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SPATIAL AND TEMPORAL CHANGES IN EXTREME AIR TEMPERATURES IN THE ARCTIC OVER THE PERIOD 1951–1990

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ABSTRACT

A detailed analysis of the spatial and temporal changes in mean seasonal and annual daily maximum (T_{max}) and minimum (T_{min}) air temperatures and diurnal temperature range (DTR) in the Arctic over the period 1951–1990 is presented. This analysis is preceded by a description of the spatial distributions of the mean seasonal and annual 40-year extreme temperatures (i.e. T_{max} and T_{min}).

The rate of decrease of the mean Arctic T_{min} is about twice as weak as the rate for T_{max} in the period 1951–1990. As a result, a decrease in DTR is observed. Not all areas of the Arctic, however, show such tendency, e.g. large parts of the Canadian Arctic do not. The increases in DTR here are more common in summer than in winter.

The decrease in DTR is related partly to increases in cloud cover, especially in the warm half-year when solar radiation is present in the Arctic. On the contrary, in the cool half-year (mainly during polar night) the day-to-day changes of temperature, governed at this time by very variable atmospheric circulation, have a greater impact than the cloudiness.

The increase in variability of T_{max} and T_{min} has not occurred in the most recent decades.

No evidence of any greenhouse warming in the Arctic over the period 1951–1990 is seen. Most of the T_{max} and T_{min} trends are not statistically significant. © 1997 by the Royal Meteorological Society, *Int. J. Climatol.* 17: 615–634, 1997

(No. of Figs: 12. No. of Tables: 5. No. of Refs: 48)

KEY WORDS: Arctic; spatial and temporal changes; time series analysis; temperatures maximum and minimum; diurnal temperature range.

INTRODUCTION

Recently it has been noticed that the phenomenon of global warming is strongly connected with an observed decrease of the mean monthly diurnal temperature range (DTR) defined as the difference between the mean monthly maximum (T_{max}) and minimum (T_{min}) temperatures. Such behaviour of the DTR is caused by the asymmetric trends of monthly mean T_{max} and T_{min}. Karl *et al*. (1991, 1993) showed that over 50 per cent of the Northern Hemisphere and 37 per cent of the global landmass, the increase of T_{min} is three times greater than that of T_{max}. It means also that global warming is caused mainly by the increase of T_{min}. In the last few years, many papers have been devoted to this subject, e.g. Frich (1992), Kukla and Karl (1993), Böhm and Auer (1994), Brázdil *et al*. (1994, 1995), Niedźwiedź and Ustrnul (1994), Jones (1995a), Karl *et al*. (1994, 1995), Przybylak (1995), including several papers presented at the 'Minimax' Workshop, held 27–30 September 1993, in Maryland and published recently in *Atmospheric Research* (e.g. Dessens and Bücher, 1995; Horton, 1995; Jones, 1995b; Kaas and Frich, 1995; Parker, 1995). These papers cite evidence that some regions of the world reveal no significant trends in DTR (e.g. Austria (Böhm and Auer, 1994); Czech Republic (Brázdil *et al*., 1994); Nordic countries (Kaas and Frich, 1995); some parts of the Arctic, mainly Canadian Arctic (Przybylak, 1995) and Antarctica (Jones, 1995a)) and some even show a significant increase in DTR (Poland (Niedźwiedź and Ustrnul, 1994); North Sea region, including the British Isles (Horton, 1995)). One can see that not all areas of the globe show negative trends of the DTR.

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Since late in the nineteenth century it has been known that the polar regions play a very important role in shaping the global climate. Both observations and modelling studies have shown also that these regions are most sensitive to climatic changes. As a consequence, warming and cooling epochs should be seen here most clearly (e.g. Polar Group, 1980; Jäger and Kellogg, 1983). Climatic models with enhanced greenhouse forcing simulate the greatest warming in the polar regions, which should be especially high near the surface and in the winter half of the year (Houghton *et al*., 1990, 1992). However, the recent warming common in most parts of the world is not present in the Arctic, as defined in *Atlas Arktiki* (1985) (see e.g. Hanssen-Bauer *et al*., 1990; Nordli, 1990; Chapman and Walsh, 1993; Kahl *et al*., 1993a,b; Przybylak and Usowicz, 1994; Karl *et al*., 1995; Weber, 1995; Przybylak, 1996). Similar situations are also evident over Iceland (Einarsson, 1991) and over Antarctica, but only since the early 1970s (Jones, 1995a), i.e. since the last phase of global warming started. This divergence between global and polar air temperature patterns is one of the most intriguing issues for a climatologist to solve. Some propositions explaining this phenomenon are given in Przybylak (1996) concerning the climate of the Arctic. In this study I focus on the behaviour of the extreme temperatures and DTR in the Arctic over the period 1951–1990.

There are two main aims of the present paper. The first is to describe the mean state and the spatial and temporal variations of the T_{max} and T_{min} in the Arctic in recent decades and the second is to check whether there is (or is not) evidence for a decreasing trend in the DTR.

The analysis presented covers the greater part of the Arctic. There are no long-term data for the inner part of Greenland and the Arctic Ocean. Because the temporal and spatial variability of air temperature over the Arctic Ocean is one of the smallest in the Arctic region, the author decided to present for this area the likely results of the variables analysed (based on data from neighbouring stations). For Greenland, such analysis would not be feasible owing to the great topographical differences occurring there.

DATA AND METHODS

Mean monthly T_{max} and T_{min} from 26 Arctic and two sub-Arctic stations (Figure 1) were available for analysis during the period 1951–1990 (except Greenland, which had data up to 1980; for the decade 1981–1990 these data exist but their quality is poor and therefore they are at present being homogenized by P. Frich from the Danish Meteorological Institute). All the data come from national Meteorological Institutes (Danish Meteorological Institute, Norwegian Meteorological Institute and Canadian Climate Centre) or other institutions (Arctic and Antarctic Research Institute at St Petersburg and National Climatic Data Center at Asheville).

The use of spatial statistical methods (e.g. as proposed by Mitchell, 1961; Alexandersson, 1986; Vincent, 1990; Gullett *et al*., 1991) to detect inhomogeneities in the Arctic air temperature series is most often impossible owing to the great distances between meteorological stations (which, among other factors, are responsible for the weak correlation of T_{max} and T_{min} between neighbouring stations) and the lack of so-called reference stations (see e.g. Vincent, 1990; Gullett *et al*., 1991). Also these methods do not allow the detection of severe errors when all stations in a study region change instruments, formulae or observation time within a few years (Frich, 1993). The quality control of the T_{max} and T_{min} is much more difficult in comparison with mean temperature from ordinary thermometers because they are more sensitive to both local conditions and ‘artificial’ changes at or near the observation site. It is also very well known that T_{min} is more sensitive than T_{max} in this respect and therefore its series may contain more inhomogeneities. It is worth noting that a major source of inhomogeneity connected with urbanization is not present in most Arctic regions.

For this study the quality control of the extreme air temperature series analysed was performed using mainly the procedures outlined by Horton (1995), which are described herein in detail. Additionally, another very simple but good method, proposed by Frich (1993) and successfully applied by Przybylak (1996), was used. This method relies on the comparison of monthly mean T_{max} and T_{min} with appropriate true monthly mean temperatures. It uses the following very well known formula:

$$\frac{T_{max} + T_{min}}{2} \approx T$$

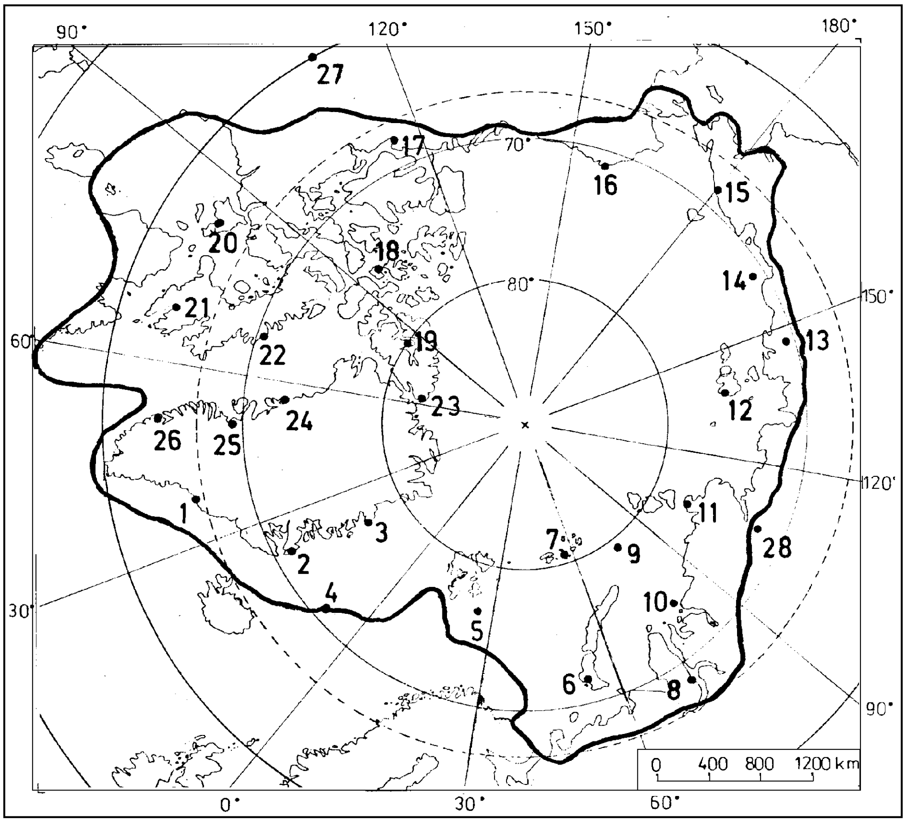

Figure 1. Location of meteorological stations used. The thick line is the border of the Arctic after *Atlas Arktiki* (1985). 1, Angmagssalik (height above sea-level ($H = 35$ m); 2, Kap Tobin ($H = 41$ m); 3, Danmarkshavn ($H = 11$ m); 4, Jan Mayen ($H = 10$ m); 5, Hopen ($H = 6$ m); 6, Malye Karmakuly ($H = 46$ m); 7, Polar GMO E. T. Krenkelya ($H = 20$ m); 8, Mys Kamenny ($H = 7$ m); 9, Ostrov Vize ($H = 18$ m); 10, Ostrov Dikson ($H = 20$ m); 11, GMO E. K. Fedorova ($H = 13$ m); 12, Ostrov Kotelny ($H = 10$ m); 13, Cokurdah ($H = 48$ m); 14, Ostrov Chetyrekhstolbovoy ($H = 6$ m); 15, Mys Szmidta ($H = 7$ m); 16, Barrow ($H = 4$ m); 17, Coppermine ($H = 24$ m); 18, Resolute A ($H = 67$ m); 19, Eureka ($H = 10$ m); 20; Coral Harbour A ($H = 64$ m); 21, Iqaluit A ($H = 34$ m); 22, Clyde A ($H = 25$ m); 23, Alert ($H = 63$ m); 24, Upernavik ($H = 63$ m); 25, Jakobshavn ($H = 47$ m); 26, Godthab ($H = 20$ m); 27, Forth Smith A ($H = 203$ m); 28; Khatanga ($H = 24$ m)

where T is the true monthly mean temperature computed from the fixed hours (usually 4, 8 or 24 measurements a day).

