Dear Editor,

We would like to thank you and the two referees for their comments to improve our paper. Based on these comments the following main changes were applied:

- The introduction was changed to constitute a global problem before highlighting the regional problem as suggested by Referee #1.

- In the section 4.1 hydrological model, we tried to explain our reasons for selecting the FLEX model as suggested by both Referees #1 and #2.

- We carefully revised the discussion of our MS and several aspects were included, for example in the section 5.2, discussing the drought analysis for two periods in addition to four decades, and adjusting the drought threshold to -1 and providing re-calculated estimates for the drought characteristics as suggested by Referee #2. Also we added a discussion part on the effects of the uncertainties on the reservoir routing scheme on downstream droughts as suggested by Referee #2.

- We also compared carefully our results with the findings of relevant literature as suggested by both Referee #1 and #2.

- We added the new figures requested by Referee #2 in the supplementary materials included Figure S1-S3 (fit of gamma and GEV distributions to calculate the drought indices), Figure S4 (Sensitivity analysis of the effects of uncertainties in the low flow reservoir routing scheme on downstream flow) and Figure S6 (Distribution of drought indices over study period).

Enclosed we added the revised MS with track chances as well as our detailed responses to the two reviewer comments. Again we would like to thank you and the two referees for handling, respectively reviewing, our MS and look forward for your assessment.

Reply to Referee #1

Response:

We would like to thank the referee for the time and effort she/he put into this review. The points she/he raises are highly relevant and addressing them will help to improve our manuscript. We hope that with these replies and the associated revision of our manuscript all issues raised can be clarified.

Comment:

The introduction section needs to be rearranged and improved. Specifically, the authors should start the introduction by stating how their research constitute a global problem before highlighting the regional problems identified in the Central Asian River Basin. I suggest that the authors start the introduction with line 66:81 (meaning it should be moved up and edited meaningfully) before discussion on the study basin. In doing so, however, the authors should ensure the literature cited are relevant in the body of work and should reflect in the discussion section.

Response:

We thank the Referee for having drawn our attention to this point and indeed we agree that it will help the logic and flow of the manuscript to start with a description of the general problem before going into regional detail. The text will be revised accordingly. Due to the scarcity of literature in this basin, we tried to cite the relevant studies in the study basin and other basins in similar environmental settings, subject to similar drought phenomena. In any case, we will extend the literature overview to provide some more context, in particular with respect to the global drought problematic.

Comment:

The methods used in this research has a good scientific foundation. Nonetheless, the reasons for selecting the hydrological model need to be highlighted in a few sentences given that other hydrological models can perform this same task. Why use this model and why is it important. The authors mentioned the scarcity of data as a limitation of this research, does this model help to alleviate the problem of data scarcity? Why do the authors choose this model over other hydrological models?

Response:

We agree that some more explanation on the choice of the model will help the reader to understand our decision. Briefly, the distributed implementation of this process-based model from the FLEX-model family was chosen as these models were previously successfully applied in climatically similar regions (e.g. Gao et al., 2014, 2017) but also many other settings worldwide (e.g. Fenicia et al., 2006; Kavetski et al., 2011; Nijzink et al., 2018; Bouaziz et al., 2018; Hulsman et al., 2020). In addition, the FLEX modelling concept is underlain by a philosophy of model customization and rigorous testing to ensure the implementation of suitable model formulations and the associated more reliable model outputs in different environments (e.g. Fenicia et al., 2011). Given the high

number of different hydrological models that have been developed over the past decades, it is indeed plausible to assume that alternative model formulations could have been used to perform the same task. Please also note that, following a multi-objective calibration strategy, i.e. simultaneously using $E_{NS,Q}$ and $E_{NS,\log(Q)}$ as calibration objectives to ensure good representation of both, high- and low flows, our model performances with respect to daily flow in all sub-basins (Table 4, Figure 4), exceed those of the studies of Hajihosseini et al. (2016) but also those of Hajihosseini et al. (2019) who assessed the monthly flow with the SWAT model in the Upper and Lower Helamand basins, respectively. We will clarify this in the revised manuscript.

Scarcity of data is a problem that is faced by *any* type of model application in science and engineering. Available observation technology cannot provide us with the necessary observations to meaningfully quantify the natural heterogeneity of the (sub-)surface properties of large scale terrestrial hydrological systems, i.e. anything larger than the hillslope scale (e.g. Beven, 2001, 2006; Hrachowitz et al., 2014; Zehe et al., 2014). To obtain the parameters of hydrological models (which represent these properties) we therefore have to resort back to the calibration of inverse model applications. Given the number of parameters in these models, data scarcity frequently limits our ability to meaningfully constrain model parameters, thereby making the modelling problem ill-posed and resulting in the well-known equifinality problem (e.g. Beven and Binley, 1992; Savenjie, 2001; Beven, 2006; Clark et al., 2011; Hrachowitz and Clark, 2017). From that perspective, our model here does not "alleviate the problem of data scarcity". Rather, we have gone quite some length to extract as much information as possible from the available data to *constrain the parameters of our model* in this data-scarce environment to avoid the adverse effects of equifinality as much as possible.

However, what are our model results finally permitted us to do, within the margins of uncertainty associated to the equifinality problem, is to estimate stream flow in periods when no observations were available at some gauging stations in the study basin. From this alternative perspective, our model can indeed be considered to "alleviate the problem of data scarcity".

Comment:

Results and Discussion: The result was well presented; however, the discussion was not adequately presented. The authors need to discuss the result by comparing or contrasting conclusions made with relevant literature. The underlying physical processes and human activities that were outlined to be responsible for the derived result should have strong theoretical underpinning. Only then can the result have a strong scientific basis and meaning. If this is not done, this section may look like a mere presentation of result. I will suggest the authors separate the result from the discussion. It will help to know where the result presentation ends and where the discussion begins.

Response:

We acknowledge and appreciate this comment. However, given the stepwise nature of the analysis we would strongly prefer to have the results of the individual steps closely associated to the interpretation and discussion thereof. However, we will extend the discussion and provide some more detailed context and comparisons with the results of previous similar studies in the revised version of the manuscript.

Comment:

It is fascinating to see that the authors presented a limitation of this research. Clarity on limitations is beneficial in outlining the extent of possible errors. However, it will be more meaningful, for instance, if the authors specifically suggest a likely better approach to help limit uncertainties inherent in the result. For instance, is there a hydrological model that does better in the data-scarce region and that will capture different processes as highlighted by the model used; otherwise could developing such a model be a way to limit uncertainties? Moreso, will comparing different meteorological dataset to determine which best captures drought (with reference to SPEI and SPI) within the study site help limit uncertainty for better policy formulation on water resource management? All these specific forward-looking documentation can be made to readers to know about possibilities and solutions since the data may inaccurately depict reality. And in-turn improve the quality of the manuscript.

Response:

We thank the Referee for the positive feedback on the limitation part of the study. As mentioned in one of the replies above, by customizing our model to the study basin and the data available there, we attempted to find an efficient trade-off between model complexity and model uncertainty. While we can of course not exclude that an alternative model formulation may do a better job, we still believe that we were able to strike a good balance between the process resolution and the resulting uncertainty in our model. This is in particular true as, after calibration for the 1971-1975 period, the daily model outputs for the gauge that recorded flow for the entire 1971-2006 period (ID8) remained rather good for the entire 35-year (!) study period (i.e. calibration period 1971-1975 and validation period 1976-2006; Table 4 and Figure 4). The best way to reduce uncertainties would be to do more observations and generate data, which could then be confronted with the model. Deficiencies in the model to reproduce this additional data could then, in an iterative process allow model improvement (e.g. Fenicia et al., 2008; Hulsman et al., 2020). However, at this stage, we believe that further model improvement will be difficult as the introduction of more complexity in the model will not be warranted by the actually available data and eventully merely lead to increased equifinality (Beven, 2006). We will clarify this in the revised manuscript.

We strongly agree with the reviewer that using alternative data sets, e.g. for precipitation, may lead to different results and interpretations. Although outside the scope of our manuscript, a future comparison of different data sets is therefore highly recommended. We will strongly emphasize this in the revised version of the manuscript.

Comment:
The grammar is sometimes not correct. The authors should allow a native speaker to help improve some sentences.
Response:
We will improve the grammar in the revised version.

Comment:
Line 97: remove "s" from occurs.
Response:
The word *occur* in this sentence refers to *precipitation*. Thus to the "s" is needed.

Comment:

Line 117-118: the sentence on this line ended with "most of the crops located in the traditionally irrigated areas (Wardlaw et al., 2013)". I feel this sentence is incomplete or rather add no additional meaning to the preceding sentence.

<u>Response:</u>
The sentence will be corrected to " … and large areas of opium poppy are grown especially in the traditionally irrigated area (Wardlaw et al., 2013)".

Comment:
Line 151: put "."After Ssn.
<u>Response:</u>
It will be corrected in the revised manuscript.

Comment:
Line 161: delete "that".
<u>Response:</u>
It will be corrected in the revised manuscript.

Comment:
Line 261: remove "the" before give.
<u>Response:</u>
It will be corrected in the revised manuscript.

Comment:
Line 294: Sentence not understandable. The authors placed Iran after requirement. The sentence needs to be rewritten to portray real meaning.
<u>Response:</u>
The sentence will be modified to "to meet irrigation demand in the downstream Helmand Valley and to satisfy the flow requirements of the Sistan River in Iran under the Iranian-Afghan Helmand River Water Treaty (1973)".

Comment:
Line 300-304: more explanation needs to be given about streamflow loses. Since soil evaporation is not enough reason as to why 60% of stream water is lost in the hyperarid region, an explanation should be provided from the literature with regards to the inferences made by the author. Since the authors outlined that deep aquifer recharge and deep infiltration may be responsible for loses, more explanation should be given with references from existing literature.
<u>Response:</u>
We would like to thank the Referee for raising our attention to this point. Addressing this comment we would like to clarify that we considered one additional parameter (K_L) to account for losses between ID7 (LHRB) and ID8 (SISP). It is true that we too confidently attributed these losses almost exclusively to deep groundwater export.

Although this of course may play a role as in many other basins worldwide (e.g. Schaller and Fan, 2009; Bouaziz et al., 2018; Condon et al., 2020), we overlooked another even much more plausible source of these losses: when the Helmand River reaches Iran, it bifurcates just upstream the gauge at SISP (ID8) into the Sistan river (SISP, ID8), which drains into the Hamun wetlands, and the completely ungauged Common Parian River, which follows the border between Iran and Afghanistan. Therefore the lumped loss factor (K_L) combines the effects of deep infiltration, soil evaporation and particularly the proportion of water which is diverted into the Common Parian River. We will correct the definition of this parameter in the revised manuscript. Unfortunately there is not sufficient additional data available in the study area to directly estimate the losses. Merely Burger (2005) in a study of the Helmand River of Afghanistan and Iran loosely mentioned some schematization errors in their project including the existence of flow diversions from the Helmand River into the Common Parian River.

Comment:
Line 350: delete "with" after was.
Response:
It will be corrected in the revised manuscript.

Comment:
Line 355-359: there is a need to corroborate the role Ep plays in intensifying or moderating drought. The authors need to discuss with relevant literature.
Response:
Thank you for your suggestion. We will discuss the role of E_P as an effective factor on droughts in the revised manuscript. However, please note that in arid environments such as in the study region, fluctuations in E_P will have a limited effect on E_A and thus on water deficits. As by definition, these regions are limited by water rather than by energy supply, E_A will not significantly change when precipitation remains stable at e.g. 400 mm yr^{-1} while E_P fluctuates from e.g. 1000 mm yr^{-1} in one year to 1300 mm yr^{-1} in the following year. In either case, most of the 400 mm yr^{-1} of available water will be evaporated as E_A. Changes in E_P will therefore be less relevant for the intensification/moderation of drought in such arid regions than changes in precipitation.

Comment:
Line 434-436: The authors should compare or contrast their findings/conclusions with relevant literature.
Response:
We agree. In the revised manuscript we will extend the discussion and better place our results into a broader context of previous studies.
Comment:
Line 437: recast sentence.
Response:
We will adjust this sentence to "Such short-term influences of hydrological drivers deficits, would be likely to manifest themselves in the evolution of ΔSDI characterized by more erratic temporal pattern".

Comment:

Line 441: remove "s" from estimate.
Response:
It will be removed in the revised manuscript.

Comment:
Line 486-487: the sentence starting with demand is not understandable, please recast.
Response:
We will change the sentence in the revised manuscript to: "Irrigation demand in the lower basin accounts with I_D ~ 13 mm y^{-1} for a substantially larger fraction of the water balance (~ 14%) than in the upper basin."

Comment:
Line 500-501: what do the authors mean by "consequence of the extension the. . ..." please recast to portray the intended meaning
Response:
we will recast this sentence to " … is largely a consequence of increases in agriculturally used area which resulted in increases of the related irrigation water demand".

Comment:
Line 508: Do not start the sentence with "this.
Response:
It will be removed in the revised paper.

References:

Beven, K. J, and Binley, A. M.: The future of distributed models: model calibration and uncertainty prediction, Hydrol. Processes., 6, 279– 298, https://doi.org/10.1002/hyp.3360060305, 1992.

Beven, K. J.: A manifesto for the equifinality thesis, J. Hydrol., 320, 18–36, https://doi.org/10.1016/j.jhydrol.2005.07.007, 2006.

Beven, K.: On hypothesis testing in hydrology, Hydrol. Processes, 15, 1655-1657, https://doi.org/10.1002/hyp.436, 2001.

Bouaziz, L., Weerts, A., Schellekens, J., Sprokkereef, E., Stam, J., Savenije, H., and Hrachowitz, M.: Redressing the balance: quantifying net intercatchment groundwater flows, Hydrol. Earth Syst. Sci., 22, 6415–6434, https://doi.org/10.5194/hess-22-6415-2018, 2018.

Burger, A. S.: Calibration of a precipitation runoff model for the Hirmand River in Afghanistan and Iran using remote sensing data, master thesis, Delft University of Technology, Faculty of civil engineering and Geosciences, 2005.

Clark, M. P., Kavetski, D., and Fenicia, F.: Pursuing the method of multiple working hypotheses for hydrological modeling, Water Resour. Res., 47, W09301, https://doi.org/10.1029/2010WR009827, 2011.

Condon, L. E., Atchley, A. L. and Maxwell, R. M.: Evapotranspiration depletes groundwater under warming over the contiguous United States. Nat. Commun. 11, 873. https://doi.org/10.1038/s41467-020-14688-0, 2020.

Fenicia, F., Kavetski, D., and Savenije, H. H. G.: Elements of a flexible approach for conceptual hydrological modeling: 1. Motivation and theoretical development, Water Resour. Res., 47, 1-13, https://doi.org/10.1029/2010WR010174, 2011.

Fenicia, F., Savenije, H. H. G., Matgen, P., and Pfister, L.: Is the groundwater reservoir linear? Learning from data in hydrological modelling, Hydrol. Earth Syst. Sci., 10, 139–150, doi:10.5194/hess-10-139-2006, 2006.

Fenicia, F., Savenije, H. H. G., Matgen, P., and Pfister, L.: Understanding catchment behaviour through stepwise model concept improvement, Water Resour. Res., 44, 1-13, https://doi.org/10.1029/2006WR005563, 2008.

Gao, H., Ding, Y., Zhao, Q., Hrachowitz, M., and Savenije, H. H. G.: The importance of aspect for modelling the hydrological response in a glacier catchment in Central Asia, Hydrol. Processes, 31, 2842–2859, https://doi.org/10.1002/hyp.11224, 2017.

Gao, H., Hrachowitz, M., Fenicia, F., Gharari, S., and Savenije, H. H. G.: Testing the realism of a topography-driven model (FLEX-Topo) in the nested catchments of the Upper Heihe, China, Hydrol. Earth Syst. Sci., 18, 1895–1915, https://doi.org/10.5194/hess-18-1895-2014, 2014.

Hajihosseini, H., Hajihosseini, M., Morid, S., Delavar, M., and Booij, M. J.: Hydrological Assessment of the 1973 Treatyon the Transboundary Helmand River, Using the SWAT Model and a Global Climate Database, Water Resour. Manag., 30, 4681–4694, https://doi.org/10.1007/s11269-016-1447-y, 2016.

Hajihosseini, M., Hajihosseini, H., Morid, S., Delavar, M., and Booij, M. J.: Impacts of land use changes and climate variability on transboundary Hirmand River using SWAT, J. Water Clim. Change [Preprint], https://doi.org/10.2166/wcc.2019.100, 18 October 2019.

Hrachowitz, M. and Clark, M. P.: HESS Opinions: The complementary merits of competing modelling philosophies in hydrology, Hydrol. Earth Syst. Sci., 21, 3953–3973, https://doi.org/10.5194/hess-21-3953-2017, 2017

Hulsman, P., Savenije, H. H. G., and Hrachowitz, M.: Learning from satellite observations: increased understanding of catchment processes through stepwise model improvement, Hydrol. Earth Syst. Sci. Discuss. [preprint], https://doi.org/10.5194/hess-2020-191, in review, 2020.

Nijzink, R. C., Almeida, S., Pechlivanidis, I. G., Capell, R., Gustafssons, D., Arheimer, B., Parajka, J., Freer, J., Han, D., Wagener, T., van Nooijen, R. R. P., Savenije, H. H. G., and Hrachowitz, M.: Constraining Conceptual Hydrological Models With Multiple Information Sources, Water Resour. Res., 54, 1-31, https://doi.org/10.1029/2017WR021895, 2018.

Savenije, H. H. G.: Equifinality, a blessing in disguise? HP Today Invited commentary, Hydrol. Process., 15, 2835–2838, 2001.

Schaller, M. F. and Fan, Y.: River basins as groundwater exporters and importers: implications for water cycle and climate modeling. J. Geophys. Res. Atmos. 114. http://dx.doi.org/10.1029/2008JD010636, 2009.

Wardlaw, R., Goes, B., Parajuli, U., and Hancock, I.: Helmand River Basin Master Plan, Phase 3, Water Resources Modelling for Helmand River Basin, Ministry of Energy and Water, Kabul, Technical Report., 3, 251 pp., 2013.

Zehe, E., Ehret, U., Pfister, L., Blume, T., Schröder, B., Westhoff, M., Jackisch, C., Schymanski, S. J., Weiler, M., Schulz, K., Allroggen, N., Tronicke, J., van Schaik, L., Dietrich, P., Scherer, U., Eccard, J., Wulfmeyer, V., and Kleidon, A.: HESS Opinions: From response units to functional units: a thermodynamic reinterpretation of the HRU concept to link spatial organization and functioning of intermediate scale catchments, Hydrol. Earth Syst. Sci., 18, 4635–4655, https://doi.org/10.5194/hess-18-4635-2014, 2014.

Reply to the comments from Referee #2

Comment:

Dear authors, I have read the manuscript "Signatures of human intervention – or not? (...)" with great interest, as I think the attribution of droughts to human or natural phenomena can be very interesting with regards to drought risk reduction policies. The study clearly states its hypotheses and is able to – with limited amount of info available model droughts in the Helmand River Basin, including both hydrological and human components. The manuscript is well written, and I particularly do like the creative graphics and the nuance at the end of the paper. While I am satisfied with the general setup and idea behind the analysis, I am left with a few concerns regarding the method, therefore I would like to recommend a major revision for this paper.

Response:

We appreciate Referee #2 for the very constructive feedback and detailed comments provided, which helped us to improve the manuscript. Please find the detailed responses to the individual comments below. We will incorporate these comments in the revised version of the manuscript. We hope that our responses to the reviewer comments clarify all issues raised.

Comment:

The use of 10year periods to conclude about large trends is questionable. As droughts are (supposed to be) an extreme event, it is possible that some decades have more droughts than others without pointing to any climate- or human-related trend. Why not dividing it only in two periods? Or only looking at average trends?

