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THE GLACIATION LEVEL AND LOWEST EQUILIBRIUM LINE ALTITUDE IN THE HIGH CANADIAN ARCTIC: MAPS AND CLIMATIC INTERPRETATION

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ABSTRACT

The glaciation level (GL) over the Queen Elizabeth Islands is highest over the main mountain areas. There are extremely steep gradients approaching 15 m km^{-1} along the northwestern margin of the archipelago where the glaciation level is very low (300 m a.s.l.). Although the glaciation level mirrors topography on a gross scale, at the finer level the relationship breaks down, probably because of the effect of the mountains on precipitation patterns. There appears to be a sharp decline in the elevation of the glaciation level between the Canadian islands and northwest Greenland. The elevation of the lowest equilibrium line altitudes (ELAs) are 100 to 200 m below the GL with a minimum elevation of 200 m a.s.l. The GL represents a theoretical surface where winter net mass accumulation is equalled by summer mass ablation. The two primary controls on the elevation and gradient are, therefore, related to the pattern of winter snow accumulation and summer snowmelt. An analysis of available climatic data (one meteorological station per 100,000 km^2) is limited by the sparcity of records and the bias of existing stations to a coastal location. Nevertheless, on the shorter time scale, fluctuations in the height of the July freezing level correlate strongly with changes in glacier ELAs. However, there is little spatial correlation between decadal maps of July freezing levels and either GL or ELA surfaces.

INTRODUCTION

The concept of a theoretical surface in a glacierized region separating ice-free mountains from those which support glaciers has a long history in the glaciological literature. Østrem (1964, 1966, 1972) has been primarily responsible for reintroducing the concept after its intermittent use from the late 19th century through to the 1940s (cf. Ahlmann, 1948). The term "glaciation limit" has been used for the critical elevation that divides glacierized from nonglacierized summits, although this term has also been used to indicate the maximum vertical or horizontal extent of glaciation. Østrem (1972, p. 77) has therefore suggested that the name be changed to "glaciation level," as used by Ahlmann (1948, Figure 31). This terminology is adopted here although "glacierization level," using the glaciological term for an ice cover would be less ambiguous.

The elevation of the glaciation level over an extensive region provides (1) an integrated, regional climatic index; (2) an important measure of the state of glacierization of the region;

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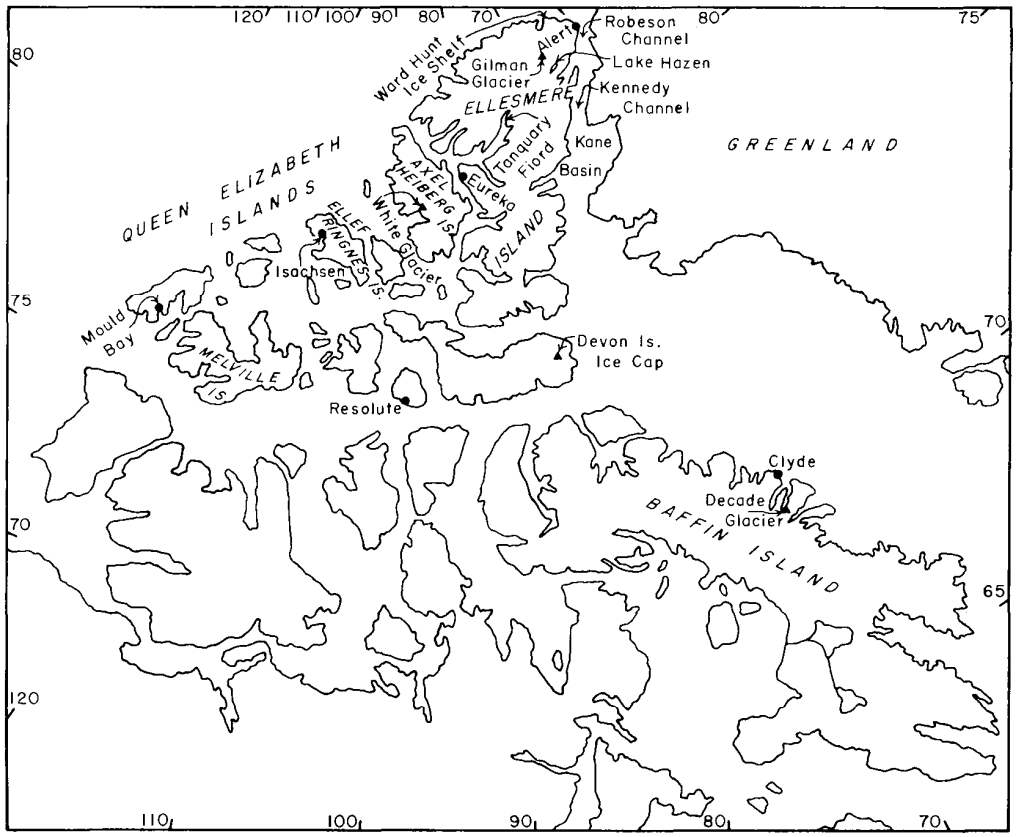


FIGURE 1. Location map for the High Canadian Arctic with place names used in this paper.

