Chugach Mountains

Introduction

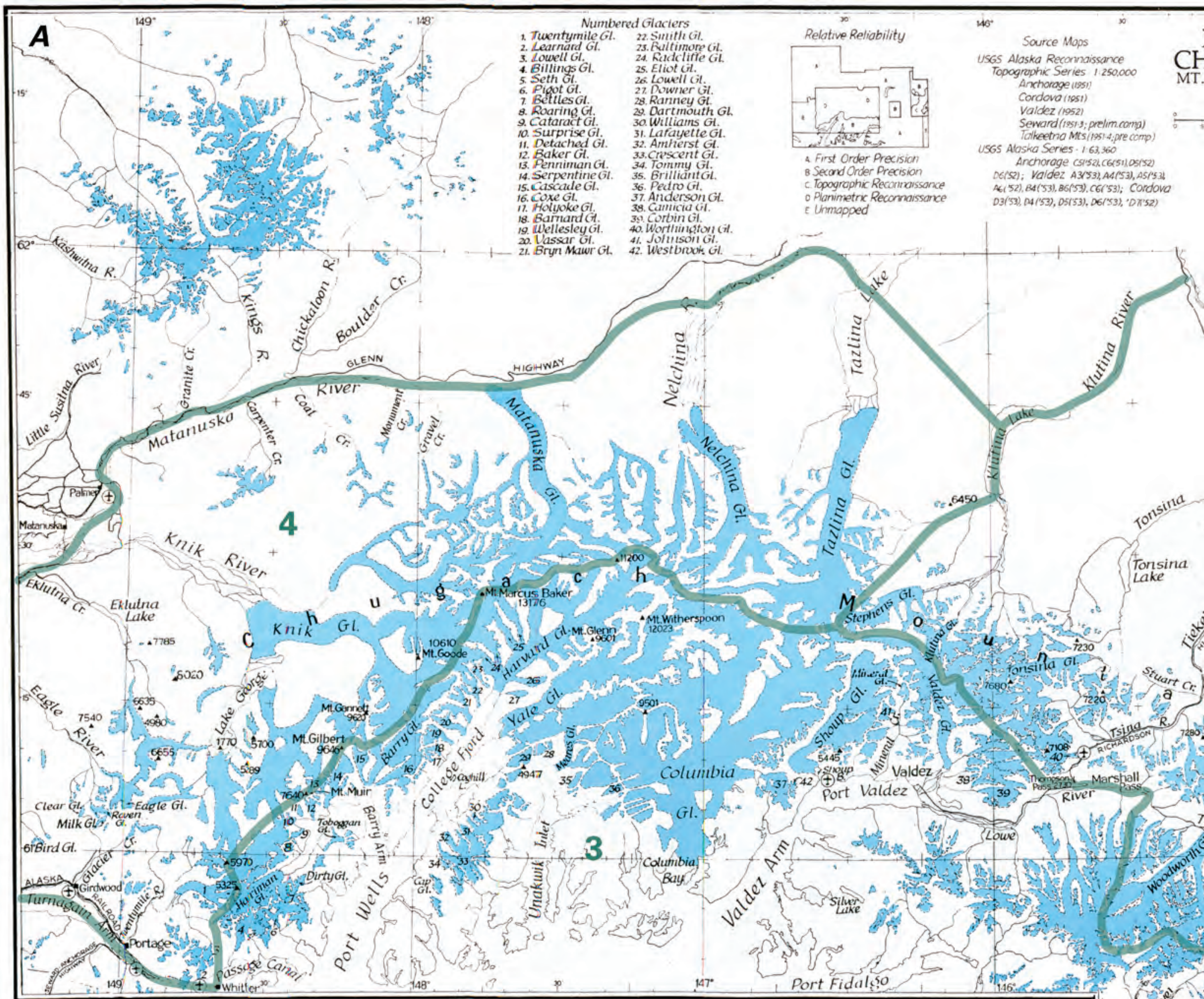
The Chugach Mountains are a 400×95-km-wide mountain range that extends from Turnagain Arm and Knik Arm on the west to the eastern tributaries of Bering Glacier, Tana Glacier, and Tana River on the east. On the north, the Chugach Mountains are bounded by the Chitina, Copper, and Matanuska Rivers. On the south, they are bounded by the northern Gulf of Alaska and Prince William Sound. The Chugach Mountains contain about one-third of the present glacierized area of Alaska (figs. 1, 2, 182) — 21,600 km^2, according to Post and Meier (1980, p. 45) — and include one of the largest glaciers in continental North America. Bering Glacier is a piedmont outlet glacier with an approximate area of 5,200 km^2 (Viens, 1995; Molnia, 2001, p.73) (table 2).

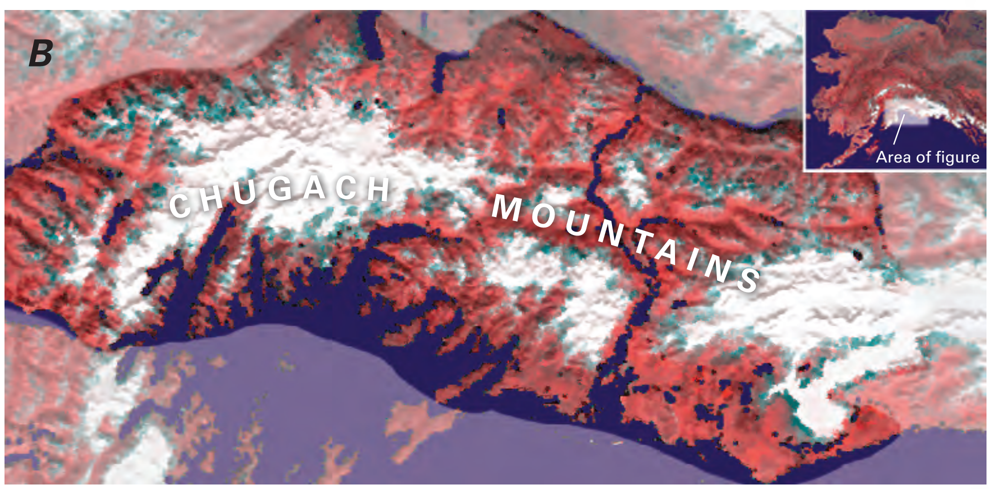
The eastern part of the Chugach Mountains is covered by a continuous series of connected glaciers and accumulation areas (Field, 1975b). Several studies have characterized this region and adjacent regions as areas experiencing a significant 20th century retreat of its glaciers (Meier, 1984; Molnia and Post, 1995; Arendt and others, 2002; Meier and Dyurgerov, 2002). A study by Sauber and others (2000) examined the effect of this regional ice loss on crustal deformation in the eastern Chugach Mountains. Recognizing that the range of annual thinning of glaciers in this region ranges from 1–6 m a^{-1}, they calculated that uplift in ablation regions of these glaciers ranges from 1 to12 mm a^{-1}, the greatest uplift being located just east of the Chugach Mountains, in the Icy Bay region. Sauber and Molnia (2004) hypothesized that continuing loss of glacier ice volume could lead to a future increase in very low magnitude earthquakes.

For ease of description, the Chugach Mountains are divided into four segments (fig. 182*A*): the Bering Glacier System Segment; the Copper River Drainage Segment, which has four subdivisions and includes glaciers that drain directly into the Copper River Delta; the Prince William Sound Segment, which has two subdivisions; and the Northwestern Chugach Mountains Segment, which also has two subdivisions.

The Bering Glacier System is bounded on the east by the St. Elias Mountains; on the south by the Gulf of Alaska; on the west by the Bering River, Canyon Creek, Carbon Mountain, and the Steller Glacier-Martin River Glacier divide; and on the north by an unnamed sinuous ridge that includes the summit of Mount Hawkins, the western Bagley Ice Valley-Tana Glacier divide, and the relatively straight ridge that includes Juniper Island. The eastern part of the Bering Glacier System is located in the western St. Elias Mountains and includes Quintino Sella and Columbus Glaciers, its eastern tributaries, both of which originate in Canada.

The Copper River Drainage Segment is bounded on the east by an irregular border that includes the Bering River, Canyon Creek, Carbon Mountain, the Steller Glacier-Martin River Glacier divide, the northern and eastern sides of Tana Glacier, and the Tana River; on the north by the Chitina River; on the northeast by the upper Copper River; and on the northwest by the Klutina River and Lake, Hallet River, and the divide between Tazlina Glacier and Stephens Glacier. On the southwest, it is bounded by the sinuous ridge that connects Mount Shouplina, Mount Cashman, Mount Mahlo, Mount Schrader, the divide between Valdez Glacier and Tonsina and Tsina Glaciers, Thompson Pass, Marshall Pass, the divide between Marshall Glacier and the unnamed glacier east of Deserted Glacier, the divide between the Gravina River basin and the north-flowing glaciers to its north, the divide between the Scott Glacier and the Rude River, including Cordova Peak, and Heney Range; and on the south by the Gulf of Alaska. This area has four subdivisions: Martin River Glacier–Martin River–lower Copper River– Bremner River–West Fork


A
Numbered Glaciers
1. Twentymile Gl.
2. Learnard Gl.
3. Lowell Gl.
4. Billings Gl.
5. Seth Gl.
6. Pigot Gl.
7. Bettles Gl.
8. Roaring Gl.
9. Cataract Gl.
10. Surprise Gl.
11. Detached Gl.
12. Baker Gl.
13. Penniman Gl.
14. Serpentine Gl.
15. Cascade Gl.
16. Coxe Gl.
17. Holyoke Gl.
18. Barnard Gl.
19. Wellesley Gl.
20. Vassar Gl.
21. Bryn Mawr Gl.
22. Smith Gl.
23. Baltimore Gl.
24. Radcliffe Gl.
25. Eliot Gl.
26. Lowell Gl.
27. Downer Gl.
28. Ranney Gl.
29. Dartmouth Gl.
30. Williams Gl.
31. Lafayette Gl.
32. Amherst Gl.
33. Crescent Gl.
34. Tommy Gl.
35. Brilliant Gl.
36. Pedro Gl.
37. Anderson Gl.
38. Canicia Gl.
39. Corbin Gl.
40. Worthington Gl.
41. Johnson Gl.
42. Westbrook Gl.
Relative Reliability
A First Order Precision
B Second Order Precision
C Topographic Reconnaissance
D Planimetric Reconnaissance
E Unmapped
Source Maps
USGS Alaska Reconnaissance Topographic Series 1:250,000
Anchorage (1951)
Cordova (1951)
Valdez (1952)
Seward (1951-3; prelim.comp.)
Talkeetna Mts (1951-4; pre comp.)
USGS Alaska Series 1:63,360
Chugach Mountains
Matanuska River
Knik River
Knik Gl.
Mt. Marcus Baker 13176
Mt. Witherspoon 12023
Mt. Glenn 9601
Mt. Goode
Mt. Gilbert
Mt. Muir
Harvard Gl.
Yale Gl.
Columbia Gl.
Matanuska Gl.
Nelchina Gl.
Tazlina Gl.
Tazlina Lake
Klutina Lake
Stephens Gl.
Valdez Gl.
Shoup Gl.
Tonsina Gl.
Tonsina Lake
Port Valdez
Valdez
Valdez Arm
Columbia Bay
College Fjord
Barry Arm
Port Wells
Unakwik Inlet
Passage Canal
Whittier
Portage
Girdwood
Turnagain Arm
Eklutna Lake
Eagle River
Palmer
Matanuska
Glenn Highway
Port Fidalgo
Marshall Pass
Thompson Pass
Lowe River
Cordova Airport
Rude River
Eastern Chugach
Bering


B
CHUGACH MOUNTAINS
Area of figure

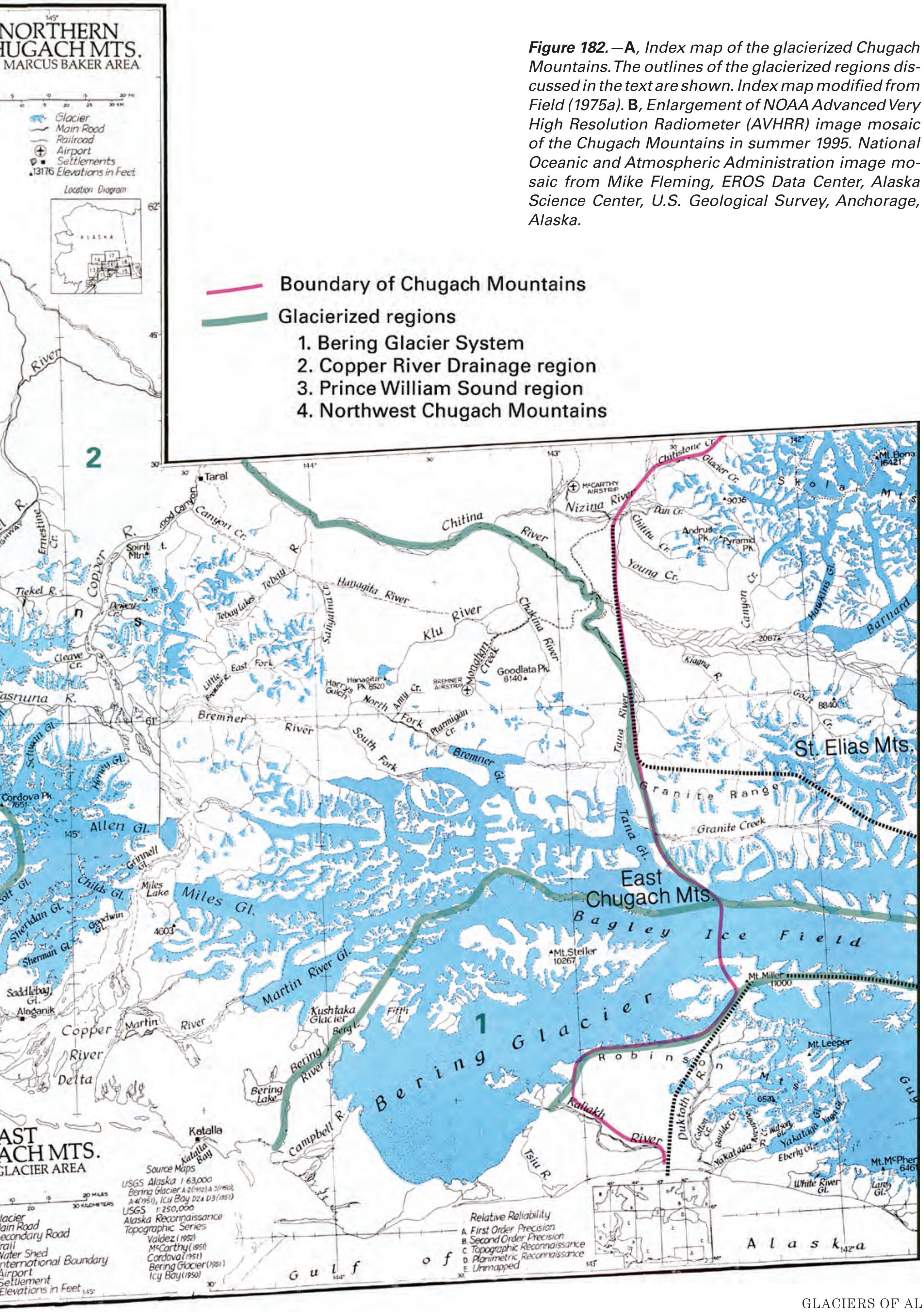
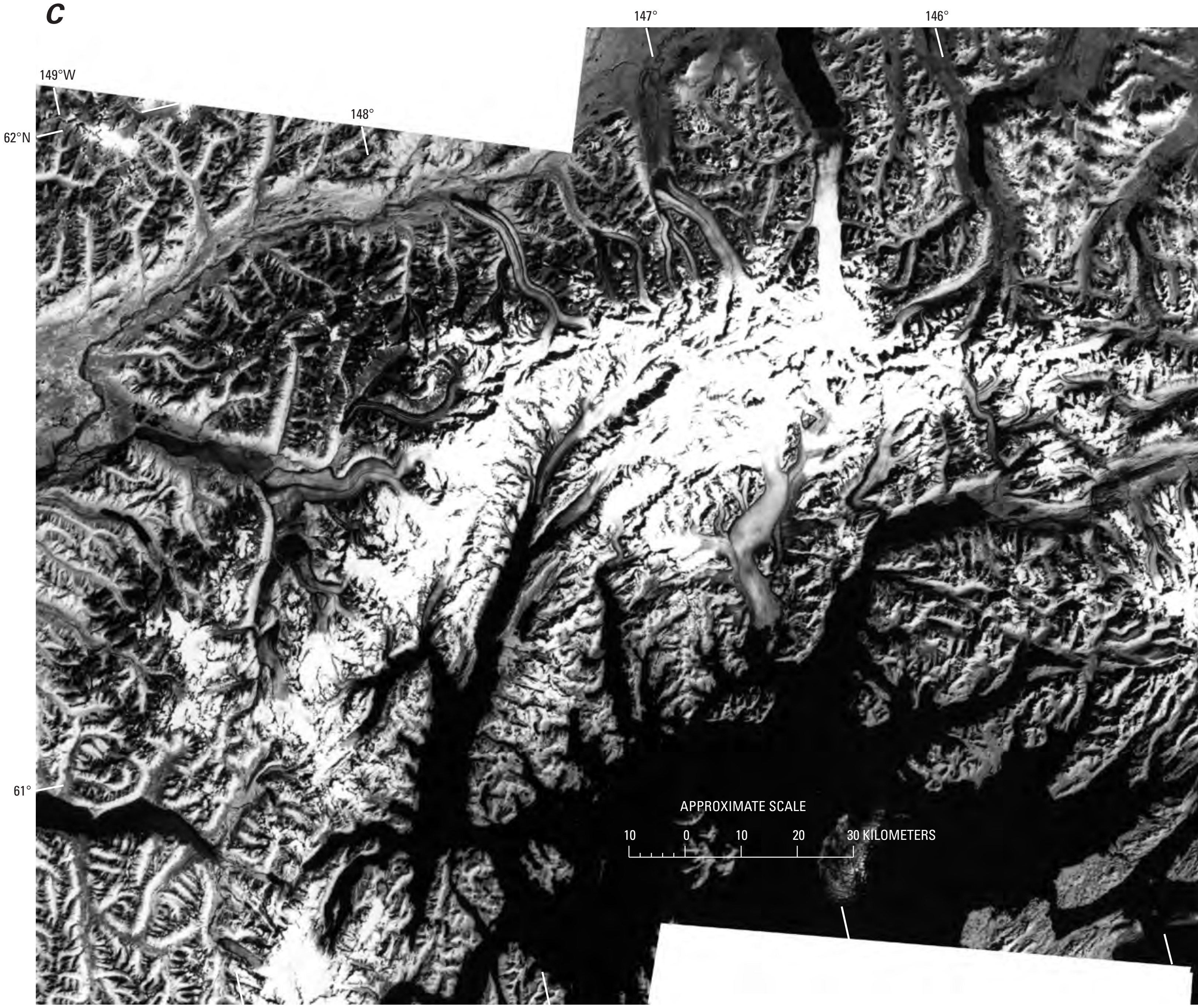


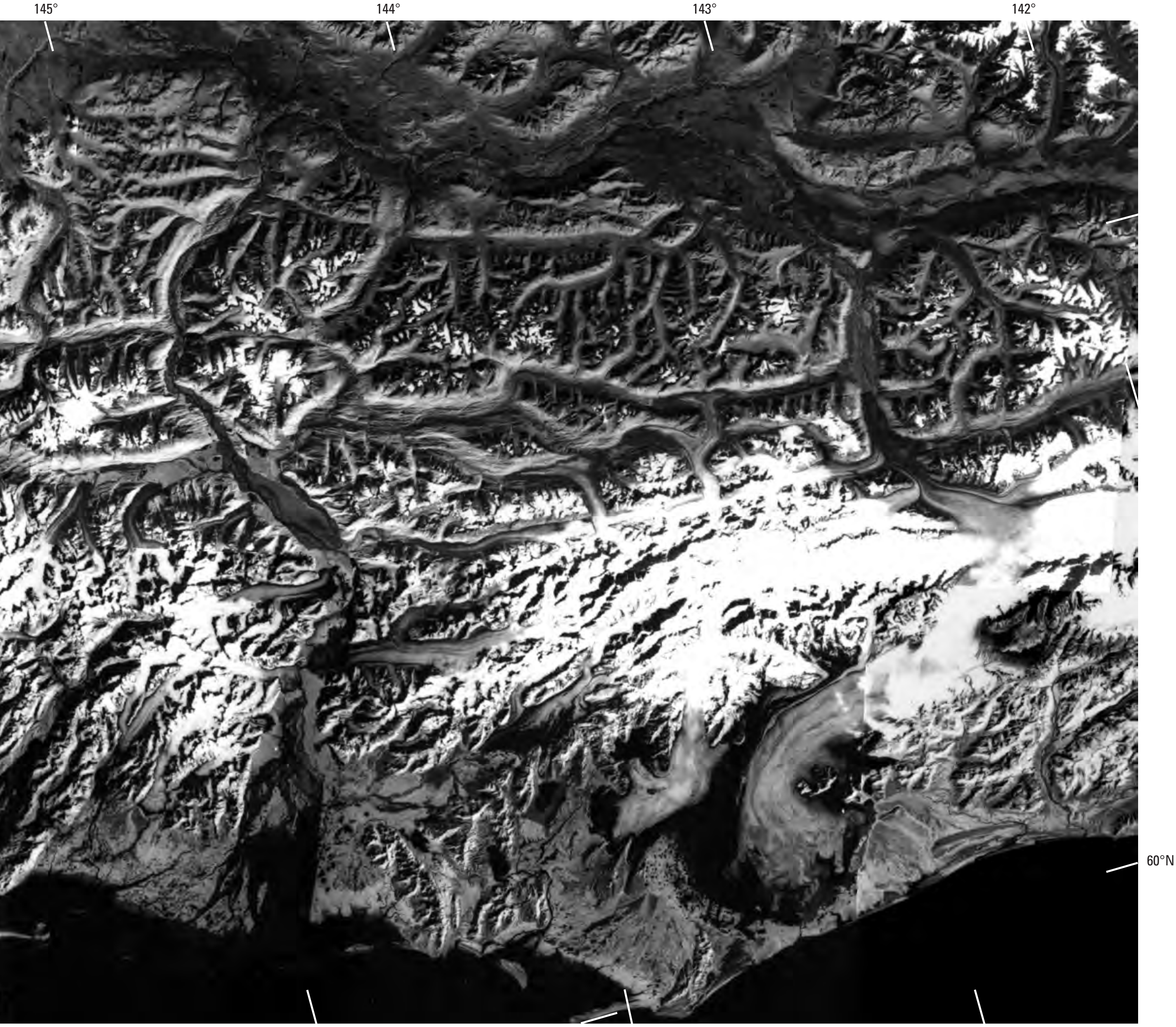
Figure 182.—**A**, *Index map of the glacierized Chugach Mountains. The outlines of the glacierized regions discussed in the text are shown. Index map modified from Field (1975a).* **B**, *Enlargement of NOAA Advanced Very High Resolution Radiometer (AVHRR) image mosaic of the Chugach Mountains in summer 1995. National Oceanic and Atmospheric Administration image mosaic from Mike Fleming, EROS Data Center, Alaska Science Center, U.S. Geological Survey, Anchorage, Alaska.*



Tana River–Tana Glacier subdivision, located east of the Copper River; the Bremner River–upper Copper River–Chitina River–Tana River subdivision, located east of the Copper River; the western Copper River Delta–lower Copper River–Tasnuna River subdivision on the west side of the Copper River; and the Cleave Creek Glacier–upper Copper River– Stephens Glacier–Tonsina Glacier Northwestern subdivision.

The Prince William Sound Segment is bounded on the east by an irregular border running from Heney Range to the divide between the Scott Glacier and the Rude River, including Cordova Peak, to the divide between the Cordova Glacier and the north-flowing glaciers to its north, to the divide between Marshall Glacier and the unnamed glacier east of Deserted Glacier, to Marshall Pass, to Thompson Pass, to the valley north of Keystone Glacier, to the perimeter of the basin of Camicia Glacier, to Mount Schrader, west of Valdez Glacier in the Chugach Icefield; on the north by the sinuous ridge that connects Mount Schrader, Mount Mahlo, Mount Cashman, Mount Shouplina, Madean Peak, Tazlin Tower, Flat Top Peak, Mount Haley, Mount Fafnir, Mount Thor,

Figure 182.—**C**, *Landsat MSS image mosaic of the Chugach Mountains. Landsat images (2956–19350, band 7; 4 September 1977; Path 68, Row 18; 1422–20212, band 7; 18 September 1973; Path 70, Row 17; 2976–19452, band 7; 24 September 1977; Path 70, Row 18; 30209–20233, band 7; 30 September 1978; Path 72, Row 17; and 30175–20345, band 7; 27 August 1978; Path 74, Row 17) are from the U.S. Geological Survey, EROS Data Center, Sioux Falls, S. Dak.*



and a number of unnamed peaks between Mount Thor and Mount Marcus Baker; on the west by the sinuous ridge that connects Mount Marcus Baker and Mount Gilbert with Passage Canal, west of Billings Glacier; and on the south by Prince William Sound. This area has two subdivisions: the Heney Range to the eastern side of Valdez Arm subdivision and the Northern Prince William Sound subdivision. Glaciers on Montague Island, the largest island in Prince William Sound, are described in the Kenai Mountains section.

The Northwestern Chugach Mountains Segment of the *Chugach Icefield* is bounded on the east by Hallet River and Klutina Lake and River and on the south by the divide between Stephens and Tazlina Glacier and by the sinuous ridge that extends west of Mount Shouplina and connects Madean Peak, Tazlin Tower, Flat Top Peak, Mount Haley, Mount Fafnir, Mount Thor, and a number of unnamed peaks between Mount Thor and Mount Marcus Baker. Then the boundary continues to the southwest, connecting with Mount Gilbert and Passage Canal, extends along Passage Canal, and continues to the north of Portage Glacier and along Turnagain Arm. The region is bounded on the west by Knik and Matanuska Rivers and on the north by the Matanuska River, Eureka Creek, and Nelchina River. This segment contains some of the larger valley glaciers of the Chugach Mountains, including Knik, Matanuska, Nelchina, and Tazlina Glaciers. This area has two subdivisions: the north-flowing, large valley glacier subdivision; and the Turnagain Arm–Western Chugach Mountains subdivision.

Landsat MSS images that cover the Chugach Mountains have the following Path/Row coordinates: 68/18, 69/17, 69/18, 70/17, 70/18, 71/17, 71/18, 72/17, 72/18, 73/17, and 74/17 (fig. 3, table 1). The Chugach Mountains contain several thousand glaciers.

Bering Glacier System Segment

Introduction

Bering Glacier is one of the most intensively studied glaciers in Alaska. In addition to historical descriptions, much of what we know about the piedmont outlet glacier is derived from the interpretation of an extensive time series of remotely sensed data sets that include the following: (1) more than 75 vertical and oblique aerial photographic data sets, containing about 10,000 photographs, acquired between 1938 and 2002; (2) more than 50 multispectral scanner (MSS), thematic mapper (TM), and enhanced thematic mapper (ETM+) Landsat images (1972 –2002); (3) about 50 digital satellite, space shuttle, and SLR or SLAR images (1978–95); and (4) about 35 hours of airborne video (1989–2000). Additionally, field observations and ice-surface measurements have provided complementary information that enhances the interpretation of remotely sensed data sets, including (1) mapping of glaciological features by the author (1974–2002); (2) time-lapse photography from in place camera systems during the latest surge (1993–96); (3) sequential photography from reoccupation of marked photo stations (1948–2002); (4) discharge measurements and flow information from telemetering stage recorders (1991–95); (5) ice-penetrating radar surveys (1990–93); (6) high-resolution marine seismic-reflection surveys of Vitus Lake, the glacier's principal ice margin lake (1991 and 1993); (7) seismic-refraction surveys of the glacier's outwash plain (1991–93); (8) dendrochronological and tree coring studies (1976–2001); (9) monitoring of movement and erosion stakes at selected terminus sites (1993–96); (10) precision location of features using differential GPS (1992–2001); and (11) sampling for water chemistry and suspended sediment load (1976–80, 1993, 1995). Lastly, a complete topographic and image map base exists for monitoring and recording changes. Topographic mapping by USGS field parties began during the first half decade of the 20th century. Subsequently, the USGS (1959, 1984) prepared 1:250,000-scale (appendix A) and 1:63,360-scale (appendix B) topographic quadrangle maps of the entire glacier using 1957 and 1972 aerial photography. The BLM produced a satellite image base map at 1:100,000-scale and a set of individual 1:63,360-scale quadrangle maps based on a 19 June 1991 Landsat TM acquisition.

The Bering Glacier System (figs. 5, 183) is one of the largest glacier systems in continental North America, with an approximate area of 5,200 km^2 (Viens, 1995; Molnia, 2001, p. 73) (table 2); it is the largest surging glacier known on Earth, outside the Greenland and Antarctic ice sheets. The eastern part of the Bering Glacier System originates in Canada in the St. Elias Mountains, at an elevation above 5,000 m, on the western flanks of the Mount Logan–King Peak massif about 25 km east of the U.S. border [see Clarke and Holdsworth 2002b]. Bering Glacier covers more than 6 percent of the glacierized area of Alaska. The ice flows west-southwestward for more than 120 km through a 7.5- to 12.5-km-wide subglacier valley in the Chugach Mountains. Before turning to the southwest, it flows through a 50-km long by 10-15-km-wide subglacier central valley and then joins the Steller Glacier to form an outlet piedmont lobe with a diameter of more than 30 km (fig. 183). At 191 km, Bering Glacier is the longest glacier in continental North America. All parts of Bering Glacier lie within 100 km of the Gulf of Alaska. Part of Jefferies Glacier, which is located in the valley north of the Bagley Ice Valley, flows to the south and contributes ice to the eastern Bagley Ice Valley. The remainder of the ice flows to the west and northwest and is the primary source of Tana Glacier.

Bering Glacier has an accumulation area of approximately 3,200 km² (Viens, 1995) (table 2), most of which is in the Bagley Ice Valley. The ablation area, which includes the southwestern part of the Bagley Ice Valley, the

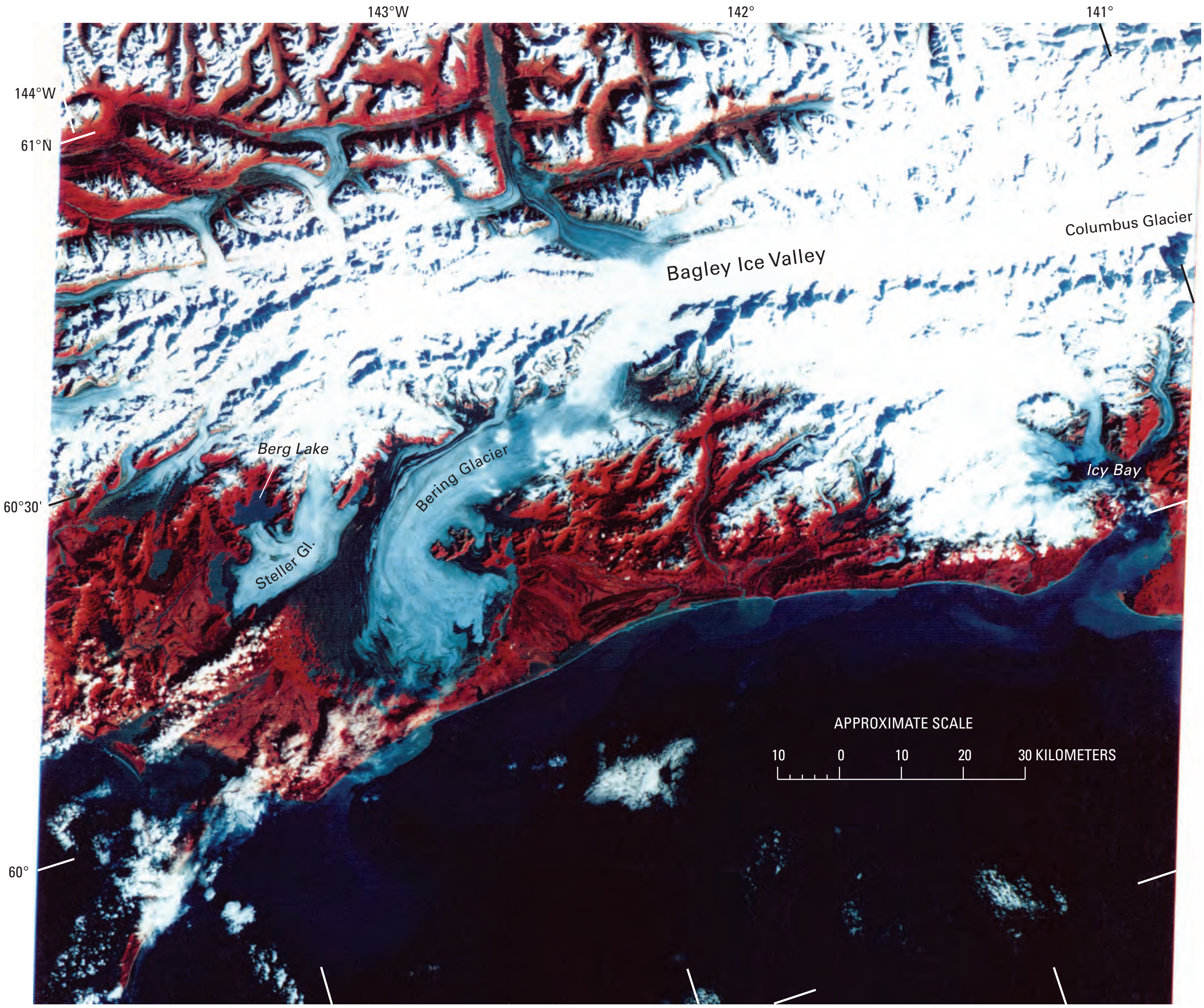


***Figure 183.**—Annotated Landsat 2 MSS image of the Bering Glacier System and Icy Bay on 23 September 1977. Bering Glacier is the largest (~5,200 km²) and the longest (~200 km) glacier in continental North America. Its accumulation zone begins north of Mount St. Elias (off the right side of this image; see fig. 109) and includes the Columbus Glacier and Bagley Ice Valley. The Bering Glacier has been thinning in recent decades, and Berg Lake and many other proglacial lakes forming around the Bering Glacier's piedmont perimeter have increased in size as the thinning progresses. The retreat has been complicated by surges, which occurred in 1957–60, 1965–67, and 1993–95, and displaced parts of the terminal lobe as much as 13 km (Post, 1972). The glaciers of Icy Bay (see St. Elias Mountains section of this volume) began retreating during the first decade of the 20th century. Landsat image (2975–19394, bands 4, 5, 7; 23 September 1977; Path 69, Row 18) and caption courtesy of Robert M. Krimmel, U.S. Geological Survey.*

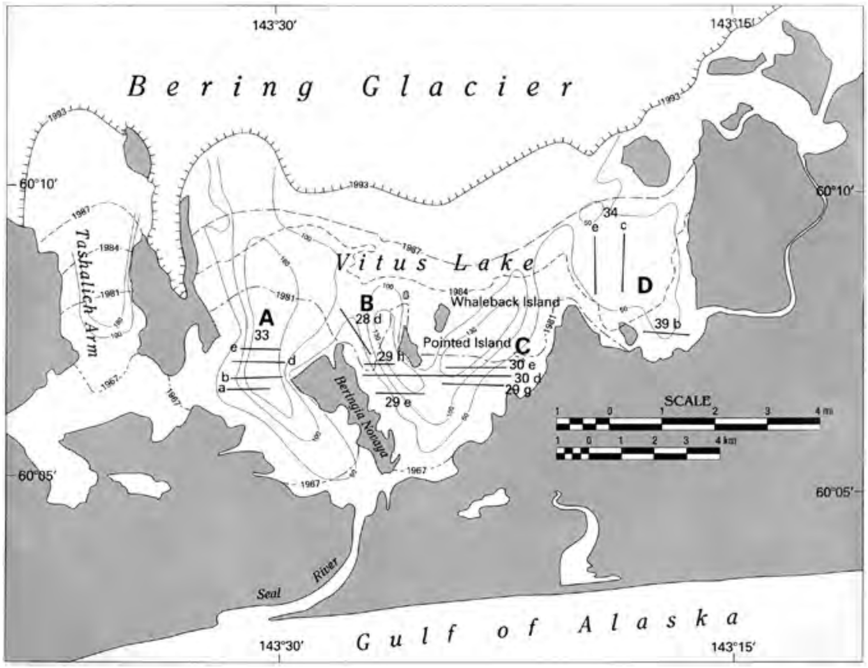


Figure 184.—Map of Vitus Lake and the 1993 margin of Bering Glacier (hachured line) showing bathymetry of the lake (in meters), positions of the retreating terminus of the glacier from 1967 to pre-surge-1993 (dashed lines), and the location of 1993 seismic profiles (short straight lines) (Molnia and others, 1996).

Bering Lobe, and a small segment of Steller Lobe and Steller Glacier, has an area of approximately 2,000 km^2. On the basis of observations made in the late 1980s and early 1990s, the ELA for the glacier ranges between 915 and 1,070 m. The *Piedmont Lobe*, with an approximate area of 900 km^2, has a maximum elevation of about 550 m and lies completely within the area of ablation. Meier and Post (1962) computed the AAR for Bering Glacier to be 0.63. Viens (1995) (table 2) calculated an AAR of 0.62 for the entire glacier, with an AAR of 0.66 for the Stellar Glacier and 0.614 for the remainder of the Bering Glacier.

The highest peaks in the St. Elias Mountains segment of the Bering Glacier System drainage are Mount Logan and Mount St. Elias. Both exceed 5,000 m in elevation. The highest peaks in the Chugach Mountains segment of the Bering Glacier drainage are Mount Miller (3,350 m), Mount Tom White (3,418 m), and Mount Steller (3,237 m).

The valley glacier section, named the Bagley Ice Valley [previously named Bagley Ice Field on maps done by Field (1975a) and Molnia (1982, 1993, 2001)], is more than 150 km in length and occupies a linear trench as much as 12 km wide (fig. 183). The eastern Bagley Ice Valley extends from northeast of the Icy Bay–Mount St. Elias region to southwest of the Tana River. The eastern margin of the Bering Lobe is fronted by Vitus Lake (fig. 184), a 20-km-long ice-marginal lake with an area of about 160 km^2. Pre-1993–95 surge bathymetric surveys of the lake's floor revealed a complex bottom morphology with at least four deep basins (informally named A, B, C, and D) having water depths of 165 m (A), 135 m (B), 135 m (C), and 85 m (D). Water depths in Tashalich Arm, the westernmost basin of the Vitus Lake system are deeper than 180 m. Two seismic reflection surveys (Carlson and others, 1993; Molnia and others, 1996) of basins A through D revealed maximum glaciolacustrine sediment-fill thicknesses of more than 110 m; depths to acoustic basement in the basins reached a maximum of 275 m below sea level (Molnia and others, 1996).

Ice Penetrating Radar (IPR) studies (Trabant and others, 1991; Molnia and Trabant, 1992) of the *Piedmont Lobe* and central valley surface conducted to determine ice thickness and morphology (fig. 185), reveal that the bed of the glacier extends as much as 350 m below sea level in places and that maximum glacier ice thicknesses exceed 800 m. These IPR studies suggest that much of the Bering Lobe, *Piedmont Lobe*, and central valley

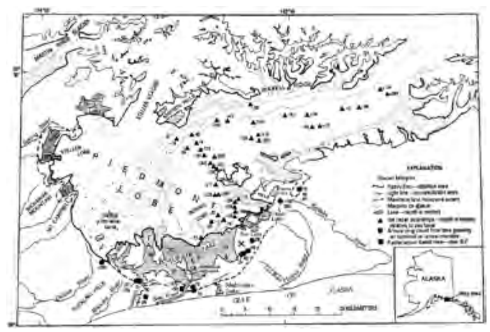


Figure 185.—Map of the lower reaches of the Bering Glacier shows geographic features and locations from which data were collected. Triangles identify locations of radio-echosounding- and precision–altimeter surveys. The numbers adjacent to the triangles are basal ice depths relative to sea level expressed in meters. Numbers in Vitus, Tsiu, and Tsivat Lakes are maximum water depths in meters. Tsiu and Tsivat Lake depths are from Fleisher and others (1993). Circles identify positions of trees growing on terminal and lateral moraines from which cores were obtained. The adjacent numbers are annual ring counts. The square adjacent to Seal River marks the location of a radiocarbon-dated spruce tree recovered from the sediments comprising the neoglacial terminal moraine. X marks the location of the Giant Log. TA is Tashalich Arm. The glacier margin and Vitus Lake are shown as they appeared in 1991. From Molnia and Post (1995, fig. 2, p. 91). A larger version of this figure is available online.

occupy a deep basin or series of channels, compatible with or even deeper than the depths determined by the high-resolution seismic reflection studies conducted in Vitus Lake. IPR measurements made as much as 60 km upglacier from the Bering Lobe terminus along the centerline of the glacier showed many areas where glacier bed depths were below sea level.

Offshore of Bering Glacier in the Gulf of Alaska is the Bering Trough (fig. 186), a sediment-floored 45-km-long submarine valley (Carlson and others, 1982). The Bering Trough has maximum water depths of 321 m,

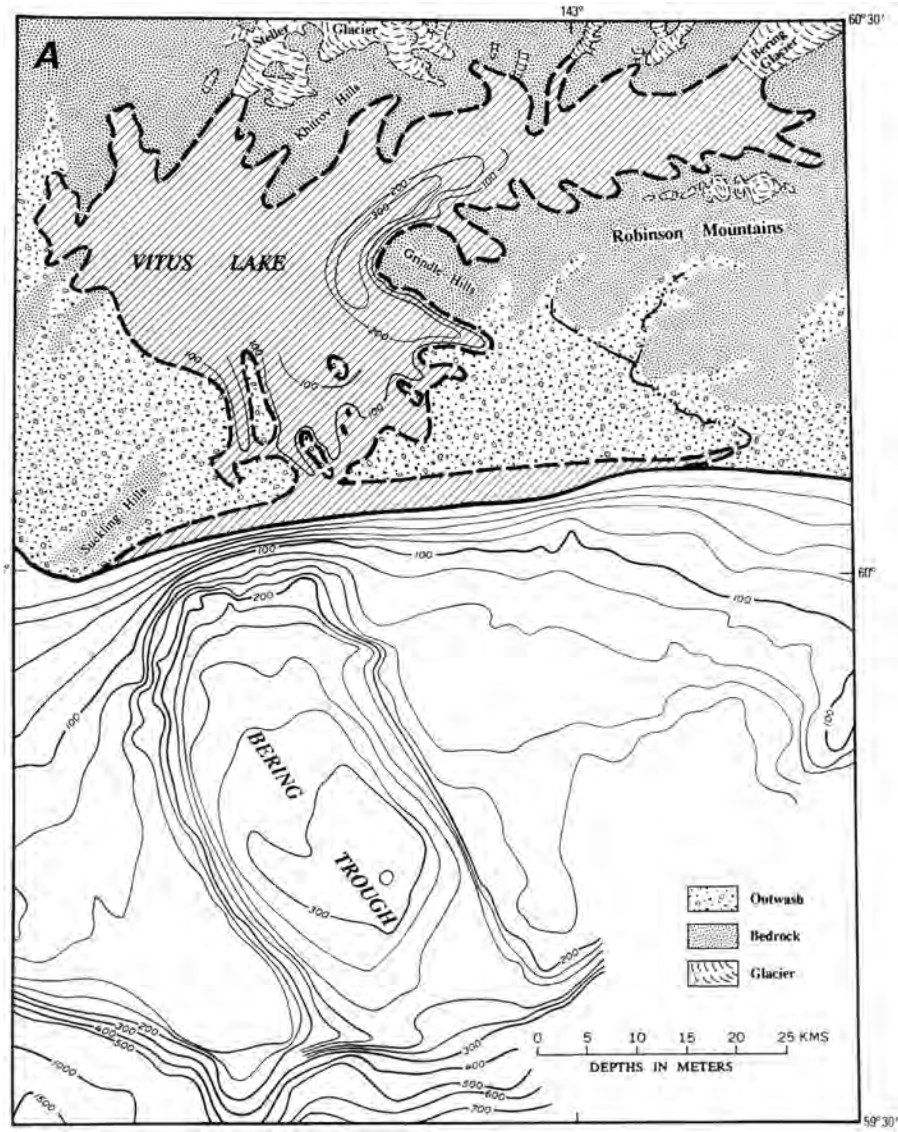


Figure 186.—**A**, *Gulf of Alaska bathymetry shows the location, morphology, and depths associated with Bering Trough. Vitus Lake is drawn to show its pre-1993 surge bathymetry (contours adjacent to its southern margin) and the maximum size that it could attain following a catastrophic retreat of the Bering Glacier. Depths adjacent to the Grindle Hills are depths below present sea level to bedrock measured by radio-echosounding traverses. The area depicted by the diagonal pattern is the area where the bed of Bering Glacier is below sea level.* **B**, *Glacially eroded morphology of the Gulf of Alaska continental shelf in the vicinity of the Bering Glacier. The trough offshore of the terminus of Bering Glacier is the Bering Trough. The numerous adjacent features attest to both the extent and the erosive power of glaciers that descended from the Chugach Mountains during the Pleistocene Epoch. Diagram by Tau Rho Alpha, U.S. Geological Survey. A larger version of this figure is available online.*

maximum bedrock depths of about 500 m, and a maximum width of 25 km. Acoustic-basement depths in Vitus Lake and IPR measurements of glacier-bed depths suggest that the sub-sea-level basins and channels that underlie Bering Glacier and Vitus Lake are landward continuations of the Bering Trough.

Vitus Lake is separated from the Gulf of Alaska by the Bering Glacier foreland, a 3-km-wide beach–outwash plain–moraine complex. The beach is composed of reworked outwash-plain sand and gravel and sand deposited by longshore transport. The outwash plain and the neoglacial moraine are composed of gravelly sand. With the exception of one small exposure of the Yakataga Formation (fig. 187) exposed along the eastern side of Tashalich Arm in 1992, there are no known bedrock outcrops in the Vitus Lake area, nor was any shallow bedrock identified in seismic-refraction profiling of the beach in 1992.

The *Piedmont Lobe* (fig. 188) consists of three main components, initially described by Post (1972). The *Piedmont Lobe* has an approximate area of 900 km^2 and comprises 17.60 percent of the total glacier system; the Bering Lobe (fig. 189), which consists of relatively debris-free, periodically surging ice, contains 6.76 percent; the Central Medial Moraine Band (CMMB), which Post (1972, p. 219) described as “A very large debris band composed of repeatedly folded medial moraines that extends across the center of the Bering Glacier lobe” (fig. 190), is composed of complexly folded,

Figure 187.*—13 August 1992 photograph of the only bedrock outcrop found along the southern perimeter of Bering Glacier. The Yakataga Formation series of beds, dipping at approximately N. 45° W., was exposed by the retreat of the Bering Glacier between 1991 and 1992. The margin of Bering Glacier can be seen in the upper left. Photograph by Bruce F. Molnia, U.S. Geological Survey.*

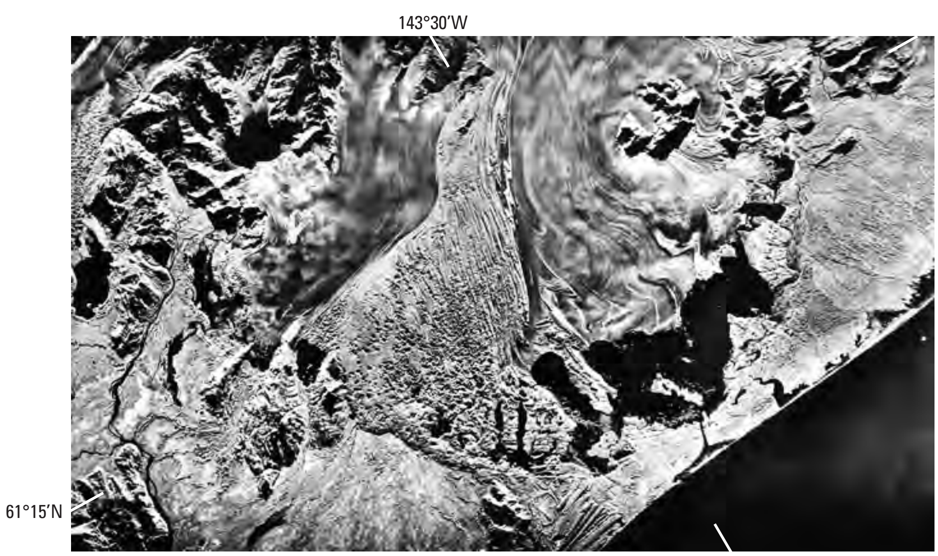


Figure 188.*—3 August 1990 X-band synthetic aperture, side-looking airborne radar image of the Bering Glacier’s* Piedmont Lobe*. The three components of the* Piedmont Lobe*, the Bering Lobe, the folded Central Medial Moraine Band, and the Steller Lobe, are visible north of Vitus Lake. About 25 folded medial moraine loops, formed by surges during the past several hundred years, can be seen in the Central Medial Moraine Band. The stagnant-ice-cored medial moraine is characterized by hundreds of thermokarst pits. From Molnia and Post (1995, fig. 3, p. 92).*

Figure 189.—*18 August 1978 AHAP false-color infrared vertical aerial photograph of part of the retreating terminus of the Bering Lobe. Nine years after the end of the 1967 surge, the terminus of Bering Glacier had retreated a maximum of approximately 1.5 km from its surge-maximum position. Two areas of the terminus are producing dozens of large icebergs. This is not a routine calving event in the true sense of the term; it is a disintegration event. Such events occur when the thinning glacier reaches a state of buoyancy and separates from its bed. As it begins to float, large pieces of ice, sometimes about 1 km in maximum dimension, separate from the terminus along old crevasse fracture planes. AHAP photograph no. L119F6063 from the GeoData Center, Geophysical Institute, University of Alaska, Fairbanks, Alaska.*

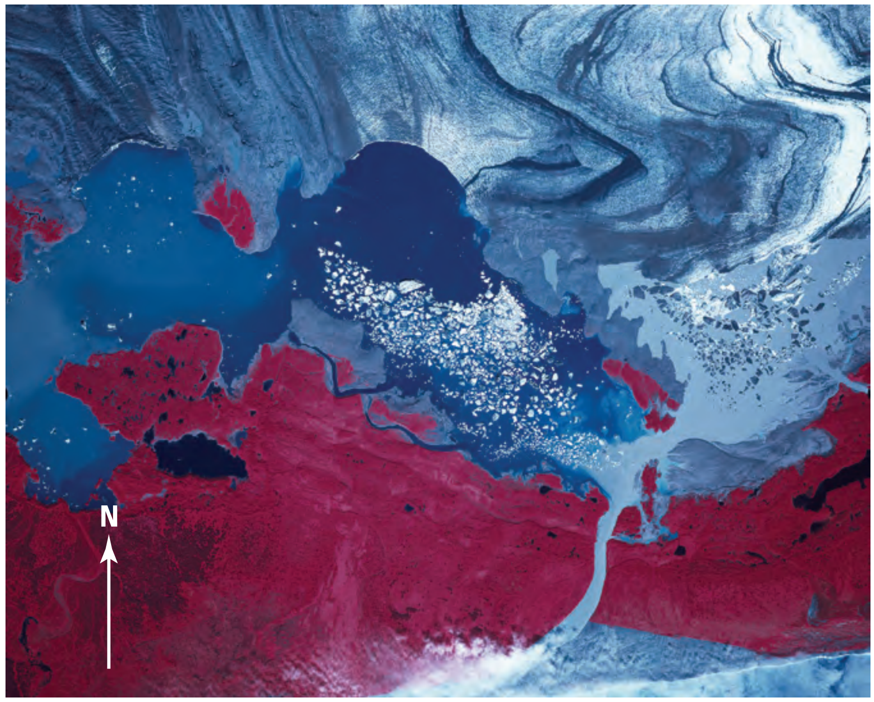


Figure 190.—*Two views of the Central Medial Moraine Band of Bering Glacier.* **A**, *Summer 1938, north-looking oblique aerial photograph of most of the northern part of the Central Medial Moraine Band. Individual folded medial moraine loops stand several meters above the bare ice surface of the glacier. Photograph no. 1894 by Bradford Washburn, Museum of Science (Boston).* **B**, *12 September 1986 north-looking oblique aerial photograph of the southwestern part of the Central Medial Moraine Band. Vegetation covers much of the southern margin of the moraine band. Most of the Steller Glacier and Steller Lobe can be seen in the background. USGS photograph no. 86–R2–229 by Robert F. Krimmel, U.S. Geological Survey.*

moraine-covered, generally stagnant ice and contains 7.36 percent; and the Steller Lobe (fig. 191), which consists of relatively debris free, active ice derived from the 50-km-long Steller Glacier and contains 3.48 percent of the total area. Tributaries of the 180 km^2 Steller Lobe cover an area of 644 km^2 (12.44 percent of the total glacier), whereas tributaries of the 350-km^2 Bering Lobe cover 3,620 km^2, or 70 percent of the total glacier area (Robert J. Viens, University of Washington, oral commun., 1993). The distance around the perimeter of the *Piedmont Lobe* of Bering Glacier, from the Bering River to the Grindle Hills, is about 75 km. Bering Glacier's *Piedmont Lobe* is about one-third smaller than that of Malaspina Glacier.

On the basis of IPR thickness measurements and bed depths (fig. 185), the CMMB appears to be composed of thin, debris-covered ice overlying a generally shallow bedrock divide between the thicker Bering Lobe and Steller Lobe.

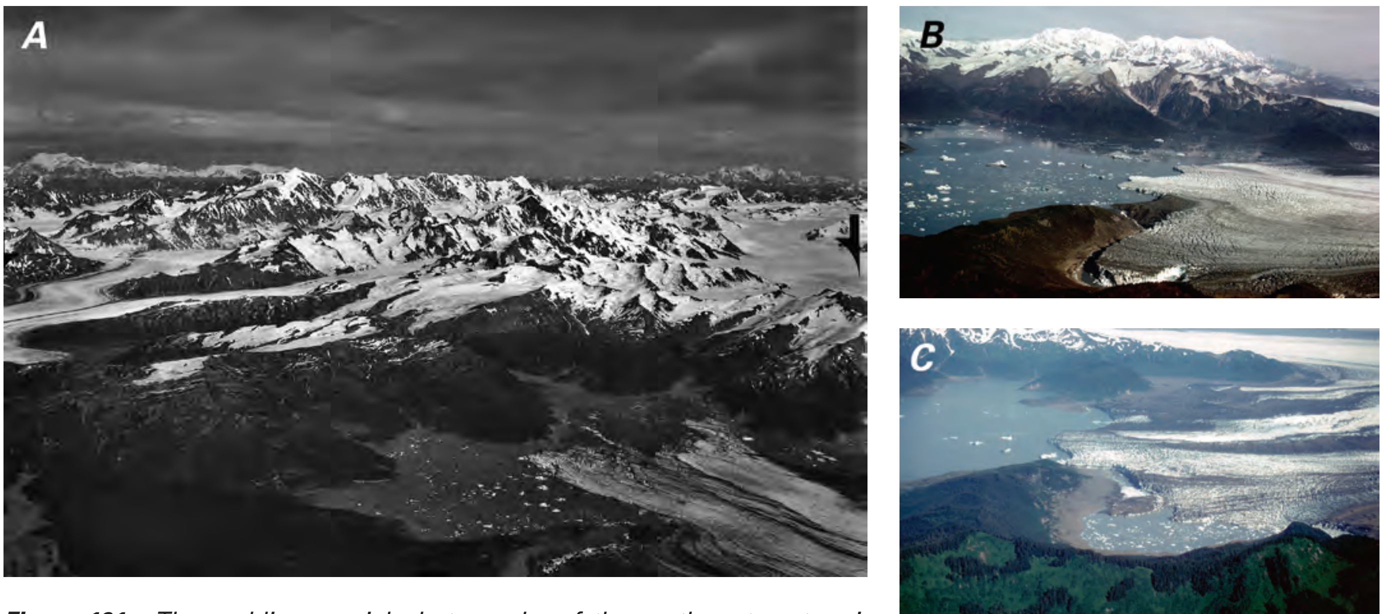


Figure 191.**—Three oblique aerial photographs of the northwestern terminus of the Steller Lobe of Bering Glacier flowing into Berg Lake.* **A, Summer 1938 north-looking photograph. The ice tongue reaches the north shore of the lake, separating it into two bodies of water. Photograph no. 2274 by Bradford Washburn, Museum of Science (Boston). Martin (1908) reported five smaller lakes here in 1905 with evidence of fluctuations in water level. By about 1940, the Steller Lobe of Bering Glacier had receded and the five lakes coalesced into one. From sometime before 1940 until 1984, there were no jökulhlaups, but Post and Mayo (1971) predicted that continued glacier thinning would allow a new jökulhlaup cycle to begin.* **B***, 6 October 1974 northeast-looking photograph of the Steller Lobe damming the southern side of Berg Lake. The Bering Glacier is an effective dam, and the only outlet for the lake is a torrent of water that overtops the low point in the unnamed ridge in the lower right of the photograph. The water runs along the margin as an ice-marginal stream. Thirty-six years earlier, the Bering Glacier covered the eastern part of the ridge. Photograph by Bruce F. Molnia, U.S. Geological Survey.* **C***, 12 August 2001 northeast-looking oblique aerial photograph of the northwestern part of the Steller Lobe flowing into the southern side of Berg Lake. In 1984, 1986, and 1994, jökulhlaups occurred and caused major flooding of wildlife habitats. Soon after June 1997, a large mass of ice grounded on the shoreline was separated from the retreating terminus of the Steller Lobe. In early 2000, a small surge began that caused a readvance of the terminus of Steller Glacier and ultimately reconnected the terminus with the grounded ice. The surge continued through the fall of 2001. The level of Berg Lake lowered through the early 21st Century and then stabilized. Photograph by Bruce F. Molnia, U.S. Geological Survey. Larger versions of* B *and* C *are available online.*

At one location within the CMMB, IPR measurements show that bedrock reaches to about 20 m above sea level and is covered by less than 100 m of ice.

