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Influence of cloudiness on extreme air temperatures and diurnal temperature range in the Arctic in 1951–1990

ABSTRACT: Detailed analysis of the influence of cloudiness on extreme air temperatures and diurnal temperature ranges (DTR) in the Arctic in 1951–1990 is presented. This analysis is preceded by a description of a cloudiness fluctuation and trends in the Arctic during the last decades. A statistically significant increase of the mean cloudiness in the Arctic occurred in winter, spring and during a year. It could be due to incursion of a polluted air to the Arctic from the lower latitudes. An overall pattern of the influence of cloudiness on the daily maximum air temperature (TMAX) and the minimum air temperature (TMIN) is roughly similar. However, sometimes there are significant differences in the anomalies for clear, partly cloudy and cloudy days. In summer even an opposite influence of cloudiness on TMAX than on TMIN was noted for the Norwegian Arctic and the southern Canadian Arctic. Relations between cloudiness and DTR, based on daily data, entirely confirm the previous conclusions based on monthly data. Therefore, an increasing cloudiness of the last decades significantly influences a decrease of DTR in the Arctic, especially during the warm half-year when a solar radiation is present. During the cool half-year (particularly at polar night) an influence of cloudiness is clearly weaker and not univocal, and probably less important than non-periodical day-to-day changes of air temperature, governed at this time by very vigorous atmospheric circulation.

Key words: Arctic, cloudiness trends, extreme air temperatures, time series analysis.

Introduction

Numerous publications published in 1990s stress a large- and regional-scale warming of the Earth's surface over the last decades, caused mainly by a rising daily minimum air temperature (TMIN), while a daily maximum air temperature

(TMAX) does not indicate any significant changes (*e.g.* Karl *et al.* 1991, 1993, 1994, 1995; Frich 1992; Kukla and Karl 1993; Böhm and Auer 1994; Weber *et al.* 1994, 1997; Dessens and Bücher 1995; Horton 1995; Jones 1995a; Kaas and Frich 1995; Brázdil *et al.* 1995, 1996; Plummer *et al.* 1995; Przybylak 1996a–c, 1997; Türkes *et al.* 1996; Easterling *et al.* 1997; Zheng *et al.* 1997). These papers confirm the conclusions of Houghton *et al.* (1992) and Karl *et al.* (1993) on the worldwide rise of the minimum land-surface air temperature since 1950, equal to about twice those of the maximum. As a result of these asymmetric trends, a statistically significant decrease of a mean monthly diurnal temperature range (DTR), defined as a difference between the mean monthly TMAX and TMIN, occurs. New analyses for the whole world (Horton 1995; Easterling *et al.* 1997) which cover approximately 42 and 54% of the global land area respectively, are in agreement with the earlier findings. Additional areas with decreasing DTR (not included in *op.cit.*) were identified by Jones (1995b) for the Antarctic and by Przybylak (1996a–c, 1997) for the Arctic. It is also noteworthy that satellite-based estimates of the tropospheric DTR in 1973–1993 (Ross *et al.* 1996) and 1979–1994 (Balling Jr. and Christy 1996) prove significant correspondence of surface and tropospheric trends of DTR.

Recent calculations of DTR using climatic models *e.g.* a one-dimensional radiative-convective model (Cao *et al.* 1992) and a general circulation model (Rind *et al.* 1989; Hansen *et al.* 1993, 1995), indicate that DTR should decrease due to doubling of carbon dioxide, and a tropospheric aerosol increases. However, comparison of DTR reduction in the last four decades, computed by models and obtained from observations, reveals large differences (Stenchikov and Robock 1995; Przybylak 1997).

What are the potential factors controlling the behaviour of DTR? The answer to this question is still open, but a significant advance was made in the last few years. Karl *et al.* (1993) presented a list of these factors and an estimation of their role in shaping DTR firstly in a more detailed way. They found the two variables related to changes in cloudiness, sky cover and ceiling height explaining the greatest portion of variance of DTR. They also presented that a cloud cover, including low clouds, has increased in many areas with a decrease in DTR. The others proposed cloudiness as the most probable factor damping DTR also *e.g.* Plantico *et al.* (1990), Frich, (1992), Henderson-Sellers (1992), Lough (1994), Dessen and Bücher (1995), Hansen *et al.* (1995), Jones (1995a), Karl *et al.* (1995), Plummer *et al.* (1995), Brázdil *et al.* (1996), and Przybylak (1997). Basing on a climate model experiment, in which impact of a wide range of radiative forcing and feedback mechanisms in a daily cycle of a surface air temperature was analysed, Hansen *et al.* (1995) concluded 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.' Stenchikov and Robock (1995), using a sophisticated radiative-convective model of the diurnal cycle, have evaluated effects of the increased CO_2 and aerosols on the diurnal cycle of surface temperature. They found feedbacks in the climate system (especially cloud and water vapour feedbacks) to be more important than forcing. Their investigations reveal that 'for all cases with warming DTR decreases, not due to the greenhouse effect of water vapour, but as a result of more intensive absorption of the solar radiation in the near infrared by water vapour and cloud water in a warmer, wetter climate independent of the type of forcing'. The other important conclusion is that the net aerosol effect may increase or decrease DTR due to accompanying drops of temperature and water vapour, which mask a direct (damping) effect of aerosol on DTR. Recently, Dai *et al.* (1997) suggest that the greenhouse gas-induced increases in thick precipitating clouds, and precipitation are better candidates to explain a decrease in DTR than an increase of a cloudiness.

This short review of investigations concerning DTR confirms a sophisticated character of mechanisms that control DTR changes during the increasing concentration of greenhouse gases and aerosols. It is however evident that an increasing cloudiness, which is observed in most of the world, is considered for a leading factor that influences a decrease of DTR during the last decades (*e.g.* Henderson--Sellers 1986, 1989, 1992; Angell 1990; Jones and Henderson-Sellers 1992; Karl and Steurer 1990; Parungo *et al.* 1994; Kane and Gobbi 1995; Houghton *et al.* 1996; Przybylak 1997).

In most of the Arctic, except for fragments of the Canadian Arctic, a significant decrease of DTR occurred, and there is a distinct coincidence of regions with trends of the decreasing DTR and the increasing cloudiness (Przybylak 1996a–c, 1997). However, a relationship between a cloudiness and DTR in the Arctic is significantly more complicated than at lower latitudes. As presented by Przybylak (1997: Table V), a statistically significant negative correlation exists mainly in summer, and in spring and autumn in some regions of the Arctic only. In winter this correlation is even positive over most of the region studied. Przybylak (1997) concluded that in the Arctic an increase in cloudiness is the most important factor influencing a decrease of DTR in the warm half-year. However, in a cool half-year the day-to-day temperature variation, ruled mainly by the atmospheric circulation is the predominant variable that damps DTR. A role of the hemispheric and the Arctic circulation patterns in shaping a behaviour of DTR is presented in another paper (Przybylak 1999).

This paper is focused on detailed relationships between cloudiness and DTR in the Arctic. An important difference between my previous works (Przybylak 1996a, 1997) and the present one is a use of daily data of DTR and cloudiness, and also an inclusion of TMAX and TMIN in the analysis. Such an approach should provide the new and more reliable results, according to the statement made by Robinson *et al.* (1995) that 'some important information regarding climate diagnostics and climate change detection are lost when monthly or longer averages are employed'.

Data and methods

Mean monthly cloud amount data were collected for 19 stations and used to the analysis of trends of cloudiness in the Arctic during the last decades (Fig. 1). In turn, daily data of TMAX, TMIN and DTR, and a mean cloudiness for 10 stations were applied for estimation of the relationships between cloudiness and the above-listed climatic variables (triangles in Fig. 1). All data come from the national meteorological surveys (Danish Meteorological Institute, Norwegian Meteorological Institute and Canadian Climate Center), Arctic and Antarctic Research Institute and the published sources: Norsk Meteorologisk Årbok (1952–1956), Meteorologisk Årbog (1956–1958) and Meteorologičeskij ežemesačnik (1968–1991). A quality control of the data was conducted by Przybylak (1996a). Each of these stations represents a different climatic region or subregion, borders of which as well as of the Arctic itself are defined after Atlas Arktiki (1985; see also Fig. 1).

Standard climatologic methods were used. The daily mean cloudiness data (C) were stratified according to three categories: clear ($C<2$), partly cloudy ($2\leq C\leq 8$) and cloudy ($C>8$) days. For each station and each category of days, mean monthly and seasonal anomalies of TMAX, TMIN and DTR were computed. Mean areal seasonal and annual values of cloudiness in the Arctic and some its parts were calculated by a simple averaging (without weighting).