Response:

We would like to thank the Referee for her/his comment. We strongly agree with the reviewer that droughts are extreme and thus by definition rare events that may extend over longer time scales. Individual decades may thus indeed exhibit varying occurrences of drought and to better understand climate- or human-induced effects on droughts longer time periods need to be considered. To do this and to limit the potential of misinterpretations, we therefore first analysed the long-term or "average" trends over the entire study period, as suggested by the reviewer and as shown in Figures 7b,d,f, 8c and 10 in the original manuscript. To provide the reader with more detail and to get a closer understanding of the underlying processes we then, in addition, zoomed in to the individual decades. This then allowed us to better analyse the overall trends and to see if there was indeed a systematic shift over time in the observed pattern or if the overall trends were merely artifacts of extreme events in the first or last decade of the study period. While the decadal considerations for the temporal analysis in section 5.2 and Figure 7 do not add substantially more information but merely support the overall trends, they provide very relevant and interesting insights in the analysis of the spatial drought pattern in section 5.3 of the original manuscript. This can, in particular, be seen in Figure 8b, where the decadal discretization reveals a clear gradual and systematic shift between upstream and downstream drought characteristics over the study period. We would therefore strongly prefer to keep the original decadal discretization as an addition to the long-term trends. In addition, we will add a Figure in the Supplementary Material showing the same analysis with a discretization of the study period into 2 periods only, as suggested by the reviewer, to illustrate also the longer-term validity of our interpretation.

Comment:

A lot of your analysis of droughts is dependent on the assumption you make regarding the reservoir routing (line164+, in particular on line 175). The routing through reservoirs during low flow, in this case the most interesting one, has a rather low R2 (0.57). Have you done a sensitivity analysis to see how this affects your results? I think this should be more prominent in the discussion. Besides, there is the assumption of human reservoir operations that are absent, assuming the outflow is not adjusted by humans, but using the empirical link with the total storage distinguishing low and high flow, but make no distinction between drought and more-than-average-Q years. The conclusions about the influence of the reservoirs on the propagation of droughts should reflect this uncertainty.

Response:

We completely agree with the reviewer that a major source of uncertainty in the analysis is the lack of more detailed data on the operation of the two reservoirs in the Helmand River Basin. Although the storage-discharge relationship is statistically significant ($p<0.001$), the effect size remains modest. It is plausible to assume that reservoir operation is more careful during drier years. However, this should, at least to some degree, be implicit in the data we used to develop the reservoir routing scheme: during dry seasons water storage in the reservoirs is lower during drought periods than during wetter periods. Of course, this may not always apply. Although it may be a potentially very valuable idea to develop an individual reservoir operation scheme for drought periods only, we unfortunately do not have sufficient data to develop such relationships that are also sufficiently robust. We thus prefer to use all available data to inform a generally applicable scheme. Please also note that the estimated water release from the reservoirs results in overall model outputs in all downstream basins that are quite consistent with the observed daily river flow, which at station SISP (ID8) is even true for the entire 35-year study period. In spite of all other sources of uncertainty throughout the modelling process, this in itself can already be seen as an indication of the plausibility of the modelled reservoir outflow. However, we fully agree with the reviewer that a more detailed sensitivity analysis of the effects of the uncertainties in the reservoir routing scheme on downstream droughts will be very informative to better understand the limitations of or results. We therefore will add a detailed discussion on the effects of such a sensitivity analysis in the revised manuscript.

Comment:

I feel I do not understand the additional parameters (such as deep infiltration losses) well: how are they parameterised? How do you know for sure this water is lost due to percolation? What is the importance and sensitivity of Snowmelt in the model? Since humans are not effective in applying irrigation water, there must be an underestimation of the water used for irrigation? I agree the end results of the hydrological bucket model are not bad (although the intra-annual variability is not very good), but with so many parameters, how sure are you that you model the correct processes? I suggest to add this to the discussion.

Response:

We would like to thank the Referee for raising our attention to this point. Addressing this comment we would like to clarify that, in absence of further quantitative information, we parameterized these losses as a constant daily loss between stations LHRB (ID7) and SISP (ID8), as specified by the additional calibration parameter (K_L). We agree with the reviewer that we too confidently attributed these losses exclusively to deep groundwater export.

Although this of course may play a role as in many other basins worldwide (e.g. Schaller and Fan, 2009; Bouaziz et al., 2018; Condon et al., 2020), we overlooked another even much more plausible source of these losses: when the Helmand River reaches Iran, it bifurcates just upstream the gauge at SISP (ID8) into the Sistan river (SISP, ID8) which then drains into the Hamun wetlands and into the completely ungauged Common Parian River, which follows the border between Iran and Afghanistan. Therefore the lumped loss factor (K_L) combines the effects of deep infiltration, soil evaporation and particularly the proportion of water which is diverted into the Common Parian River. We will correct the definition of this parameter in the revised manuscript. Unfortunately there is not sufficient additional data available in the study area to directly estimate the losses.

Snowmelt in the Helmand River Basin, mostly originating from the headwaters in the Upper Helmand (UHRB; ID1) and the Upper Arghandad River Basins (UARB; ID4), accounts for ~80% of river flow in the downstream sub-basins. As such, snowmelt is, by far, the most dominant source of river water in the region. The importance of the meltwater contribution then also leads to a well-constrained snowmelt model component as shown by rather narrow posterior distributions of the associated model parameters T_T and F_{dd} in Table 2 of the original manuscript.

We completely agree that the rather inefficient irrigation schemes in the study region may lead to an underestimation of the actual irrigation water use. With the available data, this is problematic to quantify in more detail, though. However, please note that modelled river flow reproduces the observed daily river flow at SISP (ID8) rather well over the entire 35-year study period, whereof the 30-year period 1976-2006 was a validation period without further model recalibration. Such a rather good model performance over a 30-year validation period is in itself an indicator that the hydrological dynamics of the system are reproduced by the model in a plausible way. In addition, if the modelled irrigation is an underestimation, it would nevertheless be a conservative estimate: higher real-world irrigation water demand would even further strengthen the conclusion of our analysis, that the shift form of downstream moderation to the intensification of hydrological drought over the study period is largely an effect of human intervention. We will clarify this with a more detailed discussion in the revised manuscript.

Comment:

I would like to see the goodness of fit of the gamma and GEV distributions for the SPI SPEI and SDI. They can matter a lot, a bad distribution (for some months) could potentially affect the rest of your analysis, hence other distributions could be a solution. I was wondering if you used the Stagge et al 2015 approach to deal with zero values for the Gamma? Besides, I wonder why you would use an accumulation time of 12 months – in a very vulnerable environment as you work in, I would think an accumulation time of three months is more relevant, as the 12months can balance out dry and wet periods in the different seasons. I strongly suggest to try the same analysis with an accumulation time of 3months and see if your results still hold. Then you can indeed say you balance short and long term effects. Moreover, I do not understand why you would use a standardised value of 0 to determine a drought. Often -1 is used. Further, the whole analysis now only investigates below average conditions, that maybe not lead to any impacts. I would add the same analysis but for a threshold of -1 or -1.5, to see how real extremes change over time and through space. Again, this could really affect your conclusions. Finally, I also do not really understand how you include the lag time that usually exist between meteorological

and hydrological droughts: it is logical that the SPI12Dec1987 is not consistent with the SDI12DEC1987 because droughts travel through the hydrological cycle with a certain lag time. Did you account for this?

Response:

We would like to thank the Referee for her/his suggestion. We will add the goodness of fit of the gamma and GEV distributions for the drought indices in the revised manuscript.

Interesting point. As mentioned in section 4.2 of the original manuscript and also reiterated by the reviewer in the comment above, we used 12 months accumulation periods to estimate SPI and SDI. Thus we, fortunately, did not come into the situation that we had to deal with 0-values, as the 12-months precipitation and discharge were always non-zero. The choice of the accumulation period is indeed always delicate in drought studies. For that reason, we did a preliminary analysis on multiple accumulation periods, including 3, 6, 9, 12 and 24 months. From that analysis we believe that 12-months accumulation periods can strike a good balance between the shorter and longer-term effects. Of course, no accumulation period can fully describe all drought-related processes that occur across multiple temporal (and spatial) scales. Yet, we chose the 12-months period here as we think that it can best deal with the time lag introduced by the snow accumulation and melt dynamics which are the dominant control on river flow in the study region. Overall the differences between the accumulation periods are minor and do largely not affect the relevant features of the observed pattern, as shown in the heatmaps of the 3 drought indices for the 3 and 12 months periods here below in Figures R1 and R2.
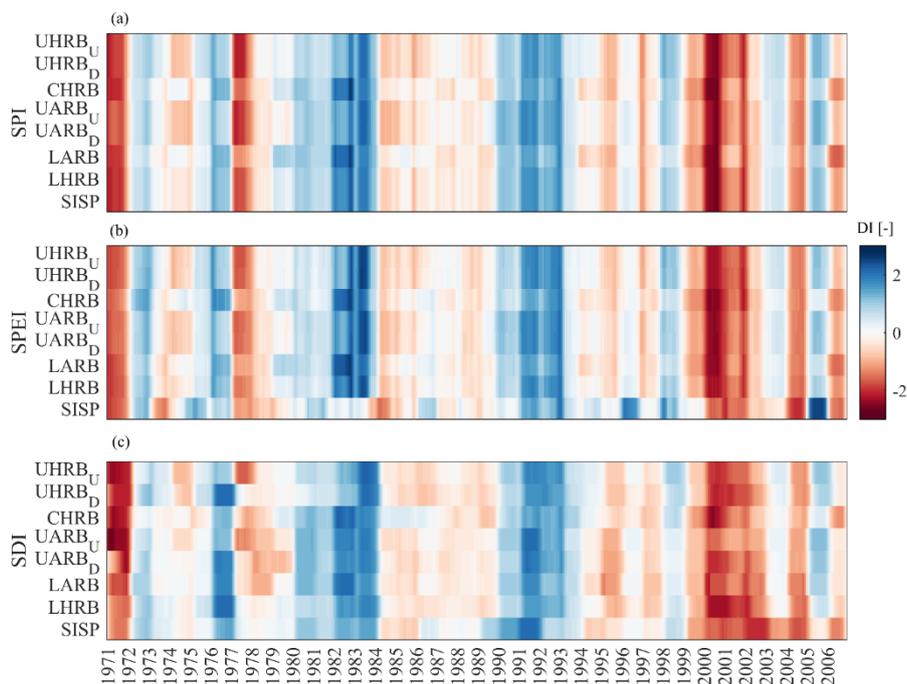

Figure R1. Time series of monthly drought indices (based on 3 months accumulation time) SPI, SPEI and SDI for the subbasins ID1 – ID8 for the 1970 – 2006 study period
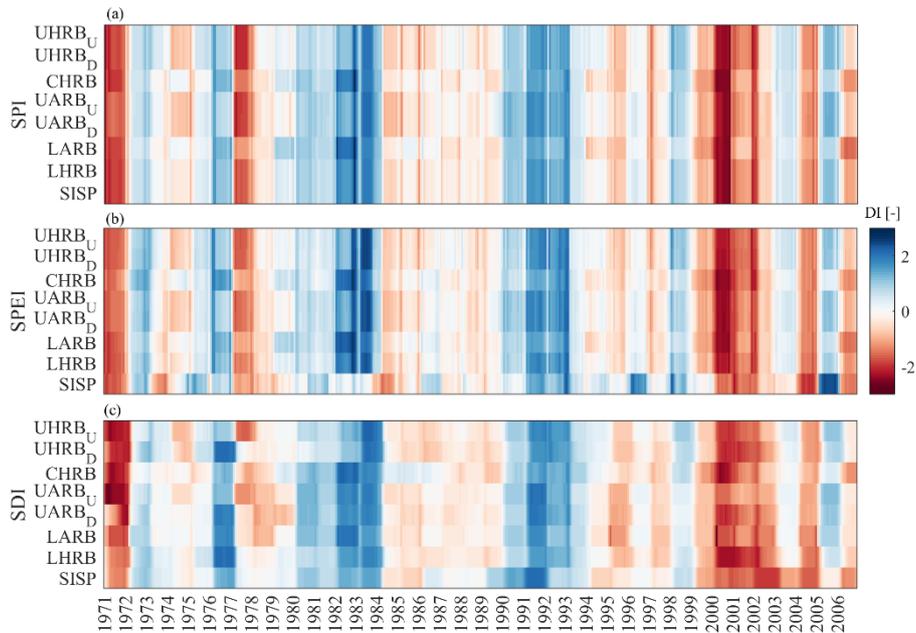
12

Figure R2. Time series of monthly drought indices (based on 12 months accumulation time) SPI, SPEI and SDI for the subbasins ID1 − ID8 for the 1970 − 2006 study period

Similarly, the choice of a drought index threshold to define droughts remains rather subjective. We agree that many studies use a threshold of -1. We here chose a threshold of 0 to provide a more comprehensive picture, including all below-average conditions. However, please note that most and the most important part of our analysis (Figures 7 − 10) do not require the definition of any threshold. Instead, we directly analyse the distributions of the individual drought indices. By doing so we avoid the need to define a binary and somewhat arbitrary decision drought yes/no. Using the full distributions has several advantages: it does not only provide a more complete picture as it includes all observed conditions, but also eliminates any further subjectivity in the analysis. However, to bring our analysis closer to what is typically done, we will adjust our drought threshold to -1 and provide re-calculated estimates of drought frequency, duration, intensity and severity in section 5.2.

We also completely agree with the last point of the reviewer here. There can be time lags of several months between meteorological and hydrological droughts. We have previously analyzed this in a preliminary cross-correlation anaylsis, which suggested a typical time lag of 3-6 months between SPI and SDI droughts (see Figure R3 here below for LHRB, ID7). However, as in this manuscript we do not compare the propagation from SPI drought to SDI drought, but rather the downstream propagation of both types of drought individually, we decided not to show the analysis on the lag times between the two as it would divert the reader from the main analysis in our manuscript without actually adding value for the interpretation.
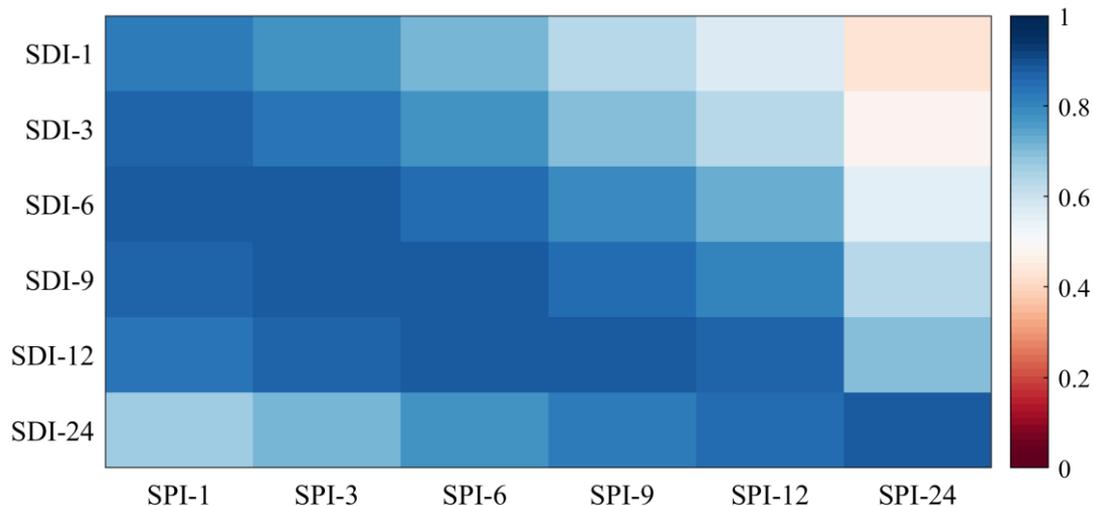
13

Figure R3. Heatmap of correlation coeffients from cross-correlation analysis between SPI and SDI for LHRB (ID7)

Comment:

In the introduction (line50+), you cite a few authors who have started to analyse droughts in Afghanistan and the HRB, but you fail to explain what your approach will add to this. Also, did they find similar results? It does not come back in your discussion sections (you cite a lot of numbers: how do they compare with other studies?).

Response:

Thank you for highlighting this important point. Addressing this comment we will provide a more detailed description of the novelty of our study and a more detailed discussion of our results in the context of previous studies which were conducted in the Helmand River Basin or similar environments. Briefly, most studies in the Helmand River Basin have so far remained limited to mere documentation and/or general assessments of mostly meteorological drought characteristics. We here extended this scope also to SPEI and SDI droughts. We particularly focus on changes in drought characteristics over time and space and attempted to explicitly and quantitatively attribute these changes to climate and human activity.

Comment:

In the abstract, I would specify human influence better. When you state "however the downstream parts of the HRB moderated the further propagation:" (l30) I would explain that this is because of the dams/reservoirs and/or land use – then it is easier to reflect on what caused the shift in this effect. Moreover, I would clearly state that you assume reservoirs without any human management.

Response:

We thank the referee for these suggestions and we will adapt the text in the abstract accordingly in the revised version of the manuscript.

Comment:

In the introduction (line70+), you refer to Mishra and Singh, but this sentence is very unclear. Please clarify that is the takeaway from this sentence.

Response:

Thank you for raising our attention to this point. We will clarify the sentence and adjust it to: "As pointed out, amongst others, by Mishra and Singh (2010) the processes underlying droughts are complex because they are dependent on many interacting processes in terrestrial hydrological systems, such as the interaction between the atmosphere and the hydrological processes which feed moisture to the atmosphere.

Comment:

I cannot find table S1 with al relevant model equations: (Only model variables)

Response:

Table S1 (Water balance and constitutive equations used in FLEX model) is in the supplementary materials of the paper.

Comment:

Why would you show the actual instead of the potential evaporation in figure 10? SPEI uses potential, so that would reflect your drought analysis better.

Response:

It is true that SPEI is typically based on potential evaporation E_P. However, we believe that in arid and thus water-limited regions fluctuations in E_P are rather irrelevant compared to fluctuations in P or E_A. In other words, there will be little difference in the partitioning of water fluxes if under the same annual precipitation of e.g. 500 mm yr^{-1}, E_P is 1000 or 1500 mm yr^{-1}, as in both cases actual evaporation E_A will be close to (or even exceed) 400 mm yr^{-1} and as therefore most of the available water will be evaporated. In contrast, more water will be evaporated as E_A (even if E_P remains stable) in years when more water is available and thus P is higher. By extension, the effects of evaporation on droughts in arid regions can only be meaningfully assessed by changes in E_A. That is also why we did not develop a strong link between SPEI and the E_A anomaly in the text as analysed in section 5.4 and figure 10, where we tried to quantify the individual role of each component of the water balance for the development of hydrological drought in the HRB. We will clarify this in the revised manuscript.

References:

Schaller, M. F. and Fan, Y.: River basins as groundwater exporters and importers: implications for water cycle and climate modeling. J. Geophys. Res. Atmos. 114. http://dx.doi.org/10.1029/2008JD010636, 2009.

Condon, L. E., Atchley, A. L. and Maxwell, R. M.: Evapotranspiration depletes groundwater under warming over the contiguous United States. Nat. Commun. 11**,** 873. https://doi.org/10.1038/s41467-020-14688-0, 2020.

Bouaziz, L., Weerts, A., Schellekens, J., Sprokkereef, E., Stam, J., Savenije, H., and Hrachowitz, M.: Redressing the balance: quantifying net intercatchment groundwater flows, Hydrol. Earth Syst. Sci., 22, 6415–6434, https://doi.org/10.5194/hess-22-6415-2018, 2018.

Mishra, A. K. and Singh, V. p.: A review of drought concepts, J. Hydrol., 391, 202-216, https://doi.org/10.1016/j.jhydrol.2010.07.012, 2010.

Marked up manuscript

Signatures of human intervention – or not? Downstream intensification of hydrological drought along a large Central Asian River: the individual roles of climate variability and land use change

Artemis Roodari[1,2], Markus Hrachowitz[1], Farzad Hassanpour[2] and Mostafa Yaghoobzadeh[3]

[1]Department of Water Management, Faculty of Civil Engineering and Geoscience, Delft University of Technology, Stevinweg 1, 2628CN Delft, Netherlands
[2]Department of Water Engineering, Faculty of Soil and Water Sciences, University of Zabol, Zabol, 98615-538, Iran
[3]Department of Water Engineering, University of Birjand, Birjand, 97175-615, Iran

Correspondence to: Artemis Roodari (artemis_roodari@yahoo.com)

Abstract. The transboundary Helmand River basin is the main drainage system for large parts of Afghanistan and the Sistan region of Iran. Due to the reliance of this arid region on water from the Helmand River, a better understanding of hydrological drought pattern and the underlying drivers in the region are critically required for effective management of the available water. The objective of this paper is therefore to analyse and quantify spatio-temporal pattern of drought and the underlying processes in the study region. More specifically we test for the Helmand River Basin the following hypotheses for the 1970-2006 period: (1) drought characteristics, including frequency and severity systematically changed over the study period, (2) the spatial pattern and processes of drought propagation through the Helmand River Basin also changed and (3) the relative roles of climate variability and human influence on changes in hydrological droughts can be quantified.

