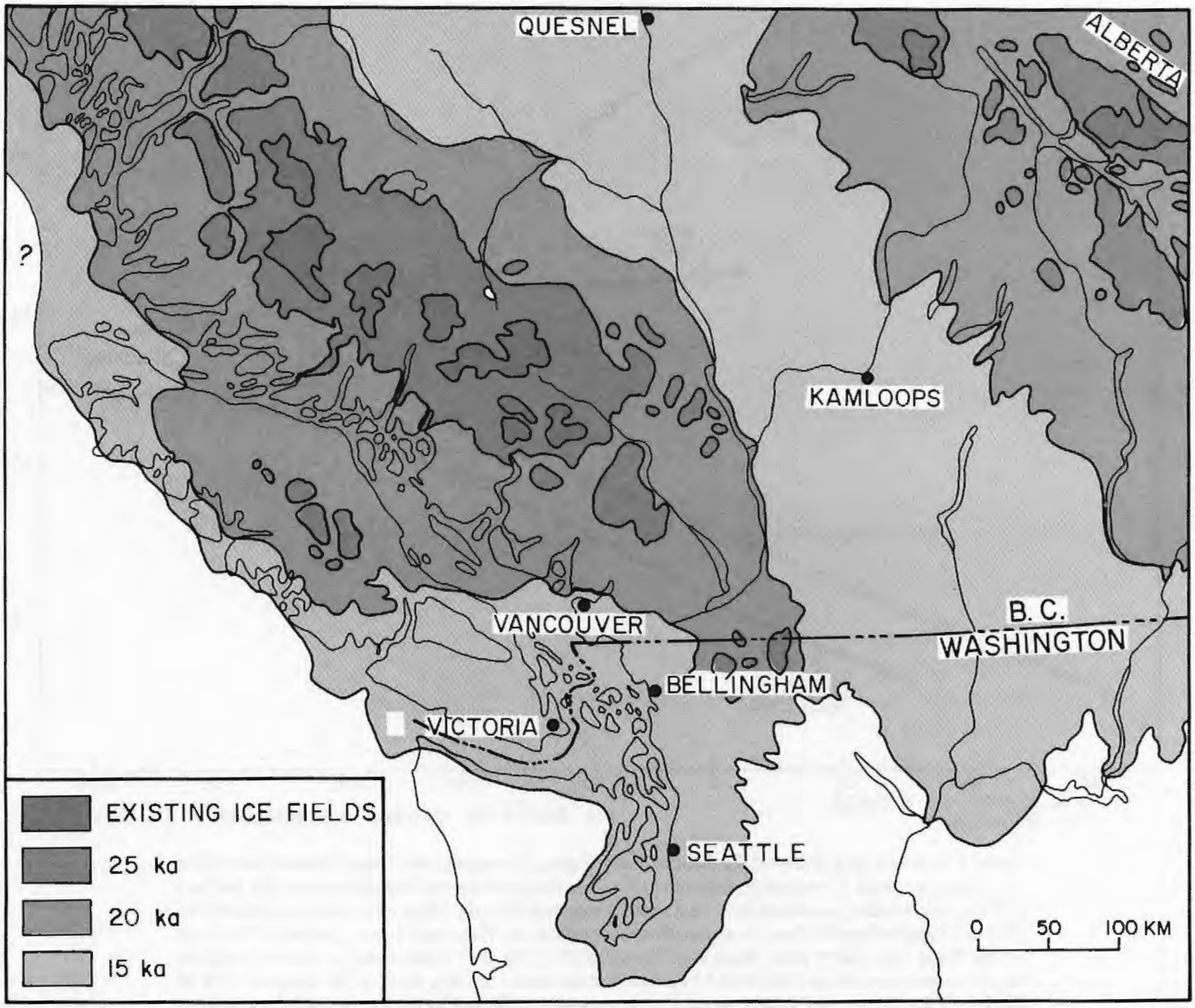

Figure 3. Extent of the Cordilleran ice sheet at 25, 20, and 15 ka (from Clague, 1981).

lar constructional landforms in the eastern Strait of Juan de Fuca (Anderson, 1968; Chrzastowski, 1980) suggest rapid retreat of the Juan de Fuca lobe, constrained by the radiocarbon dates to be at least 200 m/yr.

In the central and northern Puget Lowland, incursion of marine water via the Strait of Juan de Fuca resulted in deposition of abundant glacial-marine drift (Easterbrook, 1963; Thorson, 1980; Pessl and others, 1987; Dethier and others, written communication, 1986). Dated marine and glacial-marine sediments in this region are as old as 13.65 ±0.35 ka (BETA-1319) in the north-central Puget Lowland and 13.5 ±0.2 ka (GSC-3124) just south of the International Boundary. Deposition of these sediments continued until at least 11.30 ±0.07 ka (USGS-124) in the central Puget Lowland (Armstrong, 1981; Dethier and others, written communication, 1986).

This suite of dates marks the end of an interval that probably lasted less than 1,000 years, during which the ice lobe retreated 100 km primarily by calving, with local backwasting and stagnation (Pessl and others, 1987). Net retreat was therefore at least 100 m/yr and may have been several times more rapid. Marine sedimentation, with and without glacial input via submarine flows and ice-rafted debris (Domack, 1983), continued for at least another 1,000 yr in areas subsequently lifted above sea level by isostatic rebound.

East of the Cascade Range, ice-sheet retreat probably occurred more slowly (Fig. 4). The limiting dates on the lobes in the

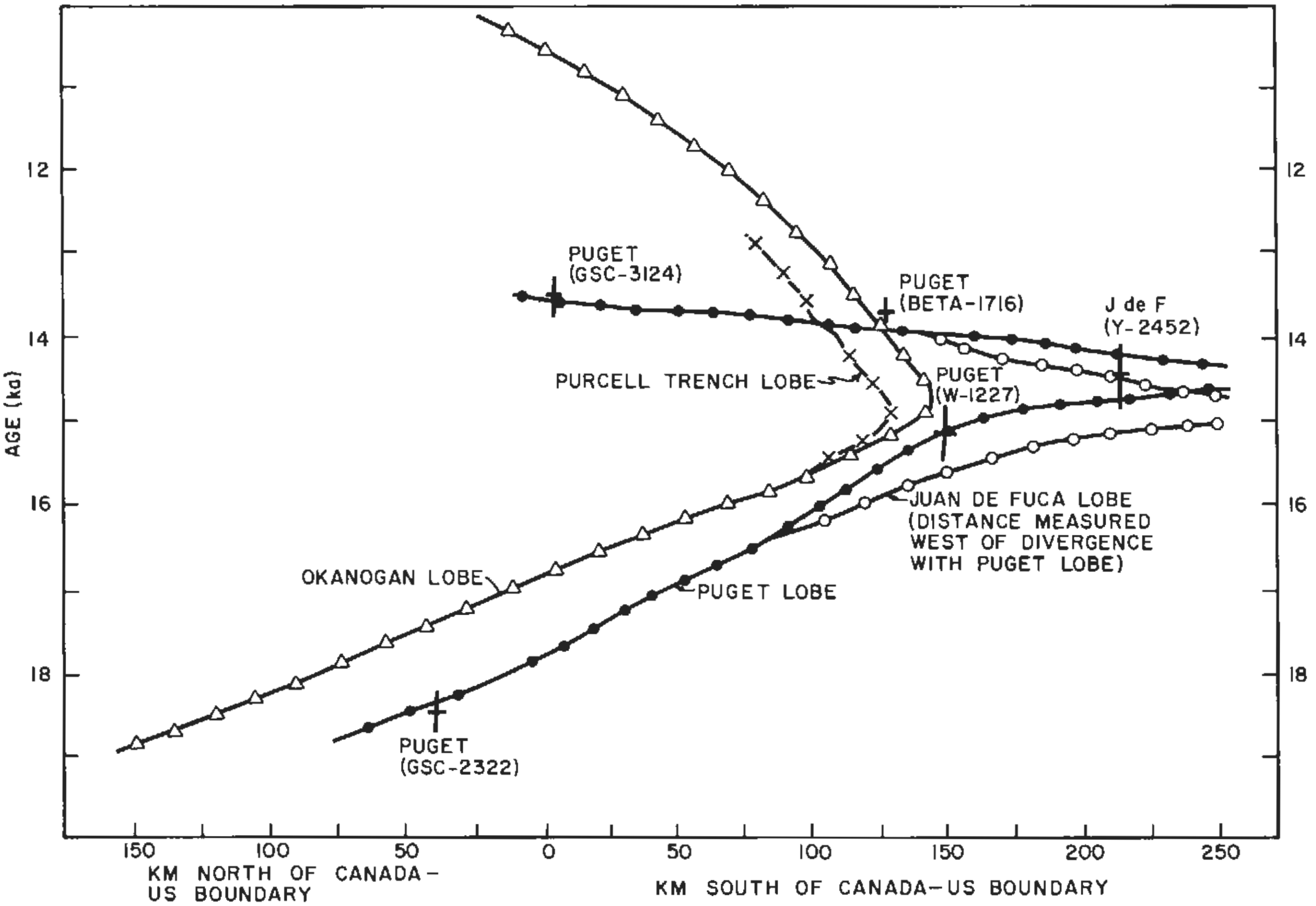

Figure 4. Advance and retreat of the Juan de Fuca, Puget, Okanogan, and Purcell Trench lobes of the Cordilleran ice sheet. Distances are measured relative to the International Boundary except for the Juan de Fuca lobe, which is measured from its western divergence from the Puget lobe. Crosses locate critical dates and reported standard errors on nonglacial deposits in the Puget and Fraser Lowlands. The curve for the Puget lobe differs from Waitt and Thorson (1983, Fig. 3-2) in assuming a relatively uniform advance rate during the period 19 to 15 ka and in more recent limiting dates on the retreat at 13.6 ka (BETA-1716; Dethier and others, written communication, 1986). Curves for the two eastern lobes (Okanogan and Purcell Trench) are from Atwater (1986).