Results and Discussion

Cloudiness and its trends

Our current knowledge concerning cloudiness in the Arctic, as already observed by Raatz (1981) and Barry *et al.* (1987), is still remarkably poor. Only a few articles have been published lately (*e.g.* Kukla and Robinson 1988; Henderson-Sellers 1989; Schweiger and Key 1992; Parungo *et al.* 1994), the first and the third of which focus on a cloudiness in the Arctic. Two other papers present mainly cloudiness in North America and for the whole world respectively, but give also some information about cloudiness in the Arctic. Exceptionally interesting are the results of Schweiger and Key (1992). For cloud climatology in the Arctic they compared the satellite-derived data from the International Satellite Cloud Climatology Project and the surface--based monthly cloud statistics from the atlas of a global cloud cover by Warren *et al.* (1986, 1988). Among previous general information on cloudiness in the studied area, the most important are the papers of Vowinckel (1962), Huschke (1969), Vowinckel and Orvig (1970), Gorshkov (1980), and Crane and Barry (1984). Crane and Barry (1984) revealed that the presented results in the above-mentioned papers, even for a mean cloudiness in the Arctic, differ significantly. Moreover, Schweiger and Key (1992) found that the satellite-derived clouds occupy generally 5–35% less than indicated by surface observations over the entire Arctic and regional differences

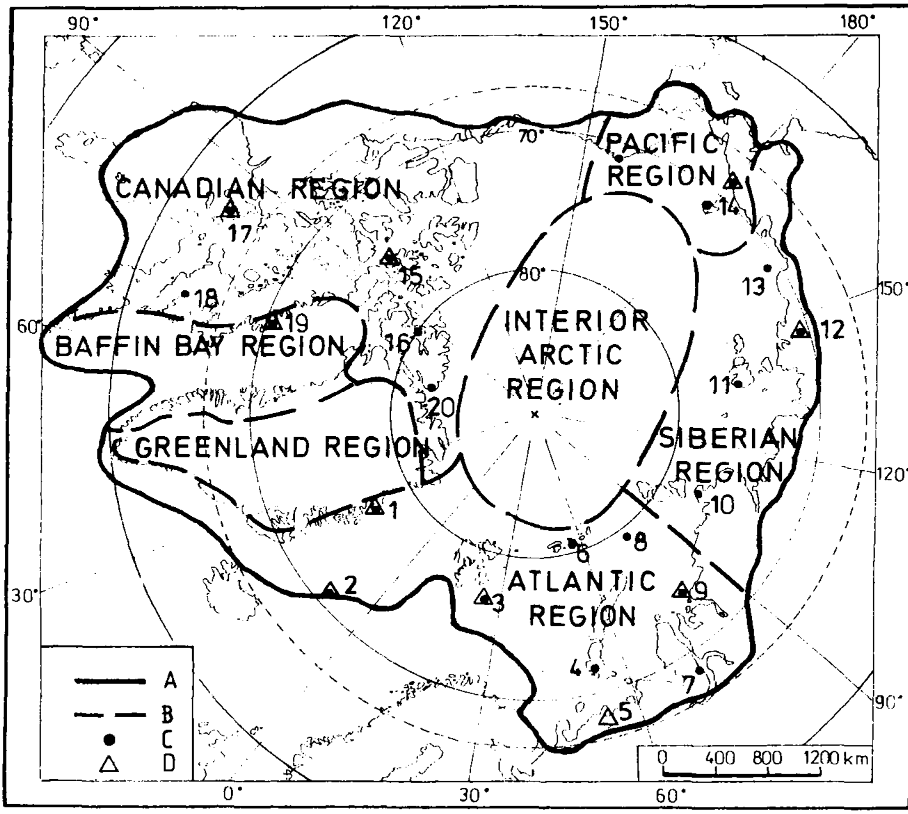


Fig. 1. Location of meteorological stations; borders of the Arctic (A) and climatic regions (B) after Atlas Arktiki (1985), C – stations with mean monthly cloud amount data used, D – stations with daily TMAX, TMIN, DTR and cloud amount data used. 1 – Danmarkshavn (H = 11 m), 2 – Jan Mayen (H = 10 m), 3 – Hopen (H = 6 m), 4 – Malye Karmakuly (H = 46 m), 5 – Naryan-Mar (H = 7 m), 6 – Polar GMO E.T. Krenkelya (H = 20 m), 7 – Mys Kamenny (H = 7 m), 8 – Ostrov Vize (H = 18 m), 9 – Ostrov Dikson (H = 20 m), 10 – GMO E.K. Fedorova (H = 13 m), 11 – Ostrov Kotelny (H = 10 m), 12 – Cokurdah (H = 48 m), 13 – Ostrov Chetyrekhstolbovoy; (H = 6 m), 14 – Mys Szmidta (H = 7 m), 15 – Resolute A (H = 67 m), 16 – Eureka (H = 10 m), 17 – Coral Harbour A (H = 64 m), 18 – Iqaluit A (H = 34 m), 19 – Clyde A (H = 25 m), 20 – Alert (H = 63 m).

may reach up to 45%. In the annual cycle these differences are 2–3 times greater from May to October than in winter. These authors stated that at present 'it is not possible to determine cloud climatology, which is the "correct" one'. Certainly, in this field our knowledge is surprisingly incomplete: such situation is very unfavourable and should be changed in the near future, because:

- polar regions are considered to be of great importance for a global climate (see *e.g.* Polar Group 1980; Arctic Climate System Study 1994), and
- a cloud cover is the major component of the Arctic climate system through its influence on both energy and moisture exchange between the elements of the system *i.e.* atmosphere, ocean, cryosphere, biosphere and lithosphere.

Table 1. Mean seasonal and annual cloudiness (a, in tenths) and their trends (b, in tenths per 10 years) in the Arctic in 1961–1990; seasons are defined as follows: winter (December–January–February is D-J-F), spring (March–April–May is M-A-M), *etc.*

Area: / Seasons:		D-J-F	M-A-M	J-J-A	S-O-N	Annual
Atlantic and Siberian regions	a	6.0	6.6	8.3	7.7	7.2
	b	0.26***	0.10	-0.05	0.00	0.08**
Pacific, Canadian, and western Baffin Bay regions	a	4.6	5.4	7.5	6.9	6.1
	b	0.00	0.12	-0.01	0.00	0.02
Pacific, western and central Canadian regions	a	4.5	5.3	7.5	6.8	6.0
	b	-0.03	-0.01	-0.10	-0.03	-0.06
Whole Arctic	a	5.5	6.2	8.0	7.4	6.8
	b	0.19***	0.13*	-0.03	0.01	0.07*

*, **, *** – trends statistically significant at the levels of 0.05, 0.01, and 0.001, respectively.

There is much less information devoted to the cloudiness trends in the Arctic, in fact only in papers of Raatz (1981) and Przybylak (1996a, 1997). Raatz (1981) used only a set of 7 stations, 2 of which are outside the Arctic (Iceland). Moreover, the other stations represent only a small fragment of the Arctic (Norwegian Arctic, eastern coast of Greenland and Alaska). Raatz (1981) did not find any trends for these parts of the Arctic and for the period 1920–1978. A more comprehensive and reliable analysis of the cloudiness trends in the Arctic during the last decades was presented by Przybylak (1996a, 1997) on the basis of 19 Arctic stations, representing almost all climatic regions in the Arctic delimited in Atlas Arktiki (1985). He found the increasing cloudiness trends for the period 1961–1990 in the European and Russian Arctic, the Baffin Bay and its vicinity *i.e.* the regions with the greatest decrease in DTR. In the rest of the Arctic, where a decrease of cloudiness occurred in the last few decades, changes of DTR were very small and oscillated around its long-term mean. The trends were computed for each analysed station (Przybylak 1996a, 1997). In this paper the areal mean seasonal and annual cloudiness trends for the whole Arctic and for some parts of it are presented (Table 1, Fig. 2).

A mean annual cloudiness in the Arctic is equal to 6.8 (Table 1), the highest in summer (8.0) and the lowest in winter (5.5). These results correspond very well with a total cloudiness in the Arctic (north of latitude 62.5°) by Schweiger and Key (1992: Fig. 1), based on data by Warren *et al.* (1986, 1988). It is noteworthy that a cloudiness in the Arctic is greater than a mean global one, and also both in the northern and the southern hemispheres (comp. Fig. 1 in Parungo *et al.* 1994). Such annual cycle of a total cloudiness is present in all analysed regions of the Arctic (Table 1, Fig. 2).

The statistically significant increase of a mean cloudiness in the Arctic was found for winter, spring and annual values. Changes of a cloudiness in summer and autumn were very small (Table 1). Only in summer cloudiness decreased in all ana-

