1.1

Diagnosis of Drought-Generating Processes

Henny A.J. Van Lanen[1], Anne F. Van Loon[2], and Lena M. Tallaksen[3]

[1]*Wageningen University, Wageningen, The Netherlands*
[2]*University of Birmingham, Birmingham, United Kingdom*
[3]*University of Oslo, Oslo, Norway*

1.1.1 Introduction

It is well known that precipitation deficits initiate drought development. In cold climates (snow and glaciers), temperature anomalies (both high and low) too may contribute to drought generation. An in-depth understanding, that is, drought diagnosis, of how climate drives precipitation and temperature anomalies, and subsequently how these anomalies propagate into soil moisture, groundwater, and streamflow deficits (catchment control), is a prerequisite for reducing the immense socioeconomic and environment impacts of droughts (e.g. Stahl et al., 2016). Comprehensive overviews on drought are provided by Wilhite (2000); Tallaksen and Van Lanen (2004); Mishra and Singh (2010); Sheffield and Wood (2011); and Van Loon (2015). This chapter builds upon these overviews and complements them by synthesising knowledge from recently finished EU projects.

The chapter starts with an introduction about key hydroclimatological processes controlling drought generation – that is, how precipitation and temperature drive drought, and which stores and fluxes are affected. Different drought types are explained, and the main drought indices that are used in this chapter are briefly described (Section 1.1.2). Section 1.1.3 gives an overview of the main atmospheric and oceanic drivers for meteorological drought (deficit in precipitation, temperature anomalies), supplemented with a detailed description of the drivers of the 2015 summer drought in Europe, and a discussion of the influence of climate change on the meteorological drought in Europe. Section 1.1.4 continues with a more comprehensive description of the influence of meteorological drought on soil water, followed by the influence on groundwater and streamflow (Section 1.1.5). Both sections focus on key processes controlling the development of droughts in the hydrological system – that is, soil moisture, streamflow, and groundwater drought – followed by the role of human influence in

Drought: Science and Policy, First Edition.
Edited by Ana Iglesias, Dionysis Assimacopoulos, and Henny A.J. Van Lanen.

modifying the drought signal. Section 1.1.6 addresses how these different droughts propagate in the hydrological system – that is, how precipitation deficits and temperature anomalies affect snow accumulation and smelt, propagating into soil water, groundwater, and streamflow (drought propagation). A recently developed hydrological drought typology explains how climate and catchment controls determine drought propagation (Section 1.1.6). The focus in this chapter is on natural processes; however, at the end of each section, human interferences and their feedbacks are briefly touched upon (Sections 1.1.3–1.1.6). Finally, the concluding remarks are given in Section 1.1.7.

1.1.2 Background

The climate in the centre and north of Europe is influenced by the westerlies of the mid-latitudes during the whole year, bringing moisture from the Atlantic Ocean. The Mediterranean region lies in a transitional climate zone, influenced by the Subtropical High-Pressure Belt during summer and the mid-latitude westerlies during winter. Hence, two main climate regions can be distinguished: a temperate climate with a dry summer season in the Mediterranean; and a temperate climate and a cold climate without any dry season in the centre and north of Europe, respectively. Within these regions, climate is modified by numerous other permanent or temporally variable global, regional, or local factors, such as soil moisture, oceanic currents, and topography. Blocking situations disturb the common eastwards movement of the mid-latitude pressure systems, that is, the westerlies. During a blocking phase, an extended, persistent, high-pressure system develops in the eastern Atlantic Ocean at the mid-latitudes that does not move eastwards or moves only very slowly (Stahl and Hisdal, 2004). As a consequence, the moisture-bringing pressure systems divert moisture to Northern Africa and Northern Fennoscandia, causing an extended dry period (precipitation deficits) in mainland Europe (Section 1.1.3). During a dry and warm summer, feedbacks between the land surface and the atmosphere may amplify the drought signal. As the soil dries out, less energy is used for evapotranspiration (latent heat flux), and the partitioning of incident solar energy changes as more energy is used for heating the air (sensible heat flux). Heat waves thus frequently accompany major droughts, as reported for Europe by Ionita et al. (2017).

The lower-than-normal precipitation, usually combined with higher temperature and associated larger potential evapotranspiration (PET), leads to a decreased net precipitation (gross precipitation minus evaporated interception water), and hence infiltration into the topsoil of the vegetated surfaces (Figure 1.1.1). In places without vegetation, the lower gross precipitation directly results in lower infiltration. Evaporation of intercepted water, especially in forests, and overland flow on sloping land is also lower.

The reduced soil infiltration, together with the often-higher atmospheric water demand (increased PET), causes a larger depletion of the soil moisture storage than normal. Consequently, lowered soil evaporation and lessened soil water uptake by vegetation results in reduced evapotranspiration in many cases. Another important effect of the more depleted soil moisture store is the lower recharge to the underlying aquifer. In catchments where interflow takes place (e.g. soils with contrasting hydraulic conductivities, slopes), the aquifer also receives less water. This leads to lower groundwater levels, and hence reduced groundwater discharge to streams and lakes. Deeper, regional

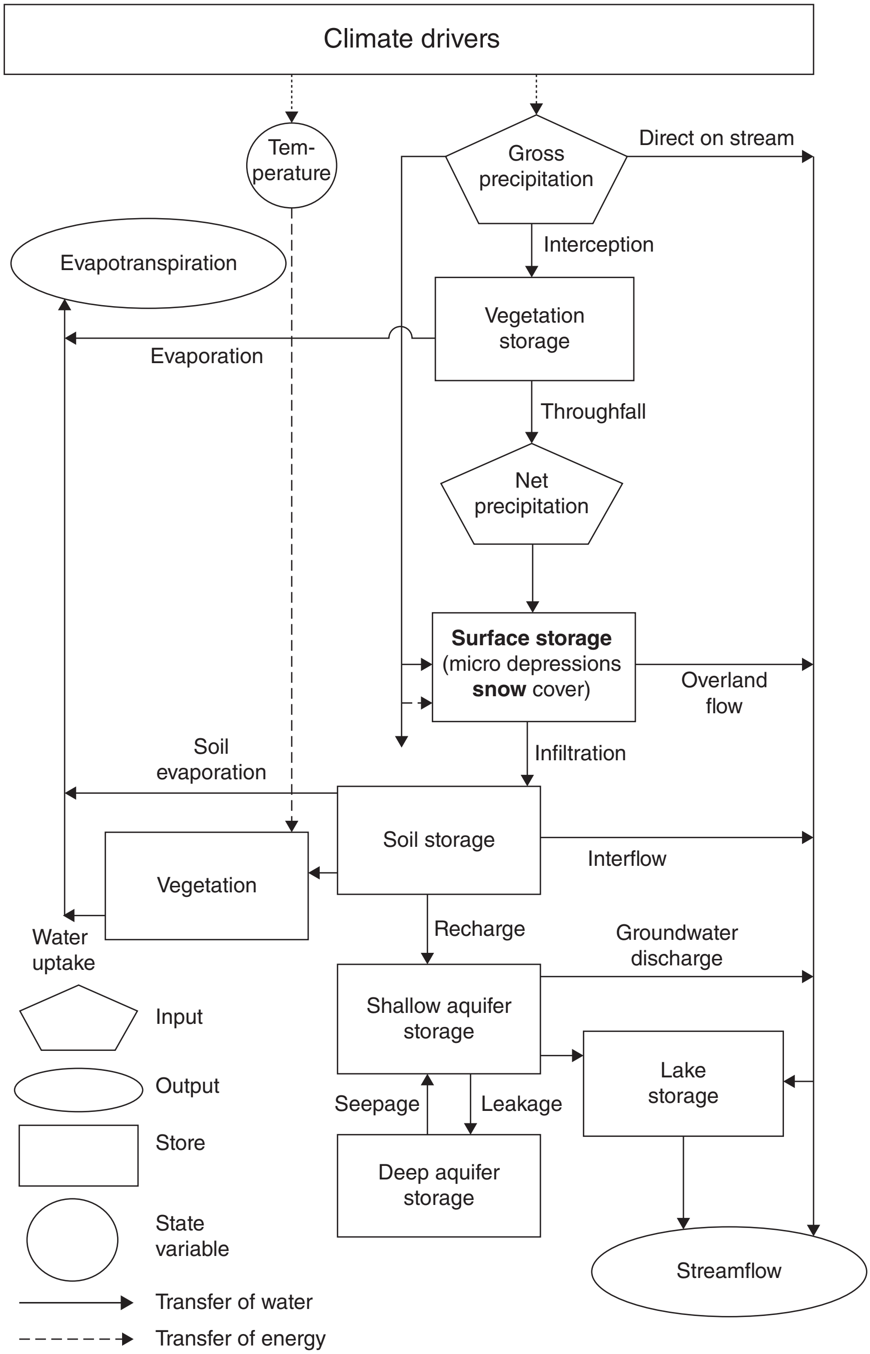

Figure 1.1.1 Water stores and fluxes affected by drought. *Source*: Derived from Van Lanen et al. (2004a).

aquifers also receive less water input (lower leakage), which may have long-lasting impacts in a wide area. Lakes can mitigate the effects of a downstream drought, because their natural role is to store surface water during wet periods and to release it during a dry period (upstream–downstream differences in streamflow). The changes in hydrological processes, which also have different time delays (i.e. response times), are more elaborated in the following sections (Sections 1.1.4 and 1.1.5).

In cold climates, irrespective of precipitation amounts, below-normal temperatures can result in earlier snow accumulation at the start of the cold season, which might lead to lower inflow to the streams. This also takes place when the cold season is longer than normal (delayed snow melt peak). In glaciated regions, cold temperature anomalies also lead to below-normal streamflow. Higher-than-normal winter temperatures may also cause below-normal summer flow, which is further elaborated in Section 1.1.6.

