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On the retreat of lagoon-terminating glaciers: sometimes fresh, sometimes salty

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Complete List of Authors:	Berry, Journey; Oregon State University, College of Earth, Ocean, and Atmospheric Sciences Pettit, Erin; Oregon State University, College of Earth, Ocean, and Atmospheric Science Rennermalm, Asa; Rutgers, Abib, Nicole; University of Alaska Southeast, Department of Natural Sciences Roth, Aurora; University of California San Diego Scripps Institution of Oceanography Kurtz, Deborah; U.S. National Park Service, Kenai Fjords National Park
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On the retreat of lagoon-terminating glaciers: sometimes fresh, sometimes salty

Journey BERRY^{1,2}, Erin PETTIT¹, Åsa K. RENNERMALM², Nicole ABIB^{1,3,5}, Aurora ROTH⁴,
Deb KURTZ⁵,

¹*College of Earth, Ocean, and Atmospheric Sciences, Oregon State University; Corvallis, OR, USA*

²*Rutgers, The State University of New Jersey; New Brunswick, NJ, USA*

³*School of Arts and Sciences - Natural Sciences, University of Alaska Southeast; Juneau, AK, USA*

⁴*Scripps Institution of Oceanography, University of California San Diego; La Jolla, CA, USA*

⁵*Kenai Fjords National Park, National Parks Service; Seward, Alaska, USA*

Correspondence: Journey Berry <berryjou@oregonstate.edu>

ABSTRACT. Draft Date July 9, 2026

Coastal lagoon-terminating glaciers flow into water bodies of variable salinity and temperature, shaped by mixing meltwater, ocean water, and ice. They remain understudied yet increasingly important as coastal glacier margins retreat globally. Lagoon salinity influences circulation through interactions with buoyant subglacial discharge. Water temperature, modulated by ocean connectivity and latent heat processes, controls thermal forcing on the ice. Thus, lagoon circulation, freshwater energetics, and retreat behavior tightly intertwine with lagoon volume and its ocean connection. Bear Glacier, fed by the Harding Icefield in Southcentral Alaska, transitioned to a glacier-lagoon as recently as 2014. Glacial lake outburst floods, tides, and storms regulate the ocean connection, contributing to the glacier's retreat. We document this evolution using historical maps, satellite imagery, local knowledge, and lagoon salinity and temperature measurements. Variable salinity reflects changes in channel flow alongside the morainal spit separating lagoon from ocean, where periodic erosion from outburst floods allows the channel to connect the two water bodies closer. Recent retreat beyond the glacier's narrowing, combined

with upstream overdeepening, saltier water, an eroding spit, and regional warming, suggests continued inland glacier retreat. This lagoon-mediated retreat style is a likely analog for future retreat of other coastal glaciers in Alaska, Patagonia, and Greenland.

INTRODUCTION

As glaciers retreat worldwide, the proglacial environments, including fjords, lakes, lagoons, and river valleys, are undergoing rapid physical change (Malles and others, 2023; Carrivick and others, 2022). In turn, these changes can accelerate or dampen future glacier retreat and ice mass loss depending on local conditions at the glacier terminus (Sutherland and others, 2020; Kochtitzky and others, 2022). Water temperature and salinity, as well as the flux of water into and out of each system via processes like subglacial discharge and glacial lake outburst floods modulate the water environment of the glacier terminus. For all glaciers that end in water, climatic drivers are decoupled from the shorter-term, rapid, and often unstable retreat by calving (Post and others, 2011). Over time, however, climate is crucial to driving glacier thinning, water storage, the evolution of a glacier, as well as freshwater input to proglacial water bodies. The number of proglacial lakes is increasing as climate change continues to cause thinning and melting, aiding the retreat of the world’s deep-bedded coastal glaciers (Zhang and others, 2024).

Most research has focused on tidewater or lake-terminating glaciers, but here we focus on the behavior of glaciers that terminate in lagoons: bodies of water with variable salinity. Tidewater and lake-terminating glaciers behave differently due to different water properties (e.g Motyka and others, 2003; Sugiyama and others, 2016; Vacek and others, 2025), but for glaciers in a lagoon water environment, salinity and other water properties can change over short periods of time. Similar to lake-terminating glaciers, lagoons are enclosed basins such that water and air temperatures have a large impact on the stability and shape of the terminus and tongue, but similar to tidewater glaciers, they can have vigorous circulation due to buoyancy of freshwater plumes in saltwater such that salinity helps determine terminus stability and position via subaqueous melting. A subaerial moraine between overdeepened valley and the ocean, thus, may form a lagoon. The resulting effect of variable properties like salinity is that the glacier may behave like a lake-terminating glacier or a tidewater glacier at different times.

Studies of the few known glaciers that terminate in coastal lagoons were pioneered in Iceland (e.g.

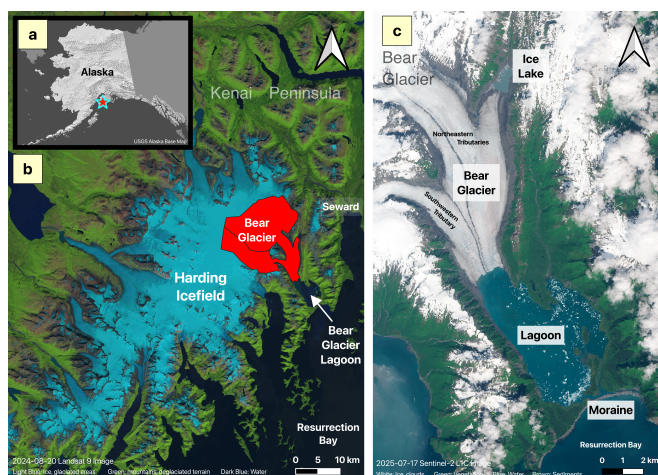


Fig. 1. a) Location of the Harding Icefield on the USGS Alaska base map. b) Landsat 9 satellite image of the Harding Icefield on the Kenai Peninsula, where light blue areas represent glaciated and green areas are ice-free terrain. Dark blue and black colors indicate water. Bear Glacier is identified in red. The town of Seward is located to the northwest of Bear Glacier. c) Fully labeled map of Bear Glacier.

Harris, 1976; Landl and others, 2003). Jökulsárlón is the most well-studied lagoon. Breiðamerkurjökull flows into Jökulsárlón and, as their research has recognized, factors such as bathymetry and lagoon shape, circulation, temperature, salinity, and climate all contribute to feedback processes within a glacier-lagoon system. Jökulsárlón now connects to the ocean through a channel engineered to minimize the chances of bridge washouts.

Here, we present the history and modern dynamics of Bear Glacier and its lagoon (59.95°N , -149.57°E) in Kenai Fjords National Park, Alaska (Figure 1 and 2). Bear Glacier Lagoon is similar in area but deeper than Jökulsárlón. In contrast to Breiðamerkurjökull, which terminates across a broad outwash plain with only part of the glacier flowing into the lagoon, the full width of Bear Glacier flows into its lagoon. Currently, Bear Glacier enters the lagoon through a deep, narrow bedrock valley less than 2 km wide) and the lagoon's connection to the ocean is natural and variable. Bear Glacier, therefore, provides an opportunity to investigate natural glacier-lagoon processes, building on the glacier-lagoon research pioneered in Iceland. We describe the evolution of Bear Glacier from tidewater to land-terminating to lake-terminating and finally, starting in 2014, to lagoon-terminating in the context of local observations since 1888 and regional ice history.

Tidewater and lake-terminating glacier endmembers

Tidewater glaciers are mountain glaciers whose terminus ends in seawater¹. Their fresh subglacial discharge water enters the salty ocean, driving vigorous circulation that entrains warm deep water toward the terminus and stimulates submarine melt and calving (Motyka and others, 2003; Post and others, 2011; Jackson and others, 2019). Though buoyant forces are present, the terminus of tidewater glaciers remains grounded because calving removes ice before the glacier reaches floatation².

Tidewater glaciers are typified by episodic rapid retreats of the glacier terminus. As described by Brinkerhoff and others (2017), the glacier advances by slowly building a morainal shoal. Rapid retreat begins when the terminus is overextended and loses the stability provided by the morainal shoal. Subglacial discharge mixes with warm ocean water to create a buoyant, rigorous plume that erodes the shoal (or moraine if above water), which opens a void between shoal and glacier (Truffer and Motyka, 2016). The new void triggers rapid retreat and mass loss until it is back to its original position. Unlike a lake-calving glacier, the flotation of a temperate tidewater glacier terminus is more unstable and the surface gradient of the terminus may be steeper.

Lake-terminating glaciers, on the other hand, often form from land-terminating glaciers retreating through bedrock overdeepenings; typically with morainal debris capping bedrock highs (as in Himalaya, etc). The presence of a proglacial lake promotes glacial retreat rates that may be over four times that of land-terminating glaciers (Sutherland and others, 2020), and those in Alaska have even shown to change more rapidly than tidewater glaciers (Sugiyama and others, 2016). Lake-terminating glaciers generally have slower calving and less subaqueous ice melt than their tidewater counterpart because of the minimal difference in salinity and temperature between the glacier and lake. Despite overall cooler temperatures and no salt content, proglacial lakes are often stratified. Subglacial discharge controlled by temperature-driven density differences can produce a floating glacier tongue and larger but less frequent tabular calving events (Trüssel and others, 2013; Sugiyama and others, 2016).

For lake-calving glaciers, the proglacial lake is more sensitive to solar heating and wind driven currents than a tidewater glacier fjord (Truffer and Motyka, 2016). A proglacial lake, therefore, has slower circulation patterns because solar surface heating is not enough to sufficiently heat the cold water column for circulation

¹The water temperatures for tidewater glaciers range from near freezing temperature of seawater (< 0°C) in Antarctica to as high as 11°C in Alaska.

