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Post-glacial relative sea level, isostasy, and glacial history in Icy Strait, Southeast Alaska, USA

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Abstract

We use the radiocarbon ages of marine shells and terrestrial vegetation to reconstruct relative sea level (RSL) history in northern Southeast Alaska. RSL fell below its present level around 13,900 cal yr BP, suggesting regional deglaciation was complete by then. RSL stayed at least several meters below modern levels until the mid-Holocene, when it began a fluctuating rise that probably tracked isostatic depression and rebound caused by varying ice loads in nearby Glacier Bay. This fluctuating RSL rise likely reflects the episodic but progressive advance of ice in Glacier Bay that started around 6000 cal yr BP. After that time, RSL low stands probably signaled minor episodes of glacier retreat/thinning that triggered isostatic rebound and land uplift. Progressive, down-fjord advance of the Glacier Bay glacier during the late Holocene is consistent with the main driver of this glacial system being the dynamics of its terminus rather than climate change directly. Only after the glacier reached an exposed position protruding into Icy Strait ca. AD 1750, did its terminus succumb – a century before the climate changes that marked the end of the Little Ice Age – to the catastrophic retreat that triggered the rapid isostatic rebound and RSL fall occurring today in Icy Strait.

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Keywords: Relative sea level; Southeast Alaska; Glacier Bay; Holocene; Late Pleistocene; Glacial isostasy; Glacial history; Neoglaciation; North Pacific; Icy Strait

Introduction

The Quaternary history of Southeast Alaska is of particular interest for two reasons. First, this region lies along the proposed coastal route for human dispersal from Asia into the Americas during the late Pleistocene (Heusser, 1960; Fladmark, 1979; Barrie and Conway, 1999; Fedje and Josenhans, 2000; Hetherington et al., 2003). Second, glaciers in Southeast Alaska respond to the same climatic and oceanographic changes over the North Pacific that affect climate and weather over large parts of North America, and so the glacial history of Southeast Alaska provides a valuable proxy record of regional climate change.

The history of relative sea level (RSL) change has direct relevance to both these topics. The feasibility of the northwest coastal route depended on the timing and extent of glacier advances along the seaward margin of the Cordilleran Glacier

Complex (Mann and Peteet, 1994; Clague and James, 2002; Cararra et al., 2003), and RSL history in Southeast Alaska contains an important component of glacial isostasy. Furthermore, proving that the coastal route was actually used during the late Pleistocene depends on locating archaeological sites of that age, and site locations may be predicted from the height of their contemporary RSLs. As regards climate history, many details of the glacial history of Southeast Alaska remain poorly understood even during the Holocene (Calkin et al., 2001; Wiles et al., 2002; Barclay et al., 2006), and reconstructions of RSL history can support and expand the existing glacial chronology there.

One of the most glacially dynamic parts of Southeast Alaska is the Icy Strait–Glacier Bay area (Fig. 1), which was repeatedly overrun during the late Pleistocene by part of the Cordilleran ice sheet complex (Mann and Hamilton, 1995). During the Little Ice Age (LIA), ca. AD 1300 to 1900, alpine glaciers draining the St. Elias Mountains coalesced in Glacier Bay and advanced into Icy Strait (McKenzie and Goldthwait, 1971). Isostatic depression under this glacier and glaciers in the surrounding

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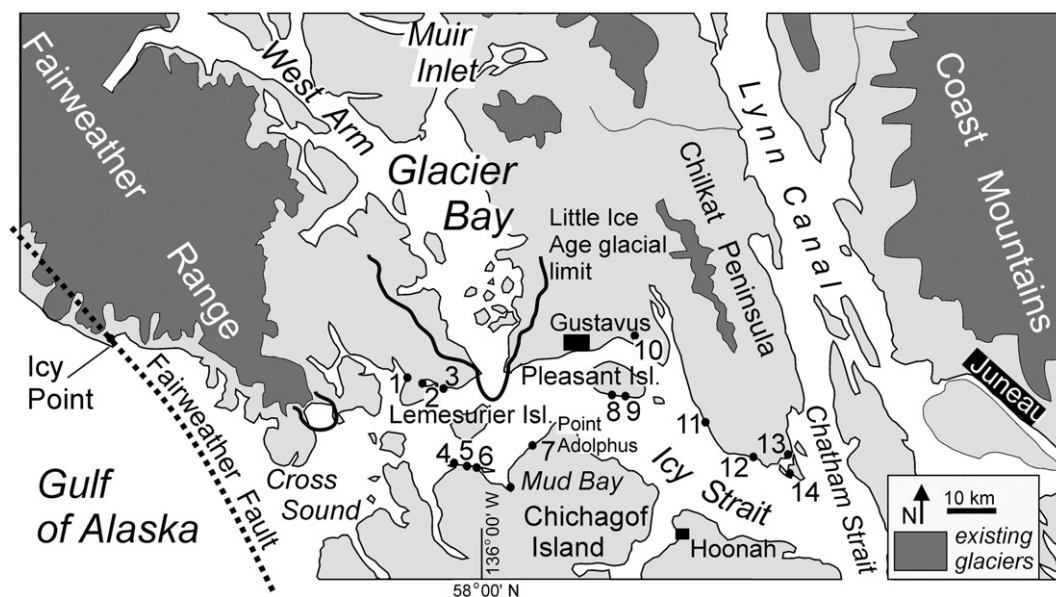


Figure 1. Location of study area in northern Southeast Alaska. Localities mentioned in text (informal names in quotations): 1 southeastern Dundas Bay, 2 Salt Chuck, 3 cove east of Salt Chuck entrance, 4 “Gull Cove Point”, 5 “Dancing Bear Creek,” 6 “Goose Island Pond,” 7 “Old Canoe Bay,” 8 “Gravel Canyon,” 9 “Black Rock,” 10 Falls Creek, 11 “Cemetery Point,” 12 Ground Hog Bay, 13 Swanson Harbor, and 14 Entrance Island. Other localities may be found on 1:63,360-scale maps or on nautical charts for the area. The Little Ice Age (LIA) limits of glaciers are from [Derksen \(1976\)](#) and A. Post and G.P. Streveler (personal communication, 2006).

ranges caused a RSL high stand of regional extent during the LIA ([Motyka, 2003](#); [Larsen et al., 2005](#)). Thus, reconstruction of the history of RSL changes in Icy Strait contributes new information to the Pleistocene and Holocene glacial history of both Glacier Bay and the surrounding region.

Study area

Coastal geomorphology

Most shorelines in Icy Strait are bedrock controlled, and the sediment on most beaches records a legacy of repeated glaciations. Along much of this coastline, rocky headlands enclose pocket beaches whose intertidal sediments are a mixture of silty clay and clastic debris up to large boulder in size. Much of this sediment is glacial in origin, having settled out of glacial-marine sediment plumes, been iceberg rafted onto the beaches during the Little Ice Age, or winnowed out of Pleistocene till and outwash. Where gravel is abundant, the beach heads often contain relict, coarse-clastic beach ridges ([Carter and Orford, 1984](#); [Neal et al., 2003](#)). Most of these are single crested, and they are now stranded above the level of spring tides and covered by dense vegetation. In contrast to bedrock-controlled shorelines along most of Icy Strait, an extensive glacial outwash plain underlies the community of Gustavus ([Fig. 1](#)), near the mouth of the Glacier Bay. The tidal range in Icy Strait between mean lower low water (MLLW) and mean high spring tide (MHST) is around 6 m.

Bedrock and tectonics

Icy Strait ([Fig. 1](#)) is a glacially eroded trough that cuts across the grain of the regional bedrock geology. Bedrock consists of metamorphosed turbidites and reefoid carbonates intruded

locally by Neogene basalts and rhyolites ([Gehrels and Berg, 1994](#)). A major active fault system, the Fairweather–Queen Charlotte fault ([Plafker et al., 1994](#); [Bruhn et al., 2004](#)), lies 50 km west of Icy Strait. The most recent great earthquakes along this fault occurred in 1899 ([Tarr and Martin, 1912](#); [Bruhn et al., 2004](#)) and 1958 ([Plafker et al., 1978](#)). Uplift associated with these historical earthquakes was localized along the Gulf of Alaska coastline and did not affect Icy Strait.

Glacial history

The radiocarbon ages of shells found in uplifted glaciomarine sediments in the Juneau area ([Miller, 1973, 1975](#)) and in Muir Inlet in upper Glacier Bay ([Fig. 1](#)) ([McKenzie and Goldthwait, 1971](#)) suggest that Icy Strait was deglaciated before ca. 13,500 cal yr BP (when corrected for the marine-reservoir effect using a $\Delta R=470$ yr). Little is known about the extent of glaciers in Glacier Bay during the early Holocene except that they terminated up-fjord from their present positions. Between about 5000 and 4200 ^{14}C yr BP (5000–4500 cal yr BP), glacial outwash buried trees growing near sea level near the intersection of the West Arm and Muir Inlet (G.P. Streveler and A. Post, personal communication, 2006), suggesting that glaciers in the West Arm were advancing at this time.

In the late Holocene, Muir Inlet was dammed twice by ice advancing down the West Arm ([McKenzie and Goldthwait, 1971](#); [Goodwin, 1988](#)). The first dam formed about 2500 ^{14}C yr BP (ca. 2600 cal yr BP) and was followed by a glacier retreat after ca. 2000 ^{14}C yr BP (ca. 1950 cal yr BP) that allowed the lake in Muir Inlet to drain. A second damming occurred ca. 1700 ^{14}C yr BP (ca. 1600 cal yr BP), and the resulting lake persisted in Muir Inlet until perhaps as late as 900 ^{14}C yr BP (ca. 850 cal yr BP) ([Goodwin, 1988](#)). Sometime after 850 ^{14}C yr

BP (ca. 800 cal yr BP), ice flowing out of the West Arm and Muir Inlet coalesced, and by AD 1500 the glacier terminus lay in the middle reaches of Glacier Bay. The LIA maximum in Glacier Bay was reached ca. AD 1750 with the glacier terminus protruding into Icy Strait (Fig. 1). In AD 1794 when the Vancouver expedition charted the area (Lamb, 1984), the glacier terminus had retreated slightly but still lay within 15 km of its LIA limit. Around AD 1800 the rapid retreat began that caused glacier recession of 120 km in less than 160 yr (Klotz, 1899; Field, 1947; Lawrence, 1958; McKenzie and Goldthwait, 1971; Clague and Evans, 1993; Koppes and Hallet, 2002).