Having these three temperature series it is possible to easily find and correct erroneous values. This method was also used to fill up a few existing gaps in the series analysed when two temperature characteristics were known (T_{max} and T or T_{min} and T). For the Arctic this method is better than that commonly used—a comparison with the data from neighbouring stations. Unfortunately, the application of this method is possible only for the areas where the mean temperatures are not calculated as $(T_{max} + T_{min})/2$. Therefore, in the Arctic we can apply it to all areas except the Canadian Arctic and Alaska.

The climatic background of the T_{max} and T_{min} in the Arctic is presented using 40-year (1951–1990) mean seasonal and annual values. For each decade the anomalies relative to the average for 1951–1990 were also computed. The linear trends of annual and seasonal T_{max} and T_{min} were calculated for each Arctic station. Additionally, the significance of these trends using Student's *t*-test was estimated. Also the share of trends in the general variance of analysed temperatures has been computed. All analysed characteristics of the extreme air temperatures in the Arctic are presented in maps. The isolines have been drawn using simple mathematical interpolation.

Some investigators are convinced that for humankind the rising frequency of extreme phenomena is more important than small changes of mean values, both connected with the greenhouse effect (see e.g. Katz and

Brown, 1992). In order to check the behaviour of T_{max} and T_{min} in this respect the standard deviations in running decades of the period 1951–1990 have been computed.

RESULTS AND DISCUSSION

Spatial and season patterns of mean T_{max} and T_{min}

The general spatial pattern of the long-term mean air temperature in the Arctic is not completely known and is based on rather old data (see e.g. Prik, 1959; Vowinckel and Orvig, 1970; CIA, 1978; Atlas Arktiki, 1985; Herman, 1986). Moreover, in these publications usually only maps with January and July mean temperatures are presented. Therefore, based on 1951–1990 means, I have made maps also for each season (DJF, MAM, JJA, and SON) and year (Przybylak, 1996). Reviewing the Arctic climatic literature I have not found any publication that presents the spatial pattern of extreme temperatures for the whole Arctic. It is certainly worthwhile to fill this gap. Rough sketches of the spatial patterns of the mean T_{max} and T_{min} (based mainly upon the period 1951–1990) are presented in Figures 2 and 3. It is seen that the thermal parameters analysed have very similar spatial patterns. The annual mean T_{min} values usually are about 5·0–7·0°C lower than T_{max} (Table I and Figure 2). The coldest part of the Arctic (excluding Greenland) is the north-eastern part of the Canadian Arctic. For example, at the Eureka station the 40-year annual means of T_{max} and T_{min} were equal to −16·4 and −23·0°C, respectively (Table I). The warmest parts of the Arctic are the southern areas adjoining the Atlantic Ocean, where mean annual T_{max} values are even positive. The spatial distributions of mean winter T_{max} and T_{min} in the Arctic are similar to the annual values, but the horizontal temperature gradients are much greater (Figure 3, upper panels). Also greater is the mean difference between T_{max} and T_{min}, which over most of the Arctic is near to 8°C.

In summer (Figure 3, lower panels) the spatial patterns of both T_{max} and T_{min} are different than in winter. The coldest part at this time is the central Arctic, especially on the Atlantic Ocean side. The 40-year mean values of T_{min} vary between −2° and 0°C and T_{max} between 0° and 2°C. Low temperatures over the Arctic Ocean are mainly maintained by the presence of melting snow and ice, which absorbs the majority of warmth arriving here. In summer the warmest temperatures are noted in the southern parts of the continental Arctic (i.e. mainly in the Russian and Canadian Arctic), where 40-year means of T_{max} and T_{min} exceed 10° and 4°C, respectively (Figure 3, lower panels).

The variability of the yearly mean T_{max} and T_{min} is greatest in the region between Spitsbergen, Franz Josef Land and Novaya Zemlya (standard deviation (σ) is greater than 1·3°C) and smallest for the greater part of central and eastern Russian Arctic, the north of Canadian Arctic, southern Greenland and probably the Arctic Ocean ($\sigma \leqslant 1{\cdot}0$°C) (Table I). The variability is usually two to three times greater in winter than in summer. The main reason of the high variability of T_{max} and T_{min} in winter, when the incoming solar radiation is lowest, is the very strong and variable atmospheric circulation bringing thermally differentiated air masses (see Przybylak, 1992). In summer a more important factor than atmospheric circulation is the insolation. Therefore, the greatest variability of both thermal parameters analysed occurs in the southern continental parts of the Arctic ($\sigma > 1{\cdot}2$°C for T_{max} and $\geqslant 1{\cdot}0$°C for T_{min}) (Table I). In this season the greater variability of T_{max} than T_{min} in the Arctic is clearly seen.

According to Przybylak (1996), in the Arctic in the period studied, the warmest decade was 1951–1960, and the coldest was the following decade. As we know for the whole globe the warmest decade was 1981–1990 (Jones, 1994). Since the mid-1970s, when a rapid change in atmospheric circulation occurred, a discrepancy is observed in the courses of global and Arctic temperatures (Przybylak, 1996). It is for this reason that I have decided to focus on the behaviour of T_{max} and T_{min} in the last decade. Spatial distributions of annual and seasonal mean 10-year anomalies of T_{max} and T_{min} in respect of the average for 1951–1990 are presented in Figures 4 and 5. As is clearly seen, over the greater part of the Arctic, positive annual anomalies of extreme temperatures prevailed. They were lower than normal mainly over the eastern part of the Canadian Arctic and Baffin Sea. An additional area of negative anomalies in the case of T_{max} occurred also over the greater part of the Atlantic region of the Arctic (Figure 4, upper panel). On the contrary, lower than normal T_{min} in this decade was observed also over some small parts of the Russian Arctic and over an area spreading from the North Pole to Greenland and Franz Joseph Land (Figure 4, lower panel). Figure 5 is the same as Figure 4 but corresponds to winter and summer. In winter the spatial distributions of the T_{max} and T_{min} anomalies are quite similar to those for the year,

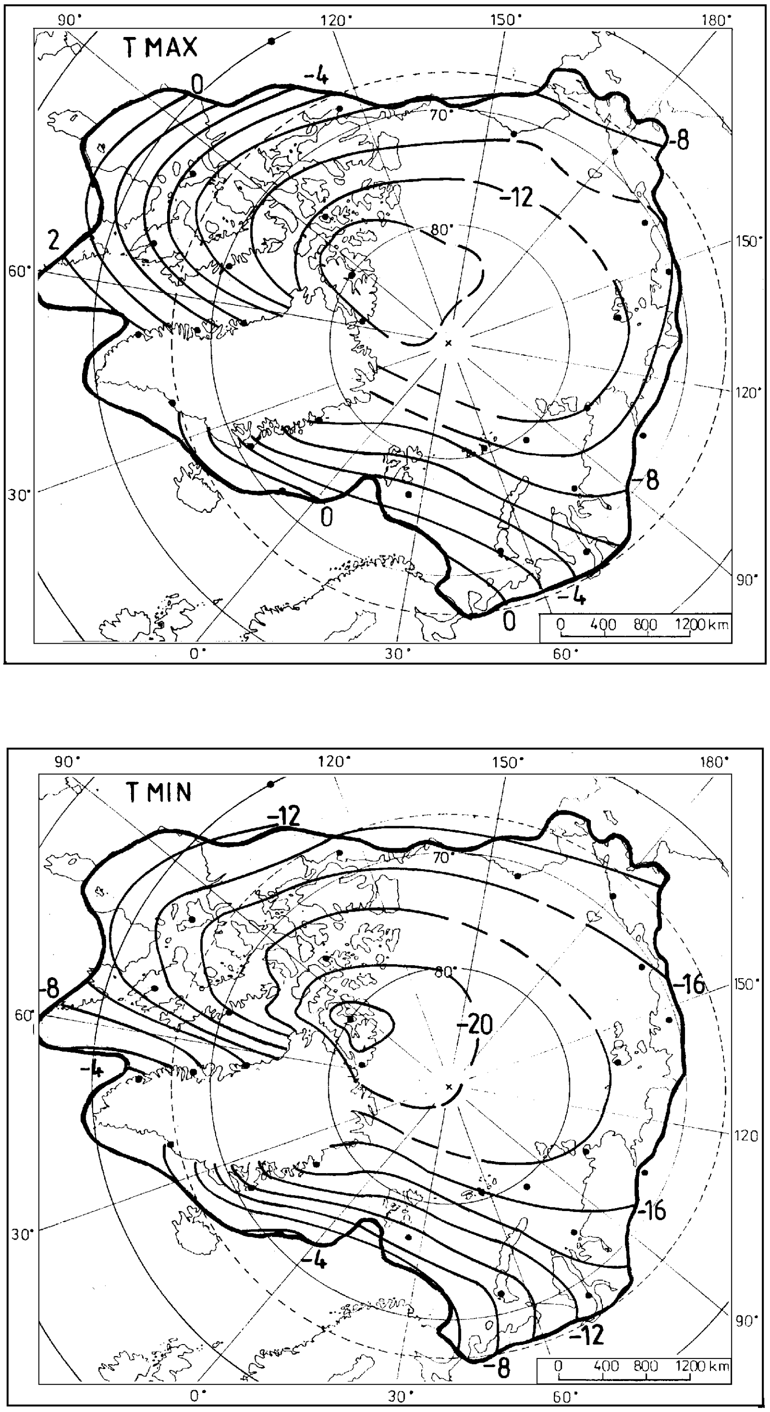

Figure 2. Spatial distribution of mean annual T_{max} and T_{min} in the Arctic (°C), 1951–1990. Note that the isotherms around the southern part of Greenland are not drawn owing to lack of data. Dots, meteorological stations, dashed lines, probable course of the isotherms