It was found that drought characteristics varied throughout the study period, but did largely show no systematic trends. The same was observed for the time series of drought indices SPI and SPEI, which exhibited considerable spatial coherence and synchronicity throughout the basin indicating that, overall, droughts similarly affect the entire HRB with little regional or local differences. In contrast, analysis of SDI exhibited significant negative trends in the lower parts of the basin, indicating an intensification of hydrological droughts. It could be shown that with a mean annual precipitation of ~250 mm y^{-1}, streamflow deficits and thus hydrological drought throughout the HRB are largely controlled by precipitation deficits, whose annual anomalies on average account for ±50 mm y^{-1} or ~20% of the water balance of the HRB, while anomalies of total evaporative fluxes on average only account for ±20 mm y^{-1}. Assuming no changes in the reservoir management practices over the study period, the results suggest that the two reservoirs in the HRB only played a minor role for the downstream propagation of streamflow deficits, as indicated by the mean difference between inflow and outflow during drought periods which did not exceed ~0.5% of the water balance of the HRB. Irrigation water abstraction had a similarly limited effect on the magnitude of streamflow deficits, accounting for ~10% of the water balance of the HRB. However, the downstream parts of the HRB *moderated* the further propagation of streamflow deficits and associated droughts because of the minor effects of reservoir operation and very limited agricultural water in the early decades of the study period. This drought moderation function of the lower basin was gradually and systematically inverted by the end of the study period, when the lower basin

eventually *amplified* the downstream propagation of flow deficits and droughts. Our results provide plausible evidence that this shift from drought moderation to drought amplification in the lower basin is likely a consequence of increased agricultural activity and the associated increases in irrigation water demand from ~13 mm y^{-1} at the beginning of the study period to ~23 mm y^{-1} at the end and thus in spite of being only a minor fraction of the water balance. Overall the results of this study illustrate that flow deficits and the associated droughts in the HRB clearly reflect the dynamic interplay between temporally varying regional differences in hydro-meteorological variables together with subtle and temporally varying effects linked to direct human intervention.

1 Introduction

There is evidence that droughts have the potential to increasingly affect human societies as well as ecosystem functioning. In a world under change, decision-makers therefore need reliable quantitative information about drought characteristics to ensure the development and implementation of effective and sustainable water management procedures. To be reliable this information needs to be based on a solid understanding of how different types of droughts propagate through different hydrological systems. While meteorological droughts are controlled by precipitation deficits only, agricultural and hydrological droughts are caused by soil moisture and runoff deficits, respectively. As pointed out, amongst others, by Mishra and Singh (2010) the processes underlying droughts are complex because they are dependent on many interacting processes in terrestrial hydrological systems, such as the interaction between the atmosphere and the hydrological processes which feed moisture to the atmosphere. Therefore, monitoring and analysis of hydrological droughts have received increased attention in recent decades (van Huijgevoort et al., 2014; Pathak and Dodamani, 2016; Weng et al., 2015; Vicente-Serrano et al., 2012; Kubiak-Wójcicka and Bak, 2018; Trambauer et al., 2014; Ahmadalipour et al., 2017; Jiao and Yuan, 2019; Moravec et al., 2019). In general, it is well-understood that both, agricultural and hydrological droughts are modulated by the interactions of climate, river basin characteristics, such as geology, as well as a human influence or any combination thereof (e.g. Van Lanen et al., 2013; Huang et al., 2016; Liu, et al., 2016; Van Loon, et al., 2019). For example, data show that reservoir operations can have both, considerable positive or negative effects on downstream hydrological drought pattern (e.g. Zhang et al., 2013; Pingue et al., 2016; Wu et al., 2017), which may politically be particularly sensitive for transboundary rivers in arid envrionments (Al-Faraj and Scholz, 2015, Wan et al., 2018).

The transboundary Helmand River system between Afghanistan and Iran is the primary contributor of water to the Hamun lake- and wetland-system in the Sistan Plain, which is the terminus of one of the largest endorheic basins in Central Asia. In this region, which is described as one of the driest, most remote deserts on Earth (Whitney, 2006), water from the Helmand River system plays a critical role not only to sustain agricultural production, hydropower generation and ecosystem stability but also for drinking water supply for some one million people living in the region, including the cities of Kandahar in Afghanistan and Zabol in Iran.

The area has recently experienced a severe, multi-year drought (1998–2004). Reduction of flow and episodic no-flow conditions in the Helmand River during this period have caused significant disruption of water supply. As a consequence, agricultural production dropped by almost 90% as compared to average no-drought conditions, further resulting in food shortage and considerable economic damage (Ebrahimzadeh and Esmaelnejad, 2013). Given the region's extreme dependence on water from the Helmand River system and the associated vulnerability to hydrological droughts, a few recent studies started to analyse droughts in Afghanistan and the Helmand River Basin (e.g. Ahmad and Wasiq, 2004; Miyan, 2015). For example, Alami et al. (2018) analyzed meteorological droughts in the Helmand River Basin using different methods and quantitatively documented the extreme drought in 2001. However, most of the research in this region focused on the application of hydrological models for the simulation of runoff to provide decision bases for integrated water management issues in the region. These studies include Hajihosseini et al. (2016), who assessed the Afghan-Iranian Helmand River Treaty (The Iranian-Afghan Helmand (Hirmand) River Water Treaty, 1973) using the SWAT model (Arnold et al., 1998) and data from the Climatic Research Unit (CRU; Harris et al., 2014). A study by Wardlaw et al. (2013) formulated a model for the development of water resources systems in the Helmand River Basin using the Water Evaluation and Planning (WEAP) model and established a list of scenarios for the future.

Similarly, Vining and Vecchia (2007) estimated future runoff conditions of the river to evaluate the effects of different reservoir operation strategies under different climate change scenarios on downstream water supply. Van Beek et al. (2008) developed methods and tools to build the capacity to sustain agriculture and ecosystems in the downstream Sistan Plain. In spite of this growing body of literature for the region, the scarcity of reliable meteorological and hydrological data so far limited systematic, quantitative analysis of spatio-temporal pattern of hydrological droughts and the underlying drivers and processes in the Helmand River Basin.

In a world under change, however, decision makers need such quantitative information about drought characteristics to ensure the development and implementation of effective and sustainable water management procedures. To be reliable this information needs to be based on a solid understanding of how different types of droughts propagate through different hydrological systems. While meteorological droughts are controlled by precipitation deficits only, agricultural and hydrological droughts are caused by soil moisture and runoff deficits, respectively. As pointed out, amongst others, by Mishra and Singh (2010) the processes underlying the latter two are complex because they are dependent on many interacting processes in terrestrial hydrological systems, such as the water and release dynamics of the subsurface as well as land-atmosphere exchange. Therefore, monitoring and analysis of hydrological droughts have received increased attention in recent decades (van Huijgevoort et al., 2014; Pathak and Dodamani, 2016; Weng et al., 2015; Vicente-Serrano et al., 2012; Kubiak-Wójcicka and Bak, 2018; Trambauer et al., 2014; Ahmadalipour et al., 2017; Jiao and Yuan, 2019; Moravec et al., 2019). In general, it is well understood that both, agricultural and hydrological droughts are modulated by the interactions of climate, river basin characteristics, such as geology, as well as human influence or any combination thereof (e.g. Van Lanen et al., 2013; Huang et al., 2016; Liu, et al., 2016; Van Loon, et al., 2019). For example, data show that reservoir operations can have both, considerable positive or negative effects on downstream hydrological drought pattern (e.g. Zhang et al., 2013;

Due to the reliance of the region on water from the Helmand River, a better understanding of hydrological drought pattern and the underlying processes in the region are critically required for effective management of the available water. Most studies in the Helmand River Basin have so far remained limited to mere documentation and/or general assessments of mostly meteorological drought characteristics. We here will extend this scope also to hydrological drought and evaluate the meteorological drought under the additional role of atmospheric water demand. The overall objective of this paper is therefore to analyse and quantify changes in spatio-temporal pattern of drought charaterisitics and the underlying processes in the study region in an attempt to quantitatively attribute these changes to climate and human interventions, respectively. More specifically we will test for the Helmand River Basin the following hypotheses for the 1970-2006 period: (1) drought characteristics, including frequency and severity systematically changed over the study period, (2) the spatial pattern and processes of drought propagation through the Helmand River Basin also changed and (3) the relative roles of climate variability and human influence on changes in hydrological droughts can be quantified.

2 Study area

The endorheic Helmand River Basin (HRB; Figure 1) covers approximately 105,000 km^2 or 15% of Afghanistan. From its source area, in the Koh-i-Baba mountains, an extension of the Hindu Kush west of Kabul, with elevations to over 4600 masl, the Helmand River system drains into the Hamun lake and wetland system in the Sistan plain of Eastern Iran, a closed inland delta with a minimum elevation of 440 masl in the south-west of the HRB, which covers 5 % of the total HRB area (Goes et al., 2016). Both, long-term mean annual precipitation ($\bar{P}$=90-480 mm yr^{-1}; Figure 1d) and potential evaporation ($\overline{E_P}$ =700-1800 mm yr^1; Figure 1e) exhibit considerable spatial variability throughout the HRB. This results in a pronounced gradient of aridity from sub-arid in the North-East to hyper-arid conditions in the South-West as expressed by the aridity index I_A ($I_A = \frac{\bar{P}}{\overline{E_P}}$ [-]; Figure 1f). Precipitation falls mostly in the winter months and in the upper basin almost always occurs as snow. In general, snowmelt generates the annual runoff peaks in early spring and sustains flow in the HRB throughout the dry summers. For the following analysis, the HRB is divided into six sub-basins (Figure 1c; Table 1): the Upper Helmand River Basin (UHRB) with the main stem of the Helmand River, the Central Helmand River Basin (CHRB) and the Upper Arghandab River Basin (UARB) as well as the Lower Arghandab River Basin (LARB) are nested in and drain into the Lower Helmand River basin (LHRB) and subsequently into the Sistan plain (SISP). The UHRB accounts for 80% of the combined inflow into the LHRB. Flow in the LHRB is influenced by the operation of two upstream reservoirs (Figure 1b; Table 1). While the reservoir at Kajakai Dam with a storage capacity of 1800 mio. m^3, located at the outflow of the UHRB, is a multi-purpose structure for electricity production, flood control and irrigation water supply, the smaller Dahla Dam, located at the outlet of the UARB into the LARB about 180 km upstream of the confluence with the LHRB, has a storage capacity of 450 mio. m^3 and is used mainly for irrigation of the lower Arghandab valley (Goes et al., 2016).

Due to the arid climate, natural vegetation is very scarce and mostly limited to seasonal grassland throughout the entire HRB. Irrigated agriculture is by far the largest consumer of water, accounting for 98 % of all abstractions (Goes et al., 2016). Except for a few recent irrigation projects in the LARB and LHRB, irrigation relies on traditional methods with irrigation canals and is thus largely confined to the valley floors along the main river channels (Figure 1b). While the irrigated area in the LARB remained somewhat stable at around 370 km^2 (~ 0.3 % of the total HRB) over the last decades, satellite imagery (Landsat 7, ETM+) shows that the total irrigated area in the LHRB more than doubled from < 800 km^2 (0.8 %) in the late 1970s to 1650 km^2 (1.6 %) in 2011 (Figure 2). More than 200 km^2 of the increase in the irrigated area is due to the conversion of seasonal grasslands to high-water-requirement poppy cultivation since the 1990s (Hajihosseini et al., 2019). By 2006 around 690 km^2 in the HRB were used for poppy cultivation (UNODC, 2006). In 2011, the main crops in the HRB were wheat (~47 %), poppy (~ 32 – 37 %), maize and beans (~16 %), with orchards in some areas (~1–4 %), and large areas of opium poppy, mostly grown in the traditionally irrigated area (Wardlaw et al., 2013). most of the crops located in the traditionally irrigated areas (Wardlaw et al., 2013).

3 Climatological and hydrological data

The HRB is characterized by poor coverage of reliable historic in-situ observations of hydro-climatic variables, particularly in the upper parts of the basin where most of the water in the HRB originates from. Analysis of Hajihosseini et al. (2016) indicated that the spatio-temporal variation of the interpolated historical precipitation and temperature in the gridded Climatic Research Unit (CRU) dataset was largely consistent with available ground observations for Afghanistan. Therefore, we here used daily precipitation and temperature estimates for the 1970–2006 study period (Figure 1a), downscaled from the monthly CRU TS 3.10 dataset (Harris et al. 2014), based on the dGen algorithm (Geng et al., 1986) that was previously also applied in other studies (e.g. Schuol and Abbaspour, 2006; Schuol et al., 2008; Hajihosseini et al., 2016). The data were available from www.2w2e.com (Ashraf Vaghefi et al., 2017) at a spatial resolution of 0.5°×0.5°.

Daily streamflow observations for the 1970-1979 period are available from the US Geological Survey (waterdata.usgs.gov) at six gauging stations throughout the HRB (IDs 1-2, 4-7; Figure 1; Table 1). Note, that there were observations available from individual gauging stations at the inlets upstream of the Kajakai Dam (ID1 – UHRB$_U$) and Dahla Dam Reservoirs (ID4 - UARB$_U$) as well as at the corresponding outlets downstream of the dams (ID2 – UHRB$_D$; ID5 – UARB$_D$). In addition, monthly flow observations for the 1970-2006 period were available at the inflow to the Sistan Plain (ID8 – SISP).

4 Methods

The analysis of the characteristics and pattern of hydrological droughts in the HRB over the recent decades in this study required a two-step approach. In a first step, the observed streamflow time series (1970-1979; Table 1) had to be extended to

cover the full 1970-2006 study period, using a hydrological model. In a second step, the modelled streamflow estimates for the 1970-2006 period at eight locations in the HRB were used to analyse hydrological droughts.

4.1 Hydrological model

We used a distributed implementation of a process-based hydrological model, based on the general concept of the FLEX model-family (e.g. Fenicia et al., 2008; Gharari et al., 2014; Bouaziz et al., 2018) to generate estimates of daily discharge from the sub-basins $UHRB_U$ (ID1), CHRB (ID3), $UARB_U$ (ID4), LARB (ID6), LHRB (ID7) and SISP (ID8). In addition, a simple reservoir routing scheme was used to estimate outflow from the two reservoirs (ID9-10), located at $UHRB_D$ (ID2) and $UARB_D$ (ID5). The distributed implementation of this model was chosen as the general model set-up was previously successfully applied in climatically similar regions (e.g. Gao et al., 2014, 2017) but also in many other settings worldwide (e.g. Fenicia et al., 2006; Kavetski et al., 2011; Nijzink et al., 2018; Hulsman et al., 2020). In general, the FLEX modelling concept applied here is underlain by a philosophy of model customization and rigorous testing to ensure the implementation of suitable model formulations and the associated more reliable model outputs in different environments (e.g. Fenicia et al., 2011).

4.1.1 Model structure at grid cell scale

The core of the model is five storage components (Figure 3) that are linked by fluxes and that conceptually represent snow storage S_{sn} [mm], interception storage S_i [mm], storage in the unsaturated root-zone S_u [mm], a fast responding component S_f [mm] that generates preferential and overland flow, and a slow responding groundwater storage S_s [mm]. A lag function represents the lag time between storm and flood peak. The snow module is based on a simple degree-day method that has been effectively applied in many conceptual models (e.g. Parajka and Blöschl, 2008; Konz and Seibert, 2010; Gao et al., 2017; Nijzink et al., 2018; Mostbauer et al., 2018). When the average daily temperature is below a threshold temperature T_t [ºC], precipitation enters the system as snowfall P_s [mm d^{-1}] and is stored in S_{sn}. When there is snow cover and the temperature exceeds T_t, snow melt M [mm d^{-1}], specified by a melt factor F_m [mm ºC^{-1} d^{-1}], sets in from S_{sn}. Precipitation falling as rain P_r [mm d^{-1}] first enters the interception reservoir S_i, specified by an interception capacity I_{max} [mm]. Water evaporates as interception evaporation E_i [mm d^{-1}] from S_i at potential rates E_p [mm d^{-1}], while water in S_i that exceeds the storage capacity I_{max} reaches the soil as throughfall P_{tf} [mm d^{-1}]. The total effective precipitation P_e [mm d^{-1}] infiltrating into the unsaturated soil root-zone S_u at any given time step is then the sum of P_{tf} and M (Gao et al., 2014). Water in the unsaturated reservoir S_u is, depending on the storage capacity S_{umax} [mm], either stored and eventually released by plant transpiration E_T [mm d^{-1}], or directly released as groundwater recharge R_s [mm d^{-1}] or preferential flow R_f [mm d^{-1}]. The response reservoirs S_f and S_s represent a fast responding storage component and a slower responding groundwater component, respectively, that both drain water to the river according to their associated time scales k_f [d] and k_s [d], so ~~that~~ that the total flow can be expressed as $Q=Q_f+Q_s$ [mm d^{-1}]. All relevant model equations are provided in Table S1 in the supplementary material.

6

4.1.2 Reservoir routing

Large reservoirs such as the Kajakai (ID9) and Dahla (ID10) Dam reservoirs in the HRB, can considerably alter downstream flow regimes (Haddeland et al., 2014; Wada et al., 2017). This has recently received growing attention and a number of studies have suggested methods to quantify reservoir outflow where reservoir operation rules are largely unknown (e.g. Coerver et al., 2018; Yassin et al., 2019). Here, the effects of the reservoirs were estimated with a simple water accounting scheme based on elevation-storage and elevation-area relationships provided in a study by Vining and Vecchia, (2007) and similar to previous work (e.g. Hanasaki et al., 2006; Wisser et al., 2010):

$$\frac{dS_r(t)}{dt} = Q_{in}(t) - Q_{out}(t) + P(t) - E_p(t), \tag{1}$$

Where S_r is the reservoir storage, P and E_p are precipitation and potential evaporation over the surface area of the reservoir at the end of the previous time step, respectively. Q_{in} is the inflow to the reservoir, Q_{out} the outflow from the reservoir. Here, the inflows Q_{in} to the two reservoirs were estimated by the hydrological models of the respective upstream sub-basins $UHRB_U$ (ID1) and $UARB_U$ (ID4). Due to the lack of more detailed data, Q_{out} was in this study estimated based on empirical storage-outflow relationships that relate modeled reservoir storage S_r (Eq.1) and Q_{in} to observations of Q_{out}, i.e. Q_{ID2} and Q_{ID5}. We decided to develop separate linear relationships for high- and low-flow seasons, i.e. January to June and July to December, respectively as preliminary analysis suggested that these were more robust than non- or piecewise-linear relationships for the entire year, as used elsewhere (e.g. Yassin et al., 2019):

$$Q_{out} = \begin{cases} a_h S_{r,t-1} + b_h Q_{in,t} + c_h & \rightarrow high\ flow\ season \\ a_l S_{r,t-1} + c_l & \rightarrow low\ flow\ season \end{cases}, \tag{2}$$

Where a [d^{-1}], b [-], c [mm d^{-1}] are coefficients and the subscripts h and l indicate high and low flow seasons, respectively. Note, that Q_{in} becomes negligible in the low flow season and the relationship collapses to a simple linear regression. Also note, that it is plausible to assume that reservoir operation is more careful during drier years than in wetter years and may have changed over the study period. Due to the lack of sufficient data, we here developed only one low flow and one high flow relationship for each reservoir over the entire study period.