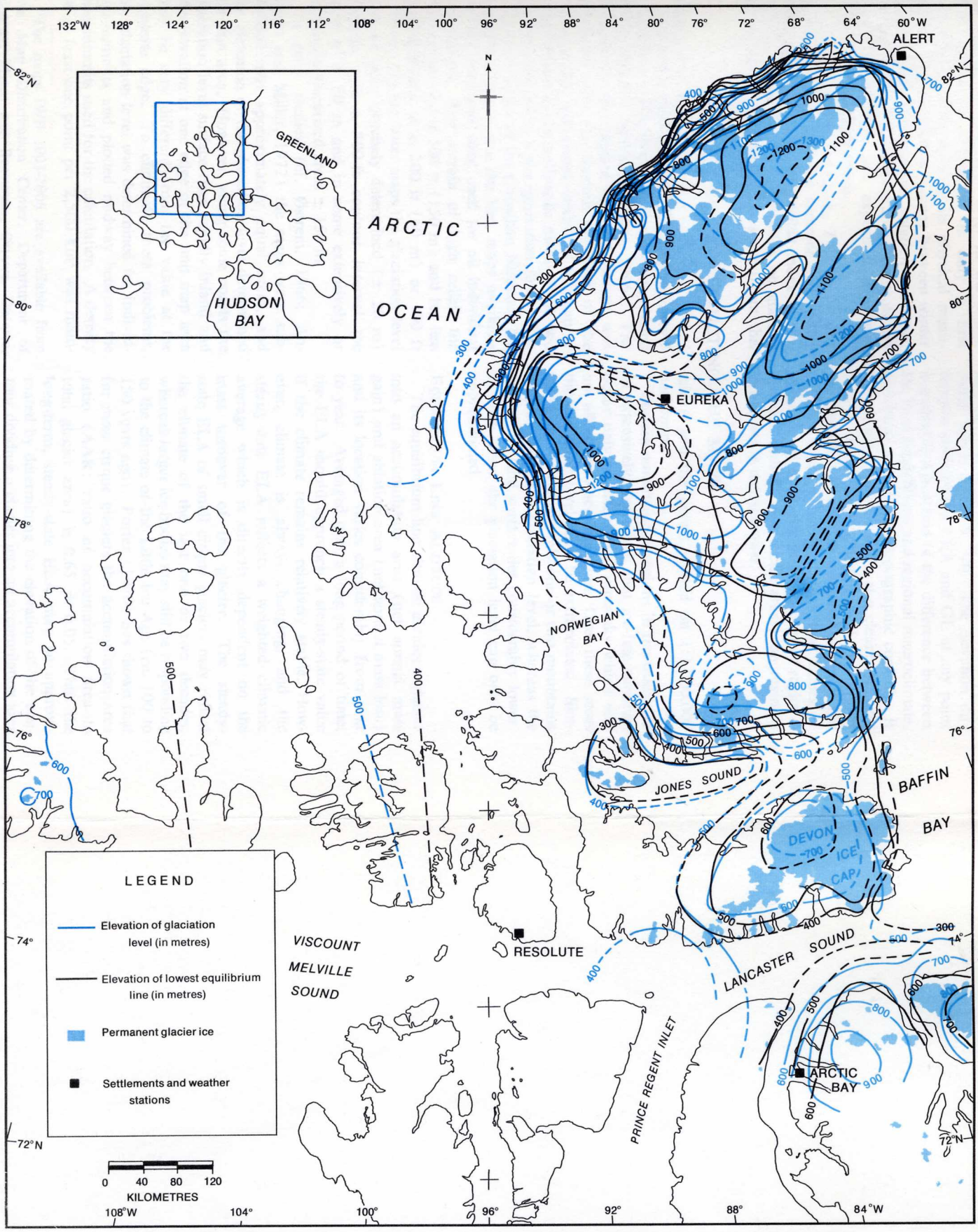
FIGURE 2 (fold-out). Map of the glaciation level and lowest ELA surfaces for the Queen Elizabeth Islands, N.W.T., Canada.

and (3) indicates the sensitivity of an area to changes in local mass balance (climatic change). However, as with other indices, the controlling physical processes need to be understood if the index is to provide information on the possible effects of some dynamic event, such as an increase in winter snowfall. Thus the glaciation level is a static representation of a complex glaciological situation; it is even probable that the glaciation level in arctic areas is not responding to present climatic conditions but to the climate of some previous decade or century (Bradley, 1972).

The following maps of equilibrium line altitudes are available: western North America—(Meier and Post, 1962); USSR—(Grosswald and Kotlyakov, 1969); Baffin Island—(Andrews and Miller, 1972); and Alaska—(Péwé and Reger, 1972). The following maps of the glaciation level are available: North Atlantic—(Ahlmann, 1948); Fennoscandia—(Østrem, 1964); southern British Columbia and Alberta—(Østrem, 1966); northern British Columbia and southeastern Alaska—(Østrem, 1972); Baffin Island—(Andrews and Miller, 1972).

PURPOSE OF THE STUDY

Our purpose was to map the glaciation level (GL) and lowest steady-state equilibrium line altitude (ELA) for the Canadian Arctic (Figures 1 and 2) covered by the 1 : 1,000,000 "Glacier Maps" IWB 1003 and IWB 1004, thus



GLACIATION LEVEL AND GLACIER EQUILIBRIUM LINE ALTITUDE.
QUEEN ELIZABETH ISLANDS, NORTHWEST TERRITORIES, CANADA.

extending similar maps for northern and southern Baffin Island (IWB 1005 and IWB 1006; Andrews and Miller, 1972).[1]

Mapping the glaciation level in the Canadian Arctic, or elsewhere, is only a first step towards understanding its physical meaning since it represents a balance between winter accumulation and summer ablation processes, modulated by local topographic controls on summit size, slope, and exposure. We may write a functional equation

$$GL = f\ (b_w,\ b_s,\ T) \qquad (1)$$

where b_w and b_s are the summer and winter balances and T is a topographic index of unknown form. The long-term ELA on glaciers lies along a line where $b_w - b_s = 0$ but this elevation will vary regionally because of regional variations in b_w and b_s and variations in local topographic conditions. The glaciation level must be related to some elevation, where, over many years, $b_w - b_s = 0$. The elevation difference between the ELA and GL at any point is primarily a function of the difference between the local topoclimate and regional macroclimate, and local and regional topographic controls. It seems that we could arrive at a clearer concept of the glaciological climatological reason for the elevation of a GL if we could eliminate the regional topographic factor. Actual values of b_w will vary primarily as a function of regional snowfall and b_s (for arctic areas) on regional variations in net radiation.

CONSTRUCTING THE MAPS

Glaciation Level

Individual calculations of the glaciation level for the Queen Elizabeth Islands were determined by the summit method (Østrem, 1966). The elevation of the highest ice-free summit was compared with the summit elevation of the lowest nearby glacierized mountain (ice cap or cirque glacier); the arithmetic mean of the two summit elevations is the glaciation level. The 1:250,000 series of the Canadian National Topographic System are the best maps available for the area and were used for all individual calculations. For terrain of high relief, the contour interval is 500 ft (150 m) and for less rugged terrain, it is 200 ft (60 m) or 100 ft (30 m). On the latter maps the glaciation level may be quite precisely determined (± 20 m) whereas with a 500-ft contour interval the error is ± 50 m and in some extensively or minimally glacerized areas ± 150 m.

In early studies (cf. Østrem, 1966; Andrews and Miller, 1972) the maps were subdivided into approximately equal areas, and the elevation of the glaciation level determined for that area. However, along some coasts the glaciation level surface rises rapidly inland and the elevation at one edge of the unit map area may be very different from the value at the opposite edge. To eliminate these problems, the glaciation level was determined for individual summits and plotted midway between the two summits used for the calculation. A density of at least one point per 2,500 km^2 was maintained and the points plotted on 1:1,000,000 base-maps.

Certain mountainous regions in the study are so extensively glacierized that ice-free summits occur only as steep-sided nunataks which will not allow snow accumulation. Over these areas the glaciation level must be extrapolated. Similarly, in a few areas only one or two mountains project above the glaciation level, whereas the rest of the land surface lies considerably lower. In such cases, the glaciation level can only be approximated.