Relative sea level history

Prior work on RSL history in the region has been limited. After deglaciation of the Juneau area ca. 13,000 ^{14}C yr BP, marine waters reached elevations as high as 230 m there (Miller, 1972). The late Wisconsin deglaciation of Muir Inlet in upper Glacier Bay was accompanied by a RSL highstand reaching >60 m above present RSL sometime prior to 10,000 ^{14}C yr BP (McKenzie and Goldthwait, 1971). In Icy Strait itself, the only previous study was reconnaissance work done by T.D. Hamilton in the course of archaeological excavations at Ground Hog Bay near Point Couverden (Ackerman et al., 1979; T.D. Hamilton, personal communication, 1983). Hamilton inferred that Icy Strait was deglaciated about 13,500 ^{14}C yr BP at a time when RSL was some 70 m higher than today. He also described a major marine transgression occurring between 400 and 250 ^{14}C yr BP. Presently, RSL is falling in Icy Strait at the rate of 10–18 mm yr $^{-1}$ because of isostatic rebound following deglaciation at the close of the LIA (Motyka, 2003; Larsen et al., 2003, 2004).

Methods

We used differential leveling with closure to measure elevations. Where dense vegetation prevented leveling, we ran multiple circuits with a surveying altimeter. Our leveling had an error of ± 2 cm, and the altimetry, even with repeated circuits, had an uncertainty of ± 50 cm (Table 1). We used mean lower low water (MLLW) as an elevation datum and estimated it by leveling to the most recent high tide line on shorelines sheltered from waves and then correcting this altitude using published tide tables (<http://tidesandcurrents.noaa.gov>).

Despite the precision of our measurements, our measurements of sea level had moderate, unavoidable uncertainties. Even along sheltered shorelines, there was a 0 to -25 cm uncertainty in the level of the last high tide due to wave run-up (Table 1). An additional ± 50 -cm error is involved in making tidal corrections because tide levels vary by that much between tide-table correction localities in Icy Strait. Relief on raised barrier beaches varies by ~ 50 cm (see also Zeeburg et al., 2001), which introduces a comparable uncertainty into the elevations of beach ridges. Observations in Icy Strait today indicate that unless drift logs are in an advanced state of water-logging and decay they are deposited at the level of MHST ± 1 m. The combined uncertainties in elevation measurements are calculated as the square root of the sum of all the squared errors (Table 1). In the data repository, we also have assigned estimated errors to the altimeter measurements reported by other workers.

The altitudinal zonation of biota provides information on ancient and recent RSL. Of the marine organisms we ^{14}C -dated (data repository), the scallop *Chlamys islandica* lives subtidally, and cockles of the genera *Clinocardium* and *Serripes* live today in Glacier Bay below the $+3$ m level. Dated *Balanus* barnacle plates could not be identified to species, but today barnacles of that genus live up to around the $+3$ m level. The lowest altitudes reached by several plant species are proxies for particular tide levels. Sitka spruce (*Picea sitchensis*) grows down to approximately the level of mean higher spring tide (MHST) along shorelines in Icy Strait (Motyka, 2003). Cow parsnip (*Heracleum lanatum*) can grow about a meter lower on shorelines than Sitka spruce (Table 1) at a level corresponding roughly to mean higher high water (MHHW).

We used the altitudes of the highest barrier beaches now forested by spruce trees and displaying immature soil profiles to estimate the height of the most recent high stand of RSL. Sitka spruce forms nearly pure stands on newly uplifted land surfaces in Southeast Alaska. Western hemlock (*Tsuga heterophylla*) typically invades the understory of spruce forests within several centuries (Fastie, 1995), and podzolic soils develop after a similar period (Ugolini and Mann, 1979). Consequently, we identify barrier beaches as having formed within the last 300 yr if they support pure stands of spruce, and if their soil profiles lack the distinctive E horizons of podzolic soils. We estimated ages for five spruce trees growing on each of eight different barrier beaches by counting annual rings in increment cores. We

Table 1
Uncertainties in elevation measurements

	Uncertainty in	Magnitude of uncertainty (m)	Type of measurement	Total uncertainty (m)
A	Differential leveling with closure	± 0.02	Leveling beach ridge crest elevation (A, C, D, E)	$+1/-0.75$
B	Altimeter circuits	± 0.5	Using altimeter to measure beach ridge crest elevation (B, C, D, E)	$+1.25/-1$
C	Geographic variability in tide-table correction	± 0.5	Leveling elevation of stump, stratigraphic section, or live plant (A, C, E)	$+0.75/-0.5$
D	Beach ridge height	± 0.5	Using altimeter to measure elevation of stratigraphic section (B, C, E)	$+1/-0.75$
E	Relation of swash line to last high tide	$-0.25/0$	Comparisons between elevations of living plants, sections, and tide levels by leveling (A, C, E; A, C, E)	$+1/-0.75$
F	Deposition of fresh log at MHST	± 1	Comparisons between elevations of living plants, logs embedded in barrier beaches, and tide levels by leveling (A, C, E; A, C, E, F)	$+1.5, -1.25$

used a modified Livingstone corer to core Goose Island Pond. Bark, root, and leaf remains were identified in the field using a hand lens.

We radiocarbon dated the outermost 1–2 cm of wood from logs and tree stumps. Radiocarbon ages are calibrated to calendar years using CALIB 5 (<http://calib.qub.ac.uk>). The ^{14}C ages of marine shells and of barnacles are corrected for the marine-reservoir effect (Stuiver and Polach, 1977; Kovanen and Easterbrook, 2002; Hutchinson et al., 2004a) using a ΔR of $=470$ yr ($\text{SD}=76$) as determined for recent marine shells along the Southern Alaskan coast (<http://calib.qub.ac.uk/marine>).

Results

Raised barrier beaches and relative sea level

Numerous barrier beaches now stranded above tide level provide evidence for a recent RSL high stand in Icy Strait. These barriers occur at cove heads and stream mouths where they dam ponds and wetlands. As detailed above, a post-AD

1700 age for these barriers is indicated by the spruce forests and poorly developed soils that cover them. Ring counts reveal that spruce trees established on the crests of these barriers as early as AD 1839, with most becoming established between that year and AD 1870. The crest altitudes of the raised barriers range from 8 to 12 m above present MLLW (Table 2). Within Icy Strait, there is no obvious relationship between the barriers' altitudes and their locations relative to Glacier Bay (Fig. 2). Instead, variations in barrier altitudes seem to relate to local wave fetch and exposure. By comparing the altitudes of barrier crests with the altitudes of the lowest living trees on their seaward slopes, we can estimate the rise in RSL that created these barriers. The average difference between the altitudes of sixteen barrier crests and of the lowest spruce growing on nearby shorelines is 3.0 m ($+1/-0.75$) (Table 2).

Radiocarbon dating of relative sea level

Twenty-eight radiocarbon ages are relevant to the RSL changes that have occurred in Icy Strait since 13,500 ^{14}C yr BP

Table 2
Elevations of Little Ice Age barrier-beach crests in Icy Strait

Site ^a	Location		Altitude (m) highest barrier crest (m) above MLLW ^b	Altitude (m) of barrier crest above lowest, living spruce ^c	Altitude (m) of barrier crest above lowest living cow parsnip ^d
	Lat (°N)	Long (°W)			
Cooper's Notch, lower Glacier Bay	58° 26.6'	135° 53.7'	9.7	2.2 (+1, -0.75)	—
Strawberry Island, Beardslee Islands, Glacier Bay	58° 30.4'	136° 0.7'	8.7	—	—
Islet east of Strawberry Island, Beardslee Islands, Glacier Bay	58° 31.5'	135° 57.3'	8.6	—	—
Swanson Harbor, Point Couverden, Chilkat Peninsula	58° 12.6'	135° 6.5'	8.7	—	3.1
Entrance Island, Point Couverden, Chilkat Peninsula	58° 11.9'	135° 6.1'	8.6	2.8	3.3
Ground Hog Bay, Chilkat Peninsula	58° 14.1'	135° 13.1'	11.7	5.1	5.6
"Cemetery Point", Chilkat Peninsula	58° 14.6'	135° 18.4'	9.0	5.1	—
Black Ole's Cabin, Big Porpoise Island	58° 19.6'	135° 27.5'	9.3	3.0	4.0
Lemesurier Island (northeast), North Passage	58° 18.2'	136° 4.7'	9.0	3.5	4.1
Lemesurier Island (northwest), North Passage	58° 18.0'	136° 5.9'	9.9	2.7	4.3
Point Gustavus	58° 23.2'	135° 55.0'	10.6	2.9	5.0
"Gravel Canyon", Pleasant Island	58° 19.6'	135° 39.1'	9.4	2.3	—
"Cruise Ship Rock", South Passage	58° 13.2'	136° 6.8'	8.7	2.1	—
"Dancing Bear Creek", South Passage, Chichagof Island	58° 13.0'	136° 5.7'	8.3	3.3 (inside creek mouth) 2.3 (outer beach)	—
Goose Island Pond, Quartz Point, Chichagof Island	58° 12.8'	136° 2.3'	9.1	3.5	4.0
Site #1, Mud Bay–Point Adolphus, Chichagof Island	58° 12.4'	135° 58.0'	9.8	1.9	2.7
Site #2, Mud Bay–Point Adolphus, Chichagof Island	58° 12.7'	135° 57.2'	10.6	—	3.8
"Old Canoe Creek" section, Point Adolphus, Chichagof Island	58° 13.0'	135° 56.6'	9.8	2.5	—
"Hooterville", Point Dundas	58° 19.1'	136° 15.4'	9.6	3.6	—
Cove east of Salt Chuck Entrance, North Passage	58° 20.6'	136° 11.1'	10.2	4.1	—
Pinta Cove, Point Adolphus	58° 15.9'	135° 45.7'	9.9	2.7	3.2
Southeastern Dundas Bay	58° 21.4'	136° 16.9'	10.5	4.9	5.4
Lena Cove, Favorite Channel, Juneau area	58° 23.6'	134° 44.9'	9.3 I (9.8 II)	—	3.9
Echo Cove, Lynn Canal, Juneau area	58° 39.5'	134° 54.5'	8.7	2.2	—
Outer Point, Douglas Island, Juneau area	58° 17.8'	134° 40.7'	10.1 I (9.7 II) (+1.25, -1)	—	6.2
Icy Point, Gulf of Alaska	58° 23.2'	137° 6.0'	11.6 (+1.25, -1)	—	—
Kaknau Creek mouth, Palma Bay, Gulf of Alaska	58° 24.0'	137° 4.4'	11.5 (+1.25, -1)	—	—

^a Informal place names in quotations.

