

FIELD NOTES ON THE ATHABASCA GLACIER AND COLUMBIA ICEFIELDS REGION

These notes were adapted from course field-trip notes. While not specifically written for the conference field trip to the Columbia Icefield, they will still give participants a good general overview of the features of the region.

Summary

These field notes give a brief overview of the climatological and glaciological setting of the Athabasca Glacier and Columbia Icefields region. During the visit we will focus on the climate, physical environment, and glacial history in the area, with emphasis on glacier advance and retreat during the Little Ice Age and 20th century. Sections 1-3 briefly outline the climatological and glaciological points that we will discuss at the Crowfoot Lake, Bow Summit, and Athabasca Glacier stops on the Icefields Parkway. Weather permitting, we will take some time at the Athabasca Glacier to travel on the glacier surface and in the forefield, where we will examine a number of glacial landforms. The recent demise of the Athabasca is vividly evident in the short walk from the Icefields Parkway to the terminus of the glacier.

What to Bring

Expect anything in October! Average October temperatures in Lake Louise are 1°C, and precipitation is just as apt to be snow as rain. On the Icefields Parkway this means that sub-freezing temperatures are likely and snow is very possible. Bring a winter jacket, toque, gloves, and sturdy footwear. There are also glorious autumn days on the Icefields Parkway in October, so hope for the best! Good running shoes or light hikers are fine for the walking that we will do. If there is too much new snow, we will limit the hiking and do most of the discussions close to the tour bus.

1. The Columbia Icefields: Climatology

A suite of climate records from Environment Canada weather stations in the Icefields vicinity is shown in Tables 1 and 2, with Calgary Airport thrown in for good measure. These are the 30-year climate normals for the period 1971-2000, with temperature data (min, max, and mean) derived from daily averages. These stations are all in valley settings, well below the elevation of the Icefields. Sunwapta Pass, which divides Jasper and Banff Parks, is at 2035 m, while the present glacier terminus is close to 2100 m and the icefall that you can see from the road is at 2700 m. The upper accumulation plateau of the Icefield extends above this to 3500 m.

Despite being at too low an altitude to really tell us what is going on in the upper reaches of the Icefield, the Lake Louise and Sunwapta climate normals give some idea of the background climatic state needed to support glaciers and icefields. Mean annual

temperatures at both sites are just below 0°C, with average daily summer (June July, August) temperatures of close to 10°C. With typical cooling rates of 6°C/km in the lower troposphere, one can approximate mean annual and summer temperatures of ca. -3°C and 7°C at Sunwapta Pass. At 3000 m, representative of the Icefield plateau, there are no long-term meteorological observations but similar extrapolation gives mean annual and summer temperatures of -9°C and 1°C. The latter number is key; it is cool enough for relatively little summer melt, while precipitation that falls at upper elevations is likely to be snow in most months of the year. Winter snow accumulations in Lake Louise and Sunwapta also illustrate the higher amount of snowfall in the continental divide region.

The sparse vegetation on Sunwapta Pass and in the vicinity of the Athabasca Glacier gives testimony to the harsh climate in the Icefields region. The Columbia Icefield straddles the continental divide, providing a relatively smooth passage for nasty wet B.C. weather systems that bring low clouds and precipitation. Average annual snow accumulations of ca. 7 m (water equivalent) are typical on the Icefield. It can snow in any month of the year; in fact, there were several snowfalls on the Icefield this past August. When clear skies prevail at Sunwapta Pass, the area is still buffeted by cold katabatic winds that drain through the Athabasca valley.

Site	Banff	Calgary	Jasper	Lake Louise	Sunwapta
Elevation (m)	1383.7	1084.1	1062.2	1524	1554.5
Latitude	51° 11'N	51° 07'N	52° 53'N	51° 26'N	52° 27'N
Longitude	115° 34'W	114° 01'W	118° 04'N	116° 13'W	117° 27'W
T _{max} (°C)	8.9	10.5	9.2	7.0	6.8
T _{min} (°C)	-2.9	-2.4	-2.6	-7.7	-7.3
T _{mean} (°C)	3.0	4.1	3.3	-0.3	-0.3
T _{summer} (°C)	13.6	15.2	14.1	11.0	10.6
Total Precipitation (mm)	472.3	412.6	398.8	569.3	488.9
Rain (mm)	296.2	320.6	296.7	265.0	284.1
Snow (cm)	234.1	126.7	138.0	304.3	205.2
Winter (DJF) Snowfall (cm)	106.6	48.7	79.1	164.7	92.0
Spring (MAM) Snowfall (cm)	67.6	47.0	26.9	59.3	55.9

Table 1. Climate normals for several sites in the Canadian Rockies, derived from daily data, 1971–2000.