Equilibrium Line Altitude

The equilibrium line altitude divides a glacier into an accumulation area (net annual mass gain) and ablation area (net annual mass loss) and its location varies considerably from year to year. Averaged over a long period of time, the ELA should approach a steady-state value if the climate remains relatively stable. However, climate is always changing, and the steady-state ELA reflects a weighted climatic average which is directly dependent on the mass turnover of the glacier. The steady-state ELA of small cirque glaciers may reflect the climate of the last one or two decades, whereas larger ice bodies may still be responding to the climate of the Little Ice Age (*ca.* 100 to 150 years ago). Porter (1970) has shown that for most cirque glaciers the accumulation area ratio (AAR: ratio of accumulation area to total glacier area) is 0.65 ± 0.05. Thus the long-term, steady-state ELA may be approximated by determining the elevation of the contour dividing a glacier into an accumulation area roughly twice the size of the ablation area.

[1]The maps IWB 1003-1006 are available from the Map Distribution Center, Department of Energy, Mines and Resources, Ottawa, Ontario, Canada.

Our determinations of the steady-state ELA were made on the lowest small cirque glacier in each map area (ca. 2500 km^2). This distinction was made to keep the measurements comparable over the region. However, in areas where cirque glaciers were absent, we used the lowest preferentially located small ice cap, and in extensively glacierized regions, the change of contour inflection from convex down glacier to concave was taken as the steady-state ELA. The ELA on local ice caps lies about 100 m above the ELA on adjacent cirque glaciers due to the orographic protection and precipitation enhancement for cirque glaciers, and the ELA contours therefore have an error term of $\pm$ 100 m. The ELA maps in this paper differ from those of Meier and Post (1962) and Grosswald and Kotlyakov (1969) in that the latter maps are based on firn line determinations from aerial photography, and these absolute values may not be directly comparable with those presented here, although both methods should produce similar patterns. The individual ELA estimates were plotted on 1:1,000,000 base-maps directly on the glacier for which the calculations were made. This method of plotting eliminates any significant positioning error.

Contouring

Preliminary contours were drawn on the basis of the initial data, and served to delineate areas where irregularities in the surface occurred, or where the surface gradient was very steep. For such areas, additional determinations were made to portray the variations in the surface more accurately. For the glaciation level, 249 individual calculations and 139 limiting values were obtained. Two hundred initial determinations were made for the ELA map with 149 additional points determined to clarify the surface pattern. Contouring of these points was straightforward, with anomalous points occurring only in areas of extensive glacierization. Over areas where no data are available, the contours are connected by interpolated lines. Figure 2 depicts the GL and the ELA surfaces.

RESULTS

The GL surface increases inland with the surface highest over major mountain areas (Figure 2). However, this pattern is complicated by the variability of the underlying topography and the location of large islands and seaways. Extremely steep gradients (15 m km^{-1}) occur along the northwestern margin of the archipelago.

Along the northwestern and western margin of the archipelago, the glaciation level surface is very low, but rises rapidly and uniformly inland, with the lowest values increasing to the south. A closed depression at the head of Jones Sound may be due to the persistence of open water in Cardigan Strait throughout the winter.

The surface rises regularly up the western flank of mountainous Axel Heiberg Island, but fails to mirror the topography west of the mountains over Eureka Sound. It is possible that the pattern of the glaciation level surface over this area is influenced by a precipitation-shadow effect on the eastern side of the Axel Heiberg mountains, but the necessary data to confirm or deny this hypothesis is not available except for the 1961 summer (cf. Müller and Roskin-Sharlin, 1967).

Between Devon Island and western Melville Island no glaciers exist. Over glacierized Melville Island the glaciation level surface is 600 to 700 m whereas on westernmost Devon Island the surface lies between 400 and 500 m. Between the two areas (500 km) the surface probably lies between 400 and 600 m but the land surface is everywhere below 450 m.

Along the eastern edges of the Queen Elizabeth Islands the surface is less regular than to the west. From Devon Island to the middle of Ellesmere Island coastal glaciation levels are low and increase inland. A pronounced embayment in the surface bordering Kane Basin extends to the coast despite a decreasing gradient on the elevation of the land. There may be an orographic control on the surface as the land, although not exceptionally high, is extensively dissected, unlike coastal areas to the west. Why the GL should be dependent upon the degree of land dissection is, however, unclear.

The ELA map is remarkably similar to the GL map, with the ELA surface, in general, lying 100 to 200 m below the GL surface. The major features on the GL map are present on the ELA map as well, with very steep gradients in the north and northwest, diminishing to the south. ELAs are lowest along coastal margins, and are highest over mountainous terrain. The lowest ELA values decrease from 400 m on northern Baffin Island to less than 200 m on northern Ellesmere Island, and must lie close to sea level in the vicinity of the Ellesmere ice shelves.

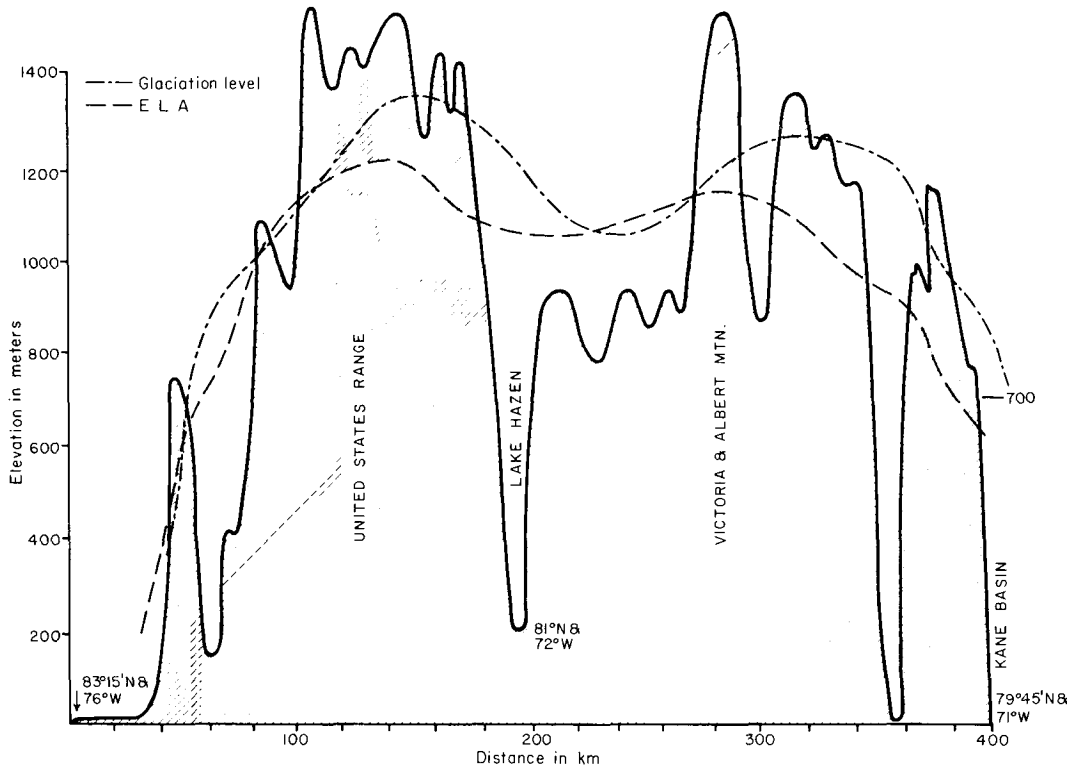


FIGURE 3. Cross section from Ward Hunt Island (83°15'N, 76°W) to John Richardson Bay (79°45'N, 71°W), northern Ellesmere Island, showing the variation in topography, glaciation level, and ELA surfaces.

