Glacier advances at the Pleistocene/Holocene transition near Mount Rainier volcano, Cascade Range, USA.

by

Jan Tillmann Heine

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Chairperson of Supervisory Committee

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Abstract

Glacier advances at the Pleistocene/Holocene transition near Mount Rainier volcano, Cascade Range, USA.

by Jan Tillmann Heine

Chairperson of the Supervisory Committee: Professor Stephen C. Porter
Department of Geological Sciences

Glaciers in the vicinity of Mount Rainier advanced twice during late glacial/early Holocene time. Radiocarbon dates obtained from lake sediments adjacent to the corresponding moraines are concordant, indicating that the ages for the advances are closely limiting. The first advance occurred before 11,300 ^{14}C yr BP (13,200 cal yr BP). During the North Atlantic Younger Dryas event, between 11,000 and 10,000 ^{14}C yr BP (12,900 and 11,600 cal yr BP), glaciers retreated on Mount Rainier, probably due to a lack of available moisture, but conditions may have remained cold. The onset of warmer conditions on Mount Rainier occurred around 10,000 ^{14}C yr BP (11,600 cal yr BP). Organic sedimentation lasted for at least 700 years before glaciers readvanced between 9800 and 8950 ^{14}C yr BP (10,900 and 9950 cal yr BP). Glaciers seem to have advanced in response to regional or local shifts in late-glacial climate. No evidence for a glacier advance during Younger Dryas time has been found. However, the Younger Dryas climatic reversal may have affected the Mount Rainier area, causing a cold, but dry, climate unfavorable to glacier advances.

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DEDICATION

I dedicate this dissertation to Barbara A. Van de Fen.

INTRODUCTION

The transition from the last glaciation to the current interglaciation (ca. 15,400 to 11,600 cal yr BP; 13,000 to 10,000 ^{14}C yr BP) has attracted considerable study in different areas of the world, as it provides the most recent example of global, rapid climatic change. Recently, high-resolution proxy records indicate that the Earth's climate systems may be unstable and fluctuate within decades or less (chapter 2). However, the mechanisms and causes of these rapid late-glacial climatic oscillations are not completely understood. To improve our understanding of the underlying mechanisms and causes, it is important to determine which climatic oscillations occurred on a local, regional, or global scales. This can be achieved by establishing and comparing well-dated local and regional climatic chronologies in different areas of the world.

While late-glacial climatic events have been resolved in detail in northwestern Europe and Greenland (chapter 2), the evidence from other areas remains less clear (e.g., reviews in Markgraf et al., 1992; Heine, 1993; Peteet, 1995). A study of glacier advances on and near Mount Rainier at the Pleistocene/Holocene transition permits reconstruction of aspects of the local climate at that time. Comparison of the climatic events in this area with other areas may provide new insights into the dynamics of climate.

1 REPORTING OF AGES AND TIMING OF PALEOCLIMATIC EVENTS

Where possible, all radiocarbon ages in the following chapters were calibrated using the CALIB program (Stuiver and Reimer, 1993), and are reported as 1σ confidence intervals (Tab. 1). This reduces the errors associated with so-called radiocarbon plateaus, times when changes in atmospheric ^{14}C concentration affected the apparent ages of contemporaneous organic matter. The corresponding ages in ^{14}C yr BP are given in parentheses. Since the CALIB program does not reach beyond 19,262 ^{14}C yr BP, older ages are reported as uncalibrated ages with their 1σ interval. In some cases, approximate ages older than 19,262 ^{14}C yr BP were calibrated after Bard et al. (1993) to allow comparisons with other timescales. However, these calibrations may be somewhat inaccurate.

Newer results point to small errors with the dataset that forms the basis of the CALIB program (Björck et al., 1996) around 12,000 cal yr BP. However, the changes are small, ca. 161 ± 80 years (Björck et al., 1996). To ensure that the results of this study can be replicated, the calibration program of Stuiver and Reimer (1993) is used in this study without modification.

Only calibrated ages allow an assessment of the duration of events. For example, calibration of the limiting ages for the Younger Dryas event (12,900 to 11,600 cal yr BP) results in a duration of 1300 cal years, which is in good agreement (within several percent) with the duration of this interval derived by counting annual layers in Greenland ice cores (Alley et al., 1993). By comparison, the corresponding ^{14}C ages (11,000 and 10,000 ^{14}C yr BP) misrepresent the length of the interval as 1000 radiocarbon years. Furthermore, reporting ages in radiocarbon years often leads to an underestimate of the standard deviation (e.g., a ^{14}C date of 10,150 ± 60 yr BP calibrates to 12,040-11,345 cal yr BP, with a 68% confidence interval of 695 years).

Table 1: Radiocarbon dates in this study

Sample	Significance	^{14}C age ±1s (yr BP)	Sample Material	Calibrated Age (1s)	Dating Method	Laboratory Number
Lowest org., exp. 4 Moraine Park	Min. date for McNeeley	6790 ± 70	Bulk peat	7640-7540	b (conv.)	QL-4774
Twig below R White R. P. (7)	Rejected (see text)	7450 ± 60	Twig (macrofoss.)	8320-8130	AMS	CAMS23064
Peat below R, White R. P. (4)	Max. date for tephra R	8760 ± 80	Bulk peat	9880-9570	b (conv.)	QL-4770
Twig in R, White R. P. (3)	Min. date for McNeeley 2	8920 ± 60	Tree branch (macrofoss.)	9980-9880	b (conv.)	QL-4765
Lowest org., above U. Dew. L. (8)	Min. date for McNeeley 2	9140 ± 100	Bulk sediments	10,280-9990	b (conv.)	QL-4769
Peat above out-wash W. R. P. (5)	Min. date for McNeeley 2	8990 ± 40	Bulk sediments	9990-9940	b (conv.)	QL-4805
Peat above out-wash, Tipsoo L (7)	Min. age for McNeeley 2	9120 ± 80	Bulk sediments	10,280-9980	b (conv.)	QL-4820
Outwash layer, White R. P. (5)	Actual date McNeeley 2	8990 ± 60	Twig (macrofoss.)	10,000-9920	AMS	CAMS23065
Peat below out-wash W.R.P. (5)	Max. date for McNeeley 2	9580 ± 50	Bulk sediments	10,900-10,480	b (conv.)	QL-4804
Peat below out-wash Tipsoo L. (7)	Max. date for McNeeley 2	10,080 ± 60	Bulk peat	11,880-11,120	b (conv.)	QL-4821
Lowest organic, Up. Dewey L. (9)	Onset of organic sed.	9460 ± 50	Bulk sediments	10,780-10,370	AMS	CAMS23066
Lowest organic, Up. Dewey L. (10)	Onset of organic sed.	9500 ± 70	Bulk peat	10,850-10,380	b (conv.)	QL-4771
Lowest organic, White R. P. (4)	Onset of organic sed.	9840 ± 70	Bulk sediments	11,010-10,990	b (conv.)	QL-4764
Lowest organic, White R. Park (5)	Onset of organic sed.	10,150 ± 60	Bulk gyttja	12,040-11,340	b (conv.)	QL-4831
Basal, White River Park (5)	Min. date for McNeeley 1	11,090 ± 120	Bulk sediments	13,130-12,880	b (conv.)	QL-4803
Basal, Upper Hen Skin Lake (6)	Min. date for McNeeley 1	11,120 ± 200	Bulk sediments	13,240-12,840	b (conv.)	QL-4774
Basal, Upper Dewey L. (9)	Min. date for McNeeley 1	11,320 ± 60	Bulk sediments	13,320-13,150	AMS	CAMS23067
Org. above tephra Lk. Mary Lea (19)	Min. date for max. stage 2	24,120 ± 340	Bulk sediments	N/A	b (conv.)	QL-4885

2 DEGLACIAL CLIMATIC OSCILLATIONS IN AND BEYOND NORTHWESTERN EUROPE

The deglacial climatic oscillations in northwestern Europe have been studied for more than 80 years (e.g., Ahlmann, 1910; Hoppe, 1948, Andersen, 1954; Huntley, 1988). The standard late-glacial stratigraphy was developed there and has been used since as a global reference.

2.1 FULL-GLACIAL TO LATE-GLACIAL ICE RECESSION IN EUROPE

The Fennoscandian Ice Sheet reached its maximum extent during oxygen isotope stage 2 between 20,000 and 18,000 ^{14}C yr BP in northern Germany and surrounding areas (Sollid and Reite, 1983; Denton and Hughes, 1981). The retreat from this position was interrupted by readvances or minor stillstands. A significant readvance occurred during the Younger Dryas interval (12,900 to 11,600 cal yr BP, 11,000 to 10,000 ^{14}C yr BP), when large parts of Scandinavia were covered by glacier ice (Fig. 1). Since the ice readvanced both in the continental climates of Finland/Russia and in the maritime climate of western Norway, Mangerud (1987) attributed this readvance to a drop in temperature. However, the magnitude of this readvance was much greater in western Norway than in eastern Scandinavia (Mangerud, 1987). During the Younger Dryas event, some cirques in western Norway, which had been deglaciated, were reoccupied by glaciers as a result of a lowering of the equilibrium line altitude (ELA) ca. 450 ± 50 m (Mangerud, 1987; Fareth, 1987). The inferred drop of summer temperatures was 5-6°C (Mangerud, 1987). These glacier advances were accompanied by intensified periglacial activity throughout northwestern Europe (Maarleveld, 1976; Larsen et al., 1984). After the Younger Dryas readvance, glaciers retreated rapidly, especially in the Norwegian fjords. In western Norway, the Jostedalsbre and some other glaciers advanced again, about 10,300-9910 cal yr BP (9100 ± 200 ^{14}C yr BP) during the Erdalen event of Preboreal time (Nesje et al., 1991). The associated ELA depression then was 325 +75/-115 m (Nesje et al., 1991, Nesje and Dahl, 1992). By comparison, the ELA depression during the Little Ice Age in western Norway was ca. 150 m (Nesje, 1989; Nesje et al., 1991).

2.2 PALEOVEGETATION STUDIES

Numerous pollen studies in northwestern Europe have attempted to reconstruct the late-glacial vegetation and climate. Rather few records are available from full-glacial time (23,000-15,400 cal yr BP; 20,000-13,000 ^{14}C yr BP), as much of northern Europe was covered by ice. The areas close to the glacier limit may not have been productive enough to allow accumulation of biogenic sediments suitable for pollen analysis (Huntley, 1988). More pollen records are available from late-glacial time (15,400-11,600 cal yr BP, 13,000-10,000 ^{14}C yr BP). During the Allerød chronozone (13,700-12,900 cal yr BP, 11,800-11,000 ^{14}C yr BP), tree birch became established over much of northern Europe (Huntley, 1988) and climatic conditions were comparable to those of the present (Mangerud, 1987; Huntley, 1988). During the following Younger Dryas interval (12,900-11,600 cal yr BP, 11,000-10,000 ^{14}C yr BP), named after the flowering plant *Dryas octopetala* which increased in abundance in deposits of Younger Dryas age, a rapid change to colder conditions occurred. Tree birch apparently disappeared, while non-arboreal taxa increased substantially in abundance, especially *Artemisia* and Cyperaceae, interpreted as evidence for a relatively open vegetation growing under relatively dry conditions (Huntley, 1988). After 11,600 cal yr BP (10,000 ^{14}C yr BP), forest spread rapidly across northern Europe. Over a time-span of about 1000 years, vegetation changed from an open, herb- and dwarf-shrub-dominated vegetation during Younger Dryas time to a dwarf-shrub, shrub, and woodland vegetation characterized by *Juniperus* and *Betula* (Huntley, 1988).

2.3 GREENLAND ICE CORES AND NORTH ATLANTIC OCEAN CIRCULATION

In Greenland, several long ice cores have yielded high-resolution paleoclimatic data, most notably the GISP II, GRIP, and Dye 3 cores. In the GISP II core, an abrupt end of the Younger Dryas is seen in the time-series of oxygen isotope ratios, which suggest a warming to Holocene conditions within a span of 50 years (Alley et al., 1993). Even more rapid were changes in dust concentration, with a transition lasting no more than 20 years (Alley et al., 1993). Snow accumulation on the central Greenland ice sheet was low during cold periods and higher during warm intervals (Alley et al., 1993). Two significant late-glacial cold events are recorded in the GISP II snow accumulation record: the Oldest Dryas interval ended 14,700 cal yr BP and the Younger Dryas interval occurred between 12,900 and 11,600 cal yr BP (Alley et al., 1993; Stuiver et al., 1995). Both during the Oldest

Dryas and the Younger Dryas intervals, the records show a gradual decrease in snow accumulation, indicating a gradual transition to colder climate during these events. The transition from the colder to the warmer climate, both at the end of the Oldest Dryas and at the end of the Younger Dryas, occurred much more rapidly, and in the case of the Younger Dryas possibly within 3 years. The magnitude of the Younger Dryas cooling in Greenland is inferred to have been ca. 7°C (Alley et al., 1993). A smaller climatic reversal occurred during the Older Dryas event around 14,100 cal yr BP (Stuiver et al., 1995).

The polar front is the boundary between polar/subpolar and temperate water masses. The North Atlantic Polar Front began to retreat northward from its full-glacial position in the eastern North Atlantic ca. 13,000 ^{14}C yr BP (Fig. 2). It readvanced southward ca. 11,000 ^{14}C yr BP at the onset of the Younger Dryas event (Ruddiman and McIntyre, 1981; Lehman and Keigwin, 1992). The associated reduction in warm-water circulation and sea-surface temperatures (SSTs) during the Younger Dryas event lasted until 12,400 cal yr BP (10,500 ^{14}C yr BP) (Lehman and Keigwin, 1992). Another reduction in circulation and SSTs occurred between ca. 10,900 and 9800 cal yr BP (9700 and 8800 ^{14}C yr BP) during the early Holocene (Lehman and Keigwin, 1992).

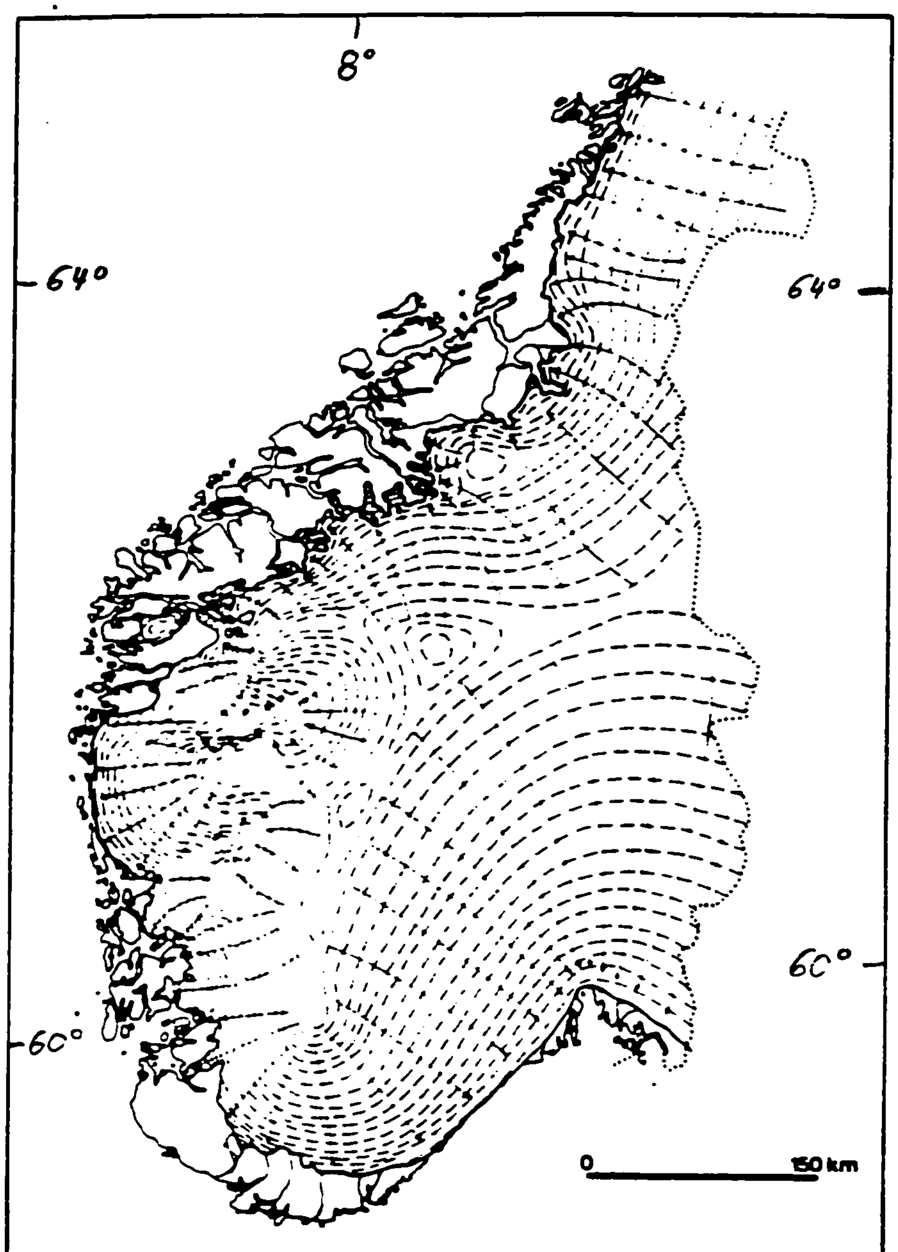


Figure 1: Overview map showing extent of Fennoscandian Ice Sheet in Norway during Younger Dryas time (from Sollid and Reite, 1983). The dashed lines probably are contour lines, however, Sollid and Reite (1983) give no information on contour intervals and elevations.

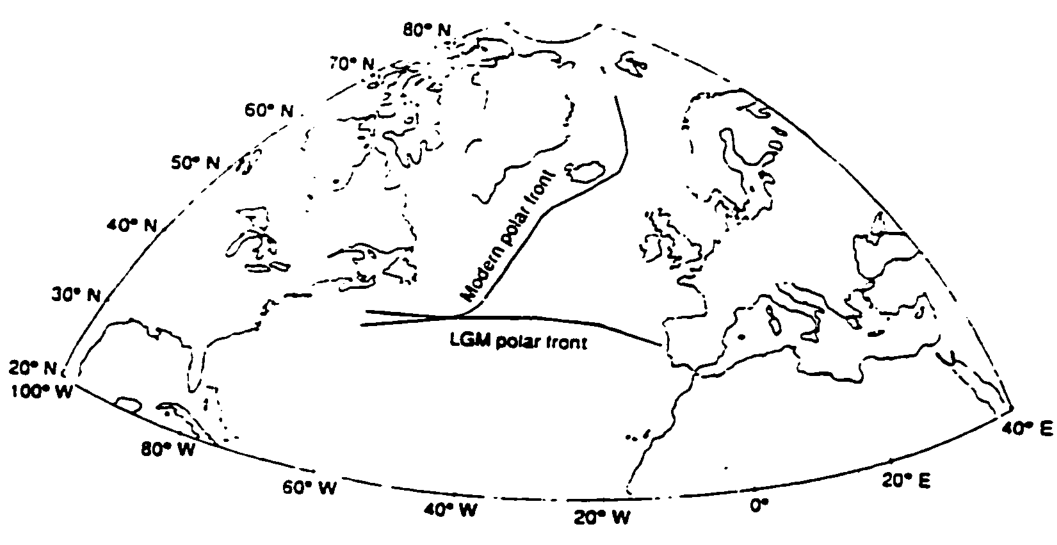


Figure 2: Map of the North Atlantic region showing the positions of the modern and reconstructed last glacial maximum (LGM) polar fronts (from Lehman and Keigwin, 1992).

3 LATE-GLACIAL ALPINE GLACIER ADVANCES IN WESTERN NORTH AMERICA

Several glacier advances in the North American cordillera may be correlative with the North Atlantic Younger Dryas climatic reversal.

3.1 CANADIAN ROCKY MOUNTAINS

Reasoner et al. (1994) dated lake sediments downstream from moraines correlated with the Crowfoot advance in the Canadian Rocky Mountains to between 13,700 and 10,970 cal yr BP (11,330 and 10,020 ^{14}C yr BP) and suggested that the Crowfoot advance was coeval with the North Atlantic Younger Dryas event. The mapped extent of the glaciers during the Crowfoot advance was only slightly greater than during the Little Ice Age, indicating that ELA depression associated with this glacier advance was minor compared to that of the European Younger Dryas event. Drift of the Crowfoot advance has been dated only indirectly and at only one site (Reasoner et al., 1994).

3.2 WIND RIVER RANGE, WYOMING

Deglaciation following the construction of the type Temple Lake moraine in the Wind River Range is radiocarbon-dated to before 14,700-12,920 cal yr BP (11,770 ± 770 ^{14}C yr BP) (Zielinski and Davis, 1987). The 2σ interval (95% confidence) for this date places the minimum date for deglaciation from the Temple Lake moraines between 15,912 and 12,016 cal yr BP, thus possibly overlapping the European Younger Dryas. Based on ^{10}Be surface exposure dating, moraines equivalent to the Temple Lake moraines have been suggested as a Younger Dryas equivalent (Gosse et al., 1995). However, the production parameters introduce a possible error of at least 10% to the surface exposure dates (Clark et al., 1995), in this case ca. 1200 years, and the standard deviation of the conventional ^{14}C ages is extremely large. Thus, the dates permit, but do not require, a correlation of the Temple Lake advance with the North Atlantic Younger Dryas event. At this time, the age of the Temple Lake advance remains poorly constrained.

3.3 NORTH CASCADE RANGE, WASHINGTON

A large system of late-glacial alpine glaciers in the North Cascades extending almost to sea level has been proposed by Kovanen and Easterbrook (1996). Field evidence for this advance remains controversial. During a Pacific Coast Friends of the Pleistocene field trip in 1996, it was suggested that the inferred terminal moraines may be ice-marginal features deposited by the Vashon advance of the Puget Lobe of the Cordilleran Ice Sheet (Fig. 3). The glacial landforms deposited by this inferred advance have not been dated directly.

3.4 SUMMARY OF EVIDENCE FOR YOUNGER DRYAS-EQUIVALENT GLACIER ADVANCES IN WESTERN NORTH AMERICA

Although the glacier advances discussed above appear to be late-glacial in age, their contemporaneity with the European Younger Dryas event has not been demonstrated. In contrast to the above results, several studies show evidence that glaciers advanced and retreated prior to the European Younger Dryas:

3.5 MISSION MOUNTAINS, MONTANA

The Piper Lake moraine in the Mission Mountains of Montana is overlain by Glacier Peak tephra G. Therefore, deglaciation from the associated advance occurred before 13,100 cal yr BP (11,200 ^{14}C yr BP) (Osborn and Gerloff, 1997). No moraines have been identified between the Piper Lake moraine and those deposited during the Little Ice Age.

3.6 SIERRA NEVADA, CALIFORNIA

After retreat from the most extensive stand of the Tioga glaciation (the local last glacial maximum), glaciers in the Sierra Nevada advanced before 13,190-13,010 cal yr BP (11,190 ± 70 ^{14}C yr BP) during the Recess Peak advance (Clark, 1995; Clark and Gillespie, 1997) to deposit a single moraine. Moraines upvalley from the Recess Peak moraines date to the Matthes advance, which occurred during the Little Ice Age. The apparent absence of moraines between those of Recess Peak and Matthes age implies that glaciers did not experience a significant period of expansion (beyond the Matthes ice limit) between ca. 13,000 and 700 cal yr BP (Clark, 1995; Clark and Gillespie, 1997).

3.7 CASCADE RANGE, WASHINGTON

Several late-glacial ice advances have been identified in the Cascade Range. In most drainage areas, a paired set of late-glacial moraines occurs. The moraines carry local names, such as Rat Creek, Hyak, and McNeeley. At Stevens Pass, a tephra that was deposited ca. 12,150 cal yr BP (11,250 ^{14}C yr BP) has been found beyond moraines that may be equivalent to the Rat Creek advance, but not on the moraines (Porter, 1978; Porter et al., 1983), implying that the moraines may postdate this tephra fall. The Hyak moraines at Snoqualmie Pass may be the equivalent of the Rat Creek moraines further northeast, based on their similar positions in the upper reaches of Cascade valleys, similar morphology and weathering characteristics (Porter, 1978). The age for the earlier of these two advances is limited by a minimum date of 13,030-12,900 cal yr BP (11,050 ± 50 ^{14}C yr BP) on wood above till behind the outer of two Hyak moraines at Snoqualmie Pass (Porter, 1976).

3.8 SUMMARY OF LATE-GLACIAL GLACIER ADVANCES IN WESTERN NORTH AMERICA

The timing and extent of glacier advances of western North America during late-glacial time remain unresolved. It remains possible that glaciers advanced during Younger Dryas time across the region, with the exception of the Sierra Nevada in California. However, it also is possible that late-glacial climate in the region was complex and that glaciers in different locations advanced and retreated at different times.

3.9 PRESENT RESEARCH

Western North America is an important region for understanding the dynamics of climate change. With the prevailing westerly winds and the prevailing ocean currents, western North America is climatically associated with the North Pacific, and climatically remote from the North Atlantic. A well-constrained late-glacial climatic history from the western

United States could not only help determine the timing and extent of late-glacial glacier advances, but also permit comparison with late-glacial events documented elsewhere, and thus provide new insights into the geographic distribution of late-glacial climatic events.

In this study, evidence is presented for glacier advances in the Cascade Range close to the Pleistocene/Holocene transition, both northeast and east of Mount Rainier and the Goat Rocks. From the glacial evidence and from sediment records in lakes, the late-glacial climate was reconstructed. Comparing the climatic events in the Cascades with the European standard will help assessing which late-glacial climatic events occurred on a global scale and which were regional in extent.

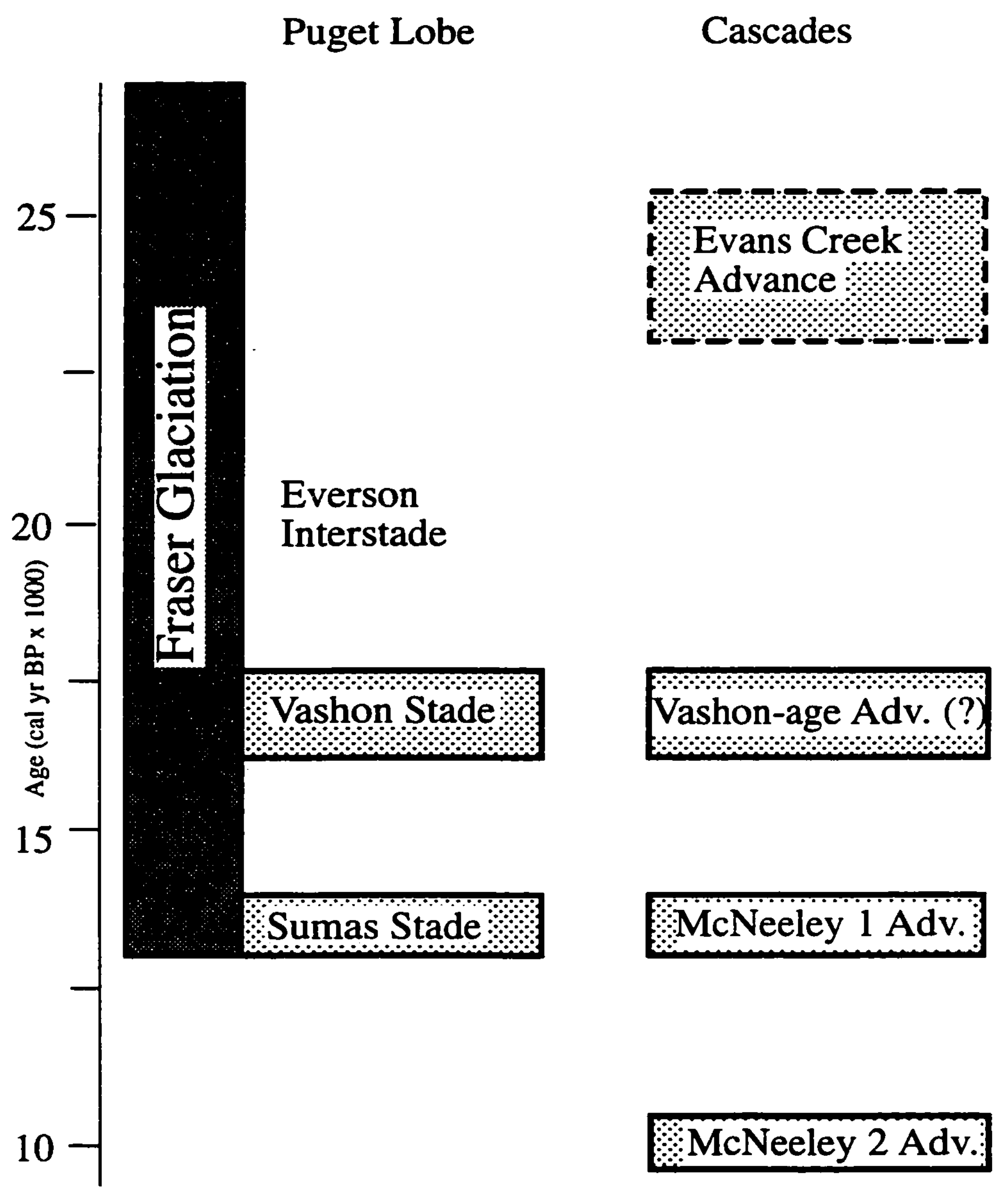


Figure 3: Glacial stratigraphy of the Puget Lowland and the Cascade Range during the last glaciation. Dashed lines indicate that the timing of an event is not well-known. Some of the above correlations are tentative.

4 LAST GLACIATION IN WESTERN WASHINGTON

4.1 VASHON ADVANCE OF THE CORDILLERAN ICE SHEET

During the Vashon Stade of the Fraser Glaciation (Fig. 3), the Cordilleran Ice Sheet advanced into the northern Puget Lowland and across Vancouver Island about 21,500 cal yr BP (18,000 ^{14}C yr BP) (Fig. 4) (Waitt and Thorson, 1983). The lowered snowlines during the Vashon Stade may have been associated with increased precipitation (Matthewes, 1991) that accompanied increased storminess along a southward-displaced polar jet stream (Barnosky et al., 1987). About 20,000 cal yr BP (17,000 ^{14}C yr BP), the ice advanced into the Strait of Juan de Fuca (Waitt and Thorson, 1983). After the Juan the Fuca Lobe reached its maximum, the Puget Lobe continued to advance and reached the Seattle area about 17,900 cal yr BP (15,000 ^{14}C yr BP) (Waitt and Thorson, 1983; Booth, 1987). The ice front rapidly advanced to its limit near Olympia between 17,380 and 16,780 cal yr BP (14,500 and 14,000 ^{14}C yr BP), followed by a rapid retreat (Waitt and Thorson, 1983; Booth, 1987).

The ice sheet readvanced during the Sumas Stade between 13,400 and 13,100 cal yr BP (11,500 and 11,200 ^{14}C yr BP) (Saunders et al., 1987), based on stratigraphic evidence and dated logs in till and outwash. Fjords bordering the Strait of Georgia were ice-free by at least 12,900 cal yr BP (11,000 ^{14}C yr BP) (Ryder and Clague, 1989). Waitt and Thorson suggested that the readvance may have been caused by a grounding of the calving ice front due to isostatic uplift rather than by a shift in regional climate (1983).

4.2 CASCADE RANGE

The maximum advance of mountain glaciers in the Cascades occurred before ca. 21,500 cal yr BP (18,000 ^{14}C yr BP) (Waitt and Thorson, 1983). No close direct radiocarbon ages on this advance exist from western Washington (Porter et al., 1983). Two moraine groups dating from the last glaciation have been identified in the Yakima valley. Recent ^{36}Cl-ages date the older to ca. 20,000-24,000 cal yr BP, and the younger to the Vashon Stade (Swanson and Porter, 1997). During the advance of the Puget Lobe, the Cordilleran ice sheet buried much of the northeastern part of the Cascade Range (Waitt and Thorson, 1983). In the Puget lowland north of the Nisqually River, no alpine end moraines of pre-Vashon age have been identified, presumably because they were overrun by the Puget Lobe (Porter et al., 1983).

The late-glacial ice advances of alpine glaciers in the Cascade Range have been discussed in chapter 3.

The maximum alpine advance of the last glaciation is termed Evans Creek on Mount Rainier (Crandell and Miller, 1974). Deposits of a more-extensive, earlier, undated advance were named Hayden Creek (Crandell and Miller, 1974). Moraines of the McNeeley advance on Mount Rainier are inferred to be late-glacial in age based on their relative position, morphology, and relationship to tephra layer R (see chapter 3) (Crandell, 1969; Crandell and Miller, 1974).

4.3 POLLEN RECORDS

Pollen evidence suggests that the climate in southern British Columbia was considerably colder and drier than today between 28,500 and 18,900 cal yr BP (25,000 and 16,000 ^{14}C yr BP) (Hicock et al., 1982; Thompson et al., 1993), when subalpine parkland and possibly alpine plant communities covered the lowlands of eastern Vancouver Island (Clague and MacDonald, 1989); the Puget lowland was occupied by a subalpine parkland with Engelmann spruce, Lodgepole pine, *Artemisia*, and grasses (Thompson et al., 1993). Temperatures and precipitation increased in coastal British Columbia after 18,900 cal yr BP (16,000 ^{14}C yr BP) (Matthewes, 1991), when non-arboreal herb-tundra covered the lowlands of the Queen Charlotte Islands (Warner et al., 1982). At 16,800 cal yr BP (14,000 ^{14}C yr BP), heaths, grass-sedge-herb meadows. In western Washington, a significant warming has been documented between ca. 14,000 and 11,600 cal yr BP (12,000 and 10,000 ^{14}C yr BP), when conifers spread rapidly (Sugita and Tsukada, 1982) and mixed communities of subalpine and lowland taxa occupied western Washington (Thompson et al., 1993).

Evidence for a widespread late-glacial cooling event in western North America that may correlate with the North Atlantic Younger Dryas event remains ambiguous (Mann and

Hamilton, 1995). Some studies indicate a cool climate during Younger Dryas time in British Columbia, where a shift from forest to open, herb-rich vegetation occurred (Matthewes et al., 1993), and in Alaska, where a pine parkland was replaced by a shrub- and herb-dominated tundra (Engstrom et al., 1990). Other vegetation studies from Oregon and Washington are interpreted as showing a uniform warming trend toward the Holocene (Tsukada et al., 1981; Worona and Whitlock, 1996; Grigg, 1996). It is possible that during Younger Dryas time, a slight drying occurred near Mount Rainier (Tsukada et al., 1981). A similar drying, probably associated with a cooling has been inferred for the California Sierra Nevada (Hemphill and Clark, 1996). Throughout western North America, climate became warmer about 11,600 cal yr BP (10,000 ^{14}C yr BP) (Mann and Hamilton, 1995). About 9980 cal yr BP (9000 ^{14}C yr BP), climate in the Pacific Northwest was drier than it has been since, with a pronounced summer drought, and also increased summer temperatures, documented by an expansion of Douglas fir and alder in open-forest habitats (Thompson et al., 1993; Worona and Whitlock, 1996).

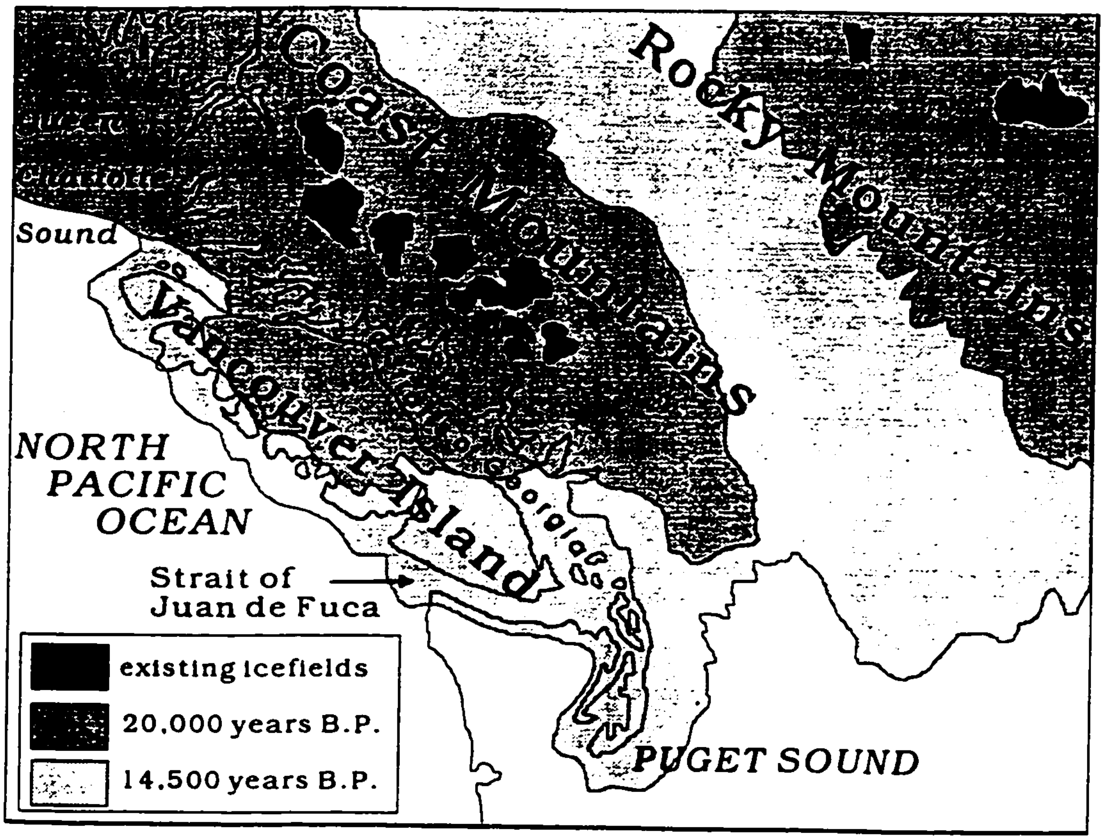


Figure 4: Extent of the Cordilleran Ice Sheet over southwest British Columbia and Washington at 20,000 ^{14}C yr BP and 14,500 ^{14}C yr BP, as well as extent of present glacier cover (from Mann and Hamilton, 1995).

5 STUDY AREA AND METHODS

5.1 SITE SELECTION

No previous in-depth study of late-glacial glacier advances had been conducted in the vicinity of Mount Rainier. The area, especially within Mount Rainier National Park and surrounding wilderness areas, is accessible and has not been developed substantially. Reconnaissance mapping (Crandell, 1969; Crandell and Miller, 1974) provides a starting point for identification of late-glacial deposits. The major valley floors surrounding Mount Rainier have been blanketed by postglacial debris flows and landslides. The densely forested character of valleys extending beyond the western and southern flanks of the mountain makes recognition of glacial features extremely difficult. By contrast, the lower peaks northeast of Mount Rainier and along the crest of the Cascade Range east of Mount Rainier have not been affected by large landslides or debris flows, and preserve numerous glacial features. Therefore, this study focuses on the late-glacial record east and northeast of Mount Rainier (Figures 5 and 6).

No glaciers occur in the main study area; current glaciers on the main cone of Mount Rainier were used for inferring comparative snowline relationships (chapter 5.4). As the full-glacial record in the main study area is very incomplete, sites from other areas in the vicinity of Mount Rainier were also examined.

The pronounced rainshadow effect of Mount Rainier makes it difficult to compare the climatic factors responsible for the glacier advances. To support the evidence from the main study area, glacial deposits in the Goat Rocks, ca. 45 km south-southeast of Mount Rainier, were examined. In this area, late-glacial terminal moraines occur within short distance of current glaciers, and the climatic factors can be compared without the complications of rainshadow effects.

5.2 BEDROCK GEOLOGY

Geologic maps of the Mount Rainier area have been compiled by Schasse (1987a, b). Further mapping is in process (D. Swanson, pers. comm., 1995). The bedrock of the western Cascade Range between the Cowlitz and White rivers consists mainly of andesitic and basaltic lavas and breccias, volcanic sedimentary rocks, and arkosic sandstones and shales of Eocene to Miocene age that have been folded and faulted. Subsequently, they were intruded by plutons, mainly granodiorite, of late Miocene age. Mount Rainier volcano was constructed during the Pleistocene on the eroded surface of volcanic and plutonic rocks of the Ohanapecosh, Stevens Ridge, and Fifes Peak formations of Tertiary age as a series of lava flows and breccias of pyroxene andesite (Fiske et al., 1963; Hammond, 1989). These lava flows formed thick intracanyon flows extending as far as 17 km radially outward from the volcano. Later erosion caused a topographic inversion, so that today, the lava flows mainly cap interfluves. On the upper slopes, thinner flows and interbedded breccia occur. Most lava flows are older than 100,000 years (Hammond, 1989). However, the most recent flows and pyroclastics that form the small summit cone complex probably are no older than 2500 years (Crandell, 1971). With the exception of debris flows, the edifice of Mount Rainier has experienced only minor changes in the last 100,000 years. For the purpose of this study, the edifice of Mount Rainier and the surrounding mountains are considered unchanged throughout the last 100,000 years. No major uplift or similar events have been documented that could influence the glacial record.