4.1.3 Model implementation at (sub-)basin scale

The model was implemented in a distributed way and the flows aggregated to the (sub-)basin scale. To limit the computational requirements, the meteorological input data, available at a spatial resolution of 0.5° x 0.5°, were averaged to run the model at a grid cell size of 1° x 1° (Figure 1). The snow (S_{sn}), interception (S_i) and unsaturated (S_u) reservoirs in each model grid cell were further stratified into 500m elevation bands to account for elevation-dependent snow dynamics and the associated differences in liquid water input to the system. The combined groundwater recharge R_s and the combined

preferential drainage R_f from all elevation zones in each model grid cell was then computed as the weighted average from all individual elevation zones, based on the areal proportion of each elevation zone (cf. Fenicia et al., 2008; Euser et al., 2015). The flow Q_i generated in each of the N grid cells of a (sub-)basin j at any time step t was subsequently routed to the (sub-) basin outlet in a convolution operation with triangular lag functions (e.g. Fenicia et al., 2011) based on lag times τ_i, proportional to the mean flow distances from the individual i cells to the outlet. In addition, irrigation demand I_D [mm d^{-1}] for agriculture was accounted for by direct river water abstractions. The aggregated flow at the outlet of a (sub-)basin was then the sum of all flows routed to the outlet minus irrigation demand, i.e. $Q_{ID_j} = \sum_{i=1}^{N} \left(Q_{i,j} * h(\tau_{i,j}) \right) - I_{D,j}$ of that specific (sub-)basin j, i.e. ID1-7. At each time step, irrigation water $I_{D,j}$ was then re-applied as input to $S_{u,i}$ in grid cells i of the corresponding subbasin j that featured agricultural use. Largely being an unregulated irrigation canal system and due to the lack of more detailed information, estimates of $I_{D,j}$ were here based on crop coefficients K_c, potential evaporation E_p and effective precipitation P_e for each day k, as well as the agriculturally used area in each year l (Allen et al., 1998), according to $I_D = K_c \left(E_p - P_e \right)_{k,l} A_l$

As a baseline, crop coefficients and the agriculturally used area were estimated based on crop pattern reported by Wardlaw et al., (2013). In that report, the irrigated areas were estimated using satellite imagery from 2010/2011. To account for land-use change over the 1970-2006 study period, the estimates were adjusted to changes in the agricultural area as extracted from available satellite imagery in 1977, 1988, and 1998.

The outflow of sub-basins UHRB$_U$, i.e.Q_{ID1}, and UARB$_U$, i.e. Q_{ID4}, were used as inflow to the Kajakai Dam (ID9) and Dahla Dam (ID10) reservoirs, respectively. The resulting estimates of reservoir outflows, i.e. Q_{ID2} and Q_{ID5} (section 4.1.2) were then used as inflows into LHRB (ID7) and LARB (ID6), respectively. In addition, LHRB (ID7) received the outflows Q_{ID3} and Q_{ID6}, while LHRB (ID7) outflow Q_{ID7} subsequently drained into SISP (ID8).

The historical absence of significant snow cover in the sub-basins ID2 and ID5-8 allowed us to omit the snow component and the related parameters from the model in these sub-basins (Figure 3) and to limit the adverse effects of equifinality (Beven, 2001). Furthermore, as agriculture is largely confined to the sub-basins LARB (ID6) and LHRB (ID7), the redistribution of river water for irrigation was only implemented in these two sub-basins. Similarly, an additional parameter k$_L$ was used to account for losses between ID7 (LHRB) and ID8 (SISP)deep infiltration losses k$_L$ was used for the sub-basins ID6-ID8. The above differences resulted in two slightly different implementations of the model in the uplands and the downstream regions of the HRB, respectively, and hereafter referred to as Model-1 and Model-2 (Figure 3). Similar implementations of this model type have in the past proven successful in a range of different environments (e.g. Prenner et al., 2018; Hulsman et al., 2020).

4.1.4 Model calibration and post-calibration evaluation

The models were run on a daily time step in all sub-basins for the entire 1970-2006 period. However, in the absence of suitable data, the models could not be calibrated for all sub-basins and over the entire period. Rather, only the models of the

five sub-basin outlets UHRB$_U$ (ID1), UARB$_U$ (ID4), LARB (ID6), LHRB (ID7) and SISP (ID8) (Table 1; Figure 3) were individually calibrated for the 1970-1975 period to time series of observed flow. Note, that all model grid cells in a given sub-basin were run with the same parameter sets but with spatially distributed hydro-climatic forcing (e.g. Ajami et al., 2004; Euser et al., 2015). To limit the effects of equfinality (Beven, 2001) and to ensure robust model implementation (Euser et al., 2013; Hrachowitz and Clark, 2017), we adopted a multi-objective (Gupta et al., 1998) calibration approach, simultaneously using the Nash-Sutcliffe Efficiency (Nash and Sucliffe, 1970) of flows ($E_{NS,Q}$) and of the logarithm of flows ($E_{NS,log(Q)}$) as objective functions. The 10 (UHRB$_U$, UARB$_U$) and 8 (LARB, LHRB, SISP) free calibration parameters, respectively, in the individual models were sampled in 10^6 realizations from uniform prior distributions following a Monte Carlo strategy. The model parameters together with their prior and posterior distributions are given Table 2. To account for trade-offs in the multi-objective calibration and uncertainties in the modelling process, we kept all parameter sets that fall into the area spanned by the pareto-optimal set of solutions as feasible (e.g. Fenicia et al., 2007; Gharari et al., 2013). For brevity, we will hereafter refer to the solution with the minimum Euclidean distance D$_E$ as the "best" solution (Hrachowitz et al., 2014):

$$D_E = \sqrt{\left(1 - E_{NS,Q}\right)^2 + \left(1 - E_{NS,log(Q)}\right)^2} \qquad (3)$$

Model uncertainty intervals were constructed from the parameter sets that were retained as feasible using D$_E$ as informal likelihood measure to weight each solution (cf. Freer et al., 1996).

In addition, storage-outflow relationships for the reservoirs (ID9-10; Eq.2) to estimate water release from the associated sub-basins downstream of the reservoirs UHRB$_D$ (ID2) and UARB$_D$ (ID5) were established as ordinary least squares estimates based on inflows from the calibrated upstream sub-basins (UHRB$_U$, ID1; UARB$_U$, ID4), Equation 1 and observations of reservoir water release in the 1970-1975 period. The parameter ranges for all solutions retained as feasible for all calibrated hydrological models and both reservoir routing schemes are given in Table 2. Note that due to physiographic similarity, the uncalibrated model for CHRB (ID3) was run with the same parameter sets as UHRB$_U$ (ID1).

The robustness of the calibrated model and its ability to reproduce the time series of daily flow with respect to $E_{NS,Q}$ and $E_{NS,log(Q)}$ in the four calibration sub-basins as well as downstream of the reservoirs was evaluated for the independent 1976-1979 test period, hereafter referred to as "validation period". In addition, the model output was evaluated against the monthly time series of flow at SISP (ID8; Table 1; Figure 1) for the entire 1976-2006 study period.

4.2 Drought indices

Three previously developed drought indices, based on the general concept of standardized deficits (e.g. Moravec et al., 2019), were used here to isolate the individual influences of different factors on hydrological drought in the HRB. The role of climatic variability and thus meteorological drought was quantified with the Standard Precipitation Index (SPI) as introduced by McKee et al. (1993), which gives information about deficits in atmospheric water supply, and with the Standardized Precipitation Evapotranspiration Index (SPEI; Vicente-Serrano et al., 2010), which describes the interaction of

precipitation and energy supply as moisture deficit $D_i = \sum_{i=1}^{T}(P_i - E_{p,i})$ and thus the additional role of atmospheric water demand. In contrast, hydrological drought was quantified with the Streamflow Drought Index (SDI; Nalbantis and Tsakiris, 2009). Differences between SPI and SPEI on the one hand and SDI on the other hand were subsequently used to analyse for potential effects of anthropogenic influences, such as irrigation water abstraction. In a parametric approach, two-parameter Gamma distributions functions were here fitted to precipitation P and flow Q and then mapped to standard normal distributions using equal probability transformations (Edwards & McKee, 1997) to estimate the dimensionless drought indices SPI and SDI, respectively (e.g. Lloyd-Hughes and Saunders, 2002; Nalbantis and Tsakiris, 2009; Mishra et al., 2018), whereas generalized extreme value (GEV) distributions were fitted to moisture deficit D to estimate SPEI for each sub-basin (Stagge et al., 2015). The goodness of fit of two-parameter gamma distributions for SPI, SDI, as well as for the GEV distribution for SPEI is provided in Figures S1-S3 in the Supplementary Material. The drought indices can be computed over different time-scales, thus leading to differences in the accumulation of deficits for the corresponding variables (e.g. McKee et al., 1993; Van Loon and Laaha, 2015). Here the drought indices were computed for each month using a time scale of the 12 preceding months as accumulation periods as these were previously found to be the most balanced time scale that gives a balance between short term and long term effects (e.g. Raziei et al., 2009; Gocic et al., 2013; Spinoni et al., 2014). All normalization was carried out relative to the full 1970-2006 study period. Droughts and their associated occurrence probabilities were classified according to the scheme suggested by McKee et al. (1993) as shown in Table 3. Since the drought indices are standardized, the same drought category thresholds were used here for all three of them.

The three drought indices were in the following used to analyse different drought characteristics. It was investigated if drought frequency, duration, severity and intensity exhibit systematic shifts over time or changes in their longitudinal propagation from upstream to downstream over the 37 year study period. Drought frequency D_F [months yr^{-1}] was here defined as the average number of months per year over a specific period in which the respective drought index, i.e. SPI, SPEI or SDI, had a value < 0 1 (Table 3). Drought duration D_D [months] was defined as the period of consecutive months with drought indices continuously < 0 1. Drought severity is defined as the total deficit D_{tot} [-] of SPI, SPEI or SDI, respectively, accumulated during all individual continuous drought periods over a specified period and, to allow comparability, normalized by the total number of months N in the time period considered, i.e. $D_S = D_{tot}/N$ [month^{-1}]. Drought intensity is expressed as the ratio $D_I = D_{tot}/D_D$ [month^{-1}] (Huang et al., 2016).

5 Results and discussion

5.1 Drought indices Model performance

The hydrological models captured the magnitudes and dynamics of daily flow relatively well when compared to observations available for both, the sub-basins upstream of the reservoirs, i.e. UHRB$_U$ (ID1; Figure 4a) and UARB$_U$ (ID4) as well as for those further downstream, i.e. LARB (ID6) and LHRB (ID7; Figure 4c). For the calibration period the "best" solutions

exceeded $E_{NS,Q} > 0.70$ and $E_{NS,log(Q)} > 0.75$ for all five calibrated sub-basins (Table 4). Similar values were found for the validation period with $E_{NS,Q} > 0.70$ and $E_{NS,log(Q)} > 0.75$. The empirical relationships to route flows through the reservoirs during high and low flow periods (Eq.2) were characterized by $R^2 = 0.80$ and 0.57, respectively for the Kajakai Dam Reservoir (ID9) and $R^2 = 0.92$ and 0.76, respectively for the Dahla Dam Reservoir (ID10). Although the storage-discharge relationships are statistically significant ($p < 0.001$), the effect size for low flow periods remains modest. However, a preliminary sensitivity analysis, based on 100 low flow time series of reservoir outflows, sampled from the 5/95th confidence intervals of the low flow storage – discharge relationships suggests that this uncertainty in the relationships has only very limited absolute effects on downstream outputs (Figure S4 in the Supplementary Material). Overall, the resulting flows at UHRB$_D$ (ID2) and UARB$_D$ (ID5) could be reproduced with $E_{NS,Q} > 0.79$ and $E_{NS,log(Q)} > 0.81$ for the calibration period and comparable performances during the validation period (Figure 4b, Table 4). The ability of these models to reproduce flow in the upstream regions resulted in a robust representation of flow in the downstream Sistan Plain (SISP; ID8) for the entire validation period 1976-2006 without further calibration (Figure 4d, Table 4). Hydrographs of sub-basins not shown in Figure 4 are provided in Figure S1 S5 in the Supplementary Material.

In general, the estimated water release from the reservoirs results in overall model outputs in all downstream basins are widely consistent with the observed daily river flow, which at station SISP (ID8) is even true for the entire 37-year study period. In spite of all other sources of uncertainty throughout the modelling process, this can be seen as an indication of the plausibility of the modelled reservoir outflow.

It could be observed that annual peak flows in spring are mostly generated by a combination of snowmelt from the high-elevation parts of the HRB, i.e. in sub-basins ID2, 3 and 4, and additional, relatively high-intensity rainfall events (Figure 4). The filling of the two reservoirs attenuates downstream flows, including the annual peaks, throughout spring and into early summer. In turn, the gradual release of water from the reservoirs sustains downstream summer and autumn flows, almost doubling long-term average low flow rates as compared to natural flow conditions (Figures 4, 5), to meet irrigation demand in the downstream Helmand Valley and to satisfy flow requirements of the Sistan River in Iran under the Iranian-Afghan Helmand River Water Treaty (1973).

Furthermore, the models adequately reproduced the losing character of the downstream sub-basins, including LHRB (ID7) and SISP (ID8). Thus, in this highly water-limited environment, these sub-basins do not generate relevant volumes of flow. Rather, most of the precipitation and, in addition, significant volumes of water entering LHRB (ID7) and eventually SISP (ID8) as flow from upstream, eventually evaporate. Besides this, streamflow draining LHRB (ID7) and crossing a hyper-arid desert region is reduced by about 60% before reaching SISP (ID8), as specified by the calibrated loss factor K_L. These streamflow reductions cannot be explained by deep infiltration losses and soil evaporation alone in this essentially vegetation-free environment. There is another even much more plausible source of these observed and modelled flow reductions: when the Helmand River reaches Iran, it bifurcates just upstream the gauge at SISP (ID8) into the Sistan river (SISP, ID8), which drains into the Hamun wetlands, and the completely ungauged Common Parian River, which follows the border between Iran and Afghanistan. The magnitudes of flow diversion are undocumented and merely Burger (2005) in a

study of the Helmand River of Afghanistan and Iran loosely mentioned potential uncertainties arising from this diversion into the Common Parian River. Therefore the lumped loss factor (K_L) combines the effects of deep infiltration (e.g. Schaller and Fan, 2009; Bouaziz et al., 2018; Condon et al., 2020), evaporation and particularly the proportion of water which is diverted into the Common Parian River.

Overall, following a multi-objective calibration strategy, i.e. simultaneously using $E_{NS,Q}$ and $E_{NS,log(Q)}$ as calibration objectives to ensure good representation of both, high- and low flows, our model performances with respect to daily flow in all sub-basins (Table 4, Figure 4), exceed those of the studies of Hajihosseini et al. (2016) but also those of Hajihosseini et al. (2019) who assessed the monthly flow with the SWAT model in the Upper and Lower Helamand basins, respectively.

5.2 Temporal pattern of drought

5.2.1 SPI

Multiple meteorological drought events in terms of SPI occurred in the HRB throughout the 1970-2006 study period (Figure 6a). An average mean drought frequency across all sub-basins of the HRB of $D_{F,SPI}$ = ~~6.3~~2.5 months year[-1] characterized the 1970-1979 decade. This is higher than in the subsequent two decades during which $D_{F,SPI}$ reached ~~5~~0.5 and ~~3.9~~0.3 months year[-1], respectively. The last part of the study period, 2000-2006, experienced more precipitation deficits again, resulting in frequent drought spells with $D_{F,SPI}$ = ~~8.4~~5.4 months year[-1]. A similar pattern was found for drought duration. While the ~~first three~~two decades in the middle of the study period experienced mean drought durations across all sub-basins between $D_{D,SPI}$ = ~~9~~1.2 and 1~~0~~.6 months, much longer droughts occurred in the first and last decades ~~2000-2006 period~~ with $D_{D,SPI}$ = 15.1 and 21.1 months, respectively (Figure 6a). Reflecting the above, the mean drought severity and intensity were also more pronounced at the beginning and towards the end of the study period, with the lowest mean $D_{S,SPI}$ = -0.~~9~~7 and $D_{I,SPI}$ = -1.~~2~~6 month[-1], respectively, in the 2000-2006 period, as compared to the highest $D_{S,SPI}$ = -0.~~2~~1 and $D_{I,SPI}$ = -0.~~3~~8 month[-1] in the wetter period between 1980 and 1999. Notwithstanding the fluctuating pattern in these drought descriptors over the study period, pairwise comparisons of the decadal distributions of basin-average annual SPI values using Wilcoxon rank-sum tests indicated that there is no significant difference between any of the decadal SPI distributions ($p > 0.05$), as also shown in Figure 7a. Correspondingly, no temporal trends in the time series of annual SPI could be detected based on Mann-Kendall tests (Kendall, 1975) for the HRB or any sub-basin therein ($p > 0.05$; Figure 7b). The outputs of the drought analysis with a discretization of the study period into 2 20-year periods resulted in equivalent interpretations (Supplementary Material Figure S6a): in spite of slightly more pronounced $D_{F,SPI}$, $D_{D,SPI}$ and $D_{S,SPI}$ in the 1990-2006 period, the differences to the 1970-1989 period are statistically not significant ($p > 0.05$).

Although these results support the findings of Miyan et al., (2015) who reported that Afghanistan experienced unusual droughts from 1995 onwards until the heavy snow falling in the 2002-2003 winter season, precipitation and the associated meteorological drought did, in spite of decadal fluctuations, not experience a systematic change in the HRB over the four study decades.

5.2.2 SPEI

The temporal pattern of drought in terms of SPEI, reflecting the combined effects of precipitation water supply and atmospheric water demand, similarly indicates the occurrence of multiple periods of severe drought in all sub-basins throughout the HRB during the 1970-2006 study period (Figure 6b). The temporal fluctuations in SPEI broadly correspond with those in SPI, suggesting that most drought events are largely controlled by water supply and thus precipitation deficits rather than by increased atmospheric water demand in this arid region. More specifically, mean drought frequency across all sub-basins decreased from $D_{F,SPEI} = $ ~~6.6~~2.8 months year^{-1} in the 1970-1979 decade to around 0.2 months year^{-1}~~, respectively,~~ in the following two decades. In the last decade of the study period, however, a pronounced increase in drought frequency to $D_{F,SPEI} = $ ~~8.9~~6.1 months year^{-1} was observed (Figure 6b). While individual drought events had average durations between $D_{D,SPEI} = $ ~~8.4~~0.8 and ~~12.9~~2.4 months in the ~~first three~~two middle decades across all sub-basins, this was substantially ~~higher~~~~increased to~~ $D_{D,SPEI} = $ 9.9 months in the first decade and even increased to ~~20.2~~15.8 months in the ~~multi-year~~extreme drought ~~in~~ of the 2000-2006 decade. Drought severity and intensity remained at relatively modest levels not falling below $D_{S,SPEI} = $ -0.~~4~~1 and $D_{I,SPEI} = $ ~~-0.6~~0.8 month^{-1}, respectively, in the 1980-1999 period. In contrast, the first and last decade were characterized by much ~~higher~~ more pronounced severity and intensity, with the lowest mean $D_{S,SPEI} = $ ~~-1.0~~0.8 and $D_{I,SPEI} = $ -1.~~1~~5 month^{-1}, respectively, occurring during the 2000-2006 period. Similar to SPI, Wilcoxon rank-sum tests showed that there is mostly no systematic and significant difference between the decadal distributions of basin-average SPEI ($p > 0.05$), with the exception of the 2000-2006 decade, during which SPEI is significantly lower than during the 1990-1999 decade for most sub-basins ($p \leq 0.05$), as shown in Figure 7c. The temporal sequence of a slight SPEI increase during the first three decades followed by a sharp decrease during the multi-year drought in 2000-2006 likewise illustrates that there is no systematic trend in the time series of SPEI in the HRB or any sub-basin therein over the study period ($p > 0.05$; Figure 7d). The results of the same analysis over two 20-year periods similarly suggest that the 1990-2006 period experienced was slightly more drought affected with somewhat more frequent, longer and more severe droughts as compared to the 1970-1989 period, yet the overall differences are statistically not significant ($p > 0.05$; Supplementary Material Figure S6b).

5.2.3 SDI

Streamflow drought, as specified by SDI, was quantified based on streamflow estimates as obtained from the best available model solution for each of the eight sub-basins. It could be observed that SDI largely follows the temporal pattern in SPI and SPEI (Figure 6c), respectively, with a relatively low lag time of ≤ 1 month in all sub-basins throughout the HRB, as suggested by a cross-correlation analysis between time series of monthly SPI, SPEI and SDI in the individual sub-basins ($r = 0.66 - 0.91$; $p < 0.05$; not shown). However, it can also be observed that, overall, SDI drought events are less pronounced than SPI and SPEI droughts occurring at around the same time. More specifically it was found that the mean drought frequency across all sub-basins fluctuated between $D_{F,SDI} = $ ~~4.7~~0.1 and ~~6.2~~2.0 months year^{-1} in the first three decades of the

study period. In the 2000-2006 decade, it experienced a marked increase to $D_{F,SDI}$ = 9.36.7 months year^{-1}. Similarly, mean drought duration was with $D_{D,SDI}$ = 30.321.0 months highest in that decade. In the other these decades the mean $D_{D,SDI}$ did not exceed 16 11.3 months. Closely reflecting the pattern of SPI and SPEI, mean drought severity and intensity across all sub-basins were most pronounced in the first and last decades with both $D_{S,SDI}$ and $D_{I,SDI}$ reaching minimum values of -0.91.0 and -1.5 month^{-1}, respectively, in the 2000-2006 period. During the wetter decades in-between $D_{S,SDI}$ and $D_{I,SDI}$ did not decrease below values of -0.2 1 and -0.5 9 month^{-1}, respectively.