²In polar regions, tidewater glaciers may form temporary floating tongues.

to be vigorous. Under these conditions, gently sloping glacial tongues form and can reach great lengths that last long periods of time (Macdonald and others, 2018). With thinning via rising air temperatures, the glacial tongue can break off in infrequent, tabular calving events (Trüssel and others, 2013).

Formation of a lagoon

Between these two end members of proglacial water bodies is a system that has received minimal attention to date: coastal lagoon-terminating glaciers. Because of intermittent connection to the ocean, lagoon salinity may range from fresh to near-seawater, changing on tidal, seasonal, and longer timescales (Gale and others, 2007). Glaciers flowing into lagoons have typically been considered primarily an Icelandic phenomenon (Harris, 1976), but glacier retreat over the last half century has created new lagoon systems. Sitkagi Lagoon, for instance, is a 3 km² proglacial body of water off of Alaska's Malaspina Glacier that now makes the glacier vulnerable to marine water intrusions and thus rapid retreat (Tober and others, 2023). There are no studies on the effects of this recent lagoonal exposure on Malaspina Glacier to date, but the consequences of introducing saltwater to a once-closed freshwater glacier system can be significant. Malaspina Glacier alone, if subjected to rapid retreat from increased contact with marine water, can contribute up to 1.4 mm to sea level rise (Tober and others, 2023).

A terminal moraine left by a retreating glacier can act as a barrier between two bodies of water. Lagoons form when a coastal terminal moraine extends above sea level. A glacier is more likely to build such a subaerial moraine when it terminates in the same location for extended periods of time (Benn and Evans, 2014). This stable terminus location could occur within a narrow section of a fjord or, alternatively, where the fjord opens to a broader bay and the glacier has insufficient ice flux to extend beyond the fjord walls.

A subaerial moraine may also form from uplift of a shallow submarine moraine after glacier retreat. In rapidly deglaciating coastal Greenland, for example, ice-sheet mass loss results in a glacial isostatic adjustment and change in local gravitational pull (together called "sea-level fingerprinting") that will lead to lowering of local sea-surface height, which may elevate shallow moraines above sea level (Mitrovica and others, 2011; Coulson and others, 2022).

The role of glacier lake outburst floods

Glaciers with ice-marginal lakes (like those in Iceland, Patagonia, the Himalayas, and Alaska) are known to be the source of damaging and devastating outburst flood events. Glacial-lake outburst floods occur

when pooled water suddenly drains after fresh glacier melt and terrestrial runoff fills the lake to a threshold depth that causes instability between the glacier ice, bedrock saddle, or moraine containing it (e.g. Post and Mayo, 1971; Fowler, 1999; Kingslake and Ng, 2013). The freshwater source for glacial-lake outburst floods can pool on top of, underneath, or adjacent to the glacier (Kjeldsen and others, 2014). The water then drains through the glacier by way of a subglacial channel. The amount of water in the lake determines the magnitude of the outburst flood, but the same lake may drain with different magnitudes depending on feedback between lake water pressure, ice overburden, local bed topography, and subglacial hydrological system (Roberts, 2005). As Kurtz and Wolken show for Bear Glacier in Alaska, regularly occurring outburst floods can scour downstream rivers or suddenly raise the water levels of a proglacial lake, damage established moraines, and, as we show here, play a key role in the dynamics of a glacier system.

Bear Glacier as a lagoon-terminating glacier

Until now, published research on Bear Glacier Lagoon treats it as a freshwater lake (Yang and others, 2020, 2022; Black and Kurtz, 2023; Wilcox and others, 2013; Larsen and others, 2015). We show here that the transition from a near-fresh to saltwater lagoon likely began in 2014, although it was still fresh in the top 100 m in 2017³. We present here the first measurable salinity greater than 1 g kg⁻¹ in 2023. Local boat captains and tour guides observed ocean water flowing into the lagoon at high tide starting in summer 2019. The lagoon’s connection is mediated through a short channel across the subaerial terminal moraine (discussed further in Section Study Site). The morphology of this channel is heavily influenced by glacial lake outburst flood events. These events occur from the sudden draining of the ice-marginal lake situated higher in the glacial drainage basin of Bear Glacier.

We further show that Bear Glacier has been undergoing more rapid retreat since 2018 (losing 9 km total since 1950, nearly 3 km since 2018), which is likely to continue as it retreats into a broad upstream overdeepening. Given current glacier retreat worldwide, lagoons may become more common as proglacial water bodies interact with seawater.

³These measurements were collected as part of the Girls* in Icy Fjords expedition. A team of nine teenage girls, including author Berry, camped for over a week exploring Bear Glacier Lagoon by kayak and foot; author Pettit led this expedition.

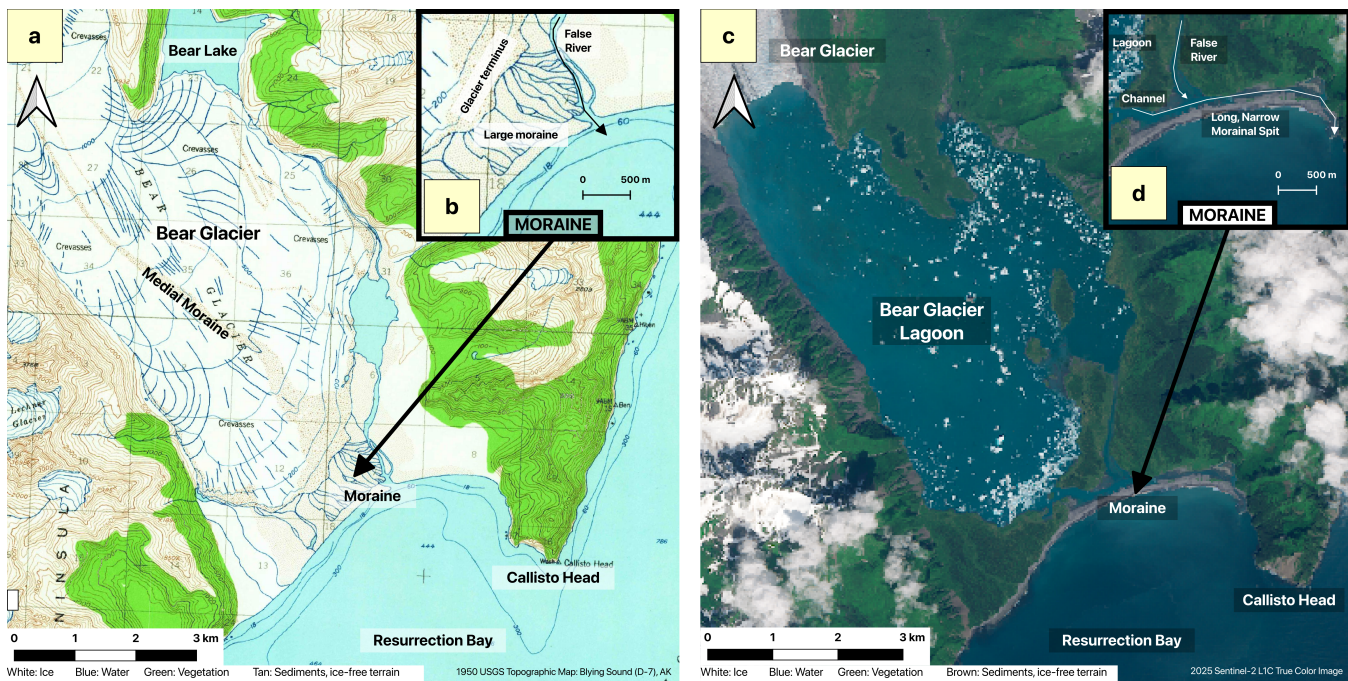


Fig. 2. a) Map of the Bear Glacier terminus from the US Geological Survey topographic map archive (1950). The Medial Moraine describes the tan speckled pattern down glacier. c) Bear Glacier terminus from Sentinel-2 imagery (2025). b, d) Insets showing close up of the morainal spit to highlight primary water routing (b) historically through False River and (d) in present day flowing through the channel parallel to the “repaired” morainal spit.

STUDY AREA

Bear Glacier is located on the eastern side of the Kenai Fjords National Park (location 59.95 °N, -149.57 °E), 19 km south of the town of Seward, Alaska (Figure 1). Bear Glacier flows southeastward from the Harding Icefield and is its largest glacier at 214 km². From this icefield, Bear Glacier is one of the very few outlet glaciers still in contact with a body of water.

The Harding Icefield is one of the largest low-elevation icefields in the United States at less than 1200 m above sea level (Adalgeirsdottir and others, 1998). The 1800 km² icefield has been thinning and losing ice mass since the mid-20th century (VanLooy and others, 2006; Kurtz and Wolken, 2019). Thinning and retreat is attributed to climate warming modulated by the Pacific Decadal Oscillation (Berthier and others, 2010). Harding Icefield mass loss has led to the publicly recognized retreat of Exit Glacier⁴, located due north of Bear Glacier (Kurtz and Baker, 2016). From the mid-1980s to 2014, the icefield lost about 195 km² in total area Yang and others (2020).

Bear Glacier has been calving, thinning, melting, and forming its own proglacial water body since the mid-20th century (Figure 2 and Adalgeirsdottir and others, 1998; Black and Kurtz, 2023). The proglacial water body, Bear Glacier Lagoon, is currently more than 23 km² in area and is separated from Resurrection Bay at its southern end by a narrow morainal spit, roughly 100 meters wide and 3.7 km in length. A river, hereafter referred to as the “channel”, is a modern fluvial formation that flows parallel to the spit and provides an intermittent connection between the southernmost end of the lagoon and Resurrection Bay (sometimes called Blying Sound), and ultimately the Pacific Ocean (Figure 2).

The known source of the glacial lake outburst floods that occasionally drain into Bear Glacier Lagoon comes from an ice-dammed lake, hereafter called “ice lake”⁵ that is located about 10 km north of Bear Glacier Lagoon. The ice lake forms where the glacier flows up an ice-free tributary valley at 275 m above sea level. The shape and location of the ice lake has shifted and its size has increased since 1984 because the glacier and tributary margin has undergone surface thinning and retreat (Wilcox and others, 2013).