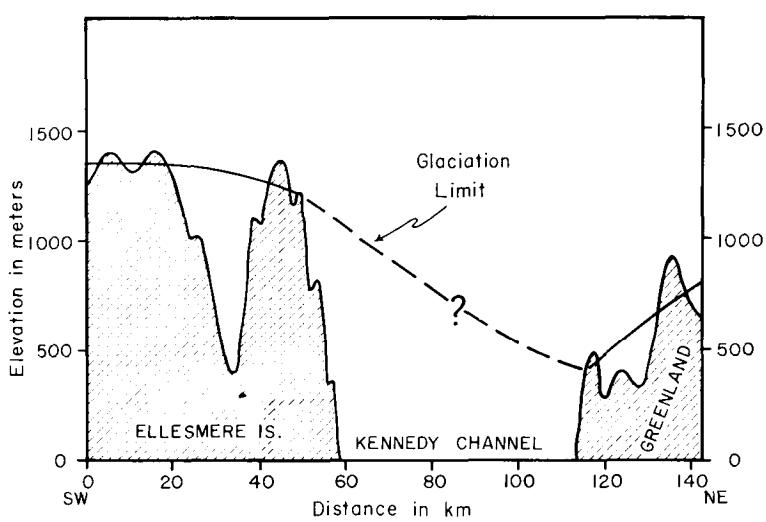


FIGURE 4. Cross section southwest - northeast, centered at 80°45'N and 67° W between Ellesmere Island and Greenland showing lowering of glaciation level across Kennedy Channel.

The north-south cross section shown in Figure 3 illustrates the general parallelism between the GL, ELA, and land surfaces, and shows an apparent trend toward coincidence of the GL and ELA surfaces in the north. This trend is apparent along the entire glacierized eastern

margin of Canada. In Labrador the ELA lies 450 m below the GL (J. T. Andrews, unpublished ms.); on southern Baffin Island it is 400 m below and decreases to 200 m below the GL over the northern part of the island, whereas on northern Ellesmere Island the surfaces are nearly coincident. The general trend of the two surfaces to approach each other at higher latitudes reflects the decreasing orographic influence on the ELA.

A southwest-northeast cross section between northeastern Ellesmere Island and adjacent Greenland (Figure 4) illustrates the dramatic lowering of the GL surface across Kennedy Channel. The GL at the eastern edge of Ellesmere Island is above 1,200 m, whereas on Greenland, less than 100 km away, glacierized hills occur below 500 m. A major climatic gradient across Kennedy Channel is difficult to conceive, and topography may be the controlling factor, although without climatic data this problem cannot be resolved. The very slow rates of mass turnover on polar glaciers may further complicate interpretation as some may still be responding to a climate of the last century.

CLIMATIC INTERPRETATION

Glaciation Level Maps

The glaciation level is a time-transgressive surface reflecting both climatic and topographic controls. In the Canadian Arctic, climatic records rarely exceed 25 years; hence any interpretation on the basis of "present" climate must assume that the climate to which the glaciation level is a response was similar to that of today. Clearly this assumption can be questioned as considerable fluctuations in temperature and precipitation have occurred over the past century (Hattersley-Smith, 1963; Bradley and Miller, 1972; Bradley, 1973a). However, it is assumed here that while absolute values may

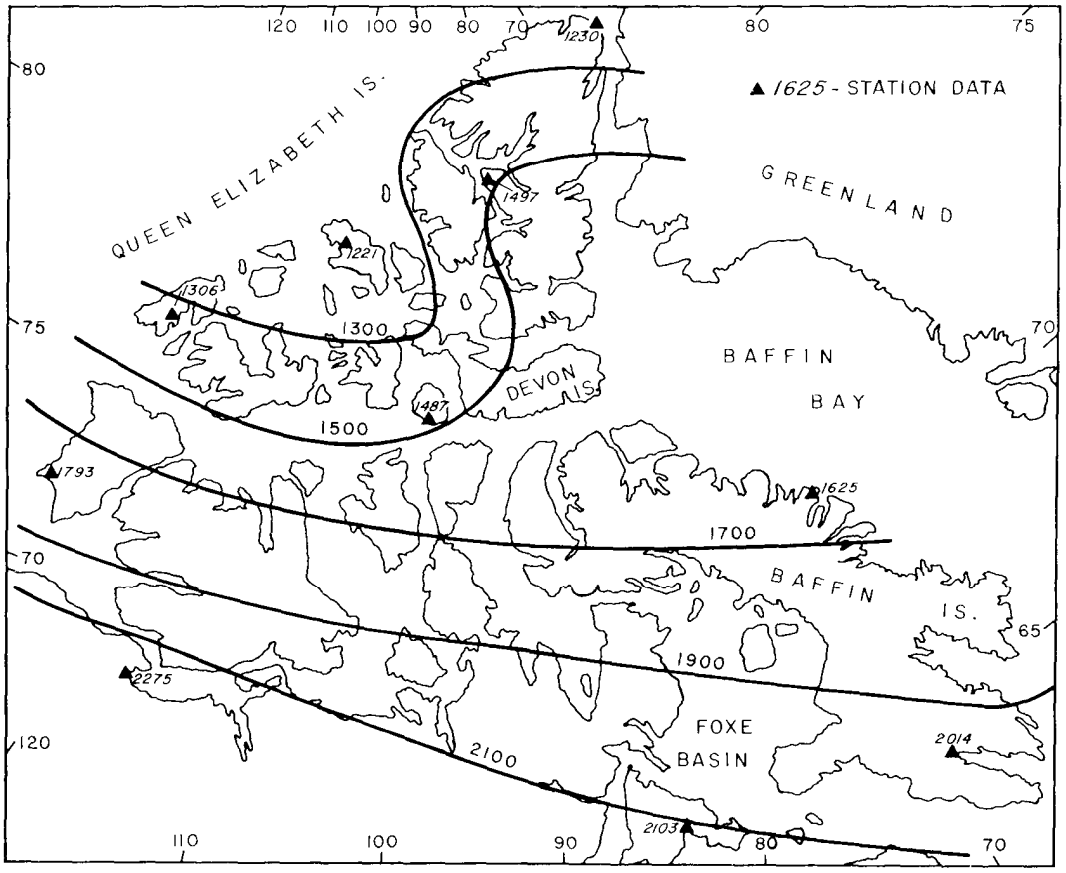


FIGURE 5. Average July freezing level 1955 to 1963 in m a.s.l.

have changed, the broad pattern of climate has been similar to that observed over the past 20 to 25 years.

