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This is Version 3.7, the final version.

Notes added on June 9, 2021:
There is an error in the caption to the graph on page 15. The yellow dot on the left is for plant material, the older one on the right is for a log. It has been left uncorrected in this copy.

Additional radiocarbon dates have been obtained since this article was published and these are recorded in detail in Addendum 2021 to *SILT* 8-13 <https://nickdoe.ca/pdfs/Webp533.pdf>. The new information affects some of the detail in this Version 3.7, particularly on page 15, but not significantly enough to justify an extensive rewrite and republishing of the article. Version 3.7 therefore remains unchanged.

Gabriola's glacial drift—an icecap?

Nick Doe

Carlson, A. E. (2011) Ice Sheets and Sea Level in Earth's Past. Nature Education Knowledge 3(10):3

The deglaciation of Gabriola

One snowy winter's day, around the year 16000 BC (14.50 ka ^{14}C BP), give or take a century or two, the ice under which Gabriola was buried at that time, reached its maximum thickness.[1] It was then about one mile thick. From then on, the ice began to decay. There was no withdrawal of the glaciers then occupying the strait; they just stopped moving and wasted away, but the ice was so thick that it wasn't until more than three thousand years later—the exact date is not certain—that Gabriola finally became ice free and re-emerged into the sunlight.

The area of research I'm going to describe in this article is the timeline of the deglaciation on land, both at the then sea level and in the island's highlands, and over the sea. How fast did the ice melt? and are these rates commensurate with what we know about climate changes that were occurring at that time, particularly during the Younger Dryas? Did Gabriola have an ice cap? how thick was it and how long did it last? In what ways did the events of that time shape the landscape we see today?

The two approaches adopted are radiocarbon dating of sites that mark the transition from the late-Pleistocene to the early-Holocene (ice age to post-ice age); and computer modelling of the deglaciation.

A computer model needs to take into account changes in climate; changes in sea level; melting rates under till-covered ice; undermelting of the ice shelf in the Salish Sea as oceanic water returned; calving of ice cliffs along the shoreline and other lateral movements; the effects of rain and snow; and micro-differences in weather at sea level and in the highlands. Even very small annual differences in environmental factors mount up over thousands of years.

The 'broad" picture

Toward the end of the last ice age, sea level in the Salish Sea was considerably higher than it is now.[2] The reason for this is not that sea level was high *per se*—eustatic (global) sea level was in fact lower than average—but it was because the height of

[1] The Vashon Stade of the Fraser (Late Wisconsin) Glaciation.

[2] Steve Earle, *The ups and downs of Gabriola*, *SHALE* 5, December 2002. pp.14–20.

the continental land relative to the floor of the Pacific Ocean had been reduced by the great weight of the Cordilleran ice sheet.

We know that local sea level remained high until some time after the waters around Gabriola had become open. There are fossil sea-shells from this time on the island that are about 100 m above the present sea level (AMSL).[3] Bones of a whale that date back to the end of the ice age have also been found here on a site about 45 m AMSL.

One of the puzzling characteristics of these former marine sites on Gabriola is that most, if not all, are buried by copious amounts of silty sand that is free of stones. These sediments are usually, but not always, topped by a thin, sparse layer of stony ablation till, or its lag-gravel remnants.[4]

Thick deposits of stone-free sand must have been accumulated by large volumes of flowing water—only flowing water can efficiently sort sand from both gravel and mud. I can only imagine that this flowing water was meltwater. There are no rivers on Gabriola, nor is there any evidence that Gabriola's climate throughout the Holocene has ever been considerably wetter than it is now.

Those sites that show no evidence of a marine layer usually have either a thick layer of bluish-grey silty-clay above the bedrock, or a layer of gleysol, which likely is a recent weathering product of the same silty-clay.[5] Such layers always contain dropstones, indicating that they are likely undermelt till from a grounded ice sheet, or were rafted on to a glacial lake by icebergs.

The conclusion must therefore be that when the sea became open, and local sea level had fallen, there were still substantial amounts of ice over the land. It was the melting of this land-bound ice that carried the sand on to the beaches and into the shallow bays left behind as the sea retreated.

This theory, for that is what it was when I started my investigations, raises two interesting points.

The first is, when did the "great melt" of the land-bound ice occur? Was it before, during, or at the end of, the 1300-year long period known as the Younger Dryas? [6]

The second is, where did all the sand and silt come from? There were at least two possibilities:

- the sand was just englacial or basal till derived from bedrock scoured by the glaciers. This bedrock would be the island's sandstone, but would include off-island bedrock further up-stream of the ice flows
- the sediments were derived from well-sorted, pre-glacial, unconsolidated sediment, primarily *Quadra Sand*.[7]

[3] AMSL = "above mean sea level" as used on modern topographical maps. The mean tide range in this area is roughly ±2.5 metres.

[4] http://www.nickdoe.ca/pdfs/Webp533.pdf

[5] Doe, N.A., *The geology of Gabriola's diatomaceous earth*, *SHALE* 24, pp.31–36, June 2010.

[6] The Younger Dryas event was a cold reversal of the Bølling-Allerød warming that followed the ending of the Fraser Glaciation. The changes from a warm climate back to an almost glacial one at the start of the Younger Dryas, and from a glacial one back to a warm one at the end of the Younger Dryas are remarkable for their abruptness. The duration of these dramatic climatic changes have been resolved to just a few decades. The phenomenon is often ascribed to short-term changes in the pattern of global ocean currents caused by huge volumes of cold freshwater that were pouring off the continents at the time.

[7] Clague, J.J., *Quadra Sand: A study of the late Pleistocene geology and geomorphic history of coastal southwest British Columbia*, Geological Survey of Canada, Paper 77-17, 1977.

The "is the sand, Quadra Sand?" issue, I will cover in another article (GD-534).

Time scales

The first thing that we have to be clear about in investigations of this sort is the units used in time scales. They can be *calendar dates* or *radiocarbon ages*.[8] I'm going to use only radiocarbon ages ^{14}C BP—meaning conventional radiocarbon ages—and calendar dates AD/BC.[9]

Life would of course be simpler if researchers only used calendar dates, but they don't with good reason. The only practical way of determining the age of ice-age material is to use radiocarbon dating. Radiocarbon dating is such a useful method that many researchers prefer to use radiocarbon ages exclusively. However, we can't do that here because we want to make calculations involving annual quantities, the amount of ice lost due to melting for example, and these require calendar years.[10]

There is of course a good database for converting conventional radiocarbon ages to calendar dates; however, it is unfortunate that the most common late-ice-age material found on the island is shell. Dating shells is problematic because the calculations involve knowing how old the carbon was at the time it was incorporated into the shell—the so-called marine reservoir constant. The results of numerous observations of this constant around the Salish Sea at the end of the ice age indicate that, unlike during more recent times, the "constant" was far from constant, and until more study has is done, its precise value for any given year remains uncertain.

[*Readers interested in the technical details of radiocarbon dating on the Pacific Coast may want to read Appendix 2 of Gabriola's glacial drift —ice-age fossil sites on Gabriola, File:* GD-533].

Quantizing the time period

For the analysis, I divided the deglaciation period from 15000 BC to 8750 BC (13.90–9.45 ka ^{14}C BP) into ten-year increments.

Climate changes

The timing of some significant climate changes during deglaciation that I will be using is as follows:[11]

Start of deglaciation (*for present purposes)*:
13900 ^{14}C BP = 1 5000 BC

End of the Oldest Dryas, start of the Bølling Interstade in Northern Europe
12485 ^{14}C BP = 12 720 BC

End of the Bølling Interstade, start of the Older Dryas (?) or start of the Allerød Interstade:
12160 ^{14}C BP = 12 050 BC

End of the Older Dryas (?) or continuation of the Allerød Interstade:[12]
11900 ^{14}C BP = 11 800 BC

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8 I am not going to use, as some researchers do, calendar or calibrated ages (cal. BP). Some define these as being calendar years before 1950 AD when the radiocarbon dating method was first developed; and some define it as years before 2000 AD. Not all researchers bother to say. Calibrated ages also depend on the database used for the calibration, and again, not all researchers explain their methodology. It is in any case, unnecessarily confusing to be using both years cal. BP and years ^{14}C BP in the same discussion.

9 The International Standard for conventional radiocarbon dates uses: a half-life of 5568 years (the Libby standard); Oxalic Acid I or II as the modern radiocarbon standard; correction for sample isotopic fractionation (δ^{13}C) to -25 ‰ relative to the ratio of ^{13}C/^{12}C in the carbonate standard VPDB (Vienna Peedee Belemnite); and 1950 AD as 0 BP.

10 Calendar ages are roughly, and on average only, 22% greater than conventional radiocarbon ages.

11 Fraser Valley dates are less well-defined and less consistent but are broadly similar: Vashon Stade to 14750 BC; Everson Interstade to 11400 BC; and Sumas Stade to 9600 BC.

End of the Allerød Interstade, start of the Younger Dryas:
11000 ^{14}C BP = 10 900 BC

End of the Younger Dryas, start of the early-Holocene:
10050 ^{14}C BP = 9600 BC

Day-to-day weather reports are, of course, not available for these times, especially not for Gabriola, so I used the following basic numbers. I have no doubt that these can be challenged, but I suspect changes would have relatively little impact on the final result.[13] We know, for example, approximately when the Strait of Georgia became ice free, and whatever climate data is used, it has to result in the predicted date being the known date of that particular event and which particular climate detail has to be changed to accomplish this is interesting, but not of crucial importance.

period	mean annual temp. ºC	max monthly mean temp.ºC	annual precipitation equivalent mm
Oldest Dryas	0.0	5.0	1000
Bølling Interstade	10.0	17.0	1100
Older Dryas	6.0	14.5	850
Allerød Interstade	10.0	17.0	1100
Younger Dryas	2.0	12.0	600
Holocene	13.0	20.0	1100

[12] Based on a 250 year duration. The Older Dryas was a brief period whose dating is not well established, and which is not present in all records.

[13] Because Gabriola's aquifers are in fractured rock, it seems likely that these were frozen up during the early stages of deglaciation and surface run-off would have been much higher than it is today.

Calculating how much ice melts as a function of air temperature

How fast glacial ice melts depends, obviously, on air temperature, but unfortunately, knowing the average annual temperature or average summer temperature, which is about all we can hope for in the late-Pleistocene, isn't enough to determine how fast the ice melted.

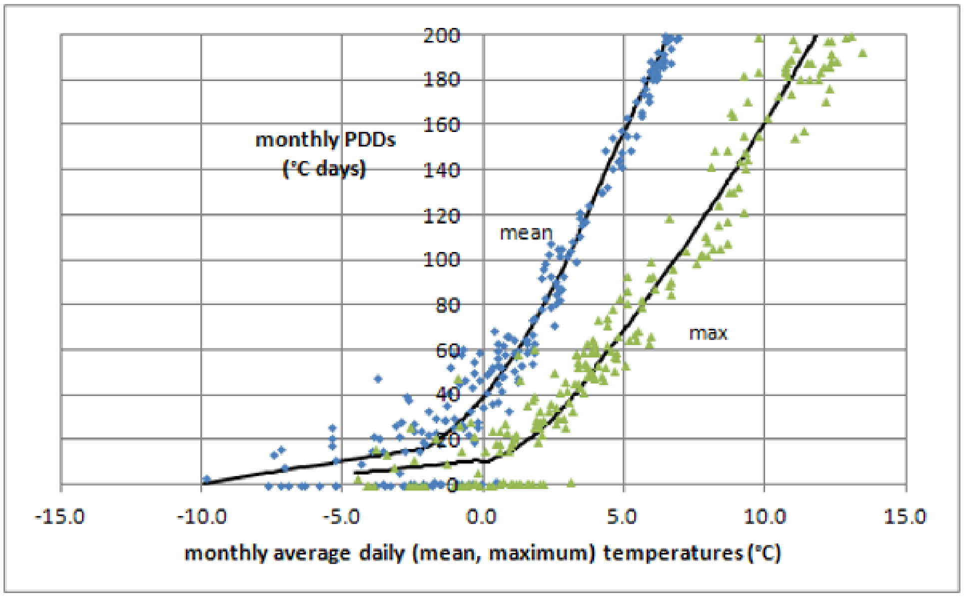
Just because the average annual temperature was, say, –4°C in a given timeframe doesn't mean there was no melting. In some circumstances where the variation about the average was low, say ±4°C, there really wouldn't have been any melting. If however the variation about the mean had been high, say ±10°C, then with a seasonal maximum of +6°C, there definitely would have been.

