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OBSAH - CONTENTS

HLAVNÍ ČLÁNKY - ARTICLES

Editorial

The Czech Geographical Society established in 1894 in Prague is one of the oldest geographical societies in Central Europe. Already since its very beginning it has endeavoured to propagate geographical findings into a larger geographical and similarly oriented scientific community. These efforts resulted in foundation of its own journal, which has been published without any interruption since 1894 – this year it is already its 113th volume. Increasing requirements on presentation of current geographic research results on international level have been at the origin of the efforts of the direction of the Czech Geographical Society and of the editors of *Geografie – Sborník ČGS* to get the journal included into the Web of Science database. At the same time they have been trying to obtain impact factor for the journal. In connection with related requirements, several leading scientists entered the body of editors of the journal, the quality of the reviewing procedure has increased and the journal has been published on a regular basis. Although it is clear that this is a long-term process, a great stimulation for a further improving of *Geografie – Sborník ČGS* is its enlisting, by Thompson Reuters society, among journals on its Web of Science since the beginning of 2008. The further development of the journal will largely depend on the quality of contributions sent to it by both Czech and foreign authors. This English copy is aimed at problems of physical geography, mainly on climatology, hydrology and geomorphology. It proves that the board of editors welcome contributions of foreign authors and their preference for an increasing part of English written articles. We hope that at time when the European geography has lost, mostly for financial reasons, some of high quality journals as for instance *Petermanns Geographische Mitteilungen, Geografie – Sborník ČGS*, journal of the Czech Geographical Society may become a significant publication forum for geographical and related sciences not only at the Central-European, but at a larger international scale.

Bohumír Janský
Editor in chief

Úvodník

Česká geografická společnost, založená v roce 1894 v Praze jako Česká společnost zeměvědná, patří mezi nejstarší geografické společnosti ve střední Evropě. Již od svého vzniku usilovala o rozšiřování geografických poznatků do širší geografické a příbuzně orientované vědecké komunity. Výsledkem těchto snad bylo založení vlastního časopisu, který od roku 1894 vychází nepřetržitě až do současné doby, v tomto roce tedy jako 113. ročník. Rostoucí nároky na prezentaci výsledků současných geografických výzkumů na mezinárodní úrovni vyvolaly snahu vedení České geografické společnosti a redakční rady časopisu *Geografie – Sborník ČGS* o zařazení do databáze „Web of Science“. Souvisejícím krokem je snaha o přidělení „impakt faktoru“ pro náš časopis. V návaznosti na požadavky s tím spojené byla provedena obměna redakční rady časopisu o významné vědecké osobnosti, bylo zkvalitněno recenzní řízení a zajištěno pravidelné vydávání jednotlivých čísel časopisu. I když je zřejmé, že jde o proces dlouhodobý, významným povzbuzením pro další zkvalitňování *Geografie – Sborníku ČGS* je její zařazení společností Thompson Reuters mezi časopisy sledované na „Web of Science“, a to počínaje prvním číslem roku 2008. Jaký bude další vývoj časopisu, se bude odvíjet zejména od toho, jak kvalitní příspěvky budou do časopisu zasílány jak českými, tak i zahraničními autory. Předložené anglické číslo je orientováno na problematiku fyzické geografie, zejména se zaměřením na klimatologii, hydrologii a geomorfologii. Je dokladem toho, že redakční rada uvítá příspěvky zahraničních autorů a bude preferovat rostoucí podíl anglicky psaných článků. Věříme, že v době, kdy byla evropská geografie ochuzena, a to zpravidla z finančních důvodů, o některé tak kvalitní časopisy jakými byli např. *Petermanns Geographische Mitteilungen, se časopis Geografie – Sborník ČGS* může stát významnou publikační tribunou pro geografické a příbuzné vědy nejen ve středoevropském, ale i v širším mezinárodním měřítku.

Bohumír Janský
šéfredaktor

GEOGRAFIE – SBORNÍK ČESKÉ GEOGRAFICKÉ SPOLEČNOSTI
ROK 2008 ● ČÍSLO 4 ● ROČNÍK 113

HEINZ WANNER, JONATHAN BÜTIKOFER

HOLOCENE BOND CYCLES: REAL OR IMAGINARY?

H. Wanner, J. Bütikofer: *Holocene Bond Cycles: real or imaginary?* – Geografie–Sborník ČGS, 113, 4, pp. 338–350 (2008). – During the Holocene (last 12,000 years) nine cold relapses were observed mainly in the North Atlantic Ocean area and its surroundings. Based on the pioneering studies by Bond et al. (1997, 2001) these events are called Bond Cycles and thought to be the Holocene equivalents of the Pleistocene Dansgaard-Oeschger cycles. The first event was the Younger Dryas (~12,000 BP; Broecker 2006), the last one was the Little Ice Age (AD 1350–1860; Grove 1988). A number of trigger mechanisms is discussed (see Table 1), but a theory for the Bond Cycles does not exist. Based on spectral analyses of both, forcing factors and climatological time series, we argue that one single process did likely not cause the Holocene cooling events. It is conceivable that the early Holocene coolings were triggered by meltwater pulses. However, the late Holocene events (e.g., the Little Ice Age) were rather caused by a combination of different trigger mechanisms. In every case it has to be taken in mind that natural variability was also playing a decisive role.

KEY WORDS: Holocene – Bond Cycles – spectral analysis – triggering processes – Little Ice Age

Introduction

The Holocene epoch, commonly considered as the recent interglacial, has sustained the growth and development of modern society. Nevertheless, the knowledge about global climate variability during this period is surprisingly sparse (Mayewski et al. 2004, Wanner et al. 2008). The Holocene climate can be considered in three main phases or time periods (Nesje and Dahl 1993, Marchal et al. 2002). The first includes the Preboreal and Boreal chronozones and lasted from about 11.6 to 9 kyr BP. The second phase, the Hypsithermal, which includes the relatively warm Atlantic chronozone, covers the period between about 9 and circa 5.7 kyr BP. In earlier papers it is also called "Altithermal" or "Holocene climate optimum". The third phase, including the Subboreal and Subatlantic chronozones, lasted from 5.7 to 0 kyr BP, and is called Neoglacial because it is characterized by several cold relapses with remarkable glacier advances in different areas of the globe. In their pioneering overview Denton and Karlén (1973) showed that the whole Holocene experienced alternating intervals of glacier advances (600–900 yr in duration) and retreats (lasting up to 1750 yr) but, according to their studies, the strongest advances occurred during the Neoglacial around 200–350, 2800 and 5300 cal yr BP (calendar years before present). The authors concluded that the solar activity was a possible trigger for these fluctuations. Observations in the northern (e.g. Dansgaard et al. 1993) and southern mid- to high-latitudes indicate that the climate during the Holocene was rather stationary. In contrast, proxy based reconstructions show that strong

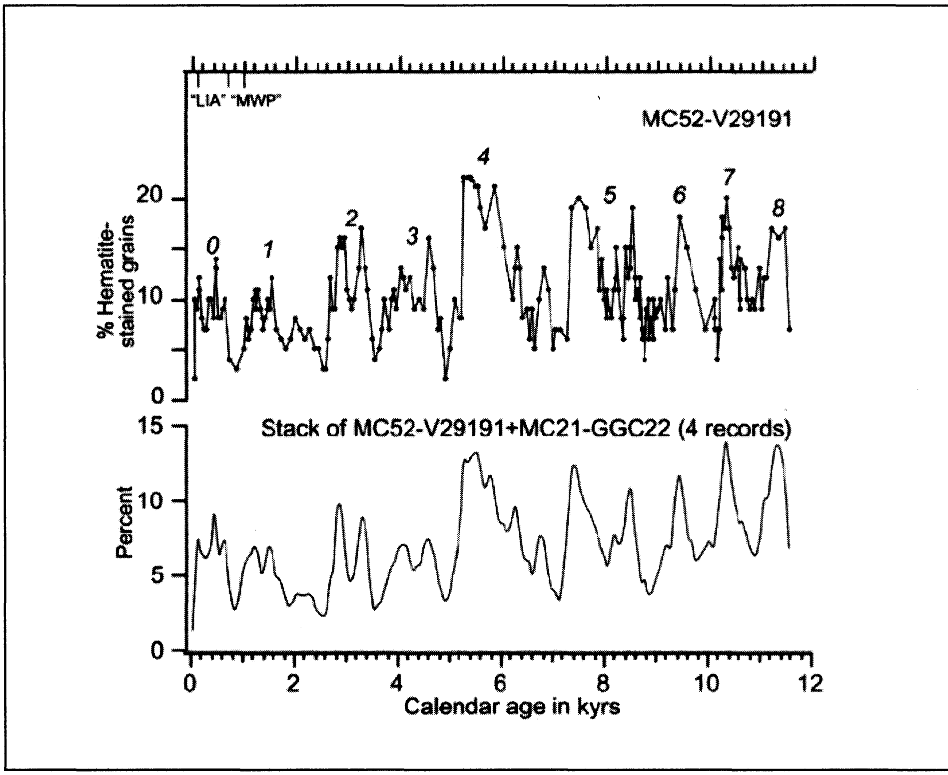

Fig. 1 – Holocene records of drift-ice as percentage variations in petrologic tracers. Upper panel: Hematite stained grains in % in two cores from the same location. Lower panel: Stack of four records at different locations. Bond et al. 1999, modified.

hydrological or circulation changes took place in the tropics and subtropics (Gasse 2000).

A larger progress in the understanding of Holocene climate variability was made by combining specific proxy data and climate models. One of the important insights of the COHMAP (Cooperative Holocene Mapping Project) was that the orbitally induced increase in solar insolation in summer between 12,000 and 6,000 cal years BP enhanced the thermal contrast between land and sea, and thus induced stronger summer monsoons generating a more humid climate in subtropical Africa as well as in west and central Asia (COHMAP Members 1988, Wright et al. 1993). These results were confirmed by the recent simulations of PMIP, the Paleoclimate Modeling Intercomparison Project (Braconnot et al. 2007).

Mainly triggered by the early studies of Denton and Karlén (1973) as well as the debate about the transition from the Medieval Warm Period (MWP) or Epoch to the Little Ice Age (LIA), several studies concentrated on the question whether the centuries or even millennia long climate variations during the Holocene were cyclic or not (O'Brien et al. 1995). The discussion was strongly stimulated by the investigations of Bond et al. (1997, 2001) who, mainly based on petrologic tracers of drift-ice in the North Atlantic (NA), postulated a "1500 year" cycle that is supposed to have persisted throughout the Holocene. These cycles were thought to be the Holocene

Table 1 – Schematic overview on 28 papers discussing the phenomenon of the Bond Cycles (Abbreviations: AO Arctic Oscillation, IOM Indian Ocean Mosoon, IRD Ice Rafted Debris, ISOW Iceland-Scotland Overflow Water, ITCZ Intertropical Convergence Zone, MOC Meridional Overturning Circulation, NADW North Atlantic Deep Water, NAO North Atlantic Oscillation, ODP Ocean Drilling Program, SST Sea Surface Temperature, THC Thermohaline Circulation).

Reference	Location	Data source(s)/parameters	Correlation with bond cycle(s)	Reported spectral peaks/length of cycle(s)	Possible mechanism
O'Brien et al. 1995	Greenland (GISP2)	Sea salt, terrestrial dust from ice core	Yes (with cycles 0, 2, 4, 5, 8)	Quasi 2600-year cycles	Enhanced meridional circulation (atmosphere)
Bianchi and McCave 1999	South Iceland basin	Sortable silt mean size	Yes (with cycles 0–6)	~1500-year periodicity	Changes in deep-water flow
Chapman and Shackleton 2000	North Atlantic off Iceland	Ocean sediment core (lightness, ^{13}C, $CaCO_3$ content)	Yes (with cycles 2, 5, 6, 7)	Several spectral peaks (e.g. 1650, 1000, 550 years)	Cycles possibly driven by variation in solar activity (with consequences for THC fluctuations)
de Menocal et al. 2000	Off Cap Blanc, Mauretania	Ocean sediment core (SST, $\delta^{18}O$, terrigenous (eolian) fraction)	Yes (with cycles 0–5, 7, 8)	About every 1500±500 years	Increased southward advection of cooler waters or enhanced regional upwelling
Jennings et al. 2002	East Greenland Shelf	Ocean sediment cores (magnetic susceptibility, $\delta^{18}O$, IRD, carbonate flux)	Yes (with cycles 1, 2, 3, 5)	–	Increased flux of polar water and sea ice (changing NAO indices?)
Fleitmann et al. 2003	Southern Oman	Stalagmite ($\delta^{18}O$)	Yes (with cycles 2–6)	Several shorter cycles	IOM monsoon precipitation responds to Bond events and to solar activity
Gupta et al. 2003	Arabian Sea	G. bulloides shells (southwest monsoon)	Yes (with cycles 1, 3, 4, 6, 7)	–	Bond events and weak Asian southwest monsoon (related to solar activity?)
Hong et al. 2003	Tibetan Plateau	Peat bog ($\delta^{13}C$ values in cellulose of 12 dominant plants)	Yes (with cycles 1, 3, 5, 7, 8)	–	Thermohaline circulation decreases, SSTs decrease in the North Atlantic and increase in the Indian Ocean, monsoon weakens
Hu et al. 2003	Arolik Lake (south-western Alaska)	Lake sediment (BSi, OC, ON, pollen, $\delta^{18}OSi$, δD_{PA})	Yes (with cycles 3, 4, 6, 7, 8)	Several spectral peaks (e.g. 590, 950 years)	Possible sun-ocean-climate link
Niggemann et al. 2003	Sauerland (Germany)	Calcitic stalagmite ($\delta^{18}O$; indicator for paleohumidity)	Yes (mainly cycles 1 and 2)	1450 years and minor peaks	Lower solar activity was probably accompanied by drier climate in Northern Europe
Oppo et al. 2003	Northeastern Atlantic Ocean	Benthic foraminifera ($\delta^{18}O$, $\delta^{13}C$)	Yes (with cycles 2, 4, 5, 6)	Millennial oscillation of $\delta^{13}C$	Enhanced and reduced NADW contribution
Risebrobakken et al. 2003	Eastern Norwegian Sea	Ocean sediment cores (^{14}C, $\delta^{18}O$)	Yes (with cycles 0, 2, 4, 6, 7)	No evidence for a clear cyclic behaviour, weak evidence for an 81-year cycle	Stronger westerlies during early to mid-Holocene gave rise to an eastward migration of subsurface Arctic water, no response to solar activity
Yu et al. 2003	Continental western Canada	Fen peat record (ash-free bulk density; indicates moisture)	Yes (with cycles 0–5)	Broad band (mean: 1785 years)	Wet periods correlate with warm periods in the North Atlantic (response to solar activity?)
Hall et al. 2004	Subpolar North Atlantic	Two ocean sediment cores (silt mean size, $\delta^{13}C$)	Yes (with cycles 0–3, 5, 6)	Broad band of quasiperiodic variability	Link between the ISOW and the surface component of the MOC
Baker et al. 2005	Lake Titicaca (South America)	Sedimentary organic matter ($\delta^{13}C$; humidity and lake level)	Yes (with cycles 1, 2, 4, 5)	Several spectral peaks (e.g. 434–454 years)	Wet conditions on the Altiplano appear to be associated with cold conditions (Bond events) in the North Atlantic
Russell and Johnson 2005	Lake Edward (Uganda/ Congo)	Sediment cores (% Mg, Mg/Ca)	Yes (with cycles 0, 1)	Enhanced power at ~725 years	Droughts in equatorial Africa occur during both, cold and warm events in the North Atlantic area, due to N-/S-displacement of the ITCZ
Gupta et al. 2005	North-western Arabian Sea	ODP core (% of G. bulloides)	Yes (with cycles 0–7)	Several significant peaks (e.g. at 1550, 152, 114, 83 years)	Monsoon minima coincide with low sunspot numbers and increased advection of drift ice in the North Atlantic
Turney et al. 2005	Northern Ireland	> 750 oak tree ring chronologies from bog and lake sites	Yes (with cycles 0, 2, 4, 5)	Broad band of variability, dominant cyclicity of ~800 years	Trees collapse due to dryness during Bond events, but no responses to changes in solar activity

Reference	Location	Data source(s)/parameters	Correlation with bond cycle(s)	Reported spectral peaks/length of cycle(s)	Possible mechanism
Wang et al. 2005	Dongge Cave (southern China)	Stalagmite ($\delta^{18}O$: monsoon precipitation)	Yes (with cycles 0–5)	1 to 5 centuries long, peaks at 159 and 206 years	Correlation with ice rafted debris in North Atlantic (influence of solar activity?)
Willard et al. 2005	Chesapeake Bay (E North America)	Pollen from sediment core (winter temperature)	Yes (mainly cycle 1 and 5)	1429, 282, 177, 148 years	Cold events in the North Atlantic are also effective in E North America
Lamy et al. 2006	Black Sea / Gulf of Aqaba	Three sediment cores inferring hydroclimatic changes (clay layer frequency, $\delta^{18}O$, terrigenous sand accumulation rate)	Partly	Cycles of ~800 and ~500 years	AO/NAO-like atmospheric variability, likely originating from solar output changes, plays a dominating role for hydroclimatic changes in the study area
Moros et al. 2006	Area off north Iceland	Phase analyses of mineralogical quartz and quartz/ plagioclase ratio	Partly	1300 years, 75–80 years	Late Holocene trend in drift ice: Increasing in the cold East Greenland Current, decreasing in warmer North Atlantic Drift
Viau et al. 2006	North America	752 pollen records from terrestrial sediments (July temperature)	Yes (with cycles 0–2, 5)	Periodicity of circa 1150 years	Possibly solar forcing?
Yu, Y. et al. 2006	Hexi Corridor (northwest China)	Lake sediment core, loess-paleosol (MS, $CaCO_3$, GS, SCR)	Yes (with cycles 0, 2, 5, 7, 8)	–	Southward expansion of northern polar vortex (NAO-), El Niño like patterns, strongly weakened Asian summer monsoon
Allen et al. 2007	Finnmark, Norway	Lake sediment cores (^{14}C ages, pollen, temperature and humidity measures, organic content)	Yes (e.g. cycles 0, 1, 5)	Several spectral peaks (e.g. 1808/09, 1639/45, 1064, 767, 348–411, 279–312, 188 years)	Complex interaction between oceanic and atmospheric circulation, solar and tidal variability
Bendle, Rosell-Melé 2007	North Icelandic Shelf	Alkenone indices (SST)	Yes (with cycles 0, 2)	–	No close correlation with Bond Cycles, rather with NAO dynamics
Li et al. 2007	White Lake (NE USA)	Two cores (^{14}C, terrestrial macrofossils)	Yes (with cycles 1–4)	–	Dry intervals correlate with cold periods in the North Atlantic Ocean
Mangini et al. 2007	Spannagel Cave (central European Alps)	Stalagmites ($\delta^{18}O$: winter precipitation)	Yes (with cycles 0, 2, 3, 4, 5)	–	Meteorological conditions in the European Alps respond synchronously to hydrographical changes in the North Atlantic

equivalents of the Pleistocene Dansgaard-Oeschger cycles (Alley 2005). The upper panel of Figure 1 shows the timeseries of hematite-stained grains (HSG) in form of a tied record from the MC-52 and VM29-191 cores from the same location (Bond et al. 2001). The lower panel shows a stack of four records from different locations in the North Atlantic area. The mechanistic explanation for the cyclic behaviour of these time series is that the high peaks with fresh volcanic glass from Iceland or Jan Mayen and HSG's from eastern Greenland are the result of the southward and eastward advection of cold, ice-bearing surface waters from the Nordic and Labrador Seas during cold periods (Bond et al. 1997). Bond et al. (2001) detected totally nine cold events during the Holocene, which were also interpreted as a consequence of a periodic expansion of the cold polar anticyclone and an enhanced meridional circulation (O'Brien et al 1995). The Bond Cycles are numbered from 0 to 8 (Fig. 1). They peaked around 400, 1,400, 2,800, 4,300, 5,900, 8,100, 9,400, 10,300 and 11,100 cal years BP. The shift from the MWP to the cold LIA with high HSG values is clearly visible in Figure 1 (upper panel). In the following we want to answer three questions:

– Where are Bond Cycles postulated in paleoclimate records?

– Do proxy time series show a similar spectral (cyclic) behaviour, and are the cycles synchronous?
– What are the possible processes causing Bond Cycles during the Holocene?

Where are Bond Cycles postulated in paleoclimate records?

The existence of Bond Cycles is reported for many areas of the globe (Bütikofer 2007). Table 1 shows a list of 28 scientific papers which refer to Bond Cycles. The series of papers starts with the publication of O'Brien et al. (1995), which was the main trigger for the deep sea research launched by the late Gerard Bond and his colleagues (Bond et al. 1997, 2001), and ends with several very recent papers. Not only the character of the data sources (e.g. isotopes or organic and anorganic compounds detected in ice cores, lake or ocean sediment cores, stalagmites, records from peat bogs) but also their spatial variety is considerable. With the exception of the Pacific area, Australia and Antarctica, multi-century to millennial scale ("Bond like") climate cycles in proxy data sets of almost all areas of the globe are interpreted to be associated with the Bond Cycle phenomenon. Yet, it is no surprise that the highest number was detected in the North Atlantic Ocean and its surroundings, followed by several studies from Arabia and Asia.

Do proxy time series show a similar spectral (cyclic) behaviour, and are the cycles synchronous?

The original data (Fig. 1) show that the cycles based on the HSG analysis by Bond et al. (1997) were quite regular during the Holocene, with a mean pacing of ~1500 years. Table 1 shows that correlations are reported between the proxy time series and all of the nine Holocene Bond Cycles. The mostly mentioned correlations with Bond Cycles in the 28 papers according to Table 1 are cycle 2 (~2800 years BP) and cycle 5 (~8100 years BP). In order to answer the question about the spectral behaviour and the synchroneity of the observed cycles, Figure 2 shows the significant (90% level) spectral peaks as described in the vast literature (black dots) and as determined by our time series analyses (open circles; Bütikofer 2007). The different dots and circles are plotted against the corresponding latitude of each record. No clear clustering is visible at all. The majority of the analyses is concentrated in the latitudes between the southern subtropics and the north pole. If we concentrate on the spectral peaks around 1500 years we recognize several dots/circles which are concentrated between 35 and 62° N. These are all located in the North Atlantic area.

Figure 3 shows the frequency distribution of all spectral peaks represented in Figure 2. As expected, the number of spectral peaks grows with decreasing timescale because the number of existing long proxy time series is limited. Small, (and non-significant) clusterings of spectral peaks occur at about 200, 500, 900 and 1500 years (dark bars in Fig. 3).

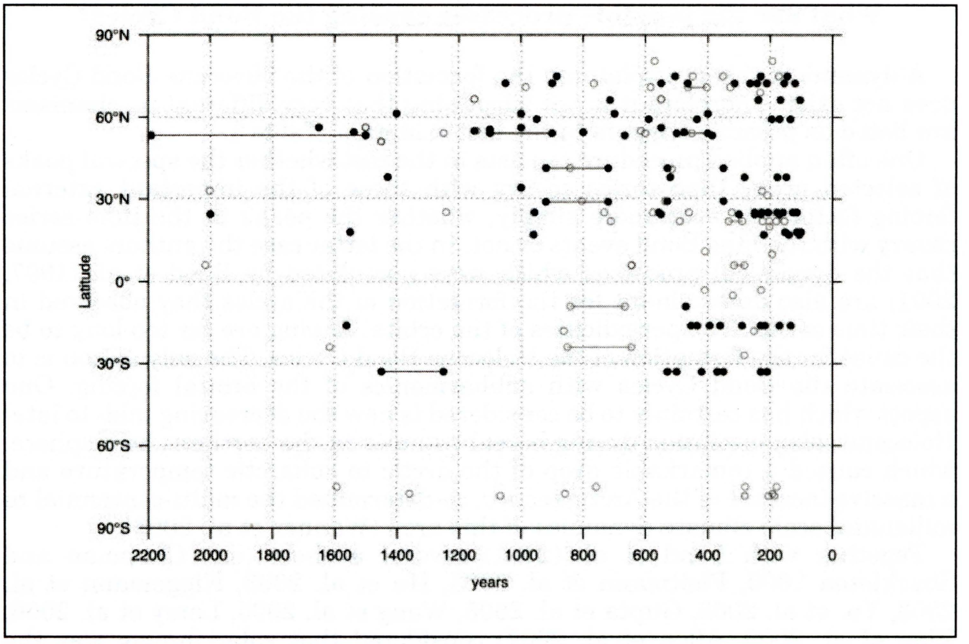

Fig. 2 – Overview of the spectral behaviour of the time series analysed by Bütikofer (2007; open circles) or of a number of referenced spectral peaks in the literature (referenced in Wanner et al. 2008; black dots). Horizontal lines represent broad peaks. The x-axes represents the times scales of the different peaks, the y-axes shows the corresponding latitude of the proxy.

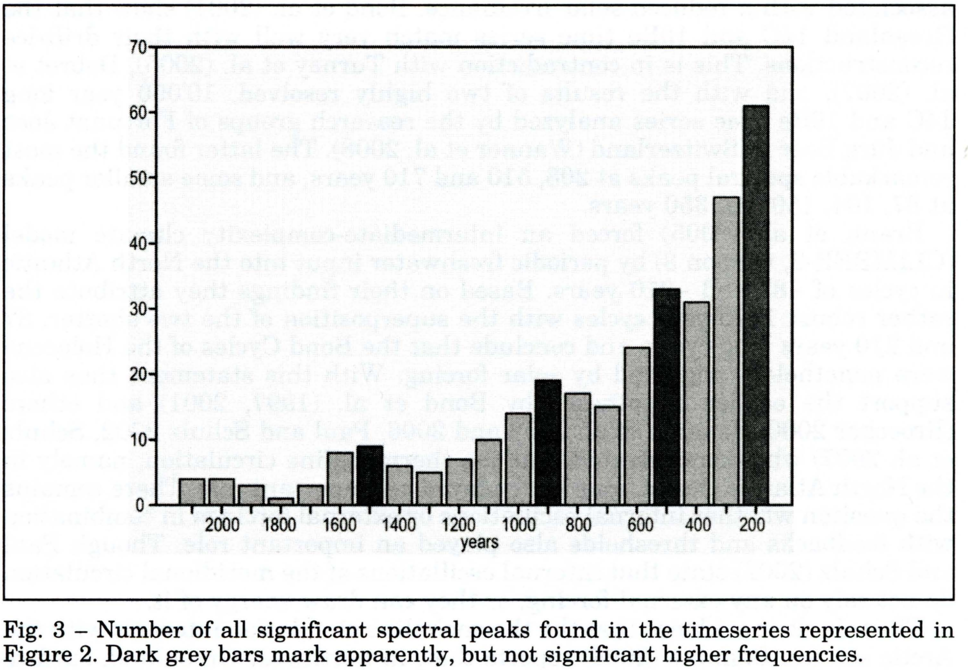

Fig. 3 – Number of all significant spectral peaks found in the timeseries represented in Figure 2. Dark grey bars mark apparently, but not significant higher frequencies.

What are the possible processes causing the Bond Cycles?

A dynamical theory explaining the formation of the Holocene Bond Cycles does not exist. Therefore it is not surprising that very different mechanisms are listed as possible triggers in the last column of Table 1.

One often applied procedure consists in the test whether the spectral peaks of selected proxy time series covary with those of the important external forcing factors or, even more simply, whether the peaks in the time series covary with the nine Bond events or not. In the latter case the authors assume that the dynamical processes which were postulated by Bond et al. (1997, 2001) are also determinant for the formation of the cycles they observed in their time series. The periodicities of the orbital forcing are far too long to be the cause for the formation of the Holocene Bond Cycles. The only option is to associate the Bond Cycles with subharmonics of the orbital forcing. One aspect which has certainly to be considered is how the decreasing mid- to late-Holocene solar insolation during boreal summer in the northern hemisphere, which caused a remarkable drop of the arctic to subarctic temperature and a massive increase of the Arctic sea ice, co-determined the multi-centennial to millennial scale climate dynamics of this area (Wanner et al. 2008).

Together with Bond et al. (2001) several authors (e.g. Chapman and Shackleton 2000, Fleitmann et al. 2003, Hu et al. 2003, Niggemann et al. 2003, Yu. et al. 2003, Gupta et al. 2005, Wang et al. 2005, Lamy et al. 2006, Viau et al. 2006, Allen et al. 2007) considered that solar forcing was the determining or at least one determining factor of their reconstructed multi-century to millennial scale climate fluctuations. Bond et al. (2001) linked their drift-ice records with the production rate of the two nuclides 14C and 10Be in Greenland ice cores. Both nuclides are related to solar wind and solar activity (or solar irradiance) in the sense that higher production rates are associated with a reduced solar irradiance. Bond et al. (2001) show that the Greenland 14C and 10Be time series match very well with their drift-ice reconstructions. This is in contradiction with Turney et al. (2005), Debret et al. (2007), and with the results of two highly resolved, 10'000 year long 14C and 10Be time series analyzed by the research groups of Fortunat Joos and Jürg Beer in Switzerland (Wanner et al. 2008). The latter found the most remarkable spectral peaks at 208, 510 and 710 years, and some smaller peaks at 87, 104, 150 and 350 years.

