Glaciers of North America–

GLACIERS OF CANADA

GLACIERS OF THE ARCTIC ISLANDS

GLACIERS OF THE HIGH ARCTIC ISLANDS

By ROY M. KOERNER

SATELLITE IMAGE ATLAS OF GLACIERS OF THE WORLD

Edited by RICHARD S. WILLIAMS, Jr., *and* JANE G. FERRIGNO

U.S. GEOLOGICAL SURVEY PROFESSIONAL PAPER 1386–J–1

Canadian High Arctic ice caps, both dynamic and stagnant, respond only very slowly to changes in climate. Ice cores contain records of environmental change during the past 100 thousand years, temperatures ranging from −20° C to +2.5° C. The last 150 years have been the warmest in the past 1,000 years, but only slight changes in area and volume (mass-balance) of ice caps are evident

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Abstract

If we exclude the massive ice sheets of Greenland and Antarctica, the Canadian High Arctic ice caps are among the oldest and largest in the world and contain long records of past environmental changes. Ice-core records indicate that these ice caps formed about 100 thousand years ago, beginning their growth as the last interglacial was coming to an end and the last glacial was beginning. However, the same ice-core records from these ice caps show that substantial climatic changes took place over their history. These changes encompass a range from a 20- degree Celsius cooling (compared to today) in the coldest parts of the last glacial to a period 10 thousand years ago that was 2.5-degrees Celsius warmer than today. Ice-core records show that the last 150 years, although still cooler than the warm period at the beginning of the present interglacial (the Holocene), are the warmest for at least the last 1,000 years. On the other hand, continuing mass-balance measurements show very few signs of warming or cooling in the eastern Canadian Arctic during the last 40 years. Satellite imagery reveals only a very slow pattern of change in the geometry of these ice caps in response to the climatic change of the last 150 years, including the very warm period from 1920 to 1960. The differences between the ablation and accumulation zones on these ice caps can be clearly seen on satellite imagery. However, because one of the modes of accumulation on these subpolar ice caps is by superimposed ice, the demarcation between the two zones cannot be clearly determined. Recent improvements in satellite sensors, however, are opening up a new approach to the mapping and study of glaciers hitherto unattainable by traditional methods.

History of Glacier Research

The Canadian High Arctic islands (Queen Elizabeth Islands, Nunavut, and Northwest Territories) (fig. 1) contain 151,057 km^2 (0.17 percent) of the world's, and 5 percent of the Northern Hemisphere's, glacierized area (Ommanney, 1970). Most of the ice masses are in the mountainous eastern part of the archipelago. The western islands are lower in elevation, and the only ice masses are on the higher parts of southwestern Melville Island.

Scientific work on the ice caps and glaciers in the Canadian High Arctic began in the 1950's with the work of the Canadian Defence Research Board (Hattersley-Smith, 1974). The Board began its work on the Ward Hunt Ice Shelf in northern Ellesmere Island in the early 1950's (Crary, 1960). In 1955, the same group established a glaciological research program on the northernmost ice caps in Canada, some of which lie north of the orbital coverage of Landsat 1–3 imagery. Toward the end of the 1950's, McGill University began work on Axel Heiberg Island (Müller, 1961, 1963a, b, 1966). The Polar Continental Shelf Project (in the Department of Energy, Mines and Resources (now Natural Resources Canada) of the Federal Government of Canada) began mapping and studying the ice cap on Meighen Island in 1959 (Arnold, 1965). In 1960, the Arctic Institute of North America prepared for its first year-round High Arctic expedition, and from 1961 to 1963, the Devon Island

Mamuscript approved for publication, 7 March 2002.

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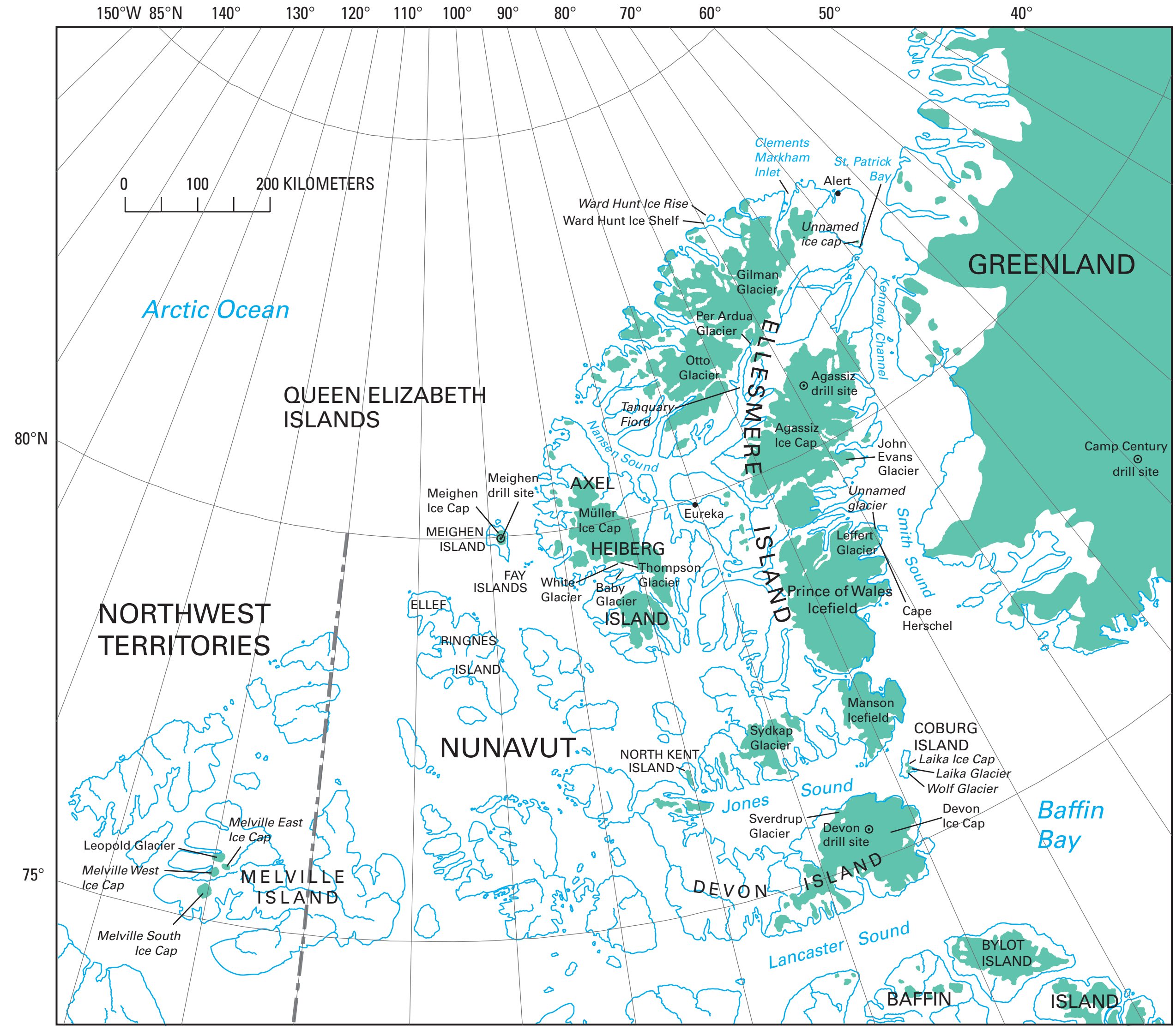


Figure 1.*—Canadian High Arctic islands. Green areas represent glacier-covered regions. The black dots within black concentric circles are ice-drilling sites.*

expedition included major investigations of the Devon Ice Cap and Sverdrup Glacier (Müller and Keeler, 1969; Hyndman, 1965; Koerner, 1970a, b).

In the 1960's and 1970's, various research programs were conducted on Coburg Island and in the following locations on Ellesmere Island: an ice piedmont near Cape Herschel (Müller and others, 1978), Leffert Glacier (Gerald Holdsworth, oral commun., 1984), Gilman Glacier (Arnold, 1968), an ice cap west of Tanquary Fiord (Hattersley-Smith and others, 1975), Ward Hunt Ice Shelf (Hattersley-Smith, 1963; Hattersley-Smith and Serson, 1970), and a small ice cap near St. Patrick Bay (Hattersley-Smith and Serson, 1973).

New programs were started in the 1980's on the Ward Hunt Ice Shelf (Jeffries and Serson, 1983) and the St. Patrick Bay ice cap (Serreze and Bradley, 1983). The Polar Continental Shelf Project (PCSP) conducted a climate-change and ice-physics program between 1964 and 1987, when surface-to-bedrock cores were drilled on Meighen, Devon, and Agassiz Ice Caps (Paterson, 1968, 1969; Koerner, 1968; Paterson and others, 1977). The program and staff transferred to the Geological Survey of Canada (GSC) in 1987. Since then, further cores have been drilled from the top of Agassiz and Devon Ice Caps and two from Penny Ice Cap on Baffin Island (Fisher and others, 1983; Fisher and others, 1998). The same program continues to monitor the mass balance on the Meighen, Devon, and Agassiz Ice Caps and the most southerly ice cap on Melville Island (Koerner and Lundgaard, 1995). Work on White and Baby Glaciers on Axel Heiberg Island has continued through the 1980's

(Arnold, 1981) and to the present day (Cogley and others, 1995, 1996). A new, and continuing, program on John Evans Glacier in central Ellesmere Island led by the University of Alberta began in 1994 (Skidmore and Sharp, 1999).

Field reports from all these studies can be found in the annual reports of glacier research conducted in Canada and published in the following journals: Canadian Geophysical Bulletins between 1961 and 1984, Arctic from 1961 to 1965, and Ice (a publication of the International Glaciological Society) to the present day.

The Ice Caps and Glaciers

We may differentiate between the larger, dynamic glaciers and the smaller, stagnant ones. The ice shelves will be discussed in the following section on "Ellesmere Island Ice Shelves and Ice Islands" by Martin O. Jeffries.

Dynamic Ice Caps and Outlet Glaciers

It was first thought, from ice-core analysis, that the larger ice caps, such as those on Devon, Axel Heiberg, and Ellesmere Islands (fig.1), predated the last interglacial period (Koerner and others, 1987). However, further work on the same and newer data have suggested that no ice caps survived that period and that they began their growth in the very early stages of the last glacial period.

The smaller ice caps on Devon Island and southern Ellesmere Island ((D) in figs. 2, 3, 4) may be much younger. These do not reach such high elevations, have smaller accumulation areas, end well above sea level, and have very few channeled outlet glaciers. Whereas some of these ice caps may be more than 10 thousand years (kilo-annum (ka)) old, it is more likely that they began their growth during the second half of the Holocene Epoch, at less than 4.5 ka B.P. The whole question of glacial history and ice-core research will be discussed more fully later.

Topography

The surface topography of almost all the large ice caps in the Canadian High Arctic is very strongly controlled by the subglacier topography. The Prince of Wales Icefield and part of the Manson Icefield in southeastern Ellesmere Island (shown in fig. 5) are good examples. Valley glaciers on the east side descend to sea level between mountainous ridges that are often ice covered. Isachsen Glacier is a good example of this (fig. 5). Some of these glaciers are a few hundred meters thick, and the bedrock in their lower reaches is below sea level (Cadogan Glacier, figs. 5 and 6). On the west side, the ice is not as thick and often forms broad, slow-moving tongues that, in places, are almost as broad as they are long (see arrows, fig. 5). The same pattern of lobate glaciers on the one side and valley outlet glaciers on the other can be found on southern Ellesmere Island (fig. 4) and along large stretches of eastern Ellesmere Island from Baffin Bay to Kennedy Channel (fig. 1). On Axel Heiberg Island, the broad, lobate ice is on the east side, and the outlet valley-type glaciers are on the west (fig. 7). However, asymmetry is not characteristic of the most northerly of the large ice caps on Ellesmere Island (for example, Agassiz Ice Cap, fig. 8), where valley outlet glaciers emerge from all sides of the ice caps. Among all these ice caps, only that on Devon Island shows anything like the symmetry of a true ice cap, and has a nunatak-free central region. Even then, this is strictly true only of its western part (fig. 2).

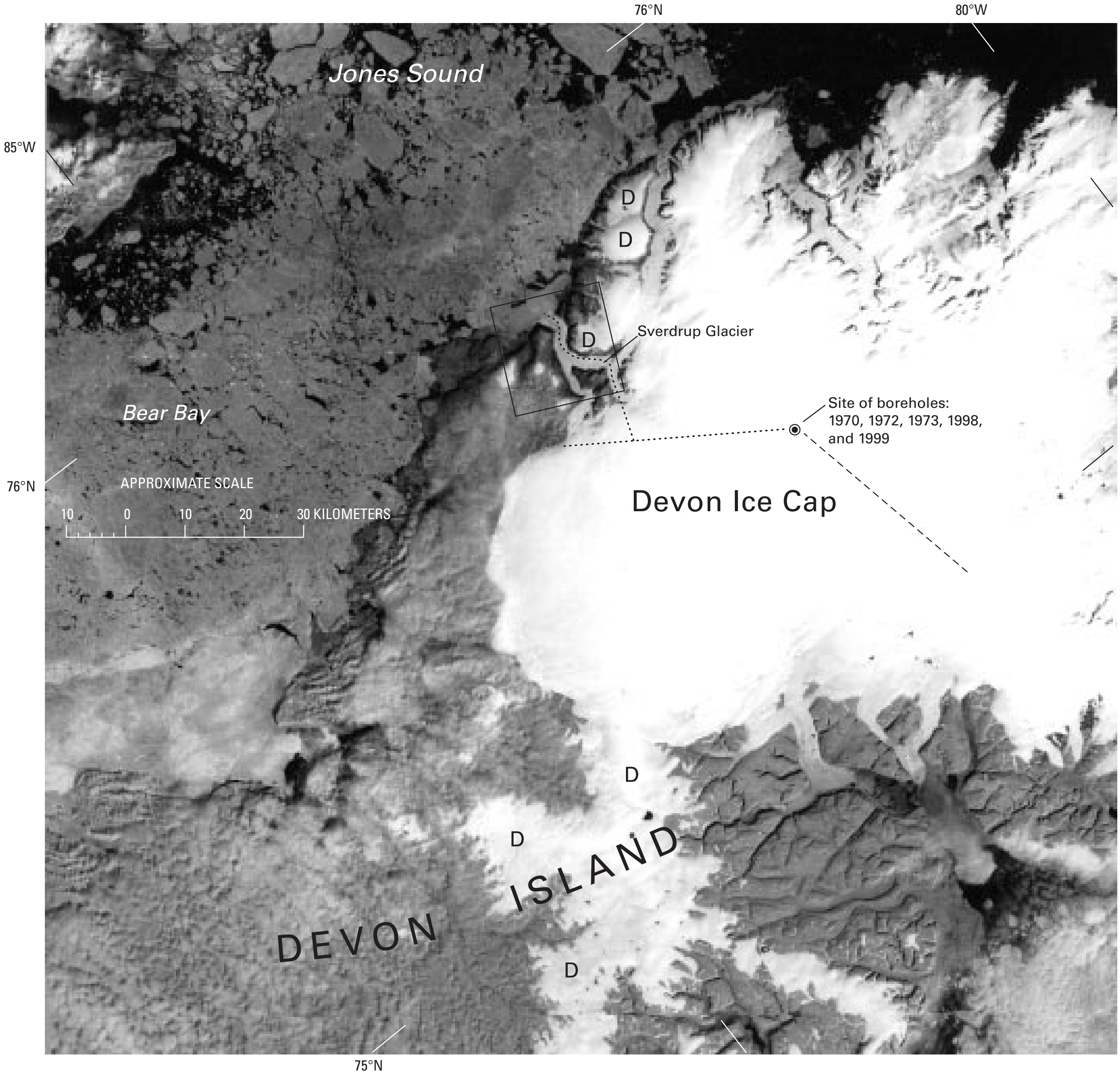


Figure 2.*—Landsat 3 MSS image showing the western part of Devon Ice Cap. The dotted lines represent the location of the stake network measured annually for mass balance; these lines (northwest side profile) and the dashed line (south side profile) are the locations of the ice-thickness profiles for Devon Ice Cap shown in figure 6. The location of the five borehole sites is shown with a concentric circle and dot symbol; the boreholes include a 230-m deep core drilled in 1970, two 299-m surface-to-bedrock cores drilled in 1972 and 1973, and nearby sites of a 305-m borehole-ice-core drilled in 1998, and one drilled in 1999. The ice caps marked with the letter (D) are dynamic and probably Holocene in origin, unlike the larger, and older, dynamic main ice cap, which has ice formed at least by 60 ka at its base. The outlined area is shown in figure 3. The Landsat image (30523–17365, band 7; 10 August 1979; Path 44, Row 6) is from the Canada Centre for Remote Sensing, Ottawa, Ontario, Canada.*