An analysis of historical data includes the following sources: 18th-, 19th-, and early 20th-century exploration maps, description of voyages, and reports of expeditions; 19th century published maps and nautical charts; late 19th and early 20th century geological reports; and early field photography obtained between 1897 and 1905. These sources provide additional information about the history of and changes to Bering Glacier over time and give a much longer term perspective on the glacier's post-"Little Ice Age" history than present-day measurements or remotely sensed data sets do. For example, a description of the physical characteristics of the surface of Bering Glacier from the late 1830s by Belcher (1843) suggests that the glacier was surging at the time of his observation. Similarly, photographs obtained by the Duke of Abruzzi's 1897 climbing expedition and a 1905 USGS geological field party clearly show the characteristics and position of parts of the margin of Bering Glacier. All of these historical data are useful in understanding the behavior of the glacier in the past.

Pre-20th Century Observations of Bering Glacier

Through the middle of the 1880s, the geography of the southeastern Chugach Mountains was very poorly known. In 1886, Lieutenant H.W. Seton Karr, a member of the New York Times Expedition, applied the name Bering Glacier to the "ice-plain" that he observed west of Icy Bay during an unsuccessful attempt to reach the summit of Mount St. Elias (Seton Karr, 1887). He was knowledgeable about glaciers and realized the significance of his discovery. Seton Karr wrote, "In this direction (west) the 'foothills' of Elias stood like islands in the enormous expanse of glacier stretching prairie-like as far as the eye could penetrate through the crystalline air towards the country of the *Atna* [editors' italics] or Copper River" (Seton Karr, 1887, p. 109–110).

Several weeks later, in early August 1886, as he sailed along the Gulf of Alaska coast more than 100 km west of Icy Bay, Seton Karr observed (1887, p. 139–140),

> I had understood that with Icy Cape the last ice along the coast was left behind. But looming twenty miles or so to the westward appears another vast ice-plain ... which sweeps down and opens fan-like on the ocean, where the coast range of "foot-hills" comes to an end. It is evidently the opening or outlet of the vast glacier-desert or ice-lake which we saw from the slopes of Mount St. Elias, lying to the northwest of the mountain. Its birthplace is an icy range that forms an enlarged continuation of the great western ridge of Elias. It is not marked or mentioned by the early navigators, all of whom mistook the true nature of these stupendous glaciers. La Perouse describing them as "snow lying upon a barren soil," and "a plain totally destitute of verdure"

Figure 192.—Part of a late 19th century U.S. Coast and Geodetic Survey chart of the Gulf of Alaska region showing only the edge of the piedmont lobe of the essentially unknown and unexplored Bering Glacier. A larger version of this figure is available online.

Six years earlier, in 1880, the Coast and Geodetic Survey (C&GS) had named this virtually unknown and unexplored coastal glacier Bering Glacier. The C&GS had chosen the name to commemorate Captain Vitus Bering, a Danish sea captain in the Russian naval service of Czar Peter the Great, who led a voyage of exploration that made a landfall in the vicinity of the glacier's terminus in 1741. As late-19th century C&GS nautical charts (fig. 192) and USGS topographic maps clearly show, nothing was known about the size, shape, or geography of most of Bering Glacier.

In 1890 and again in 1891, USGS expeditions led by Israel C. Russell attempted to climb Mount St. Elias. Although both attempts were unsuccessful, each reached elevations where Russell was able to see the upper parts of Bering Glacier. On the first, Russell observed part of what we now know as the Bagley Ice Valley, which he described as "another vast glacier extending westward to the limits of vision." (Russell, 1891, p. 141).

On his second expedition in 1891, Russell observed the source area of Bering Glacier to the north of Mount St. Elias. He wrote (Russell, 1893, p. 47):

> I was now so near the crest of the divide that only a few yards remained before I should be able to see the country to the north, a vast region which no one had yet beheld. ... I expected to see a comparatively low, forested country, stretching away

to the north, with lakes and rivers and perhaps some signs of human habitation, but I was entirely mistaken. What met my astonished gaze was a vast snow-covered region, limitless in expanse, through which hundreds and perhaps thousands of barren, angular mountain peaks projected. There was not a stream, not a lake, and not a vestige of vegetation of any kind in sight. A more desolate or more utterly lifeless land one never beheld. Vast, smooth, snow surfaces, without crevasses stretched away to limitless distances, broken only by jagged and angular mountain peaks.

At the beginning of the 20th century, Bering Glacier had nearly a dozen major drainage outlets, arranged in a radial pattern along the piedmont outlet glacier's margin. These included: Kaliakh River, the many branches of the Tsiu and Tsivat Rivers, Midtimber River, Seal River, Tashalich River, Kiklukh River, Oaklee River, Edwardes River, Campbell River, Nichawak River, Gandil River, and Bering River (Molnia and Post, 1995, p. 104, fig. 9). A significant retreat of Bering Glacier began during the first decade of the 20th century, sometime between 1905 and 1910. Since then, it has lost nearly 100 km^2 of ice from the terminus area of the *Piedmont Lobe* (fig. 193). Loss of ice by melting and calving along the terminus resulted in the development of a series of lakes along much of the glacier's perimeter (figs. 36, 194; see also a 9 June 1948 oblique aerial photograph, USGS photograph no. 119LT–72PL–R–8M–148–72RS–GS6B). Field (1975) reported that, between 1905 and 1957, as the glacier retreated back from its "Little Ice Age" maximum moraine, the percentage of the perimeter of Bering Glacier ending in ice-marginal lakes increased from about 10 percent to about 54 percent. By the early 1920s, with continued melting and retreat, several of these lakes expanded and coalesced to form Vitus Lake.

Retreat along the margin, interrupted by a series of 20th century surges, has resulted in an areal expansion of all the ice-marginal lakes. Along the margin of the Bering Lobe, the result has been an expanded Vitus Lake. To the west, along the eastern margin of the CMMB, a pair of parallel linear lakes began to develop beginning in the 1940s. Formed initially by the coalescence of thermokarst pits, these lakes have continued to lengthen and widen through the early 21st century, with the eastern lake reaching a length of approximately 5 km. These lakes and several other features of Bering Glacier have been used as experimental test sites to compare the use of spaceborne

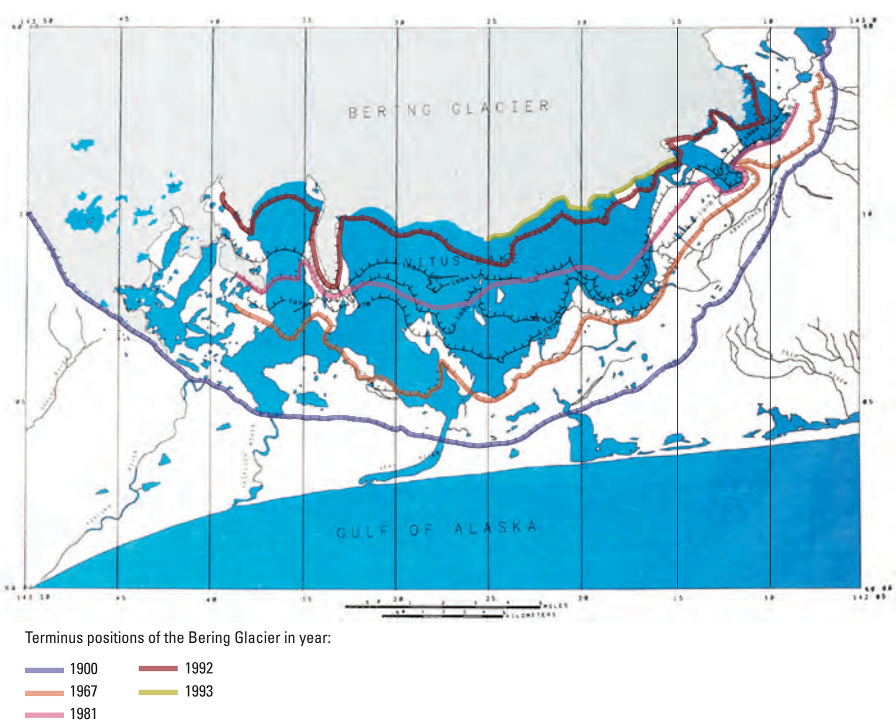


Figure 193.—*Map showing positions of the terminus of Bering Glacier between 1900 and 1993. The 1993 position shown is the pre-surge (pre-28 August 1993) position of the terminus.*

Figure 194.—12 August 1961 oblique aerial photograph of the eastern margin of the Bering Glacier showing development of ice-marginal lakes. Large icebergs can be seen separating from the disintegrating margin of Bering Glacier. View looking westerly across the southeastern part of the terminus shows a continuous ice-marginal lake of variable width fronting Bering Glacier. Photographed a year after a surge, the surging glacier had almost filled the lake northeast of the Seal River. The Seal River is at the center-left edge of the photograph. Photograph by Austin Post, U.S. Geological Survey. A 1938 oblique aerial photograph can be seen in figure 36.

Figure 195.—17 August 1979 oblique aerial photograph showing post-1967 surge retreat of the terminus of Bering Glacier. North-looking view of the southern part of Vitus Lake from offshore, east of the mouth of Seal River. In the 12 years after the end of the surge in 1967, the glacier has retreated about 3 km and thinned significantly. Bering Glacier has a very low gradient and almost no relief at its southern margin. The two large plumes of icebergs are the result of calving from rapidly disintegrating parts of the terminus of the glacier on either side of the island. Photograph by Bruce F. Molnia, U.S. Geological Survey.

ERS-1 and SIR-C synthetic aperture radar (SAR) with airborne SAR in identifying different types of glaciological features (Molnia and Molnia, 1995).

At the beginning of the 21st century, more than 90 percent of the meltwater from the Bering Lobe and the CMMB flows through Vitus Lake, entering the Gulf of Alaska through Seal River. Seal River, a 6-km-long low-gradient tidally dominated stream, flows diagonally across Bering Glacier foreland from east to west (figs. 195, 207). On 28 June 1991, the discharge of the Seal River, measured where it exits Vitus Lake, was approximately 1.4×10^3 $m^3 s^{-1}$.

20th Century Observations of Bering Glacier

G.C. Martin, a USGS geologist who mapped the Bering River coal field—a large undeveloped coal reserve located on the western side of the glacier—presented the earliest 20th century description of Bering Glacier. Martin (1905, p. 17) described the Bering as:

> a large glacier of the piedmont type, and, if considered independently, is second in size, among glaciers of this type, only to the Malaspina Glacier. It is a large field of stagnant ice which has overflooded the eastern extension of the zone of coastal foothills...and it is considered by many as merely the western lobe

> of the Malaspina Glacier. It is however, in all probability entirely separate from the latter. A great many valley glaciers coming from the Chugach Range enter it as tributaries. It is fringed along its southwestern border by a wide moraine, while the ice itself is thickly covered by rock debris for a distance of several miles from its front, and, as is in the case of the Malaspina Glacier, this covering is so thick that it is often impossible to determine the margin of the glacier.

Several of his photographs document that retreat of the Steller Lobe was underway before 1905 and that at least one ice-marginal lake had already formed (see USGS Photo Library Martin 245 photograph).

Three years later, Martin (1908) published a second description of the glacier accompanied by a 1:200,000-scale topographic reconnaissance map and a detailed 1:62,500-scale topographic map, both dated 1907, as well as a geologic map of the glacier's western margin (Martin and others, 1907). Martin (1908, p. 16) wrote: "Bering Glacier...is a huge, even-surfaced, stagnant mass of glacial ice, which is fed by many valley glaciers coming from the high mountains north of it. It is a piedmont glacier of the same general character and of about the same size as Malaspina Glacier. Portions of its surface form a good highway to some of the coal camps in the east end of the Bering River coal field, but most of it is so covered by irregular masses of rock and gravel and so much crevassed that travel over its surface is difficult and dangerous." The crevassing described above by Martin (1908) may be the result of a surge that occurred about 1900. The surface elevation of the glacier's margin, inferred from the 1907 topographic map, was quite similar to the Holocene thicknesses of Bering Glacier described by Miller (1958, 1961).

Berg Lake

On the northwestern margin of Bering Glacier, Berg Lake (also called Berg Lakes on some maps) is the largest (approximate area 30 km^2) of the ice-marginal lakes formed by the retreat of the Steller Lobe (fig. 191). In 1905, Martin (1908) mapped five separate smaller lakes, many of which showed evidence of fluctuating water levels. He reported maximum water depths of approximately 250 m and noted that the water level had recently been higher. The existence of these five lakes suggests that, by 1905, the Steller Lobe was already in retreat from its "Little Ice Age" maximum position. By the early 1940s, continued retreat of the Steller Lobe resulted in the five lakes coalescing into one. Continued retreat of the Steller Lobe, interrupted by several small pulses or minisurges such as one that occurred in 2000 and 2001, has resulted in the continued expansion of Berg Lake. Post and Mayo (1971) noted that in 1970, the lake was 207 m deep, about 40 m lower than the 1905 level, and that the area of Berg Lake had increased from about 12 km^2 in 1905 to about 28 km^2 in 1970. These changes are the result of the ongoing thinning and retreat of the Steller Lobe.

Berg Lake has a history of at least three late 20th century glacier outburst floods (jökulhlaups). In 1984, 1986, and 1994, jökulhlaups from Berg Lake caused major flooding in the Bering River drainage and severely impacted wildlife habitats. The May 1994 jökulhlaup resulted from the catastrophic draining of partially ice-dammed Berg Lake. The development of an approximately 500-m-long subglacier channel through the northwestern margin of the Steller Glacier resulted in about a 100-m lowering of the lake in a 72-hour period. An estimated 5.5×10^9 m^3 of water escaped the 9×6-km Berg Lake and drained through the Bering River during the ensuing 72 hours, completely inundating the floor of the Bering River valley from valley wall to valley wall. Because May is the time of moose calving and prime migratory bird nesting activity, especially Dusky Canada Geese (*Branta canadensis occidentalis*) and Trumpeter Swans (*Olor buccinator*), the flood had a major impact on the local ecosystem. Post and Mayo (1971) predicted that continued thinning of the glacier would result in a cycle of repeated jökulhlaups.

During August 2001, Josberger and others (2001) measured the bathymetry, temperature, and conductivity of Berg Lake. They determined that the intense vertical convection in the lake is controlled by suspended sediment. The temperature profiles from Berg Lake show a vertical structure that consists of a 10-m thick surface layer, where the temperature drops from near 9°C to approximately 4°C, the temperature of maximum density for fresh water. Below this depth, the temperature decreases to 0°C in the deepest portions of the lake, at a depth of around 75 m. Superimposed on this general unstable temperature profile are spatially variable fine-structure details that include vertical steps and temperature inversions. Although the temperature profiles indicate a highly unstable situation, the subglacier discharge has a suspended sediment load sufficient to marginally stabilize the density structure in the lake. This sediment-laden water flows out from below the glacier and spreads horizontally throughout Berg Lake. As the suspended sediment settles, vertical thermal convection yields the observed fine structure in the temperature profiles.

Surges of the Bering Glacier

The 20th century retreat of Bering Glacier has been interrupted by at least six major episodes of surging. Although rapid retreat and surging may seem incompatible, the results of these two processes have been large-scale fluctuations in the position of the terminus of Bering Glacier and in the continuing drawdown of ice from the accumulation area of the glacier. The surges occurred in: ca. 1900, ca. 1920, ca. 1938–40, 1957–60, 1965–67 (Post, 1972), and 1993–95, an approximate average of 20 years between surges (Post, 1972). The surges that occurred during the second half of the 20th century have been closely monitored. Each surge event resulted in a rapid and significant advance of the terminus of Bering Glacier and was accompanied by the transfer of a significant volume of ice into the expanding terminus area. A rapid retreat of the terminus, significantly enhanced by large-scale calving into Vitus Lake, and disintegration upglacier from the terminus also followed each surge event. The result was a glacier that was much thinner than before the surge.Far more ice was removed from Bering Glacier than melting alone could have done. The two most recent surges (1957–67, 1993–95) consisted of a pair of ice-displacement events separated by periods of stagnant ice. In the earlier surge, ice-displacement occurred from 1957 to 1960 and from 1965 to 1967, and stagnant ice occurred from late 1960 to early 1965. Maximum ice displacement exceeded 10 km (Post, 1972).

Post (1960, 1965, 1967b, 1969, 1972), Meier and Post (1969), and Post and LaChapelle (1971) have shown from direct observations of many surging glaciers that the folding (or contortion) of surface medial and lateral moraines is an expected result of surging. From analysis of vertical aerial photography, Post (1972) determined that individual chevron-folded medial moraines are the product of a separate surge event. He wrote that (Post, 1972, p. 219):

> Vertical aerial photography taken before and after surges disclose the direction and magnitude of ice flow in various parts of the piedmont lobe. The ice moved toward the terminus and expanded in a normal, radial pattern with no evidence of unusual shearing that would result in the formation of large folds. Many surging glaciers display repeated lateral displacements in their medial moraines which result from periodic surging of the main glacier past non-surging tributaries. Moraines of the Bering Glacier display small periodic irregularities of this nature. The large "accordion" folds in the moraines in the Bering Lobe are judged to be due to the combined effects of compressive flow and lateral or transverse expansion of those previously formed irregularities. The initially small pre-existing perturbations in the moraines are simply spread laterally and shortened radially into large folds as the ice spreads out.

At Bering Glacier, field observations and analysis of aerial photography and SAR images indicate that at least 25 folded moraines are located in the CMMB. With a 20- to 30-year cycle (Post, 1972), at least 500 to 700 years of periodic surging would be necessary to create the observed number of folded moraines (fig. 190).

The 1957–60, 1965–67, and 1993–95 Surges of the Bering Glacier

1957–60 and 1965–67 Surges

Following a surge that ended about 1940, Bering Glacier began to retreat. Retreat continued through 1957; Vitus Lake expanded to an approximate size of 14×5 km and an approximate area of 48 km^2. D.J. Miller's 1958 aerial photograph of a new surge developing in the eastern Bagley Ice Valley showed that the surface of Bering Glacier was lowering and that recent shearing of the glacier's margin had occurred at Waxell Ridge. The surge continued through 1960, probably reaching the terminus in 1959. Much of the margin of Bering Lobe advanced and ice reoccupied much of Vitus Lake. By 1960, the terminus of Bering Lobe had advanced to a position from 0.5 to 3.0 km beyond the 1948 margin, reducing Vitus Lake to a maximum of approximately 3.5×2.5-km and an approximate area of 3 km^2. Post (1972) reported maximum displacements of as much as 9 km that probably occurred during a 12- to18-month period. By the fall of 1960, no further terminus activity was observed. The position of the terminus changed little through the spring of 1963 (Molnia and Post, 1995, fig. 13); recession was underway by summer 1963.

A major earthquake occurring on 27 March 1964 produced many rock avalanches that fell on the surface of Bering Glacier. Four had lengths of more than 5 km, all in the Waxell Ridge area. The largest was 6.5 km long and as much as 2.5 km wide. It fell from an elevation of more than 2,000 m and was estimated to have a volume of 10^7 m^3 of rock (Post, 1967a).

Between 1965 and 1967, a smaller surge reactivated much of the terminus and resulted in up to 1.0 km of additional terminus advance. Ice was displaced by as much as 4.0 km (Post, 1972). By 1967, the glacier margin was within 1 to 3 km of its early 20th century maximum position (fig. 193).

Molnia and Post (1995) quantified the post-1967 rate of terminus retreat throughout the eastern margin of Bering Glacier and found that the terminus of Bering Glacier retreated drastically (maximum recession of as much as 10.7 km) between the end of the 1965–67 surge and the 28 August 1993 onset of terminus displacement associated with the 1993–95 surge. Retreat did not occur at a constant velocity because individual years and areas show considerable variability; the average rate of retreat was determined to be 0.43 km a^{-1}. Molnia and Post (1995, p. 112, table 1) presented their annual average retreat rates along four approximately north-south transects extending from the 1967 ice limit to the August 1993 pre-surge ice margin (transect locations shown by Molnia and Post, 1995, p. 111, fig. 15).

Annual recession rates were calculated along each of the transects and ranged from 0.04 to 1.0 km a^{-1}, except for the times when rapid calving and disintegration occurred. Because ice-margin positions were plotted from available aerial photography, intervals between adjacent dated lines were not necessarily multiples of full years. Except for the times when rapid calving was taking place, the maximum measured recession rates along each transect occurred since 1990. The largest retreat—1.0 km—occurred along transect A in Tashalich Arm between June 1992 and March 1993. The total amount of terminus retreat along each transect is 8.80 km (transect A), 10.70 km (transect B), 8.85 km (transect C), and 7.00 km (transect D). In transects A and D, initial recession rates were low (0.15 km a^{-1} or less). In transects B and C, there was an initial rapid loss of what was perhaps a floating, thin extension of ice as retreat began, followed by a much slower rate of terminus retreat.

In 1977 (transect B) and 1984 (transect D) (Molnia and Post, 1995, p. 111), two episodes of very rapid calving and disintegration occurred; each instance resulted in more than 2.0 km of ice-margin recession. These rapid recessions occurred when the thinning, retreating glacier terminus decreased to a minimum thickness, after which buoyancy would not permit it to remain in contact with the bottom. As it floated, it rapidly disintegrated. Many large icebergs separated from the terminus resulting in a very rapid retreat of the

***Figure 196.**—20 July 1993 oblique aerial photograph showing post-1967 surge retreat of the east-central* Piedmont Lobe *region of Bering Glacier. Northwest-looking view along the terminus of Bering Glacier. The glacier is rapidly calving and disintegrating along old crevasse planes. The height of the shoreline scarp at the northeast edge of Vitus Lake is less than 5 m. Because of their buoyancy, several calved blocks of ice have risen above the elevation of the surface of Bering Glacier. Photograph by Bruce F. Molnia, U.S. Geological Survey. A larger version of this figure is available online.*

***Figure 197.**—29 July 1990 east-looking view of part of a 300×600-m mass of debris-covered stagnant Bering Glacier ice located near the head of the* Taggland Peninsula. *This remnant had a maximum height of approximately 30 m. When the rapidly advancing ice of the 1993 surge encountered this stagnant ice, it reconnected to the stagnant-ice mass and pushed it forward. Note the person for scale. Photograph by Bruce F. Molnia, U.S. Geological Survey. A larger version of this figure is available online.*

margin. This produced substantial short-lived increases in the recession rate. Before the summer of 1991, the retreating margin at the head of Tashalich Arm changed from one with a low rate of iceberg production to one with a much higher rate, and the fjord completely filled with floating calved icebergs. Correspondingly, the 1992–93 retreat rate—1.0 km a^{-1}—was nearly twice as high as the next highest rate (1987–90, 0.53 km a^{-1}) and more than six times higher that the initial rate of retreat (1967–69, 0.15 km a^{-1}).

Sequential aerial photography provided additional insights into the characteristics of the post-1967 retreat, as a 17 August 1979 oblique aerial photograph by the author shows (fig. 195). For instance, rates of retreat in the ice-filled basins of Vitus Lake were much greater than they were on the adjacent islands and land areas. Similarly, as the retreat progressed (fig. 196), the surface gradient and elevation of Bering Glacier decreased significantly. On areas adjacent to Vitus Lake, large stagnant masses of glacier ice covered the land surface. Some of these masses were remobilized when they were contacted by surge-displaced land-based ice (fig. 197).

In addition to rapid retreat, Bering Glacier has also thinned significantly during the period of retreat. Between 1967 and 1993, the ablation area of Bering Lobe was virtually stagnant ice, and there was little detectable ice flow. The reduced flow from upglacier contributed to a significant thinning of the ice in the lower 80 km of the glacier. Comparing ice-surface elevations at the terminus of Bering Glacier derived from the 1:63,360-scale USGS Bering Glacier A–7 topographic map (1984), which was based on 1972 aerial photography, with 1991 elevations derived from USGS vertical aerial photography shows a thinning of approximately 150 m at the Bering Lobe terminus, an average of approximately 7.9 m a^{-1}. Elsewhere, Trabant and Molnia (1992) compared 1990–92 surface-elevation profiles made by using a precision altimeter with elevations measured from a 1972 Bering Glacier topographic map (USGS, 1972). They determined that a thinning of between 85 and 180 m occurred along the centerline of Bering Lobe. The locations of many of their precision altimeter survey points are shown on figure 185. In some places, ice loss within the subsequent 20-year period represents approximately 20 to 25 percent of the 1972 thickness of Bering Glacier at that location. Molnia and Post (1995) reported that a comparison of 1946 and 1990 vertical aerial photographs shows that almost 200 m of thinning occurred between photographs along the margin of the Bering Glacier at an elevation of 280 m south of the Grindle Hills. Molnia and Post (1995, p. 113, fig. 16) presented surface-elevation profiles of Bering Glacier for 1900, 1957, 1960s, 1972, and 1991 and a mostly below-sea-level profile of the glacier base.

The 1993–95 Surge

By late August 1993, just before the beginning of the 1993 surge-related terminus advance, approximately 20×10-km Vitus Lake occupied an area of about 70 km^2, an increase in area resulting from late 20th century recession. The 1993–95 surge produced a maximum of around 10 km of advance of the terminus of Bering Lobe. Terminus-ice displacement occurred in two discrete phases—between late August 1993 and 17 September 1994 and between early May 1995 and mid-September 1995. A 7-month period without detectable ice displacement—from 17 September 1994 to early May 1995—separated the two advances. The surge resulted in a substantial increase in iceberg production; significant changes in the size, bathymetry, hydrology, and water chemistry of ice-marginal Vitus Lake; advance over vegetation (fig. 65); and the complete or partial covering of all of the islands within Vitus Lake by advancing ice. Maximum ice-displacement rates and maximum surge-front-displacement velocities approached 100 m d^{-1} during the initial period of the surge.

The 1993–95 surge was anticipated and closely monitored. More than a year before the appearance of any evidence that a surge was beginning, the

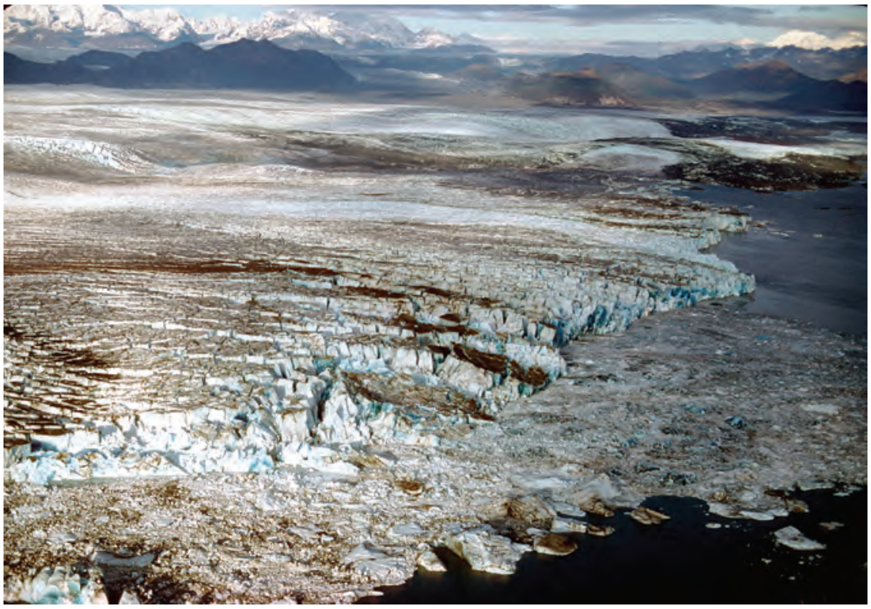


Figure 198.*—6 October 1993 northeast-looking oblique aerial photograph of the intensively fractured, disintegrating terminus of the Bering Glacier. Fracturing extends many kilometers upglacier. Several large surge-produced pressure ridges can be seen migrating towards the terminus. Photograph by Bruce F. Molnia, U.S. Geological Survey.*

tongue of ice that filled the head of Tashalich Arm began to rapidly calve icebergs, and the entire fjord was filled with floating ice within less than 3 months. An analysis of sequential vertical aerial photographs led Robert M. Krimmel (USGS, Tacoma, Washington, oral commun., 1996) to report that, during the 9-month period between 12 June 1992 and 16 March 1993, ice advanced toward Tashalich Arm at a velocity of 1.3 m d^{-1}, yet the terminus retreated about 300 m, more than 1 m d^{-1}. In the 4 months between 16 March 1993 and 10 July 1993, the velocity increased to 2 to 3 m d^{-1}. A similar rapid-calving process was characteristic of changes at several Bering Lobe ice-marginal locations when they were impacted by the arriving surge front, as shown on a 6 October 1993 oblique aerial photograph taken by the author (fig. 198).

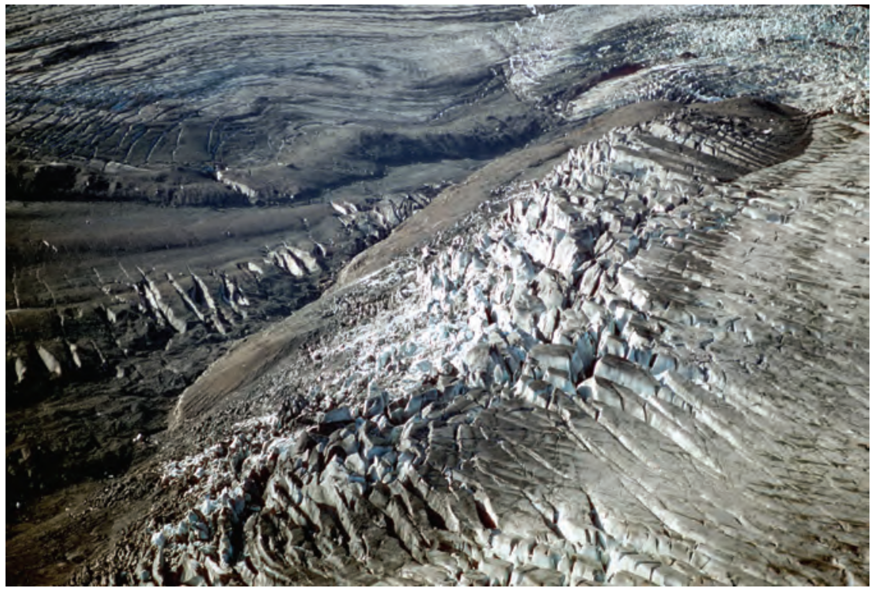
The 1993–94 Phase

In the central Bering Lobe, the first visible evidence of the onset of the surge was the development of a large pressure ridge along the southwestern edge of the Grindle Hills during the spring of 1993, although the ridge probably had been building up for many months before that observation. During the early stages of the surge, this location served as the initiation point for the fracturing and shattering of the surface of Bering Glacier. Within weeks, the upglacier surface of Bering Lobe was fractured for a distance of more than 50 km, and the thickening glacier was overtopping ridges, such as Override Ridge, in the lower Bagley Ice Valley (fig. 199).

A comparison of three sets of mosaicked vertical aerial photographs of the terminus of Bering Glacier in Vitus Lake acquired by the author on 10 July 1993, 10 September 1993, and 25 February 1994 show how various locations along the glacier's terminus responded very differently to the surge process. In some areas, the glacier's terminus rapidly advanced; in other areas, it retreated. During the first 2 months of the surge, the ice in part of the terminus adjacent to the eastern side of the peninsula that separates Vitus Lake from Tashalich Arm shattered and retreated rapidly following the arrival of the surge front, while the front advanced to the east. By November, the entire terminus was advancing rapidly into Vitus Lake.

Comparing two sets of vertical aerial photographs acquired on 17 October 1993 and 16 May 1994 showed that the terminus adjacent to the eastern side of Beringia Novaya advanced about 7.78 km in this 211-day period. The average rate of terminus displacement was about 36.7 m d^{-1}. The actual displacement was higher because a significant volume of ice calved daily from the advancing glacier margin. Cross-glacier fracturing proceeded at a much slower rate. But, by July 1994, the CMMB showed fracturing and

Figure 199.—6 October 1993 oblique aerial photograph shows surge-thickened ice at the edge of the Bagley Ice Valley expanding over an exposed bedrock ridge. The ridge, named Override Ridge by Austin Post, was also overridden by surge-thickened ice during the 1957–1960 surge. Photograph by Bruce F. Molnia, U.S. Geological Survey.



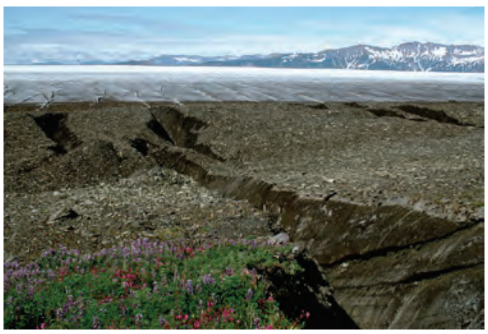


Figure 200.—Photograph of several extensional crevasses/faults located on the western side of the Central Medial Moraine Band of the Bering Glacier on 24 July 1994. Before the surge, this area was vegetation-covered stagnant ice. Photograph by Bruce F. Molnia, U.S. Geological Survey. A larger version of this figure is available online.

folding caused by both compressional and extensional stresses (fig. 200). In essence, the fractures were a complex network of parallel and subparallel crevasses that opened in the previously near-stagnant CMMB.

Robert M. Krimmel (USGS, Tacoma, Washington, written commun., 1996) analyzed nine sets of sequential vertical aerial photographs acquired between 12 June 1992 and 7 September 1994 to derive many physical characteristics of the surging Bering Glacier (12 June 1992, 16 March 1993, 10 July 1993, 10 September 1993, 15–17 November 1993, 25 February 1994, 16 May 1994, 13 August 1994, 7 September 1994). He reported velocities lower than those cited above but noted that “the highest speed measured from the successive aerial photographs was 22 m d^{-1}. This speed was measured at two locations: in the west terminus area above the entrance to Tashalich Arm and in the area a few kilometers west of Grindle Hills. These speeds were the average speed for a several month period, and it is reasonable to expect that the speed was higher for portions of the measured period.” Krimmel’s plots for seven positions of the terminus of Bering Glacier between 10 July 1993 and 7 September 1994 are shown in figure 201.

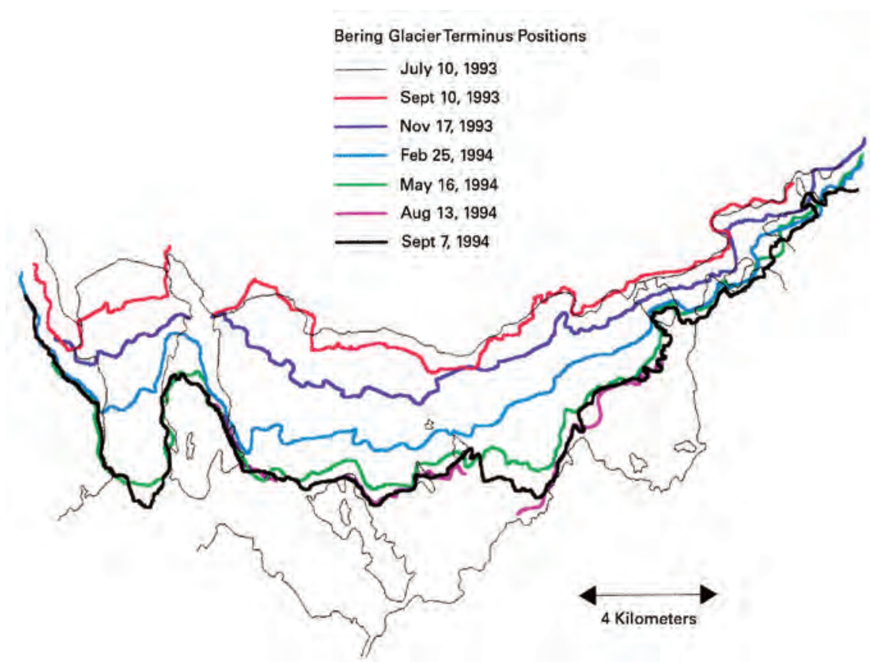


Figure 201.—Terminus positions of Bering Glacier are shown between 10 July 1993 and 7 September 1994. The July 1993 shoreline of Vitus Lake is shown as a fine black line. From data compiled by Robert M. Krimmel, U.S. Geological Survey.

Subsequent examination of sequential SAR imagery by Roush (1996) confirmed and enhanced the understanding of many of the events that occurred during the first 8 months of the surge. Roush found that the surge was in progress by 30 April 1993 and may have begun as early as 26 March 1993. Measurements showed that the surge front propagated downglacier at a mean velocity of 90 m d^{-1} between 19 May and 25 August 1993, the migrating surge front first reaching the terminus of the glacier in Vitus Lake just before 24 August 1993. Subsequently, the calving terminus advanced rapidly into Vitus Lake. On radar imagery, the advancing surge front consisted of a distributed region of undulations and elongated bulges on the surface of Bering Glacier, having heights estimated from the SAR data of 40 to 110 m and widths varying from about 0.7 to 1.5 km. Roush compared 9 August 1993 and 18 October 1993 radar images and calculated that, during this 71-day period, the average advance rate in its central area was 19 m d^{-1} and that the mean rate of advance across the entire width of the terminus was 11 m d^{-1}. As previously noted, these rates pertain to the displacement of the ice margin and do not take into account the large volume of ice calved from the end of the advancing margin during the entire period of observation. Additionally, had Roush used 24 August 1993 as a start date, his average rates would have been more than 20 percent higher.

Studies by Fatland and Lingle (1998, 2002) analyzed the surge by using differential SAR interferometry (InSAR) on two pairs of ERS-1 radar images, one pair collected during a 3-day period in January 1992 before the onset of the surge and the second during a 3-day period between 4 and 7 February 1994 at the peak of the surge. Their resulting high-resolution velocity data clearly show that the Bagley Ice Valley—southwest of Tana Glacier and northeast of Steller Glacier—experienced a 2.7-fold increase in velocity during the interval between image pairs, velocity increasing from 0.36 to 0.95 m d^{-1}. They also reported a drawdown of the surface of Bering Glacier of between 5 and 10 m and speculated that the velocity increase may have been caused by increased longitudinal stress gradients resulting from coupling to the surging main trunk of Bering Glacier. They also reported that InSAR was unsuccessful in determining velocities on the rapidly moving ice of Bering Glacier because there was no image correlation between sequential SAR images.

The advancing ice moved much more rapidly when it filled subglacier and sublacustrine fjord channels than it did when it advanced over land. Average land advance rates rarely exceeded 4 m d^{-1}. However, the advancing ice would frequently bulldoze large wedges of sediment in front of its advancing face. Many observed advances were accompanied by upward thrusting. Individual spatulate fingers of ice ranging from a few meters to more than 100 m in width built distinctive push moraines. Several of the islands in Vitus Lake were partially covered during the first phase of the surge. The northern part of 1.7-km-long Tsitus (*Arrowhead*) Island, which had only become free of retreating ice during the summer of 1992, was quickly covered by advancing ice (fig. 202*A*). By the end of the first phase of the surge, nearly 1 km of the northern end of the island was covered by advancing glacier ice (fig. 202*B*). The same lobe of ice that overtopped the north end of Tsitus (*Arrowhead*) Island filled the channel east of the island that connected the two eastern ice-marginal lakes with the main body of Vitus Lake. All of the drainage from the eastern part of Bering Glacier that previously flowed through this channel into Vitus Lake was deflected to flow through an abandoned former ice-marginal channel adjacent to the 1967 ice-maximum position, just to the east of the south end of transect D (Molnia and Post, 1995, p. 111, fig. 15).

Advancing ice slowly covered the northern end of Pointed Island. However, a much more rapidly moving tongue of ice, at least 170 m thick, advanced past the island on its western side at a rate that exceeded 20 m d^{-1}. Before the first phase of the surge ended, two small islands, named the The Wallypogs

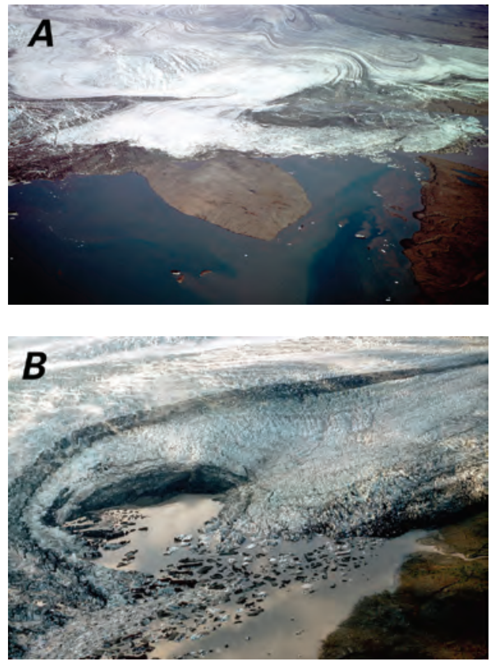


***Figure 202.**—Two oblique aerial photographs of the Tsitus (*Arrowhead*) Island area that show changes that occurred during the first and second phases of the 1993–1995 surge of the Bering Glacier. **A**, 7 October 1993 north-looking oblique aerial photograph shows the advancing margin of the Bering Glacier just as it made contact with the north end of the island. The channel on the east side of the island, which connects to the eastern marginal lakes, is still open. Between the summers of 1992 and 1993, the north end of Tsitus (*Arrowhead*) Island had become ice free. **B**, 9 July 1995 north-looking oblique aerial photograph shows that the advancing margin of Bering Glacier had covered all but the southernmost 45 m of the island. Advancing ice masses in the channels on both sides of the island advanced towards each other but never joined. This photograph depicts the maximum ice advance and ice cover of Tsitus (*Arrowhead*) Island. Photographs by Bruce F. Molnia, U.S. Geological Survey. Larger versions of these figures are available online.*

by Austin Post, were completely covered, as were the northern two-thirds of Whaleback Island. The northwestern shoreline of *Taxpayer's Bay*, the area at the southern end of transect D (Molnia and Post, 1995, p. 111, fig. 15) was also overtopped by the advancing glacier.

Robert M. Krimmel (Tacoma, Washington USGS, oral commun., 1996) compared sequential vertical aerial photographs acquired on 16 May 1994, 13 August 1994, and 7 September 1994 to determine when this phase of the surge ended. He reported that ice velocities were high for the period between 16 May 1994 and 13 August 1994. Terminus changes were minor between 13 August 1994 and 7 September 1994. By mid-September 1994, advance of Bering Glacier had ceased. Vitus Lake was so filled with the terminus of Bering Glacier and icebergs that almost no open water remained.

Surge-Produced Changes in Vitus Lake

Gray and others (1994) documented changes in the physical and sedimentary characteristics of Vitus Lake that occurred during the first year of the surge by comparing pre-surge measurements made during the July 1992 retreat phase with peak surge measurements made in July 1994. In 1992, the lake had an area of approximately 160 km^2, and about 90 percent of its surface was free of icebergs. The lake was stratified into two distinct layers, the topmost of which extended from the surface to a depth of about 7 m. Generally, specific conductance values increased from about 2,000 to about 5,000 $\mu S\ cm^{-1}$ (micro Siemens per centimeter), and water temperature decreased from about 5° to about 3°C. Specific conductance values of approximately 17,000 $\mu S\ cm^{-1}$ at a depth of 10 m increased to a maximum of approximately 32,400 $\mu S\ cm^{-1}$ at a depth of 142 m. The majority of water temperature measurements made were less than 1°C. The depth of the transition zone between the two layers corresponded to the depth of the thalweg (longitudinal outline of riverbed from source to mouth) of the Seal River. Additionally, the mean Secchi-Disk depth (measurements of water clarity) obtained from measurements made at locations more than 1 km from sediment-rich glacial water inflows was 2.3 m.

When Vitus Lake was measured in mid-July 1994, the surge had reduced its area to approximately 50 km^2, and more than 90 percent of its surface was covered by floating ice. Specific conductance values ranged from about 2,000 to about 3,200 $\mu S\ cm^{-1}$, and water temperature ranged from about 4° to about 0°C. No evidence of stratification was found. Mean Secchi-Disk depth obtained from measurements made near the head of the Seal River was 0.4 m. Visual observations of water samples collected with a Van Dorn sampling device suggested that suspended sediment concentration increased with depth, unlike it did in 1992 when the lake was essentially sediment free. Lastly, in contrast to the1992 sediment discharge rate of about 10 $kg\ s^{-1}$, the suspended sediment discharge in 1994 was around 10^4 $kg\ s^{-1}$.

The 1995 Phase

Oblique aerial photographs of the terminus region of Bering Glacier acquired in late November 1994 and again in late January 1995 showed no evidence of surge activity. Following a 7-month period beginning in September 1994 and characterized by minor retreat and near-stagnant ice, parts of the eastern Bering Glacier resumed surging; several areas of the terminus advanced about 750 m in the 13-day period between 19 May and 1 June 1995 (57.7 $m\ d^{-1}$). The first evidence that new surge activity had begun was noted on 14 April 1995 by pilots Gayle and Steve Ranney of Cordova, Alaska. They observed that a section of the winter-ice cover of Vitus Lake was being compressed into a series of accordionlike folds. They also observed numerous deep fresh cracks and rifts as well as a number of blue-water lakes forming on the surface of Bering Glacier. The lakes, fractures, and rifts were characteristic features of the 1993–94 phase of the surge.

USGS vertical aerial photography of the Bering Lobe terminus and Vitus Lake, acquired on 1 May 1995, confirmed the new fracturing and rifting as well as the numerous lakes. When they were compared with the late November 1994 vertical aerial photographs, the 1 May 1995 photographs showed that the terminus was advancing over the northern end of Beringia Novaya, the largest island in Vitus Lake, and over the north-central part of Pointed Island. The margin showed a significant increase in iceberg production, including several icebergs that were more than 0.5 km long. 19 May 1995 oblique aerial photography confirmed that the terminus was continuing to advance. On the southeastern shoreline of the lake, just west of *Taxpayer's Bay*, the advancing ice margin was redirecting the entire drainage from the eastern part of the glacier to a narrow channel. The result was several tens of meters of shoreline retreat and the development of a high bluff, as two oblique aerial photographs taken on 6 and 9 July 1995 by the author show (fig. 203).

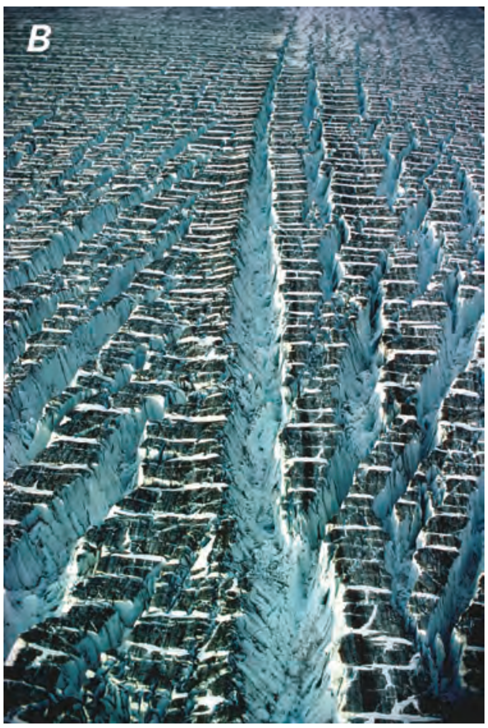
The author's 1 and 2 June 1995 visit to the glacier confirmed that not only was the surge affecting the eastern terminus region but also that it was affecting the northern part of Bering Lobe and the southern part of the Bagley Ice Valley, as much as 30 km north of the terminus (fig. 204). There, the winter 1994–95 snow surface was complexly fractured and rifted, and several large ice-surface (supraglacier) lakes reappeared in the same location where lakes had existed during the January to July 1994 period. On 1 June 1995, only the southernmost 50 m of Pointed Island remained exposed; about 750 m of the island had been covered since 19 May 1995. On Pointed Island, thousands of nesting birds were displaced or killed and their nests destroyed as they were overtopped by the advancing ice margin. On Tsitus (*Arrowhead*) Island, the total ice advance during this phase of the surge was less than 200 m.

This phase of the surge was much shorter lived than the first. However, total terminus advance was as significant as it was in the 1957–67 surge event, and advance speeds into the proglacial Vitus Lake again greatly exceeded flow speeds over islands and adjacent land areas. This advance revitalized the stagnant mass of ice to the north in the Weeping Peat Island area, and the advancing ice reconnected with the toe of the glacier that had been detached by the jökulhlaup in December 1994 (see following section). Here the total amount of advance was about 250 m. As was the case to the west, the last evidence of ice advance disappeared by mid-September 1995.

Muskett and others (2000) reported that, in June 1995, just after the onset of the second phase of the 1993–95 surge, Bering Glacier was profiled by a geodetic airborne laser altimeter according to the methods of Echelmeyer and others (1996). These results were used to compare elevation and volume

Figure 203.**—Two views show changes occurring on the southeast shoreline of Vitus Lake during July 1995.* **A, 9 July 1995 northeast-looking oblique aerial photograph of the advancing ice margin that forced the entire drainage from the eastern part of the Bering Glacier to flow into a narrow channel. A larger version of this figure is available online.* **B***, 6 July 1995 southeast-looking photograph of the shoreline retreat and the development of a high bluff. Photographs by Bruce F. Molnia, U.S. Geological Survey.*

Figure 204.—Two oblique aerial photographs showing complex fracturing and crevassing that occurred during the later part of phase one and the early part of phase two of the 1993–1995 surge. **A***, 9 May 1994 south-looking view shows a 4-km-wide section of the Bagley Ice Valley, with several types of crevasses, including rectilinear crevasses, and the development of subparallel rifts.* **B***, 24 July 1994 north-looking view across the central Piedmont Lobe shows part of a 5-km-long developing rift set in a wide section of rectilinear crevasses. The east-west-trending snow-filled crevasses represent fractures that formed during early fall 1993 and that were subsequently filled with winter 1993–1994 snow. The north-trending fractures and crevasses began forming during the spring of 1994. Photographs by Bruce F. Molnia, U.S. Geological Survey. A larger version of this figure is available online.*



changes that occurred between June 1995 and 1972–73. They found that, despite the fact that a significant volume of ice was transported to Bering Lobe by the surge in 1993 and 1994, the surface of the glacier was “generally lower in 1995 than in 1972–1973” (Muskett and others, 2000, p. F404). They estimated that the total volume lost was 41±10 km^3 of ice, which they presented as a corresponding area-average mass balance of –0.8±0.2 m a^{-1} water equivalent. Elevation change was nonuniform; surface lowering was greatest on Bering Lobe near the terminus, where a maximum of 75 to 100 m of lowering had occurred. Although they did not quantify it, Muskett and others (2000) noted an area of the Bering Lobe 13 km upglacier from the terminus where a small thickening occurred. A maximum of 50 m of thickening was seen further upglacier above the equilibrium line. They concluded that “negative mass balance predominated over the massive downglacier transport of ice caused by the 1993–1995 surge” (Muskett and others, 2000, p. F404).

Comparing 1950s map data of the glacier with data obtained during airborne profiling surveys in the middle and late 1990s showed that, on an annual basis, Bering Glacier thinned by 0.914 m a^{-1} and that its volume decreased by 1.78 km^3 a^{-1}. However, between the middle 1990s and 1999, on an annual basis, Bering Glacier thinned by 3.077 m a^{-1} and its volume decreased by 6.0 km^3 a^{-1} (K.A. Echelmeyer, W.D. Harrison, V. B. Valentine, and S. I. Zirnheld, University of Alaska Fairbanks, written commun., March 2001).

The 1994–1995 Jökulhlaup

During the spring of 1994, a number of large blue-water ice-surface (supraglacier) lakes formed at several locations on Bering Lobe and in the Bagley Ice Valley, as a 24 July 1994 oblique aerial photograph by the author shows (fig. 205). The largest of these was more than 1 km long. On 27 July 1994, a large sediment-laden jökulhlaup began at the terminus of Bering Lobe immediately east of Weeping Peat Island. During the first few hours of the jökulhlaup, hundreds of large blocks of ice, many with maximum dimensions estimated to be greater than 30 m, broke off from the glacier face and were jetted into the deep lake adjacent to the ice margin. Within 24 hours, a canyon more than 1 km long was cut into the ice margin as the point of water discharge retreated upglacier. Initial peak discharges were estimated to be greater than 10^5 m^3. In early September 1994, the retreating point of the high-volume discharge reached the northern end of Weeping Peat Island. Within days, many of the supraglacier lakes were drained.

The high-volume discharge of water, which had shifted direction nearly 90°, continued to cut a widening west-trending channel through the stagnant terminus of the glacier during the next few months and dissected the toe from the remainder of the glacier. During the first year of the flood, approximately 0.3 km^3 of sediment and ice were deposited into the ice-marginal lake system on the southeastern side of the glacier, and a channel was cut through the central area of Weeping Peat Island. The high-volume discharge continued for much of the next year and deposited a large outwash-fan delta, part of which formed in water as deep as 50 m. By late July 1995, following a lowering of the lake's level, the melting of ice blocks buried in the sediment created several dozen near-circular pit pond depressions (kettles) on the surface of the northern end of the outwash-fan delta adjacent to the ice margin. The largest had a diameter of about 20 m and a depth of 5 m. At several lakeshore and riverbank bluffs, blocks of ice buried in the sediment were exposed in profile.

By the spring of 1996, the volume of water discharging from the base of the glacier had decreased to near pre-jökulhlaup levels. As a result, the lake's surface elevation dropped several meters, and more than 6 km^2 of the outwash plain-fan delta were exposed. The stream discharging at the base of the glacier began to meander across the exposed outwash plain-fan delta. An inversion of topography occurred at many locations where ice blocks were buried in the sediment, and the tops of buried ice blocks were exposed, many reaching more than 3 m above the outwash plain-fan delta's surface (figs. 206*A, B*). Many blocks were surrounded by moats, some water

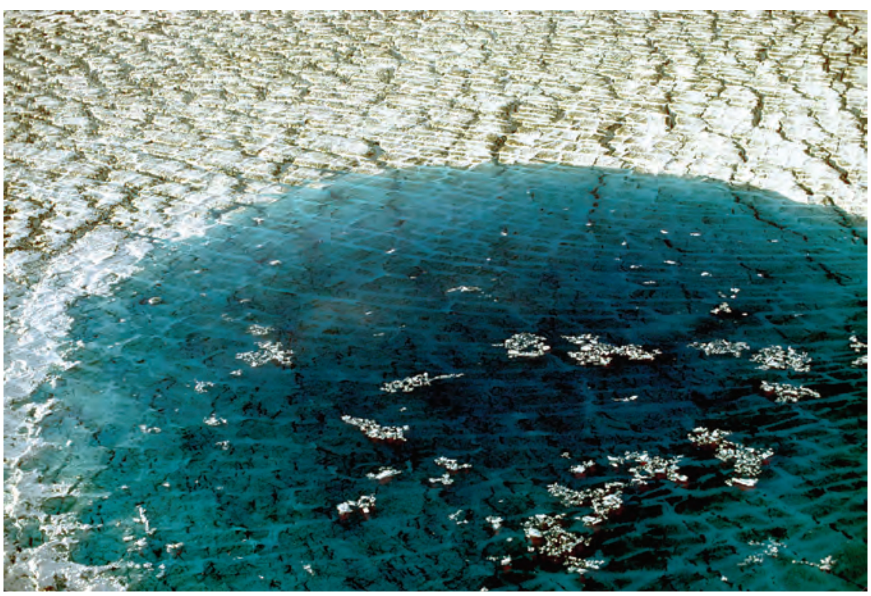
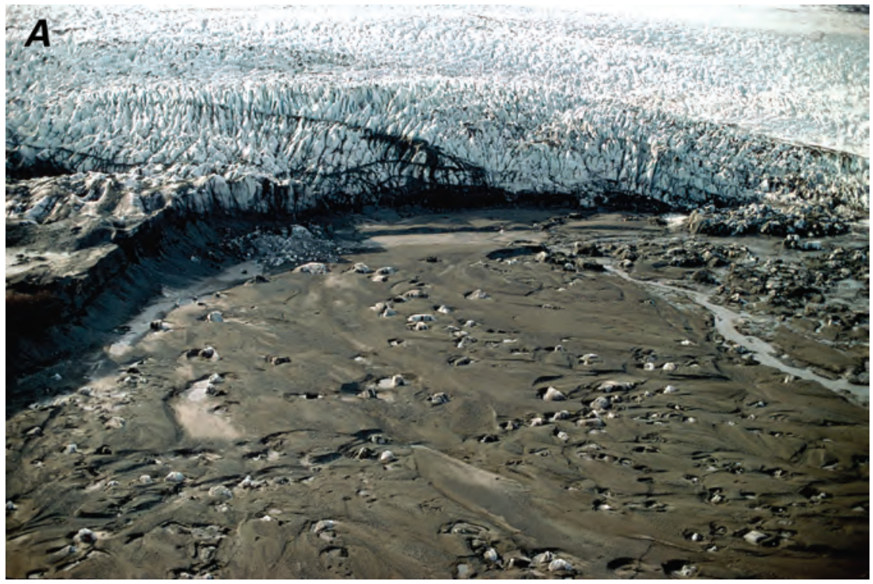
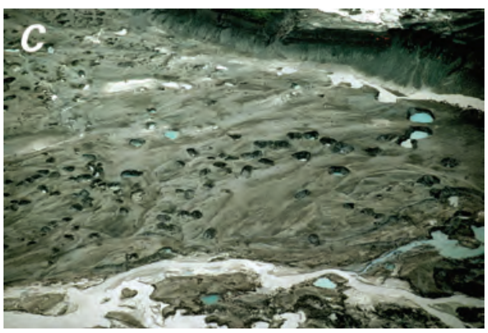


Figure 205.—*24 July 1994 north-looking pre-jökulhlaup oblique aerial photograph of a blue-water lake formed on the surface of Bering Glacier in the Bagley Ice Valley taken less than a week before the onset of the 28 July 1994 jökulhlaup. The photograph shows part of an approximately 350-m-diameter circular lake located approximately 10 km west of Juniper Island. Photograph by Bruce F. Molnia, U.S. Geological Survey.*

Figure 206.—*Three photographs showing the evolution of kettle ponds in the outwash-fan delta formed by sediment deposited during the 1994–1995 jökulhlaup.* **A***, 30 April 1996 oblique aerial photograph of the northern end of the outwash-fan delta shows two stages in the development of the kettle ponds. Meandering of the stream located to the right of center has eroded the sediment cover from many large ice boulders to its east. Their rapid melting resulted in the formation of only a few small kettle ponds. To the west of the stream, where only the top part of the ice boulders is exposed, melting over the ensuing 3 months formed many large kettle ponds.* **B***, 1 May 1996 ground photograph of a 20-m-long sediment-banded ice boulder present in* A*. Three months later, a 6-m-deep kettle pond was located at this site.* **C***, 20 July 1996 oblique aerial photograph of the same area shown in* A*. Individual kettle ponds can be correlated with exposed ice blocks in the earlier photograph. Photographs by Bruce F. Molnia, U.S. Geological Survey. Larger versions of* B *and* C *are available online.*





filled. Erosion and scour resulting from lateral stream-channel migration had removed at least 4 m of sand and gravel and left behind the newly exposed ice boulders. This erosion and downcutting were responsible for the reversal of topography.