Braun et al. (2005) forced an intermediate-complexity climate model (CLIMBER-2, version 3) by periodic freshwater input into the North Atlantic in cycles of ~87 and ~210 years. Based on their findings they attribute the rather robust 1470 year cycles with the superposition of the two shorter, 87 and 210 years long cycles and conclude that the Bond Cycles of the Holocene were nonetheless triggered by solar forcing. With this statement they also support the earlier hypotheses by Bond et al. (1997, 2001) and others (Broecker 2000, Renssen et al. 2002 and 2006, Paul and Schulz 2002, Schulz et al. 2007) who demonstrated that the thermohaline circulation, namely in the North Atlantic Ocean, may have played an important role. There remains the question whether internal oscillations or external forcings in combination with feedbacks and thresholds also played an important role. Though Paul and Schulz (2002) state that internal oscillations of the meridional circulation do not rely on any external forcing, or they can draw energy of it.

There is clear evidence that the thermohaline circulation interacts with the Arctic sea ice. Bennike (2004) reports that, in northwest Greenland, surface waters were warmer and less sea ice than at present existed from 7300 to 3700 cal yr BP. That means the sea ice started to grow after this period. In their run with the ECBilt-CLIO-VECODE model Renssen et al. (2006) showed that negative total solar irradiance (TSI) anomalies increase the probability of a local shutdown of deep convection in the Nordic Seas. This initial cooling associated with TSI reductions leads to sea-ice expansion in the area, which stratifies the water column and hampers deepwater formation.

Some authors (e.g. Briffa et al. 1998) argument that large volcanic eruptions can low summer temperatures dramatically. Robock (2000) and Fischer et al. (2006) demonstrated that one single tropical volcanic eruption leads to a climate signal which only lasts for about 2–3 years. Wanner et al. (2008) point to the fact that a higher number of large tropical volcanic eruptions has occurred during certain intervals of the last millennium, i.e. between AD 1200 and 1350 or around AD 1700 and 1800. During the last millennium these maxima of volcanic activity happened to coincide with both, low orbitally induced insolation in the Northern Hemisphere and an unusual concentration of solar activity minima (Wolf, Spörer, Maunder, Dalton) which likely led to the lowest temperatures in this area since 8000 years.

Other processes such as tidal forcing (Keeling, Whorf 1997; Berger, von Rad 2002), the influence of cosmic rays and the global electric circuit (Tinsley et al. 2007, Kirkby 2007) or changes in the Earth's magnetic field (Courtillot et al. 2007) are also hold responsible for decadal to multi-century long climate fluctuations. Finally, it has to be taken in mind that forcing signals with higher frequencies might also be transformed into low frequency ones by slow reactors such as ice caps, or by internal feedbacks. In addition, some authors even express their doubt about a direct forcing of the decadal to multi-century scale cooling events. Related to the Dansgaard-Oeschger events Ditlevsen et al. (2007) vote that a detection of periodicity relies strongly on the accuracy of the dating of these events, and Wunsch (2000) states that a broad band of quasiperiodic variability rather than any kind of significant spectral peak is typical for climate records.

Conclusions

We have demonstrated that evidence for the existence of nine ~1500 years long cold or cool events during the Holocene, called Bond Cycles, exists at least in the North Atlantic area and its surroundings (Fig. 1) where such cycles were described by different authors (Table 1). A theory explaining the formation of these cycles does not exist. This fact is reflected in the long list of possible mechanisms which can trigger Bond events (Table 1). Therefore, the scepticism of different authors voting that one single factor may not be responsible for the cold relapses during the Holocene (e.g., Schulz and Paul 2002, Moros et al. 2006, Ditlevsen et al. 2007) is justyfiable. Their reservations are further supported by the fact that the natural (external) forcing factors changed remarkably during the Holocene (Wanner et al. 2008). First of all, the orbitally induced summer insolation decreased strongly between 10 kyr BP and the present. This led to decreasing temperatures in the North Atlantic Ocean as well as in the northern continental areas of the Northern Hemisphere which favoured the growth of the Arctic sea ice and the strengthening of the cold polar surface anticyclone. These processes were possibly amplified during periods with a low solar activity and an

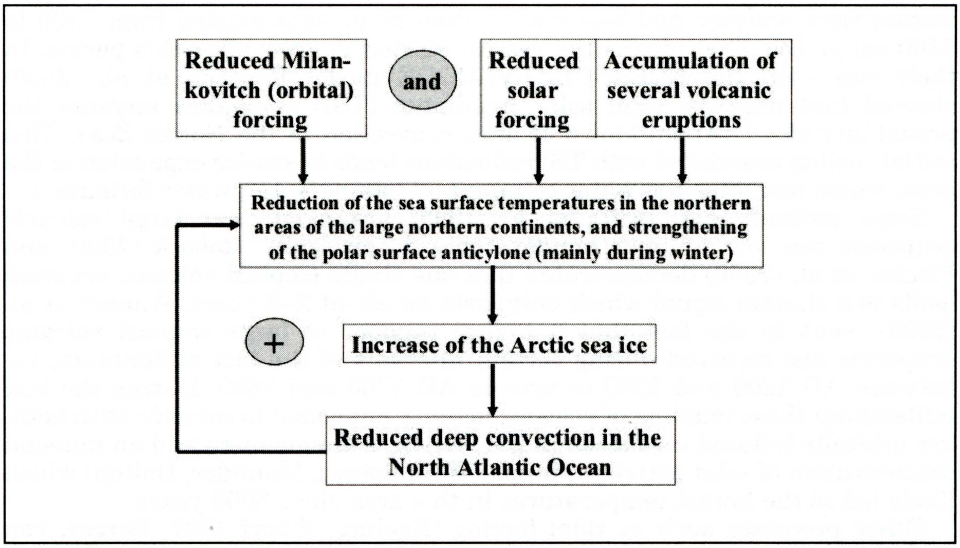

Fig. 4 – Possible mechanism for the formation of a cold event during the late Holocene (e.g., the Little Ice Age) when the summer orbtital forcing in the northern hemisphere coincided with a low solar activity and accumulations of strong volcanic.

accumulation of strong, climate-relevant volcanic eruptions. This was likely the case during the period of the Neoglacial (Denton and Karlén 1973), above all during the Little Ice Age (Wanner et al. 2008).

In form of a simplified scheme Figure 4 tries to represent a possible mechanism for the formation of a cold Holocene event. If we assume that the summer orbital insolation in the Northern Hemisphere was much stronger in the early Holocene (Wanner et al. 2008) and that the Arctic sea ice was very sparse (Bennike 2004), it is indicated that cold relapses such as the 8.2 kyr event were likely triggered by strong meltwater pulses (Barber et al. 1999, Renssen et al. 2002). The late Holocene cold events (Bond events 0, 1 and 2 in Fig. 1) were rather caused by mechanisms which are shown in Figure 4. We are fully aware that this conclusion is still speculative and must be investigated with further data analyses as well as suitable model runs.

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Shrnutí

BONDOVY HOLOCÉNNÍ CYKLY: JSOU SKUTEČNÉ NEBO FIKTIVNÍ?

Na základě analýzy petrologických indikátorů z ledových ker v severním Atlantiku dokázali Bond et al. (1997, 2001), že během holocénu, tedy během posledních 12 000 let, došlo k devíti chladným obdobím. Autoři předpokládají, že tyto jevy jsou holocenními ekvivalenty pleistocénních Dansgaard-Oeschgerových cyklů a že se objevují s určitou pravidelností s délkou cyklu cca 1 500 let. Na základě literatury (tab. 1), jakož i na základě rozšířené spektrální analýzy různých proxy řad jsme ukázali, že existence takovýchto cyklů se předpokládá v mnoha oblastech světa, ale že neexistují jasné důkazy o cyklech trvajících 1 500 let. Jednotliví autoři tedy předpokládají řadu různých spouštěcích mechanismů (např. minimální sluneční aktivita, shluky silných tropických sopečných erupcí, zpětné vazby mezi termohalinní cirkulací a dynamikou arktického mořského ledu).

Domníváme se, že Bondův cyklus byl pozorován zejména v severním Atlantiku a v přilehlých oblastech (východ Spojených států, severní Afrika, Evropa a části Asie). Protože se hraniční podmínky (jako je rozsah kontinentálního zalednění, rozložení mořského ledu, množství tepla v oceánech, rozložení vegetace) během holocénu výrazně změnily, zastáváme na základě dostupných modelových studií hypotézu, že vznik těchto chladných období byl způsoben různými procesy. Soubor jevů ranného holocénu, jako mladší dryas či 8 200 let trvající chladné období, byl vyvolán spíše vodou z tající ledové pokrývky severních kontinentů. Chladná období pozdního holocénu (Bondovy jevy 0, 1 a 2 na obr. 1) byla způsobena spíše součinností různých faktorů, jako omezeného slunečního záření, souběhu několika tropických sopečných výbuchů, posílení studené arktické anticyklony a jejich interakce s dynamikou mořského ledu a termohalinní cirkulací. V průběhu přechodného období před 7 000–4 500 let byly podmínky velice složité, takže jsou potřeba nové údaje a modelové studie. Konečně bychom neměli zapomínat, že důležitou úlohu může hrát vnitřní variabilita systému (tedy kolísání severoatlantské oscilace) a že pro klimatické záznamy je typický spíše široký pás kvaziperiodické variability než jakýkoliv druh významného spektrálního vrcholu.

Obr. 1 – Holocenní záznamy z ledových ker jako percentuální variace petrologických indikátorů. Horní panel: hematitem obalená zrna v % ve dvou jádrech ze stejné lokality. Spodní panel: soubor čtyř záznamů z různých lokalit. Podle Bonda et al. 1999, pozměněno.

Obr. 2 – Přehled spektrálního chování časových řad analyzovaných Bütikoferem (2007; otevřené kroužky) či řady spektrálních vrcholů uváděných v literatuře (odkazy viz Wanner et al. 2008; černé tečky). Vodorovné čáry představují široké vrcholy. Osa x vyjadřuje časové měřítko různých vrcholů, osa y ukazuje příslušnou zeměpisnou šířku nepřímých (proxy) dat.

Obr. 3 – Množství všech významných spektrálních vrcholů v časových řadách znázorněných na obrázku 2. Tmavě šedé sloupky označují zjevné, avšak nikoliv významně vyšší frekvence.

Obr. 4 – Možný mechanismus vzniku studeného období v pozdním holocénu (např. malé doby ledové), kdy došlo k souběhu letního orbitálního působení na severní polokouli s nízkou sluneční aktivitou a řadou silných vulkanických erupcí. V obrázku shora: redukovaný Milankovič (orbitální záření), redukované sluneční záření, souběh několika sopečných výbuchů; snížení teploty mořského povrchu v severních oblastech velkých severních kontinentů a posílení polární anticyklony (zejména v zimě); nárůst arktického mořského ledu; omezená hloubková konvekce v severní části Atlantského oceánu.

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RUDOLF BRÁZDIL, PAVEL ZAHRADNÍČEK, PETR DOBROVOLNÝ, OLDŘICH KOTYZA, HUBERT VALÁŠEK

HISTORICAL AND RECENT VITICULTURE AS A SOURCE OF CLIMATOLOGICAL KNOWLEDGE IN THE CZECH REPUBLIC

R. Brázdil, P. Zahradníček, P. Dobrovolný, O. Kotyza, H. Valášek: *Historical and recent viticulture as a source of climatological knowledge in the Czech Republic.* – Geografie–Sborník ČGS, 113, 4, pp. 351–371 (2008). – The cultivation of the vine (*Vinis vinifera*) that yields grapes for wine manufacture is strongly influenced by the weather. This relationship enables the use of historical viticultural data (e.g., the start date of the grape harvest, notes on wine quality and quantity) for the reconstruction of temperatures and weather extremes in past times. This paper summarises the basics of the relationship between viticulture and climate in the Czech Lands. We compile historical observations before AD 1500 and for the 16th–18th centuries from various types of documentary evidence. The starting dates of the grape harvest in Znojmo for 1800–1890 are used for the reconstruction of April–August temperatures in Brno. The quality of the wine from Bzenec (1800–1890), Znojmo (1802–1845) and Bohutice (1861–1912) is analysed with respect to temperatures corresponding to excellent, good, average and bad wine. Times of flowering and grape harvest are compared with temperatures at the Velké Pavlovice station for the period 1956–2007 and 1984–2007, for various grape varieties.
KEY WORDS: vine – viticulture – vintage – grape harvest – wine quality – wine quantity – temperature reconstruction – weather extremes

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1. Introduction

Cultivation of the vine Vinis vinifera in Europe, which originally came from latitudes 25–40° N, was extended to the south and north as a secondary enterprise. Viticulture in the Czech Republic is strongly influenced by the fact that it is at the northernmost extent of its range in Europe (Košťál 1958). In such conditions, its dependence on weather patterns increases enormously (Kraus 1964, 1999). The nature of the impact of the weather is closely bound to the times of year in which the weather affects the vines' stages of development such as growth, flowering and fruiting (e.g. Winkler et al. 1974, Mullins 1992). Predominantly abundant sunshine and higher air temperatures, together with sufficient precipitation, are factors positively influencing yield and quality of grapes. On the other hand, cold and rainy weather in the period of maturation, extreme winter frosts, late spring and early autumn frosts and hailstorms appear as negative factors. The occurrence of various vine diseases and pests also depends to some extent upon meteorological patterns.

Such close association of vine cultivation with the weather means that information related to viticulture may supply direct or indirect knowledge about the weather and its representation. However, the difference between strictly natural information (such as flowering date and wine quality) and culturally influenced information (such as harvest date and wine quantity) has to be taken in account. This knowledge finds application in historical climatology, in which, by means of the analysis of various types of documentary evidence, the climate and the occurrence of hydrometeorological extremes may be reconstructed for the period prior to systematic instrumental meteorological observation (Brázdil et al. 2005a).

Among European papers dealing with documentary data on vine cultivation, the study made by Le Roy Ladurie and Baulant (1980) is worthy of particular mention. They collected 103 series of vintage data from eastern and central France, western Switzerland and south-west Germany for the period 1484–1879 that could be used for temperature reconstruction. Based on this data, Burkhardt and Hense (1985) reconstructed April–July temperatures for Basle in 1484–1768. Many other papers related to the climatological use of viticultural data have originated in Switzerland (Pfister 1981, 1988; Meier et al. 2007), Austria (Lauscher 1983, 1985; Strömmer 2003) and Germany (Lauer, Frankenberg 1986; Glaser 1991). Střeštík and Verő (2000) used measurements of the lengths of grapevine sprouts from Kőzseg (Hungary) for the reconstruction of spring temperatures after AD 1740. Data from north-eastern France and Switzerland have been used to check the paleoclimatological reconstruction of the North Atlantic Oscillation Index (Souriau, Yiou 2001). Chuine et al. (2004) published a reconstruction of April–August temperatures in Burgundy for AD 1370–2003 using a process-based phenology model developed for the Pinot Noir grape (see also Le Roy Ladurie 2004, 2006, 2007).

The vintage series published by Le Roy Ladurie and Baulant (1980) were used by Menzel (2005) to demonstrate the extremity of the 2003 heat-wave in Europe in the context of the past 500 years. April–August temperatures account for 84% of the year-by-year variability in the dates for grape harvesting in Western Europe. However, Keenan (2007), basing his analysis on a paper by Chuine et al. (2004) and concentrating on the statistical point of view, argued, in terms of the 2003 patterns, that "grape-derived temperature estimates are highly unreliable".

Climatological use of information on vine cultivation in the Czech Republic has not attracted any particular interest to date. For example, a summary paper by Frolec et al. (1973) devoted to viticulture in the Czech Republic paid only little attention to the topic of climate. The most important paper is by Pejml (1974). Using data for the area of Velké Žernoseky (No. 31 in Fig. 1) from 1816 to 1896, he looked for relationships between temperatures in the vegetation period and yields of grapes and/or quality of wine. In the years with good or excellent wine, the probability of above-mean temperatures in the vegetation period achieved 73 %, while in the years with low quality the probability of below-mean temperatures was 77%. A weaker relationship emerged between temperatures and yield. More recently, Brázdil and Valášek (2005) summarised data on vine cultivation in the Czech Republic in terms of the documentary evidence that may be used for historical climatology.

The current article is an attempt to make an inventory of recent knowledge concerning vine cultivation in the Czech Republic with respect to the study of climate variability over a period of several centuries. After a description of basic climate-related wine data and the sources from which they can be extracted, information related to viticulture is analyzed for individual centuries. Series of harvest dates and wine quality as temperature proxies are

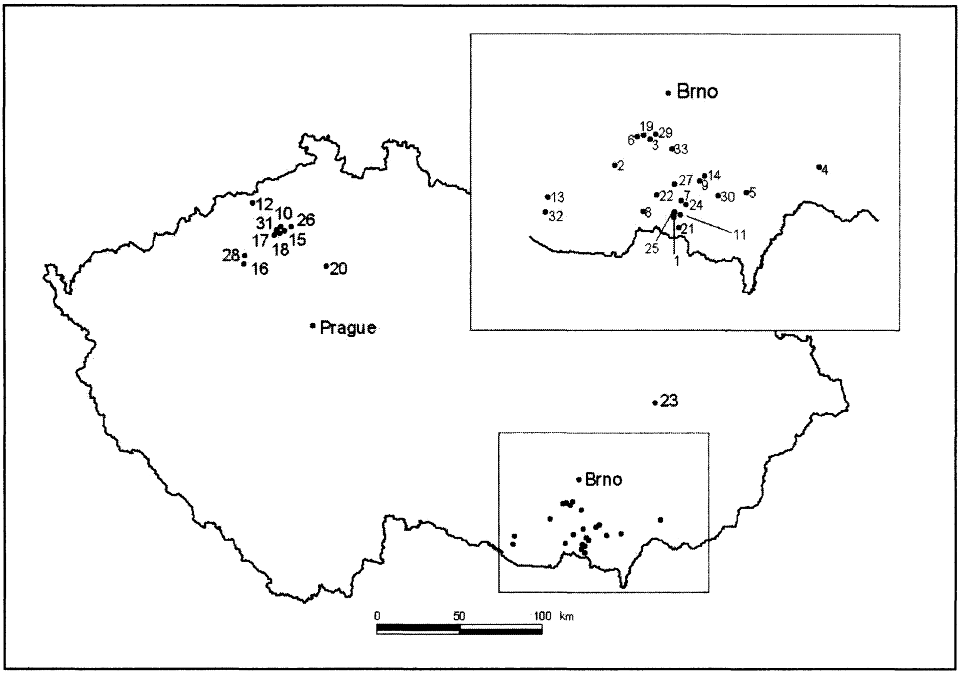

Fig. 1 – Location of places in the Czech Republic mentioned in the text: 1 – Bavory, 2 – Bohutice, 3 – Bratčice, 4 – Bzenec, 5 – Čejkovice, 6 – Dolní Kounice, 7 – Dolní Věstonice, 8 – Drnholec, 9 – Hustopeče, 10 – Kamýk, 11 – Klentnice, 12 – Krupka, 13 – Kuchařovice, 14 – Kurdějov, 15 – Litoměřice, 16 – Louny, 17 – Lovosice, 18 – Malíč, 19 – Mělčany, 20 – Mělník, 21 – Mikulov, 22 – Mušov, 23 – Olomouc, 24 – Pavlov, 25 – Perná, 26 – Ploskovice, 27 – Pouzdřany, 28 – Raná, 29 – Syrovice, 30 – Velké Pavlovice, 31 – Velké Žernoseky, 32 – Znojmo, 33 – Židlochovice.

investigated for the 19th century. A reflection of temperature variability in more recent vine phenophases (1956–2007) is given (phenophases are significant, clearly observable and periodically appearing life stages of plants, such as flowering, ripening etc., which are connected with changes of weather patterns in the course of the year). Finally, the importance of the climate for past viticulture is discussed.

2. Information on vines for the reconstruction of climate and weather extremes in the Czech Republic

The following types of information may be interpreted climatologically:

(i) Start of the grape harvest (vintage): The time at which the grape harvest begins depends on the weather patterns in the months that precede it. Close relations exist between grape harvest and flowering dates (Chuine et al. 2004). Warm, sunny weather contributes to an early start, while colder, rainy weather slows down the ripening of the grapes. The timing of the vintage contains proxy information on temperature patterns in the foregoing period, so systematic records of it may be used for quantitative temperature reconstruction. However, the start of the grape harvest does not depend entirely upon the climate; commercial decisions made by winemakers and wine merchants also play a significant part. For example, they might leave the grapes to ripen further in the vineyards until late autumn in an attempt to enhance the quality of the resulting wine. While in some cases this has led to better grapes (e.g. in 1578 in the Litoměřice area), in other years the whole yield might have been destroyed by early frosts (e.g. in 1579, *Kniha pamětní litoměřických městských písařů*). Furthermore, local traditions (for example, bans on vintage) and parochial festivities play an important role in the long-term scale.

(ii) Quality of wine: The quality of wine is often a reflection of the temperature and humidity patterns that precede the harvest. Grapes with a high sugar content in the must (the grape juice before or during fermentation) correspond with warmer and drier weather at the time of ripening and a "sweet" wine. On the other hand, low sugar content, a consequence of prevailing cold and rainy weather, leads to "dry", more astringent, or even sour wine.

(iii) Quantity of wine: A bad grape harvest and subsequent lower quantity of wine may be related to the occurrence of diseases and pests (e.g. in 1588 considerable quantities of grapes were eaten by flocks of birds (fieldfares) in the Litoměřice area – *Kniha pamětní litoměřických městských písařů*), or to the negative impacts of meteorological extremes (e.g. a hard winter, late spring and early autumn frosts, hail, etc.).

Climatologically utilisable information comes from documentary evidence or phenological observations:

a) Documentary evidence

(i) Chronicles, annals, memoirs: Personal written records made by individuals or on behalf of religious and/or commercial concerns may contain information about the start of the harvest, damage to vineyards, quality, quantity and prices of wine. For example, the *Kniha pamětní litoměřických městských písařů* of Litoměřice includes a great deal of viticultural data from 1570–1607. Similarly, the "Memory Book of Krupka" by master-tanner Michel Stüeler from 1629–1649 (Knott s.a.) cites quantity and quality of wine and damage to vineyards.

(ii) Systematic daily records: Information on vine cultivation is also reflected in the diaries of people who took systematic records that often contained incidental visual weather observations. Baron Peter Forgatsch wrote in his Brno records for 27 May 1812 that frosts in recent days had been very harmful to fruit trees and vineyards. Further, his entry for 5 September states that if the weather does not warm up, the grapes will not ripen. Eventually, there was a lot of rain, and thus a large quantity of grapes (entry for 30 September), but the quality of the wine is not mentioned (Welzl 1910).

(iii) Personal and official correspondence: This often includes information about the yield and quality of the grape harvest. For example, Zdeněk Lev of Rožmitál asked Martin Střítežský of Drast, a burgrave in Ploskovice, to buy wine for him, in a letter from Prague dated 1 March 1526. He had heard that wine was cheap in Litoměřice (Dvorský 1888). This demonstrates a good yield from the previous year's harvest, something that is confirmed by other sources: an abundance of a good wine in 1525 is mentioned by Daniel Adam of Veleslavín (Adam z Veleslavína 1590), a large yield of wine by Chronicon Magistri Georgii Pisensis (1907). Prince Otto, in a letter dated 18 October 1719 from Olomouc to his brother, Cardinal Wolfgang Hannibal, a prince of

Fig. 2 – Symbols of vintage and grape-pressing were used to portray the month of October by Johann Willenberg in the calendar for 1604. The vintage began most frequently in that month in Louny as well at other locations in the Czech Lands.

Schrattenbach, expressed pleasure at a good vintage, excellent wine and another good year for wine in Moravia (Lechner 1896).

(iv) Journalism: Newspapers often contain information about vintage or damage to vines. For example, according to the *Brünner Zeitung* of 30 September 1809, the harvest was already under way in Brno although it usually started on 6 October at the earliest. An above-average yield of grapes and high quality of wine was anticipated. Another report mentions that vines in Brno were blossoming for a second time after 12 September 1822, but that the vintage proper started on 26 September. The harvest began earlier than in the immediately previous years: 1819 – 4 October, 1820 – 17 October, 1821 – 24 October (*Mährisch-Ständische Brünner Zeitung*).

(v) Official financial records: These consist of evidence related to economic activity and the official record-keeping that accompanied it. For example, account books kept for the town of Louny, surviving for various time intervals in the 15th–17th centuries, feature wages paid every Saturday to harvesters (Fig. 2) or to other vineyard workers in the course of the year (e.g. for spreading manure, hoeing and other ground-work; Brázdil, Kotyza 2000). Because these wages reflected work done in the given weeks, interannual fluctuations in the timing of particular activities were consequences of weather patterns in the previous days or weeks. Farmers' requests for tax relief arising out of damage done to vineyards by extremes of weather (mainly hail and frost) for the Dietrichstein demesnes of Dolní Kounice and Mikulov from 1650–1849 are an important source in southern Moravia (Brázdil et al. 2003).

(vi) Technical papers: Climate-related viticultural data may also be extracted from a range of technical papers (e.g. Katzerowsky 1887; Donek et al. 1932). Košťál (1958) published grape yields for Malíč, Kamýk and Lovosice, covering various periods of the 16th–18th centuries, and mentioned the importance of meteorological factors to the value of yields. Such authors often worked with original archive sources, some of which may not have survived to this day. Many such studies were published, for example, in the journal *Vinařský obzor* (Viticultural horizon), founded in 1907.

b) Phenological observations

Phenological observations in the Czech Lands began in the 1780s thanks to the work of Director Antonín Strnad at the Prague-Klementinum observatory and have continued since (Nekovář 2008). Phenological yearbooks started to be published from 1923 under the auspices of the Agriculture Research Institute of Brno and from 1938 under the Central Meteorological Institute (later Hydrometeorological Institute) in Prague (Mihálikova 1983), but vine observations were not included in them. However, in the archives of the Czech Hydrometeorological Institute, viticultural data appear in original handwritten observations preserved from individual phenological stations. Systematic observations may be partly completed by data taken from various other documentary sources (as mentioned above) but their temporal and spatial distribution is discontinuous.

3. Historical information on the vine as a climate indicator

3.1 Climate and viticulture before AD 1500

Information on viticulture in the Czech Lands before AD 1500 is sparse and limited mainly to wine quality and quantity (Brázdil, Kotyza 1995). The first report, from AD 1122, mentions an ample yield of grapes. Further reports refer to damage to vineyards by late spring frosts (1258, 1283, 1323, 1331, June 1424, 1430, 1448, 1485, 1486) and hail (1260), differing qualities of wine (good – 1369, very bad or sour – 1254, 1259, 1335, 1487, 1491) as well as its quantity (good yield, plenty of wine – 1260, 1270, 1319, 1320, 1324, 1442, 1469, 1484, 1499, small quantity or lack of wine – 1256, 1262, 1266, 1310, 1330, 1333, 1335, 1337, 1486, 1487, 1491). In 1420, vineyards were already in blossom on 4 April after a mild winter. For 28 October 1335, only the start of the vintage is mentioned (for quotations of all sources, see Brázdil, Kotyza 1995).

For the 15th century, records of the Louny "Liber rationum" (Vaniš 1979) mention the start of grape harvest for the weeks before 8 November 1451 (the latest of the given dates), 23 October 1452, 29 October 1453, 28 October 1454, 13 October 1455, 1 November 1456, 3 October 1457, 9 October 1458, 22 October 1459, 20 October 1460, 18 October 1462, 17 October 1463, 16 October 1469, 29 October 1470 and 30 September 1471 (the earliest of the given dates).

3.2 Climate and viticulture in the 16th–19th centuries

Decadal frequencies of weather extremes with documented damage to vineyards (late spring and early autumn frosts, hail), of wine quality (excellent, good, average and bad) and wine quantity (abundant, average, low) have been calculated (Fig. 3) separately for the two key viticulture regions of the Czech Republic (the Litoměřice and Louny areas in Bohemia and southern Moravia). Despite incomplete annual evidence on the vines, the data sources described in the previous chapter, and those that follow, were used for this.

