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CONTENTS

INTRODUCTION

Climate changes have been occurring on the Earth since it came into existence; however, in relation to a life of an individual human being they were very slow. Probably they wouldn't have been noticeable for one human generation in any period of time if people had existed then. These changes were also going on in the past millennia. Some fluctuations of the climate conditions can be observed or concluded on the basis of specialist research and also on the basis of historical sources. However, so far it has never been hypothesized that the reason for those changes could have been the then human activity.

The research on climate change, similarly to other types of research within the field of natural science, is particularly difficult due to the necessity of taking into consideration the recurrence of changes and a great variability of the processes occurring in nature. According to common knowledge, the researchers often come across such opinions even published in press, a snowy winter or a cool summer contradict the climate warming. People tend to forget that annual weather observation does not tell us anything about the long-term changes. They tend to forget, for instance, extraordinarily warm January two or three years earlier or a sequence of years with significantly higher temperatures. Also it is neglected that actually we can talk about climate changes only after keeping measurement records for a few dozen years, or changes in the environment recorded throughout a few dozen of years. Turbulent times in Poland in the 20th century made obtaining long-term record data very difficult and finding the phenological observations, which would show the changes in the environment independently on human activity, is even harder. Moreover, there are no glaciers or permafrost whose melting would indicate evident temperature changes. Summing up, the climate changes in Poland are discussed mostly basing on the analyses of measurement data of various meteorological parameters. In the following monograph, a team of researchers from the Institute of Meteorology and Water Management (IMGW) in Poznań presented the analyses of variability of precipitation, air temperature and insolation on the basis of the data gathered in the period between years 1966-2005 from 14 measurement stations located in Wielkopolska region. Subsequent chapters submitted by teams of researchers from the universities in Kraków, Szczecin and Wrocław showed various methods of precipitation variability analysis regarding climate changes. Since meteorological droughts can be a natural result of unfavorable precipitation pattern, their analyses were presented in Chapter 5. Quite a good indicator of climate changes seems to be snowfall and the period of its lingering, particularly that in most of the lowlands in Poland the periods of frost intermingle with thaws. This issue was addressed by teams from Poznań, Kraków and Śląsk. The results of the research were presented in Chapters 6 and 7. It is impossible to speak of climate changes without mentioning the exchange of energy and mass between the surface and the atmosphere. The relevant research results were obtained by the team from Poznań and were presented in Chapter 8. In order to consider climate change in the near and distant future, it is necessary to simulate it. The last chapter of the following monograph was devoted to exploring the possibility of applying the WRF model in climate research.

The present publication contains only selected issues related to research on climate changes observed in Poland. However, it seems that, together with other publications on that subject, it will be a good supplement and extension of knowledge regarding the methods of investigating into climate changes and the observed results of these changes.

Dr Jacek Leśny

1. VARIABILITY OF SELECTED CLIMATIC INDICES DURING VEGETATION PERIOD IN WIELKOPOLSKA

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INTRODUCTION

Moderate climate is thought to generate one of the most favorable conditions for agriculture. This, however, is not entirely true, since it is characterized by periodical occurrences of phenomena that disrupt plant growth and result in lower yields. These phenomena are: rainfall shortages causing drought or rainfall surpluses which induces rotting process, development of plant diseases and pests; low winter temperatures combined with insufficient snow cover causing freezing of winter crops; ground-frosts that damage plants and consequently lower yields; too high air temperatures in summer that disrupt physiological processes, etc. (Atlas of climatic… 2001). At the same time, climatologists alarm about intensification of extraordinary meteorological phenomena and warming processes (Climate Change… 2007, Kożuchowski and Degirmendžić 2005, Żmudzka 2009). In the context of the observed and projected climate change, it is necessary to monitor those climate elements that are critical for agricultural production. The effects of warming developing for the last two decades are evident especially in areas of intensive agricultural production. Wielkopolska region is one of such areas. This is an industrial-agricultural region with highly developed agriculture and high yields. Its location, however, makes it the most vulnerable area to drought events in Poland (Farat R. *et al.* 1994, Farat R. *et al.* 1995). The exceptional drought vulnerability of this region is a result of several additional natural and anthropogenic factors such as dominance of light soils, deforestation, highly developed drainage network, too low retention level, etc. (Kowalczak *et al*. 2006, Kowalczak *et al.* 2007). If the observed adverse climate conditions continue to develop, it will be necessary to adapt agricultural production to new environmental conditions (Adapting to climate… 2007, Demidowicz *et al.* 1999).

One of the most evident effects of climate change that is of direct concern for a successful agricultural production in the Wielkopolska region, is rainfall shortage. This paper presents the analysis of selected climate indices characterizing water conditions for growing plants in this part of Poland.

MATERIAL AND METHODS

The study was conducted in the Wielkopolska region. Because of high compatibility of historical borders with the present administrative borders, spatial distribution of the analyzed indices is presented within the administrative borders of the Wielkopolska Voivodeship.

In order to analyze spatial distribution of climate elements, data collected at 14 synoptic stations (5 in Wielkopolska area and 9 in its closest vicinity) were used.

The eight indices characterizing three climate elements i.e. temperature, rainfall and sunshine duration were analyzed. Pluvial indices (rainfall measurements taken 1 m above the ground level) included:

- rainfall (mm),
- number of days with rainfall of ≥ 3 mm,
- number of days with rainfall ≥ 20 mm.

The analysis of thermal indices included (measurements taken 2 m above the ground level):

- air temperature (^{o}C),
- number of days with ground-frost (T_{min}<0^{o}C),
- number of hot days (T_{max}≥ 25^{o}C),
- number of very hot days (T_{max}≥ 30^{o}C).

Sunshine duration on the basis of solar radiation (h) was also examined.

The group of pluvial indices includes number of days with rainfall of ≥ 3 mm; ≥ 20 mm which are taken by many researchers as thresholds for estimation of crop growing conditions in Poland. Daily rainfall of 3 mm stops atmospheric drought (water reaches soil through fully developed plant cover) (Koźmiński and Michalska 1999, Meteorological Dictionary 2003), and daily rainfall of ≥ 20 mm has the potential to damage crops and lower the effectiveness of agrotechnical measures. The group of thermal indices includes mean values and the most frequently used characteristics describing conditions of thermal stress.

Data analyzed in this paper were collected in 1966-2005. Data series homogeneity was tested using MASH program (Multiply Analysis of Series for Homogenization) (Szentimrey 1999) within the COST 734 Action (Impacts of Climate Change and Variability on European Agriculture – CLIVAGRI (www.cost.734.eu). Totals and mean values of indices were calculated for vegetation period i.e. April-September (inclusive) for each year in 1966-2005 period. Fluctuations and trend analysis provided information on variability of the studied elements.

Change direction and dynamics of the analyzed climate elements were analyzed with linear regression equation. Statistical significance level was analyzed with Student's t-test at the significance levels of $\alpha = 0.05$ and $\alpha = 0.10$.

Because the presently observed warming manifested at the turn of 1970's and 80's (Kędziora 2008, Kożuchowski and Żmudzka 2001), in order to best capture direction and intensity level of the occurring changes, the research period was divided in two 20-year sub-periods: 1966-85 ("colder" period) and 1986-2005 ("warmer" period – i.e. currently observed warming process).

Fluctuations are defined here as changes that are characterized by successive tendencies of relatively low and then relatively high values of time series (Kożuchowski 1985). They were analyzed with the moving average method. The method of moving average used to characterize multi-year changes of climatic elements is superior to the often used method of regression analysis in that it does not impose a presumed function that approximates the actual run of the examined variable. The result of moving average method is only "smoothing" of fluctuations in period shorter than the corresponding set of the averaged data. The sets of agrometeorological indices were analyzed with 15-year moving averages, which were then examined for deviations from the 40-year average. These deviations provide principal characteristics of variability of the examined variables for the period of 1966-2005. The decision to work on 15-year averages was dictated by the fact, that the analyzed data series were to indicate, if any, main direction of the occurring changes while ignoring short term variability, and also because the series themselves were relatively short – only 40-year period. Based on the test results, 15-year period of averaging was adopted.

In case of each analyzed element, attention was paid to the period of occurrence of the highest and the lowest value of the 15-year average at each station. Statistical significance of differences between the extreme average values was analyzed using Student's t-test at the significance level of $\alpha = 0.10$ and 0.05.

RESULTS

Pluvial indices

The analysis of rainfall totals in Wielkopolska revealed statistically insignificant trend, both rising and declining, during the whole 1966-2005 period and in both subperiods. In 20-year periods, changes did not exceed 5 mm/10 yrs. In both, the 40-year period and two subperiods, majority of the analyzed area was characterized by a negative trend of rainfall totals. In case of the 40-year period, this tendency was most evident with the value of over 10 mm/10 yrs (Fig. 1). Only in the south to Wielkopolska, at Wroclaw station, did the statistically significant at the level of $\alpha = 0.05$ downward trend of rainfall occur.

Change trends of the number of days with rainfall of ≥ 3 mm and ≥ 20 mm were also statistically insignificant (Tab. 1). In case of the number of days with rainfall of ≥ 3 mm, the trend changed direction in the 20-year periods, while the trend of the number of days with ≥ 20 mm rainfall was negative and persisted in the east and south parts of the region.

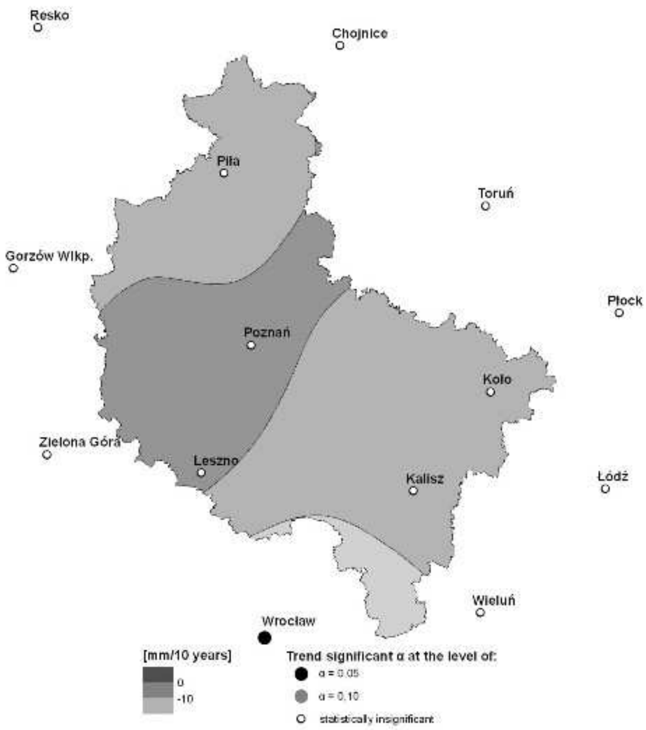

Fig. 1. Spatial distribution of linear trend coefficient of rainfall totals in 1966-2005 (mm/10years)

Fluctuation analysis with moving averages indicated different run of rainfall totals variability at each station. The majority of stations (2/3 of all stations) registered high precipitation during vegetation period at the beginning of the first 20-year period, either in the middle or by the end of the second 20-year period. In this group of stations, the minimum rainfall values were registered at the turn of the first and the second 20-year period. Regardless of the rainfall tendency during the 40-year period, the majority of stations noted downward trend of rainfall change in the last few years of the second 20-year period.

Table 1. Direction and statistical significance of changes of analyzed climate parameters

Parameter	Coefficient of regression equation								
	1966-2005			1966-1985			1986-2005		
	Trend direction	Level of statistical significance		Trend direction	Level of statistical significance		Trend direction	Level of statistical significance	
		0,10	0,05		0,10	0,05		0,10	0,05
Rainfall total	- (+)			- (+)			- +		
Number of days ≥ 3 mm	- (+)			- (+)			- +		
Number of days ≥ 20 mm	- (+)			- (+)			- +		
Mean air temperature	+			-			+		
Number of days $T_{min} < 0^oC$	-			- +			- +		
Number of days $T_{max} \geq 25^oC$	+			-			+		
Number of days $T_{max} \geq 30^oC$	+			- ((+))			+		
Sunshine duration	+			- +			+		

Statistically insignificant trend.
Trend statistically significant at the level 0.10 in some stations.
Trend statistically significant at the level 0.10.
Trend statistically significant at the level 0.05 in some stations.
Trend statistically significant at the level 0.05.

Trend direction:

"+" or "-" Same trend for the entire analyzed area.

"(+)" Positive trend for less than a half of the analyzed area.

"((-))" Negative trend for small parts of the analyzed area.

The differences between the maximum and minimum values of 15-year average were statistically significant at the $\alpha = 0.05$ level at 1/3 of all stations (south-east part of the analyzed area), and at the $\alpha = 0.10$ level at almost all other stations.

Trend of the number of days with given range of rainfall totals also showed to have different directions throughout the analyzed 40-year period. As in the case of rainfall total variability, the high numbers of days with rainfall occurred at the beginning or in the middle of the first 20-year period, and the minimum numbers of days with precipitation were registered usually at the beginning of the second 20-year period (Fig. 2).

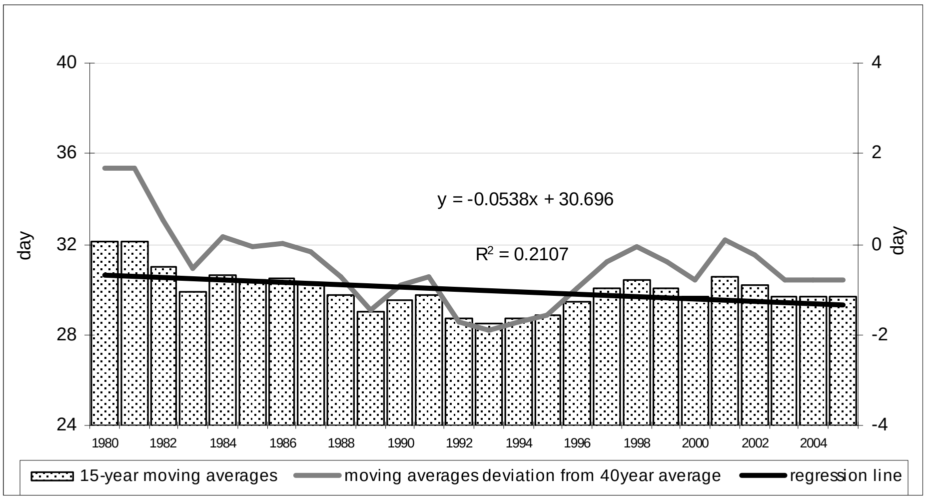

Fig. 2. Number of days with precipitation ≥ 3 mm fluctuations in 1966-2005. Kalisz synoptic station. 15-year average 1966-1980 is shown by 1980 bar and etc.

In the last few years of 1986-2005 period, decrease of the number of days with rainfall of ≥ 3 mm became visible, and the increase of days with precipitation of ≥ 20 mm that characterized this period, ceased entirely. The differences between the maximum and the minimum 15-year averages of the number of days with precipitation showed lower statistical significance than differences of rainfall totals (they were statistically significant at the level of $\alpha = 0.05$ in case of only one station, while the number of stations with the statistically significant differences at the level of $\alpha = 0.10$ oscillated between 2/3 (the number of days with ≥ 3 mm rainfall) and 1/3 (the number of days with ≥ 20 mm rainfall).

Thermal indices

Trend analysis of mean air temperature allowed to determine change trend of this climate indicator. The first 20-year period was characterized by a statistically insignificant drop of air temperature in the entire analyzed area. It was most visible in the north-west part of the region (over 0.5°C/10 yrs). The second 20-year period was characterized by a statistically significant ($\alpha = 0.05$) temperature increase which was the strongest in the west, central and east parts of the region (over 0.7°C/10 yrs) (Fig. 3). The upward trend in the second subperiod was so strong that the trend of mean air temperature for the entire 1966-2005 period was also increasing and statistically significant at the level of α = 0.05 at almost all stations. The highest increase was registered in the south of the region (with the increase of over 0.35°C/10 y).

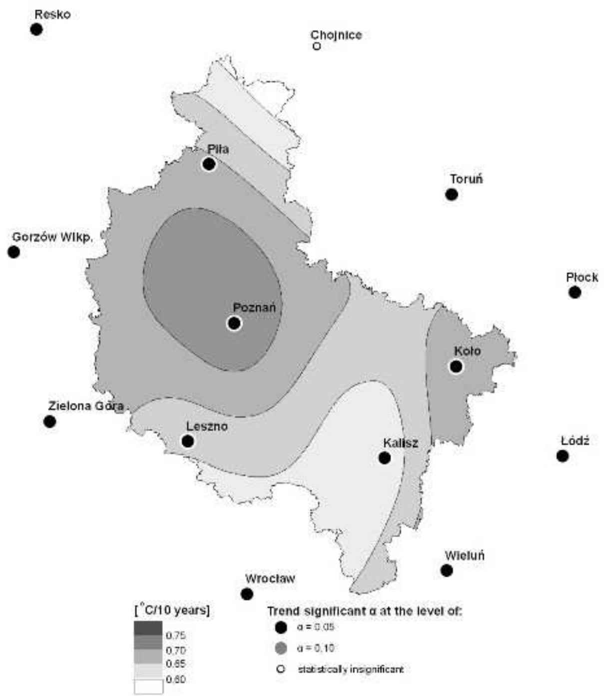

Fig. 3. Spatial distribution of linear trend coefficient of mean air temperature in 1966-2005 (°C/10years)

Direction of changes of mean air temperature is reflected in changes of the number of days with specific thermal characteristic. During the first subperiod (1966-85), there was statistically insignificant drop of the number of hot and very hot days, while in the second subperiod (1986-2005) there was an increase of the number of such days including statistically significant increase of the number of hot days at some stations. As in the case of mean air temperature, the increase of the number of hot and very hot days in the second subperiod was so evident that the 40-year trend was also increasing. In the south part of the region and its surrounding area, the statistical significance of this trend was α = 0.10 (in case of the number of hot days it was α = 0.05 at some stations).

The trend of the number of days with ground-frost was less dynamic. It was statistically insignificant in both 20-year periods with both increasing and decreasing changes in the entire analyzed area. The 1966-2005 period was characterized by a drop of the number of days with ground-frost, statistically significant at four stations located in the south-west part of the analyzed area (Tab. 1).

The run of mean air temperature trend and the number of hot and very hot days analyzed with moving averages indicated very similar variability of these indices in 1966-2005. The first 20-year period was characterized by a very small decrease (in case of the number of days with specific thermal characteristic – at first increase and then drop – Fig. 4), and significant increase of moving averages of these indices in the second 20-year period. The maximum values of moving averages occurred in the last three years of the second subperiod, while in the case of mean air temperatures, 15-year averages of the highest values ended in 2005 at all stations.

Differences between the maximum and minimum values of 15-year averages were statistically significant at the level of α = 0.05 at all stations (with one exception in case of hot days and three exceptions in case of very hot days).

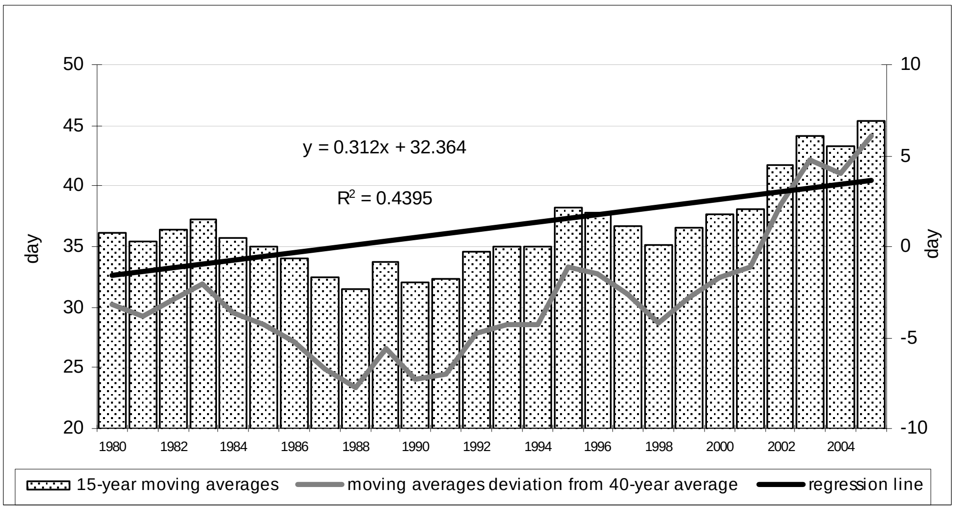

Fig. 4. Number of days with $T_{max} \geq 25^{o}C$ fluctuations in 1966-2005. Poznań synoptic station. 15-year average 1966-1980 is shown by 1980 bar and etc.

The run of the number of days with ground-frost was also similar at all stations. The maxima of moving averages occurred at the beginning of the first 20-

year period, and the minima at different times in the second 20-year period. Differences between the maximum and minimum values of 15-year moving averages were statistically significant at the level of α = 0.05 at half of the stations, and at the level of α = 0.10 at almost all other stations.

Sunshine duration

During the first 20-year period, change trends of sunshine duration were characterized by different directions and were statistically insignificant. Decreasing tendency was most visible in the south-west and north fringes of the analyzed area (over 50 h/10 years) while small increases of total sunshine duration were noted in the center and south west. In the second 20-year period the increase of sunshine hours was registered in the entire region and was statistically significant at the level of α = 0.05 in north-west and north. Analysis of data for the entire 40-year period also indicated statistically significant (α = 0.05) upward trend of sunshine duration totals in the belt spreading from south-east to north-west of the region (Fig. 5).

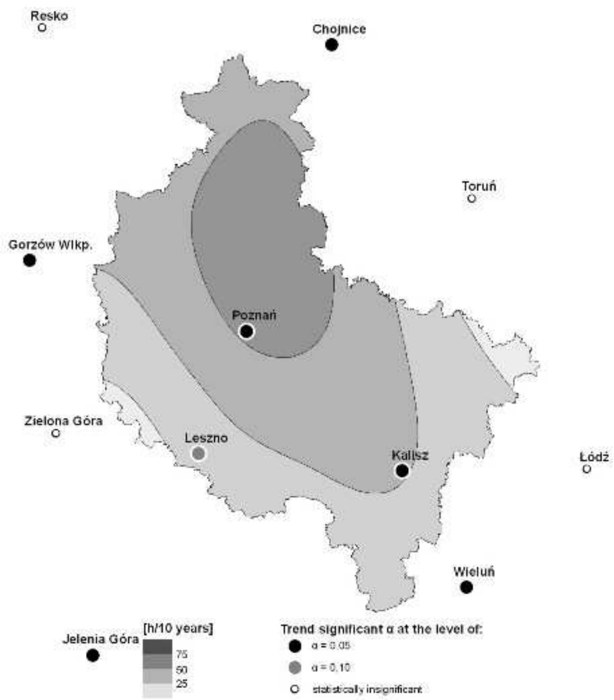

Fig. 5. Spatial distribution of linear trend coefficient of sunshine duration in 1966-2005 (h/10years)

Fluctuation analysis confirmed a decrease of sunshine duration totals in the first 20-year period at almost ½ of the stations and considerable increase in the second subperiod. All other stations registered increase of sunshine duration totals already during the first subperiod (see the example below: Fig. 6).

Minimum values in the run of moving averages occurred in the first subperiod, and maxima by the end of the second 20-year period. It may be emphasized that the differences between the maximum and the minimum values of the 15-year averages were statistically significant at the level of α = 0.05 at 90% of all stations, and at the level of α = 0.10 at the remaining stations.

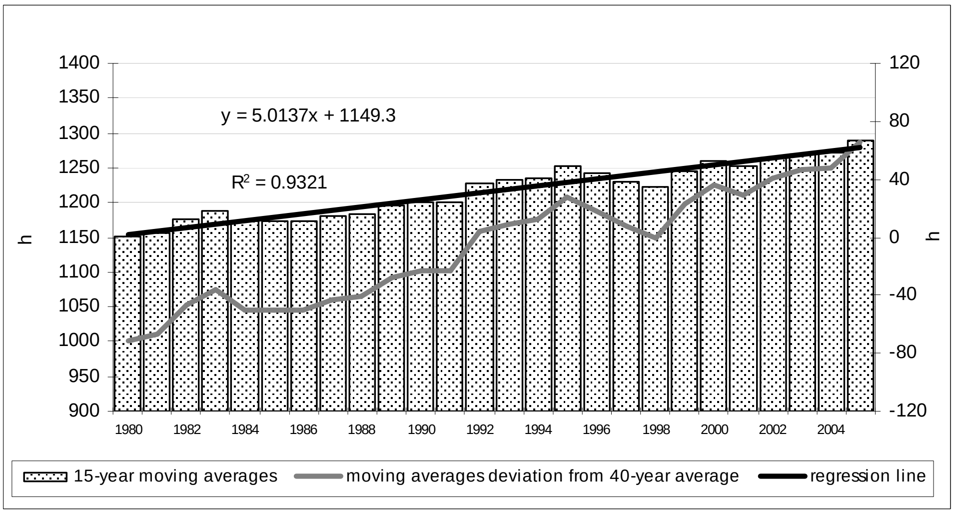

Fig. 6. Sunshine duration fluctuations in 1966-2005. Kalisz synoptic station. 15-year average 1966-1980 is shown by 1980 bar and etc.

DISCUSSION

The analysis of results presented here confirmed the results of other researchers regarding directions of changes of principal parameters of present climate in Poland. All emphasize the existence of a trend of progressing warming (Kożuchowski and Żmudzka 2001, Mager *et al*. 2009a), increasing sunshine duration (Kępińska-Kasprzak *et al*. 2008), and statistically insignificant changes of precipitation totals in different directions (Mager *et al*. 2009a, Żmudzka 2002). Directions of changes discussed by Kożuchowski and Żmudzka (2001) and

Żmudzka (2002) in the period ending in 2000 continue to develop also in the following years (see the results presented in this paper).

The impact on agriculture in the Wielkopolska region, and the entire area of Poland, of the changes of climate elements observed in the last years is difficult to be determined as either definitely positive or negative. The impact varies depending on the analyzed climate element, its change direction and the area where the change is observed. In case of Wielkopolska, a very important factor is the amount of water that is accessible to plants during vegetation period and, therefore, determines total yield. The amount of available water is, simplifying, the difference between precipitation total and runoff combined with evaporation total.

Change trends of all pluvial indices, i.e. water input, were not homogenous in the entire area – in case of each parameter in this group, there were upward and downward trends in different parts of the area with the dominance of a decreasing tendency. The observed changes were not statistically significant. Because of the general characteristic of precipitation variability, it may be assumed that the observed changes of pluvial indices reflect the situation within a given period rather than represent permanent changes.

Air temperature and sunshine duration increases observed for over two decades influenced the evaporation rates. This parameter also showed growing tendency of statistical significance throughout the entire Wielkopolska region after 1985 (Kępińska-Kasprzak and Mager 2010, Mager *et al*. 2009a, Mager *et al*. 2009b). The observed increase of evaporation coupled with insignificant changes of precipitation totals led to a decrease of available water during vegetation period.

The increase of air temperature and smaller numbers of days with ground-frost during vegetation period combined with longer sunshine durations is a positive factor that opens possibility to introduce plants that require higher temperatures. Example of this is the increase of area of maize for seed production.

Changes of the indices that determine conditions for crops may have multidirectional character. According to IPCC projections (Climate Change... 2007), the probability that the presently observed climatic trends will continue for a long time is very high. Since these changes concern also all the elements that influence the amount of water available to plants, it is necessary to stress and to raise awareness of decision makers and all stakeholders to the possibility of deterioration of water conditions for agricultural needs in large areas of central Poland.

CONCLUSION

Wielkopolska is one of the regions that are most vulnerable to water shortages in Poland. The presently observed warming, due to –among others- statistically significant air temperature and sunshine duration increase with small changes in rainfall totals, causes higher water evaporation. This process results in declining water resources available to plants during vegetation season. Water shortage occurrences can significantly contribute to crop failure. Long term climate forecasts indicate that the adverse change trends of some climate elements may continue for a longer period of time. This fact should force administration and government into taking adaptive actions that would secure further development of agriculture in these changing conditions.

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2. THE IMPORTANCE OF RESEARCH INTO FORM AND TYPE OF PRECIPITATION IN THE STUDY OF CLIMATE CHANGE*

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INTRODUCTION

Contemporary trends of air temperature and precipitation

According to a recent report published by the Intergovernmental Panel on Climate Change (IPCC, 2007) the average global air temperature grew by approximately 0.7°C during 1906-2005. The highest rates of increase were observed from 1976 onwards. In the light of research into the contemporary climate change in Europe (Klein Tank *et al*. 2002) the recent air temperature growth is indeed unprecedented. It is particularly strong in Western and Central Europe, where the growth rate was found to exceed 0.5°C per 10 years (Fig. 1). During the last ten years, the warming trend involved average annual temperatures higher than the long-term averages, e.g. in Krakow where the average temperatures of 11.0°C recorded in 2000 and 2008 were higher than in any other year since 1792 when instrument-based measurements began in the city (Fig. 2). Also notable were extremely high air temperatures recorded in the summer seasons of 2003, 2006 and 2007 (Twardosz 2009a).

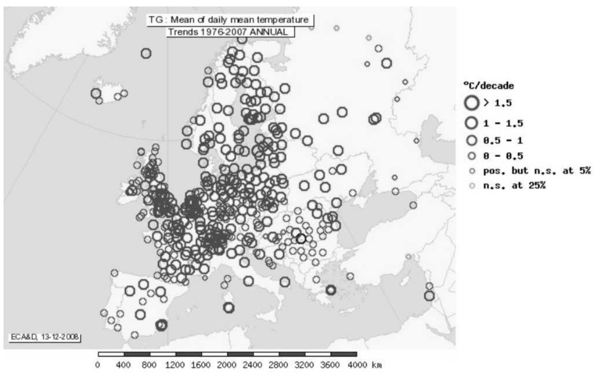

Fig. 1. Trends of mean annual temperature, 1976-2007 (source: www.knmi.nl)

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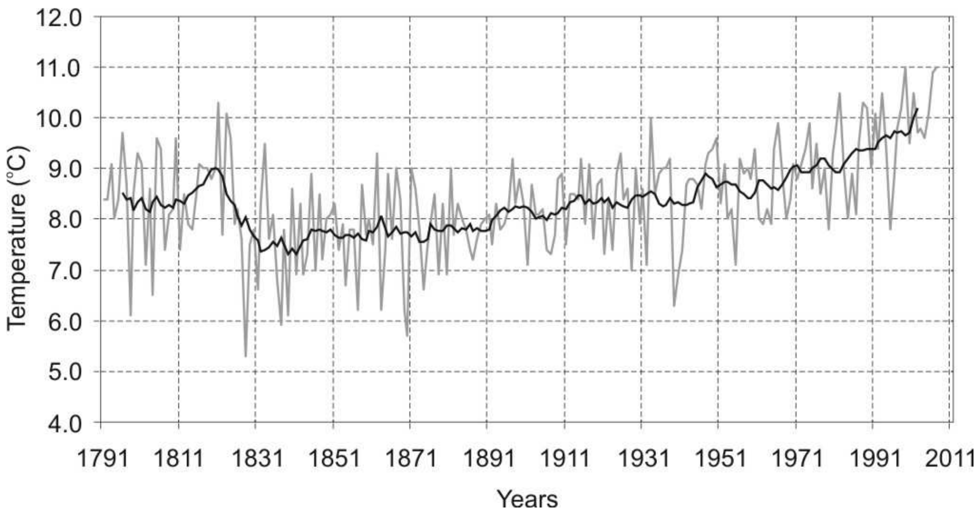

Fig. 2. Long-term variation in mean annual temperature in Krakow, 1792-2008 (source: Own study based on data from the Climatology Department of the Jagiellonian University)

Temperature is what comes to mind first when talking about contemporary climate change. Indeed, air temperature has followed a statistically significant trend to increase, both on the annual and seasonal scale. However, no such clear-cut trends are found in precipitation. Kożuchowski (2004) aptly noted that there were irregular fluctuations and shifting zones of precipitation surplus and deficit, rather than consistent trends, as illustrated by the secular variation in annual precipitation in Krakow (Fig. 3). This particular and very important climate component is characterised by a great deal of temporal and spatial variability that complicates any attempts at linking it with global warming. Bradley *et al.* (1987) notes that scenarios derived from global circulation models (GCM) carry a much greater error margin in the case of precipitation change than in the case of air temperature change.

The IPCC Report (IPCC 2007) finds that the global trend in mean annual land precipitation over the period from 1900 to 2005 is statistically insignificant. Also the latest research into long-term precipitation change at the annual and monthly scales in Central Europe revealed no statistically significant change (Niedźwiedź *et al*. 2009). All weather stations of the region have recently recorded a clear-cut reduction in precipitation after a series of very wet years in the 1990s and at the beginning of the 20th century.

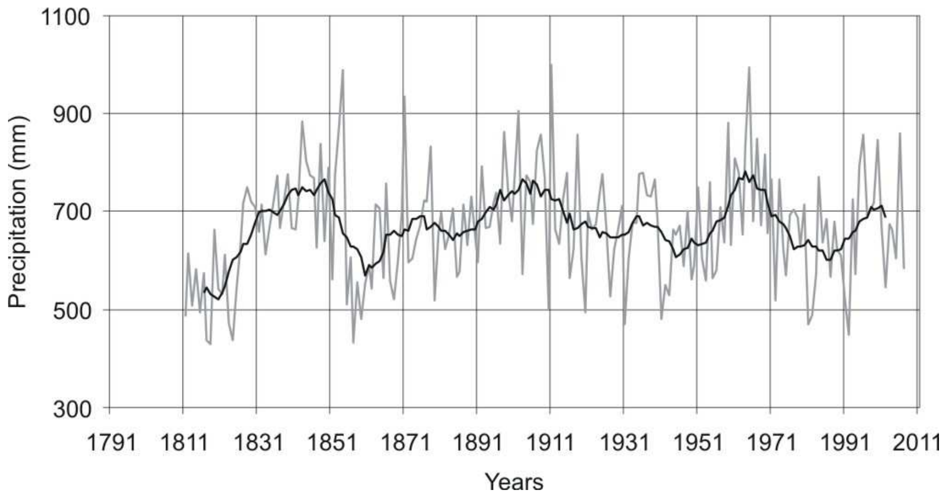

Fig. 3. Long-term variation in mean annual precipitation totals in Krakow, 1812-2008 (source: Own study based on data from the Climatology Department of the Jagiellonian University)

In Poland an observed increase in air temperature was accompanied by a trend to a decline in annual precipitation in lowland areas (Kożuchowski and Żmudzka 2003) and mild winter seasons (Kossowska-Cezak 2009). A forecast based on the HadCM2 GS model envisages an approximately 30% increase in precipitation as the region's temperatures increase (Kożuchowski 2004). However, Kożuchowski (2004) himself warned that the precipitation scenario carries a great deal of uncertainty.

Precipitation characteristics and measures for their determination

Most research into the change and variability of precipitation solely considers precipitation totals. Indeed, this is the parameter that is most readily available, because most measurements, such as those with pluviometers, only record the amount of water collected over a certain period of time (normally a 24 hour period). Other precipitation characteristics, such as the frequency of occurrence, intensity and type of precipitation, rarely find their way into published research.

Climatological research seldom involves the breakdown of precipitation by form (i.e. solid or liquid) and type (front-related, convective and other types). Indeed, coding of precipitation type requires precise information about meteorological phenomena, which is only available from round-the-clock visual observation, which is rare. For this reason few weather stations offer long-term series of data on the form and type of precipitation. Snowfall and thunderstorm precipitation is particularly significant in the study of nature, climate and hydrology. Given the right conditions, solid-state precipitation results in snow cover, which acts as a form of natural water retention (Kossowska-Cezak and Bajkiewicz-Grabowska 2008). Snow cover is linked to various dangerous phenomena, including rapid thawing that can cause violent floods affecting large areas. There is also a practical benefit associated with investigations into the changing form of precipitation in winter, because it has important consequences for human activities. Snowfall, especially when intense, can cause a wide range of inconveniences that are particularly troublesome in urban areas and for road and air traffic. Thick snow cover is known to cause building failure. In a recent Polish example the roof of the Katowice International Fair building caved in on 28 January 2006 due to a thick cover of snow and ice, killing 65 and injuring 170, including 13 foreigners. Thunderstorm precipitation tends to cause disruptions in communications and transport in summer. Overfilled storm drainage systems cannot cope with the runoff water and low-lying areas become flooded. The financial cost of such events tends to be high and there are cases in the USA reported at billions of dollars (Changnon 2001).

Thunderstorm precipitation is a result of strong convection in the lower parts of the atmosphere and tends to be short, but can be very intensive. The Polish climate is prone to the occurrence of isolated precipitation events of extremely high intensity, such as in Krakow on 9 September 1963 when nearly 100 mm of water fell within one hour, an equivalent of 14% of the mean long-term precipitation recorded in that city (Twardosz 2009b). This type of short and intense downpour falling on a limited area can cause a rapid increase in runoff and serious flood hazards. Both snowfall and thunderstorm precipitation have a great impact on the amount and timing of runoff, as well as evaporation, and therefore constitute important factors in the development of floods and the proportions of the components in the water balance.

The form of precipitation can also be classified in an indirect way. It involves an empirical relationship between the form of precipitation observed and the temperature in the lower atmosphere (Kupczyk 1997). This automatic classification is, however, marred by several deficiencies. Indeed, Kupczyk (1997) herself stated that during intermediate seasons, especially in moderate latitudes, there is much fluctuation in air temperature and in the conditions required for the formation of precipitation over the 24 hour period. Figure 4 provides a good illustration of this pattern with snowfall being dependent on the average daily temperature in

Krakow. Snowfall is recorded across a broad range of average daily air temperatures, i.e. from –19.2ºC to +12.1ºC. Only isolated snowfall events were associated with either temperature limit, but nearly 47% of the total snow fell within a narrow band of –5 to 0ºC, and 76% fell at temperatures from -5 to +5ºC. This also means that a considerable proportion, 24%, of the snow fell outside of that latter range. Intensive snowfall, defined as events with a minimum of 10 mm of snow, occurred in a narrower range of average daily temperatures of –9.7 to 3.1ºC. Nearly 74% of days with events of this intensity recorded average temperatures ranging from –2 to 2ºC.