Figure 3.—*Vertical aerial photograph taken in 1959 of Sverdrup Glacier on the northwest side of Devon Ice Cap (see outline in fig. 2). The letter (m) refers to supraglacier meltwater streams; the northernmost (m) is a moulin (where meltwater streams plunge into, and sometimes under, the glacier). The dotted line represents the location of the leveling profile measured for surface-height changes and is referred to in the text. Ice cap (D) is a dynamic ice cap that is much younger than the main Devon Ice Cap to its south and east. Ice caps (A), (B), (E), and (F) are all stagnant and probably less than 1 ka. Ice cap (C) is less than 200 or 300 years old, as are the three small ice caps, indicated by arrows, all of which have now melted and appear as bare ground in figure 2. All these ice caps are discussed in more detail in the text.*

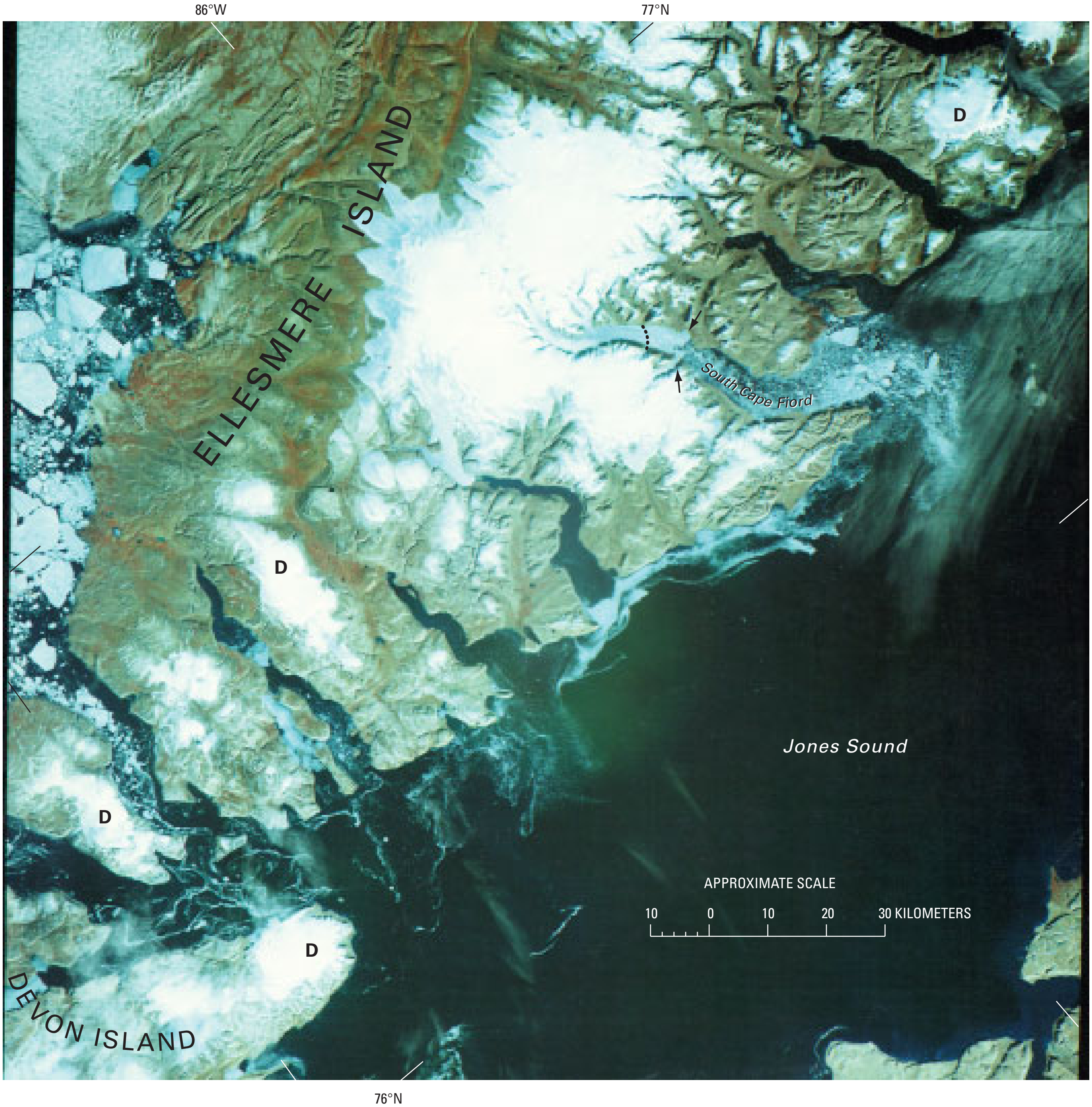


Figure 4.—*Landsat 1 MSS false-color composite image of southwestern Ellesmere Island and part of northwestern Devon Island. Ice caps marked (D) are examples of the small dynamic types discussed in the text. The arrows mark the 1957 position of a glacier in South Cape Fiord that retreated to the position marked by a dotted line sometime between 1957 and 1974. The ice between these two positions is sea ice or icebergs. The Landsat image (1760–18015, bands 4, 5, and 7; 22 August 1974; Path 48, Row 5) is from the Canada Centre for Remote Sensing, Ottawa, Ontario, Canada.*

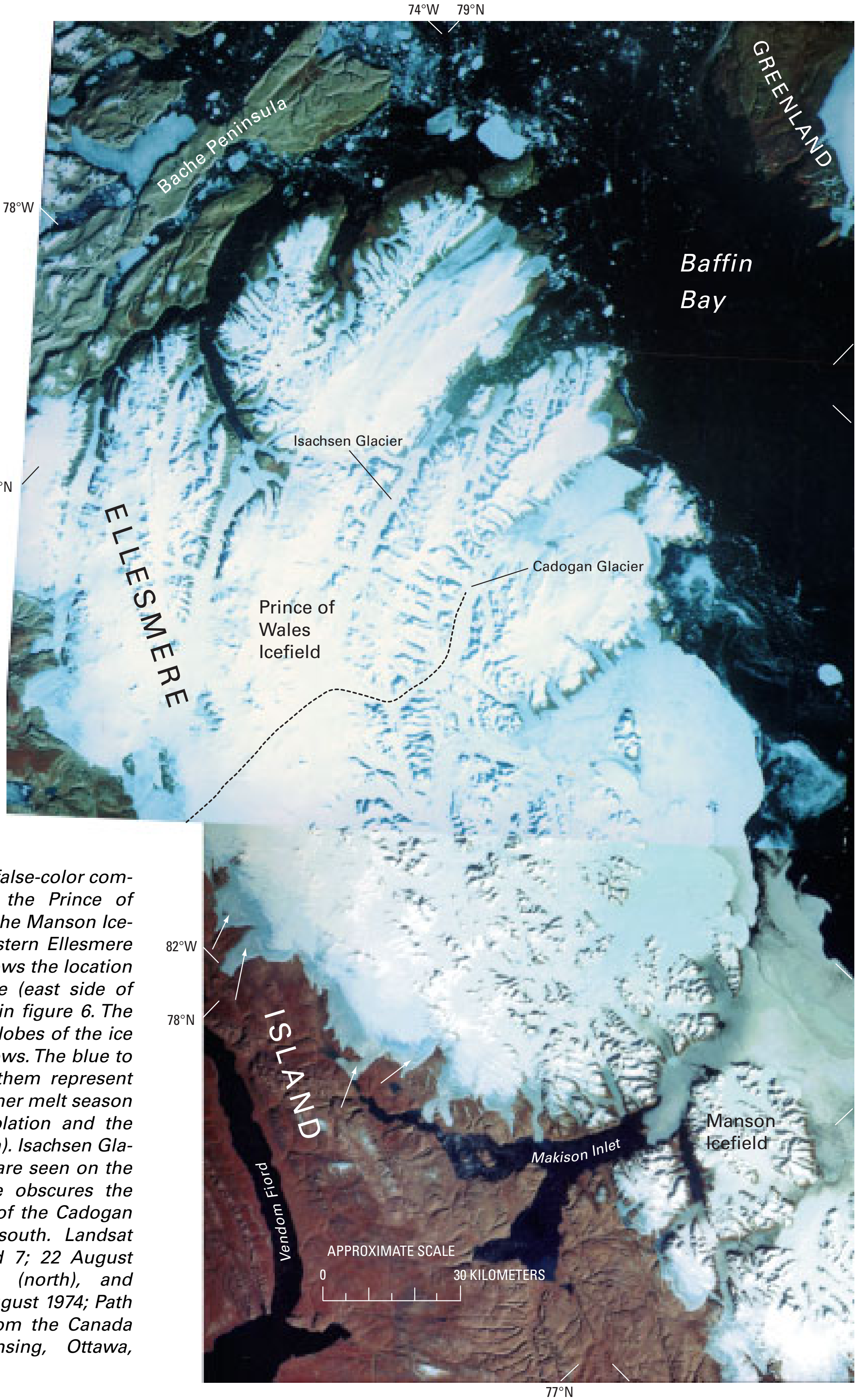


Figure 5.—*Landsat 1 MSS false-color composite image mosaic of the Prince of Wales Icefield and part of the Manson Icefield, central and southeastern Ellesmere Island. The dashed line shows the location of the ice-thickness profile (east side of central Ellesmere) shown in figure 6. The broad, slowmoving outlet lobes of the ice cap are indicated with arrows. The blue to white color tones above them represent various stages in the summer melt season (see text in "Summer Ablation and the Glacier Landscape" section). Isachsen Glacier and Cadogan Glacier are seen on the mosaic. Land-fast sea ice obscures the coastline at the terminus of the Cadogan Glacier and also to its south. Landsat images (1760–18010, band 7; 22 August 1974; Path 48, Row 3 (north), and 1758–17500, band 7; 20 August 1974; Path 46, Row 4 (south)) are from the Canada Centre for Remote Sensing, Ottawa, Ontario, Canada.*