All of the exposed ice boulders had melted by 20 July 1996, and the surface of the outwash plain-fan delta was covered with several hundred kettle holes (fig. 206*C*). Of the 90 kettles the author studied during a 2-week field investigation in mid-August 1996, the largest was about 35 m in diameter and about 5 m deep. More than a dozen had ice exposed in their walls. Most of the studied kettles were continuing to enlarge through slumping and melting of buried ice. Another stream channel migration event was underway on 17 September 1996, the date of the last 1996 site visit. Between 26 August and 17 September 1996, the stream migration had completely removed about 40 of the easternmost kettles and exposed ice boulders at the surface of several of the former kettle locations on the eastern side of the westward-migrating stream. Many other areas showed evidence of surface slumping, an indication that melting of subsurface glacier ice was continuing. When the area was observed in August 2001, there was no evidence of continuing kettle growth, and various types of vegetation were becoming established in the area between the kettles.

It was initially thought that the July 1994 jökulhlaup was a surge-ending flood event because ice velocities decreased following the onset of the jökulhlaup and ice-displacement ceased within 60 days. But it was not, because the ice resumed its advance following a 7-month pause. Perhaps the jökulhlaup was caused by a partial failure of the subglacier "barrier" or "dam" responsible for the initiation of the 1993–94 surge. However, it was the only major jökulhlaup associated with the 1993–95 surge of the glacier.

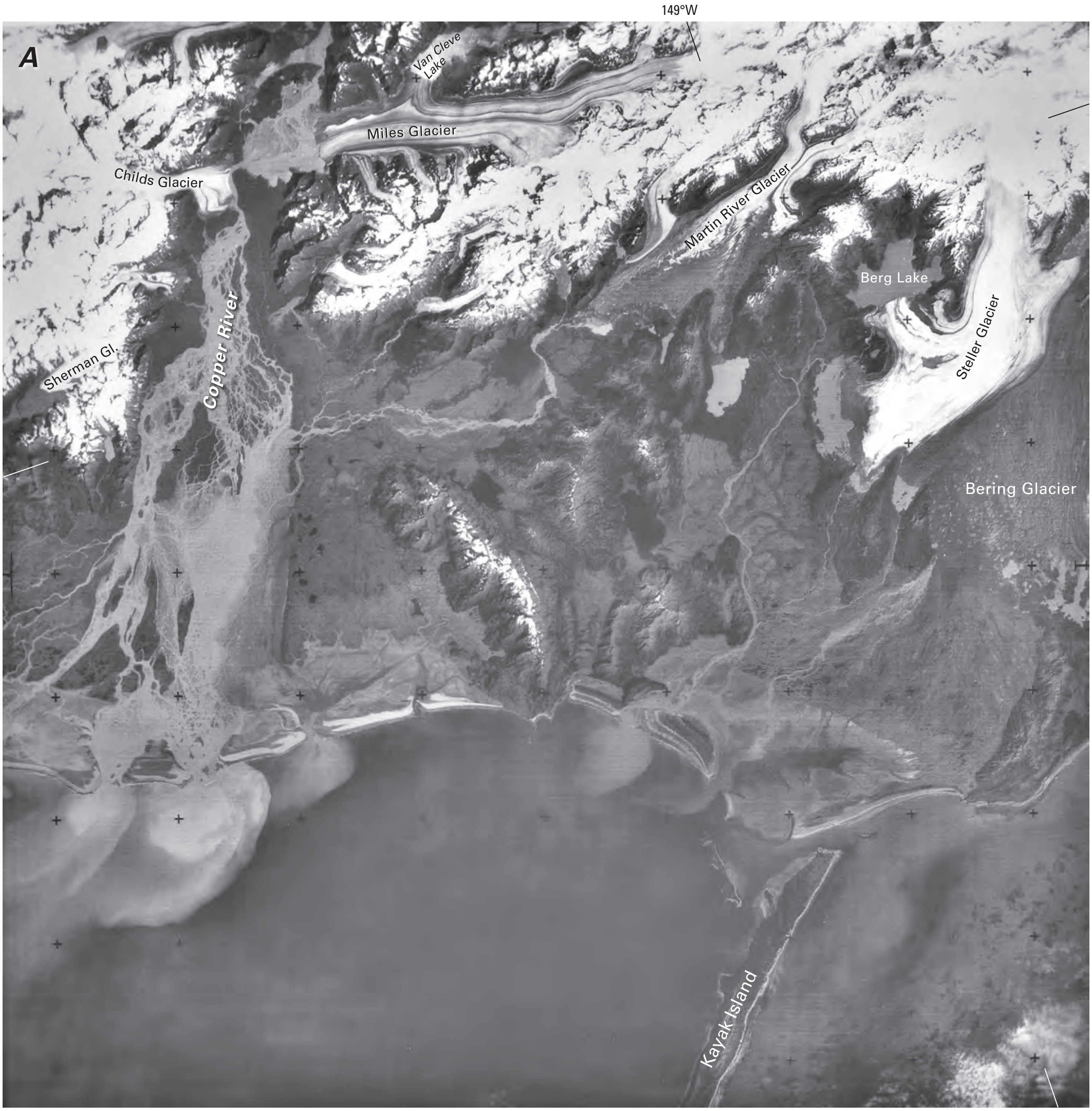
Following the onset of the July 1994 jökulhlaup, a large suspended-sediment plume began to exit Vitus Lake and entered the Gulf of Alaska through Seal River, as a 7 September 1994 vertical aerial photograph shows (fig. 207). It is likely that the origin of the plume was one or more subglacier channels located on the eastern side of the Bering Lobe. The restriction, if not total blockage, of flow through these sediment-dammed channels may have been responsible for the 1993 surge. The discharge of sediment appears to be closely tied to the mechanics of Bering Glacier's surge cycle. From the 1970s through the middle-1980s, sediment-laden water flowed from Vitus Lake through Seal River into the Gulf of Alaska. Sometime during 1985, the plume disappeared and remained absent until just after the beginning of the 1994 jökulhlaup. The sediment plume's reappearance at about the same time

Figure 207.—7 September 1994 vertical aerial photograph of Seal River shows the large suspended sediment plume exiting Vitus Lake. At the time of this photograph, the suspended sediment load exceeded 2 g L^{-1}. For much of the decade before the onset of the 1994 glacier outburst flood (jökulhlaup), the suspended sediment load dropped by more than an order of magnitude. USGS photograph no. 94–V4–85 by Robert F. Krimmel, U.S. Geological Survey.

as the jökulhlaup's occurrence adds credence to the hypothesis that Bering Glacier's surges are caused by a buildup of subglacier water pressure and water volume resulting from blockage of Bering Glacier's subglacier channels with sediment and that the jökulhlaup resulted from the failure of the subglacier sediment dam.

The Seal River sediment plume is visible on nearly every Landsat image collected during the Landsat baseline period (1972–81) (fig. 208*A*) (Post, 1976). During that period, it was the second largest and second densest sediment plume entering the Gulf of Alaska. The largest—the Copper River plume—annually transports about 10^8 metric tons of sediment into the Gulf of Alaska (Reimnitz, 1966) (fig. 208*B*). In addition to the Landsat imagery, 1979 aerial photographs of the Seal River also show the sediment plume, but it is absent from 1990 and 1992 photographs. A 22 August 2003 Moderate

► ***Figure 208.****—Two 1970s views, a 2003 view, and a 1938 view of the suspended sediment plume from the Seal River. The magnitude of the sediment plume before the middle 1980s suggests that little impediment existed to its flow through subglacial drainage channels of the Bering Glacier. The subsequent closure of the drainage channels, resulting in the middle 1980s decrease in suspended sediment load, may have been a factor in the onset of the 1993 surge. The Seal River sediment plume, west of the terminus of Bering Glacier, separates into two components: part of the plume extends offshore, and part flows along the west side of Kayak Island, merging with the Copper River suspended sediment plume. In the 1970s and early 1980s, the Seal River plume was the second largest sediment plume entering the Gulf of Alaska. For comparison, Reimnitz (1966) calculated that the Copper River plume, largest in Alaska, annually introduced approximately 10^8 kg of suspended sediment into the Gulf of Alaska.* **A***, Annotated Landsat 3 RBV image of the Copper River Delta area and environs on 24 August 1978. Seal River is to the right off the image, but part of its sediment plume is visible in the lower right corner. The Steller Lobe dams Berg Lake (fig. 191), which had jökulhlaups in 1984, 1986, and 1994. Martin River Glacier has a large stagnant terminus. Miles Glacier has receded 7.2 km since 1890 (Field, 1975, p. 321). Previously, it coalesced with Childs Glacier on the opposite bank of the Copper River. Childs Glacier surged in 1909, with an advance of several hundred meters, but has not surged since. The Sherman Glacier terminus is covered with the debris from a landslide that occurred during the 1964 earthquake. The Copper River is very turbid; its sediment plume extends far into the Gulf of Alaska. Landsat 3 RBV image (30172–20175–B; 24 August 1978; Path 71, Row 18) and caption courtesy of Robert M. Krimmel, U.S. Geological Survey, is from the EROS Data Center, Sioux Falls, S. Dak.* **B***, 12 July 1976 photograph taken from the rail of the* R.V. Sea Sounder*, approximately 3 m above the ocean surface, of the edge of the Seal River plume in the Gulf of Alaska. Photograph by Bruce F. Molnia, U.S. Geological Survey. A larger version of this figure is available online.* **C** *and* **D***, see following page.*



149°W
A
Van Cleve Lake
Miles Glacier
Childs Glacier
Martin River Glacier
Berg Lake
Steller Glacier
Sherman Gl.
Copper River
60°30′N
Bering Glacier
Kayak Island

B

Figure 208.—**C**, *22 August 2003 Moderate Resolution Imaging Spectroradiometer (MODIS) image of the Gulf of Alaska coastal area from just east of Prince William Sound (left) to Grand Plateau Glacier (right). The same area shown in* A *is located on the western third of the MODIS image. MODIS image from the National Aeronautics and Space Administration.* **D**, *Summer 1938 oblique aerial photograph of the southernmost part of the terminus of Bering Glacier, Vitus Lake, Seal River, and the Gulf of Alaska. In this post-1920s surge photograph, a large sediment plume can be seen entering the Gulf of Alaska and flowing around Kayak Island. A large calving and disintegration event is also taking place at the glacier's terminus. Photograph no. 1000 by Bradford Washburn, Museum of Science (Boston).*

Resolution Imaging Spectroradiometer (MODIS) image (fig. 208*C*) covers approximately the same area encompassed by the 24 August 1978 Landsat 3 RBV image (fig. 208*A*) and shows multiple plumes. A very large sediment plume is visible on Bradford Washburn's summer 1938 photograph of the mouth of the Seal River (fig. 208*D*).

Post-Surge Retreat of Bering Glacier

As it was during the advancing phase, the retreat rate of the terminus from the land areas was significantly less than that of the thicker tongues of ice that filled the deeper basins of Vitus Lake. Before the summer of 1996, little change was noted on the ice-covered islands within Vitus Lake or on the ice-covered part of the mainland east of Vitus Lake. By early summer 1996, the retreat of this land-based ice was underway at an average rate of less than 0.5 m d^{-1}. By 1997, multiple recessional moraines indicated hiatuses in the land-based retreat cycle at several locations.

Following the end of terminus advance in mid-September 1995, the parts of Vitus Lake that were not filled by surge-advanced ice were filled with

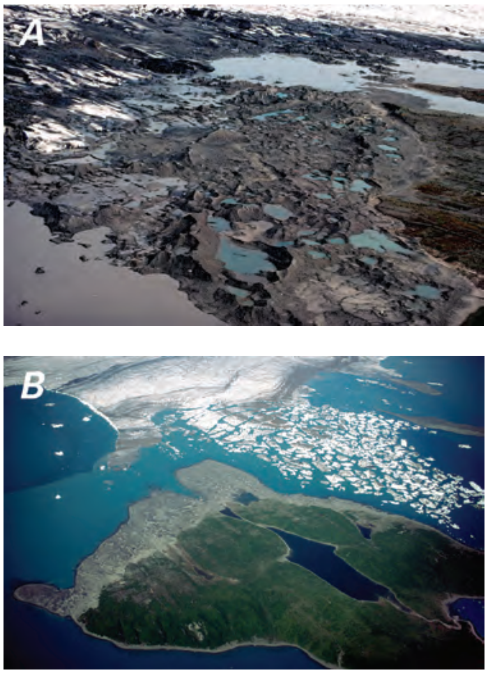


***Figure 209.**—Two oblique aerial photographs of the continued retreat of Bering Glacier from the north end of Beringia Novaya.* **A**, *6 June 1997 photograph shows the surge-maximum moraine and several recessional moraines, numerous developing kettles, and the retreating ice margin.* **B**, *12 August 2001 photograph taken more than 6 years following the cessation of surge motion within the terminus of the Bering Glacier. The boundary between vegetation and bare ground marks the 1995 maximum advance of the glacier. Continued thinning and retreat have resulted in the rapid disintegration of the terminus of Bering Glacier. Photographs by Bruce F. Molnia, U.S. Geological Survey. Larger versions of these figures are available online.*

icebergs. Almost no water was visible between the glacier's margin and the southern shore of Vitus Lake. Calving, which was ongoing during the early rapid advance of the terminus, continued as surge advances waned. The glacier slowly began to actively retreat again (fig. 209*A*), mostly through the production of many icebergs (fig. 209*B*). Without the influx of surge-transported ice, rapid retreat of the terminus from the deeper basins of Vitus Lake ensued. Both Vitus Lake and Tashalich Arm quickly developed semicircular calving embayments (fig. 40). Active calving in both areas continued through the summer of 1998. By late 1999, Beringia Novaya and Pointed Island were ice free; Whaleback Island was ice free by early 2001.

The retreat from Tsitus (*Arrowhead*) Island was much slower. Less than 50 m of retreat had occurred by 10 August 1998 (fig. 210); less than 300 m of the southern end of the island had been exposed by August 2001. The ice that had covered Tsitus (*Arrowhead*) Island was the thickest of any island in Vitus Lake. The channel to the east that had connected to the eastern ice-marginal lakes was still filled by a large volume of surge-advanced ice. Similarly, ice that had advanced over the mainland east of Vitus Lake still covered much of the land, as a 12 August 1998 oblique aerial photograph taken by the author shows (fig. 211). Before the summer of 2001, all of the ice that had advanced onto Weeping Peat Island had retreated. The surface of the eastern part of the island displayed at least four recessional moraines.

By late summer 1997, the rate of calving decreased in the Vitus Lake embayment west of Beringia Novaya. Several concentric, arcuate crevasses developed parallel to the perimeter of the terminus in the calving embayment. By 26 September 1997, intensive compressional forces rotated a single large mass of the glacier bounded by two subparallel crevasses and raised it more than 40 m above the surface of the adjacent ice (fig. 212). After about 72 hours, this massive 80×50×10-m pyramidal-shaped piece of ice calved into Vitus Lake.

Calving of the margin continued at a reduced rate from late 1997 through late 1998; however, the physical characteristics of the margin changed significantly. The height of the face of the glacier decreased by more than 50 percent

***Figure 210.**—Oblique aerial photograph of the deglacierized terrain following the post-surge retreat of Bering Glacier from the south end of Tsitus (*Arrowhead*) Island. 10 August 1998 east-looking view of the southern end of Tsitus (*Arrowhead*) Island shows the retreating terminus of Bering Glacier. As was the case on Pointed Island, several recessional moraines mark the post-surge retreat. Many grooves and furrows mark the areas of maximum advance. Photograph by Bruce F. Molnia, U.S. Geological Survey.*

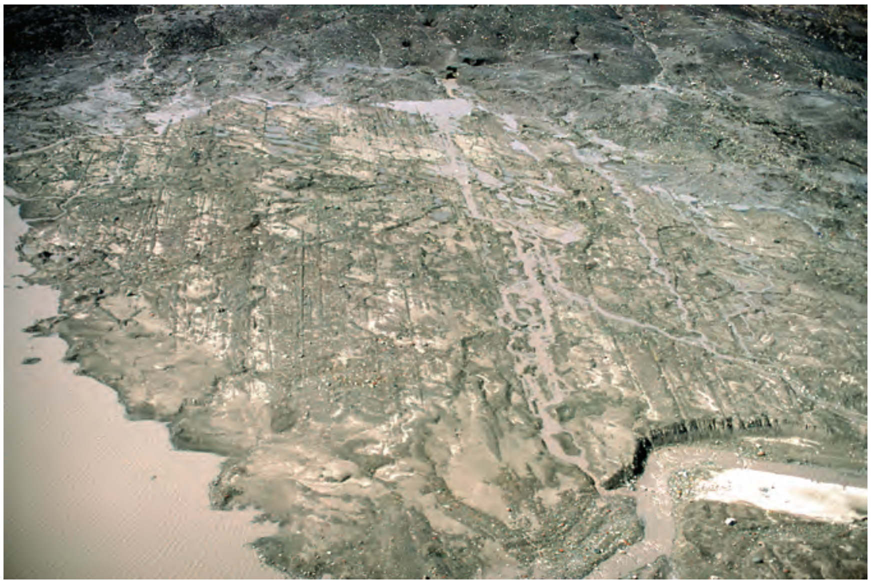


Figure 211.*—12 August 1998 north-looking oblique aerial photograph of the post-surge retreat of Bering Glacier west of Weeping Peat Island. The most obvious features are the numerous grooves and furrows. Retreat during the 3-year period since the cessation of the surge amounts to approximately 500 m. Photograph by Bruce F. Molnia, U.S. Geological Survey.*

Figure 212.*—Photograph of an uplifted pyramidal-shaped piece of ice, approximately 80×50×10 m, produced by unusual compressional stresses in the calving embayment of Bering Glacier in Vitus Lake on 26 September 1997. Southwest-looking view across the Bering Glacier west of Beringia Novaya. The uplifted triangular block of ice was not observed while being rotated. However, its debris-covered former surface has been rotated towards the west. Note the single-engine aircraft flying behind the ice block. Photograph by Bruce F. Molnia, U.S. Geological Survey.*

between 1996 and late 1998. The height of the ice cliff along much of the margin in Vitus Lake continued to decrease in 1999 and 2000. By 1999, part of the terminus had retreated more than 5 km. By 2001, continuing retreat of the glacier had caused much of the terminus to reoccupy positions similar to those held by the ice front in the pre-surge period from 1992 to 1993.

In 2000 and again in 2001, as successions of large icebergs calved from the margin and drifted into Vitus Lake, parts of the terminus were observed retreating as much as 700 m in less than 24 hours. By 12 August 2001, the surface of the glacier in the eastern terminus region closely resembled the pre-1993-surge surface of the glacier (fig. 213).

In the central part of Bering Lobe, several kilometers behind the terminus, dramatic changes occurred during the post-surge period. When the surge ended in September 1995, Bering Lobe's surface was fractured by large rifts, irregular rectangular blocks of crevasse-bounded ice, and many seracs. Local relief was as much as 20 m in places (fig. 204). Rapid melting was accelerated by the large surface area of the numerous surface features. By the end of the summer 1997, the surface of Bering Lobe had little local relief. In August 2001, the time of the last observation made by the author, many areas of the generally flat surface of Bering Lobe showed scars from the intensive deformation that the glacier had undergone during the surge. In addition, extensional fractures (fig. 200) that had opened in the CMMB closed within 24 months. Sauber and others (2000) recognized that substantial quantities of ice were transported from the accumulation area of Bering Glacier to the terminus region during the 1993–95 surge of Bering Glacier. They

Figure 213.—*12 August 2001 oblique aerial photograph of the nature of the terminus of Bering Glacier 6 years after the end of the 1995 surge. The north-looking view shows the low-relief terminus of Bering Glacier and the recently emerged Whaleback Island. With the exception of break up of the terminus on the left side of the picture, calving is minimal. The* Piedmont Lobe *has a very low gradient, and the ice face ranges in height from less than 1 m to approximately 3 m in height. Photograph by Bruce F. Molnia, U.S. Geological Survey.*

noted substantial rates of near-instantaneous uplift in the reservoir region in response. These rates ranged from 18.2±6.6 to 29.9±5.7 mm a^{-1}. Sauber and Molnia (2000) also found that seismicity in the region of the dramatic thinning increased during the surge interval relative to the pre-surge period. Following the surge, during the 2-year post-surge period of 1998 to 2000, no earthquakes greater than M2.5 occurred.

Josberger and others (2001) measured the bathymetry, temperature, and conductivity of Vitus Lake during August 2001. They found that intense vertical convection in Vitus Lake was controlled by the salt content of the water and that there was strong saline stratification in the deeper portions of the lake. Thermal diffusion across the pycnocline (zone of change in water density as function of depth) may produce frazil ice growth, whereas melting of the glacier terminus produces convection at the margin of the lake.

Figure 214.—*Two oblique aerial photographs of the effects of the surge on the ice-cored lateral moraine of the Steller Lobe, located north of Nichawak Mountain.* **A***, 12 August 2000 close-up north-looking view of the distal end of the lateral moraine. In late spring 2000, the ice-cored moraine started to disintegrate. By late summer, the Steller Lobe rapidly advanced into the adjacent lake.* **B***, By 12 August 2001, the surge was over. In a higher altitude northeast-looking view, most of the seracs and ice pinnacles visible in 2000 have melted. The rapid advance of the terminus has also ceased. Photographs by Bruce F. Molnia, U.S. Geological Survey. Larger versions of these figures are available online.*

Steller Glacier Activity

Steller Glacier did not actively surge during the 1993–95 Bering Glacier surge. In 1994, it began to develop a series of large subparallel cracks along the southern margin of the Steller Lobe, which were thought to be the result of stresses that were transmitted through the CMMB. By 1998, part of the southwestern terminus of the Steller Lobe experienced a pulse that caused it to advance several hundred meters. By 2000, much of the surface of the Steller Lobe was fractured, and the main trunk of Steller Glacier and the northern lobe of its terminus, which enters (and dams) Berg Lake, began a mini-surge. This surge continued through the late summer of 2001. The ice-cored lateral moraine on the northwestern side of the part of the Steller Lobe that ends in an unnamed lake, north of Nichawak Mountain, became broken up in 2000 and rapidly advanced into the lake (fig. 214). By fall 2001, the surge was over. The southern part of the terminus of the Steller Lobe did not show any evidence of been involved in the surge.

Glaciers of Waxell Ridge

Waxell Ridge separates the valley glacier section of Bering Glacier from the Bagley Ice Valley. Its high point is the summit of Mount Steller. Several dozen unnamed cirque and small valley glaciers, including Yushin, Betge, and Ovtsyn Glaciers, descend from the southern side of the ridge and the flanks of Mount Steller and flow toward Bering Glacier. During the last half of the 20th century and the early 21st century, every one of these glaciers has thinned and retreated. On the southern flank of Mount Steller, glaciers originate in four cirque basins. Before 1950, a single large glacier flowed out of each cirque basin and descended to the base of the mountain, connecting to one of two debris-covered valley glaciers that flowed into Bering Glacier. When observed from the air on 8 August 2001 (fig. 215), glaciers in each cirque basin had separated into many discontinuous, retreating small ice masses, and cirque glaciers from three of the four basins no longer connected to the valley glaciers.

About a dozen small, unnamed, north-flowing glaciers descend from the northern side of Waxell Ridge into Waxell Glacier to nourish the Bagley Ice Valley. The longest is about 5 km long. Although some show evidence of a little thinning, all reach the Bagley Ice Valley. When observed from the air on 12 August 2001, all were still snow covered.

Glaciers of the Southern Side of Juniper Island

Juniper Island is the 43-km-long east-west–trending ridge that separates the eastern Bagley Ice Valley from Jefferies Glacier. About a half dozen small, unnamed, south-flowing cirque glaciers descend from the ridge. All are rapidly retreating and thinning, and nearly all have lost contact with the Bagley Ice Valley (fig. 216). A conspicuous trimline, with a height of less than 50 to nearly 100 m, extends along the western third of the ridge.

Figure 215.*—8 August 2001 view of Mount Steller showing the retreat of a number of unnamed glaciers from nearly every cirque basin on the south side of the mountain. When first observed by the author in 1974, each cirque basin was covered by ice and snow, and all of the glaciers descending from Mount Steller were in contact with the Bering Glacier. Photograph by Bruce F. Molnia, U.S. Geological Survey.*

Figure 216.*—12 August 2001 oblique aerial photograph of the Juniper Island area. View of the north side of the Bagley Ice Valley showing several rapidly retreating and thinning cirque glaciers. Photograph by Bruce F. Molnia, U.S. Geological Survey.*

Figure 217.—Close-up photograph on 13 August 1992 of an in situ *mollusk (*Petricola*?) located within the beds of the Yakataga Formation that yielded a ¹⁴C-calibrated age of 4,860 yr B.P. The presence of the bivalve mollusk indicates that, at that time, the Bering Glacier must have been significantly smaller in area than it is at present. Photograph by Bruce F. Molnia, U.S. Geological Survey.*

Holocene History of Bering Glacier

Much of what we know about the Holocene history of Bering Glacier is derived from (1) radiocarbon dating (^{14}C) of subfossil wood, peat, shells, and organic debris recovered from glacial and fluvioglacial sediment exposed by late 20th century retreat of Bering Glacier (Molnia and Post, 1995) and (2) dendrochronological analysis of subfossil wood by Wiles and others (1999), who developed a floating tree-ring calendar dating to A.D. 1020. Wiles and Calkin (1994) have also shown the benefit of using tree-ring analysis to construct records of past glacial fluctuations in the late Holocene from boreal forest regions of Alaska.

Molnia and Post (1995) constructed a late Pleistocene to present-day history for Bering Glacier. Between 15,000 and 12,000 yr B.P., the large late Pleistocene glacier that covered much of the Gulf of Alaska continental shelf began to retreat (Molnia, 1986). Interpretation of ^{14}C-dated basal peat samples collected from bogs and fens developed on glacially eroded bedrock basins suggests that, by about 10,000 yr B.P. (if not earlier), the entire Gulf of Alaska continental shelf was ice free, and remnants of the large Pleistocene glacier that filled Bering Trough had retreated into the Chugach Mountains. At this time, the ancestral Bering Glacier did not have a piedmont lobe.

For about the next 5,000 years, the glacier remained in a retracted position. ^{14}C-dated core samples from bogs and from loose marine shells of about 8,000 yr B.P., all from the Tashalich Arm area, suggest that open-water marine conditions existed in the area of the piedmont lobe. An exposure of the Yakataga Formation containing an in place mollusk (*Petricola*?) emerged from under the pre-1993-surge terminus of the glacier in 1992 (fig. 217). These mollusks lived in abandoned circular borings made by another bivalve mollusk of the family Pholadidae near the limit of low tide. These shells yielded a ^{14}C-calibrated age of 4,860 yr B.P. Because the clam shells are preserved in perfect condition, it appears that they were buried quickly, possibly in outwash sand and gravel, by an advance of Bering Glacier into marine waters around 5,000 yr B.P. or by rapid sedimentation following a tectonic event. Shortly after the death of the clams, their entombing strata were uplifted above sea level.

^{14}C dates were also obtained from the top and bottom of a 5-m-thick peat deposit (fig. 218), which was composed of leaves, bark, twigs, and seeds, interbedded with thin (1 mm to 2 cm) silt layers, and located adjacent to the outcrop containing the shells. The peat deposit, which lies just above the level of the Yakataga Formation outcrop, was exposed by glacier recession

Figure 218.—13 August 1992 photograph of part of a 5-m-thick peat deposit located on the northeast side of Tashalich Arm. The ¹⁴C age of the deposit ranges from 2,970 yr B.P. (top) to 4,360 yr B.P. (bottom). Photograph by Bruce F. Molnia, U.S. Geological Survey.

between 1988 and 1993 and occurs about 0.5 km north of the shell-bed exposure described above. The top and bottom of the deposit (elevation of ~3 m) had ^{14}C ages of 2,970 yr B.P. (top) and 4,360 and 4,130 yrs B.P. (bottom).

Interpretation of the Tashalich Arm deposits and the bog chronology suggests that Bering Glacier was retracted for most, if not all, of the middle Holocene (fig. 219*A*). It certainly was not in an advanced position between 4,400 and around 3,000 yrs B.P. Molnia and Post (1995) concluded that, following uplift, the Tashalich Arm area was a lowland bog occasionally inundated by marine or fluvial waters. If a glacial advance was responsible for the burial of the Yakataga Formation outcrop, it must have been short lived. The start of the accumulation of peat less than 500 years after the death of the clams indicates that local conditions were favorable for the growth of abundant vegetation.

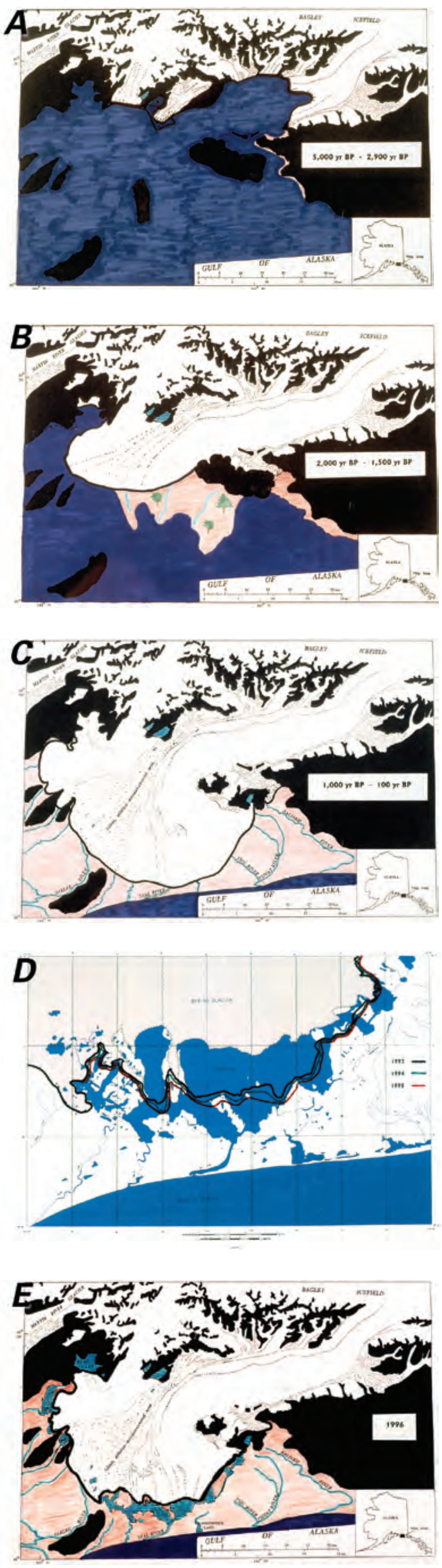
Warm (peat-forming) conditions continued for at least 1,200 years. Sometime after 2,970 yr B.P., the Tashalich Arm peats were buried by outwash sand and gravel deposits, suggestive of the first advance of Bering Glacier in more than 7,000 years. Gravel deposition terminated about 2,000 yr B.P.

Following gravel deposition, a forest, the *Major Forest Bed* (MFB) of Muller and others (1991) grew to near maturity. Although subsequent glacial erosion and overriding by glacier ice have removed much of the forest, enough remains to identify in place trunks of spruce (*Picea* sp.) and hemlock (*Tsuga* sp.), many 200 years old and 40 cm in diameter (Muller and others, 1991). The similarity in both the age and the general level of the trees, usually about 10 m (range of 4–20 m) above the level of Vitus Lake, indicates that most, if not all, of the foreland was stable at the time the forest became established. Many ^{14}C ages of tree stumps from various locations around eastern Vitus Lake indicate that the trees were growing between approximately 2,000 and approximately 1,500 yr B.P. (A.D. 34 – A.D. 539) In all, 45 tree-ring series from 36 MFB trees indicate that the MFB grew for 299 years from A.D. 277 through A.D. 525 (Wiles and others, 1999). Growth in some places was sustained until about 1,500 yr B.P. (fig. 219*B*). In some localities, silts up to 2 m thick buried and killed the trees. Floating calendar dating indicates that most or all of these trees then were killed by the deposition of up to 3 m of fine silt around their trunks during a period of 10 years or less (Wiles, written commun., 1994).

Soon thereafter, during the late 5th century through the early 6th century A.D. (around 1,400 yr B.P.), the standing trees were sheared off, presumably at the new ground level, quite probably by an advance of the glacier. As much as 20 m of layered outwash sand and gravel was deposited on top of the sheared forest, suggesting that a nearby glacier was actively advancing. Near Nichawak Mountain, a ^{14}C age of 1,250±60 yr B.P. obtained from a sheared log suggests that the margin of the Steller Lobe also was advancing at this time.

In several localities around the margin of Bering Glacier, outwash-plain sequences contain at least two more young forest layers. The more prominent, exposed on the western shore of Tashalich Arm where more than 100 stumps have been counted and in canyons on islands in Vitus Lake, grew around 900 to 1,000 years ago (fig. 219*C*). Nearby, a forest overrun by the glacier's last major advance includes some of the biggest trees exposed by glacier recession in Alaska. The largest, a spruce called *Giant Log*, has more than 375 growth rings 6 m above its roots and is approximately 1.6 m in maximum diameter. Tree-ring chronology indicates this spruce lived during the period A.D. 1256 to 1631 (Wiles and others, 1999). This log and several others nearby appear to have been directly buried by a glacier advance. Trees lived along the eastern margin of Vitus Lake from at least A.D. 1020 until A.D. 1631 The youngest vegetation found in outwash gravels is a small number of in place alder bushes (*Alnus* sp.) that yield ^{14}C ages of about 400 yr B.P.

► ***Figure 219.****—A sequence of five sketch maps that depict positions and events in the Holocene history of the Bering Glacier.* **A***, Map showing the retracted position of the terminus of Bering Glacier from approximately 5,000 yr B.P. until approximately 2,900 yr B.P. It is likely that Bering Glacier did not expand beyond the terminus position shown here at any time during the first 7,000 years after its latest Pleistocene retreat into the mountains.* **B***, Map showing the first Holocene advance of the terminus of Bering Glacier some time after approximately 2,000 yr B.P. until approximately 1,500 yr B.P. This advance is documented by several forest deposits consisting of numerous rooted trees sheared off by an ice advance.* **C***, Map showing the maximum position attained by the terminus of Bering Glacier during the Holocene. Dendrochronological evidence suggests that the Bering Glacier reached this position approximately 1,000 yr B.P. and remained at this maximum position until the first decade of the 20th century.* **D***, Sketch map showing the position of Bering Glacier's terminus before and at annual intervals (at the end of each summer) during the 1993–1995 surge. The north shore of Vitus Lake marks the area occupied by the Bering Glacier before the onset of the surge at the end of summer 1992.* **E***, Sketch map showing the position of the terminus of Bering Glacier in 1996 following the end of the 1993–1995 surge. With the exception of Arrowhead Island, which was still essentially ice covered, the 2002 position of the Bering Glacier was similar to that of the ice margin at the end of summer 1992 (*D*). Larger versions of these figures are available online.*



During the past 1,500 years, the margin of Bering Glacier has fluctuated, advancing in some places and retreating in others. The net result is that the "Little Ice Age" maximum position, reached at many locations within the past 250 to 1,250 years, represents its greatest advance for at least the last 7,000 years. This advance produced the neoglacial maximum moraine, which rims the glacier. A sheared spruce tree exposed along the western side of the Seal River, where the river cuts through the outer edge of the neoglacial maximum moraine, yields a calibrated ^{14}C age of 1,053±50 yr B.P., suggesting that the earliest formation of the end moraine occurred no more than about 1,000 years ago.

Elsewhere, growth-ring counts of trees growing on the neoglacial terminal moraine, sampled between 1991 and 1993, yield maximum dendrochronological ages of more than 250 years (near Oaklee River) and more than 180 years (south of Hanna Lake). The oldest living trees growing on the terminal moraine at the edge of the CMMB near Tashalich Arm are around 130 years old. These ages indicate that the active glacier margin may have remained close to the moraine until the middle 18th to early 19th centuries. Recession from the maximum position probably began within the last 200 to 250 years and had reached as much as 12 km before the advent of the 1993 surge.

In most cases, early 20th century positions of the terminus of Bering Glacier are available from early USGS maps and reports. Additional descriptions and maps by D.J. Miller (1958, 1961) provide details about the glacier in the 1940s and 1950s. Oblique aerial photographs taken in 1938, 1946, and 1948 and many vertical aerial photographs acquired since 1950 have been useful for mapping recent end and recessional moraines and for determining positions of the terminus. Comparing terminus positions shown on the early USGS maps with the 1993 pre-surge terminus position (fig. 219*D*) reveals that parts of the Bering Lobe have retreated as much as 12 km and thinned by more than 200 m, whereas parts of the Steller Lobe have receded a maximum of 4 km. The location of the glacier terminus during the 1993–95 surge (fig. 219*D)* and at the maximum reached before the latest retreat (fig. 219*E*) are derived from aerial photography.

Copper River Drainage Segment (Including Glaciers That Drain Directly into the Copper River Delta)

A number of large glaciers making up part of the eastern shoreline of the lower Copper River drain into it or into the Copper River Delta (see fig. 208). A braided stream, the Copper River is the largest river that drains into the Gulf of Alaska. The Copper River is very turbid and, typically, its sediment plume extends far out into the Gulf of Alaska (see figs. 208*A*, *C*). Reimnitz (1966) determined that its typical suspended sediment load exceeds 1.0×10^{8} kg a^{-1}. Hence, the Copper River transports more sediment than any other river in Alaska, including the Yukon River. About 1905, when large copper deposits were discovered in the Wrangell Mountains, a route to transport the copper ore to the port of Cordova was needed. Much of the route that was selected was within the Copper River valley and had to pass through a reach that contained many large glaciers (Post, 1976). Between 1906 and 1910, the Copper River and Northwest Railway was constructed at a cost of about 20 million dollars. Mining officials were concerned that the railroad could easily be disrupted by advances of other glaciers and by changes in the glaciers on which some of the tracks had been laid. Through the 1930s, work crews continuously struggled to maintain the track and compensate for glacier flow and floods.

Martin River Glacier–Martin River–Lower Copper River–Bremner River–West Fork Tana River–Tana Glacier Subdivision

The Martin River Glacier–Martin River–lower Copper River–Bremner River–West Fork Tana River–Tana Glacier subdivision, located east of the Copper River, is the southeastern part of the Copper River drainage segment. It contains a number of large valley glaciers, including many that drain westward into the lower Copper River and the Copper River Delta. The subdivision has a maximum length of about 95 km and a maximum width of about 95 km. In this subdivision alone, an estimated area of 1,750 km^2 is covered by glaciers, the largest being the combined Jefferies and Tana Glaciers, which feed many unnamed glaciers. Large glaciers in the north-central part of this subdivision not connected to the Bering Glacier System — such as Bremner, Fan, Martin River, Miles, and Wernicke Glaciers — are nourished by a series of interconnected accumulation areas west and north of the Bagley Ice Valley.

All glaciers (named and unnamed) listed have approximate lengths greater than 7 km. The areas of Martin River, Slide, and Johnson Glaciers are according to Field (1975b, p. 461–463). All other glacier areas have been estimated by the author. Glaciers in this subdivision include:

- Martin River Glacier, the source of the Martin River (48 km, 290 km^2)
- Slide Glacier, named for a large 1964 earthquake-produced rockslide, but previously called *Sioux Glacier* (10 km, 15 km^2)
- Johnson Glacier (11 km, 26 km^2)
- An unnamed glacier (9 km, 13.5 km^2)
- McPherson Glacier (12 km, 15 km^2)
- Miles Glacier (52 km, 225 km^2)
- Van Cleve Glacier, which drains into Van Cleve Lake, an ice-dammed lake on the northern side of Miles Glacier (20 km, 30 km^2)
- Wernicke Glacier, located at the head of the Wernicke River (33 km, 60 km^2)
- Fan Glacier, which has both an eastern and a western terminus (34 km, 60 km^2)
- An unnamed glacier (12 km, 14 km^2)
- An unnamed glacier (7 km, 10 km^2)
- Bremner Glacier, which has three distinct outlets: Middle Fork Lobe, North Fork Lobe, and Tana Lobe (44 km, 150 km^2)
- Tana Glacier (68 km, 100 km^2), including the western part of the Jefferies Glacier (55 km, 300 km^2), a complex glacier with many outlet glaciers that is contiguous with the Tana Glacier and that is an eastern source of the Bagley Ice Valley.

Ragged Mountains

Less than a dozen small unnamed glaciers exist on the western flank of the Ragged Mountains, heading at elevations between 730 and 1,000 m. The longest is less than 1 km in length. All show conspicuous evidence of thinning and retreat. Several small glaciers, shown with lengths of approximately 400 m on the USGS 1:250,000-scale Cordova topographic quadrangle map (1953) (appendix A), have disappeared.

Martin River Glacier

Martin River Glacier has a length of 48 km and an area of 290 km^2 (Field, 1975b, p. 461) and drains into the Copper River by way of its principal distributary, Martin River (see fig. 213). It has a large stagnant terminus, in places fronted by a large ice-marginal lake formed by retreat during the 20th century. Elsewhere, the terminus is composed of stagnant ice thermokarst features or supports a mature spruce forest. The terminus of Martin River Glacier has three distinct lobes separated by bedrock ridges. The 7-km-long southeastern lobe, named Kushtaka Glacier, ends on an outwash plain north of Kushtaka Lake. The 3-km-long central lobe, called the *Charlotte Lobe* by Reid and Clayton (1963), is the smallest of the lobes. The unnamed southwestern lobe is more than 7 km wide. Like nearby Bering Glacier, the terminus of Martin River Glacier is presently stagnant, downwasting, and slowly retreating. The lower 11 km of the glacier is covered by an ablation moraine. This morainic debris cover ranges in thickness from 0.3 to 6.0 m, the average thickness being 0.7 m (Reid and Clayton, 1963). Clayton (1964) attributed the ablation till to concentration of subglacial debris brought to the surface by thrusting. The glacier's 4-km-wide marginal zone hosts a number of funnel-shaped sinkholes, most supporting small thermokarst lakes. Reid and Clayton (1963) and Clayton (1964) determined that the sinkholes have an average diameter of 250 to 300 m and depths of 30 to 90 m. Many are connected and partly filled with standing water and significantly increase the discharge of meltwater at the terminus when draining (Reid and Clayton,1963).

Clayton (1964) described a multitude of thermokarst features that are present on the surface of the glacier, including ice caves, tunnels, sinking streams, dry stream beds, blind valleys, large springs, natural bridges, sinkholes coalescing to form compound sinkholes, thermokarst windows (unroofed parts of englacial or subglacial streams), and glacial uvulas (enlarged thermokarst windows where a stream flows from one side of the uvula across an exposed gravel bed and into another tunnel on its opposite side).

Martin River Glacier was observed many times from the air by the author between 1974 and 2004. During this 30-year interval, parts of the glacier have thinned by at least 30 m, and many of the sinkholes have expanded and connected. Likewise, the ice-marginal lake adjacent to the debris-covered terminus expanded perhaps 250 to 350 m by the melting of stagnant ice.

In a trend first noted by the author in 1974, many small unnamed glaciers — some former tributaries to Martin River Glacier, located adjacent to and along the northern margin of the glacier — have continued to thin and retreat. Some smaller cirque glaciers have disappeared (fig. 220).

Figure 220.*—North-looking oblique aerial photograph on 31 July 1999 of a pair of small, retreating unnamed glaciers located adjacent to the north side of Martin River Glacier. Many small cirque glaciers in the Chugach Mountains are rapidly retreating. Photograph by Bruce F. Molnia, U.S. Geological Survey.*

Slide Glacier

Slide Glacier, located in a side valley adjacent to and west of Martin River Glacier, has a length of 10 km and an area of 15 km^2 (Field, 1975b, p. 461) (fig. 221). Originally named *Sioux Glacier*, its current name is the result of several large 1964 earthquake-produced rockslides that covered approximately 17 percent of its accumulation area and approximately 90 percent of its ablation area with 2 m or more of debris (Post, 1967a). According to Post, three of the largest slides contained about 2.4×10^7 m^3 of rock. Reid (1969) found that, before the earthquake, the glacier had been shrinking. From the beginning of the 20th century until 1964, the glacier had thinned by as much as 35 m. One and one-half years after the slides occurred, differential melting had caused the slide-debris-covered portion of the glacier to be 13 m higher than the adjacent bare ice. The next year, the elevation difference was 18 m. Hence, over three summer ablation cycles (1964, 1965, 1966), the average difference in ablation rate exceeded 5 m a^{-1}. When the author observed the glacier from the air on 16 August 2000, he noted a small amount of retreat along the debris-covered terminus; a fresh trimline along the margin of the bare ice portion of the glacier suggested that it was continuing to thin.

Figure 221.*—Two north-looking oblique aerial photographs of Slide Glacier, with 1964 earthquake-produced rock avalanches covering much of its terminus area.* **A**, *6 October 1974 view shows the debris-covered terminus region standing higher than the bare ice to its north. Photograph by Bruce F. Molnia, U.S. Geological Survey.* **B**, *13 August 1990 view shows a second 1964 rockslide moving into the terminus region. Photograph by Larry Mayo, U.S. Geological Survey.*

Johnson Glacier

Johnson Glacier has a length of 11 km and an area of 26 km^2 (Field, 1975b, p. 461). An unnamed glacier located to its north has a length of 9 km (Field, 1975b, p. 462) and an area estimated by the author at around 13 km^2. Both Johnson Glacier and the unnamed glacier showed signs of thinning and retreat when the author observed them from the air on 16 August 2000. Both were fronted by ice-marginal lakes and outwash plains indicative of significant retreat during the later half of the 20th century. Johnson Glacier was actively calving large icebergs into its proglacial lake. The size of its ice-marginal lake suggests at least 2.5 km of retreat from its late 1950s terminus position.

McPherson Glacier

When Martin photographed McPherson Glacier in 1910, it had a length of 12 km (Field, 1975b, p. 462) and an area estimated by the author at approximately 15 km^2; it extended to the floor of the Sheep Creek valley. By the 1960s, retreat of more than 2 km resulted in the disappearance of its terminal tongue and the exposure of an approximately 2.4×1.0-km lake basin that served as the reservoir for an ice-dammed proglacial lake. Several times during the operation of the Copper River and Northwestern Railroad, jökulhlaups washed out part of the railroad. Similarly, one or more jökulhlaups in 1962 or 1963 washed out 1.5 km of the Copper River Highway (Post and Mayo, 1971). An additional jökulhlaup occurred in 1965 (Post, 1967a). When it was observed by the author from the air in 2000, McPherson Glacier had so significantly thinned and retreated that it could no longer impound water. Much of the floor of its former proglacial lake bed was covered with vegetation.

Miles Glacier

Miles Glacier has a length of 52 km (Field, 1975b, p. 462) and an area estimated by the author at about 225 km^2. It retreated a maximum of 7.2 km between the 1880s and the late 1960s (Field, 1975b) and approximately another 2.9 km through 2004. Late 19th and early 20th century retreat produced Miles Lake, originally an elliptical ice-marginal lake along the southwestern margin of the glacier and now a near-circular body of water approximately 8 km in diameter that fronts the entire terminus of Miles Glacier. Large numbers of icebergs calve off the terminus of the glacier and float across the lake, passing under Million Dollar Bridge as they drift downstream. To protect the bridge from Miles Glacier's icebergs, massive steel and concrete iceberg deflectors were installed upstream of the bridge's main supports when it was built in l909.

Before 1840, an advance of Miles Glacier resulted in its merging with Childs Glacier. This merger displaced the Copper River to the western side of its valley. Tarr and Martin (1914) suggested that, at that time, the river probably flowed in a tunnel or in a gorge that it cut through the glacier ice. A relict part of this expanded terminus is currently located on the western side of the river, a result of the Copper River's cutting a new channel through this stagnant moraine during the 1940s. Between l885 and 1888, the terminus of Miles Glacier advanced to a position about 125 m from the site where Million Dollar Bridge would eventually be built. From 1888 to 1908, a 20-year-period of rapid retreat ensued. Miles Glacier's terminus retreated as much as 4 km, an average retreat rate of 200 m a^{-1}. A 2-year advance began in 1908; by 1910, the terminus had advanced approximately 1.25 km, an average advance rate of 610 m a^{-1}. This advance occurred in spite of a calving event in July 1909 in which an approximately 1-km^2 section of Miles Glacier broke from the terminus and produced icebergs that completely filled Miles Lake (Tarr and Martin, 1914). When Miles Glacier was next described in 1931 by Wentworth and Ray (1936), it had retreated at least 1 km. Field (1975b) reported that, by 1957, the glacier had retreated an additional 800 to 1,800 m (see also AGS

Glacier Studies Map No. 64–3–G6) (Field, 1965) and that, between 1957 and 1968, maximum retreat was an additional 2 km.

Since 1991, the author has observed Miles Glacier from the air annually, most recently on 16 October 2002. During that period, the glacier has continued to retreat (fig. 222) and thin along its margins, losing contact with several tributaries and thinning to the point that former small ice-marginal lakes located along the glacier's northern margin no longer fill with water.

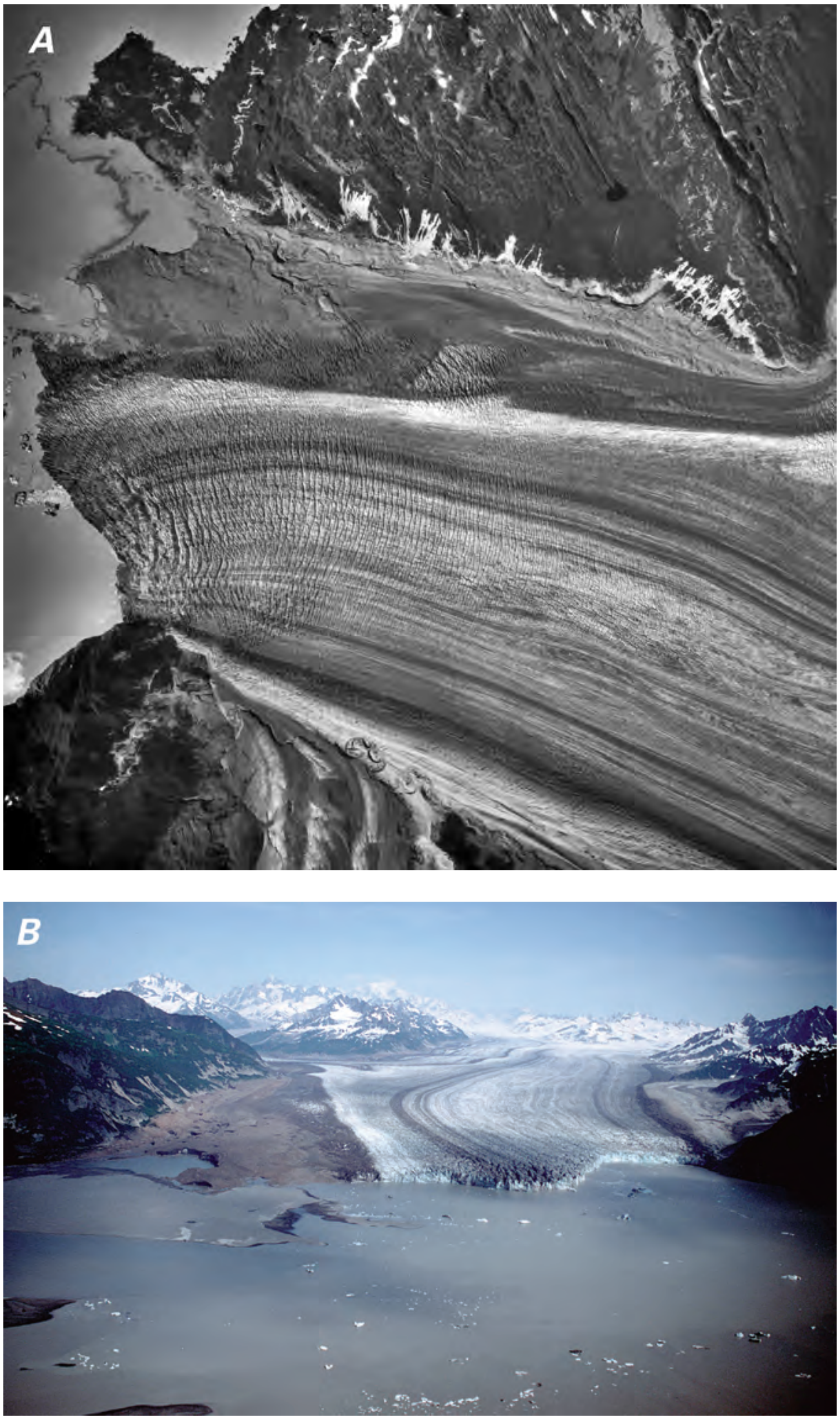


Figure 222.*—Two photographs of the terminus area of Miles Glacier in Miles Lake in the late 20th and early 21st centuries.* **A**, *13 August 1994 vertical aerial photograph of the lower 6 km of Miles Glacier shows multiple evidence of thinning and retreat. An unnamed north-flowing tributary barely reaches the margin of Miles Glacier. USGS photograph no. 94V3–175 by Austin Post, U.S. Geological Survey.* **B**, *12 August 2001 east-looking oblique aerial photograph shows that the northern part of the terminus of Miles Glacier is characterized by stagnant, debris-covered ice, whereas the southern side is dominated by an active calving margin. Photograph by Bruce F. Molnia, U.S. Geological Survey.*

Van Cleve Glacier

Van Cleve Glacier has a length of 20 km (Field, 1975b, p. 462) and an area estimated by the author at about 30 km^2; it drains into Van Cleve Lake, an ice-dammed lake located adjacent to the northern side of Miles Glacier, approximately 7 km east of its terminus (fig. 223). The proglacial lake of Van Cleve Glacier was the source of several early 20th century jökulhlaups that significantly impacted the railroad (Post and Mayo, 1971) and, more recently, the Copper River Highway. Tarr and Martin (1914) thought that calving of Miles Glacier's terminus was influenced by flooding from the lake. Twentieth century retreat of Van Cleve Glacier has lengthened the basin, but lowering the surface of both Miles and Van Cleve Glaciers has reduced the proglacial lake's capacity; it continues, however, to be a source of jökulhlaups.

Large Glaciers North of Miles Glacier

Four large glaciers—Wernicke, Fan, Bremner, and Tana—flow down the northern side of the mountains and ridges west and north of the Bagley Ice Valley. All four glaciers show evidence of continuing thinning and retreat and the presence of stagnant ice.

Wernicke Glacier

Wernicke Glacier, the source of the Wernicke River, is located at the head of an approximately 12-km-long outwash plain that ends at the Copper River. It is located in the next large valley north of Miles Glacier and has a length of 52 km (Field, 1975b, p. 462) and an area estimated by the author at about 60 km^2. Along with Fan Glacier and the three lobes of the Bremner Glacier, it drains the northern side of the unnamed mountain ridge that contains the summits of Mount Tom White (3,419 m) and Mount Hawkins (3,140 m). Although its terminus positions have been compared on oblique aerial photographs taken

Figure 223.*—7 September 1994 vertical aerial photograph of the southern part of Van Cleve Lake where it is dammed by an unnamed distributary of Miles Glacier. This ice tongue, composed of ice flowing from both upglacier and downglacier, regulates both the depth and volume of Van Cleve Lake. In the 15 years between a photograph taken by Austin Post in 1969 (USGS photograph no. 69–R2–214) and this photograph, little change has occurred in the position of the terminus of the distributary glacier, but a 25-m-high strand zone along the shoreline of the lake indicates a thinning of the ice dam. The numerous folds in the medial moraines indicate the complex flow interaction between the western and adjacent eastern parts of the ice tongue. Note the right-angle bend made by the Miles Glacier medial moraine on the western side of the ice that flows into the unnamed distributary. USGS photograph no. 94–V4–236 by Robert M. Krimmel, U.S. Geological Survey.*

by Bradford Washburn in 1938, the U.S. Army Air Forces in 1941, the USAF in 1957, and the USGS in 1964 (Field, 1975b), Wernicke Glacier has not been the subject of scientific investigations (nor have most of the glaciers in this part of the Chugach Mountains). Although Field (1975b) did not quantify the amount of change that he observed, he reported that the end of the lower limit of ice has receded upglacier as the area covered by moraine has been extended. When the author observed it from the air on 16 August 2000, the terminus region of Wernicke Glacier showed significant evidence of recession and thinning and the presence of stagnant ice. Many former tributaries have retreated to the upper reaches of their cirques or have disappeared altogether.