In many places, people try to intervene in natural processes, as described in the preceding text, to reduce drought impacts. For instance, deep soil tillage is applied to increase soil moisture supply capacity, or irrigation is applied to increase soil moisture content. Reservoirs are built to retain streamflow in a certain area to supply irrigation or water supply in general during a drought. However, this may enhance the drought downstream. On the other hand, the reservoirs can also be managed to maintain ecologically minimum flow during a drought. Van Loon et al. (2016a; 2016b) discuss the different implications that human interventions may have in a catchment, and introduce the following terms: (i) *climate-induced drought*, (ii) *human-modified drought*, and (iii) *human-induced drought*. Climate-induced drought is caused by natural climate variability, and is the focal area of this chapter; human-modified drought reflects a situation where humans have enhanced or alleviated the effects of climate-induced drought; and human-induced drought is caused entirely by people's measures (i.e. the drought should not have occurred under natural conditions). In Sections 1.1.3–1.1.6, human influences are further described.

Droughts happen in different domains of the hydrological cycle, and it is important to make distinctions between different drought types and their associated impacts (e.g. Van Lanen et al., 2016). Precipitation deficits and temperature anomalies cause meteorological drought, reduced soil infiltration results in soil water drought, and reduced recharge leads to groundwater drought. The combined effect of lower precipitation on the stream (and reduced overland flow, interflow, and groundwater discharge) is the origin of streamflow drought. Groundwater drought and streamflow drought are both referred to as 'hydrological drought' (Tallaksen and Van Lanen, 2004).

Various indices have been introduced to describe the different drought types, including their onset, duration, severity, and intensity (total deficit divided by duration). Two principal approaches are commonly used: (i) *standardised approaches* and (ii) *threshold approaches*. The *Standardised Precipitation Index* (SPI, McKee et al., 1993) and the *Standardised Precipitation–Evapotranspiration Index* (SPEI, Vicente-Serrano et al., 2010) are the most well-known standardised indices to describe meteorological drought. The SPI is a probabilistic measure that describes the number of standard deviations by which the dry event (accumulated precipitation over a given period) deviates from the median precipitation total over the same period. It can be calculated for different rainfall accumulation periods (e.g. 1–48 months, SPI-1 to SPI-48). The SPEI adds the PET and reflects the *climatic water deficit* (P-PET) over a given period. Similar standardised indices have been developed for groundwater (GRI, Bloomfield and Marchant, 2013) and runoff/streamflow (SRI, Shukla and Wood, 2008). Threshold approaches are better suited to quantify the water needed (volume) to manage and recover from a drought. It is based on defining a threshold, below which, for instance, the precipitation or streamflow is considered as a drought (Yevjevich, 1967). The threshold level method was first introduced for daily time series by Zelenhasic and Salvai (1987), and further explored by Tallaksen et al. (2009) for two catchments with varying storage properties. Both fixed and variable thresholds are used (Hisdal et al., 2004). Van Loon et al. (2010) and Van Loon (2015) describe the implementation of a daily smoothed monthly variable threshold that frequently is used. Heudorfer and Stahl (2017) comprehensively set out differences in outcome between fixed and variable thresholds when using the same time series (e.g. streamflow).

1.1.3 Climate drivers of drought

Regional drought-causing atmospheric situations are characterised by: (i) the anomalous timing of a seasonal phenomenon; (ii) the anomalous location of pressure centres and tracks of cyclones; and/or (iii) the anomalous persistence or persistent recurrence of dry weather patterns (Stahl and Hisdal, 2004). In the Mediterranean region, with its seasonal climate (Section 1.1.2), severe droughts can, for instance, be caused by longer-than-usual influence of the Subtropical High-Pressure Belt. Droughts can accordingly last for several weeks or even months. In the more humid mid-latitudes of western and northern Europe, ‘atmospheric blockings’ are the major atmospheric anomalies causing extended dry weather periods. Here, a few weeks or months with low rainfall may constitute a severe drought.

1.1.3.1 Atmospheric and oceanic drivers

Quantification of large-scale climate drivers of drought is important to understand and better manage spatially extensive and often prolonged natural hazards such as droughts, particularly their triggering mechanisms and persistence (e.g. Fleig et al., 2010, 2011). Primary drought-controlling mechanisms at the continental scale for Europe were explored by Kingston et al. (2015). Drought events were identified using SPI-6 and SPEI-6 (Section 1.1.2), both calculated using the gridded Water and Global Change forcing dataset (WATCH-WFD) for 1958–2001. Based on correlations between monthly time series of the percentage area in drought and 500 hPa geopotential height, a weakening of the prevailing westerly circulation was found to be associated with drought onset. Such conditions can be linked to variations in the East Atlantic/Western Russia (EA/WR) and North Atlantic Oscillation (NAO) atmospheric circulation patterns (Figure 1.1.2). Kingston et al. (2015) also performed an event-based composite analysis of the most widespread European droughts. It revealed that a higher number of droughts were identified by SPEI-6 than by SPI-6, with SPEI-6 drought events showing a greater variety of locations and start dates. They further concluded that differences between the atmospheric drivers of SPI-6 (associated with the NAO) and SPEI-6 events (associated with the EA/WR) reflected the sensitivity of these indices to the underlying drought type (precipitation versus climatic water balance), as well as sensitivity to the associated differences in their timing and location (northern Europe in winter vs. Europe-wide and year-round).

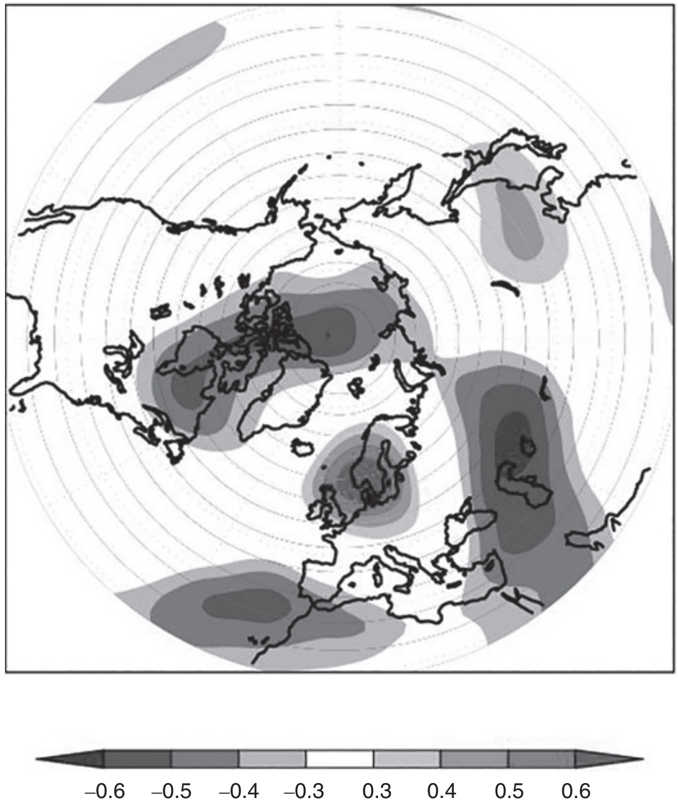

Figure 1.1.2 Correlation of May SPI-6 with December–May mean 500 hPa geopotential height. *Source*: From Kingston et al. (2015). (*See colour plate section for the colour representation of this figure.*)

Persistent dry conditions are associated with anticyclonic circulation, as reported in the preceding text for Europe, but oceanic factors, such as sea surface temperatures (SSTs), too can play a role through interaction with large-scale climatic or oceanic modes of variability, such as the North Atlantic oscillation (NAO) (Ionita et al., 2017; Kingston et al., 2013; 2015; Schubert et al., 2014).

The work of Kingston et al. (2013) illustrates the impact of antecedent sea surface temperatures and atmospheric circulation patterns on summer drought in Great Britain. However, the atmospheric bridge linking North Atlantic SST to drought development was identified as being too complex to be described solely by indices of the NAO. In Ionita et al. (2017), the key drivers of the 2015 drought event in Europe were analysed, with special emphasis on the role played by the SST and large-scale (atmospheric) circulation modes of variability, as described in the following section.

1.1.3.2 Summer drought of 2015

The summer drought of 2015 affected a large portion of continental Europe and was one of the most severe droughts since the summer of 2003, with record high temperatures in many parts of central and eastern Europe (Ionita et al., 2017). Over the summer period, there were four heat wave episodes, all associated with persistent blocking events. Upper-level atmospheric circulation was characterised by positive 500 hPa geopotential height anomalies bordered by a large negative anomaly to the north and west (i.e. over the central North Atlantic Ocean extending to northern Fennoscandia), and another centre of positive geopotential height anomalies over Greenland and northern Canada. At the same time, summer SSTs were characterised by large negative anomalies in the central North Atlantic Ocean and large positive anomalies in the Mediterranean basin (Figure 1.1.3). Ionita et al. (2017) conclude that the lagged relationship between the Mediterranean SST and summer drought conditions, especially over the eastern part of Europe, as identified in their study, holds potential for the prediction of drought conditions over Europe on seasonal to decadal timescales.

1.1.3.3 Influence of humans (Climate change)

Global warming has been most pronounced after 1970 (Hartmann et al., 2013), and Europe has warmed faster than the global mean land trend (Christensen et al., 2013). The most pronounced warming is seen in summer, notably in August (Nilsen et al., 2016). To understand the causes of regional temperature changes, it is common to separate the factors inducing the changes into *changes in atmospheric synoptic circulation* vs. *other factors*, including the so-called 'within-type changes'. Nilsen et al. (2016) revealed that changes in synoptic circulation could not account for all the observed (WATCH Forcing Dataset ERA-Interim, or 'WFDEI') warming in Europe over the period 1981–2010. Significant warming in specific months and regions, such as the large-scale warming in April, June, July, August, and November, must also be caused by other factors. Within-type changes may be caused by positive feedbacks between the land surface and the atmosphere, by forcing from greenhouse gases, or from other potential climate factors. Warming in regions with a seasonal snow cover, such as Scandinavia (Rizzi et al., 2017), is likely influenced by snow albedo feedbacks related to changes in snow cover, particular in spring when incoming radiation is high. Warming in water-limited regions, such as southern Europe during the summer, has been documented to be influenced by soil moisture–temperature feedbacks (Section 1.1.4). A general increase in temperature, and thus potential evapotranspiration, likely will enhance drought, regardless of changes in precipitation.