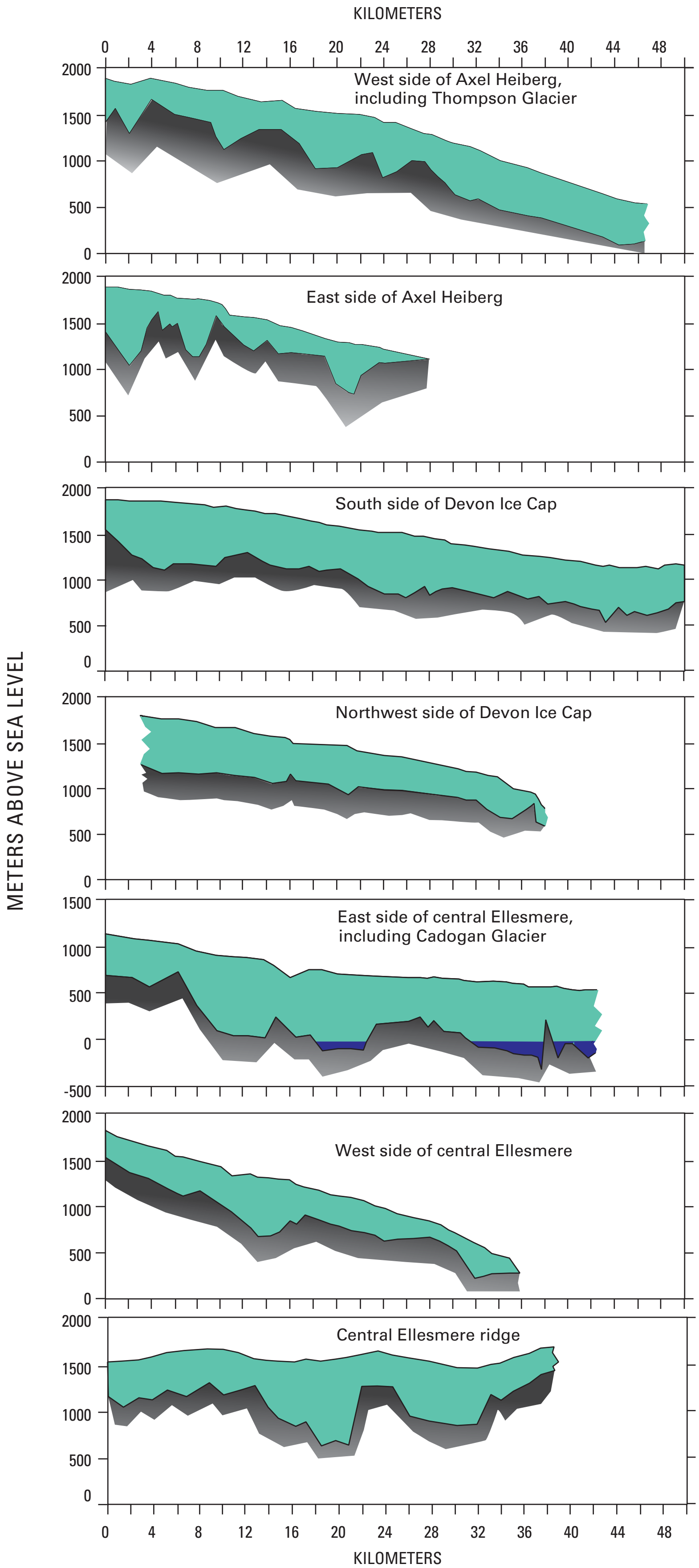
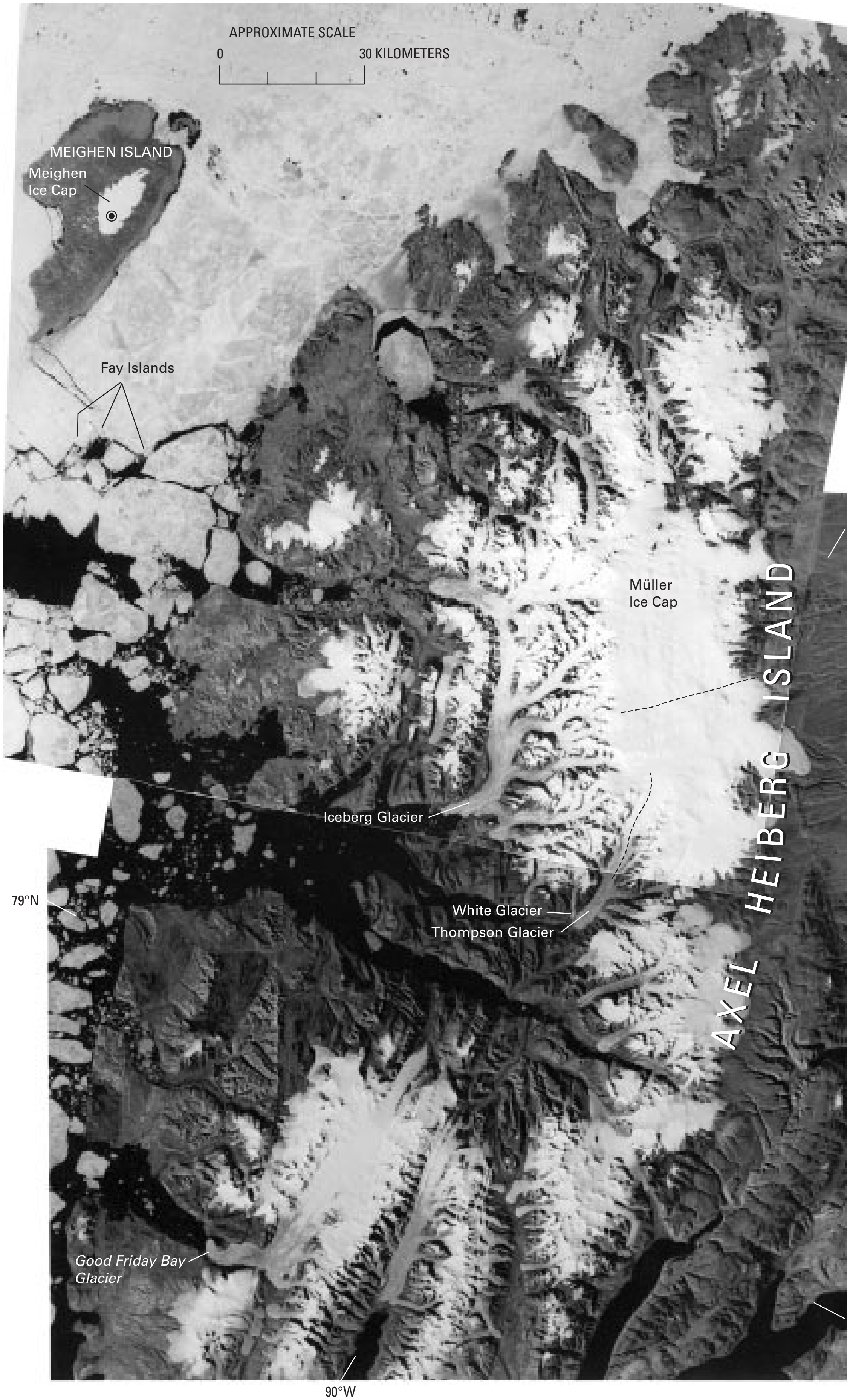


Figure 6.—*Ice-thickness profiles measured by radio-echosounding in April-May 1976 (Koerner, 1977b). The green represents ice, and the dark blue represents ice below sea level on Cadogan Glacier. See figures 2, 5, and 7 for the locations of five of the profiles. Reproduced by permission of the Canadian Journal of Earth Sciences.*

Figure 7.—*(opposite page) Mosaic of two Landsat MSS images of part of Axel Heiberg Island and Meighen Island. A concentric circle and dot symbol marks the Meighen glacier-borehole site, where a 121-m surface-to-bedrock core was drilled by the Polar Continental Shelf Project in 1965 (see also fig. 9*A*). The dashed lines on the Müller Ice Cap on Axel Heiberg Island represent ice-thickness profiles (east side and west side) shown in figure 6. Iceberg Glacier (see also fig. 13), White Glacier, which has been the site of glacier research since 1959, and Good Friday Bay Glacier, which advanced (surged?) sometime between 1952 and 1959, are all seen in the mosaic. The area northeast of the Fay Islands, between Meighen and Axel Heiberg Islands, is a region where sea ice persists throughout many summers. An incipient ice shelf formed here in the 1950's and 1960's and reached a few meters in thickness before it broke up in the middle 1970's. The Landsat images (1158–18455, band 7; 24 August 1977; Path 69, Row 1 (north) and 20950–18542, band 7; 29 August 1977; Path 66, Row 2 (south) are from the Canada Centre for Remote Sensing, Ottawa, Ontario, Canada.*


APPROXIMATE SCALE
0
30 KILOMETERS
MEIGHEN ISLAND
Meighen
Ice Cap
Fay Islands
Müller
Ice Cap
AXEL HEIBERG ISLAND
Iceberg Glacier
White Glacier
Thompson Glacier
79°N
Good Friday Bay
Glacier
90°W

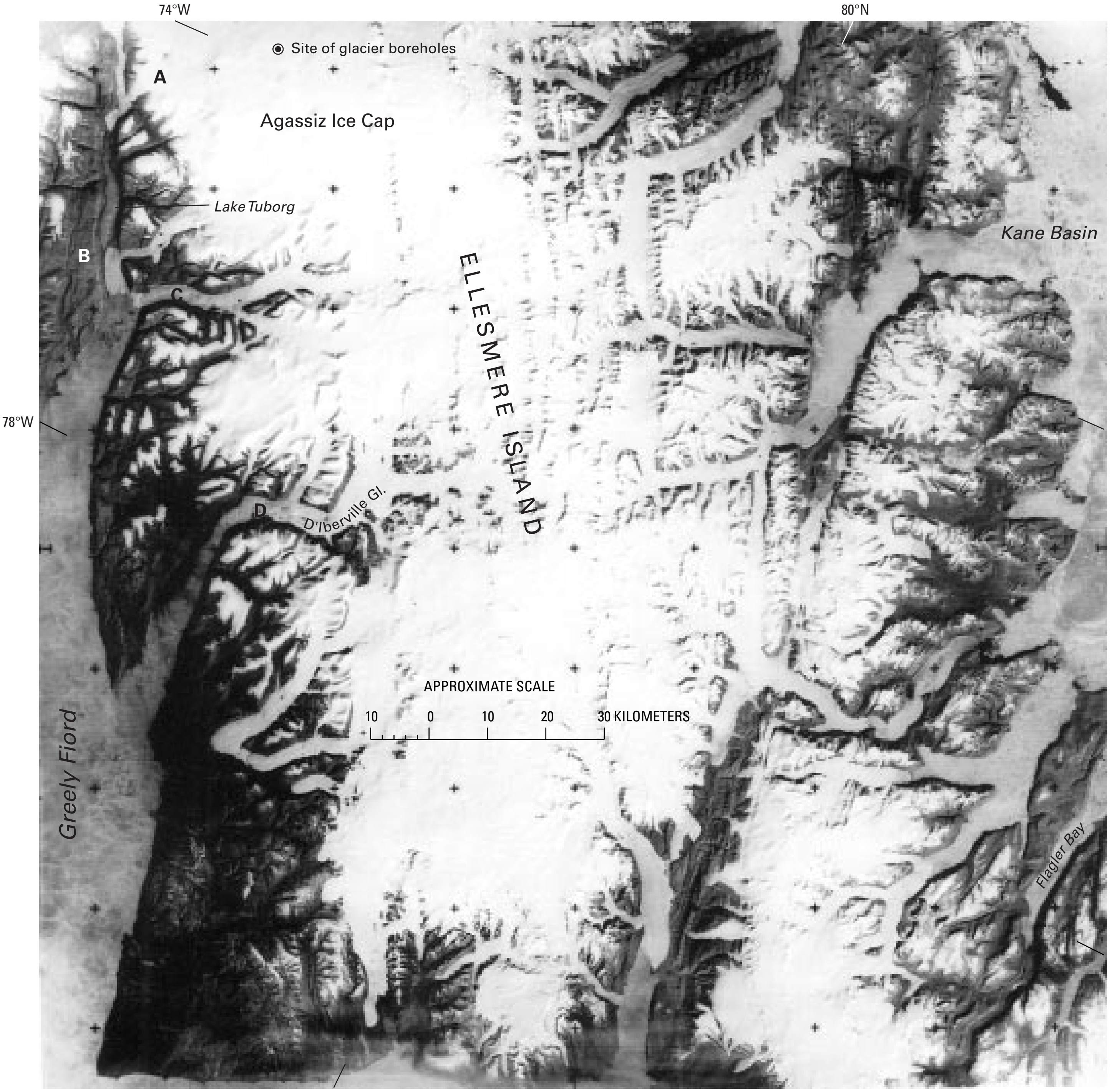


Figure 8.—*Landsat 2 return-beam vidicon (RBV) image of the Victoria and Albert Mountains, Ellesmere Island. A symbol indicates the location of six surface-to-bedrock boreholes drilled into the Agassiz Ice Cap by the Polar Continental Shelf Project: a 337-m core in 1977, a 137-m core in 1979, two 127-m cores in 1984 and 1987, and two 130-m cores in 1994. Glaciers (A), (B), (C), and (D) have either heavily crevassed or hummocky surfaces, and, therefore, very high summer ice-melt rates (see text in "Summer Ablation and the Glacier Landscape" section). Glacier (B) has advanced across the head of Greely Fiord and thereby created an ice-locked lake (Lake Tuborg). Dating of the basal waters of this lake indicate that the glacier advanced at about 3.5 ka (Hattersley-Smith and others, 1970). The Landsat image (2550–18480, band 3; 25 July 1976; Path 58, Row 1) is from the EROS Data Center, Sioux Falls, S. Dak.*

Velocity

Glacier motion has been measured by various methods. Arnold (1965, 1968, 1981) used traditional field surveying and photogrammetric techniques on Meighen Ice Cap, Gilman Glacier (Ellesmere Island), and White Glacier (Axel Heiberg Island), whereas Doake and others (1976) used radio-echosounding techniques at the top of Devon Ice Cap. The results of these and other measurements are shown in table 1. New approaches to measuring velocity, such as the Global Positioning System (GPS) of satellites and satellite radar interferometric (InSAR) techniques are quickly expanding the present slim velocity data base.

Velocities in this area are generally of the order of 10–50 m a^{-1}. However, velocity in summer can be as much as twice as high as in winter (table 1). This can be attributed to the presence of meltwater at the glacier bed in summer (Iken, 1974) and indicates that parts of some glaciers are at the melting point at their beds. Calculations of basal temperatures using known ice thicknesses and 10–15-m englacier temperatures (Müller, 1976; Paterson, 1994), as well as measured basal temperatures (Blatter, 1987), confirm the unfrozen basal condition of many High Arctic glaciers.

One consequence of relatively low glacier velocities (strictly speaking, low strain rates) is that, compared to glaciers in other areas of the world, many of these glaciers are not very crevassed. Sverdrup Glacier, although it is the major outlet for ice from the northwest side of the Devon Ice Cap, is a good example (fig. 3). Another example is the crevasse-free ice cap above, and to the east of, Sverdrup Glacier (fig. 3, (D)).

A few glaciers in the High Arctic, however, maintain high velocities, and probably comparably high strain rates, in their lower reaches and are more crevassed. Good examples are the glaciers draining the west side of Agassiz Ice Cap (fig. 8, (A), (C), and (D)). Using aerial photographs, Holdsworth (1977) calculated tongue velocities of more than 400 m a^{-1} on one of them (D'Iberville Glacier; (D), fig. 8; table 1). These glaciers act as the major outlets for Agassiz Ice Cap and drain large catchment areas.

Calving of the glaciers in the High Arctic islands has not received much attention. Although it certainly does not rule it out, very little photographic evidence exists of calving. For example, we can see very few icebergs adjacent to Sverdrup Glacier (fig. 3). Most of the icebergs that enter

TABLE 1.—*Velocity measurements (winter-summer) of selected Canadian High Arctic glaciers on Axel Heiberg Island, northern Ellesmere Island, central Ellesmere Island, and Devon Island*

[Note that Meighen Ice Cap and Melville ice caps are stagnant. See figure 1 for locations. Abbreviations: I., island; N., northern; C., central; do, ditto; Accum., accumulation area; Eq. line, equilibrium line; Abl., ablation area; <, less than; leaders (- -), not determined. Data archived at the Geological Survey of Canada, 601 Booth Street, Ottawa, Ontario K1A 0E8, Canada]

Location	Glacier	Velocity m a^{-1} winter	Velocity m a^{-1} summer	Location on glacier	Reference
Axel Heiberg I.	White	<13.0	- -	Accum.	Müller, 1963b
.....do...........	.....do..........	24.5	- -	Eq. line	Do.
.....do...........	.....do..........	21.4	44.0	Abl.	Do.
.....do...........	.....do..........	10.4	- -	Tongue	Do.
.....do...........	Thompson	47.0	51.0	Abl.	Do.
N. Ellesmere I.	Gilman	14.8	0	Accum.	Arnold, 1968
.....do...........	.....do..........	22.9	22.2	Eq. line	Do.
.....do...........	.....do..........	19.6	- -	Abl.	Do.
C. Ellesmere I.	D'Iberville	457.0	500.0	Tongue	Holdsworth, 1977
.....do...........	Leffert	40.0	- -	do	Gerald Holdsworth, oral commun., 1984
Devon Island	Devon Ice Cap	2.4	- -	Top of ice cap	Doake and others, 1976
.....do...........	Sverdrup	36.4	65.0	Abl.	Cress and Wyness, 1961

the shipping lanes south of here off the coast of Newfoundland are from the west coast of Greenland. New studies are needed to determine the importance of calving in this area.

