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Driving role of snow cover on soil moisture and drought development during the growing season in the Czech Republic

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ABSTRACT: This study presents a detailed analysis on the role of snow cover during the cold season (October–March) on soil moisture deficit and drought development during the growing season (April–September) in the lowland and highland sites in the Czech Republic. Besides daily, weekly and seasonal series of basic snow-cover characteristics [the first day and the last day of snow cover, the number of days with snow cover (DSC), snow depth and snow water equivalent (SWE)] and soil water content measurements, six drought indices have been used in this study to quantify drought. Accumulations of years with significantly below average DSC/SWE were recorded in the early 1960s, mid-1980s, late 1990s and most of the 2000s. The trend towards an earlier end date of snow cover is found in both lowland and highland sites. However, the most significant shift in the dates of early end of snow cover has been identified to occur mostly in the hilly areas while in the lowland areas, these changes are not that evident. Liquid precipitation more than solid precipitation (snowfall) during the cold season lead to weakening correlation between SWE/DSC and the subsequent early summer (April–May–June, AMJ) soil moisture. Snow-cover characteristics can significantly influence soil water saturation during the first part of the growing season, while seasonal amount of SWE can explain up to 45% of soil moisture variability during AMJ season. More than 52% of dry AMJ followed after cold seasons with poor snow, and 42% of wet AMJ season followed after cold seasons with abundant snow. The strength of correlation between drought indices and soil moisture anomalies is higher in later summer. The negative anomalous snow characteristics in conjunction with winter and AMJ drought amplify lingering impact on the depletion of soil moisture in the later summer.

KEY WORDS anomalous snow seasons; snow-cover depth; snow water equivalent; snow phenology; soil moisture; drought indices (SPI, SPEI, PDSI, scPDSI, Z-index, scZ-index)

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1. Introduction

Snow cover is known to exert a strong influence on climate. It has major effects on surface albedo and energy balance, and represents a major storage of water (Vavrus, 2007). Many of these first-order effects are described by Cohen and Rind (1991), who emphasize the major thermodynamic influences of snow cover: high albedo, high emissivity, low thermal conductivity and latent heat sink. Snow-cover duration influences the growing season of vegetation at high latitude. A shortening snow season enhances soil warming due to increased solar absorption (Lawrence and Slater, 2010). Comprehensive studies on snow variability and their spatial patterns conducted at the Northern Hemisphere (NH) scale by Brown (2000) and Dye (2002) demonstrate that the snow cover is decreasing in response to recent warming. Brown and Robinson (2011) analysed the NH spring snow-cover variability and change over 1922–2010 and concluded that spring snow-cover extent has undergone significant reductions over the past 90 years and the rate of decrease has accelerated over the past 40 years. The observed trends in snow-cover extent are being mainly driven by warmer air temperatures, particularly NH mid-latitude air temperature. Modelling studies of Betts *et al.* (2014a, 2014b) on coupling of winter transitions to snow over Canadian Prairies point out that snow acts as a fast climate switch. Changes in snow phenology in the NH were found to show a significant trend for earlier snow-cover termination in Eurasia and virtually no trend in North America that was mostly associated with local temperature trends (Peng *et al.*, 2013). The study indicates that snow cover in the NH is very sensitive to rising temperature, and that there could be a positive feedback of snow-cover change to spring temperature.

In Europe, results on snow-cover variability have been reported either at European scale (Henderson and Leathers, 2010) or at the country level such as for Switzerland (Scherrer and Appenzeller, 2006; Marty and Blanchet, 2012), Austria (Hantel *et al.*, 2000), France (Durand *et al.*, 2009) and Italy (Valt and Cianfarra, 2010).

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In Finland, trends towards shorter snow season and greater snow depth have been noted by Hyvärinen (2003) and for Estonia by Jaagus (1997). In many studies, the timing and the amount of snow cover have been related to large-scale circulation patterns. Beniston (1997) found the North Atlantic Oscillation (NAO) influencing the timing and amount of snow in the Swiss Alps. Bednorz (2002) showed that the NAO exhibited strong negative correlations with the snow-cover duration over western Poland. Also, the NAO has been shown to influence the winter precipitation and snow cover in Poland (Bednorz, 2004). For Bulgaria, Brown and Petkova (2007) associated the years having high snow accumulation with a negative NAO pattern. Studies on snow variability in Romania (Birsan and Dumitrescu, 2014) show that all snow-related parameters display significant negative correlations with the NAO index for winter. Specifically, decreased frequency of seasonal snowfalls was found for stations in the Romanian Carpathians, as well as earlier snow melt in spring rather than late snow onset in autumn (Micu, 2009). In the Czech Republic (CR), climatology of snow has been produced (Tolasz *et al.*, 2007) and a snow-cover model for agroclimatological applications has been developed (Trnka *et al.*, 2010).

The linkage between snow mass, snow disappearance and soil moisture has been studied by Shinoda (2001), who suggests that soil moisture over Eurasia recharged by snow melt has a short memory due to large overland runoff over permafrost surface, and that its direct thermal effect on land surface temperature may not be as significant as previously thought. The ecological and hydrological impacts of snow cover are important for environmental and water-resource issues (Arslan *et al.*, 2006; Peng, *et al.*, 2010). Furthermore, recent and anticipated reductions in snow cover due to future greenhouse warming are an important topic for the global change community (IPCC, 2013).

Drought indices were often adopted for estimating soil moisture. Using the Palmer Drought Severity Index (PDSI) and the Standardized Precipitation Index (SPI) for estimating soil moisture, Sims and Raman (2002) found SPI to be more representative for short-term precipitation and soil moisture variation and hence a better indicator of soil wetness. PDSI has been used to assess the summer moisture variability across Europe (Briffa *et al.*, 1994; van der Schrier *et al.*, 2006a, 2006b) and European Alpine moisture variability (van der Schrier *et al.*, 2007), and the relationship between drought and soil moisture and the effects of surface warming at global scale (Dai *et al.*, 2004). Recent studies have demonstrated the impact of soil moisture on extreme maximum temperatures in Europe (Hirschi *et al.*, 2011; Whan *et al.*, 2015). It was argued that mega-heatwave temperatures are due to combined soil desiccation and atmospheric heat accumulation (Miralles *et al.*, 2014), and it was demonstrated that evapotranspiration amplifies the European summer drought (Teuling *et al.*, 2013).

Among other drivers of soil drying in the CR, Trnka *et al.* (2015a) have identified that the water from snow melting is critical for the recharge of soil moisture at the beginning of growing season. An analysis of snow-cover model data indicates an overall decline in the volume of water in the form of snow from December to April, although changes in the first 3 months of this period were insignificant. Statistically significant trends of decreasing soil moisture content were found, notably during May and June between 1961 and 2012 (Trnka *et al.*, 2015b).

This study based on observation data aims at analysing the relationships between snow-cover characteristics and soil moisture during the growing season (April–September) in the lowland and highland sites in the CR. The focus is on identifying the role of snow cover during the cold season (October–March) on soil moisture deficit during early summer (April–May–June, AMJ) and later summer (July–August–September, JAS). To accomplish these goals, the specific objectives of the research were to assess: (1) the tendency of snow-cover characteristics, (2) the relationship between the number of days with snow cover (DSC)/snow water equivalent (SWE) and the soil moisture for each month of the early growing season and (3) the correlation between drought indices and soil moisture for each month of early summer (AMJ) and later summer (JAS).

The paper is organized as follows. The data and the methodology used are described in Section 2. Section 3 presents the results which are then discussed in Section 4. The conclusions are highlighted in Section 5.

2. Data and methods

2.1. Data

Daily, weekly and monthly series of various climatic parameters at seven representative stations in the CR have been used in this study (Figure 1(a); Table 1). The selected climatological stations meet the following criteria: (1) provide simultaneous measurements of snow, SWE and soil moisture content, (2) representative of a specific region of the CR in terms of climatic characteristics and drought severity at various altitudes and (3) provide continuous observations during the warmest and driest decades in Central Europe (1991–2014). The Doksany agrometeorological observatory provides all data sets used in this study for the longest period available in the CR (1951–2014), while for the rest of six stations the daily data for snow and soil moisture are only available for the period 1991–2014, and the monthly data for temperature and precipitation are available for the period 1961–2014. For all stations, snow characteristics of the cold season 2014/2015 were also included. All the time series have been quality controlled and homogenized by the Czech Hydrometeorological Institute (CHMI).

2.1.1. Snow data

The snow data used in this study consist of daily snow depth (cm) and weekly SWE (mm) measurements. The meteorological measurements have been carried out at the

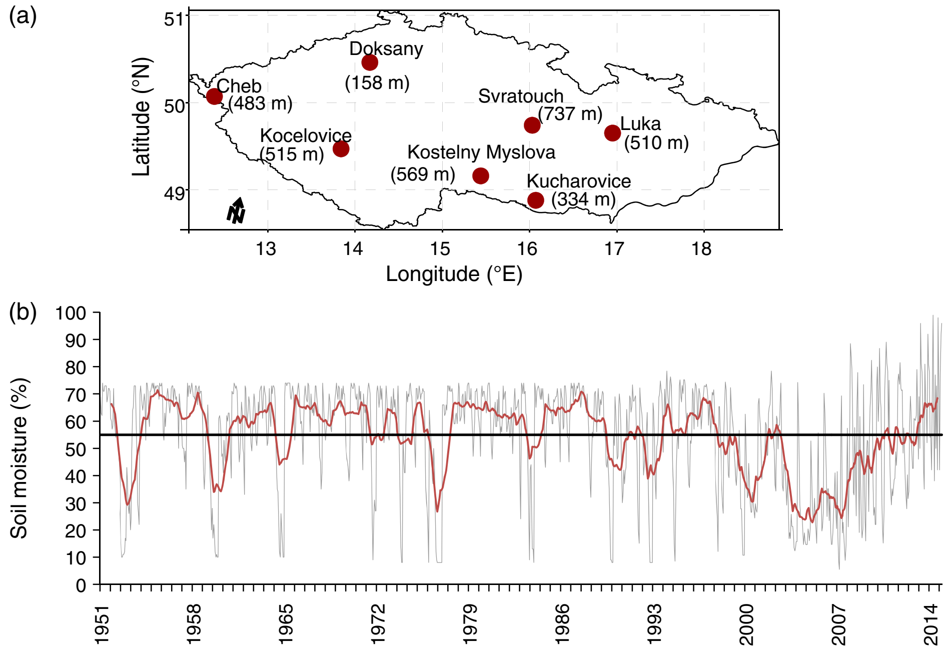

Figure 1. (a) Location of stations used for the calculation of basic snow-cover characteristics, anomalies in soil moisture and six drought indices for the territory of the CR. (b) Long-term evolution of daily soil moisture from April through September for the period 1951–2014 at Doksany. The series are smoothed with a 10-year Gaussian filter (red curve).

