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Denny M. Capps,1 Gregory C. Wiles,2 John J. Clague1 and Brian H. Luckman3

Abstract
We utilized dendrochronology and precise elevation-constrained mapping to date glacially overridden and drowned trees at the margin of Brady Glacier in southeast Alaska. This technique allowed determination of the timing of the former tidewater glacier’s last advance and consequent formation and filling of two marginal lakes. The subfossil tree-ring chronology spans the interval from ad 1370 to 1861. Brady Glacier impounded Spur Lake to an elevation of 83 m a.s.l. around 1830 and 121 m a.s.l. around 1839. Soon after, Spur Lake reached 125 m a.s.l. and began to overflow a stable bedrock sill. The glacier continued to advance, thickening by at least 77 m between c. 1844 and 1859 at a site down-glacier of Spur Lake on the opposite glacier margin. Farther down-glacier, North Trick Lake began to form by 1861 and reached its highest elevation at approximately 130 m a.s.l. when Brady Glacier reached its maximum extent around 1880. Our findings add precision to the chronology of the last advance of Brady Glacier and provide insight into the evolution of glacier-dammed lakes and calving glaciers.

Keywords
Alaska, Brady Glacier, dendrochronology, glacier-dammed lake, glacier fluctuation, tidewater glacier

Introduction
Brady Glacier is the largest glacier in the Fairweather Range of the St Elias Mountain in southeast Alaska. Recent downwasting has exposed well preserved subfossil trees that were overridden by the glacier or inundated by two lakes dammed at the glacier margin. Previous researchers have used dendrochronology to date glacier advances (Lawrence, 1950; Luckman, 1995; Wiles et al., 1999) and the formation of glacier-dammed lakes (Masiokas et al., 2010). In this study, we combine dendrochronologically established death dates with differential GPS mapping of these subfossil trees to reconstruct the last advance of Brady Glacier and the filling of the lakes.

Wastage of Alaskan glaciers accounts for the largest measured glacial contribution to recent sea level changes (Arendt et al., 2002). Over 75% of the ice volume lost from glaciers in southeast Alaska and adjacent Canada comes from tidewater and lake-calving glaciers (Larsen et al., 2007). Calving glaciers are susceptible to nonlinear dynamic instability when they terminate in water that is relatively deep compared with ice thickness. Studies that provide insight into calving glacier dynamics are therefore particularly important for forecasting future sea level rise (Meier et al., 2007). Brady Glacier was formerly a tidewater glacier and it may become one in the future because it is presently grounded well below sea level. Seven distributary lobes of Brady Glacier also have lake-calving margins that exceed 1 km in length. This study provides insight into the evolution of glacier-dammed lakes, the future contribution of Brady Glacier to sea level change, and the dynamics of other calving glaciers.

Study area
Brady Glacier is located in Glacier Bay National Park, 125 km west of Juneau, Alaska (Figure 1). It is 51 km long and 590 km² in area. The glacier has an accumulation area ratio of 0.65, with an approximate equilibrium line altitude (ELA) of 610 m a.s.l. (Viens, 1995). The glacier’s major source areas are mountain peaks up to 3467 m a.s.l. to the northwest and the glacier terminates at approximately 10 m a.s.l. to the northwest and the glacier terminates at approximately 10 m a.s.l. on a large outwash plain that extends 4.8 km into Taylor Bay. The glacier occupies a north-northwest-trending, fault-controlled valley 64 km long, that extends from Taylor Bay in the south to near the north end of Glacier Bay (Derksen, 1976). Ice from the Fairweather Range feeds glaciers flowing both south-southeast towards Taylor Bay (Brady Glacier) and north-northwest to Glacier Bay via Lamplugh and Reid glaciers (Bengtson, 1962; Derksen, 1976). The divide between south- and north-flowing...
ice is at approximately 820 m a.s.l. based on the digital elevation model (DEM) from the 2000 Shuttle Radar Topography Mission (SRTM). Ice-penetrating radar measurements near the axis of Brady Glacier indicate that the bed is at least 200 m below sea level (Barnes and Watts, 1977). Previous dendrochronology studies determined that Brady Glacier reached its current position around 1880 and observations indicate the terminus position has subsequently varied by only a few hundred meters (Bengtson, 1962; Derksen, 1976; Klotz, 1899). Field observations of the highest 'Little Ice Age' trimline near North Trick Lake indicate that the total lowering of the glacier surface in that area is in excess of 135 m.

