On the Propagation of Drought

how climate and catchment characteristics influence hydrological drought development and recovery

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On the propagation of drought

How climate and catchment characteristics influence hydrological drought development and recovery

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Voor mijn grootouders.

The scientist does not study nature because it is useful; he studies it because he delights in it, and he delights in it because it is beautiful. If nature were not beautiful, it would not be worth knowing, and if nature were not worth knowing, life would not be worth living.
Jules Henri Poincaré (1854–1912)

Contents

Chapter 1

Introduction

1.1 Drought as a natural hazard

Societies around the world are exposed to natural hazards, such as earthquakes, floods, tornadoes, volcanic eruptions, droughts, hurricanes, and storms. Hydrological extremes (floods and droughts) are natural hazards that are not confined to specific regions, but occur worldwide and, therefore, impact a very large number of people [Kundzewicz and Kaczmare, 2000]. Flooding events receive most attention, both in the news and in scientific literature, due to their fast, clearly visible, and dramatic consequences. Drought events, also called 'the creeping disaster' [Wilhite, 2000; Mishra and Singh, 2010], have a much larger spatial and temporal scale than floods. Droughts can cover extensive areas and can last for months to years, with devastating impacts on many economic sectors [Tallaksen and Van Lanen, 2004; Sheffield and Wood, 2011]. Examples of affected sectors are drinking water supply, crop production (irrigation), waterborne transportation, nature (forest fires), electricity production (hydropower or cooling water), and recreation (water quality) [e.g. Wilhite, 2000; EurAqua, 2004; Tallaksen and Van Lanen, 2004; UNDP, 2006; EEA, 2007; Sheffield and Wood, 2011; Stahl et al., 2012a; Van Vliet et al., 2012]. Drought is considered one of the most damaging natural hazards in terms of economic cost [Wilhite, 2000] and, regionally, in terms of societal problems, such as hunger, mass migration, and loss of life. In the period 1900–2010, worldwide 2 billion people were affected and more than 10 million people died due to the impacts of drought [EM-DAT, 2012; EEA, 2012]. Currently, there is increasing awareness of drought and related hazards (heat waves and wildfires), resulting in more research on the topic [Mishra and Singh, 2010] and increasing efforts to inform the general public via, for example, the European Drought Centre (EDC; www.geo.uio.no/edc/), the US Drought Monitor [Svoboda et al., 2002], the European Drought Observatory (EDO; edo.jrc.ec.europa.eu), and the Global Drought Monitor (drought.mssl.ucl.ac.uk/).

In recent years, many severe drought events occurred. In 2012, a simultaneous drought in central and southern USA and Russia induced an increase in food prices. In spring 2011, western Europe faced severe water shortage and low water levels. In 2011, a long-lasting drought triggered hunger, mass migration, and loss of life in the Horn of Africa [Viste et al., 2012]. In 2010 and 2011, Russia experienced a drought and heat wave [Grumm, 2011], resulting in widespread forest fires [Huijnen et al., 2012]. In 2010, large parts of China were affected by drought, hampering food production on a large scale [Lu et al., 2011], and in that same year Scandinavia faced drinking water shortage and hydropower production problems [Cattiaux et al., 2010]. In 2005 and 2010, the Amazon rain forest was affected by a severe lack of precipitation, resulting in a massive dying of vegetation and release of CO_2 into the atmosphere [Lewis et al., 2011]. In 2008, the Iberian peninsula had to cope with the impacts of a multi-year drought that had reduced groundwater levels and reservoir storage to a minimum [Andreu et al., 2009]. A severe continent-wide multi-year drought impacted Australia between 2002 and 2010 [McGrath et al., 2012]. In 2003 and 2006, Europe was hit by a drought that caused crop failure, navigation problems, cooling water restrictions, and loss of life due to a heat wave [Rebetez et al., 2009] (in 2003 this amounted to 70,000 heat-related deaths; Robine et al. [2008]). This enumeration of recent droughts is not exhaustive, but indicates the recurring and worldwide nature of droughts.

Drought is not a recent phenomenon. Actually, most devastating drought events occurred in the previous century. Examples are the 1976 drought in Europe, the 1930s Dust Bowl in the USA [Schubert et al., 2004], and the 1920s food crisis in Russia and China (in which more than 4 million people died, EM-DAT [2012]). Also in the paleoclimatic record, many severe 'mega-droughts' are reported that had widespread ecological and socio-economic consequences and might even be related to the collapse of civilisations [Dai, 2011; Sheffield and Wood, 2011; Kennett et al., 2012; Medina-Elizalde and Rohling, 2012; Seneviratne et al., 2012; Sivapalan et al., 2012, and references therein].

The pressing questions are: have droughts become more frequent or severe in recent decades?

And: will they become more frequent or severe in the future? Several studies investigated trends in drought occurrence, both on global and on regional scales [e.g. Lins and Slack, 1999; Hisdal et al., 2001; Seneviratne et al., 2012]. On a global scale, different trend studies yield conflicting results. Dai [2011] found increasing drought, whereas Sheffield et al. [2012] did not find a trend in global drought while using the same drought index, but different data and methodology. Overall, there are still large uncertainties regarding observed global-scale trends in drought [Seneviratne et al., 2012] and the applied methodology has a large influence on the magnitude and sometimes also on the sign of observed trends [Sheffield et al., 2012]. Seneviratne et al. [2012] summarise the regional-scale studies as follows: 'there is medium confidence that since the 1950s some regions of the world have experienced trends toward more intense and longer droughts, in particular in southern Europe and West Africa, but in some regions droughts have become less frequent, less intense, or shorter, for example, central North America and northwestern Australia.' For Europe, Stahl et al. [2010] and Stahl et al. [2012b] found a coherent picture of annual streamflow trends in both observations and multi-model ensemble results, with negative trends (lower streamflow) in southern and eastern regions and generally positive trends (higher streamflow) in western and northern regions. Additionally, a decrease in summer low flows was observed in large parts of Europe, including many regions in western Europe [Stahl et al., 2012b].

There is some consistency in model studies that these European trends will continue in the future, with the predicted impacts of climate change suggesting a dryer and warmer Mediterranean region and a northward shift of climatic regimes in Europe [e.g. Milly et al., 2005; Huntington, 2006; IPCC, 2007; Bates et al., 2008; Sheffield, 2008; Beniston, 2009]. As a result there will be an enhancement of interannual variability in the European summer climate, associated with higher risks of heat waves and droughts [e.g. Schär et al., 2004; Seneviratne et al., 2006; De Wit et al., 2007; Bates et al., 2008; Feyen and Dankers, 2009; Seneviratne et al., 2012]. In other regions around the world, there is less confidence about future drought occurrence due to larger uncertainties in model projections [Bates et al., 2008; Seneviratne et al., 2012]. Bates et al. [2008], Sheffield and Wood [2011], and Seneviratne et al. [2012] give an overview of trends and possible changes in drought occurrence in the future.

Estimates of drought impacts in recent years indicate that drought-related losses are increasing. It is difficult to isolate the impacts of climate change from changes in, for example, land use and increasing vulnerability. Important factors for increased vulnerability are population growth, concentration of people in urban areas and semi-arid regions, globalisation of food markets, and water accessibility issues. Impacts of drought are likely to increase with time as society's demands on water and environmental services increase [Wada et al., 2011]. Conflicts between water users have emerged. Worldwide drought has been a stressor for international relations in transboundary rivers [Stahl, 2005, 2008] and is expected to continue to be so in the future [De Stefano et al., 2012]. Although droughts occur everywhere, it is important to note that, in general, the most severe consequences of drought for humans occur in arid or semi-arid regions where the availability of water is already low under normal conditions, the demand often is close to or even exceeds the natural availability and society often lacks the ability to adapt to the drought hazard [Dai, 2011]. Therefore, drought management is and will increasingly be crucial.

In the European Union, the Water Framework Directive demands member states to preserve or recover a 'good status' in all water bodies [Quevauviller et al., 2012] and member states are encouraged to implement drought management measures in River Basin Management Plans [EU, 2012a]. All around the world programmes exist to save water, to rely more on desalinated water, rainwater harvesting, wastewater reuse, or even controversial methods like water transfer [Martin-Ortega et al., 2011; Shrestha et al., 2011; Grant et al., 2012; EU, 2012a,b]. The main issue is moving from short-term crisis management to long-term planning including pro-active measures [Wilhite, 2000]. To achieve the latter, increased knowledge of the physical processes underlying drought is urgently needed so that forecasting, early warning, and a quick assessment of the impacts of drought are possible.

1.2 State of the art

1.2.1 Definitions

Drought is a complex phenomenon and is therefore defined in many ways. No universal definition of drought exists. Reviews of definitions can be found in Dracup et al. [1980a], Wilhite and Glantz [1985], Hisdal [2002], Tallaksen and Van Lanen [2004], Mishra and Singh [2010], and Sheffield and Wood [2011]. The most simple definition of drought is: a deficit of water compared to normal conditions [Sheffield and Wood, 2011]. In applying this definition, the following questions arise. What are normal conditions? Do we consider water in all components of the hydrological cycle or only in some? How large must a water deficit be, or how long is it to last, in order to be called a drought? Does this definition only refer to natural processes or do human influences play a role as well?

What should be regarded as the 'normal' situation strongly depends on what the water is used for. For example, certain minimal water levels in rivers are needed for navigation and ecosystems, whereas in reservoir management deviations from the seasonal inflow cycle have serious impacts. Hence, the definition of drought is dependent on the objective of a study, which is very important when quantifying drought. In drought research, we generally focus on the atmospheric and terrestrial components of the water cycle and the linkages between them, i.e. precipitation, evapotranspiration, snow accumulation, soil moisture, groundwater, lakes and wetlands, and streamflow [Sheffield and Wood, 2011]. Furthermore, it is customary to define drought as a persistent and regionally-extensive phenomenon, although these terms are not easily quantified. Droughts are generally classified into four categories [e.g. Wilhite and Glantz, 1985; Tallaksen and Van Lanen, 2004; Mishra and Singh, 2010; Sheffield and Wood, 2011], visualised in Fig. 1.1:

- Meteorological drought refers to a precipitation deficiency, possibly combined with increased potential evapotranspiration, extending over a large area and spanning an extensive period of time.

- Soil moisture drought is a deficit of (mostly root zone) soil moisture, reducing the supply of moisture to vegetation. Soil moisture drought is also called agricultural drought, because it is strongly linked to crop failure. As soil moisture deficits have additional impacts on, for example, natural ecosystems and infrastructure [Corti et al., 2009; Van der Molen et al., 2011; Seneviratne et al., 2012], I do not use the term agricultural drought for soil moisture drought in this thesis.

- Hydrological drought is a broad term related to negative anomalies in surface and subsurface water. Examples are below-normal groundwater levels or water levels in lakes, declining wetland area, and decreased river discharge. Groundwater drought and streamflow drought are sometimes defined separately as below-normal groundwater levels [Peters, 2003; Hisdal et al., 2004; Peters et al., 2006; Mishra and Singh, 2010] and below-normal river discharge [Stahl and Demuth, 1999; Smakhtin, 2001; Fleig et al., 2006; Feyen and Dankers, 2009], respectively.

- Socio-economic drought is associated with the impacts of the three above-mentioned types. It can refer to a failure of water resources systems to meet water demands and to ecological or health-related impacts of drought.

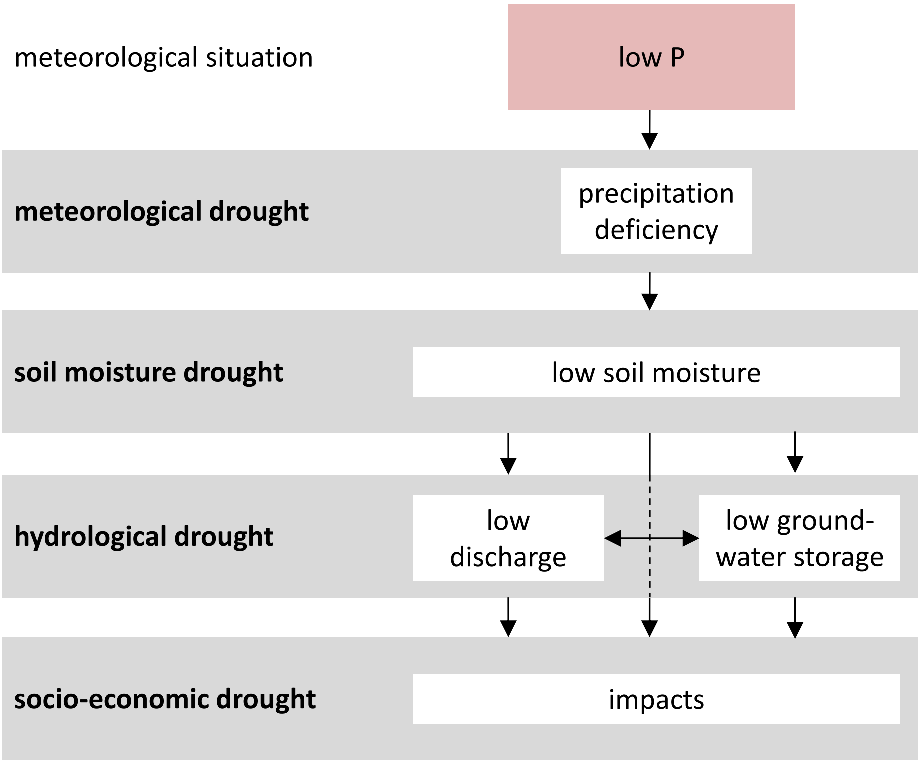

Figure 1.1: Scheme representing different categories of drought and their development (derived from Stahl [2001]; Peters [2003]). P = precipitation.

Drought should not be confused with low flow, aridity, water scarcity, or desertification, or with related hazards such as heat waves and forest fires. 'Low flow' is a frequently-used term, denoting low river discharge [Smakhtin, 2001; WMO, 2008; Laaha et al., 2013]. In this research, low flows are only considered in the section on modelling (Sect. 2.3). Low flows are often characterised by annual minimum series, which do not in all years reflect a streamflow drought. Hence, Hisdal et al. [2004] propose to distinguish between low flow characteristics and streamflow drought characteristics. 'Aridity' is the general characteristic of an arid climate and represents a (relatively) permanent condition, while drought is temporary [Mishra and Singh, 2010]. In an arid climate, drought can still occur when local conditions are even drier than normal [Stahl and Hisdal, 2004; Sheffield and Wood, 2011]. The term 'water scarcity' is used for a situation in which anthropogenic influence on the water system plays an important role in the development of below-normal water availability. Water scarcity is caused fully or in part by human activities [Seneviratne et al., 2012] and reflects conditions with long-term imbalances between available water resources and demands [e.g. Tallaksen and Van Lanen, 2004; EU, 2006a]. In this thesis, water scarcity is only studied in Ch. 3. The term 'desertification' is related to misuse or mismanagement of a region with a dry climate, leading to a reduction in vegetation cover [Kassas, 1987; Kefi et al., 2007]. Dry periods can intensify desertification. 'Heat waves' develop as a result of high temperatures. Soil moisture drought can aggravate heat waves, due to feedbacks of the land surface with the atmosphere [Seneviratne et al., 2006; Fischer et al., 2007; Vautard et al., 2007; Jaeger and Seneviratne, 2011]. The typical time scale of heat waves is in the order of weeks, whereas drought generally has durations of months to years [Mishra and Singh, 2010]. 'Forest fires' are uncontrolled fires in a wooded area. The risk of forest fire appears to increase with drought [Zumbrunnen et al., 2009], although in some regions human activities were found to be the most important driving force for forest fires [Pausas and Fernández-Muñoz, 2012].

In this research only physical processes related to drought are investigated, no socio-economical analyses were performed. Anthropogenic effects are, however, sometimes hard to neglect because they affect observed hydrometeorological variables. Anthropogenic effects on the water cycle related to drought can be direct and indirect. Direct effects are decreases of water availability by e.g. abstractions from surface water or groundwater, water diversions, and construction of reservoirs. Indirect effects are related to changes in the hydrometeorological system, leading to a decrease in water availability. For example, changes in land use can result in a faster runoff to the stream and, therefore, to lower groundwater levels. Global warming can lead to increased evapotranspiration or changes in the precipitation pattern, resulting in lower streamflow. In this research, I define drought as a below-normal water availability by natural causes only. This means that, when one works with measured data in disturbed systems, human causes must be separated from natural causes.

In this research, I use the following definition of drought, proposed by Tallaksen and Van Lanen [2004]:

> Drought is a sustained period of below-normal water availability. It is a recurring and worldwide phenomenon, with spatial and temporal characteristics that vary significantly from one region to another.

1.2.2 Drought propagation

Reasons for the occurrence of hydrological drought are complex, because they are dependent not only on the atmosphere, but also on the hydrological processes that feed moisture to the atmosphere and cause storage of water and runoff to streams [Mishra and Singh, 2010].

The atmospheric processes that are the starting point of hydrological drought development are a result of climatic variability [Stahl and Hisdal, 2004; Sheffield and Wood, 2011]. Generally, a prolonged precipitation deficiency generates less input to the hydrological system (Fig. 1.2). Depletion of soil moisture storage is related to its antecedent condition, drainage to the groundwater, and evapotranspiration from bare soil and, especially, from plants. During a dry spell, potential evapotranspiration can increase due to increased radiation, wind speed, or vapour pressure deficit (e.g. caused by a decreased moisture availability or an increased temperature). This can lead to increased actual evapotranspiration, resulting in an extra loss of water from the soil and open water bodies. In extreme drought, a lack of available soil moisture and wilting of plants can limit evapotranspiration, thus limiting a further soil moisture depletion, but possibly also limiting locally-generated precipitation, contributing to the maintenance of drought conditions. Vegetation is an important factor in modifying these feedbacks. Examples with evidence for strong feedbacks are given in D'Odorico and Porporato [2004], Teuling et al. [2005], Bierkens and Van den Hurk [2007], Dekker et al. [2007], Ivanov et al. [2008], and Seneviratne et al. [2010]. The depletion of soil moisture storage causes a decreased recharge to the groundwater system, resulting in declining groundwater levels. When pre-event groundwater levels are high, such a decrease has little effect, but when pre-event groundwater levels are low, a hydrological drought can develop. As discharge is strongly linked to storage, low groundwater levels lead to decreased groundwater discharge, which prevents aquifers from further drying, but also causes decreased streamflow [e.g. Van Lanen et al., 2004a]. These processes are summarised with the term 'drought propagation', which denotes the change of the drought signal as it moves through the terrestrial part of the hydrological cycle.

Note that in the climate community the term 'drought propagation' is sometimes used for the spatial migration of a drought event, due to atmospheric transport of anomalously warm and dry air [Joseph et al., 2009]. For example, in eastern China and western USA, a southward migration of meteorological drought was found [Hu and Feng, 2001] and in Europe, droughts starting in southern Europe were found to spread northwards [Vautard et al., 2007; Zampieri et al., 2009]. In this thesis, I use the term 'drought propagation' strictly for the translation from anomalous meteorological conditions to hydrological drought.

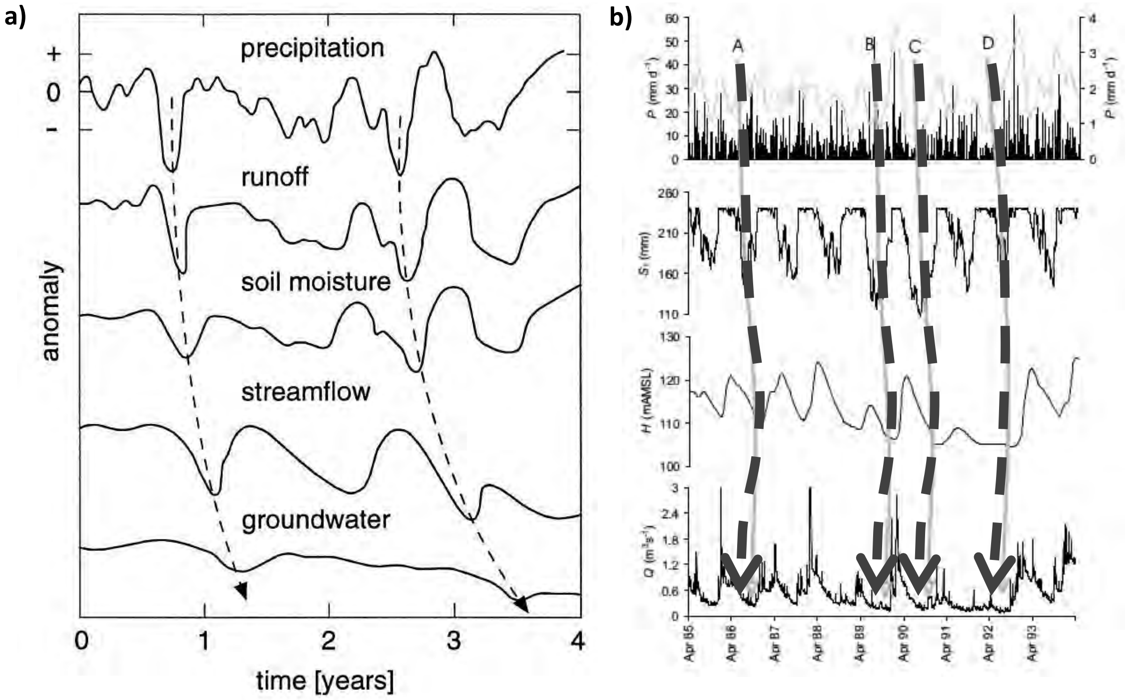

Figure 1.2: Propagation of a precipitation anomaly through the terrestrial part of the hydrological cycle for various variables, a) synthetic time series: 0 = mean, + = positive anomaly, – = negative anomaly [Changnon Jr, 1987], b) time series of the Pang catchment (UK): P = precipitation, Sr = soil moisture storage in the root zone, H = groundwater level, and Q = streamflow [Peters, 2003]. Propagation of drought events is indicated by the arrows. Note that the order of the variables is different in a) and b).

Fig. 1.2 shows the propagation of drought by means of a) synthetic time series of anomalies in different hydrometeorological variables by Changnon Jr [1987], and b) a real-world example from the Pang catchment (UK) by Peters [2003]. The general differences between the variables (in both Fig. 1.2a and b) are: many anomalies in precipitation, fewer and smaller anomalies in soil moisture, and fewer and longer anomalies in groundwater. Streamflow occupies an intermediate position in this sequence, because it is a composite of fast (direct runoff and interflow) and slow (baseflow) flow routes within a catchment. The relative position of streamflow in relation to soil moisture and groundwater is different for different areas, i.e. if a river is mainly discharging groundwater (like the Pang catchment) the streamflow drought signal is comparable to the groundwater drought signal. In Fig. 1.2a, it should also be noted that the hydrological drought of year 1 is followed by a long period with sufficient recharge to let the system recover to its original state, whereas the drought in year 3 is not compensated by sufficient recharge to assure a complete recovery of the system. The positive precipitation anomaly after the drought in year 3 is almost completely used to recover soil moisture levels and little remains for recovering streamflow and groundwater levels. If the system does not recover before the next meteorological drought develops it turns into a multi-year drought, as is apparent in the groundwater signal. This is also visible in the time series of the Pang catchment (drought C and D in Fig. 1.2b).

Propagation of drought is characterised by a number of features [Eltahir and Yeh, 1999; Peters et al., 2003; Van Lanen et al., 2004a], which are related to the fact that the terrestrial part of the hydrological cycle acts as a low-pass filter to the meteorological forcing [Kim, 1995; Marković and Koch, 2005; Rodell et al., 2010]. In Chs. 5 and 6, these features are described in relation to drought propagation in different catchments. Here, they are shortly summarised and visualised in Fig. 1.3.

- Pooling: meteorological droughts are combined into a prolonged hydrological drought.

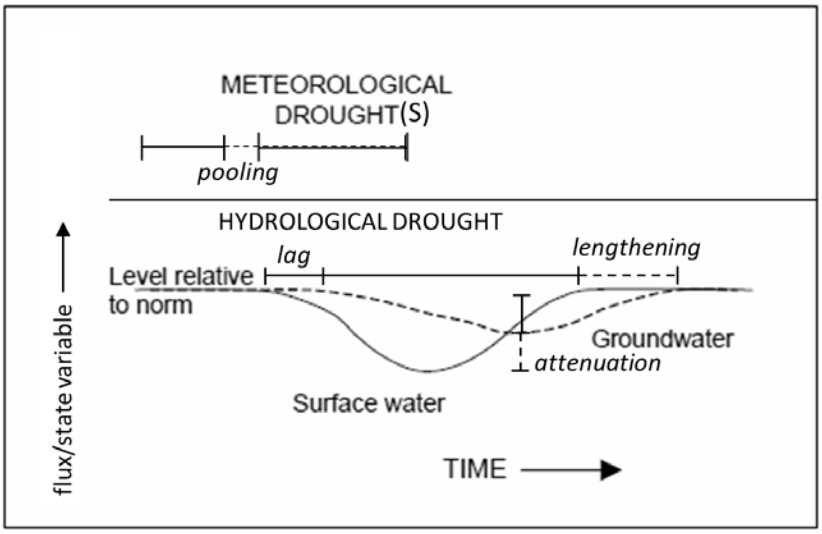

Figure 1.3: Features characterising the propagation of meteorological drought(s) to hydrological drought: pooling, lag, attenuation, and lengthening (modified from Hisdal and Tallaksen, 2000).

- Attenuation: meteorological droughts are attenuated in the stores, causing a smoothing of the maximum negative anomaly.
- Lag: a lag occurs between meteorological, soil moisture, and hydrological drought, i.e. the timing of the onset is later when moving through the hydrological cycle.
- Lengthening: droughts last longer when moving from meteorological drought via soil moisture drought to hydrological drought.

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

1.2.3 Climate control

Drought propagation is dependent on climate [Sheffield and Wood, 2011]. Various authors examined the dependency of drought characteristics on climate. In Stahl and Hisdal [2004] a broad overview is given of hydroclimatological regimes and potential for drought development in different climates around the world.

In general, hydrological droughts develop differently in relatively constant climates as compared to climates with strong seasonality. In a constant climate, the main factor for drought development is a below-normal precipitation (possibly combined with higher than normal potential evapotranspiration), as described in the previous section. In a seasonal climate, additional processes lead to the development of summer or winter droughts. In warm seasonal climates, most recharge occurs in a distinct wet season. A drought in this wet season decreases storage and can influence dry-season conditions. During the dry season, potential evapotranspiration is generally higher than precipitation, which potentially gives evapotranspiration a larger role in drought development. The role of evapotranspiration, however, is still highly uncertain. For example, Kriaučiuniene et al. [2007] found that in Lithuanian rivers (based on data starting in 1810) precipitation was more important than temperature (reflecting evapotranspiration) for the timing of dry periods in summer. Teuling et al. [2013], however, argue in favour of a large contribution of anomalies in evapotranspiring to anomalies in storage, based on observational evidence from central and western European catchments.

In seasonal climates with below-zero temperatures and snow accumulation in winter, snow-related processes play a role in drought development. Snow accumulation and frozen soils cause storage of water and prevent recharge to the groundwater, resulting in decreasing groundwater levels and streamflow throughout the winter. Early or late snow melt influences hydrological processes, namely the timing of recharge and discharge to streams [Sheffield and Wood, 2011; Huntington and Niswonger, 2012]. Frozen soils have a dual effect on drought development. On the one hand they immobilise water in the winter season, but on the other hand they can cause a fast direct runoff when snow melt and rainfall during the (early) melting period cannot infiltrate into the soil. This then leads to less recharge to the groundwater system, which can eventually enhance a summer drought in groundwater. However, many studies indicate that the effect of soil frost enhancing surface runoff during snow melt is limited, at least in forested catchments [Nyberg et al., 2001; Stähli et al., 2001; Lindström et al., 2002].

In monsoon climates, dry and wet seasons alternate, due to large-scale atmospheric processes. As this is the normal situation in these climates, I do not regard such a dry season a 'drought' (see Sect. 1.2.1). A drought occurs when the onset of the monsoon is delayed or a complete or partial failure of the monsoon takes place [Flatau et al., 2003; Schewe and Levermann, 2012]. This results in a lack of soil moisture replenishment and recharge after the dry season, causing storage to decrease to below-normal levels. In arid climates, dry periods are irregular and can last long due to erratic precipitation. Streamflow in these climates is highly dependent on groundwater discharge, showing a long recession during periods without rain [Stahl and Hisdal, 2004]. These differences in processes underlying drought development in different climates pose challenges to drought characterisation, which are discussed in Ch. 2.

1.2.4 Catchment control

The propagation of a drought in a fast responding catchment differs from that in a slow responding catchment, i.e. pooling, lag, attenuation, and lengthening of the drought signal are influenced by the catchment characteristics. Not only the hydrological variables discharge and groundwater levels themselves are related to catchment characteristics [e.g. De Wit, 2001; Uijlenhoet et al., 2001; Bidwell, 2005; Detenbeck et al., 2005], but also the dry anomalies of these variables, i.e. low flow and drought, as has been shown in many studies. For instance, Keyantash and Dracup [2004] related drought severity to surface-water storage, Engeland et al. [2006] determined regression equations between low-flow indices and catchment characteristics, Tokarczyk and Jakubowski [2006] concluded that different types of rock result in a different development of low flow. Eng and Milly [2007] evaluated from previous studies which catchment parameters show a significant relation with low-flow characteristics and found that catchment area and soil type are important. Van Lanen et al. [2004a] provide a comprehensive overview of the mechanisms by which hydrological processes and catchment characteristics influence hydrological drought. Smakhtin [2001], Demuth and Young [2004], and Laaha et al. [2013] do the same for low flows, showing the relationship between low-flow indices and catchment characteristics.

When the response time of a catchment is very long, lag times between meteorological and hydrological drought are very long as well, which can make a hydrological drought to occur in a different season than the meteorological drought that is causing it. A lack of recharge in winter can then be an important factor in causing a hydrological drought in summer in some slow responding catchments. For example, Stahl et al. [2002] concluded that, for their summer discharge, groundwater-dominated rivers in the UK mainly depend on groundwater recharge in the winter period, and Peters et al. [2006] found that in a specific groundwater-fed catchment a sequence of dry winters resulted in a multi-year drought.

For hydrological drought development, the most important catchment characteristic is the storage capacity of a catchment. Major stores in a catchment are: snow and glaciers, peat

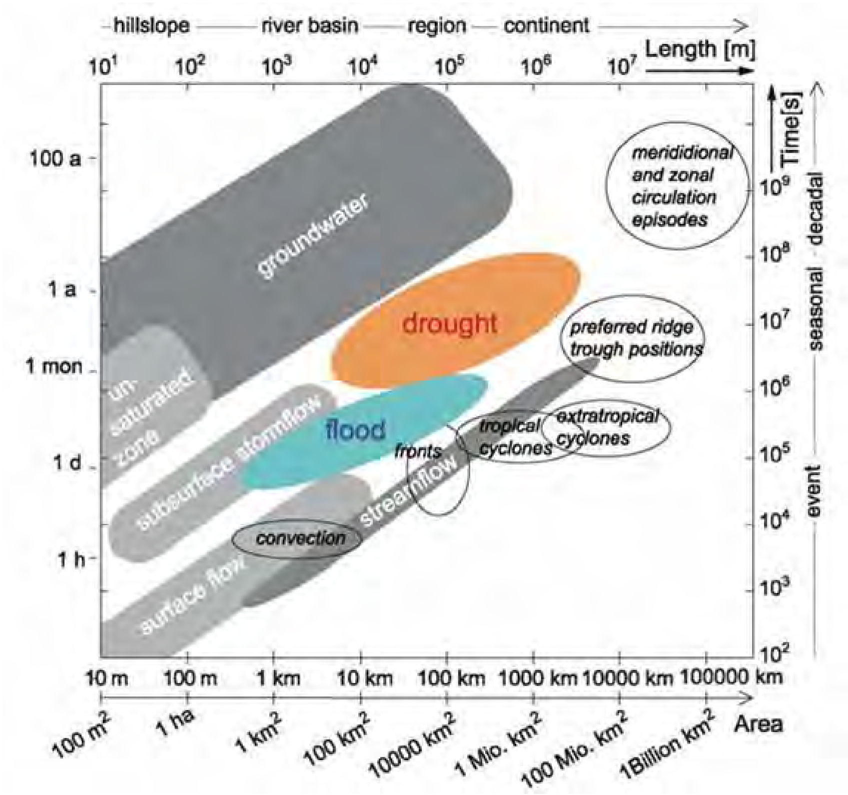

Figure 1.4: Spatial and temporal scales of hydrological processes including floods and droughts (adapted from Stahl and Hisdal [2004]).

swamps and bogs, the soil column (in particular when groundwater levels are low), the groundwater system, and lakes and reservoirs. These stores create a long memory in the hydrological system, which determines the transformation of the drought signal [Van Lanen et al., 2004a; Brutsaert, 2005]. In general, storage in a catchment is determined by factors such as the climate (in case of snow and glaciers) and the geology of the catchment (i.e. percentage of hard rock and types of rock), topography, soil (e.g. soil texture and structure), drainage network, land use, and vegetation.

Not all catchment characteristics are constant, some may change over time [e.g. Van Ogtrop and Vervoort, 2008]. Eltahir and Yeh [1999], for example, found that drainage density is dependent on groundwater level and thus on the drought state of the system. This non-linear behaviour of storage factors results in an asymmetric response of streamflow to a drought signal [Eltahir and Yeh, 1999; Van de Griend et al., 2002; Bierkens and Van den Hurk, 2007].

1.2.5 Drought scales

As was mentioned previously, droughts occur on other time and spatial scales than floods. Fig. 1.4 relates the scale of drought to typical scales of meteorological and hydrological phenomena [see also Grayson and Blöschl, 2001]. Droughts typically occur on catchment to continental scales, but there are also differences in scale between different drought types. Tallaksen et al. [2009] found that meteorological droughts are short (1–2 months) and frequently cover the whole catchment, whereas hydrological droughts have a longer duration (4–5 months) and cover a smaller area. Meteorological droughts are dependent on large-scale atmospheric drivers that usually cover a large area. In contrast, the spatial pattern of hydrological drought is more patchy, because it is more dependent on local catchment characteristics and how they change the drought signal when it propagates through the terrestrial hydrological cycle. Zaidman et al.

[2002] found the same for the 1976 drought in Europe and concluded that there was a higher level of autocorrelation in the streamflow time series than in the precipitation time series, resulting in a lower areal coverage, but higher persistence in streamflow droughts. In regions where convective thunderstorms are the dominant precipitation type and catchment conditions are relatively uniform, spatial drought patterns might be reversed, with more patchy meteorological droughts and spatially more coherent hydrological droughts [Grayson and Blöschl, 2001].

Depending on the scale, different processes are dominant. For example, in large catchments elevation differences result in a large variation in precipitation and temperature over the catchment. This leads to high spatial variability, which dampens the spatial development of hydrological drought. Also the travel time within the catchment needs to be taken into account in large catchments, as it results in a different response in upstream and downstream parts of the catchment. Peters et al. [2006] investigated the spatial distribution of drought propagation and concluded that short groundwater droughts are more severe near the stream and are attenuated at greater distances. Long periods of below-normal recharge have relatively more effect near the groundwater divide. Pandey et al. [2008] found that the upper reaches of a river in India were more prone to severe drought than the lower reaches.

Other spatial aspects of drought are synchronicity, clustering and the breaking up of drought clusters. These aspects have not been investigated in this research. I refer to Burn and DeWit [1996], Changnon [1996], Zaidman et al. [2002], Andreadis et al. [2005], Peters et al. [2006], Sheffield et al. [2009], Tallaksen et al. [2009], Santos et al. [2010], and Corzo Perez et al. [2011a].

1.3 Scientific framework

1.3.1 Background

The first research addressing changes in the drought signal due to propagation through the hydrological cycle was done in Illinois, USA, by Changnon Jr [1987] and Eltahir and Yeh [1999]. The latter were the first to use the word 'propagation' in the context of the translation from meteorological to hydrological drought. The work of Changnon Jr [1987] and Eltahir and Yeh [1999] was continued by Peters [2003] who published a study on the propagation of drought in groundwater. In recent years, drought propagation has been studied by Tallaksen and Van Lanen [2004], Peters et al. [2006], Van Lanen [2006], Tallaksen et al. [2006], Tallaksen et al. [2009], Di Domenico et al. [2010], and Vidal et al. [2010].

As mentioned in the first section of this chapter, drought receives increasing attention in research [Sheffield et al., 2012; Dai, 2013]. However, the focus is mainly on meteorological and soil moisture drought, whereas society is also severely affected by the impacts of hydrological drought, through its effects on water resources (Sect. 1.1). In studies on drought forecasting it was found that processes underlying the development of hydrological drought are still poorly understood. Mishra and Singh [2010], for example, state that 'understanding the development of hydrological drought will remain a challenge for water resources planners'. In the recent IPCC report on extremes, Seneviratne et al. [2012] write:

> The space-time development of hydrological drought as a response to a meteorological drought and the associated soil moisture drought (drought propagation, e.g. Peters [2003]) needs more attention. There is some understanding of these issues on the catchment scale [e.g. Tallaksen et al., 2009], but these need to be extended to the regional and continental scales. This would lead to better understanding of the projections of hydrological droughts, which would contribute to a better identification and attribution of droughts and help to improve global hydrological models and land surface models.

A comparable call for a better understanding of the processes related to drought, also on large scales, is given by Cloke and Hannah [2011].

In the research described in this PhD thesis I investigate the processes underlying drought propagation and their relation with climate and catchment control. I focus on the terrestrial part of the hydrological cycle. Consequently, causative mechanisms of meteorological drought development, e.g. related to blocking high-pressure systems [Fleig et al., 2010, 2011], correlation with ENSO, NAO, or other tele-connection patterns [Kingston et al., 2010], and feedbacks between the land surface and the atmosphere [Seneviratne et al., 2012], are not treated in this thesis. I first investigate drought propagation processes on the catchment scale, using catchments with contrasting climate and catchment characteristics. As Cloke and Hannah [2011] and Seneviratne et al. [2012] advocate, process knowledge of drought propagation should be extended to larger scales. Therefore, in this research, the knowledge acquired on the catchment scale is transferred to and tested on larger scales (using various large-scale methods).

1.3.2 Objective and research topics

The general objective of this PhD research is to investigate drought propagation through the terrestrial hydrological cycle, related to climate and catchment control (Fig. 1.5). More specifically, I will:

- distinguish between drought and water scarcity in order to exclude anthropogenic control;
- gain insight into hydrological processes underlying drought propagation in selected case study areas, by investigating the effects of climate control and catchment control and by developing a typology for drought propagation on the catchment scale;
- test the performance of large-scale methods (i.e. models) on reproducing drought propagation processes; and
- explore the influence of climate on drought propagation on the global scale.

There are a number of important fields where the outcomes of this research can be applied, related to prediction in time and in space: i) monthly or seasonal forecasting of hydrological drought on the basis of meteorological forecasts, ii) prediction of the effect of global change on hydrological drought, and iii) prediction of hydrological drought in ungauged basins. Improvement of the seasonal forecasting of drought is a prerequisite for adequate operational water management (e.g. reservoir operation, irrigation abstractions, or management of wetlands). For long-term water management (e.g. reservoir design or policy development like the EU Water Framework Directive [EU, 2012b]), information on larger time and spatial scales is needed. Knowledge of climate and catchment control on drought propagation processes can assist in the assessment of, for example, the effect of global change on drought patterns. In this PhD research, the effects of global change on drought propagation are not studied directly. However, the outcomes might be used to infer the effect of changes in climate control and catchment control on hydrological drought and might help to attribute trends. Lack of available data is generally a problem in water management, but especially in drought management. The outcomes of this research (which are based on gauged basins) might be extrapolated to ungauged basins, using catchment properties and regionalisation methods [Laaha et al., 2013].

1.3.3 Outline of the thesis

This thesis is organised as follows. After this introductory chapter, the reader is introduced to the general data and methods used throughout this thesis in Chapter 2. I present the case study areas that are used in the catchment-scale studies and give an overview of the observational

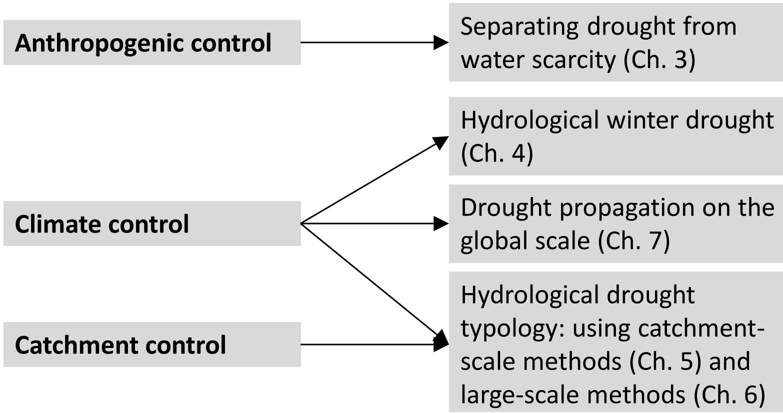

Figure 1.5: Controls on hydrological drought and the chapters in which they are studied.

data available from gauging stations in those catchments and from large-scale datasets. Next, I discuss modelling approaches and their suitability for low-flow and drought simulation and present challenges and suggestions for improvement of low-flow modelling. The last part of Chapter 2 covers an overview of drought characterisation methods, including the most widely-used drought indices. The drought characterisation method that is used in this thesis is explained in more detail.

In Chapters 3 to 7, the influence of anthropogenic control, climate control and catchment control on the development and recovery of hydrological drought are investigated, both on the catchment scale and the global scale (Fig. 1.5).

In Chapter 3, I investigate how one can distinguish between drought and water scarcity in water-stressed regions. As I do not study anthropogenic control in the remainder of this thesis, I needed to naturalise observational data and quantify natural and human influences on abnormally dry conditions.

In Chapter 4, I examine which processes cause winter droughts. Two catchments with different climates are compared. Both catchments have snow accumulation in winter, but one catchment has extremely cold winters with continuous snow cover and the other catchment has winter temperatures around zero leading to an occasional melt of the snow cover.

In Chapter 5, I present a classification of hydrological drought events into different types, which is based on processes underlying drought propagation. This hydrological drought typology is derived from and applied to case study catchments with contrasting climate and catchment characteristics, so that general rules regarding the occurrence of drought types in other catchments can be inferred.

The information on drought propagation processes, obtained in Chs. 4 and 5 using observations and traditional catchment-scale models, enables us to test how well large-scale models reproduce these drought propagation processes. In Chapter 6, drought characteristics, drought propagation features (Sect. 1.2.2) and hydrological drought typology, derived from an ensemble of large-scale models, are evaluated for the case study areas and compared with the results of the catchment-scale models obtained in Ch. 5.

In Chapter 7, a simple method is used to investigate climate control on drought propagation on the global scale. I show how seasonality in climate influences both soil moisture and hydrological drought characteristics.

A synthesis of the research presented in this thesis is given in Chapter 8. In this final chapter, the contribution of the results to the general objective is evaluated. Additionally, implications of this research for science and drought management are discussed and recommendations for future research are given.

Chapter 2

Study areas, data and methods

Prerequisites for drought research are, i) a sufficient amount of good-quality data (long measured or modelled time series), and ii) an appropriate drought identification method. As mentioned in Sect. 1.1, the specific data used and the methodology applied can have a large impact on the findings. Therefore, the data and methodology I used in this PhD research are thoroughly described in this chapter. First, I introduce the case study areas that were used in Chs. 3, 4, 5, and 6, then I summarise the data, explain the modelling approaches, and discuss the drought analysis method.

2.1 Study areas

The outcomes of this research need to be robust and representative of climate and catchment control under a variety of circumstances. Therefore, I investigated a number of catchments in Europe with contrasting characteristics in a multi-catchment analysis. This 'comparative hydrology', a term coined by Falkenmark and Chapman [1989] and reintroduced by Sivapalan [2009], suggests that 'generalisable insights' can be obtained by using data of more than one catchment.

The catchments that were investigated in this study were selected on the basis of the following criteria:

- They represent regions with contrasting climatological properties and different catchment characteristics (especially stores).
- Human influence in the catchments was limited: only the headwater area of the catchment was studied (if necessary, naturalisation of time series [Rees et al., 2004] was applied).
- Measurements of various hydrological variables were well documented (good quality, long time series, and adequate spatial distribution) and readily available (free of charge) [Van Lanen et al., 2008]. A minimum record length of 20-30 years is recommended for drought studies [Hisdal et al., 2004; Laaha et al., 2013].
- Information on the hydrological functioning of the catchment was available from local sources.

The catchments that were selected are: Narsjø in Norway, Upper-Metuje and Upper-Sázava in the Czech Republic, Nedožery in Slovakia, and Upper-Guadiana in Spain (Fig. 2.1). Characteristics of these catchments are presented in Table 2.1.

2.1.1 Narsjø

The Narsjø catchment is located in southeastern Norway (Fig. 2.1d). It is a sub-basin of the Upper-Glomma, which is the headwater catchment of the Glomma. The area of the Narsjø catchment is approximately 120 km^2 (Table 2.1). The catchment is located in a glacially formed mountainous region with rounded tops and U-shaped valleys. The altitude range is rather large, from approximately 740 to 1600 m a.m.s.l., and the average altitude is 945 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 (measured at Røros, 628 m a.m.s.l., 25 km north of the catchment) was 0.7°C, precipitation was around 590 $\mathrm{mm\,yr^{-1}}$, and potential evapotranspiration was around 300 $\mathrm{mm\,yr^{-1}}$ (Table 2.1). In winter, a continuous snow cover is present for, on average, 7 months from mid-October until the end of May, depending on the altitude [Engeland, 2002]. Measured mean discharge was around 820 $\mathrm{mm\,yr^{-1}}$, which is higher than measured precipitation due to the low elevation of precipitation gauges in the valleys around the catchment (Fig. 2.1d) in combination with an increase of precipitation with altitude.

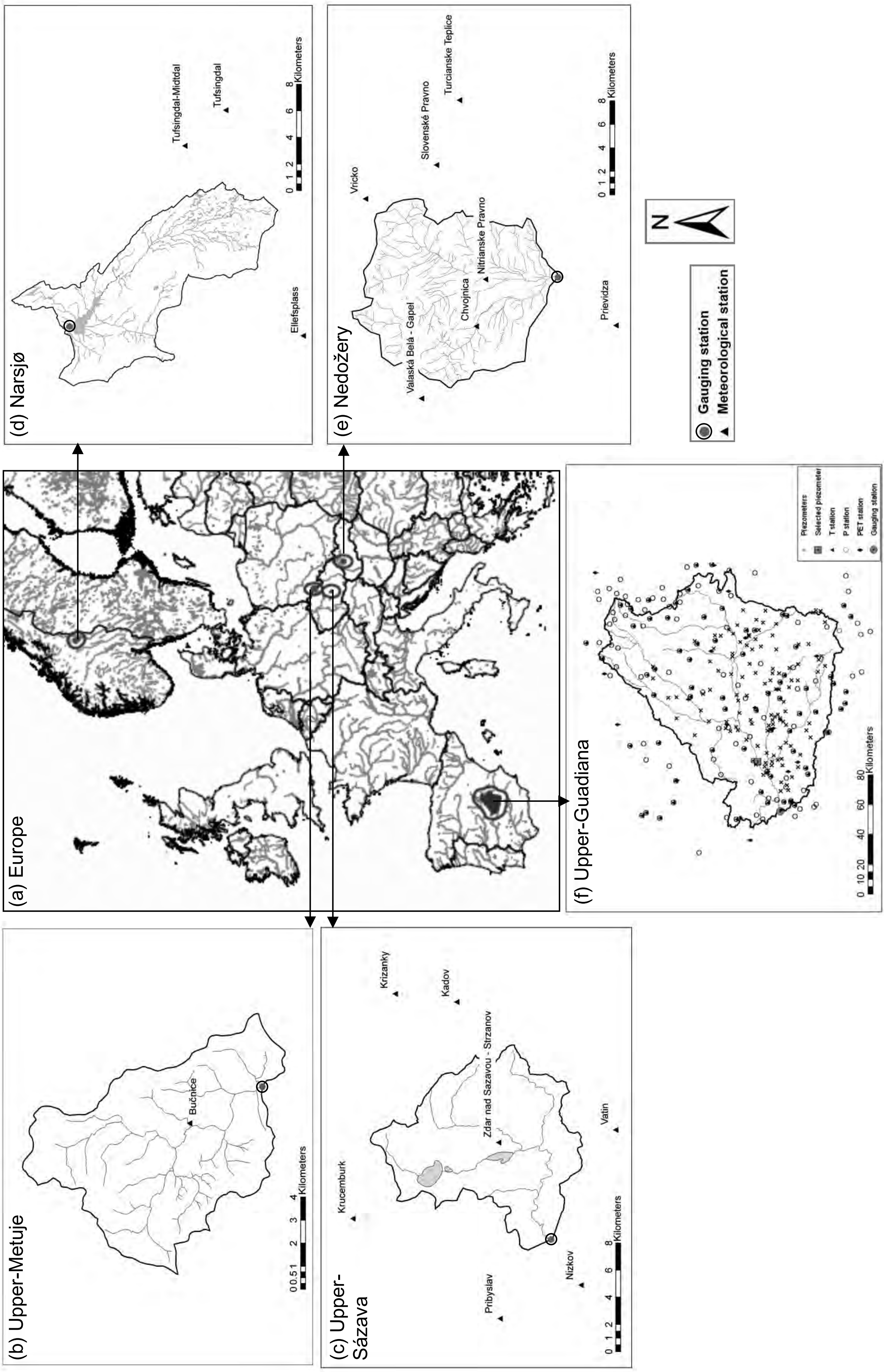

Figure 2.1: **(a)** Location of the selected catchments in Europe, including gauging stations and meteorological stations; **(b)** Upper-Metuje catchment; **(c)** Upper-Sázava catchment; **(d)** Narsjø catchment; **(e)** Nedožery catchment; and **(f)** Upper-Guadiana catchment.

Table 2.1: 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 evapotranspiration, Q = discharge

	Narsjø	Upper-Metuje	Upper-Sázava	Nedožery	Upper-Guadiana
Area [km^2]	119	74	131	181	16,479
Altitude [m a.m.s.l.][a]	945 (737–1595)	591 (459–780)	628 (487–805)	573 (288–1172)	769 (599–1100)
Topography	mountainous to lowland	high diversity of deep valleys, gentle and steep slopes	hilly landscape with flat and wide valleys, mild slopes	mountainous to lowland	hilly to flat basin
Geology	hard rock: metamorphic rocks	Cretaceous basin: porous sandstone	hard rock: metamorphic rocks	hard rock, different types	sedimentary rock: large aquifer system
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][b]	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}][b]	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}][b]	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
Stores	snow, blanket bogs, lakes (intermediate storage)	soil, groundwater (high storage)	soil, reservoirs / ponds (high storage)	soil (low storage)	groundwater, lakes, wetlands (high storage)

[a] = mean (min–max). [b] = min monthly; max monthly.

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 (Table 2.1). Narsjø is a hardrock catchment consisting predominantly of impermeable metamorphic rocks without extensive groundwater storage, which makes the catchment respond quickly to precipitation. Some delay in the response is caused by lakes, covering 3 % of the catchment, and bogs, covering 12 %. Other land cover types of the catchment are open area (61 %), forest (24 %), and only some agriculture (0.4 %) [Hohenrainer, 2008]. Human influence is very limited in the Narsjø catchment.

2.1.2 Upper-Metuje

The Upper-Metuje catchment is located in the northeastern part of the Czech Republic and partly in Poland (approximately 10 % of the catchment area) (Fig. 2.1b). It is the headwater catchment of the Metuje, which drains into the Elbe. The area of the Upper-Metuje catchment is approximately 75 km^2 (Table 2.1). The catchment is located in a hilly region of gentle slopes and wide valleys, except for some steep sandstone formations in the centre of the catchment. The altitude range is approximately 450–780 m a.m.s.l., with an average of 591 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 evapotranspiration was around 570 mm yr^{-1} (Table 2.1). 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}.

The low-flow season of Upper-Metuje is summer/autumn, and highest flows occur in March due to snow melt (Table 2.1). 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 slow responding catchment with a relatively high baseflow. Nevertheless, discharge peaks occur when storage is filled. 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.

2.1.3 Upper-Sázava

The Upper-Sázava catchment is located in central Czech Republic (Fig. 2.1c). It is the headwater catchment of the Sázava, which drains into the Vltava and, finally, into the Elbe. The area of the Upper-Sázava catchment is approximately 130 km^2 (Table 2.1). The catchment is located in a hilly region of gentle slopes and wide valleys and the altitude range is approximately 490–800 m a.m.s.l., with an average of 628 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 evapotranspiration was around 680 mm yr^{-1} (Table 2.1). 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}.

The low-flow season of Upper-Sázava is summer, and highest flows occur in March due to snow melt (Table 2.1). 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 area. Other land cover types of the catchment are forest (50 %), and cropland and grassland (40 %) [Rakovec et al., 2009]. Human influence is limited to extensive agriculture, and some groundwater extraction and sewage water release.

2.1.4 Nedožery

The Nedožery catchment is located in the Prievidza District in central Slovakia (Fig. 2.1e). It is the headwater catchment of the Nitra, which drains into the Vah and, finally, into the Danube. The area of the Nedožery catchment is approximately 180 km^2 (Table 2.1). The catchment is located in a mountainous region with steep slopes. Therefore, the altitude range is large, from approximately 290–1170 m a.m.s.l., with an average of 573 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 evapotranspiration was around 980 mm yr^{-1} (Table 2.1). In winter, a continuous snow cover is present for, on average, 4 months from December until the beginning of April, with large variations within the catchment due to elevation differences. 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 2.1). Nedožery is a hardrock catchment consisting predominantly of impermeable metamorphic rocks without extensive groundwater storage, which makes it quick in responding to precipitation. The presence of steep slopes and the absence of bogs or lakes accelerate the response. Two-thirds of the catchment is covered with forest. Other land cover types are agriculture (23 %), natural meadow (6 %), and urban area (5 %) [Oosterwijk et al., 2009]. Human influence is limited to extensive agriculture.

2.1.5 Upper-Guadiana

The Upper-Guadiana catchment is located in central Spain (Fig. 2.1f). It is the headwater catchment of the Guadiana, which flows through Spain and Portugal and discharges into the North Atlantic Ocean [Van Lanen et al., 2008]. The Upper-Guadiana catchment lies completely within the province of Castilla-La Mancha and is part of the Central Spanish Plateau. The area of the Upper-Guadiana catchment is approximately 16 480 km^2, which is considerably larger than the other catchments (Table 2.1). This larger area is chosen to rule out any significant groundwater transport over the catchment boundary and to ensure a good quality of discharge measurements [Veenstra, 2009]. The larger catchment area might cause a slight increase in lag, because of the travel time of water within the catchment (see Sect. 1.2.5), but is otherwise not expected to influence drought propagation processes, mainly due to the generally flat topography of the catchment. The altitude range is approximately 600–1100 m a.m.s.l. and, especially in the centre, topography is rather flat, sloping gently from north-east to south-west.

The Upper-Guadiana catchment has (in part of the catchment) a Mediterranean and (in part of the catchment) a 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 evapotranspiration was around 1250 mm yr^{-1} (Table 2.1). 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 evapotranspiration exceeded precipitation, resulting in a relatively low measured mean discharge of 16 mm yr^{-1} [De la Hera, 1998]. The Upper-Guadiana has a strong seasonality in meteorological forcing, with relatively high precipitation and low potential evapotranspiration in winter and relatively low precipitation and high potential evapotranspiration in summer. This results in a strong seasonality in recharge and, somewhat attenuated, also in discharge. Highest river flows occur in winter and the low-flow season is summer (Table 2.1). Additionally, interannual variation in precipitation is large. In dry years, rivers in the catchment can go dry during summer.

The Upper-Guadiana is a groundwater catchment with many areas consisting of multiple layers of sedimentary rock, forming large aquifer systems, particularly in the centre. These aquifer systems are underlain by basement gneiss cropping out along the southern and eastern boundaries [Bromley et al., 2001]. In the northeastern and southwestern part of the catchment, some relatively small groundwater units are located, named Sierra de Altomira and Campo Montiel, respectively. The centre of the catchment is underlain by the large groundwater system La Mancha Occidental. It is made up of two hydrogeological units (Miocene and Jurassic limestones), partly separated by a less permeable layer (Cretaceous) and underlain by impervious Paleozoic material [Martínez-Santos et al., 2008; Martínez-Santos and Martínez-Alfaro, 2010]. Hydrogeological boundaries between the Upper-Guadiana aquifer units are complex [Martínez-Santos and Martínez-Alfaro, 2010]; the reader is referred to the elaborate descriptions of IGME [1985] and Veenstra [2009]. The presence of these aquifer systems makes Upper-Guadiana a slow responding catchment.

A number of interconnected wetlands cause further delay in the response to precipitation. These wetlands are the main natural discharge areas of the aquifer system and show the strong groundwater–surface water interaction in the Upper-Guadiana catchment [Bromley et al., 2001; Martínez-Santos and Martínez-Alfaro, 2010]. The main wetland area, Tablas de Daimiel, is a internationally valued and protected UNESCO Biosphere Reserve and a RAMSAR-site. Additionally, three small reservoirs are present in the Upper-Guadiana catchment (Peñarroya, El Vicario, and Gasset), which were built between 1900 and 1975 and are used for a.o. irrigation and domestic water supply [Veenstra, 2009]. These reservoirs simply act as additional (surface water) storage and do not alter the hydrological regime (see CEDEX: hercules.cedex.es/anuarioaforos). Their influence can be regarded as comparable to that of the natural stores in the catchment, i.e. aquifers and wetlands. As the area covered by the artificial reservoirs is minor (0.055% of the catchment area) in comparison with the area covered by natural wetlands (approx. 2%; Sánchez-Andrés et al. [2010]) and the area underlain by aquifers (almost the entire catchment), the effects of the artificial reservoirs on the total catchment discharge was assumed to be negligible.

Land use in the Upper-Guadiana catchment is mainly agricultural. Before the 1970s, dryland farming of cereals and permanent agriculture prevailed, with only limited irrigation. Since 1970–1980, human influence in the catchment increased dramatically through intensified irrigated agriculture, causing declining groundwater levels and wetland area, and decreasing discharge [Bromley et al., 2001].

2.2 Data

In hydrology, and especially in drought analysis, availability of long time series of undisturbed, observational data is essential [Santos et al., 2002; Rees et al., 2004]. Often, however, observational records are not long enough, some variables are not monitored at all, data quality is too low, or observations are influenced by human activities. To overcome these problems hydrological models can be used to extend data series, fill gaps, and naturalise disturbed time series (see Ch. 3). However, models also require data. Models need to be forced with observed meteorological data and hydrological data are needed for calibration and validation. In this section, I describe the meteorological and hydrological data that were available for the case study areas.

2.2.1 Meteorological data

2.2.1.1 Precipitation and temperature

For all case study areas, the basic meteorological data (daily temperature and precipitation) were available from stations inside or around the catchment (Figs. 2.1 and 2.2, and Van Lanen et al. [2008]).

- For the Narsjø catchment, meteorological data were measured in three stations. Daily temperature was measured in the meteorological station Røros (25 km north of the catchment, not in Fig. 2.1d). Daily precipitation was measured in the stations Ellefsplass and Tufsingdal (the latter was moved to Tufsingdal-Midtdal in 1991; Fig. 2.2), located on either side of the catchment (Fig. 2.1d). Catchment precipitation was calculated by taking the arithmetic mean of the data of these two stations.

- For the Upper-Metuje catchment, daily mean temperature and precipitation were measured in the Bučnice meteorological station (Figs. 2.1b and 2.2; Rakovec et al. [2009]).

- For the Upper-Sázava, daily temperature data were available for two stations, Přibyslav and Svratouch. Precipitation was recorded in Přibyslav, Svratouch, Krucemburk, Žďár nad Sázavou-Stržanov, Křižánky, and Kadov (Fig. 2.1c). However, in some stations the observation period was very short, data quality was low, or there were many gaps in the time series, so, finally, records from only the two professional meteorological stations in Přibyslav and Svratouch (Fig. 2.2) were used. Catchment average temperature and precipitation were calculated using Thiessen polygons [Rakovec et al., 2009].

- For the Nedožery catchment, meteorological data were measured in a number of stations in and around the catchment (Fig. 2.1e). Daily temperature data were derived from two meteorological stations: Prievidza and Turcianske Teplice, and daily precipitation measurements from five stations: Nitrianske Pravno, Chvojnica, Vricko, Slovenské Pravno and Valaská Belá-Gapel. Catchment average temperature and precipitation were calculated using Thiessen polygons [Oosterwijk et al., 2009].

- For the Upper-Guadiana, temperature and precipitation data were taken from meteorological stations inside and around the Upper-Guadiana catchment (Figs. 2.1f and 2.2) and spatially averaged using Thiessen polygons [Veenstra, 2009].

For the large-scale studies (Chs. 6 and 7), precipitation and temperature data were obtained from the WATCH Forcing Data [WFD, Weedon et al., 2011]. The WFD dataset consists of gridded time series of meteorological variables (e.g. rainfall, snowfall, temperature, wind speed) on a daily basis for the time period 1958–2001. The data have a spatial resolution of 0.5° based on the CRU land mask. The WFD originate from modification (e.g. bias correction and downscaling) of the ECMWF ERA-40 re-analysis data [Uppala et al., 2005]. The data have been

Figure 2.2: Photos of rain gauges in and around the selected catchments.

interpolated and corrected for the elevation differences between the grids of ERA-40 and CRU. For precipitation, the ERA-40 data were first adjusted to have the same number of wet days as CRU [Brohan et al., 2006]. Next, the data were bias-corrected using monthly GPCC precipitation totals [Schneider et al., 2008] and, finally, gauge-catch corrections were applied. For temperature, the ERA-40 data were bias-corrected using CRU monthly average temperatures and temperature ranges. For more information the reader is referred to Weedon et al. [2011]. The WFD have been used to force the large-scale hydrological models of Ch. 6 and the synthetic model used in Ch. 7.

2.2.1.2 Potential evapotranspiration

Another important input variable for hydrological models is potential evaporation (PE). PE indicates the atmospheric vapour demand and is used to simulate actual evapotranspiration, an important component of the water balance. PE values can be calculated with a variety of methods that can roughly be divided in radiation-based (e.g. the Penman-Monteith method; Monteith [1965]) and temperature-based methods (e.g. the Thornthwaite method; Thornthwaite [1948]). Some controversy exists about the influence of using a different PE calculation method on hydrological modelling results and especially on drought. Dai [2011] found that global trends in a (soil moisture) drought index are not influenced by the PE calculation method, whereas Sheffield et al. [2012] found a global drying trend when they used a temperature-based

Figure 2.3: Photos of the meteorological stations of the selected catchments.

method and no trend when using a radiation-based method. Locally, results of PE calculations using different methods can vary considerably, especially in (semi-)arid regions [e.g. Er-Raki et al., 2010].