Table 1. List of the stations used in the study and their geographical and climatological features. Mean annual temperature and precipitation for the period 1971–2000.

Number	Station	Altitude (m)	Latitude (°)	Longitude (°)	Soil texture	Region	Mean annual temperature (°C)	Mean annual precipitation (mm)
1	Doksany	158	50.46	14.17	Sandy loam	Nord-West	9.1	442
2	Kucharovice	334	48.88	16.07	Clay loam	South	9.0	454
3	Cheb	483	50.07	12.39	Sandy loam	West	7.5	570
4	Luka	510	49.65	16.95	Loam	East	7.4	583
5	Kocelovice	515	49.47	13.84	Loam	West	7.6	570
6	Kostelny Myslova	569	49.16	15.44	Sandy loam	South-East	7.2	584
7	Svratouch	737	49.74	16.03	Loam	East	6.1	755

CHMI. The snow depth was measured daily at 0700 using a ruler and a snow-gauge panel according to the standard requirements of the CHMI. The SWE is a measurement of the amount of water contained in snow pack if there is an unbroken snow cover of 4 cm depth or more. It can be considered as the depth of water that would theoretically result if the whole snow pack instantaneously melts. SWE is the product of snow depth and snow density and it is presented in units of mm. Out of these data, the following parameters have been calculated: (1) DSC per month/cold season and (2) the first and the last day with snow cover.

A given day was counted as one with snow cover when the snow depth was ≥1 cm. The criterion for counting a given day as one with snow cover is a matter of regional characteristics of snow. Haiden and Hantel (1992) had used a 1-cm data set; Beniston (1997) considered snow depth thresholds from 1 cm up to 150 cm, whereas Hantel *et al.* (2000) used the threshold of at least 5 cm. This paper shall restrict the subsequent discussion to the threshold of 1 cm as a relevant choice for the lowlands of Central Europe and also conform to the CHMI standards for snow cover. The first day with snow cover was considered the first day at the beginning of the cold season in which the snow cover was ≥1 cm. The last day of snow cover was the last day at the end of the cold season when the snow cover was ≥1 cm.

2.1.2. *Soil moisture data*

Daily soil moisture content was measured in the 0–50 cm layer under grass cover. This top soil layer is mostly exposed to drought. Soil moisture content may be expressed on either a mass ratio kg kg^{-1} (kg water per kg soil) or as a volume fraction, m^3 m^{-3} (m^3 water per m^3 of bulk soil volume), respectively. In either case, the value is a dimensionless fraction and can be multiplied by 100 to express it as a percentage. In this study, volumetric soil water content (%, hereinafter) was used for soil moisture. Historical *in situ* measurements of soil moisture content in the CR are only available for a few stations, and are often very short (Možný *et al.*, 2012). In this paper, the unique longest daily soil moisture content measurements (Figure 1(b)) during the growing season at Doksany observatory covers the period 1951–2014 while at six selected stations high-quality soil moisture daily data are available for the period 1991–2014.

2.1.3. Monthly data for calculating drought indices

Monthly means of daily maximum and daily minimum air temperature have been used to calculate the monthly means of air temperature which together with monthly precipitation totals have been used to calculate a set of six most applied worldwide drought indices, described in the following section. The climatological data sets have been complemented with the values of maximum soil water-holding capacity which have been extracted from the digital soil map of the CR (Tomašek, 2000).

2.2. Methods

2.2.1. Drought indices

Besides daily soil water content measurements, a set of drought indices has been calculated in this study to quantify drought. The PDSI (Palmer, 1965) is among the most comprehensive and commonly used drought index (Dai *et al.*, 2004; van der Schrier *et al.*, 2006a, 2007; Dai, 2011). It incorporates antecedent and current moisture supply (precipitation) and demand (potential evapotranspiration) into a hydrological accounting system, and includes a two-layer bucket-type model for soil moisture calculations (Dai, 2011). A number of deficiencies have been reported for PDSI of which most significant are that (1) it is not comparable between diverse climatological regions, (2) it does not work well over mountainous and snow-covered areas, (3) it produces values that are ≥4 or ≤−4 up to 15% or more of the time and (4) it includes strong influence of calibration period (Dai *et al.*, 2004; Wells *et al.*, 2004).

To overcome most of the above mentioned drawbacks, a self-calibrated version of PDSI (scPDSI) algorithm has been developed by Wells *et al.* (2004) that replaces the empirically derived climatic characteristic and duration factors – which determine how sensitive the index is to precipitation and the lack thereof – with values automatically calculated based upon the historical climatic data of a location. The scPDSI as advanced by Wells *et al.* (2004) performed better than the original PDSI during the 20th Century over Europe (van der Schrier *et al.*, 2006a, 2007), North America (van der Schrier *et al.*, 2006b) and at global scale (Dai, 2011). However, some shortcomings of PDSI/scPDSI are still under debate in the drought researcher community (Vicente-Serrano *et al.*, 2011, 2014; Sheffield *et al.*, 2012; Beguería *et al.*, 2014) mainly because of its simplicity (Wells *et al.*, 2004) owing to the treatment of potential evaporation (the evaporative demand of the atmosphere) which is calculated from temperature data by using the empirical Thornthwaite equation (Thornthwaite, 1948), and because of its water balance model as it oversimplifies soil surface hydrological processes (Vicente-Serrano *et al.*, 2011).

In this study, the PDSI programme (http://greenleaf.unl.edu/downloads/) has been run which uses monthly temperature means, precipitation totals, available water content and station latitude as input variables, and the Thornthwaite algorithm to calculate potential evapotranspiration.

PDSI/scPDSI calculation also includes an intermediate term known as the Palmer moisture anomaly index (Z-index/scZ-index), which is a measure of surface moisture anomaly for a given month, without consideration of the antecedent conditions characteristic of PDSI. The Z-index/scZ-index may therefore be employed to track short-term deviations of soil moisture from the normal range and it is frequently used to estimate agricultural droughts, as the Z-index/scZ-index responds relatively quickly to changes in soil moisture. Dai *et al.* (2004) have demonstrated that the monthly Z-index and PDSI correlate rather well, during the warm season, with observed soil moisture departures across a number of regions on the NH.

The SPI is a simple index which was developed by McKee *et al.* (1993, 1995) to quantify precipitation deficits on multiple time scales. There are both strengths and weaknesses when using the SPI to characterize drought severity (Hayes *et al.*, 1999). The SPI has three main advantages. The first and primary strength is its simplicity. By avoiding dependence on soil moisture conditions, the SPI can be used effectively in both summer and winter. The second advantage is that SPI is not adversely affected by topography. The last advantage of SPI is its variable time scale. This temporal versatility is also helpful for the analysis of drought dynamics, especially the determination of onset and end, which have always been difficult to track with other indices. The SPI has three potential weaknesses, the first being the assumption that a suitable theoretical probability distribution can be found to model the raw precipitation data prior to standardization. A second limitation of the SPI arises from the standardized nature of the index itself; namely that extreme droughts measured by the SPI, when considered over a long time period, will occur with the same frequency at all locations. Thus, the SPI is not capable of identifying regions that may be more 'drought prone' than others. A third problem may arise when applying the SPI at short time scales to regions of low seasonal precipitation.

The Standardized Precipitation Evapotranspiration Index (SPEI) has been developed by Vicente-Serrano *et al.* (2010) to improve the original SPI concept. The SPEI is based on a monthly climatic water balance (precipitation minus evapotranspiration), which is adjusted using a three-parameter log-logistic distribution. The values are accumulated at different time scales (from 1 to 48 months) and converted to standard deviations with respect to average values. In this study, the parameterization of potential evapotranspiration was based on the monthly minimum and maximum air temperature, and extra-terrestrial radiation (Hargreaves and Allen, 2003). We used ProClimDB software (Štěpánek, 2010; http://www.climahom.eu/) to prepare the input data for the calculation of SPEI. This data was then put into the updated version of R package SPEI (Beguería *et al.*, 2014;

http://cran.r-project.org/web/packages/SPEI), and then the output was put back into the ProClimDB environment (see further description in Potopová *et al.*, 2015b). Some criticisms were addressed to SPEI as compared with PDSI arguing that SPEI does not represent soil water content (Dai, 2011). As 'the aim of the SPEI is to represent departures in climatological drought, the balance between the water availability and the atmospheric water demand, it is therefore slightly different from the drought indices that include a simplified soil moisture budget' (Vicente-Serrano *et al.*, 2015).

Hence, the six drought indices that have been calculated for this study are: PDSI and Z-index, scPDSI and scZ-index SPI and SPEI. The Z-index and scZ-index series account for moisture conditions for the current month, while the PDSI and scPDSI account for the current month's cumulative moisture conditions integrated over the last several months. The SPI and SPEI were used to examine the role of precursor moisture accumulation deficit, including the influence of dryness during the cold season on soil moisture deficit during early summer (AMJ) and later summer (JAS). To place a higher weight on past moisture conditions in the calculation of drought severity and their effect on soil moisture during AMJ and JAS, the multiscalar SPEI/SPI at 3-, 4-, 5- and 6-month lags have been used. The SPEI/SPI calculated for various lags contained the memory of moisture conditions prior to the current month. The magnitude of this memory is controlled by the time scale. To quantify the strength of relationship between SPEI (SPI) series at 3-, 4-, 5- and 6-month lags and soil moisture, Spearman rank correlation coefficients (r) have been calculated at these lags. SPEI/SPI computed over several time scales indirectly considers the effect of accumulating precipitation deficit and high potential evapotranspiration, which are critical for soil moisture. The 3-month lag incorporates moisture conditions from the current month and the preceding 2 months.

