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RECENT CLIMATIC FLUCTUATIONS OF THE CANADIAN HIGH ARCTIC AND THEIR SIGNIFICANCE FOR GLACIOLOGY

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ABSTRACT

Various measures of the character of ablation season conditions in the Canadian High Arctic (north of 74°N) are discussed based on an analysis of daily climatic data from Alert, Eureka, Isachsen, Resolute, and Thule. Melting degree day totals appear to be the most useful index of "summer warmth." An abrupt change in the summer climate of the region occurred around 1963/64. Various indices indicate a marked decrease in summer temperature after 1963. During the same period, annual precipitation in the north and northwest has increased.

Glacier mass balance is strongly controlled by summer climate; in particular, annual melting degree day totals are highly correlated with long-term mass-balance records. This enabled mass balance on the northwest sector of the Devon Island ice cap to be reconstructed back to 1947/48. Cumulative mass losses on the Devon Island ice cap from 1947/48 to 1962/63 are estimated to be ~ 3500 kg m^{-2}. However, from 1963/64 to 1973/74 a total of < 350 kg m^{-2} have been lost. Significant ice-cap growth is presently limited by low precipitation even when mean summer temperatures are very low; an occasional warm summer may therefore obliterate cumulative mass gains over many years.

The post-1963 change in summer climate appears to be related to the massive increase of volcanic dust in the upper atmosphere, primarily due to the eruption of Mt. Agung (March 1963). Subsequent eruptions may have caused the cooler conditions to persist. Volcanic dust affects solar radiation receipts and perhaps also influences the general circulation. If the high volcanic dust levels of the 1960s are responsible for reduced mass losses on High Arctic glaciers and ice caps, it is probable that other periods with high atmospheric dust levels (e.g., 1750 to 1880) had summer temperatures at least as cold as the mid to late 1960s. Conversely, the period of very negative balance on the Devon Island ice cap from 1947 to 1963 was probably typical of the period back to 1920 when the atmosphere was relatively free of volcanic dust.

INTRODUCTION

The Canadian High Arctic (north of latitude 74°N) contains the greatest concentration of land-based snow and ice outside of Greenland and Antarctica (Figure 1). Glaciation levels vary from over 1200 m above sea level on north-central Ellesmere Island to less than 300 m a.s.l. along the coast of northwestern Ellesmere Island where extensive ice shelves are presently found in sheltered fiord mouths and embayments (Miller *et al.*, 1975). Prior to 1946 the climate of this area was virtually unknown with only scattered records available from early expedition memoirs (e.g., Nares, 1878; Greely, 1888). The most complete records are those of Greely's ill-fated expedition to Fort Conger where synoptic observations were kept continuously for two

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years (August 1881 to July 1883) as part of the United States contribution to the First Polar Year. During this period daily minimum temperatures as low as -52.3°C were recorded and the daily maxima never rose above 11.7°C.

Although Greely's observations were a great contribution to polar meteorology, they were limited in duration and geographic extent. It was not until the end of World War II that a network of meteorological stations, the Joint Arctic Weather Stations (JAWS) was estab-

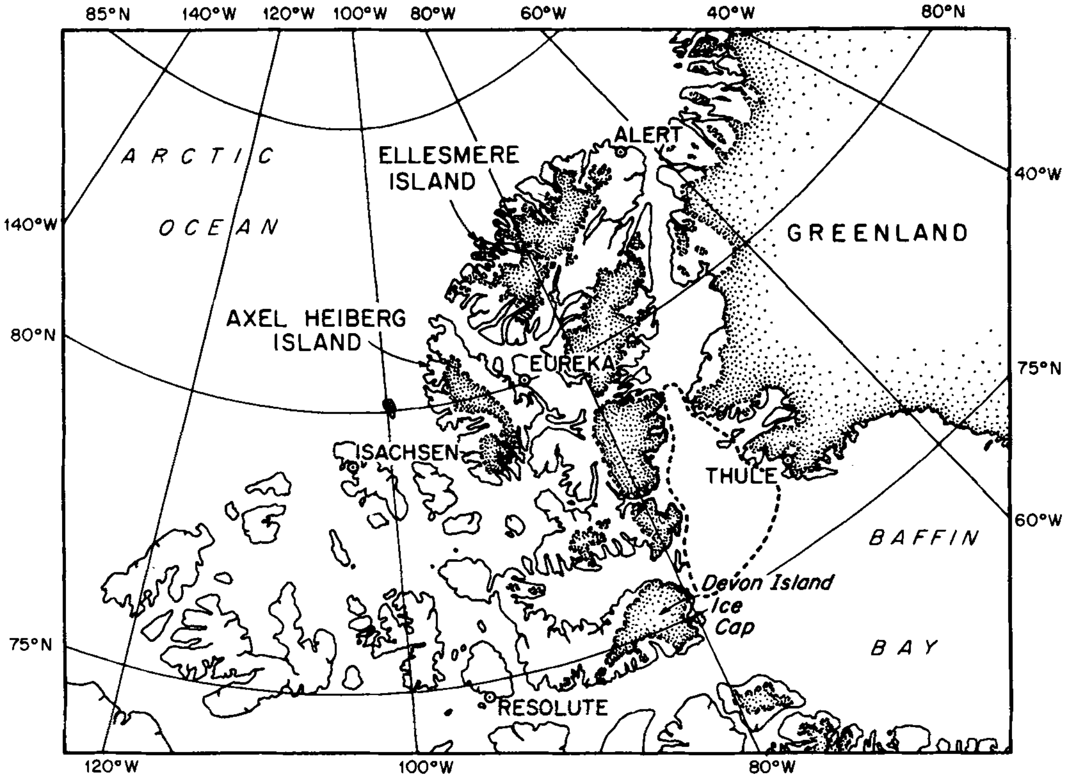

FIGURE 1. Location of principal weather stations. Ice caps are stippled. Dashed line is the average North Water extent in March (after Dunbar, 1969).

TABLE 1
Long-term High Arctic weather stations

Station number[a]	Station	Latitude (N)	Longitude (W)	Elev. (m)	Record starts	Record eds[b]
2400300	Alert	82°30′	62°20′	63	June 1950	
2401200	Eureka	80°00′	85°56′	2	May 1947	Aug 1963
2401200	Eureka	80°00′	85°56′	10	Aug 1963	
2502600	Isachsen	78°47′	103°32′	25	May 1948	
2403500	Resolute	74°41′	94°55′	17	Oct 1947	Oct 1953
2403500	Resolute A	74°43′	94°59′	64	Oct 1953	
17602-W	Thule	76°33′	68°49′	38	Oct 1946	July 1952
17605-A	Thule	75°32′	68°45′	60	Oct 1951	

[a] Station numbers are Canadian Atmospheric Environment Service index numbers except for Thule which is a U.S. Weather Bureau Station (not part of the JAWS network).