The FAO Expert Consultation on Revision of FAO Methodologies for Crop Water Requirements advised the use of the radiation-based Penman-Monteith method as the standard method for the calculation of PE [Allen et al., 1998]. In 1948, Penman combined the energy balance and the mass transfer method and proposed a formula to calculate evaporation from an open water surface [Penman, 1948]. This formula used data from standard weather records of sunshine, temperature, humidity, and wind speed. Later, Penman added the transpiration component of vegetation. In 1965, the Penman formula was reformulated by Monteith [1965] in order to make it applicable to dry, horizontal, vegetated surfaces with optimal water supply [Melsen et al., 2011].

Incoming radiation is, however, not a widely measured variable, which limits the use of the Penman-Monteith method. Incoming radiation can be approximated by a formula that uses extraterrestrial radiation and minimum and maximum temperature as input [Allen et al., 1998]. This approach is preferable to purely temperature-based methods, because it stays close to the physical processes and does not reduce to a purely empirical formula. Rakovec et al. [2009] and Veenstra [2009] investigated the effect in the Upper-Metuje, Upper-Sázava, and Upper-Guadiana catchments and found that long-term monthly averages of PE are similar when radiation data are replaced by estimates based on temperature in the Penman-Monteith method.

When used as input for a hydrological model, the Penman-Monteith method using either radiation or minimum and maximum temperature gave similar values for daily actual evapotranspiration. This is consistent with the studies of Weiss and Menzel [2008] and Vangelis et al. [2013]. Therefore, I adopted the FAO Penman-Monteith method described by Allen et al. [1998] for this research. Due to different data availability and quality in the catchments, slightly different calculation procedures were followed according to the assumptions and recommendations described by Doorenbos and Pruitt [1975] and Allen et al. [1998].

- For the Narsjø catchment, meteorological data for the calculation of the FAO Penman-Monteith PE (daily minimum and maximum temperature, wind speed, and vapour pressure) were obtained from the Røros meteorological station (Fig. 2.3).

- For the Upper-Metuje catchment, measurements of daily minimum and maximum temperature from the Bučnice meteorological station were used to calculate PE (Figs. 2.1b and 2.3; Rakovec et al. [2009]).

- For the Upper-Sázava catchment, meteorological data for the calculation of the FAO Penman-Monteith PE (daily minimum and maximum temperature, wind speed, and solar radiation) were obtained from the Svratouch station (Figs. 2.1c and 2.3; Rakovec et al. [2009]).

- For the Nedožery catchment, meteorological data for the calculation of the FAO Penman-Monteith PE (daily minimum and maximum temperature, wind speed, cloudiness, and relative air humidity) were obtained from the stations Prievidza and Turcianske Teplice (Figs. 2.1e and 2.3; Oosterwijk et al. [2009]).

- For the Upper-Guadiana catchment, PE was calculated from 18 fully-equipped meteorological stations inside and around the catchment, and was afterwards averaged using Thiessen polygons (Figs. 2.1f and 2.3; Veenstra [2009]).

The WATCH forcing data (see p. 22) include radiation in the dataset, but some inconsistencies were found between daily radiation and temperature data [Melsen et al., 2011]. Therefore, the same procedure was applied to this large-scale dataset as to the local data, i.e. replacing radiation with an approximation based on extraterrestrial radiation and minimum and maximum temperature.

2.2.2 Hydrological data

Hydrological data are needed for calibration and validation of hydrological models. For all case study areas, discharge data were available from the outlet station (Figs. 2.1 and 2.4, and Van Lanen et al. [2008]). For validation, some snow, soil moisture, and groundwater data were available.

- For the Narsjø catchment, daily discharge was recorded at the outlet of Lake Narsjø (Figs. 2.1d and 2.4). Soil moisture and groundwater levels were measured in a location close to, but outside the Narsjø catchment. This location is not fully representative for the Narsjø, but the measurements can be used to validate the temporal dynamics of the simulations [Hohenrainer, 2008]. The soil moisture and groundwater level measurements were performed at time intervals of about one week for the period 1980–2000.

- For the Upper-Metuje catchment, daily discharge was measured at the outlet of the catchment at gauging station MXII, Teplice nad Metují (Figs. 2.1b and 2.4). Groundwater observation wells with various depths are located in the centre of the catchment [Rakovec et al., 2009].

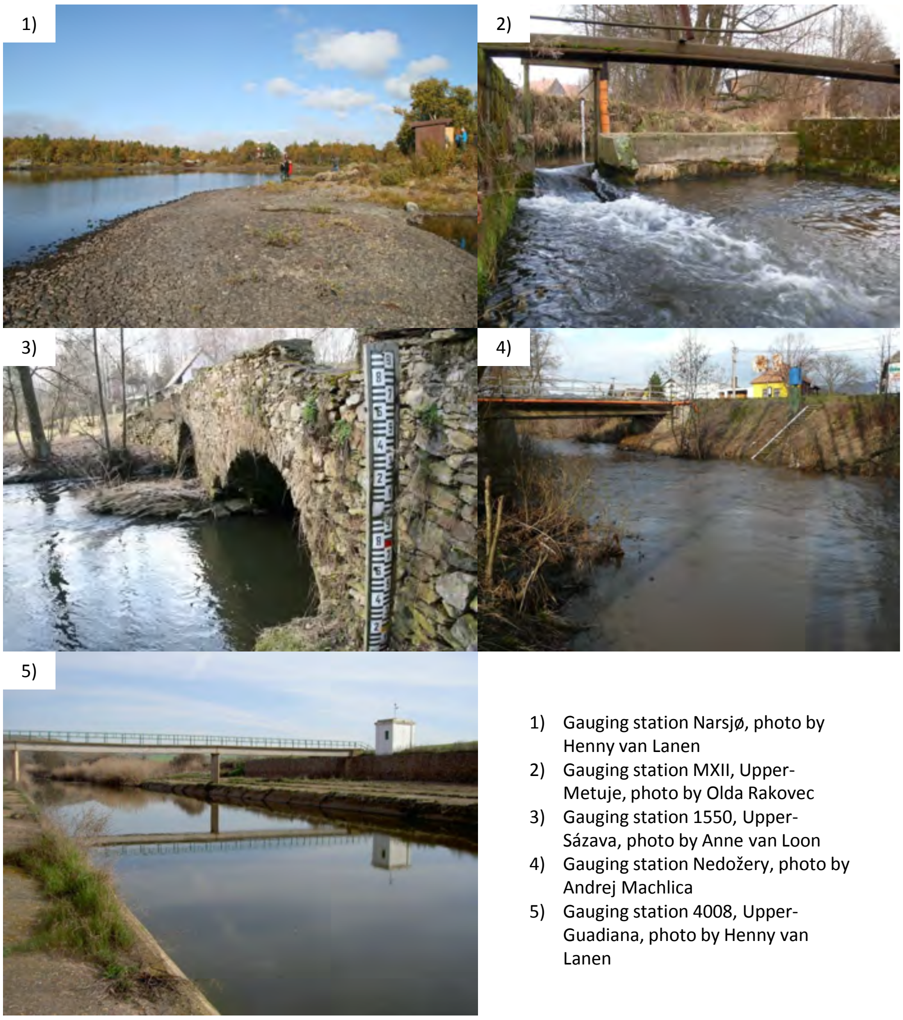

Figure 2.4: Photos of the gauging stations of the selected catchments.

- For the Upper-Sázava catchment, daily discharge was measured at gauging station 1550 (Sázava u Žďáru nad Sázavou) (Figs. 2.1c and 2.4). From the meteorological station of Žďár nad Sázavou-Stržanov daily snow height and weekly snow water storage records were available. Groundwater head observations were taken at weekly intervals at a groundwater well in the upper-reach of the catchment. The quality of the groundwater data is considered to be quite low [Rakovec et al., 2009].

- For the Nedožery catchment, daily discharge was measured at gauging station Nedožery (Figs. 2.1e and 2.4). Groundwater heads were measured in a well close to the catchment outlet and information on snow cover was provided by the Prievidza meteorological station (Figs. 2.1e and 2.3; Oosterwijk et al. [2009]).

- For the Upper-Guadiana catchment, discharge was measured at the outlet of the catchment, station 4008, Balbuena (Figs. 2.1f), by means of a specially designed measurement structure with a low-flow channel (Fig. 2.4). Discharge was measured from 1960 to 2001, on a daily basis. Some gaps and 'staircase' data [Rees et al., 2004] were present in the time series, which were linearly interpolated (5 % of the entire time series; Veenstra [2009]. In this catchment, many groundwater observation wells have been installed (approximately 120; Fig. 2.1f). Unfortunately, most of these wells have only a short period of record and data in the period before 1980 were limited. Measurements were taken on a monthly basis and gaps in the time series were linearly interpolated.

Table 2.2: Scales and model approaches used in the different chapters of this thesis

	catchment scale (case studies)	global scale (grid cells)
catchment-scale model	Chs. 3, 4, 5	
large-scale models	Ch. 6	Ch. 6
synthetic model		Ch. 7

All datasets used in this research have uncertainties (Sect. 3.4). Both the catchment-scale and large-scale data have been thoroughly checked and robust methods were used for gap filling, bias correction, and PE calculation [Rakovec et al., 2009; Oosterwijk et al., 2009; Veenstra, 2009; Weedon et al., 2011]. However, data are never perfect, especially not during drought conditions.

2.3 Hydrological modelling

If observational data of drought-related variables are not available, if the period of record is insufficient or quality is low, modelling is required [Tallaksen and Van Lanen, 2004; Mishra and Singh, 2010; Dai, 2011; Sheffield and Wood, 2011; Seneviratne et al., 2012]. Modelling is current practice in hydrology, both in science and in operational water management. Hydrological models range from simple statistical models with a few parameters via conceptual models with varying complexity to complex physically-based models (for an overview of current modelling approaches, see for example Beven [2000], Wagener et al. [2004], Matonse and Kroll [2009]). Hydrological models are usually designed to simulate average and high flows and have been shown to give good results in catchments around the world. The application of these models specifically to low-flow situations has been relatively limited [Smakhtin, 2001]. In low-flow studies the focus is mainly on statistical methods, such as indices and extreme value analysis [WMO, 2008]. However, if there are not enough data for these methods, models are used.

The choice of model is not straightforward. In this research I used three types of models (see Table 2.2):

- a conceptual, semi-distributed, catchment-scale rainfall-runoff model in Chs. 3, 4, and 5;
- an ensemble of a number of physically-based, distributed, large-scale hydrological models and land surface models in Ch. 6; and
- a conceptual, distributed, synthetic hydrological model in Ch. 7.

As the first type of model is used in more than one chapter, I present a general description of that modelling approach in this section. The other modelling approaches are treated in the chapters in which they are used. I also add sections on the improvement of the catchment-scale rainfall-runoff model for the simulation of low flows (Sects. 2.3.3 and 2.3.4).

2.3.1 HBV model

The conceptual, semi-distributed rainfall-runoff model HBV [Seibert, 1997] was chosen as the catchment-scale hydrological 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]; in catchments in Europe: Austria [Merz and Blöschl, 2004], Belgium [Van Pelt et al., 2009; Driessen

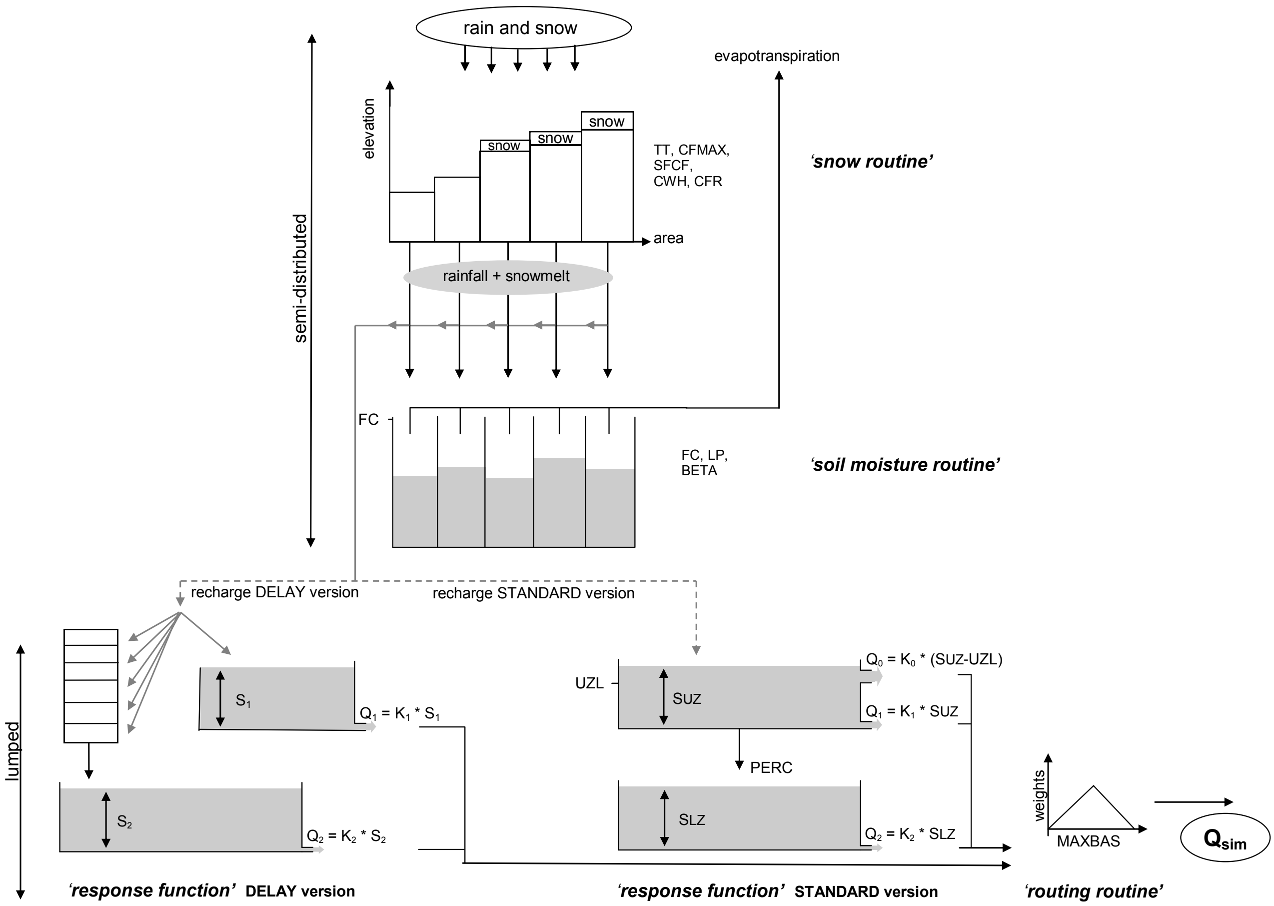

Figure 2.5: Structure of the HBV model with two versions for the response routine: on the right-hand side the STANDARD version, and on the left-hand side the DELAY version (adapted from Seibert, 2000 and Oosterwijk et al., 2009).

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] and selected catchments in Africa and South America [Lidén and Harlin, 2000]. HBV was used in various low-flow and drought studies [e.g. De Wit et al., 2007; Te Linde et al., 2008; Engeland and Hisdal, 2009; Bisterbosch, 2010; Van Huijgevoort et al., 2010]. In this research I used 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 thesis.

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. 2.5). Snow accumulation and melt are calculated by the degree-day method for a number of elevation (maximum 10) and vegetation (maximum 3) zones separately. In each of these zones groundwater recharge and actual evapotranspiration are functions of actual water storage in the soil moisture routine. 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 suited for modelling slow responding deep-groundwater catchments, I tested this version in addition to the STANDARD response routine. The DELAY response routine consists of two linear reservoirs in parallel, of which the lower reservoir is preceded by a distribution of recharge over different delay boxes (Fig. 2.5).

2.3.2 Calibration and validation of the HBV model

Parameter values of HBV were determined by calibration for each study area (Sect. 2.1) separately. Calibration focused on correctly reproducing observed discharge and for that purpose I used 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]. Clark et al. [2008] advocate the development of a new metric for model evaluation, especially for the assessment of model performance during low-flow periods [Boyle et al., 2000]. Recently, Pushpalatha et al. [2012] recommended a Nash-Sutcliffe efficiency calculated from inverse flow values. In this research the Nash-Sutcliffe efficiency of the logarithm of observed and simulated discharge (ln Reff) was used as the objective function for low-flow modelling [Seibert, 1999, 2005; Krause et al., 2005].

The entire observation period (Table 2.1) was used as the calibration period in all catchments, except Upper-Guadiana. Due to the strong human influence in that catchment after 1980 (see Sect. 2.1.5) the calibration period was restricted to 1960–1970, and the period 1970–1980 was used for validation.

In catchments for which data on snow, soil moisture, and/or groundwater were available (Sect. 2.2) a validation has been performed by means of a visual inspection of the time series and a calculation of the coefficient of determination r^2. Simulated groundwater storage first had to be converted to groundwater levels, for which I used a constant storage coefficient concept. This was done by a linear regression on observed groundwater levels (see Rakovec et al. [2009], Oosterwijk et al. [2009], and Veenstra [2009]).

Several output variables of HBV were used for further drought analysis, 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 I used only storage in the lower groundwater reservoir (SLZ or S2; Fig. 2.5), which represents deep groundwater. The reason for not including storage in the upper groundwater reservoir (SUZ or S1; Fig. 2.5) is that fast flow paths (e.g. surface runoff) are modelled through this upper reservoir; hence it does not represent real groundwater storage (Fig. 2.5).

2.3.3 Challenges in low-flow modelling

Unfortunately, low flows are often not captured satisfactorily by models [Smakhtin et al., 1998; Engeland et al., 2006; Lehner et al., 2006; De Wit et al., 2007; Kumar and Samaniego, 2008; Basu et al., 2010; Kumar et al., 2010]. 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'. As the HBV model is not designed especially for low flows, we anticipated that HBV also has difficulties in correctly simulating low flows [Bergström, 1997; Uhlenbrook et al., 1999].

According to Matonse and Kroll [2009], low flows could potentially be simulated in a better way. Recently, various attempts have been made to improve low-flow modelling using existing models. Perrin et al. [2003] improved a lumped rainfall-runoff model to match both high and low flows. Matonse and Kroll [2009] used hillslope storage models (i.e. kinematic wave hillslope storage and hillslope storage Boussinesq models) to improve groundwater flow in a small steep headwater catchment. Romanowicz [2007] used a combination of a physically-based model (TOPMODEL) and stochastic transfer functions based on a logarithmic transformation of flows. Basu et al. [2010] focused on riparian zones to improve low-flow modelling in a simple threshold-based model. Pushpalatha et al. [2011] added a routing reservoir to a conceptual rainfall-runoff model. These studies show a minor improvement in the simulation of low flows, but no approach is explicitly the best; in all approaches deficiencies remain.

According to WMO [2008], 'one of the central issues is the choice of model structure and the level of complexity included', and also Clark et al. [2008] state that 'the choice of model structure is just as important as the choice of model parameters'. In low-flow modelling, current practice is the use of *a priori* determined model structures [Fenicia et al., 2008b], although these are designed for average and high flow conditions. This general approach overlooks the fundamental differences that exist in hydrological processes that lead to high and low flows. However, alternative, more tailor-made approaches to hydrological modelling have been developed. One of them is the 'top-down' or 'downward' approach, originally introduced by Klemeš [1983], reformulated by Sivapalan et al. [2003], Farmer et al. [2003], Sivapalan [2009] and applied by o.a. Basu et al. [2010] and Tekleab et al. [2011]. Other examples include the data-based mechanistic approach (DBM) [Young and Beven, 1994; Young, 1998, 2006], the development of flexible modelling frameworks [Wagener et al., 2001; Fenicia et al., 2011], step-wise model improvement based on data [Fenicia et al., 2008a,b], and approaches that consider multiple plausible model architectures and flux parameterisations [Marshall et al., 2007; Clark et al., 2008, 2009, 2011]. Although their methodology is different, these approaches have one thing in common: they are based on systematic learning from data. The model structure is flexible, not *a priori* defined, and model components are selected and combined based on (different sorts of) data [Sivapalan et al., 2003]. Just like the traditional models, these approaches have mainly been tested to and used for the prediction of average and high flows. Applications in semi-arid regions, but still focusing on high flows, are presented in Mwakalila et al. [2001] using the DBM approach and in Jothityangkoon et al. [2001] using the 'downward' approach. Staudinger et al. [2011] did a multi-model analysis with the FUSE framework in the Narsjø catchment. They did not find an overall best model structure for both summer and winter low flows. Apart from the attempts of Romanowicz [2007] and Pushpalatha et al. [2011], no application of the 'downward' or data-based modelling approach has been performed with specific focus on low flows.

Our objective is to investigate the possibilities to improve low-flow modelling by the HBV model using a data-based approach on the basis of recession analysis for the case study areas (Sect. 2.1). Recession curves have often been used to infer general hydraulic or hydrological properties of a catchment [Brutsaert and Nieber, 1977; Troch et al., 1993; Tallaksen, 1995; Lamb and Beven, 1997; Wittenberg and Sivapalan, 1999; Fenicia et al., 2006; Kirchner, 2009]. For the estimation of low-flow characteristics or groundwater discharge, in particular, base-flow recession analysis is widely used with good results [Vogel and Kroll, 1992; Wittenberg and Sivapalan, 1999; Jothityangkoon et al., 2001; Demuth and Young, 2004; Kroll et al., 2004; Eng and Milly, 2007]. This gives us confidence that we can use recession analysis for the selection of the model structure that is most suitable for the simulation of low flows. The methodology and results of this investigation are elaborated in Appendix A.

2.3.4 Discussion on the attempt to improve the HBV model for low flows

The findings presented in Appendix A can be summarised as follows:

- Recessions of all studied catchments can be modelled using the same model structure (one non-linear reservoir) with different parameters.

- In some of the catchments, all recessions can be modelled adequately with one model structure and one fixed parameter set. For those catchments, the parameters of the outflow relationships are robust. In other catchments, this is not the case and results decrease drastically when fixing model structure and parameters.

- In all catchments, the step towards modelling the entire hydrograph including peaks (from event-based to continuous modelling) still poses problems.

- The final, most-promising model structure (a linear reservoir overflowing into a non-linear reservoir, Fig. A.5) does not perform better than the original HBV model.

So, in the end, we did not succeed in improving the HBV model for low flows. In this section I give an overview of the possible reasons for this lack of success.

Failure to improve HBV using a data-based approach might be related to the recession analysis itself. First, data quality at low flow is usually poor. Discharge data can have a 'staircase' pattern, because there is less accuracy in the low-flow reach and because weed growth and sedimentation often decrease data quality [Rees et al., 2004; WMO, 2008]. Moreover, data errors have a relatively large influence at low flows. Second, human influence can be especially pronounced during low flow situations [Wittenberg, 2003]. Wang et al. [2009], for example, found large effects of water withdrawal and return flow on base-flow recession. Finally, an increase in evaporation accelerates the rate of recession [Wittenberg and Sivapalan, 1999]. In this study, evapotranspiration effects were only implicitly taken into account, because we selected recession periods based on a criterion for recharge (see Sect. A.1.3). Despite these known difficulties, we obtained good results from our low-flow recession analysis. The parameters estimated in Sect. A.2.1 were quite robust.

We encountered most problems when we changed from event-based to continuous modelling, i.e. when we included recharge from the upper part of the HBV model (Sect. A.2.3). Possible explanations for the decrease in skill when modelling the entire hydrograph are the following.

- Stochastic fluctuations in the storage-discharge relationship lead to noise [Suweis et al., 2010], which decreases the suitability of discharge data to infer storage.

- In some catchments, recessions are very different and cannot be modelled with a fixed parameter set. Exceptions are the Narsjø and Upper-Metuje catchments that have uniform recessions, which are governed by one dominant process. In the other catchments, seasonal variation in the processes underlying recessions could be an explanation, but no seasonal variation in optimal parameters was found.

- Recharge pulses during the recession change the exponent of the power law [Birk and Hergarten, 2010]. In this study we included all recessions with less than 1.0 mm d^{-1} recharge averaged over the recession period, with the aim to focus on long recessions (Sect. A.1.3). As no input was used during recession analysis, this probably resulted in an underestimation of the recession coefficient, which resulted in difficulties in reproducing the hydrograph when recharge finally was included. The results improved when we did not fix the recession parameters beforehand, but fitted them afterwards (Sect. A.2.3). They might have improved even more when the upper part of the model would also have been changed. For example, Fenicia et al. [2008a] found that simulation of low flows improved when interception and evaporation from interception were added.

- Spatial differentiation within the catchment is important during low flows. Seibert et al. [2003], Fenicia et al. [2008b], and Basu et al. [2010] found that modelling results highly depend on catchment differentiation and that especially the riparian zone plays an important role. This problem might be tackled by the additional use of spatial information on groundwater levels in data-based model development.

Our conclusions are substantiated by the systematic approach and contrasting catchments used in this study. We approached the limits of conceptual modelling of low flows, because even with such a systematic data-based approach low-flow model results could not be improved.

Discharge data probably do not contain enough information to take decisions about the best model structure. The challenges for low-flow research lie in adequate understanding of the processes underlying low flows. Romanowicz [2007] faced the same challenge and concluded that 'further research is needed to choose the best way of predicting both low and high flows'. Kværner and Klöve [2008] recommended to model low flows and high flows separately, but we believe that for drought studies high-flow events are also important as they can prevent a hydrological drought from developing or lead to recovery from a drought situation (see Sect. 5.5.3). Staudinger et al. [2011] found that they could slightly improve winter low-flow modelling in the Narsjø catchment. Simulations of summer low flows were poorer, because various model structures (both the soil and groundwater parts) were identified to influence model performance during summer. This is consistent with our results in Appendix A. Pushpalatha et al. [2011] added a new response routine to an existing conceptual hydrological model and obtained only a slight improvement in modelling low flows at the expense of an extra model parameter.

From this analysis I conclude that the HBV model cannot be improved for the simulation of low flows using a data-based approach on the basis of recession analysis. I use the original HBV model as described in Sect. 2.3.1 in the remainder of this thesis (Chs. 3, 4, and 5; Table 2.2). The performance of an ensemble of large-scale models on drought is investigated in Ch. 6, which also includes a discussion on the suitability of this type of models for drought analysis. Ch. 7 uses a synthetic model that is based on the HBV modelling concepts for the snow and soil moisture routines (Fig. 2.5).

2.4 Drought analysis

In order to understand drought processes and impacts, drought characteristics such as the timing, duration, severity (or intensity), and spatial extent of a drought event need to be identified [Wilhite, 2000; Panu and Sharma, 2002; Tallaksen and Van Lanen, 2004; Mishra and Singh, 2010; Seneviratne et al., 2012]. Their slow onset and slow recovery, the different drought categories (Fig. 1.1) and impacted sectors (Sect. 1.1) make droughts very difficult to define (Sect. 1.2.1), giving rise to a multitude of indices. Reviews of drought indices can be found in Heim Jr. [2002], Keyantash and Dracup [2002], Hisdal et al. [2004], Niemeyer et al. [2008], Mishra and Singh [2010], Wanders et al. [2010], Dai [2011], Sheffield and Wood [2011], and Seneviratne et al. [2012]. The choice of index and its implementation are important as they can result in different conclusions, especially in the light of trends and global change (Sect. 1.1 and Burke and Brown [2008]; Sheffield et al. [2012]).

In this section I do not go into details on the multitude of existing drought indices. Instead I focus on a few widely-used indices for the characterisation of meteorological, soil moisture, and hydrological drought and provide information on how drought characteristics are determined in this thesis.

2.4.1 Drought indices

Meteorological drought indices use precipitation as an input. Because precipitation has a high spatial and temporal variability, meteorological drought indices often use monthly values [Wanders et al., 2010]. The most-used meteorological drought index is the Standardised Precipitation Index (SPI) [McKee et al., 1993; Lloyd-Hughes and Saunders, 2002]. It is based on long-term precipitation records that are fitted to a probability distribution (Fig. 2.6). This distribution is then transformed to a normal distribution, ensuring zero mean and unit standard deviation, so that regional comparison of SPI values is possible. SPI can be computed over several time scales (e.g. 1, 3, 6, 12 months or more) and thus indirectly considers effects of accumulating precip-

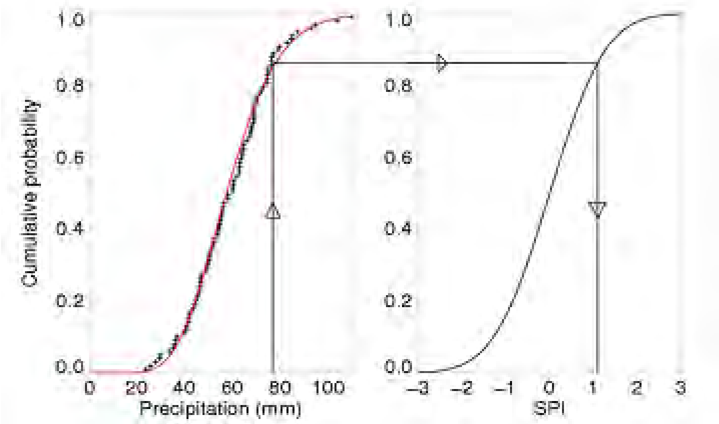

Figure 2.6: Methodology to determine the Standardised Precipitation Index (from drought.mssl.ucl.ac.uk/spi.html).

itation deficits. SPI calculated over long time scales is sometimes used as an approximation of hydrological drought [e.g. Szalai et al., 2000; Nalbantis and Tsakiris, 2009; Zhai et al., 2010].

Experts participating in a WMO drought workshop in 2009 recommended that the SPI be used by all National Meteorological and Hydrological Services (NMHSs) around the world to characterise meteorological drought [WMO, 2012]. Advantages of SPI are that its calculation results in normalised values and that it can be computed for different time scales [Sheffield and Wood, 2011]. Disadvantages of SPI are that only precipitation is considered, while other meteorological drivers might be important too [Dai, 2011]. Additionally, the length of a precipitation record and the fitted probability distribution have significant impact on the SPI values [Mishra and Singh, 2010]. Fitting a distribution in dry climates can be problematic [Wanders et al., 2010], which limits the use of SPI on a global scale. Another drawback of standardised indices like SPI is that the severity of a drought event is expressed only in relative terms, while in water resources management absolute values of the lacking amount of water with regard to 'normal' conditions (i.e. deficit volume) are needed.

Other commonly-used meteorological drought indices are the Consecutive Dry Days index (CDD) and Rainfall Deciles (RD). The CDD method considers the maximum consecutive number of days with no measurable precipitation within a considered period (i.e. a year in general; Deni and Jemain [2009]). CDDs main use is in climates with a distinct dry season, like monsoonal or savannah climates (see Sect. 1.2.3), but it is also regularly applied in research on climate change impacts [Seneviratne et al., 2012]. The RD method compares monthly aggregated data of precipitation (rain and snow) with average values extracted from long-term observations. RD is used by the Australian Drought Watch System, because it is easy to calculate [Hayes, 2000]. However, long records of data are required to obtain accurate values.

As precipitation is not the only meteorological variable influencing drought conditions, some meteorological indices also include (a proxy for) evapotranspiration. As an alternative for SPI, Vicente-Serrano et al. [2009] developed the Standardised Precipitation and Evapotranspiration Index (SPEI). SPEI considers cumulated anomalies of precipitation and potential evapotranspiration and, like SPI, fits a probability distribution and transforms it into a normal distribution [Seneviratne et al., 2012]. Another index that reflects both precipitation and evapotranspiration is the Palmer Drought Severity Index (PDSI). It has been developed by Palmer [1965] for the USA and it is applied all around the world [e.g. Burke and Brown, 2008; Dai, 2013]. It measures the departure of the moisture balance from normal conditions using a simple water balance model and can be regarded as a hydrological accounting system [Dai, 2011]. PDSI is sometimes classified as a meteorological drought index [Dai, 2011] and sometimes as a soil moisture drought index [Sheffield and Wood, 2011]. Advantages of PDSI are that it has been much used and that it is standardised, which makes comparisons between different climatic zones possible [Mishra and Singh, 2010]. Despite its worldwide application, PDSI has important shortcomings that should limit its use on the global scale: i) the calculation procedure is complex and non-transparent [Sheffield and Wood, 2011], ii) the time scale is fixed [Mishra and Singh, 2010], iii) it uses a potential evaporation method based on absolute temperature, which in some regions can have large impact (see Sect. 2.2.1.2), iv) as it is calibrated for the USA, re-calibration is needed for application to other regions [Dai, 2011], and v) snow accumulation is not accounted for and no soil moisture or vegetation control on evapotranspiration is included [Seneviratne et al., 2012].

Palmer also developed a soil moisture drought index (Z-index) and a hydrological drought index (PHDI) [Palmer, 1965], which have similar calculation procedures as the PDSI and, therefore, the same advantages and disadvantages [Wanders et al., 2010]. Recent developments in soil moisture drought indices include the Soil Moisture Deficit Index (SMDI; Narasimhan and Srinivasan [2005]) and Soil Moisture Index (SMI; Hunt et al. [2009]), but their application is still very limited [Mishra and Singh, 2010]. Sheffield et al. [2004] use an SPI-like procedure to calculate soil moisture percentiles.

Indices for the characterisation of hydrological drought use different hydrological variables (from observed or simulated data) as input. Most common is a focus on streamflow, because streamflow is most measured, most easily simulated, and of most interest to water resources management. Other variables used in hydrological drought indices include groundwater levels and lake levels [e.g. Peters, 2003; Hisdal et al., 2004; Peters et al., 2006; Tallaksen et al., 2009]. The Standardised Runoff Index (SRI) has a calculation procedure similar to SPI [Shukla and Wood, 2008], fitting a distribution to streamflow data and transforming it to a normal distribution. The limitations of SPI also apply to SRI, i.e. the length of the data record and the fitted distribution strongly influence SRI values. The Surface Water Supply Index (SWSI) is developed by Shafer and Dezman [1982]. It is calculated based on non-exceedance probabilities from historical records of reservoir storage, streamflow, snow pack, and precipitation [Wanders et al., 2010]. Contrary to PDSI and PHDI, SWSI does take into account snow accumulation and water storage. It is, however, very basin-dependent, which limits its use on the global scale [Mishra and Singh, 2010]. Low-flow indices, for example mean annual minimum flow [Stahl et al., 2010], or the Baseflow Index (BFI) [Hisdal et al., 2004] are not considered in this thesis (see Sect. 1.2.1).

Some newly-developed drought indices are derived from satellite information. These focus on vegetation and are a measure of general vegetative condition. Advantages are that satellite data provide a large spatial coverage and high spatial resolution. Disadvantage is that it is difficult to discern other influences on vegetation health [Sheffield and Wood, 2011]. For the 2005 and 2010 drought events in the Amazon rainforest (Sect. 1.1), satellite-derived information has been very useful in quantifying the extent and severity of the droughts. The European Drought Observatory uses remote sensing information in its Combined Drought Indicator (CDI[1]; Sepulcre-Canto et al. [2012]). They provide 10-day updates of the agricultural drought status in Europe by integrating the meteorological index SPI (on 1, 3 and 12-month scales), simulated soil moisture anomalies, and a vegetation stress indicator derived from satellite information.

Besides these at-site indices, some regional indices exist that quantify the spatial aspect of drought [e.g. Andreadis et al., 2005; Peters et al., 2006; Sheffield et al., 2009; Tallaksen et al., 2009]. Most of these indices calculate the portion or percentage of an area in drought. The Regional Deficiency Index (RDI), for example, divides the number of catchments in drought by the total number of catchments [Stahl, 2001; Hannaford et al., 2011] and the Regional Drought Area Index (RDAI) divides the drought area by the total area of the region [Fleig et al., 2011]. Regional indices are not considered in this thesis, as the spatial aspects of drought are not investigated (Sect. 1.2.5).

[1] edo.jrc.ec.europa.eu/documents/factsheets/factsheet_combinedDroughtIndicator.pdf

Besides these more or less complex indices, drought characteristics can also be derived directly from time series of observed or simulated hydrometeorological variables using a predefined threshold level. When the variable is below this level, the site is in drought. Drought duration, severity, and frequency can easily be calculated. This approach is called 'threshold level method' [e.g. Yevjevich, 1967; Hisdal et al., 2004; Fleig et al., 2006], but the term 'deficit index' is also used [Laaha et al., 2013], because it measures the 'lacking' volume of water below a certain threshold. This so-called deficit volume can only be calculated by the threshold level method and not by the drought indices mentioned above. This is a big advantage of the threshold level method, because deficit volume is an important drought characteristic in water resources management.

All three categories of drought (meteorological, soil moisture, and hydrological drought) can be analysed with the threshold level method. This makes comparison between variables possible, which is required when studying drought propagation. Therefore, studies on drought propagation use the threshold level method [e.g. Peters, 2003; Peters et al., 2003, 2006; Tallaksen et al., 2009; Di Domenico et al., 2010; Vidal et al., 2010]. Another advantage of the threshold level method is that it stays as close to the original time series as possible. It does not need to fit a distribution to the data (which may lead to large discrepancies, as mentioned by Vidal et al. [2010]), or use water balance computations and calibration (which greatly increases uncertainty in drought estimation, as pointed out by Seneviratne [2012]). The physical meaning of the threshold level method is very clear, which makes interpretation of the results easier. A disadvantage of the threshold level method is that no standard drought classes are calculated, so that in global drought studies standardisation is needed to prevent large differences between climate types and to enable comparison [Wanders et al., 2010]. An additional disadvantage of the threshold level method (and other drought analysis methods) for global analysis occurs in extremely dry areas with ephemeral rivers. This is due to long periods with almost no precipitation and natural zero flow, resulting in a threshold level of zero [Scanlon et al., 2006]. In arid climates, the use of a zero-day or zero-month approach (comparable to the CDD method) is more appropriate than the threshold level method. Van Huijgevoort et al. [2012b] therefore developed a new method for the characterisation of streamflow drought on large scales based on a combination of the threshold level method and the CDD method.

Studies comparing various indices found that the threshold level method performs best compared to other indices, especially for meteorological and hydrological drought [Keyantash and Dracup, 2002; Wanders et al., 2010].

2.4.2 Threshold level method

In this section, I provide details of the calculation procedures used in this thesis.

2.4.2.1 Calculation procedure

When one uses the threshold level method [Yevjevich, 1967; Hisdal et al., 2004], a drought occurs when the variable of interest (i.e. precipitation, soil moisture, groundwater storage, or discharge) is below a predefined threshold (τ). A drought event starts when the variable (x) falls below the threshold level (onset; $t = 1$) and the event continues until the threshold is exceeded again (recovery; $t = T$). Each drought event (i) can be characterised by its duration (Δ) and by some measure of the severity of the event.

The duration of a drought event is calculated with Eqs. 2.1 and 2.2.

$$\delta(t) = \begin{cases} 1 & \text{if } x(t) < \tau(t) \\ 0 & \text{if } x(t) \geq \tau(t) \end{cases} \tag{2.1}$$

in which $\delta(t)$ is a binary variable indicating a drought situation on time t, $x(t)$ is the hydrometeo-

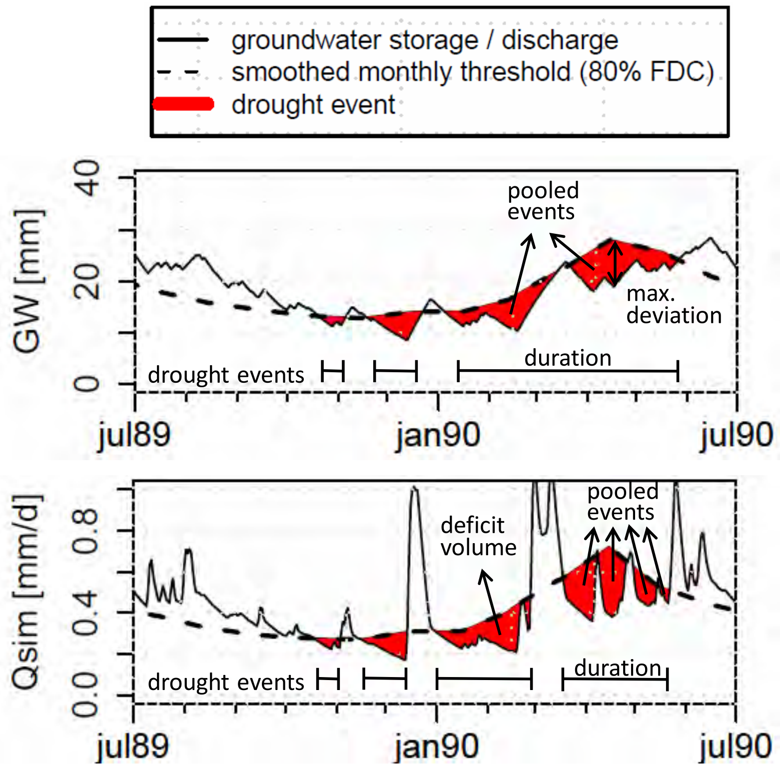

Figure 2.7: Threshold level method with variable threshold for groundwater storage (upper row) and discharge (lower row), including an illustration of pooling method and drought characteristics duration, deficit volume, and maximum deviation.

rological variable on time t, and $\tau(t)$ is the threshold level of that hydrometeorological variable on time t (t is measured in discrete time steps).

$$\Delta_i = \sum_{t=1}^{T} \delta(t) \cdot \Delta t \tag{2.2}$$

in which Δ_i is the duration of drought event i, i.e. the period for which $x(t) < \tau(t)$, $t = 1$ is the beginning of drought event i, T is the end of drought event i, and Δt is the time step of t (in this thesis: 1 day).

For fluxes (i.e. precipitation and discharge, see Sect. 2.3.2 on page 29) the most commonly used severity measure is deficit volume (D), calculated by summing up the differences between the actual flux and the threshold level over the drought period (Fig. 2.7; Hisdal et al. [2004] and Fleig et al. [2006]). Eqs. 2.3 and 2.4 show the procedure.

$$d(t) = \begin{cases} \tau(t) - x(t) & \text{if } x(t) < \tau(t) \\ 0 & \text{if } x(t) \geq \tau(t) \end{cases} \tag{2.3}$$

in which $d(t)$ is the deviation from the threshold (τ) on time t (in mm d^{-1}).

$$D_i = \sum_{t=1}^{T} d(t) \cdot \Delta t \tag{2.4}$$

in which D_i is the deficit (volume) of drought event i (in mm).

In Ch. 7 this deficit is standardised by dividing by the mean of the hydrometeorological variable.

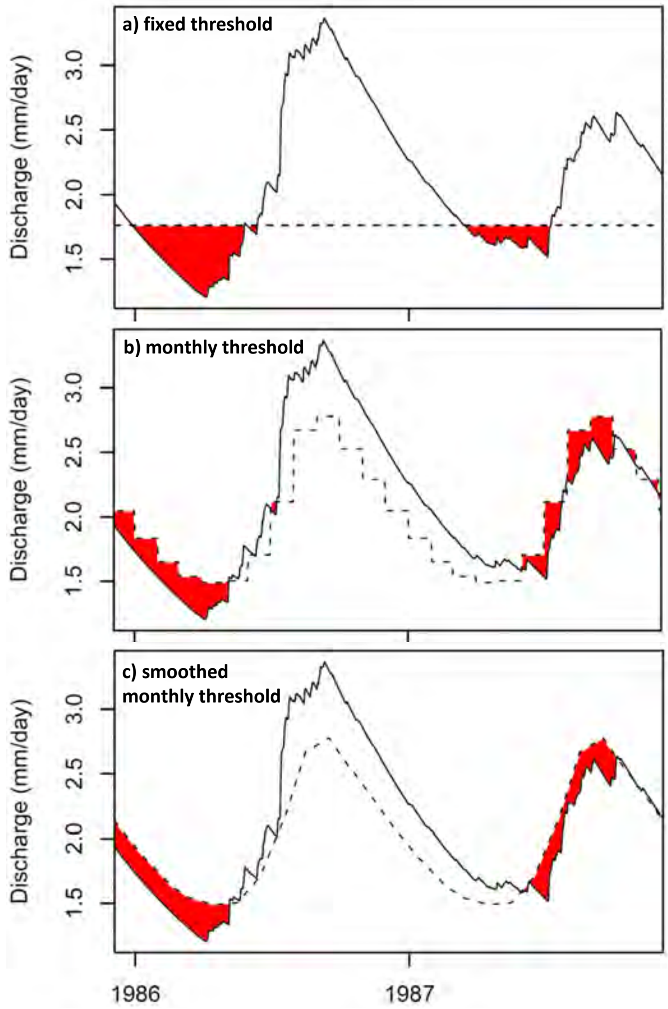

Figure 2.8: Comparison of fixed (a), monthly (b), and smoothed monthly (c) threshold level (adapted from Wanders et al. [2010]).

$$Ds_i = D_i \; / \; \overline{x(t)} \tag{2.5}$$

in which Ds_i is the standardised deficit (volume) of drought event i (in d). The physical interpretation of standardised deficit is the number of days with mean flow required to reduce the deficit volume to zero.

For state variables (i.e. soil moisture and groundwater storage, see Sect. 2.3.2 on page 29) I used the maximum deviation from the threshold (d_{max}) as the severity measure (Fig. 2.7), because the deficit volume of state variables is physically meaningless [Tallaksen et al., 2009]. This d_{max} for each drought event is calculated with Eq. 2.6.

$$d_{i,max} = \max\left(d_1(t), ..., d_T(t)\right) \tag{2.6}$$

in which $d_{i,max}$ is the maximum deviation from the threshold ($d(t)$; from Eq. 2.3) of drought event i (in mm).

2.4.2.2 Threshold level

Selection of a threshold level is crucial [Mishra and Singh, 2010]. Ideally the threshold level should be related to drought impacted sectors, e.g. irrigation water requirements, cooling water for industry, drinking water supply, reservoir operation levels, minimum water depth for

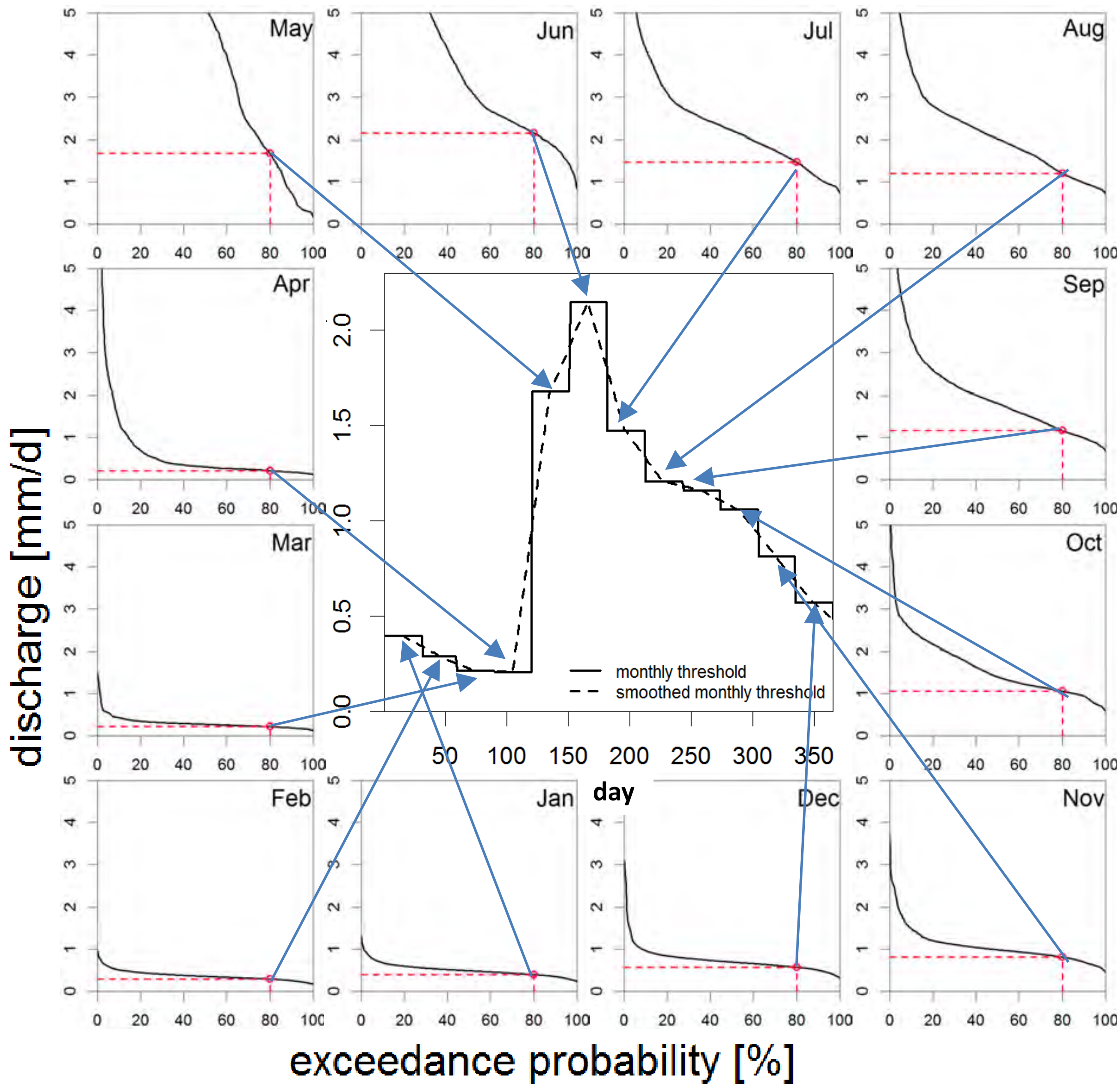

Figure 2.9: Derivation of smoothed monthly threshold level from monthly duration curves (example with discharge data, similar procedure for other variables).

navigation, or environmental flows to support stream ecology [Nathan and McMahon, 1990; Hisdal et al., 2004; Fleig et al., 2006; Sheffield and Wood, 2011; Laaha et al., 2013]. Either a fixed or a variable (seasonal, monthly or daily) threshold can be used (Fig. 2.8). 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 the definition of drought in Sect. 1.2.1). 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]. A variable threshold level was also used by e.g. Stahl [2001], Nyabeze [2004], Hirabayashi et al. [2008], Vidal et al. [2010], Hannaford et al. [2011], Prudhomme et al. [2011], Van Huijgevoort et al. [2012a], Parry et al. [2012].

I applied a monthly threshold derived from the 80th percentile of the monthly duration curves. This implies that for each month a value of a flux or state variable is chosen that is exceeded 80 % of the time in a specific month (see Fig. 2.9). The chosen 80th percentile lies within the range of 70th–95th percentile commonly used in drought studies for perennial rivers [e.g. Hisdal et al., 2001, 2004; Andreadis et al., 2005; Fleig et al., 2006; Tallaksen et al., 2009; Wong et al., 2011]. The choice of a different percentile in the calculation of the threshold level changes the magnitude of drought characteristics. For example, with a 95th percentile threshold fewer events with shorter durations and lower deficit volumes and maximum deviations are identified, whereas with a 70th percentile threshold the opposite is true. However, the relation between the drought characteristics of various hydrometeorological variables or catchments is not expected to change. This is shown by, amongst others, Woo and Tariiule [1994], Tate and Freeman [2000], and Oosterwijk et al. [2009].

As time series of precipitation contain many zero values, the 80th percentile of precipitation for most months would be zero. An option is to use a lower percentile to calculate the threshold for precipitation. However, this complicates the comparison with droughts in other variables. In this research I have, therefore, chosen to transform the precipitation time series by applying a moving average, in Chs. 4 to 6 of 30 days and in Ch. 7 of 10 days.

For all catchments except Upper-Guadiana the entire observation period (Table 2.1) was used for the calculation of the threshold. For the Upper-Guadiana catchment the threshold values were calculated based on the period 1960–1980 and applied to the entire time series in order to eliminate the strong human impact after 1980 (see Sect. 2.1.5).

The discrete monthly threshold values were smoothed by applying a centred moving average of 30 days to prevent a 'staircase' pattern and, consequently, unrealistic drought characteristics (Fig. 2.8).

2.4.2.3 Pooling, minor droughts, and average drought characteristics

In some parts of this thesis, mutually-dependent droughts were pooled using two different methods, i.e the inter-event time method in Ch. 5 and a 30-day moving average in Ch. 6 [Fleig et al., 2006]. In Ch. 5, an 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 inter-event time period is quite a subjective parameter. Tallaksen et al. [1997] and Fleig et al. [2006] tested a number of inter-event time options for a representative sample of catchments around the world (taken from a global dataset, Rees et al. [2004]). They concluded that the sensitivity curves generally started to level off around 5 days and that, for most streams, the deficit characteristics did not change substantially after 10 to 15 days, implying that a maximum of pooling was obtained. Other studies used an inter-event time period of 2 days [Engeland et al., 2004], 6 days [Tate and Freeman, 2000], and 30 days [Pandey et al., 2008]. I have chosen to use an inter-event time period of 10 days, which is quite a conservative number. This minimises the occurrence of dependent drought events, but should not include too long periods of high flow in a drought event. The choice of the inter-event time period is not expected to change the results regarding drought propagation.

The calculation of drought characteristics of the pooled drought events (visualised in Fig. 2.7) is done according to Zelenhasić and Salvai [1987], excluding the excess periods, with Eqs. 2.7, 2.8, and 2.9.

$$\Delta_p = \sum_{i=1}^{I_p} \Delta_i \tag{2.7}$$

in which Δ_p is the pooled duration of drought events $i = 1$ to I_p (in d).

$$D_p = \sum_{i=1}^{I_p} D_i \tag{2.8}$$

in which D_p is the pooled deficit of drought events $i = 1$ to I_p (in mm).

$$d_{p,max} = \max\left(d_{1,max}, ..., d_{I_p,max}\right) \tag{2.9}$$

in which $d_{p,max}$ is the maximum deviation ($d_{i,max}$) of the pooled drought events $i = 1$ to I_p (in mm).

To eliminate minor droughts, all drought events with a duration of less than 3 days (Chs. 4 and 6) and 15 days (Chs. 3 and 5) were excluded from the analysis (values up to 5 days are used by Hisdal et al., 2004; Birkel, 2005; Fleig et al., 2006, but various studies showed that minor droughts can have durations of up to 20 days; Hisdal, 2002; Fleig et al., 2005; Kaznowska and Banasik, 2011; Kim et al., 2011). In Ch. 7, no minimum duration was used.

Finally, the total number of drought events (n) and average drought characteristics were calculated for each variable and catchment or grid cell with Eqs. 2.11, 2.12, 2.13, and 2.14.

The total number of drought events was calculated as follows:

$$\alpha = \begin{cases} 1 & \text{if } \Delta_i > 0 \\ 0 & \text{if } \Delta_i = 0 \end{cases} \tag{2.10}$$

$$n = \sum_{i=1}^{I} \alpha\{\Delta_i > 0\} \tag{2.11}$$

in which n is the total number of drought events in a time series (where the drought events range from $i = 1$ to I) and α is an indicator function. In Chs. 4, 5, and 6, n is divided by the total length of the time series in years in order to obtain the number of drought events per year, which enables comparison between catchments with different length of the time series.

The average drought characteristics were calculated as follows:

$$\overline{\Delta} = \sum_{i=1}^{I} \Delta_i \;/\; n \tag{2.12}$$

in which $\overline{\Delta}$ is the average duration of all drought events in a time series. In Ch. 5, Δ_i is replaced by the pooled drought duration Δ_p (see Eq. 2.7).

$$\overline{D} = \sum_{i=1}^{I} D_i \;/\; n \tag{2.13}$$

in which $\overline{D}$ is the average deficit of all drought events in a time series. In Ch. 5, D_i is replaced by the pooled drought deficit D_p (see Eq. 2.8).

$$\overline{d_{max}} = \sum_{i=1}^{I} d_{i,max} \;/\; n \tag{2.14}$$

in which $\overline{d_{max}}$ is the average maximum deviation of all drought events in a time series. In Ch. 5, $d_{i,max}$ is replaced by the pooled drought deficit $d_{p,max}$ (see Eq. 2.9).

Chapter 3

Separating drought from water scarcity

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3.1 Introduction

Human activities impact our environment, both directly and indirectly (Sect. 1.2.1 and Wagener et al. [2010]). Indirect influence on water resources is, for example, related to climate change associated with greenhouse gas emissions. There is some confidence that in many regions around the world climate change will cause an increase in drought occurrence and severity in the 21st century [Kundzewicz et al., 2002; Bates et al., 2008; Hirabayashi et al., 2008; Mpelasoka et al., 2008; Feyen and Dankers, 2009; Dai, 2011; Kirono et al., 2011; Stahl et al., 2011a, 2012b; Seneviratne et al., 2012].

Direct human influence on water resources is also subject to change. In the future, water scarcity (the unsustainable use of water resources) likely will increase due to population growth and higher standard of living [Thomsen, 1993; Alcamo et al., 2003; Lehner et al., 2006; De Marsily, 2008; Rosegrant et al., 2009; WWDR, 2009; UNEP, 2011], and impacts of drought will become more severe in a number of regions [Kundzewicz et al., 2008; Krysanova et al., 2008; Palmer et al., 2008; Sheffield, 2008; Watts et al., 2012; Seneviratne et al., 2012].

This increasing risk of both water scarcity and drought was also notified during a meeting of the European Commission (EC) Expert Group on Water Scarcity and Drought in Venice (October 2011; www.isprambiente.gov.it/) and repeated during recent meetings (September 2012 in Athens and December 2012 in Bratislava). This group calls for an urgent debate on adapting land and water management in Europe (and elsewhere). The proposed debate requires a distinction to be made between drought and water scarcity [EU, 2012a], which is not trivial because of their interwoven nature. The terms 'water scarcity' and 'drought' are often bracketed together and used interchangeably, although they refer to quite different phenomena.

For 'water scarcity', various definitions exist. Here, we mention two internationally used definitions.

> According to the European Union [EU, 2007]: Water scarcity is defined as a situation where insufficient water resources are available to satisfy long-term average requirements. It refers to long-term water imbalances, where the availability is low compared to the demand for water, and means that water demand exceeds the water resources exploitable under sustainable conditions [EU, 2007; UNEP, 2011].

> UN-WATER [2006] and FAO [2007] define water scarcity as: the point at which the aggregated impact of all users impinges on the supply or quality of water under prevailing institutional arrangements to the extent that the demand by all sectors, including the environment, cannot be satisfied fully.

Both definitions refer to the imbalance between water availability/supply and demand. In this study, we define water scarcity as the overexploitation of water resources when demand for water is higher than water availability. Thus, we focus on the effect that human activities have on the hydrological system. 'Water shortage' or 'water stress' is sometimes used as a synonym for water scarcity [Pereira et al., 2002; Palmer et al., 2008; Taylor et al., 2009; Wada et al., 2011], but in this study we avoid these terms.

The term 'drought' is also defined in many different ways. Two examples are given here. Mishra and Singh [2010] and Seneviratne et al. [2012] provide a wider overview and distinguish between different user perspectives.

> According to EU [2007]: Droughts represent relevant temporary decreases of the average water availability, refer to important deviations from the average levels of natural water availability and are considered natural phenomena.

> According to Tallaksen and Van Lanen [2004]: Drought is defined as a sustained and spatially-extensive period of below-average natural water availability [which is more or less along the line as earlier defined by WMO, 1986].

Table 3.1: Schematic concepts for terms related to conditions of low water availability (adapted from Pereira et al. [2002])

	long time scales	short time scales
natural	ARIDITY	DROUGHT
anthropogenic	DESERTIFICATION	WATER SCARCITY

These and other conceptual definitions refer to drought being a natural hazard, i.e. caused by natural processes (climate, hydrology). Here, we follow the definition of Tallaksen and Van Lanen [2004], defining drought as a period of below-normal water availability with natural causes. We consider all water stores and fluxes in a catchment, e.g. natural storage in groundwater and wetlands and artificial storage in reservoirs. Related terms are 'aridity' and 'desertification', which refer to more permanent phenomena (Table 3.1). While drought is a consequence of climate variability (short time scales), aridity is related to the average climate (long time scales). Desertification is the longer-term equivalent of water scarcity, as it has anthropogenic causes (Sect. 1.2.1).

Unfortunately, the term 'drought' is regularly used for dry situations in which anthropogenic influence plays a significant role, e.g. FAO [2007], Lopez-Moreno et al. [2009], Taylor et al. [2009], WWDR [2009], Sheffield and Wood [2011]. Or, the other way around, the term 'water scarcity' is sometimes used for a dry situation with natural causes, e.g. WMO [2005], EEA [2012]. In most studies, the terms are defined in the right way, but in the application they are used differently.

Mixing up of the terms 'water scarcity' and 'drought' can be misleading in water management and should be avoided [Pereira et al., 2002], as there is a fundamental difference in how management can influence these phenomena. Management can combat overexploitation of water resources (water scarcity), whereas it only can adapt to climate variability (drought) by reducing vulnerability and increasing resilience through implementing pro-active measures [Wilhite, 2002; EU, 2007; Estrela and Vargas, 2012; Kossida et al., 2012; Mortazavi et al., 2012]. Palmer et al. [2008] and Schiermeier [2008] advocate a shift in the focus of water management in water-scarce regions from reducing the vulnerability to drought, to reducing the overexploitation of water resources, so from coping with natural variability to reducing anthropogenic effects.

Water management is traditionally supported by decision support tools [Andreu et al., 1996; Flug and Campbell, 2005; Mysiak et al., 2005; Carbone et al., 2008], which use indicators [Smakhtin and Hughes, 2004; Niemeyer, 2008; Wanders et al., 2010; Jaranilla-Sanchez et al., 2011] to monitor, forecast and predict. These indicators focus either on drought or on water scarcity. Decision support tools that use these indicators try to analyse drought and water scarcity separately, but what if both occur simultaneously? We often experience a combination of both phenomena: in water-scarce regions, the impact of drought is more severe [Wilhite, 2002] and in a drought situation, the management of water supply is even more crucial [Kundzewicz et al., 2002]. In these situations, making a clear distinction between water scarcity and drought is very much needed. But how to distinguish between the interlinked phenomena of water scarcity and drought?