While knowing the average summer temperature, late-July and early-August, does at least determine whether or not there was any melting at all, it leaves us guessing as to how much there was on an annual basis.

The method that I have found best[14] for estimating melt rates (mm/year) is to assume that the melt is proportional to the number of positive-degree days (PDD) in the year where PDD is the integral of air temperature as a function of time throughout the year whenever the temperature is above freezing.[15 16] So, for example, PDD =

[14] During a study of the melting of the Whistler Glacier as a result of climate change. This was a course project (GEOL 305, *Quaternary Geology*) in the fall of 2012 at Vancouver Island University led by Professor Steve Earle.

[15] Braithwaite, R.J. & Olesen, O.B., *Calculation of glacier ablation from air temperature, West Greenland*, in Oerlemans, J. (ed.), *Glacier*



Monthly positive-degree days (PDD) as a function of monthly average daily mean and maximum temperatures at Whistler airport 1977–2010.

365°C.days is equivalent to the temperature remaining at +1°C day and night for a year. The PDD method is empirical, but appears to be theoretically sound.[17]

Values for the proportionality constant linking PDD to melting rates that have been used (Reeh, 1991, p.119) are:
3 mm (SWE) per PDD for snow[18]
7 mm per PDD for ice.

To these I added:
5 mm (SWE) per PDD for névé and firn.

PDDs vs. temperatures

To get some idea of the relationship between annual PDD and temperature, I analyzed daily weather data for Whistler airport (WAE) from Environment Canada from 1977 to 2010. The airport is 658 m AMSL. The procedure was to assume that, given maximum (T_{max}) and minimum (T_{min}) daily temperatures, the variation throughout the day was:

$T_{var} = T_m + T_d \cos(2\pi t')$;

where

$T_m = 0.5(T_{max} + T_{min})$;

$T_d = 0.5(T_{max} - T_{min})$;[19] and

the angle t' is normalized time such that $t' = -½$ at midnight at the start of the day, and $+½$ at midnight at the end of the day.

The daily PDD_{day} was then:

$PDD_{day} = \int T_{var}\, dt'$;

for $-½ \le t' \le ½$; $T_{var} \ge 0$ else 0.

This has a closed-form solution;

<u>if</u> $T_{max} \le 0$ <u>then</u> $PDD_{day} = 0$;

fluctuations and climatic change, Kluwer Academic Publishers, pp.219-233, 1989.

16 Niels Reeh, *Parameterization of melt rate and surface temperature on the Greenland ice sheet*, Polarforschung 59/3: pp.113-128, 1991.

17 Atsumu Ohmura, *Physical basis for the temperature-based melt-index method*, Journal of Applied Meteorology, 40, pp.753- 761, April 2001.

18 SWE = snow water equivalent; what you get when you melt it. SWE depends on the snow density which is usually in the range of 5-20% with 10% being an average value. Snow has a higher albedo than ice and and so melts more slowly.

19 This definition implies $T_{min} = 2T_m - T_{max}$ so only T_m and T_d or T_{max} are independent variables.

else if $T_{min} \geq 0$ then $PDD_{day} = T_m$;

else if $0 \leq T_m/T_d \leq 1$ then $PDD_{day} = T_m - (T_m/\pi)\text{acos}(T_m/T_d) + (T_d/\pi)\text{sqrt}(1-(T_m/T_d)^2)$;

else $PDD_{day} = (T_m/\pi)\text{acos}(-T_m/T_d) + (T_d/\pi)\text{sqrt}(1-(T_m/T_d)^2)$.

The monthly PDD is then:
$PDD_{month} = s\ \Sigma\ PDD_{day}$
$s = 365.25/(12 \times \text{days in the month})$
and the monthly average daily mean (T_{mm}) and the monthly average daily maximum (T_{max}) are:
$T_{mm} = \Sigma T_m$ / days in the month
$T_{mx} = \Sigma T_{max}$/ days in the month

["months" throughout this article are equal periods of 30.44 days]

the trend lines for PDD_{month} *vs*. T_{mm} and T_{mx} were found to be:

if $T_{mm} < -10°C$ then $PDD_{month} = 0$;

else if $T_{mm} < -2.1°C$ then $PDD_{month} = 20.4 + 1.97T_{mm}$;

else if $T_{mm} < 4.4°C$ then $PDD_{month} = 39 + 14.9T_{mm} + 1.9T_{mm}^2$;

else if $T_{mm} < 20°C$ then $PDD_{month} = 29.3 + 24.2T_{mm} + 0.235T_{mm}^2$;

else $PDD_{month} = 30.44T_{mm}$.

if $T_{mx} < -8.2°C$ then $PDD_{month} = 0$;

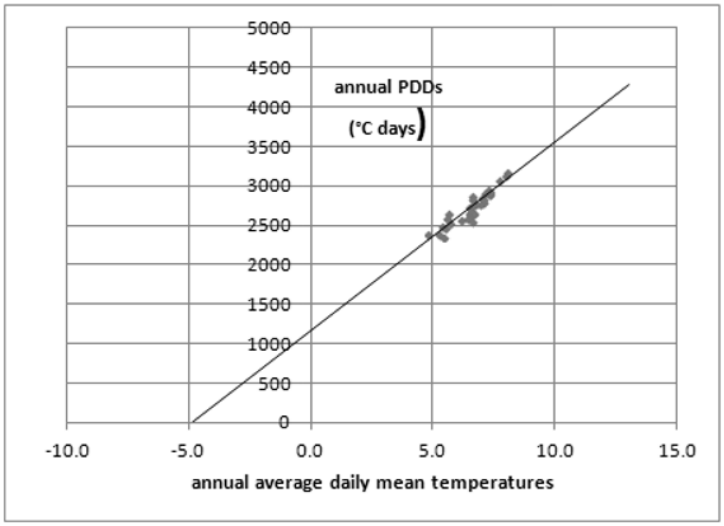
else if $T_{mx} < 0.4°C$ then $PDD_{month} = 11 + 1.33T_{mx}$;

else if $T_{mx} < 4.4°C$ then $PDD_{month} = 10 + 3.4T_{mx} + 1.8T_{mx}^2$;

else if $T_{mm} < 20.2°C$ then $PDD_{month} = -3.8 + 12.6T_{mx} + 0.39T_{xm}^2$;

else $PDD_{month} = 20.27T_{mx}$.

Having reduced the data to spreadsheet form, the next step was to find the relationship between annual PDD (PDD_{year}) and mean annual monthly mean temperatures.



Annual positive-degree days (PDD) as a function of annual average daily mean temperatures at Whistler airport 1977–2010.

$PDD_{year} = \Sigma\ PDD_{month}$

$T_{year} = \Sigma\ T_{mm}/12$.

Here we run into a problem. Although there is a fair amount of variation of T_{mm} from month to month, as a result of the seasons, there is relatively little variation in T_{year} from year to year. The weather changes but the climate does not. Hence, though we can use the T_{mm} to calculate a trend line as we did above, using the T_{year} data gives a trend line that may be applicable only over a very limited range of values. For the record it is: if $T_{year} < -4.9°C$ then $PDD_{year} = 0$ else $PDD_{year} = 1171 + 239T_{year}$. The high temperature limit is $PDD_{year} = 365.25T_{year}$ when T_{year} is more than about 9.3°C.

One solution to this problem is to analyze data from locations at a different latitude or elevation—a lot of work—but the preferred alternative was more analytical. The advantage of this second method is that predictions can be made for climatic conditions that no longer exist. The downside however is that some assumptions have to be made that may not be valid.

The first step was to produce, from the Whistler data, an empirical relationship between annual mean temperatures and monthly mean temperatures. This was found to be:

$T_{mm}(m_n) = m_{m0}+m_{m1}\cos(2\pi(m_n+\Phi_m))+m_{m2}\cos(4\pi(m_n+\Phi_{2m}))$

where:

$T_{mm}(m_n)$ is the monthly average daily mean temperature for month m_n:

m_n identifies the particular "month" (actually equal periods of 30.44 days) and is 0.5/12 for January, 1.5/12 for February, 2.5/12 for March…and 11.5/12 for December; and

$m_{m0} = 6.5°C$; $m_{m1} = 9.6°C$; $m_{m2} = 1.4°C$; $\Phi_m = –0.546$; $\Phi_{2m} = –0.149$;

We can simplify a bit. The phase factor Φ_m is only there to synchronize the month-to-month temperature variations with the actual seasons as defined by m_n, and so can be regarded as a constant. The factor m_{m2} is there because the warming of the atmosphere in spring is slower than the cooling of the atmosphere in the fall; so let's assume that the ratio m_{m2}/m_{m1} is a constant. The associated seasonal phase shift factors Φ_{2m}, can also be taken as a constant. The factor m_{m0} is synonymous with T_{year}.

This leaves us with:

$T_{mm}(m_n) = T_{year}+ m_{m1}[\cos(2\pi(m_n –0.546)) + 0.15\cos(4\pi(m_n – 0.149))]$

The factor m_{m1} reflects the seasonality of the monthly temperature throughout the year and is a useful factor to be able to vary as the difference between summer and winter was greater during the late-Pleistocene than it is now.

The definition I will use therefore is:

$m_{m1} = (T_{ja} – T_{year})/1.03$

where T_{ja} is the average daily mean temperature throughout either July or August, whichever is highest; they are usually about the same, the peak being around the end of July, beginning of August.

For the year 1997 (a random choice), the observed PDD was 2769°C.days, the annual average daily mean temperature was 6.9°C, and the highest monthly average daily mean temperature (August) was 17.2°C. The analytical method gave a PDD of 2747°C.days, which for this particular year is only -0.8% below the observed value.

An interesting exercise is to see what happens if we maintain the summer temperatures but lower the annual average daily mean temperature by 7°C to –0.1°C; the winter is much harsher.

The PDD drops to 1938°C.days. This is an expected modest drop given that most melting occurs in a short period at the height of summer.

Dropping also the highest monthly average daily mean temperature (July) by 7°C to 10.2°C has the expected greater effect. The PDD drops to 1208°C.days, which is still sufficient to melt several metres of bare ice and snow.

Calculating the surface temperature of the ice

The temperature of the surface of an ice sheet can be below the "average" temperature for the region for two reasons. One is that pooling of cold air above the ice; and the other is as a consequence of altitude.

In a study of the Whistler glacier (elev.1930 m),[20] which is in a bowl, it was found that predictions of the ice loss could only be reconciled with observations of the actual loss if the temperature used in the calculations was taken to be 1.5–3.5°C lower than that deduced from meteorological data recorded at the near-by Whistler Roundhouse (elev.1835 m). That temperatures are lower in a cwm (cirque, corrie) is a common observation.[21] One study of the phenomenon mentions temperature decreases in the bowl relative to the rim of –0.7 to –1°C/100m during the day and "during the night, a thermal belt on the valley floor 3°C colder than the top of the inversion layer".[22]

This is something to consider, but only in the very final stages of deglaciation.

Lapse rates (the decrease in temperature with elevation) present several modelling problems. Studies have shown that there is no constant lapse rate in mountainous areas [23] or above ice sheets.

Every hiker and skier knows that it is not uncommon for the top of a mountain to be warmer than in the valley. A comparison of meteorological data taken over 30 years at Whistler Airport (658 m) and Whistler Roundhouse (elev.1835 m) shows that not only does the lapse rate vary from month to month, the lapse rates for maximum and minimum temperatures are different.

The highlands of Gabriola are around peak at around 160 m AMSL, so summer maximum temperatures would be 1.5°C lower up there once sea level had returned to the current level, and assuming the Whistler data is applicable to the coast.