Following a pairwise comparison of all decadal basin-average SDI distributions, the slight, yet statistically insignificant increase of the decadal SPI and SPEI distributions (p > 0.05; Figures 7a,c) from 1970-1999 could not be observed in SDI, which remained rather stable during the first three decades (Figure 7e). In contrast, the decrease of SPEI in the last decade is reflected in correspondingly lower basin-average SDI in the 2000-2006 period (p ≤ 0.05; Figure 7e) (Li et al., 2019; Noorisameleh et al., 2019). However, the time series of basin-average SDI did not exhibit a significant trend (p > 0.05; Figure 7f). In contrast, comparison of the two 20-year periods (1970-1989 and 1990-2006) suggests higher drought frequency, longer duration and more pronounced severity and intensity, respectively, as compared to the 1970-1989 period. Based on a Wilcoxon rank-sum test, there is a systematic and significant difference between the basin-average SDI distributions for the two 20-year periods (p > 0.05, Supplementary Material Figure S6c), indicating a shift from mild to severe hydrological drought in the study area.

5.3 Spatial pattern, synchronicity and propagation of drought

5.3.1 SPI

In most years of the study period meteorological drought, as specified by SPI, exhibits considerable spatial coherence and synchronicity throughout the HRB (Figure 6a). In other words, at any given time, the entire HRB experiences similar relative precipitation deficits (or surpluses), with a median r = 0.97 (p < 0.05) as obtained from a Spearman rank correlation between the time series of SPI across all sub-basins. Regional differences in SPI remain limited to parts in the central HRB, i.e. CHRB (ID3) and LARB (ID6, Figure 6a). In contrast to the remainder of the HRB, these two sub-basins are characterized by multiple periods that are, in relative terms, more humid, such as in 1974 or 1982, but also by periods that are, in comparison, considerably drier, such as 1987 or 1994. The elevated degree of spatial coherence and synchronicity in SPI on the scale of the HRB is further illustrated by the comparison of the upstream and downstream decadal SPI distributions (Figure 8a). No significant differences (p > 0.05) between the SPI distribution of the six most upstream sub-basins (ID1-ID6) and the SPI distribution of the two most downstream sub-basins, LHRB (ID7) and SISP (ID8), could be found in any of the four decades during the study period. To provide some more explicit spatial context, the spatial distribution of SPI at the resolution of the individual model grid cells for four selected years is shown in Figure 9a-d. Compared to the SPI aggregated at the scales of the individual sub-basins (Figure 6a), this more detailed picture corroborates the level of large-scale spatial coherence, in

spite of somewhat increased local variations in SPI (Figure 9a-d). A rather rare exception is the year 1987, which was characterized by a substantial North-South gradient in SPI spatial variations and whose extent is largely masked by the aggregation of SPI to the sub-basin scale in Figure 6a.

5.3.2 SPEI

While SPEI is widely coherent (median r = 0.94, p < 0.05) and spatially broadly follows the pattern of SPI throughout large parts of the HRB, it can also be observed that inter-annual differences in atmospheric water demand, here estimated based on E_P, lead to modest, yet contrasting effects (Figure 6b). For some sub-basins and time periods characterized by comparably cool temperatures, water deficits are attenuated and SPEI thus remains higher than SPI (e.g. $UARB_U$-ID4 in 1986 or LARB-ID6 in 1989). For other sub-basins and warmer periods, increased atmospheric water demand reinforces water deficits (e.g. CHRB-ID3 in 1981). As shown in Figure 8a, the distributions of SPEI closely reflect the distributions of SPI in the first decade of the study period. In the following 1980-1989 decade as well as in the 2000-2006 decade SPEI is lower than SPI, potentially indicating the role of E_P in intensifying water deficits in these periods. In contrast, the opposite effect can be observed during the 1990-1999 decade, where rather low E_P had a moderating effect, leading to higher values of SPEI than SPI. Although these effects occur across the entire HRB, water deficits in terms of SPEI are considerably more sensitive to fluctuations in atmospheric water demand and the differences between SPEI and SPI are thus more pronounced in the downstream parts of the HRB (Figure 8a). In particular, SPEI in the hyper-arid SISP (ID8; Figure 6b) is characterized by a low degree of coherence and synchronicity compared to upstream SPEI, exhibiting both, markedly more severe water deficits (e.g. 1973, 1984 or 2003) but also more pronounced water surpluses (e.g. 1986, 1996 or 2005). Notwithstanding these varying effects of E_P on water deficits and thus on the differences between SPEI and SPI, no systematic temporal trend of E_P reinforcing/moderating water deficits could be detected. However, note that the applicability of SPEI in arid areas such as the study region may be limited (Pei et al., 2020). In such environments such fluctuations in E_P will have a limited effect on E_A and thus on water deficits as the systems are, by definition, water limited rather than energy limited. Changes in E_P will therefore be less relevant for the intensification/moderation of drought in such arid regions than changes in precipitation.

5.3.3 SDI

Hydrological drought in terms of SDI during the study period exhibited a lower degree of spatial coherence and synchronicity (Figure 6c) across the HRB. This is reflected by a lower median r = 0.85 (p < 0.05) from pairwise Spearman Rank correlations between the individual time series of SDI across all sub-basins. The spatially and temporally more heterogeneous mosaic of SDI, however, allows a few insights. The data suggest that both reservoirs, at Kajakai dam and Dahla dam, respectively, have effects on the propagation of hydrological drought. This can be seen in the differences in SDI between the sub-basins upstream ($UHRB_U$-ID1; $UARB_U$-ID4) and the associated sub-basins downstream of the dams ($UHRB_D$-ID2; $UARB_D$-ID5) in Figure 6c. In the early phase of the study period, the reservoirs had some moderating effects on the propagation of hydrological droughts, most notably for the 1977 (both dams) and 1971 (Dahla dam) droughts. The

median SDI in the 1970-1979 decade was ~0.2 higher downstream than upstream of both reservoirs ($p < 0.05$). However, over the following decades, both reservoirs largely lost their drought attenuating functions and the reservoir at Dahla dam may have even contributed to amplifying the 1999-2002 drought downstream of the dam, with a median SDI over that period being ~0.18 ($p < 0.05$) lower at the downstream $UARB_D$ (ID5) than the upstream $UARB_U$ (ID4).

While the distribution of SDI broadly follows the distributions of SPI and SPEI in the upstream part of the HRB (ID1-ID6), downstream hydrological drought is characterized by rather distinct dynamics (ID7-ID8; Figure 8a). In contrast to the basin-average time series of SDI (Figure 7f), SDI in the two downstream sub-basins exhibit clear negative trends over the four decades of the study period ($p \leq 0.05$; not shown). In addition, the data suggest that for the 1970-1979 decade the median downstream SDI ~ 0.2 is significantly higher ($p < 0.05$) than SPI, SPEI and upstream SDI, which are all characterized by a median of about -0.1 (Figure 8a). As also shown by the individual SDI distributions of all sub-basins in Figure 8b, hydrological drought is considerably attenuated and the relative river water deficits reduced compared to upstream parts of the HRB during that period. However, throughout the following two decades, the downstream moderation of hydrological drought weakens, i.e. the distributions of downstream SDI more closely reflect those of SPI, SPEI and upstream SDI (Figure 8b). This pattern then eventually fully inverts into a downstream drought amplification in the 2000-2006 decade, during which the median downstream SDI = -1.5 is significantly lower ($p < 0.05$) than not only the median upstream SDI = -0.9 but also than SPI and SPEI (Figures 8a,b). This shift from downstream drought moderation to drought amplification can be seen clearly for the four selected years in Figure 9e-h. In spite of dry meteorological conditions throughout the HRB in 1977 (Figure 9a) and severe hydrological drought in the North of the HRB, no drought occurred in the South of the study region (Figure 9e). In 1987, similarly, the increasing precipitation deficits from upstream to downstream (Figure 9b) were buffered and not reflected in the North-South gradient of SDI, indicating the wettest conditions in the most downstream part of the HRB (Figure 9f). The extreme opposite of the above two examples occurred in the last decade of the study period. In both years, 2002 and 2003, respectively, spatially relatively coherent precipitation pattern across the entire HRB (Figure 9c-d) led to severe hydrological drought in the most downstream parts of the HRB in particular at SISP (ID8; Figure 9g-h). This is particularly striking for the rather wet year 2003, in which SDI in the upstream sub-basins reflected the generally wet conditions of that year, while further downstream river water deficits developed, gradually amplifying to severe drought at SISP (ID8). Further analysis of the time series of the difference between upstream (ID1-ID6) and downstream (ID7-8) SDI (i.e. ΔSDI) shows the inversion from a negative to a positive ΔSDI over the 37 years of the study period occurred gradually and, according to a Mann-Kendall test, following a significant trend ($p < 0.05$), while the differences in SPI remain stable over time (Figure 8c) (Ma et al., 2019). This suggests that it may not be implausible to assume that the inversion of downstream hydrological drought moderation in the 1970-1979 decade into drought amplification in the 2000-2006 decade was, at least partly, an effect of systematic, longer-term shifts in the system rather than a short-term, synchronous occurrence of multiple drought-amplifying hydro-meteorological conditions, such as sustained high precipitation deficits and high atmospheric water demand (Van Loon, 2015). Such short-term influences of deficits in hydrological drivers would be likely

to manifest themselves in the evolution of ΔSDI characterized by a more erratic temporal pattern. ~~Such short-term influences would be likely to manifest themselves in a more erratic temporal evolution of ΔSDI.~~

5.4 Drought drivers and process attribution

The above drought indices provide only limited information to identify dominant drivers of droughts. To gain more understanding of the spatio-temporal pattern of hydrological drought and to eventually attribute droughts to physical processes estimate~~s~~ of the absolute magnitudes of multiple modelled hydrological fluxes, as obtained from the best available model solution for each sub-basin, are in the following analysed.

With a long-term mean annual precipitation of ~250 mm y^{-1} in the HRB, the overall magnitudes of streamflow deficits, and thus of hydrological droughts, are clearly dominated by fluctuations in precipitation anomalies (Figure 10a), with a mean absolute anomaly of around ±50mm y^{-1} for the entire HRB or ~ 20% of the long-term mean water balance. In contrast, anomalies in total evaporation E_A (here: $E_A = E_I + E_T + I_D$) exhibit much lower variability in this arid environment, with a mean absolute anomaly of about ±20 mm y^{-1}. As water supply is the limiting factor for evaporation, the highest rates of E_A occur in the wettest years (Figure 10b). Conversely, E_A has a proportionally less impact on streamflow in dry years. In general it can be seen that precipitation anomalies of ~ -50 – -100 mm y^{-1} lead to streamflow anomalies of ~ -20 – -30 mm y^{-1} (Figure 10c). Although SPEI is typically based on potential evaporation E_P, arid and thus water-limited environments are rather insensitive to fluctuations in E_P compared to fluctuations in P. In other words, there will be little difference in the partitioning of water fluxes if under the same annual precipitation of e.g. 500 mm yr^{-1}, E_P is 1000 or 1500 mm yr^{-1}, as in both cases actual evaporation E_A will be close to (or even exceed) 400 mm yr^{-1} and as therefore most of the available water will be evaporated. In contrast, more water will be evaporated as E_A (even if E_P remains stable) in years when more water is available and thus P is higher. By extension, the effects of evaporation on droughts in arid regions can only be meaningfully assessed by changes in E_A. Haung et al., (2017) mentioned that actual evaporation affected strongly the propagation time of meteorological to hydrological drought in the Wei River Basin (WRB), a typical arid and semi-arid region in China.

The modelled data suggest that during drought years, the reservoir at Kajakai Dam released slightly less water (UHRB$_D$-ID2) than it received as an inflow (UHRB$_U$-ID1), as shown in Figure 4. The mean difference between drought period inflow to and outflow from the reservoir remained stable at ΔQ ~ 0.9 mm y^{-1} throughout the four decades of the study period. This implies that there is no evidence that the reservoir neither moderated nor significantly amplified downstream propagation of streamflow deficits, underlining the very minor role of this reservoir for the drought pattern. In contrast, the modelled flow estimates for the reservoir at Dahla Dam suggest that this reservoir had some moderation effect on downstream flow deficits and thus drought propagation in the first decade of the study period (Wang et al., 2019). On average, the reservoir outflow (UARB$_D$-ID5) during drought periods in that decade exceeded the inflow (UARB$_U$-ID4) by ΔQ ~ 1.1 mm y^{-1} (Supplementary Figure S1). However, this difference gradually decreased over time and eventually converged towards zero in the 2000-2006 period. In spite of uncertainties arising from data and the modelling process, this nevertheless indicates the possibility that the Dahla Dam reservoir has lost its, albeit very minor, drought-moderating function over the study period.

For further analysis the HRB was separated into an upper and a lower basin. The upper basin comprises $UHRB_D$ (ID2), CHRB (ID3) and LARB (ID3), which together drain into the lower basin, here defined as LHRB (ID7) only and thus for clarity of presentation excluding SISP (ID8). As illustrated by Figure 10, and consistent with the spatial analysis of drought indices in Section 5.3, the general pattern of anomalies corresponds well between the upper and the lower basin, suggesting a considerable level of spatial coherence and drought synchronicity. However, reflecting the evolution of ΔSDI (Figure 8c), a subtle but gradual shift in the difference between streamflow anomalies of the upper and lower basins from, on average, -9.4 mm y^{-1} in the 1970-1979 decade to 5.5 mm y^{-1} in the 2000-2006 period is evident (Figure 10c). Thus, while anomalies were less negative/more positive, therefore indicating proportionally "more" water, in the lower than in the upper basin at the beginning of the study period, the opposite was true at the end of the study period. This entails that in the first decade of the study period streamflow deficits from the upper basin were to some degree attenuated in the lower basin. This effect was gradually reduced and finally completely inversed in the last decade of the study period. During the 2000-2006 period streamflow anomalies from the upper basin were systematically amplified in the lower basin. The absence of a similar systematic shift in the difference of precipitation anomalies between the upper and the lower basin (Figure 10a) strongly suggests alternative reasons for the above effects. Mianabadi et al. (2020) similarly indicated that in the lower Helmand River Basin water availability issues cannot be attributed to the changes in precipitation in the downstream area itself.

The analysis of the relative contributions of different water fluxes from the upper and lower basins, respectively, as well as their evolution over time as estimated from the models allowed some more detailed insights into these patterns. The combined water balance of all three sub-basins of the upper basin for the 1970-1979 period (Figure 11a) shows that of the mean annual precipitation P ~ 202 mm y^{-1} of the upper basin, 28% drained away as streamflow (Q ~ 56 mm y^{-1}) and the remainder of 72% was released as combined evaporative fluxes (E_A ~ 146 mm y^{-1}). While transpiration (E_T ~ 130 mm y^{-1}) and interception evaporation (E_I ~ 9 mm y^{-1}) played a role throughout the entire upper basin, irrigation demand (I_D ~ 7mm y^{-1}) was limited to the agriculturally used areas of the LARB (ID6) sub-basin and thus only accounted for ~ 4% of the water balance of the upper basin. The flow partitioning of the lower basin for the same period exhibited a considerably different pattern. It can be seen in Figure 11a that of the available water in the lower basin (~ 97 mm y^{-1}), i.e. precipitation over the LHRB (ID7) sub-basin plus the combined inflow from the upper basin, 51% (Q ~ 49 mm y^{-1}) is drained as streamflow and 49% are released as evaporative flux (E_A ~48 mm y^{-1}). In comparison to the upper basin, irrigation demand in the lower basin is with I_D ~ 13 mm y^{-1} a substantially larger fraction of the water balance (~ 14%) than in the upper basin. ~~irrigation demand accounts with ~ 14% for a substantially larger fraction of the water balance in the lower basin (I_D ~ 13 mm y^{-1}).~~

During the 2000-2006 period (Figure 11b), the upper basin received slightly less precipitation (P ~ 179 mm y^{-1}) than in the 1970-1979 period. However, the relative contributions of the different fluxes remained rather stable over time. The fraction of water drained as streamflow slightly decreased to 25% (Q ~ 44 mm y^{-1}), while the fraction of evaporative fluxes

correspondingly increased to 75 % (E_A ~ 135 mm y^{-1}) of the water balance of the upper basin with similar increases for all three evaporative components (Figure 11b). In contrast, substantial shifts in the flux partitioning can be observed for the lower basin (Figure 11b). In spite of a reduction of available water to ~ 71 mm y^{-1} in the 2000-2006 period, the evaporative release (E_A ~ 49 mm y^{-1}) reached the same level as in the 1970-1979 decade. As illustrated by Figure 11b, the high levels of evaporative release were sustained by significant absolute and relative increases of irrigation demand to I_D ~ 23 mm y^{-1} or 32% of the water available in the lower basin (or ~10% of the water balance of the entire HRB). This, in turn, resulted in a reduction of streamflow to Q ~ 22 mm y^{-1}, equivalent to a reduction from 51% of the water balance in the 1970-1979 decade to 31% in the 2000-2006 decade. The increases of I_D and the corresponding decreases in Q are directly related to increases in the agricultural area over the study period (Figure 2). It is therefore plausible to assume that the inversion of the function of the lower basin from moderation to amplification of flow deficits and the associated droughts is largely a consequence of increases in an agriculturally used area which resulted in increases of the related irrigation water demand (AghaKouchak et al., 2015; Van Loon et al., 2016; Haile et al., 2019). In addition, inefficient irrigation schemes in the study region may lead to an underestimation of the actual irrigation water use. Therefore higher real-world irrigation water demand would even further strengthen the results, that the shift form of downstream moderation to the intensification of hydrological drought over the study period is largely an effect of human intervention. Margariti et al. (2019) found that human activities prolonged drought durations in all the European catchments they analysed. Ma et al., (2019) showed that human inventions are likely to have changed the positive correlation between meteorological and hydrological droughts to negative in the semiarid Heile River Basin, especially during warm and irrigation seasons.

Overall, the magnitudes of flow deficits and the associated hydrological droughts are largely driven by precipitation deficits across the HRB. The two reservoirs in the HRB had a very minor effect on the propagation of flow deficits, with levels not exceeding 0.5% of the water balance of the HRB in the study period. Burger (2005) stated that the extreme drought in the Sistan region between 1999 and 2004 was not caused by the Afghan reservoirs, but rather this period was extremely dry in the whole catchment but in the future, the drought might be worsened due to extend the Kajakai reservoir. In contrast, the increase of agricultural area, mostly in LHRB (ID7), led to an increase of the basin-wide irrigation water demand (i.e. from LARB-ID6 and LHRB-ID7) from ~ 7% to ~ 12% of the water balance of the HRB. While at the scale of the entire HRB this remains of minor relevance for flow deficits, and thus hydrological drought, it led throughout the study period to a continuous and gradual change in the downstream propagation of flow deficits from moderation to amplification. This clearly underlines the argument by Haile et al. (2019), indicating that human activities such as expansion of cultivation, overexploitation of water resources, particularly for irrigation demands have an impact on altering the hydrological processes which are directly linked to drought. Our results further illustrate that flow deficits and droughts in the HRB clearly reflect the dynamic interplay between temporally varying regional differences in hydro-meteorological variables together with subtle and temporally varying effects linked to direct human intervention (Jahanzaib et al., 2020; Jiang et al., 2019; Saeidi et al., 2018, Wan et al., 2017).