Finding a distinction between natural and human influence is not trivial, as is proved in hydrology [Kassas, 1987; Panda et al., 2007; Apaydin, 2010; Hoogland et al., 2010; Kauffman and Vonck, 2011], and in other fields like, for example, vegetation science and land degradation [Pechmann et al., 1991; Evans and Geerken, 2004; Wessels et al., 2007; De Beurs et al., 2009]. A (water) manager needs to have information on the situation that would have occurred without human influence, the so-called 'naturalised situation'. This information needs to be quantitative, so that the manager can adequately assess past effects and future scenarios. And because both

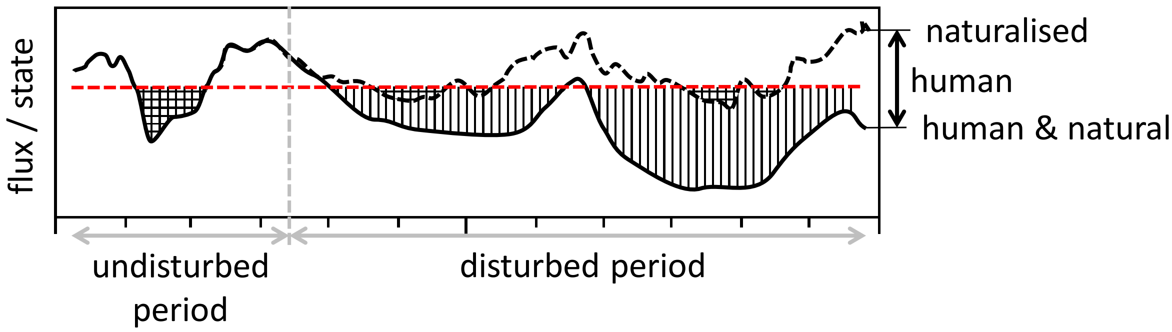

Figure 3.1: Conceptual figure of a system that is affected by both natural and anthropogenic influences (solid line), of which the 'naturalised situation' (dashed line) needs to be known to quantify the anthropogenic effect. The red horizontal line indicates a threshold below which the system is negatively impacted. The gridded surfaces indicate anomalies with natural causes (drought) and the vertically-striped surfaces indicate anomalies with human (water scarcity) causes.

the natural and the human impacts show temporal variability (e.g. seasonal), this information also needs to be on a transient basis, as time series of state variables and fluxes characterising the water system. This is depicted in the conceptual graph in Fig. 3.1, in which the solid line represents the observed situation and the dashed line represents the naturalised situation, which is unknown. In the first part of the graph, where the dashed line still coincides with the solid line, human influence on the system is negligible. This period is called the 'undisturbed period'. The period in which the lines deviate is called the 'disturbed period'. In the 'disturbed period', the difference between the dashed and the solid lines represents the human influence. These lines can represent hydrological variables (states, fluxes), such as soil moisture, groundwater storage, streamflow. The red line in Fig. 3.1 can be regarded as the average normal situation or a fixed threshold below which the system is negatively impacted, e.g. environmental minimum flow, critical reservoir level, water temperature relevant for cooling water release regulations. By using such a threshold, quantitative information can be derived on the relative impact of human influence on anomalies (i.e. deviations from the threshold). In the disturbed period in Fig. 3.1, the natural situation (without disturbance) would have led to four short periods with minor effects (the gridded surfaces show minor deviations from the red line), while the disturbed system (anthropogenic, including natural) resulted in two long periods with large effects (the vertically-striped surfaces show large deviations from the red line).

Several studies have tried to find a proxy for the naturalised situation of the hydrological system. Lorenzo-Lacruz et al. [2010], for example, compared inflow and outflow of reservoirs. Inflow stands for the climatic signal, outflow stands for the climatic signal minus the anthropogenic signal. Barco et al. [2010] compared observed groundwater data (natural + anthropogenic) with the ENSO signal (natural) to filter out anthropogenic effects. Mair and Fares [2010] compared the trend in measured streamflow (natural + anthropogenic) to the trend in precipitation (natural). These studies are very valuable, but they do not take into account the intrinsic non-linearity of the system. For example, reservoir processes were not included in the study of Lorenzo-Lacruz et al. [2010], and Barco et al. [2010] and Mair and Fares [2010] did not take into account the transformation of the climate signal in the subsurface. Additional non-linearity is introduced by the fact that more abstraction takes place in dry years [Custodio, 2002] and by changes in feedback between artificially lowered water levels and the atmosphere (evaporation and precipitation), which in turn have consequences for human interventions.

Due to this non-linearity (also visible in Fig. 3.1), quantification of the naturalised situation is not a simple modification of the observed signal using known human influences (i.e. subtraction or addition of a constant value); some sort of manipulation of the data (i.e. system modelling) is always needed. Even in the ideal situation, with perfect observed data of all anthropogenic

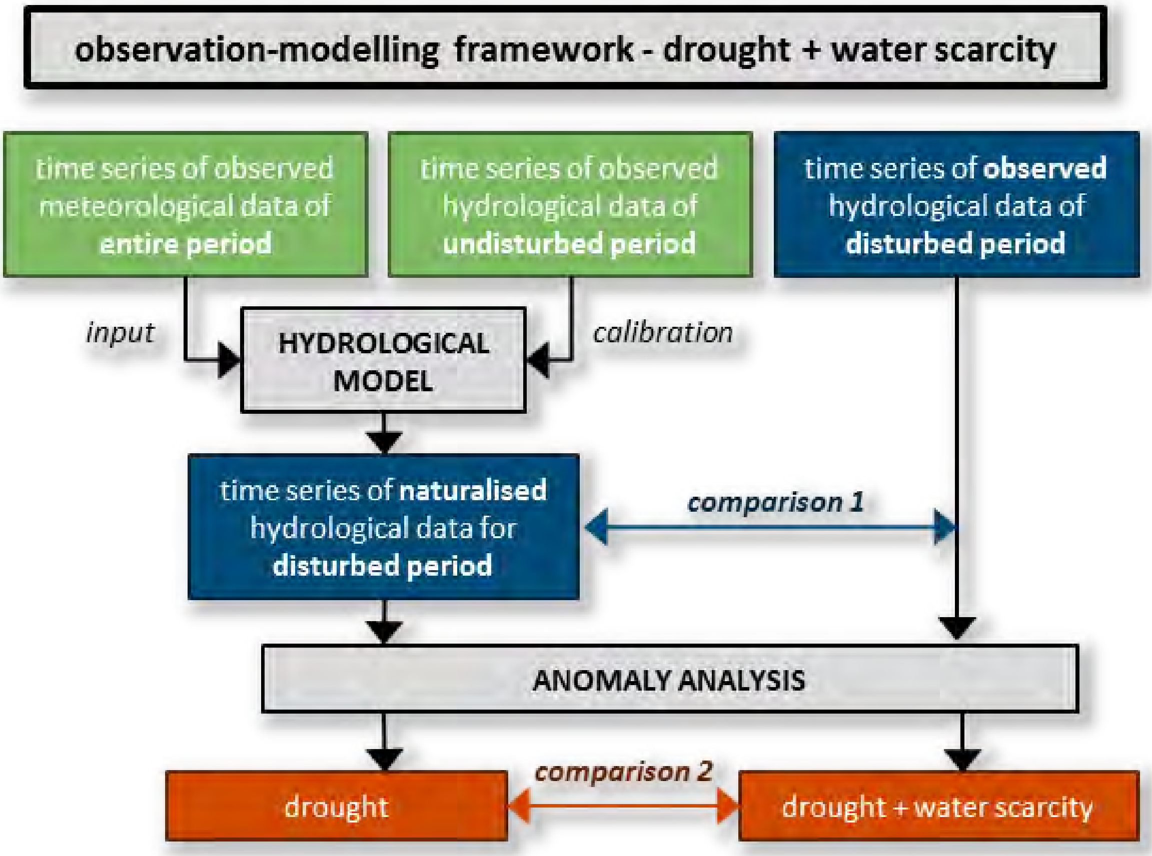

Figure 3.2: Observation-modelling framework, as proposed to distinguish drought and water scarcity.

effects on water resources (e.g. abstractions, reservoir releases, irrigation and sewage water return flows), some kind of system model is needed to obtain time series of naturalised state variables and fluxes (e.g. groundwater heads, river flow).

This chapter provides a methodology to use both observations and modelling to simulate the naturalised situation in order to separate drought and water scarcity, complying with the urgent call of the EC Expert Group on Water Scarcity and Drought. The proposed observation-modelling framework is illustrated by the application of a conceptual hydrological model in a case study area in Spain. The framework is generic by nature and can be applied to other regions and with other models. Its novelty lies in the comparison of anomalies in hydrological variables between the disturbed situation and the naturalised situation. The case study area in Spain is chosen because of its semi-arid climate and frequent (multi-year) droughts combined with vast groundwater abstraction for large-scale irrigated agriculture. This combination results in a high risk of water scarcity in the area with potentially large economic consequences [Gil et al., 2011; Estrela and Vargas, 2012]. The decreasing trend in observed streamflow and increasing trend of drought deficit volume in large parts of Spain [Hisdal et al., 2001; Stahl et al., 2012b; Lorenzo-Lacruz et al., 2012] and projections of a drier climate and a higher drought frequency in Spain in the future [Lehner et al., 2006] make water management an increasingly pressing issue in this area.

The outline of this chapter is as follows. First, in Sect. 3.2 the observation-modelling framework is explained. In Sect. 3.3 the case study area in Spain is introduced. In Sect. 3.4, the application of this framework to the case study is presented and discussed. Finally, the research is discussed and concluded in Sects. 3.5 and 3.6.

3.2 Observation-modelling framework

The observation-modelling framework that we propose as a tool to make the distinction between drought and water scarcity is depicted in Fig. 3.2.

The basic requirements of this framework are the availability of observations of hydrometeorological variables for both the undisturbed and the disturbed period. These observations serve as input and calibration data for the hydrological model, which is the central point of the framework. The model simulates the naturalised situation of the hydrological system based on meteorological forcing as input and hydrological data of the undisturbed period for calibration (Fig. 3.2). Which variables are required depends on the choice of the hydrological model. The meteorological forcing data contain at least the variables temperature, precipitation, and reference evaporation, either from local measurement station(s) or from large-scale forcing dataset(s). The hydrological variables needed for calibration can be discharge and/or groundwater level.

Various hydrological model types can be chosen as model in the framework, e.g. a distributed or lumped model, a physically-based model, a conceptual model, or even a stochastic model [Beven, 2000; Wagener et al., 2004], as long as it is capable of reproducing the natural situation, especially during low flow and drought. This is not straightforward, as is shown by Smakhtin [2001] and Staudinger et al. [2011] and in Sects. 2.3.3 and 2.3.4. But if the model and the calibration method are chosen well, the model can be used to simulate the situation that would have occurred without human influences in the disturbed period (represented by the dashed line in Fig. 3.1).

Calibration against hydrological data can be parsimonious, depending on the natural variability of the catchment under consideration, so that only a minimal amount of observed hydrological data of the undisturbed period is needed [Seibert and Beven, 2009]. The use of separate calibration and validation periods is recommended, as model performance outside the calibration range can then be evaluated.

The naturalised time series of discharge and/or groundwater can then be compared to the observed time series of discharge and/or groundwater (represented by the solid line in Fig. 3.1). This so-called ‘comparison 1’ (Fig. 3.2) gives a visualisation and quantification of the time-varying human influence on the flux/ state variable considered. If uncertainties in observed and simulated time series are high, multiple time series should be provided representing the sensitivity ranges of both variables. The comparison of these ranges then gives an indication whether the difference between natural and human influences (‘signal’) is larger than the sensitivity of the observations and simulations (‘noise’).

Even more important in the framework for separating drought and water scarcity is the anomaly analysis (Fig. 3.2). This analysis extracts anomalies from time series of (observed or naturalised) hydrological variables, both state variables and fluxes. In this way we can investigate deviations from normal conditions (represented by the red line in Fig. 3.1). In the undisturbed period, anomaly analysis on both observed and simulated time series gives drought events. In the disturbed period, anomaly analysis on simulated (= naturalised) time series gives drought events (represented by the gridded surfaces in Fig. 3.1), whereas anomaly analysis on observed time series gives the combined effect of drought and water scarcity (represented by the vertically-striped surfaces in Fig. 3.1). By comparing these events (‘comparison 2’ in Fig. 3.2), the relative contribution of human and natural effects on anomalies can be visualised and quantified.

Just as different hydrological models can be used, the specific method used for anomaly analysis within the observation-modelling framework can vary as well. Anomaly analysis methods are usually called drought analysis methods (Sect. 2.4), but the term ‘drought’ is reserved for anomalies with natural causes. As we also study anomalies with human causes in this chapter, we use the more general term ‘anomaly analysis’. Some possibilities for anomaly analysis methods are the threshold level method, the Sequent Peak Algorithm [SPA, Hisdal et al., 2004; Fleig et al., 2006]. Even hydrological (streamflow/groundwater) drought indicators [Niemeyer et al., 2008; Mishra and Singh, 2010; Wanders et al., 2010] can be used, as long as they are transient and based on deviation from normal conditions like, for example, the standardised indices of Vidal et al. [2010].

The application of this theoretical framework to a specific case study is demonstrated in Sect. 3.4.

3.3 Case study

The Upper-Guadiana catchment in Spain was used as an example in this chapter. The Upper-Guadiana has been studied intensively in a number of EU-funded projects, e.g. EFEDA, GRAPES, ARIDE, NeWater, WATCH, MEDIATION. Consequently, background information about the catchment is widely available. Here we give a summary of the most important characteristics of the area.

3.3.1 Catchment characteristics

For a description of the Upper-Guadiana catchment see Sect. 2.1.5.

3.3.2 Hydrological drought

In the first half of the 1980s and of the 1990s, severe multi-year drought events in precipitation (meteorological droughts) have occurred in the Upper-Guadiana catchment (Fig. 3.3). In the period 1960–1980, annual average precipitation was 483 mm, while in the next decades (1980–2000) annual average precipitation had decreased to 403 mm. The presence of aquifer systems and wetlands influences the propagation of a meteorological drought into a hydrological drought [Peters et al., 2005; Van Lanen et al., 2004a]. Hydrological drought events in the Upper-Guadiana catchment are generally very long. Drought characteristics of discharge are more comparable to those of groundwater storage than to those of soil moisture, which is a sign of the strong coupling between groundwater and discharge and the slow response to precipitation (see also Ch. 5). Due to the large storage capacity in the aquifer system and wetlands in Upper-Guadiana (Sect. 2.1.5), meteorological droughts are also often attenuated after a period of high precipitation and do not develop into a hydrological drought.

3.3.3 Land use and human influence

Land use in the Upper-Guadiana catchment is mainly agricultural. Main crops are grapes (vineyards) and cereals, both in rainfed and irrigated agriculture [Aldaya et al., 2010]. Before the 1970s, dryland farming of cereals and other non-irrigated crops prevailed. Some irrigation from surface water and groundwater was applied, but this was done in a very traditional way using the Arabic system called 'norias', which did not exceed 300 km^2, less than 2 % of the Upper-Guadiana catchment. A lot of land remained unused and was occupied by natural vegetation [Acreman, 2000]. Abstraction was very limited in that time (always below 50 million $m^3\,yr^{-1}$) and no effect of abstraction on the natural system (e.g. groundwater levels) was observed [Veenstra, 2009].

Since 1970–1980, agriculture intensified and human influence (i.e. irrigation and artificial drainage) in the catchment increased dramatically. In the period 1974–1988, borehole discharge from La Mancha Occidental (Sect. 2.1.5) increased from 200 to 688 million $m^3\,yr^{-1}$, which is about 90 % of the abstracted volume in the whole Upper-Guadiana catchment (Fig. 3.3). Currently, a significant proportion of the agricultural area is irrigated [38 %; Aldaya et al., 2010],

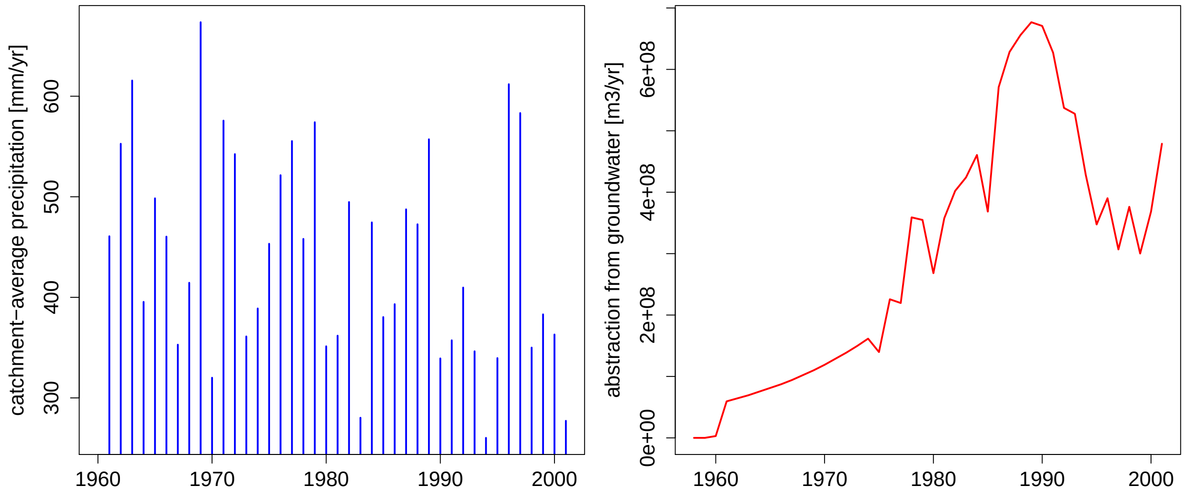

Figure 3.3: Catchment-average precipitation (precipitation data provided by AEMET) and abstractions in the Upper-Guadiana catchment (abstraction data provided by CSIC, based on information from the Guadiana Water Authority (GWA), from the Castilla-La Mancha Local Government, from Martínez Cortina [2001], and from an estimation based on piezometric data fitting by the University of Castilla-La Mancha (UCLM); abstraction figure comparable to Bromley et al. [2001]).

accounting for approximately 95 % of the total consumptive groundwater use [Carrasco, 2002; CHG, 2006]. However, there is a lack of reliable information concerning borehole abstractions, due to private and therefore uncontrolled groundwater development [Bromley et al., 2001]. Less than half of the approximately 40,000 wells in the La Mancha Occidental area are legally registered [Martínez-Santos et al., 2008].

In recent decades, groundwater levels showed a significant decline (Fig. 3.4). In the La Mancha Occidental aquifer (Sect. 2.1.5), the average regional drawdown of the water table was 22 m, with a maximum of over 50 m [Bromley et al., 2001; Martínez-Santos et al., 2008]. This drawdown resulted in a total disconnection of groundwater and surface water by the early 1980s [Martínez-Santos and Martínez-Alfaro, 2010] and therefore declining wetland area, decreasing discharge, changes in water quality, decreasing evaporation, and spontaneous combustion of peatlands [Martínez-Santos et al., 2008; Veenstra, 2009; Varela-Ortega et al., 2011]. The La Mancha Occidental aquifer was officially declared overexploited in 1994 (Cobelas et al. [1996] in Alvarez-Cobelas et al. [2001]), which gave rise to various social pressures [Aldaya et al., 2010]. Although the intensification of agriculture and the increase in groundwater abstraction in Upper-Guadina started around 1970, the effect on the hydrological system only became clearly visible after 1980 (pers. comm. Vicente Navarro and Miguel Candel; UCLM, Spain). An indicator for the effects of both irrigation abstraction and multi-year drought (see previous section) on the hydrological system is the disconnection of groundwater and surface water around 1983, which is mentioned in Fig. 3.4 (1983: springs of Guadiana dry). The hydrological situation before 1980 is regarded as 'undisturbed', meaning that streamflow and aquifer levels, even in the presence of minor human influences, resemble the behaviour of the natural system. Therefore, in the remainder of this chapter, we refer to the period before 1980 as the 'undisturbed period', and to the period afterwards as the 'disturbed period'.

The human influence on the observed water table drawdown in Upper-Guadiana has been put forward in several papers, e.g. Cruces de Abia and Martínez Cortina [2000], Bromley et al. [2001], Alvarez-Cobelas et al. [2001], Custodio [2002], Conan et al. [2003], Martínez-Santos et al. [2008], Zorrilla et al. [2010], and Navarro et al. [2011]. Martínez-Santos et al. [2008] gave a graphical overview of processes related to water level drawdown and measures for recovery (Fig. 3.4) and concluded that drying of wetlands and springs is caused by 'mounting water demands for irrigation, aggravated by EU-subsidies (CAP)'. In Fig. 3.4 also some natural

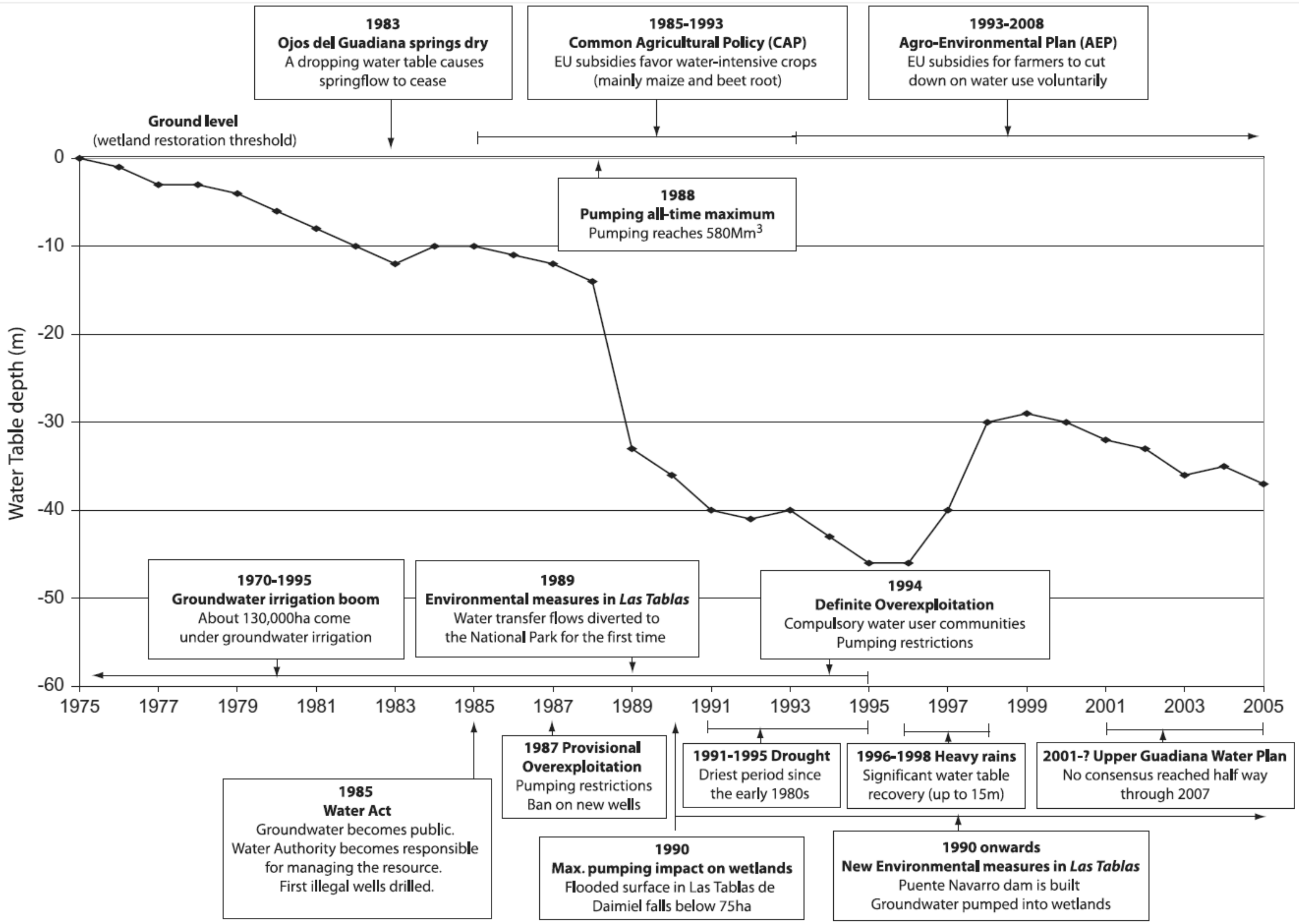

Figure 3.4: Evolution of the water table in the La Mancha Occidental aquifer in relation to water, agricultural, and environmental policies and other representative events [from Martínez-Santos et al., 2008, reprinted with permission from the publisher Wiley].

processes are included, such as a drought period in 1991–1995 and heavy rains in 1996–1998 (see also Fig. 3.3). The relative contribution of these natural processes to water table drawdown and recovery is, however, nowhere clarified in the paper of Martínez-Santos et al. [2008]. The implementation, since 1987, of policy measures to reduce pumping have not resulted in a corresponding reduction in the rate of decline of groundwater levels (Fig. 3.4). Bromley et al. [2001] mentioned a lack of recharge (i.e. drought) as one of three possible options to explain this lack of recovery, but also stated that it is difficult to be certain of the main cause. The opposite happened after 1995, when a recovery of water tables by approximately 15 m was observed (Fig. 3.4). This created optimistic claims about the applied policy, i.e. more pumping restrictions and the implementation of an Agro-Environmental Plan [Menendez, 2001]. Martínez-Santos et al. [2008], however, mentioned that the effects of the policy were masked by one of the most significant rainfall episodes of the century (Fig. 3.3). After 1999, groundwater levels again declined (Fig. 3.4). Was this the effect of natural or of human causes? No decisive answer has been given up to now.

The uncertainty in the relative contribution of natural and anthropogenic effects on groundwater level changes results in vague statements that are of little use to water managers. Some examples can be found in Bromley et al. [2001]:

- 'High abstraction rates can be sustained in future when rainfall exceeds the long-term average.'

- 'After a stop of abstractions, recovery of the water table to its natural condition will take longer in a period of below-average rainfall.'

This shows that making the distinction between drought (natural causes) and water scarcity (human causes) is, also in Upper-Guadiana, essential for water management, as is pointed out by other authors as well [Hernández-Mora et al., 2001; Conan et al., 2003].

3.4 Application of the observation-modelling framework to the Upper-Guadiana basin

In this section we demonstrate the application of the proposed observation-modelling framework (Sect. 3.2 and Fig. 3.2) to the Upper-Guadiana basin. In this example we used HBV as the hydrological model and the variable threshold level method as the anomaly analysis method.

3.4.1 Observations

The observation-modelling framework uses both meteorological and hydrological observations. The meteorological data needed as input for the hydrological model consist of temperature, precipitation, and potential evaporation. All meteorological data used in this study were available for the period 1960–2001 on a daily time scale (see Sect. 2.2 and Fig. 3.5).

The observed hydrological data of the Upper-Guadiana catchment consist of discharge and groundwater levels. Discharge was measured at the outlet of the catchment (gauging station Balbuena, station no. 4008, lat: 38.9, long: -4.1) at a specially designed measurement structure with a low-flow channel (Fig. 2.4). The gauging station is located just below the confluence with the Jabalón tributary and the discharge of the Jabalón (gauging station Puente Morena, station no. 4103, lat: 38.9, long: -4.0) is subtracted from the discharge measured at Balbuena. The gauging station indicated in Fig. 3.5 represents the fictive gauging station of the Upper-Gaudiana catchment excluding the Jabalón tributary (approximately, lat: 39.0, long: -4.0). Discharge was measured by the Guadiana Water Authority and data were available from an online database maintained by CEDEX (hercules.cedex.es/anuarioaforos) for the period 1960–2001, on a daily

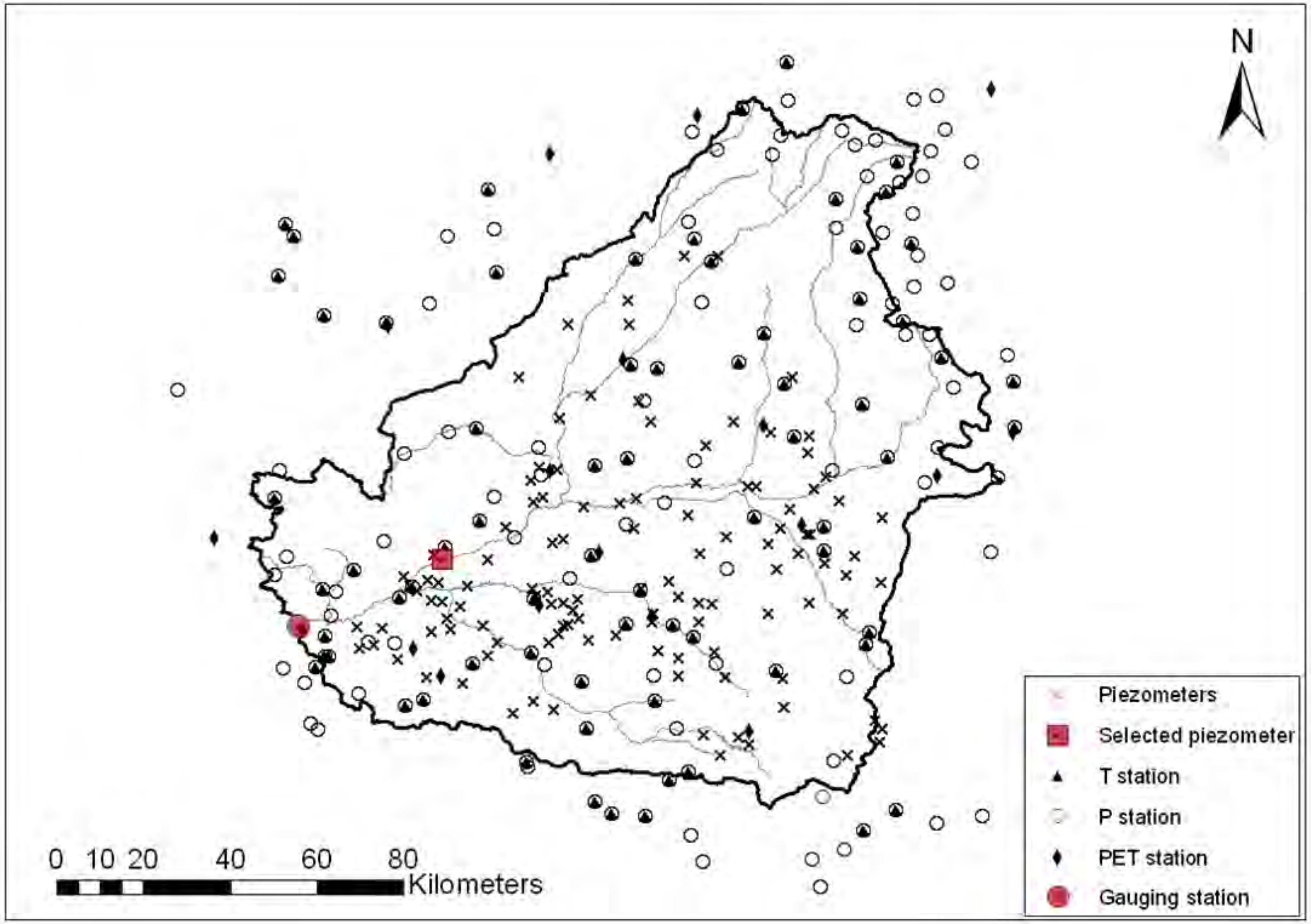

Figure 3.5: Upper-Guadiana catchment, including gauging station, piezometers, and meteorological stations (T = temperature, P = precipitation, PET = potential evapotranspiration).

basis (Sect. 2.2.2). Some gaps and 'staircase' data [Rees et al., 2004] were present in the time series, which were linearly interpolated (5% of the entire time series; Veenstra [2009]).

Groundwater levels are measured at a number of locations in the catchment (approximately 120, Fig. 3.5) by the former Geological Survey of Spain (IGME) and made available through the research project ARIDE [Demuth and Stahl, 2001]. As an example, here we used only one representative groundwater measurement station (station no. 1929-70002, lat: 39.2, long: -3.6). This station did not show the largest drawdown due to abstractions, but was chosen because it had a long and relatively continuous time series that, after calibrating the model on observed discharge, matched very well with simulated catchment average groundwater storage. The selected station was located near the wetland Tablas de Daimiel, quite close to the catchment outlet (Fig. 3.5), and measured groundwater levels in the large La Mancha Occidental aquifer system (Sects. 2.1.5 and 3.3). Measurements were taken in the period 1973–1997 on a monthly basis. Gaps in the time series were linearly interpolated.

In the observation of both discharge and groundwater uncertainties arise, which are difficult to quantify [Sweet et al., 1990; Rees et al., 2004; Di Baldassarre and Montanari, 2009]. In an attempt to make a first rough estimate of the uncertainty in the observational series, we applied a method to add noise to the time series of both groundwater and discharge. This method kept the auto-correlation of the time series intact and only added random noise of 20 % of the standard deviation of the variable. For observed discharge this was done multiplicative and for groundwater this was done additive.

The observed discharge and groundwater data of the undisturbed period (1960–1980 for discharge and 1973–1980 for groundwater) were used for calibration and validation of the hydrological model. The observed discharge and groundwater data of the disturbed period (1980–2001 for discharge and 1980–1997 for groundwater) were used for comparison with naturalised data from the hydrological model (see Fig. 3.2).

3.4.2 Hydrological model

The conceptual, semi-distributed, rainfall-runoff model HBV (see Sect. 2.3.1) was chosen as the example hydrological model in the framework (Fig. 3.2). In this study, we used the DELAY response routine, because it was found to be more suitable for simulating the hydrological regime of Upper-Guadiana than the STANDARD response routine (Fig. 2.5). The HBV model does not explicitly simulate human influence. Any human influence in the undisturbed period that impacts flow is included in the calibrated parameters.

The undisturbed period was subdivided in 10 years of data for calibration (1960–1970) and 10 years for validation (1970–1980). The calibration procedure (described in Sect. 2.3.2) was applied twenty times and the best result, based on the Nash-Sutcliffe efficiency using logQ (ln Reff) and visual inspection, was selected for further anomaly analysis.

The HBV model simulates discharge in mm d^{-1} (i.e. per unit area) and groundwater storage in mm. Simulated discharge can be used directly for comparison with observed discharge. Simulated groundwater storage first had to be converted to groundwater levels using a constant storage coefficient concept. This was done by a linear regression on observed groundwater levels for the undisturbed period, which for groundwater observations was limited to 1973–1980 (Sect. 3.4.1).

The time series of simulated and observed discharge and groundwater levels for the undisturbed period are shown in Fig. 3.6. The blue lines give the observed groundwater level and discharge and the red lines represent the selected model simulations for both variables. The ranges (semi-transparent blue and red surfaces) provided in Fig. 3.6 can be seen as a first attempt to visualise observational uncertainty (see Sect. 3.4.1) and model parametric uncertainty (see above).

Discharge had a clear seasonal variability with high values in winter and low values in sum-

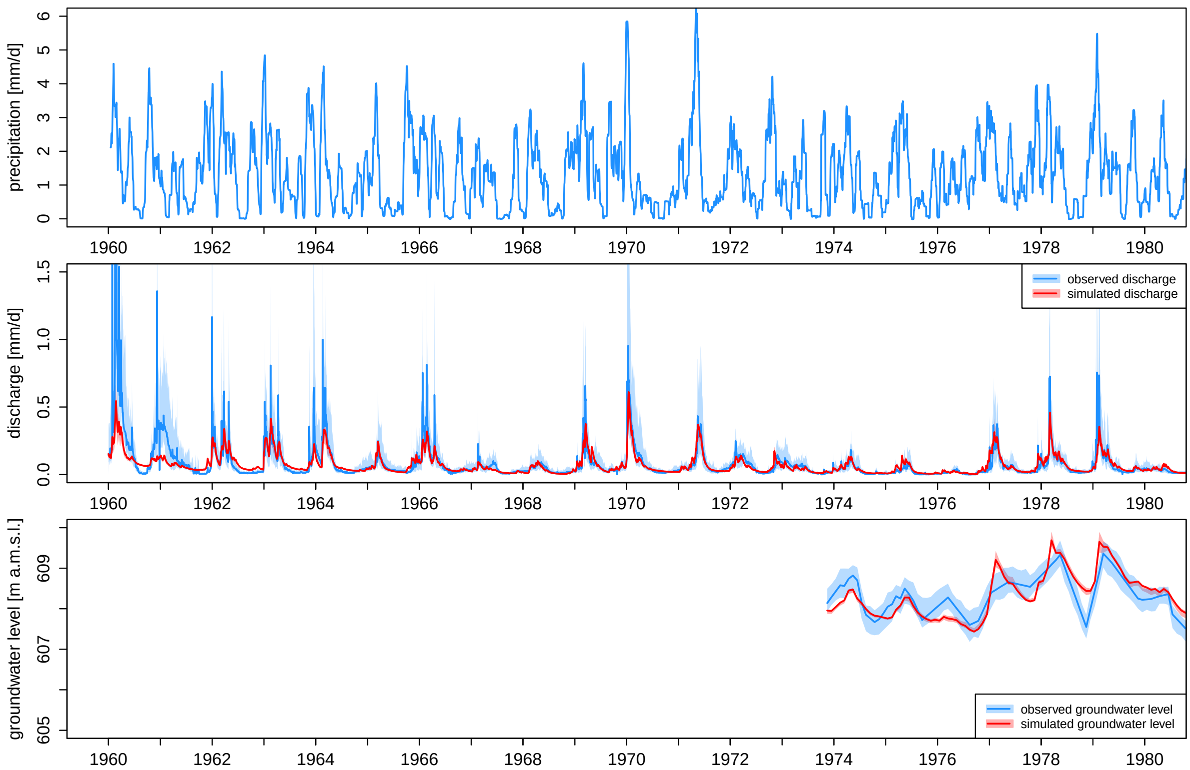

Figure 3.6: Time series of precipitation (30 day moving average), discharge, and groundwater level of Upper-Guadiana in the undisturbed period (calibration and validation period); ranges give a rough indication of observational uncertainty (20 possible time series of noised observations) and simulation uncertainty (20 calibration results).

mer that was reproduced by the model (Fig. 3.6 - middle row). The interannual variability in discharge, i.e. wet years in the beginning of the 1960s and dry years halfway the 1970s, was also simulated well. Observed discharge was slightly more peaky than simulated discharge, but simulations are within the range of 'noised' observations. The $\ln$ Reff for the calibration period was 0.64, which is quite a good result. For the validation period it was slightly lower (0.47), but still reasonable. The $\ln$ Reff for the entire undisturbed period was 0.54. A scatter plot of simulated versus observed discharge, transformed to log-scale (Fig. 3.7), shows that daily discharge values are scattered along the 1:1-line, except for the lowest low flows for which uncertainties in both observations and simulations are highest.

Simulated groundwater storage was validated using the coefficient of determination (r^2), based on comparison with observations in part of the validation period (Sect. 3.4.1). This r^2 had a value of 0.83, which is high. Two other groundwater wells that we investigated had r^2 values of 0.54 and 0.43, which is lower, but still acceptable. The remaining observation wells had a too short period of record in the undisturbed period. In groundwater, both seasonal and interannual variability were reproduced by the model (Fig. 3.6 - lower row). For example, the period with lower discharge halfway the 1970s was reflected by a lower groundwater level. However, simulated groundwater levels were not always within the range of 'noised' observations and seemed to have a slightly lower temporal variability than observed levels, which is, however, much influenced by the assumption of a constant storage coefficient and the optimal parameters chosen in the regression.

In general, the good performance of the model in the undisturbed period gave confidence in the ability of the model to simulate the naturalised hydrological situation in the disturbed period.

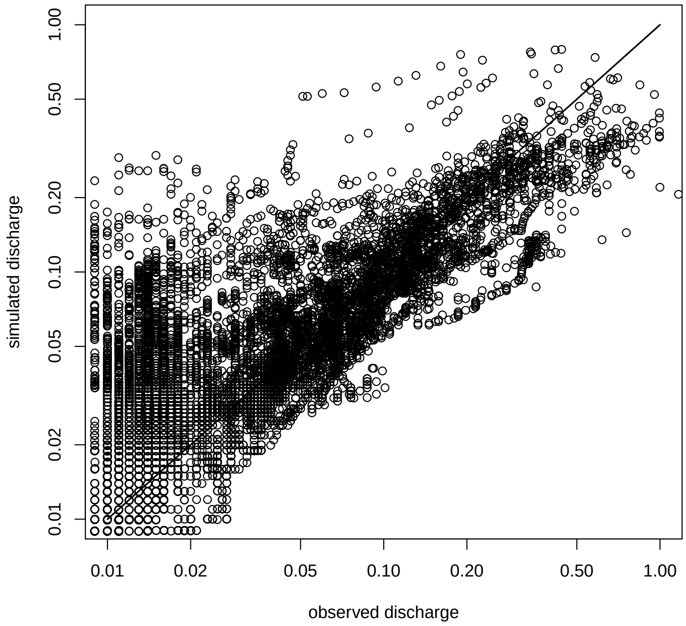

Figure 3.7: Simulated vs. observed discharge in the undisturbed period transformed to log-scale (incl. the 1:1-line).

3.4.3 Comparison 1 - Time series

As a next step, the calibrated model was run for the disturbed period (1980–2000) and used for a comparison with observed data (Sect. 3.2 and Fig. 3.2 - comparison 1). Time series of simulated and observed discharge and groundwater level in the disturbed period are shown in Fig. 3.8. The simulated time series can be regarded as the naturalised situation, so that the difference between observed and naturalised time series shows the anthropogenic effect in the catchment. Just as in Fig. 3.6, the ranges provided in Fig. 3.8 can be seen as a first attempt to include the influence of observational uncertainty and model parametric uncertainty. From Fig. 3.8 we can conclude that the difference between the naturalised and the human-influenced situation ('signal') is much larger than the imposed uncertainty of the observations and simulations ('noise').

In the beginning of the 1980s, observed discharge was still very similar to naturalised discharge (Fig. 3.8 - middle row). By the end of the 1980s, all discharge peaks that are visible in the simulations (red line) were completely absent from the observations (blue line), due to increased abstraction. In 1990, a precipitation peak resulted in a discharge peak in both simulations and observations, but in the observations this peak was very short-lived. Afterwards, in the period 1990–1996, naturalised discharge (red line) was very low and decreased to zero, whereas observed discharge (blue line) was zero during the entire period. In the naturalised time series (red line), the dry period ended in 1996, followed by three years of high discharge. Again, observed discharge also showed peaks (blue line), but these were much shorter and less high than in the naturalised time series. In 1999, after these relatively wet years, observed discharge immediately reduced to zero again.

Although observed groundwater levels did not span the entire disturbed period, a similar pattern as for discharge was observed when comparing naturalised and observed groundwater levels (Fig. 3.8 - lower row). Observed groundwater levels in the period until 1990 (blue line) were below naturalised levels (red line) and showed a decreasing trend. In 1990, observed groundwater levels showed a peak, which made observed and naturalised groundwater levels

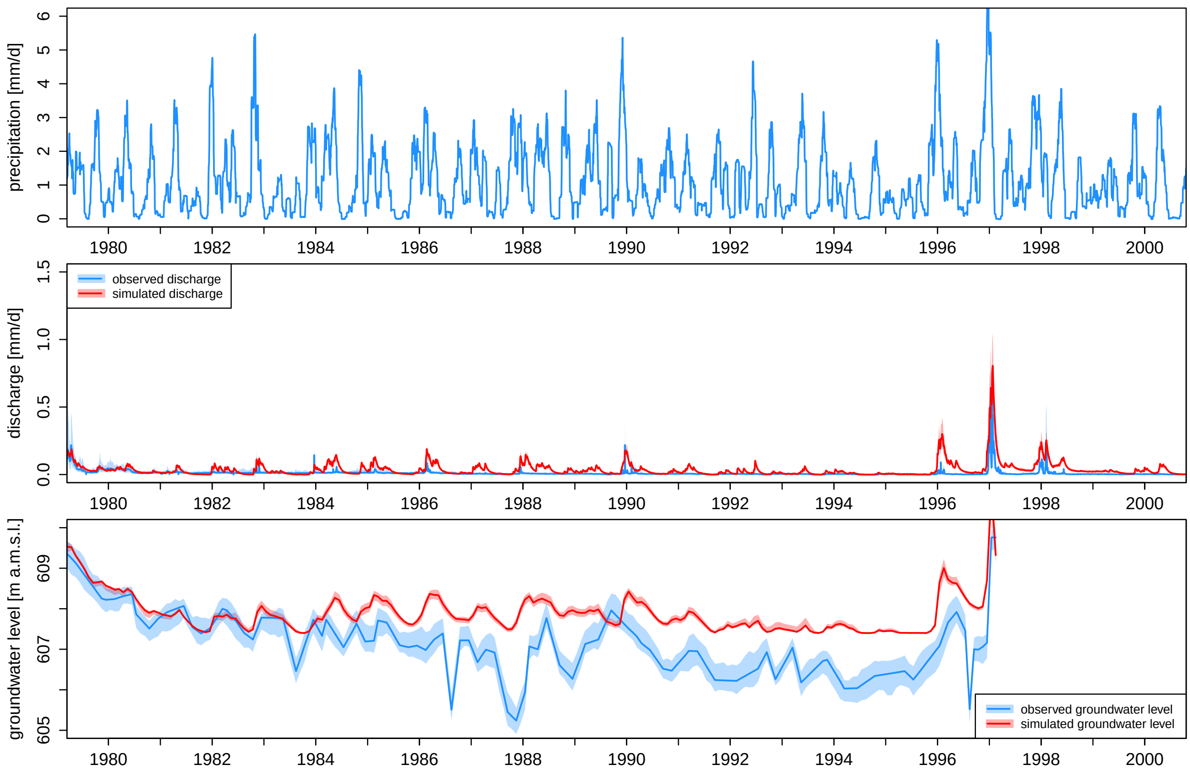

Figure 3.8: Time series of precipitation (30 day moving average), discharge, and groundwater level of Upper-Guadiana in the disturbed period; ranges give a rough indication of observational uncertainty (20 possible time series of noised observations) and simulation uncertainty (20 calibration results).

more or less similar for a short period. Afterwards, observed groundwater levels dropped quickly to low levels again (blue line), while also in the simulations a decreasing trend was visible (red line). In 1997, observed and naturalised groundwater levels were equal again. Zorrilla [2009] and Lopez-Gunn et al. [2011] reported that this period with high groundwater levels lasted for a short period, just as in discharge, and afterwards groundwater levels decreased again. Due to the limited period of record of the groundwater observations, we cannot confirm this.

To quantify the time-varying human influence on groundwater levels in the Upper-Guadiana, we plotted the difference between observed and naturalised groundwater level in Fig. 3.9. We can see that, in the disturbed period, on average, this difference increased almost linearly in time, which indicates a more or less constant decrease of storage due to abstraction. Exceptions are the wet years 1990 and 1997, in which high precipitation resulted in some groundwater recharge and, additionally, abstraction was probably limited because natural soil water supply through precipitation was sufficient.

3.4.4 Anomaly analysis

As the anomaly analysis method in the framework we applied the threshold level method (see Sect. 2.4.2). A fixed threshold can be used just as in the conceptual figure (Fig. 3.1). But in this example we applied a variable threshold, which is assumed to better represent the strong seasonal variability in the Upper-Guadiana (Sect. 2.1.5). The threshold values were calculated based on the undisturbed period (1960–1980) and applied to the entire time series. A different threshold was calculated for observed and simulated variables. For groundwater levels, the monthly threshold values were used directly because groundwater observations were available on a monthly time scale.

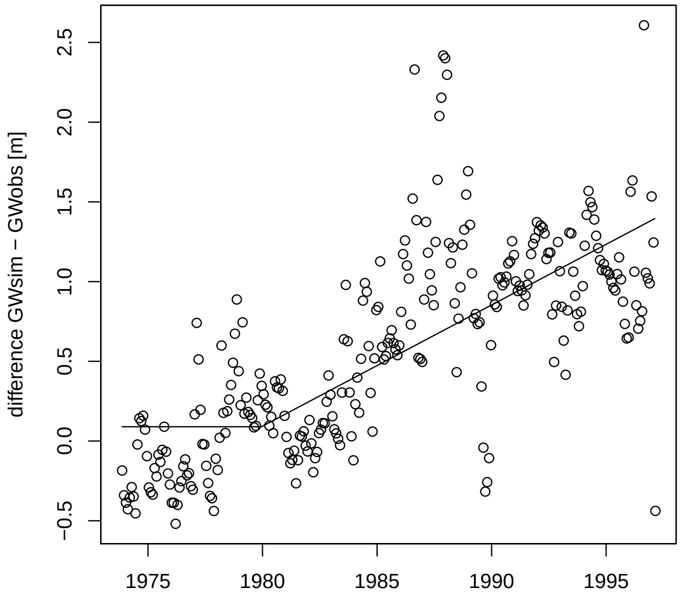

Figure 3.9: Difference between observed and naturalised time series of groundwater levels in undisturbed (horizontal regression line) and disturbed period (tilted linear regression line).

3.4.5 Comparison 2 - Anomalies

Anomaly analysis on hydrological variables of the naturalised situation and the human-influenced situation enables us to distinguish between drought and water scarcity (Sect. 3.2 and Fig. 3.2 - comparison 2).

To evaluate the distribution of the differences between drought and water scarcity over time, we studied time series again. The comparison of time series of anomalies in naturalised and observed discharge (Fig. 3.10) clearly shows that a number of severe droughts occurred in the disturbed period 1980–2001 (especially the multi-year drought of 1992–1996; Fig. 3.10 - upper row), but that water scarcity resulted in much longer dry periods (e.g. 1990–1996), dry years that did not occur in the naturalised situation (e.g. 1984–1985), and failure in recovery from drought (e.g. 1996–1998; Fig. 3.10 - middle row). Discharge can be used to evaluate the effect of the water scarcity in terms of the duration of anomalies. As discharge is bounded by zero, it can, however, not be used to evaluate the effect on the severity of anomalies. Groundwater levels give a better representation of differences in severity between drought and water scarcity.

The naturalised groundwater levels (Fig. 3.11 - upper row) showed long, but not so severe events, whereas observed groundwater levels (Fig. 3.11 - middle row) dropped deeply below the threshold, both in the relatively wet period (1985–1990) and in the relatively dry period (1990–1996).

Anomaly analysis also allows for further quantification of natural and human influences (Table 3.2). In Upper-Guadiana, the number of anomalies in the observed situation was three to four times lower than in the naturalised situation and mean duration was four to six times longer, in discharge and groundwater, respectively. The longest anomaly in observed discharge was more than ten years (122 months), whereas in naturalised discharge it was only two years (25 months). The deficit volume also showed large differences between naturalised and observed discharge. On average, deficit volume was 2.6 mm for drought and 14 mm for the com-

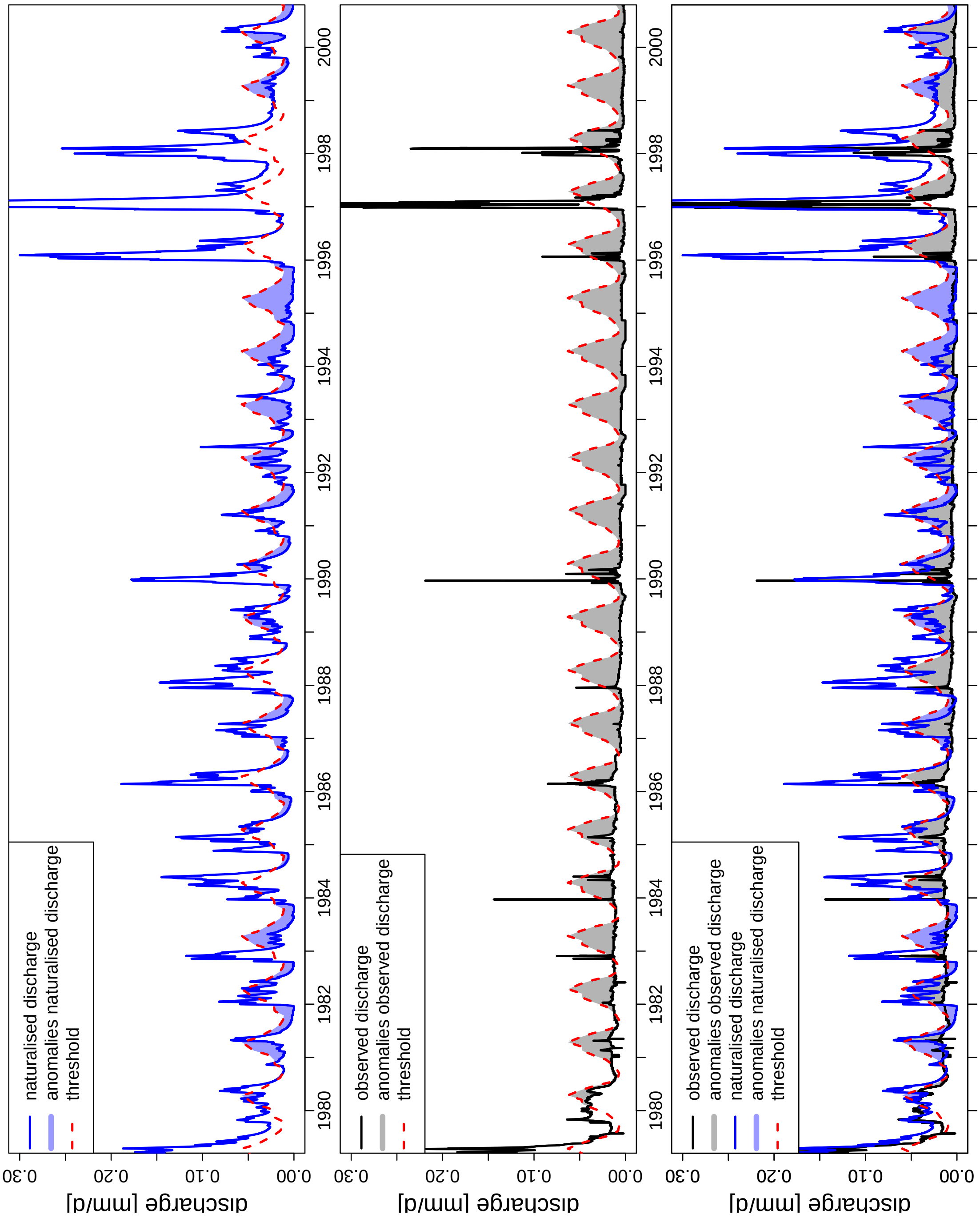

Figure 3.10: Anomalies in discharge in Upper-Guadiana in the disturbed period, using a 80 % monthly threshold (moving average 30 days), the inter-event time method for pooling, and a minimum drought duration of 15 days; upper row: naturalised discharge (blue areas indicate drought), middle row: observed discharge (grey areas indicate drought + water scarcity), lower row: comparison observed and naturalised discharge (blue areas indicate drought, grey areas indicate water scarcity).

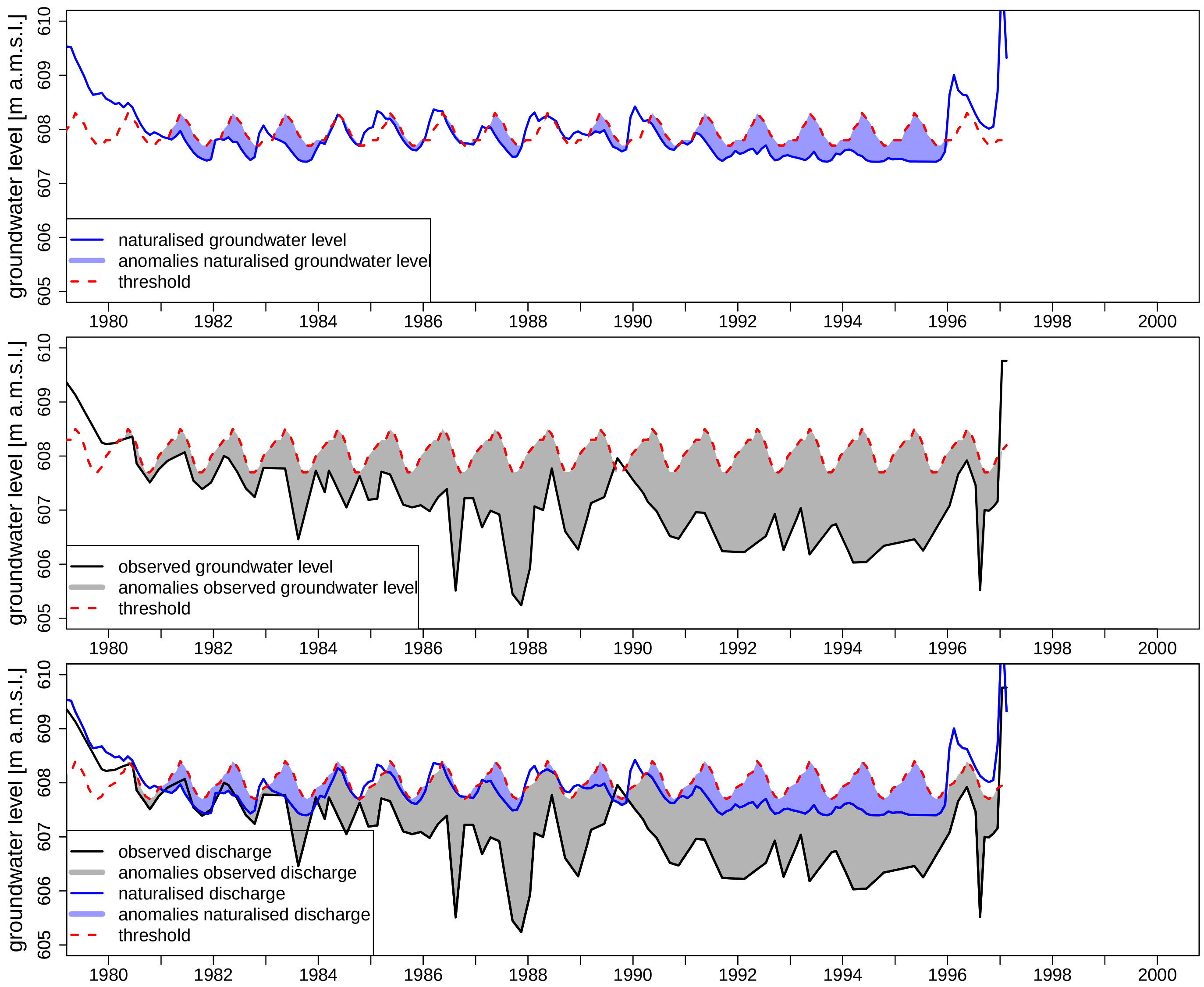

Figure 3.11: Anomalies in groundwater level in Upper-Guadiana in the disturbed period, using a 80 % monthly threshold; upper row: naturalised groundwater level (blue areas indicate drought), middle row: observed groundwater level (grey areas indicate drought + water scarcity), lower row: comparison observed and naturalised groundwater level (blue areas indicate drought, grey areas indicate water scarcity).

Table 3.2: General anomaly characteristics (using a 80 % monthly threshold) for the observed and the naturalised hydrological variables discharge (Q) and groundwater level (GW) of the disturbed period 1980–2001

	no. of anomalies [-]	duration [months]		deficit [mm]		max.deviation [m]	
		mean	max	mean	max	mean	max
naturalised Q	29	5.4	25	2.6	17.5	-	-
observed Q	*10*	*20*	*122*	*14.2*	*92.4*	-	-
naturalised GW	12	11	61	-	-	0.34	0.9
observed GW	*3*	*66*	*112*	-	-	*1.68*	*2.56*

bined effect of drought and water scarcity. Maximally, the deficit volume increased from 18 mm for the naturalised situation to 92 mm for the influenced situation (Table 3.2). This means that human influence added on average 12 mm (198×10^6 m^3) to the deficit volume, and in the maximum anomaly event even 75 mm (1236×10^6 m^3), which is more than four times as much as the natural effect.

For groundwater, the difference in maximum duration was lower than for discharge, five years (61 months) for naturalised and almost ten years (112 months) for observed groundwater levels (Table 3.2). However, the difference in mean and maximum max.deviation from the threshold indicates the large human influence on anomalies in groundwater. On average, drought events in groundwater had a 0.3 m deviation from the threshold and the combined events (natural and human influence) had a 1.7 m deviation. Maximally, the deviation from the threshold increased from 0.9 m for the drought events to 2.6 m for the combined events (Table 3.2). Consequently, the net effect of human influence (i.e. water scarcity) on the water table of the selected observation well in Upper-Guadiana was on average 1.3 m, with a maximum of 1.7 m, which is four and two times as much as the natural effect (i.e. drought).

The statistics in Table 3.2 only give information on the average and most extreme situation, but anomaly characteristics can of course be calculated for each anomaly separately. Another way of evaluating the evolution over time is provided in the lower rows of Figs. 3.10 and 3.11, in which the blue areas indicate drought and the grey areas water scarcity. It can be concluded that water scarcity resulted in a disappearance of the high-flow period in winter (Fig. 3.10 - lower row), even in relatively wet years (e.g. 1987–1989), and a non-linear response of the groundwater level, with some severe anomalies in for example 1986, 1987–1988, and 1996 (Fig. 3.11 - lower row).

3.5 Discussion

In this study, we proposed to be very clear on the use of the terms ‘drought’ and ‘water scarcity’ and, for this purpose, we spoke of ‘anomaly analysis’ instead of ‘drought analysis’. As mentioned in Sect. 3.1 in many studies, these terms are not used consistently. One example is the chapter on Human Influences [Van Lanen et al., 2004b] in the Hydrological Drought textbook [Tallaksen and Van Lanen, 2004]. On page 389–392 ‘drought’ characteristics, such as ‘drought’ duration, are calculated for a human-influenced system with groundwater abstractions. Similarly, Lopez-Moreno et al. [2009] claim to study dam effects on ‘drought’ magnitude and duration. And there are more examples. In these situations the term ‘drought’ should be avoided and the term ‘anomaly’ should be used instead. Sheffield and Wood [2011] use the term ‘drought’ for the situation in which both natural and human influences occur and talk about ‘climate-induced drought’ to denote the natural hazard (p. 30). For the sake of consistency, the term ‘human-induced drought’ might then be used for the situation in which a below-average water availability is caused by human influence. We are, however, not in favour of this concept as the definition of drought then becomes much less clear. Another terminology is suggested in EEA [2012]. In this report, a common term for the situation in which drought and water scarcity coexist is introduced, namely the abbreviation 'DWS'. We encourage the use of this term instead of 'drought' for situations in which both natural and human influences play a role. The most important, however, is being consistent in defining and using the terms throughout the whole study.