[b] Where station relocation occurred; otherwise station has operated uninterrupted at the same location to the present day.

lished in the High Arctic under the cooperative sponsorship of the United States and Canadian governments (Buss, 1971; Table 1). Unfortunately, these stations are widely spaced (Figure 1) and they are all located close to sea level. The climate of the mountainous interiors of the islands remains virtually unknown except for limited observations compiled by glaciological expeditions, primarily during the ablation season (cf. Sagar, 1960; Lotz, 1961; Havens, 1964; Müller and Roskin-Sharlin, 1967; Holmgren, 1971). Although the primary purpose of the JAWS network was to provide synoptic observations (surface and upper air) for aviation purposes, their continuous operation for almost 30 yr has provided a basis for looking at the broad-scale climate and climatic fluctuations of the area and the importance these fluctuations may have for the mass balance of snow and ice bodies in the region.

TEMPERATURE

THE ABLATION SEASON

In the High Arctic, where temperatures are well below freezing for most of the year, the character of the ablation season is of critical importance to glacier mass balance. A common index of ablation season conditions is the mean maximum temperature (T_{max}) of the warmest month, July. Lotz and Sagar (1962) showed that ablation on the Gilman Glacier (north-central Ellesmere Island) was closely related to daily maximum temperatures, particularly in July when 75% of ablation occurred (Sagar, 1960). Furthermore, Bradley (unpublished data) found that mass balance and equilibrium-line elevations on the White Glacier, Axel Heiberg Island, are closely related to July maximum temperatures at Eureka and Isachsen. Although at all stations the year-to-year variability of July temperature is high, careful examination of the records shows that temperatures fell markedly in the early 1960s, generally around 1963/64 (e.g., Figure 2). This corresponds to abrupt changes in July freezing-level heights throughout the Canadian Arctic Archipelago; freezing levels averaged as much as 500 m lower in the decade following 1963 than in the preceding decade (Bradley, 1973a). Table 2 shows that the actual changes in July T_{max} at each station ranged from 1.1 to 2.7°C, and that at

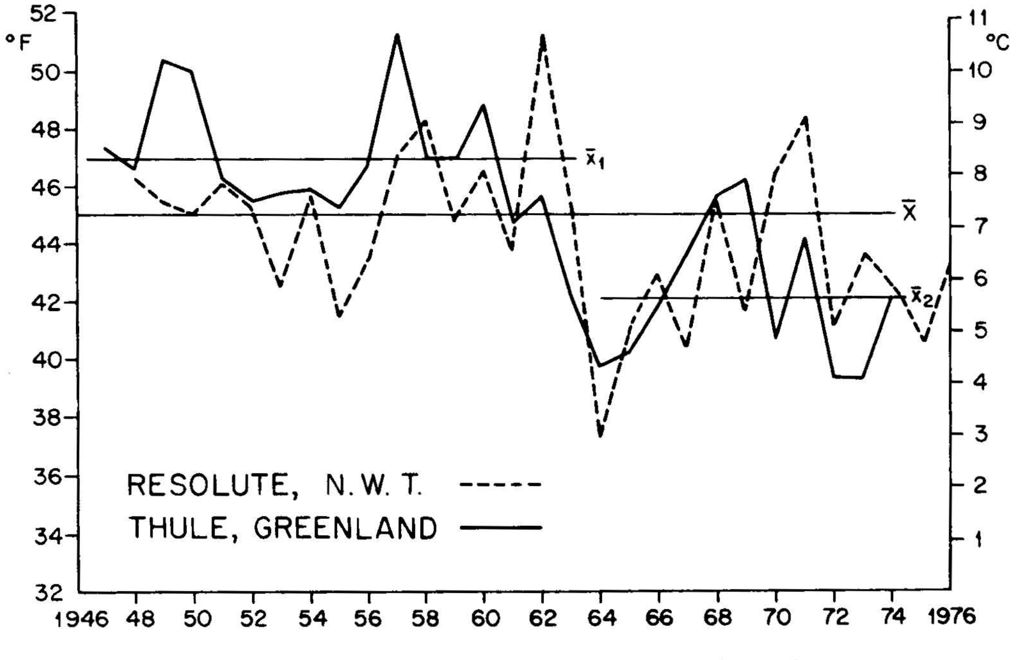

FIGURE 2. Mean July maximum temperatures at Thule and Resolute. Line $\bar{x}_1$ denotes average temperature at Thule up to 1963; $\bar{x}_2$ denotes average 1964 to 1974. $\bar{X}$ is overall mean at Thule.

TABLE 2

Change in mean July maximum temperature (°C)

Station	Period	Start of record to 1963 (*A*)	s_A	1964 to end of record (*B*)	s_A	Change (*B*-*A*)
Alert	1950-76	7.2	2.4	5.9	1.4	-1.3
Eureka	1947-76	8.9	1.4	7.8	1.2	-1.1
Isachsen	1948-76	6.3	2.1	4.8	1.6	-1.5
Thule	1947-74	8.3	1.2	5.6	1.4	-2.7
Resolute	1948-76	7.5	1.3	6.0	1.6	-1.5

TABLE 3

Change in mean July absolute maximum temperature (°C)

Station	Period	Mean, start of record to 1963 (*A*)	s_A	Mean, 1964 to end of record (*B*)	s_B	Change, (*B*-*A*)
Alert	1950-76	15.8	2.6	13.5	2.3	-2.3
Eureka	1947-74	15.3	2.7	13.1	2.6	-2.2
Isachsen	1948-76	13.7	3.7	11.2	2.9	-2.5
Thule	1947-74	14.0	1.6	12.6	2.0	-1.4
Resolute	1948-76	13.9	2.1	12.2	2.4	-1.7

TABLE 4

Average duration of melt season[a]

Station	First of 2 consecutive $T_{max} > 0^\circ$C	s (days)	Latter of 2 consecutive $T_{max} > 0^\circ$C	s (days)	Average length of melt season (days)	s (days)
Isachsen	7 June	6	31 Aug.	7	86	10
Alert	2 June	11	1 Sep.	9	92	16
Resolute	4 June	9	7 Sep.	9	96	13
Eureka	30 May	10	6 Sep.	8	100	12
Thule	21 May	12	20 Sep.	9	123[b]	13

	Extreme Length of Season[c]	
	Longest (days)	Shortest (days)
Isachsen	101 (1957)	70 (1956, 1967, 1972)
Alert	120 (1968)	66 (1956)
Resolute	120 (1952)	69 (1956)
Eureka	126 (1955)	76 (1956)
Thule	147 (1952)	90 (1961)

[a] Averages for period of record shown in Table 1.

[b] An average of 5 to 6 additional days per year experienced $T_{max} > 0^\circ$C (ranging from 0 to 14 additional days per year).

[c] Year is given in parentheses.

the northernmost stations interannual variability was generally lower in the period after 1963. Further analysis of daily data shows that this change in mean July T_{max} was strongly influenced by changes in the frequency of extremely warm days. At Alert, for example, there were 32 days with daily maxima $\geq 15.5°C$ during the period 1950 to 1963 (an average 2.3 day yr^{-1}) whereas in the period 1963 to 1976 there have been only 4 such days in total. At Isachsen, there were an average of 8 day yr^{-1} from 1948 to 1963 when daily maxima were $\geq 10°C$; from 1964 to 1976 there were less than 4 such days per year, on average. At all stations, average absolute monthly maxima in July have fallen since 1963 by up to 2.5°C (Table 3).