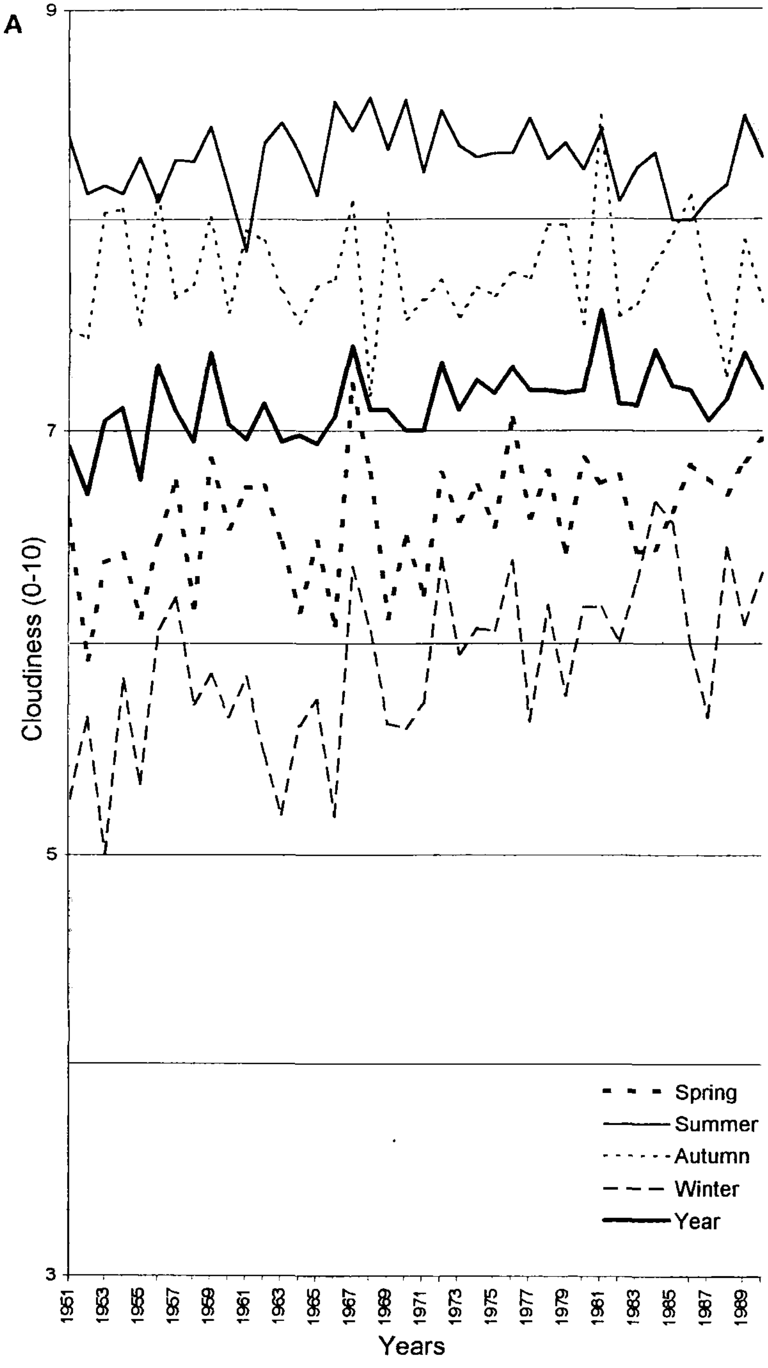
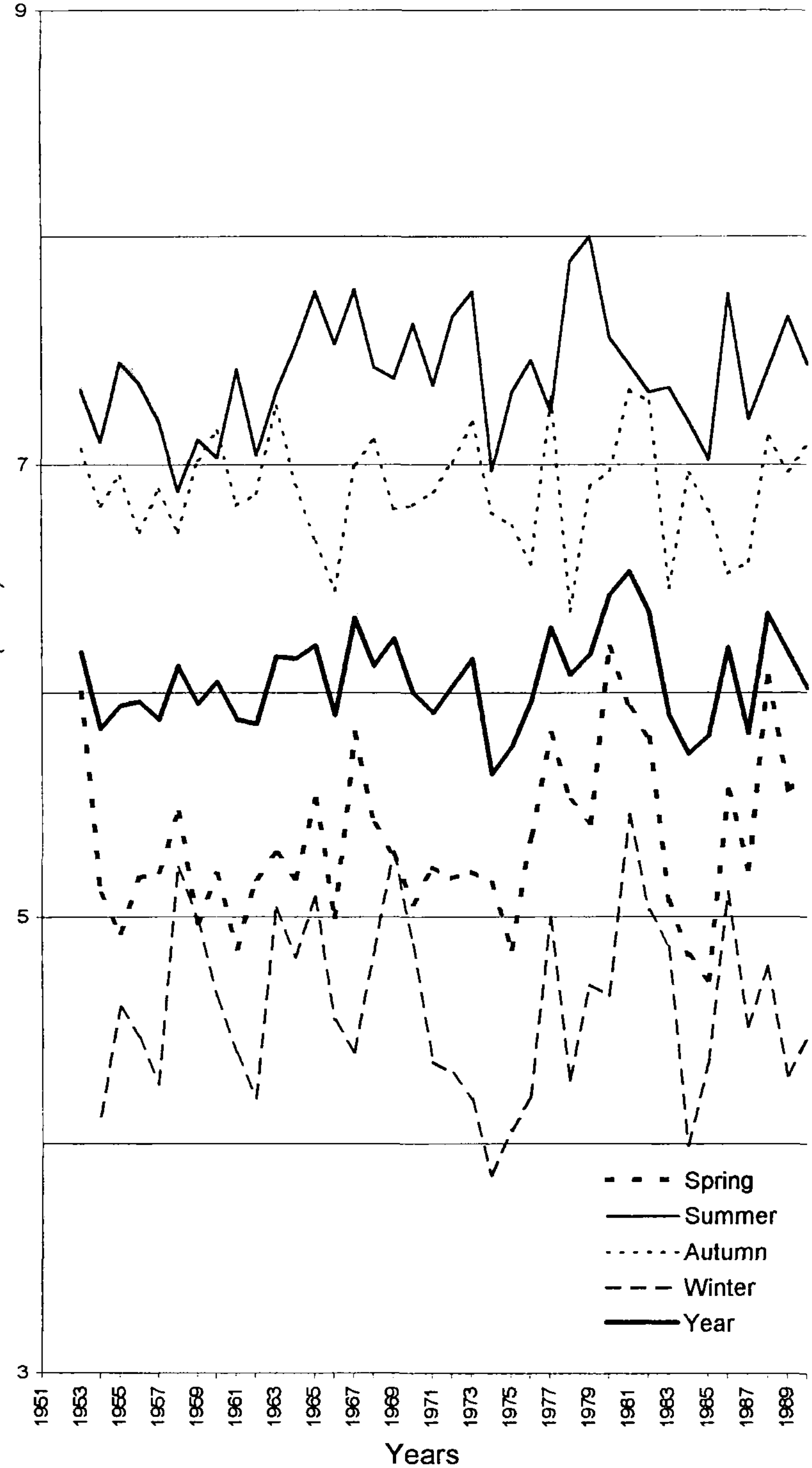


Fig. 2. Year-to-year courses of mean seasonal (DJF, MAM, JJA and SON) and annual cloudiness (in tenths) in 1951–1990. **A**. Atlantic and Siberian regions. **B**. Pacific, Canadian and western part of Baffin Bay regions. **C**. Whole Arctic.

B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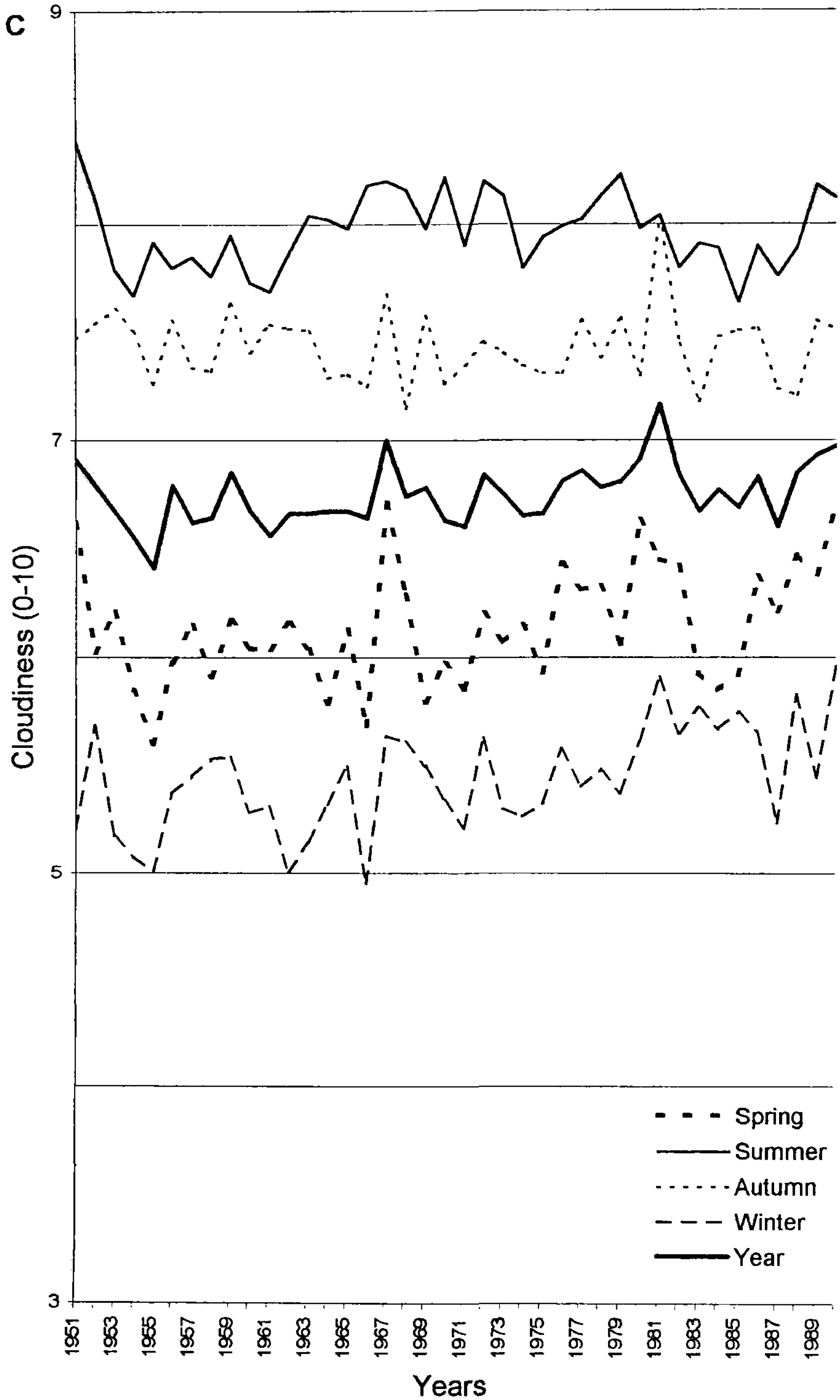
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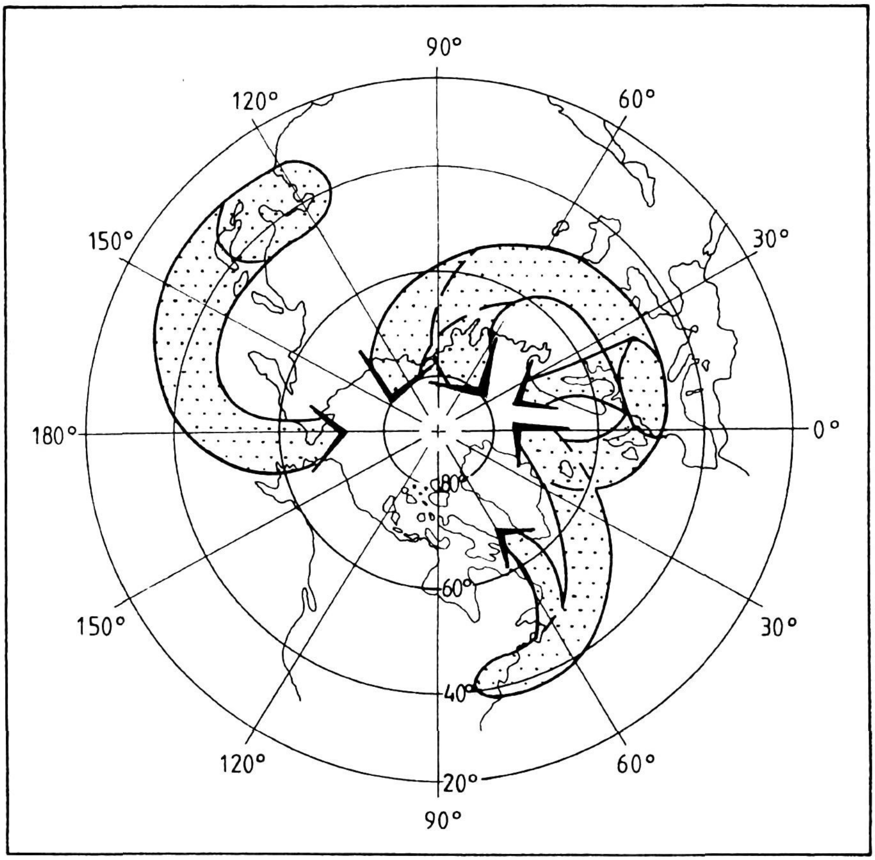

