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A process-based typology of hydrological drought

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Abstract

Hydrological drought events have very different causes and effects. Classifying these events into distinct types can be useful for both science and management. We propose a classification of hydrological drought types that is based on the governing drought propagation processes. In this classification six hydrological drought types are distinguished, i.e. (i) *classical rainfall deficit drought*, (ii) *rain-to-snow-season drought*, (iii) *wet-to-dry-season drought*, (iv) *cold snow season drought*, (v) *warm snow season drought*, and (vi) *composite drought*. The processes underlying these drought types are a result of the interplay of temperature and precipitation at catchment scale in different seasons. As a test case, about 125 groundwater droughts and about 210 discharge droughts in five contrasting headwater catchments in Europe have been classified. The most common drought type in all catchments is the *classical rainfall deficit drought* (almost 50 % of all events), but in the selected catchments these are mostly minor events. If only the five most severe drought events of each catchment are considered, a shift towards more *rain-to-snow-season droughts*, *warm snow season droughts*, and *composite droughts* is found. The occurrence of hydrological drought types is determined by climate and catchment characteristics. The typology is transferable to other catchments, incl. outside Europe, because it is generic and based upon processes that occur around the world. A general framework is proposed to identify drought type occurrence in relation to climate and catchment characteristics.

1 Introduction

Hydrological (groundwater and discharge) drought events are severe natural disasters, in damage comparable to large-scale floods and earthquakes. Due to their long duration and large spatial extent, droughts have significant economic, social, and environmental impacts (EU, 2006, 2007; Sheffield and Wood, 2011). Especially in vulnerable regions like Asia and Africa, the total number of people affected by drought is very high (up to 300 million people per event; CRED, 2011) and droughts result in famine and loss of life (ISDR, 2007), as happened recently in the Horn of Africa (FEWS-NET, 2011; UN, 2011). Droughts in developed countries primarily result in economic loss: in the USA on average 6 to 8 billion USD per year (Andreadis et al., 2005; Below et al., 2007) and in the EU more than 100 billion EUR in the period 1976–2006 (EU, 2006, 2007). According to recent drought studies (EU, 2006, 2007; Sheffield, 2008; Feyen and Dankers, 2009; Dai, 2011), there is an increasing trend in drought extent and population affected by drought, which makes drought research and management a pressing issue.

Drought is defined as a sustained and regionally-extensive period of below-average natural water availability. It is a recurring and worldwide phenomenon, with spatial and temporal characteristics that vary significantly from one region to another (Tallaksen and Van Lanen, 2004). A prolonged lack of precipitation (also called meteorological drought) can propagate through the hydrological system and affect soil moisture, resulting in soil moisture drought, as well as groundwater and discharge, resulting in hydrological drought (Tallaksen and Van Lanen, 2004; Mishra and Singh, 2010).

This so-called propagation of drought from meteorological to hydrological drought is characterised by a number of features (Eltahir and Yeh, 1999; Peters et al., 2003; Van Lanen et al., 2004; Van Loon et al., 2011b), visualised in Fig. 1:

– meteorological droughts are combined into a prolonged hydrological drought (pooling);

– meteorological droughts are attenuated in the stores (attenuation);

– a lag occurs between meteorological, soil moisture and hydrological drought (lag);

– droughts get longer moving from meteorological to soil moisture to hydrological drought (lengthening).

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These features are controlled by catchment characteristics and climate. Lag and attenuation are governed by catchment control, and pooling and lengthening by both catchment and climate control (Van Lanen et al., 2004).

Compared to other natural disasters, knowledge of drought still has large gaps (Smakhtin, 2001; Mishra and Singh, 2010). Most focus of drought research is on finding the best drought index (e.g. Bonacci, 1993; Heim, 2002; Keyantash and Dracup, 2002; Ntale and Gan, 2003; Mpelasoka et al., 2008; Niemeyer, 2008; Wanders et al., 2010), but hydrological droughts have very different causes that cannot be captured by a single index (Wanders et al., 2010). Besides by a rainfall deficit, hydrological droughts can also be caused by low temperatures and snow accumulation (Van Lanen et al., 2004; Van Loon et al., 2010). In 2006 and 2010, for example, cold and dry winters have resulted in severe problems with drinking water and electricity production in Norway (NRK, 2010).

For drought management it is very important to distinguish between different types of hydrological drought, because these different types need different preventing measures and coping mechanisms. In addition, drought research could benefit from a common terminology and further study of the processes underlying drought. Therefore, one of the most important scientific challenges is related to the diversity of causative mechanisms of hydrological drought around the world (Marsh et al., 2007). Currently, there is no generally-accepted classification scheme (Wilhite and Glantz, 1985; Lloyd-Hughes and Saunders, 2002), like there is for floods (Merz and Blöschl, 2003). Hydrological drought classification is mainly done for sectors (e.g. socio-economic drought; Mishra and Singh, 2010) and based on drought severity (Dracup et al., 1980; Rossi et al., 1992; McKee et al., 1993, 1995; Lloyd-Hughes and Saunders, 2002; Smakhtin and Hughes, 2004), but not based on processes. For meteorological droughts, some process-based classifications have been developed (Phillips and McGregor, 1998; Fowler and Kilsby, 2002; Mishra and Singh, 2010), but hydrological drought events are either defined in very general terms and analysed only by their statistics (Andreadis et al., 2005; Fleig et al., 2006; Sheffield and Wood, 2007; Sheffield, 2008; Sheffield et al., 2009)

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or a single drought event with its underlying processes is described in detail (Santos et al., 2007; Trigo et al., 2010; Li et al., 2010). A more generally applicable typology of hydrological drought is needed, both for process understanding of drought and for improvement of drought forecasting.

In this paper, we propose a general hydrological drought typology based on the underlying processes of drought propagation. The objectives of this study are: (i) to describe hydrological drought types and provide examples, (ii) to classify hydrological drought events in five contrasting catchments, (iii) to find the most common and most severe drought types in catchments with different climate and catchment characteristics, and (iv) to relate these drought types to catchment and climate control.

The outline of the paper is as follows. First, the study areas are introduced (Sect. 2) and the hydrological model and drought analysis method are explained (Sects. 3.1 and 3.2). Then, the results of the hydrological modelling and drought analysis are described (Sects. 4.1 and 4.2) and a hydrological drought typology is presented (Sect. 4.3). In Sect. 4.4, hydrological drought events in the study areas are classified into drought types, both for all drought events and for the five most severe events per catchment. Additionally, some examples are given of a situation where a hydrological drought did not develop from a meteorological drought. Finally, in Sects. 5 and 6, results are discussed and summarised and a general framework is presented that shows the occurrence of drought types in relation to climate and catchment characteristics.

2 Study areas

The five catchments used in this study are natural headwater catchments in Europe with contrasting climate and catchment characteristics (Fig. 2a; Van Lanen et al., 2008).

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2.1 Narsjø

The Narsjø catchment is located in South-eastern Norway (Fig. 2d). It is a sub-basin of the Upper-Glomma, which is the headwater catchment of the Glomma. The area of the Narsjø catchment is approx. 120 km^2 (Table 1). The catchment is located in a glacially formed, mountainous region with rounded tops and U-shaped valleys. The altitude range is rather large with approx. 740–1600 m a.m.s.l. (Engeland, 2002). The Narsjø catchment has a subarctic climate with mild summers and very cold winters (Köppen-Geiger climate Dfc). In the observation period 1958–2007, measured mean annual temperature was 0.7 °C, precipitation was around 590 mm yr^{-1}, and potential evaporation was around 300 mm yr^{-1} (Table 1). In winter, a continuous snow cover is present for, on average, 7 months, from mid-October until the end of May, dependent on altitude (Engeland, 2002). Measured mean discharge was around 820 mm yr^{-1}, which is higher than measured precipitation due to the low elevation of precipitation gauges (Fig. 2d) in combination with an increase of precipitation with altitude. The low-flow season of Narsjø is winter, when recharge is zero because of snow accumulation, and highest flows occur in May due to snow melt. Narsjø is a hardrock catchment consisting predominantly of impermeable metamorphic rocks without extensive groundwater storage, which makes the catchment quickly responding to precipitation. Some delay in the response is caused by lakes, covering 3 % of the catchment, and bogs, covering 12 % (Van Loon et al., 2010). Other land cover types of the catchment are open area (61 %), forest (24 %), and only little agriculture (0.4 %) (Hohenrainer, 2008). Human influence is very limited in the Narsjø catchment.

2.2 Upper-Metuje

The Upper-Metuje catchment is located in North-eastern Czech Republic and partly in Poland (approximately 10 % of the catchment area) (Fig. 2b). It is the headwater catchment of the Metuje, which drains into the Elbe. The area of the Upper-Metuje catchment is approx. 70 km^2 (Table 1). The catchment is located in a hilly region of

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gentle slopes and wide valleys, except for some steep sandstone formations in the centre of the catchment. The altitude range is approx. 450–780 m a.m.s.l. The Upper-Metuje catchment has an oceanic climate with mild summers and winters (Köppen-Geiger climate Cfb). In the observation period 1982–2005, measured mean annual temperature was 5.9 °C, precipitation was around 750 mm yr^{-1}, and potential evaporation was around 570 mm yr^{-1} (Table 1). In winter, a continuous snow cover is present for, on average, 4 months, from December until the beginning of April. Measured mean discharge was around 320 mm yr^{-1} (Table 1). The low-flow season of Upper-Metuje is summer/autumn, and highest flows occur in March due to snow melt (Rakovec et al., 2009). Upper-Metuje is a groundwater catchment consisting of multiple sandstone layers, alternating with less permeable sediment layers, that form a large, multiple aquifer system. This makes it a slowly responding catchment with a relatively high baseflow. Nevertheless, discharge peaks occur when storage is filled (Van Loon et al., 2010). Land cover of the catchment mainly consists of cropland and grassland (51 %), and forest (46 %) (Rakovec et al., 2009). Human influence is limited to extensive agriculture.

2.3 Upper-Sázava

The Upper-Sázava catchment is located in Central Czech Republic (Fig. 2c). It is the headwater catchment of the Sázava, which (finally) drains into the Elbe. The area of the Upper-Sázava catchment is approx. 130 km^2 (Table 1). The catchment is located in a hilly region of gentle slopes and wide valleys and the altitude range is approx. 490–800 m a.m.s.l. The Upper-Sázava catchment has an oceanic climate with mild summers and winters (Köppen-Geiger climate Cfb). In the observation period 1963–1999, measured mean annual temperature was 6.8 °C, precipitation was around 720 mm yr^{-1}, and potential evaporation was around 680 mm yr^{-1} (Table 1). In winter, a continuous snow cover is present for, on average, 4 months, from December until the beginning of April. Measured mean discharge was around 290 mm yr^{-1} (Table 1). The low-flow season of Upper-Sázava is summer, and highest flows occur in March due to snow melt

(Rakovec et al., 2009). Upper-Sázava is a hardrock catchment consisting of impermeable metamorphic rocks and sedimentary rocks with limited groundwater storage, which gives it an intermediate response to precipitation. A significant delay is caused by lakes, covering around 2 % of the catchment (Van Loon et al., 2010). Other land cover types of the catchment are forests (50 %), and cropland and grassland (40 %) (Rakovec et al., 2009). Human influence is limited to extensive agriculture, and some groundwater extraction and sewage disposal.

2.4 Nedožery

The Nedožery catchment is located in Central Slovakia (Fig. 2e). It is the headwater catchment of the Nitra, which (finally) drains into the Danube. The area of the Nedožery catchment is approx. 180 km^2 (Table 1). The catchment is located in a mountainous region with steep slopes. Therefore, the altitude range is large with approx. 290–1170 m a.m.s.l. The catchment has a humid continental climate with warm summers and cool winters (Köppen-Geiger climate Dfb). In the observation period 1974–2006, measured mean annual temperature was 7.6 °C, precipitation was around 870 $mm\,yr^{-1}$, and potential evaporation was around 980 $mm\,yr^{-1}$ (Table 1). In winter, a continuous snow cover is present for, on average, 4 months, from December until the beginning of April, with large variation within the catchment due to elevation. Measured mean discharge was around 350 $mm\,yr^{-1}$. The low-flow season of Nedožery is summer, and highest flows occur in March due to snow melt (Table 1). Nedožery is a hardrock catchment consisting predominantly of impermeable metamorphic rocks without extensive groundwater storage, which makes it quickly responding to precipitation. The presence of steep slopes and absence of bogs or lakes even accelerates the response (Van Loon et al., 2010). Two-thirds of the catchment is covered by forest. Other land cover types are agriculture (23 %), natural meadow (6 %), and urban area (5 %) (Oosterwijk et al., 2009). Human influence is very limited in this catchment.

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2.5 Upper-Guadiana

The Upper-Guadiana catchment is located in Central Spain (Fig. 2f). It is the headwater catchment of the Guadiana. The area of the Upper-Guadiana catchment is approx. 16 480 km^2, which is considerably larger than the other catchments. This is chosen to rule out any significant groundwater transport over the catchment boundary (Table 1). The catchment is part of the Central Spanish Plateau. The altitude range is approx. 600–1100 m a.m.s.l. and especially in the centre topography is rather flat. The Upper-Guadiana catchment has a Mediterranean and semi-arid climate with very warm summers and mild winters (Köppen-Geiger climate Csa, Csb and Bsk; Acreman, 2000). In the observation period 1960–2001, catchment-average measured mean annual temperature was 14.1 °C, precipitation was 450 mm yr^{-1}, and potential evaporation was around 1250 mm yr^{-1} (Table 1). In winter, no continuous snow cover is present. Only in very cold years some snow accumulation occurs in the highest parts of the catchment. Potential evaporation exceeds precipitation, resulting in a relatively low measured mean discharge of 16 mm yr^{-1} (de la Hera, 1998). The low-flow season of Upper-Guadiana is summer due to a lack of recharge in this period, and highest flows occur in winter (Table 1). Upper-Guadiana is a groundwater catchment consisting of various areas with multiple layers of sedimentary rock (mainly gravel, limestone), forming large aquifer systems. This makes it a slowly responding catchment with a relatively high baseflow. A number of interconnected wetlands cause further delay in the response to precipitation. Land use in the Upper-Guadiana catchment is mainly agricultural. Since 1970–1980, agriculture intensified and human influence (i.e. irrigation) in the catchment increased dramatically, causing declining groundwater levels and wetland area and decreasing discharge (Veenstra, 2009).