^b Error=(+1, -0.75) unless otherwise noted.

^c Error=(+1, -0.75) unless otherwise noted.

^d Error=(+1, -0.75).

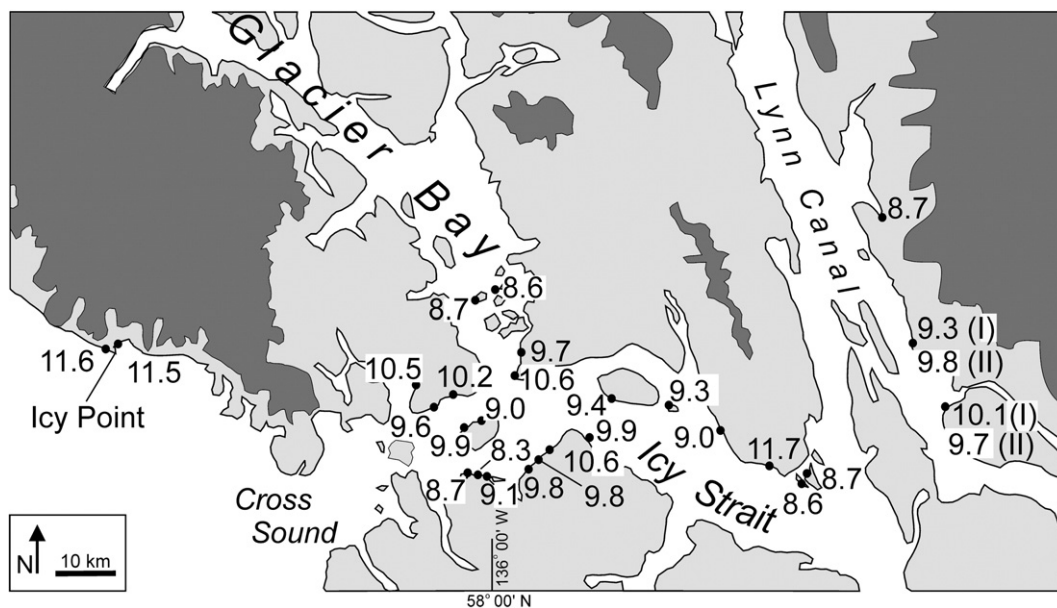


Figure 2. Altitudes (m) of vegetated, raised barrier-beach crests above present mean lower low water (MLLW). Along the eastern shore of Lynn Canal, two raised barriers occur in some places, and the older (inner) and outer (younger) paired ridges are labeled (I) and (II), respectively.

(14,270–15,190 cal yr BP) (Beta-62751) (Data Repository). To save space, we present here the five most informative and complicated stratigraphic settings associated with these dates and describe the remaining dates and their stratigraphies to the data repository.

Ground Hog Bay

The higher of two terraces (T1) at Ground Hog Bay (Fig. 1) is cut into bedrock and overlain by a veneer of clay-rich diamict, gravel, and colluvium (Fig. 3). The narrower lower terrace (T2) is a constructional feature of sand and gravel associated with the creek mouth. Both terraces exhibit mature podzolic soils and are covered by forests dominated by old-growth western hemlock. East of the creek mouth, a prominent barrier beach is covered by a youthful spruce forest. We studied a section through T1 at whose base is 40 cm of sandy diamict containing faceted and striated cobbles and rare shell debris. This basal unit coarsens upwards and is overlain by 30 cm of boulder diamict containing barnacles still attached to stones. Barnacle plates from this unit date to $13,070 \pm 60$ ^{14}C yr BP (13,800–14,480 cal yr BP) (Beta-86362) (data repository). The unit containing the barnacles is in turn overlain by 35 cm of boulder diamict with fewer fines than the lower units. An erosional boundary at the top of the boulder diamict is overlain by 80 cm of massive and planar-bedded sand and gravel.

The glacial-marine diamict in the base of this section probably records deposition in deep water distant from a glacier margin. The coarsening upward trend probably reflects the winnowing of fine sediment by wave action, culminating in a lag of boulders immediately underlying the beach deposits forming the upper half of the section. The barnacles dating to 13,070 ^{14}C yr BP could have been reworked from older sediment and deposited below their contemporary RSL, so their present altitude of 11.5 m above MLLW indicates only that the

associated RSL lay >8.5 m higher than today (data repository). Hamilton also found barnacles and shells in glacial-marine sediment on T1, which he dated to $13,350 \pm 100$ ^{14}C yr BP (14,110–15,030 cal yr BP) (SI-2114) (Ackerman et al., 1979; T.D. Hamilton, written communication, 1983).

“Old Canoe Bay”, Mud Bay–Point Adolphus

The upper of two terraces (T1) is incised into bedrock, and the lower terrace (T2) is composed of unconsolidated sediment and occurs only near creek mouths (Fig. 4). Both terraces have distinctly older forest vegetation and soils than the raised barrier beach bordering the shore. Two meters of glacial-marine diamict overlying bedrock is exposed at the base of a section through the seaward edge of T1. The diamict is overlain above an erosional contact by 1.2 m of planar-bedded sand containing marine shells. Some of these bivalves are still paired, but none are in growth position. One of these shells dates to $13,240 \pm 90$ ^{14}C yr BP (14,000–14,910 cal yr BP) (Beta-86377) (data repository). This lower sand unit coarsens slightly upward into planar-bedded, medium sand lacking shells. Above it is 1.5 m of crudely stratified, seaward-dipping gravel.

The basal diamict in this section probably represents ice-distal, glacial-marine deposits. The shell-rich sand unit probably records a shallow, subtidal environment. Assuming the mollusk we dated lived somewhere between 3 m above its contemporary MLLW and tens of meters below it, RSL ca. 13,240 ^{14}C yr BP (14,000–14,910 cal yr BP) was >11 m higher than today (data repository).

Eastern Dundas Bay

At this site, beach meadows slope down to mud flats on the seaward side of a gravel barrier now standing 10.5 m above present MLLW. Two lines of conifer stumps protrude from the surface of this mud flat. Between 60 and 100 stumps were

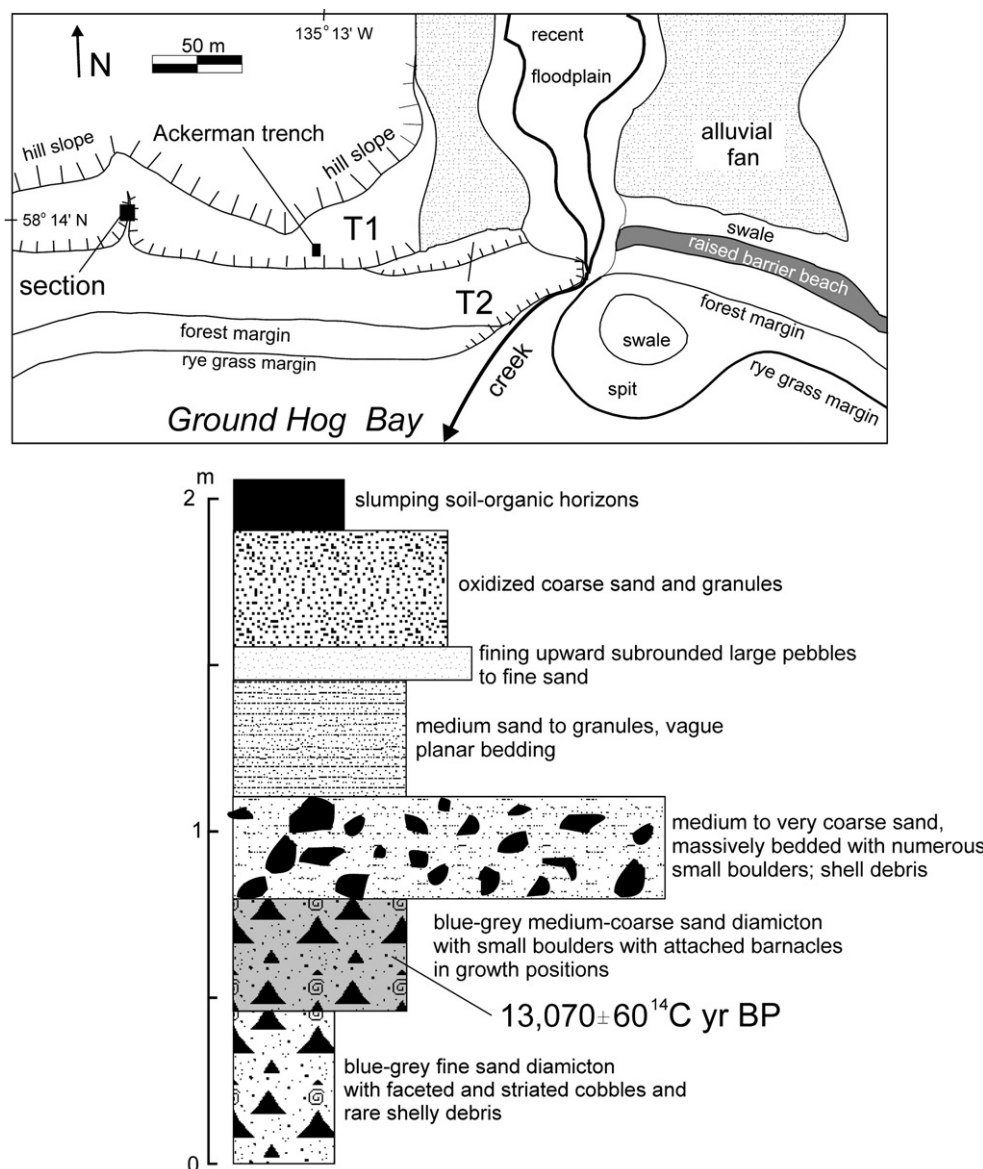


Figure 3. Setting and stratigraphy of late Pleistocene glacial-marine deposits on Terrace T1 at Ground Hog Bay.