2.2.2. *Regression analysis*

The estimation of linear slope of a trend in the dates of onset and termination of snow cover, DSC and SWE is conducted with the nonparametric Kendal–Theil method (also known as Theil–Sen slope estimate), which is suitable for a linear trend in the variable x and is less affected by non-normal data and outliers (Helsel and Hirsch, 1995). As the main goal of this paper is to assess the driving role of snow cover on soil moisture during the growing season, the r was chosen as an indicator to explain the strength of the relationship between snow cover during the cold season and soil moisture deficit during the AMJ. Greater values of DSC and SWE occur in regions where more snowfall occurs regularly, and snowmelt during the cold season is less frequent. Hence, DSC and SWE for the cold season were used as main snow-cover parameters to explain the inherent mechanism through which snow-cover anomalies would produce soil moisture deficit or excess during the subsequent AMJ. In addition to that, it should be taken into consideration the fact that the snow accumulation expressed in terms of SWE is lower during the cold season in lowlands, and therefore precipitation during the cold season should be examined as an additional factor driving the soil moisture. The main statistical method applied to establish the relationship between snow characteristics and soil moisture during the growing season, and soil moisture and various drought indices is a linear regression model. In the regression model, soil moisture is the dependent variable while snow characteristics and drought indices are the independent (predictor) variables. The model performance was evaluated with the coefficient of determination (R^2). The suitability of the linear regression model ($y = a + b \times x$) was established in agreement with all these statistical indicators. In our analysis, the R^2 value shows the percentage of variability in the soil moisture values (y) explained by the x (snow characteristics and/or previous drought) variable, whereas the r measures the degree of association between the dependent (y) and independent (x) variables. Spearman's r is a nonparametric rank-based correlation coefficient used to estimate the monotonic association between two random variables. It is computed from the difference d between the ranks of independently sorted variables x and y (Kottegoda and Rosso, 1997).

3. Results

3.1. The annual cycle of snow-cover characteristics

The multiannual means of monthly snow-cover depth, DSC and SWE have been calculated and their annual cycle for seven stations situated in lowland and highland regions in the CR are illustrated in Figure 2. The snow-cover depth is mainly dependent on air temperature and the amount of winter precipitation and its characteristics. Besides, it is also affected by the station exposure to the prevalent air streams and solar radiation by the terrain, elevation and vegetation characteristics (Tolasz *et al.*, 2007). At the most snowy stations, the mean monthly snow-cover depth exceeds 50 cm from October to March, whilst the snow depths rarely exceed 20 cm at lowland locations with less snow. The seasonal variation in the snow depth average at eastern and western highland stations is contrasting (Figure 2, upper panel). At lowland stations, the snow depth mean increases moderately and reaches its maximum by mid-January, while for the eastern highlands its maximum is reached around the end of January and mid-February. For the western highlands, the snow depth starts to increase as early as October, its curve starts to climb more sharply at the end of November, attains its maximum around mid-February and then decreases by the end of March. Therefore, we can conclude that excessive snowfall in the early winter tends to reduce the absorbed solar radiation in winter by increasing the surface albedo, thus resulting in persistence of colder temperatures.

As shown in the middle panel of Figure 2, DSC during the cold season ranges from 34 to 52 in lowlands, reaches 66 at stations with elevations between 510 and 569 m a.s.l, and exceeds 90 at the snow-rich highland station (Svratouch). At lower elevation stations (Doksany and

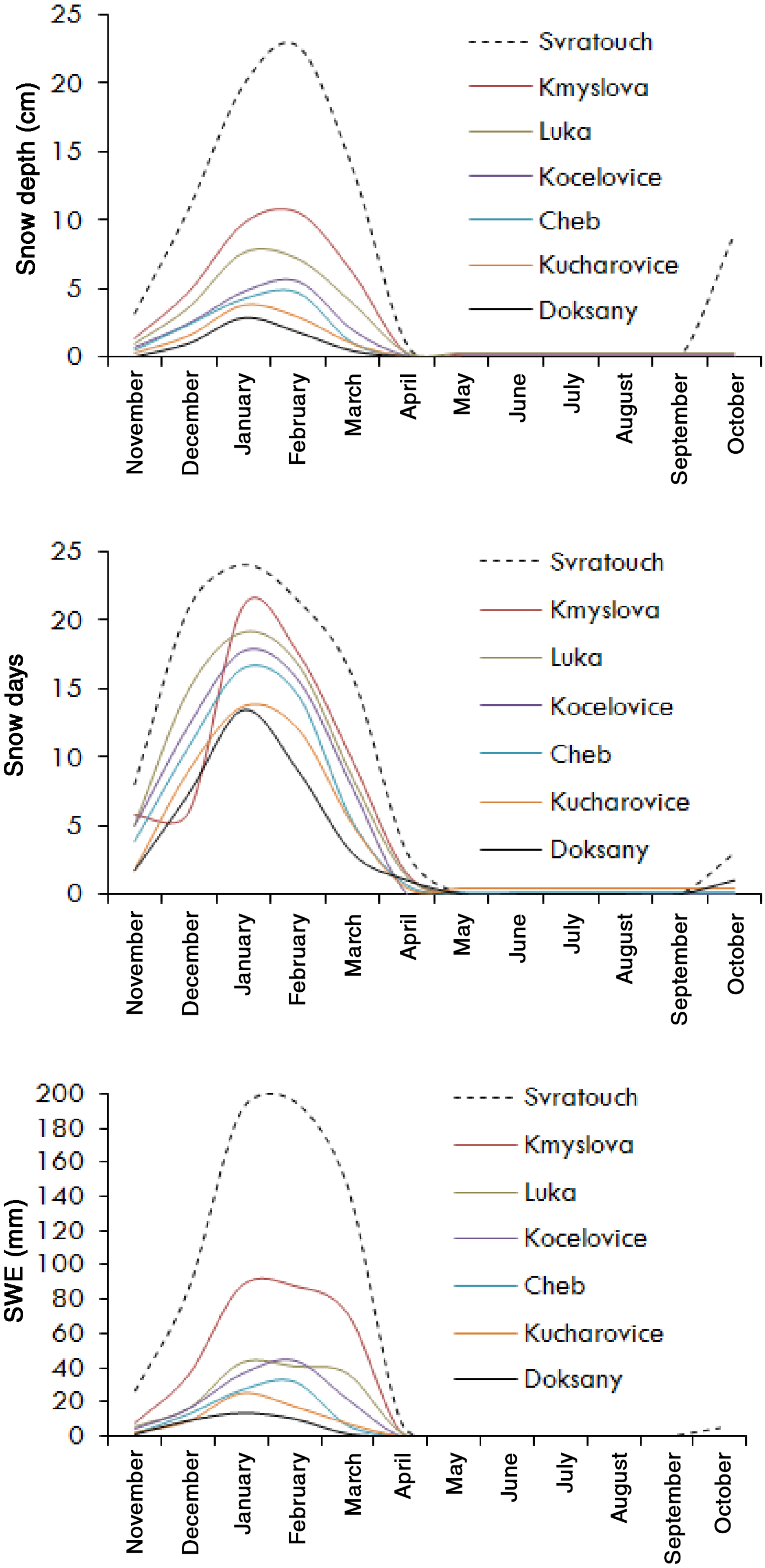

Figure 2. The multiannual cycle of the snow-cover depth (cm), DSC and SWE (mm) for seven stations situated in lowland and highland regions in the CR.

Kucharovice), January has the highest number of DSC, while at more snowy stations at higher elevation (Svratouch and Kostelny Myslova) there are no large differences between DSC during individual winter months.

The amount of SWE at the beginning and at the end of snow season is an indicative of both accumulation and speed of snow onset/melt (Figure 2, lower panel). At the beginning and at the end of the snow season, the mean values of SWE are greater at sites where snowmelt and snow accumulation occur more rapidly. Similarly, low values of SWE at the beginning or at the end of the snow season are indicative of less snow accumulation or slower snow onset/melt. There are stations with two peaks of SWE, the first maximum reached around mid-January and the second maximum around the beginning of March. The next group of stations presents only one peak of SWE in January, while for the remaining stations the maximum value of SWE occurs in February. Therefore, one can conclude that a clearly distinct behaviour of SWE is shown between the lowland stations and eastern and western highland sites.

3.2. Temporal evolution of snow-cover characteristics

Snow-cover characteristics, such as snow-cover depth, DSC and SWE, show very similar temporal evolution. The temporal evolution of snow-cover depth shows that a short snow period during 1951–1964 is followed by a more consistent snowy period during 1965–1974 and 1978–1987, and after later 1980s a notable drop towards unprecedented snow-free years during the 1990s and 2000s is shown. From 1991 to 2015, at the snow-richest highland site (Svratouch) very high year-to-year variability in snow depth cover is shown, with the most impressive example being the snowiest season of 2005/2006 and the poorest snow season of 2006/2007, and then followed by the snowiest season of 2009/2010 to the most deficient snow season of 2011/2012.