Of course, this climate does offer a hospitable environment for glaciers to thrive. In fact, it is an ideal mix of cold continental and moist maritime conditions; most months of the year are cold enough for snow, while prevailing westerlies nourish the icefield with moist Pacific air masses. Summer heat can still be intense, with temperatures reaching 20°C. This can be punishing on the outlet glaciers, which experience several m of summer melt. The net annual accumulation vs. loss of snow on a glacier is known as its *mass balance*. The mass balance is key to determining whether a glacier will grow or shrink in a given year. Studies relating regional meteorological data with long-term mass balance observations on the Peyto Glacier, an outlet of the Wapta Icefield, indicate a strong correlation between summer temperature and annual mass balance (Young, 1981; Yarnal, 1984). It has long been thought that summer temperatures are the primary control of glacier health in the Rockies, typical of continental climates. As discussed in a bit more detail below, this view has recently been questioned thanks to solid evidence of a strong maritime (North Pacific) influence on mass balance in the Rockies.

Lake Louise, AB (1524 m)													
	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Year
T (°C)	-13.8	-10.2	-5.4	0.5	5.6	9.4	12.0	11.6	6.8	1.0	-7.9	-13.7	-0.3
R (mm)	0.1	0.0	2.3	5.7	35.9	53.5	57.3	54.3	39.4	15.2	1.3	0.0	265.0
S (cm)	52.9	40.2	33.2	19.7	6.4	0.3	0.0	0.0	3.2	20.0	56.9	71.6	304.3
P (mm)	53.0	40.2	35.4	25.4	42.3	53.8	57.3	54.3	42.5	35.2	58.2	71.6	569.3
Sunwapta, AB (1555 m)													
	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Year
T (°C)	-12.3	-2.1	-5.1	0.7	5.7	9.3	11.4	11.0	6.4	0.9	-7.9	-12.9	-0.3
R (mm)	0.0	0.0	0.0	3.3	27.3	57.2	67.4	70.3	38.6	17.2	2.7	0.0	284.1
S (cm)	30.9	24.6	28.2	20.6	7.1	1.3	0.3	0.0	3.9	18.1	33.6	36.5	205.2
P (mm)	30.9	24.6	28.2	23.6	34.4	58.6	67.7	70.3	42.5	35.3	36.2	36.5	488.9
Jasper, AB (1061 m)													
	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Year
T (°C)	-9.8	-6.3	-1.2	4.3	9.1	12.8	15.0	14.5	9.8	4.5	-4.0	-9.2	3.3
R (mm)	4.5	2.8	5.1	12.0	28.7	54.7	60.1	59.0	35.9	22.1	8.3	3.4	296.7
S (cm)	30.5	18.3	16.9	8.6	1.4	0.3	0.0	0.2	1.9	8.0	21.6	30.3	138.0
P (mm)	26.9	16.0	17.6	18.8	29.9	55.0	60.1	59.1	37.3	28.7	24.5	24.8	398.8

Table 2. *Monthly climate normals at Lake Louise, Sunwapta, and Jasper, 1971-2000 data. Temperature data represents monthly averages and rainfall (R), snowfall (S), and precipitation (P) represent monthly totals.*

Figure 1 illustrates the 20th century mean-annual temperature trend in Banff, Alberta, based on Environment Canada observations that date to 1894. With the exception of recent winter temperatures, it is difficult to visually detect trends in this plot. Indeed, available meteorological records from the Canadian Rockies do not indicate as much 20th century warming as is seen elsewhere in Canada (e.g., the MacKenzie Valley). However, it is also a matter of perspective;

20th century warming in Banff is of order 0.7°C and is difficult to detect due to the fact that interannual variability is greater than this. The 1990s are telling, however, as seen in Table 3. Banff was 0.3°C warmer in the 1990s relative to the average 1960–1990 temperatures, and January temperatures in Banff were 2.1°C warmer than the 1960–1990 average. Winter warming dominates the climate change pattern in the Rockies, with large impacts on total snowfall in the region in the 1990s.