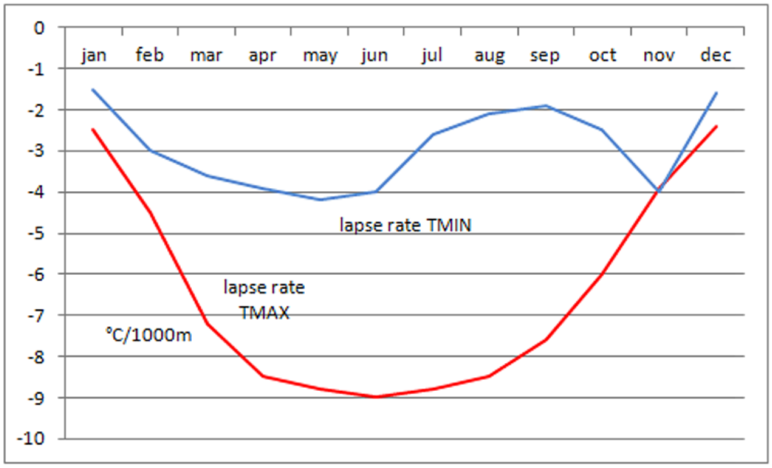
While the complexity of the lapse rate data might be a concern, I chose to ignore the complexity because Gabriola is not very high, and lower maximum temperatures are more important than lower minimum temperatures, and the lapse rate of the maximum temperatures is easier to model.

[20] This was a course project (GEOL 305, *Quaternary Geology*) in the fall of 2012 at Vancouver Island University led by Professor Steve Earle.

[21] There is on Gabriola a former lake, now filled with sediment and used for market gardening (the Commons) and it sits in a cwm-like depression. A gardener there tells me it is not uncommon for the morning temperature in the fields to be 2°C below the temperature at the building that sits on the upper rim overlooking the area.

[22] Carla Mora, *Spatial and seasonal characteristics of cold-air pools in the upper Zêzere valley (Portugal),* Geophysical Research Abstracts, 12, EGU2010-4267, 2010.

[23] Stahl, K., Moore, R.D., Floyer, J.A., Asplin M.G., & McKendry, I.G., *Comparison of approaches for spatial interpolation of daily air temperatures in a large region with complex topography and highly variable station density*, Agriculture and Forest Meteorology 139, pp.224-236, 2006.



Lapse rates for monthly average daily maximum and minimum temperatures based on a comparison of temperatures at Whistler Airport and the Whistler Roundhouse at the top of the mountain. The "standard" fixed lapse rate of –8 °C/1000m is evidently strongly modified in winter by inversions.

My proposal for dealing with lapse rates is therefore for every 100 m difference in height, subtract the following from the monthly average daily mean:

$T_{mm}' = T_{mm} - L_{month}$

where L_{month} is
0.20 (Jan.); 0.38 (Feb.); 0.54 (Mar.); 0.62 (Apr.); 0.65 (May); 0.65 (Jun.); 0.57 (Jul.); 0.53 (Aug.); 0.48 (Sep.); 0.43 (Oct.); 0.40 (Nov.); 0.20 (Dec.) ºC/100m.

Does this make a difference? Here's an example with previously calculated numbers recalculated for a site 50m higher. In this example, PDD drops from 2747°C.days to 2676°C.days. A modest drop, but sufficient for there to be about 0.5 metres less ablation of ice per year in the highland.

Calculating the effect of debris on ablation rate

One of the unusual aspects of the deglaciation of the Strait of Georgia is that the glaciers occupying the strait did not retreat at the end of the last ice age; they simply stopped flowing and wasted away. This process must have led to an accumulation of debris on the surface of the ice as ablation progressed, and this debris would have substantially altered the melting rate by both lowering the albedo of the surface, and when sufficiently thick, acting as a thermal blanket.[24]

A useful summary of observations of these effects is in a paper by Nakawo & Rana.[25] The magnitude of the thermal blanketing effect has also been tested in the laboratory [26] and modelled.[27]

An empirical formula that roughly replicates their data is:

ablation (cm/day)
$= 6[\exp(-0.06d)+0.13d.\exp(-0.1(d - 0.07)^2)]$
where d is the thickness in cm of the debris

To use this, we:
1. replace the factor 6 (ablation for t=0) with PDD. As explained above, I am going

[24] There is a good account of how stagnant ice that is buried by accumulated glacial drift can persist for hundreds, perhaps thousands, of years despite a warm climate in E.C. Pielou, *After the ice age—the return of life to glaciated North America*, Chicago Press, pp.175–184, 1991.

[25] M. Nakawo & B. Rana, *Estimate of ablation rate of glacier ice under a supraglacial debris layer*, Geografiska Annaler, 81 A4, pp.695–701, 1999.

[26] N. Reznichenko et al., *Effects of debris on ice-surface melting rates—an experimental study*, Journal of Glaciology, 56 (197), pp.384–394, 2010.

[27] L. Nicholson & D. Benn, *Calculating ice melt beneath a debris layer using meteorological data*, Journal of Glaciology, 52 (178), pp.463–470, 2006.

to assume that ablation is directly proportional to positive-degree days (PDD);

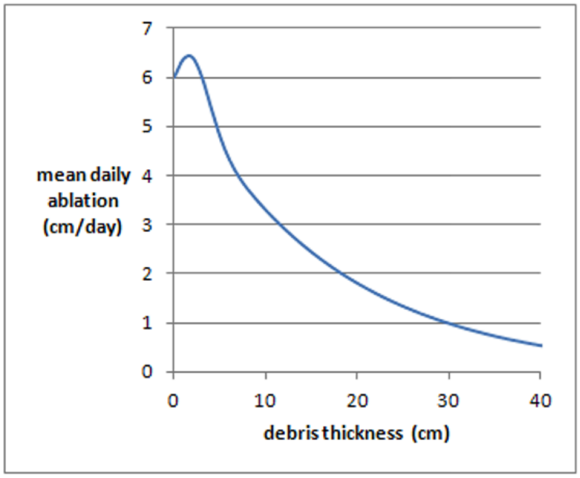
2. change the debris thickness units to metres;

Hence:
PDD′ = PDD [exp(-6D)+13.3D.exp(-0.1(100D–0.07)^2)]

where:
PDD′ is the PDD count for the month or year after correction for a debris layer D metres thick;
PDD is the PDD count for the month or year.

1997	based on formula		PDD with debris (thickness)	
	T_{mm}	PDD	0.1m	0.4 m
Jan.	–2.8	15	8	1
Feb.	–0.5	32	18	3
Mar.	2.8	95	52	9
Apr.	6.3	191	105	17
May	10.2	301	165	27
Jun.	14.3	424	232	38
Jul.	17.2	516	283	47
Aug	17.1	512	281	46
Sept.	13.2	391	214	35
Oct.	6.8	206	113	19
Nov.	0.7	50	28	5
Dec.	–2.7	15	8	1
year	6.9	2747	1508	249

If this analysis does represent reality, despite the numerous simplifications, the presence of accumulating ablation till would have very significantly slowed the melting, so much so that the total amount of supraglacial till and englacial till distributed throughout the ice sheet must have been less than a metre thick. This conforms to the



Observed change in melt rate as a function of the thickness of the debris on the surface of a glacier in Nepal. (Nakawo & Rana, 1999). Note that although debris may increase the melting rate by lowering the albedo of the surface, the effect is confined to only very thin layers. The dominant effect is the provision of insulation by the debris.

observations in the field that frequently show that loosely-consolidated ablation till often contains what appears to be a significant basal-till component in the form of stones with facets and sometimes facets with striae. Debris that is basal till would not have contributed to the slowing of the melting rate until the very end of the deglaciation.

In places on the ice when the debris cover was over a metre thick, ablation effectively ceased. In the real world, this must have created a very uneven ice surface, and debris would eventually have tumbled down ice cliffs into "valleys" on the surface where the debris was thinner. This process would have evened out the horizontal distribution of the surface debris.

One way of modelling a non-uniform vertical distribution of till in the ice is to imagine the concentration of basal and supraglacial till varies according to some relationship such as:

$C(d) = C_b(d) + C_s(d)$

where:
$C(d)$ is the total concentration of till (metres per metre) at depth d (metres);
$C_b(d)$ is the concentration of basal till at depth d (metres);
$C_s(d)$ is the concentration of supraglacial till at depth d (metres);
$C_b(d) = C_{maxb} \exp(-d/D_{0b})$;
$C_s(d) = C_{maxs} \exp(-(D-d)/D_{0s})$;
D is the total initial depth of the ice (metres);
d=D at the surface, d=0 at the bottom;
D_{0b} is a constant reflecting the thickness of the basal till layer (metres). 63% of the basal till is less than this constant distance from the bottom of the ice sheet;
D_{0s} is a constant reflecting the thickness of the supraglacial till layer (metres). 63% of the supraglacial till is less than this constant distance from the surface of the ice sheet;
$C_{maxb} = B_b/(D_{0b}(1 - \exp(-D/D_{0b})))$;
$C_{maxs} = B_s/(D_{0s}(1 - \exp(-D/D_{0s})))$;
where:
B_b is the thickness of the basal till after all the ice has gone (metres); and
B_s is the thickness of the supraglacial till after all the ice has gone (metres).

loss of ice due to ablation = L_A
loss of ice due to undermelting = L_U

basal till accumulated at the surface =
$C_{maxb}D_{0b}\exp(-D/D_{0b})(\exp(L_A/D_{0b}) - 1)$

basal till lost due to undermelting =
$C_{maxb}D_{0b}(1 - \exp(-L_U/D_{0b}))$

supraglacial till accumulated at the surface =
$C_{maxs}D_{0s}(1 - \exp(-L_A/D_{0s}))$

supraglacial till lost due to undermelting =
$C_{maxs}D_{0s}\exp(-D/D_{0s})(\exp(L_U/D_{0s}) - 1)$

Information on the distribution of till within an ice column is not available. If we assume that the till is uniformly distributed, it is hard to match the results of the simulation with reality. It also carries the unlikely supposition that the total amount of the till in an ice column is greater over the sea than it is over land because, assuming an initially level surface, the ice column is thicker over the sea than it is over the land.

For all of the simulations, I assumed that the total amount of till in each ice column was the same, and was independent of the underlying bedrock topology. One of the consequences of this assumption is that the till density over the sea is slightly less than the till density over land, and the till density over the highlands is slightly more than the till density over land at sea level. All the while the ice column over the sea is not being undercut, the ice over the sea melts very slightly faster than over the land, and the ice over the highlands melts slightly slower than the ice over land at sea level. The ice surface thereby gradually moves from being completely flat to having a topology similar to the bedrock. This is arguably reasonable, but only as long as we neglect lateral movement of the ice.

Calculating the effect of precipitation on the ablation rate

Rain and snow have multiple effects on the melting of an ice sheet.

Rain may either run off; infiltrate the snow and refreeze; or make its way into the ice and refreeze but also warm it. It may also fill gaps in overlying low-permeability rock-avalanche material, as discussed above, thereby improving the effectiveness of the debris as a thermal blanket and retarding melting, but it can also reduce the albedo of the surface debris thereby increasing warming and melting.

To say the rain "runs off" of course assumes that it has somewhere to run to. One can imagine that on a thick, fairly-flat static ice sheet, the rain might just pool on the surface rather than make its way to the bedrock through crevasses and moulins.

Without further analysis, which would require far more work than I'm offering here, I just assumed that rain has no net effect. It is true that it can freeze on the ice thereby adding to the thickness of the ice, but it also warms the ice as it does so, and this, one could argue (qualitatively), compensates for the fact that heat is later required to remelt it.

Snow at the surface initially melts and infiltrates the snow beneath it. The melting rate is lower than for ice because of the high albedo. Below the surface, it refreezes and densifies the snow creating névé and then firn (I make no distinction for present purposes). When the thickness of the névé exceeds a certain fraction of the snowpack, the melting snow ceases to refreeze and instead runs off. [28] A further complication is that snowpack acts as a thermal blanket and suppresses the diurnal temperature variations of the debris on the ice surface. This reduces the melting of the ice beyond what it would have been without the snow cover.[29]

For a spreadsheet analysis, we need to know the monthly precipitation, and how much of this is snow as a function of average monthly daily mean temperature (T_{mm}). Simplifying yet again, I reckoned:

if T_{mm}>2ºC, no snow; else if T_{mm}<0ºC all snow; else proportion of snow = (2–T_{mm})/2.

Of the snow, 100% of the SWE (snow water equivalent) requires melting.

The method used was subtract from the available PDD at the snow-albedo rate. If the PDD for the month remains positive after the subtraction, we don't need to increase the thickness of the ice by the thickness of the névé because melting the névé has already been accounted for (the névé requires heat to melt it, but it has already received heat from the snow meltwater percolating into it).