Thickness

The ice thickness of Canadian High Arctic ice caps has been measured by radio-echosounding techniques (Hattersley-Smith and others, 1969; Paterson and Koerner, 1974; Koerner, 1977b; Oswald, 1975; Narod and Clarke, 1983). Unpublished work was done as well by the Scott Polar Research Institute at the top of Devon Ice Cap in 1974; the work is available as maps at the GSC. Some of the results (from Koerner, 1977b) are shown in figure 6, where the asymmetry in thickness of some of the ice caps can be seen. This is largely attributable to high snow-accumulation rates on the slopes facing Baffin Bay, where relatively thick ice sheets have built up. Although not shown in Koerner (1977b), areas several kilometers square were sounded at the tops of Agassiz Ice Cap and ice caps on Axel Heiberg and central Ellesmere Islands. In general, the data show that at the tops of most of the ice caps the ice is about 100–300 m thick. Downslope, the thickness may reach 1,000 m in channeled areas. However, the most common thickness is approximately 500 m, even in valley glaciers. Some of the glaciers, in their lower reaches, flow over bedrock that is below sea level (for example, Cadogan Glacier, figs. 5 and 6). In 1995, traverses were flown over many of the Canadian ice caps by a National Aeronautics and Space Administration (NASA) P–3 aircraft using ice-penetrating radar (160 MHz) and geodetic airborne laser altimetry. In Spring 2000, J.A. Dowdeswell (oral commun.) mapped glacier thicknesses by using airborne radio-echosounding surveys of the Queen Elizabeth Islands' ice caps. In May 2000, NASA, using a chartered Canadian Twin Otter aircraft, resurveyed ice caps on Ellesmere Island, Axel Heiberg Island, Devon Island, and Baffin Island, Nunavut. When published, all of these data will greatly extend our knowledge of ice-cap volumes in the Queen Elizabeth Islands. The NASA surveys also constitute a valuable baseline for monitoring the changing geometry of ice caps in an era of predicted global warming. The NASA ice-penetrating-radar data are available on the NASA website at [http://tornado.rsl.ukans.edu/1995.htm].

Stagnant Ice Caps

The smallest ice caps have no outlet glaciers. Traditional surveying and also ice-fabric analysis indicate that they are stagnant. Examples of these ice caps are shown in figure 3 ((A), (B), (E), and (F)). The first work on these ice caps was done on Meighen Ice Cap (fig. 9) by Arnold (1965), Koerner (1968), and Paterson (1969). Surveying of markers, over a 2-year period, by Arnold (1965) showed no evidence for movement. Ice-fabric analysis of an ice core from the same ice cap suggested that it has been stagnant throughout its history (Koerner, 1968). Similar ice-fabric work at the edge of a smaller ice cap on Devon Island (fig. 3, (B); R.M. Koerner, unpublished data, 1984) suggests that it too has always been stagnant. It, therefore, seems reasonable to consider that ice caps of similar and smaller size are, and always have been, stagnant. Their past and present stagnant nature puts limits on the maximum dimensions that they may have reached during their lifetime. The oxygen isotope ($\delta^{18}O$) values of the basal ice in Meighen Ice Cap are not sufficiently negative to suggest that the ice is of Pleistocene age (Koerner and Paterson, 1974). Furthermore, it is unlikely that any of these ice caps would have been large enough to survive the

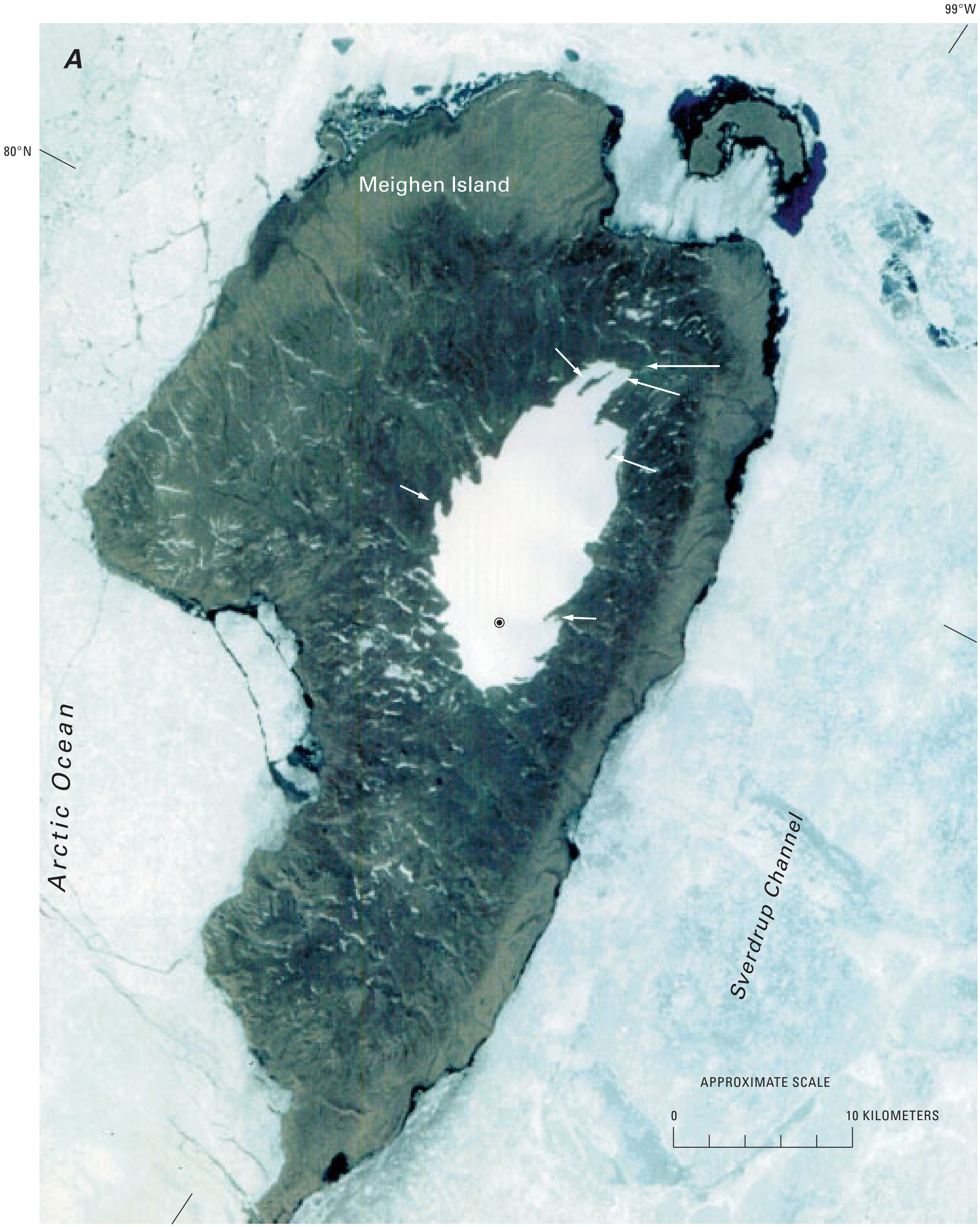
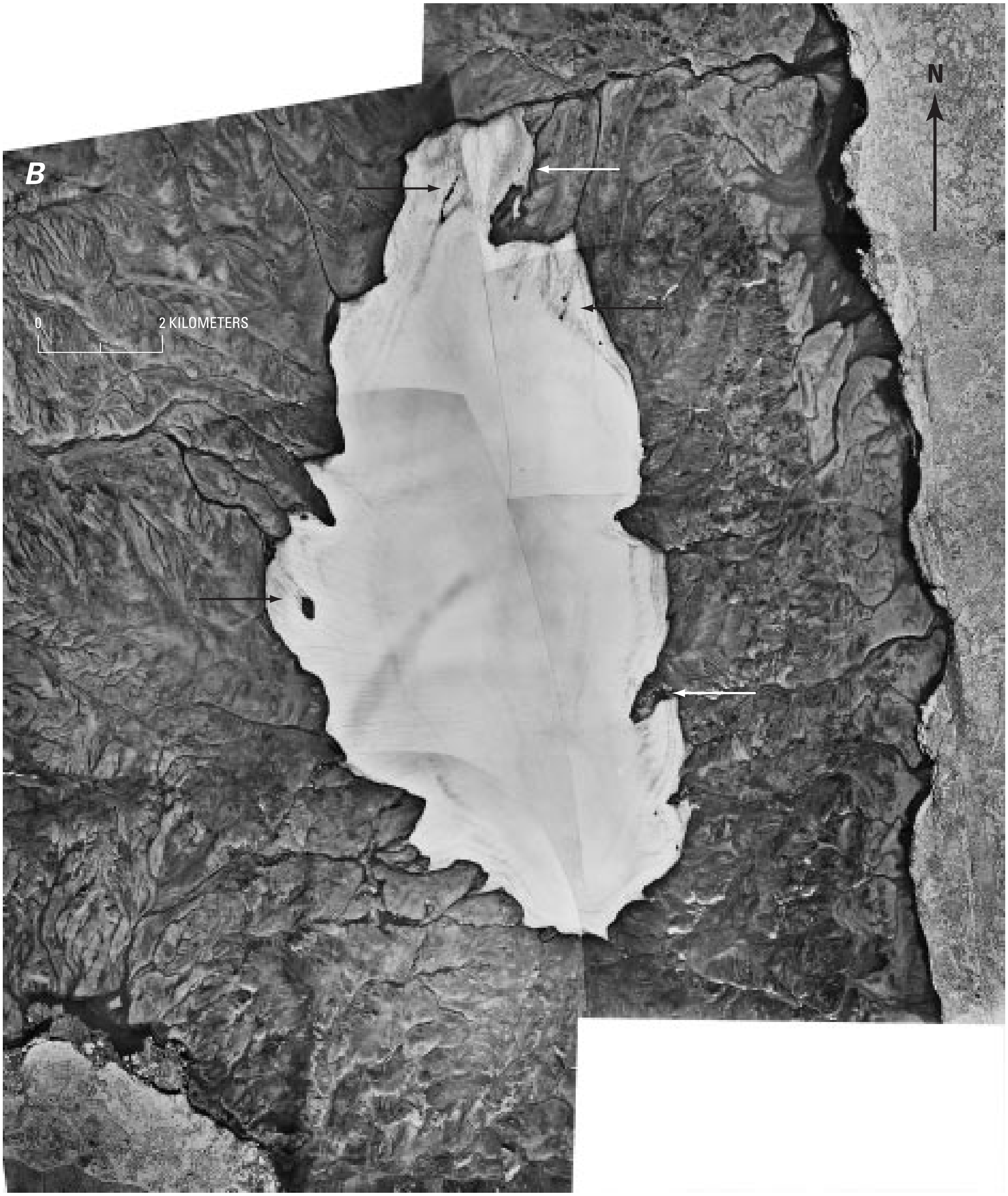


Figure 9.—*Landsat image from 1977 and an aerial photographic mosaic from 1959 showing changes in the Meighen Island Ice Cap.* **A,** *Part of an enlarged Landsat 1 MSS false-color composite image. The site of a 121-m surface-to-bedrock drill hole drilled in 1965 is shown with the concentric circle and dot symbol. Compare the margins and nunataks marked by arrows with the same locations in figure 9*B. *Landsat image (11858–18455, bands 4, 5, and 7; 24 August 1977; Path 69, Row 1) is from the Canada Centre for Remote Sensing, Ottawa, Ontario, Canada.* **B,** *see following page.*



warm period between about 9 ka and 4.5 ka (Koerner and Fisher, 1990). They are, therefore, most probably less than 4.5 ka.

Ice caps smaller than Meighen Ice Cap (for example., fig. 3, ((A), (B), (E), and (F)) must be much younger than 4.5 ka, probably less than 1 ka. The smallest (e.g., fig. 3, (C) and the three ice caps indicated by arrows) may be only 200 to 300 years old; perhaps they began their growth about 300 years ago or during the “Little Ice Age” about 200 years ago. They are mostly less than 1 km in diameter and are only 10–20 m thick. The three small ice caps marked by arrows in figure 3 had melted completely by the end of 1962.

Figure 9*B—*Vertical aerial photographic mosaic taken on 5 August 1959. Arrows show areas of glacier recession in later years. Note the banding on the ice surface along the east margin that is caused by annual cycles of accumulation in a cold period, which have been later crosscut by ablation as the climate became warmer.*

The surface topography of stagnant ice caps depends on the geometry of the underlying bedrock topography, their age, and their mass-balance history. The surface profile of a stagnant ice cap is, in part, an integration of all the annual mass-balance gradients since its inception (where the gradient is the change in mass balance with elevation). For example, on Meighen Ice Cap (figs. 7, 9), the north-facing slopes are much less steep than those facing south. Mass-balance measurements (IAHS/UNESCO, 1985) show that the north slopes of Meighen Ice Cap are losing ice at a much greater rate than the rest of the ice cap (Koerner and Lundgaard, 1995). These slopes are unlikely to have been able to withstand long periods of warm climate (such as the one ending at about 1 ka) and are probably much younger than the southern part of the ice cap.

The annual cycle of ablation and accumulation on these ice caps exposes the sedimentary ("annual growth") layers at the surface. These bands are well depicted in figure 3 ((A), (B), and (C)), and above parts of the east margin of Meighen Ice Cap (fig. 9*A*). The light-colored bands are composed of fine-grained ice. This ice forms during cold summers, when less melting and less percolation take place than usual. An incompletely soaked, low-density, bubbly layer is then formed. Conversely, the dark bands consist of coarse-grained ice formed during warmer summers, when the surface snow layer becomes completely saturated before refreezing. An almost bubble-free layer is then formed (Koerner, 1970a). Originally, each of these surface layers may cover large areas of the ice cap. Subsequently, melting toward the outer parts of the ice cap removes part of the layers, while accumulation further in buries the rest. This type of banding should not be confused with foliation bands found on valley glaciers (Hambrey and Müller, 1978) (for example, Sverdrup Glacier, fig. 3). Foliation is a tectonic feature formed by dynamic processes in moving ice.

Mass Balance

The mass balance of a glacier is the difference between the amount of precipitation accumulating on the glacier and the amount that leaves as melt or ice calving throughout the year. The glaciers where measurements have been made between 1957 and the present (table 2) are shown in figure 1. Some of these records are more than 35 years old and are among the longest high-latitude records in the world. Mass-balance techniques are described by Østrem and Brugman (1991). However, some of these techniques do not apply to subpolar glaciers, where some of the summer snowmelt refreezes in the firn. Here, we describe the methods used by the Geological Survey of Canada on the Queen Elizabeth Islands' glaciers. The reader is referred to Geografiska Annaler (1999) for an up-to-date reference on modern methods of measurement and modeling and to Jania and Hagen (1996) for a review (with summary annual data) of Arctic mass balance covering Alaska, Canada, Greenland, Iceland, Svalbard, northern Scandinavia, and the Russian Arctic.

Winter Balance

To obtain winter balance, poles are drilled into the ice and firn in both the accumulation and ablation areas; these are used as reference points. Accumulation comes as snow throughout the winter. Each spring, snow depths are taken by using a depth probe to the easily recognized, end-of-melt-season, firn layer underneath. These depths, along with density measurements, give the winter balance. The length of the reference poles is measured at the same time each year.