These recent temperature and snowfall trends in the Rockies largely reflect what is going on throughout the world. The 1990s capped the warmest century, globally, in several thousand years, leading to glacial retreat throughout the world’s mountain regions.

Site	ΔT_{JAN} (°C)	ΔT_{JUL} (°C)	ΔT_{ANN} (°C)	ΔS (%)	ΔP (%)
Banff	+2.1	-0.6	+0.3	-12%	+ 3%
Lake Louise	+1.8	0.0	+0.3	-23%	-16%
Jasper	+2.7	-0.3	+0.6	-12%	+ 4%

Table 3. *Temperature, snowfall, and total precipitation trends in the Canadian Rockies in the 1990s (Environment Canada, 2003).*

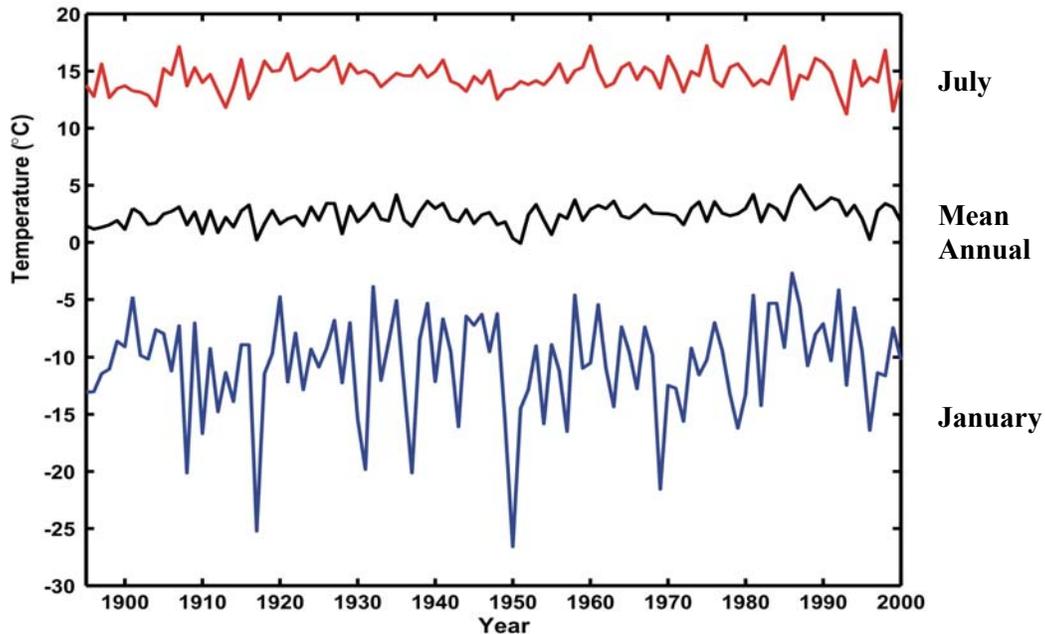


Figure 1. *20th century temperature data from Banff, AB. Only the winter temperatures show a discernible warming in recent decades.*

The Peyto Glacier mass balance record reflects these temperature trends. Annual mass balance observations on the Peyto have been carried out by the federal government since 1966, the

longest continuous observation in western Canada. There have been only 7 years with a positive net mass balance in the 35-year record: 1966–1968, 1973, 1974, and 1976 (Figure 2). So yes, your parents are telling the truth when they bemoan the snowy years of old. The 2002-2003 mass balance season has been another punishing year on the Icefields and throughout the Canadian Rockies: a slightly below-normal snowpack and a Dantean summer.