Fig. 3. Major sources and pathways for transport of pollutants between mid-latitudes and the Arctic after Jaworowski (1989).

lysed regions of the Arctic, although it was not statistically significant. The greatest increase of cloudiness was noted in the Atlantic and Siberian regions. Trends for a winter (0.26/10 yrs) and a year (0.08/10 yrs) are statistically significant. Pacific and western and central parts of the Canadian region indicate an insignificant decrease of cloudiness, noted in all the seasons. Comparing these cloudiness trends in the Arctic with similar data for the whole world and the hemispheres (see Parungo *et al.* 1994), an increase of a cloudiness between the fifties and the eighties was greater in the Arctic than for the world, and slightly smaller than for the Northern Hemisphere. Parungo *et al.* (1994) found that for the 30-year period (1952–1981) a change of cloudiness was equal to 1.6% for the whole world, 2.3% for the Northern Hemisphere, and 1.2% for the Southern Hemisphere (if referred to 1952). In the Arctic the year 1981 is not good for such a computation, because of a very great positive anomaly of a cloudiness (0.42), being the highest cloudiness in the Arctic in 1951–1990. Therefore, the two periods i.e. 1952–1980 and 1956–1989 were selected here. An increase of cloudiness in these periods was equal to 1.8 and 2.1%, respectively. Two facts given below are good evidence to connect an increasing cloudiness in the Arctic to a probable significant content of the aerosols, brought into the Arctic atmosphere by incursions of a polluted air from the lower latitudes:

- An increased cloudiness occurred in this part of the Arctic where the air pollution is the greatest (Barrie 1986: Fig. 4). Barrie (1986) wrote that "Eurasian SO_2 sources that are available to the Arctic in winter, are more than double those in North America (54×10^6 vs. 24×10^6t y^{-1}). The difference is even greater in the region to the north of 60° N latitude (6.3 vs. 0.013 Mt. y^{-1})". Confirmation of the Barrie's findings also provides the analysis of the major sources and pathways for transport of pollutants between a temperate zone and the Arctic (Fig. 3).
- The statistically significant increase of a cloudiness occurred only in winter and spring *i.e.* during the seasons when the Arctic atmosphere indicated an increased man-made pollution, originating mainly from mid-latitudinal sources in Eurasia (Barrie 1986). Particular pollution is either absent for the rest of the year (June–November), or it is present at much lower concentration than in winter (*op. cit.*).

Worth noticing are also the results of Stanhil (1995). He computed the trends of a global irradiance in the Arctic and concluded that the statistical significant decrease of the global irradiance in 1950–1994 was caused by incursions of a polluted air to the Arctic.

Relations between cloudiness, extreme air temperatures and diurnal temperature ranges

There are few publications analysing relationships between a cloudiness and the air temperature in the Arctic using daily data (*e.g.* Steffensen 1982; Hanssen-Bauer *et al.* 1990; Przybylak 1992). The first two papers present this relation for the Norwegian Arctic stations but using only a mean daily air temperature. Presumably only Przybylak (1992) made a more comprehensive analysis including also TMAX and TMIN for the Hornsund station in Spitsbergen.

This problem was examined more carefully and for the whole Arctic, in order to confirm or reject some previous conclusions concerning relations between cloudiness and DTR (*cf.* Przybylak 1997) but based on monthly data. Daily data allow finding more reliable and complete relationships.

The daily data sets (TMAX, TMIN, DTR and C) were obtained for relatively very long periods, 40 years (1951–1990) for the Norwegian Arctic stations and 24 years (1967–1990) for the Russian ones. All days, as mentioned earlier, were classified as clear, partly cloudy or cloudy. For these three types of days, mean seasonal and annual anomalies of TMAX, TMIN and DTR were computed (Tables 2–4). Mean annual courses of the studied climatic elements were calculated for clear, partly cloudy and cloudy days and they are presented in Figs 4–6.

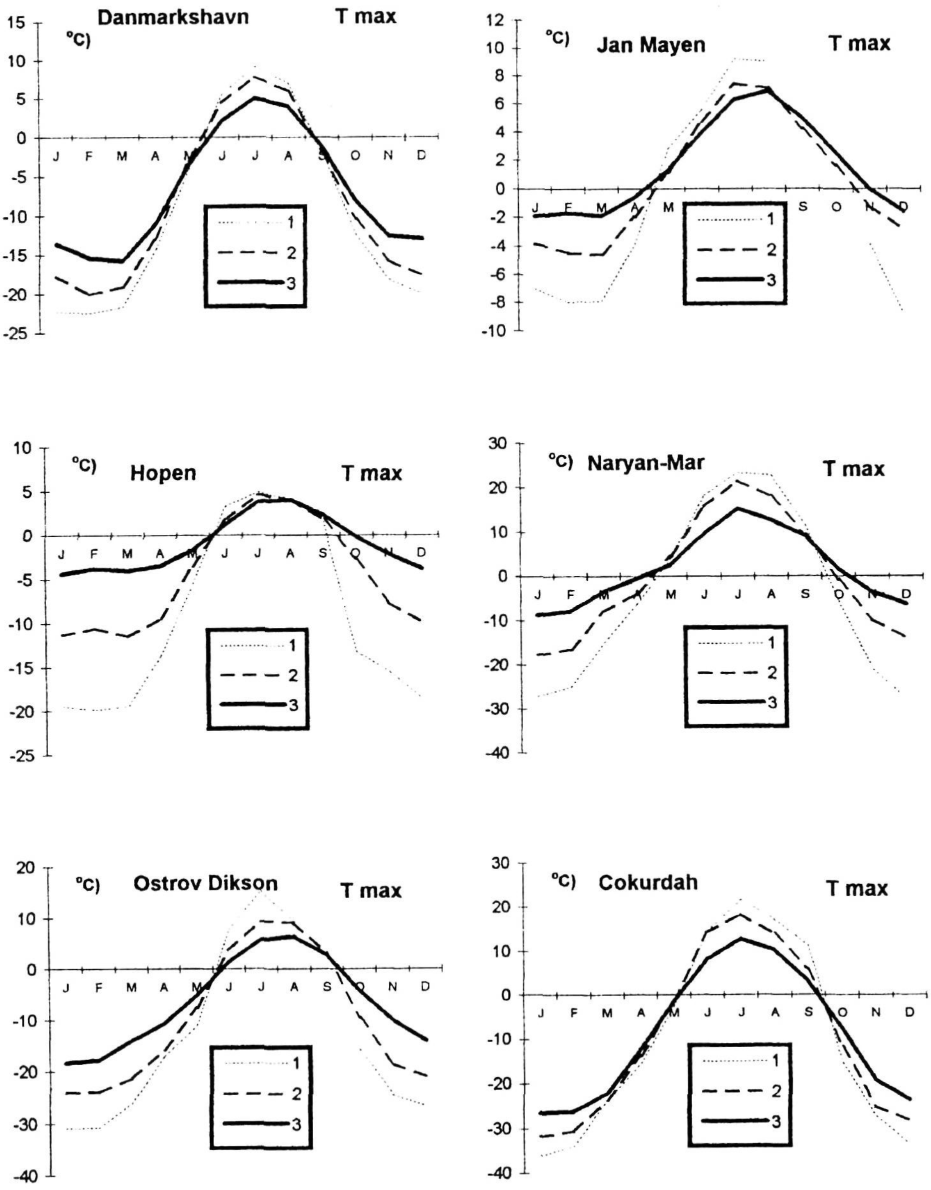

Fig. 4. Mean annual courses of TMAX on clear (1), partly cloudy (2) and cloudy (3) days at 10 Arctic stations representing majority of the climatic regions and subregions after Atlas Arktiki (1985).