TABLE 2.—*Glaciers in the High Arctic where mass-balance measurements have been carried out*

[Abbreviations: No., number; NW, northwest side; For explanation of italicized names, see text footnote 2 on p. J136]

No.	Glacier	Latitude North	Longitude West	Period observed	Years missed
	Ellesmere Island[1]				
1	Ward Hunt Ice Shelf	83°7'	73°30'	1960–85	1977, '78, '79
2	*Ward Hunt Ice Rise*	83°7'	74°10'	1958–85	1977, '78, '79
3	Gilman Glacier	82°6'	70°37'	1957–69	
4	*Unnamed ice cap*	81°57'	64°12'	1966–76	1968, '69
5	Per Ardua Glacier	81°31'	76°27'	1964–71	
6	Agassiz Ice Cap	80°	75°	1977–present	
7	Leffert Glacier	78°41'	75°01'	1979–80	
8	*Unnamed glacier*	78°39'	74°55'	1979–80	
	Coburg Island[1]				
9	*Laika Glacier*	75°53'	79°5'	1973–80	1976, '77, '78
10	*Laika Ice Cap*	75°55'	79°9'	1974–80	1976, '77, '78
11	*Wolf Glacier*	75°54'	79°12'	1979–80	
	Axel Heiberg Island[1]				
12	White Glacier	79°26'	90°40'	1959–present	1980, '81, '82, '83
13	Baby Glacier	79°26'	90°58'	1959– present	1973, 1978–88
	Meighen Island[1]				
14	Meighen Ice Cap	79°57'	99°08'	1959–present	1972, '79
	Melville Island[1]				
15	*Melville South Ice Cap*	75°25'	115°10'	1963–present	1968, '72, '75–'79
16	*Melville West Ice Cap*	75°38'	114°45'	1963–73	1968, '72
17	*Melville East Ice Cap*	75°39'	114°29'	1963–73	1968, '72
18	Leopold Glacier	75°49'	114°49'	1963–73	1968, '72
	Devon Island[1]				
19	Devon Ice Cap (NW)	75°20'	82°30'	1961–present	1968
	Baffin Island[2]				
20	Lewis Glacier[3]	70°26'	74°46'	1963–65	
21	Barnes Ice Cap	70°10'	74°46'	1963–84	
22	Decade Glacier	69°38'	69°50'	1965–73	1972
23	*Akudnirmiut Glacier*	67°35'	65°15'	1971–72	

[1] See figure 1.

[2] See figure 1 in “Glaciers of Baffin Island” section.

[3] See figures 6 and 12 in “Glaciers of Baffin Island” section.

Maps of the winter balance in the Queen Elizabeth Islands (fig. 10) show a very pronounced accumulation gradient running from the southeast to the northwest across the islands. This gradient should not be interpreted in terms of a Baffin Bay moisture source. It is rather that the major water-vapor trajectories are from the southeast (Koerner, 1979), and they probably have a source that is quite distant. Calculations, based on the stable isotope relationships of precipitation in the area, suggest that only 8 percent of the precipitation on the top of Devon Ice Cap is from Baffin Bay. The pattern of snow accumulation is associated with the growth of more dynamic glaciers in the southeast of the Queen Elizabeth Islands, that is, those facing Baffin Bay (fig. 1). This is the only area in the islands where major ice caps reach sea level. Elsewhere, they end on plateaus well above sea level and reach sea level only in the form of channeled outlet glaciers, Snow accumulation, in grams per square centimeter per year

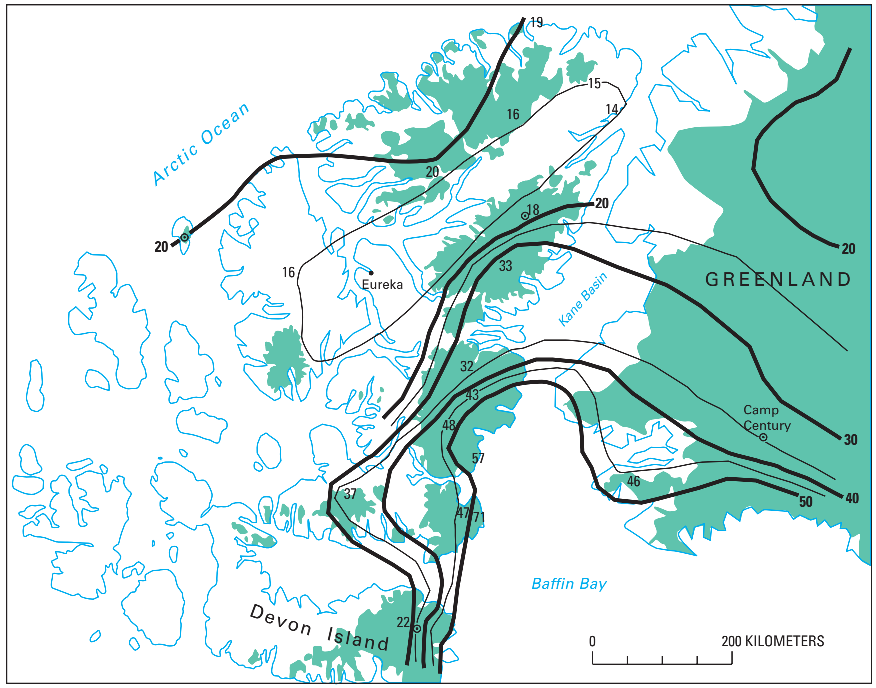


***Figure 10.**—Winter balance in the Queen Elizabeth Islands based on pit and shallow-core analysis done in 1974. Values in Greenland are from Benson (1961) or Mock (1968). The green represents ice-covered areas, and the concentric circle and dot symbol represents ice-core sites.*

such as Sverdrup Glacier (figs. 2, 3). On the other hand, the precipitation-shadow area in north-central Ellesmere Island is very dry (fig. 10). The weather station Eureka (figs. 1, 10) sits in the middle of this area and is well known for its clear skies. Indications of slightly higher snow accumulation exist close to the Arctic Ocean, perhaps associated with northerly air-mass trajectories, rather than an Arctic Ocean moisture source.

Annual and Summer Balance

The ablation season normally covers part or all of the months of June, July, and August. The amount of melt is, in general, inversely related to elevation, and about 1–3 m of ice ablates at sea level on the glaciers each summer. However, in approximately 9 summers out of 10, melting extends right to the tops of all the ice caps, where the meltwater refreezes within the annual surface-snow layer. The effects of melting on the glacier landscape will be discussed later in the "Summer Ablation and the Glacier Landscape" section.

Measurement of the summer balance in the *ablation area,* on an annual basis, consists of measuring the length of a number of poles drilled into the ice. Their changing length gives the amount of ice melted each year. Since the introduction of automatic weather stations a few years ago, the progress of melt throughout each summer season can be recorded at a few sites (fig. 11). An ultrasonic sounder measures the distance from the sounder to the ice surface. As the ice melts, the distance increases, which thereby gives the ablation rate. Because all the winter snow melts in the *ablation area* (that is, where the **annual** balance is negative), the annual balance is the sum of winter snow and total ice melt. Very occasionally, such as in the summer of 1962, the *ablation area* may creep into the firn zone. In this case, the annual balance is the sum of snow and firn melt, although the latter is very difficult to assess, as we explain below.

The measurement of mass balance in the *accumulation area* of subpolar glaciers is not simple. In general terms, snowmelt and icemelt on *temperate glaciers* form part of the ablation process. The meltwater,

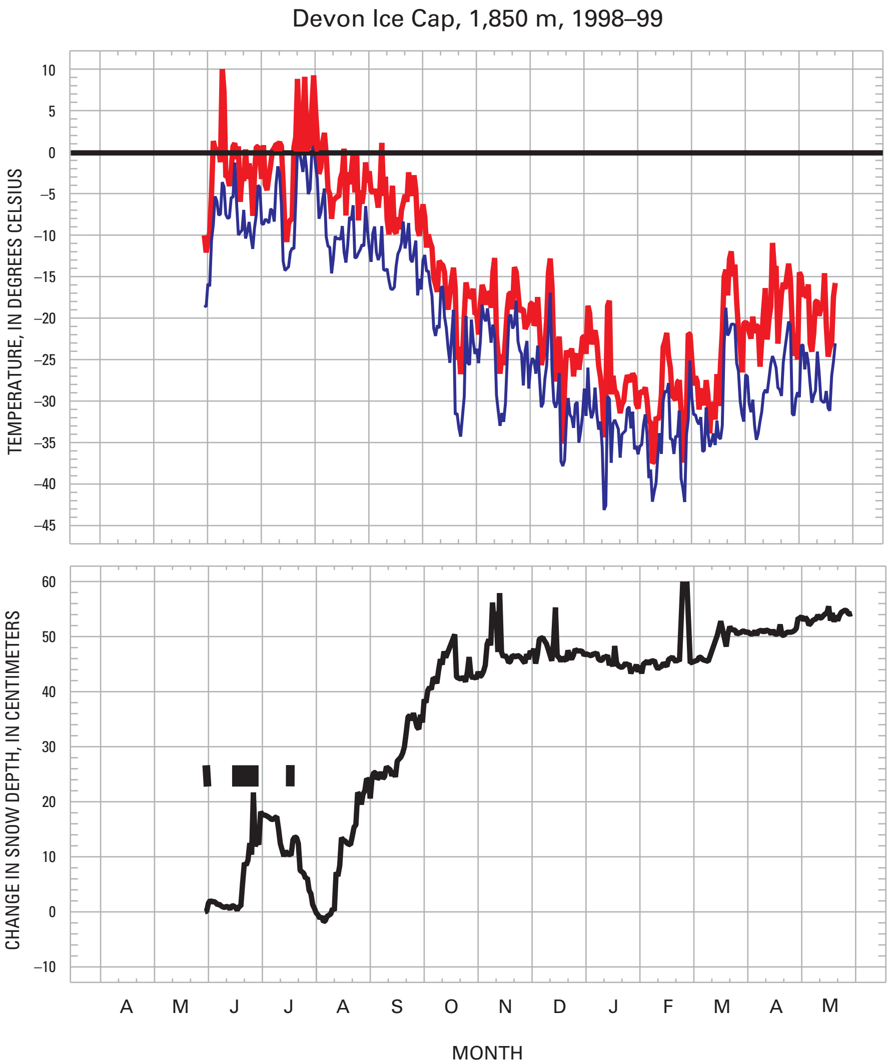


Figure 11.*—Temperatures and snow-depth changes as measured at an automatic weather station on Devon Ice Cap at 1,850 m in 1998–99. The blocks in late May, June, and July mark periods of snow accumulation during the melt season. Lowering of the snow surface in this period is due to snow melt in what was then the percolation zone. The lowering does not represent mass loss because the snowmelt refreezes at depth, in this case within the annual snow layer. The red and blue temperature traces represent maximum and minimum daily temperatures.*

whether it runs off the surface of the glacier or drains through the firn, leaves the glacier. This is not the case on subpolar glaciers where meltwater either refreezes on the ice surface under the snow or firn, or within the firn itself. Benson (1961) divided the accumulation area of polar ice caps and ice sheets into facies zones. In order, with decreasing elevation, he defined *dry-snow, percolation,* and *wet-snow facies,* as well as *superimposed ice zones.* No melting takes place at any time of year in the *dry snow zone*, which covers large parts of Greenland and Antarctica. However, High Arctic islands' ice caps lie below this zone, except in very cold summers (for example, 1965). In the *percolation facies,* melting refreezes within the current annual snow layer. The upper parts of the *accumulation areas* of the Canadian High Arctic ice caps generally fall within this zone, except in warm summers like those of 1962 and 1998 (fig. 12). In the *wet-snow facies,* melt will percolate through more than the annual layer of new snow on the surface. In warm summers, this zone may cover the entire firn *accumulation area*. In the *superimposed ice zone,* the annual snow layer melts and part of it refreezes on the ice surface beneath (Schytt, 1949, 1955; Koerner, 1970a).

On *temperate glaciers*, the equilibrium line separates an *accumulation area*, where the surface is firn, from the *ablation area*, where the surface is ice. Satellite imagery can be used to map these areas as they have very different reflective properties (Williams and others, 1991). On a subpolar glacier, the *accumulation area* includes the *superimposed ice zone*. This means that ice extends, at the surface, beyond the *ablation area* and into the *accumulation area*; hence, the equilibrium line is at the down-glacier margin of *superimposed ice zone*. Satellite imagery cannot, therefore, at present, clearly differentiate the demarcation between the *accumulation* and *ablation areas* on subpolar glaciers.

Mass-balance-measurement techniques should be unique to each of Benson's facies zones in the *accumulation area*:

(1) In the superimposed ice zone, the additional layer of ice has to be measured at the end of each melt season. This is normally done by measuring the increased height of ice on poles drilled into the ice. However, the pole may channel meltwater down to the ice under the winter snow and thereby cause a local increase in superimposed ice formation. An area around the pole has to be checked to account for this.

(2) Measuring annual accumulation in the wet-snow facies is more difficult. It is seldom possible to determine the depth to which meltwater has percolated. Consequently, no well-defined annual layer exists. Densification of the firn and snow, due to meltwater percolation and refreezing, makes pole measurement highly inaccurate. This is not always taken into account (Cogley and Adams, 1998). The practice at the GSC has been to use trays buried deeply enough in year 1 to catch meltwater percolating down in year 2. This method is adequate in most years. However, in heavy melt years (for example, 1962 and 1998 in fig. 12), the percolation tray may overflow, or, at the very least, seriously affect the melt process itself. The use of automatic weather stations (fig. 11) increases the accuracy of the mass-balance measurement, particularly in very warm summers. The automatic weather stations records the snow accumulation, on an hourly or daily basis, throughout the year. In this case, it is the part of the record between spring and the end of the melt season (that is, the summer balance) that is important. Because the automatic weather station only records the change in height of the ultrasonic snow sounder above the snow surface, the snow density

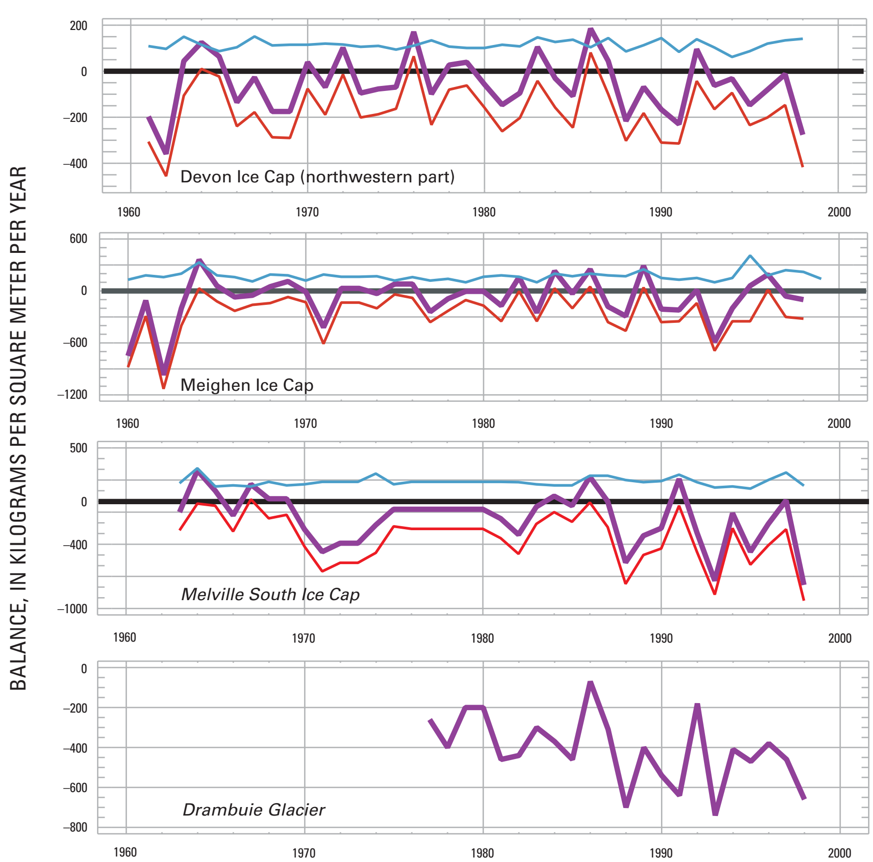


Figure 12.—*Mass-balance records from the northwestern part of Devon Ice Cap, Meighen Ice Cap,* Melville South Ice Cap, *and* Drambuie Glacier *on the northeast side of Agassiz Ice Cap. The top (blue) line is the winter balance, the middle (purple) line is the net balance, and the lowermost (red) line is the summer balance. The figure shows that net balance is driven by summer melting rather than by winter snow accumulation, which is relatively consistent from year to year. The* Drambuie Glacier *time series is not a true mass balance because it does not include area. It is the sum of the specific balance at each of the poles, measured each year. Although the system [used for the* Drambuie Glacier*] does not represent mass balance for the ice cap and glaciers, it nonetheless serves to monitor climate change (see Koerner, 1986). The values from* Drambuie Glacier *are more negative than on the other glaciers because the measurements do not extend to the top of the ice cap. Data are from the Geological Survey of Canada.*

remains unknown. Measurements over four decades have shown that new snow, which is almost always associated with wind drift, has a density of between 0.1 and 0.2 g cm^{-2}. A density within this range is chosen according to the nature of the automatic weather station record. The method introduces a 5–10 percent error into the annual accumulation measurements between the percolation line and the upper limits of the superimposed ice zone.