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3 Methodology

Long time series of observations of all hydrometeorological variables were not available for the selected catchments, hence modelling was needed. Simulating low flows is a challenge. Smakhtin (2001) describes a number of difficulties in the modelling of low flows and Staudinger et al. (2011) state that "low flows are often poorly reproduced by commonly used hydrological models, which are traditionally designed to meet peak flow situations". For that reason we used a model that has proven to be robust in low-flow situations (Te Linde et al., 2008; Driessen et al., 2010), and a calibration criterion that is especially focused on low flows (both described in Sect. 3.1). On the simulated hydrometeorological variables we performed a drought analysis with the well-known threshold level method. This method and the way we applied it are explained in Sect. 3.2.

3.1 Hydrological model HBV

The conceptual, semi-distributed, rainfall-runoff model HBV (Seibert, 1997) is chosen as model for this research. The original HBV model was developed in the early 1970s by Bergström (1976, 1995). Afterwards, different versions of HBV have been developed for both research and operational management. Although it was originally developed for Scandinavian conditions, the HBV model has been widely used in general modelling studies (Lindström, 1997; Uhlenbrook et al., 1999; Perrin et al., 2001; Oudin et al., 2005) and in catchments in Europe: Austria (Merz and Blöschl, 2004), Belgium (Van Pelt et al., 2009; Driessen et al., 2010), Germany (Uhlenbrook et al., 1999; Nützmann and Mey, 2007), Sweden (Seibert, 1999; Seibert et al., 2003), and Ireland (Wang et al., 2006), and in other areas around the world, for example the Hindukush-Karakorum-Himalaya region (Akhtar et al., 2008), selected catchments in Africa and South-America (Lidén and Harlin, 2000). In this research we use the HBV model version developed by Seibert (1997, 2005). Seibert called it "HBV light", but for reasons of brevity it is referred to as "HBV" in the rest of this paper.

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HBV simulates daily discharge from daily precipitation and temperature, and monthly or daily estimates of potential evaporation. The model consists of four routines, i.e. a distributed snow routine and soil moisture routine, a lumped response routine, and a routing routine (Fig. 3). Snow accumulation and melt are calculated by the degree-day method for a number of elevation (max. 10) and vegetation (max. 3) zones separately. In each of these zones, groundwater recharge and actual evaporation are functions of actual water storage in the soil box. Subsequently, the lumped response function, in the STANDARD version consisting of two linear reservoirs in series, transforms recharge into discharge. Finally, channel routing is computed by a triangular weighting function. Further description of the model can be found in Seibert (2000, 2005).

Since according to Seibert (2000, 2005) the DELAY response routine is better suitable for modelling slowly responding deep-groundwater catchments, we used this version besides the STANDARD response routine for the Upper-Guadiana catchment. The DELAY response routine consists of two linear reservoirs in parallel, of which the lower reservoir is preceded by distribution of recharge over different delay boxes (Fig. 3).

The HBV model was forced with observed meteorological data of the selected catchments. Temperature and precipitation data were taken from meteorological stations inside or around the catchment (Fig. 2), and, if needed, averaged using Thiessen polygons. An altitude correction was applied to get correct input data for the elevation zones. Potential evaporation was calculated using the FAO Penman-Monteith method described by Allen et al. (1998). Due to different data availability and quality in each catchment, slightly different calculation procedures were followed according to the assumptions and recommendations described by Doorenbos and Pruitt (1975) and Allen et al. (1998).

Parameter values of HBV were determined by calibration. Calibration was done on observed discharge using the genetic calibration algorithm described by Seibert (2000). The agreement between simulated and observed discharge was evaluated by the Nash-Sutcliffe efficiency (Nash and Sutcliffe, 1970) based on the logarithm of

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observed and simulated discharge (ln Reff) (Seibert, 1999, 2005). The Nash-Sutcliffe efficiency based on the logarithm of observed and simulated discharge is regarded as the best objective function for low-flow modelling (Krause et al., 2005). The entire observation period (Table 1) was used as a calibration period for all catchments except Upper-Guadiana. Due to the strong human influence in that catchment after 1980 (see Sect. 2.5), the calibration period was restricted to the period 1960–1970, and the period 1970–1980 was used for validation.

For further drought analysis we used several output variables of HBV, i.e. catchment average precipitation (elevation corrected) in mm d^{-1}, soil moisture storage in mm, groundwater storage in mm, and discharge in mm d^{-1}. For groundwater storage we used only storage in the lower groundwater reservoir (ULZ, see Fig. 3), which represents deep groundwater. The reason for not including storage in the upper reservoir is that the fast flow paths in HBV (e.g. surface runoff) are modelled through this upper reservoir, hence it does not represent real groundwater storage (Fig. 3).

3.2 Drought analysis

To determine droughts from the simulated time series, the threshold level method (Yevjevich, 1967; Hisdal et al., 2004) was applied. With this method, a drought occurs when the variable of interest (e.g. precipitation, soil moisture, groundwater storage, or discharge) is below a predefined threshold (Fig. 4). A drought event starts when the variable falls below the threshold level and the event continues until the threshold is exceeded again. Each drought event can be characterised by its duration and by some measure of the severity of the event. For fluxes (e.g. precipitation and discharge) the most commonly used severity measure is deficit volume, calculated by summing up the differences between actual flux and the threshold level over the drought period (Hisdal et al., 2004; Fleig et al., 2006). For state variables (e.g. soil moisture and groundwater storage), we used the maximum deviation from the threshold (max. deviation) as severity measure (Fig. 4).

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Both a fixed and a variable (seasonal, monthly, or daily) threshold can be used. In this study, a variable threshold was chosen, as seasonal patterns are then taken into account. For drought management, not only the yearly recurring (summer or winter) low-flow period is important, but any deviation from the normal seasonal pattern (see definition of drought in Sect. 1). Furthermore, a variable threshold shows deficiencies in the high flow season that can lead to a drought in the low-flow season (Hisdal and Tallaksen, 2000). We applied a monthly threshold derived from the 80th percentile of the monthly duration curves. This implies that, for each month, the value of a flux or state variable is chosen, that is exceeded 80 % of the time in a specific month. For the Upper-Guadiana catchment the threshold values were calculated based on the period 1960–1980 and applied to the entire time series, to eliminate the strong human impact after 1980 (see Sect. 2.5). For the other catchments, the entire period was used for the calculation of the threshold. The discrete monthly threshold values were smoothed by applying a centred moving average of 30 days. After application of the threshold level method, mutually dependent droughts were pooled using the inter-event time method (Fleig et al., 2006). A inter-event time period of 10 days was used for all catchments, based on the range given by Tallaksen et al. (1997) and Fleig et al. (2006).

The calculation of drought characteristics of the pooled drought events (visualised in Fig. 4) is done according to Zelenhasić and Salvai (1987):

- pooled duration = $\text{duration}_i + \text{duration}_{i+1} + \ldots$
- pooled deficit volume = $\text{deficit volume}_i + \text{deficit volume}_{i+1} + \ldots$
- pooled max.deviation = $\max(\text{max.deviation}_i, \text{max. deviation}_{i+1}, \ldots)$

where: i is a hydrological drought event and $i+1$ is the following hydrological drought event.

To eliminate minor droughts, all drought events with a duration less than 15 days were excluded from the analysis (values up to 5 days are used by Hisdal et al., 2004; Birkel, 2005; Fleig et al., 2006; Van Loon et al., 2011a, but various studies showed

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that minor droughts can have durations up to 20 days; Hisdal, 2002; Fleig et al., 2005; Kaznowska and Banasik, 2011; Kim et al., 2011). Of the remaining drought events, a few were found to be not real drought events, but rather artefacts of the method used. A very sharp increase in discharge in combination with a gradually rising threshold level can result in a few days of below-threshold levels. This happens in catchments with a pronounced difference between wet and dry season, such as catchments with a pronounced snow melt peak or catchments with a monsoon climate. These events are not related to a rainfall deficit or temperature difference (so not caused by meteorological anomaly as defined by Stahl and Hisdal, 2004), but are purely a consequence of the smooth threshold level in combination with a sharp increase in groundwater storage or discharge. Therefore, in this research, we do not consider these events as drought, but rather as anomaly. In this research, such anomalies were only found in the Narsjø catchment (4 % of all events in groundwater and 7 % of all events in discharge). This is due to the very sharp increase in discharge during the snow melt season. In the other catchments with snow (Upper-Metuje, Upper-Sázava, and Nedožery) no such anomalies were found, because winters are less severe in those catchments, resulting in a less abrupt transition from winter to summer. As we did not study catchments with a monsoon climate, we did not find anomalies related to a sudden increase in precipitation. In the rest of this paper, anomalies are disregarded and focus is only on droughts.

4 Results

In this section, model performance will be described briefly (Sect. 4.1). Next, model results are used to study drought characteristics (Sect. 4.2) and to develop a typology of hydrological drought based on underlying processes (Sect. 4.3). Finally, hydrological drought events in the study catchments are attributed to the defined types (Sect. 4.4).

4.1 HBV model results and hydrological regime

After calibration, all selected catchments are modelled reasonably well with HBV (Table 2). In general, ln Reff values are (slightly) higher than Reff values, because calibration was based on ln Reff.

For the Narsjø catchment, model results show the highest ln Reff (0.90; Table 2). This is due to the very regular seasonal pattern of discharge, dominated by yearly recurring winter low-flow conditions, that can be captured quite well with a rainfall-runoff model like HBV (Van Loon et al., 2010).

For the Upper-Metuje, Upper-Sázava, and Nedožery catchments, ln Reff lays around 0.65 (Table 2). This is lower than the value for the Narsjø catchment because seasonal variation is much more irregular in these catchments. Still, performance of the HBV model in these catchments is acceptable for further drought analysis (Van Huijgevoort et al., 2010; Van Loon et al., 2010).

For the Upper-Guadiana catchment, the numbers in Table 2 are obtained with the DELAY version of the HBV model (Sect. 3.1 and Fig. 3) for the calibration and validation period (1960–1980). Model results of the STANDARD version, which was used for the other catchments, show a lower ln Reff than those of the DELAY version (0.51 instead of 0.71). A visual inspection of time series of the two model versions confirmed that the DELAY version reproduced recessions best. It showed less peaky behaviour and no zero-flows as compared to the STANDARD version. Therefore, the results of the DELAY version were used for further analysis in the Upper-Guadiana catchment.

4.2 Drought characteristics

General drought characteristics were determined from the HBV modelling results and observed discharge (Table 3). The drought events of simulated and observed discharge show similar characteristics (especially regarding number of drought events and mean duration), again indicating the reasonable performance of the HBV model on low flows. Only in the Upper-Guadiana catchment drought characteristics of simulated

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discharge deviate significantly from those of observed discharge. The reason is that drought characteristics of this catchment were calculated for the entire period (1960–2001), including the period with strong human influence. The drought characteristics of observed discharge reflect this disturbed situation, while those of simulated discharge represent a situation without human influence (as HBV does not simulate human influence, because it is calibrated on natural flows).

Table 3 confirms what is known about propagation in drought characteristics (Di Domenico et al., 2010; Van Loon et al., 2011b):

– Drought events become less and longer when moving from precipitation via soil moisture to groundwater storage, so the number of droughts decreases and the duration increases.

– Drought events in discharge have drought characteristics comparable to soil moisture, because they reflect both fast and slow pathways in a catchment.

– In fast reacting systems (like Narsjø and Nedožery), discharge drought characteristics are more comparable to those of precipitation (more and shorter); in slowly reacting systems (like Upper-Metuje and Upper-Guadiana) discharge drought characteristics are more comparable to those of groundwater storage (less and longer).

– Deficit volumes are higher for droughts in precipitation than for discharge droughts, because precipitation is higher and more variable, resulting in higher threshold values and a larger deviation from the threshold.

– Mean max.deviation is higher for soil moisture droughts than for droughts in groundwater, because soil moisture values are much more variable, while in groundwater the signal is smoothed. In the drought characteristics of the Narsjø catchment this effect is not visible, because soil water storage is limited in this catchment due to very coarse, shallow soils.

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The Narsjø and Nedožery catchment have similar drought characteristics because they are both fast reacting (Table 3). Narsjø is a bit slower (less, but longer groundwater droughts), due to the presence of bogs and lakes that slightly delay the response to precipitation. The Upper-Metuje and Upper-Sázava catchment have similar drought characteristics because they are both slowly reacting (Table 3). Upper-Metuje has an aquifer system with high storage and Upper-Sázava has many lakes that delay the response. The Upper-Guadiana catchment has very long hydrological droughts (groundwater drought events of, on average, longer than two years; Table 3). This is due to its very slow response to precipitation caused by the presence of extensive aquifer systems and wetlands, and to its dry climate.

The numbers in Table 3 show some differences between catchments that indicate propagation processes, but for a thorough insight into drought generating mechanisms time series of all hydrometeorological variables need to be studied in detail.