5.3 POSTGLACIAL DEPOSITS

TEPHRA LAYERS

At least 22 layers of tephra (volcanic ash and coarser airfall pyroclastic debris) have been found among the postglacial deposits on and near Mount Rainier (Mullineaux, 1974). Each layer records a separate eruptive event. Eleven tephra layers were erupted from Mount Rainier, ten originated on Mount St. Helens 80 km south-southeast of Mount Rainier, and one originated from Mount Mazama, Oregon (at the present site of Crater Lake), 440 km south of Mount Rainier.

The oldest recognized and dated tephra layer in the study area is Mount Rainier layer R (Mullineaux, 1974). Layer R occurs as a layer of brown, coarse lapilli northeast of the volcano (Fig. 7), and forms an 8- to 50-cm-thick layer over most of the study area (Mullineaux, 1974; Heine, 1996). Layer R is easily identifiable in the field, as it is the only coarse tephra that occurs below the distinctive Mazama ash (layer O) in the Mount Rainier area (Mullineaux, 1974).

Because tephra layer R is used as a key marker horizon for this study, its age is important. Previously, the age of layer R had been constrained only by one minimum radiocarbon date of 9990-9450 cal yr BP (8750 ± 285 ^{14}C yr BP) (Mullineaux, 1974). Several additional radiocarbon ages obtained during this study further limit the age of layer R (Table 1). A twig near the top of the layer (core 5, White River Park, Fig. 8) dates to 9980-9880 cal yr BP (8920 ± 60 ^{14}C yr BP). Because this twig may have been pushed down a few centimeters into the tephra by the drilling action of the corer, its date is treated here as a minimum date for the tephra layer. Maximum ages are provided by an AMS date on a small twig below layer R (core 9, White River Park, Fig. 8) of 10,000-9920 cal yr BP (8990 ± 60 ^{14}C yr BP) and a bulk date for gyttja below layer R (core 8, White River Park, Fig. 8) of 9880-9570 cal yr BP (8760 ± 80 ^{14}C yr BP). The seeming contradiction between the uncalibrated ^{14}C values of the minimum date of 8920 ± 60 ^{14}C yr BP and the maximum date of 8760 ± 80 ^{14}C yr BP does not occur with the calibrated ages, which overlap at 9880 cal yr BP. The calibrated 1σ intervals of all limiting ages for tephra layer R overlap at 9880 cal yr BP (8850 ^{14}C yr BP). As a result, this date is used for tephra layer R in this study.

After the deposition of layer R, Mount Rainier did not erupt a widespread tephra layer for about 2500 years (Fig. 9). A renewed phase of activity during the middle Holocene deposited 8 of the 11 postglacial tephras from Mount Rainier, at an average rate of one every 300 years. Over the last 4000 years, only two tephra layers have been deposited. The youngest, layer X, is ca. 150 years old (Mullineaux, 1974). Although other, minor eruptions may have occurred, their deposits have not yet been identified.

All but one of the tephra layers erupted from Mount Rainier are associated with explosive eruptions of molten lava. Color, grain size, layer thickness, and stratigraphic position permit field identification of most tephras where the sequence is fairly complete (Mullineaux, 1974). Whereas the content and refractive index of iron-magnesium minerals are useful for determining the source of a tephra layer (Mullineaux, 1974), many of the Mount Rainier tephras have a similar character (Mullineaux, 1974; C. Newhall, pers. comm., 1996). This makes identification of minor tephra layers and correlation of these layers from site to site difficult, especially in the case of previously unrecognized layers.

In addition to layer R, two well-dated, widespread, and distinctive ash layers occur throughout the field area and can be used as marker horizons for dating and stratigraphic correlation. Layer O (the Mazama ash) was erupted from Mount Mazama. In lake sediments, it is a distinct, gray silt, whereas in soil profiles it has oxidized to a yellowish-brown color of varying hues. The color of these tephra layers varies considerably due to oxidation. Layer O was deposited ca. 7640 cal yr BP (6850 ^{14}C yr BP) (Bacon, 1983). Layer Yn from Mount St. Helens was erupted ca. 3700 cal yr BP (3400 ^{14}C yr BP) (Mullineaux, 1974). It consists of light-yellowish-brown lapilli of varying hues. The grain size and layer thickness of layer Yn decrease from the southwest to the northeast across the Mount Rainier area. Other tephra layers occur, but were not used in this study as marker horizons.

Debris flows and other mass-wasting deposits

A number of volcanic debris flows have inundated areas on and around the volcano during postglacial time (Scott et al., 1995). The floors of several large valleys originating on Mount Rainier are mantled partially by debris-flow deposits (Crandell, 1971). For example, the Osceola Mudflow has left thick deposits in the White River valley, and the National Lahar blankets most older deposits in the upper reaches of the Nisqually valley (Scott et al., 1995).

Numerous landslides have occurred throughout the area. Most are shallow localized slides, but a number of the debris flows mentioned above originated in deep-seated landslides on the main cone of Mount Rainier (Scott et al., 1995). A number of rockfalls and debris avalanches have fallen onto glaciers on Mount Rainier. The rockfall of December 1963 onto Emmons Glacier caused a short-lived glacier advance (Driedger and Kennard, 1984).

5.4 METHODS

MAPPING MORAINES AND OTHER GLACIAL LANDFORMS

Moraines and glacial landforms were mapped in the field and confirmed by inspecting aerial photographs. Moraines were identified based on morphology and lithology, and grouped based on their relative positions with respect to the cirques and other moraines. Soil profiles on the moraines were exposed uncommonly. Where exposed, the soils at the top of the late-glacial moraines were poorly developed and displayed thin A/C profiles. The low degree of soil development probably is an indication of continuous erosion of the moraines. Therefore, soil development is not very useful as a relative-age criterion for the moraines in the study area. The lack of exposures makes the identification of catena developments along the eroding moraine slopes impossible.

DISTINGUISHING GLACIAL DEPOSITS FROM NONGLACIAL DIAMICTONS

Mass-wasting deposits often resemble glacial deposits. For example, coarse debris-flow deposits can be mistaken for till (Davis and Osborn, 1987). To distinguish between the two, several characteristics were used in this study. Debris-flow deposits usually are clast-supported and gradually become finer upward. In contrast, most moraines are matrix-supported and lack sorting. Debris flows often can be traced upstream from their distal margins, whereas moraines in many cases are associated with outwash deposits downstream. However, debris flows also can be associated with alluvium downstream. Additionally, debris flows originating on Mount Rainier tend to consist mainly of andesite, while moraines often are polylithologic and contain granitic clasts from intrusions near the volcano. All of these distinctions, however, are not universally applicable. Based on a single exposure, it often is extremely difficult to determine the origin of a deposit. However, large debris flow deposits only occur in the large valleys that originate on Mount Rainier itself, whereas the drainages originating on the smaller peaks northeast of Mount Rainier and along the crest of the Cascades east of Mount Rainier have not been affected by large debris flows. In the main study area, debris flow deposits are, at most, localized deposits. No debris flow deposits were identified in the main study area.

The deposits of small, shallow landslides usually are composed of angular to subangular rocks with variable amounts of matrix, and thus easily distinguished from glacial drift. Lateral moraines may be composed of similar sediment, but they usually form linear ridges parallel to the orientation of the valley, whereas landslides in most cases deposit lobes perpendicular to the valley orientation. However, in some cases, landslides originate in till deposits on the valley sides. The associated landslide deposits in such cases include subrounded to subangular sediments. In these cases, landslide deposits might be misinterpreted as terminal moraines. For this study, several indicators were used to distinguish between landslide deposits and moraines. After the completion of field mapping, the moraines were identified on aerial photographs. Ideally, a terminal moraine resembles a crescent-shaped ridge, whereas a landslide deposits a lobe. The slopes above the presumed moraines were examined for traces of landslide headscarps. In one case, a possible moraine located in the heavily forested American River valley was later identified as a landslide deposit based on a number of landslide headscarps on the adjacent valley walls. Because these headscarps often are revegetated, they can be difficult to identify, especially when several millennia have passed since the landslide occurred.

Within the study area, single moraines sometimes cannot be identified with certainty. For this study, the relative positions of the presumed moraines in several valleys were compared, as well as relevant age criteria (based on sediment cores behind the moraines) . At most one or two significant landslides occur in each valley, and so it is extremely unlikely that the same number of contemporaneous landslides would occur in several valleys in approximately the same relative position (with respect to each other and to the cirque headwall), and feature the same sediments and shape as a moraine.

DATING GLACIER ADVANCES

The late-glacial moraines on Mount Rainier were not dated previously, and no organic matter was found in the moraines during the present study. Therefore, alternative approaches were adopted to obtain a chronology.

Tephrochronology

Tephra layer R (9880 cal yr BP, 8850 ^{14}C yr BP) is a useful marker horizon for dating the late-glacial moraines. In addition, layers O and Yn provide useful stratigraphic markers at ca. 7640 and 3700 cal yr BP (6850 and 3400 ^{14}C yr BP), respectively.

Lake stratigraphy

Additional age control for the moraines was obtained by coring lake sediments. Basal sediments from a lake upstream from a moraine yield a minimum date for moraine deposition (Fig. 10). The sediments in a lake basin downstream from a moraine may in some cases contain an alluvial layer associated with the upstream ice advance (Fig. 10). Such layers commonly contain less organic matter than nonglacial gyttja, as silty sediments washed out of the glacier dilute the organic sediment input into the lake (Karlén and Matthews, 1992). As a proxy for organic content, loss on ignition was determined by measuring the weight loss incurred by burning a dry sample at 550°C. By dating organic matter incorporated in the glacigenic sediments, direct ages for the glacier advance can be obtained. Additional bracketing ages can be obtained by dating organic matter below and above the glacigenic layer.

Radiocarbon dating

The well-constrained tephrochronology (Fig. 9) was used for correlation and dating of the lake-sediment cores. In addition, 18 radiocarbon ages were obtained from organic matter in the cores (Table 1). Four of these ages were obtained by Accelerator Mass Spectrometry (AMS), and 12 were obtained using the conventional β-counting method. Radiocarbon analyses were performed by the Quaternary Isotope Laboratory of the University of Washington (β-counting) and Lawrence Livermore National Laboratories (AMS dating).

All but one of the radiocarbon ages are in stratigraphic order and are consistent with the ages of the dated tephra layers. Ages obtained from adjacent layers, both by conventional β-counting on bulk gyttja and AMS on macrofossils, are in sequence (core 9, Fig. 29), indicating that the conventional bulk ages are not affected by contamination with younger or older carbon. Furthermore, the minimum date for tephra layer R (9980-9880 cal yr BP, 8920 ± 60 ^{14}C yr BP, core 5, Fig. 8) was obtained from a tree branch (macrofossil), while the youngest maximum date (9880-9570 cal yr BP, 8760 ± 80 ^{14}C yr BP, core 8, Fig. 8) was obtained from a bulk sediment sample. Both ages are from adjacent meadows at White River Park (Fig. 6), one downstream from the other. Since deglaciation, the two meadows have been hydrologically connected. If the bulk date was contaminated with older carbon, it would have to be older than the overlying macrofossil date (in the sense of a composite stratigraphy; the actual ages are from different cores), yet it is slightly younger (the two ages overlap within the calibrated 1σ intervals, resolving the apparent radiocarbon inversion; see discussion of tephra layers above). Furthermore, contamination of the samples due to a hardwater effect is unlikely because the lakes occur in small drainages entirely within granitic and andesitic bedrock.

One AMS radiocarbon date for a small piece of wood below tephra layer R of 8320-8140 cal yr BP (7450 ± 60 ^{14}C yr BP) (core 7, not shown) was rejected, because it was significantly younger than the minimum ages for R. All other ages for layer R are in good agreement, suggesting that the small piece of wood in core 7 may have been pushed downward in the section during coring. It is very unlikely that tephra layer R in this core was misidentified, as it occurs below the distinctive, gray silty layer O (Mazama ash), and no other coarse tephra layer older than O has been found in the study area (Mullineaux, 1974; J. Vallance, pers. comm., 1997). Furthermore, the other dates for layer R in this study are from cores in the same and an adjacent meadow, where layer R occurs in the same stratigraphic position.

Coring lake sediment

Previous researchers were unable to penetrate the numerous and thick (up to 120 cm) tephra layers in the lake sediments with piston corers (Dunwiddie, 1983). For this study, an Eijkelkamp soil auger, which provides 15-cm-long core sections, was used for coring in bogs and meadows. The core sections were analyzed, recorded, and sampled in the field before removal from the auger. The drilling, rather than pushing, action of the soil auger ensures that little sediment is lost. In some cases, several redundant cores were taken in the same lake or meadow while attempting to reach the basal sediments. To save time in the field, the stratigraphy above the Mazama ash was not recorded in some of the redundant cores, as it is of marginal interest to this study.

Three cores were taken with a percussion corer based on the design of Reasoner (1986, 1993). This corer consists of a PVC pipe, with an outer diameter of 7.5 cm, which is driven into the sediments. Some core loss occurs at the transition from thick, cohesive tephra layers to soft, incohesive gyttja, as the cohesive tephra tends to "plug up" the opening of the core barrel; this displaces the underlying soft gyttja sideways away from the core barrel. Once the core head has passed the tephra layers, the core barrel widens and no further core loss should be expected. The oldest tephra layer (layer O, ca. 7640 cal yr BP, 6850 ^{14}C yr BP) penetrated with the percussion corer (Upper Dewey Lake, core 15; Lower Dewey Lake, core 19; Tipsoo Lake, core 13 and 14) was deposited more than 2000 years after deglaciation. The basal sediments, which were of primary interest in this study, in all cores occur at least 10 cm below the oldest tephra layer, and thus probably are not affected by such core loss.

Topographic reconstruction of former glaciers

Topographic maps of glaciers in the study area were reconstructed based on the position of the moraines and headwalls. Glacier thickness h was estimated using

$$h = \tau_b / (\rho g \sin \alpha),$$

with τ_b the shear stress across the base of the ice column, ρ the density of the ice, g the acceleration due to gravity, and α the surface slope of the glacier (Paterson, 1994, p. 239-240). For the reasonable approximation of ice as a perfectly plastic material, τ_b is 100 kPa, and $\tau_b / \rho g$ = 11 m (Paterson, 1994, p. 240).

The basal slope of a glacier was used as an approximation for the surface slope. For valley glaciers, a correction factor between 0.5 and 0.9 for the effect of the valley sides is inserted in the denominator on the right-hand side (Paterson, 1994, p. 240). This correction factor is estimated based on the ratio of glacier thickness to valley width (Paterson, 1994, p. 269). For this study, a parabolic valley cross-profile was assumed. The value of the correction factor depends on the ratio of valley-depth to valley-width (Paterson, 1994, p. 240).

As an example, the McNeeley 1 glacier at White River Park was reconstructed based on the location of the moraine and the headwall location (Fig. 43A). The glacier margin at the headwall is selected based on observations of current glaciers occupying similar cirques in the Cascades. No lateral moraines were found, but the lateral extent of the glacier is reasonably well-defined by the outline of the cirque. The slope of the valley floor near the center of the glacier is ca. 40% or 0.4, with sin $\alpha = 0.37$. A correction factor of 0.8 was used, because the glacier is confined only on one side by the cirque wall. The resulting glacier thickness h is calculated to be 37 m (or 121 ft) at the center of the glacier. Based on this, contour lines are drawn that result in an approximate reconstruction of the glacier topography.

The reconstructions of the glacier topography have several potential errors: The location of the glacier margin on the headwall and the cirque/valley sides are not well-known. The surface slope of the glacier is only approximated by the basal slope, and underestimated in the terminal zone. The correction factor for the valley sides is estimated.

A maximum and a minimum reconstruction of the glacier allow estimating the potential error associated with the glacier reconstructions (Fig. 11). The maximum reconstruction assumes that the glacier occupied the entire cirque, all the way to the headwall crest. The smallest reasonable value for the basal slope (0.25) was used for the reconstruction of the glacier thickness, with a correction factor of 0.7 for the effect of the valley sides. The resulting glacier thickness is 65 m (212 ft) at the ELA. The minimum reconstruction assumes that the glacier only occupied the cirque floor (Fig. 11). The steepest value for the basal slope (0.5) was used for the reconstruction of the glacier thickness, with no correction factor for the effect of the valley sides. The resulting glacier thickness is 25 m (81 ft). For small cirque glaciers, the ice margins of the glacier for the minimum and maximum reconstructions lie within 50 m. The calculated elevation of the glacier surface varies by about 50 m between the minimum and maximum reconstructions.

For large valley glaciers, the potential error is larger. In these cases, a reconstruction of the complex glacier topography can result in significant errors. However, the surface slope of a large valley glacier more closely approximates the average valley slope, except in the terminal zone. In these cases, an error of > 50 m for the glacier thickness seems possible. None of the late-glacial glaciers in the study area were large valley glaciers.

EQUILIBRIUM LINE ALTITUDES

The equilibrium line altitude (ELA), along which the glacier net mass balance is zero, reflects the climatic parameters that control the extent of a glacier. Comparing the ELAs of different glacier advances provides a measure of the changes in climate that are responsible for these advances. However, the ELA for a glacier advance cannot be converted easily into the underlying factors of winter snowfall and summer temperature.

ELAs were calculated for the glaciers reconstructed in this study, as well as for current glaciers. To determine the ELAs, the accumulation area ratio (AAR) method was used for all but the current glaciers (see below), with a an AAR of 0.65 taken as the ratio of accumulation area:total glacier area (Porter, 1975). The accumulation area at any time depends on the mass balance of the glacier. The above empirical ratio usually is valid for temperate valley glaciers that are in equilibrium (Paterson, 1994). Because end moraines usually are deposited by glaciers that are close to being in equilibrium, it is assumed that when glaciers on Mount Rainier deposited moraines, they were in equilibrium. A number of other factors can influence the above ratio. If a substantial portion of the accumulation area is shaded, yet the ablation area is not, the ablation area also will be smaller than predicted by the above method. These problems were taken into account in the present study and are mentioned where applicable.

The accuracy of these ELA reconstructions depends on the accuracy of the reconstructed glacier topography. The glacier area, and especially the thickness of the glacier at the ELA, are important. For the late-glacial glaciers, such as the glacier at White River Park (Fig. 11), the potential errors associated with the reconstruction of the glacier topography are relatively small. In the case of the glacier at White River Park, which is representative of the late-glacial glaciers that are the focus of this study, the possible error (ELA of maximum reconstruction minus ELA of minimum reconstruction) is less than 40 m (Fig. 11). Changing the accumulation-area ratio to 0.6 in the maximum reconstruction or to 0.7 in the minimum reconstruction introduces an additional potential error of ca. 15-20 m for the ELA. For the late-glacial glacier at White River Park, the highest likely ELA is 1865 m, whereas the lowest is 1790 m. Thus, the overall possible error for the ELA reconstruction is less than 75 m. The above reconstructions deliberately maximize the error, the actual error probably being considerably less. However, for the large valley glaciers of the last glaciation, the associated errors are somewhat larger. No lateral moraines have been found, and the thickness of the glacier has to be estimated based on basal shear stress. The associated error for the glacier thickness is estimated as ± 50 m, which results in an error for the ELA of possibly ± 80 m, or more.

Because no glaciers or recent moraines occur in the main study area, recent ELAs were mapped for this study on the large glaciers of Mount Rainier. The recent ELAs then were extrapolated into the main study area. The ELAs on Mount Rainier were mapped from aerial photographs taken in late September and early October of 1971, 1972, and 1973. The firn limit at the end of the melt season approximates the ELA for a steady-state temperate glacier (Paterson, 1994, p. 11). The mapped firn limits for the three consecutive years were in good agreement and their average probably approximates the ELA for those years. However, the winter of 1971/72 experienced anomalously high snowfall (ca. 180% of long-term mean), while during the winters 1970/71 and 1972/73, the snowfall was close to the long-term mean. Possibly the extremely high snowfall of 1971/72 influenced the reconstruction and yielded somewhat lower ELAs than the long-term mean. Other approaches (accumulation-area ratio, toe-to-headwall elevation ratio) did not yield accurate results for these glaciers because of their complex geometries, even when using the same aerial photographs.

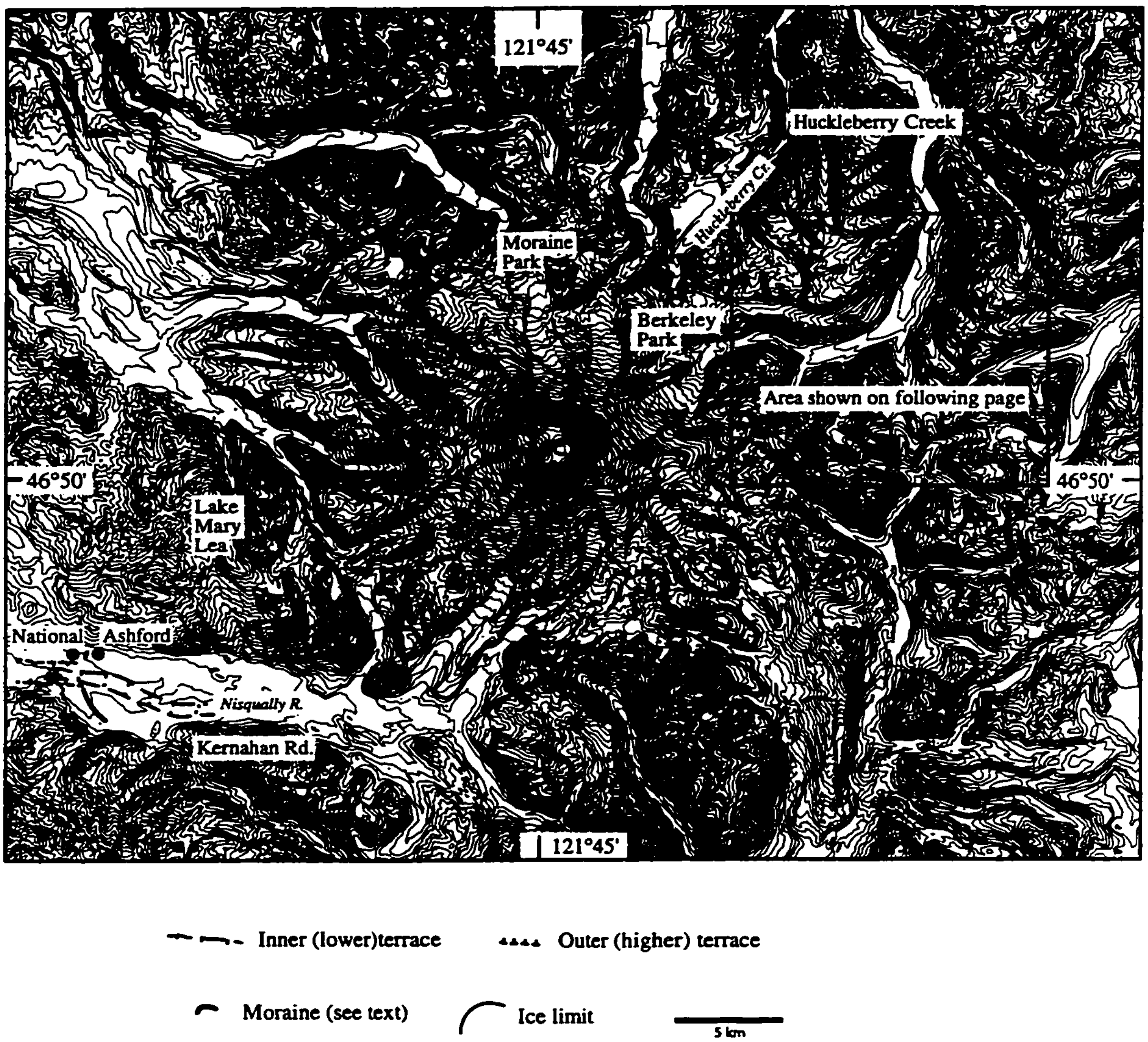


Figure 5: Overview map of study sites and glacial deposits on Mount Rainier. Contour interval is 50 m.

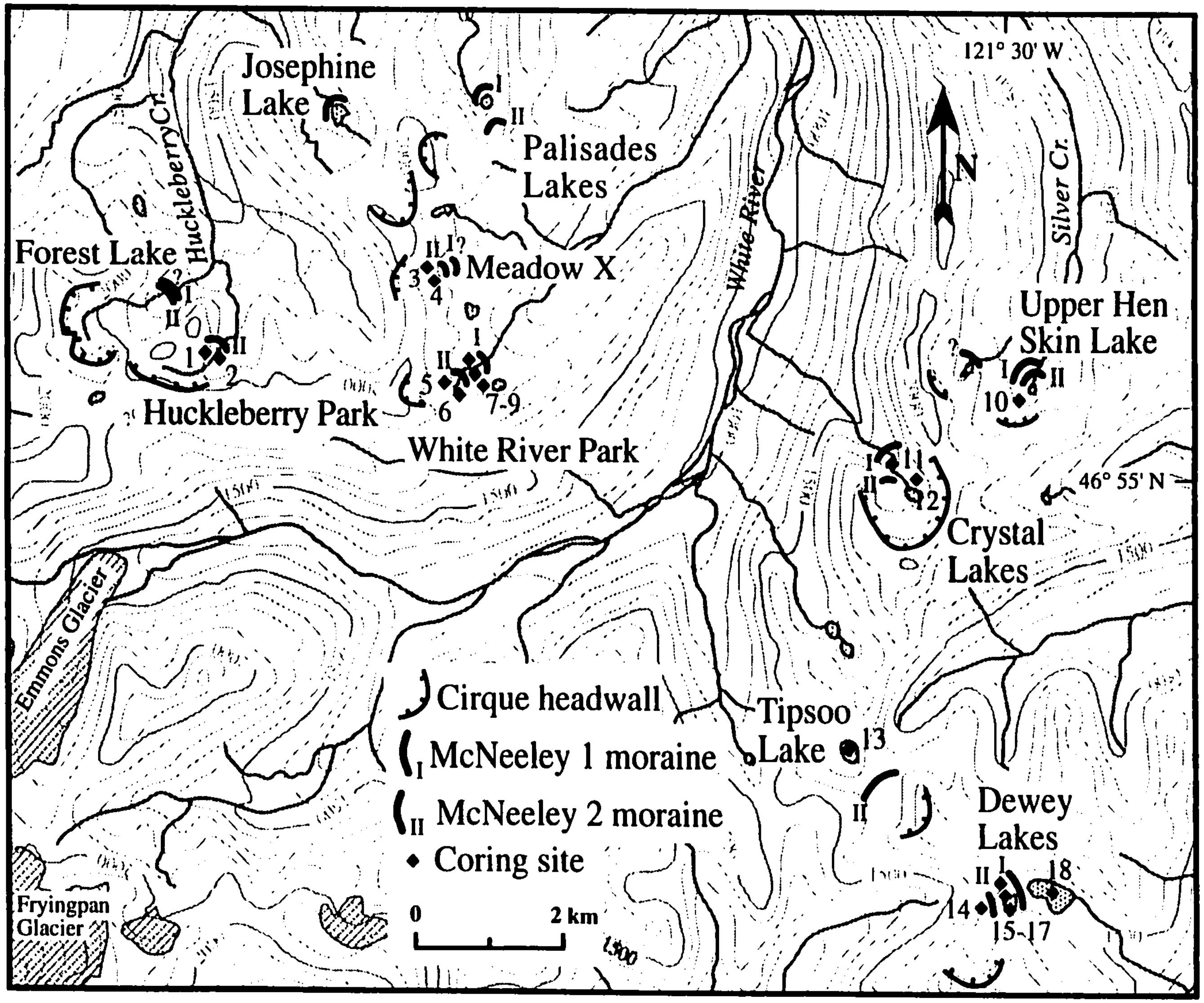


Figure 6: Overview map of glacial landforms in the main study area.

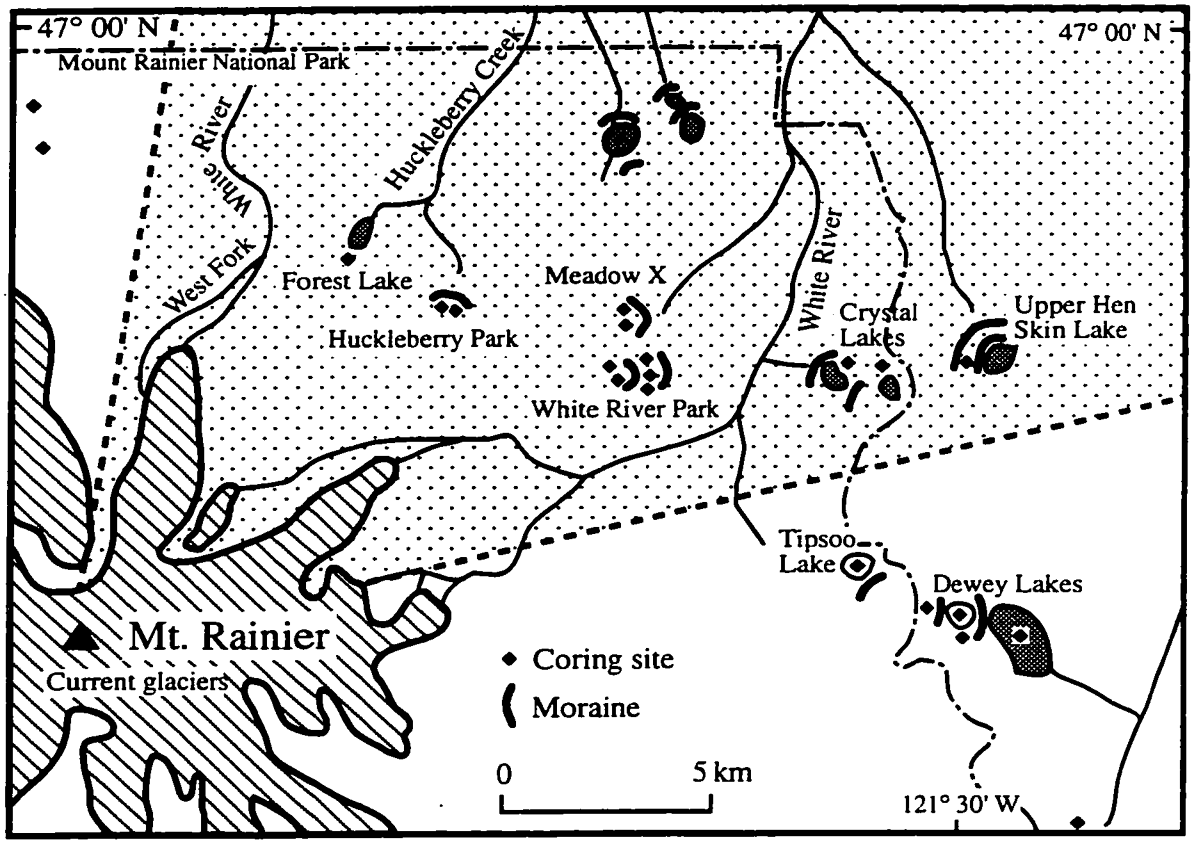


Figure 7: Distribution of tephra layer R based on cores taken in meadows, bogs, and lakes. This distribution pattern is in good agreement with the one inferred by Mullineaux (1974).

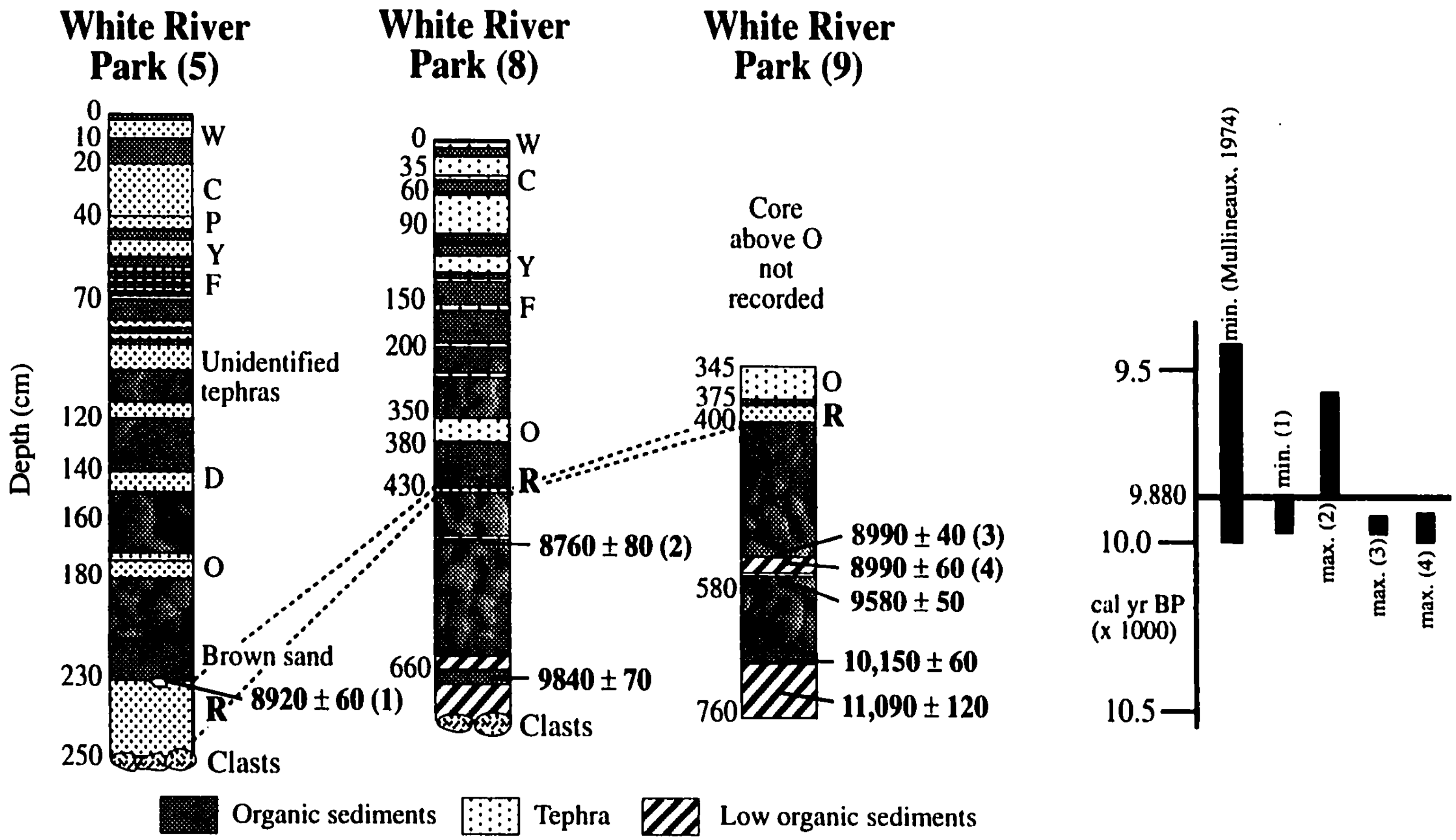


Figure 8: Radiocarbon ages constraining the age of tephra layer R in stratigraphic sections at White River Park (Table 2). Core numbers in parentheses. Uncalibrated radiocarbon ages in bold type. Capital letters designate tephra layers. On the right, calibrated bracketing ages for layer R. Bars reflect 1-sigma intervals. The ages overlap at 9880 cal yr BP (8850^{14}C yr BP).

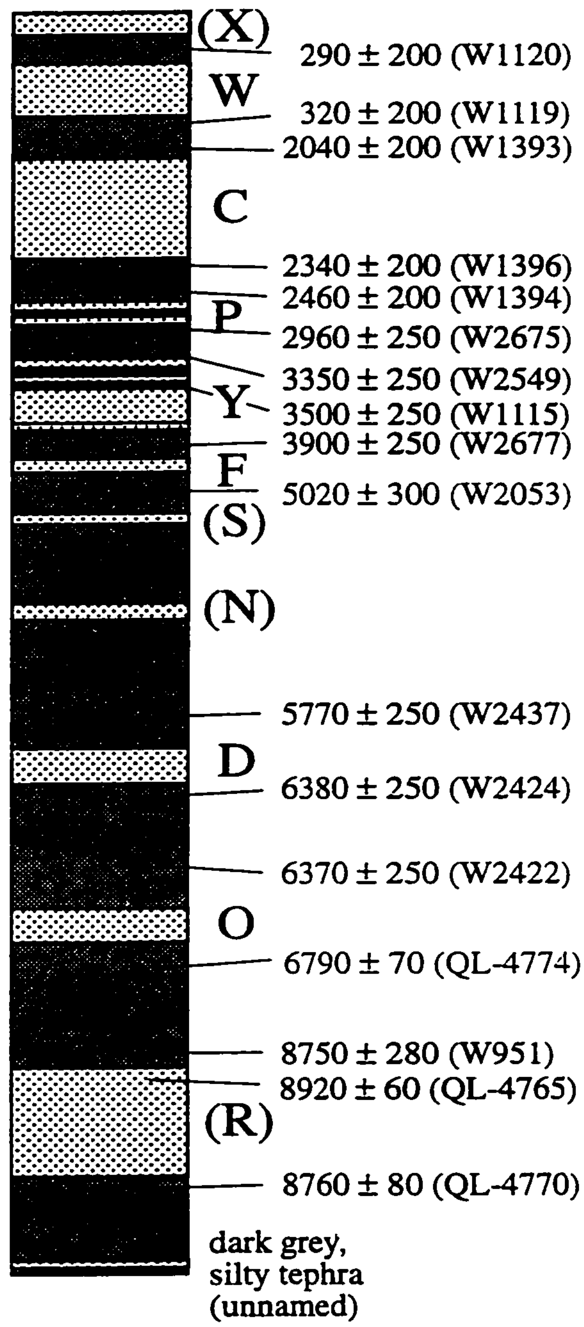


Figure 9: Composite Holocene lake/bog section in the study area with tephra layers (light stipple pattern), designated informally by letters (Mullineaux, 1974), and limiting radiocarbon dates (Mullineaux, 1974, and this study). Numbers in parentheses are laboratory numbers. Tephra layers in parentheses do not occur over the entire field area. Layers W, P, and Y are Mount St. Helens tephras; layer O was erupted from Mount Mazama.