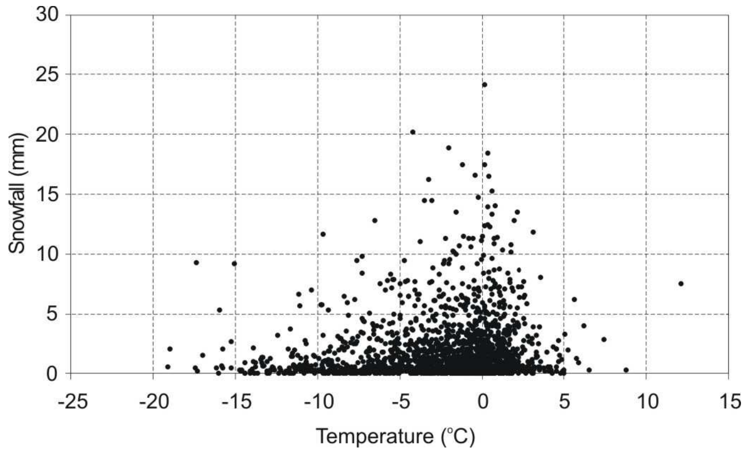

Fig. 4. Snowfall (mm) dependence on daily mean air temperature (ºC), Krakow, 1958-2008

Research hypotheses

The current period of global warming, which has been observed for more than the last ten years, provokes a question as to whether the increasing air temperature has not been accompanied by any changes in the frequency and volume of snowfall and of thunderstorm rainfall. In answering this question the author focuses mainly on finding dependencies between various characteristics of precipitation and air temperature. These relationships are not yet well understood or documented, asserts Davis *et al*. (1999).

It seems correct to assume that as air temperature increases significantly, the frequency of snowfall should diminish to be replaced by liquid precipitation in wintertime and that convective precipitation should become more frequent in summer. This might sound like very straightforward research hypotheses to make, but precipitation, regardless of its form or type, depends on more than just air temperature. Among the numerous factors shaping precipitation the most important are water vapour concentration and local considerations, such as various ground effects. Additionally, as was aptly pointed out by Ye (2008), the complexity of the climate system and feedbacks that link its various components make understanding the relationship between precipitation type and frequency difficult, especially at high latitudes where the air temperature of the cool seasons often drops below freezing point.

Many climatologists have highlighted the point that snowfall and snow cover constitute important components of the climatic system that are sensitive to change (Hantel *et al.* 2000). Hantel *et al.* (2000) found that a 1°C increase of temperature can reduce the snow cover duration in the Austrian Alps by approximately four weeks. This makes snow a good indicator for the monitoring of global changes (Namias 1985, Dobrowolny 1993, Jaagus 1997, Huntington Hodgkins 2004). Research into the change of thunderstorm precipitation offers a way to verify a hypothesis often quoted in climatological literature predicting a growing frequency of intensive precipitation and an increase in precipitation in the temperate zone. Also both snowfall and thunderstorm precipitation are important components in the water circulation cycle (Changnon 2001). Thus the benefits of research into the form and type of precipitation also include their impact on the hydrological cycle, as well as involvement in climate change. Potential advantages or disadvantages of the forecast growth of precipitation in Poland under the influence of climatic warming will depend on their form and intensity. A similar question was posed by Ye (2008) in relation to the forecast precipitation changes in the high latitudes.

RELATIONSHIPS BETWEEN PRECIPITATION AND AIR TEMPERATURE IN KRAKOW

Source material and methodology

The objective of the study was pursued and verification of the hypothesis was approached using the long-term records of daily precipitation in Krakow. The main series covered 51 years of observations taken between 1958 and 2008 at the weather station situated at 206 m a.s.l. in the botanic gardens in the city centre. The complete record of the weather station is much greater, but although these include visual observations of atmospheric phenomena, such as the form and type of precipitation, which started in 1792, and instrument-based measurements, available since 1849, this data still requires verification. Records of the form and type of precipitation at weather stations in moderate latitudes are difficult to obtain. The largest body of research into the change in frequency of snowfall over time is available for mountains, i.e. where most of the precipitation is solid-state (e.g. Førland, Hanssen-Bauer 2003, Przybylak 2002).

This study uses daily precipitation totals and details of their form and type derived from information noted about the phenomena which accompanied precipitation. With this input it was possible to determine the number of days with snowfall and the number of days with thunderstorm precipitation, as well as the sum of the total precipitation on these days, broken down by month. Relationships were sought between the monthly precipitation totals and average air temperature using linear regression. With regards to snowfall the analysis was carried out separately for the winter months (December-February) and for the winter season as a whole. The investigation of thunderstorm precipitation was limited to the summer season (June-August) because of their low frequency of occurrence. Other parameters examined in the study included trends of change in air temperature and precipitation characteristics, including their statistical significance, which were studied in relation to the period 1958-2008.

Snowfall

In Krakow snowfall accounts for only 19% of the total days with precipitation per year and about 50% in winter. The snowiest months are January and February, when more than half of all precipitation days were associated with snowfall, and December lagged somewhat behind (Twardosz 2007). During the remaining months of the cool half of the year, there were decidedly fewer days with snowfall and they varied wildly from year to year (Fig. 5). The overall period when there is a potential for snowfall is relatively long and runs from the second decade of October to the second decade of May.

Figure 6 illustrates the relationship between snowfall totals and snowfall days in Krakow and the average air temperature in winter and over the winter months. Both characteristics of solid precipitation are inversely correlated with air temperature, which means that as temperature increases, so snowfall frequency and volume drop. A 1^{o}C increase in the average temperature in winter corresponded to a 3.5 day drop in snowfall days and a 6.1 mm drop in snowfall total. The strongest statistical relationships of snowfall days and totals were obtained for the winter season and the weakest for February. The impact of air temperature on precipitation is much stronger with regard to snowfall frequency than to volume. In winter, air temperature variability explained 54% of the variance of days with solid precipitation, but only 35% of the variance of snowfall totals.

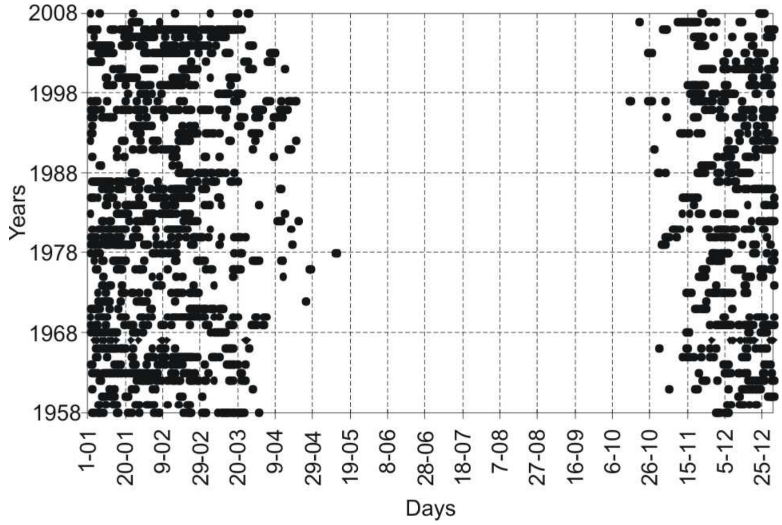

Fig. 5. Occurrence of days with snowfall, Krakow, 1958-2008

Research by other authors, including Davis *et al*. (1999) and Karl *et al.* (1993), shows that the correlation of snowfall totals and air temperature may be either positive or negative. The two studies ascertain that it is the location of the weather station, especially in terms of latitude and altitude, that decides which way the correlation goes and determines the range of average monthly temperatures. At stations where the average monthly air temperature is relatively high the correlation is negative and vice-versa. This pattern stems from the Clausius-Clapeyron formula that defines the conditions of water vapour condensation. Warmer air holds more water vapour and there is more snowfall both in terms of volume, as demonstrated by Ye (2008), and frequency.

Over the period studied of 1978-2008, the average wintertime temperature in Krakow showed a trend towards an increase at a rate of 0.4°C per 10 years (Fig. 7). However the great range of temperature fluctuation means that this trend only has a level of statistical significance of p = 0.10 and does not reach a level of statistical significance of p = 0.05.The lack of a significant trend in temperature corresponds to a lack of trend in precipitation, whether in terms of overall or snowfall totals (Fig. 8), or the number of days with any precipitation and with snowfall (Fig. 9) in wintertime. Solid-state precipitation as a proportion of overall precipitation follows a declining trend in the winter season (Fig. 10), but it is insignificant at 0.05, just like the situation with temperature.

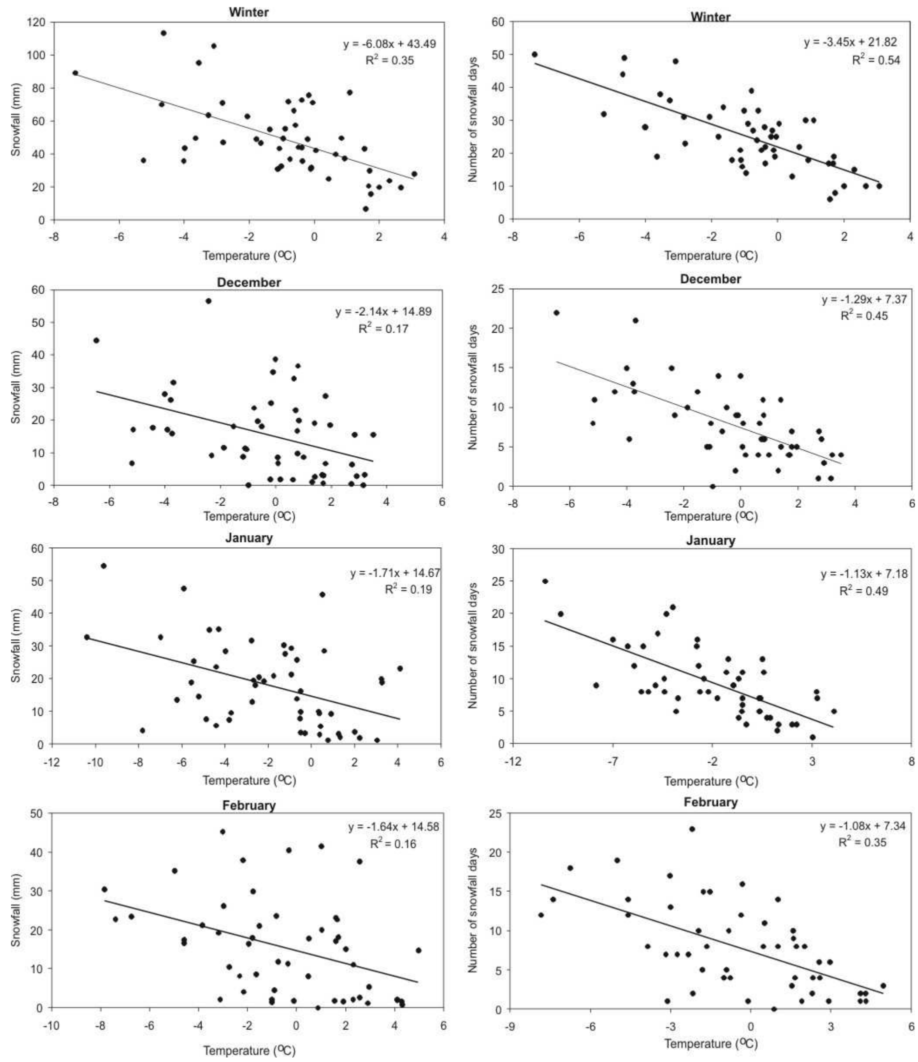

Fig. 6. Dependence of snowfall (mm) and number of days with snowfall on monthly/ seasonal mean air temperature (ºC), Krakow, 1958-2008

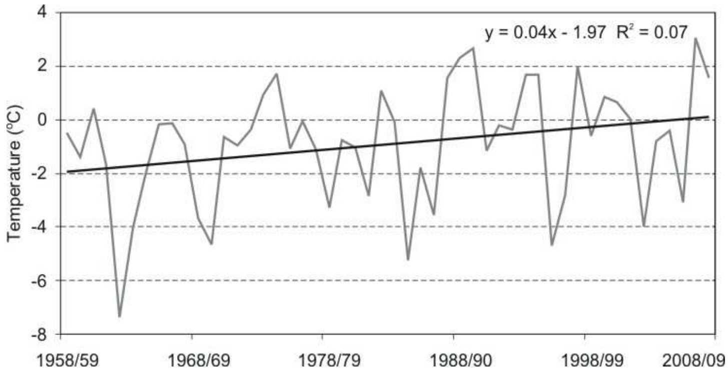

Fig. 7. Variation of mean air temperature in winter, Krakow, 1958-2008

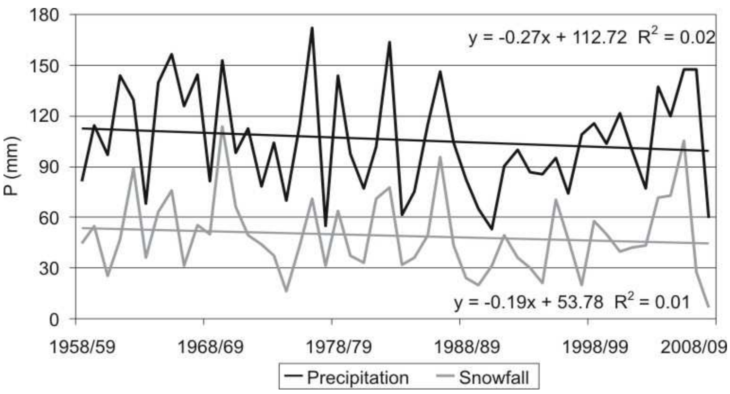

Fig. 8. Variation of precipitation totals and snowfall in winter, Kraków, 1958-2008

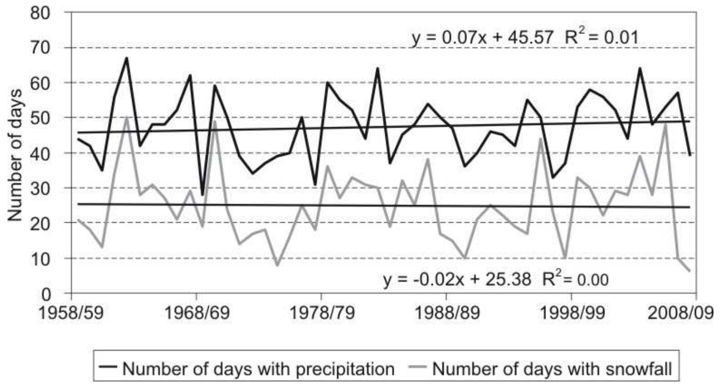

Fig. 9. Variation in the number of days with precipitation and number of days with snowfall in winter, Krakow, 1958-2008

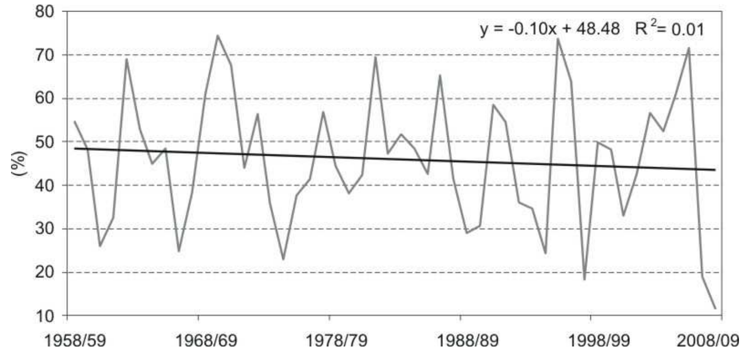

Fig. 10. Variation in the percentage share of snowfall in the winter totals, Krakow, 1958-2008

Other researchers have also found that snowfall trends in different areas varied in strength and vector, and did not always reach statistically significant levels (Ke *et al.* 2009, Laternser and Schneebeli 2003).

Thunderstorm precipitation

In Krakow thunderstorm precipitation accounts for 12.4% of precipitation days and is observed in all seasons with the highest incidence in the warm months (Fig. 11). In July, thunderstorms are recorded on 35% of days with precipitation and thunderstorm-related precipitation accounts for 50% of the month's total precipitation.

In summer, there is a negative, albeit statistically insignificant, correlation between thunderstorm precipitation total and the season's average temperature (Fig. 12) and there is no correlation between temperature and the number of thunderstorm precipitation days. Only 2% of the variance of total thunderstorm precipitation can be explained by variation in summertime temperature, which only attests to the scale of the variability of that precipitation. This shows that contrary to all appearances, there is no simple relationship between thunderstorm precipitation and temperature. While convection increases as temperature grows, this does not translate into a greater frequency and output of thunderstorm precipitation. This may be attributable to the fact that, in the Polish climate, the hottest summers are also the driest ones and dryness obviously does not favour the development of thunderstorms. Although the statistical dependence of precipitation on temperature is insignificant, Figure 12 shows that the highest thunderstorm totals, exceeding 200 mm, are confined to a narrow average temperature band of 17-18°C. As temperature increases beyond that range, and especially above 19°C, the air humidity drops which results in a decline in both the frequency and quantity of thunderstorm precipitation. In addition to temperature, the concentration of water vapour is an important factor in the development of high-intensity thunderstorm precipitation. In Poland the greatest precipitation totals are recorded during frontal thunderstorms (Twardosz 2009b) which occur most frequently in cyclonic circulation arriving from west and north-west. The greatest negative anomalies of air temperature occur in summer associated with these circulation types (Niedźwiedź 1981).

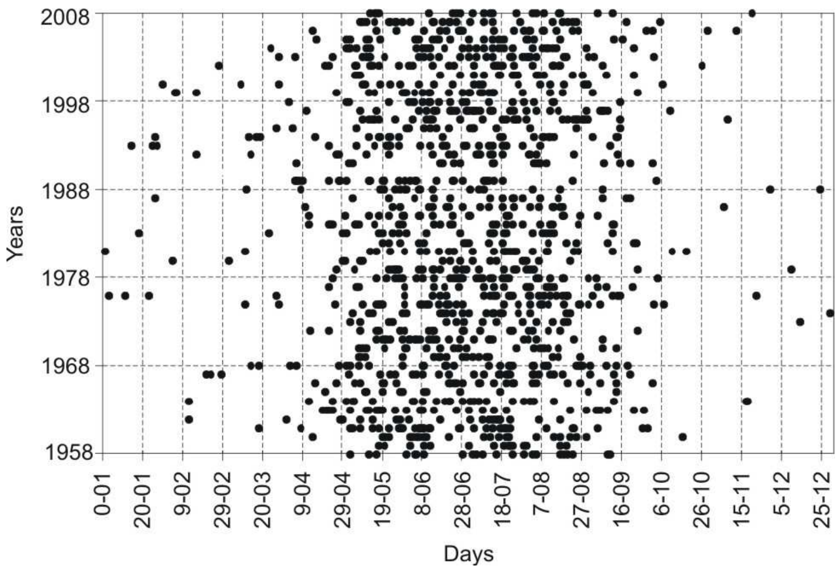

Fig. 11. Occurrence of days with thunderstorm precipitation, Krakow, 1958-2008

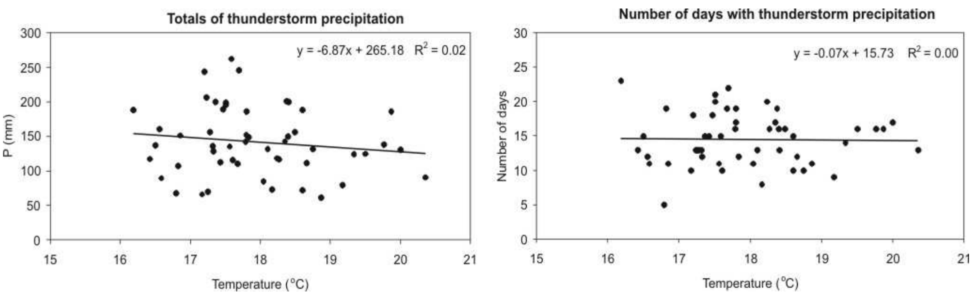

Fig. 12. Dependence of thunderstorm precipitation totals (mm) and number of days with thunderstorm precipitation on summer mean air temperature (ºC), Krakow, 1958-2008

During the study period, the average summer temperature followed a statistically significant trend to an increase at the rate of 0.3°C per 10 years (Fig. 13). This temperature increase was particularly rapid during the last 20 years. It was accompanied by a decline in summertime precipitation at the rate of 17.5 mm per 10 years, which was significant at 0.05 (Fig. 14).

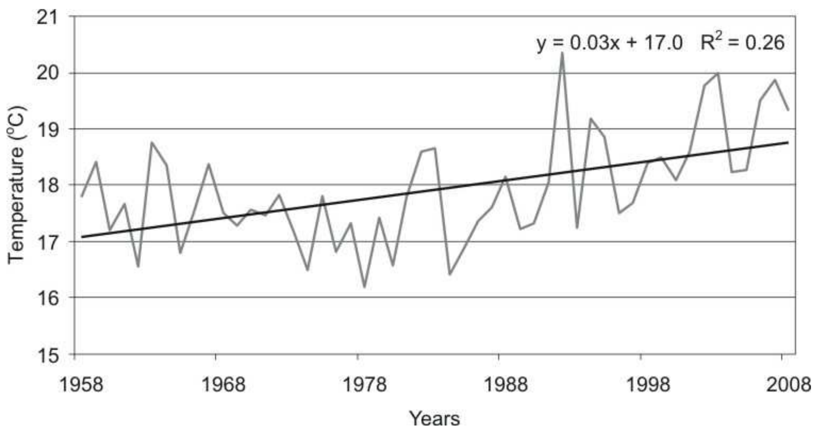

Fig. 13. Variation of mean air temperature in summer, Krakow, 1958-2008

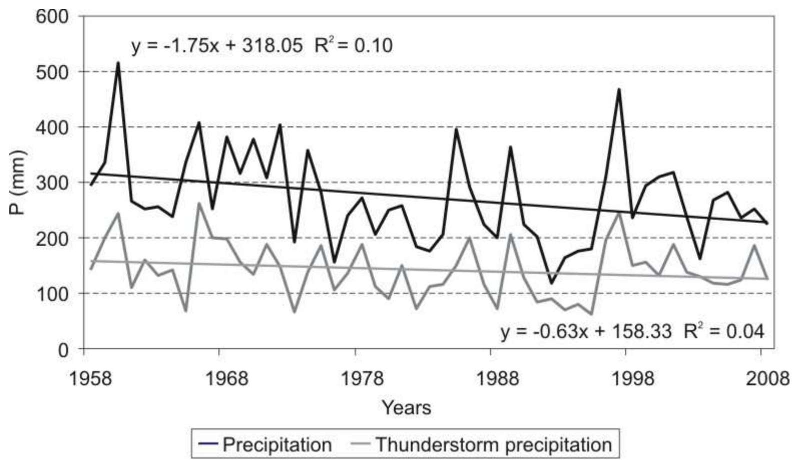

Fig. 14. Variation of precipitation totals and thunderstorm precipitation totals in summer, Krakow, 1958-2008

In his study focusing on the city of Łódź Kożuchowski (2004) found that the precipitation totals in the summer months showed a significant correlation with average air temperature. The correlation was negative, which means that as temperature increased precipitation dropped. This pattern was also found to be true in Krakow. During the study period, thunderstorm precipitation totals were falling at the rate of 6.3 mm per 10 years, but the trend was insignificant, which is also confirmed by other research (Bielec-Bakowska and Lupikasza 2009). Ultimately then one cannot define the influence of air temperature on summer precipitation with a simple linear relationship. In general it is true, however, that with higher temperatures the climate becomes drier and because of this the amount of precipitation declines.

CONCLUSIONS

1. The investigation of the form and type of precipitation has great importance for the study of contemporary climate change, as well as for many practical reasons.
2. This paper has shown that air temperature has an influence on the frequency and quantity of precipitation and that the influence is stronger in winter than in summer. The study has also demonstrated that the influence is a complex one and that it also depends on other factors, especially on the concentration of water vapour in the air.
3. There is a generally negative correlation between snowfall and thunderstorm precipitation and air temperature. The statistically strongest relationships between precipitation and temperature were found in the frequency of precipitation in winter. They were weaker, albeit still significant, in the case of precipitation totals. Although statistical relationships between thunderstorm precipitation and temperature are clearly weaker, a considerable drop in thunderstorm precipitation totals and frequency is visible as temperature increases, which is attributable to decreasing humidity.
4. If the upward trend of air temperature continues, it should be expected that southern Poland will receive less snowfall and as a result less water. There will probably also be less thunderstorm precipitation, while an intensification of convection may lead to the development of other dangerous phenomena, such as tornados, which have already been observed.
5. Further study of long-term precipitation trends is necessary concerning the relationship of form and type to thermal change using longer time-series, which should at least show secular variation. Research on the change and variability of

precipitation should also include consideration of atmospheric circulation, as this not only determines the thermal conditions, but also the humidity of air masses.

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3. SEASONAL VARIABILITY OF PRECIPITATION ON THE POLISH BALTIC SEA COAST

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INTRODUCTION

Precipitation is one of the most important parameters of the hydrological cycle. It is also very variable. These two characteristics generate high scientific interest. Meteorologists, climatologists and agriculturalists endeavor to study characteristics of precipitation.

The main aim of this study is the analysis of annual variability of precipitation totals and rainfall frequencies expressed as the number of days with precipitation on weather stations representative for the Polish Baltic Sea coast. Characteristic features of the coastal zone that differ it from the Polish interior were considered in particular in this study. Also the influence of variable circulation conditions on precipitation was considered. Both temporal and spatial differentiation of rainfall hindered the search for patterns in the variability courses.

Seasonal variability of precipitation on the Polish Baltic Sea coast have been addressed by several studies. Ones of the first ones were papers characterised precipitation patterns in particular seaside provinces, or one-time Szczecin Province (Koźmiński *at al.* 1977), Koszalin (Koźmiński *at al.* 1982), Słupsk (Koźmiński *at al.* 1986) and Gdańsk one (Koźmiński *at al*. 1986). The next important study focused on yearly variability of precipitation in the north-western part of Poland was the Ewert's paper (1984). A yearly course of precipitation totals in the west part of the coastal zone were analyzed by Fortuniak (1996) and Koźmiński, Michalska and Czarnecka (2007). In turn variability of pluvial conditions in the area of on the west seaside were described by Miętus and Filipiak (2002) and Malinowska and Filipiak (2002). The yearly courses of precipitation's frequency and totals with special regard to precipitation ≥ 10mm were analyzed by Kirschenstein, too (2004). The important paper was also the study of Miętus with contributors (2005). That work presents the analysis of pluvial conditions in the Polish coast and the neighboring area based on the data collected 30 stations. Next studies pertain the subject discussed in the present paper were studies of Świątek (2003, 2009) and Girjatowicz (2008, 2009).

MATERIAL AND METHODS

Mean daily precipitation totals from the period 1954-2003 were studied. Data from the following weather stations were utilized: Szczecin, Świnoujście, Koszalin, Łeba, Hel, Gdynia and Elbląg. Data were retrieved from published and unpublished materials of the Institute of Meteorology and Water Management. Homogeneity of data series was determined by means of SNHT (Standard Normal Homogenity Test) (Alexanderson 1986). Monthly precipitation totals for Gdynia and Łeba were found to be inhomogeneous, therefore proper corrections were calculated.

Monthly values of precipitation totals and numbers of days with precipitation were presented in a relative form – as percentages in corresponding annual values. The reason for such form of data presentation was the uneven number of days in particular months. Annual courses of daily precipitation totals were smoothed by moving averages. To obtain this, irregularity index was calculated as the average deviation of monthly precipitation totals. Pluvial continentality can also be described by semi-period of precipitation. Precipitation semi-period is a ratio of the number of days from the 1st of April to the day on which the current precipitation total equals half of the annual precipitation total, to the half of number of days in a year treated as 365,25 days. A value of precipitation semi-period equal to 100% indicates about the same precipitation totals in both warm and cold half-years. The lower the value of the index the greater is the prevalence of precipitation during the warm half-year, and consequently the higher the degree of continentality.

Due to the high dependence of precipitation on atmospheric circulation, its variability was related to the intensity of zonal air-mass advection through determination of annual courses of precipitation sum variability in the so-called circulation epochs.

RESULTS

The highest precipitation sums at the studied stations occur in July (except for Świnoujście and Łeba). It is typical for Poland (Mrugała 2001), but over the Baltic Sea maximum precipitation is observed in August (Meier 2006). In Świnoujście precipitation is the highest in June (Fig. 1a), and in Łeba it is the highest in September (Fig. 1d). Precipitation totals are the lowest in February (Szczecin, Świnoujście, Koszalin and Gdynia) or in March (Łeba, Hel and Elbląg). Low precipitation sums in March are caused by the maximal number of unfrontal days in Poland in this month (Parczewski 1965), in connection to the equal temperatures of moderate latitudes of the Atlantic Ocean and Europe. The absence of thermal differences blocks the western air masses advection.

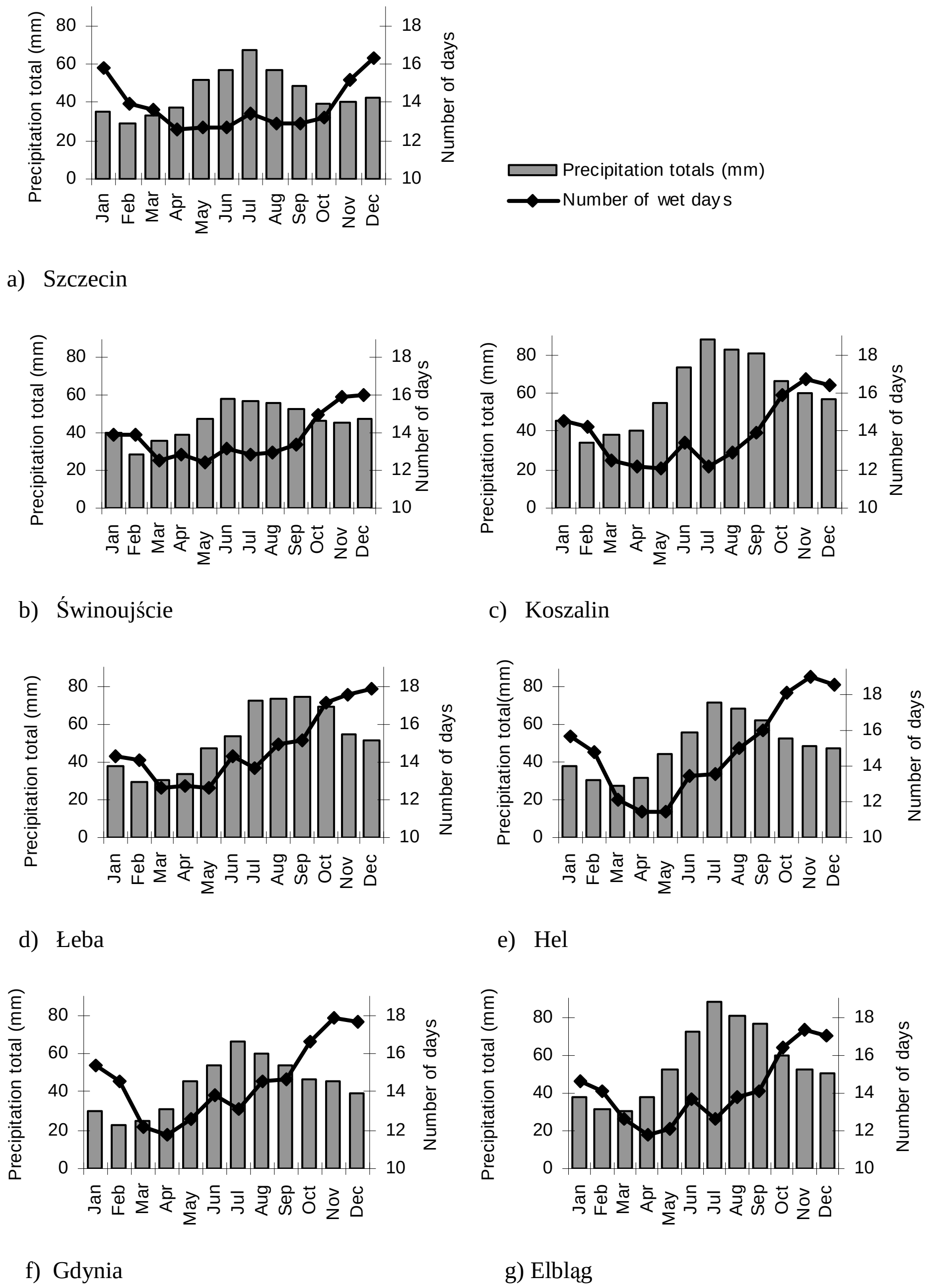

Fig. 1. Mean monthly precipitation totals and mean monthly number of days with precipitation on particular stations

The highest number of days with precipitation occurs usually in November (in Szczecin in December). This is associated with stronger cyclonic activity over the Baltic Sea in this period (Miętus 1998). Besides, November and December are the months of the annual maximum activity of the Icelandic Low. The fewest wet days occur in May, which is caused by very low pressure gradient over Europe (Schönwiese and Rapp 1997), high pressure in the Southern Baltic region and the minimum of the annual course of the Icelandic Low activity. The lowest number of days with precipitation in Szczecin occurs in May and in August. In Świnoujście, Koszalin, Hel and Elbląg it's observed in May. In Łeba the number of wet days is the lowest in May and March, and in Gdynia in March and April. At studied stations (except for Szczecin and Elbląg situated further from the sea than other stations) precipitation totals in November are higher than 1/12 of the average annual total, which is characteristic for pluvial conditions of the seacoast, but not typical for the climate of Poland (Mrugała 2001) and most of Europe, with the exception of the Mediterranean and Atlantic stations (Karagiannidis *et al.* 2008, Groisman *et al.* 1999).

Annual courses of mean monthly precipitation totals and the frequency of days with precipitation demonstrate clearly asynchronous variability (especially in Szczecin). It is caused by longer time of precipitation continuance in the cold part of a year than in the warm one (Madany 1973, Twardosz 2005), combined with many lows and frontal activities. In summer, rainfall is heavier because of convection processes, but its duration, frequency and consequently number of wet days are lesser.

Summer precipitation totals equal as much as 150% of the winter values in Świnoujście, and up to 200% in Gdynia and Elbląg (Fig. 2). The dominance of summer precipitation over winter one increases toward the east. The reason for this is the increase of distance from the Atlantic Ocean and consequently the increase in the degree of pluvial continentality. The obvious prevalence of autumn precipitation over the spring values (rainfall in an autumn is heavier and much more frequent than in a spring) is caused by the influence of the Baltic Sea on pluvial conditions at the studied stations (Figs. 1. and 3.). Higher temperature of the sea and air masses over it than over the land favors the appearance of rainfall, especially in the narrow belt of the coast (Ewert 1984).

Differences between precipitation totals in a summer and a winter (Fig. 2.) show the most distinct seasonal variability of rainfall in Elbląg. The greatest prevalence of autumn precipitation totals over spring values is observed in Łeba. It is associated with the more continental character of pluvial conditions in Elbląg and more oceanic features of precipitation in Łeba, in comparison to other stations located in the coastal zone. Other indices were also used to obtain a more comprehensive understanding of the pluvial continentality. One of them is the irregularity index presented in Figure 4.

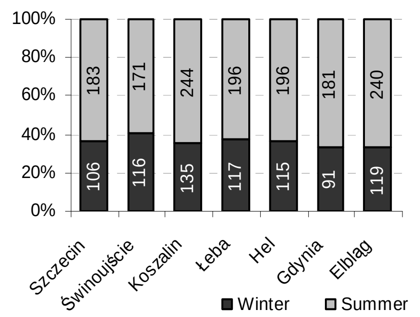

Fig. 2. The comparison of mean precipitation totals in a summer and in a winter
All measurements in millimeters

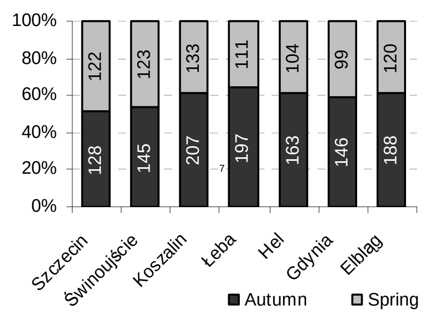

Fig. 3. The comparison of mean precipitation totals in an autumn and in a spring
All measurements in millimeters

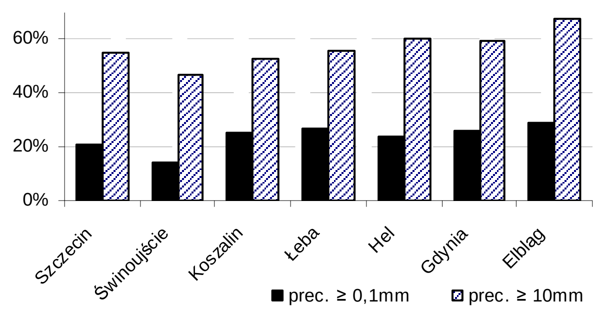

Fig. 4. Irregularity index calculated for monthly precipitation sums ≥ 0,1 mm per day and ≥ 10 mm per day (heavy precipitation)

The most regular monthly precipitation totals are noted in Świnoujście. The value of irregularity index for this station is one of the lowest in Poland (Kożuchowski and Wibig 1988). It is caused by both the Atlantic Ocean and Baltic Sea influences. The highest value of the index among coastal stations was obtained for Elbląg. Irregularity of heavy precipitation is significantly (two or three times) higher.