Presently, the glacier dams at least ten large lakes, six of which are primarily subaerial and four subglacial. These lakes, and other smaller ones, are in different stages of evolution – incipient, stable and non-draining, and periodically draining. This study focuses on North Trick and Spur lakes (Figure 2). Brady Glacier dams North Trick and adjacent South Trick lakes on its southwest side. In the past, North Trick Lake drained subglacially into South Trick Lake and then through either a stable outlet to the southwest into the ocean via Annoksek Creek or subglacially to the terminus (Figure 2) (Derksen, 1976; Post and Mayo, 1971). A new body of water, informally named East Trick Lake, has formed as the glacier has retreated through calving (Figure 2). In August 2007, North and South Trick lakes were dammed at approximately 30 m a.s.l. by moraines and East Trick Lake was glacier-dammed. However, East Trick Lake's surface elevation varies through time. For example, in a 15 August 2010 Landsat image, East Trick and North Trick Lake were connected by a narrow channel through the moraine, but South Trick Lake remained isolated. Brady Glacier dams Spur Lake on its southeast side (Figure 2). At its maximum level, this lake drains over a bedrock sill to the southeast into Dundas Bay via the Oscar River (informal name). Presently, Spur Lake does not fill to this level, but drains catastrophically along a subglacial path to the outwash plain to the south.

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**Figure 1.** Brady Glacier and its surrounding area (after National Park Service, 2009)
The dominant tree species in the area are Sitka spruce (*Picea sitchensis*), western hemlock (*Tsuga heterophylla*), and mountain hemlock (*Tsuga mertensiana*). Sitka spruce and western hemlock are common at low to middle elevations, whereas mountain hemlock is largely restricted to the subalpine zone (Pojar and MacKinnon, 1994).

**Previous studies at Brady Glacier**

Derksen (1976) summarized the late Holocene history of Brady Glacier based on observations dating back to the 1700s and radiocarbon ages from subfossil wood in moraines beyond the maximum extent of the last advance. He postulated an earlier advance...
beginning no later than 1870 cal. yr BP and culminating 1170–1290 cal. yr BP. The glacier receded at least 24 km north of its present terminus by 645–725 cal. yr BP, but began to advance shortly thereafter based on radiocarbon ages from subfossil wood in moraines (Bengtson, 1962; Derksen, 1976).

Historical observations of the glacier terminus before the twentieth century are subject to interpretation (Derksen, 1976). Joseph Whidbey, Captain George Vancouver’s lieutenant, made the first recorded observations of the terminus in stormy weather on 10 July 1794. The chart produced from his observations roughly reproduces the coastline (Vancouver, 1844), but the perspective is distorted. Klotz (1899) reviewed Vancouver’s map and observations and placed the terminus approximately 12 km north of its present location (Figure 2). Later Derksen (1976) disputed Klotz’s interpretation and placed the 1794 terminus near its present location (Figure 2). The next recorded observations are based on a map produced by Mikhail Tebenkof in 1849 (Tebenkof, 1981). Davidson (1904) reproduced this map and placed the terminus north of the present margin, but not as far north as Klotz’s interpretation of the 1794 terminus (Figure 2). In 1880 John Muir and Samuel Hall Young visited the area. Muir explored lower Brady Glacier and made notes, but inclement weather limited his view. Muir did note that the glacier was advancing, and he commented on the presence of Spur and North Deception lakes (Figure 2). He did not, however, say anything about the location of the glacier terminus. Young, who stayed near camp while Muir explored the glacier, talked with a Hoohn chief camped at the terminus. The chief complained that, ‘The icy mass had been for several years traveling towards the sea at the rate of at least a mile every year’. A few years later, the chief reported that the glacier had begun to retreat (Young, 1915). The changing glacier and inconsistent identification of landforms and reference points complicate interpretation of these early observations.

The terminus location has changed little from the time of the first detailed map in the late nineteenth century. A 1:160 000-scale topographic map of the glacier was made based on survey observations from several summits in 1894 (Klotz, 1899). On this map, the terminus is near its present location and North Trick, South Trick, and Spur lakes are clearly indicated. Subsequent maps from 1907, 1948, 1961, 1971, and 1982 and Landsat imagery from the 1990s and 2000s show that the location of the terminus has varied no more than a few hundred meters from the position it occupied in 1894.