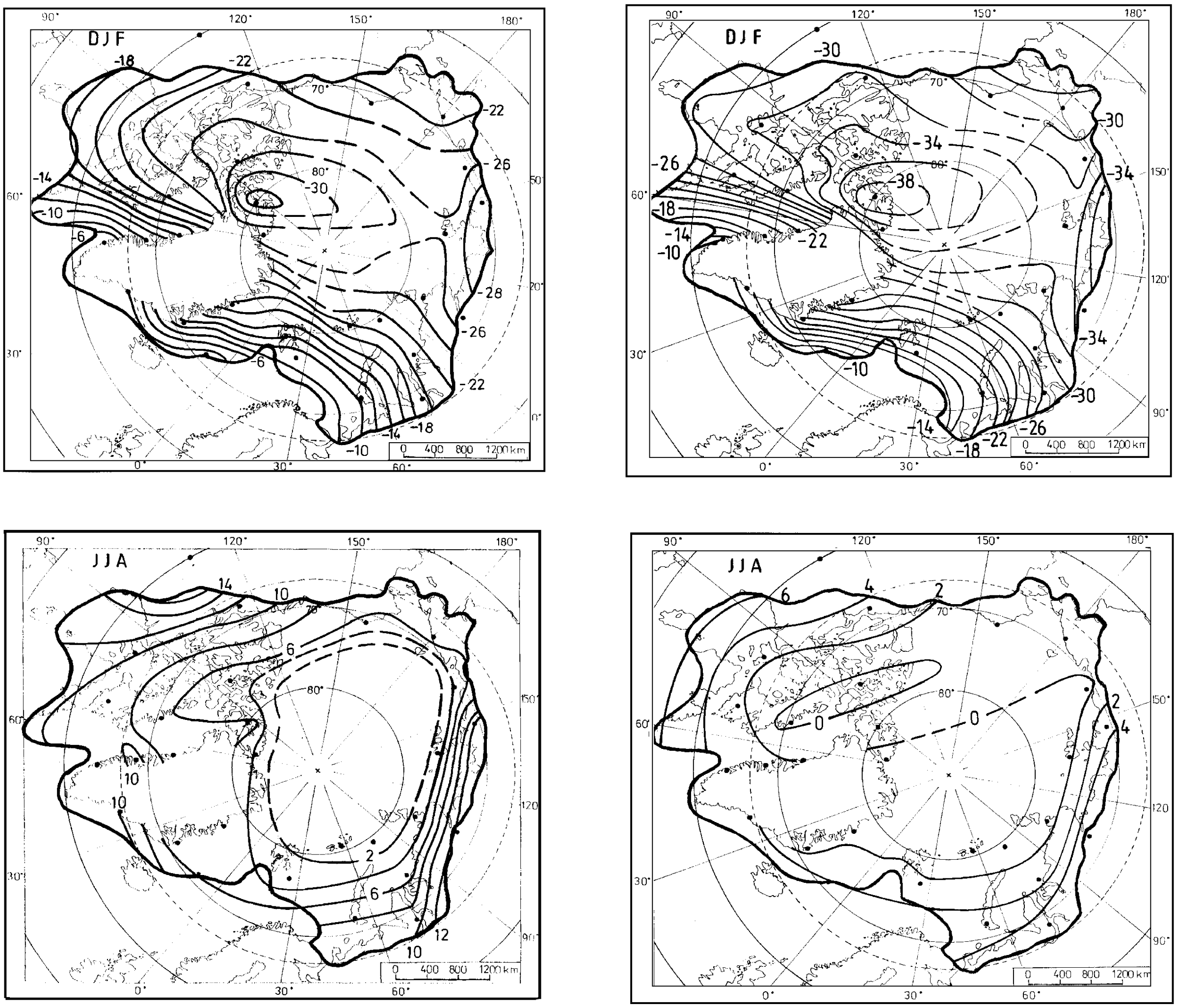

Figure 3. Same as Figure 2, but for winter (DJF) and summer (JJA): T_{max}, left panels; T_{min}, right panels

especially in the case of T_{max} (see Figures 4 and 5, upper panels). The extreme temperatures below normal occurred in the parts of the Arctic adjoining the Atlantic Ocean. In summer the spatial patterns of the anomalies of T_{max} and T_{min} are different in comparison with the annual ones (Figures 4 and 5). Both thermal characteristics show negative anomalies, mainly in the Russian Arctic. A below normal T_{max} value was present also around Greenland. In this season the greater increase in T_{min} than T_{max} is very clearly seen (Figure 5, lower panel).

Observed trends

T_{max} and T_{min}. An analysis of the trends of mean annual T_{max} and T_{min} (Figure 6) over the periods 1951–1990, 1961–1990, and 1971–1990 (not shown) revealed that the initially negative trends occurring in the greater part of the Arctic (1951–1990) were later replaced by positive ones (1961–1990, 1971–1990). In the period 1951–1990 the increasing trends of T_{max} and T_{min} were noted only over the southern part of the Russian Arctic (Figure 6). Both the thermal parameters show the greatest warming over Alaska and its vicinity (more than 0·2°C per decade). On the other hand, the greatest cooling was observed over the Atlantic region of the Arctic (especially in

Table I. Mean seasonal (DJF and JJA) and annual T_{max}, T_{min} and DTR (m, in °C) and their standard deviations (σ, in °C) in the Arctic over the period 1951–1990

Station	T_{max}						T_{min}						DTR					
	DJF		JJA		Annual		DJF		JJA		Annual		DJF		JJA		Annual	
	m	σ	m	σ	m	σ	m	σ	m	σ	m	σ	m	σ	m	σ	m	σ
Angmagssalik[a]	−3·7	1·7	10·0	0·8	2·5	0·8	−10·7	1·8	2·5	0·6	−4·5	0·9	7·0	1·0	7·5	0·7	6·9	0·6
Kap Tobin[a]	−12·1	2·1	4·9	0·6	−4·4	1·0	−20·3	2·8	−0·5	0·6	−11·2	1·2	8·3	1·9	5·4	0·7	6·8	1·0
Danmarkshavn[b]	−18·5	1·9	5·1	0·7	−8·7	0·9	−27·1	1·8	−0·3	0·6	−15·7	0·8	8·6	1·4	5·4	0·7	7·0	0·9
Jan Mayen	−2·6	1·6	5·9	0·7	1·1	0·9	−7·8	2·2	2·2	0·7	−3·2	1·1	5·2	0·8	3·7	0·4	4·3	0·4
Hopen	−9·2	3·1	3·1	0·8	−3·5	1·7	−15·9	3·8	0·1	0·8	−8·3	2·0	6·7	1·1	3·1	0·3	4·8	0·5
Malye Karmakuly	−10·4	2·7	8·1	1·4	−2·3	1·3	−17·4	3·0	3·1	1·0	−8·0	1·3	7·0	0·8	4·9	0·6	5·7	0·3
Polar GMO E. T. Krenkelya[c]	−21·0	2·9	1·3	0·5	−10·9	1·3	−28·3	2·6	−1·5	0·4	−16·3	1·3	7·4	0·8	2·8	0·2	5·4	0·3
Mys Kamenny	−19·9	3·1	9·9	1·5	−5·7	1·3	−28·2	2·8	4·4	1·2	−12·7	1·3	8·3	1·0	5·5	0·6	7·0	0·4
Ostrov Vize	−21·8	3·1	0·9	0·4	−11·5	1·4	−29·1	2·9	−1·7	0·4	−16·9	1·4	7·3	0·9	2·6	0·3	5·4	0·4
Ostrov Dikson	−21·5	2·9	5·7	1·4	−8·8	1·3	−28·8	2·8	1·1	1·1	−14·8	1·2	7·3	0·7	4·6	0·6	6·0	0·3
GMO E. K. Fedorova	−24·8	2·6	2·3	0·8	−12·4	1·1	−31·5	2·3	−1·4	0·5	−17·9	1·1	6·7	0·7	3·7	0·5	5·5	0·4
Ostrov Kotelny	−26·2	1·6	3·6	1·3	−12·2	1·0	−32·8	1·5	−0·6	0·9	−17·9	0·9	6·6	0·5	4·2	0·6	5·7	0·4
Cokurdah	−29·9	1·5	12·6	1·5	−10·1	0·9	−37·0	1·4	3·7	1·1	−17·8	0·9	7·1	0·5	9·0	0·7	7·7	0·3
Ostrov Chetyrekhstolbovoy	−25·1	1·7	4·2	1·1	−10·9	0·8	−31·3	1·5	−0·4	0·8	−16·2	0·8	6·2	0·7	4·6	0·6	5·4	0·3
Mys Szmidta	−21·4	2·1	6·5	1·3	−8·4	1·0	−28·7	1·7	0·7	0·7	−14·9	0·9	7·3	0·9	5·8	0·7	6·5	0·3
Barrow	−22·6	2·4	5·5	1·4	−9·6	1·2	−29·3	2·1	0·1	0·9	−15·6	1·1	6·7	0·9	5·4	0·6	6·0	0·4
Coppermine	−24·5	2·2	11·4	1·6	−7·6	1·1	−32·3	1·9	3·5	1·2	−15·3	1·0	7·9	0·8	7·9	0·9	7·8	0·5
Resolute A	−27·8	2·0	4·5	1·2	−13·4	1·0	−35·0	1·7	−0·5	1·0	−19·7	0·9	7·2	0·6	4·9	0·5	6·3	0·3
Eureka	−32·9	1·9	6·2	1·1	−16·4	1·1	−40·0	1·6	0·9	0·8	−23·0	1·0	7·0	0·7	5·3	0·6	6·6	0·4
Coral Harbour A	−24·0	2·6	10·1	1·4	−7·4	1·2	−32·3	2·4	2·1	1·0	−15·7	1·2	8·4	0·7	8·0	0·7	8·3	0·3
Iqaluit A	−20·5	3·5	9·5	1·1	−5·6	1·3	−28·9	3·3	2·5	0·7	−13·2	1·4	8·4	0·7	7·0	0·5	7·6	0·3
Clyde A	−22·8	2·8	6·3	1·2	−8·7	1·1	−30·5	2·7	−0·4	0·8	−16·1	1·1	7·6	1·0	6·7	0·7	7·4	0·6
Alert	−27·9	1·4	3·8	1·0	−14·7	0·9	−35·6	1·1	−1·4	0·6	−21·5	0·7	7·7	0·8	5·2	0·7	6·7	0·4
Upernavik[a]	−13·6	2·8	6·9	1·3	−4·3	1·0	−19·3	2·4	1·9	1·0	−9·6	0·9	5·8	0·9	4·9	0·7	5·2	0·5
Jakobshavn[a]	−8·8	2·9	10·3	1·0	−0·4	1·1	−16·6	3·1	3·3	0·7	−7·7	1·2	7·8	0·9	7·0	0·5	7·3	0·5
Godthab[a]	−3·7	2·1	9·3	1·1	2·0	0·9	−9·4	2·0	3·2	0·7	−3·4	0·7	5·7	0·7	6·1	0·6	5·4	0·4

[a] Data for 1951–1980.
[b] Data for 1955–1990.
[c] Data for 1958–1990.

the case of T_{max}), Greenland, Baffin Sea and the eastern part of the Canadian Arctic. As can be seen from Figure 6, T_{max} shows more pronounced cooling than T_{min}. In the annual values, negative trends of T_{max} occurred in 80 per cent of the stations versus only 52 per cent in the case of T_{min}. The cooling for both parameters was greater in the second part of the year. These results indicate different behaviour of T_{max} and T_{min} in the Arctic in comparison with most other regions of the globe (Karl *et al.*, 1991, 1993), which have mainly positive trends. As shown in Przybylak (1996) and partly in Figure 7, such a situation is explained by the fact that in the Arctic the warming that occurred in the period 1920–1960 was much greater than in other parts of Northern Hemisphere. Moreover, the magnitude of this warming was not exceeded here in the 1980s as it was for the Northern Hemisphere and the globe as a whole.

