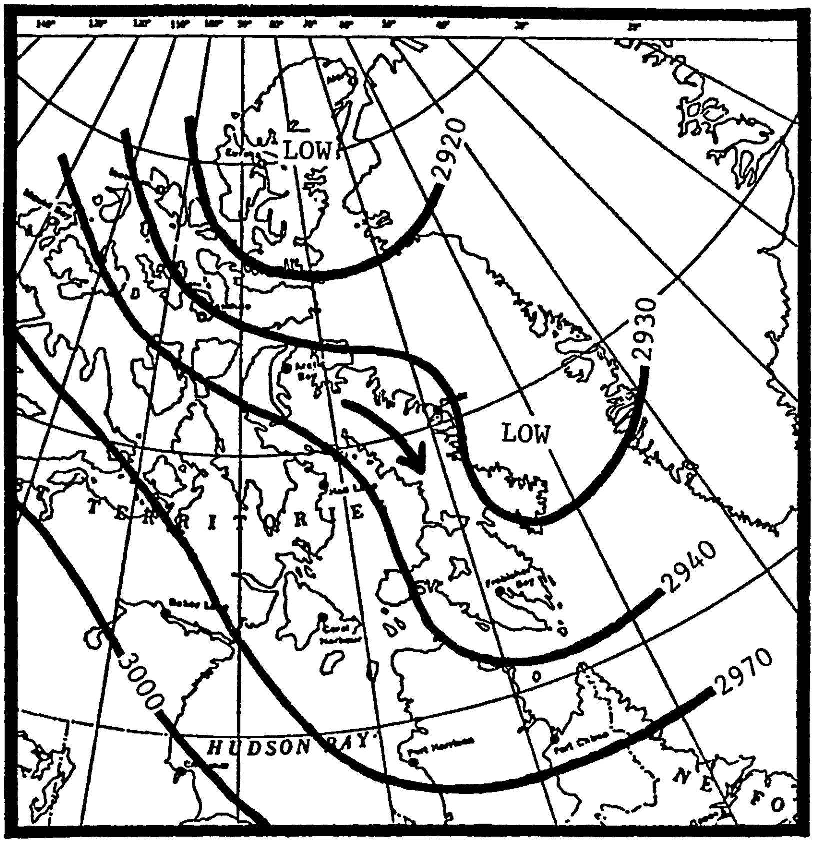
TEMPERATURE AND CIRCULATION ANOMALIES IN THE EASTERN CANADIAN ARCTIC SUMMER 1946-76

Richard A. Keen



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ABSTRACT

Baffin Island and nearby regions of the eastern Canadian Arctic have been the focus of many recent and ongoing paleoclimatic and glaciological studies. The region is very suitable for these studies because of the sensitivity of regional ice and snow conditions to small changes in summer temperature (which average a few degrees above freezing). Baffin Island's location beneath one of the major features of the general atmospheric circulation, a trough in the upper westerlies, raises the possibility that regional climate variations may be closely linked to changes in the global-scale climate and circulation. The existence of such a link would greatly enhance the importance of Baffin Island paleoclimatic and glaciological studies.

In this study, thirty summers (1946-76) of surface temperature and pressure, and upper air data for the Baffin region, the Arctic, and the extratropical Northern Hemisphere, are analyzed to establish the nature and significance of the regional-global summer climate and circulation links.

The variability of Baffin area summer temperatures is shown to be statistically significant on time scales greater than two years and to be closely correlated with the variability of the arctic zonal average summer temperature. The most significant climatological event at Baffin Island during the thirty summers of record is the cooling of the early 1960's; this event is given special emphasis in this study.

From an analysis of daily surface synoptic pressure patterns, subjective surface cyclone tracks, and 500mb positive vorticity flux, it is found that colder summers at Baffin Island are associated with an increase of cyclone activity over Baffin Bay and an intensification of the arctic front storm track across northern Canada. Colder summers are also associated with eastward displacements of the upper (500mb and 700mb) Baffin trough, which in turn are associated with stronger upper westerly winds at the latitude of the trough and over most of the Northern Hemisphere. Stronger westerly winds are shown to be related to greater meridional temperature gradients associated with general hemispheric coolings.

Because of the regional-global scale correlations found in this study, Baffin Island appears to be a significant indicator of larger-scale summer climate conditions and of the nature of the general circulation over time scales from 2 to 10 years. These correlations do not weaken within these time scales, and may very well extend to the time scales of interest to paleoclimatologists and glaciologists.

PREFACE

As part of a larger program concerned with climatic change and sea ice conditions in the eastern Canadian Arctic, supported during 1972-77 by the National Science Foundation, Division of Polar Programs (GV-28218), Richard Keen undertook to examine the mechanisms linking the atmospheric circulation and summer temperatures in the area. His analysis, using various synoptic climatological approaches, demonstrates both regional and global-scale interactions over time scales of one to ten years, involving the eastern North American tropospheric trough, the strength of the hemispheric westerly winds, and the meridional atmospheric temperature gradient. In view of the sensitivity of the land and sea ice in the eastern Canadian Arctic to small climatic fluctuations, the results of this study should be of wide interest to environmental scientists working on problems of climatic change and the polar regions.

R.G. Barry
Project Director

ACKNOWLEDGEMENTS

Many individuals and organizations helped make this completed dissertation possible. Foremost among these is Dr. Roger G. Barry, who presented the initial questions that inspired this study and whose encouragement and advice certainly aided its progress. Comments and suggestions by Drs. John Andrews, David Greenland, John Hobbs, and Uwe Radok helped improve the final draft, and are greatly appreciated.

The vorticity flux computations were made possible by a generous grant of computing time from the National Center for Atmospheric Research and by the assistance of NCAR staff, particularly R. Jenne, W. Spangler, and G. Walters. Margaret Eccles did most of the tedious computer programming that led to the synoptic type catalog, and was always helpful when other computing problems were encountered.

The NCAR CDC 7600 computer was also an invaluable aid in preparing many of the figures. Linda Crane meticulously typed the final manuscript.

Support for much of this venture was provided by National Science Foundation Grant GV-28218 and by the Veteran's Administration.

A debt of gratitude is certainly owed to all those who helped bring this project to its completion, but it is to my parents, without whom this endeavor would have never begun, that I would like to dedicate this finished dissertation.

TABLE OF CONTENTS

LIST OF TABLES

LIST OF FIGURES

FIGURE PAGE

FIGURE PAGE

CHAPTER ONE

INTRODUCTION

Statement of the Problem

The monthly, seasonal, and annual mean climatic conditions of the Arctic (north of $60^{\circ}N$) are subject to relatively large inter-annual variations. Northern Hemisphere maps of the standard deviation of July and December mean surface temperatures (Lamb, 1972, p. 280) show the greatest variability, $2^{\circ}C$ in July and $5^{\circ}C$ in December, over the high latitude regions of North America and Asia; a similar distribution is given for the occurrence of extreme ($10^{\circ}C$ or greater) monthly temperature anomalies (from Gedeonov, in Barry and Perry, 1973, pp. 200-201). The standard deviation of monthly mean sea level pressure (maps in Lamb, 1972, pp. 274-275) is likewise greatest in high latitudes. This effect is partly a result of the influence of the greater persistence of daily pressures at high latitudes on the random variance of monthly means (Madden, 1976); the remaining variance, however, presumably due to climate fluctuations, is statistically significant over much of the Arctic, especially in the North American sector.

On longer time scales, sizeable trends are apparent in the climate of the Arctic. Van Loon and Williams (1976) show that the $70^{\circ}N$ zonal mean summer surface temperatures have a downward trend of about half a degree over 30 years. Locally, the trend exceeded $2^{\circ}C$ over the 30 years. From 1958 to 1963, the mean temperature of the 300 to 700mb layer in the north polar region fell by $0.66^{\circ}C$, and has since remained fairly constant (Angell and Korshover, 1975, 1978); the 1958-63 cooling extended to the surface and into the stratosphere (Starr and Oort, 1973). On a regional scale, the cooling over the Canadian Arctic Archipelago from 1955-63 to 1964-72, resulting in a lowering of the July freezing level by 250 to 300 meters, has been shown to be statistically significant at the 0.1% level (Bradley, 1973a).

Many more examples of interannual variability and short-term trends exist in the Arctic, but those cited above suffice to illustrate that these fluctuations do exist, that they tend to be larger in the Arctic than elsewhere, and that they are not mere statistical fluctuations. The latter point is especially important, because it implies that there is some degree of determinism behind these climate variations, and that the physical causes should be identifiable over the same time scales (year-to-year, short-term trends) as the climate fluctuations.

Baffin Island, the largest island in the Canadian Arctic Archipelago, straddles the Arctic Circle just west of Greenland. Although the interannual variability of temperatures around Baffin Island is no greater than elsewhere in the Arctic, the average summer temperatures are close to freezing, especially over the higher elevations; consequently, ice and snow conditions are very sensitive to small climate changes (Andrews _et al_., 1972).

Dunbar (1972) notes that Baffin Bay sea ice conditions deteriorated markedly with the cooling of the 1960's. On land, permanent snowbanks grew in size during the 1960's, reversing the trend of the preceding decade (Bradley and Miller, 1972). On a longer time scale, permanent snow cover over much of north central

Baffin Island has decreased from 50 to 37 percent since the Little Ice Age (Locke and Locke, 1977); the decrease was from 70 to 2 percent over some upland areas close to the present snowline (Ives, 1962). The lower snowlines during the Neoglacial appear to be associated with summer temperatures approximately 1.5°C lower than present values (Williams, 1978a). Even at present, large areas of the Island are very close to glaciation. Glaciation level data published by Andrews and Miller (1972) and Bradley and Miller (1972) indicate that a summer cooling of 0.7°C to 3.0°C would, if persistent, subject much of Baffin Island to renewed glaciation. The reflection of solar radiation from the expanded snow cover could help maintain (and perhaps expand) the colder, snowier conditions, with potential consequences for larger-scale glaciation (Williams, 1978b). Thus, as Andrews _et al_. (1972) noted, "Baffin Island is highly suitable for long-term climatic monitoring and continued extensive study as a guide to the past, present, and future environmental changes."

The paleoclimatic significance of Baffin Island is enhanced by its location at one of the major circulation features of the Northern Hemisphere, a trough in the upper westerlies. Palynological studies by Nichols _et al_. (1978) and Andrews _et al_. (1979) suggest that wind shifts associated with changes in the position of the trough affect the import of exotic pollen into Baffin Island. A link between the trough position and summer climate conditions (and therefore snow conditions) on Baffin Island is noted by Brinkmann and Barry (1972) and Bradley (1973b, 1974). Bradley (1973a) proposes that a hemisphere-wide change in atmospheric circulation patterns around 1961-62 (Namias, 1969; Dzerdzeevski and Sergin, 1972) is responsible for the concurrent cooling around Baffin Island. Dzerdzeevski (1963) suggests that "all changes in temperature and precipitation -- if they are taken (observed) for long enough periods -- are connected with changes in frequency and duration of large-scale circulation patterns." Therefore, both snow cover and pollen studies may aid in reconstruction of the past history of the general circulation, as well as of the local climate.

Before any deductions about the general circulation can be made from Baffin Island paleoclimatic studies, the connection between the large-scale circulation and the climate of Baffin Island during the period of instrumented record must be established. Summer is the season of exotic pollen transport, and summer temperatures are one of the two prime determinants (along with winter snowfall) of the snow mass balance (Andrews and Barry, 1972). As is demonstrated in this study, summer temperature is also the climatic parameter most closely associated with sea ice conditions in Baffin Bay. Summer precipitation is of less paleoclimatic importance, and is also more difficult to analyze. Most Arctic summer precipitation falls during a few events (Fogarasi, 1972) and is highly spatially variable for individual events (Barry and Jackson, 1969); as a result, long term trends are not necessarily spatially consistent (Bradley, 1973b). Precipitation trends are also obscured by changes in snowfall measurement procedures and by the varied catch efficiencies of the different snow guages employed during the period of record (Goodison, 1978). Therefore, precipitation is not examined in this study, and the connection sought will be that between Baffin summer temperatures and the general circulation, this link being the one of greatest interest to climate studies in the eastern Canadian Arctic.

In concluding a report of the Institute of Arctic and Alpine Research (INSTAAR) on climate studies in eastern Baffin Island, Barry and Jacobs (1974) state: "Global scale processes are ultimately responsible for the major year to year differences in regional conditions. There is therefore a clear need to examine the nature of the regional/global interactions." Establishment of the nature of the regional/global interactions is essential not only to reconstructing the general circulation from Baffin Island paleoclimatic indicators, but also to the reverse question of understanding, modeling, and perhaps even forecasting the response of these indicators (particularly, snow cover and glaciation) to changes in the nature of the global atmospheric circulation.

Organization of the Study

Establishment of a link between the summer climate of Baffin Island and the general circulation must be done in steps if the physical nature of the link is to be understood. The organization of this study reflects these steps, starting with an assessment of the observed Baffin Island summer temperature fluctuations. The observed fluctuations are then compared with changes in synoptic scale circulation patterns, which are then linked to the variability of regional and hemispheric scale circulation regimes.

The time scales of the climate variations analyzed in this study range from interannual to inter-decadal. Because the longer time scales are of greatest interest to paleoclimatologists, and because the most significant feature of the summer climate in the Baffin area over the past thirty years is the cooling of the 1960's, special emphasis is given to circulation changes associated with this cooling. Bradley (1973b) shows that the 1960's cooling is not a unique phenomenon, but rather is typical of the decadal scale fluctuations that have occurred on Baffin Island during this century.

Because several different data sources and analytical techniques are used in this study, their discussion is presented in the relevant chapters.

In Chapter Two the observed spatial and temporal variability of summer average temperatures in the Arctic, and in particular around Baffin Island, are presented and discussed. The statistical significances of the fluctuations are noted. Baffin area temperature variations are then compared with those elsewhere in the Arctic and with a derived Arctic average.

Chapter Three examines the climate fluctuations of the Baffin area in terms of synoptic scale surface pressure patterns. Interannual shifts in regional cyclone/anticyclone activity, and their connection with summer temperature variations, are discussed.

In Chapter Four the observed changes in summer temperatures and cyclone/anticyclone activity are related to changes in the position and strength of the Baffin trough. The geographic and dynamic origins of the shifts in cyclone activity are investigated using surface cyclone and 500mb positive vorticity flux data for the North American - Western North Atlantic sector. The vorticity flux analysis is then expanded with reference to zonal and hemispheric changes in storm activity.

Chapter Five correlates fluctuations of the regional scale circulation to those of the Northern Hemispheric general circulation. Using data for the 500mb level, the longitude of the Baffin trough is related to the strength of the sub-polar westerlies, which in turn are related to the strength and latitudinal distribution of the Northern Hemisphere zonal westerlies (i.e., the intensity and structure of the north circumpolar vortex), and thence to the hemispheric average temperature. Possible causes of the largest scale circulation anomalies are discussed.

Chapter Six summarizes the results and states the nature of the essential physical links between the interannual variabilities of Baffin area summer temperatures and of the general circulation of the Northern Hemisphere.

CHAPTER TWO

OBSERVED INTERANNUAL VARIABILITY OF SUMMER TEMPERATURES

As indicated in the introductory chapter, the summer climates of Baffin Island and of the entire Arctic have undergone sizeable interannual fluctuations and longer-term trends over the period of instrumented record. In this chapter the magnitudes and spatial scales of Baffin area summer temperature variations are discussed, to provide a basis for comparison with circulation anomalies in later chapters. It is determined which of the Baffin temperature fluctuations during the period of record (1943-76) are climatologically significant and warrant further study. A brief discussion of Baffin Bay sea ice conditions is presented as environmental evidence of the climate fluctuations. Finally, Baffin area temperatures are compared with those elsewhere in the Arctic and with a derived arctic average temperature, to determine the extent to which Baffin temperature fluctuations are regional and how far they are representative of conditions across the rest of the Arctic.

Data and Analysis

The temperatures discussed in this chapter are summer mean surface air temperatures, summer being June, July, and August for averaging purposes. The station data are derived from a variety of sources, including NCAR data files, Monthly Climate Data for the World, and the Monthly Record of Meteorological Observations in Canada.

Many of the data presented in this and later chapters are smoothed by a 1-2-1 running mean. This particular smoothing function is selected because it is an economical approximation to a Gaussian-shaped function. Gaussian smoothing functions have a sharp frequency cut-off, but without the negative amplitudes at some higher frequencies, which is characteristic of many smoothing functions, including straight running means (Holloway, 1958). The frequency response of the 1-2-1 function is listed in Table 2.1.

TABLE 2.1. Effects of 1-2-1 Smoothing Function on Periodic Components of a Time Series

Frequency ($year^{-1}$)	Period (years)	Response
0.0	∞	1.00
0.1	10.0	.91
0.2	5.0	.66
0.3	3.3	.35
0.4	2.5	.10
0.5	2.0	.00

Since summers are sampled at yearly intervals, frequencies greater than 0.5 are indistinguishable from those the same amount less than 0.5. The effective length of the smoothing is 2 years.

A consideration in estimating the significance of interannual variations is the "natural variability" of the seasonal average (Madden, 1976). The natural variability differs from the standard error of the mean in that the number of effectively independent samples during the season differs from the total number of days (92), due to persistence effects. At Broughton Island, on the shores of Baffin Bay, the standard deviation of daily temperatures within a summer season (seasonal trend removed) is 3.1°C. The standard error for a 92-day season would be 0.3°C. However, the lag autocorrelations of the daily temperatures, calculated for 12 summers and then averaged, shows a great deal of persistence (Table 2.2).

TABLE 2.2. Lag Autocorrelations of Broughton Island Summer Temperatures

Lag (days)	Autocorrelation
0	1.00
1	.69
2	.40
3	.28
4	.22
5	.16
10	.14
15	.09
20	.12

The e-folding decay time of the autocorrelation, derived from a best fit to lags 0 to 5, is 2.5 days. For the 12 individual summers, this e-folding decay time is found to be insignificantly correlated with the average summer temperature; thus, greater persistence does not appear to be a feature of anomalously warm or cool summers.

Twice the e-folding decay time, 5.0 days, is the effective time between independent sample values (Leith, 1973). Madden (1976) found the same figure -- 5.0 days -- for July surface pressure variations at the same location. With a daily standard deviation of 3.1^{o}C and 18.4 effectively independent samples during a 92-day summer season, the natural variability at Broughton Island is 0.7^{o}C. The standard deviation of 16 summer average temperatures at Broughton is 1.0^{o}C; half of the total interannual variance is therefore due to "natural variability", and half due to physically significant climate fluctuations.

The 1-2-1 smoothing of Broughton summer averages reduces the total interannual variance by 40 percent; this 40 percent belongs to the high frequency component of the interannual variability. The natural variability, due to the occurrence of individual spells of weather within a calendar (June 1 to August 31) summer, is random for climatological purposes. Therefore it is contained in much of the high frequency "noise" removed by the smoothing. Very little "signal" is lost to the smoothing, since summer to summer persistence and the quasi-biennial cycle appear to have little influence around Baffin. At Broughton, the autocorrelations of 16 summer average temperatures are -.01, .09, and .02 for lags 1, 2, and 3 years, respectively. At nearby Clyde, with a longer (31-year) record, the respective autocorrelations are .19, .09, and -.04. For the rest of this study, the 1-2-1 smoothed values will be given greater emphasis than the annual values, since the smoothing more clearly reveals climatologically significant variations. It should be remembered, though, that the two-year effective length of the smoothing function cuts the sample size in half.

Baffin Area Temperature Variations

Summer average temperatures over Baffin Bay, averaged between three coastal stations (Clyde, Upernavik, and Egedesminde) for 1943-76, are plotted in Fig. 2.1. The longer-term variations are made more apparent by the 1-2-1 smoothed values (solid line). The natural variability of this 3-station average should approximate the 0.7°C natural variability for Broughton Island, since the interannual standard deviations for both Broughton and the 3-station average are 1.0°C. Keeping in mind the natural variability and the effect of the smoothing functions, significant events in the summer temperature history of Baffin Bay appear to be the cool summers of the mid-1940's, the warm spell from 1948 to 1960 (especially warm around 1949 and 1960), the cold summers after 1963, and the exceptionally cold summers of the early 1970's.

A time series alone does not give the whole story; the spatial variations are also important. Figs. 2.2a and 2.2b map the summer temperature anomaly around Baffin Island for the extreme summers of 1960 and 1972, relative to the published 1941-70 climatological normals. In 1972, the entire region of West Greenland and the eastern Canadian Arctic was well below normal, with the greatest anomalies around Foxe Basin. At most stations on the map the summer was the coldest on record. The 1960 pattern was less monolithic -- at several stations scattered across the

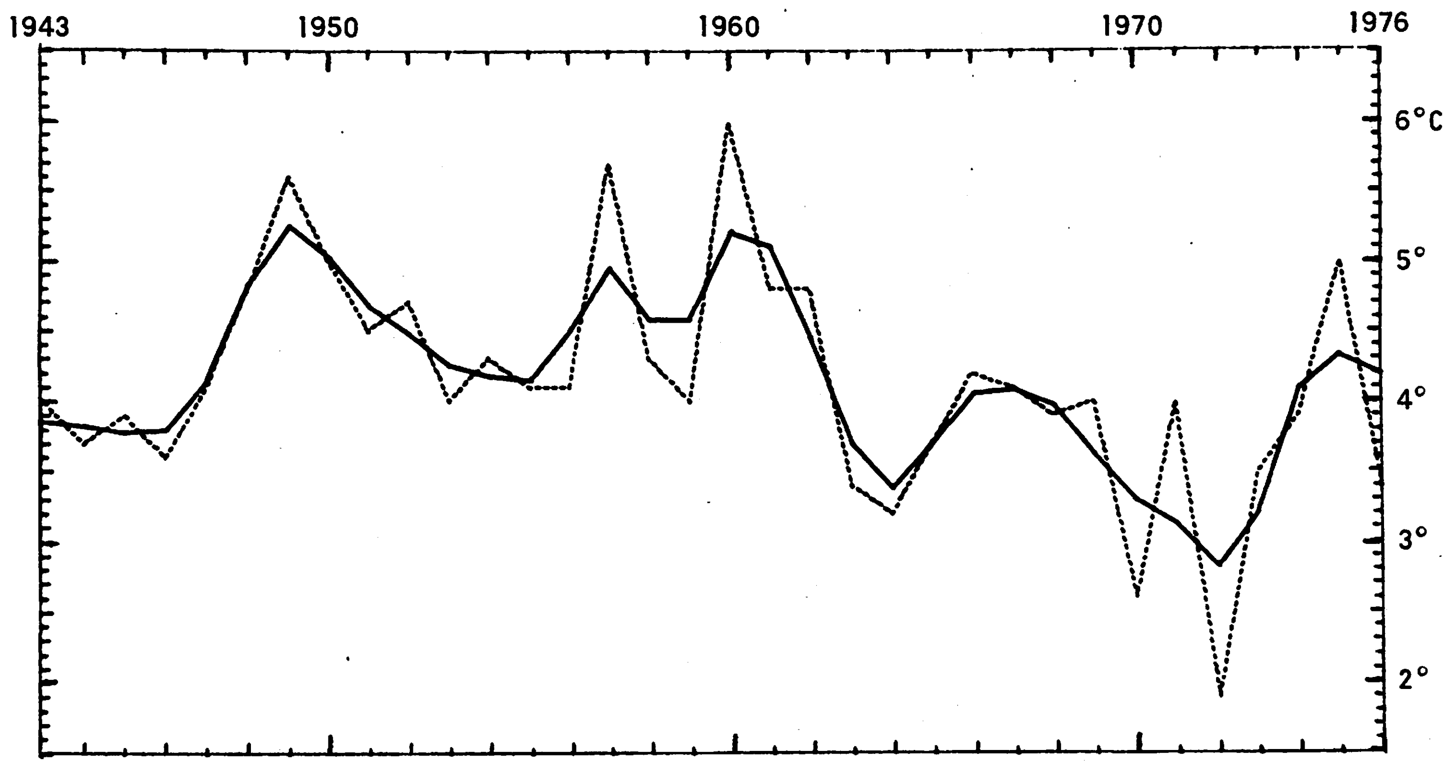
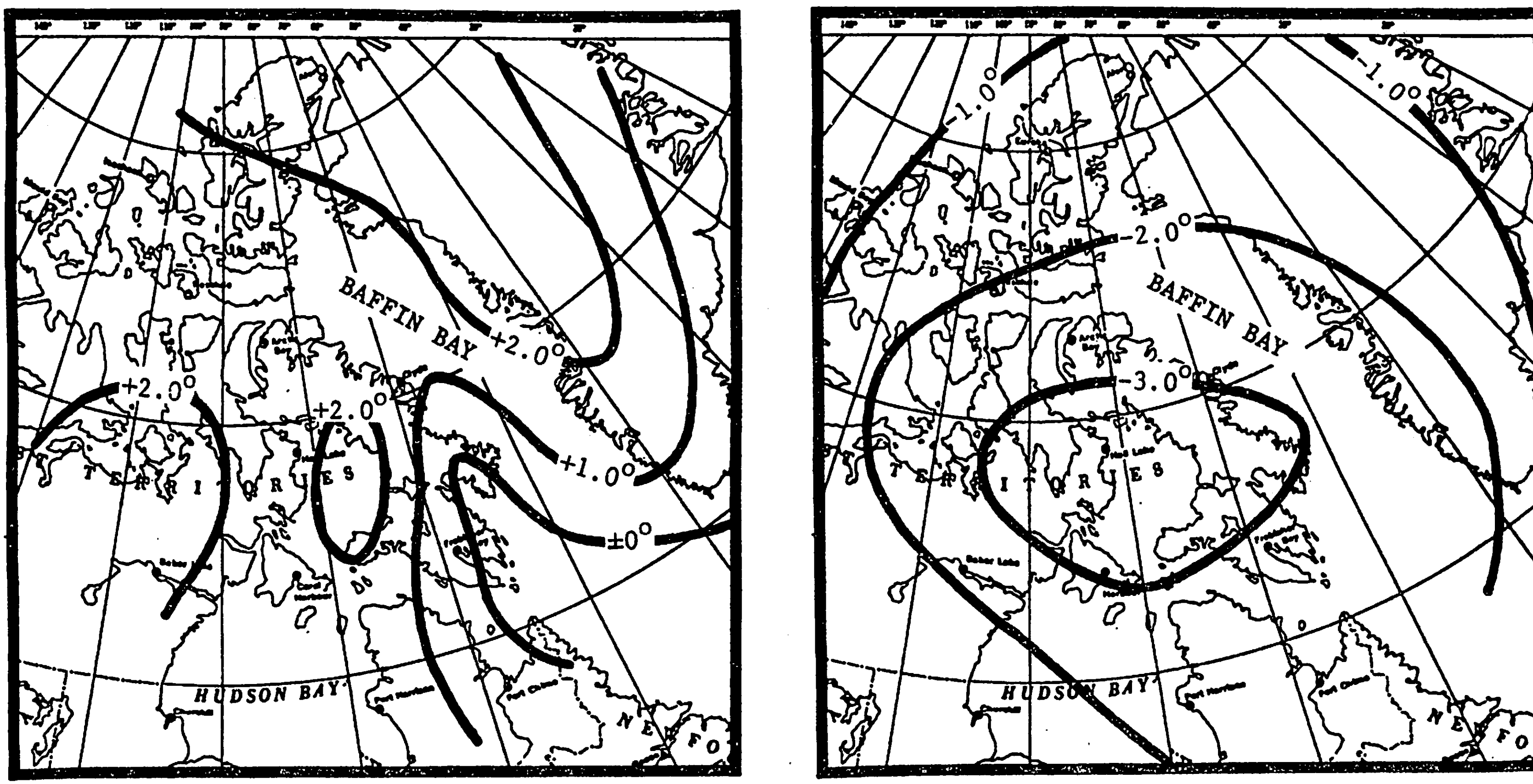


FIG. 2.1. Summer temperatures around Baffin Bay, 1943-76: average of Clyde, Upernavik, and Egedesminde. Yearly values, dotted line; 1-2-1 smoothed values, solid line.



Summer 1960 Summer 1972

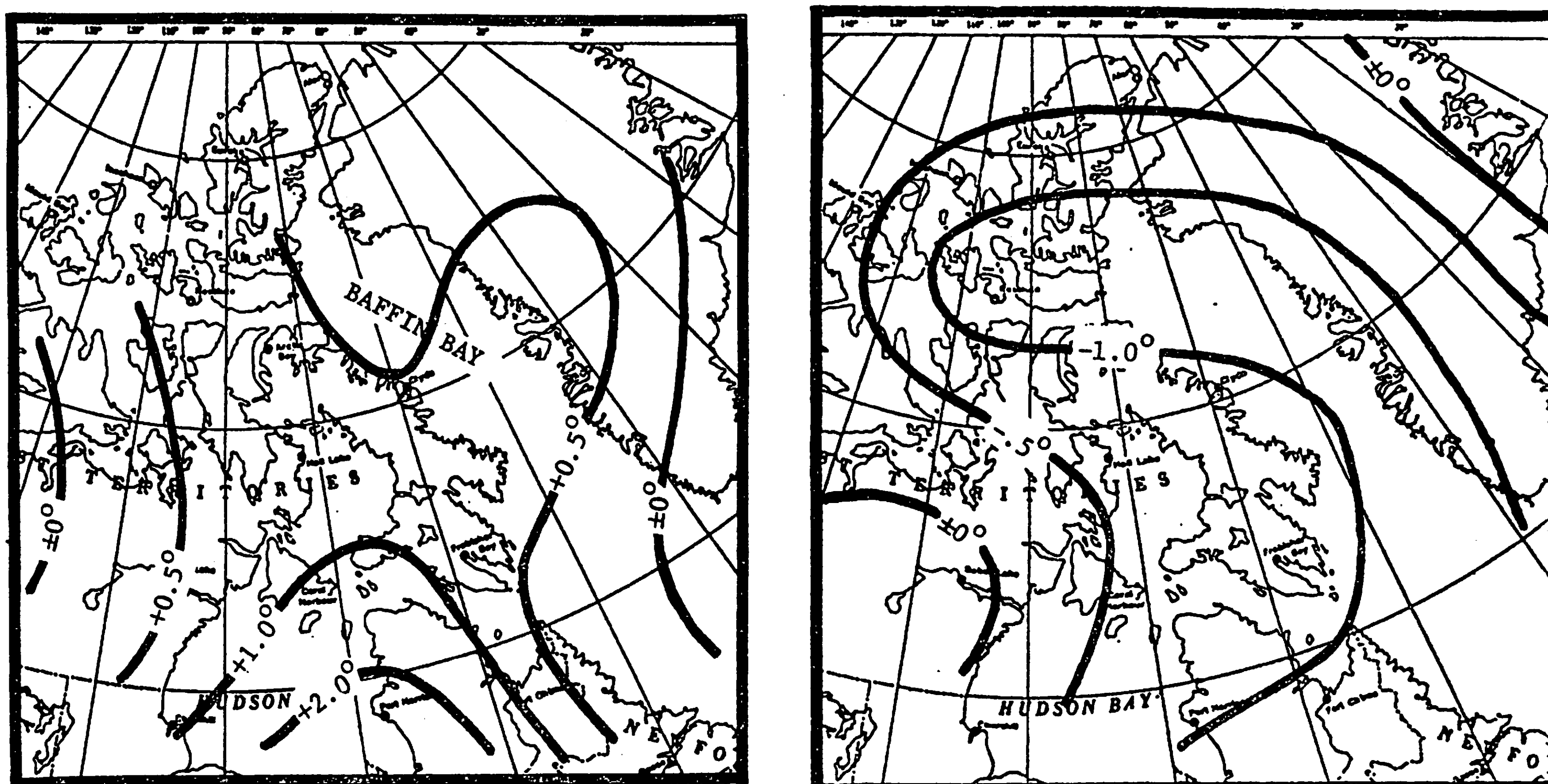
FIG. 2.2. Summer temperature anomalies around Baffin Island for 1960 (left) and 1972 (right): departures from 1941-70 climatological normals, in ^{o}C.

region it was the warmest summer on record -- and again, most extreme around Foxe Basin. It will be shown later in this chapter that this is not an unusual occurrence -- that the Foxe Basin area temperature varies in phase with the regional average more than does any other sub-region of the eastern Canadian Arctic. Elsewhere the 1960 temperatures were nearer normal, while the southeast tip of Baffin Island was slightly below normal for the summer. The spatial scale of the anomalies does not appear to differ much from those at middle latitudes, as illustrated by the monthly temperature departure maps for the United States published in the Monthly Weather Review.

In a sense, the two extreme summers, 1960 and 1972, marked the climaxes of two extreme short-term climate epochs. During the period 1963-73, the highest smoothed temperature in Fig. 2.1 was below the lowest value for 1948-62. The decades 1951-60 and 1964-73 are representative of these contrasting epochs, and analysis of averages and differences of various climatological parameters for these decades will constitute much of this study. Between the two decades, the Baffin Bay average summer temperature dropped 1.1^{o}C, from 4.6^{o} to 3.5^{o}. The t-value for the difference is 3.42, implying significance at the 99.9 percent level.

The period 1943-47 can be seen from Fig. 2.1 to be another cold epoch, averaging 3.8^{o}. The 0.8^{o} rise from this period to 1951-60 is significant at the 98 percent level. However, this warming cannot be analyzed in detail, due to a general lack of data.

The spatial extents of these two summer temperature changes are shown in Figs. 2.3a and 2.3b. In both cases, the area of



1943-47 to 1951-60

1951-60 to 1964-73

FIG. 2.3. Change in average summer temperatures around Baffin Island, 1943-47 to 1951-60 (left) and 1951-60 to 1964-73 (right), in °C.

cooling (or warming) covers the entirety of Baffin Island and Baffin Bay, as well as the other Canadian Arctic islands and West Greenland. Another expression of the spatial extent of the temperature fluctuations is a count of the number of stations (out of 10) on or near Baffin Island reporting summer temperatures below the 1941-70 climatological norms (Fig. 2.4). All ten of these stations are in the sector 58^{o}-80^{o}N, 50^{o}-100^{o}W, which is the same area analyzed for daily sea-level pressure patterns in Chapter Three. The main features noted for the time series in Fig. 2.1 are also apparent in Fig. 2.4, and are therefore all regional in extent, and not localized to Baffin Bay. The average number of stations with below normal temperatures for 1943-47, 1951-60, and 1964-73 are 7.0, 3.6, and 7.1, respectively. Again, the difference between the first two time periods has a 98 percent significance, and between the last two, 99.5 percent.

Baffin Bay Ice Conditions

The significance of the climate fluctuations can also be indicated by the associated environmental effects, such as the severity of sea ice conditions. Sea ice variations may have a considerable effect on arctic and global climate as a feedback mechanism, not to mention their effect on human activities in the area. However, the question of relative importance of the mutual feedback mechanisms between sea ice and summer temperatures is beyond the scope of this study, and ice conditions are presented here as further evidence of the climate fluctuations.