4.3 Typology of hydrological droughts

Based on an in-depth analysis of time series of hydrometeorological variables of the study catchments, a hydrological drought typology is proposed that uses the diversity of drought generating mechanisms as basic principle.

The following hydrological drought types are distinguished:

- classical rainfall deficit drought
- rain-to-snow-season drought
- wet-to-dry-season drought
- cold snow season drought
- warm snow season drought
- composite drought

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For each of these drought types, generating mechanisms are described and examples are presented.

4.3.1 Classical rainfall deficit drought

The *classical rainfall deficit drought* is caused exclusively by a prolonged lack of rainfall (meteorological drought) that propagates through the hydrological cycle and develops into a hydrological drought.

Some examples are shown in Fig. 5 with droughts in summer, spring and winter in different catchments. In the first example (Fig. 5a, Narsjø catchment), a meteorological drought in May–July 1992 (3rd panel) caused drought in soil moisture, groundwater storage, and discharge (4th, 5th, and 6th panel). The hydrological drought event ended with high precipitation in July–August 1992 (3rd panel). In the second example (Fig. 5b, Nedožery catchment), a meteorological drought in April–June 2000 and one in August 2000 (3rd panel) both caused a soil moisture drought (4th panel) and a hydrological drought (groundwater storage and discharge; 5th and 6th panel), with a small peak in between due to rainfall in July 2000 (3rd panel). The hydrological drought event ended with high precipitation in autumn (September–November 2000; 3rd panel). In the third example (Fig. 5c, Upper-Guadiana catchment), a meteorological drought in winter (February–March 1988; 3rd panel) caused only a minor drought in soil moisture (4th panel) and a hydrological drought (groundwater storage and discharge; starting in March 1988; 5th and 6th panel). The drought in soil moisture and discharge ended with rainfall in spring (March–June 1988; 3rd panel), but the drought in groundwater storage continued because recharge was not sufficient (5th panel).

The *classical rainfall deficit drought* can occur in any season, in any catchment (quickly and slowly responding), and in any climate region (Köppen-Geiger climate types A, B, C, D, and E), as long as precipitation falls as rain (snow related droughts are treated in Sects. 4.3.2, 4.3.4 and 4.3.5). A *classical rainfall deficit drought* can have all possible durations, deficit volumes, and max.deviations, mainly dependent on the rainfall deficit(s) that caused it and on the antecedent storage in the catchment. In the

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examples in Fig. 5, durations range from 28 to 245 days, max.deviations from 2.9 to 10.7 mm, and deficit volumes from 0.45 to 28 mm. *Classical rainfall deficit droughts* can show all propagation features (i.e. pooling, lag, attenuation, and lengthening; see Sect. 1), mainly dependent on catchment characteristics. Pooling, for example, often occurs. The examples in Fig. 5 show a clear propagation of one meteorological drought into one hydrological drought, but in many cases more meteorological droughts are pooled and it is harder to point out the exact rainfall deficits that caused a specific hydrological drought. In the examples in Fig. 5, lag (groundwater: 9–44 days, discharge: 7–39 days) and attenuation of the drought signal are visible in all catchments, and lengthening of the drought period is striking in the Nedožery catchment (Fig. 5b) and especially in the groundwater storage of Upper-Guadiana catchment (Fig. 5c).

The *classical rainfall deficit drought* is a very common hydrological drought type. As it occurs all around the world, it has been described and analysed by many different authors. Some examples are Tallaksen and Van Lanen (2004); Stahl and Hisdal (2004); Smakhtin and Hughes (2004); Fleig et al. (2006).

4.3.2 Rain-to-snow-season drought

The *rain-to-snow-season drought* is caused by a rainfall deficit (meteorological drought) in the rain season (usually summer and/or autumn) that continues into the snow season (usually winter). The meteorological drought ends with precipitation, which, however, falls as snow because the temperature has dropped below zero. Consequently, soil moisture and groundwater stores are not replenished by recharge in the rain season, the season in which recharge normally takes place. Therefore, the initial value of the normal winter recession is lower than normal and groundwater storage and discharge stay below the threshold level until the snow melt peak next spring.

Two examples of the *rain-to-snow-season drought* are shown in Fig. 6. In the first example (Fig. 6a, Narsjø catchment), the meteorological drought in July, August and September 1968 (3rd panel) directly resulted in a soil moisture drought (4th panel) and hydrological drought (5th and 6th panel). The precipitation peak that started

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mid-October (3rd panel) mainly fell as snow (2nd panel) because temperatures had dropped below zero (1st panel). Some replenishment of the soil moisture store took place and the soil moisture drought disappeared (4th panel), but the groundwater system remained in drought until the snow melt peak of May 1969 (5th panel). In the second example (Fig. 6b, Upper-Sázava catchment), two meteorological droughts of July and September–October 1969 (3rd panel) caused groundwater storage (5th panel) and discharge (6th panel) to decrease below threshold level. Part of the precipitation of November 1969 and almost all that of February 1970 (3rd panel) fell as snow (1st and 2nd panel). Therefore, the hydrological drought did not end, but continued until the snow melt period in April 1970 (6th panel). In the groundwater subsystem, the drought even continued longer, until July 1970 (not shown).

The *rain-to-snow-season drought* occurs in catchments with a clear snow season, which can be catchments at high latitude or at high elevation (Köppen-Geiger climate types D and E, and some subtypes of C). These catchments have a low-flow season in winter due to the continuous snow cover that hampers recharge. Durations of *rain-to-snow-season droughts* are long (almost up to a year; in the examples of Fig. 6, 279 and 147 days for drought in discharge) and deficit volumes can be high (partly due to the long durations; in the examples of Fig. 6, 54 and 11 mm for drought in discharge). As can be seen from the examples in Fig. 6, lengthening is the main drought propagation feature defining *rain-to-snow-season droughts*. Other drought propagation features also occur (e.g. pooling and lag in Fig. 6b), but are less important than lengthening.

The *rain-to-snow-season drought* has previously been described by Van Loon et al. (2010) under the name Type 1 winter drought. Pfister et al. (2006) mention historical evidence of a hydrological winter drought event in 1540 that might have been of this type. In other studies, these multi-season droughts are mostly filtered out, because they complicate statistical analysis (Hisdal et al., 2001; Fleig et al., 2006).

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4.3.3 Wet-to-dry-season drought

The *wet-to-dry-season drought* is governed by the same principle as the *rain-to-snow-season drought*, only in this case no snow is involved, but a very high potential evaporation in the dry season. The *wet-to-dry-season drought* is caused by a rainfall deficit (meteorological drought) in the wet season (usually winter) that continues into the dry season (usually summer). The meteorological drought ends with precipitation, which, however, is completely lost to evapotranspiration because potential evaporation in this season is higher than precipitation. Consequently, soil moisture and groundwater stores are not replenished by recharge in the wet season, the season in which recharge normally takes place. Therefore, the initial value of the normal summer recession is lower than normal and groundwater storage and discharge stay below the threshold level until the next wet season.

Two examples of the *wet-to-dry-season drought* are shown in Fig. 7 (both Upper-Guadiana catchment; in the other studied catchments the potential evaporation is not sufficiently high). In the first example (Fig. 7a), one large meteorological drought in the wet season (April–June 1987; 3rd panel) caused discharge to drop below the threshold level (6th panel). Groundwater was already in drought (5th panel) as remnant of a previous dry period. The rainfall event of June–July 1987 (3rd panel) did not result in recovery from the hydrological drought, because it was partly lost to evapotranspiration and partly used for replenishment of soil moisture (4th panel). The hydrological drought continued until December 1987 (6th panel), when rainfall was high (3rd panel) and potential evaporation lower than in summer. In the second example (Fig. 7b), a number of small meteorological drought events in the wet season (between November 1998 and May 1999; 3rd panel) resulted in a soil moisture drought in the wet season (4th panel) and a decrease in groundwater storage and discharge to below-threshold levels (5th and 6th panel). In both examples, the hydrological drought continued throughout the dry season, until the first recharge in the following wet season (November–December).

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The *wet-to-dry-season drought* occurs in catchments with a clear wet and dry season (Köppen-Geiger climate subtypes A – monsoon climate, B – steppe climate – and C – Mediterranean climate). Durations are long (half a year to a year; in the examples of Fig. 7, 222 and 243 days for drought in discharge), and deficit volumes can be high in wet climates and often stay low in semi-arid climates because of the low threshold level (in the examples of Fig. 7, 3.0 and 2.7 mm for drought in discharge). Just as *rain-to-snow-season droughts*, lengthening is the main drought propagation feature defining *wet-to-dry-season droughts*. Other drought propagation features also occur (e.g. pooling and lag in Fig. 7b), but are less important than lengthening.

The *wet-to-dry-season drought* has previously been described by Van Lanen et al. (2004); Stahl and Hisdal (2004); Trigo et al. (2006); Santos et al. (2007); Trigo et al. (2010); Kim et al. (2011).

4.3.4 Cold snow season drought

The *cold snow season drought* is caused by an abnormally low temperature in the snow season (winter), possibly, but not necessarily, in combination with a meteorological drought in that same season. Three subtypes are distinguished, subtype A and B in cold climates and subtype C in temperate climates.

Subtype A In climates with temperatures well below zero and a continuous snow cover in winter (Köppen-Geiger climate types D and E), a below-normal winter temperature only influences the beginning and end of the snow season. If temperatures are low during the beginning of winter, temperatures drop below zero earlier in the year than normal and precipitation falls earlier as snow. This causes the normal winter recession period to start earlier than normal. When the initial value of the recession of soil moisture, groundwater storage, and discharge is high enough this will not lead to drought (see Sect. 4.4.3), but when storage and discharge were already low, groundwater storage and discharge can go below threshold levels during winter. An example is shown in Fig. 8a (Narsjø catchment). In this case, temperature decreased below zero two weeks early, in the beginning of October instead of the end of October 1960 (1st

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panel), and the precipitation of October fell as snow (2nd and 3th panel). The recession of groundwater storage and discharge started earlier than normal and the values dropped just below threshold level from November 1960 to February 1961 (5th and 6th panel). The hydrological drought ended with some snow melt in March 1961, caused by high temperatures (1st panel). *Cold snow season droughts – subtype A* usually have a long duration (several months), but a low deficit volume and small max.deviation because groundwater storage and discharge are just below the threshold level. In the example in Fig. 8a, durations are 83 and 93 days for groundwater storage and discharge, respectively, and deficit volume of discharge is only 1.6 mm. Drought propagation features are not applicable, because this type of hydrological drought is not caused by a meteorological drought (P-control), but only by a temperature anomaly (T-control).

Subtype B If, in the same cold climates, temperatures are low at the end of winter, snow melt is later than normal. A late snow melt leads to below-threshold levels when groundwater storage and discharge stay low while threshold levels increase. An example is shown in Fig. 8b (Narsjø catchment). In this case, temperature stayed below zero until the beginning of May instead of mid-April (three weeks later than normal; 1st panel) and snow melt was delayed (2nd panel). The threshold levels started to increase by mid-April, while groundwater storage and discharge still showed a recession (5th and 6th panel). When temperature finally increased above zero in the beginning of May (1st panel), snow melt (2nd panel) ended the hydrological drought (5th and 6th panel). *Cold snow season droughts – subtype B* can have high deficit volumes (in the example 15.2 mm), but only short durations, in the order of a few weeks (in the example about three weeks). This type of drought is mostly confined to discharge and is usually not found in groundwater. Again, drought propagation features are not applicable. This specific case of *cold snow season drought* should not be confused with a snow melt anomaly, which does not have an abnormal temperature pattern, but is only caused by the very sharp increase in discharge in combination with a gradually rising threshold level (see Sect. 3.2).

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Subtype C In climates with temperatures around zero and some snow accumulation in winter (Köppen-Geiger climate types C and some subtypes of D), the effect is different. In these climates, the snow season normally provides recharge to the groundwater system, due to occasional and partial melt of the snow cover. So, the normal winter situation is one of increasing storage and discharge. If, however, winter temperatures decrease to values well below zero and no melting of snow takes place, recharge decreases to zero. If low temperatures persist a hydrological drought can develop. This is clearly visible in Fig. 8c (Upper-Metuje catchment). In December 1995 to April 1996 temperatures were lower than normal (on average $-3.9\,^{\circ}$C instead of $-0.4\,^{\circ}$C; 1st panel) and snow accumulation was higher than normal (2nd panel). The lack of recharge caused a decrease in groundwater storage and discharge, leading mid-February to drought in discharge (6th panel) and mid-March to drought in groundwater (5th panel). The drought ended with snow melt. A *cold snow season drought – subtype C* typically has a duration of a few weeks to months (in this example 60 days in groundwater and 47 days in discharge) and an intermediate deficit volume (in this example 4.4 mm). Again drought propagation features are not applicable, although the reaction of groundwater can be different from that of discharge (delayed and attenuated, like in Fig. 8c).

Stahl and Demuth (1999) and Pfister et al. (2006) mention a cold winter as a reason for drought, but do not describe underlying processes. Van Lanen et al. (2004) discuss causative mechanisms of *cold snow season drought – subtypes A, B and C*.

4.3.5 Warm snow season drought

The *warm snow season drought* is caused by an abnormally high temperature in the snow season (winter), in some cases in combination with a rainfall deficit (meteorological drought) in that same season. Two subtypes are distinguished, subtype A in cold climates and subtype B in temperate climates.