Although these data indicate a significant change in the temperature of the warmest month of the year, it should not be assumed that this aptly summarizes the entire picture of ablation-season conditions, because the total melt season duration may vary widely from year to year.

LENGTH OF THE MELT SEASON

Although temperatures $>0°C$ may occasionally be recorded in mid-winter months during periods of extreme warm air advection (Thomas and Titus, 1958) daily records indicate that the melt season (when daily maxima are above 0°C) is generally a discrete and uninterrupted period. To avoid those infrequent warm days in mid-winter (e.g., 26 February 1965 when the daily maximum temperature at Alert was +1.1°C) the "melt season" was defined as the length of time from the first two consecutive days with daily maxima $>0°C$ in the spring to the last two such days in the fall.[1] The average duration of the melt season is shown in Table 4. The shortest melt season on average (86 days) is experienced at Isachsen in the northwestern sector of the Queen Elizabeth Islands, where mean summer temperatures are also lowest. Average daily temperature at Isachsen during the melt season is only 1.6°C. Although additional climatic data from this region are sparse, it seems likely that the low glaciation levels along northern Axel Heiberg and northwestern Ellesmere Island are in part the result of these very brief, cool ablation seasons (cf. Arnold, 1965; Alt, 1975). By contrast, at the northern end of Baffin Bay, melt seasons average almost 50% longer than at Isachsen, with Thule having a melt season of approximately 4 mon. In addition, above freezing temperatures have been recorded at Thule in every month of the year (averaging five to six such occurrences per year) and temperatures as high as 6.1°C have been recorded in February (e.g., 24 February 1965). Such extremes are clearly related to Thule's position at the northern end of Baffin Bay where both warm air advection from the south and the presence of the North Water (Dunbar, 1969) ameliorate the mid-winter climate (Figure 1).

Of particular interest in Table 4 is the melt season duration at Eureka which averages 2 weeks longer than at Isachsen, about 380 km to the east-southeast, and 4 days longer than at Resolute, over 600 km to the south. It can indeed be claimed that Eureka is "The Garden Spot of the High Arctic" (Eureka Weather Service Office Motto)! This relatively long melt season is presumably due to its more continental location (Figure 1) away from cool maritime influences and their associated coastal fogs and low cloud which so frequently affect the northwestern margins of the archipelago (Alt, 1975).

It is clear from Table 4 that at all stations, except Thule, the melt season closely approximates the months of June, July, and August and hence it might be assumed that the mean temperature of these months would be a useful index of ablation season conditions. At Alert, for example, mean temperature of the melt season (1951 to 1974) was +1.6°C, whereas the June through August mean for the same period was +1.3°C. However, Table 4 also shows that both the onset and termination of the melt season may vary considerably from year to year. In extremely cool summers (e.g., 1956 and 1972) the duration of the melt season may be 25% shorter than average. Thus, at Alert in 1956 the melt season did not begin until 15 June, a week before the summer solstice, and was over by 19 August. In other years, the melt season length may be up to 30% longer than average. In 1968, for example, the melt season at Alert was almost twice as long as in 1956, yet 1956 had the warmest July mean maximum temperature on record and 1968 had one of the coldest. Hence, neither the mean temperature of the period June to August, nor July T_{max}, nor the length of the melt season can be assumed to closely reflect overall ablation season temperature conditions from one year to

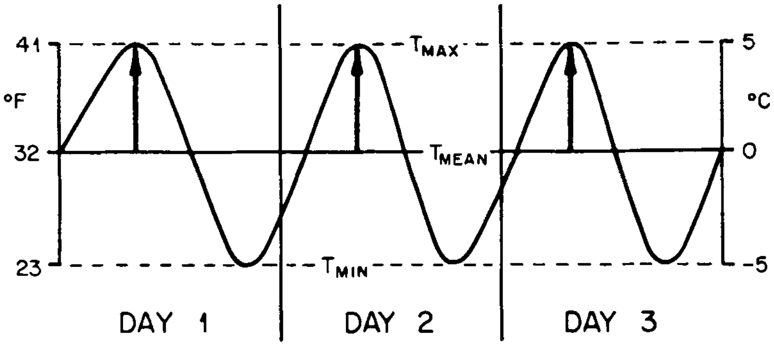

FIGURE 3. Schematic diagram illustrating the importance of computing melting degree day totals on T_{max} and T_{min} separately. If data were computed on T_{mean} in the period illustrated, zero *MDD* totals would be registered when in reality temperatures were >0°C for much of the time. ($\Sigma MDD_X = 15°C$; $\Sigma MDD_N = 0$.)

TABLE 5

Average annual melting degree day totals (°C)

Station	Record length	MDD_X	s	MDD_N	s	$MDD_X + MDD_N$	s
Isachsen	1948-76	347	106	58	39	405	142
Alert	1950-76	408	103	62	28	470	128
Resolute	1948-76	456	111	91	52	547	159
Eureka	1947-76	600	99	155	51	755	142
Thule	1947-74	639	165	167	77	806	225

the next, and a more suitable index must be sought.

MELTING DEGREE DAYS

A useful index of summer ablation may be given by the annual melting degree day total. A daily melting degree day total (*MDD*) is the difference between 0°C and the daily temperature when the latter is above 0°C. The index is commonly computed on the daily mean temperature, (Max + Min)/2, but this has some limitations (Arnold and MacKay, 1964). A more meaningful index is given by computing the index for maximum and minimum temperatures (MDD_X and MDD_N) separately (Figure 3). The index takes into account both melt season length and the accumulated warmth of the season.

Table 5 gives long-term mean annual MDD_X and MDD_N totals for the stations under consideration. Isachsen has the lowest total with only 405 melting degree days (MDD_X and MDD_N). Alert and Resolute have surprisingly similar *MDD* totals considering Alert is more than 900 km farther north (a mean difference of only 77 [MDD_X + MDD_N]). Once again Eureka stands out as exceptionally warm in the summer — almost double the *MDD* total at Isachsen and almost equal to that at Thule. This high total is due not only to higher daily maxima but also to more frequent cases of minimum temperatures remaining above the freezing point. At Eureka and Thule temperatures may remain continuously above 0°C for several weeks. Similarly, daily maximum and minimum temperatures at Tanquary Fiord from 1963 to 1967 were considerably warmer than at Eureka, 220 km to the southwest (Barry and Jackson, 1969), suggesting that a quite extensive area around Fosheim Peninsula and Greely Fiord benefits from continental heating during summer months. In this regard, it is of interest that glaciation levels are extremely high in the Eureka/Tanquary Fiord area (~1100 m). By contrast, to the west and northwest, mountain summits as low as 300 m support ice caps (Miller *et al.*, 1975).