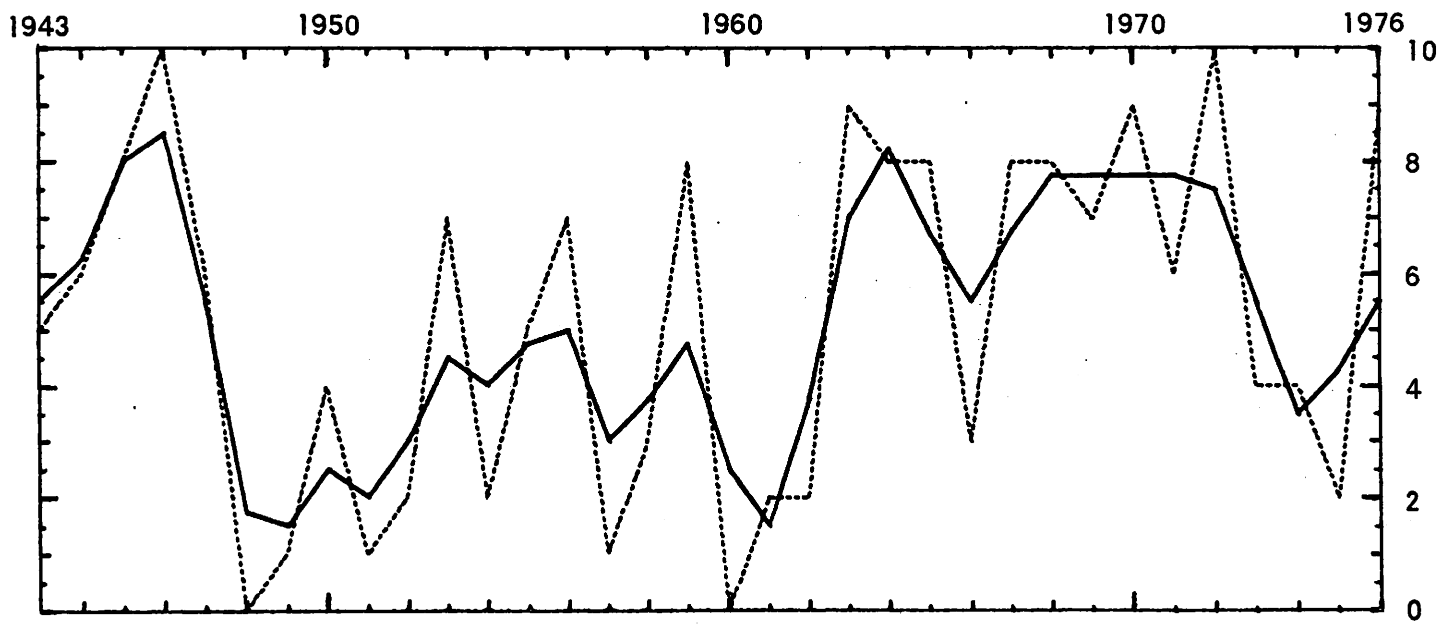


FIG. 2.4. Number of stations (out of 10) reporting average summer temperatures below the 1941-70 climatological norm, 1943-76. Yearly values, dotted line: 1-2-1 smoothed values, solid line.

In Baffin Bay, pack ice formation normally starts in mid-October, and reaches maximum extent in April, when the Bay is ice-covered except for two open areas: the so-called North Water at the northern end of the Bay, and in the southern part of the Bay along the Greenland coast (Dey et al., 1979). During break-up, the last area to clear of ice is usually north of the Cumberland Peninsula in southeastern Baffin Island (Wittmann and Burkhardt, 1973; Weaver, 1974; Crane, 1978). The total clearing usually occurs by late September, but the date can vary considerably, and in some years total clearing of the ice never occurs (Dunbar, 1972). Because of the long record available, the date of total clearing is selected as the indicator of sea ice conditions for this analysis; the dates for 25 seasons (updated from Dunbar, 1972) are presented in Fig. 2.5. Once again the major features of the Baffin Bay temperature series (Fig. 2.1) are readily apparent. Most important, of course, is the dramatic deterioration of ice and climate conditions around 1963. During the decade preceding that year, the ice had cleared by mid-September every year but one; the decade following 1963 had seven occasions when the ice failed to clear.

The date of ice clearing is given the numerical value of 1 for the first week in August, through 8 for the last week in September, and 9 and 10 for total clearing not occurring at all; a Spearman (rank) correlation between these values and those of Baffin Bay summer temperatures yields a coefficient of .62 for the annual values and .74 for the 1-2-1 smoothed values. The higher correlation for the smoothed values is not surprising, since the smoothing reduces the number of independent samples, and tends to

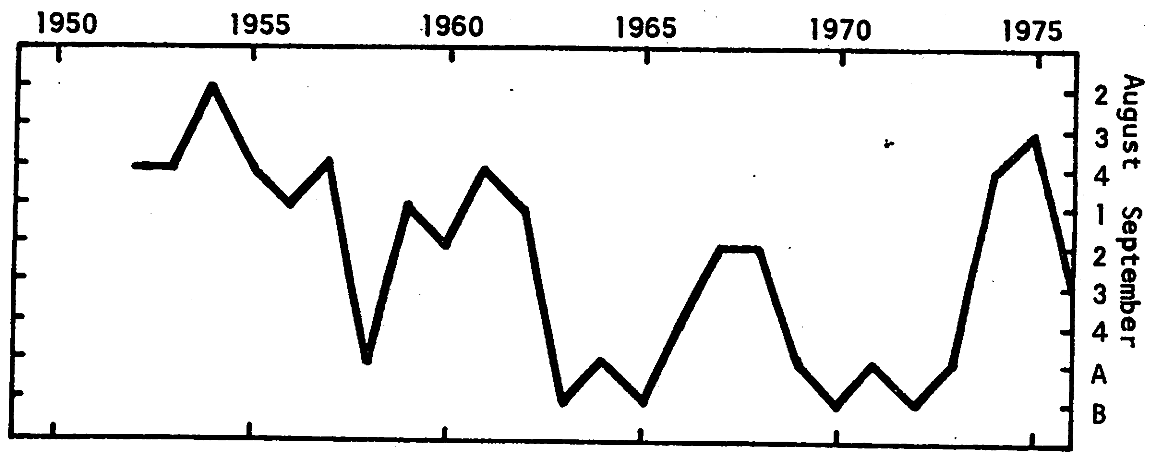


FIG. 2.5. Date of total clearing of ice from Baffin Bay, by week, 1952-76. A and B indicate ice did not clear, with, respectively, small and appreciable amounts remaining.

average out the component of ice breakup variability due to factors acting at least partially independently of mean summer temperature, such as sunshine, wind, storm events, and thawing during May and September. The latter may have been particularly important for the 1952, 53, and 54 thawing seasons, when total clearing occurred very early during summers of about average warmth. All three summers were preceded by very warm springs, however, with April-May departures of +3°C.

Winter (September-May) temperatures are found to be uncorrelated (r = .07) with subsequent summer temperatures on the annual to decadal time scales, and their inclusion with summer temperatures in a stepwise regression adds less than a percent to the correlation with date of ice clearing. This confirms findings by Barry and Jacobs (1974) for Baffin Bay (based on three years of data) and by Rogers (1978) for the Beaufort Sea, namely, that summer temperatures are the dominant determinant of the severity of the ice season.

Arctic Average Temperatures

To some extent, the Baffin area would be expected to participate in the general climate variability of the Arctic as a whole, with an additional component of variability on the regional scale. To compare the Baffin temperature fluctuations with those elsewhere in the Arctic, and with the Arctic average, a uniform data set covering the Arctic is needed. For this purpose summer average temperatures from 37 stations (see map, Fig. 2.6) have been interpolated to yield values at 10° longitude intervals at

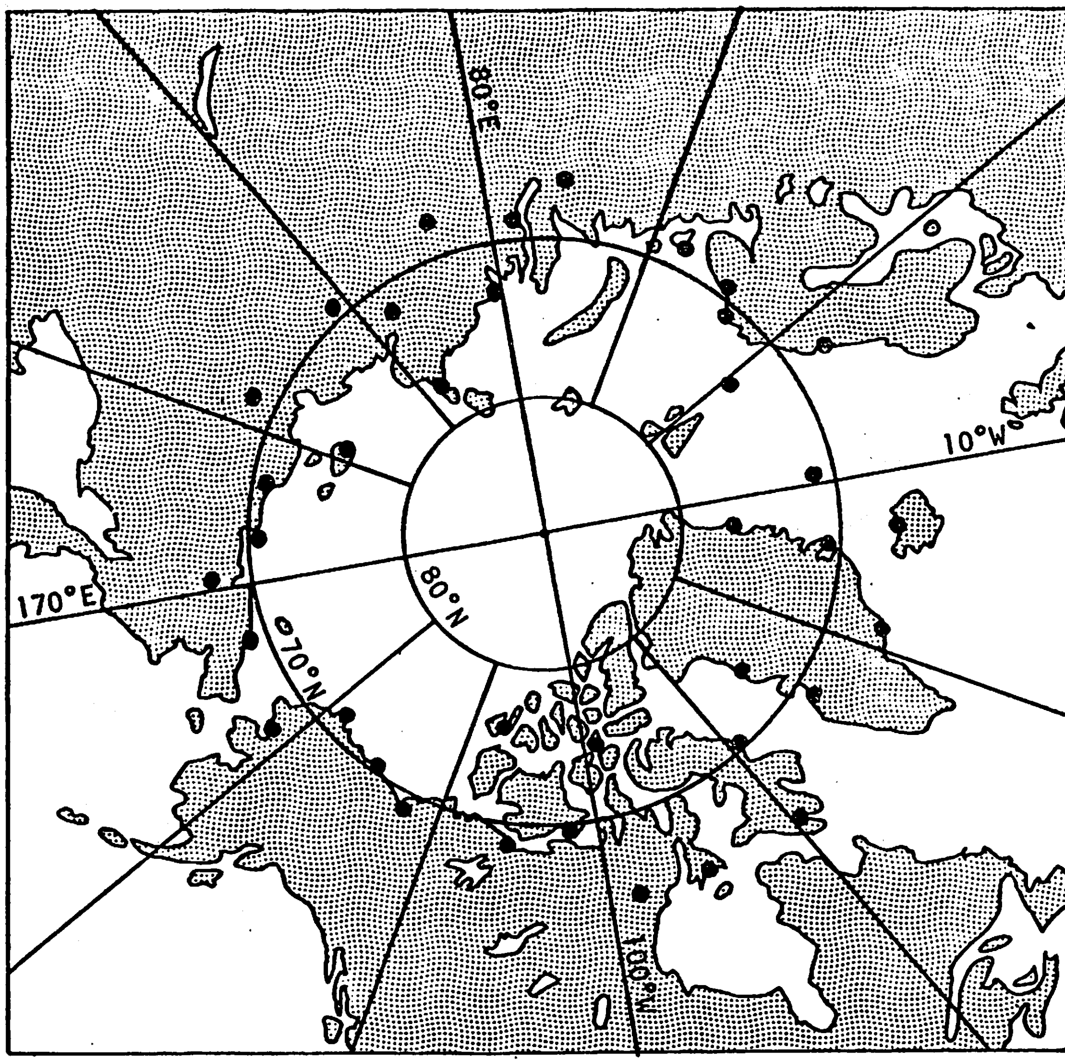


FIG. 2.6. Locator map of 37 stations used in determining summer surface temperatures at $70^{\circ}N$.

70°N latitude for the period 1951-76. Most of the stations are at coastal or island locations, but a close correlation between the derived surface temperature variations and those of the lower troposphere (Chapter Five) indicates that these stations are representative of conditions across the Arctic.

The temperature at each latitude-longitude point is a weighted average of the three nearest stations that form a triangle enclosing the point (to avoid undue bias towards temperature anomalies on one side of the point). The weighting factor used is $W = (1.0\text{-}d/1900\text{km})^2$, where d is the distance between the station and the point. This weighting factor is equal to $r^2(d)$, where r(d) is the linear regression value of r vs. d for correlations between 48 pairs of stations around the Arctic. Missing data are rare, and usually limited to only one of the three months in a summer average. When encountered, missing values have been replaced by a value interpolated from nearby stations, using the same method as above.

The average distances from the lat-long point to the three stations used in the interpolation are 225, 480, and 710 km; the respective normalized weighting factors are .45, .32, and .23. This interpolation results in an effective spatial smoothing of the data. The effect of this smoothing on different longitudinal wavenumbers of the temperature variation at 70°N (Holloway, 1958) is shown in Table 2.3. The half-wavelength represents the longitudinal extent of a positive or negative anomaly associated with that wave-number. Temperature anomalies such as the major features in Figs. 2.2 and 2.5 would, because of their extent, be retained at

TABLE 2.3. Spatial Smoothing Effect of the Interpolation of Station Temperatures to 70°N Latitude-Longitude Points

Wavenumber	$\lambda/2$ (deg)	$\lambda/2$ (km)	Amplitude Response
0	--	----	1.00
2	90	3423	.96
4	45	1711	.83
6	30	1141	.64
8	23	855	.41
10	18	685	.17
12	15	570	.00

nearly full amplitude, while small-scale features, such as the small area of below average temperatures over extreme southeast Baffin Island during summer 1960 (Fig. 2.2a), would be virtually eliminated by the smoothing.

The interpolated temperature data are displayed in Figs. 2.7 through 2.11. The 26-year means and standard deviations of the summer temperatures vs. longitude are shown in Fig. 2.7. The influence of the long wave pattern of the Northern Hemisphere upper air circulation (Fig. 4.1) is readily apparent, with the highest summer temperatures associated with ridges (at 40°E, 110°E, and 130°W longitude), and the lower temperatures with troughs (80°E, 170°E, and 60°W longitude). The warmest summers occur around Lapland and in north central Siberia, probably because of the relatively warm offshore waters for the former and the large land

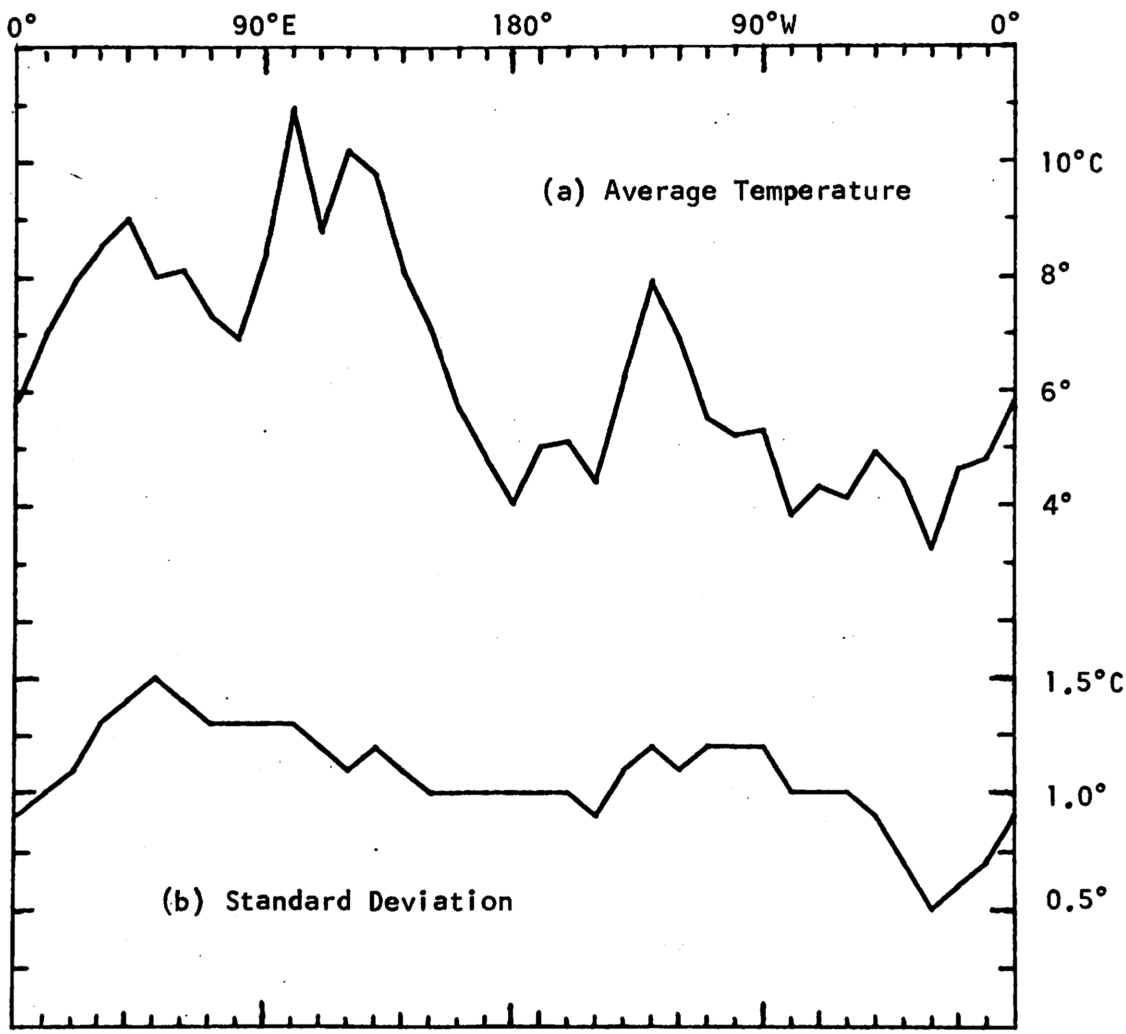


FIG. 2.7. Longitudinal variation of summer temperature at 70°N, 1951-76: 26-year average (top) and interannual standard deviation (bottom).

mass at 70°N for the latter. The standard deviations are somewhat higher for those longitudes with continental land areas to the south and colder water to the north -- so changes in wind direction can have large effects on the temperature. The high variability around the Barents Sea may be due to the mean trough-ridge system in that area not being as "fixed" in position by geographical factors as the major troughs along the eastern edges of Asia and North America (Namias, 1958). The effect of changes in the position and intensity of the long wave pattern on the temperatures of Baffin Island are examined in detail in Chapter Four.

Yearly summer temperatures at five different longitudes, and for the Arctic as a whole (zonal average at 70°N), are shown in Fig. 2.8. Two of the longitudes -- 50°W (West Greenland) and 80°W (Foxe Basin) -- are chosen to illustrate, along with Fig. 2.1 (60°W -- Baffin Bay), the relation between temperature trends on opposite sides of Baffin Bay. The other longitudes are approximately equally spaced around the rest of the Arctic, at 150°W (north coast of Alaska), 130°E (northeast Siberia), and 40°E (northeast Europe), and are located near mean ridges in the upper westerlies (Fig. 4.1).

The correlations between summer temperatures across Baffin Bay are generally good, particularly for the longer-term fluctuations. Dunbar (1954) has proposed that West Greenland should respond more quickly to cooling trends than would Baffin Island due to the West Greenland Current, but this effect is not apparent in Fig. 2.8. On a yearly basis, the correlation between summer temperatures at 50°W and at 80°W is 0.50. Among the other

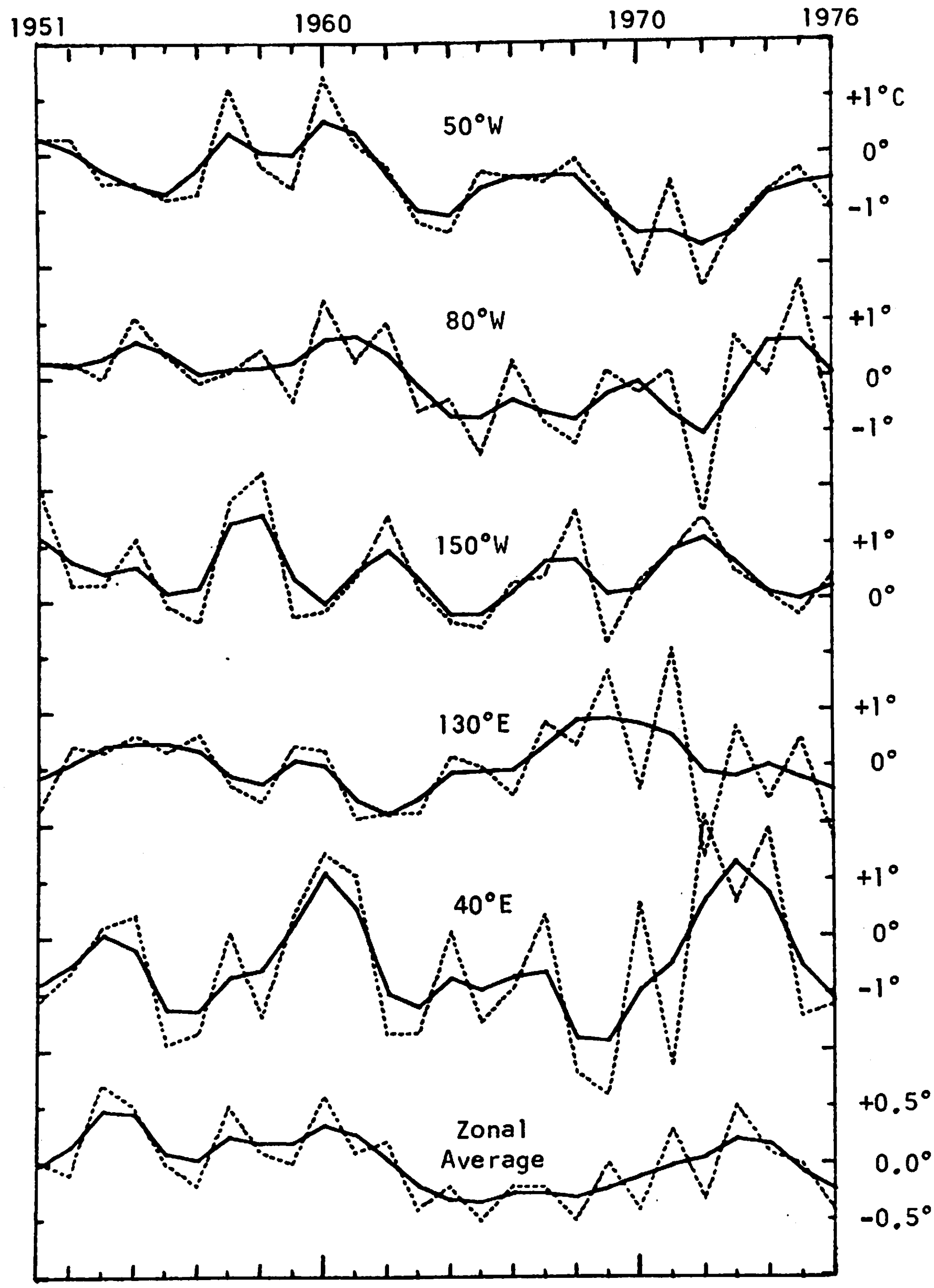


FIG. 2.8. Summer surface temperatures at 70°N, for five different longitudes and zonal average, 1951-76. Values are departures from long-term average. Yearly values, dotted line; smoothed values, solid line.

three longitudes, correlations in the year-to-year variability are generally lacking, but a widespread tendency towards lower temperatures after the early 1960's is apparent. The decline is more distinct for the arctic average summer temperature; the drop of 0.4°C between 1951-60 and 1964-73 passes the t-test at the 99 percent level, and is consistent in sign and magnitude with the zonal cooling found for this latitude by other researchers (van Loon and Williams, 1976; Walsh, 1977).

The summer temperatures around the Arctic at 70°N are displayed differently in Fig. 2.9. The contoured temperature values are departures of the interpolated summer temperature from the 26-year average for each longitude, and are time-smoothed by a 1-2-1 running mean. Some coherencies in longitude and time are noticeable, the most pronounced being, of course, the widespread cooling of the early 1960's. There also appears to be a tendency apparent in the figure for temperature anomalies to drift in longitude from year, with two separate drifts, eastward at 10°/year and westward at 25°/year.

Further inspection of Fig. 2.8 shows that of the six longitudes presented, the temperatures at 80°W (Foxe Basin) appear to have the closest correlation with the zonal mean. Fig. 2.10 confirms that this is the case, and that the correlation exceeds 0.5 (99.5 percent significance) over the entire Baffin Bay -- Keewatin sector. Examination of the correlation matrix between longitudes shows that 80°W is positively correlated with all longitudes except 0° to 40°E, and is strongly ($r > 0.5$) correlated with the entire sector from 50°W to 120°W -- from West Greenland

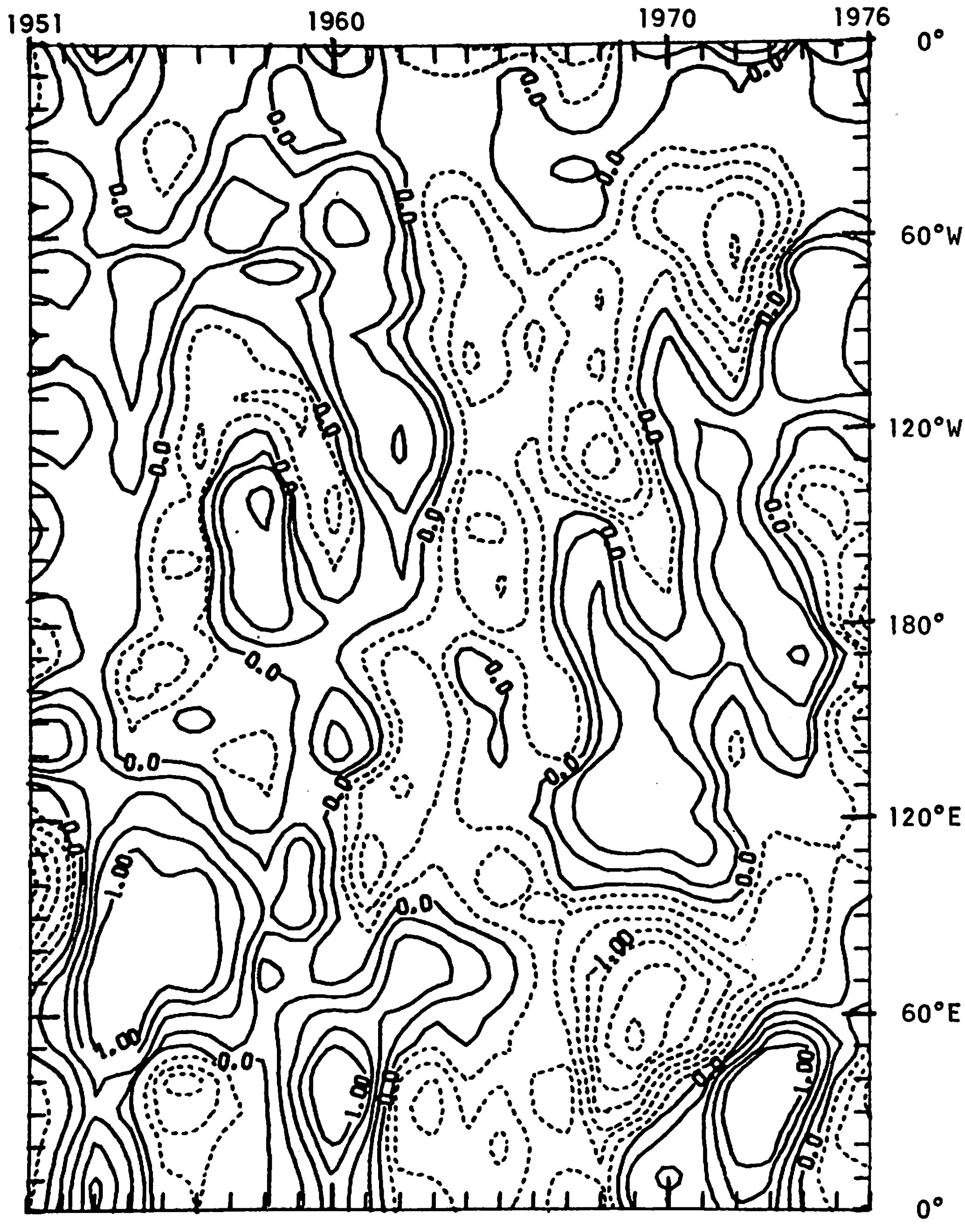


FIG. 2.9. Summer surface temperature anomalies at 70°N, 1951-76. Values shown are departures from 1954-76 average at each longitude, and are smoothed in time by a 1-2-1 running mean. Dotted contours indicate negative departures; contour interval, 0.25°C.

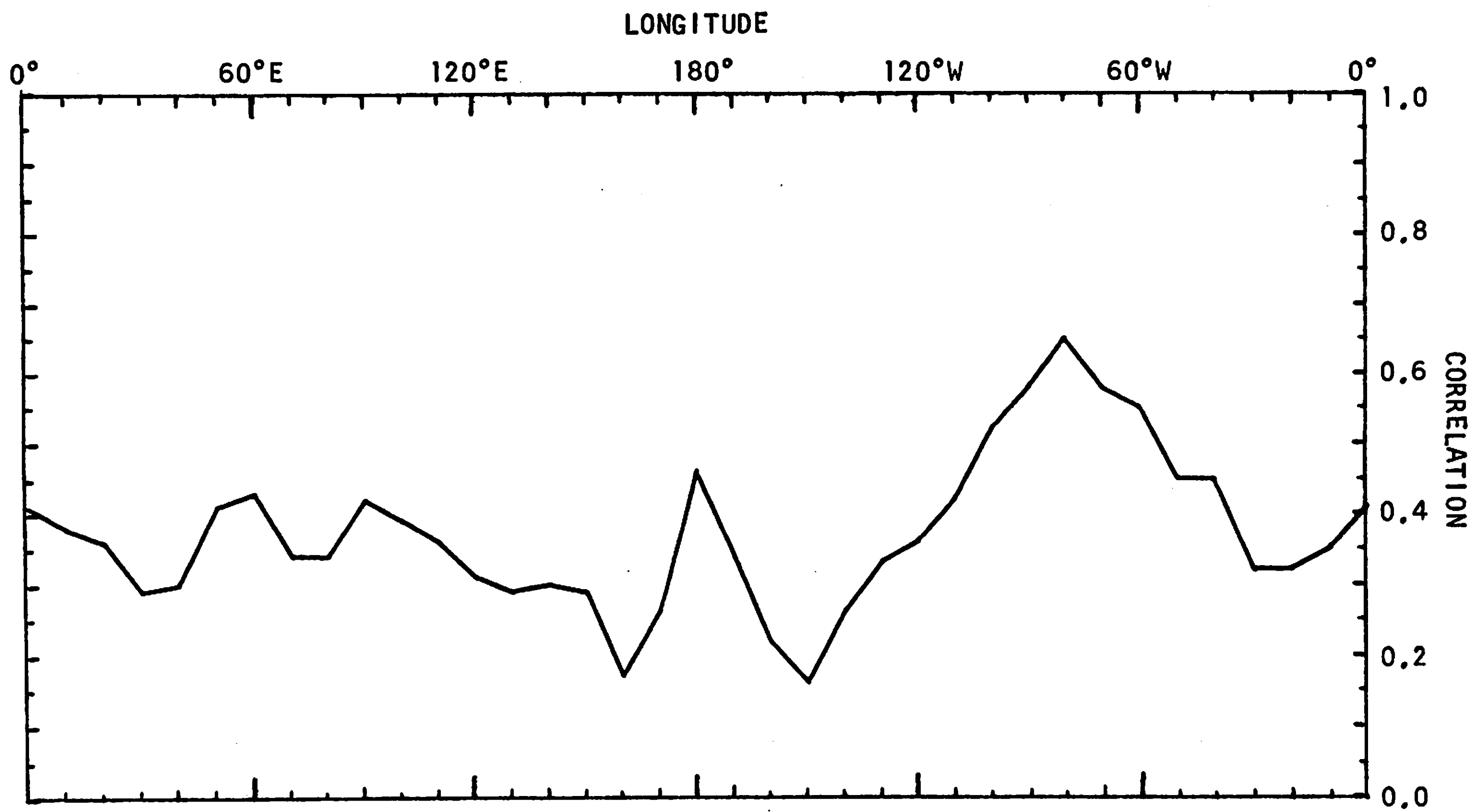


FIG. 2.10. Correlation of summer temperatures at different longitudes at 70°N with zonal average, 1951-76.

to Banks Island. Indeed, all longitudes in the range $10^{\circ}W$ to $140^{\circ}W$ (North Atlantic to the Yukon) are positively correlated with each other. Baffin Island is at the center of this large coherently varying sector. Walsh (1977) found a similar result from an eigenvector analysis of monthly average arctic surface temperatures (all seasons), with variance centers of the first two eigenvectors located, respectively, over Baffin Island and the District of Mackenzie.

The extension of these correlations to the decadal time scale is indicated by Fig. 2.11. While all but three rather small sectors experienced a cooling between 1951-60 and 1964-73, most of the effect is concentrated in two sectors: northwest Siberia, and Baffin Island-West Greenland. Baffin is positively correlated with the zonal mean temperature on both the yearly and decadal time scales, with variations about three times those of the zonal mean. The zonal average and regional components of temperature variability operate in unison around Baffin Island more than anywhere else in the Arctic, making the Baffin temperature a sensitive indicator of summer conditions across the Arctic as a whole. Paleoclimatic studies of this area will take on added significance if this correlation extends to longer time scales.

Some of the enhanced response of Baffin area temperatures to zonal temperature variations may be due to the increased albedo of expanded ice and snow cover during cold summers, but in Chapter Four it is shown that much of the response can be explained by

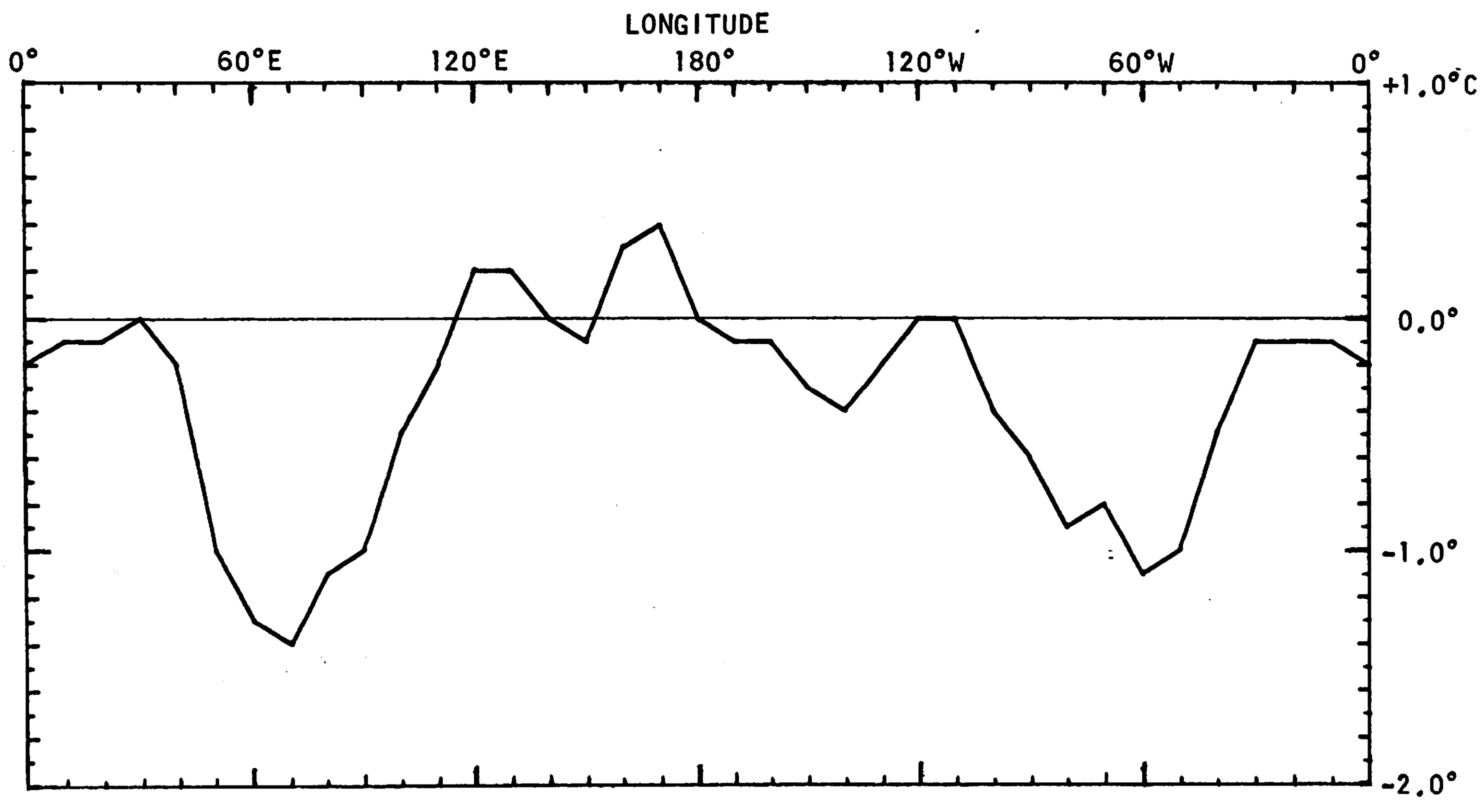


FIG. 2.11. Longitudinal variation of decadal summer surface temperature change at $70^{O}N$, 1951-60 to 1964-73.

changes in the regional circulation patterns. The next chapter examines the observed Baffin summer temperature variations in terms of synoptic pressure systems affecting the area.