northwest interior indicate 180 km of advance and retreat during the interval 17.24 to 11.18 ka (see above; Mullineaux and others, 1978; Clague, 1981; Mehringer and others, 1984). They permit an interpretation of slower ice retreat in the east (as little as 30 m per year) but do not constrain a hypothesized lag in eastern ice-sheet retreat (Waitt and Thorson, 1983; e.g., see Atwater, 1986, Fig. 28).

Sumas interval and final deglaciation of the lowlands. Glacial sediment in the eastern Fraser Lowland near the International Boundary was initially interpreted as recording a widespread readvance of Cordilleran ice during the Sumas stade of Armstrong and others (1965). Dates on its maximum position range from 11.70 ±0.15 ka (L-3313) to 11.40 ±0.17 ka (GSC-1695), with final deglaciation of the lowland complete by 11.1 ka (Armstrong, 1981). This later work, however, has questioned the climatic significance of the Sumas interval, suggesting instead that the readvance marked only the glacier's response to isostatic emergence of the terminus by reduction in the rate of calving.

Elsewhere in the Pacific Northwest, evidence for a glacial readvance ca. 11.5 ka is ambiguous. East of the Cascade Range the oldest nonglacial date north of 49°N latitude is only 11.00 ±0.18 ka (GSC-909; Lowden and Blake, 1970, p. 71). Additional evidence suggests that retreat across the International Boundary in most areas followed deposition of Glacier Peak tephra-layer G at 11.18 ±0.15 ka (WSU-2668; Mehringer and others, 1984; Clague, 1981, p. 17). Recessional stillstands or readvances of the eastern Cordilleran lobes are recorded by glacial deposits (Waitt and Thorson, 1983, p. 67) but remain undated and uncorrelated.

Alpine glaciers in the Cascade Range show widespread evidence of a late-glacial readvance, the Rat Creek advance of Page (1939; summarized in Porter and others, 1983, p. 86). Originally inferred to predate 11.05 ±0.05 ka (UW-321) and to postdate the 12-ka Glacier Peak tephra-layer M, this advance was correlated provisionally with the Sumas interval by Porter (1978). Additional data on tephra distribution and age, however, cast doubt on both the upper age constraint and its applicability here; they instead suggest only a pre–11.2 ka age that may have no correlation to the regional climate of Sumas time (Porter and others, 1983, p. 87; Mehringer and others, 1984).

Fraser-age lowland climate

Plant-fossil record. The climate of the Pacific Northwest during the Fraser glaciation has been reflected not only in the growth and wastage of glaciers but also by the changing influx of pollen and plant macrofossils into lakes and wetlands of the region. Such data provide an independent picture of temperature and precipitation during this period. The value of this information to glacial geology and reconstruction lies in the spatially discrete and easily dated nature of the samples, which, when correlated among multiple sites over a region, provide a fairly complete picture of the near-glacier climate. The response of local plant communities to climate should also be more rapid than that of a continent-scale ice sheet (Wright, 1984).

Yet as a companion to a study of a glacier's response to climate, the pollen record also has two significant limitations. First, it may be influenced by the physical proximity of the ice, implying a set of climatic conditions that exist only locally. Second, the conditions of temperature and precipitation over the main body of the ice sheet may correlate only poorly with those recorded at or beyond its margins.

Despite these potential problems, a reasonably consistent picture of glacial-age climate has emerged from palynologic study throughout the region that provides a climatic framework to complement the previously described chronology of ice advance and retreat. Barnosky's (1984) synthesis of data from sites in south-central Washington identifies several climatically distinct periods in the Puget Lowland during the Fraser glaciation (see also Barnosky and others, this volume): (1) 26 to 19 ka: cold and dry parkland-tundra environment; (2) 19 to 17 ka: cold and drier than before; (3) 17 to 15 ka: possible warming, indicated by fossil-plant and insect data at some sites (4) 15 to 12.5 ka: cool and increasingly wet maritime conditions; and (5) post-12.5 ka: widespread warming; possibly beginning as much as 1,000 years earlier east of the Cascades (Barnosky, 1985).

Most of these climatic conclusions are in broad accord with direct records of glacial growth and wastage. Both the Evans Creek and Vashon maxima coincide with cold ("glacial") conditions, and the final wastage of Cordilleran ice roughly coincides with the late-glacial warming. Some of the details of these two chronologies, however, fit less readily. The main western lobe of the Cordilleran ice sheet must have advanced south into the Fraser Lowland during either the period of enhanced aridity before 17 ka or immediately following it, during a time of possibly increased warmth in the southern Puget Lowland (the "unnamed interstade" of Barnosky, 1981, 1984). To explain this apparent contradiction, Hicock and others (1982) suggest an accompanying increase in precipitation. Retreat of the Juan de Fuca and Puget lobes at ca. 15 ka appears to have occurred during a time of increasingly cool and wet climate, conditions that intuitively should *favor* ice-sheet stability and growth. The initial and most rapid retreat of ice west of the Cascade Range preceded general warming by several thousand years, an asynchrony that is difficult to explain without postulating a lag in the vegetational response of 2,000 yr or more (see also Barnosky and others, this volume).

East of the Cascades, pollen-recorded climatic fluctuations are quite sparse. Barnosky's reconstruction for Fraser time, based on a single site in the Columbia Basin, only specifies a long interval of periglacial steppe or tundra from 23.5 to 10 ka. Glacier fluctuations also are not as well constrained as those west of the Cascades, rendering any present effort to make a detailed comparison of eastern and western chronologies nearly meaningless.

Global climate models. Numerical simulations of late Quaternary climate provide a supplementary view of some of the external controls on ice-sheet growth and decay. All available simulations share the disadvantage of low spatial resolution; the entire region of the Cordilleran ice sheet is typically represented by only a few grid points. Nevertheless, these efforts provide an independent view of probable large-scale climatic conditions during glacial time and help identify certain determinants of ice-sheet behavior.

Model simulations of the ice-age climate, typically at ca. 18 ka, generally agree with one another in their characterization of conditions in the Pacific Northwest region (e.g., Gates, 1976; Manabe and Hahn, 1977; Kutzbach, this volume). Relative to present conditions here, sea-surface temperatures estimated from ocean-core data were 2° to 4°C colder (CLIMAP, 1976), land temperatures were less than 5° colder beyond the ice-sheet limit and up to 20° colder toward its interior (Manabe and Hahn, 1977, Fig. 20; Kutzbach and Guetter, 1986, Figs. 5 and 10), and both precipitation and runoff differed by less than 1 mm/day (Manabe and Hahn, 1977, Figs. 8b and 9; Kutzbach and Guetter, 1986, Figs. 9 and 14). The primary deglacial change in the region was an increase in temperature (Kutzbach, this volume); neither precipitation nor prevailing winds show dramatic change during the period 18 to 12 ka.

THE PHYSICAL BEHAVIOR OF THE CORDILLERAN GLACIERS

The ice build-up phase

The alpine-glacial maximum. Alpine glaciers from the mountains flanking the Puget and Fraser Lowlands advanced to their maximum position during the Fraser glaciation, probably about 20 ka. Although the main Cordilleran ice sheet was still several hundred kilometers and 5,000 years away from its maximum stand, this widespread alpine-glacial maximum must record a period of regional climate change that favored ice growth. Their subsequent, pre-Vashon retreat in turn required no more than a minor warming or drying.

The most useful parameter to quantify the effect of climate on a glacier is the glacier's equilibrium-line altitude (ELA). This value defines the altitude at which the yearly gain of snow equals its yearly loss by ablation. Above the ELA lies the accumulation zone, where a year's supply of snow is not completely melted; below lies the ablation zone, where a net yearly loss of mass occurs. The rate at which mass is added or subtracted from the glacier is a function of position on the glacier and is proportional to glacier-surface altitude relative to the ELA (the "balance gradient"; cf. Schytt, 1967).