The difference in the climate characteristics of the north-western lowland (Doksany), southern lowland (Kucharovice), eastern (Svratouch, Luka and Kostelny Myslova) and western (Cheb and Kocelovice) hilly areas could be primarily responsible for the large station-to-station discrepancy in the trends of DSC/SWE (Figure 3). Interesting to note is the most pronounced decreasing trend of DSC (5.3 and 11.7 days decade^{-1}) was observed for the highland snowiest stations (Luka and Svratouch). During 1991–2015, DSC shows a statistically significant decreasing trend in the southern lowland and western hilly areas and slightly increasing trend but not statistically significant in the eastern part of the country. As for the SWE, the results show that during the 64-year study period, the long-term decreasing trend of SWE is not statistically significant although decline in the water accumulation in the form of snow is shown from October to March. After 1990s, insignificant upward trends of SWE are observed in western hilly areas and southern lowland. This could be explained by the fact that the increasing trends of DSC/SWE after 1990s have been caused by several snowiest cold seasons recorded at the end of the study period. Moreover, the number of cold seasons with high snow depth/DSC in the decade 2001–2010 is higher than in the decade 1991–2000 because of precipitation shortages and warmer conditions during the latest.

The average dates of the onset and termination of snow cover together with the earliest and the latest dates of these events are included in Table 2. There is an apparent relationship between the onset and the end dates of snow cover and the station altitude: the higher altitude, the earlier onset date and the later end date of snow cover. From 1951 to 2015, the average dates of the onset and termination of snow cover are December 5 and March 13, respectively. The earliest onset of snow cover occurred in 1973 (October 20), which was 45 days earlier than the average and the latest onset was recorded in 2012 (February 10), which was 66 days later than the average date. The earliest end date of snow cover occurred in 1974 (January 7), which was 66 days earlier than the average. For the snow-richest highland site, the average dates of onset and end of snow cover are November 8 and April 10, respectively. The earliest date for the onset of snow cover and latest end date for snow cover were 1 October 1996 and 23 May 2004, respectively.

The linear trends in the onset and termination dates of snow cover have also been analysed (Figure 4). The onset date of snow cover was recorded towards the end of the year, while the end date of snow cover occurs progressively earlier than the end of the cold season shifting towards the beginning of the year. The trend towards an earlier end date of snow cover is found in both lowland and highland sites. However, the most significant shift in the dates of termination of snow cover has been identified to occur mostly in the hilly areas; in the lowland areas, these changes are not that evident. From 1951 to 2015, the trend of the end date of snow cover shows the shifts towards earlier dates by 3 days decade^{-1}. The shift of the end date of snow cover during the last 24 years was more pronounced. For the western highland stations, the trend of the shift varies between 11.0 and 16.0 days decade^{-1}, whereas for the eastern highland stations it varies between 10.0 and 17.0 days decade^{-1}. These shifts represent a decrease of snow-cover duration and earlier snow melt. Similar results but at continental scale were reported by Peng *et al.* (2013), with a trend towards an earlier end date of snow cover at 70% of the Eurasian sites, and 28% of the sites showing an advance of snowmelt in spring as fast as 5 days decade^{-1}. Therefore, if seasonal snow accumulation is melting earlier in the season, evapotranspiration can be enhanced earlier, leading to an additional depletion in soil moisture during the early growing season.

3.3. The link between snow cover and soil moisture

Results showing the strength of relationship between the DSC and soil moisture, and between the SWE and soil moisture for each month of the early growing season, and the entire AMJ are presented in Table 3. Here, we

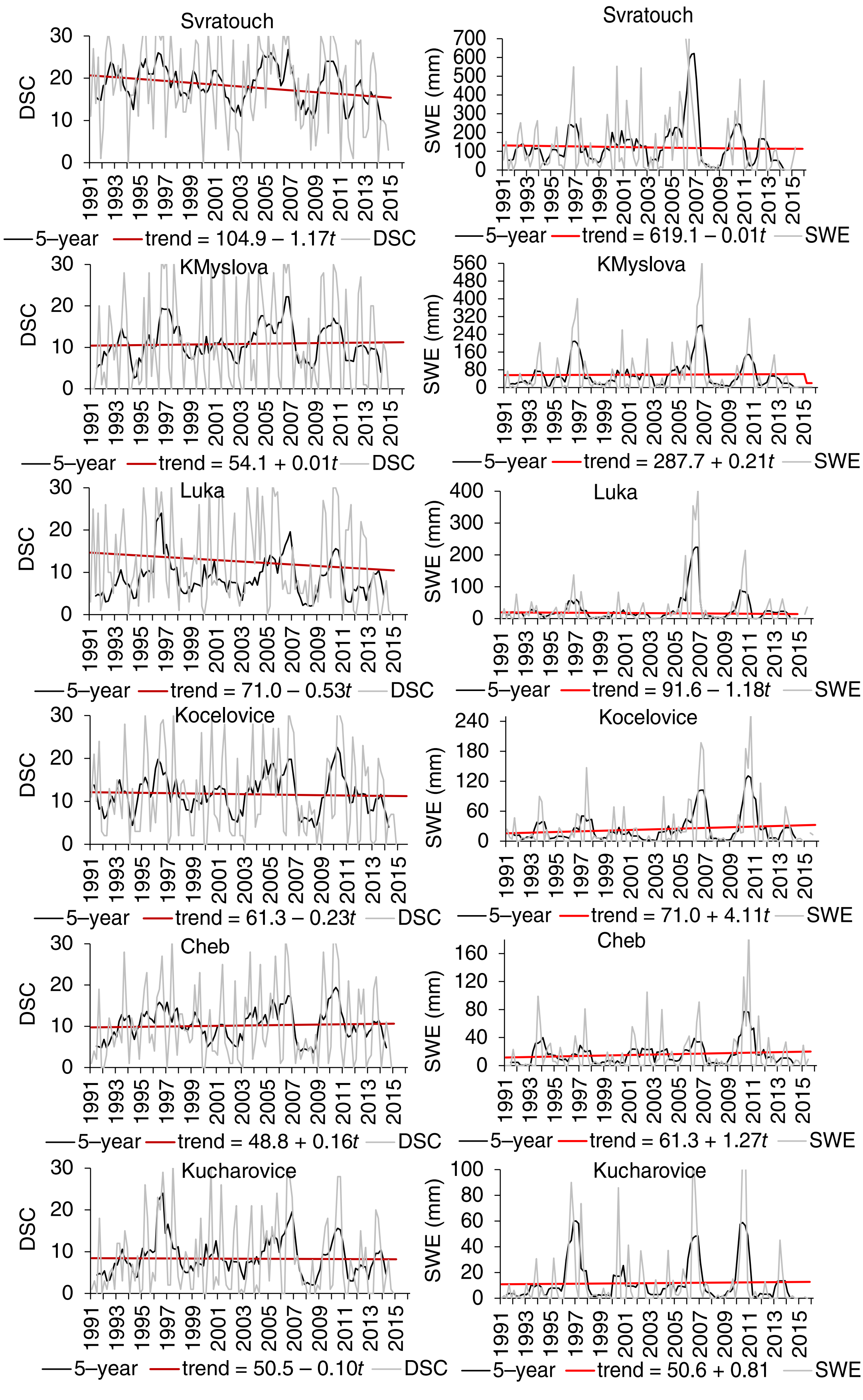

Figure 3. Temporal evolution of DSC/SWE (grey line) and 5-year low-pass filter (black line) from November through March for the period 1991–2015 (1951–2015 Doksany). The figures also quantify trends of seasonal values of DSC/SWE (seasonal sum of day year^{-1} and mm year^{-1}) (red line). The significance of trends was tested using Student's one-tailed *t*-test at the 95% level.

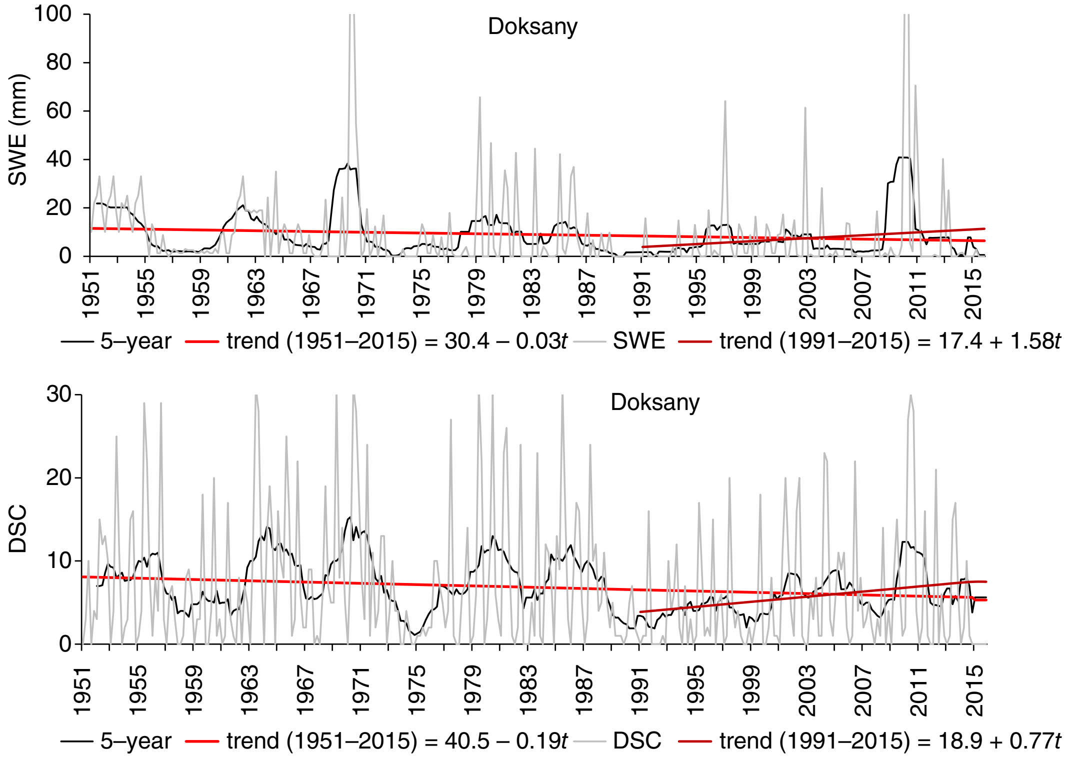

Figure 3. Continued.