CHAPTER THREE

SYNOPTIC CLIMATOLOGY

Introduction

The area around Baffin Island, especially the waters to the northeast (Baffin Bay) and to the southeast (Davis Strait and the Labrador Sea), is a region of frequent and persistent cyclonic activity in all seasons of the year. For much of the year most of the cyclones are slow-moving cold tropospheric lows far removed from their frontal origins, but during the summer frontal wave cyclones often penetrate the region (Hare and Hay, 1974). However, the regional circulation is not always cyclonic, since a preferred track for polar anticyclone outbreaks passes west of Baffin Island (Klein, 1957), and these systems sometimes pass over Baffin Island (Reed and Kunkel, 1960). The influence of synoptic scale surface circulation systems on daily weather is shown by Bradley (1974), who found characteristic temperature differences of about 5°C over eastern Baffin Island between days with a low over Davis Strait (cold) and days with a high over Davis Strait and a low to the west (warm). Although the frequencies of occurrence of different synoptic patterns varies from summer to summer, their correlations with summer average temperatures are not always consistent. For example, anticyclonic conditions are usually warm on a daily basis, but their frequency increased during the late 1960's as summer temperatures decreased (Barry _et al_., 1975). Despite these problems and potential limitations, the synoptic climatological approach to the question of interannual variability in the Arctic has been shown to be valid and useful by Alt (1978) and Crane (1978), who examined, respectively, mass balance fluctuations of the Devon Island ice cap and variations in sea ice extent in the Davis Strait - Labrador Sea area.

In this chapter a catalog of objectively derived synoptic surface pressure pattern types for the Baffin region is utilized to extend the synoptic climatological studies over the period 1946-74. First, the daily summer temperature characteristics of the different synoptic types are established, then the interannual variability of synoptic circulation features (cyclones and anticyclones) and its connection to temperature variability is discussed. It is shown that the synoptic pressure patterns do not explain the interannual variability of local temperatures; therefore the interpretation of these patterns will emphasize their role as atmospheric circulation features rather than as determinants of local temperatures.

The Synoptic Pressure Pattern Types

The classification of Baffin area pressure patterns is based on normalized surface pressure distributions over 37 grid points in the sector 58^{o}-80^{o}N, 50^{o}-100^{o}W; the normalization is done for each day by dividing the grid point departures from the 37-point mean by the standard deviation across the grid. From a cross-comparison of daily pressure patterns in a 58-month sample, 28 "key days" were determined to be typical of the variety of pressure patterns that can occur around Baffin Island. The primary criterion for comparison is the root-mean-square difference of normalized grid pressures between two days; a lower r.m.s. difference means a closer fit between the two patterns. Daily gridded pressures from January 1, 1946 to August 31, 1974 are classified by their best r.m.s. fit to one of the key days; if the best r.m.s. difference of normalized pressures is greater than 1.0, the day is not classified. Only 2 percent of the days are not classified. Details of the typing scheme are provided by Barry and Keen (1978).

The pressure patterns for the 28 key days are shown in Fig. 3.1, arranged in groupings based on the location of low pressure centers. The contour interval is 0.5σ, σ being the standard deviation of the 37 grid point pressures on the key day. It should be noted that these synoptic pressure pattern types give no indication of the intensity of the pressure features on the grid, but the reality of these features on days with weak pressure fields is verified by a check of the original surface charts for days with very small σ (3mb or less).

It is important to know the degree of variation that can occur within the set of days classified as a particular synoptic type to assess the significance of the differentiation between types. Using the entire 29 years of data, the percentage frequency of occurrence of the grid's lowest pressure at each grid point for each synoptic type was computed. The results for five representative types are displayed in Fig. 3.2, and show that the lows tend to concentrate in a small portion of the grid -- typically, 50 percent of the occurrences occupying 2 to 4 grid points. Thus there is little overlap between similar types, in terms of the location of the dominant low pressure feature.

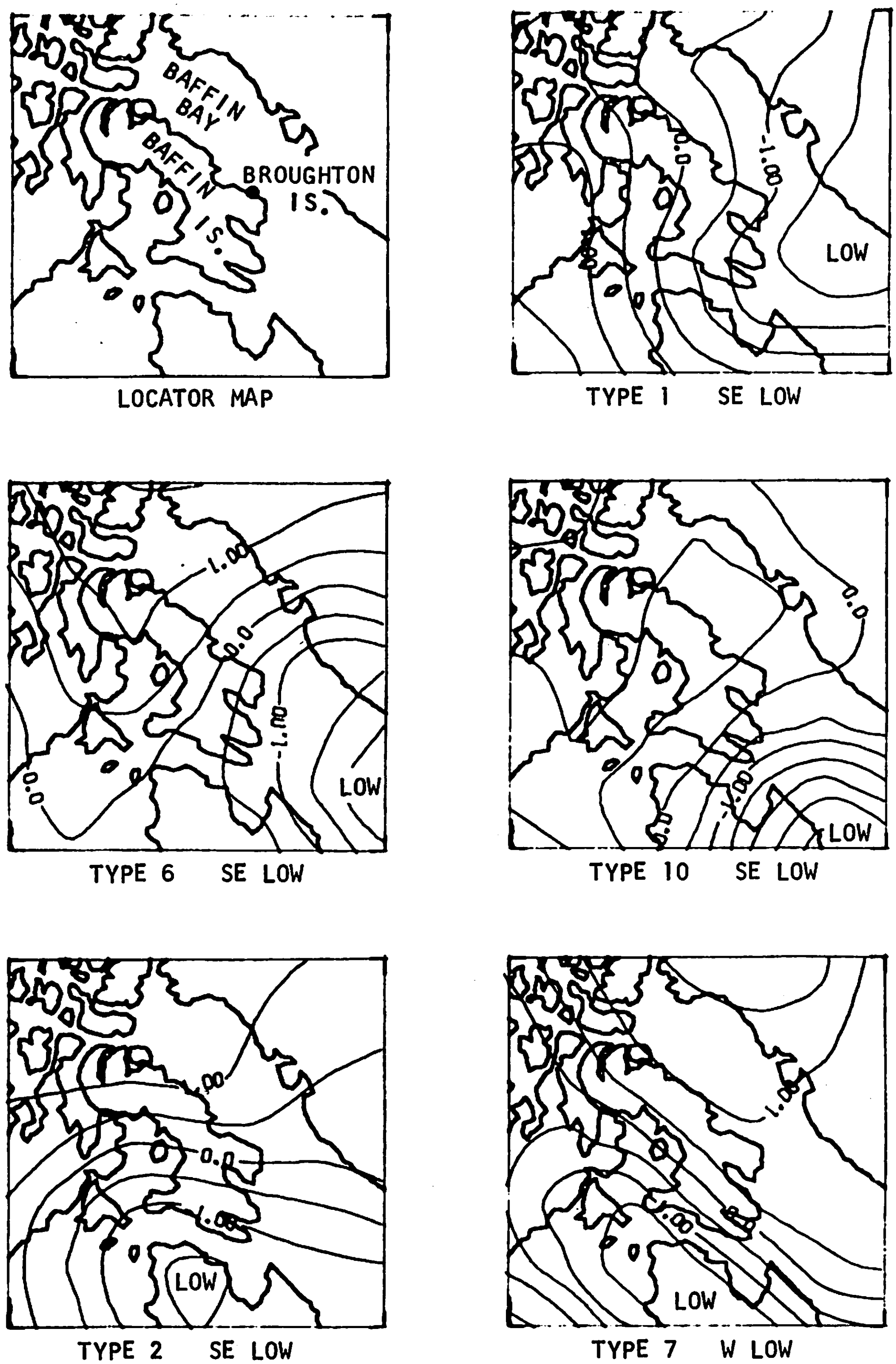
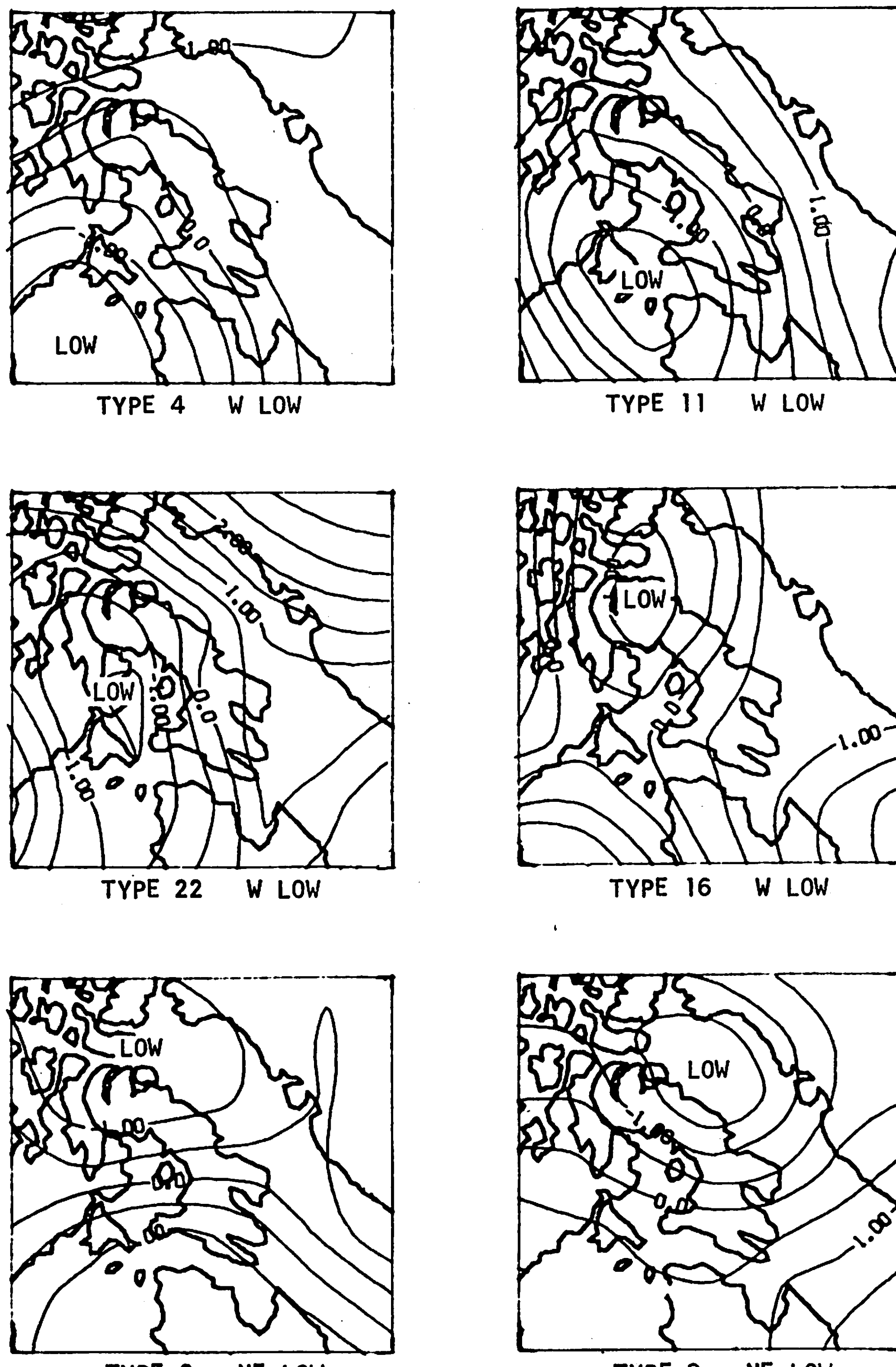


FIG. 3.1. The Baffin Island synoptic pressure pattern types: normalized pressure maps for the key days.



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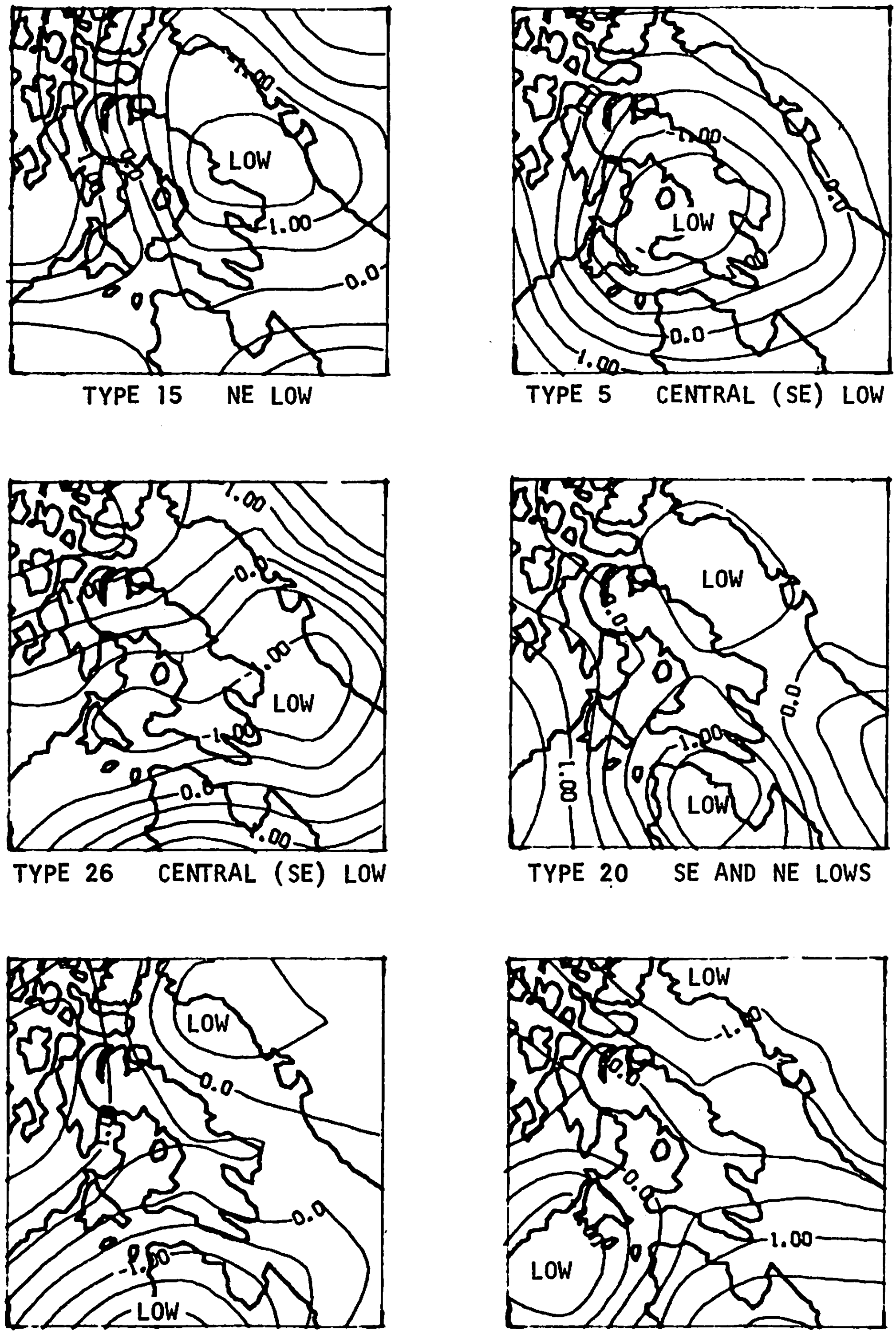
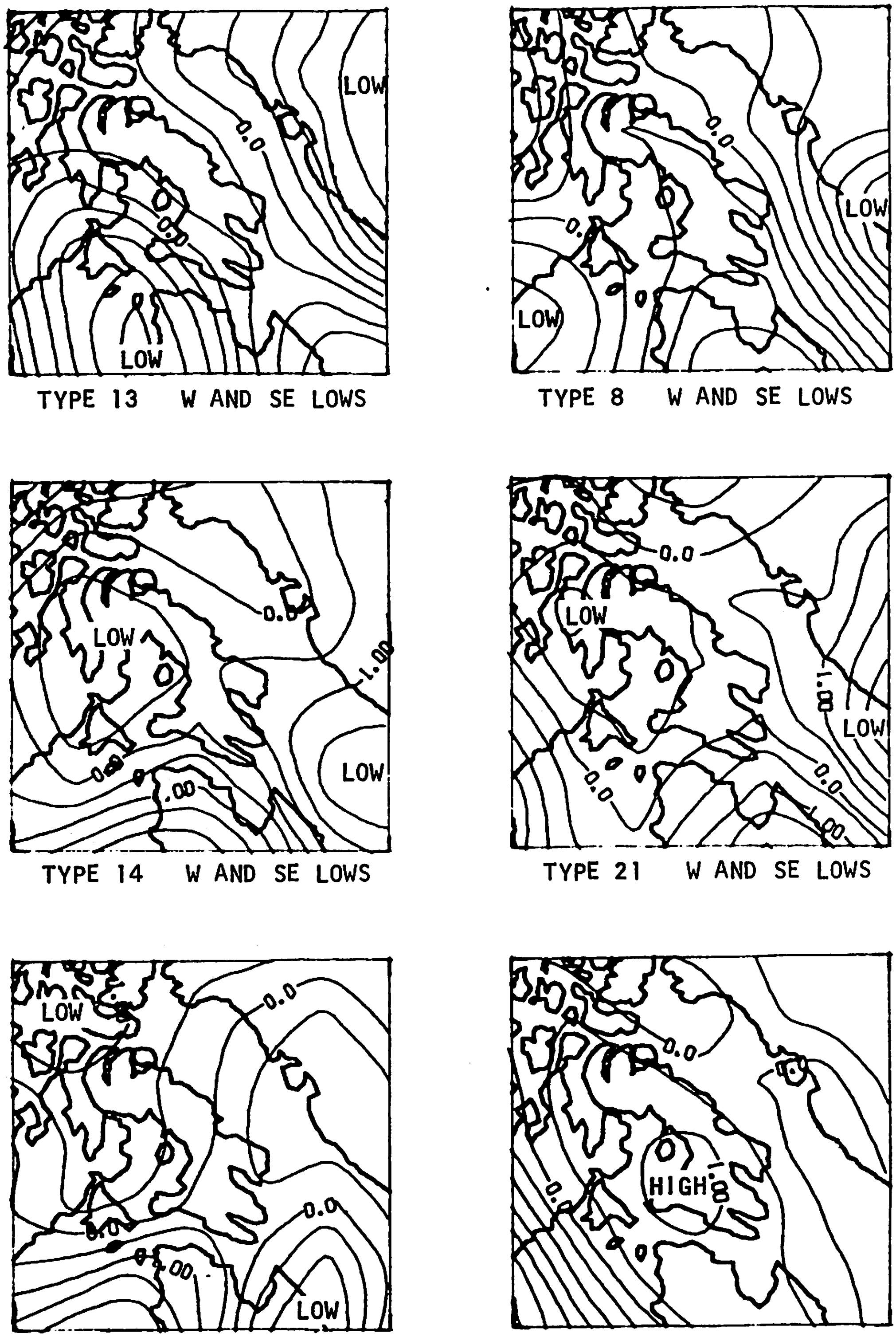


FIG. 3.1. The Baffin Island synoptic pressure pattern types: normalized pressure maps for the key days.



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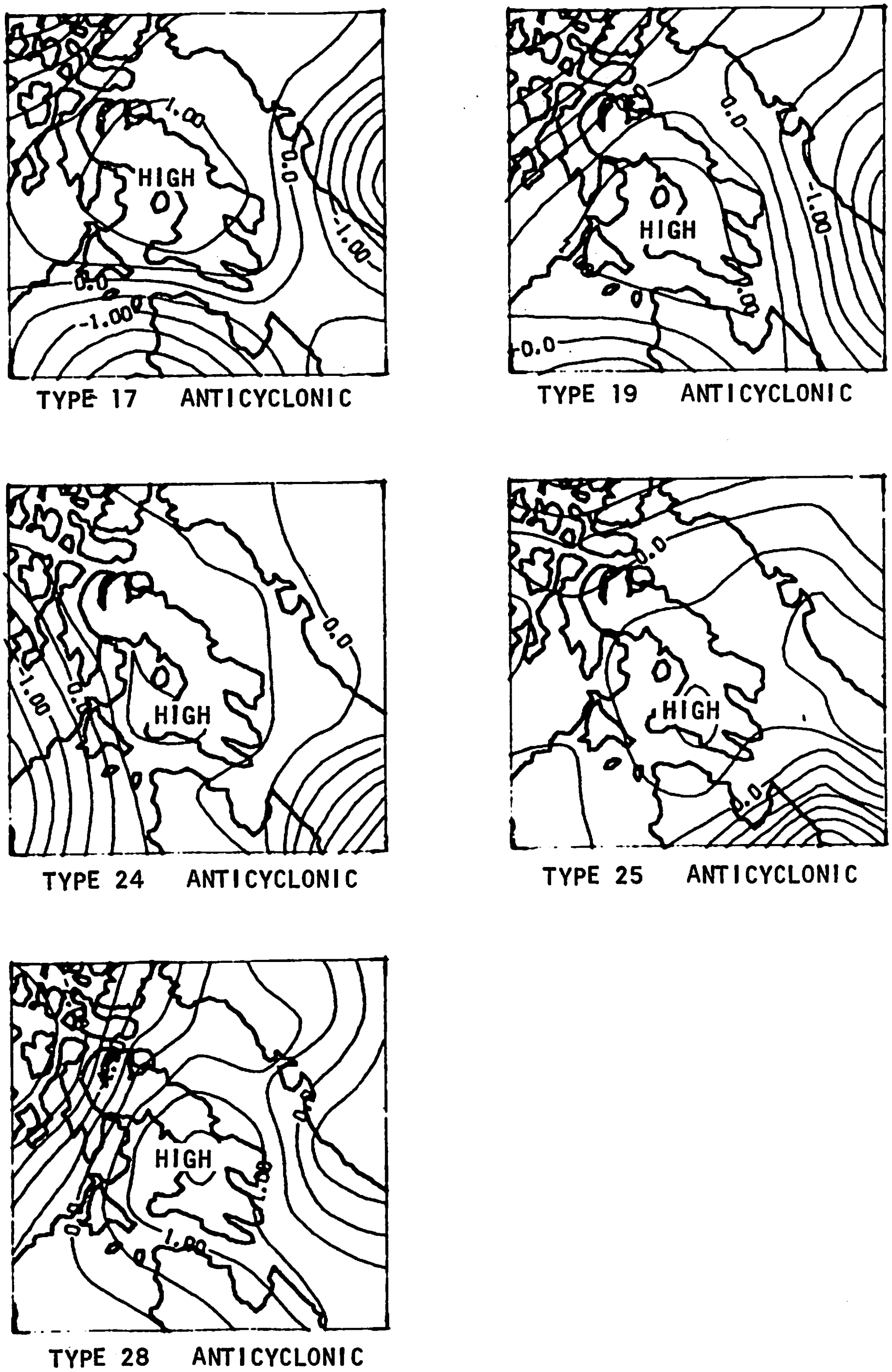
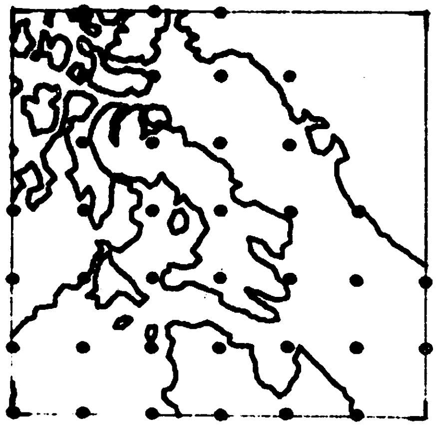


FIG. 3.1. The Baffin Island synoptic pressure pattern types: normalized pressure maps for the key days.



Grid point map

	0	0	0			
	0	0	0	0		
	0	0	0	0		
0	0	0	0	0	1	
0	0	0	0	1	4	21
0	0	0	0	1	5	52
4	1	0	0	1	7	

Type 6 SE Low

	0	0	0			
	0	0	0	0		
	0	0	0	0		
3	0	0	0	0	0	
6	11	2	0	0	0	0
13	15	7	0	0	0	0
7	22	10	2	0	0	

Type 4 W Low

	6	6	4			
	3	12	21	9		
	2	9	16	3		
0	0	2	1	2	0	
0	0	0	0	0	0	0
0	0	1	0	0	0	0
0	0	1	1	0	0	

Type 9 NE Low

	1	0	0			
	0	1	1	1		
	0	5	7	2		
1	1	14	15	8	2	
0	1	9	21	7	0	1
0	0	1	0	0	0	0
0	0	0	0	0	0	

Type 5 Central (SE) Low

	6	1	2			
	0	0	8	1		
	0	1	4	3		
0	0	0	0	2	0	
0	0	0	7	3	0	0
0	0	1	14	9	0	0
0	0	2	19	13	4	

Type 20 SE and NE Lows

FIG. 3.2. Percentage distribution of occurrence of the lowest pressure on the Baffin synoptic type grid for five representative types, 1946-74. For a random distribution the value at each grid point would be 2.7 percent.

The complete catalog of daily types for January 1946 through August 1974 is listed in Appendix A. Sixty-two days with missing grid point data have been typed subjectively, using hand-analyzed maps on file at NCAR. These days are included in the catalog.

One of the purposes of this study is to determine how much of the interannual variability of the Baffin area summer climate can be explained by changes in the frequency of synoptic events. A necessary first step is to find the thermal characteristics of each synoptic type, in terms of the daily temperature anomaly to be expected on a given day with that particular pressure pattern.

The climate station selected for this comparison is Broughton Island, located on the northeast shore of the Cumberland Peninsula. Broughton Island was the site of INSTAAR's fast-ice breakup field study (Barry, Jacobs et al., 1978) from 1971 through 1975. At an elevation of 581 meters, the station is occasionally above the low-level inversion that blankets much of the Arctic, and should therefore be more responsive to changes in the synoptic scale flow pattern than a station nearer sea level.

The daily departures of the Broughton Island temperature from the long-term daily average are computed for each summer (June-July-August) day during the period of record (1959-1970) and summarized by synoptic type. The daily departure is based on the annual cycle of "normal" daily temperatures given in Table 3.1. Within each month, the daily values are calculated from a quadratic interpolation function (Temp. = $a+bt+ct^2$) fit to the average values on the 15th of the given ($t = 0$), preceding ($t = -1$), and following

TABLE 3.1. Computed Daily Normal Temperatures (^{o}C) at Broughton Island

Month:	1	2	3	4	5	6	7	8	9	10	11	12
Day: 1	-23	-25	-25	-21	-12	-4	3	6	1	-5	-11	-19
2	-23	-25	-25	-21	-11	-3	3	5	1	-5	-12	-19
3	-23	-25	-25	-20	-11	-3	4	5	1	-5	-12	-19
4	-23	-25	-25	-20	-11	-3	4	5	0	-5	-12	-20
5	-24	-25	-25	-20	-10	-3	4	5	0	-5	-12	-20
6	-24	-25	-25	-20	-10	-3	4	5	-0	-6	-13	-20
7	-24	-25	-24	-19	-9	-2	4	5	-0	-6	-13	-20
8	-24	-25	-24	-19	-9	-2	4	5	-0	-6	-13	-20
9	-24	-25	-24	-19	-9	-2	5	5	-1	-6	-14	-20
10	-24	-25	-24	-18	-9	-2	5	4	-1	-6	-14	-21
11	-24	-25	-24	-18	-8	-2	5	4	-1	-7	-14	-21
12	-24	-25	-24	-18	-8	-1	5	4	-1	-7	-14	-21
13	-24	-25	-24	-17	-8	-1	5	4	-1	-7	-15	-21
14	-24	-25	-24	-17	-7	-1	5	4	-2	-7	-15	-21
15	-24	-25	-24	-17	-7	-1	5	4	-2	-7	-15	-21
16	-24	-25	-24	-17	-7	-1	5	4	-2	-8	-15	-22
17	-24	-25	-24	-16	-6	-0	5	3	-2	-8	-16	-22
18	-24	-25	-23	-16	-6	-0	6	3	-2	-8	-16	-22
19	-25	-25	-23	-16	-6	0	6	3	-3	-8	-16	-22
20	-25	-25	-23	-15	-6	0	6	3	-3	-9	-16	-22
21	-25	-25	-23	-15	-6	1	6	3	-3	-9	-17	-22
22	-25	-25	-23	-15	-5	1	6	3	-3	-9	-17	-22
23	-25	-25	-23	-14	-5	1	6	2	-3	-9	-17	-22
24	-25	-25	-22	-14	-5	1	6	2	-4	-9	-17	-23
25	-25	-25	-22	-14	-5	2	6	2	-4	-10	-18	-23
26	-25	-25	-22	-13	-4	2	6	2	-4	-10	-18	-23
27	-25	-25	-22	-13	-4	2	6	2	-4	-10	-18	-23
28	-25	-25	-22	-13	-4	2	6	1	-4	-11	-18	-23
29	-25		-21	-12	-4	3	6	1	-4	-11	-18	-23
30	-25		-21	-12	-4	3	6	1	-5	-11	-19	-23
31	-25		-21		-4		6	1		-11		-23
AVERAGE	-24.2	-24.9	-23.5	-16.6	-7.2	-.6	5.0	3.6	-1.9	-7.7	-15.2	-21.4

(t = 1) months. As a first approximation, the values for the 15th of each month were set equal to the 1941-70 climatological normals for the entire month (published in Monthly Record of Meteorological Observations in Canada). Subtracting c/12 from the constant coefficient a brings the integrated mean value of the quadratic into line with the climatological average. The linear coefficients b were then adjusted slightly (in all cases by less than 1°C) for the mean temperature computed for the end of one month to equal that for the beginning of the next month.

The "core-less" winter is evident in the resulting tabulation of daily temperatures, with the computed mean remaining between -23°C and -25°C for 90 days, from December 24 through March 23. The spring warming (from the derivative of the quadratic) reaches its maximum rate of 0.4°C/day in early May; the autumn cooling is less rapid, peaking at 0.3°C/day in early November. The mean temperature is above freezing for 79 days in summer, with peak warmth on July 27.

Table 3.2 lists the mean, standard deviation, and standard error of the summer temperature anomalies for each type, along with the number of occurrences during the twelve summers. A correction of +0.2°C was applied to the mean anomalies for all types before inclusion in Table 3.2, since the twelve summers analyzed averaged that much below the 1941-70 climatological mean. The root mean square value of the mean temperature anomalies listed in Table 3.2, weighted by the relative frequency of each synoptic type, is 1.4°C. The standard deviation of daily temperatures within a summer season (seasonal trend and interannual variability removed) is

TABLE 3.2. Broughton Island Temperature Characteristics of Synoptic Types, June-August 1959-70

Type	Mean Temperature Anomaly	Standard Deviation	Standard Error	Number of Occurrences 1959-70
3	.5°C	3.4°C	.4°C	80
9	.0	3.3	.6	32
15	-1.1	3.4	.6	29
20	-.6	3.0	.4	49
1	-1.6	3.0	.3	127
2	-1.3	2.9	.3	110
5	-1.3	2.4	.3	58
6	-1.4	2.5	.4	38
10	-.5	3.0	.4	64
26	-2.7	2.7	.9	8
8	-1.0	2.7	.7	16
13	.5	3.0	.9	12
14	.3	2.7	.6	20
18	.6	2.4	.7	14
21	1.4	2.5	.9	7
4	.7	3.0	.3	90
7	.5	3.4	.5	43
11	1.7	2.8	.4	50
16	3.6	3.2	.6	24
22	2.7	3.3	.9	12
23	-.4	3.1	.5	47
27	1.9	3.6	.9	18
12	.5	3.5	.6	30
17	1.4	2.6	.6	22
19	.7	3.2	.8	17
24	.4	2.3	.6	17
25	2.1	2.8	.8	13
28	3.5	4.5	.9	27
No Type	.8	2.8	.5	30

3.1°C; thus, the 28 synoptic types account for $1.4^2/3.1^2$, or 19 percent, of the daily variance of Broughton Island temperatures within a summer season. Therefore, there is a large within-type temperature variance, possibly due to variable cloud cover and radiative effects, changes in the local advective processes during the summer resulting from seasonal changes of snow and ice cover, and air mass differences such as those between cold-core lows and frontal wave cyclones. However, the between-type temperature differences are statistically and (as is shown in this section) physically significant, and are large enough to justify examining their role in the variability of average summer temperature.

Although this study deals specifically with summer temperatures, a comparison of the summer and winter thermal characteristics of each type gives some insight into the physical processes responsible for these characteristics. The comparison is presented in Fig. 3.3. The standard error for a given type is typically 0.5°C in summer, and 0.5 to 2.0°C in winter, depending on the frequency of the type. Types 1, 6, 10, 2, 7, 4, 11, 22, 16, 3, 9, and 15 (see maps, Fig. 3.1) are similar in that each type is dominated by a single cyclonic center near one edge of the grid, and form a sequence with the cyclone progressively displaced clockwise with respect to Baffin Island. The seasonal nature of the source regions of air flow to Broughton Island is readily apparent. In winter, the warmest synoptic types (7, 4, 11, 22) are those that bring southerly flow from the open waters of the Labrador Sea, while the coldest types of the sequence (1, 3, 6, 9, 15) subject Broughton Island to flow from the northeast or

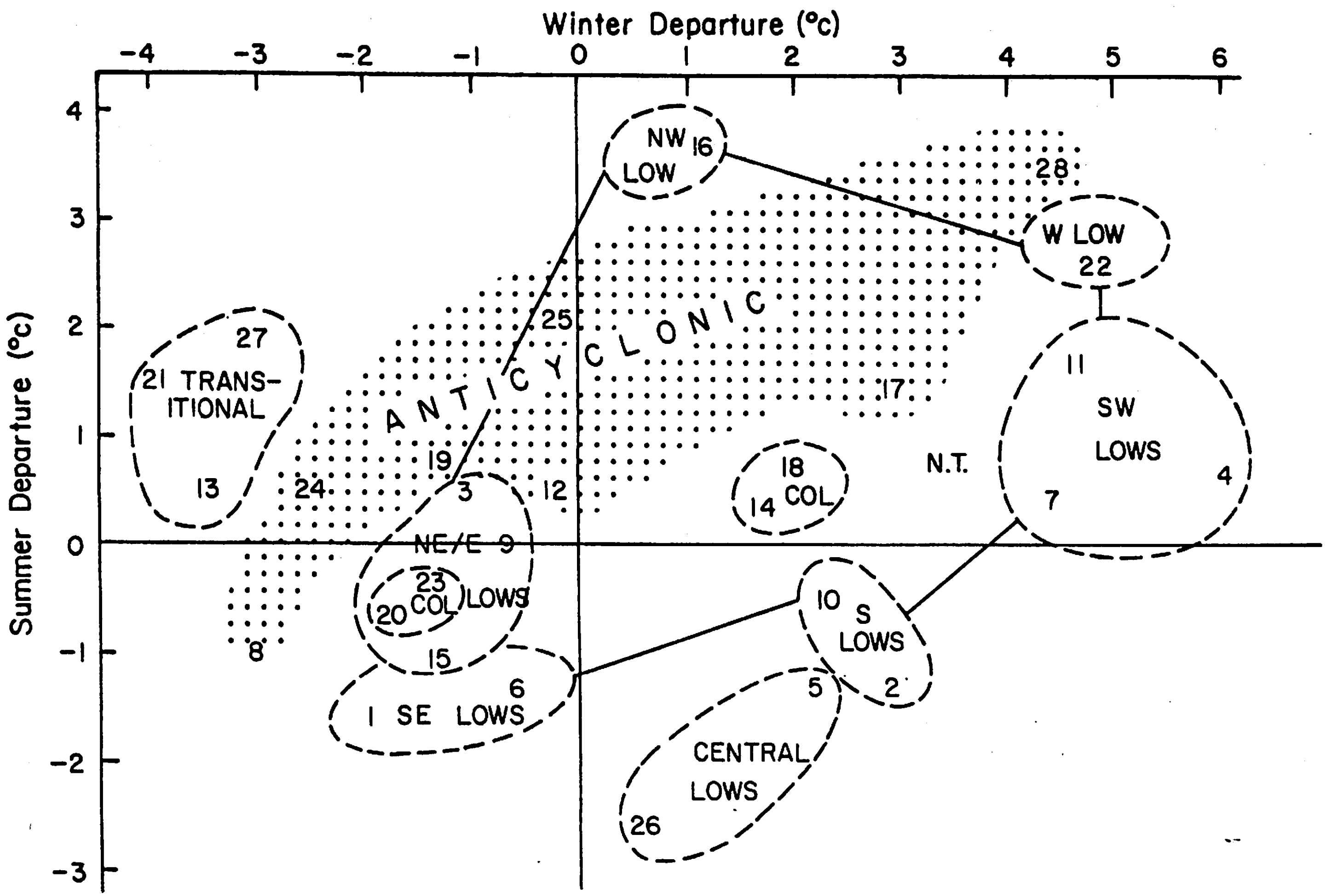


FIG. 3.3. Comparison of summer and winter Broughton Island temperature characteristics of the Baffin synoptic pressure pattern types, 1959-70.

northwest. The warmest of these types in summer (16, 22) are those that bring flow from the land masses to the south and southwest; the coldest (1, 2, 6, 10, 15) bring cool air from the relatively cold waters (and ice) of Baffin Bay.