The frequency of meteorological droughts in Europe has increased since 1950 in parts of southern and central Europe, whereas droughts have become less frequent in northern Europe and parts of eastern Europe (EEA, 2017; Stagge et al., 2017), consistent with climate change projections (e.g. Stagge et al., 2015). Trends in drought severity (based on a combination of indices, including SPI and SPEI) also show significant increases in the Mediterranean region and parts of central and southeast Europe, and decreases in northern and parts of eastern Europe (EEA, 2017; Spinoni, et al., 2015; 2016). Stagge et al. (2017) document an increasing deviation in European drought frequency using SPI and SPEI, derived from instrumental records (1958–2014). Notably, they conclude that increases in temperature and reference evapotranspiration have enhanced droughts in southern Europe while counteracting increased precipitation in northern Europe.

Based on an ensemble from the EURO-CORDEX community project, Stagge et al. (2015) project that the frequency and duration of extreme meteorological droughts

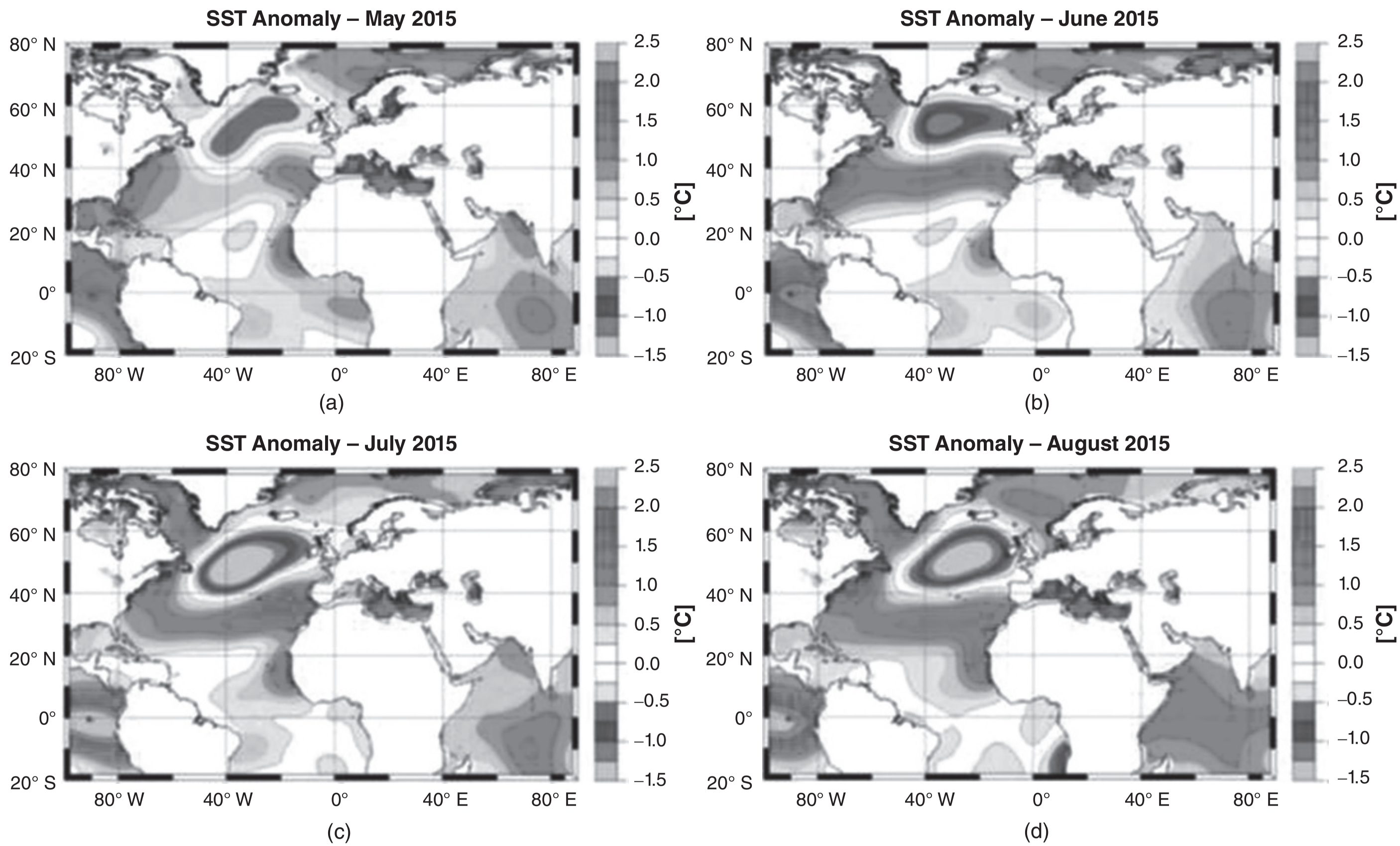

Figure 1.1.3 Monthly SST anomalies: (a) May 2015; (b) June 2015; (c) July 2015; and (d) August 2015; computed relative to the 1971–2000 period. *Source*: From Ionita et al. (2017). (*See colour plate section for the colour representation of this figure.*)

(SPI-6 < −2) will significantly increase in the future with respect to the baseline period (1971–2000). These projections show the largest increases in frequency for extreme droughts in parts of the Iberian Peninsula, southern Italy, and the eastern Mediterranean region, especially towards the end of the century (Stagge et al., 2015). Drought projections that also consider PET (e.g. SPEI) showed substantially more severe increases in the areas affected by drought than those based on precipitation alone (e.g. SPI) (EEA, 2017).

1.1.4 Soil moisture drought processes

The influence of a lower-than-normal infiltration into the soil surface is described in the following text, including the effects on evapotranspiration and recharge to the underlying aquifer. An example illustrates the change in processes, that is, the development of a soil water drought. The section concludes with how human interferences affect soil water drought. For a comprehensive description of drought-relevant flow-generating processes in the unsaturated zone, readers also are referred to Van Lanen et al. (2004a) and Sheffield and Wood (2011).

1.1.4.1 Processes

Lower-soil infiltration (Figure 1.1.1) influences *actual evapotranspiration* (ET_a) through a reduction in soil moisture. *Soil water storage* (SM) depletes quicker than normal, particularly because PET usually is higher. Several cases can be distinguished:

(i) Soils with high *soil moisture supply capacity* (SMSC) in humid climates
(ii) Soils with low SMSC in humid climates
(iii) Soils with high SMSC in dry climates
(iv) Soils with low SMSC in dry climates

In case (i), soils are more depleted, because of the high SMSC. SM does not reach the *critical soil moisture* (SM_c) level implying that the ET_a equals PET. Water losses to the atmosphere are higher during a drought than normal, at least in the first phase of the drought, because of the increased PET. Teuling et al. (2013) illustrate, for four catchments in west and central Europe, that ET_a increases, contributing to the aggravation of the drought. In case (ii), in the early phase of the drought, evapotranspiration might be higher – see case (i) – but this is followed by a situation where soil moisture becomes lower than the critical level ($SM < SM_c$), which leads to an earlier reduction of the PET ($ET_a < PET$). This causes vegetation stress, resulting in biomass reduction (e.g. reduced crop yield). Only rarely does the soil moisture level *not* reach the wilting point for plants (SM_w) – that is, soil conditions that the plants fully dry out. Case (iii) is probably similar to case (ii): the higher PET first results in higher losses to the atmosphere, which are then followed by a situation where $ET_a < PET$. In case (iv), $ET_a < PET$ is the dominant situation that eventually leads to an earlier full depletion of soil moisture ($SM = SM_w$; SMSC fully used).

The above-mentioned anomalies in precipitation and evapotranspiration affect soil moisture change (SM’). In case (i), the lower precipitation and higher PET have to be fully compensated by SM'. This results in a soil water drought. Later, in a wet period, when the rainfall is higher than normal, SM' has to be fully balanced by a reduction in recharge to the underlying aquifer, RCH' = SM'. In cases (ii) and (iii), the earlier reduction of the PET results in a situation where SM' is smaller than the sum of the anomalies in precipitation and PET. The soil water drought that develops is relatively smaller than in case (i). In case (iv), there is an impact on the ET_a that is lower due to the full depletion of SMSC earlier. A soil water drought develops, but finishes at the time of full depletion. SM at the start of a following wet period does not deviate from normal conditions (SM' = 0). This implies that, in this case, the recharge is not affected. In other words, a meteorological drought leads to a temporary soil water drought, but it does not result in a hydrological drought.

Soil water drought is driven by precipitation and PET anomalies, but it is constrained by the SMSC (the soil moisture determines to what level the anomalies in precipitation and PET can be compensated). Furthermore soil water drought is conditioned by the climate. In humid climates, soils are more likely to be fully replenished than in dry climates. In cold climates , temperature anomalies play a role too. Earlier snow accumulation or later snow melt than normal results into below-normal soil infiltration. This may imply the development of a soil water drought or the continuation of an ongoing drought. Likewise, seasonality, as a climate characteristic, contributes to how a soil water drought develops and recovers, and how it affects drought characteristics such as drought duration (DD) and deficit of soil moisture drought (DSMD) (Van Loon et al., 2014). For drought management, it is relevant to know whether a deficit quickly increases or not during a drought event. Long-lasting droughts can have either a small or large deficit. In a modelling experiment, Van Loon et al. (2014) investigated the relation between DD and DSMD for over 1000 grid cells distributed across 27 Köppen–Geiger climate types for the period 1958–2001. They used the variable threshold approach. For each grid cell, the correlation (R^2) between DD and DSMD of the soil moisture drought events was calculated (Figure 1.1.4).