Brady Glacier was previously a tidewater glacier. Several early explorers noted its change from a tidewater glacier to one terminating on an outwash plain. Vancouver, in his account of the 1794 exploration, wrote that, ‘... further progress was now stopped by an immense body of compact perpendicular ice, extending from shore to shore’, which indicates a tidewater terminus. By 1880, however, Muir commented, ‘No icebergs are discharged from it, as it is separated from the water of the fiord at high tide by a low, smooth mass of outspread, overswept moraine material ... The front of the glacier, like all those which do not discharge icebergs, is rounded like a brow, smooth-looking in general views ...’ (Muir, 1915). Therefore, sometime between 1794 and 1880 the terminus evolved from a calving margin into a non-calving margin. Post and Motyka (1995) described how Taku Glacier, 25 km northeast of Juneau, evolved in much the same way. When the first detailed map of that area was created in 1890, Taku Glacier was calving into a tidal basin that it had overdeepened during its last advance. By 1931, a morainal shawl had formed that substantially reduced mass loss through calving. The glacier’s rate of advance increased from about 60 m/yr initially to 150 m/yr during 1929–1937. As Taku Glacier continued to advance onto the tidal flat, its terminus spread out and became convex in map view (Post and Motyka, 1995). Brady Glacier’s calving margin was probably similar to that of the Taku in the late 1800s and other present tidewater glacier margins, i.e. perpendicular to or nearly perpendicular to the margins of the fjord in map view. As Brady Glacier built up a stable subaerial shoal and calving diminished, its margin probably became increasingly convex until it resembled the present terminus (Figure 2). The bathymetry of the previous fjord below the shoal is unknown, although it likely affected this evolution. Since Muir’s observations in 1880, the outwash plain has grown and now extends almost 5 km beyond the present terminus of the glacier.

Sample sites and chronology development

We sampled rooted subfossil trunks of glacially overridden trees at the glacier margin and inundated trees at North Trick and Spur lakes in August 2007. The overridden trees, exposed by glacier downwasting in recent decades (Figure 3a), are located approximately 500 m northeast of North Trick Lake above the present glacier margin. Most of the inundated trees are located in the water near the west shore of North Trick Lake (Figure 3b) and on the southeast shore of Spur Lake (Figure 3c). We collected increment cores or discs from rooted subfossil trees at a range of elevations at all three study sites. Samples were collected as low as possible on the trunks and from the least decomposed wood. The location and elevation of the base of each sampled tree were determined with differential GPS. At North Trick Lake we collected samples from partially submerged, erect trunks directly above the water surface and determined the elevation of the samples above the tree base using a weighted measuring tape. Where available, we selectively sampled trees with preserved bark to ensure recovery of the outermost ring. Very few trees had bark. If no bark was present, the sample was classified as possibly lacking outer rings. However, trees that obviously had abraded or rotten outer layers were not sampled.

Cores and discs were prepared for analysis and crossdated using standard dendrochronological techniques (Stokes and Smiley, 1968). We measured ring widths to the nearest micron using a Velmex measuring system and developed a floating ring-width chronology from all ring-width series using COFECHA (Grissino-Mayer, 2001; Holmes, 1983). Calendar dating of the floating tree-ring chronology was accomplished by comparison with unpublished tree-ring chronologies of living mountain and western hemlock from sites on Excursion Ridge and the Bearrack Mountains, 65 and 55 km to the east (Figure 1), both within Glacier Bay National Park (Wiles, unpublished data, 2006). Dating was confirmed by visual comparison of marker rings (Stokes and Smiley, 1968). Samples are archived at Simon Fraser University’s Centre for Natural Hazard Research.

Results

The final chronology is based on radii from 35 subfossil trees at the glacier margin, nine adjacent to Spur Lake, and four in North Trick Lake. Preservation of subfossil wood ranged from good at the glacier margin to poor at North Trick Lake, where no trees had preserved bark. The tree-ring series from all three sites provide a
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composite tree-ring chronology spanning 492 years, from AD 1370 to 1861. Correlation of the Brady Glacier chronology with the Excursion Ridge ($R = 0.66$, $n = 344$) and Beartrack ($R = 0.54$, $n = 347$) master chronologies is significant well above the 99% confidence level (Figure 4). The oldest and longest-lived tree grew at 144 m a.s.l. at the glacier margin site, died around 1839 and was at least 469 years of age. Therefore, this glacier margin site had been ice-free since at least 1370. We obtained the pith in only a few samples, thus the innermost sampled ring only provides a minimum date for establishment of the tree in most cases. Few sampled trees had surviving bark; consequently the outermost sampled ring only provides a limiting date for the death of the tree. The limiting outer ring dates of individual samples at Spur Lake range from 1830 at 83 m a.s.l. to 1839 at 121 m a.s.l., at the glacier margin site they range from 1844 at 125 m a.s.l. to 1859 at 202 m a.s.l., and at North Trick Lake they range from 1817 at 16 m a.s.l. to 1861 at 10 m a.s.l. (Figure 5).