The trends of mean summer T_{max} and T_{min} are more similar to the annual ones than the trends of winter (Figures 6 and 8). In summer in the period for 1951–1990, negative trends of the extreme temperatures prevailed, while in winter positive trends prevailed. The difference in magnitudes of trends is much greater in winter, ranging from −0·6 to 0·6°C per decade. It is worth noting that spatial distributions and magnitudes of trends are very similar for both thermal parameters analysed (see Figures 6 and 8).

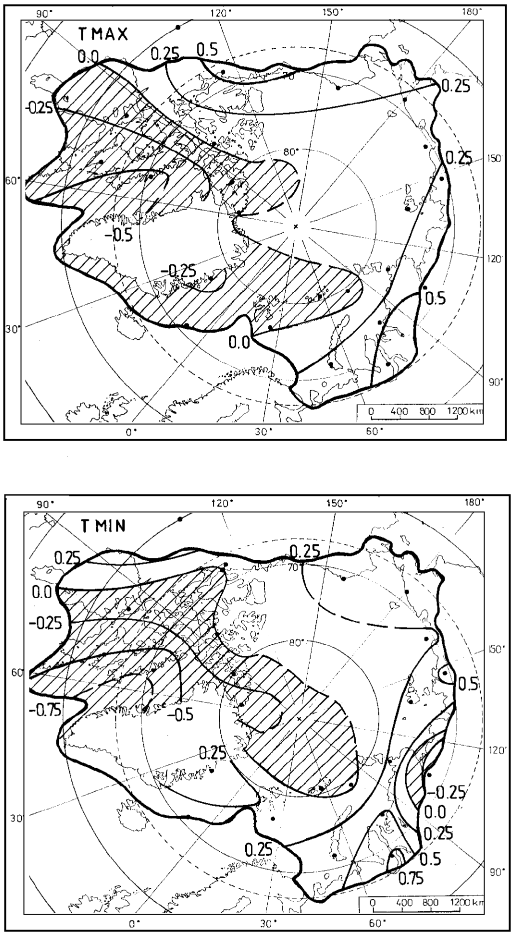

Figure 4. Spatial distributions of the annual anomalies of T_{max} and T_{min} for the decade 1981–1990, with the 1951–1990 mean for the Arctic (°C). Negative anomalies are hatched. Note also that the isopleths around the southern part of Greenland are not drawn owing to lack of data for the period 1981-1990, but the probable sign of the anomalies in this area (positive or negative) was estimated taking into account all the information available from neighbouring stations as well as the behaviour of the mean air temperatures (Przybylak, 1996). Dots, meteorological stations; dashed lines, probable course of the isopleths

The pattern of distribution of trends in the Arctic changed for the period 1961–1990. During this time a considerable domination of positive trends of T_{max} and T_{min} is seen, but cooling still occurs over the eastern part of Canadian Arctic, Baffin Sea and probably over most of Greenland (in the case of T_{max})—Figure 6. The greatest change in trend (between periods 1951–1990 and 1961–1990) occurred over the Atlantic region of the Arctic (from −0·2°C per decade to about 0·2–0·4°C per decade). The positive trends in the period analysed are attributable to the fact that the Arctic in the 1960s was as its coldest, at least since 1920. From the mid-1970s, T_{max} and T_{min} for most of the Arctic show a lack of trends or small trends (Figure 7).

The majority of trends, for all periods analysed, are not statistically significant. The calculations have also shown that very rarely the linear trends explain more than 10 per cent of the general variance of T_{max} and T_{min}.

Based on the aforementioned results one can conclude that the anthropogenic warming projected by GCM outputs is not seen in the Arctic in the last decades. According to Przybylak (1996), this could be due to: (i) a delay in reaction of the Arctic climate system, which has considerable inertia due to large water masses and sea and land ice; (ii) the influence of natural factors (mainly of the change in atmospheric circulation that occurred in the mid-1970s—since this time there is a clear rising frequency of zonal circulation (see e.g. Kożuchowski, 1993; Hurrell, 1995; Przybylak, 1996)), which, although leading to cooling of the Arctic, considerably reduces or

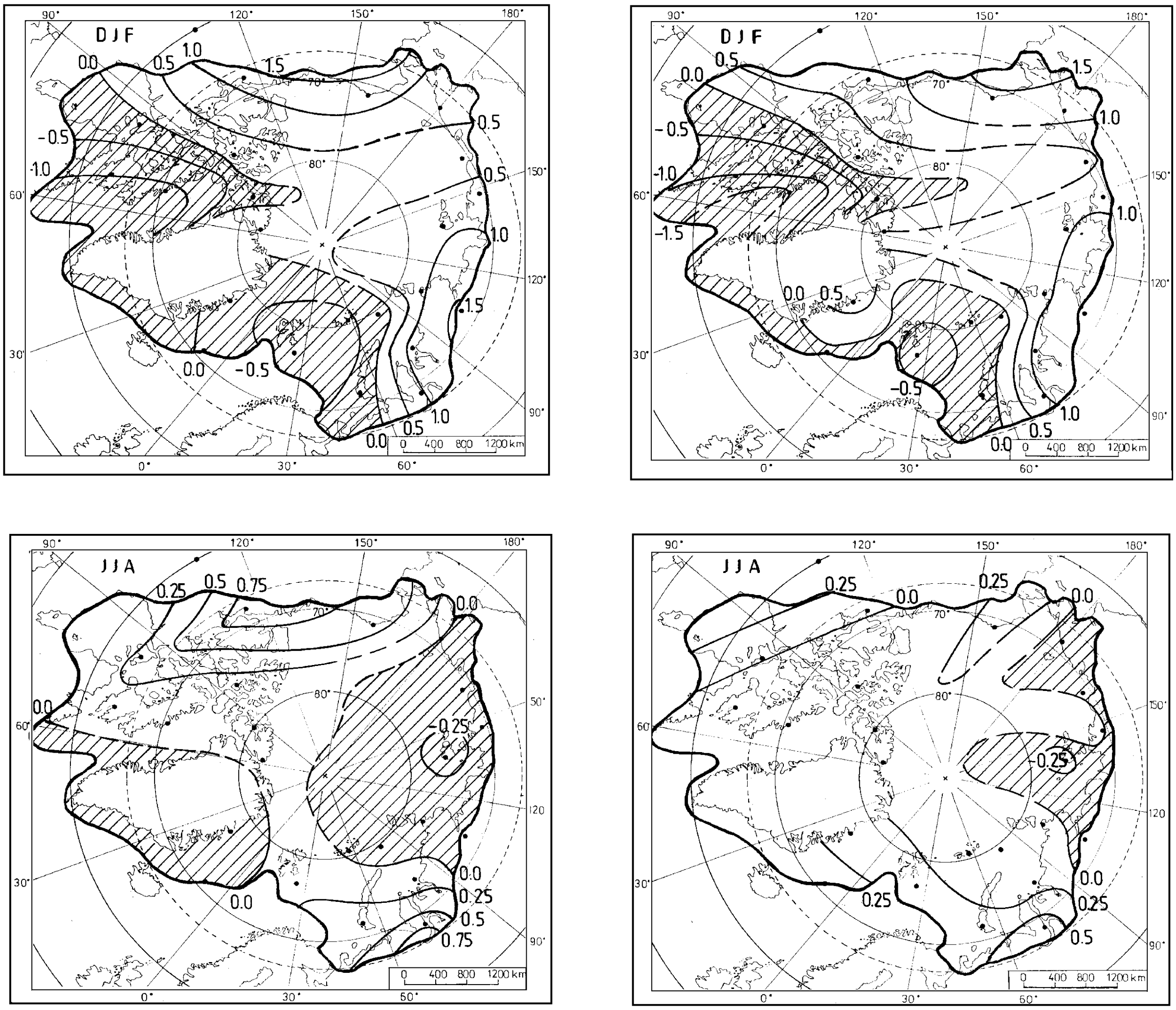

Figure 5. Same as Figure 4, but for winter (DJF) and summer (JJA): T_{max}, left panels; T_{min}, right panels

completely removes the warming caused by greenhouse gases; (iii) the influence of anthropogenic sulphate aerosols, which, as investigations by Santer *et al*. (1995) have shown, is very strong, especially in the high latitudes—in large parts of the Arctic, reduction of air temperature connected with this aerosol is greater than warming caused by the enhanced greenhouse effect of CO_2; and (iv) combination of these three factors.

Diurnal temperature range. The differential rate of changes of T_{max} and T_{min} in the Arctic can be seen in Figures 6–8. In most of the Arctic, increasing (or less decreasing) trends of T_{min} are more pronounced than those of T_{max}. Such asymmetric trends of the extreme air temperatures lead to a decrease in the DTR (Table II). For the period 1951–1990 this decrease has occurred with a frequency of 76 per cent for winter, spring and summer, 64 per cent for autumn, and 72 per cent for the year (Table III). Spatial distributions of the mean annual and seasonal DTR trends in the period 1951–1990 are presented in Figures 9 and 10. It is seen that positive trends of both annual and seasonal DTR in the Arctic occurred during this time mainly in the Canadian Arctic. They are more common in summer than in winter, but in the latter season the increase of the DTR was also noted over a small fragment of the western part of Russian Arctic (Figure 10). Only every second or third station (from all stations with a decreasing trend in the DTR) shows a statistically significant decrease in the DTR (Tables II and III).