Fan Glacier

Fan Glacier bifurcates and has both an eastern and a western terminus; it has an approximate length of 34 km (Field, 1975b). The western terminus is located at the head of the South Fork of the Bremner River; the eastern terminus drains into the Middle Fork of the Bremner River. Fan Glacier has received little scientific investigation. However, a comparison by Field (1975b, p. 323) of terminus positions as depicted on oblique aerial photographs taken by Bradford Washburn in 1938, the U.S. Army Air Forces in 1941, the USAF in 1957, and the USGS in 1964 found a "recession from outer moraines, appreciable lowering of the ice surface, and progressive stagnation in the terminal area." Field (1975b, p. 323) also reported that "all the smaller glaciers, some of which were formerly tributaries, also show evidence of marked recession." A similar comparison by the author of the USGS Bering Glacier 1:250,000-scale topographic quadrangle map (1959) (appendix A), an AHAP false-color infrared vertical aerial photograph acquired on 18 August 1978 (fig. 224*A*), and oblique aerial photography of the terminus area of the glacier obtained by the author on 16 August 2000 shows significant evidence of recession and thinning and the presence of stagnant ice (fig. 224*B*).

East of the eastern terminus of Fan Glacier are several unnamed glaciers that formerly were tributaries to a connected Fan Glacier–Middle Fork Lobe of the Bremner Glacier ice mass (fig. 225). The western unnamed glacier

***Figure 224.**—Two aerial photographs of Fan Glacier showing evidence of ongoing retreat, thinning, and stagnation. **A**, 18 August 1978 AHAP false-color infrared vertical aerial photograph of the debris-covered western terminus of Fan Glacier. All of the north-flowing glaciers in the photograph, including those that are former tributaries of Fan Glacier located on the south side of the glacier and those in the unnamed mountain ridge to the north, are actively retreating. Note the trimline and area of stagnation on the north side of Fan Glacier. AHAP photograph no. L111F6217 from the GeoData Center, Geophysical Institute, University of Alaska, Fairbanks, Alaska. **B**, see facing page.*

Figure 224B.—*16 August 2000 east-looking oblique aerial photograph of the debris-covered western terminus of Fan Glacier. The northern and southern margins show evidence of recent thinning. Numerous thermokarst features characterize the stagnant-ice terminus. Photograph by Bruce F. Molnia, U.S. Geological Survey.*

▼ **Figure 225.**—*18 August 1978 AHAP false-color infrared vertical aerial photographic mosaic of the debris-covered Bremner Glacier with its three distinct outlets: Middle Fork Lobe, North Fork Lobe, and Tana Lobe. All of the termini of Bremner Glacier show evidence of stagnation, thinning, and active retreat. Four former tributaries to the Middle Fork Lobe are shown to its south. Note the elevated lateral moraine and trimline north of the North Fork Lobe. AHAP photograph nos. L111F6212 and L111F6214 from the GeoData Center, Geophysical Institute, University of Alaska, Fairbanks, Alaska.*

previously flowed into the eastern Fan Glacier terminus; it has a length of 13 km (Field, 1975b, p. 462) and an area estimated by the author of about 14 km^2. Adjacent to it on the east is another unnamed glacier that terminates between the glaciers; it has a length of 9 km (Field, 1975b, p. 463) and an area estimated by the author at about 10 km^2. To the east of this latter unnamed glacier are two smaller unnamed glaciers. When the author observed them from the air on 16 August 2000, all showed significant evidence of recession and thinning and the presence of stagnant ice.

Bremner Glacier

Bremner Glacier has three distinct distributaries (Middle Fork Lobe, North Fork Lobe, and Tana Lobe); it has a length of 44 km (Field, 1975b, p. 463) and an area estimated by the author at about 150 km^2. Like adjacent Fan Glacier, it has received little scientific investigation. A similar comparison by Field (1975b, p. 323) of terminus positions depicted on oblique aerial photographs taken by Bradford Washburn in 1938, the U.S. Army Air Forces in 1941, the USAF in 1957, and the USGS in 1964 found a "recession from outer moraines, appreciable lowering of the ice surface, and progressive stagnation in the terminal area." Field (1975b, p. 323) also reported that "all the smaller glaciers, some of which were formerly tributaries, also show evidence of marked recession." Likewise, a similar comparison by the author of the USGS Bering Glacier 1:250,000-scale topographic quadrangle map (1959) (appendix A), an AHAP false-color infrared photograph acquired on 18 August 1978 (fig. 225), vertical aerial photographs, and oblique aerial photographs of the terminus area of the glacier obtained by the author on 16 August 2000 shows significant evidence of recession and thinning and the presence of stagnant ice.

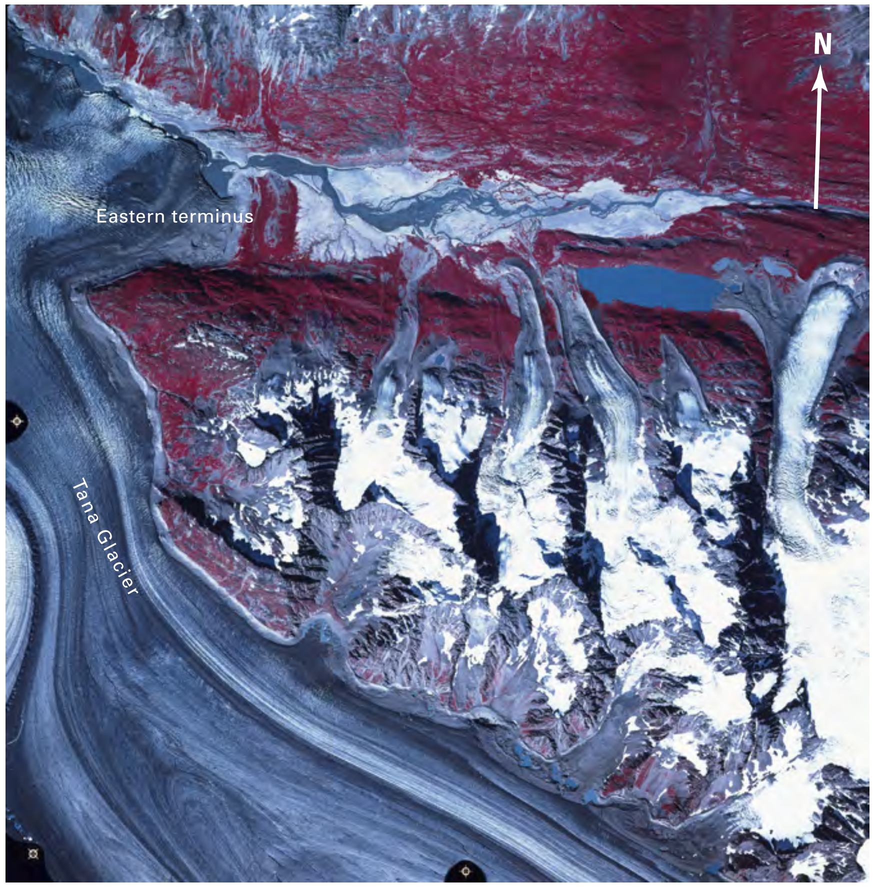
Tana Glacier

Tana Glacier has a length of 68 km (Field, 1975b, p. 463) and an area estimated by the author at about 100 km^2, including the western part of Jeffries Glacier. Its primary sources are the valley north and east of Juniper Island, including the western end of an accumulation area of the Jefferies Glacier; the western Bagley Ice Valley; and an unnamed east-flowing tributary that drains the eastern part of the accumulation area that feeds the Bremner Glacier. Much of the lower glacier is flanked by prominent trimlines and freshly exposed lateral moraine. Brabb and Miller (1962) reported that, during the late Wisconsinan, Tana Glacier may have been more than 200 m thicker than it was in the late 1950s.

The glacier has two distributaries. The primary northern terminus is stagnant and is characterized by a network of debris-covered thermokarst pits and open water. The eastern terminus is similar, showing multiple evidence of stagnant ice, thinning, and retreat, as an 18 August 1978 AHAP photograph shows (fig. 226). In 1975, Field (1975b, p. 324) reported that aerial photographs of the glacier obtained in 1938, 1957, and 1960 "show no appreciable change in the position of the terminus, but there appears to have been some surface lowering and progressive stagnation." Things changed significantly during the later 20th century. Observations of the terminus area made by the author from the air between 1995 and 2001 confirm continued thinning and the presence of stagnant ice in both termini and an increase in the volume of water associated with the two stagnant lobate termini. Behind the termini, the lower 6 km of the glacier is covered by moraine, and the glacier ice appears virtually stagnant. Much of the margin of Tana Glacier shows significant elevated moraines and trimlines, indicative of substantial late 20th century thinning.

Through the first one-third to one-half of the 20th century, the eastern terminus of the glacier dammed the flow of water from Granite Creek, forming Barkley Lake, an ice-dammed lake (Moffitt, 1918) that occupied the lower part of the Granite Creek valley. Annual draining of the lake was through

Figure 226.—18 August 1978 AHAP false-color infrared vertical aerial photograph of the retreating, partially debris-covered terminus of the eastern distributary of Tana Glacier. The former bed of Barkley Lake is visible just east of the distributary terminus, as are a number of former tributaries. AHAP photograph no. L113F6174 from the GeoData Center, Geophysical Institute, University of Alaska, Fairbanks, Alaska.



an ice-marginal channel. By the late 1940s, thinning and narrowing of the glacier resulted in unimpeded flow of Glacier Creek. Stone (1955) reported that, before 1951, the lake was "continuously empty for a minimum of three to five years." A significant jökulhlaup occurred in July 2003.

The Bremner River–Upper Copper River–Chitina River–Tana River Subdivision (Located East of the Copper River)

Every glacier in this subdivision is unnamed. All are less than 7 km long, with the exception of an unnamed glacier located at the head of the Little Bremner River (fig. 227), with a length of 17 km (Field, 1975b, p. 463) and an area estimated by the author at approximately 40 km^2. Many of the glaciers in this area were observed from the air during several overflights made by the author in August 2000; all showed evidence of thinning and retreat.

Glaciers in this area exist in several locations. About two dozen small unnamed glaciers are located west of Tana River and north of North Fork Lobe and Tana Lobe of Bremner Glacier. There are also about 100 small unnamed glaciers from west of Hanagita Peak to the eastern side of Little Bremner River and Tebay River. The largest glacier between Little Bremner River and Tebay River is the previously mentioned 17 km unnamed glacier. It was actively retreating when observed in 2000. There are several other medium sized glaciers, including an unnamed glacier located at the head of Dewey Creek. The earliest photographs of glaciers in this area were made during the first half of the second decade of the 20th century and show that retreat was underway (USGS Photo Library Moffit photograph 517). In the northwestern part of this subdivision, about 40 small unnamed glaciers drain into Canyon Creek and its tributaries or Copper River. All glaciers observed have been retreating for most, if not all, of the 20th century and showed significant evidence of thinning and retreat.

Figure 227.—Photograph by F.H. Moffitt, taken between 1910 and 1915, showing early 20th century evidence of thinning and retreat of an unnamed glacier located at the head of the Little Bremner River. USGS Photo Library photograph Moffitt 515.

The Western Copper River Delta–Lower Copper River–Tasnuna River Subdivision (Located on the West Side of the Copper River)

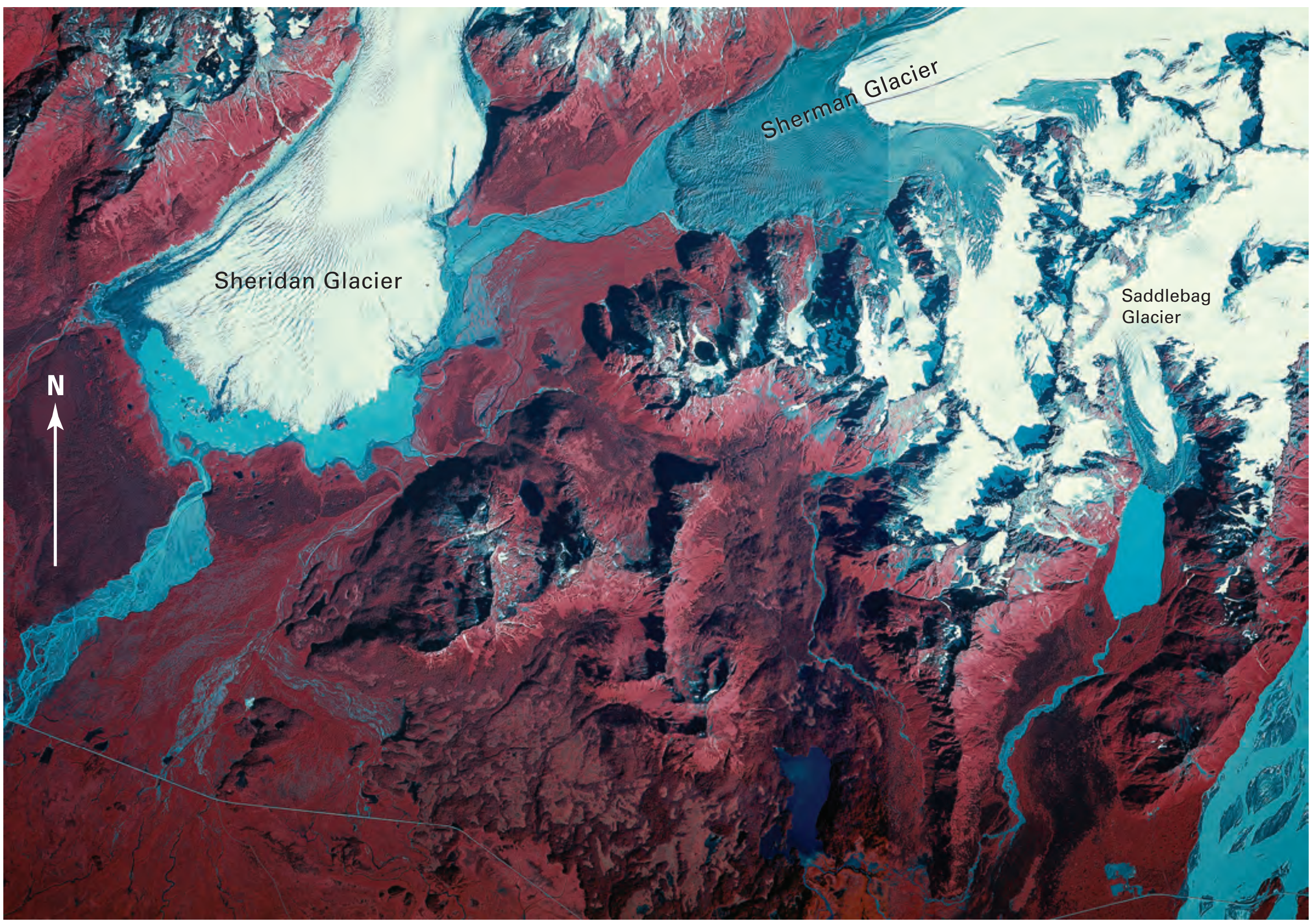
This subdivision contains a number of south-, east-, and north-flowing valley glaciers. Several have played a significant role in the early 20th century history of mineral development and extraction of ore. The Copper River and Northwest Railway was constructed on approximately 9 km of Allen Glacier, about 0.5 km of moraine-covered ice of Grinnell Glacier, and about 0.75 km of the terminal moraine of Heney Glacier. Many jökulhlaups from more than a dozen glaciers in this subdivision disrupted railroad operations. As recently as the summer of 2001, a flood of unknown origin washed out a segment of the Copper River Highway.

In this subdivision, the lengths (>8 km) and areas of glaciers are from measurements made by Field (1975b, p. 467–468) and include: Sherman Glacier (13 km, 55 km^2), Sheridan Glacier (24 km, 101 km^2), Scott Glacier (24 km, 160 km^2), Marshall Glacier (8 km, 8 km^2), Tasnuna Glacier (13 km, 28 km^2), Woodworth Glacier (23 km, 185 km^2), Schwan Glacier (23 km, 131 km^2), Heney Glacier (19 km, 72 km^2), Allen Glacier (31 km, 230 km^2), Childs Glacier (19 km, 100 km^2), and Goodwin Glacier (9 km, 13 km^2).

Saddlebag Glacier

Saddlebag Glacier is the southeasternmost glacier in this subdivision. Figure 228, a 13 August 1982 Alaska AHAP photomosaic of this area, shows the geographic relationship of Saddlebag Glacier and adjacent Sherman and Sheridan Glaciers. When it was photographed by the author on 12 August 2001, the glacier was located at the head of an approximately 3.5-km-long ice-marginal lake, dammed by a large end moraine that was probably deposited

Figure 228.*—13 August 1982 AHAP false-color infrared vertical aerial photographic mosaic of the area between the Copper River (east) and the Glacier River (west) that flows from the Sheridan Glacier. Shown are Saddlebag Glacier (the south-easternmost glacier in this subdivision), Sherman Glacier, and Sheridan Glacier. All show evidence of thinning and retreat. A large 1964 earthquake-generated rock avalanche covers the terminus of Sherman Glacier. AHAP photograph nos. L115F1516 and L115F1517 from the GeoData Center, Geophysical Institute, University of Alaska, Fairbanks, Alaska.*



during the “Little Ice Age” maximum. Field (1975b, p. 345) reported that, in 1899, the terminus of the glacier occupied the lake basin but was “already beginning to withdraw from the end moraine.” Retreat was slow, and an ice tongue filled much of the basin through the late 1950s. Between 12 June 1959, when it was photographed by the USFS, and 25 August 1960, when it was photographed by Post, “a massive break-up of the tongue” occurred and the glacier retreated about 1.5 km. Between 1960 and 2001, approximately 2.0 km of additional retreat occurred, accompanied by significant thinning.

Sherman Glacier

Sherman Glacier (figs. 58, 228, 229) has a length of 13 km and an area of 55 km^2 (Field, 1975b, p. 468). It and adjacent Sheridan Glacier (fig. 228) both drain into the delta of the Copper River through the Glacier River. The surface of the lower 5 to 6 km of Sherman Glacier is covered with rock debris deposited from a massive landslide that occurred during the 27 March l964 Alaska “Good Friday” Earthquake (figs. 58, 228). The powerful earthquake triggered many large rock and debris avalanches that were deposited on the surfaces of more than 50 glaciers in the Chugach Mountains. The Sherman Glacier landslide was among the largest of all, covering an area of about 8.5 km^2, about one-third of the glacier’s ablation area (Post, 1967b). The volume of debris that was deposited on the glacier surface was about 1×10^7 m^3 (Bull, 1969) (fig. 229*A*). When the author visited on 12 August 2000, the debris-covered ice stood about 35 m higher than the bare ice immediately upglacier from it. In the 36 years since the landslide, the 1.3- to 8-m-thick debris layer had insulated the ice beneath it and thus retarded its ablation.

Before the landslide, Sherman Glacier had been retreating. Tuthill and others (1968) described both an outer moraine, dating from about 1880, and an inner moraine, dating from about 1910, within 1.2 km of the 1964 terminus position. In the years between 1910 and 1959, the glacier retreated 970 m, yielding an average retreat rate of about 20 m a^{-1}. During the 5-year period before the earthquake (1959–64), the glacier retreated approximately 230 m, yielding an average retreat rate of approximately 46 m a^{-1}.

Bull (1969) speculated that a surge of this previously receding glacier would occur during the 1970s. He expected rates of advance of 50 to 100 m a^{-1} that would persist for perhaps a decade. As of the author’s last observation from the air in August 2004, the predicted surge had not occurred. Bull and Marangunic (1967, 1968) noted that, in the year following the slide, the ice surface below the slide was lowered between 8 and 10 m. They attributed much of this loss to anomalously high air temperatures resulting from the heating of downglacier air as it passed over the surface of the debris slide. Thus, the heat reflected by the slide contributed to increased melting in the bare ice in the ablation area. By 29 August 1984, 20 years after the landslide, almost all of the bare ice below the slide had melted away, and the debris cover had been carried downvalley about 2 km (fig. 229*C*). The terminus was in nearly the same position in 1984 as it was in 1964, but the debris-covered ice was significantly thicker. All of the landslide debris was derived from one mountain, Shattered Peak. Shreve (1966, 1968) examined the mechanics of the slide and concluded that the debris had moved downglacier on a cushion of air at velocities of up to 75 km h^{-1}. Part of the debris glided over a ridge about 130 m high.

Since 1984, the position of the terminus has changed only minimally, with stream erosion and melting of stagnant ice resulting in a few meters of annual retreat. However, many of the tributaries of Sherman Glacier are actively retreating. By 2002, one unnamed tributary on its northern side had separated into seven smaller glaciers.

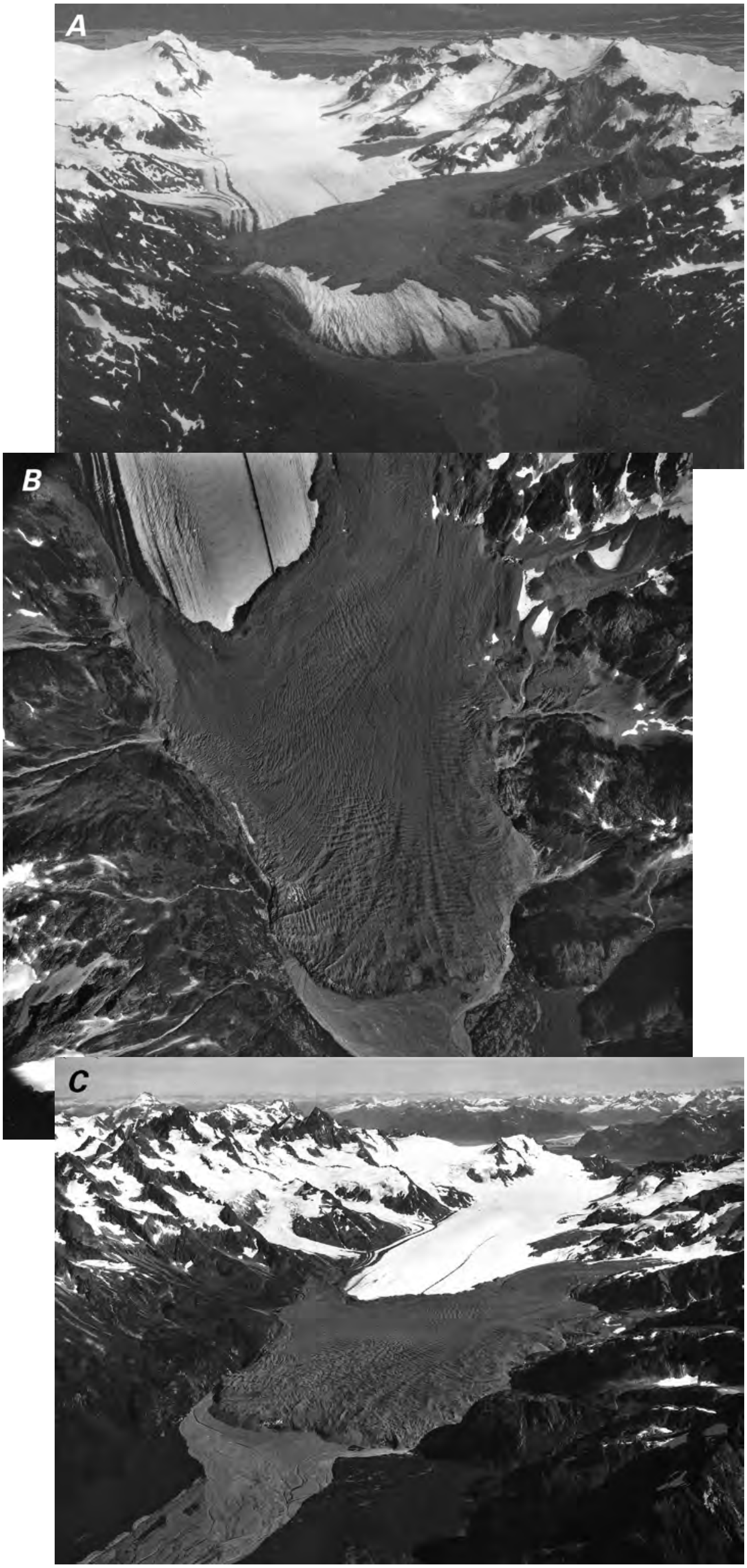


Figure 229.—*Three aerial photographs of the Sherman Glacier.* **A**, *Oblique aerial photograph taken 24 August 1964 of a major rock avalanche that occurred during the 1964 Alaska earthquake and covered about 35 percent of the ablation area (Post, 1967b). The volume of the 1.3 to 8-m-thick debris was about 1×10^7 m^3 (Bull, 1969). The debris layer acted to insulate the buried ice and thus retard ablation, and Bull (1969) speculated that an advance of this previously receding glacier (Field, 1975, p. 347) would occur during the 1970s. Photograph no. K642–108 by Austin Post, University of Washington.* **B**, *22 August 1979 vertical aerial photograph of the expanse of the terminus of Sherman Glacier. In the 15 years since the earthquake, all of the bare ice between the rockslide and the terminus melted away. USGS photograph no. 79–V2–186 by Austin Post, U.S. Geological Survey.* **C**, *By 29 August 1984, the debris cover had been carried down valley by the moving ice about 2 km. Although the terminus was in nearly the same position in 1984 as in 1964, the debris-covered ice was much thicker in 1984. USGS photograph no. 84–R1–222 taken on 29 August 1984 by Austin Post, U.S. Geological Survey. Photographs and caption courtesy of Robert M. Krimmel, U.S. Geological Survey. Photographs taken by the author in August 2000 show little change in the position of the terminus. A lowering of the base level in nearby Sherman Lake has lowered the surface of the outwash plain in front of the glacier. The elevation difference between the covered and bare ice is currently about 35 m.*

Sheridan Glacier

The terminus of Sheridan Glacier is only about 2.5 km to the west of Sherman Glacier. However, Sheridan Glacier, with a length of 24 km and an area of 101 km^2 (Field, 1975b, p. 468), is a much larger glacier. The terminus of Sheridan Glacier is a small retreating piedmont lobe with a width of more than 5 km (fig. 58). At the beginning of the 21st century, nearly all of the margin of the retreating and thinning glacier was fronted by a single large ice-marginal lake (figs. 228, 230).

In 1965, Tuthill and others (1968) examined a series of concentric recessional moraines and related deposits located as much as 2.5 km beyond the present margin of the glacier; by analyzing these deposits, they were able to construct a late Pleistocene-Holocene history for Sheridan Glacier and adjacent Sherman Glacier. Tuthill and others (1968) determined that, at the end of the Pleistocene Epoch, the two glaciers were joined and terminated about 5 km south of the 1965 position of the terminus of Sheridan Glacier. Subsequently, but at an unknown date, Sheridan Glacier retreated to a point north of its present margin. Sheridan Glacier remained in a retracted position until less than about 2,000 years ago, when it readvanced. Sheridan Glacier remained at this location until about A.D. 300, after which it underwent a series of fluctuations in position, culminating in an advance that produced a moraine in about A.D. 1700 (300 years ago). This moraine is the outermost of the concentric recessional moraines that they examined. At least four younger moraines exist between the A.D. 1700 moraine and the present-day margin of Sheridan Glacier.

Tarr and Martin (1914) reported that Seton Karr visited the glacier in 1886 and noted a fresh moraine resulting from a recent advance or stillstand. In 1910, Tarr and Martin (1914, p. 390) observed the glacier and noted that "The presence of thick mature forest up to the very edge of the bulb [bulb-shaped terminus] indicates that the glacier has not been more extensive for a score of years, perhaps for a century." When observed twenty-one years later, Wentworth and Ray (1936) described the margin as having retreated between 100 and 125 m and noted that the surface of Sheridan Glacier was showing signs of thinning produced by ablation. An average retreat rate for this interval would be about 5 to 6 m a^{-1}.

Field (1975b, p. 350), on the basis of his own observations and on the analysis of aerial photographs collected by Bradford Washburn in 1938, the U.S. Army Air Forces in 1941, and the USAF in 1950, concluded that "the massive moraine nearest the present terminus was formed in the early or middle 1930's." In the approximately 30 years between the date of the formation of the moraine and the middle 1960s, Field (1975b, p. 350) reported that "recession has varied from around 200 m on the ridges to perhaps 500 m. in the lake basins." Hence, during this period, retreat rates averaged between 7 and 17 m a^{-1}. Without quantifying the amount of change, Field (1975, p. 351) reported that, through 1971, the glacier "showed considerable further lowering of the ice surface and some recession of the terminus since 1965 and 1968."

Field (1975b, p. 351) described the evolution of the ice-marginal lakes fronting Sheridan Glacier through 1971. He stated that "The lakes were only beginning to form by the recession of the terminus in 1931." Through examination of the USGS 1:63,300-scale Cordova C-4 topographic quadrangle map (1953) (appendix B), he determined that, by 1950, five lakes had formed: one at the western end of the glacier, draining into the Glacier River; two 1-km-wide lakes along the front of the middle part of the terminus; and two small lakes along the eastern margin of the glacier. All had separate outlets. On the basis of his 1965 observations, Field (1975b) noted that the five lakes still maintained their unique existence but that the level of one of the eastern lakes had dropped when an ice dam was removed. By 1971, the drainage of all of the lakes entered the two middle lakes, generally englacially, and all of Sheridan and Sherman Glaciers' drainage flowed through the middle lake's outlet to the Gulf of Alaska.

Recent changes in the evolution of the ice-marginal lake system were studied by Bailey and others (2000). They found that, by 1981, the ice front in the center of the glacier had receded and allowed the two middle lakes and the western lake to combine. This merging was accompanied by about 10 m of downcutting of the outlet through the A.D. 1700 moraine. By 22 August 1979 (fig. 230), the basins of the eastern lakes had filled with sediment, and they had completely disappeared. When photographed by the author in October 1974, an isolated, small eastern lake still remained. Increased availability of sediment after 1964 from Sherman Glacier may have been a factor in the subsequent infilling of the basin.

Unlike any other part of Sheridan Glacier, the southwesternmost margin is debris covered, possibly from a 1964 earthquake-generated landslide. This debris mantle has protected the terminus in this area from retreat. Repeated visits by the author to this part of the glacier between 1974 and 2002 have documented that the rate of thinning is reduced and the rate of retreat is slower than that of the bare ice to the east. Before 1974, a small lake had formed at the western margin of the debris-covered ice. By 2002, it had a maximum length of around 600 m.

On 17 August 2000 (fig. 231), the author observed from the air that a significant amount of calving was occurring along the eastern margin of the glacier. Continued thinning of the glacier may have resulted in a floating ice tongue that was rapidly disintegrating.

A comparison of 1950s map data of the glacier with data obtained during annual airborne profiling surveys in the middle 1990s indicates that Sheridan Glacier thinned by 0.725 m a^{-1} and had a volume decrease of 0.0729 km^3 a^{-1} (K.A. Echelmeyer, W.D. Harrison, V.B. Valentine, and S.I. Zirnheld, University of Alaska Fairbanks, written commun., March 2001).

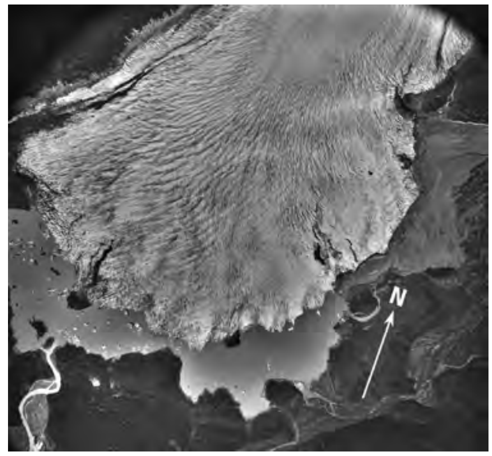


Figure 230.—22 August 1979 vertical aerial photograph shows the terminus of the Sheridan Glacier. The large trimline on the west side of Sheridan Glacier documents more than 50 m of 20th century thinning. USGS photograph no. 79–V2–190 by Austin Post, U.S. Geological Survey. A larger version of this figure is available online.

Scott Glacier

Scott Glacier (fig. 232) has a length of 24 km and an area of 160 km^2 (Field, 1975b, p. 468). Its approximately 2-km-wide terminus is fronted by a braided outwash plain that stretches from valley wall to valley wall. Part of the outwash plain is incised, probably a result of tectonic uplift caused by the 1964 earthquake. When Field (1975b) compared aerial photographs of the terminus area made by Bradford Washburn in 1938 and by Austin Post in the 1960s, he noted that the terminus had retreated about 200 m (a rate of ~7 m a^{-1}) and that a wider marginal zone had developed. When the author photographed the terminus area on 17 August 2000 (fig. 232*B*), the entire ice margin was covered by sediment, the result of in place melting of at least

Figure 231.—17 August 2000 oblique aerial photograph of the ice-marginal lakes developed in front of Sheridan Glacier. East-looking view across the terminus of Sheridan Glacier shows the lake bounded by an early 20th century recessional moraine to the south and the ice margin to the north. A large amount of calving is occurring. The small basin located on the southwest corner of the Sheridan Glacier has yet to connect with the main lake. Photograph by Bruce F. Molnia, U.S. Geological Survey.

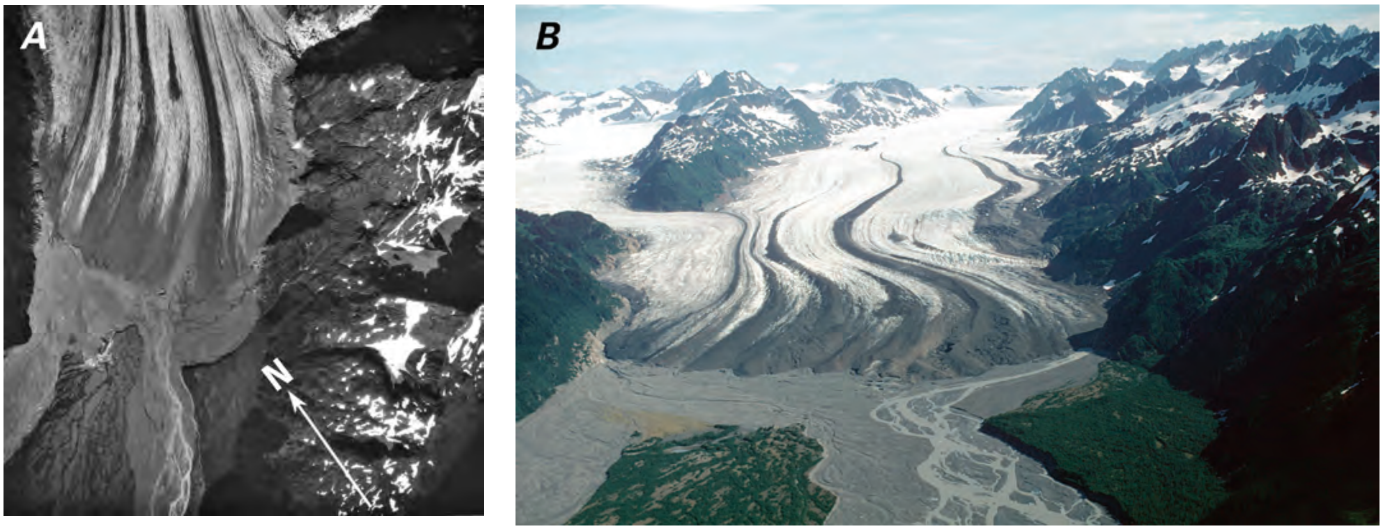


Figure 232.—*Two aerial photographs of Scott Glacier.* **A**, *24 July 1987 vertical aerial photograph of the retreating and thinning terminus of Scott Glacier and the proximal part of its braided outwash plain. Trimlines and several thermokarst pits developing in the terminus region are indicators of the declining health of the glacier. As with Sherman Glacier, the elevation of the surface of the outwash plain of Scott Glacier is being lowered to compensate for 1964 earthquake uplift. USGS photograph no. 87–V2–083 by Robert M. Krimmel, U.S. Geological Survey. A larger version of this figure is available online.* **B**, *17 August 2000 north-looking oblique aerial photograph shows the terminus of the Scott Glacier and its adjacent outwash plain. The looping medial moraines that have developed since 1987 are a result of variations in flow between Scott Glacier and its unnamed tributary on the left. Elevated medial moraines, projecting beyond the leading edge of the terminus, are indicators of the continued retreat and thinning of Scott Glacier. Photograph by Bruce F. Molnia, U.S. Geological Survey.*

15 medial moraines, all of which stood above the bare ice surface. The height of the marginal zone had more than doubled since the 1960s.

A comparison of 1950s map data of the glacier with data obtained during annual airborne profiling surveys in the middle 1990s indicated that Scott Glacier thinned by 0.672 m a^{-1} and had a 0.0112-km^3 a^{-1} volume decrease (K. A. Echelmeyer, W. D. Harrison, V.B. Valentine, and S.I. Zirnheld, University of Alaska Fairbanks, written commun., March 2001).

Glaciers Draining into the Tasnuna River

South and east of Marshall Pass, six north-flowing valley glaciers drain into the Tasnuna River (from west to east): Marshall Glacier, Tasnuna Glacier, two unnamed glaciers, Woodworth Glacier, and Schwan Glacier. All rapidly retreated during the second half of the 20th century. Several were documented to be retreating at even earlier dates.

Woodworth Glacier

When Schrader (1900) photographed Woodworth Glacier (length 23 km, area 185 km^2) (Field, 1975b, p. 467) in 1898, its terminus was located in the Tasnuna River. By August 1938, when it was photographed again by Bradford Washburn (fig. 233), the glacier had retreated 800 to 900 m. By 1950, the

Figure 233.—*August 1938 near-vertical aerial photograph of an esker, several recessional moraines, and fluted topography on the outwash deposits in front of the retreating terminus of the Woodworth Glacier. Photograph by Bradford Washburn, Museum of Science (Boston), negative no. 1825.*

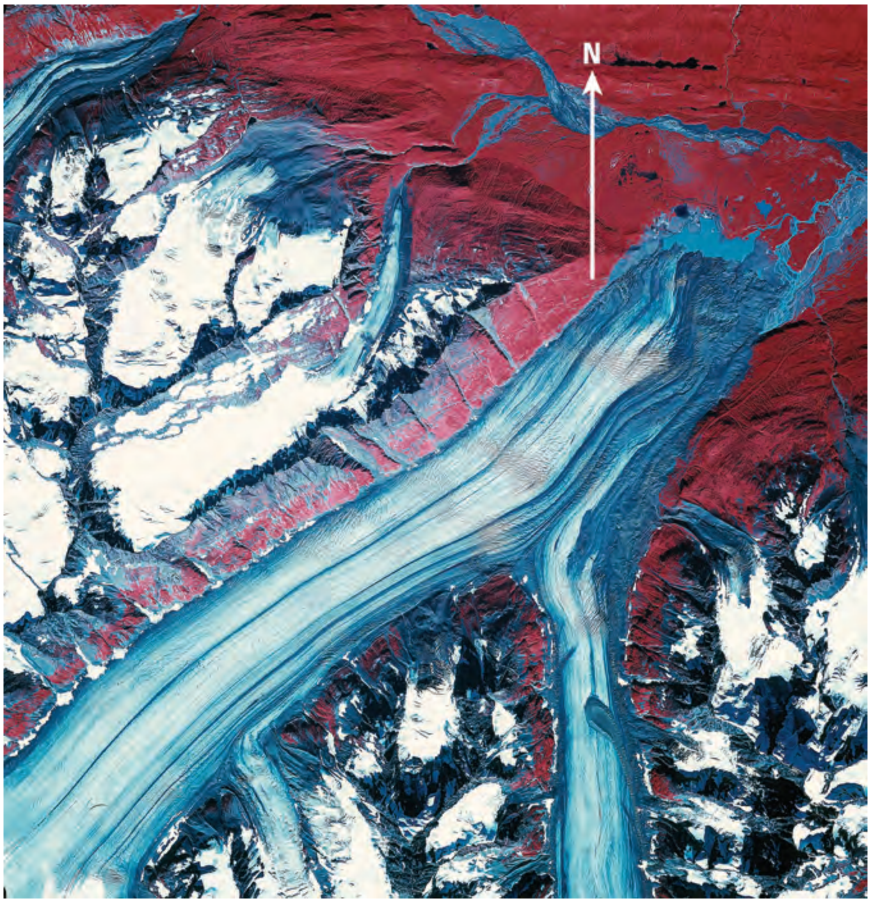


__Figure 234.__—25 August 1978 AHAP false-color infrared vertical aerial photograph of the lower reaches of Woodworth Glacier, including the glacier's retreating terminus and the developing ice-marginal lake along its western terminus. Also shown are the retreating and thinning Tasnuna Glacier (upper left) and Tasnuna River (upper right). The area photographed by Bradford Washburn (fig. 233) is located north of the ice-marginal lake and is covered by vegetation. Many of the north- and west-facing tributaries on the eastern side of the Tasnuna Glacier no longer make contact with the trunk glacier. AHAP photograph no. L110F7219 from the GeoData Center, Geophysical Institute, University of Alaska, Fairbanks, Alaska.

date of the photography used to prepare the USGS Valdez A-3 1:63,360-scale topographic quadrangle map (appendix B), the glacier had retreated another 500 m; Field (1975b) reported that there was little change from this position through 1964. By 25 August 1978, approximately 300 m of additional retreat occurred, and an ice-marginal lake had begun to form (fig. 234). When the author observed the glacier from the air on 3 September 2002, it had retreated approximately an additional 800 m.

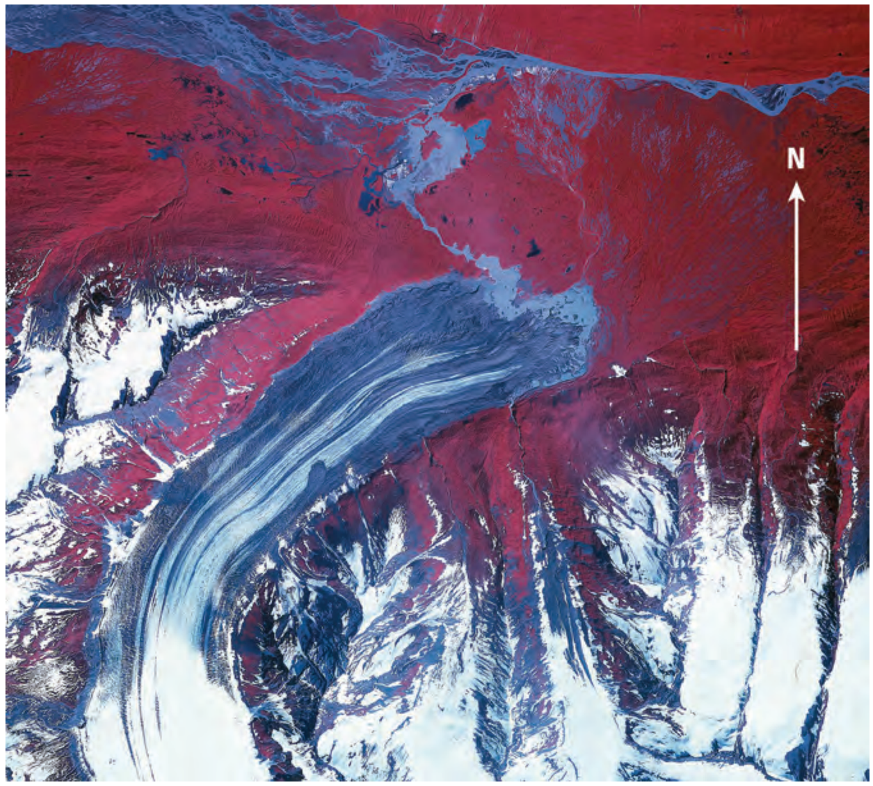
Schwan Glacier

Schwan Glacier has a length of 23 km and an area of 131 km^2 (Field, 1975b, p. 467). Its 1950 position was approximately 2.5 km behind a 19th century end moraine that marks its most recent maximum position, probably achieved during the "Little Ice Age." Through the late 1930s, retreat totaled about 2 km, but a small moraine visible in Bradford Washburn's 1937 and 1938 aerial photographs suggests that the glacier margin was stable or perhaps even slightly advancing (Field, 1975b). By 1941, the terminus had retreated about 50 m. When the glacier was next photographed in 1950, approximately 350 m of additional retreat had occurred. Through 1964, the glacier retreated an additional 300 m. By 9 July 1978, approximately 250 m of additional retreat had occurred, and an ice-marginal lake had begun to form (fig. 235). When the author observed the glacier from the air on 3 September 2002, it had retreated an estimated additional 600 m, and the ice-marginal lake fronted the entire terminus.

Small Glaciers Between Schwan and Heney Glaciers

East of Schwan Glacier and just west of the confluence of the Tasnuna and Copper Rivers, about six small, unnamed, north-facing retreating glaciers occur in steep U-shaped valleys separated by distinctive arête ridges. Most have retreated so much that the lower half of their valleys are ice free.

Figure 235.*—9 July 1978 AHAP false-color infrared vertical aerial photograph of the lower reaches of the Schwan Glacier, including the glacier's retreating terminus and the developing ice-marginal lake. Also shown are several unnamed retreating and thinning glaciers to its east and west. As with the adjacent Woodworth Glacier, many of its tributaries no longer make contact with the Schwan Glacier. Tasnuna River can be seen flowing across the upper part of the photograph. AHAP photograph no. L110F6220 from the GeoData Center, Geophysical Institute, University of Alaska, Fairbanks, Alaska.*



Heney Glacier

Heney Glacier has a length of 19 km and an area of 72 km^2 (Field, 1975b, p. 467). In 1910, Lawrence Martin visited the glacier and observed that a large part of the terminus was covered by a thick forest and fronted by a terminal moraine. He determined "that the last expansion of the bulb of Heney Glacier was about a century ago and that for over 76 years there has been no period of activity capable of breaking up the outer portion of the bulb" (Tarr and Martin, 1914, p. 449). He also observed that the central part of the terminus showed no recession from this moraine but that the flanks showed 800 to 1,400 m of retreat (Tarr and Martin, 1914).

In 1910, the Copper River and Northwest Railroad was constructed on the surface of this moraine (Tarr and Martin, 1914, pl. CLXXVIII, ff p. 464). Before Martin's next visit in 1911, a pulse or minor surge event shattered much of the stagnant ice in the terminus region, rendering it impassable. However, it did not cause terminus advance and did not disrupt the railroad track. Field (1975b) reported that little change occurred through 1937. Between 1937 and 1950, about 200 m of retreat occurred; approximately 300 to 350 m of additional retreat occurred through 1964. When the author observed the glacier from the air in 2002, it was estimated to have retreated more than 1.4 km, and an ice-marginal lake fronted the entire terminus.

Allen Glacier

Allen Glacier, located about 11 km north of Childs Glacier, has a length of 31 km and an area of 230 km2 (Field, 1975b, p. 468). Unlike Childs and Miles Glaciers, which presently calve icebergs directly into the Copper River, the retreating terminus of Allen Glacier is as much as 3 km from the river and is fringed by an estimated 15-km-long arcuate end moraine that separates the river from a broad area of vegetation- and moraine-covered stagnant ice. The first map showing the position of the terminus of Allen Glacier was made by D.C. Witherspoon for the USGS in 1900. It was revised in 1906 and published by Tarr and Martin in 1914. When Lawrence Martin visited and mapped the glacier in 1910 (Tarr and Martin, 1914, pl. CLXXIV, ff p. 448), he found alders

(*Alnus* sp.) as old as 67 years growing on this moraine. About 9 km of the bed of the Copper River and Northwest Railroad was constructed on stagnant ice along the river's edge, which Tarr and Martin described as "the end of a living glacier" (Tarr and Martin, 1914, p. 445). During construction, Tarr and Martin observed that "The ballast beneath the ties and rails of the railway actually rests upon the ice, not upon an abandoned moraine as at Heney Glacier" (Tarr and Martin, 1914, p. 445). In 1912, the northeastern part of the terminus began to advance; by 1913, as much as 800 m of advance had occurred (Wentworth and Ray, 1936). In 1931, Wentworth and Ray (1936, p. 923) visited the glacier and noted that "A more marked stagnation had set in, and the area of active ice movement had been thrown back, perhaps several thousand feet, from the northeast margin of 1910." This advance had only a minimal impact on the railroad; for more than 20 years, trains crossed the glacier margin daily in both directions.

The glacier was photographed by Bradford Washburn in 1938 and by the USN in 1957. In the 47 years between Martin's 1910 map and the 1957 photograph, the northern part of the terminus retreated between 400 m and 1 km, while the middle and southern part of the terminus retreated more than 2 km. Additionally, the glacier had thinned more than 100 m over most of the lobate terminus and at the head of the piedmont lobe, where the main trunk flows from its valley (Field, 1975b). Post (1967a) reported that, beginning in the early 1960s, the glacier again began to advance. Between 1963 and 1964, he noted that the glacier advanced 300 m. Similarly, between 1964 and 1965, the glacier advanced an additional 300 m. This advance reactivated much of the distal part of the stagnant ice area.

The 1964 earthquake caused three large rock avalanches that fell onto the surface of Allen Glacier. The largest, termed "Allen I," had an area of about 2 km^2 and a length of about 3 km. Before 29 July 1965, another rockslide—larger and longer than any that took place during the earthquake—occurred. This slide had an area of about 7.5 km^2 and a length of about 7.5 km and consisted of material that was probably loosened by the 1964 earthquake (Post, 1967a).

In the 36 years between Post's 1966 observations, 17 August 2000 (fig. 236), and 17 October 2002 (the author's most recent visit), the glacier is estimated to have retreated more than 3 km and thinned at least 125 m. Even the upper reaches of the glacier showed significant thinning.

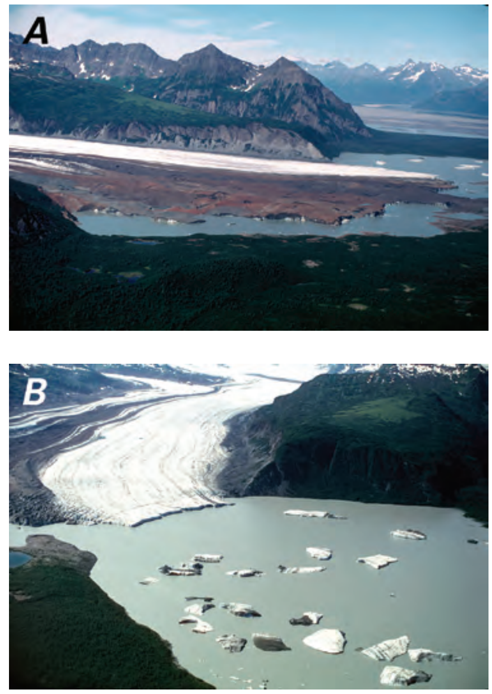


Figure 236.—*Two 17 August 2000 oblique aerial photographs of Allen Glacier show several lines of evidence confirming the rapid thinning and retreat of the glacier.* **A**, *Northeast-looking view across part of the terminus of Allen Glacier shows the debris-covered southern part of the terminus and the bare ice northern part. The large trimline documents more than 100 m of recent thinning.* **B**, *View of the northern part of the ice-marginal lake fronting Allen Glacier. The ongoing thinning of the glacier appears to have resulted in flotation of part of the terminus, accompanied by the production of large icebergs. Photographs by Bruce F. Molnia, U.S. Geological Survey. Larger versions of these figures are available online.*

Grinnell Glacier

According to Tarr and Martin (1914, p. 437), the "last great expansion" of Grinnell Glacier occurred before 1892, with little change occurring between 1891 and 1900. Minor advances occurred in 1907 and between 1909 and 1911. In 1910, the terminus of Grinnell Glacier was located about 500 m from the Copper River and about 600 m from the northern margin of Miles Glacier's stagnant northern lobe, which is located on the eastern side of the Copper River. Railroad track that had been laid over 400 m of stagnant ice was not affected by the early 20th century advances.

The next visitors to the glacier were Wentworth and Ray in 1931. They reported that, "Between 1911 and 1931, there was a general retreat of the active portion of the glacier and continued reduction and melting of the stagnant ice" (Wentworth and Ray, 1936, p. 921). They also noted that trees up to 25 cm in diameter were growing on the ice-cored moraine.

Field (1975b) reported that several hundred meters of retreat occurred in the 28-year interval between photographs by Bradford Washburn in 1938 and an AGS field party's visit in 1966. In the 36 years between Austin Post's 1966 observations and the author's 17 October 2002 aerial observations, Grinnell Glacier is estimated to have retreated 1 km and thinned at least 100 m. The southeastern tributary also separated from the main trunk. Fresh bedrock

Figure 237.—17 August 2000 northwest-looking oblique aerial photograph of much of Grinnell Glacier. Several areas of freshly exposed bedrock around the perimeter of each distributary lobe of the terminus of Grinnell Glacier and a fresh trimline along the margin of the glacier are evidence of the rapid thinning and retreat of the glacier. Photograph by Bruce F. Molnia, U.S. Geological Survey. A larger version of this figure is available online.

around the terminus, seen in a 17 August 2000 oblique aerial photograph taken by the author, suggests a recent rapid retreat (fig. 237).

Childs Glacier

Childs Glacier, located immediately west of Million Dollar Bridge, has a length of 19 km and an area of 100 km^2 (Field, 1975b, p. 468), and makes up the western bank of the Copper River for a distance of about 6 km. U.S. military expeditions led by W.R. Abercrombie in 1884 and H.T. Allen in 1885 provided the first descriptions and photographs (Allen, 1887) of Childs Glacier (Abercrombie, 1900; Allen, 1900).

Most of the early information concerning Childs Glacier, however, resulted from Tarr and Martin's expeditions between 1909 and 1911 (Tarr and Martin, 1914). Although the glacier's "Little Ice Age" chronology is not well dated, several moraines as high as 100 m and located as much as 1.5 km beyond the present margin may represent late 18th century or early 19th century maxima.

Aside from a small advance between 1905 and 1906, the glacier's retreat was slow between 1884 and 1909. A surge that began in the spring of 1909 and lasted through the summer of 1910 was carefully documented by railway engineers, who were concerned about the potential impact of the advancing glacier terminus on the construction of the Million Dollar Bridge across the Copper River. The surge produced a terminus advance of about 600 m that reduced the distance between the ice margin and the bridge by more than 50 percent. A conspicuous moraine formed by this advance is visible approximately 450 m from the bridge. Tarr and Martin (1914) cited maximum flow rates of at least 40 m d^{-1} between 29 July and 6 August 1910 and 9 to12 m d^{-1} between May and October. Much ice was lost to calving and river erosion.

Wentworth and Ray (1936) visited the glacier in 1931 and reported that it had retreated between 150 m and 200 m. Field (1975b) summarized the results of more than a dozen visits and photographic observations between 1938 and 1971. In the mid-1930s, an advance of the glacier built "a massive moraine...on the land front at the northern end of the terminus" (Field, 1975b, p. 343). By 1950, however, a retreat of about 60 m occurred; between 1950 and 1959, an additional 200 m of retreat occurred. These retreats were accompanied by a reduction in flow rate and a reduction in the height of the ice margin.

By 1961, Childs Glacier was advancing, and its ice face had thickened to approximately 65 m. Between 1959 and 1961, the glacier advanced as much as 100 m. In the 7 years between 1961 and 1968, the glacier advanced approximately 150 m further. Little change was noted in 1971, although the terminus was within approximately 10 m of the mid-1930s moraine.

Since then, the glacier has thinned and retreated. This ongoing trend was noted by the author in multiple observations made between 1981 and 2004. Aerial photographs taken on 22 August 1979 (see oblique aerial photograph no. 79V2–179 by Austin Post, U.S. Geological Survey), 13 August 1994 (fig. 238*A*), 17 August 2000 (fig. 238*B*), and 17 October 2002 show a large arcuate calving embayment forming along the Copper River. The maximum amount of retreat that occurred between 1971 and 2004 was approximately 300 m. Field (1965) provided a map showing terminus positions between 1912 and 1961 (AGS Glacier Studies Map No. 64–3–G1).

Figure 238.— **A**, *13 August 1994 vertical aerial photograph showing the terminus of Childs Glacier. Between 22 August 1979 (USGS photograph no. 79V2–179 by Austin Post, U.S. Geological Survey) and 13 August 1994, the calving embayment on the west side of the Copper River deepened by approximately 150 m. Note the elongated mass of landslide debris that is moving down the center of the glacier. USGS photograph no. 94–V3–179 by Robert M. Krimmel, U.S. Geological Survey.* **B**, *17 August 2000 northwest-looking oblique aerial photographic mosaic showing the full terminus of Childs Glacier. In addition to a deepening of the calving embayment, the moraines on the southern half of the terminus have become more convoluted and offset. Photograph by Bruce F. Molnia, U.S. Geological Survey. A larger version of this figure is available online.*

Goodwin Glacier

When Tarr and Martin (1914) observed the terminus of Goodwin Glacier, it was located immediately adjacent to a forest. This observation led them to conclude that the glacier had not been more advanced from its current position for a significant period of time. Field (1975b) noted that there were no reports of any advance of Goodwin Glacier during the 20th century. Additionally, Field (1975b, p. 344) compared aerial photographs made in 1938, 1950, and 1961 and stated that “the lower one kilometer of the glacier is moraine-covered which effectively masks most of the ice edge.” His analysis showed that “no appreciable change [had occurred] in recent decades, except extension of the alders on the ice-free and ice-cored parts of the moraines. Slow recession marked by progressive stagnation of the terminus is thus indicated.” Aerial observations by the author between 1990 and 2001 documented continued downwasting of the glacier, separation of the debris-covered stagnant ice part of the terminus from older vegetation-covered ice-cored moraine, and the formation of several ice-marginal lakes.