visible in 1995; most were 10–30 cm in diameter, and all were rooted in a peaty soil. The anatomy of the bark fragments indicates that these trees were Sitka spruce. One stump dated to 1570 ± 70 ^{14}C yr BP (1320–1610 cal yr BP) (Beta-86369) and another to 1640 ± 60 ^{14}C yr BP (1400–1700 cal yr BP) (Beta-86370) (data repository). Probing revealed that these stumps were rooted on low, parallel ridges separated by a swale approximately 50 cm deep. A pit dug into the crest of the shoreward ridge (Fig. 5) showed that these tree stumps were rooted in a layer of sedge peat that buried the gullied surface of a deposit of blue–gray, silty sand. Our pit intersected part of a steep-sided gully. Sedge leaves and the cone of a western hemlock tree from the peaty sediment filling this gully dated to 4440 ± 60 (Beta-86371) and 4420 ± 50 ^{14}C yr BP (Beta-86372) (4870–5290 and 4860–5280 cal yr BP, respectively) (data repository). The peat-forming sedge was *Carex aquatilis*, a species

that does not tolerate salt water. The *in situ* roots of another salt-intolerant taxon, horsetail (*Equisetum* sp.), were also present.

This site in eastern Dundas Bay records two separate episodes when RSL was lower than today. The first occurred ca. 4430 ^{14}C yr BP (4860–5290 cal yr BP) and the second ca. 1600 ^{14}C yr BP (1320–1700 cal yr BP). The blue–gray sand exposed in the pit wall represents an upper-intertidal mud flat deposited before 4440 ^{14}C yr BP (4870–5290 cal yr BP). This gully probably was cut as RSL fell and streams incised into the abandoned tidal flat. Then a freshwater sedge marsh spread over the low-gradient surface and infilled the gully with peat. MLLW probably was below the -3.9 m level ca. 4430 ^{14}C yr BP (4860–5290 cal yr BP) (data repository). Sometime after that, RSL rose across the site; it then fell again prior to ca. 1640 ^{14}C yr BP (1400–1700 cal yr BP), allowing trees to colonize the beach ridges as drainage conditions improved. On

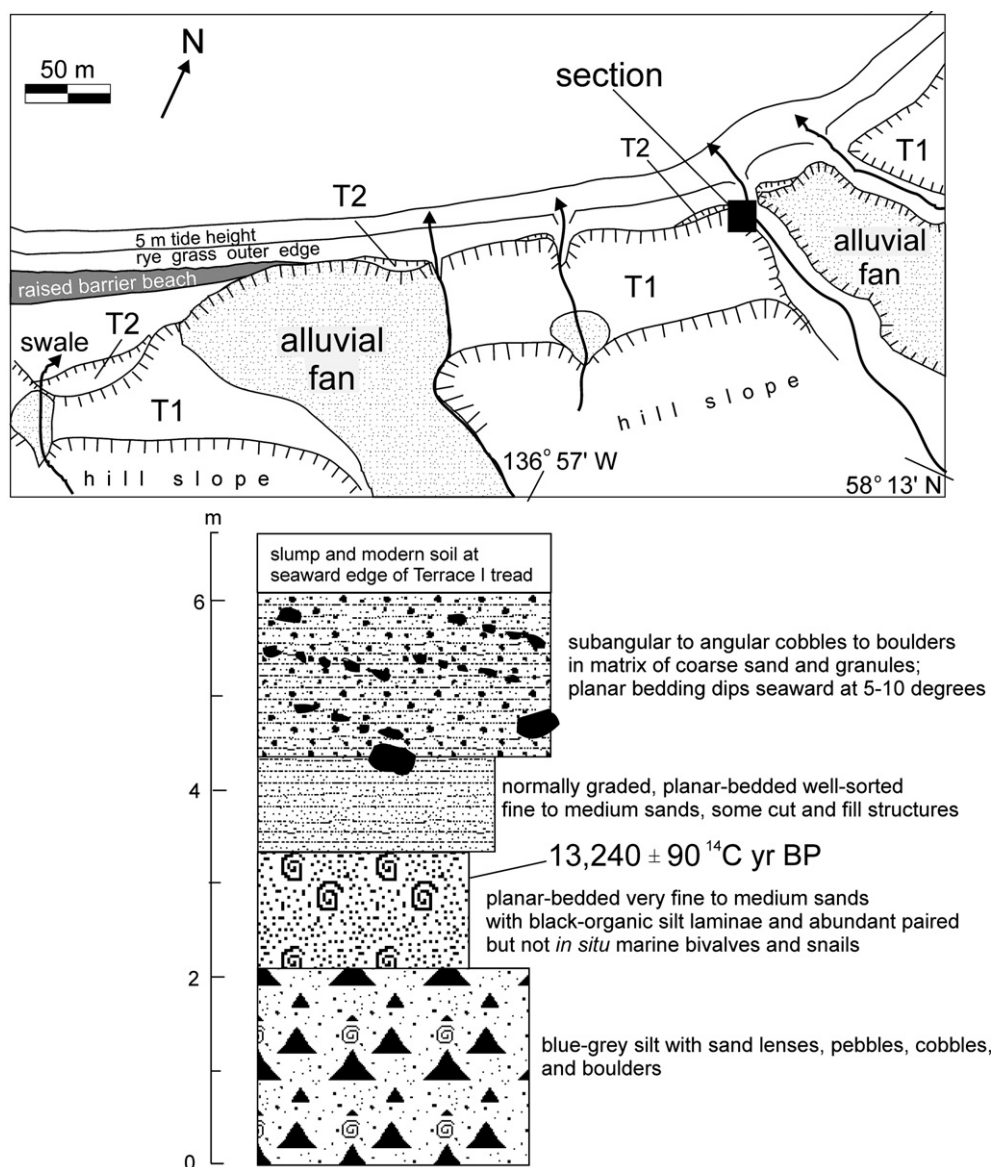


Figure 4. Setting and stratigraphy of late Pleistocene glacial-marine deposits at Old Canoe Bay between Point Adolphus and Mud Bay.

the present shoreline, spruce trees grow down to the +5.6 m level. It follows that when a youthful forest grew on the crests of the ridges, between 1640 ^{14}C yr BP (1300–1700 cal yr BP) and 1570 ^{14}C yr BP (1320–1610 cal yr BP), MLLW was at least 2.8 m lower. The absence of tree stumps in the intervening swale suggests that the water table was high at that time. After 1570 ^{14}C yr BP (1320–1610 cal yr BP), RSL again rose over this site, killing these trees.

Goose Island Pond

Two small ponds are dammed behind raised barrier beaches on Goose Island. We took two sediment cores, 2 m apart, through the floating mat at the margin of the larger, southern pond (Fig. 6). These cores penetrate a near-surface layer of fen (freshwater) peat overlying 50 cm of blue–gray, silty gravel containing fragments of marine shells. Under this are 2.5 m of sedge peat and then 60 cm of algal mud (gytija). Below the

gytija are 30 cm of horizontally laminated, cream-colored volcanic ash. More gytija lies beneath this tephra below an abrupt boundary. In the lowest 5 cm of the gytija underlying the tephra, we found the seeds of *Arctostaphylos* sp., a dwarf shrub that grows today as an early colonizer of newly deglaciated surfaces in Glacier Bay. Different samples of these seeds from the two different cores dated to $12,030 \pm 80$ (13,730–14,060 cal yr BP) (Beta-186945) and $12,200 \pm 40$ ^{14}C yr (13,940–14,180 cal yr BP) (Beta-124175) (data repository). More blue–gray, silty diamicton with shell fragments lies below the lower gytija unit.