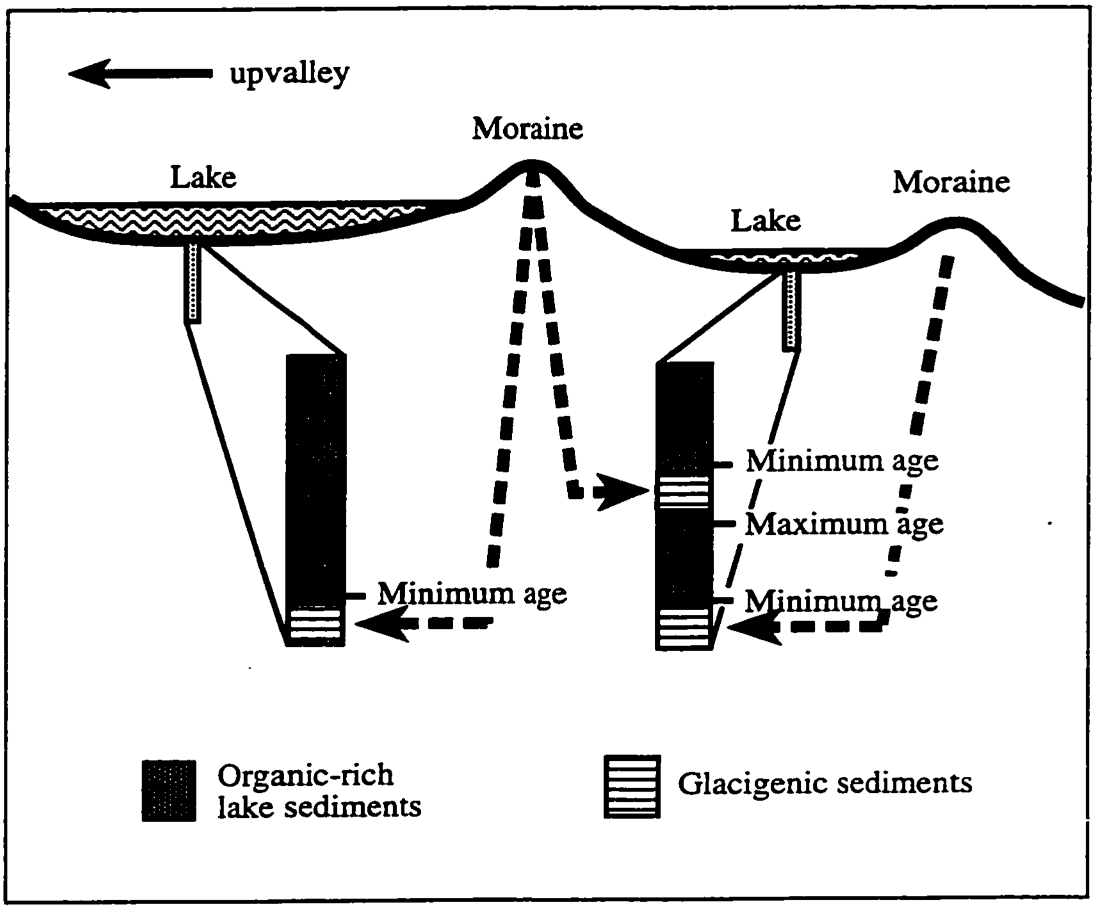
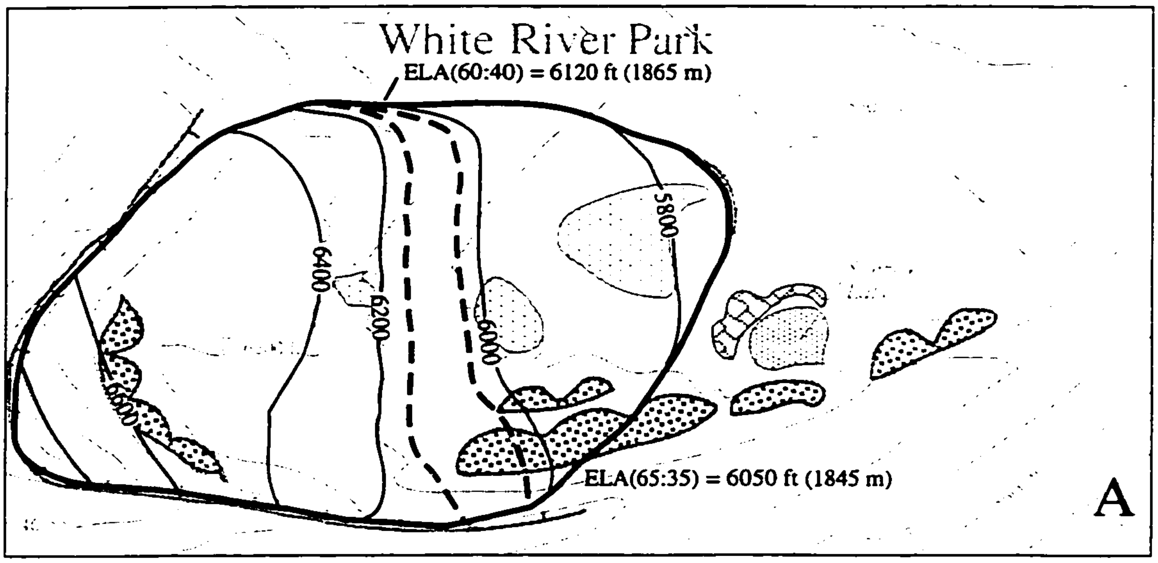


Figure 10: Idealized stratigraphy of lakes associated with moraines. Dashed arrows indicate relationships of lake sediments with moraines. Coring a lake upvalley from a moraine yields minimum ages for the moraine. In lakes downvalley from a moraine, a glacigenic layer may occur, which is contemporaneous with the moraine, and bracketing ages can be obtained.



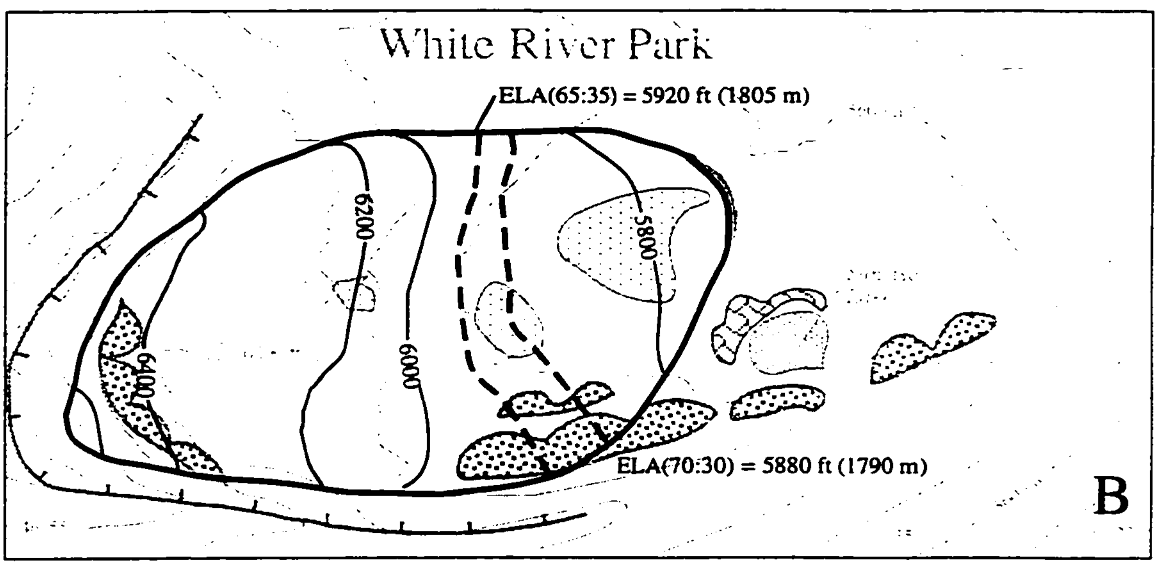


Figure 11: Maximum (A) and minimum (B) reconstructions of glacier topography at White River Park. Contours are in feet. Dashed line is ELA. Numbers in parentheses are accumulation area ratio and ELA in m, respectively. Additional legends in Figure 19.

6 EXTENT OF PRESENT AND FORMER GLACIERS AROUND MOUNT RAINIER

6.1 CURRENT GLACIERS

Although no glaciers occur within the study area, Mount Rainier supports 25 major glaciers (Driedger and Kennard, 1984). The glaciers and perennial snowpatches cover ca. 9.2×10^6 m^2. The ice volume is ca. 4.4×10^9 m^3 (Driedger and Kennard, 1984).

6.2 MAXIMUM GLACIER EXTENT DURING THE LAST (FRASER) GLACIATION ON MOUNT RAINIER

The till of the last (Fraser) glaciation (Evans Creek Drift) was mapped by Crandell (1969) and Crandell and Miller (1974). Prior to the present study, no age control existed for this unit from the field area. The deposits representing the maximum advance of ice during the last glaciation were mapped in three areas to compare their extents with those during the late-glacial periods of glacier advance. The term "last glacial maximum" (LGM) is avoided here, as this implies a correlation with the last maximum of the Laurentide and other ice sheets around 20,000 ^{14}C yr BP.

HUCKLEBERRY CREEK

Huckleberry Creek originates in the Huckleberry Basin (see chapter 6.3) near McNeeley Peak (Fig. 5). A major tributary drains the Forest Lake area (see chapter 6.3). A large ridge extends on the right (south) shore perpendicular to the Huckleberry Creek valley downstream from the confluence with Ada Creek at ca. 960 m elevation (Fig. 12). The primary vegetation in this part of the valley consists of mature old-growth trees. Pits created by upending of the root structure of fallen trees provide exposures of the sediments on the ridge. The exposed sediment of the ridge consists of nonsorted, subrounded to subangular cobbles in a sandy and silty matrix. No bedrock was found on the ridge or in the bed of Huckleberry Creek at this location. Thus, it is unlikely that the ridge is formed by an intrusive dike. No lava flows occur in the area, and no landslide scars are visible on airphotos and topographic maps that might explain the origin of the ridge as a landslide deposit. However, the area is densely forested, and it is difficult to follow the ridge across the valley. The ridge does not seem to continue on the left (north) side of the stream. Tentatively, the ridge is interpreted as a moraine deposited by a glacier in the Huckleberry Creek valley. No other moraines have been identified upstream between this moraine and the late-glacial moraines (see chapter 6.3). No age control exists for the advance at this site other than the relative position of the moraine.

NISQUALLY RIVER VALLEY

Moraines

The Nisqually River originates at the terminus of Nisqually Glacier on the south side of Mount Rainier. Crandell and Miller (1974) mapped the type Evans Creek moraine as a terminal moraine between National and Ashford (Fig. 13). This landform is exposed in a roadcut along State Route 706 (Fig. 15). The till in the lower part of the section consists of cobbles in a highly indurated, sandy and silty matrix, and is interpreted in this study as a lodgment till, rather than the till of a single terminal moraine (Crandell and Miller, 1974). The upper part of the section consists of loose cobbles and pebbles in a sandy and silty matrix, with reworked brownish tephra interspersed throughout the layer. This layer is interpreted as reworked till. It is possible that this reworked layer contains the National lahar.

A ridge on the south side of the Nisqually River near the town of National at ca. 475 m elevation (Fig. 13) may be a moraine. It trends at an angle to the axis of the valley, with a slight crescent shape toward the center of the valley. The ridge is about 900 m long, ca. 50 m wide, and ca. 10 m high. It is longer and more linear than a landslide deposit. The bed of the Nisqually River ca. 150 m from the end of the ridge is deeply incised. No bedrock crops out in the channel. Thus, it is unlikely that the ridge is formed by bedrock. It is reasonable to infer that the ridge is a moraine deposited by the Nisqually glacier. No exposure was found in the ridge. A pit dug at the top of the ridge exposed 50 cm of coarse, well-sorted sand, very similar to the sands deposited by the National Lahar in this area (Scott et al., 1995). This coarse sand blankets most deposits in the valley outside the modern floodplain.

Further evidence for a former ice limit near the town of National comes from a west-sloping trimline that is visible from a distance on the upper slopes of the northern valley side. Projected downvalley, it intersects the valley floor near National, consistent with an ice limit, as inferred by Crandell and Miller (1974) and as suggested by the terminal moraine found in this study.

Numerous debris flows, rockfalls, landslides, and other mass-wasting deposits have obscured many of the glacial landforms in the Nisqually River valley. No moraines were found between the moraines mapped near National and the Neoglacial moraines within ca. 1 km of the terminus of Nisqually Glacier, nor in any tributary valleys.

Outwash terraces

In the present study, two terraces were identified in the Nisqually River valley (Fig. 13). Their surfaces are blanketed by coarse sandy lahar deposits similar to those overlying the inferred moraine near National (see above). Numerous gravel pits and streambanks expose rounded to well-rounded, stratified cobble-pebble gravel, interpreted as outwash. In each terrace, the mean grainsize decreases downvalley. In some places, stratified silts and clays are interstratified with the cobble-pebble gravel. The terraces were mapped in the present study by establishing the elevation of the terrace remnants several tens of meters from the terrace edge with a precision altimeter (+/- 3 m) (Fig. 14), and by comparing the terrace elevation on both sides of the river using a surveying level. Each terrace probably formed during recession following a glacier advance, when outwash gravels were deposited in front of the glacier as a valley train. After the glacier retreated, Nisqually River cut its bed into the plain, leaving the terraces on the valley sides.

The inner (lower) terrace originates near Kernahan Road (Fig. 13) at ca. 545 m and almost continuously extends downvalley past National. No moraines have been identified at or near Kernahan Road. The outer (upper) terrace originates at National (ca. 500 m downstream from Ashford) and is consistent with an ice-front position marked by the moraine near National.

No dating control exists on the glacial deposits in the Nisqually valley other than their relative positions. Relative dating methods, such as soil development, cannot be used on these deposits, as they are blanketed by a coarse sandy lahar.

Lake Mary Lea

Lake Mary Lea is located ca. 22 km west of the summit of Mount Rainier, on a private tree farm. The area has been logged repeatedly. Lake Mary Lea lies in a well-preserved cirque (Fig. 16), facing north-northwest. The cirque floor is at 1325 m elevation. Ca. 1 km south of Lake Mary Lea, another cirque faces north-northwest. The cirque floor is at 1310 m elevation.

Both cirques feature a single, ca. 3-m-high, round-crested ridge at the threshold of the cirque, interpreted as moraines. Till is exposed in streamcuts where the ridges have been eroded by creeks draining the cirques. Based on the elevation of the cirques, the ridges are assumed have been deposited during full-glacial time. The ridges cannot be compared to the other suspected full-glacial moraines, as they are much smaller in scale. The ridges at Lake Mary Lea and the neighboring cirque are similar to the late-glacial moraines described in chapter 6.3. Considering that they occur in a gentler terrain and at significantly lower altitudes than the late-glacial moraines, it is conceivable that they are considerably older than the late-glacial moraines having a similar appearance. Several cores (cores 18 - 20, Fig. 17) in the cirques failed to penetrate a tephra layer that is several meters thick. The tephra layer itself has not been recognized previously, and its origin is unknown. A radiocarbon age of grass and other organic matter on top of the tephra layer of 24,120 ± 340 ^{14}C yr BP (approximately 27,500 cal yr BP) provides a minimum date for the glacier advance that deposited the moraine. Because till was not reached in this core, this date may not closely date ice retreat from the moraine.

Summary of evidence for the maximum glacier extent of the last glaciation on Mount Rainier

The preserved glacial landforms deposited by the maximum advance of the last glaciation on and near Mount Rainier are few and insufficiently dated. The moraine record most likely is incomplete. The outwash terraces in the Nisqually valley probably provide a more complete record of the glacier advances during the last glaciation. Two sets of outwash terraces imply two glacier advances. In other parts of the Cascades, two distinct glacier advances have been identified during the last glaciation (Porter, 1976). Swanson and Porter (1997) obtained ^{36}Cl dates for boulders on moraines in the Wenatchee and Yakima river valleys. The ^{36}Cl dates indicate that mountain glaciers in the southern North Cascades advanced during the Evans Creek and Vashon advances. Since these sites are located only ca. 30-60 km northeast of the Nisqually valley, it is likely that the two outwash terraces in the Nisqually valley were deposited during the Evans Creek and Vashon advances, respectively.

The age of the moraine at Ada Creek in the Huckleberry valley is unknown. Assuming glacier advances around Mount Rainier were synchronous, the moraine at Ada Creek may have been deposited at the time of the Vashon advance, as indicated by the apparent lack of moraines between the moraine at Ada Creek and the late-glacial moraines. Alternatively, the moraine may have been deposited during the Evans Creek advance, and the Vashon moraine is not preserved or was not identified.

The moraines at Lake Mary Lea were deposited before ca. 27,500 cal yr BP (24,000 ^{14}C yr BP). Because till was not reached in the cores, this date might not closely date the retreat of glaciers from these moraines. If the moraines were deposited during the Evans Creek advance, it appears that this advance culminated before ca. 27,500 cal yr BP (24,000 ^{14}C yr BP) in this cirque. However, it is possible that the moraines at Lake Mary Lea were deposited by an advance predating the Evans Creek maximum (detailed discussion in chapter 8).

6.3 EXTENT OF GLACIERS IN THE VICINITY OF MOUNT RAINIER DURING LATE-GLACIAL TIME

Moraines inferred to correlate with Crandell and Miller's (1974) McNeeley drift were mapped within or beyond cirques throughout the field area (Figs. 5 and 6). In most valleys, two moraines occur in close proximity. The older one here is designated McNeeley 1 and the younger McNeeley 2. With one exception, no lateral moraines are preserved. This probably is due to relatively unstable slopes in the cirques occupied by the small McNeeley glaciers, which caused any lateral moraines to erode, or be buried by sliderock and colluvium. No boulders were found on most moraines.

Huckleberry Basin

The southwestern part of Huckleberry Basin is a large cirque (Fig. 18), which formed on the north side of the Sourdough Mountains. The cirque is oriented to the east-northeast, and its floor lies between 1830 and 1980 m (6000 and 6500 ft) elevation.

A number of ridges occur on the south side of the basin. The ridges are about 3 m high, have sharp crests and lack trees. The ridges appear very fresh, comparable to moraines deposited elsewhere on Mount Rainier during the Little Ice Age. Behind the innermost ridges, tephra layer F, deposited ca. 5700 cal yr BP (5000 ^{14}C yr BP), was penetrated at the base of a core. The age of the sediments behind the ridge contrasts sharply with the fresh morphology of the ridge itself. No tephras older than several centuries were found on the ridge. The ridge probably was deposited as a protalus rampart during the Little Ice Age. Unlike a flowing glacier, the perennial snow cover causing the formation of a protalus rampart does not disturb the sediments it overlies.

East of McNeeley Peak, a crescent-shaped ridge extends across the entire valley of Huckleberry Creek at an elevation of ca. 1770 m (5800 ft). The ridge is about 2 m high and has a rounded crest. Where Huckleberry Creek cuts through the ridge, nonsorted, loose, subrounded to subangular cobbles and pebbles in a sandy and silty matrix are exposed in the streambank. No bedrock is evident in the streamcut. The sediment in the ridge most likely is till. Based on the sediment, crescent shape, elevation, and position in the cirque, the ridge is interpreted as a McNeeley terminal moraine. A second round-crested ridge is visible downstream from this moraine on airphotos, but is not identifiable in the field because of dense forest cover. While this may be a McNeeley terminal moraine, the lack of field evidence precludes inclusion in this study.

White River Park

The cirque at the head of White River Park (Fig. 19) faces northeast, and its floor is between 1830 and 1890 m (6000 and 6200 ft) elevation.

Sunrise Lake is dammed by hummocky terrain that originated as a series of landslides from the steep valley walls above. The deposits are not moraines, as they are located on the cirque floor and oriented parallel to the expected direction of glacier flow in the cirque.

They also are too far from the valley walls to be protalus ramparts. Numerous protalus ramparts occur along the cirque walls.

Two ridges occur on the basin floor of White River Park. The lower ridge, at 1710 m (5610 ft) elevation, is very subtle, no more than 2 m high, and has a rounded crest. In a streamcut near the ridge, loose, nonsorted, subangular to subrounded cobbles in a sandy and silty matrix are exposed. No bedrock is exposed in the streamcut. The sediment of the ridge is interpreted as till. Based on the sediment, position in the cirque, elevation, and shape, the ridge is interpreted as a McNeeley 1 terminal moraine. The moraine dammed a lake that has filled with sediments to become a bog.

A well-preserved, round-crested ridge, located at ca. 1750 m (5740 ft) elevation, is interpreted as a McNeeley 2 moraine, based on its crescent shape, its position in relation to the McNeeley 1 moraine and its elevation, as well as the absence of bedrock nearby. The McNeeley 2 moraine dams a bog.

Meadow X

The meadow north of Clover Lake (Fig. 19) was named informally Meadow X for this study. The cirque at the head of the creek in Meadow X is located on the east side of the Sourdough Mountains (Fig. 19). The cirque faces east-northeast, and its floor lies between 1770 and 1860 m (5800 and 6100 ft) elevation.

Meadow X is dammed by a pronounced ridge at ca. 1720 m (5640 ft) elevation. The ridge is about 4 m high and has a rounded crest. It is interpreted as a McNeeley moraine, based on its position in the cirque, its elevation, its crescent shape, and the absence of bedrock nearby. Based on the stratigraphy of sediments in two cores from Meadow X (chapter 6.4), the moraine is assigned to the McNeeley 2 advance. Downstream from this ridge, hummocky terrain occurs at ca. 1700 m (5580 ft) elevation, with till exposed in stream cuts. The hummocky terrain may be the remnant of a McNeeley 1 moraine.

Palisades Lakes

On the southeast side of the Palisades, a large cirque is oriented to the east (Fig. 20). The elevation of the cirque floor is ca. 1860 m (6100 ft). A glacier filled the cirque, and then split into two lobes, one descending toward Sunrise Creek, while the other filled the basin of Upper Palisades Lake and then descended toward Lost Creek. A small lake near Lower Palisades Lake in the Lost Creek valley is dammed by a round-crested ridge at ca. 1650 m (5410 ft) elevation. The ridge is ca. 3 m high. In a streamcut at the outlet of the lake, loose, nonsorted, subangular to subrounded cobbles in a sandy and silty matrix are exposed. No bedrock occurs in the streamcut. Another small lake ca. 50 m upstream is dammed by a round-crested ridge at ca. 1665 m (5460 ft) elevation. The ridge is ca. 2 m high. Based on the sediments, the location, the elevation, and the absence of bedrock nearby and in the streamcut, the ridges are interpreted as McNeeley 1 and McNeeley 2 moraines, respectively.

JOSEPHINE LAKE

Josephine Lake is located in a long, shallow valley that originates in a cirque on the northwest side of Marcus Peak (Fig. 20). The cirque faces to the northwest, and its floor lies between 1920 and 2010 m (6300 to 6600 ft) elevation.

Josephine Lake is dammed by a sharp-crested ridge at ca. 1650 m (5410 ft) elevation. The ridge is ca. 5 m high. Loose, nonsorted, subrounded to subangular cobbles in a sandy and silty matrix are exposed in a stream cut at the outlet of the lake. No bedrock occurs in the streamcut. Based on its sediments, its crescent shape, and the absence of bedrock nearby, the ridge is interpreted as a McNeeley moraine. No other moraines were found in the Josephine Creek valley. Therefore, it cannot be determined whether the moraine was deposited by the McNeeley 1 or McNeeley 2 advance. It is possible that the McNeeley 2 advance in this drainage was more extensive than the McNeeley 1 advance, and that the McNeeley 1 moraine is not preserved as a result. It is possible that only one late-glacial moraine was deposited in this drainage area. A small lake ca. 300 m upstream from Josephine Lake is dammed by a landslide from the Palisades.

CRYSTAL LAKES

Upper and Lower Crystal lakes are located within a steep-walled compound cirque on the west side of an unnamed peak on the crest of the Cascades (Fig. 21). Lower Crystal Lake lies at 1660 m (5450 ft) elevation. Above the lake, the cirque consists of three parts, carved out by three small glaciers that coalesced at the lake. The western cirque faces north-

northeast, and its floor lies between 1820 and 1920 m (6000 and 6300 ft) elevation. The central cirque faces north, and its floor lies at ca. 1860 m (6100 ft) elevation. The eastern cirque faces northwest, and its floor lies at ca. 1800 m (5900 ft) elevation. Taluses and protalus ramparts occur along the base of the cirque headwalls.

Lower Crystal Lake is dammed by a round-crested ridge, ca. 4 m high, at ca. 1660 m (5450 ft) elevation. The creek draining the lake has cut several meters into the ridge, and loose, nonsorted, subangular to subrounded cobbles are exposed in a sandy and silty matrix. The sediment composing the ridge most likely is till. This, combined with the crescent shape of the ridge, its location in the cirque, and its elevation point to a McNeeley moraine. Upper Crystal Lake is dammed by a bedrock sill. No moraines are preserved between Upper and Lower Crystal lakes. The steep terrain associated with the bedrock sill makes moraine preservation unlikely. A crescent-shaped ridge in the central cirque was identified on airphotos, and confirmed on the ground. The ridge is located at ca. 1700 m (5580 ft) elevation. It is sharp-crested and ca. 3 m high. Based on its position in relation to the moraine damming Lower Crystal Lake and its elevation, this ridge is interpreted as a McNeeley moraine. Based on their positions relative to each other, the moraine damming Lower Crystal Lake is interpreted as a McNeeley 1 moraine, whereas the moraine upstream is interpreted as a McNeeley 2 moraine.

Crystal Mountain

The Crystal Mountain ski area, located in the valley of Silver Creek, encompasses a number of cirques. The landscape of the developed part of the ski area, mostly north of Elizabeth Creek, has been altered substantially to develop ski runs. The southern part mostly remains close to its natural state. The cirque at the head of Upper Hen Skin Lake faces northeast (Fig. 22). Its floor lies between 1830 and 1950 m (6000 and 6400 ft) elevation. Several lakes are dammed by ridges of subangular to subrounded diamicton, interpreted as till. However, roads and chair lifts were built near Elizabeth Lake, and only remnants of what may have been a more-complete record of glacier advances remains. A paired set of sharp-crested ridges dams Upper Hen Skin Lake and an adjacent meadow, on which this discussion focuses.

A long, round-crested ridge extends along the meadow north of Upper Hen Skin Lake and curves southeastward to enclose the meadow at ca. 1690 m (5550 ft) elevation. Loose, nonsorted, subangular to subrounded cobbles in a brown sandy and silty matrix are exposed in a streamcut at the outlet of the meadow. No bedrock is visible in the streamcut. The sediments probably are till. Based on the sediments, the location, and the crescent shape, the ridge is interpreted as a lateral and terminal moraine of a McNeeley 1 glacier.

Upper Hen Skin Lake is dammed by a crescent-shaped ridge at ca. 1700 m (5580 ft) elevation. The ridge is ca. 3 m high and features a rounded crest. Loose, nonsorted, subangular to subrounded cobbles in a sandy and silty matrix are exposed where a creek draining the meadow north of Upper Hen Skin Lake (see above) has cut laterally into the ridge. The sediments are interpreted as till. The ridge is interpreted as a McNeeley 2 terminal moraine, based on its location, crescent shape, and sediment composition.

Tipsoo Lake

A cirque, facing west-northwest, occurs on the west side of the crest of the Cascade Range near Chinook Pass (Fig. 23). The cirque floor lies between 1700 and 1830 m (5600 and 6000 ft) elevation. The cirque contains several taluses and protalus ramparts. Tipsoo Lake, located immediately downstream from the cirque, is dammed by a bedrock ridge.

A small unnamed lake on the floor of the cirque is dammed by a crescent-shaped ridge at ca. 1630 m (5350 ft) elevation. The ridge is about 4-6 m high and has a rounded crest. State Route 410 provides a long exposure of the ridge in a roadcut. The ridge is composed of loose, nonsorted, subangular to subrounded cobbles in a sandy and silty matrix. Based on the sediments, the crescent shape, and the location, the ridge is interpreted as a McNeeley moraine. The stratigraphy of sediments in a core from Tipsoo Lake (see chapter 6.4) suggests that the moraine was deposited by the McNeeley 2 advance. Other moraines or glacial landforms may not be preserved in the area, which may be due to the steep terrain, as well as grading in connection with the construction of a parking lot.

Dewey Lakes

A cirque is located in the headwaters of Dewey Creek on the north-northeast side of an unnamed peak on the crest of the Cascades (Fig. 24). The cirque is not very well developed and faces northeast. The cirque floor is sloping and lies between 1710 and 1770 m (5600 and 5800 ft) elevation.

Upper Dewey Lake is dammed by a large, round-crested ridge at ca. 1560 m (5120 ft) elevation. The ridge is 2-3 m high, and has numerous rounded boulders at the surface up to 1.5 m in diameter. A streamcut through the ridge exposes loose, nonsorted, subrounded to subangular cobbles in a sandy and silty matrix. The sediment probably is till. No bedrock is exposed in the streambed. Based on the sediments, the crescent shape, and the location, this ridge is interpreted as a McNeeley 1 moraine.

A meadow above Upper Dewey Lake is dammed by a crescent-shaped ridge at 1570 m (5150 ft) elevation. No exposures were found, but the shape of the ridge and the lack of a steep slope above the ridge preclude landslides as the origin of this feature. Thus, it is likely that the ridge is a moraine, deposited by the McNeeley 2 advance, as it is located upstream from the other McNeeley moraine.

Lower Dewey Lake is dammed by a bedrock sill. The glacier that occupied the valley of Dewey Creek spilled over the divide on the crest of the Cascade Range and deposited moraines that now dam Upper Dewey Lake and a meadow above Upper Dewey Lake.

OTHER SITES

Remnants of late-glacial moraines have been identified in a number of other places. Because the stratigraphy at these sites is not entirely clear, they were not used for paleoclimatic reconstruction.

Clover Lake

The cirque upstream from Clover Lake is located between the cirques at White River Park and Meadow X (Fig. 19). It faces east-northeast, and its floor is between 1830 and 1890 m (6000 and 6200 ft) elevation. Clover Lake is dammed by a ridge. In several exposures, loose, nonsorted, subrounded to subangular cobbles in a sandy and silty matrix are exposed. The sediment at the surface of the ridge probably is till. However, in a streamcut at the outlet of Clover Lake, bedrock is visible. Thus, it cannot be determined whether the ridge is a moraine or a bedrock ridge mantled with till. No other moraines have been found in the drainage.

Forest Lake

Forest Lake is dammed between rockslide deposits. Two cirques occur upvalley from Forest Lake (Fig. 6). The western cirque faces north-northeast, and its floor lies at ca. 1980 m (6500 ft) elevation. The eastern cirque faces north-northeast, and its floor lies at ca. 1950 m (6400 ft) elevation. Glaciers formed in both cirques and converged upstream from Forest Lake. Two round-crested ridges occur downstream from the lake. The west fork of Huckleberry Creek has cut into one of the ridges, and loose, nonsorted, subangular to subrounded cobbles in a sandy and silty matrix are exposed. No bedrock is exposed in the streambed. The sediment probably is till. The ridges probably are McNeeley 1 and McNeeley 2 moraines. However, the proximity of the ridges to landslide deposits argues against including them in this study.

Berkeley Park

A large cirque on the north side of Burroughs Mountain at Berkeley Park (Fig. 5) faces north, and its floor lies at ca. 1950 m (6400 ft) elevation. A prominent moraine shortly below the cirque threshold was mapped as a McNeeley moraine by Crandell and Miller (1974).

Moraine Park

A cirque at Moraine Park, northeast of Carbon Glacier, faces west-northwest (Fig. 5), and its floor lies at ca. 1790 m (5870 ft) elevation. Three closely spaced ridges occur in the cirque. These ridges were mapped by Crandell and Miller (1974) as McNeeley moraines. The ridges are crescent-shaped, about 3 m high, and feature rounded crests. A creek has eroded the meadow inside the innermost moraine (exposure 5), where loose, nonsorted, subangular cobbles are exposed in a sandy and silty matrix. The ridges probably are moraines deposited during the McNeeley advances.

6.4 AGE OF MCNEELEY DEPOSITS IN THE VICINITY OF MOUNT RAINIER

HUCKLEBERRY PARK

McNeeley 2 glacier

Near its source spring, Huckleberry Creek has cut its channel about 1 m into a meadow (Fig. 18) and flows on cobble-rich till (exposure 1, Fig. 25). The till is overlain by tephra R. The cirque must have been largely deglaciated by the time layer R was deposited.

The McNeeley moraine in the Huckleberry Basin dams a former lake that has filled with sediments and now is a bog. Core 2 from the bog reached tephra layer R, but did not penetrate this layer due to influx of groundwater. In core 1 (Fig. 26) about 10 cm of gray, organic-poor silt lies above rocks at a depth of 250 cm, presumably till, that could not be penetrated. The silt in turn is overlain by layer R. Thus, the glacier retreated from the McNeeley moraine before tephra layer R was deposited.

WHITE RIVER PARK

McNeeley 1 glacier

The McNeeley 1 moraine at White River Park dams a bog (Fig. 19). Core 7 from the bog reached the Mazama ash (O), but failed to penetrate it due to influx of groundwater. Core 8 (Fig. 27) could not penetrate a basal stone layer, presumed to be till, at a depth of 700 cm. The core displays all Holocene tephra layers deposited in the area. A layer of low-organic, gray silt overlies the basal till. Above the gray silt, a ca. 10-cm-thick layer of organic sediments follows. A thin layer of low-organic sediments overlies the oldest organic sediments (see discussion of McNeeley 2 advance below). Above this layer, sediments throughout the section are high in organic content with the exception of the tephra layers.

The onset of organic sedimentation in core 8 was dated to 11,010-10,980 cal yr BP (9840 ± 70 ^{14}C yr BP). Sedimentation rates in the early Holocene were high, with 180 cm deposited within ca. 1100-1300 years between the onset of organic sedimentation and the next layer sampled for radiocarbon dating (9880-9570 cal yr BP; 8760 ± 80 ^{14}C yr BP). The overlying 50 cm between the level of the upper radiocarbon date and tephra layer R may also reflect rapid accumulation, as the date of R (9880 cal yr BP, 8850 ^{14}C yr BP) and the radiocarbon date are virtually identical (within 1σ).

The stratigraphy of core 9 (Fig. 27) is similar to that of core 8 (Fig. 27). However, at a depth of 760 cm (the length of the available corer) the core had not reached basal till. The thick layer of low-organic, gray silt above the till was reached, indicating that till probably was not far below the bottom of the core.

A radiocarbon date near the base of core 9 of 13,130-12,880 cal yr BP (11,090 ± 120 ^{14}C yr BP) provides a minimum date for deglaciation from this site behind the McNeeley 1 moraine. A layer of low-organic sediments was deposited during the early Holocene, and is interpreted as glacigenic sediments associated with the McNeeley 2 advance upstream (see below).

McNeeley 2 glacier

The McNeeley 2 moraine at White River Park (Fig. 19) dammed a lake that filled with sediments to become a meadow. Two cores were raised. Both contain almost identical stratigraphies (cores 5 and 6, Fig. 28), and both reached a stone layer, presumed to be basal till, that could not be penetrated, at a depth of 260 and 250 cm, respectively. The stone layer is overlain directly by tephra layer R in core 5. In core 6, about 5 cm of dark gray silt separate R from the till.

A ca. 5-cm-long tree branch without bark was imbedded near the top of tephra layer R in core 5. The branch has a radiocarbon date of 9980-9880 cal yr BP (8920 ± 60 ^{14}C yr BP), which indicates that the site was deglaciated before 9880 cal yr BP (8860 ^{14}C yr BP).

A ca. 20-cm-thick, light-colored, silty layer occurs in core 9 downstream from the McNeeley 2 moraine (Fig. 29). Loss-on-ignition (LOI) is a proxy for organic-matter content. LOI is low at the base of the core (<5%), then increases to ca. 23% with the onset of organic sedimentation. In the light-colored layer, LOI decreases to 12%, then increases again in the overlying gyttja to ca. 25%.

The low organic content at the base of the core is consistent with other cores taken behind McNeeley 1 moraines. The second, upper layer with low organic content is interpreted as sediments deposited by meltwater from the McNeeley 2 glacier, which diluted the higher-organic sediments normally accumulating in the lake. At the same time when this silty, low-organic layer was deposited, no sedimentation occurred in the lake upstream from the McNeeley 2 moraine at White River Park (cores 5 and 6, Fig. 28). When high-organic sedimentation resumed in the downstream lakes, sedimentation commenced in the upstream lake. A glacier probably occupied the cirque and the area covered by the upstream lake while the silty, low-organic layer was deposited in the downstream lake. Similar silty, low-organic sediment layers have been found in downstream lakes in southern Norway, where they were correlated with upstream glacier advances (Karlén and Matthews, 1992).

The layer immediately overlies a dark fine ash, which probably is correlative with the ash below a similar glacigenic layer in core 13 from Tipsoo Lake (see below). A twig in the glacigenic layer in core 9 was AMS-dated to 10,000-9920 cal yr BP (8990 ± 60 ^{14}C yr BP), and probably provides a direct date for the McNeeley 2 advance. A minimum radiocarbon date of 9990-9940 cal yr BP (8990 ± 40 ^{14}C yr BP) and a maximum radiocarbon date of 10,900-10,480 cal yr BP (9580 ± 50 ^{14}C yr BP) also support an early Holocene date for the McNeeley 2 advance at this site. The organic sedimentation below the glacigenic layer suggests that the cirque was largely ice-free before glaciers reoccupied the cirque during the McNeeley 2 advance. Thus, the McNeeley 2 moraine is regarded as having been deposited by a readvance and not by a recessional stillstand.

Several alternative explanations might be suggested for the upper layer low in organic matter. It could stem from increased sediment input due to unstable slopes, or be caused by lower productivity of the vegetation around the lake. However, if the sediment input increased, so would sedimentation rates. Yet the sedimentation rate in the low organic layer is significantly lower than in the layers above and below. Furthermore, pollen studies from lowland sites in the Pacific Northwest (chapter 4.3) indicate that climate was warmer and more productive during the early Holocene. However, none of these studies are from alpine sites at comparable altitude. Finally, landslides would produce a coarser sediment input into the lake than the fine silty sediments that occur in core 9.

MEADOW X

McNeeley 1 glacier

No locations suitable for coring exist between the possible McNeeley 1 moraine and a McNeeley 2 upstream moraine (see below).

McNeeley 2 glacier

Meadow X is dammed by a McNeeley 2 moraine (Fig. 19). Two cores (core 3 and 4, Fig. 28) provide minimum ages for the McNeeley 2 glacier advance. Both cores show a similar stratigraphy. At a depth of 405 cm in core 4, the corer reached a stone layer, presumed to be basal till, that could not be penetrated. The stone layer is overlain by 3 cm of brown, low-organic, clayey sediments, overlain by tephra layer R, which is ca. 15 cm thick in core 4. In core 3, basal till at a depth of 230 cm is overlain directly by tephra layer R. Thus, sedimentation at Meadow X started shortly before deposition of tephra layer R (ca. 9880 cal yr BP, 8850 ^{14}C yr BP). The site must have been ice-free by that time.

PALISADES LAKES

No cores were taken in this drainage, and no age control exists for the moraines.

JOSEPHINE LAKE

The lake was not cored, and no age control exists for the moraine.

CRYSTAL LAKES

McNeeley 1 glacier

Lower Crystal Lake (Fig. 21) was not cored, but a core was taken in a meadow adjacent to the lake (core 11). Even though the coring location is within the fallout area of tephra layer R (Fig. 7), R was not found in the core. Either the core did not reach the oldest sediments, or R is missing, indicating an interruption of sedimentation or a period of erosion at this site in a marginal location. R is present in the upstream core near Upper Crystal Lake (see below). This excludes the possibility that Lower Crystal Lake was deglaciated after R was deposited.

McNeeley 2 glacier

In a core from a bog adjacent to Upper Crystal Lake (core 12, Fig. 26), a stone layer was reached at a depth of 400 cm and could not be penetrated. This stone layer is interpreted as McNeeley 2 till. No moraine is preserved between Upper and Lower Crystal lakes, but a moraine in another part of the cirque indicates that McNeeley 2 glaciers reached a position between the two lakes (chapter 6.3). The stone layer is overlain by ca. 4 cm of low-organic, gray silt, in turn overlain by tephra layer R (ca. 8 cm thick). Above layer R, organic-rich lake sediments occur, as well as all Holocene tephra layers. The lake sediments give way to peat in the upper part of the core. Sedimentation in Upper Crystal Lake began shortly before tephra layer R fell out (9880 cal yr BP, 8850 ^{14}C yr BP). The site must have been deglaciated by this time.

UPPER HEN SKIN LAKE

McNeeley 1 glacier

A meadow between the McNeeley 1 and McNeeley 2 moraines at Upper Hen Skin Lake (Fig. 22) was cored (core 10, Fig. 27). A stone layer, presumably basal till deposited by the McNeeley 1 glacier, was reached at a depth of 480 cm and could not be penetrated. The till is overlain by low-organic, gray silt, the color of which darkens toward the top. At 420 cm, a transition occurs to a dark gyttja. At 400 cm, a fine, dark-gray ash occurs as a thin, ca. 0.5-cm-thick band, which may correlate with the dark-gray ash found below the glacigenic layers at Tipsoo Lake (see below) and at White River Park (see above). Tephra layer R occurs above this dark gray tephra as a ca. 1-cm-thick band. A radiocarbon date from the gray silt near the bottom of the core of 13,240-11,840 cal yr BP (11,120 $\pm$ 200 ^{14}C yr BP) provides a minimum date for glacier retreat from the McNeeley 1 moraine.

McNeeley 2 glacier

Five attempts to core the small meadow upstream from Upper Hen Skin Lake were unsuccessful in penetrating a layer of wood, which occurs at a depth of ca. 3 m.

Tipsoo Lake

McNeeley 2 glacier

Tipsoo Lake is located downstream from a McNeeley moraine (Fig. 23). Core 13 (Fig. 29) from Tipsoo Lake could not penetrate a stone layer at a depth of 80 cm. This layer may be till from the McNeeley 1 advance. The stone layer is overlain by several centimeters of gray silt. About 5 cm of dark-gray gyttja follow, overlain in turn by a dark-gray fine sandy tephra layer. Above the tephra layer, a distinct, 6-cm-thick layer of gyttja slightly lighter in color with abundant oxidized Fe/Mg minerals occurs, followed in turn by dark gyttja and the Mazama ash (O). The light-colored layer is unique in the entire core, and no similar layer has been observed in any other core in this study, with the exception of the glacigenic layer in core 9 from White River Park (see above). Low LOI values (<5%) in the bottom gray layer are consistent with similar layers in all cores spanning late-glacial sedimentation. The abrupt transition to dark gyttja coincides with in an equally abrupt increase in LOI to more than 25%. The lighter-colored layer with the Fe/Mg minerals shows a distinct decrease in LOI to 12%. This stratigraphy is very similar to the one of core 9 at White River Park. In the same manner as the correlated layer in that core, the low LOI in this layer is interpreted as influx of glacigenic sediments from the upstream McNeeley 2 glacier. The Fe/Mg mineral grains may be the result of abrasion on the granitic rocks in the cirque floor upstream. A minimum radiocarbon date of 10,280-9990 cal yr BP (9140 ± 100 ^{14}C yr BP) and a maximum date of 11,880-11,120 cal yr BP (10,080 ± 60 ^{14}C yr BP) bracket the age of the McNeeley 2 advance at this site. In addition, the dark gray tephra layer immediately below the glacigenic layer probably is correlative with the dark gray tephra layer occurring directly below the glacigenic layer at White River Park (see above). Since the geochemical signatures and mineral contents of all tephras erupted from Mount Rainier are similar, no unequivocal correlation of the tephras has been made. However, the radiocarbon ages corroborate the inferred correlation.