5.5 Uncertainties, unresolved questions and limitations

All the above results are necessarily conditional on a range of uncertainties and choices made during the modelling process (Beven, 2006; Hrachowitz and Clark, 2017). This is in particular relevant in the HRB, where detailed and reliable data are scarce. It entails further, that although the results of this study are largely consistent with the available data, the data themselves may inaccurately reflect reality. In addition, whereof sufficient quality, the available data may not have sufficient detail to accurately represent the underlying mechanistic processes and/or changes thereof over time in a model. Two major sources of uncertainty, due to the lack of detailed and high-quality data need to be explicitly highlighted for this study. First, the routing of flows through the two reservoirs in the HRB was estimated with a simple empirical relationship (Eq.2) based on data from the 1970-1979 period, under the assumption that this relationship did not change over time. In reality, reservoir operation rules may have changed over the study period, yet this cannot be clarified with the available data. However, even if such changes occurred, their impact is likely limited, as model evaluation at SISP (ID8) showed that adequate model performances were achieved throughout the entire study period (Table 3, Figure 4).

A second not completely resolved issue is the observed and modelled considerable reduction of stream flow between LHRB (ID7) and SISP (ID8). The loss of ~ 60% of streamflow as the river crosses the desert region between Afghanistan and Iran can plausibly be explained by the combined effects of evaporation, deep infiltration losses and, most importantly, river water diversion into the completely ungauged Common Parian River that bifurcates from the Helamnad River just upstream of SISP (ID8). In the model it was represented by an unspecified loss factor K_L. A clearer mechanistic interpretation was not warranted by the available data. Another cause that cannot be completely ruled out is a potentially low quality of the available streamflow data either at LHRB (ID7), at SISP (ID8), or both of them.

We explicitly reiterate here that although this modelling study allowed robust insights into the pattern of drought characteristics, including changes in downstream drought propagation over time, the absolute magnitudes of variables reported herein remain, for the above reasons, highly uncertain. These magnitudes should therefore, under no circumstances and without more detailed data and understanding of the underlying processes, be used for direct policy advice in this arid environment where the transboundary nature of the HRB makes water scarcity a highly sensitive issue.

Clearly, the most reliable way forward to reduce uncertainties in flow estimations in the the HRB is to do more observations and generate data, which can then be confronted with the model. Deficiencies in the model to reproduce these additional data will then, in an iterative process, allow model improvement (e.g. Fenicia et al., 2008; Hulsman et al., 2020). However, at this stage, further model improvement is problematic as the introduction of more complexity in the model will not be warranted by the available data and eventually merely lead to increased equifinality (Beven, 2006). Although beyond the scope of this study, a future comparison of alternative data sets to inform the model and an analysis of the associated potential differences in model results will be highly valuable to ensure reliable interpretations of the HRB and thus to limit uncertainty for better policy advice on water resource management in the HRB.

6 Conclusions

In combined data analysis and modelling study in the transboundary Helmand River Basin (HRB) we analysed spatial patterns of drought and changes therein over the 1970-2006 study period, based on the drought indices SPI, SPEI and SDI, as well as on absolute water deficits. The results provide some evidence that:

(1) Drought characteristics varied throughout the study period. In general, the 2000-2006 and partly the 1970-1979 periods were drier than the decades in between. Depending on the drought index, mean drought duration reached D_D ~ ~~20~~ 10 – ~~3~~20 months and mean drought intensity D_I ~ -1.~~0~~5 month^{-1} in these drier periods, as compared to D_D ~ ~~8~~0 – ~~16~~2 months and D_I ~ -0.~~3~~1 – ~~-0.6~~1.0 in the 1980-1999 period.

(2) The basin-average decadal distributions of the drought indices largely exhibited no statistically significant differences, with the exception of significantly lower SPEI and SDI in 2000-2006 compared to the preceding decades. In addition, no systematic trend over time was detected for any of the basin-average drought indices.

(3) All three drought indices exhibit considerable spatial coherence and synchronicity across the HRB throughout the study period. This indicates that in most cases droughts similarly affect the entire HRB with little regional or local differences.

(4) The overall magnitudes of streamflow drought in the HRB are consistently controlled by precipitation deficits, while the effects of the two reservoirs, as well as water abstraction for irrigation on flow deficits, remain minor during drought years, accounting for only 0.5 % and ~10%, respectively, of the water balance of the HRB.

(5) The downstream parts of the HRB moderated the further propagation of streamflow deficits and the associated droughts in the early decades of the study period. This drought moderation function of the lower basin was gradually and systematically inverted by the end of the study period, when the lower basin eventually amplified the downstream propagation of flow deficits and droughts.

(6) The shift from drought moderation to drought amplification in the lower basin is very likely a consequence of agricultural activity and the associated increased irrigation water demand in spite of being only a minor fraction of the water balance.

Overall the results of this study illustrate that the flow deficits and associated droughts in the HRB clearly reflect the dynamic interplay between temporally varying regional differences in hydro-meteorological variables together with subtle and temporally varying effects linked to direct human intervention.

References

AghaKouchak, A., Feldman, D., Hoerling, M., Huxman, T., and Lund, J.: Water and climate: Recognize anthropogenic drought, Nature News, 524, 409–411, 2015.

Ahmad, M. and Wasiq, M.: Water resources development in Northern Afghanistan and its implications for Amu Darya Basin (English), World Bank working paper series, Washington, D.C., World Bank Group. 36, 66 pp., 2004.

Ahmadalipour, A., Moradkhani, H., and Demirel, M. C.: A comparative assessment of projected meteorological and hydrological droughts: Elucidating the role of temperature, J. Hydrol., 553, 785-797, https://doi.org/10.1016/j.jhydrol.2017.08.047, 2017.

Ajami, N. K., Gupta, H., Wagener, T., and Sorooshian, S.: Calibration of a semi-distributed hydrologic model for streamflow estimation along a river system, J. Hydrol., 298, 112–135, https://doi.org/10.1016/j.jhydrol.2004.03.033, 2004.

Alami, M. M. and Tayfor, G.: Meteorological Drought Analysis by Different Methods in Helmand River Basin, Afghanistan, Int. J. Sci. Eng. Res., 8, 738-744, https://doi.org/10.14299/ijser.2018.02.004, 2018.

Al-Faraj, F. A. M. and Scholz, M.: Assessment of temporal hydrologic anomalies coupled with drought impact for a transboundary river flow regime: The Diyala watershed case study, J. Hydrol., 517, 64-73, https://doi.org/10.1016/j.jhydrol.2014.05.021, 2014.

Allen, R., Pereira, L., Raes, D., and Smith, M.: Crop evapotranspiration - Guidelines for computing crop water requirements, FAO Irrigation and drainage paper No. 56, 300 pp., 1998.

Arnold, J., Srinivasan, R., Muttiah, R., and Williams, J.: Large area hydrologic modeling and assessment part I: model development, J. Am. Water Resour. Assoc., 34, 73–89, https://doi.org/10.1111/j.1752-1688.1998.tb05961.x, 1998.

Ashraf Vaghefi, S., Abbaspour, N., Kamali, B., and Abbaspour, K. C.: A toolkit for climate change analysis and pattern recognition for extreme weather conditions – Case study: California-Baja California Peninsula, Environ. Modell. Software, 96, 181-198, https://doi.org/10.1016/j.envsoft.2017.06.033, 2017.

Beven, K. J.: A manifesto for the equifinality thesis, J. Hydrol., 320, 18–36, https://doi.org/10.1016/j.jhydrol.2005.07.007, 2006.

Beven, K.: On hypothesis testing in hydrology, Hydrol. Processes, 15, 1655-1657, https://doi.org/10.1002/hyp.436, 2001.

Bouaziz, L., Weerts, A., Schellekens, J., Sprokkereef, E., Stam, J., Savenije, H., and Hrachowitz, M.: Redressing the balance: quantifying net intercatchment groundwater flows, Hydrol. Earth Syst. Sci., 22, 6415–6434, https://doi.org/10.5194/hess-22-6415-2018, 2018.

Burger, A. S.: Calibration of a precipitation runoff model for the Hirmand River in Afghanistan and Iran using remote sensing data, master thesis, Delft University of Technology, Faculty of civil engineering and Geosciences, 77p, 2005.

Coerver, H. M., Rutten, M. M., and van de Giesen, N, C.: Deduction of reservoir operating rules for application in global hydrological models, Hydrol. Earth Syst. Sci., 22, 831–851, https://doi.org/10.5194/hess-22-831-2018, 2018.

Condon, L. E., Atchley, A. L. and Maxwell, R. M.: Evapotranspiration depletes groundwater under warming over the contiguous United States. Nat. Commun. 11, 873. https://doi.org/10.1038/s41467-020-14688-0, 2020.

Ebrahimzadeh, I. and Esmaelnejad, M.: Climate change and the role of recent droughts on agricultural economy of Sistan, Romanian Review of Regional Studies., IX, 11-22, 2013.

Edwards, D.C. and McKee, T.B.: Characteristics of 20th Century Drought in the United States at Multiple Time Scales, Department of Atmospheric Science, Colorado State University, Fort Collins, Climatology Report 97-2, 1997.

Euser, T., Hrachowitz, M., Winsemius, H., and Savenije, H.: The effect of forcing and landscape distribution on performance and consistency of model structures, Hydrol. Processes, 29, 3727–3743, https://doi.org/10.1002/hyp.10445, 2015.

Euser, T., Winsemius, H. C., Hrachowitz, M., Fenicia, F., Uhlenbrook, S., and Savenije, H. H. G.: A framework to assess the realism of model structures using hydrological signatures, Hydrol. Earth Syst. Sci., 17, 1893–1912, https://doi.org/10.5194/hess-17-1893-2013, 2013.

Fenicia, F., Kavetski, D., and Savenije, H. H. G.: Elements of a flexible approach for conceptual hydrological modeling: 1. Motivation and theoretical development, Water Resour. Res., 47, 1-13, https://doi.org/10.1029/2010WR010174, 2011.

Fenicia, F., Savenije, H. H. G., Matgen, P., and Pfister, L.: A comparison of alternative multiobjective calibration strategies for hydrological modeling, Water Resour. Res., 43, 1-16, https://doi.org/10.1029/2006WR005098, 2007.

Fenicia, F., Savenije, H. H. G., Matgen, P., and Pfister, L.: Is the groundwater reservoir linear? Learning from data in hydrological modelling, Hydrol. Earth Syst. Sci., 10, 139–150, doi:10.5194/hess-10-139-2006, 2006.

Fenicia, F., Savenije, H. H. G., Matgen, P., and Pfister, L.: Understanding catchment behaviour through stepwise model concept improvement, Water Resour. Res., 44, 1-13, https://doi.org/10.1029/2006WR005563, 2008.

Freer, J., Beven, K., and Ambroise, B.: Bayesian Estimation of Uncertainty in Runoff Prediction and the Value of Data: An Application of the GLUE Approach, Water Resour. Res., 32, 2161-2173, https://doi.org/10.1029/95WR03723, 1996.

Gao, H., Ding, Y., Zhao, Q., Hrachowitz, M., and Savenije, H. H. G.: The importance of aspect for modelling the hydrological response in a glacier catchment in Central Asia, Hydrol. Processes, 31, 2842–2859, https://doi.org/10.1002/hyp.11224, 2017.

Gao, H., Hrachowitz, M., Fenicia, F., Gharari, S., and Savenije, H. H. G.: Testing the realism of a topography-driven model (FLEX-Topo) in the nested catchments of the Upper Heihe, China, Hydrol. Earth Syst. Sci., 18, 1895–1915, https://doi.org/10.5194/hess-18-1895-2014, 2014.

Geng, S., Devries, F., and Supit, I.: A simple method for generating daily rainfall data, Agr. Forest Meteorol., 36, 363-376, https://doi.org/10.1016/0168-1923(86)90014-6, 1986.

Gharari, S., Hrachowitz, M., Fenicia, F., and Savenije, H. H. G.: An approach to identify time consistent model parameters: sub-period calibration, Hydrol. Earth Syst. Sci., 17, 149–161, https://doi.org/10.5194/hess-17-149-2013, 2013.

Gharari, S., Shafiei, M., Hrachowitz, M., Kumar, R., Fenicia, F., Gupta, H. V., and Savenije, H. H. G.: A constraint-based search algorithm for parameter identification of environmental models, Hydrol. Earth Syst. Sci., 18, 4861–4870, https://doi.org/10.5194/hess-18-4861-2014, 2014.

Gocic, M. and Trajkovic, S.: Analysis of precipitation and drought data in Serbia over the period 1980–2010, J. Hydrol., 494, 32-42, https://doi.org/10.1016/j.jhydrol.2013.04.044, 2013.

Goes, B. J. M., Howarth, S. E., Wardlaw, R. B., Hancock, I. R., and Parajuli, U. N.: Integrated water resources management in an insecure river basin: a case study of Helmand River Basin, Afghanistan, Int. J. Water Resour. Dev., 32, 3-25, http://dx.doi.org/10.1080/07900627.2015.1012661, 2016.

Gupta, H. V., Sorooshia, S., and Yapo, P. O.: Toward improved calibration of hydrologic models: Multiple and noncommensurable measures of information, Water Resour. Res., 34, 751-763, https://doi.org/10.1029/97WR03495, 1998.

Haddeland, I., Heike, J., Biemans, H., Eisner, S., Flörke, M., Kanasaki, N., Konzmann, M., Ludwig, F., Masaki, Y., Schewe, J., Stacke, t., Tessler, Z. D., Wada, Y., and Wisser, D.: Global water resources affected by human interventions and climate change, PNAS, 111, 3251-3256, https://doi.org/10.1073/pnas.1222475110, 2014.

Haile, G. G., Tang, Q., Sun, S., Huang, Z., Zhang, X., & Liu, X.: Droughts in East Africa: Causes, impacts and resilience. Earth-Science Reviews, 193, 146–161, 2019.

Haile, G. G., Tang, Q.,cLi, W., Liu, X., and Zhang, X.: Drought: Progress in broadening its understanding, Wiley Interdisciplinary Reviews: Water, 7, e1407, https://doi.org/10.1002/wat2.1407, 2020.

Hajihosseini, H., Hajihosseini, M., Morid, S., Delavar, M., and Booij, M. J.: Hydrological Assessment of the 1973 Treatyon the Transboundary Helmand River, Using the SWAT Model and a Global Climate Database, Water Resour. Manag., 30, 4681–4694, https://doi.org/10.1007/s11269-016-1447-y, 2016.

Hajihosseini, M., Hajihosseini, H., Morid, S., Delavar, M., and Booij, M. J.: Impacts of land use changes and climate variability on transboundary Hirmand River using SWAT, J. Water Clim. Change [Preprint], https://doi.org/10.2166/wcc.2019.100, 18 October 2019.

Hanasaki, N., Kanae, S., and Oki, T.: A reservoir operation scheme for global river routing models, J. Hydrol., 327, 22-41, https://doi.org/10.1016/j.jhydrol.2005.11.011, 2006.

Harris, I., Jones, P. D. Osborn, T. J., and Lister, D. H.: Updated high-resolution grids of monthly climatic observations – the CRU TS3.10 Dataset, Int. J. Climatol., 34, 623-642, https://doi.org/10.1002/joc.3711, 2014.

Hrachowitz, M. and Clark, M. P.: HESS Opinions: The complementary merits of competing modelling philosophies in hydrology, Hydrol. Earth Syst. Sci., 21, 3953–3973, https://doi.org/10.5194/hess-21-3953-2017, 2017.

Hrachowitz, M., Fovet, O., Ruiz, L., Euser, T., Gharari, S., Nijzink, R., Freer, J., Sanenije, H. H. G., and Gascuel-Odoux, C.: Process consistency in models: The importance of system signatures, expert knowledge, and process complexity, Water Resour. Res., 50, 7445–7469, https://doi.org/10.1002/2014WR015484, 2014.

Huang, S., Huang, Q., Chang, J., and Leng, G.: Linkages between hydrological drought, climate indices and human activities: a case study in the Columbia River basin, Int. J. Climatol., 36, 280-290, https://doi.org/10.1002/joc.4344, 2016.

Hulsman, P., Savenije, H. H. G., and Hrachowitz, M.: Learning from satellite observations: increased understanding of catchment processes through stepwise model improvement, Hydrol. Earth Syst. Sci. Discuss. [preprint], https://doi.org/10.5194/hess-2020-191, in review, 2020.

Jehanzaib, M., Ali Shah, S., Yoo, J., and Kim, T.: Investigating the impacts of climate change and human activities on hydrological drought using non-stationary approaches, J. Hydrol., 588, 125052, https://doi.org/10.1016/j.jhydrol.2020.125052, 2020.

Jiang, S., Wang, M., Ren, L., Xu, C., Yuan, F., Liu, Y., Lu, Y., and Shen, H.: A framework for quantifying the impacts of climate change and human activities on hydrological drought in a semiarid basin of Northern China, Hydrol. Processes, 33, 1075-1088, https://doi.org/10.1002/hyp.13386, 2019.

Jiao, Y. and Yuan, X.: More severe hydrological drought events emerge at different warming levels over the Wudinghe watershed in northern China, Hydrol. Earth Syst. Sci., 23, 621–635, https://doi.org/10.5194/hess-23-621-2019, 2019.

Konz, M. and Seibert, J.: On the value of glacier mass balances for hydrological model calibration, J. Hydrol., 385, 238-246, https://doi.org/10.1016/j.jhydrol.2010.02.025, 2010.

Kubiak-Wójcicka, K. and Bąk, B.: Monitoring of meteorological and hydrological droughts in the Vistula basin (Poland), Environ. Monit. Assess., 190, 691, https://doi.org/10.1007/s10661-018-7058-8, 2018.

Li, Y., Wang, Z., Zhang, Y., Li, X., and Huang, W.: Drought variability at various timescales over Yunnan Province, China: 1961–2015. Theor Appl Climatol 138, 743–757, https://doi.org/10.1007/s00704-019-02859-z, 2019.

LIoyd-Hughes, B. and Saunders, M. A.: A drought climatology for Europe, Int. J. Climatol., 22, 1571-1592, https://doi.org/10.1002/joc.846, 2002.

Liu, Y. Ren, L., Zhu, Y., Yang, X., Yuan, F., Jiang, S., and Ma, M.: Evolution of Hydrological Drought in Human Disturbed Areas: A Case Study in the Laohahe Catchment, Northern China, Adv. Meteorol., 2016, 1-12, https://doi.org/10.1155/2016/5102568, 2016.

Ma, F., Luo, L., Ye, A., and Duan, Q.: Drought Characteristics and Propagation in the Semiarid Heihe River Basin in Northwestern China, Journal of Hydrometeorology, 20, 59-77, https://doi.org/10.1175/JHM-D-18-0129.1, 2019.

Margariti, J., Rangecroft, S., Parry, S., Wendt, D. E., and Van Loon, A. F.: Anthropogenic activities alter drought termination. Elem. Sci. Anth., 7, 1-20, https://doi.org/10.1525/elementa.365, 2019.Vicente-Serrano, S. M., Beguería, S., and López-Moreno, J. I.: A Multiscalar Drought Index Sensitive to Global Warming: The Standardized Precipitation Evapotranspiration Index, J. Climate, 23, 1696–1718, https://doi.org/10.1175/2009JCLI2909.1, 2010.

McKee, T. B., Doesken, N. J., and Kleist, J.: The relationship of drought frequency and duration to time scales, in: Eighth Conference on Applied Climatology, Anaheim, California, 17-22 January 1993, 1993.

Mianabadi, A., Davary, K., Mianabadi, H. Karimi, P.: International Environmental Conflict Management in Transboundary River Basins. Water Resour. Manage. 34, 3445–3464, https://doi.org/10.1007/s11269-020-02576-7, 2020.Wang, Y., Yang, J., Chang, J., and Zhang, R.: Assessing the drought mitigation ability of the reservoir in the downstream of the Yellow River, Science of The Total Environment, 646, 1327-1335, 2019.

Mishra, A. K. and Singh, V. p.: A review of drought concepts, J. Hydrol., 391, 202-216, https://doi.org/10.1016/j.jhydrol.2010.07.012, 2010.

Mishra, V., Shah, R., Azhar, S., Shah, H., Modi, P., and Kumar, R.: Reconstruction of droughts in India using multiple land-surface models (1951–2015), Hydrol. Earth Syst. Sci., 22, 2269–2284, https://doi.org/10.5194/hess-22-2269-2018, 2018.

Miyan, M. A.: Droughts in Asian Least Developed Countries: Vulnerability and sustainability, Weather Clim. Extremes, 7, 8-23, https://doi.org/10.1016/j.wace.2014.06.003, 2015.

Moravec, V., Markonis, Y., Rakovec, O., Kumar, R., and Hanel, M.: A 250-Year European Drought Inventory Derived From Ensemble Hydrologic Modeling, Geophys, Res. Lett., 46, 5909-5917, https://doi.org/10.1029/2019GL082783, 2019.