Subtype A In climates with temperatures well below zero and a continuous snow cover in winter (Köppen-Geiger climate types D and E), a higher winter temperature again only influences the beginning and end of the snow season. If temperatures are high during the beginning of winter, more precipitation will fall as rain instead of snow and a drought in the snow season will be less likely (see Sect. 4.4.3). However, if temperatures are high at the end of winter, snow melt is earlier than normal. An early snow melt leads to an early peak in discharge, resulting in lower discharge values in the following normal snow melt period. Discharge can drop below the (high) threshold level. If a rainfall deficit occurs in the spring season, it can aggravate this *warm snow season drought*. In the example in Fig. 9a (Narsjø catchment), temperature increased to above zero three weeks early, end of March 2004 instead of mid-April (1st panel), resulting in an early snow melt (2nd panel). Consequently, the peak in discharge (normally in June) was advanced to April–May and in June a hydrological drought developed (6th panel), because threshold levels were high and discharge already decreased after the snow melt peak. So, a *warm snow season drought – subtype A* can develop without a meteorological drought (although precipitation was not extremely high in May 2004; Fig. 9a). The reason is the normally-occurring pronounced snow melt peak in this catchment that is clearly reflected in the threshold level. *Warm snow season droughts – subtype A* usually have short durations (in the example in Fig. 9a, 25 days). Deficit volumes can be high (in the example 8.2 mm) due to the high threshold level during the snow melt peak. A *warm snow season drought – subtype A* is mostly confined to discharge and is usually not found in groundwater. Again drought propagation features are not applicable, because this type of hydrological drought is not caused by a meteorological drought (P-control), but only by a temperature anomaly (T-control).

Subtype B In climates with temperatures around zero and some snow accumulation in winter (Köppen-Geiger climate types C and some subtypes of D), the effect is different. In these climates the snow season normally provides recharge to the groundwater system, due to occasional and partial melt of the snow cover. If, however, winter temperatures rise above zero and the snow cover melts completely, no snow store is left that can provide recharge. If, at the same time, a meteorological drought occurs, a hydrological drought can develop. Two examples of this case of the *warm snow* *season drought* are shown in Fig. 9. In the first example (Fig. 9b, Upper-Sázava catchment), the warm and dry period of February–March 1974 (1st and 3rd panel) caused a complete melt of the snow cover (2nd panel) and afterwards a lack of recharge to groundwater. Consequently, a hydrological drought developed (5th and 6th panel) that continued until the high rainfall period in the spring of 1974 (3rd panel). In the second example (Fig. 9c, Nedožery catchment), the high temperatures of December 1989 to March 1990 (1st panel) also led to a complete melt of the snow cover (2nd panel). The meteorological drought of December 1989–January 1990 (3rd panel) could therefore trigger a soil moisture (4th panel) and hydrological drought (5th and 6th panel). The rainfall peak in March 1990 (3rd panel) caused a quick reaction in discharge (6th panel), but did not end the drought that continued until May–June 1990. That spring, no snow melt peak occurred because the snow cover had already melted in December (2nd panel). So, contrary to the *rain-to-snow-season drought*, the *cold snow season drought – subtypes A–C*, and the *warm snow season drought – subtype A* that are also winter droughts (Sects. 4.3.2, 4.3.4, and 4.3.5), the *warm snow season drought – subtype B* is not ended with a snow melt peak, because snow cover already melted before. A *warm snow season drought – subtype B* can continue into summer. Durations can be long and deficit volumes high. *Warm snow season droughts – subtype B* can show all propagation features (i.e. pooling, lag, attenuation, and lengthening; see Sect. 1), mainly dependent on catchment characteristics.

The *warm snow season drought – subtype A* has previously been described by Van Lanen et al. (2004), and *subtype B* by Van Loon et al. (2010) under the name Type 2 winter drought.

4.3.6 Composite drought

A *composite drought* combines a number of drought generating mechanisms. In this hydrological drought type a number of drought events (of the same or different type) in distinct seasons cannot be distinguished any more. The main feature of the *composite*

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drought is that the system has not recovered from a hydrological drought event, when the next event starts.

Examples of the *composite drought* are shown in Fig. 10. The first example (Fig. 10a, Upper-Metuje catchment) shows two *classical rainfall deficit droughts* in subsequent summers (1982 and 1983, 3rd panel) that are combined into one hydrological drought (5th and 6th panel). The drought in groundwater started in July 1983 and lasted for 440 days. The drought in discharge was interrupted by some small rainfall peaks in December 1982 and January 1983, and a snow melt peak in April 1983, but every time it returned to below-threshold levels afterwards. In total, the drought in discharge had a net duration of 330 days and a deficit volume of 22.2 mm. The hydrological drought ended with high precipitation events by the end of 1984. In the second example (Fig. 10b, Upper-Sázava catchment), the hydrological drought that lasted from December 1989 to August 1991 (5th and 6th panel) was caused by two *warm snow season droughts – subtype B* in the winter of 1989–1990 and 1990–1991 (1st, 2nd and 3rd panel) and a *classical rainfall deficit drought* in the summer of 1990 (3rd panel). The precipitation peaks in between caused small discharge peaks that interrupted the hydrological drought, but afterwards discharge returned to its low level. In the third example (Fig. 10c, Upper-Guadiana catchment), a large number of *classical rainfall deficit droughts* (3rd panel) and *wet-to-dry-season droughts* (3rd and 4th panel) in subsequent years are combined into a very long hydrological drought (5th and 6th panel). The drought in groundwater lasted for 2126 days (March 1990 until January 1995). In discharge, a number of separate drought events can still be distinguished, for example a *wet-to-dry-season drought* from February to October 1990, and a *classical rainfall deficit drought* from December 1990 to March 1991.

Composite droughts only occur in catchments with a long memory, so catchments with considerable storage. This storage can be in e.g. aquifers, bogs, lakes. *Composite droughts* can occur in all climates, but are most likely in (semi-)arid climates (Köppen-Geiger climate type B) due to the irregular rainfall in these climates. The drought types that are combined differ per catchment and climate zone. *Composite droughts*

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have long to very long durations (often multi-year) and deficit volumes are high (for the examples in Fig. 10, 20–40 mm in total). The main drought propagation feature defining *composite droughts* is pooling and this type of drought is especially pronounced in groundwater and less in discharge.

The *composite drought* has previously been mentioned by Marsh et al. (2007) and analysed by Van Loon et al. (2011a) under the name Multi-year drought.

4.4 Occurrence of hydrological drought types in study catchments

4.4.1 All drought events

Some of the hydrological drought types defined in Sect. 4.3 occur in all catchments, others only in one or two of the studied catchments. That is because some hydrological drought types are specific for a certain climate type (e.g. *rain-to-snow-season drought* and *wet-to-dry-season drought*) or for a certain catchment type (e.g. *composite drought*). Table 4 shows that the *classical rainfall deficit drought* occurred in all studied catchments and the *wet-to-dry-season drought* only in one (Upper-Guadiana). The other drought types occurred in more than one of the studied catchments, but in different percentages.

Drought events in groundwater and discharge show a comparable distribution over the drought types (Table 4). Droughts in discharge only show up in more categories than droughts in groundwater because the total number of droughts in discharge is higher (Table 3), resulting in higher possibility for different drought types. In groundwater these drought events have grown together and formed a *composite drought*. Consequently, the percentage of *composite droughts* in groundwater is, in general, higher than that of discharge (Table 4; exception Upper-Sázava). Furthermore, *warm-snow-season droughts* are more clearly visible in discharge than in groundwater, because these droughts are easily attenuated in the stores.

The *classical rainfall deficit drought* occurred in all studied catchments with percentages often around 50 % (Table 4). This is the most common hydrological drought type in these catchments. Only in the groundwater drought events of the Upper-Guadiana catchment the *classical rainfall deficit drought* is not recognisable any more because it is included in *composite droughts*.

The *rain-to-snow-season drought* occurred only in catchments with a clear snow season, i.e. Narsjø, Upper-Metuje, Upper-Sázava, and Nedožery. Percentages are relatively low (7 to 19 %; Table 4).

The *wet-to-dry-season drought* occurred only in Upper-Guadiana, because that is the only studied catchment with a clear dry season in which potential evaporation exceeds precipitation (Cs and Bs climate; Table 1).

The *cold snow season drought* occurred in all studied catchments, but with varying percentages. The 3 % of the Upper-Guadiana catchment reflect only one event in the time series of 42 yr. This was an extremely cold winter (1970–1971) with considerable snow accumulation. The large number of *cold snow season droughts* in the Narsjø catchment are due to an early start of the snow season (subtype A) or a late end (subtype B). The *cold snow season droughts* in Upper-Metuje, Upper-Sázava and Nedožery catchments are mostly due to a lack of recharge in winter (subtype C) and sometimes to a late end of the snow season (subtype B).

The *warm snow season drought* is not represented in the Upper-Guadiana catchment, because of its warm climate. In the Narsjø catchment, some *warm snow season drought – subtype A* occurred, but only in discharge. In the catchments with temperatures around or just below zero in winter (e.g. Upper-Metuje, Upper-Sázava, Nedožery), most *warm snow season drought* were found (around 20 % occurrence). These were all *subtype B* droughts.

The *composite drought* occurred in slowly responding catchments, with the highest percentage in Upper-Guadiana (67 % for groundwater droughts) and lower percentages in Upper-Metuje and Upper-Sázava (7 to 19 %). Upper-Guadiana had very long droughts that span over different seasons and even years (Table 3) due to the long memory in its extensive groundwater system.

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A few events are not included in Table 4 (causing percentages of some catchments not to add up to 100 %). In the Narsjø catchment these omitted events are classified as anomalies (and thus disregarded, see Sect. 3.2) and in the Upper-Guadiana catchment a few events were unidentifiable, because they were a remnant drought from low storage in groundwater that did not have a clear cause in precipitation or temperature. In these events, discharge returned to a drought situation after a small peak caused by a rainfall event or snow melt.

If drought characteristics of all discharge drought events in the five studied catchments are grouped by drought type (Fig. 11), some drought types stand out. Especially *rain-to-snow-season droughts*, *wet-to-dry-season droughts*, and *composite droughts* show a distinct pattern with short duration and high deficit volume for *rain-to-snow-season droughts*, and long duration and low deficit volume for *wet-to-dry-season droughts* and *composite droughts*. *Classical rainfall-deficit droughts*, *cold snow season droughts*, and *warm snow season droughts* show large overlap. Most events of these types are have relatively short durations and low to intermediate deficit volumes. Hence, although processes underlying these drought types are different, drought characteristics are comparable.

In Fig. 12, the same discharge drought events are plotted with more detail (one plot for each drought type and a different colour for quickly and slowly responding catchments). For each drought type, the events in slowly responding catchments have, in general, somewhat longer durations and lower deficit volumes than those in quickly responding catchments. *Wet-to-dry-season droughts* and *composite droughts* were only found in slowly responding catchments. *Composite droughts* do not occur in quickly responding catchments. *Wet-to-dry-season droughts* presumably do occur in quickly responding catchments, but in this study no quickly responding catchment with semi-arid climate was included.

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4.4.2 Most severe drought events

Because Table 4 includes many small drought events that affect the distribution over the drought types, we selected the five most severe drought events for each catchment. The selection was done based on max.deviation for groundwater and on deficit volume for discharge. Table 5 shows that the distribution of hydrological drought events over the different drought types changes significantly after this selection. The *classical rainfall deficit drought* is represented less in most catchments (in total for all catchments together, from 22 to 12 % in groundwater, and from 43 % to 32 % in discharge; not shown). The *cold snow season drought* disappeared almost completely from the list, because this drought type usually has low deficit volumes. A large part of the most severe drought events are *rain-to-snow-season droughts* (up to 80 % for the Narsjø catchment). The reason is that these droughts are usually very long and can build up a large deficit volume. For the same reason *composite droughts* are more represented in the most severe drought events.

When drought events are classified according to their duration and the five longest drought events are selected, the distribution over the drought types is similar to Table 5 (not shown).

Based on Table 5, we can conclude that the most severe hydrological droughts are:

- in snow catchments: *rain-to-snow-season drought* and *warm snow season drought*
- in semi-arid climates: *wet-to-dry-season drought*
- in quickly responding catchments: *classical rainfall deficit drought*
- in slowly responding catchments: *composite drought*

The *cold snow season drought* occurs regularly, but is usually not severe.

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4.4.3 Non-drought development

Up to now we only discussed situations in which meteorological droughts developed into hydrological droughts. For process understanding and drought management, it is also relevant to study situations when a hydrological drought did not develop. Why did a rainfall deficit not propagate through the hydrological cycle? Which processes are involved that buffer or counteract the drought?

In snow climates a number of processes can prevent a hydrological drought to develop. One example is the situation where a rainfall deficit in the spring season coincides with the snow melt period. In that case no hydrological drought will develop, because water availability is very high. If this same rainfall deficit would have occurred a few months later, a *classical rainfall deficit drought* would have developed. On the other hand, a warm winter and an early snow melt could lead to a *warm snow season drought – subtype A*, but not if it is combined with very high rainfall amounts during the normal snow melt season (Sect. 4.3.5). A warm winter can also have another effect in snow climates, namely a late start of the snow season (Sect. 4.3.5). This can prevent a *rain-to-snow-season drought* from developing. An example is shown in Fig. 13a (Narsjø catchment). The rainfall deficit in September 2000 (3rd panel) resulted in just below-threshold levels in groundwater storage and discharge (5th and 6th panel). If temperatures would have dropped below zero in October, like they normally do, the precipitation peak in October–November 2000 (3rd panel) would have fallen as snow and groundwater storage and discharge would have stayed below the threshold until the next snow melt season. In this case, however, temperature dropped below zero only at the end of November (1st panel), hence the aforementioned precipitation peak could alleviate the hydrological drought, and the meteorological drought did not develop into a *rain-to-snow-season drought*.