While alternative explanations cannot be excluded for this layer (see discussion under White River Park), it seems likely that the low-organic layers in both core 9 at White River Park and core 13 at Tipsoo Lake are the result of glaciers occupying the cirques upstream from the lakes.

Tipsoo Lake apparently lies outside the fallout area of tephra layer R (Fig. 7).

Dewey Lakes

McNeeley 1 glacier

Upper Dewey Lake is dammed by a McNeeley 1 moraine (Fig. 24). Three cores from the lake display a similar stratigraphy, with the exception of the basal sediments. Core 16 and 17 (Fig. 30) reached a stone layer, interpreted as basal McNeeley 1 till, that could not be penetrated. Core 15 was taken parallel to core 16 and features the same stratigraphy, but ends shortly above the stone layer as the depth of the core was restricted by the available corer. Core 17 was taken from the center of the lake with the percussion corer. The core barrel was driven 300 cm into the lake bottom, but only 140 cm of core were retrieved because of core loss in the upper parts of the core (chapter 5.4). However, because the core loss occurs in conjunction with thick, cohesive tephra layers, the basal sediments should not be affected (chapter 5.4). The core loss does not affect the interpretation of minimum ages. A fine gray ash layer, less than 3 mm thick, found in both core 15 and 16 near the margin of the lake basin is missing in core 17.

An oxidized, low-organic layer occurs above the basal rock layer (core 17, Fig. 30). A radiocarbon date of 13,320-13,150 cal yr BP (11,320 ± 60 ^{14}C yr BP) provides a minimum date for retreat from the McNeeley 1 moraine. Sediment influx changed to unoxidized organic gyttja about 10,780-10,370 cal yr BP (9460 ± 50 ^{14}C yr BP). A date for the basal organic sediments in core 16 of 10,850-10,380 cal yr BP (9500 ± 70 ^{14}C yr BP) suggests that the oldest sediments in the lake basin were not reached in this core from a very marginal location. The lake level may have been slightly lower for several millennia after deglaciation. The coring site might not have been covered by water, in which case no sediments were deposited.

McNeeley 2 glacier

A meadow above Upper Dewey Lake is dammed by a McNeeley 2 moraine (Fig. 24). Core 14 (Fig. 26) from this meadow reached a stone layer, presumably till, that could not be penetrated. Overlying the stone layer are 5 cm of gray, low-organic silt. This layer in turn is overlain by dark gyttja. A radiocarbon date of 10,280-9990 cal yr BP (9140 ± 100 ^{14}C yr BP) provides a minimum date for ice retreat from the McNeeley 2 moraine. Tephra layer R did not reach this area (Fig. 7).

Both in a core from Lower Dewey Lake and in the cores from Upper Dewey Lake (cores 15-17, Fig. 30), no glacigenic sediments from the upstream glacier advances were found (McNeeley 1 for Lower Dewey Lake and McNeeley 2 for Upper Dewey Lake). It is possible that during both advances, glaciers spilled over the divide only for a short time, long enough to deposit a moraine, but not long enough to wash enough sediments into the relatively large lakes to form distinct sediment layers.

OTHER SITES

Upstream from Forest Lake (Fig. 6), the west fork of Huckleberry Creek has cut into a meadow (exposure 3, Fig. 31). This exposure may the same that Mullineaux (1974, p. 13) described as the McNeeley site. Till is exposed at the base of the section, overlain by ca. 5 cm of gyttja, which in turn is overlain by tephra layer R. This site was deglaciated before layer R was deposited about 9880 cal yr BP (8850 ^{14}C yr BP).

Upstream from the moraine at Berkeley Park (Fig. 5), tephra layer O (Mazama ash) is exposed in a streamcut, indicating that the site was deglaciated before 7640 cal yr BP (6850 ^{14}C yr BP).

In a streamcut in the innermost moraine at Moraine Park (Fig. 5), basal till is overlain by peat (exposure 4, Fig. 31). A radiocarbon date from the lowest part of the peat layer of 7640-7540 cal yr BP (6790 ± 70 ^{14}C yr BP) is a minimum date for glacier recession. However, the date is from sediments on a moraine, where erosion is likely to have occurred. It is likely that deglaciation occurred several millennia earlier when the other McNeeley sites were deglaciated.

6.5 DISCUSSION OF THE MCNEELEY GLACIER ADVANCES

GLACIER EXTENT

Glaciers advanced twice near the Pleistocene/Holocene transition, first during the McNeeley 1 advance, then again during the McNeeley 2 advance. The moraines of these two advances are preserved in numerous drainages (Fig. 6). In most cases, the McNeeley 1 moraine is located a few tens of meters downvalley from the McNeeley 2 moraine. In several cases, only one McNeeley moraine is preserved. The other McNeeley moraine is not preserved, possibly due to postglacial erosion, or because the McNeeley 2 glaciers may have overrun the moraines deposited by McNeeley 1 advance. The glaciers of the McNeeley 1 advance and those of the McNeeley 2 advance were very similar in size (Fig. 32).

TIMING OF GLACIER ADVANCES

Of 42 cores obtained for this study, 18 reached basal till and display the entire postglacial stratigraphy. The well-dated tephrochronology, with layer R as a marker horizon, and 18 radiocarbon ages constrain the ages of the McNeeley 1 and McNeeley 2 glacier advances (Table 1).

McNeeley 1 advance

The age of the McNeeley 1 advance is constrained by three minimum ages between 13,320 and 12,880 cal yr BP (1-sigma interval) (11,320 ± 60 and 11,090 ± 120 ^{14}C yr BP) (Fig. 33). These radiocarbon ages are from separate drainages, yet they are statistically identical at the 2σ confidence level. This implies that sedimentation in the lakes started shortly after deglaciation. If there was a substantial lag between deglaciation and onset of sedimentation, the basal ages probably would scatter more widely, unless the lag was caused by a regional climatic condition that prevented any sedimentation in the lakes. Further support for a short lag comes from the nearby lakes dammed by the younger McNeeley 2 advance, where sedimentation started within less than 350 years after deglaciation (Heine, in press). Therefore, it is reasonable to treat these minimum ages as close minimum ages and to hypothesize that glaciers retreated from the McNeeley 1 advance shortly before approximately 13,200 cal yr BP (11,300 ^{14}C yr BP).

McNeeley 2 advance

Tephra layer R, deposited about 9900 cal yr BP (8900 ^{14}C yr BP) (Heine, 1996), almost directly overlies the McNeeley 2 till in all four lakes cored within the fallout area and within the glacial limit of the McNeeley 2 advance (three of the four cores are shown in Fig. 34). The absence of organic sediments below layer R in these cores implies that layer R was deposited within decades or possibly a few centuries after deglaciation. An additional minimum date of 10,280-9990 cal yr BP (9140 ± 100 ^{14}C yr BP) outside the fallout area of R (above Upper Dewey Lake, core 14, Fig. 34) further constrains the age of the McNeeley 2 moraines.

An alternative interpretation for the age of the McNeeley 2 advance is that the onset of sedimentation consistently lags deglaciation by hundreds or thousands of years. However, within the fallout area of layer R, the tephra layer in all cirques overlies the basal till with no more than 5 cm of inorganic sediments in between, yet layer R is found in all cirques. This indicates that the onset of sedimentation was almost simultaneous in all cirques, probably within decades or possibly several centuries. The small amount of inorganic sediments below layer R most likely does not reflect 1000 or more years of sedimentation at the same time when at a distance of 50 m or less, the downstream lakes (dammed by the McNeeley 1 moraines) experienced organic sedimentation at the highest rates throughout the record. Because the upstream lakes are sediment traps for a considerably larger drainage area than the downstream lakes, sedimentation rates in the upstream lakes would be expected to exceed those of the downstream lakes. A climate that severely limited both erosion and organic productivity could explain a consistent lag in sedimentation in the upstream lakes. However, the downstream lakes would have been affected by this climatic event as well, yet they experienced high rates of organic sedimentation.

In two drainages, cores were obtained from two lakes downstream and within 150 m of McNeeley 2 moraines (cores 9 and 13, Fig. 35). A distinct, light-colored sediment layer occurs in the lower part of each core. Loss-on-ignition measurements (a proxy for organic content) decrease in this layer in each core by more than 50% compared to the overlying and underlying sediments (from >25% to ca. 12%). In core 13 from Tipsoo Lake, small Fe/Mg mineral grains occur only in this layer, which may be evidence for glacial abrasion on the granitic bedrock in the cirque. Based on the combined evidence of color, organic content, and occurrence of lithic grains, I interpret these layers as glacier-derived sediments, and therefore as evidence for a glacier advance upstream. A dark gray, fine tephra layer, probably originating from Mount Rainier, occurs immediately below the glacigenic sediments and allows correlation of the two cores.

The core from White River Park (core 9) has a relatively high sedimentation rate (760 cm in ca. 13,000 yr) and good resolution. The glacigenic layer is AMS-dated directly from a twig within the layer to 10,000-9920 cal yr BP (8990 ± 60 ^{14}C yr BP). The age of the advance is also bracketed by a maximum date of 10,900-10,480 cal yr BP (9580 ± 50 ^{14}C yr BP) and a minimum date of 9990-9940 cal yr BP (8990 ± 40 ^{14}C yr BP). The sedimentation rate in Tipsoo Lake (core 13) is lower by an order of magnitude than at White River Park (core 9); therefore the resolution is not as good as in core 9 from White River Park. However, the ages from Tipsoo Lake confirm an early Holocene date for the sediment layer deposited by the advance, with a maximum date of 11,880-11,120 cal yr BP (10,080 ± 60 ^{14}C yr BP) and a minimum date of 10,280-9980 cal yr BP (9120 ± 80 ^{14}C yr BP). The above ages indicate that glaciers advanced during the McNeeley 2 advance some time after 10,900 cal yr BP (9650 ^{14}C yr BP). Glaciers retreated about 10,000 cal yr BP (9000 ^{14}C yr BP).

The McNeeley 2 moraines most likely were deposited by a glacier readvance, rather than a recessional stillstand. In both cored lakes downstream from McNeeley 2 moraines, sediments rich in organic matter occurred before the lakes returned to low organic sedimentation during the McNeeley 2 advance. In both cases, the cirques are small, and must have been largely ice-free to enable the deposition of organic sediments.

An alternative interpretation of the sediment record is possible: If the lag in all cirques was similar in duration, yet not induced by climate (since the lakes immediately downstream do not show evidence for a lack of sedimentation), and if the silty, low-organic sediment layers in both cored lakes downstream from the McNeeley moraines are unrelated to glacier activity, but were caused by early Holocene landslides or similar events in both cirques, then the glacier advance could be older than inferred above. Following this interpretation, the minimum ages from the upstream lakes are minimum ages, while the ages from the downstream lakes are meaningless. Thus, the McNeeley 1 advance would have occurred at some time before ca. 10,280-9990 cal yr BP (9140 ± 100 ^{14}C yr BP), as indicated by the oldest minimum date from above Upper Dewey Lake.

6.6 EXTENT OF GLACIERS IN THE GOAT ROCKS AREA

The Goat Rocks (Fig. 36) are the remnants of a volcanic cone, located about 45 kilometers south-southeast from Mount Rainier on the crest of the Cascades. The andesitic lava flows of the Goat Rocks were deposited ca. 3 to 1 million years ago (Schasse, 1987b). Schasse (1987b) mapped the extent of the Evans Creek till based on weathering characteristics. The highest elevation of the Goat Rocks, Tieton Peak (2368 m) rises no more than 500 to 600 m above the average elevation of the crest of the Cascade Range in this area. Therefore, the only significant rainshadow effect in this area is associated with the linear crest of the Cascade Range. The valleys of the North Fork of the Tieton River and Conrad Creek (Fig. 37) are aligned northeast-southwest, and west-east, respectively, parallel to the predominant moisture flux in the area. It is possible to compare the extent of glaciers during late-glacial time with the current glaciers.

North Fork of the Tieton River

The North Fork of the Tieton River originates in a glaciated cirque near Tieton Peak (Fig. 36). In late 1996, the glacier in the cirque was about one third the size mapped on the USGS 1:24,000 topographic map (Old Snowy Mountain quadrangle), which was compiled and field checked between 1959 and 1963.

Last (Fraser) glaciation

No moraines were found in the valley of the North Fork of the Tieton River between the McNeeley moraines (see below) and the confluence of the North Fork and Hell Creek, where a wide, round-crested ridge extends across the North Fork valley at 940 m elevation (Fig. 36). The ridge is ca. 10 m high and ca. 50 m wide. Exposures in a roadcut and in streambanks of Hell Creek display loose, nonsorted, rounded to subangular pebbles, cobbles, and boulders in a sandy and silty matrix. Dark andesite and porphyritic rhyolite with andesitic phenocrysts are the dominant lithologies. Numerous boulders are more than

2 m in diameter. The ridge is unlikely to have been deposited as an alluvial fan at the break in slope where Hell Creek enters the valley of the North Fork of the Tieton River. Outwash typically displays better sorting than the sediment in the ridge. A landslide deposit would not feature as many subrounded boulders, unless it was reworked glacial drift. Furthermore, rhyolite does not crop out near the ridge nor in the Hell Creek drainage, but it does occur in the headwaters of the North Fork of the Tieton River (Schasse, 1987b). Based on this evidence, the ridge is interpreted as a moraine deposited during the last glaciation.

McNeeley 1 glacier

At the confluence of the North Fork of the Tieton River with an unnamed creek draining the McCall Basin, the outer of two ridges of hummocky terrain occurs on both sides of the river at 1340 m elevation (Fig. 37). No bedrock is visible in the beds of either stream, implying that the ridge is not a dike resistant to weathering. Subrounded cobbles occur at the surface of the ridge. Where the unnamed creek has cut into the ridge, nonsorted, loose, subrounded to subangular cobbles in a sandy and silty matrix are exposed. Landslides have occurred in this area, but all clearly identifiable landslide deposits (with associated headscarps on the valley walls above) consist of angular and subangular sediments. Thus, the sediment forming the ridge is interpreted as till. The ridge itself probably is a moraine deposited by the McCall glacier. Based on its relative position in the valley in relation to other moraines and on its elevation, this moraine probably was deposited during the McNeeley 1 advance.

McNeeley 2 glacier

About 200 m upstream from the McNeeley 1 moraine at 1350 m elevation, an 8- to 10-meter-high, round-crested ridge occurs on the left (northwest) side of the North Fork of the Tieton River (Fig. 37). The streambed adjacent to the ridge consists of well-rounded cobbles. No bedrock occurs in the stream bed. Immediately downstream from the ridge, loose, nonsorted, subangular cobbles and boulders in a coarse sandy matrix are exposed in the streambank. Upstream from the ridge, recent landslides have deposited angular clasts. The sediment associated with the ridge is distinct from both the angular landslide deposits and the rounded fluvial gravel. The deposit is interpreted as till. Based on the sediments, the position in relation to other moraines in the valley, and the elevation, the ridge is interpreted as a moraine deposited by the McCall glacier during the McNeeley 2 advance.

Conrad Creek

Conrad Creek originates between Gilbert Peak and Tieton Peak (Fig. 37). In late 1996, the glacier in the cirque at the head of Conrad Creek was about one third the size mapped on the USGS 1:24,000 topographic map (Old Snowy Mountain quadrangle), which was compiled and field checked between 1959 and 1963. This indicates that the glacier has been receding in recent decades.

McNeeley glaciers

In the Conrad Creek valley two adjacent ridges occur perpendicular to the valley trend within a short distance at ca. 1340 m elevation (Fig. 37). The round-crested ridges are about 4-6 m high and ca. 25 m wide. Numerous rounded to subangular dark andesitic boulders up to 3 m in diameter occur on the surface of the ridges. Where a tributary creek has cut into the ridge, nonsorted, rounded to subangular cobbles and boulders in a sandy and silty matrix are exposed. No bedrock is visible in the ridges, nor in the bed of Conrad Creek at the site, reducing the likelyhood of a dike or bedrock strike ridge resistant to erosion forming the ridges. No traces of landslides occur uphill from the ridges. The ridges are longer, narrower, and more linear than landslide deposits. Debris flow deposits generally trend parallel to the orientation of the valley, not perpendicular to the valley axis. The ridges probably are moraines deposited by the unnamed glacier between Gilbert Peak and Tieton Peak. Based on their elevation compared to the current glaciers, the moraines are interpreted as deposits of the McNeeley 1 and McNeeley 2 advances, respectively.

Discussion of glacier advances in the Goat Rocks

The ice limit during the maximum advance of the last glaciation (Evans Creek?) has been mapped near the east end of Rimrock Lake, several miles downstream from the Hell Creek moraine (Schasse, 1987a). The moraine at Hell Creek was deposited by a later readvance or recessional stillstand. Based on the ^{36}Cl ages on boulders from moraines in the North Cascades that show an advance contemporaneous with the Vashon advance of the

Cordilleran Ice Sheet (Swanson and Porter, 1997), and an earlier one likely of Evans Creek age, the moraine at Hell Creek tentatively is correlated with the Vashon advance.

Two McNeeley advances deposited moraines in the valleys of both the North and the South Fork of the Tieton River, indicating that these advances were not limited to the immediate vicinity of Mount Rainier.

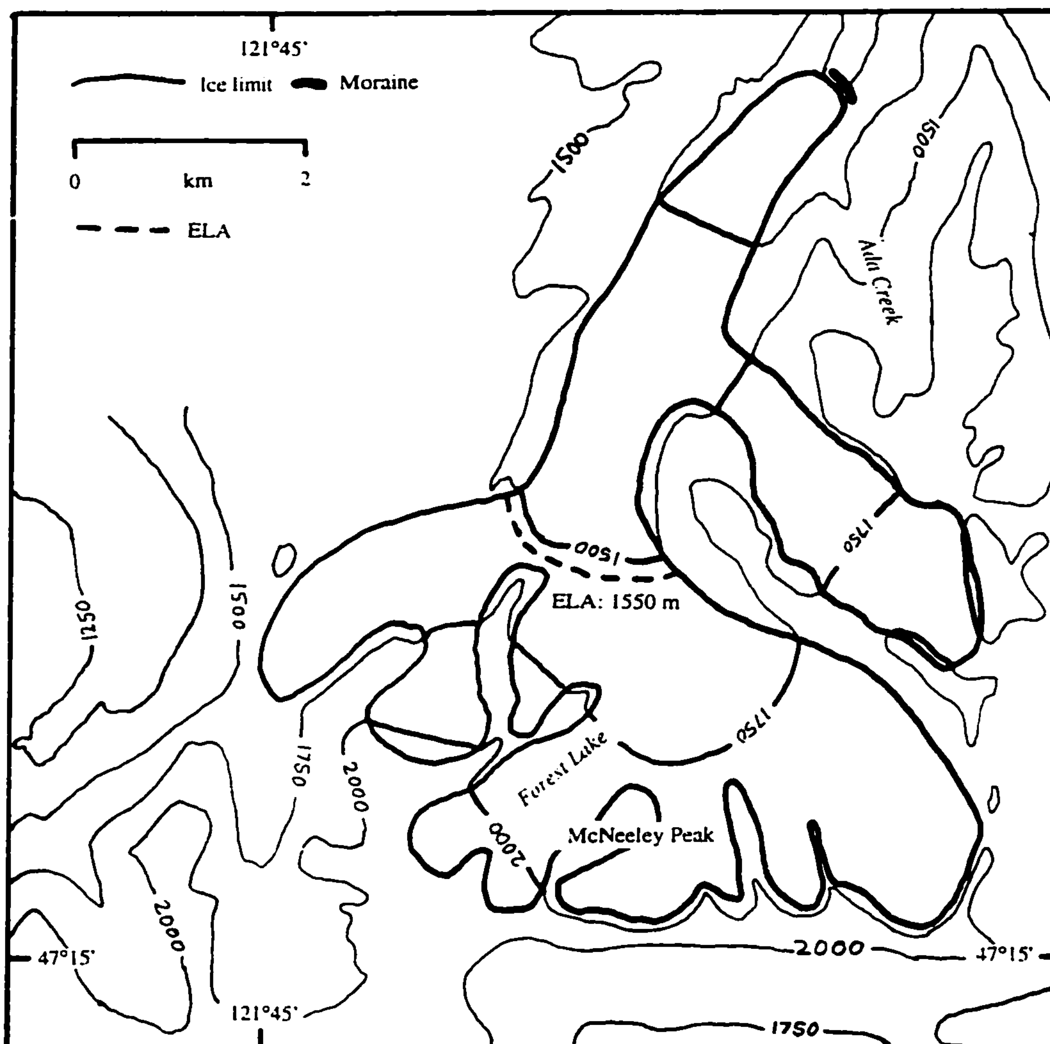


Figure 12: Glacier extent in the Huckleberry valley duringthe last glaciation. The age of the glacial deposits has not been determined. Contour interval: 250 m.

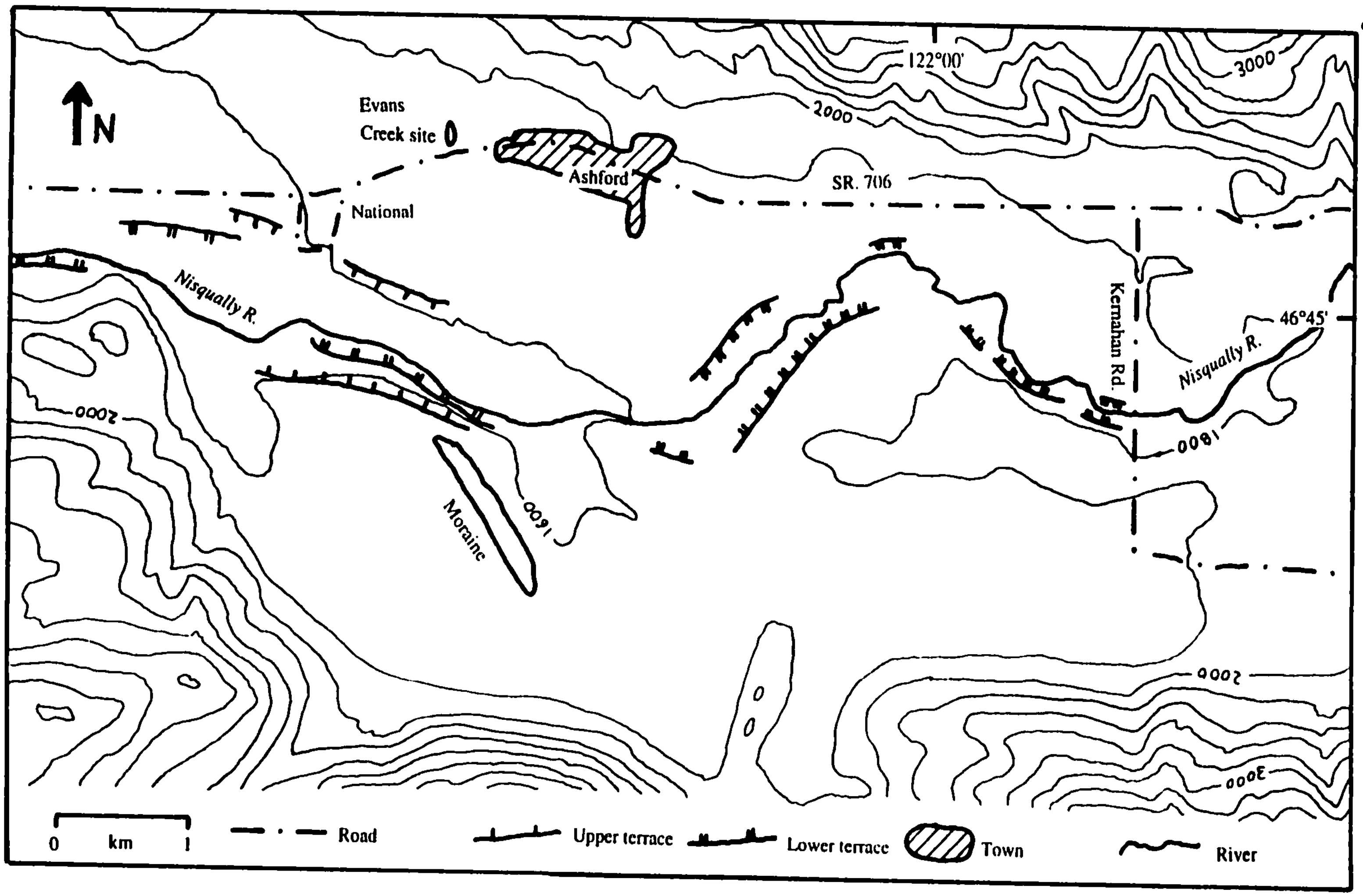


Figure 13: Trace of terrace margins in the Nisqually River valley. Tick marks point toward lower (younger) terrace surface. Contour interval: 200 ft (61 m).

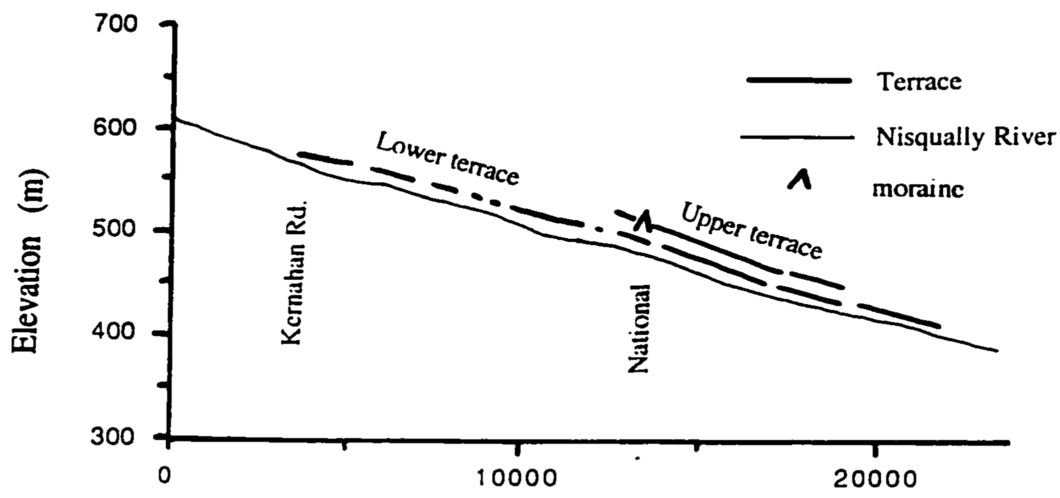


Figure 14: Long-profiles of outwash terraces in the Nisqually River valley. Dashed lines indicate that terraces have eroded. Terraces are blanketed by coarse sandy lahar deposits. The mapped surface is the top surface of the landform, not the top surface of the outwash. The thickness of the lahar is ca. 60-120 cm in all exposures.

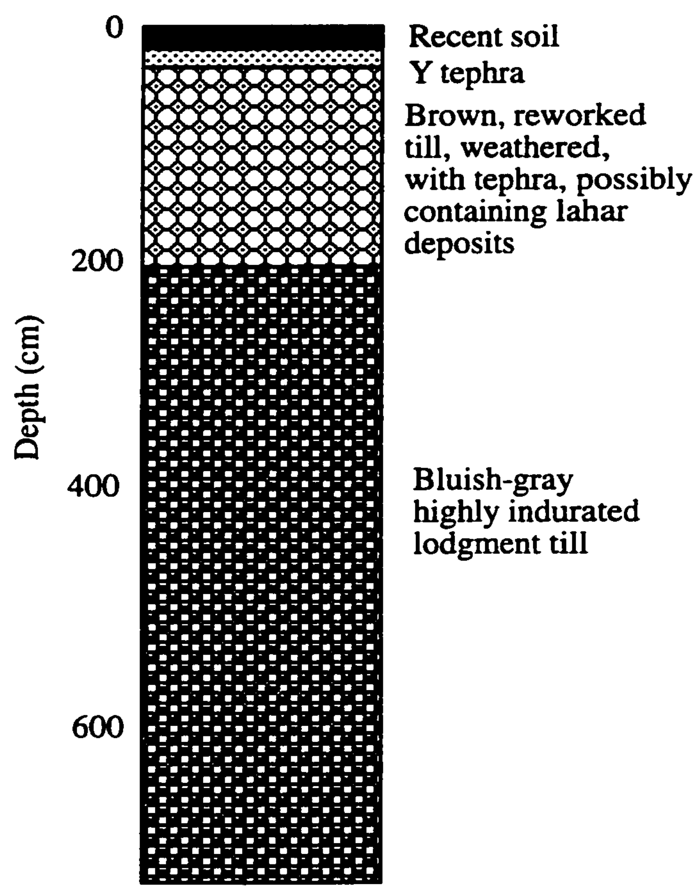


Figure 15: Exposure 2 at Evans Creek type moraine. The induration of the lowest unit increases with depth.

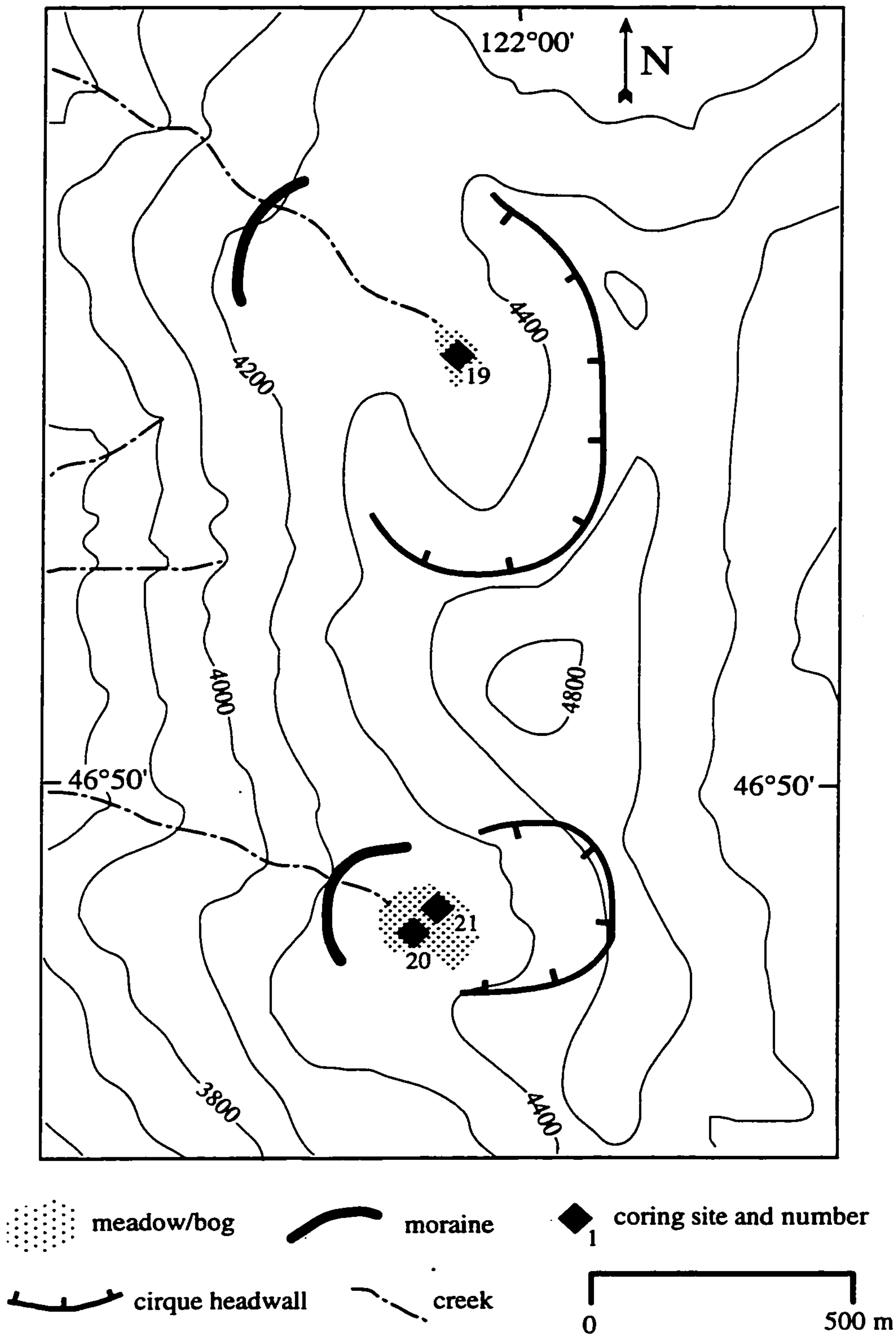


Figure 16: Map of glacial landforms at Lake Mary Lea. Numbers on contour lines are elevation in feet.

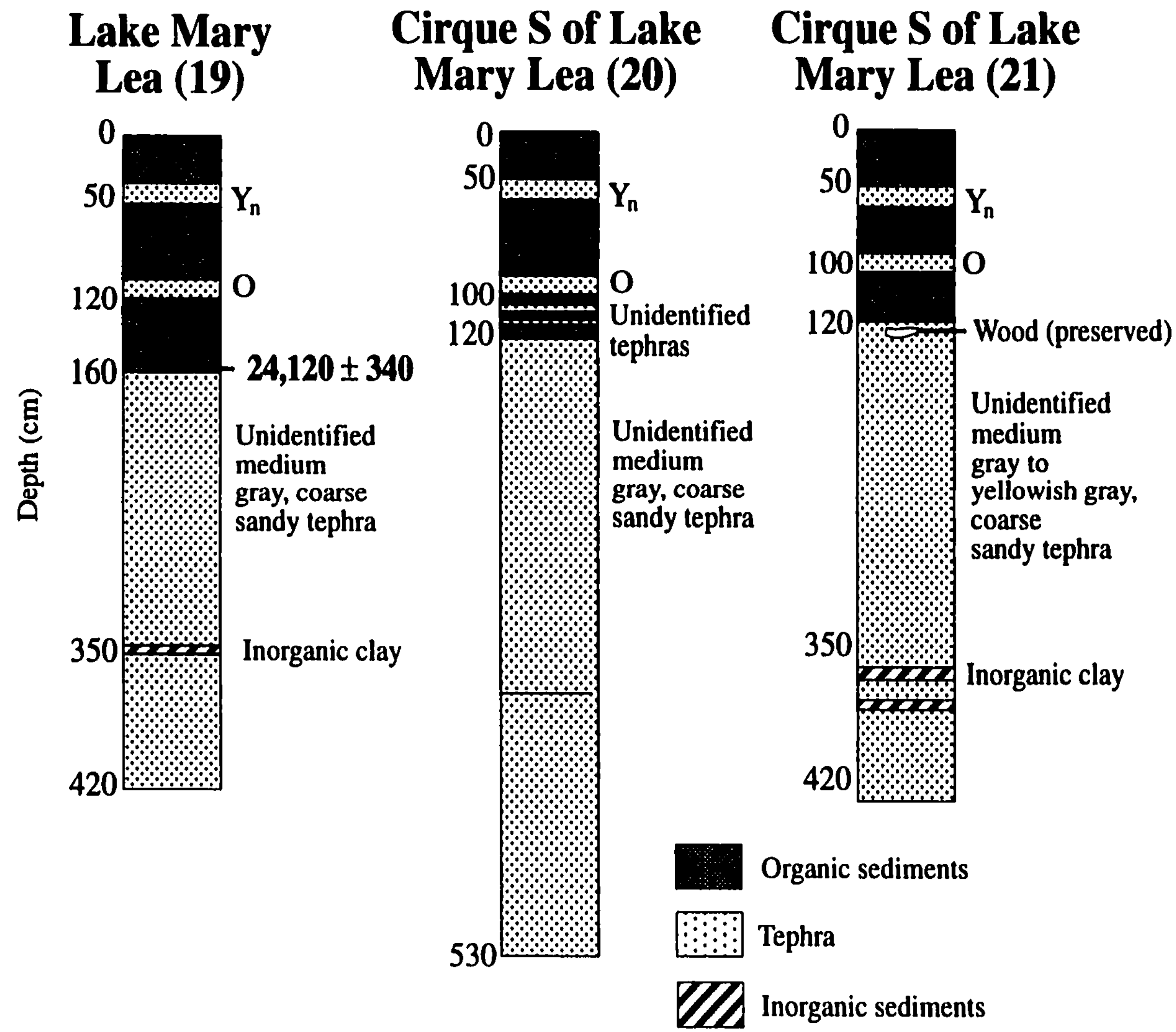


Figure 17: Cores at Lake Mary Lea and an adjacent cirque. Letters designate tephra layers. Radiocarbon age in bold.

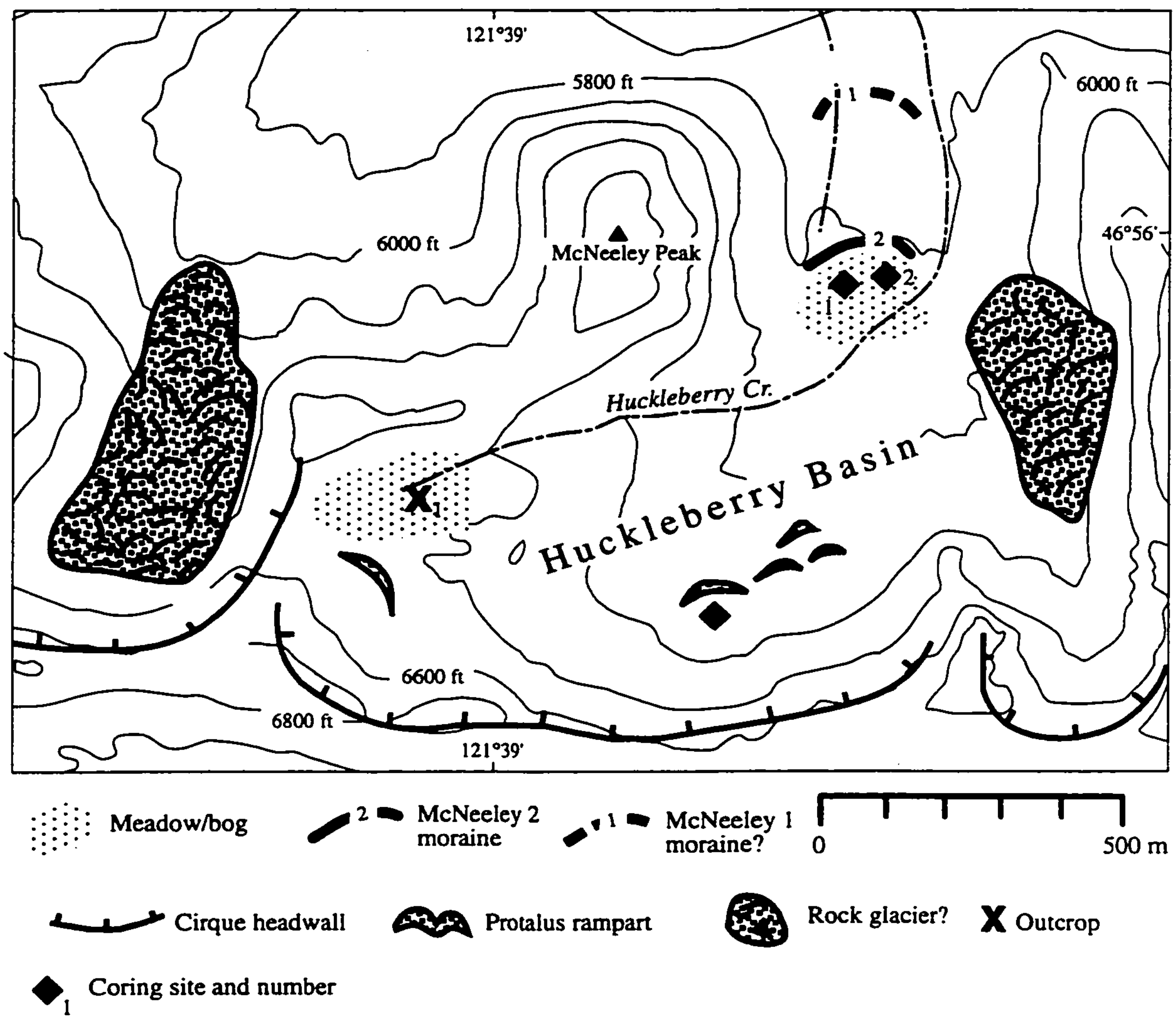


Fig. 18: Map of glacial landforms at Huckleberry Basin.