The ELA of a glacier can be estimated by several methods. A regional relationship of height to mass balance has been compiled from data on seven modern Pacific Northwest maritime glaciers (Fig. 5; Meier and others, 1971; Porter and others, 1983). If a glacier lies in a similar climatic regime and its boundaries and ice-surface contours can be reconstructed at equilibrium conditions, this curve can be used directly to find an ELA that brings the glacier into balance (see below). A simpler approximation is to assume a ratio of the accumulation area to the total glacier area (the accumulation-area ratio, or AAR) of 0.60 ± 0.05, observed on modern maritime glaciers (Porter, 1975). Finally, the regional ELA can also be estimated by the glaciation threshold method (e.g., Porter, 1977).

These techniques have been applied to various regional and local areas in the Cascade Range and Olympic Mountains (Porter, 1977; Williams, 1971; Waitt, 1977; Booth, 1986b, 1987b). Median ELA values in northern and central Washington lie near 1,000 m for Evans Creek time, about 900 m lower than present. A pronounced east-west decrease in ELA across the Cascade Range, inferred to reflect greater precipitation on the Pacific side of the mountains, complicates this picture. At latitude 47°30′N, for example, the glaciation threshold descends westward from over 1,500 m at the Cascade crest to 750 m at the western rangefront, a gradient of about 10 m/km (Porter, 1977, Fig. 7).

Glaciers that occupied the west-draining alpine valleys probably integrated the effect of traversing a range of local ELA values. Inspection of Porter's (1964, 1977) data suggests that such composite values probably lay in the range of $1{,}100 \pm 100$ m for glaciers that extended from the crest of the range to their base in the Puget Lowland. Those glaciers that headed in subsidiary ridges west of the main Cascade crest, such as documented by Williams (1971) and Booth (1987a, 1987b), experienced wetter conditions and had ELA values about 100 to 200 m lower. Conversely, those glaciers that did not extend as far west as the Cascade rangefront (e.g., the reconstructed Skagit Valley glacier in Waitt, 1977, Fig. 2) had median ELAs one to several hundred meters higher.

The rate and magnitude of alpine-glacier retreat is poorly documented. Porter (1976) suggests that moraines in the southern North Cascade Range lying 20 km upvalley of the Fraser ice-sheet maximum may indicate the minimum amount of retreat that occurred during the millennia intervening before the ice-sheet maximum, equivalent to the removal of a layer of ice 300 to 400 m thick (Porter, 1976, Fig. 6). In the time interval available, this would have necessitated an annual deficit of 1 m of ice or less, corresponding to an approximate 100-m rise in the ELA (Fig. 3; also Østrem, 1975). An ELA rise of this magnitude represents only about 10 percent of the maximum Fraser-age depression. Even with an additional rise in the ELA of about 100 m to account for these reduced glaciers (Porter, 1976, Fig. 6), the climatic warming or drying represented by post–Evans Creek glacial retreat represents but a small fraction of the maximum Fraser-age climatic change. The potential effect of isostatic depression by the ice lobe in the adjacent Puget Lowland does not substantially affect this conclusion (see below).

Ice advance along the Georgia Depression. *Introduction.* While most alpine glaciers in the Pacific Northwest probably reached their late-glacial maxima about 20 ka, the Cordilleran ice sheet was still 5,000 years from reaching its farthest southern limit. Plausible explanations for this asynchrony include differences in glacier response times or in terminus position relative to sea level. Although the ice sheet typically is characterized as advancing steadily down the Georgia Depression during this 5,000-year interval, no evidence supports this assumption (Clague, 1981, p. 10–11). Such data, however, might be recorded only along the axis of the Georgia Depression in sediments that either are now under hundreds of meters of water or were completely eroded away after their deposition.

Response times. To any perturbation in mass balance, such as that caused by a change in temperature or precipitation, a glacier will respond by a net increase or decrease in its mass that ultimately results in movement of its terminus. Climatic change and the resulting glacier adjustments, however, are not related simply. The flowing ice averages the effects of short-term climatic fluctuations (Kuhn, 1981), and the response at the terminus lags behind the long-term climatic trend. Because the geologic record typically documents only the position of the terminus through time, distortions and delays characterize this geologic reflection of paleoclimate.

Nye (1960, 1963, 1965a, 1965b) has attempted to quantify most comprehensively the relationship between climatic change and glacial fluctuation. He considers a two-dimensional glacier resting on a sloping planar bed. The discharge of ice at any cross section depends on the position downglacier, ice thickness, and ice-surface slope. By considering a small perturbation of the mass balance, he shows that the disturbance should move along the glacier's length as a kinetic wave—a "wave" of constant discharge per unit of cross-sectional area that moves through the glacier.

Nye's theory predicts not only changes in thickness and discharge but also response times needed to accomplish these

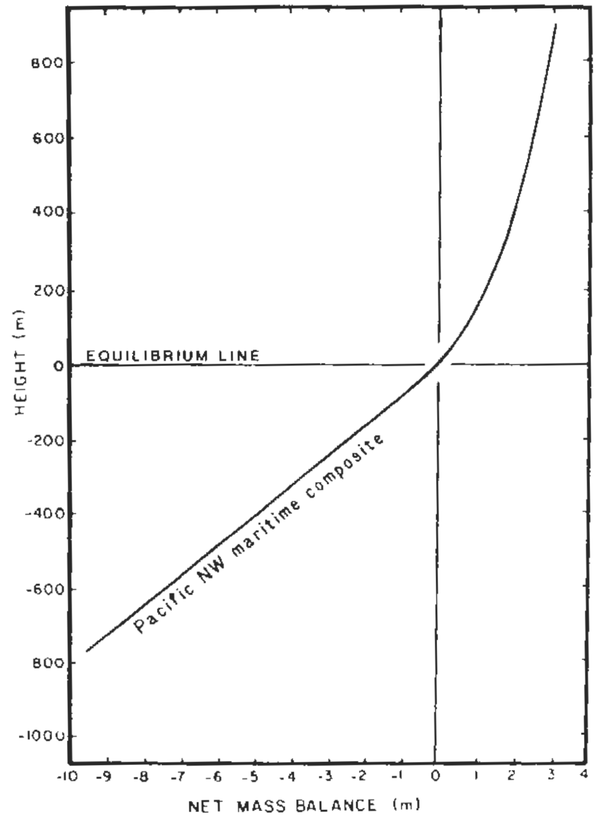

Figure 5. Height–mass balance relationship compiled from seven modern maritime glaciers in the Pacific Northwest and Alaska. Height is referenced to the equilibrium-line altitude; net mass balance is in cubic meters of equivalent water volume per square meter of ice surface per year. Original data are from Pacific Northwest glaciers (Meier and others, 1971), summarized in Porter and others (1983).

changes by making several simplifying assumptions. The response time is equated with how long such a wave takes to traverse the length of the glacier. This velocity in turn depends on the assumed velocity of the ice at the terminus and on the diffusion of the kinematic wave as it propagates downglacier. Later calculations for actual glaciers, however, suggest that the predicted response time may be significantly too long (Meier and Tangborn, 1965; Nye, 1965a; Johannesson, 1986).

Despite uncertainties in this analysis, the theory predicts that response times should increase linearly with glacier length and inversely with ice velocity. These results are intuitively reasonable and provide a quantitative basis to compare disparate glacial advance-retreat histories.

An alternative calculation of response time is presented by Oerlemans and van der Veen (1984), whose calculations for a simplified ice sheet are based on "typical" values of the ice thickness, ice velocity, and mass balance. Weertman's (1957) expression for sliding velocity is used to express the velocity as a function of ice thickness and length. Changes in ice thickness are then evaluated by incremental changes in the mass balance, defining a "relaxation time" in which the glacier responds. Although the magnitude of the relaxation time depends on poorly quantified sliding parameters, the form of its dependence on physical parameters of the glacier is fairly simple to obtain. Because most ice-sheet profiles can be approximated by a parabola (e.g., Mathews, 1974; Thorson, 1980; Paterson, 1981, p. 157), ice thickness is generally proportional to the square root of the distance from the terminus. Substituting this relationship in the expression for relaxation time predicts that relaxation time is proportional simply to the length of the glacier. This result resembles Nye's, in that the response time depends linearly on length. An implicit dependence on velocity also exists via the sliding parameter.

A third method of evaluating response times exists (Johannesson, 1986; Johannesson, Raymond, and Waddington, written communication, 1986). A particular flow law is not assumed; rather, mass is simply assumed to be conserved over the glacier as a whole. A uniform perturbation over the glacier is imposed on the equilibrium steady-state mass balance. In response, the glacier grows (or shrinks) to a new length but is assumed to change only a little in height. The change in volume to achieve the new equilibrium shape divided by the rate of gain (or loss) of ice volume defines a minimum response time, which can be expressed simply as the maximum ice-sheet height divided by the ablation rate at the terminus. This is not a "dynamic" response time, because no formulation of ice velocity is included. Yet the calculated time should both rank glaciers by relative rate of response and specify the minimum time needed to achieve final equilibrium. The expression resembles the preceding two by a dependence of response time on glacier length, here proportional to the square root of glacier length for parabolic ice-surface profiles, and an implicit dependence on the ice velocity, which is driven by the rate of ablation for an ice sheet at or near equilibrium.