Maximum temperatures and cloudiness

Influence of a cloudiness on TMAX is opposite in warm (June–September) and cool (October–May) half-year periods (Table 2, Fig. 4). In summer the highest

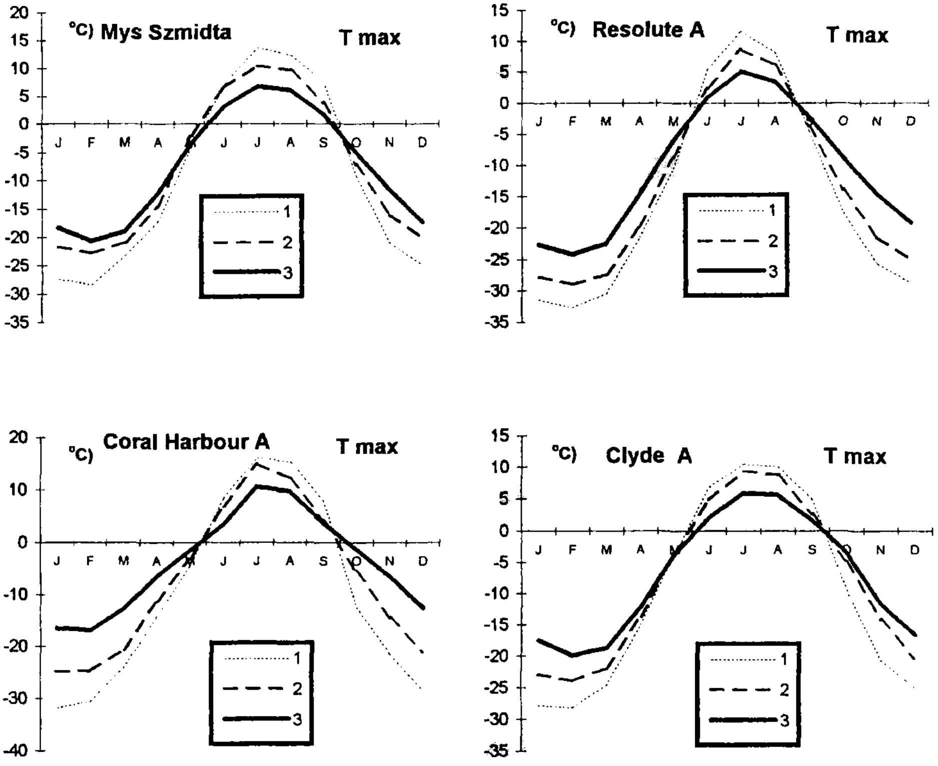

Fig. 4 *continued.*

TMAX is connected with clear days and the lowest with cloudy days. Positive anomalies during clear days (cf. Table 2) are especially high (3–7°C) in the most continental part of the Russian and Canadian Arctic. They are much lower in the western and central parts of the Atlantic region (1–2°C). An increase of a cloudiness in summer leads to a cooling of the whole Arctic, but especially of these parts of the Arctic which are located near its southern border and are characterised by a high continentality: the stations Naryan-Mar, Cokurdah and Coral Harbour A. For these stations the mean differences of TMAX between clear and cloudy days vary from 5 to 7°C, while in the Norwegian Arctic (maritime climate) from 1 to 2°C only (see Table 2 and Fig. 4). In the cool half-year an influence of cloudiness on TMAX is opposite to the one in summer *i.e.* an increasing cloudiness leads to a warming of the Arctic. The positive anomalies of TMAX during the cloudy days are the highest in winter (above 4°C) almost in the whole Arctic, except for the regions represented by the stations Jan Mayen and Mys Szmidta. It is noteworthy that most of the Arctic (excluding Siberian and the western part of Atlantic region) has higher positive anomalies on cloudy days in spring rather than in autumn. On clear days the highest negative anomalies occurred in autumn, except for the stations Jan

Table 2. Mean seasonal anomalies of TMAX (in °C) in the Arctic on clear (1), partly cloudy (2) and cloudy (3) days in 1951–1990; DAN – Danmarkshavn, JAN – Jan Mayen, HOP – Hopen, NAR – Naryan-Mar, DIK – Ostrov Dikson, CZO – Cokurdah, SZM – Mys Szmidta, RES – Resolute A, COR – Coral Harbour A, CLY – Clyde A; 1 – C<2; 2 – 2≤C≤ 8; 3 – C>8; Mean – mean seasonal TMAX; seasons are defined as follows: winter (December–January–February is D-J-F), spring (March–April–May is M-A-M), *etc.*

Season	Element	DAN#	JAN	HOP	NAR*	DIK*	CZO*	SZM*	RES**	COR**	CLY**
D-J-F	1	-3.1	-5.2	-10.1	-14.1	-8.4	-4.8	-5.9	-3.2	-6.3	-4.3
	2	0.0	-1.1	-1.4	-3.8	-1.7	-0.6	-0.4	0.5	0.4	0.3
	3	4.5	0.9	5.2	4.9	4.8	4.0	2.5	5.8	8.9	4.8
	Mean	**-18.5**	**-2.6**	**-9.2**	**-12.4**	**-21.4**	**-29.7**	**-21.2**	**-27.8**	**-24.0**	**-22.6**
M-A-M	1	-2.3	-0.2	-8.9	-7.2	-8.7	-4.6	-6.1	-5.6	-6.0	-4.5
	2	0.1	-1.1	-1.9	-0.9	-3.5	-0.3	-1.0	-1.0	-0.6	-0.3
	3	2.2	0.7	3.8	1.7	4.0	3.6	3.1	7.8	6.3	4.4
	Mean	**-11.7**	**-1.0**	**-6.7**	**-1.9**	**-12.9**	**-12.7**	**-12.6**	**-18.4**	**-11.5**	**-13.1**
J-J-A	1	2.0	1.7	1.0	6.1	7.1	4.7	3.9	3.4	2.8	2.6
	2	0.8	0.5	0.2	3.4	2.8	2.9	2.1	1.4	1.3	1.4
	3	-1.7	-0.2	-0.1	-2.8	-0.9	-2.2	-1.3	-1.3	-2.4	-1.8
	Mean	**5.3**	**5.9**	**3.2**	**15.2**	**5.3**	**12.6**	**6.7**	**4.4**	**10.1**	**6.3**
S-O-N	1	-3.7	-3.1	-12.4	-7.1	-15.4	-10.9	-6.6	-10.8	-12.4	-10.3
	2	0.1	-0.7	-2.5	-1.5	-6.2	-3.5	-2.0	-2.6	-0.9	-0.9
	3	3.0	0.3	1.6	1.2	3.6	4.1	1.5	5.8	3.6	2.1
	Mean	**-9.4**	**2.2**	**-1.4**	**1.2**	**-6.3**	**-9.7**	**-5.8**	**-12.0**	**-4.4**	**-5.5**

* – in 1967–1990, # – in 1955–1990, ** – in 1953–1990.

Mayen and Naryan-Mar. In the annual course the lowest differentiated influence of cloudiness on TMAX occurs at the turn of May/June and September/October when the described relations between a cloudiness and TMAX change rapidly from one mode to another (see Fig. 4).

Minimum temperatures and cloudiness

Generally speaking, an influence of a cloudiness on TMIN is roughly similar to the one on TMAX, but there exist also several important differences (see Tables 2 and 3, and Figs 4 and 5). One of them is the opposite influence of cloudiness on TMIN than on TMAX in summer in the Norwegian Arctic and in the southern Canadian Arctic. During this season TMIN is higher on cloudy than clear days (see Table 3 and Fig. 5). Moreover, positive anomalies of TMIN in the rest of the Arctic are significantly (2–3 or more times) lower than anomalies of TMAX (compare Tables 2 and 3, and Figs 4 and 5).