Two other types -- 5 and 26 -- are also dominated by a single low, centrally located over Baffin Island. The associated flow at Broughton Island is generally from the waters to the south and east, rendering these types warm in winter, cool in summer. The associated cloud cover and precipitation from a nearby low center would tend to suppress summer temperatures further at Broughton.

Several types are characterized by two lows on the grid. For some of these, the temperature anomaly is close to the average of the anomalies for the two types with similarly located single lows. For example, type 14 has two lows in the positions occupied by single lows in types 6 and 22, and is located midway between the two types in Fig. 3.3. Likewise, type 18 acts like a combination of types 10 and 16.

In other cases one of the two lows appears to dominate the Broughton Island temperature anomaly. Types 20 and 23 both have lows to the northeast and south of Baffin, but from their position in Fig. 3.3, it is clear that the northeast low controls the temperature characteristics. The lows are to the east and west of Baffin Island in types 8, 13, 21, and 27; in all four cases, the eastern low (similar to a type 1 cyclone) appears to dominate. The winter temperatures tend to be colder, and the summer days slightly warmer, with these four types than with a single eastern low. A count of the day-to-day transitions between types over the entire 29 years shows that types 8, 13, 21, and 27 are preceded by eastern lows (types 1, 3, 6, 9, 15) 59 percent of the time. Low winter temperatures, then, are probably due to cold air pushing southward behind the eastbound low, while in summer the clearing behind the storm allows radiational heating.

Radiational effects are quite pronounced with the six anticyclonic types (12, 17, 19, 24, 25, 28): all are warm types in summer. The spread of temperatures in the winter appears to be controlled by the history of the anticyclone. The warmest anticyclone type, 28, is most frequently preceded by type 3, implying an anticyclone moving up from the south. The cold type 24 anticyclone usually follows types 6, 7, and 10, and therefore moves south from the polar regions.

One inconsistency that remains difficult to explain is the large difference (5.5^{o}C) in winter thermal anomalies between types 14 and 21, since they are morphologically very similar, having lows to the southeast and northwest of Baffin Island. Aside from this one case, the temperature anomalies associated with each type agree qualitatively with what would be expected from the influences of radiation and synoptic-scale advection.

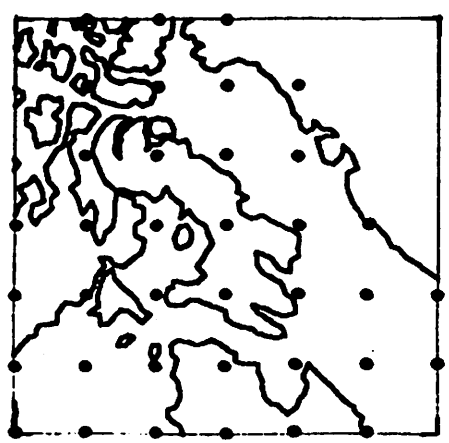
It is possible that the temperature characteristics of some of the types may be modified by local circulation mechanisms. The föhn effect -- due to subsidence of westerly flow from the highlands immediately west of Broughton Island -- may elevate the mean temperature anomalies of the Baffin Bay cyclonic types 3 and 9 (LeDrew, 1976). Baffin Bay cyclones may therefore be colder types over areas not subject to the föhn effect than the Broughton temperature anomalies would indicate. Possible sea breezes have been noted in the Arctic (Moritz, 1977); their cooling effect at coastal stations should be most noticeable on days with sunshine and light synoptic flow, i.e., anticyclonic days. However, the anticyclonic types are the warmest types at Broughton, and it would appear that sea breezes play a minor role in moderating the temperatures at Broughton Island.

Type Groupings

Because the large number of type classifications is unwieldy for many uses, it was decided to group the 28 synoptic types into a smaller number of basic types. The main criterion for grouping is the presence (or absence) and location of cyclonic centers. In most cases emphasis is given to a strong and distinctive circulation feature that appears on the grid, even if its influence does not cover all, or even most, of the grid area.

One obvious grouping is of the relatively infrequent anticyclonic types, in which cyclonic influences are limited to the very edges of the grid. The grouping of cyclonic types is not so simple, but the three basic groups chosen are those with cyclones to the northeast (NE), southeast (SE), and west (W) of Baffin Island. Three more groups are characterized by the concurrence of two cyclonic centers. None of the 28 original types have three cyclones on the grid.

The choice of the three basic cyclonic groups, and the assignment of the original types to these groups, is not entirely arbitrary. Fig. 3.4a shows the percentage distribution of lowest



Grid point map

	3	1	1			
	1	1	3	2		
	1	1	2	2		
2	1	1	1	1	4	
1	2	2	2	2	2	15
2	1	3	2	2	1	11
5	3	3	4	3	6	

(a) 29-year average

	8	3	3			
	2	4	7	5		
	0	4	2	3		
3	0	-1	-1	2	-2	
-5	-7	-1	-3	-1	-2	1
-2	-6	-7	-3	-2	0	-7
8	-1	6	2	-3	-11	

(b) Change, 1951-60 to 1964-73

FIG. 3.4. Percentage distribution of occurrence of the lowest pressure on the Baffin synoptic type grid at each grid point, computed for all days, 1946-74 (center); decadal change in percentage distribution, summer 1951-60 to 1964-73 (right).

pressures across the grid for all 10460 days from 1946 to 1974. Some concentration can be seen towards the northeast, southeast, and southwestern corners of the grid, forming a natural division of cyclonic activity into three sectors. In individual cases, some guidance is provided by the thermal character of the type, as indicated by its position in Fig. 3.3. For example, types 5 and 26, with their lowest pressures located near the center of the grid, could have been assigned to any of the three basic cyclonic groups. They are classed as SE lows on the basis of their thermal influence.

The seven groups, their constituent types, their frequency of occurrence in summer, and their average summer temperature anomaly (with standard error) are given in Table 3.3.

TABLE 3.3. The Seven Synoptic Type Groups

	Member Types	% Summer Frequency	Temp. Anomaly
NE lows	3, 9, 15	13.8	+0.1±0.3
SE + NE lows	20	3.8	-0.6±0.4
SE lows	1, 2, 5, 6, 10, 26	37.6	-1.3±0.2
W + SE lows	8, 13, 14, 18, 21	5.7	+0.2±0.4
W lows	4, 7, 11, 16, 22	19.9	+1.3±0.2
NE + W lows	23, 27	5.7	+0.2±0.4
Anticyclonic	12, 17, 19, 24, 25, 28	11.0	+1.5±0.3
Untyped		2.5	+0.8±0.5

These seven groups can be reduced further, by assigning those types with two cyclones to the basic cyclonic types. For example, one such group would include all days with a low SE of Baffin, whether or not another cyclone was present elsewhere on the grid. A summary of the frequency and average summer temperature anomaly for these groups is given in Table 3.4.

TABLE 3.4. The Four Basic Synoptic Type Groups

	Average Number of Days Per Summer	Average Percent of Days in Summer	Temp. Anomaly
All NE lows	21.4	23.3	0.0±0.2
All SE lows	43.3	47.1	-1.1±0.1
All W lows	28.8	31.3	+0.9±0.2
Anticyclonic	10.1	11.0	+1.5±0.3

During a typical (29-summer average) 92-day June-July-August period, one cyclone will dominate the grid on 66 days, two on 14 days, and none on 10 days, with 2 days untyped. Although the 28 original types are reduced to four groups, the primary features of the types -- their circulation features and temperature characteristics -- have been retained. This final reduced grouping is the one mainly used for the rest of this study.

Interannual Variability of Synoptic Type Frequencies

The summer frequencies of the four basic synoptic type groups from 1946 to 1974 are plotted in Fig. 3.5. The frequencies exhibit a large variability, with the cyclonic groups varying by a factor of 2 or 3 between extreme summers, and the anticyclonic (AC) group by a factor of 10. Trends are revealed by the 1-2-1 smoothed data (heavy line); among the notable features are the decline of SE cyclones and the dramatic increase of anticyclonic days after 1968, and a distinct out-of-phase relationship between the number of days with NE and W cyclones which remain, respectively, at above, and below, average values since the early 1960's. The correlations between the summer frequencies of the four groups, on a yearly and smoothed basis, are listed in Table 3.5.

TABLE 3.5. Correlation Between Summer Frequencies of Grouped Baffin Synoptic Types, 1946-74

(Yearly values to upper right, smoothed values to lower left.)

Type	NE	SE	W	AC
NE		-.18	-.66**	.12
SE	-.35		-.09	-.58**
W	-.76**	.28		-.36*
AC	.31	-.80**	-.48*	

*90% confidence of significance; **99% confidence.

On the yearly basis, there are near zero correlations between groups AC and NE and between SE and W, increasing to slightly positive values for the 2-year smoothing. On both time scales, the AC and NE groups are strongly anticorrelated with the SE and W types.

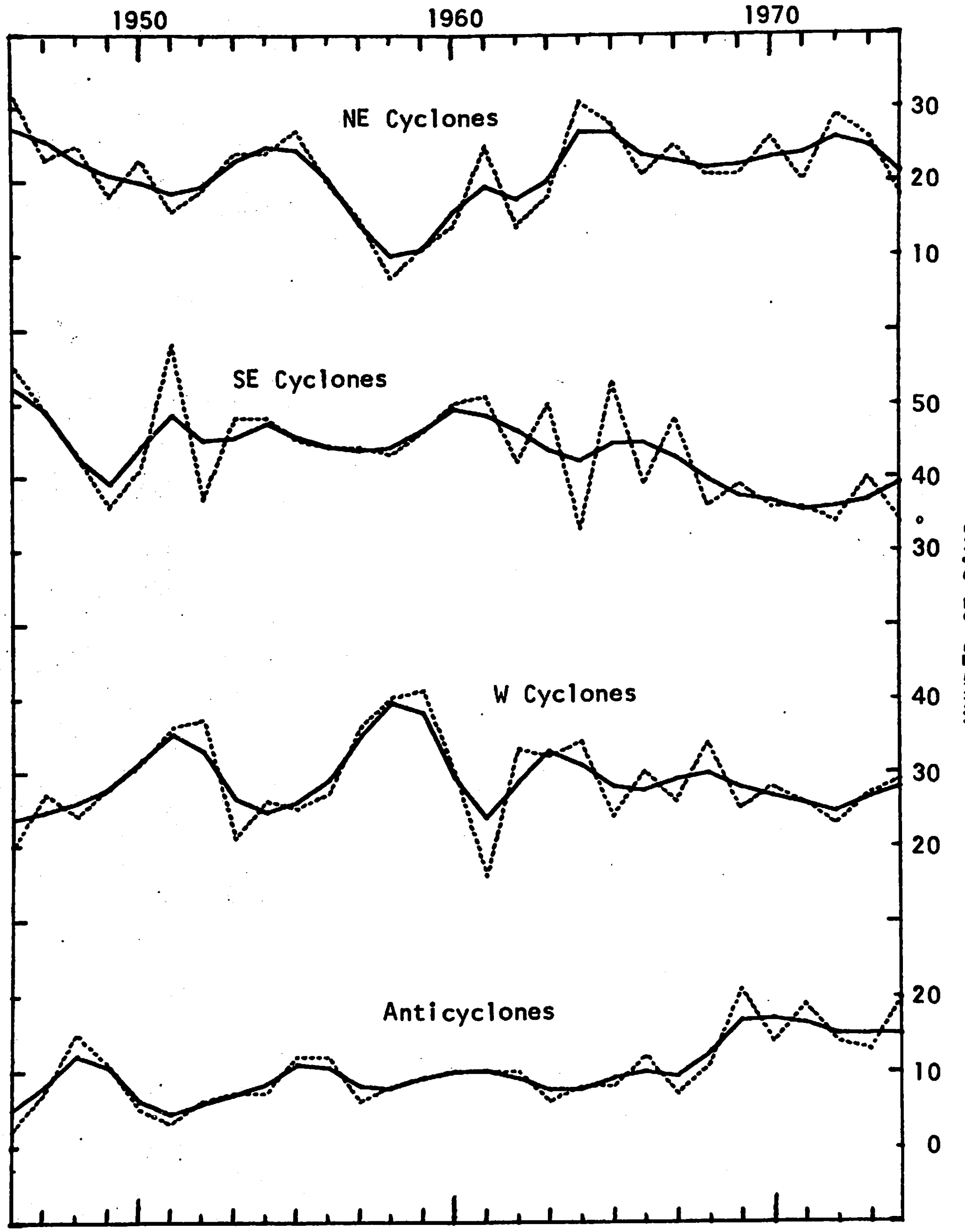


FIG. 3.5. Number of days per summer with each of the four basic synoptic types, 1946-74. Yearly values, dotted lines; smoothed values, solid lines.

Changes in frequencies of the 28 synoptic types on a longer time scale, between the climatically extreme decades of 1951-60 and 1964-73, are listed in Table 3.6. Eight types gained ten or more days in frequency between the decades. Of these, four (3, 9, 15, 23) are NE cyclonic types, and four (12, 17, 25, 28) are anti-cyclonic. Of the five types that lost ten or more, two (4 and 7) are W cyclonic, and three (1, 2, 10) are SE cyclonic. The decadal differences for the grouped types are listed in Table 3.7.

Table 3.7 confirms that the NE and AC types gained in frequency between the two decades at the expense of the SE and W types. The values for all groups differs from the number of days in ten Junes, Julys, and Augusts because days with two cyclones are credited to two different groups; the change in this frequency indicates a very slight increase in the occurrence of days with two cyclones. The correlations given in Table 3.5 can be seen to extend to the decadal time scale: the near-zero correlations of yearly frequencies of AC and NE types, and of SE and W types, becomes more strongly positive as the time scale increases, while the anticorrelation between AC and NE types and SE and W types remains strong on the one, two, and ten-year bases.

Some authors (Dzerdzeevski and Sergin, 1972; Lamb, 1977) have suggested an increase in meridional circulation and blocking activity over the past 30 years. It is therefore of interest to determine whether the decadal changes in frequency (days of occurrence) of grouped synoptic types are due to changes in the number of systems affecting Baffin, or to changes in their average duration. The number of spells of consecutive days with a

TABLE 3.6. Decadal Change in Summer Frequencies of Synoptic Types, in days, 1951-60 to 1964-73

Type	Group	1951-60	1964-73	Change
3	NE	63	82	+19*
9	NE	20	37	+17**
15	NE	14	25	+11*
20	SE+NE	34	37	+3
1	SE	116	104	-12
2	SE	120	86	-34*
5	SE	42	41	-1
6	SE	26	31	+5
10	SE	57	43	-14
26	SE	10	6	-4
8	W+SE	14	13	-1
13	W+SE	6	8	+2
14	W+SE	15	8	-7
18	W+SE	18	13	-5
21	W+SE	5	4	-1
4	W	103	64	-39**
7	W	48	37	-11
11	W	37	38	+1
16	W	18	17	-1
22	W	9	8	-1
23	NE+W	28	50	+22**
27	NE+W	18	17	-1
12	AC	20	34	+14**
17	AC	15	28	+13**
19	AC	9	16	+7*
24	AC	19	12	-7
25	AC	3	13	+10**
28	AC	14	24	+10*
No Type		19	24	+5

*90% confidence; **95% confidence.

TABLE 3.7. Decadal Change in Summer Frequencies of Grouped Synoptic Types, 1951-60 to 1964-73

Group	1951-60	1964-73	Change	Change (Percent of 1946-74 Average)
NE	177	248	+71**	+33
SE	463	394	-69**	-16
W	319	277	-42*	-15
AC	80	127	+47**	+47
All Groups	1039	1046	+7	+1

*95% confidence; **99% confidence.

particular grouped type, and the mean duration of these spells, for the two decades are listed in Tables 3.8 and 3.9. Tables 3.7, 3.8, and 3.9 are summarized in Table 3.10.

The decrease in days with SE and W cyclonic types between the two decades appears to be due to the same number of systems spending, on the average, less time in the grid area, while the increase in NE cyclone days results from more systems of unchanged duration. There were more anticyclones in the latter decade, each with a tendency to stay longer. The cyclonic types all tended to move (or evolve) more rapidly during the colder 1964-73 decade, and on the whole, weather systems changed more frequently. There appears to have been an increase in the number of cyclones moving into, or forming within, the Baffin grid area between the two decades, despite the net decrease in the number of summer days with cyclonic types.

Most of the additional cyclones during the latter decade were over Baffin Bay, which raises the question: what is their origin?

The key day pressure pattern on the day preceding a NE cyclone gives an indication of the direction the cyclone came from. A count of the transitions between synoptic types over 29 summers reveals that of the 12.6 NE cyclone events during a typical summer, 7.6 of the storms moved in from the south and southeast, 3.0 from the southwest and west, and 2.0 from the north and northwest -- the latter representing cyclones of Arctic Ocean origin, such as a system analyzed by LeDrew (1976). Between the decades 1951-60 and 1964-73, the frequency of these storms from the northwest rose from 1.4 per summer to 2.2, an increase of 0.8. However, the number of

TABLE 3.8. Decadal Change in Number of Continuous Spells (1 Day or Longer) of Grouped Synoptic Types, 1951-60 to 1964-73

Group	1951-60	1964-73	Change	Change, Percent of 1946-74 Average
NE	105	149	+44**	+35
SE	159	157	-2	-1
W	130	134	+4	+3
AC	58	70	+12	+19
All Groups	452	510	+58*	+12

*90% confidence; 99% confidence.

TABLE 3.9. Decadal Change in Average Duration of Continuous Spells of Grouped Synoptic Types, in Days, 1951-60 to 1964-73

Group	1951-60	1964-73	Change	Change, Percent of 1946-74 Average
NE	1.7	1.7	0.0	-1
SE	3.0	2.6	-.4*	-15
W	2.5	2.1	-.4**	-17
AC	1.5	2.0	+.5	+24
All Groups	2.3	2.1	-.2**	-10

*90% confidence, **95% confidence.

TABLE 3.10. Decadal Change, in Percent of 1946-74 Average, of Number of Spells, Duration of Spells, and Number of Days of Occurrence of Grouped Synoptic Types, 1951-60 to 1964-73

Group	Number of Spells	Average Duration	Number of Days
NE	+35*	-1	+33*
SE	-1	-15	-16*
W	+3	-17*	-15*
AC	+19	+24	+47*
All Groups	+12*	-10*	+1

*95% confidence.

all Baffin Bay cyclone events rose by 4.4 per summer, and most of these additional storms must have entered Baffin Bay from the west, south, or southeast. This is not necessarily inconsistent with the lack of increase in number of SE and W cyclones; it could indicate that more of these southern cyclones turned north into Baffin Bay during 1964-73, rather than heading eastwards and leaving the grid.

With regard to the increase in anticyclone longevity, it should be noted that most of the gain is due to one anomalous summer, 1969. That summer saw two anticyclonic spells last 9 days apiece, contributing to a seasonal total of four spells averaging 5.3 days long. Removing these values from the figuring yields a 1964-73 decadal average anticyclone duration of 1.6 days, just 0.1 day over the 1951-60 value. It would therefore appear that the increase in anticyclone days is primarily due to more of these systems moving into the area, rather than from more stagnating or blocking anticyclones.

Since the frequencies of anticyclones and of NE (Baffin Bay) cyclones are positively correlated, one would suspect that a common weather sequence is for an anticyclone to follow a Baffin Bay cyclone. However, during the period 1946-68 (a period of low to average anticyclone frequency), this sequence happened an average of only 1.5 times per summer, while anticyclones followed SE and W cyclones an average of 4.1 times. During 1969-74, however, the average summer frequency of Baffin Bay cyclone - anticyclone sequences jumped to 3.3, while the SE and W cyclones preceded anticyclones 4.6 times. This latter period was one of frequent anticyclones, and contributed to most of the increase noted earlier for the 1964-73 decade. It appears that the additional anti-cyclones were mostly those following Baffin Bay cyclones.

The decadal change in cyclone day frequencies across the Baffin grid is summarized in Fig. 3.4b. For each difference, the average number of days with lows at each grid point was reconstructed from the distribution of lows for each of the 28 types (Fig. 3.2) and the decadal frequency of those types. Fig. 3.4b displays graphically the shifts in cyclone activity discussed earlier. It also shows that while days with southern and western cyclones generally decreased, the frequencies of those at the extreme southern and southwestern grid points increased.

The net effect of the shifts in cyclone activity on the seasonally averaged sea-level pressure pattern is shown in Fig. 3.6*. The pattern is consistent with the regional relocation of cyclone centers, with rising pressures across most of the southern half of the region, and falling pressures to the north and extreme southwest. The time intervals (1950-59 and 1965-69) represented in Fig. 3.6 do not coincide exactly with those analyzed for Fig. 3.4b, but it can be seen from Fig. 3.5 that the cyclone redistributions for the two sets of time periods should be similar. The pattern of pressure change is very similar to that of the 1942-72 pressure trend mapped by van Loon and Williams (1976).

The changes in seasonal vectorially averaged surface geostrophic wind direction and speed over the Cumberland Peninsula from 1946 through 1974 are displayed in Fig. 3.7. The June-July-

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* From data provided by R. Jenne and W. Spangler, NCAR.

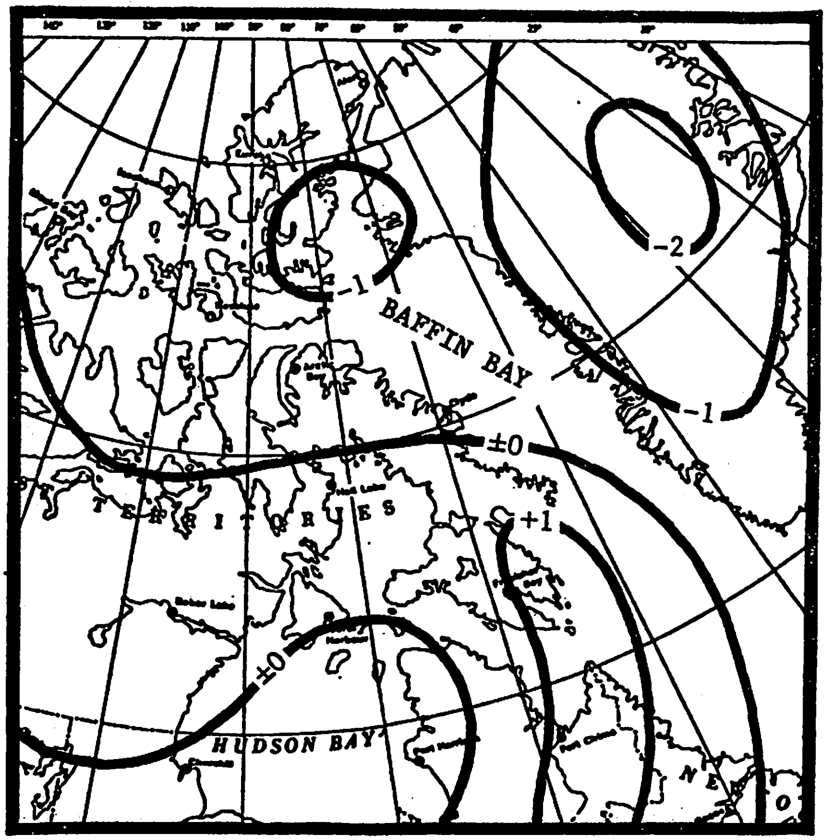


FIG. 3.6. Change of average summer sea-level pressure around Baffin Island, 1950-59 to 1965-69, in millibars.

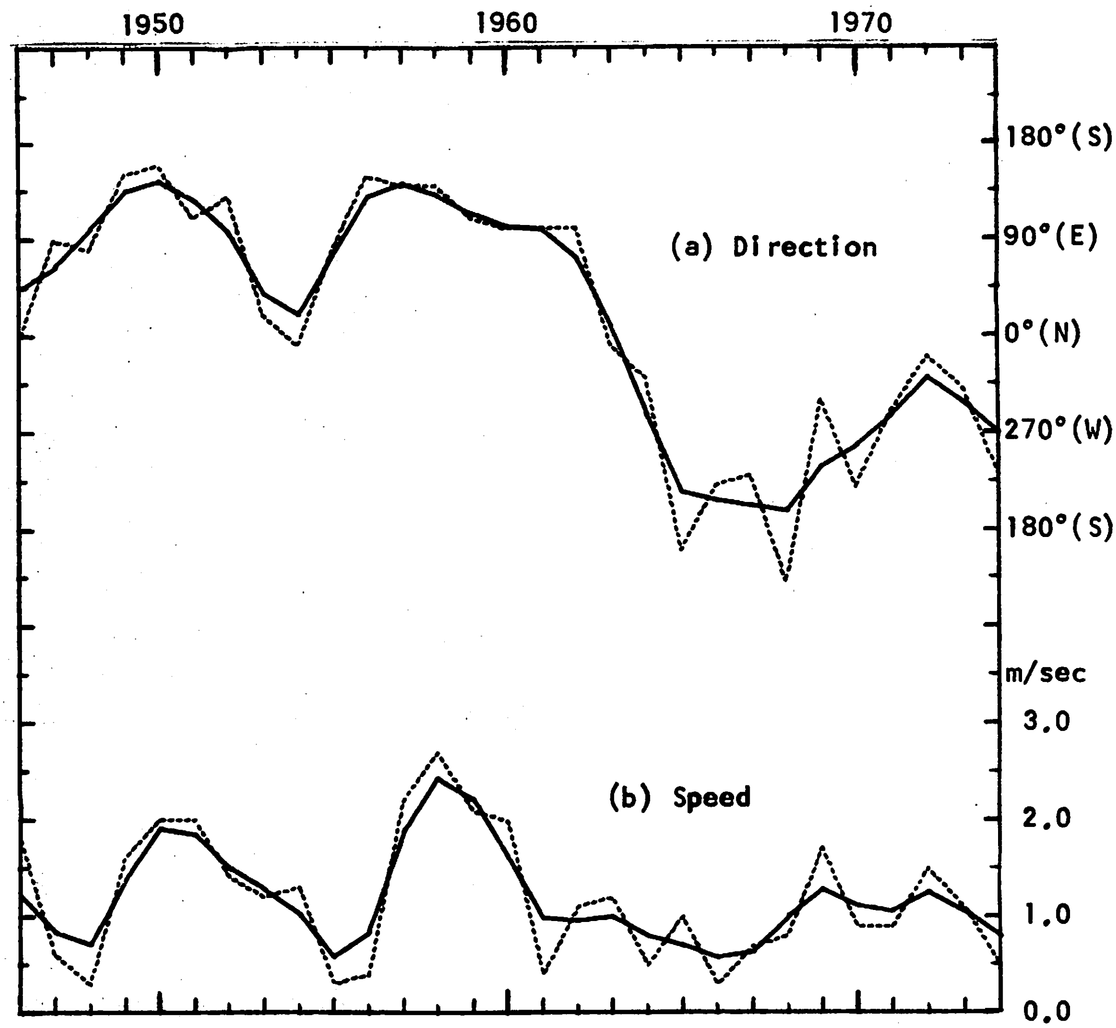


FIG. 3.7. Resultant geostrophic sea-level winds near the Cumberland Peninsula, southeast Baffin Island, summer 1946-74. Annual values, dotted lines; smoothed values, solid lines.

August mean winds were computed from four grid points covering a 390 km square area centered near Broughton Island. A dramatic wind shift -- from predominately easterly to predominately westerly -- is apparent around 1962-65. The vectorially averaged summer winds for 1951-60 were from 110^{o} east of north at 1.2 m/sec; for 1964-73 they had reversed to 290^{o} at 0.6 m/sec (standard errors of ±25^{o} and ±0.3 m/sec). Comparison with Fig. 2.1 shows that the warmest summers (1949-50, 1957-62) are those with prevailing easterly surface geostrophic winds; the coldest summers (1963-64, 1970-73) occur when the prevailing wind is most nearly westerly. This is in direct contradiction to the relation between temperature and wind direction found for the individual synoptic types. Apparently the advective processes of the synoptic types, which influence the day-to-day temperature changes, are overwhelmed on the seasonal average by the advection associated with the prevailing winds. This is not surprising, considering that over the course of a season, a 1 m/sec prevailing wind can transport air from far more distant -- and different -- source regions than can a 10 m/sec synoptic scale wind acting over a typical synoptic type lifetime of two days. The effects of large-scale advection will be covered in more detail in the next chapter.

With the influences of advection on temperature being so different on the daily and seasonal bases, it should be instructive to look at the correlation between synoptic type frequencies and seasonal temperatures. Table 3.11 lists these correlations for yearly and 1-2-1 smoothed Baffin Bay summer temperatures for 1949-74, as well as the agreement (+) or disagreement (-) of sign for the decadal (1951-60 to 1964-73) changes in average temperature and type frequencies. The mean Broughton Island temperature anomaly is also noted for each type. The sign of the correlation for each type is the same on all three time scales, but in most cases bears little resemblance to the daily temperature characteristics of the type. Only the W cyclones have a consistent relationship with temperature on time scales from a few days to ten years.

TABLE 3.11. Correlations Between Summer Temperature and Frequencies of Grouped Synoptic Types on Different Time Scales

Type Group	Yearly	Yearly (1-2-1 Smoothed)	Decadal	Broughton Island Mean Daily Temperature Anomaly
NE	-.54**	-.64*	-	0.0°
SE	.31	.65*	+	-1.1°
W	.20	.33	+	$+0.9^{\circ}$
AC	-.31	-.60*	-	$+1.5^{\circ}$

*95% confidence; **99% confidence.

As noted earlier, the occurrence of Baffin Bay (NE) cyclones followed by anticyclones was most frequent during 1969-74, coincident with the coldest summers of the period of record. Summer anticyclones tend to be warm and Baffin Bay cyclones neutral, but perhaps the combination -- a Baffin Bay low pulling a polar high southwards from the ice-covered Arctic Ocean -- is a cold situation. It is also possible that during cold summers, more of the Baffin Bay cyclones are cold-core cyclones rather than 'warm' frontal wave systems. However, the synoptic typing scheme used in this study takes no account of the thermodynamic differences between cold lows and frontal waves or of the intensity of the systems, and therefore cannot distinguish between structurally different cyclones. A possible refinement of the typing scheme would use cyclone movement rates (slower for deep cold lows than for frontal waves) or a two-level (surface and 500mb) typing scheme which, by noting the degree of similarity between the surface and upper patterns, would distinguish between cold-core and frontal cyclones, as well as between migrating and blocking anticyclones.

As a final test of the ability of synoptic type frequencies to explain some of the interannual variance of summer temperatures, the frequencies and Broughton Island temperature characteristics of each of the 28 types were combined to reconstruct the average summer temperatures for Broughton. The results are shown in Fig. 3.8, along with the observed temperatures. The two quantities have little in common, the correlation being but .15. On a decadal basis, the synoptic types "predict" a summer warming of 0.1°C between 1951-60 and 1964-73. The Broughton station did not operate before 1959, but as noted in Chapter Two, the entire region experienced a sharp cooling, as exemplified by Clyde (-0.8°C), Frobisher Bay (-0.7°C) and Egedesminde, Greenland (-1.0°C). As a predictor, the four grouped types did even worse, "predicting" a warming of 1.2°C between the two decades.

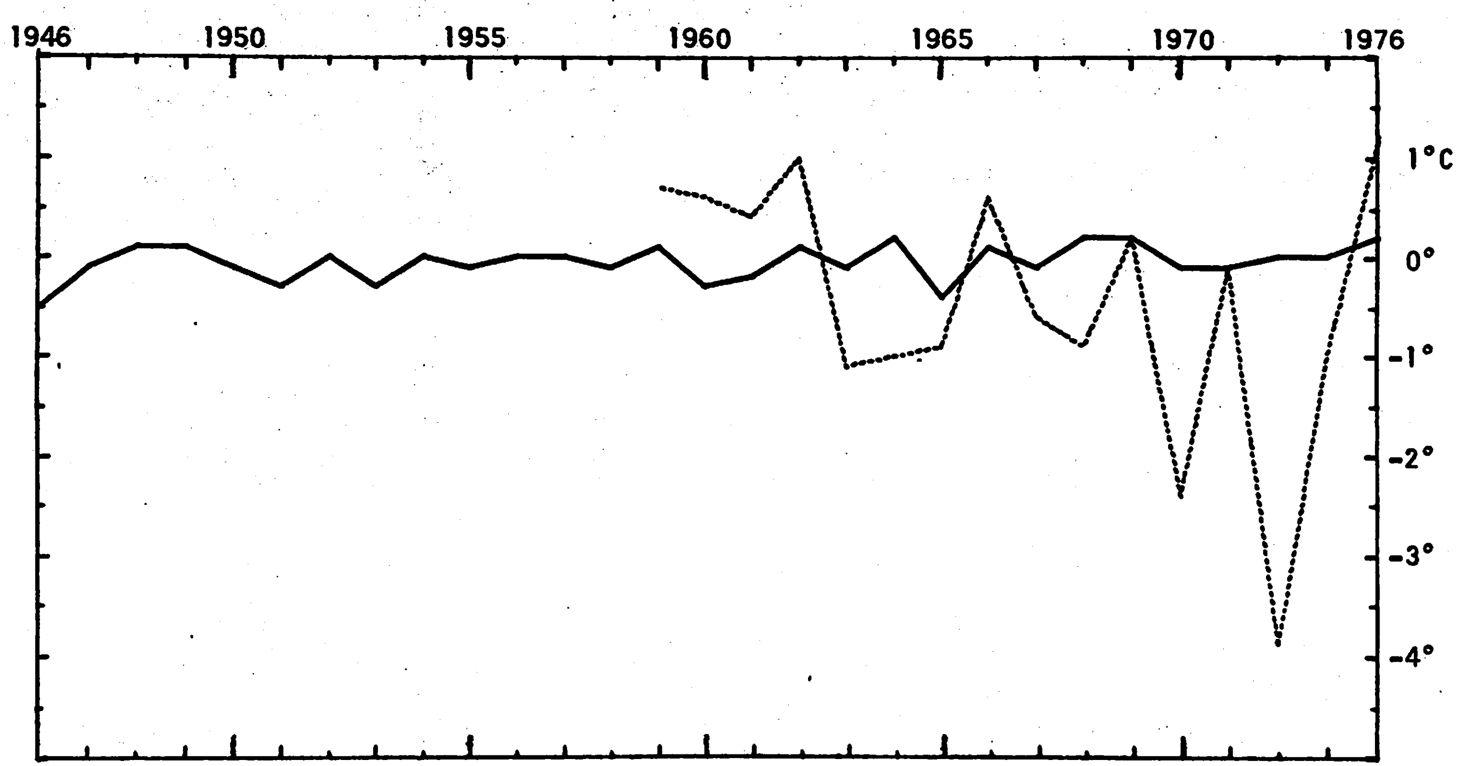


FIG. 3.8. Broughton Island summer temperatures, reconstructed from synoptic type frequencies (solid line), compared with observed temperatures (dotted line).

It is clear from Table 3.11 that the synoptic type frequencies show significant correlations with the interannual temperature variations. However, the correlations are in the wrong sense for the frequency variations to be invoked as the causal factor, and it appears that both seasonal temperature and type frequency are responding to a more pervasive influence. The nature of this influence will be the subject of the next chapter.

CHAPTER FOUR

REGIONAL CIRCULATION INFLUENCES

Fluctuations of Baffin summer temperatures and of the associated surface cyclone activity were the subjects of the previous two chapters. In the last chapter it was suggested that the interannual temperature fluctuations are largely due to the large scale advective processes associated with the long standing waves of the global circulation, as found by Brinkmann and Barry (1972) for summer surface temperature anomalies in Keewatin and Labrador-Ungava. Since the existence of a trough on a mean upper air chart is largely due to the net effect of cyclones on the normal west to east flow (Namias, 1958), a close connection between fluctuations of surface cyclones and upper troughs would also be expected.