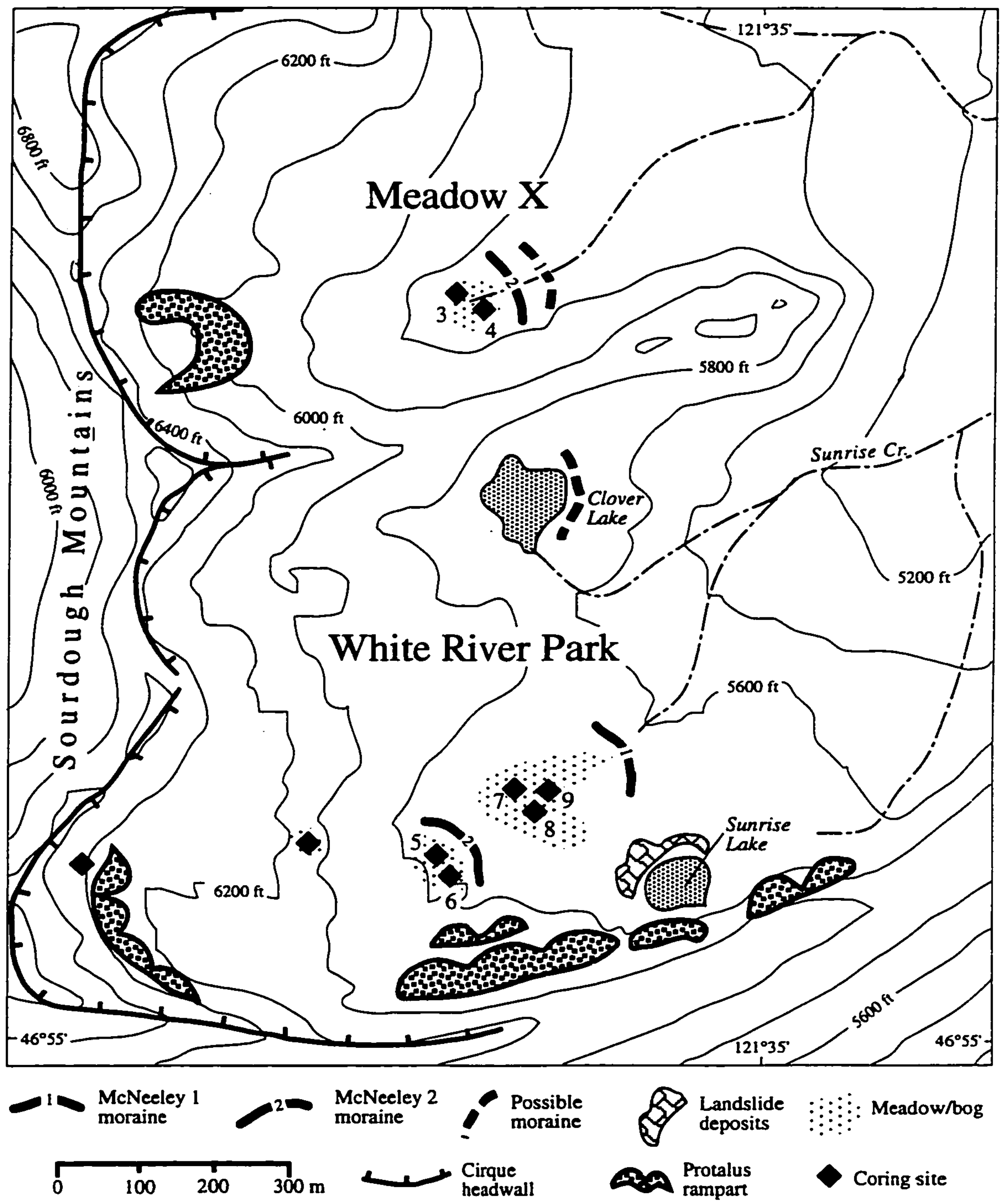


Figure 19: Map of glacial landforms at White River Park. Numbers on contour lines are elevations in feet. The ridge damming Clover Lake may be a bedrock ridge mantled with till (see text).

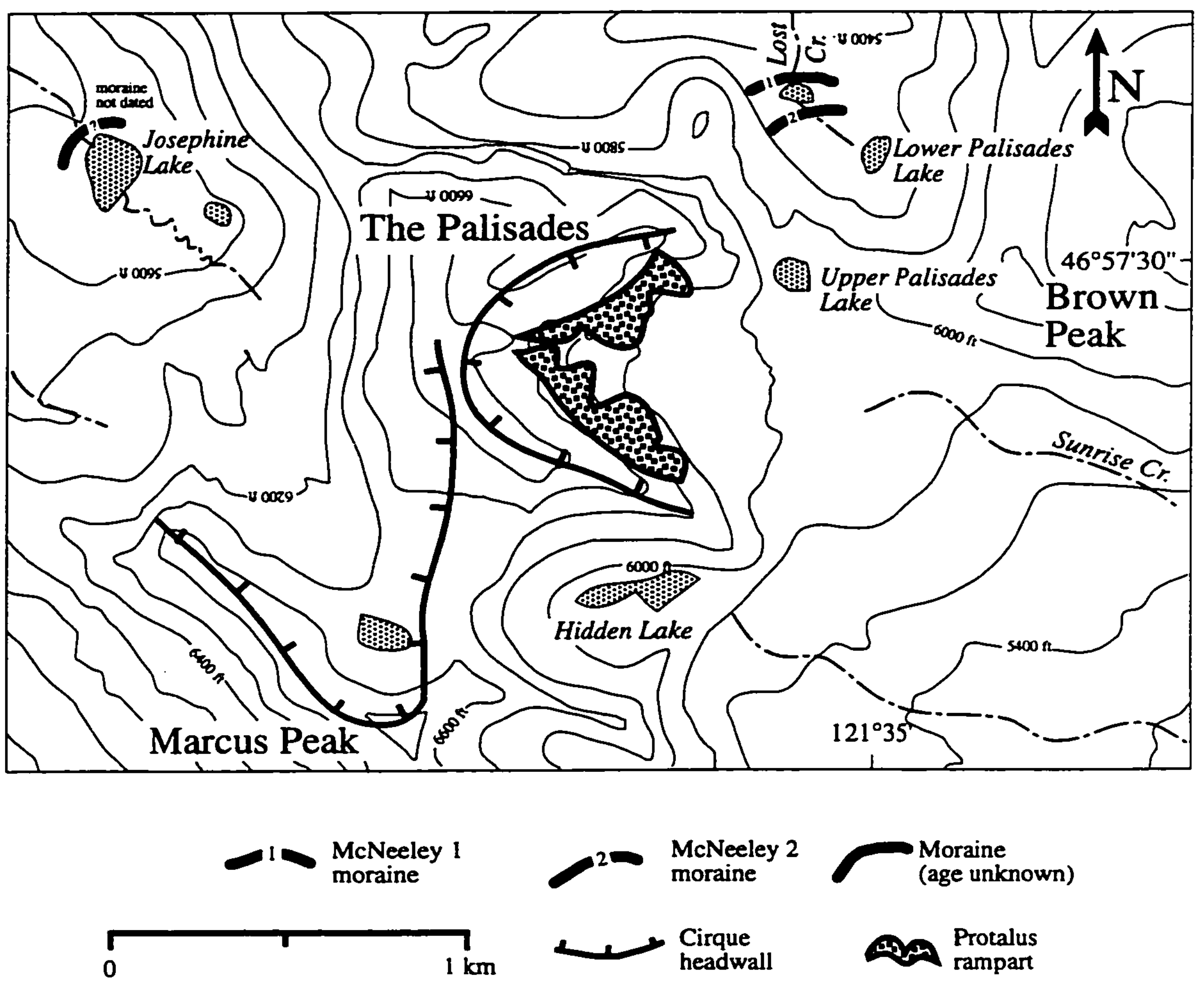


Figure 20: Map of glacial landforms at the Palisades.

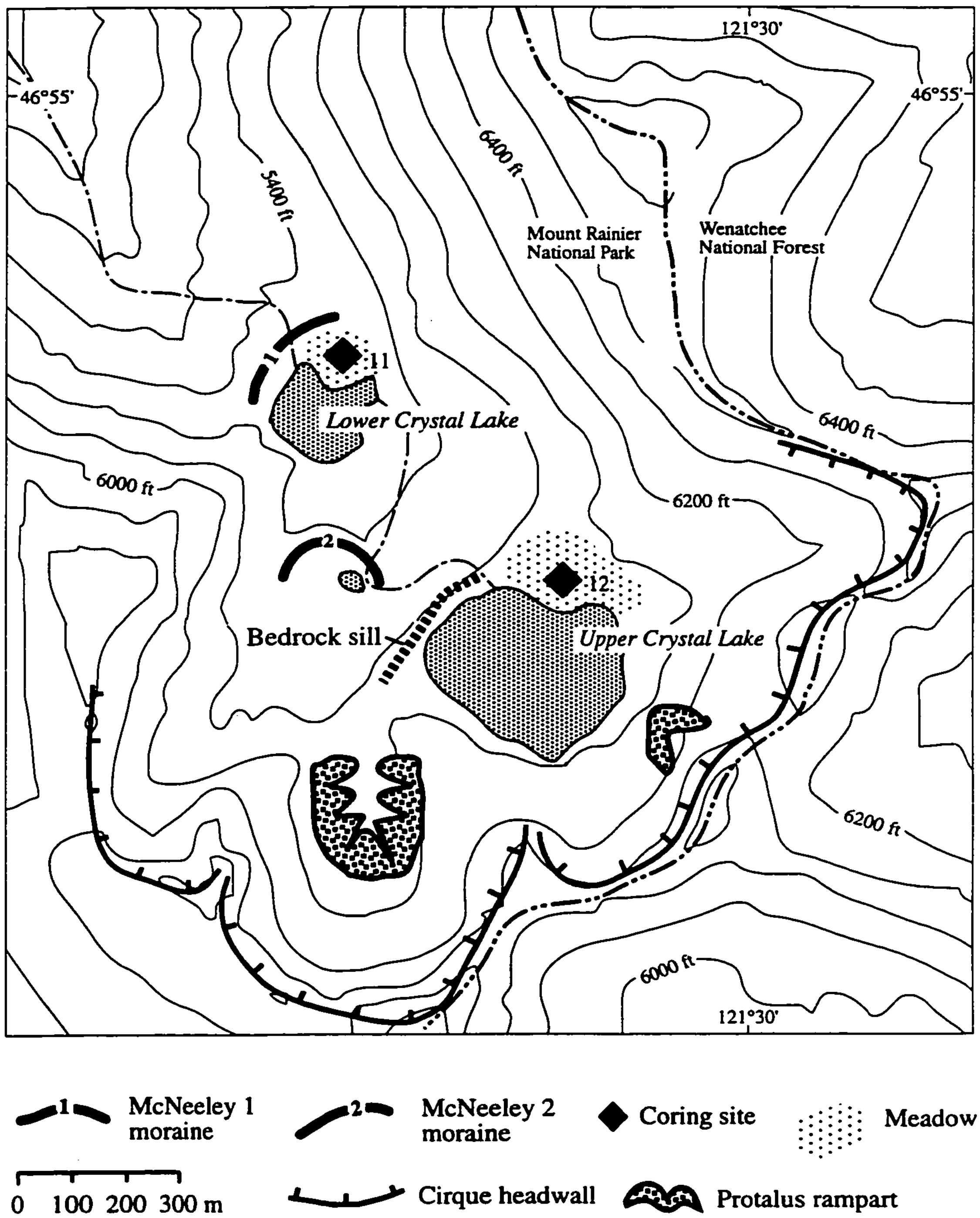


Figure 21: Map of glacial landforms at Crystal lakes.

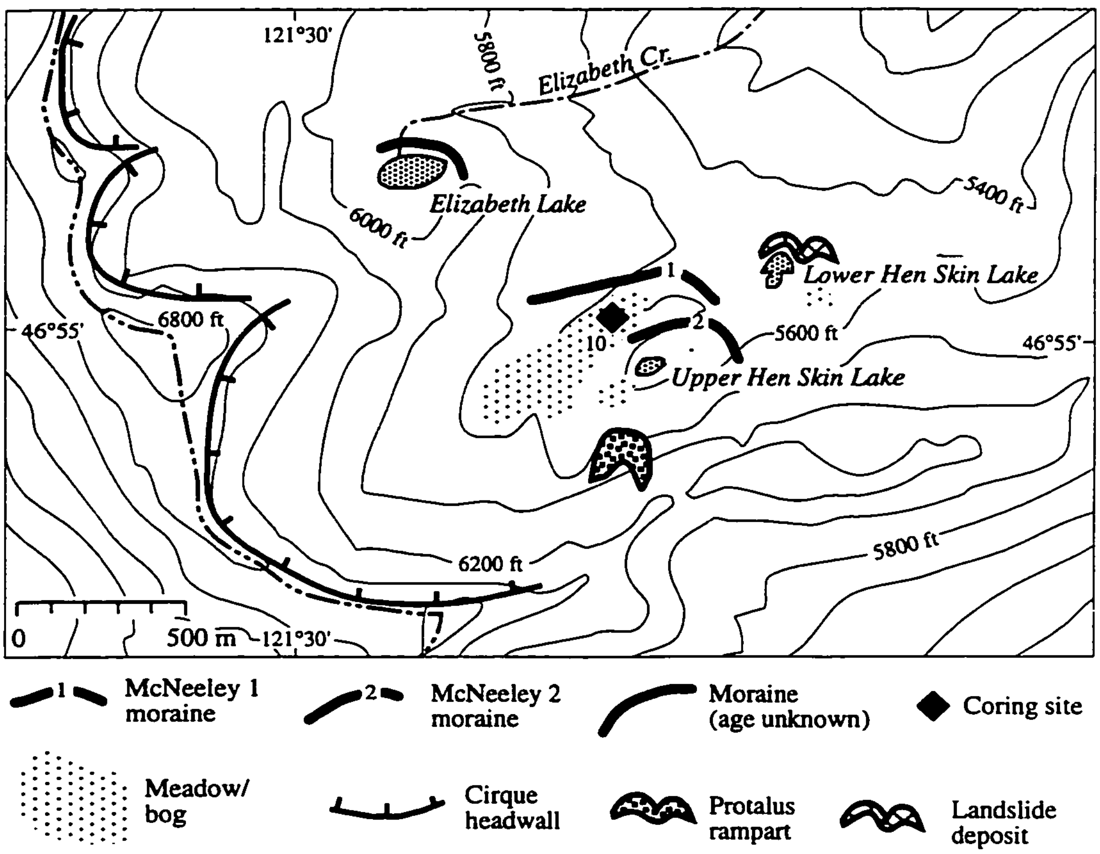


Figure 22: Map of glacial landforms at Crystal Mountain.

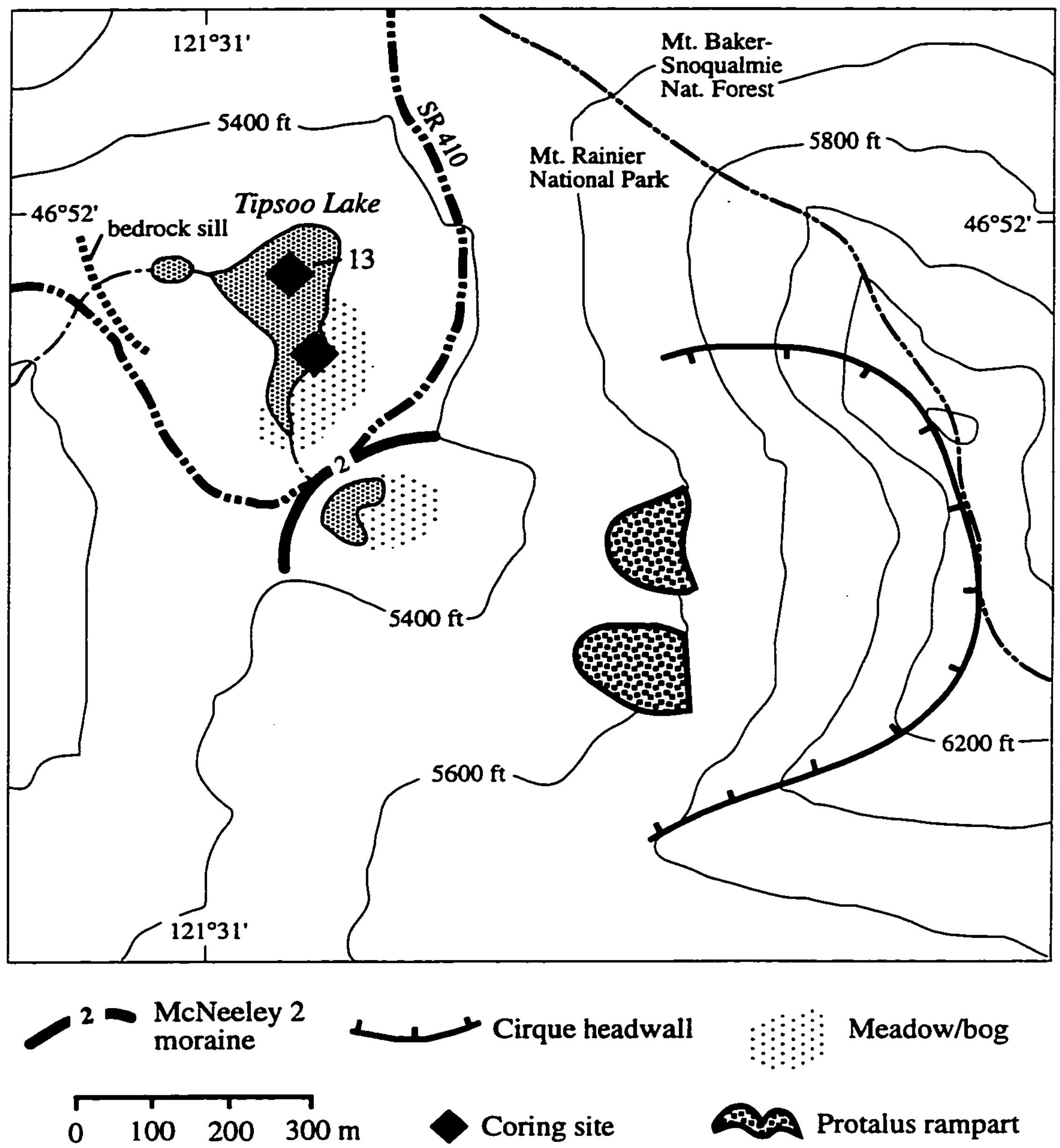


Figure 23: Map of glacial landforms at Tipsoo Lake.

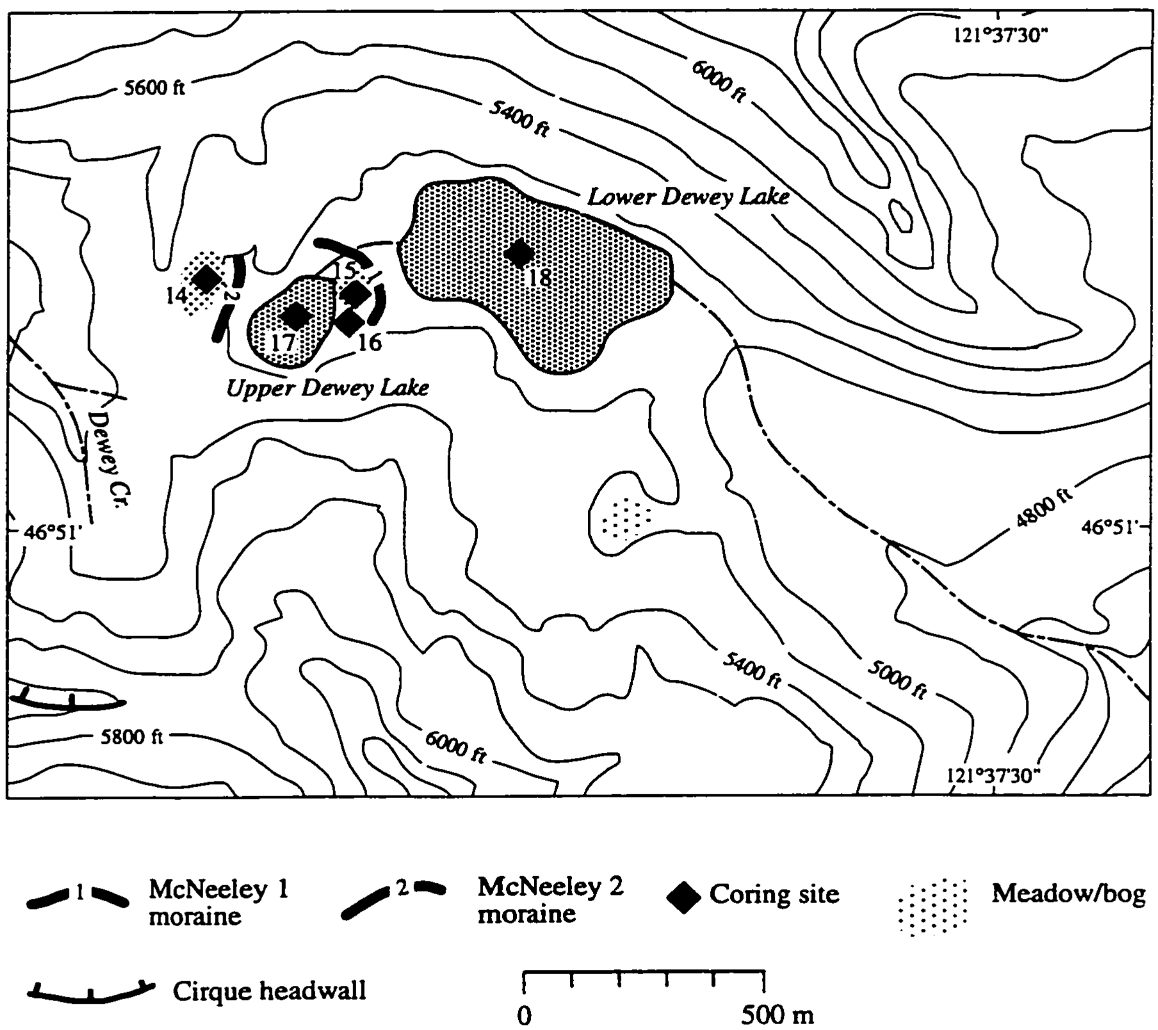


Figure 24: Map of glacial landforms at Dewey Lakes.

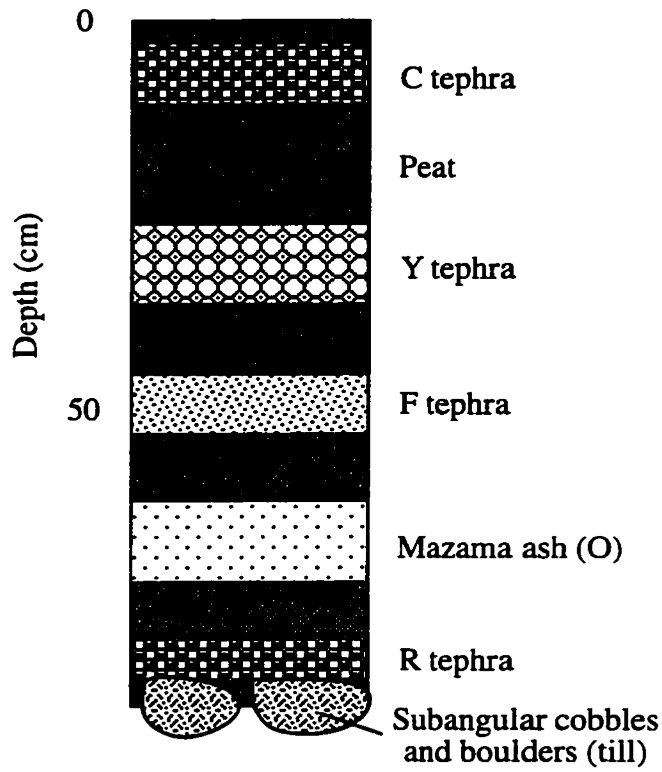


Figure 25: Stratigraphy at exposure 1, Huckleberry Park (Figure 17).

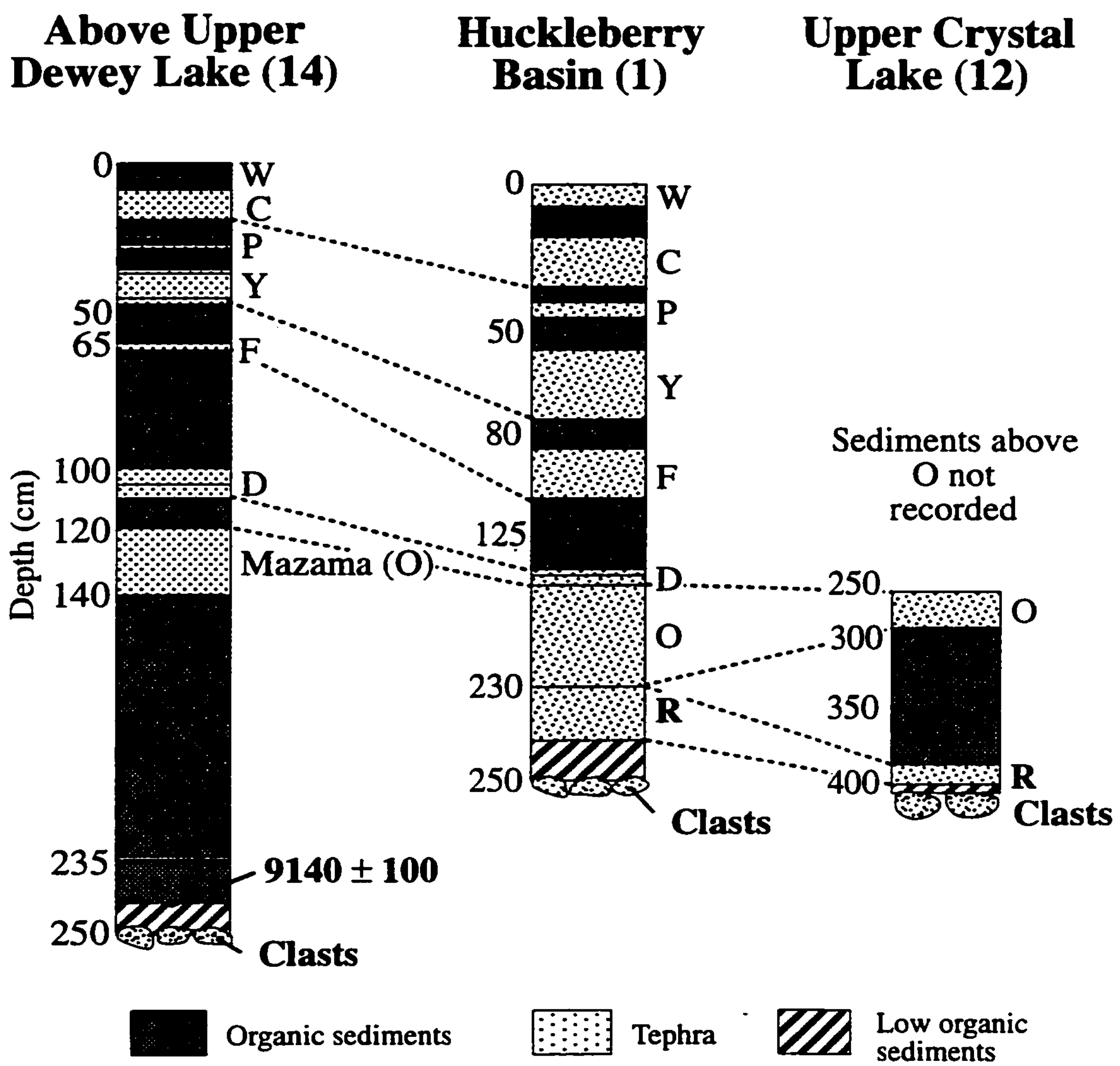


Figure 26: Cores taken behind McNeeley 2 moraines. Radiocarbon ages in bold type. Letters designate tephra layers. Numbers in bold type are radiocarbon ages.

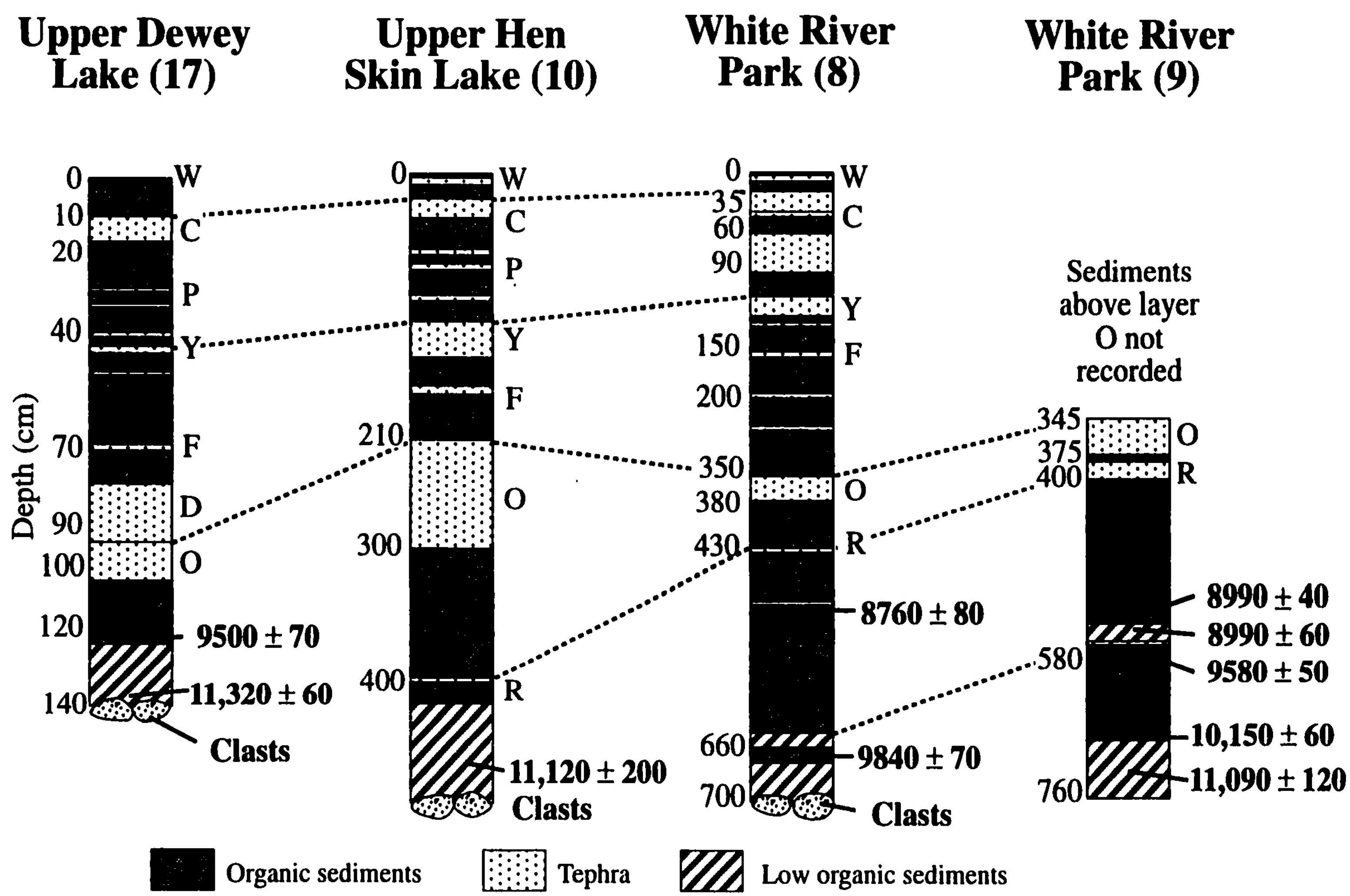


Figure 27: Cores taken behind McNeeley 1 moraines. Capital letters designate tephra layers. Numbers in bold type are radiocarbon ages.

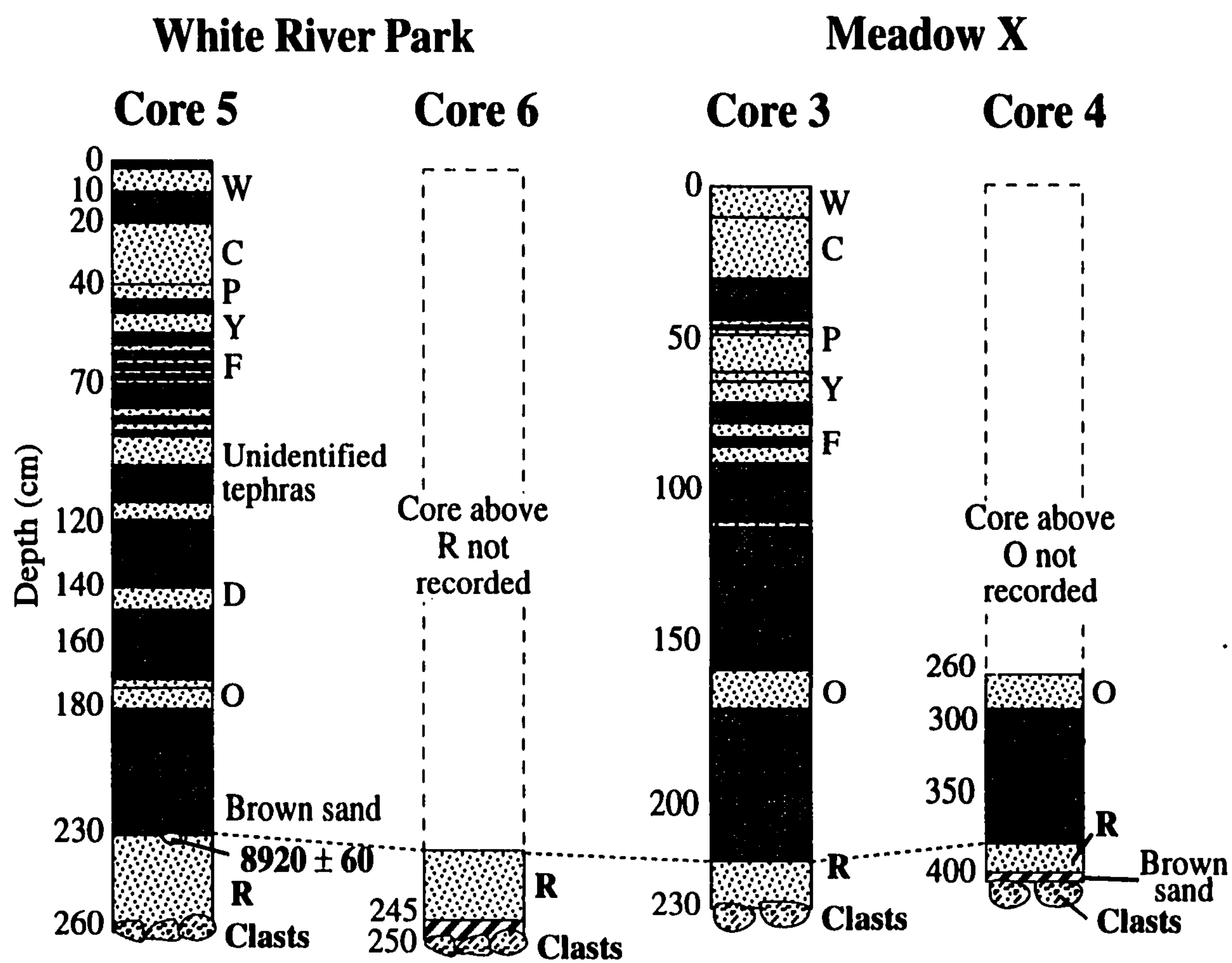


Figure 28: Cores taken behind McNeeley 2 moraines. Letters designate tephra layers. Bold numbers are radiocarbon ages. For legends, see Figure 27.

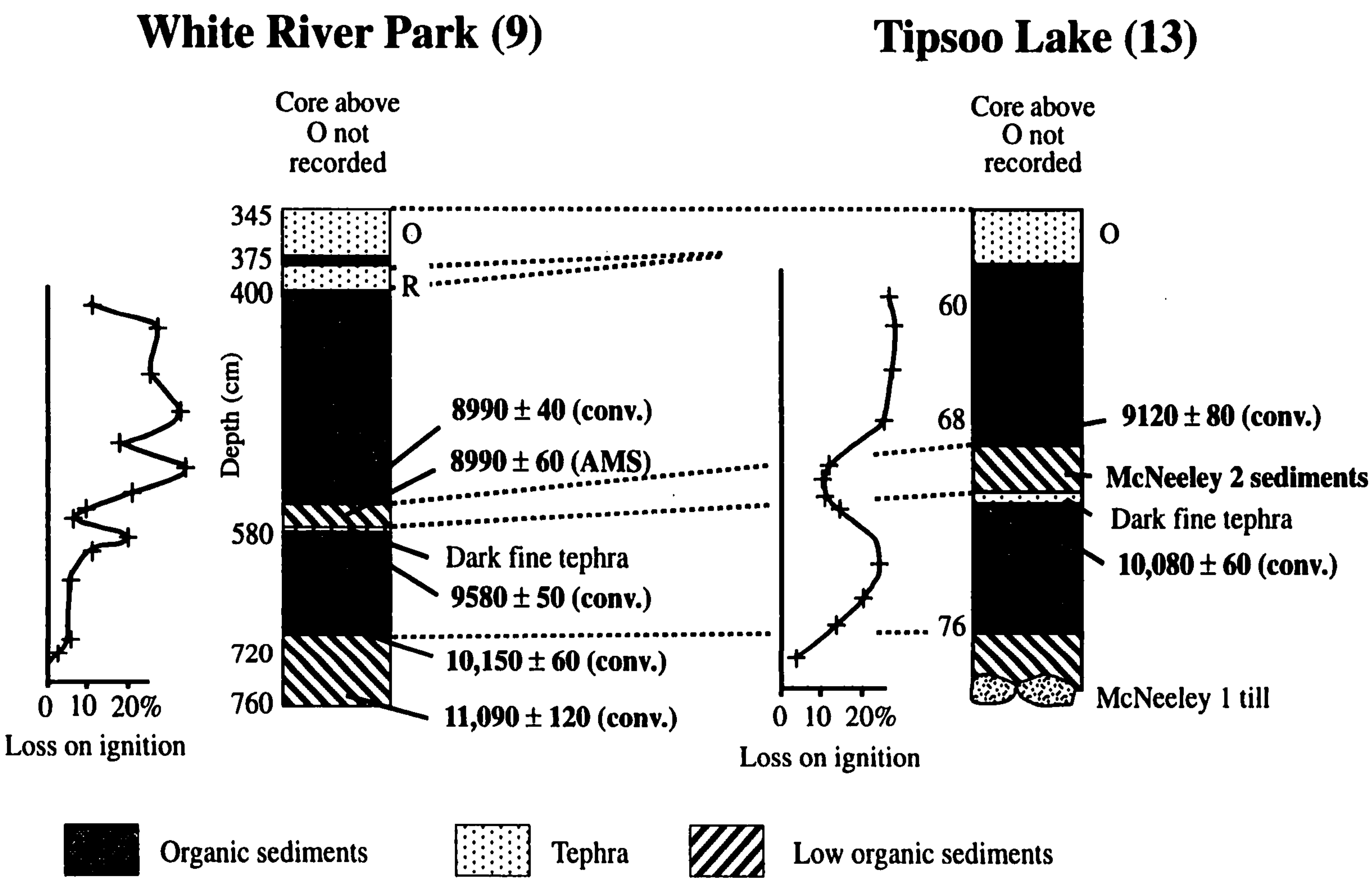


Figure 29: Cores taken downstream from McNeeley 2 moraines. The upper layer of "low organic sediments is interpreted as evidence for the McNeeley 2 glacier advance upstream (see text). Letters designate tephra layers. Bold numbers are radiocarbon ages. "Conv." designates conventional ages, "AMS" designates AMS ages.

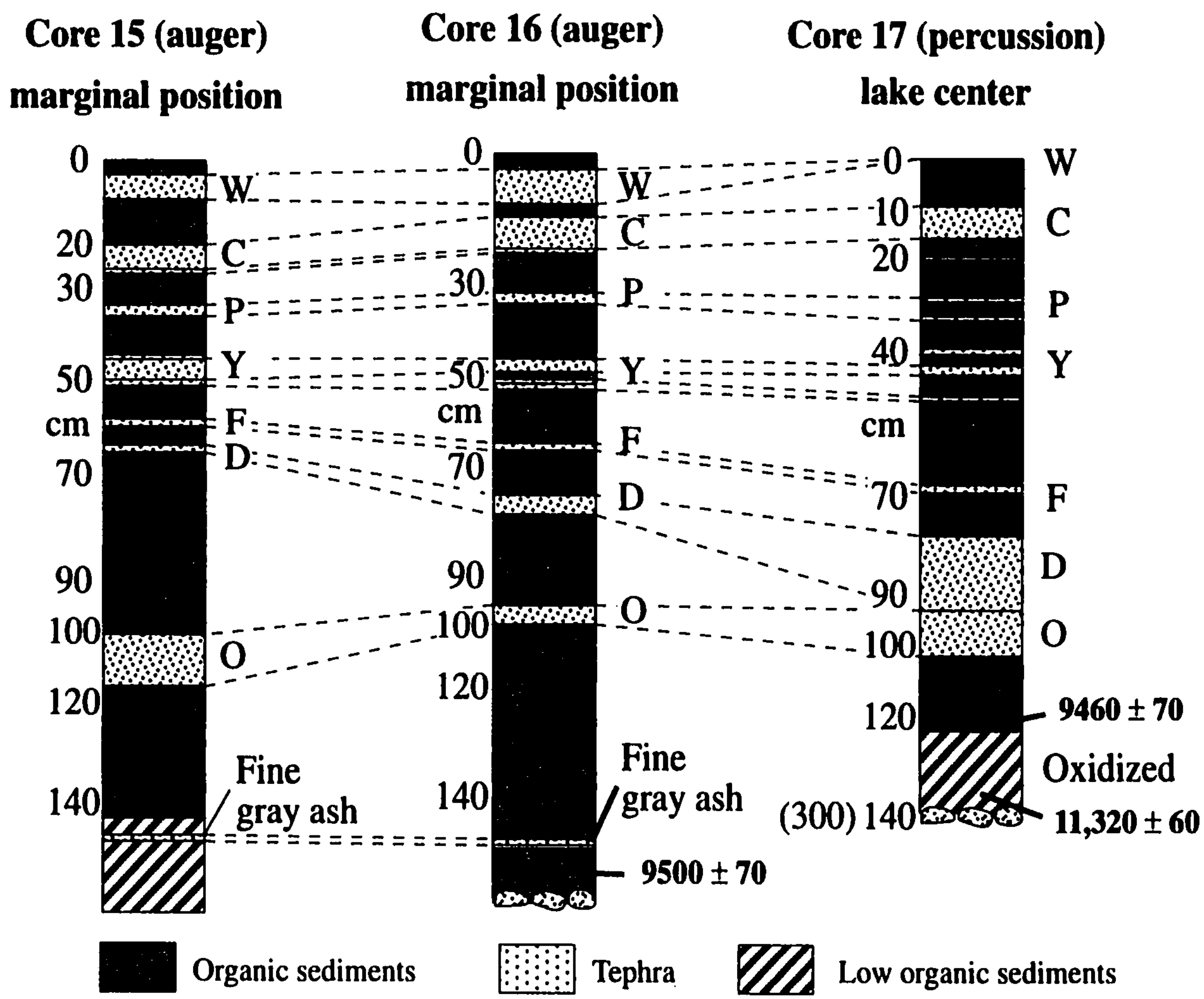


Figure 30: Cores from Upper Dewey Lake. Upper Dewey Lake is dammed by a McNeeley 1 moraine. Capital letters indicate tephra layers. Bold numbers are radiocarbon ages. Core 17 experienced core loss in the upper layers (see text).