We considered all water stores and fluxes in a catchment, i.e. artificial storage in reservoirs in addition to natural storage in groundwater and wetlands (Sect. 3.1). This is consistent with the way drought is defined by the EU Mediterranean countries, but contrasts with the official EC definition that 'drought is a period of below-normal natural water availability' [EU, 2007]. When one uses catchment-averaged data (e.g. discharge) and a lumped conceptual model in the observation-modelling framework, the distinction between water stored in wetlands and water stored in artificial reservoirs cannot be made. In this study we investigated whether the causes of anomalies were natural or man-made, we did not investigate whether this anomaly occurred in natural or man-made water stores.

For the analysis of water scarcity several indicators exist [Savenije, 2000; Kummu et al., 2010]. The Water Exploitation Index (WEI) is widely used within the European Union to report and compare the water scarcity situation in different countries and river basins [EU, 2007]. WEI calculates the ratio of annual total freshwater abstraction to the long-term annual average total renewable resource [EEA, 2012; Kossida et al., 2012]. Disadvantages of the WEI (and of many other water scarcity indicators) are that good-quality data are often not available and that it is often based on aggregated spatial and temporal information [Savenije, 2000]. A fixed WEI on country level is meaningless for countries like Spain, where water scarcity conditions in the North are much different from those in the South and both abstractions and available resources change over time (as is shown in this research). This diversity can be included in the WEI by calculating the index on river (sub)basin spatial scale and on yearly or monthly time scale, but then data availability can become limiting. Furthermore, non-linearity is not taken into account in WEI and other indicators, which is a clear limitation of these approaches [Savenije, 2000]. WEI refers to average conditions, while during droughts it is likely that abstraction will be higher and water availability lower. What water managers need in order to be able to assess whether the water system still remains in a healthy (ecological) state, even under extreme conditions, is quantitative transient information on drought and water scarcity in state variables and fluxes [EU, 2012a; Kossida et al., 2012]. The observation-modelling framework proposed here is needed to complement indicators like the WEI.

The proposed observation-modelling framework can be used in the same way as was done in the example of Upper-Guadiana to quantify the distinction between drought and water scarcity in the hydrological system (e.g. groundwater and discharge), but it can also be applied to the impacts of drought and water scarcity, such as crop yields and river water temperature [e.g. Van Vliet et al., 2012]. Comparison 1 is a well-known method in hydrology, used for impact assessment of, for example, land use change [e.g. Van Lanen et al., 2004b]. Comparison 2 adds quantitative, transient information on anomalies, which is regarded as a novel approach. According to Vincent [2004] and Reed et al. [2006], a management framework that addresses the human-climate-terrestrial interactions impacting our river basins is clearly needed. As mentioned before, the observation-modelling framework (Fig. 3.2) can be adapted by using a different model. Clearly, hydrometeorological variables required as input and for calibration depend on the chosen model. Also the specific method used for anomaly analysis can vary. When applying the threshold level method, choices have to be made about the character (fixed or variable) and level of the threshold, pooling method, etc. (Sect. 2.4.2). In water management the choice of threshold level is dependent on the requirements for a healthy (ecological) state of the system, e.g. environmental minimum flow. The only prerequisites for using the framework in a more general sense are that observed data are available and that a model concept exists to estimate the naturalised situation in the disturbed period. If those conditions are met, both comparisons in Fig. 3.2 can be done, i.e. the comparison of the raw time series (comparison 1) and the comparison of anomalies (comparison 2).

In the Upper-Guadiana high-quality data of all abstractions were not available. Therefore we applied the framework by selecting the rainfall-runoff model HBV that requires only observed time series of meteorological variables (temperature, precipitation, potential evaporation; Sect. 2.3.1) and hydrological variables (discharge and groundwater levels; Sect. 2.3.2). In the ideal situation observed data of all anthropogenic effects on water resources would be available in addition to hydrometeorological variables, i.e. time series of all human influences like abstractions, reservoir releases, etc. However, groundwater pumping is generally one of the least measured variables in the hydrometeorological system [Martínez-Santos and Martínez-Alfaro, 2010; Ruud et al., 2004] and intensive groundwater extraction is usually carried out with little or no planning or control [Llamas and Martínez-Santos, 2005; Martínez-Santos and Martínez-Alfaro, 2010]. Pumping data, when measured, are usually not available in high spatio-temporal detail due to privacy regulations. Abstraction data can in some cases be estimated from data on irrigated area, type of crops, and unit water consumptions. Satellite data are also increasingly used [e.g. Droogers et al., 2010]. Data on reservoir operations are generally difficult to obtain. If more and better observational data are available, the distinction between drought and water scarcity becomes more reliable, because more advanced models can be used and the uncertainty related to observations and simulations decreases. The hydrological model used in the framework can then be more physically based [in the Upper-Guadiana we would, for example, use the PROOST model; Slooten et al., 2010; Jódar Bermúdez et al., 2011], which has the advantage that less calibration is needed and that spatially-distributed groundwater heads are obtained. The disadvantage of the framework is that some information of the undisturbed situation is always needed for calibration and/or validation of the model, even when using physically-based models [Savenije, 2009]. So the old saying 'monitoring is crucial' [Taylor et al., 2001; Svoboda et al., 2001; Reed et al., 2006; Lovett et al., 2007] certainly applies to the task of making the distinction between water scarcity and drought.

A challenge for the proposed framework lies in a region where very different types of human influences play a role simultaneously. In that case the observation-modelling framework can only be used to calculate the net anthropogenic effect and to separate it from natural effects. The separation of different human influences (that can even counteract each other) can only be done if reliable data of one (or more) of the human influences are available and the model applied in the framework can simulate one (or more) of these human influences.

In this chapter we showed an example of the application of the observation-modelling framework in the Upper-Guadiana catchment in Spain. We found that, in the past decades, human influences impacted the hydrological system on average four times as much as natural influences. These results correspond to earlier qualitative assessments of the effect of drought and water scarcity in Upper-Guadiana [Custodio, 2002] and adjacent basins [Lorenzo-Lacruz et al., 2010], but add the for water management very relevant quantification of the relative importance of human and natural influences on a transient basis. As mentioned in the introduction, water management can combat overexploitation of water resources [water scarcity, e.g. Garrido et al., 2006], whereas it only can adapt to climate variability [drought, e.g. Harou et al., 2010]. In the Upper-Guadiana catchment focus should mainly be on the former, with special attention to the latter during extreme events (when the human influence on anomalies in groundwater is 'only' twice the natural effect). This is in line with the recommendations of Palmer et al. [2008] and Schiermeier [2008], who advocate a shift in the focus of water management from reducing the vulnerability to drought to reducing the overexploitation of water resources.

Spanish rates of groundwater withdrawal to recharge are, however, even lower than those of, for example, southern and southwestern USA [Custodio, 2002] and parts of Australia [Habermehl, 2008]. As in those regions recent severe (multi-year) droughts [Leblanc et al., 2009; Roderick, 2011; Kogan et al., 2013] have worsened water scarcity problems, water managers in the USA, Australia, and other regions with similar problems could benefit from the application of an observation-modelling framework, comparable to the one proposed in this study.

3.6 Conclusions

In this chapter we proposed an observation-modelling framework to quantify the effect of natural (drought) and human influence (water scarcity) on anomalies in time series of groundwater and discharge. The main parts of this framework are a hydrological model that can simulate the 'naturalised' situation in the disturbed period and anomaly analysis that can identify anomalies in the time series of state variables and fluxes. The basic requirements of the framework are observed data: meteorological data of the entire period to run the hydrological model, hydrological data of the undisturbed period to calibrate the hydrological model, and hydrological data of the disturbed period for comparison with the naturalised situation. Two comparisons can be made to quantify the relative effect of human and natural influences, i.e. comparison 1 on the raw time series to identify human influence, and comparison 2 on the anomalies obtained from anomaly analysis to distinguish between water scarcity and drought. The combination of comparison 1 and comparison 2 is the innovative part of this research.

We demonstrated the application of the proposed observation-modelling framework in a heavily-influenced groundwater catchment in Spain by using a conceptual hydrological model and the variable threshold level method. For this catchment we found that the difference between observed and naturalised groundwater levels increased almost linearly in time, which indicates a more or less constant decrease of storage due to unsustainable abstraction. Exceptions are some very wet years, in which recharge to the groundwater occurred and abstraction was limited. Anomalies in observed discharge and groundwater (i.e. the situation with both natural and human influences) were three to four times less frequent and, on average, four to six times longer than those in naturalised discharge and groundwater. The relative effect of both natural and human influences was quantified using the severity measures deficit volume for discharge and maximum deviation from the threshold for groundwater. These measures showed that human influence in the Upper-Guadiana catchment was, on average, four times as large as natural influence, in both discharge and groundwater. Due to non-linearity in groundwater response this decreased to two times as large during extreme events.

The proposed observation-modelling framework gives a water manager a tool to distinguish between natural and human effects on anomalies and adapt his/her management accordingly. This would imply adapting to the natural variability (drought) and reducing unsustainable use of water resources (water scarcity).

Chapter 4

Hydrological winter drought

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4.1 Introduction

Hydrological droughts occurring in winter can have profound effects on water resources, both in winter and in the subsequent summer. Drought research, however, mainly focuses on summer droughts. Drought is a sustained and regionally extensive occurrence of below-average natural water availability. It can thus be characterised as a prolonged deviation from normal conditions of the natural system (climate and hydrology), which is reflected in variables such as precipitation, soil moisture, groundwater and streamflow [Tallaksen and Van Lanen, 2004]. In literature, winter droughts are defined as a drought occurring 'when the temperature is below the freezing point and water in the catchment is stored as snow and ice' [Tallaksen and Hisdal, 1997; Hisdal et al., 2001; Fleig et al., 2006]. In this chapter winter drought is defined as a drought in the winter season. The winter season is limited to the months in which monthly average temperature is below zero and monthly precipitation falls (partly) as snow. Winter droughts are, following the definition of drought above, regarded as a prolonged deviation from 'normal' winter conditions. A drought in precipitation may affect all variables in the hydrological cycle, e.g. soil moisture storage, groundwater storage, and discharge. This process is called propagation of drought [Peters et al., 2006] and is dependent on climate and catchment characteristics (Sects. 1.2.2–1.2.4). To understand the processes underlying drought development and propagation in winter, data on several hydrometeorological variables are needed, either measured or modelled (Sect. 2.2). Our objective is to increase understanding of winter drought development in snow-affected regions in Europe. To reach this objective, winter droughts in two different climate regions were investigated using observed and modelled data from two catchments in Europe (in Norway and Slovakia).

4.2 Study areas

We studied two small, contrasting, snow-affected catchments in Europe: Narsjø in Norway, and Nedožery in Slovakia (see Sects. 2.1.1 and 2.1.4, Fig. 2.1, and Table 2.1).

In both catchments, seasonal variation in temperature is on average slightly above 20°C, and seasonal variation in precipitation shows a maximum in summer and a minimum in winter (Table 2.1). Both catchments show a peak in discharge during the snow melt season, but the peak in Narsjø is more pronounced and occurs later than in Nedožery. The low-flow season of Narsjø is winter (due to precipitation accumulating as snow) and that of Nedožery is summer (due to a high climatic deficit). Both catchments are dominated by hard rock. Nedožery shows the expected quick response to rainfall, but in Narsjø the rainfall response is delayed due to the presence of many bogs and lakes.

4.3 Methods and materials

As a common model for both catchments we used the semi-distributed rainfall-runoff model HBV (Sect. 2.3.1). Droughts in precipitation, soil moisture, groundwater storage, and discharge were evaluated using the threshold level method (Sect. 2.4.2).

4.4 Results

Model results are used to explain the hydrological regime in both catchments and to discuss processes underlying winter droughts.

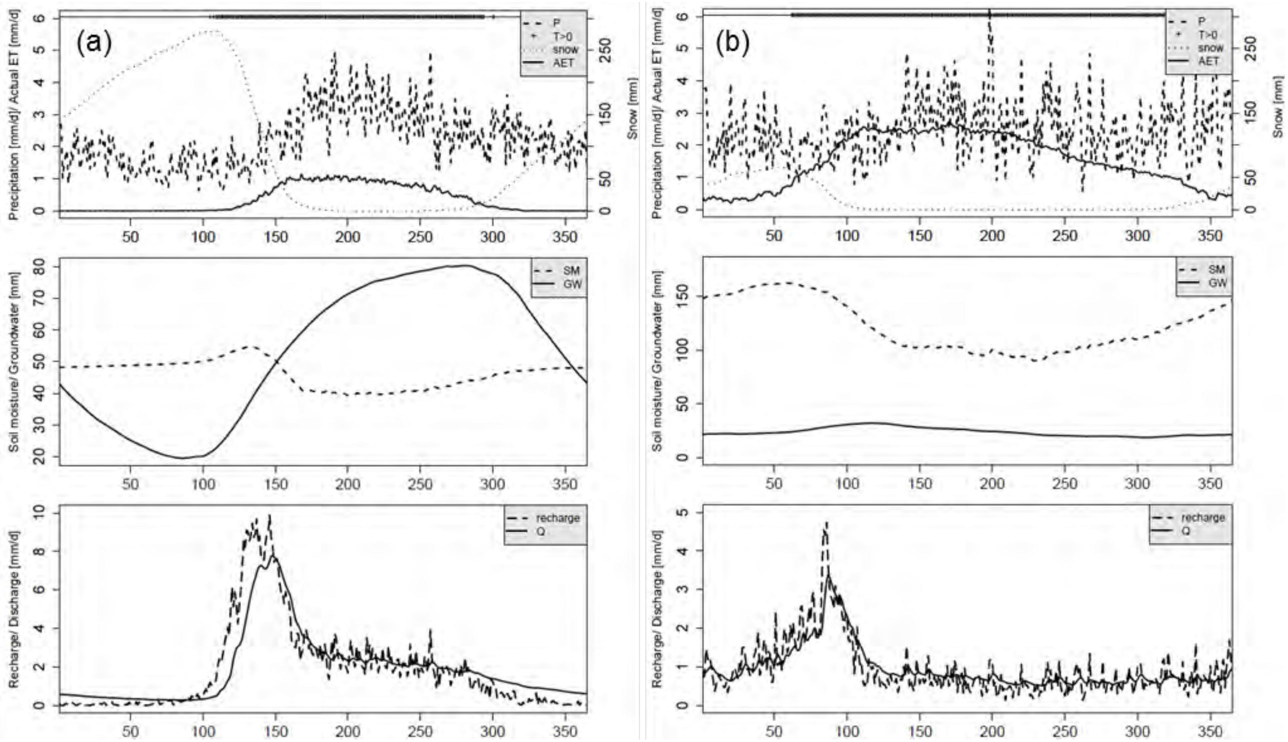

Figure 4.1: Long-term daily averages of HBV model output: (a) Narsjø catchment (1958–2007); (b) Nedožery catchment (1974–2006). Top row: precipitation (P), temperature (T, – = below zero, + = above zero), actual evapotranspiration (AET), and snow accumulation (snow); middle row: soil moisture storage (SM) and groundwater storage (GW); lower row: recharge and discharge (Q). Note the different scale of the y-axes.

4.4.1 HBV model results and the hydrological regime

Both catchments are modelled reasonably well with HBV. For Narsjø, model results show a high ln Reff (0.90). This is due to the very regular hydrological regime, dominated by yearly recurring winter low-flow conditions that can be captured quite well with a simple rainfall-runoff model like HBV. For Nedožery, however, ln Reff (0.68) is lower than for Narsjø, because the hydrological regime is less regular and more determined by a fast response to rainfall. Model results in Nedožery are therefore more dependent on the quality and representativeness of precipitation measurements. More detailed calibration and validation results can be found in Appendix B.

The hydrological regimes of Narsjø and Nedožery show clear differences (Fig. 4.1). In Narsjø, temperatures drop below zero for on average six months (175 days), while in Nedožery only three months (90 days) have negative temperatures. This temperature difference influences evapotranspiration in summer, which is on average three times higher in Nedožery than in Narsjø, and snow accumulation in winter, which rises to almost 300 mm Snow Water Equivalent (SWE) in Narsjø and stays at approx. 60 mm SWE in Nedožery. Lower temperatures in Narsjø also cause a later maximum snow accumulation and later snow melt than in Nedožery. Maximum snow accumulation in Nedožery occurs, on average, in the middle of February, while in Narsjø the snow pack grows until the middle of April. Contrary to what the long-term averages in Fig. 4.1 suggest, daily temperature can temporarily rise above zero during the winter months and cause some snow melt in both catchments. In Narsjø, this temporary melt is limited to a few days per winter season and never causes a complete melt of the snow cover. In Nedožery, however, these periods of relatively high temperature can last for weeks and can cause a complete melt of the snow cover during the winter period, at least in the lower parts of the catchment. The effect is clearly visible in the soil moisture regime. In Narsjø, simulated soil moisture is nearly constant during winter, because there is no inflow into the soil (precipitation accumulates as snow). In Nedožery, however, simulated soil moisture increases during the winter months, due to occasional and partial melting of the snow cover. This process is also reflected in the recharge regime. In Narsjø recharge is zero during winter and starts to rise only when snow melt starts in April. In Nedožery recharge is lowest in summer and increases during the winter season. The discharge regime in both catchments follows recharge quite closely. The snow melt peak in Narsjø is slightly delayed and smoothed due to the influence of bogs and lakes.

4.4.2 (Winter) drought analysis and classification

General drought characteristics were determined from the HBV modelling results. In Nedožery the average number of droughts per year is higher than in Narsjø for all variables (5–30 % higher), but droughts are 10–25 % shorter and mean deficit and mean intensity of discharge droughts are up to 70 % lower (absolute numbers not shown). This again indicates the fast response of the Nedožery catchment.

Study of the processes underlying the development of winter droughts reveals two types of winter drought occurring in the snow-affected catchments Narsjø and Nedožery. In Narsjø a typical winter drought (Type 1) is found that develops when the winter season is preceded by a dry summer. The mechanisms behind this drought type can be demonstrated using the example of the 1968–1969 drought (Fig. 4.2a). A meteorological drought in late summer 1968 (start: 9 July, duration: 67 days, deficit: 53 mm) caused low soil moisture, groundwater storage, and discharge. Some precipitation in November led to a recovery of the precipitation and soil moisture droughts, but no recharge to groundwater occurred because a large part of the precipitation fell as snow and the rest replenished soil moisture. Consequently, the starting point of the winter recession of groundwater storage and discharge was lower than normal. Both variables stayed below the threshold for 248 and 259 days, respectively, until snow melt started in the spring of 1969 and recharge replenished groundwater storage. This type of winter drought (Type 1) occurred five times in the period 1958–2007, so on average once every 10 years. Deficits are large (up to 50 mm) over a long period of time (over 8 months), so water resources can be negatively impacted.

In Nedožery Type 1 winter droughts occurred three times in the studied 33 year period, but were not as long and intensive as in Narsjø because snow melt and/or rain periods occur frequently during the winter season. Investigation of the processes underlying other winter droughts in Nedožery revealed an unexpected type of winter drought (Type 2). An example of a Type 2 winter drought is the 1989–1990 drought (Fig. 4.2b). During the winter months December–April, temperature in Nedožery was higher than normal and precipitation lower than normal (except for March). This resulted in a limited snow cover (Fig. 4.2b). The meteorological drought that developed was relatively short and intensive, with a deficit of 29 mm over a period of 64 days, while the corresponding discharge drought (adding up all dependent droughts in the period, like in Eqs. 2.7 and 2.8, although in this chapter no formal pooling procedure was applied, see Sect. 2.4.2) reached a total duration of 143 days and a deficit of only 17 mm. The distinctive feature of the Type 2 winter drought are the interruptions of the drought (especially visible in discharge), caused by the interplay between temperature and precipitation. Soil moisture and groundwater storage also show small peaks during the drought close to the threshold level (Fig. 4.2b), but the signal is smoother than that of discharge. This difference between storage and discharge signals is due to the fast runoff of snow melt and rain to the stream, resulting in low recharge to the groundwater storage. In a climate like that of Nedožery, characterised by winters with (catchment average) temperatures around zero and regular snow accumulation and melt, the combined effect of temperature and precipitation is important for drought development. Type 2 winter droughts are found three times in Nedožery in the period 1974–2006, so they occur on average once every 10 years. Sometimes, impacts of these winter droughts are still visible during the subsequent summer. Type 1 winter droughts end due to

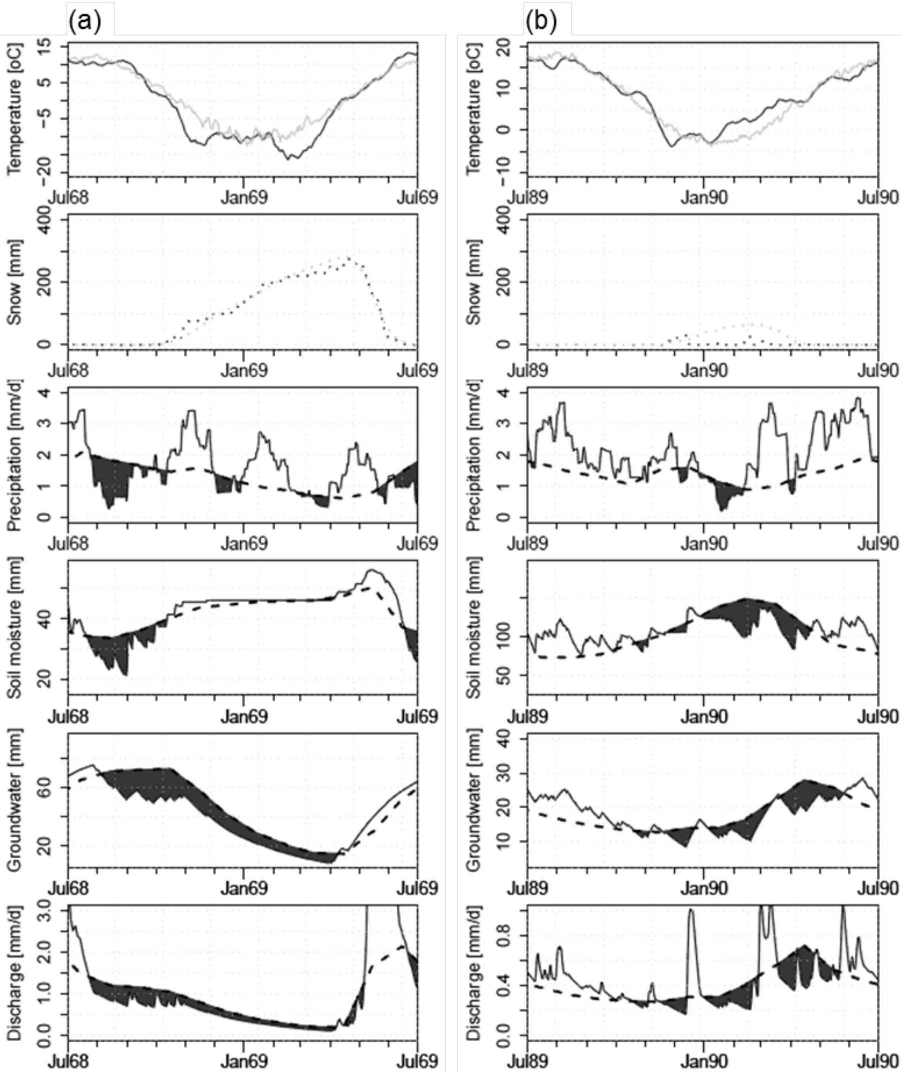

Figure 4.2: Winter drought propagation: (a) example of Type 1 winter drought in the Narsjø catchment; (b) example of Type 2 winter drought in the Nedožery catchment. Top row: 30-day moving average temperature (grey = long term average); second row: snow accumulation; third row: 30-day moving average precipitation, dashed line = monthly precipitation threshold; fourth row: soil moisture storage, dashed line = monthly soil moisture threshold; fifth row: groundwater storage, dashed line = monthly groundwater threshold; lower row: discharge, dashed line = monthly discharge threshold.

snow melt. But in the case of Type 2 winter droughts, recharge from snow melt is often too low to end the drought and it continues into summer (and can be enhanced by a precipitation drought in summer), which can have a very negative impact on water resources.

4.5 Discussion

Based on research in two snow-affected catchments in Europe (in Norway and Slovakia), we found two distinct types of winter droughts: Type 1 winter droughts, which are caused by late summer droughts that continue into winter, and Type 2 winter droughts, which develop when snow cover disappears (in part of the catchment) and precipitation is lower than normal. So far, Type 2 winter droughts have not been described in literature, while Type 1 winter droughts have previously been discussed [Tallaksen and Hisdal, 1997; Hisdal et al., 2001; Fleig et al., 2006]. These studies focused on summer droughts, but had some problems eliminating winter droughts from their analysis. In many time series they observed summer droughts that continue into winter (= Type 1). However, the processes behind the development and propagation of this type of drought have not been treated as extensively as in this study before.

We argue that Type 2 winter droughts occur widespread over Europe, because large parts of Europe have mild winters with some snow accumulation, and temperatures around zero degrees. However, the occurrence of the described types of winter drought is not discrete. Transition zones exist where processes interact and the types of winter drought cannot easily be discriminated. These transitions can be topographic, so changing with elevation (for example, in central Europe, e.g. Alps, Carpathians), or non-topographic, so changing with latitude or distance to the coast (e.g. in southern and western Norway). Global warming will likely influence the occurrence of winter drought types in Europe, because cold climates (with Type 1 winter droughts) might change into milder climates (with Type 2 winter droughts). The consequence is that winter droughts will more often continue into summer, with larger impact on water resources.

In this research the processes underlying two types of winter droughts were studied based on rainfall-runoff modelling using the HBV model (Sect. 2.3.1). Conceptual rainfall-runoff modelling was needed because not all data for detailed mechanistic understanding were measured in the studied catchments. Although efficiency of model results was satisfactory, results were not perfect and some uncertainties remain [Wagener et al., 2004]. These uncertainties do not allow for a detailed analysis of the exact start and end date, and deficit of a single drought. However, it is feasible to draw general conclusions about more generic drought processes. Furthermore, care should be taken when interpreting drought in simulated time series of soil moisture and groundwater storage as these cannot be considered to represent true catchment averages. It is not straightforward to compare drought characteristics in variables of different nature, here fluxes and storages, as discussed in Tallaksen et al. [2009].

For drought analysis, we used the well-known threshold level method with a few adaptations (Sect. 2.4.2). To find winter droughts a fixed threshold is not suitable [Hisdal et al., 2004], therefore we used a variable threshold. A monthly threshold is regarded as most convenient, because it has enough detail to pick out seasonal differences, but is not too much influenced by individual measurements, as a daily threshold would be. We used a centred 30-day moving average to get rid of the ‘staircase’ pattern of the monthly threshold and jumps between the months (Fig. 2.8). This smoothed monthly threshold method has proved to be a good tool to investigate the processes underlying the development and propagation of winter droughts.

4.6 Conclusions

Two types of winter droughts could be discriminated by studying winter drought development in two snow-affected catchments in Europe (Narsjø in Norway and Nedožery in Slovakia). In Narsjø, where winters are extremely cold and water is stored as snow throughout the winter season (Köppen climate Dfc), Type 1 winter droughts are found. These droughts develop when, due to a lack of summer rain, the starting point of winter recession is lower than normal, and groundwater and streamflow drop below the monthly threshold. Type 1 winter droughts last long and have large deficits, so impact on water resources can be large. However, they are relieved by the snow melt peak in spring. In Nedožery (catchment average) winter temperatures are around zero degrees and periods of snow accumulation alternate with periods of snow melt (Köppen climate Dfb). Type 1 winter droughts can occur in Nedožery, but are hard to discriminate and generally are short-lived. Additionally, modelling results for Nedožery revealed a type of winter drought that is determined by both temperature and precipitation (Type 2). When winter temperatures are temporarily above zero and snow accumulation is limited, a continuous snow cover does not develop. When, in addition, precipitation (rain) is below average, a hydrological drought develops. This Type 2 winter drought has not been studied before, but is probably more widespread in snow-affected regions in Europe and beyond (i.e. in mountainous regions and other transition areas at higher latitudes). Because this type of winter drought can continue into summer, impacts on water resources can be large. Especially if, due to global warming, catchments that previously experienced a continuous snow cover during the winter season change to a climate like that of Nedožery, then the occurrence of Type 2 winter droughts can become more widespread.

Chapter 5

Hydrological drought typology

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5.1 Introduction

Hydrological drought events are severe natural disasters, in damage comparable to large-scale floods and earthquakes (Sect. 1.1). Due to their long duration and large spatial extent, droughts have significant economic, social, and environmental impacts [EU, 2006b, 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, economic loss due to drought on average amounts to 6 to 8 billion USD per year [Andreadis et al., 2005; Below et al., 2007] and in the EU it was estimated at more than 100 billion EUR in the period 1976–2006 [EU, 2006b, 2007]. According to recent drought studies [EU, 2006b, 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.

Compared to that of other natural disasters, knowledge of drought still has large gaps [Smakhtin, 2001; Mishra and Singh, 2010]. The focus of drought research mostly is on finding the 'best' drought index [e.g. Bonacci, 1993; Heim Jr., 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 (Chs. 1 and 4, and Van Lanen et al. [2004a]). In 2006 and 2010, for example, cold and dry winters have resulted in severe problems with drinking water supply 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 for hydrological droughts [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 is based on drought severity [Dracup et al., 1980a; Rossi et al., 1992; McKee et al., 1993, 1995; Lloyd-Hughes and Saunders, 2002; Smakhtin and Hughes, 2004], but not 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] or a single drought event with its underlying processes is described in detail [e.g. 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 propagation and for improvement of drought forecasting and management.

In this chapter we propose a general hydrological drought typology based on the underlying processes of drought propagation. These governing processes were derived from time series investigation (observed and/or simulated) and drought analysis in selected catchments with contrasting characteristics. Therefore, the resulting typology is applicable to other catchments around the world where observed and/or simulated hydrometeorological data are available. The objectives of this study are: (i) to describe hydrological drought types and provide examples, (ii) to show the application of the drought typology by classifying hydrological drought events in five contrasting catchments, (iii) to find the most common and most severe drought types in

Table 5.1: Nash-Sutcliffe values per catchment; Reff = based on discharge values, ln Reff = based on the logarithm of the discharge values

	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

catchments with different climate and catchment characteristics, and (iv) to relate these drought types to catchment and climate control.

The outline of the chapter is focused on the hydrological drought typology, which is presented in Sect. 5.4 and applied in Sect. 5.5. The defined drought types are the result of detailed studies of drought events in five contrasting study areas (Sect. 5.2), which were analysed using a hydrological model (Sect. 5.3.1) and a drought analysis method (Sect. 5.3.2). Finally, in Sects. 5.6 and 5.7, 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.

5.2 Study areas

The five catchments used as the basis for this study are natural headwater catchments in Europe with contrasting climate and catchment characteristics (see Sect. 2.1 and Van Lanen et al. [2008]).

5.3 Modelling and drought analysis

5.3.1 Hydrological modelling

As a common model for both catchments we used the semi-distributed rainfall-runoff model HBV (Sect. 2.3.1). After calibration, all selected catchments were modelled reasonably well with HBV (Table 5.1). In general, ln Reff values were (slightly) higher than Reff values, because calibration was based on ln Reff (Sect. 2.3.2). This indicates a good performance of the model on low flows, in particular when the lack of success in improving the model is considered (Sects. 2.3.3 and 2.3.4). Further validation of the HBV model results, including graphs and tables of simulated vs. observed discharge and groundwater, are given in Appendix B. The results of calibration and validation of the HBV model justify the use of simulated fluxes and state variables for drought analysis.

5.3.2 Drought analysis

Droughts in precipitation, soil moisture, groundwater storage, and discharge were evaluated using the threshold level method (Sect. 2.4.2).

A few drought events 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 (thus not caused by meteorological anomaly, as defined by Stahl and Hisdal [2004]), but are purely a consequence of the smooth threshold level in

Table 5.2: General drought characteristics using an 80 % monthly threshold (moving average 30 days), the inter-event time method for pooling, and a minimum drought duration of 15 days for the hydro-meteorological variables simulated with HBV and observed discharge for all selected catchments

		No. of droughts [per year]	Mean duration [day]	Mean deficit [mm]	Mean maximum 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*	–

combination with a sharp increase in groundwater storage or discharge. Therefore, in this research we did not consider these events as drought but rather as artefact. In this research such artefacts 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 artefacts 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 artefacts related to a sudden increase in precipitation. In the remainder of this chapter these artefacts are disregarded and the focus is only on droughts.

General drought characteristics of all study catchments are displayed in Table 5.2. The drought events of simulated and observed discharge showed similar characteristics (especially regarding the number of drought events and their mean duration), again indicating the reasonable performance of the HBV model on low flows. Only in the Upper-Guadiana catchment did drought characteristics of simulated discharge deviate significantly from those of observed discharge. In this catchment observations and simulations cannot be compared, as is explained in Sect. 2.3.1. The reason is that drought characteristics of this catchment were calculated for the entire observation period (1960–2001), including the period of strong human influence (Sect. 2.1.5 and Ch. 3). 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 5.2 confirms what is known about propagation in drought characteristics (Sect. 1.2.2):

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

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

- In fast responding systems (like Narsjø and Nedožery), discharge drought characteristics are more comparable to those of precipitation (more and shorter); in slow responding systems (like Upper-Metuje and Upper-Guadiana) discharge drought characteristics are more comparable to those of groundwater storage (fewer 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 maximum 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.

The Narsjø and Nedožery catchments have similar drought characteristics because they are both fast reacting (Table 5.2). Narsjø is a bit slower (fewer, but longer lasting groundwater droughts) due to the presence of bogs and lakes that slightly delay the response to precipitation. The Upper-Metuje and Upper-Sázava catchments have similar drought characteristics because they are both slow reacting (Table 5.2). 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, more than two years; Table 5.2). 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 (Sect. 2.1.5).

The numbers in Table 5.2 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.

5.4 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 the 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*.

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

5.4.1 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. 5.1 with droughts in summer, spring, and winter in different catchments. In the first example (Fig. 5.1a, Narsjø catchment), a meteorological drought in May–July 1992 (3rd row) caused a drought in soil moisture, groundwater storage, and discharge (4th, 5th, and 6th row). The hydrological drought event ended by high precipitation in July–August 1992 (3rd row). In the second example (Fig. 5.1b, Nedožery catchment), a meteorological drought in April–June 2000 and one in August 2000 (3rd row) both caused a soil moisture drought (4th row) and a hydrological drought (groundwater storage and discharge; 5th and 6th row), with a small peak in between due to rainfall in July 2000 (3rd row). The hydrological drought event ended by high precipitation in autumn (September–November 2000; 3rd row). In the third example (Fig. 5.1c, Upper-Guadiana catchment), a meteorological drought in winter (February–March 1988; 3rd row) caused only a minor drought in soil moisture (4th row) and a hydrological drought (groundwater storage and discharge; starting in March 1988; 5th and 6th row). The drought in soil moisture and discharge ended by rainfall in spring (March–June 1988; 3rd row), but the drought in groundwater storage continued because recharge was not sufficient (5th row).

The *classical rainfall deficit drought* can occur in any season, in any catchment (fast responding or slow 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. 5.4.2, 5.4.4 and 5.4.5). A *classical rainfall deficit drought* can have all possible durations, deficit volumes, and maximum deviations, mainly dependent on the rainfall deficit(s) that caused it and on the antecedent storage in the catchment. In the examples in Fig. 5.1, durations range from 28 to 245 days, maximum deviations from 2.9 to 10.7 mm, and deficit volumes from 0.45 to 28 mm. *Classical rainfall deficit droughts* can exhibit all propagation features (i.e. pooling, lag, attenuation, and lengthening; see Sect. 1.2.2), mainly dependent on catchment characteristics. Pooling, for example, often occurs. The examples in Fig. 5.1 show a clear propagation of one meteorological drought event into one hydrological drought event, but in many cases more meteorological drought events are pooled and it is more difficult to point out the exact rainfall deficits that caused a specific hydrological drought event. In the examples in Fig. 5.1, 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. 5.1b) and especially in the groundwater storage of the Upper-Guadiana catchment (Fig. 5.1c).

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 Stahl and Demuth [1999]; Tallaksen and Van Lanen [2004]; Stahl and Hisdal [2004]; Smakhtin and Hughes [2004]; and Fleig et al. [2006].

5.4.2 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 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 of the next spring.

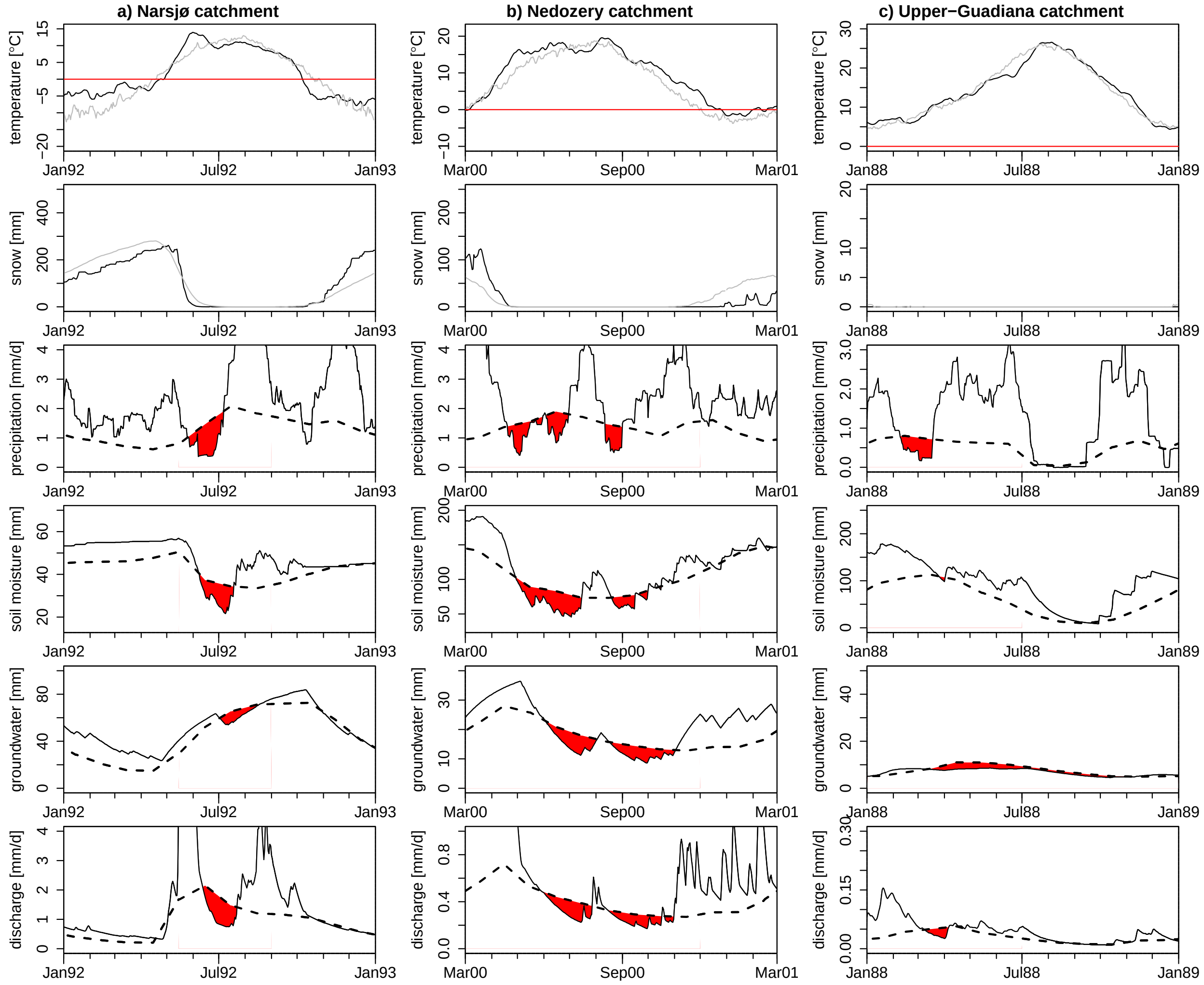

Figure 5.1: Examples of *classical rainfall deficit drought* type: **(a)** Narsjø catchment 1992–1993, **(b)** Nedožery catchment 2000–2001, **(c)** Upper-Guadiana catchment 1988 (all rows: grey line = long-term average of displayed variable, dashed line = smoothed monthly 80 %-threshold of displayed variable, red area = drought event referred to in text; upper row: black line = 30-day moving average of observed temperature, red line = 0 degrees; second row: black line = simulated snow accumulation; third row: black line = 30-day moving average of observed precipitation; fourth row: black line = simulated soil moisture; fifth row: black line = simulated groundwater storage; lower row: black line = simulated discharge).

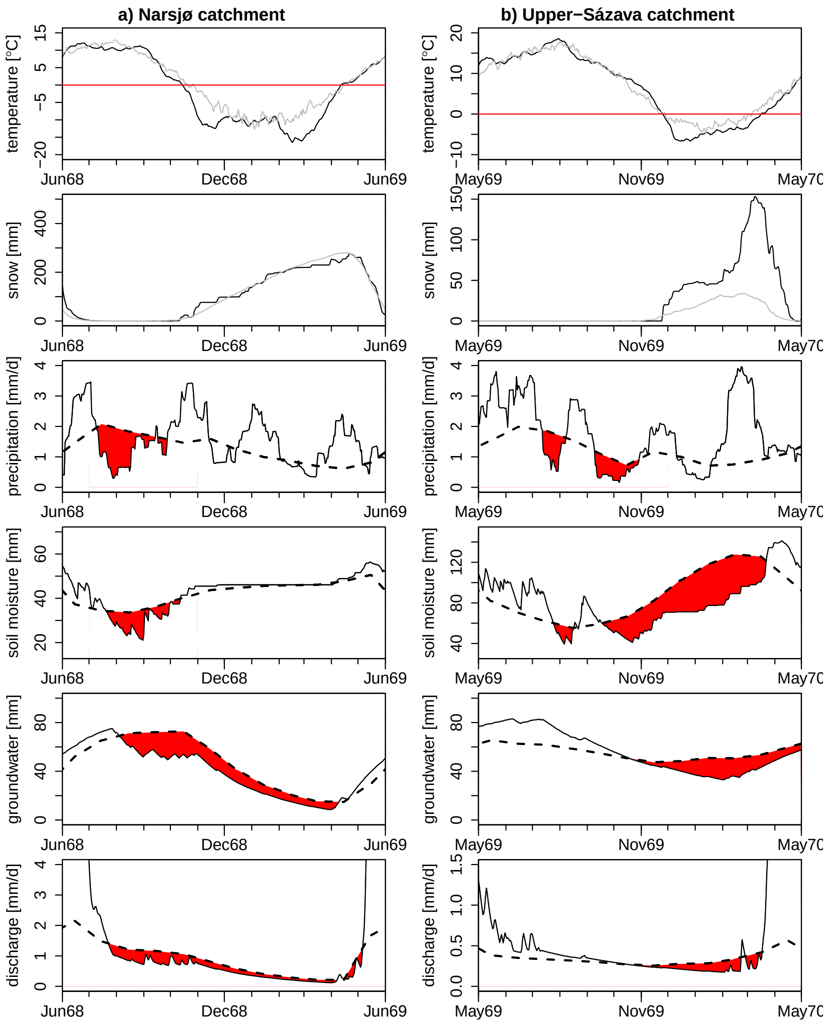

Figure 5.2: Examples of *rain-to-snow-season drought* type: **(a)** Narsjø catchment 1968–1969, **(b)** Upper-Sázava catchment 1969–1970 (legend: see Fig. 5.1).

Two examples of the *rain-to-snow-season drought* are shown in Fig. 5.2. In the first example (Fig. 5.2a, Narsjø catchment), the meteorological drought in July, August and September 1968 (3rd row) directly resulted in a soil moisture drought (4th row) and hydrological drought (5th and 6th row). The precipitation peak that started mid-October (3rd row) mainly fell as snow (2nd row) because temperatures had dropped below zero (1st row). Some replenishment of the soil moisture store took place and the soil moisture drought disappeared (4th row), but the groundwater system remained in drought until the snow melt peak of May 1969 (5th row). In the second example (Fig. 5.2b, Upper-Sázava catchment), two meteorological droughts of July and September–October 1969 (3rd row) caused groundwater storage (5th row) and discharge (6th row) to decrease below threshold levels. Part of the precipitation of November 1969 and almost all that of February 1970 (3rd row) fell as snow (1st and 2nd row). Therefore, the hydrological drought did not end, but continued until the snow melt period of April 1970 (6th row). In the groundwater system, the drought even continued longer, until July 1970 (not shown).

The *rain-to-snow-season drought* occurs in catchments with a clear snow season, notably catchments at high latitude or 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. 5.2, 279 and 147 days for drought in discharge) and deficit

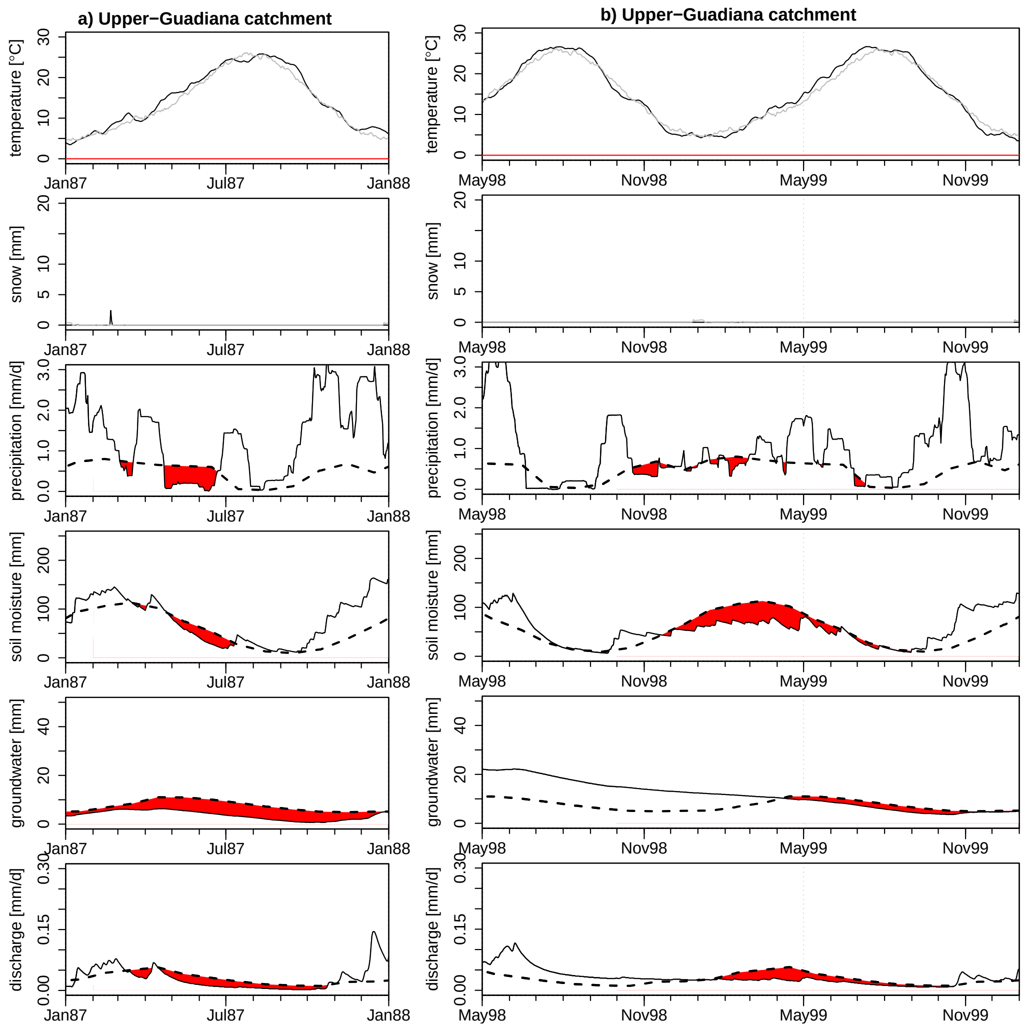

Figure 5.3: Examples of *wet-to-dry-season drought* type: **(a)** Upper-Guadiana catchment 1987, **(b)** Upper-Guadiana catchment 1998–1999 (legend: see Fig. 5.1).

volumes can be high (partly due to the long durations; in the examples of Fig. 5.2, 54 and 11 mm for drought in discharge). As can be seen from the examples in Fig. 5.2, 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. 5.2b), but are less important than lengthening.

The *rain-to-snow-season drought* has previously been described in Ch. 4 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].

5.4.3 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. 5.3 (both Upper-Guadiana catchment; in the other studied catchments the potential evaporation is not sufficiently high to cause this type of drought). In the first example (Fig. 5.3a), one large meteorological drought in the wet season (April–June 1987; 3rd row) caused discharge to drop below the threshold level (6th row). Groundwater was already in drought (5th row) as a remnant of a previous dry period. The rainfall event of June–July 1987 (3rd row) did not result in recovery from the hydrological drought, because the precipitation was partly lost to evapotranspiration and partly used for replenishment of soil moisture (4th row). The hydrological drought continued until December 1987 (6th row), when rainfall was high (3rd row) and potential evaporation was lower than in summer. In the second example (Fig. 5.3b), a number of small meteorological drought events in the wet season (between November 1998 and May 1999; 3rd row) resulted in a soil moisture drought in the wet season (4th row) and a decrease in groundwater storage and discharge to below-threshold levels (5th and 6th row). In both examples, the hydrological drought continued throughout the dry season, until the first recharge in the following wet season (November–December).

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 (six months to a year; in the examples of Fig. 5.3, 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. 5.3, 3.0 and 2.7 mm for drought in discharge). Just as in *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. 5.3b), but are less important than lengthening.

The *wet-to-dry-season drought* has previously been described by Tate and Freeman [2000]; Van Lanen et al. [2004a]; Stahl and Hisdal [2004]; Trigo et al. [2006]; Santos et al. [2007]; Pandey et al. [2008]; Trigo et al. [2010]; and Kim et al. [2011].

5.4.4 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 the 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 values of the recession of soil moisture, groundwater storage, and discharge are high enough, this will not lead to drought (see Sect. 5.5.3); but when storage and discharge are already low, groundwater storage and discharge can drop below threshold levels during winter. An example is shown in Fig. 5.4a (Narsjø catchment). In this case, temperature dropped below zero two weeks earlier than normal, in the beginning of October instead of the end of October 1960 (1st row), and the precipitation of October fell as snow (2nd and 3rd row). 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 row). The hydrological drought ended by some snow melt in March 1961, caused by high temper-

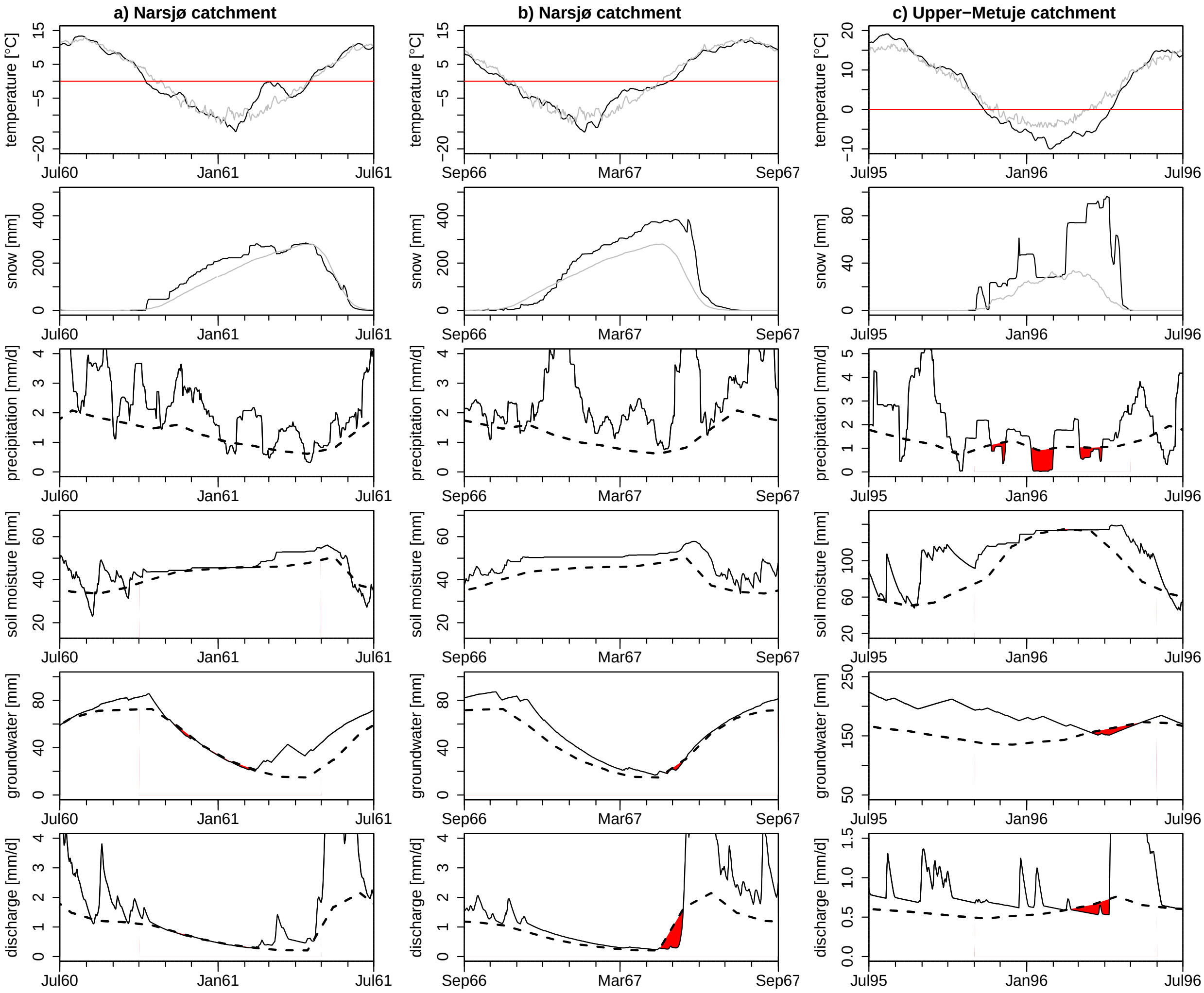

Figure 5.4: Examples of *cold snow season drought* type: **(a)** Narsjø catchment 1960–1961, **(b)** Narsjø catchment 1966–1967, **(c)** Upper-Metuje catchment 1995–1996 (legend: see Fig. 5.1).

atures (1st row). *Cold snow season droughts-subtype A* usually have a long duration (several months), but a low deficit volume and small maximum deviation because groundwater storage and discharge are just below the threshold level. In the example in Fig. 5.4a, 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 occurs 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. 5.4b (Narsjø catchment). In this case, temperature stayed below zero until the beginning of May instead of mid-April (three weeks later than normal; 1st row) and snow melt was delayed (2nd row). Threshold levels started to increase by mid-April, while groundwater storage and discharge still showed a recession (5th and 6th row). When temperature finally reached values above zero in the beginning of May (1st row), snow melt (2nd row) ended the hydrological drought (5th and 6th row). *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 the artefacts described in Sect. 2.4.2. These artefacts do not have an abnormal temperature pattern, but are only caused by a very sharp increase in discharge in combination with a gradually rising threshold level.

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. Thus, 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. 5.4c (Upper-Metuje catchment). From December 1995 to April 1996 temperatures were lower than normal (on average -3.9 °C instead of -0.4 °C; 1st row) and snow accumulation was higher than normal (2nd row). The lack of recharge caused a decrease in groundwater storage and discharge, leading to drought in discharge (6th row) in mid-February and to drought in groundwater (5th row) in mid-March. The drought ended by 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. 5.4c).

Stahl and Demuth [1999] and Pfister et al. [2006] mention a cold winter as a reason for drought, but do not describe the underlying processes. Van Lanen et al. [2004a] discuss causative mechanisms of various *cold snow season droughts*.

5.4.5 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, as in the *cold snow season drought*, only influences the beginning and the end of the snow season. If temperatures

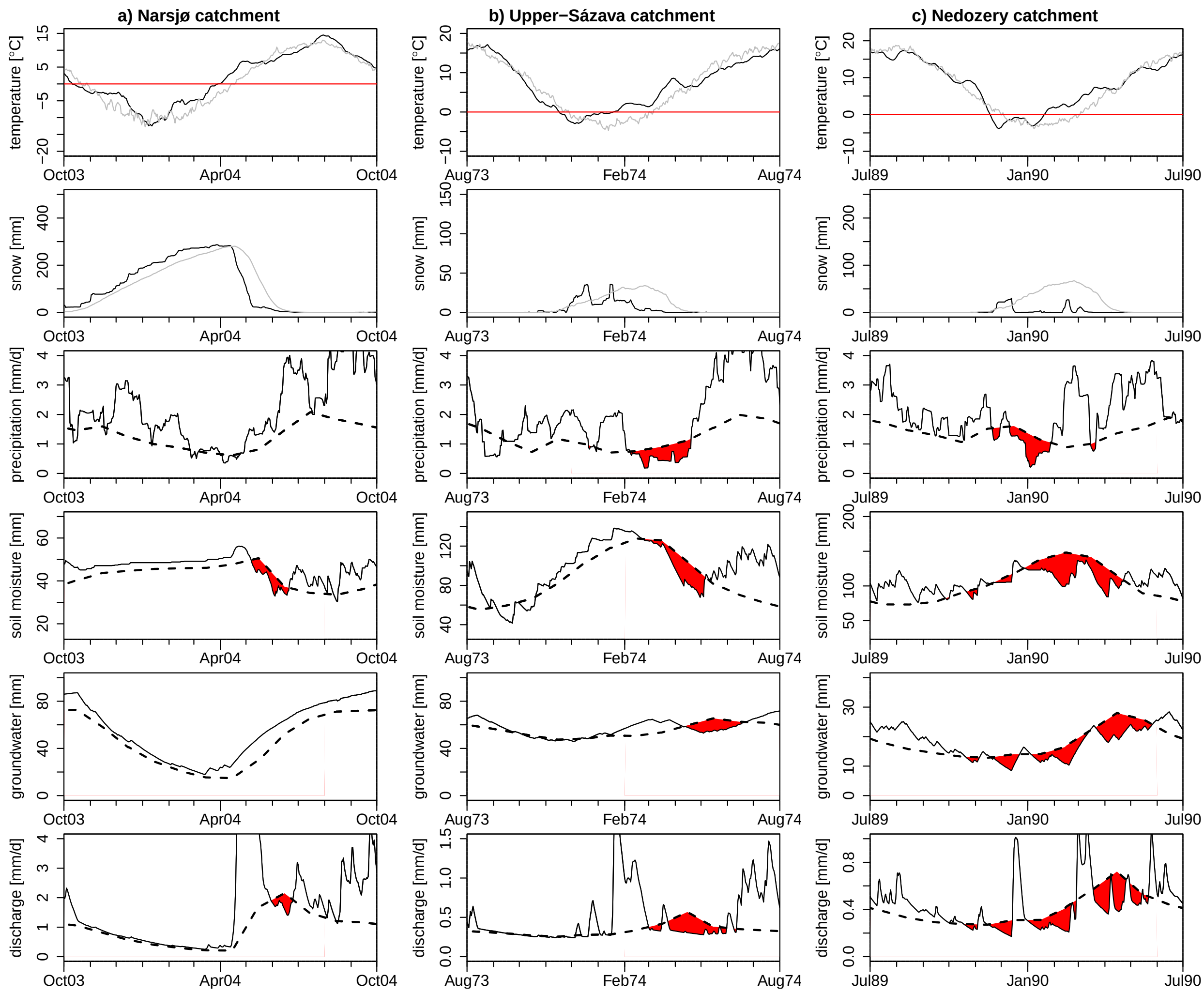

Figure 5.5: Examples of *warm snow season drought* type: **(a)** Narsjø catchment 2003–2004, **(b)** Upper-Sázava catchment 1973–1974, **(c)** Nedožery catchment 1989–1990 (legend: see Fig. 5.1).

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. 5.5.3). 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. 5.5a (Narsjø catchment), temperature increased to above zero three weeks earlier than normal, at the end of March 2004 instead of mid-April (1st row), resulting in an early snow melt (2nd row). Consequently, the peak in discharge (normally in June) was advanced to April–May and in June a hydrological drought developed (6th row), because threshold levels were high and discharge already decreased after the snow melt peak. Thus, a *warm snow season drought-subtype A* can develop without a meteorological drought (although precipitation was not extremely high in May 2004; Fig. 5.5a). The reason is the normally-occurring pronounced snow melt peak in cold climates that is clearly reflected in the threshold level. *Warm snow season droughts-subtype A* usually have short durations (in the example in Fig. 5.5a, 25 days). Deficit volumes can be high (in the example 8.2 mm) due to the high threshold level. 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 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. 5.5. In the first example (Fig. 5.5b, Upper-Sázava catchment), the warm and dry period of February–March 1974 (1st and 3rd row) caused a complete melt of the snow cover (2nd row) and afterwards a lack of recharge to groundwater. Consequently, a hydrological drought developed (5th and 6th row) that continued until the high rainfall period in the spring of 1974 (3rd row). In the second example (Fig. 5.5c, Nedožery catchment), the high temperatures of December 1989 to March 1990 (1st row) also led to a complete melt of the snow cover (2nd row). The meteorological drought of December 1989–January 1990 (3rd row) therefore triggered a soil moisture (4th row) and hydrological drought (5th and 6th row). The rainfall peak in March 1990 (3rd row) caused a quick reaction in discharge (6th row), 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 row). 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. 5.4.2, 5.4.4, and 5.4.5), the *warm snow season drought-subtype B* is not ended by a snow melt peak, because snow cover had already melted earlier. A *warm snow season drought-subtype B* can continue into summer. Durations can be long and deficit volumes can be high. *Warm snow season droughts-subtype B* can show all propagation features (i.e. pooling, lag, attenuation, and lengthening; see Sect. 5.1), mainly dependent on catchment characteristics.

The *warm snow season drought-subtype A* has previously been described by Van Lanen et al. [2004a], and *subtype B* in Ch. 4 under the name Type 2 winter drought.