The silty diamicton in the upper part of the Goose Island Pond cores records the recent entry of marine waters into the pond basin at a time when glaciomarine sediments were abundant in the waters of Icy Strait. This RSL high stand flooded a formerly freshwater basin as indicated by the freshwater fen peat and algal mud. Lower in the cores, the color, texture, and

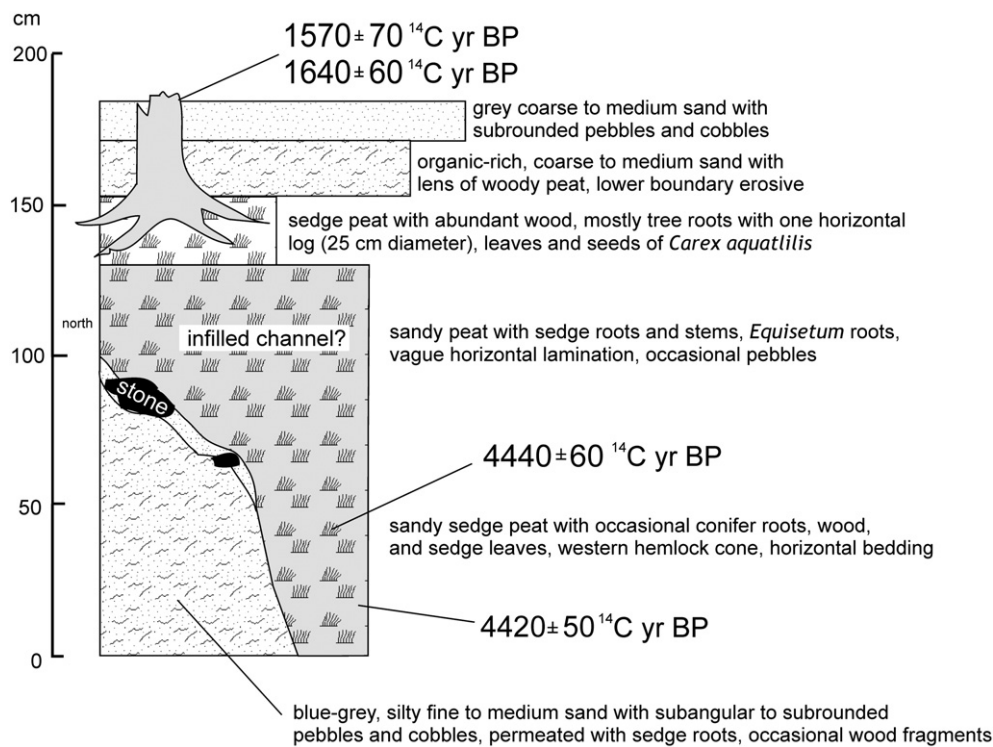


Figure 5. Stratigraphy of pit in the intertidal zone, southeastern Dundas Bay.

approximate thickness of the tephra matches the one erupted from Mount Edgecumbe near Sitka ca. 11,000 ^{14}C yr BP (ca. 12,420–13,260 cal yr BP) (Riehle et al., 1992) (data repository). *Arctostaphylos* grows on open ground in immature soils and is not salt tolerant, so its presence suggests a treeless setting recently exposed by falling RSL and/or retreating glaciers. The *Arctostaphylos* seeds must have washed into the pond from the adjacent land surface. The elevation of the transition from glaciomarine to freshwater sediment (+2.35 m) provides an important index point for RSL in these cores. Comparison with

modern tide levels (data repository) suggests that ca. 12,030–12,180 ^{14}C yr BP (13,730–14,160 cal yr) MLLW was at least 3 m lower than today. RSL remained below the sill elevation of Goose Island Pond (+6 to +7.5 m) until the recent transgression occurred that deposited the glaciomarine sediment high in the core.

Goose Island beach pit

We dug a 2.25-m deep pit on the shore face adjacent to Goose Island Pond (Fig. 7). The lowest 30 cm of sediment exposed in this pit is massively bedded, inorganic silt and sand. This is

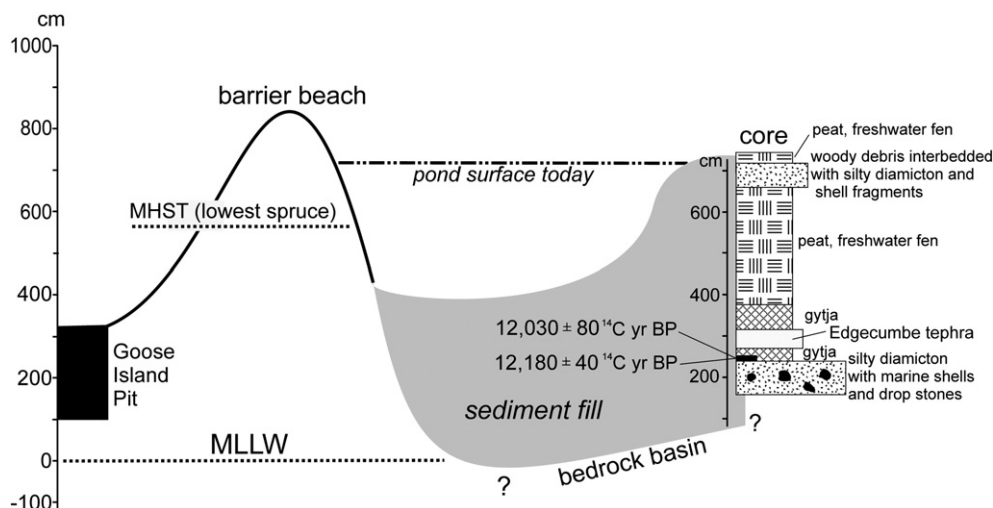


Figure 6. Goose Island Pond showing stratigraphy of the sediment cores and their relationship to the Goose Island pit, present MLLW level and the barrier-beach crest.

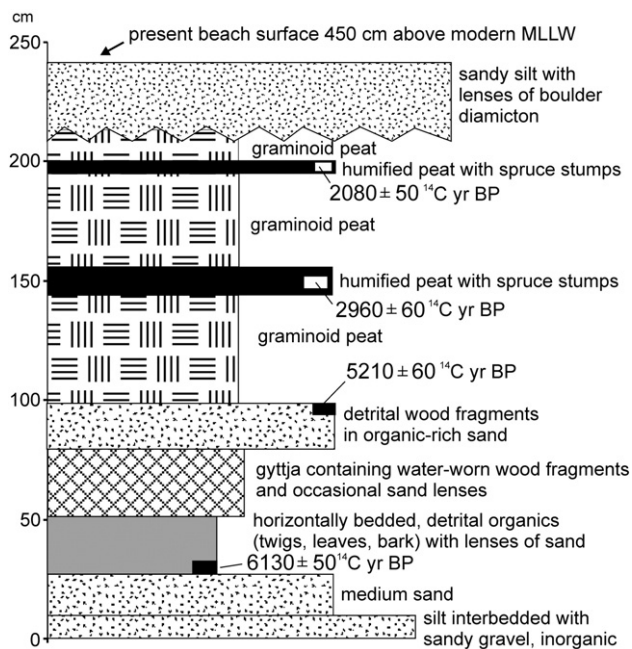


Figure 7. Stratigraphy of the Goose Island pit in the beach adjacent to Goose Island Pond.

overlain by 20 cm of horizontally bedded, detrital organic material composed of bark fragments, twigs, and conifer needles. A twig near the base of this unit dates to 6130 ± 50 ^{14}C yr BP (6890–7160 cal yr BP) (Beta-145850) (data repository). Overlying this organic unit along a conformable boundary is 40 cm of gyttja containing water-worn wood fragments and occasional sand lenses. A 10-cm layer of medium sand containing wood fragments overlies this gyttja. Wood from this sand unit dates to 5210 ± 60 ^{14}C yr BP (5770–6180 cal yr BP) (Beta-145851).

Overlying the sand is the first of three layers of crudely stratified sedge peat that preserve the stem and leaf fragments of the salt-tolerant sedge, *Carex lyngbyei*. The lowest of these sedge-peat units is 45 cm thick. Above a conformable boundary, the lowest peat layer is overlain by 10 cm of well-decomposed muck containing small-diameter (<5 cm) conifer stumps and roots. The anatomy of the bark fragments and roots indicates that these trees were Sitka spruce. A spruce root from this humified peat layer dates to 2960 ± 60 ^{14}C yr BP (2960–3330 cal yr BP) (Beta-145852). This forest layer is overlain conformably by 50 cm of layered, graminoid peat, which in turn is overlain by another layer of well-decomposed muck containing small-diameter spruce stumps and branches. A twig from this upper forest layer dates to 2080 ± 50 ^{14}C yr BP (1900–2300 cal yr BP) (Beta-145853). More sedge peat separates this upper forest layer from the crudely stratified sand and silt mixed with bouldery diamicton that forms the surface of the beach today.

The lack of marine shells in the silt and sand at the section's base suggests deposition in freshwater. The layer of detrital organics dated to 6130 ^{14}C yr BP (6890–7160 cal yr BP) probably was deposited in a back-beach setting, and the overlying deposit of algal mud confirms the presence of a freshwater pond. Along this shoreline today, MHHW (cow

parsnip level) is about 5.1 m above MLLW. If we assume that the lowest altitude a freshwater pond could have existed was at the level of MHHW, then ca. 6130 ^{14}C yr BP (6890–7160 cal yr BP), MLLW must have been below the -4.1 m level (data repository). RSL was still relatively low ca. 5210 ^{14}C yr BP (5770–6180 cal yr BP) when detrital wood was deposited in a sand unit lacking marine shells. A back-beach depositional setting is again suggested.

Some time after about 5200 ^{14}C yr BP, the area became a salt marsh, most likely in response to a rise in RSL. This marsh was invaded by spruce trees ca. 2960 ^{14}C yr BP (2960–3330 cal yr BP), suggesting that RSL had fallen below the -3.3 m level. Judging from the small diameter of these tree stumps, this regression was short-lived, perhaps only several decades in duration, at least at this height on the paleo-shoreline. It could have lasted significantly longer at higher levels on the shore face. Salt marsh then covered the site again, suggesting a rise in RSL. This was followed ca. 2080 ^{14}C yr BP (1900–2300 cal yr BP) by a second, short-lived fall in RSL that allowed spruce trees to invade the marsh again. MLLW lay below the -2.8 m level at this time (data repository); then RSL rose again. The capping layer of silt and sand represents the recent RSL high stand that also flooded Goose Island Pond with glaciomarine sediment (Fig. 6).

Discussion

Relative sea level history in Icy Strait

The late Pleistocene

Three of the eight ^{14}C dates that are older than $12,200$ ^{14}C yr BP can be used to estimate the level of their contemporary RSL (Fig. 8). Hamilton (personal communication, 1983) dated shells and barnacles from a section approximately 30 m above present sea level near Hoonah to $13,420 \pm 130$ ^{14}C yr BP (14,130–15,150 cal yr BP) (SI-2082) (data repository). The shells he dated were associated with what he termed “marsh grass,” which we infer to have been eel grass, *Zostera* sp., a plant that typically grows within several meters of MLLW. If this is correct, the shells Hamilton dated were living near the level of their contemporary MLLW. The exact error associated with Hamilton's altimetry is unknown, but we assign it an error of ± 5 m.