It was pointed out in the introduction that the GL is a function of summer and winter balances and of topography. The GL is an integrated response to seasonal climatic variations and any climatic interpretation is faced with the problem of distinguishing between the major controls. This would be a difficult task even in an area where widespread climatic records are kept, but the problem is compounded in the Canadian Arctic where there is, on average, only one meteorological station per 100,000 km^2.

Flint (1971) states that "a large number of measurements show that the altitude of the regional snowline coincides with, or is slightly lower than, the summer (or July) mean O°C isotherm." Other work (e.g. Østrem, 1966, 1972) has indicated that winter precipitation is a major control. There is thus a dichotomy of opinion between those who favor summer balance as critical and those who favor winter balance although there is no *a priori* reason for believing that only one of these positions is always true. A great deal depends on the regional variation in accumulation and summer conditions. If Flint's statement is correct, then glaciation levels in the Canadian Arctic Archipelago should reflect, broadly, the pattern of July freezing level (FL) heights (cf. Titus, 1968). Figure 5 shows the mean July freezing level for 1955 to 1963 based on an analysis of daily upper air soundings taken at 0000 GMT (0350 GMT in 1955). In the western archipelago, a fairly constant decrease in elevation with latitude is observed, with the average freezing level height decreasing from approximately 2,500 m on the mainland coast to about

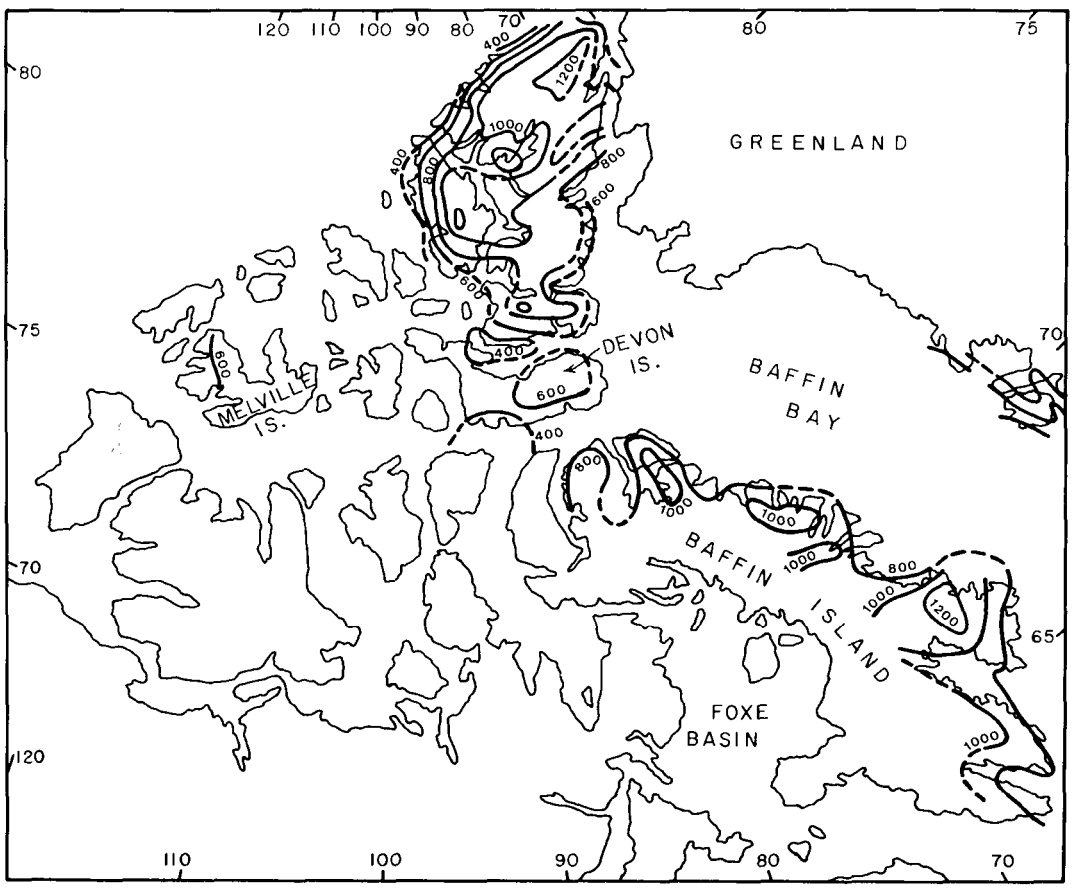


FIGURE 6. Generalized glaciation level map for the Canadian Arctic Archipelago. Contour interval 200 m.

1,200 m in the vicinity of Ellef Ringnes Island. In the eastern archipelago, however, higher freezing level elevations for a given latitude are observed, reflecting the influence of warm air advection into the eastern region via Baffin Bay. A similar pattern is observed for summer months of 1955 to 1963 and during the period 1964 to 1972, though absolute values are considerably different (Bradley, 1973b, 1975).

Figure 6 shows a generalized map of the GL for eastern North America: a comparison of Figures 5 and 6 shows there are few similarities between GL and FL heights in July. Large parts of the Baffin coastal mountains have GLs similar in elevation to northern Ellesmere Island (>1,000 m) and higher than on Devon Island. Figure 6 does show, however, that minimum glaciation levels on Devon, Ellesmere, and Axel Heiberg islands are lower than on Baffin Island. On Baffin Island glaciation levels are nowhere less than 500 m whereas on Devon, Ellesmere, and Axel Heiberg islands GLs are found as low as 300 m. Further, minimum GLs are also lower on western Ellesmere and western Axel Heiberg islands (down to approximately 200 m on Meighen Island) compared with eastern Ellesmere and Devon islands (475 m). It appears, therefore, that glaciation levels are lower in the region where freezing levels also tend to be lowest (viz. the northwestern archipelago), but in detail, the overall spatial correlation is poor. One problem may be that freezing level maps are based on only a few points compared to the detailed mapping of the glaciation level. However, upper air patterns are fairly coherent and are unlikely to show great spatial variations. Certainly, if the July (or summer) FL was the single major control on the GL, then a stronger broadscale relationship might be expected. For the upper air stations in the eastern archipelago the differences between GL and FL heights in July are shown in Table 1. Although there are few points for comparison, it is clear that the July FL is considerably higher than the GL in the south and much less so in the north. Again this suggests that FL height is not the single major control on the GL.