In slowly responding catchments, attenuation is a well-known drought propagation feature. Meteorological drought events are often attenuated in the stores and no hydrological drought develops. An example is shown in Fig. 13b (Upper-Guadiana

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catchment). The rainfall deficit in February 1961 (3rd panel) led to a drought in soil moisture (4th panel) and to a decrease in groundwater levels and discharge (5th and 6th panel), but high groundwater storage prevented both variables to fall below threshold level. If antecedent storage would have been low, a *wet-to-dry-season drought* would have developed, like in the examples in Fig. 7. Attenuation of a meteorological drought can also occur in quickly responding catchments, but only after a very wet period (e.g. after extensive rainfall or snow melt). The rainfall deficit in September–October 1985 in Fig. 13c (Nedožery catchment; 3rd panel) would have developed into a *classical rainfall deficit drought*, but due to the very wet condition of the catchment after the extensive rainfall in the previous months (5th and 6th panel), the recession of groundwater storage and discharge did not drop below the threshold level.

Also a combination of processes can prevent a meteorological drought to develop into a hydrological drought. The example in Fig. 13d (Upper-Metuje catchment) could have become *warm snow season drought* (above-zero temperatures in the snow season, melt of the snow cover, and additionally a rainfall deficit in January 1989), but the snow melt peak had increased groundwater storage and discharge to such high levels that the warm and dry winter did not have much effect.

From these examples we learn that both precipitation and temperature, and antecedent storage in the catchment are important factors that can prevent a hydrological drought from developing.

5 Discussion

5.1 Typology

In this paper we proposed a hydrological drought typology based on drought propagation processes. Table 6 summarises the governing processes of the six hydrological drought types.

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Because division into types is based on the interpretation of time series of hydrometeorological variables, the boundaries between drought types are not sharp. Subjective choices cannot be avoided, for example when several processes are involved in the development of a hydrological drought. This is not a major drawback, as the typology should be used for process understanding, to study differences between catchments, and as a general tool for drought management. Therefore, the exact number of drought events of certain type for a specific catchment is not relevant, but rather the general occurrence of drought types in a catchment and the drought type of the most severe drought events. We propose that for events where more processes play a role, the dominant one determines the drought type.

The drought propagation features on which the typology is based, are determined by climate and catchment control (see Sect. 1). In Sects. 4.2, 4.3, and 4.4, these controls have already been used to describe drought characteristics, different hydrological drought types, and the occurrence of these types in the study catchments. In the following sections, catchment and climate control and their relation with the defined hydrological drought types are discussed in more detail.

5.2 Catchment control

For drought propagation, catchment control is very important. Lag and attenuation, but also pooling and lengthening, are determined by catchment characteristics like geology (Vogel and Kroll, 1992; Mishra and Singh, 2010), area (Rossi et al., 1992; Byzedi and Saghafian, 2009), mean slope, percentage of lakes and forest (Demuth and Young, 2004). These propagation features occur in all hydrological drought types but show up most prominently in *composite droughts*. In Sect. 4.4, we saw that *composite droughts* only occur in slowly responding catchments and that this drought type is amongst the most severe events. The governing factor is a catchment's reaction to precipitation, which is mainly determined by the amount of storage in the catchment. This storage can be in groundwater (like in Upper-Metuje and Upper-Guadiana catchments), in lakes (like in Upper-Sázava catchment), or in bogs (like in Narsjø catchment).

Very striking is that, in catchments with high storage, where a very smooth discharge signal is expected, peaks in discharge still often occur as a reaction to a precipitation event (see Figs. 9 and 10). These peaks interrupt the drought event, but do not lead to full recovery from the drought. After the peak, discharge returns to its very low values. This was also found by Woo and Tariiule (1994), who state that "brief inter-event streamflow rises will seldom ameliorate a drought event". Pooling is therefore a crucial step in drought analysis to prevent separation of drought events that are actually caused by the same process.

Figure 14 shows that the *composite drought* is the only drought type that is primarily controlled by catchment characteristics (the x-axis in Fig. 14). The other drought types are mainly controlled by climate (the y-axis in Fig. 14).

5.3 Climate control

The effect of climate on hydrological drought types is divided into the influence of general climatology and the influence of the weather pattern.

General climatology The general climatology determines the occurrence of specific drought types in certain regions (Stahl and Hisdal, 2004; Sheffield and Wood, 2007) and is governed by climatic variables like mean annual temperature and mean annual precipitation (Rossi et al., 1992; Demuth and Young, 2004). The occurrence of drought types in climate regions is indicated in Sect. 4.3, Table 6, and Fig. 14. *Classical rainfall deficit droughts* occur in all climates and *wet-to-dry-season droughts* only in climates with strong seasonal variation. The three snow-related drought types occur in a similar range of climates from temperate to continental and polar (Fig. 14).

The hydrological drought typology is developed using five catchments with different climate in Europe. These catchments are indicted in Fig. 14, based on their climate and catchment characteristics. Because the typology is based on generally observable processes, it can also be used in catchments that fall outside the reach of the studied catchments (for example in the lower-left corner of Fig. 14). Adding more catchments with different climate and catchment characteristics to the framework of Fig. 14 is an

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interesting way forward in drought research. Focus can then be on e.g. tropical climates and quickly responding catchments in steppe or monsoon climates. This can be achieved using data of real catchments or synthetic data, following the approach of Van Lanen et al. (2011). This newly-developed approach also allows for a better quantifcation of the effect of catchment and climate control on drought propagation and drought typology.

Weather pattern The weather pattern determines the development of a hydrological drought event of a certain type in a certain catchment. Precipitation and temperature are key variables. Table 6 shows whether the hydrological drought types are determined by precipitation (P-control), temperature (T-control), or a combination of precipitation and temperature (P and T-control).

By studying hydrological droughts in different catchments, we found that the influence of precipitation is different in different regions. In (semi-)arid climates, for example, long-term precipitation amounts are important. Rainfall in these climates is little and very irregular. A relatively dry period can last for years or decades (Vicente-Serrano and López-Moreno, 2006), leading to very low storage. *Composite droughts* are the result. Also in other catchments we found that droughts tend to cluster in time: periods with little drought events are alternating with periods with many drought events, which is consistent with other studies (Stahl and Hisdal, 2004; Uhlemann et al., 2010). In Central Europe, for example, the first half of the 1980's, the 1990's, and the 2000's were dry periods and the periods in between were relatively wet (Tallaksen and Van Lanen, 2004). This clustering of meteorological droughts is important for propagation. An isolated meteorological drought might be attenuated in the stores (Sect. 4.4.3), but a number of successive meteorological droughts decrease storage and a severe hydrological drought can develop. In that light, not only low precipitation events are important for the development of hydrological drought. Also high precipitation events should be included in drought analysis, as they can prevent a drought from developing due to high storage in the catchment (see Sect. 4.4.3), or cause the end of a drought (in case of drought types not related to snow, e.g. Sect. 4.3.1).

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A sustained lack of precipitation is usually governed by large-scale circulation patterns. Therefore, many studies that focus on hydrological drought include atmospheric circulation patterns, e.g. correlation with ENSO (Kingston et al., 2010; Lavers et al., 2010), weather types (Phillips and McGregor, 1998; Fowler and Kilsby, 2002; Fleig et al., 2010, 2011), blocking high-pressure areas (Stahl and Demuth, 1999; Stahl, 2001; Stahl and Hisdal, 2004; Pfister et al., 2006). These large-scale circulation patterns determine the timing of a precipitation event and whether it is high or low, which is crucial for drought development.

Temperature is also determined by large-scale circulation patterns (Domonkos et al., 2003; Xoplaki et al., 2003). But, because the development of snow-related hydrological drought types is very sensitive to a narrow temperature range around zero, elevation also plays an important role in those drought types. Two catchments in the same region can have different drought type occurrence when they have a different elevation. For example, in the higher catchment a *rain-to-snow-season drought* can develop because precipitation already falls in the form of snow, while in the lower catchment the hydrological drought is ceased due to rainfall. Synchronicity of droughts within a region, therefore, mainly happens with drought types that are precipitation controlled (i.e. *classical rainfall deficit drought* and *wet-to-dry-season drought*) and less with those that are temperature controlled (i.e. *rain-to-snow-season drought*, *cold snow season drought*, and *warm snow season drought*). In catchments with a large elevation range, variability of drought development within the catchment can occur, as the timing of when and for how long temperatures decrease below zero is variable within the catchment. This effect is the reason that discharge peaks can occur when the catchment-average temperature is still below zero.

In this study, potential evaporation is found not to be a major factor governing the development of different hydrological drought types. The reason is that even in situations when potential evaporation is higher than normal, actual evaporation is low due to lack of available water. In regions with very high water availability (e.g. some subtypes of Köppen-Geiger climate type A) an increase in potential evaporation might have

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more influence (Van Lanen et al., 2004). For the presented drought typology, potential evaporation is only important in climatic perspective: in catchments with a season in which potential evaporation is higher than precipitation, *wet-to-dry-season droughts* can occur.

In many papers a distinction is made between summer and winter droughts. The term summer drought is mostly used referring to a *classical rainfall deficit drought*. The term winter drought, however, is less clear. It covers a number of drought types (*rain-to-snow-season drought*, *cold snow season drought*, *warm snow season drought*, or even *classical rainfall deficit drought*) and drought generating processes are not well addressed if winter drought is defined as a drought in the winter half year (Pfister et al., 2006).

Climate change will probably lead to a change in occurrence of drought types (Feyen and Dankers, 2009), because in a higher temperature regime the Köppen-Geiger climate regions will shift to higher latitudes and higher elevations and the associated hydrological drought types will shift along. This can have strong implications for drought management. For example, a drought type that normally ends with a snow melt peak might change into a drought type that can continue into summer (Van Loon et al., 2010).

6 Conclusions

In this paper, we propose a general hydrological drought typology based on underlying processes of drought propagation. The typology can be used in research and management. Drought research could benefit from a common terminology, which can also guide further study of the processes underlying drought. Drought management is supported because different drought types need different preventing measures and coping mechanisms. The hydrological drought types that have been distinguished are: (i) *classical rainfall deficit drought*, (ii) *rain-to-snow-season drought*, (iii) *wet-to-dry-season*

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drought, (iv) *cold snow season drought*, (v) *warm snow season drought*, and (vi) *composite drought*.

- *Classical rainfall deficit droughts* are caused by a rainfall deficit (in any season) and occur in all climate types.
- *Rain-to-snow-season droughts* are caused by a rainfall deficit in the rain season, after which the hydrological drought continues into the snow season because temperatures have decreased below zero, and occur in catchments with a pronounced snow season.
- *Wet-to-dry-season droughts* are caused by a rainfall deficit in the wet season, after which the hydrological drought continues into the dry season, when potential evaporation is much higher than precipitation, and occur in catchments with pronounced wet and dry seasons.
- *Cold snow season droughts* are caused by a low temperature in the snow season. In catchments with a very cold winter, *subtypes A and B* occur, which are caused by an early beginning of the snow season and a delayed snow melt, respectively. In catchments with temperatures around zero in winter, *subtype C* occurs, which is caused by a lack of recharge due to snow accumulation.
- *Warm snow season droughts* are caused by a high temperature in the snow season. In catchments with a very cold winter, *subtype A* occurs, which is caused by an early snow melt. In catchments with temperatures around zero in winter, *subtype B* occurs, which is caused by a complete melt of the snow cover in combination with a subsequent rainfall deficit.
- *Composite droughts* are caused by a combination of hydrological drought events (of the same or different drought types) over various seasons and can occur in all climate types, but are most likely in (semi-) arid climates and slowly responding catchments.

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About 125 groundwater droughts and about 210 discharge droughts of five contrasting headwater catchments in Europe have been classified using the developed topology. The most common drought type in all catchments is the classical rainfall deficit drought (almost 50 % of all events), but these are mostly minor events. If only the five most severe drought events of each catchment are considered, a shift towards more *rain-to-snow-season droughts*, *warm snow season droughts*, and *composite droughts* is found. The occurrence of drought types is determined by climate and catchment characteristics. The typology is transferable to catchments outside Europe, because it is generic and based upon processes that occur around the world. A general framework is proposed that enables to identify the occurrence of hydrological drought types in relation to climate and catchment characteristics.

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Table 1. Catchment characteristics of the selected catchments Narsjø (Norway), Upper-Metuje and Upper-Sázava (Czech Republic), Nedožery (Slovakia), and Upper-Guadiana (Spain); obs. period = observation period, T = temperature, P = precipitation, PET = potential evaporation, Q = discharge.

	Narsjø	Upper-Metuje	Upper-Sázava	Nedožery	Upper-Guadiana
area [km^2]	119	73.6	131	181	16,479
altitude [m a.m.s.l.]*	945 (737–1595)	591 (459–780)	628 (487–805)	573 (288–1172)	769 (599–1100)
climate type [-]	Dfc	Cfb	Cfb	Dfb	Csa, Csb and Bsk
obs. period	1958–2007	1982–2005	1963–1999	1974–2006	1960–2001
T [°C]	0.7	5.9	6.8	7.6	14.1
[°C]**	Jan: −10.1; Jul: 11.9	Jan: −3.9; Jul: 15.5	Jan: −3.2; Jul: 16.3	Jan: −2.8; Jul: 17.5	Jan: 5.1; Jul: 25.0
P [mm yr^{-1}]	594	746	717	873	450
[mm $month^{-1}$]**	Mar: 27; Jul: 81	Apr: 42; Jul: 92	Feb: 36; Jun: 92	Feb: 52; Jun: 96	Jul: 9; Dec: 54
PET [mm yr^{-1}]	296	574	684	981	1250
Q [mm yr^{-1}]	820	321	291	352	16
[mm d^{-1}]**	Mar: 0.29; May: 8.0	Oct: 0.66; Mar: 1.9	Aug: 0.48; Mar: 1.7	Aug: 0.42; Mar: 2.1	Sep: 0.009; Feb: 0.11

* = mean (min – max)
** = min monthly; max monthly

Table 2. Nash-Sutcliffe values per catchment.