The most informative RSL index points during the late Pleistocene are the altitudes of the two raised deltas along Falls Creek near Gustavus (data repository). The surface of the highest delta is at 56.5 m elevation and contains shells dating to $12,960 \pm 60$ ^{14}C yr BP (13,750–14,150 cal yr BP) (Beta-222030). The surface of the lower delta is located at 29.5 m elevation and contains shells dating to $13,150 \pm 80$ ^{14}C yr BP (13,870–14,760 cal yr BP) (Beta-133352). In both cases, the dated mollusks lived in the deltas' bottomset beds. The other five late Pleistocene dates on shells or barnacles provide only minimum-limiting estimates on RSL between $12,980$ and $13,500$ ^{14}C yr BP (13,690–15,190 cal yr BP). The maximum possible age span between the establishment of the upper Falls Creek delta at 56.5 m and the fall of RSL below the sill elevation of Goose Island Pond is only 420 cal yr, which indicates that RSL was

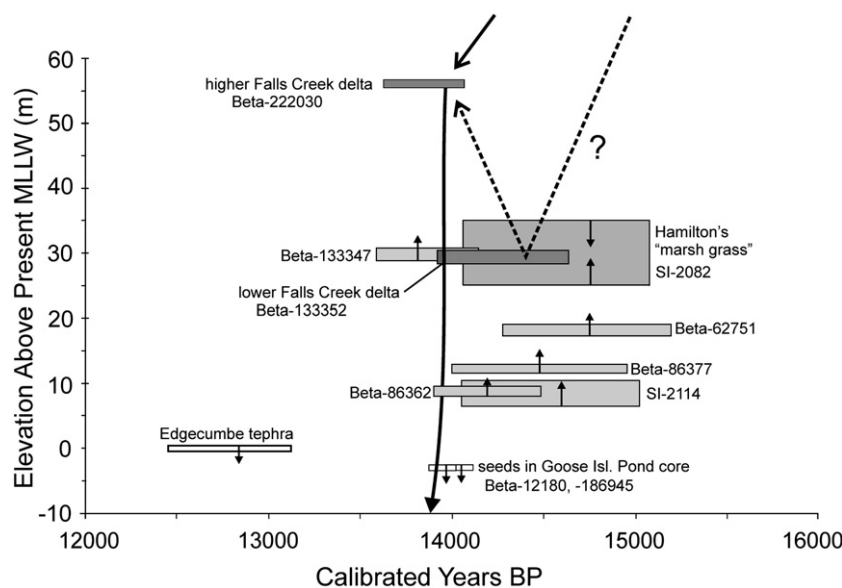


Figure 8. Synthesis of RSL history in Icy Strait 15,500 to 12,500 cal yr BP. In the data repository, we describe how each date is corrected, calibrated, and interpreted in terms of RSL. Boxes around date symbols show 2σ age errors (horizontal) and elevation measurement errors (vertical). Two different reconstructions for the course of RSL change (heavy and dashed lines) are given for the existing data. Small arrows show the sense of the limiting ages provided by the radiocarbon ages.

dropping very rapidly around 13,900 cal yr BP. The presence of un-reworked Edgumbe tephra near high tide line at Gull Cove Point (data repository) indicates that RSL had fallen below the -0.3 m level before this tephra fell sometime between 11,200 and 10,800 ^{14}C yr BP (12,420–13,260 cal yr BP) (Engstrom et al., 1990; Riehle et al., 1992; Hansen and Engstrom, 1996).

It is interesting that the only uplifted marine terraces in Icy Strait that are cut into bedrock are pre-Holocene in age. We found laterally discontinuous, uplifted, wave-abrasion platforms near Black Rock on Pleasant Island (Fig. 1), at Ground Hog Bay (Fig. 3), and at Old Canoe Bay (Fig. 4). In all three cases, a pre-Holocene age is indicated by the presence of glacial striae and/or glacial sediments on the bedrock tread of the uplifted terrace. The lower terraces at Ground Hog Bay and Old Canoe Bay are constructional features of sand and gravel that probably represent the surfaces of deltas deposited during RSL fall in the late Pleistocene. The discontinuous occurrence of the raised bedrock terraces probably reflects their differential erosion by overriding glaciers.

The Holocene

Our inability to find any stratigraphy relating to RSL between 11,000 and 6000 ^{14}C yr B.P. suggests that RSL lay at least several meters below its present level during the early Holocene. The record resumes ca. 6130 ^{14}C yr BP (6890–7160 cal yr BP) in the Goose Island pit (Fig. 7) when MLLW lay below the -4.1 m level (Fig. 9). At ca. 5210 ^{14}C yr BP (5770–6180 cal yr BP), RSL was still below the -3.4 m level. The peat-filled channel in Dundas Bay suggests that MLLW was below the -2.27 to -4.75 m level 4420 to 4870 ^{14}C yr BP (4860–5290 cal yr BP). A rise in RSL of perhaps 1 m occurred prior to ca. 2960 ^{14}C yr BP (2960–3330 cal yr BP), which allowed a sedge marsh to grow at the site of the Goose Island

pit. At 2960 ^{14}C yr BP, spruce trees were growing there on the surface of the former marsh (Fig. 7), suggesting that RSL had fallen below the -3.3 m level. Then RSL rose, and sedge marsh covered the site again. The presence of a spruce stump near the Goose Island pit dating to 2410 ^{14}C yr BP (2340–2710 cal yr BP) (data repository) suggests that RSL level during this high stand was at least 2 m lower than today. RSL probably fell below the -2.8 m level again when spruce trees re-established themselves on the former marsh surface ca. 2080 ^{14}C yr BP (1900–2300 cal yr BP). A subsequent rise in RSL is indicated by the sedge peat that buries these tree stumps in the Goose Island pit (Fig. 7).

Five dates on tree stumps define a fall in RSL to between -1 and -4 m between 1400 and 1900 ^{14}C yr BP (1300–1900 cal yr BP) (Figs. 9, 10) (data repository). After this low stand, RSL rose to around the $+0.4$ m level, depositing a drift log in barrier-beach gravels at Cemetery Point ca. 1310 ± 60 ^{14}C yr BP (1070–1310 cal yr BP) and inundating trees along the Mud Bay River ca. 1320 ± 60 ^{14}C yr BP (1080–1340 cal yr BP). RSL then probably fell to around the -2.4 m level and deposited a drift log in the barrier beach at Dancing Bear Creek ca. 800 ± 60 ^{14}C yr BP (660–900 cal yr BP). It then rose once more, burying forest and marsh under a barrier beach east of the Salt Chuck between 660 ± 60 ^{14}C yr BP (540–690 cal yr BP) and 510 ± 60 ^{14}C yr BP (460–650 cal yr BP) (Fig. 9).

The forest-covered gravel barrier beaches that are now widespread in Icy Strait were deposited during the RSL high stand that accompanied isostatic depression during the LIA (McKenzie and Goldthwait, 1971; Motyka, 2003). The oldest of the thirty-nine spruce trees we dated from barrier crests germinated in AD 1841. There can be a significant lag between the uplift of a land surface above sea level and tree establishment (Bégin et al., 1993), and Motyka (2003) infers that in Southeast Alaska this lag ranged from several years during the

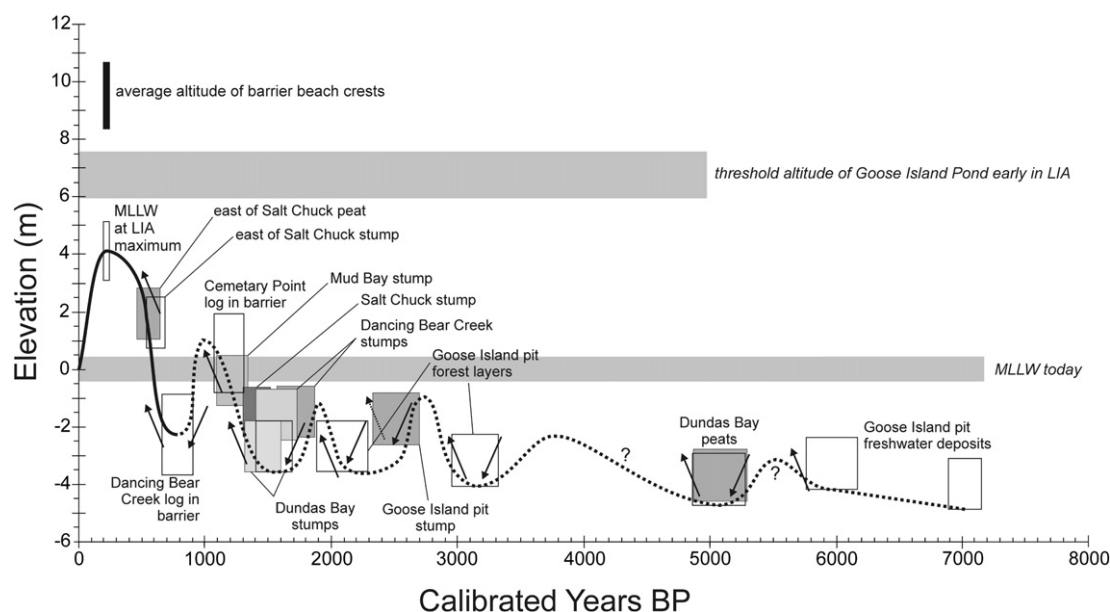


Figure 9. The last 7000 yr of RSL change in Icy Strait plotted on a time scale of calibrated years BP. Boxes around date symbols show 2σ age errors (horizontal) and elevation measurement errors (vertical). In the data repository, we describe how each date is interpreted in terms of RSL. Arrows show sense of RSL change indicated by the depositional environment of the dated material. Marine waters re-entered the basin of Goose Island Pond only during the LIA.

last century to as much as 40 yr early in the nineteenth century. Subtracting 40 yr, from the age of the oldest tree observed suggests that the crests of the LIA barrier beaches were abandoned by the sea ca. AD 1800. In the Juneau area, Motyka (2003) estimates that post-LIA rebound began between AD 1770 and 1790.