3.2.1 The 16th century

The greater part of viticultural data for the 16th century comes from the account books for Louny and its estates (Brázdil, Kotyza 2000) and from the *Kniha pamětní litoměřických městských písařů* for Litoměřice. Only little data is available for Moravia, mainly from the administrator of the Žerotín estate

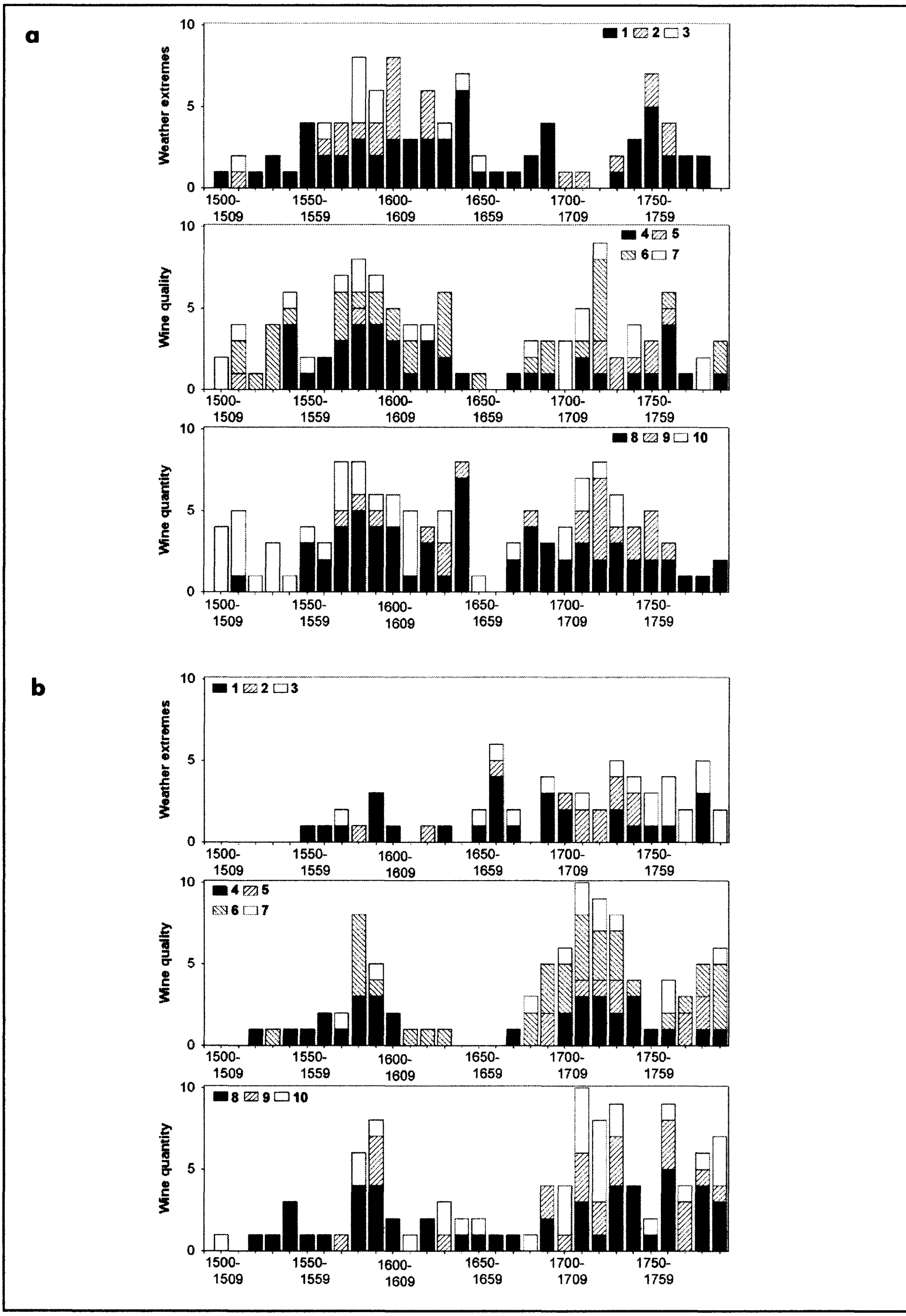

Fig. 3 – Decadal frequencies of weather extremes damaging vines (1 – late spring frost, 2 – early autumn frost, 3 – hail), wine quality (4 – bad, 5 – average, 6 – good, 7 – excellent) and wine quantity (8 – low, 9 – average, 10 – abundant) in viticultural regions of Bohemia (a) and Moravia (b) in the 16th–18th centuries.

in Židlochovice, Matyáš Matuška from Topolčany (Ondrůj 1977). Otherwise, there are many years for which no information related to viticulture is available.

Looking at the frequency of various aspects of viticulture in Bohemia, several decades stand out (Fig. 3a). The start of the grape harvest was recorded for only 19 years of the 16th century. In terms of wine quality, four years with good wine in the 1530s and four years with sour wine in the 1540s, 1580s and 1590s should be mentioned. Small quantities of wine were recorded in the 1580s (for five years) followed by the 1570s and 1590s (four years). The 1500s–1510s and 1530s proved more favourable from this point of view, with abundant harvests and no year of sour wine. Damage by late spring frosts occurred in four years of the 1550s and hail did damage in the 1580s with the same frequency. In southern Moravia, the 1580s–1590s were remarkable for their frequency of low yields (four years each) and sour wine (three years), but also for five years with good wine in the 1580s (Fig. 3b).

The foregoing information about viticulture in the Czech Lands relates well to the known summer climate development in Central Europe during the 16th century, with its cooler and wetter deterioration after 1560 (Pfister, Brázdil 1999). On the other hand, the exceptionally warm and dry year of 1540 was remarkable for an excellent wine, remembered in Central Europe for several centuries to come (Glaser et al. 1999).

3.2.2 The 17th century

Rich viticulture data are available for the first three decades of the 17th century thanks to the Louny account books, particularly their "Registers of Incomes and Expenditures" and "Registers of Raná Farm" (Brázdil, Kotyza 2000). For example, the times at which the harvest started are available for 22 years in the 1600–1632 period. The earliest harvest was recorded on 8 September 1616, after a very warm and dry summer, by Louny chronicler Pavel Mikšovic (the week preceding 24 September on Raná Farm). On the other hand, the latest recorded dates were in the weeks before 15 November 1608 and 16 November 1619 (Brázdil, Kotyza 2000). Increased damage to vineyards by spring and autumn frosts in the 1600s and 1620s is worthy of note (Fig. 3a). While in 1614 the production of beer in Litoměřice came to a halt because of a glut of wine (Donek et al. 1932), Pavel Mikšovic recorded in 1627 that no-one in Louny could remember such sour wine for 30 years (Brázdil, Kotyza 2000).

The following two decades of the 17th century may be characterised from the records of Michel Stüeler of Krupka (Knott s.a.). While the 1630s were characterised by good wine in four years, the 1640s were almost barren. May frosts damaged the vines every year in 1641–1645. Moreover, Stüeler mentions only a small quantity of wine for 1640, 1646–1647 and no wine at all for 1648–1649 (Brázdil et al. 2004) – a total of seven relatively barren years in the 1640s.

Wine quality information in Moravia is absent for the 1640s–1660s, but reports of frost damage appear again in the 1660s (Fig. 3b). Vineyards in many parts of southern Moravia were heavily damaged by frosts on 18 May 1662. On 5–6 September 1664, a heavy frost affected vineyards in the Mikulov area, with less damage to the vineyards of the aristocracy compared to those of their subjects. Frost damage to vineyards around various villages in southern Moravia is further mentioned for 7 May 1666, before 19 May 1667 and again on 2 May 1668. Moreover, hail heavily damaged vineyards in

Mušov and Pouzdřany on 21 July 1664 (Brázdil et al. 2003).

Further data to the end of the 17th century are incomplete to a greater or lesser extent. In the last two decades the number of low grape yields increased in Bohemia (Fig. 3a). Despite some damage to vineyards by spring frosts in the 1690s, three years of good wine were recorded in southern Moravia (Fig. 3b).

3.2.3 The 18th century

For Moravia, information about yields of grapes, quality and prices of wine for 1704–1743 comes from a Jesuit of the Olomouc college, priest for the Čejkovice demesne (Hlavinka 1908). For several years of this period this can be supplemented by the "Memory Book" for Bzenec (Hanák 1922), the *Kronika Hustopečí* (Chronicle of Hustopeče) as well as chronicles relating to several other places in southern Moravia. Viticulture data from Bohemia is mainly derived from the records of Anton Gottfried Schmidt, a Litoměřice town clerk (Katzerowsky 1887).

The 1700s–1730s and 1790s may be considered the most fruitful decades for Moravia; wine both excellent and good was recorded in four–six of their years (Fig. 3b). But every decade in the 1700s–1730s also threw up two or three years with sour wine. A similar situation occurred in Bohemia, where excellent wine occurred once and good wine five times in the 1720s, but only sour wine was available there in four years of the 1760s (Fig. 3a). Wine quantity in Moravia fluctuated between five years with abundant harvests in the 1720s and five years with small harvests in the 1760s. The unpleasant 1760s correspond to records kept by Jan Josef Albrecht, a Mělník scribe, who mentioned bad yields in 1757–1765 (Teplý 1902). In Bohemia the harvest was average in five years of the 1720s.

As in previous centuries, many sources document damage to vineyards. For example, on 15 May 1712, hail damaged vineyards in the villages of Klentnice, Pavlov, Perná and the Drnholec demesne (Brázdil et al. 2003). During the 1750s, spring frost damage to vineyards was recorded in Bohemia for five years and autumn frost damage for two other years (Fig. 3a). On the night of 25/26 May 1796, hail totally or partially destroyed vineyards in Dolní Kounice, Mělčany, Syrovice and Bratčice (Brázdil et al. 2003).

3.2.4 The 19th century

Viticulture in the 19th century is better documented than that of previous centuries. A series of vintage beginnings exists for Znojmo, compiled from two sources:

- the protocols of the town council for the period 1800–1864 (*Vinobranní, vinice a víno*)
- announcements of the beginning of the vintage by the town council in the *Znaimer Wochenblatt* for the period 1865–1890.

Some years are missing from this series: 1801–1803, 1805–1806, 1809–1810, 1851, 1865–1866, 1876 and 1883. This is due in particular to missed harvests in years of crop failure. For example, the yield was negligible in Bzenec and Znojmo in 1805, the wine was sour and poured away in front of the wine cellars. In 1866, the vines were totally frozen by heavy May frosts in several places (Haase 1873, Anonymous 1908). Altogether, the starting dates for harvests are available for a total of 76 years.

Viticultural data from Znojmo were compared with the Brno temperature series, which is a homogeneous long-term series starting in 1799 (Brázdil et al. 2005b, Štěpánek et al. 2006). Brno temperatures are highly correlated with the recent Kuchařovice meteorological station, located close to Znojmo: 0.98 for April and May, and 0.97 from June to September (period 1961–2007). This means that Brno is sufficiently representative for this analysis.

To assess the potential of Znojmo vintage data for climate reconstruction, its relation to temperatures from the Brno series was studied using correlation coefficients between the two variables. The highest correlation was found for April–August (0.57), followed by May–July (0.56). Adding September or August temperatures to previous months does not have any significant influence on changes in the correlations. This is due to the decisive role played by late-spring temperatures, while August and September temperatures strongly influence the sugar content rather than the timing of the grape harvest (Chuine et al. 2004; see also Fig. 6). The strong relationship between the beginning of the grape harvest and April–August temperatures in the Znojmo region is in agreement with a similar study performed for Switzerland (Meier et al. 2007).

For application of the linear regression model, a calibration/verification exercise between Znojmo vintage data (predictor) and Brno temperatures (predictand) was performed, separating the whole period into two parts with 38 years always available (i.e. 1800–1847 and 1848–1890). Then the linear regression model was calculated for the first sub-period and the temperatures obtained were independently verified using data from the second period, and vice versa. The accuracy of the quantitative reconstruction, for both the calibration and the verification periods, was evaluated by squared correlation coefficient r^2 (representing the variance explained by the statistical model), with reduction of error RE and coefficient of efficiency CE (Cook et al. 1994). Using these statistics, the reconstructed temperatures may be compared with those measured (Table 1).

The revealed April–August temperature variance of between 38% and 47% is comparable with that of April–September temperatures (36%) reconstructed from tree rings in northern Bohemia (Brázdil et al. 1997). Although the sub-periods used for the two calibrations/verifications are relatively short, positive values of RE and CE for both of them support the reliability of the linear regression model. Thus for the final reconstruction of April–August temperatures, the whole overlapping period 1800–1890 was used. The suitability of the regression model was further tested with Durbin-Watson statistics (DW). A value of DW = 1.44 for one independent variable and for 38 years does not indicate significant positive autocorrelation in residuals and proves suitability for the linear regression model used. Temperature measurements from Brno have been expressed in the form of anomalies with respect to the 1961–1990 reference period. A 95% confidence interval was also expressed for each reconstructed value (Fig. 4). The unexplained variance of the linear regression model may be attributed to other natural factors influencing vine growth, flowering and fruiting (such as other climate elements, pests, diseases, etc.) and human activities reflected in grape harvesting.

It follows from Figure 4 that there is a good correspondence between measured and reconstructed temperatures, especially from the 1830s to the 1870s. There is also a lack of coherence in the 1880s. The largest differences between measured and reconstructed temperatures are related to the years with the most extreme April–August temperatures with respect to the

Table 1 – Measures of reconstruction skill of Brno April–August temperatures. Vintage data from Znojmo are used as an independent variable in the model: r^2 – squared correlation coefficient, RE – reduction of error, CE – coefficient of efficiency, MSE – mean square error.

Characteristics	1800–1847	1848–1890	1800–1890
r^2	0.382	0.468	0.329
RE	0.320	0.401	0.339
CE	0.351	0.353	–
MSE	0.533	0.307	0.428

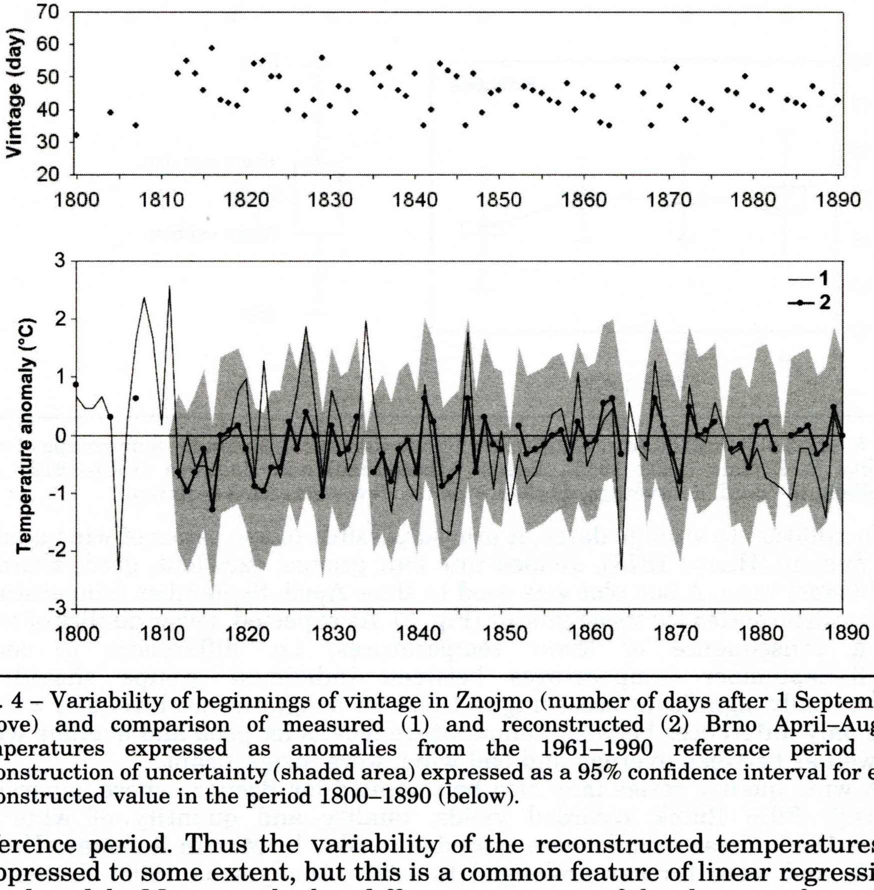

Fig. 4 – Variability of beginnings of vintage in Znojmo (number of days after 1 September) (above) and comparison of measured (1) and reconstructed (2) Brno April–August temperatures expressed as anomalies from the 1961–1990 reference period and reconstruction of uncertainty (shaded area) expressed as a 95% confidence interval for each reconstructed value in the period 1800–1890 (below).

reference period. Thus the variability of the reconstructed temperatures is suppressed to some extent, but this is a common feature of linear regression-based models. Moreover, higher differences in some of the shorter sub-periods or in individual years can be attributed to the fact the beginning of vintage may be influenced by other factors. For instance, the latest beginning of vintage in the series processed is 1816, often referred as "the year without a summer" as a result of the eruption of Tambora in 1815 (Písek, Brázdil 2006). Relatively higher differences appear between measured and reconstructed data in several years. For example, the harvest started significantly later in 1822 and 1827 (very good wine) and earlier in 1864 (bad wine), i.e. there was either a tendency to leave grapes on the vine for longer or harvest earlier (Anonymous 1908).

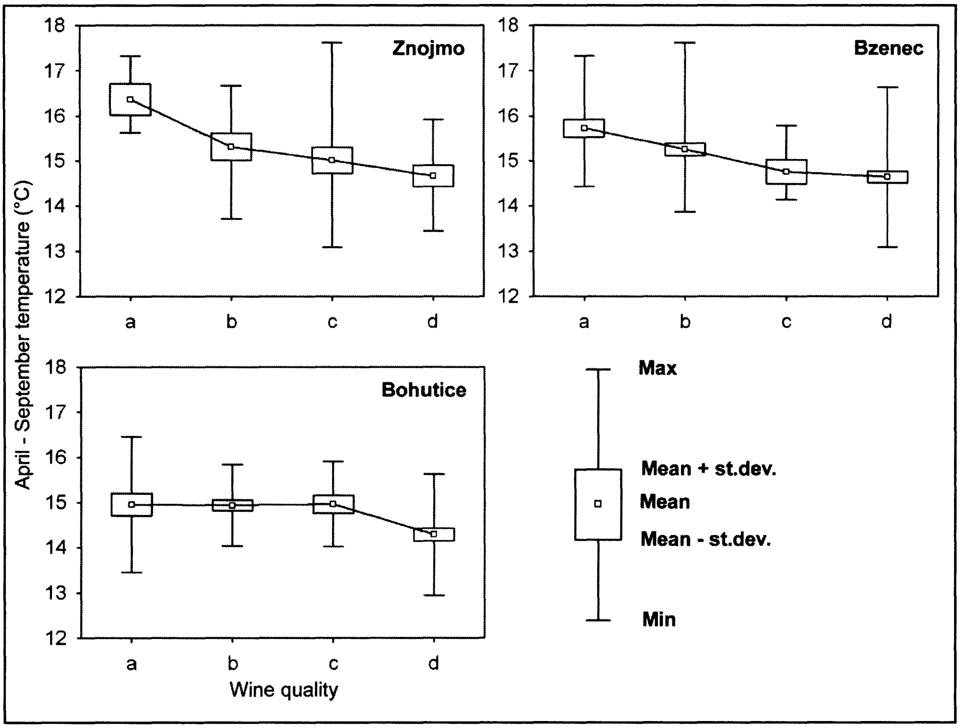

Fig. 5 – Box plot of Brno April–September temperatures calculated for four groups of wine quality in Znojmo (1802–1845), Bzenec (1800–1890) and Bohutice (1861–1912): a – excellent, b – good, c – average, d – bad wine (st. dev. – standard deviation).

In addition to vintage dates, it proved possible to use series of wine quality for Znojmo (Haase 1873), divided into four groups: excellent, good, average, bad (sour) wine. A box plot was used to show April–September temperatures in the Brno series for these groups (Fig. 5). As expected, lower quality of wine is a consequence of lower temperatures, i.e. differences in mean April–September temperatures between individual groups should be statistically significant according to t-test for a significance level of $\alpha = 0.05$. The only differences between temperature means for good and average wine, as well as between average and bad wine, were insignificant.

A wine quality series may also be compiled for Bzenec, where supervisor (horný) Filip Rúbik recorded yields, quality and quantity of wine for 1796–1840. These records were extended to 1849 by Martin Richman (Hanák 1922) and were further prolonged to 1899 (Anonymous 1908). Using the Bzenec data for the period 1800–1890, it was possible to perform the same analysis as for Znojmo (Fig. 5). This shows that differences in mean Brno April–September temperatures were insignificant between neighbour categories (e.g. excellent and good, good and average, average and bad), but significant with step over one category (e.g. excellent and average or bad, good and bad). The low quantity of sour wine in Bzenec for the every year of 1850–1860 is noteworthy (Anonymous 1908).

Another wine quality series occurs in the financial records kept by a farmer Josef Svoboda for 1861–1912 in Bohutice (Anonymous 1913). Surprisingly, there is no difference in mean August–September temperatures to correspond

Table 2 – Correlation coefficients of mean monthly temperatures with vintage of various grape varieties in Velké Pavlovice during the period 1984–2007 (missing years: *Modrý Portugal* – 1985, 1987, 1991; *Veltlínské zelené* – 1987, 1989, 1991; *Frankovka* – 1987, 1988, 1991, 1997). Statistically significant coefficients according to the t-test for the α = 0.05 significance level are shown in bold.

Grape variety	Months						
	Apr.	May	June	July	Aug.	Sep.	Apr.–May
Modrý Portugal	**–0.47**	**–0.77**	**–0.66**	–0.18	–0.41	–0.09	**–0.79**
Veltlínské zelené	–0.43	**–0.54**	–0.40	–0.11	–0.40	–0.07	**–0.61**
Frankovka	–0.45	**–0.69**	–0.44	–0.01	–0.34	–0.06	**–0.72**

Grape variety	Months						
	May–June	June–July	July–Aug.	Aug.–Sep.	Apr.–June	May–July	June–Aug.
Modrý Portugal	**–0.81**	**–0.49**	–0.39	–0.36	**–0.81**	**–0.72**	**–0.55**
Veltlínské zelené	**–0.53**	–0.30	–0.33	–0.33	**–0.56**	–0.47	–0.40
Frankovka	**–0.63**	–0.25	–0.22	–0.29	**–0.68**	**–0.49**	–0.34

Grape variety	Months						
	July–Sep.	Apr.–July	May–Aug.	June–Sep.	Apr.–Aug.	May–Sep.	Apr.–Sep.
Modrý Portugal	–0.32	**–0.78**	**–0.72**	**–0.48**	**–0.79**	**–0.65**	**–0.72**
Veltlínské zelené	–0.26	**–0.55**	**–0.52**	–0.35	**–0.58**	–0.47	**–0.54**
Frankovka	–0.19	**–0.58**	**–0.51**	–0.30	**–0.61**	–0.46	**–0.54**

with excellent, good or average wine. On the other hand, all three categories differ from sour wine to a statistically significant degree (Fig. 5).

Viticulture in the 19th century was also, of course, negatively affected by weather extremes. For example, as reported for Bzenec, in 1814 the majority of vineyards were frozen by spring frosts and everything that survived was frozen in August. For 1820, the record says "all vineyards frozen to tobacco and there was no wine". Further, frost damage in Bzenec was also mentioned for 1821 and 1823–1825. In 1825, as well as spring frosts, hail occurred three times, so bad that "it was not possible to recognise where the vineyards were". On the other hand, in 1834 grapes were already being harvested on 20 September and "the wine was so good that nobody remembered having wine of such unheard virtue: it was strong, sweet and clean" (Hanák 1922).

4. Phenology of the vine and recent warming: Velké Pavlovice 1956–2007

Phenological yearbooks published since 1923 contain no information about viticulture (e.g. Fenologická ročenka 1953–1963). This means that the relationships between vine phenophases and temperatures for a couple of decades may be studied only for the few stations preserved in the archives of the Czech Hydrometeorological Institute. Changes in observation practice may be another complication in obtaining such long series. The basic instructions for observers were laid down in 1956 (Pifflová et al. 1956) and include the following observed variables for the vine *Vinis vinifera*: start of pruning, emergence of buds and blossom, general flowering, general yellowing of leaves, and full maturity (beginning of vintage). This was replaced by new guidelines in 1984 (Valter 1981).

Looking at changes in these instructions and the type of data in documentary sources, it was possible to use only series for the beginning of blossoming and the harvest for the *Frankovka* (Lemberger) grape in Velké Pavlovice for 1956–2007. The monthly temperatures at the climatological station were first checked for relative homogeneity by Standard Normal Homogeneity Test (Alexandersson 1986) and then adjusted with respect to the non-homogeneity year of 1975. The start of blossoming correlates best with April–June temperatures (–0.79). Much weaker, but still statistically significant, are correlations of vintage with temperatures combined for different months (May –0.46, May–June –0.45, March–June and April–June –0.44, etc.; Fig. 6).

Correlations comparable with the *Frankovka* variety may be found for the *Veltlínské zelené* (Grüner Veltliner) variety and much better correlations for the *Modrý Portugal* (Blauer Portugieser) variety in the common period 1984–2007 (Table 2). This demonstrates differences in the sensitivity of individual grape varieties to temperature patterns in the vegetation period (Kraus 1964, 1999). The *Modrý Portugal* variety needs a sum of temperatures

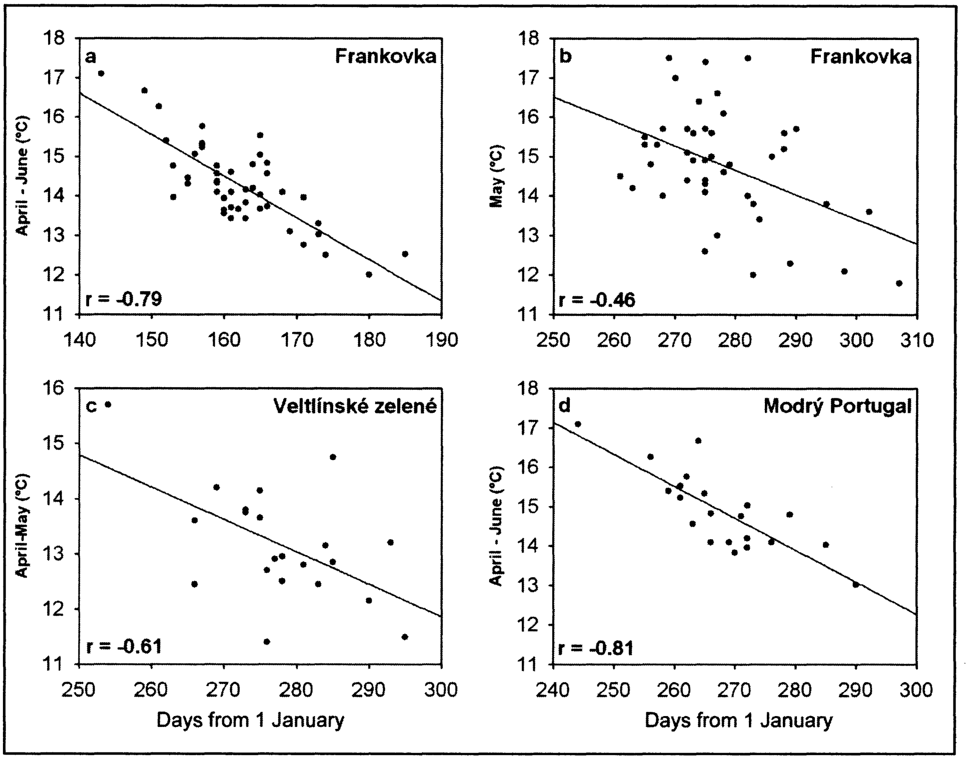

Fig. 6 – Correlations between temperatures and grapevine phenophases for various grape varieties at Velké Pavlovice: a) beginning of flowering, b) beginning of vintage – both 1956–2007, c) and d) beginning of vintage, 1984–2007; r – correlation coefficient.

≥10°C in the vegetation period of between 2,000 and 2,250°C for maturation. These values are reached early, and the beginning of vintage depends on the time at which they are achieved. On the other hand, the *Vetlínské zelené* and particularly *Frankovka* varieties need temperature sums between 2,500 and 2,750°C; they ripen later and can even be left longer in the vineyard for improvement of grape quality before harvesting. This means that changes in vintage beginnings are even more dispersed, sensitive to viticultural practices, and less directly related to temperatures.

5. Discussion and conclusions

Until the second part of the 19th century, viticulture was an essential source of livelihood for people in many settlements. For example, the village of Kurdějov turned to Empress Maria Theresa in 1766 with a request that its grain debt be relieved (Nosek 1908). The lord of the Pavlovice demesne had lent the grain in 1762, and the village intended to repay the debt from the abundant grape harvest of that year. However, this offer was twice refused. It emerged in the following part of the request that three barren wine years, 1763–1765, followed and in 1766 the vines were sorely afflicted by hail and grape mildew (*Peronospora* spp.), to the extent that the estimated damage amounted to thousands of gulden. Based on these facts, postponement of debt instalments or their cancellation was requested (Nosek 1908). Another example of the importance of viticulture is related to heavy frost damage from 10 to 15 May 1831, in which vineyards froze in Bavory, Dolní Věstonice and Pouzdřany. These settlements turned to the regional office in Brno with a request for tax reduction. Although they acknowledged that frost damage per se did not give them right to tax relief, they based their request on the basic importance of viticulture to the villages in the Mikulov area (Brázdil et al. 2003).