Pluvial continentality can also be expressed by means of semi-period of precipitation (Fig. 5).

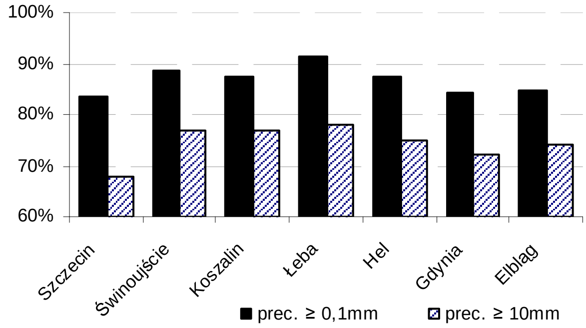

Fig. 5. Semi-period of precipitation

The Table 1. shows a semi-period of precipitation in more clear form. The days when precipitation totals calculate from the 1st of April reach a half of a yearly precipitation total are presented in it.

Table 1. Dates, on which half of an annual precipitation total (counted from the 1st of April) is reached at each station

	Szczecin	Świnoujście	Koszalin	Łeba	Hel	Gdynia	Elbląg
Prec. ≥ 0.1 mm	31 VIII	9 IX	7 IX	14 IX	7 IX	1 IX	2 IX
Prec. ≥ 10 mm	3 VIII	19 VIII	18 VIII	21 VIII	16 VIII	10 VIII	13 VIII

At the majority of stations situated on the Polish Baltic Sea coast, half of the annual precipitation total is achieved in the beginning of September (in Szczecin on the last day of August). The semi-period is longer in Łeba, where it lasts at least until the 15th of September. In the remaining part of Poland, the precipitation semi-period is shorter. Małopolska has the shortest semi-period – fall on the beginning of August (Kożuchowski and Wibig 1988). Similar relationships between particular stations also concern occurrences of heavy precipitation. Intensive rainfall is much more concentrated in a summer, hence the values of precipitation semi-period are lower. Specific seasonal course of precipitation in Łeba indicates a high degree of pluvial oceanity of this station, which is caused by the exposure of this part of the coastline for the offing sea.

Subsequent analyses were aimed at a more detailed presentation of the seasonal precipitation variability. They were focused on stations located at the opposite ends of the Polish Baltic Sea coast (Świnoujście and Elbląg), and the station displaying very high degree of pluvial oceanity (Łeba).

Two maxima of precipitation totals are observed in Świnoujście (Fig. 6). The highest sums of rainfall first time appear in the latter part of June and early part of July and second time at the end of August and at the beginning of September. The lowest values occur in the latter part of February.

The annual courses of variability in precipitation totals shown in Figures 6. – 14. were smoothed by 30-day moving averages.

The principal difference between the annual courses of precipitation totals in Łeba and Świnoujście is a decrease in the first precipitation maximum (in June in Świnoujście and in July in Łeba) and an increase in the second one (Fig. 7.). This causes the period of precipitation concentration to be shifted toward autumn.

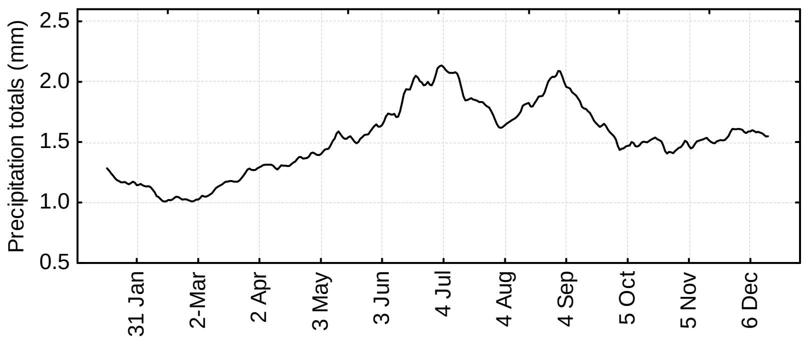

Fig. 6. Mean daily precipitation totals in Świnoujście in succeeding days of the year

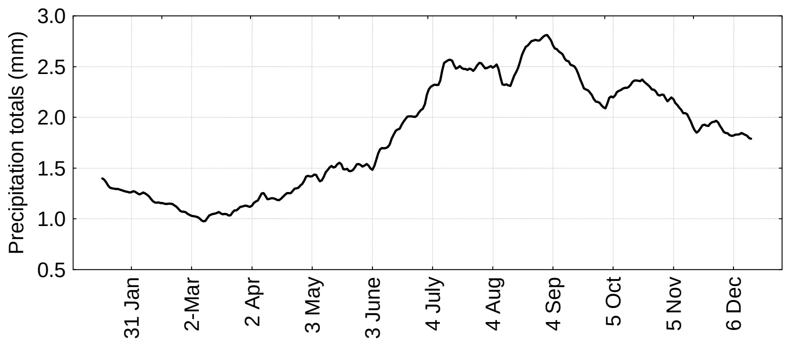

Fig. 7. Mean daily precipitation totals in Łeba in succeeding days of the year

Two maxima of precipitation totals are observed in Elbląg (like in Świnoujście) (Fig. 8.).

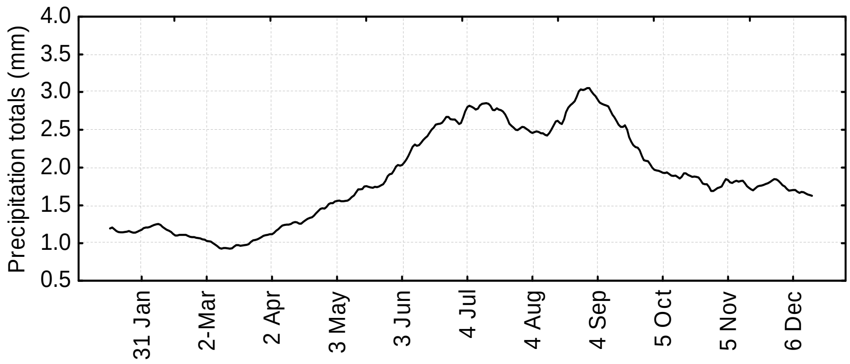

Fig. 8. Mean daily precipitation totals in Elbląg in succeeding days of the year

Annual variability of precipitation in particular circulation conditions

The presented patterns (double maxima in the annual course of precipitation totals in particular) were investigated in order to determine whether they are actual features of pluvial conditions or whether they result from differences in circulation conditions over subsequent multi-annual periods.

Intervals during which definite macroforms of circulation (associated with the zonal flow described by Zonal Index and North Atlantic Oscillation) prevailed over Europe and the whole Northern Hemisphere were classified for the period 1901-1998 by Degirmendžić *et al*. (2000). The least intense zonal flow and the most meridional advection simultaneously occurred from 1957 to 1970. A relatively strong eastern flow was observed during the period 1971-1986. Between the years 1987 and 1998 the strongest western circulation prevailed. Figures 9-14 present annual courses of daily precipitation totals in particular periods called "circulation epochs" by Degirmendžić et al. (2000) in Świnoujście and Elbląg.

As indicated by above graphs, the circulation conditions prevailing in the period 1987-1998 had the strongest influence on the occurrence of two precipitation total maxima in Świnoujście. The occurrence of two maxima in Elbląg is caused by other factors: an increased intensity of easterlies in the period 1971-1986, and an extraordinary intensity of westerlies in the period 1987-1998.

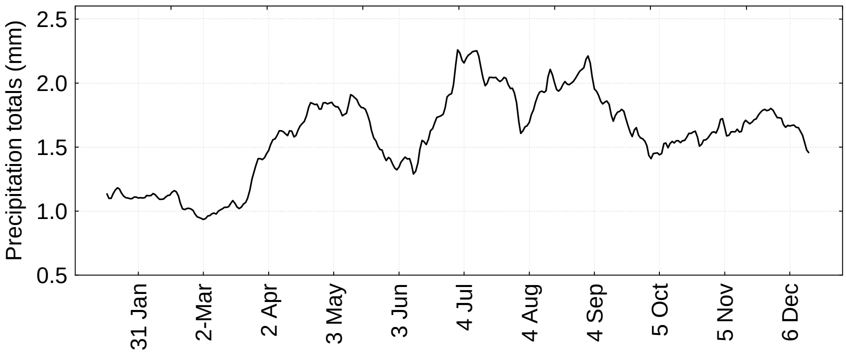

Fig. 9. Mean daily precipitation totals in Świnoujście during the period of a reduce the intensity of the zonal air mass advection (1957-1970)

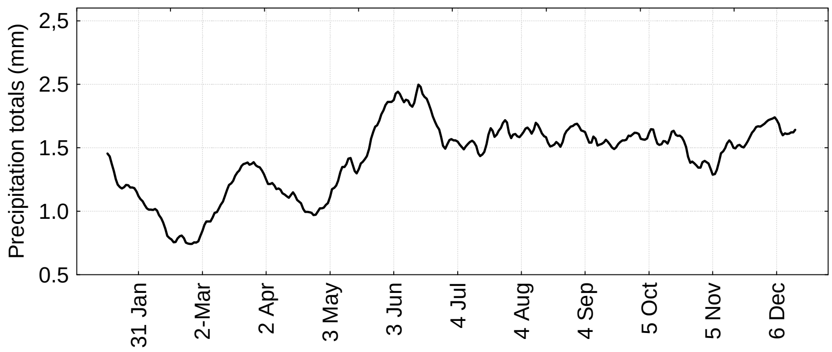

Fig. 10. Mean daily precipitation totals in Świnoujście in the period of an exceedingly intensity of the eastern air mass advection (1971-1986)

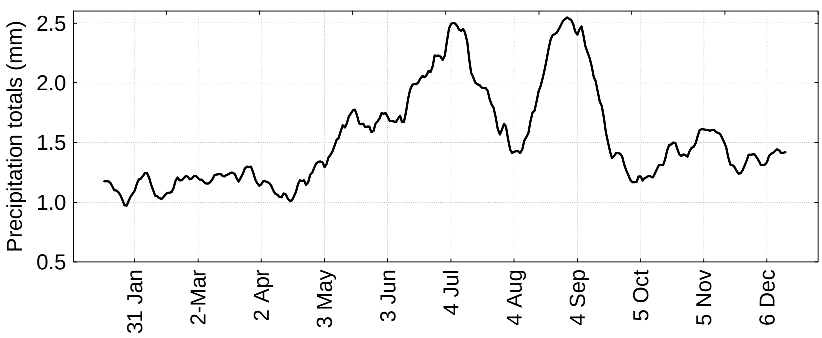

Fig. 11. Mean daily precipitation totals in Świnoujście in the period of exceedingly intensity of the western air mass advection (1987-1998)

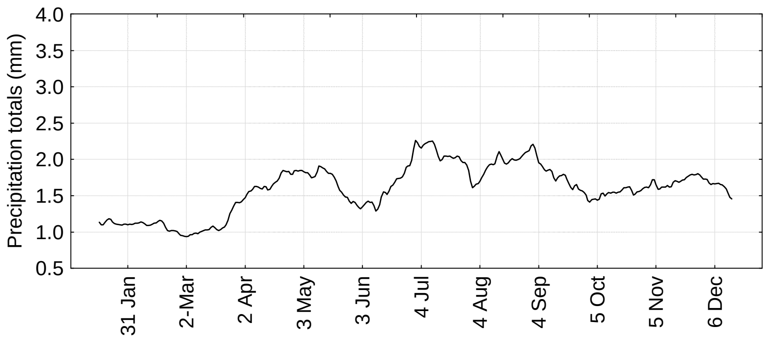

Fig. 12. Mean daily precipitation totals in Elbląg in the period of a reduce the intensity of the zonal air mass advection (1957-1970)

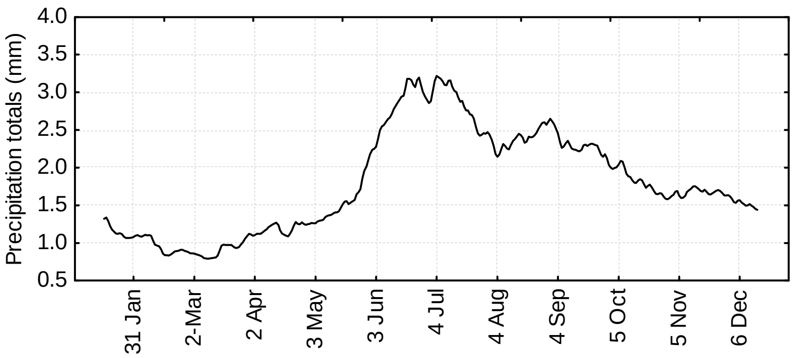

Fig. 13. Mean daily precipitation totals in Elbląg in the period of an exceedingly intensity of the eastern air mass advection (1971-1986)

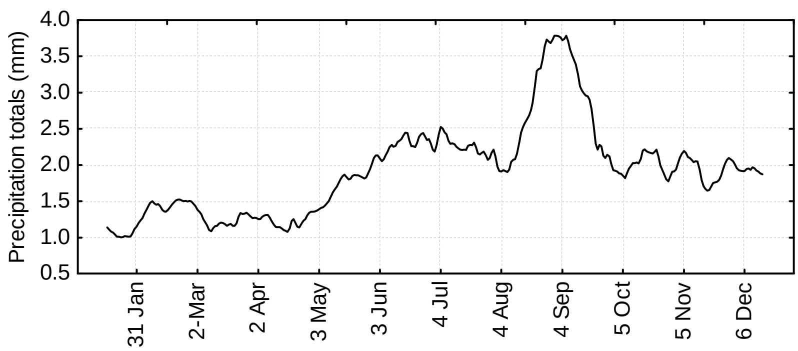

Fig. 14. Mean daily precipitation totals in Elbląg in the period of an exceedingly intensity of the western air mass advection (1987-1998)

CONCLUSIONS

1. The annual course of atmospheric precipitation at the stations located along the Polish Baltic Sea coast shows a strong influence of the Baltic Sea. Because of this, the study area differs from the remaining area of Poland with respect to the following: relatively low seasonal variability of rainfall; only very slight prevalence of warm half-year precipitation totals over the cold half-year values (especially in Łeba); significant prevalence of autumn precipitation over spring one (especially in terms of number of wet days); shift of the most intense precipitation period toward the autumn, and – consequently – a relatively low degree of pluvial continentality.

2. The occurrence of two precipitation maxima in summer, interrupted by a local minimum, in some cases makes the actual feature of pluvial climate of individual stations. At other stations however, the occurrence of two maxima results from averaging of different annual cycles describe periods with various circulation conditions.

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4. PRECIPITATION VARIABILITY IN THE MIDDLE ODRA RIVER BASIN IN THE YEARS 1951-2005

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INTRODUCTION

Precipitation is a meteorological element characterized by a significant temporal and spatial variation. Precipitation variability is an important index of climate changes, which is crucial to the natural environment as well as to socio-economic activity.

Research on long-term precipitation totals shows that tendencies of changes of this element of the climate have a very pronounced seasonal structure (Radziejewski *et al.* 2002, Kożuchowski 2004). In the Polish lowlands, the increase in air temperature occurred alongside an insignificant increase in total annual precipitation during the second half of the 20th century. Seasonally, an insignificant increase in spring and autumn precipitation totals were noted, and a corresponding decrease during the summer and winter periods were also observed (Żmudzka 2002). However, the month of March presented a statistically significant increase in monthly precipitation totals (Degirmendžić *et al.* 2004).

It has been shown that precipitation trends are not persistent. The length of the analysed period strongly influences the value of a given trend and significant trends may appear within shorter time periods of research (Niedźwiedź and Twardosz 2004, De Jongh *et al.* 2006).

A high spatial variation in precipitation changes, both on a local and regional scale, was also noted (Schönweise and Rapp 1997, Gonzalez-Hidalgo *et al.* 2001, New *et al*. 2001). However, in the Central Europe, precipitation changes are weaker than in other parts of the Continent (IPCC 2001).

The purpose of this paper is to determine the features of variability and trends in annual and seasonal precipitation total changes in the middle Odra River basin in the years 1951 to 2005.

The middle Odra River basin is characterised by a high variability in precipitation totals. The precipitation conditions of this area are influenced by circulation factors, the varied relief and elevation. The lowest annual precipitation totals (500-540 mm) were noted along the lowest part of middle Odra River valley as well as in the lower Kaczawa River valley. In the uplands located to the north-east and the south-west of the Odra valley axis, average annual precipitation totals normally increase proportionally to the hypsometric lie of the land. The Western Sudeten area, which is exposed to the flow of moist marine air, has the highest precipitation total, locally above 1350 mm (Głowicki *et al*. 2004).

MATERIAL AND METHODS

The source material for this analysis is comprised of precipitation data from 6 IMGW measuring stations, located in various physiographic regions of the middle Odra River basin (Fig. 1). Precipitation data set based on monthly totals covers the years 1951-2005. The data used in this study were taken from published precipitation yearbooks (years 1954 to 1981), as well as the archive of the Wrocław IMGW Branch.

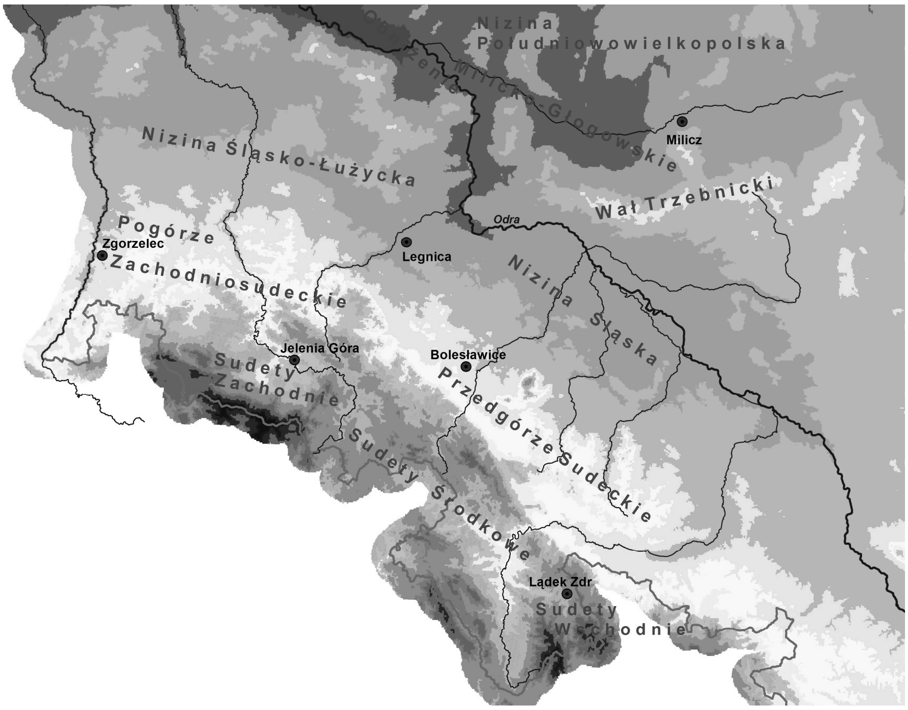

Fig. 1. Geographical location of measuring stations used in the study

The middle Odra River basin is characterised by a significant variation in environmental conditions. The study area, according to Kondracki's regionalization (1988), is comprised of the following physiographic regions: the Milicz-Głogów Depression, the Trzebnica Ridge, the Silesian and the Silesian-Lusatian Lowlands, the Sudetes Foreland, the Western Sudetes Foothills, as well as Western, Middle and Eastern Sudetes Mts. The location of the study area within several physiographic entities provides significant physiographic variability as well as strongly pronounced landscape zones.

The study of precipitation variability was carried out with the use of 30-year moving window, based on the analysis of average value, standard deviation, as well as 10% and 90% quantiles (Otop and Kuchar 2008). The 30-year moving windows were based on the precipitation series from the years 1951 to 2005. In the analysed 55-year period, 26 moving windows containing 30-year observational periods were created. The first period spanned the years 1951 to 1980, the last – 1976 to 2005. The statistical precipitation index values determined for the 30-year moving windows formed time series, which were then described through a linear function. The statistical significance of the determined linear regression equations at the $\alpha = 0.05$ level was tested through the Student's t-test. The analysis has been made for seasonal precipitation totals: spring (March-May), summer (June-August), autumn (September-November), winter (December-February) and for annual totals.

The empirical distribution of the precipitation totals was approximated through the gamma distribution with the shape and scale parameters (Pruchnicki 1987), estimated with the maximum likelihood method (Sneyers 1990). The values of the 10% and 90% quantiles of annual and seasonal precipitation totals were determined through the gamma cumulative distribution.

RESULTS AND DISCUSSION

Average annual precipitation totals from the analysed stations in the middle Odra River basin for the years 1951 to 2005 were rather varied. The lowest precipitation totals were noted in the Silesian-Lusatian Lowlands (Legnica – 535 mm). Significantly higher totals were observed in the Sudetes Mountains (Lądek Zdrój – 852 mm). Extreme values of annual precipitation totals differed significantly from the average ones (Tab. 1). Maximum totals amounted to 150-135% of the 55-year average and minimum totals were 71-56% average. A low degree of stability in precipitation totals is one of the characteristic features of the Polish climate. Seasonal precipitation totals have much higher variation than annual ones. The highest variation was observed in the autumn. The highest and the lowest precipitation totals were observed in the autumn, with respectively from 228% to 17% of the 55-year average.

In the study area, the highest percentage (ca. 40%) of the total annual precipitation is recorded in the summer. Spring precipitation is on average 23% of the annual total and is higher than the autumn precipitation total. Precipitation in the winter season is the lowest and it is about 15% of the annual total.

Between the years 1951 and 2005, average annual values of the precipitation variability coefficient for the analysed stations in the middle Odra River basin evened-out to 17-19% (Tab. 1). The variability coefficient reached its highest value in the autumn season (36% on average). A high value of the variability coefficient is a typical climate feature of the study area as it has been previously mentioned (Kosiba 1948). High precipitation variability often causes abnormal conditions characterised by precipitation totals which are either excessively low or high.

The 10% quantile values for the annual precipitation distribution in the study area in the years 1951-2005 are characterized by a high variability, from 410 mm to 678 mm. Similarly, the 90% quantile values are widely varied, ranging from 668 mm to 1037 mm. The 10% and 90% quantile values of the seasonal precipitation totals for the years 1951-2005 are shown in Table 1.

The course of average annual precipitation totals in the 30-year moving windows proves their tendency to decrease in the years 1951 to 2005, and linear approximation of average totals proves the occurrence of significant decreasing tendencies (Fig. 2). An increasing tendency in seasonal precipitation totals can only be observed in the winter, and only in the measuring stations (Lądek Zdrój, Jelenia Góra) located in the Sudetes Mountains and in the Sudetes Foreland – Bolesławice station (Tab. 2). The analysis of the change in the values of the standard deviation of the annual precipitation totals in moving windows showed that this measure of dispersion is increasing. The linear trend of the value of standard deviation annual precipitation is positive at the majority of the analysed stations, which proves their rising dispersion in relation to the average value. The observed trends in the changes of this variability index of seasonal precipitation are characterized by various sign of trend coefficient, the most emphasized one being the summer precipitation, which tended to increase its variability. The standard deviation of the winter precipitation totals also increased but only at the measuring stations located in the Sudetes mountains.

Table 1. Statistical characteristics of annual and seasonal precipitation totals in the middle Odra River basin in the years 1951-2005

Period	Station	Altitude (m a.s.l.)	Mean (mm)	δ (mm)	V (%)	Max		Min		Quantiles (mm)	
						Total (mm)	Year	Total (mm)	Year	10%	90%
Winter	Milicz	115	97	29	30	159	1967	37	1978	62	135
	Legnica	122	76	24	32	145	1987	39	1996	47	108
	Zgorzelec	203	109	33	30	190	1975	50	1964	70	153
	Bolesławice	240	78	25	32	139	1987	25	1990	48	111
	Jelenia Góra	342	102	30	29	175	1975	50	1991	66	141
	Lądek Zdrój	461	128	36	28	229	1992	61	1961	84	176
Spring	Milicz	115	127	37	29	225	1965	54	1990	82	177
	Legnica	122	126	38	30	234	1965	56	1993	81	175
	Zgorzelec	203	147	40	27	270	1961	70	1976	99	200
	Bolesławice	240	137	32	23	232	1965	80	1981	98	179
	Jelenia Góra	342	164	43	26	278	1967	92	1976	112	221
	Lądek Zdrój	461	204	51	25	387	1965	112	2002	142	271
Summer	Milicz	115	217	63	29	372	2001	107	1992	141	301
	Legnica	122	222	74	33	435	1977	101	1983	134	320
	Zgorzelec	203	233	68	29	378	1957	95	1982	151	323
	Bolesławice	240	247	78	32	425	1964	113	1983	154	351
	Jelenia Góra	342	279	84	30	559	1997	147	1990	178	391
	Lądek Zdrój	461	342	111	32	757	1997	150	1992	210	489
Autumn	Milicz	115	125	48	38	238	1963	23	1959	69	189
	Legnica	122	111	42	37	204	1976	23	1959	62	166
	Zgorzelec	203	139	49	35	232	1956	40	1982	81	204
	Bolesławice	240	123	43	34	201	1998	34	1959	73	180
	Jelenia Góra	342	143	50	35	298	1956	24	1959	84	209
	Lądek Zdrój	461	178	62	35	406	1952	49	1982	105	260
Year	Milicz	115	566	103	18	764	2001	363	1953	438	702
	Legnica	122	535	101	19	795	1977	361	2003	410	668
	Zgorzelec	203	629	114	18	838	1981	351	1982	487	779
	Bolesławice	240	586	100	17	837	2001	362	1969	462	716
	Jelenia Góra	342	688	119	17	1005	1977	462	1990	540	845
	Lądek Zdrój	461	852	141	17	1282	1997	606	1982	678	1037

V – variability coefficient, δ – standard deviation.

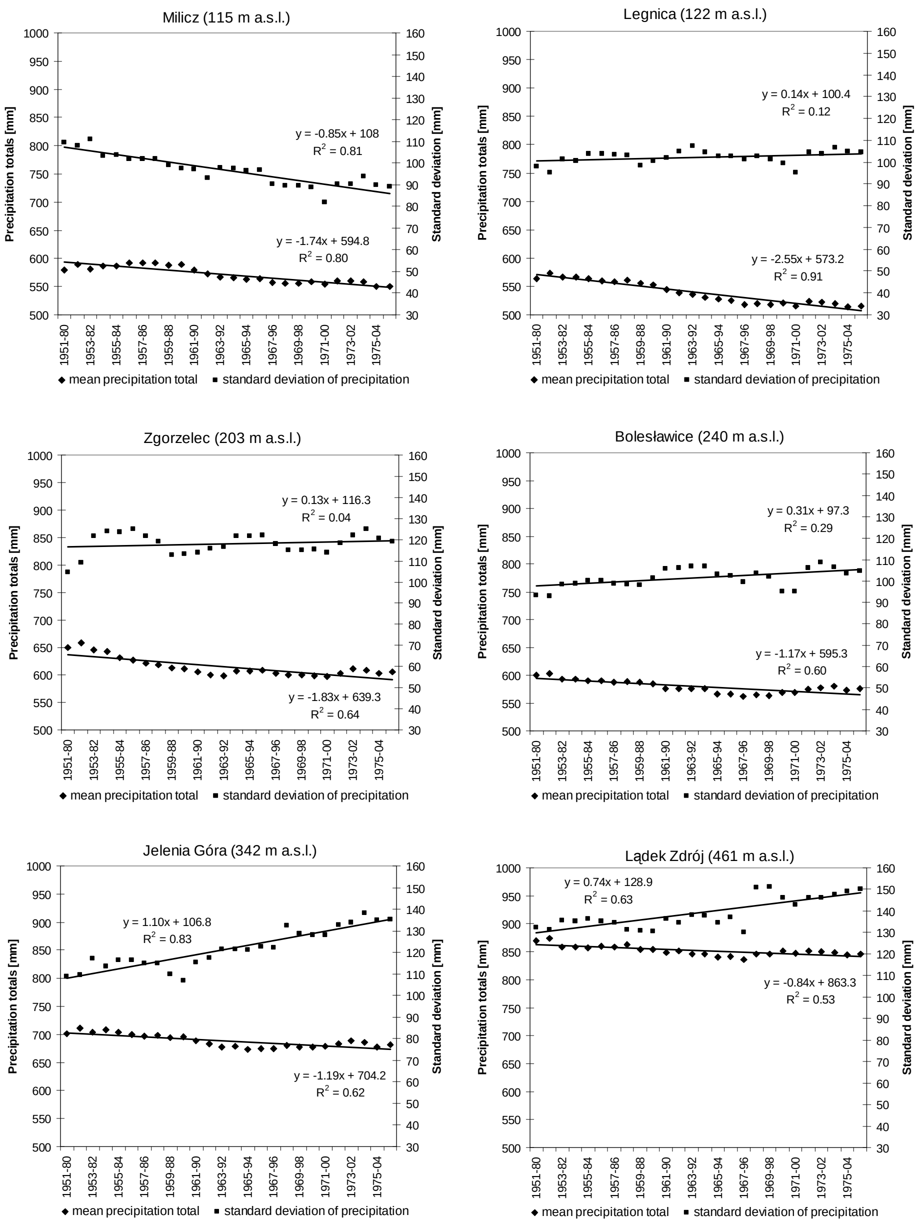

Fig. 2. Average and standard deviation of annual precipitation totals in the middle Odra River basin of 30-year moving windows from 1951-2005 and their linear trend

Table 2. Linear regression equations and determination coefficients (R^2) for statistical characteristics of seasonal precipitation totals in 30-year moving windows from 1951-2005 in the middle Odra River basin

Station	Mean		Standard deviation		Quantile 10%		Quantile 90%	
	Linear trend	R^2	Linear trend	R^2	Linear trend	R^2	Linear trend	R^2
Winter								
Milicz	–0.23x+101*	0.56	–0.16x+32*	0.82	–0.04x+62	0.02	–0.45x+144*	0.89
Legnica	–0.41x+83*	0.87	0.04x+26	0.09	–0.39x+52*	0.87	–0.41x+117*	0.58
Zgorzelec	–0.05x+107	0.04	–0.07x+36*	0.17	0.03x+65	0.00	–0.15x+155*	0.29
Bolesławice	0.07x+76*	0.17	0.08x+25*	0.23	–0.02x+47	0.01	0.17x+109*	0.24
Jelenia Góra	0.11x+101*	0.33	0.27x+28*	0.69	–0.19x+68*	0.35	0.48x+138*	0.74
Lądek Zdrój	0.23x+123*	0.46	0.29x+32*	0.70	–0.10x+84*	0.22	0.62x+165*	0.64
Spring								
Milicz	–0.50x+132*	0.62	–0.47x+41*	0.76	0.03x+83	0.01	–1.13x+187*	0.74
Legnica	–0.96x+140*	0.89	–0.34x+41*	0.61	–0.56x+91*	0.69	–1.42x+194*	0.86
Zgorzelec	–0.78x+156*	0.89	–0.33x+43*	0.66	–0.39x+104*	0.62	–1.22x+214*	0.86
Bolesławice	–0.44x+142*	0.84	–0.33x+38*	0.76	–0.05x+96	0.06	–0.88x+192*	0.84
Jelenia Góra	–0.84x+173*	0.89	–0.83x+53*	0.89	0.11x+109	0.10	–1.95x+244*	0.92
Lądek Zdrój	–0.42x+207*	0.72	–0.78x+60*	0.79	0.48x+134*	0.52	–1.48x+287*	0.83
Summer								
Milicz	–0.54x+226*	0.48	0.45x+55*	0.60	–1.02x+159*	0.86	0.08x+299	0.01
Legnica	–0.67x+230*	0.47	0.15x+76*	0.16	–0.51x+139*	0.54	–0.82x+331*	0.34
Zgorzelec	–0.64x+233*	0.36	–0.21x+73*	0.24	–0.39x+145*	0.27	–0.92x+330*	0.36
Bolesławice	–0.69x+252*	0.37	0.11x+76	0.05	–0.79x+160*	0.61	–0.54x+353	0.12
Jelenia Góra	–0.39x+283*	0.23	0.94x+75*	0.74	–1.39x+192*	0.92	0.89x+382*	0.29
Lądek Zdrój	–0.7x+358*	0.52	1.22x+92*	0.74	–2.0x+246*	0.90	0.96x+479*	0.30
Autumn								
Milicz	–0.46x+136*	0.61	–0.61x+54*	0.89	0.35x+70*	0.27	–1.47x+212*	0.80
Legnica	–0.50x+121*	0.89	–0.43x+46*	0.80	0.30x+60*	0.32	–1.51x+191*	0.88
Zgorzelec	–0.33x+143*	0.56	–0.34x+52*	0.56	0.05x+82	0.01	–0.80x+212*	0.68
Bolesławice	–0.12x+126*	0.20	–0.47x+48*	0.91	0.60x+65*	0.66	–1.03x+196*	0.80
Jelenia Góra	–0.05x+147	0.03	–0.54x+56*	0.71	0.64x+79*	0.51	–0.92x+224*	0.69
Lądek Zdrój	0.06x+176	0.01	–0.07x+56	0.02	0.13x+109*	0.24	–0.04x+250	0.00

* – statistically significant at 0.05 level.

Focusing on quantiles, the analysis of the changes in the 10% quantile values in the 30-year moving windows of annual precipitation totals for the years 1951-2005 showed a statistically significant decrease, whereas the 90% quantile values had varying sign of the trend coefficient. With the exception of the autumn, there were decreasing tendencies in the 10% quantile of seasonal totals. Spring and autumn precipitation showed a decreasing tendency in the 90% quantile value (Tab. 2). The course of 10% and 90% quantile values of the annual precipitation totals in the 30-year moving windows for the years 1951-2005, including the approximated trend line in the analysed measuring stations in the middle Odra River basin, is shown in Figure 3.

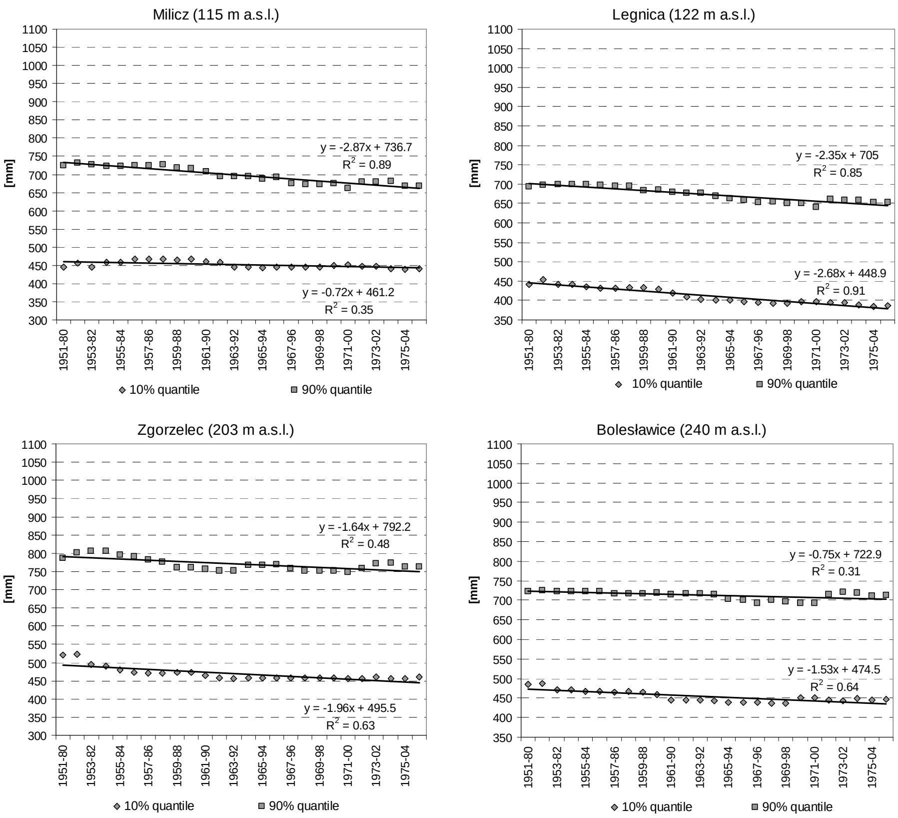

Fig. 3. Values of 10% and 90% quantiles of annual precipitation totals in the middle Odra River basin in 30-year moving windows from 1951-2005 and their linear trend

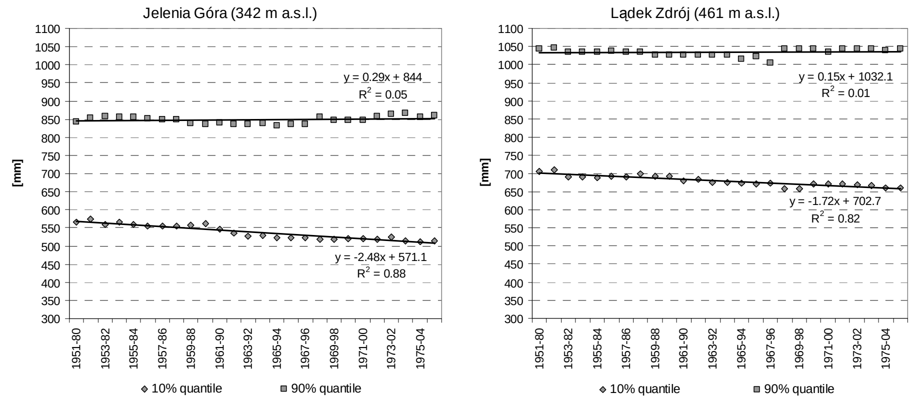

Fig. 3. Cont. Values of 10% and 90% quantiles of annual precipitation totals in the middle Odra River basin in 30-year moving windows from 1951-2005 and their linear trend

Set trends in the changes of average precipitation totals in the 30-year moving windows are not fully comparable with the results of other research conducted in the area of precipitation change tendencies in Poland. This is due to differences in methodology, as well as differences in the time periods studied. However, the results show a certain degree of similarity between the tendencies of change described in the present paper and precipitation variability research conducted in Poland in the second half of the 20th century. The tendencies of change in lowland part of the study area shows homogenous sign of trend coefficient for summer and winter periods with the results obtained for averaged precipitation totals from the Polish lowlands (Żmudzka 2002). However, differences in the methodological approach of these two studies made the comparison of value in their regression equations unable. Only a comparison of their direction has been possible.