In the 21st century, particularly since the 2010s, the morainal spit has been eroded—sometimes enough to breach it—by the glacial lake outburst floods draining and temporarily increasing the water levels in the lagoon. The normal outflow of water is northeast along the channel to the easternmost end of the spit

⁴Kenai Fjords National Park has a popular access point from the Seward Highway that highlights the glacier’s retreat since 1917.

⁵Some locals call this lake “Upper Lake.”

just north of Callisto Head. The breached spit creates a more direct opening for water overflow to leave the lagoon to the ocean (Black and Kurtz, 2023). After a breach, storms and wave action work to repair it in the winter months (Kurtz and Wolken, 2019).

A secondary river locally referred to as False River flows south, perpendicular to the morainal spit and channel. When False River has flowing water, its confluence with the channel is at “T” intersection (Figure 2b). False River does not provide the primary modern connection between lagoon and ocean; it ceases flow at low lagoon water levels, but it is important for understanding Bear Glacier in its historical context.

Published History of Bear Glacier

Bear Glacier once flowed as a tributary into a much larger glacier that advanced down Resurrection Bay and out towards the continental shelf edge during the Last Glacial Maximum, about 15,000 to 13,000 years before the present (Mann and Hamilton, 1995). The earliest documented terminus locations exist as dated moraines, lichen, and tree rings (Wiles and Calkin, 1994). Dated isotopes constrain Bear Glacier’s terminus location to the morainal outwash plain that separated it from the ocean as recently as 1888. The oldest published map and images of Bear Glacier date to July 1909, where surveyors described the glacier as low in elevation, relatively stable, and resting on a gravel flat where high tides reached the terminus center (Grant and Higgins, 1913). We obtained available Seward and Blying Sound 15’ Quadrangle maps circa 1950 from the US Geological Survey to compare to the current moraine morphology (Figure 2). The earliest accounts of lakes are from the 1950s USGS Blying Sound Quad maps. Bear Lake (which expanded to form the current lagoon) formed in the northern end of Bear Glacier terminus; the water from Bear Lake flowed south along the terminus and fed a secondary lake and eventually False River. The upper ice lake was also first documented in the Blying Sound Quad map set (not shown here).

Discussions on Bear Glacier from 1950 to the early 2010s are predominantly in the context of its elevation change, as well as ice mass loss of the entire Harding Icefield. Yang and others (2020) report the Harding Icefield thinned at a rate of $0.61 \pm 0.12 \text{ m a}^{-1}$ during this time, while VanLooy and others (2006) calculate slightly lower icefield thinning rates, with accelerated thinning at $0.72 \pm 0.13 \text{ m a}^{-1}$ starting in the 1990s. By 2014, the icefield lost approximately 12% of its area coverage (Yang and others, 2020). Bear Glacier shows relatively greater elevation change and area loss than the Harding Icefield overall. Adalgeirsdottir and others (1998) find a thinning rate of 0.86 m a^{-1} from 1950 to 1994. VanLooy and others (2006) estimate

an average thinning rate of Bear Glacier to be 0.8 m a^{-1} from 1950 to 1996, then 1 m a^{-1} from 1996-2000, equating to a 10 km^2 area loss.

Nearly all prior studies label Bear Glacier as lake-terminating. Adalgeirsdottir and others (1998), however, categorized this glacier as both tidewater and lake, detailing that it was once a tidewater glacier but terminated in a lagoon at the time of publication. While Bear Glacier reached the ocean in its earliest history, there are no published salinity measurements to distinguish the stated lagoon from a freshwater body of proglacial melt between the Little Ice Age (c.1850) and 2017 (our study).

Wilcox and others (2013) were the first to document glacial lake outburst floods from the ice lake draining into the lagoon. They created a time series of ice lake area changes to reflect the drainage. The first estimated area of 0.18 km^2 was measured in 1985. Satellite imagery and moored pressure transducers continue the area-based outburst flood record up to 2012. We use this record as part of the early outburst flood history.

The Kenai Fjords National Park division of the National Parks Service published information on the outburst floods in 2014 and 2018 (Kurtz and Wolken, 2019). During both flood events, they document that the lagoon exceeded its capacity and excess water eroded (“breached”) the morainal spit. The Kenai Fjords National Park captured the 2018 outburst flood with time lapse, showing the full drainage of the ice lake and icebergs from the calving ice margin left behind.

DATA AND METHODS

Here we integrate available data with new measurements and observations to investigate the dynamics of Bear Glacier. These dynamics include glacier loss, changes in the landscape geomorphology, lagoonal bathymetry, and oceanographic properties. We document the current rapid retreat of Bear Glacier from satellite record. We also connect the satellite observations to measurements of salinity, water and air temperature, as well as bathymetry from measured (via oceanographic instruments) and infer (via satellite imagery) evidence during active periods of change. Per regulations, our field methods and research footprint were limited (e.g., restricted motorized boat use) to minimize the impact on the landscape and visitor experience within Kenai Fjords National Park.

Satellite Imagery

We digitized the terminus of Bear Glacier and the ice lake using Landsat 1-9, Sentinel 2, and Aster optical imagery (Lea, 2018). The first available imagery of the southern Alaskan region is from Landsat 1 in September 1972. The last image we use is from Sentinel-2 in January 2025. Spatial resolution ranges from 80 m/pixel for Landsat 1 and 2 to 10 m/pixel in Sentinel-2 and some Aster images. We trace the ice extent, lagoon shoreline, moraine spit and channel geometry, and ice-lake area from 1972 to early 2025. We set a reference line across the glacier at the narrowest point of the valley walls to allow better comparisons of the terminus changes over time relative to the terminus in the 1950s USGS 15' Quad map. We quantify the length of the glacier using three flow lines: the glacier centerline closely following the medial moraine, a western flow line, and an eastern flow line, because the glacier terminus does not retreat uniformly. Additional data we use to help reconstruct Bear Glacier's history include glacier velocities at the location intersecting the centerline and our terminus reference line (59.9938 °N, -149.6297 °E) from 1985 to early 2025 (Gardner and others, 2025, NASA's MEAsURES ITS LIVE collection,).

Our ice lake area estimates derive from vectors of the ice lake perimeter. The time range of the area estimates span from October 2012 to January 2025, extending the record started by Wilcox and others (2014). We use between 1 and 5 satellite images for each year to create the time series. For certain years we use additional imagery to capture significant lake area changes associated with a glacial lake outburst flood. There is a data gap between 1994 and 1998 due to a lack of satellite imagery during that time as a result of funding issues, poor resolution, and the failed launch of the Landsat 6 satellite. Imagery is more consistent following the launch of Landsat 7 in 1999 and ASTER in 2000.

For the moraine and channel analysis, we first identify whether or not there was a breach in the normally continuous spit, and then measure the length of the channel and its opening width to Resurrection Bay (called here a "breach opening"). The normal pathway is the channel flowing parallel to the full length of the spit, and exiting to the bay where the moraine meets the cliff of the bedrock ridge Callisto Head. We identify a "breach" when the channel cuts through the spit and water flows in a direct path from the lagoon to the ocean. The continuous full length of the spit without the channel cutting through it we refer to as "repaired". Only satellite images where Bear Glacier's features are unobstructed by clouds, image striping, or poor resolution are used. This dataset starts in 1980 to exclude earlier history when the False River was the primary connection to Resurrection Bay (1950-1979).

Lagoon Bathymetry

In August of 2024, we collected bathymetry of the lagoon using the Lowrance Elite Fishing System 7 (FS7) sonar system, a consumer fish finder with frequencies of 83 and 200 kHz for deep bathymetry reading at wider ranges. Due to the restriction on motorized boats entering the channel and lagoon, we traversed by kayak and operated a small remote controlled boat (BlueBoat from BlueROV). The lagoon bottom was auto-detected, meaning the brightest reflection was chosen as the lagoon floor for each sonar profile. Erroneous picks in the auto detection algorithm may come from strong stratification, wildlife like fish and seals, icebergs, or suspended sediment in the water column; we manually removed these obvious outliers along each pathway. We corrected the sounding speed of all profiles to 1427 m s^{-1} based on the lagoon's temperature and salinity at the time of data collection. Uncertainty in the depth is $<0.50\text{ m}$ due primarily to small variations between sound speed profiles. The 2024 bathymetry supplements those collected at Bear Glacier in June 2009 by the University of Cincinnati (Lowell and others, 2010). They collected bathymetry in the southern half of Bear Glacier Lagoon using the Hummingbird 798 SI combo, also a consumer fish finder that uses the same frequencies, attached to a pontoon boat with a remote-controlled motor. Their sampling resolution was approximately 1 m at its maximum setting of 150 m. Their depth was calculated from two-way travel times and assumed that the sound speed is approximately 1500 m s^{-1} . In combining these data with our 2024 data, we corrected this sounding speed to 1414.5 m s^{-1} using the 2017 average temperature and salinity in the lagoon and the assumption that the lagoon was completely (or very close to) fresh before 2017.

Water Salinity and Temperature

For the years 2017, 2023, and 2024, we collected depth profiles of conductivity (salinity), temperature, and depth (CTD). In 2017, we collected a profile in Bear Glacier's lagoon using a Sontek Castaway CTD logger⁶. We lowered the CTD logger by fishing line from the side of a kayak, and the profile ended at the local bottom depth. The Sontek Castaway is limited to 110 m depth and has lower resolution than a research-grade CTD. For profiles in July 2023 and August 2024, we used an RBR Concerto to collect the profiles. We present all profiles and the depth-averaged temperature and salinity here in units of in-situ temperature ($^{\circ}\text{C}$) and absolute salinity (g/kg); however, a full analysis of the temperature and salinity is beyond the scope of this paper.