Failure of vine cultivation was not confined to individual places. Landsteiner (1999) demonstrated a sudden failure in wine production in Central Europe from the second half of the 1580s to the end of the 16th century. Series of bad grape harvests started in Switzerland in 1585, continued in 1586 in Württemberg and in 1587 affected Lower Austria and western Hungary. Lower wine production was followed by a definite reduction of viticulture income to the Habsburg state treasury. Moreover, higher wine prices and lower sweetness of the product led the public in Lower Austria to switch from wine to beer consumption. This data correlates well with information from Litoměřice and Židlochovice, where only small quantities of wine were recorded in 1585–1589 and 1591–1594, as well as explicitly sour wine in 1587, 1588, 1591, 1592, 1594 and 1597. In contrast, excellent wine was mentioned only in 1590, good wine in 1586 and 1599. Abundant yields occurred in 1586, 1590 and 1596, while in 1595 and 1597 the yield of grapes was characterised as average (Brázdil, Kotyza 2002).

All of this is summed up well by Hanák (1919), referring to the evaluation of historical yields and wine quality in Bzenec: "From records it follows that there are rather few years with excellent wine. For this reason, caution and patience should be prominent [virtues] for every wine merchant; caution (enough) to keep a good wine long enough, so that in the favourable year he will be able to fill barrels with good vintage, and the patience not to be faint-hearted in years of bad harvest and not to neglect vineyards, or even to

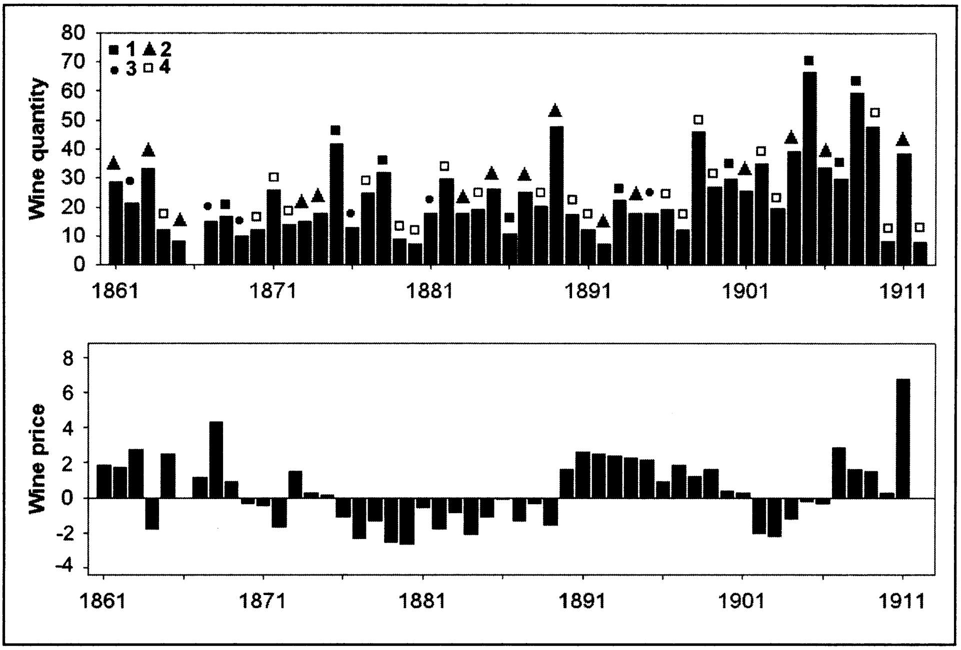

Fig. 7 – Fluctuation of detrended wine prices (gulden) at Bohutice for 1861–1912 (below) in comparison with wine quantity (hl) and quality: 1 – excellent, 2 – good, 3 – average, 4 – bad (above).

abandon them, but wait, since a favourable year will come which will compensate him for the costs of bad years."

Socio-economic factors do not stand alone; weather extremes, quantity and quality of wine were also reflected in the prices of wine, as documented for Bohutice in 1861–1912 (Fig. 7). To avoid inflation, prices were expressed as deviations from the corresponding linear trend. The highest positive deviations (expensive wine) were recorded in 1911 (probably also a consequence of the previous year, when frozen vines as well as cold and rainy weather occurred at blossoming time and the grapes were infected with mould) and in 1868 (bad harvest in the previous three years). Also interesting is a sudden jump from relatively cheap wine in 1876–1889 to more expensive wine in 1890–1901. No wine was available in 1866 due to May frosts, and in 1912 due to hail and mould. An even more important failure of vineyards started in 1866, when heavy frosts in this and following years destroyed many vineyards located at lower elevations. Another negative factor for viticulture might be related to the occurrence of phylloxera disease, which first appeared in 1872 in Austria and started to spread to further parts of the Austrian-Hungarian empire. As a consequence of this development, farmers around Znojmo and along the Dyje River eventually turned to the cultivation of cucumbers and other vegetables (Anonymous 1923).

It follows from this article that much data related to viticulture containing direct or proxy information about the weather and meteorological extremes exists in the Czech Republic. However, a great deal of this information is of a fragmentary character from the temporal and spatial points of view. The possibility that these gaps may be filled in the course of further archive research remains open. Such fragmentation provides a reason for the relative unavailability of continuous series of climate-relevant viticultural data for selected places (e.g. beginnings of grape harvest) that might be used for temperature reconstruction in the Czech Republic, with no chance of following the quantitative temperature reconstructions known from other countries (e.g. Chuine et al. 2004; Meier et al. 2007). The best viticulture data for the 19th century appears somewhat unpromising for temperature reconstructions since several long-term instrumental temperature series already exist. On the other hand, data on the impacts of weather extremes on grape production are already significantly completing and verifying the chronologies of extreme events available in the historical-climatological database of the Institute of Geography, Masaryk University, Brno, and may also be used in research into historical viticulture.

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Vinobranní, vinice a víno. SOkA Znojmo, fond Katalog staré spisovny města Znojma, inv. č. XVIII.1, karton 404.

Shrnutí

VINAŘSTVÍ V MINULOSTI A V SOUČASNOSTI JAKO ZDROJ KLIMATOLOGICKÝCH POZNATKŮ V ČESKÉ REPUBLICE

Pěstování vinné révy (Vinis vinifera) a vinařství v České republice je významně ovlivněno počasím. Závislost na počasí umožňuje využít informace o pěstování vinné révy (např. začátek vinobraní, množství a kvalita vína) jako nepřímý indikátor pro rekonstrukci teploty vzduchu a výskytu povětrnostních extrémů v minulosti. Tato studie shrnuje základní poznatky o vztazích mezi vinařstvím a klimatem v českých zemích pro oblast Čech (zejména Lounsko a Litoměřicko) a jižní Moravy během několika posledních století opírajíce se o různé typy dokumentárních pramenů (kroniky, paměti, deníkové záznamy, osobní a oficiální korespondence, noviny, záznamy ekonomické povahy, odborné práce). Zatímco před rokem 1500 jsou informace tohoto typu spíše sporadické v závislosti na dochovaných pramenech, pro 16.–18. století již bylo možné stanovit dekádové četnosti výskytu povětrnostních extrémů se škodami na vinné révě (pozdní jarní a časné podzimní mrazy, krupobití), kvality vína (špatné, průměrné, dobré, výborné) a množství vína (málo nebo žádné, průměrně, nadbytek) ve vinařských oblastech Čech a jižní Moravy (obr. 3). V řadě roků během těchto tří století však jakékoliv informace o pěstování vinné révy chybí úplně. Data začátků vinobraní ze Znojma, získaná z protokolů městské rady nebo oznámení v novinách Znaimer Wochenblatt pro období 1800–1890 byla využita pro rekonstrukci teplot vzduchu dubna–srpna v Brně s použitím lineárního regresního modelu (obr. 4). Parametry rekonstrukce (tab. 1) sice ukazují na využitelný potenciál této metody pro teplotní rekonstrukce, dostupné řady počátků vinobraní pro období před začátkem systematických teplotních měření v českých zemích jsou však zatím spíše sporadické. Pro Bzenec (1800–1890), Znojmo (1802–1845) a Bohutice (1861–1912) byly použity řady kvality vína (výborné, dobré, průměrné, špatné či kyselé) k vyjádření jejich závislosti na teplotě vzduchu (obr. 5). Zatímco v případě Bzence a Znojma bylo možné prokázat statisticky významné rozdíly mezi průměrnými teplotami odpovídajícími jednotlivým kategoriím kvality vína, pro Bohutice se takto lišily pouze první tři kategorie od špatného či kyselého vína. Začátek květu vinné révy a vinobraní v závislosti na teplotě vzduchu je analyzován pro Frankovku, Veltlínské zelené a Modrý Portugal podle fenologických pozorování ve Velkých Pavlovicích pro období 1956–2007, resp. 1984–2007, na bázi průměrných měsíčních teplot (obr. 6). U teplotně náročnějších odrůd vinné révy jako je Veltlínské zelené a Frankovka je vazba začátku vinobraní na teplotě vzduchu předchozích měsíců podstatně slabší než u méně náročné odrůdy Modrý Portugal (tab. 2). Pro Bohutice je demonstrováno kolísání cen vína v letech 1861–1912 v porovnání s množstvím vína a jeho kvalitou (obr. 7).

Obr. 1 – Místa v České republice zmiňovaná v textu: 1 – Bavory, 2 – Bohutice, 3 – Bratčice, 4 – Bzenec, 5 – Čejkovice, 6 – Dolní Kounice, 7 – Dolní Věstonice, 8 – Drnholec, 9 – Hustopeče, 10 – Kamýk, 11 – Klentnice, 12 – Krupka, 13 – Kuchařovice, 14 – Kurdějov, 15 – Litoměřice, 16 – Louny, 17 – Lovosice, 18 – Malíč, 19 – Mělčany, 20 – Mělník, 21 – Mikulov, 22 – Mušov, 23 – Olomouc, 24 – Pavlov, 25 – Perná, 26 – Ploskovice, 27 – Pouzdřany, 28 – Raná, 29 – Syrovice, 30 – Velké Pavlovice, 31 – Velké Žernoseky, 32 – Znojmo, 33 – Židlochovice.

Obr. 2 – Symboly sklizně a lisování vinných hroznů byly použity Johannem Willenbergem pro znázornění měsíce října v kalendáři na rok 1604. Vinobraní začínalo v tomto měsíci nejčastěji na Lounsku stejně jako v jiných místech českých zemí.

Obr. 3 – Dekádové četnosti výskytu povětrnostních extrémů se škodami na vinné révě (1 – pozdní jarní mráz, 2 – časný podzimní mráz, 3 – krupobití), kvality vína (4 – špatná, 5 – průměrná, 6 – dobrá, 7 – výborná) a množství vína (8 – málo nebo žádné, 9 – průměrné, 10 – mnoho) ve vinařských oblastech Čech (a) a jižní Moravy (b) v 16.–18. století.

Obr. 4 – Variabilita začátků vinobraní ve Znojmě (počet dnů od 1. září; nahoře) a porovnání měřených (1) a rekonstruovaných (2) teplot vzduchu dubna–srpna v Brně v podobě anomálií od referenčního období 1961–1990 a míra nejistoty rekonstrukce (šedě) vyjádřená jako 95% interval spolehlivosti pro každou rekonstruovanou hodnotu v období 1800–1890 (dole).

Obr. 5 – Krabicový graf teplot vzduchu dubna–srpna v Brně počítaných pro čtyři skupiny kvality vína ve Znojmě (1802–1845), Bzenci (1800–1890) a Bohuticích (1861–1912): a – výborná, b – dobrá, c – průměrná, d – špatná.

Obr. 6 – Korelace mezi teplotami vzduchu a fenofázemi vybraných odrůd vinné révy ve Velkých Pavlovicích: a) začátek květu, b) začátek vinobraní – obojí v období 1956–2007, c) a d) začátek vinobraní, 1984–2007; r – korelační koeficient.

Obr. 7 – Kolísání detrendovaných cen vína (zlaté) v Bohuticích v letech 1861–1912 (dole) v porovnání s množstvím vína (hl) a jeho kvalitou: 1 – výborná, 2 – dobrá, 3 – průměrná, 4 – špatná (nahoře).

Tab. 1 – Míry vhodnosti rekonstrukce teplot vzduchu dubna–srpna v Brně. Data o vinobraní ze Znojma jsou použity jako nezávisle proměnná v modelu: r^2 – kvadrát korelačního koeficientu, RE – redukce chyby, CE – koeficient citlivosti, MSE – střední kvadratická chyba.

Tab. 2 – Korelační koeficienty mezi průměrnými měsíčními teplotami vzduchu a začátky vinobraní pro vybrané odrůdy vinné révy ve Velkých Pavlovicích v období 1984–2007 (chybějící roky: Modrý Portugal – 1985, 1987, 1991; Veltlínské zelené – 1987, 1989, 1991; Frankovka – 1987, 1988, 1991, 1997). Statisticky významné koeficienty podle t-testu na hladině významnosti $\alpha = 0.05$ jsou vytištěny tučně.

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ZBIGNIEW W. KUNDZEWICZ, DAMIAN JÓZEFCZYK

TEMPERATURE-RELATED CLIMATE EXTREMES IN THE POTSDAM OBSERVATION RECORD

Z. W. Kundzewicz, D. Józefczyk: *Temperature-related climate extremes in the Potsdam observation record.* – Geografie–Sborník ČGS, 113, 4, pp. 372–382 (2008). – This paper examines temperature-related climate extremes in the unique long-term gap-free record at the Secular Meteorological Station in Potsdam. Increasing tendencies in daily minimum temperature in winter and daily maximum temperature in summer, as well as monthly means of daily minimum temperatures in winter months and of daily maximum temperatures in summer months are illustrated. Also the numbers of hot days and of summer days (with maximum daily temperature exceeding 30 °C and 25 °C, respectively) have been increasing. In agreement with warming of winter minimum temperatures, the numbers of frost days (with minimum daily temperature below 0 °C) and of ice days (with maximum daily temperature below 0 °C) have been decreasing. However, low correlation coefficient and huge scatter illustrate strong natural variability, so that the occurrence of extremes departs from the general underlying tendency.
KEY WORDS: climate extremes – climate variability – climate change – temperature – trend.

Introduction

As noted in IPCC (2007), warming of the global climate system is unequivocal. This is now evident from observations of increases in air temperatures, which show clear trends at a range of scales, from local, via regional, to continental, hemispheric, and global. Most of the observed increase in global mean air temperature since the mid-20th century is very likely due to the observed increase in anthropogenic greenhouse gas concentrations. The updated 100-year linear trend (1906 to 2005) shows a 0.74 °C (0.56 °C to 0.92 °C) global mean temperature increase, while the linear warming trend over the last 50 years (0.65 °C) is nearly twice as strong as that for the last 100 years (IPCC 2007). Global temperature changes are accompanied by changes in other climatic variables.

However, besides conducting the trend detection studies, the research community has been carefully watching temperature-related records of climate extremes in different categories, such as maximum and minimum daily, monthly, seasonal, and annual temperatures. Absolute record values of maximum or minimum temperature do not necessarily match the tendencies present in the long-term time series. Even if a clear rising trend of global mean annual temperature is unabated, the highest global mean annual temperature on record occurred already nine years ago, in 1998 (related to a strong El Niño phase). This is so, despite the fact that among 13 globally warmest calendar years in the global instrumental observation record, available since 1850, there are 12 years from the last 13 years. Each of the years 2001–2007 belongs to the set of second-warmest to eighth-warmest years (Brohan et al. 2006, Kundzewicz 2008).

Beside the data at larger scales, it is of much interest to examine long time series of good-quality observation records, wherever available, looking for changes at regional and local scales.

The secular meteorological station in Potsdam

The data set used in the present paper stems from the Secular Meteorological Station in Potsdam (Germany), located at the south-west of the town (co-ordinates: 52°23'N, 13°04'E, elevation 81 m a.s.l.), approximately 600 m away from the built-up area. It is a notable station, world-wide, with an uninterrupted observation programme carried out since January 1, 1893. The station was established with the purpose to serve for a longer time (for ages, since the word saeculum means age in Latin). Comprehensive information about the station, as well as a wealth of long-term observation records can be found in public domain at the web portal: http://www.klima-potsdam.de/. Open access to the long time series of good-quality climatic observations at the Potsdam observatory encourages scientists to analyze these data holdings. This is true not only for Germany, but also for the neighbour country, Poland, where no hydrometeorological data are in public domain and the prohibitively high cost charged by the national hydrometeorological service is not affordable to most scientists (cf. Kundzewicz et al. 2007).

Among the many variables that have been measured at the station are: air and ground temperature, air pressure, global radiation, relative humidity, water vapour pressure, wind speed, precipitation, cloudiness, snow cover, frost depth, sunshine hours, such weather events as haze and storm. Considerable efforts have been made to keep the observation conditions homogeneous, by maintaining the station location, conditions of the environment, methods and principles of instrumental observation.

In the present contribution, valuable long gap-free records at Potsdam are examined in the context of temperature-related climate extremes, understood here as the maximum or minimum values of selected climate indices at a pre-defined time interval (e.g. 12 consecutive months, year, season, month, day). Among the variables tackled are temperature (minimum, maximum); the number of frost days (with minimum daily temperature below 0 °C) and ice days (with maximum daily temperature below 0 °C); the number of hot days (with maximum daily temperature exceeding 30 °C) and summer days (with maximum daily temperature exceeding 25 °C).

Annual and 12-month temperature records

The mean annual temperature observed at Potsdam shows a clearly increasing trend (Kundzewicz et al. 2007). There have been seven calendar years on record with mean annual temperature in excess of 10 °C (Table 1). In 2007, the mean annual temperature record of 2000 (10.47 °C) was not exceeded, but the value recorded for 2007 was only marginally lower (10.46 °C).

Even if the instantaneous temperature value is measured with 0.1 °C accuracy, Table 1, presenting mean annual temperature, calculated from

Table 1 – Warmest years at the Potsdam secular meteorological station

Rank	Year	Mean annual temperature (°C)
1	2000	10.47
2	2007	10.46
3	1934	10.44
4–5	1989, 1999	10.26
6–7	1990, 2006	10.17

daily mean temperatures, contains values with 0.01 °C resolution. This allows, for instance, ordering of the years 2000 and 2007. Leaving 0.1 °C resolution one would not distinguish between the mean annual temperature in 2000 and 2007.

There is a strong random component in climate extremes that illustrates natural variability and weather vagaries. Occurrence of a record high mean monthly temperature does not necessarily mean that the highest daily maximum temperature in this month is record high. For example, July 2006 was the warmest July on record, as far as the monthly mean temperature is concerned (23.69 °C). However, the highest daily maximum temperature during this month was 35.9 °C, that is below the highest daily maximum temperature of 36.8 °C, observed during a much less warm July 2007, with mean monthly temperature being 18.05 °C only.

Even if a record of a mean temperature during a calendar year dates back to 2000 and has not been exceeded since, looking at the mean temperature of any consecutive 12-month period, that commences on the 1st of any month (rather than on January 1), one can find a new record. Until 2007, the warmest 12 consecutive months in the history of observations at the Potsdam Secular Meteorological Station were recorded from July 1999 to June 2000, with a mean temperature of 10.70 °C. On 31 January 2007, this record went up to 10.83 °C, on February 28, 2007 – to 11.16 °C. On March 31, 2007 the

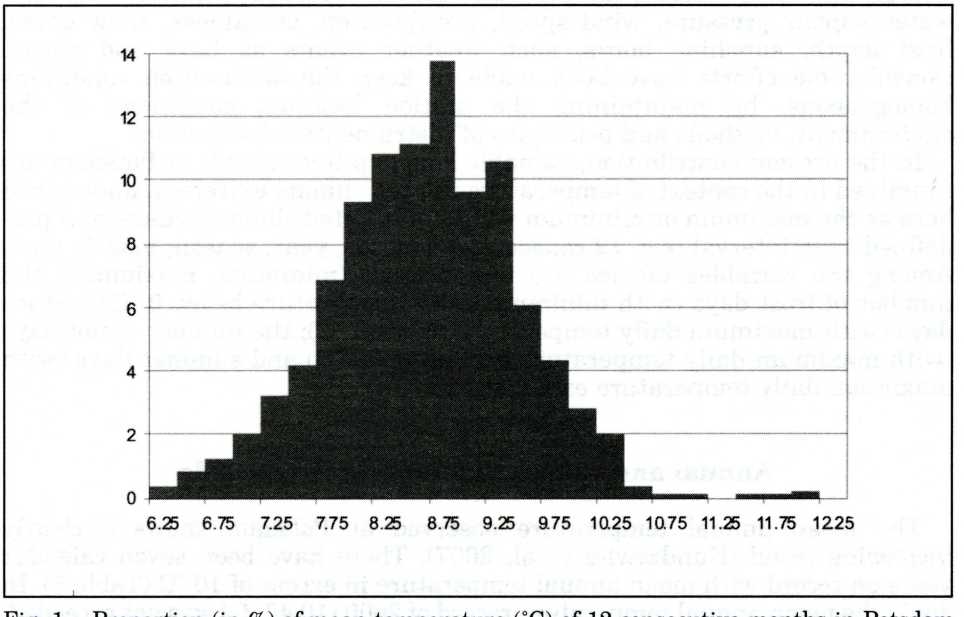

Fig. 1 – Proportion (in %) of mean temperature (°C) of 12 consecutive months in Potsdam (after Kundzewicz et al., 2007). All 12-month periods from 1 January 1893 to 30 June 2007 were considered. Value of 2 corresponding to the interval 7.0–7.25 means that 2 % of all values of 12-month mean temperature belong to this interval.

Table 2 – Review of records of temperature-related climate extremes. Data from www.klima-potsdam.de.

Variable	Date of record	Value of record	Other recent entries in the list of 5
Maximum daily temperature	9.8.1992	39.1 °C	Rank 4 reached in 1994
Minimum daily temperature	11.2.1929	–26.8 °C	Most recent entry (rank 4) in 1969
Longest hot[1] period	23.7. to 6.8.1969	15 days	Rank 2 reached in 1994
Longest cold[2] period	21.1. to 26.2.1947	37 days	Rank 4 reached in 1996
First hot day in year	22.4.1968	31.8 °C	Ranks 2 and 3 reached in 2000 and 1996, respectively
Last hot day in year	20.9.1947	32.9 °C	No entries after 1975 in top 5
First frost[3] day in year	2.10.1957	–0.1 °C	This is the most recent entry in top 5
Last frost day in year	20.5.1952	–0.8 °C	This is the most recent entry in top 5
First summer[4] day in year	30.3.1968	25.7 °C	Among entries in top 5: 1985 and 1989
Last summer day in year	10.10.1995	25.3 °C	Rank 3 in 1985. Attention – rank 2 was observed in 1893!
Daily maximum sum of precipitation	8.8.1978	105.7 mm	Rank 2 in 2002
Longest dry spell	19.9.–20.10.1949	32 days	Rank 4 in 1996 (winter)
Longest wet spell	2.2.–6.3.1970	33 days	This was the most recent entry in top 5
Maximum snow depth	6.3.1970	70 cm	All five entries in the top 5 refer to the same month (March 1970)
Longest period with closed snow cover	1 December 1969 to 23 March 1970	113 days	Rank 5 refers to Dec 1978 – March 1979

Notes: [1] Hot day is understood as a day with $T_{max} \geq 30$ °C; [2] Cold (ice) day is understood as a day with $T_{max} < 0$ °C; [3] Frost day is understood as a day with $T_{min} < 0$ °C; 4 Summer day is understood as a day with $T_{max} \geq 25$ °C.

record increased vigorously (by 0.48 °C) to 11.64 °C, because the cold March 2006 did not count any more in calculations, and on April 30, 2007 it rose strongly again (by 0.27 °C) to 11.91 °C, as April 2007 – warmest on record – replaced less warm April 2006. Further, on May 31, 2007 the record went up to 12.04 °C, and on June 30, 2007 – to 12.09 °C (Kundzewicz et al. 2007). This latter figure is higher than the record before 2007 (see Fig. 1) by a very large increment (1.39 °C). All six outlier-like values (between 10.75 and 12.25) in the tail of the distribution of the 12-month mean temperature presented in Figure 1 occurred in 2006–2007.

It was shown in Kundzewicz et al. (2008) that in 2006-2007, the records of mean temperature over consecutive 12 months have been exceeded also at the national, continental, and hemispheric scales.

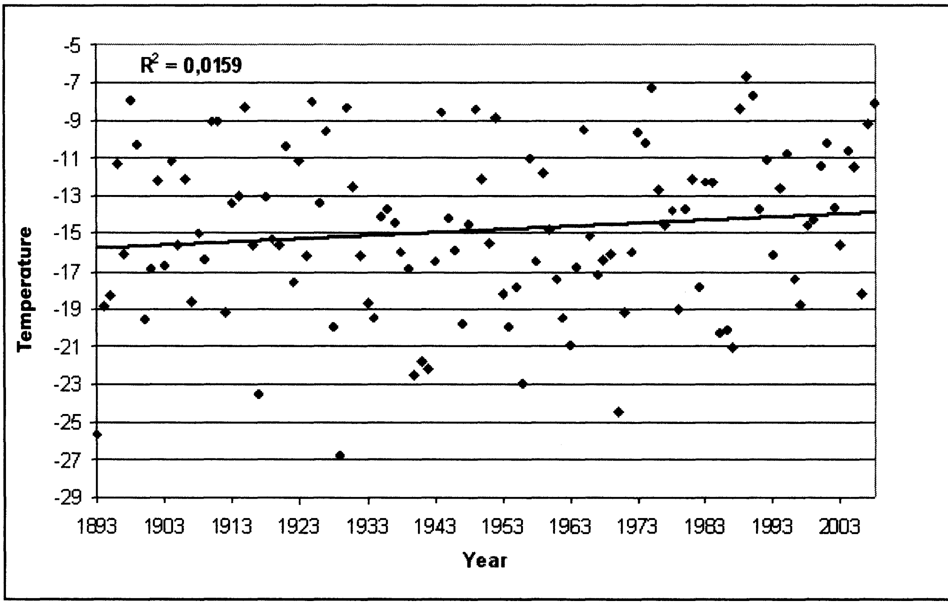

Fig. 2 – Lowest minimum daily temperature (°C) for winters from 1893 to 2008

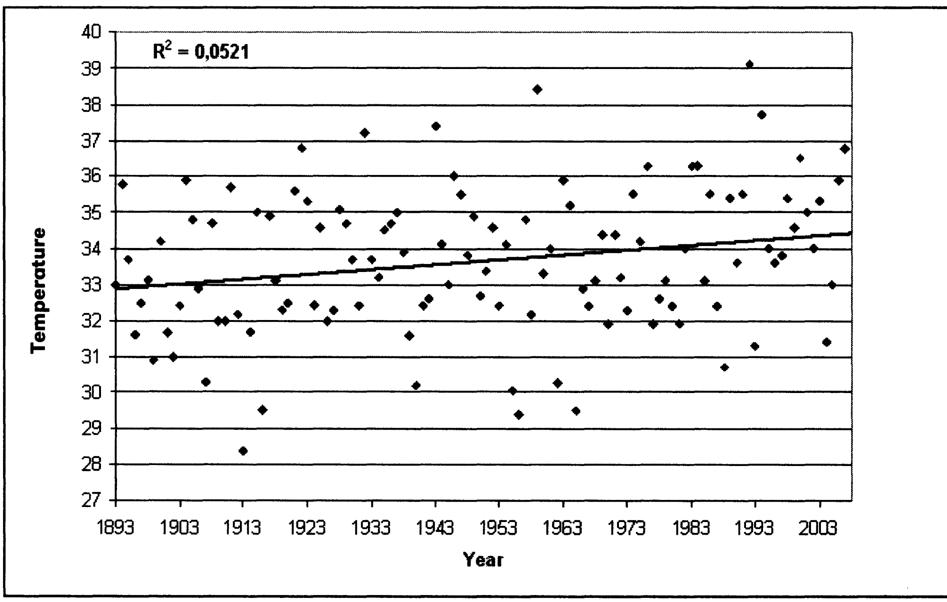

Fig. 3 – Highest maximum daily temperature (°C) for summers from 1893 to 2007

Review of records

Careful look at the table of records from Potsdam (Table 2; data from www.klima-potsdam.de) allows to conclude that cold extremes have been rare

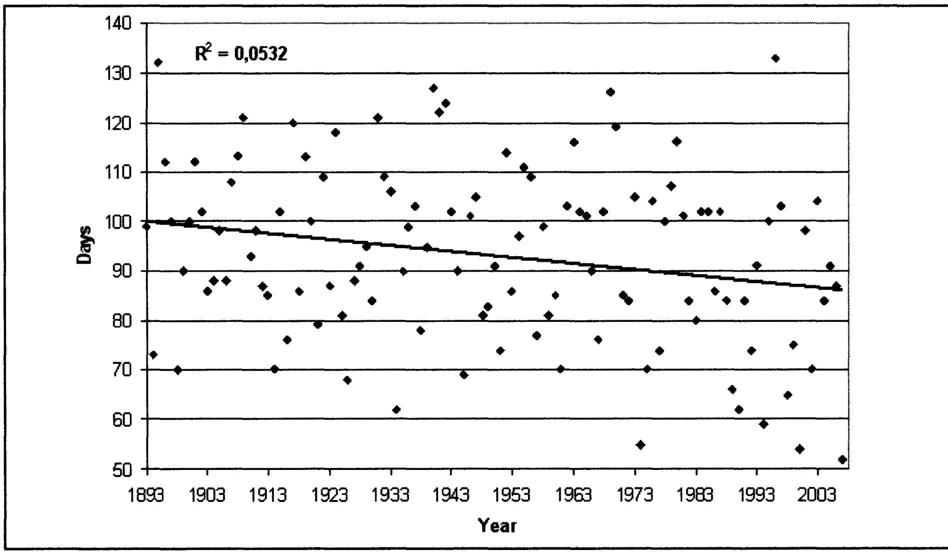

Fig. 4 – Number of frost days in individual winters from 1893 to 2008

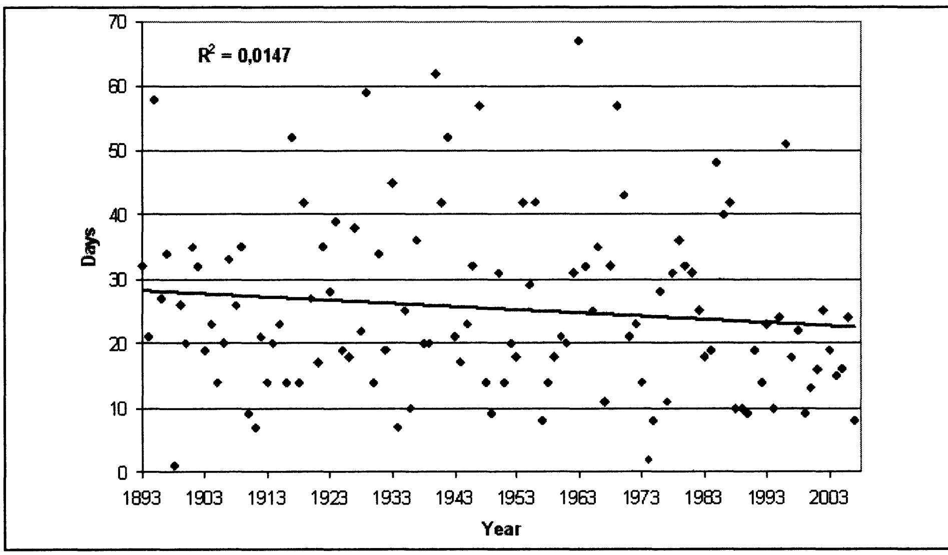

Fig. 5 – Number of ice days in individual winters from 1893 to 2008

in recent decades, while warm extremes occur more often. For example, the most recent entry in the list of five lowest values of minimum daily temperature dates back to 1969 (rank 4), while two entries in the list of five highest values of maximum daily temperature occurred after 1990 (ranks 1 and 4).