CONCLUSION

This research shows some feature of the precipitation regime, especially the tendencies of seasonal and annual precipitation changes in the study area of the middle Odra River basin.

Between the years 1951 and 2005, based on the analysis of 30-year moving windows, the systematic decrease of annual and summer precipitation totals occurred alongside an increase in their variability, which was expressed by the standard deviation value. Moreover, 10% quantile values were characterized by a downward tendency, which indicates an increase in precipitation deficiency during negative anomaly periods. As far as the spring and autumn periods are concerned, a downward tendency of both the precipitation totals and their standard deviation values were observed. On the other hand, the trends of the 10% quantile values varied, and showed a downward tendency in the spring at the majority of the analysed stations, but upward in the autumn. In the winter though, precipitation totals from the stations located in the Sudeten mountains showed an increase in totals, variability and also in 90% quantile value.

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5. CHARACTERISTICS OF METEOROLOGICAL DROUGHTS IN WROCŁAW-SWOJEC IN THE YEARS 1964-2006

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INTRODUCTION

Meteorological drought is a long time period characterized by the lack of precipitation, low air moisture and high air temperature. Droughts are tightly connected with climate. The data from reports concerning climate changes warn against extreme conditions, including drought. The latter is related to water deficit, causing damage to natural environment and economics, as well as danger for population. Drought is a complex condition, difficult to be defined. It is usually assessed retrospectively from past records, when the process had already been finished. Drought indices in both Polish and international literature are described as a single value concerning specified region. Due to the fact that drought is often treated as a result of many factors influencing each other, our work contains the description of two drought indices: standardized precipitation index (SPI) and standardized climatic water balance (SCWB).

METHODS

In our work, based on a homogenic material collected for many years, there is a verification of two drought indices, defined with aid of precipitation and potential evapotranspiration. The first one is SPI (Standardized Precipitation Index). It was calculated based on half-year (IV-IX) precipitation sums in the years 1964-2006.

Since periodical (monthly, half-year) precipitation sums are gamma distributed data (Kaczmarek 1970), they were converted to a normal distribution using different data transformations, after which the chi- squared goodness of fit test could be used to verify normality of the data. Then transformed, normally distributed data were standardized to normal standard distribution N(0,1). For such data the Standarized Precipitation Index (SPI index) is calculated. The SPI index was introduced by McKee, Doesken, Kleist (1993) and is useful to quantitative assessment of precipitation deficit in a various time scale.

$$SPI = \frac{z - \bar{z}}{\sigma_z}, \tag{1}$$

where $\bar{z}$ and σ_z denotes the mean value and standard deviation respectively of normalized sequence of precipitation sums.

In the Table 1 the classification system of drought intensity is introduced:

Table 1. Drought classification according to the standardized precipitation index (SPI) and corresponding probabilities

Value of SPI	Drought category	Probabilities
SPI ≤ –2.0	extreme	P(SPI<–2) = 0.023
–2.00<SPI ≤ –1.50	severe	P(–2<SPI<–1.5) = 0.0443
–1.50<SPI ≤ –1.00	moderate	P(–1.5<SPI<–1) = 0.092
–1.00<SPI< 0.0	mild	P(–1<SPI<0.0) = 0.341

SPI classification is used to define the intensiveness of drought periods for any time scale. Every drought is characterized by a constantly negative SPI value, whereas SPI positive value is equivocal to the end of drought. Table 1 shows the probability of appearance of every type of drought. The probability of extreme drought, assuming that the distribution is normal, is 0.02, of mild drought – 0.34. The probability of lack of drought is 0.5.

The description of meteorological drought should contain, apart from precipitation, the information about other meteorological factors, like temperature, sunshine, the saturation, the vapour pressure deficit, the wind speed. The above mentioned markers compose the process of potential evapotranspiration (Musiał 2002, Gąsiorek *et al.* 2008), defined in this work by the Penman method. Therefore, the second marker to characterize drought is climatic water balance (CWB) (Łabędzki 2006). Potential evapotranspiration and precipitation, together with the measurement of climatic water balance (Rojek 1987), may point at the climatic excess or deficit of precipitation.

$$CWB = P - ET_0 \tag{2}$$

where: *CWB* – climatic water balance, P – precipitation sum in examined time period, ET_0 – vapotranspiration according to the Penman method.

The first step of investigation is to calculate the potential evapotranspiration according to the Penman method. The value of potential evapotranspiration is described by the following equation:

$$ET_0\ (mm) = \frac{\frac{\Delta}{\gamma}(R_n + G) + E_a}{\left(1 + \frac{\Delta}{\gamma}\right)} \frac{n}{28.34} \tag{3}$$

where: R_n – net radiation (W m^{-2}); G – soil heat flux density (W m^{-2}); E_α – aerodynamic or ventilation term (W m^{-2}); Δ – mean rate of change of saturated vapour pressure with temperature, (hPa K^{-1}); γ – psychrometric constant (0,655 hPa K^{-1}); n – number of days in period for which the calculations were done (decade, month). In the Penman equation the following information has to be taken into account: all outgoing energy fluxes are negative, whereas those toward the active surface are positive.

Since climatic water balance (CWB), proposed by Łabędzki (2006), is not standardized, it cannot be compared with SPI as a drought index. Random variable, the values of which are sums of climatic water balance using various time steps, is usually normally distributed. Normally distributed sequences of precipitation sums were converted through standardization to standardized normal distribution N(0,1). This standardization allows to obtain the values of standardized climatic water balance SCWB.

$$SCWB = \frac{CWB - \overline{CWB}}{s_{CWB}} \tag{4}$$

where $SCWB$, $\overline{CWB}$ and s_{CWB} denotes the standardized climatic water balance, mean climatic water balance and standard deviation of climatic water balance, respectively.

It is assumed, that the introduced classification of drought system, based on SCWB, is the same as the classification for SPI (Tab. 1).

RESULTS

The characteristics of years 1964-2006 in Wrocław-Swojec was commenced by the presentation of relations between the precipitation sums and the sums of potential evapotranspiration in the IV-IX period. In the years 1964-2006 the values of potential evapotranspiration sums far exceeded those of precipitation sums (Fig. 1). The above mentioned relations are the result of constant precipitation deficit in the Wrocław-Swojec region.

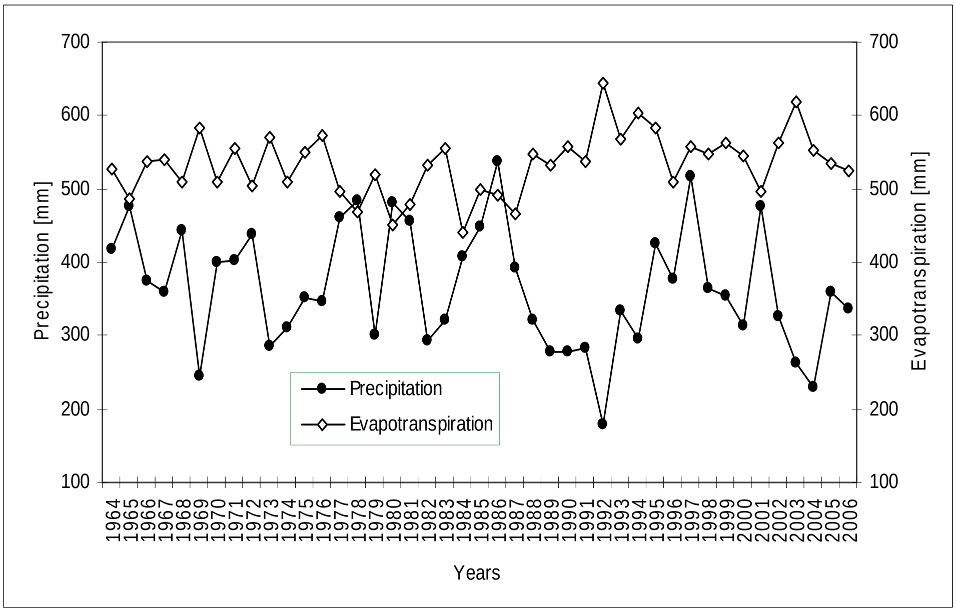


Fig. 1. Precipitation sums and potential evapotranspiration in the IV-IX period in Wrocław-Swojec in the years 1964-2006

Drought is always related to the temperature increase and the precipitation deficit, hence the subsequent analysis concerned the change in mean temperature within the warm half-year in the years 1964-2006. The regression equation, calculated for the mean temperature within the IV-IX period of the years 1964-2006 has the following form: Y = 0.042 X – 68.53 and is statistically significant on significance level $\alpha = 0.05$. Slope of regression line, equal to 0.042, indicates that the mean warm half-year in the 1964-2006 period has an increasing tendency with a temperature growth equal to 0,042^{o}C per year. The regression line with 95% confidence interval for the mean temperature in the warm half-year is presented in Figure 2.

In order to calculate the SPI index (Łabędzki *et al.* 2002, Paulo *et al.* 2006, McKee *et al.* 1993, 1995), the precipitation sums of the IV-IX period in the years 1964-2006 were analyzed.

The analysis of changes in the half-year precipitation sums was performed in a similar way. The regression equation calculated for the half-year precipitation sums in the 1964-2006 period has the following form Y = –1.6X + 3540 and indicates the decreasing tendency of this meteorological factor in the analyzed period (Fig. 3).

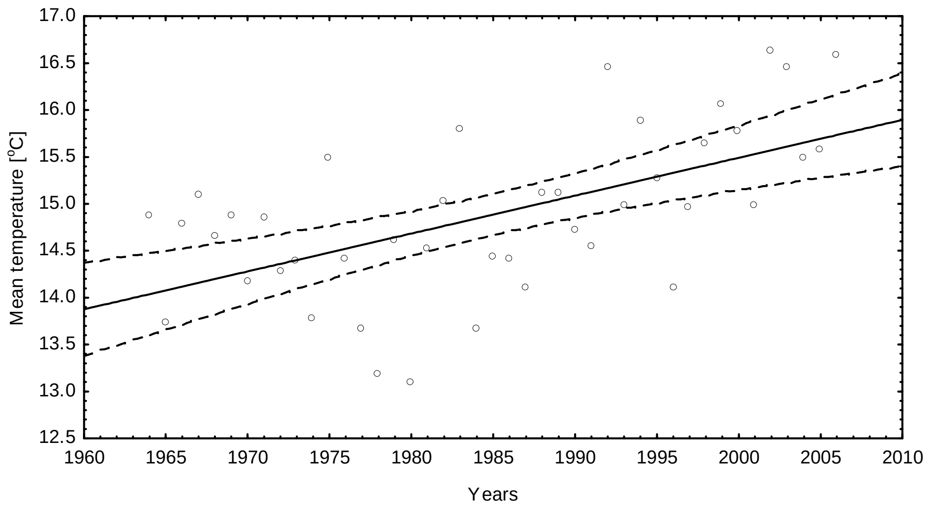


Fig. 2. Mean temperature in the IV-IX period in the years 1964-2006 in Wrocław-Swojec

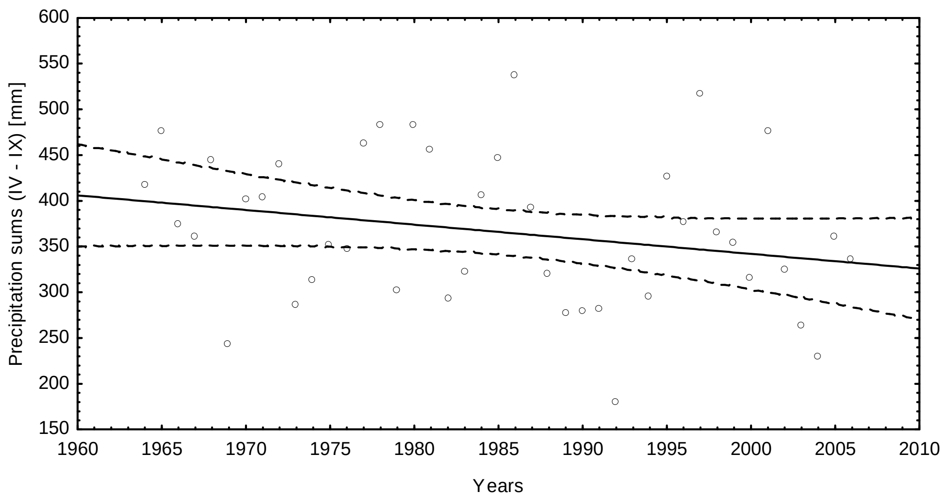
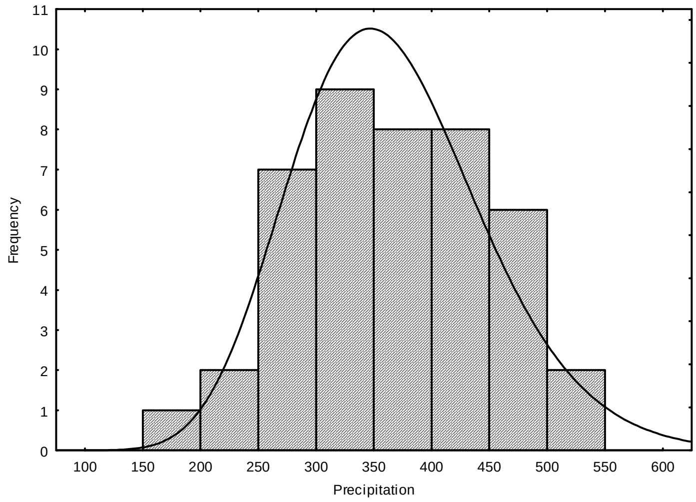
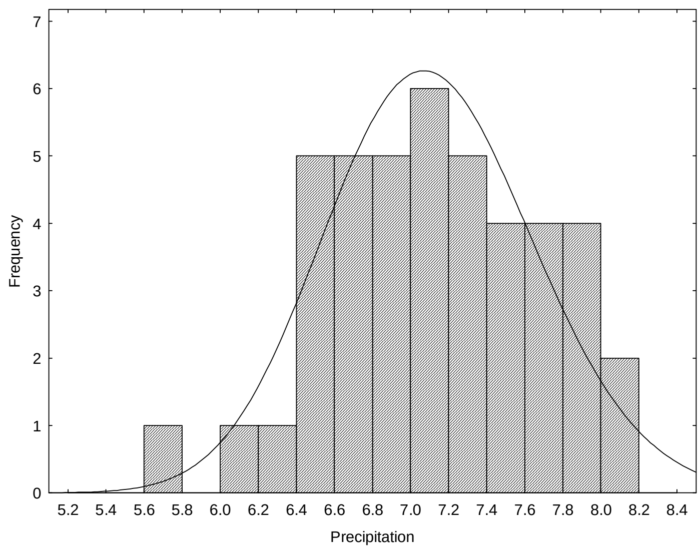


Fig. 3. Precipitation sums in the IV-IX period in the years 1964-2006 inWrocław-Swojec

Since half-year precipitation sums are gamma distributed data (Fig. 4a) they were converted to a normal distribution using different data transformations (Fig. 4b-d).

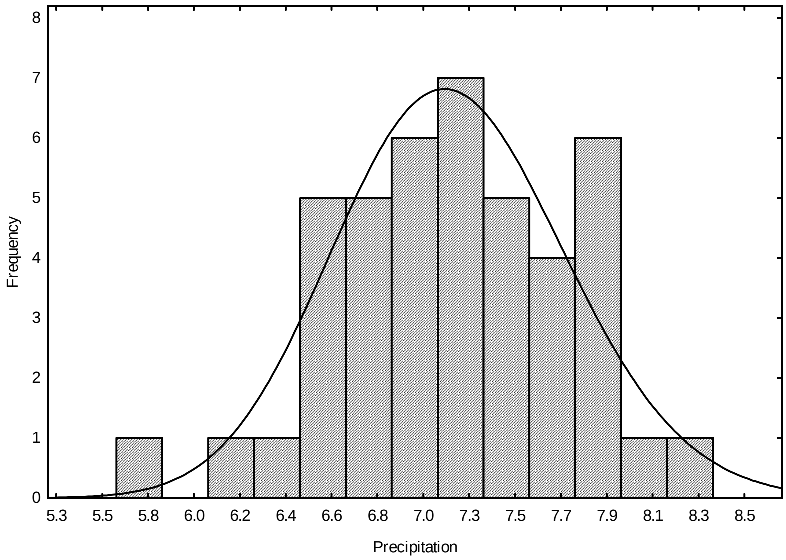


a)

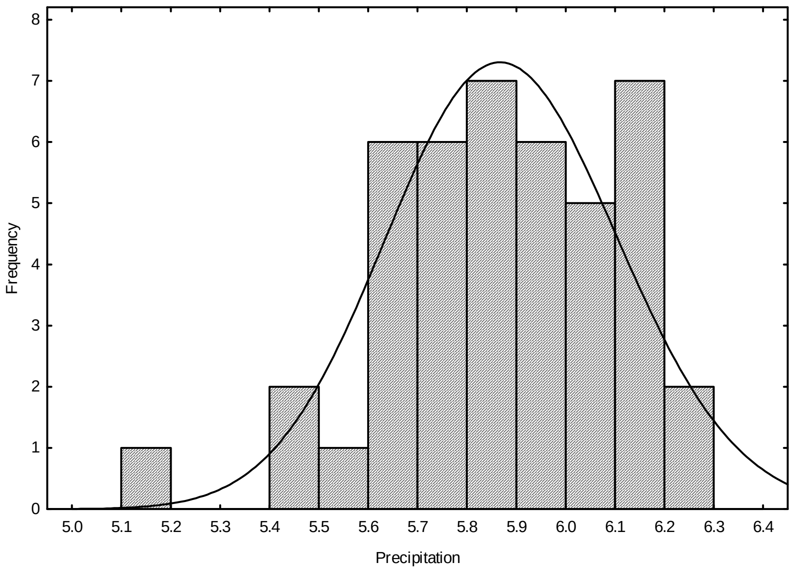


b)

Fig. 4. Fitting distributions with the half-year precipitation sums in years 1964-2006 in Wrocław-Swojec:
a). the original Gamma distribution (p = 0.65),
b). normal distribution due to the transformation $Y = \sqrt[3]{P}$ (p = 0.87),
c). normal distribution due to the transformation $Y = \sqrt[3]{P+10}$ (p = 0.95),
d). normal distribution due to the transformation $Y = \ln P$ (p = 0.80).



c)

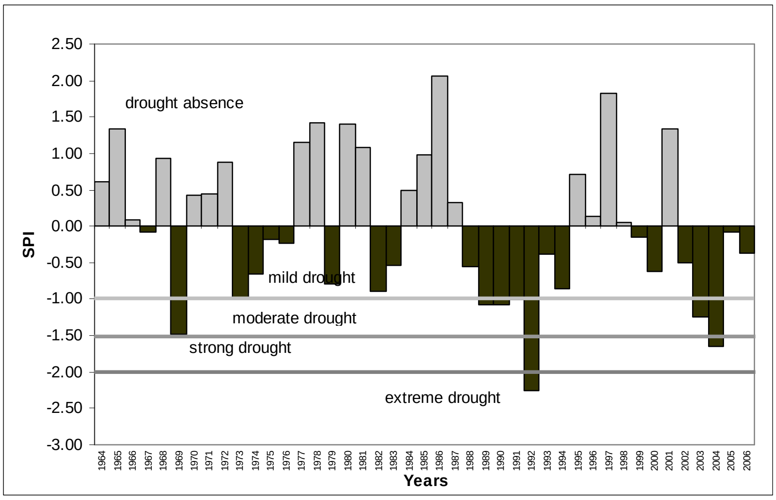


d)

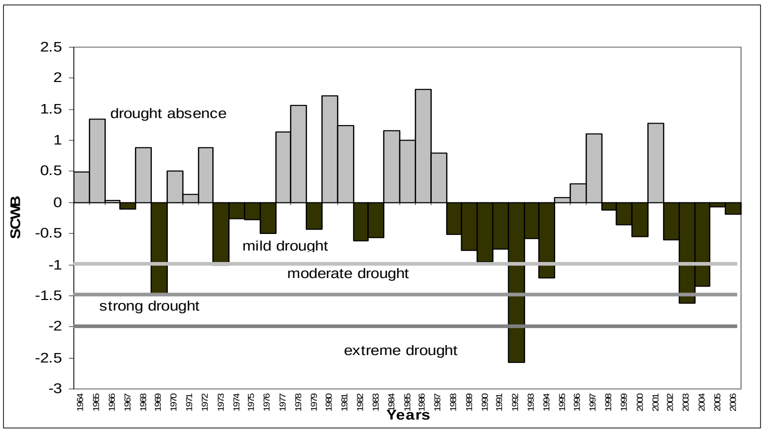
Fig. 4. Cont. Fitting distributions with the half-year precipitation sums in years 1964-2006 in Wrocław-Swojec:
a). the original Gamma distribution (p = 0.65),
b). normal distribution due to the transformation $Y = \sqrt[3]{P}$ (p = 0.87),
c). normal distribution due to the transformation $Y = \sqrt[3]{P+10}$ (p = 0.95),
d). normal distribution due to the transformation $Y = \ln P$ (p = 0.80).

The calculated SPI index defined the given half-year in different categories of drought or its absence. Random variable, the values of which are the half-year (IV-IX) sums of climatic water balance (CWB), is normally distributed ($CWB_{IV\text{-}IX}$ ~ N(-124; 124). Thus, for the calculation of SCWB for the warm half-year only standardization of the random variable was needed, and those standardized values classified the droughts according to the criteria put in the Table 1.

The drought classification, according to SPI and SCWB, and intensity in the warm half-year of the 1964-2006 period, is shown in the Figures 5a and 5b.



a)



b)

Fig. 5. Drought indices for the (IV-IX) period in the years 1964-2006 for Wrocław-Swojec a) SPI; b) SCWB

There is one-to-one correspondence between precipitation and SPI. Precipitation is a quantile of the same order of the distribution of the half-year precipitation sums as the corresponding SPI index in the standardized normal distribution. The same situation takes place in the case of SCWB.

The calculated quantile values for half-year precipitation sums can be interpreted in the following way: extreme drought in the warm half-year in Wrocław-Swojec appeared when the precipitation sum was less than 217 mm (Tab. 2). Strong drought was present during years 1964-2006 in those half-year periods, where the values of precipitation sums were between 217 and 283 mm. When precipitation sum was higher than 360 mm, there was no drought.

Table 2. Drought classification with the use of SPI in the IV-IX period in the years 1964-2006 in Wrocław-Swojec

Drought category				
Drought absence	Mild drought	Moderate drought	Strong drought	Extreme drought
SPI≥0	–1≤SPI<0	–1.5≤SPI<–1.0	–2.0<SPI<–1,5	SPI≤–2
Quantiles (mm)				
P≥360	283≤P<360	249≤P<283	217<P<249	P≤217
Clasification (IV-IX) periods in years 1964-2006				
1964,1965,1966 1968,1970,1971 1972,1977,1978 1980,1981,1984 1985,1986,1987 1995,1996,1997 2001	1967,1973,1974, 1975,1976,1979, 1982,1983,1993, 1994,1998,1999, 2000,2002,2005, 2006	1969,1989, 1990,1991, 2003	2004	1992

The calculated quantile values for half-year sums of climatic water balance in years 1964-2006 in Wrocław-Swojec can be interpreted in the following way: extreme drought in the warm half-year appeared when the climatic water balance sum for that period was less than 399 mm (Tab. 3). When the sum of climatic water balance was lower than 167 mm, drought was absent.

Table 3. Drought classification with the use of SCWB in the IV-IX period in the years 1964-2006 in Wrocław-Swojec

Drought category				
Drought absence	Mild drought	Moderate drought	Strong drought	Extreme drought
SCWB≤0	–1≤SCWB<0	–1.5≤SCWB<–1.0	–2.0<SCWB<–1.5	SCWB≤–2
quantiles (mm)				
CWB≤-167	–283≤CWB<–167	–341≤CWB<–283	–399<CWB<–341	CWB≤–399
Clasification (IV-IX) periods in years 1964-2006				
1964,1965,1966, 1968,1970,1971, 1972,1977,1978, 1980,1981,1984, 1985,1986,1987, 1995,1996,1997, 2001	1967,1974,1975, 1976,1979,1982, 1983,1988,1989, 1990,1991,1993, 1998,1999,2000, 2002,2005, 2006	1969,1973, 1994,2004	2003	1992

CONCLUSIONS

1. In the years 1964-2006 in Wrocław-Swojec, the values of potential evapotranspiration sums, calculated by the Penman method, far exceeded those of the precipitation sums in the IV-IX period. Those relations point at the constant precipitation deficit in the analyzed region.
2. The mean air temperature in the IV-IX period in the years 1964-2006 in Wrocław-Swojec was characterized by the increasing tendency, whereas the half-year precipitation sums in the examined period tended to decrease.
3. According to the SPI classification criteria, the drought characterization in the years 1964-2006 in Wrocław-Swojec is as follows: 47% of the IV-IX periods were devoid of drought, in 37% of cases the drought was mild, 12% presented the moderate drought, and 2% were classified as strong and extreme drought. The SCWB classification gave similar results.
4. The quantification of two indexes, SPI and SCWB, proposed for the drought identification, gives almost identical results. Therefore, the authors suggest the evaluation of drought based on SPI, requiring only the precipitation data.

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6. SNOW COVER OCCURRENCES IN POLAND

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INTRODUCTION

Snow cover, which appears during winter seasons in Polish climatic conditions, is a good indicator of thermal and precipitation conditions of winter seasons. Because of the specific physical characteristics of snow cover such as high albedo, low thermal conductivity and high emission capacity, it is also an important factor that determines thermal conditions of both ground-level air and the ground on which it lingers. Economic aspect of snow cover is also of considerable concern, especially in the fields such as agriculture, transportation, construction industry, tourism and winter recreation. Snow cover, therefore, became the focus of study for many researchers, which finds its reflection in literature on the subject. The history of observation of snow cover is much shorter than observation of other climate elements such as air temperature or atmospheric pressure. The beginning of systematic measurements of snow cover in Poland is dated in the last decade of the 19th century (Kosińska-Bartnicka 1924), while studies that included the entire area of Poland started many years after the Second World War (Bednorz 2001).

MATERIALS AND METHODS

The purpose of this investigation was to characterize climatic conditions of snow cover occurrences in non-mountainous regions of Poland in 1951-2008 (57 winter seasons). The research, therefore, omits Sudetes range and Carpathians Mountains. The analyzed data was collected from 83 meteorological stations, which are part of the meteorological observation network of the Institute of Meteorology and Water Management (Fig. 1). Snow cover occurrences were characterized with the following indices:

- number of days with snow cover (≥1 cm) in October-May season,
- frequency of days with snow cover (≥1 cm) in period of December-March,

- number of days with snow cover thickness of >5 cm, >10 cm, >20 cm in October-May season,
- duration of snow cover occurrence,
- the maximum thickness of snow cover in October-May season,
- mean snow cover thickness in December-March,
- total thickness of snow cover in October-May season.

Fig. 1. Distribution of meteorological stations

The parameters were analyzed for entire seasons (October-May) or periods (December-March) without breaking them into smaller units (e.g. months). Numerical values of the analyzed parameters from selected stations representing particular region are presented in tables.

RESULTS

Number of days with snow cover (≥1 cm) in October-May season

The most common parameter of snow cover occurrence is the number of days with snow cover in a season. This simple parameter gives a reliable indication of climatic variability of the analyzed element. The number of days with snow cover in season shows long term fluctuations from below 35 days in the Szczecinska Lowland and the lower Odra River to over 90 days in the north-east part of the country (Fig. 2). The mean value of this parameter for data from 83 stations is little less than 61 days (Tab. 1). Spatial variability of this indicator is rather high, since snow cover lingers almost three time longer in Suwalki and east Podlasie regions than in north-west fringes of Poland.

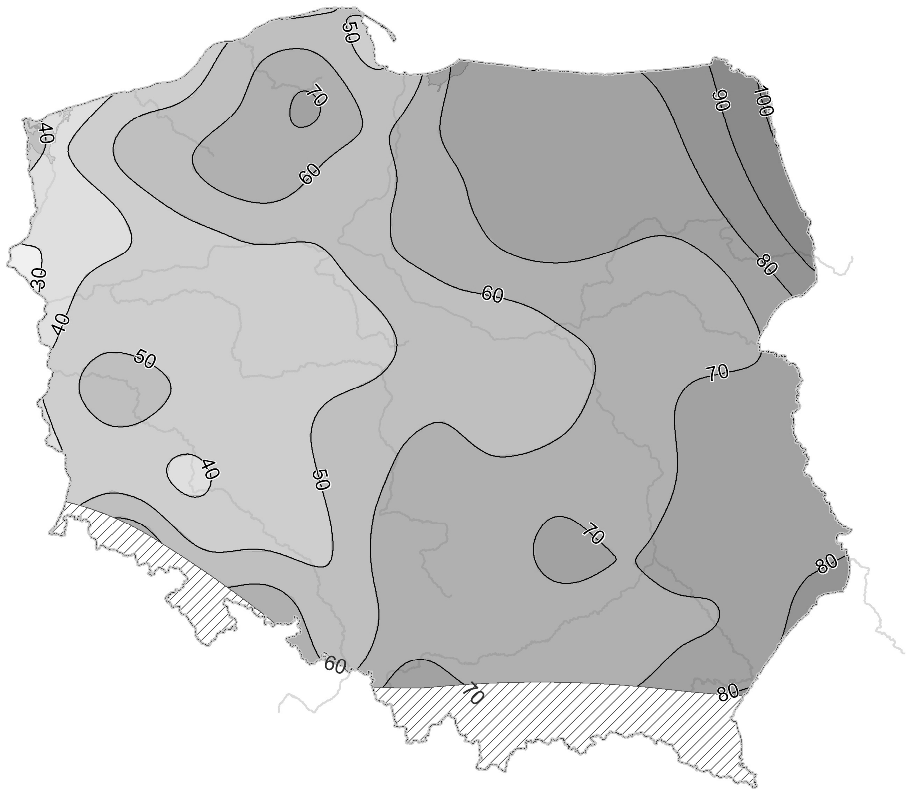

Fig. 2. Mean annual number of days with snow cover (≥1 cm) in October-May season in 1951-2008

Table 1. Number of days with snow cover (*sc*): totals and above the specified threshold values in 1951-2008

Region	Exemplary stations	No. of days with snow cover (October-May)				Days with (%) *sc*≥1 cm (Dec-Mar)
		≥1 cm	>5 cm	>10 cm	>20 cm	
Pomerania, Warmia, Masurian Lakeland, Suwalki Region	Świnoujście	43.2	20.1	11.4	3.9	33.6
	Resko	51.6	26.5	15.0	5.7	39.0
	Ustka	51.9	28.9	17.5	6.6	39.9
	Hel	50.9	32.0	20.6	8.6	40.2
	Chojnice	66.1	35.8	20.2	7.8	50.7
	Olsztyn	76.0	51.8	33.9	13.7	57.0
	Suwałki	91.4	66.5	49.8	27.7	67.6
Lowlands	Słubice	34.8	14.2	6.7	0.9	27.1
	Poznań	46.8	19.5	10.8	2.2	36.3
	Zgorzelec	45.7	23.4	14.1	4.4	35.0
	Wrocław	43.4	21.1	10.3	2.2	33.2
	Racibórz	56.2	28.7	14.5	3.0	42.1
	Łódź	60.6	32.7	19.2	7.3	45.8
	Warszawa	57.4	29.8	17.2	7.2	43.9
	Białystok	83.1	56.5	39.1	18.1	62.7
	Terespol	72.5	46.1	28.1	9.4	55.5
Uplands and Carpathian piedmont basins	Kielce	72.6	46.2	27.5	9.8	54.5
	Lublin	75.6	48.8	30.8	11.8	57.0
	Kraków	63.6	36.6	23.9	8.2	48.0
	Rzeszów	69.3	42.2	26.3	8.3	52.1
Poland (83 stations, non-mountainous area)	Mean value	**60.8**	**35.5**	**21.6**	**8.3**	**46.1**
	Min	33.2 (Przelewice)	12.5 (Przelewice)	6.1 (Przelewice)	0.9 (Słubice)	26.0 (Przelewice)
	Max	91.4 (Suwałki)	67.7 (Białowieża)	50.9 (Białowieża)	27.7 (Suwałki)	67.6 (Suwałki)

In each particular winter season, the number of days with snow cover could deviate considerably from the mean value. In the least snowy winters, in the west part of the country, the number of such days was less than 10, and even less than 5 at some stations. In the more snowy regions, however, during winters with extremely low snow, the number of days with snow cover did not drop below 30. During more snowy winters, the duration of snow cover could be more than three times longer than the norm in regions with the least snow in Poland. The longest periods with snow cover during winter season lasted over 135 days with the highest values of about 145 and more (the maximum – 156 days in Białowieża).

Frequency of days with snow cover (≥1 cm) in December-March periods

During the period of December-March, the average percentage of days with a snow cover was 46% in Poland. The lowest values of this parameter were registered in the area of the lower Odra River region – less than 30% and the highest in the Suwalki and east Podlasie regions, where days with snow cover accounted for the 2/3 of the entire period (December-March). During the least snowy winters in regions with higher snow cover conditions, the percentage of days with snow was no lower than 20%. During the most snowy winters, snow cover was present 100% of days, while it did not reach 90% at many stations located in the southern parts of Poland. This is explained by the fact that spring season approaches from the south and south-west of Poland as well as a shorter time of snow cover presence in March.

Number of days with snow cover of >5 cm, >10 cm, >20 cm in October-May season

The mean number of days with snow cover of >5 cm was 35.5 for the entire Poland which is 58% of the total number of days with snow cover (Tab. 1). In case of this parameter. the spatial distribution contrasts were very visible. In the lower Odra River region, the number of such days was, on the average, less than 15, while in the north-east fringes of Poland, it increased to over 65. Days with snow cover of >5 cm constituted c. 40% of all days with snow cover in the west fringes of Poland, and over 70% in the north-west and piedmont areas. In the western parts of Poland, the number of such days was 20-25, in central Poland 25-35, and in highlands and Mazurskie Lakeland – 45-55.

The mean number of days with snow cover of >10 cm was 21.6 for the area of Poland, which was 36% of the entire snow cover duration (Tab. 1). Spatial distribution contrasts of long term mean values of this parameter were similar to those of snow cover of >5 cm, however, the most snowy winters could be clearly identified. The number of days with snow cover of >10 cm on the average varied from 6-7 in the area of the lower Odra River up to 50 in the north-west fringes of Poland. The number of such days was 10-15 in the Wielkopolska region and the Silesian Lowland, 15-20 in the central Poland and ca. 25 or more in the highlands. Year-to-year variation of this parameter was very high and, in some seasons, days with snow cover of >10 cm did not occur at all. Statistically, in the west parts of the analyzed area, snow cover was no deeper than 10 cm almost every second winter. In the south-east and in inner-mountain and piedmont areas such winters were very rare.

The mean number of days with snow cover of >20 cm was 8.3 for the area of Poland, which was 14% of the total number of days with snow cover.

Duration of snow cover occurrences

In Poland, with the exception of mountain areas, snow cover appears, on the average, from November to March (inclusive), and in Masurian Lakeland, Suwalki and Podlasie regions, up to the first decade of April. There were, however, incidents where snow cover appeared already in October or as late as in April, and even May.

On the long-term average, snow cover first appeared in north-east fringes of Poland between November 10 and 15 (Tab. 2, Fig. 3). In the north part of the Masurian Lakeland, Podlasie and Suwalki regions as well as in the south-east part of the Lubelska Highland, snow cover was observed no later than November 20. By the end of November, snow cover presence was registered in all other parts of Poland with the exception of parts of the Silesian Lowland, Wielkopolska, Kujawy and Szczecinska Lowland. On the average, the first observations of snow cover in the area of the lower Warta and Noteć rivers and west parts of the Szczecinska Lowland were made after December 5. The difference between the first appearance of snow cover in the north-east of Poland and north – west edges was, therefore, almost a month.

Dates of the first snow cover occurrences in particular years could differ considerably from long-term averages even by several weeks. In general, snow cover could appear as early as the first decade of October in lowlands, but these were extremely rare incidents, e.g. appearance of snow cover in Gniezno on the October 3rd 1998, while the average for this part of Poland is December 6.

Similar differences could be observed in case of the latest dates of the first appearance of snow cover. Here, however, climatic regularities can be noticed.

Late dates of first snow cover occurrences were almost always correlated with winters with little snow. In Słubice, snow cover did not appear at all in one single season (1991/92) – the only such case in the area of Poland. In general, the latest first snow cover occurrences were dated in January, and in case of few stations with smaller snow cover conditions in climatic scale, in February.

Similar spatial regularities were observed with disappearance dates of snow cover. In regions, where it appeared at the latest dates, it disappeared at the earliest time. In the south-west of the Szczecinska Lowland and on the west coast, the average date of the last snow cover occurrence was c. March 10 or a little earlier. (Tab. 2, Fig. 4).