Decadal and multi-decadal synoptic-scale meteorological conditions in the Pacific Ocean (the *Pacific Decadal Oscillation*, or PDO) appear to present a major control on snow-bearing weather systems in the Rockies (Hodge et al., 1999). Mike Demuth, a Geological Survey of Canada researcher that has carried out the Peyto Glacier mass balance observations for more than a decade, feels that annual snow accumulation – hence the PDO – is the primary control on Peyto mass balance trends, rather than summer temperatures (Demuth and Keller, 1997). A recent analysis by Shea et al. (submitted) indicates that for snow accumulation, spring precipitation is the major control on glacier distribution in the Canadian Rockies. This is particularly true for the eastern slopes, where spring is the wet season and most spring and autumn precipitation falls as snow at altitudes above 2000 m. This important control on total annual snow accumulation has been overlooked because lower altitude climate stations (e.g., Lake Louise) receive much of this precipitation as rain.

These analyses that point to the importance of snow accumulation in glacier health in the Rockies. Good snow years, such as 1999, can offset hot summers to a point. However, the summers of recent years have been hot enough to trump even excellent winters, giving a net loss of mass. The Peyto Glacier lost mass in 1999 and 2002, two years with profusive spring and early-summer snowfall, due to the intensity of the summer melt season. We may therefore be entering a climatic regime where continental effects dominate the maritime influences on glaciers in the Canadian Rockies. Linear correlation analyses on the winter, summer, and net annual mass balance in Figure 2 suggest that the summer melt season and the winter accumulation season weigh in almost equally on the net annual mass balance, with both influences important ($r = 0.77$ for the summer balance and $r = 0.67$ for the winter balance).

2. The Columbia Icefields: Glaciology

The Icefields Parkway drive offers views of dozens of glaciers – active, perennial ice masses. There are a variety of glacier classifications in general usage, based primarily on glacier morphology. Most of the ice masses that we will see are cirque, niche, or outlet glaciers. The first two glacier types are common in the Rockies, generally occupying sheltered north-facing aspects. The sedimentary stratigraphy of the region also offers many opportunities for differential erosion and cliff-top or cliff-side niches. The Athabasca Glacier is an archetypical outlet glacier, draining a large reservoir of ice that we only get hints of from the highway: the Columbia Icefield. The Columbia Icefield, like the Wapta Icefield to the south, sprawls over a broad area of high-elevation summits, which provide a large area of net snow accumulation. These two regions are icefields (rather than ice caps) because the underlying topography controls the morphology and dynamics of the ice mass. A true ice cap (as found in Iceland or the Canadian Arctic, for instance) over-rides the underlying topography to form an ice dome that effectively creates or dominates the topography in a region.

We will stop at the toe of the Athabasca Glacier, the much-visited and photographed outlet of the Columbia Icefield that is about 1 km from the parkway. The forefield area, between the glacier terminus and the parkway, is littered with recessional terminal moraines from glacier retreat in the 20th century. When explorers first visited the Sunwapta Pass area, the Athabasca actually

extended downvalley to the position of the present-day visitor centre, blocking straightforward passage through the valley. The current parkway would not have been feasible and the original route climbed to the backside of Wilcox Peak, a very pleasant hiking traverse if you get the chance. After strolling through the forefield, we will climb up onto the glacier for a quick look at some features of the glacier in the ablation (melting) zone. The toe of the Athabasca is gently-sloping, permitting comfortable walking in hiking boots or good runners on most summer days. The glacier surface can be slippery if it is misty, if it has been freshly rained on, or if it is early morning and it froze overnight. By late morning or afternoon, in summer, surface water is melting once again and the surface ice becomes hospitably rough and soft.

In the vicinity of the glacier terminus, the most evident surface features on the glacier will be meltwater forms and crevasses. Crevasse fields develop by brittle failure of ice that is under tension. This can arise as a result of undulating bed topography (like the cracked-up icefalls high above), velocity/stress gradients in the ice, or shearing along valley walls where friction holds back the ice. Surface meltwater drains down the ice in *supraglacial* surface channels, or it dives below the surface into crevasses and *moulins*. Moulins, which can be found further up from the terminus, are vertical drainage “pipes” that can reach frightening dimensions. Water that disappears into crevasses and moulins may or may not reach the base of the glacier.