The observed link between the location and strength of the quasi-permanent Baffin trough, regional surface temperature, and regional cyclone activity is the topic of the first section of this chapter. The second section examines changes of surface cyclone tracks over North America and the western Atlantic that lead to changes in Baffin area cyclone activity. In the third section the concept of 500mb vorticity flux as the upper level reflection of surface cyclone activity is introduced, and is applied to the question of storm track variations. The vorticity flux analysis is extended to the extratropical Northern Hemisphere in the fourth section.

Influences of the Baffin Trough

Surface temperatures are closely related to the average temperature of the lower troposphere, as measured by the thickness of the 1000-500mb layer -- this being the main parameter used in numerical weather prediction to forecast surface temperature. The flow pattern at the 700mb level, lying at the middle of the 1000-500mb layer, should be a good indicator of advection in the lower troposphere, and is used for this purpose in the analyses of monthly weather and circulation published in the Monthly Weather Review.

The normal (1953-76 average) summer upper air circulation pattern (Fig. 4.1) is characterized in the Arctic by two major troughs in the westerlies, one near the Bering Straits and a somewhat stronger one over Baffin Island, and one or two lesser troughs over the North Atlantic to central Siberian sector. Fig. 4.1 shows the 500mb height pattern, but the same features are evident at 700mb. The longitude of the troughs is subject to change from year to year; over Baffin Island the prevailing upper winds can change from southerly to northerly as the trough axis shifts from west to east of its average position over the Island. This effect is illustrated in Figs. 4.2 to 4.5.

Fig. 4.2 shows the average July 700mb height pattern for 1943-47, 1951-60, and 1964-73. The data are from Henry and Armstrong (1949), Titus (1967), and the Monthly Climate Data for the World. During 1951-60 (a warm decade at Baffin) the trough remained, on the average, west of Baffin Island, with resulting southerly flow. For the other two (colder) time periods, the

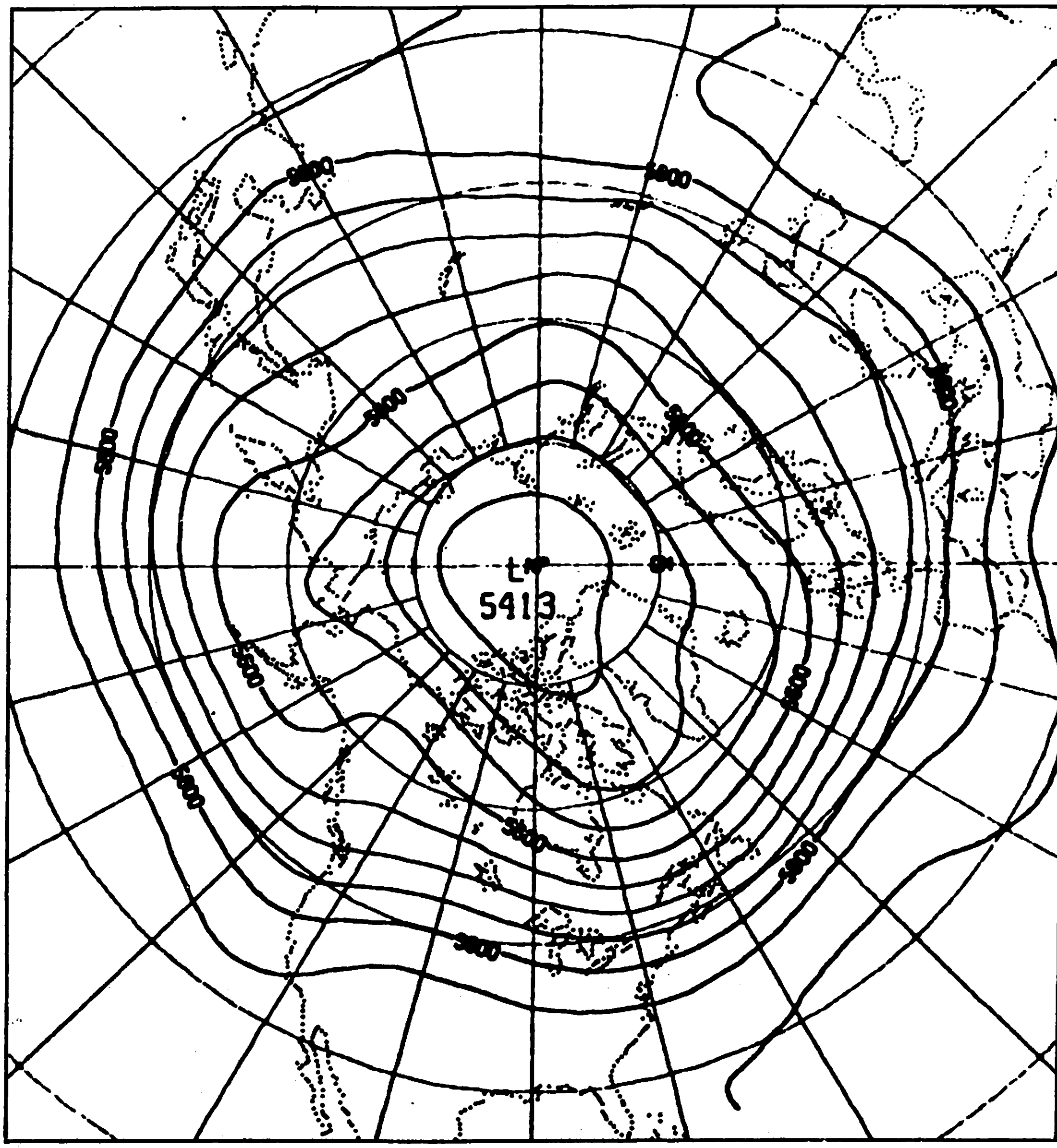
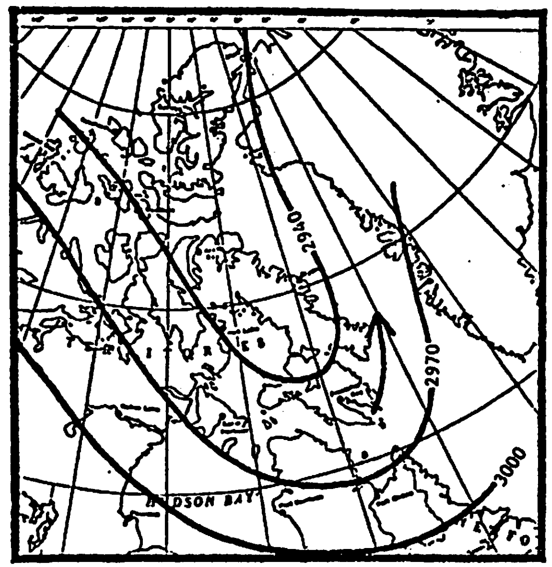
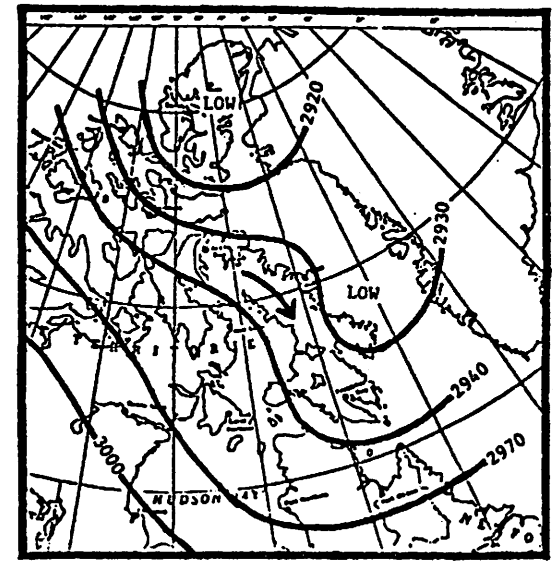


FIG. 4.1. Average summer 500mb heights, June-August, 1953-76. Contour interval, 50 meters.

(a) 1943-47



(b) 1951-60



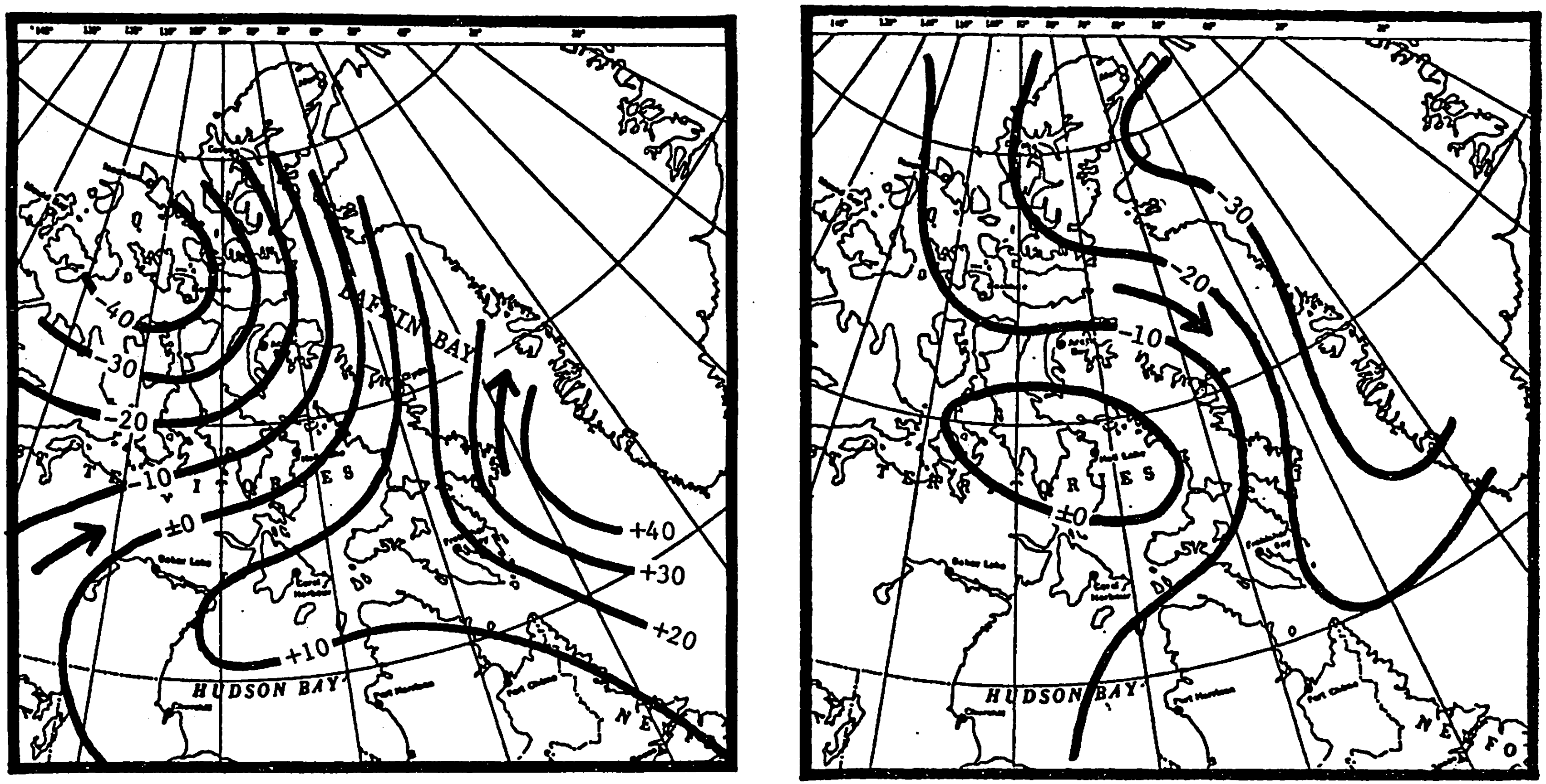
(c) 1964-73

FIG. 4.2. Maps of average July 700mb heights around Baffin Island (in meters), 1943-47 (left), 1951-60 (center), and 1964-73 (right).

average trough was stronger (lower 700mb heights) and shifted eastward over Baffin Bay, with northerly flow to Baffin Island. In all three cases the position of the trough and direction of flow is consistent with the temperature anomalies.

The effect is more clearly seen by comparing the differences in 700mb height (Fig. 4.3) with the maps of temperature changes (Fig. 2.3) between the same time periods. The standard deviation of individual July heights around a 10-year mean is typically 30 meters; the standard error is therefore about 10 meters. In both cases depicted in Fig. 4.3 height changes several times the standard error are noted, and in both cases the change in advection is qualitatively consistent with the change in temperature. The axis of greatest warming between 1943-47 and 1951-60 lies directly over Baffin Island, coincident with the greatest shift to southerly advection from the North American continent. Between 1951-60 and 1964-73, the change to more northerly flow is strongest in a corridor from the polar ice cap, over Ellsmere Island, then southeastward across Baffin Bay. This corridor coincides with the area of maximum cooling.

The upper circulation changes on shorter time scales are shown by the interannual wanderings of the Baffin 700mb low center (Fig. 4.4). The low is a remarkably persistent feature on the mean monthly 700mb charts published in the Monthly Weather Review; only three of the 90 summer months from 1949 to 1978 failed to show a closed cyclonic center in the trough near Baffin. Each year the June, July, and August latitudes and longitudes of the low were averaged; the 1-2-1 smoothed values are plotted in Fig. 4.4. The



(a) 1943-47 to 1951-60

(b) 1951-60 to 1964-73

FIG. 4.3. Change of average July 700mb heights around Baffin Island (in meters), 1943-47 to 1951-60 (left) and 1951-60 to 1964-73 (right).

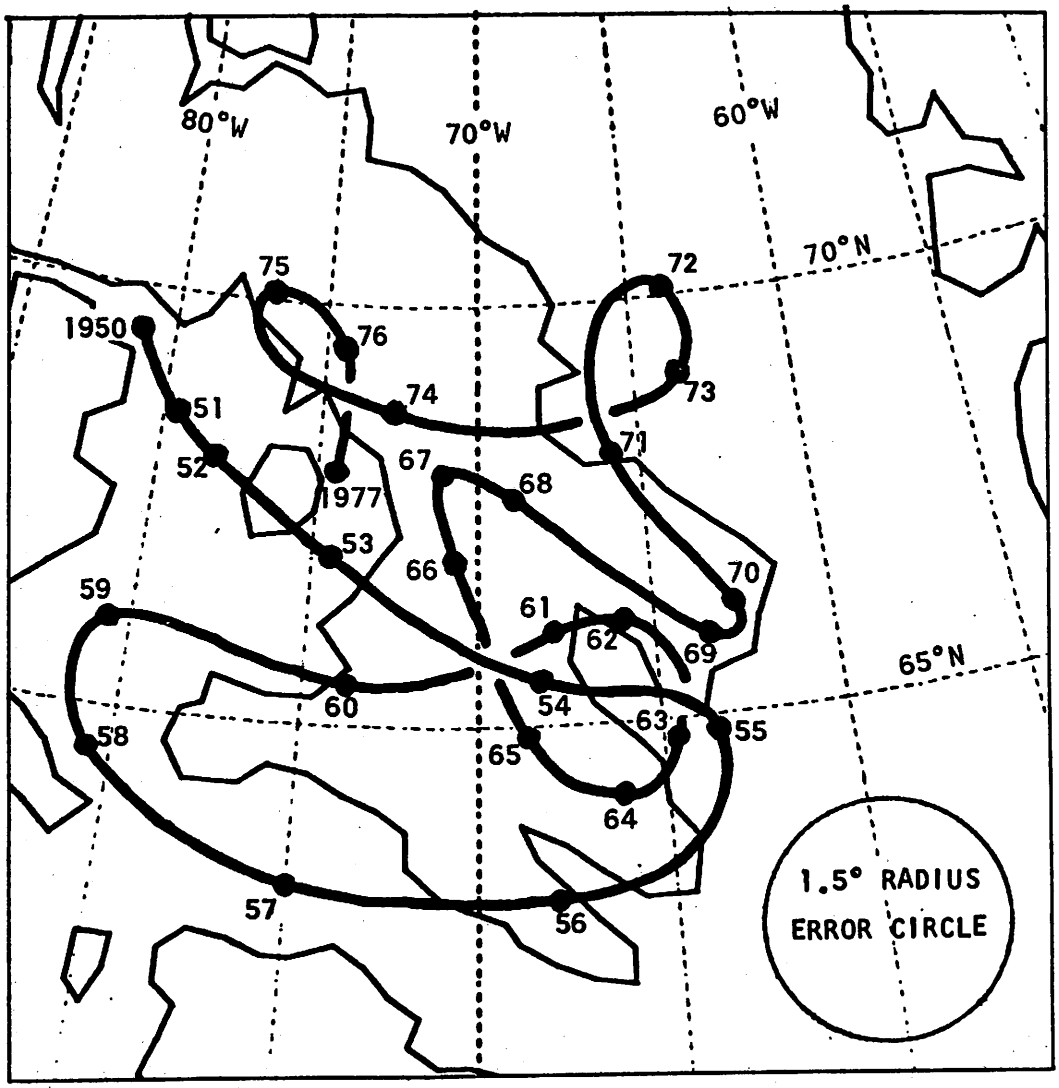


FIG. 4.4. Location of the 700mb Baffin low, summer 1950-77. The positions are averages of the June, July, and August latitudes and longitudes for each year, smoothed by a 1-2-1 running mean. The uncertainty of each position estimate is indicated by the circle at lower right.

subjectively estimated location of maximum curvature of the flow was used for the three months (July of 1961, 1964, and 1969) the low center was absent. The standard error of an individual smoothed seasonal position of the low, derived from the average standard deviation of the three monthly positions within a summer and the average interannual change of the seasonal mean position, using n = 6 months, is 1.6° in latitude and 3.3° in longitude. The corresponding error circle, shown in Fig. 4.4, has a radius of 1.5 latitude degrees.

The warmest summers over Baffin Bay (Fig. 2.1) occur when the 700mb low is well to the west or southwest of its average position, as around 1950, 1957-60, and 1975. Colder summers are associated with the low displaced to the east and northeast, as around 1955, 1964, 1970, and especially 1972. The association is somewhat easier to perceive when just the longitude is considered. A linear regression of the 30 sets of seasonally-averaged latitudes and longitudes gives the relation: West longitude = $32^{\circ} + 0.6 \cdot$latitude, reflecting the tilt of the long-term average trough. The longitudinal displacements from this mean trough axis for 1949-76 are shown in Fig. 4.5. The similarities with Fig. 2.1 are obvious, with the correlation between the longitude of the low and Baffin Bay temperatures being .57 for yearly values, and .79 for smoothed values. The greatest discrepancy from this correlation was during 1960-62, when the warm summers persisted despite an eastward movement of the 700mb low; these anomalous summers deserve closer study in a future investigation.

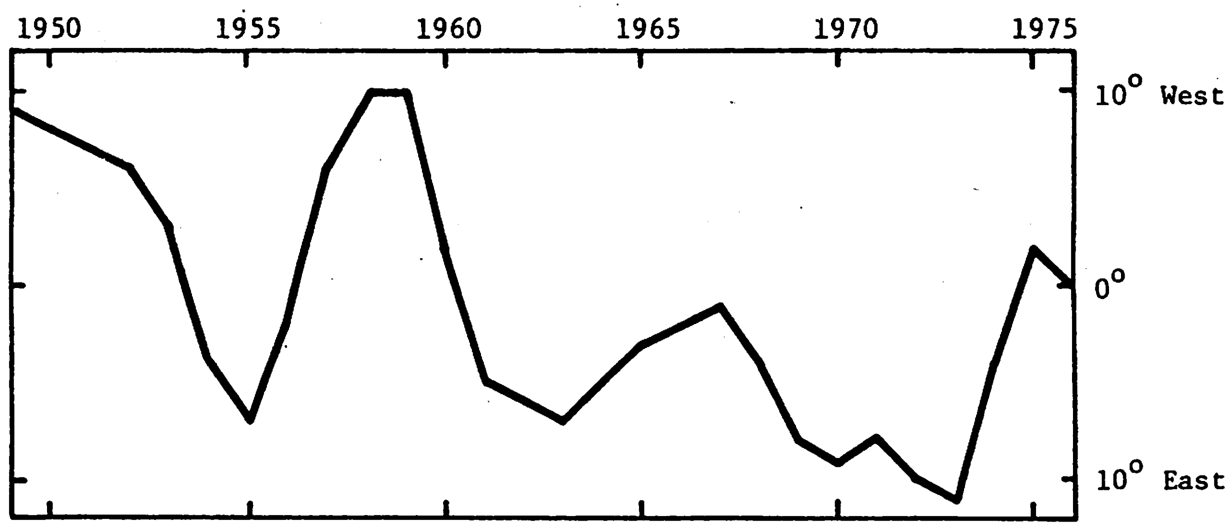


FIG. 4.5. Displacement of the 700mb Baffin trough from the long-term average position, in degrees of longitude, 1949-76. Values are smoothed by a 1-2-1 running mean.

The excursions of the 700mb low also compare favorably with the shifts in surface cyclone activity seen in Fig. 3.5. It is not surprising to find that a 700mb low displaced to the west and southwest is associated with more days with cyclones west of Baffin Island, and to the east and northeast with more Baffin Bay cyclones. No relation between the 700mb low and southeast cyclones is obvious. The 1969 jump in days with anticyclonic synoptic types occurred as the 700mb low shifted to its most northeasterly displacements of the 30-year period, apparently with a corresponding eastward shift of the polar anticyclone track. For decadal averages, the tendency for the 700mb low to stay to the southwest during 1951-60 and to the northeast during 1964-73 is reflected in the decadal change in the distribution of surface lows (Fig. 3.4b). The longitudinal displacement of the 700mb low between the two decades was 9.4° eastward, with a 99.9 percent significance. The correlations between the displacement of the 700mb low (west positive), Baffin Bay summer temperatures, and numbers of days with different grouped synoptic types are summarized in Table 4.1. All correlations are consistent among the different time scales. Another correlation effective on all three time scales is a tendency for the 700mb low to be deeper (stronger) when it is displaced eastwards, by 10 meters per 15° of longitude.

Thus, the following regional climate and circulation scenario appears to be effective on time scales ranging from interannual to inter-decadal: Cool summers over Baffin Bay are accompanied by a sizeable increase in cyclonic activity over the Bay, and by a decrease in the number of days with cyclones to the south and

TABLE 4.1. Correlations of Baffin Bay Temperatures and Synoptic Type Frequencies With 700mb Trough Longitude

	Yearly	1-2-1 Smoothed	1951-60 to 1964-73
Baffin Bay Temperature	.57**	.75**	+
NE Cyclones	-.61**	-.69**	-
SE Cyclones	.14	.46	+
W Cyclones	.62**	.69**	+
Anticyclones	-.53**	-.68*	-

*95 percent significance; **99 percent.

southwest. The decrease in southwest cyclones is particularly significant, for these systems bring warm southerly and southeasterly winds to the region. The net effect is a shift to the north and east of cyclone activity, large enough to reverse the prevailing surface wind direction over southern Baffin Bay from easterly to westerly. The associated eastward displacement of the upper trough increases the number of anticyclones brought into the region by the northwesterly flow on the trough's western flank.

Surface cyclones and upper troughs are intimately connected; the next topic of this chapter concerns the origins and tracks of cyclones associated with the shifting upper trough.

Surface Cyclone Tracks

The major cyclone tracks for mid-summer, derived from a gridded count of July cyclone tracks for 1951-70 (from Reitan, 1974; similar tracks are given by Klein (1957) and McKay *et al.*, 1970) are shown in Fig. 4.6. The relative frequencies of cyclones along each of the four tracks that could affect Baffin, as inferred from the mean June and July occurrence of cyclogenesis in the regions centered by an "x" in Fig. 4.6 (also from Reitan, 1974) are: 7 cyclones per June-July from the District of Mackenzie region, 19 from the Montana area, 14 from the east coastal region of the United States, and 13 from the Labrador Sea. An analysis of the June-July gridded data (from Reitan, personal communication) reveals that the frequency of cyclogenesis decreased in three of the areas between 1951-60 and 1964-73, with a slight increase in the number of Mackenzie cyclones, from 6 to 7 per June-July.

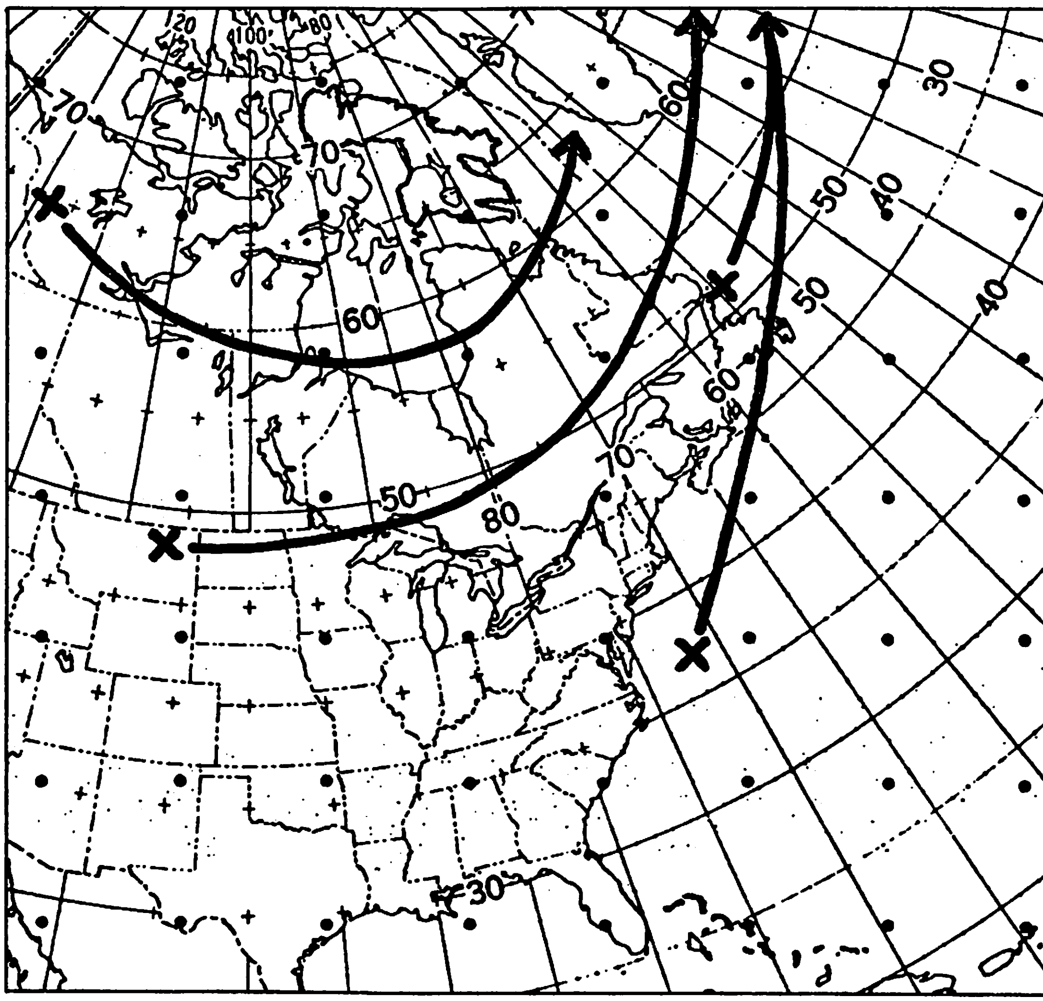


FIG. 4.6. Major July storm tracks, from Reitan (1974). "X" marks favored locations for cyclogenesis.

Cyclogenesis over Baffin Bay remained constant at about 1 per June-July. The change of the cyclone tracks between the two decades, in terms of numbers of tracks per June-July crossing a grid square, is shown in Fig. 4.7. A general decline in storm activity by about 20 percent is evident across the entire area, with the exception of a small path tracking eastwards from the Mackenzie cyclogenesis area. The data in Fig. 4.7 seem to contradict the results found in Chapter Three, namely, a finding of no change in the number of cyclones passing south of Baffin Island, and an increase of their number over Baffin Bay. The first apparent contradiction can be resolved by noting that the number of cyclone days at the extreme southern edge of the synoptic typing grid -- at about the same location as the Mackenzie cyclone track -- increased between the decades (Fig. 3.4b). The second contradiction -- concerning Baffin Bay cyclones -- is knottier. It is not resolved by considering only June and July synoptic type data; both June-July and June-July-August counts of continuous spells of Baffin Bay cyclones show an increase of about 40 percent; the cyclone track count for the two grid squares encompassing Baffin Bay decreased 30 percent. The general drop in mean sea-level pressure over Baffin Bay (Fig. 3.6) precludes the possibility that the Baffin Bay cyclones were weaker during 1964-73 than earlier, and therefore less likely to be tracked. However, there may have been some subtle change in the criteria for tracking these cyclones, although no such procedural change is noted on the basic data (the storm track maps published monthly in the Monthly Weather Review and Climatological Data, National Summary). However, it is clear

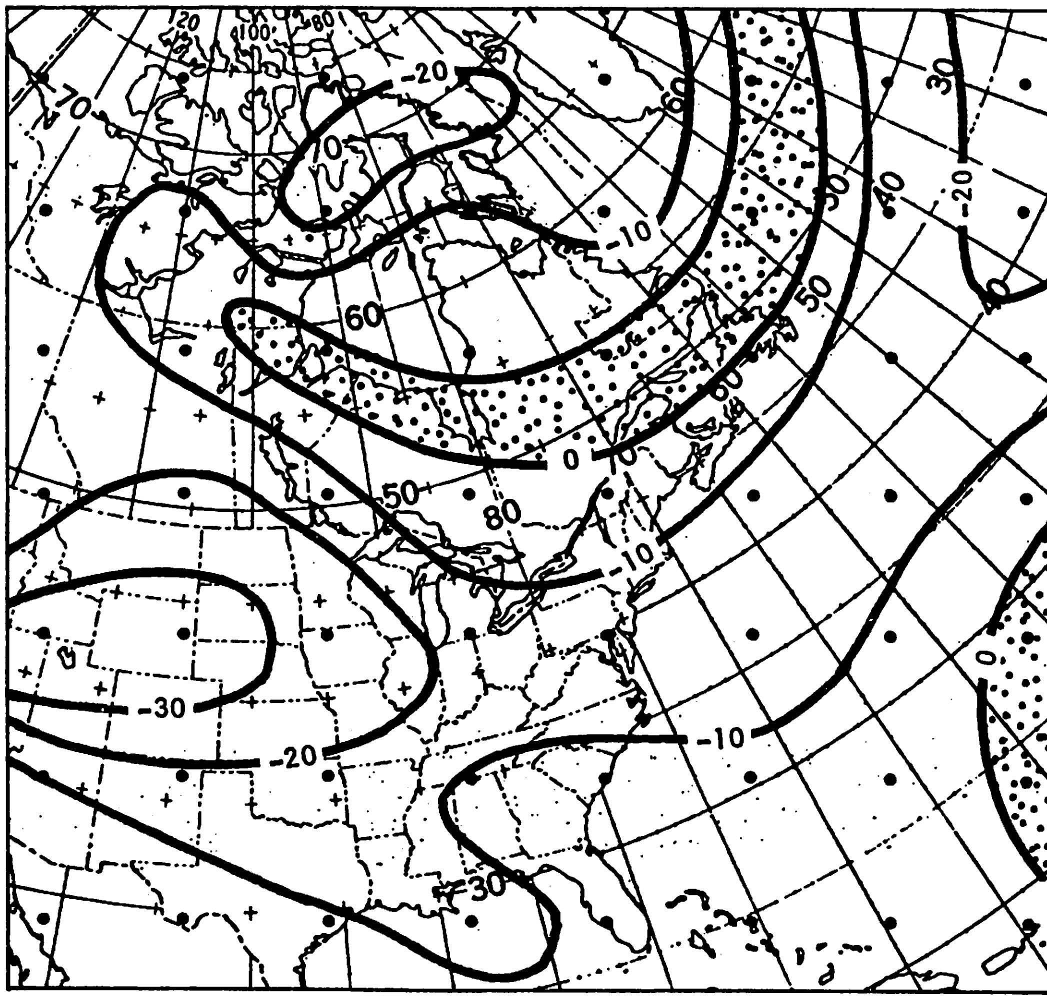


FIG. 4.7. Decadal change in summer (June and July) surface cyclone activity across North America and the western North Atlantic, 1951-60 to 1964-73, in terms of numbers of storm tracks crossing grid squares. Large dots indicate centers of grid squares; stippled areas are those of increased storm activity in 1964-73.

another approach is needed to locate the origin of the additional Baffin Bay cyclones.

Dzerdzeevski (1968, 1970) developed a subjective classification scheme for the Northern Hemisphere atmospheric circulation, based on the distribution of areas of surface cyclone and anticyclone activity. The primary criteria for his types (or Elementary Circulation Mechanisms, ECM) are the number and geographic longitudes of middle latitude cyclone and anticyclone tracks. Dzerdzeevski's catalog extends from 1899 through 1969, and each day is classed as one of his 41 types (of which 31 occur during the summer months). Thirty-one is a cumbersome number of data sets to analyze, and was reduced to three by grouping all ECM's with similar cyclone-anticyclone activity in the North American-Western Atlantic sector. The three grouped ECM's are:

A. Northbound cyclones over the western Atlantic and the east coast of North America; polar anticyclone outbreaks over the eastern Canadian Arctic. Includes Dzerdzeevski ECM's 8a, 9b, 12a, 12bz, 12bl, 12cl, 12d.

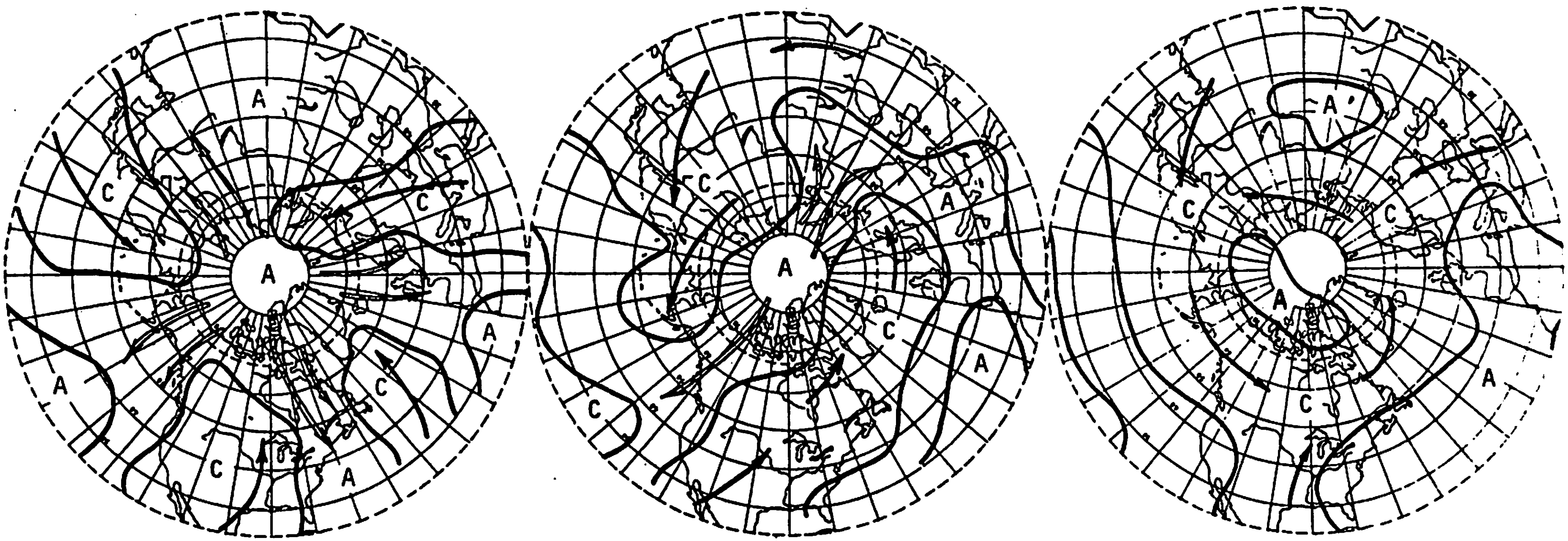
B. Cyclones over central and eastern North America; polar outbreaks over western Canada. Similar to A, but the pattern is shifted westward. Includes ECM's 7al, 7bl, 9a, 10a, 10b, 11b.