	Reff	ln Reff
Narsjø	0.77	0.90
Upper-Metuje	0.51	0.69
Upper-Sázava	0.59	0.63
Nedožery	0.64	0.68
Upper-Guadiana	0.54	0.71

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Table 3. General drought characteristics using a 80 % monthly threshold (moving average 30 days), the inter-event time method for pooling, and a minimum drought duration of 15 days for the hydrometeorological variables simulated with HBV and observed discharge for all selected catchments.

		no. of droughts [per year]	mean duration [day]	mean deficit [mm]	mean max.deviation [mm]
Narsjø	catchment precipitation	1.8	34	13.6	–
	soil moisture	1.1	59	–	7.4
	groundwater storage	0.9	68	–	7.3
	simulated discharge	1.2	56	11.7	–
	observed discharge	*1.2*	*54*	*17.5*	–
Upper-Metuje	catchment precipitation	1.7	33	14.2	–
	soil moisture	1.2	45	–	15.2
	groundwater storage	0.6	112	–	11.3
	simulated discharge	1.0	60	3.2	–
	observed discharge	*1.2*	*53*	*4.5*	–
Upper-Sázava	catchment precipitation	2.0	30	12.5	–
	soil moisture	1.3	47	–	18.3
	groundwater storage	0.5	139	–	8.1
	simulated discharge	1.1	62	3.6	–
	observed discharge	*1.1*	*58*	*5.6*	–
Nedožery	catchment precipitation	1.6	34	16.5	–
	soil moisture	1.4	43	–	22.4
	groundwater storage	1.1	59	–	5.3
	simulated discharge	1.3	50	4.6	–
	observed discharge	*1.4*	*45*	*4.5*	–
Upper-Guadiana	catchment precipitation	2.0	40	10.9	–
	soil moisture	1.2	77	–	21.9
	groundwater storage	0.2	756	–	5.9
	simulated discharge	1.0	154	2.2	–
	observed discharge	*0.7*	*253*	*5.5*	–

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Table 4. Drought types of all drought events per catchment (groundwater and discharge).

		classical rainfall deficit drought	rain-to-snow-season drought	wet-to-dry-season drought	cold snow season drought	warm snow season drought	composite drought
Narsjø	groundwater	28 %	13 %	–	54 %	–	–
	discharge	32 %	10 %	–	47 %	5 %	–
Upper-Metuje	groundwater	50 %	19 %	–	13 %	–	19 %
	discharge	52 %	7 %	–	15 %	19 %	7 %
Upper-Sázava	groundwater	58 %	11 %	–	11 %	11 %	11 %
	discharge	36 %	2 %	–	21 %	24 %	14 %
Nedožery	groundwater	57 %	8 %	–	14 %	22 %	–
	discharge	53 %	9 %	–	14 %	23 %	–
Upper-Guadiana	groundwater	–	–	33 %	–	–	67 %
	discharge	50 %	–	35 %	3 %	–	5 %

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Table 5. Drought types of 5 most severe drought events per catchment (groundwater and discharge).

		classical rainfall deficit drought	rain-to-snow-season drought	wet-to-dry-season drought	cold snow season drought	warm snow season drought	composite drought
Narsjø	groundwater	20 %	80 %	–	–	–	–
	discharge	20 %	80 %	–	–	–	–
Upper-Metuje	groundwater	20 %	40 %	–	–	–	40 %
	discharge	60 %	20 %	–	–	–	20 %
Upper-Sázava	groundwater	20 %	40 %	–	–	–	40 %
	discharge	20 %	20 %	–	–	40 %	20 %
Nedožery	groundwater	–	20 %	–	40 %	40 %	–
	discharge	40 %	20 %	–	–	40 %	–
Upper-Guadiana	groundwater	–	–	–	–	–	100 %
	discharge	20 %	–	40 %	–	–	20 %

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Table 6. Drought propagation processes per hydrological drought type and occurrence in Köppen-Geiger major climate types.

hydrological drought type	governing process(es)	P-control/T-control	climate type
Classical rainfall deficit drought	rainfall deficit (in any season)	P-control	A,B,C,D,E
Rain-to-snow-season drought	rainfall deficit in rain season, drought continues into snow season	P and T-control	C,D,E
Wet-to-dry-season drought	rainfall deficit in wet season, drought continues into dry season	P and T-control	A,B,C
Cold snow season drought	low temperature in snow season, leading to:		
subtype A	early beginning of snow season	T-control	D,E
subtype B	delayed snow melt	T-control	C,D
subtype C	no recharge	T-control	C,D
Warm snow season drought	high temperature in snow season, leading to:		
subtype A	early snow melt	T-control	D,E
subtype B	in combination with rainfall deficit, no recharge	P and T-control	C,D
Composite drought	combination of a number of drought events over various seasons	P and T-control	A,B,C,D,E

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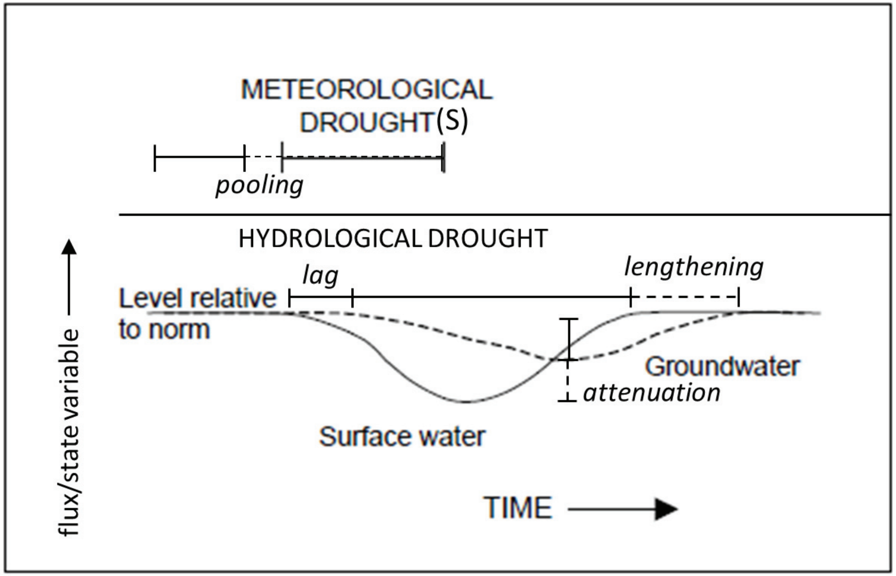

Fig. 1. Features characterising propagation of meteorological drought(s) to hydrological drought: pooling, lag, attenuation, and lengthening (modified from Hisdal and Tallaksen, 2000).

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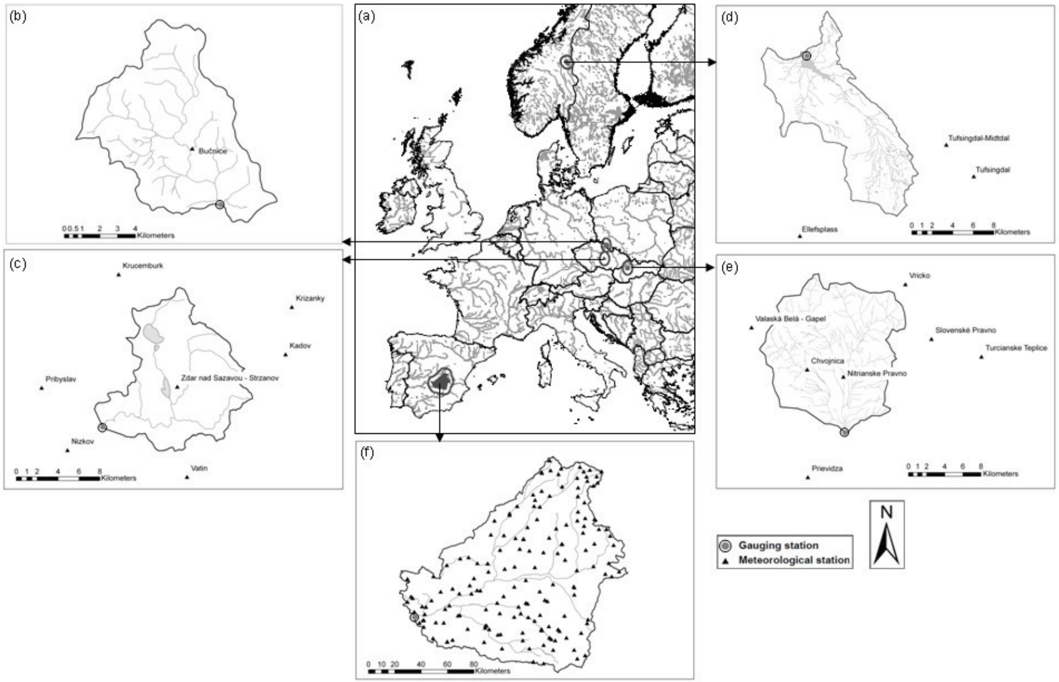

Fig. 2. **(a)** Location of the selected catchments in Europe, including gauging station and meteorological stations; **(b)** Upper-Metuje catchment; **(c)** Upper-Sázava catchment; **(d)** Narsjø catchment; **(e)** Nedožery catchment; and **(f)** Upper-Guadiana catchment.

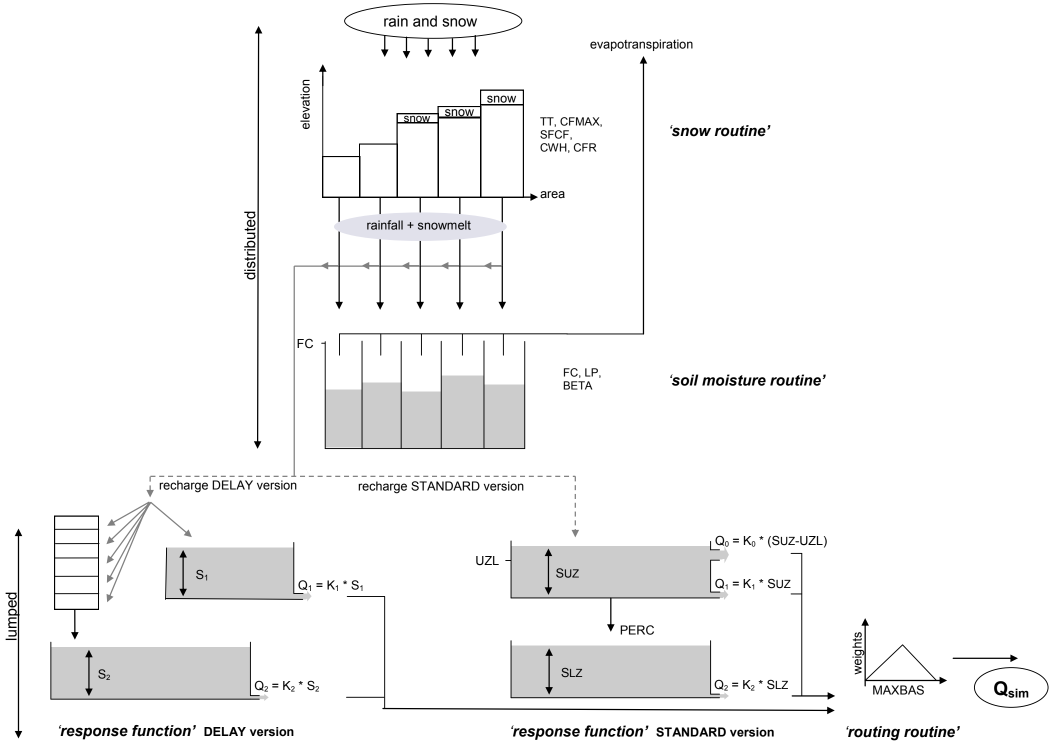

Fig. 3. Structure of the HBV model with two versions for the response routine: on the right-hand side the STANDARD-version, and on the left-hand side the DELAY-version (adapted from Seibert, 2000).

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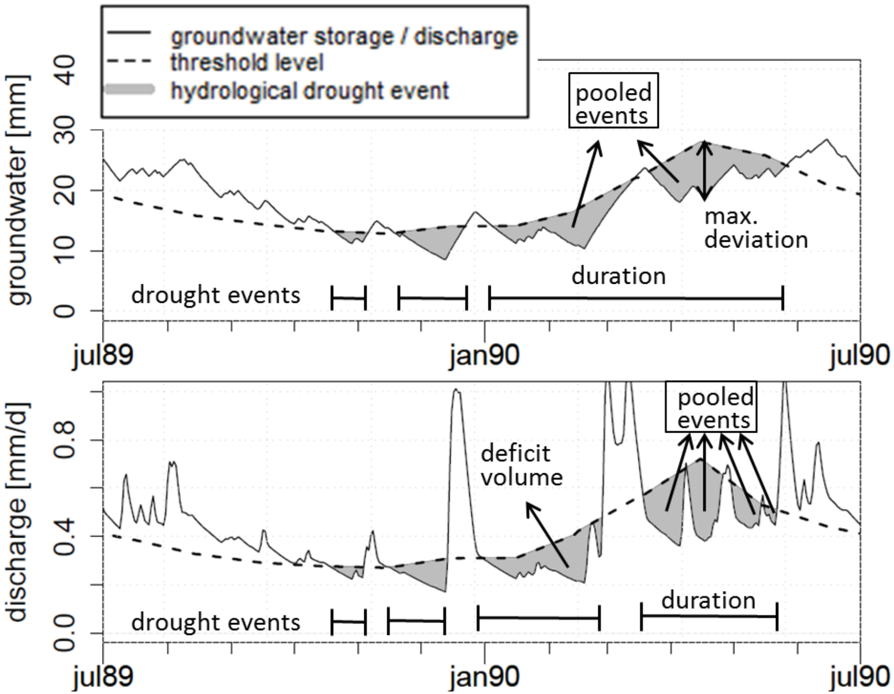

Fig. 4. Threshold level method with variable threshold (80th percentile of monthly duration curve, smoothed by 30-day moving average) for groundwater storage (upper panel) and discharge (lower panel), including an illustration of pooling method and drought characteristics duration, deficit volume, and max.deviation.