Cleave Creek Glacier–Upper Copper River–Stephens Glacier–Tonsina Glacier Northwestern Subdivision

Glaciers in this subdivision that have lengths greater than 8 km and areas determined by Field (1975b, p. 465–467), include Cleave Creek (11 km, 31 km^2), two unnamed glaciers (9 km, 20 km^2; 10 km, 22 km^2), Tsina (10 km, 37 km^2), Tonsina (17 km, 47 km^2), two more unnamed glaciers (11 km, 19 km^2; 8 km, 16 km^2), Klutina (11 km, 45 km^2), and Stephens (16 km, 56 km^2). Worthington Glacier, a popular tourist destination, is also located in this subdivision.

Cleave Creek Glacier

Cleave Creek Glacier is the largest and only named glacier in an upland area located between the Copper, Tasnuna, Tiekel, and Tsina Rivers. All of the glaciers in the complex are actively retreating, and some have completely disappeared since being mapped in the 1950s. At the southeastern end of the area, about a half-dozen small, north-flowing, unnamed, steeply entrenched former tributaries to a previously expanded Cleave Creek Glacier are rapidly thinning and retreating as shown on an 8 August 1996 oblique aerial photograph by the author (fig. 239). The terminus of Cleave Creek Glacier has narrowed to approximately one-third of its 1950s width and exposed a surface of grooved sediment and bedrock.

Worthington Glacier

Worthington Glacier is a small mountain glacier about 6 km long. Its bifurcated terminus, located about 2 km from the Richardson Highway, shows conspicuous evidence of recent retreat. A large lateral and terminal moraine complex surrounds the glacier. A large ice-marginal lake dammed by this moraine is located between the moraine and the southern lobe of the glacier. When the author visited on 3 September 2002, the retreating ice margin no longer reached the shore of the lake. During the later part of the 20th century, the distal end of the northern terminus separated into three separate distributaries. The site of a State of Alaska glacier interpretive center, the glacier has thinned so much that visitors need to walk downhill from a 1960s parking lot to reach the terminus.

A surveying party under the leadership of Austin Post mapped Worthington Glacier at 1:10,000 scale with a 5-m contour interval during the IGY in

Figure 239.*—8 August 1996 south-looking oblique aerial photograph of several of the former tributaries of the retreating Cleave Creek Glacier. All are rapidly thinning and retreating. The lower reaches of the middle glacier of the three unnamed, steeply entrenched former tributaries have lost contact with the upper accumulation area. Currently, ice and snow reach the lower area through avalanching. Photograph by Bruce F. Molnia, U.S. Geological Survey.*

1957. Part of the AGS's Nine Glacier Maps Project, the accompanying text (American Geographical Society, 1960, p.19, 21) described Worthington as "descending eastward from a broad névé field between peaks rising to more than 2,010 meters (6,600 feet) to a bifurcated terminus, one lobe of which extends a short distance into the broad valley of Ptarmigan Creek at an elevation of 660 meters (2,165 feet).... This glacier was receding in 1957, continuing a recession which evidently had been in progress for many years, as evidenced by a large bare area, terminal and recessional moraines. Moderately crevassed and almost completely free of ablation moraine, this glacier is rather typical of the heavy accumulation, rapid velocity, and high ablation glaciers typical of the region. The greatest snow depth noted on the glacier in the first half of July 1957, was less than 3 meters (10 feet)...."

Viereck (1967) visited the glacier in 1957 and identified a major moraine located about 600 m beyond the 1957 ice margin and a set of inner moraines about 200 m from the 1957 terminus. He concluded that the glacier occupied the outer moraine sometime between 1837 and 1857 and the inner moraines sometime beginning in the 1930s.

The glacier was revisited in 1961, 1964, and 1966 by AGS parties that included William Field and was photographed by Austin Post for the USGS in 1968 and 1971. Analysis of these data by Field (1975b) show that the glacier retreated approximately 400 m between the middle 19th century and the middle 1930s (~4–5 m a^{-1}); approximately 150 m between the middle 1930s and 1957 (~7 m a^{-1}); approximately 50 m between 1957 and 1961 (~12 m a^{-1}); at least 35 m between 1961 and 1964 (~12 m a^{-1}); and at least 35 m between 1964 and 1966 (~17 m a^{-1}). Retreat rates were similar through 1971.

On the basis of a comparison of the USGS map of the glacier based on 1949 and 1950 aerial photography and data obtained during an airborne profiling survey conducted on 31 May 1994, Echelmeyer and others (1996) determined that the terminus of the north lobe of Worthington Glacier retreated about 250 m, and the terminus of the south lobe had retreated about 380 m, an average of about 310 m during the 44 to 45 years between data sets; the average annual retreat rate was about 7 m a^{-1}. The area of Worthington Glacier decreased from 8.8 to 8.6 km^2 in 1994, a decrease of around 2.3 percent; however, its thickness increased an average of 11.1 m, and its volume increased 0.97×10^8 m^3. The details of the thickening are actually more complicated than these values suggest. According to Sapiano and others (1998), during the 37 years from 1957 to 1994, the glacier thinned by more than 40 m at elevations of less than 800 m, thinned at all elevations below 1,200 m, and thickened at most locations above 1,200 m.

Between the middle 1990s and 1999, Worthington Glacier thinned by 0.948 m a^{-1} annually, and its volume decreased by 0.00856 km^3 a^{-1} (K.A. Echelmeyer, W.D. Harrison, V.B. Valentine, and S.I. Zirnheld, University of Alaska Fairbanks, written commun., March 2001). In the 1990s, Harper and others (1998) conducted studies of crevasses patterns and longitudinal strain rate. Between 1975 and 2001, the author estimates that the individual termini have retreated about 300 m, and the lower part of the glacier has thinned at least 25 m.

Tonsina Glacier

Little is known about Tonsina Glacier, which has a length of 17 km and an area of 47 km^2 (Field, 1975b, p. 466). A comparison of the position of the glacier's terminus as shown on the USGS Valdez 1:250,000-scale topographic quadrangle map (1960) (appendix A), which is based on early 1950s surveys; a 25 August 1978 AHAP false-color infrared vertical aerial photograph (fig. 240), and observations from the air by the author on 31 August 2000 and 3 September 2002 show that the glacier has retreated about 2 km, and thinned by more than 100 m in the 50 years since the 1950s surveys. All of the other glaciers in the area have also retreated significantly, including Tsina Glacier

Figure 240.—25 August 1978 AHAP false-color infrared vertical aerial photograph of the terminus of Tonsina Glacier and a number of adjacent small, unnamed cirque glaciers. All glaciers show evidence of ongoing thinning and retreat. AHAP photograph no. L107F7055 is from the GeoData Center, Geophysical Institute, University of Alaska, Fairbanks, Alaska.

approximately 5 km to the south, many small unnamed cirque glaciers, and an unnamed glacier located about 2.5 km to the north.

Two unnamed glaciers, each having a length of about 4 km and an area of about 7 km^2, are located between Tonsina and Klutina Glaciers. As are so many other glaciers in the north-central part of the Chugach Mountains, these two are actively thinning and retreating. Both have retreated as much as 2.5 km during the approximately 50-year interval between the time when they were mapped in the early 1950s and when they were observed from the air and photographed by the author in August 1993, on 31 August 2000, and on 2 September 2002.

Klutina and Stephens Glaciers

Field (1975b, p. 466) reported that Klutina Glacier (fig. 241) has a length of 11 km and an area of 45 km^2; Stephens Glacier has a length of 16 km and an area of 56 km^2. Both glaciers were photographed by Bradford Washburn in 1937 and Austin Post in 1964. Comparison shows that both glaciers retreated between 300 and 500 m in the 27-year interval between the two sets of photographs. Individual comparisons of the positions of each glacier's terminus as shown on the USGS Valdez 1:250,000-scale topographic quadrangle map (1960) (appendix A), which is based on early 1950s surveys, and observations made by the author from the air on 30 August 2000 (fig. 241) and 3 September 2002 show that each glacier has thinned significantly and retreated between 2 and 5.5 km in the ensuing 50 years, although little evidence of change is visible in the accumulation areas. When the author observed the glaciers from the air on 3 September 2002, both were actively retreating and thinning. Each had elevated lateral and medial moraines, and each had separated from former tributaries. The terminus of Klutina Glacier was about 4 km upvalley from a conspicuous end moraine, perhaps marking its "Little Ice Age" maximum.

Figure 241.—30 August 2000 oblique aerial photograph of Klutina Glacier, looking north; the former western tributary is now more than 800 m from Klutina Glacier and nearly 200 m above it. Photograph by Bruce F. Molnia, U.S. Geological Survey. A larger version of this figure is available online.

Glaciers draining into the Uranatina River and Haley Creek are also retreating and thinning. Many ice-free cirques suggest that a significant number of glaciers completely disappeared during the 20th century. When the author observed the glaciers from the air on 31 August 2000, less than 30 percent of the cirques examined still had glacier ice.

Prince William Sound Segment—Heney Range to the East Side of Valdez Arm Subdivision

Glaciers in this subdivision that have lengths of 8 km or more and areas determined by Field (1975b, p. 469–470) include two unnamed glaciers (9 km, 26 km^2; 9 km, 22 km^2), Cordova (13 km, 45 km^2), an unnamed glacier (9 km, 13 km^2), Bench (8 km, 8 km^2), Deserted (17 km, 36 km^2), Wortmanns (14 km, 55 km^2), Keystone (9 km, 20 km^2), an unnamed glacier (15 km, 30 km^2), and Valdez (34 km, 158 km^2).

Shephard Glacier

Shephard Glacier is a small unstudied glacier located at the head of Power Creek northeast of Cordova. Although it was snow covered to within about 200 m of its terminus when the author observed it from the air on 8 August 2000, the terminus was fronted by a barren zone about 1 km in width. The bedrock became deglacierized following the mapping of the glacier in the early 1950s.

Glaciers of the Rude River Drainage

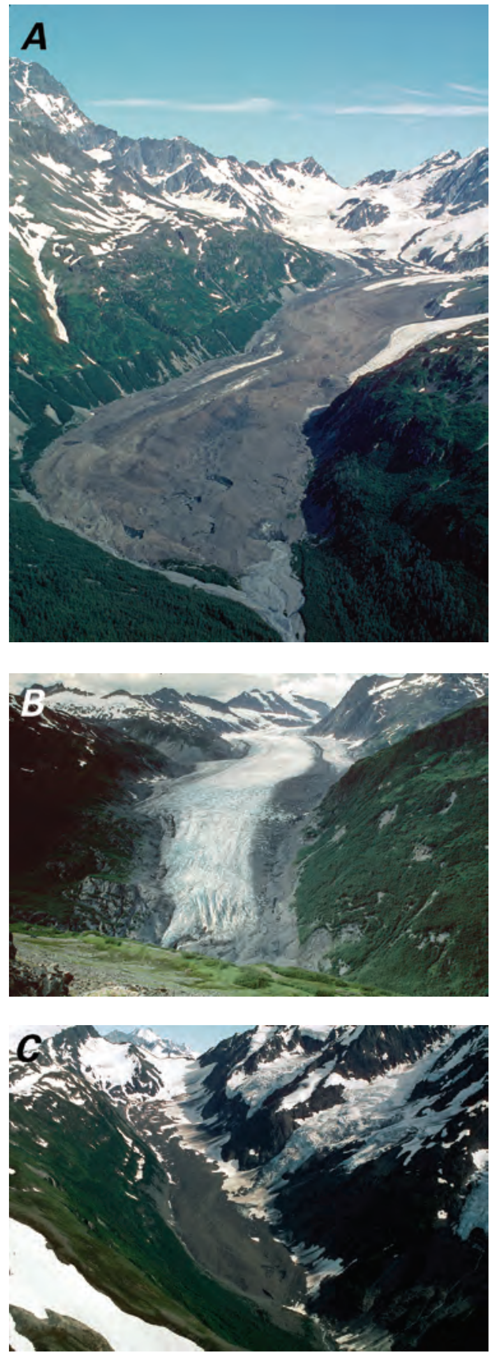
South of 2,350-m-high Cordova Peak, several unnamed debris-covered valley glaciers show signs of thinning, retreat, and stagnation. The largest, an unnamed glacier located at the head of the eastern fork of the Rude River (fig. 242*A*), has a length estimated by the author of approximately 8 km. More than half its length is mantled by a debris cover that is estimated to be as much as 1 m in thickness. Another unnamed glacier (fig. 242*B*) is located approximately 3 km to the northwest. Unlike its neighbor, its terminus is debris free. It is surrounded by an apron of abandoned moraine and freshly exposed bedrock, the result of late 20th century thinning and retreat. An unnamed west-flowing glacier located at the head of a northeastern fork of the Rude River has more than 80 percent of its length covered by debris. What is unusual about this glacier (fig. 242*C*) is that three of its four southern tributaries are in rapid retreat and terminate hundreds of meters above the surface of the glacier; the fourth descends all the way to the valley floor and connects with the thinning main trunk.

Cordova Glacier

Field (1975b) reported that the terminus of Cordova Glacier, which has a length of 13 km and an area of 45 km^2 (Field, 1975b, p. 469) and is located at the head of the western fork of the Rude River, dammed a lake that was 1.8 km long and an average of 600 m wide in 1950. The distal end of the lake was adjacent to the terminus of an unnamed glacier that has an approximate length of 8 km. The unnamed lake has been described by both Stone (1963b) and Post and Mayo (1971). Field (1975b, p. 353) speculated — correctly — that, "Since all the glaciers in this part of the central Chugach appear to be shrinking, it is likely that some year the ice-dam will not form and this lake will disappear." Although the date of the last formation of the lake is unknown, much of its former bed was covered by vegetation when the author observed the former area of the lake from the air on 8 August 2000. The terminus of Cordova Glacier had thinned and retreated so much that it was entrenched behind a large terminal and several large recessional moraines at least 1.5 km from the eastern wall of the former lake.

Glaciers South of Lowe River

To the north and northwest of Cordova Glacier, more than a dozen glaciers flow northward and drain into the Lowe River. An unnamed glacier located at the head of a northeast-flowing creek west of Browns Creek and another unnamed glacier located in the Browns Creek drainage both show evidence of ongoing retreat. Like so many other glaciers in this subdivision, they have large areas of freshly deglacierized bedrock exposed around their margins.



◀ ***Figure 242.****—Three 8 August 2000 oblique aerial photographs of unnamed glaciers within the Rude River drainage.* **A***, East-looking view of the debris-covered lower portion of an unnamed glacier at the head of the east fork of the Rude River. Note the vegetated, elevated lateral moraine along the north side of the glacier. Thermokarst pits and many other stagnation features are visible.* **B***, North-looking view of an unnamed glacier on the southwest flank of Cordova Peak, with its debris-free retreating terminus. Note the apron of abandoned moraine and freshly exposed bedrock.* **C***, East-looking view of an unnamed glacier. Of its four southern tributaries, three are in rapid retreat and terminate hundreds of meters above the surface of the glacier, whereas the fourth descends all the way to the valley floor and connects with the thinning main trunk. Photographs by Bruce F. Molnia, U.S. Geological Survey. Larger versions of these figures are available online.*

Four of the larger glaciers to the east of Browns Creek—Wortmans, Bench, Heiden, and Deserted Glaciers—were photographed by Bradford Washburn in 1938 and again by Post in 1960. Field (1975b, p.353) characterized all four glaciers as having "terminal and marginal barren zones in the 1950s and 1960 which indicate recent net recession."

Wortmanns Glacier

The largest of the four glaciers is Wortmanns Glacier (fig. 243), a retreating wishbone-shaped glacier that is 14 km long and has an area of 55 km^2 (Field, 1975b, p. 469). Field (1975b) reported that, in 1950, its terminus was about 1.5 km behind an outer moraine of unknown age. When the terminus was photographed by the AHAP Program on 25 August 1978 (fig. 243), it had retreated another 250 m. When the author observed it from the air, it had retreated approximately an additional 300 m (August 1996), 350 m (8 August 2000), and 400 m (3 September 2002). A comparison of 1950 map data of the glacier and data obtained during an airborne profiling survey in the middle 1990s, showed that Wortmanns Glacier thinned by an average of 0.5228 m a^{-1} and decreased in volume by 0.0305 km^3 a^{-1} (K.A. Echelmeyer, W.D. Harrison, V.B. Valentine, and S.I. Zirnheld, University of Alaska Fairbanks, written commun., March 2001).

Bench Glacier

Bench Glacier, which has a length of 8 km and an area of 8 km^2 (Field, 1975b, p. 469), is located immediately to the east of Wortmanns Glacier. It was photographed by the AHAP Program on 25 August 1978 (fig. 243) and observed from the air by the author in August 1996, on 8 August 2000, and on 3 September 2002. Characterized by Field (1975b) as retreating, its terminus had retreated an additional 50 m by 1978. When the author observed it from the air and photographed it in 2000 and 2002, the terminus had retreated at least another 275 m. Comparison of 1950 map data of the glacier with data obtained during an airborne profiling survey in the middle 1990s showed that Bench Glacier thinned by an average of 1.168 m a^{-1} and decreased in volume by 0.0138 km^3 a^{-1} (K.A. Echelmeyer, W.D. Harrison, V.B. Valentine, and S.I. Zirnheld, University

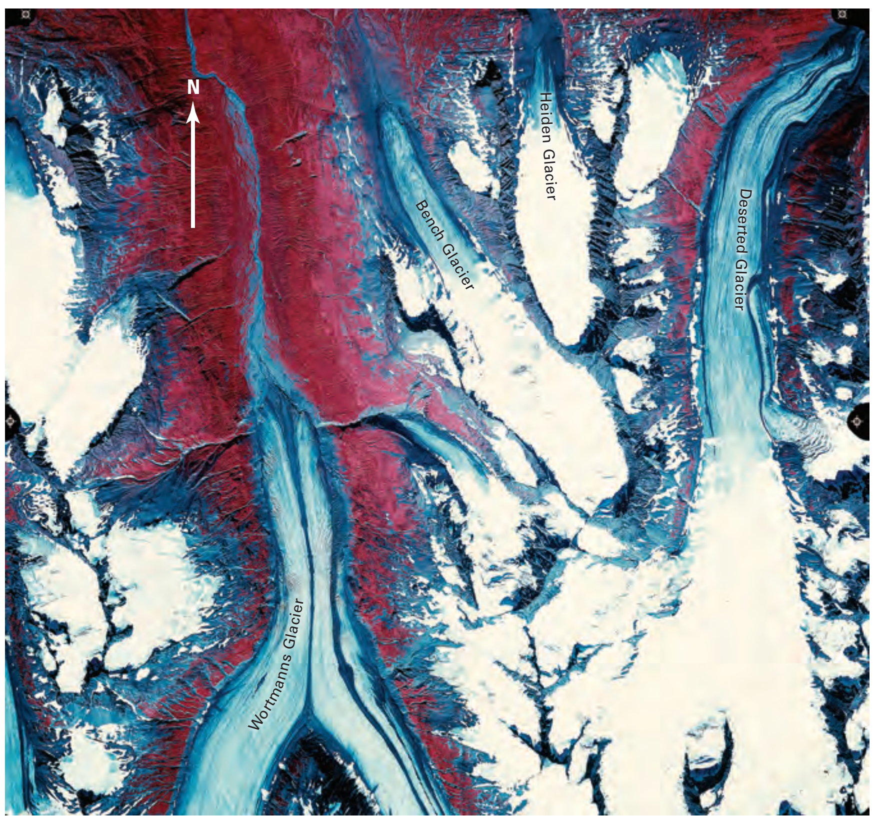


Figure 243.*—25 August 1978 AHAP false-color infrared vertical aerial photograph of the thinning and slowly retreating terminus of Wortmanns Glacier, Bench Glacier, Heiden Glacier, all but the terminus of Deserted Glacier, and several small retreating cirque and valley glaciers. AHAP photograph no. L110F7288 from the GeoData Center, Geophysical Institute, University of Alaska, Fairbanks, Alaska. Every glacier shown presents evidence of thinning and retreat.*

of Alaska Fairbanks, written commun., March 2001). Glacier hydrology studies of Bench Glacier, including jökulhlaups associated with the glacier, have been carried out by Howard and others (1996) and Anderson (2003).

Heiden Glacier

Heiden Glacier, the smallest of the four glaciers south of Lowe River, was photographed by the AHAP Program on 25 August 1978 (fig. 243) and was observed from the air by the author in August 1996, on 8 August 2000, and on 3 September 2002. Between 1978 and September 2002, the terminus retreated at least 400 m.

Deserted Glacier

The longest glacier of the four is Deserted Glacier, which is 17 km long and 36 km^2 in area (Field, 1975b, p. 469). A comparison of its position on the USGS Cordova D–5, 1:63,360-scale topographic quadrangle map (1953) (appendix B), which is based on 1950 photography, with observations made by the author from the air, including oblique aerial photographs of the glacier taken on 8 August 2000 and on 3 September 2002, indicates that the terminus of the glacier retreated by at least 1.6 km. Similarly, a comparison of 1950 map data of the glacier with data obtained during an airborne profiling survey in the middle 1990s shows that Deserted Glacier thinned by an average of 0.816 m a^{-1} and had a volume decrease of 0.03 km^3 a^{-1} (K.A. Echelmeyer, W.D. Harrison, V.B. Valentine, and S.I. Zirnheld, University of Alaska Fairbanks, written commun., March 2001).

Glaciers North of Lowe River and West of Keystone Canyon

Three named glaciers—Corbin, Rubin, and Camicia Glaciers—and about 10 unnamed glaciers are all west-flowing glaciers and ultimately drain into Valdez Arm. All of these glaciers and the adjacent Valdez and Hogback Glaciers have been remapped by the USGS in a revision of the USGS 1:63,360-scale Valdez A–6 topographic quadrangle map (1953) (appendix B). Although it was printed in 2002, the revised map has a 1993 date because it was based on 1993 imagery. Since the 1953 edition of the map was compiled, every one of these glaciers has undergone conspicuous thinning and retreating. Aerial observations made on 8 August 2000 and 3 September 2002 show that every glacier was characterized by trimlines well above the ice surface and that there was an apron of vegetation-free bedrock adjacent to the terminus of each glacier.

Valdez Glacier

Valdez Glacier (fig. 244), which is 34 km long and 158 km^2 in area (Field, 1975b, p. 470), is located 6 km northeast of the pre-l964 Alaska earthquake location of the town of Valdez. Before Valdez was relocated about 4 km to the west, the glacier presented a significant flood threat to the town's inhabitants (Field, 1975b). At the end of the 19th century, when Schrader (1900) first described the glacier, it was thinning and retreating. This was further documented by a series of measurements made by L.S. Camicia, a local optician, between 1901 and 1905 and again in 1908. Reid (1909) and Tarr and Martin (1914) provide summaries of Camicia's measurements. Reid (1909, p. 670) stated that "Dr. L.S. Camicia has been keeping a record since 1901 of the position of the Valdez Glacier, Prince William Sound, Alaska. A stone monument was made on the moraine in front of glacier and the distance to the ice determined. He found the following variations, measurements having been made in June of each year: 1901–2, a retreat of 39 feet; 1902–1904, 165 feet; 1904–1905, 138 feet. The next observation was made in October 1908; as the monument had been destroyed, he estimated its position as well as he could, and found a retreat since the last observation of 244 feet, making a total retreat from 1901 to 1908 of 568 feet." Tarr and Martin (1914) also commented on Camicia's measurements.

These early observations, along with Moffit's 1905 photograph (fig. 244*A*), documented that, with the exception of a short-lived advance in 1906, the

Figure 244.— Two photographs showing changes at Valdez and Camicia Glaciers during the 20th century. **A***, 1905, northeast-looking photograph by F.H. Moffit across the terminus of Valdez Glacier toward Camicia Glacier. Bare bedrock on the west side of the glacier and a trimline on the east side document 19th century thinning of the glacier. Prospectors Peak can be seen above Camicia Glacier. USGS Photo Library photograph, Moffit 344.* **B***, 8 August 2000 oblique aerial photograph of the termini of Valdez and Camicia Glaciers. Valdez Glacier is fronted by an ice-marginal lake approximately 2 km in diameter. Camicia Glacier retreated approximately 5 km during the 20th century. Photograph by Bruce F. Molnia, U.S. Geological Survey. Larger versions of these figures are available online.*

glacier was in retreat at the beginning of the 20th century. Camicia's measurements continued until 1911, during which time the glacier retreated 179 m (average retreat of 18 m a^{-1}). According to Grant and Higgins (1913), the 1906 advance was as much as 76 to 91 m. Field (1932) visited the glacier in 1931 and determined that an additional 390 m of retreat had occurred in the 20 years since 1911 (average retreat of 19.5 m a^{-1}). Field (1937) again visited the glacier in 1935 and determined that an additional 45 m of retreat had occurred between 1931 and 1935 (average retreat of 11 m a^{-1}). Therefore, in the 34-year-period from 1901 to 1935, the terminus of Valdez Glacier retreated 614 m (average retreat of 18 m a^{-1}).

Field (1975b) reported that, in the 25 years between 1935 and 1961, the terminus retreated an additional 325 m (average retreat of 13 m a^{-1}). He stated that "Since there was a great quantity of stagnant ice in the terminal part of the glacier during the 1950s and 1960s, it was difficult to identify the actual terminus and the position of the lower end of the active ice" (Field, 1975b, p. 363). This active ice edge was 150 to 175 m further back from the terminus. By 1971, when Austin Post photographed the edge of the retreating active ice for the USGS, it was about 175 m from Field's 1964 position. When the author photographed the terminus from the air on 8 August 2000 (fig. 244*B*) and on 3 September 2002, it had retreated north of the valley of its former eastern tributary, Camicia Glacier. Valdez Glacier was fronted by an ice-marginal lake about 2 km in diameter and was calving, producing large tabular icebergs.

The history of Valdez Glacier is closely tied to the gold rush at the turn of the 20th century. Between 1898 and 1900, the glacier served as a major highway for prospectors traveling from the port of Valdez to the gold fields of the interior (Tarr and Martin, 1914, pl. XCV, ff p. 240). The Valdez Glacier route bypassed the rugged and dangerous route through Keystone Canyon and Thompson Pass (Schrader, 1900) and provided a relatively easy 29-km-long route to the head of Klutina Glacier and from there to Klutina Lake. About 1910, mining began in the bedrock exposed on both walls of the valley of Valdez Glacier. 20th century thinning of the glacier left these mining claims more than 100 m above the present ice surface. Hence, during the 20th century, Valdez Glacier retreated more than 2 km and thinned by more than 100 m.

A comparison of 1950s map data of the glacier with data obtained during a geodetic airborne laser altimeter profiling survey in the middle 1990s, shows that Valdez Glacier thinned by an average of 1.23 m a^{-1} and decreased in volume by 0.201 km^3 a^{-1} (K.A. Echelmeyer, W.D. Harrison, V.B. Valentine, and S.I. Zirnheld, University of Alaska Fairbanks, written commun., March 2001).

An unnamed glacier located north of Prospectors Peak (1975 length estimated by the author of about 12 km, area of about 42 km^2) and Camicia Glacier were both connected to Valdez Glacier in the early 20th century. By the beginning of the 21st century, Camicia Glacier had retreated about 4.2 km, and the unnamed glacier had retreated about 2.5 km. Both are incorrectly mapped on the 1993 revision of the USGS Valdez A–4 topographic quadrangle map (appendix B).

Camicia Glacier separated from Valdez Glacier soon after the beginning of the 20th century. By 1909 (Tarr and Martin, 1914), an ice-marginal lake dammed by the flank of Valdez Glacier separated the two glaciers. In 1931, when Wentworth and Roy (1936) observed the glaciers, the lake still was present. Field (1975b) reported that, by 1950, Valdez Glacier had narrowed to the point where no lake was evident. Field also reported that a comparison of two maps—the first surveyed in 1911 and 1912 and the second based on 1950 photography—showed that, between 1912 and 1950, the terminus of Camicia Glacier retreated 2.75 km and thinned by 335 m (average retreat of 72 m a^{-1}, average thinning of 9 m a^{-1}). Thinning and retreat were continuing when both the unnamed glacier and Camica Glacier were observed from the air by the author on 3 September 2002.

Prince William Sound Segment—The Northern Prince William Sound Subdivision

Northern Prince William Sound contains one of the greatest concentrations of calving tidewater glaciers in Alaska (figs. 41, 245, tables 2, 3). The Chugach Mountains fjords that have tidewater glaciers are located on the northern and

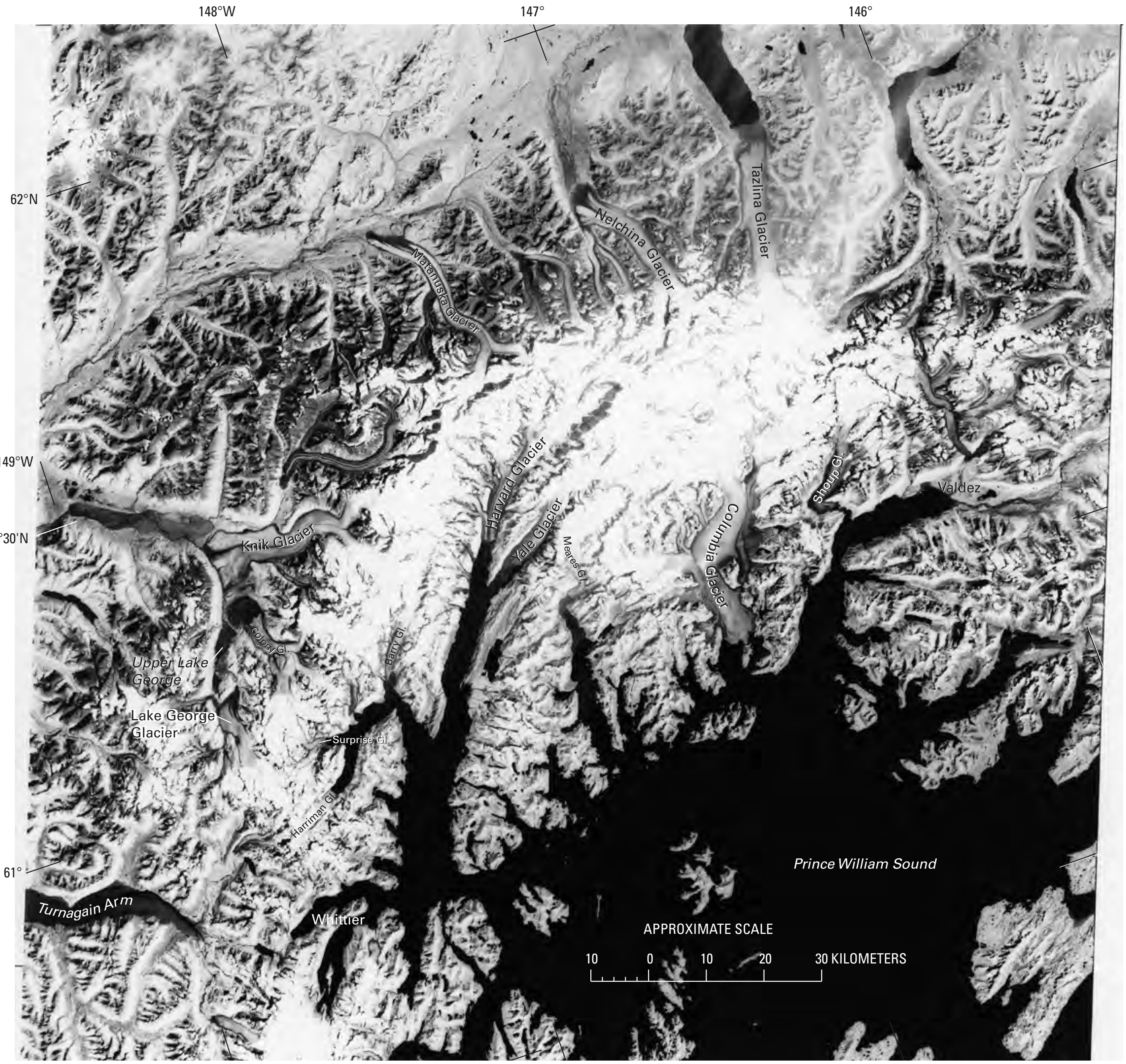


Figure 245.—Annotated Landsat 3 MSS image of the western Chugach Mountains on 26 August 1978 showing the location of many of the tidewater and valley glaciers in the region. Of the tidewater glaciers, Surprise and Barry Glaciers were slowly retreating; Harriman, Harvard, and Meares Glaciers were advancing slowly; and Yale and Columbia Glaciers were retreating rapidly (Meier and others, 1980). Shipping lanes used by tankers transporting oil from Valdez are within 20 km of Columbia Bay at the terminus of Columbia Glacier. Numerous small icebergs enter these lanes, creating a possible hazard to shipping. Knik Glacier dammed the outflow from the Lake George basin annually from 1918 to 1962 and in 1964 and 1965, resulting in annual glacier-outburst floods (jökulhlaups). Since 1966, no ice dam has formed (Post and Mayo, 1971). Matanuska and Nelchina Glaciers are both slowly retreating. Landsat image (30174–20290, band 7; 26 August 1978; Path 73, Row 17) and caption courtesy of Robert M. Krimmel, U.S. Geological Survey.

northwestern sides of Prince William Sound and include the following bays, inlets, and fjords: Shoup Bay, Columbia Bay, Unakwik Inlet, Harvard Arm and Yale Arm of College Fiord, Barry Arm, Harriman Fiord, and Port Wells. Glaciers in this subdivision that have lengths of 8 km or more and areas determined by Field (1975b, p. 470–472) [lengths and areas calculated later by Viens (1995) (table 2) are shown in brackets] include: Shoup (30 km, 146 km^2 [29 km, 156 km^2]), Columbia (66 km, 1,370 km^2 [59.5 km, 1,121 km^2]), Meares (25 km, 130 km^2 [25.7 km, 142 km^2]), Amherst (8 km, ~10 km^2, estimated by the author), Yale (35 km, 220 km^2 [32.2 km, 194 km^2]), Harvard (39 km, 500 km^2 [39.4 km, 524 km^2]), Smith [10 km, 20 km^2], Bryn Mawr [8 km, 26 km^2], Coxe (11 km, 19 km^2 [11.3 km, 20 km^2]), Barry (24 km 75 km^2 [26.5 km, 95 km^2]), Cascade (9 km, 15 km^2 [8 km, 15 km^2]), Serpentine (10 km, 23 km^2 [10 km, 30 km^2]), Surprise (13 km, 66 km^2 [12.1 km, 80 km^2]), Harriman (13 km, 49 km^2 [12.9 km, 60 km^2]), and Billings (8 km, 12 km^2). Information about many of the glaciers in the Prince William Sound region has been presented by Lethcoe (1987).

Shoup Glacier

At the end of the 20th century, Shoup Glacier (Post and Viens, 2000, pl. 2) was located approximately 6 km from the mouth of Shoup Bay, a sinuous embayment located on the northwestern shore of Valdez Arm. At its "Little Ice Age" maximum position, approximately 250 years ago, its terminus was astride the Valdez Arm shoreline, where it deposited a terminal moraine (Post and Viens, 2000). Hence, in the 250-year period between 1750 and 2000, Shoup Glacier retreated about 6 km, an average annual retreat of 24 m a^{-1}.

Originally named Canyon Creek Glacier by Abercrombie in 1898 (Orth, 1971), Shoup Glacier was first photographed the same year by Schrader (1900). At that time, it was located at the head of a shallow 2.5-km-long bay where it discharged into tidewater (2.5 km of retreat in 148 years, an average retreat of ~17 m a^{-1}). Tarr and Martin (1914) commented that, before Schrader's visit in 1898, there were no known previous observations of the glacier by either the Russians or by Whidbey, who visited Valdez Arm in 1794. Tarr and Martin (1914) further stated that the glacier also was photographed by the C&GS in 1901, by Grant in 1905 and 1908 (fig. 246), by the NGS in 1909 and 1910, and by Bradford Washburn on 14 June 1937. The area containing the glacier was surveyed in 1911, 1912, and 1916 by J.W. Bagley, C.E. Giffin, and R.H. Sargent for the USGS topographic quadrangle mapping program. Grant and Higgins (1913, p. 15), commented that Shoup glacier is "of economic importance in that it furnishes ice for Valdez and Fort Liscum, the detached bergs being lifted upon barges and being taken to these towns."

Figure 246.*—Two photographs of Shoup Glacier showing the position of its terminus near the start* (A) *and at the end of the 20th century* (B)*. Note the position of the triangular snow accumulation just behind the terminus in* A*. This feature is common to both photographs.* **A***, 13 July 1908 photograph by U.S. Grant of the terminus of Shoup Glacier (Grant and Higgins, 1911). A push moraine is located along much of the terminus, suggesting that Shoup Glacier was advancing when it was photographed. The terminus is fronted by an outwash fan delta.* **B***, 8 August 2000 north-looking oblique aerial photographic mosaic showing most of the area exposed by 20th century retreat. Note the large barren area around the margin. The proximal edge of the 1908 outwash fan delta is located on the right edge of the photograph. Photograph by Bruce F. Molnia, U.S. Geological Survey. Larger versions of these figures are available online.*

Field (1975b) stated that, from 1898 through 1957, there was relatively little change in the position of the glacier's terminus; the largest change was a recession of less than 200 m. The most significant change was more than 100 m of downwasting of the stagnant terminus. To complicate matters, small advances occurred between 1898 and 1901 (Tarr and Martin, 1914), between 1910 and 1914 (Keen, 1915a), and between 1935 and 1942 (Field, 1975b).

By 1961, the glacier had begun to retreat. Field (1975b) stated that, during the next 7 years, the glacier retreated between 500 and 600 m. In an unpublished map of the terminus of Shoup Glacier based on 1957 aerial photography and 1961 and 1964 field observations (American Geographical Society Glacier Studies Map No. 64–3–G5), Field showed that, at that time, the terminus of Shoup Glacier was separated from Shoup Bay by the terminal moraine, an outwash plain, and a tidal flat. He also showed that a 250-m-diameter tidal lake, the precursor to the present lake, had already formed.

A bathymetric map (Post and Viens, 2000, pl. 2) shows the position of the terminus of Shoup Glacier in 1916, 1966, 1978, 1982, and 1990. In 1990, the terminus sat at the western end of a 1.9-km-long lake named *Shoup Basin* by Post and Viens (2000). The eastern end was bounded by the moraine deposited by the glacier between 1957 and the early 1960s. When the author photographed the glacier from the air in August 1996, on 8 August 2000 (fig. 246*B*), and on 3 September 2002, he noted evidence of continued thinning and retreat. The terminus of Shoup Glacier was located a maximum of about 850 m behind its 1990 position, and the southwestern one-third of the margin of the terminus was fringed by a large area of bare bedrock and stagnant, debris-covered ice. In the 34 years between 1968 and 2002, the glacier retreated approximately another 2.85 to 3.15 km. During the 20th century, the glacier retreated a maximum of approximately 3.6 km.

In the early 1990s (Viens, 1995) (tables 2, 3), Shoup Glacier had a length of about 29 km, an area of 156 km^2, a width at its face of 1.3 km, and an AAR of 0.61. Its accumulation area was about 95 km^2, and its ablation area was about 62 km^2.

Columbia Glacier

The largest glacier in Prince William Sound, Columbia Glacier is located at the head of Columbia Bay (figs. 43, 44, 45, 46), a fjord with depths greater than 300 m. Previously called *Fremantle Glacier*, *Live Glacier*, and *Root Glacier*, Columbia Glacier had a pre-retreat length of 66 km and an area of 1,370 km^2 (Field, 1975b, p. 470). The ice speed at the terminus ranged between 5 and 6 m d^{-1}. Named in 1899 by the Harriman Alaska Expedition for Columbia University, it has been rapidly retreating and calving large quantities of icebergs since 1979. Until early 1979, part of the terminus of Columbia Glacier was grounded on Heather Island and an adjacent submarine moraine that protected the terminus from significant loss of ice through calving, as a 24 August 1978 AHAP photograph (fig. 247) shows. But, in January 1979, the glacier retreated from Heather Island and lost contact with its submarine terminal moraine. Thus, Columbia Glacier began a “drastic irreversible, retreat” (Meier and others, 1980, p. 10) that is well documented in this volume in the section authored by Robert M. Krimmel, “Columbia and Hubbard Tidewater Glaciers.” Columbia Glacier has since retreated more than 12 km and thinned as much as 400 m. By 2001, the length had decreased to about 54 km, and, according to Meier and others (2001), the velocity at the terminus in-

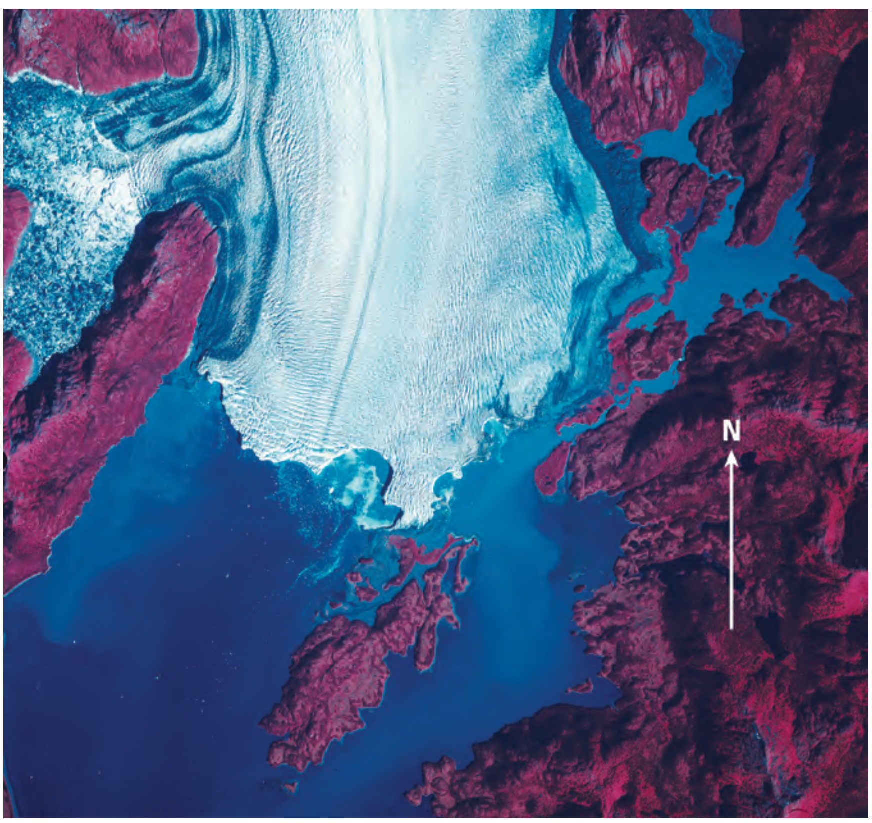


Figure 247.*—24 August 1978 AHAP false-color infrared vertical aerial photograph of the terminus of Columbia Glacier. Note that although part of the terminus of the glacier is in contact with Heather Island (at center of terminus), some retreat is underway along the southeastern margin. AHAP photograph no. L110F6623 from the GeoData Center, Geophysical Institute, University of Alaska, Fairbanks, Alaska.*

creased nearly fivefold to 25 m d^{-1} (more than 9 km a^{-1}). Columbia Glacier has its source on the southern side of the Chugach Mountains at maximum elevations that exceed 3,500 m. Four principal tributaries combine to form a south-flowing main trunk nearly 5 km wide.

Because of Columbia Glacier's proximity to the port of Valdez, which is the southern terminus of the Trans-Alaska Pipeline, icebergs produced by its retreat pose a real threat, especially to oil tankers transiting shipping lanes in Prince William Sound. These concerns resulted in a USGS comprehensive, multi-year investigation into the glacier's dynamics and an attempt to quantify the potential impacts of increased iceberg production. In addition to the summary provided by Krimmel in this volume, many of these data were released by the USGS in September 2001 (Krimmel, 2001). [Editors' note: According to Mark F. Meier (written commun., 2004), Columbia Glacier — a tidewater glacier in the retreat phase of the cycle of tidewater glacier stability (fig. 42) — is the best-documented and most thoroughly studied glacier in Alaska. Studies done on Columbia Glacier are comparable to studies of the surge-type Variegated Glacier. More than 120 aerial photogrammetric survey missions provided the data to compile an accurate record of its changing velocity and strain-rate fields. Studies of Columbia Glacier provide much of the scientific knowledge about calving of icebergs from a grounded glacier terminus.] Because Krimmel's section of this volume describes the late 20th century retreat of the tidewater glacier, the remainder of this discussion will focus on the early history of Columbia Glacier, which was the subject of much attention at the beginning of the 20th century.

Vancouver's 1794 voyage provided the first mention of the glacier as well as the first map showing its location (Vancouver, 1798). Tarr and Martin (1914) mentioned three other early maps — Applegate's in 1887 (Davidson, 1904), Mahlo's in 1898 (USGS, 1899), and Schrader's in 1900 (Schrader, 1900) — that also show the location of the terminus. All show the glacier at about the same location, leading Tarr and Martin (1914, p. 257) to conclude that "The only fact indicated by these maps is that the glacier was not strikingly different in the years of observation between 1794 and 1900." It must be kept in mind, however, that none of these earlier (pre-1970s) observations were made in the context of tidewater glacier stability (fig. 42), a theory first set forth by Post (1975).

Tarr and Martin (1914, p. 259) reported that, in 1898, the first water depth sounding of approximately 90 m (50 fathoms) near the face of Columbia Glacier was made by the steamship *SS Dora*, captained by A.O. Johansen. In 1899, the Harriman Alaska Expedition arrived in Columbia Bay and named and studied the glacier. From 25 to 28 June 1899, USGS geologist Grove Karl Gilbert (Gilbert, 1904) photographed and mapped the lower 15 km of the glacier (figs. 248, 249, 250), documenting an advance that occurred about 1892 and a retreat underway in 1899.

Grant visited and photographed the glacier in 1905. He was joined by Higgins in both 1908 and 1909 (Grant and Higgins, 1910, 1913). Their collaboration on 23 and 24 June 1909 produced a map of the terminus that compares the 1899 and 1909 terminus positions. They concluded that the glacier retreated between 1899 and 1905, advanced between 1905 and 1908, and advanced even more through June 1909. Tarr and Martin (1914) visited the glacier for the NGS in 1909, twice in 1910, and again in 1911 and observed that the glacier continued to advance through their 1911 visit. Their topographic map shows the terminus of an unnamed eastern branch, most of whose length is separated from the main trunk of the glacier by the Nunatak, making contact with the glacier just below the 1,000-ft contour line (Tarr and Martin, 1914). Gilbert (1904) showed that the two were in only minimal contact on his 1899 map. Tarr and Martin quantified the changes that they and Gilbert observed: (1) 1892–July 1899, retreat of 243 m (800 ft), an average retreat of about 35 m a^{-1}; (2) July 1899–24 June 1909, advance of about 150 m

Figure 248.—*Photographs by USGS geologist Grove Karl Gilbert of the lower Columbia Glacier.* **A**, *25 June 1899 photograph of the Harriman Alaska Expedition's vessel* S.S. George W. Elder *in Columbia Bay, with the grounded terminus (on submarine moraine) of Columbia Glacier in the background. USGS Photo Library photograph Gilbert 339.* **B**, *26 June 1899 photograph from the highest point on Heather Island, a 107-m-high hill, of the lower reaches of Columbia Glacier. Note the debris-covered eastern margin. USGS Photo Library photograph Gilbert 343.*

(500+ ft), an average advance of about 15 m a^{-1}; (3) 23 August 1909–5 July 1910, advance of 213 m (700 ft), an average advance of about 245 m a^{-1}; (4) 5 July 1910–5 September 1910, advance of 33 m (~110 ft), an average advance of about 200 m a^{-1}; and (5) 5 September 1910–21 June 1911, advance of less than 30 m (<100 ft), an average advance of about 38 m a^{-1}. Hence, during the 19-year period from 1892 thru 1911, Columbia Glacier advanced approximately 290 m, an average advance of about 15 m a^{-1}.

In 1910, Bean (1911) investigated the "submerged lands" of Prince William Sound. He described his study as follows: "The aim of this thesis is to study the soundings of this embayment (Prince William Sound) and from this study outline the history of the fiords, and determine as far as possible the extent of former glaciers and their influence upon preglacial valleys. Especial attention will be paid to that part of the fiord which is now submerged" (Bean, 1911, p. 2). Working in July and August 1910 as part of an NGS Expedition under the direction of Lawrence Martin, Bean took about 300 soundings in about 150 miles of Prince William Sound using a 20-lb lead weight obtained from the C&GS. He took soundings for his cross sections every 1/8 to 1/2 mi. Along the long axes of the fjord, he spaced soundings at 1- or 2-mi intervals. A sounding taken approximately 400 m in front of Columbia Glacier showed 182 m (600 ft) of water (Tarr and Martin, 1914). Bean found that depths increased with distance from the ice margin—275 m at a point 5.6 km from the glacier and 310 m at a point 8 km from the glacier—and that water depths

Figure 249.—*25–28 June 1899 map by USGS geologist Grove Karl Gilbert of the lower 15 km of Columbia Glacier. Contours are in feet. Published as Plate XI in Gilbert (1904).*

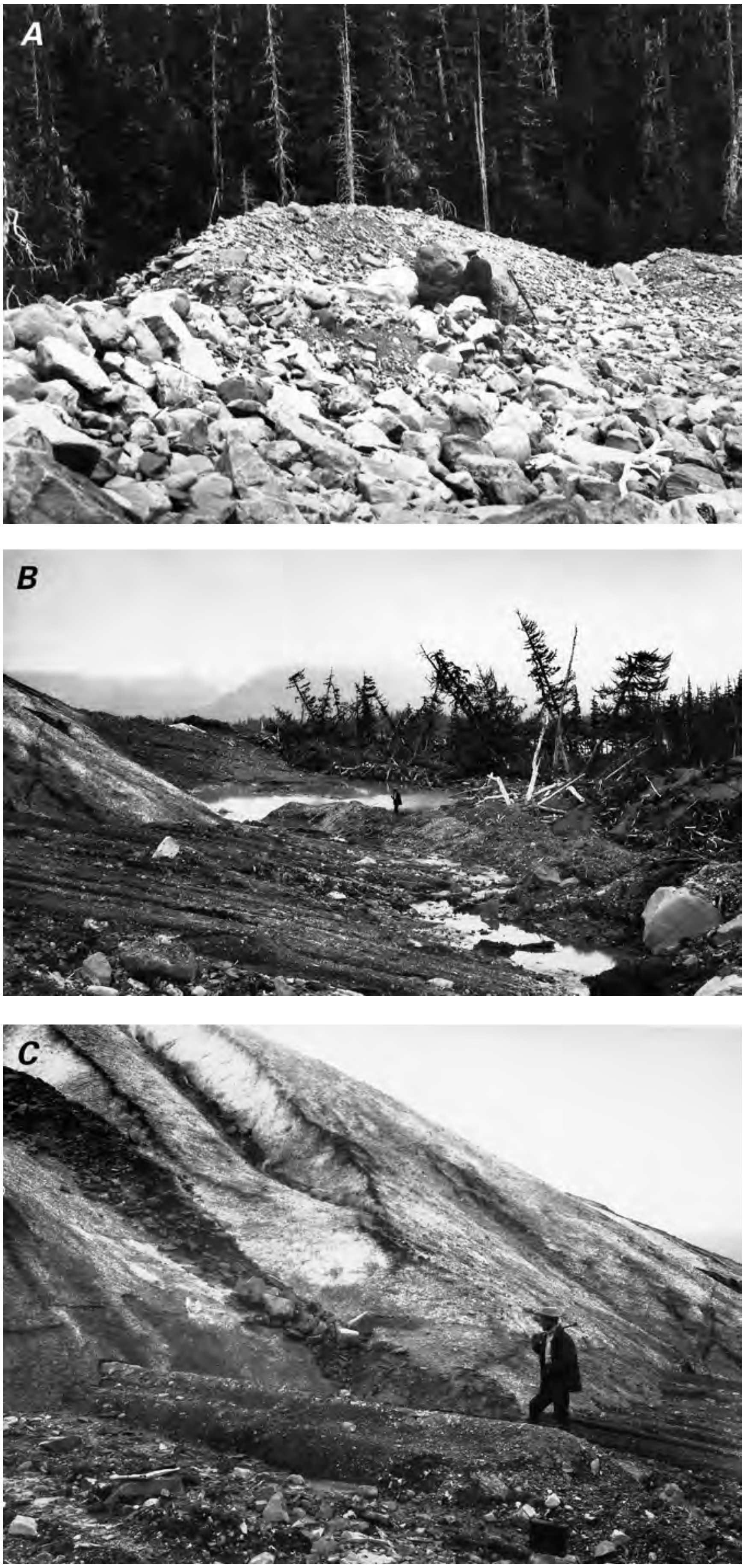


Figure 250.—*Three 26 June 1899 ground photographs by USGS geologist Grove Karl Gilbert, with Professor Charles Palache (Harvard University geologist) for scale, showing late 19th century features at the terminus of Columbia Glacier.* **A**, *An arcuate lobe of a push moraine formed by an advance of Columbia Glacier before 1892. Immediately adjacent to the moraine are a number of trees that were killed by the advance. USGS Photo Library photograph Gilbert 356.* **B**, *A recessional moraine formed after 1892. Note the trees that were killed by the advance of the glacier. Between the moraine and Columbia Glacier are numerous flutes exposed by the retreating ice. USGS Photo Library photograph Gilbert 354.* **C**, *The retreating margin of Columbia Glacier and an approximately 0.75-m-high flute exposed by the retreating ice. USGS Photo Library photograph Gilbert 352.*

increased uniformly by about 30 m with every 1,600-m increment away from the face of the glacier. A complete bathymetric survey of Columbia Glacier's margin and approaches was published by Post (1978).

According to Field (1975b), no additional observations were made of the glacier until 1931. In 1931, Field (1932) attempted to reoccupy many of the earlier photographic stations of Gilbert, Grant and Higgins, and Tarr and Martin. He reported that "All the earlier moraines and trimlines had been destroyed by an advance which was then tentatively assigned to the early 1920s and which was later determined by vegetation studies to have occurred at various points along the terminus between 1917 and 1922..." (Field, 1975b, p. 369). He did not specify the magnitude of the advance. Following the 1917 to 1922 advance, the glacier began to retreat. By 1931, Field (1932) measured 275 m of recession at the western margin of the glacier, about 60 m of recession on Heather Island, and from 120 to 250 m of recession along the land-based eastern margin.

This retreat was quickly followed by another advance. Field (1937) revisited Columbia Glacier in 1935 and measured 1,045 m of advance at the western margin of the glacier, 72 m of advance on Heather Island, and 25 m of advance on the land-based eastern margin. Field (1937) described a massive moraine deposited by this advance on Heather Island as a prominent feature that protrudes through the 1920 moraine.

Maynard M. Miller visited Columbia Glacier in 1947 and described a recession of several hundred feet since the 1935 advance (Miller, 1948a). Field (1975b) reported that an advance of 100 m or more was underway between 1947 and 1949. Eight years later, he observed that the glacier was slowly retreating from the moraine formed by the late 1940s advance.

Field (1975b) reported that observations of Columbia Glacier were made by a variety of individuals in 1960, 1961, 1963, 1964, 1965, 1966, 1967, 1968, 1969, 1970, and 1971. During this period, Columbia Glacier experienced several small advances between 1957 and 1961, recession or minimal change between 1961 and 1968, a slight advance of a few tens of meters on land between 1968 and 1969, and an advance on Heather Island and on the eastern land-based terminus between 1969 and 1971. Field (1975b, p. 369) remarked that "Each of the advances since the maximum of 1920 have in most places not extended as far as the previous one, so that their end-moraines have been preserved." Most of the fluctuations in terminus position described here are for the land-based part of Columbia's terminus where the moraine records serve as near-permanent markers to document change. Field (1975b, p. 370) noted that the marine-based termini on either side of Heather Island "have tended to fluctuate with the land termini, but on the whole have remained remarkably stable, in so far as appreciable net change is concerned. Their general appearance in the 1960s was essentially the same as in 1899."

Post (1975), Field (1975b), and Sikonia and Post (1979) described observed changes in the tidal ice-cliff termini of Columbia Glacier and suggested that the increasing size and frequency of seasonal embayments in the terminus might be a precursor to a rapid retreat. Retreat from the terminal moraine shoal, over which water depths are about 20 m or less, formed deep embayments in the ice margin during the summer and fall. Until 1979, winter and spring readvances were normally sufficient to refill the embayment and recover stability on the submarine moraine. Field (1975b) noted that, in a 1938 Washburn photograph, deep embayments existed on both sides of Heather Island. He speculated that, if they had joined north of Heather Island, "they would have probably caused a wide retreat of the whole terminus" (Field, 1975b, p. 370–71). A complete inventory of years in which embayments formed does not exist. However, they were observed in 1938, 1939, 1940, 1941, 1960, 1961, 1964, 1965, 1968, and throughout the 1970s. The lack of embayments between the middle 1940s and 1960 is more likely owing to an absence of observation rather than to an absence of embayments.

The 1964 embayment was 900 m deep and 1,300 m wide, whereas the 1971 embayment was 1,200 m deep and 800 m wide (Field, 1975b). Sequential changes at the terminus of Columbia Glacier from 1966 to 2000 are shown in figure 251. Figure 252 shows different views of the margin of the Columbia Glacier on 8 August 2000. The author began observations from the air in 1974 and has continued them into the 21st century (fig. 252).