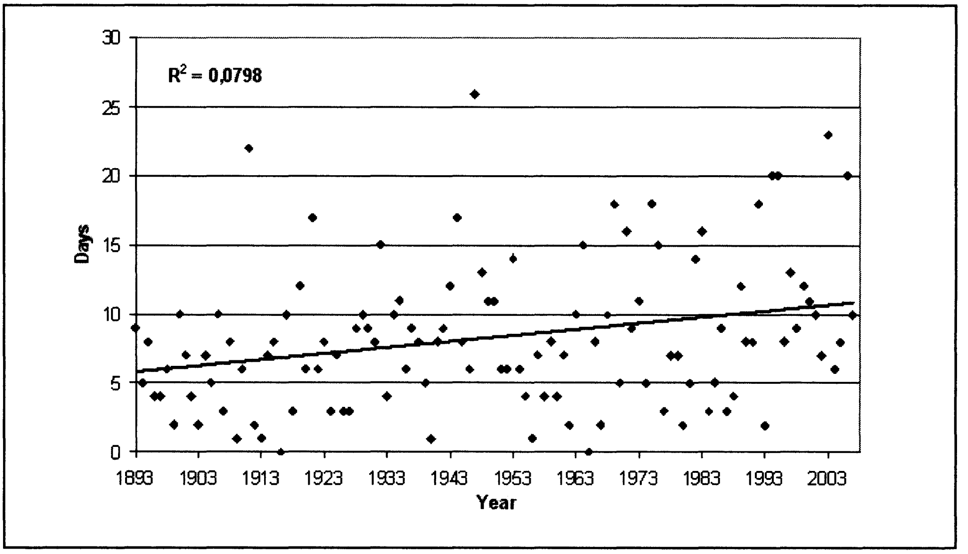

Fig. 6 – Number of hot days in individual years from 1893 to 2007

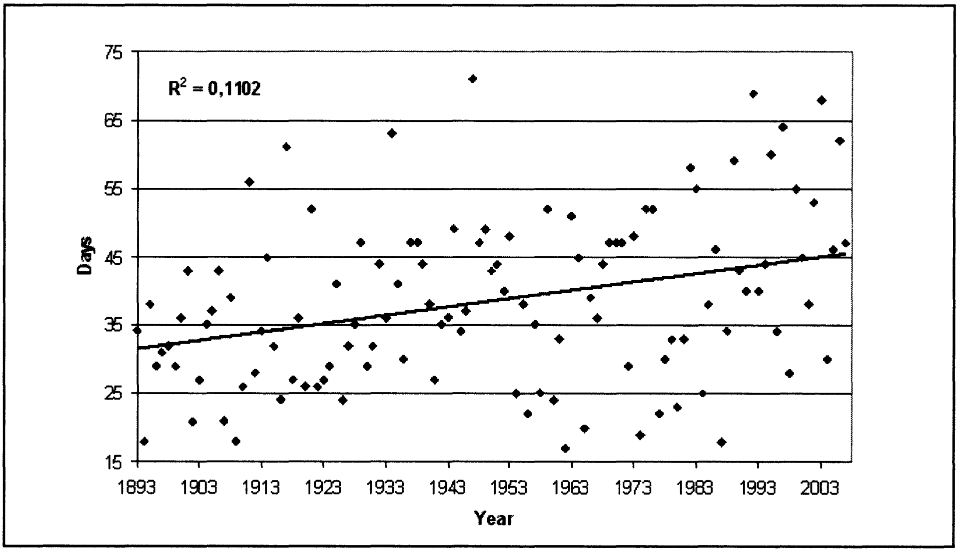

Fig. 7 – Number of summer days in individual years from 1893 to 2007

Indices of temperature-related climate extremes

Trends in daily minimum temperature in winter (December, January, February) and daily maximum temperature in summer (June, July, August), as well as monthly means of daily minimum temperatures in winter and monthly means of daily maximum temperatures in summer are examined next.

Table 3 – Monthly means of daily maximum temperature in summer

	June		July		August	
	Year	Temperature (°C)	Year	Temperature (°C)	Year	Temperature (°C)
Warmest	1917	27.15	2006	30.3	1997	28.09
2nd warmest	1992	26.59	1994	29.26	1944	27.97
2nd coldest	1984	18.39	1954	19.55	1941	19.79
Coldest	1923	15.62	1898	19.33	1902	19.54

Table 4 – Monthly means of daily minimum temperature in winter

	December		January		February	
	Year	Temperature (°C)	Year	Temperature (°C)	Year	Temperature (°C)
Warmest	2006	2.80	2007	2.24	1998	2.41
2nd warmest	1974	2.49	1975	2.13	1990	2.15
2nd coldest	1933	–6.85	1893	–12.76	1956	–13.35
Coldest	1969	–8.99	1940	–13.52	1929	–15.43

Figures 2 and 3 illustrate variations of the lowest minimum daily temperature in winter and of the highest maximum daily temperature in summer, for each of the years 1893–2008. Although the rising trend is visible, the scatter is very strong and overshadows the signal. The correlation coefficient attains low values. For lowest minimum daily temperature in winter the correlation coefficient takes considerably lower values than for highest maximum daily temperatures in summer.

The warming trend for the minimum temperature in winter is stronger than for the maximum temperature in summer (attention: the temperature scales in Figs. 2 and 3 are different). The number of frost days (defined as days with minimum temperature below 0 °C) and ice days (defined as days with maximum temperature below 0 °C) have been decreasing with time (Figs. 4 and 5), but in individual years, departures from the overall decreasing trend are very strong. For example, within the last 15 years both the highest value (133 frost days in 1996) and the lowest value (52 frost days in 2007) on record have been observed.

Figures 6 and 7 illustrate, respectively, the numbers of hot days, defined as days with maximum daily temperature exceeding 30 °C and the numbers of summer days (with maximum daily temperature exceeding 25 °C), for individual years. Increasing tendency is clearly seen in both figures and the values of correlation coefficient are significantly higher than in all the time series of indices considered earlier in this paper.

The records of monthly means of daily maximum temperature in summer and of daily minimum temperature in winter are presented in Tables 3 and 4, respectively. These tables, containing the two highest and the two lowest values on record, agree with the warming tendency, as expected. Warm extremes get gradually more frequent and cold extremes – less frequent. Among the two warmest temperature values for each summer and winter months, there are 8 (out of 12) entries since 1990 and only two before 1950. However, variability of monthly means is very high, so that high or low values occur throughout the time period analyzed, independently on the general tendency. For instance, in 1917, the highest monthly mean of daily maximum temperature of June (27.15 °C) was observed, even if the climate was clearly colder than now. Only six years later, in 1923, the lowest monthly mean of daily maximum temperature of June (15.62 °C) was observed. Tables 3 and 4 show that individual outliers-like lowest values may be very much below the second lowest values. For instance, the two coldest Decembers had mean daily minima of –8.99 °C in 1969 and –6.85 °C in 1933 (difference of 2.14 °C). The two coldest Februaries with mean daily minima of –15.43 °C (1929) and –13.35 °C (1956), show a difference of 2.08 °C. The two Junes with lowest daily maxima of 15.62 °C (1923) and 18.39 °C in 1984 manifest an even larger difference of 2.77 °C.

Conclusions

Climatic time series show strong natural variability (irregular oscillations), which is superimposed on a gradual trend accompanying the warming signal. Extremes get more extreme – says the IPCC report. Such tendency refers to both observations and even more so to projections, but there is a strong random component, so that heat records are not broken every year. Indeed, hot extremes occurred in old times, while cold extremes occur also now (albeit less frequently).

As reported by Trenberth et al. (2007), in the last 50 years there has been a significant decrease in the annual occurrence of cold nights (falling below the 10th percentile from the control reference period) in winter. The distributions of minimum and maximum temperatures have not only shifted to higher values, consistent with overall warming, but the cold extremes have warmed more than the warm extremes. More warm extremes imply an increased frequency of heat waves. There has been a global trend towards fewer frost days associated with the warming.

The present paper indicates that these global and general findings also hold for a specific, long-term, high-quality observation record. However, it shows that the natural variability is very strong and that extremes in individual year may largely differ from the dominating tendency. It can be clearly seen that an extreme value of such temperature-related indicators as maximum temperature in summer, minimum temperature in winter, number of hot days, and number of summer days may have occurred in the remote past, when the level of warming (as indicated by the linear regression) was much lower than now. Similarly, despite the warming, high values of the number of frost days and ice days may occur recently, largely exceeding the low value resulting from the decreasing tendency.

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Shrnutí

TEPLOTNÍ KLIMATICKÉ EXTRÉMY ZAZNAMENANÉ PŘI POZOROVÁNÍCH V POSTUPIMI

Pozorované teploty vzduchu, které vykazují na všech úrovních vzestupný trend, svědčí o nesporném oteplování globálního klimatického systému. Globální lineární teplotní trend je v posledních padesáti letech silný (0,65 °C). Pozorované vzestupy průměrné teploty vzduchu od poloviny 20. století jsou většinou pravděpodobně důsledkem pozorovaného vzestupu koncentrace skleníkových plynů, produkovaných člověkem.

Příspěvek doplňuje rozsáhlé souhrnné výsledky uvedením tendencí dlouhodobých záznamů velmi kvalitních pozorování. Zkoumá klimatické extrémy teploty vzduchu v jedinečném, dlouhodobém a souvislém pozorování na meteorologické stanici v Postupimi, prováděném od ledna 1893 do února 2008. Průměrná roční teplota vzduchu zaznamenaná v Postupimi vykazuje zřetelný vzestupný trend. I když rekordní průměrná teplota v kalendářním roce byla dosažena v roce 2000 a nebyla od té doby překonána, průměrná teplota jakéhokoliv následujícího dvanáctiměsíčního období začínajícího prvním dnem kteréhokoliv měsíce (spíše než 1. lednem) je vždy rekordní. Rekordní dvanáctiměsíční průměrná teplota před rokem 2007 byla 10,70 °C, ale na konci června 2007 dosáhla 12,09 °C a výrazně překonala předchozí hodnotu (o 1,39 °C).

Jsou prokázány rostoucí minimální denní teploty v zimě (prosinec, leden, únor) a maximální denní teploty v létě (červen, červenec, srpen), jakož i rostoucí měsíční průměry minimálních denních teplot v zimních měsících a maximálních denních teplot v letních měsících. Rostou také ukazatele extrémního tepla, jako počet horkých dní (s maximálními denními teplotami překračujícími 30 °C) a letních dní (s maximálními denními teplotami nad 25 °C). V souvislosti se zvyšováním minimálních zimních teplot se snižují indikátory extrémního chladu, jako je počet mrazových dní (s minimálními denními teplotami pod bodem mrazu) a ledových dní (s maximálními denními teplotami pod bodem mrazu). V určitých letech se však objevují značné odchylky od tohoto obecně poklesového trendu. Například během posledních 15 let byly pozorovány jak nejvyšší hodnoty (133 mrazových dní v roce 1996), tak nejnižší hodnoty (52 mrazových dní v roce 2007). Nízké hodnoty korelačního koeficientu a široký rozptyl, který zastiňuje tento signál, svědčí o silné přirozené variabilitě a o existenci extrémních odchylek od obecné tendence.

Pro nejnižší minimální denní teploty v zimě dosahuje korelační koeficient výrazně nižší hodnoty než pro nejvyšší maximální denní teploty v létě. Tendence růstu minimálních teplot v zimě je silnější než tendence zvyšování maximálních teplot v létě. Nejen to, že se rozložení minimálních a maximálních teplot posunulo k vyšším hodnotám, což svědčí o globálním oteplování, ale i teplota extrémních chladen vzrostla více než teplota extrémního tepla. Více teplotních extrémů znamená vyšší frekvenci vlivů horkých vln.

Klimatické řady svědčí o silné přirozené variabilitě (nepravidelné výkyvy), které se přidávají k postupnému trendu oteplování. Jak konstatuje zpráva Mezivládního panelu pro klimatické změny (IPCC), extrémy se stávají ještě extrémnějšími. I v minulosti

samozřejmě docházelo k extrémnímu horku, ale extrémní chladna jsou nyní výrazně méně častá, i když se samozřejmě mohou objevovat i dnes. Vzhledem k silné náhodné komponentě nejsou teplotní rekordy překonávány každým rokem.

Předložená práce uvádí, že tyto globální a obecné poznatky platí také pro konkrétní dlouhodobá a vysoce kvalitní pozorování. Ukazuje se však, že přirozená variabilita je velice silná a že extrémy se mohou v jednotlivých letech velmi odchylovat od převládajícího trendu. Je jasně vidět, že velice vysoké hodnoty takovýchto extrémů spojených s teplotou, jako jsou maximální teploty v létě, minimální teploty v zimě, počet horkých dní a počet letních dní, se mohly objevovat i v dávné minulosti, kdy úroveň oteplování (indikovaného lineární regresí) byla výrazně nižší než dnes. Podobně i navzdory oteplování se i dnes (i když ne příliš často) může objevit vysoký počet mrazových a ledových dní, který vysoce přesahuje nízké hodnoty, které plynou z uvedené poklesové tendence.

Obr. 1 – Podíl (%) průměrné teploty vzduchu (°C) ve dvanácti po sobě následujících měsících v Postupimi (podle Kundzewicze et al., 2007). Zkoumána byla všechna dvanáctiměsíční období od 1. ledna 1893 do 30. června 2007. Hodnota 2 odpovídající intervalu 7,0–7,25 znamená, že 2 % všech hodnot dvanáctiměsíčních průměrných teplot náleží do tohoto intervalu.
Obr. 2 – Nejnižší minimální denní teploty vzduchu v zimě od roku 1893 do roku 2008.
Obr. 3 – Nejvyšší maximální denní teploty vzduchu v létě od roku 1893 do roku 2007.
Obr. 4 – Počet mrazových dní během jednotlivých zim v období 1893–2008.
Obr. 5 – Počet ledových dní během jednotlivých zim v období 1893–2008.
Obr. 6 – Počet horkých dní během jednotlivých let v období 1893–2007.
Obr. 7 – Počet letních dní během jednotlivých let v období 1893–2007.

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BOHUMÍR JANSKÝ, JAN KOCUM

PEAT BOGS INFLUENCE ON RUNOFF PROCESS: CASE STUDY OF THE VYDRA AND KŘEMELNÁ RIVER BASINS IN THE ŠUMAVA MOUNTAINS, SOUTHWESTERN CZECHIA

B. Janský, J. Kocum: *Peat bogs influence on runoff process: case study of the Vydra and Křemelná River basins in the Šumava Mountains, southwestern Czechia.* – Geografie–Sborník ČGS, 113, 4, pp. 383–399 (2008). – Specific part of wide complex of preventive measures against floods and extreme droughts could be procedures realized in river headstream areas. In order to increase a water retention in headwaters the detailed analysis of peat bogs hydrological function needs to be carried out. Suitable conditions for the research realization at present is related to an existence of several automatic water level gauges and utilization of modern equipment and methods in experimental catchments of the Otava River headstream area (Šumava Mts., southwestern Czechia), representing the core zone of a number of extreme floods in Central Europe. Thorough analyses of extreme runoff phases show more distinct discharge variability of streams draining peat land localities. For the retention potential assessment the detailed measurement of potential accumulation reservoirs, bathymetric mapping of bog pools and the detailed analysis of snow conditions as an important component of a rainfall-runoff process in headwaters is being pursued. The final part of the paper is consisted of suggestions of several unforceable measures implementation that could contribute to reduction of peak flows and to increase of water resources during extreme droughts in future.
KEY WORDS: retention potential – headstream area – flood protection – upper Otava River basin – runoff variability – drought – peat bogs hydrological function.

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1. Introduction

In context of catastrophic floods and extreme droughts in recent years there is an urgent need of solving of flood protection issues and measures leading to discharge increase in dry periods, not using just classical engineering methods but also untraditional practices. There is a new strategy focusing on gradual increase of river catchment retention capacity including the realization of measures as runoff retardation and water retention increase in headstream areas.

In order to enhance the retention potential in headwaters the detailed analysis of peat bogs hydrological function needs to be done. The peat bogs influence on runoff conditions and other hydrographic and climatic characteristics is being assessed by detailed comparison of hydrological regimes in subcatchments with different peat land proportion. The research is concentrated to the upper Otava River basin (Vydra and Křemelná River basins) representing a territory with frequent occurrence of flood events and with a high heterogeneity in terms of physical-geographic and social-economic aspects. We can reason about the peat bogs influence on hydrological process also with respect to its affecting of water quality, respectively to ionic structure of water in periods of high or low discharges (Novák 1955, 1959; Onderíková, Štěrbová 1956; Oulehle, Janský 2003).

The problem of peat bogs hydrological function has not been so far fully solved despite a number of domestic and foreign projects and broad debates among experts. Opinions on such issues vary as it is evident in the literature that has been dealing with these questions already in the second half of the 19th century. The detailed analysis of various approaches was made by Ferda (1960). The so called "theory of sponge", which was in a literature acknowledged approximately from the last 60s, supposed the importance of peat land for its significant water retention and discharge regulating capability during high rainfall totals and its discharge heightening and runoff balancing in dry periods. From the last 70s studies that infirm the peat bogs retention function appear. They assert that only possible way to increase their retention capacity is to lower groundwater level by means of a drainage. Then these ameliorative hits were realized in a number of mountainous areas in Czechia. Since that time, the problem of drainage, respectively dyking of former drainage channels, has become the field for broad debates within a literature (Conway, Millar 1960; Burke 1967; McDonald 1973; Moklyak et al. 1975; Baird 1997; Holden et al. 2001; etc.). The detailed study of the literature representing both opinion poles was carried out by Holden et al. (2004).

Results of studies dealing with this research subject proved that water courses draining peat land areas show significant discharge variability and that the peat land influence on runoff regime balance had been overestimated in the past. It was found out that winter snow precipitations have a relatively low influence on discharge increase in summer period while summer rainstorms play a very significant role in this sense. While filling peat bogs up, runoff values increase rapidly. As well, during longer droughts, peat land does not play any positive role in hydrological terms, i.e. they do not feed water courses draining them. On the contrary, the past research projects state the hydrological regime improvement after peat bogs drainage and ameliorating. Peat land influence on water quality in water courses is assessed as unambiguously negative while intensity of affection is related to its area and volume in a catchment. The problem of pollution is further intensified in water reservoirs located in former peat land and moor areas.

Mentioned topic is currently studied in the upper Otava River basin (Janský, Kocum 2007a, 2007b and Kocum, Janský 2007a, 2007b). Outcomes of the research is considered to be used within the realization of specific effective flood protection measures in cooperation with all concerned institutions (including flood warning system, etc.). Assessment of peat bogs revitalizing measures influence of chosen localities on its hydrological regime change is also one of the project goals. Every single element of rainfall-runoff process, especially snow conditions in the study area, needs to be completely studied. In order to increase the retention potential in this area a qualified reference of measures being implemented at present by the Šumava Mts. National Park Management in connection with former ameliorative channels (made during communist regime) dyking needs to be done. The influence of peat bog localities on runoff process is assessed by detailed comparison of Vydra River and Křemelná River hydrological regimes (Otava River main sources with significantly different peat land proportion in their catchments). Profile Otava River – Rejštejn (catchment area 336.5 km^2, T. G. Masaryk Water Research Institute – WRI GIS layers), long-term mean discharge Q_a=7.56 m^3.s^{-1} (Czech Hydrometeorological Institute – CHMI data) is the closing profile of the study area.

2. Methodology

More than 30 years ago the first results related to peat bogs hydrological function were presented within the study of CHMI in Prague (Ferda, Hladný, Bubeníčková, Pešek, 1971). In this project drainage and ameliorating of peat bog beds is recommended with regard to an improvement of their hydrological function. According to results from domestic and foreign literature it is stated that maximum discharges could be markedly reduced this way as a result of groundwater level decrease and consequently of extension the depth of peat bog surface layer for capturing causal rainfall totals. It is hereat adverted to other positive effects such as increase of forest stand accretion on drained areas (Vidal, Schuch 1963; Huikari 1963; Robertson, Nicholsen, Hughes 1963). This study is so far the last paper dealing with hydrological regime and water chemism in the upper Otava River basin focused on peat bog habitation.

In recent years the Otava River headstream area has become the study catchment of the research consisting partly in bathymetric mapping of organogenous lakes (bog pools, so called Šumava Mts. Moors) including specification of their main physical parameters and chemical composition, but especially in the initiation of thorough monitoring of Vydra and Křemelná River runoff regimes inclusive of assessment of various measures leading to their source areas retention potential increase. Very favourable conditions for the realization of this project currently bear on a better accessability to the study area and lengthening data time series but also on using quite modern equipment and methods.

At the end of 2005 six water level laths were installed in chosen profiles (Roklanský Brook, Modravský Brook, Filipohuťský Brook, Vchynicko-tetovský Floating Channel – Rechle, Křemelná River above Prášilský Brook, Prášilský Brook above Křemelná River) in order to initiate hydrological observations. Till recent years water level values had been read constantly by local observers in a one day step (during melting process in spring period twice a day) in these profiles. Since summer period 2006 sixteen automatic ultrasound and hydrostatic pressure water level gauges with dataloggers for continual monitoring of water level fluctuation (11 in Vydra River basin, 5 in Křemelná River basin; Fig. 1) were subsequently installed beyond these profiles. Furthermore 4 water stage recorders within the CHMI water stage recorders network (Otava River – Rejštejn, Křemelná River – Stodůlky, Vydra River – Modrava, Hamerský Brook – Antýgl) and two profiles controlled by ČEZ group (Czech electropower company; Vchynicko-tetovský Floating Channel – Rechle and Mechov) became parts of the research network system. Besides, one meteorological station (observing precipitation, air temperature and humidity, solar radiation, wind speed and direction and soil temperatures) and

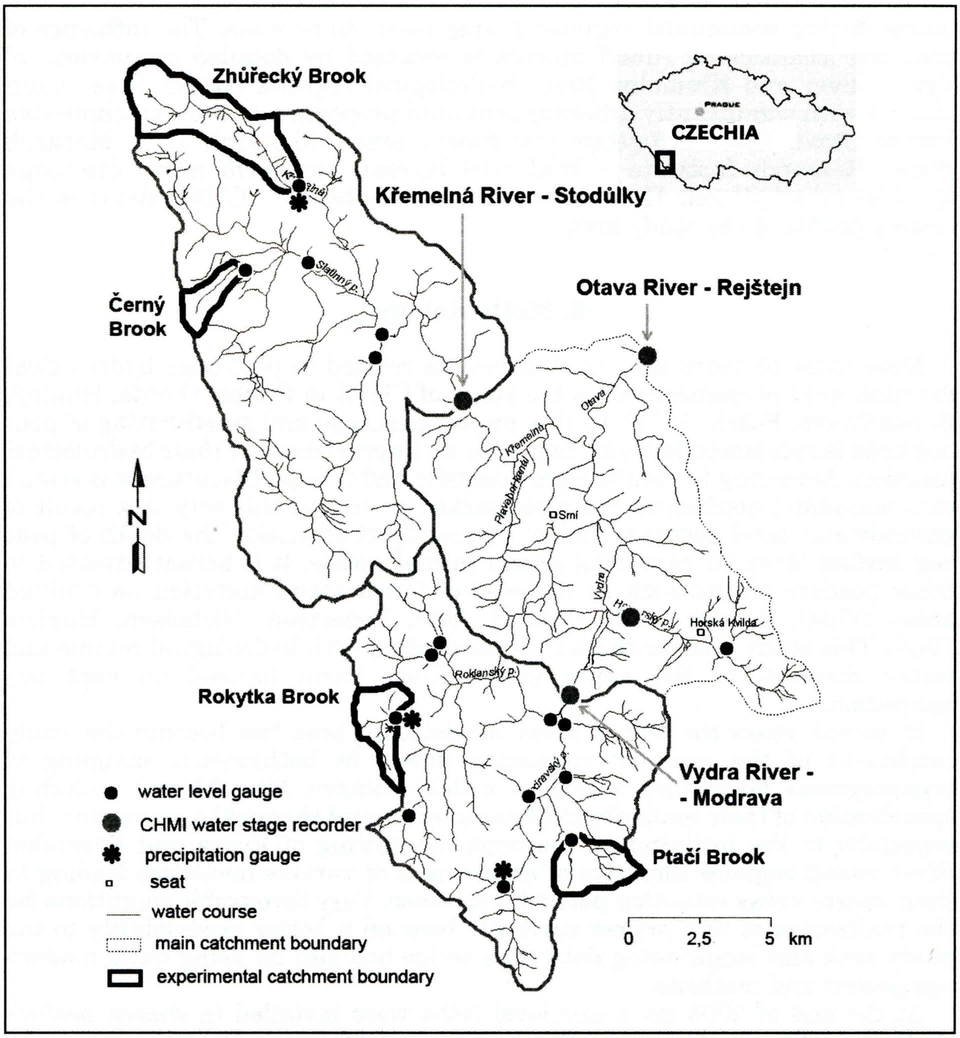

Fig. 1 – Localization of the Vydra River and the Křemelná River study basins with CHMI water stage recorders and experimental subcatchments with automatic water level gauges and shuttle precipitation gauges within the headstream area of Otava River (Otava River–Rejštejn closing profile)

two other shuttle precipitation gauges measuring in 10 minutes step were installed in the upper part of the Vydra River basin (Rokytka Brook; measures since 18 September 2006 excepting several months in winter periods) and Křemelná River basin (Zhůřecký Brook; measures since 29 March 2007). From technical reasons the amount of snowfall during winter periods cannot be measured meanwhile in mentioned profiles.

Measuring set from Fiedler–Mágr Company including registering and controlling unit of M4016 type, ultrasound or pressure sensor and GSM module for data transmission by means of GPRS network is used for continual water level monitoring in 10 minutes step with 1 mm accuracy. Data transmission in a one day interval or shorter depending on the dynamics of hydrological situation allows its operative solution and also regular control of whole measuring set function. In given profiles with installed water level gauges periodical discharge measurements using hydrometric propeller are carried out in order to construct an accurate consumption curves. Meanwhile, totally about 250 discharge measurements have been done.

Analyses of snow conditions, being a very significant element of rainfall-runoff process in Czech headstream areas, were carried out in the last two winter periods. Snow cover height and snow water equivalent (SWE) monitoring is done by point measurements with specific spatial distribution considering altitude, exposition, slope and vegetation cover. Acquired data are then digitalized and interpolated using suitable methods in GIS software (ArcMap, MapInfo, Surfer) so spatial distribution of snow reserves could be assessed. Information about accumulation dynamics is logged on the base of several measurements during winter period. Snow cover height and SWE measurement is carried out by means of the snow hydrometer SM 150–50 and exact position and altitude of measurement points is determined using GPS60CSx and GPS Leica.