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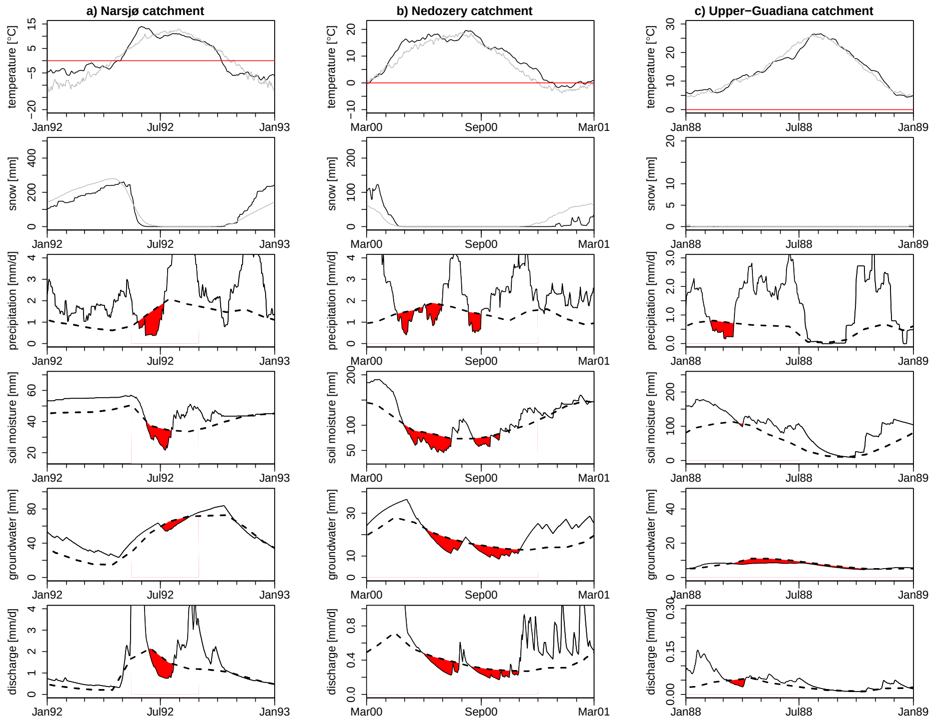

Fig. 5. Examples of *classical rainfall deficit drought* type: **(a)** Narsjø catchment 1992–1993, **(b)** Nedožery catchment 2000–2001, **(c)** Upper-Guadiana catchment 1988 (all panels: grey line = long-term average of displayed variable, dashed line = smoothed monthly 80 %-threshold of displayed variable, red area = drought event referred to in text; upper panel: black line = 30-day moving-average of observed temperature, red line = 0 degrees; second panel: black line = simulated snow accumulation; third panel: black line = 30-day moving-average of observed precipitation; fourth panel: black line = simulated soil moisture; fifth panel: black line = simulated groundwater storage; lower panel: black line = simulated discharge).

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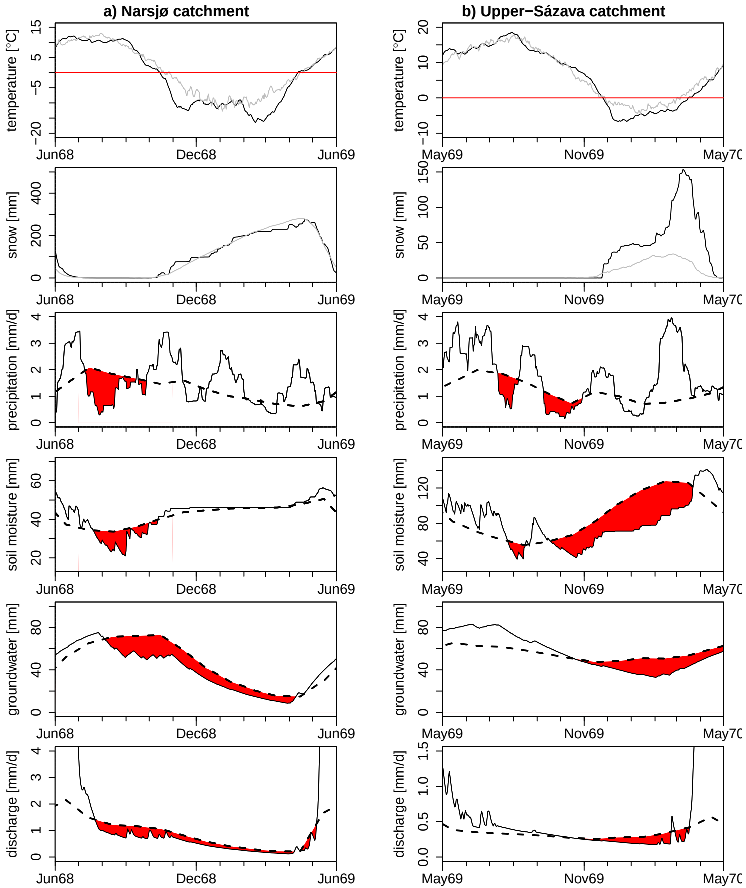

Fig. 6. Examples of *rain-to-snow-season drought* type: **(a)** Narsjø catchment 1968–1969, **(b)** Upper-Sázava catchment 1969–1970 (legend: see Fig. 5).

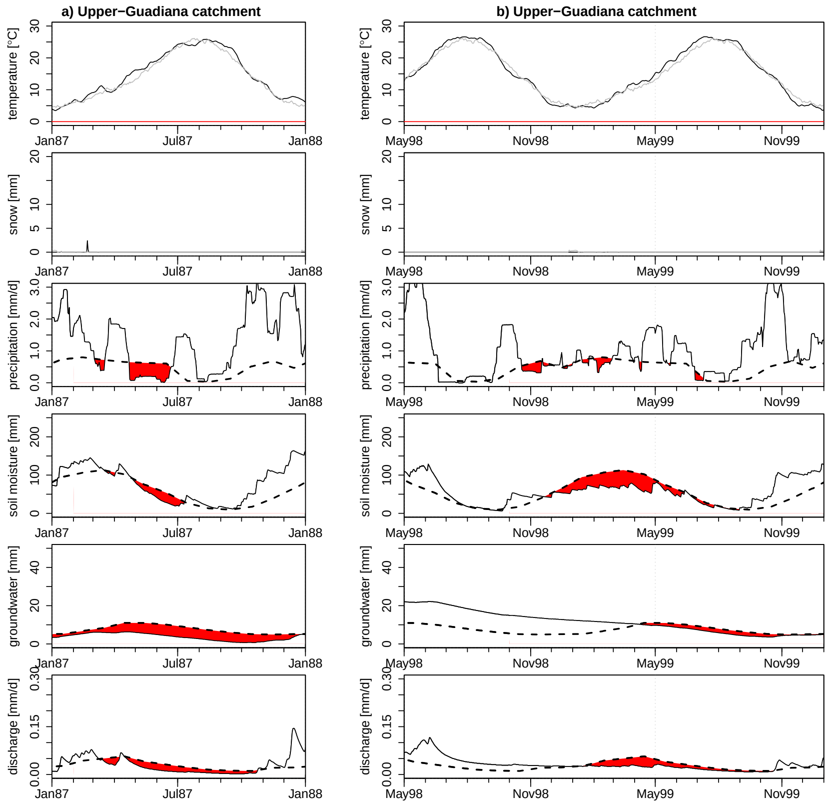

Fig. 7. Examples of *wet-to-dry-season drought* type: **(a)** Upper-Guadiana catchment 1987, **(b)** Upper-Guadiana catchment 1998–1999 (legend: see Fig. 5).

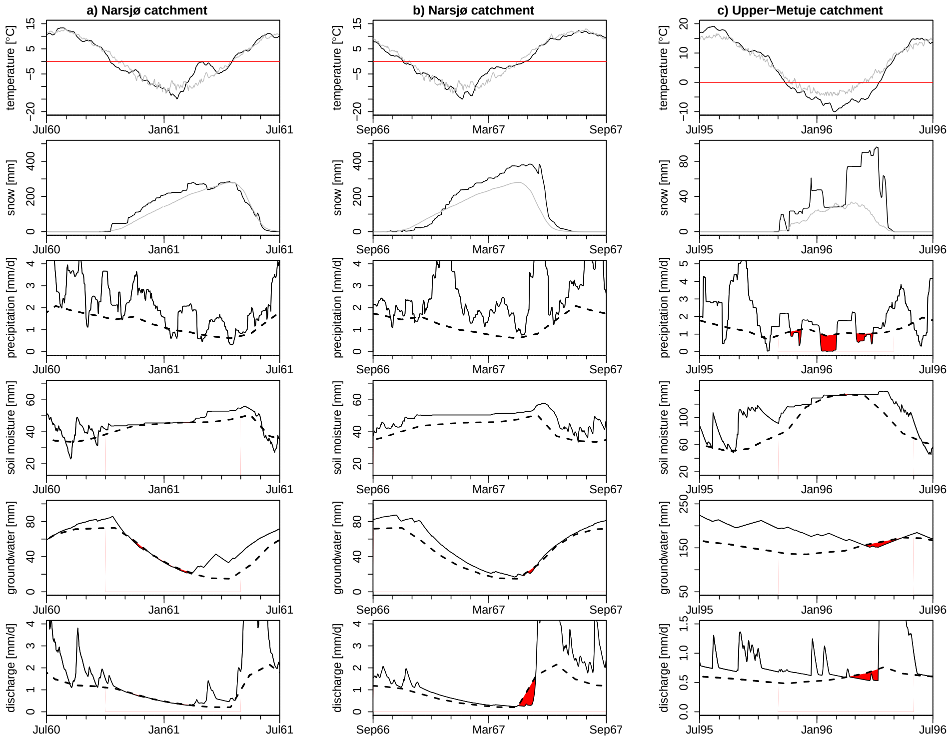

Fig. 8. Examples of *cold snow season drought* type: **(a)** Narsjø catchment 1960–1961, **(b)** Narsjø catchment 1966–1967, **(c)** Upper-Metuje catchment 1995–1996 (legend: see Fig. 5).

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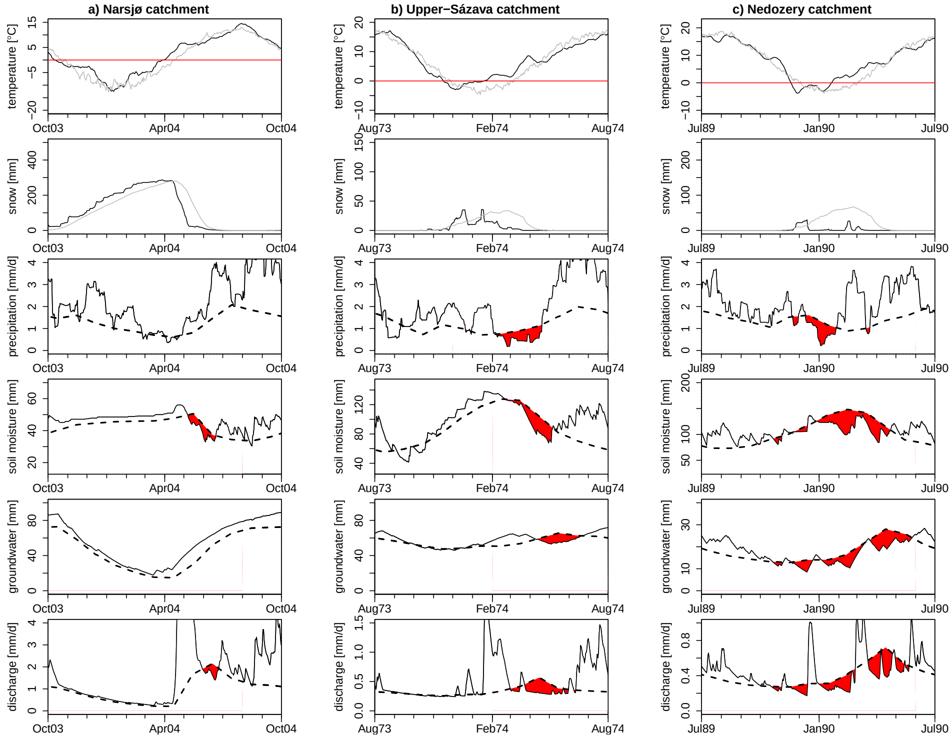

Fig. 9. Examples of *warm snow season drought* type: **(a)** Narsjø catchment 2003–2004, **(b)** Upper-Sázava catchment 1973–1974, **(c)** Nedožery catchment 1989–1990 (legend: see Fig. 5).