Mostbauer, K., Kaitna, R., Prenner, D., and Hrachowitz, M.: The temporally varying roles of rainfall, snowmelt and soil moisture for debris flow initiation in a snow-dominated system, Hydrol. Earth Syst. Sci., 22, 3493–3513, https://doi.org/10.5194/hess-22-3493-2018, 2018.

Nalbantis, I. and Tsakiris, G.: Assessment of Hydrological Drought Revisited, Water Resour. Manage., 23, 881–897, https://doi.org/10.1007/s11269-008-9305-1, 2009.

Nash, J. E. and Sutcliffe, J. V.: River flow forecasting through conceptual models: part I —A discussion of principles, J. Hydrol., 10, 282–290, https://doi.org/10.1016/0022-1694(70)90255-6, 1970.

Nijzink, R. C., Almeida, S., Pechlivanidis, I. G., Capell, R., Gustafssons, D., Arheimer, B., Parajka, J., Freer, J., Han, D., Wagener, T., van Nooijen, R. R. P., Savenije, H. H. G., and Hrachowitz, M.: Constraining Conceptual Hydrological Models With Multiple Information Sources, Water Resour. Res., 54, 1-31, https://doi.org/10.1029/2017WR021895, 2018.

Noorisameleh, Z., Khaledi, S., Shakiba, A., Zeaiean Firouzabadi, P., Gough, W. A., and Qader Mirza, M. M.: Comparative evaluation of impacts of climate change and droughts on river flow vulnerability in Iran, Water Science and Engineering, https://doi.org/10.1016/j.wse.2020.05.001, 2020. (In Press).

Parajka, J. and Blöschl, G.: Spatio-temporal combination of MODIS images – potential for snow cover mapping, Water Resour. Res., 44, 1-13, https://doi.org/10.1029/2007WR006204, 2008.

Pathak, A. A. and Dodamani, C. B. M.: Comparison of two hydrological drought indices, Perspect. Sci., 8, 626-628, https://doi.org/10.1016/j.pisc.2016.06.039, 2016.

Pei, Z.; Fang, S.; Wang, L.; Yang, W.: Comparative Analysis of Drought Indicated by the SPI and SPEI at Various Timescales in Inner Mongolia, China, Water, 12, 1925, https://doi.org/10.3390/w12071925, 2020.

Piqué, G., Batalla, R. J., and Sabater, S.: Hydrological characterization of dammed rivers in the NW Mediterranean region, Hydrol. Processes, 30, 1691–1707, https://doi.org/10.1002/hyp.10728, 2016.

Prenner, D., Kaitna, R., Mostbauer, K., and Hrachowitz, M.: The Value of Using Multiple Hydro-meteorological Variables to Predict Temporal Debris Flow Susceptibility in an Alpine Environment, Water Resour. Res., 54, 6822-6843, https://doi.org/10.1029/2018WR022985, 2018.

Raziei, T., Saghafian, B., Paulo, A. A. Pereira, L. S., and Bordi, I.: Spatial patterns and temporal variability of drought in western Iran, Water Resour. Manage., 23, 439–455, https://doi.org/10.1007/s11269-008-9282-4, 2009.

Saidi, H., Dresti, C., Manca, D., Ciampittiello, M.: Quantifying impacts of climate variability and human activities on the streamflow of an Alpine river. Environ. Earth Sci. 77,690, https://doi.org/10.1007/s12665-018-7870-z, 2018.

Schaller, M. F. and Fan, Y.: River basins as groundwater exporters and importers: implications for water cycle and climate modeling. J. Geophys. Res. Atmos. 114. http://dx.doi.org/10.1029/2008JD010636, 2009.

Schuol, J. and Abbaspour, K. C.: Calibration and uncertainty issues of a hydrological model (SWAT) applied to West Africa, Adv. Geosci., 9, 137–143, https://doi.org/10.5194/adgeo-9-137-2006, 2006.

Schuol, J., Abbaspour, K. C., Yang, H., Srinivasan, R., and Zehnder, A. J. B.: Modeling blue and green water availability in Africa, Water Resour. Res., 44, 1-18, https://doi.org/10.1029/2007WR006609, 2008.

Spinoni, J., Naumann, G., Carrao, H., Barbosa, P., and Vogt, J.: World drought frequency, duration, and severity for 1951–2010, Int. J. Climatol., 34, 2792–2804, https://doi.org/10.1002/joc.3875, 2014.

Stagge, J.H., Tallaksen, L.M., Gudmundsson, L., van Loon, A.F., and Stahl, K.: Candidate distributions for climatological drought indices (SPI and SPEI), Int. J. Climatol., 35, 4027-4040, http://dx.doi.org/10.1002/joc.4267, 2015.

Trambauer, P., Maskey, S., Werner, M., Pappenberger, F., van Beek, L. P. H., and Uhlenbrook, S.: Identification and simulation of space–time variability of past hydrological drought events in the Limpopo River basin, southern Africa, Hydrol. Earth Syst. Sci., 18, 2925–2942, https://doi.org/10.5194/hess-18-2925-2014, 2014.

United Nation Office on Drugs and Crime (UNODC), Afghanistan Opium Survey, Government of Afghanistan, Ministry of Counter Narcotics, Kabul, 150 pp., 2006.

Van Beek, E., Bozorgi, B., Vekerdy, Z., and Meijer, K.: Limits to agricultural growth in the Sistan Closed Inland Delta, Iran, Irrig. Drain. Syst., 22, 131-143, https://doi.org/10.1007/s10795-008-9045-7, 2008.

Van Huijgevoort, M. H. J., van Lanen, H. A. J., Teuling, A. J., and Uijlenhoet, R.: Identification of changes in hydrological drought characteristics from a multi-GCM driven ensemble constrained by observed discharge, J. Hydrol., 512, 421-434, https://doi.org/10.1016/j.jhydrol.2014.02.060, 2014.

Van Lanen, H. A. J., Wanders, N., Tallaksen, L. M., and Van Loon, A. F.: Hydrological drought across the world: impact of climate and physical catchment structure, Hydrol. Earth Syst. Sci., 17, 1715–1732, https://doi.org/10.5194/hess-17-1715-2013, 2013.

Van Loon, A. F. and Laaha, G.: Hydrological drought severity explained by climate and catchment characteristics, J. Hydrol., 526, 3-14, https://doi.org/10.1016/j.jhydrol.2014.10.059, 2015.

Van Loon, A. F., Stahl, K., Di Baldassarre, G., Clark, J., Rangecroft, S., Wanders, N., Gleeson, T., Van Dijk, A. I. J. M., Tallaksen, L. M., Hannaford, J., Uijlenhoet, R., Teuling, A. J., Hannah, D. M., Sheffield, J., Svoboda, M., Verbeiren, B., Wagener, T., and Van Lanen, H. A. J.: Drought in a human-modified world: reframing drought definitions, understanding, and analysis approaches, Hydrol. Earth Syst. Sci., 20, 3631–3650, https://doi.org/10.5194/hess-20-3631-2016, 2016.Van Loon, A. F., Rangecroft, S., Coxon, G., Breña Naranjo, J. A., Van Ogtrop, F., and Van Lanen, H. A. J.: Using paired catchments to quantify the human influence on hydrological droughts, Hydrol. Earth Syst. Sci., 23, 1725–1739, https://doi.org/10.5194/hess-23-1725-2019, 2019.

Vicente-Serrano, S. M., López-Moreno, J. I., Beguería, S., and Lorenzo-Lacruz, J.: Accurate Computation of a Streamflow Drought Index, J. Hydrol. Eng., 17, 318-332, https://doi.org/10.1061/(ASCE)HE.1943-5584.0000433, 2012.

Vining, K. C. and Vecchia, A. V.: Water-Balance Simulations of Runoff and Reservoir Storage for the Upper Helmand Watershed and Kajakai Reservoir, Central Afghanistan, U.S. Agency for International Development, Sci. Invest. Rep., 2007–5148, 24 pp., 2007.

Wada, Y., Reager, J. T., Chao, B. F., Wang, J., Lo, M. H., Song, C., Li, Y., and Gardner, A. S.: Recent changes in land water storage and its contribution to sea level variations, Surv. Geophys., 38, 131-152, https://doi.org/10.1007/s10712-016-9399-6, 2016.

Wan, W., Zhao, J., Li, H.-Y., Mishra, A., Hejazi, M., Lu, H., Demissie, Y., and Wang, H.: A holistic view of water management impacts on future droughts: A global multimodel analysis. Journal of Geophysical Research: Atmospheres, 123, 5947–5972. https://doi.org/10.1029/2017JD027825, 2018.

Wan, W., Zhao, J., Li, H.-Y., Mishra, A., Ruby Leung, L., Hejazi, M., … Wang, H.: Hydrological drought in the anthropocene: Impacts of local water extraction and reservoir regulation in the U.S. J. Geophys. Res. Atmos., 122, 11,313–11,328. https://doi.org/10.1002/ 2017JD026899, 2017.

Wardlaw, R., Goes, B., Parajuli, U., and Hancock, I.: Helmand River Basin Master Plan, Phase 3, Water Resources Modelling for Helmand River Basin, Ministry of Energy and Water, Kabul, Technical Report., 3, 251 pp., 2013.

Weng, B., Zhang, P., and Li, S.: Drought risk assessment in China with different spatial scales, Arabian J. Geosci., 8, 10193-10202, https://doi.org/10.1007/s12517-015-1938-9 , 2015.

Whitney, J. W.: Geology, Water, and Wind in the Lower Helmand Basin, Southern Afghanistan, U.S. Agency for International Development, Sci. Invest. Rep., 2006–5182, 50 pp., 2006.

Wisser, D., Fekete, B. M., Vörösmarty, C. J., and Schumann, A. H.: Reconstructing 20th century global hydrography: a contribution to the Global Terrestrial Network- Hydrology (GTN-H), Hydrol. Earth Syst. Sci., 14, 1–24, https://doi.org/10.5194/hess-14-1-2010, 2010.

Wu, J., Chen, X., Yao, H., Gao, L., Chen, Y., and Liu, M.: Non-linear relationship of hydrological drought responding to meteorological drought and impact of a large reservoir, J. Hydrol., 551, 495-507, https://doi.org/10.1016/j.jhydrol.2017.06.029, 2017.

Wu, J., Chen, X., Yu, Z, Yao, H, Li, W., and Zhang, D.: Assessing the impact of human regulations on hydrological drought development and recovery based on a 'simulated-observed' comparison of the SWAT model, J. Hydrolo., 577, 123990, https://doi.org/10.1016/j.jhydrol.2019.123990, 2019.

Yassin, F., Razavi, S., Elshamy, M., Davison, B., Sapriza-Azuri, G., and Wheater, H.: Representation and improved parameterization of reservoir operation in hydrological and land-surface models, Hydrol. Earth Syst. Sci., 23, 3735–3764, https://doi.org/10.5194/hess-23-3735-2019, 2019.

Zhang, D., Liu, X., Liu, C., and Bai, P.: Responses of runoff to climatic variation and human activities in the Fenhe River, China, Stochastic Environ. Res. Risk Assess., 27, 1293-1301, https://doi.org/10.1007/s00477-012-0665-y, 2013.

Table 1. Summary of sub-basins and reservoirs in the Helmand River Basin. UHRB, CHRB and LHRB denote the Upper, Central and Lower Helmand River Basins, respectively, UARB and LARB are the Upper and Lower Arghandab River Basins, respectively and SISP is the Sistan Plain. The subscripts U and D are inflows into the reservoir upstream of the dams and outflows from the reservoirs downstream of the dams, respectively.

Location ID	Sub-basin Symbol	Station name	Latitude (°)	Longitude (°)	Average Elevation (m)	Precipitation (mmy^{-1})	Discharge (mmy^{-1})	Aridity Index (-)	Observation Period (daily/monthly)	Reservoir Storage (10^6 m^3)
1	UHRB$_U$	Dehraout – Kajakai Dam inflow	32.42	65.28	2865	300	130	0.29	1970–1979 (d)	----
2	UHRB$_D$	Kajakai Dam outflow	32.19	65.06	2798	300	111	0.29	1970–1979 (d)	----
3	CHRB	-	-	-	1994	204	-	0.13	-	
4	UARB$_U$	Upper Arghandab – Dahla Dam inflow	31.50	65.52	2830	254	91	0.22	1970–1979 (d)	----
5	UARB$_D$	Dahla Dam outflow	31.57	66.02	2776	254	88	0.22	1970–1979 (d)	----
6	LARB	Qala-i-Bust	31.30	64.23	2509	225	32	0.13	1970–1980 (d)	----
7	LHRB	Char Burjak	30.17	62.02	2610	229	43	0.16	1970–1979 (d)	----
8	SISP	Sistan	30.49	61.46	2584	250	15	0.16	1970–2006 (m)	----
9		Kajakai Dam Reservoir	32.19	65.10	963	----	----		----	1800
10		Dahla Dam Reservoir	31.52	65.54	1070	----	----		----	450

Table 2: Uniform prior and posterior distributions of model parameters for the calibrated models. The posterior column distributions show the parameter values of the best available parameter set as well as the 5/95th percentile of feasible solutions (in brackets). Note that loss factor K_L had negligible influence and was thus set to 0 for the models of ID6 and ID7 to keep to the principle of model parsimony.

ID	Sub-basin symbols	Parameter	Prior distribution	Posterior distribution
1	UHRB$_U$	I_{max} [mm]	0-2	0.13 (0.11-0.55)
		C_e [-]	0.2-1	0.44 (0.36-0.57)
		S_{umax} [mm]	40-800	250 (112-550)
		β [-]	0.2-3	1.08 (0.68-1.55)
		P_{max} [mm d^{-1}]	0.009-1	0.67 (0.65-0.70)
		T_{lag} [d]	2-7	3.12 (3.00-3.84)
		K_f [d^{-1}]	0.01-0.1	0.07 (0.06-0.08)
		K_s [d^{-1}]	0.0009-0.01	0.001 (0.001-0.002)
		T_{th} [°C]	-2.5 – 2.5	-1.12 (-1.42 -0.69)
		F_{dd} [mm °C^{-1} d^{-1}]	0-3	0.38 (0.27-0.51)
4	UARB$_U$	I_{max} [mm]	0-2	0.45 (0.10-0.83)
		C_e [-]	0.2-1	0.84 (0.41-0.84)
		S_{umax} [mm]	40-800	200 (100-430)
		β [-]	0.2-3	1.73 (0.93-2.27)
		P_{max} [mm d^{-1}]	0.009-1	0.47 (0.15-0.47)
		T_{lag} [d]	2-7	2.41 (2.00-3.01)
		K_f [d^{-1}]	0.01-0.1	0.07 (0.03-0.08)
		K_s [d^{-1}]	0.0009-0.01	0.001 (0.001-0.003)
		T_{th} [°C]	-2.5 – 2.5	-1.35 (-1.50 -1.14)
		F_{dd} [mm °C^{-1} d^{-1}]	0-3	0.85 (0.39-1.99)
6	LARB	I_{max} [mm]	0.1-3	1.66 (0.97-2.15)
		C_e [-]	0.1-1	0.23 (0.18-0.33)
		S_{umax} [mm]	40-600	455 (200-515)
		β [-]	0.1-3.00	2.76 (1.56-2.82)
		P_{max} [mm d^{-1}]	0.01-0.1	0.04 (0.03-0.05)
		T_{lag} [d]	2-7	3.45 (2.12-4.18)
		K_f [d^{-1}]	0.01-1.00	0.02 (0.01-0.02)
		K_s [d^{-1}]	0.0009-0.01	0.009 (0.008-0.01)
		K_L [-]	0.00	0.00 (0.00-0.00)
7	LHRB	I_{max} [mm]	0.1-3	1.58 (0.27-1.85)
		C_e [-]	0.1-1	0.19 (0.11-0.35)
		S_{umax} [mm]	40-600	515 (220-585)
		β [-]	0.1-3.00	2.81 (1.86-2.88)
		P_{max} (mm/day)	0.01-0.1	0.03 (0.02-0.05)
		T_{lag} [d]	3-10	6.61 (3.42-7.12)
		K_f [d^{-1}]	0.01-1.00	0.03 (0.02-0.05)
		K_s [d^{-1}]	0.0009-0.01	0.009 (0.005-0.01)
		K_L [-]	0.00	0.00 (0.00-0.00)
8	SISP	I_{max} [mm]	0.1-3	1.58 (0.27-1.85)
		C_e [-]	0.1-1	0.19 (0.11-0.35)
		S_{umax} [mm]	40-600	515 (220-585)
		β [-]	0.1-3.00	2.81 (1.86-2.88)
		P_{max} [mm d^{-1}]	0.01-0.1	0.03 (0.02-0.05)
		T_{lag} [d]	3-10	6.61 (3.42-7.12)
		K_f [d^{-1}]	0.01-1.00	0.03 (0.02-0.05)
		K_s [d^{-1}]	0.0009-0.01	0.009 (0.005-0.01)

		K_L [-]	0-1	0.34 (0.33 – 0.36)
9	$UHRB_D$	a_h [d^{-1}]	-	0.27 (0.13-0.40)
		b_h [-]	-	0.64 (0.50-0.77)
		c_h [mm d^{-1}]	-	-173 (-332- -15)
		a_l [d^{-1}]	-	0.13 (0.09-0.17)
		c_l [mm d^{-1}]	-	217 (173-262)
10	$UARB_D$	a_h [d^{-1}]	-	0.21 (0.11-0.31)
		b_h [-]	-	0.86 (0.79-0.93)
		c_h [mm d^{-1}]	-	-58 (-84- -32)
		a_l [d^{-1}]	-	0.26 (0.11-0.42)
		c_l [mm d^{-1}]	-	25 (17-33)

Table 3. Classification of standardized drought indices DI used in this study (SPI, SPEI and SDI).