(3) Mass-balance measurements in the *percolation facies* are simpler. Pole measurements are used as a guide so that the annual layer is readily recognized from its an annual sequence of refrozen melt-soaked firn overlying unsoaked firn. Depth and density measurements then give the mass balance.

Results

The mass-balance results from the GSC program are shown in figure 12. It is very clear that the net annual balance is driven by the summer melting. Each series shows that the period of record has been one of very slightly negative balance. Alt (1978, 1979) has analyzed some of these records and related them to synoptic conditions in the Arctic Islands. However, no statistically significant trend is found in the various time series (Koerner and Lundgaard, 1995; Koerner, 1996; Cogley and others, 1995, 1996). The same lack of trend extends eastward to the glaciers in Svalbard (Dowdeswell and others, 1997; Hagen and Liestøl, 1990). This contrasts with the substantial warming over the same period in the western Arctic. Regional differences in climatic change should not be unexpected, and they emphasize the need for monitoring the climate closely. As we have seen, glaciers are particularly valuable in this respect.

Summer Ablation and the Glacier Landscape

Melting produces a variety of impressive supraglacier, englacier, and subglacier features (Maaq, 1969). Melt streams can be seen in figure 3, where they are marked with a letter ‘m’. These melt streams may be up to 15-m wide and flow in channels as deep as 30 m that have vertical, or even overhanging, sides. Some even flow in tunnels just below the ice surface (for example, just downstream of the confluence of the two glaciers forming Sverdrup Glacier in fig. 3). Most of the meltwater finds its way to the margins of the glaciers, where large streams flow between the rock wall and glacier side and commonly through subglacier and englacier tunnels. Many streams disappear down moulins, especially near the terminus (for example, the northernmost ‘m’ on Sverdrup Glacier in fig. 3). Streams flow at high velocities because of the low friction between water and ice and the smooth nature of the channels. Consequently, they are able to transport boulders greater than 1 m^3 in size. Several such boulders can be found perched along the courses of abandoned streambeds.

The heavily crevassed glaciers draining the west side of Agassiz Ice Cap in northern Ellesmere Island (fig. 8, (A), (C), (D)) have much higher melting rates in their ablation areas than typical High Arctic glaciers. This is suggested by unpublished measurements by the GSC on one of these glaciers (fig. 8, (A)), as well as on highly hummocked areas of glaciers on southeastern Devon Island (Koerner, 1970b). Not only do these glaciers have relatively high specific ablation rates, which are related to their increased mesorelief and lower albedo, but they also have a greater surface area as a result of their crevassed and hummocked relief. Thus, ice loss due to melting may be up to an order of magnitude higher in their ablation zones than on glaciers elsewhere. An example of a highly hummocked glacier, Iceberg Glacier, which undergoes intense summer melting, is shown in figure 13.

A late August 1974 Landsat 1 multispectral scanner (MSS) false-color composite mosaic image (fig. 5) illustrates a High Arctic landscape at the height of the summer melt season. Various shades of blue and gray through to white (most clearly seen on the west side of the ice cap) are related to the progress of melting at different elevations. At the top of the ice cap, the snow is still highly reflective, which shows that it has melted very little, if at all. Lower down, it changes to a light gray, which indicates snow in an advanced state of melt. On the west side of the ice cap, the light-gray tone abruptly changes to either a dark gray blue or to blue. The blue tone is probably new superimposed ice, formed either during the melt season when the image was taken or during the previous melt season. The dark gray blue is older glacier ice that was formed higher up the flowline by firn compaction. The gray part of the tone becomes more pronounced closer to the glacier terminus. The light-gray tone (wet snow) extends right down to sea level on the east side of the ice cap. This is because the snow, which accumulates to much greater depths there, has not yet completely melted to expose the blue glacier ice underneath. The heads of Makinson Inlet and Vendom Fiord are already free of sea ice. The melting of the sea ice is hastened by runoff of snowmelt from ice-free terrain.

Figure 13.*—High-angle, oblique aerial photograph of Iceberg Glacier, Axel Heiberg Island (see fig. 7 for location), taken on 23 July 1964. The highly hummocked surface is associated with intense melting.*

Ice Caps and Climatic Change

Early attempts to derive climate-change records from ice caps were based on the number and thickness of ice layers and on changing stratigraphy in snow-firn pits (Hattersley-Smith 1960; Müller, 1963b; Koerner, 1970b). Hattersley-Smith (1960) detected a sharp warming of climate in the 1920's by using this method.

The Ice-Core Record

Ice cores allow climate-change studies to be carried out to much greater depth (that is, a longer length of time). As snow accumulates year by year in the accumulation area of ice caps, it traps within it atmospheric aerosols, which include cosmic and terrestrial particles (for example, dust and tephra), heavy metals, pollen, and acids. Atmospheric gases also are trapped. The oxygen-18/oxygen-16 ratio of the snow ($^{18}O/^{16}O$) and the changing percentage of ice layers in the core give a proxy record of past temperatures. At the top of the ice caps, where ice movement is mainly vertical, a continuous record of these variables exists that covers the entire history of the ice cap. This is where ice cores in the High Arctic islands have been drilled. The location of each deep drill site in the Canadian High Arctic is shown in figure 1.

One of the most important contributions of ice-core analysis to climate-change studies is a temperature proxy. This is a temperature derived from a variable measured in the core. The stable isotope-temperature relationship depends largely on the temperature of formation of precipitation; the lower the condensation temperature, the more negative the ^{18}O values of precipitation. The isotope record is either in terms of hydrogen or oxygen isotopes in water: hydrogen-deuterium-oxygen (HDO) or $H_2{}^{18}O$, both with respect to Standard Mean Ocean Water (SMOW); that is, δD or $\delta^{18}O$. This is translated into temperature on well-founded empirical relationships between surface temperatures and the δD or $\delta^{18}O$ composition of the snow (Dansgaard and others, 1973). Limitations apply to this temperature proxy. The most serious are changing sources of water vapor, variations in the length of the seasonal fractions of snowfall, and changing elevation of the drill site. However, except for the areas facing Baffin Bay, the Queen Elizabeth Islands' ice caps bear only a very weak relationship with elevation (Koerner, 1979). Ice layers that form in the snow pack in summer can be used as a *summer* temperature proxy (Koerner, 1977a). The warmer the summer, the greater the number or thickness of ice layers that form. Changing concentrations of ice layers in the ice cores can then be related to summer climate changes in the past (Koerner and Fisher, 1990).

The first task of ice-core analysis is the determination of a time scale. This may be done by counting annual cycles of some of the included "contaminants" down selected lengths of the core, by detecting acid layers from known volcanic events, by picking up well-known climatic events in the oxygen-isotope record, or, theoretically, from a knowledge of the flow law of ice. A combination of these methods has been used to develop a time scale for the Canadian High Arctic ice cores. This time scale is accurate to about 5 percent for the last 5,000 years and to about 10 percent for ice that formed at 10 ka to 5 ka. In ice older than 10 ka, the time scales have been derived from comparisons between oxygen-isotope and particulate profiles in the Canadian and the more accurately dated Greenland ice cores. This part of the record is contained in the lowermost 5–10 m of the core, where flow over uneven bedrock can produce distortions and gaps in the stratigraphy (Paterson and others, 1977).

The first deep ice core drilled from the Queen Elizabeth Islands was from the Meighen Ice Cap (figs. 1, 9*A*). Analysis of the 121-m core, which is from the highest part of the ice cap close to its southern edge (Paterson, 1968, 1969; Koerner, 1968; Koerner and others, 1973; Koerner and Paterson, 1974), originally suggested that the deepest ice is of the order of 4.5 ka. However, because of the difficulty of establishing a time scale in an ice cap formed dominantly of superimposed ice (Dowdeswell and others, 1990; Koerner, 1997), all one can say is that it began to form in the latter part of the Holocene Epoch.

Records from some of the other cores are shown in figure 14, which has three records from Canada, which shows a –20 degree Celsius cooling (compared with today) in the coldest parts of the last glacial (Fisher and others, 1983, 1998), four from Greenland (Dansgaard and others, 1985; Johnsen and others, 1992), and one from Antarctica (Jouzel and others, 1993). Ice deposited during glacials is differentiated from interglacials on the basis of the presence of higher concentrations of microparticles and major ions in glacial-period ice, its more negative $\delta^{18}O$ values, and its finer grained ice texture. Despite quite different thicknesses, the main interglacial-glacial sections are common to each ice core. However, only the Vostok and Summit cores from Antarctica and central Greenland, respectively, contain ice older (fig. 14, (D)) than that deposited during the last interglacial (C) (Jouzel and others, 1993; Johnsen and others, 1992).

Basal ice in the Canadian cores (fig. 14, (C)) has high pollen concentrations and less negative oxygen-isotope $\delta^{18}O$ values, consists of much clearer (less bubbly) ice, and, on Agassiz Ice Cap, contains large amounts of dirt (Koerner and others, 1988). Such a signature, as well as its position at the base of the ice cores, suggests that it was deposited in the early growth stages of the ice cap, during the last interglacial, when the climate was entering the Wisconsinan glacial. Ice older than this must have melted during the main part of the same interglacial time, as it did at the sites of the Camp Century and Dye 3 cores in Greenland (Koerner, 1989). Because the Canadian ice-core sites are in the central and highest parts of their respective ice caps, it suggests that the Canadian High Arctic islands were ice free during some part of that same interglacial (Koerner, 1989). Similarly, it has been suggested (Koerner, 1989; Cuffey and Marshall, 2000) that the Greenland ice sheet, although it was not completely removed, must have been substantially smaller at that time.

The Holocene part of some of these records is shown in figure 15. The Penny Ice Cap and Agassiz Ice Cap $\delta^{18}O$ records show an early thermal

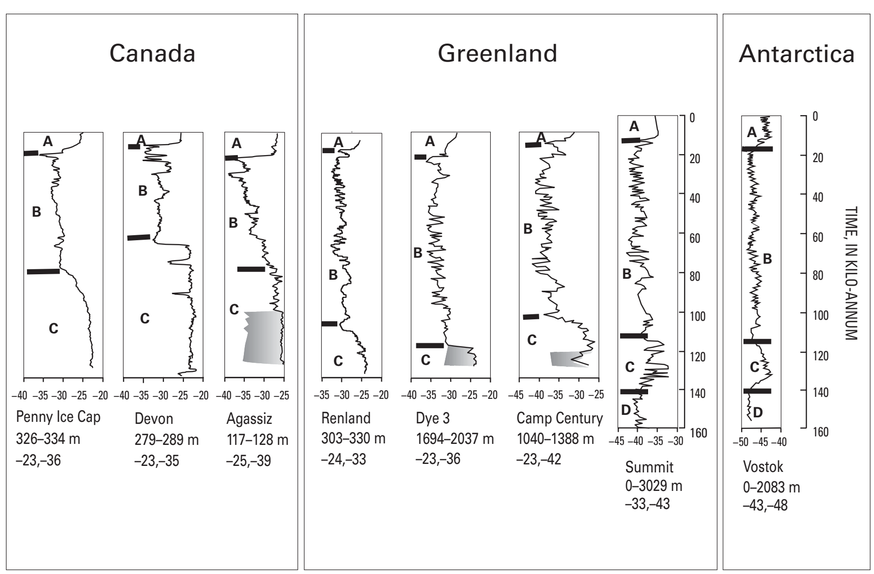


Figure 14.—*$\delta^{18}O$ profiles from Canadian ice caps (Fisher and others, 1983, 1998), Greenland ice sheet (Dansgaard and others, 1985; Johnsen and others, 1992), and Antarctic ice sheet (Jouzel and others, 1993). The $\delta^{18}O$ values are in parts per thousand with maximum and minimum values listed under each ice-core section. The shaded part at the base represents ice that has a high silt content. (A), Holocene; (B), Würm/Wisconsinan glacial; (C), Eemian/Sangamonian interglacial (Summit, Vostok), or early growth ice dating either from late Eemian/ Sangamonian interglacial or early Würm/Wisconsinan glacial (Penny, Devon, Agassiz, Dye 3, Camp Century); (D), earlier Pleistocene. See text for location of sections. The Summit and Vostok cores are plotted on a time scale (right-hand side of the diagram). For ice older than 110 ka, the Summit time scale is inaccurate due to dynamic disturbance of the ice at those depths. The other six cores are plotted on individual linear depth scales, where the depth interval is noted for each profile.*

maximum. However, it is not as early as that shown by the ice-layer record from Agassiz Ice Cap (percent of melt). The difference may be due to the effect of massive runoff of meltwater from the Laurentide ice sheet at this time. This water, which had very negative [oxygen-isotope] $\delta^{18}O$ values, and which may have pushed the water-vapor source farther south, is likely to have had the effect of making the [oxygen-isotope] $\delta^{18}O$ values too negative (Koerner, 1988; Fisher, 1992). The salt record from Penny Ice Cap also shows a very early Holocene maximum. Salt has been shown to have an inverse correlation with sea-ice extent in Baffin Bay (Grumet and others, 2001). The Holocene salt record, therefore, suggests that the extent of sea ice has been slowly increasing in Baffin Bay from an early Holocene minimum. Dyke and others (1996) concluded, from ^{14}C-dated fossil bones in the Arctic Islands, that Bowhead whale populations were at a maximum in the early Holocene. Together, the $\delta^{18}O$, ice-layer, and salt records suggest a varying, but overall, cooling trend has taken place throughout most of the Holocene Epoch, particularly during the last 5,000 years. The modern