Figure 251.—*Four aerial photographs show changes in the terminus region of Columbia Glacier between 1966 and 2000.* **A**, *3 September 1966 oblique aerial photograph of the lower reaches of Columbia Glacier. Note that part of the terminus of Columbia Glacier is in contact with Heather Island. USGS photograph no. 664–18 by Austin Post, U.S. Geological Survey.* **B**, *15 August 1981 north-looking oblique aerial photograph of the lower reaches of Columbia Glacier. Note that the terminus of the glacier is retreating and has lost contact with Heather Island and the mainland to the east. Note also the calving embayment. Photograph by Bruce F. Molnia, U.S. Geological Survey. A larger version of this figure is available online.*

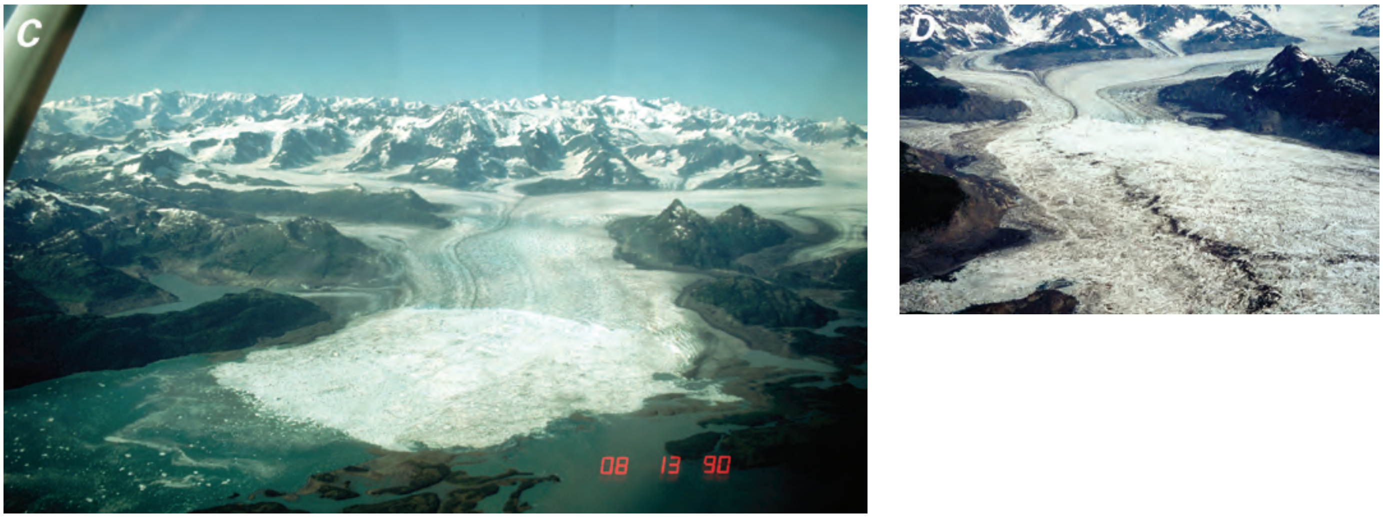


Figure 251.—**C**, *13 August 1990 north-looking oblique aerial photograph of the lower reaches of Columbia Glacier. Note that the terminus of the glacier is actively retreating and that the newly exposed part of Columbia Bay bounded on the south by Heather Island is filled by approximately 5 km of icebergs that are contained by the Heather Island terminal moraine. Photograph by Larry Mayo, U.S. Geological Survey.* **D**, *8 August 2000 north-looking oblique aerial photograph of the lower reaches of Columbia Glacier. Note that the terminus has retreated approximately 12 km and that the newly exposed part of Columbia Bay is filled with icebergs. Photograph by Bruce F. Molnia, U.S. Geological Survey. A larger version of* D *is available online.*

Figure 252.—*Three 8 August 2000 oblique aerial photographs of features along the margin of Columbia Glacier.* **A**, *Northwest-looking view of the retreating and thinning eastern Land Lobe of Columbia Glacier.* **B**, *North-looking view of a lake and an unnamed ice remnant left behind by the rapidly retreating main trunk. The lake at the upper right separates the remnant of ice from the southern margin of the retreating Land Lobe that can be seen on* A. *Iceberg-filled Columbia Bay is to the left.* **C**, *Southeast-looking view of Kadin Lake, an ice-dammed lake on the west side of the Columbia Glacier, currently dammed by an unnamed remnant of ice left behind by the rapidly retreating main trunk. The remnant is actively calving icebergs into Kadin Lake and Columbia Bay. The ice remnant of* B *can be barely seen at the top left of the photograph. Photographs by Bruce F. Molnia, U.S. Geological Survey. Larger versions of these figures are available online.*

A comparison of 1950s map data of the glacier with data obtained during geodetic airborne laser altimeter profiling surveys in the middle and late 1990s showed that, on an annual basis, Columbia Glacier thinned by 1.44 m a^{-1} and decreased in volume by 1.54 km^3 a^{-1}. However between the middle 1990s and 1999, Columbia Glacier thinned by an average of 7.42 m a^{-1} and decreased in volume by 7.64 km^3 a^{-1} (K.A. Echelmeyer, W.D. Harrison, V.B. Valentine, and S.I. Zirnheld, University of Alaska Fairbanks, written commun., March 2001). In the early 1990s (Viens, 1995) (tables 2, 3), Columbia Glacier had a length of about 60 km, an area of about 1,120 km^2, a width at its calving terminus of 8 km, and an AAR of 0.67. Its accumulation area was about 753 km^2, and its ablation area was about 368 km^2.

Meares Glacier

Meares Glacier is located at the head of Unakwik Inlet, a 32-km-long northward-trending fjord between College Fiord and Columbia Bay. The first European explorer to visit the inlet was Whidbey on 4 June 1794 (Vancouver, 1798). Although he did not sail close enough to observe the glacier and determine its position, floating pieces of ice that he noted in the fjord suggested that Meares Glacier was a tidewater glacier at that time. Meares Glacier was first investigated by Grant and Higgins (1910, 1911a, b, 1913) in 1905 and 1909 and again by Lawrence Martin in 1910 (Martin, 1913). At the beginning of the 21st century, Meares Glacier was an advancing, iceberg-calving, tidewater glacier (fig. 253).

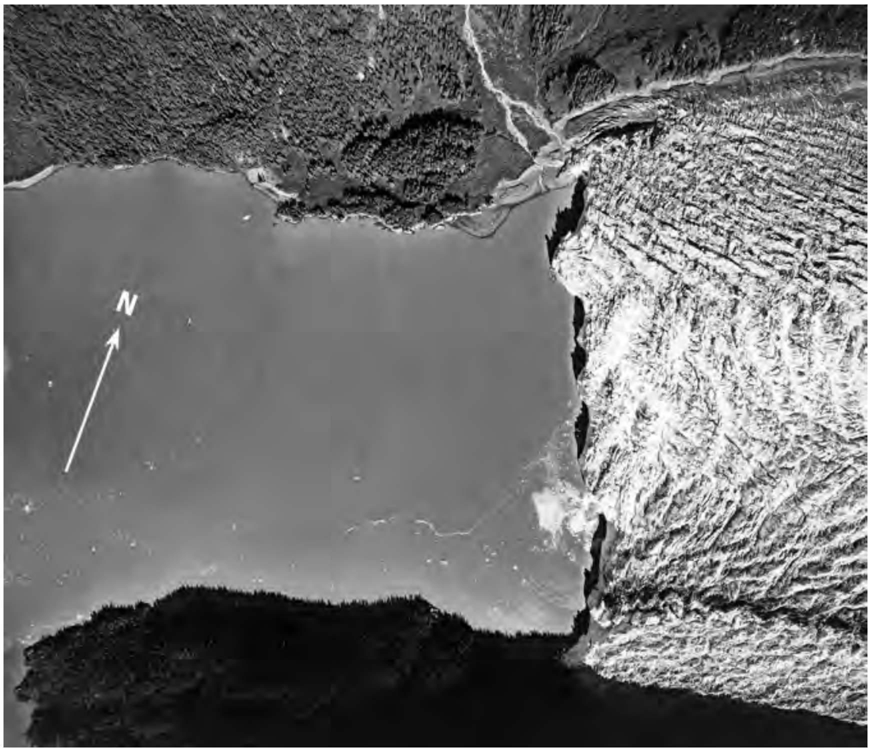


Figure 253.—*17 August 1999 vertical aerial photograph of the terminus of Meares Glacier. Before the date of the photograph, Meares Glacier advanced into Ranney Creek (to the left of the terminus), diverting its waters to the south. Subsequent calving and retreat of the ice at the western margin resulted in the stranding of a block of ice along the edge of Ranney Creek. Subsequently, the ice was incorporated into the advancing terminus of the glacier. Photograph no. R3–FL7–FR119 from the U.S. Bureau of Land Management.*

Meares Glacier has been advancing since it was first observed in 1905. Field (1975b) summarized the reported changes of the terminus of Meares Glacier as follows: 1910–31, advance of 250 to 300 m, annual advance of 12 to 14 m a^{-1}; 1931–66, advance of 200 to 425 m, an annual advance of 6–12 m a^{-1}. Hence, in the 56 years between 1910 and 1966, the terminus of Meares Glacier advanced between 500 and 675 m, an average advance of 9 to12 m a^{-1}. When the author visited by boat on 3 September 2000, the terminus of Meares Glacier was slowly advancing and pushing down trees. However, the author's observations from the air in 1999, 2000, and 2002 showed some evidence of thinning along the margins of the terminus, especially on its eastern side. Between 1910 and 2000, the terminus advanced about 1.1 km, an average advance of 12.2 m a^{-1} (fig. 254). In the 36-year-period between 1966 and 2002, the glacier advanced a maximum of about 550 m, for an average annual advance of about 15 m a^{-1}. A large terminal moraine about halfway between the glacier and the mouth of the fjord represents the maximum neoglacial advance of Meares Glacier and presents a significant hazard to navigation.

In 1910, Bean (1911) determined that much evidence of former glacial erosion existed in the submerged lands downfjord from the glacier margins. At Meares Glacier he found that "The depth of water ranges from 534 feet about a half mile south of the glacier to over 1,000 feet near the entrance of the inlet. There is much topographic evidence of former glaciation. The fiord walls are straight and steep, with all spurs truncated, there are numerous hanging valleys from which streams cascade to the sea, and there is abundant rounding of knobs, grooving, plucking and striation. The soundings prove quite conclusively that the Meares Glacier formerly extended much farther, and eroded this great valley."

Bean (1911, p. 19–20) also determined that Meares Glacier was not floating but was grounded on its bed: "On the assumption that the water does not shallow from the nearest sounding to the glacier, it seems certain that the Meares is not afloat even now, but has great erosive power. Since the height of the ice front is at least 200 feet and the depth of the water is 534 feet, there is a total thickness of 734 feet. If glacier ice floats in sea water with one-sixth of its volume above water, the total thickness of the glacier would have to be 1,200 feet or 456 feet more than it is, in order to be afloat. Five hundred and thirty-four

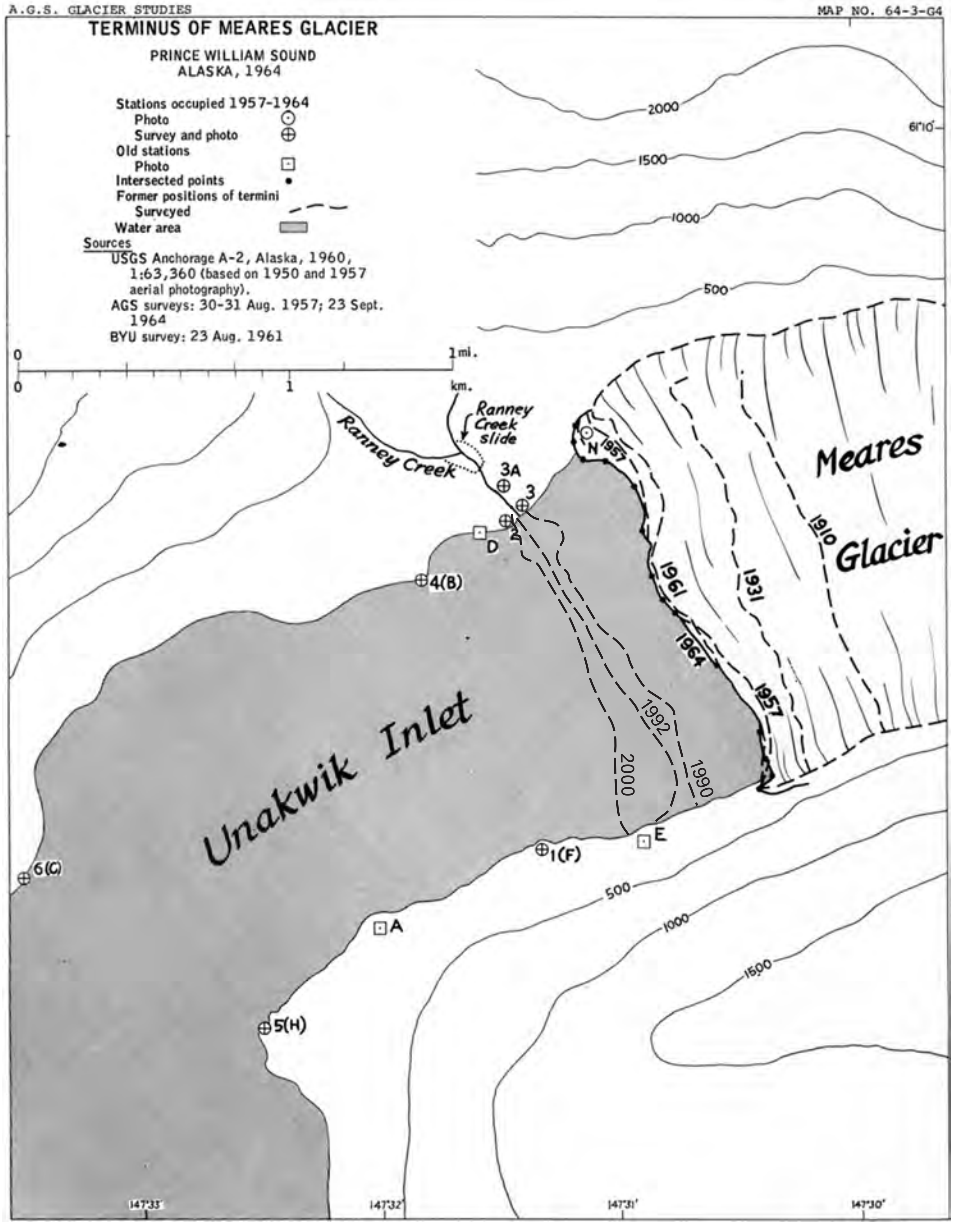


__Figure 254.__—Map of the terminus of Meares Glacier showing fluctuations in the position of its terminus from 1910 to 2000. Base map is derived from AGS Glacier Studies Map no. 64–3–G4 (Field, 1965). Year 1990, 1992, and 2000 positions (approximate) are based on observations made by the author.

feet of water is sufficient to float a mass of ice with but 107 feet above water. Hence the Meares Glacier is at present resting on the valley floor with the weight of 93 cubic feet of ice or about 4,650 pounds per square foot...."

In the early 1990s (Viens, 1995) (tables 2, 3), Meares Glacier had a length of 26 km, an area of 142 km^2, a width at its face of 1.2 km, and an AAR of 0.86. Its accumulation area was 121 km^2, and its ablation area was 20 km^2.

Pedro and Brilliant Glaciers

Pedro Glacier and Brilliant Glacier, both located on the eastern side of Unakwik Inlet, are two small retreating valley glaciers that were tributaries to a much larger Meares Glacier at the time when it was extended to the position where it deposited its middle fjord moraine. Tarr and Martin (1914) described Brilliant Glacier as descending to an elevation near sea level and being separated from the inlet by a 2.2-km-long outwash plain and delta. By the 1950s (Field, 1975b) observed that its terminus had retreated to an elevation of between 200 and 250 m and that the outwash plain complex had extended to be about 2.5 km long. When the author observed it from the air in 1999, 2000, and 2002, the terminus was located at an elevation of about

350 m, and the margin of Brilliant Glacier showed much evidence of continuing retreat.

Ranney and Baby Glaciers

Ranney Glacier, a hanging glacier located on the northwestern wall of the Unakwik Inlet, is also retreating. The 1964 earthquake generated a large rockslide that covered much of the lower part of the glacier as it continued down below the terminus to the delta of Ranney Creek (Post, 1967a), the stream that heads at the glacier's terminus. Ranney Glacier has been retreating since it was first observed. Similarly, Baby Glacier—a 2-km-long valley glacier—drains into the western side of the inlet. Like Ranney Glacier, it has been retreating throughout the 20th and early 21st centuries.

Glaciers of College Fiord

College Fiord (figs. 255, 256), located at the northern end of Port Wells, is about 40 km long and 5 km wide at most. At the beginning of the 21st century, it contained five calving tidewater glaciers: Harvard Glacier and Yale Glacier (fig. 34) (AGS Glacier Studies Map No. 64–3–G10) (Field, 1965), in two arms (Harvard Arm and Yale Arm) at the fjord's head, and Smith, Bryn Mawr, and Wellesley Glaciers on its western wall. Named non-tidewater glaciers in the fjord and in the eastern side valleys include Cap, Tommy, Crescent, Amherst, Lafayette, Williams, Muth, Dartmouth, Downer, Lowell, Eliot, Radcliffe,

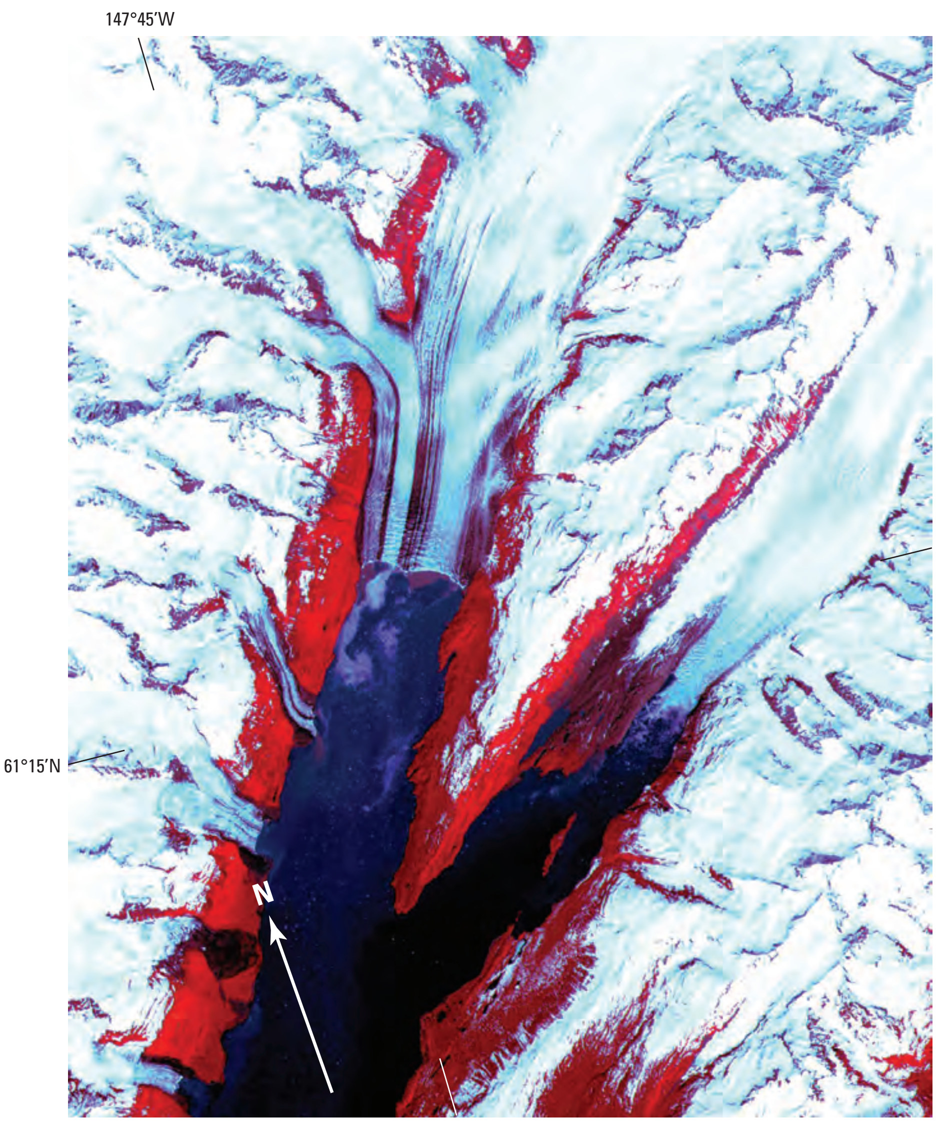


Figure 255.*—24 June 2000 Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) image of upper College Fiord. The ASTER multispectral sensor on the Terra spacecraft has 15-m-pixel multispectral resolution and is another type of satellite image used to monitor changes in Alaskan glaciers. ASTER image (AST_LIB.003:2012687866) from the U.S. Geological Survey, EROS Data Center, Sioux Falls, S. Dak.*

Figure 256.—*8 August 2000 north-looking view of College Fiord shows all five of the tidewater glaciers located on the west side. Photograph by Bruce F. Molnia, U.S. Geological Survey.*

Baltimore, Vassar, Barnard, and Holyoke Glaciers. With the exception of previously named Cap and Tommy Glaciers, nearly all of these glaciers were named by the geologists of the Harriman Alaska Expedition of 1899, by Grant and Higgins between 1908 and 1912, and by Lawrence Martin in 1910 for Ivy League colleges and eastern universities with which they had affiliations. Included are all seven of the "Seven Sisters" colleges at that time, women's colleges associated with what were then all-male Ivy League colleges, about half of the Ivy League colleges, two other woman's colleges, and two past presidents of Harvard University. Baltimore Glacier was named by Grant and Higgins in 1908 for the Woman's College of Baltimore, now Goucher College; Eliot Glacier was named by Martin in 1910 for Charles William Eliot, former president of Harvard (Tarr and Martin, 1914). Downer Glacier was named for the Milwaukee-Downer College for Women by Martin in 1910; Lowell Glacier, first reported by the USGS in 1915, took its name from the first non-native settler in the Seward area (Orth, 1967).

Most of College Fiord's glaciers have shown a general trend of slow recession since first being mapped at the beginning of the 20th century. Harvard Glacier, however, has been slowly advancing for nearly 100 years. Yale Glacier retreated rapidly with more than 6 km of recession during the 20th century. The 1964 earthquake caused rockslides and avalanches on Smith, Vassar, Harvard, and Yale Glaciers.

Cap and Tommy Glaciers

Cap Glacier, an unnamed glacier, and Tommy Glacier are the three southernmost glaciers that ultimately drain into College Fiord. Cap Glacier, located on the northern wall of its valley, is a retreating hanging glacier surrounded by exposed bedrock. Its outlet stream enters Davis Lake. The unnamed glacier, located at the head of the Avery River, is a small retreating valley glacier. Tommy Glacier, located on the northern wall of the Avery River valley, is a retreating hanging glacier surrounded by exposed bedrock. When the author observed them from the air on 8 August 2000, all showed evidence of continued retreat.

Crescent and Amherst Glaciers

In 1957, Crescent and Amherst Glaciers were visited by an AGS field party (AGS Glacier Studies Map No. 64–3–G3) (Field, 1965) that included Field and Viereck (Field, 1975b). This party established photographic stations and mapped the glacier termini. The termini of both glaciers had been photographed previously from the deck of the Harriman Alaska Expedition vessel *SS George W. Elder* by Merriam in June 1899. Viereck's 1957 investigation found an end moraine that was 800 m beyond the 1957 margin of Crescent Glacier and a trimline about 60 m above the 1957 ice surface. On the basis of dendrochronology, Viereck concluded that the moraine had formed about 125 years earlier, or ca. 1830. A second moraine 125 m from the 1957 terminus was dated to the middle 1930s. The author has photographed the glacier several times from the air between 1975 and 2002. When he observed it on 5 September 2000, the terminus of Crescent Glacier had retreated about 1.5 km from its 1957 position. Hence, in the 170 years since abandoning its probable "Little Ice Age" maximum position, the glacier has retreated almost 2.3 km, yielding an average retreat rate of 13.5 m a^{-1}. Between 1830 and 1957, the glacier retreated at an average rate of 6.3 m a^{-1} and thinned an average of 0.47 m a^{-1}. Between 1957 and 2000, the average rate of retreat was 34.9 m a^{-1}.

The 1957 AGS visit to Amherst Glacier was the first on record. Other than subsequent aerial reconnaissance, it is still the only recorded visit to date. Field (1975b) summarized the 1914–71 history of the glacier by analyzing maps and aerial photography. He reported that, when the glacier was first mapped, it ended on an outwash plain. By 1935, when Bradford Washburn photographed it, the terminus had retreated and caused the formation of an ice-marginal lake. Aerial photographs taken in 1941 by the U.S. Army Air Forces, in 1950 and 1957 by the USAF, in 1964 by the C&GS, in 1965 by the USGS, and in 1966 and 1971 by the AGS all showed continuing retreat and expansion of the lake. This trend was continuing when the author observed Amherst Glacier on 14 July 1978 (fig. 257).

Field (1975b) noted that an end moraine is located on the outwash plain about 1.5 km in front of the 1957 terminus. Unlike the terminal moraine of Crescent Glacier, it could not be dated. The distance between the position of the terminus when it was first surveyed in 1914–16 and again in 1950 is between 500 and 600 m, suggesting a retreat at an average rate of about 15 m a^{-1}. Between 1935 and 1950, the glacier retreated about 200 m, suggesting a retreat at an average rate of about 13 m a^{-1}. Between 1950 and 1957, the glacier retreated about 100 m, suggesting a retreat at an average rate of about 14 m a^{-1}. Field (1975b) speculated that the undated end moraine could be contemporary with the moraine at adjacent Crescent Glacier. When the author observed the terminus of Amherst Glacier from the air on 4 August 2000 and 3 September 2002 and from a boat on 3 September 2000, it had retreated about 1.4 km from its 1957 position. Therefore, since first being mapped in 1914–16, the glacier has retreated about 2.0 km, with an average retreat rate of approximately 24 m a^{-1}. Because the glacier margin is ringed by an ice-marginal lake, part of this retreat may have been caused by calving.

Figure 257.*—14 July 1978 east-looking photograph of Amherst Glacier. A trimline flanks each lateral margin, and an apron of bedrock surrounds the terminus, which has been retreating at a rate of approximately 15 m a^{-1} for most of the 20th century. Photograph by Bruce F. Molnia, U.S. Geological Survey. A larger version of this figure is available online.*

Lafayette Glacier

Lafayette Glacier is another small west-flowing valley glacier that has received minimal scientific attention. Baird and Field (1951) reported that, during the 2-year period between observations in 1947 and 1949, the glacier's terminus retreated 85 m. Similarly, Field (1975b) reported that a comparison of Bradford Washburn's 1935 photograph and a vertical aerial photograph taken in 1957 by the USAF shows a retreat of 600 to 700 m. When the author observed the glacier from a boat on 3 September 2000 and from the air on 3 September 2002, it had retreated about another 1.2 km from its 1957 position. Hence, in the 67 years between 1935 and 2002, the glacier retreated about 1.85 km, yielding an average retreat rate of about 28 m a^{-1}.

Glaciers Draining into Coghill Lake

Coghill Lake, which drains into College Fiord through the Coghill River, is a 5-km-long lake that is a part of the sediment-filled eastern arm of College Fiord. The Coghill Lake and Coghill River basin are separated from College Fiord by a ridge with maximum elevations exceeding 710 m. A number of small west-flowing glaciers descend from the ridge that separates College Fiord and Unakwik Inlet. The named glaciers are Williams, Muth, and Dartmouth Glaciers.

Williams Glacier.—Williams Glacier is an approximately 2-km-long unstudied glacier that is sheltered from sunlight by the ridge that separates College Fiord from Unakwik Inlet. When the author observed it from the air on 8 August 2000, it showed evidence of retreat.

Muth Glacier.—In l908, Grant and Higgins gave the name Muth Glacier to a previously unnamed glacier located on the ridge between Unakwik Inlet and College Fiord. However, it is uncertain which glacier they actually named. The USGS Anchorage A–2 1:63,360-scale topographic quadrangle map (1960) (appendix B) shows Muth Glacier as a westward-flowing glacier that drains into Coghill Lake. However, Bean (1911), who mapped the bathymetry of Port Wells and College Fiord in 1910, showed Muth Glacier as being on the eastern side of this divide, approximately 7.5 km to the south. Bean's Muth Glacier is an eastward-flowing glacier within the Unakwik Inlet drainage, draining into Jonah Bay. Both glaciers show signs of continuing retreat and thinning.

Dartmouth Glacier.—Dartmouth Glacier is a 3.5-km-long southwest-flowing unstudied valley glacier that has been retreating since it was first photographed in 1941. The 1941 terminus sits nearly 800 m behind a trimline that marks a major previous undated expansion of the glacier. Field (1975b) stated that, between 1941 and 1966, the terminus retreated about 750 m, an annual average rate of 30 m a^{-1}. When the author photographed the terminus from the air in August 2000, it was approximately 1.3 km behind the 1966 terminus. During this 34-year interval, the average rate of retreat was around 38 m a^{-1}. Between 1941 and 2000, the glacier retreated about 2.1 km, an average rate of retreat of about 35 m a^{-1}.

Bathymetry of College Fiord

Bean (1911) performed the first bathymetric survey of College Fiord in 1910. He found that the fjord walls were very steep both above and below the water line, that all spurs were truncated, and that water depths in the fjord ranged from 38 to 245 m, being approximately 194 m deep about 300 m from the 1910 face of Harvard Glacier, and 86 m deep about 540 m from Yale Glacier (Tarr and Martin, 1914, pl. CXXVI, ff p. 320). This shallowing of the water is now known to be the result of a submarine moraine at the grounded terminus of an advancing tidewater glacier, in conformance with the theory of tidewater glacier stability (Post, 1975) (fig. 42).

Glaciers of Yale Arm

Yale Glacier.—Yale Glacier, an active calving tidewater glacier, sits at the head of Yale Arm. It originates in snowfields around Mount Witherspoon, which has a summit elevation of more than 3,500 m. In 2002, the terminus of Yale Glacier consisted of two parts—a southeastern tidewater calving terminus with an estimated 15- to 20-m-high ice face and an approximate width of 1 km and a land-based northwestern terminus about 1.2 km wide that lies within a basin of bedrock bosses and knobs that it previously covered. Yale Glacier has retreated nearly 8 km from its "Little Ice Age" maximum position adjacent to College Point at the southern end of Dora Keen Ridge, which separates Yale Arm from Harvard Arm.

In 1887, Yale Glacier was first mapped by Applegate (Davidson,1904) from a location adjacent to Coghill Point. In late April 1898, it was visited and photographed by Mendenhall (see USGS Photo Library photograph Brooks 1114) and visited by Castner (Glenn, 1899). Castner walked over much of the terminus on snowshoes. That summer, the glacier was mapped again by Glenn (1899). In 1899, it was also observed by the Harriman Alaska Expedition (Tarr and Martin, 1914, pl. CXXI, ff p. 320) and mapped again by Gannett (Gannett, 1899; Gilbert, 1904). Detailed investigations were performed in 1905 and 1909 by Grant and Higgins (1910, 1911a, b, 1913) and by Martin (1913) and Bean (1911) in 1910. In 1910, field surveys were conducted under the direction of Martin for the NGS, with topography by W.B Lewis and F.E. Williams and soundings by E.F. Bean (Tarr and Martin, 1914, pl. CXXII, ff p. 320). The resulting contour map, published by Tarr and Martin (1914), depicted the lower reaches of Yale, Downer, Lowell, Harvard, Eliot, Radcliffe, Baltimore, Smith, Bryn Mawr, Vassar, and Wellesley Glaciers. Not only did Martin's observations serve as the basis for his doctoral dissertation, but the map produced by his field party is also the first precise map made of the upper College Fiord region.

Glenn (1899, p. 19–20) described Yale Glacier and adjacent Harvard Glacier as follows:

> The day was dry and clear. Directly in our front was the most imposing sight we had yet seen—I might add more imposing than any we saw during the season. Glistening in the sun were two large glaciers, which we named the "Twin Glaciers," the pair being separated by a short ridge of hogback that runs down to salt water. In front of the one on our right (Yale) the sea ice extended for over three miles, while in front of the other (Harvard) this sea ice extended at least twice that distance. The ice was covered with snow several feet in depth. We soon discovered that it would bear up the weight of a man and that we could make no headway against it with the boat. Each of these glaciers is what is termed "live" or "working" glaciers. The front of each was an almost perpendicular mass of ice, from which immense pieces were constantly breaking off and falling into the sea with a great roaring noise, due principally to the action of the tides.

Cooper's (1942) observations of the area around the glacier in 1935 led him to conclude that Yale Glacier had been both thicker and longer quite recently. Development of alder (*Alnus* sp.) thickets about 200 m above the southeastern side of the glacier led Cooper to conclude that the terminus of Yale Glacier had previously reached the mouth of Yale Arm, about 3.2 km in front of its 1935 terminus. Twenty-two years later, Field (1975b) reported that Viereck, a member of the 1957 AGS field party, examined a trimline that he found on the northeastern side of the glacier. On the basis of dendrochronological information that he collected, he concluded that Yale Glacier had reached its most recent maximum position sometime 130 to 150 years before his study (between 1807 and 1827). He also concluded that no larger glacier advance had occurred since at least 1650.

The distance between this 1807–27 limit and the 1910 position of the terminus ranges from 3.5 to 5.0 km. The average rate of retreat during this interval ranges from a minimum of 34 m a^{-1} to a maximum of 60 m a^{-1}. Vegetative evidence indicates most of the 19th century retreat occurred before 1860. Grant and Higgins (1913, p. 29) stated that "The presence of a mature alder thicket close to the ice front indicates that the glacier is now very near its maximum advance in a period of perhaps 50 or more years."

Tarr and Martin (1914) presented evidence that the position of the terminus was relatively unchanged between 1899 and 1910. In 1910, Martin observed as much as 230 m of change in the position of the terminus, an indication that a strong advance was underway. In one location, Tarr and Martin (1914) reported that trees at least 33 years old were being overtopped. When Field (1932) visited Yale Glacier in 1931, he found evidence of a pre-1931 advance of as much as 400 m. Following a short period of retreat, Field (1937) noted that, by 1935, a maximum advance of about 60 m had occurred. On the basis of 1949 photography, Baird and Field (1951) concluded that, aside from thinning, the position of the terminus at most locations was similar to its 1935 position. The major exception was a recession of several hundred meters along the southeastern margin of the terminus.

In addition to using the 1949 photography, Field (1975b) compared aerial photographs of Yale Glacier made by the USAF in 1950, 1954, and 1957 and by the AGS in 1957 with field observations and photographs made by a group from Brigham Young University in 1961 and by the AGS in 1964, 1966, and 1971. He concluded that, between 1950 and 1957, the southern part of the glacier's terminus advanced from 50 to150 m, so that "At the southeastern end, the 1957 terminus was within 50 m of its 1899 position" (Field, 1975b, p. 385). Retreat began quickly; between 200 and 450 m of retreat occurred between 1957 and 1961. By 1966, the terminus had retreated about 1,350 m, yielding an average annual retreat rate of about 150 m a^{-1}. Rapid retreat continued, and, by 1971, the terminus had retreated approximately 2.5 km from its 1957 location. By 1964, retreat had exposed a 2-km-long bedrock island, parts of which had been exposed previously as bedrock ledges in the 1899–1910 period.

Field (1975b) commented that, in 1964, the height of the iceface was measured at 112 m. In all of his observations, which at that time spanned a period approaching 40 years, he had not seen as high an ice face (nor has the author in more than 35 years of observations).

From an analysis of the same photographs and field observations, Field (1975b) reported that the recession of the northwestern part of the terminus was significantly slower than that of the active tidewater southeastern part; between 0.5 and 1.0 km of retreat occurred between 1910 and 1961. This part of the terminus entered tidewater until the 1940s. Between the 1940s and 1957, much of this margin retreated onto land. However, the USGS Anchorage A–2, 1:63,360-scale topographic quadrangle map (1960) (appendix B), which was based on 1957 photography, still shows the presence of a tidewater embayment on the northwestern margin of Yale Glacier.

Sturm and others (1991) summarized historic data on the retreat of Yale Glacier and provided new information about its retreat between 1974 and 1990. Their findings are based on aerial photography acquired in 1987, 1989, and 1990 and on a Landsat image acquired in 1985. They determined that, between 1974 and 1990, the southern side of the glacier retreated 2,070 m, and the northern side retreated 1,640 m. Hence, maximum retreat rates ranged from 103 to129 m a^{-1}. Specifically, for the southern side, they measured 1,380 m of retreat between 1974 and 1978 (an average retreat of 345 m a^{-1}), 260 m of retreat between 1978 and 1985 (an average retreat of 37 m a^{-1}), 320 m of retreat between 1985 and 1987 (an average retreat of 160 m a^{-1}), and 120 m of retreat between 1987 and 1990 (an average retreat of 40 m a^{-1}). For the northern side, they measured 920 m of retreat between 1974 and 1978 (an average retreat of 230 m a^{-1}), 280 m of retreat between 1978 and 1985 (an average retreat of 40 m a^{-1}), 440 m of retreat between 1985 and 1987 (an average retreat of 220 m a^{-1}), and no retreat between 1987 and 1990. Hence, at times, the retreat of the land-based northern part of the terminus exceeded the rate of retreat of the tidewater southern terminus.

The author has observed and photographed the glacier more than a dozen times between 1974 and 2004. During that period, the southeastern tidewater terminus has retreated between 2.7 and 3.1 km, and the northwestern grounded part of the terminus has retreated a maximum of about 2 km. This retreat has been accompanied by significant thinning of the entire terminus area. During the 26-year period between 1974 and 2000, including observations on 15 August 1978 (fig. 258), 13 August 1994 (oblique aerial photograph no. 94VE–88 by Austin Post, USGS), and 17 August 1999 (BLM photograph no. Yale GL.R3, FR–186), Yale Glacier retreated at an average annual rate that ranged between 74 m a^{-1} (western side) and 110 m a^{-1} (eastern side).

In the early 1990s (Viens, 1995) (table 2), Yale Glacier had a length of 32.2 km, an area of 194 km^2, a width at its face of 1.0 km, and an AAR of 0.79. Its accumulation area was approximately 153 km^2, and its ablation area was approximately 41 km^2.

Unnamed Glaciers of Yale Arm.—On the southern side of Yale Arm opposite Dora Keen Ridge, several unnamed west- to north-flowing retreating hanging glaciers descend from the flanks of 1,687-m-high Mount Castner. Two unnamed glaciers were pre-20th century tributaries to Yale Glacier. A third unnamed glacier was still connected to Yale Glacier in 2001. A fourth unnamed glacier has retreated more than 1 km since being photographed in 1957.

Glaciers of Harvard Arm

Harvard Glacier.—Harvard Glacier, located at the head of Harvard Arm, is the second largest and second longest glacier in Prince William Sound. It is an active calving tidewater glacier with a face that is approximately 2.5 km wide. It is fed by more than 20 major tributaries that originate in snow- and ice fields on the flanks of some of the highest peaks of the Chugach Mountains, including 4,017-m-high Mount Marcus Baker, many unnamed summits, Mount Thor, Mount Gilbert Lewis, and Mount Elusive, all having summits higher than 3,500 m. In 2004, the terminus of Harvard Glacier, which has advanced nearly 3 km since first being photographed in the late 19th century, was advancing. Harvard Glacier has four present or former named tributaries: Downer and Lowell Glaciers on its eastern side and Radcliffe and Elliot Glaciers on its western side.

Like Yale Glacier, Harvard Glacier was first mapped from a distance by Applegate in 1887 (Davidson, 1904). Martin (1913) stated that this map and earlier maps are not detailed enough to determine pre-1887 changes. During July 1899, the glacier was observed by the Harriman Alaska Expedition and mapped again by Gannett (Gannett, 1899; Gilbert, 1904). Detailed investigations were performed in 1905 and 1909 by Grant and Higgins (1910, 1911a, b, 1913), and by Martin (1913), and Bean (1911) in 1910. Between 1899 and 1905, the glacier changed very little. However, during the next 4 years, an advance occurred that Grant and Higgins (1911a) reported as being from 400 to 800 m but that Martin (1913) reported as being only about 200 m. Between 1909 and 1910, the glacier advanced another 30 to 45 m. The contour map of upper College Fiord published by Tarr and Martin (1914) shows the relationship of Harvard Glacier and its southern tributaries (Downer, Lowell, Eliot, and Radcliffe Glaciers) and Baltimore Glacier.

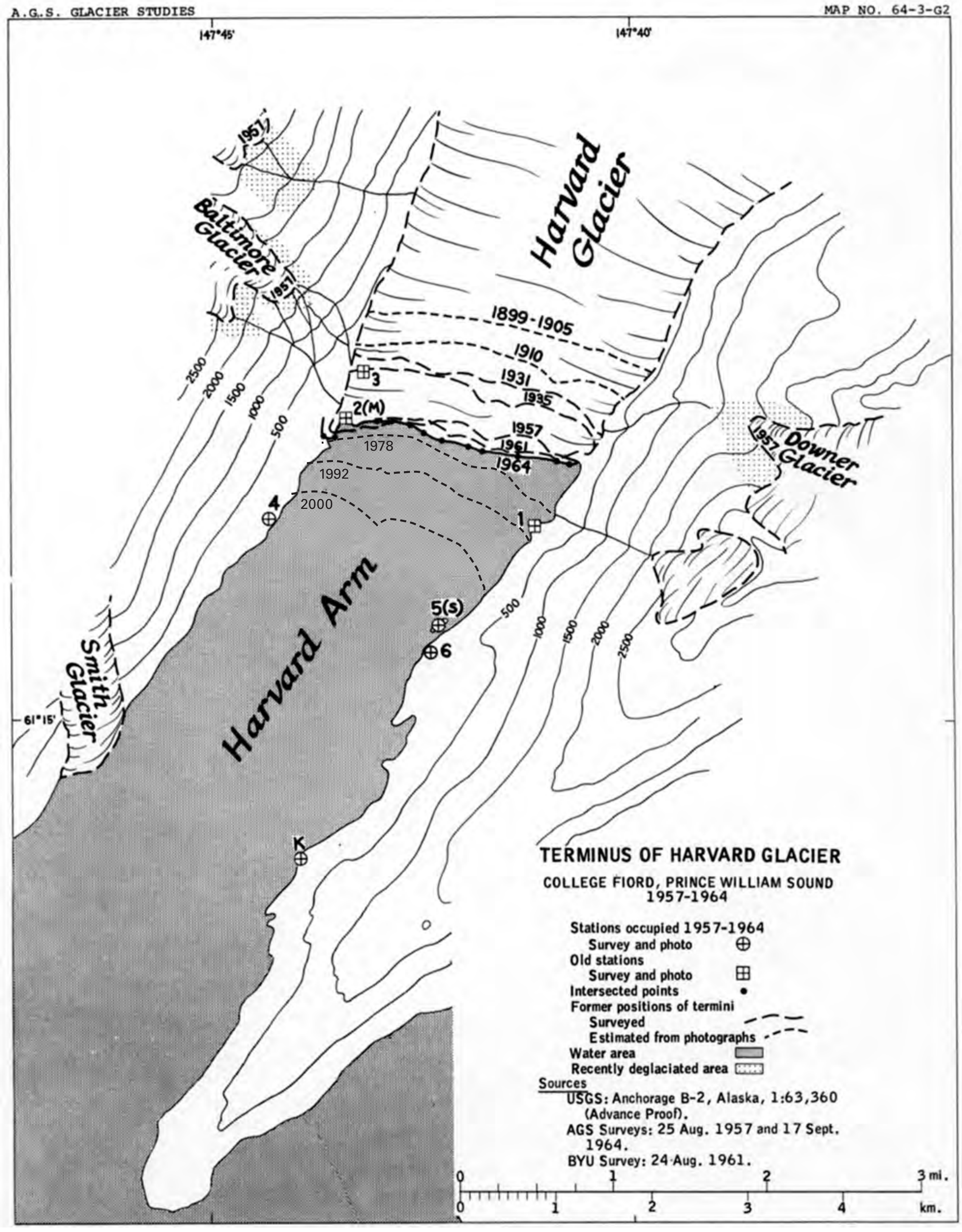
Dora Keen visited and mapped the glacier's terminus position in 1914 and reported that the eastern side of the margin showed some recession, although the western side continued to advance (Keen, 1915a). Her observations were made as she attempted to be the first woman to traverse the western Chugach Mountains from south to north. Her route over the lower 26 km of Harvard Glacier (Keen, 1915b) permitted her to map and observe Harvard Glacier and many of its larger tributaries.

Field monitored changes in Harvard Glacier for more than 50 years. A map that he prepared showing positions of its terminus between 1899 and 1964 has been updated by the author to 2000 (fig. 259). Field first visited Harvard

Figure 258.*—View of the retreating Yale Glacier on 15 August 1978. Northeast-looking view of ice extending to the right (east) edge of the photograph and newly emergent bedrock to the left of center. Photograph by Bruce F. Molnia, U.S. Geological Survey. A larger version of this figure is available online.*

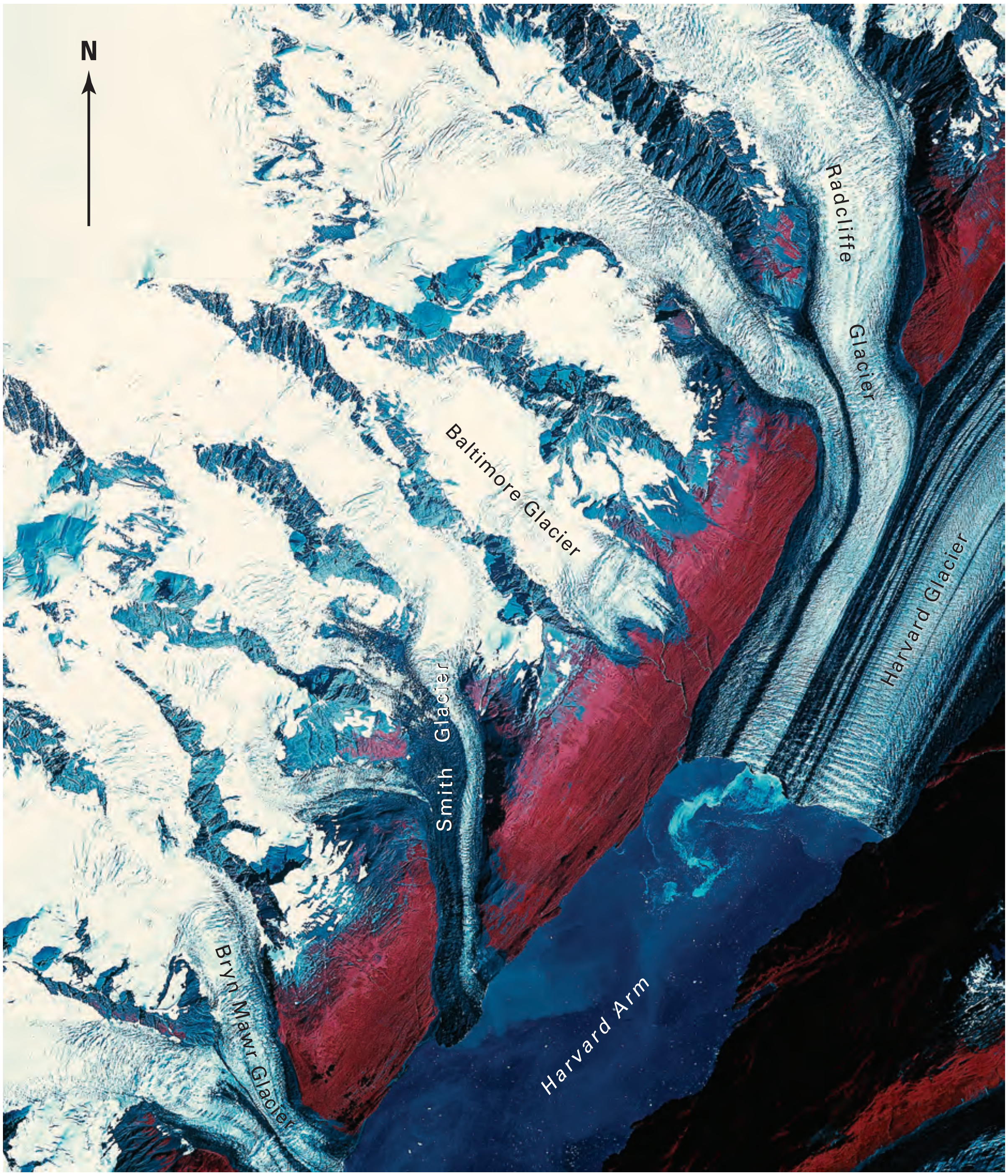
► ***Figure 259.****—Map of Harvard Glacier showing fluctuations in the position of its terminus from 1899 to 2000. Base map is derived from AGS Glacier Studies Map no. 64–3–G2 (Field, 1965). Year 1978, 1992, and 2000 positions (approximate) are based on observations made by the author. Positions of the termini of Smith Glacier and several other glaciers are also shown.*

Glacier in 1931 (Field, 1932, p. 378–379) and determined that, in comparison with its 1905 position, the terminus had advanced "at least 2000 feet (~610 m) at the eastern margin and somewhat less on the western." Field (1937) visited the glacier again in 1937 and determined that, between 1931 and 1937, the glacier had advanced another 30 to 60 m. A decade later and again in 1949, D.M. Brown (1952) visited the glacier and documented a continuing advance, although he did not quantify the amount of change.



Field (1975b) compared aerial photographs of the glacier made by the USAF in 1950, 1954, and 1957, and the USN in 1957, the C&GS in 1964, and Austin Post between 1960 and 1971 with field observations and photographs made by a group from Brigham Young University in 1961 and the AGS in 1957, 1964, 1966, and 1971. He concluded that, between 1899 and 1971, the terminus of Harvard Glacier advanced about 1.4 km and stated that this location is "its most advanced position since at least the seventeenth century" (Field, 1975b, v. 2, p. 388). Field based this conclusion on Cooper's (1942) observation that, in 1935, the advancing terminus of the glacier was pushing into a forest of spruce trees (*Picea* sp.) having as many as 246 growth rings. Since then, the glacier has continued to advance. Additional aerial photographs have been acquired by many agencies, including the AHAP Program on 25 August 1978 (fig. 260) and the BLM in 1999 (see BLM vertical aerial photograph no. Harvard GL.R3, FR–208 acquired on 17 August 1999).

Figure 260.—*25 August 1978 AHAP false-color infrared vertical aerial photograph of upper Harvard Arm and the Harvard Glacier. AHAP photograph no. L107F7037 from the GeoData Center, Geophysical Institute, University of Alaska, Fairbanks, Alaska.*



Sturm and others (1991) summarized historical data on the changes in Harvard Glacier and provided new information about its advance during the second half of the 20th century. These new findings are based on data presented by Brown and others (1982) and Meier and Post (1987). Sturm and others (1991) determined that, between 1899 and 1989, the western side of the glacier advanced 1,520 m; the eastern side advanced 2,056 m. Hence, maximum long-term advance rates ranged from 16.9 to 22.9 m a^{-1}. Specifically, for the western side, there was no change between 1899 and 1905, 270 m of advance between 1905 and 1910 (an average advance of 54 m a^{-1}), 300 m of advance between 1910 and 1931 (an average advance of 14 m a^{-1}), 100 m of advance between 1931 and 1935 (an average advance of 25 m a^{-1}), 490 m of advance between 1935 and 1957 (an average advance of 22 m a^{-1}), 170 m of advance between 1957 and 1961 (an average advance of 43 m a^{-1}), 40 m of retreat between 1961 and 1964 (an average retreat of 13 m a^{-1}), 80 m of advance between 1964 and 1976 (an average advance of 7 m a^{-1}), no change between 1976 and 1978, and 150 m of advance between 1978 and 1989 (an average advance of 14 m a^{-1}). For the eastern side, there was no change between 1899 and 1905, 370 m of advance between 1905 and 1910 (an average advance of 74 m a^{-1}), 240 m of advance between 1910 and 1931 (an average advance of 11 m a^{-1}), 160 m of advance between 1931 and 1935 (an average advance of 40 m a^{-1}), 130 m of advance between 1935 and 1957 (an average advance of 6 m a^{-1}), 320 m of advance between 1957 and 1961 (an average advance of 80 m a^{-1}), 140 m of advance between 1961 and 1964 (an average retreat of 47 m a^{-1}), 250 m of advance between 1964 and 1976 (an average advance of 21 m a^{-1}), no change between 1976 and 1978, and 446 m of advance between 1978 and 1989 (an average advance of 41 m a^{-1}).

The author has observed and photographed Harvard Glacier more than a dozen times between 1974, 14 August 1978 (fig. 261*A*), 31 August 2002 (fig. 261*B*), and 2004. During that period, the tidewater terminus has advanced about 1.0 km. This advance has been accompanied by some thinning along the entire margin of the terminus area. Hence, during the 27-year

Figure 261.—*Two photographs showing details of the advancing Harvard Glacier.* **A**, *14 August 1978 north-looking view shows the advancing terminus of Harvard Glacier and its two primary northwestern tributaries, Radcliffe and Elliot Glaciers.* **B**, *31 August 2000 south-looking view from a point approximately 10 km north of the terminus of Harvard Glacier shows that the volume of ice delivered by the entering Radcliffe Glacier is sufficient to deflect the central medial moraine complex of Harvard Glacier to the east. Photographs by Bruce F. Molnia, U.S. Geological Survey. A larger version of* B *is available online.*

period between 1974 and 2001, the glacier has advanced at an average annual rate of approximately 37 m a^{-1}. In the early 1990s (Viens, 1995) (tables 2, 3), Harvard Glacier had a length of 40 km, an area of 524 km^2, a width at its face of 2.4 km, and an AAR of 0.81. Its accumulation area was 423 km^2, and its ablation area was 101 km^2.

Radcliffe Glacier.—Radcliffe Glacier (fig. 260) is a primary tributary of Harvard Glacier that joins it from the west about 6 km above its terminus. It is composed of two principle east-flowing tributaries that coalesce about 2 km above its juncture with Harvard Glacier. When its ice was combined with the main trunk of Harvard Glacier, it made up approximately one-third of the width of the terminus in 2000. Since first being observed in the later part of the 19th century, Radcliffe Glacier has always been connected to Harvard Glacier.

Gilbert (1904, p. 84) described the 1899 relationship of Radcliffe Glacier to Harvard Glacier as follows: "The Radcliffe joins the Harvard so close to the water front that it does not become fully merged with the greater stream, but merely coalesces at one edge on its way to the sea. A conspicuous medial moraine of the Radcliffe maintains its high declivity quite to the water's edge, and the cliff where the Radcliffe ends is notably lower than the confluent cliff along the front of the Harvard."

The 1910 position of Radcliffe Glacier and its relationship to Harvard Glacier are shown on Tarr and Martin's (1914, pl. CXXVI, ff p. 320) topographic map of glaciers in College Fiord. When Martin observed Radcliffe in 1910, it occupied a larger part of the face of Harvard Glacier than it did at the beginning of the 21st century. Tarr and Martin (1914, p. 298) stated that, "The position of the medial moraine formed by the north lateral moraine of the Radcliffe Glacier, indicates that in 1910 the Radcliffe was almost as strong as the main Harvard Glacier above the junction; for this medial moraine reaches the terminal cliff of Harvard Glacier nearly in the middle, showing that at least in the upper layers the ice from Radcliffe Glacier compresses the Harvard stream to half of its normal width." When the author observed the glacier in September 2000, a recently exposed bare area along its northern margin was evidence of minor thinning. Otherwise, Radcliffe Glacier appeared to be an active and healthy glacier.

Baltimore Glacier.—Baltimore Glacier (fig. 260), a small retreating hanging glacier on the western side of Harvard Arm, just south of Radcliffe Glacier, was named by Grant and Higgins in 1908 for the Woman's College of Baltimore, now Goucher College. Its 1910 position is shown on Tarr and Martin's (1914, pl. CXXVI, ff p. 320) topographic map of glaciers in College Fiord. When Tarr and Martin observed it in July 1910, they described it as "a clean white ice mass, with no moraine except small medials near the terminus. Half a mile from the end it bifurcates, the north lobe being larger and ending lower (1,004 feet) than the south lobe. A barren zone about its terminus indicates that the glacier is less extensive than it was a few years ago...." (Tarr and Martin, 1914, p. 300).

Field (1975b) reported that a significant amount of retreat occurred during the period through 1931. However, between 1931 and 1935 (Field, 1937), the glacier advanced. From 1935 through 1938, it retreated. This retreat was followed by an advance between 1938 and 1941. Little change occurred through 1954. The glacier resumed its retreat through 1964. Field (1975b) reported no appreciable change between 1964 and 1971 and concluded that, during the 47-year-period 1910 through 1957, the glacier retreated between 500 and 600 m and thinned between 120 and 150 m. The average annual rate of retreat ranged from 10.6 to 12.7 m a^{-1}, whereas the average rate of thinning ranged from 2.6–3.2 m a^{-1}.

Observations by the author were initiated in 1974 and have continued through 2004. During that 30-year-period, the terminus separated into a number of fingerlike projections, retreated as much as 600 m, and thinned appreciably.

Smith Glacier.—Smith Glacier, an iceberg-calving tidewater glacier located about 2.5 km south of Harvard Glacier, has a length of 10 km (Field, 1975b, p. 471) (fig. 260). It was first described by the Harriman Alaska Expedition in 1899 (Gannett, 1899; Gilbert, 1904, p. 86), as follows: "Smith Glacier reaches the fiord three or four miles from the Radcliffe, and is of the same order of magnitude. Fed by several tributaries among the crests of the range, it gathers in a high mountain valley and then descends in magnificent cascades down the mountain front to the sea. In the last part of its course it has scarcely any valley, the outer surface of the ice being flush with the face of the mountain; and there is no flattening of its profile as it reaches the water."