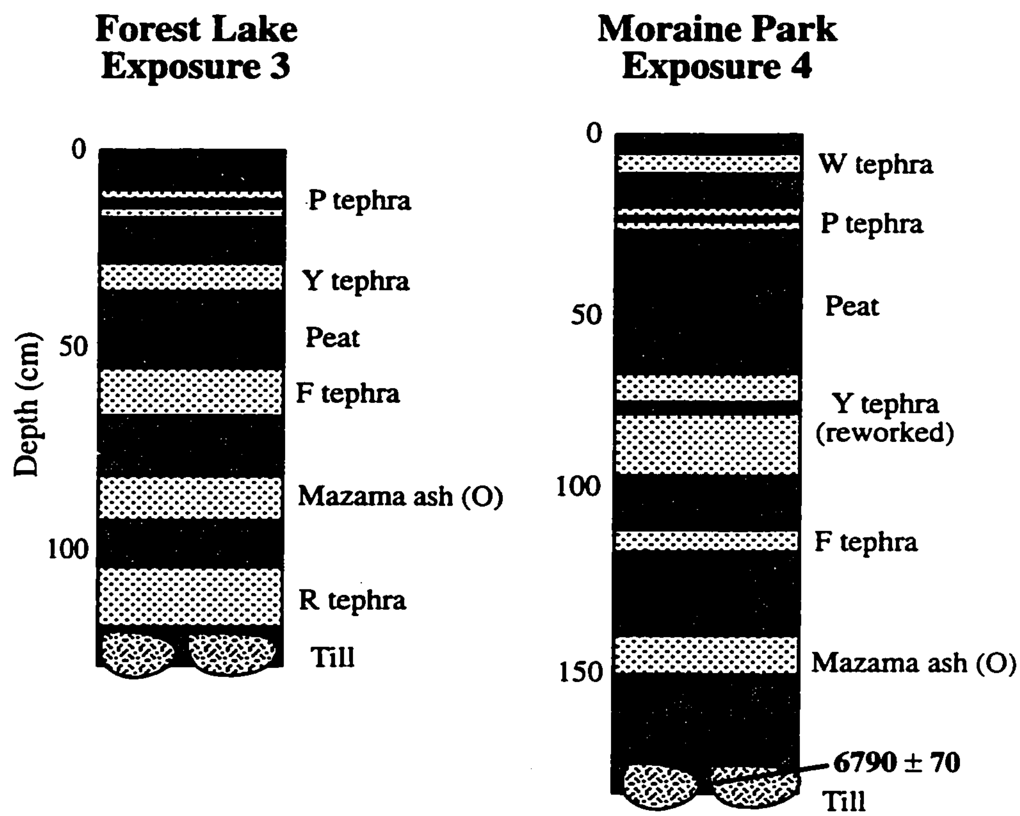


Figure 31: Exposures at Forest Lake and Moraine Park. These sites were not used for paleoclimatic reconstruction. For explanation of the sedimentary units, refer to Figure 29. Bold numbers are radiocarbon ages.

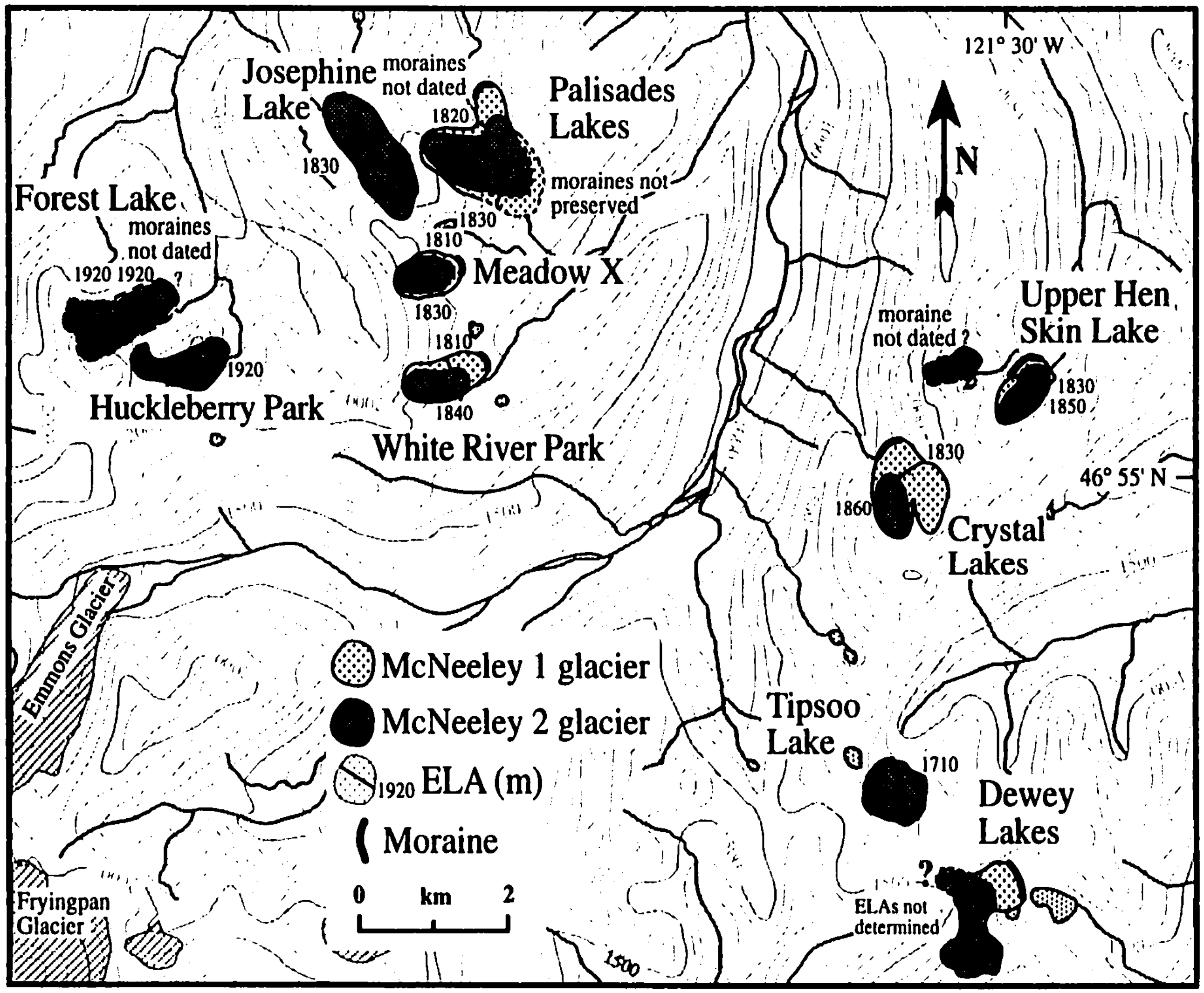


Figure 32: Reconstruction of McNeeley glaciers in the main study area. Other glaciers may have existed in the area, but are not shown.

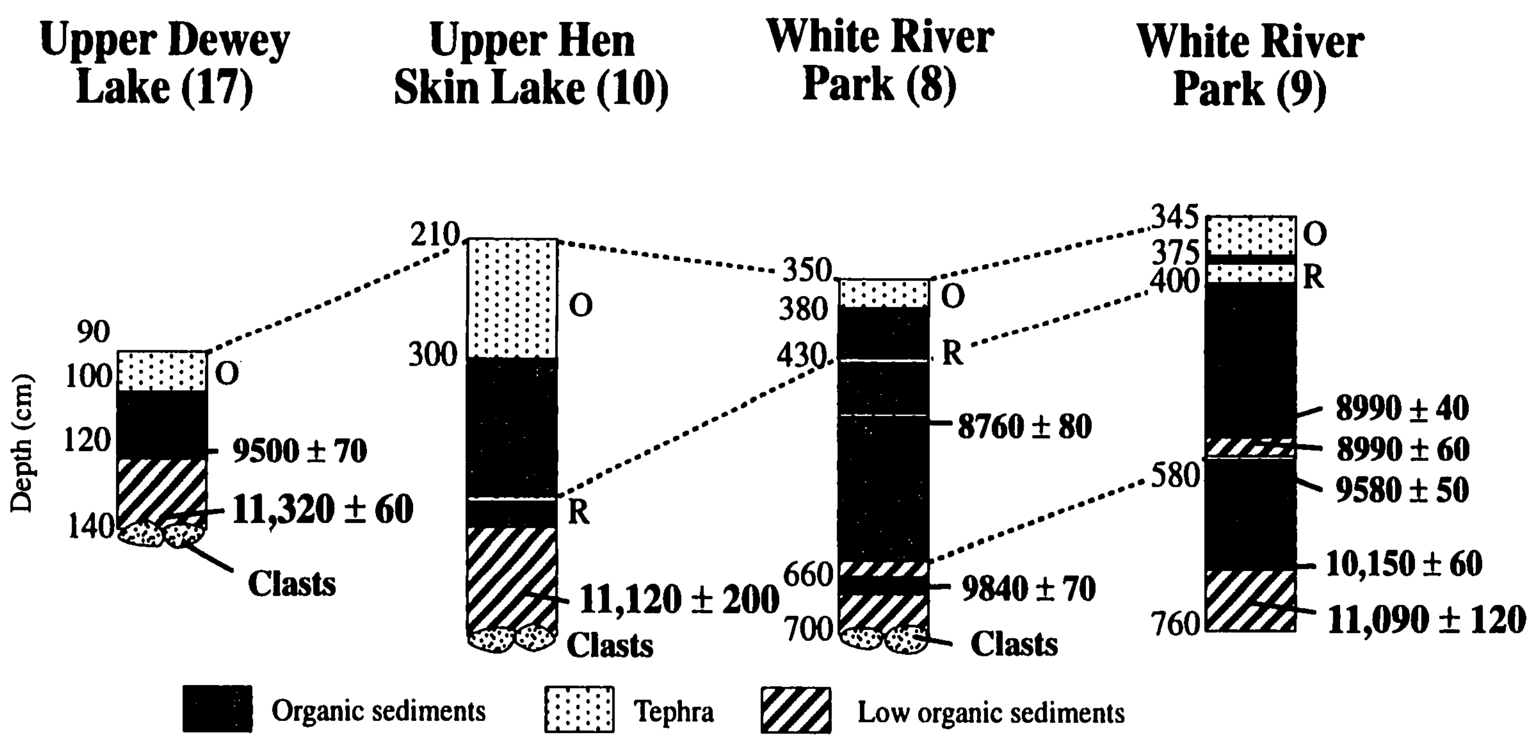


Figure 33: Sub-Mazama stratigraphy of cores taken behind McNeeley 1 moraines. Capital letters designate tephra layers. Bold numbers are radiocarbon ages.

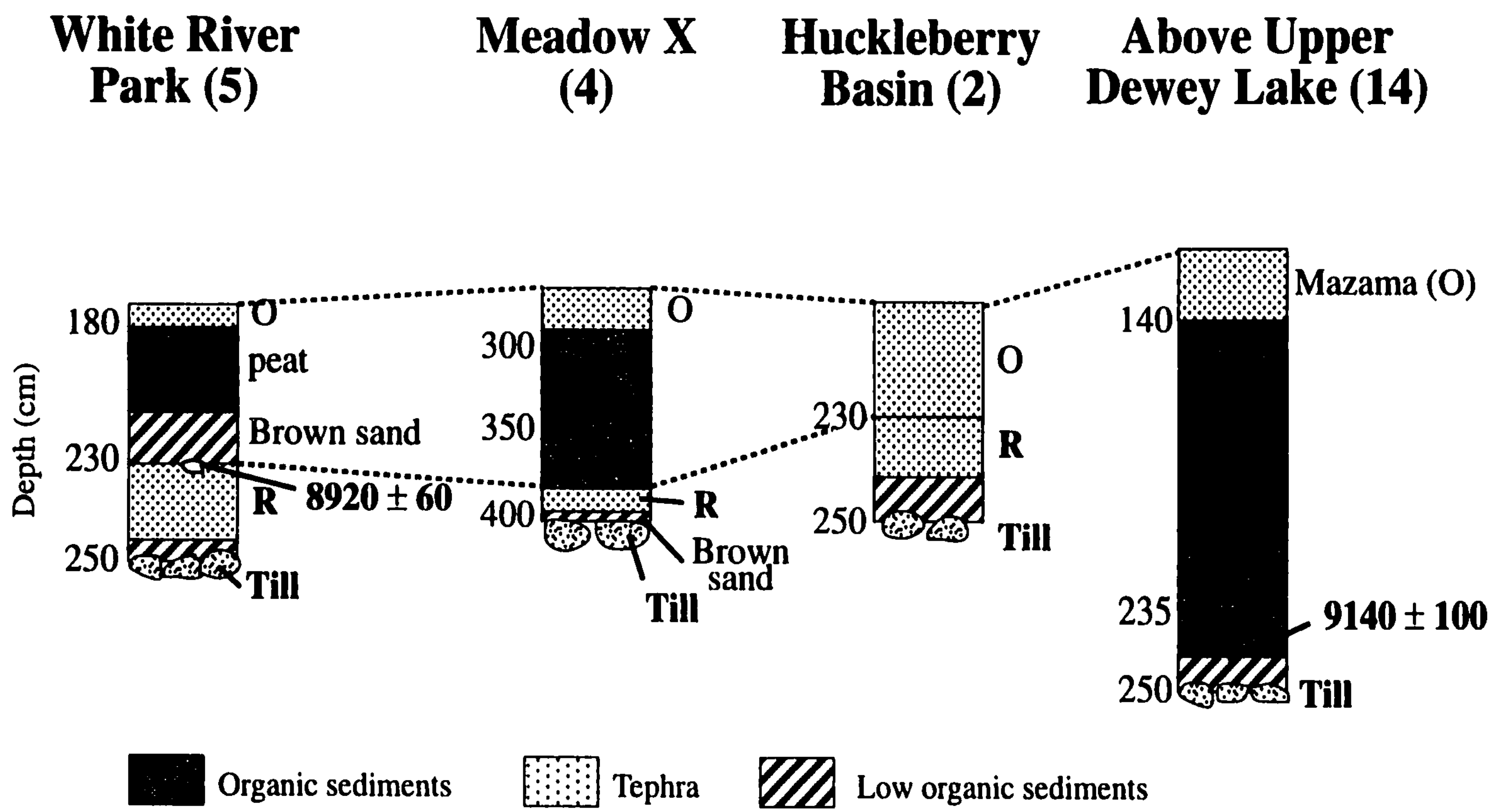


Figure 34: Sub-Mazama stratigraphy of cores taken behind McNeeley 2 moraines. Capital letters designate tephra layers. Bold numbers are radiocarbon ages.

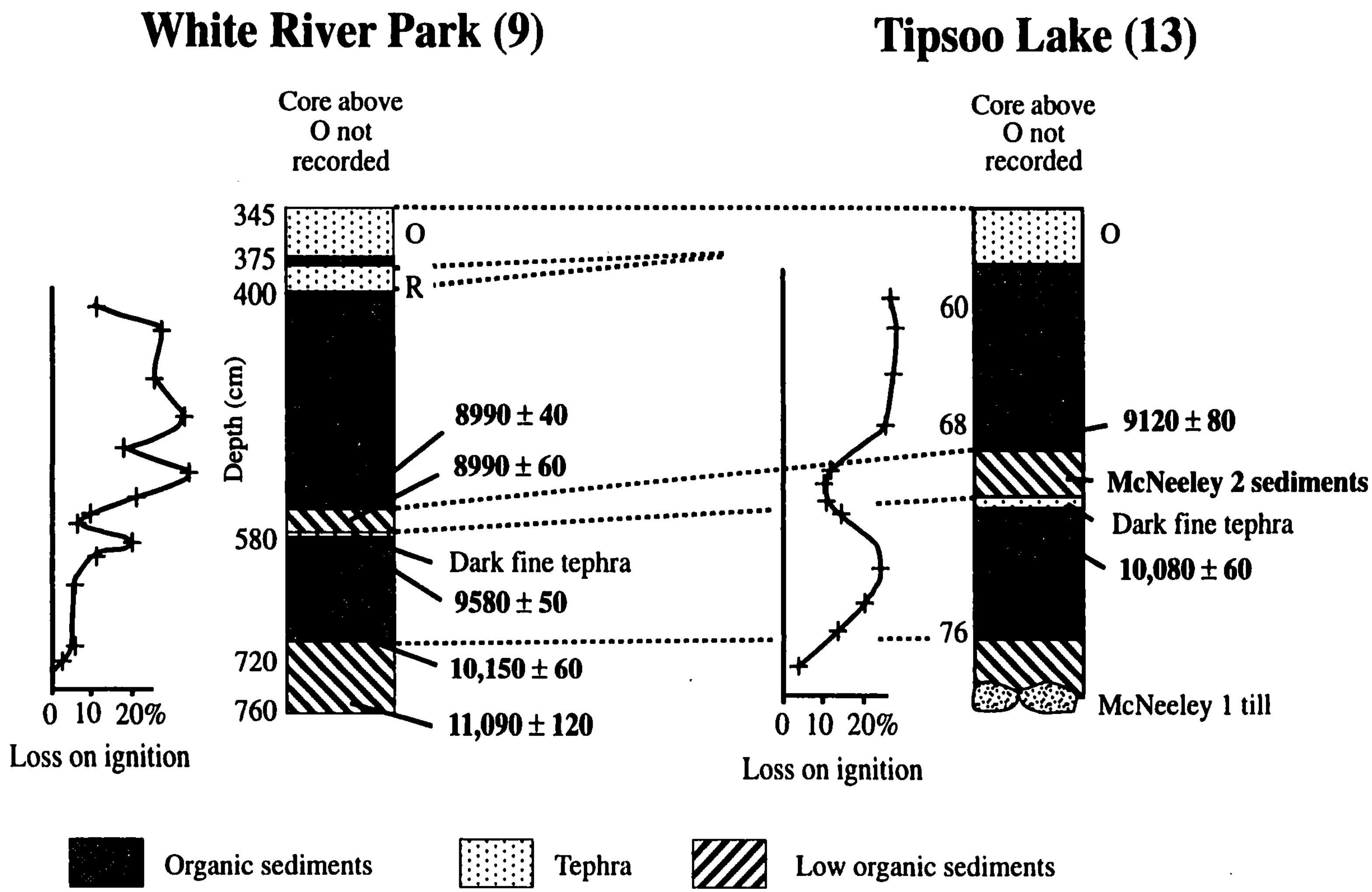


Figure 35: Cores taken downstream from McNeeley 2 moraines. The upper layer of "low organic sediments is interpreted as evidence for the McNeeley 2 glacier advance upstream (see text). Letters designate tephra layers. Bold numbers are radiocarbon ages.

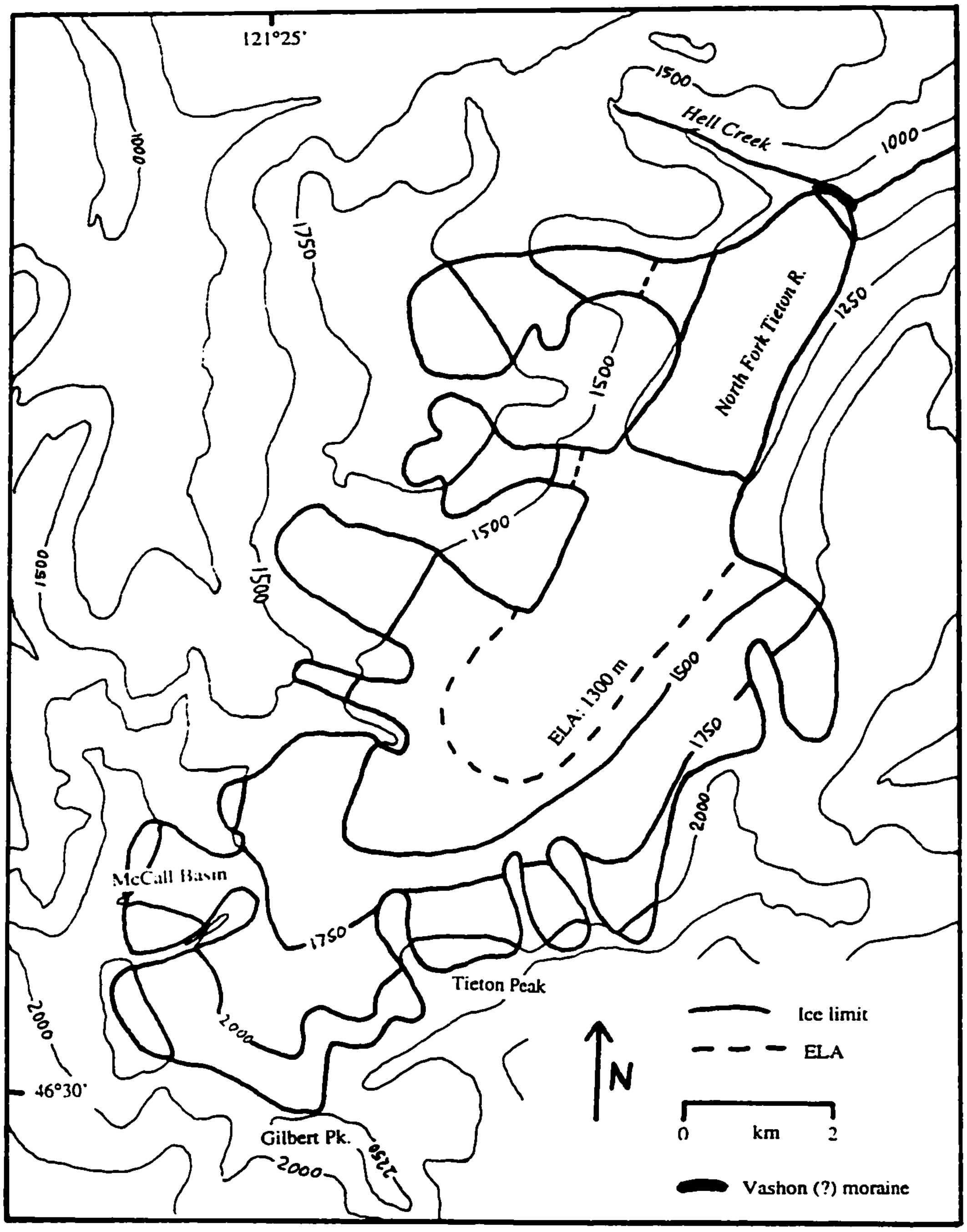


Figure 36: Map of full-glacial deposits and glacier extent in the North Fork of the Tieton River valley (Goat Rocks) during oxygen isotope stage 2. The age of the glacial deposits has not been determined. Contour interval 250 m. Other glaciers may have existed in the area, but are not shown.

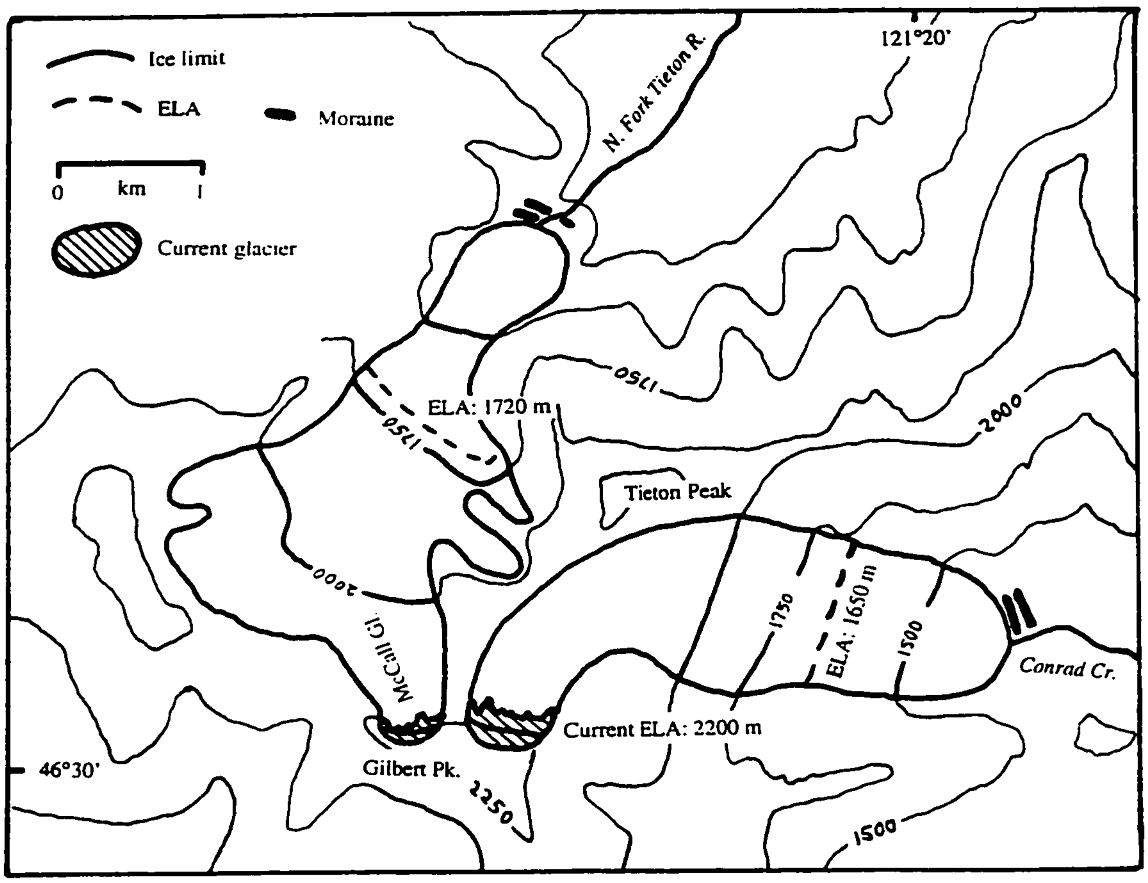


Figure 37: Glacier extent in the Goat Rocks during late-glacial time, and at present. Contour interval 250 m. Other glaciers may have existed in the area, but are not shown.

7 EQUILIBRIUM LINE ALTITUDES

7.1 PREVIOUS STUDIES IN THE CASCADE RANGE

Equilibrium line altitudes (ELAs) have been reconstructed for the glacial sequence in the Yakima valley ca. 10 km northeast of Mount Rainier (Porter, 1976). The oldest moraine group, named Bullfrog, was associated with an ELA lowering of at least 850 m compared to today's (Porter, 1976; Porter et al., 1983). The Domerie moraines are associated with an ELA depression of 750 m (Porter, 1976; Porter et al., 1983). The ELA depression during the Hyak advances was ca. 600 m (Porter et al., 1983; Hurley, 1996). Since the above ELA values do not take into consideration glacio-isostatic effects associated with the Cordilleran Ice Sheet, the ELA depressions of the Bullfrog and Domerie moraines may have been greater than the moraines suggest (Porter, 1976). In the Oregon Cascades, Scott (1977) found an ELA depression of 950 m for the maximum advance of the last glaciation. Porter (1977) suggested an ELA depression of ca. 950 m for this advance throughout the Cascade Range.

7.2 MOUNT RAINIER

PRESENT

The 4392-m-high cone of Mount Rainier creates a significant rainshadow effect. Current annual precipitation in the study area ranges between ca. 1800 and more than 2400 mm (E. Stein, unpubl. data). This results in significant ELA variations within the study area. The ELAs on Mount Rainier range between 2330 and 2500 m (Fig. 38), based on mapping of the late-summer snowline in three consecutive years (Chapter 5.4). ELAs were projected into the study area based on the mapped ELAs (Fig. 38) and current precipitation values. However, considering the steep gradient of ELA values within the study area, the extrapolated values may be unreliable.

FULL-GLACIAL TIME

Nisqually valley

The ELA associated with the inferred Evans Creek advance in the Nisqually River valley was ca. 920 m (Fig. 39). The ELA of the inferred Vashon advance was ca. 1000 m (Fig. 40). Extrapolating the current ELAs (Fig. 38) toward the location of the Evans Creek and Vashon ELAs, respectively, results in ELA depressions of ca. 1080 and 1000 m, respectively. However, Nisqually Glacier is located in an area of large variations in precipitation on a small scale. This was even more pronounced at times when the glacier extended further downvalley and thus further from the orographic effect of Mount Rainier, which presumably increases precipitation on this side of the edifice. While the accumulation area ratio method to determine ELAs is an empirical method and has been shown to work well in various mountain glacier environments (S. Porter, pers. comm., 1997), the steep precipitation gradient may result in errors for the reconstructed ELA.

Lake Mary Lea

The reconstructed ELA for the glaciers at Lake Mary Lea was ca. 1340 m, and 1325 m in the adjacent cirque (Fig. 41). If the moraines were deposited by the Evans Creek advance, the ELAs seem high compared to those in the Nisqually valley. However, the cirques are oriented to the west, and are very shallow. The glaciers were exposed to the sun throughout most of the day, resulting in greater ablation and higher ELAs. The current ELA values at Lake Mary Lea are too unreliable to calculate meaningful ELA depressions.

Huckleberry Creek

Assuming the ridge at Ada Creek is a moraine, the reconstructed ELA for the Huckleberry Creek Glacier is ca. 1550 m (Fig. 12). Since the age of the moraine is not closely constrained, it remains unknown whether this ELA reflects the Evans Creek or the Vashon advance. The current ELA values (Fig. 38) at Huckleberry Creek are too unreliable to calculate meaningful ELA depressions.

McNeeley glaciers

McNeeley 1 advance

The reconstructed ELAs for the McNeeley 1 advance range from ca. 1700 and 1830 m. ELA values were 1820 m at Lower Palisades Lake (Fig. 42), 1810 m at Meadow X (Fig. 43) and at White River Park (Fig. 43), and 1830 m at Upper Hen Skin Lake (Fig. 45). At Crystal Lakes, the complex topography of the associated three glaciers (Fig. 46) renders reconstructions of the McNeeley 1 ELA difficult, but an ELA of ca. 1830 m seems likely. Reconstructing the glaciers at Dewey Lakes (Fig. 48) is difficult due to the complex topography. An exact calculation of the associated ELAs was not attempted, but the ELA probably was lower than 1700 m to enable the glaciers to deposit the moraines.

McNeeley 2 advance

The reconstructed ELAs for the McNeeley 2 advance range from 1710 to 1920 m. ELA values were 1920 m at Huckleberry Basin (Fig. 49), 1840 m at White River Park (Fig. 44), 1830 m at Meadow X (Fig. 44), at Josephine Lake (Fig. 42), and at Lower Palisades Lake (Fig. 42), 1850 m at Upper Hen Skin Lake (Fig. 45), 1860 m at Crystal Lakes (Fig. 47), and 1710 m at Tipsoo Lake (Fig. 50). ELA values at Dewey Lakes (Fig. 48) were not calculated (see above), but the ELA probably was lower than 1720 m to enable the glaciers to deposit the moraines.

Sites not used for paleoclimatic reconstruction

If the ridge damming Clover Lake is a moraine, the ELA for the associated glacier was ca. 1840 m (Fig. 44). If the interpretation of the moraines downstream from Forest Lake is correct, the associated ELA was ca. 1920 m for both McNeeley glaciers (Fig. 32). The ELA associated with the moraine at Berkeley Park was ca. 1920 m. The glaciers that deposited moraines at Moraine Park had an ELA of ca. 1770 m.

While these sites were not used for paleoclimatic reconstruction, the ELAs and dating at these sites are in good agreement with the findings of this study.

Discussion of McNeeley ELAs

The ELAs for the McNeeley 1 and McNeeley 2 advances are very similar (Fig. 32). In some drainages, the McNeeley 1 ELAs are up to 40 m lower, whereas in others the

McNeeley 2 advance may have overrun the McNeeley 1 moraines, indicating that the McNeeley 2 ELAs were lower than the McNeeley 1 ELAs in those cirques. In others, both McNeeley ELAs were very similar, and the McNeeley 2 moraines were deposited very close to the McNeeley 1 moraines.

ELAs vary slightly with cirque orientation. Larger variations in ELA are due to the location of the accumulation area of the glacier in the rainshadow of Mount Rainier. The lowest ELAs occurred at Tipsoo Lake and at Dewey Lakes, where moisture that was diverted by the edifice of Mount Rainier may have been funneled through the Chinook Pass area. The highest ELAs occurred in the rainshadow of Mount Rainier at Huckleberry Park. The relatively high ELA at Crystal Lakes can be explained by the westward orientation of the cirque, and exposure to the afternoon sun. The relatively high ELA at Upper Hen Skin Lake may be explainable by a rainshadow effect in the lee of the crest of the Cascade Range.

During McNeeley time, a pronounced rainshadow existed northeast of Mount Rainier, centered in the Huckleberry Park area. The late-glacial ELA pattern (Fig. 32) mirrors today's projected ELAs (Fig. 38). Currently, most precipitation originates during storms associated with southwesterly winds. Apparently, the precipitation events during the McNeeley 1 and McNeeley 2 glacier advances were associated with similar wind directions.

The McNeeley 1 ELA was comparable to the McNeeley 2 ELA. The ELAs for both the McNeeley 1 and McNeeley 2 advances ranged between 1710 and 1920 m, depending on the cirque's aspect and position within the rainshadow of Mount Rainier. Projection of the current ELAs into the study area results in estimated modern ELA values between 2150 and 2350 m. The ELAs of the McNeeley advances were approximately 400-500 m lower than at present. The ELA depressions during full-glacial time were approximately 950 m (inferred Vashon) and approximately 1050 m (Evans Creek). Thus, the McNeeley ELA depression was ca. 42 to 53% of the ELA depression during the inferred Vashon advance.

The above comparison of the McNeeley ELAs with current and full-glacial (Evans Creek) ELAs may be unreliable because it is based on a projection of the current ELAs into the study area. In the rainshadow of Mount Rainier, large variations in precipitation occur on a small scale, which do affect ELAs. Current annual precipitation in the study area ranges from approximately 1800 to more than 2400 mm. No modern glaciers or Neoglacial moraines exist in the study area to allow direct comparison with the McNeeley glaciers.

7.3 GOAT ROCKS

The glaciers in the Goat Rocks included in this study were, or are, oriented approximately in the same direction. The crest of the Cascades is the only significant topographic obstacle that creates a rainshadow effect in the area. As a result, there probably is a small rainshadow east of the crest of the Cascades.

PRESENT GLACIERS (1996)

In late 1996, the extent of two glaciers was mapped in the field (Fig. 37), and the ELAs were calculated using the accumulation-area ratio method. The ELA of the McCall glacier in the headwaters of the North Fork of the Tieton River was ca. 2200 m. The ELA of the glacier in the cirque at the headwaters of Conrad Creek was ca. 2200 m. Other glaciers are shown on the topographic maps of the area, but since the extent of the glaciers has decreased considerably since the maps were compiled in the 1950s, the mapped extent cannot be used to estimate the current ELAs.

LAST (FRASER) GLACIATION

Evans Creek advance

The glacier extent of the Evans Creek advance is poorly known. Furthermore, the complex geometry of several coalescing glaciers makes ELA calculations difficult.

Vashon advance

The ELA of the inferred Vashon glacier that occupied the North Fork of the Tieton River valley was ca. 1300 m (Fig. 36). Without taking into account the ELA gradient across the Cascade Range, the ELA depression between the present ELA and that of full-glacial time is ca. 900 m. However, considering that ELAs rise to the east, the actual ELA depression probably was somewhat larger. For the purposes of this study, the Vashon ELA depression is estimated as ca. 950 m. These reconstructions assume that the present ELA gradient parallels the ELA gradient during Vashon time. However, considering the relatively short distance of ca. 12 km between the inferred Vashon glacier terminus and the present glaciers, the above estimate for the ELA depression seems reasonable.

McNeeley glaciers

The reconstructed ELA for both McNeeley glaciers in the valley of the North Fork of the Tieton River was ca. 1720 m (Fig. 37). The ELA of the glaciers during both McNeeley advances in the South Fork of the Tieton River was ca. 1650 m (Fig. 37). The McNeeley moraines lie within 2.5 km of the present glaciers. On such a small scale, the errors associated with ELA gradients probably are negligible compared to the errors associated with the determination of the ELAs of current and past glaciers. The ELA depression during the McNeeley advances was ca. 480-550 m. With the reconstructed ELA depression of 950 m for the inferred Vashon advance in the Goat Rocks, the ELA depression during late-glacial time was close to half that of the ELA depression during the inferred Vashon advance.

The ELA depressions calculated for the former glaciers in the Goat Rocks are similar to the ELA depressions calculated for the areas around Mount Rainier. The McNeeley ELAs in the study area near Mount Rainier range between 1710 and 1920 m. These values are somewhat higher than those calculated for the Goat Rocks (1650-1700 m). This difference probably is due to the uncertainty in accounting for the rainshadow effect of Mount Rainier. The late-glacial ELA depressions in the Goat Rocks are better constrained than those in the vicinity of Mount Rainier, because current glaciers and McNeeley moraines occur in close proximity in the Goat Rocks.

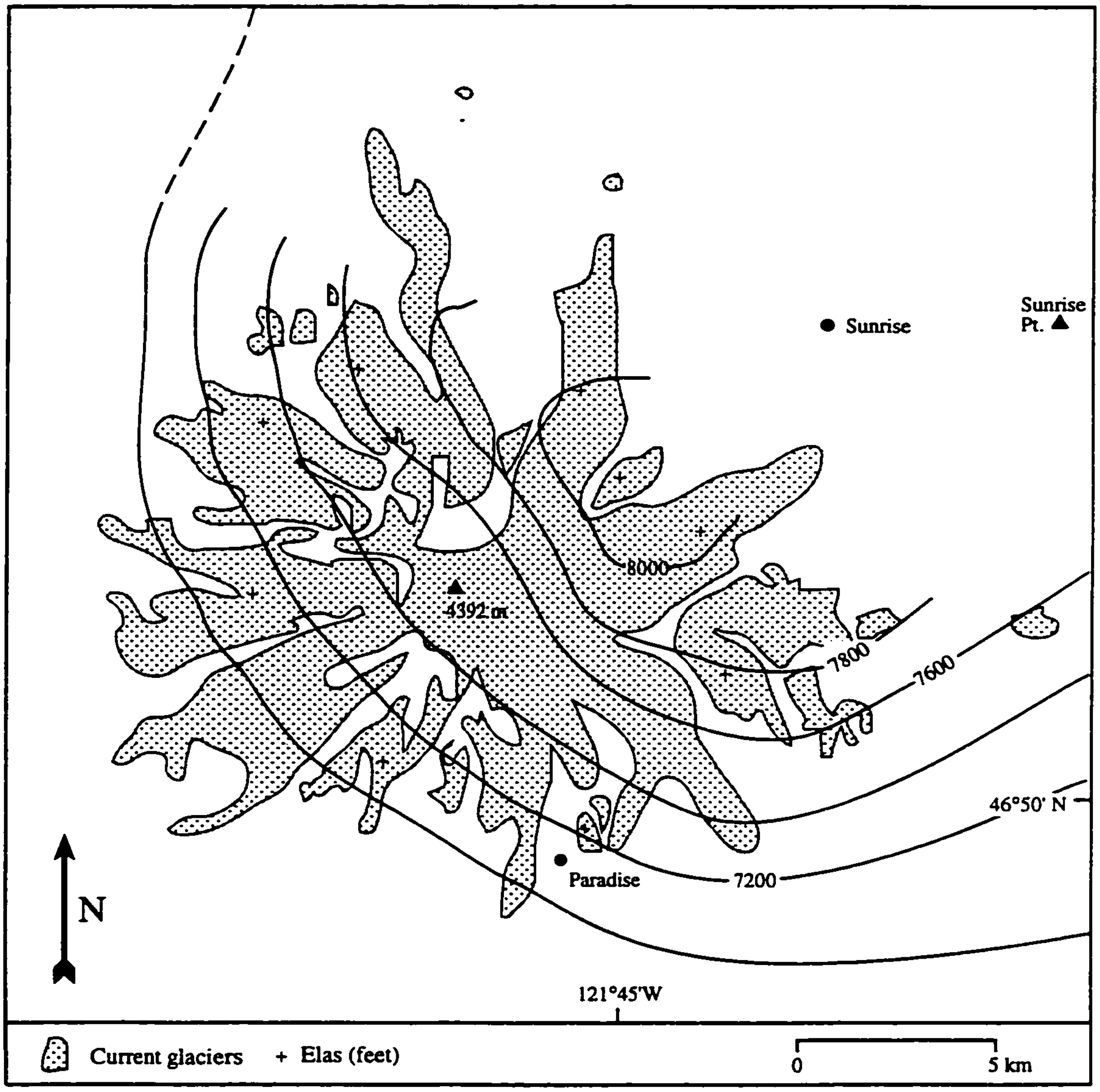


Figure 38: Surface of potential equilibrium line altitudes of recent glaciers on Mount Rainier. Reconstruction is based on snowlines in late September/early October during three consecutive years 1971-1973 (see text). Toward northeast, values are based on extrapolation and therefore may be unreliable. The small glaciers north of Mount Rainier are located in shaded cirques and do not accurately reflect the ELA of the surrounding area.