Opportunities to verify any of these methods are sparse. In one example, this last method predicts a response time of 20 yr using four years of data from 100-m-thick South Cascade Glacier (Meier and Tangborn, 1965). From the same data, Nye (1962) calculated a response time by yet another technique of correlating measured accumulation rates with observed changes in glacier thickness (a method possible only where continuous measurements are available). He determined a response time of 10 yr, which suggests that model predictions are correct at least to an order of magnitude.

In summary, all three approaches relate the response time to the glacier's size and its rate of mass transfer. The first two explicitly include dynamic considerations, although the glaciologic theory underpinning such formulation is still only approximate. The last technique estimates minimum response times and also permits simple comparison of such estimates among different glaciers.

During early Fraser time, typical ice-sheet thicknesses probably approached or exceeded 1,000 m (see below), and ablation rates at the terminus, by analogy with modern glaciers, ranged between 5 and 10 m/yr. The response-time formulation of Johannesson (1986) therefore yields minimum values of 100 to 200 yr. Alpine glaciers, by comparison, were as much as an order of magnitude less in thickness (e.g., Porter, 1976) but experienced nearly equivalent ablation rates at their termini during their probable advance into major trunk valleys and perhaps beyond the rangefront down to lowland altitudes. Their predicted response times were thus in the range of several tens of years.

Geologic evidence supports these predictions. Thorson's (1980) reconstructed profile of the Puget lobe closely approximates the "ideal" parabolic form of a glacier with uniform basal shear stress throughout. He argues by analogy that the Puget lobe was at or close to equilibrium at maximum. Because the southernmost 100 km of advance and retreat occurred in less than 1,500 years, the inferred equilibrium ice profile at maximum requires a substantially shorter response time.

The Evans Creek/Coquitlam advance of alpine glaciers at ca. 20 ka preceded the maximum stand of the main Cordilleran ice sheet by up to 5,000 yr, at least several times longer than the ice sheet's predicted response time. Thinner, shorter alpine glaciers fluctuated more rapidly in response to changing climate, and thus their 20-ka maximum probably marks the most extreme Fraser-age ELA lowering. These simplified calculations demonstrate, however, that the much later maximum advance of the Cordilleran ice sheet, culminating after nearly 10,000 yr of ice buildup in central and coastal British Columbia and flow into the southern lowlands (Clague, 1981), cannot simply be attributed to a sluggish response to the *same* climatic episode.

Environments of glacier termini. Differences in the physical environments of the alpine glaciers and the Cordilleran ice sheet may also have contributed to their asynchronous development. During Evans Creek/Coquitlam time, eustatic sea level was close to its late Pleistocene minimum, 120 m below present (see below), and the alpine glaciers emerging onto the Puget and Fraser Lowlands all terminated on land.

In contrast, many parts of the Georgia Depression are now 200 to 300 m below sea level, and thus any ice sheet attempting to advance directly into this tidewater environment would have experienced greatly increased ablation rates from calving. This part of the ice sheet probably could not have maintained a rate of advance equivalent to its nontidewater counterparts in the Cascade Range and Olympic Mountains. Advance rates would have been controlled less by net mass balance, which could have been driven quickly to zero or negative values by calving, than by the rate at which proglacial outwash infilled the water-filled depressions, thereby allowing the now-grounded ice front to advance. Stratigraphic evidence is consistent with this inference; Clague (1976, 1977) argues that the advance outwash of the ice sheet in British Columbia (Quadra Sand) probably filled the Georgia Depression an indeterminant distance in front of the advancing margin. Mullineaux and others (1965) arrived at the identical conclusion for the correlative deposit (the Esperance Sand Member of the Vashon Drift) in the Puget Lowland.

Advance of the eastern Cordilleran ice sheet. Advance of the lobes east of the Cascades was generally synchronous with that of the Puget and Juan de Fuca lobes to the west (Waitt and Thorson, 1983) and not with the Evans Creek/Coquitlam alpine-glacial advance. Because none of the eastern lobes reached tidewater, this delay with respect to the alpine glaciers must reflect differential response of the ice sheet as a whole to climate and not simply a difference in terminal environments. Either (1) the supply of moisture was limited so that significant growth was possible only by smaller alpine glaciers at 20 ka (Hicock and others, 1982), (2) the climatic change was not uniform over the region, being more pronounced in the south where alpine-glacial fluctuations were recorded, or (3) the change that drove the alpine-glacial advance at ca. 20 ka was sufficiently brief (i.e., centuries) that only these shorter, faster-reacting glaciers had time to respond fully.

Evidence for and against these options is limited and circumstantial. Porter (1977) argues for a change in Evans Creek precipitation of less than 30 percent relative to present, weakening support for the first alternative; global climate models similarly suggest little relative change in precipitation or adjacent sea-surface temperatures throughout this period. Existing data and climate-model reconstructions are too coarse in scale to support (or reject) the second alternative. The third option, although unsupported by the detail of present vegetation or climate models, would in fact be best recorded only by a system that could respond rapidly to such change. The record of alpine-ice advance and ELA depression during this time, augmented by analysis of likely glacier-response times, thus may offer the best available evidence for a substantial, but necessarily brief, period of regional climatic cooling at 20 ka.

The Vashon stade—Ice advance to maximum

Introduction. Whereas the Cordilleran ice sheet reached its maximum extent during the Vashon stade (ca. 15 ka), alpine glaciers experienced a readvance (Porter, 1976) that was both generally smaller than their own earlier Evans Creek/Coquitlam advance and dwarfed by the rate and magnitude of the advance of the main Cordilleran ice sheet. The relative sizes of Evans Creek/Coquitlam alpine glaciers, Vashon-age alpine glaciers, and the ice sheet at maximum stand reflect most closely the ELAs for these various glaciers. The ELA during Evans Creek/Coquitlam time is deduced from direct evidence of glaciated cirques (see above; Porter, 1977). The ELA during Vashon time requires more circuitous analysis of the ice-sheet equilibrium itself because, of all the glaciers in the Cordillera, only the ice sheet's position can be well established during this time.

Reconstruction of the southwestern Cordilleran ice sheet. Method. Not only the regional ELA but also many of the mechanics of ice-sheet behavior can be derived from the physical dimensions and mass balance of the ice sheet, including its rates of advance and retreat, magnitude of isostatic loading, and response time to climatic change. A reconstruction of these param-

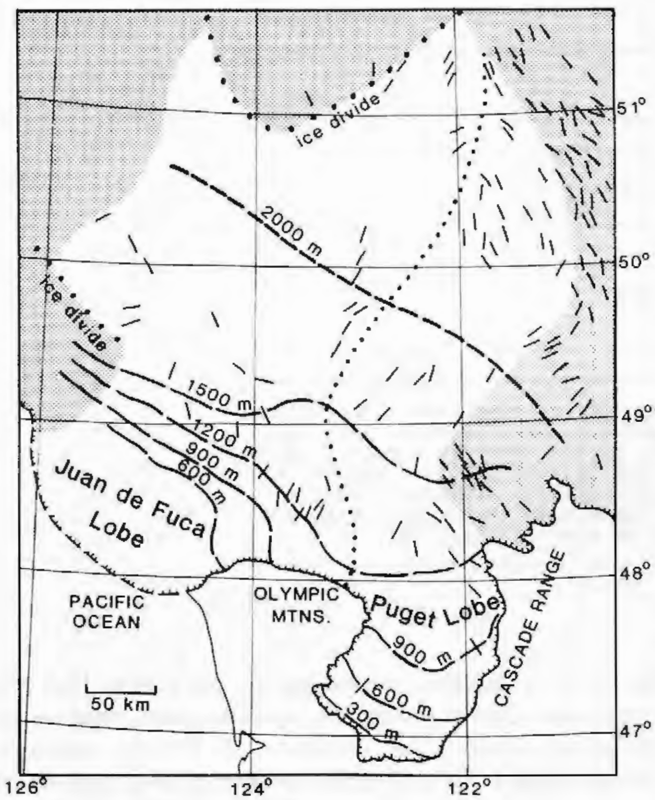

Figure 6. Reconstruction of the Puget and Juan de Fuca lobes of the Cordilleran ice sheet at maximum stage. Short lines show orientation of representative striations and other glacial lineations; hachured line shows maximum extent of ice. Heavy lines are contours of ice-surface altitude above modern sea level, uncorrected for glacial-age sea level or isostatic depression. Unshaded ice-covered area includes all ice of the Puget and Juan de Fuca lobes. Ice of these two lobes is inferred to have been separated along the dotted line. Sources of data include Wilson and others (1958), Prest and others (1968), Thorson (1980), Heller (1980), Dethier and others (1981), Fred Pessl, Jr. (personal communication, 1983), and Booth (1987a, 1987b, and unpublished data).

eters is possible only in a few special instances where sufficient data are available. The southwest part of the Cordilleran ice sheet is one such example.