⁶as part of Girls* in Icy Fjords expedition

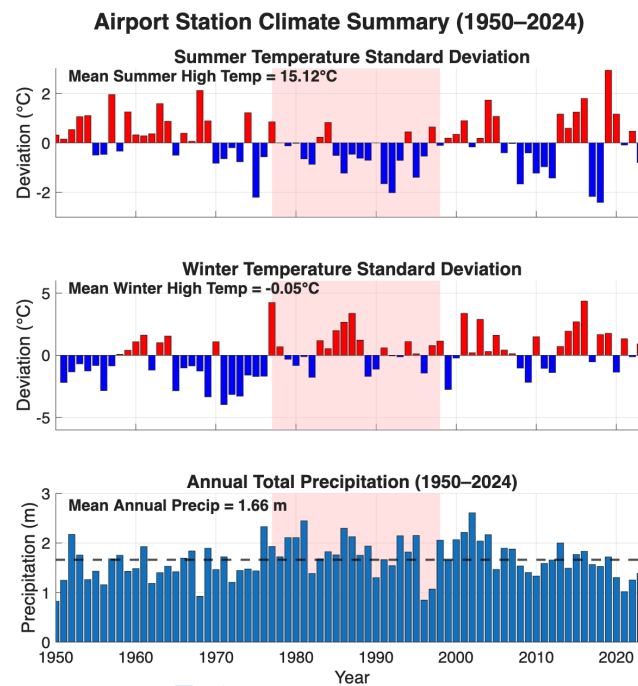


Fig. 3. Top: Standard deviations of the annual average daily maximum temperature during the summer from the 74-year mean (1950–2024). Middle: Standard deviations of the annual average maximum temperature during the winter from the 74-year mean. Summer is defined as the months between June and September, and winter is defined as December to February. Bottom: Annual cumulative precipitation from 1950–2024. The dotted line represents the average annual precipitation of the time record. The shaded region on all graphs is the duration of the Pacific Decadal Oscillation warm phase (1977–1998).

Air Temperature and Precipitation

A time series of daily temperature, precipitation, and other climate variables is available from the Global Historical Climatology Network Daily (Menne and others, 2012). Seward Airport⁷ and Seward 8NW⁸ are the two weather stations closest to Bear Glacier. We use the Seward Airport station here as it contains a longer and more consistent climate record from 1950 to 2024. Figure 3 shows the annual average high temperature during the summer and winter as the yearly standard deviation from the 74-year mean, and the precipitation as an annual cumulative.

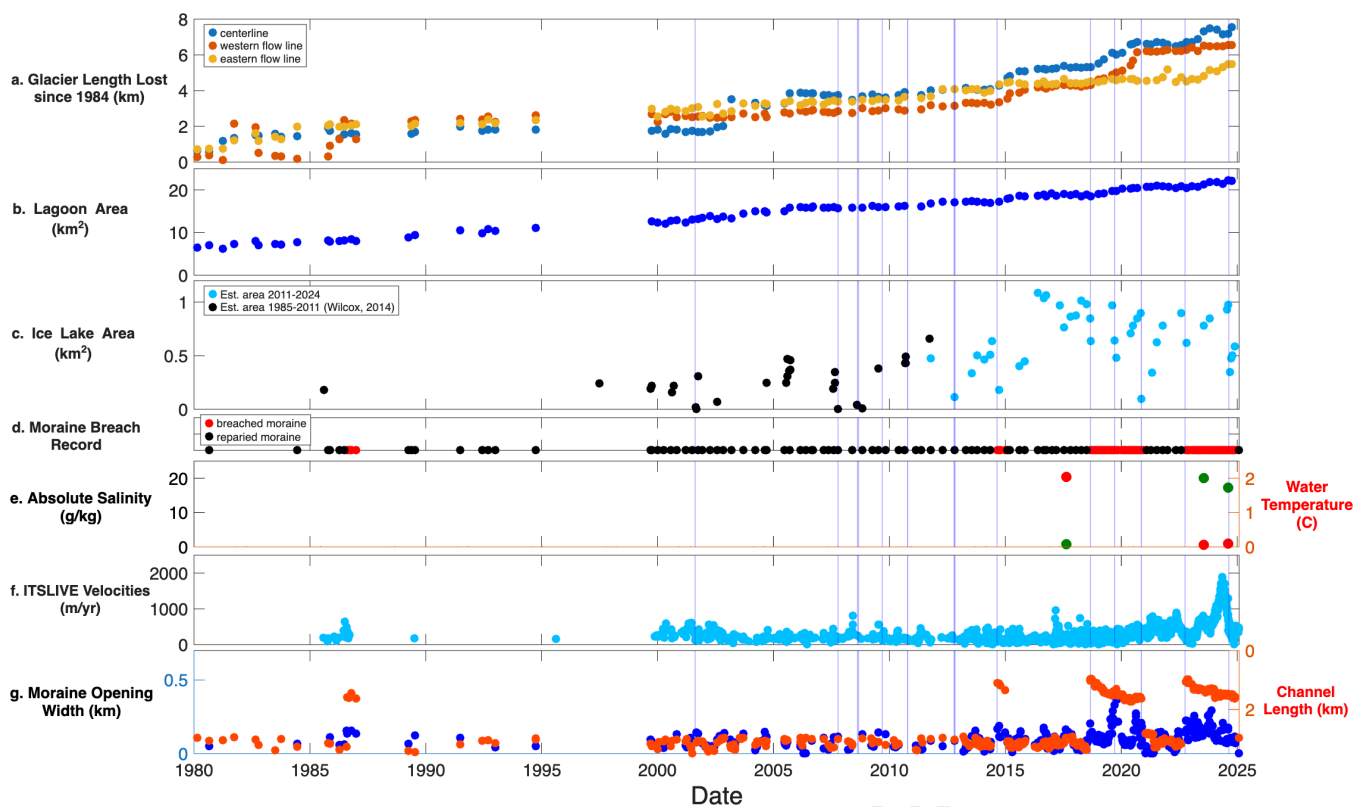


Fig. 4. From top to bottom: time series plots of a) the terminus length lost along each flow line, b) lagoon area, c) ice lake area, d) record of morainal spit breaches and repairs, e) depth-averaged water temperature and absolute salinity, f) ice surface velocities from the terminus reference line center (59.9938°N , -149.629°E), and g) channel length and moraine breach opening width, from 1980 to 2024. The axis for channel length is reversed to highlight periods of shorter channel length. Blue vertical lines identify documented outburst floods.

RESULTS

We summarize the behavior of the Bear Glacier system in Figures 4 and Figure 5 which show the temporal and spatial evolution of the glacier terminus, lagoon, and ice lake. Figure 4 spans over 52 years of satellite history. From top to bottom, these display the terminus length lost along the three flow lines, the area of the lagoon, the area of the ice lake, a temporal record of when the moraine was eroded and repaired, salinity and water temperature measurements, glacier speeds at the terminus, and the channel's length and opening width. Figure 5 shows the glacier terminus behavior by means of the spatial pattern of retreat, highlighting periods of relative stability.

Bear Glacier Terminus

Bear Glacier has lost over 7.5 km along its centerline, 6.5 km in along the western side, and 5.7 km along its eastern extent between the earliest available Landsat image in 1972 and 2024 (Figure 4a). This equates to 23.5 km² in ice area loss. The terminus from 1978 to 2002 shows lateral narrowing from the eastern and western sides of the glacier while maintaining length along the glacier centerline. In 2003, the centerline of the terminus rapidly lost 2.7 km in length. From 2005 onward, the retreat pattern is stair-stepped involving all three flow lines. In Figure 5, this pattern is shown spatially, with periods that the terminus remained relatively stable between rapid retreat: the first stable period shows the terminus laterally narrowing while remaining in the same position in the lagoon from 1978 to 2002. The following stable periods were from 2005 to 2013, then from 2015 to 2018, and most recently from 2020 to early 2022 (Figure 5).

Ice Lake

A total of 12 outburst floods have been documented, including the ones previously published by Wilcox and others (2014) (Table 1 and blue vertical lines in Figure 4). Eight of the 12 outburst floods occurred between August and September, three occurred October, and one in November. The earliest outburst floods were recorded in early September 2001 and recurred in the late summer to early fall months of the years 2007-2010 and 2012. From the data available, the 2001, 2007, 2008, and 2012 outburst floods left the ice lake with no water visible by satellite (area of 0 km²), whereas the floods in 2009 and 2010 only

⁷The ID for the Seward Airport weather station is GHCND:USW00026438.

⁸The ID for the Seward 8NW weather station is GHCND:USC00508375.

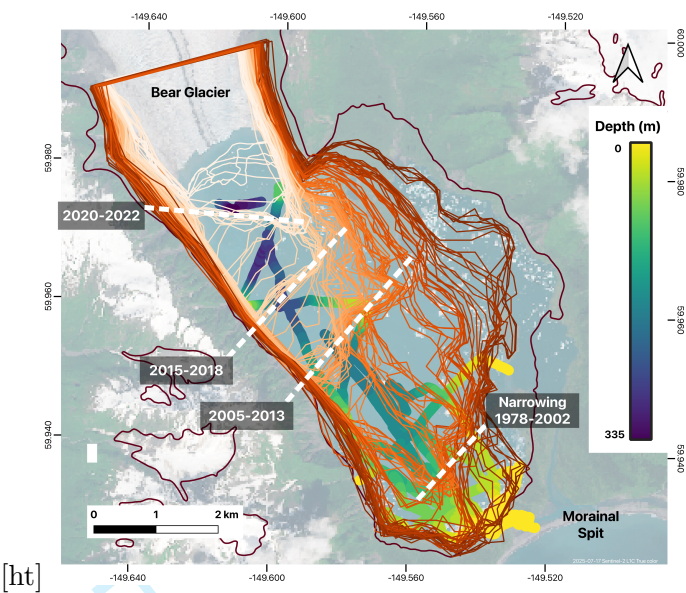


Fig. 5. Glacier terminus positions in 1950 (brown), and from 1972 (red) to 2024 (beige). White dotted lines with dates show periods when the terminus was stable. Bathymetry is shown underneath the terminus positions, where dark blues classify greater depths. 1978-2002 shows narrowing of the terminus prior to retreat. Background image is from Sentinel 2 in July 2025, projected using UTM zone 6N Coordinate Reference System of the 1984 WGS ensemble.

partially drained the lake⁹. Decreases in the area of the ice lake correlate to the outburst flood events, which is why there is significant variability (Figure 4c).