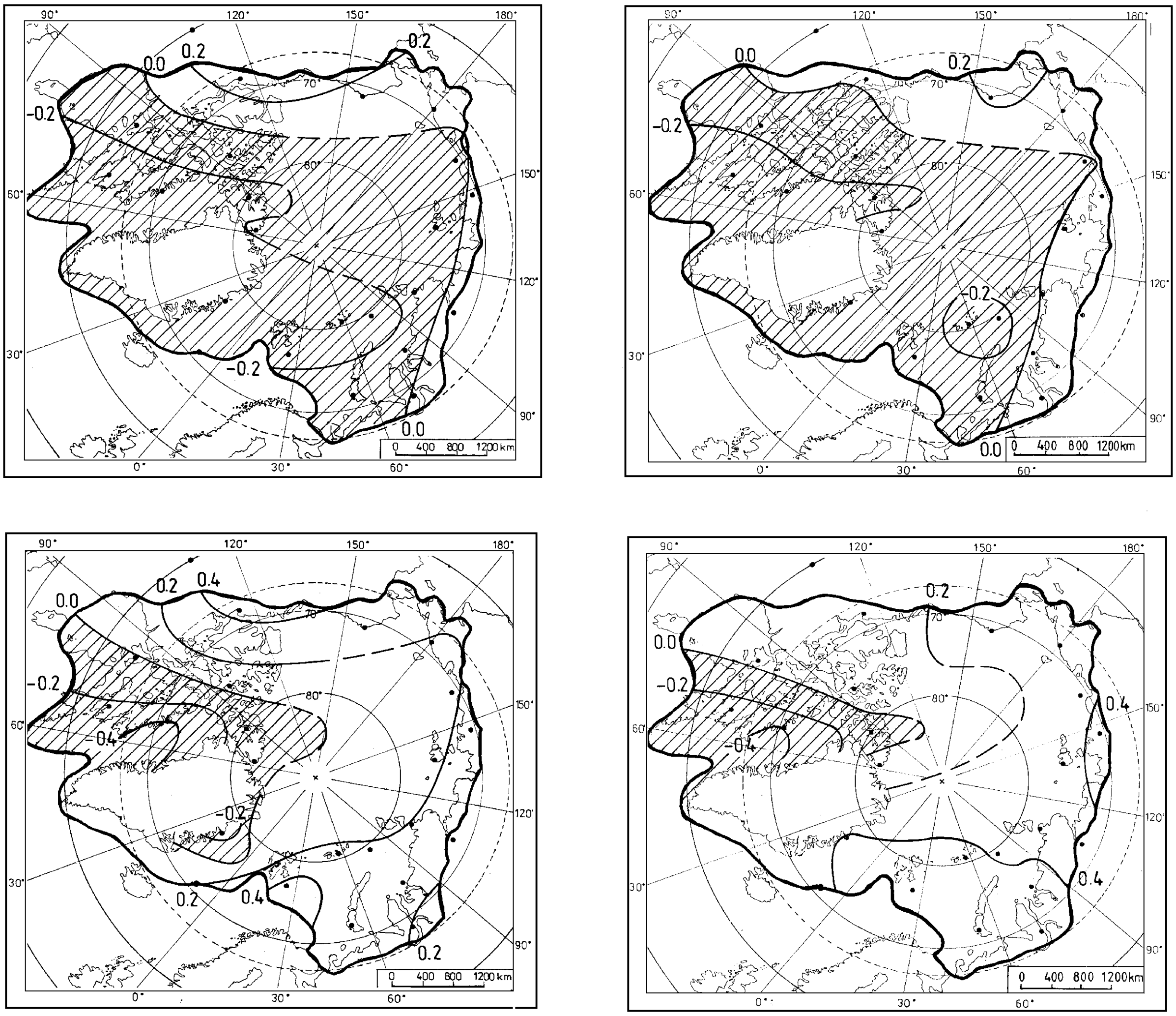

Figure 6. Spatial distribution of the mean annual T_{max} (left panels) and T_{min} (right panels) trends (°C per decade) in the Arctic over the period 1951–1990 (upper maps) and 1961–1990 (lower maps). Negative trends are hatched. Note also that the isopleths around the southern part of Greenland are not drawn owing to lack of data, but the probable general trends in this area (positive or negative) were estimated taking into account all the information available from neighbouring stations as well as the behaviour of the mean air temperatures (Przybylak, 1996). Dots, meteorological stations; dashed lines, probable course of the isopleths

In the period 1961–1990, asymmetric trends of the extreme air temperatures are more strongly marked than in the previously analysed period (Table III). During this time only three stations, Barrow, Coppermine, and Iqaluit A, show an increase in mean annual DTR. From Table II and Figure 9 one can also see that the decreases in the DTR in other parts of the Canadian Arctic are smaller than in the rest of the Arctic.

In the period 1971–1990 the decrease in the mean annual DTR trends occurred in only 58 per cent of stations (Tables II and III, and Figure 9). An increase in DTR, beside the Canadian Arctic, was also noted over a large part of the Norwegian Arctic. The decrease in the DTR was slightly more common in the cool half-year than in the warm half-year (Tables II and III). In summer a greater part of the Arctic even shows an increase in DTR. Also, the number of statistically significant negative trends in the DTR has dropped dramatically, especially in winter and spring (Table III). Summarizing the above results, one can note that in most of the Arctic a decrease in the DTR is observed, similar to other parts of the world (Karl *et al*., 1991, 1993). However, only about 30 to 55 per cent of them (depending on the season) are statistically significant. This conclusion is confirmed also by the comparison of Figure 3 in Karl *et al*. (1993) and Figure 7, which includes the fluctuations of the mean annual

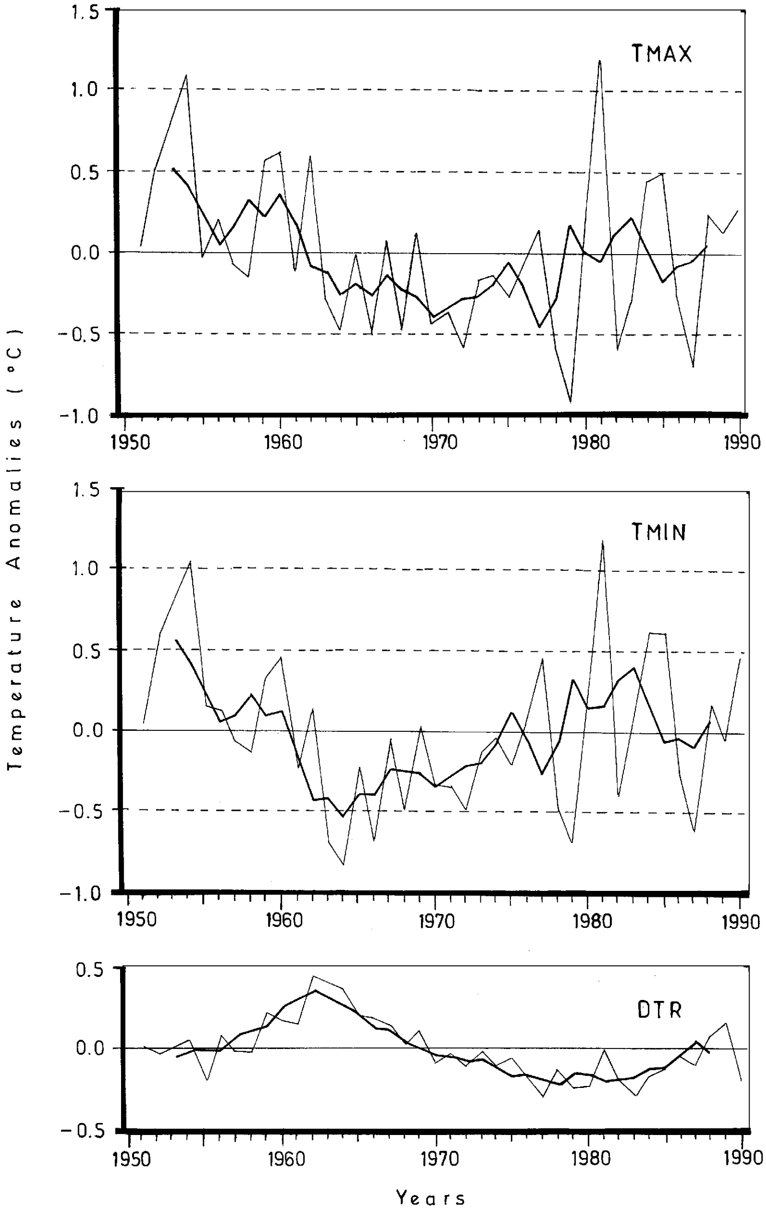

Figure 7. Time series of the air temperature anomalies of the annual mean maximum (T_{max}), minimum (T_{min}), and diurnal temperature range (DTR) in the Arctic (based on data from 26 stations). Smooth curve is a weighted 5-year running mean

anomalies of the DTR for the Arctic. Except for the first few years, a very good correspondence is apparent in these curves. It is worth noting also that the magnitude of changes in the DTR was similar, and from the early 1960s to 1980s a drop equal to about 0·5°C occurred. Such a reduction in the DTR in the Arctic is, however, greater than simulated by climatic models. The experiments carried out by Hansen *et al*. (1995) suggest that the changes in the DTR in the Arctic, relevant to the global warming of 0·5°C, should oscillate from −0·1 to −0·3°C (for the case when aerosols and clouds are uniformly distributed over land) or from +0·1 (central Arctic) to −0·3°C (near-continental parts of the Arctic) for the real distributions of the tropospheric sulphate aerosols. The question of what are the reasons of this discrepancy between observed data and model projections, remains open.