5.4.6 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 a different type) in distinct seasons cannot be distinguished any more. The main feature of the *composite drought* is that the system

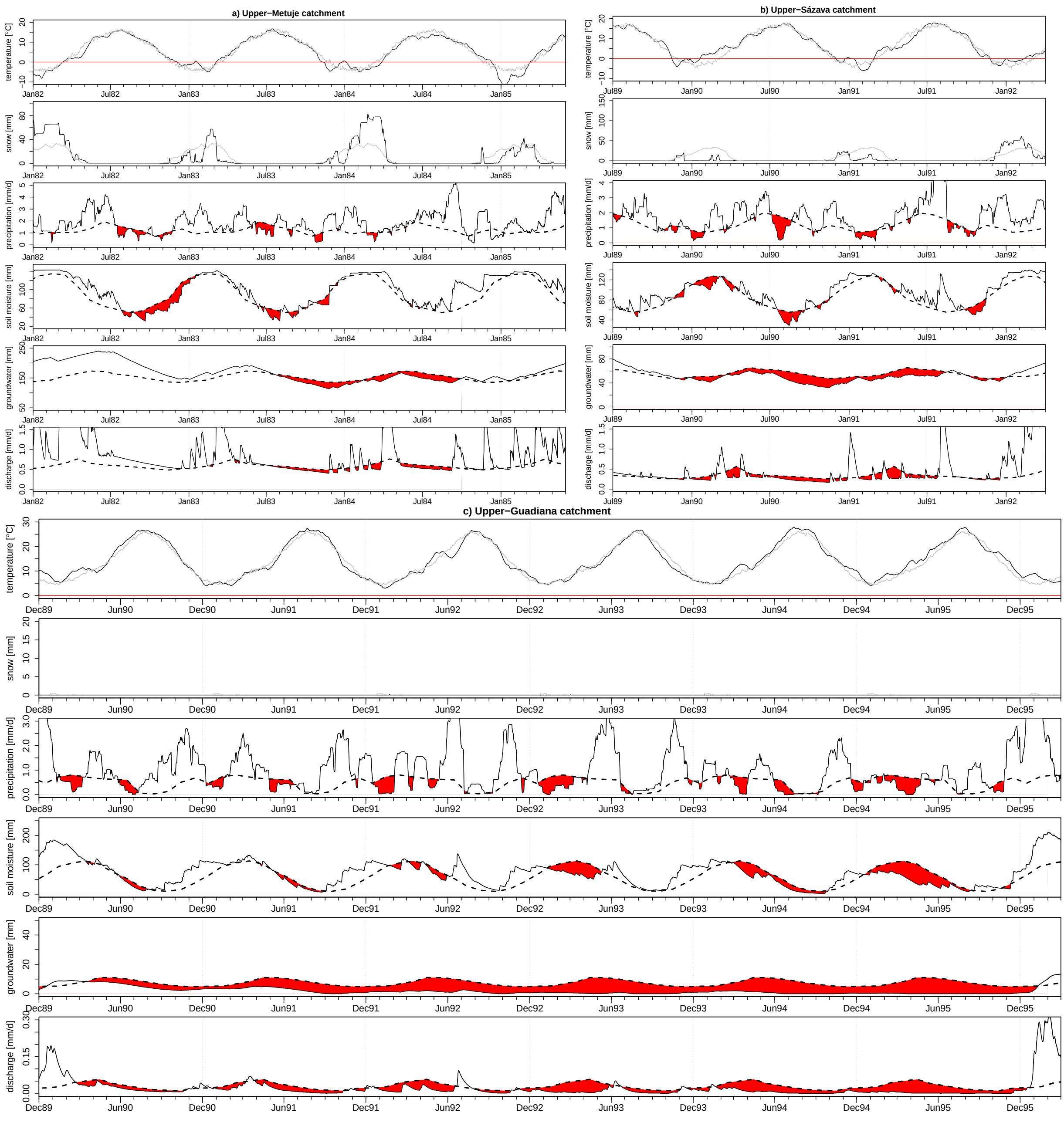

Figure 5.6: Examples of *composite drought* type: **(a)** Upper-Metuje catchment 1982–1985, **(b)** Upper-Sázava catchment 1989–1992, **(c)** Upper-Guadiana catchment 1989–1995 (legend: see Fig. 5.1).

has not yet recovered from a hydrological drought event when the next event starts.

Examples of the *composite drought* are shown in Fig. 5.6. The first example (Fig. 5.6a, Upper-Metuje catchment) shows two *classical rainfall deficit droughts* in subsequent summers (1982 and 1983, 3rd row) that are combined into one hydrological drought (5th and 6th row). The drought in groundwater started in July 1983 and lasted 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 by high precipitation events by the end of 1984. In the second example (Fig. 5.6b, Upper-Sázava catchment), the hydrological drought that lasted from December 1989 to August 1991 (5th and 6th row) was caused by two *warm snow season droughts-subtype B* in the winter of 1989–1990 and 1990–1991 (1st, 2nd and 3rd row) and a *classical rainfall deficit drought* in the summer of 1990 (3rd row). 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. 5.6c, Upper-Guadiana catchment), a large number of *classical rainfall deficit droughts* (3rd row) and *wet-to-dry-season droughts* (3rd and 4th row) in subsequent years are combined into a very long hydrological drought (5th and 6th row). The drought in groundwater lasted 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, that is to say 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 to occur in (semi-)arid climates (Köppen-Geiger climate type B) due to the irregular rainfall pattern in these climates. The drought types that are combined differ per catchment and per climate zone. *Composite droughts* have long to very long durations (often multi-year) and deficit volumes are high (for the examples in Fig. 5.6, 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 Bierkens and Van den Hurk [2007] and Marsh et al. [2007], and analysed by Van Loon et al. [2011a] under the name Multi-year drought.

5.5 Application of the hydrological drought typology in the study catchments

As an example of the application of the hydrological drought typology, we classified drought events in the study catchments (Sect. 5.2). Knowledge of the occurrence of drought types in a catchment is valuable information for water managers. In water management, it is not only useful to know the typology of all drought events, but especially the typology of the most severe events and also the development of non-drought events (the situations where a meteorological drought did not result in a hydrological drought).

5.5.1 Typology of all drought events

Some of the hydrological drought types defined in Sect. 5.4 occurred 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 5.3 shows that the *classical rainfall deficit drought* occurred in all studied catchments and the *wet-to-dry-season*

Table 5.3: 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 %

Table 5.4: Drought types of the five 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 %

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 showed a comparable distribution over the drought types (Table 5.3). Droughts in discharge only showed up in more categories than droughts in groundwater, because the total number of droughts in discharge was higher (Table 5.2), resulting in a higher possibility for different drought types to occur. In groundwater, these drought events have grown together and formed a *composite drought*. Consequently, the percentage of *composite droughts* in groundwater was, in general, higher than that of discharge (Table 5.3; exception Upper-Sázava). Furthermore, *warm snow season droughts* were 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 5.3). This is the most commonly occurring hydrological drought type in these catchments. Only in the groundwater drought events of the Upper-Guadiana catchment, the *classical rainfall deficit drought* was not recognisable any more because it was 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 5.3). 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 types; Table 2.1).

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 caused by an early start of the snow season (*subtype A*) or a late end (*subtype B*). The *cold snow season droughts* in

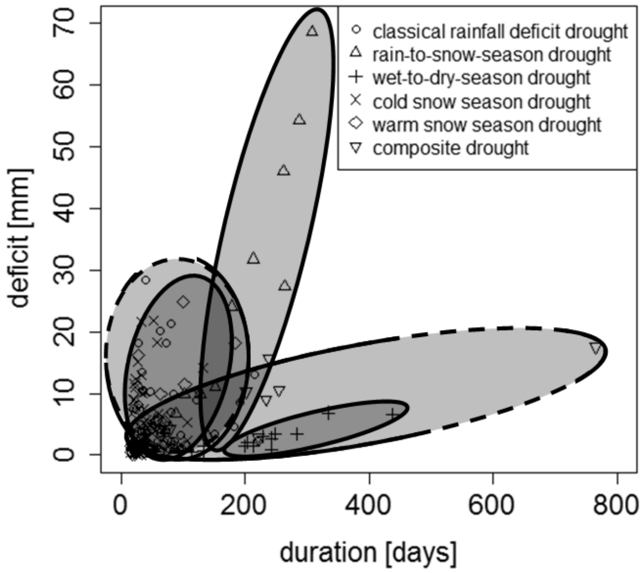

Figure 5.7: Drought duration and deficit volume of all discharge drought events grouped per hydrological drought type (ellipses are added to more clearly identify groups of events with similar drought type; dashed lines indicate an approximation based on a single event).

the Upper-Metuje, Upper-Sázava, and Nedožery catchments are mostly due to a lack of recharge in winter (*subtype C*) and sometimes due to a late ending 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 (i.e. Upper-Metuje, Upper-Sázava, Nedožery), highest percentages of *warm snow season droughts* were found (around 20 % occurrence). These were all *subtype B* droughts.

The *composite drought* occurred in slow 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 several seasons and even years (Table 5.2) due to the long memory in its extensive groundwater system.

A few events are not included in Table 5.3 (causing percentages of some catchments not to add up to 100 %). In the Narsjø catchment these omitted events are classified as artefacts (and thus disregarded, see Sect. 2.4.2.3) 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.

If drought characteristics of all discharge drought events in the five catchments studied are grouped by drought type (Fig. 5.7), 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 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. 5.8, the same discharge drought events are plotted in more detail (one plot for each drought type and a different colour for fast responding and slow responding catchments). For each drought type, the events in slow responding catchments have, in general, somewhat longer

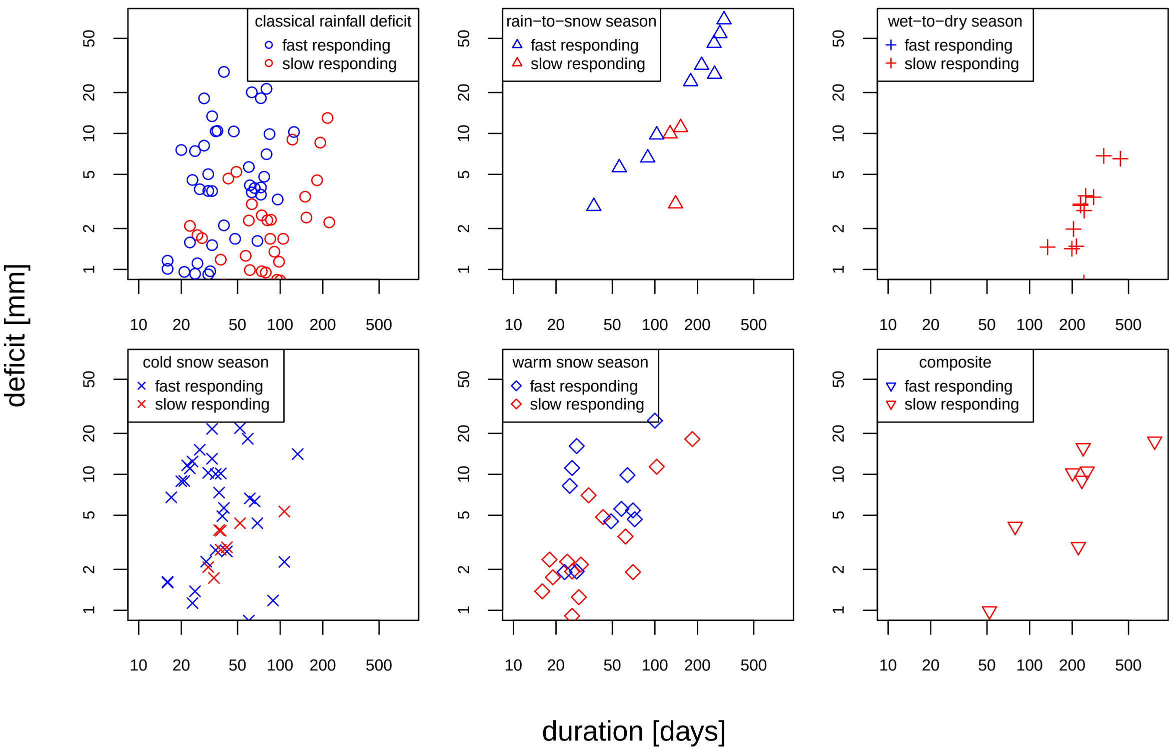

Figure 5.8: Drought duration and deficit volume of all discharge drought events grouped per hydrological drought type, on log-log scale, differentiating between fast responding and slow responding catchments (fast responding: Narsjø and Nedožery catchments; slow responding: Upper-Metuje, Upper-Sázava, and Upper-Guadiana catchments).

durations and lower deficit volumes than those in fast responding catchments. *Wet-to-dry-season droughts* and *composite droughts* were only found in slow responding catchments. *Composite droughts* do not occur in fast responding catchments. *Wet-to-dry-season droughts* presumably do occur in fast responding catchments, but in this study no fast responding catchment with semi-arid climate was included.

5.5.2 Typology of the most severe drought events

Because Table 5.3 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 based on maximum deviation for groundwater and on deficit volume for discharge. Table 5.4 shows that the distribution of hydrological drought events over the different drought types changed 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-lasting and can build up a large deficit volume. For the same reason *composite droughts* are represented more in the most severe drought events.

If drought events had have been classified according to their duration and the five longest drought events selected, the distribution over the drought types would have been similar to Table 5.4 (not shown).

Based on Table 5.4, 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 fast responding catchments: *classical rainfall deficit drought*;
- in slow responding catchments: *composite drought*.

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

5.5.3 Non-drought development

Up till 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 in which 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 from developing. One example is when 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 had 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. 5.4.5). A warm winter can also have another effect in snow climates – namely a late start of the snow season (Sect. 5.4.5). This can prevent a *rain-to-snow-season drought* from developing. An example is shown in Fig. 5.9a (Narsjø catchment). The rainfall deficit in September 2000 (3rd row) resulted in just-below-threshold levels in groundwater storage and discharge (5th and 6th row). If temperatures would have dropped below zero in October, like they normally do, the precipitation peak in October–November 2000 (3rd row) 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 row), 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 slow responding catchments, attenuation is a well-known drought propagation feature (Fig. 1.3). Meteorological drought events are often attenuated in the stores and no hydrological drought develops. An example is shown in Fig. 5.9b (Upper-Guadiana catchment). The rainfall deficit in February 1961 (3rd row) led to a drought in soil moisture (4th row) and to a decrease in groundwater levels and discharge (5th and 6th row), but high groundwater storage prevented both variables from falling below threshold level. If antecedent storage would have been low, a *wet-to-dry-season drought* would have developed, as in the examples in Fig. 5.3. Attenuation of a meteorological drought can also occur in fast 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. 5.9c (Nedožery catchment; 3rd row) would have developed into a *classical rainfall deficit drought*, but due to the very wet condition of the catchment after extensive rainfall in the previous months (5th and 6th row), the recession of groundwater storage and discharge did not drop below the threshold level.

Also a combination of processes can prevent a meteorological drought from developing into a hydrological drought. The example in Fig. 5.9d (Upper-Metuje catchment) could have become a *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.

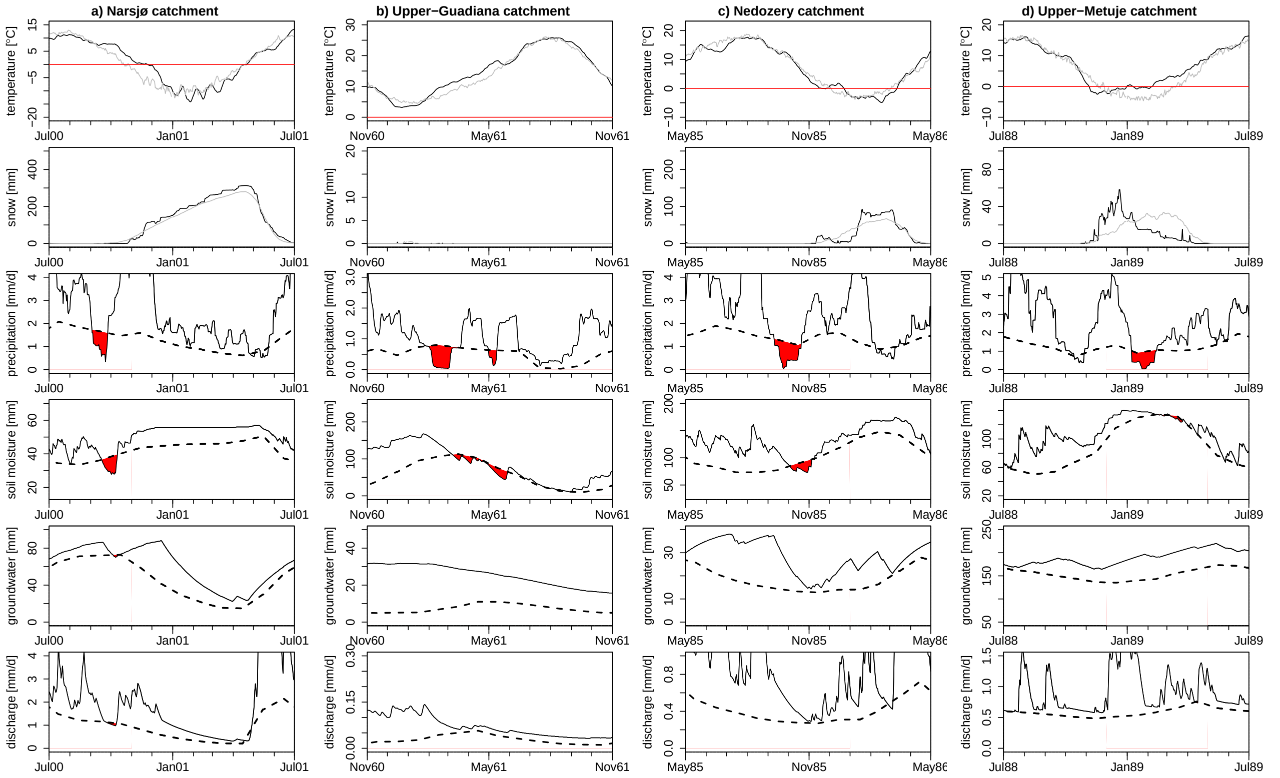

Figure 5.9: Examples of non-drought events: **(a)** Narsjø catchment 2000–2001, **(b)** Upper-Guadiana catchment 1960–1961, **(c)** Nedožery catchment 1985–1986, **(d)** Upper-Metuje catchment 1988–1989 (legend: see Fig. 5.1).

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.

5.6 Discussion

5.6.1 Typology

In this chapter we proposed a hydrological drought typology based on drought propagation processes. Table 5.5 summarises the processes that govern the six hydrological drought types.

Because the division into different 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 event. 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 a 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 several 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. 5.1). In Sects. 5.3.2, 5.4, and 5.5, 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.

Table 5.5: 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	D, E
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/or T control	A, B, C, D, E

5.6.2 Catchment control

For drought propagation catchment control is very important (see Sect. 1.2.4). 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, and the percentage of lakes and forest [Demuth and Young, 2004]. These propagation features are represented in all hydrological drought types, but show up most prominently in *composite droughts*. In Sect. 5.5 we saw that *composite droughts* only occur in slow responding catchments and that this drought type is among 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).

It is very striking 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. 5.5 and 5.6). 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.

Fig. 5.10 shows that the *composite drought* is the only drought type that is primarily controlled by catchment characteristics (the x-axis in Fig. 5.10). The other drought types are mainly controlled by climate (the y-axis in Fig. 5.10).

5.6.3 Climate control

The effect of climate on hydrological drought types (see Sect. 1.2.3) can be distinguished in 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 Sects. 5.4 and 5.5, Table 5.5 (last column), and Fig. 5.10 (y-axis). *Classical rainfall deficit droughts* occur in all climates and *wet-to-dry-season droughts* only in climates with strong seasonal variation in precipitation. The three snow-related drought types occur in a similar range of climates from temperate to continental and polar (Fig. 5.10).

The hydrological drought typology is developed using five catchments with different climates in Europe. These catchments are indicated in Fig. 5.10, based on their climate and catchment characteristics. The data of the studies mentioned in Sect. 5.4 could not be included in Fig. 5.10, because insufficient information on catchment and climate control was provided in these papers.

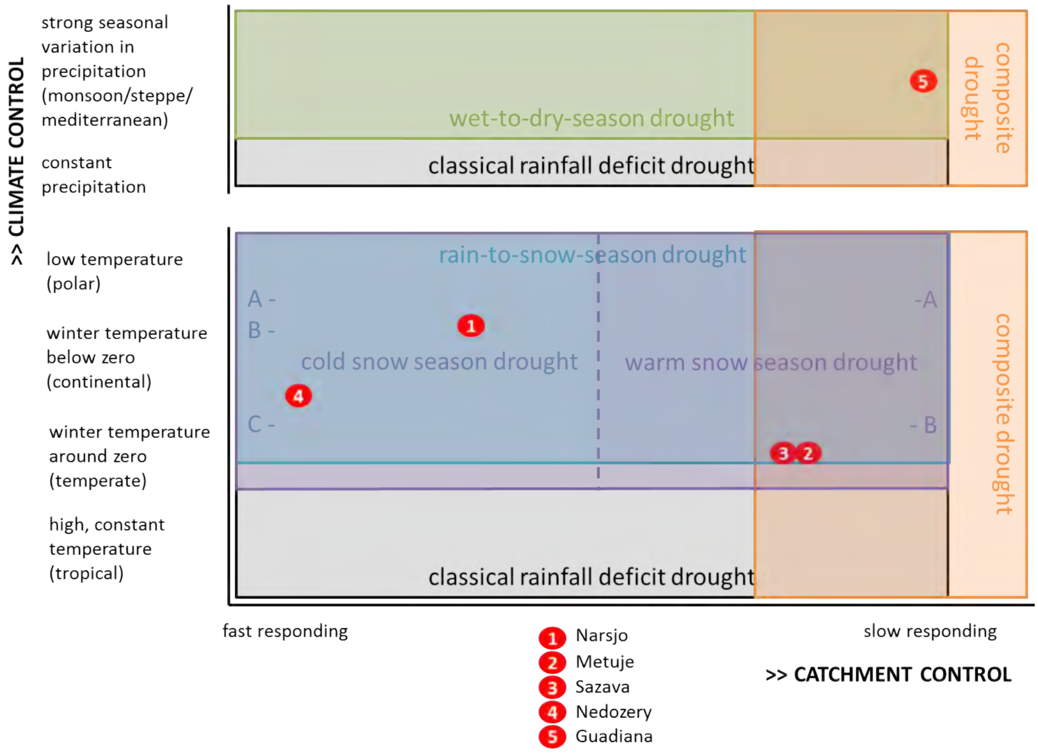

Figure 5.10: Hydrological drought (sub-)type occurrence in relation to catchment and climate control. Catchment control is indicated by a slower response of discharge to precipitation when moving from left to right on the x-axis. Climate control is indicated by describing temperature and precipitation regimes relevant for drought development: temperature on the lower part of the y-axis, precipitation on the upper part of the y-axis [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 on the basis of their climate and catchment characteristics (see Sect. 2.1); for explanation of the drought (sub-)types see Table 5.5.

Because the typology is based on generally observable processes, it can be used in catchments that fall outside the reach of the catchments studied here (for example in the upper-left part of Fig. 5.10). Adding more catchments with different climate and catchment characteristics to the framework of Fig. 5.10 is an interesting way forward in drought research. Focus can then be on e.g. tropical climates and fast responding catchments in steppe or monsoon climates. This can be achieved by using data of real catchments or synthetic data, following the approach of Van Lanen et al. [2012]. This newly-developed approach also allows for a better quantification of the effect of catchment and climate control on drought propagation and drought typology (see Ch. 7).

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 5.5 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 few drought events alternate with periods with many drought events, which is consistent with the results of other studies [Stahl and Hisdal, 2004; Uhlemann et al., 2010]. In central Europe, for example, the first half of the

1980s, the 1990s, and the 2000s 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. 5.5.3), 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. 5.5.3), or they can cause the end of a drought (in case of drought types not related to snow, e.g. Sect. 5.4.1).

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], and 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 a 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 falls in the form of snow, while in the lower catchment the hydrological drought ceases due to rainfall. Synchronicity of droughts within a region, therefore, mainly happens in the case of drought types that are precipitation controlled (i.e. *classical rainfall deficit drought* and *wet-to-dry-season drought*) and less in the case of 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. A large elevation range is also the reason that discharge peaks can occur when the catchment-average temperature is still below zero.

In this study, potential evaporation was 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 a lack of water available for evaporation. In regions with very high water availability (e.g. some subtypes of Köppen-Geiger climate type A) an increase in potential evaporation might have more influence [Van Lanen et al., 2004a]. For the presented drought typology, potential evaporation is only important in a 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 used principally to refer 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 that occurs in the winter half year [Pfister et al., 2006].

Climate change will probably lead to a change in the 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 is normally ended by a snow melt peak might change into a drought type that can continue into summer (see Ch. 4).

5.7 Conclusions

In this chapter we proposed a general hydrological drought typology based on the processes underlying 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 are distinguished are: (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*.

- *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 they 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 they occur in catchments with pronounced wet and dry seasons.
- *Cold snow season droughts* are caused by low temperatures in the snow season. In catchments with very cold winters, *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 high temperatures in the snow season. In catchments with very cold winters, *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 to occur in (semi-)arid climates and slow responding catchments.

About 125 groundwater droughts and 210 discharge droughts of five contrasting headwater catchments in Europe have been classified using the developed typology. The most common drought type in all catchments was the *classical rainfall deficit drought* (almost 50 % of all events), but these are mostly minor events. When only the five most severe drought events of each catchment were considered, a shift towards more *rain-to-snow-season droughts*, *warm snow season droughts*, and *composite droughts* was 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 identification of the occurrence of hydrological drought types in relation to climate and catchment characteristics. Herewith, we hope to contribute to process understanding of drought propagation and the improvement of drought forecasting and management all around the world.

Chapter 6

Drought propagation in large-scale models

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6.1 Introduction

Drought studies on global or continental scales increasingly make use of large-scale models, both land-surface models (LSMs) and global hydrological models (GHMs) [Andreadis et al., 2005; Lehner et al., 2006; Sheffield and Wood, 2008; Mishra and Singh, 2011; Wang et al., 2011; Stahl et al., 2012b]. However, little is known of the performance of large-scale models in simulating drought development in the large variety of climate zones and catchments around the world [Gudmundsson et al., 2012]. Simulating low flow and drought is a challenge, even for catchment-scale models (see Sect. 2.3). So the question is how well do large-scale models perform for low flows and drought? A evaluation of large-scale models is needed to estimate the uncertainty related to drought simulation using large-scale models and to guide further improvement of these models. Some first steps in the evaluation of drought simulation by large-scale models are set by Prudhomme et al. [2011]; Stahl et al. [2011a, 2012b], and Gudmundsson et al. [2012]. They looked at trends and general patterns/statistics of low flows, but most of them did not take into account actual timing and duration of drought events. Only Prudhomme et al. [2011] investigated timing and duration of drought events. However, like Stahl et al. [2011a, 2012b] and Gudmundsson et al. [2012], they focused solely on runoff. Drought propagation from meteorological to hydrological drought was not taken into account. Hence the simulation of processes underlying hydrological drought development (i.e. drought propagation, Fig. 1.1) by large-scale models is not yet evaluated. With this study we take a first step towards filling this gap. A correct simulation of these processes is needed, so that we know that large-scale simulations are robust when extrapolating to data-scarce regions [e.g. Stahl et al., 2012b] or to the future [e.g. Gosling et al., 2011; Corzo Perez et al., 2011b].

In this study drought is defined as a sustained and regionally extensive period of below-normal water availability (Sect. 1.2.1). We focus on the development of hydrological drought, which is a drought in groundwater and/or discharge (Fig. 1.1). Hydrological drought is a recurring and worldwide phenomenon, with spatial and temporal characteristics that vary significantly from one region to another [Tallaksen and Van Lanen, 2004]. Some of the most studied drought characteristics are the number of droughts, drought duration and drought deficit [Hisdal et al., 2004; Fleig et al., 2006; Sheffield and Wood, 2011]. Not only drought characteristics vary per region, there is also a variation in the way a drought propagates from a precipitation and/or temperature anomaly to a hydrological drought around the world (Tallaksen and Van Lanen [2004]; Mishra and Singh [2010] and Ch. 5). The flow chart in Fig. 1.1 demonstrates the propagation of drought and the way it depends on meteorological factors [similar illustrations can be found in e.g. Changnon Jr, 1987; Tallaksen and Van Lanen, 2004; Sheffield and Wood, 2011]. On the basis of the results of Chs. 4 and 5 the effect of temperature in cold climates can be added, hence a distinction can be made between rain and snow seasons in these climates. Despite these different ways that a hydrological drought can develop from the meteorological situation, some drought propagation features are common to all hydrological droughts [Eltahir and Yeh, 1999; Peters et al., 2003; Van Lanen et al., 2004a; Van Loon et al., 2011b]: pooling, attenuation, lag and lengthening (see Sect. 1.2.2). These drought propagation features manifest themselves in different ways dependent of catchment characteristics and climate [Van Lanen et al., 2004a, 2012]. This results in different hydrological drought types, dependent of the interplay between precipitation, temperature and catchment characteristics. In Ch. 5 we distinguish six hydrological drought types: (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 elements of drought propagation, i.e. drought characteristics (Sect 2.4.2), drought propagation features (Sect. 1.2.2) and drought typology (Sect. 5.7), can be used as tools to evaluate the simulation of drought propagation by large-scale models. In hydrology often a single large-scale model is used with its specific advantages and disadvantages [e.g. Lehner et al., 2006;

Sheffield and Wood, 2007; Döll et al., 2009; Hurkmans et al., 2009; Mishra and Singh, 2010; Sutanudjaja et al., 2011]. In several studies, however, the multi-model ensemble of a number of large-scale models was closer to observations than most participating models individually, both in general hydrological studies [e.g. Gao and Dirmeyer, 2006; Guo et al., 2007] and in low-flow and drought research [e.g. Gudmundsson et al., 2012; Stahl et al., 2011b]. Therefore, in this study, we investigated a multi-model ensemble, as was previously done in some other drought studies [Wang et al., 2009, 2011; Gudmundsson et al., 2012; Stahl et al., 2012b; Van Huijgevoort et al., 2012a]. The aim of this chapter is explicitly not to compare individual models or model approaches, but to see whether large-scale models in general can reproduce drought propagation. Therefore the outcomes of individual models are not shown; only the multi-model ensemble with ranges of daily minimum and maximum is presented.

The objective of this study is to evaluate the simulation of drought propagation in large-scale hydrological models. To reach this objective we used a global meteorological dataset (Sect. 6.2.1.1), hydrological data from an ensemble of ten large-scale models (Sect. 6.2.1.2), we selected a number of case study areas with contrasting climate and catchment characteristics (Sects. 6.2.2.1 and 6.2.2.2) and we studied drought development in those areas in detail (Sects. 6.2.2.3 and 6.2.2.4). Focus hereby is not on individual drought events, but on general phenomena, i.e. (i) drought characteristics (Sect. 6.3.1), (ii) drought propagation features (Sect. 6.3.2), and (iii) drought typology (Sect. 6.3.3). Individual drought events of specific case study areas are only included as examples to illustrate these general phenomena. In Sect. 6.4 we discuss our methodology and results and in Sect. 6.5 we summarise and conclude this study.

6.2 Data and methods

In this study we used data from a large-scale meteorological dataset and from a suite of large-scale hydrological models. These large-scale data were extracted and post-processed in a number of steps. Subsequently, drought analysis was performed on the hydrometeorological data and the hydrological drought typology was applied to the results.

6.2.1 Large-scale data

6.2.1.1 Meteorological data

The large-scale meteorological data used in this study were obtained from the WATCH Forcing Data [WFD, Weedon et al., 2011] described in Sect. 2.2.1.1. We used the WFD time series of temperature and precipitation to investigate drought propagation. The WFD have also been used to force the large-scale hydrological models [Haddeland et al., 2011] from which output data were used in this study.

6.2.1.2 Hydrological data

The large-scale hydrological data used in this study were obtained from large-scale hydrological models that participated in the model intercomparison project (WaterMIP) of WATCH (www.eu-watch.org), which is described by Haddeland et al. [2011]. Data of ten large-scale hydrological models have been provided, i.e. GWAVA, H08, HTESSEL, JULES, LPJmL, Mac-PDM, MATSIRO, MPI-HM, Orchidee, and WaterGAP (Table 6.1). All models were run at 0.5° spatial resolution for the global land area for a 38-yr period (1963–2000), with a 5-yr spin-up period (1958–1962).

Based on the type of model (LSM/GHM) and its development history, the large-scale models use different variables from the WFD as input (Table 6.1) and have different schemes for calculating evapotranspiration, snow accumulation and melt, and runoff [Haddeland et al., 2011;

Table 6.1: Main characteristics of the participating models [derived from Haddeland et al., 2011]

Model name[a]	Input variables (from WFD)[b]	Output variables[c]	Reference(s)
GWAVA	P, T, W, Q, LWn, SW, SP	SM, Q_{sub}, Q_{total}	Meigh et al. [1999]
H08	R, S, T, W, Q, LW, SW, SP	SM, Q_{sub}, Q_{total}	Hanasaki et al. [2008]
HTESSEL	R, S, T, W, Q, LW, SW, SP	SM, Q_{sub}, Q_{total}	Balsamo et al. [2009]
JULES	R, S, T, W, Q, LW, SW, SP	SM, Q_{sub}, Q_{total}	Best et al. [2011]; Clark et al. [2011]
LPJmL	P, T, LWn, SW	SM, GW, Q_{sub}, Q_{total}	Bondeau et al. [2007]; Rost et al. [2008]
Mac-PDM	P, T, W, Q, LWn, SW	GW, Q_{sub}, Q_{total}	Arnell [1999]; Gosling and Arnell [2011]
MATSIRO	R, S, T, W, Q, LW, SW, SP	SM, Q_{sub}, Q_{total}	Takata et al. [2003]; Koirala [2010]
MPI-HM	P, T	SM, Q_{sub}, Q_{total}	Hagemann and Gates [2003], Hagemann and Dümenil [1998]
Orchidee	R, S, T, W, Q, SW, LW, SP	SM, Q_{sub}, Q_{total}	De Rosnay and Polcher [1998]
WaterGAP	P, T, LWn, SW	SM, GW, Q_{sub}, Q_{total}	Alcamo et al. [2003]

(a) Model names written in bold are classified as LSMs; the other models are classified as GHMs.
(b) R: Rainfall rate, S: Snowfall rate, P: Precipitation (rain or snow distinguished in the model), T: air temperature, W: Wind speed, Q: Specific humidity, LW: Longwave radiation (downward), LWn: Longwave radiation (net), SW: Shortwave radiation (downward), SP: Surface pressure.
(c) SM: Soil moisture storage, GW: Groundwater storage, Q_{sub}: Subsurface runoff, Q_{total}: Total runoff (subsurface runoff + surface runoff).

Gudmundsson et al., 2012]. LSMs and GHMs were run on different time steps and, after simulation, sub-daily data were aggregated to daily data. The model time step is not expected to influence drought simulation, in contrast with model structure, which is of paramount importance (see Sect. 6.4.2).

Human impacts such as reservoir operation and water withdrawals for agriculture or drinking water were not included in the model output we used for this study (i.e. natural situation). The large-scale models have not been calibrated for WaterMIP, except WaterGAP, for which correction factors were applied in some major river basins [e.g. Alcamo et al., 2003; Hunger and Döll, 2008]. More details of the models can be found in Haddeland et al. [2011] and Gudmundsson et al. [2012], or in the references listed in Table 6.1.

Output variables used in this study include the main water balance states and fluxes on daily time scale: soil moisture storage (SM), groundwater storage (GW), subsurface runoff (Q_{sub}), and total runoff (Q_{total} = surface runoff + subsurface runoff). Soil moisture data were only available for nine models, groundwater storage only for three models (see Table 6.1). In the models that explicitly simulate groundwater storage, subsurface runoff reflects baseflow. In the other models subsurface runoff is drained from the soil storage and reflects a slow runoff component.

6.2.2 Methodology

6.2.2.1 Extraction of data for case study areas

To investigate whether drought propagation from an anomaly in precipitation/temperature (meteorological situation in Fig. 1.1) to groundwater/runoff (hydrological drought in Fig. 1.1) is adequately reproduced by large-scale models, time series of model results need to be studied. Only a limited number of case study areas can be studied in detail, and prior knowledge of drought propagation in the selected case study areas is essential for a proper evaluation of the models. For example, Gudmundsson et al. [2012] concluded that the limitation of their study was the loss of information due to spatial aggregation in data processing. Therefore, in this study, a limited selection of case study areas was used that corresponds to catchments that have been studied in previous research (Van Huijgevoort et al. [2010] and Chs. 4 and 5). These catchments are restricted to Europe, but the conclusions drawn with regard to the studied catchments have a wider validity, because of their contrasting climate and catchment characteristics and the general phenomena that were studied.

Table 6.2: Grid cell characteristics of the selected catchments Narsjø (Norway), Upper-Metuje and Upper-Sázava (Czech Republic), Nedožery (Slovakia), and Upper-Guadiana (Spain)

grid cell	Narsjø	Upper-Metuje	Upper-Sázava	Nedožery	Upper-Guadiana
latitude	62.25	50.75	49.75	48.75	39.25
longitude	11.75	16.25	15.75	18.75	−3.75
area of catchment within grid cell	72 %	100 %	91 %	100 %	14 %
area of grid cell covered by catchment	6 %	4 %	6 %	9 %	99 %
altitude [$m\,a.m.s.l.$]	785	446	461	580	740

From the gridded large-scale meteorological and hydrological datasets mentioned in the previous section, we selected five case study areas for a detailed drought propagation research. These are the Narsjø catchment in Norway, the Upper-Metuje and Upper-Sázava catchment in Czech Republic, the Nedožery catchment in Slovakia, and the Upper-Guadiana catchment in Spain. These case study areas correspond to natural (or naturalised) headwater catchments in Europe with contrasting climate and catchment characteristics (see Sect. 2.1).

One grid cell completely covers the Upper-Metuje catchment. The same holds for the Nedožery catchment, whereas for the Narsjø and Upper-Sázava, of the two grid cells covering the catchment, the one with the highest coverage was used (72 % and 91 %, respectively; Table 6.2). Of the grid cells covering the Upper-Guadiana catchment, the one closest to the outlet of the catchment was used, representing 14 % of the catchment (Table 6.2). A number of other grid cells from this catchment were also studied (including one with a Bsk-climate instead of a Csa-climate), but the results were not significantly different. The time series of hydrological variables, the drought characteristics, and the conclusions drawn with regard to the performance of the large-scale models in simulating drought propagation processes all were similar.

We are aware that caution should be taken when comparing large-scale models with observations on the scale of one single grid cell. In this study we therefore did not compare model output with observations. Instead, we studied the most important processes underlying drought propagation in the example catchments and compared the results with what is generally known of drought propagation and with the results of catchment-scale models, described in Ch. 5. Comparisons of large-scale model(s) with observations have been performed previously by Van Loon et al. [2011b] and Stahl et al. [2011a, 2012b]. Van Loon et al. [2011b] did a qualitative assessment of the regime of the ensemble mean of a comparable set of large-scale models for four of the case study areas that were also used in this study. They concluded that the most important characteristics of those regimes, i.e. low flows and snow melt peaks, were reproduced by the large-scale models. This gives confidence that large-scale models can be used for drought analysis in these case study areas. Stahl et al. [2011a] compared anomaly indices in a large number of small catchments in Europe, some being represented by a single grid cell and some by more than one grid cell (up to nine cells). They found no significant correlations of anomaly indices with area, and thus ruled out a scaling effect. Hence, small catchments can be represented by a single grid cell, as long as the elevation difference between model and observations is not too high [in Stahl et al., 2011a, less than 300 m].

6.2.2.2 Post-processing

We processed the data of the selected case study areas through a number of steps:

1. interpolation of NA-values of leap days,

2. standardisation of state variables SM and GW by dividing the data by the long-term average [needed because of huge inter-model differences in reference level Wang et al., 2009],

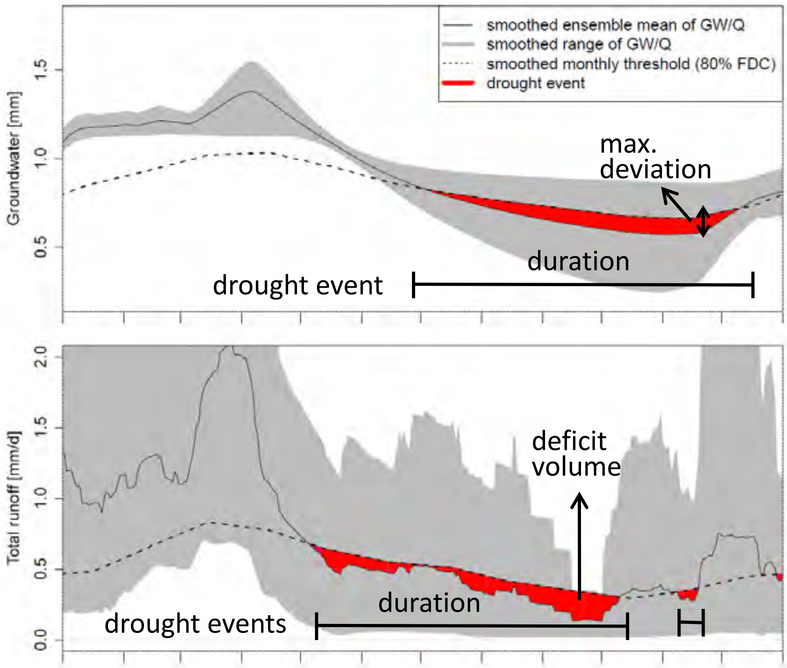


Figure 6.1: Threshold level method with variable threshold (80th percentile of monthly duration curve, smoothed by 30-day moving average) for groundwater storage (GW; state variable; upper row) and total runoff (Q; flux; lower row), including an illustration of drought characteristics duration, deficit volume, and maximum deviation (see Fig. 2.7).

3. calculation of the ensemble mean of all models for SM, GW, Q_{sub}, and Q_{total} (nine models for SM, three for GW, and ten for Q_{sub} and Q_{total}; see Table 6.1),

4. calculation of the daily maximum and minimum value of all models for SM, GW, Q_{sub}, and Q_{total} to determine model range,

5. smoothing the daily ensemble mean, maximum, and minimum of SM, GW, Q_{sub}, and Q_{total} by applying a 30-day centred moving average [the necessity of smoothing when using large-scale models was demonstrated by Van Loon et al., 2011b].

6.2.2.3 Drought analysis

Droughts were identified using the variable threshold method (see Sect. 2.4.2 and Fig. 6.1). This method was applied to all hydrometeorological variables, i.e. smoothed precipitation (from WFD), and the smoothed ensemble mean of SM, GW, Q_{sub}, and Q_{total} (from the large-scale hydrological models). The smoothing (Sect. 6.2.2.2, step 5) was used as a pooling method [Hisdal et al., 2004; Fleig et al., 2006]. The drought characteristics duration, deficit volume, and maximum deviation (Fig. 6.1) are used to illustrate drought propagation [Di Domenico et al., 2010].

6.2.2.4 Typology of hydrological droughts

The hydrological drought typology developed in Ch. 5 was used to study drought propagation processes. This typology (see Table 5.5) was developed using a catchment-scale model that was calibrated against observations. Table 5.5 also includes a column on the influence of precipitation (P) and temperature (T) control on the development of each hydrological drought type. *Classical rainfall deficit droughts* are the only hydrological drought type that is completely

Table 6.3: General drought characteristics using a 80 % monthly threshold (moving average 30 days) and a minimum drought duration of 3 days for the hydrometeorological variables derived from WFD and simulated with the large-scale models for all selected case study areas

		no. of droughts [per year]	mean duration [day]	mean deficit [mm]	mean max.deviation [mm]
Narsjø	precipitation	4.6	16	4.3	–
	soil moisture	1.4	53	–	0.04
	groundwater storage	1.0	70	–	0.07
	subsurface runoff	1.3	57	4.0	–
	total runoff	1.8	42	4.3	–
Upper-Metuje	precipitation	4.9	14	6.1	–
	soil moisture	1.5	45	–	0.05
	groundwater storage	1.0	70	–	0.07
	subsurface runoff	1.0	69	4.6	–
	total runoff	2.5	28	3.8	–
Upper-Sázava	precipitation	4.6	16	6.3	–
	soil moisture	1.4	48	–	0.05
	groundwater storage	0.7	106	–	0.09
	subsurface runoff	0.6	117	7.8	–
	total runoff	2.3	30	3.7	–
Nedožery	precipitation	4.7	15	5.9	–
	soil moisture	1.7	41	–	0.04
	groundwater storage	0.7	99	–	0.07
	subsurface runoff	1.0	66	3.3	–
	total runoff	2.9	24	2.7	–
Upper-Guadiana	precipitation	3.4	19	4.2	–
	soil moisture	1.3	53	–	0.08
	groundwater storage	0.5	159	–	0.11
	subsurface runoff	0.7	107	0.94	–
	total runoff	2.0	36	0.81	–

governed by P control. *Cold snow season droughts* (all subtypes) and *warm snow season droughts-subtype A* are hydrological drought types that are completely governed by T control. *Rain-to-snow-season droughts* and *wet-to-dry-season droughts* are initiated by P control and continued by T control. *Warm snow season droughts-subtype B* are initiated by T control and continued by P control. In the case of *composite droughts*, it depends on the hydrological drought types that are combined whether only P control, only T control, or a combination of P and T control plays a role (Ch. 5).

The application of the drought typology is based on expert knowledge (as in Ch. 5). In the part of this study dealing with typology, subsurface runoff (Q_{sub}) was used as proxy for groundwater, because groundwater storage data were only supplied by three out of ten large-scale models (see Table 6.1).

6.3 Results

In this section, we present the results of the analysis of the large-scale models on drought characteristics, drought propagation features, and drought typology, and we link these results to earlier work on drought propagation. This exercise can be regarded as an evaluation of the large-scale models.

6.3.1 Drought characteristics

General drought characteristics were determined from the large-scale model ensemble mean for all five case study areas (Table 6.3). These drought characteristics reflect aspects of drought propagation and differences in climate:

- Drought events became fewer and longer when moving from precipitation via soil moisture to groundwater storage, i.e. the number of droughts decreased from 3–5 per year to 0.5–1 per year and the duration increased from around 15 days to 70–160 days. The decrease in the number of droughts can be seen in Fig. 6.2e, in which there were more drought events in precipitation (2nd row) than in groundwater (4th row) due to attenuation, and the increase in duration is visualised in Figs. 6.2c and 6.3b and c, in which drought events in precipitation (2nd row) were (more and) shorter than those in groundwater (4th row).

- Drought events in total runoff had drought characteristics in between those of precipitation and groundwater, because total runoff reflects both fast and slow pathways in a catchment. This is visualised in Figs. 6.2 and 6.3, in which the signal of total runoff (lower row) is a reflection of the signals of subsurface runoff (5th row) representing slow pathways and precipitation (2nd row) representing fast pathways.

- Deficit volumes were higher for droughts in precipitation than for droughts in total runoff, because precipitation is higher and more variable, resulting in higher threshold values and a larger deviation from the threshold (compare 2nd and lower rows in Figs. 6.2 and 6.3). The exception was Narsjø, which had a slightly lower variability in precipitation and a slightly higher variability in total runoff than the other case study areas, resulting in a similar mean deficit (i.e. 4.3 mm; Table 6.3).

- Drought characteristics of subsurface runoff were comparable to those of groundwater storage (although a different number of large-scale models was used to calculate the average of both variables; see Table 6.1). In Figs. 6.2 and 6.3, the 4th and 5th row have a comparable number and duration of drought events. In some case study areas, e.g. Narsjø and Nedožery, droughts in subsurface runoff were only slightly more and shorter than those in groundwater storage (Table 6.3). The similarity of both variables also justifies the use of Q_{sub} as a proxy of groundwater storage in the remainder of this research.

- Due to its semi-arid climate Upper-Guadiana had slightly fewer and longer meteorological droughts than the other case study areas (Table 6.3).

These results correspond to those of earlier work on drought propagation [Peters et al., 2003; Tallaksen and Van Lanen, 2004; Di Domenico et al., 2010] and Ch. 5.
The drought characteristics in Table 6.3 also showed some unexpected behaviour:

- Mean maximum deviation was lower for soil moisture droughts than for droughts in groundwater. This was expected to be the other way around (like in Hohenrainer, 2008 and Ch. 5) and is probably due to the standardisation of the values of soil moisture and groundwater (Sect. 6.2.2.2, step 2).

- The drought characteristics of total runoff were in between those of precipitation and soil moisture in all case study areas, while a differentiation between fast and slow responding systems was anticipated. The drought characteristics of total runoff in the slow responding systems Upper-Metuje, Upper-Sázava, and Upper-Guadiana were expected to be more comparable to those of groundwater storage/subsurface runoff (fewer and longer droughts, like in Ch. 5). In the Upper-Sázava and Upper-Guadiana case study areas, mean duration of droughts in groundwater storage and subsurface runoff was relatively long, as was expected (106 and 117 days and 159 and 107 days, for Upper-Sázava and Upper-Guadiana, respectively), but total runoff did not reflect a substantial groundwater influence as the mean duration of droughts in total runoff was short (30 and 36 days, respectively). This is visualised in Figs. 6.2 and 6.3, in which there were more and shorter drought events in total runoff (lower row) than in groundwater (4th row).

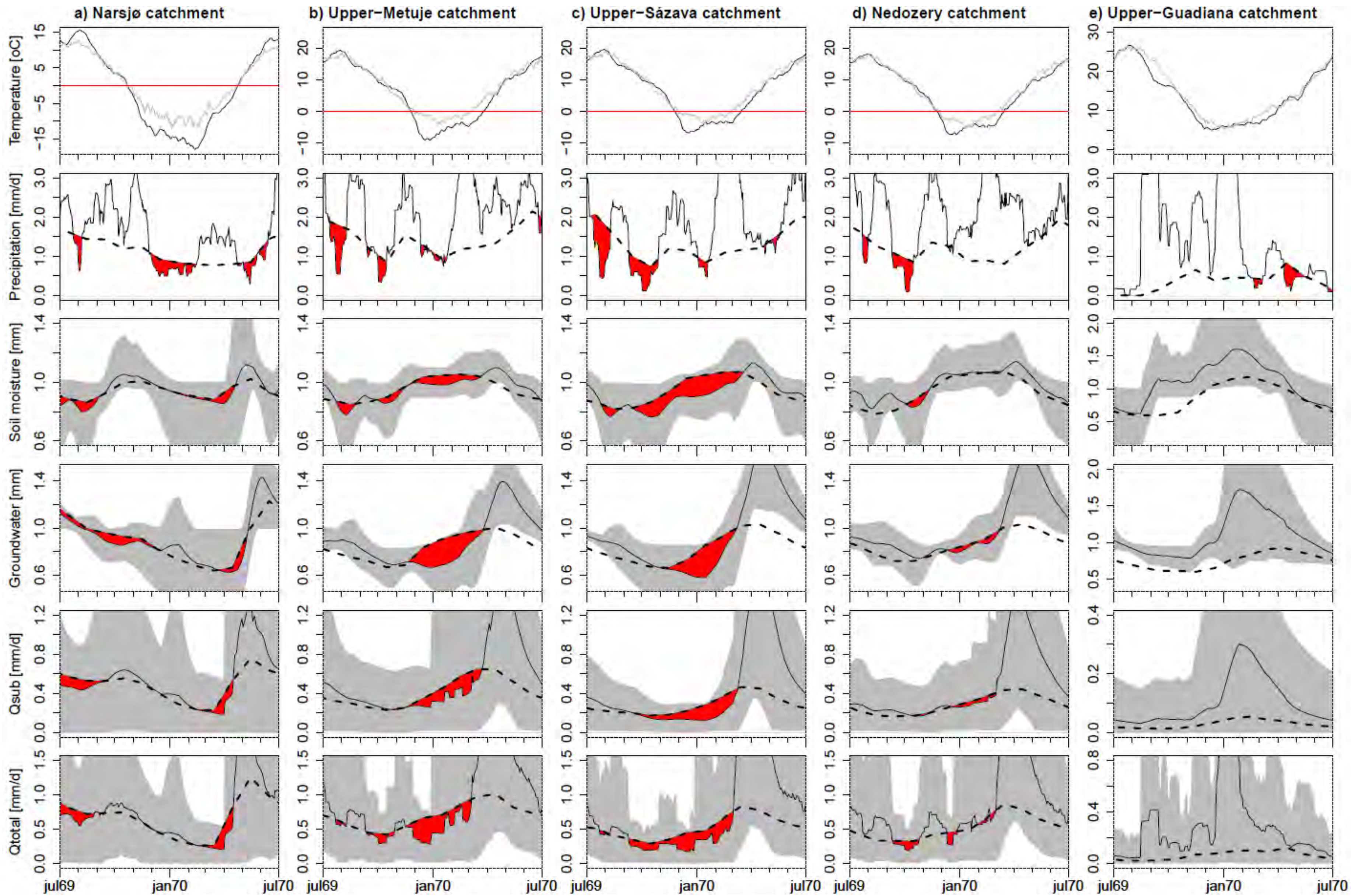


Figure 6.2: Example of drought events in all case study areas, Narsjø, Upper-Metuje, Upper-Sázava, Nedožery, and Upper-Guadiana in 1970 (all rows: black, solid line = time series of meteorological variable (30-day moving-average of WFD temperature and precipitation) or ensemble mean of hydrological variable (see y-axis), grey area = range of individual models, dashed line = smoothed monthly 80 %-threshold of displayed variable, red area = drought event; upper row: grey line = long-term average of WFD temperature, red line = 0°C.

- Average groundwater drought duration in Nedožery was longer (99 days) than groundwater drought duration in Upper-Metuje (70 days). Nedožery was anticipated to have shorter groundwater droughts, due to the lack of storage in the catchment and as a result a fast reaction to precipitation (Sect. 2.1.4 and Oosterwijk et al., 2009), and Upper-Metuje was anticipated to have longer groundwater droughts, due to storage in the extensive aquifer system and as a result a slow reaction to precipitation (Sect. 2.1.2 and Rakovec et al., 2009). Upper-Guadiana was expected to have even longer groundwater droughts than the average of around 160 days, because multi-year droughts are common in this catchment due to its semi-arid climate and large storage in extensive aquifer systems and wetlands (Sect. 2.1.5 and Peters and Van Lanen, 2003). In Ch. 5, average duration of groundwater droughts in Upper-Guadiana was more than 750 days.

In conclusion, the ensemble mean of the large-scale models showed a reasonable reproduction of general drought characteristics in the case study areas. Propagation processes were clearly reflected. In general, the ensemble mean of the large-scale models reproduces droughts in fast responding systems better than those in slow responding systems. In slow responding systems too many short hydrological droughts were simulated.

6.3.2 Drought propagation features

For a more thorough insight into drought generating mechanisms we also investigated time series of meteorological data of the WFD and hydrological data of the large-scale models for the propagation features mentioned in Sect. 1.2.2. From a visual inspection of the total time series of precipitation (examples in 2nd row in Figs. 6.2 and 6.3) and total runoff (examples in lower row in Figs. 6.2 and 6.3) we learned that the shape of the signal of the ensemble mean total runoff was quite similar to the precipitation signal. Recessions, which are an indication of catchment processes, were not visible in the time series of total runoff and only slightly in groundwater storage. With regard to the drought propagation features, the ensemble mean of the large-scale models showed:

- Very little lag: the start of a hydrological drought almost coincided with the start of the associated meteorological drought. The lag between a drought in precipitation and total runoff was estimated to be on average between 4 and 15 days (dependent on catchment), while using a catchment-scale model it was estimated to be between 24 and 51 days for the same catchments (Ch. 5). A study of the hydrological drought response time to weather-type occurrence in northwestern Europe showed even larger values, varying between 45 and 210 days, dependent on basin storage properties [Fleig et al., 2010]. The absence of a lag in the ensemble mean of large-scale models can be partly explained by the fact that we studied single grid cell runoff, for which no routing was applied. If we would have studied the routed discharge of a large number of grid cells (i.e. a larger catchment), a larger lag would be expected. We checked this hypothesis by studying the routed discharge of the Upper-Guadiana case study area, because it is the largest catchment with highest routing effects expected there. When switching from single grid cell runoff to routed discharge, the lag between precipitation and discharge increased from 4 days to 11 days, which is still considerably less than the lag of 24 days produced by a catchment-scale model (Ch. 5).

- Very little lengthening: also the end of a hydrological drought almost coincided with the end of the associated meteorological drought, because a precipitation peak immediately caused a higher runoff in the large-scale model simulations. Exceptions are some cases in winter with temperatures below zero in which snow accumulation took place (e.g. in Upper-Metuje and Upper-Sázava, Fig. 6.2b, c). Furthermore, sometimes during a dry series of years recovery from drought was slightly slower than during a wet series of years.

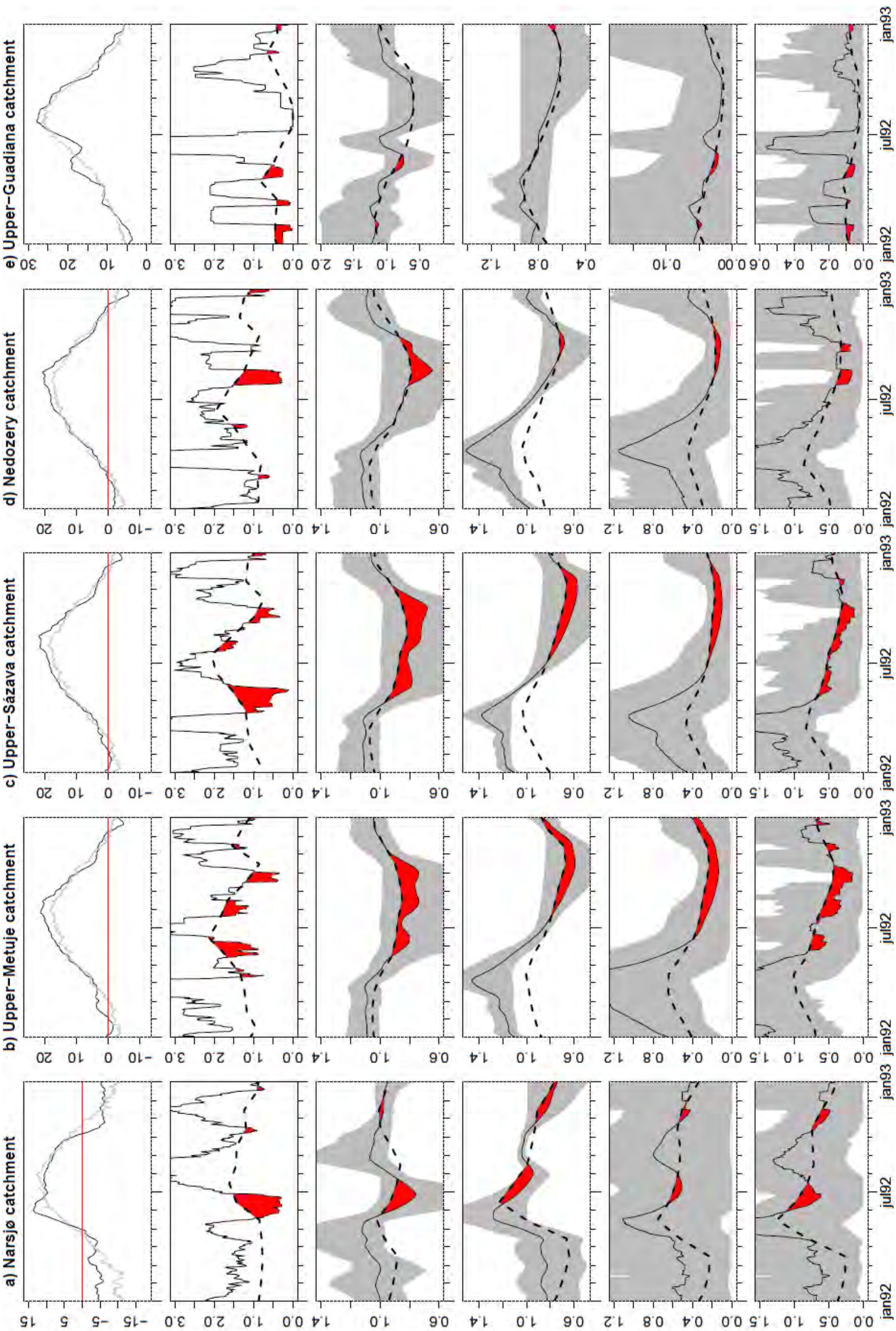


Figure 6.3: Example of drought events in all case study areas, Narsjø, Upper-Metuje, Upper-Sázava, Nedožery, and Upper-Guadiana in 1992 (legend: see Fig. 6.2).

- Almost no pooling: most meteorological droughts resulted in a separate hydrological drought (compare precipitation, 2nd row, and total runoff, lower row, in Figs. 6.2 and 6.3). Only in some cases meteorological droughts merged into one long hydrological drought (e.g. the drought events in Upper-Sázava; see Figs. 6.2c and 6.3c, lower row).

- Some attenuation: during a multi-year period of on average high precipitation, short meteorological drought events were filtered out (e.g. in Upper-Guadiana in 1970; see Fig. 6.2e, lower row). Prudhomme et al. [2011] also found that the non-occurrence of extremes is generally simulated in the correct period by a number of large-scale models.

In conclusion, the ensemble mean of the large-scale models showed a poor reproduction of drought propagation features in the case study areas. Total runoff reacted immediately to precipitation. Meteorological droughts directly led to hydrological droughts (little lag and only some attenuation) and a precipitation peak immediately ended a hydrological drought (little lengthening or pooling).

6.3.3 Typology

Additionally, we applied the drought typology of Ch. 5 to the large-scale model results. Many hydrological drought events were unidentifiable (5 % of all events for Upper-Metuje up to 28 % for Narsjø, Table 6.4, last column), meaning that no anomaly in precipitation or temperature could be found that caused the hydrological drought event. Many of these unidentifiable drought events occurred in the snow season. The snow-related drought types (i.e. *rain-to-snow-season drought*, *cold snow season drought* and *warm snow season drought*, Sect. 6.2.2.4) were clearly more difficult to distinguish using the ensemble mean of the large-scale models than using catchment-scale models (that were used to develop the typology). In Narsjø, for example, a precipitation deficit during winter (with temperatures well below zero and precipitation falling as snow, Table 6.2) occasionally initiated a hydrological drought during that same winter. This should not have occurred, because if temperatures are below zero, a lack of snowfall should not influence winter runoff, but only snow accumulation.

6.3.3.1 Classification of all hydrological drought events in the case study areas

Table 6.4 gives the percentages of all drought events in total runoff and subsurface runoff (proxy for groundwater storage; Sects. 6.2.2.4 and 6.3.1) in all five case study areas that were attributed to a certain hydrological drought type. The following can be noted:

- Drought events in subsurface runoff and total runoff had very similar hydrological drought types. The exception is *composite drought*, which did not occur in total runoff in some case study areas (e.g. Upper-Sázava).