Climate types with significant precipitation in all seasons (denoted by 'f' in the Köppen–Geiger climate type acronyms) are characterised by high correlations – for example, temperate and cold climates without a dry season and warm summer, or tropical rainforests (e.g. Af, Cfb, Dfb) that cover large parts of Europe and Southeast United States. Soil moisture droughts can occur in all seasons as a response to precipitation anomalies, and drought duration and deficit have a strong linear relation ($R^2 >$ 0.88). Climate types with a dry summer or winter ('s' or 'w' in the climate type acronyms) have strong seasonality. These are characterised by a divergent density field, and the associated lower R^2 (<0.68). A typical example is the tropical savanna climate (Aw) – for example, in South Brazil, Southeast Asia, North Australia, and large areas around the equator in Africa. Drought events occur both in the wet and dry season. In the wet season, the variable threshold is rather big, and soil moisture does not reach wilting point ($SM > SM_w$), which allows large deficits (SM') to develop. On the contrary, in the dry period, the SM is already close to wilting point under normal conditions (variable threshold is small). During a drought, SM reaches wilting points and the change in soil moisture is minor, which means that only small deficits develop. Hence, in climate types such as Aw, droughts of the same duration can either have a large deficit (wet season drought) or a small deficit (dry season drought) leading to low R^2.

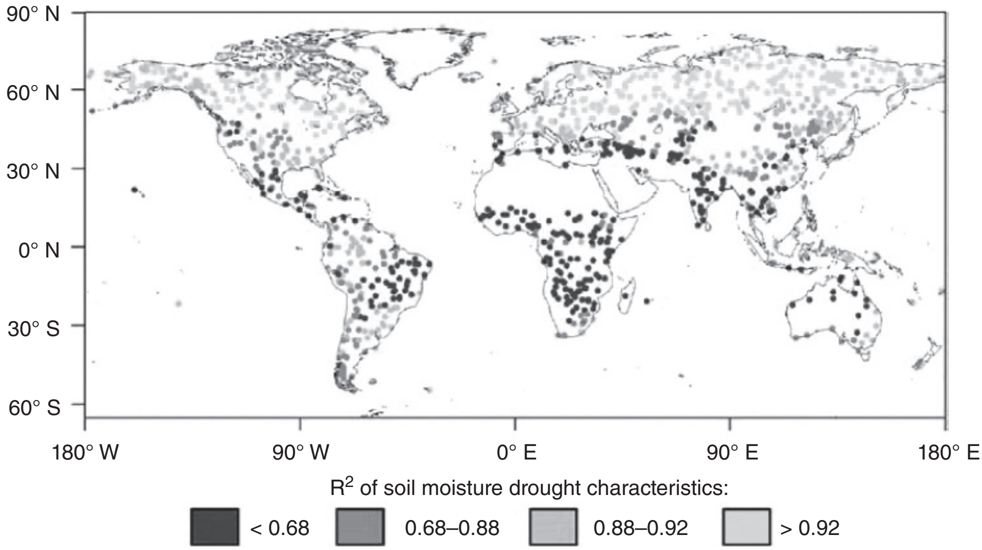

Figure 1.1.4 Spatial distribution of the correlation between DD and DSMD events for the period 1958–2001. *Source*: From Van Loon et al. (2014). (*See colour plate section for the colour representation of this figure.*)

1.1.4.2 Human influences

There are several studies that have investigated the impact of global warming on soil water moisture – in other words, whether soil water drought has changed over time. Long time series of soil moisture are required, but observed series with a continental or global coverage are lacking. Time series derived from satellite products are still rather short (De Jeu and Dorigo, 2016). Hence, large-scale changes of soil moisture have mainly been investigated using modelled and reanalysis data (e.g. Dai, 2012; Sheffield et al., 2012). These studies, based on the Palmer drought severity index (PDSI, Palmer, 1965), report apparently conflicting results on how drought is changing under climate change. Trenberth et al. (2014) explain that these disparities are caused by: (i) the different ways of calculating PET – that is, the more physically based Penman–Monteith concept (PDSI_PM) versus the only-temperature-driven Thornthwaite approach (PDSI_Th) (Van der Schrier et al., 2011); (ii) the weather data from global datasets (PDSI_PM uses more weather data than PSDI-Th, but some of these data are less reliable); and (iii) the differences in the global precipitation datasets. They conclude that the use of PDSI to assess the impact of global warming on soil water drought should be treated with care. Orlowsky and Seneviratne (2013) used the threshold-based soil moisture anomaly (SMA), which is a more comprehensive approach than PDSI, to explore the impact of global warming on soil water drought. The outcome of about 30 Climate Model Intercomparison Project Phase 5 (CMIP5) Global Climate Models (GCMs) for the period 1979–2009 was used, which was tested against three global monthly precipitation datasets. They found no significant changes in SMA in 12 major regions across the world for this historic period.

Irrigation is an important and direct human activity that considerably influences soil water drought. Over 70% of human need for water is intended for irrigation purposes, aimed at raising biomass production. The abstracted volume increased from ~500 km^3/year to ~4000 km^3/year over the last 100 years (e.g. Oki and Kanae, 2006). In the Northern Hemisphere, extensive irrigated areas are present in several countries between 20° and 50° latitudes, and similar regions are found in the Southern Hemisphere in South America, South Africa, and Australia (FAO, 2015). In these irrigated areas, soil water drought is substantially reduced or even eliminated.

1.1.5 Hydrological drought processes (Groundwater and streamflow)

The influence of a lower-than-normal recharge from the soil (Section 1.1.4) affects groundwater storage and, consequently, groundwater flow to the streams. Some examples are presented in the following text to illustrate the change in processes – that is, the development of a hydrological drought in response to a deficit in recharge. How human interferences affect hydrological drought is described towards the end. For a comprehensive description of drought-relevant flow-generating processes in water-saturated stores (such as groundwater, lakes, and wetlands), readers are also referred to Van Lanen et al. (2004a).

1.1.5.1 Groundwater

Processes A below-normal recharge (Section 1.1.4.1) leads to a more rapid decline of the groundwater heads, which causes a lower-than-normal groundwater storage in the shallow aquifer (Figure 1.1.1). Groundwater table response to a meteorological drought may occur months later (termed 'delay'; see Section 1.1.6) if the water table is deep – that is, several tens of meters or more below the soil surface (Van Lanen, 2004a).

In case a deep aquifer occurs, the faster-than-normal water table decline of the shallow aquifer is counteracted by either a lower leakage or higher seepage, which implies that the groundwater drought develops slower than in an area with only a shallow aquifer. The change in leakage or seepage, however, also affects the groundwater storage in the deep aquifer, where also a groundwater drought is induced. In multiple-aquifer systems, groundwater drought develops slowly, and recovery can last long. The effect of responsiveness of groundwater systems on drought has been explored by Van Lanen et al. (2013) in a modelling experiment. Responsiveness depends on aquifer properties, such as the thickness of the aquifer, hydraulic conductivity, and storativity. The distance between the streams to which the groundwater flows also plays a role; the larger the distance, the lower the responsiveness. In a modelling experiment, 1495 grid cells across the world that were proportionally distributed over the major climates of the globe (Köppen–Geiger) were selected. Hydrological characteristics for each grid cell were derived from the time series of simulated groundwater discharge obtained with a hydrological model that was forced with 44 years of observed weather data. Summary statistics of the number of droughts are given in Figure 1.1.5. Fast-responding groundwater systems have a substantially higher number of droughts as

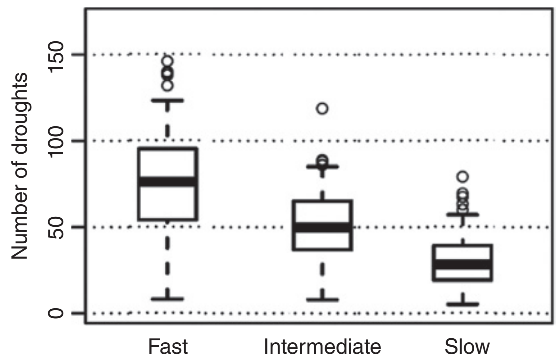

Figure 1.1.5 Summary statistics (box: 25, 50 and 75 deciles; whiskers: 5 and 95 deciles) of the number of hydrological droughts of all drought events (1495 grid cells) for three different groundwater systems. The circles represent extreme events beyond 95 decile. *Source*: Derived from Van Lanen et al. (2013).

compared to the slow-responding ones. Of the major climates (polar climates excluded), (semi-)arid climates have the lowest number of droughts. However, the droughts last longer there.

Bloomfield and Marchant (2013) also point at the importance of hydrogeology for groundwater drought development. They studied the long time series of a groundwater drought index from 14 different observation wells across the United Kingdom. They concluded that, in granular aquifers, drought duration is mainly determined by the effects of intrinsic aquifer properties on saturated groundwater flow, which can lead to long droughts (drought duration and frequency are negatively correlated in general). However, in fractured aquifers, drought duration is usually shorter and more connected to the temporal distribution of the recharge. In general, fractured aquifers respond faster than granular aquifers, implying that these are more affected by the recharge pattern. The role of *hydrogeology*, that is, how groundwater drought affects streamflow drought development, is explained in the next section (Section 1.1.5.2).

Human influences Groundwater abstraction is one of the best examples of human influence on the groundwater system. Abstraction usually leads to the enhancement of natural groundwater drought; exceptions occur under irrigated fields with leakage losses. The effect of abstraction on groundwater drought also depends on whether it is a permanent extraction (e.g. public water supply) or non-permanent (supplementary irrigation), and clearly on the distance to the well field. Van Lanen et al. (2004b) elaborated on the effects of both permanent and non-permanent abstractions on groundwater drought in a Dutch lowland river basin. The total abstraction of the non-permanent abstraction is half of the abstraction of the permanent well, and occurs only in the wet season. It is obvious that the permanent abstraction leads to more severe drought (expressed as the cumulative deviation between the groundwater table and the threshold). However, there are non-linear effects making the severity of the permanent abstraction more than twice as extreme as the non-permanent one. The main reason for this is that, during a drought, even under natural conditions, a large part of the drainage system (e.g. ditches, small streams) in a lowland area does not carry water. Under these conditions, abstracted groundwater cannot come from reduced groundwater flow to the surface water, but it comes from a decrease of groundwater storage, which means larger drawdowns of the water table than in normal or wet periods. The influence of this partly dry drainage network is larger for permanent abstractions than for non-permanent ones.