**Discussion**

**Chronology of last glacier advance and lake damming**

Limiting kill dates of the trees allow us to make inferences about the vertical and horizontal extent of the glacier through time. The trees in the Spur and North Trick lake basins likely were killed when they were inundated by the water dammed by the advancing glacier. Given that glacier-dammed lakes typically drain before lake levels reach 90% of the height of the dammed ice (Thorarinsson, 1939; Tweed and Russell, 1999), we assume that the glacier would need to be at least 10% higher in elevation than the associated lake level. The earliest outer ring date related to the most recent advance of Brady Glacier comes from Spur Lake. The lowest sampled tree at 83 m a.s.l. has a limiting date of 1830. To dam the lake to this elevation, the glacier would need to be at least 92 m in elevation at the southern ice-margin of the basin at that time. The highest sampled tree in the Spur Lake basin at 121 m a.s.l. has a limiting date of 1839. This elevation is only 4 m below the stable outlet at 125 m a.s.l. To dam the lake to 121 m a.s.l., the glacier must have been at least 134 m in elevation at the southern ice-margin of the basin. This equates to a thickening of 42 m between 1830 and 1839.

We can calculate horizontal glacier advance from vertical thickening data if we assume a constant glacier gradient. The average gradient of Brady Glacier below the divide separating north- and south-flowing ice on the International Boundary Commission’s 1907 topographic map is 3.9° (International Boundary Commission, 1923). Assuming the gradient was the same in the 1830s, an increase in thickness of 42 m would correspond to a horizontal advance of approximately 600 m. This advance, while rapid compared to many glacier advances, is substantially slower than the rate of the Taku Glacier advance of 150 m/yr from 1929 to 1937 (Post and Motyka, 1995).

Lake level data provide an estimate of the elevation of the ice margin at a given time but it is more difficult to estimate the state of the terminus in the main valley. Most calving glaciers have steep, near-vertical to vertical faces (Molnia, 2008); therefore we can safely assume that Brady Glacier terminated in a near-vertical front several tens of meters high when it was advancing into the sea. For example, the calving front of

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**Figure 3.** Sample sites: (a) glacier margin; view west (person circled for scale). (b) North Trick Lake; view east with high strandlines denoted by arrow and red dotted line. (c) Spur Lake; view northwest. Photographs taken from viewpoints indicated on Figure 2.
Hubbard Glacier, 230 km northwest of Brady Glacier, is over 100 m high, 8 km wide, and reaches 125 m a.s.l. within a few hundred meters of the terminus. The calving fronts of Johns Hopkins and Margerie glaciers, 25 and 35 km northwest of Brady Glacier, are over 75 m high, 1.5 and 2.0 km wide, and also reach elevations of 125 m a.s.l. within a few hundred meters of their termini. If Brady Glacier, approximately 4.5 km wide at the study sites, had a 125 m high calving front when it dammed Spur Lake in 1830, it would quickly raise the level of the lake to 83 m a.s.l. or more. If Brady Glacier was not calving and had a less steep front, several years could have passed between the initial damming of the lake and it eventually reaching 83 m a.s.l. We know from observations cited in the introduction that the glacier was calving in 1794 but not in 1880. Thus, we assume that the terminus was at an intermediate stage between these two states when it reached Spur Lake. We infer that a shoal had begun to form, which limited mass loss through calving, and the glacier front became slightly convex in map view (Figure 6).