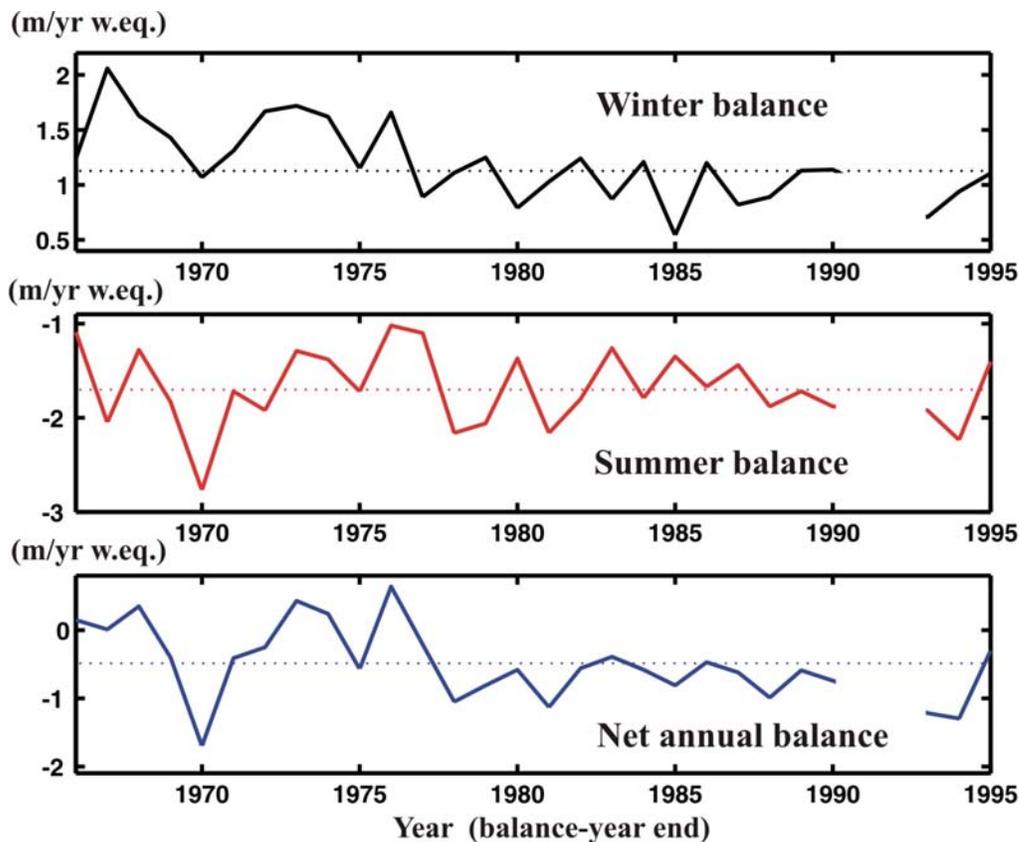


Figure 2. Mass balance observations on the Peyto Glacier, 1966–1995, adapted from Demuth and Keller (1997). Winter balance is the net accumulation of snow during the previous winter and spring, measured in May or June of each year. Summer balance is the total amount of summer mass loss due to surface melt, measured in September of each year – the end of the “balance year.” Net annual balance is the sum of the winter and summer balance. All variables

are averaged over the area of the glacier and are expressed in m of water equivalent. Dashed lines indicate the 30-year means. No data was collected in 1990-91 and 1991-92.

The Athabasca Glacier and the Saskatchewan Glacier, a major outlet of the Columbia Icefield, which lies to the south, have been the sites of some seminal work in glaciology in the 1960s and early 1970s. Stan Paterson, who literally wrote the textbook on glacier dynamics, worked on the Athabasca in the late 1950s and early 1960s as part of his Ph.D. work at UBC. Fragments of his field efforts can still be found on the glacier – aluminum pipes and scattered nuts and bolts. Working in the area from the glacier terminus to the base of the lower ice-fall, Paterson drilled to the glacier bed with a hot-point (thermal) drill, essentially melting his way down through 200–350 m of ice. Through drilling and through a variety of geophysical methods (e.g., gravity anomalies), Paterson mapped out glacier thickness in the region. His results have subsequently been confirmed by ice radar and GPR studies in the 1990s. The deepest part of the Athabasca appears to lie in the vicinity of the Brewster snowcoach turnaround, in a bedrock overdeepening that gives a relatively flat ice surface and ice thicknesses of 300–350 m. Ice has thinned in the region by roughly 10% since the early 1960s.