The detailed survey of potential spaces for capturing causal rainfall totals and runoff retardation and successive assessment of potential accumulation reservoirs efficiency is carried out using automatic total geodetic station Levica TCRP1202 R1000.

3. Results

The main part of the project is the assessment of a hydrological regime in the Otava River basin headstream area including an evaluation of various measures for its retention potential enhancement and comparison of runoff variability in chosen subcatchments with regard to the peat land occurrence. In order to reach the goal the assessment and comparison of runoff variability partly in the Vydra and Křemelná River basins state profiles with relatively long time series, partly in experimental subcatchments with installed automatic ultrasound or hydrostatic pressure water level gauges, is carried out.

3.1. Runoff regime analysis in the Vydra and Křemelná River basins

In order to compare hydrological regimes in basins of Otava River two main sources, Vydra and Křemelná Rivers, from the runoff variability point of view, the data of mean daily discharges in water stage recorders within the CHMI basic network are used. Two of these profiles are discussed: Vydra River – Modrava (catchment area 90.17 km^2; WRI GIS layers), long-term mean discharge 3.483 m^3.s^{-1} (CHMI data), peat land proportion 38 % (Ferda, Hladný, Bubeníčková, Pešek 1971) and Křemelná River – Stodůlky (134.11 km^2, 3.722 m^3.s^{-1}, 5 %; Fig. 1). With respect to the fact that time series in studied profiles are not of the same length, the time period when both water stage recorders were in function is assessed. Accordingly the period 1 November 1999 – 31 October 2006 was analysed. This relatively short period could appear to be non-representative, yet basic statistical analyses of daily, monthly and annual time series were made and runoff was on the base

Table 1 – Runoff variability characteristics in the Vydra River - Modrava and Křemelná River – Stodůlky profiles (1 November 1999 – 31 October 2006 period).

Profile	Vydra R.-Modrava	Křemelná R.-Stodůlky
catchment area (km^2)	90.17	134.11
long-term mean discharge ($m^3.s^{-1}$)	3.483	3.722
minimum discharge ($m^3.s^{-1}$)	0.763	0.880
maximum discharge ($m^3.s^{-1}$)	55.100	64.600
specific runnoff ($l.s^{-1}.km^{-2}$)	38.5	27.8
minimum specific runnoff ($l.s^{-1}.km^{-2}$)	8.5	6.6
maximum specific runnoff ($l.s^{-1}.km^{-2}$)	611.0	481.7
yearly discharge volume (km^3)	0.1098	0.1174
runoff height (mm)	1,218	875
median	2.030	2.410
dispersion	15.859	15.153
mean divergence from mean value	2.471	2.405
authoritative divergence	3.982	3.893
decil divergence (Q_d)	0.601	0.502
coefficient of variability C_v (Q_d)	1.143	1.046
coefficient of variability C_m (Q_m)	0.421	0.392
monthly discharge variability coef. K_r	5.048	3.686

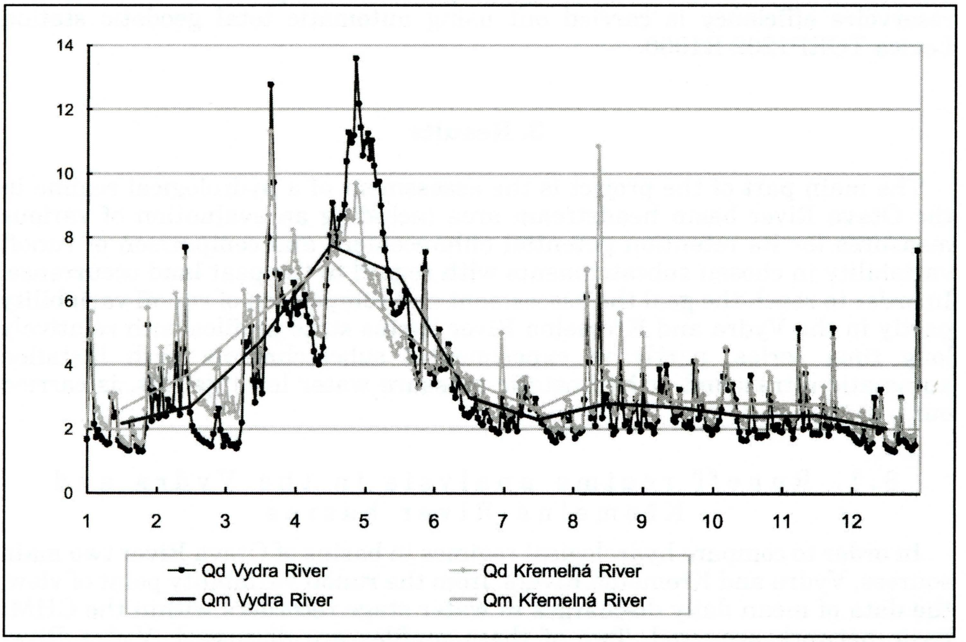

Fig. 2 – Long-term mean daily (Q_d) and monthly discharges (Q_m) in the Vydra River–Modrava and Křemelná River–Stodůlky profiles in 1 November 1999 – 31 October 2006 period. Axis x – months, axis y – discharge ($m^3.s^{-1}$).

of its variability graphically defined from the daily and monthly discharges point of view (Fig. 2) and described by characteristics presented in Table 1. On the basis of these outcomes it could be presumed that runoff variability appears to be slightly higher in the case of the Vydra River – Modrava profile. This fact is demonstrated especially by K_r coefficient used for runoff assessment from the monthly discharge variability point of view. Higher

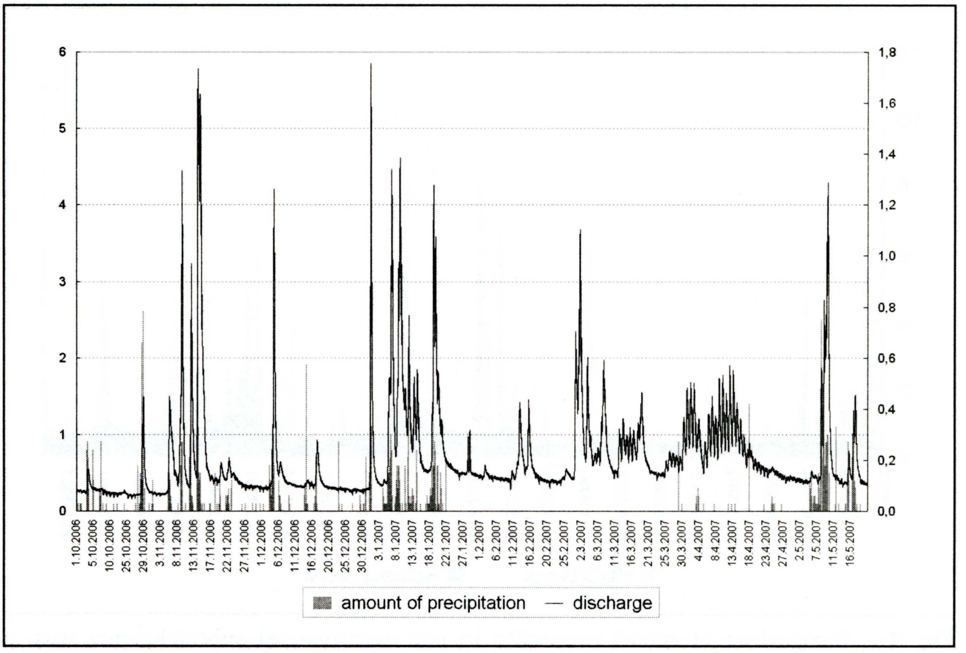

Fig. 3 – Outcome from the automatic ultrasound water level gauge and shuttle precipitation gauge – runoff reaction to causal amount of precipitation in the Rokytka Brook profile (Vydra River headstream area) in 1 October 2006 – 20 May 2007 period (in 1 January 2007 – 28 March 2007 period precipitation gauge was out of service due to the technical reasons). Axis y left – amount of precipitation (mm), axis y right – discharge ($m^3.s^{-1}$).

runoff fluctuation is accordingly described in the profile closing the catchment with more significant peat land proportion. It is also evident by comparison of ratios of the maximum (April) and minimum (December) mean monthly discharge in the studied profiles which reach up to 3.83 in the Vydra River – Modrava profile, respectively 3.35 in the Křemelná River – Stodůlky profile.

3.2. Runoff variability in experimental subcatchments

In consequence of present short period of water level fluctuation monitoring using automatic water level gauges only partial results are kept at disposition. Preview of one of outcomes from the ultrasound water level gauge and shuttle precipitation gauge is presented in the Figure 3. It shows the discharge fluctuation of Rokytka Brook (Vydra River headstream area; catchment area 3.721 km^2; WRI GIS layers) in relation to the amount of precipitation in 1 October 2006 – 20 May 2007 period. Significant discharge increase during spring period as a result of snow melting process in the catchment is very distinct. Nevertheless, striking runoff fluctuation (between 0.2 and 0.5 $m^3.s^{-1}$) was registered also within the day.

As it was mentioned, in order to assess the peat bog localities and peat forming soils influence on the runoff regime variability two subcatchments within the upper Otava River study basin with significantly different peat land proportion were chosen. In the upper part of the Rokytka Brook

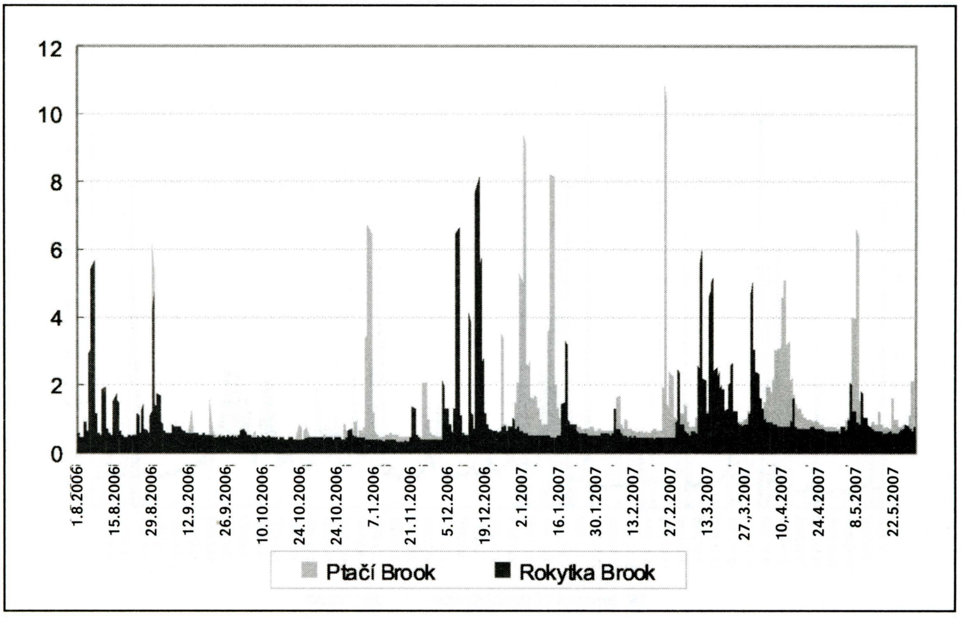

Fig. 4 – Comparison of runoff variability in two experimental subcatchments with a different peat land proportion (Rokytka Brook, Ptačí Brook) in 1 August 2006 – 31 May 2007 period. Axis y – Q/Q_p.

catchment closed by the profile with installed automatic water level gauge a large complex of so called "Rokytecké Moors" is situated (approximately 55% peat land proportion; Fig. 1). Much more sporadic occurrence of peat bog beds is fixed to the Ptačí Brook catchment (about 10 % peat land proportion; catchment area 4.063 km^2; WRI GIS layers). Slightly higher runoff variability in the case of Rokytka Brook is quite distinct from Figure 4. In doing so, discharge variability is besides absolute value of culmination defined especially by a peak flow frequency. Different rate of discharge of both water courses in monitored profiles is taken into account using Q/Q_p ratio, where Q is actual 10-minutes discharge and Q_p is mean discharge from the serie of all 10-minutes discharges from the whole monitoring period.

Through the exact study of runoff ascending and descending phases, concretely through the analysis of runoff reaction to causal rainfall (interval between maximum 10-minutes amount of precipitation and corresponding peak flow) during several rainfall situations within the monitoring period, more significant peak flow retardation in the Zhůřecký Brook profile (Fig. 1; about 4:40 hours in average; catchment area 13.946 km^2; WRI GIS layers) compared to the Rokytka Brook profile (about 3:20 hours) was determined. It signifies higher water retention potency in the catchment with distinctively lower peat land proportion. Mentioned claims necessarily demand stronger reliance in terms of longer data time series and detailed analyses of a larger number of namely extreme rainfall situations.

To sum it up, detailed analyses of extreme runoff phases show higher frequency of peak flows and their shorter reaction to causal rainfall in case of highly peaty areas, therefore more distinct runoff variability of streams draining peat land localities and peat forming soils. Continuously much more

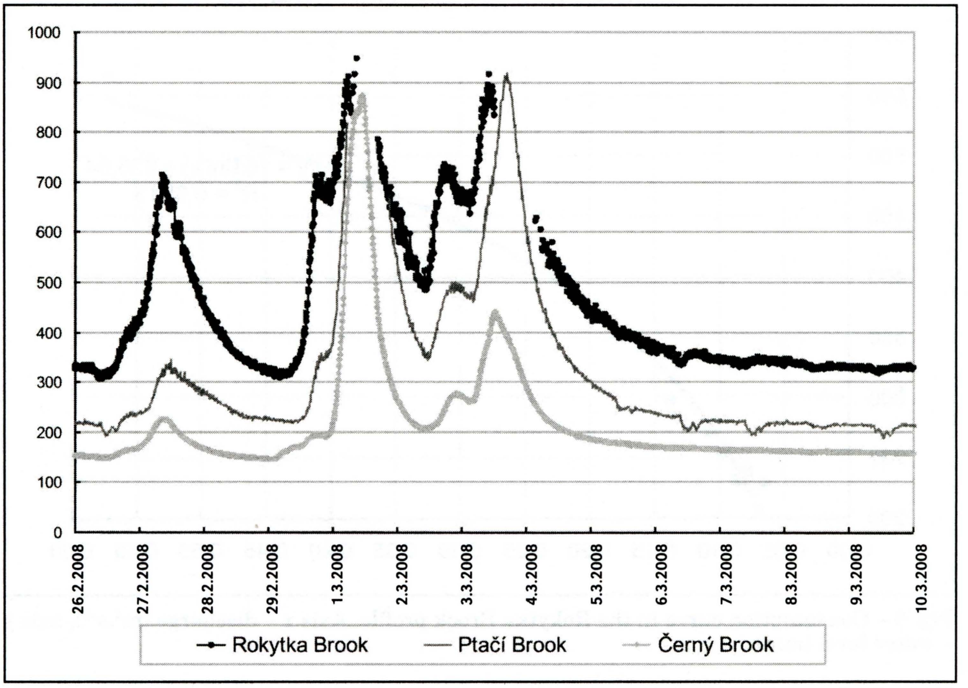

Fig. 5 – Water level fluctuation in three model subcatchments with different peat land proportion during the extreme flood situation in March 2008. Axis y – water level (mm).

detailed studies of hydrological and climatic data time series need to be done, especially reaction analyses of runoff from several peat bog localities in relation to rainfall duration, intensity and spatial distribution in monitored catchments by means of thorough study of its ascending and descending phases.

Hydrological monitoring in above mentioned watercourses is completed by the ionic balance including carbon and oxygen isotopes balance observing in 2008 hydrological year (cooperation with Czech Geological Survey) in order to make the precise separation of runoff phases by means of anion deficiency. Twice a month the atmospheric deposition samples and water samples are subscribed in relation to the monitored discharges.

3.3. Extreme flood event analysis in model subcatchments

In order to compare the runoff conditions of areas with different physical-geographic parameters during an extreme runoff situation, flood event at the beginning of March 2008 was partly analysed in three experimental subcatchments. Their choice was concentrated mainly on the peat land proportion in their territories. To reach the goal, within these three localities one catchment with very high proportion of peaty areas (Rokytka Brook; approx. 55 %) and one with very low value (mineral Černý Brook catchment; about 5 % proportion of peat land; catchment area 2.419 km^2; WRI GIS layers) were chosen. The distribution of moor areas within the catchment of Ptačí Brook is also quite low (see Fig. 1).

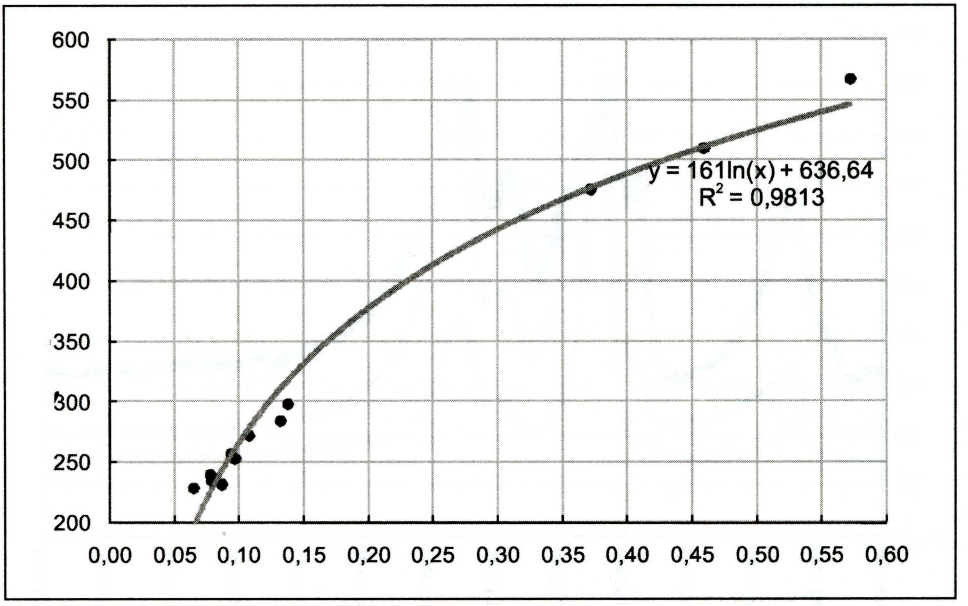

Fig. 6 – Consumption curve in the Rokytka Brook profile. Axis x – discharge ($m^3.s^{-1}$), axis y – water level (mm).

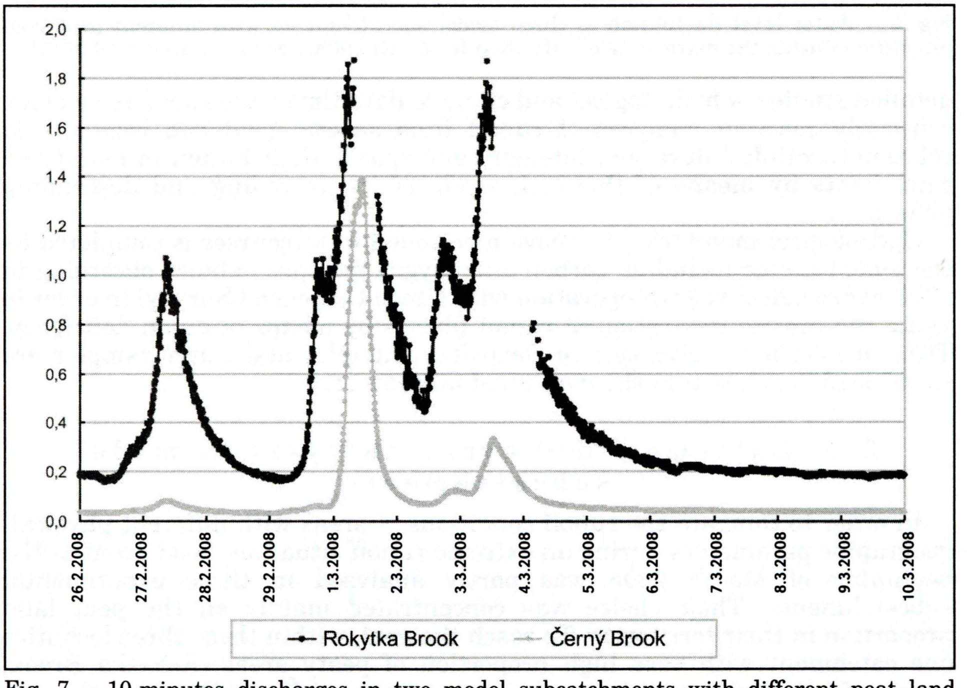

Fig. 7 – 10-minutes discharges in two model subcatchments with different peat land proportion during the extreme flood event in March 2008 (highly peaty area is represented by the Rokytka Brook catchment). Axis y – discharge ($m^3.s^{-1}$).

The extreme amount of precipitation which affected a large territory of Europe early in the year 2008, known as Emma atmospheric low pressure, became very intensive especially in this study area. The runoff reaction on such an extreme rainfall totals in headwaters is generally very rapid and is demonstrated even by the graph in Figure 5. Water level fluctuation in three study profiles with installed automatic gauges for discharge monitoring quite apparently show very similar references mentioned above in the paper - higher runoff variability in the case of a highly peaty area, even during such an extrem hydrological situation. Corresponding discharge values calculated using consumption curves (Fig. 6) in two out of three observed profiles are even more evident (see Fig. 7). Consumption curves are described by considerably high reliability values. The existence of periods with missing discharge data is caused by an enormous extremity of this event.

3.4. Snow conditions analysis

Even snow conditions in a catchment, as it was already mentioned, form very important phenomenon in our mountainous river headstream areas. Their detailed analysis represents a necessary basis for correct assessment of runoff formation in these areas and for truthful integration of this intricately quantificating element into hydrological processes modelling.

In February 2007 and February and March 2008 the detailed snow survey in the experimental subcatchments closed by profiles with automatic water level gauges (Rokytka Brook, Ptačí Brook and Černý Brook catchments) in the upper part of the Vydra River and Křemelná River basins, was carried out. Using snow laths and snow hydrometers the snow cover height and SWE was mapped in order to describe conditions for snow accumulation dynamics in this territory. For example, in the Ptačí Brook catchment with an area of 4.063 km^2 44 point measurements were done (it corresponds to the measurement density of approximately 11 points/km^2 of an area). Character of snow cover occurrence is characteristic by a very significant time and spatial variability (Fig. 8). For the process of point data interpolation the IDW (Inverse Distance Weighted) method, where various values of input parameters were tested, was used. While the snow cover height in February 2007 in the lowest part of the Vydra River catchment (Modrava – 980 m a.s.l.) reached up to about 30 centimetres, values in the highest parts of the catchment (1,330 m a.s.l.) were oscillating around 90 cm. CHMI station Churáňov (1,118 m a.s.l.) was measuring in the time of a field survey around 30 cm of the snow cover. The situation one year later (February 2008) was even more variable. While the snow cover thickness was moving around 30 to 50 cm in the lowest parts of the above mentioned catchment, values measured in the source area close to the border with Germany were reaching up to about 150 cm (Kocum, Janský 2008).

Snow conditions survey in experimental subcatchments confirmed the existence of a considerable difference in the amount of accumulated snow storage not just in relation to an altitude but also to a vegetation cover, especially between open and forest areas. It indicates the circumstance which is difficult to be afflicted by measurements carried out usually on state meteorological stations. Along with this fact, very significant variability of snow storages on different localities of a similar character and altitude was proved. It gives an evidence about a considerable complexity of the snow cover accumulation process and its significant dependance on a number of factors.

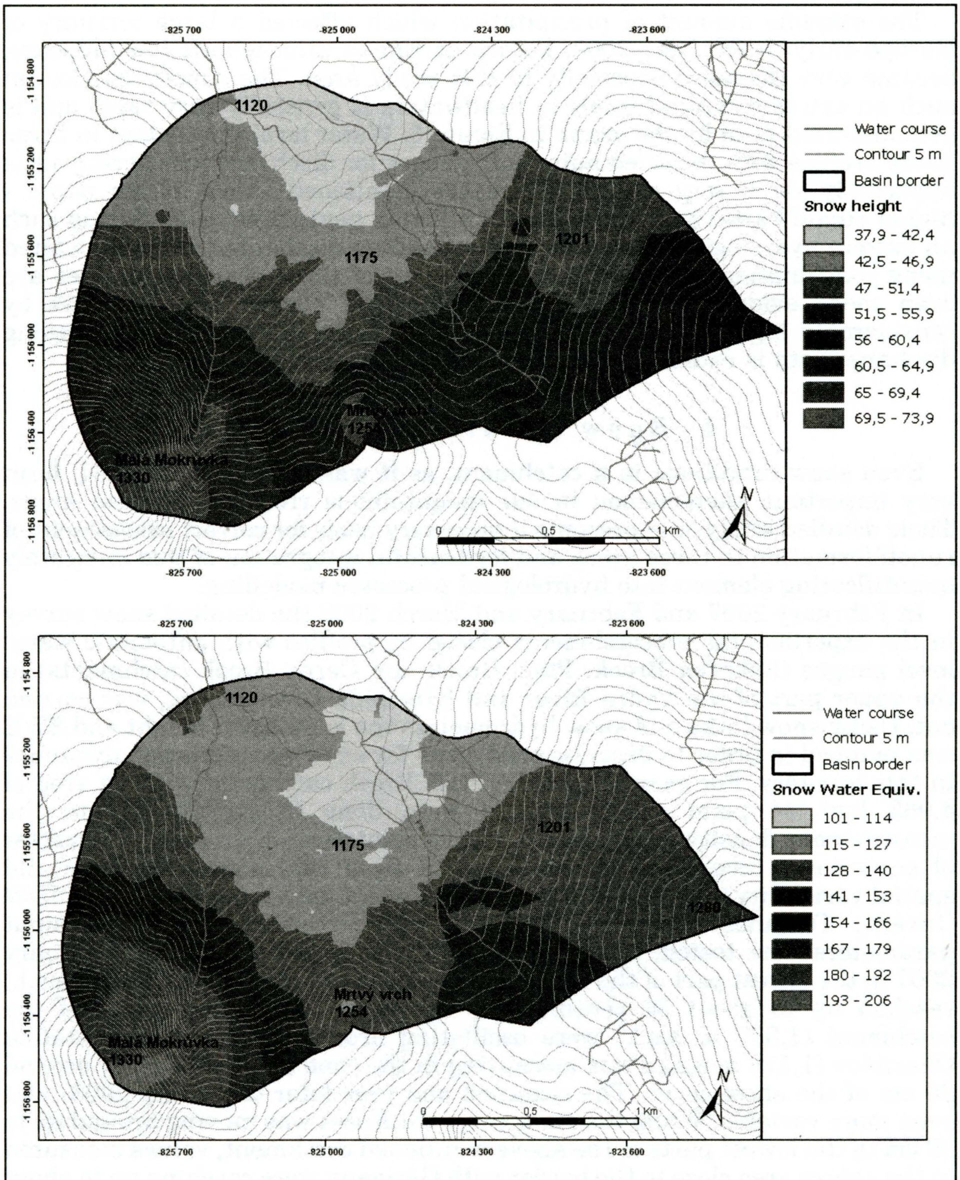

Fig. 8 – Spatial distribution of snow cover height (top, in cm) and snow water equivalent (down, in mm) in the Ptačí Brook catchment on the 13 February 2007 (M. Jeníček). Data: Database DMÚ25 – VGHMÚř (http://geoportal.cenia.cz), DIBAVOD WRI. Coordinate system: S–JTSK.

On the base of the runoff analysis in the Rokytka Brook profile it could be stated that 2007 hydrological year was very specific from the occurrence and amount of snow storages point of view, in the same way as the snow melting runoff. Above-average air temperatures, especially in January 2007, did not make the creation of usual amount of snow storages possible and represented the cause of markedly above-average runoff in this month. Year 2008 was from the character of winter and spring runoff point of view much closer to an average. Because of the fact that last two winter periods were in term of snow storages below the average and therefore inconvenient to detailed analyses and qualified conclusions the thorough field survey and monitoring of snow cover and its evolution dynamics in following snow period, especially during the process of snow melting in spring months, is considered to be continuing. In following years remote sensing methods for snow cover estimation is planned to be used.

The above described facts are the most important aspects that are necessary to be taken into account within the process of calculating the total water volume retained in snow storages. It is needed to consider the fact that on the base of data from CHMI climatic stations it is not possible to assess the real state of snow conditions in the study catchment in order to determinate the most accurate prognosis of runoff from snow cover. Partial analyses of correlation between a snow cover height, respectively SWE and altitude together with other physical-geographic factors acknowledge our hypothesis that spatial distribution of a snow cover including all its parameters is very variable, especially in mountainous basins.

4. Conclusions

All of the issues related to various possibilities and measures leading to river headstream areas retention capacity increase should be discussed by experts in various fields taking into account objectives and priorities of a regional and local significance (Buček, 1998; Knapp, 2000; Kolejka, 2003). Such a discussion could result for example in the introduction of suitable landscape elements or gradual modification of land use in areas playing various roles within the flood control (Kovář, Sklenička, Křovák, 2002). However, this cannot be applied to national nature reserves that should be left free of any human interventions.