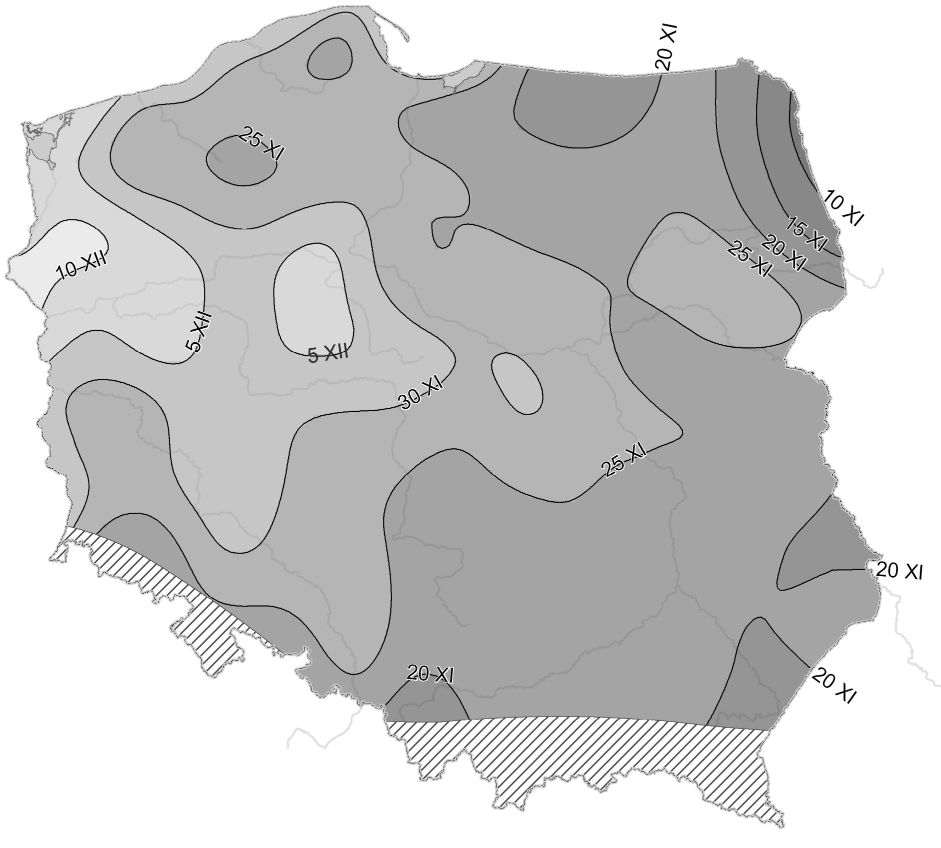

Fig. 3. The average dates of the first occurrences of snow cover in 1951-2008

Table 2. Average and extreme dates of snow cover (*sc*) occurrences in 1951-2008

Region	Exemplary stations	First occurrence			Last occurrence		
		Avg.	The earliest	The latest	Avg.	The earliest	The latest
Pomerania, Warmia, Masurian Lakeland, Suwalki Region	Świnoujście	**07.12**	24.10 (2003/04)	17.02 (1991/92)	**16.03**	14.12 (1989/90)	23.04 (1987/88)
	Resko	**28.11**	23.10 (2003/04)	23.01 (2006/07)	**23.03**	20.12 (1988/89)	11.05 (1977/78)
	Ustka	**03.12**	24.10 (2003/04)	15.02 (2007/08)	**22.03**	18.12 (1989/90)	02.05 (1984/85)
	Hel	**10.12**	02.11 (2006/07)	28.01 (1997/98)	**21.03**	18.12 (1989/90)	26.04 (1954/55)
	Chojnice	**26.11**	22.10 (1978/79)	21.01 (1972/73)	**29.03**	21.12 (1988/89)	10.05 (1952/53)
	Olsztyn	**21.11**	07.10 (2002/03)	26.12 (2000/01)	**05.04**	25.02 (1973/74)	29.04 (1975/76)
	Suwałki	**16.11**	14.10 (2002/03)	19.12 (2000/01)	**04.04**	02.03 (1990/91)	05.05 (1979/80)
Lowlands	Słubice	**07.12**	29.10 (1997/98)	No *sc* (1991/92)	**14.03**	No *sc* (1991/92)	10.05 (1952/53)
	Poznań	**03.12**	25.10 (1997/98)	21.01 (1972/73)	**18.03**	08.01 (1988/89)	02.05 (1984/85)
	Zgorzelec	**01.12**	04.11 (1980/81)	06.01 (1951/52)	**23.03**	07.01 (1988/89)	28.04 (1975/76)
	Wrocław	**01.12**	27.10 (1997/98)	21.01 (1982/83)	**19.03**	12.12 (1989/90)	29.04 (1975/76)
	Racibórz	**24.11**	15.10 (1971/72)	29.12 (2006/07)	**26.03**	09.01 (1988/89)	10.05 (1952/53)
	Łódź	**25.11**	13.10 (1973/74)	13.01 (1951/52)	**31.03**	23.02 (1990/91)	11.05 (1977/78)
	Warszawa	**27.11**	13.10 (1973/74)	30.12 (1982/83)	**28.03**	24.02 (1990/91)	03.05 (1969/70)
	Białystok	**19.11**	13.10 (1973/74)	26.12 (2000/01)	**02.04**	27.02 (1973/74)	28.04 (1983/84)
	Terespol	**22.11**	13.10 (1973/74)	24.12 (1954/55)	**22.03**	16.02 (1988/89)	21.04 (1958/59)
Uplands and Carpathian piedmont basins	Kielce	**23.11**	26.10 (1997/98)	21.12 (1953/54)	**26.03**	09.01 (1988/89)	27.04 (1981/82)
	Lublin	**23.11**	13.10 (1973/74)	22.01 (1951/52)	**03.04**	20.02 (1965/66)	11.05 (1977/78)
	Kraków	**24.11**	20.10 (2007/08)	03.01 (1955/56)	**22.03**	08.01 (1988/89)	27.04 (1981/82)
	Rzeszów	**22.11**	26.10 (1997/98)	10.01 (2000/01)	**28.03**	09.02 (1973/74)	07.05 (1956/57)
Poland (83 stations, non-mountainous area)		**27.11**	03.10 Nowy Sącz (1972). Gniezno (1998)	24.02 Szczecin. Przelewice (1988)	**26.03**	22.11 Szczecin (1988)	20.05 Aleksandrowice (1965)

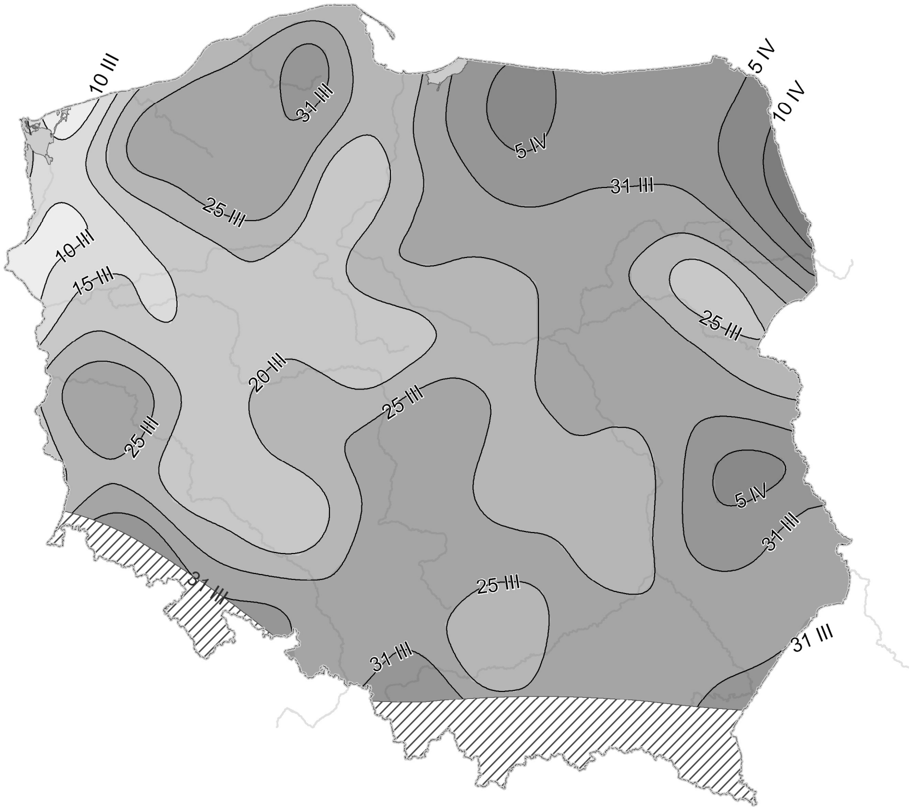

Fig. 4. The average dates of the last occurrences of snow cover in 1951-2008

On the average, between March 10 and 20, snow cover disappeared completely in the entire area of the Szczecinska Lowland, in the northern part of Lubuskie region, west and partially central Wielkopolska, Kujawy and large part of the Silesian Lowland. The average time of the last occurrences of snow cover in the remaining parts of the country was the third decade of March. The outmost eastern parts of Poland registered the last appearance of snow cover in the beginning of April. As is the case with the average time of the first snow cover appearance, there were also differences of average time of the last time of snow cover occurrence between north-west and north-east edges of Poland.

Temporal difference between the extremely early and extremely late dates of the last snow cover occurrence was even higher than the difference between extreme dates of the first snow cover occurrences. The majority of stations registered the latest snow cover occurrence at the beginning of May, and in case of only few stations in the third decade of April.

The dates of extremely early last snow cover occurrences were almost always correlated with winters with low snow cover conditions making it is a good indicator of the local climate.

The mean duration of snow cover occurrence, i.e. time between long-term mean date of first and last snow cover occurrence varied from c. 90 days or less in the Szczecinska Lowland to 130-140 days in the Masurian Lakeland and north-east part of Poland (Tab. 3, Fig. 5). On the average, the mean duration of snow cover exceeded 120 days in the areas east of the Wisła River.

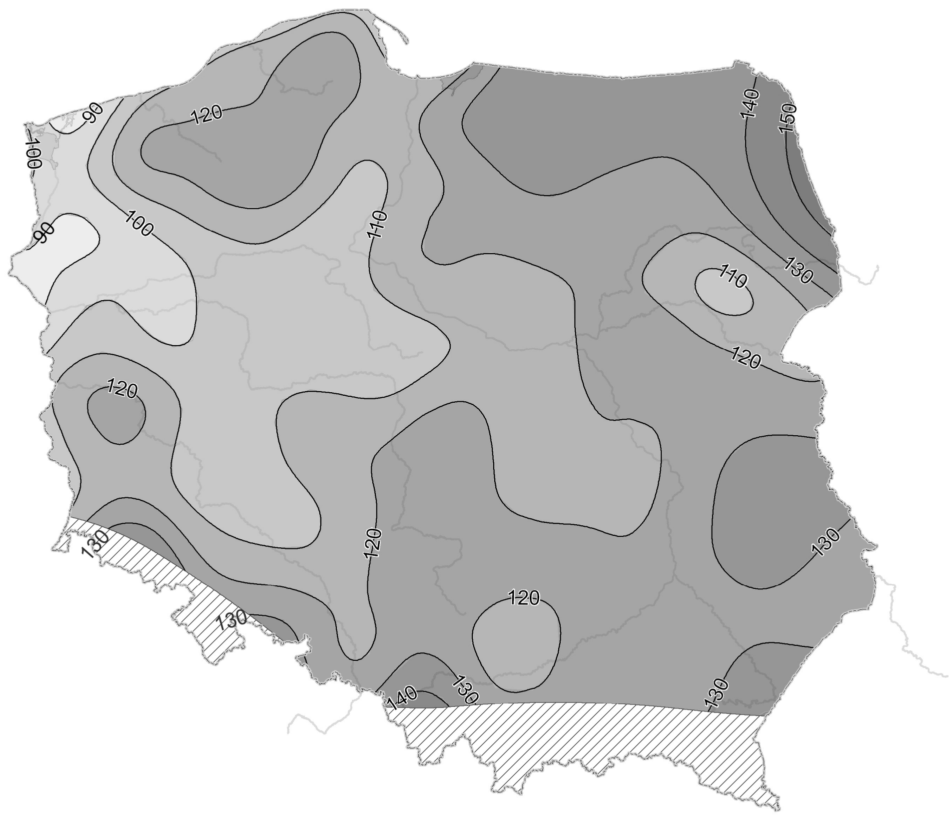

Fig. 5. Mean duration of snow cover occurrences (snow period) in days in 1951-2008

Table 3. Duration of snow cover occurrences, i.e. duration of snow season, in 1951-2008

Region	Exemplary stations	Duration of *sc* occurrences (days)		
		Avg. (A)	Potential[1] (B)	Difference (A)−(B)
Pomerania, Warmia, Masurian Lakeland, Suwalki Region	Świnoujście	99	181	82
	Resko	115	200	85
	Ustka	109	190	81
	Hel	101	175	74
	Chojnice	123	200	77
	Olsztyn	135	204	69
	Suwałki	139	203	64
Lowlands	Słubice	97	193	96
	Poznań	105	189	84
	Zgorzelec	112	175	63
	Wrocław	108	184	76
	Racibórz	122	207	85
	Łódź	126	210	84
	Warszawa	121	202	81
	Białystok	134	197	63
	Terespol	120	190	70
Uplands and Carpathian piedmont basins	Kielce	123	183	60
	Lublin	131	210	79
	Kraków	118	189	71
	Rzeszów	126	193	67
Poland (83 stations, non-mountainous area)	Mean value	**119**	**196**	**77**
	Min	84 (Dziwnów)	165 (Dziwnów)	–
	Max	145 (Aleksandr.)	222 (Aleksandr.)	–

[1] – potential duration of snow cover occurrences is the number of days between the earliest first date and the latest last date of snow cover occurrence in season in a multiyear period.

The maximum thickness of snow cover in October-May season

The deepest registered snow covers in Poland, outside the mountainous areas, usually oscillated around several dozen centimeters. Snow cover thickness reached at least 50 cm at 82% of all stations usually in February (39% of stations) or March (37%), and less often in January (22%).

However, a much better indicator of snow conditions is the average of maximum seasonal snow cover depth rather than the absolute maximum values. This parameter informs what maximum snow cover thickness can be expected during a season. Spatial distribution of this indicator confirms general characteristics of snow cover in Poland as described above. The smallest value of the maximum snow cover thickness (<15 cm) during season were registered in the west regions of Poland. Mean maximum seasonal snow cover thickness increased towards the north-east, the east and south-east. It exceeded 20 cm east of the Wisła River, Pomeranian Lakeland and highlands. In the east part of the Pomerainian Lakeland, the Masurian Lakeland, Podlasie region, Lubelska Highland, the south and east Małopolska, snow cover thickness, on the average, reached at least 25 cm. The highest values of >30, and even >35 cm, were noted in the Suwalki and Białystok regions (Tab. 4, Fig. 6).

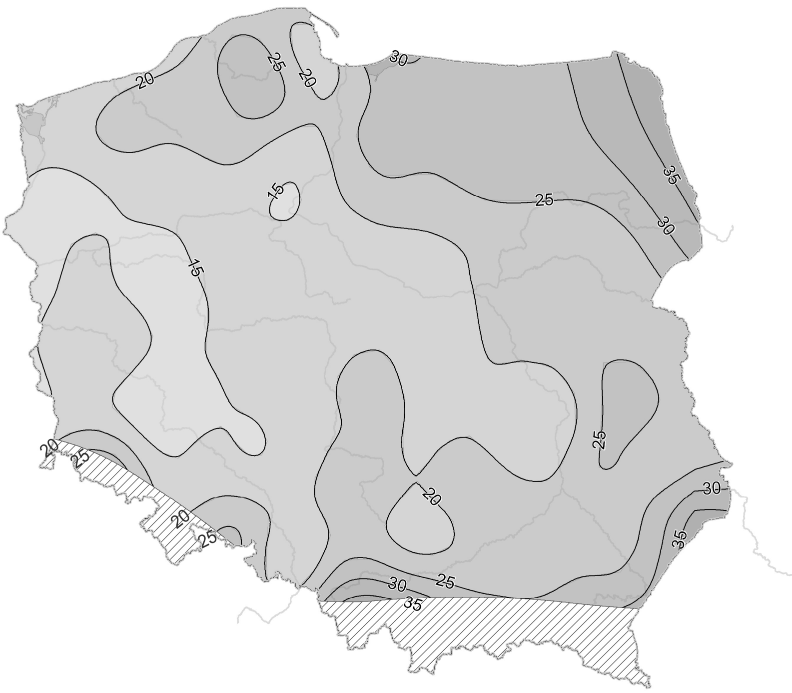

Fig. 6. Maximum averages of seasonal snow cover thickness (cm) in October-May season in 1951-2008

Table 4. Maximum snow cover thickness (cm) in 1951-2008

Region	Exemplary stations	Avg. from Max (October-May)	Absolute Max (date)	
Pomerania. Warmia, Masurian Lakeland, Suwalki Region	Świnoujście	17.4	52	(3.03.2005)
	Resko	20.4	56	(5.03.1965)
	Ustka	20.3	55	(9.01.1985)
	Hel	23.2	56	(6.03.1987)
	Chojnice	22.1	66	(2.03.1970)
	Olsztyn	26.5	67	(4.03.1970)
	Suwałki	34.6	84	(15.02.1979)
Lowlands	Słubice	12.5	42	(5.03.1965)
	Poznań	14.4	46	(5.03.1970)
	Zgorzelec	19.2	52	(17.01.1979)
	Wrocław	15.2	42	(7.02.1963)
	Racibórz	17.1	41	(19.02.1952)
	Łódź	21.1	78	(1.02.1979)
	Warszawa	21.1	70	(31.01.1979)
	Białystok	29.6	78	(19.01.1970)
	Terespol	23.9	65	(19.01.1970)
Uplands and Carpathian piedmont basins	Kielce	22.5	58	(20.02.1952)
	Lublin	24.8	50	(25.01.1979)
	Kraków	24.0	85	(6.02.1963)
	Rzeszów	24.0	53	(12.03.2005)
Poland (83 stations, non-mountainous area)	Mean value	**22.0**	**60**	–
	Min	12.5 (Słubice)	37	(29.01.1979, Legnica)
	Max	38.1 (Maków Podh.)	100	(18.03.1962, Maków Podh.)

Mean value of snow cover thickness in December-March

The mean long-term snow cover thickness for all days in December-March (4 month) period oscillated from below 2 cm in the Szczecinska Lowland and the north part of the Lubuskie region to over 10 cm in the north-east edges of Poland (Tab. 5). In the west of the country, the mean values were within the range of 2-3 cm.

Table 5. Mean thickness snow cover of all days in December−March period and total thickness for the entire season (October-May) in 1951-2008

Region	Exemplary stations	Mean thickness *sc* (Dec-Mar) (cm)			Total thickness *sc* (Oct-May) (cm)		
		avg.	max (season)	min (season)	avg.	max (season)	min (season)
Pomerania, Warmia, Masurian Lakeland, Suwalki Region	Świnoujście	**2.9**	18.5 (1978/79)	0.0 (1988/89, 1989/90)	**354.0**	2194 (1978/79)	5 (1988/89)
	Resko	**3.7**	23.0 (1978/79)	0.1 (1974/75, 1988/89)	**463.8**	2725 (1978/79)	17 (1974/75)
	Ustka	**4.0**	19.6 (1969/70)	0.1 (1988/89, 2007/08)	**497.6**	2437 (1969/70)	15 (2007/08)
	Hel	**4.6**	25.7 (1986/87)	0.1 (1972/73)	**566.3**	3116 (1986/87)	11 (1972/73)
	Chojnice	**5.2**	28.9 (1969/70)	0.4 (2007/08)	**644.1**	3557 (1969/70)	52 (1960/61)
	Olsztyn	**7.5**	36.9 (1969/70)	0.6 (1972/73)	**938.3**	4610 (1969/70)	88 (1972/73)
	Suwałki	**12.3**	46.1 (1969/70)	0.9 (2007/08)	**1538.0**	5960 (1969/70)	140 (2007/08)
Lowlands	Słubice	**1.7**	9.7 (1978/79)	0.0 (1991/92, 2007/08)	**217.5**	1147 (1978/79)	0 (1991/92)
	Poznań	**2.6**	18.9 (1969/70)	0.0 (1988/89)	**320.1**	2341 (1969/70)	2 (1988/89)
	Zgorzelec	**3.3**	16.1 (1978/79)	0.1 (2007/08)	**409.8**	1916 (1978/79)	21 (1974/75)
	Wrocław	**2.5**	13.0 (1962/63)	0.1 (1974/75, 2007/08)	**315.8**	1522 (1962/63)	11 (1974/75)
	Racibórz	**3.4**	16.7 (1962/63)	0.2 (2007/08)	**435.0**	1976 (1962/63)	48 (1974/75)
	Łódź	**4.7**	31.1 (1978/79)	0.3 (1974/75)	**590.2**	3694 (1978/79)	39 (1974/75)
	Warszawa	**4.3**	26.2 (1969/70)	0.2 (2007/08)	**530.1**	3206 (1969/70)	27 (2007/08)
	Białystok	**9.2**	48.2 (1969/70)	0.8 (1974/75)	**1137.7**	5867 (1969/70)	102 (1974/75)
	Terespol	**6.5**	34.2 (1969/70)	0.3 (1974/75)	**803.4**	4083 (1969/70)	71 (1974/75)
Uplands and Carpathian piedmont basins	Kielce	**6.1**	27.6 (1969/70)	0.6 (2007/08)	**765.1**	3337 (1969/70)	84 (1974/75)
	Lublin	**6.7**	22.2 (1978/79)	0.7 (1960/61)	**842.9**	2648 (1978/79)	89 (1960/61)
	Kraków	**5.3**	32.1 (1962/63)	0.2 (1993/94)	**667.1**	3838 (1962/63)	62 (1993/94)
	Rzeszów	**5.6**	22.8 (1962/63)	0.5 (1974/75)	**706.9**	2708 (1962/63)	57 (1974/75)
Poland (83 stations, non-mountainous area)		**5.1**	21.2 (1969/70)	0.5 (1974/75)	**635.8**	2629 (1969/70)	63 (1974/75)

In the Pomerania Lakeland, the mean snow cover thickness was higher than 4 cm, in Warmia and Masurian Lakeland – 6-9 cm, central Poland – 3-4 cm, central east of Poland – from 4.5 cm to 6.5 cm, and in the Lubelska Highland it increased to over 6 cm. During the most snowy seasons, the values of this parameters were several times higher.

Total thickness of snow cover in October-May

Daily totals thickness (depth) of snow cover give a very good synthesis of snow conditions in a season because this parameter is determined by both the number of days with snow cover and its thickness. Characteristics of temporal and spatial variability of this indicator reflects almost perfectly the mean thickness of snow cover in December-March period calculated for all days of this period. The mean long-term values total thickness of snow cover oscillated between 200 cm in the south of the Szczecinska Lowland and north of the Lubuskie region to over 1 500 cm in the north-east fringes of Poland (Tab. 5, Fig. 7). Spatial variability

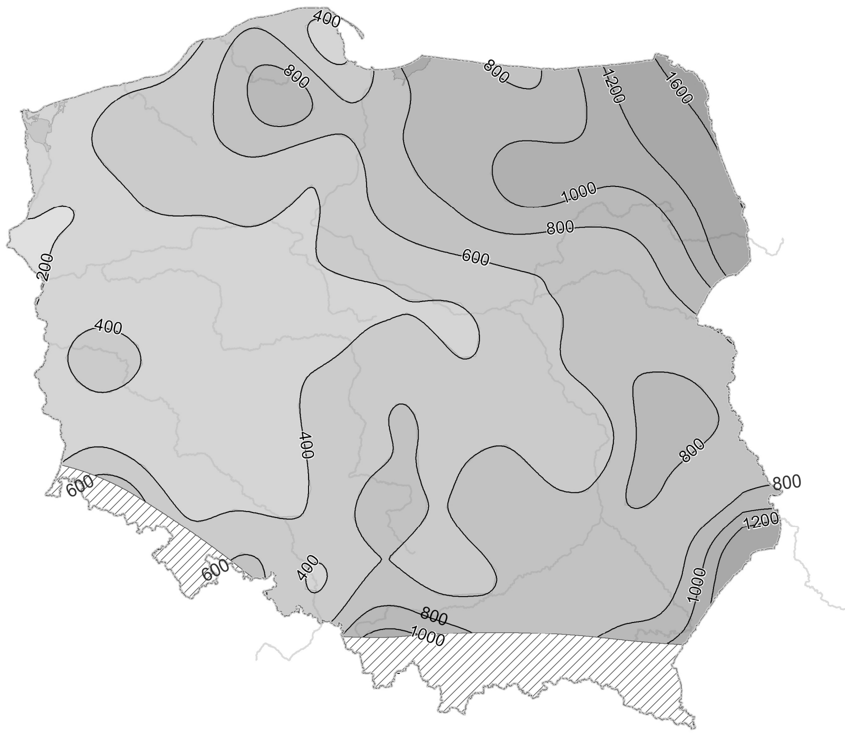

Fig. 7. Mean annual total thickness of snow cover (cm) in 1951-2008

confirms general trends presented here. Snow index intensified as the influence of oceanic climate weakens (and continental climate influence grows) and with growing altitude above the sea level. The increase thickness total of snow cover was clearly visible towards north-west, and to lesser degree, south-east and east directions. Mean seasonal total thickness of snow cover in Suwalki was over seven times higher than in Słubice.

Long-term variability of this indicator was also very high. During the snowiest winters, its values rose to 300% to even over 850% of long-term mean values (on the average, during the most snowy winter season, it was 520% of the norm in Poland). The absolute minima of this indicator during the least snow winter were less than 10%, 6% on the average. In each particular season, total thickness of snow cover deviated from the norm by 100% on the average in Poland.

During the most snowy winter seasons, the totals of snow cover could reach over 1000 cm in the area of Poland and rise over 6000 cm in the north-east (e.g. winter of 1969/70). In the least snowy winters, these totals oscillated from few centimeters in areas with the least snow in climatic scale to over several dozen centimeters in the most snowy areas (e.g. winter of 1974/75).

Long term variability of snow cover occurrences

The study of the analysis results showed that the most reliable parameters of snow cover conditions of winter were:

- the number of days with snow cover,
- total thickness of snow cover,
- Paczos snow cover index (Paczos 1982) (Tab. 6, Fig. 8).

This finding is supported by the fact that many other researchers use these indices for classification of snow cover conditions in winters (Paczos 1982, Chrzanowski 1986).

The highest variability in particular winter seasons shows total thickness of snow cover and the least – indicator of the number of days with snow cover. Paczos snow cover index, which combines two first indices, shows moderate variability. In case of the most snowy winters, the total thickness of snow cover was c. 390% of norm, for the least snowy it was 12-14% of the norm. In case of the number of days with snow cover, the maxima reached almost 200% of norm, and the minima – c. 25%.

This means that the most snowy winters characterized by indices describing snow cover occurrences and its thickness are easier identified than using only the indicator describing the number of days with snow cover.

Winters with the most intense snow cover conditions were: 1969/70, 1962/63, 1978/79, 1986/87, 2005/06, and 1995/96. Winters with the least snow were: 1974/75, 1988/89, 1989/90, 2006/07, 2007/08, and 1960/61.

Table 6. Winters with the most and the least snow cover conditions in Poland (averages from 83 stations) in 1951/52–2007/08 depending on the selection of snow cover indices

a) the most snowy winters

No.	Number of days with snow cover of ≥1cm in Oct-May		Mean[1] thickness of snow cover (cm) in Dec-Mar		Total thickness of snow cover (cm) in Oct-May		Paczos snow cover index	
1.	1969/70	121.3	1969/70	21.2	1969/70	2628.7	1969/70	5.21
2.	1995/96	108.3	1978/79	18.3	1978/79	2185.8	1978/79	4.34
3.	2005/06	101.3	1962/63	16.3	1962/63	1935.6	1962/63	4.28
4.	1962/63	95.8	1986/87	13.3	1986/87	1601.1	2005/06	3.84
5.	1978/79	90.9	2005/06	10.8	2005/06	1315.2	1986/87	3.78

b) the least snowy winters

No.	Number of days with snow cover of ≥1cm in Oct-May		Mean[1] thickness of snow cover (cm) in Dec-Mar		Total thickness of snow cover (cm) in Oct-May		Paczos snow cover index	
1.	1974/75	20.3	1974/75	0.5	1974/75	63.0	1988/89	0.52
2.	2006/07	22.4	2007/08	0.7	1960/61	107.5	1989/90	0.61
3.	1988/89	22.6	1988/89	0.7	1988/89	110.9	1974/75	0.61
4.	1989/90	26.3	1989/90	0.8	2007/08	122.2	2006/07	0.71
5.	1960/61	27.3	1960/61	0.9	1972/73	150.7	2007/08	0.73

[1] mean value of all days in December-March period.

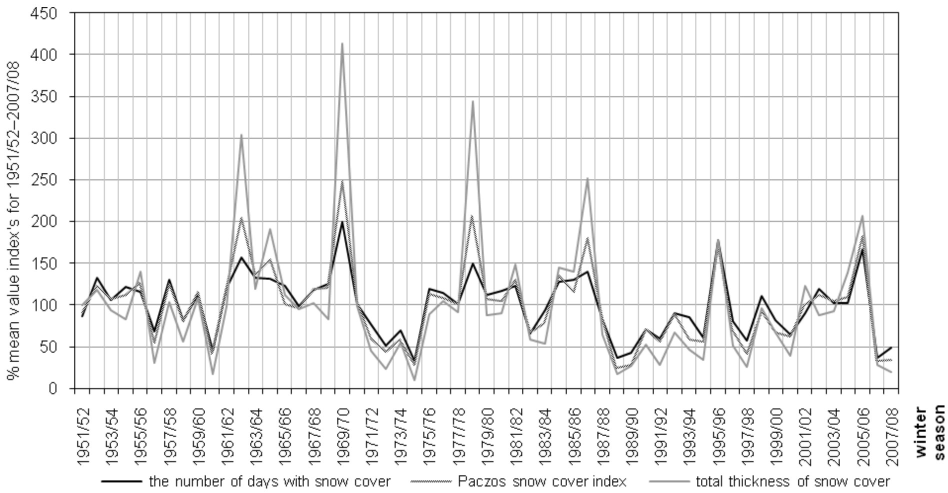

Fig. 8. Snow cover conditions in winters according to selected indices of snow cover occurrences in Poland (mean values from 83 stations) in 1951-2008

CONCLUSIONS

1. Among the analyzed snow cover indices, the most useful ones are those indicating the number of days with snow cover occurrences and the total thickness of snow cover during a season (October-May).
2. Snow cover tends to last longer and be thicker as the influence of the oceanic climate lessens, i.e. the snowiness increases from the west towards the east. The altitude above the sea level has a modifying effect on snowiness.
3. Spatial distribution of snow cover occurrences shows high variability in Poland (e.g. the number of days with snow cover in the north east is three times higher than in the Szczecin Lowland on the average).
4. Potentially, winters in the west of Poland may be snowless, however the probability of such winter is very low (in the period of 1951/52-2007/08). A snowless winter was registered once on one station only (Słubice 1991/92). The maximum number of days with snow cover in the Polish lowlands may be ca. 145 days or more (almost 4 months).
5. The extremely early dates of the first snow cover occurrence in a season and extremely late dates of its appearance are not good indicators of snowiness characteristics of a climate. On the average, snow cover in Poland may be ex-

pected to linger for ca. 54% of days in a year, from the beginning of November till the end of April.

6. The maximum absolute thickness of snow cover (which is not a good indication of snowiness characteristics of a climate) reaches over 50 cm in the large part of Poland, however it never exceeds 1 meter (60 cm on the average). In fact, the average maximum snow cover thickness is much smaller, from 15 cm in the west to 25-30 cm in the east of Poland.

7. The total thickness of snow cover in a season, despite being a rather abstract value, is a very good indicator of climate snowiness because it integrates the fact of snow cover occurrence and snow cover thickness. During the most snowy seasons, the values of this parameter are 5 times higher than the long term average values which is a good reflection of the variability of climatic snowiness of winters in Poland.

8. In the analyzed multi-year period (of over 50 years), there were several exceptionally snowy winters (occurring statistically every 10 years). Generally, the snow cover thickness parameters show relatively high variability while the variability of snow cover occurrences indicators is lower. The winters with the most snow are the result of uninterrupted lingering of snow and consequently its accumulation and better resistance to melting (due to freezing).

9. The most snowy winters occurred in: 1969/70, 1962/63, 1978/79, 1986/87, 2005/06 and 1995/96. The winters with the least snow were: 1974/75, 1988/89, 1989/90, 2006/07, 2007/08 and 1960/61.

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7. SNOWFALL IN KRAKOW AND ITS LINK TO ATMOSPHERIC CIRCULATION DURING THE PERIOD 1951-2008*

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INTRODUCTION

Snowfall and snow cover constitute important components of the climate system and are sensitive to change in this system, especially thermal change. For this reason they are regarded as leading indicators of climate change (Huntington et al. 2004, Ke *et al*. 2009**)**. Studies of long-term snowfall mostly tend to focus on mountain-located weather stations or high latitudes (e.g. Przybylak 2002, Førland and Hanssen-Bauer 2003**,** Łupikasza 2008) where most precipitation is solid. In moderate latitudes there is a limited access to details of precipitation type. Snowfall variability in time is not very well researched, especially in Central Europe, even though snowfall accounts for much of the annual precipitation in this region. The only Polish studies on this matter relate to the city of Krakow (Twardosz 2002-2003, 2003).

Some of the latest research into the variability and totals of snowfall in Canada and the US (Groisman and Easterling 1994**,** Kunkel *et al*. 2007) and in China (Ke *et al*. 2009) found trends of various scales and directions which were not always statistically significant. In the Canadian Arctic significant trends in snowfall totals were found during the period of 1950-1995, but without a significant change in their share of the total precipitation (Przybylak 2002). In the Norwegian Arctic the share of snowfall in total precipitation was on the decline during the last decades of the 20th century (Førland and Hanssen-Bauer 2003).

Numerous studies found atmospheric circulation to have a significant impact on precipitation, especially at moderate latitudes in the northern hemisphere. The NAO influence on precipitation in Europe is well documented (e.g. Wibig 2001), just as is the influence of regional indicators of atmospheric circulation on precipitation in Central Europe (e.g. Niedźwiedź *et al*. 2009). There are also studies

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that found a significant impact of NAO on snowfall in eastern USA (Hurrell 1996, Hurrell and Dickson 2004, Durkee *et al*. 2007) and on the number of days with snow cover in Eastern Europe (Bednorz 2004) and in Poland (Falarz 2007).

The objective of this study is to investigate long-term variability of snowfall in Krakow using the number of days and total snowfall and to identify a relationship between snowfall and atmospheric circulation at the macro and mezo synoptic scales.

DATA AND METHODS

This study used a 58-year record of precipitation measurements and details of their type from Krakow (1951-2008). Days with snowfall were defined at ≥0.1 mm. The paper consists of three main sections. The first section discusses the long-term variability of the date of the first and last day of snowfall and the duration of the snowfall season. The snowfall season (or the duration of the snowfall occurrence) is defined as the number of days between the first and last snowfall of the year. In Krakow the potential snowfall season, defined as the number of days between the date of the first and last snowfall measured over the entire study period, lasts from October to May. The second section discusses trends in the number of snowfall days and snowfall totals over a snowfall season and in each individual month from November to March. These months were selected because their snowfall frequency provided sufficient data for linear regression analysis.

A simple least squares linear regression was used to calculate the magnitude and sign of trends, while their significance was tested with the non-parametric Mann-Kendall test. The statistical significance of precipitation trends is usually lower as compared to other climate elements due to its large spatial and temporal variability. Therefore, a lower significance level than most other studies was used (Rapp 2000, Hänsel 2009). All the snowfall parameters had their trends calculated over the whole study period 1951-2008 (referred to as long-term trends) and in moving 30-year periods, i.e. 1951-80, 1952-81 and so on until 1979-2008 (referred to as short-term trends). In this way it was possible to assess the stability of the trends and to identify short-term, but significant, changes. Monthly short term trend magnitudes are expressed in percentage of average snowfall characteristics from the whole studied period (Fig. 2B and 3B), thus obtaining relative trends. The relative trend magnitudes are comparable regardless of monthly differentiation in average values of snowfall characteristics and various units of these characteristics (number of days, mm). Both trend and relative trend magnitudes are given for monthly characteristics of snowfall within the text.

In the final section, the study identifies the impact of atmospheric circulation on the number of snowfall days and snowfall totals. The North Atlantic Oscillation index (Hurrell 1995) was used as the measure of macrosynoptical circulation while the mezosynoptical circulation was described by regional types and indicators of atmospheric circulation determined using a classification for southern Poland by Niedźwiedź (1981, 2000, 2009).

The NAO index was defined by Hurrell (1995) as the normalized pressure difference between a station on the Azores and one on Iceland. We applied an extended version of the index based on the normalised pressure difference between Gibraltar and Reykjavik (Jones *et al*. 1997), updated on the Climate Research Unit website (http://www.cru.uea.ac.uk/cru/data/nao.htm). For the purpose of this analysis, three simple indices were derived through modification of the method proposed by Murray and Lewis (1966) using the number of days with specific circulation patterns (Niedźwiedź 2000), i.e. western (W) circulation, southern (S) circulation and cyclonicity (C). The Western circulation index (W), expressing the intensity of westerly circulation within the zone, was measured by the summation of scores (points) for days with different directions of air flow as follows: +2 for W, +1 for NW and SW types, –2 for E, and –1 for types NE and SE. Positive values of this index occur when there is a distinct predominance of air advection from the west, while negative values point to a strong easterly air flow. A recent modification of this index involved dividing the value of the scores by double the number of days in the month and the result was expressed as a percentage. The percentage values of the W index range from –100% (eastern advection on all days) to +100% (western advection on all days). The index of Southerly circulation (S) was calculated by the summation of scores allocated as follows: +2 for type S, +1 for types SW and SE, –2 for type N, and –1 for types NW and NE. Hence, high positive values of index S point to an intensive advection of air masses from the south, while negative ones point rather to advection from the north. The percentage values of the S index range from -100% (northern advection on all days) to +100% (southern advection on all days).