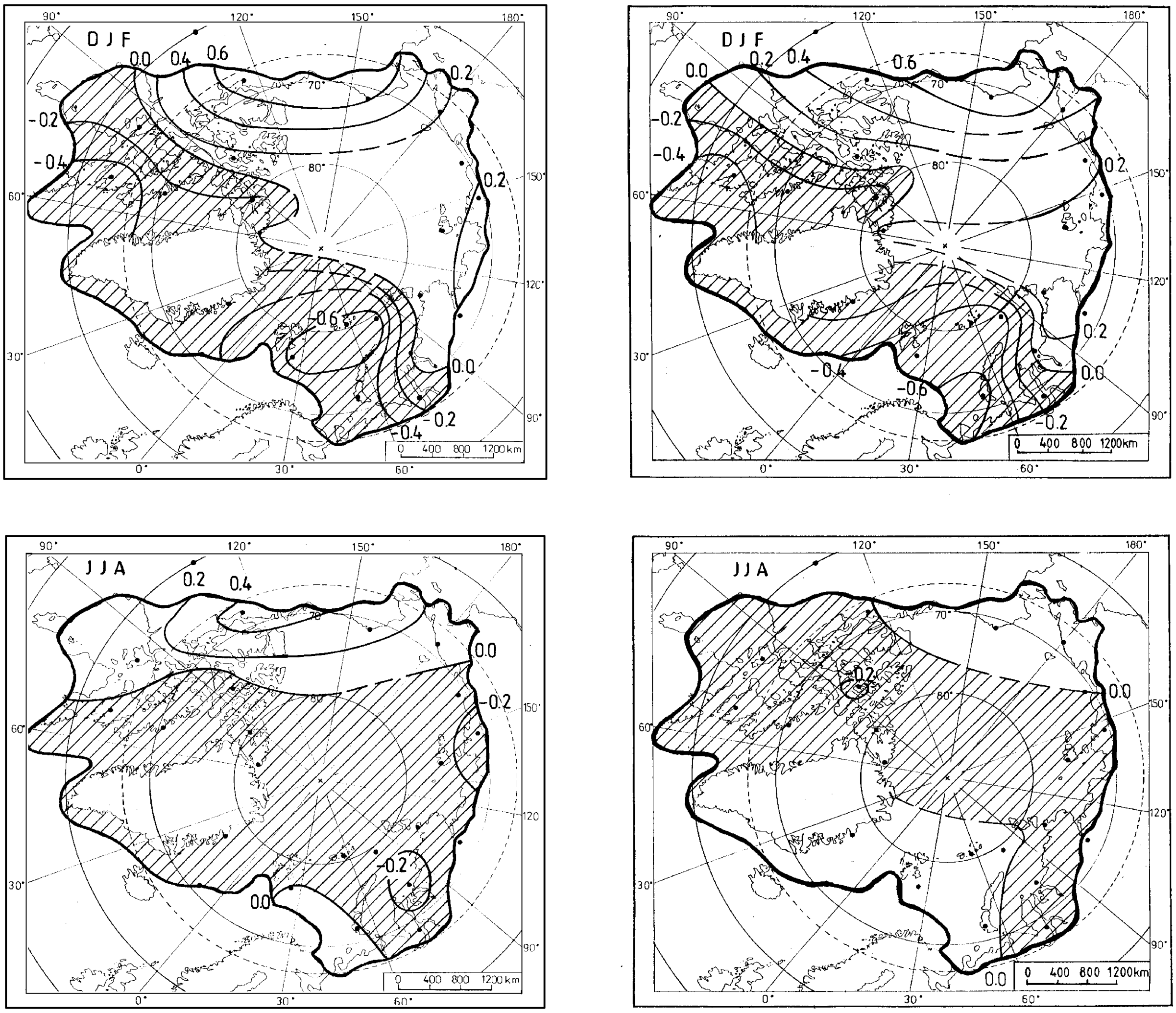

Figure 8. Same as Figure 6, but for winter (DJF) and summer (JJA), 1951–1990: T_{max}, left panels; T_{min}, right panels

What are the causes of the damping in the DTR in the last decades? The answer is rather difficult. According to Karl *et al*. (1991, 1993), the most probable factor is increase of cloudiness, which explains, together with the changes of clouds mean ceiling height, the greatest portion of the variance in DTR. Frich (1992) also presents similar view. This was recently reliably confirmed by Hansen *et al*. (1995), who used a global climate model to investigate the impact of a wide range of radiative forcing and feedback mechanisms in the daily cycle of surface air temperature. They found that 'Only an increase of continental cloud cover, possibly a consequence of anthropogenic aerosols, can damp the diurnal cycle by an amount comparable to observations'. Other climatic factors (e.g. snow cover, mean wind speed) are much less important. The influence of local effects, such as urbanization, irrigation, and desertification, is also weak, and in the case of the Arctic is practically absent.

The second group of factors seriously influencing DTR (from one-third to one-half of the observed damping of the diurnal cycle) is, according to Hansen *et al*. (1995), the increase of greenhouse gases and tropospheric aerosols.

Taking into account these findings I have decided to investigate the relations between cloudiness and T_{max} and T_{min} in the Arctic. The behaviour of cloudiness in the region studied was analysed using data from 19 stations. The increasing trends in cloudiness over the period 1961–1990 have been found over the European and Russian

Table II. Mean seasonal (DJF, MAM, JJA and SON) and annual trends of the DTR (°C per 10 years) in the Arctic

Station	1951–1990					1961–1990					1971–1990				
	DJF	MAM	JJA	SON	Annual	DJF	MAM	JJA	SON	Annual	DJF	MAM	JJA	SON	Annual
Angmagssalik[a]	0·32	−0·46**	−0·28	0·22	−0·06	−0·78*	−0·73*	−1·08***	−0·54*	−0·80***	−0·52	0·49	−1·34**	−0·90	−0·60**
Kap Tobin[a]	0·67	0·22	0·07	0·13	0·28	−1·24*	−0·86	−0·58*	−0·71**	−0·82**	0·01	−1·66	−1·25**	−1·02*	−0·80
Danmarkshavn[b]	−0·54*	−0·47**	−0·48***	−0·74***	−0·57***	−0·89**	−0·67**	−0·55**	−1·00***	−0·78***	−0·57	−0·60	−0·18	−0·39	−0·44
Jan Mayen	−0·02	−0·23*	−0·22***	−0·20**	−0·16**	−0·26	−0·18	−0·19*	−0·08	−0·18*	−0·08	−0·22	0·11	0·18	0·04
Hopen	−0·06	−0·17	−0·08*	−0·00	−0·07	−0·22	−0·25*	−0·01	−0·03	−0·12	0·16	−0·32	0·02	0·38	0·10
Malye Karmakuly	0·08	−0·01	−0·03	−0·01	0·01	−0·16	−0·21	−0·08	−0·12	−0·14*	−0·33	−0·37	0·19	−0·53*	−0·23
Polar GMO E. T. Krenkelya	–	–	–	–	–	−0·30	−0·10	−0·06	−0·14	−0·15*	−0·39	0·34	0·11	0·22	0·06
Mys Kamenny	−0·04	−0·09	−0·22**	−0·21*	−0·13*	−0·27	−0·28*	0·11	−0·29	−0·23**	−0·03	−0·19	0·14	−0·50	−0·14
Ostrov Vize	−0·17	−0·06	−0·12**	0·04	−0·08	−0·32	−0·01	−0·25***	−0·03	−0·15	−0·50	0·20	−0·19*	0·25	−0·07
Ostrov Dikson	−0·14	−0·04	−0·17*	−0·07	−0·10*	−0·27	0·01	−0·21	−0·01	−0·11	−0·13	0·26	0·22	−0·19	0·04
GMO E. K. Fedorova	−0·19	−0·20**	−0·15*	−0·04	−0·15**	−0·15	−0·22	−0·20	−0·04	−0·15*	−0·01	−0·22	−0·14	0·08	−0·09
Ostrov Kotelny	−0·19**	−0·14	−0·02	−0·13	−0·12*	−0·13	0·00	−0·08	−0·28*	−0·12	0·32*	0·84*	−0·06	0·03	0·27*
Cokurdah	−0·04	0·02	−0·12	−0·16*	−0·08	−0·22*	−0·20	−0·22	−0·30**	−0·24***	−0·34	−0·47*	−0·47	−0·37	−0·42**
Ostrov Chetyrekhstolbovoy	−0·14	0·03	−0·10	−0·04	−0·06	−0·16	0·01	−0·11	−0·12	−0·10	−0·30	0·14	−0·64**	0·01	−0·19
Mys Szmidta	−0·23	−0·17	−0·08	−0·03	−0·13**	−0·32	−0·24	0·12	−0·15	−0·14*	−0·58	0·25	−0·10	−0·03	−0·08
Barrow	−0·17	−0·10	0·13	−0·06	−0·05	−0·03	−0·14	0·32*	−0·00	0·04	0·34	0·34	0·36	0·37	0·37*
Coppermine	0·30**	0·20*	0·49***	0·26***	0·31***	0·60***	0·17	0·48*	0·18	0·36***	0·66*	0·58*	1·11**	0·07	0·58***
Resolute A	−0·05	−0·06	0·06	0·08	0·01	−0·17	−0·08	0·09	−0·00	−0·05	0·20	−0·16	−0·24	−0·28	−0·14
Eureka	0·14	−0·08	−0·03	0·05	0·02	0·06	−0·19*	−0·02	−0·17	−0·08	0·63*	0·17	0·11	−0·18	0·17
Coral Harbour A	−0·06	0·03	0·17	0·01	0·04	−0·04	−0·25	0·09	−0·09	−0·09	0·40	−0·25	−0·31	−0·51	−0·20
Iqaluit A	−0·07	0·14	0·07	0·06	0·05	0·00	0·21*	0·06	0·07	0·09	0·29	−0·08	0·02	0·05	0·07
Clyde A	0·01	−0·23	−0·02	0·13	−0·03	−0·08	−0·08	−0·01	0·09	−0·02	1·03**	0·91*	0·02	0·50	0·60**
Alert	−0·22*	−0·33***	−0·10	−0·02	−0·17***	−0·18	−0·22**	−0·09	−0·20	−0·18*	0·09	−0·15	0·01	−0·27	−0·07
Upernavik[a]	−0·14	−0·34*	−0·20	−0·18*	−0·22	−0·68	−0·58	−0·75**	−0·62***	−0·67**	0·45	0·30	0·38	−0·61*	0·18
Jakobshavn[a]	−0·71***	−0·23	−0·18	−0·43***	−0·39***	−0·85**	−0·47	−0·30	−0·48**	−0·53**	−1·71*	−2·56	0·23	−1·39**	−1·38**
Godthab[a]	−0·23	−0·29**	−0·32**	−0·47***	−0·34***	−0·88**	−0·50*	−0·41	−0·54**	−0·60***	−1·08	0·20	−0·34	−0·17	−0·41

*, **, *** Trends statistically significant at the levels of 0·05, 0·01, and 0·001, respectively.
[a] Data for 1951–1980.
[b] Data for 1955–1990.

Table III. Mean seasonal (DJF, MAM, JJA, and SON) and annual frequency of occurrence (in per cent) of decreasing trends in the DTR for selected Arctic stations

Period	DJF	MAM	JJA	SON	Annual
1951–1990					
a	76	76	76	64	72
b	26	37	42	44	56
1961–1990					
a	88	81	73	88	88
b	26	33	32	35	61
1971–1990					
a	54	50	46	58	58
b	7	8	33	27	20

(a) All cases with decreasing trend.
(b) Only statistically significant decreasing trends (frequency is computed relative to all cases with decreasing trend).