If there is insufficient PDD to melt the snow, then the PDD for the month become negative. In calculating this I assumed that whatever PDD was available melted the snow, and the remainder of the snow became névé. The negative value of PDD then reflected the fact that névé is easier to melt than snow, but harder to melt than pure ice.

The proportion of the annual precipitation falling each (equal-day) month at Whistler from 1977 to 2010 was found to be: 0.130 (Jan.); 0.091 (Feb.); 0.081 (Mar.); 0.061 (Apr.); 0.055 (May); 0.048 (Jun.); 0.033 (Jul.); 0.038 (Aug.); 0.056 (Sep.); 0.118 (Oct.); 0.159 (Nov.); 0.130 (Dec.).

The average annual precipitation is 1240 mm; at Nanaimo in the 20th century, it was 1120 mm.

Undermelting (basal melting) of sea ice

Undermelting of the ice shelf would have played a crucial role in the opening up of the Strait of Georgia during deglaciation before the adjacent land was ice free. Unfortunately, although there are many

[28] Braithwaite R.J. & Thomsen H.H, *Run-off conditions at Paakitsup Akuliarusersua, Jakobshavn, estimated by modelling*, Groll. Geol. Ulldersogelse Gletschcr-Hydrol. Medd. 3/1984.

[29] Complexities such as these are discussed in L. Nicholson L. & Benn D.I., *Properties of natural supraglacial debris in relation to modelling sub-debris ice ablation*, Earth Surface Processes and Landforms, Wiley, 2012.

papers on the topic,[30] "calculating" the rate of ablation by undermelting is not possible without knowing the temperature of the sea beneath the ice, and this is as unknown as is the melting rate itself.

Many studies of Antarctic ice shelves are of limited application here because ocean currents down there are different from what they would have been in the strait. Tidal currents here are strong, water exchange with the open Pacific Ocean is limited, and the strait is relatively shallow.[31] About all one can say is that the temperature was likely close to freezing (–1.8ºC to +2ºC), and the rate was probably somewhere in the 0.1–10 m/year range although values as high as 20 m/year (6 cm/day) have been observed around Antarctica. In early trials of the model it was found that undermelt rates had to be greater than 4 m/year to avoid predictions that the shallow parts of the strait become ice free before the deeper central parts, which is probably not a realistic scenario.

A critical point in the deglaciation of the Strait of Georgia would have been when the base of the glacier became free of the floor of the strait. Theoretically, the ice would have lifted off when the level of the Pacific Ocean above the floor of the strait was equal to 0.89 times the thickness of the ice in the strait.[32]

[30] Holland, P.R. & Jenkins A., *The Response of Ice Shelf Basal Melting to Variations in Ocean Temperature*, Journal of Climate, 21, pp.2558–2572, 2008.

[31] Mueller, RD, Padman L, Dinniman MS, Erofeeva SY, Fricker HA, King MA, *Impact of tide-topography interactions on basal melting of Larsen C Ice Shelf, Antarctica*, Journal of Geophysical Research-Oceans, 117, 2012.

[32] Less than the thickness because floating ice in the sea rises above the surface.

The possible consequences of undermelting are shown in the accompanying three diagrams.

The first curve shows melting at a constant rate starting at around 15000 BC. The ice is gone shortly after 13000 BC. The sea and the land become ice-free at the same time.

The middle curve shows melting at the same constant rate starting at around 15000 BC, but when ice lift-off occurs at the mean depth of the strait around 13000 BC, undermelting of the sea ice begins. It is also assumed that there is a uniform distribution of till in the ice with a total depth of 0.1 m.[33] The rate of undermelting shown here (1.2 m/year) has been chosen so that the sea becomes ice free at the same time it did in the first scenario. Storms (depressions) can also hasten the break up of sea-ice by wave action and heightened tides.

[33] Not including lodgement till which does not affect the melting rate.

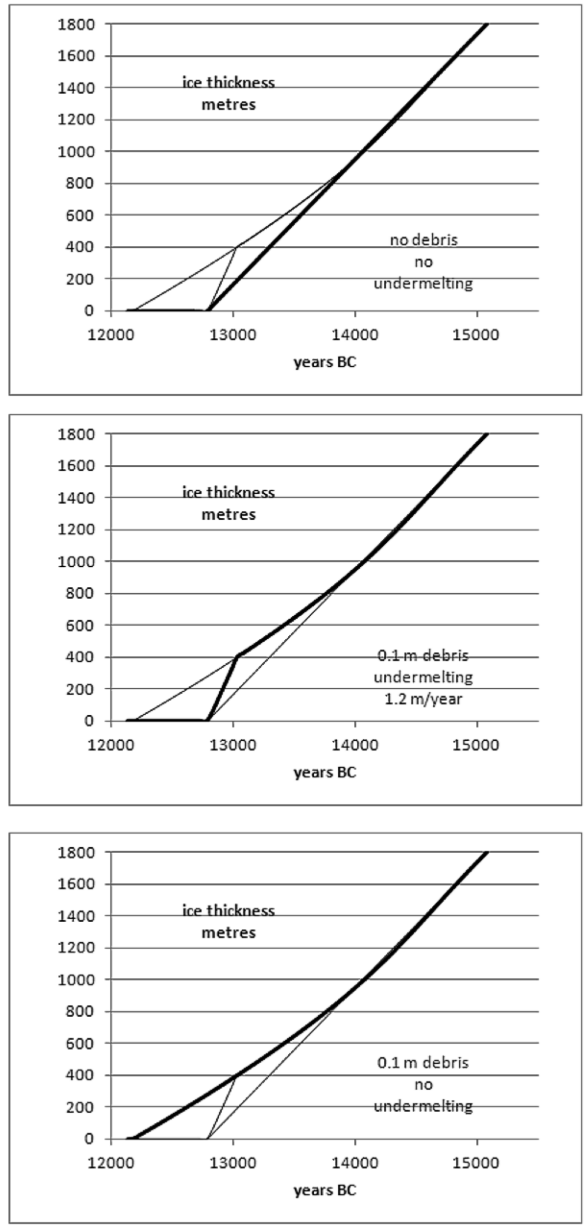
The third curve shows the same melting scenario as the second curve, except that there is no undermelting. The land does not become ice free until shortly before 12000 BC, 650 years after the sea has had become so.

Fall in local sea level

The timing of the fall in local sea level from its ice-age level on Gabriola of about 110 m ASML is a useful parameter, even if only to help distinguish between glaciomarine and glaciofluvial deposits. Two of the best reconstructions of sea-level change that I am aware of [34] [35] however require some modification to be applicable to Gabriola because Vancouver Island was not only depressed during the ice age, it was also tilted with the north end of the Salish Sea being deeper than the south end. The interpolated curve for Gabriola based on distances from the locations of the observations reported in the two published curves was entered into the program in table form.

Calving and other lateral movement of the ice

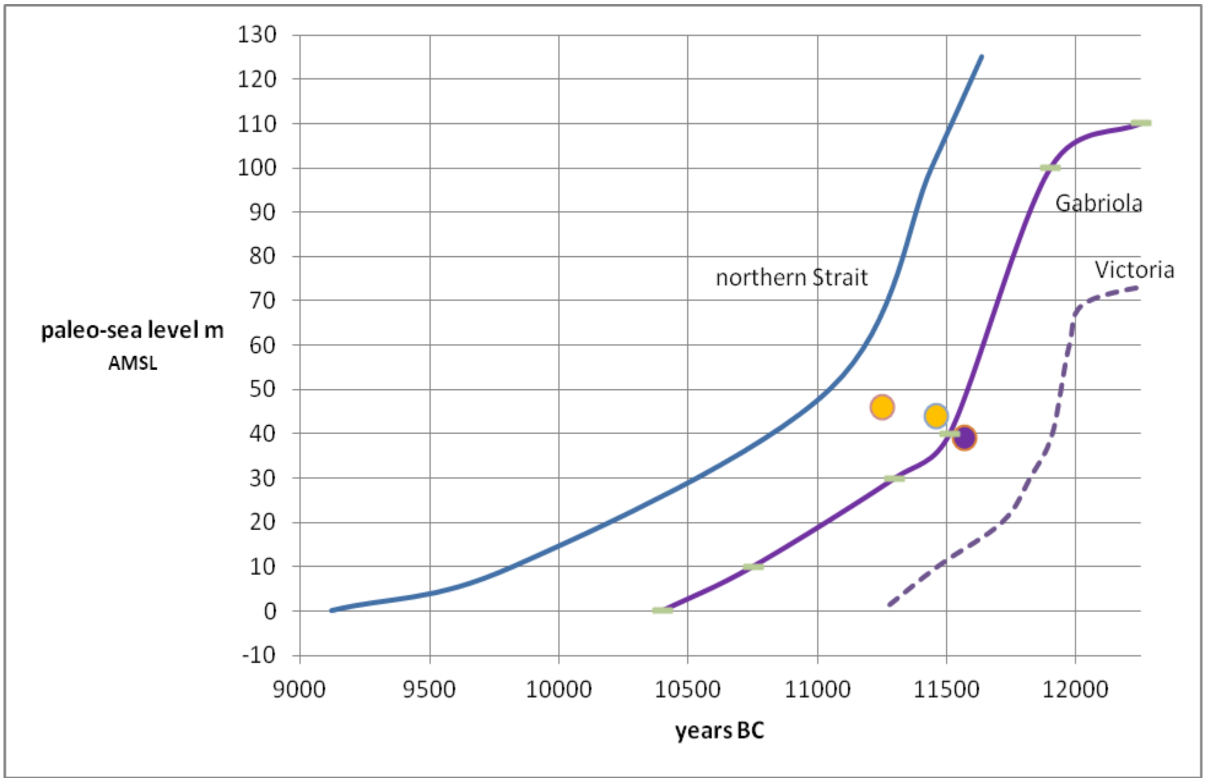
The analysis of the ice thickness over the ocean and the ice thickness over adjacent land can easily lead to predicted differences in surface heights of hundreds of metres. This is not realistic. When the height of vertical cliffs become greater than, say 70 metres, they calve or flow horizontally in some way so as to reduce the height.



Hypothetical ice decay over land and sea with (1) no englacial till or undermelting of sea ice; (2) englacial till and undermelting of sea ice; (3) englacial till over land. See text for details.

[34] T. James, E.J. Gowan, Ian Hutchinson, John J. Clague, & J.Vaughn Barrie, *Sea-level change and paleogeographic reconstructions, southern Vancouver Island, British Columbia, Canada*, Quaternary Science Reviews 28, pp.1200–1216, 2009.

[35] T. James, Ian Hutchinson, J. Vaughn Barrie, K. Conway, & Darcy Mathews, *Relative sea level change in the northern Strait of Georgia, British Columbia*, Géographie physique et Quaternaire, 59, pp.113–127, 2005.