The third index, the Cyclonicity index (C), was found to have the highest impact on precipitation patterns in Central Europe (Niedźwiedź and Twardosz 2004). This index was calculated by the summation of scores (points) for days with particular forms of circulation as follows: +2 for centre of cyclone Cc or trough Bc, +1 for other cyclonic situations Nc (northern cyclonic), NEc, Ec, SEc, Sc, SWc, Wc, and NWc, –1 for anticyclonic advection types Na (northern anticyclonic), NEa, Ea, SEa, Sa, SWa, Wa, and NWa, –2 for centre of anticyclone Ca or wedge Ka. Thus positive values of the C index mean a measure of domination of cyclonic types of circulation over anticyclonic ones during the particular period (month, season, year). The percentage values of the C index range from –100% (a cyclone centre Cc or cyclone trough Bc on all days) to +100% (an anticyclone centre Ca or an anticyclone wedge Ka on all days). Index C for Poland is not well correlated with the NAO (Niedźwiedź 2000), and can therefore be useful as an additional tool in explaining the variability of precipitation.

The relation between snow precipitation and circulation indices was investigated by calculating Pearson's and Spearman's correlation coefficients, separately for the snowfall season (October-May), winter season (December-February) and each month.

SNOWFALL CHARACTERISTICS

The data of the first and last snowfall

In Krakow the first snowfall occurs on average on 20 November. The earliest first snowfall was recorded on 15 October 1997 and the latest on 22 December 1953, which means that there is a range of 69 day. During the study period, the dates of first snowfall followed statistically significant trends (significance level p = 0.097), as a result of which the first snowfall occurs now approximately 10 days earlier than it did in the mid-20th century (the trend is 1.7 days per 10 years). Over the study period, the first snowfall dates varied very little (only 14 days) between the late 1960s and late 1970s and then, beginning in the early 1980s, this variability rose significantly (Fig. 1A). Short term trends were not statistically significant (Fig.1B).

The last snow of the season falls on average on 29 March. The earliest last snowfall was recorded on 7 February 2006 and the latest on 12 May 1977 giving a range of 95 days. In the long-term, the last day with snowfall followed a different variation pattern than the first day. There were no significant long-term changes, but strong short-term trends (Fig.1A). In the 30-year periods from the beginning of the study period until virtually the end of the 1980s, these trends were statistically significant and pushed the end of the snowfall season towards spring at a rate of up to 13 days per 10 years. From the end of this period, the trend directions continued until the end of 1990 s, but was no longer significant. The trend magnitudes dropped rapidly from the 30-year periods starting in 1959

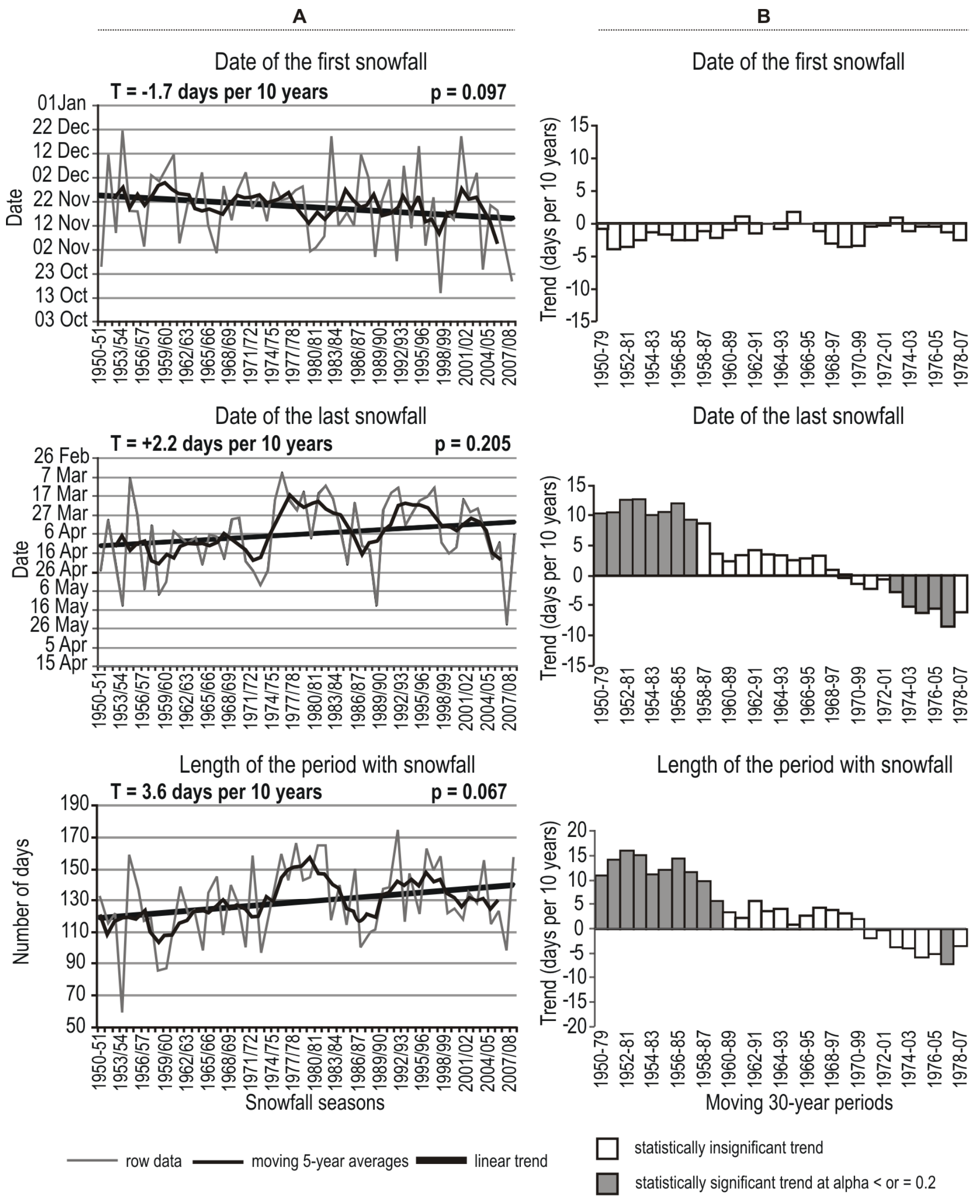

Fig. 1. Temporal variability and trends in characteristics of annual course of snowfall
a – temporal variability and long-term trend, b – trend magnitudes for each of the moving 30-year periods within the period 1951-2008, p – significance level, T – index change per 10 years

to just approximately three days per 10 years. The trend then reversed during the 30-year periods of 1968-99 and the dates were increasingly earlier. These trends had an average rate of six days per 10 years and were statistically significant from the mid 1970s until the end of the study period. The mid-1970s included a landmark season in 1976/77 when the last day with snowfall began to move very consistently towards winter (Fig. 1A). This tendency seems compatible with the observed growth in air temperature, especially visible in February and April (Piotrowicz 2007).

Duration of the snowfall season

The average snowfall season in Krakow lasts 136 days. During the period studied, the longest season lasted 176 days (28 October 1991 – 20 April 1992) and the shortest had only 58 days (22 December 1953 – 17 February 1954). There is a long-term increasing trend which is statistically significant ($p = 0.067$) over the entire study period, but the actual period during which the snowfall season was extending only lasted from the beginning of the period until the early 1980 s, when it had grown from 121 to 166 days (i.e. by 45 days). At the beginning of the 1980s, the trend reversed and the snowfall seasons began to shrink, albeit at a much slower rate than they had previously expanded (Fig.1A). The character of the changing duration of snowfall season is well reflected by trends identified over the moving 30-year periods, which eliminate the shorter than 30 year fluctuations and generalise the long-term picture. In all 30-year periods of the 20th century, the snowfall season expanded (Fig.1B). Initially, these trends had a magnitude of up to 15 days per 10 years, but it gradually diminished to two days per 10 years in the period 1970-99. These changes were statistically significant until the end of 1980s. During the 1970s, the trend reversed and the snowfall season began to shrink. While the average magnitude of these decreasing trends was four days per 10 years, in a clear majority of cases the changes were statistically insignificant (Fig. 1B).

Number of days with snowfall

In Krakow snow mainly falls in the winter months (December-February). Peak snowfall is recorded in January (nine snowfall days on average), but is also quite frequent in November (3.4 days) and March (4.8 days). Snow fell very rarely in April (1.2 days) and October (0.3 days) and there were only two cases of snowfall in May.

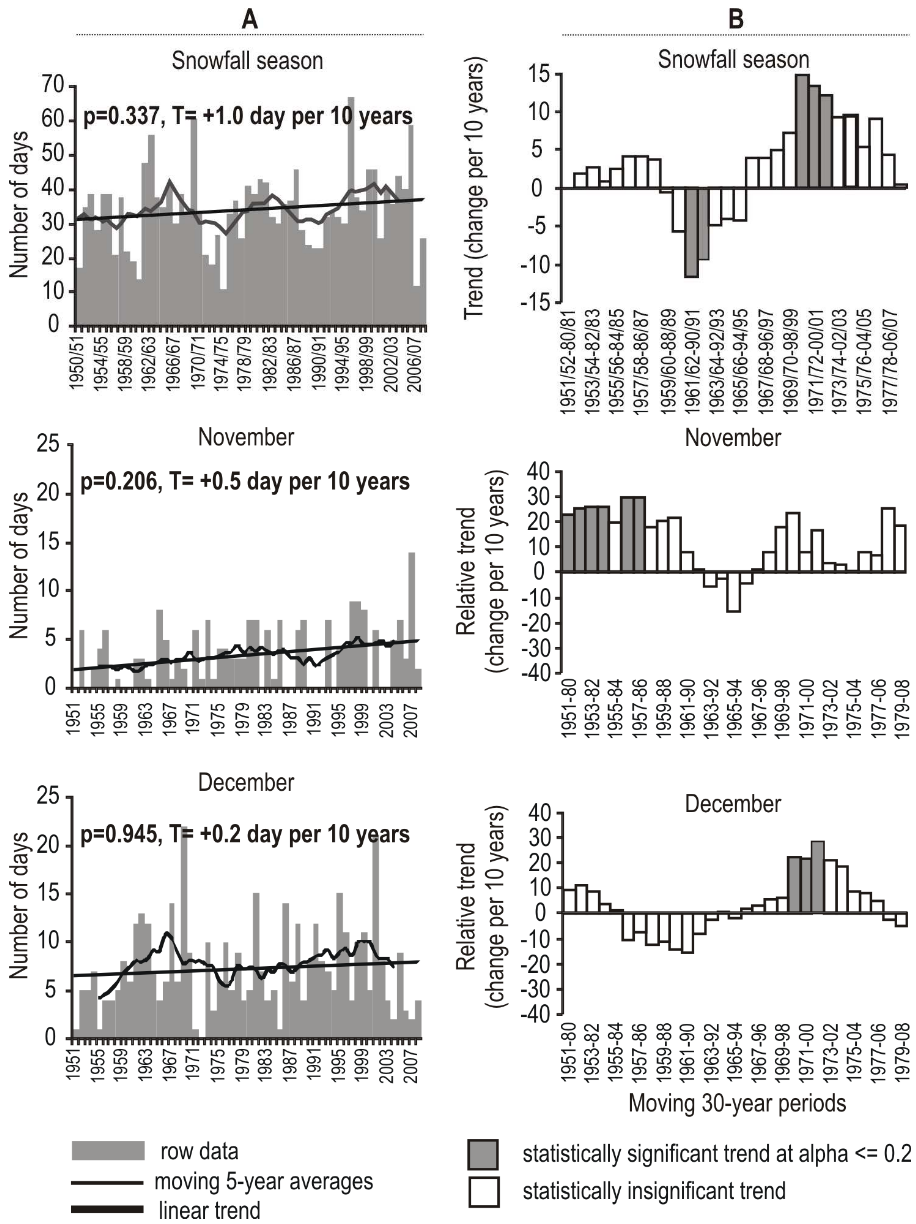

Fig. 2. Temporal variability and trends in the number of days with snowfall
a – variability and long-term trends, b – short term trends (monthly short term trend magnitudes are expressed in % of average index value from the period 1951-2008), p – statistical significance of linear trend, T-index change per 10 years

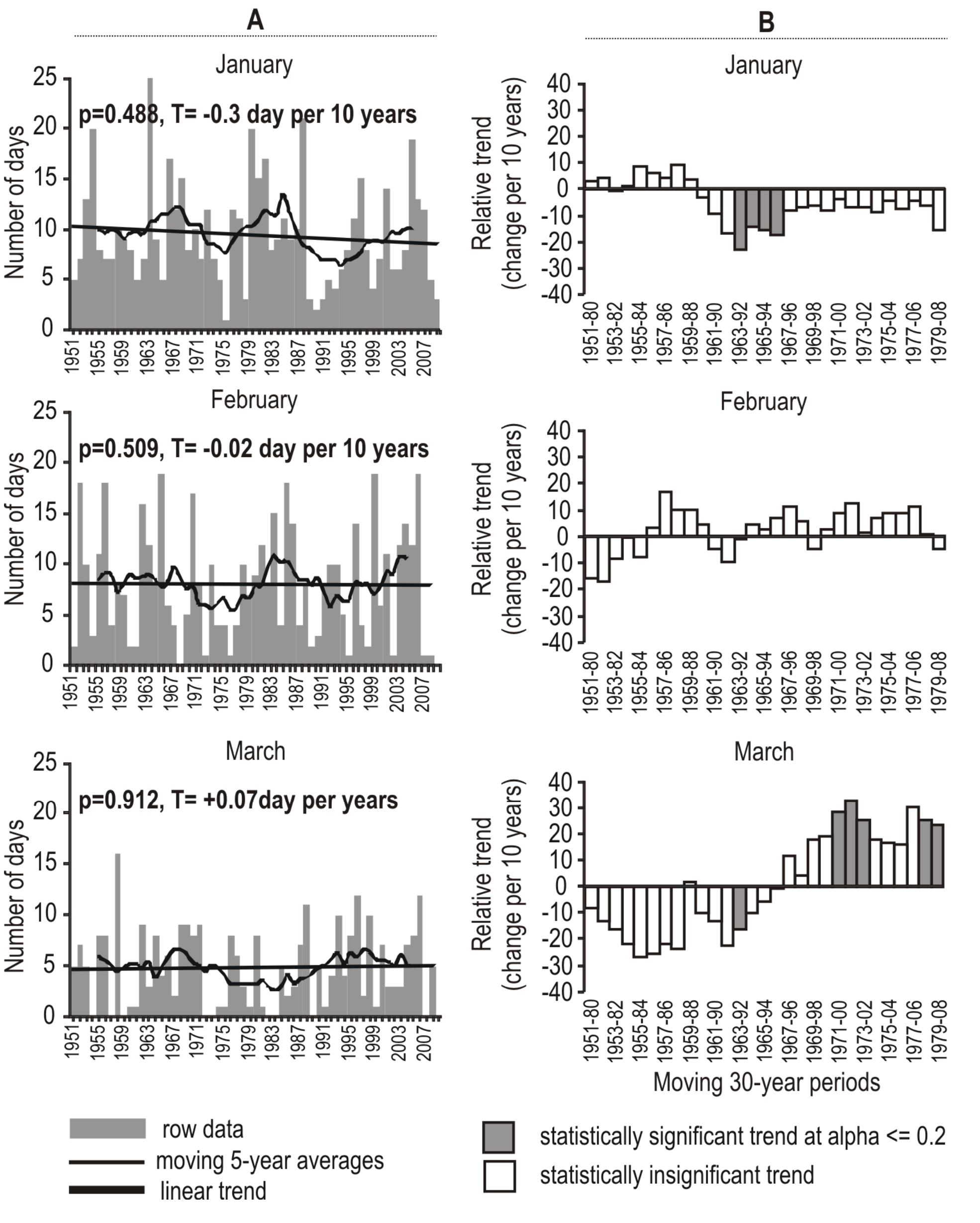

Fig. 2. Cont. Temporal variability and trends in the number of days with snowfall
a – variability and long-term trends, b – short term trends (monthly short term trend magnitudes are expressed in % of average index value from the period 1951-2008), p – statistical significance of linear trend, T-index change per 10 years

The average number of days with snowfall during the season (October-May) was 34. The largest number of 67 days with snowfall was observed in the season 1995/96, closely followed by 61 days in 1969/70, 59 days in 2005/06 and 56 days in 1962/63. The lowest numbers of 11 days with snowfall was recorded in 1974/75, followed by 12 days in 2006/07, 14 days in 1960/61 17 days in 1950/51. The number of days with snowfall grew by approximately one day per 10 years during the study period, but this was not statistically significant like most of the short-term trends (Fig. 2A, B). Despite a lack of significant trends in the number of days with snowfall in the majority of moving 30-year periods, it is worth noting that these trend directions changed often and that their magnitudes gradually increased (Fig. 2B). Positive trends were noted during the periods 1950/51-87/88 (0.8 days per 10 years on average) and 1966/67-2007/08 (2.6 days per 10 years), while negative trends occurred during the periods from 1959/60 to 1994/95.

Long-term monthly trends in the number of days with snowfall varied: they were increasing in November, December and March and decreasing in January and February. None of these trends were statistically significant (Fig. 2A). Only some of the 30-year trends were statistically significant. In January, most of these trends showed a decrease and the statistically significant ones had an average magnitude of –1.6 days (17.4% of average) per 10 years and were recorded during the period 1962-1994. In February, the trends varied and were insignificant. In March, they showed a decrease during the periods from 1951 to 1995 and an increase during the periods from 1967-1996 until the end of the study period. The latter trends varied from 1.2 days (25.5% of average) per 10 years during the periods 1973-2002 and 1978-2007 to 1.6 days (32.9% of average) per 10 years during the period 1972-2001 (Fig. 2B.). In November, trends to an increase dominated and were significant (0.8 days (25.6% of average) per 10 years) during the initial 30-year periods (1951-86 except 1955-94) (Fig. 2B). In December the short-term trends followed the overall patterns of the snowfall season. Significant changes in that month involved an increase in the number of snowfall days by an average of 1.7 days (24.1% of average) per 10 years in the periods 1970 to 2001 (Fig. 2B).

Snowfall totals

The highest monthly snowfall total is recorded during the winter months, in the same manner as the number of snowfall days, including on average 18.5 mm in January, 15.0 mm in December and 16.7 mm in February. The highest monthly total of 65.6 mm was recorded in February 1952. The average seasonal snowfall is 72.0 mm, while the extreme values varied from a minimum of 25.3 mm in 1974/75 to a maximum of 148.1 mm in 1969/70 producing a range of 122.8 mm. Other seasons with high snowfall included: 1999/01 (137.5 mm), 1995/96 (132.5 mm) and 2005/06 (131.9 mm). Seasons with low snowfall included 1960/61 (27.5 mm), 1990/91 (31.9 mm) and 2006/07 (34.2 mm). The positive trend discovered during the study period was not statistically significant (Fig. 3A). There were several short-term trends, but they were also mostly not significant because of a considerable degree of change between subsequent snowfall seasons (Fig. 3B). There was, however, a visible regularity in the short-term trends, which was to a certain degree linked to the temporal variability of the trends in the number of days with snowfall. Snowfall totals were generally found to be on the increase during both the early and late 30-year periods (1951-1987 and 1970-2008). The magnitude of the trends was clearly higher in the later periods. In the middle of the study period, i.e. in the 30-year periods between 1959 and 1998, the snowfall totals were decreasing. Some of these changes were statistically significant and their magnitude was up to approx. –10.0 mm (14.3% of average) per 10 years (Fig. 3B).

Snowfall totals decreased over the study period during all of the months studied apart from November, but only in December was the trend weakly significant ($p = 0.150$). There are two to three characteristic periods in the long-term monthly totals. In the winter months (December-February), the short-term trends were mostly towards a decrease. In January, this trend was consistent in 30-year periods from 1961-2004, including a statistically significant change of –3.8 mm (20.1% of average) per 10 years in the 30-year periods in 1962-1995. In February and December, most of the short-term trends were not statistically significant and showed a decrease (Fig. 3B). With regards to March, there were some periods, beginning in 1971, in which there was a statistically significant trend to growth. November had the largest number of statistically significant short-term trends and these were observed until the end of the 20th c. (Fig. 3B). The statistically significant increasing trends observed in the periods 1951-1982 and in 1956-85 averaged at 2.9 mm (34.9% of average) per 10 years, while the figure for the periods 1964-1996 was –2.1 mm (24.5% of average) per 10 years.

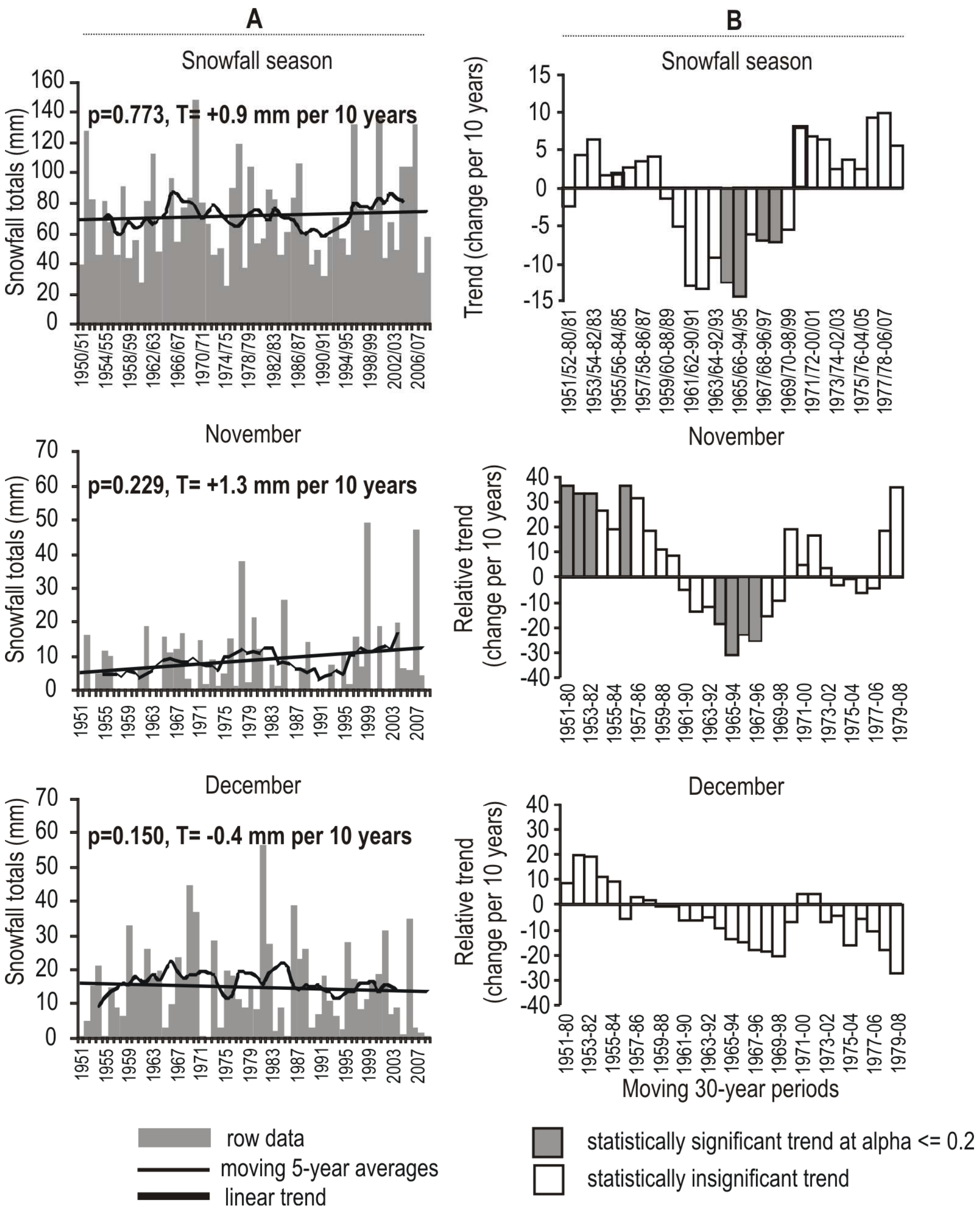

Fig. 3. Temporal variability and trends in the snowfall totals
a – variability and long-term trends, b – short term trends (monthly short term trend magnitudes are expressed in % of average index value from the period 1951-2008), p – statistical significance of linear trend, T– index change per 10 years

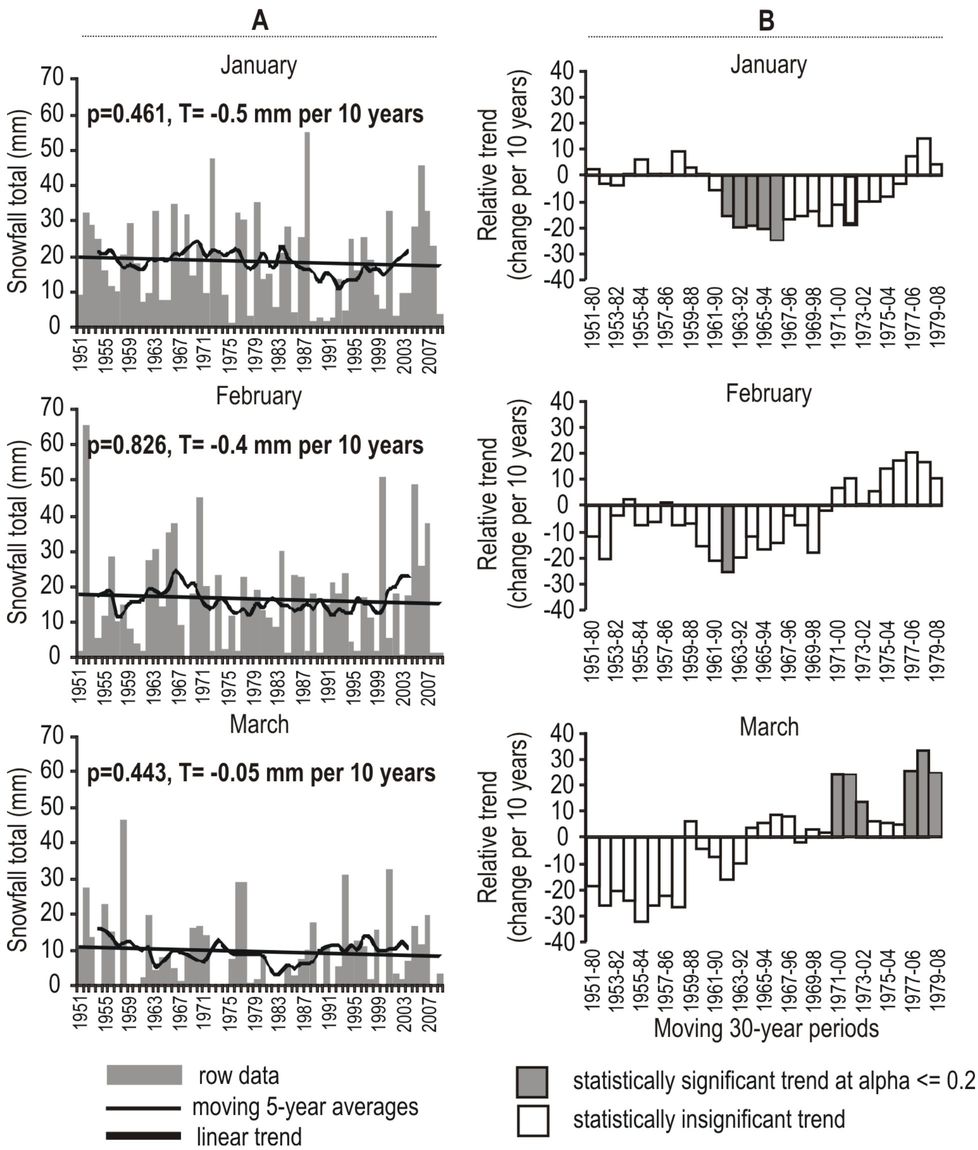

Fig. 3. Cont. Temporal variability and trends in the snowfall totals
a – variability and long-term trends, b – short term trends (monthly short term trend magnitudes are expressed in % of average index value from the period 1951-2008), p – statistical significance of linear trend, T – index change per 10 years

SNOWFALL AND ATMOSPHERIC CIRCULATION

Snowfall studies tend to underline the importance of the North Atlantic Oscillation (NAO) in determining snowfall frequency (e.g. Durkee *et al*. 2007). Other studies also identify synoptic situations that favour snowfall, especially intensive snowfall (Spreitzhofer 1999a, b, Bednorz 2008).

The NAO index represents westerly air flow and significantly determines snowfall occurrence in Krakow. Correlation coefficients between NAO and snowfall characteristics are statistically significant ($\alpha = 0.05$) in all of the months and snowfall seasons studied (Tab. 1). Larger correlation coefficients were obtained in the case of days with snowfall. Between November and March, the NAO index explained ca. 27% of the variability of days with snowfall and ca. 20% of snowfall totals.

The phase of the NAO index had a significant impact on the snowfall characteristics in question. Figure 4 shows monthly records of snowfall days and totals (expressed as a percentage of their long-term average) during the negative, neutral and positive phases of the NAO. Both variables were significantly higher than the long-term average in all of the months during the negative NAO phase. The influence of this NAO phase on snowfall in Krakow is demonstrably greater in February and March. In March, snowfall was ca. 64% more frequent during that phase than on average and even in December, when the NAO influence is the weakest, there is ca. 30% more snowfall than the long-term average. During the positive NAO phase, snowfall is less frequent than average, including by ca. 17.8% in December and by 34.3% in November. Also during the neutral phase, the snowfall frequency is lower than average, but the difference is much smaller than in the case of the positive phase. In January, the snowfall frequency during the neutral phase is identical to the overall average (Fig. 4). The relationship between snowfall totals and NAO phases follows a very similar pattern to that of snowfall frequency. The amount of snowfall during the negative phase is higher than the monthly averages by between 39% (December and January) and 69% (March). During the positive and neutral phases, the amount is less than average. It is noteworthy that the difference between the average number of snowfall days during the negative and neutral phases is the greatest in January (by 26.3% of the long-term average), while with the snowfall total it is greatest in February and March (by 28.0% and 23.2% of the long-term average, respectively).

Table 1. Correlation coefficient between snowfall characteristics (number of snowfall days and snowfall totals) and the circulation indices
NAO – North Atlantic Oscillation, W – western circulation index, S – southern circulation index, C – cyclonicity index, P – Pearson correlation coefficient, S – Spearman correlation coefficient, bolded values mean statistically significant correlation at α<0.05

Circu-lation index	Correlation coefficient	Number of snowfall days						
		Nov	Dec	Jan	Feb	Mar	Nov-May	Dec-Feb
NAO	Sp	**–0.375**	**–0.354**	**–0.453**	**–0.428**	**–0.347**	**–0.376**	**–0.443**
	P	**–0.363**	**–0.349**	**–0.512**	**–0.448**	**–0.382**	**–0.431**	**–0.489**
W	Sp	–0.124	**–0.305**	**–0.495**	–0.224	–0.111	–0.179	**–0.438**
	P	–0.090	**–0.430**	**–0.540**	**–0.280**	–0.170	**–0.338**	**–0.558**
S	Sp	**–0.308**	–0.051	**–0.264**	**–0.554**	**–0.347**	–0.093	–0.256
	P	**–0.330**	–0.120	–0.230	**–0.580**	**–0.340**	–0.052	–0.171
C	Sp	–0.021	0.042	0.166	0.095	0.119	–0.065	–0.038
	P	0.020	0.040	0.130	0.130	0.140	–0.090	0.034
		Snowfall totals						
NAO	Sp	**–0.414**	**–0.302**	**–0.339**	**–0.452**	–0.252	**–0.332**	**–0.467**
	P	–0.237	–0.260	**–0.337**	**–0.373**	–0.237	**–0.342**	**–0.432**
W	Sp	–0.160	–0.173	**–0.283**	–0.115	–0.039	–0.151	**–0.333**
	P	–0.060	–0.205	**–0.340**	–0.100	–0.070	–0.206	**–0.384**
S	Sp	**–0.314**	–0.114	–0.250	**–0.494**	**–0.349**	0.030	–0.099
	P	**–0.330**	–0.090	–0.160	**–0.510**	**–0.330**	0.025	–0.098
C	Sp	0.003	0.153	**0.301**	0.257	0.135	0.094	**0.300**
	P	–0.080	0.220	0.250	**0.350**	0.100	0.080	**0.329**

Looking at regional atmospheric circulation indices and considering the seasonal values (Nov-May, Dec-Feb), the strongest correlation is achieved with the western circulation index W. For example in winter (Dec-Feb), 31% of the variance of the number of snowfall days can be explained by the variability of this index. Monthly number of days with snowfall significantly depends on the W index in January and December. The relationships between snowfall totals and the W index are significant only in January. An inflow of air from west lowers both the frequency and totals of snowfall. The relationships between the snowfall characteristics and the S index are also negative. Snowfall dependency on the S index is statistically significant only in case of monthly characteristics. The S index significantly reduces the snowfall both frequency and totals in November, February and March. This index explains nearly 34% of the variance in the number of days with snowfall in February. The most frequent positive correlation of snowfall, if rarely significant, is with the Cyclonicity index (C). The C index does not significantly influence the frequency of snowfall in any of the periods considered. In principle, it significantly increases the snowfall totals only in winter (Dec-Feb) (Tab. 1). Increased cyclone activity over Central Europe is generally conducive to precipitation, regardless of its type (Bednorz 2008, Niedźwiedź *et al*. 2009, Spreitzhofer 1999a, 2000).

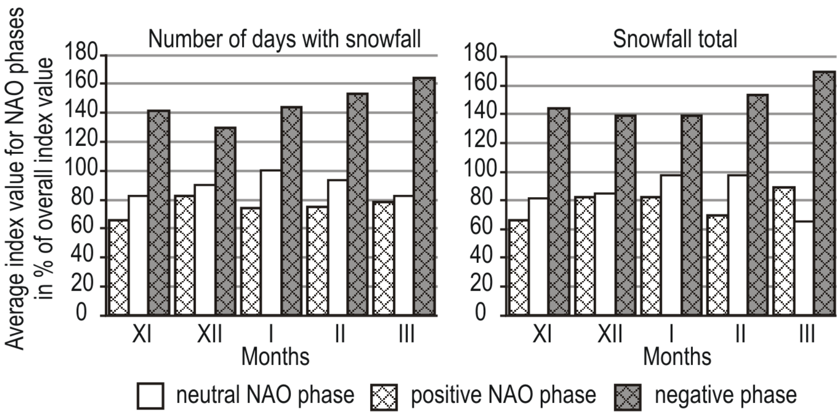

Fig. 4. Average values of snowfall characteristic (number of days, totals) for NAO phases
These average values are expressed in percentage of overall seasonal number of days with snowfall and snowfall totals respectively, from the period 1951-2008

CONCLUSIONS

1. Using a long and homogeneous record of daily precipitation in Krakow the study has shown that snowfall changes and the degree, sign and significance of that change vary between the parameters investigated. Atmospheric circulation, both at a macro and mezosynoptical scales, was found to have a significant impact on snowfall change.

2. The dates of the first snowfall and snowfall season duration changed considerably. Essentially the change involved an increasingly early start of the snowfall season and its effective extension by ca. 3.6 days per 10 years. The date of the last snowfall did not follow any significant long-term trend, but there were some short term trends. The latter trends were significant from the beginning of the study period, when the last snowfall occurred in late February or in early March, to the end of the 1970s, when the last snowfall occurred in the first half of May.

3. Over the study period, there is an insignificant increasing trend in the number of snowfall days and in snowfall totals. The lack of a statistically significant trend in snowfall in Krakow is generally compatible with the results of similar research in various parts of the world where a great variability of trends was shown with regard to this element of climate (e.g.: Karl *et al*. 1993, Ke *et al*. 2009, Kunkel *et al*. 2007). It may be surprising, however, that there is a positive trend of these parameters which would normally be associated with higher latitudes including Canada northwards of 55°N (Groisman and Easterling 1994) and those areas of Eurasia where the average temperature stays at or above –8°C (Ye 2008).

4. Despite the lack of a significant long-term trend there were short term trends during 30-year periods which were significant in certain months. In November, both the number of snowfall days and the snowfall total tended to increase significantly during the short-term periods of 1951-1986. These trends were reversed during the consecutive 30-year periods of 1961-1998 when the snowfall frequency was falling significantly and during the 30-year periods of 1964-1996 when the snowfall total also diminished significantly. A falling trend was also recorded in January during the consecutive 30-year periods of 1963-1996. March, on the other hand, displayed significant increasing trends in both totals and frequency of snowfall during the final 30-year periods of the study period (starting in early 1970s).

5. The number of days and snowfall totals in Krakow depend significantly on the NAO phase. During the negative NAO phase, the two parameters are between 30% and nearly 70% higher than their monthly averages. This influence is at its lowest in December, but grows rapidly as the snowfall season progresses to peak in March.

6. Among the regional circulation indices the W index has the greatest influence on both the frequency and amount of snowfall in Krakow in wintertime (December-February) (negative correlation $r = -0.558$ with the number of snowfall days). The C index has a significant impact on the snowfall total (positive correla-

tion r = +0.329). The S index has a significant impact on monthly characteristics of snowfall (in November, February and March).

7. The study of the role of atmospheric circulation in shaping snowfall requires further research using longer observation records.

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8. SEASONAL WATER USE EFFICIENCY RUN AT RZECIN WETLAND*

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INTRODUCTION

Global climate changes have recently become one of the most discussed topics in scientific circles and in the world of politics and the media alike. They raise many controversies as a result of poor recognition and understanding of mechanisms and phenomena governing the climate on our planet. Among many questions related to climate changes is the question how different ecosystems will react to them? What changes will occur in the intensity of mass and energy fluxes exchanged between ecosystems and the atmosphere?