In 2014, the data shows a transition in behavior. The average area of the ice lake at its maximum before any area decrease (when the lake is filled) prior to 2014 was an estimated 0.43 km². After 2014, however, the area of the filled ice lake doubled to an average area of 0.93 km². Additionally, all 7 outburst floods dated 2014 and later were partial drainage events as opposed to full drainage events. 2014 and 2020 were the only two years when the ice lake drained nearly all of its water. Large icebergs remained at the bottom of the lake after some drainage events, seen in satellite imagery. In addition to the 12 reductions in ice lake area that correspond to an outburst flood, there is an additional reduction in ice lake area between May and August of 2016 outside of the documented outburst floods. The 2016 area decrease of 0.053 km² is relatively small compared to the ones correlated with recorded outburst floods. Brown plumes near the terminus in satellite imagery suggest this may have been a partial drainage event sometime during the summer that was not large enough to cause a breach or any change in the spit morphology. We also cannot

⁹The area of the lake after drainage is not explicitly described in the dataset from Wilcox and others (2014). Their combination of visual inspections, NDWI analysis, and local pressure transducer data from 1985-2012 determined whether the lake was visible, draining, or drained.

rule out that the discharged water may be from an additional water storage area within Bear Glacier's subglacial system.

Bear Glacier Lagoon

As the glacier retreated, the lagoon grew in size (Figure 4c). As of 2024, the area of the lagoon is 22.3 km²¹⁰, which is more than triple the size of its 1984 area of 6.9 km². The lagoon also is deeper towards the current terminus, reaching a maximum observed depth of 335 m as of 2024. A deep central trough consistently over 200 m depth follows the former glacier centerline; the bathymetry in the western side of this trough is steep and bounded by a ridge, the eastern side slopes more gently towards the historic 1950 terminus that resembles a half-piedmont lobe (Figure 2a).

The temperature and salinity measurements through depth are shown for the years 2017, 2023, and 2024 (Figure 6a, 6b). The lagoon was overall warmer and fresher in 2017 than it was in 2023 and 2024. The 2017 profile is relatively uniform through depth compared to the sharply curved 2023 and 2024 profiles. The lagoon temperature decreased from 2.05 °C throughout the water column in 2017, to near 0 °C at depths greater than 6 m in 2023. The 2017 average salinity of the lagoon shows nearly fresh water at 0.816 g kg⁻¹. In 2023, the salinity measured 20 g kg⁻¹. The 2024 salinity was lower than in 2023 but the lagoon was still brackish at 17 g kg⁻¹. The 2023 salinity was the highest recorded across all years of salinity data obtained. The 2023 and 2024 temperature-salinity profiles in respect to water mixing reference lines (Figure 6c, 6d) show that surface lagoon waters are more strongly influenced by the glacier, while deeper waters may be dominated by the introduction of denser seawater. A more in depth analysis of the lagoon's water properties and mixing are reserved for future work.

Moraine

The length and width of the channel are compiled in Figure 4g starting in 1980. Figure 4d identifies the dates of spit breaches in red. Aside from a short breach and channel shortening in 1986, the breaches are a generally newer phenomenon recurring since 2014. The channel shortens to 1.013 ± 0.089 km at the initiation of a breach and lengthens to 3.132 ± 0.146 km as the spit repairs. Conversely, the "breach opening" widens to 0.221 ± 0.036 km immediately following an outburst flood and narrows as the spit repairs. The closure of the breach and restoration of the spit typically takes just over two years (28 months) to return

¹⁰The glacier area loss is larger than the lagoon size because the glacier narrowed as it thinned due to valley wall geometry.

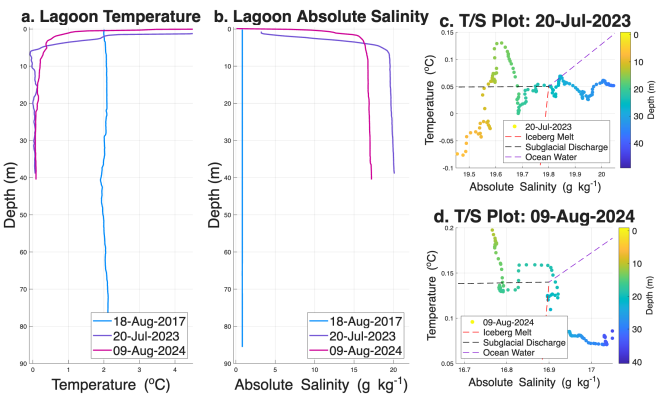


Fig. 6. a) Temperature profiles taken in Bear Glacier Lagoon in 2017 (blue), 2023 (purple), and 2024 (pink). b) Absolute salinity profiles taken in 2017, 2023, and 2024. Panels a and b show the lagoon in 2017 to be warmer and fresher than that in 2023 and 2024. c) Temperature-salinity plot of the 2023 profile. d) Temperature-salinity plot of the 2024 profile. Points in panels c and d are colored by depth from shallowest (yellow) to deepest (blue). Dotted lines represent Gade-style references of iceberg melt (red), subglacial discharge (black), and ocean water (purple) as modifiers of lagoon mixing (Gade, 1979).

to normal stages or even close off connection to Resurrection Bay entirely. We suspect wave action and longshore drift as the primary mechanisms for this restoration.

Due to erosion, sedimentation, and variable water levels within the highly variable channel, we do not have a time series of channel depths. From conversations with local boat captains, the channel can be as shallow as 1 m and is only accessible by smaller jet boats. We conducted a limited survey of channel bathymetry in 2024 before the outburst flood. The channel at its shallowest cross section at that time had a depth of 2.09 m.

In Figure 7, we show the transition from False River to the modern channel between 1977 and 1982 through satellite imagery, which is coincident with a period of warmer weather, associated with the Pacific Decadal Oscillation (Figure 3).

DISCUSSION

Transition from tidewater to land-terminating glacier

Reconstructions of Bear Glacier’s history show the glacier terminating just before the ocean on flat gravel land in 1909 (Grant and Higgins, 1913). The size of the modern morainal spit relative to the depth of the lagoon upstream of the terminus (Figure 5), as well as descriptions of the Bear Glacier face partly

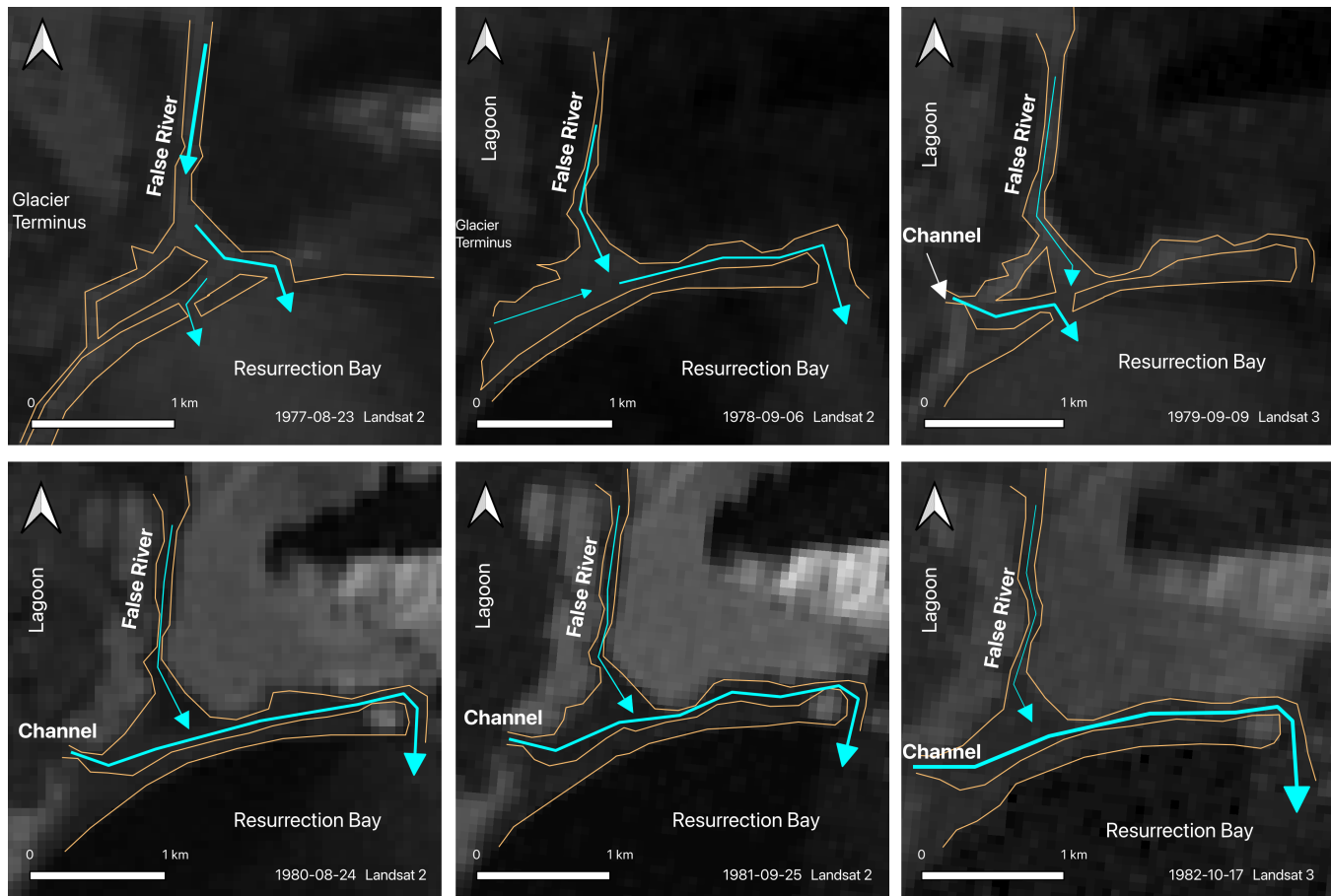


Fig. 7. Morainial spit and adjacent water flow from 1977-1982 showing a switch in the dominant passageway out to the ocean. The orange lines outline False River, the channel, and moraine from Landsat 2 and Landsat 3 images. The blue arrows show the direction of water flow from the southeast end of the lagoon, through False River and the channel, and out to Resurrection Bay. Thicker arrows indicate more dominant flow through the outlets.