Height of the LIA high stand

The exact height of the RSL high stand in Icy Strait during the LIA remains uncertain. The ongoing regime of falling RSL

in Icy Strait means that there are no barrier beaches forming today whose altitudes we can compare with the LIA ones. Based on comparisons between the altitudes of barrier crests and the altitudes of the lowest spruce trees growing on adjacent shore faces (Table 1), the height of the LIA high stand was $3 (+1/-0.75)$ m. Such a comparison between landforms and vegetation, however, is problematic: Did the crest altitudes of LIA barrier beaches equilibrate to MHSST (the level of lowest spruce trees) or to MHHW (the approximate level of the lowest cow parsnip plants)? The average sea level condition represented by MHHW is perhaps the more

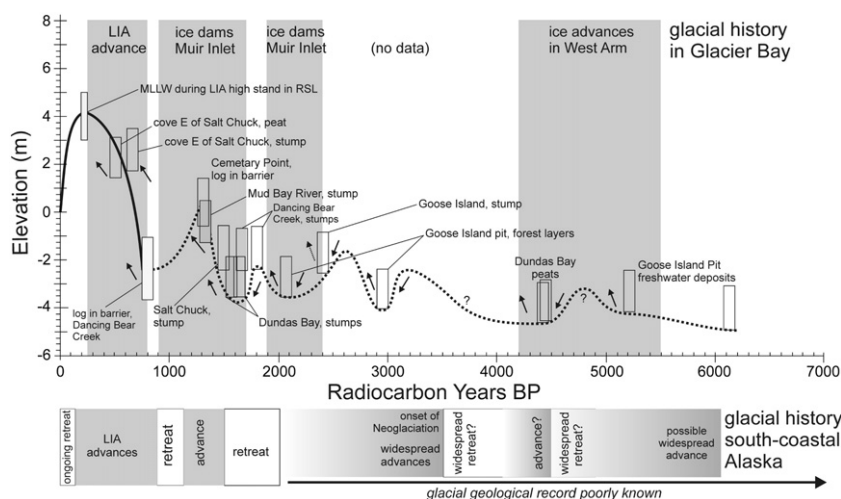


Figure 10. The last 7000 yr of relative sea level change in Icy Strait plotted on a time scale of radiocarbon years BP to facilitate comparison with published records of glacial history. Background shading indicates timing of glacier fluctuations in Glacier Bay from Goodwin (1988). The 4200–5500 ^{14}C yr BP period of glacier advance in the West Arm is based on unpublished ^{14}C dates (G. P. Streveler and A. Post, personal communication, 2006). Glacial history in southern coastal Alaska and coastal British Columbia comes mainly from Calkin et al. (2001), Wiles et al. (2002), Barclay et al. (2006), Reyes et al. (2006), and Osborn et al. (2007). During the interval following 2500 BP, which is the best-known part of the glacial geological record, low stands in RSL in Icy Strait coincided with periods of glacial retreat, while RSL high stands coincided with periods of glacier advance. Both the RSL and glacial records are more poorly known prior to 2500 ^{14}C yr BP, but a similar correspondence between RSL and glacial isostasy is suggested.

reasonable datum, in which case the LIA transgression reached approximately the +4 m level in Icy Strait between AD 1750 and 1800.

Larsen et al. (2005) estimated the height of the LIA transgression between +3 and +5.7 m based on four study sites in Icy Strait. They identified emergent shorelines of post-LIA age using the same criteria of forest composition, tree age, and degree of soil formation that we used, but they equated the marine limit with the base of an erosional scarp. This method involves its own set of inaccuracies, the chief one being that mass movement commonly obscures wave-cut inflection points and can lead to an overestimate of the altitude of the scarp base. We suspect that the true height of the LIA transgression in central Icy Strait lies in the range of +4 m and ± 1 m. Greater precision than that is moot given the variability in wave run-up along this geographically complex coastline.

Discounting tectonism

RSL change was of such large magnitude (>50 m) and so rapid (>10 cm yr⁻¹) in the aftermath of late Pleistocene deglaciation that tectonism had only minor effects on RSL changes in Icy Strait between ca. 16,000 and 13,000 cal yr BP. Discounting the effects of tectonism during the Holocene is less straightforward because the elevation changes caused by glacial isostasy are much smaller. At least over the last 2000 yr, however, multiple lines of evidence suggest that RSL has probably been mainly controlled by isostatic rebound rather than tectonic uplift. The evidence includes:

- 1) the geographical pattern of uplift rates revealed by historical tide-gage records (Hicks and Shofnos, 1965; Hudson et al., 1982; Larsen et al., 2003) corresponds to the former distribution of ice loads in the region;
- 2) geodynamical modeling of ice loading demonstrates that the observed amount and rate of uplift can be accounted for by isostatic rebound following the retreat of Glacier Bay ice that started ca. AD 1750 (Clark, 1977; Larsen et al., 2005);
- 3) the distribution of uplift over the last 200 yr does not correspond to the geography of known faults (Motyka, 2003; Larsen et al., 2005); and
- 4) the timing of RSL change between AD 100 and AD 1200 and during the LIA correlates with the timing of glacier retreat in Glacier Bay, at least as we understand that history now (McKenzie and Goldthwait, 1971; Motyka, 2003).

But what about the role of tectonism earlier in the Holocene? In Icy Strait, dated tree stumps and fen peats intercalated with marine sediment in the present intertidal zone suggest there was no net submergence after 6000 cal yr BP (Fig. 9). Evidence against coseismic, tectonic uplift comes from the lack of raised marine terraces in Icy Strait, in striking contrast to the seaward flank of the St Elias Mountains where extensive flights of terraces have been created by movements of the Fairweather fault in post-glacial times (Derksen, 1976; Plafker et al., 1994; Bruhn et al., 2004). Moreover, we find no definitive stratigraphic evidence in Icy Strait for RSL changes that

occurred during earthquakes. Such evidence would consist of peats overlain by intertidal silt and sand above abrupt and laterally extensive stratigraphic boundaries (Hamilton and Shennan, 2005) or coarse, clastic layers deposited in marshes by tsunamis (Nelson et al., 1996; Pinegina and Bourgeois, 2001).

On the other hand, it is difficult to disentangle glacial isostatic effects from the changes in RSL driven by earthquake-deformation cycles (Thatcher, 1984; Mann and Crowell, 1996; Nelson et al., 1996), even by meticulous analysis of salt-sensitive diatoms (Hamilton and Shennan, 2005; Shennan and Hamilton, 2006). In what follows, we assume that tectonism did not significantly influence RSL history in Icy Strait during the Holocene and, following the stabilization of global eustatic sea level ca. 8000 cal yr BP (Peltier and Fairbanks, 2006), that the main driver of RSL change in Icy Strait has been glacial isostasy. Nevertheless, we caution the reader that this assumption needs further testing.

Other possible factors in Holocene RSL change

Our reconstruction of RSL history relies heavily on the elevation of the remains of terrestrial vegetation, particularly trees. What if the vertical distribution of trees was controlled by the presence of back-barrier, freshwater swamps earlier in the Holocene? Rising RSL is probably the most common cause for the onshore and upslope movement of coarse, clastic barriers like the relict ones now found in bay heads in Icy Strait (e.g., Carter and Orford, 1988; Jennings et al., 1998; Orford et al., 2006; de la Vega-Leinert et al., in press); however, barrier beaches can also exhibit complex dynamics in the face of changes in sediment supply and wave regime (Carter and Orford, 1988; Orford et al., 1988). Could some of the barriers in Icy Strait have moved upslope in the past without being pushed there by rising RSL?

At least during the LIA, the large number of these raised barrier systems, each occupying its own bay head and separated by bedrock points from others, makes it doubtful that changing sediment supply caused by shifts in patterns of long-shore drift could have affected their dynamics to produce the degree of consistency in barrier age and elevation that we observe. It does seem likely that wave action was greater at the height of the LIA, particularly along shorelines facing northward into the mouth of Glacier Bay, but we currently lack the data to separate out such local wind effects from RSL changes caused by isostatic adjustments. As described in the next section, the correspondence between RSL and glacial history during the late Holocene tends to support the inference that glacial isostasy was the main controller of RSL change in Icy Strait.

Glacial history

The late Pleistocene

The oldest marine shells in Icy Strait constrain the timing of deglaciation at the end of the Pleistocene. Icy Strait was probably deglaciated during the interval 14,270–15,190 cal yr BP (Beta-62751), which is our oldest date on a marine shell

(Fig. 8). The Queen Charlotte Islands 600 km to the south (Blaise et al., 1990; Barrie and Conway, 1999; Fedje and Josenhans, 2000) and the Kodiak archipelago 1100 km to the west (Mann and Peteet, 1994) show a roughly similar timing of deglaciation. In contrast, the deglaciation of southern Puget Sound began several millennia earlier (Porter and Swanson, 1998; Clague and James, 2002).

Some of this difference in timing relates to position relative to the former ice margin. Icy Strait is located in a mid-fjord position, and the final deglaciation of the outer continental shelf of the Gulf of Alaska was probably underway ca. 13,400 ^{14}C yr BP (15,500–16,300 cal yr BP) (Mann and Peteet, 1994; Barrie and Conway, 1999). The timing and extent of late Pleistocene glacial ice on the continental shelf west of Icy Strait would have affected the suitability of the Pacific coastline as a route for human migration (Heusser, 1960; Fladmark, 1979), and we speculate that people could already have moved south along the outer coastline of Southeast Alaska prior to the 14,000–15,200 cal yr BP deglaciation of Icy Strait.