- Many drought events were classified as *classical rainfall deficit drought* (in total for all case study areas together, 48 % in subsurface runoff and 62 % in total runoff). Especially Upper-Sázava and Upper-Guadiana had many *classical rainfall deficit droughts*.

- As expected, *wet-to-dry-season droughts* only occurred in the case study area with a semi-arid climate (Upper-Guadiana) and snow-related droughts (*rain-to-snow-season drought*, *cold snow season drought*, and *warm snow season drought*) only in regions with a continuous snow cover in winter (all except Upper-Guadiana).

- *Composite droughts* were found in all case study areas, but with low percentages. They did not only occur in regions with a slow response to precipitation (Upper-Metuje, Upper-Sázava, and Upper-Guadiana), but also in Narsjø and Nedožery (regions which typically

Table 6.4: Hydrological drought types of all hydrological drought events per catchment (subsurface runoff and total runoff)

		classical rainfall deficit drought	rain-to-snow-season drought	wet-to-dry-season drought	cold snow season drought	warm snow season drought	composite drought	un-identifiable
Narsjø	Q_{sub}	25 %	13 %	–	15 %	19 %	2 %	27 %
	Q_{total}	31 %	9 %	–	12 %	15 %	4 %	28 %
Upper-	Q_{sub}	53 %	3 %	–	13 %	23 %	5 %	5 %
Metuje	Q_{total}	63 %	–	–	14 %	17 %	1 %	6 %
Upper-	Q_{sub}	63 %	4 %	–	4 %	8 %	17 %	4 %
Sázava	Q_{total}	71 %	2 %	–	7 %	9 %	–	11 %
Nedožery	Q_{sub}	50 %	10 %	–	20 %	5 %	5 %	10 %
	Q_{total}	62 %	2 %	–	14 %	7 %	–	15 %
Upper-	Q_{sub}	65 %	–	19 %	–	–	4 %	12 %
Guadiana	Q_{total}	75 %	–	8 %	–	–	–	17 %

Table 6.5: Hydrological drought types of the five most severe hydrological drought events per catchment (subsurface runoff and total runoff), selection based on deficit volume

		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	Q_{sub}	20 %	20 %	–	–	40 %	20 %
	Q_{total}	–	40 %	–	–	20 %	40 %
Upper-	Q_{sub}	–	20 %	–	20 %	20 %	40 %
Metuje	Q_{total}	20 %	–	–	60 %	20 %	–
Upper-	Q_{sub}	–	20 %	–	–	–	80 %
Sázava	Q_{total}	20 %	40 %	–	–	40 %	–
Nedožery	Q_{sub}	20 %	20 %	–	–	20 %	40 %
	Q_{total}	80 %	–	–	–	20 %	–
Upper-	Q_{sub}	40 %	–	20 %	–	–	20 %
Guadiana	Q_{total}	60 %	–	40 %	–	–	–

have only limited storage and show a quick response to precipitation). In Nedožery these *composite droughts* were two events in subsurface runoff for which different hydrological drought types in different seasons were not interrupted by a recharge peak. One example in Nedožery, in which *warm snow season droughts* and *classical rainfall deficit droughts* were combined, is shown in Fig. 6.4a. This is a phenomenon that can occur in reality, but that was not expected in this specific case study area because of its quick response to precipitation. In Narsjø, *composite drought* events were related to a missing snow melt peak due to a severe meteorological drought in winter (e.g. the winter of 1996; see Fig. 6.4b, 2nd row). This phenomenon has not been found earlier in observations or catchment-scale models for the respective catchment (Chs. 4 and 5 and Van Loon et al. [2011b]), nor in other European catchments [Hannaford et al., 2011; Prudhomme et al., 2011]. In these studies, winter drought events in cold climates always ended by snow melt, even after winters with limited snow cover. It is therefore unknown whether these simulations with the large-scale models reflect a phenomenon that occurs in reality.

- Only a small number of *composite droughts* occurred in Upper-Guadiana and Upper-Metuje, while those case study areas reflect catchments with extensive aquifer systems and were therefore expected to have more *composite droughts* (in Ch. 5 *composite droughts* were 17 % of all groundwater drought events in Upper-Metuje and 67 % in Upper-Guadiana).

In Narsjø and Upper-Guadiana, the interplay between precipitation and temperature was not always according to expectations, leading to an unforeseen distribution over the hydrological drought types in Table 6.4. In Narsjø, runoff peaks and hydrological droughts developed during winter, although temperatures were well below zero. This has two consequences. First, drought events starting in summer/autumn were ended by a runoff peak in winter and could therefore not develop into a *rain-to-snow-season drought*, but were classified as *classical rainfall deficit droughts* (see the drought in groundwater, 4th row, and the minor event in subsurface runoff and total runoff, 5th and lower row, in November 1974 in Fig. 6.4c). Second, *warm snow season droughts-subtype B*, or *classical rainfall deficit droughts* developed in Narsjø during winter (see the drought in subsurface runoff and total runoff, 5th and lower row, in March 1975 in Fig. 6.4c), while these were expected to occur only in catchments with winter temperatures around or above zero (Sect. 6.2.2.4). This is because in winter, despite the temperatures well below zero, runoff still reacted immediately to precipitation, and therefore a lack of precipitation in winter could start a hydrological drought.

A similar process was observed in Upper-Guadiana. In summer, when potential evapotranspiration is much higher than precipitation, recharge and runoff should be zero because all precipitation normally is lost to evapotranspiration. In the ensemble mean of the large-scale models, however, runoff peaks still occur in Upper-Guadiana in summer. Consequently, drought events did not extend into the dry season and were classified as *classical rainfall deficit droughts* instead of *wet-to-dry-season droughts* (see the runoff peak in July 1987 in Fig. 6.4d, lower row).

6.3.3.2 Classification of the five most severe hydrological drought events in selected case study areas

For each case study area the five most severe drought events were selected on the basis of deficit volume (as in Ch. 5). This changed the distribution over the hydrological drought types (compare Tables 6.4 and 6.5).

The *classical rainfall deficit drought* occurred less often in most case study areas (in total, for all case study areas together, occurrence decreased from 48 % to 16 % in subsurface runoff, and from 62 % to 36 % in runoff). The exception is total runoff in Nedožery, where four of the five most severe drought events were of the *classical rainfall deficit* type. The *cold snow season drought* disappeared almost completely from the list, because this hydrological drought type usually has low deficit volumes. These shifts are in line with Ch. 5.

If we compare Table 6.5 with Table 5.4 in Ch. 5, we note some differences between the typology of severe drought events using catchment-scale and using the ensemble mean of large-scale models:

- In general, more of the most severe drought events were *classical rainfall deficit droughts* and *warm snow season droughts* (on average in total runoff, 36 % *classical rainfall deficit droughts* using large-scale models vs. 32 % using a catchment-scale model, and 20 % *warm snow season droughts* using large-scale models vs. 16 % using a catchment-scale model). Differences between catchments were large. For example, Upper-Metuje had fewer *classical rainfall deficit droughts* using the large-scale models instead of a catchment-scale model (20 % instead of 60 % in total runoff), whereas Nedožery had more (80 % instead of 40 % in total runoff).

- Fewer of the most severe drought events were *rain-to-snow-season droughts* (for example, in Narsjø 20 % and 40 %, instead of 80 % using a catchment-scale model).

- The distribution of *composite droughts* was different. Severe drought events of this type did not only occur in slow responding catchments, but in all catchments (in subsurface runoff).

If drought events had have been classified according to their duration (instead of deficit volume) and the five longest drought events selected, the distribution over the hydrological drought types would have been only slightly different from the one in Table 6.5 (not shown). Intense but short-lived drought types such as *warm snow season droughts* would have occurred slightly less often, and long but non-intense drought types like *rain-to-snow-season droughts* and *wet-to-dry-season droughts* would have occurred slightly more often.

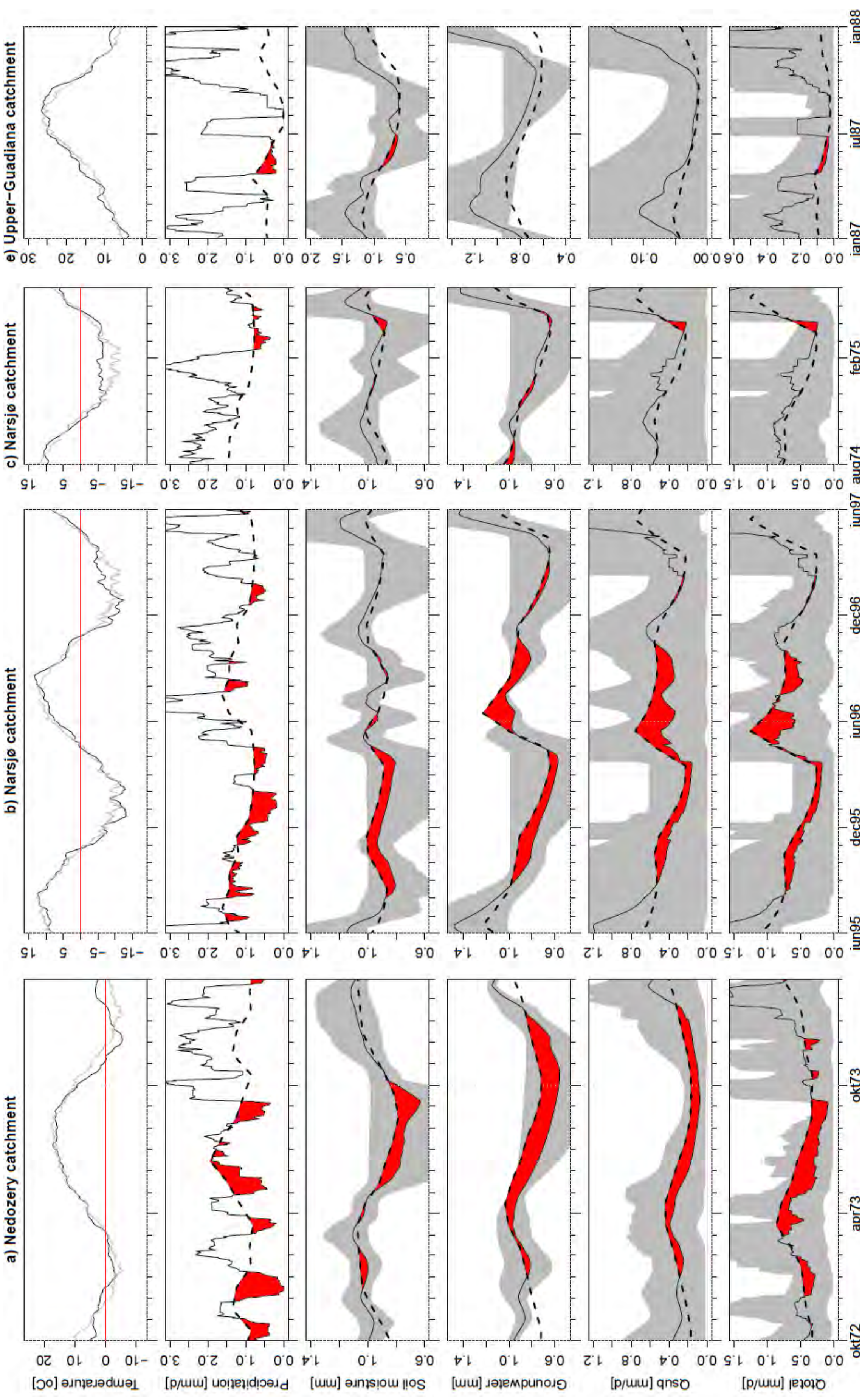


Figure 6.4: Examples of problems occurring in drought simulation; **(a)** *composite drought* in Nedožery (1973), **(b)** *composite drought* in Narsjø (1995–1997), **(c)** drought in winter in Narsjø (1974–1975), and **(d)** drought in summer in Upper-Guadiana (1987) (legend: see Fig. 6.2).

In conclusion, the ensemble mean of the large-scale models showed a reasonable reproduction of drought typology in the case study areas. All hydrological drought types of Ch. 5 were represented in the ensemble mean of the large-scale models, and in the climate type in which they were expected. The distribution of the hydrological drought types had some mismatches, e.g. a high percentage of *classical rainfall deficit droughts* in all case study areas, a low percentage of *composite droughts* in slow responding case study areas, unexpected occurrence of *composite droughts* in fast responding case study areas, a low percentage of *rain-to-snow-season droughts* in cold climates and *wet-to-dry-season droughts* in semi-arid climates.

6.4 Discussion and recommendations for the improvement of large-scale models

In this research the central question was how well large-scale models reproduce drought propagation. Before we answer that question (Sect. 6.4.2) and give some recommendations for the improvement of the models that are based on our analysis (Sect. 6.4.2.3), we first discuss the limitations of our methodology (Sect. 6.4.1).

6.4.1 Methodology

We used a specific set of large-scale models for our analysis, but we could have chosen other or more models (GHMs and LSMs). The time series of the individual models and therefore the ranges of the hydrological variables shown in Figs. 6.2, 6.3, and 6.4 would have been different. However, we expect that the ensemble mean of the models would not change significantly, because the models in our selection are representative of the range of existing large-scale models [e.g. Haddeland et al., 2011; Harding et al., 2011]. They have very different model structure and parametrisations, and therefore show very different responses. Unfortunately, no overall 'best' large-scale model exists. Some models perform, for example, very good in temperate regions, but bad in cold climates; others perform good in cold climates, but very bad in tropical regions. The same is true for fast and slow responding physio-geographic regions. For drought propagation studies in small uniform regions, i.e. with similar climate and catchment characteristics, it would be possible to select the large-scale model that performs best in that region. But for drought studies on continental or global scales, where conditions and therefore model results are extremely variable, such a choice cannot be made and the best solution is to use a multi-model ensemble (as was suggested earlier by various authors; see Sect. 6.1). As this study aims to test these large-scale applications, we follow this approach.

The model spread is an indication of model structure uncertainty in the multi-model ensemble. Parametric uncertainty in the individual models has not been investigated in this study. A single simulation was used for all models. We do, however, expect that parametric uncertainty is substantial. The large-scale models were not (or only minimally) calibrated (Sect. 6.2.1.2), because (i) observed and simulated variables and scales do not match (for example simulated grid cell runoff vs. observed catchment discharge, or scarce point-measurements of groundwater vs. simulated total subsurface storage); (ii) the models are assumed to include all important physical processes; and (iii) parameters of the models were derived from large-scale maps of e.g. vegetation and soil properties. As a result of both model structure and parametric uncertainty, the simulation of soil moisture and hydrological droughts is far more uncertain than simulation of meteorological droughts. The simulation of state variables in particular has a high uncertainty, as was recently reported by Samaniego et al. [2012]. In this study, however, the standardisation of the state variables SM and GW (Sect. 6.2.2.2) and the use of a relative threshold (percentile of flow duration curve; Sect. 2.4.2) account for biases in the absolute value of the states. Further issues regarding the effect of model structure and parametric uncertainty on drought propagation will be discussed in the next section (Sect. 6.4.2).

We tested the ensemble mean of the large-scale models in five case study areas. An extrapolation to more and other case study areas would be interesting, especially to regions outside Europe (e.g. tropical and arid regions in Africa and Asia). The analysis of drought characteristics can be done on a high number of grid cells with different climates using the method of Van Huijgevoort et al. [2012b]. The analysis of drought propagation features and the classification of hydrological droughts into types require visual inspection and expert knowledge. Therefore, it would be more difficult to study these drought-related aspects in a much larger sample of case study areas.

In classifying hydrological droughts into types we found a large number of unidentifiable droughts (Table 6.4). For the remaining events the meteorological anomaly/ anomalies causing the drought event was/were found by visual inspection of time series of all hydrometeorological variables. Quantification of this relationship between meteorological and hydrological drought has barely been investigated and has proved to be very difficult. To our knowledge the best effort is elaborated in the recent paper of Wong et al. [2013]. They found that copulas have more potential to link a hydrological drought to preceding meteorological drought(s) than classical linear correlation techniques.

Our aim was to include only natural headwater catchments in our study. The Upper-Guadiana, however, is far from natural, as groundwater extraction for irrigation has increased dramatically since the 1980s [e.g. Bromley et al., 2001]. The resulting hydrological situation is a combination of drought (natural causes) and water scarcity (anthropogenic causes). Therefore, the observed hydrological time series of this case study area were naturalised using the method described in Ch. 3. We compared drought propagation in the large-scale models (which did not simulate anthropogenic influences for this exercise; see Sect. 6.2.1.2) with drought propagation in these naturalised time series. An undisturbed catchment would have been better suited for our study, but finding an undisturbed groundwater-dominated catchment in a semi-arid climate with sufficient good quality data is not trivial.

In this study, we used the variable threshold to identify droughts. There are many other ways to calculate droughts using a kind of threshold approach (see Sect. 2.4), e.g. standardised precipitation index (SPI) and standardised runoff index [SRI; Lloyd-Hughes and Saunders, 2002; Shukla and Wood, 2008], regional deficiency index [RDI; Stahl, 2001; Hannaford et al., 2011], fixed threshold level method [Hisdal et al., 2004], cumulative precipitation anomaly (CPA), and soil moisture deficit index (SMDI) [e.g. Wanders et al., 2010]. These approaches give different numbers for the drought characteristics for a specific hydrometeorological variable (i.e. the numbers in Table 3.2), but the conclusions regarding propagation are not expected to change when one of these other methods is used. For example, Peters et al. [2006] and Tallaksen et al. [2009] use a fixed threshold in the Pang catchment (UK) instead of a variable threshold. They found drought propagation processes (e.g. lag, lengthening) that are comparable to the ones found in studies that used a variable threshold. An important reason to choose the variable threshold level method is that it enables comparison with the catchment model studies described in Ch. 5.

For our analyses, we used grid cell precipitation and runoff. The use of average catchment precipitation instead of grid cell precipitation would not have led to different results in the drought analysis. There are two reasons for this. First, the differences between observed catchment precipitation and grid cell precipitation for the studied case study areas were small, as was demonstrated by Van Huijgevoort et al. [2010, 2011]. Second, meteorological droughts have a large spatial extent and frequently cover a large region, as was demonstrated by Peters et al. [2006] and Tallaksen et al. [2009] (see Sect. 1.2.5), so there is little chance of missing a meteorological drought event by using a slightly different spatial coverage. As river routing has a considerable influence on discharge characteristics in large catchments, we tested the use of simulated streamflow at the outlet instead of grid cell runoff for the Upper-Guadiana case study area. Upper-Guadiana is the only studied area that is large enough to encompass more than one grid cell. We found that the lag between meteorological drought and hydrological drought increased slightly, but that the shape of the time series did not change at all. Our conclusions regarding the lack of attenuation and multi-year droughts are also valid when using streamflow at the outlet. We expect this to be consistent in other regions as well.

6.4.2 Evaluation of simulation of drought propagation by large-scale models

We investigated three different aspects of drought propagation: drought characteristics, drought propagation features and drought typology. In general these drought propagation aspects indicated a reasonable simulation of hydrological drought development in contrasting catchments in Europe, but we also found important deficiencies. Some drought propagation processes were clearly not simulated well by the ensemble mean of the large-scale models. These difficulties are all related to a too strong coupling between precipitation and discharge, which results in an immediate reaction of runoff to precipitation. This should not occur in certain climates types, i.e. semi-arid climates in summer and cold climates during the frost season, and in catchments with considerable storage. Hence the difficulties arise from deficiencies in the simulation of processes related to temperature and storage.

6.4.2.1 Temperature

The drought events simulated by the ensemble mean of the large-scale models are mainly governed by P control, and less by T control (Table 5.5). This resulted in an overestimation of the occurrence of the hydrological drought type that is predominantly caused by P control, i.e. *classical rainfall deficit drought*, and an underestimation of the occurrence of hydrological drought types that are (partly) caused by T control, i.e. *rain-to-snow-season drought*, *wet-to-dry-season drought*, *cold snow season drought*, *warm snow season drought*, especially *subtype A* (see Table 5.5 and Sect. 6.2.2.4). This is mainly due to the fact that droughts and discharge peaks were simulated in periods in which no drought or peaks were expected. Discharge peaks in winter in cold climates and in summer in semi-arid climates end drought events prematurely and therefore largely influence drought characteristics (shorter than anticipated) and drought typology (fewer *rain-to-snow-season droughts* and *wet-to-dry-season droughts* than anticipated). Hence the deficiencies of large-scale models in the reproduction of drought propagation processes are related to simulation of snow (low temperature) and evapotranspiration (high temperature).

Large-scale models are known to have difficulties with the correct simulation of snow accumulation [Feyen and Dankers, 2009; Haddeland et al., 2011; Stahl et al., 2011b, 2012b]. Prudhomme et al. [2011] and Stahl et al. [2012b] experienced problems in drought simulation in regions with winter temperatures close to zero. Their conclusion is confirmed in this study. Additionally, we also encountered problems in regions with winter temperatures well below zero, which is inconsistent with Prudhomme et al. [2011], who concluded that droughts in Scandinavia were reproduced well. One reason for incorrect snow simulation is related to elevation. Prudhomme et al. [2011] and Stahl et al. [2012b] found a larger error of drought simulation in mountainous areas. In these areas the grid cell elevation often deviates from the actual elevation of a catchment [Gudmundsson et al., 2012]. This difference influences both snowfall (simulated by WFD or by some of the large-scale models themselves; see input data in Table 6.1) and snow accumulation and melt (simulated by the large-scale models). According to Chs. 4 and 5, elevation plays an important role in drought propagation, because the development of snow-related hydrological drought types is very sensitive to a narrow temperature range around zero. This is comparable to floods, for which a critical zone for snow melt was found by Biggs and Whitaker [2012]. Subgrid variability, which is not captured by the large-scale models, results in a deviation in elevation between large-scale models and observations/catchment-scale models, and therefore in a deviation in drought typology. A higher resolution for the large-scale models might solve this issue, as was argued by Wood et al. [2011]. They explicitly mention snow (melt) simulation as one of the challenges that can be overcome using hyperresolution models. Besides elevation, this should also improve the simulation of the effect of the aspect of slopes and, therefore, the exposure to radiation and wind on snow melt in mountainous areas [e.g. Blöschl et al., 1990, 1991; Kustas et al., 1994; Grayson et al., 2002]. In climate modelling the benefits of higher resolution models are proved, e.g. by Hagemann et al. [2009].

Another temperature-related problem in large-scale models is the simulation of evapotranspiration (ET). The methodology used for the calculation of ET varies considerably between models [Haddeland et al., 2011] and can cause significant differences in model results [Gosling and Arnell, 2011; Stahl et al., 2012b]. The importance of ET for drought development has been demonstrated by Melsen et al. [2011] and Teuling et al. [2013]. One reason for deficiencies in the simulation of ET can be the lack of ET from wetlands and surface water [Gosling and Arnell, 2011]. Gosling and Arnell [2011] also mention that their model does not include transmission loss along the river network or evaporation of infiltrated surface runoff. This is a common issue in GHMs, which generally leads to an overestimation of runoff in dry catchments. Another reason may be related to groundwater storage. Van den Hurk et al. [2005] state that larger storage in model reservoirs results in sustained summertime evaporation. As many large-scale models have little storage, summertime evaporation is probably underestimated and discharge peaks can occur during summer in semi-arid climates. Also Bierkens and Van den Hurk [2007] and Lam et al. [2011] point towards the role of groundwater storage in the simulation of evaporation, especially related to the convergence of groundwater in wet discharge zones.

6.4.2.2 Storage

The effect of storage on hydrological drought development has been demonstrated by many authors [e.g. Peters et al., 2003; Van Lanen et al., 2004a; Tallaksen et al., 2009; Hannaford et al., 2011; Van Loon et al., 2011a; Van Lanen et al., 2012]. Therefore, the correct simulation of storage is important if large-scale models are to be used in hydrological drought analysis. Additionally, storage is important in climate change impact assessment. A more realistic storage capacity leads to smaller changes in both wintertime and summertime monthly mean runoff, and as a result to less extreme impacts of climate change [Van den Hurk et al., 2005]. Storage acts as a buffer to climate change.

Currently, storage is not simulated well in the ensemble mean of the large-scale models, resulting in insufficient variability between fast and slow responding areas. In slow responding areas, the reaction of runoff to precipitation is too fast, resulting in deficiencies in the reproduction of drought characteristics (shorter than anticipated), drought propagation features (little lag, lengthening, pooling, and attenuation), and drought typology (few *composite droughts*). The fast reaction of runoff to precipitation corresponds to the findings of, for example, Gosling and Arnell [2011]; Haddeland et al. [2011]; Stahl et al. [2012b]; Gudmundsson et al. [2012]. Based on their analysis of spatial cross-correlation patterns and runoff percentiles, Gudmundsson et al. [2011, 2012] conclude that discharge during dry conditions is largely influenced by terrestrial hydrological processes (catchment storage and release), in contrast to floods, which are mostly determined by forcing data. Stahl et al. [2012b] and Gudmundsson et al. [2012] found that these terrestrial hydrological processes are poorly replicated in the simplified storage schemes of large-scale models. Most models release too much of the incoming precipitation too quickly [Gudmundsson et al., 2012], and simulated droughts are interrupted more frequently than in observations [Stahl et al., 2011a]. Therefore, models perform best in regions where the runoff response to rainfall is more direct [Stahl et al., 2011a] or in very wet climates, where storage does not play an important role.

Hence both climate control (temperature) and catchment control (storage) on drought propagation are not simulated correctly by the ensemble mean of the large-scale models. This indicates a limited suitability of large-scale models when extrapolating to the future [e.g. Gosling et al., 2011; Corzo Perez et al., 2011b], in which drought propagation is governed by climate control, and to data-scarce regions [e.g. Stahl et al., 2012b], in which drought propagation is governed by climate control and catchment control.

6.4.2.3 Recommendations

Although representation of hydrological processes is better in large-scale hydrological models than in global climate models [GCMs; Hagemann and Dümenil, 1998; Van den Hurk et al., 2005; Sperna Weiland et al., 2010], there is still space for improvement of large-scale hydrological models for a correct reproduction and prediction of drought propagation across the globe. Simulation of evapotranspiration, snow accumulation, and storage in large-scale models should be improved to decrease the uncertainty in hydrological drought simulation.

For the improvement of the simulation of evapotranspiration a better understanding and a better representation of local-scale hydrological processes in dry regions of the world are essential [Gosling and Arnell, 2011; Lam et al., 2011]. Furthermore, re-infiltration and evaporation of surface runoff should be implemented in large-scale models.

First steps on the improvement of snow simulation have been set by Cherkauer et al. [2003], who improved the VIC model for cold areas, and Dutra et al. [2010] and Balsamo et al. [2011], who improved snow simulation in TESSEL. However, despite major advances Lettenmaier and Su [2012] note that 'there remain important problems in parameterisation of cold land hydrological processes within climate and hydrology models'.

First steps on the improvement of storage simulation have been set by Sutanudjaja et al. [2011] and Tian et al. [2012], who coupled a groundwater model (MODFLOW and AquiferFlow) to a land surface model (PCR-GLOBWB and SiB2). An important limitation is that these couplings are still offline, not allowing for dynamic feedbacks between groundwater storage, soil moisture, and evapotranspiration [Sutanudjaja et al., 2011]. Another difficulty is that in large-scale models parameters are representative of typical rather than locally realistic hydrogeological conditions [Gosling and Arnell, 2011; Gudmundsson et al., 2012]. For more locally (or at least, regionally) realistic subsurface runoff simulation using large-scale models, two steps must be taken. First, storage should be represented in the models in a better way, e.g. by including more groundwater reservoirs into the models or by online coupling with a groundwater model; second, higher-resolution large-scale datasets on storage properties should be derived in order to achieve more realistic model parameters for this groundwater part of large-scale models. This is necessary even in hyperresolution models, because there will always be sub-grid variability that needs parametrisation of processes [Beven and Cloke, 2012]. It is important to evaluate model results not only against observed discharge, but also against observations of state variables like snow accumulation, soil moisture and groundwater storage.

An encouraging note is that not all models have the same difficulties in simulating temperature and storage effects on drought propagation (see the model range in Figs. 6.2 and 6.3). For example, at least one model in the suite of large-scale models used in this study had extremely slow recessions, in other words a very slow reaction to precipitation [as was also demonstrated previously by Gudmundsson et al., 2012]. The drawback of this lies in the fact that a single large-scale hydrological model is often used globally, independent of the representativeness of the model for that specific region. Models with a fast reaction to precipitation are also used in slow responding systems and vice versa [e.g. Prudhomme et al., 2011]. In a comparable way models that have difficulties simulating snow accumulation processes are applied in cold regions and models that have difficulties simulating evapotranspiration processes are applied in semi-arid regions [e.g. Feyen and Dankers, 2009]. Therefore, in agreement with Stahl et al. [2012b] and Gudmundsson et al. [2012], we still advise the use of a multi-model ensemble of a number of large-scale model for drought studies, because in that way flashy and smooth hydrographs of very different large-scale models are averaged out. According to Beven and Cloke [2012], ensemble simulation is one methodology for taking into account the lack of knowledge on parametrisation of sub-grid processes.

Large-scale modellers can learn from each other, as has been shown by WaterMIP of the WATCH-project. More Model Inter-comparison Projects (MIPs) are needed that focus on hydrology instead of climate [e.g. Gates et al., 1999; Meehl et al., 2000, 2007; Covey et al., 2003; Friedlingstein et al., 2006]. Therefore, expectations for the recently started Inter-Sectoral Impact Model Intercomparison Project, ISI-MIP[1], are high [Schiermeier, 2012].

6.5 Conclusions

This study showed that drought propagation processes in contrasting catchments in Europe are reasonably well reproduced by an ensemble mean of ten large-scale models. However, results also indicated a limited suitability of large-scale models when extrapolating to the future and to data-scarce regions, because both climate control (temperature) and catchment control (storage) on drought propagation are not simulated correctly by the ensemble mean of the large-scale models.

The ensemble mean of the large-scale models was well able to simulate general drought propagation processes in drought characteristics; i.e. drought events became fewer and longer when moving from precipitation via soil moisture to groundwater storage, and drought characteristics of discharge were in between. Furthermore, the correct hydrological drought types were generally simulated in the correct climate type, i.e. *classical rainfall deficit droughts* in all climates, *wet-to-dry-season droughts* only in semi-arid climate, and snow-related droughts in areas with a continuous snow cover in winter.

However, challenges still exist in catchments with cold or semi-arid climates and catchments with large storage in aquifers or lakes. The immediate reaction of runoff to precipitation in the large-scale models, even in winters with below-zero temperatures and summers with high evapotranspiration, resulted in many short droughts in total runoff, and consequently in an overestimation of *classical rainfall deficit droughts* and an underestimation of *wet-to-dry-season droughts* and snow-related droughts. The still limited representation of storage in the large-scale models is reflected in the absence of a differentiation in drought characteristics of total runoff between fast responding and slow responding systems. Furthermore, almost no *composite droughts* were simulated for the slow responding case study areas, while many multi-year drought events were expected in these systems. The flashiness of the hydrograph of the ensemble mean of the large-scale models also showed up clearly in the drought propagation features. Drought events in the ensemble mean had very little lag and lengthening, almost no pooling and only some attenuation.

In general we anticipate that the simulation of hydrological drought has a significantly higher uncertainty than the simulation of meteorological drought. Potential improvement of hydrological drought simulation in large-scale models lies in the better representation of hydrological processes that are important for drought development. These processes are evapotranspiration, snow accumulation and especially storage. Besides the more explicit inclusion of storage in large-scale models, parametrisation of storage processes also requires attention, for example through a global-scale dataset on aquifer characteristics, improved large-scale datasets on other land characteristics (e.g. soils, land cover), and calibration/evaluation of the models against observations of storage (e.g. in snow, groundwater).

[1] www.pik-potsdam.de/research/climate-impacts-and-vulnerabilities/research/rd2-cross-cutting-activities/isi-mip.

Chapter 7

Drought propagation on the global scale

Under review at Geophysical Research Letters as: Van Loon, A.F., Tijdeman, E., Wanders, N., Van Lanen, H.A.J., Teuling, A.J., and Uijlenhoet, R.: *Seasonality modifies the propagation of meteorological drought into hydrological drought*.

7.1 Introduction

Drought is a natural disaster, resulting in severe economic and societal problems around the world [Seneviratne et al., 2012]. Although meteorological drought (i.e. a precipitation deficit) is most studied, agricultural or soil moisture drought and hydrological drought are primarily causing impacts that affect society [Dai, 2011]. These types of drought are therefore of paramount importance for water resources management.

Hydrological drought (i.e. a below-normal water availability in aquifers, lakes, reservoirs, and/or streams) has a variety of causes ranging from precipitation deficiency to prolonged frost conditions [Sheffield and Wood, 2011]. The translation of a drought signal from deviating meteorological conditions into soil moisture and/or hydrological drought is called drought propagation. Drought propagation strongly depends on climate and catchment characteristics (Ch. 5) and, consequently, hydrological drought characteristics show a pronounced variation around the globe.

Many studies investigated drought propagation on the catchment scale [Eltahir and Yeh, 1999; Peters et al., 2006; Tallaksen et al., 2009] and the regional scale [Vidal et al., 2010; Hannaford et al., 2011; Stewart et al., 2011]. These studies have resulted in a more thorough understanding of the processes underlying drought propagation. However, the geographical reference is limited to the catchment or region under study and a further generalisation and extension of process knowledge to the global scale is needed, as is advocated by Mishra and Singh [2010] and in the recent IPCC report on extremes by Seneviratne et al. [2012] (see Sect. 1.3.1). The approach introduced by Van Lanen et al. [2012] using a synthetic model is well suited for this purpose, because it allows for the isolation of effects and easy sensitivity analyses.

The main aim of this study is to investigate patterns in drought characteristics (of meteorological, soil moisture, and hydrological drought) associated with the effects of seasonality in climate on drought propagation.

7.2 Methodology

7.2.1 Data and model

We studied drought propagation across climate types through a controlled modelling experiment. Time series of climate data from the WATCH global re-analysis dataset (see Sect. 2.2.1) have been used as driving force for a conceptual hydrological model that combines a degree-day snow accumulation model, a soil water balance model, and a groundwater model based on linear reservoir theory. No channel routing was included; final model output was daily subsurface discharge.

The hydrological model was run for the period 1958–2001 for a large number of randomly selected grid cells of $0.5^\circ \times 0.5^\circ$ across the world. To ensure that different climates are well represented, we used the Köppen-Geiger climate classification [Geiger, 1961] and excluded extremely dry (desert, BW) and extremely cold (glacier, EH) climates. This resulted in a total number of 1271 selected cells. Van Lanen et al. [2012] demonstrated that this number is more than sufficient as a representative sample to study drought characteristics on a global scale.

In order to isolate climate effects, which are the focus of this study, from effects of catchment properties, all grid cells were assigned the same catchment characteristics based on representative conditions for land use, soil, and groundwater. Parameters of the model were chosen in accordance with recommendations for an average situation, e.g. a threshold temperature for snowfall of 0°C and a linear reservoir coefficient of 250 days [Seibert, 2000; Van Lanen et al., 2012]. The adopted approach generated possible time series for precipitation (P), soil moisture storage (SM), and subsurface discharge ($Qsub$) in every grid cell.

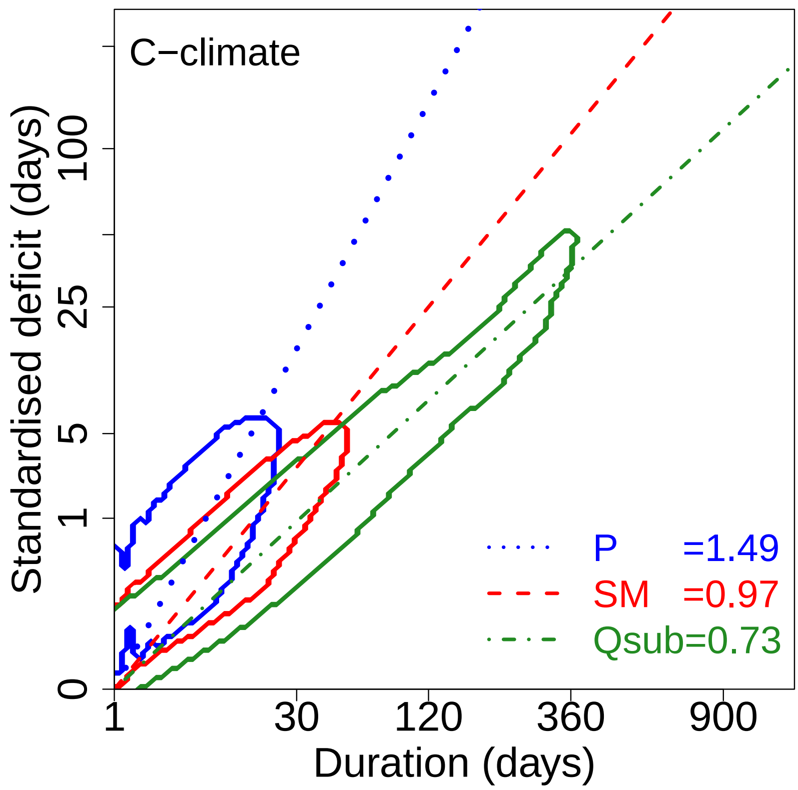


Figure 7.1: Contour plot of the 90 % probability density fields of transformed drought duration and standardised deficit in precipitation (blue), soil moisture (red) and subsurface discharge (green) for the temperate climate (C-climate), including linear regression lines and the slope of these lines (α).

7.2.2 Drought analysis and density fields

From these time series we identified droughts using the threshold level approach [Hisdal et al., 2004; Fleig et al., 2006]. To reflect seasonality we used a monthly varying threshold based on the 80th percentile of monthly duration curves of P (after applying 10-day moving average), SM, and $Qsub$. The discrete monthly threshold values were smoothed with a centred moving average of 30 days (Sect. 2.4.2). Such a time-varying procedure is similar to that applied in many other drought studies [e.g. Hisdal et al., 2004; Vidal et al., 2010] and is assumed to be robust for the analysis of drought propagation, especially when taking into account seasonality.

Drought characteristics that were analysed in this study are duration (Eq. 2.2) and standardised deficit (deficit divided by the long-term mean of the variable; Eq. 2.5). Although standardised deficit has no physical meaning for state variables (Sect. 2.4.2), we used it as a measure for the severity of SM droughts as well, as it enables relative comparison with droughts in P and $Qsub$. We transformed the data by taking the fourth root of both duration and standardised deficit to ensure a similar range for both variables. A logarithmic transformation would exclude zero deficit values. Similar to Kim et al. [2003] and Wójcik et al. [2006], non-parametric kernel density estimators [Wand and Jones, 1995] were adopted to determine smoothed bivariate probability fields of the drought characteristics, i.e. the joint probability density field of the transformed drought duration and standardised deficit. Based on this field, the area representative for 90 % of the drought events was selected.

As an example, the contour lines of the 90 % probability density fields of droughts in P, SM, and $Qsub$ of the temperate climate (C-climate) are presented in Fig. 7.1. The density fields of all variables have an elongated shape and a clear orientation to one direction, which we henceforth call a linear pattern. The variable $Qsub$ exhibits a longer density field than P and

SM, indicating that many droughts in $Qsub$ have a longer duration and a higher standardised deficit. As reported in the literature [Peters et al., 2006; Tallaksen et al., 2009], this is a result of propagation of drought through the hydrological cycle and is caused by pooling and lengthening. Pooling is the merging of meteorological droughts into a prolonged hydrological drought and lengthening refers to the situation that droughts become longer moving from meteorological via soil moisture to hydrological drought (Sect. 1.2.2).

To quantify the orientation of the density fields, we fitted a linear regression line through the drought events within the density field (visualised by the dashed lines in Fig. 7.1). We used a fixed starting point of (1,0), which is the theoretical minimum for a drought with the applied drought identification method. The slope of the regression line (α) was used for comparison between hydrometeorological variables and climate types. The smaller this α is, the smaller the increase of standardised deficit with duration. In the example in Fig. 7.1, α decreases from 1.49 for P to 0.97 for SM and 0.73 for $Qsub$. This means a reduced increase in standardised deficit with duration when moving through the hydrological cycle, which is also a result of drought propagation [Peters et al., 2006]. It is caused by attenuation, i.e. the damping effect of stores (soil moisture, groundwater) on the drought signal (Sect. 1.2.2).

More detailed information about the methodology can be found in Van Lanen et al. [2012] and Tijdeman et al. [2012].

7.3 Results

7.3.1 Seasonality effects on precipitation drought

Fig. 7.2 displays density fields of a selection of subclimate types. The orientation of the precipitation density fields in these subclimates is very similar, indicating no apparent climate effect on P drought characteristics. A slight difference in shape is visible in climate types with a strong seasonality in precipitation (e.g. BSh- and Cwa-climates). In these climates, high variability in precipitation between wet and dry seasons causes more variation in standardised deficit and, therefore, a slightly wider density field of P drought characteristics.

7.3.2 Seasonality effects on soil moisture drought

The soil moisture density fields of some subclimates (e.g. Af-, Cfb-, and Dfb-climates in Fig. 7.2) exhibit a linear pattern, comparable to that of the example in Fig. 7.1. Some other subclimates (e.g. Aw-climate) show a divergent pattern in the SM density field, i.e. at longer durations the density field becomes wider.

This difference in drought characteristics of soil moisture is related to seasonality in precipitation, as indicated by the second letter in the climate type name. The letter 'f' refers to significant precipitation in all seasons [Geiger, 1961]. In these climate types the soil moisture threshold is always high and, as a consequence, a meteorological drought results in a SM drought with a large increase of standardised deficit with duration, comparable to Fig. 7.1. The letters 's' and 'w' denote climates with a dry summer or winter, respectively [Geiger, 1961]. In these climate types, the threshold has a strong seasonality (Fig. 7.3a), which has implications for SM drought development. A meteorological drought in the wet season (Jun-Nov in the example in Fig. 7.3a) results in a SM drought with a large increase of standardised deficit with duration, because the threshold level is high enough to ensure that SM droughts are not limited by the wilting point, but can develop freely (SM1 in Fig. 7.3a). In the dry season (Nov-Jun in the example in Fig. 7.3a), however, the threshold level is low and SM droughts are bounded by the wilting point (SM2 in Fig. 7.3a). This causes limited deviation from the threshold and smaller standardised deficit values than expected based on drought duration. Occurrence of drought events in wet and dry seasons results in a divergent pattern of drought characteristics.

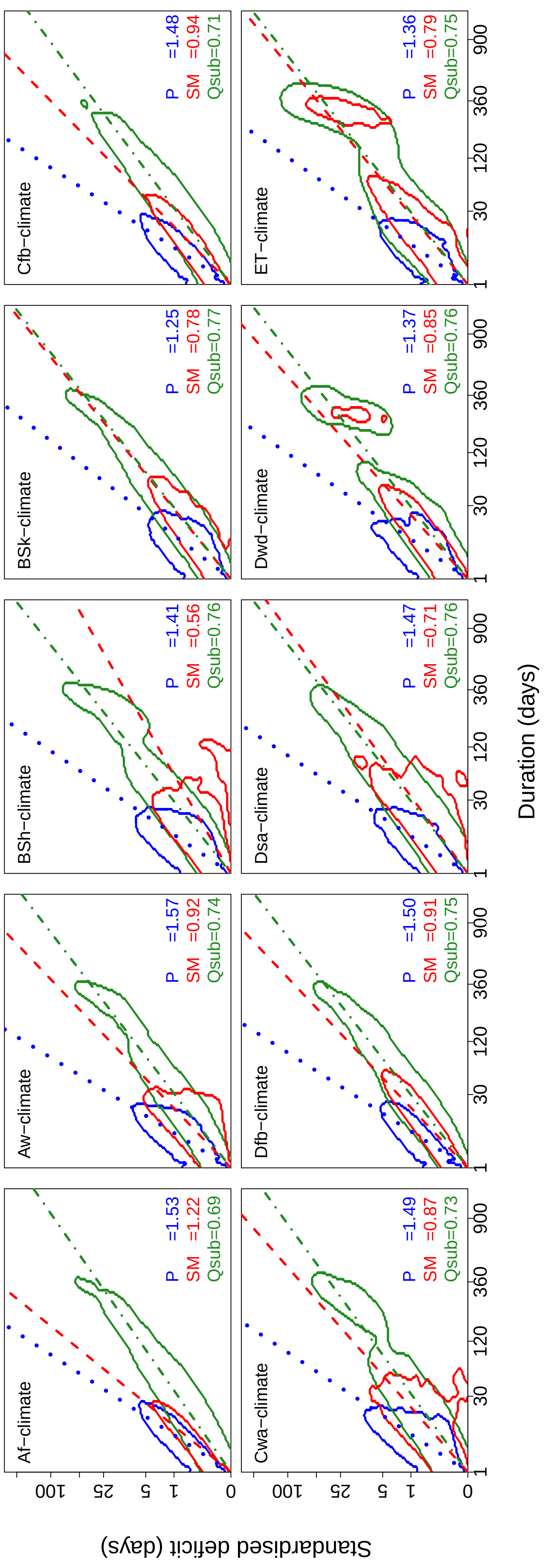


Figure 7.2: Contour plot of the 90 % probability density fields of transformed drought duration and standardised deficit in precipitation (blue), soil moisture (red) and subsurface discharge (green) for a selection of subclimates, including linear regression lines and the slope of these lines (α).

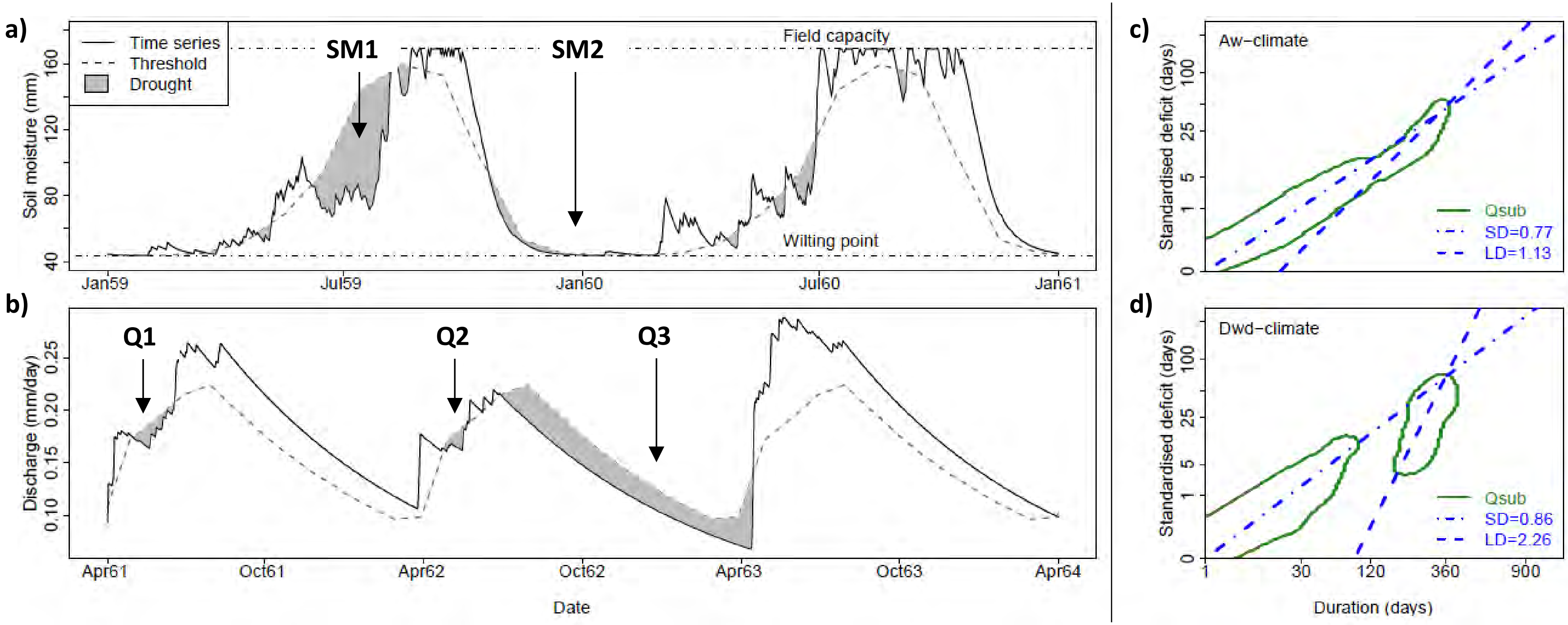


Figure 7.3: Time series of a) soil moisture for a typical savannah climate (Aw), and b) subsurface discharge for a typical continental climate (Dwd), and $Qsub$ density fields of the c) Aw-climate and d) Dwd-climate, including regression lines for short-duration (SD) and long-duration (LD) $Qsub$ droughts.

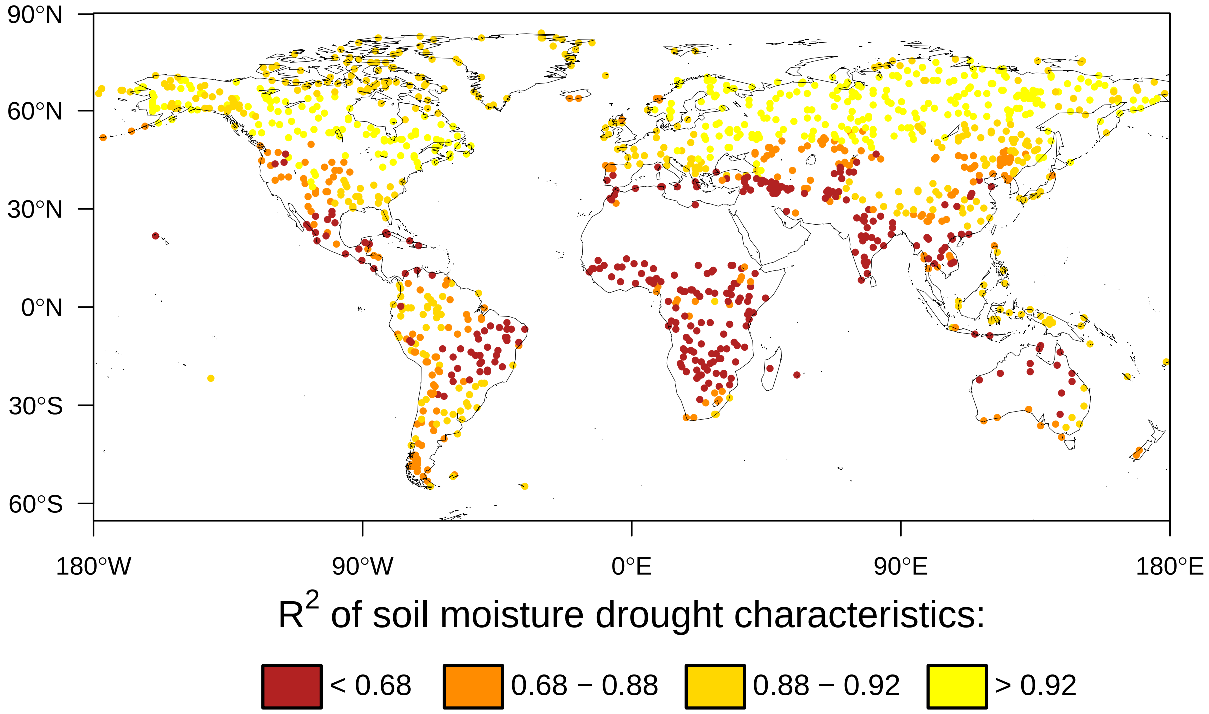


Figure 7.4: Spatial distribution of the correlation between transformed duration and standardised deficit of the soil moisture drought events within the density fields (quantified by the fraction of explained variance R^2); subclimate types are distributed evenly over the four classes.

In some climate types (e.g. the BSh-climate and climate types with the letter 'a' as third letter, denoting a hot summer [Geiger, 1961], such as the Cwa-, and Dsa-climates), a separate part of the SM density field near the x-axis is visible (Fig. 7.2). In these climates seasonality in precipitation is complemented with a strong seasonality in temperature, resulting in high evapotranspiration, lower threshold values and consequently, lower standardised deficit of SM droughts in the dry season.

The pattern of the SM density field is quantified by the correlation of the SM drought events, so that a high correlation reflects a more clustered SM density field and a low correlation reflects a more divergent density field. The spatial distribution of this correlation is plotted in Fig. 7.4. We clearly see highest correlations in climates with significant precipitation in all seasons, especially at higher latitudes on the northern hemisphere, and lowest correlations in climates with strong seasonality, e.g. the monsoonal, savannah and Mediterranean climates in most of Africa, southern Asia, Brazil, and Middle America. The warmer and more seasonal the subclimate, the more limited drought deficit development in SM in the dry season, and the more divergent the pattern of the SM density field.

7.3.3 Seasonality effects on hydrological drought

The density field of subsurface discharge of many subclimates exhibits a linear pattern (e.g. Af-, Cfb-climates in Fig. 7.2), indicating regular hydrological drought development. A striking feature in some other subclimates is a change in direction of the $Qsub$ density field around 120 days (Aw-, BSh-, Cwa-, Dwd-, ET-climates). Two distinct modes can be distinguished with different sensitivity of deficit with duration, i.e. longer $Qsub$ droughts had a larger increase in standardised deficit with duration than shorter $Qsub$ droughts.

This bimodal pattern in $Qsub$ drought characteristics is related to seasonality in both precipitation and temperature, influencing drought propagation. In subclimates with a distinct dry

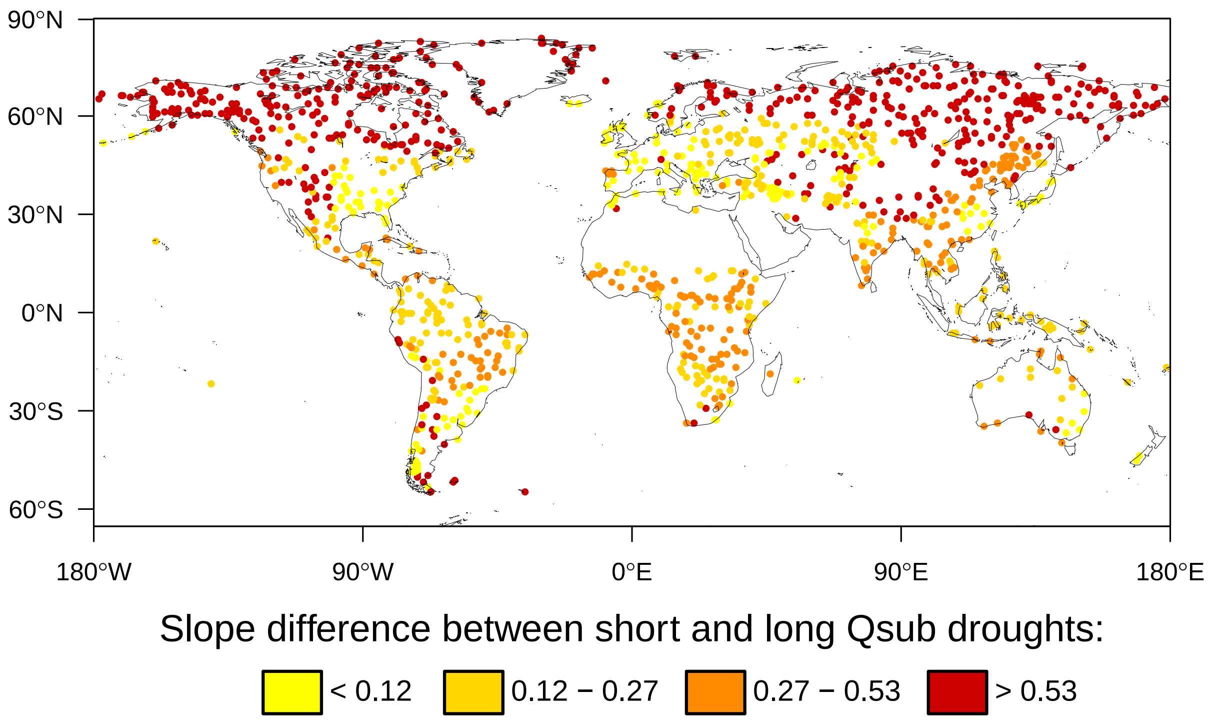


Figure 7.5: Spatial distribution of the differences between the slope of the weighted linear regression line (α) of long and short duration droughts; subclimate types are distributed evenly over the four classes.

season, a hydrological drought in the wet season that ends by a precipitation peak in that same season results in a short drought with drought characteristics exhibiting a linear pattern. A hydrological drought that does not end in the wet season continues throughout the dry season, because chances of recovery during the dry season are extremely low. Low precipitation and high evapotranspiration in the dry season rarely result in recharge to the groundwater and $Qsub$ stays below the threshold until a precipitation peak in the next wet season. The resulting long multi-season droughts (*wet-to-dry-season drought* in Ch. 5) have a larger increase of standardised deficit with duration.

A comparable process is observed in cold climates, where seasonality in temperature is dominant and recovery from hydrological drought is prevented by snow accumulation in winter. An example of drought events in a continental climate is given in Fig. 7.3b. Hydrological droughts Q1 and Q2 end in summer due to high rainfall. Hydrological drought Q3 does not end in summer, but continues throughout the winter. Chances of recovery of a hydrological drought during winter are extremely low, because all precipitation falls as snow and no recharge takes place. Therefore, the recession of subsurface discharge stays below the threshold until the snow melt peak in spring (May 1963 in Fig. 7.3b). These long multi-season droughts (*rain-to-snow-season drought* in Ch. 5) also have a larger increase of standardised deficit with duration, especially when snow melt is delayed.

The relation between deficit and duration within the different modes is quantified by calculating the slope of the linear regression line of short-duration droughts (<6 months; with fixed starting point (0,1)) and long-duration droughts (>6 months; with free starting point, see Figs. 7.3c and d). The difference between these slopes is a measure of the strength of the bimodal pattern, so that in subclimates with stronger bimodality this α-difference is larger (e.g. 1.4 in Fig. 7.3d) than in sublimates with a less pronounced bimodality (e.g. 0.36 in Fig. 7.3c). The spatial distribution of the α-difference is plotted in Fig. 7.5. Highest values are found in the cold regions in the high northern latitudes and in mountainous areas, like the Himalayas, the Andes, and the Rocky Mountains. The steppe, savannah, and Mediterranean subclimates (mainly on the southern hemisphere) show intermediate values, and the temperate subclimates of Europe and the eastern part of the United States have lowest values. In cold seasonal subclimates, the α-difference between short- and long-duration droughts is larger than in dry seasonal subclimates, indicating a more pronounced bimodal pattern in the $Qsub$ density field in cold seasonal subclimates than in dry seasonal subclimates. This leads to the conclusion that there is a lower chance of recharge in winters in cold climates than in dry seasons in warm climates.

7.4 Conclusion and discussion

This research has shown that in climates with strong seasonality in precipitation and/or temperature:

1. Precipitation drought characteristics are comparable to those in climates without seasonality.

2. Soil moisture droughts in the wet season can develop freely and have a large deficit, whereas soil moisture droughts in the dry season are limited by the wilting point and have a small deficit. This results in a divergent pattern in the density field of soil moisture drought characteristics.

3. Droughts in subsurface discharge are prolonged, in cold seasonal climates by snow accumulation in winter and in dry seasonal climates by low precipitation and high evapotranspiration in the dry season. This results in a larger increase of deficit with duration and, therefore, in a bimodal pattern in the density field of subsurface discharge drought characteristics.

Maps representing a quantification of patterns of droughts in soil moisture and subsurface discharge show that the effect is more pronounced when the climate is more seasonal. The effect of seasonality on soil moisture drought is mostly visible in the warm seasonal climates (like monsoonal, Mediterranean and semi-arid climates) and the effect of seasonality on subsurface discharge drought is most pronounced in the cold seasonal climates (such as Boreal climates). These findings are consistent with smaller-scale drought studies (Ch. 5 and Vidal et al. [2010]) and with observational evidence from recent severe drought events, like the 2011 drought in the Horn of Africa [Viste et al., 2012] and the 2009–2010 winter drought in Europe [Cattiaux et al., 2010] and Central Asia [Davi et al., 2010].

This increased process understanding of hydrological drought development on the global scale has some important implications for water management in seasonal climates. It is, for example, known that the drought characteristics duration and deficit are related (as reported by e.g. Dracup et al. [1980b]; Woo and Tariiule [1994]; Hisdal et al. [2004]), but this research has shown that in seasonal climates the relation is not uniform and sensitivity of drought deficit changes with duration. Additionally, we provide an argument for taking into account terrestrial processes on hydrological drought development in water management. We found that seasonality effects on soil moisture and subsurface discharge drought characteristics cannot be explained by meteorological processes alone. Frequently, however, meteorology-based drought indices (e.g. Standardised Precipitation Index, SPI, or Palmer Drought Severity Index, PDSI) are used as proxy for hydrological drought [Nalbantis and Tsakiris, 2009; Zhai et al., 2010; Dai, 2011]. The effects of high evapotranspiration and snow accumulation on hydrological drought development are not accounted for in these indices. Due to the non-linear response of groundwater and streamflow to the meteorological situation in climates with strong seasonality, hydrological drought characteristics cannot be derived straightforwardly from meteorological drought characteristics.

Chapter 8

Synthesis

The general objective of this PhD research project was to investigate drought propagation through the terrestrial part of the hydrological cycle, related to climate and catchment control (see Sect. 1.3). This aim was achieved by taking the following steps:

- distinguishing between drought and water scarcity to exclude anthropogenic control by means of an observation-modelling framework, as described in Ch. 3;
- gaining insight into the hydrological processes that underlie drought propagation on the catchment scale (Chs. 4 and 5) by studying the effects of climate type and catchment characteristics on drought propagation, and developing a hydrological drought typology;
- testing the performance of large-scale models on their ability to reproduce drought propagation processes (Ch. 6); and, having established that these large-scale models had important deficiencies,
- using a synthetic model to explore the influence of climate control on drought propagation on the global scale, as described in Ch. 7.

The overall conclusion of this thesis is that, although drought is a complex, non-linear phenomenon with drought characteristics varying as a function of climate type and catchment characteristics, generic patterns can be derived that reflect different hydrological processes underlying drought propagation. These processes result in different hydrological drought types that are shown to play a role both on the catchment scale and on the global scale.

In this last chapter I put the outcomes of this research into their scientific context, discuss the methods used, summarise the implications (for both science and management) and give recommendations for further research.