Østrem (1972) has mapped the GL for British Columbia and southeast Alaska and shows there is an inverse relationship between winter precipitation and the glaciation level. Depressions in the general GL surface are related to local influences of increased winter precipitation. This supports his earlier hypothesis (Østrem, 1966) that precipitation is a major control on the GL. Unfortunately, there are few data to investigate such a hypothesis in the Canadian Arctic. Figure 7 shows the average water equivalent of snow cover on March 31 for all of Canada (B. F. Findlay pers. comm., 1974). A similar pattern to the July freezing level map is observed with lowest values over the northwestern archipelago. Much higher values are found over Labrador-Ungava. However, it should be noted that nearly all stations recording precipitation north of 60°N are coastal, at or close to sea level. The relatively high values south of Knob Lake may only reflect the much higher station elevations in this area (Schefferville, 513 m; Nitchequon, 515 m) compared to stations farther north. If Østrem's hypothesis applies to the Arctic islands, then areas having the highest glaciation levels would experience lowest (winter) precipitation. It should be noted, however, that in the High Arctic, summer precipitation accounts for ca. 40% of the annual totals (Jackson, 1961) and thus must be considered as relevant to the problem of interpreting the glaciation level.

TABLE 1

Comparison of glaciation level (GL) and mean July freezing level (FL) (1955 to 1963) at sites of upper air stations

Station	July FL (m)	GL (m)	Difference in elevation (m)
Frobisher	1,970	1,000	970
Clyde	1,593	840	753
Eureka	1,470	1,100	370
Alert	1,205	700	505

The highest glaciation levels are found over the mountainous interior regions of Axel Heiberg and northern Ellesmere islands. Two gradients must therefore be considered; distance from the coast and elevation, both of which increase inland. Unfortunately, in the arctic islands the only precipitation data along either gradient are from glaciological studies at widely scattered points. Table 2 lists summer precipitation for 1961 in the area of western Axel Heiberg Island (Müller and Roskin-Sharlin, 1967). Precipitation shows an altitudinal increase in the first 200 m but does not change noticeably above that elevation (cf. 21.9 cm at 208 m and 23.2 cm at 1,920 m). In contrast, Figure 8 shows average specific accumulation derived from firn studies at three sites; two on

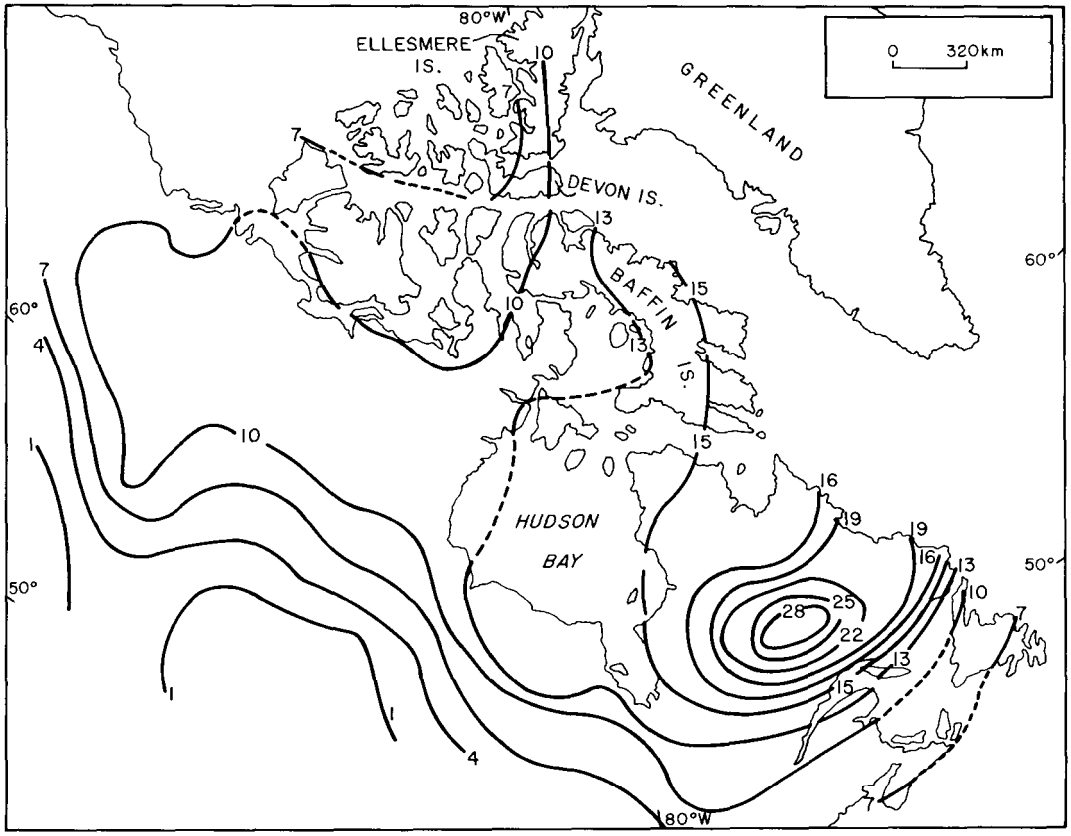


FIGURE 7. Average water equivalent of snow cover on March 31 (cm) from B. F. Findlay (pers. comm., 1974).

McGill Ice Cap, Axel Heiberg (Müller *et al.*, 1963), and one on Gilman Glacier, Ellesmere Island (Hattersley-Smith, 1963). Although Gilman Glacier is far from Axel Heiberg, these data suggest that precipitation increases with elevation at a rate of approximately 6 cm · 100 m^{-1}. In addition, winter balances at the two McGill Ice Cap sites (Upper Ice I and II) have been obtained for two winters; 1960-61 and 1961-62 (Müller *et al.*, 1963) and these values are also plotted on Figure 8, and show a similar accumulation gradient. Despite its inland position, the Gilman Glacier accumulation does fit in fairly well with an interpolated value based on the two McGill Ice Cap sites. Table 3 lists winter balance by altitudinal zones on the White Glacier, Axel Heiberg Island (Müller *et al.*, 1963). Although values fluctuate considerably, a general increase in precipitation with elevation is shown.

The data suggest that precipitation does increase with elevation over small distances. How-

TABLE 2

Total precipitation (June 1 to August 25, 1961)[a]

Station	Elevation (m)	Precipitation (m)
Base Camp	190	15.7[b]
Lower Ice	208	21.9[b]
Upper Ice II	1920	23.2[b]
Eureka	7	5.7[c]
Isachsen	30	11.9[c]
Meighen Ice Cap	241	12.5[c]

[a]Data from Müller and Roskin-Sharlin, 1967, p. 49.

[b]Snow course, adjusted for snow density in cm w.e.

[c]Water equivalent.