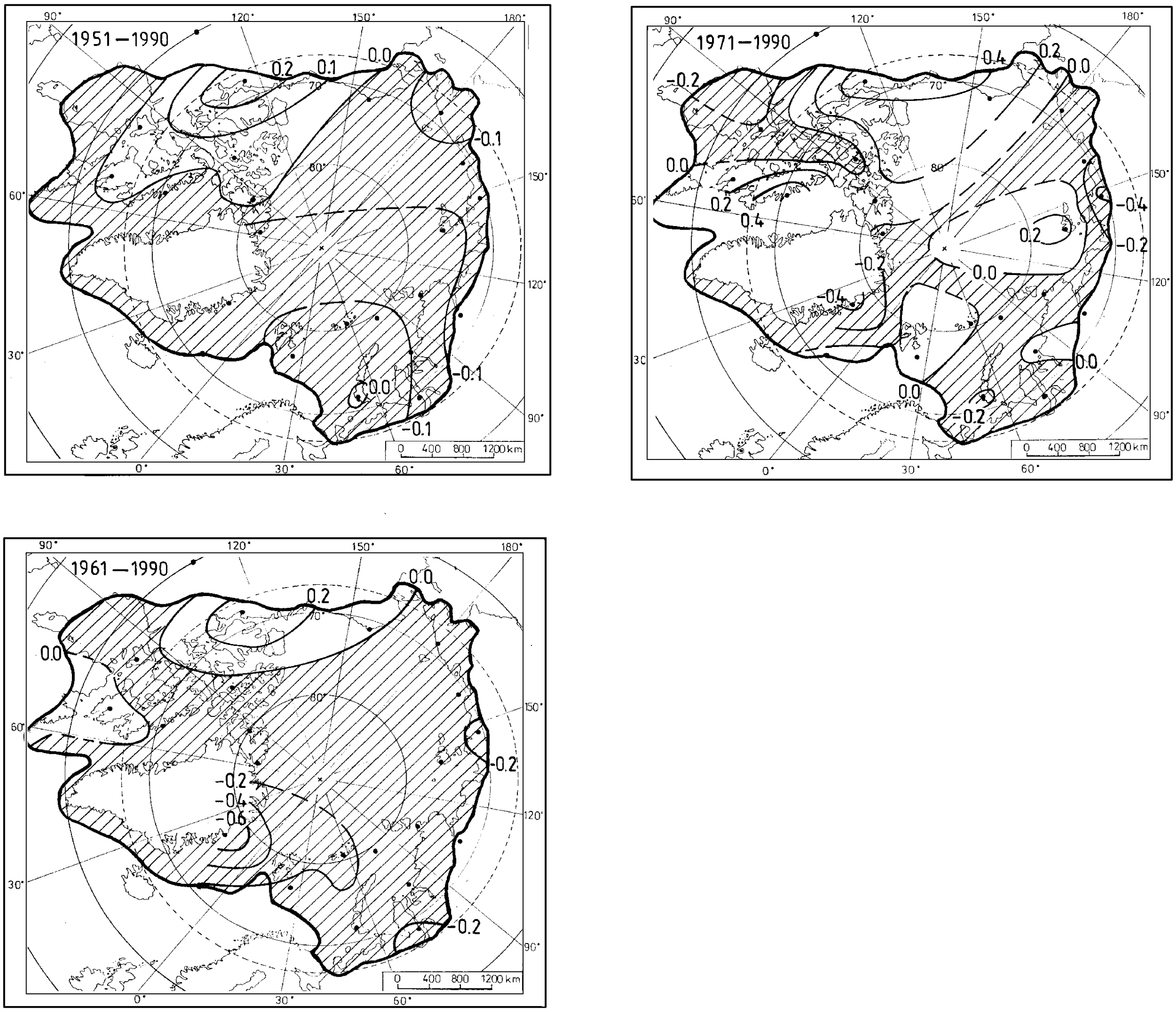

Figure 9. Spatial distribution of the mean annual DTR trends in the Arctic (°C per decade). Negative trends are hatched. Note also that the isopleths around the southern part of Greenland are not drawn owing to lack of data. The probable general trends in this area (positive or negative) were estimated based upon the DTR trends in the nearest stations. Dots, meteorological stations; dashed lines, probable course of the isopleths

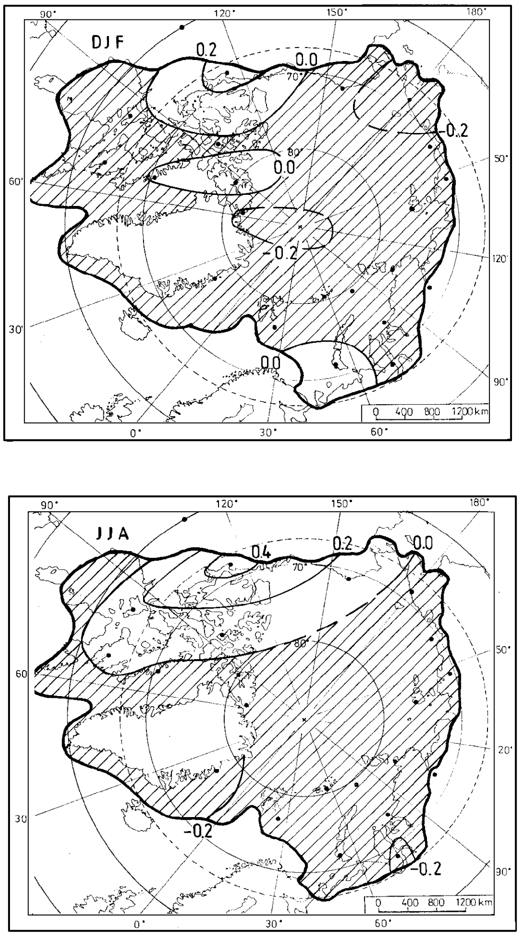

Figure 10. Spatial distribution of the mean winter (DJF) and summer (JJA) DTR trends in the Arctic (°C per decade) over the period 1951–1990. Key as in Figure 9

Arctic as well as over Baffin Sea and its vicinity (Table IV and Figure 11), that is, in the regions that are characterized by the greatest decrease in the DTR. It is important to notice that these regions of the Arctic lie nearest the main sources of emissions of anthropogenic aerosols. So, it is possible that the increase of cloudiness here is connected with the rise of aerosol concentration. In the rest of the Arctic, where a decrease of cloudiness occurred in the last decades, the changes of DTR are very small and oscillate around its long-term mean. These conclusions are also confirmed by the computed mean DTR. For the part of the Arctic where increases of cloudiness are observed the trend of annual DTR is equal to −0·16°C per decade, whereas in areas with a decrease in cloudiness it is 0·03°C per decade. This suggests that one of the most important factors influencing the decrease in DTR in the Arctic is the increase of cloudiness. However, it must be noted that the relationship between cloudiness and extreme air temperatures in the Arctic is not so clear and simple as at lower latitudes. A statistically significant negative correlation exists mainly in summer and only in some Arctic regions in spring and autumn (Table V). In winter that correlation is even positive over most of the region studied, because at this time the cloudiness depends strongly upon atmospheric circulation. A vigorous cyclonic activity (especially in the Atlantic region of the Arctic) causes the inflow of warm and humid air masses from middle latitudes to the Arctic. As determined by Przybylak (1992), for the Hornsund station (Spitsbergen), the mean extreme temperatures are greater (the differences are especially marked in the cool half-year) on cloudy than on clear days (except summer months, especially in the case of T_{max}). Przybylak (1992) found also that mean DTR in Hornsund

Table IV. Seasonal (DJF, MAM, JJA, and SON) and annual linear trends (a) of cloudiness (in tenths per 10 years) in the Arctic over the period 1961–1990 and their share (in per cent) in general variance of cloudiness (b)

Station		DJF	MAM	JJA	SON	Annual	Station		DJF	MAM	JJA	SON	Annual
Danmarkshavn	a	−0·01	0·02	−0·25	−0·20	−0·09	Cokurdah	a	0·02	−0·05	0·04	−0·08	−0·02
	b	0·01	0·09	12·52	7·31	4·57		b	0·12	0·55	0·46	2·11	0·31
Jan Mayen	a	0·10	0·02	0·04	−0·02	0·04	Ostrov Chetyrekhstolbovoy	a	0·05	0·12	−0·12	−0·04	0·00
	b	3·09	0·28	1·07	0·67	2·47		b	0·29	1·91	4·81	0·50	0·00
Hopen	a	−0·04	0·18	−0·02	0·01	0·04	Mys Szmidta	a	−0·02	−0·16	−0·24*	−0·08	−0·13
	b	0·15	7·67	0·35	0·03	0·67		b	0·10	4·48	18·60	1·44	8·10
Malye Karmakuly	a	0·05	−0·05	−0·24	−0·05	−0·07	Resolute A	a	0·01	0·03	−0·09	−0·13	−0·05
	b	0·69	0·50	11·98	0·60	4·94		b	0·01	0·22	2·35	3·47	2·06
Polar GMO E. T.	a	0·80***	0·23	0·03	0·06	0·27***	Eureka	a	−0·12	−0·04	−0·10	0·05	−0·06
Krenkelya	b	53·80	10·87	0·99	1·59	54·72		b	2·71	0·38	1·81	0·61	1·68
Mys Kamenny	a	−0·11	−0·13	−0·12	−0·06	−0·13	Coral Harbour A	a	−0·10	0·07	−0·02	−0·10	−0·04
	b	1·80	4·46	2·70	1·03	11·47		b	2·08	1·37	0·16	2·18	1·16
Ostrov Vize	a	0·58**	0·22	0·01	0·06	0·20*	Iqaluit A	a	0·08	0·41**	0·13	−0·03	0·16
	b	24·91	5·67	0·07	0·68	17·99		b	0·63	24·95	5·67	0·18	8·86
Ostrov Dikson	a	0·60**	0·16	0·04	0·14	0·23***	Clyde A	a	−0·06	0·34	0·05	0·08	0·10
	b	29·13	4·69	0·95	3·82	35·56		b	0·24	10·98	0·59	2·26	2·55
GMO E. K.	a	0·55**	0·27*	−0·04	0·20	0·26***	Alert	a	0·10	0·26	0·17	0·14	0·17*
Fedorova	b	27·11	19·08	0·70	12·88	35·95		b	1·52	9·72	8·00	4·49	13·79
Ostrov Kotelny	a	0·46**	0·19	−0·03	0·03	0·17*							
	b	24·00	6·68	0·59	0·32	15·55							

*, **, *** Trends statistically significant at the levels of 0·05, 0·01, and 0·001, respectively.