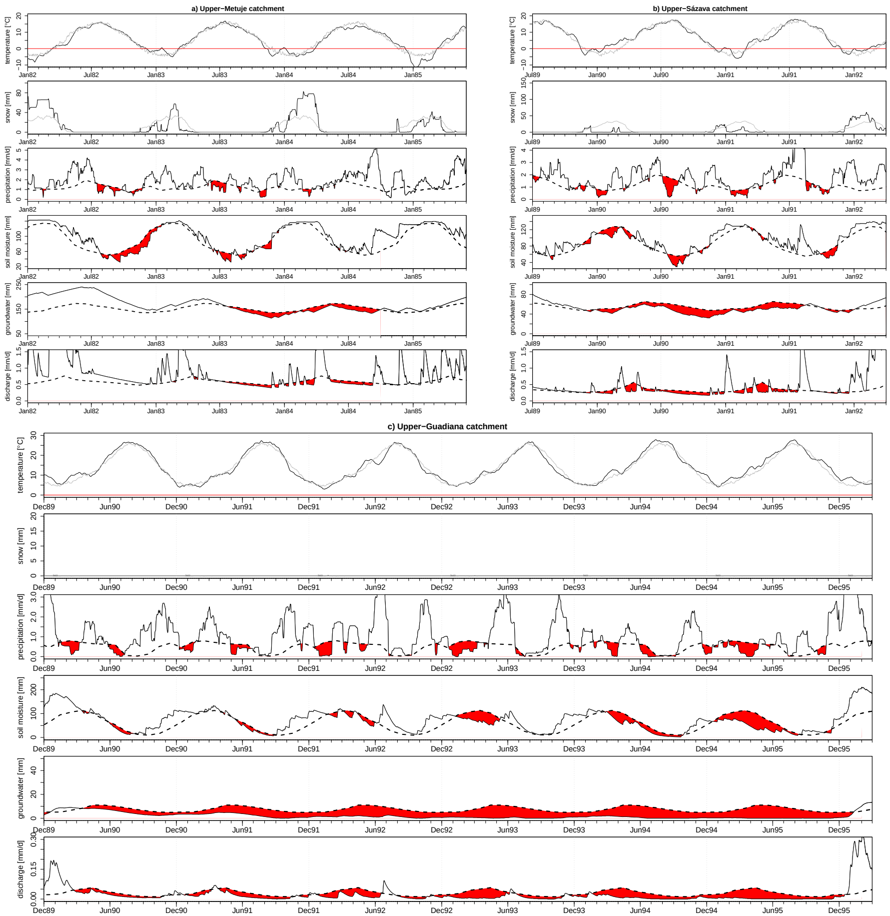

Fig. 10. Examples of *composite drought* type: **(a)** Upper-Metuje catchment 1982–1985, **(b)** Upper-Sázava catchment 1989–1992, **(c)** Upper-Guadiana catchment 1989–1995 (legend: see Fig. 5).

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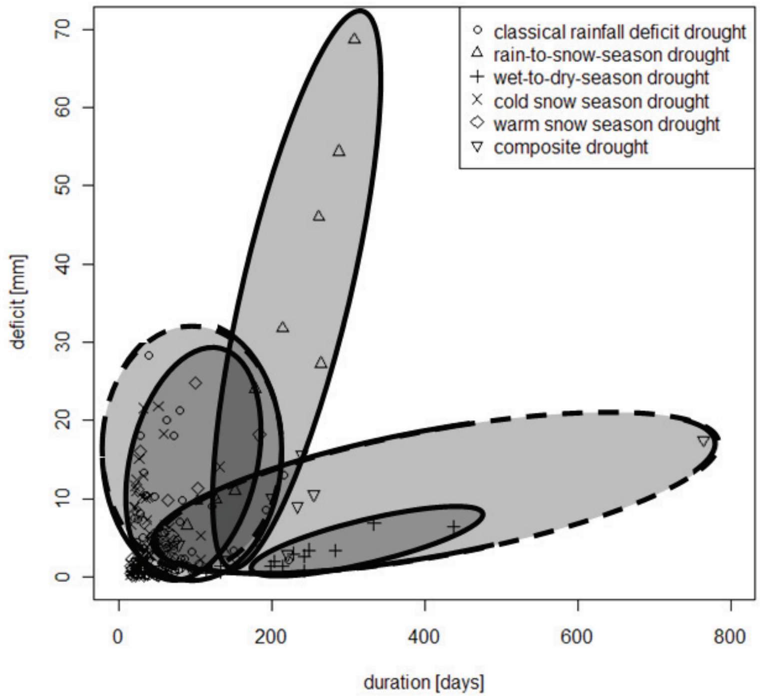

Fig. 11. Drought duration and deficit volume of all discharge drought events grouped per hydrological drought type (ellipses are added to more clearly identify groups of events with similar drought type; dashed lines indicate an approximation based on a single event).

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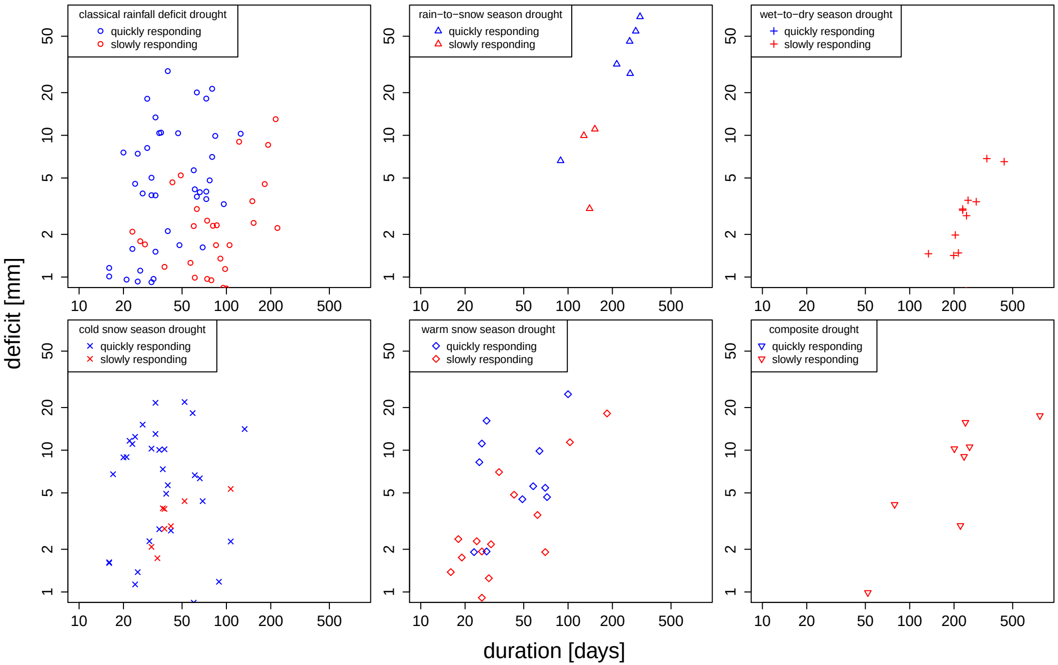

Fig. 12. Drought duration and deficit volume of all discharge drought events grouped per hydrological drought type, on log-log scale, differentiating between quickly and slowly responding catchments (quickly responding: Narsjø and Nedožery catchments; slowly responding: Upper-Metuje, Upper-Sázava, and Upper-Guadiana catchments).

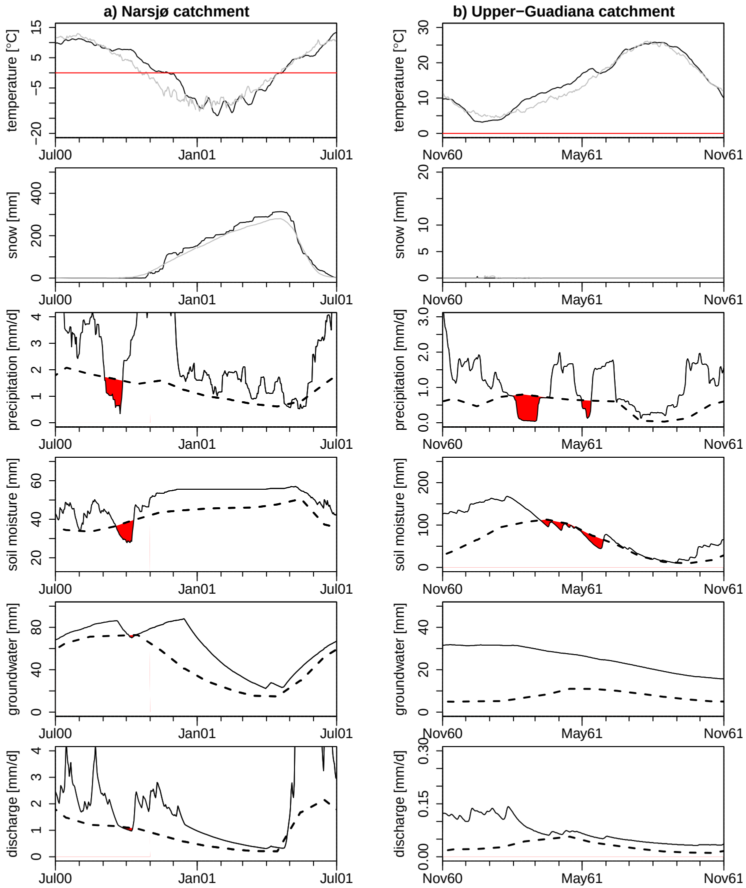

Fig. 13a,b. Examples of non-drought events: **(a)** Narsjø catchment 2000–2001, **(b)** Upper-Guadiana catchment 1960–1961 (legend: see Fig. 5).

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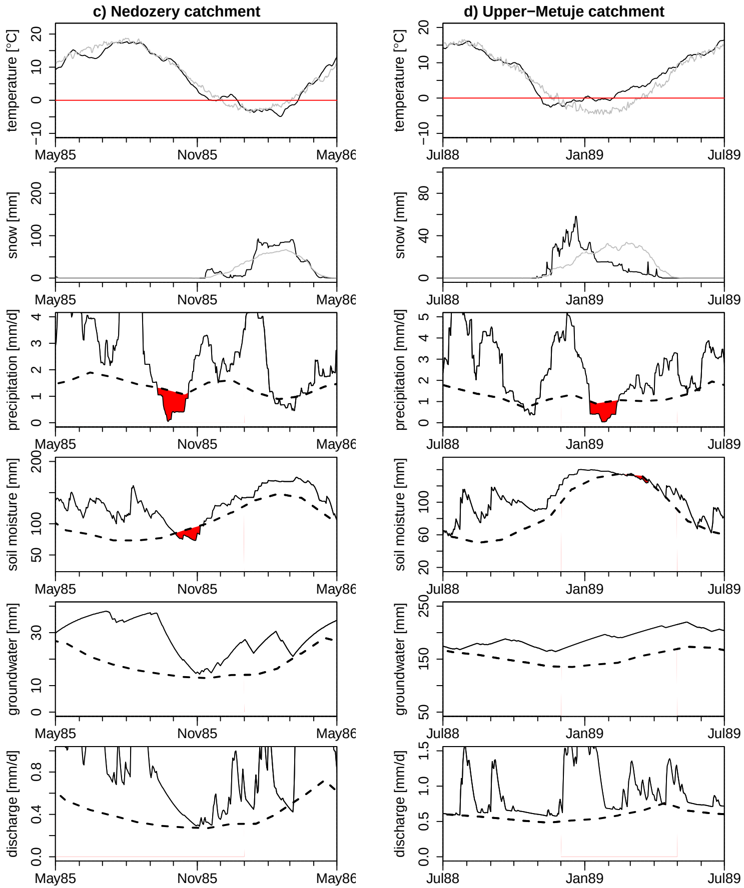

Fig. 13c,d. Examples of non-drought events: **(c)** Nedožery catchment 1985–1986, **(d)** Upper-Metuje catchment 1988–1989 (legend: see Fig. 5).

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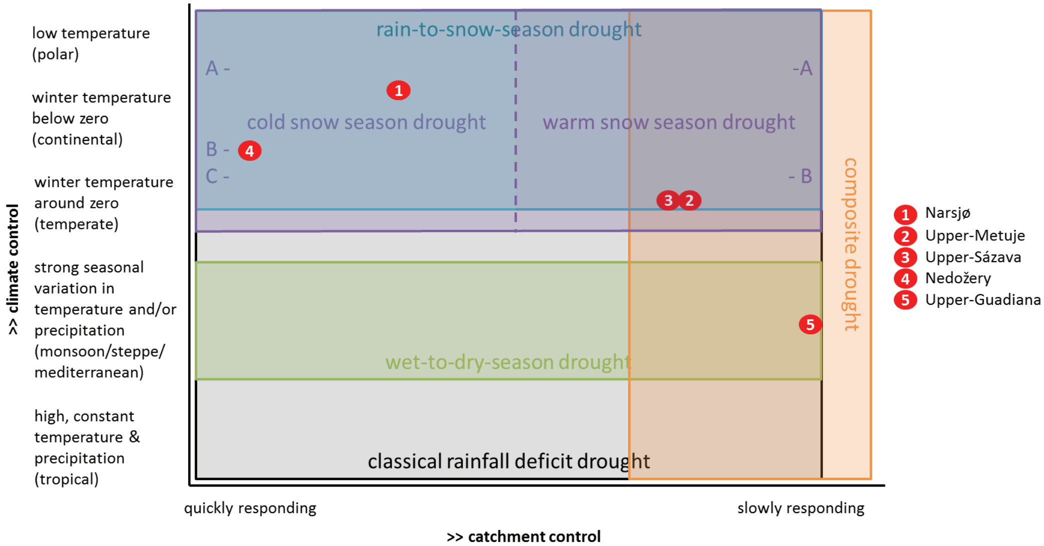

Fig. 14. Hydrological drought (sub-)type occurrence in relation to catchment and climate control; catchment control is indicated by a slower response of discharge to precipitation when moving from left to right on the *x*-axis, climate control is indicated by describing temperature and precipitation regimes relevant for drought development (desert and glacier climates are not included as is it not relevant to speak of droughts in these climates, WMO, 2008); the five study catchments are included based on their climate and catchment characteristics (see Sect. 2); for explanation of the drought (sub-)types see Table 6.

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Hydrological drought typology

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