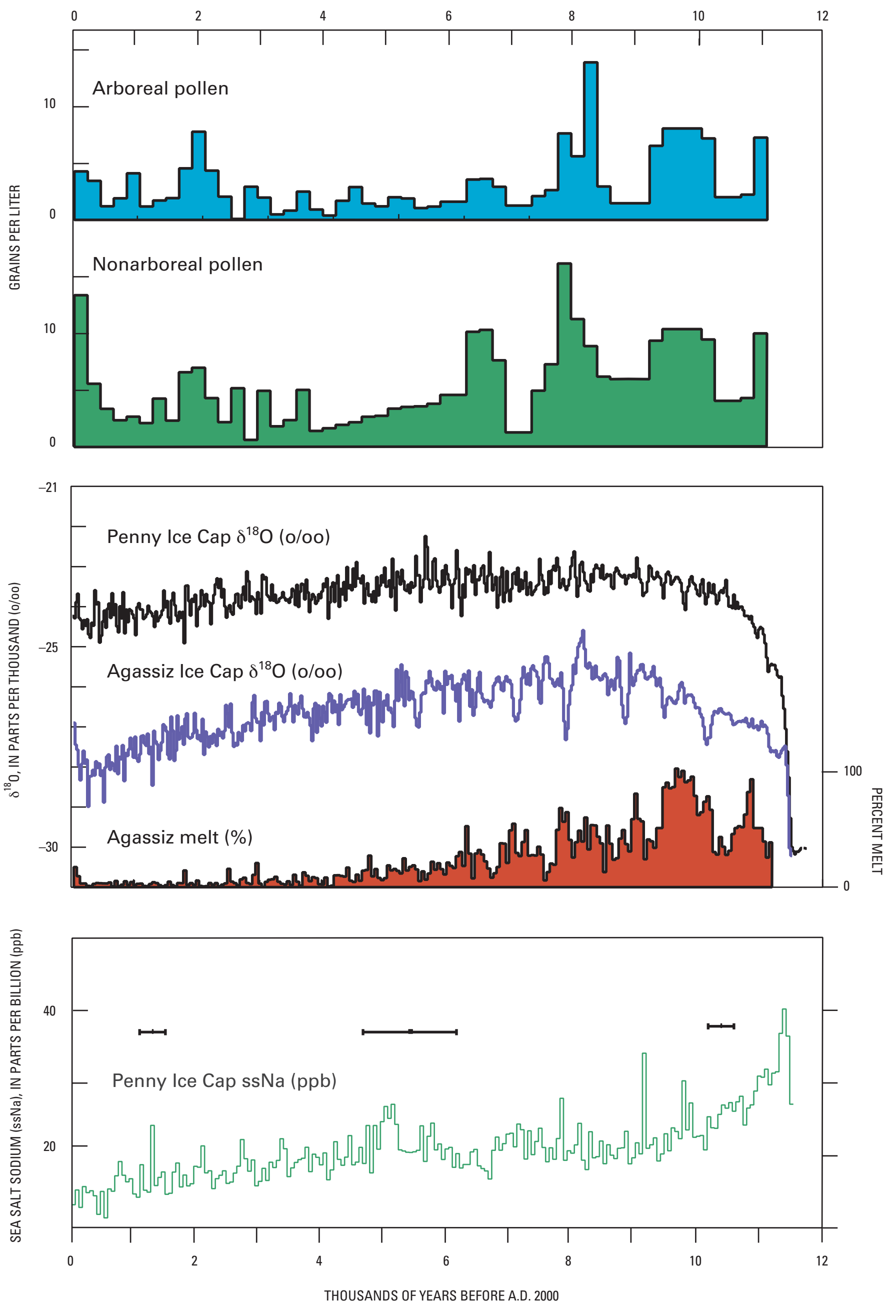


Figure 15.—*Paleoclimate records from the Holocene section of ice cores from Penny and Agassiz Ice Caps. A higher percent melt from the ice-layer record or less negative $\delta^{18}O$ values signify a warmer climate. Conditions in the early and middle Holocene were warmer than today by about 2°C and show an early Holocene thermal maximum. The sea-salt record from the Penny Ice Cap also shows a very early Holocene maximum. Salt has an inverse correlation with sea-ice extent in Baffin Bay. Therefore, the record suggests a gradually increasing sea-ice extent in Baffin Bay from an early Holocene minimum. Three horizontal error bars indicate the range of uncertainty in the sea-salt record during these intervals of time. The Holocene pollen record also suggests warmer conditions in the early Holocene and in the last 2,000 years.*

warming during the last 100–150 years appears to have produced the warmest period of at least the last 1,000 years. It follows, but highlights, a cold period ("The Little Ice Age") ending about 150 years ago. The "Little Ice Age," coming at the end of an overall cooling trend, may be nothing more than a slight cooling variation within that trend.

The Holocene pollen record, although it shows higher pollen concentrations in early Holocene ice that suggest warmer conditions, also shows increasing concentrations in the last 2,000 years. Pollen is probably not a direct temperature proxy. Whereas the other records suggest that the thermal maximum was in the early Holocene, it was a time when the Laurentide ice sheet still covered a large part of Canada. What is a pollen source today was, therefore, covered by ice at that time. An alternate early Holocene source may have been northwestern Canada, where the tree line reached farther north than today. Whether the late Holocene pollen increase represents a return to a western air-mass trajectory and pollen source has not yet been determined.

Ice-core records from Severnaya Zemlya and Svalbard to the east (Koerner, 1997) show similar Holocene $\delta^{18}O$ and ice-layer records: an early thermal maximum followed by a trend of decreasing temperatures (Koerner, 1997). It has already been shown that glacier mass balance is slightly negative today, which the ice-core records indicate is not as warm as the early Holocene (fig. 15). It suggests, therefore, that ice caps and glaciers must have had more strongly negative balances in the early Holocene than today. Only two ice cores from the Severnaya Zemlya and Svalbard group of islands show any evidence of the Pleistocene ice seen in the Canadian cores (fig. 14). The evidence for Pleistocene ice from one of these cores, Vavilov Ice Cap (Stiévenard and others, 1996), is questionable. Probably most of the Pleistocene ice caps at these sites melted during the period of negative balance in the early Holocene; only the larger ice caps in the Canadian islands and Academii Nauk Ice Cap on Severnaya Zemlya survived. Increasingly colder conditions, particularly during the last 5,000 years, have promoted regrowth of glaciers. They reached their maximum extent about 150 years ago (Müller, 1966). It was at this time that the desperate, but unsuccessful, attempts to open up the Northwest Passage were made. Success was achieved by the Norwegian explorer Roald Amundsen early this century when summer conditions promoted more open sea- ice conditions (Koerner, 1977a; Alt, 1985).

Aerial Photographic and Satellite Image Evidence of Glacier Fluctuations

Although ice cores are valuable in revealing climatic changes in the past, they do not tell much, on their own, about past changes in the dimensions of the ice caps. They show how climate changes over a broad range of wavelengths and amplitudes with respect to time and temperature. In terms of the response of glaciers to climatic changes, the waxing and waning of continental ice sheets is a response at one end of the spectrum, and the growth and disappearance of stagnant ice caps is a response at the other end of the same spectrum. Nye (1963) and Jóhannesson and others (1989) considered the problem of response times of glaciers, times that may vary from a few years to many thousands of years. In the Canadian High Arctic islands, one might expect a response time of about 2,000 years on the largest dynamic ice caps without valley outlets, such as, for example, the west side of Devon Ice Cap (fig. 2), but one might expect an immediate response on stagnant ice caps. Thus, a small glacier may be advancing in response to a recent climatic change while a larger one is retreating in response to a climatic change that ended a thousand years before. Relating these geometry changes to those of climate is further complicated because of our poor knowledge of the ice volume and bedrock topography of the Queen Elizabeth Islands' ice caps. Extensive radio-echosounding will help enormously in this respect (J.A. Dowdeswell, oral commun., 1999).

To examine changes in the areal extent of the Queen Elizabeth Islands' glaciers and ice caps, we compare vertical aerial photography taken in the 1950's with Landsat imagery taken in the 1970's. The Landsat MSS imagery, because of its 79-m pixel resolution, can only detect changes at the terminus of glaciers over a 20 year timespan, if the changes are greater than an average, cumulative change of 4 m a^{-1} (Williams and others, 1997).

Dynamic Ice Caps and Outlet Glaciers

On the basis of Landsat imagery shown in figures 2, 4, 5, and 7 and the vertical aerial photograph in figure 3, no significant marginal changes have been found on the larger ice caps. Some measurable changes are evident in glacier terminus positions that are mostly "increased-ablation" retreats in response to the warmth of the 1950's (Bradley and England, 1978). Because the time period between the aerial photography and the Landsat images includes part of the very warm period from 1920 to 1960, it may appear surprising that relatively few measurable changes are seen at the margins of the small glaciers. Similarly, why are some of the glaciers not advancing in response to the cooler climate of the "Little Ice Age" (about 1550 to 1850)? Continuing mass-balance measurements also show very few signs of warming or cooling in the eastern Canadian Arctic during the last 40 years.

The answer may be found in the low *activity index* of the High Arctic glaciers. That is, the accumulation and ablation rates are relatively small when compared with those of alpine glaciers or the maritime-nourished glaciers of Alaska. Changes are, therefore, likely to be small and not detectable by the use of 79-m pixel resolution imagery in a time period of only a few decades. The higher resolution satellite imagery, such as the Landsat Enhanced Thematic Mapper (ETM+) (15-m pixels), the Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) (15-m pixels), and IKONOS (1-m pixels), that is available will provide a better opportunity to detect changes.

A few documented cases do exist, however, of glaciers that have advanced dramatically in the High Arctic islands. One of them, the Otto Glacier in northern Ellesmere Island (fig. 1), has been described as a surging glacier (Hattersley-Smith, 1969). This glacier advanced 2–3 km between 1950 and 1959 and a farther 2–3 km between 1959 and 1964. The lower part of this advancing tidewater glacier was afloat. However, because the Otto Glacier is both a tidewater glacier and a surging glacier, the advance may not be related to climate (Post, 1975).

Another case of a possible surging glacier is the *Good Friday Bay Glacier*[2] (fig. 7). An advance of about 2 km began on this glacier between 1952 and 1959. Although some of the surface characteristics of this glacier suggest a surge (Müller, 1969), climatic forcing cannot be ruled out.

Figure 4 illustrates a 6.5-km retreat of a valley glacier that flows into South Cape Fiord on southwestern Ellesmere Island. The retreat took place between 1957 and July 1974. The retreat is associated with increased crevassing of the surface toward the terminus. However, no changes to surface level can be seen along the glacier. We cannot say whether the glacier is retreating *from* a more common position or *to* a more common position. None of the nearby glaciers has advanced or

[2] The names in this section conform to the usage authorized by the Secretariat of the Canadian Permanent Committee on Geographic Names (CPCGN); URL address: [http://GeoNames.NRCan.gc.ca/]. The Website is maintained by the Secretariat through Geomatics Canada, Natural Resources Canada, and combines the CPCGN server with the Canadian Geographical Names Data Base (CGNDB). Variant names and names not listed in the CPCGN/CGNDB are shown in italics.

retreated during the same period. On first inspection, the Landsat image in figure 4, taken only a month after a set of aerial photographs that showed the retreat, appears to show the terminus in its old, advanced position. Prior knowledge that the glacier has retreated, together with very careful examination of the image, shows that the ice in front of the terminus is either sea ice or icebergs.

Stagnant Ice Caps

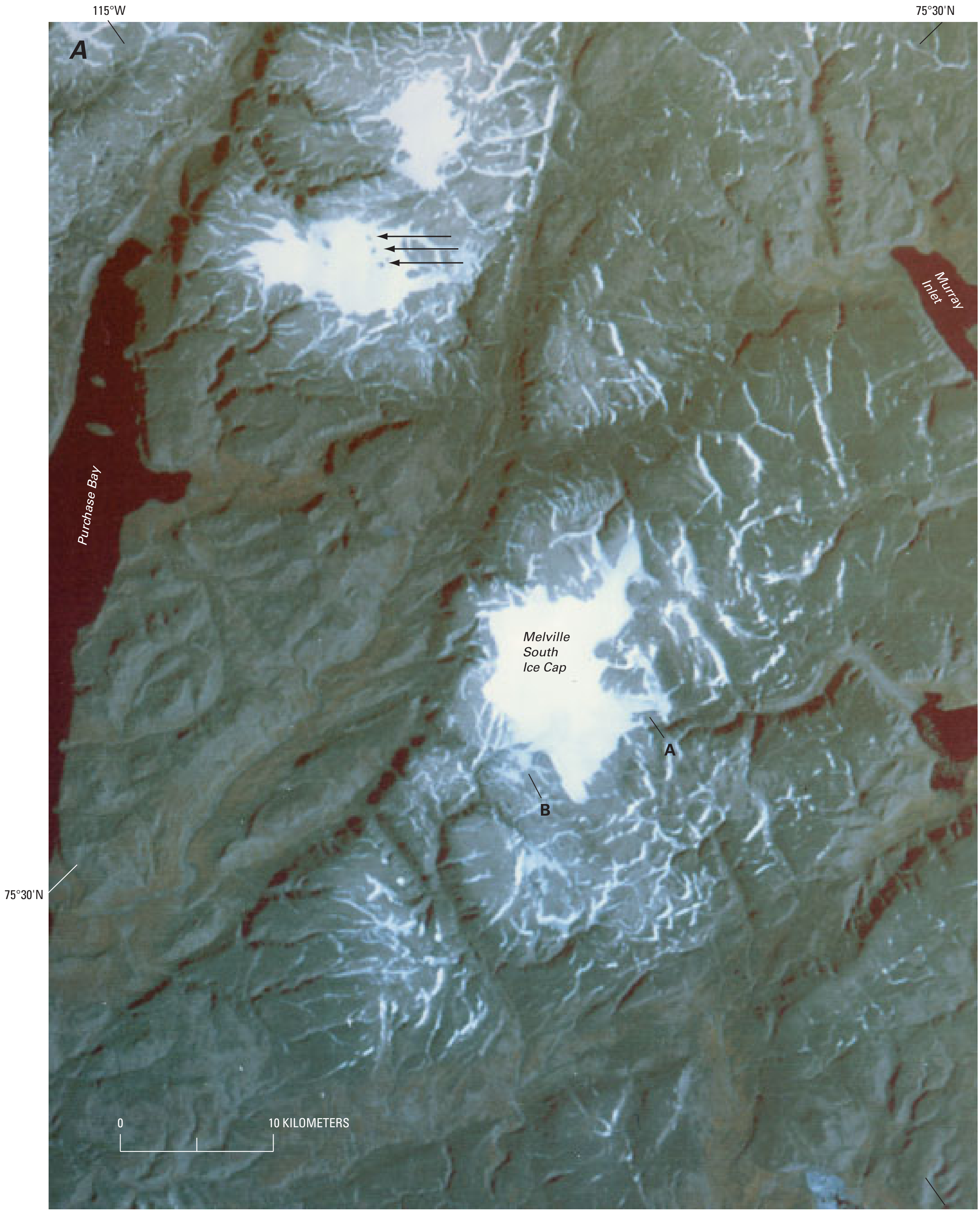
Stagnant ice caps respond immediately to climatic change. A warming trend will always cause marginal retreat of these ice caps. However, retreat at its margins does not necessarily indicate that the ice cap has a negative balance. As long as the equilibrium line altitude (ELA) is low enough, the ice cap may retreat despite having a positive balance. Because the ice cap is stagnant, the margin will "advance" only if the ELA lies beyond the margin. Although Meighen Ice Cap does not have an overall positive balance, it does have a slightly positive balance at its highest elevations. Retreat of its margins, accompanied by thickening in its central parts, is steepening its slopes. Theoretically, steeper slopes could cause greater basal shear stress and change it from a stagnant to a weakly dynamic ice cap.