C. Zonal flow across North America, with cyclones moving west to east at higher latitudes. Also cyclones moving northeastward across the continent at middle latitudes. Includes ECM's 1a, 1b, 2a, 2b, 2c, 3, 4a, 4b, 4c, 5b, 5c, 6, 8bl, 8bz, 8cl, 8dl, 8dz, 13ℓ.

The schematic diagrams for one representative ECM of each group are in Fig. 4.8 (from Dzerdzeevski, 1970). In general, the higher numbered ECM's are the more meridional (higher wavenumber) circulation patterns.

The number of June-July-August days with each of the three grouped ECM's (Fig. 4.9) underwent marked changes around 1960. Between 1951-60 and 1963-69 (unfortunately, the catalog ends in 1969) the group A days increased by an average of 19.0 per summer, while groups B and C decreased by 9.5 days each. The changes are all significant at the 99 percent level. Thus, the meridional group A and B days gained at the expense of the zonal group C days, and among the meridional days the axis of cyclone activity shifted eastward, from the North American continent to the Atlantic coastal region. These changes are consistent with the eastward shift and intensification of the 700mb trough at the same time. This eastward displacement of storm activity is also suggested by the cyclone track counts (Fig. 4.7), which shows greater decreases in activity over the continent than over the western Atlantic at middle latitudes. Dickson and Namias (1976) note an eastward shift of winter storm activity off the east coast of the United States in the early 1960's.

Apparently the increase in meridional circulation patterns was a hemispheric phenomenon; Dzerdzeevski and Sergin (1972) have divided the Northern Hemisphere into six sectors, and note an increase in the number of days per summer with meridional ECM's between the 1950's and 1960's for all six sectors. The North American sector gained 18 days/summer between the two decades; the North Atlantic sector, 9 days; and the six sector average, 13 days.



ECM 12a (Group A) ECM 10b (Group B) ECM 2b (Group C)

FIG. 4.8. Schematic maps of representative examples of each of the three groups of Dzerdzeevski elementary circulation mechanisms (ECM). A and C on the maps denote, respectively, regions of anticyclonic and cyclonic activity; arrows indicate the motions of these systems. Maps adapted from Dzerdzeevski (1970).

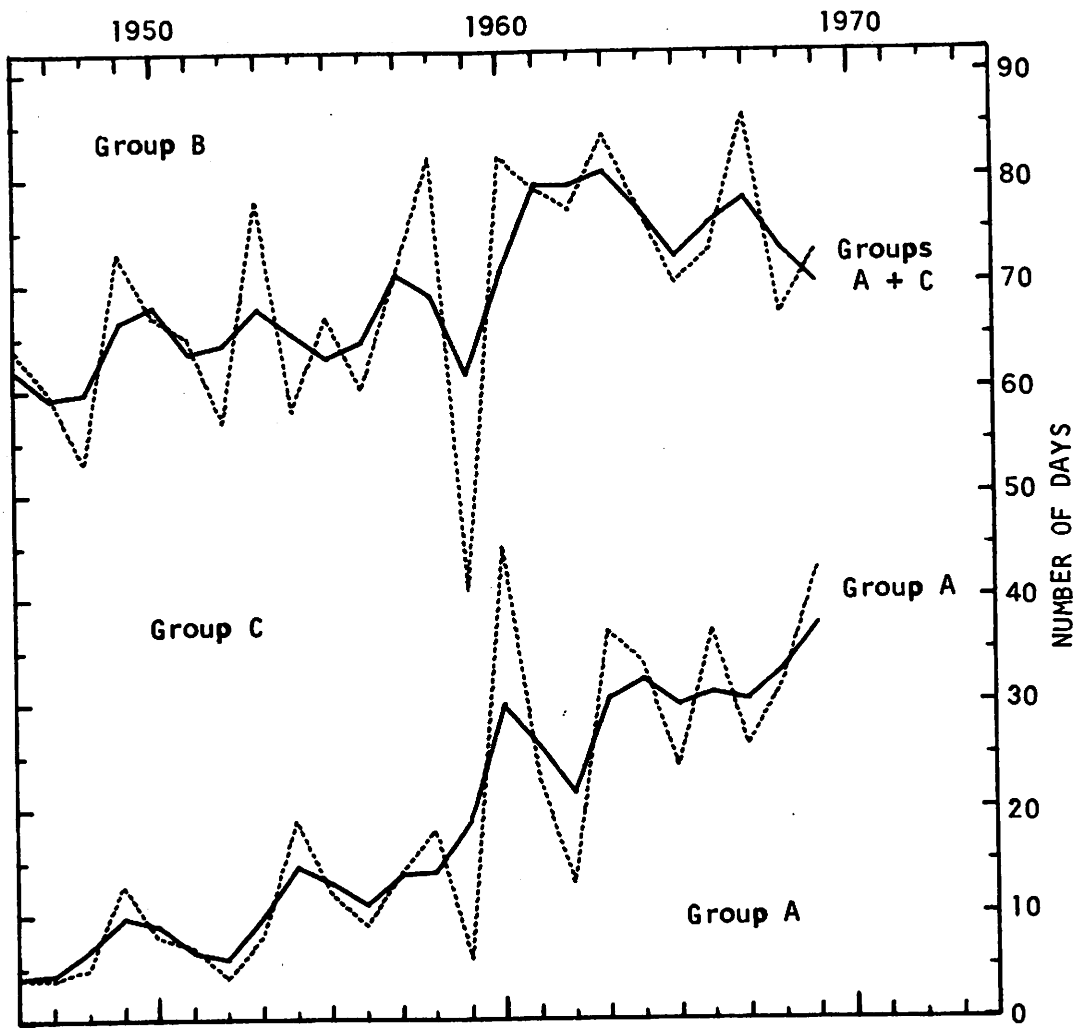


FIG. 4.9. Number of days per summer with each of the three grouped Dzerdzeevski elementary circulation mechanisms (ECM), 1946-69. Annual values, dotted lines; smoothed values, solid lines.

Dzerdzeevski and Sergin observe similar changes in other seasons, and mark 1962 as the end of a 30-year long epoch of predominantly zonal circulation.

The analysis of Dzerdzeevski's ECM's links an eastward shift of middle latitude cyclones to the eastward and northward shift of Baffin area cyclone activity, but leaves open the question of whether the two events are directly connected, i.e., are the additional Baffin Bay cyclones of west Atlantic origin?

Vorticity Flux

Surface cyclone centers are reflected in the upper levels by disturbances in the large-scale flow called short baroclinic waves (Palmén and Newton, 1969, Chapter Six). While these upper short waves rarely develop the closed circulations that are common with their surface counterparts, they are easily identifiable in computed vorticity fields as small areas of high positive vorticity. Maps of the 500mb vorticity field are among the most valuable tools currently used by meteorological forecasters and analysts in predicting and explaining the formation, growth, and decay of storms. It was therefore anticipated that a climatological analysis of the 500mb vorticity field might reveal some of the atmospheric dynamics responsible for the observed fluctuations in surface cyclone activity.

Since the approach of this study is to compare sums and averages of summer mean conditions, analysis of mean vorticity fields would add little insight except to confirm the shifts of the mean trough. In order to identify the source regions of cyclones affecting Baffin, the flux of vorticity must be considered. Relative, rather than absolute, vorticity was chosen for the computations because it is a direct measure of the intensity of the disturbance, and it was decided to separate the relative vorticity flux into its positive and negative components, to avoid the mutual cancellation by successive short wave troughs and ridges.

The calculations were done on the NCAR CDC 7600 computer, and the basic data source was the NCAR file of daily NMC gridded 500mb heights for 1946 to 1977. For each day (1200 GMT) during that period, geostrophic wind and relative vorticity were computed at each point on the grid, except at the very edge.

Relative vorticity is computed from:

$$\zeta = (g/fd^2)\cdot(z_2+z_3+z_4+z_5-4z_1) \qquad (4.1)$$

where z_1 is the 500mb height at the grid point where ζ is being computed; z_2, z_3, z_4, and z_5 are the heights at the four nearest points on the rectangular grid (Fig. 4.10); d is the grid spacing, ranging from 306 km at 30°N, to 381 km at 60°N, and 408 km at the north pole (Jenne, 1970).

The components of the geostrophic wind, in grid coordinates, are:

$$U = -(g/2fd)\cdot(z_5-z_3) \;;\quad V = (g/2fd)\cdot(z_4-z_2) \qquad (4.2)$$

Thus there is an effective smoothing of ζ, U, and V over an area about 700 to 800 km across at arctic latitudes.

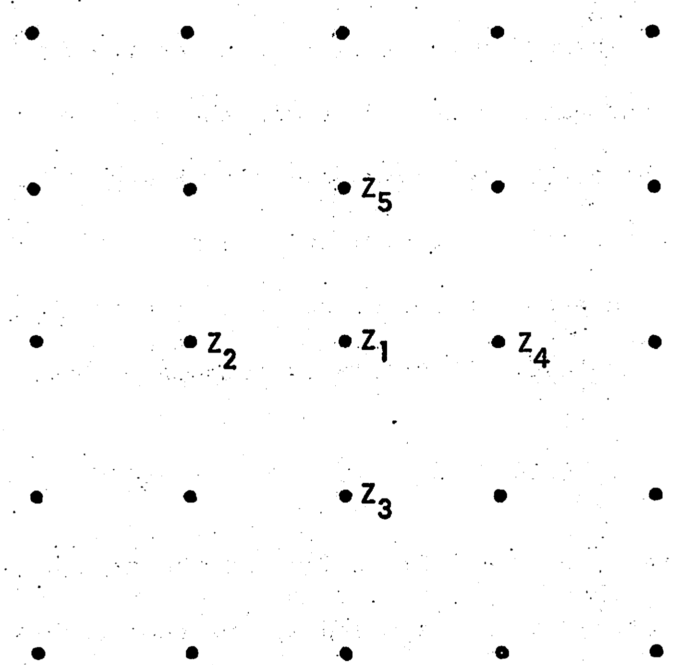


FIG. 4.10. Sample NMC Grid

The positive and negative relative vorticity fluxes at each point are vectorially summed over each season, then divided by the number of days with data in that season. No summer from 1946 to 1976 has more than six days of missing data. Multiplication of ζ by $\vec{V}$ gives greater weight to the faster moving systems for the instantaneous (daily) values of vorticity flux, but these systems affect a given point for proportionately fewer days. Therefore, the net accumulated vorticity flux is proportional to the integrated number, size, and intensity of those systems passing over a point, and is independent of the speed at which the systems move. Since 500mb is near the level of non-divergence, the implicit assumption that the vorticity field is moving with the wind velocity is approximately correct.

Once the hemispheric vorticity flux vector field is known, the distribution of sources and sinks (corresponding to cyclone formation and decay, in the case of positive vorticity) can be calculated from the quantity $\frac{\partial \zeta}{\partial t} + \nabla \cdot (\zeta \vec{V})$. For a seasonal average, $\frac{\partial \zeta}{\partial t}$ is the local change in vorticity between the first and last days of the season, divided by the length of the season; it is typically one or two orders of magnitude smaller than the other term, the vorticity flux divergence. In the geostrophic approximation, $\nabla \cdot \vec{V}$ is not computed, and the computed quantity reduces to $\frac{\partial \zeta}{\partial t} + \vec{V} \cdot \nabla \zeta = \frac{D\zeta}{Dt}$.

The relative vorticity equation (e.g., Godske <u>et al</u>., 1957) is:

$$\frac{D\zeta}{Dt} = -(\zeta + f)\ \nabla \cdot \vec{V} - \frac{f}{a} V_y + \begin{array}{l}\text{(other source/sink terms;} \\ \text{frictional dissipation,} \\ \text{conversion of horizontal} \\ \text{vorticity)}\end{array} \tag{4.3}$$

The computed quantity, then, represents the net effect of all source and sink terms in the generation of positive and negative vorticity (computed separately), averaged over a season.

The flux divergence is computed as:

$$\nabla\cdot(\zeta\vec{V}) = (\zeta_4 U_4 - \zeta_2 U_2 + \zeta_5 V_5 - \zeta_3 V_3)/2d \qquad (4.4)$$

the subscripts referring to the points in Fig. 4.10.

One typical moderately strong short wave trough with an average vorticity of 10^{-4} sec^{-1} over a length of 1000 km will contribute $13x10^{-6}$ m/sec^2 to the seasonally averaged positive vorticity flux. Computed scalar magnitudes of vectorially averaged positive vorticity flux in the middle latitude storm belt in summer are about $200x10^{-6}$ m/sec^2 (Figs. 4.11, 4.12), representing the equivalent of 15 moderate short waves in a 92-day season. A typical positive vorticity source region (Fig. 4.13) has a point generation rate of $50x10^{-12}$ sec^{-2}, which, over a 92-day season, yields a point vorticity output of $4x10^{-4}$ sec^{-1}, equivalent to four moderate short waves.

The relationship between surface (1000mb) and 500mb vorticities is given by the equation:

$$\zeta_{500} = \zeta_{1000} + \frac{g}{f}\nabla^2 h \qquad (4.5)$$

where h is the thickness of the 1000mb-500mb layer. 500mb vorticity maxima would therefore be displaced towards the axis of cold air (or thickness trough) from the center of the associated surface cyclone. In general, the upper vorticity maximum would be north and west of the surface low at the onset of development, moving to

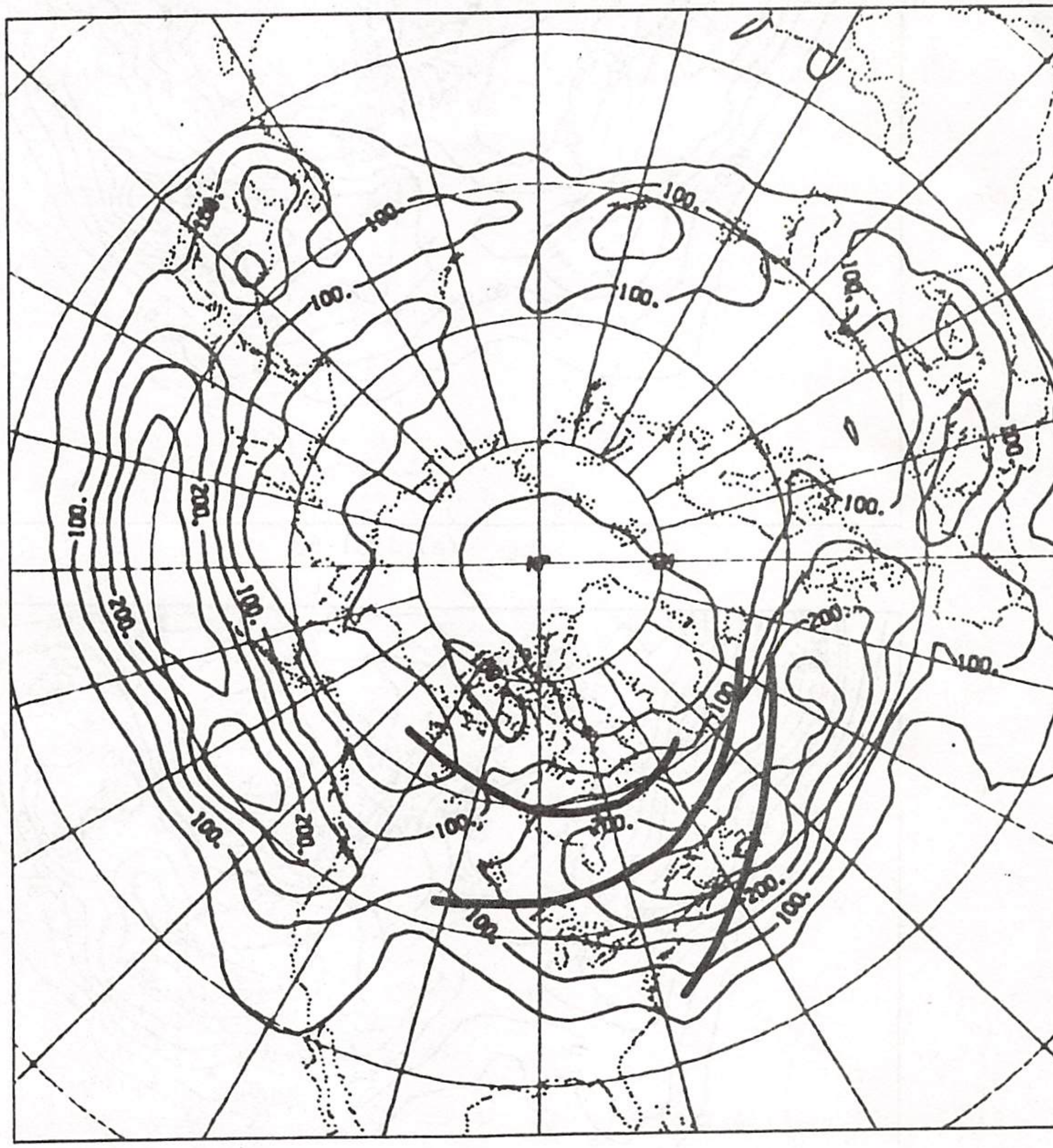
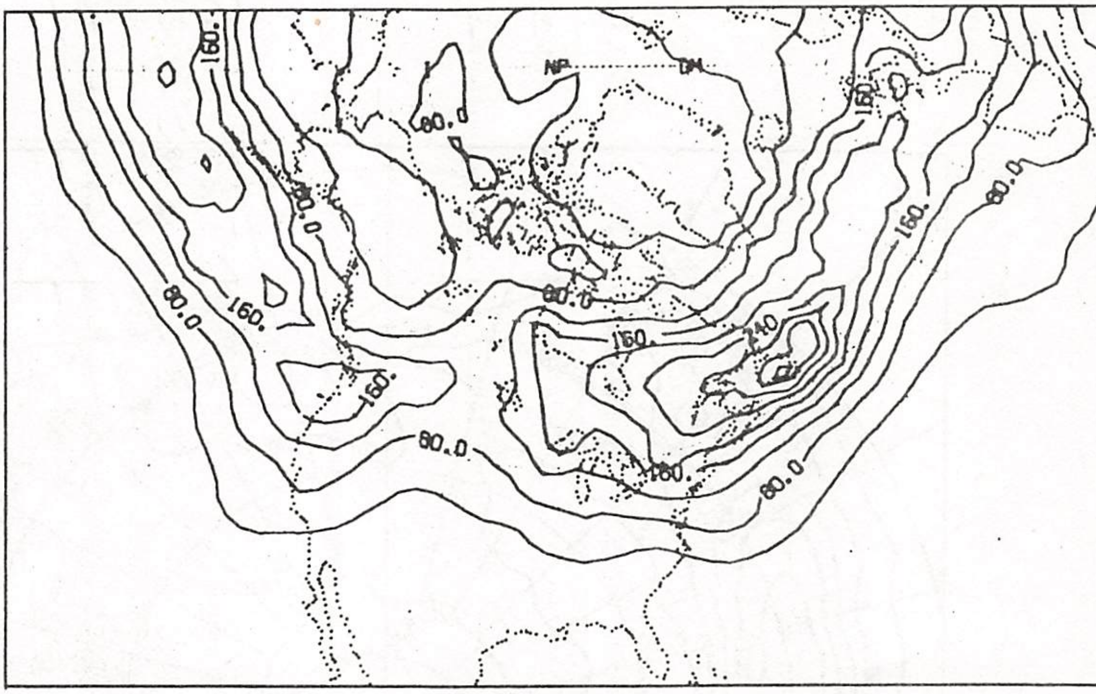
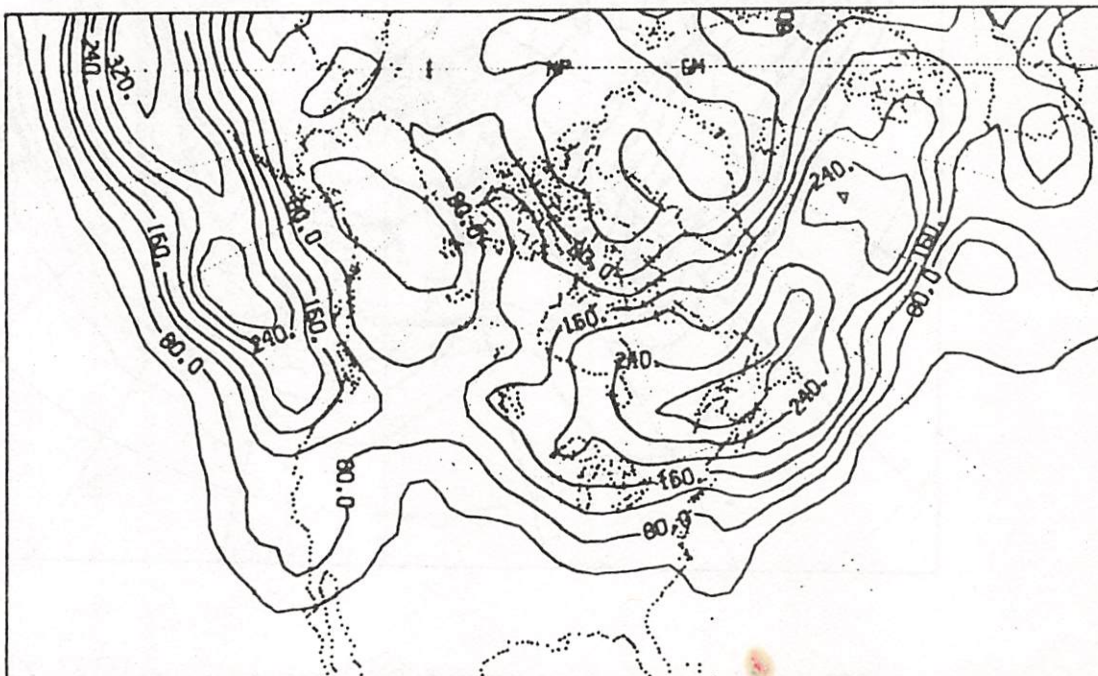


FIG. 4.11. Average summer 500mb positive vorticity flux, June-August, 1953-76. Heavy lines are surface storm tracks from Fig. 4.6. Contour interval, 50×10^{-6} m/sec^2.

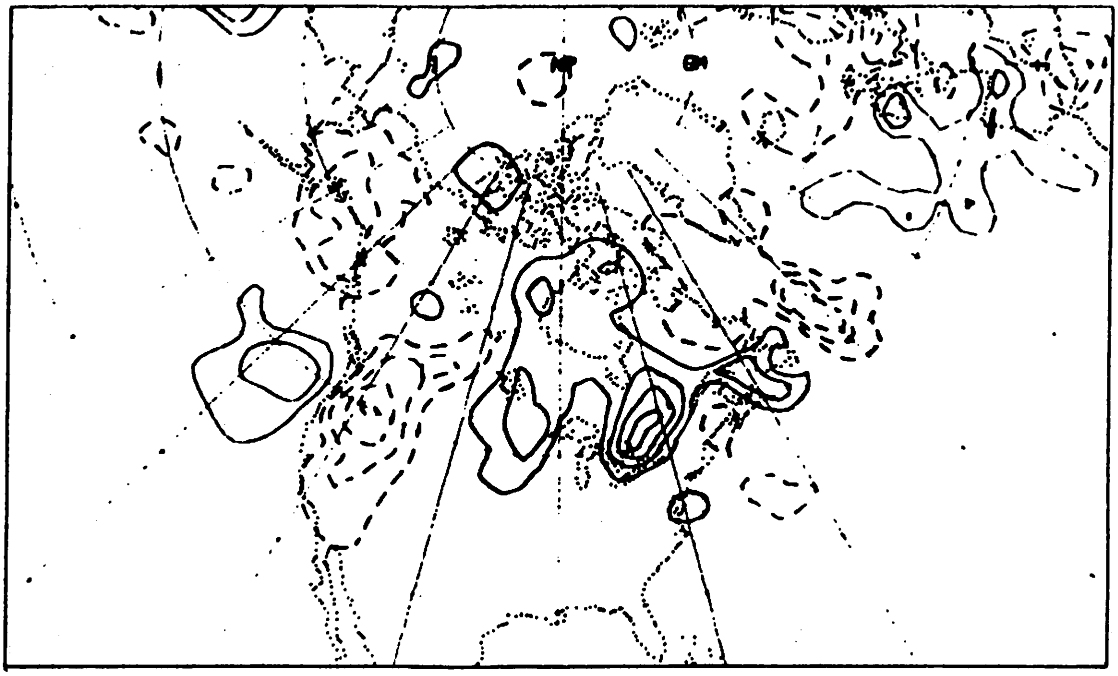


(a) 1951-60

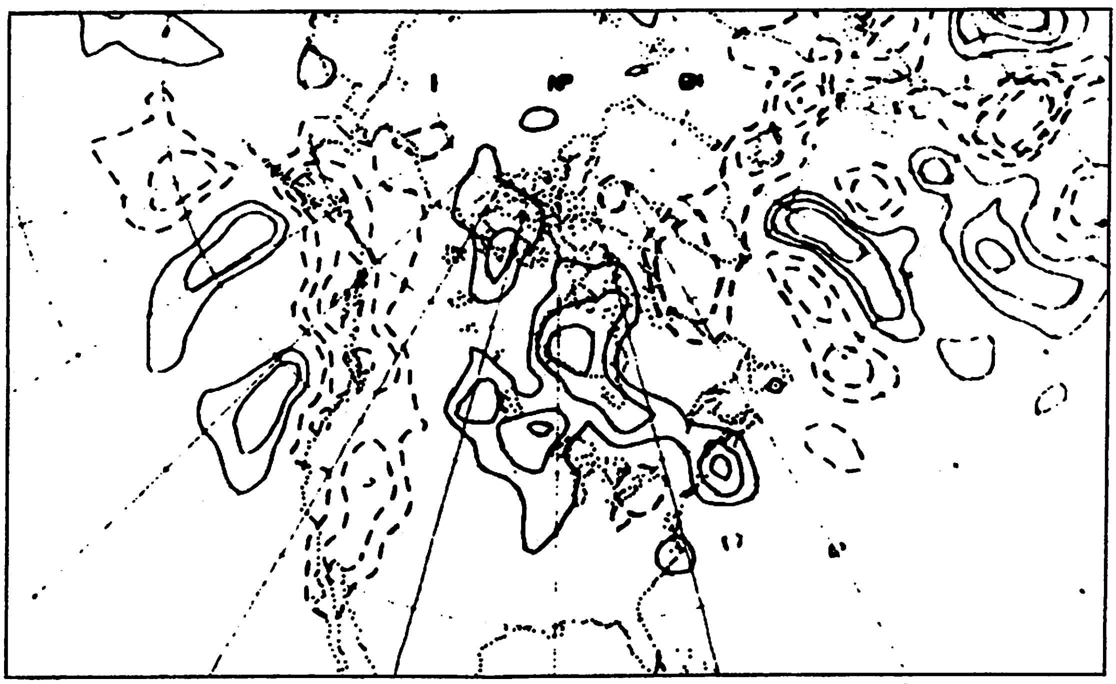


(b) 1964-73

FIG. 4.12. Average summer 500mb positive vorticity flux, 1951-60 (top) and 1964-73 (bottom). Contour interval, 40×10^{-6} m/sec^2.



(a) 1951-60



(b) 1964-73

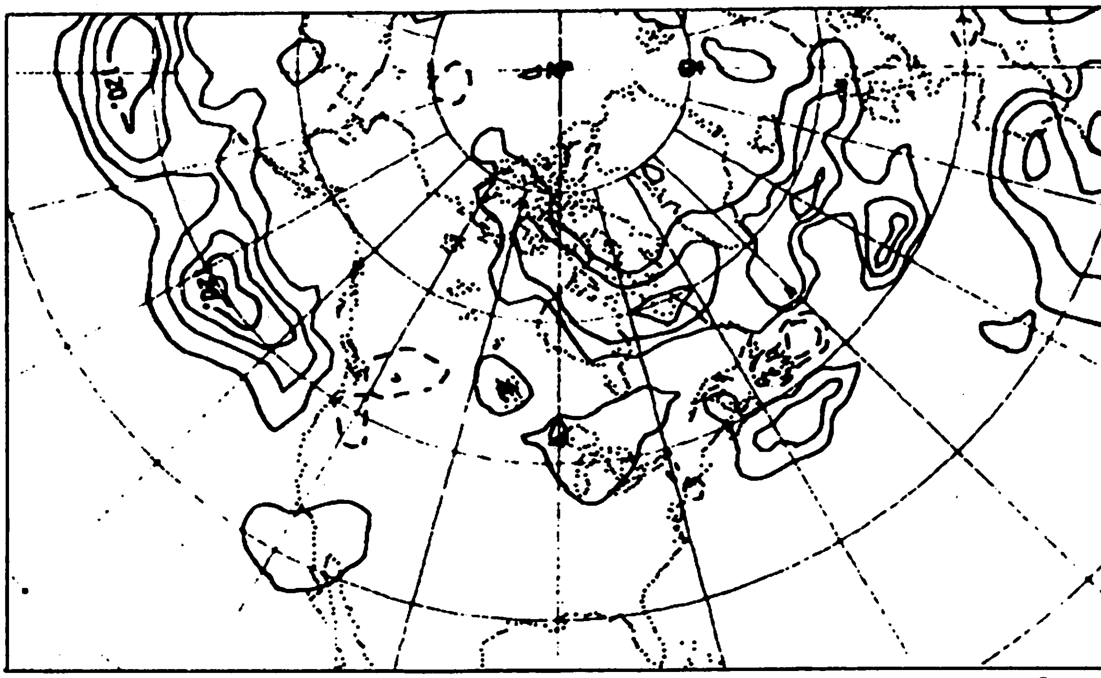
FIG. 4.13. Average summer positive vorticity flux divergence, 1951-60 (top) and 1964-73 (bottom). Contour interval, 30×10^{-12} sec^{-2}; solid contours indicate positive values (vorticity sources); dashed contours indicate negative values (vorticity sinks). Zero contours omitted for clarity.

the south and west of the surface low as the cold air is pulled southwards by the developing system (Sutcliffe and Forsdyke, 1950). When the upper trough overtakes the surface cyclone the system decays (Pettersen, et al., 1955). A typical path of an upper short wave would therefore begin north of the track of the surface low, cut southwards across the surface track as both progress eastwards, and finally rejoin the surface track at the end. This is illustrated in Fig. 4.11, where the July storm tracks from Fig. 4.6 are superimposed on a map of the 24-summer average positive vorticity flux. The effect is most pronounced with the surface cyclone tracks originating in Montana and the Mackenzie area. These cyclone tracks start some 5° to 10° south of their respective positive vorticity streams, but over the east coast of Canada they are north of the vorticity stream. However, there appears to be no tendency for the surface and 500mb disturbances to rejoin before the surface tracks end (by reaching the edge of their gridded analysis area) in the North Atlantic. The relation is more distinct on winter maps.

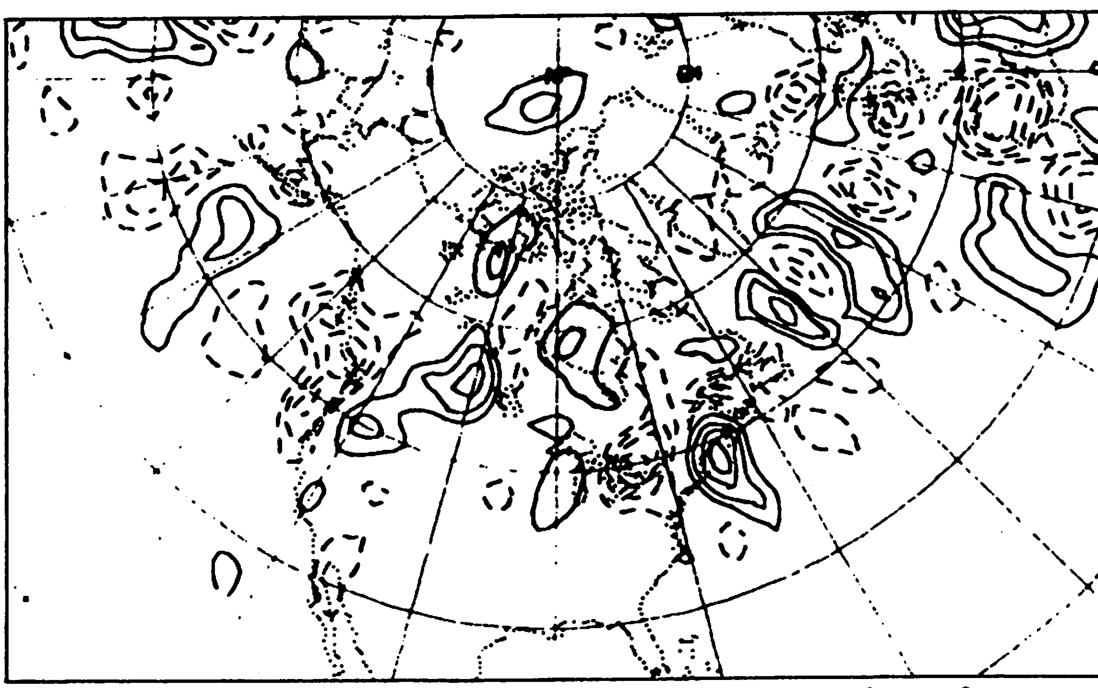
Decadal changes in the streams of positive vorticity flux can be seen in maps of the scalar magnitude of the positive vorticity flux for the decades 1951-60 and 1964-73 (Fig. 4.12). Immediately apparent is the 50 percent increase in the intensity of the Pacific positive vorticity stream. While the peak magnitude of the North America-Atlantic stream shows no increase, the stream broadens considerably to the north and south and total flux increases in the second period. The northward expansion is associated with higher fluxes (locally doubled) over the southern end of Baffin Island and a northward spur up the west coast of Greenland. This spur is clearly related to the increase in Baffin Bay cyclone activity.

The difference of scalar positive vorticity flux between the two decades is shown in Fig. 4.14a. A general increase is evident across the Northern Hemisphere. The increases over the western Atlantic and extreme eastern North America are greater than those over the interior of the continent, in agreement with the Dzerdzeevski ECM shifts noted earlier in this chapter. The greater frequencies of meridional ECM's -- with greater north-south excursions of cyclone activity -- may be responsible for the broader vorticity stream across the Atlantic. The slight decreases (or reduced increases) of vorticity flux near the axis of the Atlantic stream would then be due either to the spreading of activity over a greater area, or to the greater tendency of meridional components of vorticity flux to cancel out over a seasonal average than would be expected with zonal components.

It is apparent from Fig. 4.14a that it is not the northbound systems of middle latitude origin that are responsible for the increase in Baffin Bay cyclone activity, but rather systems that originate within the Arctic. The additional vorticity flux into Baffin Bay can be traced backwards to the Beaufort Sea area, from where it heads southeast, then turning east as it passes just south of Baffin Island. As happens frequently to surface cyclones approaching from the west (Reed and Kunkel, 1960; Putnins, 1970), a small fraction of the additional vorticity flux is then deflected into Baffin Bay by the blocking effect of Greenland. The location of this vorticity stream associates it with the surface cyclone



(a) Positive vorticity flux. Contour interval, 30×10^{-6} m/sec^2.



(b) Flux divergence. Contour interval, 30×10^{-12} sec^{-2}.

FIG. 4.14. Decadal change in magnitude of positive vorticity flux (top) and flux divergence (bottom), summer 1951-60 to 1964-73. Dashed contours indicate lower values in 1964-73; zero contours omitted for clarity.

track originating in the Mackenzie district -- the only one of the four surface storm tracks that gained in number of events between 1951-60 and 1964-73. The gain in these surface cyclones is about one per summer, while the increase in vorticity flux is about 60×10^{-6} m/sec^2, roughly equivalent to four moderately strong short waves per summer. This suggests that most of the vorticity flux increase is due to stronger, rather than more, systems.

Over Baffin Bay the vorticity flux approximately doubled between the two decades, a far greater increase than the 35 percent gain in the number of surface systems. This suggests that much of the vorticity flux increase is due to stronger systems at 500mb, which is consistent with the suggestion made in Chapter Three that more of the cyclones were deep cold lows during the colder 1964-73 decade.