Because the terminal configuration of the Puget lobe is well determined and a plausible mass-balance relationship is available, the ELA and basal sliding velocity of this lobe can be calculated (Booth, 1986b). For a glacier in equilibrium, net accumulation above any transect perpendicular to flow must be transferred by ice discharge through that transect to replenish the net ablation downglacier from it. This mass-balance method requires a reconstruction of the physical boundaries of the ice mass, topographic contours of its surface, and a relationship between specific net balance and altitude on the ice sheet. An equilibrium-line altitude (ELA) can then be found that brings the reconstructed glacier as a whole into balance. The rate of ice flux through any cross section is then easily calculated.

Reconstruction of the ice sheet. Ice limits for the southern boundary of the ice sheet on land have been rather accurately determined from extensive geologic mapping (Fig. 6; see Thorson, 1980, and Waitt and Thorson, 1983, for recent summaries). The Juan de Fuca lobe terminus was probably at tidewater and its locale subsequently submerged (Clague, 1981); it has been inferred (Alley and Chatwin, 1979) to coincide with the edge of the continental shelf southwest of Vancouver Island. Flow-direction indicators (striations, elongated topography) define a consistent pattern of ice flow and permit discrimination between ice in the Puget–Juan de Fuca portion of the Cordilleran ice sheet and that which lay to the northwest and east. Ice-surface contours shown on Figure 6 largely follow Thorson (1980) for the Puget lobe and Wilson and others (1958) over Canada. This present compilation attempts to reconcile these and other reconstructions (e.g., Waitt and Thorson, 1983), following several basic principles: the ice-surface contours should lie perpendicular to the direction of flow, flow lines should not converge or diverge without commensurate changes in ice thickness or net balance, and the local shear stress (proportional to the product of ice thickness and slope) should vary only gradually along the glacier's length.

Mass-balance calculations. The local mass balance of modern Pacific Northwest maritime glaciers is plotted in Figure 5 as a function of height above or below the equilibrium line (Meier and others, 1971; Porter and others, 1983). These data allow quantitative estimates of ice discharge through transverse cross sections of any glacier having an equivalent height–mass balance relationship.

If the ELA is known independently, the balance of any glacier can be calculated by defining areas on the ice surface with a representative altitude for each, converting each area into a rate of accumulation or ablation by using the height–mass balance curve and the given ELA, and integrating over the glacier surface to determine the predicted state of equilibrium or nonequilibrium for the chosen ELA value. The areas of each region shown in Figure 6 that are bounded by contour lines, ice limits, and lobe-dividing flow lines are assigned representative mid-point altitudes. In Table 1 the net balance is calculated for three alternative vlaues of the ELA.

Conversely, if equilibrium is assumed, the ELA can be determined by choosing an arbitrary initial ELA value and iterating by the same procedure. Equilibrium for the Puget–Juan de Fuca portion of the Cordilleran ice sheet, assumed to approximate conditions during ice maximum, is attained with an ELA of about 1,225 m. The balance contribution from calving along the tidewater terminus of the Juan de Fuca lobe is ignored in this analysis; based on measured calving rates from modern glaciers (Brown and others, 1982), this process could have consumed at most the excess yearly mass of a 25-m drop in the ELA.

Sliding velocity. On any glacier the ice accumulated above the equilibrium line must be transported into the ablation area. The flux will be greatest at the equilibrium line and decrease

TABLE 1. NET BALANCE VALUES IN 10^{10} m^3/a OF EQUIVALENT WATER VOLUME*

Altitude interval (m)	ELA = 1250 m J de F	ELA = 1250 m Puget	ELA = 1200 m J de F	ELA = 1200 m Puget	ELA = 1150 m J de F	ELA = 1150 m Puget
0-300	} -12.9	-1.2	} -12.3	-1.1	} -11.6	-1.0
300-600		-2.4		-2.2		-2.1
600-900	-2.0	-2.4	-1.8	-2.1	-1.6	-1.9
900-1200	-1.1	-1.5	-0.8	-1.1	-0.5	-0.8
1200-1500	0.7	0.8	0.8	0.9	0.9	1.1
1500-2000	5.4	2.0	5.6	2.1	5.9	2.1
>2000	9.5	4.3	9.8	4.4	10.1	4.5
NET BALANCE	-0.4	-0.4	+1.3	+0.9	+3.2	+1.9

*Positive values indicate net accumulation; negative values are net ablation. Calculations are based on areas between contours (Fig. 6) and specific net mass balance relative to ELA shown in Figure 5. Both lobes are brought into balance by an ELA between 1250 and 1200 m. An ELA value of 1225 m is therefore assumed for all subsequent calculations. Note these values differ from those used in the discussions of the ice volume by a density factor of approximately 10%.

through successive downglacier cross sections, because increasing volumes of glacier mass are lost through surface ablation. The inferred pattern of ice flow in the terminal part of the Juan de Fuca lobe is sufficiently complex to introduce large lateral variability in this flux. The simpler flow pattern of the Puget lobe, however, is well suited to flow calculations based on accumulation and ablation volumes. With an ELA of 1,225 m, 7.9×10^{10} m^3 of ice must have crossed the equilibrium line each year for the Puget lobe to remain in a balanced state (Table 2). The cross-sectional area at the equilibrium line is 1.2×10^8 m^2, giving an average ice speed of 660 m/yr. As only about 10 m/yr can be accounted for by internal deformation of the ice (based on the reconstructed ice thickness and surface slope; see Paterson, 1981, p. 87), the sliding velocity is approximately 650 m/yr, or 98 percent of the total velocity. Average rates of this order are estimated downglacier as well; for example, where the ice-surface altitude is 900 m (about the latitude of Seattle), the average ice speed is still over 500 m/yr.

Each element in this reconstruction is sufficiently constrained that the final conclusion, that is, basal sliding rates of several hundred meters per year, is scarcely affected by the uncertain precision of this analysis (see Booth, 1986b, for a more detailed discussion). The mass-balance relationship is least verifiable, because of the uncertainties involved in scaling up an empirical relationship derived from modern valley glaciers to a part of a subcontinental ice sheet. The most probable error is that high-altitude accumulation rates are overestimated, suggested by observational and theoretical evidence of an "elevation desert" high on an ice sheet (e.g., Budd and Smith, 1981; however, their data from a polar area may not be equally applicable to a maritime regime). To assess these uncertainties, the consequences of an imposed yearly accumulation limit of 2 m/yr can be evaluated. In response, the ELA of the reconstructed Puget lobe must drop by 100 m to remain in equilibrium. Yet this lowered ELA falls outside the range of geologically plausible values based on observed convergence and divergence of flow-line indicators (Booth, 1986b); this contradiction suggests a lower limit for the accumulation rates far upglacier. Using a variety of ELAs constrained by such geologic data, alternate ice velocities can be calculated and are shown in Table 3. Sliding velocity at the equilibrium line was consequently at least 500 m/yr and exceeded the internal deformation rate by nearly two orders of magnitude.

Implications for ice-sheet behavior. Predicted high mass flux, expressed as high sliding rates, is consistent with the rapid advance and brief occupancy of the Puget lobe in western Washington during the Vashon stade. Limiting radiocarbon dates in the Seattle area (Rigg and Gould, 1957; Mullineaux and others, 1965) require that the ice sheet advanced and then retreated more than 80 km during a period of 1500 ± 600 ^{14}C yr. If the advance rate is assumed to have been half that of the retreat (Weertman, 1964), the ice margin advanced at least 80 to 200 m/yr. Although these rates are high relative to many modern glaciers, they represent only a fraction of the equilibrium mass-transfer rate of the ice sheet. The Vashon stade is thus a modest perturbation in a much larger and longer depression of the ELA. For example, a rise of the ELA that recovers only 3 percent of the maximum Fraser-age ELA depression (900 m; Porter, 1977), from 1,225 m to 1,250 m, would generate a 5-km^3/yr deficit of the ice-maximum Puget lobe (Table 1), and, consequently, about 50 m/yr of retreat (using a typical cross-sectional area of $10^8 m^2$; see Table 2). The sensitivity of the lobe to small climatic changes demonstrates the plausibility of rapid ice-front movements without equally precipitous climatic changes.