Van Loon and Van Lanen (2013) investigated the impact of groundwater abstraction for irrigation on groundwater drought in the Upper Guadiana river basin in Spain. They compared the observed groundwater levels, in the period during which the hydrological system is affected by the abstractions, with naturalised groundwater levels (Figure 1.1.6). The naturalised groundwater levels were obtained by running a hydrological model without groundwater abstractions. The model was calibrated with data from the period before irrigation expanded. Further, both time series of groundwater levels were compared against the variable threshold (Section 1.1.2) to obtain anomalies. The anomalies were used to distinguish between climate-induced drought versus human-modified and human-induced droughts (Section 1.1.2). The variable threshold curve (Figure 1.1.6) is identical for every year, and has been derived from monthly cumulative frequency distributions of the groundwater level before 1980 (almost no disturbance).

Climate-induced droughts occur when the naturalised and observed groundwater levels more or less coincide, and are below the threshold (e.g. 1991–1993, Figure 1.1.6), whereas human-modified droughts happen when both the observed and naturalised groundwater levels are below the threshold (e.g. 1992–1995). Human-induced droughts are rarer, and take place when the naturalised groundwater level is above the threshold and the observed level below (e.g. 1986 and 1996). Until 1983, the observed and naturalised groundwater levels more or less coincided, meaning that there was almost no human disturbance, and only minor climate-induced drought (1981–1982). However, this changed from 1984 onwards, and particularly since 1986. There is a clear increase of the human impact (groundwater abstraction) leading to the enhancement of groundwater drought and the development of human-modified drought. Van Loon and Van Lanen (2013) report that the number of groundwater

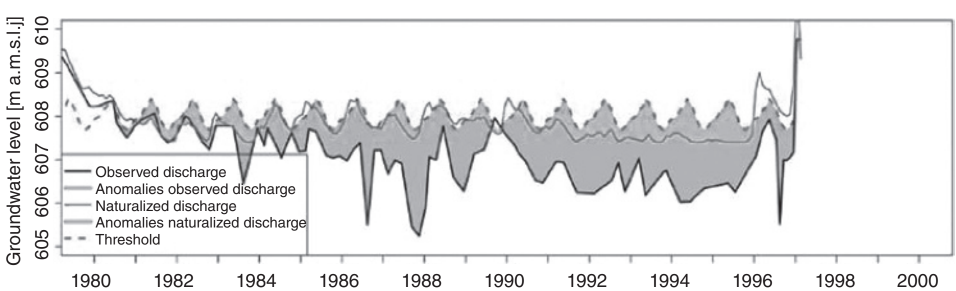

Figure 1.1.6 Anomalies in groundwater level in Upper Guadiana during the period 1980–1997 (disturbed period), using a variable 80% monthly threshold, and derived from the observed and naturalised groundwater levels. Blue areas can be attributed to climate, and grey areas to humans. *Source*: Derived from Van Loon and Van Lanen (2013). (*See colour plate section for the colour representation of this figure.*)

droughts has decreased by a factor of four, and that the duration of groundwater droughts has increased by a factor of six, owing to the pooling of droughts associated with groundwater abstraction.

1.1.5.2 Streamflow

Processes A below-normal groundwater discharge due to groundwater drought (see the subsection titled 'Processes' under Section 1.1.5.1) leads to lower stream levels and associated streamflow than normal (hydrological drought). In addition to groundwater discharge, overland flow and interflow (Figure 1.1.1), and other stores (such as lakes and wetlands), feed a stream. Substantial interflow is rather rare, but overland flow occurs in many catchments after intensive rainfall or snowmelt on frozen soil, particularly in areas with slopes and soils with a low infiltration capacity. During a meteorological drought, overland flow is lower than normal or absent, meaning that less rainwater reaches the stream quickly. During a drought, a stream is fully dependent on the groundwater discharge, and drainage from lakes and wetlands, if occurring in a region, which gradually becomes lower than normal. Overland flow can play a role in a temporary recovery from a drought; long-lasting droughts are interrupted by flow peaks. Streamflow drought does not recover unless the groundwater discharge (and lake or wetland drainage) becomes normal or above normal.

Response of streams to meteorological drought has much in common with groundwater response (see the subsection titled 'Processes' under Section 1.1.5.1), as long as overland flow does not play a dominant role. The number of streamflow droughts decreases, going from flashy to slowly responding catchments. The duration shows an opposite signal (Van Lanen et al., 2004a; Tallaksen et al., 2009).

The link between recharge deficit and streamflow response was investigated by Stoelzle et al. (2014) in several almost-natural catchments in Southwest Germany with different hydrogeological frameworks (hydraulic conductivities, storage capacities). Similar to Bloomfield and Marchant (2013), they found that hydrological drought in karstic and fractured aquifers rather quickly responds to drought in recharge. Specific recharge events could be traced back in the hydrological drought. Hydrological drought in granular aquifers responds more slowly to recharge deficits. Patterns in hydrological drought are more dominated by subsurface characteristics than by specific events in recharge.

Streamflow drought in Austria was studied using data from 44 catchments covering a wide spectrum of geo-climatic settings (Van Loon and Laaha, 2015). Drought duration is primarily driven by storage (accumulation and release) in the catchment. Large accumulation and slow release mean slow responsiveness, expressed by a high *Base Flow Index* (BFI, part of annual volume in the river coming from stores, e.g. groundwater). Clearly, drought duration in streamflow is also affected by the length of dry spells (in rainfall) or cold spells (in temperature). The streamflow deficit volume is more connected with seasonal storage in the snow pack and glaciers, which in Austria is linked to the elevation and the associated mean annual precipitation. Haslinger et al. (2014) add a dynamic component to the discussion on streamflow drought development in not-too-large catchments (<650 km^2). Under still rather wet conditions, droughts are largely determined by climate forcing, and can be predicted reasonably well by a simple meteorological index. However, when more dry conditions develop, the predictive skill decreases, and underground storage becomes more dominant. In the United Kingdom, research has shown that streamflow drought termination characteristics are also related to elevation and catchment annual precipitation, resulting in shorter drought terminations in wet upland catchments (Parry et al., 2016).

Lakes also impact downstream drought by storing upstream surface water in wet periods (rise of lake level) and releasing it (decline of level) during dry periods. The more the lake volume can change, the more upstream droughts are pooled, leading to less droughts downstream of the lake, but these can become longer during long dry periods (Van Lanen et al., 2004a). Wetlands have a similar role in alleviating and enhancing downstream streamflow drought.

Human influences In many parts of the world, humans have influenced streamflow drought (e.g. surface water abstraction, groundwater abstraction, water transfers, land use change; Van Lanen et al., 2004b). Similar to the influence on groundwater levels (see the subsection titled 'Human Influences' under Section 1.1.5.1), Van Loon and Van Lanen (2013) investigated the impact of groundwater abstraction on streamflow drought in the Upper Guadiana river basin. The observed flow, which was strongly impacted after 1980, was compared against the naturalised flow and the variable threshold (similar to Figure 1.1.6). The latter two were derived from the flow in the undisturbed period. The human-modified streamflow drought due to groundwater abstraction appeared to be four times larger than the natural, that is, the climate-induced drought.

In the Bilina catchment (Czech Republic), large-scale mining activities triggered water transfer from an adjacent catchment, implying that, in the disturbed period, the observed flow was usually higher than under natural conditions. Van Loon and Van Lanen (2015) show that streamflow drought was substantially alleviated. The frequency dropped by 85%, and the mean deficit volume by almost 90%. The streamflow drought in the catchment that supplied the water was not analysed, but due to the lower flow there likely the drought has been enhanced (depending on the ratio of water supply relative to the total discharge).

Surface water reservoirs impact discharge downstream of the dam. Usually, the high flows decrease substantially, whereas the low flows are higher. However, if water is abstracted directly from the reservoir downstream, low flows cannot be sustained during drought. López-Moreno et al. (2009) carried out a comprehensive drought analysis of the impact of the Alcántara reservoir, which was built on the Tagus River in the late 1960s. The reservoir is one of the largest in Europe, and it affected the river's flow in Spain and downstream in Portugal. Since the construction of the dam, the streamflow drought became more severe, particularly in the Portuguese part of the river, before tributaries enter the main river. Rangecroft et al. (2016) analysed the influence of the Santa Juana dam (a reservoir in northern Chile built for downstream water supply) on streamflow drought. They compared the upstream–downstream relation in the pre- and post-dam periods using a set of drought identification methods, including standardised and threshold approaches (Section 1.1.2). The dam caused a reduction in the average duration and deficit volume downstream (−42 and −86%, respectively). Hence, the dam alleviated downstream drought, although the study also showed that the capacity was insufficient to fully alleviate the severe multi-year droughts.

1.1.6 Drought propagation

1.1.6.1 Climate–hydrology links

Meteorological, soil moisture, and hydrological drought cannot be regarded separately. They are part of the same interconnected system in which one drought type influences the other by changing evapotranspiration, infiltration, and runoff processes (Figure 1.1.1). This transfer of drought through the terrestrial part of the hydrological cycle is called *propagation* (Peters et al., 2003; Van Loon, 2015).

Because a catchment can be regarded as a low-pass filter, the drought signal will undergo changes when it propagates from meteorological to hydrological drought (Figure 1.1.7). A soil moisture drought or hydrological drought starts later than a meteorological drought ('delay'), has a longer duration ('lengthening') and is of lower intensity ('attenuation'). Multiple meteorological droughts can also grow together into one hydrological drought ('pooling'). These four propagation processes are dependent on catchment characteristics and climate – that is, more storage in the catchment and more seasonality in climate lead to more delay, lengthening, attenuation, and pooling. The result is fewer but longer droughts in discharge than in precipitation (Peters et al., 2003; 2006; Tallaksen et al., 2009; Vidal et al., 2010; Van Loon and Van Lanen, 2012; Fendeková and Fendek, 2012; Van Loon and Laaha, 2015).