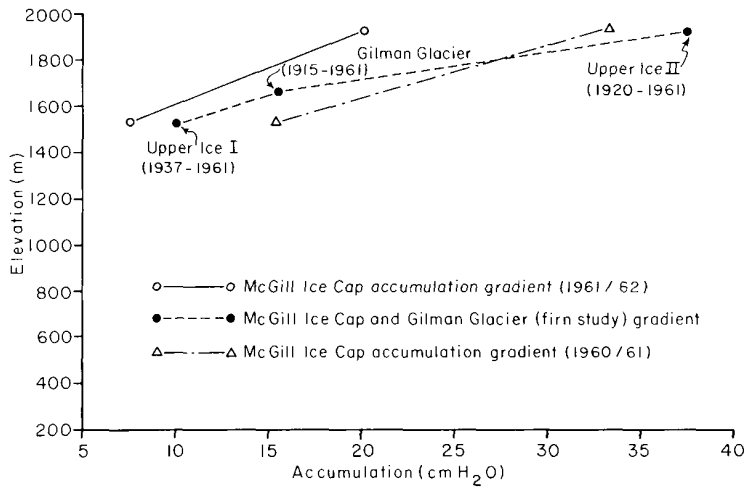


FIGURE 8. Accumulation-elevation relationships, McGill Ice Cap, western Axel Heiberg Island, and Gilman Glacier, central Ellesmere Island. Data from Hattersley-Smith (1963) and Müller *et al.* (1963).

ever, these data are not sufficient to satisfactorily investigate the question of coastal versus inland variations in precipitation. Although some data from coastal sites, such as Isachsen and Alert, are available, the glaciation level at these points can only be given as minimum estimates. No precipitation data at all are available from northwestern Ellesmere Island where comparison with an inland site might shed some light on the problem. It is interesting to note that winter balance on the Ward Hunt Ice Rise, northern Ellesmere Island, 1958-59 to 1968-69 averaged 18.7 cm w.e. (Hattersley-Smith and Serson, 1970) compared to an average mean annual accumulation 140 km inland on the Gilman Glacier (1915 to 1961) of only 15.9 cm (at 1,600 m). However, precipitation at Lake Hazen (161 m) from August 1957 to July 1958 was only 2.5 cm compared to 11.5 cm at Alert and 5.3 cm at Eureka for the same period. This suggests a very marked decrease of precipitation inland for sites of similar elevation. A study of winter accumulation along a transect from the coast of northwestern Ellesmere Island inland to Lake Hazen would add greatly to our knowledge of climate in this region and would be of considerable value in interpreting the GL.

In conclusion, the climatic controls on the GL surface are not easily resolved and are masked by the effect of local topographic controls. The July FL height does not adequately explain the GL pattern, and although precipitation may be a major control, the available evidence is not adequate to resolve this hypothesis. Precipitation does seem to increase with elevation over short distances but may decrease along a gradient inland from the coast. Such a gradient would have to be greater than the altitudinal precipitation gradient to explain the pattern of glaciation levels. Further, the coastal/inland gradient would have to be very steep (i.e. precipitation decreasing rapidly) across the northwestern coast of Ellesmere Island to explain the rapid increase in the glaciation level which in some areas rises 400 m in less than 10 km. This steep glaciation level gradient may be related to the presence of local coastal fogs and/or persistent temperature inversions, or to frequent subsynoptic scale events which have not yet been identified. Such events have been noted in Tanquary Fiord, central Ellesmere Island (Barry and Jackson, 1969). Meteorological observations from northwestern Ellesmere Island are needed before this problem can be solved.

TABLE 3

Net specific winter balance accumulation, White Glacier, Axel Heiberg Island[a]

Elevation range (m)	Winter balance 1959-60	1960-61	1961-62
1,400-1,500	16.5	35.8	3.3
1,300-1,400	18.5	29.1	1.4
1,200-1,300	14.8	30.6	
1,100-1,200		23.8	
1,000-1,100		23.5	
900-1,000		17.3	

[a]Data are given in cm water equivalent from Müller *et al.* (1963, p. 21).

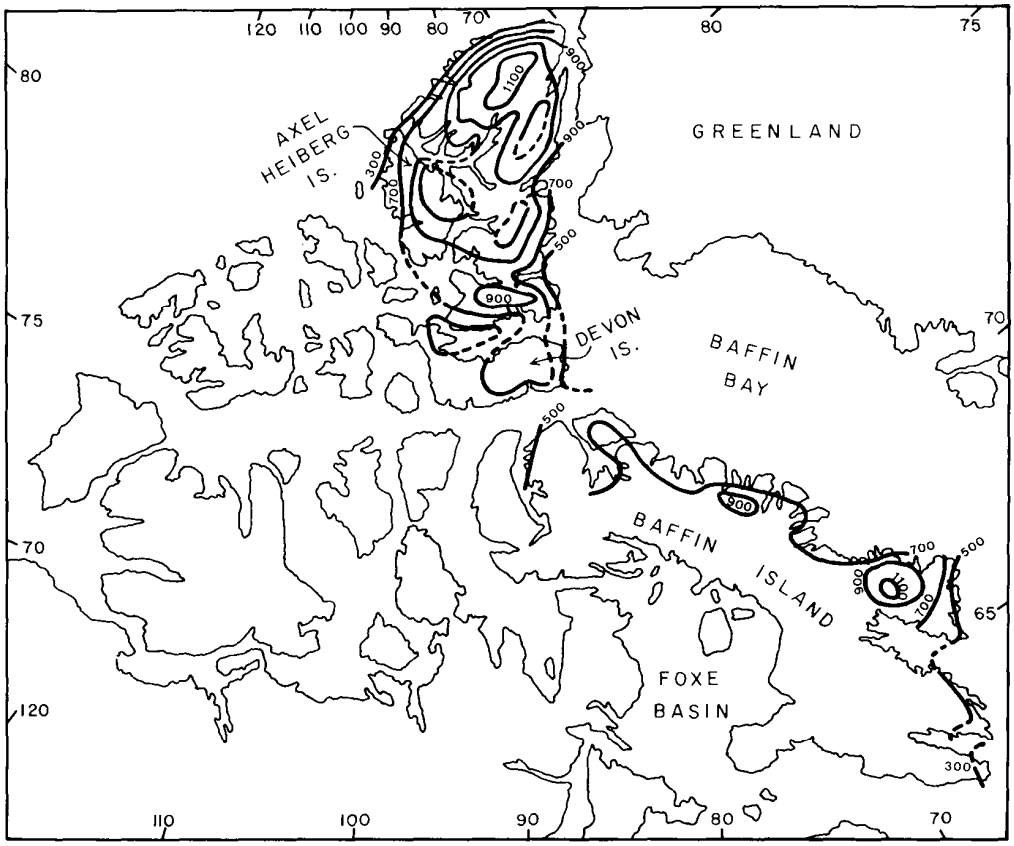


FIGURE 9. Generalized steady-state ELA map for the Canadian Arctic Archipelago. Contour interval 200 m.