Near the level of non-divergence, short waves would move with the speed and direction of the winds associated with the quasi-stationary long waves; this is the concept of "steering" (Palmén and Newton, 1969, Chapter Six). Fig. 4.15 shows the change in geostrophic 500mb wind speed between 1951-60 and 1964-73. The winds increase by about 2 m/sec, or 25 percent, along the storm track passing southwest, south, and southeast of Baffin Island, in rough agreement with the conclusion in Chapter Three that the W and SE cyclonic types move 15 to 17 percent faster during the second decade. The greater northward component of the steering flow over western Greenland shunts more of these systems towards Baffin Bay, confirming the suggestion made in the previous chapter.

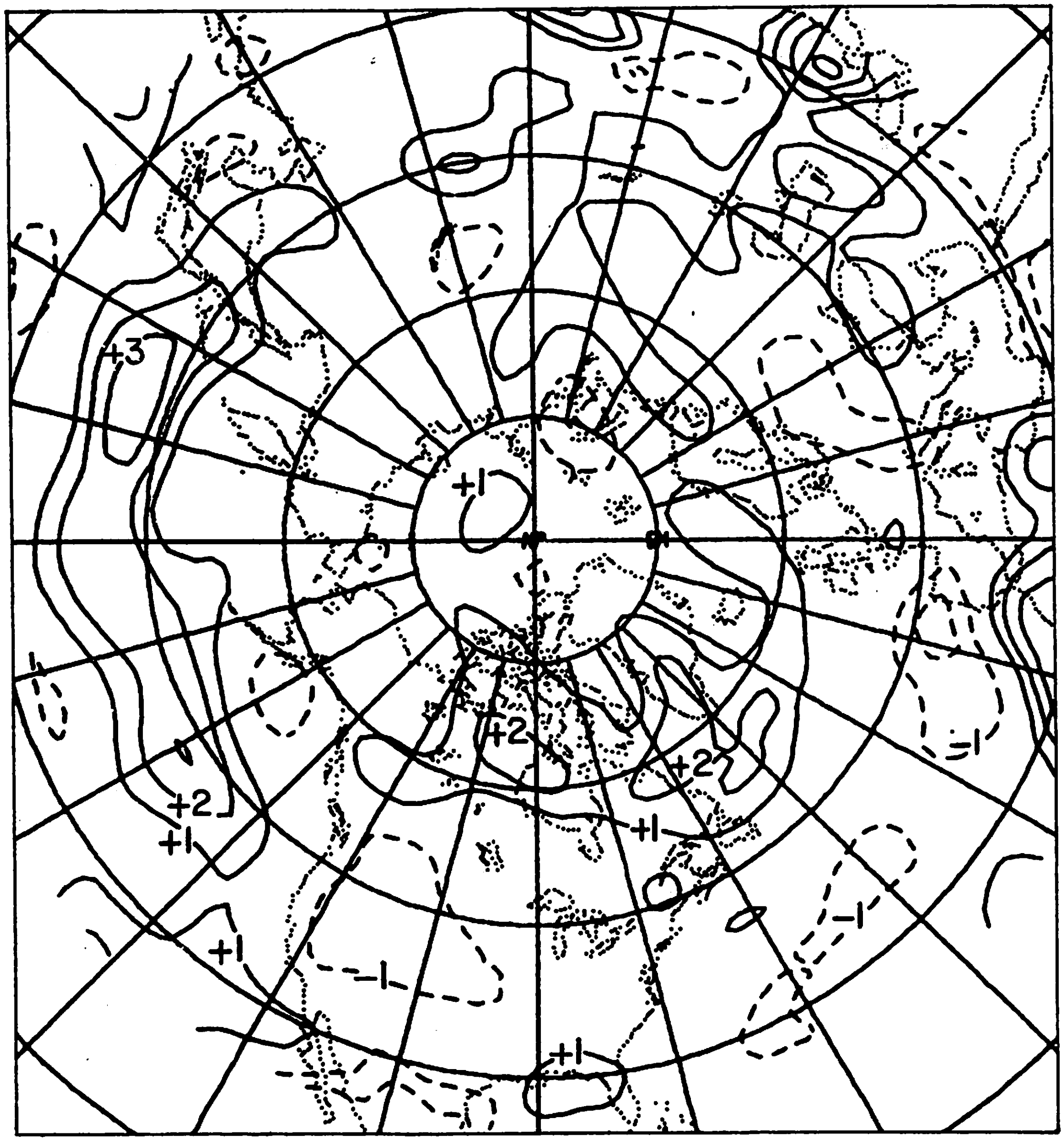


FIG. 4.15. Decadal change of 500mb geostrophic wind speed, summer 1951-60 to 1964-73. Contour interval, 1 m/sec; dashed contours indicate decreased winds in 1964-73. Zero contour omitted for clarity.

The positive vorticity sources and sinks for the two decades are shown in Fig. 4.13, and the decadal difference in Fig. 4.14b. Immediately noticeable is the eastward relocation of a major vorticity source, from Ontario to the New England coastal waters, another indication of the longitudinal shift of storm activity noted before. Of more immediate interest to the Baffin area is the movement to the southeast and intensification of the vorticity sources over Hudson Bay and the western Arctic Archipelago. These are the sources of the increased vorticity flux to Baffin Bay, and are located where a vorticity stream originating in Siberia passes over the mean summer position of the arctic front (Bryson, 1966; Barry, 1967; Hare and Hay, 1974) in northern Canada.

According to Pettersen et al. (1955), the majority of extratropical cyclones develop "when and where an area of appreciable positive vorticity advection in the middle and upper troposphere becomes superimposed upon a low-level frontal system." Strong regions of positive vorticity generation (and of surface cyclone formation) should normally occur, then, where a stream of positive vorticity flux intensifies in passing over a surface baroclinic zone. The arctic front in North America is such a region (Hare, 1968). Since the flux of vorticity reaching the arctic front in northwest Canada does not increase between the two decades, the intensified vorticity generation along the front is likely due to the front being stronger during the later, and colder, decade. The increase from 2.6°C to 3.5°C of the summer mean surface temperature difference between 120°W and 80°W at 70°N (from Figs. 2.7 and 2.11) between the two decades provides further evidence for a stronger arctic front.

While it appears that the eastward shift of cyclone activity around Baffin Island is associated with a general eastward shift of activity in this sector of the hemisphere, the actual storm systems that affected Baffin appear to come from the northwest.

The above discussion is restricted entirely to positive vorticity flux and its sources and sinks. Negative vorticity flux maps have also been produced, but are found to add little to the insight gained from analyzing positive vorticity. Polar anti-cyclones are low-level phenomena, without marked counterparts at 500mb. The flux due to short wave ridges follows, not surprisingly, the same paths taken by the flux due to short wave troughs. Negative vorticity fluxes would greatly exceed the positive vorticity fluxes in two situations: around the northern edges of the subtropical highs, and around warm blocking highs. The map of negative vorticity flux differences between the decades was examined for evidence of increased blocking in the Arctic, but none was found.

Hemispheric Cyclone Activity

From Fig. 4.14a it is clear the extratropical Northern Hemisphere experienced a general increase in storminess (as measured by positive vorticity flux) between 1951-60 and 1964-73; the regions of net decrease are few and small. The magnitudes of the changes can be appreciated by comparing the decadal zonal averages of positive vorticity flux (Fig. 4.16a). The zonal averages increase at every latitude between 25° and 80°N, and are significant

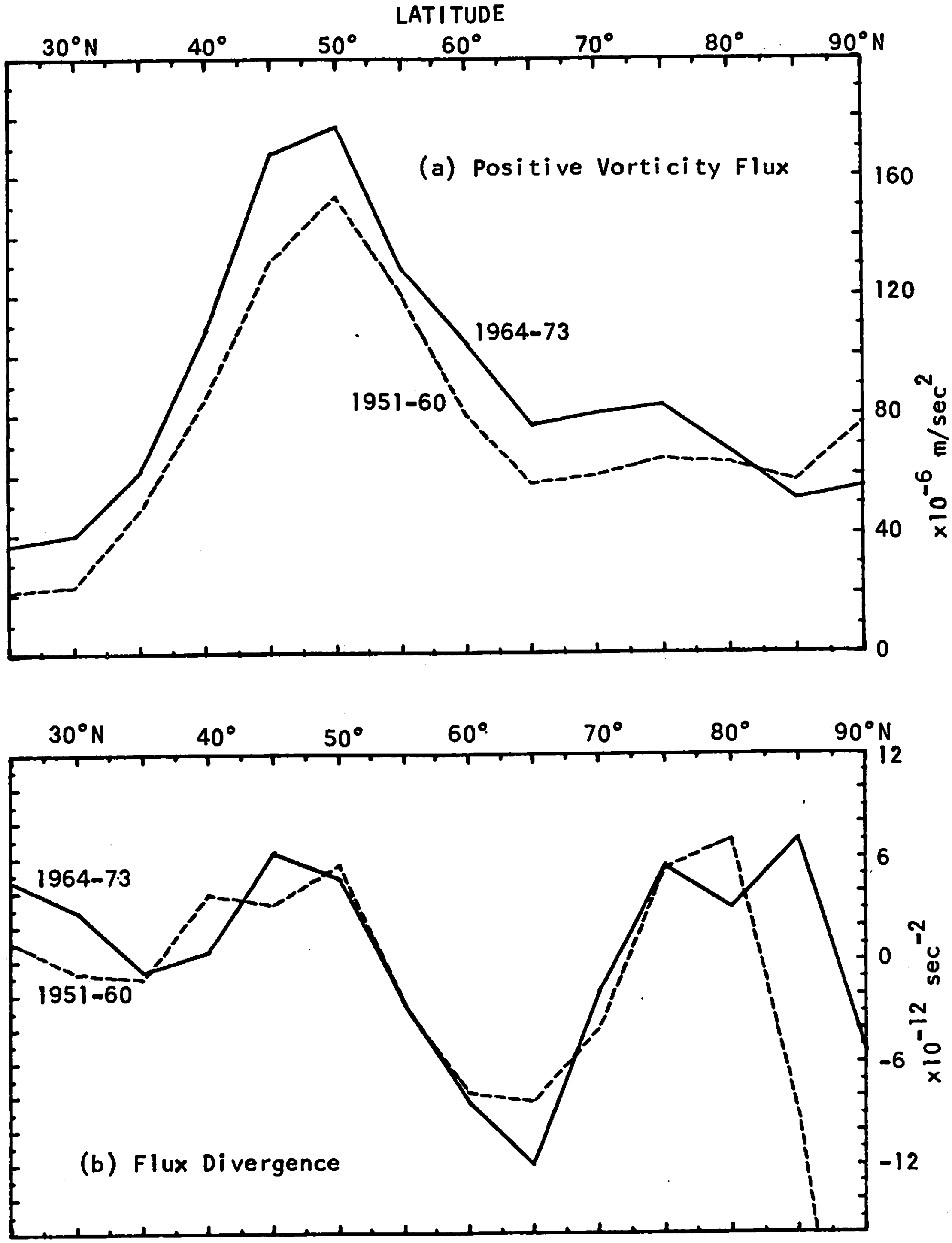


FIG. 4.16. Decadal means of zonally averaged positive vorticity flux (top) and flux divergence (bottom), summer 1951-60 (dashed line) and 1964-73 (solid line).

at the 99.9 percent level at all latitudes south of 75^{o}N, with the exception of 55^{o}N. The greatest gains are just north and south of 55^{o}N. The increase at 55^{o}N is relatively small because that latitude lies north of the Pacific vorticity stream, and near the center of the Atlantic stream (which, as noted earlier, has its greatest increases at its edges). The area-weighted average for the hemisphere north of 25^{o}N increases from 74.2×10^{-6} m/sec^2 to 93.2×10^{-6} m/sec^2, an increase of 26 percent.

While the area-weighted hemispheric average of vorticity flux source and sink terms cannot change over long time intervals, latitudinal redistributions of the zonally averaged sources and sinks can occur, and are shown in Fig. 4.16b. During both decades there is net vorticity generation south of 50^{o}N and from 75^{o} to 80^{o}N, while net vorticity dissipation occurred in the range 55^{o} to 70^{o}N. Thus, there is a net flux of vorticity from both north and south into the zonal sink region between 55^{o} and 70^{o}N, with a general intensification of both the zonal source and sink processes between the two decades.

The temporal variations of zonally averaged positive vorticity flux are displayed in Fig. 4.17. Immediately apparent is the sharp increase occurring at low latitudes around 1960 and reaching the Arctic two years later. The jump in flux magnitude is equally dramatic for all seasons of the year; in winter it occurred simultaneously around 1960 at all latitudes. The suddenness of the change leads to suspicions that the cause may be in the data, not the atmosphere. Indeed, 1960 was the year the National Meteorological Center switched from hand to computer analysis of

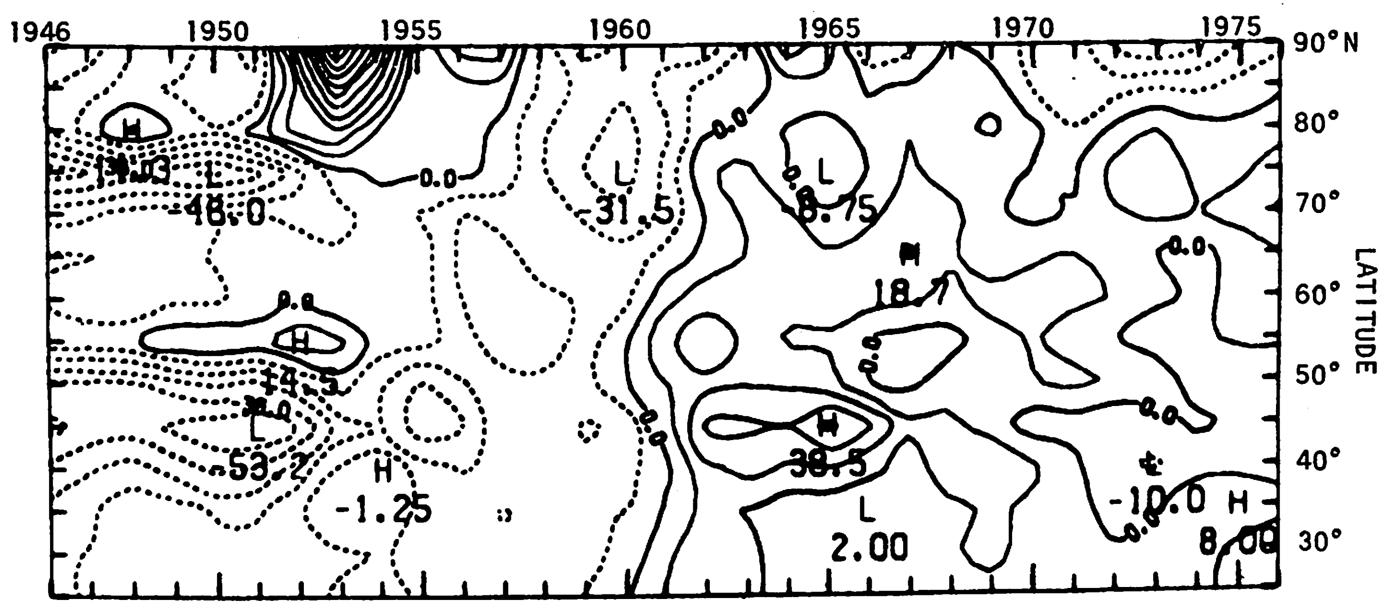


FIG. 4.17. Departure of zonally averaged positive vorticity flux from the 1954-76 average at each latitude, summer 1946-76, smoothed in time by a 1-2-1 running mean. Contour interval, 9×10^{-6} m/sec^2; dashed contours indicate negative values.

its upper air data. There was a resulting suppression of high wavenumber features by machine smoothing at first, followed by a recovery of these features as the NMC improved its analytical models. This effect is apparent in the 700mb mid-latitude meridional index, which decreased by 10 percent during 1960-61 (relative to the nine preceding years), then recovered (Wahl, 1972). Computed vorticity would be expected to show a similar decrease, but instead the vorticity flux increased, even over data-dense regions such as North America, where changes in interpolation procedures would have a smaller effect. Namias (1969) notes an increase, commencing in 1961, of cyclonic activity over the central and western portions of the North Pacific Ocean, another region of increased vorticity flux seen in Fig. 4.14a. It therefore appears the sudden onset of higher positive vorticity fluxes, coinciding with a sharp increase in the frequency of meridional Dzerdzeevski ECM's, is real. However, the large negative vorticity flux anomalies prior to 1954 may be due to sparse data.

Summary

The aim of this chapter is to identify the regional-scale circulation anomalies responsible for the temperature and synoptic weather type variations discussed in the previous two chapters. It has been demonstrated that both average summer temperatures and the occurrence of basic synoptic circulation patterns are closely linked to the average summer longitude of the Baffin trough. Eastward displacements of the trough are associated with more cyclone activity to the north and east of Baffin Island, more anticyclones passing over the Island, and cooler summers than are westward trough displacements.

Transport of exotic pollen to Baffin Island should also be closely linked to the location of the trough. Increased pollen transport would be expected to occur during periods of westward trough displacement (warm summers), with prevailing southerly winds aloft and more frequent synoptic events with southerly flow (W cyclones). Conversely, eastward trough displacements should cut off the influx of exotic pollen. Pollen records for the Cumberland Peninsula published by Andrews et al. (1978) and Nichols et al. (1978) show that episodes of heavy pollen influx -- such as during recent decades -- have been brief and infrequent (recurring at 200-250 year intervals) during the past 2000 years, implying that the trough has spent much of the past two millenia east of Baffin Island. However, studies correlating pollen transport with observed synoptic events and trough displacements are needed to verify any paleoclimatic reconstruction of the trough position from exotic pollen data.

Analysis of surface cyclone tracks, Dzerdzeevski hemispheric circulation types, and 500mb positive vorticity flux shows a general eastwards shift of middle-latitude storm activity over North America, in phase with the eastward shift of the Baffin trough between 1951-60 and 1964-73. However, the additional cyclone activity over Baffin Bay appears to originate with an intensification of the storm track running from the Mackenzie district to the

Baffin region. The increased cyclogenesis over the Mackenzie area is apparently related to a stronger arctic front. A general increase in storminess across the extratropical Northern Hemisphere accompanied these regional circulation changes between the two decades.

CHAPTER FIVE

HEMISPHERIC CIRCULATION INFLUENCES

In the preceding chapters of this study, interannual and and inter-decadal variability of Baffin area temperature and cyclone/anticyclone activity have been correlated with shifts of the Baffin trough and its associated storm tracks. In this chapter the hemispheric scale influences on these regional scale circulation changes are examined.

The chapter begins with a treatment of the Baffin trough as a standing Rossby wave in the sub-polar westerlies to determine the relation between trough position and zonal wind speed. The sub-polar westerlies are the northern fringe of the belt of westerlies encircling the Northern Hemisphere; the second section investigates the changes in strength and latitudinal distribution of the hemispheric westerlies (i.e., the intensity and structure of the north circumpolar vortex) that alter the speed of the sub-polar westerlies. Finally, the nature of the hemispheric westerlies are related to hemispheric scale tropospheric temperature changes, and possible causes of the largest scale fluctuations are discussed.

The Baffin Trough as a Standing Rossby Wave

The Rossby wave theory (Rossby, 1939) states that the length of a long standing wave in the atmosphere should be proportional to the square root of the speed of the westerly winds. If some wave feature upstream (west) of the Baffin trough, such as the Alaska-Yukon ridge or the Bering trough (Fig. 4.1), remains fixed in position, an increase in the westerly wind speed at that latitude would lengthen the wave, and push the Baffin trough eastward. The longitude of the Baffin 700mb trough is given in Fig. 4.5; the speed of the summer sub-polar westerly 700mb winds, averaged between 55° and $70^{\circ}N$, 0° to $180^{\circ}W$, is given in Fig. 5.1. Also shown in Fig. 5.1 are the mid-latitude westerlies (35° to $55^{\circ}N$) and the subtropical (20° to $35^{\circ}N$) easterly winds. These wind values are seasonal averages of the standard zonal indices, smoothed by a 1-2-1 running mean. The linear regression between the sub-polar wind speed and the Baffin trough longitude is 12.1° eastward displacement for each 1 m/sec increase in wind speed ($r = .74$), in qualitative agreement with the Rossby theory.

Inspection of Fig. 5.1 reveals a general out-of-phase relationship between the fluctuations of the sub-polar westerlies and of the subtropical easterlies, with a correlation of -.62, while the mid-latitude winds remain fairly constant. The area weighted average zonal wind over the whole latitude range, 20° to $70^{\circ}N$, is very constant, with the 1-2-1 smoothed values always between 3.1 and 3.4 m/sec. The implication here is that the variations in the sub-polar westerlies are due mostly to latitudinal excursions of the entire zone of westerlies (i.e., contractions and expansions of the circumpolar vortex), rather than reflecting changes in the total strength of the winds. However, the averaging of the zonal winds over 15° to 20° wide latitude bands may have removed some telling evidence that would refute this idea.

Seasonally and zonally averaged 500mb zonal winds at 5° latitude intervals from $25^{\circ}N$ to the north pole have also been

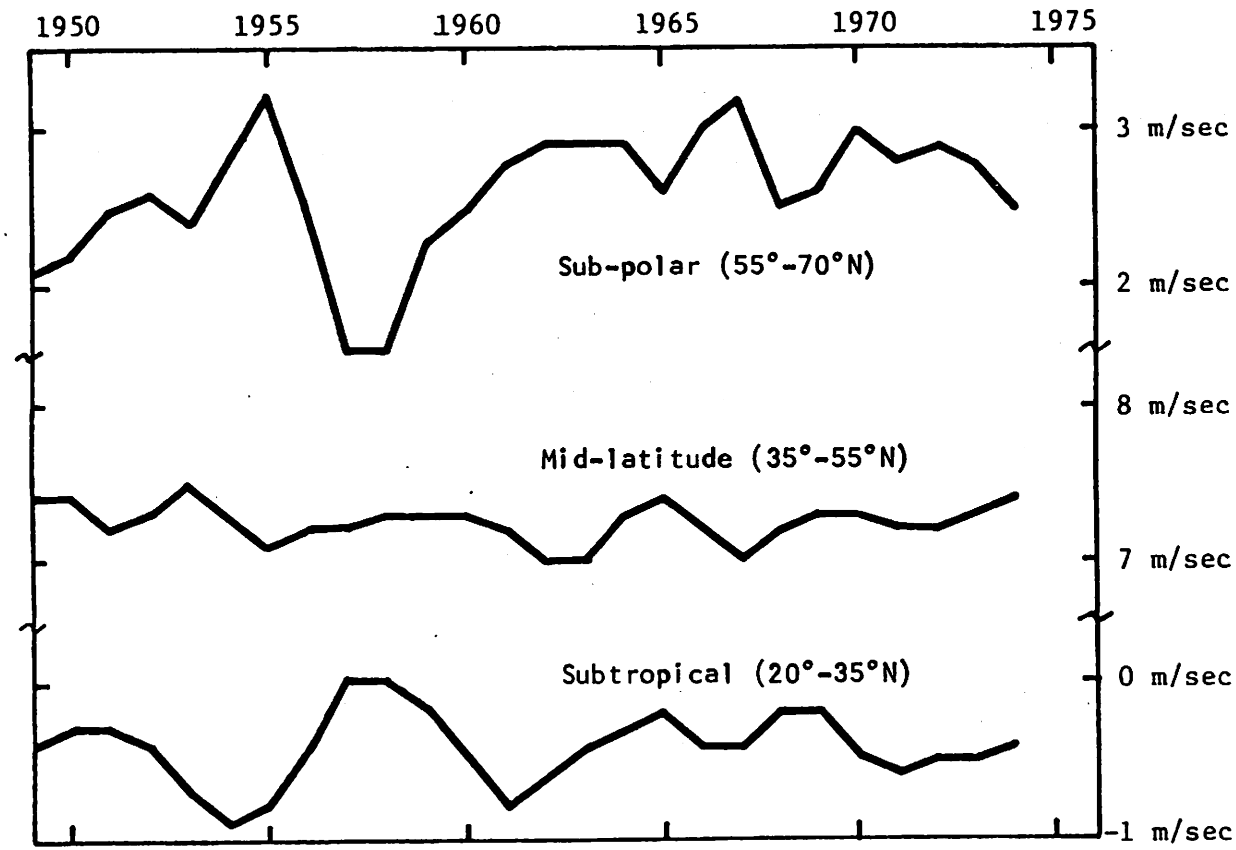


FIG. 5.1. Summer 700mb zonal wind speeds (westerly positive) averaged across three latitude bands, 1949-74, smoothed by a 1-2-1 running mean.

computed for 1946-76. Northern Hemisphere 500mb height maps have been computer plotted for each summer, enabling the position of the mean trough to be read. With this data base, a refined attack on the question of hemispheric scale influences on Baffin Island climate is possible, and since 500mb is near the level of non-divergence, the Rossby wave theory can be applied quantitatively.

The equation for the wavelength of a small amplitude sinusoidal standing wave (from Godske, et al., 1957) is:

$$L_s^2 = 360^2 U/2\Omega\ a \cdot \cos^3 \phi \tag{5.1}$$

where L_s is the stationary wavelength in degrees of longitude;

ϕ is the average latitude of the streamline being considered;

U is the average wind speed along the streamline, or approximately the zonal wind at latitude ϕ;

Ω is the rotation rate of the earth, 7.27×10^{-5} sec^{-1}; and

a is the radius of the earth, 6.371×10^6 m.

The incremental change of L_s for a change of U is given by:

$$\frac{\partial L_s}{\partial U} = \frac{L_s}{2U} \tag{5.2}$$

Placing 31-year average summer values of the zonal wind at three latitudes into the Rossby wave equation gives the following values of L_s and $\frac{\partial L_s}{\partial U}$:

ϕ	U (m/sec)	L_s (deg)	$\frac{\partial L_s}{\partial U}$ (deg/m$\cdot$sec^{-1})
67.5°	4.6	105°	11.7
72.5°	4.8	158°	16.4
77.5°	3.8	230°	30.2

The length of the wave associated with the Baffin trough, measured between it and the next upstream trough over the Bering Straits on the long-term average summer 500mb map (Fig. 4.1), is about 110°. This corresponds to $\phi = 68^{\circ}$, which can be seen from Fig. 4.1 to be the approximate average latitude of the streamline that passes through the Baffin and Bering troughs at 65°-70°N. Both troughs are caused in part by cyclone activity along the east coasts of their respective continents, but at arctic latitudes, the Baffin trough is the stronger of the two. This may be due to a longitudinal resonance between the forcing (cyclone activity) over Baffin and that at the upstream Bering trough.

The longitudes of the seasonal mean 500mb trough, averaged between 65° and 70°N for 1946-76 are shown in Fig. 5.2a. By comparison with Fig. 4.5, it can be seen that the longitudes of the 500mb and 700mb troughs correlate quite well ($r = .80$), although the 1964-73 average position of the 500 mb trough is 4.7° east of the 1951-60 average, a shift of only half that of the 700mb trough.

The correlation between trough longitudes and zonal wind strength is best for the zonal wind between 75°-80°N, where $r = .68$; for the 1-2-1 smoothed values of winds and longitudes, the correlation is best ($r = .74$) for the 70°-75°N and 75°-80°N zonal winds. These latitudes are somewhat higher than the average streamline latitude inferred from Fig. 4.1, but this is because the 24-year average in Fig. 4.1 smooths out the trough-ridge pattern of individual summers. Likewise, the 1-2-1 smoothing moves the relevant latitude a little farther south.

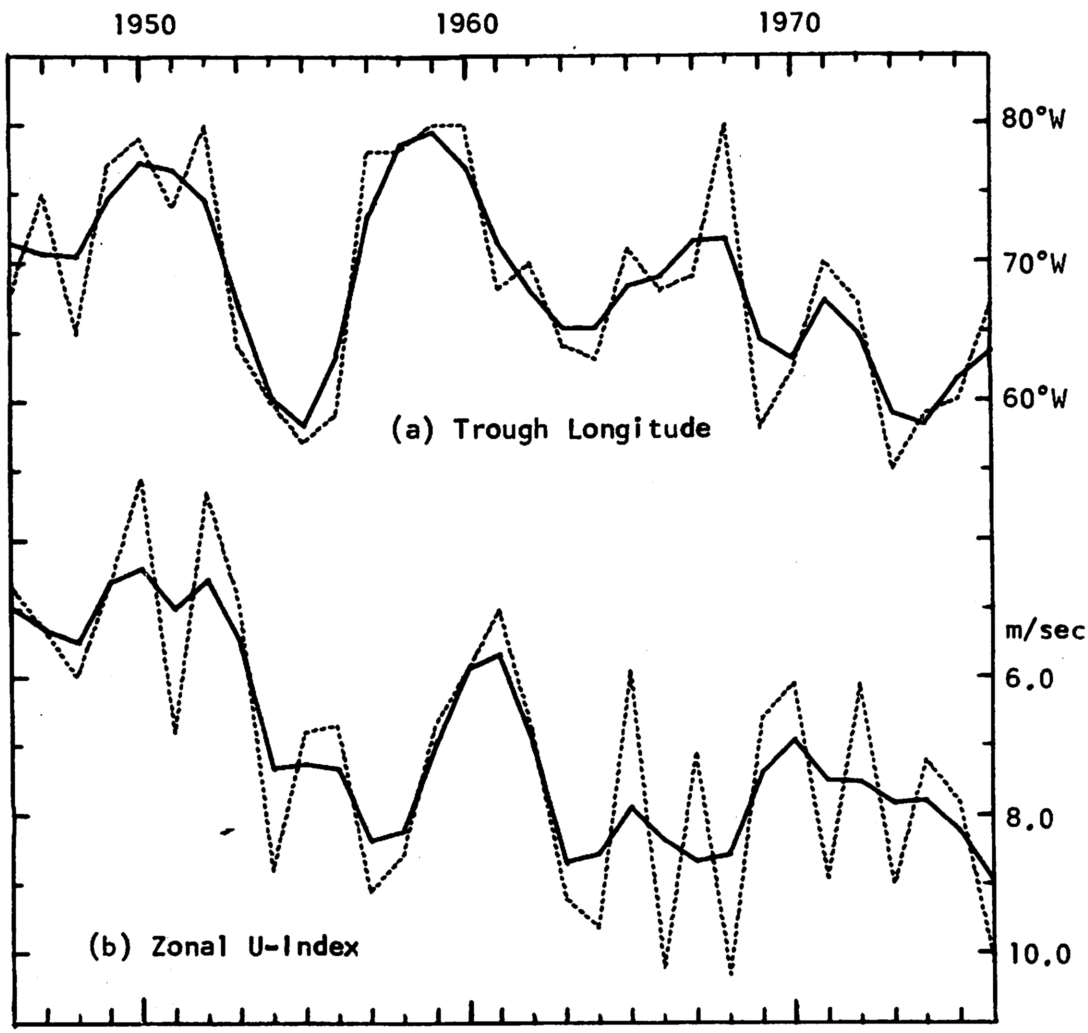


FIG. 5.2. 500mb Baffin trough longitude (top) and zonal U-index (defined in text), summer 1946-76. Annual values, dotted lines; smoothed values, solid lines.

The linear regression relation between the trough's longitudinal excursions and changes in the zonal wind are $3.7^{\circ}/\text{m}\cdot\text{sec}^{-1}$ (75°-80°N) on an annual basis, and $5.0^{\circ}/\text{m}\cdot\text{sec}^{-1}$ (75°N) for the 1-2-1 smoothed values. The position of the Bering trough, read from the seasonal mean 500mb maps, is also found to move slightly eastward with stronger zonal winds, by $1.5^{\circ}/\text{m}\cdot\text{sec}^{-1}$ on an annual basis and $0.7^{\circ}/\text{m}\cdot\text{sec}^{-1}$ for the smoothed values. The resulting wavelength changes are $2.3^{\circ}/\text{m}\cdot\text{sec}^{-1}$ and $4.3^{\circ}/\text{m}\cdot\text{sec}^{-1}$ for the annual and smoothed data; these figures are one-twelfth and one-fifth of their respective theoretical wavelength changes. This would imply that local influences, perhaps the ridging and occasional blocking induced in the tropospheric westerlies by the presence of Greenland (Namias, 1958; Lamb, 1972), prevent the Baffin trough from responding freely and fully to changes in the wavelength between it and the upstream Bering trough.

Other factors that could obscure the wind-wavelength relation are longitudinal variations of zonal wind speed from the zonal average, "anchoring" of the wave by a feature less than a wavelength upstream (such as the Alaska-Yukon ridge), and departure of the actual wave from the assumptions of small amplitude, sinusoidal shape, and stationarity. If the mean wave represents the average of moving, rather than stationary, Rossby waves, U is replaced by (U-c) in equations 5.1 and 5.2, where c is the zonal propagation speed of the wave (positive eastward). Equation 5.2 then becomes:

$$\frac{\partial L}{\partial U} = \frac{L}{2(U-c)} \qquad (5.3)$$

which results in smaller wavelength changes for negative c (retrograde waves). Long waves often do move westward (Palmén and Newton, 1969, Chapter Six); whether this is the explanation of the small observed wavelength changes is a question that calls for a separate study of daily wave patterns within a set of summer seasons.

From the general decrease of average streamline latitude with increased averaging time, one would expect the decadal trough shift to agree with the zonal wind change at 70°-75°N. However, while the trough moved 4.7° eastward between 1951-60 and 1964-73, the zonal wind at that latitude <u>decreased</u> by .05 m/sec (Fig. 5.3a). The trough shift does agree in sign and magnitude with the zonal wind increases at 65°-70°N and 75°-80°N, with 4.7° being close (within 1°) to the full wavelength changes for both latitude zones. The weaker zonal wind at 70°-75°N is due to the fact that decadal deepening of the Baffin trough was greatest at 65°-70°N (see Fig. 5.4), causing a weakening of the zonal flow north of 70°N near the longitude of Baffin Island. The relevant streamline passes south of 70°N, and therefore misses the region of weaker westerlies, even though these weaker westerlies go into the zonal average. In this case the zonally averaged zonal wind is not a good approximation to the streamline average wind speed. From Fig. 4.15 it can be seen that the average wind speed for the streamline passing through the Baffin trough at 65°-70°N increased by about 0.3 m/sec. Inserting this increase into the wave equation for 72.5°N gives a wavelength increase of 4.9°, close to the observed trough shift of 4.7°. As noted on decadal 500mb charts, the average position of the Bering trough remained unchanged. It

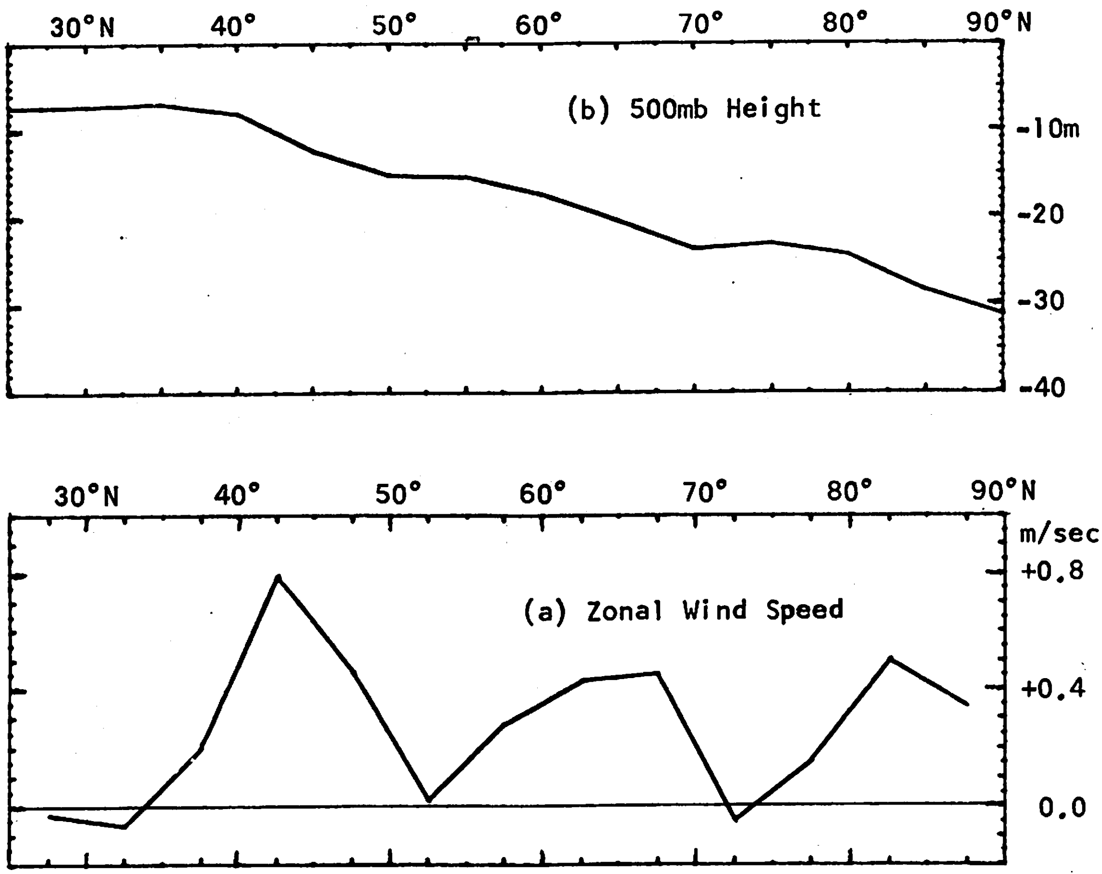


FIG. 5.3. Decadal change of zonally averaged 500mb height (top) and zonal wind speed (bottom), 1951-60 to 1964-73.