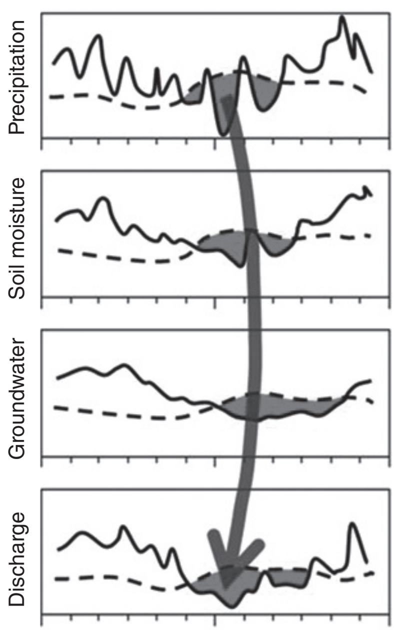

Figure 1.1.7 Propagation of drought through the terrestrial hydrological cycle, from meteorological drought to soil moisture drought to hydrological drought, showing pooling, delay, lengthening, and attenuation. (*See colour plate section for the colour representation of this figure.*)

1.1.6.2 Hydrological drought typology

Different climate seasonality and catchment storage result in clearly distinguishable hydrological drought types with different causing factors, development, and termination processes (Van Loon and Van Lanen, 2012; Van Loon et al., 2015). The most common hydrological drought type that occurs around the world is the *classical rainfall deficit drought*, in which a meteorological drought propagates through the hydrological system to form a groundwater drought or streamflow drought (Figure 1.1.7). In extremely seasonal climates, such a classical rainfall deficit drought developing in the high-flow season (summer in seasonally snow-covered catchments and winter in seasonally dry catchments) might continue into the low-flow season because all precipitation falls as snow or is lost to evapotranspiration. This results in *rain-to-snow-season drought* and *wet-to-dry-season drought*, respectively, which tend to have long durations (Van Loon and Van Lanen, 2012).

In snow- and ice-dominated catchments, anomalies in temperature play a role in drought development besides precipitation deficits (Van Loon and Van Lanen, 2012; Haslinger et al., 2014; Van Loon et al., 2015). In these studies, several drought types have been distinguished – for example, *glaciermelt drought*, *snowmelt drought*, *warm snow season drought*, and *cold snow season drought*. The variety in hydrological drought types related to snow and seasonality in cold climates makes it very important to consider snow and ice in drought monitoring and management (e.g. Staudinger et al., 2014; Haslinger et al., 2014; Harpold et al., 2017).

Finally, *composite droughts* are multi-year hydrological droughts that undergo extreme pooling, as a result of which different drought types in different seasons and years are combined into one long drought event (Van Loon and Van Lanen, 2012).

1.1.6.3 Human influences

Besides natural factors, each of the drought propagation characteristics and hydrological drought types is also influenced by anthropogenic factors. Human activities, intentionally or unintentionally, influence catchment storage and transfer processes, thereby enhancing or alleviating drought (Van Loon et al., 2016a). For example, reservoir building results in more pooling and attenuation of drought (e.g. Rangecroft et al., 2016), and groundwater abstraction leads to growing together of droughts into composite droughts (e.g. Van Loon and Van Lanen, 2013; see also the subsection titled 'Human Influences' under Section 1.1.5.2).

In most drought research till date, the climate system is seen as the driver of drought, and the societal system as the receiver (Figure 1.1.8, top). In reality, these systems are intertwined, and feedbacks occur with both the climate system and the societal system (Figure 1.1.8, bottom). For example, during the *Millennium Drought* in Australia, river regulation and abstraction influenced drought propagation and made the streamflow drought almost twice as severe as in natural circumstances (Van Dijk et al., 2013). In the recent multi-year drought in California, the presence and management of reservoirs decreased streamflow drought deficit by 50% in some areas, but irrigation water use increased streamflow drought deficit by 50–100% in others (He et al., 2017). In the future, the impact of human water use on changes in streamflow drought severity is expected to be as important as the effects of climate change in many regions around the world (Wanders and Wada, 2015). By including these feedbacks in drought analysis, we

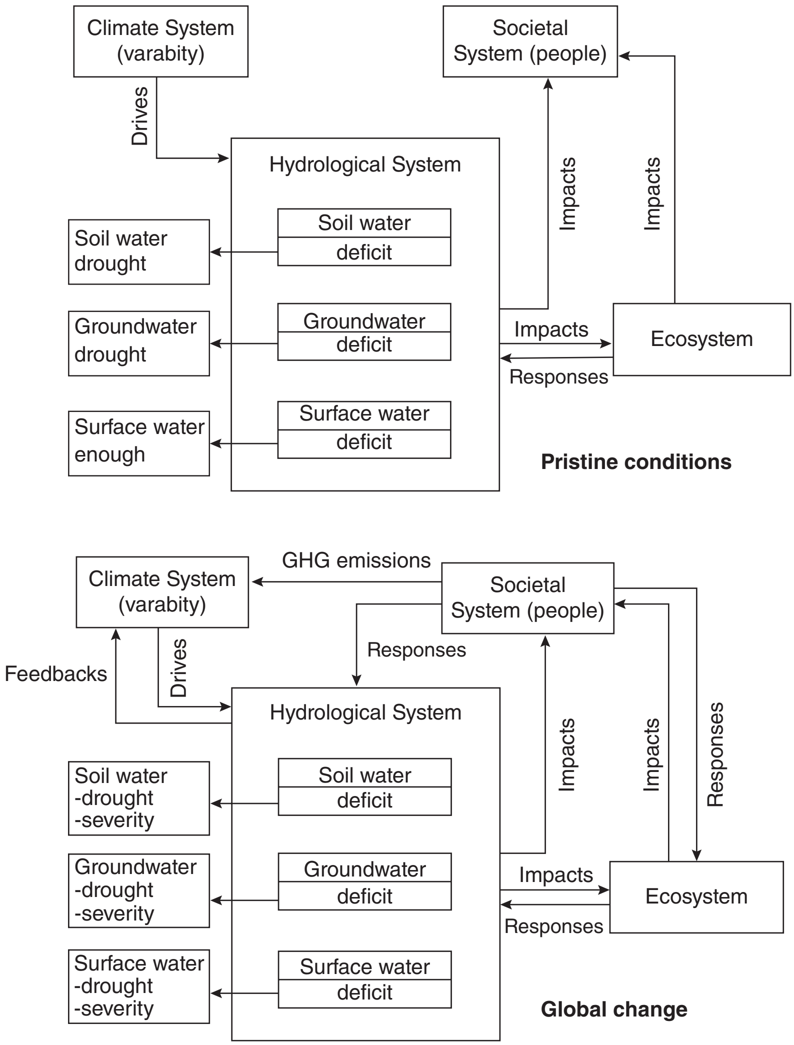

Figure 1.1.8 Relationships between the climate system, hydrological system, ecosystem, and societal system. Top: pristine conditions (representing a unidirectional drought propagation). Bottom: global change conditions (representing a multi-directional drought propagation characterised by feedbacks).

will not only increase our understanding of drought, but also be able to manage drought more effectively (Van Loon et al., 2016b).

1.1.7 Concluding remarks and outlook

1.1.7.1 Conclusions

In the centre and north of Europe, blocking situations that disturb the common eastwards movement of the mid-latitudes pressure systems, that is, the westerlies, are the main cause for meteorological drought development. In the Mediterranean region, severe droughts can be caused by the longer-than-usual influence of the Subtropical High-Pressure Belt. SSTs also play a role through their interaction with large-scale climatic or oceanic modes of variability.

Documented trends of drought over Europe show apparently conflicting outcomes (both drying and wetting trends are reported; see also Chapter 1.2 in this book, titled 'Recent Trends in Historical Drought'). Improved knowledge regarding the diagnosis of drought-generating processes can expose the underlying reasons – for example, regional geophysical settings, drought types, drought characteristics, and drought index.

Soil water drought has disappeared or been reduced in irrigated areas, but it is largely unclear what happened with the groundwater drought and surface drought in the irrigated area itself and beyond (downstream areas).

Hydrogeology (e.g. aquifer characteristics) plays an important role in the development of groundwater drought, and consequently streamflow drought is affected through drought propagation. Catchments with slowly responding groundwater systems, such as granular aquifers, have a lower number of droughts as compared to the fast-responding ones (e.g. fractured or consolidated karstic aquifers).

The development of the observation-modelling framework enabled us to better distinguish between climate-induced and human-modified/human-induced drought. So far, the framework has mainly been used for studying the effects of groundwater abstraction and reservoirs in a limited number of geo-climatic settings.

Climate seasonality and catchment storage trigger various drought generation processes, including termination, which lead to clearly distinguishable hydrological drought types. Similar to the flood typology, drought types can help us better understand the impacts associated with drought.

1.1.7.2 Outlook

Research on drought-generating processes should move from a focus on only natural processes to studies that integrate the physical system (climate–hydrology) and the societal system, including feedbacks. By intertwining the two systems in drought analysis, our understanding improves, from which drought management will benefit.

Improved knowledge on the diagnosis of drought-generating processes may enhance our abilities for monthly and seasonal drought forecasting. Teleconnections, including the lagged relationship between the SSTs and drought conditions, holds the potential for drought prediction on seasonal to decadal timescales over Europe. In addition to further improvements of weather/climate models to forecast drought, statistical models that use teleconnections need to be investigated further.

Improved knowledge on the processes underlying drought should be used to better explain the reasons for the apparently contradicting trends in drought, and may help to attribute to underlying causes for the trends (e.g. decadal climate variability, land use change, abstractions).

Although groundwater is an important water resource that is often intensively exploited during drought, the development of groundwater drought is still not extensively studied. More focus is required on the role of hydrogeology in drought generation.

Hydrological drought (groundwater, streamflow) as a response to water use for irrigation needs to be further investigated in the irrigated and downstream areas.

Application of the observation-modelling framework to distinguish between climate-induced and human-modified/human-induced drought needs more attention – for example, for better coverage of geo-climatic settings and human interferences that affect drought generation.