Table 2. Statistical characteristics of the first day and the last day of snow cover for the period 1951–2015 for Doksany station, and 1991–2015 for the rest of stations.

stations	The onset date of snow cover			The end date of snow cover		
	Average	Earliest	Latest	Average	Earliest	Latest
Doksany	5 December	20 October 1973	10 February 2012	13 March	7 January 1974	23 April 1981
Kucharovice	3 December	28 October 2013	24 January 2007 and 24 January 2014	13 March	1 February 2014	7 April 2002
Cheb	20 November	27 October 1998 and 27 October 2013	26 December 2015	25 March	7 February 2014	22 April 1991
Luka	20 November	15 October 2010	27 December 2001	28 March	6 February 2014	20 April 1991
Kocelovice	16 November	8 October 1995	26 December 2015	25 March	7 February 2014	20 April 1997
Kostelny Myslova	16 November	7 October 1995	25 December 2015	29 March	9 February 2014	21 April 1991
Svratouch	8 November	1 October 1996	16 December 2001	10 April	24 March 1999	23 May 2004

describe these results in more details. The correlation of DSC–soil moisture is relatively weak in the lowland of the north-western region of the CR, whereas in the arable lowland in the south of the country the highest correlation between the cold season DSC/SWE and subsequent early summer soil moisture content is shown during April–May ($0.43 < r < 0.52$). In the eastern highland and western highland, larger values of DSC/SWE during the cold season lead to higher values of subsequent soil moisture in early summer ($0.28 < r < 0.49$), whereas smaller values of DSC/SWE correspond to lower soil moisture. On the other hand, the results also show weak correlation between SWE and soil moisture in late spring and early summer for the snow-richest highland site. It turns out that SWE resulted from snow accumulation during the cold season is only partly stored in the active soil layer.

The snow cover–soil moisture relationship gradually diminishes during April–June, and it appears that soil moisture is mostly affected by summer rainfall. Both for

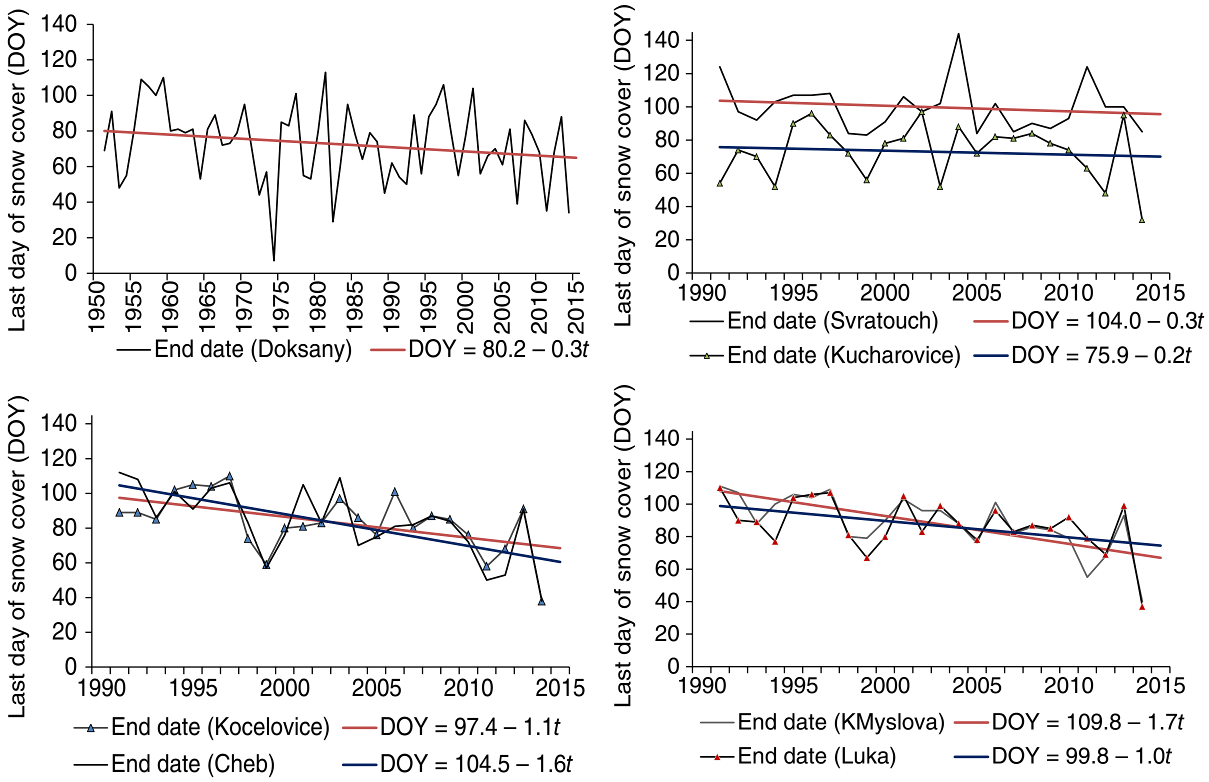

Figure 4. Statistically significant trends at 95% level in dates of termination (the last day of snow cover) of snow cover from 1951/1952 to 2014/2015 (1991/1992–2014/2015).

Table 3. Correlation coefficients between the seasonal DSC and soil moisture during the period 1991/1992–2014/2015 and between mean SWE during cold season and soil moisture for early summer (AMJ).

	Doksany	Kucharovice	Cheb	Luka	Kocelovice	Kostelny Myslova	Svratouch
DSC–soil moisture							
April	0.40*	0.52**	0.38*	0.35*	0.33*	0.46*	0.26*
May	0.36*	0.47*	0.27*	0.22	0.27*	0.37*	0.26*
June	0.26*	0.36*	0.18	0.26*	0.21	0.37*	0.22
AMJ	0.35*	0.49*	0.30*	0.30*	0.30*	0.49*	0.30*
SWE–soil moisture							
April	0.29*	0.47*	0.28*	0.30*	0.29*	0.38*	0.20
May	0.36*	0.43*	0.21	0.22	0.33*	0.31*	0.18
June	0.34*	0.37*	0.12	0.20	0.19*	0.25*	0.16
AMJ	0.35*	0.47*	0.21	0.28*	0.31*	0.37*	0.20

*90% confidence level, **95% confidence level and non-significant without asterisks.

lowland and highland regions, soil moisture in AMJ was found to be correlated with precipitation during the cold season ($0.22 < r < 0.46$). It turns out that liquid precipitation, more than solid precipitation (snowfall), lead to weakening correlation between SWE/DSC and the subsequent early summer soil moisture.

Snow-cover characteristics may significantly influence soil water saturation for the first part of the growing season, up to 45% ($0.23 < R^2 < 0.45$) of the variability in soil moisture; however, summer rain offsets the potential soil moisture deficit. The analysis of the results presented here reveals that (1) the strength of the link between the snow cover and soil moisture varies from April to June, and depends on the topographic position of the station; (2) DSC/SWE appears to especially influence the soil moisture during April–May (small amount of SWE leads to diminished soil water saturation) and (3) exclusively for the station Svratouch, regardless of which snow-cover parameter is used, the strength of snow cover–soil moisture correlation is very weak. There are two possible assumptions of these results. The first is that during April–May the soil moisture depends on preceding season snowmelt and precipitation accumulation, as well as on the amount of available soil water from the preceding autumn. The second is that the correlation results for snow-richest highland sites suggest that although snow accumulation accounts for a large percentile of annual precipitation amount, little of the moisture released through snow melting is maintained in soil because of rapid snowmelt, runoff and outflow.