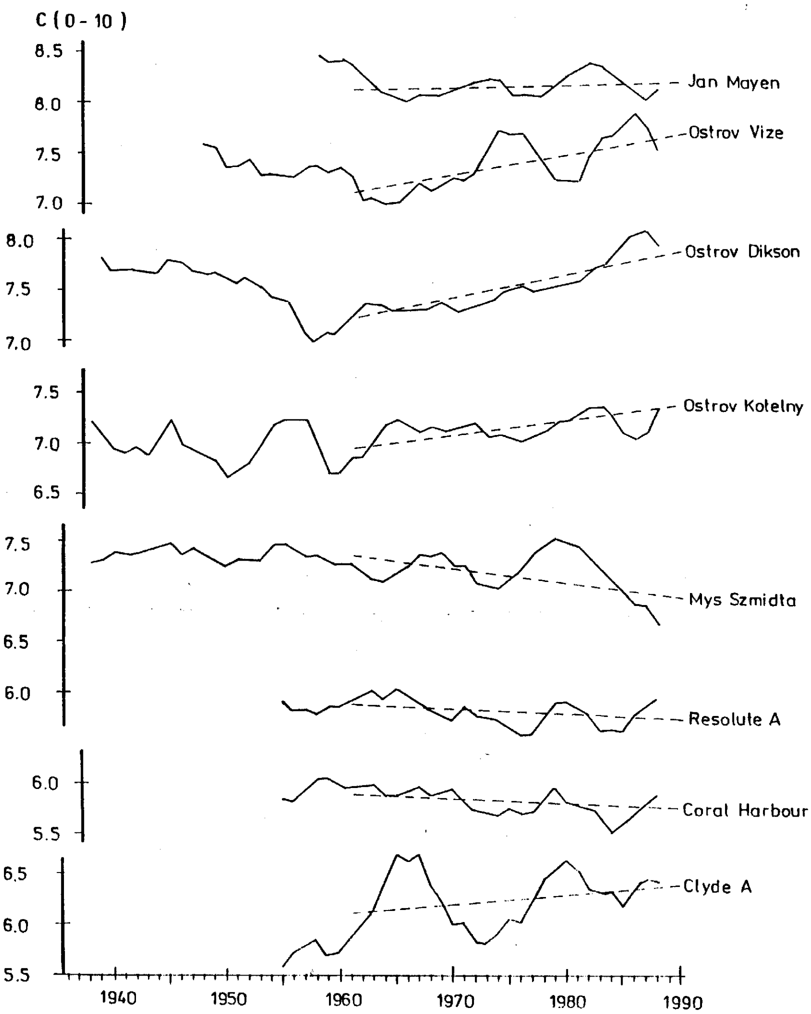

Figure 11. Courses of the 5-year running means of cloudiness (solid line) and their linear trends over the period 1961–1990 (dashed line) in selected Arctic stations

Table V. Correlations between observed seasonal (DJF, MAM, JJA, and SON) and annual mean DTR and mean cloudiness at each of the 10 stations representing different climatic regions of the Arctic

Station	Period	DJF	MAM	JJA	SON	Annual
Danmarkshavn	1955–1980	0·03	0·19	0·15	−0·04	0·33
Jan Mayen	1956–1990	−0·32	−0·20	−0·34*	−0·06	−0·26
Hopen	1956–1990	−0·30	−0·07	−0·16	−0·62***	−0·50**
Ostrov Vize	1951–1990	0·16	−0·17	−0·49**	−0·18	0·13
Ostrov Dikson	1951–1990	0·09	−0·08	−0·55***	−0·16	−0·12
Ostrov Kotelny	1951–1990	0·05	0·04	−0·55***	−0·08	−0·13
Mys Szmidta	1951–1990	−0·13	−0·50***	−0·38**	−0·71***	0·02
Resolute A	1953–1990	0·17	−0·10	−0·59***	−0·20	−0·21
Coral Harbour A	1953–1990	0·30	−0·56***	−0·66***	−0·47**	−0·40*
Clyde A	1953–1990	0·37*	−0·23	−0·14	−0·17	0·02

*, **, *** Coefficients of correlation statistically significant at the levels of 0·05, 0·01, and 0·001, respectively.

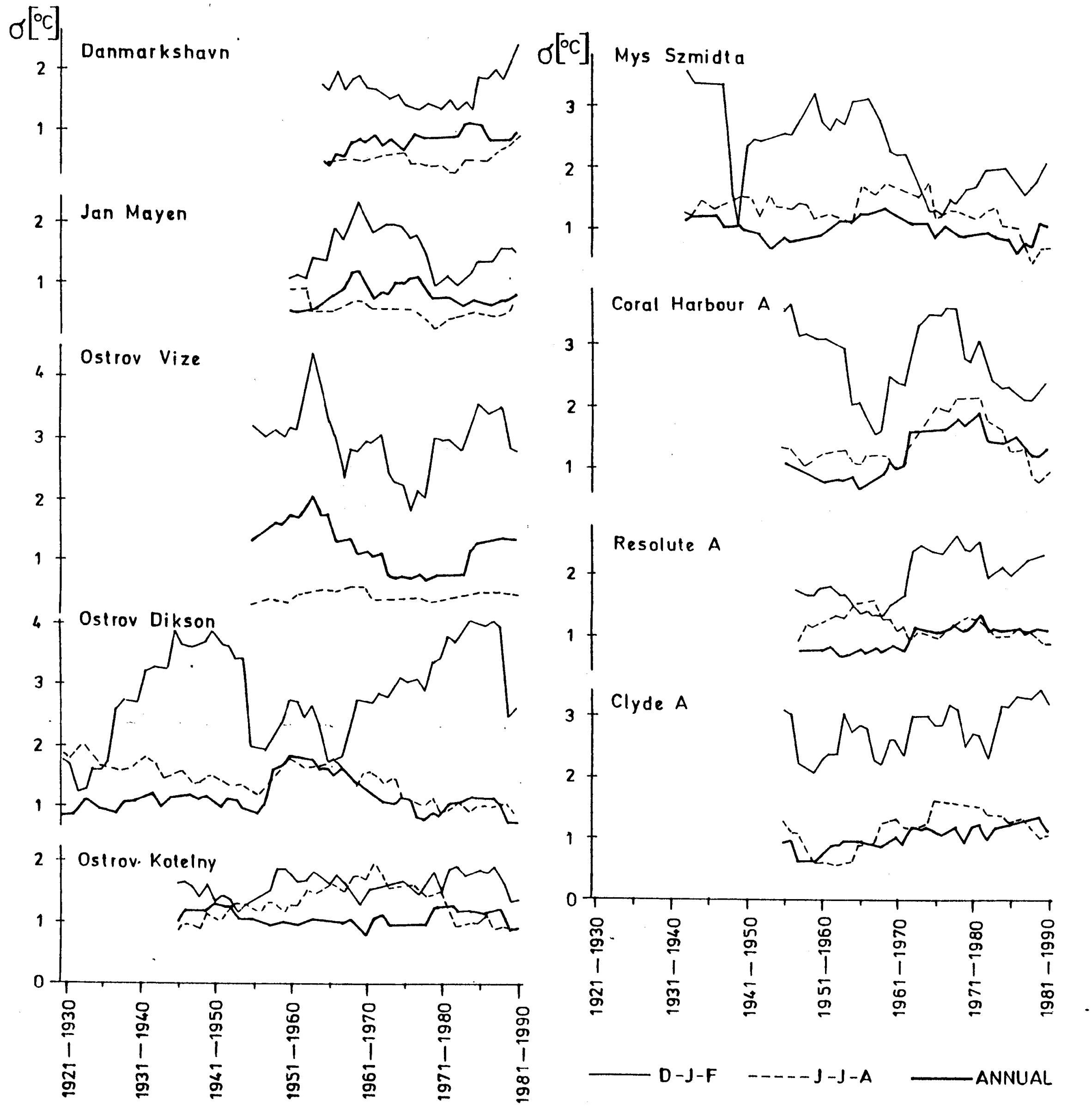

Figure 12. Standard deviations of mean winter (D–J–F) , summer (J–J–A) and annual T_{max} for running decades in selected Arctic stations

is greatest during clear days in spring, autumn and, especially, in summer. The opposite is true for winter. Thus in winter factors other than an increase in cloudiness must affect the decrease in the DTR. It is well known from many studies (e.g. Baranowski, 1968; Przybylak, 1992) that the DTR in the Arctic in winter, early spring and late autumn (when the solar radiation is low or not present) is shaped mainly by non-periodic day-to-day changes of air temperature. In turn, these fluctuations of temperature are controlled largely by the thermal advection associated with synoptic-scale cyclones and anticyclones. The results presented in Figure 8b in Karl *et al.* (1993) confirm this conclusion, showing that the partial correlation coefficient between day-to-day changes of temperature and DTR grows when solar radiation (and thus temperature) decreases. As also can be seen from this figure, in such solar conditions this variable is equally important, as is the added influence of an increase in cloud cover and decrease in its ceiling height. These results are based on data taken from mid-latitudes. Of course, for the Arctic, where during winter the incoming solar radiation is significantly lower than in the mid-latitudes or even not present (polar night), the day-to-day changes of temperature must play a considerably greater role in decreasing the DTR.

Fluctuations of variability of winter, summer and annual T_{max} and T_{min} have been analysed using standard deviations on running decades. Because the changes of variability of both the thermal parameters are very similar, I present here in graphic form only the results for T_{max} (Figure 12). In the whole Arctic—excluding regions around the stations Ostrov Vize and Clyde A—the increase in T_{max} and T_{min} variability in the last 10–20 years is not seen. In contrast, many areas even show a decrease in σ (e.g. Ostrov Dikson, Mys Szmidta, Coral Harbour A, Resolute A) or no distinct changes (Jan Mayen, Ostrov Kotelny). The highest values of dispersion of winter and annual extreme temperatures in recent years occur only at Clyde A (1980–1989). In most of the remaining areas of the Arctic the maximum of variability occurred either in the 1950s or 1970s. The regions of the Arctic that are strongly influenced by cyclonic circulation (Jan Mayen, Ostrov Vize) have annual values more similar to winter values than to summer σ. The opposite is true for the most continental areas.

CONCLUSIONS

Trends in T_{max} and T_{min} in the Arctic show insignificant changes in recent decades. The signs of their trends depend upon placement of the starting point, from which the trends are computed (see Przybylak and Usowicz, 1994). Although the changes in the temperature extremes are small, a significant decrease in the DTR over a large part of the Arctic is seen. The results of this analysis show that in the Arctic one of the most important factors influencing the decrease in DTR is an increase in cloudiness. The effect of this factor is especially strong in the warm half-year. In the cool half-year the dominant variable damping the DTR are day-to-day changes of temperature governed mainly by atmospheric circulation. It follows that the relationship between cloudiness and extreme air temperatures in the Arctic is not so clear and simple as at lower latitudes. An increase in variability of both parameters analysed has not occurred in the most recent years. These findings confirm Kahl *et al.*'s (1993a) statement that in the Arctic there is an absence of evidence for greenhouse warming in the period 1951–1990.

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