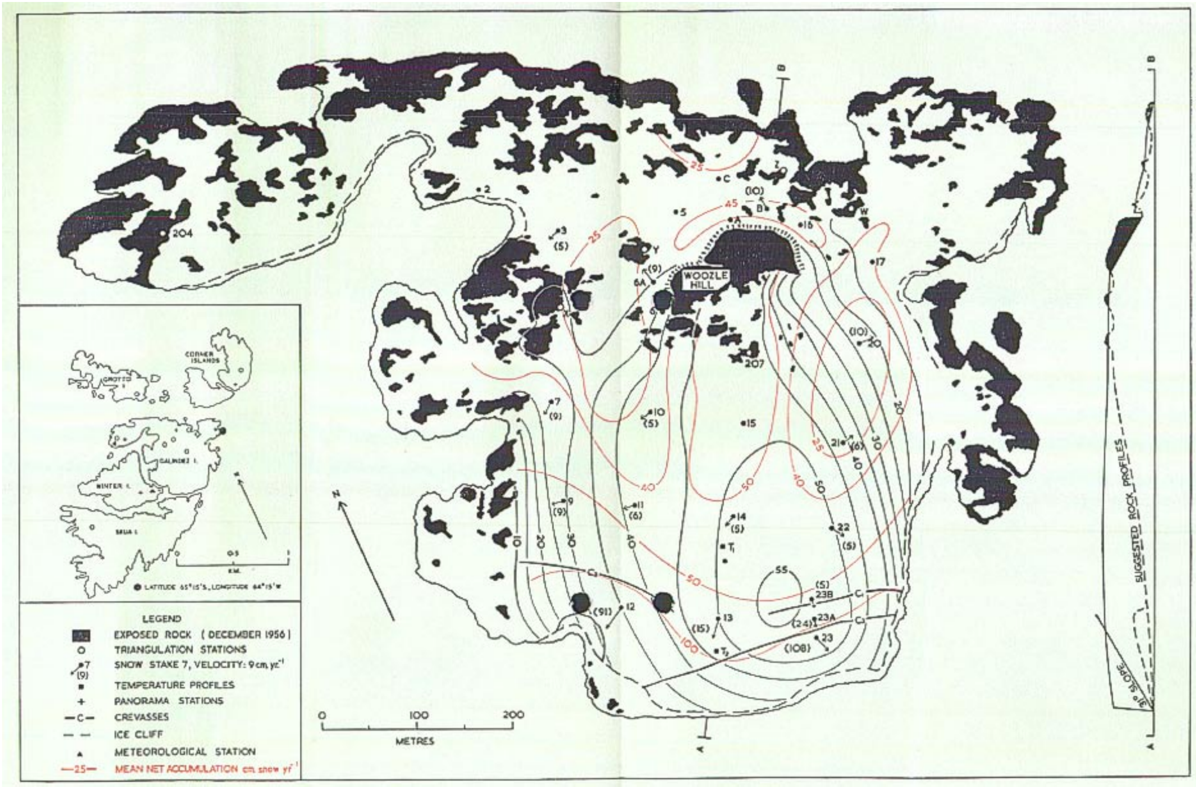


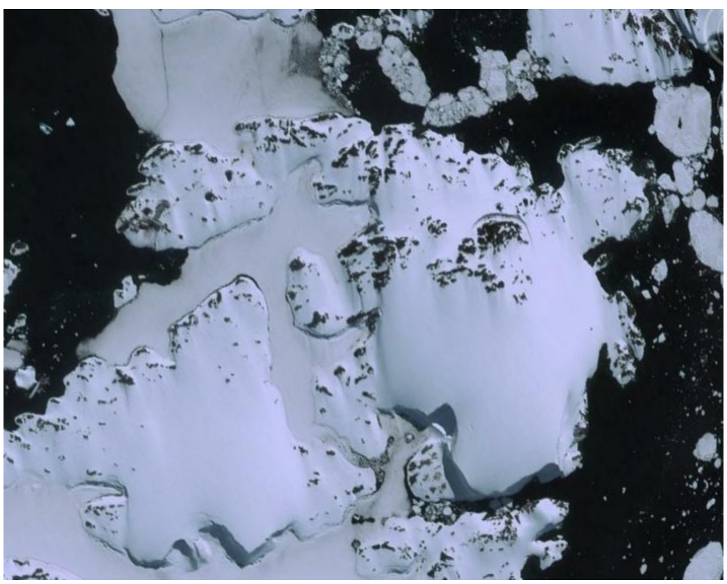
Sea level changes in the Victoria area and at the north end of the Strait of Georgia, with a conjectured curve for Gabriola. The two yellow circles are for a log (*left*) and plant material (*right*), and the blue circle with red rim is for a seashell sample, all three from Gabriola. The dashed line is the James et al. (2009) estimate for Victoria with a marine reservoir correction of 950 years. The solid blue line is their estimate for the northern strait (2005), also with a marine reservoir correction of 950 years. Levels below the current sea level are ignored.

The physics of this process is complicated and not something I could tackle here. Instead, for each time increment, I calculate the cliff heights that would be predicted if no calving or lateral movement were to take place, and if these heights exceed a given fixed value, then the height of the top of the cliff is reduced until it doesn't.

For most simulations, I used a fixed limit of 50–70 metres,[36] that is the height of the surface of the ice on land adjacent to the seashore was not permitted to exceed whatever the chosen value was above either the height of the surface of the ice over the sea, or the height of the ice-free sea before isostatic rebound commenced (110 m). Similarly, the height of the surface of the ice on highlands was not permitted to exceed whatever the chosen value was together with the bedrock height, (usually 50 m, corresponding to 160 m AMSL), above either the height of the surface of the ice over the sea, or the height of the ice-free sea before isostatic rebound commenced (110 m).

[36] Ice shelves and ice tongues can be hundreds of metres thick, but only if the water is deep enough to support them. See Bassis, J.N., *Bound to fail*, Lecture Notes, University of Michigan, 2013. Ice begins to deform under its own weight at 50 m.





Photograph from Google Earth (65°15'S,64°15'W, Dec.14, 2004). Map and notes from Thomas, R.H., *Studies on the ice cap of Galindez Island, Argentine Islands*, British Antarctic Survey Bulletin, 2, pp.27–41, December 1963. See also Sadler I., *Observations on the ice caps of Galindez and Skua Islands, Argentine Islands, 1960–66*, British Antarctic Survey Bulletin, 17, pp.21–49, December 1968.

Galindez Island, Argentine Islands, supports an ice cap, and provides an interesting example of what Gabriola might have been like during the final stages of deglaciation. Galindez is small, 1.1 × 0.7 km, but Gabriola would only have been about 25% of its present size and split into two islands when sea level was 110 m AMSL.

Galindez has a 50 m cap, but also exposed bedrock which absorbs solar radiation and causes extensive sub-surface melting. Ice cliffs along the shore are up to 50 m high; calving occurs at intervals of 5–10 years. Sea ice is gone by mid-summer. Run-off flows over the ice surface to collect in gullies and forms gushers at intervals along the coast. Run-off also escapes into cracks in the ice to emerge from the cliffs high above the sea. The cap has crevasses up to 30 m deep. Mean annual temperature is –5ºC and mean summer temperature around 0ºC with daily max to +8ºC. Rain is not uncommon. Precipitation is about 380 mm/year.

An ice-free strait

Obtaining a precise date as to when the Strait of Georgia became "ice-free" is not possible for several reasons,[37] but choosing a date, and getting the model to conform to it, is a useful way of restricting the values of some of the variables.

Some fairly-late dates in the literature are:

> "…most of the strait was ice free…" [38]

11300 ^{14}C BP=11250 BC

> "…A beach on Gabriola at 40 m AMSL." [39]

11641 ^{14}C BP = 11567 BC

> "…by 13.7 cal kyr BP Puget Sound, Juan de Fuca Strait and Strait of Georgia were ice free…" [40]

11830 ^{14}C BP =11 750 BC

Examples of earlier dates are:

> "…A beach on Gabriola (no RC date, based on sea level) at 104 m AMSL…" [41]

ca. 12100 ^{14}C BP = 12000 BC

> "…Paired pelecypod valves from northern strait…" [42]

12360 ^{14}C BP = 12385 BC

> "…shell fragments, Maltby Lake, southern Vancouver Island.…" [43]

12370 ^{14}C BP = 12394 BC

> "…ice-distal facies…containing many shell fragments.…" [44]

12000–12500 ^{14}C BP = 11897–12779 BC

> "…A calving embayment began to develop in the Strait of Georgia about 13,000 ^{14}C yr BP, and the Strait was completely deglaciated shortly thereafter.…" [45]

<13000 ^{14}C BP = <13280 BC

"…Saanich Inlet and Courtenay on Vancouver Island and Bowen and Texada islands northward in the Strait of Georgia were freed of ice before 12500 ^{14}C BP (12780 BC) by calving of a floating ice shelf.…" [46]

For better or worse, my adopted date for an "ice-free" strait was 12400 BC (12370 ^{14}C BP) when applied to points in the strait that are now 155 m deep (the current mean depth). Sea level at that time around Gabriola would have been about 109 m AMSL.

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[37] What does "ice-free" mean? All of the strait, or just the southern half? An opening, but only in deep water, or no ice in every fiord, shallow channel, and protected, embayment? Does pack ice count? Dates are often from marine organisms which require their radiocarbon age to be corrected for a marine reservoir constant whose value is not accurately known. Bivalves can be from deep water, or intertidal, and might have colonized the area some time after the ice had gone, or they may have lived beneath an ice shelf. And so on…

[38] Barrie, J.V., and K.W. Conway, *Contrasting glacial sedimentation processes and sea-level changes in two adjacent basins on the Pacific margin of Canada, in* Dowdeswell, J., and O'Cofaigh, C., eds., *Glacier-Influenced Sedimentation on High-Latitude Continental Margins*, Geological Society of London, Special Publication # 203,pp..181-194 2002.

[39] GD-533 Appendix 6.

[40] E.J. Gowan, *Glacio-isostatic adjustment modelling of improved relative sea-level observations in southwest British Columbia, Canada*, M.Sc. thesis, University of Manitoba, 2005.

[41] GD-533 Site 3. 900 year MRC.

[42] T. James et al. 2005 (*ibid*)

[43] T. James et al. 2009 (*ibid*)

[44] Guilbault J.P. et al., *Paleoenvironments of the Strait of Georgia, British Columbia, during the last deglaciation: microfaunal and microfloral evidence*, Quaternary Science Reviews 22, pp.839–857, 2003.

[45] Clague J.J. & James T.S., *History and Isostatic effects of the last ice sheet in southern British Columbia*, Science Direct, 21, pp.71–87, 2002.

[46] Wilson M.C., Kenady S.M., & Schalk R.F., *Late Pleistocene* Bison antiquus *from Orcas Island, Washington, and the biogeographic importance of an early postglacial land mammal dispersal corridor from the mainland to Vancouver Island*, Quaternary Research 71, pp.49–61, 2009.

Early trials of the model predicted the strait becoming "ice-free", as defined, several decades after the deepest part of the strait at 447m. Shallower parts could also have taken several decades to have become ice-free. However, the break-up of the ice would have been far more complicated than modelled here.

An ice-free island

If obtaining a date as to when the Strait of Georgia became "ice-free" is hard, then so is figuring out when the land on the island became ice-free. One can envisage a landscape of patches of bare ground with isolated plants and wetlands in hollows full of reeds and rushes, while nearby are great towers of dirty ice still only slowly melting away.

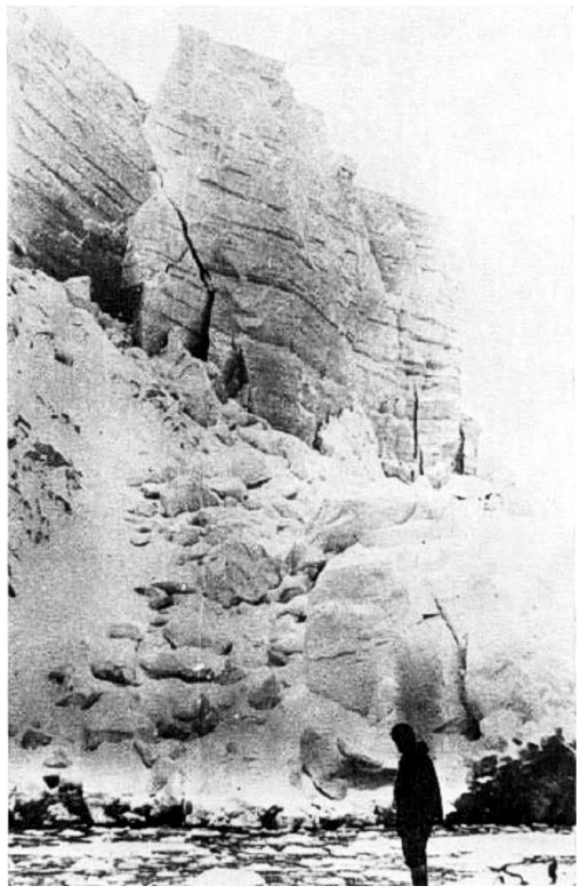
Radiocarbon dating gives us a clue as to when the earliest vegetation appeared on Gabriola. Three results have been obtained as follows:[47]

plant from McGuffies Swamp
11640 ^{14}C BP =11533 BC
log from Somerset pit
11590 ^{14}C BP =11460 BC

plant from Somerset pit
11310 ^{14}C BP =11251 BC

This data actually gives us two constraints on deglaciation. The McGuffies Swamp establishes that there was some vegetation on Gabriola before *ca*. 11530 BC, perhaps closer to 11590 BC (11700 ^{14}C BP), which is a date when a considerable increase in pollen grains has been observed in marine cores from the southern strait and bison were already present on southern Vancouver Island and the San Juans.