covered by wave and wind action at high tide, suggest that it behaved as a tidewater glacier while it built the moraine (possibly as recently as the Little Ice Age). The moraine formed between two mountain ridges: Bear Glacier Point to the west (not in frame) and Callisto Head to the east. Beyond the moraine, Resurrection Bay widens to the ocean and provides few stabilizing bedrock features on the northeast side. This geometry suggests that the two ridges acted as a pinning point for the glacier because it struggled to advance into the wider bay after the retreat of the main trunk glacier that flowed down Resurrection River Valley and out Resurrection Bay. Bear Glacier, therefore, would have built the moraine through successive advances to Callisto Head, and by rarely extending beyond this point, the glacier ultimately terminated on the subaerial morainal outwash plain that is now the morainal spit. While this history of Bear Glacier is inferred from a sparse data set, we summarize it here because the processes involved in its evolution are important to note for understanding which glaciers globally may be more likely to become lagoon-terminating with future ice loss.

Transition from land-terminating to lake terminating glacier

The glacier’s retreat in the early and mid-20th century first formed Bear Lake and a second meltwater pool to its northeast and east (Figure 2). We assume here that these ponds were composed solely of glacier melt and were therefore fully freshwater. In the 1950s, the lakes drained through the morainal spit via the False River, which cut through the moraine directly to reach Resurrection Bay. While most of the glacial meltwater flowed through the False River, there were also small meltwater streams crossing the low lying moraine in other places.

We turn to the geometry of the glacier and its flow from the Harding Icefield to better understand the early formation of Bear Lake around the 1950s (Figure 1). The medial moraine in the 1950s USGS map separates the glacier such that the northeastern tributary glaciers feeds the northeastern terminus and the southwestern tributary feeds the southwestern terminus. The northeastern tributaries are sourced directly from the Harding Icefield through a wide valley, while the southeastern tributary funnels ice from the icefield through narrower valleys and at higher elevations. The subglacial hydrological systems may also vary on either side of the medial moraine because of the differences in tributary geometry, especially near the equilibrium line altitude, where the subglacial hydrological systems typically become more active. The size of the northeastern tributaries and their ablation area, as well as the two lakes flowing into the False River, suggest that a larger percentage of meltwater flowed through the northeastern section of the

glacier.

Satellite imagery shows that water flow shifted southward, rerouting drainage along the terminus towards the moraine between 1977 and 1980 (Figure 7). We infer that this reorganization was caused by enhanced ice melt that increased water levels at Bear Lake. The lake expanded across most of the large, low-lying outwash plain in front of the terminus. The Pacific Decadal Oscillation brought warmer winters and greater precipitation during this time (Figure 3). Because the southern outwash plain at the time had a lower elevation than False River, the expanded lake drained to the south, instead of through False River. The final barrier to the ocean (the modern morainal spit), however, required the water to flow northeast along glacier side of the spit until it exited to the ocean at the bedrock ridge near Callisto Head. This drainage configuration is what we see today. False River now intermittently dries up under low lagoon levels. As the glacier continued thinning, the glacier retreated from the moraine and the terminus gradually narrowed from the eastern side (Figure 5). Glacier retreat created more water to form a deeper, more extensive, proglacial Bear Lake, and the terminus behavior changed.

The first visible change after 1980 we identify in the satellite record shows that the terminus formed a long, narrow tongue by retreating from the southwest valley wall in addition to the ongoing retreat along the northeast margin. A catalyst to this retreat was flooding in 1986. The glacier's terminus, surface velocity, and morainal spit signatures looked similar to that of a modern outburst flood event (Figure 4a,d,f,g); however, there is not enough ice lake area data to verify this. The long narrow ice tongue existed from around 1980 through 2002. Yang and others speculated that this ice tongue was floating, and we concur: our bathymetry (Figure 10) shows the tongue persisted within a trough 100 to more than 300 m deep. We can infer that this narrow glacier ice was unlikely to be thick enough to be grounded everywhere. The presence of a floating ice tongue, which is common for lake-terminating glaciers but not for tidewater glaciers, points to Bear Glacier terminating in freshwater at least until 2002.

Between 2003 and 2005, the floating terminus retreated 2.5 km before stabilizing in 2005. The first documented outburst flood happened in the fall of 2001 and did not reoccur until 2007 (Wilcox and others, 2014); thus, we cannot link the 2.5 km terminus retreat and stabilization between 2003 and 2005 to an outburst flood. However, warmer and wetter conditions in the early 2000s may have prompted greater thinning, and eventually, weakening and breakup of the ice tongue (Fig 3). The average high temperature during these summers was 0.5–2 °C above the 74-year mean, and winter high temperatures have been 0.5–5 °C over the mean since the start of the Pacific Decadal Oscillation warm phase in 1977. We suspect that

the terminus stabilized in 2005 because the terminus reached a shoal that may have provided some back stress (lighter bathymetry under 2005-2013). The terminus stayed at this location until 2013.

Transition from lake to lagoon

Outburst floods and subglacial hydrology

While the first documented outburst flood was in 2001 (and one that may have occurred in 1986), the outburst floods have been recurring more regularly since 2007. Comparing the 2000s to the late 2010s and 2020s, the ice lake area and magnitude of these outburst floods have increased. There are three factors that modulate outburst flood magnitude and frequency: the size of the ice lake that sources the floods, the thickness of the glacier ice over the drainage path, and the organization of the subglacial water system. The size of the ice-marginal lake (Figure 2b) is constrained by the shape of the valley and the geometry of the ice. The glacier flows up into this valley after turning the corner around the ridge separating it from the main trunk of the glacier. There, it acts as a small lake-calving glacier margin. In 2014, after an outburst flood, the ice margin of the lake collapsed, which doubled the size of the ice lake (from 0.43 km² in 2001-2012, to 0.93² in 2015-2024). We document this ice margin change using satellite imagery over a 2-year period for comparison before and after the 2014 outburst flood (Figure 8). The increase in lake area may have allowed more water to be stored in the ice lake, resulting in more water draining during an outburst flood.

The thickness of the glacier ice helps determine the behavior of the ice lake and outburst flood activity in two ways. First, ice thickness is directly related to the height of the dam containing the ice lake, which controls how much water the lake can hold (Wolfe, 2008). Depending on the ice thickness, the lake may either store more water over time, or reach a threshold to where the ice dam thins, allowing it to float easier, and the lake gradually stores and releases smaller amounts of water (Veh and others, 2023). In the case of Bear Glacier, the 2014 margin collapse expanded the ice lake and allowed for more water storage and thus stronger outburst floods; however, we speculate from 2018-2024 ice lake area that the lake may be approaching an ice thickness threshold. We do not have bed topography under the glacier that flows into the ice lake, but because the ice is thinning, outburst floods may be slightly smaller but more frequent in the near future. Given that all outburst floods after 2014 have breached the spit, we suspect that a small decrease in lake water would still be enough to cause a breach, meaning more frequent outburst floods would leave the moraine breached for longer periods of time. Secondly, water from the ice lake drains

into the lagoon through a subglacial channel system that expands by melting from turbulent water flow and shrinks by viscous creep closure. Outburst floods commonly occur when there is an imbalance in the viscous ice closure compared to melt (Fowler, 1999). Thinner ice reduces the overburden pressure, which can also initiate easier flotation of the ice and thus more frequent drainage.

Finally, the organization of the subglacial water system matters. Figure 4f shows the speed of the glacier surface through time. The glacier accelerates in spring or early summer due to a decrease in the effective water pressure (increasing water pressure relative to ice overburden) which allows it to slide more effectively. Eventually, the subglacial water systems reorganize to more efficiently transport water to the terminus by channelizing the flow. The channelization of the subglacial water slows down the glacier because the lubrication provided by the water is concentrated in subglacial channels rather than being distributed. The episodic outburst floods from the ice lake to the lagoon happen after this reorganization to channelized flow. The velocity data from the recent decade along with documented outburst floods show that the outburst floods occur as the glacier flow speed rapidly falls or reaches its annual minimum around late summer to early fall. Furthermore, the peak spring speed has been increasing over the past decade (Figure 4f), reaching its fastest at over 2000 m a^{-1} in 2024. This increase in maximum speed each spring implies that the effective pressure at the base of the glacier is getting progressively lower each spring. Lower effective pressure can result from either increased water pressure (due to earlier and more enhanced spring warming) or due to lower ice pressure (due to thinning of the ice). It is likely a combination of these.

When the ice lake level is high enough to provide sufficient hydraulic gradient to drive flow into the channelized subglacial hydrological system, this water flow will trigger the instability that widens the subglacial channel and allow the ice lake to drain rapidly. These outburst floods do not happen every year, and when they do happen they do not always erode the spit enough to breach. A more in-depth analysis of the subglacial hydrology and outburst flood triggers must be further conducted, including the idea that large late summer and early fall storms may be a potential trigger for some outburst flood drainage events.