In winter an influence of cloudiness on both TMAX and TMIN is very similar, but negative anomalies on clear days are in the most Arctic lower in the case of TMIN. In spring an influence of cloudiness is significantly greater on TMIN than

Table 3. Mean seasonal anomalies of TMIN (in °C) in the Arctic on clear (1), partly cloudy (2) and cloudy (3) days in 1951–1990; DAN – Danmarkshavn, JAN – Jan Mayen, HOP – Hopen, NAR – Naryan-Mar, DIK – Ostrov Dikson, CZO – Cokurdah, SZM – Mys Szmidta, RES – Resolute A, COR – Coral Harbour A, CLY – Clyde A; 1 – $C<2$; 2 – $2\leq C\leq 8$; 3 – $C>8$; Mean – mean seasonal TMIN; seasons are defined as follows: winter (December–January–February is D-J-F), spring (March–April–May is M-A-M), *etc.*

Season	Element	DAN#	JAN	HOP	NAR*	DIK*	CZO*	SZM*	RES**	COR**	CLY**
D-J-F	1	-3.3	-5.2	-8.4	-12.9	-6.9	-3.9	-6.1	-2.5	-4.9	-3.9
	2	-0.3	-1.3	-1.9	-5.5	-1.8	-1.1	-1.1	0.2	-0.3	-0.3
	3	5.4	1.0	5.2	6.1	4.6	4.1	3.4	5.8	9.2	6.1
	Mean	**-27.1**	**-7.8**	**-15.9**	**-21.3**	**-28.5**	**-36.7**	**-28.2**	**-35.0**	**-32.3**	**-30.4**
M-A-M	1	-3.7	-1.2	-8.8	-10.7	-9.1	-5.8	-8.9	-6.1	-7.5	-5.8
	2	-0.2	-1.6	-2.4	-2.7	-4.0	-0.9	-2.1	-1.3	-1.4	-1.1
	3	4.1	1.0	4.3	3.6	4.6	5.3	5.2	9.0	9.2	7.0
	Mean	**-19.7**	**-5.5**	**-12.3**	**-10.4**	**-19.8**	**-21.3**	**-20.0**	**-25.5**	**-21.2**	**-22.4**
J-J-A	1	0.8	-0.7	-0.3	1.8	4.4	1.6	0.9	1.5	-0.2	0.3
	2	0.0	-0.6	-0.3	1.1	1.1	0.9	0.5	0.3	0.0	0.0
	3	-0.3	0.2	0.1	-0.9	-0.4	-0.7	-0.3	-0.3	0.0	-0.1
	Mean	**-0.3**	2.2	0.1	6.4	0.9	3.7	0.9	-0.5	2.1	-0.4
S-O-N	1	-4.2	-2.4	-13.0	-9.6	-15.1	-11.6	-8.9	-11.4	-12.5	-11.5
	2	-0.3	-0.9	-3.1	-3.3	-7.4	-4.6	-4.0	-3.4	-2.2	-2.2
	3	4.2	0.4	1.9	2.4	4.1	4.9	2.6	6.9	5.3	3.6
	Mean	**-15.6**	**-1.6**	**-5.1**	**-5.2**	**-11.3**	**-15.8**	**-11.3**	**-17.8**	**-11.5**	**-11.3**

* – in 1967–1990, # – in 1955–1990, ** – in 1953–1990.

on TMAX. Negative (positive) anomalies of TMIN on clear (cloudy) days are clearly greater than TMAX. It means that an increasing cloudiness results in a much greater rise of TMIN than TMAX during this season. Similar situation is also present in autumn, although it is expressed slightly weaker than in spring.

Diurnal temperature ranges and cloudiness

Basing on the above-presented results, a general pattern of influence of cloudiness on TMAX and TMIN seems to be quite similar. However, the existing differences in magnitudes of this influence (expressed by anomalies) are significant during some seasons and should cause appropriate changes of DTR in case of increasing or decreasing cloudiness in the Arctic.

An influence of cloudiness on DTR is presented (Table 4, Fig. 6). These data clearly show that on the annual basis an increasing cloudiness leads to a decreased DTR. This influence is the highest in summer, then in spring and also in autumn, except for the region of Jan Mayen. In winter the situation is much more complicated, because the highest positive anomalies of DTR occur on the partly cloudy days in the whole Arctic. Further increase of cloudiness leads to a de-

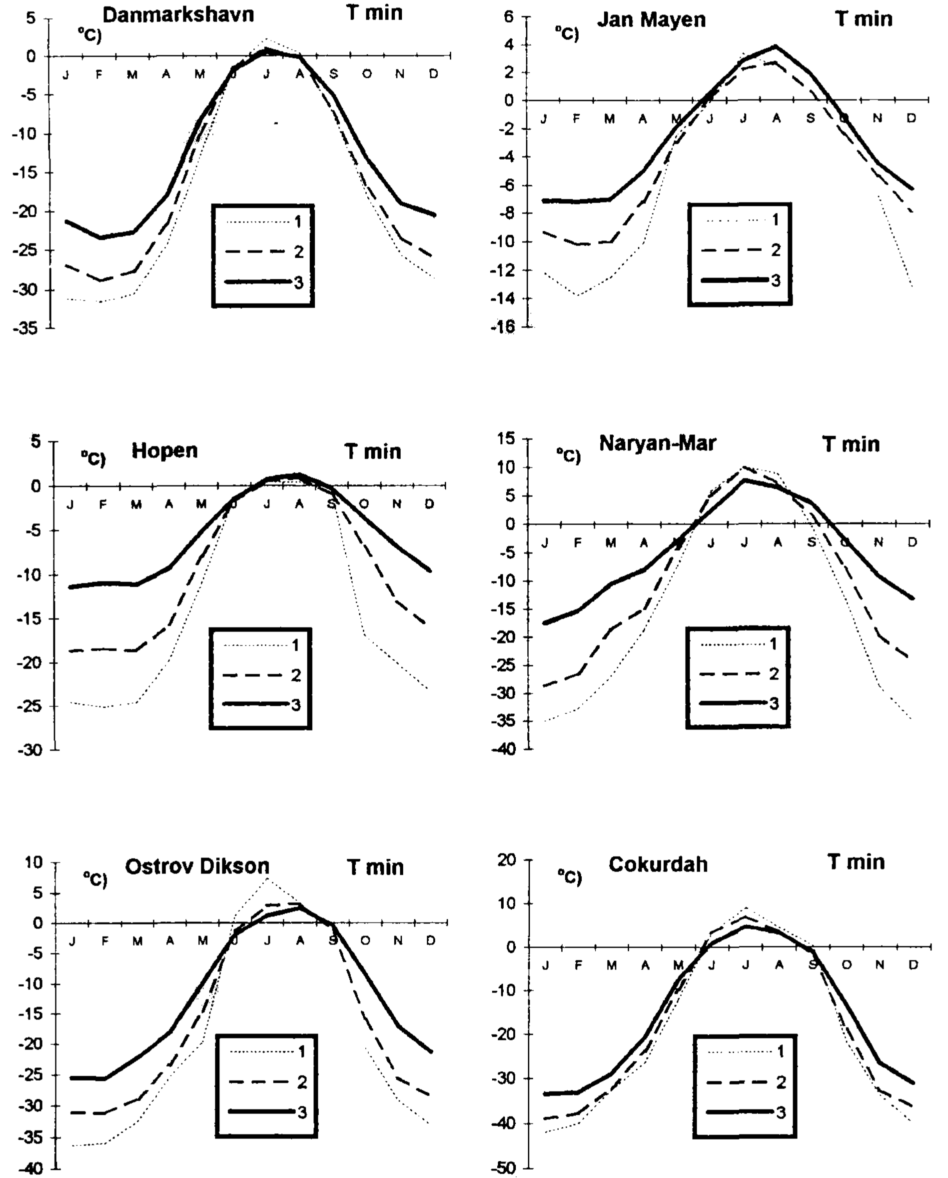

Fig. 5. Mean annual courses of TMIN on clear (1), partly cloudy (2) and cloudy (3) days at 10 Arctic stations representing majority of the climatic regions and subregions after Atlas Arktiki (1985).

crease of DTR. Slightly positive anomalies of DTR occur also in some parts of the Arctic, both on clear (Danmarkshavn and Mys Szmidta) and cloudy (Ostrov Dikson and Cokurdah) days. It is noteworthy that in the Arctic where cyclonic activity prevailed (Atlantic, Pacific and Baffin Bay regions), the anomalies of DTR during cloudy days were lower than during clear days (see Table 4 and Fig.