Figure 16*A* shows three of the ice caps on Melville Island (fig. 1). The small nunataks on one of the two more northerly of the Melville Island ice caps (arrows in fig. 16*A*) show no measurable change between a 1957 aerial photograph and the 1977 Landsat image. Some retreat is evident at (A) and (B) on the southern ice cap (compare with fig. 16*B*). Retreat elsewhere may be hidden by a late snow cover at the time the Landsat image was taken; some snow still remains on the surrounding plateau. However, even if a general retreat is present at the margins, it must be quite small.

The situation is a little different on Meighen Ice Cap (fig. 9). The nunataks marked by arrows (fig. 9*A*) are more exposed in the 1977 Landsat imagery than in earlier (1959) aerial photography (fig. 9*B*). In one case, a nunatak near the west margin that is seen on the earlier aerial photograph is part of the surrounding plateau in the later Landsat image. The ice margin has also retreated at the locations indicated by arrows. From aerial photographs, Arnold (1965) found slow wastage at the margins of Meighen Ice Cap for the period 1950–59. An oblique aerial photograph taken on 18 July 1950, and shown in Arnold (1965), shows a continuous ice cover where the Landsat image shows a long narrow nunatak at the northern end.

Clear changes can be seen on the smallest stagnant ice caps when comparing aerial photographs of the 1950's and the satellite imagery of the 1970's. Comparing figures 2 and 3 of the area close to the northwest margin of the Devon Ice Cap, we can see that three small ice caps (fig. 3, arrows) have disappeared by the 1970's, when they were replaced by bare ground (fig. 2). Another small ice cap (fig. 3, (C)) just above the terminus of Sverdrup Glacier has almost disappeared, and a slightly larger one to the southwest (fig. 3, (E)) has been separated from a section located in a nearby gully.

These are not especially dramatic changes, but they do suggest that the climate of the 20-year period between the 1950's and 1970's was warmer than the climate under which these ice caps formed. The records of negative balance from Meighen Ice Cap and Melville Island ice caps would certainly apply to these smaller stagnant ice caps. It seems most likely that they began growth during the colder parts of the last 1,000 years and are presently out of phase with modern warming. Many more may disappear within the next 50 years if the climate does not cool again.


115°W
75°30'N
A
Murray
Inlet
Purchase Bay
Melville
South
Ice Cap
A
B
75°30'N
0
10 KILOMETERS

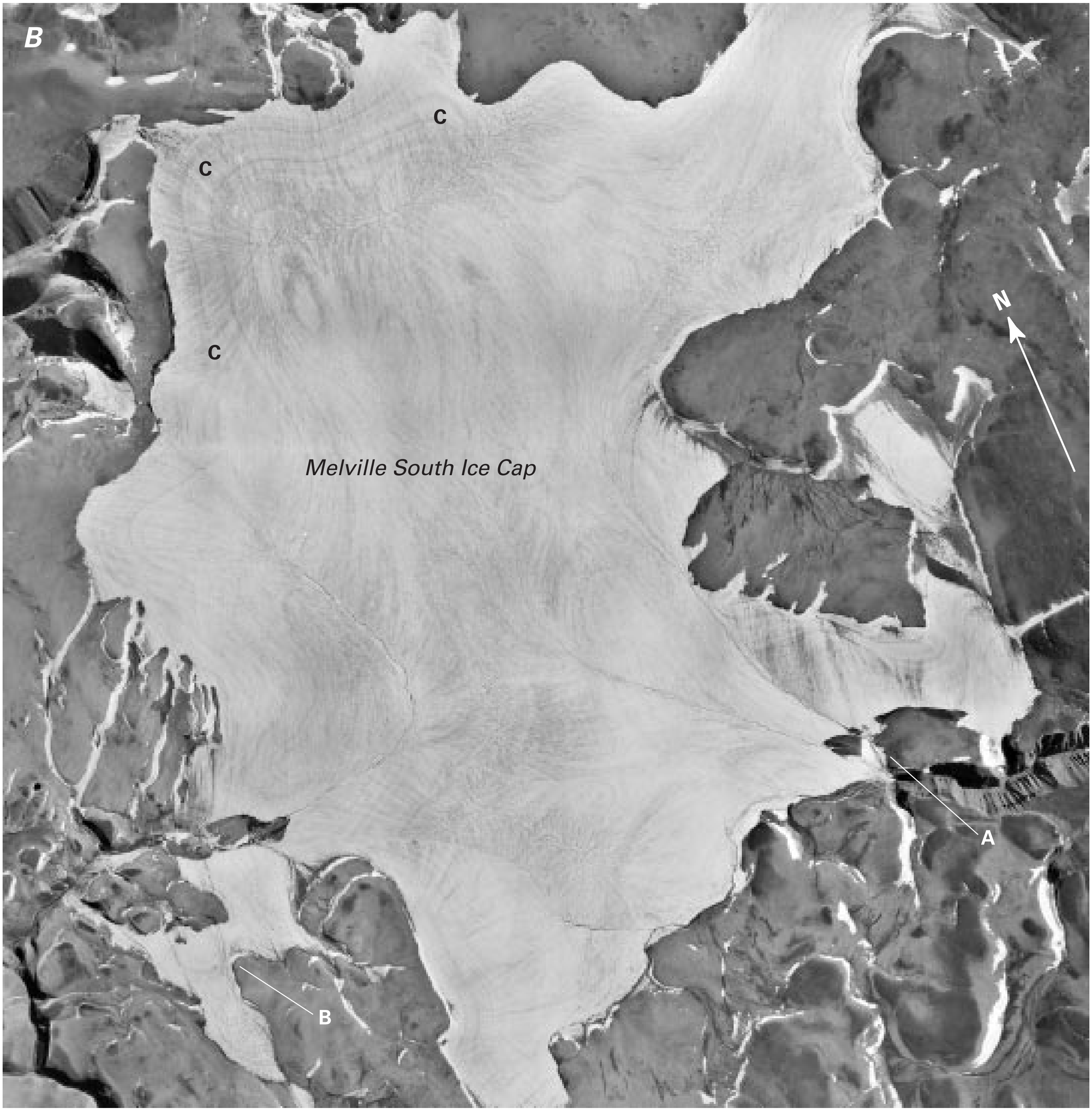


Figure 16.—*Landsat image from 1977 and an aerial photograph from 1957 showing slight changes in the Melville Island ice caps.* **A**, *(opposite page) Part of an enlargement of a Landsat image of three of the Melville Island ice caps. The nunataks at the top of the figure (indicated by arrows) show very little change in area compared to an aerial photograph taken 20 years earlier. A slight retreat can be seen on the southern ice cap at points (A) and (B) when compared with figure 16*B*. The Landsat image (11855–18310, bands 4, 5, and 7; 21 August 1977; Path 66, Row 6) is from the Canada Centre for Remote Sensing, Ottawa, Ontario, Canada.* **B**, *Vertical aerial photograph of* Melville South Ice Cap *taken in 1957. The ice margins at (A) and (B) should be compared with the same points in figure 16*A. *Note the "banding" on the ice surface at (C) formed by outcropping of dust layers.*

Evidence of Changes in Glacier Volume

Changes of glacier thickness may be assessed by more traditional techniques. The absolute, or relative, change of glacier elevation is measured photogrammetrically, by surveying across the glacier between fixed points off the glacier, by radio-echosounding, by gravimetric methods, or by changing borehole length. These surveys measure the difference between the vertical velocity of the glacier and the ice-melt or accumulation rate for the period between the surveys. The vertical velocity in this case is a long-period, or even paleobalance, rate, representing a period of unknown length.

Arnold (1968, 1981) discussed at length the techniques used for glacier surveys. Briefly, one has to survey the surface of a glacier on at least two occasions sufficiently separated in time to detect vertical (height) changes. If markers on the glacier move, any resurveys must be made to the original coordinates. Sets of changing vertical angles, made by the use of traditional survey methods, determine height changes of the glacier surface. Modern photogrammetric techniques can give the same results; the effort involved is less, but the expense greater. Arnold (1968, 1981) has done work of both kinds on Gilman Glacier (fig. 1) and on White Glacier (figs. 1, 7).

Another method involves leveling along a straight line across a glacier between two fixed points on the valley walls. When repeated, comparison of the two surveys gives the change of elevation of the glacier surface. This method has been used on Sverdrup Glacier (fig. 3, dotted line; Cress and Wyness, 1961).

Very careful radio-echosounding, repeated after an interval of a few years, can be used to detect glacier-thickness changes (Nye and others, 1972). Precision gravimeters can also be used to achieve the same results (Bentley, 1975). The latter technique has been used at the top of Devon Ice Cap (Winter, oral commun., 1976). It is also possible to use repeated measurements of the length of surface-to-bedrock boreholes to achieve the same results (Paterson, 1976).

Arnold's (1968) measurements in the accumulation area of Gilman Glacier (fig. 1) found no significant changes during the 1957–67 period. Similarly, gravimeter measurements on Devon Ice Cap detected no change for the 1971–76 period (Peter Winter, oral commun., 1976). In contrast, Paterson's (1976) measurements of change in borehole length indicated a slight thickening of the highest part of this ice cap. Arnold (1968) measured an average surface lowering of 1.66 m (0.17 m a^{-1}) for 1957–67 in the ablation area of Gilman Glacier. He also measured an average surface lowering of 8.2 m (0.83 m a^{-1}) for the period 1960–70 on White Glacier (Arnold, 1981). The lowering of Gilman Glacier and White Glacier took place during periods that included the warmer early 1960's, including the summer of 1962 that had the most negative glacier mass balances on record (fig. 12). Farther south, on Sverdrup Glacier, leveling in 1965 and again in 1975 detected no significant height changes of the glacier 10 km from the terminus and 300 m above sea level.

The most effective method for measuring ice-thickness change, however, consists of repeated geodetic airborne laser altimetry surveys (Krabill and others, 1995). One such survey was done by NASA in 1995 and was repeated in 2000 along flightlines suggested by the GSC. The results of this survey will give broad spatial coverage of change in ice-cap thickness [relative change based on change in surface elevation] in the Canadian High Arctic islands from northern Ellesmere Island to southern Baffin Island.

Conclusions and Recommendations

The Landsat imagery (having 79-m pixel resolution) used in this chapter is barely adequate to detect changes in the margins of the larger ice caps during a 10- to 20-year period. Partly, this is due to the low *activity index* of High Arctic glaciers. Both snow-accumulation and ice-melt rates are small compared to those of alpine glaciers. Changes in both snowfall and ice melt might also be small. Unlike Greenland and Antarctica, where similarly low activity indices apply, the glaciers in the High Arctic are relatively small, so that the kinds of dramatic changes, like massive ice-shelf calving are not present. Stagnant ice caps, on the other hand, do show changes. Their response to climatic change has no lag, unlike dynamic glaciers, and marginal retreat can be directly related to summer climate even on an annual basis. A few of the smaller stagnant ice caps, detectable in the earlier aerial photography, disappeared between the 1950's and 1970's.

Satellite imagery is continually improving, however, and future coverage having improved resolution, such as coverage by the Landsat 7 (ETM+) (15-m pixel resolution), Ikonos (1-m pixel resolution), and the Global Land Ice Mapping System (GLIMS) ASTER (15-m pixel resolution), should prove to be valuable new tools for glacier monitoring in terms of changes in glacier area. The aerial photography taken in the 1940's and 1950's will, therefore, serve as an invaluable baseline for future comparative purposes. The recent application of satellite synthetic aperture interferometry techniques is enabling the mapping of velocity fields on glaciers, ice caps, and ice sheets (Mohr and others, 1998). Now, by combining interferometry and local mass-balance and ice-radar techniques, it is proving possible to calculate surface-elevation changes, as well as the velocity fields on glaciers (Reeh and others, 1999).

Recent application of airborne laser altimetry surveys (having 0.1-m vertical accuracy) to the Greenland ice sheet (Krabill and others, 1995) and smaller circumpolar ice caps (Garvin and Williams, 1993) has already shown its value. A repeat of a 1995 series of flightlines over the Canadian High Arctic islands and Baffin Island in 2000 gave, for the first time, an extensive measure of the changes in ice-cap thickness there. However, any changes on a dynamic glacier relate to climatic change that has taken place in the past, as well as in the present. It may prove difficult to relate those changes of thickness to any particular period of climatic change, although they are directly relevant to the part played by glaciers in sea-level change.

One limitation of satellite imagery over subpolar ice caps is the problem of detecting the equilibrium line. This is because superimposed ice forms part of the accumulation process. Consequently, the accumulation region includes a zone where the surface is ice. This ice has the same characteristics as the ice *below* the equilibrium line, which is superimposed ice that has flowed down the ice cap in the past from its zone of formation. With improvements in the sensors carried on satellites expected to be launched in the future, the situation of satellite monitoring of glaciers will improve. However, it will be essential to conduct field-based glacier research for accurate image interpretation.

The use of automatic weather stations is becoming more applicable and popular in the practice of glacier research and monitoring. These stations are proving especially valuable in terms of correcting for meltwater percolation and refreezing in the accumulation area. However, the effect of rime-ice and hoarfrost accumulation on automatic weather station sensors must be assessed. GSC research is showing that wind and radiation recorded during large parts of the winter (when hoarfrost develops) and summer (when rime ice forms) produce barely usable and, at worst, misleading results. Research should be directed to keeping the sensors free of these deposits.

The importance of glacier calving and basal-ice melt by the sea under floating glacier tongues in the High Arctic islands is poorly understood. Examination of aerial photography suggests that calving is not a very important factor in mass balance, and basal melt under the glacier tongues is even less so. However, this is an area deserving of future research.

Acknowledgments

I am greatly indebted to reviewers of this chapter, Drs. Martin O. Jeffries, Julian A. Dowdeswell, and W.S.B. Paterson. Dr. Paterson, in particular, suggested that I undertake a major revision of the entire chapter and made many very constructive suggestions. Although I did not agree to all the reviewers' suggestions, the chapter is the better for the time and effort that they put into their technical reviews.

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