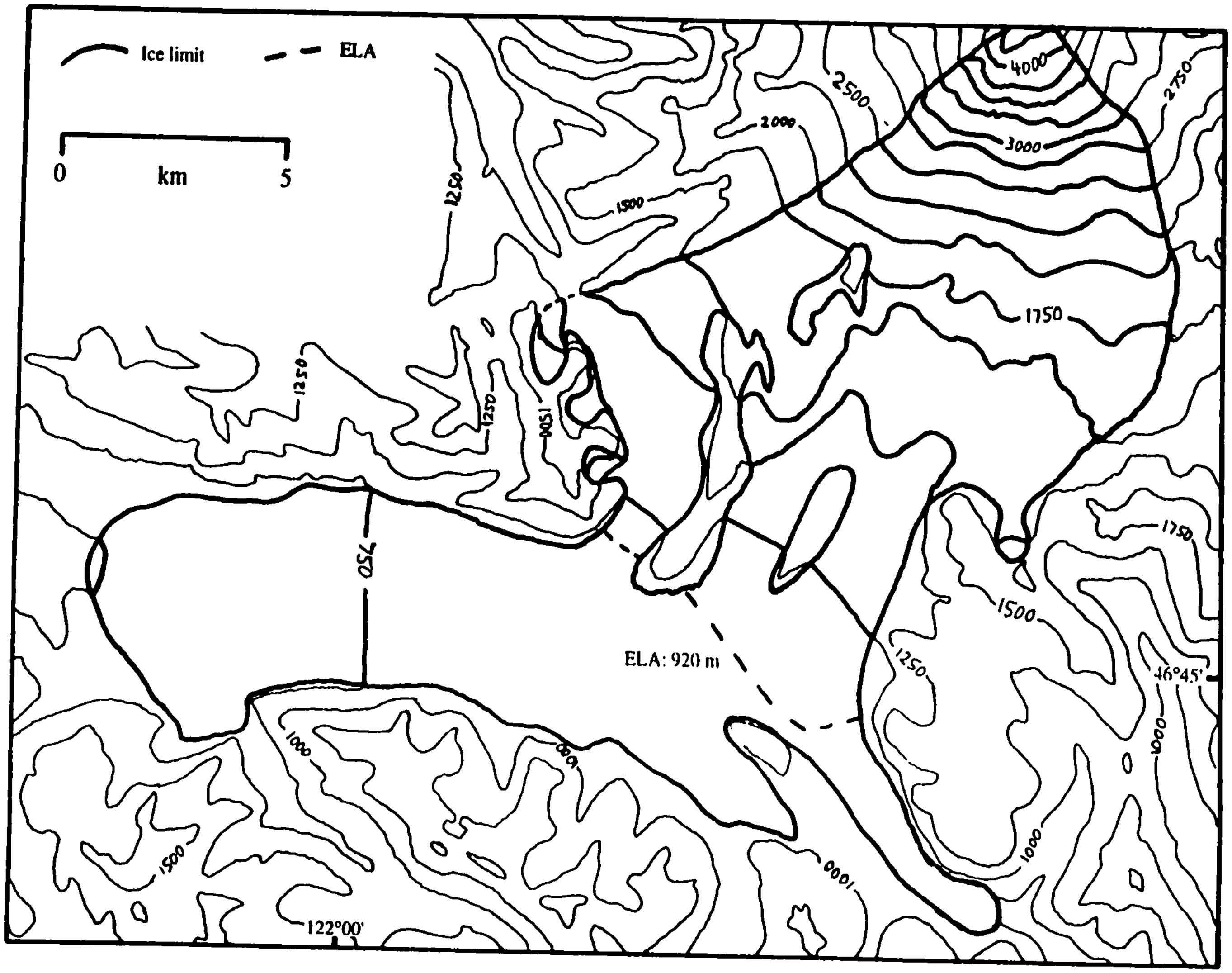


Figure 39: Extent of the Nisqually Glacier during the Evans Creek advance. Contour interval 250 m. Other glaciers may have existed in the area, but are not shown.

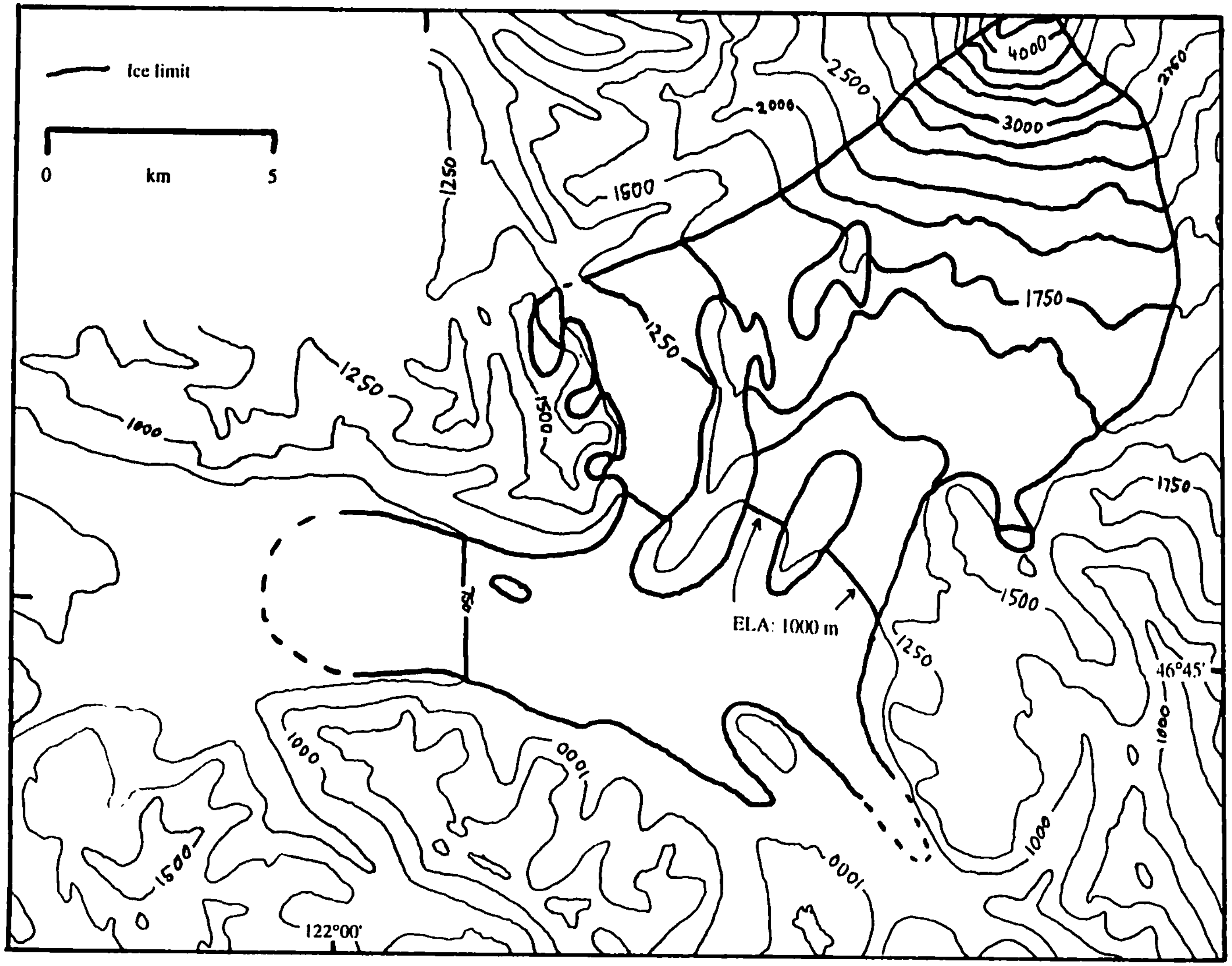


Figure 40: Extent of the Nisqually Glacier during the inferred Vashon advance. Contour interval 250 m. Other glaciers may have existed in the area, but are not shown.

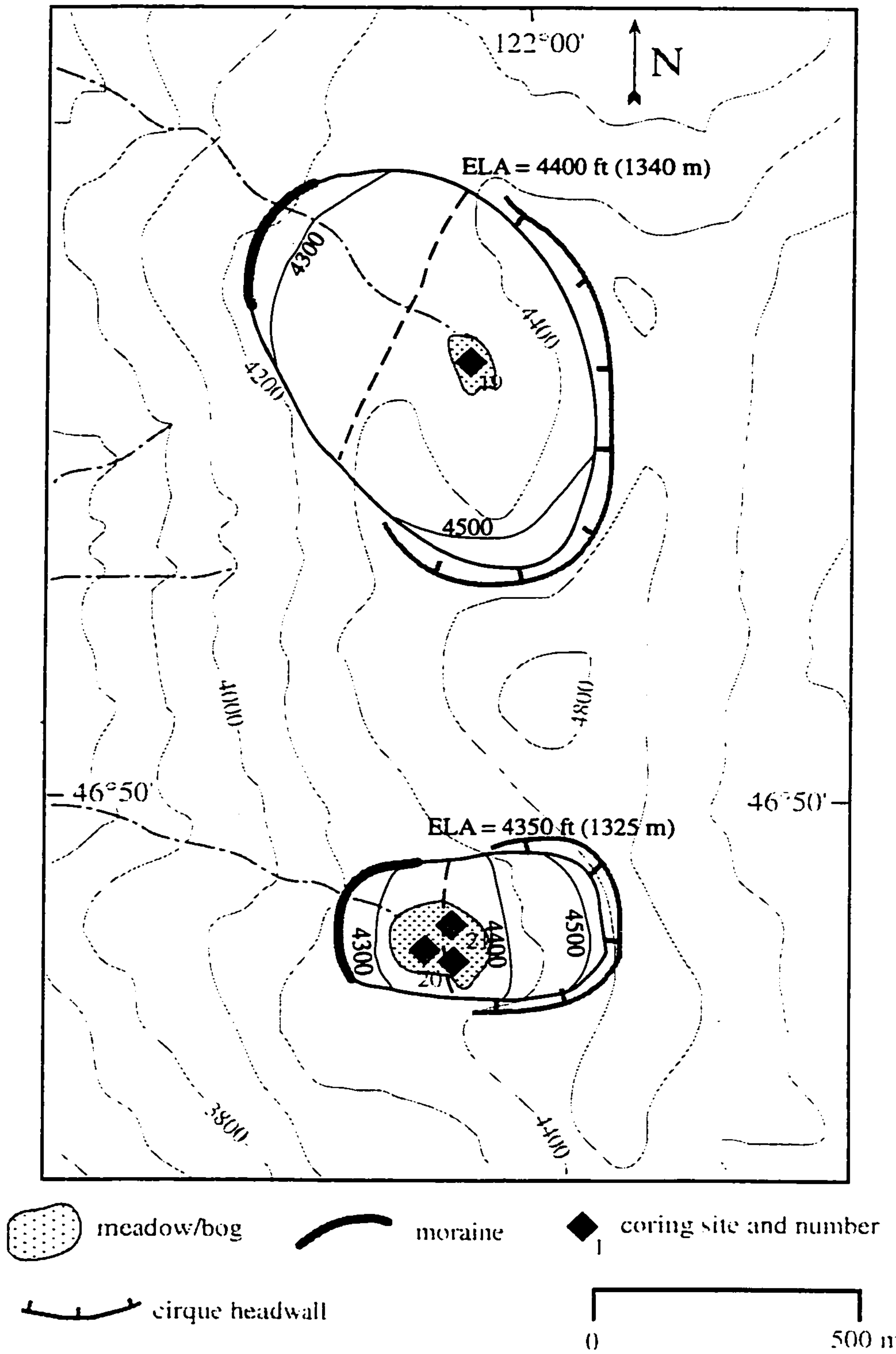


Figure 41: Reconstruction of glaciers at Lake Mary Lea. Dashed line is ELA. Contour interval: 100 ft.

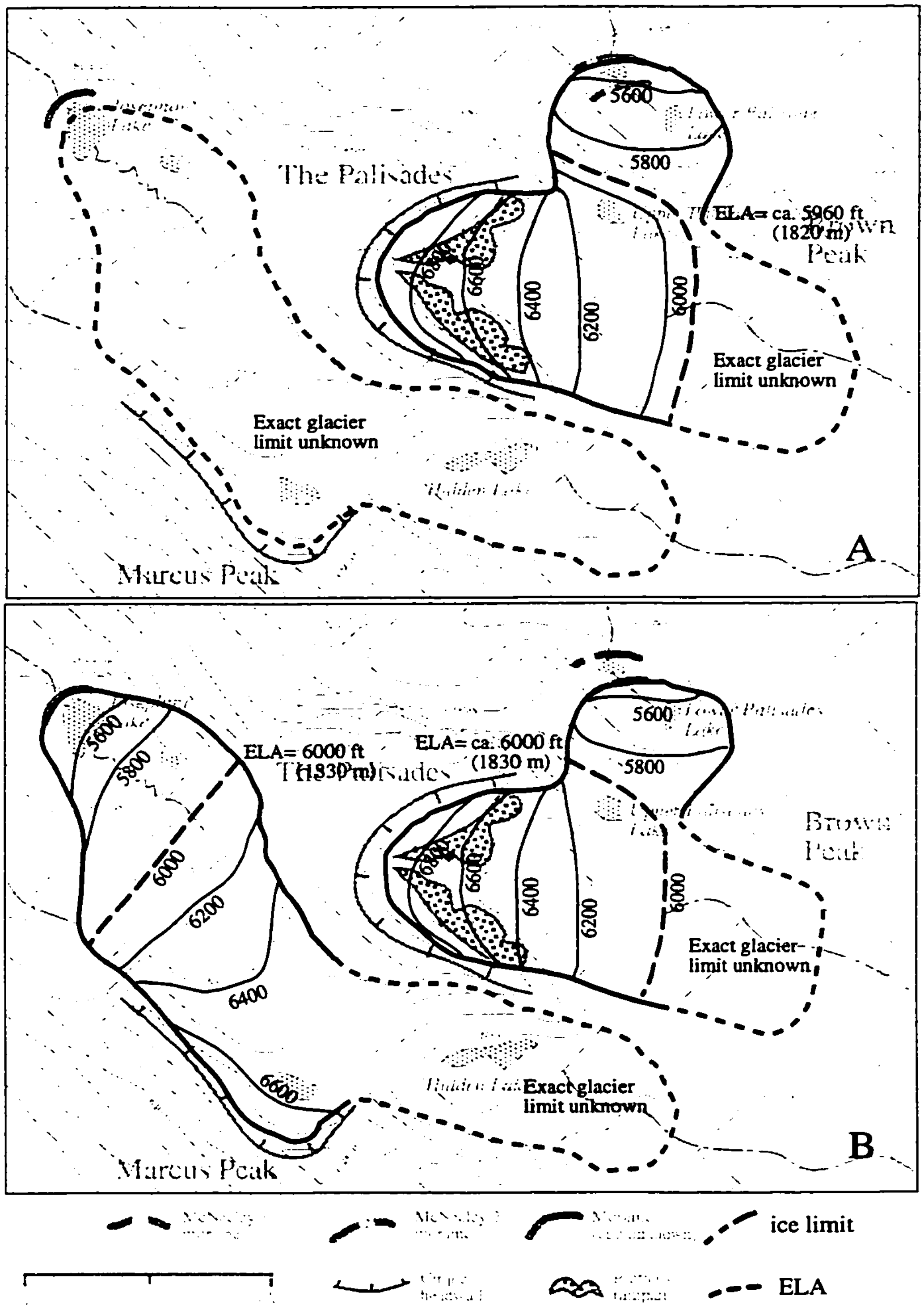


Figure 42: Reconstruction of McNeeley 1 (A) and McNeeley 2 (B) glaciers at the Palisades. Numbers on contour lines are elevation in feet.

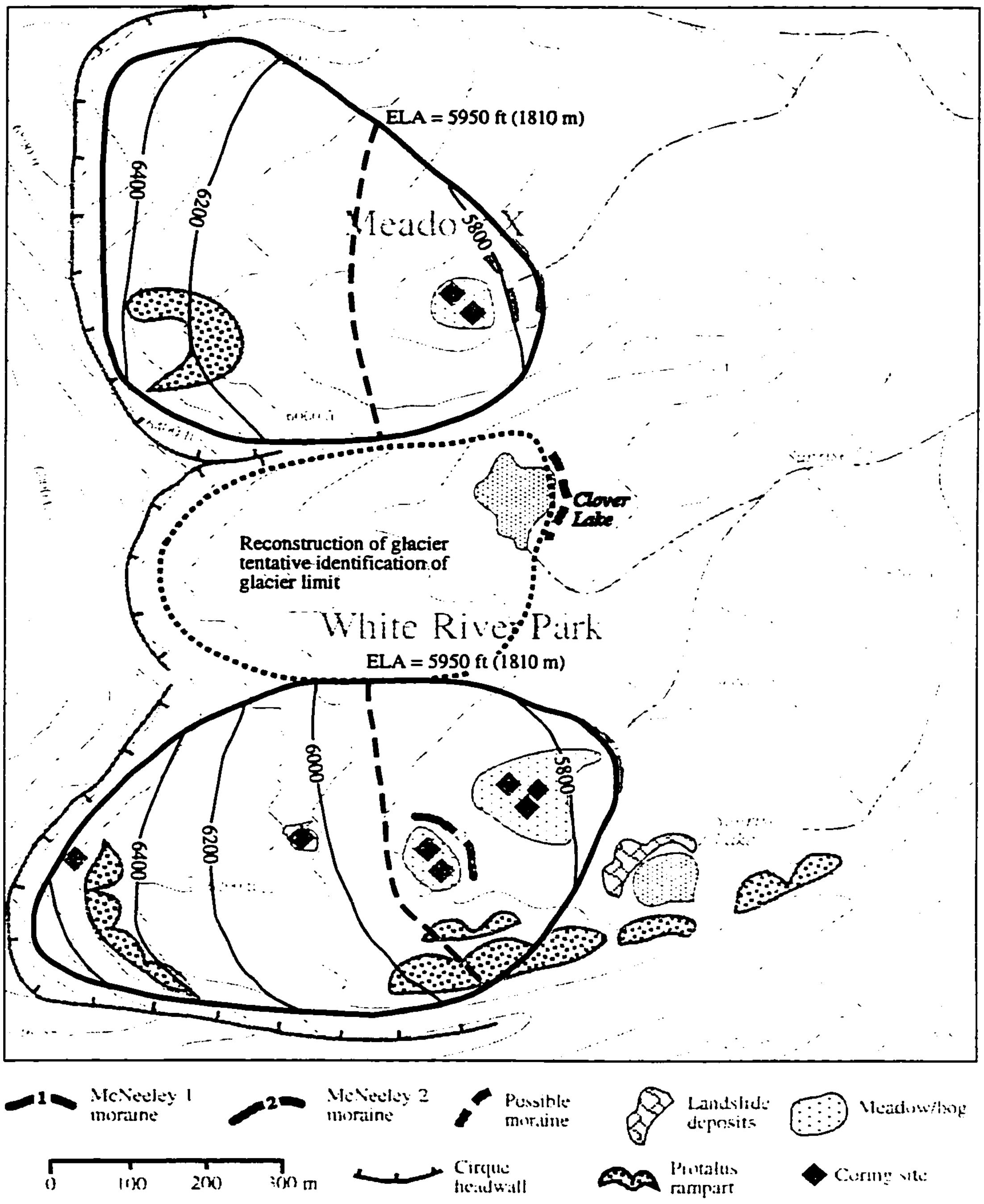


Figure 43: Reconstruction of McNeeley 1 glacier at White River Park. Dashed line is ELA. Numbers on contour lines are elevation in feet.

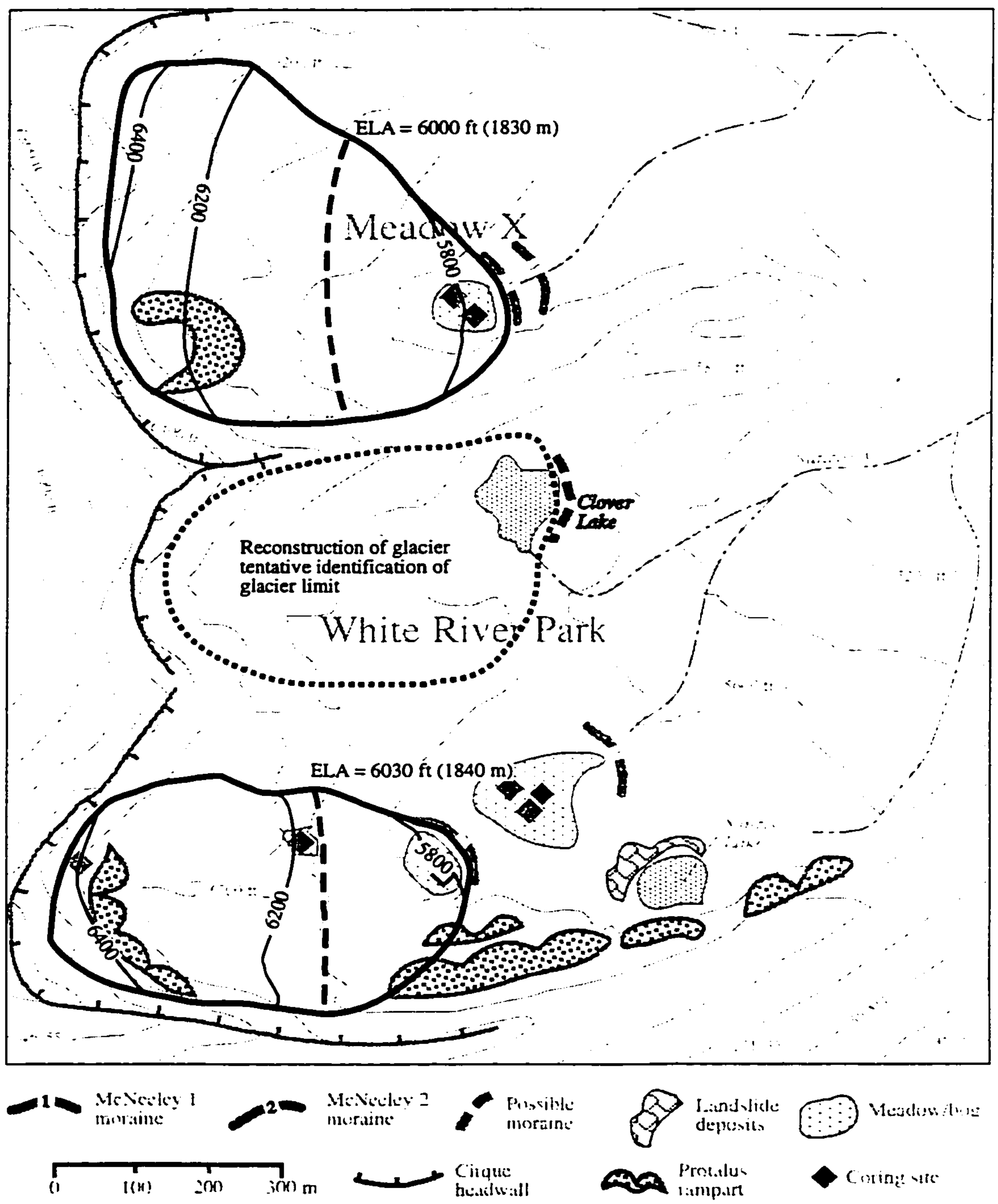


Figure 44: Reconstruction of McNeeley 2 glacier at White River Park. Dashed lines across glaciers are ELA. Numbers on contour lines are elevation in feet.

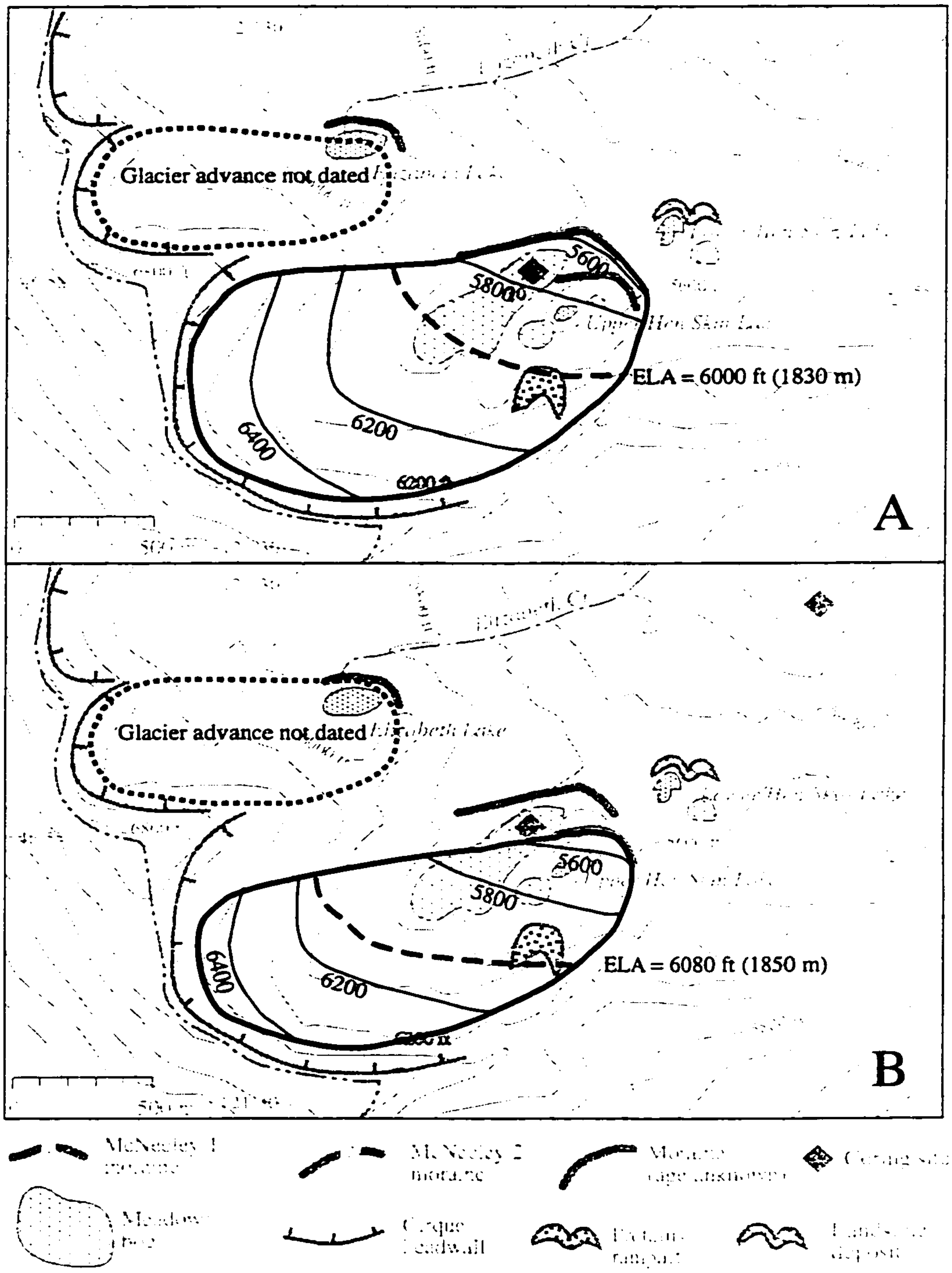


Figure 45: Reconstruction of McNeeley 1 (A) and McNeeley 2 (B) glaciers at Crystal Moutain. Dashed lines across glaciers are ELAs. Numbers on contour lines are elevation in feet.

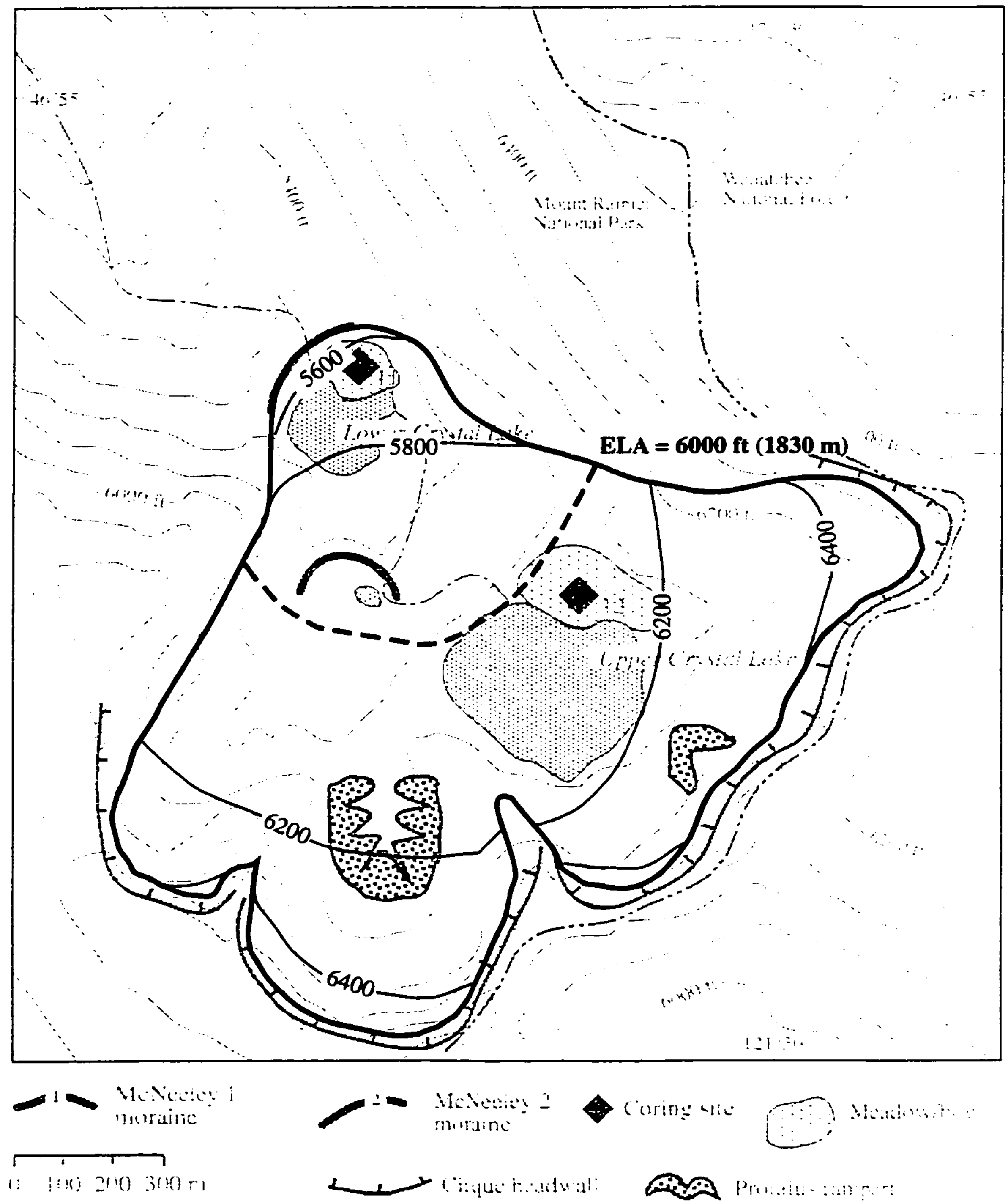


Figure 46: Reconstruction of McNeeley 1 glacier at Crystal Lakes. Dashed line is ELA. Numbers on contour lines are elevation in feet.

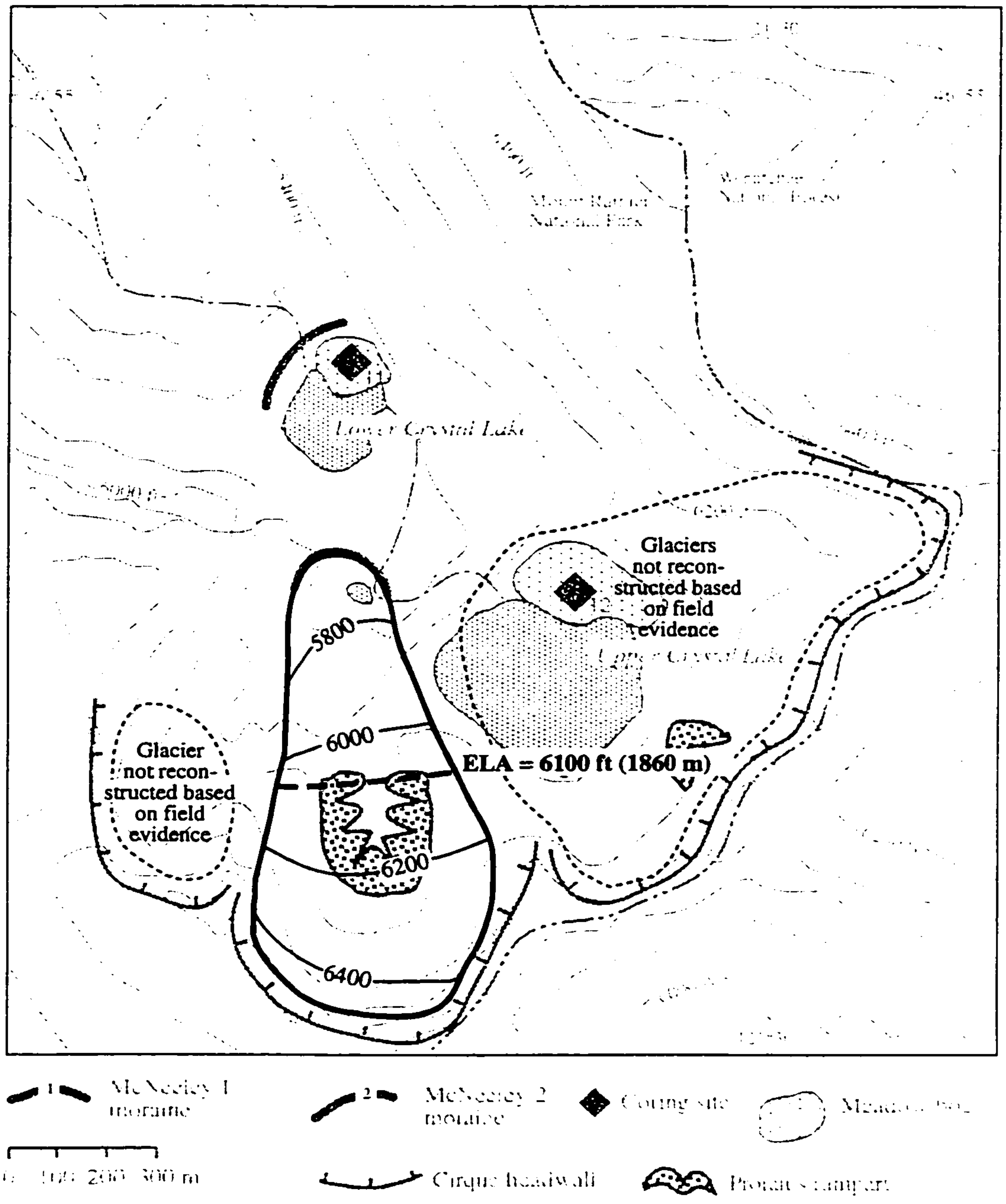


Figure 47: Reconstruction of McNeeley 2 glacier at Crystal Lakes. Dashed line across glacier is ELA. Numbers on contour lines are elevation in feet.

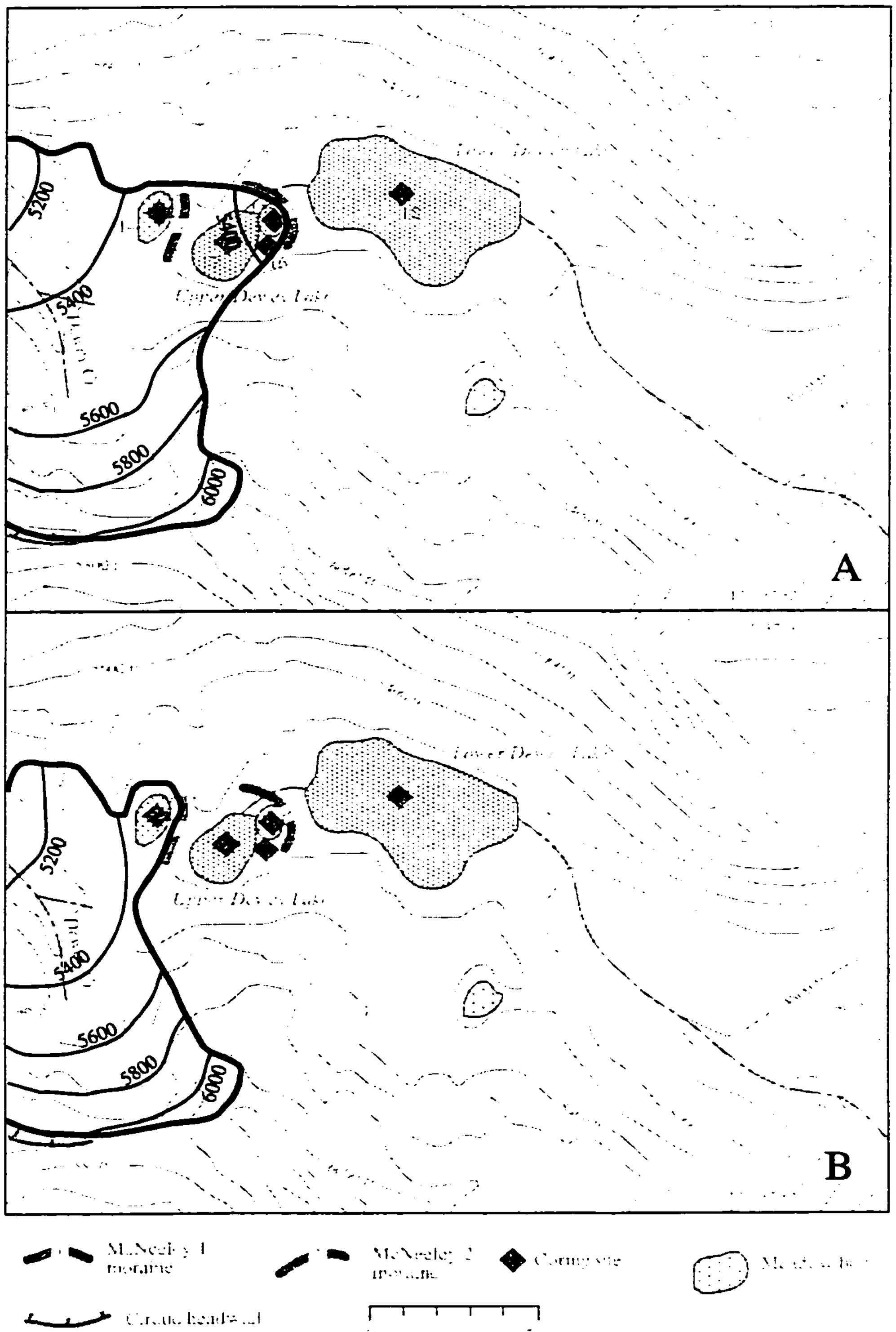


Figure 48: Reconstruction of McNeeley 1 (A) and McNeeley 2 (B) glaciers at Dewey Lakes. Numbers on contour lines are elevation in feet.