The glacier margin site provides evidence for continued thickening and advance of Brady Glacier after the formation and filling of Spur Lake. The limiting date derived from the lowest sample at 125 m a.s.l. is 1844. If the calving front was 125 m high when it reached the site, it would have killed the tree on arrival in 1844. However, if the glacier was not calving and had a less steep front, several years would have passed between the arrival of the ice front at sea level and the death of the tree at 125 m a.s.l. This date and location are in agreement with Davidson’s reproduction of Tebenkof’s 1849 map of Brady Glacier (Figure 2). The limiting date derived from the highest sample at 202 m a.s.l. is 1859. Again, assuming a glacier gradient of 3.9°, an increase in thickness of 77 m corresponds to a horizontal advance of approximately 1100 m over 15 years. We could not find adequate subfossil wood to sample above 202 m a.s.l., although the trimline marking the maximum advance of the glacier is approximately 58 m higher, at 260 m a.s.l. We therefore conclude that the glacier continued to advance for some time after 1859. No scarred or killed trees were found at the highest trimline to provide an absolute date for this maximum position. As in the case of Spur Lake, we cannot determine the exact location of the glacier front at any particular date, we only can estimate its marginal height and terminus location. We hypothesize that the glacier had further developed its stabilizing shoal and its terminus had become more convex in map view (Figure 6).
The final evidence for the timing of glacier advance comes from North Trick Lake. Although we sampled 17 trees in and around North Trick Lake through a vertical range of over 100 m, we were able to crossdate only four samples. These yielded limiting death dates ranging from 1817 at 16 m a.s.l. to 1861 at 10 m a.s.l. The 1861 sample was best preserved and thus provides the most secure limiting date. Because this critical sample is at 10 m a.s.l., glacier advance past this point would have dammed the lake regardless of the frontal profile. An 1861 death date is consistent with the advance recorded at the glacier margin site and Spur Lake. The lake probably rose to the elevation of the highest lake strandline (approximately 130 m a.s.l.) in two to three decades based on the rate of glacier thickening at the other two sites. Thus, we conclude that the glacier continued to thicken and advance until at least 1880, which is consistent with previous tree-ring dates (Bengtson, 1962; Derksen, 1976).

Comparison with Glacier Bay

The piedmont glacier in Glacier Bay attained its maximum size around 1750, completely filling the bay and extending a short distance beyond its mouth (Connor et al., 2009). About 1780, the glacier began a dramatic retreat. Lamplugh Glacier remained connected to the main ice in Glacier Bay until at least 1894 (Klotz, 1899), but no later than 1906 (Molnia, 2008). Glacier Bay ice had retreated over 100 km by 1925 (Molnia, 2008), which is the greatest historic retreat of a glacier on record. Yet part of this retreat occurred at a time when Brady Glacier was advancing.

The ice divide between Brady and Lamplugh-Reid glaciers was probably highest in the mid- to late-1700s, when Glacier Bay ice was thickest. During its last maximum around 1750, ice in Glacier Bay at the present termini of Lamplugh and Reid glaciers was 1000 m a.s.l., 200 m higher than the present ice divide (Bengtson, 1962). As ice in Glacier Bay thinned, Lamplugh and Reid glaciers also thinned, drawing down ice from Brady Glacier. The ice divide thus moved to the south, drawing more ice to the north and away from Brady Glacier. Comparison of an annotated map by Klotz (1899) and the SRTM DEM from 2000 indicates that the divide between the glaciers has migrated approximately 3 km south in the past century.

The concurrent advance of Brady Glacier and retreat of ice in Glacier Bay seems counterintuitive, especially given that the two were connected via Lamplugh and Reid glaciers, as outlined above. We argue that a change in environment at the terminus of Brady Glacier was more important as a determinant of glacier change than the lowering and migration of the ice divide at its head. Past research has demonstrated that the activity of tidewater glaciers is controlled more by calving dynamics than climate (Motyka and Begét, 1996; Post, 1975). As Brady Glacier constructed a stabilizing outwash plain in the nineteenth century, calving ceased. This decrease in mass loss facilitated the further advance of the glacier. It constructed a morainal embankment, perhaps at a sill, near its present terminus and, from that position, constructed an outwash plain. Brady Glacier has thinned significantly in the twentieth century (Larsen et al., 2007), but has

Figure 6. Inferred terminus of Brady Glacier at three times during the nineteenth century, based on death dates of trees killed by overriding glacier ice and inundated in lakes that were dammed by the advancing glacier. Landsat Thematic Mapper 7 multispectral satellite image – 2000 U.S. National Park Service
maintained and enlarged its outwash plain, preventing the reestablishment of a calving front.