In field experiments that are still cited today, Paterson tapped a number of his boreholes with hollow aluminum tubing and examined the deformation of the ice over the period 1959–1962 by repeat inclinometry of the boreholes. These were the first field observations of the internal deformation in glaciers, adding great insight to the freshly-emerging understanding about the way in which ice deforms and flows (ice rheology). A general view of the internal deformation profile in ice is given in Figure 3. Ice flows by “creep deformation,” in an ongoing downhill transfer of ice under the force of gravity. The flow of ice is not unlike that of a river, but motion takes place on a much longer time scale. The best way to understand it is to think of ice as a very high-viscosity fluid, something akin to the last remnants of a two-year old carton of blackstrap molasses that reside in the back of your fridge.

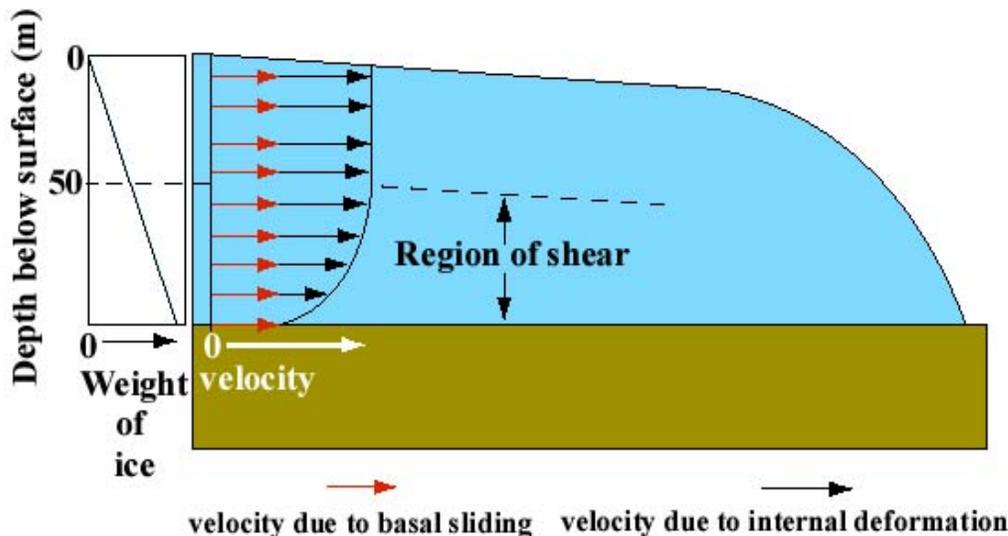


Figure 3. Schematic of the two primary flow mechanisms for ice: internal deformation and basal sliding. The latter can occur through decoupled sliding over the bed (“surfing”) or through deformation of subglacial sediments.

Ice acts as a non-linear fluid with relatively well-established behaviour. The creep/deformation velocity of ice, v_c is proportional to ice thickness to the fourth power and ice surface slope to the third power:

$$v_c \propto H^4 (\Delta h/\Delta x)^3 \quad (1)$$

where H is the ice thickness, h is the surface topography, and x is the downslope distance. $\Delta h/\Delta x$ is a simple expression for the surface slope: change in height over change in x .

As seen in Figure 3, glacier ice can also make its way downhill by two other mechanisms: (i) decoupled sliding of the glacier over the bed, v_s , and (ii) deformation or failure of the subglacial sediment, v_d . Both of these subglacial flow processes have been observed and inferred in many places, and these mechanisms can lead to high ice velocities in some ice masses – up to several km/yr. Both of these basal flow mechanisms require extremely high subglacial water pressures, approaching flotation. This too is a common occurrence in places where the glacier bed does not drain very readily (e.g., overdeepenings); surface meltwater travels through *crevasses* or moulins, reaching the base of the glacier in thin, temperate ice masses like the Athabasca. If a subglacial drainage network is well-established, water may flow quickly through (as was the case for a fortunate tourist on the Athabasca who flushed through a moulin and came out the front of the glacier, unscathed). In other sections of the glacier, water may pond and reach extremely high pressures. If these exceed flotation, the glacier can be lifted up to effectively surf over the bed.