To answer this question, scientists from different countries have been carrying out measurements of mass and energy fluxes exchanged between different ecosystems and the atmosphere for many years. Most frequently measured parameters are apparent heat, water vapor, and since the early of 1990s, also carbon dioxide fluxes (Kędziora and Olejnik 1996, Lee *et al.* 1996). In order to estimate the volume of this exchange number of measurement techniques have been introduced, but in recent years, the most developed is the eddy covariance method. The theoretical bases of this method have been known for many years (Reynolds 1895, Swinbank 1951), however, the high technical requirements expected from the equipment that is collecting, storing and processing the measurement data have caused that only the development of modern electronics allowed its wide implementation in field conditions (Wofsy 1993).

In the past 20 years, the eddy covariance (EC) method has been spread around the world, but the high costs of equipment constantly limits its range of application. In Poland, continuous measurements of CO_2 and H_2O fluxes measured with the EC method were started on wetland ecosystem just in the year 2003. A lot of research projects, using the networks created on the basis of this type of greenhouse gases monitoring stations, were established much earlier in the world. They were used,

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and most of them continue to serve this function, for continuous CO_2 and H_2O fluxes measurements (e.g. Boreas, CarboEurope and others) (Aubinet *et al.* 2000).

The results of the measurements of water vapor and carbon dioxide fluxes exchanged between Rzecin wetland and the atmosphere in 2004 are presented in this paper.

The aim of this paper is to show the seasonal variations of water-use efficiency (*WUE*). Presumably, the value of this coefficient varies cyclically with a change in the quantity of radiation and temperature.

MATERIALS AND METHODS

Study site

The results presented in the paper are based on measurements carried out on peat-bog from January till December of the year 2004. The wetland area is located close to the Rzecin village in the north-western part of the Wielkopolska Region, about 80 km far from Poznan in the north-west direction (52° 45'N, 16° 18'E, at an altitude of 54 meters above sea level). The area of the bog is about 140 ha. About 34 ha of wetland is occupied by Rzecin lake, which is in the phase of strong overgrowing. Currently, this bog has a temporary character and the main plant species found there are as follows: *Sphagnum sp, Dicranum sp., Carex sp., Phragmittes communis, Typha langifolia, Vaccinium oxicoccus, Drosera rotundifolia, Potentilla palustris, Ranunculus acris, Menyantes trifoliata,* etc. (Wojterska 2001).

The average annual air temperature and precipitation is 8.5°C and 526 mm, respectively. Westerly winds prevail (Farat *et al.* 2004).

The measurement tower was placed in the middle of the bog (Fig. 1). Before selecting the exact location where the measurement tower should be constructed, the interaction area was analyzed by means of the stochastic model of Lagrangian (Kljun 2002).

The required stability of the measuring instruments was achieved by using a platform built on 36 pine piles with the average length of 10 m each. A steel tower with the square cross-section of 1 m and the height of 2 m was attached to this platform. A 1.5 m high mast was mounted in the middle of it, on top of which the EC system is installed. The whole construction has the height of 4.5 m.

After building a stable platform and the measuring mast, all sensors were installed on it. The entire system was supplied with electricity, transmission cables and a data acquisition module. All data were collected through the system of data transmission and power control monitored by the main steering computer located about 500 m away from the measuring tower.

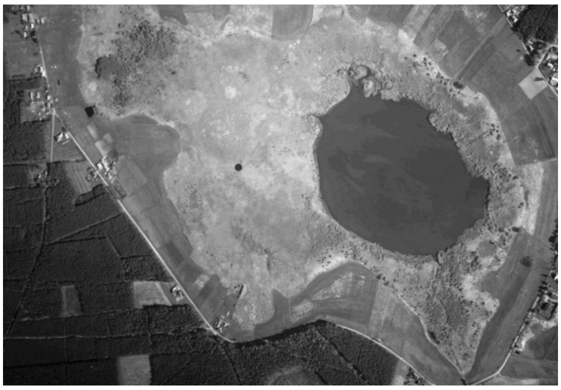

Fig. 1. The aerial photo of the measurement site. The black dot – the location of the Eddy Covariance measuring tower

Theory of the eddy covariance method

In order to estimate CO_2 and H_2O fluxes the eddy covariance method (EC) was used. The measuring system consists of two basic elements: an ultrasonic anemometer and a gas analyzer. These devices have to perform measurements with the frequency of 10 Hz or higher, in order to meet the demands of the method. In practice, the most commonly used frequency is 20 Hz. In this way the measurement system provides data on the fluctuations of vertical air movements and concentration values of the studied gas.

The eddy covariance method is highly recognized by researchers because of the simplicity with which it expresses the fluxes values obtained from the collected data. The mathematic, general notation of this method can be presented as follows:

$$F = \underbrace{\overline{w}\overline{\rho}}_{I} + \underbrace{\overline{w'\rho'}}_{II} \tag{1}$$

where: F – flux of any scalar value (µmol m^{-2} h^{-1}),

$\overline{\rho}$ – the average value of scalar quantity (temperature (K) for energy or substance density (µmol m^{-3}) for gases,

ρ' – fluctuations of the scalar value,

$\overline{w}$ – the average value of vertical component of wind velocity (m s^{-1}),

w' – fluctuations of vertical component of wind velocity (m s^{-1}).

From equation (1) follows that the total vertical flux of any scalar value is the sum of the average vertical mass flux (term *I*) and turbulent flux (term *II*) (Moncrieff *et al.* 1997, Webb *et al.* 1980). The second term of equation (1) is the covariance of the vertical component of a wind velocity and the scalar value, whose flux we are interested in this term determines the average value of the product of momentary deviations of the vertical component of a wind velocity and the scalar value from their mean value and describes the turbulent mass and energy exchange between the active surface and the atmosphere. The first term of equation (1) is a mass flow caused by the vertical, non-zero wind velocity component. As it follows from the above equation, to calculate the flux one has to assume a time period in which the covariance and the mean of measured values should be calculated. This is the so called integration time, which is usually assumed to be 30 minutes.

During the calculation of fluxes several commonly used and recommended improvements were applied. One of them is the Moor amendment (Moore 1986), consisting of shifting of series (relative to each other) of the vertical component of wind velocity values and the scalar values. The aim is to find the largest covariance values obtained for each shift between these series. This leads to correcting an error in the flux calculation resulting from the fact that the instruments included in the system are located at some distance from each other (tens of centimeters). Spatial separation of instruments means that, depending on the wind direction, the studied portion of the air first reaches first instrument and after a while another one. This causes some shift in the time of both measurement series and thus leads to underestimation of the flux value (lower value of covariance).

The eddy covariance method has some limitations. One of them is an incorrect estimation of fluxes during stable atmospheric conditions, when there is no turbulences in the atmosphere (or turbulences are not sufficient), because in such moments occurs a phenomenon known as flux non-stationarity. This means that the spot measurement is unrepresentative for the surveyed fluxes (Baldocchi 2003). Fluxes estimated based on the parallel measurements carried out at a different height above the surface of the studied ecosystem would be completely different than those calculated based on the EC measurements. This is caused by the fact that the developed boundary layer is not yet fully formed. The measurement data collected in such period should be excluded, as for this data the theory of covariance cannot be applied.

In order to determine the time when there was sufficient turbulence in the atmosphere, the test on the threshold value of friction velocity was performed (Gu *et al.* 2005, Urbaniak 2006). Data measured at the time when the value of friction velocity exceeded the previously established threshold value for that place were considered correct and used in further analyses. At the same time the data were put to the stationarity tests according to the procedure described by Foken and Wichura (1996). This procedure involves the calculation of the covariance of the vertical component of a wind velocity and the scalar value (e.g. CO_2 and H_2O) for each five minutes of every half an hour. If the average of the covariance obtained in this way did not differ more than 30% from the covariance calculated for the entire half an hour, it was considered that, in the given half an hour the flux was stationary and such data were qualified for further analysis.

Experimental setup

The concept of the measuring system applied in this study was fully developed and implemented in the Meteorology Department of the Poznan University of Life Sciences (Olejnik *et al.* 2001).

The measurement system consists of an acoustic anemometer used to measure fluctuations in the wind velocity (R3 Gill Instruments, Lymington, United Kingdom) and a spectrometric gas analyzer with an open path (LI7500 LI-COR Inc. Lincoln, NE, USA) used to measure fluctuations in the concentrations of the carbon dioxide and water vapors in the atmosphere (Instruction Manual LI-7500 2001, Omnidirectional… 2004). This instrument relates the contents of the measured gas to the volume of the measurement path. This means that factors, such as barometric pressure and air temperature affect the measured values. In order to overcome this problem the Webb, Pearman, Leuing (WPL) correction was applied (Webb *et al.* 1980).

The anemometer, thanks to the three-dimensional measurements of wind vector (the independent measurement of horizontal and vertical components) allows also to measure the fluctuations of the air temperature (Kaimal and Gaylor 1991).

The atmospheric pressure was measured by a fast responding sensor (Electronic Kest, Poznan, Poland), which allowed to do the correction of CO_2 and H_2O concentrations in respect to fluctuations of air pressure.

The data from the above mentioned instruments were transmitted in an analog way to an analog-digital converter (Data-logger KEST16, Kest Electronic, Poznan, Poland), which processed at the rate of 40 Hz the data and sent them to a PC using a standard RS422 interface. The program for data acquisition saved them on this computer's hard drive by creating successive files containing all the data collected in a given half an hour.

In addition to the eddy covariance system measurements, basic meteorological parameters were measured simultaneously. Air temperature (*Ta*) and relative humidity (*RH*) were measured by means of HMP90Y sensor (Vaisala Helsinki, Finland). These measurements were conducted at four heights: 0.50, 1.50, 2.50 and 4.50 m above the ground level. Moreover, the measurements of soil temperature (*Ts*) were carried out, using semiconductor thermometers installed at the following depths: 0.02, 0.04, 0.06, 0.10, 0.20 and 0.50 m below the ground surface. The measurements of the short and long wave radiation of the active surface were performed by CNR1 sensor (Kipp & Zonnen, Delft, The Netherlands). Additionally, a separate CM3 sensor (Kipp & Zonnen, Delft, The Netherlands) was used to measure the total radiation (*Rg*) and net radiation (*Rn*) (NRLITE Kipp & Zonnen, Delft, The Netherlands).

Photosynthetic photon flux density (PPFD) was measured by the following sensors: SKP 215 (Skye Instruments Ltd., UK) and BF3H (Delta-T Devices Ltd., Cambridge, UK). All instruments used for radiation measurements were installed at the height of 2.50 m above the ground. The exception is BF3H sensor that was installed at the height of 3.20 m. The soil heat flux was measured with four soil plaits HFP01SC (Hukseflux Thermal Sensors, Delft, The Netherlands) installed at the depth of 0.02 m. Precipitation was measured with two instruments: a Hellman rain-gauge, which recorded a daily precipitation and RG2-M cradle rain-gauge (ONSET, Pocasset, MA, USA) equipped with a heating element of autonomous design (heating switched on at temperatures below 5^{o}C), which allowed carrying out automatic precipitation measurements also in winter period. Data from this rain-gauge were recorded on a HOBO recorder (Onset, Pocasset, MA, USA). Other data from the above mentioned sensors were recorded on KEST32 datalogger (Electronic Kest, Poznan, Poland).

The measuring system, in spite of its modernity, does not ensure 100% data coverage during measurements. One of the reasons was explained earlier (no turbulence), in addition, there are failures such as power outage and data transmission interruptions, which increase the number of gaps in the database. The overview of seasonal and daily flows of measured fluxes requires supplementing these gaps.

Measurements of carbon dioxide fluxes

In the case of completion of the gaps in the CO_2 fluxes data, a procedure relying on the empirical models describing the exchange of this gas was used (Falga *et al.* 2001). This procedure is based on the division of the measurement period into day and night sessions and applying the appropriate formula in the case of each of these periods (Urbaniak 2006). Michaelis-Menten (1913) model was used in order to estimate the daily flux, following Carrara and the team (Carrara *et al.* 2004):

$$NEE = \frac{-\alpha\, PPFD}{1-(PPFD/2000)+(\alpha\, PPFD/GEP_{opt})} + R_{day} \qquad (2)$$

where: NEE – net ecosystem exchange (μmol CO_2 m^{-2} s^{-1}), calculated from the formula:

$$NEE = F_c + S_c \qquad (3)$$

where: F_c – turbulent flux of CO_2 measured using EC system (μmol m^{-2} s^{-1}),
S_c – the quantity of CO_2 stored below the measuring instruments (storage) (μmol m^{-2} s^{-1}).

The remaining elements in the formula 2 are:
α – the ecosystem growth rate (μmol CO_2 m^{-2} s^{-1}),
$PPFD$ – Photon Flux Density of photosynthetic active radiation (μmol quantum m^{-2} s^{-1}),
GEP_{opt} – gross ecosystem production for the "optimal" PPFD,
R_{day} – ecosystem respiration during the day (μmol CO_2 m^{-2} s^{-1}).

Whereas for the nocturnal period the ecosystem respiration model of Schlentner and Van Cleve (1985) was used following Fang and Moncrieff (Fang and Moncrieff 2001):

$$F_{RE,\,night} = \frac{a}{a+b^{-\left(\frac{T_C-10}{10}\right)}} + c \qquad (4)$$

where: $F_{RE,night}$ – flux density of nocturnal ecosystem respiration (mol m^{-2} s^{-1}),
T_C – air or soil temperature (°C),
a, b, c – estimated parameters of the model.

The respiration model was used not only to estimate the net flux at night, but also, after substituting the obtained values to the formula 2 as R_{day}, to assess the ecosystem respiration during the day. As a result of the used procedure it became possible to draw up the course of the CO_2 fluxes comprising the balance of this gas exchange (Fig. 2). For easier understanding of the values of the individual fluxes they are presented in this figure in (mg m^{-2} s^{-1}), (1 mol (CO_2) = 44.01 g (CO_2)).

Relations between individual fluxes included in the carbon balance of the ecosystem (Fig. 2) can be described using the following equation:

$$-NEE = NEP = GEP - R \qquad (5)$$

where: NEP – net ecosystem productivity,
GEP – gross ecosystem photosynthesis (or productivity),
R – ecosystem respiration.

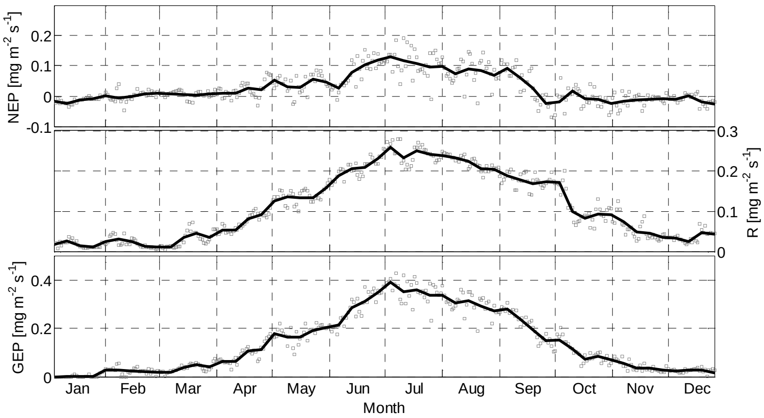

Fig. 2. The annual course of NEP, R and GEP fluxes on Rzecin wetland in 2004. Points indicate daily averages, lines weekly averages (sign directions of CO_2 fluxes as in the equation (5))

Measurements of water vapor fluxes

To fill the gaps in the water vapor fluxes a procedure based on Penman and Penman – Monteith formulas were used. Original Penman formula was applied to determine the impact on the evaporation of the factors other than energy. This allowed to estimate the conductivity of the surface and aerodynamic resistance, which were then used in Penman – Monteith formula. The applied procedure for completion of the gaps in the H_2O fluxes is rather complicated that is why its full description is presented below together with the chart that shows the successive steps in the adopted procedure (Fig. 3).

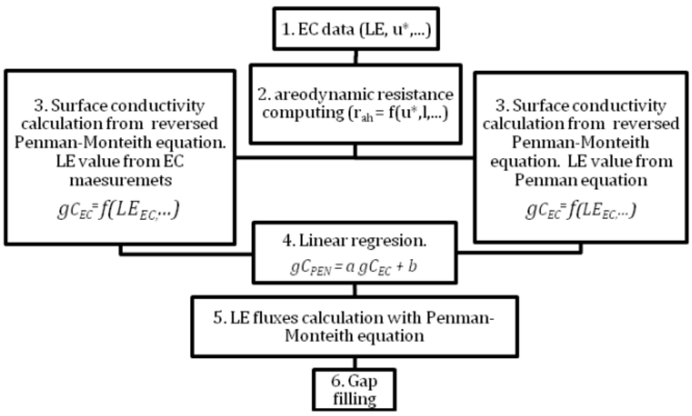

Fig. 3. The diagram of the procedure of filling the gaps in the database of latent heat flux values (LE)

In the first place (step 1) the data gaps of H_2O fluxes in the database were found. Then (step 2) the aerodynamic resistance was calculated by means of (6) and (7) formulas (Gash *et al.* 1999). The first one (6) was used, when the data from the anemometer included in the EC system were available. The second formula (7) was used when the data from the anemometer were not available, and only the horizontal wind velocity measured by another anemometer was accessible.

$$r_{ah} = (l/u* + 2/\mathrm{k} + \Psi/\mathrm{k})/u* \tag{6}$$

where:

- l – horizontal wind velocity (m s^{-1}),
- $u*$ – friction velocity (m s^{-1}),
- k – von Karman constant amounting to 0.41,
- Ψ – parameter describing the state of atmospheric stability,

and:

$$r_{ha} = \frac{\log(z\text{-}d/z_{0m}) \cdot \log(z\text{-}d/z_{0h})}{l\,\mathrm{k}^2} \tag{7}$$

where:

- z – height of the wind velocity sensor placement (m),
- d – height of the zero-plane raising (m),
- z_{0m} – surface roughness parameter for the momentum (m),
- z_{0h} – surface roughness parameter for water vapor and heat (m).

Step 3 involved calculating the surface conductivity for the periods when the EC system measurements were available, by transforming Penman-Monteith formula. The same was done with the values calculated by means of Penman formula, which were used to calculate the conductivity by inverted Penman-Monteith formula. The next step (4) involved calculating the conductivity values for the periods when evaporation data gaps occurred. It required the linear regression analyses with the conductivity values, calculated on the basis of the collected material, as the independent data and the values calculated with Penman formula as the dependent data. The values of the surface conductivity obtained in this way were substituted for Penman-Monteith formula (step 5), and thus, the values of modeled latent heat flux (*LE*) were obtained, which then were used to fill the gaps in the data (step 6).

The above mentioned steps allowed conducting fairly accurate and precise estimation of the water vapor fluxes exchanged between Rzecin wetland surface area and the atmosphere The course of seasonal water vapor flux presented in the form of its energetic equivalent – latent heat flux density (*LE*) for 2004, after filling the gaps, is shown in Figure 4. According to the principle adopted in the description of *LE* flux, the flux outgoing from the surface has a positive value, whereas it assumes a negative value when the water vapor condensate on the surface of the ecosystem.

The data obtained from measurements and modeling were then used to calculate the water use efficiency (*WUE*) by the studied ecosystem, in the period of the whole year. This indicator is defined as the ratio of net carbon assimilated in the form of CO_2 fluxes to evapotranspiration (formula 8); both fluxes have been determined in mass units (Chen *et al.* 2002).

$$WUE = \frac{F_C}{F_{H_2O}} \tag{8}$$

where:

Fc – net flux of C as CO_2 (g m^{-2}),

F_{H_2O} – water vapor flux (kg m^{-2}).

In this case it is also assumed that F_{H_2O} has a positive value when the ecosystem surface gets rid of water (evapotranspiration) and the contrary when it gains water (condensation).

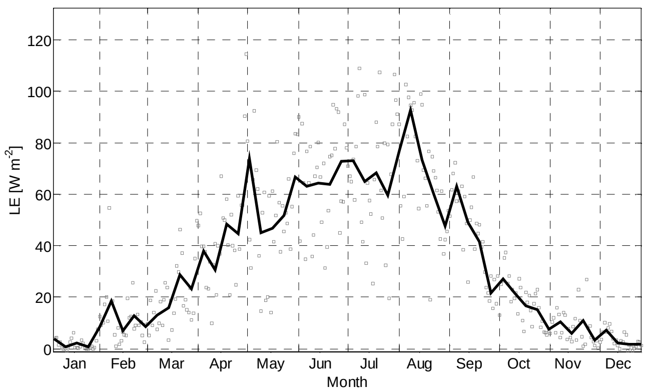

Fig. 4. The course of annual LE fluxes on Rzecin wetland in 2004. Points indicate the daily averages and lines weekly averages

This indicator combines well the variability of both studied fluxes and provides the information on how much water must evaporate from the ecosystem to allow it to assimilate a unit of carbon (C-CO_2). This means that the calculation of this indicator makes sense only for the periods during which the absorption of CO_2 occurred. Therefore, only the data from the periods in which both *NEP* flux and F_{H_2O} were positive were used for *WUE* calculations. This reduces significantly the amount of the data, since such situation takes place only during daytime. The data from the periods when the above conditions were met, are presented in

Figures 5 and 6, for the fluxes of daily *NEP* (C-CO_2) and F_{H2O} respectively. The results of the calculations of *WUE* index are shown in Figures 7 and 8.

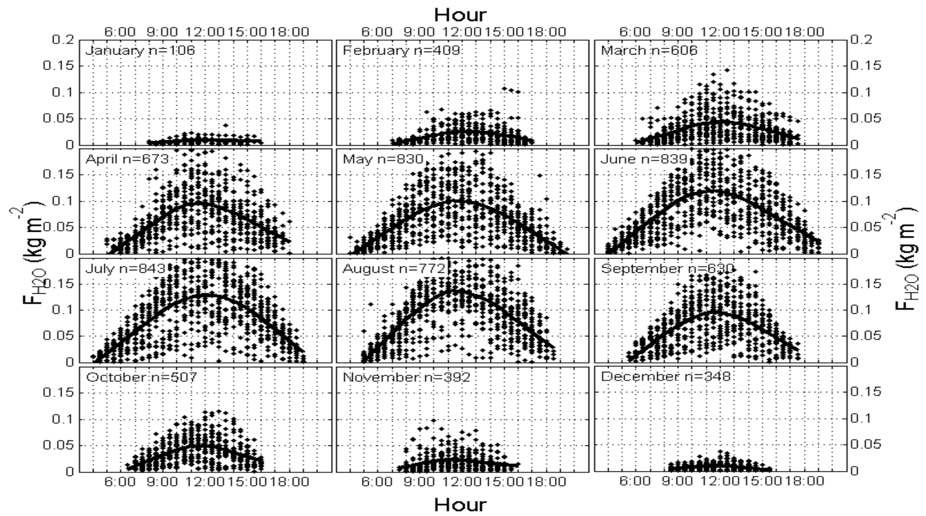

Fig. 5. Daily trends of water vapor flux over the wetland in Rzecin. The trends over the measurement periods are presented using Friedman smoothing, a technique that selects the appropriate degree of smoothness based on cross-validation

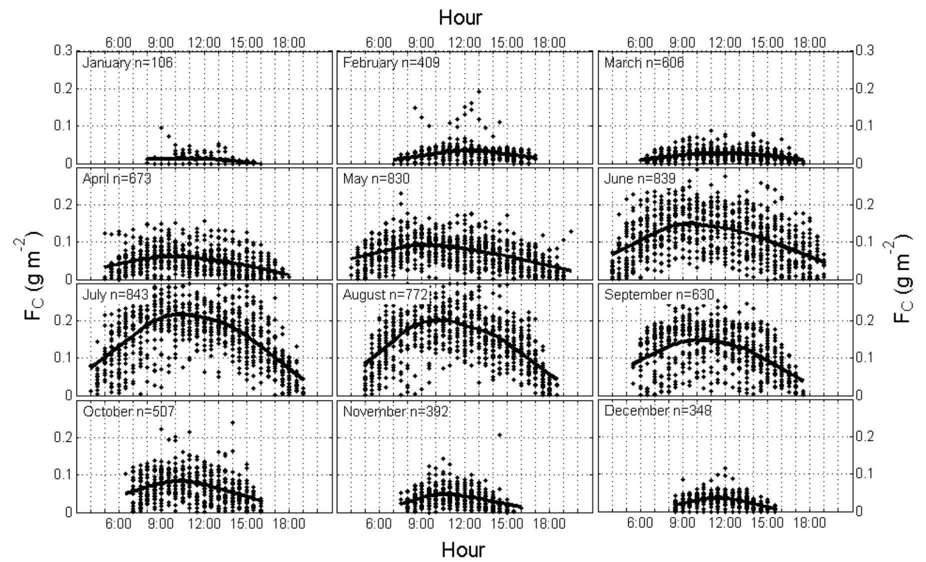

Fig. 6. Daily trends of net flux of C as CO_2 (Fc) over the peat-bog in Rzecin. The trends over the measurement periods are presented using Friedman smoothing, a technique that selects the appropriate degree of smoothness based on cross-validation

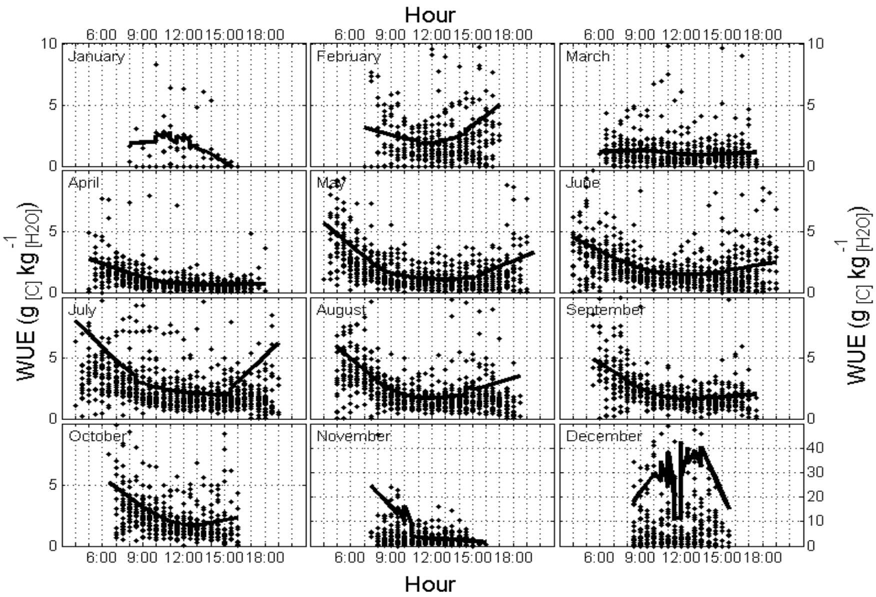

Fig. 7. Daily changes in water-use efficiency (WUE) at Rzecin wetland in 2004. The trends over the measurement periods are presented using Friedman smoothing, a technique that selects the appropriate degree of smoothness based on cross-validation (Chen *et al.* 2002)

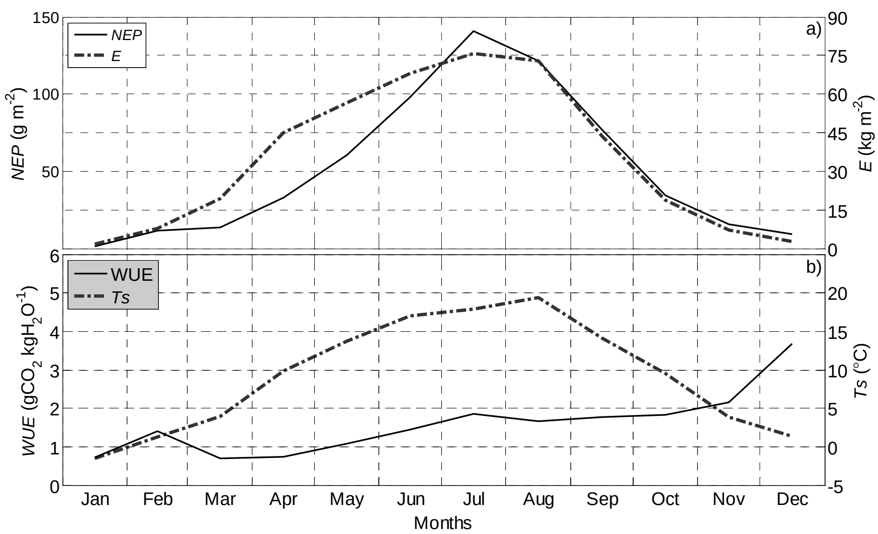

Fig. 8. The monthly values of the net ecosystem production (NEP) and evapotranspiration (E) (a) and the ratio NEP/E (WUE) and soil temperature (Ts) at 2 cm (b)

RESULTS AND DISCUSSION

For the purposes of this study the fluxes were calculated on the basis of the data obtained directly from the EC system, these calculations involved the triple rotation and WPL correction. In this way the quality requirements for the calculations of mass fluxes were met, which have been described in the literature over the past few years (Aurel 2005, Baldocchi *et al.* 1998, *Black et al.* 1996, Lafleur 2003).

The gaps in the measurement data or data that do not meet the above criteria were supplemented by values obtained in the process of application of carefully selected and parameterized simulation models of water vapor and carbon dioxide fluxes (Aurel 2005, Falge *et al.* 2002).

Then, the data that contain positive values of *NEP* and F_{H_2O} were selected from the daily data material. The flux values selected in this way were converted into mass units and *WUE* values were calculated for each half an hour.

The analysis of the daily *WUE* courses was carried out for individual months. Simultaneously, the analysis of changes in the dynamics of daily runs of *NEP* and F_{H_2O} values for each month was carried out.

The monthly breakdown is a subjective way of dividing the time series, but it makes it easier to trace the seasonal variations in the studied fluxes and the coefficient.

The net flux of CO_2 during the day, is the resultant of photosynthesis process and the ecosystem respiration. The process of photosynthesis depends mainly on the amount of ambient heat, intensity of photosynthetic photon flux density (*PPFD*), environment chemistry (e.g. nutrient availability), water availability and the development phase of plants. By contrast, the ecosystem respiration depends on the temperature of plants and the topsoil, soil moisture and the available quantity of organic matter in the soil.

In turn, the evaporation process is dependent primarily on the availability of water, energy, and the amount of water vapor deficiency in the air (called vapor pressure deficit (*VPD*). Hence, one can say that evaporation is a physical process, rather than a biological-physical one. But in the case of ecosystems covered with vegetation, physiological processes of plants (transpiration) have a significant influence on the evaporation process. In the case of a wetland, it appears that transpiration is a small portion of the total flux of water vapor observed over the surface of the bog.

Both the daily course of H_2O (Fig. 5) and C-CO_2 (Fig. 6) show a shape which indicates the dependence on the amount of available radiant energy reaching the surface ecosystem. Each curve shows the elevation in the afternoon, but its shape is different for each month and it may be caused by different degree of vegetation development.

Daily evapotranspiration

The shapes of average daily evaporation curves (Fig. 5) seem to confirm the supposition that the process of evapotranspiration on the bog in Rzecin is affected mainly by the physical factors, such as the amount of available radiant energy, *VPD*, air temperature and surface area. The effect of plants' activity on the intensity of evapotranspiration is rather inconspicuous in this figure and this may be either due to the comfortable water conditions (no limitations of available water) in which the plants grow, or the fact that evaporation from the surface soil (water hollows) is significantly greater than the process of evaporation from the plants' surface.

Daily net ecosystem production (*NEP*)

Interestingly it seems that the smoothed curves describing the value of the C-CO_2 flux in the first half of the day are inclined at a greater angle than in the afternoon (Fig. 6).

This asymmetry is most pronounced during the summer months. The first factor limiting photosynthesis is generally the amount of reaching *PPFD* radiation, however, the shape of curves in this situation should be symmetrical in the respect of 12:00 o'clock - just as it is in the case with evaporation. It seems that other, not physical but rather physiological factors limit the ability of the ecosystem to absorb CO_2 from the air in the afternoon.

One of the most probable reasons may be running out of nutrients, at the end of the day, since the environment in which the process of photosynthesis takes place is relatively poorly fertile.

Another reason may be the water stress. This is, however, very unlikely because the wetland belongs to the environments rich in water and the depletion of it in one day is rather impossible. An additional proof of this negation is the symmetrical shape of the daily evaporation curves. If the environment was exposed to the water stress in the afternoon then the value of F_{H_2O} would be strongly reduced and thus, a clear asymmetry of the daily evaporation curve would appear.

Another very convincing explanation is the increase in the intensity of the ecosystem respiration (*R*). Since it depends mainly on temperature and humidity of the surface, then the increase in the temperature during the day may cause the escalation of this process, which in turn affects the balance of CO_2 exchange.

Daily *WUE* course

The nature of the difference described above, between the shape of the daily trends of these fluxes is reflected in *WUE* values and the shape of the trend-line of this factor. Figure 7 shows how this ratio changes during the day and each month. One can see that in the months from April to October the trend curves reflect the relationship of these fluxes in a characteristic way which is described by *WUE*.

High *WUE* coefficient values give evidence of more efficient use of water by the ecosystem during photosynthesis. In the discussed months, we observe higher *WUE* values in the morning and evening, however, the latter are lower. This is understandable because of the previously discussed trends in the courses of C-CO_2 and H_2O fluxes.

The high values of *WUE* in winter months may seem surprising. This is due to small values of evapotranspiration that occurs in these conditions. Photosynthesis is also hindered by the presence of peat moss, which makes the ecosystem very quickly switch from being an emitter into an absorber of the atmospheric CO_2. This is because the assimilation process of these plants occurs throughout the whole year. The possibility of photosynthesis outside the period of vegetation is the inherent feature of wetlands. This means that in a favorable moment, when there is a correspondingly large radiation flux and the temperature rises above 0°C, the ecosystem assimilates CO_2. This process has less inertia than the evapotranspiration, hence, the high values of *WUE* in the cold months. Moreover, at that time, the moments when the two fluxes take positive values are not as frequent as in the warm part of the year. Figures 5, 6 and 7, show the names of the months and the number of half-an-hour periods in a given month, which fulfilled the criteria adopted for *WUE* calculation.

For cold months like January, February, November and December the number of such half-hours was about 2-2.5 times lower, and in January six - seven times lower than during the rest of the year. This may be the cause of such irregular shapes of daily *WUE* trend curves in January, November and December. However, despite these doubts, it is worth noticing that the peat-bog can grow in difficult weather conditions, and it is doing so using the water efficiently. *WUE* reaches particularly high values in November and December, when they exceed the summer values by several times. The appropriate explanation for this fact seems to be that the vegetation of the peat-bog, which has evolved over the vegetation period is still functioning quite well in these months. Whereas the decrease in the solar radiation and consequently the temperature has influenced the evapotranspiration to a greater extent than it influences *NEP* fluxes.

Figure 8 was prepared to better illustrate the seasonal variability of the particular fluxes and *WUE*. The figure illustrates that in the first quarter of the year *WUE* value ranges from 0.8 to about 1.5 g(C-CO_2) kg^{-1} (H_2O) (gram of carbon in the form of carbon dioxide per kilogram of water). The clearly outlined growth of this coefficient in February may be caused by the response of the ecosystem to increasing *PAR* radiation over time. However, the amount of energy that the peat-bog has received since the beginning of the year is still negligible. The situation is worsened by high moisture level of the environment, and thus high specific heat. The peat-bog is heated very slowly which in February may be reflected by the increased *WUE*. On the other hand, the value of both fluxes is so small during cold months (Fig. 8a) that any temporary changes in weather conditions significantly affect the average value of both fluxes (Fig. 5, 6), which means that it is difficult to find appropriate explanation for *WUE* behavior under such conditions.

In the next month the peat-bog heats up, so does the air above it which increases water vapor deficiency (*VPD*), stimulates evaporation and makes *WUE* decrease.

In the second quarter of the year one can see an increase in *WUE* factor from 0.8 to nearly 2 g(C-CO_2) kg^{-1}(H_2O). The beginning of the vegetation season in this period causes major changes in the environment. As a result of intensive biomass growth, the intensity of photosynthesis increases. Clearly it is beginning to dominate over the ecosystem respiration which is reflected in higher values of net flux. A clear predominance of photosynthesis, over respiration is conspicuous from March to July (Fig. 8a). The simultaneous increase in temperature causes also increase in evaporation, but this process does not intensify as fast as photosynthesis – hence, the increase in *WUE*. In the months from July until October *WUE* remains relatively constant at about 2.8 g(C-CO_2) kg^{-1}(H_2O). At the same time one can observe a decrease in the value of both fluxes illustrated in Figure 8. It is caused by a decrease in PAR radiation on one hand, and the temperature decrease on the other (Fig. 8b). This figure clearly shows as the water vapor flux responds to the changes in the temperature of the peat-bog surface. Photosynthesis remains at a higher level than in the first half of the year, when the radiation and temperature conditions were similar, due to the better developed vegetation.

Further decrease in the temperature in November and December causes even greater reduction in evaporation, whereas the net flux of CO_2 still remains above zero. Only a drop in temperature, followed by frost, which occurs more and more often results in a reduction of photosynthesis. This is because part of the plants (vascular) lose their assimilation system in this way. Therefore, in January *WUE* factor is significantly lower than at the end of the year. Apart from that, the peat-bog has dried out during the year, which hampers the process of evaporation, while having a limited impact on the intensity of photosynthesis. Both factors cause a large increase in *WUE*.