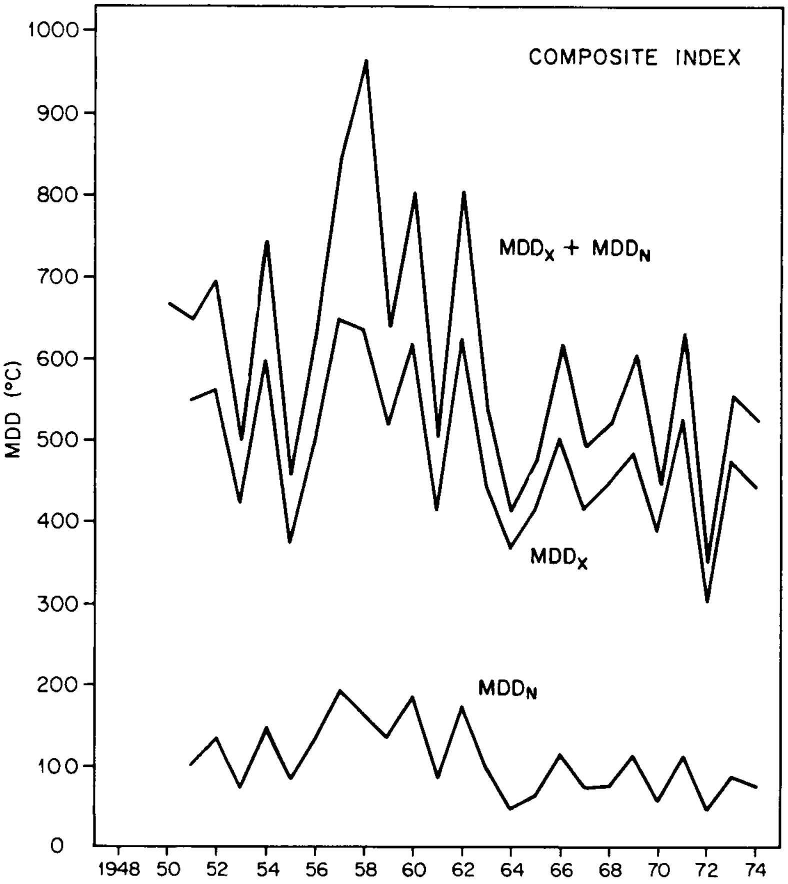

FIGURE 4. Annual averages of MDD_X, MDD_N and (MDD_X + MDD_N) at Alert, Eureka, Isachsen, Resolute, and Thule.

TABLE 6

Change in average annual melting degree day totals

(MDD_X + MDD_N, °C)

Station	Start of record	Mean to 1963 (*A*)	s_A	Mean, 1964 to 1976 (*B*)	s_B	*B/A* (%)
Isachsen	1948	458	153	339	97	74
Alert	1951	505	137	440	113	87
Resolute	1948	616	157	462	118	75
Eureka	1947	808	164	702	88	87
Thule	1947	937	162	604[a]	153	64

[a] 1964-74 inclusive.

Of particular interest for glacier mass-balance studies in the area is the annual melting degree day total since records began (Figure 4). Again these data indicate that an abrupt change in the summer climate of the region occurred between 1962 and 1964. At all stations, interannual variability is high, but *MDD* totals since 1963 have fallen significantly (Table 6). This is most apparent at Thule where mean MDD_X totals (1964 to 1976) were only 65% of the average from 1947 to 1963. MDD_N totals (1964 to 1974) were only 56% of 1947 to 1963 averages with the greatest changes occurring in the months of June and July. Furthermore, the variability of interannual *MDD* totals has been markedly lower since 1964 than in the preceding 16 yr (Table 6).

PRECIPITATION

MEASUREMENT PROBLEMS

Precipitation in Arctic areas is difficult to measure accurately and it has been suggested that absolute amounts of snowfall are greatly underestimated (Hare and Hay, 1971). Furthermore, in the early 1960s, Canadian precipitation measuring procedures were altered following the widespread introduction of "type 3" Nipher shielded precipitation gauges (Potter, 1964). Prior to this change in instrumentation (which varied from station to station)[2] snowfall was measured every 6 h and assumed to have a density of 0.1 (i.e., 10 mm of snow = 1 mm of precipitation). After 1960 (or the date type 3 Nipher gauges were installed, if after 1960) snow was collected in the gauge and melted to obtain a water equivalent measurement.[3]

This change in recording procedure introduces a significant discontinuity into the precipitation record as indicated by an analysis of snow densities in the period after the procedural change (Table 7). Mean monthly snow densities are generally less than 0.1, the figure formerly assumed. For the winter months, November to April, densities average ~15% less than 0.1 (cf. Walker and Lake, 1975). Snow densities are lowest overall at Eureka, the most continental locality. The significance of this is that prior to the record-

TABLE 7

Monthly mean snow densities (to June 1975) after introduction of water-equivalent recording procedure[a]

Station	Period starts	Jan.	Feb.	Mar.	Apr.	May	Jun.	Jul.	Aug.	Sep.	Oct.	Nov.	Dec.
Alert	7/63	.092	.091	.096	.093	.070	.095	.104	.097	.079	.080	.096	.094
Isachsen	5/66	.101	.099	.095	.099	.089	.106	.098	.099	.093	.090	.099	.098
Eureka	9/63	.084	.083	.087	.090	.084	.091	.100	.096	.088	.087	.084	.085
Resolute	5/63	.093	.094	.094	.088	.082	.098	.104	.098	.091	.090	.097	.095

[a] Density = (Total precipitation - rainfall)/snowfall. See also Notes 2 and 3.

TABLE 8

(a) *Percentage of days per month with precipitation total recorded as a trace*

Station	Period	J	F	M	A	M	J	J	A	S	O	N	D
Alert	May 1951-June 1975	40	34	27	23	25	24	17	15	19	25	30	36
Eureka	May 1947-June 1975	46	46	42	30	25	23	21	20	30	35	44	48
Isachsen	May 1948-June 1975	48	50	41	40	36	42	30	29	34	40	44	54
Resolute	Oct 1947-Dec 1972	49	51	43	39	40	39	28	30	41	39	52	54
Thule	Nov 1951-Dec 1970	32	31	32	28	34	32	21	21	29	31	35	30

(b) *Comparison of presently recorded annual precipitation (B) and annual precipitation assuming traces = 0.127 mm (A)*

Station	Period	Traces = 0.127 mm (A)	Traces = 0 (B)	A/B (%)
Alert	May 1951-June 1975	169.7	157.5	108
Eureka	May 1947-June 1975	76.2	60.4	126
Isachsen	May 1948-June 1975	124.8	105.2	119
Resolute	Oct 1947-Dec 1972	153.8	134.3	114
Thule	Nov 1951-Dec 1970	153.0	138.8	110

ing of snowfall water equivalents, total (measurable) precipitation amounts (i.e., amounts greater than Trace) were overestimated relative to later measurements and in considering precipitation variations over time this factor must be taken into account.

A further problem with precipitation data concerns the frequent occurrence of Traces in the records of Arctic weather stations. Jackson (1960) has pointed out that the practice of recording precipitation every 6 h may lead to underestimation of daily total precipitation (i.e., four 6-h Traces = 1 [daily] Trace, rather than any finite total). Traces are commonly recorded for the daily precipitation total, particularly in mid-winter months (Table 8a). If a daily total, recorded as a trace, is assumed to be half of the lowest amount normally recorded (i.e., 0.005 inches, 0.127 mm) annual precipitation totals increase by +8 to +26% (Table 8b). This is clearly an important difference and it seems likely that annual precipitation amounts in the High Arctic are being underestimated (at least in part) because of the high frequency of traces regardless of the time period under discussion. An adjustment based on this fact may account for some of the discrepancy between recorded annual precipitation and measured runoff noted by Hare and Hay (1971).