Increased salinity

Saltwater exchange into the lagoon can happen from storms and during high tides. This saltwater exchange depends on the shape and depth of the channel, as well as the lagoon level relative to sea level. Reverse flow is more favorable when the lagoon water level is relatively low and sea level is high. Even when the surface water is flowing out to sea, however, salt water intrusion into the lagoon can happen if the

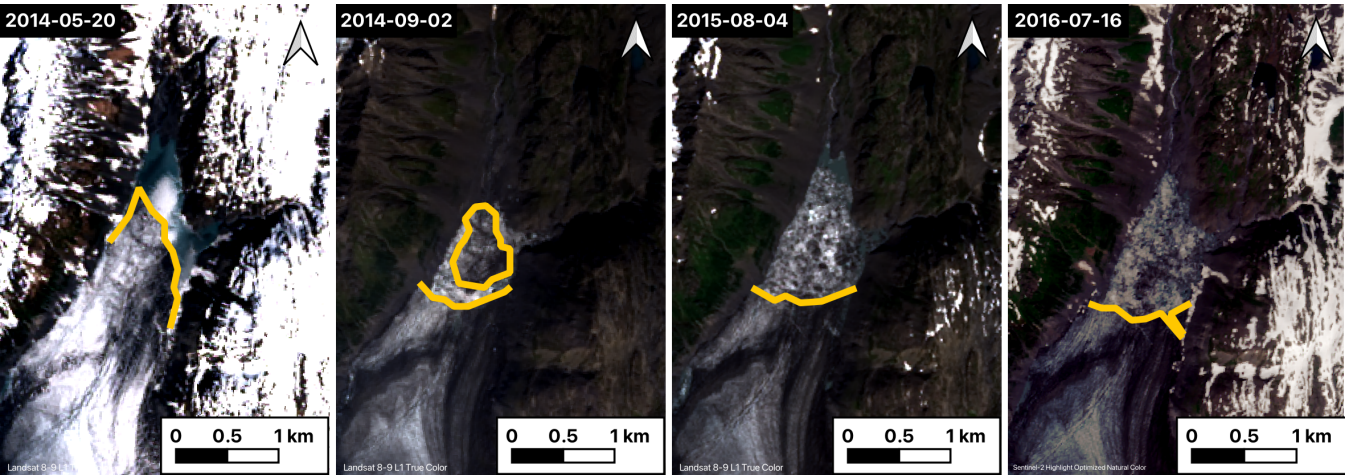


Fig. 8. Satellite imagery of the ice lake between 2014 and 2016. The yellow line follows the ice margin, which collapses and breaks away after the 2014 outburst flood (seen in the 2014-09-02 image).

channel geometry and flow dynamics are such that a dense salt wedge can flow upstream underneath the surface freshwater layer, as is known to happen in estuaries (Gale and others, 2007). The 2023 and 2024 temperature-salinity plots point to either reverse-flow or salt-wedge saltwater intrusions, but more analysis is needed to determine the dominant processes.

We can say that the erosion of the spit is an important control on saltwater exchange. The erosive power during breach events likely cuts a channel significantly deeper than the normal channel flow. We do not have direct measurements of channel depth, except from a few eyewitness accounts by boat captains who have driven up the breached channel and observed reverse flow at the highest tides during summer. In winter, lagoon levels will naturally lower as the snowpack stores local precipitation and there are minimal sources for water within the glacier subglacial hydrological system. Thus the winter months are the most likely time when saltwater intrusion occurs.

The erosion and breach of the spit after the 2014 outburst flood was the first since 1986, and the first recorded one associated with a documented outburst flood. The spit remained breached for 16 months (compared to only 4 months in 1986). The first known salinity measurement in Bear Glacier’s lagoon was collected by us in 2017 at 0.816 g kg^{-1} maximum (observed at 100 m depth due to limited instrument capabilities, Figure 6). We assume that the observed salinity is representative of the full depth of the lagoon. While it is possible that the lagoon water was somewhat stratified and some dense water pooling at deeper depths, mixing due to iceberg calving (Shaya and others, 2025), winds, and other sources for lagoon kinetic energy make it unlikely. We also assume that a very small volume of ocean water infiltrated

the lagoon following the 2014 outburst flood. The lagoon was still near-fresh in 2017: perhaps not fully freshwater, but far from the salinity of ocean water. Flushing of the lagoon by precipitation and glacial melt water between 2014 and 2017 make it difficult to quantify the change in salinity caused by the 2014 breach.

The 2018 outburst flood was the largest in Bear Glacier's recorded history; it breached the moraine and likely significantly affected salinity. Our next salinity measurement, however, was not collected until 2023, 5 years later, and one year after the 2022 outburst flood, which also left the spit breached. This measurement shows the salinity in the lagoon at over 21 g kg^{-1} , 60% that of seawater. The mean temperature in 2023 was 0.5°C , lower than when it was a freshwater lake, due to the more vigorous mixing stimulated by buoyancy and melting icebergs that consume latent heat from the ocean water. There was no outburst flood in 2023. Our observations the following year in 2024 showed slight freshening along with more cooling (to 0°C) compared to 2023; this implies that flushing by precipitation and glacial melt exceeds infiltration of ocean water.

Although no historical salinity data at Bear Glacier exists prior to 2017, we infer that the salinity increase between 2017 and 2023 shows a shift in the lagoon from over 50 years as a fresh water lake to brackish lagoon. As stated above, this saltwater exchange flow can occur if the lagoon's water level decreases to near or below that of the ocean, such that the export of water slows or reverses flow direction. We surmise that this lowering of the lagoon level compared to the ocean promotes backflow of saltwater into the lagoon and sediments into the breach, which then slowly repairs the breach through wave and tidal processes moving from southwest to northeast.

Because of our observation of breaches initiating increased salinity, the transition to a saltwater lagoon or even a tidewater fjord could occur if erosion caused by a flood event or storm were to deepen the thalweg of the channel permanently. We rule this out so far because of the observed mobility of sediment and other processes acting to repair the spit. Local surfers have noticed southwest to northeast movement of the beach via ocean currents and waves as the longshore transport repairs the spit over the winter months. Note that the repair is not immediate and the spit has historically shown to recover over the course of 2-3 winter seasons. After the 2020 and 2024 outburst floods, for instance, coastal sediment transport repaired the moraine and returned the channel flow through its normal routing in 4 months.

We conclude, therefore, that the freshwater from glacier melt and precipitation, and saltwater from the ocean, are both important inputs that vary on differing timescales. We hypothesize that the lagoon likely

freshens both immediately after an outburst flood and during summer ablation and snow melt season when it progressively flushes out salts, aided by high precipitation events. The lagoon gets saltier in the winter, particularly after a breach in the spit, when lagoon water levels lowers and reverse flow and wave processes dominate. Confirming this hypothesis will require additional observations and modeling beyond the scope of this paper.

Glacier terminus behavior

Saltwater has turned the lagoon from near-fresh to brackish in less than 10 years (from 2014 to 2023), and we can correlate this to the changes in how the glacier retreats. Modern lagoon dynamics since 2014 show that the glacier retreats the most after an outburst flood and breach in the spit (Figure 4c and 4d). We stated previously that the 1978-2002 narrowing of Bear Glacier’s terminus is characteristic of a lake-terminating glacier floating ice tongue before calving off from 2003-2005. After 2005, the terminus has not formed an obvious, detectable ice tongue. The glacier stabilizing from 2005-2013 marked the first plateau of a stair-step retreat pattern. The next “step” formed after rapid retreat from 2013-2015, and when the terminus stabilized from 2015-2018, which is also when the first salinity measurement was taken (2017). Since then, the glacier has undergone periods of stability and active retreat that alternate every 3-4 years. The stepped retreat pattern, in addition to our understanding that the terminus at this time sits above an overdeepened trough in the bathymetry, displays characteristics more like that of a tidewater glacier.

The future of Bear Glacier: more rapid retreat and widening valley

Our final objective is to provide insight into the future of Bear Glacier: continued glacier retreat will open the lagoon to deeper bathymetry and wider glacier area. As of the date of submission of this paper (spring 2026), the terminus of the glacier has retreated back to the narrowest point of the valley. The deepest recorded bathymetry is over 335 meters below mean sea level located 500 m in front of the terminus. We expect the valley to deepen underneath the ice upstream because a 2010 airborne radar survey of the Harding Icefield revealed bed topography as deep as 450 m below sea level (Figure 9). The radar data combined with relatively smooth glacier surface topography suggest that bedrock bumps or ridges on which the glacier can stabilize may be rare. The glacier may not stabilize again until it has retreated another 10 km. This retreat process, coupled with frequent outburst floods from the ice lake, a prolonged breached spit, a resulting saltier lagoon, and a faster flowing glacier may cause more rapid retreat and mass loss

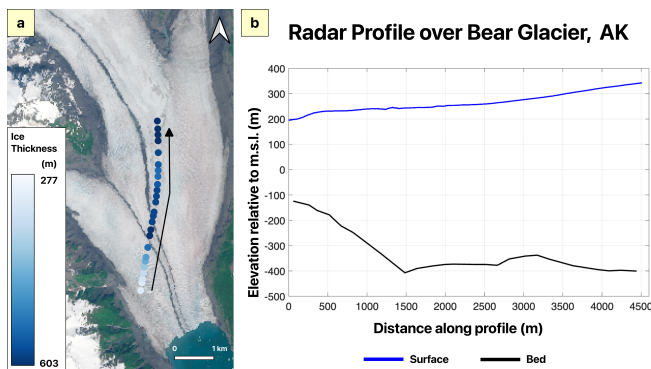


Fig. 9. a) Airborne radar data collected in 2010 that passes over Bear Glacier (Truffer, 2014). The arrow indicates the direction of the corresponding profile. b) Elevation profile along the 2010 Bear Glacier radar survey line. The blue line is the ice surface elevation and black line shows the glacier bed, both relative to sea level.

similar to that of a tidewater glacier. Retreat of this glacier threatens reconfiguration of Harding Icefield flow and amplified mass loss for other glaciers stemming from the Harding Icefield, affecting regional climate feedbacks, ecosystems, and recreation.