**Differential preservation of subfossil wood**

Preservation of subfossil wood at the three sites differs considerably and does not appear to be related to the age of the wood. The trees in the Spur Lake basin died first and are moderately well preserved; those at the glacier margin site were killed next and are the best preserved. The highest trees in North Trick Lake were the last to die, yet are the most poorly preserved; no cores from these higher trees were dateable because of poor preservation.

Differential local conditions are thought to be responsible for the differential preservation of wood at the three sites. At the glacier margin site, the subfossil wood was covered by ice and sediment from the time that it was overridden until it became exposed during twentieth-century glacier recession. Burial of the trees minimized weathering and erosion. Barclay et al. (2009) describe similar preservation of trees buried by Tebenkof Glacier in southeastern Alaska over 1700 years ago.

Water covered the subfossil wood for differing amounts of time at the two lake sites. Spur Lake filled to near its stable outlet in 1839, or shortly thereafter, and the glacier continued to thicken for approximately 40 years. With a thick glacier dam and a stable outlet, Spur Lake does not appear to have been drained by jökulhlaup until recently. Available evidence from maps, aerial photographs, satellite imagery, and published and unpublished research show Spur Lake filled to overflow prior to 2002. The first evidence of the lake below its stable outlet is a Landsat image acquired in March 2002. Since then, jökulhlaups have periodically drained the lake, the lake has not filled to overflow, and vegetation is becoming established below the highest strandline. Stranded icebergs in the basin in 2007 (Figure 3c) show that jökulhlaups have continued. The lake level defined by the stranded icebergs was only 89 m a.s.l., 36 m below the stable outlet. This history suggests that the subfossil wood we sampled in Spur Lake has only been exposed to air for a short time and thus was relatively well preserved. Clague and Shilts (1993) described similar preservation of trees in two stable landslide-dammed lakes where the trees were preserved for at least 800 years.

Most of the trees in North Trick Lake were killed after 1861 as the glacier continued to advance and raise the lake to successively higher levels. By 1929, the date of the first U.S. Navy aerial photographs, the glacier was downwasting. In these images, vegetation is colonizing the area between the lakeshore and the highest strandline. Aerial photographs taken in 1948 show a lower lake and at least three, progressively less-densely vegetated strandlines above the shoreline. South Trick Lake had drained catastrophically to the terminus before the 1948 photographs were taken, which allowed North Trick Lake to drain to lower levels. Therefore, most of the trees in the basin have been subaerially exposed discontinuously for at least 59 years. Post and Mayo (1971) and Derksen (1976) state that North Trick Lake had a long history of jökulhlaups. The lowest non-glaciated outlet of the basin is approximately 247 m a.s.l. or about 117 m higher than the highest strandline. Without a stable outlet, the lake would fill and drain quasi-periodically throughout its life cycle. The highest and longest exposed subfossil wood at this site was the most decomposed, whereas the trees in the present moraine-dammed lake are the best preserved because they have been exposed to subaerial processes for the shortest time.

In both North Trick and Spur lakes, there is a clear concordance in height of the tops of standing, formerly inundated trees (Figure 3). Most of the inundated trees in Spur Lake are cropped at approximately 125 m a.s.l., and those in North Trick Lake do not extend above approximately 35 m a.s.l. We postulate that the height of these cropped trees results from trimming by winter lake ice at elevations where the lakes were maintained for relatively long periods of time.

**Summary and conclusions**

We developed a tree-ring chronology spanning the period from AD 1370 to 1861 from subfossil wood at three sites bordering Brady Glacier and used it to document the history of glacier fluctuations and glacier-dammed lake evolution. Our data confirm that the study sites were not occupied by the glacier or lakes for at least 450 years prior to 1830. Spur Lake was impounded by the advancing glacier no later than 1830 and had filled to an elevation of approximately 121 m a.s.l. by 1839. The glacier continued to advance, reaching a site about 1.4 km farther south by 1844, and dammed North Trick Lake another 1.1 km down valley by 1861. Brady Glacier did not reach its maximum extent until at least the 1860s, and probably not until 1880. We postulate that this advance was made possible by establishment of a morainal shoal, similar to the behavior observed at Taku Glacier in the 1900s, which caused a major reduction in ablation because of cessation of terminal calving. This study is the first to apply elevation-constrained mapping and dendrochronology to date glacier fluctuations and the evolution of ice-marginal lakes.

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