EQUILIBRIUM LINE ALTITUDE MAPS

A generalized lowest steady-state equilibrium line altitude map for the arctic archipelago is shown in Figure 9. The pattern is very similar to the glaciation level map, and many of the comments regarding the climatic controls on the glaciation level are pertinent to the equilibrium line altitude. Although the GL and steady-state ELA do not vary much from year to year, the annual ELA varies in response to changes in the seasonal balances. On the White Glacier, Axel Heiberg Island, and the northwest sector of the Devon Ice Cap, ELAs have been measured for a number of years (Müller, 1966, 1967, 1968, 1970; Koerner, 1970). Correlation coefficients between ELAs on the White Glacier and various climatic variables at Eureka and Isachsen (the nearest permanent weather stations) strongly suggest that summer balance is the critical control on ELAs with freezing levels at Eureka "explaining" 94% of the variance (Bradley, 1975). Figure 10 shows the graph of Eureka freezing levels in July against ELAs on the White Glacier. It is interesting to note that the mean ELA for the 9-year period is 905 m with a corresponding mean July freezing level at Eureka of 1,220 m (i.e. 315 m above the ELA). Murray (1952, 1959) has shown that at an elevation of approximately 250 m below the freezing level the probability of precipitation falling as rain or snow is 50%. When the freezing level is lower, the percentage of snowfall increases and vice versa. A similar relationship has been established for the Canadian Arctic by R. G. Barry (pers. comm., 1973). Thus the relative proportions of summer precipitation falling as rain or snow on the glacier may be important to the height of the ELA.

ELAs on the northwest sector of the Devon Ice Cap (Koerner, 1970) have also been com-

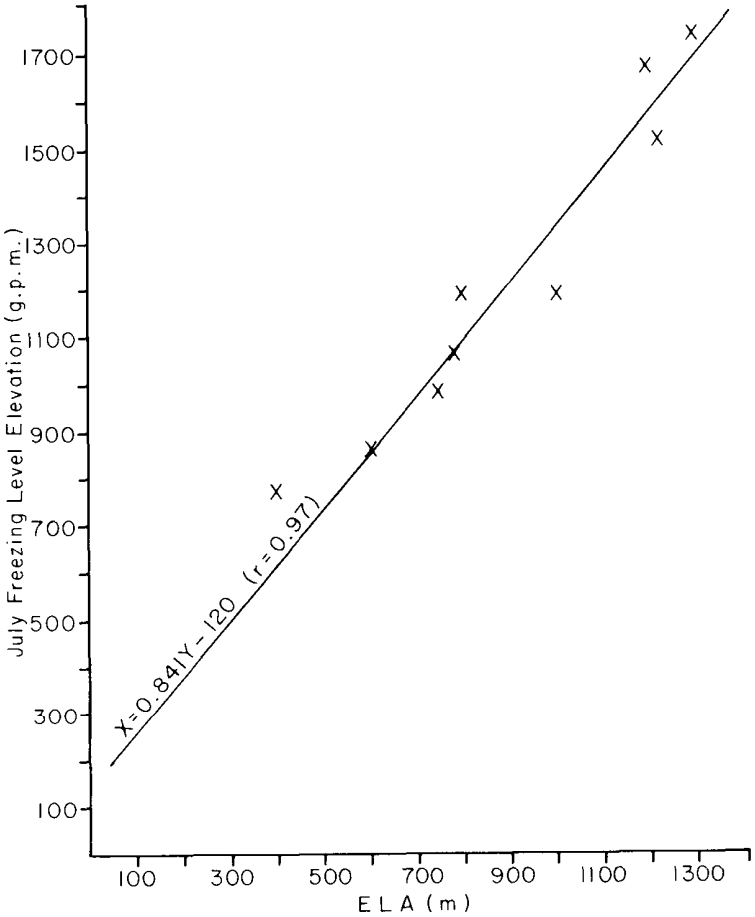


FIGURE 10. Relationship between the ELA on the White Glacier, Axel Heiberg Island, 1959-60 to 1968-69 (excluding 1967-68) and Eureka freezing level. Data from Müller (1966, 1967, 1968, 1970).

pared to mean July freezing levels at Clyde and at Resolute. Again very high correlation coefficients were obtained (0.96 and 0.90, respectively; significant at the 1% level) and very similar slopes were given (Bradley, 1975). The July freezing level at Resolute was, on average, 312 m above the ELA, a figure close to that for the White Glacier ELA and freezing level at Eureka in July. At Clyde, 750 km southwest of the Devon Ice Cap, the freezing level is on average 482 m above the ELA. Only one other ELA record of useful length is available for the arctic islands—that of the Decade Glacier, eastern Baffin Island. This record shows virtually no relationship to July freezing level nor winter or summer precipitation at Clyde, only 120 km away. On the basis of this limited sample, all that can be said is that the ELA on high latitude glaciers may be controlled mainly by summer ablation, which is greatly dependent on freezing level height. Farther south the relationship appears to be more complex and may not be easily explained.

It is clear that the relationship between climate and the glaciation level and even steady-state ELAs in the arctic islands is by no means a simple relationship, as is suggested by studies elsewhere (e.g. Østrem, 1972). The glaciation level may indeed be an "integrated regional climatic index" but with our present knowledge of the climate of the Arctic it is impossible to say just what climatic information it reflects. There is good evidence that the yearly ELA in two areas is closely related to the July freezing level height but that has been of little help in explaining the broad-scale ELA patterns. Perhaps our basic assumption, that the climate to which the present glaciation level is a re-

sponse had a similar pattern to that of today, is incorrect. If we assume that the glaciation level we are observing today is essentially reflecting the glaciological response to a climate 100 to 150 years ago (i.e. to the Little Ice Age) then it is possible that the major controls on the glaciation level in the past were not the same as they are currently. Thus if more precipitation was carried into high latitudes (particularly from the west) and the coastal/inland precipitation gradient was much greater than today then this could explain the high glaciation level in the mountainous interior of Ellesmere and Axel Heiberg islands compared with the coastal regions. At present this gradient may be insignificant and not reflect the true causative controls on the present glaciation level distribution. Similarly, whereas at present the major control on the ELA might be the July freezing level elevation, in the past the ELA may have been controlled by other factors. One thing is clear in trying to interpret the glaciation level in climatic terms; our knowledge of the present climate of the arctic islands and particularly of the mountain regions is inadequate. Precipitation data are particularly sparse for high inland sites. Only with more field data will a meaningful climatic interpretation be possible.

CONCLUDING REMARKS

The glaciation level and ELA surfaces are necessarily time-transgressive and may not be strictly in phase with current climatic conditions. Ice masses, whose margins at low elevations are sustained by the continuous movement of ice from the accumulation zone down into the ablation zone have a time lag in their response to climatic change. The larger the accumulation area, and the greater distance for ice transport to the periphery, the longer the response time. For large ice caps 50 to 150 years may elapse before a climatic fluctuation is felt at the margin, whereas smaller cirque glaciers may require only 5 to 10 years to respond to the same change. On the other hand, snowbanks and thin ice caps whose perimeters are governed solely by direct precipitation respond essentially concurrently with climate. The time lag will provide an average over anomalous balance years, but the variation in response times of different sized glaciers may mask or amplify the trends of the *true* present glaciation level.

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