PRECIPITATION VARIATIONS

Table 9 shows monthly precipitation as a percentage of annual totals. At all stations, most precipitation falls in the 4 mon July to October (~57 to 66% of annual totals) and a significant proportion of this falls in the form of snow during the melt season, increasing albedo and retarding ablation. In order to investigate precipitation fluctuations through time, snowfall amounts during the period when densities were assumed to be 0.1 were adjusted to the densities given in Table 7. The data were then combined into two seasonal precipitation totals: July to October and November to June. Although abrupt changes in the precipitation record are not as apparent as in the temperature record (perhaps due to the greater variability of interannual precipitation amounts) the data have been divided into comparable periods (Table 10). At all stations, November to June precipitation has increased since the early 1960s although overall amounts have remained quite low (<80 mm). In the summer and early winter months (July to October) precipitation amounts have also increased at Alert and

TABLE 9

Monthly precipitation as a percentage of annual total and mean annual precipitation

	Resolute (Oct 1947-Dec 1972)	Isachsen (May 1948-June 1975)	Eureka (May 1947-June 1975)	Alert (July 1950-June 1975)	Thule (Nov 1951-Dec 1970)
			Monthly precipitation		
J	2.1	2.7	4.9	4.5	8.0
F	2.5	2.4	4.1	3.2	8.3
M	2.3	2.3	3.4	4.6	3.7
A	4.3	4.1	3.7	4.6	4.0
M	6.5	8.7	5.0	6.9	8.3
J	9.0	7.7	6.6	8.8	4.6
J	18.8	19.5	20.6	12.8	13.5
A	23.3	20.5	16.4	17.4	11.2
S	12.8	15.7	16.2	17.5	10.4
O	10.9	9.9	11.3	9.5	11.5
N	4.0	4.4	4.3	5.3	9.2
D	3.5	2.3	3.5	5.0	7.4
			Mean annual precipitation		
(mm)	134.3	105.2	60.4	157.5	138.8
(inches)	5.3	4.1	2.4	6.2	5.5

TABLE 10

Seasonal precipitation totals (snowfall, adjusted for density and rainfall) (mm)

July-October								
Station	Period	% of $\bar{x}$ annual precip.	Start of record to 1963 (*A*)	s_A	1964 to 1976	s_B	Change (*B*-*A*)	*B*/*A* (%)
Alert	1951-76	57	75	29	96	27	+21	128
Isachsen	1948-76	66	66	26	75	36	+ 9	114
Eureka	1948-76	65	41	25	36	10	- 5	88
Resolute	1948-76	66	86	23	86	32	0	100
Thule[a]	1952-74	47	67	41	60	21	- 7	90

November-June								
Station	Period	% of $\bar{x}$ annual precip.	Start of record to 1962/3(*A*)	s_A	1963/4 to 1975/76(*B*)	s_B	Change (*B*-*A*)	*B*/*A* (%)
Alert	1950/1 -75/6	43	50	18	80	24	+30	160
Isachsen	1948/9 -75/6	34	28	14	48	17	+20	171
Eureka	1947/8 -75/6	35	18	10	25	5	+ 7	139
Resolute	1947/8 -75/6	34	43	14	47	17	+ 4	109
Thule[a]	1951/2 -74/5	53	66	38	80	32	+14	121

[a] Thule data not adjusted for density; rainfall is not recorded separately, so for much of the year snowfall density cannot be calculated.

Isachsen, but no change was recorded at Resolute; at Eureka and Thule average precipitation declined. At all stations, annual precipitation totals have increased but the changes are only of significance at Alert (+ ~30 mm). These increases, if representative of the northern and northwestern margins of the Queen Elizabeth Islands, are important for glacier mass balance in such an arid environment, but when the fall in summer temperature is also considered, it is clear that this recent change in the climate of the Canadian High Arctic is particularly noteworthy.

RECENT CLIMATIC FLUCTUATIONS AND GLACIER MASS BALANCE

It has been reported elsewhere (Bradley and England, 1978) that mass balance on the Devon Island ice cap is closely related to melting degree day totals at Thule and Resolute. The relationship is particularly good with average annual MDD_N values at these stations (Figure 5) but a similarly high correlation is found with MDD_X values ($r = 0.86$) and the composite MDD_X index for all High Arctic stations shown in Figure 4 ($r = 0.80$). The correlation appears to be consistent regardless of variations in winter snowfall though in some years above average winter accumulation is followed by relatively cool summers (and vice versa) which tends to reinforce the relationship noted. Since 1963, a marked change in the frequency of positive balance years on the Devon Island ice cap has occurred, resulting in a net mass loss on the northwestern sector of the ice cap of <350 kg m^{-2} from 1963/64 to 1973/74. Using reconstructed mass-balance data based on the regression in Figure 5 we can compare this figure with an estimated net mass loss of ~3500 kg m^{-2} from 1947/48 to 1963/64 (Bradley and England, 1978). During this earlier period,

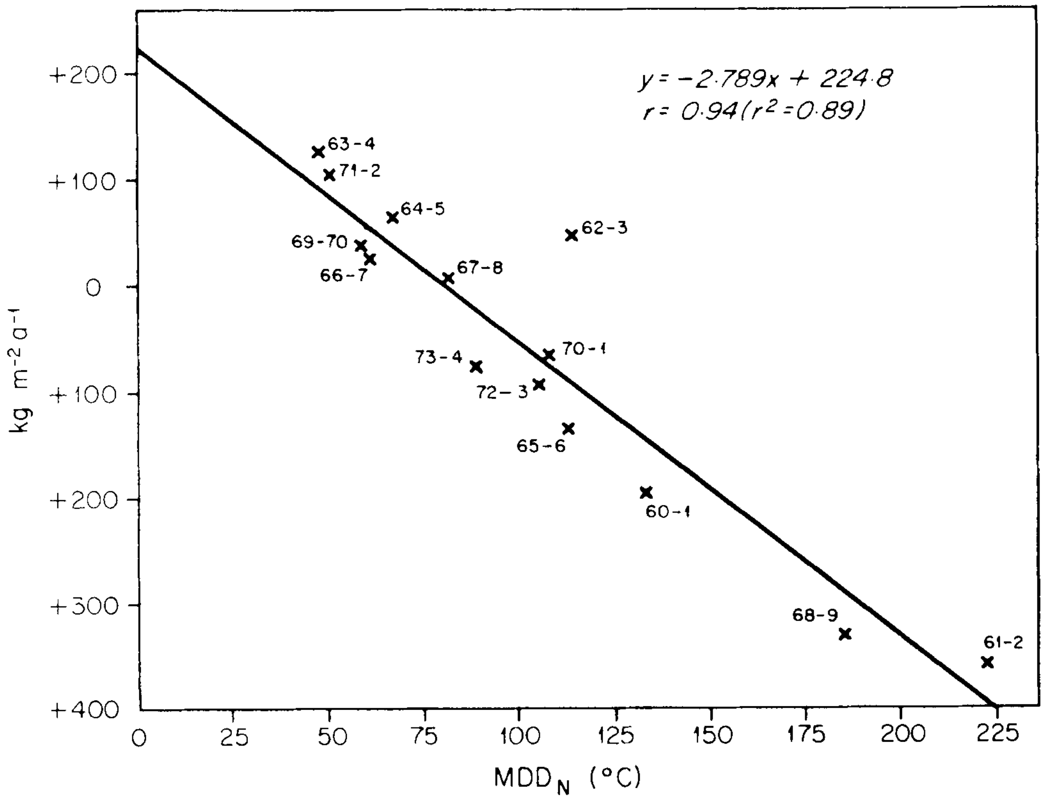

FIGURE 5. Regression of mass balance on the northwest sector of the Devon Island ice cap (Koerner, 1977) and average of T_{min} melting degree day totals at Thule and Resolute.