The other two figures establish that deglaciation of Gabriola was not complete before *ca*.11250 BC. This is because the plant material was buried under a glaciofluvial deposit. At what date all of Gabriola became completely ice free in summer is an open question. It might have been shortly after 11250 BC (11310 ^{14}C BP), say 11200 BC (11245 ^{14}C BP), or it might conceivably be, even if unlikely, only after the abrupt end of the Younger Dryas in 9600 BC (10050 ^{14}C BP).



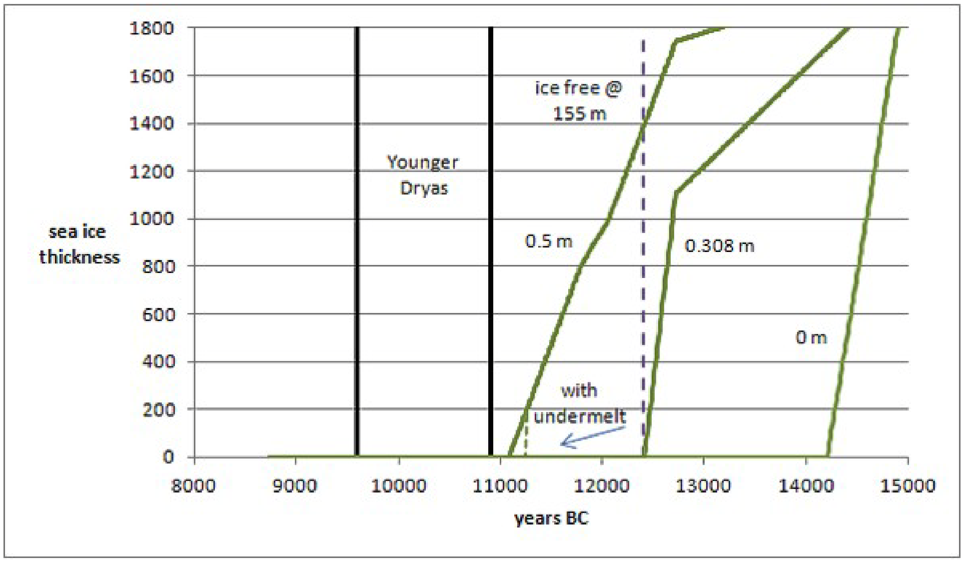
Ice cliff along the shore of Galindez Island, May 1966.

Photograph by J.A. Thoday.

[47] Details are in http://www.nickdoe.ca/pdfs/Webp533.pdf

Some results

Graph 1 marine supraglacial debris



1. marine supraglacial debris

A computer simulation of the thickness of ice over the sea for three thicknesses of supraglacial debris (D_{0s}=1) (0m, 0.308m, and 0.5m).

The black dotted vertical line corresponds to 12400 BC, which is when the strait is reckoned to have been deglaciated at positions corresponding to its present-day average depth of 155 m. The maximum depth was taken to be 447 m.

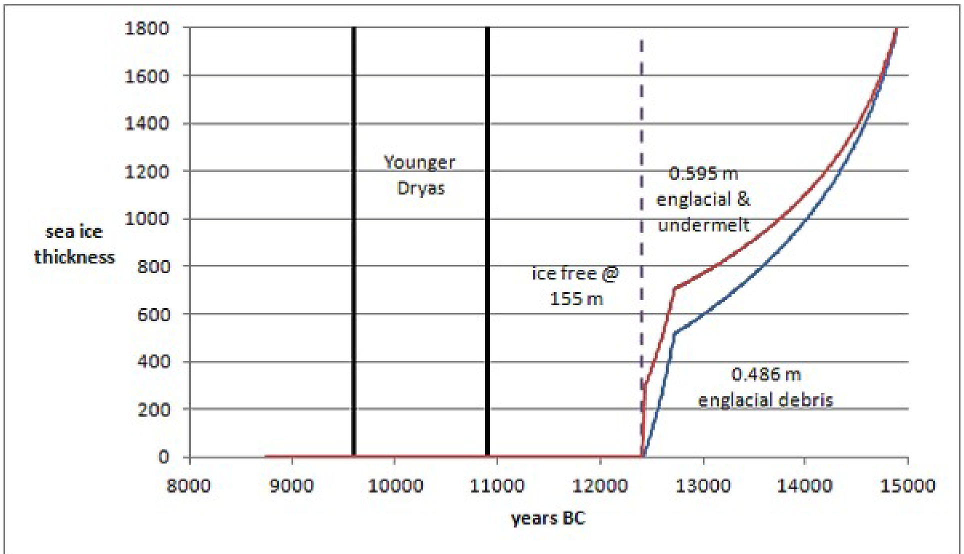
The maximum stable ice-cliff height was set at 50 m but there was no calving in this scenario. Ice surface cooling was set at 1ºC.

Undermelting by the sea as the ice lifted off the ocean floor plays little part in the melting in this scenario, even with an undermelt rate, as shown for 0.5 m debris, of 20 m/year.

It is impossible to reconcile the computed and observed "ice-free" date without supposing that there was some debris on the ice. If this supraglacial debris was of varying thickness, as it undoubtedly was, some areas of the strait would have been ice-free at very different dates from others.

If this simulation is correct, one has to imagine a chaotic situation with huge icebergs, extensive dirty pack ice, and open areas of sea existing at the same time.

Graph 2 marine englacial debris



2. marine englacial debris

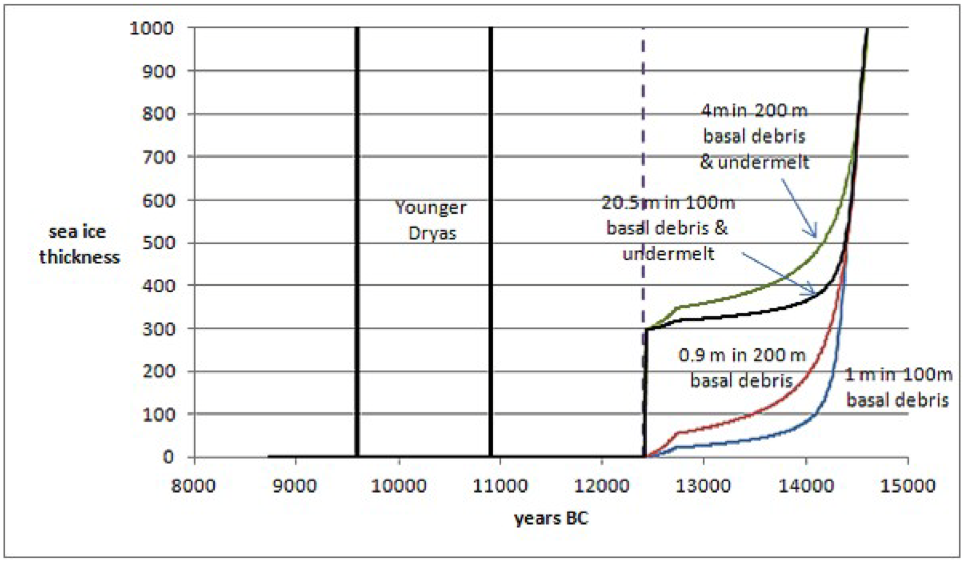
A computer simulation of the thickness of ice over the sea for two thicknesses of englacial debris (D_{0s}=10000) (0.595 m and 0.486 m). The black dotted vertical line corresponds to 12400 BC, which is when the strait is reckoned to have been deglaciated at positions corresponding to its present-day average depth of 155 m.

The maximum stable ice-cliff height was set at 50 m but there was no calving in this scenario.

The effect of englacial debris, the amount and distribution of which was variable and is unknown can be offset in the model by the undermelting rate, also an unknown. Reasonable values of these unknowns can easily match the required 12400 BC date for the sea becoming ice free.

If this simulation is correct, it is possible to reconcile an ice-free date for the sea of 12400 BC with supraglacial debris alone, or, as here, englacial debris distributed throughout the ice column.

Graph 3 marine basal debris



3. marine basal debris

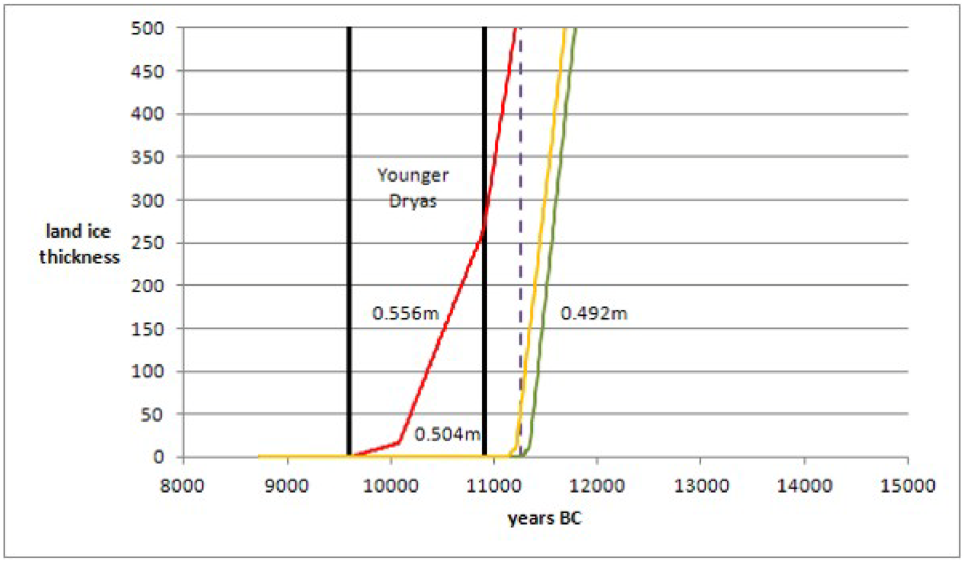
A computer simulation of the thickness of ice over the sea for four thicknesses of basal debris and no supraglacial and no significant englacial debris beyond the basal layer.

When there is no undermelting by the sea, melting of the sea ice proceeds rapidly until the ice is so thin that basal debris is at the surface and provides a thermal blanket. One curve (blue) shows the situation when 1.056 m of basal debris is concentrated in the bottom 100m of the ice (D_{0b} = 100). A second curve (red) shows the situation when 0.915 m of basal debris is concentrated in the bottom 200m of the ice (D_{0b} = 200).

When undermelting occurs, the situation is different. The undermelting rapidly removes basal debris so it never becomes a thermal blanket. Once undermelting has started, the thickness of the ice collapses and the debris is dumped as undermelt till. One curve (green) shows the situation when 4.03 m of basal debris is concentrated in the bottom 200m of the ice (D_{0b} = 200). A second curve (black) shows the situation when 20.55 m of basal debris is concentrated in the bottom 100m of the ice (D_{0b} = 100). The value of the undermelt rate is not critical (20m/year was used).

These first three simulations show that it is not possible to arrive at a single debris thickness or a single distribution using only the ice-free date for the sea.

Graph 4 land supraglacial debris



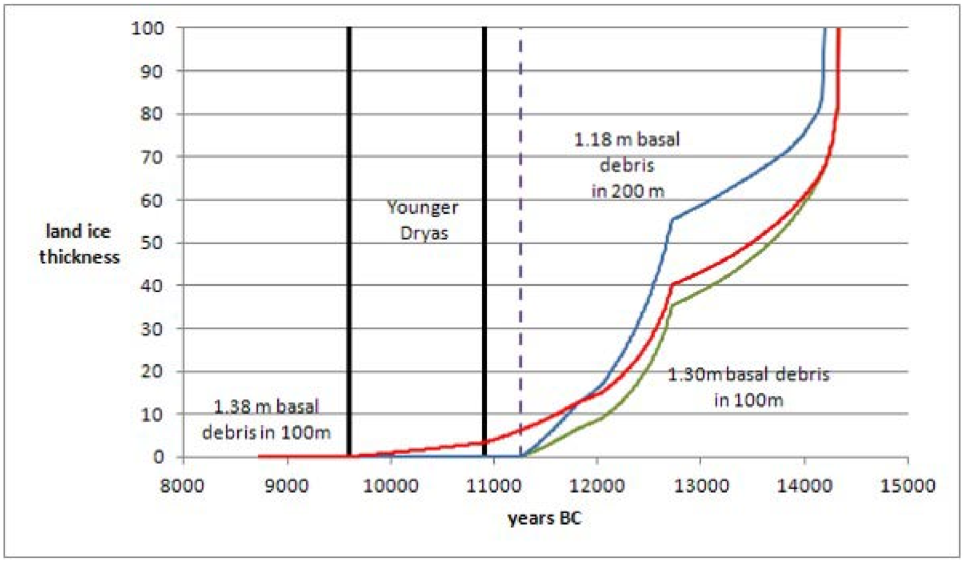
4. land supraglacial debris

A computer simulation of the thickness of ice over land. In this study, the land reference was taken to be the present 110 m AMSL, which was the late-Pleistocene sea level maximum for Gabriola. Simulations were also done for land 50 m higher (160 m AMSL), but it nearly all cases the difference in elevation made little difference.