8.1 Scientific framework

Like any coupled sociohydrometeorological system, drought is so complex that it cannot be studied by investigating all parts separately. A holistic view is needed. Sivapalan et al. [2011], along the line of Harte [2002], advocated a synthesis of the Darwinian approach and the Newtonian approach in hydrology. The Darwinian approach is based on ecology and values holistic understanding of the behaviour of the landscape. In hydrology this would translate into: analysing individual catchments and trying to understand what is happening in terms of hydrological processes [Sivapalan et al., 2011]. In the Darwinian approach the system is regarded as complex and the science is mainly descriptive [Harte, 2002]. The Newtonian approach is based on physics and is exemplified in hydrology by the use of detailed process-based models that are run on large scales [Sivapalan et al., 2011]. In the Newtonian approach the system is regarded as simple and based on universal laws and the science is mainly predictive [Harte, 2002]. The synthesis of the Darwinian approach and the Newtonian approach aims to combine complex interdependencies and simplicity in universal laws. Elements of this synthesis in earth system science are [Harte, 2002]:

1. **Embracing the science of place.** Place-centred studies provide the best means of understanding processes, of identifying the actual mechanisms at work. Information obtained from studies on the catchment scale provides the basis for generalisation [Harte, 2002]. Studies like the ones described in Chs. 4 and 5 are examples of detailed small-scale studies that teach us about the mechanisms of drought propagation in different climates and different catchments.

2. **Searching for patterns.** Although the system is complex, general patterns can be found, which enable us to extend our insights from small scales to larger scales [Grayson et al., 2002]. These general patterns can improve our ability to predict [Harte, 2002]. An example is the hydrological drought typology developed in Chs. 4 and 5 and applied in Ch. 7. Such a typology is based on expert knowledge through inspection of hydrometeorological time series and, consequently, contains approximations and exceptions. Nevertheless it contributes significantly to our understanding of the complex system. A second search for patterns is presented in Ch. 3. In this chapter, a methodology is developed on the basis of a case study and generalised in order to obtain a wider applicability.

3. **Using simple models.** In earth system science, fully-coupled, all-inclusive, global models are increasingly being used [Van den Hurk et al., 2011; Wood et al., 2011] and a growing number of scientists have little doubt that these models are the future. However, the ever more complex models are generally unfalsifiable [Oreskes et al., 1994; Blöschl, 2001; Harte, 2002]. They are so complex that most hydrologists and drought researchers can only work with their outcomes and do not know the details of the models' structures and parametrisations [Van Vuuren et al., 2012]. Therefore, deviating results of models cannot be attributed to differences in structure or parametrisation anymore. Haddeland et al. [2011], Stahl et al. [2012b], Gudmundsson et al. [2012], Van Huijgevoort et al. [2012a] did not succeed in exactly pinpointing the causes of the differences and deficiencies in (drought) simulation by large-scale models; nor did I (in Ch. 6). In Sect. 6.4.2.3 I advocate the necessity of making improvements in large-scale models. However, the solution for hydrology might not lie in these complex models at all, especially not for process understanding and attribution studies, because no matter how detailed they are, there will always be something missing [Van Vuuren et al., 2012]. Furthermore, all the physical processes in these models need to be parametrised, and boundary conditions and initial conditions of the models need to be set, which requires an enormous amount of information that is generally not available [Beven and Cloke, 2012]. Another drawback is that it will still take quite some time before these kind of models are operational, because they have to overcome both conceptual and computational difficulties. Therefore, simple (parsimonious) mechanistic models, which capture the essence of the problem and the catchment's most dominant controls [Blöschl, 2001; Grayson and Blöschl, 2001] and can be used for hypothesis-testing and sensitivity analyses, might get us further [Harte, 2002]. An example is the synthetic hydrological model used in Ch. 7.

Harte [2002] suggested that particularity and contingency, which characterise the ecological sciences, and generality and simplicity, which characterise the physical sciences, are all necessary to understand the complex earth system. I argue that this also applies to drought.

Along the same line, Blöschl [2001] called for the use of the concept of 'similarity' in hydrology to reconcile statistical descriptions and observation-based process interpretations and, recently, Sivapalan et al. [2012] introduced the concept of 'socio-hydrology', a new science of people and water. The focus of socio-hydrology should be on 'observing, understanding and predicting future trajectories of co-evolution of coupled human-water systems' [Sivapalan et al., 2012]. Socio-hydrology can be studied using models, but not all model approaches are equally suitable. Van Vuuren et al. [2012] gave an overview of model approaches that couple the earth system and the human system, ranging from simple one-way exchange of information, via the representation of the human system within earth system models and vice versa, to fully-coupled online models. All approaches have both advantages and disadvantages, nicely summarised by Van Vuuren et al. [2012]. They state that 'the trend in model development seems to suggest that there is a "natural" evolution from the first, most simple, approach towards the last, most complex, approach', but that the strengths and weaknesses of the approaches indicate 'which type of integration is most suitable to address a given scientific question'. The conclusion is that the more complex approaches are only 'useful if potentially strong feedbacks are involved and the processes are rather well established', and that 'while full integration can potentially deal with feedbacks and consistency issues, it also leads to rather complex models and little flexibility in exploring uncertainty'. Van Vuuren et al. [2012] gave a number of examples of the optimal approach in different fields of study related to climate change. The work presented in this thesis can be regarded as an example related to drought propagation, in which the lack of process understanding and the need for determining uncertainties still argue for a simple one-way coupled approach. However, strong feedbacks exist, both within the natural system (Sect. 1.2.2) and within the human system (Ch. 3), which calls for a more sophisticated approach at a later stage.

8.2 Discussion

8.2.1 Definition

In Chs. 2 and 4, I argued that, since no universal definition of drought exists, any study on drought should start with providing a clear definition and should use that definition consistently throughout the study. Here, I used the definition that drought is a sustained and regionally-extensive period of below-normal water availability with natural causes only. This definition immediately raises the question of how to deal with the omnipresent human influence on the water system. For this part of my research, I did not follow the advise of Harte [2002], Sivapalan et al. [2011], and Sivapalan et al. [2012] to investigate the sociohydrometeorological system as a whole. As still too little is known about the hydrological (i.e. natural) processes underlying drought propagation (see Sect. 1.3), I decided to study these natural aspects in more depth. Hence, I have selected case study areas that are little affected by human influence (see Sect. 2.1). For one of the case study areas, significant human influence in the second half of the observation period was unavoidable. These anthropogenic impacts (water scarcity) first had to be separated from the natural effects (drought), before drought propagation could be studied in this case study area. In Ch. 3 I showed that, with regard to direct anthropogenic effects, this can be achieved by using an observation-modelling framework. Indirect human influence on the hydrological system (e.g. variations in feedback induced by land-cover change or climate change) also influences drought occurrence, but this can only be investigated with sophisticated models that are capable of performing scenario analyses including feedbacks [e.g. Van Lanen et al., 2004b; Hurkmans et al., 2009; Van Vuuren et al., 2012], which was not done in this thesis.

8.2.2 Indices

Various indices are being used in drought studies (Sect. 2.4.1). In this PhD research the focus was explicitly not on finding the 'best' drought index. Instead I applied a method that has been proved to be suitable for drought propagation research, taking into account seasonal variation and diversity in the underlying processes, namely the variable threshold level method (Sect. 2.4.2). Just like any other method, this method has advantages and disadvantages and, with the results presented in this PhD thesis, it can be evaluated and compared with other methods.

The main advantage of the variable threshold level method is that it stays as close as possible to the actual time series of hydrometeorological variables, so that the deficit volume of a drought event can be determined. The disadvantages are related to the choice of the threshold level, the method of calculation of the threshold level, and subjective choices in pooling and exclusion of minor droughts. In water management and applied research, the choice of the threshold level should be related to the sectors impacted by drought, e.g. critical point (when evapotranspiration becomes water-limited; Van Lanen et al. [2012]) for soil moisture drought, ecological minimum flow for hydrological drought, etc. [Hisdal et al., 2004; Laaha et al., 2013]. In most research, however, choices usually have to be made independent of a specific water management issue. One option is to study a range of threshold values, so that water managers can pick the one of their interest [e.g. Woo and Tariiule, 1994; Tate and Freeman, 2000]. Another option is to use only one threshold value while being aware of the fact that drought characteristics should be evaluated on the basis of the relative difference between variables or catchments, instead of on the basis of their absolute values (Sect. 6.4.1). According to Fleig et al. [2006], there is no single threshold level that is preferable to another and the selection of a specific threshold level remains a subjective decision. Although the choice of a pooling procedure and its parameters is also quite subjective, it is important to apply pooling. In Sect. 5.6 I found that, even in slow responding catchments, small discharge peaks occur on top of a smooth recession or low-flow signal. As these peaks do not alleviate the drought event, the drought identification method should not result in splitting up the drought event in several events that are mutually dependent.

In Sect. 3.5 the Water Scarcity Indicator (WEI) was discussed. Its value is recognised, but the WEI suffers from a lack of good-quality data on the appropriate temporal and spatial scales, just as many other water scarcity indicators [Savenije, 2000]. In Sect. 7.4 the use of meteorological drought indices like SPI and PDSI (see Sect. 2.4.1) as indicator for hydrological drought was criticised. Of course, general patterns of long-term (multi-year) meteorological indices (e.g. SPI12/SPI36) reflect patterns of long-term hydrological drought on a global scale, because in the end the long-term terrestrial hydrological cycle is driven by precipitation. This is a scale issue. In Sect. 1.2.5 and Fig. 1.4 I have shown that droughts typically occur on catchment to continental scales and with a duration of a few weeks to a year (a few years). On larger spatial scales and longer time scales only large-scale circulation patterns are relevant for hydrology (Fig. 1.4). All other processes are averaged out. On long time scales and large spatial scales, non-linearities due to temperature-related and storage-related processes can be neglected.

Within-year hydrological droughts are strongly dependent on these temperature-related and storage-related processes. Within-year variability is often more relevant for water resources management than year-to-year variability, both on the river basin scale and larger scales. Averaging the meteorological signal over long time scales can partly mimic these processes. Averaging leads to an attenuation of the signal and a pooling of short meteorological droughts, but does not produce the drought propagation features lag and lengthening (see Sect. 1.2.2 and Fig. 1.3). Especially lengthening is of paramount importance for seasonal drought types, i.e. *rain-to-snow-season drought* (Sect. 5.4.2) and *wet-to-dry-season drought* (Sect. 5.4.3).

Another important point is that these long-term meteorological indices cannot predict all the types of non-drought events that were described in Sect. 5.5.3. In that section some examples were given of situations in which a hydrological drought does not develop, e.g. i) a rainfall deficit in spring coinciding with snow melt peak, ii) a late start of the snow season, iii) high antecedent storage conditions, iv) a combination of temperature-related and storage-related processes. Although storage-related processes can sometimes be mimicked by averaging the precipitation signal over a long time period, temperature-related processes are not included. Strong arguments exist to include at least evapotranspiration in hydrological drought indices, which is expected to play a more prominent role in the future [Dai, 2011, 2013]. But also snow accumulation and snow melt and storage-related processes should be incorporated, because they cannot be captured by a simple statistical manipulation of precipitation time series. This PhD research has shown that non-linearities dominate the development of hydrological drought (see Ch. 7) and advocates the incorporation of these non-linearities in drought indices that are used for water resources management and planning.

In view of the above, hydrology is still relevant in drought research, as is confirmed by other authors. Bates et al. [2008] state that 'lakes respond in a very non-linear way to a linear change in climatic inputs'. Lavers et al. [2010] conclude that in the UK 'there are weaker atmospheric links with river flow compared to precipitation, reflecting the non-linearity of the rainfall-runoff transformation and the importance of basin properties as a modifier of climate inputs'. Teuling et al. [2013] also stress 'the need for a correct representation of evapotranspiration and runoff processes in drought indices'. Joetzjer et al. [2012] contradict this by showing that in the Amazon and Mississippi basins, 12-month meteorological drought indices (e.g. SPI, SPEI, and PDSI) have a high correlation with the 12-month hydrological drought index SRI. The reasons that their conclusion deviates from the previously cited references and my own results are that, i) the SPI and SRI have some disadvantages related to the fitting of a probability distribution function (see Sect. 2.4.1), ii) the SRI used by Joetzjer et al. [2012] is based on basin-averaged gridded runoff simulated by a large-scale model, which is shown in Ch. 6 to closely resemble the precipitation time series, iii) Joetzjer et al. [2012] did not study within-year droughts, so the effects of seasonality was completely disregarded, and iv) there is little snow accumulation in the Amazon and Mississippi basins, thus their results cannot be extrapolated to catchments in cold regions.

8.2.3 Data and models

In drought research the use of long time series of undisturbed good-quality data is essential (Sect. 2.2), either obtained from observations or from simulations. Both contain uncertainties. Observational uncertainty is hard to quantify (Sect. 2.2), but an estimate of the range of the uncertainty of hydrological observations is given in Ch. 3. The uncertainty of meteorological observations was not quantified in this thesis. All meteorological data in this study were carefully selected and quality-checked to reduce uncertainty, which does not guarantee perfect data, but at least it is the best available. Although the same was done for hydrological data, the discharge data of the selected catchments appeared to contain insufficient information to be suitable for data-based model improvement (Sect. 2.3.4 and Appendix A). This might, however, be due to the methodology that was applied rather than to the quality of the data.

Uncertainty in hydrological model results originates from input data uncertainty [McMillan et al., 2011], calibration data uncertainty [McMillan et al., 2010], and model uncertainty [Ajami et al., 2007]. In studies of input data uncertainty and calibration data uncertainty focus has been on average and high flows [e.g. McMillan et al., 2010, 2011]. There is little knowledge of the relative importance of these different sources of uncertainty during low flow and drought. In this thesis I also did not explicitly quantify input data and calibration data uncertainties, but I can state that more and better data lead to better results, for example, in making the distinction between drought and water scarcity (Ch. 3) and in calibrating and validating a hydrological model (e.g. Sects. 4.4 and 6.4).

Model uncertainty can be subdivided in structural uncertainty (i.e. related to model structure) and parametric uncertainty (i.e. related to model parameters and their identification). Model structural uncertainty was investigated in Sect. 2.3 and Ch. 6 and model parametric uncertainty was explored in Ch. 3.

In this thesis I have used three types of models (see Sect. 2.3 and Table 2.2):

- the conceptual, semi-distributed, catchment-scale rainfall-runoff model HBV in Chs. 3, 4, and 5;
- an ensemble of a number of physically-based, distributed, large-scale hydrological models and land surface models in Ch. 6; and
- a conceptual, distributed, synthetic hydrological model in Ch. 7.

The aim of this thesis was not to compare the results of all individual models with observations, but to use the model that is considered most suitable for drought analysis. As the dominant processes are scale-dependent (Sect. 1.2.5), different models were needed on the catchment scale and on the global scale. Only the catchment-scale model has been compared directly with observations (in Sects. 4.4, 5.3.1, 5.3.2, and Appendix B) and it proved to be quite successful in simulating low flows and drought under a variety of circumstances. The simulation of low flows by this model could not be improved on the basis of a systematic investigation of observed recessions (Sect. 2.3.4 and Appendix A).

Some critical remarks can be made on the structure and performance of the HBV model. Due to the focus on low flows during calibration (by using the Nash-Sutcliffe based on the logarithm of discharge as objective function), the model might have compromised the simulation of high flows. This is not the case in the catchments Upper-Metuje, Nedožery, and Upper-Guadiana, for which model biasses of 3–7 mm yr^{-1} were found. In the Narsjø catchment, however, the bias between simulated and observed discharge was as high as 137 mm yr^{-1}. The reason is that the high snow melt peak in this catchment was generally underestimated by the model. Several explanations for this difference are possible, either related to deficiencies in the model process representation or in the model parametrisation. Snow simulation in HBV is done with the degree-day method (Sect. 2.3.1), an empirical relationship between snow accumulation and melt and air temperature, and formulations for melt water storage and refreezing. Despite its simplicity, the degree-day method has been proved to be a powerful tool for melt modelling, often outperforming energy balance models on the catchment scale [Hock, 2003]. The two main shortcomings of the method (inaccuracy at high temporal resolution and high spatial variability) do not apply to this research as I did not consider sub-daily time scales and used a semi-distributed version of HBV, which allows for a spatial variation of melt rates with elevation and a number of slope/azimuth-factors to account for differences in the exposure of slopes to radiation [Blöschl et al., 1991; Kustas et al., 1994; Hock, 2003; Seibert, 2005].

HBV does not simulate soil frost and soil moisture depletion in winter with physically-based energy-balance equations like, for example, the COUP model does [Gustafsson et al., 2004; Colleuille et al., 2007]. From a comparison between the results of the HBV and COUP model in Norway, I estimated that both effects probably cancel each other out, since a correct simulation of soil frost would lead to a faster response of discharge to snow melt and a more realistic simulation of soil moisture depletion would lead to a slower response. In comparing the results with observations (Appendix B), I did not encounter problems in simulating the timing of the snow melt peak. The model parametrisation is probably of more importance in the underestimation of the snow melt peak. In the calibration, parameter values were chosen to optimise the simulation of low flows. Apparently, for the Narsjø catchment, this was at the expense of a correct simulation of high flows. As Hinzman and Kane [1991] and Seibert [2000] demonstrated, HBV can correctly simulate the snow melt peak, as long as the calibration is focused on average and high flows. Modelling high and low flows simultaneously in catchments with highly variable discharge and different governing processes in different seasons continues to be a challenge [Staudinger et al., 2011].

The ensemble of large-scale models was not compared with observations (Ch. 6), because I studied only five grid cells (out of more than 67,000) and, on that scale, validation against observations is not recommended (Sect. 6.2.2.1 and Blöschl [2001]). I rely on previous studies that concluded that the ensemble mean of a number of models is closer to the observations than any of the individual models (Sect. 6.1). Hurkmans et al. [2008] found that a large-scale model performed better than a water-balance model in simulating high flows for the Rhine basin. In Ch. 6 I concluded that this is not the case for low flows. As the catchment-scale model that was used in this thesis was calibrated to the specific catchments (Sect. 2.3.2) and the large-scale models were not (they are based on large-scale datasets to determine parameter values; Sect. 6.2.1.2), it is not surprising that my conclusion deviates from the one of Hurkmans

Table 8.1: Drought propagation processes (including development and recovery) per hydrological drought type and subtype (adapted from Table 5.5). P = precipitation and T = temperature

Hydrological drought type	Governing process(es)	Development	(lack of) Recovery
Classical rainfall deficit drought	Rainfall deficit (in any season)	P control	P control
Rain-to-snow-season drought	Rainfall deficit in rain season, drought continues into snow season	P control	T control
Wet-to-dry-season drought	Rainfall deficit in wet season, drought continues into dry season	P control	P and T control
Cold snow season drought	Low temperature in snow season, leading to:		
Subtype A	Early beginning of snow season	T control	T control
Subtype B	Delayed snow melt	T control	T control
Subtype C	No recharge	T control	T control
Warm snow season drought	High temperature in snow season, leading to:		
Subtype A	Early snow melt	T control	P control
Subtype B	In combination with rainfall deficit, no recharge	P and T control	P control
Composite drought	Combination of a number of drought events over various seasons	P and/or T control	P control
No drought	Combination of factors, incl. pre-event soil moisture and hydrological conditions	P and/or T control	-

et al. [2008]. Physically-based models could potentially lead to better simulation results in my catchments as well, because their representation of hydrological processes is more realistic. However, as various authors indicate, that is no guarantee for success as there is a difference in scale between the model and its physics, which makes spatial aggregation necessary [Kim, 1995; Blöschl, 2001].

The outcomes of the synthetic model (Ch. 7) were also not directly compared with observations of the case study areas, because the model was not meant to generate time series of hydrometeorological variables that are unique for a specific grid cell in the way large-scale models do [Van Lanen et al., 2012]. Furthermore, no routing was included in the model structure and no fast runoff component was simulated [Tijdeman et al., 2012]. The synthetic model is simple, but it includes the most important processes for hydrological drought development. Its structure and some of the processes included are based on the HBV model, which ensures consistency between the methodology used in Ch. 7 and Chs. 4 and 5. This is important, as it allows us to extend the understanding of the processes from small scales to large scales, as advocated by Harte [2002] and Seneviratne et al. [2012] (Sects. 1.3.1 and 8.1).

In drought research, some uncertainty is added by the drought identification method, which is described in the previous section (Sect. 8.2.2). On the basis of the evaluations of the different sources of uncertainty described above, the quantification of hydrological drought is regarded as more uncertain than the quantification of meteorological drought (Sect. 6.4.1). In contrast, the high temporal variation in precipitation might result in erratic behaviour that is apparent in meteorological drought and is filtered out in hydrological drought. This is again related to the scale issue mentioned previously [Blöschl and Sivapalan, 1995]. As hydrological droughts generally occur on larger time scales than meteorological droughts, whereby the terrestrial hydrological cycle acts as a low-pass filter of the highly variable meteorological inputs (Sect. 1.2.5), errors in the meteorological forcing are filtered out. This is especially true during dry conditions (more than during floods) because the relative contribution of slow pathways in a catchment to discharge is higher during drought.

According to Oreskes et al. [1994], (hydrological) models cannot be verified to represent the 'true' system. As there is often a 'lack of full access, either in time or space, to the phenomena of interest' models are 'useful for guiding further study', but researchers should not forget that they are 'representations that may resonate with nature, but are not the truth' [Oreskes et al., 1994].

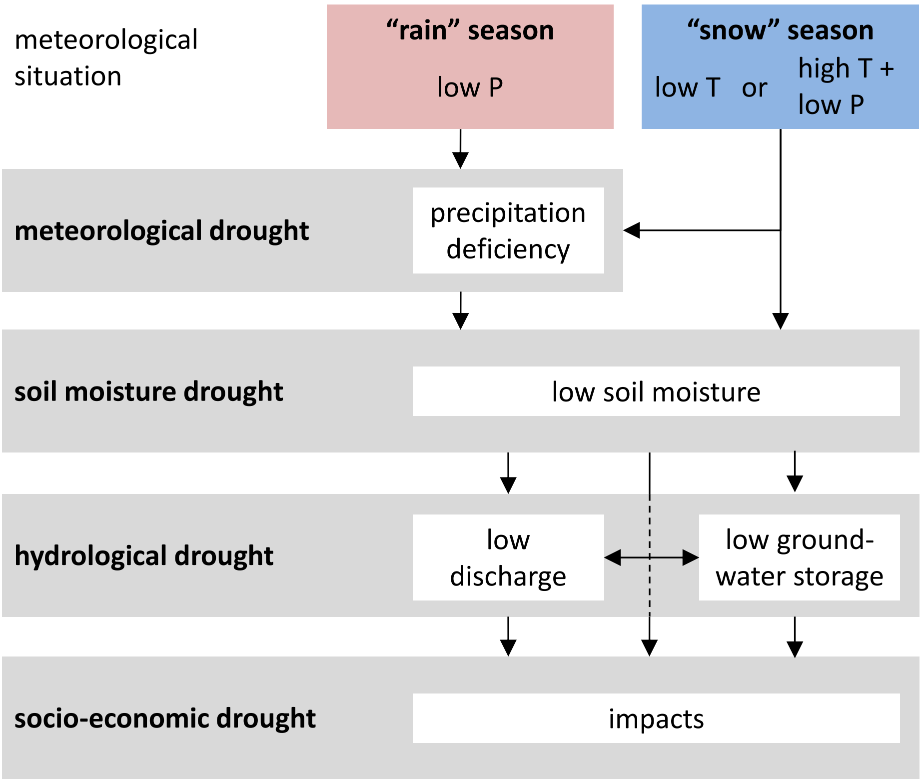


Figure 8.1: Scheme representing different types of drought and their development (based on Fig. 1.1, updated to include seasonality and snow-related processes).

8.2.4 Drought propagation processes

The basic drought propagation processes, e.g. fewer and longer events moving from meteorological drought via soil moisture drought to hydrological drought, an attenuated deficit in hydrological drought compared to meteorological drought, as well as differences between catchments with contrasting climate and catchment characteristics (Sect. 1.2.2), are reproduced by the catchment-scale model (Sect. 5.3.2), by the ensemble of large-scale models (Sect. 6.3.1), and by the synthetic model (Sect. 7.2.2). In this thesis I stressed the importance of the large diversity of the processes underlying drought propagation (e.g. related to temperature and storage), playing a role both on the catchment scale and on the global scale. This diversity in processes is not always reproduced well by all model approaches (Sect. 6.3).

The table summarising these processes for each hydrological drought type (Table 5.5) is repeated here for reasons of clarity, with minor adaptations (Table 8.1). The factors causing the different hydrological drought types are related to precipitation (P control), temperature (T control), or a combination of both. Above-normal evapotranspiration is not found to be the cause of hydrological drought (Sect. 5.6.3). Evapotranspiration can aggravate a drought event [Teuling et al., 2013] and, in a dry season, can prevent recovery (Sect. 5.4.3), but it is not found to be the cause of hydrological drought in the case study areas.

On the basis of the findings presented in this thesis, the drought propagation flow chart of Ch. 1 (Fig. 1.1) should be extended to include seasonality and the influence of temperature on drought development (Fig. 8.1). Fig. 1.2, which actually represents the development of only *classical rainfall deficit drought* events, can also be extended to yield synthetic time series of hydrometeorological variables for each individual hydrological drought type (Fig. 8.2). In this figure the processes summarised in Table 8.1 and Fig. 8.1 are visualised in time series. Time series of the *composite drought* are not included, because this drought type is caused by

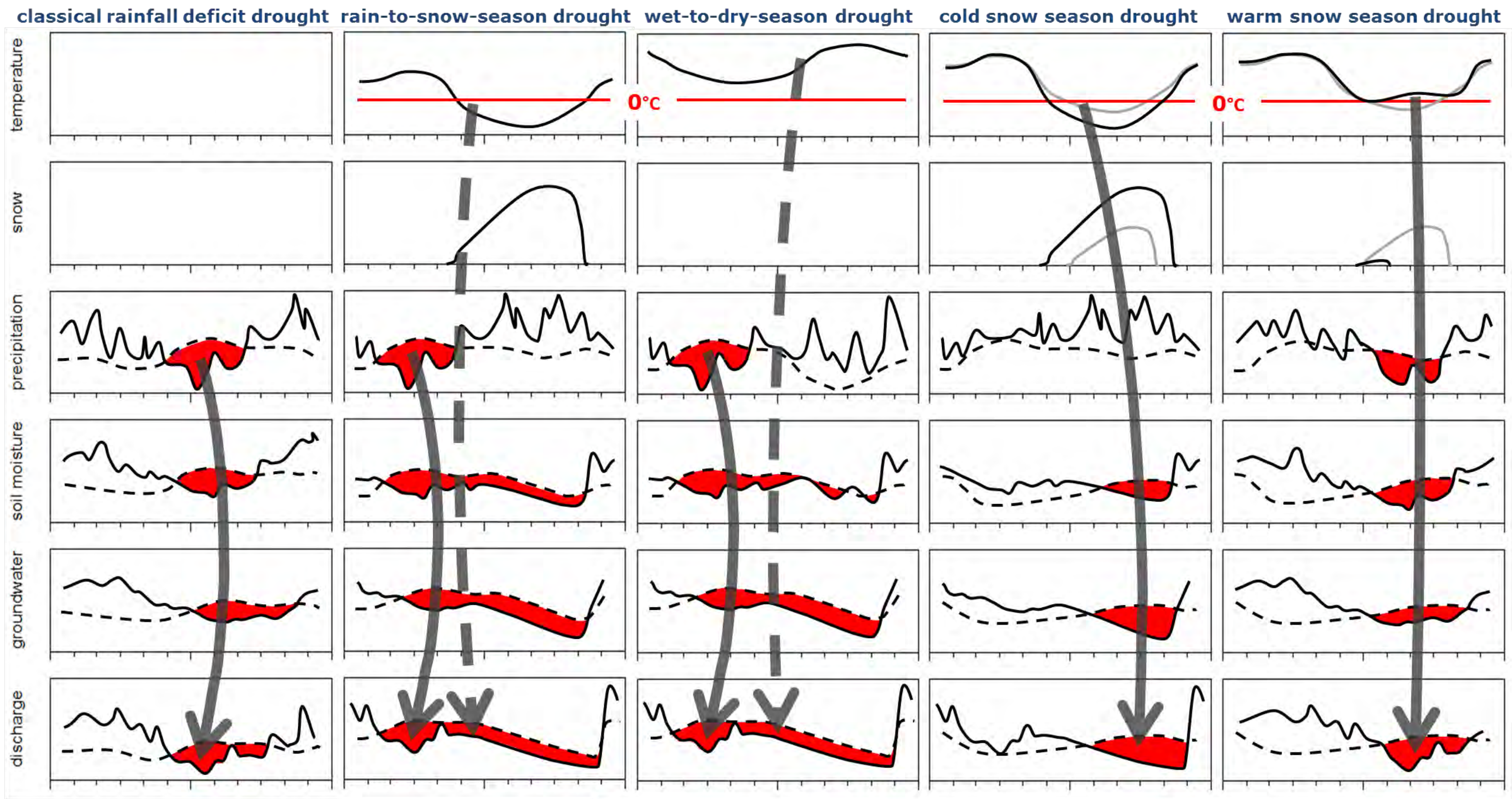


Figure 8.2: Synthetic time series representing the propagation of a meteorological anomaly (precipitation and/or temperature) through the terrestrial hydrological cycle per hydrological drought type. The black lines are the time series of each hydrometeorological variable, the grey lines in the upper two rows are long-term averages of temperature and snow, the dashed lines represent the threshold levels, and the red surfaces indicate drought events. Propagation of drought events is indicated by the arrows, dashed arrows represent a lack of recovery of the hydrological drought (meteorological drought ceased). The *cold snow season drought* is *subtype C* and the *warm snow season drought* is *subtype B* (Table 8.1). Please note that the arrows in this figure are curved differently than in Fig. 1.2a because the order of groundwater and streamflow has been switched.

a combination of processes that is different for every drought event, so no general synthetic time series can be composed. As in Fig. 1.2, the discharge drought signal in Fig. 8.2 resembles either that of soil moisture or that of groundwater, depending on the relative contribution of fast and slow pathways in the catchment (Sect. 1.2.2), which is also reflected in the characteristics of discharge drought events in the case study areas (Sect. 5.3.2). The arrows in Fig. 8.2 can therefore be curved differently for different catchments.

On the basis of Fig. 8.2 I conclude that the flow chart in Fig. 8.1 is a strong simplification of the complex processes underlying the development of hydrological drought and does not consider non-drought development, (non-)recovery of hydrological drought, feedbacks with the atmosphere, etc. In fact, every hydrological drought type should have its own flow chart, similar to Fig. 8.1, indicating both development and recovery of that specific hydrological drought type, as described in detail in Ch. 5. The relevant processes of all hydrological drought types are included in the flow chart of Fig. 8.3. Time is plotted on the x-axis and the y-axis represents a vertical cross-section of the terrestrial hydrological cycle, including the atmosphere and the subsurface. The development and recovery of a drought event are indicated in the grey boxes. Meteorological conditions and catchment characteristics determine the pre-event storage conditions of soil moisture and groundwater/streamflow (*P1* and *P2*). When these are high (*D1a* and *D2a*), no drought develops (Sect. 5.5.3 and Table 8.1). When they are low (*D1b* and *D2b*), the meteorological conditions mentioned at the top of the figure can either directly cause a drought in soil moisture and groundwater/streamflow (*cold snow season drought*; T control only, *M2*), or they can do so via a meteorological drought (other types; P control, or P and T control, *M1* and *M3*). The pre-event storage conditions need not be similar for soil moisture and groundwater/streamflow, i.e. a hydrological drought can develop without a soil moisture drought or vice versa (dashed arrows in the left grey box).

Recovery (*R1a* and *R2a*) of a *cold snow season drought* is effectuated by snow melt (*M5*), recovery of a *classical rainfall deficit drought* or a *warm snow season drought* is caused by a rainfall peak (*M4*). A *classical rainfall deficit drought* that does not recover due to below-zero temperatures or a dry season (*R1b* and *R2b*) develops into a *rain-to-snow-season drought* or a *wet-to-dry-season drought*, respectively. Recovery (*R1a* and *R2a*) of a *rain-to-snow-season drought* is caused by snow melt (*M5*) and a period of high rainfall (*M4*) leads to recovery of a *wet-to-dry-season drought*. When no recovery takes place because of attenuation of rain (or snow melt) in the catchment storage (*R1b* and *R2b*) and hydrological drought types merge, a *composite drought* develops (Sect. 5.4.6 and Table 8.1). Finally, the *composite drought* recovers as well (*R1a* and *R2a*), usually due to above-normal rainfall (*M4*). Hydrological drought events can be prolonged several times by different processes, e.g. first by evapotranspiration in the dry season and then by attenuation of precipitation in the stores. This 'loop' that some hydrological drought types need before recovery is denoted in Fig. 8.3 by the circling arrows in the recovery box (right grey box). After full recovery, the post-event conditions of soil moisture and groundwater/streamflow (*P3* and *P4*) determine what will happen afterwards. These conditions are the pre-event conditions for a next drought event (*P1* and *P2*).

8.2.5 Drought characteristics and typology

Drought characteristics of some hydrological drought types have a distinct pattern. *Composite droughts*, for example, always have long durations (Sect. 5.4.6). Although *classical rainfall deficit droughts*, *cold snow season droughts*, and *warm snow season droughts* have diverse, and even contrasting, causative mechanisms, their drought characteristics largely overlap (e.g. Fig. 5.7), which makes it difficult to distinguish them on the global scale, where they result in a linear pattern at short durations (Sect. 7.3.3). The occurrence of *wet-to-dry-season droughts* or *rain-to-snow-season droughts* in a certain climate type causes a steeper relation between deficit and

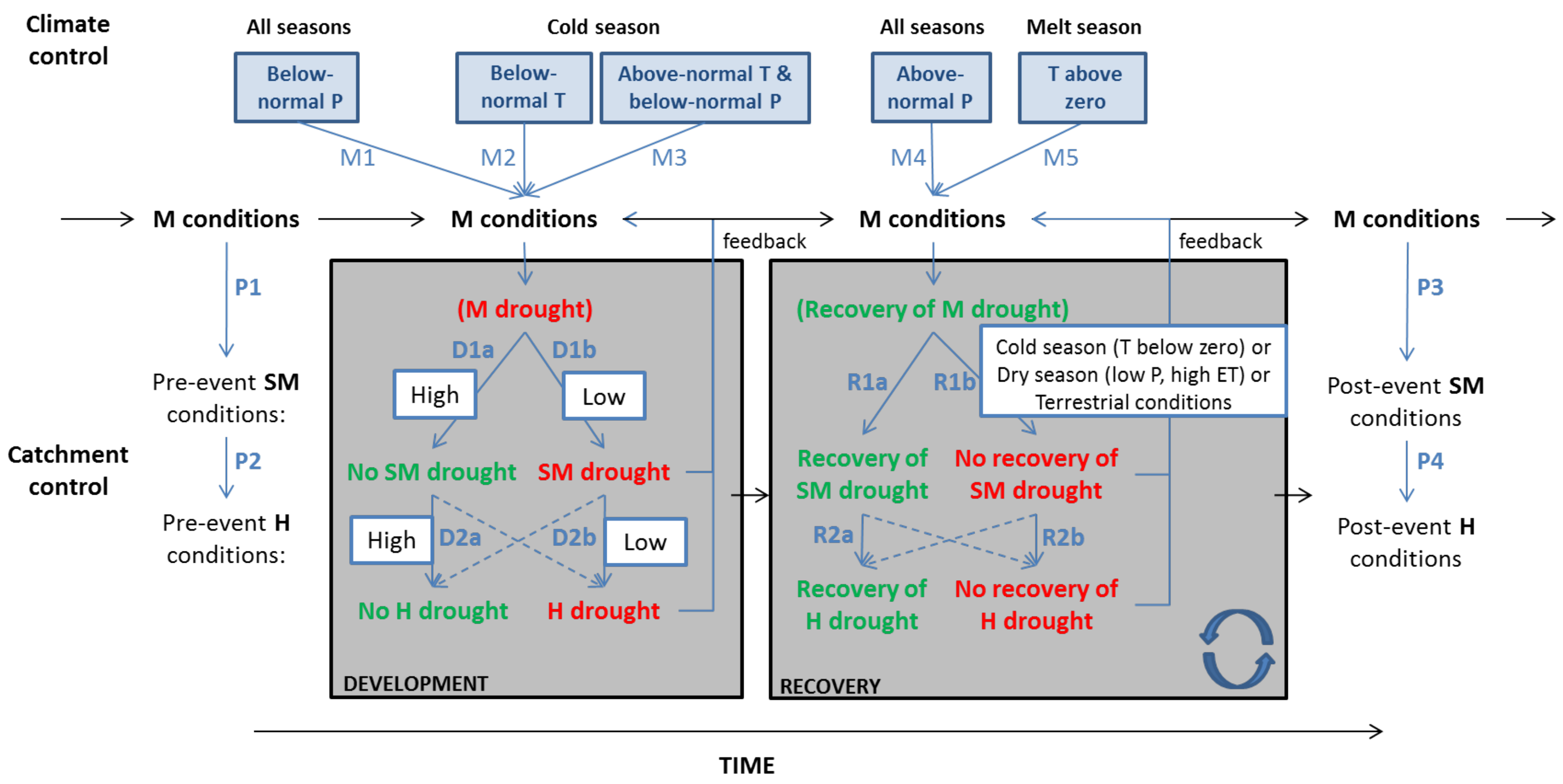


Figure 8.3: Extended scheme representing development and recovery of different hydrological drought types (including feedbacks). M = meteorological, SM = soil moisture, H = hydrological, black arrows represent time, blue arrows represent causal relationships (codes are explained in the text), the brackets around '(recovery of) M drought' indicate the fact that T control can cause a hydrological drought without a meteorological drought, the circling arrows in the recovery box indicate the fact that hydrological drought events can be prolonged several times by different processes. The meteorological conditions leading to development and recovery of drought are dependent of the hydrological drought type (for a description of the hydrological drought types see Ch. 5 and Table 8.1), so not all combinations are possible.

duration at longer durations (causing a bimodal pattern in the drought characteristics of (subsurface) discharge; Sect. 7.3.3). Such a steeper relation could also be observed in the drought characteristics of these drought types on the catchment scale (e.g. Fig. 5.8). When plotting the points of Fig. 5.8 in the transformed probability density fields of the corresponding climate type (Fig. 7.2 and Tijdeman et al. [2012]; for the climate type of the studied catchments see Table 2.1), the drought characteristics derived from the catchment-scale analysis mostly ended up within the 90% probability density field of drought characteristics derived from the large-scale analysis (not shown), confirming a similarity in patterns between the catchment scale and the global scale. Contrary to the pattern on the global scale (Sect. 7.3.3), the catchment-scale pattern could not be quantified yet due to the very low number of points (drought events) at longer durations.

I do not claim that I was able to detect all hydrological drought types that might exist in the world. I did, for example, not find a drought in the snow melt peak due to below-normal snow accumulation in the course of the winter season. This might be related to the use of a monthly threshold level, because a month is longer than the duration of the usually short snow melt peak, so that the peak discharge always exceeds the threshold for a short period. Using a daily threshold level might solve this problem, but studies using a daily-varying threshold [Hannaford et al., 2011; Prudhomme et al., 2011] did not reveal such a drought type in Europe either (see Sect. 6.3.3). A below-normal snow melt peak is important in regions where snow melt is needed, for example, to fill up reservoirs.

Another intriguing drought type was anticipated to occur in monsoonal climates (Sect. 1.2.3). I demonstrated that a *wet-to-dry-season drought* (a hydrological drought at the end of the wet season continuing into the dry season) occurs in that climate type (see Sect. 7.3.3) and dry periods within the monsoon season can be regarded as a *classical rainfall deficit drought*, but another drought type might be related to a complete failure of the monsoon or variability in the start of the monsoon [Bhuiyan et al., 2006; Ratnam et al., 2010]. Such a *'dry-to-wet-season drought'* was not found in the catchment-scale studies (Ch 5), because I did not investigate a catchment with a monsoon climate (Sect. 2.1), nor in the global-scale study (Ch 7), because such climate conditions are probably very rare or its droughts characteristics are similar to those of other hydrological drought types so that they do not stand out in the bivariate probability density fields of Fig. 7.2.

The hydrological drought typology developed in this thesis complements the well-known flood typology of Merz and Blöschl [2003]. The flood typology is also based on expert knowledge using different types of spatial information and time series. The names chosen for the hydrological drought types in Ch. 5 reflect the underlying processes, as was also done by Merz and Blöschl [2003] for the flood types. Two out of five flood types were (partly) governed by T control, whereas for the drought typology T control played a role in four to five out of the six types [Van Loon and Van Lanen, 2012]. And these temperature-controlled drought types also ranked higher than the precipitation-controlled drought types in the selection of the most severe drought events in the case study areas (Sect. 5.5.2). This indicates that temperature-based processes might be more important for hydrological drought than they are for floods.

8.3 Implications and recommendations

8.3.1 Implications for drought research

As stated in Ch. 1, this research mainly aimed at enhancing the understanding of the mechanisms underlying drought propagation. Consequently, its main scientific implication is increased process understanding, which is achieved by the detailed process studies on the catchment scale in Chs. 4 and 5 and their application on the global scale in Ch. 7. I developed a typology of hydrological drought (Ch. 5) that can be a useful tool in drought research. Using this typology, hydrological droughts with different causative mechanisms can be studied separately. For instance, trend studies need to investigate drought events of one type only (e.g. only summer low flows caused by a rainfall deficiency in Stahl et al. [2010, 2012b]), because it facilitates attribution and prevents the cancelling out of effects. Furthermore, climate models and hydrological models can be improved on the basis of their ability to capture the hydrological drought typology, with the aim of a better reproduction of the different processes underlying drought propagation.

In this thesis I demonstrated the importance of temperature for drought development and recovery (resulting in hydrological droughts due to, for example, a longer snow season, a complete snow cover melt in winter or a lengthening of a hydrological drought until next spring; Fig. 8.3). I hope that researchers will pay more attention to the correct representation of drought, especially in the cold regions of the world (i.e. at high latitudes and altitudes). Contrary to expectation (most people think that drought impacts on society are limited to semi-arid regions), cold region droughts can result in major damage. Examples are problems with electricity production and drinking water supply in Scandinavia [e.g. Cattiaux et al., 2010] and livestock mortality and economic loss in regions like Mongolia [Davi et al., 2010; Sternberg, 2010]. It is not without a reason that people in Mongolia have a local name for *rain-to-snow-season droughts*, namely 'Dzud', defined as a summer drought followed by a severe winter [Shestakovich, 2010], and that special aid programs exist for Mongolia because this type of drought generally causes serious loss of livestock [Humanitarian Appeal, 2010; UNICEF, 2010].

As already mentioned in Sect. 8.2.2, an important outcome of this PhD research is the significance of non-linearities in drought propagation. In Ch. 3 I found non-linearities in the response of groundwater storage to abstraction and drought. In Ch. 5 I concluded that the influence of climate and catchment characteristics on drought propagation results in a non-linear relation between hydrological drought and the underlying meteorological conditions. In Ch. 7 non-linear patterns of drought characteristics were found in seasonal climates. Hence, in addition to the non-linear behaviour due to storage mentioned in Sect. 1.2.4, I found non-linear effects as a result of climate control. I therefore strongly emphasise that drought researchers, both in the climate community and the hydrological community, should pay more attention to these non-linearities.

8.3.2 Implications for drought management

The outcomes of this study have a wider applicability than to drought research. Although the focus of this scientific research was not on the implementation of the results in water management, some conclusions can be useful for water managers. In general, it was found that it is important to distinguish between drought and water scarcity (Ch. 3) and between different hydrological drought types (Chs. 4 and 5), because they require different management approaches (Sects. 3.5 and 5.7).

Water managers can use the observational-modelling framework (developed in Ch. 3) to distinguish between drought and water scarcity. In this research I used this framework only to eliminate anthropogenic effects, which allowed me to focus on the natural effects (i.e. drought) in the remainder of this thesis. However, in Ch. 3 some examples were given that show that the framework can also be used to quantify the relative importance of drought and water scarcity and to study changes over time. In the same way, the effects of water management or policy measures to reduce water scarcity can be separated from changes in, for example, climatic inputs. If reliable data are available and a hydrological model is applied that can simulate the relevant human influences (e.g. SIMGRO; Van Lanen et al. [2004b]), the observation-modelling framework can also be used to break down the combined anthropogenic influence into separate parts such as, for example, reservoir operation, groundwater abstraction and land use change. This is relevant for water management because it helps in the attribution of anomalies to different causes.

The hydrological drought typology (developed in Ch. 5) can be applied by water managers to distinguish between different hydrological drought types, provided that data (observed or simulated) of precipitation, temperature and at least one hydrological variable (e.g. groundwater level, discharge) are available. As different hydrological drought types require different measures of prevention and adaptation, water managers can focus their efforts on the most severe or most frequent drought type in the river basin. The typology also provides a tool for drought management and policy making on larger scales. For example, the occurrence of hydrological drought types in ungauged basins can be predicted on the basis of information on climate and catchment characteristics (e.g. using the schematic diagram of Fig. 5.10). Furthermore, the effects of land use change (leading to a change in the response of a catchment to precipitation) and climate change can be qualitatively assessed using the relation between climate and catchment characteristics and the hydrological drought typology. For example, a shift in climate leads to a shift in the occurrence of drought types. This might be important, for example, in regions where winter droughts change from drought types that always end with a snow melt peak to drought types that continue into the summer (see Sects. 4.5 and 5.6). Policy-makers should consider different drought types when designing guidelines for drought management as part of the integrated river basin management plans, just as for example flash floods and other flood types [Merz and Blöschl, 2003] are treated differently in the EU Flood Directive [EU, 2007]. With the elaborated hydrological drought typology developed in this thesis, river basin management, which in many places needs to balance between the two hydrological extremes flood and drought, obtains the appropriate tool to take both extremes into account equally [Van Loon and Van Lanen, 2012].

On the basis of the results of this research I recommend that large-scale models should be used with caution in regions where drought types other than *classical rainfall deficit droughts* play an important role, i.e. in cold and seasonal climates and in catchments with high storage. Large-scale models are suitable in drought research for a number of selective purposes. In Sect. 8.2.2 I argued that on time scales longer than a year non-linear seasonal processes are averaged out and the terrestrial hydrological cycle is primarily driven by precipitation. As discharge simulated by (an ensemble of) large-scale models is strongly related to the precipitation input signal, the models are expected to reproduce year-to-year variability in discharge and, consequently, dry periods on time steps larger than a year. However, one can argue whether, in this case, sophisticated physically-based models are needed when probably a statistical manipulation of the precipitation time series would lead to similar results (see Sect. 8.2.2).

The term 'synchronicity' of hydrological droughts denotes the simultaneous occurrence of hydrological droughts in different parts of a region, e.g. a continent or the globe. In line with the findings in Sects. 5.6 and 6.4.2.1 this synchronicity of hydrological droughts within a region can be adequately investigated with (an ensemble of) large-scale models. The reasons are that synchronicity is expected to be much stronger for hydrological drought types that are governed by P control than for types that are determined by T control (see Table 8.1) and that large-scale models do reproduce P control correctly (Sect. 6.4.2). For these applications the use of a multi-model ensemble is preferred instead of one individual model (see Sect. 6.4).

In drought management indices are often used because they reduce a complex problem to a single number. However, my findings underline that great caution must be taken in using these indices, both for water scarcity and for hydrological drought. The most important recommendation for water managers is that they should not use only meteorological drought indices like SPI and PDSI for hydrological drought analyses (see Sect. 8.2). Additionally, care is needed when different hydrological drought types or droughts in different seasons are studied with the same index. Not all indices are able to capture the different causative mechanisms of drought, which is essential in statistical analysis (e.g. trend studies) and in the assessment of the impacts of, for example, climate change on drought. For water resources management and planning (both on the catchment scale and on larger scales), I advise the use of a variable threshold level method, as has been done by, amongst others, Hannaford et al. [2011], preferably integrated with the consecutive dry day method [Van Huijgevoort et al., 2012b], and to apply this method to as many hydrometeorological variables as possible (i.e. precipitation, soil moisture, groundwater storage, discharge) in order to obtain adequate information on drought propagation.

8.3.3 Further research

On the basis of the outcomes of this project I suggest a number of further steps to be taken in hydrological drought research. First of all, there should be more focus on the use of **observational data**. Experimental projects and satellite missions, including large-scale river flow archives [Hannah et al., 2011], FLUXNET data [Baldocchi et al., 2001] and the Gravity Recovery and Climate Experiment [GRACE; e.g. Swenson et al., 2003; Tapley et al., 2004] provide a wealth of observational data on larger scales, of which the potential for drought research should be explored more intensely. For example, the processes underlying the different hydrological drought types could be explored in observational data.

Second, there are still some intriguing issues in the hydrological processes underlying drought propagation that remain to be understood, e.g.:

- **Catchment control.** While this research focused mainly on climate control (Fig. 1.5), it would be interesting to investigate the possibility of extending the hydrological drought typology to include a greater number of catchment-controlled types and explore the occurrence of these types on larger scales. According to Van Lanen et al. [2012] the influence of catchment characteristics on hydrological drought characteristics is as large as the influence of climate.

- **The effect of evapotranspiration.** The controversy concerning the role of evapotranspiration (called the 'drought paradox' by Teuling et al. [2013]) needs to be investigated in more detail.

Finally, it would also be of great interest to explore how the results of this research can be applied to, for example, **hydrological drought forecasting** and the prediction of the occurrence of hydrological drought types in ungauged basins and in the future. As large-scale models cannot be used to perform this task (Sect. 6.4), simple conceptual model approaches that use the relation between drought propagation and climate and catchment control might be a more appropriate tool (Sect. 8.1). In this thesis the hydrological drought typology was applied based on expert knowledge. Development of an algorithm that can determine hydrological drought types from time series would allow for a quicker and, consequently, wider application of the typology. It is, however, a challenge to identify an appropriate classification approach to capture the expert knowledge used in Ch. 5.

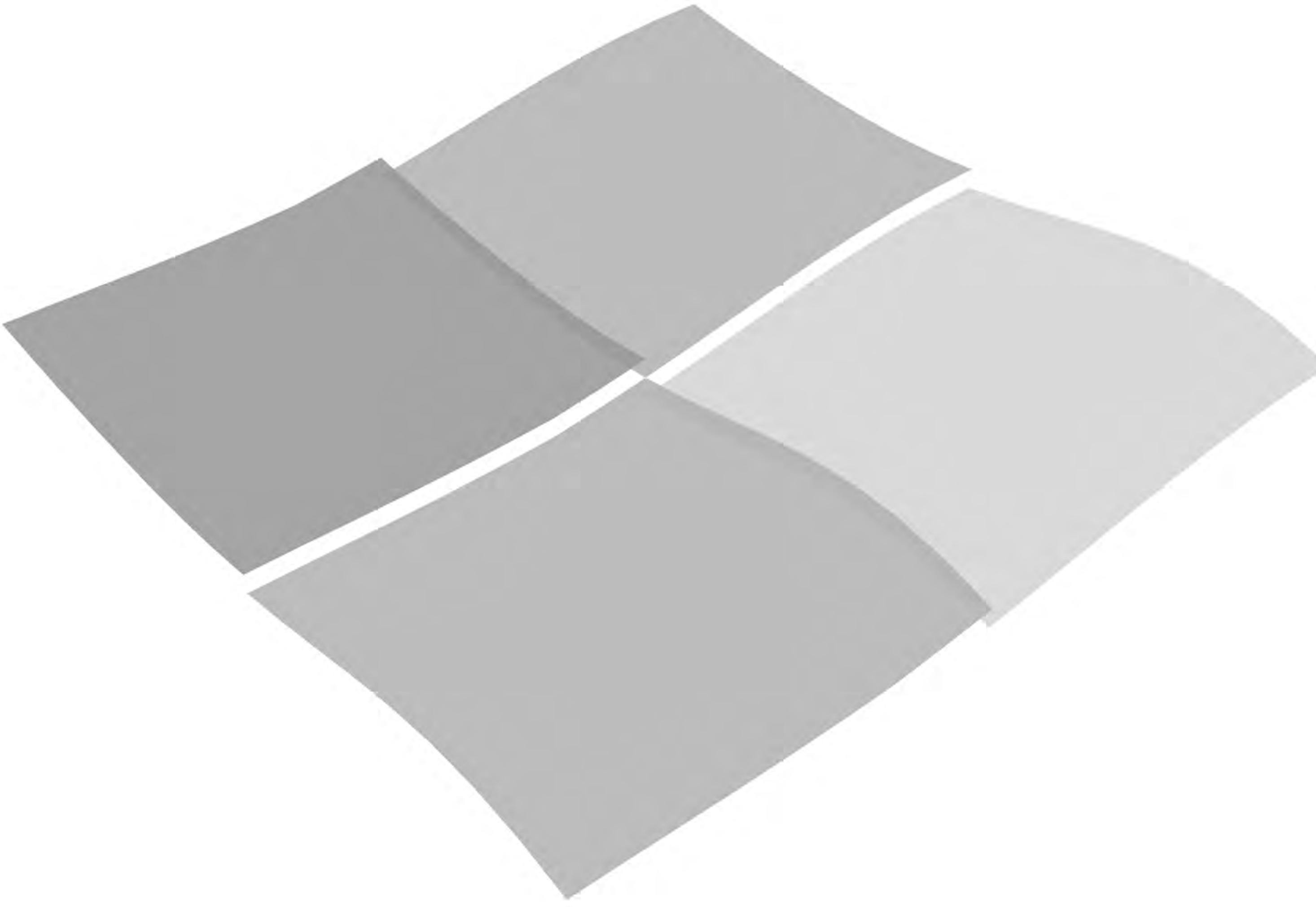
8.3.4 Perspectives

As mentioned before, the social aspects of the coupled sociohydrometeorological system were not studied in this research. Adding the human dimension to drought research could be the right way to bridge the gap between the social and the natural sciences [Sivapalan et al., 2012]. Vincent [2004] states that the interplay of nature-technology-society is important both in the light of generating knowledge and awareness, and in order to resolve conflicts that may arise in situations of water scarcity [Sonnett et al., 2006].

Various studies [e.g. Stahl, 2005; D'Odorico et al., 2010; Hoekstra et al., 2012] used a minimalist modelling framework along the lines of a combination of Newtonian and Darwinian approaches to study the socio-hydrology of drought [Harte, 2002; Sivapalan et al., 2012]. Stahl [2005] performed a statistical analysis and found that the pattern of water-related international relations in arid to sub-humid climates is determined by hydroclimatic variability and population density, and in humid areas by socioeconomic and political factors. D'Odorico et al. [2010] showed that a complete globalisation of water resources would reduce societal resilience to drought. Societies are most able to cope with exceptional drought in a world of so-called 'water solidarity', in which water resources are generally used locally, except in cases of drought-induced famine [D'Odorico et al., 2010]. Hoekstra et al. [2012] performed global analyses of monthly water scarcity using the concept of water footprint and comparing this footprint (per river basin) to natural runoff. Analyses of the relation between the physical causes and dimensions of drought and its impacts are a promising way forward [EU, 2012a], as was shown in Stahl et al. [2012a].

One step further is bridging the gap between science (both natural and social) and policy [Kossida et al., 2012]. Quevauviller et al. [2012] put forward arguments for the strengthening of the links between the scientific and the policy-making communities by discussing the implementation of the EU Water Framework Directive [EU, 2000] with a wide ranch of experts and stake-holders. Quevauviller et al. [2012] see the interaction between science and policy as a two-way process on different levels (EU, national, and regional) that requires a constant dialogue and a mediator mechanism to come to optimal results. My ultimate dream is that in the future a wealth of scientific knowledge and tools will contribute to a successful forecasting of and adaptation to a drought situation, and thus mitigate the extreme economic and social impacts mentioned in the introduction of this thesis (Sect. 1.1).

Appendix A: HBV model improvement



Based on: Van Loon, A.F., Torfs, P.J.J.F., Van Lanen, H.A.J., Uijlenhoet, R., Brauer, C.C., Teuling, A.J.: *Downward (data-based) approach to modelling low flows*, not published.

This annex shows the methodology and results of the attempts to improve the HBV model for low flows.

A.1 Methods

A.1.1 Background

The basis of this research is the conceptual rainfall-runoff model HBV (Sect. 2.3.1). For low-flow modelling not all components of the model need to be considered. Low flows are mainly driven by base flow. Therefore, we can limit ourselves to the improvement of the sub-surface part of the model. To determine the characteristics of this part of the model, we used an analysis of long recessions in observed discharge.

In this appendix we follow the set-up proposed by Jothityangkoon et al. [2001] and Fenicia et al. [2008b]. Both papers present their results in a stepwise manner, based on the idea of moving from a simpler to a more complex model structure in confrontation with data. In contrast to the work of Jothityangkoon and Fenicia and co-workers, however, we do not focus on one specific test catchment, but on a number of contrasting catchments in Europe (Sect. 2.1). Our hypothesis is that, based on a multi-catchment analysis of long recession periods in observed discharge, a general model structure for the sub-surface part of a rainfall-runoff model and robust model parameters can be found that lead to a better simulation of low flows.

A.1.2 General framework

The general methodological framework of this research stems from ideas of the 'downward approach' [Sivapalan et al., 2003]. To test our hypothesis on the improvement of low-flow modelling we used a systematic data-driven approach. The boundary conditions of our methodology are:

1. The model structure should be data-based, but not purely black-box. As the final model should be applicable for process understanding and low-flow prediction, the balance of mass is the guiding principle. We adopted the approach of Romanowicz [2007], where 'model structures are accepted only when they can be interpreted in a physically meaningful manner'.
2. The model structure should be parsimonious. We start with the simplest model structure possible, and systematically increase complexity to overcome deficiencies in low-flow simulation.
3. The focus is specifically on low flows, but with the aim not to compromise high flows. In the final model, correct timing of peaks is not considered important, but correct volumes of high flows are.
4. We used a number of contrasting catchments (Sect. 2.1), so that the identified model structure is not site-specific. Our goal is to find a single model structure with different parameters for the different catchments.

We used the HBV model (Sect. 2.3.1) as the basis for this research. We replaced the sub-surface compartment of HBV (the part that generates the slow flow component; the 'response function' in Fig. 2.5) with an adapted model structure. In other words, we took the recharge simulated with HBV and used this as the input to another model for the sub-surface part. According to Seibert (pers.comm., 2009), the obtained parameters of the snow and soil moisture routines of HBV are robust when changing the lower part of the model. The structure of the alternative model for the subsurface part of HBV is based on recession analysis of long recessions in observed discharge. In the final model, we allow for as little parameter calibration as possible, so we used recession analysis not only to determine model structure, but also parameter values (like in Jothityangkoon et al. [2001]).

A.1.3 Recession analysis and modelling

Recession analysis is widely used in many areas of hydrological research. For applications to low flows, Tallaksen [1995] and Smakhtin [2001] give a comprehensive overview.

For the selection of recession periods from the time series of observed discharge of the five studied catchments, an objective method was used. First, we performed a 30-day moving average on the observed discharge data. This was needed because of the relatively low data quality in the low-flow range [Rees et al., 2004]. Next, we defined a recession period as a period with decreasing or constant discharge. Minor increases in discharge (0.1 %), which can occur as a result of measurement error, especially under low-flow conditions, were allowed. Furthermore, during recession periods recharge simulated with HBV should be less than $1.0\,\mathrm{mm\,d^{-1}}$. Recharge was taken instead of precipitation, because we wanted to consider both summer and winter low flows and precipitation is not a useful variable in catchments with snow conditions in winter. Finally, we selected only recession periods longer than 20 days, because our focus is on long low-flow periods and not on short recessions in between high flow events. Some examples

Table A.1: Recession characteristics of the selected catchments Narsjø (Norway), Upper-Metuje and Upper-Sázava (Czech Republic), Nedožery (Slovakia), and Upper-Guadiana (Spain)

catchment	time period	no. of recessions	no. of recessions per year	avg. length	max. length
Narsjø	1958–2007	63	1.26	138	226
Upper-Metuje	1982–1999	50	2.78	49	137
Upper-Sázava	1962–2000	107	2.74	37	102
Nedožery	1974–2006	74	2.24	43	123
Upper-Guadiana	1960–1979*	41	2.05	87	250

* undisturbed period: rest of time series is highly impacted by human influence

are presented in Fig. A.1. For the five studied catchments this resulted in a total of 335 recession periods with a average length of 65 days (for recession information of the studied catchments separately, see Table A.1).

In the process of finding a new model structure for simulating low flows, a number of steps were followed. In this paragraph the steps are described in general. Details are presented in Sect. A.2.

Step 1. Fitting of the selected recession periods:
Different model structures were tested on the selected recession periods. We started with the most simple structure, one linear reservoir, and progressively more complex outflow functions and combinations of reservoirs were evaluated. First, each recession was fitted individually. Subsequently, when good fits were achieved, four recessions were fitted simultaneously, using a moving window through the recessions of each catchment. Based on the results of these fits, the best-performing model structure (for all studied catchments) was selected.

Step 2. Fixing the parameters for each catchment:
As Sivapalan et al. [2003] stated: 'model parameters ideally are also derived from the data analysis'. In this research parameters for the sub-surface part of the rainfall-runoff model were also determined from recession analysis. Again, a systematic approach was adopted in which first one parameter was fixed and the other(s) fitted, then a second parameter was fixed and the other(s) fitted, until all parameters were fixed (except for the starting volume of the reservoir). This was done for each catchment individually, because our goal was to use the same model structure with different parameters for the contrasting catchments.

Steps 1 and 2 can be described as event-based modelling. Event-based modelling has up to now mainly been used for flood events [Jain and Indurthy, 2003; Maneta et al., 2007; Bahat et al., 2009], but it can also be applied to low flows.

Step 3. Include periods in between recessions:
In this step we go from event-based to continuous modelling. The objective is to make sure that the simple structure selected on the basis of recession analysis can reasonably capture the total hydrograph (volume of peak, not exact timing). Different pre-processors were tested and the way of fitting parameters of these pre-processors was examined.

A.1.4 Model fitting and evaluation

In this section, we present some tools that were used in all steps of the recession analysis and modelling introduced in Sect. A.1.3. All fitting was done using an R-based fitting tool [Torfs et al., 2010]. Both the starting volume of the reservoir(s) and the parameter(s) of the outflow relationship(s) were fitted. The data used for fitting was observed discharge, smoothed by a moving average of 30 days, and log-transformed. Log-transformed discharge data is used for calibration in other studies, e.g. Seibert [1999] and Romanowicz [2007], to give relatively more weight to low flows. In recession analysis, fitting on log-transformed discharge puts more emphasis to the tail of the recessions. Tests showed that fits to low flows were better than when using untransformed discharge data.