Where a pervasive layer of soft sediment underlies the glacier, high water pressures can also dilate and weaken the sediment, causing viscous flow or plastic failure (shearing) of the sediment layer below the ice. These two subglacial flow modes, called basal sliding and basal sediment deformation, respectively, result in motion at the ice–bed interface that gives “plug flow” – a uniform vertical velocity profile. This basal motion will augment any internal deformation to give a net surface velocity that is a superposition of all three modes of flow: $v_{net} = v_c + v_s + v_d$.

Surface velocities can be measured through repeat surveying of flow markers or stakes that are drilled into the ice. Such measurements are made much easier these days by the advent of high-resolution differential GPS. Paterson made early estimates of ice velocity on the lower Athabasca through monitoring of his borehole array (old-fashioned surveying) over the period 1959–1962. Velocities of up to 75 m/yr are seen at the base of the lower icefall. Flow rates in the flatter bench of the lower glacier, in the vicinity of the snow-coach tours, are in the neighbourhood of 35 m/yr, and ice velocity at the terminus has been measured at 15 m/yr.

Interestingly, one of Paterson's aluminum poles remained intact until the early 1990s, and a back-trajectory of its displacement in the period 1960–1993 indicates an average velocity of 33 m/yr in the region. Paterson measured a flow rate of 34 m/yr in the same region in the early 1960s, implying little or no slowdown of the glacier. This is not consistent with the undeniable thinning that has occurred in the region, which would cause creep deformation velocities to decrease as per Eq. (1). This observation suggests that most motion in the area is likely due to basal sliding over the bed, which would not be expected to decrease with thinning.

3. Holocene Glacial History of the Region

There is not enough time to delve deeply into the environmental history of the Icefields region. A huge amount of work has been done in this area, and the Little Ice Age studies by a number of workers here arguably represent the most productive scientific research to come out of the region. Brian Luckman of the University of Western Ontario, London, is foremost in the group of

investigators that have worked here, and some useful information on his dendrochronological work is on display at the Icefields Visitor Centre (indeed, I would suggest skipping the glaciology at the Centre and going straight for the paleoenvironmental reconstructions). A number of review articles by Luckman and by Gerry Osborn (Geology, University of Calgary) give good overviews of the glacial and climatic history of this region.

To condense the story to four paragraphs, it is fair to say that little is known about the extent, thickness, or dynamical form of glacial cover in the Icefields area during the peak of the last Ice Age. Cordilleran ice would have filled the main valleys from ca. 30,000 to 15,000 years ago, flowing southwards along the Parkway corridor. Ice in the region was probably under topographic control still, rather than acting as part of a thick Cordilleran Ice Sheet that overwhelmed the peak in the region. Ice retreated from its glacial maximum position at some unknown time (probably driven by the Bölling-Alleröd, a global warm interval that is evident in Figure 4a, ca. 14 kyr BP).

There is evidence of a re-advance of local glacier systems at about 11 kyr BP. This is tentatively (but convincingly) correlated with the Younger Dryas cold interval seen in the Greenland ice core data of Figure 4a. The Younger Dryas is a period of brief return to full-glacial conditions which interrupted the termination of the last major glaciation. While not fully understood, it is now known to be global in scope. In the Canadian Rockies this period of glacial activity is known as the *Crowfoot advance*. The type section is named from the Crowfoot Glacier area, which we will view from the highway stop. Ice advanced from some retracted position ca. 11.3-10.7 kyr BP. Crowfoot-age terminal moraines are also found on the flanks of Wilcox Peak above the Columbia Icefields visitor centre, beyond the Athabasca Glacier's Little Ice Age limits.