Present outcomes from automatic water level gauges installed in the study basin of upper Otava River persuade us of the fact that data measured this way make it possible to assess peat bogs hydrological function very in detail. Especially comparison of those parts of catchment where revitalizing measures took place, respectively other parts where ameliorative adjustments of mountainous peat bogs were implemented in the last 70s, needs to be carried out. Continual records of water level and corresponding discharge values offer an extraordinary database for detailed analyses of flood waves ascending and descending phases, respectively for assessment of peat bogs and peat forming soils influence on runoff process during dry periods. Qualified conclusions from field survey can be formulated after analyses of longer data time series. Nevertheless, partial outcomes from present studies quite conclusively present more distinct runoff variability in profiles closing catchments with very significant peat land proportion mainly with respect to higher frequency of peak flows. Negative effect of peat bog localities on river headstream area hydrological regime was confirmed by thorough study and comparison of runoff reaction to causal rainfall totals in experimental catchments as well. Longer reaction interval adverting to more significant causal rainfall amount retention in the catchment was determined in the case of the catchment with less peat land proportion.

The problem of peat bogs hydrological function depends on a number of factors, especially on its type, health state, rate of anthropogenic impact, etc. The assessment of peat land hydrological regime and considering the influence of chosen physical-geographic factors is currently being carried out. At present the determination of peat bog revitalizing measures influence on their runoff regime is in process. The peat land influence on hydrological regime is being considered also with respect to the ionic content and carbon and oxygen isotopes balance of water in periods of low or high discharges. First results support above mentioned claims and outcomes. In addition to considering dyking of former drainage channels and focusing on recovery of vegetation health state having a positive influence on retention capability in a catchment the possible former accumulation reservoirs (used for wood floating in the past) restoration with potential function as dry polders is in the process of evaluating (Janský 2006; Janský, Kocum 2008). Using complex system of hydrological models with semi-distributed approach the simulation of runoff process and the assessment of the effectiveness of these reservoirs could be made. Implementation of such unforceable measures could contribute to reduction of peak flows and to increase of water resources during extreme droughts in future. In addition, the running of mentioned profiles with installed automatic gauges within the flood warning system in cooperation with CHMI is cogitated.

Partial results from the snow conditions survey in representative subcatchments confirmed the existence of a considerable variability in the amount of accumulated snow storage not just in relation to an altitude but also to a vegetation cover and other factors. Along with this fact, acquired data in the form of graphical outcomes prove very significant variability of snow storages on different localities of a similar altitude and character which gives an evidence about a high complexity of the snow cover accumulation process and its significant dependance on a number of physical-geographic factors. The above described facts represent the most important aspects that are necessary to be taken into account within the calculating of the total water volume retained in snow storages. Detailed field survey and monitoring of snow conditions including remote sensing methods are planned to be continuing during the next winter period.

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Shrnutí

VLIV RAŠELINIŠŤ NA ODTOKOVÝ PROCES: PŘÍPADOVÁ STUDIE POVODÍ VYDRY A KŘEMELNÉ NA ŠUMAVĚ (JIHOZÁPADNÍ ČESKO)

V souvislosti s výskytem katastrofálních povodní v povodí Odry a Moravy v roce 1997 a v povodí Labe v letech 2002 a 2006 se objevuje naléhavá potřeba řešení širokého komplexu otázek protipovodňové ochrany, a to nejen klasickými inženýrskými metodami, ale i netradičními postupy. V popředí zájmu společnosti se objevuje nová strategie ochrany před povodněmi zaměřená na postupné zvyšování retenční kapacity povodí včetně jeho pramenných oblastí. Všechny otázky související s rozmanitými možnostmi a opatřeními k dosažení tohoto cíle by měly být posouzeny na základě kvalifikované diskuze odborníků různého zaměření vzhledem k cílům a prioritám, které mají význam nadregionální, místní či lokální (např. Buček 1998, Knapp 2000, Kolejka 2003). Přirozený odtokový proces je ovlivňován člověkem již u samého jeho zrodu, tedy v pramenných oblastech toků, kde lze realizovat četné postupy související se zpomalováním odtoku a zvyšováním retence vody v území (Janský 2006; Janský, Kocum 2008).

Pro zvýšení retenčního potenciálu zdrojových částí povodí je třeba podrobně analyzovat hydrologickou funkci horských vrchovišť spočívající v detailním porovnání odtokových režimů v povodích s rozdílným stupněm zrašelinění a kvalifikovaně posoudit význam opatření souvisejících s hrazením původních melioračních kanálů. Velmi dobré podmínky pro realizaci tohoto výzkumu poskytují povodí Vydry a Křemelné na Šumavě (jihozápadní Česko), kde byly instalovány automatické ultrazvukové a tlakové hladinoměry a klimatické stanice s telemetrickým přenosem dat (obr. 3). Otázka vlivu rašelinišť na odtokový proces, především pak vlivu revitalizačních opatření rašeliništních lokalit na hydrologický režim toků, které je odvodňují, není přes řadu tuzemských i zahraničních projektů a mnohé spory vedené v odborných kruzích doposud uspokojivě vyřešena. Tzv. „houbová teorie“, která byla v domácí i světové literatuře uznávána přibližně do 60. let 20. století, předpokládala, že rašeliniště zadržují vodu při vysokých srážkách (povodňových průtocích) a naopak v období sucha průtoky nadlepšují a podílejí se tak na vyrovnávání odtoku. Od 70. let 20. stol. se objevují práce, které retenční funkci rašelinišť zpochybňují a za jedinou možnost zvýšení jejich retenční kapacity doporučují snížení hladiny podzemní vody pomocí odvodnění. Tyto meliorační zásahy byly poté provedeny v řadě horských oblastí Česka. Problematika odvodnění, resp. hrazení původních melioračních kanálů odvodňujících rašeliniště, se od té doby stala podnětem pro široké diskuse v rámci zahraniční i tuzemské odborné literatury (Holden et al. 2004).

Na podkladě dosavadních výsledků, grafických výstupů i statistických analýz, lze usoudit, že variabilita odtoku je vyšší v případě profilů uzavírajících povodí s významným podílem zrašelinělých a rašeliništních půd než závěrových profilů povodí s minimálním zrašeliněním. Fakt, že existence rašelinišť má negativní vliv na odtokový proces, zvláště během extrémních hydrologických situací jako jsou povodně a období sucha, byl dokázán jednak detailní statistickou analýzou průměrných denních a měsíčních průtoků ve státních profilech Vydra–Modrava a Křemelná–Stodůlky (obr. 1, 2 a tab. 1), a jednak porovnáním odpovídajících si vzestupných a poklesových fází odtoku v experimentálních profilech s různým stupněm zrašelinění (obr. 4). Variabilita odtoku je přitom kromě absolutní hodnoty kulminace definována především četností výskytu kulminačních průtoků. Rozdílná vodnost toků ve sledovaných profilech je zohledněna použitím poměru Q/Q_p, přičemž Q je okamžitý desetiminutový průtok a Q_p je aritmetický průměr z řady desetiminutových průtoků za celé sledované období. Podrobným rozborem jednotlivých fází odtoku, konkrétně analýzou reakce odtoku na příčinnou srážku (doba mezi výskytem max. desetiminutového úhrnu srážek a odpovídajícím kulminačním průtokem) během několika srážkových situací v rámci sledovaného období, bylo zjištěno významnější zpoždění kulminace odtoku a tedy vyšší schopnost retence vody v povodí s výrazně nižším zastoupením rašelinišť. Podobných výsledků bylo dosaženo při analýze příčinných srážek a odtokové odezvy u několika epizod povodňových průtoků ve třech dílčích experimentálních povodích – Rokytky, Ptačího potoka a Černého potoka (obr. 5 a 7). Extremita odtoku byla i zde přímo závislá na stupni zrašelinění daného povodí. Hydrologický monitoring v modelových povodích je doplněn i sledováním iontové bilance rašelinišť včetně bilance uhlíku a izotopů kyslíku za účelem detailní separace jednotkového hydrogramu pomocí aniontové deficience.

Důležitým fenoménem tvorby odtoku v pramenných oblastech českých toků je v celkové roční bilanci sezónní sněhová pokrývka, která představuje v prostoru a čase poměrně těžko kvantifikovatelný prvek. Charakter jejího výskytu se vyznačuje vysokou mírou časové a prostorové variability (obr. 8). Sněhoměrná pozorování v experimentálních povodích potvrdila existenci významného rozdílu v množství akumulovaného sněhu nejen v závislosti na nadmořské výšce, ale rovněž na vegetačním pokryvu, zejména mezi otevřenými plochami a lesem. Jedná se o okolnost, kterou lze jen obtížně postihnout použitím výsledků měření prováděných běžně na meteorologických stanicích. Zároveň byla dokázána i velmi výrazná variabilita v množství sněhu na různých stanovištích obdobného charakteru ve srovnatelné nadmořské výšce, což svědčí o značné složitosti procesu akumulace sněhové pokrývky a jejím významném ovlivnění značným množstvím faktorů. Z analýzy odtoku v profilu Rokytka–pod Rokyteckou slatí vyplývá, že hydrologický rok 2007 byl z hlediska výskytu a množství sněhové pokrývky značně specifický, stejně jako odtok z následného jarního tání. Nadprůměrné teploty vzduchu zejména v lednu 2007 neumožnily tvorbu obvyklého množství zásob sněhu a byly příčinou výrazně nadprůměrného odtoku v tomto měsíci. Rok 2008 se z pohledu chodu odtoku během zimního a jarního období daleko více blížil průměrnému stavu. Výsledky sněhových kampaní by měly výrazně pomoci při zpřesnění odhadů zásob vody vyskytujících se ve sněhové pokrývce a pro následné simulace odtoku z ní za účelem precizace hydrologických předpovědí.

Obr. 1 – Lokalizace povodí Vydry a Křemelné se státními profily ČHMÚ a experimentálních povodí s instalovanými automatickými hladinoměrnými a srážkoměrnými zařízeními v rámci pramenné oblasti Otavy (závěrový profil Otava–Rejštejn). V legendě shora: hladinoměr PřF UK, limnigrafická stanice ČHMÚ, srážkoměr PřF UK, sídlo, vodní tok, hranice hlavního povodí, hranice experimentálního povodí.

Obr. 2 – Dlouhodobé průměrné denní (Q_d) a měsíční průtoky (Q_m) v profilech Vydra–Modrava a Křemelná–Stodůlky v období 1.11.1999–31.10.2006 Osa x – měsíce, osa y – průtok ($m^3.s^{-1}$).

Obr. 3 – Výstup z automatického ultrazvukového hladinoměru a člunkového srážkoměru – reakce odtoku na příčinnou srážku v profilu Rokytka (pramenná oblast Vydry) v období 1.10.2006–20.5.2007 (v období 1.1.2007–28.3.2007 byl srážkoměr z technických důvodů mimo provoz). Osa y vlevo – úhrn srážek (mm), osa y vpravo – průtok ($m^3.s^{-1}$).

Obr. 4 – Porovnání variability odtoku ve dvou experimentálních povodích s rozdílným stupněm zrašelinění (Rokytka, Ptačí potok) v období 1.8.2006–31.5.2007. Osa y – Q/Q_p.

Obr. 5 – Kolísání hladiny toku ve třech modelových povodích s rozdílným stupněm zrašelinění během extrémní povodňové situace v březnu 2008. Osa y – vodní stav (mm).

Obr. 6 – Konzumpční křivka v profilu Rokytka. Osa x – průtok ($m^3.s^{-1}$), axis y – vodní stav (mm).

Obr. 7 – Desetiminutové průtoky ve dvou modelových povodích s odlišným stupněm zrašelinění během extrémní povodňové události v březnu 2008 (vysoce zrašeliněné povodí je reprezentováno povodím Rokytky). Osa y – průtok ($m^3.s^{-1}$).

Obr. 8 – Prostorové rozložení výšky sněhové pokrývky (nahoře, v cm) a vodní hodnoty sněhu (v mm) v povodí Ptačího potoka dne 13.2.2007 (autor: Michal Jeníček). Horní mapa: v legendě shora: výška sněhové pokrývky v povodí Ptačího potoka, vodní tok, vrstevnice po 5 m, hranice povodí, výška sněhové pokrývky (cm). Dolní mapa, v legendě shora: vodní hodnota sněhu v povodí Ptačího potoka, vodní tok, vrstevnice po 5 m, hranice povodí, vodní hodnota sněhu (mm). Data: Databáze DMÚ25 – Vojenský geografický a hydrometeorologický úřad (http://geoportal.cenia.cz), DIBAVOD (VÚV T.G.M.). Souřadnicový systém: S–JTSK.

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PIOTR MIGOŃ

MAIN FEATURES OF GEOMORPHOLOGY OF THE SUDETES RE-ASSESSED IN THE LIGHT OF DIGITAL ELEVATION MODEL

P. Migoń: *Main features of geomorphology of the Sudetes re-assessed in the light of digital elevation model.* – Geografie–Sborník ČGS, 113, 4, pp. 400–416 (2008). – The Sudetes as a geomorphological region are distinguished by complicated spatial pattern of high- and low-altitude terrains and variable mean slope gradients across the range. Several conceptual models have been proposed to account for this variability, emphasizing the significance of planation surfaces, intramontane basins, climate-controlled landform generations, or differential uplift and subsidence. An analysis of a digital elevation model and maps derived from the model have allowed for re-assessment of some of those hypotheses and concepts. It confirms that differential tectonics explains best the morphological layout of the Sudetes, but its effects are superimposed on a variety of rock – landform relationships. Neither the model emphasizing the occurrence of tiered levels of relict planation surfaces, nor one assuming the widespread existence of distinctive landforms of tropical morphogenesis find support in the light of region-wide DEM analysis. The general landform pattern of the western part of the Sudetes differs from the one in the eastern part, the difference being the abundance of intramontane basins in the former.
KEY WORDS: mountain geomorphology – DEM – geodynamics – the Sudetes.

Introduction

The Sudetes (Fig. 1) constitute the north-eastern rim of the Bohemian Massif and attain the highest elevation (1,603 m a.s.l.) within it, surpassing the other marginal mountain ranges of Krušné hory Mts. and Šumava Mts. by 150–350 m. They are also distinguished by their internal differentiation into a number of separate geomorphological units, rising to contrasting altitudes and very much different from each other in terms of landform inventories and morphometric parameters of relief. For example, a high-altitude plateau of vast extent which typifies Krušné hory Mts. and continues uninterrupted for many kilometres along the main water divide is missing in the Sudetes. On the other hand, deep intramontane basins, often elongated or rhomboidal in outline, are a very specific feature of geomorphology of the Sudetes, which hardly has an equivalent in other mountain terrains of the Bohemian Massif. Differences in height between the basin floors and summit surfaces of the surrounding ranges not uncommonly approach 1 000 m, which almost equals the relative relief of the Sudetes as a whole, as related to the lowlands to the north and south of the mountains.

Numerous attempts have been made to explain this peculiar large-scale geomorphology and to decipher the long-term evolution of landscape of the Sudetes. Several conceptual models have been presented by Polish (Jahn 1953, 1980; Klimaszewski 1958; Walczak 1968) and Czech researchers

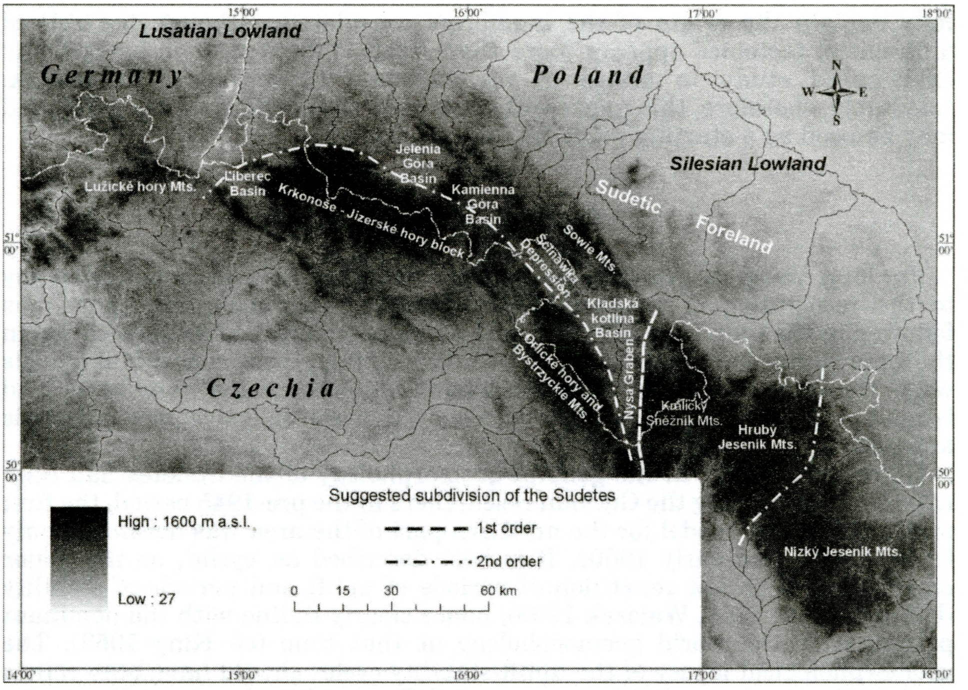

Fig. 1 – General hypsometry of the Sudetes, generated from the digital elevation model

(Czudek, Demek 1970; Kopecký 1972; Demek 1975, 1982), although the latter were devised for the Bohemian Massif as a whole, and these provided a framework adopted by subsequent workers in detailed studies in specific parts of the Sudetes. However, limited trans-boundary cooperation existed until the early 1990s and particular geomorphic models have been proposed using the evidence from either Polish or Czech side of the Sudetes, but hardly from both sides of the state border at the same time.

In the last 20 years or so an interest in gross geomorphology of the Sudetes diminished, as reflected in the decreasing number of publications addressing the issues of general geomorphology and the long-term landform evolution. Research efforts have been shifted to medium and minor landforms such as structural landforms or the geomorphic legacy of cold climate conditions, to small-scale indicators of neotectonics, or to detailed studies of late Quaternary history, chiefly of mountain glaciation. Perhaps the growing realization of insufficient database, subjectivity and over-reliance of the validity of initial assumptions have all contributed to this declining interest in region-wide geomorphology (see Migoń 1998). However, progress in research on Cainozoic deposits around the Sudetes and the advent of digital elevation models allows for the return to the 'classic' subject and for re-visiting of certain concepts and models. This paper aims to offer a new look at the gross pattern of landforms within the Sudetes, using mainly the results of recent studies of a digital elevation model built for the entire Sudetes, across the state boundary. As a result of this exercise, a few well-embedded concepts of long-term landform evolution have been challenged. At the same time it needs to be emphasized that the approach presented here is insufficient to fully explain the origin of the landforms considered, although the role of differential tectonics appears more than likely. Likewise, it is beyond any DEM-based study to arrive at an adequate chronological framework. Inevitably, therefore, this paper will leave several questions unanswered and may be used as a starting point for future research.

Previous concepts

For long, the gross geomorphology of the Sudetes has been considered a key to the recognition of an evolutionary pathway this mountain range has followed in the Cainozoic, after the last phase of widespread sedimentation in the Late Cretaceous came to an end. Unfortunately, these conceptual models were based on a rather intuitive identification of major landforms and landscape facets, hardly supplemented by a more objective morphometric analysis.

Since the interest in the general geomorphology of the Sudetes had been rather limited among the German researchers in the pre-1945 period, the first major conceptual model for the northern part of the area was developed only in the 1950s and early 1960s. It is best described as 'cyclic', as the major emphasis was on the repetition of periods of uplift and periods of stability (Klimaszewski 1958, Walczak 1968), hence clearly in line with the dominant paradigm in the world geomorphology at that time (cf. King 1953). The geomorphological legacy of the uplift/stability cycles should have been relicts of planation surfaces of different ages at different altitudes. According to the model, a relict Palaeogene surface exists along the water divides, whereas younger, Miocene and Pliocene surfaces occur at lower altitudes. Indeed, tiered planation surfaces were claimed to be present in all geomorphic units in the Polish Sudetes. They were deemed to be the most important landscape facets of the Sudetes, which can be identified in each morphotectonic unit and correlated across the entire range.

A different approach was presented by Jahn (1980), who highlighted the role of changing climates throughout the Cainozoic and saw their legacy in the occurrence of a distinctive generation of palaeo-tropical landforms in the Sudetes. Although he continued to interpret the Sudetes in terms of progressive development of 'relief horizons' (= equivalents of planation surfaces from the earlier model, but evidently of higher relief), the emphasis shifted towards geomorphic features which were considered consistent with tropical relief of the present-day. Among them, intermontane basins acquired the major importance and were explained as products of concentrated deep weathering (etching) in initial topographic depressions. A notable comment was offered in this context: 'The Sudetic basins being the most important forms in the morphology of the mountains were formed in intertropical conditions. Their dependence on tectonics and lithology is indirect' (Jahn 1980: 21; italics from the present author). Step-like long profiles of rivers, tors and shield inselbergs, and sandstone plateaux would have been further geomorphic features inherited from Palaeogene to Miocene times and contemporaneous tropical and subtropical climates. Likewise, Czudek (1977) related to the concept of relief generations as advocated by J. Büdel and interpreted the watershed plains of Nízký Jeseník Mts. as mid-Cainozoic basal surfaces of weathering (etchsurfaces), inherited from warm and humid morphoclimatic conditions.

Rather surprisingly, neither the proponents of the cyclic model of the 1950s and 1960s nor Jahn later, attempted to decipher the morphotectonic structure of the Sudetes. Differential tectonics in the Tertiary was invoked by Klimaszewski (1958) and Walczak (1968), but no details were given and its spatial pattern remained unclear. This apparent neglect of endogenic factors stayed at odds with the parallel approach to the geomorphology of the Sudetes among the Czech researchers. Kopecký (1972) and then Demek (1975) strongly emphasized unequal uplift of the Sudetes towards the end of the Tertiary and correlated the most elevated mountain ranges with the most uplifted blocks. The view of Kopecký (1972, 1986), who argued for fold-like deformation of the Palaeogene surface and its survival almost intact, was heavily criticised (Ivan 1990) and largely abandoned. However, the concept by Demek (1975), in which fragmentation of the Sudetes into separate blocks, up- and down-faulted, is a key ingredient, has been adopted as a general framework to analyse the geomorphology of the Sudetes, although 'mega-folding' is not completely ruled out, particularly with regard to the southern foreland of the mountains (Demek 2004).

Somewhat parallel, an inquiry into the nature of the early Tertiary to Miocene surfaces followed. In their seminal paper, Czudek and Demek (1970) demolished the idea of 'peneplain' survival in the Bohemian Massif, including the Sudetes, arguing for the existence of an etchplain instead. Relicts of a surface moulded by deep weathering in the Tertiary and subsequent stripping have been described from the rolling to hilly terrain of Nízký Jeseník Mts. (Czudek 1977, 1995), the granite inselberg area around Žulova in the Sudetic Foreland (Demek 1976), numerous parts of the Sudetic Foreland in Poland (Migoń 1997), and their more problematic counterparts from the more elevated massifs, including the summit parts of Krkonoše Mts. (see Migoń, Pilous 2007).

Back to the geomorphic research in Poland, studies carried out in the 1990s largely disassociated from the former views and interpreted several individual mountain massifs as block-faulted terrain units, broken into arrays of tectonic steps and half-grabens (Migoń 1991; Krzyszkowski, Pijet 1993; Migoń, Potocki 1996; Sroka 1997). A similar line of inquiry was followed by Ivan (1997, 1999). Results of morphotectonic analysis from these different massifs, taken together, have strongly suggested that the Sudetes are indeed a horst-and-graben type of relief, thus confirming the rather general view by Demek (1975), but remained to be integrated range-wide.

The very last developments concern the re-assessment of the role of lithology and variable rock resistance as factors contributing to the geomorphological diversity of the Sudetes. An independent evaluation of rock resistance, based on measurements of rock strength of representative rock complexes, indicates clearly that many facets of the present-day geomorphic landscape of the Sudetes may be satisfactorily explained by rock resistance contrasts but these seem to be subordinate to the more general relief pattern (Placek, Migoń 2007, Placek 2007).

Digital Elevation Model

Geomorphometric studies have long been considered as an important component of geomorphology, its main advantage being the ability to show landforms more objectively, via different primary and derived morphometric characteristics, and to pursue comparative analysis. Digital Elevation Models (DEM) are now standard tools in geomorphological research but their use in the Sudetes has been limited so far. One of the early examples was the geomorphological analysis of Lužické hory Mts. by Chvátalová (2000). Badura and Przybylski (2005) analysed shaded 3-D relief models of the Sudetes derived from a 1:50,000 DEM and highlighted several intriguing features of highly problematic origin, including possible meteoritic impacts (Przybylski, Badura 2004). In another study, the 3-D images assisted the morphotectonic analysis of the mountain front related to the Sudetic Marginal (Boundary) Fault (Badura et al. 2003), and later to delimit triangular facets along the mountain front (Badura et al. 2007). Likewise, Wojewoda (2007) employed DEM analysis in the morphotectonic studies of the central part of the Middle Sudetes. However, until recently DEMs have not been used to analyse the Sudetes as a whole, except for a recently published study of the distribution of low-gradient surfaces (Placek et al. 2007). This is in contrast to the wide use of DEMs in geomorphological studies elsewhere, where they proved extremely helpful in highlighting landscape features previously unaccounted for (e.g. Kuhlemann et al. 2005, Miliaresis 2006) or in supporting comparable quantitative analysis of large-scale landforms (e.g. Matmon et al. 2002).

Re-assessment of the large-scale morphology of the Sudetes offered in this paper is based on a DEM purposefully created in the Department of Geography and Regional Development, University of Wrocław[1]. The scale and spatial resolution were dictated by the aims of the study and the intention to identify landform patterns which would be significant regionally rather than locally. For the Polish side of the Sudetes, the DEM has been generated using the ArcMap software, from analogue topographic maps in 1:25,000 scale using manual vectorization method. The contour lines, with 25 m spacing, important elevation points and all drainage lines have been digitized. Subsequently, vectors were interpolated by the Topo-to-Raster tool to create a 50 m resolution raster. Then, the model has been supplemented for the Czech and German side by data from the DTED (Digital Terrain Elevation Data), available in 30 m resolution. The resolution and geographic coordinate system have been standardized and both models have been merged. The 50 m resolution and 25 m contour line have been chosen as sufficient and appropriate for further analysis at a regional scale. Gradient maps have been derived automatically, using the Spatial Analyst procedure (Surface Analysis tool in ArcGIS).

Main features of relief

Although geographical and geomorphological regionalization traditionally divides the Sudetes into a great number of smaller units and subunits (Czudek et al. 1972; Gilewska 1991; Kondracki 1994; Balatka, Kalvoda 2006), the regional DEM suggests a basic, bipartite division of the north-eastern rim of the Bohemian Massif. The western part includes an area from the Labe River in the west, through Lužické hory Mts., Krkonoše – Jizerské hory block

[1] DEM for the Sudetes was prepared by Wiesława Żyszkowska and Agnieszka Placek, within a project "Main features of geomorphology of the Sudetes in the light of geomorphometry and rock resistance, using GIS", coordinated by the present author and supported by the Ministry of Science and Higher Education through a research grant no. 3 P04E 021 23.

into the middle part of the range, as far as the mountain front separating the Nysa Graben and Králický Sněžník Mts. (Fig. 1). The eastern part includes the Králický Sněžník Mts., Hrubý Jeseník Mts., and continues to the Moravská brána Gate. The main differences between these two major units are the general style of relief and the percentage of low-altitude terrain within and around the most elevated blocks.

The West Sudetes (in the sense of the subdivision introduced above) are composed of a number of isolated elevated massifs of different size, separated by basins and intramontane troughs. The Krkonoše – Jizerské hory Massif and the massif of Orlické hory Mts. – Bystrzyckie Mts. stand out as the largest, compact blocks of roughly rectangular shape (Fig. 1). Evident mountain fronts provide geomorphic boundaries to these units, particularly on the northern side of the former and the eastern side of the latter. Across these lines, altitudes decrease from 900–1,100 m a.s.l. to as low as 350–450 m a.s.l. The respective southern and western boundaries of these two massifs are less clear, but associated with long stretches of deeply incised valleys, largely absent on the opposite sides. Several other massifs in the West Sudetes rise to 900–1,000 m a.s.l., but their spatial extent is much smaller and the outline less regular. A very specific and unique type of geomorphology is presented by the Lužické hory Mts. in the extreme west. Here, the relief is dominated by isolated hills and their clusters rising above the surroundings by as much as 300–400 m.

In between the elevated massifs numerous basins of different size occur. The largest of them is located in the eastern part of the unit. Traditionally, Kladská kotlina Basin, the Nysa Graben, and the Ścinawka Depression are distinguished here, but the relief map shows that they merge into one major topographic depression of triangular shape, pointing towards NE. In the western direction, more basins can be identified, including the distinctive rhomboidal Jelenia Góra Basin and the triangular Liberec Basin. However, other basins, such as the low-lying terrain around Kamienna Góra, are highly irregular in outline. Finally, the West Sudetes distinguish themselves by a wide belt of marginal uplands at 300–500 m a.s.l., particularly in the NW part. It is only the NE part where, along the mountain front related to the Sudetic Marginal Fault, the mountainous terrain of the Sowie Mts. falls down steeply to the Sudetic Foreland. In the remaining area the very location of the boundary of the Sudetes as geomorphological unit is problematic and somewhat arbitrary.