the estimated average annual mass loss on the ice cap was thus 6 to 7 times greater than during the later period. Nevertheless, although there has been a change towards more frequent positive balance years, an individual negative balance year (such as 1968/69) may obliterate the total mass gains of all the positive years together. This point has been examined from a synoptic climatological standpoint by Alt (1978) who determined that if more than one year dominated by "Anticyclonic types" occurs in a decade, the mean mass balance of the decade will be negative.

Further consideration of the relationship between Devon Island ice cap mass balance data and various indices of "summer warmth" indicate that the apparent linear relationship observed using data from recent years may not be appropriate when greater extremes are considered. Studies with other climatic indices indicate that a curvilinear relationship may give a more appropriate model, as shown schematically in Figure 6. Mass balance is presently limited in a positive direction by a low amount of accumulation, whereas negative mass balance is relatively unlimited, i.e., the ice cap may lose substantial amounts of snow, firn, and ice, but obviously it can never gain more mass than it accumulates in any one balance year. In recent years, mean accumulation at 1800 m on the summit of the Devon Island ice cap is ~220 kg m^{-2} yr^{-1} and the mean for the northwest sector of the ice cap is probably even less than this (Koerner, 1966); modal annual mass balance above the firn line (1600 to 1800 m) is 175 kg m^{-2} (Alt, 1978). Consequently, the linear relationship observed between high *MDD* values and mass balance must change with very low *MDD* values. In recent years, for example, a mean of 510 (MDD_X + MDD_N) at Thule and Resolute corresponded to a net balance of 200 kg m^{-2}. In years when *MDD* values rose to 810 *MDD*, a very negative mass balance was recorded (~ -350 kg m^{-2} yr^{-1}). If *MDD* values fell to 210, however, one would only expect a net balance of ~ +185 kg m^{-2} yr^{-1} (Figure 6). Even if summer temperatures were maintained at the post-1963 level, significant growth of the ice cap is thus unlikely without increased accumulation.

RECENT CLIMATIC CHANGE AND VOLCANIC DUST

An important aspect of the recent change in the summer climate of the Canadian High Arctic is the relatively abrupt shift which occurred around 1963/64. This is particularly well illustrated in Figure 2 and it was noted in upper air data for the Canadian Arctic by

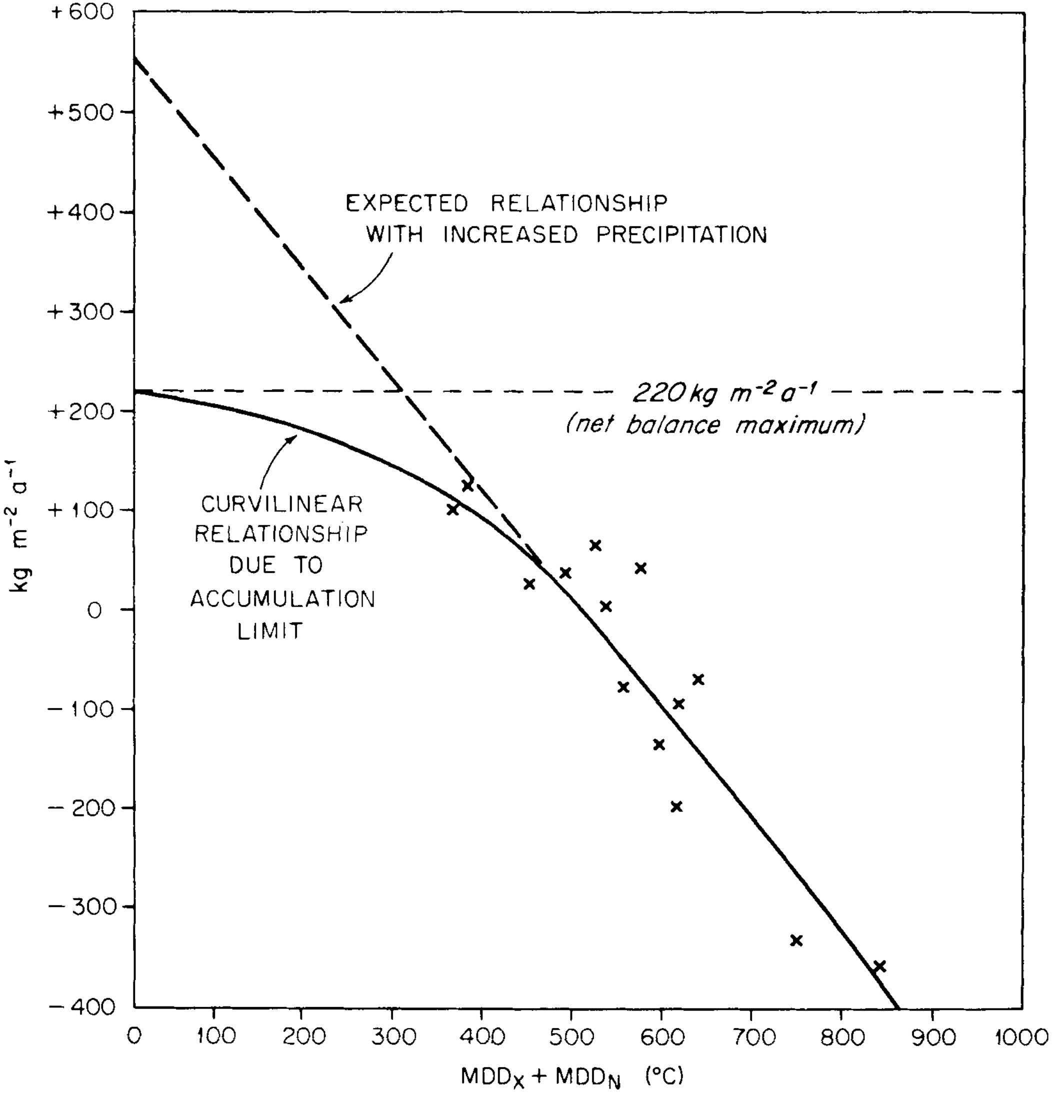

FIGURE 6. Schematic diagram illustrating the probable relationship between mass balance and melting degree day totals at extremely low *MDD* values under the accumulation conditions of recent years. Accumulation at the summit of Devon Island ice cap averages ~220 kg m^{-2} yr^{-1} and may be less for the northwest sector. For the relationship to remain approximately linear, where an "average" positive balance year would be approximately equal in magnitude to an "average" negative balance year, precipitation amounts would have to be considerably higher.