CONCLUSIONS

1. The ecosystem water use efficiency (*WUE*) factor described in this work is a parameter describing the whole ecosystem in terms of efficient water management in the context of the atmospheric CO_2 absorption. Its seasonal course is a little surprising. One would expect that, similarly to temperature values, it will grow in the first half of the year, and decrease in the second. It is a result of, occurring in autumn, vegetation season of the plant mass developed during the summer.
2. Moreover, since photosynthesis theoretically should not occur during winter, outside the vegetation period, in such period the index should take the value of zero. However, the presence of "primitive" plants such as sphagnum mosses causes that the assimilation of CO_2 from the atmosphere takes place in Rzecin peat-bog even outside the classic vegetation period. This fundamentally alters *WUE* behavior during winter and completely modifies the assumptions as to its annual course.
3. Daily *WUE* trends show how important for the value of this index is evaporation, that is physical vaporization. Reduction of the temperature of the peat-bog surface and air causes an immediate reduction in the water vapor flux. However, decreasing of the intensity of gas exchange by vegetation contributes to this flux insignificantly.
4. The curves describing *WUE* courses during the day are asymmetrical in respect of noon, and this feature results from the reduced capacity for CO_2 absorption by the plants in the afternoon. The reason for such *WUE* course remains unknown, but it seems very likely that this is not caused by the water stress of plants since the daily curves for the evapotranspiration process do not show that asymmetry. One of the most likely explanations is the intensification of the ecosystem respiration or depletion of nutrients available in such poor environment as the peat-bog.
5. Further research and development of the measurement capabilities of the measuring station in Rzecin should bring us closer to answering the question about the reason for this asymmetry.

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9. WRF MODEL IN THE CLIMATE RESEARCH*

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INTRODUCTION

The Weather Research and Forecasting (WRF) model (http://www.mmm.ucar. edu /wrf/ users/docs/arw_v2.pdf), (Michalakes *et al*. 2004), is a mesoscale, nonhydrostatic (Janjic *et al*. 2001), (Ooyama 1990) numerical weather prediction and atmospheric simulation system. It was created as a collaborations of many research groups e.g. National Center for Atmospheric Research's (NCAR), National Centers for Environmental Prediction (NCEP). High flexibility and portable code of WRF Model gives an easy installations process on massively-parallel supercomputers and laptops.

WRF Model is a regional model (Michalakes *et. al*. 1999), thus for the simulation of real case studies it requires initial and boundary conditions from general circulation models (GCM) or reanalysis. The outputs from the GCM have lower resolution then WRF model domains and the data are in grib (GRIdded Binary) format that's why it is necessary to change it in a section called Preprocessing System (WPS) – Figure 1. The names of necessary fields from GCM models to start WRF simulations are presented in Vtable file in WPS.

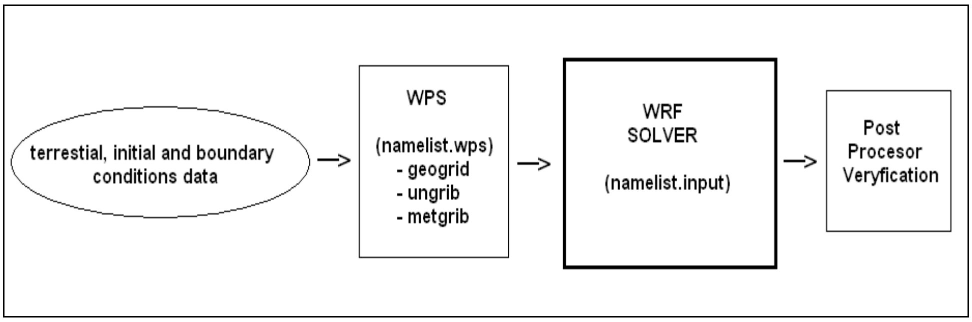

Fig. 1. Simplified idealized WRF system components

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WPS consists of geogrid.exe, ungrib.exe and metgrid.exe programs. The geogrid defines the simulation domains, and interpolates various terrestrial data sets e.g. soil categories, land use category, terrain height, monthly albedo to the model grids. The size and resolution of domains are set in namelist.wps file. The program ungrib reads grib file form GSM, decodes the data, and rewrites them in a simple WRF format called the intermediate-format. The program metgrid interpolates horizontally the intermediate-format meteorological data that are extracted by the ungrib onto the simulation domains defined by the geogrid.

The outputs of WPS are needed by WRF SOLVER. The WRF SOLVER is the key component of the modeling system which consists of several initialization programs for idealized, and real-data simulations. For the real data simulations main outcomes are generated by two programs called real.exe and wrf.exe. The real.exe program interpolates meteorological fields from WPS to WRF eta levels (Laprise, 1992). The wrf.exe initiates the computing process.

VERIFICATION OF THE WRF MODEL

The most important thing before using any regional model is its adaptation to the climatic conditions of specific areas. Many experiments were conducted to obtain the best parameterizations of microphysics, shortwave radiation and convection for the WRF outputs for Poland area. In the first step there seven Polish stations (Chojnice, Kalisz, Katowice, Łódż, Poznań, Rzeszów, Zakopane) were chosen to compare real data with model outputs – Figure 2.

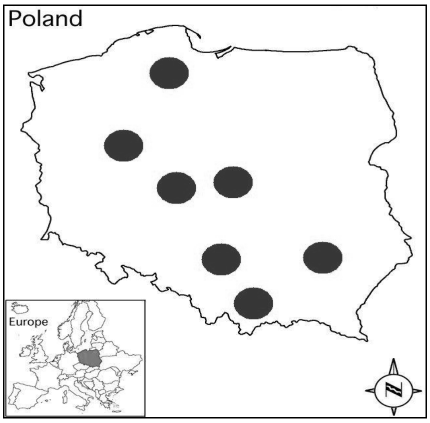

Fig. 2. Geographical positions of meteorological stations

For all stations 63 experiments were prepared with mixed parameterizations (7 – parametrizations of microphysics: Kessler, Purdue Lin, WSM3, WSM5, WSM6, Eta GCP, Thompson; 3 – parametrizations of shortwave radiation: Dudhia, Goddard, GFDL SW; 3 – parametrizations of convective and shallow clouds: Kain – Fritsch, Betts – Miller – Janjic, Grell - Devenyi). The experiments were prepared for four different, specific months: warm January 1989, cold January 2006, wet July 1997, hot and dry July 2006.

To compare real and model data two meteorological parameters were chosen namely mean daily temperature and daily total precipitation. In the case of real data the values of temperature and precipitation were connected with characteristic point – geographical position of the meteorological station. In the model it is difficult to obtain the data in the same point as a real data. It depends on resolution of the model. In all experiments the WRF grid was 30 km, that's why the values for temperature and precipitations for selected stations were calculated as a mean value from four nearest grid points – Figure 3.

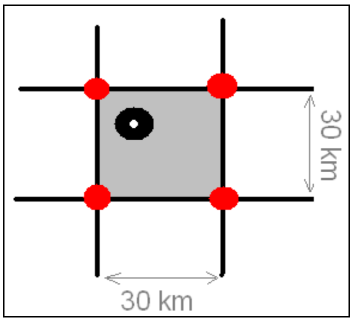

Fig. 3. The grid of the WRF model. red points – grid points, black/white point – position of the selected station

In the first step real and model values of daily mean temperature were compared. The comparison was performed for all selected stations, four selected months and 63 mixed sets of microphysics, shortwave radiation and convection parameterizations. In all cases the minimal value in January 2006 was not achieved, but local trends of real and model data were similar – Figure 4. Generally this step excludes all sets with third (GFDL SW) shortwave parameterization – the differences between real and model data were the highest using this scheme.

To compare real and model data five statistics: minimal and maximal values, mediana, the first and third quartile were also used. In this step similar result was obtained – excluding the GFDL SW shortwave parameterization. The values of all statistics for model outputs were lower then real data when this scheme was used – Figure 5.

The correlation coefficients for real and model data for all experiments were also calculated. The highest values of the correlation coefficient were for following configuration of parameterizations: Kessler's scheme (Kessler 1969) as a parameterization of microphysics, Dudhia's (Dudhia 1989) and Goddard's (Chou and Suarez 1994) schemes as a parameterizations of short wave radiation, Grell-Devenyi's (Grell and Devenyi 2002) ensemble scheme as a parameterization of convective and shallow clouds.

The similar comparisons were prepared for total precipitation. Maximal and minimal values of total precipitation, mediana, the first and third quartile, monthly total precipitation were compared. The correlation coefficients between real and model data were also calculated. Comparison for total precipitation gave similar results – the similar sets of parameterization schemes – as in the case of temperature.

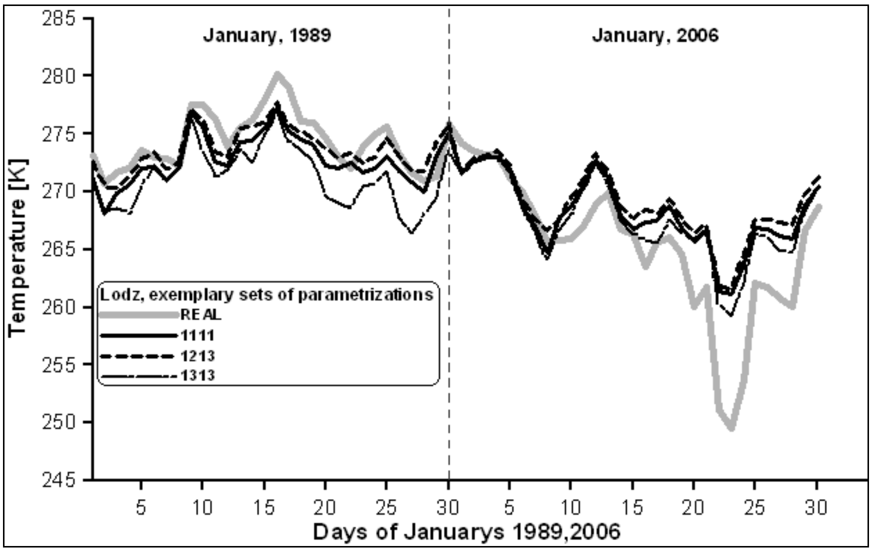

Fig. 4. Exemplary comparison of real and model data for Łódź, Januarys 1989, 2006. Description of legend: the first number - parameterization of microphysics, the next numbers: parameterization of shortwave radiation, boundary layer and convective and shallow clouds.

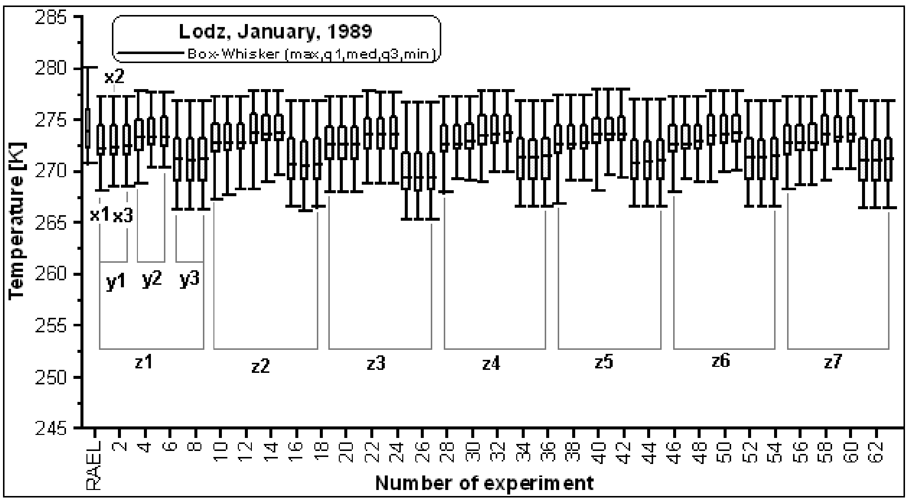

Fig. 5. Exemplary comparison of statistics (top: maximal value, quartile 3, mediana, quartile 1, minimal value) real and model data for Łódź, January, 1989. Green box – real data, red boxes – model outputs for established set of parameterization: z1,…z7 – parameterizations of microphysics, y1,…,y3 – parameterizations of shortwave radiation, x1,…,x3 – parameterizations of convective and shallow clouds.

WEATHER FORECASTING

The WRF model was used to create weather forecast – Figure 6. The base of this experiment were GFS outputs (Global Forecast System) created by NCEP – National Weather Service. GFS is global circulation model and runs four times per day (00, 06, 12 and 18 UTC). The resolution of GFS model depends on the time period of forecasting and takes values from 55 km to 110 km.

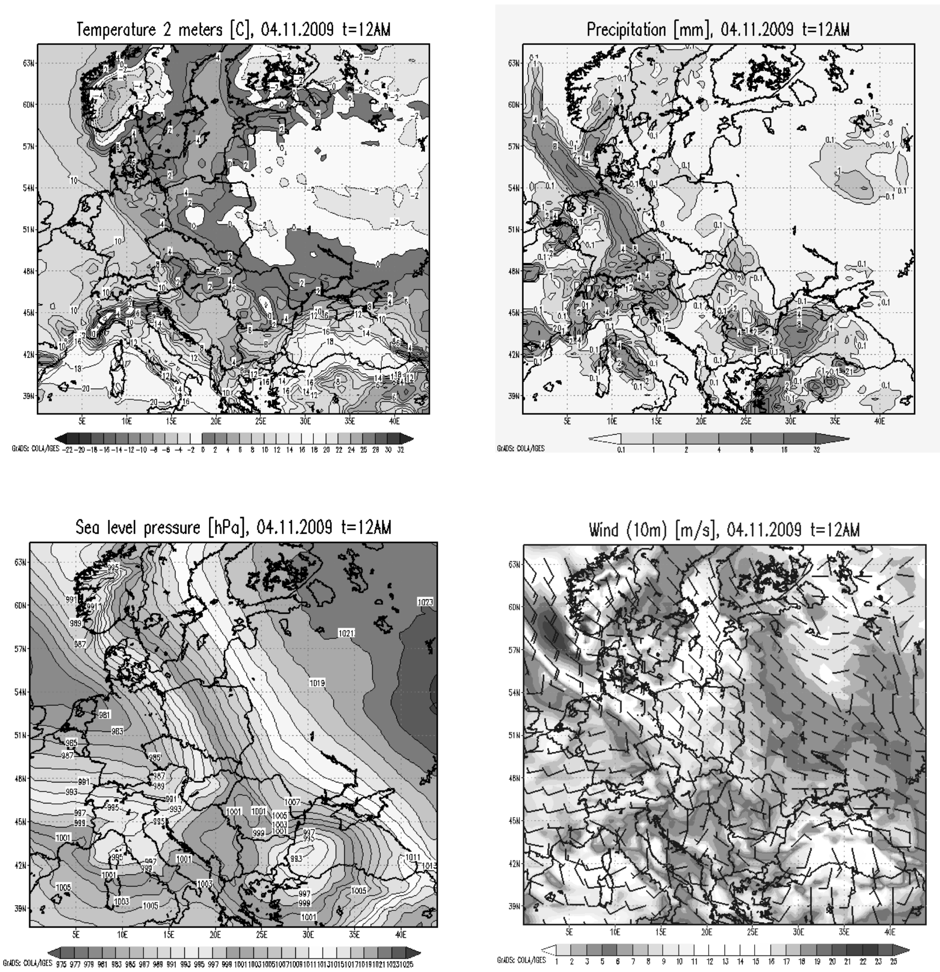

Fig. 6. Exemplary, modeled meteorological fields (air temperature, total precipitation, sea level pressure, wind speed/direction) using WRF model

After installation process the WRF model is ready to assimilate the GFS data. That's why it is necessary to download the GFS data from the NCEP server before initiation of forecasting.

The 7 days weather forecasts for central Europe are prepared at Department Meteorology and Climatology University of Łódź. The resolution of the domain is 30 km. The forecast of four chosen meteorological parameters: air temperature at 2 meters, precipitation, sea level pressure and wind speed at 10 meters. The 7 days weather forecast is actualized every 3 days and the results are presented at the Department Meteorology and Climatology's website: http://nargeo.geo.uni.lodz.pl/~meteo/stronki/ forecast.html.

LONG TERM SIMULATIONS – ASSIMILATION OF ECHAM5 MODEL

ECHAM5 model is the 5th generation of the ECHAM general circulation model developed at the Max Planck Institute for Meteorology in Hamburg. This GCM is widely used for climate simulations, in particular for the Fourth Assessment Report of the Intergovernmental Panel on Climate Change (IPCC-AR4).

In this part of the study it is explained how to use WRF to produce dynamical downscaling simulations from ECHAM5 output.

The files in Table 1 contain 3D fields at different levels and 2D fields for model variables, together with constant 2D fields – Table 1. Among 3D variables are: the geopotential height (GPH), temperature (STP), relative humidity (RELHUM) and the two components of the horizontal wind (U,V) defined at 1000, 925, 850, 775, 600, 500, 400, 300, 250, 200, 150, 100, 70, 50, 30 and 10 mbar. 2D variables include the above variables defined close to the surface: 2 meters temperature (TEMP2), 2 meters relative humidity (or dew point, DEW2, not used at the moment), 10 meters wind components (U10M, V10M). Other 2D variables are the surface pressure (APS), the mean sea level pressure (MSLP), surface temperature (TSURF), sea surface temperature (TSW, not used), soil temperature at 5 levels (TSOIL1, TSOIL2, TSOIL3, TSOIL4 and TSOIL5), soil-water flag (WS), and seas-ice flag (SEAICE). Constant fields are the orography or the geopotential at the surface (GEOSP), the land-sea flag (SLF) and the FAO land use categories (FAO). ECHAM5 outputs are stored in the Hamburg World Data Center for Climate WEB database: http://cera-www.dkrz.de/WDCC/ui/Index.jsp

Contrary to assimilation of GFS outputs by WRF model, inside which Vtable file for GFS is ready to use, assimilation of ECHAM5 outputs is more difficult. It is necessary to create a new Vtable file especially for ECHAM5 model – Table 2.

Table 1. The list of necessary files to start simulation. EH5_OM_A2_1 – name of ECHAM5 experiment, APS, DEV2, GPH, MSLP, RELHUM, SEAICE, STP, TEMP2, TSOIL, U, V, WS, FAO, GEOSP, SLF – names of metrological fields, 1-1460 – time of ECHAM5 experiment in 6 hours steps for 2001 year

EH5_OM_A2_1_APS_1-1460.grb

EH5_OM_A2_1_DEW2_1-1460.grb

EH5_OM_A2_1_GPH10_1-1460.grb (GPH30, 50 ,70, 100, 150, 200, 250, 300, 400, 500, 600, 700, 775, 850, 925, 1000)

EH5_OM_A2_1_MSLP_1-1460.grb

EH5_OM_A2_1_RELHUM10_1-1460.grb (RELHUM30, 50 ,70, 100, 150, 200, 250, 300, 400, 500, 600, 700, 775, 850, 925, 1000)

EH5_OM_A2_1_SEAICE_1-1460.grb

EH5_OM_A2_1_STP10_1-1460.grb (STP30, 50 ,70, 100, 150, 200, 250, 300, 400, 500, 600, 700, 775, 850, 925, 1000)

EH5_OM_A2_1_TEMP2_1-1460.grb

EH5_OM_A2_1_TSOIL_1_1-1460.grb

EH5_OM_A2_1_TSOIL_2_1-1460.grb

EH5_OM_A2_1_TSOIL_3_1-1460.grb

EH5_OM_A2_1_TSOIL_4_1-1460.grb

EH5_OM_A2_1_TSOIL_5_1-1460.grb

EH5_OM_A2_1_TSURF_1-1460.grb

EH5_OM_A2_1_U10_1-1460.grb (U30, 50 ,70, 100, 150, 200, 250, 300, 400, 500, 600, 700, 775, 850, 925, 1000)

EH5_OM_A2_1_V10_1-1460.grb (V30, 50 ,70, 100, 150, 200, 250, 300, 400, 500, 600, 700, 775, 850, 925, 1000)

EH5_OM_A2_1_WS_1-1460.grb

EH5_OM_CONST_FAO_1-1.grb

EH5_OM_CONST_GEOSP_1-1.grb

EH5_OM_CONST_SLF_1-1.grb

In this Vtable file there is some inconsistency that derive from ECHAM5 output that has to be fixed by modifying WRF preprocessing code. That's why it is necessary to add some improvements into WPS code. One of the most important improvements is connected with the orography geopotential (or geopotential height). It is read by WRF at every time step, while in ECHAM5 it is defined only once for a generic date, since it is a constant field. To fix this the WRF preprocessor has to read the geopotential file as many times as necessary and define the field date accordingly. Moreover WRF preprocessing use soil geopotential height while in ECHAM5 output only soil geopotential is provided. To fix this two lines were inserted in the Vtable file, one for the variable SOILGEOP, that corresponds to the actual ECHAM5 variable, and a line for SOILHGP with a dummy parameter code (000), that is not read from file and corresponds to the variable name inside the WRF pre-processor. The last improvement is connected with soil moisture fields. The soil moisture fields are not initialized from ECHAM5 output. Due to problems in the interpretation of ECHAM5 soil moisture field the initial value of soil moisture is prescribed arbitrarily (as it is usually done in regional modeling).

At the end the WRF preprocessing can be used with exactly the same procedure as for the normal use.

Table 2. Exemplary Vtable file to assimilate ECHAM5 outputs

GRIB1 Param	Level Type	From Level1	To Level2	metgrid Name	metgrid Units	metgrid Description	GRIB2 Discp	GRIB2 Catgy	GRIB2 Param	GRIB2 Level
130	100	*		TT	K	Temperature	0	0	0	100
131	100	*		UU	m s-1	U	0	2	2	100
132	100	*		VV	m s-1	V	0	2	3	100
157	100	*		RH	%	Relative Humidity	0	1	1	100
156	100	*		HGT	m	Height	0	3	5	100
167	1	0		TT	K	Temperature at 2 m	0	0	0	103
165	1	0		UU	m s-1	U at 10 m	0	2	2	103
166	1	0		VV	m s-1	V at 10 m	0	2	2	103
168	1	0		DEWPT	K	Dew point temp	0	2	2	103
134	1	0		PSFC	Pa	Surface Pressure	0	3	0	1
151	1	0		PMSL	Pa	Sea-level Pressure	0	3	0	1
140	1	0		SM010200	kg m-3	Soil Moist 10-200 cm below gr layer	2	0	13	1
144	112	0	10	SM000010	kg m-3	Soil Moist 0-10 cm below gr layer (Up)	2	0	192	106
207	111	3		ST000010	K	T 0-10 cm below ground layer (Upper)	0	0	0	106
207	111	19		ST010040	K	T 10-40 cm below ground layer (Upper)	0	0	0	106
207	111	78		ST040100	K	T 40-100 cm below ground layer (Upper)	0	0	0	106
207	111	268		ST100200	K	T 100-200 cm below ground layer (Bottom)	0	0	0	106
210	1	0		SEAICE	proprtn	Ice flag	10	2	0	1
194	1	0		LANDSEA	proprtn	Land/Sea flag (1=land, 0 or 2=sea)	2	0	0	1
129	1	0		SOILGEOP	m	Terrain field of source analysis	2	0	7	1
000	1	0		SOILHGT	m	Terrain field of source analysis	2	0	7	1
169	1	0		SKINTEMP	K	Skin temperature (can use for SST also)	0	0	0	1

CONCLUSIONS

1. The WRF model, as an example of regional model, is characterized by very high flexibility. It can be used for weather forecasting and long term simulations. In the first case all simulations were done using GFS outputs, but WRF is also ready to use many other GCM's outputs e.g. NAM 104 and 212 grids, the NAM AWIP format, the NCEP/NCAR Reanalysis archived at NCAR, RUC (pressure level data and hybrid coordinate data), and AFWA's AGRMET land surface model.
2. The most important thing was to adapt the WRF model to Polish climate conditions. Based on experiments carried out using WRF simulation and real data the best set of parameterization schemes was established. As a parameterization of microphysics was taken Kessler's scheme (a warm-rain scheme used commonly in idealized cloud modeling studies), as a parameterizations of short wave radiation was taken Goddard's scheme (two-stream multi-band scheme with ozone from climatology and cloud effects) and Grell-Devenyi's ensemble scheme as a parameterization of advection (multi-closure, multi-parameter, ensemble method with typically 144 sub-grid members). The results of weather forecasting, which are prepared upon established set of physics parameterizations, are published on the website: http://nargeo.geo.uni.lodz.pl/~meteo/stronki/forecast.html.
3. The WRF model was also used to prepare long term simulation. In this case ECHAM5 outputs were used. The list of necessary files to start long term simulation and the way to download it were explained. The WRF model is not ready to use ECHAM5 data, that's why it was necessary to add some improvements into WPS code connected with reading the data from ECHAM5 outputs by preprocessing system. In this case the new Vtable file was craated and the code connected with geopotencial hight and moisture fields reading by WPS was changed..

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10. SUMMARY

The present publication addresses selected issues regarding the research on climate changes observed in Poland.

The first chapter of the work analyses the variability of precipitation, air temperature and insolation on the basis of the data collected during the period of 1966-2005 from 14 meteorological stations located in Wielkopolska region and in the neighboring area. The precipitation was examined regarding its amount and the number of days with the precipitation of minimum 3 and 20 mm, the number of days with frost and minimal temperatures of 25 and 30^{o}C were analyzed too. The dynamics of the elements of climate mentioned above was examined by means of linear regression method adopting two values of statistical significance 0.05 and 0.10. The chapter presents the directions of change in temperature and humidity which lay basis for assumptions about the scale of long and short-term climate warming in the area of Wielkopolska Lowland.

Chapter 2 analyses the types of precipitation and snow cover, considering them as important components of climate system which are vulnerable to climate changes. The data regarding precipitation and temperature during the period of 1958-2008 collected from the meteorological station in Kraków were used for the purpose of the above analysis. The study presents the connection between the snowfall and storm precipitation and the level of air temperature, among others.

The next chapter uses the records of mean daily precipitation during the period of 1954-2003 gathered from 7 stations located in the Baltic coast. The annual variability patterns of precipitation and the occurrences of weak and heavy precipitation were analyzed. Using selected characteristics (annual amplitudes of monthly precipitation, irregularity index, precipitation semi period) the following parameters were determined: the level of pluvial continentalism and the season of maximum concentration of precipitation, as well as the dependence of the atmospheric precipitation on regional advection of air masses.

Chapter 4 was also devoted to precipitation analysis, it used monthly the results of the atmospheric precipitation measurements conducted in the period of 1951-2005 in 6 meteorological stations located in the middle section of the Oder catchment area. The work examined the variability of precipitation and trends of the changes in different time ranges on the basis of the analysis of 30-year moving sequences. The statistical measurements of precipitation were presented (means, variation coefficient, standard deviation and values of quantiles 10% and 90%), as well as the linear functions illustrating the trends in precipitation

changes both during 4 seasons and annually. The results indicate that there is a downward trend in the precipitation but the variability is on the increase; the amount of precipitation and its variability increase in winter, whereas they tend to decrease in spring and autumn.

The amount of precipitation influences the amount of water in the environment, its shortage leads to drought which can be described using various indexes. Two of them were used in the following chapter: one is the standard precipitation index SPI, which was estimated on the basis of precipitation during the warm half-year (IV-IX) as well as particular months, and the other is the standardized index of climatic water balance. The research used meteorological data from Wrocław-Swojec station, gathered during the period of 1964-2006. The obtained results indicate heavy precipitation shortage in the region of Wrocław. The biggest negative climatic water balance was observed in 1992 and in the very year both indexes defined drought as extreme. Statistically confirmed increase of air temperature in the years 1964-2004, as well as the downward trend in precipitation, makes the water shortage even worse. The classification of droughts according to two indexes SPI and SCWB gave almost the same quantitative results only the intensity was differently qualified.

The next two chapters were devoted to the length of snow cover lingering and snowfall. Chapter 6 examined the data gathered in the period of 1951-2008 from 83 meteorological stations in the area of the whole Poland. The number of days with different thickness of snow cover during two seasons was analyzed: October – May and December – March, as well as the frequency of occurrence of snow cover, length of its lingering, and the medium and maximum thickness of the snow cover. Chapter 7 uses the observations from the meteorological station of the Jagiellonian University gathered in the period of 1951-2008. Long-term changes in the dates of occurring and declining of snow precipitation, the length of the snowy period and also the number of days with snow precipitation and the level of precipitation were analyzed. The trends of changes in the whole winter season and individual months from November to March were studied, both for the 58-year research period and for shorter 30-year periods. The work estimated also the dependence of the number of days and the level of snow precipitation on the index of the regional atmospheric circulation and NAO. It is interesting that, despite noticeable warming, the length of the snowy season in Kraków increases by nearly 4 days every 10 years.

A very important scientific task is to determine how the climate changes influence various ecosystems. One of the possible methods of addressing this question are the measurements of the fluxes of mass and energy exchanged between the surface and the atmosphere, particularly the measurements of greenhouse gases such as CO_2 and H_2O. Unit 8 presented in great detail the Eddy Covariance method used to measure the exchange of the gases between a peat bog and the atmosphere; additionally, the water use efficiency index (WUE) was determined. The measurements were conducted in 2004 in Rzecin. The WUE is a good ecosystem index helpful to estimate the amount of water needed for evapotranspiration and assimilation of CO_2 from the atmosphere by a given ecosystem. The presented results show the seasonal variability of this index. In the first quarter of the year, the WUE ranged from 0.8 to about 1.5 g (C-CO_2) kg^{-1}(H_2O) (a gram of carbon in the form of carbon dioxide per a kilogram of water), while during the full vegetation period, it was 2.8 g (C-CO_2) kg^{-1}(H_2O). Surprisingly, it turned out that during the autumn-winter period this coefficient continued to grow, which appears to be specific only for wetland environments.

The aim of the last chapter of the monograph was to show the capabilities of the Weather Research and Forecasting (WRF) model. The model was used to prepare the weather forecast for countries in the central Europe, especially for the territory of Poland. The paper shows the flowchart for the WRF Modeling System Version 2.2 and the short description of the main blocks. It describes the way of assimilation of the WRF model to Polish climate conditions. It shows the way of preparing climate simulations by means of outputs of ECHAM5 model including different scenarios of CO_2 emissions.

Keywords: climate changes, variability of precipitation, air temperature and insolation, type of precipitation and snow cover, components vulnerable to climate change, standard precipitation index SPI, fluxes of mass and energy exchanged between the surface and the atmosphere, eddy covariance method, water use efficiency index, assimilation of WRF Modeling System to Polish climate condition, climate simulation

11. STRESZCZENIE

BADANIA ZMIAN KLIMATU

Niniejsza publikacja zawiera wybrane zagadnienia związane z badaniami nad zmianami klimatycznymi obserwowanymi w Polsce.

W 1 rozdziale pracy analizowano zmienność opadów, temperatury powietrza i usłonecznienia na podstawie danych zebranych z 14 stacji synoptycznych rozmieszczonych w Wielkopolsce i na terenach przyległych za okres 1966-2005. Dla opadów rozpatrywano ich sumy oraz liczbę dni z opadami o wartościach co najmniej 3 i 20 mm, analizowano liczbę dni z przymrozkami i temperaturami co najmniej 25 i 30^{o}C. Dynamikę zmian wymienionych wyżej elementów klimatu badano za pomocą regresji liniowej przyjmując dwa poziomy istotności statystycznej 0,05 i 0,10. W rozdziale prezentowane są kierunki zmian czynników termicznych i wilgotnościowych, na podstawie których można snuć przypuszczenia co do wielkości ocieplenia klimatu w bliższej lub dalszej przyszłości na terenie Niziny Wielkopolskiej.

W rozdziale 2 wykonano analizę rodzajów opadów i pokrywy śnieżnej, uznając je za ważne komponenty systemu klimatycznego, wrażliwe na zmiany klimatu. W tym celu wykorzystano dane dotyczące opadów i temperatury, ze stacji w Krakowie z lat 1958-2008. W opracowaniu pokazano m.in. związek opadów śniegu oraz opadów burzowych z wysokością temperatury powietrza.

W kolejnym rozdziale wykorzystano średnie dobowe sumy opadów z lat 1954-2003 zebrane dla 7 stacji położonych w strefie wybrzeża Bałtyku. Analizowano roczne przebiegi zmienności opadów oraz częstości opadów słabych i silnych. Za pomocą wybranych charakterystyk (roczne amplitudy miesięcznych sum opadów, wskaźnik nierównomierności i półokres opadowy) określono stopień kontynentalizmu pluwialnego i porę maksymalnej koncentracji opadów, a także zależność opadów atmosferycznych od regionalnej adwekcji mas powietrza.

Rozdział 4 również poświęcono analizie opadów, wykorzystano miesięczne wyniki pomiarów sum opadów atmosferycznych z 6 stacji meteorologicznych, położonych w dorzeczu środkowej Odry z lat 1951-2005. W pracy rozpatrywano zmienności opadów i tendencji zmian w różnych skalach czasowych na podstawie analizy 30-letnich ciągów ruchomych. Przedstawiono miary statystyczne opadów (średnie, współczynnik zmienności, odchylenie standardowe i wartości kwantyli 10% i 90%), a także funkcje liniowe opisujące tendencje zmian opadów w 4 sezonach i roku. Z badań wynika, że dla lata oraz całego roku zachodzi tendencja malejąca opadów, ale rośnie ich zmienność, zimą rosną sumy opadów i ich zmienność, a wiosną i jesienią obserwuje się tendencję malejącą opadów i ich zmienności.

Konsekwencją sumy opadów jest ilość wody w środowisku, jej niedobór prowadzi do suszy, którą można opisywać różnymi wskaźnikami. W kolejnym rozdziale wykorzystano dwa wskaźniki: wskaźnik standaryzowanego opadu SPI, który szacowano na podstawie sumy opadów z półrocza ciepłego (IV-IX), jak i z poszczególnych miesięcy oraz standaryzowany wskaźnik klimatycznego bilansu wodnego. Do badań wykorzystano dane meteorologiczne ze stacji Wrocław-Swojec z lat 1964-2006. Uzyskane wyniki wskazują na głęboki niedobór opadów w rejonie Wrocławia. Największy ujemny klimatyczny bilans wodny wystąpił w roku 1992 i w tym właśnie roku oba wskaźniki zdefiniowały suszę jako ekstremalną. Potwierdzony statystycznie wzrost temperatury powietrza w latach 1964-2004 oraz spadkowa tendencja sum opadów, pogłębia deficyt wilgotnościowy. Klasyfikacja susz według dwóch wskaźników SPI i SCWB dawała niemal identyczne wyniki ilościowe, jedynie różnie kwalifikowała ich intensywność.

Kolejne dwa rozdziały poświęcono długości zalegania pokrywy śnieżnej oraz opadom śniegu. W rozdziale 6 wykorzystano dane z lat 1951-2008 zebrane na 83 stacjach meteorologicznych z terenu Polski. Analizowano liczbę dni z pokrywą śnieżną o różnej grubości w dwóch sezonach: październik-maj oraz grudzień-marzec oraz częstość pojawiania się pokrywy śnieżnej, czas jej zalegania, a także średnią i maksymalną grubość. W rozdziale 7 wykorzystano obserwacje ze stacji meteorologicznej Uniwersytetu Jagiellońskiego z lat 1951-2008. Analizowano długookresowe zmiany w datach pojawiania się i zanikania opadów śniegu, długości okresu śnieżnego, liczby dni z opadami śniegu oraz wysokości opadów. Badano trendy zmian w całym sezonie zimowym i w poszczególnych miesiącach od listopada do marca, zarówno w 58-letnim okresie badań, jak i w krótszych okresach 30-letnich. W pracy oceniono także zależność liczby dni oraz wysokości opadów śniegu od wskaźnika regionalnej cyrkulacji atmosferycznej oraz NAO. Interesujące jest, że pomimo zauważalnego ocieplenia, w Krakowie zwiększa się długości trwania sezonu śnieżnego o blisko 4 dni na 10 lat.

Bardzo ważnym zadaniem dla nauki jest określenie jak zmiany klimatyczne wpływają na różne ekosystemy. Jedną z możliwych dróg odpowiedzi na to pytanie są pomiary strumieni energii i masy wymienianych pomiędzy podłożem a atmosferą, a w szczególności pomiary strumieni gazów szklarniowych takich jak CO_2 i H_2O. W rozdziale 8 szczegółowo zaprezentowano metodę kowariancji wirów (Eddy Covariance) do pomiarów wymiany tych gazów między torfowiskiem a atmosferą, dodatkowo też wyznaczono wskaźnik efektywności wykorzystania wody (WUE). Pomiary prowadzono w 2004 roku w Rzecinie. WUE jest dobrym wskaźnikiem ekosystemowym do oceny ilości wody potrzebnej do ewapotranspiracji i asymilacji CO_2 z atmosfery przez ekosystem. Zaprezentowane rezultaty pokazują sezonową zmienność tego czynnika. W pierwszym kwartale roku WUE wynosiło od 0.8 do około 1.5 g $(C\text{-}CO_2)\cdot kg^{-1}(H_2O)$ (gramów czystego węgla w ditlenku węgla na kilogram wody), w okresie rozwiniętej wegetacji było to 2.8 g $(C\text{-}CO_2)\cdot kg^{-1}(H_2O)$. Zaskakujące okazało się że w okresie jesienno zimowym wskaźnik ten kontynuował wzrost co wydaje się specyficzną właściwością obszarów podmokłych.

W ostatnim rozdziale monografii omówiono właściwości Modelu Badania Pogody i Prognozowania (WRF). Model ten został użyty do przygotowania prognoz pogody dla państw Europy środkowej, w szczególności dla terytorium Polski. W rozdziale omówiono schemat funkcjonowania modelu WRF w wersji 2.2 wraz z krótkim opisem poszczególnych jego bloków. Przedstawiono sposób asymilacji modelu do Polskich warunków klimatycznych. Opisano też ścieżkę przygotowania symulacji klimatu z wykorzystaniem wyników modelu ECHAM5 zawierającego różne scenariusze emisji CO_2.

Słowa kluczowe: zmiany klimatu w Polsce, zmienność opadów, temperatury powietrza i usłonecznienia, rodzaje opadów i pokrywy śnieżnej, komponenty wrażliwe na zmiany klimatu, standardowy indeks opadów SPI, strumienie energii i masy wymieniane pomiędzy podłożem a atmosferą, metoda kowariancji wirów, wskaźnik efektywności wykorzystania wody (WUE), asymilacja modelu WRF do polskich warunków klimatycznych, modelowanie warunków klimatycznych