Two possible interpretations of the trajectory of RSL change in Icy Strait are possible between ca. 15,200 and 13,800 cal yr BP (Fig. 8). In the first, RSL may have fallen to the level (29 m) of the lower Falls Creek delta ca. 14,500 cal yr BP. After a stillstand that was long enough (perhaps several decades) to allow delta formation, RSL then rose again to 57 m, where it stabilized long enough to deposit the upper Falls Creek delta ca. 14,000 cal yr BP. It then fell rapidly to below modern RSL before 13,700 cal yr BP. This first trajectory suggests the occurrence of a glacial readvance that triggered renewed isostatic depression in northern Southeast Alaska between roughly 14,500 and 14,000 cal yr BP. Note, however, that this “readvance” trajectory is constrained only by the 14,130–15,150 cal yr BP age of T.D. Hamilton’s “marsh grass” date (SI-2082, data repository); if either the elevation, age, or our interpretation of the significance of this sample is in error, then this RSL trajectory should be rejected. To our knowledge, there is no other evidence for a glacial readvance in northern Southeast Alaska during this time interval.

The second possible RSL trajectory is simpler and more probable. It consists of RSL falling unidirectionally from 57 m to below the level of modern MLLW before 13,700 cal yr BP (Fig. 8). In this scenario, the lower Falls Creek delta formed after the higher one.

In either scenario, RSL fell very rapidly between the formation of the upper Falls Creek delta and the isolation of the Goose Island pond from marine waters. The longest possible duration for this RSL fall would have been 420 yr based on the 2 σ calibrated age limits of the constraining ^{14}C dates (Beta-222030 and Beta-186945). This maximum age span suggests that RSL was falling at a rate $>14\text{ cm yr}^{-1}$. In fact, the age span was very likely <420 yr, and the rate of RSL change was probably $>30\text{ cm yr}^{-1}$ during this period, which is an order of magnitude faster than the ongoing rate in Icy Strait today in the aftermath of glacier retreat from the LIA maximum. Comparably rapid rates of RSL occurred during the late Pleistocene along other parts of the North American Pacific coastline where

mantle viscosity was relatively low and where the rapid calving retreat of glaciers cut short the amount of restrained rebound that occurred under gradually thinning ice loads (Mathews et al., 1970; Dethier et al., 1995; Clague and James, 2002; Hutchinson et al., 2004b).

RSL history during the late Pleistocene and early Holocene in Icy Strait contrasts sharply with RSL history in the Queen Charlotte Islands 600 km to the south. There RSL fell for the first two millennia after deglaciation, then rose during the next four to five millennia due to the passage of the crustal forebulge that had developed during the last glacial maximum in response to prolonged loading under the thick ice sheet lying to the east (Fedje and Josenhans, 2000; Clague and James, 2002; Hetherington et al., 2004). The absence of any detectable rise in RSL in Icy Strait ca. 12,000–8000 ^{14}C yr BP (ca. 14,100–8600 cal yr BP) suggests that a migrating forebulge was not involved in the deglacial geodynamics of northern Southeast Alaska, at least not at mid-fjord locations. The RSL history in Icy Strait probably most resembles that of southern Vancouver Island (Clague and James, 2002), where land emergence culminated in the early Holocene with shorelines lying below today’s RSL. After ca. 10,000 cal yr BP, but before eustatic sea level stabilized ca. 8000 cal yr BP, residual glacio-isostatic rebound on both southern Vancouver Island and in Icy Strait may have been roughly compensated by eustatic sea level rise, with the result that RSL was relatively stable.

Of course, a major difference between the two regions is that in Icy Strait this conjectured tracking of eustatic sea level was interrupted after ca. 6000 cal yr BP by repeated isostatic adjustments to local glacier fluctuations. Geographic variability in RSL history along recently deglaciated coastlines like those of the northeast North Pacific underlines the complexity of RSL changes over relatively small distances that result from local differences in ice loading, glacial history, and earth rheology (Dethier et al., 1995; Hetherington et al., 2004; Shennan et al., 2006).

The Holocene

The Holocene portion of the Icy Strait RSL record provides information about glacial history, probably mostly that history in nearby Glacier Bay. RSL seems to have remained below the -4 m level from before 12,000 to ca. 3600 ^{14}C yr BP (ca. 14,000 to ca. 4500 cal yr BP) (Figs. 9, 10). Glacier expansion in the West Arm of Glacier Bay between ca. 5500 and 4200 ^{14}C yr BP (A. Post and G. Streveler, personal communication, 2006) may be recorded by a rise in RSL in Icy Strait between ca. 4500 and 5000 ^{14}C yr BP (ca. 4600–5000 cal yr BP). Neoglaciation began in northern Southeast Alaska ca. 3500 ^{14}C yr BP (ca. 3700 cal yr BP) (Mann and Hamilton, 1995; Calkin et al., 2001). Indeed the RSL record from Icy Strait suggests that the interval 1600 to 3500 ^{14}C yr BP (ca. 1400–3700 cal yr BP) was a time of heightened but fluctuating ice loading (Figs. 9, 10). Specifically, the RSL curve suggests three episodes of high RSL separating two episodes of low RSL during this interval.

The youngest of these high stands occurred ca. 1800 ^{14}C yr BP (ca. 1600 cal yr BP) and, assuming some uncertainty in the

limiting ages proposed by Goodwin (1988), this event may record isostatic depression under the glacier system that advanced down the West Arm to dam the mouth of Muir Inlet between 1900 and 2400 ^{14}C yr BP (ca. 1800–2500 cal yr BP) (Figs. 9, 10). The RSL low stand we infer for the interval 1500–1700 ^{14}C yr BP (ca. 1400–1700 cal yr BP) probably correlates with the glacier retreat that caused the penultimate drainage of Lake Muir (Goodwin, 1988). The glacier expansion that dammed Muir Inlet again sometime between 1700 and 900 ^{14}C yr BP probably caused the RSL rise that we infer to have occurred in Icy Strait between 1600 and 1300 ^{14}C yr BP (ca. 1100–1500 cal yr BP). The fall in RSL to around the -2 m level that is indicated by the log emplaced in the barrier beach at Dancing Bear Creek ca. 800 ^{14}C yr BP (ca. 660–900 cal yr BP) coincides with the retreat of West Arm ice from the mouth of Muir Inlet around this same time (Goodwin, 1988). On the other hand, a recent synthesis of glacial history in Glacier Bay by C. Connor, G.P. Streveler and A. Post (personal communication, 2007) disagrees with Goodwin's (1988) interpretation and sees no clear evidence for glacial retreat at this time. Either the log in the Dancing Bear Creek barrier was reworked from older deposits during the LIA rise in RSL, or we are missing the evidence for the corresponding glacier retreat in Glacier Bay.

After 800 ^{14}C yr BP (660–900 cal yr BP), RSL in Icy Strait rose to its all-time high during the Holocene, pushing gravel barrier beaches onshore as LIA ice advanced towards the mouth of Glacier Bay. The ca. AD 1800 age of trees colonizing the LIA marine limit both in Icy Strait and in the Juneau area (Motyka, 2003) is compatible with a glacier maximum reached ca. AD 1750, followed by glacier retreat that was initially slow enough to delay the onset of rapid isostatic rebound until after AD 1800.

The inferred glacial history of Glacier Bay that we obtain by combining Goodwin's (1988) work on the history of ice-dammed lakes in Muir Inlet with the RSL record from Icy Strait is broadly similar to that of other glaciers in the region (Fig. 10, lower panel). Evidence for glacier retreat is widespread along the North Pacific coastline between 1500 and 2000 ^{14}C yr BP (ca. 1400–2200 cal yr BP) at the time of a RSL low stand in Icy Strait. A widespread, pre-Medieval Warm Period advance of glaciers began ca. AD 200–300 all along the North American Pacific coastline (Reyes et al., 2007) and culminated between ca. AD 500–900 during a RSL high stand in Icy Strait (Fig. 10). Numerous glaciers in the region underwent retreat around AD 1000 at the beginning of the Medieval Warm Period (MWP), and the RSL low stand we infer in Icy Strait ca. AD 1100–1300 (ca. 700–900 ^{14}C yr BP) may record these MWP retreats.

Our RSL proxy for glacial history in Glacier Bay provides some interesting insights into glacier dynamics there. A striking feature of RSL history in Icy Strait since the mid-Holocene is the gradual, fluctuating rise in RSL that culminated ca. AD 1800 (Fig. 9). This RSL history is consistent with a glacial history that consisted mainly of a progressive advance of ice down Glacier Bay, building and recycling a massive terminal moraine shoal ahead of itself for ca. 6000 yr. This slow advance

of ice towards Icy Strait was interrupted by episodes of retreat that drove the terminus off this moraine shoal for unknown distances up-fjord.

Correlations with other glaciers in the region suggest that climate changes triggered these terminus fluctuations, which in turn altered isostatic depression and hence RSL. Nonetheless, the progressive down-fjord expansion of glacial ice in Glacier Bay over many millennia is consistent with the main driver of this glacier system being the dynamics of its terminus. The terminus dynamics of calving glaciers involve feedback interactions between the rate of terminal moraine construction, water depth at the grounding line, iceberg calving, glacier flow rate, and RSL (Meier and Post, 1987; Post and Motyka, 1995; Van der Veen, 1996). As in other calving glacier systems (Mann, 1986; Powell, 1991), the effects of climate change on the overall glacier system become slaved to the slower, larger-scale dynamics of the glacier's terminus, which may then modulate the system's overall responses to climate change. If the terminus is securely anchored on a terminal moraine shoal, climate change can cause only a brief and modest retreat or thinning of the glacier; however, if the terminus lies in an insecure position, as at a fjord mouth, even minor changes in climate can potentially trigger a fjord-emptying, catastrophic calving retreat. In the case of Glacier Bay, it was only after the glacier and its terminal moraine shoal entered Icy Strait ca. AD 1750 that the glacier became susceptible to catastrophic calving retreat, which in this case proved irreversible before the LIA ended ca. AD 1880.

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Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at [doi:10.1016/j.yqres.2007.12.005](https://doi.org/10.1016/j.yqres.2007.12.005).

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