The lagoon-terminating glacier cycle

We generalize our findings by conceptualizing a cycle driven by outburst flood based on the observed dynamics at Bear Glacier. An outburst flood initiates the cycle, draining glacial melt, runoff, and collected precipitation from the ice lake into the lagoon through the subglacial drainage system. The sudden influx of water into the lagoon causes the lagoon water levels to rise (or flood). Excess water flowing into the channel overtops the weak point in the morainal spit, stimulating erosion that leads to a full breach of the spit, which forms a direct connection between Bear Glacier Lagoon from Resurrection Bay. Saltwater from Resurrection Bay enters the lagoon via backflow, most significantly during high tides and low lagoon levels. Over a period of 1-2 years, coastal processes act to repair the breach. As the spit recovers, the channel gradually lengthens and the breach opening narrows until it closes off or returns to its original estuary position at the far northeastern end of the spit by Callisto Head, which is also furthest point away from the lagoon. In this “repaired” channel configuration, the influx of saltwater is small or negligible. It is possible (and we observed in the Sentinel-2 satellite record in January 2021) that the channel can be completely blocked with no water flowing in either direction until glacier melt or precipitation raises the lagoon levels again. The combined effects of the outburst floods, subsequent breaching of the spit, and influx of saltwater precede periods of elevated calving and large retreats of the terminus. Vigorous

circulation of fjord waters is known to cause ice melt and calving (e.g. Truffer and Motyka, 2016). We hypothesize, then, that a significant increase in salinity will increase near-glacier currents, increase glacier melt, and trigger retreat to a new stable terminus position.

CONCLUSIONS

Here we show how Bear Glacier, its lagoon, and the ice lake have evolved over 74 years of history. What was initially a small, likely freshwater pond from Bear Glacier melt in the mid-20th century has evolved into a large, saltier lagoon of 15-20 g kg⁻¹ as of 2024. The salinity of the lagoon is set by the backflow of ocean water through an eroded morainal spit, frequently breached by glacial lake outburst floods from the upstream ice lake.

Understanding the patterns and connections between the flow Bear Glacier, the outburst floods from the ice lake, and the future retreat of Bear Glacier, can set a foundation for other emerging lagoon-terminating glacier systems in the Arctic and sub-Arctic. A section of Alaska’s Bering Glacier terminates in Vitus lake that had a measured salinity of 18 psu in 2001 (Josberger and others, 2006). A section of Malaspina Glacier, located in southeastern Alaska, flows into Sitkagi Lagoon. San Rafael Glacier in southern Chile evolved recently into a lagoon-terminating glacier (C. Escauriaza and M. Williams, personal communication). Importantly, as we have shown here, it may not be easy to distinguish between freshwater lake or brackish lagoon without salinity measurements. As more icefields and ice sheets retreat (particularly in Greenland and the high Arctic), more lagoons may form.

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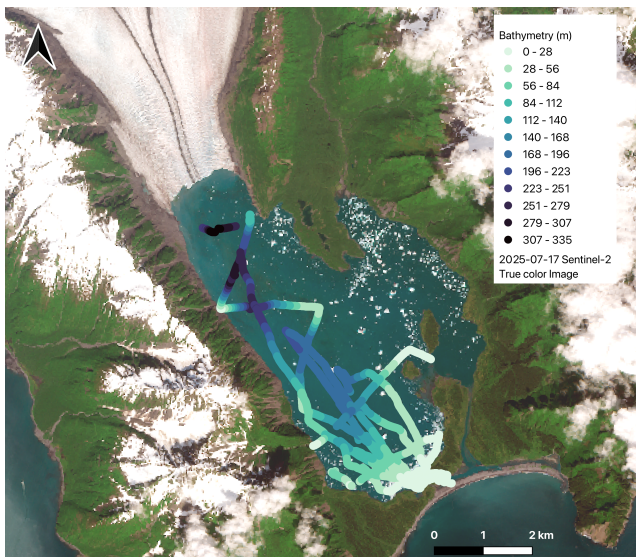


Fig. 10. Bathymetry within Bear Glacier Lagoon from terminus to channel.

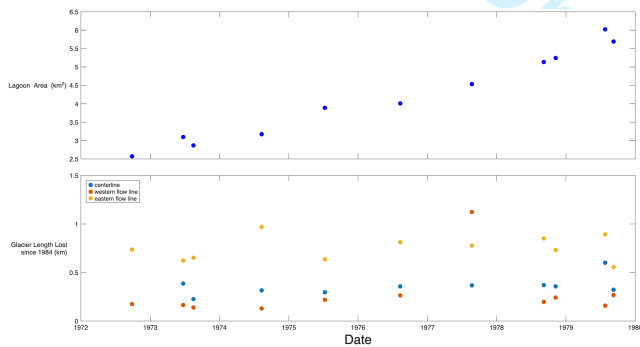


Fig. 11. Area of the lagoon (top) and the length of glacier lost from each flow line (bottom) from 1972-1980.

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732 APPENDIX

Table 1. Table of the ice lake imagery used and area from 1985-2011. Need to figure out how to highlight rows to identify GLOFs (the main command isn't working)

Date	Satellite	Area (km ²)
1985-08-10	Landsat 5 bands 1-7	0.18
1997-06-26	DOQQ (aerial photo, orthorectified)	0.24
1999-09-08	Landsat 7 bands 1-8	0.19
1999-09-26	Landsat 7 bands 1-8	0.22
2000-08-11	Landsat 7 bands 1-8	0.16
2000-09-09	Color compressed Landsat 7TM	0.22
2001-08-21	Landsat 7 bands 1-8	0.02
2001-09-06	Landsat 7 bands 1-8	0
2001-09-29	Landsat 7 bands 1-8	0.31
2002-07-30	Landsat 7 bands 1-8	0.07
2002-11-12	Color compressed Aster L1B	lake visible
2004-09-14	Landsat 7 bands 1-8	0.25
2005-07-15	Landsat 7 bands 1-8	0.25
2005-08-06	Aerial or ground-based photo	lake visible
2005-08-08	Landsat 5 bands 1-7	0.31
2005-08-08	Color compressed IKONOS 2	0.47
2005-08-09	Color compressed Aster L1B	lake visible
2005-09-01	Landsat 7 bands 1-8	0.36
2005-09-18	Landsat 5 bands 1-7	0.37
2005-09-25	Landsat 5 bands 1-7	0.46
2007-07-28	Landsat 7 bands 1-8	0.19
2007-08-26	Landsat 7 bands 1-8	0.25
2007-08-30	Landsat 5 bands 1-7	0.35
2007-10-09	Landsat 7 bands 1-8	0
2008-08-08	Landsat 7 bands 1-8	0.04
2008-08-19	Aerial or ground-based photo	lake drained
2008-08-30	Aerial or ground-based photo	lake drained
2008-10-27	Landsat 7 bands 1-8	0.01
2009-06-03	Aerial or ground-based photo	lake visible
2009-07-09	Landsat 5 bands 1-7	0.38

Date	Satellite	Area (km ²)
2009-09-04	Aerial or ground-based photo	lake drained
2010-07-21	Aerial or ground-based photo	lake visible
2010-08-05	Aerial or ground-based photo	lake visible
2010-08-23	Color compressed Aster L1B	lake visible
2010-08-31	Landsat 5 bands 1-7	0.43
2010-09-11	Aerial or ground-based photo	lake visible
2010-09-12	Color compressed SPOT ortho tile	0.49
2010-09-16	Aerial or ground-based photo	lake visible
2010-09-16	Landsat 5 bands 1-7	0.43
2010-10-08	Field work pressure transducer	lake draining
2011-08-11	Aerial or ground-based photo	lake visible
2011-09-26	Landsat 5 bands 1-7	0.66
2011-09-26	Aerial or ground-based photo	lake visible
2011-10-12	Landsat 5	0.473
2012-10-22	Landsat 7	0.112
2013-07-20	Landsat 7	0.336
2013-10-09	Landsat 7	0.502
2014-01-30	Landsat 7	0.466
2014-05-04	Landsat 7	0.511
2014-06-05	Landsat 7	0.634
2014-09-26	Landsat 7	0.179
2015-08-12	Landsat 7	0.401
2015-10-31	Landsat 7	0.448
2016-05-26	Landsat 7	1.084
2016-08-30	Landsat 7	1.037
2016-10-04	Sentinel 2	1.064
2017-05-05	Sentinel 2	0.968
2017-07-12	Sentinel 2	0.763
2017-10-19	Sentinel 2	0.862
2018-01-20	Sentinel 2	0.873
2018-04-14	Landsat 7	1.013
2018-07-04	Sentinel 2	0.979
2018-08-28	Sentinel 2	0.845
2018-09-05	Sentinel 2	0.636

Date	Satellite	Area (km ²)
2019-08-07	Landsat 7	0.972
2019-09-14	Sentinel 2	0.641
2019-10-12	Sentinel 2	0.483
2020-05-27	Sentinel 2	0.711
2020-07-01	Sentinel 2	0.782
2020-09-10	Aster	0.849
2020-10-29	Sentinel 2	0.897
2020-11-15	Sentinel 2	0.097
2021-04-26	Sentinel 2	0.342
2021-07-15	Sentinel 2	0.627
2021-10-26	Sentinel 2	0.778
2022-07-28	Sentinel 2	0.898
2022-10-21	Sentinel 2	0.621
2023-07-20	Sentinel 2	0.778
2023-10-26	Sentinel 2	0.850
2024-07-19	Sentinel 2	0.931
2024-08-13	Sentinel 2	0.976
2024-09-07	Sentinel 2	0.345
2024-09-27	Sentinel 2	0.473
2024-10-18	Sentinel 2	0.504
2024-11-19	Sentinel 2	0.584