Isostatic depression. As the Puget lobe advanced to its maximum position, it may have affected neighboring alpine gla-

TABLE 2. ICE DISCHARGE AND VELOCITY THROUGH TRANSVERSE SECTIONS OF THE PUGET LOBE*

Contour Interval (m)	Ice Width (10^5 m)	Average Thickness (10^3 m)	Cross-sectional Area (10^8 m^2)	Ice Discharge (10^{10} m^3/yr of ice)	Ice Velocity (m/yr)
2000	1.1	1.0	1.1	4.8	430
1500	1.1	1.1	1.2	7.0	580
1200	1.0	1.2	1.2	7.9	660
900	1.3	0.9	1.2	6.4	540
600	1.1	0.7	0.8	3.8	470
300	1.0	0.4	0.4	1.2	310
TERMINUS	1.0	0	0	0	0

*ELA = 1225 m. Sections are taken along contour lines shown on Figure 6. Ice velocity through these sections will be overwhelmingly by basal sliding (see text).

ciers by isostatically depressing the Cascade and Olympic mountains (Thorson, 1981). Any glacier on this depressed terrain would sink relative to its predepression ELA (Porter and others, 1983, p. 87). Depending on the areal extent of the depression, particularly the involvement of features affecting orographic snowfall, the ELA may have remained fixed relative to the glacier (i.e., no change in mass balance) or it may have remained fixed relative to a regional, undepressed datum (i.e., a decrease in net mass balance, reflected by an apparent rise in ELA on the glacier's surface). The actual effect would probably lie between these extremes. Thorson (1979, Fig. 31) estimates the rebound following deglaciation of the Cascade Range in areas not covered by the Puget lobe; it ranges from 0 to 80 m. He also argues (see below) for a maximum depression during full-glacial conditions that was at most twice this value. If this isostatic depression caused an effective rise in alpine-glacier ELA in the Cascade Range, the rise was negligible in the southern part of the Cascades to at most 160 m in the central Northern Cascades. ELA rise can be translated into changes in glacial length for an assumed accumulation-area ratio (0.6; Porter, 1975) and measured average valley slope. For example, in the South Fork Snoqualmie valley, a maximum ELA rise of 80 m and an average valley slope of 0.01 yields:

(80 m)/(0.01) = 8,000 m lateral (upvalley) ELA shift; thus
(8,000 m)/(0.6) = ~10 km upvalley retreat of terminus.

These values for ELA rise and terminus retreat account for only a fraction of the change in the likely position of the South Fork Snoqualmie glacier terminus between Evans Creek and Vashon times (Porter, 1976, Fig. 6), in spite of representing the *maximum* estimate of relative ELA rise due to isostatic depression for this area. Just north, in the Middle Fork Snoqualmie River valley, Williams' (1971) reconstruction of the Evans Creek and Vashon-age alpine glaciers is even less well explained by isostatic depression by the ice sheet. On the east slope of the Cascade Range, this effect would have been further diminished because of the increased distance from any major ice sheet, yet an equivalent substantial rise in inferred post–Evans Creek ELAs remains (Porter, 1976). Thus the influence of ice-sheet depression on the mass balance of contemporaneous alpine glaciers was probably small and primarily limited to the west slope of the north-central Cascade Range.

TABLE 3. ICE VELOCITIES AT THE EQUILIBRIUM LINE OF THE PUGET LOBE FOR VARIOUS ESTIMATES OF THE ELA

ELA	Average Ice Velocity (m/yr)	Ice Velocity Due To Internal Deformation (m/yr)	% of Total Flow Due to Basal Slip
1125	500	11	98
1200	610	11	98
1225	660	11	98
1500	900	13	99

Fraser-age ELA variation. Barring major impact of ice-sheet depression on the ELAs for alpine glaciers, the reduced extent of these mountain glaciers during the Vashon maximum can only reflect a climatically controled rise in the ELA of a few hundred meters or less relative to Evans Creek/Coquitlam time. This Vashon-age alpine-glacier ELA does in fact compare quite well with the regional value of 1,200 to 1,250 m calculated for the Puget lobe at maximum stage. On the basis of likely response times and the reconstructed ice-sheet profile, both alpine glaciers and the ice sheet approached or attained equilibrium form as the Vashon stade came to its climax. Their respective limits thus accurately reflected a regional, long-term climatic depression of the ELA relative to postglacial conditions. Over the entire Fraser glaciation, the magnitude of this depression was apparently exceeded only once in the Pacific Northwest, during the earlier and short-lived Evans Creek/Coquitlam interval.

The final retreat of the Cordilleran ice sheet

Isostatic and eustatic changes. The process and rate of the retreat of the southwestern Cordilleran ice sheet was strongly

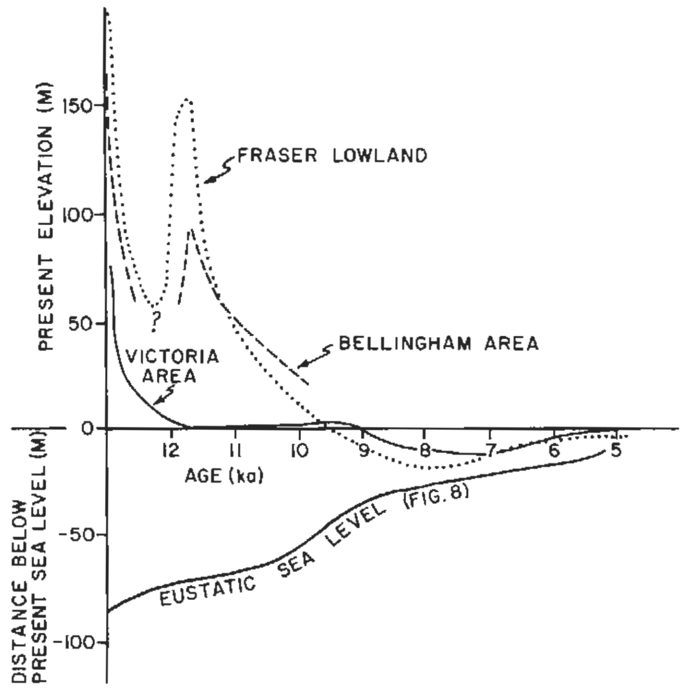

Figure 7. Proposed regional uplift curves for the Victoria, Fraser, and northern Puget Lowland areas (from Mathews and others, 1970, Fig. 4). The eustatic sea level curve in the lower section of the diagram is part of the data for the entire glacial period shown in Figure 8. The vertical distance between eustatic and shoreline curves at any given age indicates the amount of isostatic rebound that has subsequent occurred.

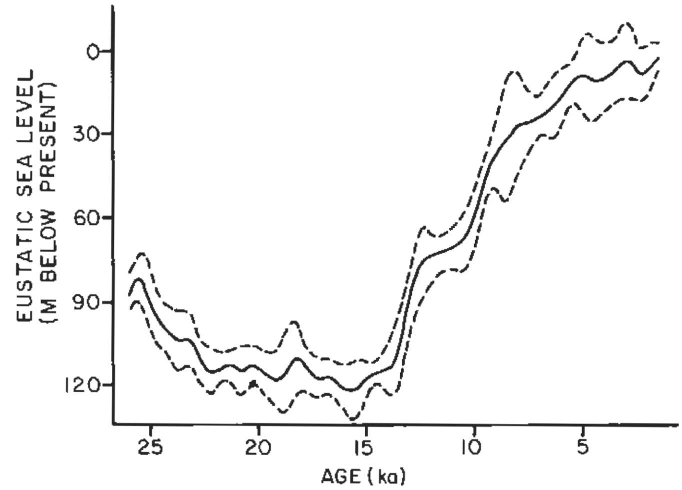

Figure 8. Proposed changes in sea level during the last glaciation, based on a maximum lowering of 120 m (modified from Mix, this volume). Dashed lines show error bounds on the best-fit curve.

influenced by concurrent changes in relative sea level. Dated marine deposits in the Puget and Fraser lowlands have yielded several regional uplift curves for the late-glacial and postglacial interval (Fig. 7). These curves reflect the combined influence of both eustatic rise (Fig. 8) and isostatic adjustment, plus an indeterminant component due to tectonic movement.

Shoreline data in the central and southern Puget Lowland are consistent with simple, monotonic uplift following deglaciation (Thorson, 1979, 1981). The rate of this uplift is unknown, and no known deposits lying between the maximum marine limit and present sea level suggest temporary cessation of relative uplift. Total rebound increases from south to north with a gradient of about 9 m/km and correlates extremely well with the reconstructed mass distribution of the Puget lobe at its maximum stage (Thorson, 1979, Fig. 28). South of the observed Everson-age marine limit, which is exposed above modern sea level only north of 47°40′N latitude (just north of Seattle; Thorson, 1979, Fig. 22), deltas associated with temporary lowland lakes that were dammed by the retreating ice extend this uplift gradient and allow its southward extension to a line of zero deformation near the ice-maximum limit (the data are insufficiently precise to identify any crustal bulge beyond the ice margin).

The uplift at Victoria also appears to have been monotonic, fitting into the pattern of the regional gradient with a maximum value of 75 m on shorelines cut immediately after deglaciation. Virtually all of the rebound here occurred in the period 13 to 11.5 ka (Mathews and others, 1970), thus averaging about 5 cm/yr. Because the rebound rate is generally presumed to be proportional to the magnitude of the isostatically uncompensated mass, initial rates were probably several times faster.