The established series of drought types (drought typology) should be further tested in other geo-climatic settings. Furthermore, robustness of the typology needs investigation – for example, do drought types change due to humans interferes, such as abstractions, reservoirs, global warming.

Acknowledgements

Funding was provided by the EU WATCH project (WATer and global CHange), contract 036 946, and the EU project DROUGHT–R&SPI (Fostering European Drought Research and Science–Policy Interfacing), contract 282 769. In the last phase, support came from the EU H2020 ANYWHERE (contract no. 700 099). The work was partly funded by the Dutch NWO Rubicon project 'Adding the human dimension to drought' (reference number: 2004/08338/ALW). The study is also a contribution to the UNESCO IHP-VII programme (Euro FRIEND-Water project) and the Panta Rhei Initiative of the International Association of Hydrological Sciences (IAHS).

References

Bloomfield, J.P. and Marchant, B.P. (2013). Analysis of groundwater drought building on the standardised precipitation index approach. *Hydrology and Earth System Sciences* 17: 4769–4787. doi: 10.5194/hess-17-4769-2013.

Christensen, J., Krishna Kumar, K., Aldrian, E. An S.I., Cavalcanti, I., De Castro, M., Dong, W., Goswami, P., Hall, A., Kanyanga, J., Kitoh, A., Kossin, J., Lau, N.C., Renwick, J., Stephenson, D., Xie, S.P., and Zhou, T. (2013). *Climate Phenomena and Their Relevance for Future Regional Climate Change*. Book section 14. Cambridge: Cambridge University Press, ISBN 978-1-107-66182-0: Cambridge, UK and New York, NY, 1217–1308. doi: 10.1017/CBO9781107415324.028.

Dai, A. (2012). Increasing drought under global warming in observations and models. *Nature Climate Change* 3: 52–58. doi: 10.1038/nclimate1633.

De Jeu, R. and Dorigo, W. (2016). On the importance of satellite observed soil moisture. *International Journal of Applied Earth Observation and Geoinformation* 45: 107–109. doi: 10.1016/j.jag.2015.10.007.

EEA (European Environmental Agency) (2017). Climate Change, Impacts and Vulnerability in Europe 2016. An Indicator-based Report. *EEA Report No. 1/2017, European Environmental Agency*, Copenhagen. doi: 10.2800/534806.

FAO (2015). AQUASTAT Main Database, Food and Agriculture Organization of the United Nations (FAO). Available from: http://www.fao.org/nr/water/aquastat/main/index.stm, website accessed on 22/03/2018.

Fendeková, M. and Fendek, M. (2012). Groundwater drought in the Nitra River Basin –identification and classification. *Journal of Hydrology and Hydromechanics* 60 (3): 185–193.

Fleig, A.K., Tallaksen, L.M., Hisdal, H., Stahl, K., and Hannah, D.M. (2010). Inter-comparison of weather and circulation type classifications for hydrological drought development. *Physics and Chemistry of the Earth* 35: 507–515. doi: 10.1016/j.pce.2009.11.005, 2010.

Fleig, A.K., Tallaksen, L.M., Hisdal, H., and Hannah, D.M. (2011). Regional hydrological droughts and associations with the objective Grosswetterlagen in north-western Europe. *Hydrological Processes* 25: 1163–1179. doi: 10.1002/hyp.7644, 2011.

Harpold, A.A., Dettinger, M., and Rajagopal, S. (2017). Defining snow drought and why it matters. *Eos* 98. doi: 10.1029/2017EO068775.

Hartmann, D.L., Klein Tank, A.M.G., Rusticucci, M., Alexander, L.V., Brönnimann, S., Charabi, Y., Dentener, F.J., Dlugokencky, E.J., Easterling, D.R., Kaplan, A., Soden, B.J., Thorne, P.W., Wild, M., and Zhai, P.M. (2013). Observations: atmosphere and surface. In: *Climate Change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change* (ed. T.F. Stocker, D. Qin, G.-K. Plattner, M. Tignor, S.K. Allen, J. Boschung, A. Nauels, Y. Xia, V. Bex, and P.M. Midgley). Cambridge, UK and New York, NY, USA: Cambridge University Press.

Haslinger, K., Koffler, D., Schöner, W., and Laaha, G. (2014). Exploring the link between meteorological drought and streamflow: effects of climate–catchment interaction. *Water Resources Research* 50: 2468–2487. doi: 10.1002/2013wr015051.

He, X., Wada, Y., Wanders, N., and Sheffield, J. (2017). Human water management intensifies hydrological drought in California. *Geophysical Research Letters* 44: 1777–1785. doi: 10.1002/2016GL071665.

Heudorfer, B. and Stahl, K. (2017). Comparison of different threshold level methods for drought propagation analysis in Germany. *Hydrology Research* 48 (5): 1311–1326. doi: 10.2166/nh.2016.258.

Hisdal, H., Tallaksen, L.M., Clausen, B., Peters, E., and Gustard, A. (2004). Drought characteristics. In: *Hydrological Drought: Processes and Estimation Methods for Streamflow and Groundwater; Developments in Water Science, Volume 48* (ed. L.M. Tallaksen and H. Van Lanen), 139–198. Amsterdam: Elsevier.

Ionita, M., Tallaksen, L.M., Kingston, D.G., Stagge, J.H., Laaha, G., Van Lanen, H.A.J., Chelcea, S.M., and Haslinger, K. (2017). The European 2015 drought from a climatological perspective. *Hydrology and Earth System Sciences* 21: 1397–1419. doi: 10.5194/hess-21-1397-2017.

Kingston, D.G., Fleig, A.K., Tallaksen, L.M., and Hannah, D.M. (2013). Ocean–atmosphere forcing of summer streamflow drought in Great Britain. *Journal of Hydrometeorology* 14: 331–344. doi: 10.1175/JHM-D-11-0100.1.

Kingston, D.G., Stagge, J.H., Tallaksen, L.M., and Hannah, D.M. (2015). European-scale drought: understanding connections between atmospheric circulation and meteorological drought indices. *Journal of Climate* 28: 505–516. doi: 10.1175/JCLI-D-14-00001.1.

López-Moreno, J.I., Vicente-Serrano, S.M., Beguería, S., García-Ruiz, J.M., Portela, M.M., and Almeida, A.B. (2009). Dam effects on droughts magnitude and duration in a transboundary basin: the Lower River Tagus, Spain and Portugal. *Water Resources Research* 45: W02405. doi: 10.1029/2008WR007198.

Mckee, T.B., Doesken, N.J., and Kleist, J. (1993). The relationship of drought frequency and duration to time scale. In: *Proceedings of 8th Conference on Applied Climatology*, Anaheim, California, 17–22 January 1993. Boston: American Meteorological Society, 179–184.

Mishra, K.K. and Singh, V.P. (2010). A review of drought concepts. *Journal of Hydrology* 391: 202–216.

Nilsen, I.B., Stagge, J.H., and Tallaksen, L.M. (2016). A probabilistic approach for attributing temperature changes to synoptic type frequency. *International Journal of Climatology* 37 (6): 2990–3002. doi: 10.1002/joc.4894.

Oki, T. and Kanae, S. (2006). Global hydrological cycles and world water resources. *Science* 313: 1068–1072. doi: 10.1126/science.1128845.

Orlowsky, B. and Seneviratne, S.I. (2013). Elusive drought: uncertainty in observed trends and short- and long-term CMIP5 projections. *Hydrology and Earth System Sciences* 17: 1765–1781. doi: 10.5194/hess-17-1765-2013.

Palmer, W.C. (1965). Meteorological drought, *Weather Bureau, Research Paper No. 45,* U.S. Department of Commerce, Washington, DC.

Parry, S., Wilby, R.L., Prudhomme, C., and Wood, P.J. (2016). A systematic assessment of drought termination in the United Kingdom. *Hydrology and Earth System Sciences* 20: 4265–4281. doi: 10.5194/hess-20-4265-2016.

Peters, E., Torfs, P.J.J.F., Van Lanen, H.A.J., and Bier, G. (2003). Propagation of drought through groundwater – a new approach using linear reservoir theory. *Hydrological Processes* 17(15): 3023–3040.

Peters, E., Bier, G., Van Lanen, H.A.J., and Torfs, P.J.J.F. (2006). Propagation and spatial distribution of drought in a groundwater catchment. *Journal of Hydrology* 321(1): 257–275.

Rangecroft, S., Van Loon, A.F., Maureira, H., Verbist, K., and Hannah, D.M. (2016). Multi-method assessment of reservoir effects on hydrological droughts in an arid region. *Earth System Dynamics Discussions.* doi: 10.5194/esd-2016-57.

Schubert, S.D., Wang, H., Koster, R.D., Suarez, M.J., and Groisman, P. (2014). Northern Eurasian heat waves and droughts. *Journal of Climate* 27: 3169–3207. doi:10.1175/JCLI-D-13-00360.1.

Rizzi, J., Nilsen, I.B., Stagge, J.H., Grini, K., and Tallaksen, L.M. (2017). Five decades of warming: impacts on snow cover in Norway. *Hydrological Research* (Submitted).

Sheffield, J. and Wood, E.F. (2011). *Drought: Past Problems and Future Scenarios*. London: Earthscan.

Sheffield, J., Wood, E.F., and Roderick, M.L. (2012). Little change in global drought over the past 60 years. *Nature*, 491: 435–438. doi: 10.1038/nature11575.

Shukla, S. and Wood, A.W. (2008). Use of a standardized runoff index for characterizing hydrologic drought. *Geophysical Research Letters* 35 (2): L02,405.

Spinoni, J., Naumann, G., Vogt, J., and Barbosa, P. (2015). European drought climatologies and trends based on a multi-indicator approach. *Global and Planetary Change* 127: 50–57.

Spinoni, J., Naumann, G., Vogt, J., and Barbosa, P. (2016). *Meteorological Droughts in Europe. Events and Impacts Past Trends and Future Projections*. Luxembourg, EUR 27748 EN: Publications Office of the European Union. doi: 10.2788/450449.