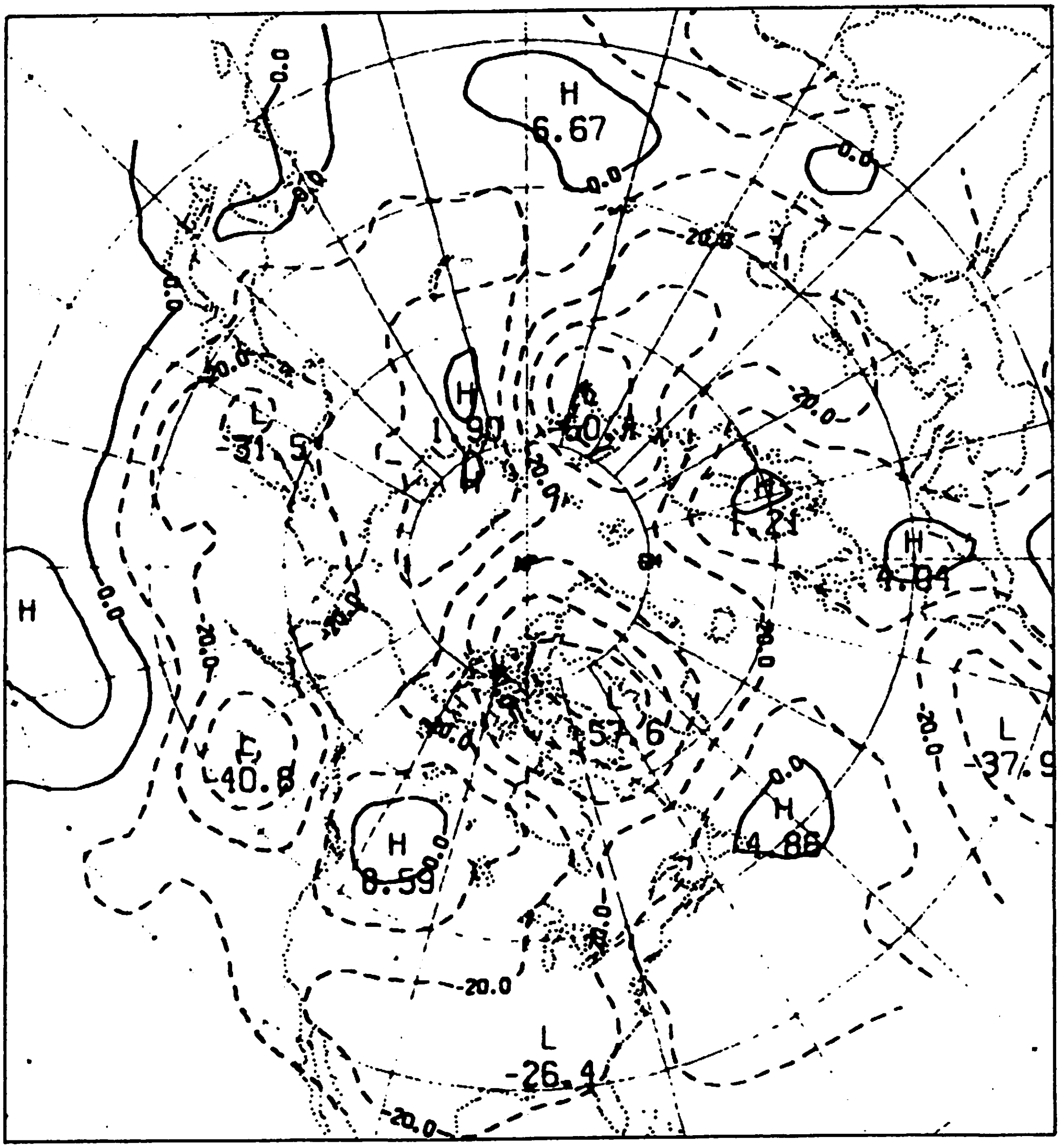


FIG. 5.4. Decadal change of summer 500mb heights, 1951-60 to 1964-73. Contour interval, 10 meters; dashed contours indicate lower heights in 1964-73.

therefore appears that the Baffin trough responds more fully to Rossby wavelength changes between it and the statistically fixed Bering trough on the decadal time scale than it does on shorter time scales; this may be due to averaging out of the moderating effects noted earlier or to the smaller variability of decadal mean trough longitudes (hence, less interference from Greenland).

Influence of the Northern Hemispheric Westerlies

How do the sub-polar zonal winds relate to changes in the zonal winds farther south? Table 5.1 presents the correlation matrix for the zonal winds in 5° latitude bands between 35° and 80°N. The winds north and south of this range are not included because they offer no significant correlations. The coefficients at lower left in Table 5.1 are for correlations of 1-2-1 smoothed values; those at the upper right are for the yearly departures from the smoothed running mean. This is done to separate the time scales. The smoothed values emphasize fluctuations with periods greater than 4 years, the annual values those with periods less than 4 years. Coefficients significant at the 95 percent level, except for the positive coefficients between adjacent latitudes, are marked with an (*).

The short-period (mostly summer-to-summer) wind fluctuations between 50° and 70°N show strong negative correlations with winds 15° farther south. This includes the zone of maximum westerlies (40°-50°N), and the correlations represent contractions and expansions of the polar vortex. However, its effects stop at 70°N, south of the zone of westerlies that govern the Baffin trough. This

TABLE 5.1. Correlations Between Zonal Winds at Different Latitudes:

Smoothed Values at Lower Left; Yearly Departures from Smoothed Values at Upper Right

Latitude	35	40	45	50	55	60	65	70	75	80
35										
			.79	.37	-.57*	-.76*	-.41*	-.05	-.18	-.18
40										
		.79		.74	-.31	-.83*	-.53*	-.20	-.26	-.18
45										
		.45	.70		.37	-.49*	-.73*	-.56*	-.22	.01
50										
		-.46*	-.49*	.22		.53	-.18	-.53*	-.10	.13
55										
		-.31	-.37	.15	.84		.64	.13	.06	.05
60										
		-.07	-.07	-.01	.24	.64		.73	.15	-.15
65										
		.04	.21	-.25	-.58*	-.27	.49		.58	.12
70										
		-.04	.11	-.38	-.67*	-.49*	.04	.79		.77
75										
		-.02	.01	-.39	-.54*	-.37	-.06	.58	.89	
80										

*95 percent confidence.

short-term component will therefore receive no further treatment.

The smoothed values show that the zonal winds between 65° and 80°N, which control the Baffin trough, are strongly anticorrelated with the 50°-55°N winds, which are in turn anticorrelated with the winds between 35° and 45°N. A similar latitudinal relation is seen in the decadal change of zonal wind, except that the 65°-80° band is narrowed to 65°-70°N by the effect of the deepened Baffin trough discussed earlier. Apparently, when the zone of westerlies as a whole strengthens, the winds between 50° and 55°N weaken, or strengthen less. The zonally averaged scalar wind speed at 50° and 55°N behaves the same way (a separate correlation matrix was computed for this parameter), so the effect cannot be caused by a stronger mean wave pattern at that latitude directing more of the flow into non-zonal directions. The wave patterns at 50°-55°N, as measured by the ratio of zonally averaged 500mb scalar wind to zonally averaged west wind, are no stronger during 1964-73 than during 1951-60, although the area-weighted average of this ratio over the hemisphere north of 25°N increases slightly from 1.12 to 1.15.

In Fig. 4.16a it can be seen that 50°-55°N is near the zone of maximum zonally averaged vorticity flux, and is the zone of maximum northward vorticity transport, as inferred from the latitudinal gradient of vorticity generation/dissipation (Fig. 4.16b). It was suggested in the discussion of vorticity flux that the small decadal increase in zonally averaged vorticity flux at 55°N (relative to other latitudes) may have been due to more meridionally moving transient eddies, which would result in rapidly (daily) fluctuating meridional components and reduced zonal components. The meridional components would tend not to be evident on the seasonal mean, due to cancelling of the daily fluctuations. There are more transient eddies during the later decade, and the evidence suggests that they are more meridional in their motions. This implies that the negative correlation between the 50^o-55^oN zonal wind and the winds at higher and lower latitudes has the same cause on the 1-2-1 smoothed time scale as proposed above for the decadal time scale.

The magnitude of the component of the zonal wind variability represented by the out-of-phase behavior of the winds at 50^o-55^oN can be indicated by a single zonal U-index, $U(40^o\text{-}45^o N)-U(50^o\text{-}55^o)+U(70^o\text{-}75^o N)$. This index is a measure of the intensity of meridional transient eddy activity at mid-latitudes, if this activity is actually the cause of the out-of-phase behavior of the 50^o-55^oN zonal winds. The value of this index for summer 1946-76 is plotted in Fig. 5.2b. The correlation between the 500mb trough longitude and this index ($r = -.43$) is not as good as it is with the 70^o-75^oN zonal winds alone ($r = -.74$), but use of the index provides a perspective on the relation between the trough and the nature of the general circulation. It is clear that simple expansions and contractions of the circumpolar vortex do not significantly affect the 500mb trough. The anticorrelation found earlier between 700mb sub-polar (55^o-70^oN) and subtropical (20^o-35^oN) zonal winds is also found at 500mb, but it is slight ($r = -.32$ between 1-2-1 smoothed values of zonal winds at 25^o-30^oN and 60^o-65^oN), and does not extend far enough north to affect the Baffin trough. It appears that the trough responds to increases in the total strength of the zonal circulation, where the zonal winds increase generally except in the band 45°-60°N, just north of the maximum westerlies. In this band the increase of zonal components may be suppressed or negated by the increased effect of transient eddies.

Any net strengthening of the westerlies would, of necessity, be associated with an increase in the subtropics to pole 500mb height gradients. Fig. 5.3b shows the change of zonally averaged 500mb height between 1951-60 and 1964-73; the 40° to 90°N height difference increased by 24 meters, or 6 percent. The zonally averaged 500mb height fell over the entire range of latitudes, and the geographic distribution of these height falls is shown in Fig. 5.4. The ubiquity of the height falls suggests another phenomenon concurrent with the strengthened gradients: a hemispheric cooling.

Effects of Hemispheric Average Tropospheric Temperature

The thickness of the 1000-500mb layer would decrease by 20 meters for a drop of 1°C in the average temperature of the layer. Thus, if the surface pressure is unchanged, a lower tropospheric cooling of 1°C for each 20 meters of 500mb height fall is implied. The surface pressure changes around Baffin between the decades (Fig. 3.6) are generally less than 1mb (or, in terms of the height of the 1000mb level, about 10 meters), and a map published by van Loon and Williams (1976) shows 1942-72 surface pressure trend generally less than 2mb over the Northern Hemisphere, with more areas showing pressure (and therefore 1000mb height) rises than falls. Thus, the general fall of 500mb heights between the decades appears due to widespread cooling, rather than to a mass evacuation of air from the Northern Hemisphere. This contention is further supported by the geographic agreement between the height falls and the surface temperature changes at 70°N (Figs. 5.4 and 2.11), and by observed declines in annual mean thickness during the same period (Dronia, 1974; Angell and Korshover, 1975, 1977, 1978; Yamamoto et al., 1975; Harley, 1978).

On the premise that 500mb height changes are primarily caused by lower tropospheric temperature changes (at the rate of 20 meters per 1°C), the data presented in Fig. 5.5 can be interpreted as showing the latitudinal differences in summer temperature changes between 1946 and 1976. The most pronounced features are the general cooling around 1954, after which the polar regions promptly rewarmed, followed by another general cooling in the early 1960's. This latter cooling is not simultaneous at all latitudes.

Area-weighted averages of the seasonal height departures from the long-term mean, calculated for the polar cap north of 70°N and for the hemisphere north of 25°N, are presented in Fig. 5.6. The polar cap heights correlate well ($r = .75$) with the observed zonally averaged surface temperature at 70°N (Fig. 2.8), lending further credence to the link between 500mb height and temperature. The polar and hemispheric averages are very closely correlated ($r = .88$), with the standard deviation being twice as great for the polar heights. The decadal drop in polar cap average height was also twice that of the hemispheric average (24 meters vs. 13 meters).

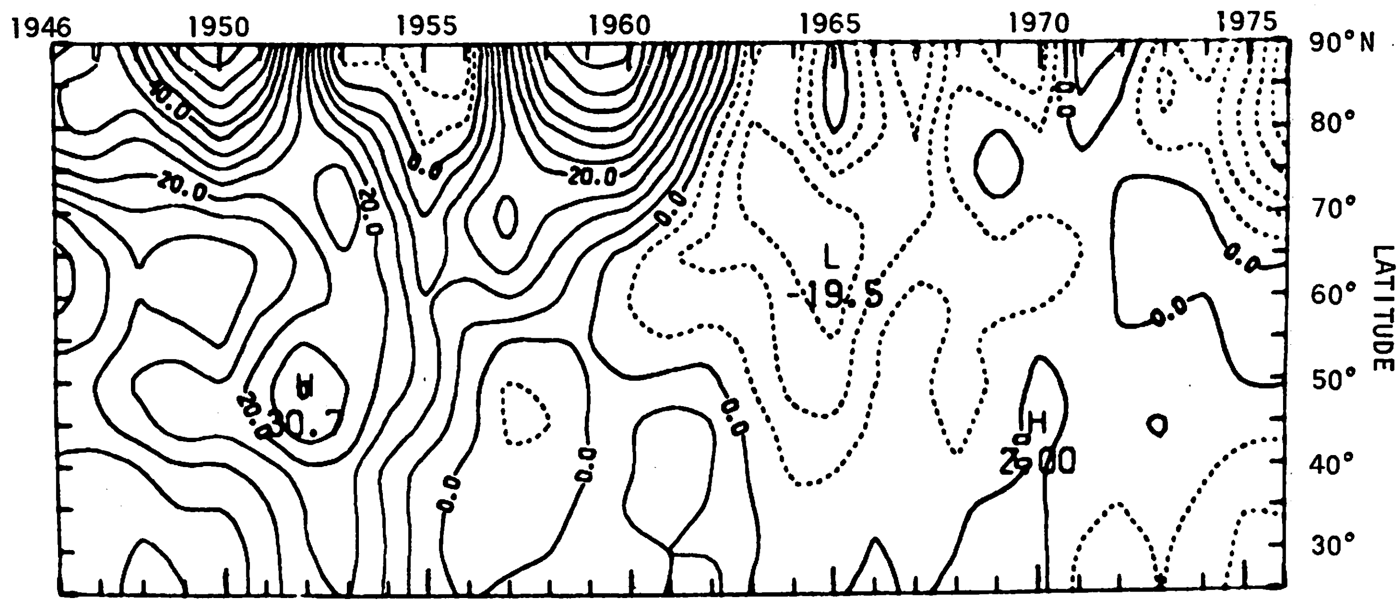


FIG. 5.5. Departure of zonally averaged 500mb height from 1954-76 average at each latitude, summer 1946-76, smoothed in time by a 1-2-1 running mean. Contour interval, 5.0 meters; dashed contours indicate negative values.

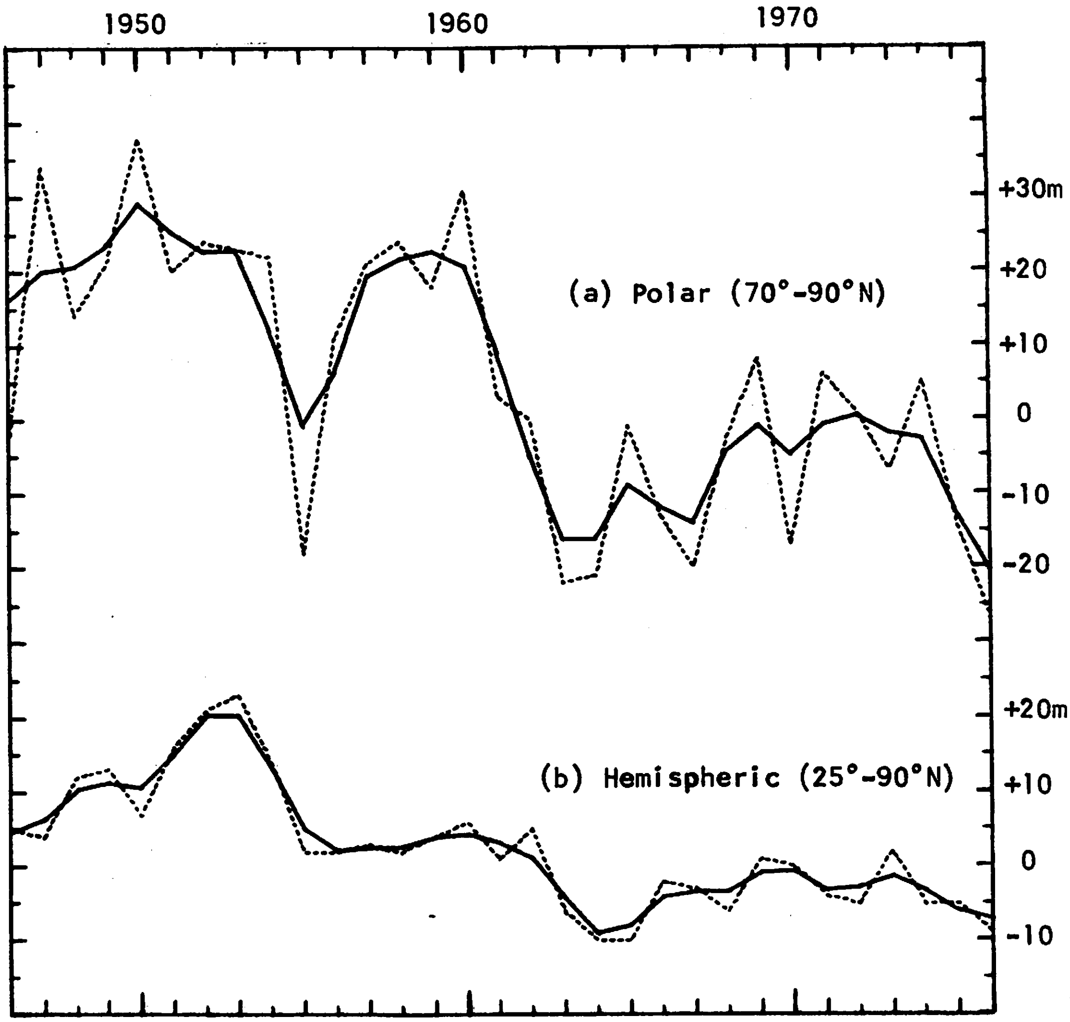


FIG. 5.6. Area-averaged 500mb height for the polar regions north of 70°N (top) and for the hemisphere north of 25°N (bottom), summer 1946-76: departures from 1954-76 mean. Annual values, dotted lines; smoothed values, solid lines.

Thus, the latitudinal height gradient (and strength of westerly winds) tends to increase as the hemisphere cools, at the rate of about a 10 percent increase per 1° of cooling. The 500mb zonal U-index defined earlier, and shown in Fig. 5.2b, correlates well with decreasing hemispheric average heights (r = .75). Therefore, increased meridional transient eddy activity just north of the zone of maximum winds appears to accompany the stronger polar vortex associated with the cooling. These eddies reduce the zonal wind increase at mid-latitudes, but not at sub-polar latitudes, and the Baffin trough shifts eastward.

Summary and Discussion

The Baffin trough is shown to respond to changes in the speed of the sub-polar westerlies, in approximate quantitative agreement with the Rossby wave theory. The trough shifts eastward with stronger sub-polar westerlies, which in turn are associated with general strengthenings of the hemispheric westerlies on longer time scales (periods greater than 4 years). On shorter time scales (summer to summer), latitudinal excursions of the belt of maximum westerlies (i.e., contractions/expansions of the polar vortex) are a major component of the Northern Hemispheric general circulation variability, but this process does not extend far enough north to significantly affect the Baffin trough.

General strengthening of the hemispheric westerlies is associated with hemispheric cooling of the lower troposphere, which, being greater in the Arctic, results in greater pole-to-subtropics gradients of 500mb height. A strengthened arctic front would be one manifestation of the enhanced latitudinal 500mb height gradients. Concurrent with strengthenings of the general circulation are widespread increases of storm activity (as measured by positive vorticity flux), with apparent rises in the meridionality of these transient eddies.

As noted by Oort and Rasmusson (1971) and van Loon (1979), most of the northward transport of heat in the extratropical atmosphere is accomplished by transient eddies. Mean eddies (standing waves) contribute little to the total transport in summer, especially at higher latitudes. The general hemispheric increase of vorticity flux, and particularly the increase of meridional vorticity flux implied by the increased latitudinal gradients of vorticity flux divergence (Fig. 4.16b), suggest that the meridional transfer of sensible heat also increased from 1951-60 to 1964-73. Data provided by van Loon (personal communication) show that the summer 700mb transient eddy meridional heat flux north of 40°N increased by about 20 percent between the two decades. Although small, the northward flux of heat by mean eddies at the surface also increased, showing a general upward trend from 1942 to 1972 (van Loon and Williams, 1976). During the cooling period 1958-63, the total northward energy transport across 60°N by all eddies increased by about 4 percent on both annual and summer bases (Oort, 1974).

One might expect that increased heat transports into the Arctic would be associated with higher temperatures, but the earth-atmosphere system seems to be quite effective at negating this effect. Monthly values of total energy (mostly sensible heat) flux across 60^{o}N have a standard deviation (seasonal trend removed) of $4x10^{14}$W (Oort, 1974), about one-seventh of the annual average monthly flux. This standard deviation corresponds to an anomalous heating (or cooling) of the arctic atmosphere of 0.1^{o}C/day, or 3^{o}C over the course of a month, assuming other factors remained constant. Of course, these other factors -- namely, net radiation loss to space and net heat flux from the surface -- do not remain constant, since the inferred 3^{o}C monthly anomaly is an order of magnitude greater than the observed monthly anomalies.

The radiative and surface heat flux processes act as an effective negative feedback to the variations in atmospheric heat transports into the Arctic on a monthly basis (Oort, 1974). On the longer time scales represented by the Arctic cooling of the 1960's, the increased heat transport appears to be the feedback mechanism, responding to increased latitudinal 500mb height gradients accompanying an enhanced arctic heat sink. The lower temperatures of the 1960's probably result from a change in the net radiation and surface heat flux balance of the Arctic.

In summer the surface heat fluxes are primarily those due to oceanic transport of heat into the Arctic and to loss of heat from the atmosphere to the polar seas as ice melt. In both cases the heat exchange rates between the surface and atmosphere for the region north of 60^{o}N are about $3x10^{14}$W (Palmén and Newton, 1969; Vowinkel and Orvig, 1970; VonderHaar and Oort, 1973; Oort, 1974), compared to a five-year (1958-62) increase in atmospheric heat transport across 60^{o}N of $2x10^{14}$W. The surface fluxes would therefore need to change by factors of nearly 100 percent to induce the observed increase in atmospheric transport.

The summer average net radiation over the Arctic north of 60°N is about 20×10^{14}W, and represents the difference between the absorbed solar radiation (about 60×10^{14}W) and the outgoing infrared radiation (about 80×10^{14}W). These values are approximate averages of values given by Houghton (1954), Palmén and Newton (1969), and Oort (1974). If changes in the vertical profile of infrared emissivities are neglected, the Stefan-Boltzmann radiation equation (radiative flux = σT^4) may be applied to the average arctic atmospheric temperature decrease of 0.5°C during 1958-63 to yield a decrease of 0.6×10^{14}W in outgoing infrared radiation. The amount of solar radiation absorbed by the earth-atmosphere system north of 60°N would therefore have dropped by about 2.6×10^{14}W, or 4 percent, from 1958 to 1963 to maintain a balance between the net radiation and the atmospheric heat transport, if surface heat flux changes are indeed small.

A four percent decrease in absorbed solar radiation could be due to a decline in solar output or to an increase of the arctic planetary albedo (due to more snow, ice, or cloud cover). A stratospheric layer of volcanic dust would diminish the amount of solar radiation absorbed by the surface and lower atmosphere, although its effects on the radiation balance of the stratosphere could moderate the expected surface cooling (Schneider and Mass, 1975). Several authors (Dronia, 1974; Angell and Korshover, 1975, 1977, 1978; Yamamoto _et al._, 1975; Bradley and England, 1977; Handler, 1979) have suggested that the cooling of the early 1960's was caused by volcanic dust injected into the stratosphere by the eruption of Mt. Agung in March, 1963. However, Figs. 2.8 and 5.6a show that the arctic cooling began in 1961 and was complete by summer 1963, and that the bulk of the hemispheric cooling (Fig. 5.6b) occurred between the summers of 1962 and 1963, in agreement with summer surface and tropospheric temperature data presented by Starr and Oort (1973) and Angell and Korshover (1978). The volcanic dust veil did not arrive at mid-northern latitudes until September or October, 1963 (Meinel and Meinel, 1963; Volz, 1964), and although dust from Agung and subsequent eruptions may be a plausible cause for the cool summers of 1964 and later, it is an unlikely agent for initiating the cooling and circulation changes of the early 1960's. A climate model incorporating the effects of volcanic dust and variable solar activity (Schneider and Mass, 1975) predicts variations of global surface temperature that are qualitatively similar to the variations of Northern Hemisphere summer 500mb height (Fig. 5.6b), but the problem of the precise timing of the major cooling events (1953-55 and 1962-63) remains. Namias (1969) attributes the "special climatic regime" of the 1960's to a hemispheric response to a self-perpetuating North Pacific pressure and sea surface temperature anomaly that formed in the fall of 1961. Although Namias' theory concerns only the circulation changes and not the hemispheric cooling, the possibility does exist that the cause of both lies wholly within the ocean-atmosphere system.

Although the question of the ultimate cause of the observed global circulation and temperature changes has not been answered, the joint occurrence of cooler arctic summers with stronger hemispheric meridional heat transports (implied from greater transient eddy activity) during the 1960's suggests that the circulation changes are a response to the temperature changes. Variations in the hemispheric and arctic radiation balances are possibly the major cause of the observed temperature changes, but their nature and origin -- solar, terrestrial, or atmospheric -- remains unresolved.

CHAPTER SIX

CONCLUSIONS

Throughout the course of this study the aim has been to answer one basic question: what is the nature of the connection, if any, between the interannual and longer-term variability of Baffin area summer temperatures and of the atmospheric circulation? In pursuit of the answer, the investigation expanded to ever larger scales of circulation, from synoptic scale systems to the regional circulation, and eventually to the general circulation of the extratropical Northern Hemisphere. The next step -- to examine the connections between the tropical and northern extratropical circulations -- is beyond the scope of this study. There can be no doubt that the tropics strongly influence the circulation at higher latitudes, since it is the angular momentum transferred from the earth to the atmosphere by the tropical Hadley circulation that drives the westerlies (Palmén and Newton, 1969, Chapter One). In an analysis of selected winters since 1854, Meehl (1978) notes that the Icelandic low tends to be stronger, and the Aleutian low weaker, when several tropical circulation indices indicate an intensified and expanded Hadley circulation. A similar investigation of the summer circulation would be an invaluable extension of the results found in this study.

Several different data sets and analytical techniques have been employed in this study; some of the techniques are tried and tested and others are refinements of existing methods. One approach is new. In Chapter Four the concept of 500mb vorticity flux as the upper reflection of surface cyclone activity is introduced and applied to an analysis of storm tracks. The vorticity flux approach removes much of the subjectivity inherent in previous storm and storm track climatologies, and certainly has applications beyond those included in this study.

Changes in synoptic surface pressure patterns are shown to be responsible for some of the daily temperature variability, but an attempt to explain the interannual summer temperature variability in terms of synoptic type frequencies is unsuccessful. However, treating the synoptic types as circulation features whose frequency of occurrence varies with average summer temperature (but not as a causal factor) is a useful approach to understanding the circulation changes that do control the summer climate of the Baffin area.

The results of the investigation may be summarized as a scenario describing the connection between Baffin area summer climate and the climate and general circulation of the extra-tropical Northern Hemisphere. The conditions described are those that result in a cool summer, or spell of summers, around Baffin Island. The scenario begins with a widespread cooling of the extratropical troposphere. The cooling is more than twice as great in the polar regions than over the subtropics, resulting in stronger than usual meridional temperature and 500mb height gradients. The stronger height gradients lead to stronger westerlies at most latitudes, including the sub-polar regions. The stronger sub-polar winds cause a lengthening of the standing waves in the westerlies, and the Baffin trough shifts eastward as the wavelength between it and the upstream Bering trough increases.

One manifestation of the stronger meridional temperature gradients is a stronger arctic front across Canada. More frequent, stronger (probably), and faster-moving storms form along this front in northwestern Canada and move along their usual track south of Baffin Island. There is a general increase in storminess elsewhere in the hemisphere, also because of enhanced temperature gradients. Upon approaching the southern tip of Greenland some (more than usual) of these storms are shunted northwards towards Baffin Bay by the steering effect of the stronger, eastward-displaced Baffin trough. The increased cyclonic activity over Baffin Bay helps maintain the strength and displacement of the trough.

At the same time, more northerly flow on the western flank of the trough over Baffin Island brings in cold air from the polar ice cap. The track of southbound polar anticyclones over north-central Canada is shifted eastward with the trough, and more of these systems pass over Baffin Island. The air masses affecting Baffin Island are from colder source regions, and low average summer temperatures result, along with reduced exotic pollen influx.

This scenario is found to be effective on time scales ranging from 2 to 10 years. No marked link between the general circulation and Baffin area temperatures is found on the shorter time scale represented by the differences between consecutive summers; this time scale is apparently subject to considerable climatologically random effects.

On the 2-year time scale the links are strong enough that the correlation between the hemispheric average temperature and Baffin area average temperature is quite good (r = .55). The correlation between the hemispheric temperature and one of the Baffin area environmental indicators, the date of total clearing of sea ice from Baffin Bay, is equally good (r = .56). The date of ice clearing is shown in this study to be closely correlated (r = .74) with the regional temperature. All three correlations are significant, and the point is made that the circulation links noted above are real enough to effect an end result (Baffin area climate) that is quite responsive to the initial input (hemispheric temperature).

Because the summer climate of the Baffin area does have a clear link to the hemispheric-scale climate and circulation, the following points can be made about the global significance of Baffin Island climate studies:

1) Since Baffin area summer temperatures are well correlated with those of the Arctic as a whole and those of the entire extra-tropical Northern Hemisphere, Baffin Island is a significant indicator of larger-scale summer climate conditions.

2) The regional atmospheric circulation near Baffin Island and its effect on regional temperatures are closely linked to variations in the general circulation of the Northern Hemisphere. Therefore, Baffin Island is also a significant indicator of the nature of the general circulation.

3) Ice and snow conditions on and near Baffin Island are very sensitive to small changes in summer temperature, enhancing

Baffin Island's usefulness as an indicator of past, present, and future climate and circulation variations.

4) These correlations are effective over time scales from 2 to 10 years, and do not weaken with increasing time scale within this range (some appear to strengthen). Therefore these correlations may very well extend to the time scales of interest to paleoclimatologists and glaciologists.

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APPENDIX

CATALOG OF BAFFIN ISLAND SYNOPTIC PRESSURE PATTERN TYPES, JANUARY 1, 1946 TO AUGUST 31, 1974

Zero entry indicates unclassified days.

YEAR	MO \ DAY	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	25	26	27	28	29	30	31
1946	1	1	21	1	1	8	15	10	1	19	6	1	1	1	1	1	1	21	1	1	1	3	12	8	5	5	10	1	1	1	1	1
	2	1	1	1	14	8	2	2	1	27	23	21	14	13	13	2	1	1	3	22	2	6	10	1	1	0	2	2	11			
	3	24	4	4	7	5	9	11	1	1	1	26	26	6	6	6	6	19	8	1	26	15	10	1	8	1	1	8	4	2	1	1
	4	3	9	5	1	1	3	0	6	1	3	0	4	5	20	1	6	1	5	3	18	27	4	4	2	2	1	12	2	6	1	
	5	1	10	1	1	1	5	0	4	2	10	2	2	23	5	15	20	1	1	21	5	12	2	10	8	2	2	6	6	6	6	10
	6	10	1	1	15	15	20	2	5	11	24	2	2	10	1	1	1	10	10	10	1	8	5	5	5	4	1	1	1	15	1	
	7	20	2	10	1	3	3	9	9	27	23	23	2	1	1	5	1	1	15	1	1	8	9	27	11	20	1	27	23	20	3	3
	8	27	1	27	4	7	25	2	2	10	20	1	11	7	4	11	2	1	1	9	20	3	3	3	3	3	5	1	10	10	1	23
	9	23	2	11	11	9	5	18	1	1	15	20	1	15	15	6	6	26	2	1	15	4	2	6	4	4	2	7	13	14	20	
	10	24	0	6	12	2	1	1	3	12	2	2	4	11	9	9	5	24	6	1	6	1	15	23	8	4	2	1	1	1	8	17
	11	1	8	8	2	1	6	2	6	2	1	5	5	11	1	8	23	2	5	5	21	26	4	7	7	4	4	11	2	1	1	
	12	1	8	6	1	1	8	1	1	8	6	6	6	2	2	6	1	5	20	1	3	15	20	6	1	2	2	1	1	1	3	1
1947	1	1	1	14	1	1	15	5	1	1	1	1	1	1	3	7	5	5	1	15	13	2	1	5	5	16	18	23	2	2	2	6
	2	6	6	22	2	2	4	17	10	2	2	2	10	1	3	1	6	4	4	7	7	2	2	2	2	2	2	23	10			
	3	6	17	4	23	2	10	10	1	1	1	1	1	1	1	1	1	2	2	5	6	2	4	4	20	1	2	2	2	22	6	1
	4	1	1	1	1	1	1	1	1	5	5	1	14	3	21	2	6	2	6	1	1	1	8	3	0	10	1	1	10	2	5	
	5	3	27	4	20	1	1	10	14	10	1	1	5	10	1	8	1	1	1	1	1	14	13	12	4	4	4	2	2	2	4	23
	6	2	10	14	3	0	8	15	15	5	5	11	10	10	18	28	6	1	3	3	3	27	5	15	15	12	4	4	4	4	2	
	7	1	0	24	4	11	11	2	5	1	23	20	3	[illegible]	20	3	15	1	5	2	5	11	11	5	18	14	18	3	9	5	9	18
	8	22	5	0	15	24	17	13	10	26	22	18	5	5	1	1	1	3	20	19	21	1	21	1	1	13	1	3	11	6	15	2
	9	6	13	20	1	1	26	10	1	15	12	13	6	12	23	2	2	1	3	3	27	11	5	9	1	1	3	27	5	5	5	
	10	9	3	26	26	4	2	1	1	3	3	27	13	7	24	11	6	6	12	4	7	10	1	1	1	1	27	4	20	6	8	8
	11	5	1	4	1	1	8	8	4	4	4	4	17	4	21	24	23	17	6	15	1	8	22	7	2	4	2	11	5	11	2	
	12	2	21	21	6	6	10	2	6	6	6	6	6	6	5	6	6	23	10	6	6	6	1	14	8	13	15	1	9	2	6	14
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YEAR	MO \ DAY	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	25	26	27	28	29	30	31
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