Through 1909, each margin of the glacier was marked by a narrow barren zone (Tarr and Martin, 1914). However, between 1909 and 1910, the glacier began to advance and widen. By July 1910, the barren zones had been overridden by advancing ice, which was pushing into the adjacent alders (*Alnus* sp.). The 1910 position is shown on Tarr and Martin's (1914, pl. CXXVI, ff p. 320) topographic map of glaciers in College Fiord. Tarr and Martin described this advance as follows: "Along the advancing margin the alders were being destroyed in three ways,—by actual overriding of the spreading glacier, by stream encroachment, and by ice-block avalanches which rolled some distance out into the forest knocking down and breaking off shrubs and removing their bark 6 or 8 feet above the ground (Tarr and Martin, 1914, pl. CXVI, ff p. 320) [Editors' note: a ground photograph taken on 22 July 1910]. It was impossible to tell exactly how much the tidal terminus of the glacier had moved forward since 1909 but there was undoubtedly several hundred feet of advance, accompanying the spreading on the north and south margins..." Tarr and Martin (1914, p. 301–302) observed "a flat tidal terminus extending a short distance out into the fiord, where Gilbert says there was none in 1899. The extreme southern edge of ice cliff was a black, crevassed precipice, the lateral moraine of 1899 having been pushed forward into the sea."

Smith Glacier was next observed by Keen (1915a) in 1914. She reported no change from 1910. Field (1932) visited the glacier in 1931, and noted a significant amount of retreat. Upon his return in 1935, Field (1937) noted an increase in terminus activity. He later reported that, except for several minor fluctuations in the position of the terminus, "the glacier did not change appreciably from 1931 to 1971" (Field, 1975b, v. 2, p. 392). Observations by the author were made first in 1974, repeated on 15 August 1978 (fig. 262*A*) and 8 August 2000 (fig. 262*B*), and have continued through 2004. During that 30-year-period (1974–2004), the terminus has maintained a calving face in the fjord, but the barren zone along the margins of the glacier has increased significantly in width. The extent of the terminus in 2004 was larger than that of the terminus in 1910 and was at least 150 m more advanced than the terminus in 1899.

In the early 1990s (Viens, 1995) (tables 2, 3), Smith Glacier had a length of 9.7 km, an area of 20 km^2, a width at its face of 400 m, and an AAR of 0.81. Its accumulation area was 16 km^2, and its ablation area was 4 km^2.

Bryn Mawr Glacier.—Bryn Mawr Glacier, an iceberg-calving tidewater glacier located about 2.5 km south of Smith Glacier, is 8 km long (Field, 1975b, p. 471) (fig. 260); it is formed by the joining of two eastward-flowing tributaries about equal in size. When the Harriman Alaska Expedition visited it in 1899, Bryn Mawr's terminus reached tidewater. Gilbert (1904, p. 88) noted that it is somewhat larger than Smith Glacier. He described its tidewater terminus as follows: "As tide is reached, there is a tendency to flatten the profile, and the central portion of the stream becomes nearly or quite horizontal for a few hundred feet before breaking off in the terminal cliff."

According to Grant and Higgins (1910, 1911a, b), little change occurred through 1905. By 1909, they noted that the glacier had advanced as much as

Figure 262.—Two photographs show changes of Smith Glacier during the last quarter of the 20th century and the beginning of the 21st century. Although Smith Glacier continues to maintain its tidewater terminus, it is clearly thinning and retreating. **A**, *15 August 1978 northwest-looking view of most of Smith Glacier. A trimline on the left and newly emergent bedrock at several locations document the recent thinning of the glacier. Its tidewater calving terminus is located close to its 20th century maximum position.* **B**, *8 August 2000 northwest-looking oblique aerial photograph shows nearly all of Smith Glacier (left half of photograph). Emergent bedrock lies along much of the lower lateral margins of both sides of the glacier and at the southern side of its terminus. Note the difference in the position of the southern versus the northern part of the terminus. The northern part of the terminus has retreated several hundred meters since the 1970s; the southern part shows little change. The tidewater terminus of Harvard Glacier is on the right; Radcliffe Glacier is the large tributary glacier that merges with Harvard Glacier. Photographs by Bruce F. Molnia, U.S. Geological Survey. A larger version of* B *is available online.*

150 m. Tarr and Martin (1914, p. 302–303) reported that, when they visited in 1910, the glacier was advancing: “On each side of the glacier a small stream emerges from the ice, and at the time of our visit the borders of the glacier were encroaching on these stream courses. All along its northern margin the Bryn Mawr Glacier was advancing into the forest, where it was killing spruces up to 5 inches in diameter, suggesting that the glacier had not been so large for a half century or thereabouts.” Bryn Mawr’s 1910 position is shown on Tarr and Martin’s (1914) topographic map of glaciers in College Fiord.

By 1931, Field (1932) reported that the glacier had retreated more than 450 m. Upon his return in 1935, he noted that the glacier had advanced and had spread laterally as much as 60 m (Field, 1937). Brown (1952, p. 43) reported that, by 1949, the glacier had readvanced and recovered the area lost before 1931: “Laterally, the tongue has elongated southward over six hundred feet and is tearing up trees with greater than fifty annual rings.” Field (1975b) again visited the glacier in 1957 as part of an AGS field party and measured about 100 m of retreat from the late 1940s maximum position. According to Field (1975b, v. 2, p. 393), the late 1940s advance “was the greatest on the southern margin since at least the 1860s.” Retreat continued until sometime between 1961 and 1964 (AGS Glacier Studies Map No. 64–3–G8) (Field, 1965). Between 1964 and 1971, the last observation reported by Field (1975b), the glacier underwent a small readvance. In summary, during the 72-year period between the observations of the Harriman Alaska Expedition in 1899 and 1971, the position of the terminus of Bryn Mawr Glacier fluctuated as much as 575 m, reaching maxima in 1910–14 and again in 1949. By 1971, the glacier was again retreating, although it was still less than 50 m from its 1949 position.

Observations by the author were made first in 1974 and repeated on 8 August 2000 (fig. 263) and have continued through 2004. During the 30-year-period between 1974 and 2004, the terminus first advanced and then retreated. It has continued to maintain a calving face in the fjord, although bedrock is beginning to crop out at sea level along the face of the glacier. The position of the terminus in 2004 is less advanced than in 1978, and at least 250 m more retreated than that of the glacier in 1910. The position of the 1978 terminus is approximately the same as it was in 1949.

In the early 1990s (Viens, 1995)(tables 2, 3), Bryn Mawr Glacier had a length of 7.6 km, an area of 26 km^2, a width at its face of 900 m, and an AAR of 0.84; its accumulation area was 22 km^2, and its ablation area was 4 km^2.

Vassar Glacier.—Vassar Glacier was a tidewater glacier when it was first visited by the Harriman Alaska Expedition in 1899, but it is no longer. Gilbert (1904, p. 88) described it as a cascading glacier similar to Smith and Bryn Mawr Glaciers "but of smaller size and less direct in its course. It is cumbered, especially in its lower part, by rock debris, and close inspection was necessary to determine that it was actually tidal." Early 20th century observations were made in 1905 and 1909 by Grant and Higgins (1910, 1911a, b, 1913) and by Martin in 1910 (Martin, 1913). Between 1899 and the summer of 1909, the glacier was ringed by a barren zone and showed very little change. However, by 21 July 1910, an advance was underway. Tarr and Martin (1914, p. 304) reported that, "At the time of our visit the glacier touched tidewater along the whole portion of the front between the flanking alluvial fans, but with a low, sloping moraine-veneered margin along the southern half, and with a low, dirty, nearly vertical cliff in the northern half... portions of the barren zone were covered. That on the north side near sea level was not completely overridden by July 21, 1910, but higher on the fiord wall it was almost covered... The southern edge of the glacier at sea level was also obviously advancing... Near sea level the glacier extended right up to the forest which included mature spruces." The amount of the 1910 advance and its duration are unknown but the advance is estimated to be about 30 to 45 m. The contour map of upper College Fiord, published by Tarr and Martin (1914, pl. CXXVI, ff p. 320), shows the 1910 position and extent of Vassar Glacier.

Vassar Glacier was next observed by Keen (1915a) in 1914. She reported that the glacier was continuing to advance but did not quantify the amount of change from its 1910 position. Field (1932, 1937) visited the glacier in 1931 and again in 1935. He stated that changes from 1910 were not appreciable, but he was unable to make any quantitative comparison of change. However, he noted that between 1931 to 1935, there was evidence of shrinkage in the terminus area and of lateral expansion above the terminus.

Field (1975b) stated that the glacier was photographed in 1941 by the U.S. Army Air Forces and numerous additional times through 1968. He reported that during this period, "the moraine-covered terminal lobe has slowly diminished in volume and has shrunk in height and area since the turn of the century. This shrinkage is especially noticeable in a comparison of vertical aerial photographs taken between 1941 and 1968. Alders are steadily encroaching along the margins and on the moraine-covered, apparently stagnant, ice of the terminal lobe" (Field, 1975b, v. 2, p. 396).

Observations by the author were begun in 1974 and have continued through 2004. When the author observed Vassar Glacier on 15 June 1978 (fig. 264), vegetation was well established near the shoreline on the stagnant debris-covered terminus. It was impossible to tell if any ice remained in the terminus area. During the next 26 years, the width of the cascading lower portion of the glacier diminished by about 30 percent. In 2004, retreat and thinning were apparent along all of the bare-ice margins of the glacier.

Figure 263.*—8 August 2000 oblique aerial photograph shows bare bedrock along each lateral margin and the retreated central part of the terminus of Bryn Mawr Glacier. Photograph by Bruce F. Molnia, U.S. Geological Survey. A larger version of this figure is available online.*

Figure 264.*—15 June 1978 photograph shows the debris-covered terminus of Vassar Glacier and a large area of exposed bedrock in the middle of the glacier. Vassar Glacier shows evidence that it has thinned along each lateral margin and retreated, especially along the south side of its terminus. Photograph by Bruce F. Molnia, U.S. Geological Survey. A larger version of this figure is available online.*

Wellesley Glacier.—Wellesley Glacier, a tidewater glacier (fig. 41) situated at the head of a small inlet, has a terminus that is about1 km wide. The inlet has water depths of 6 to 39 m. The shores of the inlet are composed of an undated older breached terminal moraine. Wellesley Glacier, then surrounded by a large barren zone, was first described by the Harriman Alaska Expedition in 1899 (Harriman, 1902). Gilbert (1904, p. 88) characterized it as flowing "with gentle grade through a mountain trough joining the fiord at right angles, and then cascades to the sea, into which it plunges without notable modification of profile. Beyond it are small glaciers occupying alcoves on the mountain front but ending far above the water."

When Martin next visited in 1910 (Tarr and Martin,1914), the position of the terminus was essentially unchanged from 1899. Its location is shown on their contour map of upper College Fiord. However, Tarr and Martin (1914, p. 305) presented several lines of evidence that an advance was underway: "Although at the time of our visit there was a very much larger barren zone around the glacier terminus than around any other ice tongue in College Fiord, it was then actively advancing, and the northern and southern margins had partly covered the lateral barren zone previously exposed."

The glacier was next observed by Keen (1915a) in 1914, who reported that a slight recession had occurred from its 1910 position. Field (1932) visited the glacier in 1931 and found that the 1931 terminus position was advanced beyond the 1910 position. However, he described evidence suggesting that the glacier had first retreated before its advance. Between 1931 and 1935 (Field, 1937), there was evidence of lateral expansion above the terminus, but no evidence of change in the terminus region. Between 1935 and 1968, the glacier changed very little; it experienced a small advance between 1968 and 1971 (Field, 1975b).

The author began his observations in 1974 and continued them through 2004. When he first observed the southern margin of the glacier, it was surrounded by a large vegetation-free, barren bedrock area; the northern margin was composed of a broad area of stagnant, debris-covered ice. By 8 August 2000 (fig. 265), retreat and thinning were apparent along all margins of the glacier, and bedrock was exposed at several locations along its face.

In the early 1990s (Viens, 1995) (tables 2, 3), Wellesley Glacier had a length of 7.2 km, an area of 16 km^2, a width at its face of 550 m, and an AAR of 0.78; its accumulation area was 12 km^2, and its ablation area was 4 km^2.

Barnard Glacier.—Barnard Glacier is one of the "small glaciers occupying alcoves on the mountain front but ending far above the water" mentioned by Gilbert (1904, p. 88). This hanging glacier had an extensive barren zone around its margin when first described, but an advance that began after 1899 and before 1910 reduced the size of the zone. Tarr and Martin (1914, p. 306) stated that Barnard Glacier, with two ice tongues, "has not descended much farther toward the fiord for a century or more; but a barren zone between the ice and forest, present in 1899 as well as in 1910, proves that it has been retreating in recent years. Between 1899 and 1910 there was an advance of the south lobe down the lip of the hanging valley, and a slight advance of the north lobe.... We are inclined to believe that the advance was still going on in 1910." Keen (1915a) noted no change of the margin's position in 1914. Between 1914 and 1937, when the glacier was photographed by Bradford Washburn, it had retreated. Retreat continued through Field's (1975b) last data set, which was based on 1971 aerial photography. Field (1975b, p. 397) described this retreat as a "very considerable net recession of the terminus during that interval."

The author began his observations in 1974 and continued them through 2004. During this 30-year period, the glacier has retreated and thinned significantly. The author's initial observation of the perimeter of the glacier indicated that it was surrounded by a large vegetation-free, barren bedrock area and that an abandoned lateral moraine was developing a dense vegetative cover. In 2004, retreat and thinning were continuing along all margins of the glacier.

Figure 265.*—8 August 2000 west-looking oblique aerial photograph shows the calving, retreating terminus of Wellesley Glacier. The debris-covered northern part of the terminus projects beyond the bare ice. Barren zones are visible along both sides of the glacier. Photograph by Bruce F. Molnia, U.S. Geological Survey. A larger version of this figure is available online.*

Holyoke Glacier.—Holyoke Glacier heads in a large cirque. In 1910, when the glacier was first described by Tarr and Martin (1914), it was fed by two tributaries and surrounded by a large barren zone. At that time, Tarr and Martin (1914, p. 306) reported that "No distinct signs of recent advance were seen." By 1914, Keen (1915a) noted that an advance was underway. Field (1975b) reported that recession was underway in 1935 and that observations in 1957 and again in 1966 indicated that it was continuing. He reported that the glacier advanced more between 1966 and 1971. The author's observations were initiated in 1974 and continued through 2004. During this 30-year period of observation, the glacier has retreated and thinned significantly. When the author first observed it, he noted no evidence of the advance underway in 1971. By 2001, the perimeter of the glacier was surrounded by a large vegetation-free, barren bedrock area and the glacier had separated into two distinct ice masses. Retreat and thinning were apparent along all margins of the glacier.

Barry Arm and Harriman Fiord

Barry Arm and Harriman Fiord (fig. 266) are connected fjords that comprise the northwestern extension of Port Wells. Between them, they host five tidewater glaciers—Barry, Coxe, Cascade, Surprise, and Harriman Glaciers (fig. 41)—and several former tidewater glaciers, including Serpentine (tables 2, 3) (Viens, 1995) and Cataract Glaciers. With the exception of Harriman Glacier, all of the glaciers in this area have long histories of 20th and early 21st century retreat. Before and including the 1899 visit by the Harriman Alaska Expedition, Barry Arm and the glaciers at its head were mapped with varying degrees of accuracy by Vancouver in 1794 (Vancouver, 1798), Applegate in 1887 (Davidson, 1904), Glenn (1899) in 1898, Castner (1899) in 1898, Mendenhall (1900) in 1898, and the Harriman Alaska Expedition in 1899 (Gannett, 1899; Gilbert, 1904; Tarr and Martin, 1914). Formally named Barry Arm Port Wells by Glenn in 1898 for Col. Thomas Barry, Assistant Adjutant General of the U.S. Army, Barry Arm is a 15-km-long fjord that extends from Barry Glacier at its northern end to Pakenham Point at its mouth.

When they were mapped at the end of the 19th century, three connected glaciers (from west to east)—Cascade, Barry, and Coxe Glaciers—were located at the head of Barry Arm. In 1899, when the Harriman Alaska Expedition explored Barry Arm, Barry Glacier was retreating from a recent, although undated, maximum position. At that time, its terminus extended

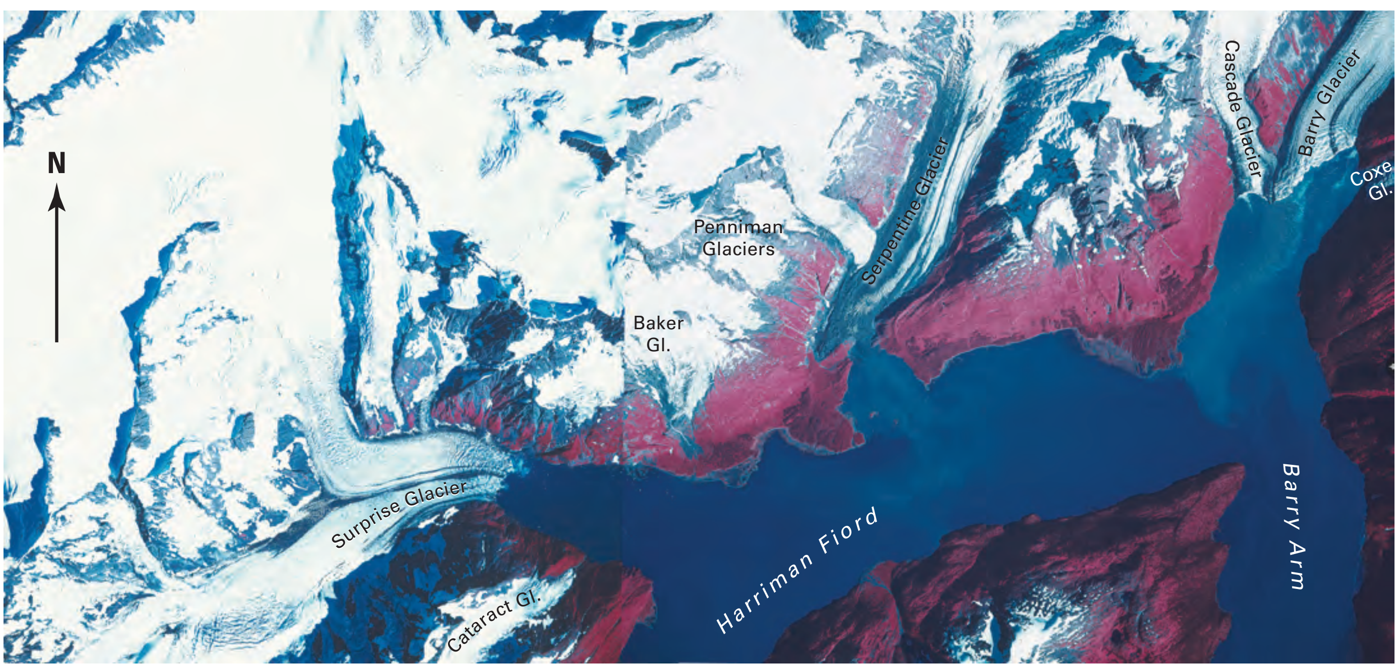


***Figure 266.**—24 August 1978 AHAP false-color infrared vertical aerial photographic mosaic of lower Harriman Fiord and Barry Arm. The positions of the termini of Cascade, Barry, Coxe, Surprise, Serpentine, Penniman, Baker, and Cataract Glaciers and several unnamed glaciers indicate that all have retreated since first observed in 1899. AHAP photograph nos. L109F6595 and L109F6597 from the GeoData Center, Geophysical Institute, University of Alaska, Fairbanks, Alaska.*

a significant distance down the fjord, nearly completely closing what then was the unknown entrance to Harriman Fiord. When the *SS Elder* reached a location just to the south of the point now known as Point Doran, which juts into Barry Arm from the southwest, the local pilot felt it was no longer safe to pass and returned control of the ship to its captain, Peter Doran. The following description, presented by Muir in 1911 (Lethcoe, 1987, p. 86) summarized what followed:

> Then Mr. Harriman asked me if I was satisfied with what I had seen and was ready to turn back, to which I replied: "Judging from the trends of this fiord and glacier, there must be a corresponding fiord or glacier to the southward, and although the ship has probably gone as far as it is safe to go, I wish you would have a boat lowered and let me take a look around that headland into the hidden half of the landscape." "We can perhaps run the ship there," he said and immediately ordered the captain to "go ahead and try to pass between the ice wall and the headland." The passage was dangerously narrow and threatening, but gradually opened into a magnificent icy fiord about twelve miles long, stretching away to the southward. The water continuing deep, as the soundline showed, Mr. Harriman quietly ordered the captain to go right ahead up the middle of the new fiord. "Full speed, sir?" inquired the captain. "Yes, full speed ahead," The sail up this majestic fiord in the evening sunshine, picturesquely varied glaciers coming successively to view, sweeping from high snowy foundations and discharging their thundering wave-raising icebergs, was, I think, the most exciting experience of the whole trip.

The result was the discovery that would become Harriman Fiord, an inlet previously unknown to all but possibly indigenous seal hunters.

Glaciers of Barry Arm

Barry Glacier.—Tarr and Martin (1914, p. 321–322) presented the following compilation of early descriptions of Barry Glacier:

> Mendenhall relates that in 1898 Barry Glacier was more extensive than either Harvard or Yale Glaciers. Glenn ... said that it was "one of the most formidable as well as the most interesting" of the glaciers that they saw. Coming from it they saw "immense icebergs that had evidently broken off from the glacier. Many of these were from ten to twenty times as large as our boat." When they were near Yale Glacier ... a "noise caused by the falling of the immense ice floes from Barry Glacier, could be heard like the rumbling of distant thunder and which seemed to shake the mountains on either side of us." In May 1898, Castner went up Barry Arm to "within a half mile of the sea end of the great glacier." Photographs were taken and the interesting manufacture of icebergs watched. The later consisted of a breaking off and tumbling into the sea of tons of blue ice from the face of the glacier, accompanied by the roar of a Niagara, as the berg started on its ocean voyage, eventually to melt and become a part of the tides which now carried it away.

Some confusion exists about where the terminus of Barry Glacier was located during the later part of the 19th century and about its behavior. Applegate's 1887 sketch map of Port Wells shows a glacier with two tributaries filling the entire width of the fjord at a location about 10 km north of Pakenham Point but no evidence of an entrance to Harriman Fiord.

The NGS's expedition to Alaska mapped Barry Glacier in 1910 (fig. 267). An arcuate shallow moraine located north of and within 250 m of Point Doran was mapped by Grant and Higgins (1911a, fig. 7) and by Bean (1911) in 1910. It has shoaling depths as shallow as 5.5 m and spans the width of the fjord. Although its age is unknown, it is probably the "Little Ice Age" maximum terminal moraine of an expanded Barry Glacier. Whether Applegate's mapped terminus position for Barry Glacier and the location of this submarine moraine are coincident is unknown. However, given the geometry of the fjord and the distance from which he observed its head, he could have easily erred in depicting the length of the fjord and the actual 1887 location of the terminus of Barry Glacier.

When the Harriman Alaska Expedition visited Barry Glacier on 26 June 1899, it did not observe the discharge of large icebergs or hear the thunderous noises described the previous year. Gannett's 1899 map of Barry Glacier shows its terminus about 1 km north of Point Doran. Gannett (1899) used the name *Washington Glacier*, although it was not officially adopted. At the same time, C. Hart Merriam of the U.S. Biological Survey photographed the glacier, and Gilbert (1904) described a barren zone on the eastern margin of

Figure 267.—*Map of Harriman Fiord and Barry Arm showing the position of the combined termini of Cascade, Barry, and Coxe Glaciers as mapped by the National Geographic Society's 1910 Expedition (Martin, 1911; Tarr and Martin 1914). Soundings (in fathoms) were performed by E.F. Bean.*

Barry Glacier that he suggested was evidence that retreat from a recent, earlier maximum position was already underway. Grant and Paige visited Barry Arm in 1905; Grant and Higgins (1913) visited in 1910. All noted that Barry Glacier was still connected to Coxe and Cascade Glaciers but that, in the decade between 1899 and 1909, it had retreated approximately 2.6 km. The face of the combined glacier was about 75 m high in 1905 and in 1908 but had diminished to only about 37.5 m in 1909. The height of the exposed barren zone along the margin of the glacier was approximately 120 m, indicative of an average rate of retreat of about 260 m a^{-1} and an average rate of thinning of about 12 m a^{-1}. Tarr and Martin (1914) described a large ice mass that had detached from the retreating glacier and was left stranded on the eastern wall of the fjord in 1910. They described its size as being 400×200 m and extending to elevations of 75 to 85 m. The detached ice mass was located more than 3 km from the face of Barry Glacier in 1910.

A summary of Barry Glacier's behavior through 25 July 1910 was presented by Tarr and Martin (1914): (1) Before Whidbey's 8 June 1794 visit, unknown; (2) 8 June 1794 until Applegate's visit in June 1887, unknown; (3) June 1887 until Castner's visit on 6 May 1898, unknown amount of advance; (4) 6 May 1898 until the Harriman Alaska Expedition on 26 June 1899, retreat of approximately 805 to 1,210 m; (5) 26 June 1899 until Grant and Paige's visit on 20 August 1905, retreat of about 1,930 m; (6) 5 August 1905 until Grant's visit of 20 August 1908, retreat of about 650 m; (7) 11 August 1908 until Grant and Higgins' 29 June 1909 visit, retreat of about 805 m; and (8) 29 June 1909 until Martin's 25 July 1910 visit, retreat of about 150 to 300 m. Hence, in the 3,733 days (10.2 years) between 6 May 1898 and 25 July 1910, Barry Glacier retreated between 4,304 and 4,895 m (as much as 1.3 m d^{-1} or nearly 480 m a^{-1}).

Bertrand Johnson, a USGS geologist, visited Barry Arm in 1913 and again in 1914 (Johnson, 1917). He noted the continuing retreat of Barry Glacier and the separation of Coxe Glacier from Barry in 1913. Between 1910 and 1913, ongoing retreat of the eastern side of Barry Glacier amounted to about 2 km; on the western side, it was about 750 m. Along the retreating western margin, bedrock was exposed at the base of the ice for the first time. By 1914, the eastern part of the terminus had retreated another 500 m.

In the ensuing 90 years, the glacier has been observed, monitored, and photographed many times. Individuals providing information about Barry Glacier include Keen in 1914 (Keen, 1915a) and again in 1925; Field in 1931 (Field, 1932); Field and Cooper in 1935 (Field, 1937; Cooper, 1942); Robert E. Fellows of the USGS in 1943 (Field, 1975b); and Brown in 1947 and again in 1949 (Brown, 1952). Barry Glacier was photographed by Bradford Washburn in 1937 and 1938, the U.S. Army Air Forces in 1941 and 1942, the USAF in 1950 and 1957, the USN in 1957, Austin Post between 1960 and 1983, and Robert M. Krimmel between 1984 and 1990, among others. These observations document that Barry Glacier had a small advance between 1925 and 1931, advanced more in 1935 and 1937, retreated between 1938 and 1943, and changed little through 1947. A slight recession occurred during the 1950s, followed by an advance beginning about 1961. Since then, the location of the western part of Barry's terminus has remained within a few hundred meters of its 1914 location; the eastern part of the margin has retreated several hundred meters, as two oblique aerial photographs taken by the author on 8 August 1981 (fig. 268*A*) and 8 August 2000 (fig. 268*B*) show; the entire glacier has continued to thin.

Although the amount of exposed bedrock has fluctuated substantially (Lethcoe, 1987), at no time has it been reported that the glacier has retreated above sea level. The author has noted substantial exposures of bedrock along the base of the glacier several times. In September 2000, two small bedrock outcrops were noted near the central face of the terminus.

In the early 1990s (Viens, 1995) (tables 2, 3), Barry Glacier had a length of 26.5 km, an area of 95 km^2, a width at its face of 2.2 km, and an AAR of 0.74; its accumulation area was 70 km^2, and its ablation area was 25 km^2.

Coxe Glacier.—Coxe Glacier separated from the retreating Barry Glacier in 1913. Between 1913 and the late 1930s, it retreated about 2.5 km; its terminus retreated into its fjord, where it maintained a tidewater terminus (figs. 41, 266). Since then, its terminus has fluctuated: a small advance in the 1941–49 period (Brown, 1952), a period of retreat between 1952 and 1957 (Field, 1975b), and a small advance between 1979 and the late 1980s

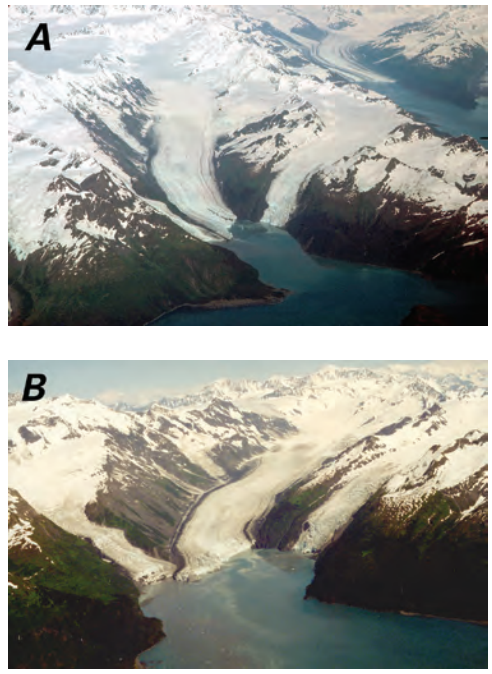


Figure 268.**—Two north-looking oblique aerial photographs showing changes in Cascade (left), Barry (center), and Coxe (right) Glaciers between 1981 and 2000.* **A, 8 August 1981 photograph shows the tidewater calving termini of all three glaciers and much of their accumulation areas. The most extended part of the combined Barry and Cascade Glaciers is the central part of their combined termini. Harvard Glacier can be seen in the distance.* **B***, 8 August 2000 photograph shows the tidewater calving termini of all three glaciers and much of their areas of accumulation. The continued retreat and thinning of Cascade Glacier has exposed bedrock along the western part of its terminus. It is thinner than Barry Glacier at their junction. The east side of Barry Glacier retreated several hundred meters since an earlier visit by the author in 1992. Coxe Glacier also thinned and retreated. Photographs by Bruce F. Molnia, U.S. Geological Survey. Larger versions of these figures are available online.*

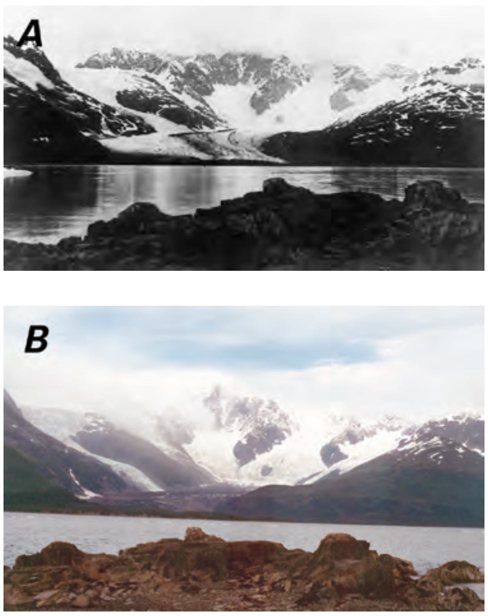


Figure 269.*—A pair of photographs taken from the east shore of Harriman Fiord show changes in Serpentine Glacier between 1899 and 2000.* **A***, 1899 Harriman Alaska Expedition photograph (Harriman, 1902) by Edward Curtis from the east shore of Harriman Fiord shows most of the terminus and source area of Serpentine Glacier. The dark area at the foot of the glacier is an older terminal moraine, dating from circa 1800, that forms much of the entrance to the inlet in front of Serpentine Glacier.* **B***, 6 September 2000 photograph from the same vantage point shows that Serpentine Glacier has thinned by approximately 70 to 100 m and has retreated approximately 2.5 km. The terminus is now debris covered and barely visible above the terminal moraine. Photograph by Bruce F. Molnia, U.S. Geological Survey. Larger versions of these figures are available online.*

(Lethcoe, 1987). During the past 17 years, the terminus has retreated slightly but maintained some contact with tidewater. During the second half of the 20th century, Coxe Glacier has thinned significantly and narrowed appreciably, producing wide bedrock exposures along both lateral margins of the glacier. When the author visited the glacier on 5 September 2000, bedrock was exposed along the base of the entire southern half of the terminus and at the northernmost edge of the terminus at the base of the glacier. In the early 1990s (Viens, 1995) (tables 2, 3), Coxe Glacier had a length of 11 km, an area of 20 km^2, a width at its face of 1.1 km, and an AAR of 0.74. Its accumulation area was 15 km^2, and its ablation area was 5 km^2.

Cascade Glacier.—Cascade Glacier is a steep south-flowing valley glacier that joins the western edge of Barry Glacier at tidewater (figs. 41, 266, 267, 268). During the last few years of the 20th century the two glaciers barely touched. At the beginning of the 20th century, Cascade Glacier was the westernmost tributary to the extended Barry Glacier. By 1913 (Johnson, 1917), Cascade Glacier had become independent from the rapidly shrinking Barry Glacier, although it still made contact with its western margin. At times, retreat of the terminus of Cascade Glacier exposes a large bedrock area just above sea level. During much of the 20th century, fluctuations in the glacier's margin have covered and uncovered this bedrock. Field (1975b) reported that bedrock exposed in the middle 1930s was covered by a minor advance that continued through about 1950 and was followed by recession through 1957. An advance that began in 1968 resulted in the tidewater terminus of Cascade Glacier pushing into Barry Glacier in 1971. In both 1981 (fig. 268) and 1992, when the author visited the eastern margin of Cascade Glacier at its junction with Barry Glacier, it was the most extended part of the terminus. At the start of the 21st century, Barry and Cascade Glaciers remained barely connected. In the early 1990s (Viens, 1995) (tables 2, 3), Cascade Glacier had a length of 8 km, an area of 15 km^2, a width at its face of 600 m, and an AAR of 0.89; its accumulation area was 14 km^2, and its ablation area was 2 km^2.

Glaciers of Harriman Fiord

Harriman Fiord contains more than a dozen glaciers. The largest and longest is Harriman Glacier at its head. With the exception of Harriman Glacier, all of the glaciers in the fjord have retreated since first being observed.

Serpentine Glacier.—In the early 1990s, Serpentine Glacier had a length of 10 km, an area of 30 km^2, and an AAR of 0.70; its accumulation area was 21 km^2 and its ablation area was 9 km^2 (Viens, 1995) (tables 2, 3). In 1899, when the glacier was first photographed by Curtis (fig. 269*A*) and described by Gilbert (1904), it was tidal, although much of its margin was fronted by a developing moraine. Gilbert (1904, p. 93) described the glacier as follows: "It is a broad stream, of low grade, fed by four or five tributaries descending steeply from amphitheaters in the encircling mountains. Though it reaches the sea, it yields few bergs, but is building a moraine barrier along most of its front...." When the author visited on 6 September 2000 (fig. 269*B*), the northern part of the terminus was surrounded by a subcircular-shaped outwash plain about 500 m in diameter; the southern part of the terminus was fronted by a vegetated ridge composed of ablation till and moraine.

Following the Harriman Alaska Expedition's visit in 1899 (Harriman, 1902), Serpentine Glacier was observed in 1905 and again in 1909 by Grant and Higgins (1910, 1911a, b, 1913), and by Martin and the NGS party, who produced a map of its terminus in 1910 (Tarr and Martin, 1914, fig. 45, p. 327). Their studies showed that, in 1910, the terminus of Serpentine Glacier was situated at the head of a 1.6-km-long inlet bounded at its junction with the main part of Harriman Fiord by a breached lobate terminal moraine approximately 2.5 km long. On the basis of dendrochronological evidence, Martin recognized that this terminal moraine consisted of two separate nested moraines that he concluded were formed by advances in the early 1800s and again in about 1870. Grant and Higgins (1913) observed little change in the terminus position from 1899 until their first visit in 1905 and about 400 m of terminus retreat between 1905 and 1909. Piecing this information together suggests that the glacier reached a maximum—perhaps its "Little Ice Age" maximum—about 1800. For the next 70 years, its position may have fluctuated, but, in 1870, its terminus was near the 1800 maximum. By 1900, the glacier had retreated about 1.2 km, with an average retreat rate of about 40 m a^{-1}. From 1900 until 1905, the position of the terminus of Serpentine Glacier was stable. Approximately 400 m of retreat followed between 1905 and 1909, with an average retreat rate of about 100 m a^{-1}. Little change was observed in 1910.

Keen (1915a) visited the glacier in 1914 and found no changes from Martin's 1910 position. Field (1932) observed the glacier in 1931 and noted a recession of about 400 m from its 1910–14 position. Field and Cooper visited in 1935 (Field, 1937) and noted a slight advance since 1931. Bradford Washburn's 1938 photograph of the terminus showed a continuation of this advance. U.S. Army Air Forces photographs taken in 1941 show the glacier retreating but close to its 1931 position. When Brown visited in 1947 and again in 1949 (Brown, 1952), Serpentine Glacier was advancing. A comparison of an aerial photograph acquired by the USAF in 1950 with the 1941 photograph shows that the terminus was between 200 and 300 m forward of the 1941 position. Between 1957, when it was photographed by the USN, and 1971, when it was photographed by the USGS, the glacier had retreated, although it was less than 50 m. A comparison of the USGS 1960 Anchorage A-4 1:63,360-scale topographic quadrangle map (appendix B), which was based on the 1957 photography, with oblique aerial photographs taken by the author on 8 August 2000 (fig. 270) shows that the lower part of the glacier has retreated approximately 400 m and is now covered by a thick brown-and-black debris cover. The northern perimeter of the glacier consists of stagnant ice-cored moraine.

Compilation and analysis of this information suggests that, between 1910 and 2000, Serpentine Glacier retreated and thinned significantly and that its stagnating debris-covered terminus was located about 2.1 km behind the moraine marking the "Little Ice Age" maximum position. From 1900 until 1914, there was no change in the position of the terminus of Serpentine Glacier. This period of stability was followed by about 400 m of retreat, beginning after 1914 and lasting until 1931, with a minimum average retreat rate of about 23.5 m a^{-1}. Little change was observed between 1931 and 1941. By 1950, the

Figure 270.*—8 August 2000 oblique aerial photograph of Serpentine Glacier. The lower 4 km of the glacier is mantled by thick debris. Note that the terminus of Serpentine Glacier is fronted by a outwash plain fan delta and is no longer at tidewater. Note also that, of the glaciers visible above the terminus, only the northernmost on the right, an unnamed tributary glacier, makes contact with Serpentine Glacier. Photograph by Bruce F. Molnia, U.S. Geological Survey.*

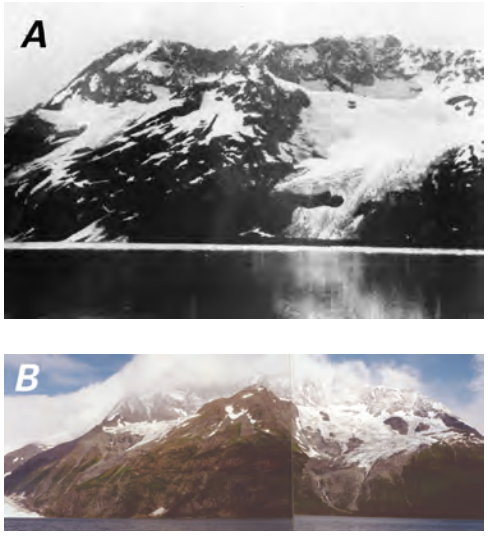


***Figure 271.**—Two northwest-looking views showing changes in Baker and Detached Glaciers between 1909 and 2000.* **A,** *1909 photograph by Grant and Higgins shows Baker Glacier (right) and Detached Glacier (left). Baker Glacier descends to near sea level, but Detached Glacier terminates at approximately 300 m above sea level. (Published as Plate XXI A, Grant and Higgins, 1913.)* **B**, *6 September 2000 photographic mosaic from nearly the same vantage point as* A *shows the changes that occurred in both glaciers during the 91 years between photographs. Baker Glacier thinned everywhere and retreated a maximum of more than 1 km. Although Detached Glacier has not retreated as much, it also thinned significantly. Photographic mosaic by Bruce F. Molnia, U.S. Geological Survey. Larger versions of these figures are available online.*

terminus of Serpentine Glacier had advanced between 200 and 300 m, yielding an average advance rate of between about 22 and 33 m a^{-1}. Little change was observed between 1950 and 1957. From 1957 until 1971, the glacier's terminus position retreated minimally, with an average retreat rate of about 7 m a^{-1} (AGS Glacier Studies Map No. 64–3–G9) (Field, 1965). From 1971 until 2000, the terminus of Serpentine Glacier retreated about 400 m, and ice became stagnant along most of its terminus area. The average rate of retreat during the latter part of the 20th century was about 14 m a^{-1}.

Penniman Glaciers, Baker Glacier, and Detached Glacier.—The Penniman Glaciers, Baker Glacier, and Detached Glacier are all small cirque and hanging glaciers on the eastern and southern flanks of Mount Muir. All have retreated significantly during the 20th century. Baker and Detached Glaciers were investigated and photographed by many scientists, including the Harriman Alaska Expedition in 1899 and Grant and Higgins in 1905 and 1909 (fig. 271*A*) (Grant and Higgins, 1910, 1911a, b, 1913, pl. XXIA), described and mapped by Martin and the NGS party in 1910 (Tarr and Martin, 1914), and photographed by the author on 6 September 2000 (fig. 271*B*).

The two Penniman Glaciers were named and described in 1914 by Keen (1915a). Since 1914, both have retreated as much as 700 m, with most of the retreat coming at the end of the 20th century. Field (1975b) reported that, between 1914 and 1931, the glaciers experienced a small retreat. Between 1931 and 1935, they advanced slightly. A significant, although not quantified, retreat occurred through 1961, with little change through 1971. During the 30 years since being first observed by the author in 1974, the glaciers have retreated about 500 m.

Martin investigated a large end moraine located adjacent to the shoreline below Baker Glacier (Tarr and Martin, 1914, fig. 47, p. 332). He discovered that some time between the end of the 18th century and the early part of the 19th century, an expanded Baker Glacier reached the shore of Harriman Fiord with a bulb-shaped terminus that was approximately 800 m in diameter. Between the time of the early 18th century maximum and 1910, the glacier had retreated about 400 to 500 m. Since then, Baker Glacier has retreated as much as 1.2 km. However, there have been several intervals of significant advance. Between 1910 and 1914 (Keen, 1915a), Baker Glacier advanced about 300 m to the foreland at the base of Mount Muir. By 1925, when Keen again visited the glacier, a substantial retreat was underway. When the glacier was next observed in 1931 and 1935 by Field (1932, 1937), it was again advancing. Between 1931 and 1935, this advance was as much as 50 to 60 m. Between 1935 and 1949, there was minimal change (Field, 1948; Baird and Field, 1951). Field (1975b) reported that, between 1949 and 1964, there was some recession, followed by a period of no appreciable change through 1971. During the author's 30-year period of observation from 1974 to 2004, Baker Glacier has retreated about 700 m.

Similarly, at the end of the 20th century, Detached Glacier was rapidly retreating. Grant and Higgins (1913) described a barren zone present in 1905 and 1909 below the hanging terminus of the glacier, which they interpreted to indicate that Detached Glacier was recently much larger and was connected to an extended and larger Surprise Glacier. Martin, who visited the glacier in 1910, found no evidence of change since the 1899 visit of the Harriman Alaska Expedition (Tarr and Martin, 1914). Similarly, Keen (1915a) saw no evidence of change through 1914. A significant amount of retreat occurred through 1931, followed by about 30 m of advance between 1931 and 1935 (Field, 1937). By 1957, the glacier was approximately 1.5 km long, and about 200 m of the lower hanging tongue of the glacier had disappeared. Field (1975b) reported that, between 1957 and 1961, there was some recession, followed by a decade of no appreciable change through 1971. During the author's 30 years of observation, from 1974 to 2004, Detached Glacier has retreated about 500 m.

Surprise Glacier.—Surprise Glacier, located at the head of Surprise Inlet, is so named because it was the first glacier the Harriman Alaska Expedition saw when it entered Harriman Fiord. The Harriman Alaska Expedition photographed the glacier's tidewater terminus and much of its lower reaches from a distance, and Gannett (Gilbert, 1904) prepared a map showing its terminus position but did not conduct any detailed examination of its margin. A detailed analysis, however, was first accomplished in 1905 and again in 1909 by Grant and Higgins (1913). They noted a large barren zone extending about 160 m beyond Surprise Glacier's 1899 terminus position, suggesting that the glacier was in retreat when Gannett had mapped it. Grant and Higgins (1913) noted that the retreat was continuing in 1905 and that, by 1909, the total recession was about 1.8 km. Bedrock began to appear at the base of the southern part of the margin at that time.

By the time Martin visited the glacier in 1909 (Tarr and Martin, 1914), retreat totaled 2 km. Martin also noted vegetation-free marginal zones up to 90 m wide above the location of the 1909 terminus. Field (1932) reported that little change had occurred in the terminus position through 1931. But when Field (1937) observed the glacier in 1935, a slight advance had taken place. Minor fluctuations occurred, with both short periods of advance and retreat noted through 1971.

The author made several observations of the glacier between 1974 and 2004. During this 30-year period, the terminus retreated about 300 m, thinned, and narrowed, as an 8 August 2000 oblique aerial photograph taken by the author shows (fig. 272). A small bedrock outcrop began to become visible at the base of Surprise Glacier along its southern side, and several tributaries thinned, retreated, and even lost contact with the main glacier. In the early 1990s, Surprise Glacier had a length of 12.1 km, an area of 80 km^2, a width at its face of 1.2 km, and an AAR of 0.80; its accumulation area was 64 km^2, and its ablation area was 16 km^2 (Viens, 1995) (tables 2, 3). Stairway Glacier is the name that the Harriman Alaska Expedition gave to the large southeast-flowing tributary that contributes a substantial amount of ice to the northern side of the trunk of Surprise Glacier. Part of Surprise Glacier and all of the glaciers in the northern half of Harriman Fiord are shown in figure 266, a 24 August 1978 AHAP false-color infrared vertical aerial photograph.

Cataract Glacier.—Cataract Glacier, a 2.5-km-long northeast-flowing hanging glacier that drains into the southern side of Surprise Inlet, had a tidewater terminus that was in contact with Surprise Glacier when it was first observed by the Harriman Alaska Expedition in 1899. Grant and Higgins (1913) visited the glacier in 1905 and 1909, photographing it in 1909

Figure 272.*—8 August 2000 northwest-looking oblique aerial photograph shows Surprise Inlet and Surprise and Cataract Glaciers. Cataract Glacier (far left), which had a tidewater terminus in 1910, retreated approximately 600 m. Photograph by Bruce F. Molnia, U.S. Geological Survey.*

***Figure 273.**—Two southwest-looking photographs showing changes at Cataract Glacier during the 91-year period from 1909 to 2000.* **A**, *1909 photograph by Grant and Higgins shows the terminus of Cataract Glacier reaching tidewater, although some bedrock is exposed at sea level (Grant and Higgins, 1913).* **B**, *6 September 2000 view of Cataract Glacier shows that its terminus retreated approximately 700 m and is now located more than 200 m above sea level. Exposed bedrock completely surrounds the ice margin. Photograph by Bruce F. Molnia, U.S. Geological Survey. Larger versions of these figures are available online.*

(fig. 273*A*). They noted that it experienced no appreciable change through 1909. They described "a narrow bare zone along the west side of the glacier, but the extent to which the shrubs have encroached upon this zone indicates that the ice stream has not in recent years (perhaps 25 years) been much larger than at present" (Grant and Higgins, 1913, p. 37).

Cataract Glacier was advancing and overriding shrubs and willows (*Salix* sp.) along the margin when Martin visited the glacier in 1910 (Tarr and Martin, 1914). In 1914, Keen (1915a) visited the glacier and observed that the terminus of the glacier had not changed its position since Martin's visit. Eleven years later, in 1925, she observed that the glacier had experienced a small recession since 1914 but that its terminus was still at tidewater. In 1931, Field (1932) noted that the terminus had retreated from tidewater but that it was thickening at higher elevations. By 1935 (Field, 1937), the glacier had readvanced to tidewater but was showing signs of narrowing along its margins. Retreat was underway again when Bradford Washburn photographed it in 1938. This retreat, greater than 500 m, continued through Field's last reported observation of the glacier in 1968 (Field, 1975b). The author made several observations of the glacier between 1974 and 2004, and photographed it on 6 September 2000 (fig. 273*B*). During this 30-year-period, the terminus retreated about 300 m, thinned, and narrowed.

Roaring Glacier.—Roaring Glacier, a hanging glacier with a reconstituted glacier at its base, has retreated since 1899. Gilbert (1904, p. 96) described why it was named Roaring Glacier as follows: "Roaring Glacier, between the Cataract and the Harriman, owes the peculiarity suggesting its name to an abrupt change of grade. From a comparatively gentle slope it passes to one so steep that loose masses find no lodgement, and as its movement steadily projects its end beyond the point of inflection, fragments of ice break away and tumble down the steep incline, to gather in a heap far below, where they lie until melted." On the basis of the size of the ice and snow accumulation at its base, Grant and Higgins (1913) concluded that the glacier was less active in 1905 than it was in 1899 and more active in 1909. In August 1910, the terminus of Roaring Glacier was between 200 and 300 m above the fjord. By 1957, it was approximately 500 m above the beach. During that 47-year period from 1910–1957 (Field, 1975b), the glacier retreated an additional 300 m. Field also reported that little additional change occurred through 1971. Between 1974 and 2004, the interval of the author's observations, the glacier thinned and retreated another 50 to 100 m up the face of its steep bed. A 12 July 1978 photograph shows the glacier in retreat (fig. 274).

Harriman Glacier.—As was the case with all of the other glaciers in Harriman Fiord, Harriman Glacier was first described by scientists of the Harriman Alaska Expedition in 1899. Gilbert (1904, p. 94–95) described Harriman Glacier as follows: "The valley containing Harriman Glacier is a continuation of the main trough of the fiord and holds the same general southwest trend. The glacier curves toward the west and then toward the south, disappearing from view at a distance of nearly ten miles. As the most distant portion seen has a gentle slope and lies far below the bordering mountains, it is probable that the sources are still several miles beyond. Its general width is about a mile and a half, but its high-grade tributaries are so thick-set as practically to coalesce, especially on the south-east side, giving a broad expanse of nearly continuous ice and snow. This expanse, fully commanded from the water, makes the view of the glacier a most impressive spectacle."

Gilbert (1904, p. 95) described a "detrital bank," probably an outwash fan, delta, or moraine along the "eastern" (southern) side of the terminus. He stated: "Above this bank the frontal cliff is low and irregular, but elsewhere it is lofty, ranging in height from 200 to 300 feet. From such a cliff an active discharge of bergs might be assumed, but our parties encountered only a moderate quantity of floating ice near the head of the fiord." According to tidewater glacier cycle theory (fig. 42), the absence of bergs indicates that

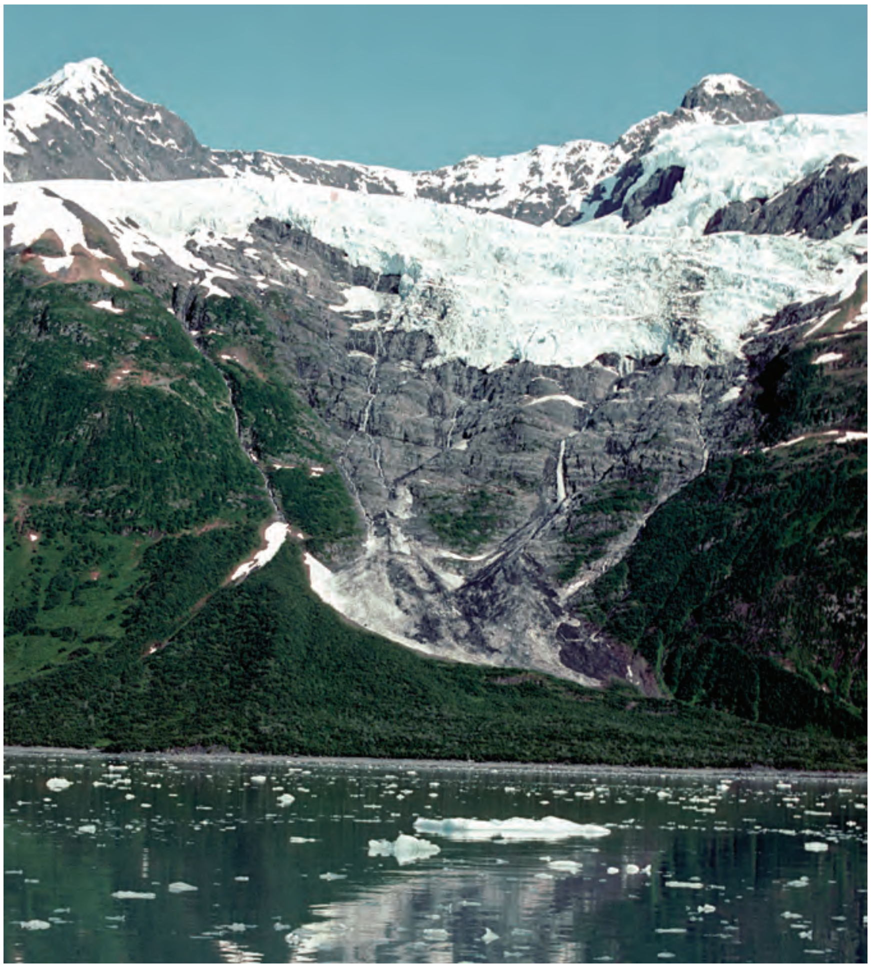


Figure 274.—12 July 1978 photograph shows the retreating terminus of Roaring Glacier. Avalanching built a small reconstituted glacier at the base of the near-vertical rock face. Photograph by Bruce F. Molnia, U.S. Geological Survey.

the terminus of the glacier was grounded and sitting on the back side of its submarine moraine. Gilbert (1904, p. 95) could not determine if the glacier was advancing or retreating: "The glacier is not closely approached by forest growth, but shrubs were seen on the shore of the fiord within a few hundred yards of the ice. If the ice is diminishing, the recent retreat of the glacier front would appear not to have been rapid." In 1899, Gilbert and Curtis made several photographs of the terminus.

In 1905 and again in 1909, Grant and Higgins (1913) visited Harriman Glacier and determined that the southern part of the terminus had retreated about 105 m between 1899 and 1905 and another approximately 210 m through 1909. They noted no change in the position of the northern part of the terminus during the same decade. Martin visited the glacier in 1910 (Tarr and Martin, 1914, p. 335) and reported that, on 10 August 1910, the western margin had advanced about 210 m beyond its 1909 position: "The eastern [southern] margin also advanced, coming forward the whole distance that it had retreated from 1899 to 1909. In 1910 it seemed to be slightly beyond the 1899 position and there was much thickening of the eastern terminus...." Gilbert's detrital bank was not visible, and iceberg production was much greater than in 1899.

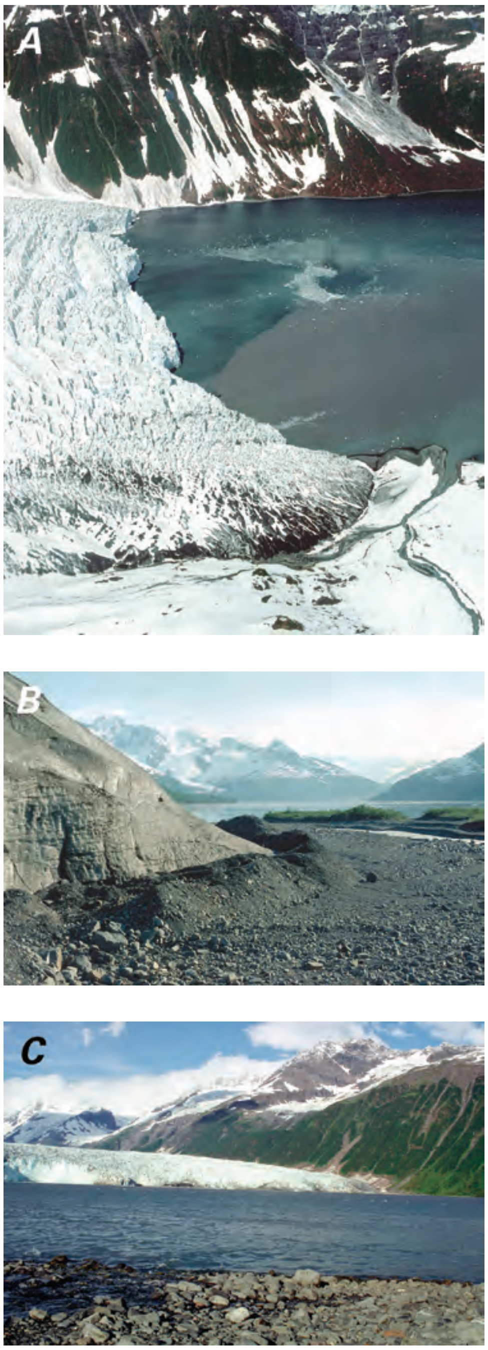
Martin (Tarr and Martin, 1910) also noted that, in addition to the advance and thickening of the part of the eastern margin on the shores of the bay, the glacier was advancing on the land. Annual plants in the barren zone were being buried, and a ridge of push moraine a foot or two high lay at the base of a lofty, uncrevassed ice cliff. On 1 August 1910, the cliff was about 48 m from an older terminal moraine that marked the maximum of a former advance, doubtless before 1899. An analysis of vegetation in the fjord led Cooper (1942) to conclude that Harriman Glacier had not advanced beyond the moraine that it was approaching in 1910 for at least 500 years.

Keen (1915a) noted that Harriman Glacier was continuing to advance in 1914. By 1931 (Field, 1932), the terminus on the southern side of the glacier had advanced approximately 975 m from its 1909 position, an average annual advance of about 44 m a^{-1}. Four years later, Field (1937) returned and measured another 47 m of advance, for a total of about 1,022 m between 1909 and 1935. Field measured 470 m of advance on the northern side of the terminus during this 25-year period.