Similarly to the HBV modelling (Sect. 2.3.2), we wanted to evaluate the agreement between simulated and observed discharge during recession periods using the Nash-Sutcliffe efficiency [Nash and Sutcliffe, 1970]. The normal Nash-Sutcliffe efficiency is based on a comparison of the performance of a model with a 'no model' case. Nash and Sutcliffe [1970] used the mean of the observed discharges as the 'no model' case. However, for evaluating recessions, which have a comparable shape, i.e. decreasing, the mean is not a representative case to compare the model with. Any reasonable fit will perform better than the mean. Therefore, we decided to choose another 'no model' case, namely exponential decay. Our 'no model' case was calculated by an exponential function based on the difference between the logarithm

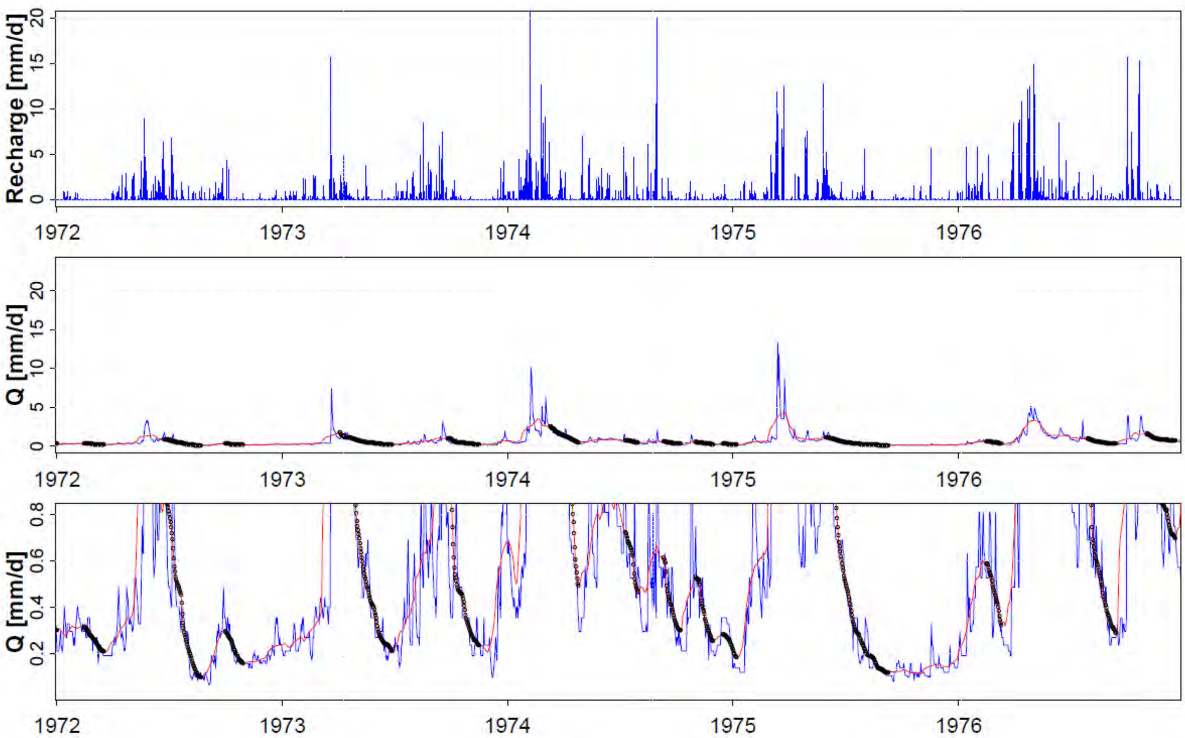


Figure A.1: Example of selected recessions from the Upper-Sázava catchment; upper panel: recharge, middle panel: discharge, lower panel: lower range of discharge (blue line = observed discharge, red line = 30 days moving average of observed discharge, black dots = recession periods).

of the first and the last observed discharge value in the recession period, divided by the duration of the recession period (Fig. A.2). The logarithm was again used to put relatively more weight to low flows. We refer to this adapted Nash-Sutcliffe coefficient as NS_{exp}.

A.2 Search for an improved HBV model

As described before, this research follows the set-up proposed by Jothityangkoon et al. [2001] and Fenicia et al. [2008b]. That means that, in line with the 'downward' approach, the description of the results follows the steps taken in the process of model adaptation. In each step we systematically describe the summarised results for all catchments.

A.2.1 Step 1. Fitting of the selected recession periods

We tested different model structures (number and configuration of reservoirs and outflow relationship(s)) on the selected recession periods (Sect. A.1.3). We started with fitting each recession period individually.

To start simple, we first used a single linear reservoir. The linear reservoir is widely acknowledged as the starting point for all conceptual rainfall-runoff models [Eriksson, 1971] and is found to be the simplest way of reproducing the hydrological behaviour of a catchment [O'Kane, 2006]. Especially for modelling baseflow or hydrograph recessions, the linear reservoir is much used [Tallaksen, 1995; Fenicia et al., 2006; Lehner et al., 2006]. The outflow function of a linear reservoir is:

$$Q = a \cdot S \tag{A.1}$$

in which Q is discharge (in mm d^{-1}), S is storage in the reservoir (in mm), and a is a parameter (in d^{-1}). In this appendix the linear reservoir function is denoted as LIN. Many authors pointed out non-linearities in the hydrological system [Tallaksen, 1995; Fenicia et al., 2006; Clark et al., 2009; Botter et al., 2010]. In our catchments we noticed that many recessions had a characteristic shape, namely a steep beginning and a long tail. This corresponds to the findings of other studies, in which recessions showed a declining slope during low-flow periods [De Wit et al., 2007; Basu et al., 2010; Birk and Hergarten, 2010]. A single

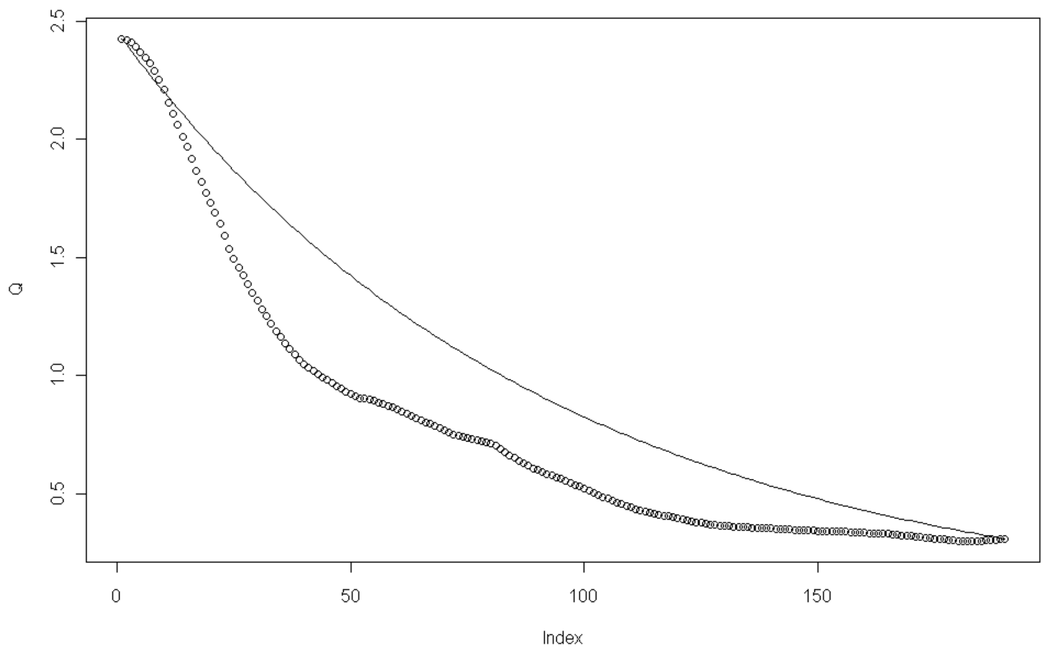


Figure A.2: Example of observed discharge (o) and the 'no model' case based on exponential decay (-).

linear reservoir is not suitable for modelling this non-linear behaviour. Therefore, we also tested two different non-linear reservoirs, with outflow functions:

$$Q = a \cdot S^b \tag{A.2}$$

and

$$Q = aS \cdot e^{-bS} \tag{A.3}$$

in which Q is discharge (in mm d^{-1}), S is storage in the reservoir (in mm), and a and b are parameters (a: d^{-1}, b: - in Eq. A.2 and mm^{-1} in Eq. A.3). The first (conventional) non-linear reservoir function (Eq. A.2, also known as 'power law' function) is denoted as NONLIN_CONV, and the second (Eq. A.3, also known as 'gamma' function) as NONLIN_EXP. In the remainder of this section these abbreviations are used.

When all recessions in all studied catchments are taken into account (Table A.2a), the linear reservoir gives a good fit (NS_{exp}= 0.58–0.72) and both non-linear reservoir functions give an even better fit (NS_{exp} = 0.66–0.87). In four catchments the NONLIN_EXP fit gave best results, i.e. Narsjø Upper-Metuje, Upper-Sázava, and Nedožery. Only in Upper-Guadiana the NONLIN_CONV fit gave better results. However, for Upper-Sázava and Nedožery the differences between both non-linear fits are very small.

In a test phase of this research, many other non-linear reservoir functions were explored, like $Q = a \cdot e^{bS}$ and $Q = aS \cdot (1 + bS^c)$. The results did not show an improvement in the fit of the recessions. Some authors found threshold behaviour in recessions [Basu et al., 2010]. To test the threshold behaviour in our catchments we tried to fit a reservoir with more than one outflow. This configuration did not lead to an improvement of fits and was therefore disregarded in this study.

As a next step, we explored the effect of multiple reservoirs, because the non-linear response of a catchment can also be obtained through the sum of linear hillslope responses [Hall et al., 2012; Tallaksen, 1995; Clark et al., 2009; Harman et al., 2009]. This approach is used in many conceptual hydrological models, among which the HBV model (Fig. 2.5). We tested different combinations of reservoirs:

- two LIN reservoirs (2LIN);
- one LIN reservoir and one NONLIN_CONV reservoir;
- one LIN reservoir and one NONLIN_EXP reservoir.

The reservoirs were placed in parallel. Moore [1997] stated that, in recession analysis, it is not possible to distinguish between two reservoirs in parallel or in series. In this research serial configurations of the

Table A.2: Average NS_{exp} values for a) single reservoir fits, recessions fitted separately; b) multiple reservoir fits, recessions fitted separately; and c) single reservoir fits, four recessions fitted simultaneously

model	Narsjø	Upper-Metuje	Upper-Sázava	Nedožery	Upper-Guadiana
a)					
LIN	0.72	0.65	0.58	0.60	0.65
NONLIN_CONV	0.83	0.68	0.66	0.67	0.76
NONLIN_EXP	0.87	0.70	0.67	0.67	0.67
b)					
2 LIN	0.83	0.64	0.62	0.63	0.74
1 LIN and 1 NONLIN_CONV	0.86	0.75	0.67	0.70	0.80
1 LIN and 1 NONLIN_EXP	0.88	0.78	0.68	0.71	0.77
c)					
LIN	0.61	0.27	0.05	0.18	0.42
NONLIN_CONV	0.78	0.29	0.16	0.31	0.58
NONLIN_EXP	0.84	0.48	0.21	0.34	0.48

reservoirs have not been tested.

When all recessions in all catchments are considered (Table A.2b), the same pattern in NS_{exp} values can be observed as in Table A.2a, i.e. for Narsjø Upper-Metuje, Upper-Sázava, and Nedožery the NONLIN_EXP fit gave best results, whereas for Upper-Guadiana the NONLIN_CONV fit gave better results. Contrary to the findings of Tallaksen [1995], Clark et al. [2009], and Hall et al. [2012], the double linear reservoir fit is (slightly) worse than the single non-linear reservoir fit (compare row 1 in Table A.2b with row 2 in Table A.2a) for all catchments.

Overall, the NS_{exp} is higher for the multiple reservoir fits than for the single reservoir fits (0–15 % higher). However, in the multiple reservoir fit more parameters are fitted, almost twice as many. Therefore, we prefer the single reservoir fit with less degrees of freedom [Beven, 2000; Wagener et al., 2004].

After the individual fits we fitted four recession periods simultaneously (using a moving window through the recessions of each catchment) to check the robustness of selected single reservoir model structure. This assessment can be regarded as preliminary work for Step 2.

Results of all recessions in all catchments are shown in Table A.2c. Again, the same pattern in NS_{exp} values can be observed as in Table A.2a, i.e. for Narsjø Upper-Metuje, Upper-Sázava, and Nedožery the NONLIN_EXP fit gave best results, whereas for Upper-Guadiana the NONLIN_CONV fit gave better results. The four-recession fit has lower NS_{exp} values (3-90 % decrease) than the individual fit. However, some interesting differences can be observed between the catchments. In the Narsjø catchment fits decreased only slightly. This indicates that recessions have similar shape and can be modelled with the same model structure and parameters. In the Upper-Sázava and Nedožery catchments fits decreased considerably. These catchments apparently have a large diversity in recession shape, thus they cannot be modelled with the same model structure and parameters. The Upper-Metuje and Upper-Guadiana catchments are in between: fits decreased, but not drastically.

On the basis of the results of Step 1, the single NONLIN_CONV reservoir (Eq. A.2) is selected as the best model structure. It is the most simple configuration with good results in all catchments. The slightly better performance of the NONLIN_EXP reservoir in some catchments does not justify the use of this more complex model structure. From Step 1 we can conclude that recessions of all studied catchments can be modelled adequately using the same model structure (a single non-linear reservoir).

A.2.2 Step 2. Fixing the parameters for each catchment

Not only model structure, but also parameters for the selected model structure were determined from recession analysis. This was done for each catchment individually. The single NONLIN_CONV reservoir has two parameters that were fitted (Eq. A.2). We started by fixing the b parameter. For each catchment, the arithmetic average of the fitted b parameters of all recessions was taken. Subsequently, all recessions

Table A.3: NS_{exp} for the single NONLIN_CONV reservoir fits of all recessions (335) in all studied catchments, fixing parameter a and b

model	Narsjø	Upper-Metuje	Upper-Sázava	Nedožery	Upper-Guadiana
free a and b	0.78	0.29	0.16	0.31	0.58
fixed b, free a	0.77	0.29	0.14	0.29	0.42
fixed a and b	0.76	0.27	0	0.07	0.15

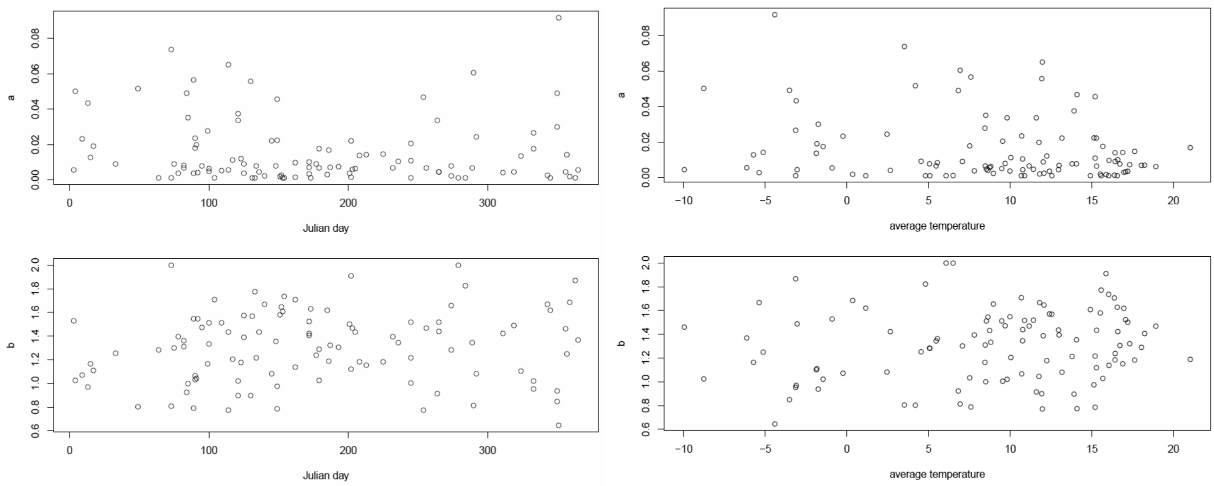


Figure A.3: Scatter plot of parameters a and b against Julian day and average temperature during the recession period, for the Upper-Sázava catchment.

were again fitted with only one free parameter. This resulted in a comparable fit; NS_{exp} values decreased only slightly (Table A.3). We continued by fixing the a parameter at its arithmetic average. Subsequently, all recessions were fitted with fixed parameters, so only the starting volume of the reservoir was fitted. On average this resulted in lower NS_{exp} values (Table A.3), but differences between catchments are large (2–100 % decrease). The Narsjø and Upper-Metuje catchments show good results, NS_{exp} values decrease only slightly and visual inspection shows a comparable fit for all recessions. So, in these catchments, all recessions can be modelled with the same parameters. For the Upper-Sázava, Nedožery, and Upper-Guadiana catchments this is not the case. The large decrease in NS_{exp} values (Table A.3) indicates that in those catchments recessions are very different and should not be modelled with the same parameter set.

This observation corresponds with earlier findings in the literature that a time series including different hydrological processes cannot be modelled with the same parameter set [e.g. Staudinger et al., 2011]. However, the good results for Narsjø and Metuje show that this does not hold for all catchments. Long recessions in those two catchments are governed by one dominant process. In the Narsjø catchment long recessions occur almost always in winter when water is stored as snow and ice. The shape of the recession is determined by the release of water by bogs and lakes in the catchment. In the Upper-Metuje catchment, the dominant process is aquifer discharge.

For the Upper-Sázava, Nedožery, and Upper-Guadiana catchments the processes underlying long recessions are apparently very variable. A plausible explanation would be a seasonal variation. However, we could not find any seasonal dependence of parameter values [Tallaksen, 1995; Griffiths and Clausen, 1997; Uijlenhoet et al., 2001]. No correlation was found between day of the year and parameters, nor between temperature and parameters. An example is presented for the Upper-Sázava catchment (Fig. A.3), which had the lowest NS_{exp} (Table A.3) for the fit with fixed parameters.

For continuous modelling the use of fixed parameters is inevitable. Even for the catchments that showed bad results, a best parameter set should be chosen (Table A.4). A striking conclusion from Table A.4 is that the slow responding catchments Upper-Metuje and Upper-Guadiana have lower b parameters than the fast responding catchments.

A.2.3 Step 3. Include periods in between recessions

The goal of this third step was to select the most simple structure that can reasonably capture the total hydrograph. Now we move from event-based to continuous modelling. In the previous steps the starting

Table A.4: Fixed parameters for the single non-linear reservoir fits (NONLIN_CONV) in all studied catchments

catchment	a	b
Narsjø	0.0019	1.4
Upper-Metuje	0.0049	1.2
Upper-Sázava	0.0079	1.3
Nedožery	0.0051	1.3
Upper-Guadiana	0.014	1.2

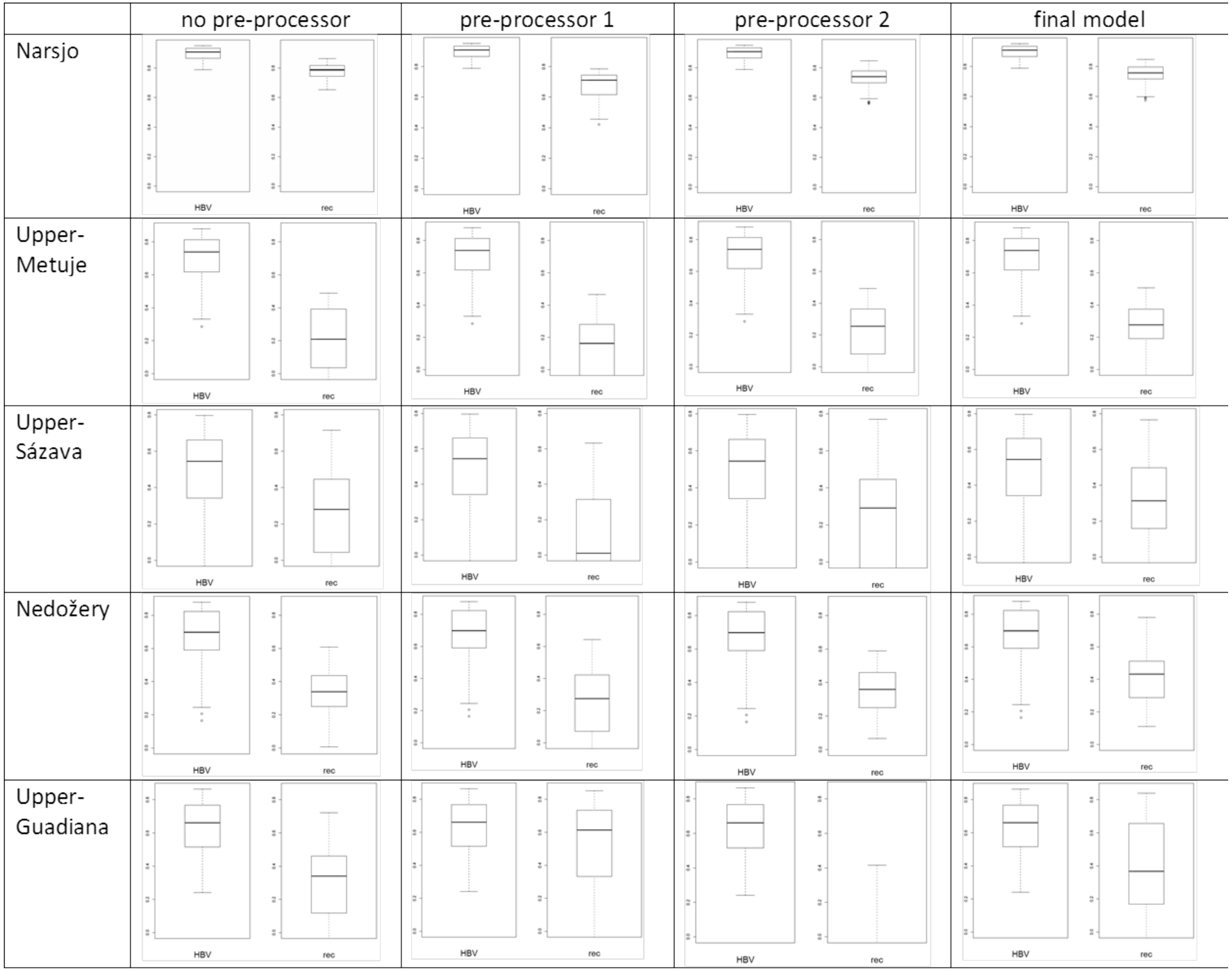


Figure A.4: Boxplots of NS_{exp} values for all selected recessions (fitted per four recessions simultaneously) and options for pre-processors, comparison with HBV.

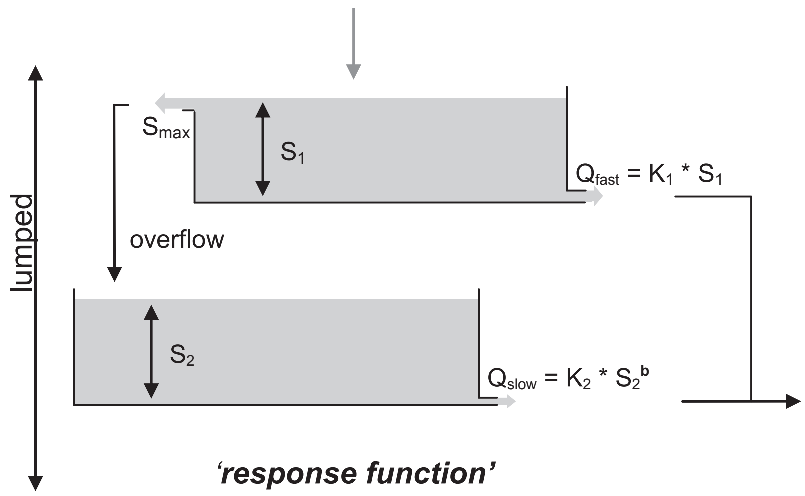


Figure A.5: Structure of the adapted HBV model, in which the response function of HBV is replaced by a linear reservoir overflowing into a non-linear reservoir (compare with Fig. 2.5).

volume of the reservoir(s) was fitted and only the recession periods were regarded. In this step, also the periods in between recessions were considered and the recharge taken from the upper part of the HBV model forms the inflow of the reservoir. To check whether this will not cause problems we tested (for every recession) if there was enough recharge in the period before the recession to fill the reservoir up to the fitted starting volume of Step 2. That was the case for most of the recessions, which gave us enough confidence to continue. We had to divide the recharge from HBV in a fast component flowing directly to the stream and a slow component flowing into our selected NONLIN_CONV reservoir. We tried different pre-processors to get this division right and we evaluated the results by comparing the Nash-Sutcliffe values based on the logarithm of discharge (Sect. 2.3.2) of the same period (including four recessions and the periods in between) and by visual inspection of hydrographs. Again, we moved through the time series by four recessions (and the periods in between) at a time.

We started simple by using a fixed fraction that divides the recharge from HBV into flow entering the NONLIN_CONV reservoir and rest flow. For this configuration there was only one parameter that needed calibration. Results were showing many small peaks on top of the low flows during long recessions and Nash-Sutcliffe values were lower than the ones of HBV (Fig. A.4 - first column). Therefore, we tried a pre-processor that could delay the inflow of recharge into the NONLIN_CONV reservoir. We chose a linear reservoir, that would only need two parameters for calibration (pre-processor 1). This did remove the small peaks, but the fit with the low flows decreased considerably, both at visual inspection of hydrographs and considering Nash-Sutcliffe values (Fig. A.4 - second column). The only exception was the Upper-Guadiana catchment, where fits improved. Consequently, we tried an overflowing reservoir (pre-processor 2). Thus, the inflow of the NONLIN_CONV reservoir was not determined by the outflow of the linear reservoir. Instead, this outflow was routed directly to the river as fast flow and only the water in the reservoir above a certain threshold was used as inflow to the NONLIN_CONV reservoir. This option gave slightly better results, again with the exception of the Upper-Guadiana catchment (Fig. A.4 - third column). Nash-Sutcliffe values were still lower than those of the HBV model, but the shape of the recessions was modelled quite well (not shown).

The difficulties that we encountered were that the beginning of recessions could not be modelled as well as during event-based fitting (Steps 1 and 2) and that in continuous modelling also inflow into the reservoir occurred during the recession period. This of course influenced the parameters of the recession. Therefore, we tried calibrating the parameters of the NONLIN_CONV reservoir in this last step, instead of using the pre-determined parameters from Step 2 (Table A.4). This improved the results considerably for the Upper-Sázava and Upper-Guadiana catchments, but for Narsjø Upper-Metuje and Nedožery catchments NS_{exp} values were comparable (Fig. A.4 - last column). However, for all catchments NS_{exp} values were still lower than those of the HBV model.

Choosing a 'best' model structure is not straightforward, because all model structures generate good results in some periods and poor results in other periods. However, we decided that, based on the work presented in this appendix, the single non-linear reservoir combined with a overflowing linear reservoir gave best overall results in all catchments. If this model structure would have been implemented in HBV light (Sect. 2.3.1), the response function of the model would have looked like Fig. A.5.

Appendix B: HBV model validation

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For drought studies it would be most desirable to have long (tens of years), complete time series of observed fluxes and state variables. Unfortunately, these data were not available for this study and generally are very rare, in particular for sufficiently contrasting catchments. The main purpose of using a model in this research was the simulation of state variables (snow accumulation, soil moisture, groundwater storage) for which no long time series exist. For the Upper-Guadiana catchment modelling was also required to naturalise the disturbed time series (Ch. 3). In this appendix we present a validation of the HBV model (Sect. 2.3.1) on discharge and groundwater by comparing simulations with observations using graphs of time series and annual and monthly values of the 50th and 80th percentile of the duration curves. Additionally, a summary of the validation of snow and soil moisture from previously published reports is given.

For the Narsjø catchment model results showed the highest ln Reff (0.90; Table 5.1). This is due to the very regular seasonal pattern of discharge, dominated by yearly recurring winter low-flow conditions (Fig. B.1 – upper row), that can be captured quite well with a rainfall-runoff model like HBV.

This regular seasonal pattern is also visible in the groundwater levels in the Narsjø catchment (Fig. B.2 – upper row). Groundwater in this catchment had a good fit to observations, as can be seen from the percentiles in Table B.1. The coefficient of determination, r^2, was quite high with 0.72, and visual comparison indicated a good ability of the model to reproduce the general dynamics of the groundwater table (Fig. B.2 – upper row). Simulated soil moisture percentiles showed a reasonable agreement to the percentiles of observations (Table B.1, upper rows), although the coefficient of determination was quite low ($r^2 = 0.35$). This low value can be explained by deviations in winter, i.e. decreasing observed values vs. constant simulated values (not shown). This is partly because the TDR probes measured available water content, which is lower than stored water content due to soil frost [Hohenrainer, 2008], and partly because HBV does not simulate outflow from the soil moisture store when evaporation is zero (Fig. 2.5). Hohenrainer [2008], who used the HBV model with similar settings, calibration procedure and objective function, stated that the onset and duration of drought periods were captured reasonably well by the model, justifying the use of simulated soil moisture and groundwater series for drought analysis.

For the Upper-Metuje, Upper-Sázava, and Nedožery catchments, ln Reff was around 0.65 (Table 5.1). This is lower than the value for the Narsjø catchment, because seasonal variation is much more irregular in these catchments (Fig. B.1 – 2nd, 3rd and 4th rows). Figure B.1 shows that the hydrographs of Upper-Metuje and Nedožery are better reproduced than that of Upper-Sázava. However, the yearly and monthly percentiles of Upper-Sázava are still reasonable (Table B.1).

For the Upper-Metuje catchment a validation against observed groundwater levels was performed. The coefficient of determination was high ($r^2 = 0.79$) and the yearly and monthly percentiles show similar values (Table B.1). Visual comparison indicated a good ability of the model to reproduce the general dynamics of the groundwater table (Fig. B.2 – 2nd row).

For the Upper-Sázava catchment both snow storage and groundwater simulations were validated. For groundwater the coefficient of determination was quite low ($r^2 = 0.46$). This is probably due to the lack of representativeness of the groundwater well for groundwater storage in the entire catchment. Actually, most of the catchment consists of crystalline rock, whereas the groundwater well is located in sedimentary rocks. Furthermore, some measurement problems were recorded at this well [Rakovec et al., 2009]. This results in deficiencies in reproducing the time series of observed groundwater levels (Fig. B.2 – 3rd row), but the yearly and monthly percentiles are still very similar (Table B.1). The reason for this difference is that an incorrect simulation of the timing of high and low flows is not reflected in the percentiles in Table B.1, while it has a large impact on the coefficient of determination. For snow the coefficient of determination was reasonable ($r^2 = 0.57$). The general pattern of the simulation agrees well with observed values (not shown, see Rakovec et al., 2009).

For the Nedožery catchment both snow storage and groundwater simulations were validated. For groundwater the coefficient of determination was high ($r^2 = 0.74$) and the yearly and monthly percentiles also showed similar values (Table B.1). Visual inspection of the time series of observed and simulated groundwater levels showed that the general dynamics of the groundwater table were reproduced rather well (Fig. B.2 – 4th row). For snow visual comparison between simulated and observed snow cover showed that the model was able to simulate snow in the correct period and with the correct volume (not shown, see Oosterwijk et al., 2009).

For the Upper-Guadiana catchment the numbers in Table 5.1 were obtained with the DELAY version of the HBV model (Sect. 2.3.1 and Fig. 2.5) for the calibration and validation period combined (1960–1980). Model results of the STANDARD version, which was used for the other catchments, showed 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 best reproduced recessions. It showed less peaky behaviour and no zero-flows as compared to the STANDARD version (Fig. B.1 – lower row). Therefore, the results of the DELAY version were used for further analysis in the Upper-Guadiana catchment. In the other catchments Nash-Sutcliffe values and visual inspection of time series revealed that the DELAY version had less agreement with observations (not shown). The good results of the Upper-Guadiana model in the calibration and validation period (both undisturbed, see Table 2.1) justify the extrapolation of the model to the disturbed period (i.e. naturalisation of disturbed time series for the period after 1980; Ch. 3).

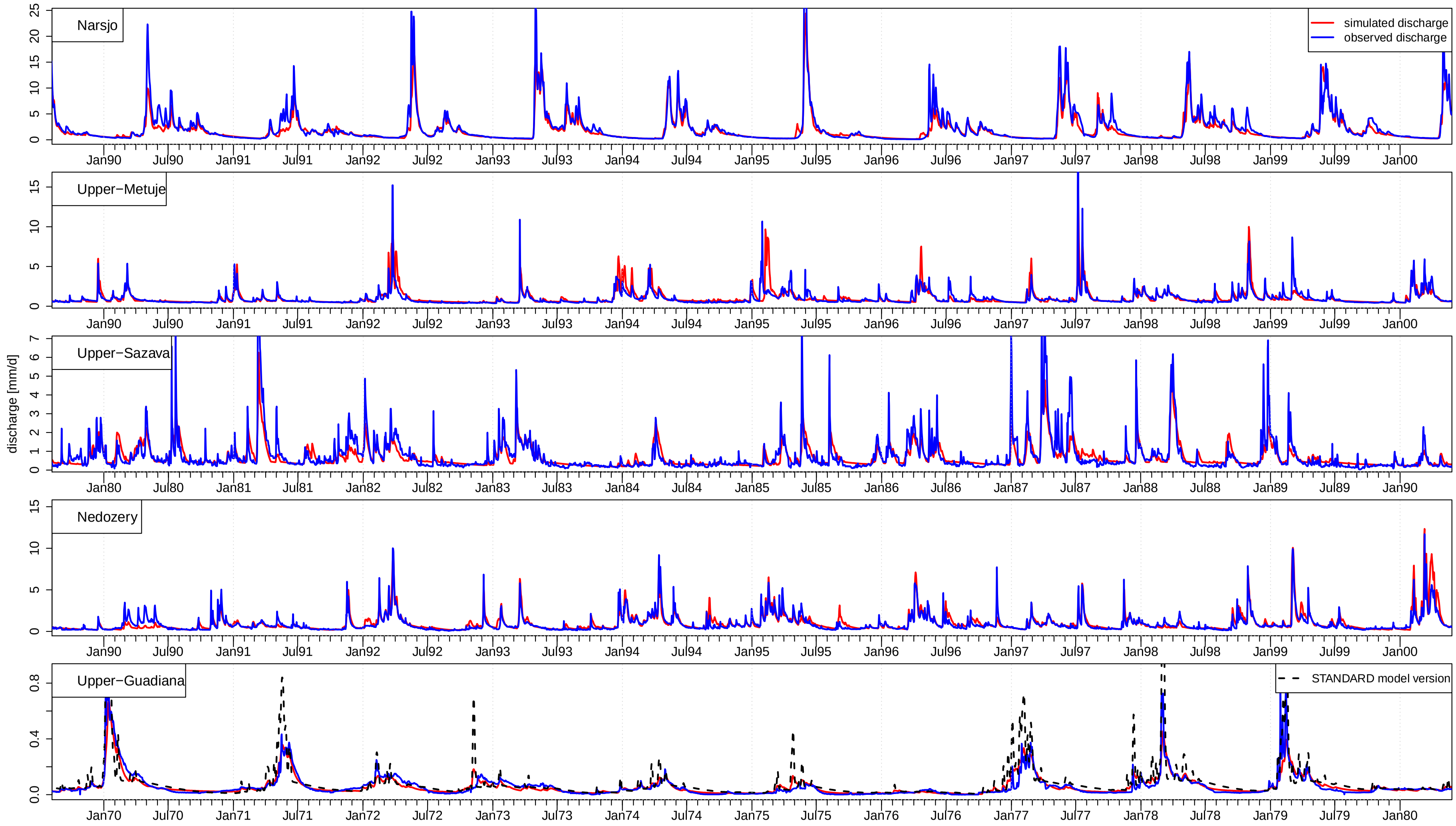

Figure B.1: Time series of observed and simulated discharges for all catchments and different time periods: upper row Narsjø catchment 1990–2000, 2nd row Upper-Metuje catchment 1990–2000, 3rd row Upper-Sázava catchment 1980–1990, 4th row Nedožery catchment 1990–2000, and lower row Upper-Guadiana catchment 1970–1980.

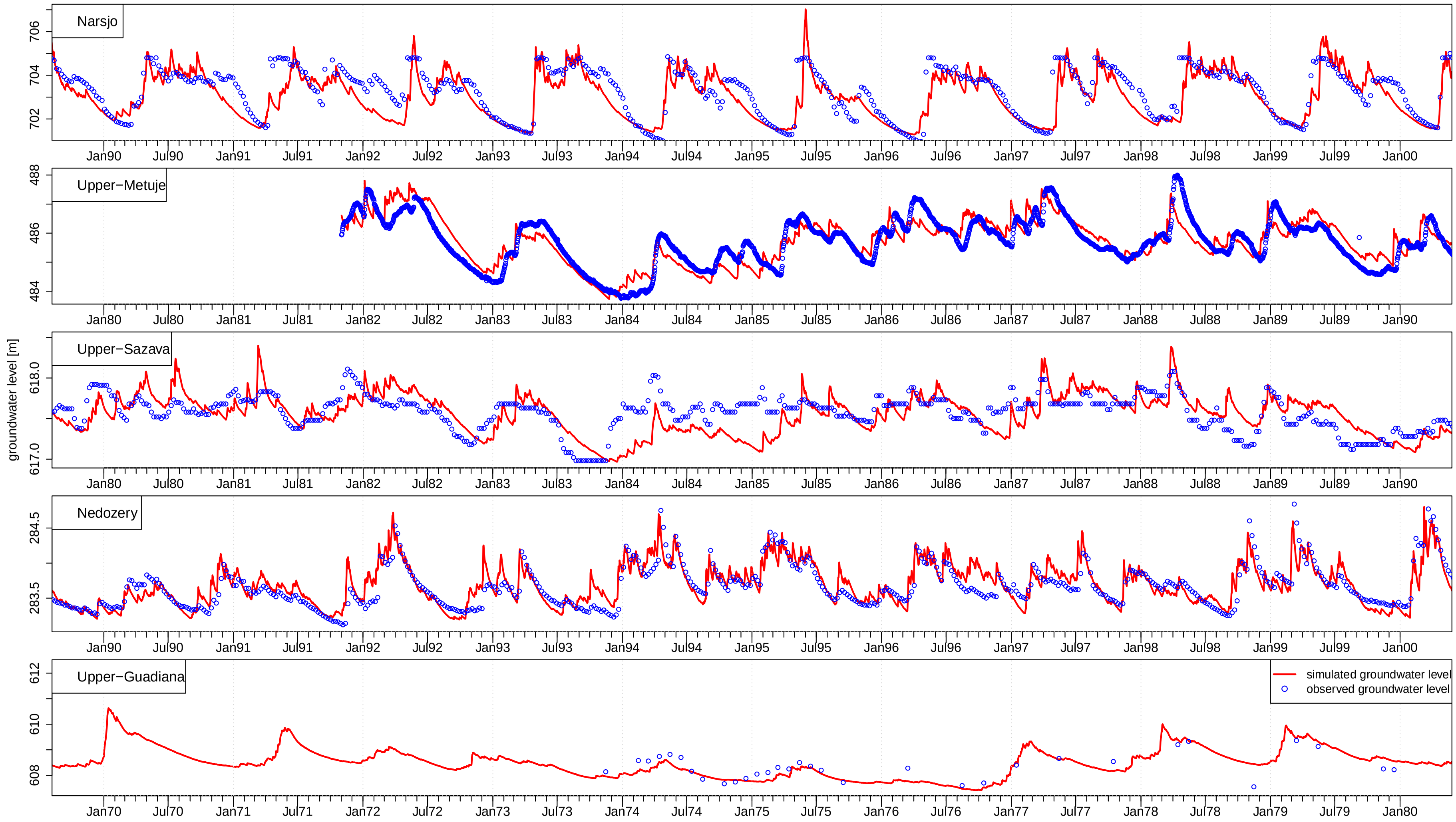

Figure B.2: Time series of observed and simulated groundwater levels for all catchments and different time periods: upper row Narsjø catchment 1990–2000, 2nd row Upper-Metuje catchment 1980–1990, 3rd row Upper-Sázava catchment 1980–1990, 4th row Nedožery catchment 1990–2000, and lower row Upper-Guadiana catchment 1970–1980.

Table B.1: Annual and monthly values of the 50th and 80th percentile of the duration curves of soil moisture (only Narsjø), groundwater and discharge

			Annual	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
Narsjø	SM_{sim}	50 %	0.2926	0.3013	0.3032	0.3051	0.3182	0.3393	0.2755	0.2553	0.259	0.2646	0.2814	0.2932	0.2979
		80 %	0.2584	0.2839	0.2858	0.287	0.297	0.3144	0.2322	0.2141	0.2092	0.2254	0.2503	0.2721	0.2783
	SM_{obs}	50 %	0.28	0.25	0.255	0.26	0.29	0.49	0.34	0.29	0.24	0.28	0.28	0.28	0.26
		80 %	0.22	0.15	0.16	0.16	0.2222	0.464	0.278	0.23	0.15	0.238	0.26	0.24	0.172
	GW_{sim}	50 %	703.1	702.1	701.8	701.6	701.6	703.9	703.9	703.8	703.9	703.9	703.7	703.1	702.5
		80 %	701.8	701.9	701.6	701.5	701.5	702.9	703.5	703.3	703.4	703.4	703.3	702.8	702.2
	GW_{obs}	50 %	703.5	702.7	702.3	701.9	701.8	704.8	704.4	703.8	703.5	703.6	703.8	703.7	703.6
		80 %	702.1	702	701.7	701.4	701.2	704.7	704.1	703.5	702.4	702.2	703.3	703.3	702.6
	Q_{sim}	50 %	1.04	0.487	0.354	0.271	0.3005	5.152	3.191	2.213	2.013	1.809	1.402	0.985	0.6945
		80 %	0.388	0.396	0.2892	0.2148	0.207	1.68	2.149	1.471	1.205	1.161	1.058	0.8114	0.5698
	Q_{obs}	50 %	1.046	0.45	0.342	0.279	0.284	6.56	4.353	2.229	1.895	1.91	1.883	1.164	0.706
		80 %	0.36	0.36	0.279	0.223	0.223	2.511	2.52	1.227	0.883	1.101	1.141	0.868	0.553
Upper-Metuje	GW_{sim}	50 %	485.6	485.4	485.6	486.1	486.1	486	485.8	485.6	485.5	485.3	485.2	485.2	485.1
		80 %	484.9	484.5	484.9	485.1	485.6	485.5	485.4	485.2	485	484.8	484.6	484.4	484.4
	GW_{obs}	50 %	485.6	485.4	485.6	486.2	486.8	486.4	486	485.5	485.2	485.2	485.2	484.9	485.2
		80 %	484.7	484.3	484.9	485.1	486	485.7	485.5	485	484.7	484.4	484.2	484	484.2
	Q_{sim}	50 %	0.687	0.845	0.8085	1.334	1.181	0.724	0.6785	0.661	0.625	0.6285	0.594	0.5955	0.653
		80 %	0.563	0.535	0.587	0.6508	0.7618	0.651	0.614	0.593	0.5698	0.541	0.508	0.483	0.514
	Q_{obs}	50 %	0.686	0.803	0.8405	1.291	1.186	0.773	0.645	0.602	0.566	0.581	0.557	0.582	0.648
		80 %	0.523	0.546	0.557	0.743	0.8936	0.654	0.5494	0.523	0.4898	0.492	0.47	0.474	0.5116
Upper-Sázava	GW_{sim}	50 %	617.5	617.5	617.5	617.7	617.7	617.7	617.6	617.6	617.5	617.5	617.4	617.4	617.4
		80 %	617.3	617.2	617.2	617.4	617.5	617.5	617.4	617.4	617.4	617.3	617.2	617.2	617.2
	GW_{obs}	50 %	617.5	617.6	617.6	617.7	617.7	617.6	617.5	617.5	617.5	617.4	617.3	617.5	617.6
		80 %	617.3	617.4	617.4	617.5	617.5	617.4	617.4	617.3	617.2	617.2	617.2	617.2	617.4
	Q_{sim}	50 %	0.426	0.411	0.5965	0.954	1.024	0.4895	0.427	0.4115	0.4045	0.37	0.344	0.339	0.3845
		80 %	0.316	0.2834	0.336	0.4134	0.5678	0.3814	0.35	0.333	0.317	0.297	0.27	0.2548	0.277
	Q_{obs}	50 %	0.494	0.58	0.6745	1.218	1.08	0.632	0.441	0.366	0.3455	0.402	0.355	0.375	0.5285
		80 %	0.27	0.263	0.3102	0.4666	0.5936	0.329	0.2686	0.2296	0.211	0.237	0.213	0.237	0.296
Nedožery	GW_{sim}	50 %	283.7	283.7	283.7	283.8	283.9	283.8	283.7	283.7	283.6	283.6	283.5	283.6	283.6
		80 %	283.5	283.5	283.5	283.7	283.8	283.7	283.6	283.5	283.5	283.5	283.4	283.4	283.5
	GW_{obs}	50 %	283.7	283.7	283.8	283.9	283.9	283.8	283.7	283.6	283.6	283.5	283.5	283.5	283.6
		80 %	283.5	283.5	283.6	283.7	283.8	283.7	283.6	283.5	283.4	283.4	283.3	283.4	283.4
	Q_{sim}	50 %	0.588	0.568	0.6425	1.403	1.283	0.671	0.5965	0.548	0.448	0.4545	0.39	0.4575	0.521
		80 %	0.361	0.3114	0.4132	0.5584	0.7214	0.5274	0.4418	0.386	0.3264	0.292	0.277	0.2708	0.31
	Q_{obs}	50 %	0.598	0.682	0.7815	1.559	1.425	0.823	0.577	0.448	0.355	0.326	0.365	0.46	0.601
		80 %	0.328	0.446	0.4604	0.8234	0.9148	0.572	0.3888	0.287	0.221	0.212	0.239	0.298	0.368
Upper-Guadiana (1960–1980)	GW_{sim}	50 %	608.2	608.3	608.4	608.4	608.4	608.4	608.3	608.1	608	NA*	607.8	607.9	608
		80 %	607.7	607.8	607.8	607.8	607.9	607.9	607.8	607.7	607.6	NA*	607.5	607.6	607.7
	GW_{obs}	50 %	608.3	608.2	608.3	608.4	608.7	608.8	608.5	608.2	607.7	NA*	607.7	607.9	608.1
		80 %	607.9	608.1	608.2	608.3	608.4	608.6	608.4	608.2	607.7	NA*	607.7	607.7	607.9
	Q_{sim}	50 %	0.044	0.0735	0.091	0.118	0.094	0.065	0.05	0.036	0.02801	0.028	0.028	0.032	0.037
		80 %	0.023	0.0268	0.043	0.05	0.057	0.04	0.031	0.022	0.015	0.012	0.011	0.021	0.022
	Q_{obs}	50 %	0.04	0.0755	0.098	0.136	0.103	0.076	0.051	0.025	0.014	0.013	0.016	0.022	0.036
		80 %	0.015	0.035	0.047	0.048	0.063	0.051	0.036	0.016	0.008	0.007	0.01	0.015	0.021

* = not enough groundwater observations to determine percentiles for Guadiana in September.

For the Upper-Guadiana catchment a validation against observed groundwater levels was performed in part of the undisturbed period for which data was available. In this catchment many groundwater observation wells have been installed. Some of the wells showed quite a poor correlation with simulated values, but the well with best correlation had an r^2 value of 0.83. Visual comparison indicated a good ability of the model to reproduce the general dynamics of the groundwater table, although the data points in the undisturbed period were limited (Fig. B.2 – lower row). Table B.1 also showed that intra-annual variation in groundwater levels was reproduced well by the model.

In this catchment the lack of good-quality data was the main problem in modelling; there were many gaps in the observations, observed discharge data on the low-flow reach had a 'staircase' pattern and human influence reduced the period available for calibration. Despite these limitations and the large size of the Upper-Guadiana catchment and its complex interaction between groundwater and surface water (rivers and wetlands), the simple conceptual model HBV performed surprisingly well. Probably, the HBV model was in the range of 'optimal model complexity' for the given availability of data [see Fig. 1 in Grayson et al., 2002].

In summary, we can conclude that the performance of the HBV model in the study catchments is acceptable for drought analysis, as was also found by Van Huijgevoort et al. [2010], and hence for the identification of different hydrological drought types.

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Abstract

Drought is a severe natural disaster resulting in high economic loss and huge ecological and societal impacts. In this thesis drought is defined as a period of below-normal water availability in precipitation (meteorological drought), soil moisture (soil moisture drought), or groundwater and discharge (hydrological drought), caused by natural variability in climate. Drought propagation is the change of the drought signal as it moves from anomalous meteorological conditions to a hydrological drought through the terrestrial part of the hydrological cycle. The objective of this PhD research is to investigate the processes underlying drought propagation and their relation with climate and catchment characteristics, both on the catchment scale and on the global scale.

The catchment-scale studies are based on five headwater catchments in Europe with contrasting climate and catchment characteristics. In one of these case study areas, anthropogenic influence on the water system was significant, resulting in severe water scarcity. As I only study natural processes in this thesis, there was a need to separate drought (as defined in this thesis) from human-induced water scarcity in this case study area. I proposed an observation-modelling framework that consists of a hydrological model to simulate the 'naturalised' situation and an anomaly analysis method to quantify drought and water scarcity events. Both the time series and the anomaly characteristics of the 'disturbed' and 'naturalised' situation were compared to quantify human and natural influences on the hydrological system.

After simulation of hydrometeorological variables of all case study areas with a conceptual hydrological model and drought identification with the variable threshold level method, time series and characteristics of drought events were analysed. I classified the drought events into six hydrological drought types that are the result of the interplay of temperature, precipitation, evapotranspiration and storage in different seasons. The most common hydrological drought type develops as a result of a rainfall deficit. However, in the development of the most severe hydrological drought events temperature and storage-related processes play an important role, for example through a lack of recovery of the drought.

As I aimed to investigate drought propagation also on larger scales, I tested an ensemble mean of a number of large-scale models (both land-surface models and global hydrological models) on their ability to reproduce the drought propagation processes found in the case study areas. The large-scale models did simulate general aspects of drought propagation (e.g. fewer and longer drought events in discharge than in precipitation), but the above-mentioned effects of temperature and storage-related processes were only partly reproduced. In the large-scale model ensemble, daily runoff reacted almost immediately to changes in precipitation, resulting in important deficiencies in drought simulation in cold and semi-arid climates and regions with large storage. For the time being, this limits the use of large-scale models for the study of processes underlying drought propagation on a global scale.

Consequently, I used a synthetic conceptual hydrological model to study drought propagation on the global scale. I focused on climate control by isolating forcing effects from effects of catchment properties. The drought characteristics (duration and

deficit combined) of both soil moisture and subsurface discharge exhibited strongly non-linear patterns in seasonal climates. The non-linear effects in soil moisture drought were caused by the fact that the development of soil moisture droughts in warm seasonal climates is limited by the wilting point. Hydrological droughts in both warm and cold seasonal climates showed a strong increase of deficit with duration due to a lack of recovery in the dry season or snow season, respectively. This effect was strongest in cold seasonal climates, which indicates that for the development and recovery of within-year hydrological drought temperature is an important factor.

The overall conclusion of this research is that, although drought is a complex, non-linear phenomenon with drought characteristics varying with climate type and catchment characteristics, generic patterns can be derived that reflect the different hydrological processes underlying drought propagation. These processes result in different hydrological drought types that are shown to play a role both on the catchment scale and on the global scale. The non-linear effects of snow and storage-related processes on drought are not incorporated sufficiently in the currently-used large-scale models and drought indices. Possible future steps include more focus on catchment control, in particular the representation of storage, and the role of temperature and evapotranspiration. Additionally, the findings of this research can be applied to hydrological drought forecasting, prediction in ungauged basins, and prediction under global change.

Samenvatting

Droogte is een ernstige natuurramp; het veroorzaakt veel economische schade en heeft enorme ecologische en sociale gevolgen. In dit proefschrift definieer ik droogte als een periode van benedennormale waterbeschikbaarheid in de neerslag (meteorologische droogte), in het bodemvocht (bodemvochtdroogte) of in het grondwater en de rivierafvoer (hydrologische droogte), die wordt veroorzaakt door de natuurlijke variabiliteit in het klimaat. De voortplanting van droogte is de ontwikkeling van bodemvochtdroogte en hydrologische droogte uit afwijkende meteorologische condities en treedt op binnen het terrestrische deel van de hydrologische cyclus. In dit onderzoek worden de processen bestudeerd die ten grondslag liggen aan de voortplanting van droogte en de relatie van die processen met het klimaat en met stroomgebiedskarakteristieken, zowel op stroomgebiedsschaal als op wereldschaal.

De deelstudies op stroomgebiedschaal maken gebruik van vijf stroomgebieden in de bovenloop van Europese rivieren. Het klimaat en de stroomgebiedskenmerken verschillen sterk per gebied. In een van deze stroomgebieden was de menselijke invloed op het hydrologisch systeem zeer groot, wat leidde tot ernstige watertekorten. Omdat ik in dit proefschrift alleen natuurlijke processen bestudeer en dus de menselijke invloed zo veel mogelijk wil uitsluiten, moest in dit stroomgebied onderscheid worden gemaakt tussen droogte zoals deze in dit onderzoek wordt gedefinieerd, en door de mens veroorzaakte watertekorten. Daartoe heb ik een waarneming-modelkader voorgesteld waarin een hydrologisch model om de 'vernatuurlijkte' situatie te simuleren wordt gecombineerd met een methode om afwijkingen te analyseren. Om de menselijke en de natuurlijke invloeden op het hydrologisch systeem te kwantificeren vergeleek ik de tijdreeksen en afwijkingen van de 'door mensen verstoorde' situatie met die van de 'vernatuurlijkte' situatie.

Nadat de hydrometeorologische variabelen van alle studiegebieden waren gesimuleerd met een conceptueel hydrologisch model en droogtes gekwantificeerd met de variabele drempelwaardemethode, heb ik de tijdreeksen en de droogtekarakteristieken geanalyseerd. Ik vond daarin hydrologische droogtes met verschillende oorzaken, die ik heb geclassificeerd in zes typen. Deze droogtetypen zijn het resultaat van de wisselwerking tussen temperatuur, neerslag, verdamping en berging in verschillende seizoenen. Het meest voorkomende hydrologische droogtetype wordt veroorzaakt door een neerslagtekort. Maar in de ontwikkeling van de ernstigste hydrologische droogtes spelen de temperatuur en bergingsgerelateerde processen een belangrijke rol, bijvoorbeeld door een gebrek aan herstel in het hydrologisch systeem.

Omdat ik mij ook tot doel had gesteld de voortplanting van droogte op grotere schaal dan die van stroomgebieden te onderzoeken, heb ik het gemiddelde van een aantal grootschalige modellen (zowel landoppervlaktemodellen als grootschalige hydrologische modellen) getest op hun vermogen om de processen van de voortplanting van droogte zoals ik die heb gevonden in de studiegebieden te reproduceren. Hoewel de grootschalige modellen de algemene aspecten van de voortplanting van droogte goed simuleerden (bijvoorbeeld dat er minder hydrologische droogtes zijn dan me-

teorologische droogtes, maar dat ze langer duren), werden de bovengenoemde effecten van temperatuur en bergingsgerelateerde processen maar voor een deel gereproduceerd. De modelgemiddelde dagelijkse afvoer reageerde bijna onmiddellijk op veranderingen in de neerslag, wat resulteerde in belangrijke tekortkomingen in het nabootsen van droogtes in koude en semi-aride klimaattypen, alsmede in gebieden met veel berging. Dit beperkt vooralsnog de mogelijkheden om grootschalige modellen te gebruiken in onderzoek naar de processen die ten grondslag liggen aan de voortplanting van droogte op wereldschaal.

Vervolgens heb ik een synthetisch conceptueel hydrologisch model gebruikt om de voortplanting van droogte op wereldschaal te bestuderen. Ik heb me gericht op de invloed van het klimaat door de effecten van meteorologische invoergegevens te isoleren van de effecten van stroomgebiedskenmerken. De droogtekarakteristieken (een combinatie van de duur van droogtes en hun deficiet) van zowel bodemvocht als ondergrondse afvoer lieten sterk niet-lineaire patronen zien in seizoensgebonden klimaten. De niet-lineaire effecten in bodemvocht ontstonden doordat de ontwikkeling van bodemvochtdroogtes in warme seizoensgebonden klimaten beperkt wordt door het verwelkingspunt. Hydrologische droogtes in zowel warme als koude seizoensgebonden klimaten lieten een sterke toename van het deficiet met de duur zien, wat werd veroorzaakt doordat het hydrologisch systeem zich in het droge seizoen, respectievelijk het sneeuwseizoen niet herstelt van een droogte. Dit effect was het sterkst in de koude seizoensgebonden klimaten, wat aangeeft dat de temperatuur een belangrijke factor is voor het ontstaan en het herstel van hydrologische droogtes die korter dan een jaar duren.

De algemene conclusie van dit onderzoek is dat, hoewel droogte een complex, niet-lineair fenomeen is met karakteristieken die variëren per klimaattype en stroomgebiedskenmerken, er generieke patronen gevonden kunnen worden die een goede afspiegeling zijn van de verschillende hydrologische processen die ten grondslag liggen aan de voortplanting van droogte. Deze processen resulteren in verschillende typen hydrologische droogte, waarvan ik heb laten zien dat ze een rol spelen zowel op stroomgebiedsschaal als op wereldschaal. De niet-lineaire effecten van sneeuw en bergingsgerelateerde processen op droogte zijn niet voldoende geïntegreerd in de huidige grootschalige modellen en droogte-indices. Mogelijke vervolgstappen na dit onderzoek zijn onder andere een grotere focus op de invloed van stroomgebiedskarakteristieken, in het bijzonder het effect van berging, alsmede de rol van temperatuur en verdamping. De resultaten van dit onderzoek kunnen worden toegepast bij het voorspellen van hydrologische droogte uit meteorologische condities, ook in stroomgebieden waar weinig metingen beschikbaar zijn, en bij mondiale milieuverandering.

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Wageningen, April 2013

Anne F Van Loon

Curriculum Vitae

Anne Frederike Van Loon was born in Arnhem, the Netherlands, on 25 May 1981. She attended secondary school at the Stedelijk Gymnasium Arnhem, from which she graduated cum laude in 1999. She followed the BSc programme 'Soil, Water and Atmosphere' at Wageningen University, the Netherlands, and specialised in hydrology and soil science. In 2002 she graduated cum laude. She attended the 4-month international course 'From Mountain to Fjord - Geology and Ecology of Western Norway' in Sogndal, Norway. In 2003 she worked as a researcher for the NGO PROFAFOR in Ecuador on a project related to the effects of forestry plantations in the Andes on soil properties, hydrology and erosion.

When she returned in the Netherlands, Anne continued her studies at Wageningen University in the MSc programme 'Hydrology and Water Quality'. Her first thesis project on the topic of Hydrogeology was entitled 'Water flow and tidal influence in a mangrove system in Saigon-Dong Nai river delta, Vietnam'. It included three months of field work in a mangrove area in Vietnam and extended data analysis, and resulted in a peer-reviewed paper in Aquatic Botany. Her second thesis project on the topic of Hydraulics was entitled 'Fine sediment transport in the Western Wadden Sea'. It was conducted at WL Delft Hydraulics (now: Deltares) in Delft, the Netherlands, and aimed at estimating water and sediment transport in a delta area using a 3D model. To complete her studies Anne did an internship at the Royal Dutch Meteorological Institute (KNMI; De Bilt, the Netherlands). Her work at KNMI focused partly on improving the land surface scheme of a climate model and partly on the communication of climate scenarios to end users in the water sector. In 2006 Anne received her MSc diploma cum laude.

After finishing her studies, Anne joined the research and consulting agency FutureWater (Wageningen, the Netherlands). She worked on various projects such as the BSIK project 'Tailoring', related to modelling the effect of climate change on hydrology for water managers in the Netherlands, and the Partners for Water project 'WatManSup, Integrated Water Management Support Methodologies: a case study in Kenya and Turkey'. In this period, Anne built up experience with project management and knowledge transfer in national and international projects.

In 2007 Anne returned to Wageningen University, first as research assistant to write a 'Position Paper' on the potential and scope of collaboration between IBM and the climate and water communities, and later as a PhD student within the EU project WATCH. She published four papers in peer-reviewed journals (and one more under review), a number of technical reports, presented her work at various scientific conferences and had a share in the BSc and MSc education of the chair group Hydrology and Quantitative Water Management.

More information on the author and the publications related to this thesis can be found at www.researchgate.net/profile/Anne_Van_Loon/.

Cover design by Maarten van der Velde (www.ideogram.nl).

The image on the back cover depicts a *classical rainfall deficit drought*, in which a lack of precipitation causes decreasing soil moisture (leading to cracks) and lowering of the groundwater table. The image on the front cover represents a *cold-snow season drought*, in which below-normal temperatures in winter cause snow accumulation and soil frost resulting in decreasing soil moisture and lowering of the groundwater table. The graphics on the chapter title pages show decreasing groundwater levels.

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SENSE PhD Courses

- Environmental Research in Context
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- Techniques for Writing and Presenting Scientific Papers
- Teaching methodology and skills for PhD-students
- Theatervaardigheden in het onderwijs
- Brain-based teaching

Management and Didactic Skills Training

- Organisational support of EU projects WATCH, Xerochore and DROUGHT-R&SPI
- Lecturer, course development and practical supervisor for the courses *Hydrogeology* and *Hydrological Processes in Catchments*
- Lecturer for the course *Integratievak bodem, water en atmosfeer*
- Supervision of seven MSc theses, two BSc theses and one internship

Oral Presentations

- *Modelling drought propagation in WATCH test basins using HBV.* WATCH WB4 meeting, October 2009, Isegran, Norway
- *Winter droughts: Their development and impact in different parts of Europe.* WIMEK-SENSE symposium 'Soil, Water and Atmosphere Interactions', March 2010, Wageningen
- *Understanding hydrological winter drought in Europe.* Sixth World FRIEND conference, October 2010, Fez, Morocco
- *A new perspective on hydrological drought: a process-based classification into different drought types.* EGU2012-2429, EGU General Assembly, April 2012, Vienna, Austria
- *Ideas on using knowledge of climate and catchment control on drought propagation around the world to predict hydrological drought development in ungauged basins.* UNESCO-FRIEND meeting, October 2012, Payerbach, Austria
- *Catchment and climate control on global hydrological drought across the world.* Completion of the IAHS Decade on Prediction in Ungauged Basins, October 2012, Delft

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