Bradley (1973a). Also of interest in this regard is the recent study (Dronia, 1974) of 1000 to 500-mb thickness for the area shown in Figure 7. Although Dronia's observations were biased towards high latitudes (by using data from a uniform latitude-longitude grid) the results show a marked change in the climate of the Northern Hemisphere around 1963/64, with predominantly negative 1000 to 500-mb thickness departures (from the

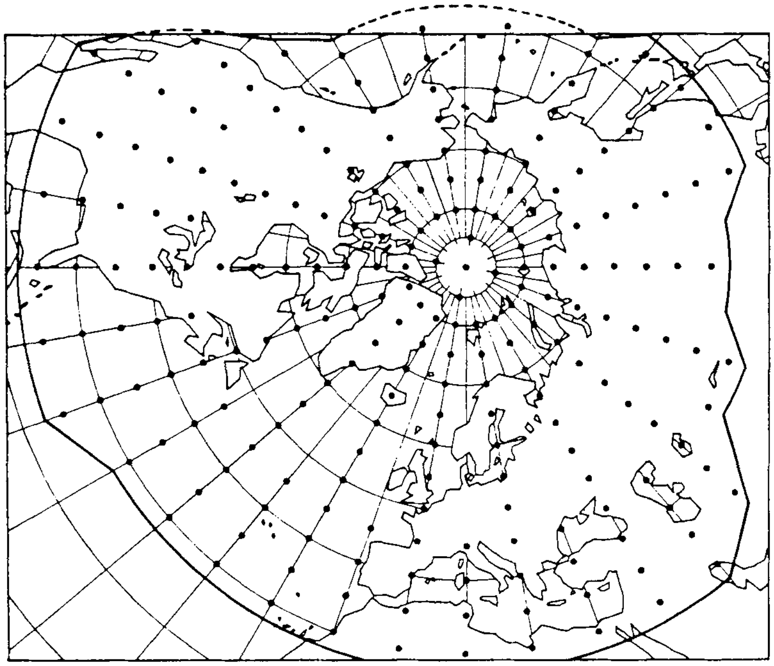

FIGURE 7. Grid network of 1000 to 500-mb thickness values used in study by Dronia (1974). (Reproduced with permission of Gebrüder Borntraeger from *Meteorol. Rundsch.* 27, 1974, Abb. 1.)

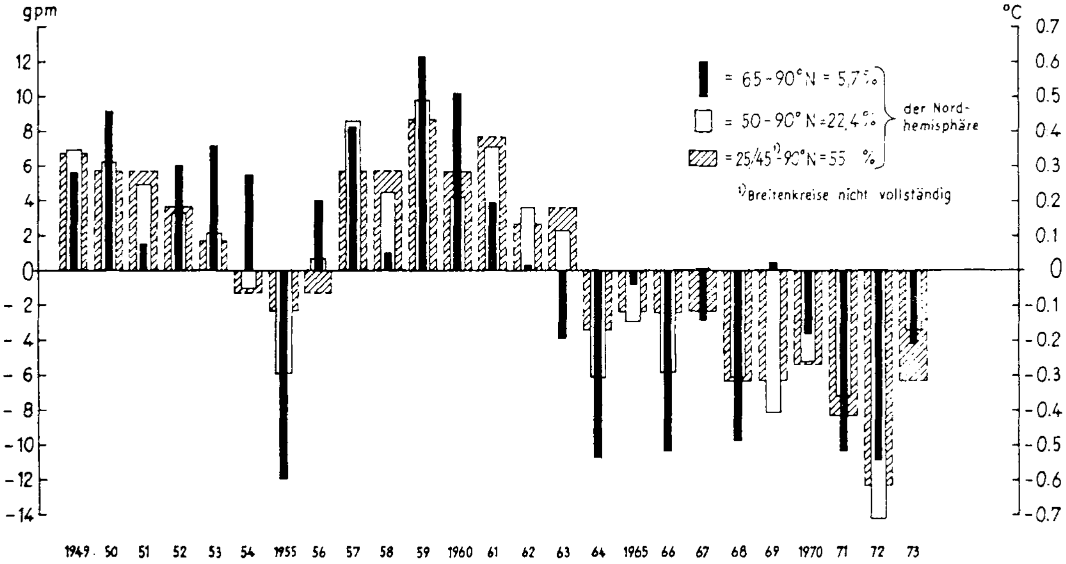

FIGURE 8. Annual values of 1000 to 500-mb thickness over the grid network shown in Figure 7 for the latitude zones indicated. (Reproduced with permission of Gebrüder Borntraeger from Dronia in *Meteorol. Rundsch.* 27, 1974, Abb. 2.)

1949 to 1973 mean) after 1963 (Figure 8). Similar changes have been noted in other global climatic indices by Kukla *et al.* (1977).

It has been suggested by several authors that this fall in tropospheric temperature was due to volcanic dust from the eruption of Mt. Agung (Yamamoto *et al.,* 1975; Newell and Weare, 1976; Angell and Korshover, 1977; Hansen *et al.*, 1978). This was one of the largest eruptions since 1883 when Krakatau (6°S, 105°E) injected large amounts of dust into the stratosphere. Lamb (1972) estimates a World Dust Veil Index (DVI) of ~800 as a result of the Mt. Agung eruption (compared to a DVI of 1000 for Krakatau). It is of interest that several other major eruptions have occurred since 1963; these include Awu, Philippines (1966; DVI ~200); Fernandina, Guatemala (1968, DVI ~200); Mt. Hudson, Chile (1971, DVI ~250) and Fuego, Guatemala (1974, DVI ~200) (Lamb, 1977). As a result of these eruptions, the annual DVI for the Northern Hemisphere since 1963 has been greater than at any time since 1915 (Figure 9).

Is there a relationship between the recent climatic fluctuation observed in the Canadian Arctic and this influx of volcanic dust into the stratosphere? Bradley and England (1978) suggest that this is indeed the case, basing their argument partly on theoretical and partly on observational grounds. As noted by Lamb (1970) the greatest climatic effect of a stratospheric dust veil is likely to be observed at high latitudes in summer months when solar radiation passes through the greatest depth of the atmosphere and the surface is illuminated continuously. Volcanic dust also tends to persist in the atmosphere longer at high latitudes than elsewhere. Solar radiation measurements for the Canadian Arctic are extremely limited and generally absent before 1961. Comparison of solar radiation receipts in the period before and after the eruption of Mt. Agung is thus not very helpful. Selection of only clear days is also difficult because of the small sample of days involved.[4] Diffuse radiation receipts at Resolute, averaged for all summer days, reached the maximum levels ever recorded in 1964 (the year following the eruption of Mt. Agung) and subsequently declined (Bradley and England, 1978). Similarly, direct radiation receipts were at a minimum in 1964 and increased somewhat erratically thereafter. This increase in diffuse and decrease in direct radiation is a typical volcanic dust "signal" and has been noted elsewhere in the world, albeit on cloud-free days (Dyer and Hicks, 1965). The Resolute record is difficult to interpret because it is extremely dependent on cloud-cover variations which closely affect diffuse and direct solar radiation receipts. At the same time, volcanic dust veils are known to affect the general circulation of the atmosphere in particular by changing the equator-pole temperature gradient, although the magnitude of this effect may vary due to a variety of factors (Defant, 1924; Lamb, 1970). A frequent effect, however, is for an increase in the strength of the circulation (probably due to a stronger equator/pole temperature gradient) which would tend to increase the advection of moisture into the Arctic, thereby increasing cloudiness and diffuse radiation receipts. This effect is thus superimposed on any change in diffuse radia-