Field (1975b) reported that, by 1957, the terminus of Harriman Glacier had advanced about another 100 m. The southern side had advanced about 150 m by 1961. By 1964, when it was next observed, the southern margin had retreated about 25 m, and the fjord-based ice cliff had retreated about 100 m. Field (1975b) attributed this retreat to the 2.4 m of regional subsidence caused by the 1964 Alaska earthquake. By 1966, most of the terminus had resumed advancing. By 1971, the year of Field's last recorded observation, the terminus on the north had advanced 30 to 50 m, and the ice cliff on the southern side had increased in height.

Field (1975b) summarized the changes of Harriman Glacier's southern margin between 1899 and 1971 as follows: (1) From the unknown date of its pre-20th century maximum to 1899, a recession of about 350 m; (2) from 1899 to 1905, a recession of about 100 m, yielding an average rate of retreat of about 17 m a^{-1}; (3) from 1905 to 1909, a recession of about 200 m, yielding an average rate of retreat of about 50 m a^{-1}; (4) from 1909 to 1910, an advance of about 300 m, yielding an average rate of advance of about 300 m a^{-1}; (5) from 1910 to 1931, an advance of about 675 m, yielding an average rate of advance of about 32 m a^{-1}; (6) from 1931 to 1935, an advance of about 47 m, yielding an average rate of advance of about 12 m a^{-1}; (7) from 1935 to 1957, an advance of about 97 m, yielding an average rate of advance of about 4 m a^{-1}; (8) from 1957 to 1961, an advance of about 151 m, yielding an average rate of advance of about 38 m a^{-1}; (9) from 1961 to 1964, a retreat of about 28 m, yielding an average rate of retreat of about 9 m a^{-1}; (10) from 1964 to 1966, an advance of about 26 m, yielding an average rate of advance of about 13 m a^{-1}; and (11) no change between 1966 and 1971. Hence, between 1899 and 1971, the net change on the southern margin of the glacier was an advance of 1,268 m. For the northern margin, Field reported an advance of 750 to 850 m.

The author observed Harriman Glacier more than a dozen times between 1974 and 2004. Between 1974 and 1984, the glacier was advancing. With the exception of the extreme southern margin of the terminus, which was fronted by a delta, the entire terminus was tidewater in 1976 (fig. 41) and consisted of a vertical face with three semicircular calving embayments, as an 8 June 1976 oblique aerial photograph taken by the author (fig. 275*A*) shows. On 15 July 1978, a push moraine adjacent to the southern margin indicated that the glacier was recently advancing (figs. 64, 275*B*), while the northern half of the terminus still maintained a vertical iceberg-calving face. Lethcoe (1987)



◄ ***Figure 275.**—Three photographs show characteristics of Harriman Glacier during its late-20th century advance.* **A***, 8 June 1976 north-looking oblique aerial photograph shows the width of the terminus of Harriman Glacier. Note the near-vertical character of the ice face, except along the southern edge of the terminus, and the three calving embayments cut into the margin.* **B***, 15 July 1978 photograph shows a push moraine formed by a small advance of the terminus earlier in the year. Similar push moraines were seen in both 1998 and 2000 in front of the central part of the terminus, indicating that, even with a cessation of annual advance, winter flow may result in seasonal advances of the terminus.* **C***, In this 6 September 2000 photograph, the terminus of Harriman Glacier is more than 750 m more advanced than in 1899 and approximately 50 m forward of its 1978 position and is beginning to change its profile from subrounded to more vertical. By 2002, the northern terminus was near vertical in profile. Photographs by Bruce F. Molnia, U.S. Geological Survey. Larger versions of these figures are available online.*

reported that, in 1979 and 1981, two large calving embayments developed in the terminus, with radii of more than 150 m. Similar embayments developed at Bering and Columbia Glaciers before their recent rapid retreats. In 1984, the northern margin of the glacier was adjacent to a triangular-shaped vegetation-free area that served as a major accumulation area for snow sliding off the northern wall of the valley. Lethcoe (1987) stated that, by 1986, both embayments had refilled, and the margins of the glacier advanced about 15 m, causing the abandonment of a kittiwake *(Rissa* sp.) rookery.

By 6 September 2000, the profile of nearly the entire terminus of Harriman Glacier had changed from vertical to subrounded (fig. 275*C*). With the exception of one calving embayment located adjacent to the delta along the southern margin of the glacier, the entire face of the terminus is fronted by an exposed sediment plain. In 1999, the embayment had a radius of more than 120 m. Since 1984, the position of the northern part of the terminus of the glacier had receded slightly from the snow chute, and a barren zone was developing along the northern margin. Several aerial and ground observations made by the author between 1999 and 2004 confirm that no recent advance of the terminus has occurred. Although the elevation of the terminus has continued to decrease, no evidence of significant retreat along the southern margin or the face of the glacier has been observed. In August 1999, a push moraine was observed about 5 m in front of the glacier along part of the face fronted by the sediment plain. In the early 1990s, Harriman Glacier had a length of 12 km, an area of 60 km^2, a width at its face of 1.9 km, and an AAR of 0.79. Its accumulation area was 48 km^2, and its ablation area was 13 km^2 (Viens, 1995) (tables 2, 3).

Dirty and Wedge Glaciers.—Dirty and Wedge Glaciers are two small northwest-flowing ice tongues that descend the southern wall of Harriman Fiord. Both have retreated more than 1 km since they were first observed. In 1899, Dirty Glacier terminated close to the shoreline. By 1909, the terminus was about 400 m upvalley from the beach (Grant and Higgins, 1913). By 1935 (Field, 1937), the glacier's debris-covered terminus was about 600 m from the beach, with bare ice about 200 m further upglacier. Field (1975b) reported that an additional 425 m of retreat had occurred by 1961. Hence, between 1899 and 1961, the terminus of Dirty Glacier retreated about 1.25 km, yielding an average annual retreat rate of about 20 m a^{-1}. Between 1961 and 2000, the terminus retreated about an additional 250 m. When the author observed the glacier from the air in 2000, it was continuing to thin and retreat and was on the verge of separating into two distinct ice tongues.

In 1899, Wedge Glacier had a steep upper section and a low-relief section terminating close to sea level. The terminus area changed little through 1935 (Field, 1937). By 1961 (Field, 1975b), the low-relief lower part of the glacier had melted away. Between 1950 and the middle 1960s, Field estimated that the glacier had retreated about 500 m (Field, 1975b). Through the end of the 20th century, the terminus retreated an additional 750 to 800 m. When the author observed it from the air in August 2000, all that remained of the glacier was a small ice mass hanging on the valley wall below the divide between Harriman Fiord and Bettles Bay.

Toboggan Glacier.—Toboggan Glacier, a small retreating valley glacier flowing from Mount Doran toward the southern side of Harriman Fiord was visited and photographed by Grant and Paige in 1905 and was visited by Grant and Higgins in 1909 and by Martin in 1910. Martin (Tarr and Martin, 1914) mapped both moraine segments and a trimline that was formed by a 19th century advance of the glacier that reached to the shore of the fjord, a distance of about 400 to 500 m beyond the 1910 terminus. He speculated that this advance had occurred in the 1830s. A second moraine located within the outer moraine represents an advance dating to ca. 1880. The 1905 terminus position was about 325 m behind the outer moraine (Grant and Higgins, 1913). Between 1905 and 1909, the glacier fluctuated, advancing about

120 m and then retreating about 195 m. When Toboggan Glacier was photographed in 1909, a trimline was located more than 80 m above the glacier on both sides of its valley. Between 1909 and 1910, the glacier retreated an additional 23 m. Between 1910 and 1931 (Field, 1932), the glacier retreated an additional 282 m, for an average annual rate of about 13.5 m a^{-1}. By 1935 (Field, 1937), the glacier had retreated another 22 m.

During the next 22 years, Toboggan Glacier continued to retreat. It had lost an additional 416 m when it was observed by an AGS party that included Field (1975b) in 1957 at the start of the IGY. Four years later, in 1961, a Brigham Young University field party measured an additional 64 m of retreat, which placed the terminus more than 775 m behind its 1910 position (Field, 1975b). When the author observed the glacier in 1978, 2000, and 2004, it was continuing to retreat, losing at least 325 m of its length through 2000. Hence, in the 90-year period between 1910 and 2000, the terminus of Toboggan Glacier retreated about 1.1 km, for an average rate of 12.2 m a^{-1}.

Glaciers of Port Wells

Bettles and Pigot Glaciers.—Bettles and Pigot Glaciers are two small retreating glaciers that drain into Port Wells. Field (1975b), reported that Bettles Glacier retreated about 1 km between 1910 and 1950. Since then, it has retreated at least 1 km.

Pigot Glacier has a similar history, retreating about 1.5 km in the 40-year period between 1910 and 1950. Since 1950, it also has retreated at least 1 km. Debris from a large landslide (Post, 1967a) covers the terminus, as an 8 August 2000 oblique aerial photograph made by the author (fig. 276) shows. This debris could be the product of slides that occurred during earthquakes in the 1940s and 1964.

Figure 276.—*8 August 2000 oblique aerial photograph, looking northwest, of Pigot Glacier, which has retreated more than 2 km since the beginning of the 20th century. Debris from a large landslide covers the terminus. Photograph by Bruce F. Molnia, U.S. Geological Survey.*

Glaciers of Passage Canal

Seth and Billings Glaciers.—Seth and Billings Glaciers are two small, poorly studied but actively retreating glaciers that drain into Passage Canal. Field (1975b) noted that Seth Glacier retreated about 1.6 to 1.8 km between 1910 and 1950. Since then, it has retreated about an additional 1.5 km.

Field (1975b) reported that Billings Glacier retreated about 1.5 km between 1910 and 1971. At that time, the glacier was separating into a pair of termini, located on either side of an emerging bedrock ledge. Continued retreat through the author's last observation on 8 August 2000 (fig. 277) has accentuated the separation and has also revealed a light-colored bedrock apron, exposed around the margin of the glacier. This light-colored bedrock, possibly aplite, is unlike any other rock unit in the immediate area. Since 1910, the glacier has retreated approximately 1.8 km.

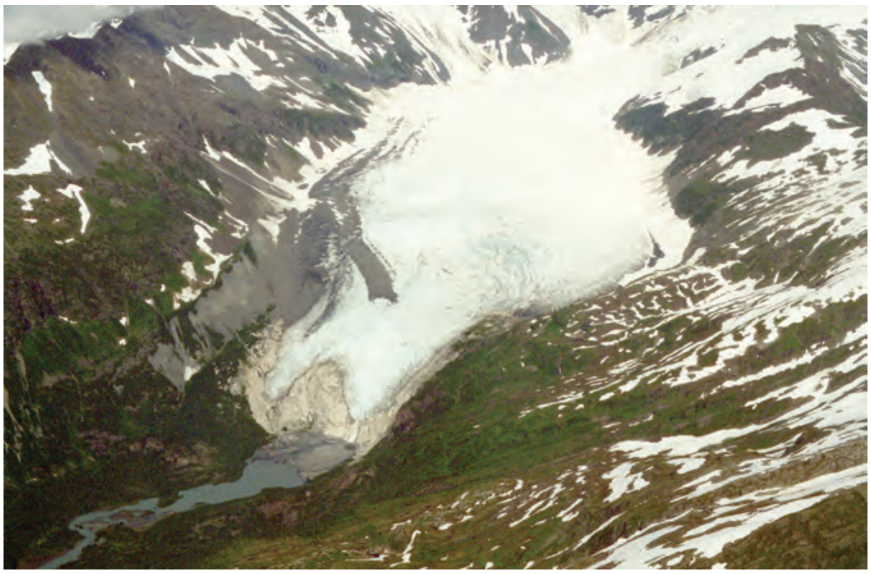


Figure 277.*—8 August 2000 north-looking oblique aerial photograph of the retreating and thinning Billings Glacier. Note the elevated lateral moraine and the light-colored outcrop of bedrock, possibly aplite, around the retreating terminus of the glacier. Photograph by Bruce F. Molnia, U.S. Geological Survey.*

Northwestern Chugach Mountains Segment—North-Flowing Large Valley Glacier Subdivision

The western part of this subdivision, which is the northwesternmost part of the Chugach Mountains (fig. 278), includes a number of small unnamed retreating glaciers located at the heads of Wolverine, Carpenter, Coal, and Monument Creeks and several unnamed creeks, all draining into the westward-flowing Matanuska River. All glaciers showed evidence of significant retreat and thinning when the author observed them from the air in September 2000 and again in September 2002. Some have extensive debris-covered termini, and some have newly vegetated, formerly barren zones around their margins. In this subdivision, glaciers lengthen to the east.

Glaciers that have lengths greater than 8 km and areas determined by Field (1975b, p. 465, 475) include (from west to east): an unnamed glacier (16 km, 74 km^2), an unnamed glacier (12 km, 21 km^2), Matanuska Glacier (46 km, 324 km^2), Powell Glacier (26 km [author's estimate]), Nelchina Glacier (39 km, 328 km^2), an unnamed glacier (9 km, 16 km^2), an unnamed glacier (8 km, 16 km^2), and Tazlina Glacier (47 km, 398 km^2). Each glacier shows significant evidence of thinning and retreat.

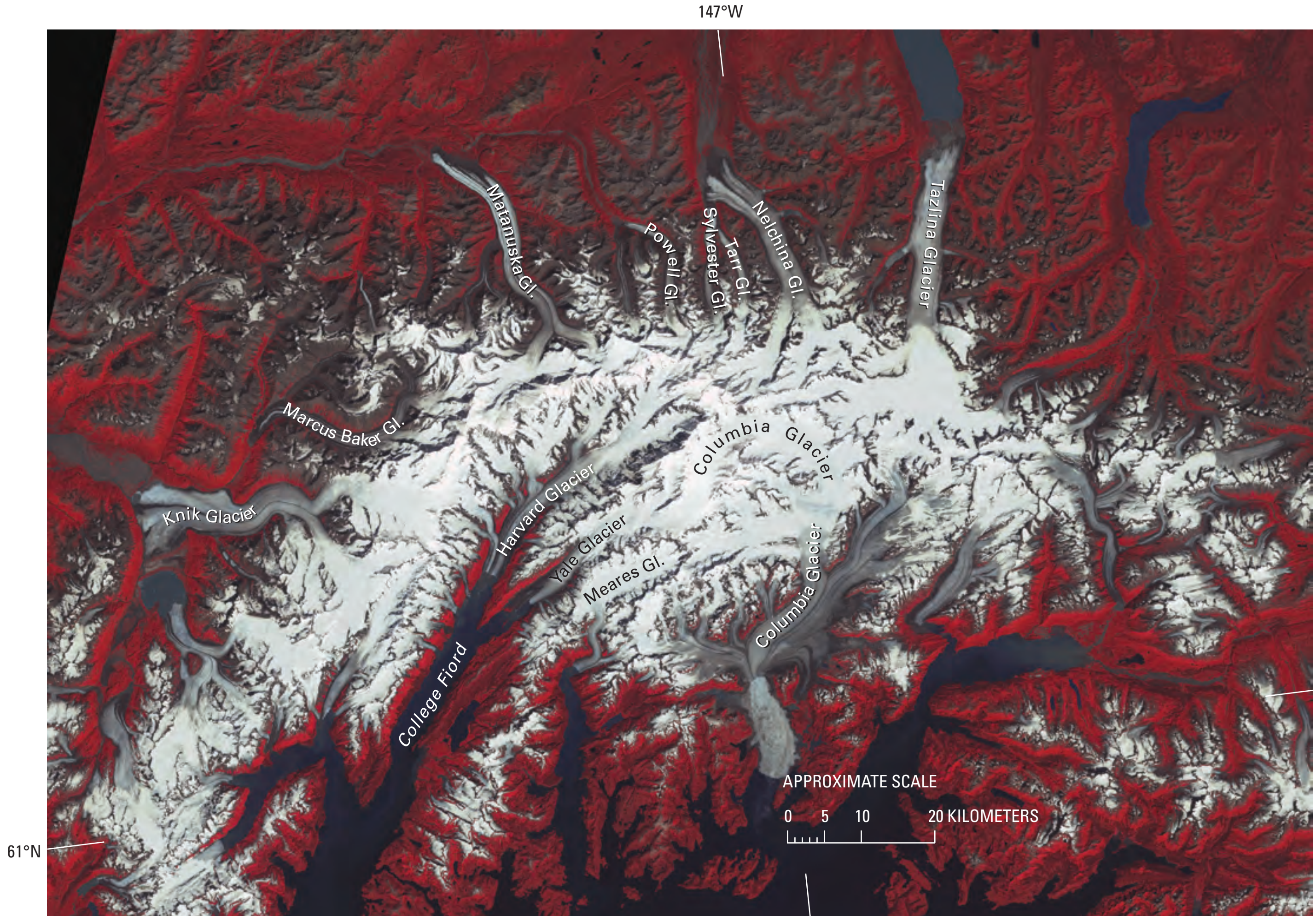


***Figure 278.**—1 August 2002 Landsat 7 ETM+ image of the northwestern part of the Chugach Mountains. Shown is the area that includes the large glaciers on the northwest side of the Chugach Mountains and northwestern College Fiord, including Columbia, Meares, Yale, Harvard, Knik, Marcus Baker, Matanuska, Powell, Sylvester, Tarr, Nelchina, and Tazlina Glaciers. Landsat 7 ETM+ image (7067017000221350; 1 August 2002; Path 67, Row 17) from the U.S. Geological Survey, EROS Data Center, Sioux Falls, S. Dak.*

Matanuska Glacier

Matanuska Glacier, which is 46 km long and visible from the Glenn Highway (fig. 279), drains more than 600 km^2 of the central Chugach Mountains between Mount Marcus Baker and Mount Thor. The glacier ranges in width from about 2.2 km near its source to about 5 km near its terminus. Most of the terminus area consists of debris-covered ice, the outer part of which is stagnant. On the western side of the glacier is a zone of active ice with a debris-free surface. Several times during the 20th century, the terminus of Matanuska Glacier has advanced and overridden stagnant ice and ablation moraine.

Nineteenth and early 20th century information about changes in Matanuska Glacier is very limited. In 1898, Mendenhall (1900) observed and photographed the terminus of the glacier and noted that a forest was growing on the surface of the debris-covered, stagnant eastern margin. Reid (1909) noted that the glacier was retreating in 1905. In 1954, Williams and Ferrians (1961) duplicated photographs taken by Mendenhall from Glacier Point. They determined that, in the ensuing 56 years, little horizontal retreat had occurred but noted that enlargement of areas of ablation moraine indicated considerable thinning of the ice. They also determined that a small advance had recently occurred over a till ridge on the west side of the glacier. Stone (1955) dated this advance at about 1945. Williams and Ferrians (1961) described a moraine located about 400 m in front of the terminus that is less than 200 years old. They also mapped earlier Holocene advances about 4,000 and 8,000 years ago that had left conspicuous moraines about 1.4 km and about 8 km, respectively, beyond the present terminus.

Figure 279.—30 August 2000 south-looking oblique aerial photograph shows that much of the terminus of Matanuska Glacier is debris covered. Both the east and west margins of the glacier show evidence of recent thinning and narrowing. Photograph by Bruce F. Molnia, U.S. Geological Survey.

Field (1975b) examined aerial photography of the terminus acquired in 1938, 1941, 1948, 1957, and 1964. He reported that, in the 19 years between Bradford Washburn's 1938 photograph and a 1957 USAF photograph, the terminus "is without significant change" (Field, 1975b, v. 2, p. 453). However, between 1957 and 1964, "slight recession can be seen" (Field, 1975b, v. 2, p. 453). Recession continued through the late 1970s. Between 1969 and 1974, Lawson (1979) reported but did not quantify ongoing retreat. During the summer of 1979, this retreat was interrupted, as the terminus advanced more than 30 m in 60 days. A comparison of Lawson's 1969 photograph and a BLM vertical photograph acquired on 17 August 1999 (BLM vertical aerial photograph No. R3–FL6–FR–15) shows about 1 km of terminus retreat during the 30-year period of coverage. When the author observed the glacier from the air on 30 August 2000 (fig. 279), the glacier showed multiple signs of continued thinning and retreat.

Since the l970s, research has been conducted at the western terminus of the Matanuska Glacier by scientists from the Cold Regions Research and Engineering Laboratory of the U. S. Army Corps of Engineers and by the academic community. Topics include sediment transport and deposition (Lawson, 1979), present-day formation of basal ice (Strasser and others, 1996), and short-pulse radar analysis of subglacier structure (Arcone and others, 1995).

Nelchina Glacier

Although it has been photographed from the air since 1938, Nelchina Glacier, which is 39 km long and 328 km^2 in area (Field, 1975b, p. 465), has not been the subject of any published field investigations. Field (1975b, v. 2, p. 357) reported that, in the 28 years between Bradford Washburn's 1938 photograph and a 1964 photograph of the glacier by Austin Post, "there was no conspicuous change in the position of the terminus. However, the lateral barren zone on the eastern side of the lower part of the glacier appears to have widened, indicating a lowering of the ice surface. Above this barren zone a conspicuous lateral moraine appears to mark the limit of a recent maximum in which the terminus was more advanced, perhaps as much as 500 m."

Between 1964 and 2000, Nelchina Glacier continued to thin and retreat. When the author observed it from the air on 30 August 2000 (fig. 280), several small ice-marginal lakes had developed adjacent to the eastern and central terminus regions of the glacier. Elevated lateral moraines and distinctive trimlines were clearly visible. As much as 2 km of retreat had occurred in

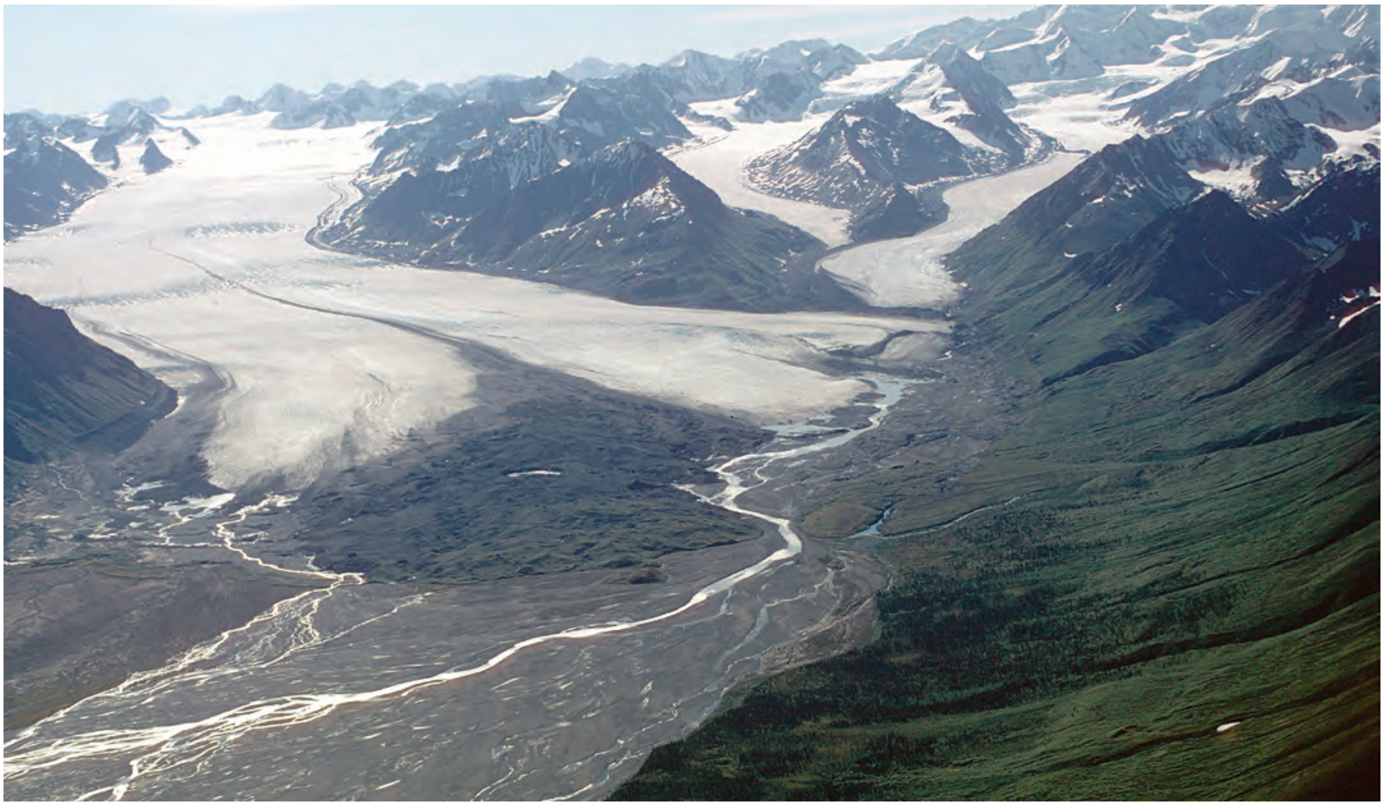


Figure 280.—30 August 2000 south-looking oblique aerial photograph of the thinning and retreating terminus of Nelchina Glacier and its western tributaries, Sylvester and Tarr Glaciers. The absence of any deflection of the moraine where Sylvester and Tarr Glaciers join the Nelchina Glacier suggests that they no longer contribute ice to the Nelchina Glacier. Much of the central part of the terminus of Nelchina Glacier is debris covered and extends beyond the bare ice lobes on either side. Both the east and west margins of Nelchina Glacier show evidence of recent thinning. Photograph by Bruce F. Molnia, U.S. Geological Survey.

the 36 years since the 1964 photograph by Austin Post. Sylvester and Tarr Glaciers, the two western tributaries to Nelchina Glacier, both have thinned significantly as well. At their junction with Nelchina Glacier, their flow has decreased so much that they no longer deflect the medial moraine that separates them from the western side of Nelchina Glacier.

Tazlina Glacier

Tazlina Glacier, another glacier that was not studied before the IGY, was visited in 1957 by an AGS field party that included Field and Viereck. Viereck (1967) examined the vegetation and confirmed that the terminus had not been more than 2.5 km further advanced since at least 1450 A.D. A terminal moraine located about 1.5 km from the 1957 terminus was formed between 1800 and 1820. Viereck (1967) estimated recession during the first half of the 20th century to be at a rate of 15 to 21 m a^{-1} and to be between 24 to 28 m a^{-1} since the middle 1940s. The author compared the USGS 1960 Valdez C-7 1:63,360-scale topographic quadrangle map, compiled in 1950, with a 1993 Landsat image. The comparison showed that an ice-marginal lake filling the area of post-1950s retreat formed along much of the terminus of Tazlina Glacier, leaving most of its former outwash plain as a sandflat at the head of Tazlina Lake. When the author observed the glacier from the air on 30 August 2000, continued retreat had all but separated it from its outwash plain (fig. 281). However, the upper part of the glacier, at an elevation of about 1,500 m, showed no evidence of thinning or retreat. Observations on the same day to the east of Tazlina Glacier showed large glacial cirques occupied only by small ice patches, another sign of ongoing glacier retreat.

Comparing 1950s map data of Tazlina Glacier with data obtained during geodetic airborne laser altimeter profiling surveys in the middle 1990s showed that, on an annual basis, Tazlina Glacier thinned by 0.687 m a^{-1}, had a volume decrease of 0.252 km^3 a^{-1}, and retreated 17 m a^{-1} (K. A. Echelmeyer, W. D. Harrison, V. B. Valentine, and S. I. Zirnheld, University of Alaska Fairbanks, written commun., March 2001).

Figure 281.—30 August 2000 oblique aerial photograph of Tazlina Glacier. The south-looking view of the terminus region shows that continued retreat had all but separated the glacier from its middle 20th century outwash plain. Note the 1950s end moraine. Photograph by Bruce F. Molnia, U.S. Geological Survey.

Northeastern Chugach Mountain Segment—The Turnagain Arm–Western Chugach Mountains Subdivision

In this subdivision, glaciers that have lengths greater than 8 km and areas determined by Field (1975b, p. 473–475) include Marcus Baker Glacier (39 km), an unnamed glacier (9 km), Knik Glacier (49 km, 380 km^2), Gannett Glacier (14 km, 24 km^2), Colony Glacier (29 km, 237 km^2), Lake George Glacier (24 km, 88 km^2), an unnamed glacier (8 km, 15 km^2), another unnamed glacier (12 km, 43 km^2), Whiteout Glacier (15 km [author's estimate]), an unnamed glacier at the head of Troublesome Creek (11 km, 20 km^2), Twentymile Glacier (15 km, 32 km^2), Eagle Glacier (14 km, 49 km^2), Eklutna Glacier (13 km, 31 km^2), Hunter Creek Glacier (9 km, 14 km^2), and Metal Creek Glacier (9 km, 12 km^2).

Marcus Baker Glacier

Marcus Baker Glacier, which is 39 km long (Field, 1975b, p. 475) and has an area estimated by the author of about 20 km^2, is located at the head of Grasshopper Valley and may have been a "Little Ice Age" tributary glacier to Knik Glacier, joining it from the north. Today, its terminus is more than 6 km from Knik Glacier. Marcus Baker Glacier originates on the western flank of Mount Marcus Baker (4,107 m), one of the highest peaks in the Chugach Mountains. Although it has not been the subject of any scientific investigations, its terminus was photographed by Bradford Washburn in 1938 and the U.S. Army Air Forces in 1941. At that time, the lower 1.75 km of the glacier was covered by debris and showed little evidence of retreat (Field, 1975b). When it was mapped in the 1950s, the terminus of Marcus Baker Glacier had begun to retreat. By 1996, an outwash plain about 2 km in length fronted the terminus. When the author photographed the glacier from the air on 30 August 2000, it had retreated about 2.5 km from its 1950s terminus position and showed conspicuous evidence of thinning along both margins (fig. 282).

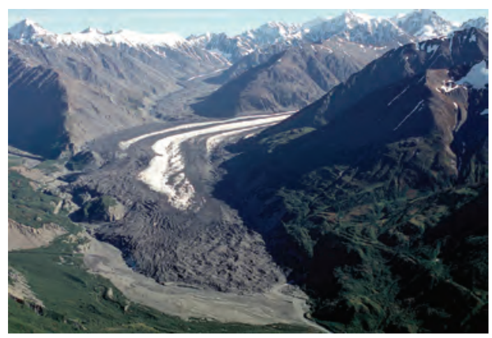


Figure 282.—30 August 2000 north-looking oblique aerial photograph of the debris-covered terminus area of Marcus Baker Glacier. The unnamed tributary that enters from the north has separated from Marcus Baker Glacier. Note the difference in elevation between the ice surface and the vegetation on the bedrock knob on the west side of Marcus Baker Glacier, evidence of recent thinning. Photograph by Bruce F. Molnia, U.S. Geological Survey. A larger version of this figure is available online.

Knik Glacier

Knik Glacier (fig. 6), which is 49 km long and 380 km^2 in area (Field, 1975b, p. 474), terminates in a small piedmont lobe, the perimeter of which is more than 16 km, as a 30 August 2000 oblique aerial photograph taken by the author shows (fig. 283). Several moraines located up to 1 km from the terminus document the fact that Knik Glacier was larger in the early 20th century. At various times during the past, the southern part of the terminus, which flows

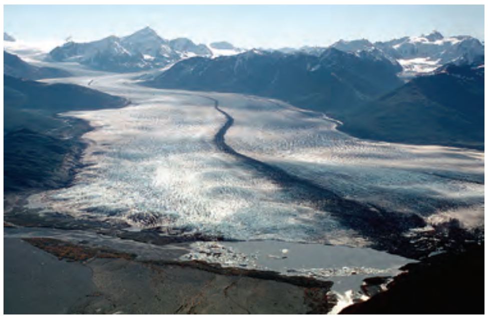


Figure 283.—30 August 2000 east-looking oblique aerial photograph of the northern part of the terminus of Knik Glacier. The perimeter of this small piedmont lobe is more than 16 km. Thinning and retreat of the terminus have reduced the chance that the volume of water in the glacier-dammed Inner Lake George (off the photograph to the right) would reach a critical level and trigger a jökulhlaup. Photograph by Bruce F. Molnia, U.S. Geological Survey. A larger version of this figure is available online.

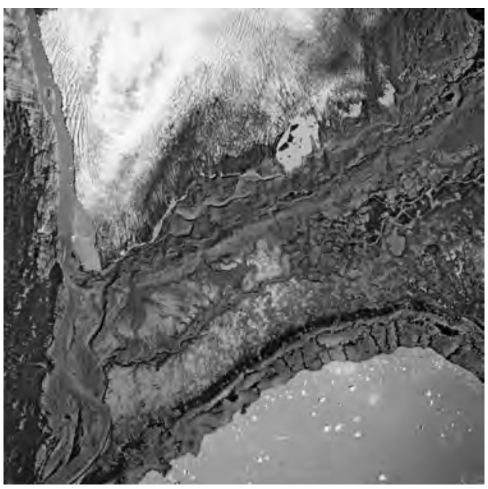


Figure 284.—17 August 1999 vertical aerial photograph of the southwestern part of the terminus of Knik Glacier. North is at the top of the photograph. The Gorge is visible on the west side of the photograph. Inner Lake George is at the bottom. Photograph no. R3–FL2– FR–224 from the U.S. Bureau of Land Management. A larger version of this figure is available online.

towards the eastern flank of Mount Palmer, would occasionally make contact with the mountain, blocking meltwater flow, forming a lake, and creating flood potential. Hulsing (1981, p. 10) reported that, before 1900, one flood destroyed three Native villages: “No previous damage along the Knik River had been recorded although, according to the Indians living in the area, the lake emptied once every 15 or 20 years.”

In 1915, Capps (1916, p. 169) visited Knik Glacier and described the terminus: “No facts were observed that would indicate any great amount of recent retreat of Knik Glacier. ... The presence of ... bushes and trees so near the ice front shows conclusively that the glacier is now almost as far advanced as it has been for many years.” He also provided some details about the lake formed by Knik Glacier: “Natives and prospectors report a lake many miles long that occupies a valley along the east side of the southwest fork of Knik Glacier. This lake fills gradually and at intervals of six or seven years breaks out through Knik Glacier and sends great floods of water down Knik River and Knik Arm” (Capps, 1916, p. 169). Capps reported that such a flood occurred in September 1915, causing significant damage to a railroad being constructed across Knik Arm.

After 1915, the lake formed and emptied more regularly. Every winter between 1915 and 1966, with the exception of 1963, the southwestern terminus of Knik Glacier (fig. 284) advanced against the eastern side of Mount Palmer, closing the 10-km-long ice-marginal drainage channel known as The Gorge. This blockage impounded meltwater flowing from Lake George Glacier, Colony Glacier, and the southeastern part of the terminus of Knik Glacier and prevented it from entering the Knik River. Meltwater would pond behind the ice dam each spring and form a large lake (Lake George) that covered an area of as much as 75 km^2. By late June or early July, the lake would overtop its ice dam and begin a flood that would last for about 2 weeks. During the peak of the jökulhaup, discharge could be as much as 9.45×10^6 l s^{-1} (1.5×10^8 gal min^{-1}), as predicted by Meier in an unpublished report submitted to the U.S. Department of the Interior in the late 1950s (Mark F. Meier, written commun., 2004). Since 1966, owing to late 20th century thinning and retreat (fig. 285), Knik Glacier has failed to seal the flood channel (Trabant and Mayo, 1980; Mayo and Trabant, 1982). During the early 1970s, the channel had a width of about 300 m (Post and Mayo, 1971). When the author observed it from the air on 12 April 2002, the width was less than 100 m, but the height of the ice wall was significantly smaller.

In 1951, Stone (1955) visited the glacier and studied evidence of higher lake levels. He found evidence of an undated maximum lake level as much as 45 m higher than the 1951 level and a 1940s level as much as 2 m higher. Stone (1955, p. 43–44) concluded that “Knik Glacier’s two ice faces appear to have been in about the same position for at least the past 50 years and the lake probably has existed for as much as 75 years.”

Photographs of the glacier obtained between 1938 and 1968 (Field, 1975b) document that, between 1938 and 1957, the terminus retreated between 150 and 200 m and that its location was 600 to 1,000 m from its most advanced position. An additional 200 m of retreat had occurred by 1964, with 100 m more by 1968. Hence, in the 30 years between 1938 and 1968, the terminus of Knik Glacier retreated between 450 and 500 m, an average retreat of about 15 m a^{-1}, the greatest retreat being at the end of the interval. Since 1968, the glacier has continued to thin and retreat. By 2000, when the author photographed it from the air, the terminus had retreated about 600 m from its 1968 position.

Figure 285.—30 August 2000 south-looking oblique aerial photograph of the westernmost part of Knik Glacier and the entire length of The Gorge. The Lake Fork of the Knik River flows through The Gorge. Inner Lake George is in the background with Colony Glacier behind it. Although Knik Glacier is closer to the bedrock wall of Mount Palmer now than it has been in the past few decades, the height of the terminus is too low to impound much water because the terminus is fractured and may even be floating, permitting water to drain through the Lake Fork of the Knik River. Photograph by Bruce F. Molnia, U.S. Geological Survey.

Gannett Glacier

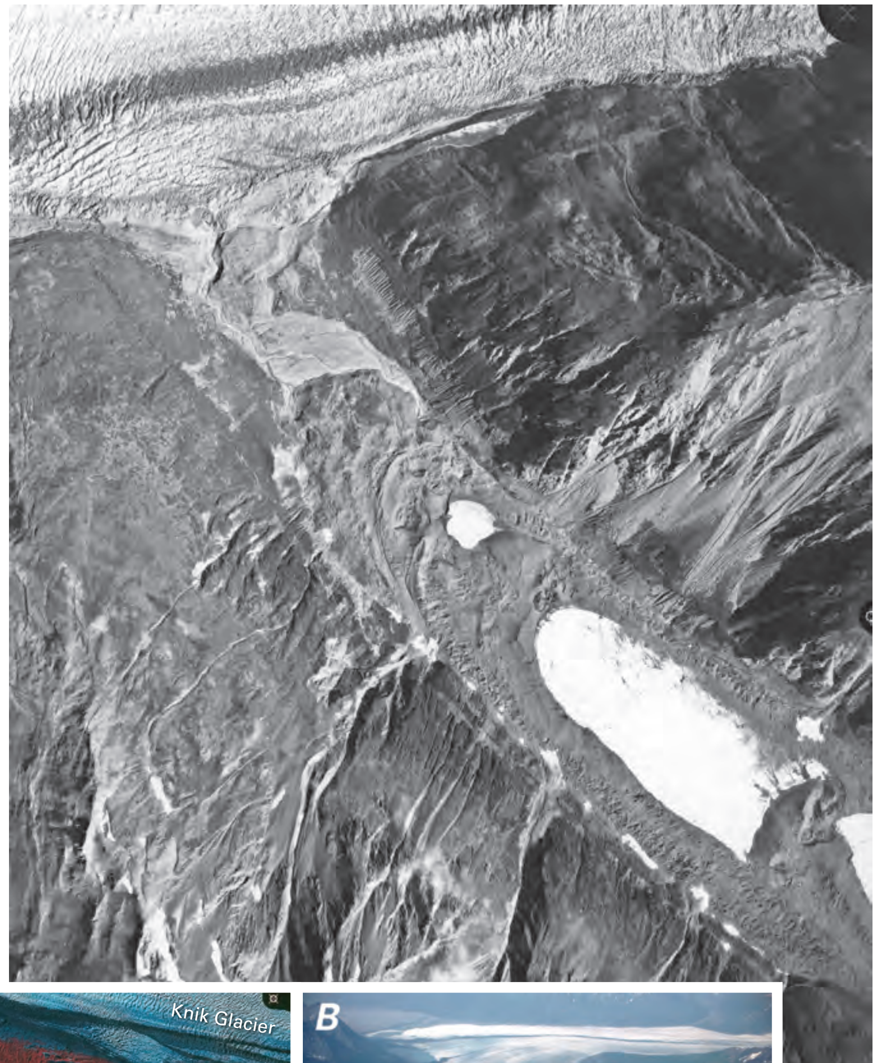
Gannett Glacier, named because of its proximity to Mount Gannett, has a length of 14 km and an area of 24 km^2 (Field, 1975b, p. 475). It flows in a northwestwardly direction toward Knik Glacier but ends more than 2 km short of merging with it. According to Field (1975b), relatively fresh trimlines seen in the valley of Gannett Glacier on Bradford Washburn's 1938 photograph of these two glaciers suggests that they were formerly connected. On the USGS Anchorage B-7 1:63,360-scale topographic quadrangle map (1960) (appendix B), which was based on mid-1950s photography, the terminus of Gannett Glacier was about 500 m from the margin of Knik Glacier. By 17 August 1999, the closest point of the debris-covered terminus of Gannett Glacier was about 2.0 km from Knik Glacier, and exposed ice was about 3.9 km from Knik Glacier (fig. 286).

Colony Glacier

Colony Glacier, which has a length of 29 km and an area of 237 km^2 (Field, 1975b, p. 474), fronts on and calves into ice-marginal Inner Lake George. The northern side of the lake, which has an approximate diameter of 6 km, is dammed by Low Ridge, a subcircular terminal moraine that formed around the bulbous terminus of a formerly extended Colony Glacier. Bradley and others (1972) stated that this moraine has an age of less than 200 radiocarbon years. The location of the northern part of Low Ridge is about 1.2 km south of the southernmost terminal moraine previously deposited by Knik Glacier. This interglacier area between moraines is the bed of Lower Lake George, which has been essentially dry since at least the middle 1950s, at which time the terminus of Colony Glacier had retreated approximately 5 km from Low Ridge and had thinned about 150 m. In 1951, Stone (1955) examined the moraine area and concluded that the last 3.5 km of retreat had occurred in the previous 35 years (since 1916).

The termini of Colony and Knik Glaciers were photographed by the AHAP Program on 25 August 1978 (fig. 287*A*). By that time, the terminus of Colony Glacier had retreated a maximum of 2 km from its middle 1950s position. The author visited Colony Glacier in 1998 and 1999 and photographed it from the air on 15 August 2000 (fig. 287*B*) and in July 2001. Between August 1978 and July 2001, the glacier retreated at least 1.5 km.

Figure 286.—*17 August 1999 vertical aerial photograph of the retreating and thinning terminus of Gannett Glacier, a former tributary to the Knik Glacier. Landslide debris covers much of the terminus. North is at the top of the photograph. Photograph no. R3–FL1–FR–219 from the U.S. Bureau of Land Management.*



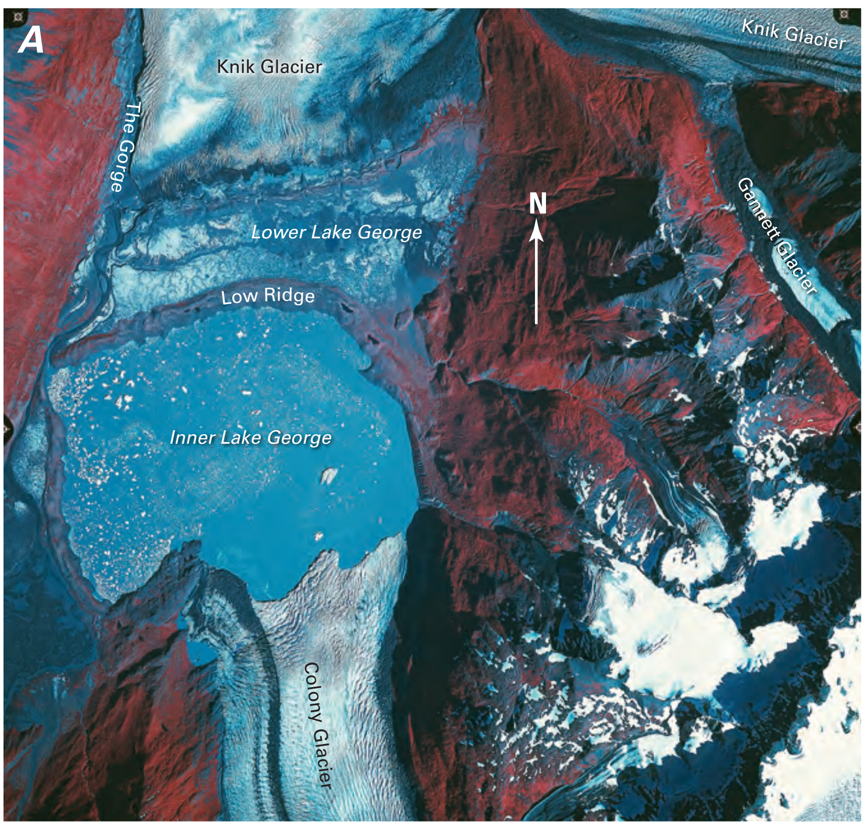


Figure 287.—*Two aerial photographs of Colony Glacier and the surrounding area.* **A**, *25 August 1978 AHAP false-color infrared, vertical aerial photograph of the terminus of Colony Glacier, Inner Lake George, the dry bed of Lower Lake George, the southwestern part of Knik Glacier, The Gorge, and the terminus of Gannett Glacier. The bedrock projection that is in contact with the debris-covered western terminus of Colony Glacier is Colony Point. AHAP photograph no. L107F7031 from the GeoData Center, Geophysical Institute, University of Alaska, Fairbanks, Alaska.* **B**, *15 August 2000 high-altitude north-looking oblique aerial photograph of the terminus of Colony Glacier and Inner Lake George. In the 22 years between photographs, Colony Glacier retreated at least 1.5 km. Photograph by Bruce F. Molnia, U.S. Geological Survey. A larger version of* B *is available online.*

Lake George Glacier

The western part of Low Ridge serves as the dam of the presently dry Upper Lake George, a former proglacial lake bed that is approximately 15 km long and extends from near the terminus of the retreating Lake George Glacier to the western side of Inner Lake George. Lake George Glacier has a length of 24 km and an area of 88 km^2 (Field, 1975b, p. 474). Field (1975b) reported that, in 1957, the terminus of Lake George Glacier was located about 800 m behind a moraine representing an undated recent maximum. When the author observed the terminus in 1998 and 1999, it had retreated another 2.5 km from its 1957 location, and a triangular-shaped ice-marginal lake had formed in front of the terminus.

Glaciers West of Knik Glacier, Upper Lake George, and Lake George Glacier

West of a line connecting Knik Glacier, Upper Lake George, and Lake George Glacier are a number of small unnamed glaciers and a dozen more with names, including Hunter Creek Glacier (fig. 288). All are retreating, and all were observed from the air by the author in August 2000. An unnamed glacier at the head of Troublesome Creek and Whiteout Glacier are two

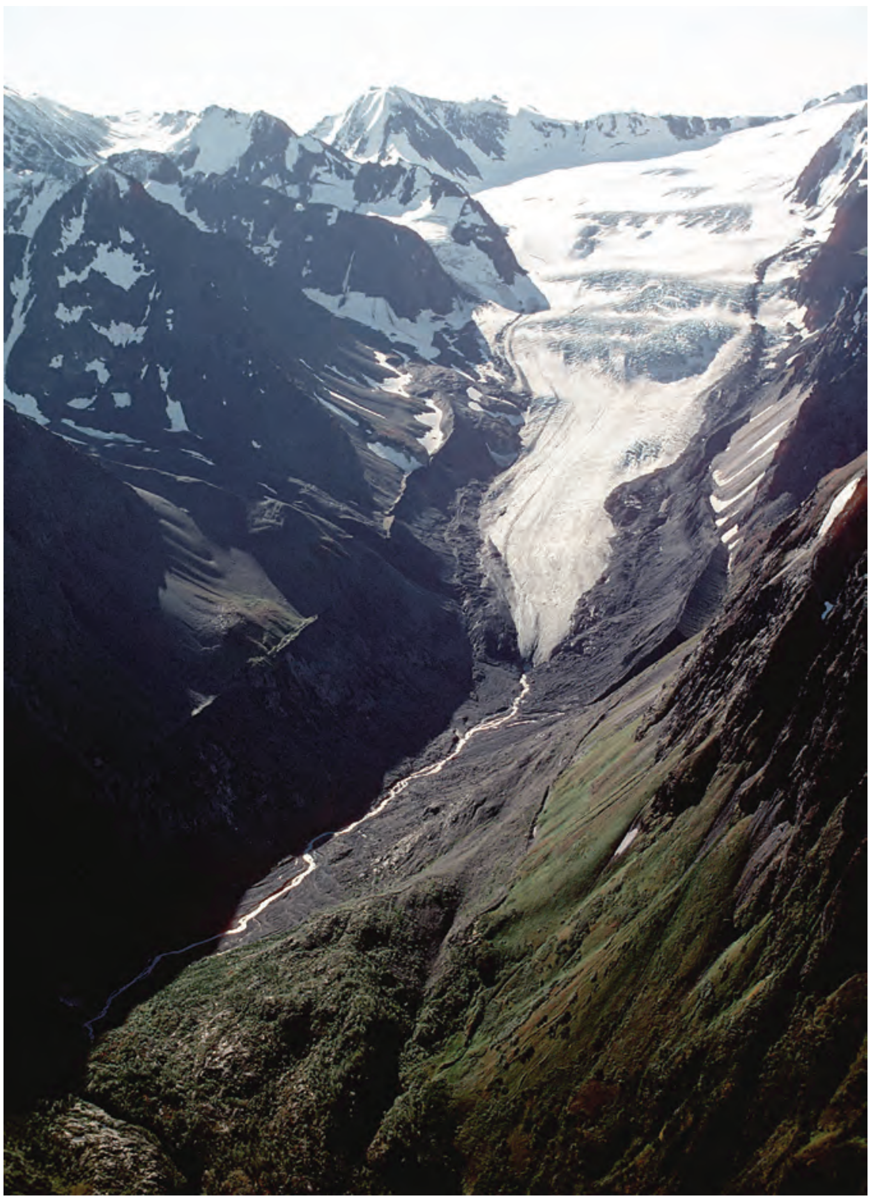


Figure 288.*—30 August 2000 oblique aerial photograph of Hunter Creek Glacier made during an aerial survey of the western Chugach Mountains area west of a line connecting Lake George Glacier and Knik Glacier. All of the glaciers observed were retreating. The south-looking view of Hunter Creek Glacier shows its terminus is approximately 2 km from the terminal moraines that mark its most recent, undated advances. Elevated lateral moraines and trimlines document that this glacier was recently much thicker. Photograph by Bruce F. Molnia, U.S. Geological Survey.*

eastward-flowing glaciers that drain into the bed of Upper Lake George. In 1957, both the unnamed glacier and Hunter Creek Glacier were about 1.5 km from terminal moraines that marked their most recent undated advances. By 30 August 2000 (fig. 288), both glaciers had retreated approximately another 1.0 to 1.5 km upvalley.

Glaciers of the Eagle River and Crow Creek Drainages

Milk, Crow, Raven, Clear, Flute, Icicle, and Organ Glaciers are located along the drainage of Eagle River or Crow Creek.

Eagle Glacier

Eagle Glacier at the head of the Eagle River is the largest glacier in the drainage. Its retreating terminus continues to expose bedrock as it thins and narrows. At the foot of its valley is a former ice-marginal lake with an outwash fan delta at its head. The lake is the product of post-1915 retreat. In 1915, Capps (1916) photographed (see USGS Photo Library photograph Capps 732) and described the terminus area of Eagle Glacier. He reported that several crescentic lines of terminal moraines were located in the valley below the glacier and that the glacier had retreated more than 1 km in the recent past. His photograph shows that an ice-marginal lake of unknown size was located adjacent to the terminus. According to Field (1975b), the location of the terminus in 1915 corresponds to the distal end of the present lake. The most distal moraine, probably corresponding to the "Little Ice Age" maximum position of Eagle Glacier is about 1.5 km beyond the 1915 moraine. A second younger large undated moraine is located about 400 m closer to the glacier. A 1957 photograph shows the terminus position at that time to be about 2.9 km from the outermost moraine. Between 1915 and 1931, Eagle Glacier retreated about 225 m. By 1938, when Bradford Washburn photographed the glacier, another 175 m of retreat had occurred, placing the terminus of the glacier in the emerging lake basin. Another 750 m of retreat occurred in the 12 years between 1938 and 1950. By 1957, Eagle Glacier had retreated another 250 m and receded above the proximal end of the lake, ending on an outwash plain. When the author photographed it from the air on 30 August 2000 (fig. 289), the retreating and thinning terminus was more than 2 km from the distal end of the lake.

Figure 289.*—30 August 2000 oblique aerial photograph shows the location of the retreating terminus of Eagle Glacier with a delta at the end of the 2-km-long lake. The terminus is more than 2 km from the end of the lake where it was located in 1915. Photograph by Bruce F. Molnia, U.S. Geological Survey.*

Eklutna Glacier

Eklutna Glacier, which covers about 50 percent of the West Fork of the Eklutna Lake drainage basin, is 13 km long and has an area of 13 km^2 (Field, 1975b, p. 473). When it was mapped by the USGS in 1957 (Anchorage B–6 1:63,360-scale) (appendix B), the terminus was located at an elevation of about 350 m. Thirty-one years later, during field observations made in 1988 by Brabets (1993), the terminus had retreated to an elevation of about 700 m. The East Fork Eklutna Creek basin has an area of about 100 km^2, of which 20 percent is glacierized. The West Fork Eklutna Creek basin has an area of approximately 67 km^2, about 50 percent of which is glacierized. As with Eagle Glacier, Eklutna Glacier was visited and photographed by Capps in 1915 (see USGS Photo Library photograph Capps 698). At that time, its terminus was situated at the head of an outwash plain about 2 km above the proximal end of Eklutna Lake. The lake occupies an elongated, glacially steepened depression dammed by a older terminal moraine of Eklutna Glacier, produced by a pre-19th century advance of the glacier. Field (1975b) stated that the glacier retreated about 1 km between 1915 and 1957, for an average rate of about 24 m a^{-1}. Eklutna Glacier was photographed by the BLM on 17 August 1999 (BLM vertical aerial photograph nos. R3–FR–272 and R3–FR–274). When the author observed it from the air in August 1996 and again on 30 August 2000 (fig. 290), the retreating and thinning terminus was about 7 km from the head of Eklutna Lake.

Figure 290.*—30 August 2000 oblique aerial photograph of the location of the retreating terminus of Eklutna Glacier. The glacier has retreated approximately 5 km since photographed in 1915. Photograph by Bruce F. Molnia, U.S. Geological Survey. A larger version of this figure is available online.*

Twentymile Glacier

Twentymile Glacier is a southwest-flowing retreating glacier that is the primary source of the Glacier River, which is the eastern source of Twentymile River. At the end of the 20th century, the terminus of Twentymile Glacier calved icebergs into a large ice-marginal lake. Its southeastern side and its northwestern side—each the product of different tributaries—have retreated at different speeds throughout its observed history. Twentymile Glacier has received little scientific attention, but it has been photographed and observed from the air since 1938. Field (1975b) examined photographs collected at six different times (1938, 1941, 1950, 1957, 1964, 1971) and derived a history of recent changes in Twentymile Glacier. When Bradford Washburn photographed it in 1938, the terminus of Twentymile Glacier was about 1.2 km from a prominent undated terminal moraine. This moraine was about 2.5 km upvalley from an even older terminal moraine, located adjacent to the south side of Carmen Valley, which Field (1975b) speculated dates from approximately 1870 or earlier. The 1938 position of the terminus of Twentymile Glacier is separated from dense vegetation by a 200-m-wide barren zone. This zone suggests that, sometime after 1920, the glacier resumed retreat after a period of perhaps several decades of stability. An ice-marginal lake was also beginning to form adjacent to the southeastern part of the terminus. By 1941, the lake fronted the entire ice face, and the terminus continued to retreat.

Between 1938 and 1941, the southeastern side of Twentymile Glacier retreated 150 m, while the northwestern side of the glacier retreated 100 m. The average annual rates of retreat were 50 m a^{-1} and 33 m a^{-1}, respectively. Between 1941 and 1950, the southeastern side retreated 550 m, while the northwestern side retreated 100 m. The average annual rates of retreat were 61 m a^{-1} and 11 m a^{-1}, respectively. Between 1950 and 1957, the southeastern side retreated 350 m, while the northwestern side retreated 400 m. The average annual rates of retreat were 50 m a^{-1} and 57 m a^{-1}, respectively. Between 1957 and 1964, the southeastern side retreated 300 m, while the northwestern side retreated 100 m. The average rates of retreat were 43 m a^{-1} and 14 m a^{-1}, respectively. Between 1964 and 1971, the southeastern side retreated 300 m, while the northwestern side retreated 200 m. The average rates of retreat were 43 m a^{-1} and 28 m a^{-1}, respectively. For the entire 33-year-period

Figure 291.—18 July 1978 north-looking oblique aerial photograph of the thinning and retreating terminus of Twentymile Glacier. Note that the proglacial lake has a length of approximately 2 .5 km and that the northwestern side of the glacier extends several hundred meters beyond the southeastern side. Photograph by Bruce F. Molnia, U.S. Geological Survey.

from 1938 to 1971, the southeastern side of the glacier retreated 1,650 m, while the northwestern side retreated 900 m. The average rates of retreat were 50 m a^{-1} and 27 m a^{-1}, respectively. By 18 July 1978, the lake was about 2.5 km long (fig. 291). By 2000, when it was last observed from the air by the author, the lake had a maximum length of about 4 km.

Summary

During the entire period of the Landsat baseline (1972–1981), Meares and Harvard Glaciers were advancing. Bryn Mawr, Harriman, and Columbia Glaciers advanced during the early part of the baseline. Smith Glacier was stable for most of the baseline period, with the position of its termini fluctuating from year to year. Available evidence suggests that all other valley and outlet glaciers in the Chugach Mountains were thinning and retreating.

At the end of the 20th century, Meares and Harvard Glaciers were still advancing. The terminus of Harriman Glacier was stable. All other valley and outlet glaciers in the Chugach Mountains were thinning or retreating or had become stagnant.