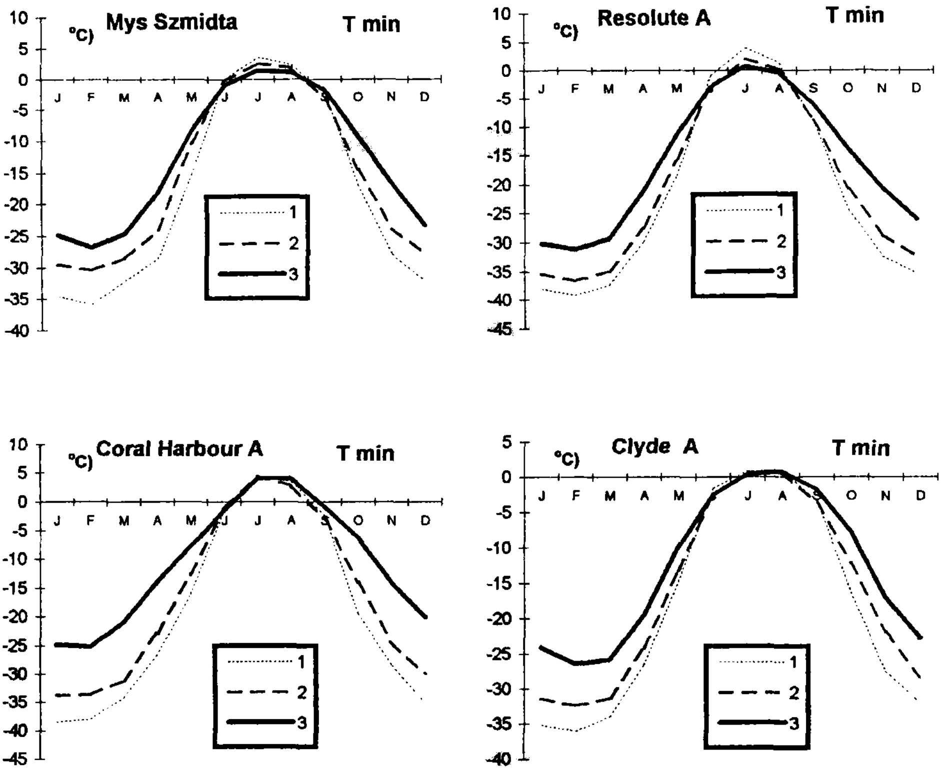

Fig. 5 *continued.*

6). An opposite relation occurs in the Siberian and Canadian regions (with prevailing anticyclones) where radiation plays a significantly greater role than in the previously mentioned regions. On cloudy days there (Tables 2–3), the anomalies of TMAX and TMIN are nearly the same, but on clear days negative anomalies of TMAX are significantly greater. In the regions where much cloudiness is transported together with warm and humid air masses from the lower latitudes by a synoptic-scale cyclonic activity, an influence of a cloudiness on TMAX and TMIN is different, mainly on cloudy days (opposite than in the Canadian and Siberian regions). This is a consequence of the very similar influence of advection of warm air masses on TMAX and TMIN. On the other hand, a high cloudiness connected with this advection reduces a loss of the long-wave outgoing radiation to the space. This radiation flux is relatively more important during a night than a day, and the resulting positive anomalies of TMIN are higher than TMAX (see Tables 2–3).

This analysis of the daily cloudiness and DTR data confirms entirely the previous conclusions drawn from monthly data (Przybylak, 1997, 1999).

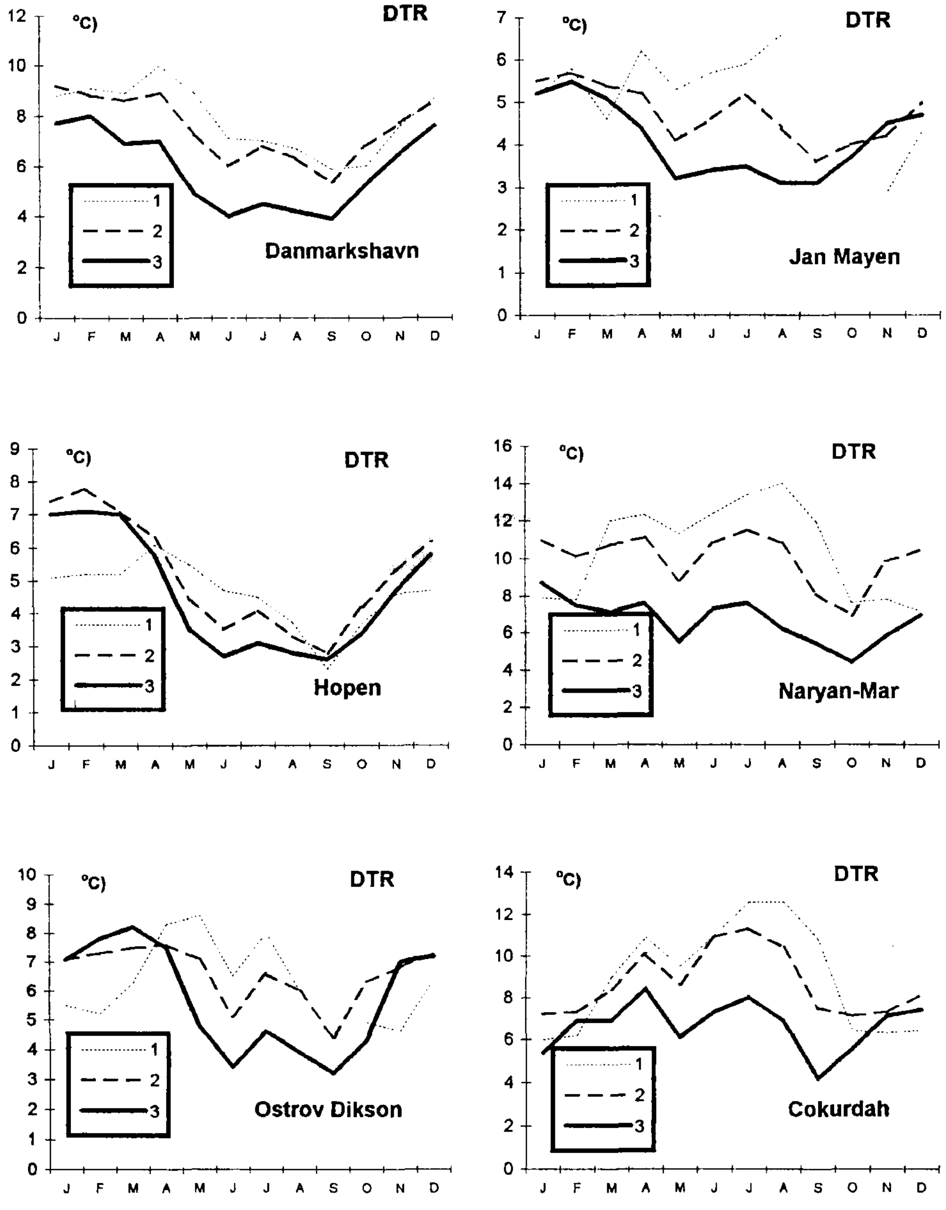

Fig. 6. Mean annual courses of DTR on clear (1), partly cloudy (2) and cloudy (3) days at 10 Arctic stations representing majority of the climatic regions and subregions after Atlas Arktiki (1985).

Conclusions

Our knowledge about the cloudiness in the Arctic is still incomplete. There are great differences between surface-based and satellite-derived cloud climatology.

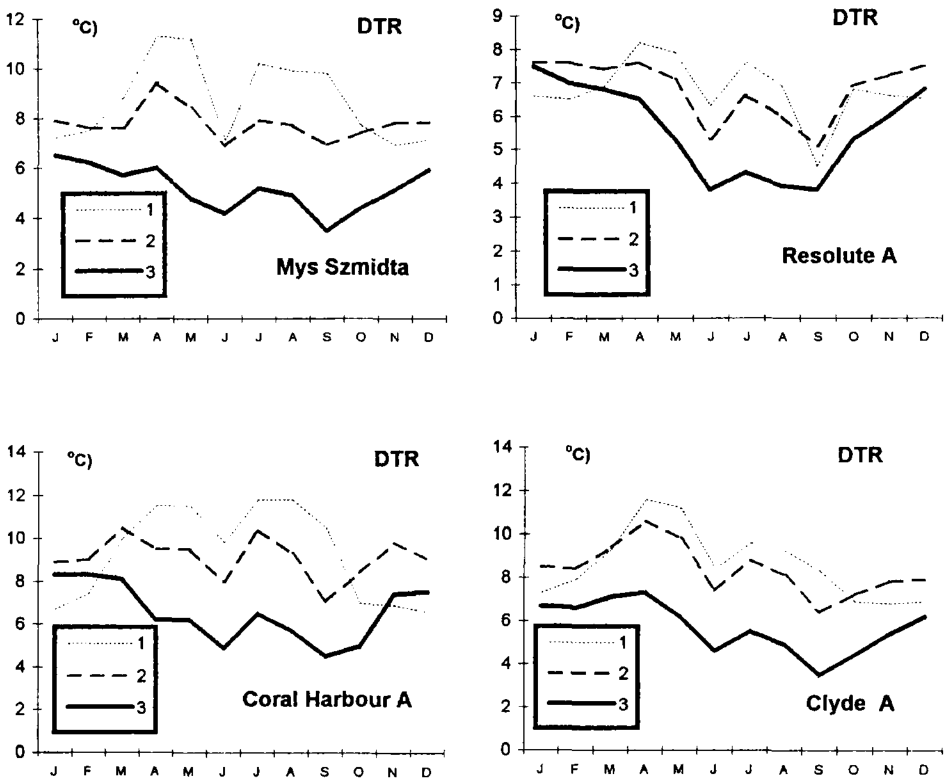

Fig. 6 *continued.*

At present we do not know even which climatology is correct. Particularly few papers were devoted to the recent trends in cloudiness in the Arctic.

A mean cloudiness in the Arctic (based on the average data from 19 stations) for the period 1961–1990 is equal to 6.8. The highest cloudiness predominates in summer (8.0) and the lowest in winter (5.5). The statistically significant increase of a mean cloudiness in the Arctic was present in winter, spring and for the year. The greatest increase of cloudiness occurred in the Atlantic and the Siberian regions. Only in summer an insignificant decrease of cloudiness occurred in all analysed regions of the Arctic. Both geographic (Atlantic and Siberian regions) and time (winter and spring) patterns of the increasing cloudiness during the recent decades support the conclusion that they can be caused by the aerosols brought into the Arctic by incursion of a polluted air from the lower latitudes.