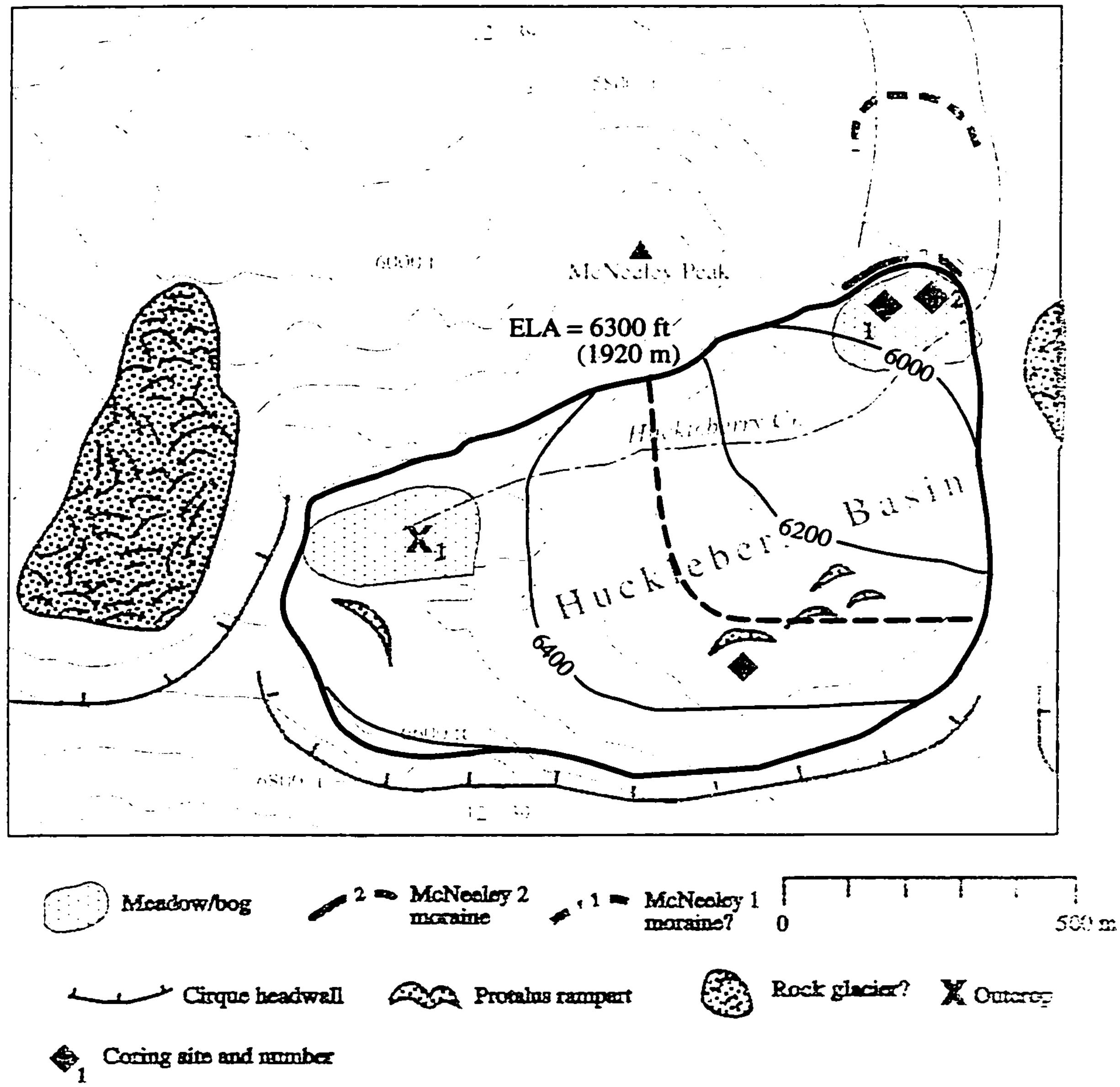


Figure 49: Reconstruction of McNeeley 2 glacier at Huckleberry Park. Dashed line is ELA. The ELA is curved due to the complex shape of the cirque. Numbers on contour lines are elevation in feet.

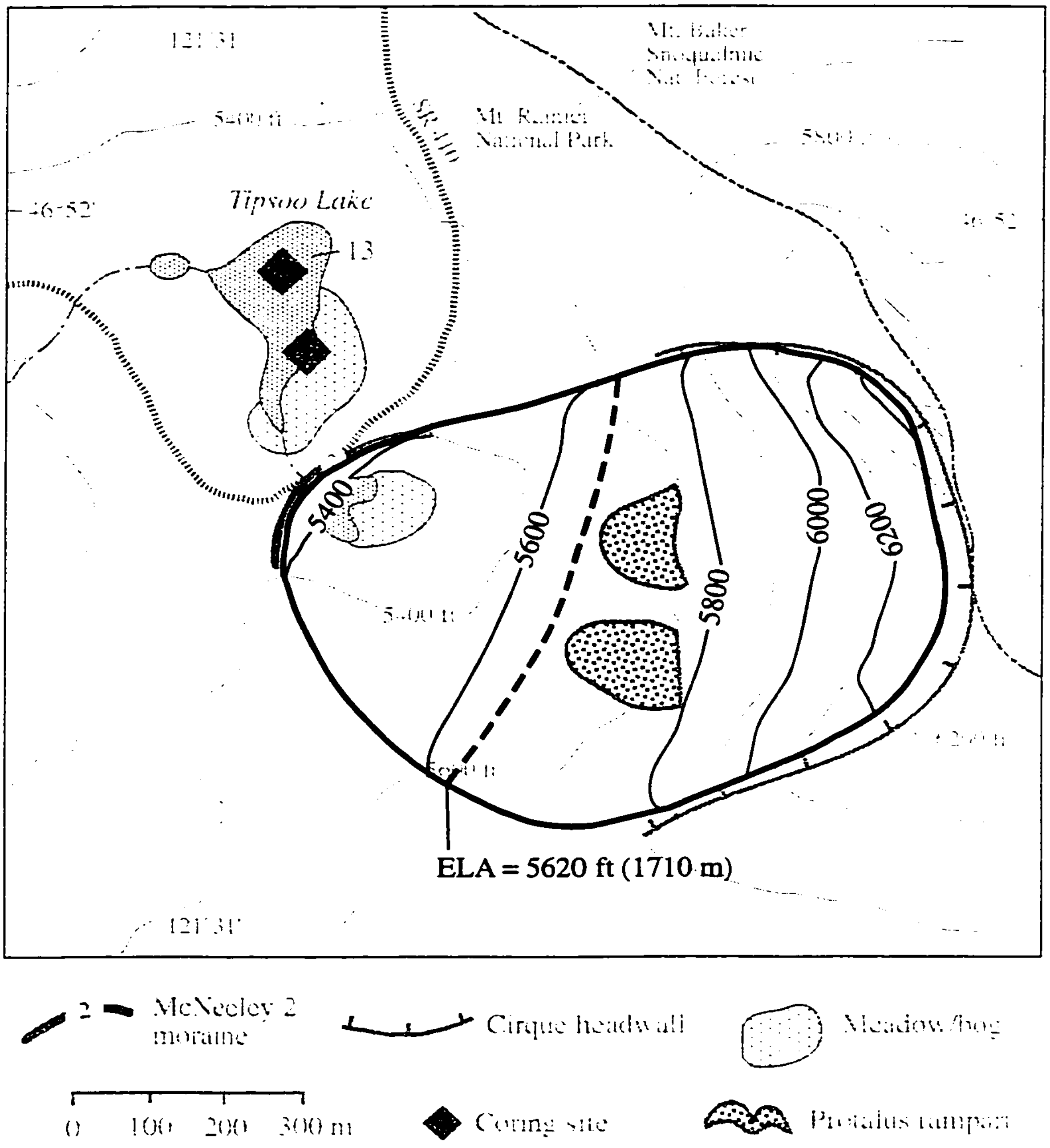


Figure 50: Reconstruction of McNeeley 2 glacier at Tipsoo Lake. Dashed line is ELA. Numbers on contour lines are elevation in feet.

8 RECONSTRUCTION OF CLIMATE ON AND NEAR MOUNT RAINIER

8.1 FULL-GLACIAL ICE ADVANCES

Two glacier advances have been documented on and near Mount Rainier during full-glacial time. The older glacier advance has been termed the Evans Creek advance (Crandell and Miller, 1974).

The evidence from Lake Mary Lea (chapter 6.2) poses a number of questions. The cirques at Lake Mary Lea have been deglaciated at least since ca. 27,500 cal yr BP (24,000 ^{14}C yr BP). If they were glaciated last during the Evans Creek advance, then this advance occurred earlier than the previously assumed date of ca. 23,000 cal yr BP (20,000 ^{14}C yr BP) (Porter et al., 1983). Three new ^{36}Cl dates for a moraine of possible Evans Creek age in the North Cascades average about 24,000 cal yr BP (Swanson and Porter, 1997). It is possible that the dated boulders have eroded, and that the ^{36}Cl dates in effect are minimum dates for the Evans Creek moraine. However, possible errors with the radiocarbon dating, the tentative calibration (Bard et al., 1993), and the ^{36}Cl ages may permit reconciliation of these dates by suggesting a date of 28,000 cal yr BP for deposition of the Evans Creek moraines.

On the other hand, possibly the moraines at Lake Mary Lea were deposited prior to the Evans Creek advance, and the cirques were not glaciated during Evans Creek time. The cores at Lake Mary Lea did not reach basal till, and it is unknown how much time elapsed between deglaciation and deposition of the dated layer. If the cirques at Lake Mary Lea were not glaciated during the Evans Creek advance, the ELA at Lake Mary Lea then must have been higher than 1325 - 1340 m to prevent glaciation of the cirques. Compared to the Evans Creek ELA of Nisqually Glacier (920 m), this seems very high. However, this might be explained by the location outside the orographic influence of Mount Rainier, which depressed the ELA of the Nisqually Glacier, and by the shallow cirques, which are exposed to the sun throughout most of the day.

Based on ^{36}Cl ages from the North Cascades (Swanson and Porter, 1997), it is inferred that the second advance of alpine glaciers, including that of the Nisqually Glacier, occurred at the same time as the Vashon advance of the Cordilleran Ice Sheet in the Puget Lowland, about 15,000-14,000 ^{14}C yr BP (Fig. 51).

Both the Evans Creek and the inferred Vashon advances were the result of an ELA depression of approximately 1000 - 1080 m in the Nisqually River valley, a value which is similar to the ELA depression of the inferred Vashon advance in the Goat Rocks of ca. 950 m. This is in good agreement with the "full-glacial" ELA depression of 950 m previously suggested for the Cascade Range (Scott, 1977; Porter, 1977).

The ELA of the glacier that deposited the undated moraine at Ada Creek in the Huckleberry Creek valley, north of Mount Rainier, was only ca. 300 m lower than the late-glacial McNeeley ELAs in this area. By comparison, the inferred Vashon ELA in the Goat Rocks was ca. 400 - 500 m lower than the McNeeley ELAs, and the Evans Creek ELA probably was similar to, or slightly lower than it was for the Nisqually Glacier. Assuming a stage 2 age for the moraine at Ada Creek, the area must have received anomalously low snowfall or experienced anomalously high temperatures during the glacier advance that deposited the moraine at Ada Creek, compared to late-glacial time when the McNeeley moraines were deposited. It is unlikely that temperatures gradients, which depend very much on altitude and adiabatic lapse rates, have changed significantly during the past 30,000 years. A regional anomaly in precipitation could have been caused by a more pronounced rainshadow effect of Mount Rainier. The age of the moraine has not been established. If the moraine was deposited earlier than the Vashon advance, then all subsequent advances must have been less extensive, and must have had higher ELAs. However, it cannot be excluded that the ridge at Ada Creek is not a moraine, but a landslide deposit (see chapter 6.2). The dense forest cover makes unambiguous identification of moraines difficult. No moraines in similar locations have been identified in other drainage areas, which could support the interpretation of the ridge at Ada Creek as a moraine.

8.2 MCNEELEY 1 ADVANCE

Glaciers retreated following the McNeeley 1 advance before 13,200 cal yr BP (11,300 ^{14}C yr BP) (Figs. 51 and 52). The associated ELA depression was ca. 50% of the full-glacial ELA depression.

8.3 THE INTERVAL BETWEEN 13,000 - 11,600 CAL YR BP (11,200 - 10,000 ^{14}C YR BP)

After deglaciation, sedimentation rates in all cored lakes in the vicinity of Mount Rainier remained low (Fig. 33), as did organic-matter content (<5% loss on ignition). Ca. 11,600 cal yr BP (10,000 ^{14}C yr BP), sedimentation rates and organic content increased dramatically. For example, in core 8 (Fig. 33), average sedimentation rates were ca. 18 cm/1000 yr during late-glacial time, and increased to ca. 230 cm/1000 yr after 11,600 cal yr BP (10,000 ^{14}C yr BP). In contrast, deglaciation following the McNeeley 2 advance saw a rapid return to organic sedimentation. The low-organic sedimentation cannot be explained as a result of dilution caused by increased sediment input from unstable slopes, as the sedimentation rates are too low. Instead, it is interpreted as evidence of a poorly developed vegetation cover. Because a lag in renewed organic sedimentation of about 1500 years (based on the calibrated ages) is unlikely to be explained by a lack of soil development or by plant migration from lower elevations into the study area, it is more reasonable to infer that the low-organic sedimentation resulted from a climate unfavorable to the recolonization by vegetation. When the glaciers retreated following the McNeeley 1 advance, the climate may have been too dry for glaciers to remain extended, yet too cold for significant vegetation to become established in the deglaciated terrain. Palynological evidence from two lakes in the Puget lowland near Mount Rainier also shows a slight drying and cooling during this interval (Tsukada et al., 1981), but further paleovegetation studies are necessary to determine the nature of the climate at that time.

8.4 EARLY HOLOCENE

Significant vegetation became established in the vicinity of Mount Rainier by about 11,600 yr BP (10,000 ^{14}C yr BP), when sedimentation in all cored lakes became more organic-rich (>25% loss-on-ignition) (Fig. 52). Sedimentation rates increased dramatically, and reached a maximum during the early Holocene. Organic-rich sedimentation lasted for at least 700 years. The radiocarbon ages for the onset of Holocene organic-rich sedimentation scatter considerably between sites. This may be due to problems with radiocarbon of dating organic matter formed around 11,600 cal yr BP (10,000 ^{14}C yr BP), a time of rapidly rising $\Delta^{14}C$ values (Björck et al., 1996), which resulted in one of the most pronounced radiocarbon plateaus (see chapter 1) yet identified. It is almost impossible to assign precise ^{14}C ages to this interval (Björck et al., 1996).

8.5 MCNEELEY 2 ADVANCE

Glaciers readvanced during the McNeeley 2 event between 10,900 and 10,000 cal yr BP (9800 and 9000 ^{14}C yr BP) (Figs. 51 and 52). After deglaciation following the McNeeley 2 advance, organic sedimentation resumed with little apparent lag, implying that after deglaciation, conditions rapidly became favorable for plant growth.

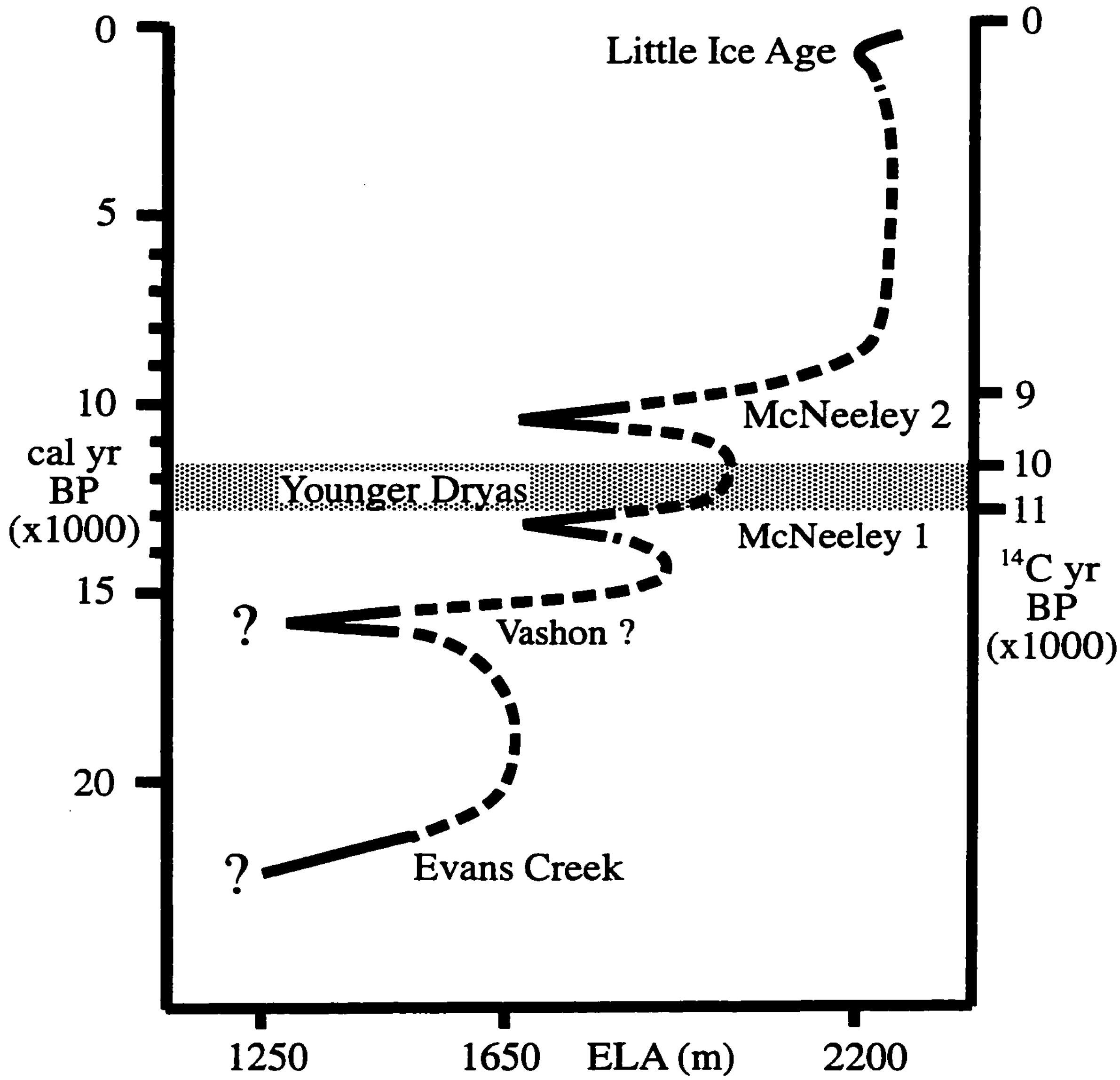


Figure 51: Fluctuations of ELAs over time in the vicinity of Mount Rainier and in the Goat Rocks, based on the interpretations of the glacial deposits in this study. Dashed lines show time periods when glacier extent is unknown. Question marks indicate that the timing of the advance is not firmly established.

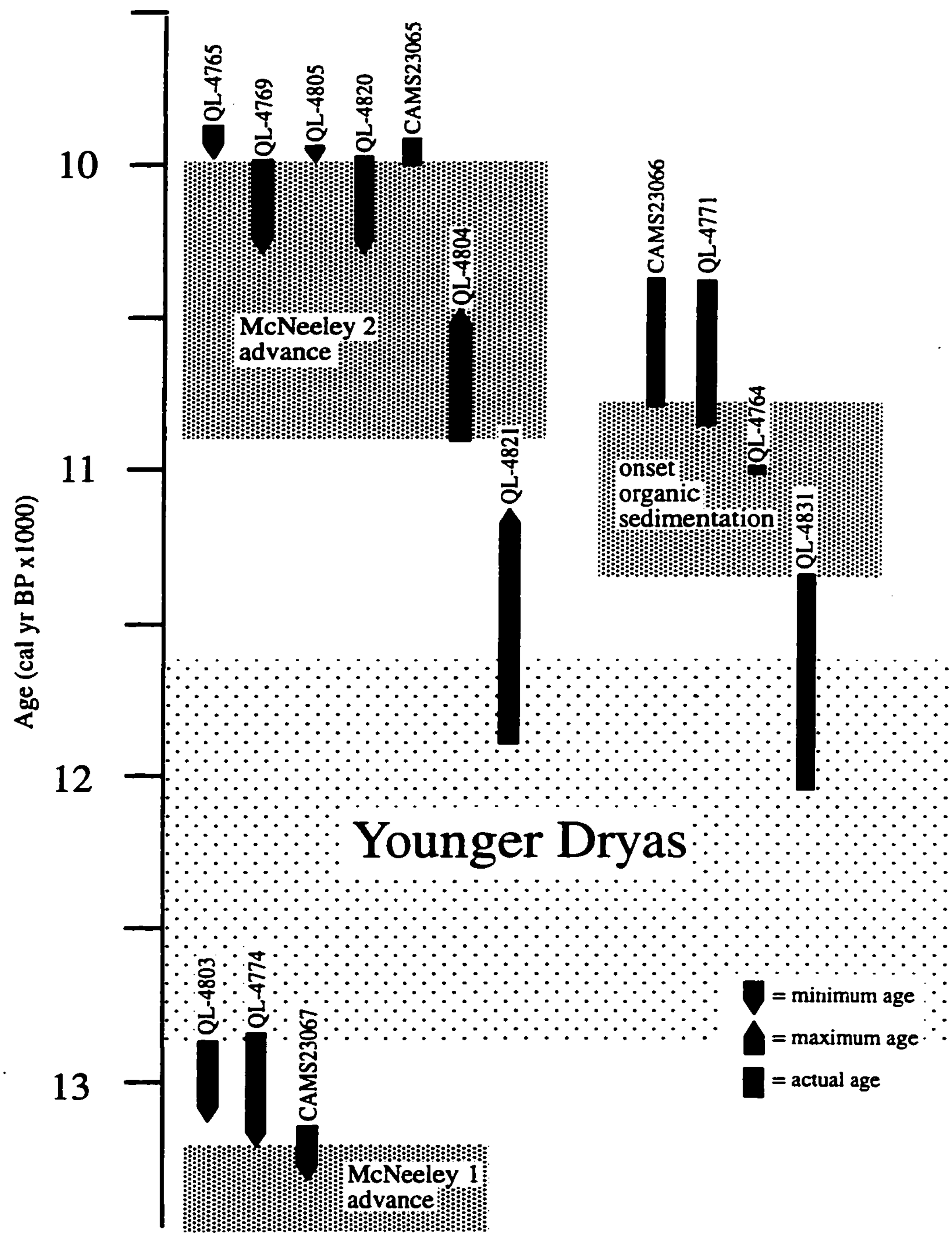


Figure 52: Radiocarbon ages used in this study (Table 2). Bars reflect calibrated 1-sigma intervals. Numbers are laboratory numbers. The large 1-sigma intervals between 11,500 and 11,000 cal yr BP reflect a radiocarbon plateau (see text). The McNeeley 2 advance additionally is constrained by tephra layer R (9880 cal yr BP) (Fig. 34).

9 EXAMINATION OF THE GLACIAL EVIDENCE WITH REGARD TO THE NORTH ATLANTIC YOUNGER DRYAS

The extensive discussion on the geographic extent of Younger Dryas-age glacier (e. g., Markgraf et al., 1992; Heine, 1993; Peteet, 1995; Osborn et al., 1995; Heine and Heine, 1996) warrants a separate assessment of whether the glaciers might have advanced near of Mount Rainier during Younger Dryas time (12,900-11,600 cal yr BP; 11,000-10,000 ^{14}C yr BP). The calibrated 2σ intervals of all limiting ages for the McNeeley 1 and McNeeley 2 glacier advances were examined to obtain the 95% confidence level for possible correlations (Fig. 53).

The McNeeley 1 advance probably predates the European Younger Dryas event, with glaciers retreating before 13,000 cal yr BP (11,200 ^{14}C yr BP). Even if the oldest minimum date for deglaciation is rejected, there is little possible overlap between the McNeeley 1 advance and the North Atlantic Younger Dryas event. Clearly, glaciers had retreated by 12,750 cal yr BP (10,850 ^{14}C yr BP) at the latest.

The McNeeley 2 advance apparently postdates the European Younger Dryas event. Both the maximum date of 10,950-10,420 cal yr BP (9580 ± 50 ^{14}C yr BP) and the direct date for inferred glacigenic sediments of 10,040-9890 cal yr BP (8990 ± 60 ^{14}C yr BP) postdate the European Younger Dryas event, even at the 2σ (95% confidence) level. Thus, by this interpretation, glaciers advanced during the McNeeley 2 advance no earlier than 10,950 cal yr BP (9680 ^{14}C yr BP), and glaciers retreated from moraines as late as 10,040-9890 cal yr BP (8990 ± 60 ^{14}C yr BP).

The limiting ages for the McNeeley 2 advance are from glacigenic sediments in lakes downstream from the McNeeley 2 moraines. It is possible that these sediments are not related to the McNeeley 2 glacier advance upstream, and that the minimum ages for the McNeeley 2 advance consistently lag deglaciation. In this case, the McNeeley 2 advance could be older than assumed above, and correlate with the European Younger Dryas climatic reversal. However, if glaciers retreated from the McNeeley 2 moraines at the end of the Younger Dryas interval at ca. 11,600 cal yr BP (10,000 ^{14}C yr BP), yet organic sedimentation in all cores behind the McNeeley 2 moraines started only after deposition of tephra layer R at 9880 cal yr BP (8850 ^{14}C yr BP), a consistent lag of 1600 years has to be explained. During this time, all other lakes in the study area (which are dammed by McNeeley 1 moraines) experienced high sedimentation rates of dark, organic sediments. It is unlikely that the lakes behind the McNeeley 2 moraines received no significant sediment input during this time, especially since these lakes are directly upstream from the lower lakes (dammed by the McNeeley 1 moraines), and therefore are the first sediment trap below the cirque headwalls for the drainage. It is likely, however, that no sedimentation occurred because the lakes had not yet formed. If the lag was caused by residual dead ice, it is unlikely that dead ice persisted in all cirques for 1600 years. More scatter in the minimum ages would be expected, with some lakes having organic sedimentation before deposition of layer R, and others, where layer R was deposited onto the dead ice and would be missing as a discrete layer from the lake sediments.

In summary, it seems likely that glaciers in the vicinity of Mount Rainier during Younger Dryas time were less extensive than before (McNeeley 1) and after (McNeeley 2). However, the climate during Younger Dryas time seems to have been unfavorable for the establishment of vegetation, which is interpreted as evidence for a cold and dry climate (see chapter 8).

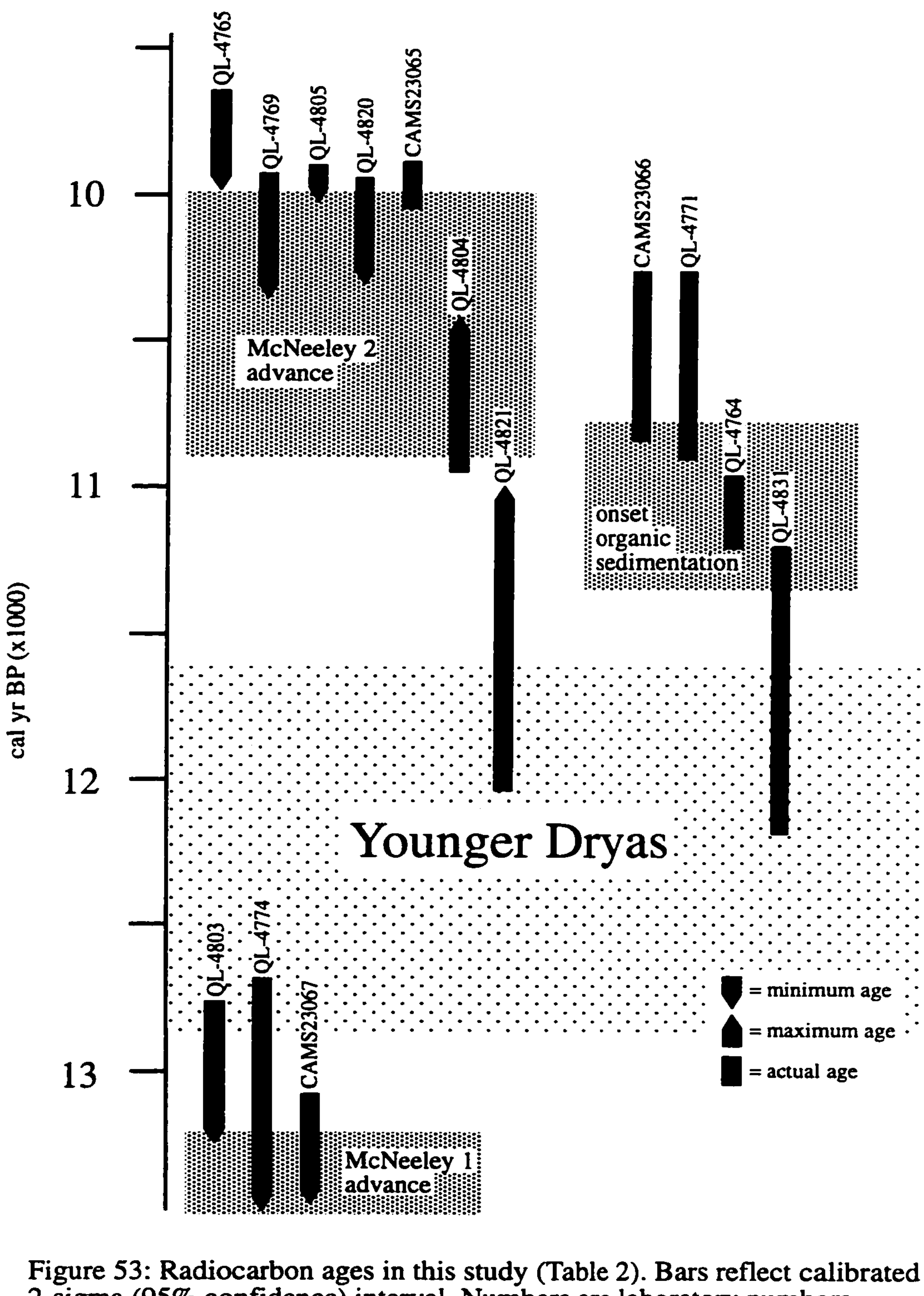


Figure 53: Radiocarbon ages in this study (Table 2). Bars reflect calibrated 2-sigma (95% confidence) interval. Numbers are laboratory numbers.

10 REGIONAL CORRELATIONS AND CLIMATE ALONG THE WEST COAST OF THE UNITED STATES AND CANADA

The McNeeley 1 advance occurred before 13,200 cal yr BP (11,300 ^{14}C yr BP) (Figs. 51 and 52). On Stevens Pass, Rat Creek-equivalent may be correlative with this advance moraines (Porter, 1978; Porter et al., 1983). Additional dating control for a possibly correlative moraine is provided by a minimum date of 13,030-12,890 cal yr BP (11,050 ± 50 ^{14}C yr BP) behind the outermost Hyak moraine on nearby Snoqualmie Pass (Porter, 1976). The ELA depression during the Hyak advance is comparable to the ELA depression for the McNeeley 1 moraines on Mount Rainier (Hurley, 1996). The two glacier advances probably are correlative. The McNeeley 1 advance probably occurred at the same time as the Sumas readvance of the Cordilleran Ice Sheet, between 13,600 and 13,100 cal yr BP (11,500 and 11,200 ^{14}C yr BP) (Saunders et al., 1987) (Fig. 3). If this is the case, this probably indicates that the Sumas readvance was caused by a shift in regional climate, rather than by a grounding of the Fraser Valley Lobe of the Cordilleran Ice Sheet due to isostatic uplift, as suggested by Waitt and Thorson (1983). The Recess Peak advance in the Sierra Nevada of California ended before 13,190-13,010 cal yr BP (11,190 ± 70 ^{14}C yr BP) (Clark, 1995; Clark and Gillespie, 1997), and may correlate with the McNeeley 1 advance. Deglaciation from the Piper Lake moraine in the Mission Mountains of Montana occurred before 13,100 cal yr BP (11,200 ^{14}C yr BP) (Osborn and Gerloff, 1997). The advance that deposited this moraine also may correlate to the McNeeley 1 advance. Considering the poor dating control for the Temple Lake moraines in Wyoming, it is possible that they also correlate with the McNeeley 1/Sumas/Recess Peak phase of glacier advance. However, because most of the advances are constrained by minimum dates, further studies from other areas in western North America are needed to determine whether this advance occurred throughout the region.

The apparent lag until vegetation became established following glacier retreat from the McNeeley 1 moraines may be the result of a cold and dry climate (Figs. 51 and 52), which could be correlated to the North Atlantic Younger Dryas event. A palynological study from alpine lakes in the Sierra Nevada (Hemphill and Clark, 1996) and evidence from several palynological sites in the Pacific Northwest (Matthewes et al., 1993) also suggest a cooler and drier climate during this period. However, other palynological studies show a uniform warming with no climatic reversal (Tsukada et al., 1981; Worona and Whitlock, 1995; Grigg, 1996). Since many of these study sites may not have been at or near boundaries of species distribution during that time, they may fail to record a small, brief event. Alternatively, Grigg (1996) speculates that sites from inland locations might not record a Younger Dryas signal, as they would have been more sensitive to the effects of increasing summer insolation, and less sensitive to changes in adjacent ocean conditions.

I interpret from the available evidence that the McNeeley 2 advance occurred during the early Holocene (Figs. 51 and 52). Early Holocene glacier advances have been suggested in the North Cascade Range, but the stratigraphy of one study (Begét, 1981) has been questioned (Davis and Osborn, 1987), while the date of the glacier advance in another study is a minimum tephrochronological date only (Waitt et al., 1982). It is possible that these advances occurred during the latest Pleistocene. However, no conclusive evidence for a major glacier readvance during Younger Dryas time has emerged from western North America so far. Glacier extent during the Crowfoot advance in the Canadian Rockies, which may have occurred during Younger Dryas time (Reasoner et al., 1994), was only slightly larger than at present. Dating control for the Temple Lake moraines in the Wind River Range, suggested as evidence for a Younger Dryas-equivalent advance (Zielinski and Davis, 1987; Gosse et al., 1995), remains problematic.

Pollen records from lowland sites in the Puget Lowland and Oregon suggest an early Holocene climate characterized by warmer and much drier summers than at present (Thompson et al., 1993; Sea and Whitlock, 1995; Worona and Whitlock, 1995). Such a climate does not appear to be consistent with glacier advances, unless the winters were extremely snowy. More paleovegetational research from alpine sites may help resolve this apparent discrepancy. It is possible that increased winter precipitation caused a glacier advance (due to increased snowfall) without affecting lowland vegetation, while the inferred summer drought may have affected the glaciers to a much lesser extent than it affected the vegetation. Earth-orbital variations predict a nigh seasonality at this time, with a maximum of summer insolation and a minimum in winter insolation (Berger et al., 1984). This raises the possibility of colder, snowier winters and drier summers that could resolve the apparent discrepancy between an early Holocene glacier advance and the pollen evidence.

11 COMPARISON WITH THE NORTHWESTERN EUROPEAN RECORD

The McNeeley 1 advance predates the European Younger Dryas event. It seems to have occurred during the time of the Allerød warm period of northwestern Europe. However, considering that the ages on the McNeeley 1 advance are minimum ages, it also is possible that the advance correlates with the preceding Older Dryas event in Europe.

The timing of the possible cold and dry event following the McNeeley 1 advance correlates with that of the Younger Dryas event in the Greenland ice cores, the end of which was marked by a rapid environmental change at the onset of the Holocene (Alley et al., 1993; Stuiver et al., 1995) and northwestern Europe (Björck et al., 1996). However, whereas conditions during Younger Dryas time apparently were dry and cold in the vicinity of Mount Rainier, the available records do not show a cold reversal after a warmer climate at the beginning of this interval. It is possible that conditions remained cold after ice retreat from the McNeeley 1 moraines, rather than warming at first, and then cooling again. In this case, a reduction in snowfall may have caused the glacier retreat.

Whereas the timing of the McNeeley 2 advance is similar to that of the Erdalen event in northwestern Norway (Nesje et al., 1991), speculations about a causal relationship between the two events seem premature. If a climatic event dating to this time can be demonstrated in other areas of the world, or if a link between northwestern Norway and the Cascade Range is suggested, a possible correlation can be examined further.

The available evidence does not suggest that the glacial sequence in the Cascade Range parallels the one in northwestern Europe. Most notable is the likely absence of a glacier advance in the vicinity of Mount Rainier during the Younger Dryas event. It is possible that the last glacial maximum of the glaciers in Scandinavia and northern Europe, which occurred around 20,000 ^{14}C yr BP, is not reflected on Mount Rainier as a phase of glacier advance. If the moraines at Lake Mary Lea (chapter 6.2) were deposited by an Evans Creek advance, which was the maximum advance of the last glaciation, then this advance occurred before 24,000 ^{14}C yr BP. However, the interpretation of the results from Lake Mary Lea is based on only one minimum radiocarbon date in one small cirque.

Glaciers in the vicinity of Mount Rainier may have reacted primarily to changes in available moisture. In this case, it may not be surprising that the alpine glacier advances are not in phase between the Cascades and western Europe. This does not preclude the possibility that climatic events in both areas were driven by the same underlying mechanisms. It is likely that late-glacial climate in the western United States was influenced mainly by events in the north Pacific Ocean. These events in turn may have been related to the North Atlantic events (which could have caused a slight Younger Dryas cooling and drying in the Cascades).

12 CONCLUSIONS

Glaciers in the vicinity of Mount Rainier advanced twice near the Pleistocene/Holocene transition. Glaciers retreated from the McNeeley 1 moraines before 13,200 cal yr BP (11,300 ^{14}C yr BP). Therefore, this advance predates the European Younger Dryas event. This phase of glacier advance seems to be a regional phenomenon occurring from California in the south to British Columbia in the north along the North American coastal cordillera, and most likely was caused by a regional shift to a moister and/or colder climate. Following retreat from this advance, significant vegetation probably did not become established around alpine lakes until ca. 11,600 cal yr BP (10,000 ^{14}C yr BP). Based on the low organic content of cores between 13,000 and 11,600 cal yr BP (11,100 and 10,000 ^{14}C yr BP), cold and dry conditions apparently existed in the area of Mount Rainier during Younger Dryas time. It probably was drier than in northwestern Europe, where glaciers expanded, while they retreated in the vicinity of Mount Rainier. It remains uncertain whether the two events were caused by the same mechanisms. The younger McNeeley 2 advance likely occurred during the early Holocene, between 10,900 and 10,000 cal yr BP (9800 and 9000 ^{14}C yr BP), at a time when palynological evidence from this region suggests warm, dry summers. This apparent discrepancy may reflect a strong seasonality, with colder winters with more snowfall, which led to positive mass balances and glacier advance.

The results of this study suggest that climate in the vicinity of Mount Rainier at the last glaciation/interglaciation transition probably was dissimilar from that in northwestern Europe. The glacier advances on Mount Rainier may have been the result of regional fluctuations in climate. However, it cannot be excluded that the same mechanisms were driving a pattern of rapid climatic change in both regions. The cold conditions on Mount Rainier during Younger Dryas time may indicate that the imprint of the Younger Dryas climatic reversal was global, but that it did not cause an equal climatic response everywhere.

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VITA

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Education

University of Washington
Ph.D. in Geology, 1992-1997

University of Bonn, Germany
Master's degree in Geography, 1990-1992
Minors: Soil Sciences and Mathematics
Overall grade A.
Thesis: *The Younger Dryas climatic reversal - Did it happen in South America?*

The University of Texas at Austin
Graduate coursework in Physical Geography, 1989-90
Exchange Student with scholarship
GPA 3.86.

University of Würzburg, Germany
Bachelor's degree in Geography, 1987-89
Minors: Geology and Mathematics.

Honors and awards

NASA Fellowship in Global Change	1994-97
Geological Society of America Outstanding Student Research Award	1994
The University of Texas at Austin Scholarship	1989-90
Test of English as a Foreign Language: score 670 (out of 677)	1989
Test of Written English: score 4.0 (out of 4.0)	1989

Internships

Remote Sensing Department 1991

Deutsche Agentur für Luft- und Raumfahrt (German branch of the European Space Agency), Munich.

Worked on a number of remote sensing projects.

Radiocarbon Laboratory 1991

Niedersächsisches Landesamt für Bodenforschung (Lower Saxony State Department of Soil Research), Hannover

Prepared and evaluated ^{14}C samples, other pedologic analyses.

Gobabeb Research Station, Namib Desert 1988

Namibia Department of Nature Conservation

Park Ranger. Aided in biological research. Independent research on sand dune morphology.

Community involvement

Washington Hydrological Society: active member
Northwest Bicycle Federation: volunteer education program coordinator
Citizenship: United States of America

Publications

Journal articles

Heine, J. T., in press. Timing, extent, and climatic significance of glacier advances at the Pleistocene/Holocene transition near Mount Rainier, Washington. Quaternary Science Reviews.

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