Data from the northern Puget and Fraser Lowlands suggest at least one interruption of such uplift. Citing dated shorelines and stratigraphic sequences near Bellingham, Washington, Easterbrook (1963) proposed a midrecessional resubmergence of 150 to 200 m prior to final emergence. Mathews and others (1970) proposed an equivalent pattern for the Fraser Lowland, with the maximum marine limit now at altitude 175 m and a 100-m transgression and regression at 12 ka. Although Thorson (1979) doubted whether the radiocarbon and stratigraphic evidence demanded such a complex history, subsequent scrutiny (Armstrong, 1981, p. 24–25; Clague, 1981, 1983) reaffirmed that at least one marine transgression, dated at 11.50 ±0.13 ka (BETA-1324) near Bellingham and indicating at least 35 m of resubmergence (Dethier and others, written communication, 1986), interrupted isostatic emergence of the Fraser and northern Puget Lowlands.

The pattern of maximum emergence in these northern lowlands continues the pattern inferred for the south. Emergence began as the ice thinned; the now-highest glacial-marine drift and shore deposits were formed immediately after deglaciation, at ca. 13.5 ka in the north. Sea-level altitudes increase to the north on a gradient of 1.25 to 1.50 m/km (Dethier and others, written communication, 1986) to a maximum reported value of 200 m just north of Vancouver (Clague, 1983, Fig. 3). Uplift was largely complete by 9 ka (Mathews and others, 1970), giving average rates of 4 cm/yr in the Fraser Lowland. This value, however, was temporarily exceeded in the northern lowlands because (1) initial rates probably decayed with time as the magnitude of the uncompensated mass decreased, and (2) the inferred transgression and regression required the net uplift to be accomplished in a shorter time. Initial rates also decrease from north to south; 30 cm/yr calculated on eastern Vancouver Island contrasts with 10 cm/yr in the northern Puget Lowland and with only 1 to 2 cm/yr in the central Puget Lowland (Dethier and others, written communication, 1986).

Relative sea-level changes were even more complex along the northern British Columbia coast. Here the mainland was uplifted following deglaciation, but the outer islands, beyond the limit of the main ice sheet, subsided during deglaciation until 10 to 9.5 ka (Clague, 1983, p. 335). The rate of subsidence was about 1.25 cm/yr during the period 12 to 8.5 ka and contrasted with a mainland emergence rate several times higher (Clague, 1983, Fig. 13).

Eustatic sea level. Estimates of maximum global ice volume during the late Pleistocene range from 75 m of sea-level lowering (Clark and others, 1978) through 110 to 115 m (Paterson, 1972), 120 m (Curray, 1965), and 130 m (Flint, 1971). Recent studies tend to favor eustatic sea level near –120 m, attained about 18 ka (Chappell, 1981; Duplessy and others, 1981).

Summaries in Ruddiman and Duplessy (1985) and Mix (this volume) suggest that global deglaciation did not occur uniformly. Although there is no consensus for the exact timing of eustatic sea-level change or of fluctuations in the rate of eustatic rise, a growing number of well-dated ^{18}O curves, analyzed in light of the problems and ambiguities raised by earlier studies, are in remarkably good agreement (Berger and others, 1985; Mix and Ruddiman, 1985). They indicate a global decrease in ice volume beginning about 14 ka, with a possibly more rapid 1,000 to 2,00-year interval of sea-level rise centered at about 13 ka (Fig. 8).

Isostatic rebound. *Isostatic response time.* From the available data in the Fraser Lowland, Thorson (1979, p. 130) calculated values of 800 to 1,600 years as the time necessary for the crust to compensate for one-half of the imposed ice load. Thus the crustal response was probably grossly in phase with the several-thousand-year interval of advance and retreat of the Cordilleran ice sheet. However, isostatic depression probably did not fully compensate for the imposed loads during the glacier's relatively brief maximum stand (Thorson, 1979, p. 139).

Victoria area. In the Victoria area, the shoreline data depict simple monotonic uplift during the period 13 to 11 ka. Global $\delta^{18}O$ data (Fig. 8) indicate that eustatic sea level may have risen up to 20 to 30 m during the interval 14 to 13 ka. The magnitude of the rebound following deglaciation equals the sum of the present marine limit (e.g., 75 m at Victoria) plus the magnitude of eustatic sea-level lowering at the time of deglaciation. This value probably ranged from about 90 to 110 m. Total rebound in the Victoria area, therefore, was probably 160 m or more, with an average uplift rate of 10 cm/yr (Mathews and others, 1970), several times greater than the predicted rate of sea-level rise.

Northern lowlands. In the northern Puget Lowland and in the Fraser Lowland, deglaciation has left a more complex record. Maximum marine-limit altitudes of 200 m imply total rebound in the range 270 to 310 m. Average rebound rates were several times greater than those of eustatic sea-level rise. Thus the only plausible mechanism for the inferred late-Everson transgression and regression is a temporary change in not only the rate but also the direction of crustal movement. Dethier and others (written communication, 1986) propose a migrating forebulge (Clark and others, 1978) to explain these data. Predictions of the magnitude of such a bulge, however, suggest that at maximum it would only have been about one-tenth that of the initial deformation (Walcott, 1972) and should further decay as it migrates with the retreating ice margin. Its effects, therefore, should have been even more pronounced farther south. In combination, these criteria do not fit well the requirements near Bellingham for at least 35 m of relative subsidence, although changes inferred from $\delta^{18}O$ data of Mix and Ruddiman (1985) for this period may account for nearly half of this change. The 100-m resubmergence postulated for the Fraser Lowland by Mathews and others (1970) appears to be well beyond the expected range of either isostatic or eustatic adjustment. If these data are correct, otherwise unrecognized tectonic movements must have occurred during this time. In contrast, the submergence noted by Clague (1983) along the northern British Columbia coast was sufficiently slow that eustatic rise alone, with or without the added effects of a colapsing forebulge, is sufficient to explain the data there.

In summary, deglaciation of the Puget and Fraser Lowlands occurred during a time of eustatic sea-level rise, at a rate of up to a few centimeters per year, at about 14 to 13 ka. Over the following several thousand years, isostatic rebound elevated the lowland areas at average rates of several centimeters per year, typically several times faster than concurrent rates of sea-level rise. Total isostatic rebound increased nearly linearly from zero at or near the southern ice margin to perhaps 300 m north of Vancouver, 250 km farther north. Resubmergence of at least 35 m probably interrupted the isostatic uplift of the northern Puget Lowland and the Fraser Lowland at about 11.5 ka. According to present geophysical theories, migration of a collapsing forebulge cannot fully account for such a drastic reversal in the postglacial isostatic rebound and suggests that tectonism may have been active as well.

Retreat in the southern and near-marginal areas of the Puget Lowland. As the margins of the Puget lobe first retreated from their Vashon-age maximum positions, a sequence of recessional meltwater deposits was laid down by streams issuing from the ice and in lakes impounded by the ice (Bretz, 1913). The pattern and preservation of these deposits owes much to the physiography of the southern Puget Lowlands. A drainage divide, nearly coincident with the Vashon-age ice limit, separates southward and (ultimately) westward flow into the Chehalis River and then into the Pacific Ocean from predominantly northward flow into Puget Sound. During the deglaciation, thick ice over the central and northern sound blocked northward drainage, impounding lakes whose waters spilled southward over the drainage divide into the Chehalis River. Because of the north-south grain of the topography in the lowlands, itself largely created by glacial erosion, several such postglacial lakes occupied adjacent valleys simultaneously. These lakes were interconnected by temporary spillways, with the lowest, Glacial Lake Russell, draining south into the Chehalis River for as long as the ice blocked drainage routes to the north (Fig. 9). Such lakes include Hood, Puyallup, Sammamish, Cedar, and Snoqualmie (Bretz, 1913; Thorson, 1980). Deposits associated with the progressive opening of channels by retreating ice can be identified and correlated with specific spillways from the southern edge of the ice-occupied lowland (Crandell, 1963; Lea, 1984) at least as far north as 48°10′N latitude in the eastern Puget Lowland (Booth, 1987a), 150 km north of the southern ice limit. Equivalent deposits in the east- and west-draining alpine river valleys from the Olympic Mountains and Cascade Range also are present this far north along the periphery of the Lowland (Thorson, 1980; Booth, 1986a, 1987b).

In total, the pattern of these deposits requires that the early retreating Puget lobe largely maintained an active, nonstagnant ice front. Discrete spillways separated by less than a few kilometers in the Snoqualmie and Skykomish river valleys (Booth, 1987b) demonstrate that stagnation zones were probably of negligible extent during this time, an inference supported by the general paucity of extensive upland dead-ice topography (Crandell, 1963; Thorson, 1980) in all but the most southerly plains of the Puget Lowland.