A conducted analysis revealed that influence of a cloudiness on TMAX is different during warm (June–September) and cool (October–May) half-year seasons. In the first period the highest TMAX occurs on clear days and the lowest one on cloudy days. The opposite is true for the second period. It seems interesting that on

Table 4. Mean seasonal anomalies of the DTR (in °C) in the Arctic on clear (1), partly cloudy (2) and cloudy (3) days in 1951–1990; DAN – Danmarkshavn, JAN – Jan Mayen, HOP – Hopen, NAR – Naryan-Mar, DIK – Ostrov Dikson, CZO – Cokurdah, SZM – Mys Szmidta, RES – Resolute A, COR – Coral Harbour A, CLY – Clyde A; 1 – C<2; 2 – 2≤C≤8; 3 – C>8; Mean – mean seasonal DTR; seasons are defined as follows: winter (December–January–February is D-J-F), spring (March–April–May is M-A-M), *etc.*

Season	Element	DAN#	JAN	HOP	NAR*	DIK*	CZO*	SZM*	RES**	COR**	CLY**
D-J-F	1	0.2	0.0	-1.7	-1.2	-1.5	-0.8	0.3	-0.7	-1.4	-0.5
	2	0.3	0.2	0.4	1.7	0.2	0.6	0.8	0.3	0.6	0.5
	3	-0.8	-0.2	-0.1	-1.3	0.2	0.1	-0.8	-0.1	-0.3	-1.3
	Mean	**8.6**	**5.2**	**6.7**	**8.9**	**7.1**	**7.0**	**7.0**	**7.2**	**8.3**	**7.8**
M-A-M	1	1.3	1.0	0.0	3.5	0.3	1.2	2.8	0.4	1.5	1.1
	2	0.3	0.5	0.6	1.8	0.5	0.5	1.1	0.3	0.8	0.7
	3	-1.9	-0.3	-0.5	-1.9	-0.5	-1.7	-2.1	-1.3	-2.9	-2.6
	Mean	**8.0**	**4.5**	**5.6**	**8.5**	**6.9**	**8.6**	**7.4**	**7.1**	**9.7**	**9.3**
J-J-A	1	1.3	2.3	1.2	4.3	2.8	3.1	3.0	1.9	3.1	2.3
	2	0.8	1.0	0.5	2.3	1.7	2.0	1.6	1.1	1.3	1.4
	3	-1.4	-0.4	-0.2	-1.8	-0.4	-1.5	-1.0	-0.9	-2.4	-1.7
	Mean	**5.6**	**3.7**	**3.1**	**8.8**	**4.4**	**8.9**	**5.8**	**4.9**	**8.0**	**6.7**
S-O-N	1	0.5	-0.7	0.6	2.4	-0.2	0.7	2.1	0.6	0.2	1.2
	2	0.4	0.2	0.6	1.8	1.2	1.1	1.9	0.9	1.3	1.5
	3	-0.8	0.0	-0.3	-1.2	-0.5	-0.7	-1.2	-1.1	-1.7	-1.5
	Mean	**6.2**	**3.8**	**3.7**	**6.4**	**5.0**	**6.1**	**5.5**	**5.8**	**7.1**	**5.8**

* in 1967–1990, # – in 1955–1990, ** – in 1953–1990.

clear days the highest negative anomalies occurred in autumn almost in the whole Arctic.

Generally speaking, an influence of cloudiness on TMIN is roughly similar to the one on TMAX. Some important differences were however noted *e.g.* the opposite influence of cloudiness on TMIN and TMAX in summer in the Norwegian and the southern Canadian Arctic. In the rest of the Arctic there were sometimes significant differences in the anomalies computed for the three analysed cloudiness daily types.

Results of this analysis prove that an increasing cloudiness in the Arctic leads to a decrease of DTR. This relation is the strongest in summer, weaker in spring and also in autumn, except for the western part of Atlantic region represented by the station Jan Mayen. In winter the relations are much more complicated; the highest positive anomalies of DTR occur during the partly cloudy days. The slightly positive anomalies occur also in some parts of the Arctic, both on clear (Danmarkshavn and Mys Szmidta) and cloudy (Ostrov Dikson and Cokurdah) days. Parts of the Arctic with the highest cyclonic activity (Atlantic, Pacific and Baffin Bay regions) indicate the anomalies of DTR on cloudy days lower than on clear days. The opposite is true for the Siberian and Canadian regions where the anticyclonic activity prevailed. So, during this season the factors other than an increasing cloudiness *e.g.* non-periodic day-to-day changes of air temperature controlled by thermal advection associated with synoptic-scale cyclones and anticyclones, must play also a significant role, probably greater than a cloudiness (cf. Przybylak 1999).

The results confirm entirely the earlier conclusions (Przybylak 1997, 1999) that have been derived from the analysis of the monthly cloudiness and DTR data. Additional analyses of relations between cloudiness and extreme air temperatures presented in the paper allowed estimating more reliably the factors that determine changes of DTR in the Arctic.

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Streszczenie

Przedstawiono związki między zachmurzeniem (C) a temperaturami ekstremalnymi (TMAX i TMIN) oraz dobową amplitudą temperatury powietrza (DTR) w Arktyce w okresie 1951–1990. Analizę trendów zachmurzenia w Arktyce w okresie ostatnich kilku dekad wykonano wykorzystując średnie miesięczne dane z 19 stacji arktycznych. Natomiast związki statystyczne między zachmurzeniem i wspomnianymi charakterystykami termicznymi obliczono na podstawie danych dobowych zgromadzonych dla 10 stacji reprezentujących wszystkie regiony klimatyczne Arktyki (fig. 1). Większość danych uzyskano bezpośrednio z instytutów meteorologicznych poszczególnych państw arktycznych (Danii, Kanady i Norwegii) oraz Instytutu Naukowo-Badawczego Arktyki i Antarktyki w Sankt Petersburgu i Narodowego Centrum Danych Klimatycznych w Asheville. Kontrola jakości analizowanych danych została przeprowadzona przez Przybylaka (1996a).

Średnie sezonowe i roczne wielkości zachmurzenia dla Arktyki jako całości oraz dla jej poszczególnych regionów klimatycznych zostały obliczone przez zwykłe uśrednianie danych ze stacji (bez stosowania wag). Na podstawie średniego dobowego zachmurzenia zaklasyfikowano, osobno dla każdej z analizowanych stacji, każdy dzień badanego wielolecia do jednej z 3 kategorii: dni pogodnych (C<2), dni chmurnych (2≤C≤8) i dni pochmurnych (C>8). Następnie dla każdej stacji i dla każdej z wyróżnionych kategorii dni obliczono średnie miesięczne i sezonowe anomalie TMAX, TMIN i DTR.

Średnia wieloletnia roczna wartość zachmurzenia w Arktyce wyniosła 6,8 (tab. 1). Najwyższe zachmurzenie wystąpiło w lecie (C = 8,0), a najniższe w zimie (C = 5,5). Statystycznie istotny wzrost zachmurzenia w Arktyce w ostatnich kilku dekadach wystąpił w zimie, wiosną i średnio dla roku (tab. 1, fig. 2). Zmiany zachmurzenia w lecie i jesienią były natomiast niewielkie. Największy wzrost zachmurzenia odnotowano w regionach: atlantyckim i syberyjskim. Średnie zimowe i roczne wartości wykazały tutaj statystycznie istotny wzrost. Spadek zachmurzenia, statystycznie jednak nieistotny, wystąpił w regionie pacyficznym oraz w centralnej i wschodniej części regionu kanadyjskiego. Zarówno rozkład geograficzny, jak i czasowy obserwowanych zmian zachmurzenia w Arktyce wskazuje, iż ich przyczyną może być transport zanieczyszczonego powietrza z niskich szerokości geograficznych zgodnie z przedstawionym schematem (fig. 3).

Przeprowadzona analiza wykazała, że wpływ zachmurzenia na TMAX jest różny w ciepłej (VI–IX) i chłodnej (X–V) porze roku. W pierwszym okresie najwyższe TMAX występowały w dniach pogodnych a najniższe w dniach pochmurnych (tab. 2, fig. 4). Odwrotne relacje stwierdzono w drugim okresie. Wpływ zachmurzenia na TMIN w Arktyce jest, z grubsza biorąc, podobny jak w przypadku TMAX (por. tab. 2 i 3 oraz fig. 4 i 5). Występują jednak pewne różnice: na przykład w Arktyce Norweskiej i południowej części Arktyki Kanadyjskiej stwierdzono odwrotny wpływ zachmurzenia na TMAX i TMIN w lecie.

Wyniki analizy wykazały, że wzrost zachmurzenia prowadzi do spadku DTR (tab. 4, fig. 6). Jest to najlepiej widoczne w lecie, słabiej na wiosnę i w jesieni. Natomiast w zimie relacje są bardziej złożone. Najwyższe pozytywne anomalie stwierdzono dla dni chmurnych. Niewielkie pozytywne anomalie DTR mogą występować zarówno w dniach pogodnych (Danmarkshavn i Mys Szmidta), jak i pochmurnych (Ostrow Dikson i Czokurdach). Obszar Arktyki o najsilniejszej działalności cyklonalnej (regiony: atlantycki, pacyficzny i Morza Baffina) posiadają niższe anomalie w dniach pochmurnych niż pogodnych. Odwrotne relacje występują natomiast w regionach syberyjskim i kanadyjskim, gdzie przeważają układy antycyklonalne. Świadczy to, że o DTR decydują w zimie inne czynniki niż zachmurzenie. Wydaje się, że jednym z ważniejszych są nieokresowe zmiany temperatury powietrza z dnia na dzień, warunkowane bardzo zmienną w tym czasie cyrkulacją atmosferyczną.