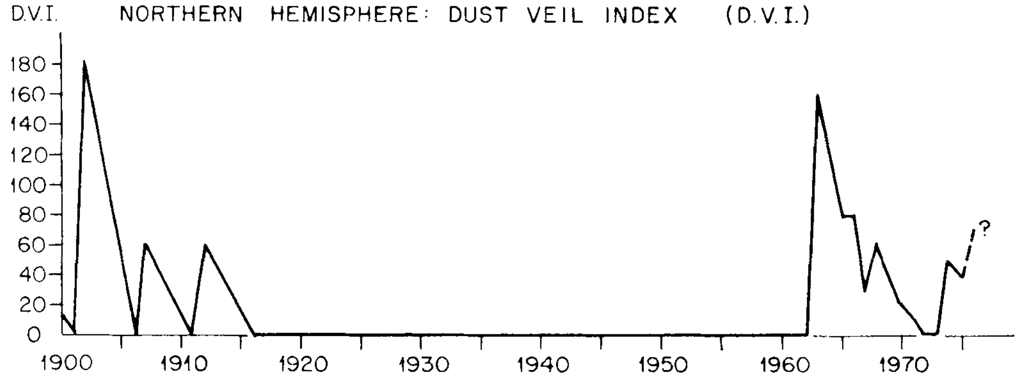

FIGURE 9. Annual values of Dust Veil Index (DVI) for the Northern Hemisphere (data from Lamb, 1977). Note that dust may persist longer at high latitudes (Lamb, 1972). Value for 1976 is tentative.

tion which might be attributable to the immediate effect of the dust layer itself. It is clear, however, that a step-like change in summer climate did occur around 1963/64 and that this has had the most significant impact on summer warmth and glacier mass balance in the High Arctic for over 30 yr, and perhaps as long as 60 yr. It has been noted by Bradley (1973b) that marked increases in circumpolar summer temperatures took place between ~1915 and 1928, a period when the stratosphere became relatively dust free after the frequent volcanic eruptions at the beginning of the 20th century. It seems plausible that the summer temperature conditions of the "post-Agung" period are typical of conditions characteristic of the late 19th and early 20th centuries. This is supported by glaciological observations on northern Ellesmere Island (Hattersley-Smith, 1963) where an analysis of firn stratigraphy on the Gilman Glacier suggested that the summers of the mid-1960s were the coldest since 1925.

The hypothesized volcanic dust effect does not, of course, preclude other factors which may have influenced climatic variability, and no doubt many such effects are superimposed on the climatic records discussed. It is this complex interaction of cause and effect which makes identifying the impact of any single factor extremely difficult. Nevertheless, the authors feel that both surface and upper air data support their contention that increased volcanic dust in the upper atmosphere has played a major role in determining summer climate and glaciological conditions in the High Arctic in recent years.

SUMMARY AND CONCLUSIONS

Analysis of climatic data for the last 25 to 30 yr from the Canadian High Arctic and northwestern Greenland indicates that a significant climatic change occurred around 1963/64. This involved a lowering of mean July freezing level heights by up to 500 m; a decrease in mean July maximum temperatures (at the surface) of up to 2.7°C; a marked decrease in annual melting degree day totals (down to as low as 65% of pre-1963 values); a concomitant increase in mean annual precipitation in northern and northwestern areas (up to 140% of pre-1963 levels). These conditions resulted in much reduced, net mass losses on glaciers in the region after 1963 than in the preceding period. No evidence for a return to pre-1963 conditions is yet apparent.

Regression equations relating mass balance on the Devon Island ice cap to annual melting degree totals indicate that mass balance is strongly controlled by summer temperature conditions. Winter precipitation at adjacent weather stations may vary by a factor of 3 but this appears to have little effect on the relationship between mass-balance and melting degree days. Reconstruction of Devon Island ice cap mass balance to 1947 indicates that the climatic fluctuation of the early 1960s is highly significant for mass balance. From 1947 to 1963 the ice cap lost an estimated 3500 kg m^{-2} whereas from 1964 to 1974 cumulative mass losses were <350 kg m^{-2}. However, although positive balance years since 1963 have occurred as frequently as negative balance years, mass losses during negative balance years are generally much larger than mass gains in positive balance years. It is suggested that this is due to low accumulation amounts on the Devon Island ice cap under present climatic conditions and that significant growth of the ice cap is unlikely without marked increases in accumulation, even if lower summer temperatures persist.

The change in climate of the area was abrupt and coincides with the input of volcanic dust into the stratosphere after the eruption of Mt. Agung in 1963. There is global evidence of a cooling effect as a result of this eruption and the cooling effect is likely to be maximized and to persist longest at high latitudes. Volcanic dust may influence ablation season conditions directly by affecting solar radiation receipts, and indirectly via its impact on the strength of the general circulation. If Agung dust (and dust from subsequent eruptions) was responsible for the recent climatic fluctuation (and its persistence) it seems likely that other periods following major volcanic eruptions may have experienced similar conditions favoring positive mass balance. Conversely, the climate of the period 1947 to 1963 was probably typical of the entire period from about 1920 to 1963 when the atmosphere was relatively free of dust.

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NOTES

[1] Anomalies early and late in the year were avoided by specifying that there must be another occurrence of above freezing temperatures within one month (after the event in the spring, before the event in the fall) before it could be counted as either initiating or terminating the melt season. At Thule this criterion was changed to 10 days because of the higher frequency of occasional isolated warm days in the early spring and late fall.

[2] July 1963 at Alert, 1962 at Isachsen (month unknown), 1952 or 1953 at Resolute (exact date unknown), and September 1963 at Eureka (cf. Potter, 1964).

[3] Monthly snow density analysis indicates recording procedures did not change on the dates indicated by Potter (1964) at Isachsen and not until May 1963 at Resolute. At Isachsen it appears that procedural changes were not initiated until May 1966. Furthermore, at all stations, the procedure does not appear to have been consistently adhered to; densities of 0.1 were commonly used for a run of months, or even years (e.g., at Eureka from October 1968 to August 1970), and then a change to values $\neq 0.1$ commonly occurred. In computing values for Table 8 all density values after the date given in the table were used; hence it is likely that the densities in that table are maximum estimates.

[4] Although the data are sparse, mean July diffuse radiation totals for "clear sky" days (cloud cover at 0000 GMT $\leq 2/10$) as a percentage of total radiation was 25% for 1961 to 1963 and 43% for 1964 to 1972, an average increase of $\sim 18\%$.

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