The simulation was done for three thicknesses of supraglacial debris (D_{0s}=1) (0.492 m, 0.504 m, and 0.556 m) and no basal or englacial debris. The black dotted vertical line indicates 11250 BC (11310 ^{14}C BP) when it is known from radiocarbon dating that deglaciation was underway but not complete. The green line shows thickness for areas where deglaciation was complete by 11250 BC. The red line shows thickness where deglaciation was not complete until the end of the Younger Dryas. Note however that in this scenario, the thickness of the ice at 11250 BC would have been 500 m, which is impossibly high. The yellow line shows deglaciation being complete by 11200 BC, and the thickness of the ice at 11250 BC (11310 ^{14}C BP) as being a plausible 50 m.

This interesting result suggests that the glaciofluvial deposits on Gabriola overlaying former beaches were not due to an abrupt change in climate at the end of the Younger Dryas. Deglaciation of the land appears to have occurred very shortly after deglaciation of the adjoining sea, and to have occurred during the warm Bølling-Allerød Interstade that preceded the Younger Dryas.

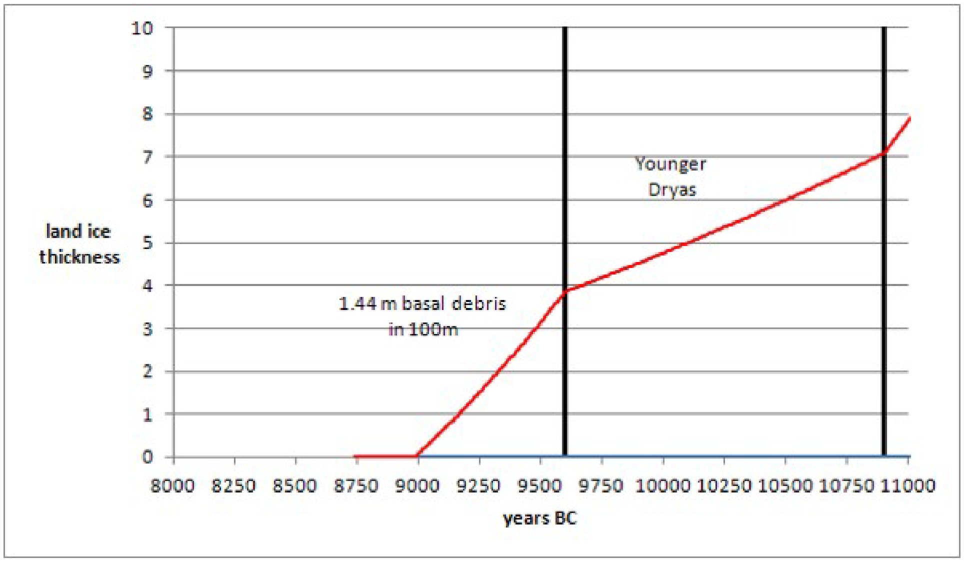
Graph 5 land basal debris



5. land basal debris

A simulation of the thickness of ice over land now at 110 m AMSL for 3 thicknesses of basal debris (1.30 m, 1.18 m, and 1.38 m) and no supraglacial debris. The black dotted vertical line corresponds to 11250 BC (11310 ^{14}C BP), which it is when it is known that deglaciation was underway but not complete. The green line shows deglaciation with debris concentrated in the bottom 100m of the ice (D_{0b} = 100) for areas where deglaciation was complete by 11250 BC (11310 ^{14}C BP). The blue line shows a similar scenario with debris concentrated in the bottom 200m of the ice (D_{0b} = 200). The red line shows a scenario where ice persisted until the end of the Younger Dryas.

Graph 6 ice cover throughout the Younger Dryas?

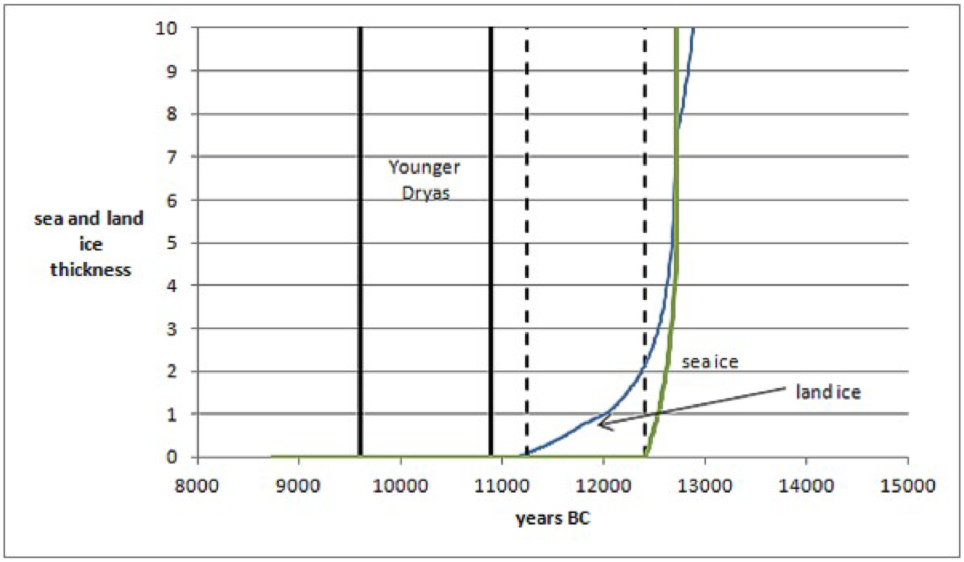


6. land basal debris in the Younger Dryas

A simulation of the thickness of ice over land now at 110 m AMSL for a basal debris thickness sufficient to protect the ice from melting during the Younger Dryas.

Although it seems quite possible for this to have happened in places, the rotten and buried ice would have been thin (note the thickness scale), and its melting at the end of the Younger Dryas would not have been fast enough to create significant glaciofluvial deposits despite the abruptness of the climate change. For ice to have persisted, it would have had to have had a relatively thick debris cover, and this thick debris cover would have prevented a quick response to climate change.

Graph 7 land and marine



7. land and marine, supa- and basal debris

A	total debris	1.60	1.70	1.80
B	sea ice free date	12500 (+100)	12400	12210 (–190)
C	land ice free date	11870 (+620)	11250	9490 (–1760)
D	% supra	16.9	17.9	18.9
E	sea ice free date	12420 (+20)	12400	12320 (–80)
F	land ice free date	11280 (+30)	11250	11220 (–30)
G	undermelt rate	.001	.00231	.003
H	sea ice free date	12120 (–280)	12400	12470 (+70)
J	land ice free date	11250 (0)	11250	11250 (0)

K	supra-depth	1000	10000	20000
L	sea ice free date	12140 (–260)	12400	12390 (–10)
M	land ice free date	11180 (–70)	11250	11250 (0)
N	basal-depth	9	9.38	10
P	sea ice free date	12430 (+30)	12400	12350 (–50)
Q	land ice free date	11310 (+60)	11250	11150 (–100)
R	sea depth	10	155	447
S	sea ice free date	12320 (-80)	12400	12430 (+30)
T	land height		0 110m AMSL	50 160m AMSL
U	land ice free date		11250	11220 (–30)
V	melt rate at 11250 BC		0.003	

Table notes

All dates are years BC. 50 m calving.

A (m) the same for sea and land

C the scenario is unrealistically sensitive to debris depths

G (m/year) these undermelt rates are unrealistically low

J undermelt rates can affect land ice because calving reduces land ice thickness if the surface level of the marine ice falls too far below the surface of the land ice. In this scenario, calving is not significant.

K values of D_{0s} in metres. Large values of the supraglacial depth D_{0s} likes these signify uniformly distributed englacial debris rather than surface debris

N values of D_{0b} in metres (m). Small values like these imply a high concentration of debris on the sole of the glaciers

R (m) positions in the strait where these are present-day sea depths. 155 m is the average depth.

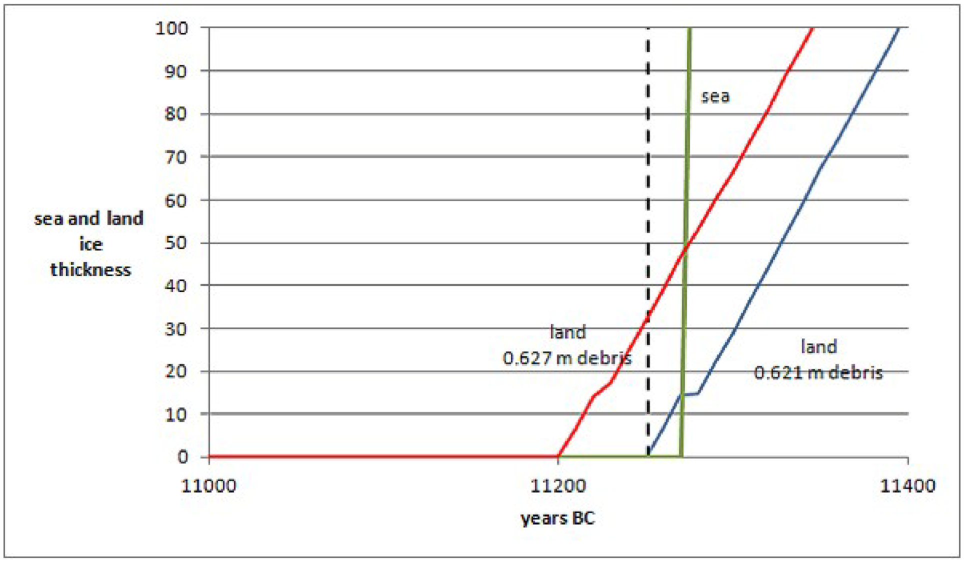
V (m/year) land rate. This is far too low to have created large glaciofluvial deposits

Comments

It was possible to find a combination of factors that produced both an ice free date over the sea at 12400 BC and an ice free date over the land at 11250 BC (11310 ^{14}C BP); however, the stagnant ice that endured buried under a blanket of basal debris would have been very thin. This scenario provides no explanation for the large glaciofluvial deposits observed at sites on Gabriola after the sea had left. There is effectively no melting at 11250 BC (V) which is when glaciofluvial deposits were not yet in place.

The undermelt rate of sea ice (G) is improbably low.

Graph 8 land and marine



8. land and marine, supa- and basal debris

A	total debris	0.627	0.621	0.42
B	sea ice free date		11260 (–1140)	12400 (0)
C	land ice free date	11200 (–50)	11250	12400 (+1150)
D	% supra	89	90	91
E	sea ice free date	11320 (–1080)	11260 (–1140)	11200 (–1200)
F	land ice free date	11310 (+60)	11250	11190 (–60)
G	undermelt rate	25	50	100
H	sea ice free date	11260 (–1140)	11260 (–1140)	11260 (–1140)
J	land ice free date	11250 (0)	11250	11250 (0)

K	supra-depth	230	330	530
L	sea ice free date	11030 (–1370)	11260 (–1140)	11620 (–780)
M	land ice free date	11020 (–230)	11250	11600 (+350)
N	basal-depth	0.2	2.0	20.0
P	sea ice free date	11260 (–1140)	11260 (–1140)	11260 (–1140)
Q	land ice free date	11250 (0)	11250	11240 (–10)
R	sea depth	10	155	447
S	sea ice free date	11230 (–1170)	11260 (–1140)	11310 (–1090)
T	land height		0 110m AMSL	50 160m AMSL
U	land ice free date		11250	11320 (+70)
V	melt rate at 11250 BC		0.64	

Table notes

All dates are years BC. 70 m calving.