The geomorphological style of the East Sudetes (again, in the sense of the above subdivision) is very much different. Intramontane basins, which are so abundant in the west, largely disappear and those very few which exist (e.g. triangular basin around the town of Jeseník) are less distinctive. Consequently, the elevated terrain is much more close together and the passes separating individual massifs are located above 700 m a.s.l. At the same time, all these massifs are considerably dissected, with individual valleys reaching far into the cores of the mountains (e.g. in Králický Sněžník Mts. and along the eastern side of Hrubý Jeseník Mts.). Major topographic boundaries appear to strike N–S to NNE–SSW and delimit rectangular blocks.

Nízký Jeseník Mts. represents a geomorphological landscape in its own, without parallels elsewhere in the Sudetes. It is typified by an extensive, gently rolling plateau rising to 600–800 m a.s.l. Fluvial dissection does occur and clearly proceeds from the topographic margins of the area inward, but so

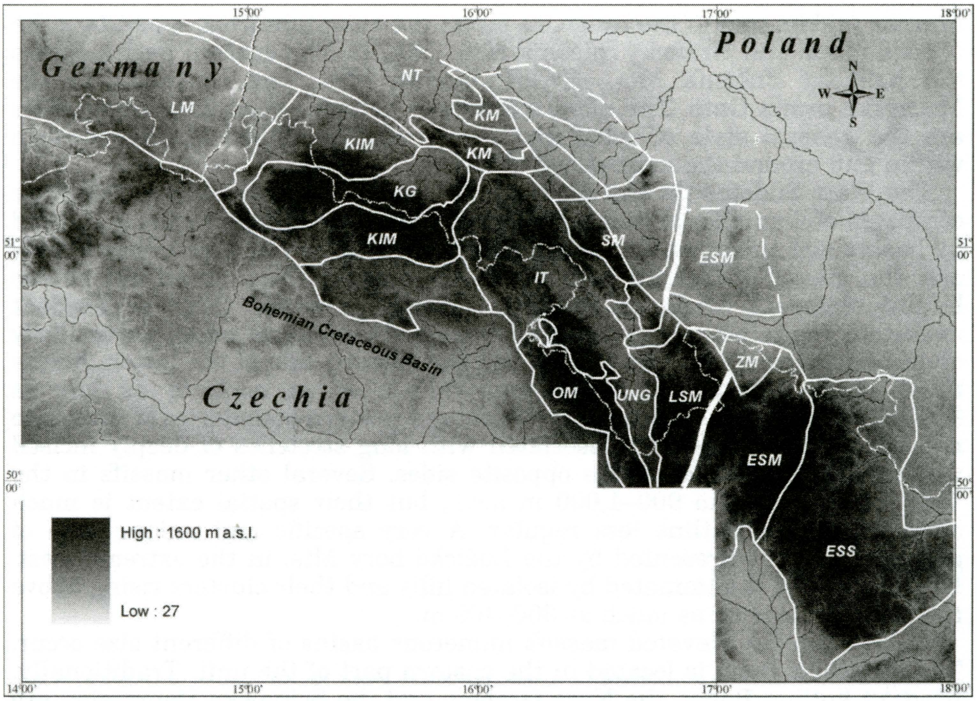

Fig. 2 – Main geological boundaries within the Sudetes superimposed onto the digital elevation model. The thick line indicates the boundary between the West Sudetic and East Sudetic terranes (after Żelaźniewicz 2005), broken lines mean an onlapping cover of Cainozoic sediments. Only major geological units are explained. LM – Lausitz Massif, KIM – Karkonosze-Izera Metamorphic Massif, KG – Karkonosze-Izera Granite Massif, NT – North-Sudetic Trough, KM – Kaczawa Metamorphic Unit, IT – Intra-Sudetic Trough, SM – Sowie Mts. Massif, OM – Orlica Metamorphic Unit, UNG – Upper Nysa Graben, LSM – Lądek-Śnieżnik Metamorphic Units, ESS – East Sudetic sedimentary Fold-and-thrust Belt.

far it failed to transform the plateau into a ridge-and-valley topography. Another feature of the East Sudetes is the rather limited extent of low-altitude (<400 m a.s.l.) uplands and foothills.

Interestingly, the above subdivision into two major units mirrors the general geological subdivision of the Sudetes, although the boundary between the two compartments is not identical (Fig. 2). The geological boundary between the West and East Sudetes, made by the Staré Město Zone, is located more to the east (Żelaźniewicz 2005) and goes through the mountainous terrain. Several SSW–NNE trending valleys roughly follow this structural line, but overall this important geological boundary is poorly revealed in the regional morphology and crosses water divides in many sections. Likewise, the prolongation of this zone in the northern foreland of the Sudetes does not correspond with any general change in morphology.

In addition to the division of the Sudetes as presented above, the existence of further morphological entities of lower order may be inferred from the relief and altitude map (Fig. 1). The West Sudetes appear to have a higher, inner part and a lower, outer part, facing the Silesian and Lusatian Lowland. Within the former, the proportion of high-altitude terrain is much higher, whereas in the latter it is only the Sowie Mts. where elevation exceeds 1,000 m a.s.l., and

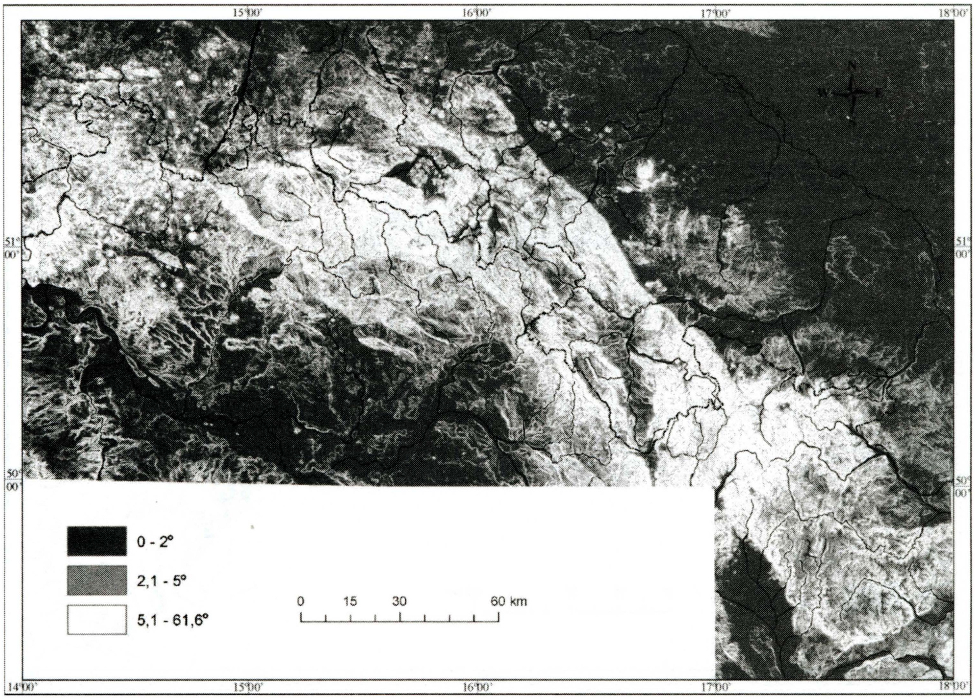

Fig. 3 – Spatial distribution of low-gradient slopes in the Sudetes and their surroundings. Mean slope angle calculated for 50x50 m squares.

even this only by 15 m. The dividing line may be traced along the western shoulder of the Nysa Graben, north-eastern scarp of the Cretaceous sandstone plateau (Stolové hory Hill country and Broumovské stěny Hill country), into the Kamienna Góra Basin, along the northern margin of the Krkonoše – Jizerské hory Massif and towards the western tip of Ještěd. Most of large intramontane troughs and depressions occur east and north of this line.

The East Sudetes in turn can be further subdivided into a major high in the north-west and an extensive low in the south-east. The boundary between the two runs roughly SW–NE, along the eastern footslope of Hrubý Jeseník Mts. and Zlatohorská vrchovina Hill country.

High-gradient and low-gradient areas

The potential of slope gradient maps in geomorphology is manifold. In the context of the large-scale geomorphology of the Sudetes, they have been used mainly to delimit objectively the extent of landforms which have been claimed highly significant for the entire area: planation surfaces and fault-generated escarpments.

The problem of the so-called planation surfaces in the Sudetes has been discussed at length elsewhere (Placek et al. 2007). Hence, only a summary of the main findings is presented here. Perhaps the most important observation derived from a partial map of slope gradients within the range 0–5° (Fig. 3) is the very limited extent of low-gradient surfaces at high elevation. Their

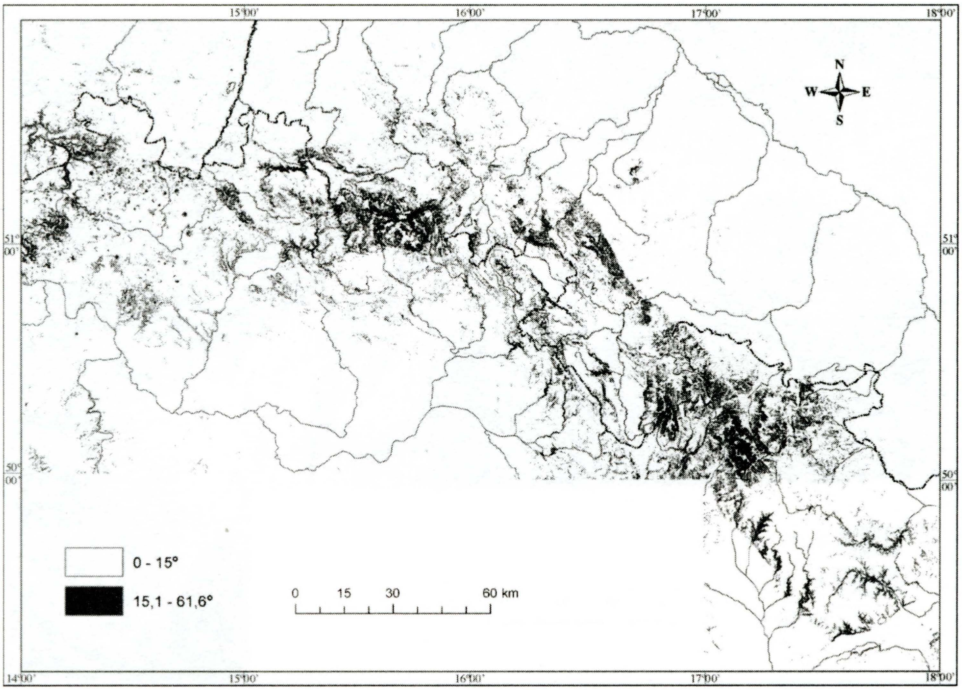

Fig. 4 – Spatial distribution of high-gradient slopes in the Sudetes and their surroundings. Mean slope angle calculated for 50x50 m squares.

absence is particularly pronounced in the most elevated part of the East Sudetes, from Králický Sněžník Mts. in the west to the Hrubý Jeseník Mts. In the West Sudetes the extent of 0–2° surfaces within the high altitude terrain is also small, but if the 0–5° is considered, then certain areas begin to show up. These include the Jizerské hory Mts., Stolové hory Hill country and, especially, Bystrzyckie Mts. Rather surprisingly, at the resolution of the DEM used, the visually impressive summit plains of Krkonoše Mts. almost fail to be revealed.

By contrast, low-gradient surfaces appear distinctively at low elevations, within intramontane basins, along some of the major rivers, and across marginal uplands. Altogether, they occupy c. 15 per cent of the Sudetes. The most extensive area of their occurrence is in the north-west. However, this is also the least elevated part of the Sudetes, with indistinct northern boundary, which has been the area of net deposition rather than erosion since the Miocene. Therefore, much of the observed flatness of the terrain is due to deposition of fluvial, lacustrine, glacial and outwash sediments, and less because of long-term denudation of the solid rock. Further areas with widespread occurrence of low-gradient watershed surfaces are the transitional uplands between the Krkonoše – Jizerské hory Massif and the Bohemian Cretaceous Basin, although these level terrain units are separated by deeply incised valleys of various tributaries of Jizera, Labe, Úpa and Orlice, and the upland of Nízký Jeseník Mts. The latter constitutes the most extensive area of low gradient in watershed position in the entire Sudetes.

Steep slopes (>15°) occur in different parts of the Sudetes, but with varying abundance (Fig. 4). To some extent, the >15° gradient map is a negative image of the 0–5° map, in the sense that the Krkonoše Mts. and the area from Králický Sněžník Mts. in the west to the Hrubý Jeseník Mts. in the east are revealed as belonging mostly to this class of slope gradient. Hence, in the Sudetes the highest altitudes coincide with the steepest average gradients. A few smaller areas scattered across the Sudetes show up in a similar way (e.g. Ještědský hřbet Ridge, Kamienne Mts. and Sowie Mts.).

Another distinctive spatial pattern of steep slopes is that of rather narrow elongated zones, which in some instances can be traced over tens of kilometres. These indicate two different landscape elements. Some are sinuous escarpments supported by gently dipping sedimentary rocks and are best seen in the central part of the Sudetes. Others are associated with plateau margins and are steep segments connecting the plateaux with the floors of intramontane basins. The northern margin of Jizerské hory Mts. and the western boundary of the Nysa Graben provide instructive examples. Interestingly, the north-eastern mountain front associated with the Sudetic Marginal Fault is not as evident on the gradient map as might be expected. In particular, the visibility of its north-western sector is very poor, indicating a much subdued escarpment and hence, probably little ongoing uplift. This, however, stands in contrast to the very low sinuosity (Krzyszkowski et al. 1995; Badura et al. 2003, 2007), typically taken as a proxy of very active tectonics along the mountain front (Bull, McFadden 1977).

Yet another setting in which considerable gradients occur is along fluvial valleys incised into the plateaux or sloping surfaces. These are most evident in the Czech part of the Sudetes, chiefly along the foot of Orlické hory Mts. and inward from the mountain fronts of the Nízký Jeseník Mts. Steep valley sides may line the valley floors by as long as 15–20 km.

Tectonic, lithologic and climatic influences: discussion based on DEM analysis

The present-day morphology of the Sudetes is a transient product of protracted geomorphological evolution, the onset of which is usually placed at the Cretaceous/Palaeogene boundary and related to the ultimate withdrawal of the Cretaceous sea (Klimaszewski 1958, Walczak 1968, Demek 1975, Jahn 1980). During the ensuing 65 or so million years the area experienced changes in stress field and tectonic regime, as well as multiple changes of climatic conditions, from warm and humid to very cold in the Pleistocene. Hence, the contemporary landform pattern is assumed to reflect tectonic, lihologic and climatic influences. This general assumption has never been challenged, but there have been widely divergent opinions expressed as to which controlling factors have been dominant and which have been rather subordinate.

The analysis of gross topography of the Sudetes strongly suggests that the prime control on the landform pattern is tectonic. Several points can be made here. First, the regional pattern of high altitude and low altitude areas is poorly correlated with lithological boundaries and variable rock resistance, although exceptions do exist (Placek et al. 2007). The most notable one is the presence of the residual volcanic range of the Kamienne Mts. in the central part of the Sudetes which creates local relief up to 400–500 m in the apparent absence of differential uplift and subsidence. On the other hand, the Krkonoše – Jizera a granite massif is divided into a low altitude area in the north-east and an elevated plateau in the south-west, with the relative relief well in excess of 500 m. Second, there is a correlation between altitude and the degree of dissection, which is consistent with a general rule that more uplift (in terms of both amplitude and rate) enhances erosion, which, followed by mass movement, leads to the steady-state mountainous landscape with little previous topography preserved (e.g. Adams 1985). All areas in the Sudetes rising above 1,200 m a.s.l. may be interpreted in this way. The best preserved planation surfaces in the mountain-top setting are associated with less elevation, from 600–800 m a.s.l. (Nízký Jeseník Mts.) to 1,000 m a.s.l. (Jizerské hory Mts., Bystrzyckie Mts.). Third, in the spatial distribution pattern of steep slopes large rectangular or rhomboidal structures are seen, which strengthens their interpretation as uplifted blocks delimited by fault zones. DEM analysis gives little support to the 'cyclic' concept of relief development of the Sudetes which relied on the alleged step-like occurrence of planation surfaces at different elevations.

Referring to the division of the Sudetes into the eastern and western part, introduced earlier in this paper, again tectonics is suspected to be the factor behind it. However, why grabens and other subsidence basins are fairly abundant in the West Sudetes and so scarce in the East Sudetes cannot be answered satisfactorily here. One possible reason is the decreasing strength of the lithosphere in the West Sudetes, related to regional variations in the heat flow. In the west it is by c. 20 mW/m^2 higher than in the eastern part of the Sudetes (Jarosiński, pers.comm.). Lithology may be an additional factor and several basins in the West Sudetes, particularly the smaller ones, have formed in weaker bedrock (Placek et al. 2007). Likewise, considerable lithological contrasts between adjacent geological units may bear on the mechanical heterogeneity of the area, which has been revealed in the 'neotectonic' period. The elevated and only marginally dissected terrain of Nízký Jeseník Mts. is puzzling, given its predominant lithology which is rather weak slate. No parallel landscape exists elsewhere in the Sudetes. Its relatively high altitude may be related to the forebulge position in respect to the Carpathians, although it is hardly consistent with the magnitude of lithospheric flexure actually decreasing westwards. On the other hand, as a lithologically homogeneous unit (broadly speaking) this crustal block may have been less prone to localized differential uplift and subsidence.

It remains an open question whether the observed topographic pattern does have any cause-and-effect relationship to the pre-Variscan and Variscan tectonic history, and to the terrane distribution in the Sudetes, as argued for Demek (2004). One problem here is that terrane boundaries, particularly in the West Sudetes, are poorly agreed (Aleksandrowski 2003, Żelaźniewicz 2005) and hence, difficult to correlate with morphology. In addition, the patterns of post-Variscan (late Carboniferous – Cretaceous) sedimentation and tectonic history in the intramontane troughs of the Sudetes apparently disregard the supposed terrane boundaries. Consequently, it remains unclear why should they not have been active in the late Palaeozoic and active again in the late Cainozoic. Thus, the influence of the very ancient tectonic features is, at best, indirect and secondary.

Lithology and local structure play an important, but rather subordinate part, accounting for second-order geomorphic features such as volcanic residual hills, inselbergs in the Sudetic Foreland, cuesta escarpments and numerous river gorges (Placek, Migoń 2007). It is also worth noting that the most elevated parts of the Sudetes are mainly supported by rocks which are mechanically strong. It is thus probable that higher rock strength is crucial in sustaining high relief and delaying erosional dissection, particularly in the granite area of Jizerské hory Mts.

The presence of climate-controlled relief generations is difficult to confirm or reject through an analysis of DEM alone. However, if the strong case for the role of differential tectonics is taken into account (also Sroka 1997; Badura et al. 2007), coupled with the results of other studies emphasizing the role of variable rock resistance at the medium scale (Placek, Migoń 2007), then the relevance of the climatic geomorphology framework to explain the present-day geomorphology of the Sudetes becomes highly dubious. The consequence of the scarcity of pre-Quaternary weathering residuals and deposits in the Sudetes and the uncertainties associated with dating of those which occur (Jahn et al. 2000), is our highly limited ability to verify and/or falsify the hypotheses of tropical inheritance.

The former emphasis on intramontane basins due to tropical deep weathering and their alleged NW–SE alignment to form the morphological axis of the Sudetes (Jahn 1980) is justified only for the West Sudetes. In the East Sudetes, no comparative features exist and the presence of inherited, palaeo-tropical landforms is uncertain[2]. In addition, the basins likely have more than one origin. The floor of the Jelenia Góra Basin has indeed formed with a substantial contribution from deep weathering (also Migoń 1992, 1999), but straight courses of topographic boundaries of the basin strongly point to the relative subsidence. Basins in the central part of the Sudetes, including the Kladská kotlina Basin and connected troughs to the south and north-west, do not bear any clear evidence for deep weathering. Rock-landform relationships suggest long-term rock-controlled denudation (Placek et al. 2007), superimposed on the more general pattern of uplift and subsidence. The best example of a basin due to subsidence is the Nysa Graben (Ranoszek 1999). One common evolutionary scheme for intramontane basins may simply not exist. The retrospective analysis of Jahn's (1980) paper shows how questionable conclusions can be reached if the transboundary position of the Sudetes is forgotten.

Conclusions

Geomorphology, as any other science, to maintain its status and scientific nature needs constant updating, critical testing and revisions of previously proposed concepts and models, however well they are embedded in the literature and minds of the geomorphologists. Revisions and re-assessments are usually done after new discoveries have been made, new paradigms have emerged, or new tools supporting research have been made available. This paper has offered an assessment of but one aspect of the geomorphology of the Sudetes, which is the general landform pattern within its geographical limits. The critical re-appraisal of previous concepts has been prompted by two circumstances: one is the increasing availability of digital elevation models and software used to work with them; another one is the strong personal feeling of the author that geomorphological research of the Sudetes can only

[2] Granite inselbergs around Žulová, considered as landforms belonging to the tropical generation of Palaeogene age (Demek 1976) are located in the Sudetic Foreland, not in the Sudetes proper.

advance if the mountains are considered as one entity stretching across the territory of Czechia, Poland, and Germany. Much of the research in the past was sadly restricted to an area on just one side of the border.

The DEM-based analysis of the large-scale geomorphology of the Sudetes allowed for a new look at the old problems as well as for a more objective test of certain concepts introduced in the past. The main conclusions from this exercise are the following:

– The general landform pattern of the Sudetes appears related to differential uplift and subsidence, as assumed by Czech geomorphologists already a few decades ago, however superimposed on a variety of rock – landform relationships arising from variable rock strength and resistance.
– The geomorphological styles of western and eastern part of the Sudetes differ from each other, the primary difference being the abundance of intramontane basins in the former. This may point to more effective Cainozoic extension in the western part, but reasons for this remain obscure.
– The most elevated (>1,000 m a.s.l.) massifs within the Sudetes are associated with the highest mean slope gradients. They are likely to be the loci of surface uplift, within the regional horst-and-graben structure of the Sudetes.
– Neither the model emphasizing the occurrence of tiered levels of relict planation surfaces, nor one assuming the widespread existence of distinctive landforms of tropical morphogenesis find support in the light of region-wide DEM analysis.
– Low-gradient surfaces at high elevation are of very limited extent in the Sudetes. However, they are considerably more widespread at 600–1,000 m a.s.l., particularly in Nízký Jeseník Mts., located in the Carpathians forebulge zone.
– Straight mountain fronts, usually taken as a proxy of active tectonics, are not necessarily associated with high slope gradients. Reasons for this apparent anomaly may reside in mechanical weakness of the footwall and its susceptibility to denudation and erosion, or in relatively low intensity of contemporary uplift.

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Shrnutí

HLAVNÍ RYSY GEOMORFOLOGIE SUDET PŘEHODNOCENÉ PODLE DIGITÁLNÍHO ELEVAČNÍHO MODELU

Sudety jako geomorfologická oblast Českého masivu se vyznačují komplikovaným prostorovým uspořádáním reliéfu o vysoké a nízké nadmořské výšce a různými průměrnými sklony svahů v celém rozsahu. V minulosti bylo navrženo několik konceptuálních modelů k vyhodnocení této rozmanitosti se zdůrazněním významu zarovnaných povrchů a cyklického vývoje reliéfu, mezihorských pánví jako pozůstatků tropické morfogeneze, vzniku tvarů reliéfu následkem klimatu či různých zdvihů a poklesů v neotektonickém období. Lze konstatovat, že v minulosti bylo učiněno jen velice málo pokusů o vytvoření uceleného modelu geomorfologického vývoje Sudet jako celku. Rámce používané českými a polskými gemorfology se liší, což vede k rozdílné interpretaci na pohled podobných tvarů reliéfu.

Cílem této práce je poskytnout nový pohled na formy reliéfu Sudet, zejména za použití výsledků posledních studií digitálního elevačního modelu vytvořeného pro celé Sudety z obou stran státních hranic. Pro polskou stranu Sudet byl vytvořen digitální výškový model (DEM) za pomoci software ArcMap na základě analogových topografických map o měřítku 1:25 000 pomocí ruční vektorizační metody. Byly digitalizovány vrstevnice po 25 m, významné výškové body a všechny vodní toky. Poté byly vektory interpolovány pomocí Topo-to-Raster a byl vytvořen rastr o rozlišení 50 m. Potom byly do modelu pro českou a německou stranu přidány údaje z digitálních modelů terénu (DTED) o rozlišení 30 m. Zvolené rozlišení a systém geografických souřadnic byly standardizovány a oba modely byly sloučeny. Jako dostačující a vhodné pro následnou analýzu na úrovni regionu bylo zvoleno rozlišení 50 m a vrstevnice po 25 m. Mapy sklonů byly odvozeny automaticky s využitím postupu Spatial Analyst (Surface Analysis v ArcGIS). Výsledkem tohoto postupu je možnost využití vhodných konceptů dlouhodobého vývoje reliéfu.

Hlavní druhy reliéfu Sudet jsou jasně spojeny s odlišným zdvihem a poklesem, což tvrdili čeští geomorfologové již před několika desítkami let. Analýza digitálního elevačního modelu pro celý region nepodpořila ani zdůrazňování výskytu uspořádaných úrovní reliktních zarovnaných povrchů, ani tvrzení o hojném výskytu zřetelných forem reliéfu tropické morfogeneze. Navíc analýza na základě digitálního elevačního modelu ukazuje, že geomorfologické složení západní a východní části Sudet se navzájem liší, přičemž základním rozdílem je hojnost mezihorských pánví v západní části. Mohou svědčit o účinnější třetihorní extenzi západní části. Nejvyšší masivy Sudet (přesahující nadmořskou výšku 1 000 m) vykazují nejvyšší průměrné sklony svahů a jsou pravděpodobně místy výzdvihu povrchu uvnitř regionální hrásťové a příkopové stavby Sudet. Existuje tedy korelace mezi nadmořskou výškou a stupněm disekce, která odpovídá obecnému pravidlu, že větší výzdvih (jak rozsahem tak stupněm) posiluje erozi, po které následují svahové pohyby, což vede k vytvoření stálého stavu horské krajiny, která si zachovala jen málo ze své předchozí topografie. Tímto způsobem mohou být vysvětleny všechny části Sudet nad 1 200 m n. m. Naopak mírně svažité povrchy jsou daleko častější v nadmořských výškách 600–1 000 m, zejména v Nízkém Jeseníku v předhůří Karpat.

Na závěr je třeba zdůraznit, že tento přístup nemůže plně objasnit původ studovaných forem reliéfu či zjistit jejich stáří. Otevírá však nové cesty výzkumu a předkládá pracovní hypotézy k dalšímu ověření. Přesvědčivým způsobem zejména ukazuje, že geomorfologický výzkum Sudet může pokročit pouze tehdy, když tato pohoří budou brána jako jeden celek prostírající se na území Česka, Polska a Německa.

Obr. 1 – Celková hypsometrie Sudet vytvořená z digitálního elevačního modelu. Navržené rozdělení Sudet.

Obr. 2 – Hlavní geologické hranice v Sudetech superponované na digitální elevační model. Tlustá čára označuje hranice mezi západosudetskými a východosudetskými terány (podle Żelaźniewicze 2005), přerušované čáry označují přesahující příkrov třetihorních sedimentů. Vysvětleny jsou pouze hlavní geologické jednotky. LM – Lužický masiv, KIM – Krkonošsko-jizerský metamorfní masiv, KG – Krkonošsko-jizerský granitový masiv, NT – Severosudetská brázda, KM – Kaczawská metamorfní jednotka, IT – Vnitrosudetská brázda, SM – masiv Sovích hor, OM – Orlická metamorfní jednotka, UNG – Hornoniská sníženina, LSM – metamorfní jednotka Lądek-Śnieżnik, ESS – Východosudetský sedimentární vrásno-přesmykový pás.

Obr. 3 – Prostorové rozložení svahů s nízkým sklonem v Sudetech a jejich okolí. Průměrný sklon svahu je počítán pro čtverce o straně 50 m.
Obr. 4 – Prostorové rozložení svahů s vysokým sklonem v Sudetech a jejich okolí. Průměrný sklon svahu je počítán pro čtverce o straně 50 m.

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HLAVNÍ ČLÁNKY – ARTICLES

ROZHLEDY „ REVIEWS

DISKUSE „ DISCUSSION

ZPRÁVY – REPORTS

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V přehledu uvádíme v abecedním pořadí recenzenty, kteří v roce 2008 recenzovali jednotlivé příspěvky v jednotlivých rubrikách časopisu Geografie – Sborník ČGS. Redakční rada jim všem děkuje za spolupráci.

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