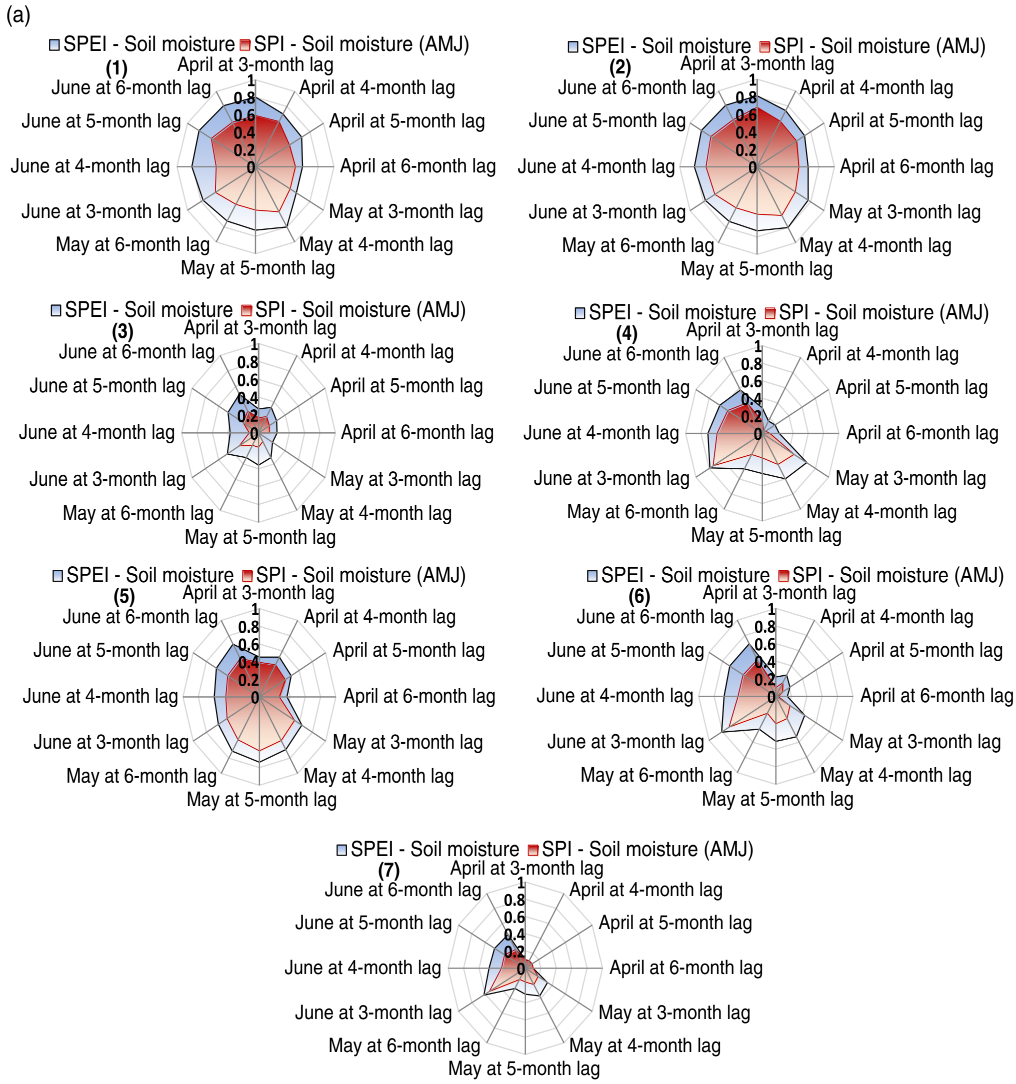

Figure 5. (a) Correlation coefficients between monthly SPEI/SPI at 3-, 4-, 5- and 6-month lags and soil moisture of early summer (AMJ) at seven stations: (1) Doksany, (2) Kucharovice, (3) Cheb, (4) Luka, (5) Kocelovice, (6) Kostelny Myslova and (7) Svratouch. SPEI3 (SPI3) was calculated for April (March–February), May (April–March) and June (May–April). The 4-month lag integrates the moisture conditions for the current month and the preceding 3 months, and the SPEI4 (SPI4) was calculated for April (March–February–January), May (April–March–February) and June (May–April–March). The 5-month lag represents moisture conditions during the current month and the preceding 4 months, and the SPEI5 (SPI5) was calculated for April (March–February–January–December), May (April–March–February–January) and June (May–April–March–February). The 6-month lag corresponds to the accumulation moisture conditions from the current month and the preceding 5 months, and the SPEI6 (SPI6) was calculated for April (March–February–January–December–November), May (April–March–February–January–December) and June (May–April–March–February–January). (b) Correlation coefficient between monthly SPEI/SPI at 3-, 4-, 5- and 6-month lags and soil moisture of later summer (JAS) at seven stations: (1) Doksany, (2) Kucharovice, (3) Cheb, (4) Luka, (5) Kocelovice, (6) Kostelny Myslova and (7) Svratouch.

3.4. The link between drought and soil moisture

In this section, the role of precursor moisture accumulation deficit, including the influence of drought during the cold season on averaged soil moisture deficit during AMJ and JAS, is analysed. To this end, the correlation coefficients between monthly SPEI/SPI at 3-, 4-, 5- and 6-month lags and mean AMJ and JAS soil moisture were plotted for each station in Figure 5. The results point out on the following aspects (Figure 5(a)): (1) the correlation between SPEI and AMJ soil moisture is higher than that between SPI and AMJ soil moisture, especially in arable lands; this is presumably due to lowland climatic characteristics (low rainfall and higher potential evapotranspiration); (2) both drought indices are highly correlated with AMJ soil moisture in June at a 3-month lag in the eastern highland; (3) the link between drought and soil moisture strongly depends on 3–4 month time scales and (iv) the magnitude of correlation coefficients displays an evident east–west gradient within the CR and clearly delimits the regions with the highest correlation (soil moisture highly depends the on current and preceding moisture conditions) for arable lowlands, moderate correlation for the western highlands and less correlation for the eastern highlands.

For arable lands, the link between drought during winter-spring and early summertime soil moisture is

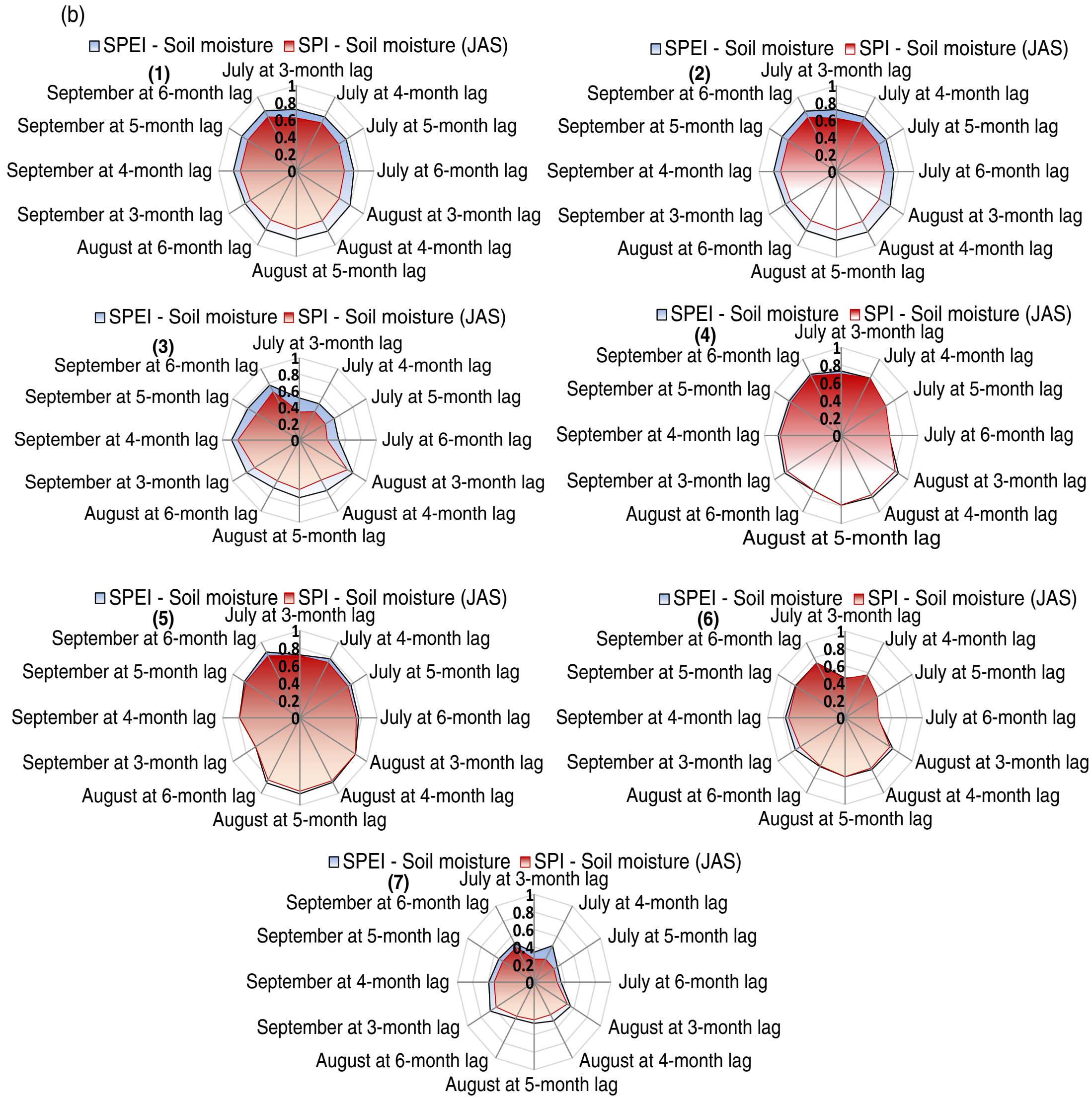

Figure 5. Continued

strong, and explains between 35 and 56% of the variability of soil water content ($0.35 < R^2 < 0.56$). The highest correlation was founded in April for the SPEI at 3-, 4- and 5-month lags ($0.75 < r < 0.81$) and in May for SPEI at 4-month lag ($r = 0.91$). At larger time scales (6-month lag), the shortage of moisture during the cold period of the year led to decrease in soil moisture at the beginning of the growing season and contributed to the persistence of drought by later summer (e.g. autumn-winter-spring-summer drought of 2011/2012 and 2014/2015). Also, the decrease of soil water content during AMJ impacts on crop productivity in the later season. Summer crops will be exposed to progressively drier conditions in May and June, together with rather low amounts of soil available water content between July and September (Potopová *et al.*, 2015b; Trnka *et al.*, 2015b).

For western highlands, the AMJ soil moisture is highly correlated with droughts at longer time scales; strong correlation is shown between winter-spring drought – SPEI/SPI at 5- to 6-month lags in May and June ($0.43 < r < 0.62$) – and subsequent AMJ soil moisture. For the eastern highlands, a weak correlation between SPEI/SPI from November to April and AMJ soil moisture was found, whilst the strongest correlation was found in June at 3-month lag ($r = 0.62–0.79$).

Similarly, in order to assess the link between drought at various time scales and the soil moisture during JAS, the correlation coefficients between SPEI/SPI at 3-, 4-, 5- and 6-month lags and JAS soil moisture have been calculated (Figure 5(b)). As such, these lags capture the drought conditions for the entire growing season. The analysis of these correlation coefficients highlights the following: (1) the correlation between multiscalar drought indices and JAS soil moisture is significantly higher than that for AMJ period; (2) similar strength of correlation is shown for both SPEI and SPI in highlands, while in lowlands SPEI shows higher correlation than SPI and (3) both indices detected the highest correlation in August at 3- and 4-month lags in western and eastern highlands. Drying soil in these regions could also be attributed to a

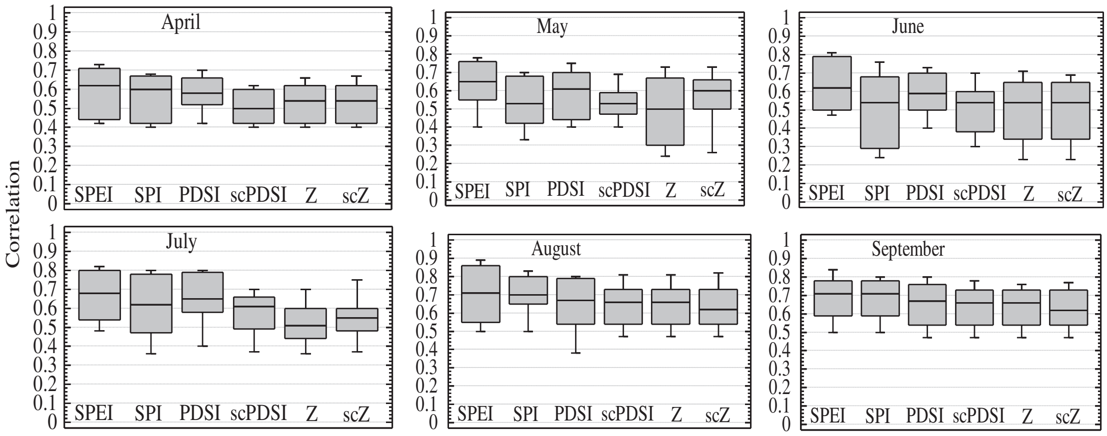

Figure 6. Box plots summarizing the correlation coefficients between the time series of drought indices (SPEI-6, SPI-6, PDSI, scPDSI, Z-index and scZ-index) and the corresponding series of standardized soil moisture data for seven stations situated in lowland and highland regions in the CR.

combination of factors such as increasing temperature and potential evaporation which is not balanced by the changes in precipitation.