A (m) the same for sea and land

B relative to 12400 BC. Unrealistically late date for sea when land date is right

C unrealistically early date for land when sea date is right

G (m/year)

E not nearly early enough

H not nearly early enough

K values of D_{0s} in metres

N values of D_{0b} in metres. Small values like these imply a high concentration of debris on the sole of the glaciers

R (m) positions in the strait where these are present-day sea depths. 155 m is the average depth.

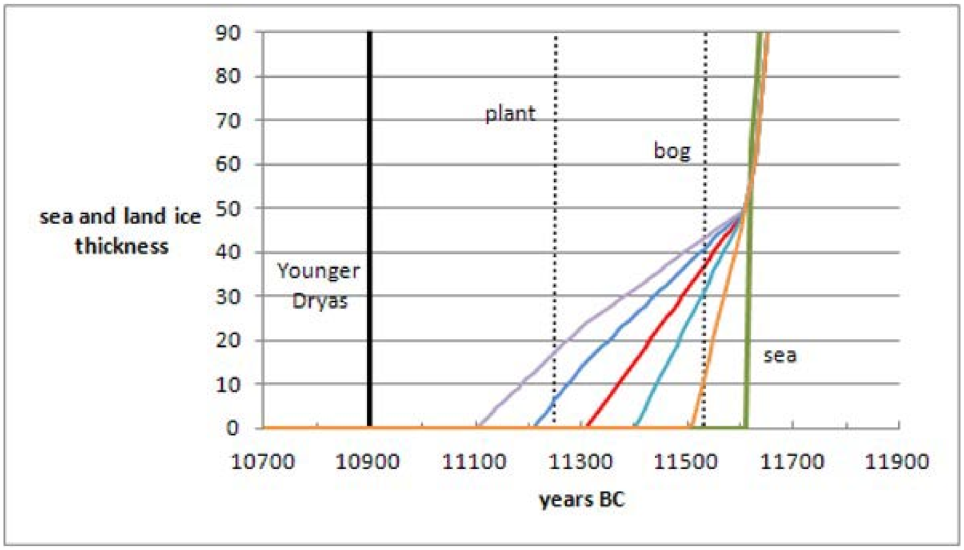
U high land deglaciated before low land

V (m/year) not high, but plausible

Comments

It was possible to find a combination of factors that produces both ice-free land *ca* 11250 BC (11310 ^{14}C BP) and a high melt rate over land (V), which would satisfy the need for incomplete deglaciation at 11250 BC and extensive flooding shortly after due to rapid melting of substantial amounts of ice. However, the flaw in this scenario is that the sea would not have been ice-free (B) until long after 12400 BC.

Graph 9 effect of variation of debris cover thicknesses



9. land debris variation

This simulation shows how variations in the thickness of the supraglacial debris alone over the land could account for the spread of dates for early post-glacial plant material found on the island. The thicknesses are for land now at 110 m AMSL.

The maximum stable ice-cliff height was set at 50 m. The ice free date over the sea was set at 11600 BC at a not-critical 155 m depth.

Unlike in the previous scenarios (1–8), the thickness of the debris over the sea was allowed to be different from the thickness over the land. Also unlike in the previous scenarios (1-8), the Allerød Interstade climate was conjectured to have warmed to that postulated for the early Holocene (post-Younger Dryas) climate between 11300–10900 BC.

The dotted vertical lines in the graph indicate results of radiocarbon observations in calendar dates on Gabriola. The line on the right is for plant material from a "bog" at an elevation well above 110 m AMSL. The line on the left is for terrestrial "plant" material deposited on a former beach after the sea had left, but before the area was inundated with glaciofluvial sand from higher elevations implying that although the area was ice free, land at other locations was not.

The land supraglacial debris thicknesses from left to right are:

purple 0.928 m ice-free at 11100 BC;
blue 0.88 m ice-free at 11200 BC;
red 0.82 m ice-free at 11300 BC;
turquoise 0.7 m ice-free at 11400 BC; and
orange 0.64 m ice-free at 11500 BC.

The marine supraglacial debris cover was 0.476 m. The undermelt rate was 10 m/year, but this value could be varied widely with only trivial effect on results.

There were insignificant differences in the ice thickness at 110 m AMSL and 160 m AMSL.

Discussion

The ice-age sites on Gabriola (described in a companion paper)[48] fall into the following categories:[49]

(A0) peat over clay with little or no stony surface till (Sites 1, 2, 5, 8, 9). These sites are found both below and above the late-Pleistocene sea level high

(B0) stony till over sand over clay over marine shell-rich silty sand (Sites 10, 12)

(B1) stony till over sand over marine shell-rich silty sand (Sites 4, 6, 7, 11, 14)

(B2) stony till over sand (Site 3)

(B3) stony till over clay (Site 13).

The stony till is patchy and may be absent in places.

The presence of overlying stony ablation till (B0–B3) suggests much of the glaciofluvial till (sand) was deposited subglacially or below debris-laden rafts of ice.

The presence of a thick sand deposit (B2) above the late-Pleistocene high of 110 m suggests a source of the sand above that elevation, but not as high as 130 m (A0, Site 2). This is probably sourced from pre-glacial deposits, quite possibly Quadra Sand.

Sand deposits always overlay marine deposits (B1) suggesting they are glaciofluvial sediments deposited by meltwater after the sea had retreated from the area. Sandy glaciofluvial deposits may well have been pre-existing deposits that had escaped being swept away by the ice. Preservation of such deposits can occur when ice flows obliquely across a valley or a broad gulley, across a local depression, or down over a steep cliff (leaving a cavity at the foot of the cliff as does a waterfall). Ice moved over Gabriola at differing times from the northwest and from the northeast, making south-facing valleys most likely to have escaped being scoured by either of these flows. The MoTI Pit on Gabriola does just that.

The simulations described in this paper suggest that these glaciofluvial deposits were not a result of warming at the end of the Younger Dryas. They rather likely occurred in the waning days of the Bølling-Allerød Interstade *ca*.11250–10900 BC (11310–11000 ^{14}C BP).

Deglaciation of Gabriola and the surrounding sea was largely complete before the onset of the Younger Dryas, although it cannot be ruled out that some stagnant ice buried under a thick layer of debris persisted locally into the Younger Dryas.

Deglaciation of the land occurred only a few years (less than a century, perhaps only a few decades) after the sea had become ice free.

The final stages of the deglaciation of the ice over land were likely rapid, indicating a warm climate among other possibilities.

Precipitation runoff may have added substantially to meltwater runoff because, back then, evapotranspiration on the island would have been low because vegetation was sparse; and the island's aquifers in fractured bedrock may have been still choked with glacial silt and ice.[50]

[48] http://www.nickdoe.ca/pdfs/Webp533.pdf

[49] Not including lodgement till which is rare and of obvious origin.

[50] A modern water budget shows 44% of the island's annual precipitation being lost to evapotranspiration, and only 12% lost as visible runoff over the surface. There is thus potential in different circumstances for a substantial increase in runoff volume without any increase in precipitation. http://www.nickdoe.ca/pdfs/Webp214c.pdf

Supporting evidence for a warming climate toward the end of the Bølling-Allerød is sparse. However, in a study of lacustrine sediments from Castor Lake in north central Washington, it was noted that there was an almost doubling of organic carbon in the study's Zone II which started in 11274 BC (13224 cal BP). Possibly causes are "increasing flux and preservation of terrestrial organic matter…" and climate "warming with increased lake and watershed productivity".[51]

Other possible causes (besides climate change) of the flooding during the final stages of deglaciation are the collapse of ice dams; the warming of the ice due to absorption of solar radiation from recently deglaciated bedrock; and instabilities in deposits created by withdrawal of the sea.

The debris on, in, and under the ice was mostly supra- or englacial debris with a non-critical vertical distribution; however, as the debris-laden surface of the ice transitioned from being high above the island to being draped over the island—and it transitioned from being 2-dimensional to 3-dimensional—crevasses would have developed, and the horizontal distribution of the debris would have been greatly altered. Crevasses would have played the role of gullies in an ice-free environment, and would have greatly increased the intensity of the meltwater runoff locally. This complex phenomenon was not modelled but effects that may be attributed to it have been observed.

The total debris thickness was not more than one metre, as is observed.

[51] Thornburg, J.D. *The Younger Dryas transition observed in lacustrine sediments from Castor Lake, Washington*, B.Sc. thesis, College of Earth and Mineral Sciences, Pennsylvania State University, 2006.

Undermelt rates of sea ice once the ice became ungrounded were approximately 20 m/year, but could have been significantly higher or lower within this order of magnitude.

It is likely that debris depths were very variable from place to place. The sea likely eroded undermelt debris by forming tidal channels rather than uniformly weathering the debris to a level surface. The sea likely opened up small sheltered beaches, leaving mounds of stagnant ice remaining in the immediate high ground surrounding these beaches.

The sea likely brought pack ice laden with supraglacial debris to the island that was subsequently grounded on beaches as sea level fell. Conversely, in some places, the sea removed debris-laden ice leaving the shore depleted of ablation debris.

The sea would have been relatively calm as wind energy would have been dissipated by moving pack ice around rather than creating waves.

Shortcomings of the model

The most significant shortcoming of the simulation is its failure to reconcile the dates of deglaciation of the strait and of the land. The elapsed time used in the simulations—12400 BC for the sea, and 11250 BC for the land —amount to 1150 years and this is too high.

Although it is possible to get the simulation to match these dates, by for example, supposing that the debris cover over the strait was less than the debris cover over the land so that the ice over the sea melted faster; these tactics fail to address an essential problem. This is that it is unrealistic to suppose that the height of the ice cliffs along the then sea shore were much more than 50 metres high, and the indications are that this ice melted fairly quickly and inundated former beaches with copious amounts of sand after sea level had fallen. Cross-bedding in these glaciofluvial sand deposits suggest a flow of surface water far greater than is seen today on the island; implying that the melting of the land ice was accomplished in a fairly short time span, possibly less than a century. The simulation, when adjusted to meet the 12400/11250 BC dates, however predicts extremely low melt rates for land ice and very thin ice cover of only a few metres. Efforts to persuade the simulation to produce higher melt rates by supposing a warming climate *ca* 11250 BC failed to produce significant improvement in these results.

One possibility that would make the 12400 BC date too early is that the marine reservoir correction used in dating shellfish and foraminifers is too small. Most calibrations use 800–900 years, but other authors have suggested something like 1100 years would be a better estimate. This would delay the date of marine deglaciation by 300 radiocarbon years or so,[52] but not much more.

Another possibility is that the island together with Mudge Island and Vancouver Island was a bight covered in dead ice based on these islands and grounded on surrounded shallows.[53]

Is the land deglaciation date of 11250 BC too late perhaps? This is based on the dating of plant material, and there might have been bare ground for a century or two before that. But that is not enough to close the gap substantially. If supraglacial debris is thick enough on stagnant ice, it can support vegetation including trees.[54]

Conclusion

The simulation does a good job of predicting the rate of disappearance of the ice from 15000 BC onward, but fails in the very last stages of deglaciation because of its simplicity in assuming that debris cover, which acts as a thermal blanket, was uniform. The reality is more likely that it was very varied.

Was there an ice cap on Gabriola? Possibly yes, but it was short-lived and chaotic, and some high ground was exposed nunatuk-style early on. The earliest post-glacial radiocarbon date obtained on the island is from the highest elevation, and the model suggests elevation had very little effect on deglaciation rates.

The abrupt end of any cap was not due to climate change at the end of the Younger Dryas, but to the warming climate at the end of the Bølling-Allerød Interstade *ca* 11200 BC and the exposure of bedrock to the sun in the final stages of deglaciation. ◊

[52] Also about 300 calendar years at this time.

[53] A suggestion that this is what resulted in more glacial conditions around Vancouver, Denman, and Hornby Islands *ca* 12020 ^{14}C BP (11900 BC) is contained in Guilbault J.P. et al. 2003.

[54] Perhaps this was the origin of the western yew found at Site 12, http://www.nickdoe.ca/pdfs/Webp533.pdf

Articles:

Sea level changes.