Stagge, J.H., Rizzi, J., Tallaksen, L.M., and Stahl, K. (2015). Future Meteorological Drought: Projections of Regional Climate Models for Europe. *DROUGHT R&SPI Technical Report No. 25*, University of Oslo, Norway.

Stagge, J.H., Kingston, D., Tallaksen, L.M., and Hannah, D. (2017). Observed drought indices show increasing divergence across Europe. *Proceedings of the National Academy of Sciences* (Submitted).

Stahl, K. and Hisdal, H. (2004). Hydroclimatology (chapter 2). In: *Hydrological Drought. Processes and Estimation Methods for Streamflow and Groundwater; Developments in Water Science, Volume 48* (ed. L.M. Tallaksen and H.A.J. Van Lanen), 19–51. Amsterdam: Elsevier Science B.V.

Stahl, K., Kohn, I., Blauhut, V., Urquijo, J., de Stefano, L., Acacio, V., Dias, S., Stagge, J.H., Tallaksen, L.M., Kampragou, E., Van Loon, A.F., Barker, L.J., Melsen, L.A., Bifulco, C., Musolino, D., de Carli, A., Massarutto, A., Assimacopoulos, D., and Van Lanen, H.A.J. (2016). Impacts of European drought events: insights from an international database of text-based reports. *Natural Hazards and Earth System Sciences* 16: 801–819. doi:10.5194/nhess-16-801-2016.

Staudinger, M., Stahl, K., and Seibert, J. (2014). A drought index accounting for snow. *Water Resources Research* 50 (10): 7861–7872. doi: 10.1002/2013WR015143.

Stoelzle, M., Stahl, K., Morhard, A., and Weiler, M. (2014). Streamflow sensitivity to drought scenarios in catchments with different geology. *Geophysical Research Letters* 41 (17): 6174–6183. doi: 10.1002/2014GL061344.

Tallaksen, L.M. and Van Lanen, H.A.J. (2004). *Hydrological Drought: Processes and Estimation Methods for Streamflow and Groundwater*. No. 48 in Development in Water Science, Amsterdam: Elsevier Science B.V.

Tallaksen, L.M., Hisdal, H., and Van Lanen, H.A.J. (2009). Space–time modelling of catchment scale drought characteristics. *Journal of Hydrology* 375 (3): 363–372.

Teuling, A.J., Van Loon, A.F., Seneviratne, S.I., Lehner, I., Aubinet, M., Heinesch, B., Bernhofer, C., Grünwald, T., Prasse, H., and Spank, U. (2013). Evapotranspiration amplifies European summer drought. *Geophysical Research Letters* 40: 2071–2075. doi:10.1002/grl.50495.

Trenberth, K.E., Dai, A., Van der Schrier, G., Jones, P.D., Barichivich, J., Briffa, K.R., and Sheffield, J. (2014). Global warming and changes in drought. *Nature Climate Change* 4: 17–22. doi: 10.1038/nclimate2067.

Van der Schrier, G., Jones, P.D., and Briffa, K.R. (2011). The sensitivity of the PDSI to the Thornthwaite and Penman–Monteith parameterizations for potential evapotranspiration. *Journal of Geophysical Research* 116: D03106.

Van Dijk, A.I.J.M., Beck, H.E., Crosbie, R.S., De Jeu, R.A.M., Liu, Y.Y., Podger, G.M., Timbal, B., and Viney, N.R. (2013). (2001–2009): Natural and human causes and implications for water resources, ecosystems, economy, and society. *Water Resources Research* 49 (2): 1040–1057. doi:10.1002/wrcr.20123.

Van Lanen, H.A.J., Fendeková, M., Kupczyk, E., Kasprzyk, A., and Pokojski, W. (2004a). Flow generating processes (chapter 3). In: *Hydrological Drought. Processes and Estimation Methods for Streamflow and Groundwater; Developments in Water Science, Volume 48* (ed. L.M. Tallaksen and H.A.J. Van Lanen), 53–96. Amsterdam: Elsevier Science B.V.

Van Lanen, H.A.J., Kašpárek, L., Novický, O., Querner, E.P., Fendeková, M., and Kupczyk, E. (2004b). Human influences (chapter 9). In: *Hydrological Drought. Processes and Estimation Methods for Streamflow and Groundwater; Developments in Water Science, Volume 48* (ed. L.M. Tallaksen and H.A.J. Van Lanen), 347–410. Amsterdam: Elsevier Science B.V.

Van Lanen, H.A.J., Wanders, N., Tallaksen, L.M., and Van Loon, A.F. (2013). Hydrological drought across the world: impact of climate and physical catchment structure. *Hydrology and Earth System Sciences* 17: 1715–1732. doi: 10.5194/hess-17-1715-2013.

Van Lanen, H.A.J., Laaha, G., Kingston, D.G., Gauster, T., Ionita, M., Vidal, J.-P., Vlnas, R., Tallaksen, L.M., Stahl, K., Hannaford, J., Delus, C., Fendekova, M., Mediero, L., Prudhomme, C., Rets, E., Romanowicz, R.J., Gailliez, S., Wong, W.K., Adler, M.-J., Blauhut, V., Caillouet, L., Chelcea, S., Frolova, N., Gudmundsson, L., Hanel, M., Haslinger, K., Kireeva, M., Osuch, M., Sauquet, E., Stagge, J.H. and Van Loon, A.F. (2016). Hydrology needed to manage droughts: the 2015 European case. *Hydrological Processes* 30: 3097–3104. doi: 10.1002/hyp.10838.

Van Loon, A.F. (2015). Hydrological drought explained. *WIREs Water*. doi: 10.1002/wat2.1085.

Van Loon, A.F. and Van Lanen, H.A.J. (2012). A process-based typology of hydrological drought. *Hydrology and Earth System Sciences* 16: 1915–1946. doi: 10.5194/hess-16-1915-2012.

Van Loon, A.F. and Van Lanen, H.A.J. (2013). Making the distinction between water scarcity and drought using an observation-modeling framework. *Water Resources Research* 49: 1483–1502. doi:10.1002/wrcr.20147.

Van Loon, A.F., and Laaha, G. (2015). Hydrological drought severity explained by climate and catchment characteristics. *Journal of Hydrology* 526: 3–14. doi:10.1016/j.jhydrol.2014.10.059.

Van Loon, A.F. and Van Lanen, H.A.J. (2015): Testing the observation-modelling framework to distinguish between hydrological drought and water scarcity in case studies around Europe. *European Water* 49: 65–75, www.ewra.net/ew/issue_49.htm.

Van Loon, A.F., Van Lanen, H.A.J., Hisdal, H., Tallaksen, L.M., Fendeková, M., Oosterwijk, J., Horvát, O., and Machlica, A. (2010). Understanding hydrological winter drought in Europe. In: *Global Change: Facing Risks and Threats to Water Resources* (ed. E. Servat, S. Demuth, A. Dezetter, T. Daniell, E. Ferrari, M. Ijjaali, R. Jabrane, H. Van Lanen, and Y. Huang), 189–197. Wallingford: IAHS Publ. No. 340.

Van Loon, A.F., Tijdeman, E., Wanders, N., Van Lanen, H.A.J., Teuling, A.J., and Uijlenhoet, R. (2014). How climate seasonality modifies drought duration and deficit. *Journal of Geophysical Research – Atmospheres* 119 (8): 4640–4656. doi: 10.1002/2013JD020383.

Van Loon, A.F., Ploum, S.W., Parajka, J., Fleig, A.K., Garnier, E., Laaha, G., and Van Lanen, H.A.J. (2015). Hydrological drought types in cold climates: quantitative analysis of causing factors and qualitative survey of impacts. *Hydrology and Earth System Sciences* 19: 1993–2016. doi: 10.5194/hess-19-1993-2015.

Van Loon, A.F., Gleeson, T., Clark, J., Van Dijk, A., Stahl, K., Hannaford, J., Di Baldassarre, G., Teuling, A., Tallaksen, L.M., Uijlenhoet, R., Hannah, D.M., Sheffield, J., Svoboda, M., Verbeiren, B., Wagener, T., Rangecroft, S., Wanders, N., and Van Lanen, H.A.J. (2016a). Drought in the Anthropocene. *Nature Geoscience* 9 (2): 89–91.

Van Loon, A. F., Gleeson, T., Clark, J., Van Dijk, A.I.J.M., Stahl, K., Hannaford, J., Di Baldassarre, G., Teuling, A.J., Tallaksen, L.M., Uijlenhoet, R., Hannah, D.M., Sheffield, J., Svoboda, M., Verbeiren, B., Wagener, T., Rangecroft, S., Wanders, N., and Van Lanen, H.A.J. (2016b). Drought in a human-modified world: reframing drought definitions, understanding and analysis approaches. *Hydrology and Earth System Sciences* 20 (9): 3631–3650. doi: 10.5194/hess-20-3631-2016.

Vicente-Serrano, S.M., Beguería, S., and López-Moreno, J.I. (2010). A multi-scalar drought index sensitive to global warming: the Standardized Precipitation Evapotranspiration Index – SPEI. *Journal of Climate* 23 (7): 1696–1718. doi: 10.1175/2009JCLI2909.1.

Vidal, J. P., Martin, E., Franchistéguy, L., Habets, F., Soubeyroux, J.M., Blanchard, M., and Baillon, M. (2010). Multilevel and multiscale drought reanalysis over France with the Safran-Isba-Modcou hydrometeorological suite. *Hydrology and Earth System Sciences* 14 (3): 459–478.

Wanders, N. and Wada, Y. (2015). Human and climate impacts on the 21st century hydrological drought. *Journal of Hydrology* 526: 208–220.

Wilhite, D.A. (Ed.) (2000). Droughts as a natural hazard: concepts and definitions. In: *Drought, A Global Assessment, Vol I and II, Routledge Hazards and Disasters Series*. London: Routledge.

Yevjevich, V. (1967). An objective approach to definition and investigations of continental hydrologic droughts. *Hydrology Papers 23*, Colorado State University, Fort Collins, USA.

Zelenhasic, E. and Salvai, A. (1987). A method of streamflow drought analysis. *Water Resources Research* 23 (1): 156–168.