As drought develops, a higher correlation is shown between early summer SPEI/SPI and JAS soil moisture. This result can be explained by the fact that precipitation and temperature anomalies in early summer initiate soil moisture anomalies in later summer which, in turn, favour the persistence of drought during the entire growing season. In addition, one should take into consideration that soil moisture depends on many other factors, including field water-holding capacity and precipitation frequency and intensity (Trenberth, 2011). A strong relationship between SPEI and soil moisture could explain how winter drought in conjunction with higher temperature and higher evapotranspiration during the growing season could substantially contribute to decreasing soil moisture in both lowland and hilly areas.

Increasing drought severity should be evident in increasing impacts on systems sensitive to drought, including soil moisture. In turn, the stock of soil moisture can modulate drought effect. To summarize the assessment of the performance of different drought indices for monitoring drought impacts on soil moisture, the correlation coefficients between the monthly time series of six drought indices (SPEI-6, SPI-6, PDSI, scPDSI, Z-index and scZ-index) and the monthly series of the standardized soil moisture from April to September for seven stations situated in lowland and highland regions in the CR have been calculated (Figure 6). Results suggest that for the selected stations in the CR and depending on the analysed month, strong differences arise when comparing the SPEI, the SPI and the Palmer drought indices, with the first two indices outperforming the latter in the majority of cases. PDSI appears to perform better compared to scPDSI for monitoring surface soil moisture deficit. The correlation between the time series of Z-index and soil moisture is much alike to that for scZ-index. However, it is interesting to note that correlations between the PDSI and soil moisture from April to June were higher than for the SPI. Dai *et al.* (2004) showed that the correlation with the soil moisture data is higher for the PDSI than precipitation, suggesting that the PDSI better represents surface moisture conditions than precipitation alone (SPI), whereas the higher correlations with the SPEI ($0.55 < r < 0.89$) than the SPI ($0.50 < r < 0.79$) were found in the warmest months (July and August), in which evapotranspiration rates are the highest.

The correlation between six drought indices and soil moisture in AMJ are statistically significant at the 95% level ($0.33 < r < 0.79$) in both lowlands and highlands. However, the strength of correlation between drought indices and soil moisture anomalies is higher during JAS. This is consistent with the results shown for the correlations of the SPEI and SPI at various time scales and total AMJ/JAS soil moisture. The risk of decreased soil moisture in late summer can be mostly because of increased probability of drought during early growing season. The results of this analysis confirm again the legitimacy of increasing evaporative demand in AMJ (Trnka *et al.*, 2015a, 2015b); it is not balanced by a similar increase in precipitation, thus leading to a higher rate of soil moisture withdrawal by plants, which leaves less water for the latter part of the growing season than for the previous.

3.5. Anomalous snow seasons and their influence on early summer soil moisture

A snow season was defined as anomalous when the value of DSC and SWE were greater (less) or equal than one standard deviation ($\geq +1.0\sigma / \leq -1.0\sigma$), respectively. The years with negative/positive snow anomaly which were followed by subsequent negative/positive soil moisture content ($\leq -1.0\sigma / \geq +1.0\sigma$) and dry/wet early summer season (identified by PDSI, scPDSI, Z-index and scZ-index) have been selected and ranked in Table 4. Because DSC and SWE have approximately similar temporal evolution, they show simultaneously the same type of anomaly. In addition, the percentage of dry early summers preceded by winters with negative anomalies of DSC/SWE has been calculated (not shown) to support the results

Table 4. The top 5 years with lowest and highest values of DSC/SWE, AMJ soil moisture and driest/wettest AMJ seasons.

Lowest		Highest		Driest	Wettest
DSC/SWE	Soil moisture	DSC/SWE	Soil moisture		
1951–2014 (Doksany $h = 158$ m)					
2006/2007	*2007*	**2009/2010**	**1965**	*2004*	**1965**
2002/2003	1953	**1969/1970**	**1970**	*2005*	**2010**
1991/1992	2005	**1964/1965**	**2010**	*2003*	**1955**
1951/1952	2011	**1968/1969**	1955	*2007*	**1995**
2013/2014	2003	**1980/1981**	1987	*1992*	**2013**
Lowland 1991–2014					
2013/2014	*2003*	**1995/1996**	**2010**	*2012*	**2010**
2006/2007	*2014*	**2005/2006**	**2006**	*2014*	**1996**
2011/2012	2000	**1996/1997**	**2013**	*2007*	**2013**
2002/2003	*2007*	**2019/2010**	1995	*2000*	2004
1997/1998	*2012*	**2012/2013**	2004	1992	**2006**
West highlands 1991–2014					
1991/1992	*2000*	**2009/2010**	2013	*2014*	**2013**
2006/2007	*1998*	**2005/2006**	**2010**	2004	**2010**
1997/1998	*2014*	**1992/1993**	**2002**	*1998*	**2008**
2013/2014	*1992*	**2000/2001**	**2001**	*2000*	1995
1999/2000	*2007*	**2001/2002**	**2008**	2012	**2001**
East highlands 1991–2014					
2002/2003	*2003*	**2005/2006**	**2010**	*2012*	**2010**
1999/2000	*2000*	**2009/2010**	1995	*2003*	**2013**
2013/2014	*2012*	**2004/2005**	**2013**	*2000*	**1996**
1997/1998	*2007*	**1997/1996**	**1996**	*2014*	**1995**
2006/2007	*2009*	**2002/2013**	**2006**	1994	**2006**

Italic/bold fonts indicate early summer drought/wetness following the season with negative/positive snow anomalies and lowest/highest AMJ soil moisture and regular fonts indicate no association between DSC/SWE and AMJ soil moisture

of this analysis. Temporal evolution of standardized values of SWE and AMJ soil moisture, and averaged values of PDSI, scPDSI, Z-index and scZ-index for AMJ season are displayed in Figure 7. The negative/positive anomalies of SWE are associated with negative/positive anomalies of the following early summer soil moisture. This feature is more consistent during the periods when (1) both SWE and soil moisture anomalies (low available water capacity) were highly negative; such cases were recorded during 1951–1954, 1989–1998 and 1999–2009 and (2) both SWE and soil moisture anomalies were highly positive; such cases were recorded during 1961–1970 and 2010–2013. AMJ soil moisture anomaly reached historical low values during the period 1999–2009 in both lowlands and highlands (2000, in eastern highlands; 2003, in southern lowland and eastern highlands; 2007, in north-western lowland).

Table 4 summarizes the top 5 years with the lowest and the highest values of DSC/SWE during the cold season, the lowest and the highest AMJ soil moisture and the driest/wettest early summer seasons in lowland, western and eastern highlands. The negative snow anomalies associated with negative AMJ soil moisture anomalies could have played a complementary role in accelerating and wide spreading of AMJ droughts such as during 1992, 1998, 2012 and 2014. In lowlands, out of 24 cold seasons, 12 were identified as being snow free (1991/1992, 1992/1993, 1994/1996, 1997/1998, 1998/1999, 2000/2001, 2006/2007, 2007/2008, 2010/2011, 2011/2012, 2013/2014 and 2014/2015) and only 5 as anomalous snow abundant (1995/1996, 1996/1997, 2005/2006, 2009/2010 and 2012/2013). The anomalous snow abundant season of 2005/2006 in lowland, with a heavy snow cover that lasted up to the second half of March and was followed by an abrupt thaw in combination with heavy rainfalls in the spring of 2006, led to a sudden increase of soil moisture in AMJ. On the other hand, the extremely poor snow seasons of 2006/2007, 2013/2014 and 2014/2015, associated with the shortest snow-cover duration and low DSC/SWE negative anomalies, were followed by the most persistent dryness during the early summers of 2007, 2014 and 2015, respectively. Large negative snow anomalies, for instance, have been recorded during the cold season 2013/2014 due to high temperature anomalies (up to 3.5 °C higher than the average of the reference period 1971–2000) and precipitation deficit (up to 50%) in north-western and south-eastern lowlands. The spatially averaged snow cover at country level for the cold season 2013/2014 only represented 27% of the long-term average. The lowest amount of snow accumulation was observed in southern lowland (around 12%), where the maximum snow depth did not exceed 3 cm. Those abnormally low snow anomalies led to the development of abnormally dry episodes during winter-early spring and early summer. The snow-free season of 2014/2015 had also contributed to the development of drought conditions in the subsequent growing season over the CR.