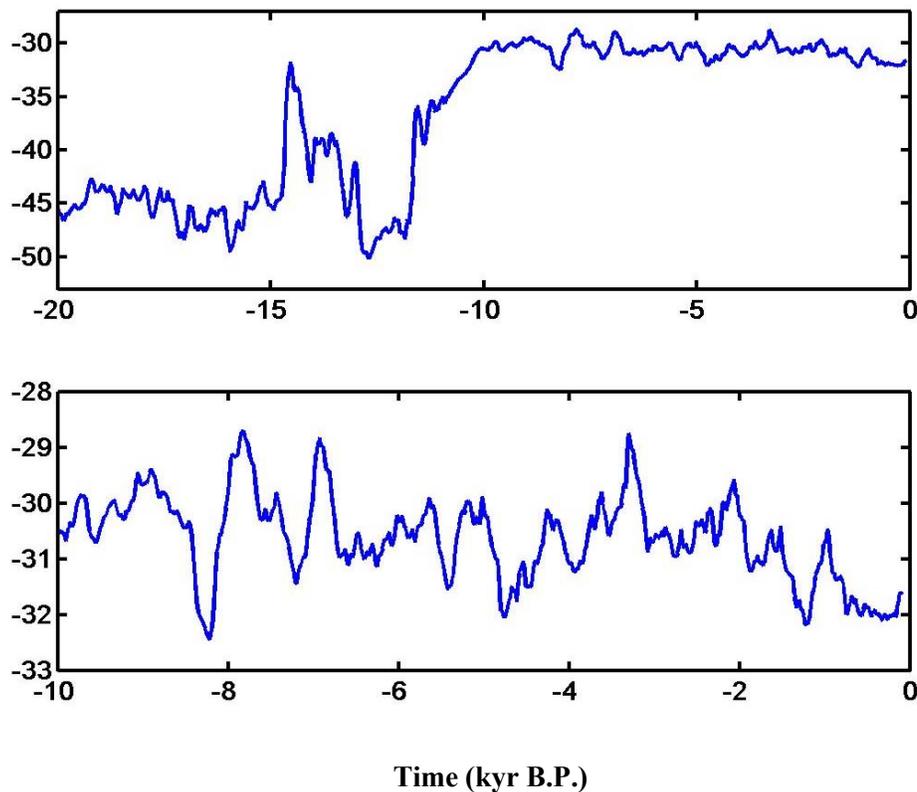


Figure 4. Air temperature reconstructions from the GISP2 ice core, central Greenland. (a) The last 20 kyr, depicting the Bölling-Alleröd warm interval, ca. 14 kyr BP, the Younger Dryas, and the comparatively stable Holocene period. (b) The last 10 kyr, illustrating the general cooling trend of the last 3000 years.

Lake records from throughout the region (e.g., Crowfoot Lake, Lake O'Hara, Bow Lake, Wilcox Pass) indicate that the valley bottoms were ice-free by ca. 10 kyr BP. Pollen records and organic layers in lake sediment and bog cores indicate that vegetation had reasserted itself at this time. Little is known about the next few millennia, but they appear to have been warm, dry, pleasant times in the region; pollen records indicate a higher treeline and the presence of several species that are absent today. Some remarkable bits of evidence for warmer conditions melted out of the Athabasca and Dome Glaciers in the 1980s: bits of gnarled wood that date to ca. 7900 and 6000 years ago. The wood fragments indicate mature pine forest well above the current glacier positions. It is not clear how retracted the outlet glaciers were, or to what extent the Columbia Icefields melted away during this time.

Beginning about 3000 years ago, the glaciers began a *neoglacial readvance* that culminated in the Little Ice Age. There were probably numerous advance-retreat phases during this time. The Little Ice Age maximum position, reached around 1840-1850 at the Peyto, Crowfoot, Saskatchewan, and Athabasca Glaciers, is the furthest extent of the glaciers in the entire Holocene period (the last 10 kyr, Fig. 4b). Detailed signatures of glacial activity prior to the Little Ice Age have therefore been destroyed to a large degree. The age of trees that have been over-ridden or tilted by glacier advance indicates a widespread return to cool, wet climate conditions beginning 3000 years ago.

The first phase of this renewed glacial activity, from 3000-2500 years ago, is known as the *Peyto* advance. It is recognized by tree stumps that have recently emerged in the forefield of the retreating Peyto and Saskatchewan Glaciers. These tree stumps are still in growth position and they make a remarkable sight, worth a hike in to these sites. Periods of modest advance and retreat likely followed, with some differences from site to site. The Little Ice Age period was seen throughout the region, known locally as the *Cavell* advance. The outermost terminal moraines that are readily visible in all outlet glaciers in the region date to the Cavell. In many places the transition from forest to rubble is wonderfully abrupt, as vegetation is just beginning to reclaim these glacial deposits.

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I have not thoroughly cited these notes, as they are intended to be casual, but the listing below offers a sample of the extensive literature available on the glaciological, climatological, glacier mass balance, and environmental history of the Columbia Icefields region.

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