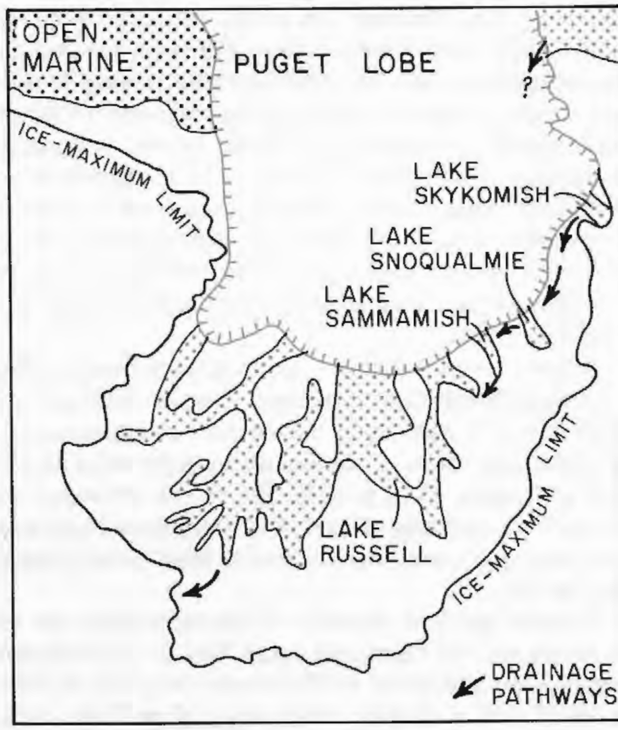

Figure 9. Glacial lakes during deglaciation of the Puget Lowland (after Thorson, 1980, Fig. 9C).

Retreat in the central and northern Puget Lowland. In contrast to the grounded, systematically retreating ice margin in the southern Puget Lowland, the Puget and Juan de Fuca lobes farther north calved extensively into tidewater (Armstrong and Brown, 1954; Easterbrook, 1963; Anderson, 1968). Even during the advance of the western Cordilleran ice lobes (17 to 15 ka), local and global conditions already were determining the character of their brief maximum stand and rapid subsequent retreat. Eustatic sea level may have already begun rising from its minimum glacial level (Ruddiman and Duplessy, 1985). Simultaneously, the earth's crust beneath and adjacent to the Puget Lowland and Strait of Juan de Fuca was sinking in response to the increased ice load. These factors, in combination, created a progressively less favorable environment for this part of the ice sheet independent of all climatic factors, by increasing water depths along the extensive seaward ice margin.

The magnitude and significance of these effects can be evaluated more precisely. Brown and others (1982) summarized calving-rate data from twelve modern Alaskan glaciers and derived the simple relationship:

$$v_c = 27.1\ h_w,$$

where v_c is the calving rate in m/yr and h_w is the water depth in meters. Eustatic sea level is estimated to have risen at most a few cm/yr around 15 to 14 ka; by analogy to later rates of isostatic rebound, the rate of crustal depression was probably several centimeters per year near the margin of the ice lobes. The net result would have been a deepening of tidewater conditions of up to 10 cm/yr. Deeper water would translate into an *acceleration* of the calving rate by 2 to 3 m/yr per year, assuming only that a minor fluctuation of the terminus occurred to remove the ice from any protective shoal that may have been deposited. Over the interval of several hundred to a thousand years that spanned the maximum stand of the ice (ca. 15 ka), the Juan de Fuca lobe would thus have increased its calving rate by perhaps 1,000 m/yr. An average terminus thickness of 150 m is suggested by the median value of observed Alaskan tidewater glaciers (Brown and others, 1982); the ice front was 150 km wide (Fig. 2). The product of these factors yields an increase in mass-loss rate during this interval, due solely to probable water-depth changes at the terminus, of approximately 10^{10} m^3/yr. This additional loss represents a significant fraction of the yearly mass budget of the Juan de Fuca lobe (Table 1) and is alone of sufficient magnitude to remove most of the approximately 10^{13} m^3 of this lobe occupying the Strait of Juan de Fuca and adjacent continental slope in less than 1,000 yr. The rate of retreat represented by these values is also the required order of magnitude for the grounding line of the Puget lobe to have retreated to its radiocarbon-dated (tidewater) positions during the Everson interval. This retreat rate is several times faster than retreat rates observed on modern nontidewater glaciers throughout the world (International Association of Hydrological Sciences, 1977). It is also over 10 times faster than the inferred retreat rate of eastern parts of the Cordilleran ice sheet (Fig. 4).

Rapid retreat of the western ice lobes was therefore a likely consequence of environmental changes at their termini, with or without concurrent, equivalent climatic amelioration. The speed of the Juan de Fuca lobe's retreat, in particular, is suggested geologically by the absence of recognized terminal or recessional features on the continental shelf or along the Strait of Juan de Fuca. This initial eastward retreat influenced the adjacent Puget lobe as well, because the opening of the Strait to ocean water led to accelerated decay of the western margin of the retreating Puget lobe (Thorson, 1980).

Climatic factors and the eastern Cordilleran ice sheet. Despite convincing evidence and consistent reconstruction of rapid ice-sheet retreat along the extensive tidewater margins of the western ice lobes, the inferred histories of both marine and nonmarine parts of the Cordilleran ice sheet are remarkably synchronous. The terminal zone of the Puget lobe was not exposed to tidewater; its early retreat history, probably initiated within a few hundred years of that of the Juan de Fuca lobe, does not reflect any marine influence. Within the broad constraints of available radiocarbon dates, the retreat of all lobes, marine and nonmarine alike, was approximately synchronous (Waitt and Thorson, 1983). Because terminal environments varied so widely among different parts of the ice sheet, this synchrony must reflect a regional climatic change at about 14.5 ± 0.5 ka. The *rates* at which individual lobes retreated, however, were closely tied to terminal conditions and correlated quite well with the presence or absence of a calving margin. Because the pollen record does not record an equivalent climatic change for 2,000 yr or more, either the lag in vegetation response was substantial or the ice sheet may have responded initially to a drier accumulation area rather than a warmer terminus.

SUMMARY

Three fundamental questions arise from the 20,000-yr history of the latest advance and retreat of the Cordilleran ice sheet. First, why was the alpine-glacial maximum at ca. 20 ka not matched by an equivalent maximum stand of the main ice sheet? Second, why had the alpine glaciers withdrawn from their farthest limits during the ice-sheet maximum 5,000 yr later? Finally, what environmental factors determined the timing and rate of final deglaciation?

Variation in the regional ELA, based on glacier reconstruction, and the physical behavior and environment of the ice-sheet terminus address these questions. During Evans Creek/Coquitlam time (20 ka), ELAs of individual alpine glaciers on the west slopes of the Cascade Range averaged close to 1,000 m. In contrast, the ELA of the southwestern part of the Cordilleran ice sheet at its maximum was probably near 1,200 m. Alpine glaciers responded much more rapidly to changes in climate than did the ice sheet, with characteristic times of about 10 to 100 yr. In contrast, the ice sheet required at least several hundred years to respond fully. The tidewater terminus of the western part of the ice sheet contrasted with land-based alpine glaciers and the largely land-based eastern ice-sheet lobes. Nevertheless, ice-sheet advance and retreat of all lobes coincided to within a millennium, although the rates of retreat were in some cases dramatically different.

The alpine-glacier maximum at ca. 20 ka reflects the lowest depression of the regional ELA during the last glaciation. It roughly coincided with a time of maximum glacial advance elsewhere in North America. The more limited extent of the western ice-sheet terminus at this time may express the slowing effect of a marine trough down which the glacier advanced. Yet the equivalent, limited extent of the eastern lobes virtually requires that this episode of lower ELA was simply too short for lobes of the ice sheet to respond fully. The predicted duration is therefore hundreds rather than thousands of years, consistent (though not required) by 2,000-yr limiting dates on the Coquitlam advance.

The ice-sheet maximum at 15 ka reflected a subsequent, long-term ELA depression over the region. The near-parabolic profile of the Puget lobe suggests that these conditions persisted for 1,000 yr or more; both the ice sheet and alpine glaciers attained equilibrium forms. Yet the reconstructed ELA value for the Puget lobe is over 100 m higher than during Evans Creek/Coquitlam time. Thus the relatively reduced extent of alpine glaciers simply reflects the shrinkage of glaciers under somewhat less favorable long-term climatic conditions.

The list of plausible determinants on ice-sheet retreat is severely limited by the relative timing of the individual lobes' wastage. In spite of a variety of terminal environments, including marine tidewater, lacustrine, and land-based on both sides of the Cascade Range, initial retreat began everywhere within 1,000 yr and progressed rapidly without substantial pause. Once initiated, rates of retreat correlated well with the terminal environments of individual lobes; the tidewater western ice sheet wasted back several times more rapidly than its eastern counterpart. However, only a regional climatic change adequately accounts for the fundamental relationship among the disparate ice-sheet lobes. The delayed expression of warming climate in the pollen record, up to 2,000 yr later, may indicate a lag in vegetation response but also suggests that the ice sheet may have responded initially not to any significant warming at its terminus but instead to reduced accumulation in its headward reaches.

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