Snow seasons alternate between positive and negative anomalies during the last decade, as do the following

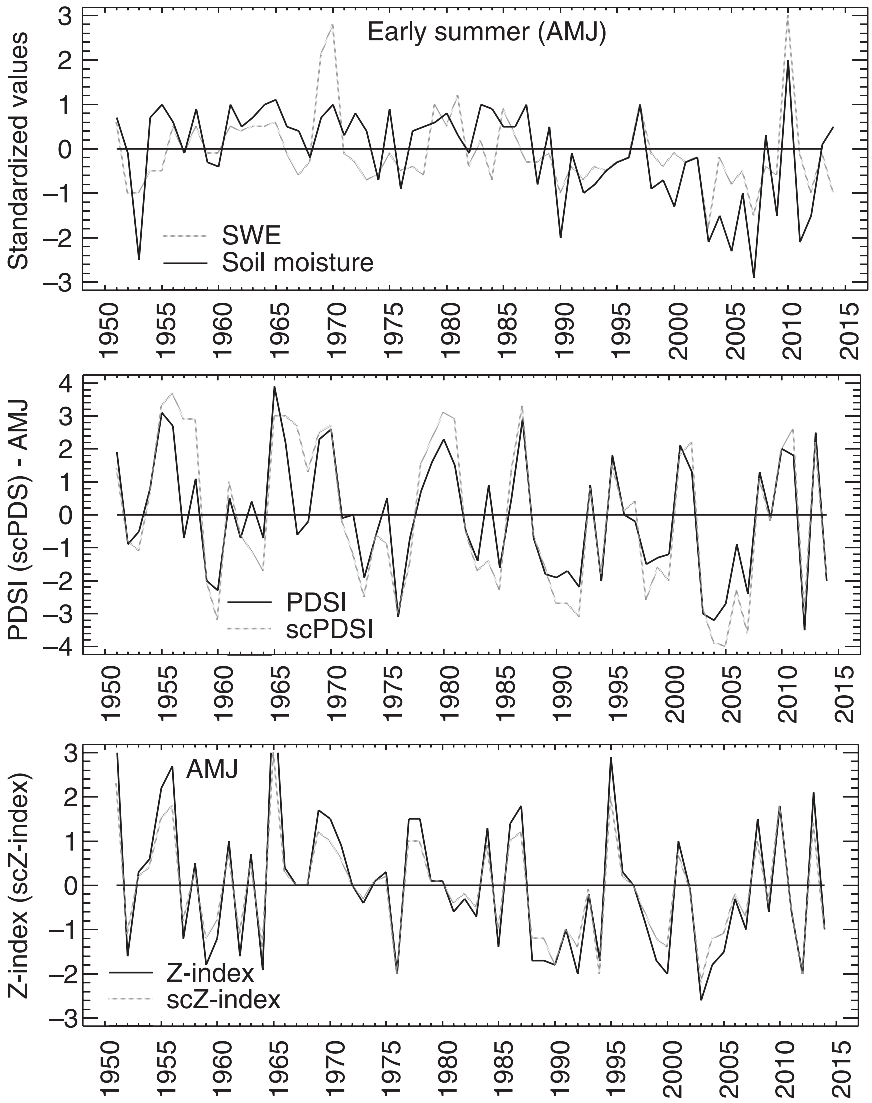

Figure 7. Temporal evolution of standardized values of mean SWE in the cold season and soil moisture in early growing season (AMJ) (top); early summer mean values of PDSI, scPDSI, Z-index and scZ-index (middle and bottom) from 1951/1952 to 2014/2015 at Doksany.

growing seasons. The dry AMJ season of 2012 was followed by a wet AMJ of 2013 and then by a dry AMJ of 2014. The abundant snow during the cold season of 2012/2013 was followed by a wet AMJ season which was more striking during May–June in western highlands. The analysis of snow anomaly characteristics during the cold season and the soil characteristics of the following AMJ season shows that >52% of dry AMJ seasons followed after cold seasons with poor snow, and 42% of wet AMJ seasons followed after cold seasons with abundant snow. That means that other soil moisture driving factors which were not analysed in this study should be further taken into consideration.

4. Discussion

In this section, the findings of this paper are discussed in context with earlier results reported at national, regional and global levels. However, the inclusion of snow processes in the development of drought impacts has been poorly considered in the scientific literature. This study analyses the relationship between the snow melt dates, SWE, climate drought indices and soil moisture in the CR. The analyses of multiannual cycle and temporal evolution of basic snow-cover characteristics across the study region illustrates a strong seasonal variation and significant differences in snow-cover depth, DSC and SWE between lowland stations and eastern and western highland sites. The snow lacking season would further amplify drought during the following growing season in lowlands and west highlands, whereas in east highlands this link was less pronounced. In the lowlands, before the growing season, soil is recharged via spring snowmelt. Moreover, high evapotranspiration and a lack of summer precipitation result in declining available soil water from April until September. Available soil water usually falls below field capacity during the late summer months (Možný *et al.*, 2012). Soil begins to recharge in the autumn once evapotranspiration decreases. This mechanism has been well identified by the correlation coefficients between multiscalar drought indices and soil moisture (Figure 5). The results also present conclusive evidence that the negative anomalous snow characteristics in conjunction with winter and AMJ drought amplify lingering impacts on the depletion of soil moisture in the later summer, and consequently reduce the latent heat flux and its ability to cool the soil surface (Bamzai and Shukla, 1999; Löffler, 2005). Despite the comprehensiveness of this analysis, other soil–atmosphere feedbacks not considered in the study may have influenced the variability of soil moisture (D´Odorico and Porporato, 2004; Mo and Lettenmaier, 2015). Strong soil–atmosphere feedbacks have been reported to contribute to the amplification of summer temperature extremes and consequently to a higher incidence of heat waves (Miralles *et al.*, 2014; Whan *et al.*, 2015).

Our results show positive correlation between the SPI and soil moisture, which are in line with the results reported by Quiring and Kluver (2009) and Wu and Kinter (2009) attesting the appropriateness of SPI to assess the soil moisture. Several studies dealing with drought–soil moisture relationship (Sims and Raman, 2002; Dai *et al.*, 2004) demonstrate that the strongest correlation between PDSI and soil moisture is shown in late summer, and the weakest correlation is shown in spring, when snowmelt is an important contributor to soil moisture. The results of this study are in line with the above-mentioned papers.

Regional aspects of the global warming signal have been reported for the CR by Brázdil *et al.* (2009), by pointing out statistically significant increasing trends in winter temperature. This finding supports the results of this study regarding the observed decreasing trend in the end date of the snow cover and increasing trend in the onset date of snow cover. These results also support concerns related to the potential decrease of snow-cover duration and earlier snow melt, and are consistent with other studies dealing with vegetation phenology (i.e. trend towards earlier dates) (Trnka *et al.*, 2015b), prolonged frost-free period (Potop *et al.*, 2014) and the effect of climate change on the growing season (Potopová *et al.*, 2015a) in the CR. Conclusive evidence has been found to show that these changes are most pronounced at higher elevations, where the shifts in both onset and end dates of snow cover towards later and earlier dates, respectively, have also been observed. Several authors have linked these changes to an increasing trend in air temperature (Vicente-Serrano *et al.*, 2007; Choi *et al.*, 2010; Brown and Robinson, 2011; Betts *et al.*, 2014a). They show that higher temperatures in November would delay the snow fall in autumn and higher temperatures during March–April would produce an earlier snowmelt.

Some of the locations showing a decreasing trend in DSC do not show such trend in SWE, as such indicating that the same amount of snow tends to fall within a considerably shorter interval. Increasing in snowfall intensity could be explained by an increased moisture-holding capacity of the warmer atmosphere (Birsan and Dumitrescu, 2014). As indicated by Kyselý and Huth (2006) and Cahynová and Huth (2014), a considerable increase in the persistence of atmospheric circulation types in Europe occurred during 1990s, which is reflected in the increase in the occurrence of climate extreme events. High correlation between NAO index and DSC over the CR was found by Brázdil *et al.* (2009), when the western circulation (positive NAO index) brings warmth and rainfall over Central Europe and negatively influences the duration of snow cover.

5. Summary and conclusions

New aspects and additional results compared to other studies related to this issue in the CR have highlighted the links between snow cover and soil moisture, and between multiscalar drought indices and soil moisture based on observed daily data. Two multiscalar drought indices were used to examine the role of precursor moisture accumulation deficit, including the influence of dryness during the cold season on soil moisture deficit during AMJ and JAS. Simultaneous correlation between the time series of averaged values of PDSI, scPDSI, Z-index and scZ-index and soil moisture during AMJ (JAS) have been used to bring new insights on the issue. The novelty of this study also consists in the extension of the study period up to now (1951–2015 and 1991–2015) for the temporal evolution and trend of snow-cover characteristics (the first day of snow cover and the last day of snow cover, DSC, snow cover duration, snow depth and SWE). The results emphasize on the need for a larger volume of observed daily data to better investigate the role of snow cover on drought propagation in soil during early summer and gradually in later summer. The conclusions can be summarized as follows:

(1) At all selected stations, similar temporal evolution of DSC has been observed during the period 1991–2000, when the number of snow days was less frequent. Large differences in snow-cover characteristics among stations during the period 2001–2015 have been observed. This analysis also highlights that after the 2000s cold seasons with much or little snow are seldom grouped together, and in every second case after a year with below-average snow cover follows a year with above-average snow. Conversely, clusters of successive years with below-average snow cover have been recorded during the 1990s.

(2) The most significant shift in the early dates of snow-cover termination has been identified to occur mostly in the hilly areas while these changes are not that evident in the lowland areas. The most pronounced decreasing trend of DSC was observed for the highland snowiest stations. Although the trends in the seasonal SWE are not statistically significant, an overall decline in the volume of snow melt water was found, notably both in early and late cold seasons, when rain prevails the snow.

(3) Small differences in the magnitude of correlation between the SPI/SPEI and soil moisture have been detected, but larger differences were found for the correlation coefficients between the Palmer drought indices (PDSI, scPDSI, Z-index and scZ-index) and soil moisture. The strength of correlation between drought indices and soil moisture anomalies is higher in later summer than in early summer. This result can be explained as an effect of negative anomalous snow characteristics in conjunction with winter and AMJ drought which amplifies lingering impacts on the depletion of soil moisture in the later summer.

These results may foster further research on other driving factors not yet considered in this study that may influence the soil moisture variability in the lowland, highland and mountain sites in the CR.

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