Classification	DI [-]	Probability [-]
No drought	$DI > 0$	0.501
Mild drought	$-1 \leq DI < 0$	0.341
Moderate drought	$-1.5 \leq DI < -1$	0.092
Severe drought	$-2 \leq DI < -1.5$	0.044
Extreme drought	$DI < -2$	0.023

Table 4. Model performance metrics for calibration and validation in all study sub-basins. The values include the best performing model as well as the range of all solutions retained as feasible (in brackets)

Location ID	Sub-basin Symbol	Calibration period (1971 – 1975)		Validation period (1976–1979)	
		$E_{NS,Q}$	$E_{NS,log(Q)}$	$E_{NS,Q}$	$E_{NS,log(Q)}$
1	UHRB$_U$	0.82 (0.82-0.83)	0.91 (0.90-0.91)	0.80 (0.79-0.80)	0.86 (0.86-0.87)
2	UHRB$_D$	0.79 (0.78-0.80)	0.81 (0.79-0.82)	0.79 (0.79-0.80)	0.85(0.84-0.86)
4	UARB$_U$	0.83 (0.83-0.84)	0.85 (0.85-0.86)	0.73 (0.72-0.73)	0.78 (0.78-0.89)
5	UARB$_D$	0.89 (0.88-0.90)	0.92 (0.91-0.92)	0.74 (0.74-0.75)	0.80 (0.79-0.81)
6	LARB	0.70 (0.69-0.71)	0.73 (0.71-0.74)	0.81 (0.80-0.83)	0.83 (0.81-0.86)
7	LHRB	0.82 (0.81-0.83)	0.85 (0.83-0.86)	0.84 (0.82-0.86)	0.88 (0.86-0.91)
		(1971 – 1975)		(1976 – 2006)	
8	SISP	0.88 (0.86-0.89)	0.89 (0.87-0.89)	0.73 (0.68-0.74)	0.75 (0.74-0.77)
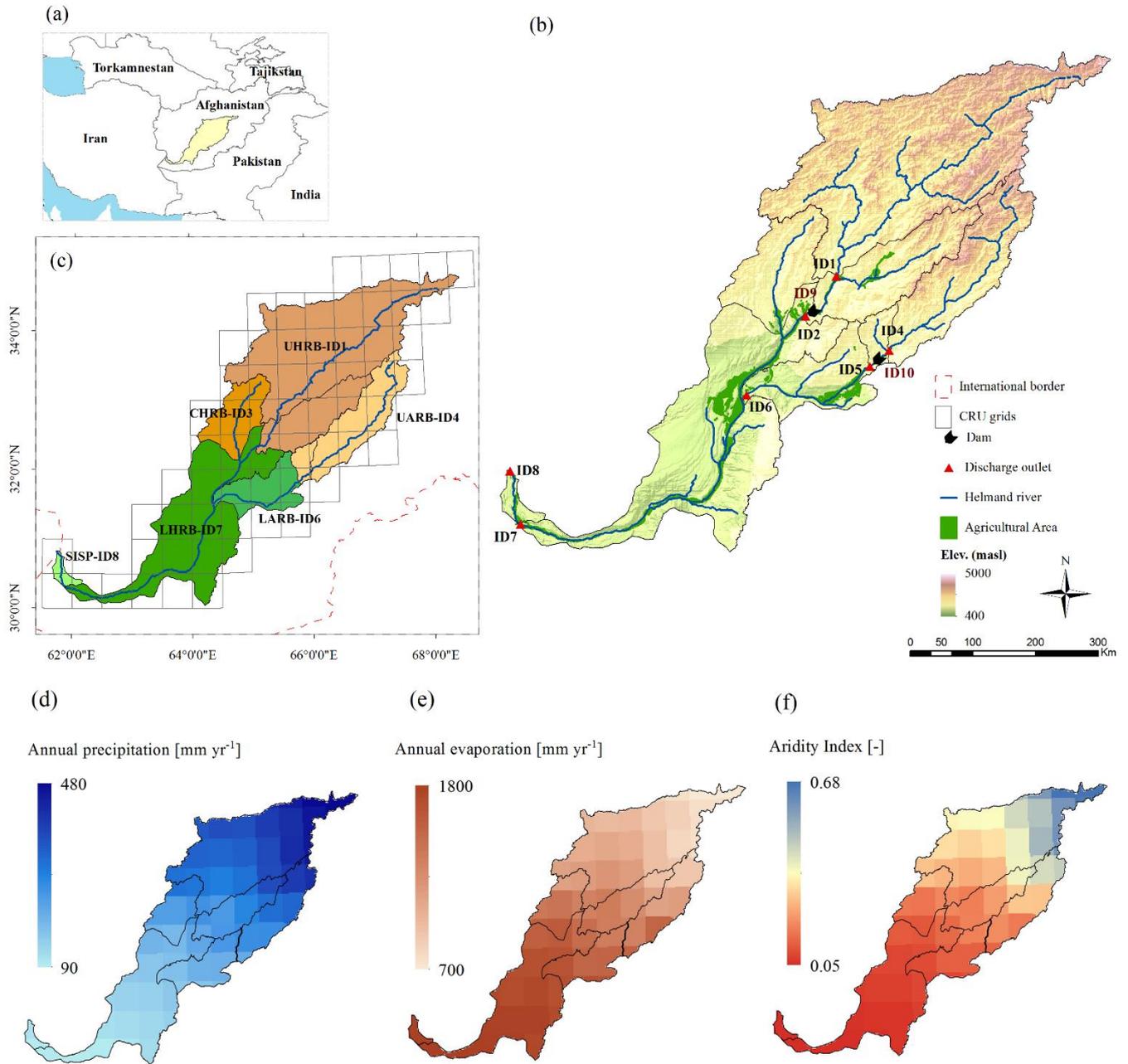
Figure 1. (a) The location of Helmand River Basin (HRB) in central Afghanistan, (b) elevation map of the HRB, also indicating the sub-basin boundaries, the locations of the sub-basin outlets and the agriculturally used area (as of 2006), (c) outline of the sub-basins analysed in this study, including the grid cells of CRU precipitation data used, (d) long-term mean annual precipitation P [mm y^{-1}], (e) long-term mean annual potential evaporation E$_P$ [mm y^{-1}] and (f) the aridity index I$_A$= P/E$_P$ [-]
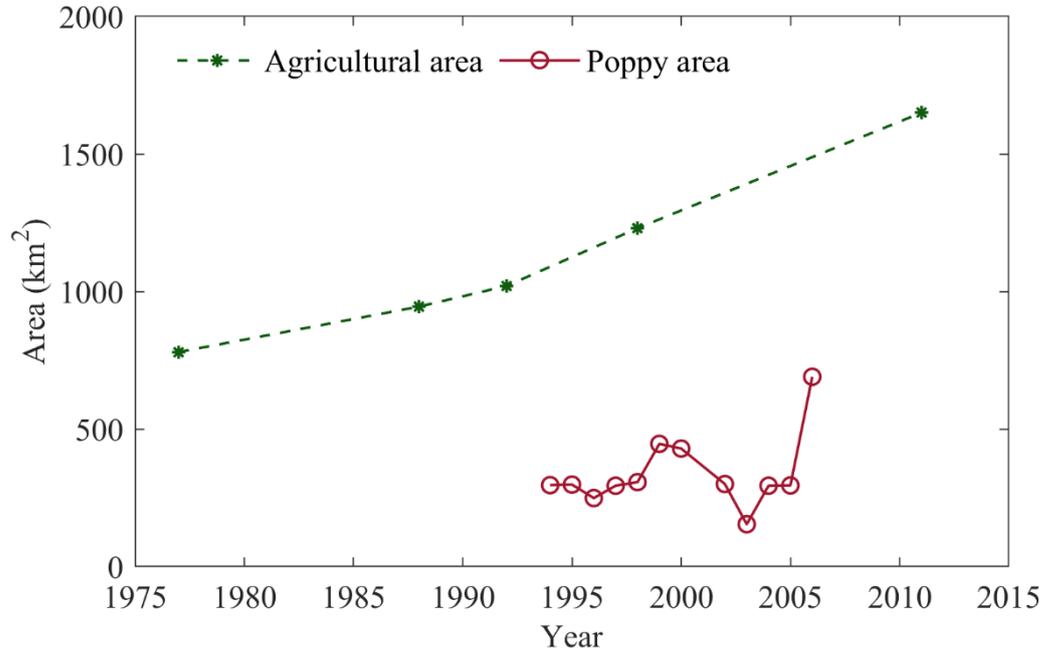
Figure 2. Evolution of total agricultural area in the HRB (1976-2011) and Poppy cultivated area thereof (1994-2006).
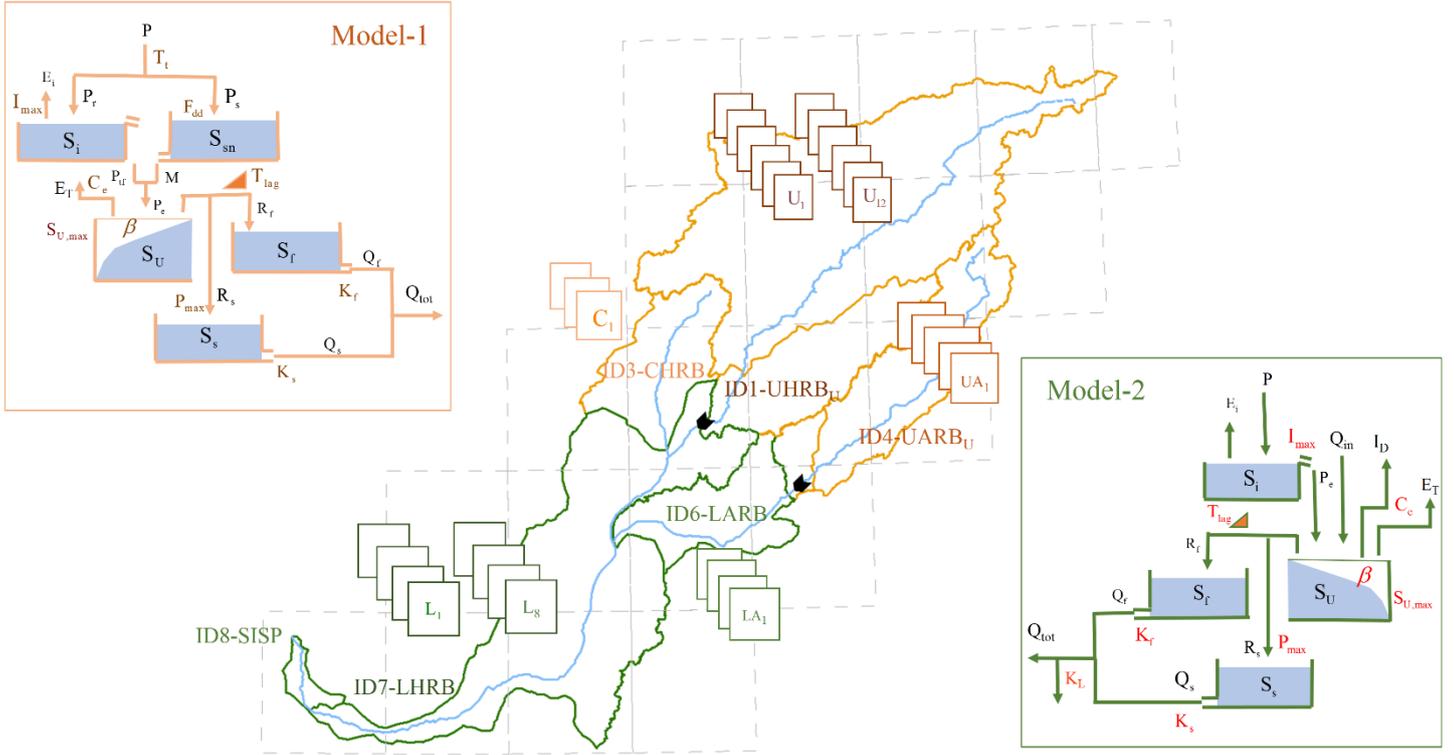
Figure 3. The distributed model structure consisting of parallel components (the structure of Model-1 used for UHRB, CHRB and UARB, Model-2 is used in LHRB, LARB and SISP) and 32 units of CRU grid cell, representing one subbasin each, characterized by an individual parameter set. Variables: P total precipitation [mm d^{-1}], P_s snowfall [mm d^{-1}], P_r rainfall [mm d^{-1}], M snowmelt [mm d^{-1}], P_{tf} throughfall [mm d^{-1}], P_e effective precipitation [mm d^{-1}], E_T transpiration [mm d^{-1}], E_i interception evaporation [mm d^{-1}], R_f recharge of fast reservoir [mm d^{-1}], R_s recharge of slow reservoir [mm d^{-1}], I_D irrigation demand for LHRB and LARB, respectively., $Q_{in} = Q_{UHRB_U} + Q_{LARB} + Q_{CHRB}$ [mm d^{-1}], Q_f runoff from fast reservoir [mm d^{-1}], Q_s runoff from slow reservoir [mm d^{-1}], Q_{tot} = total runoff [mm d^{-1}], S_{sn} storage in snow reservoir [mm], S_i storage in interception reservoir [mm], S_U storage in unsaturated reservoir [mm], S_f storage in fast reservoir [mm], S_s storage in slow reservoir [mm]. Parameters: T_t threshold temperature [℃], F_{dd} melt factor [mm ℃$^{-1}$ d^{-1}], I_{max} interception capacity [mm], $S_{U,max}$ storage capacity in unsaturated reservoir [mm], β shape parameter [-], P_{max} percolation capacity [mm d^{-1}], Ce runoff generation coefficient [-], K_f storage coefficient of fast reservoir [d^{-1}], K_s storage coefficient of slow reservoir [d^{-1}], K_L loss factor [-] and, T_{lag} lag time [d].
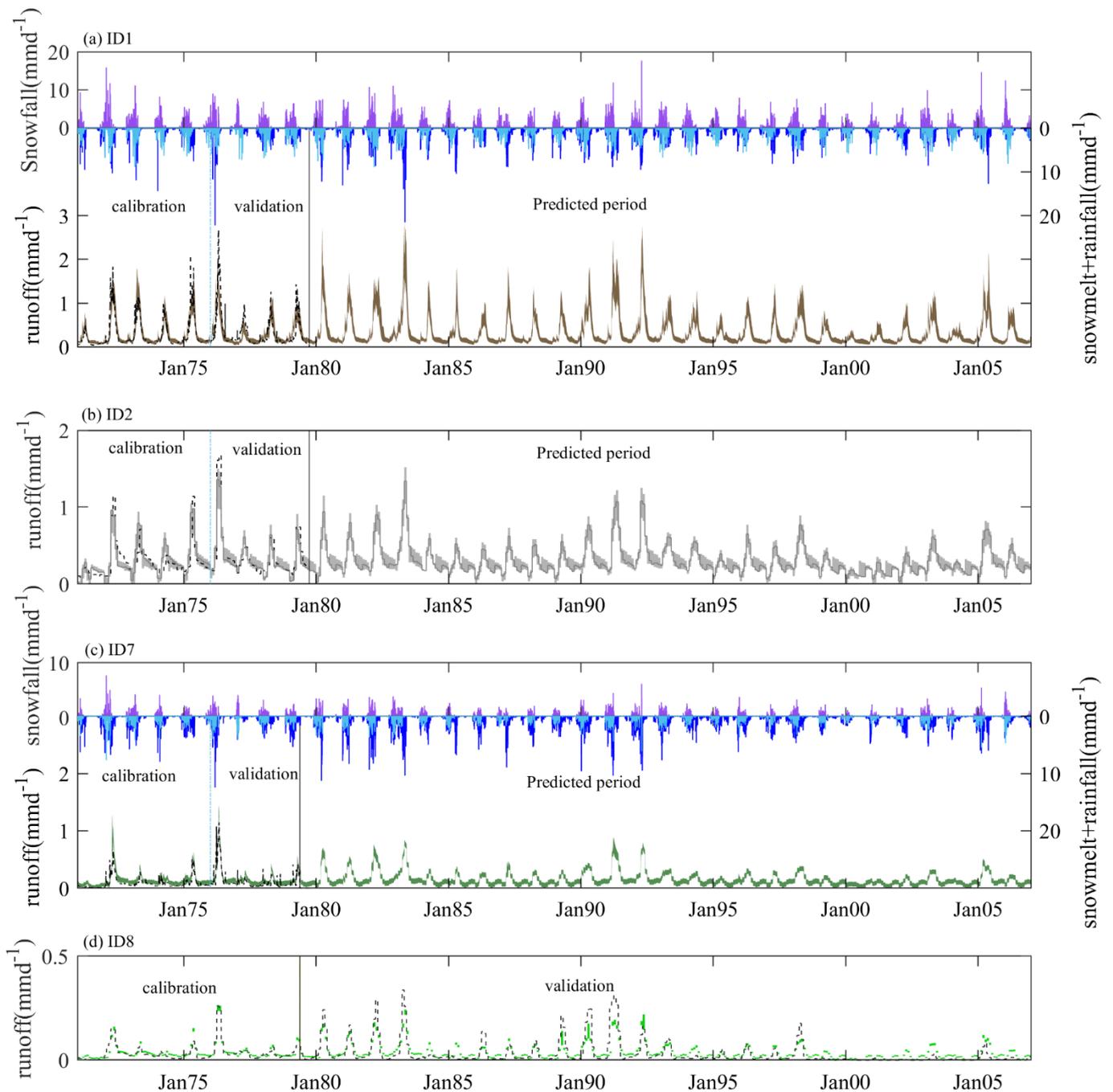
Figure 4. Precipitation and streamflow in UHRB$_U$ (ID1), UHRB$_D$ (ID2), LHRB (ID7), and SISP (ID8). The purple bars show the modelled snowfall P$_S$ [mm d^{-1}], the dark blue bars the modelled snowmelt M [mm d^{-1}] and the light blue bars the modelled rainfall P$_R$ [mm d^{-1}]. The dashed black lines indicate the observed runoff and the shaded areas the uncertainty ranges of modelled runoff during calibration, validation and prediction periods
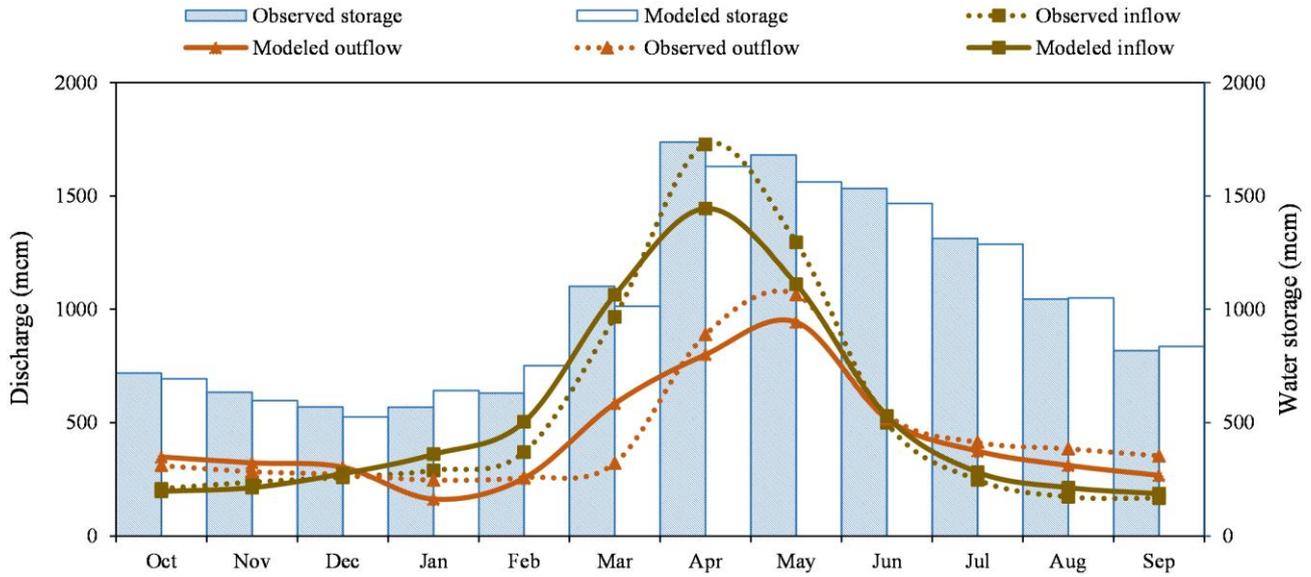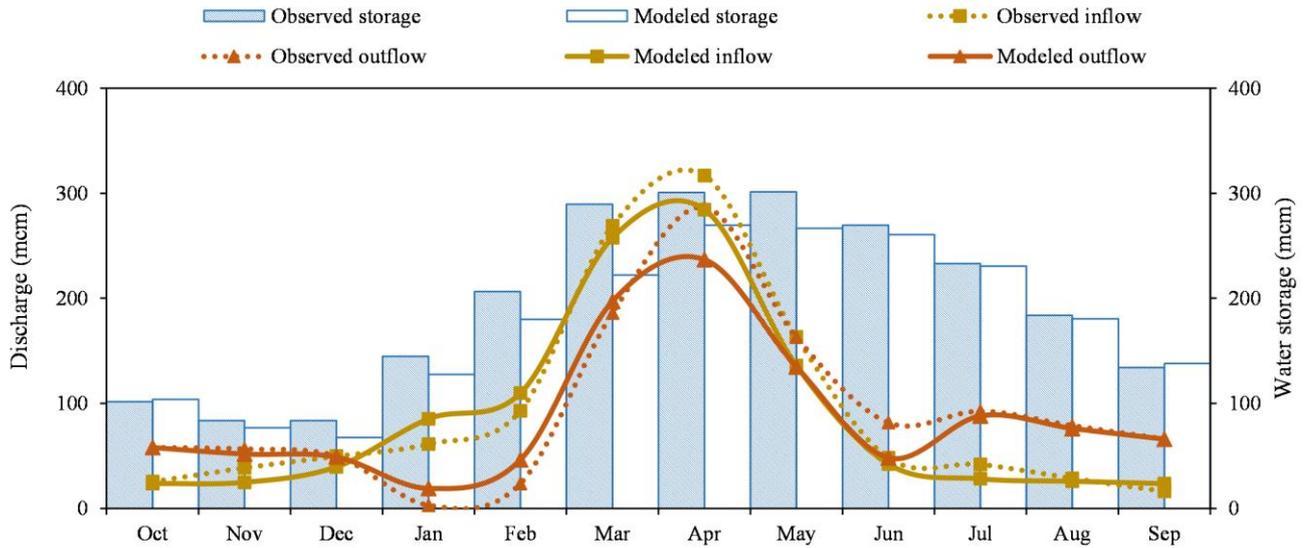
Figure 5. Mean observed and modeled inflow, outflow, and storage volume at (a) the Kajakai and (b) Dahla Dam reservoirs during the 1970–1979 period
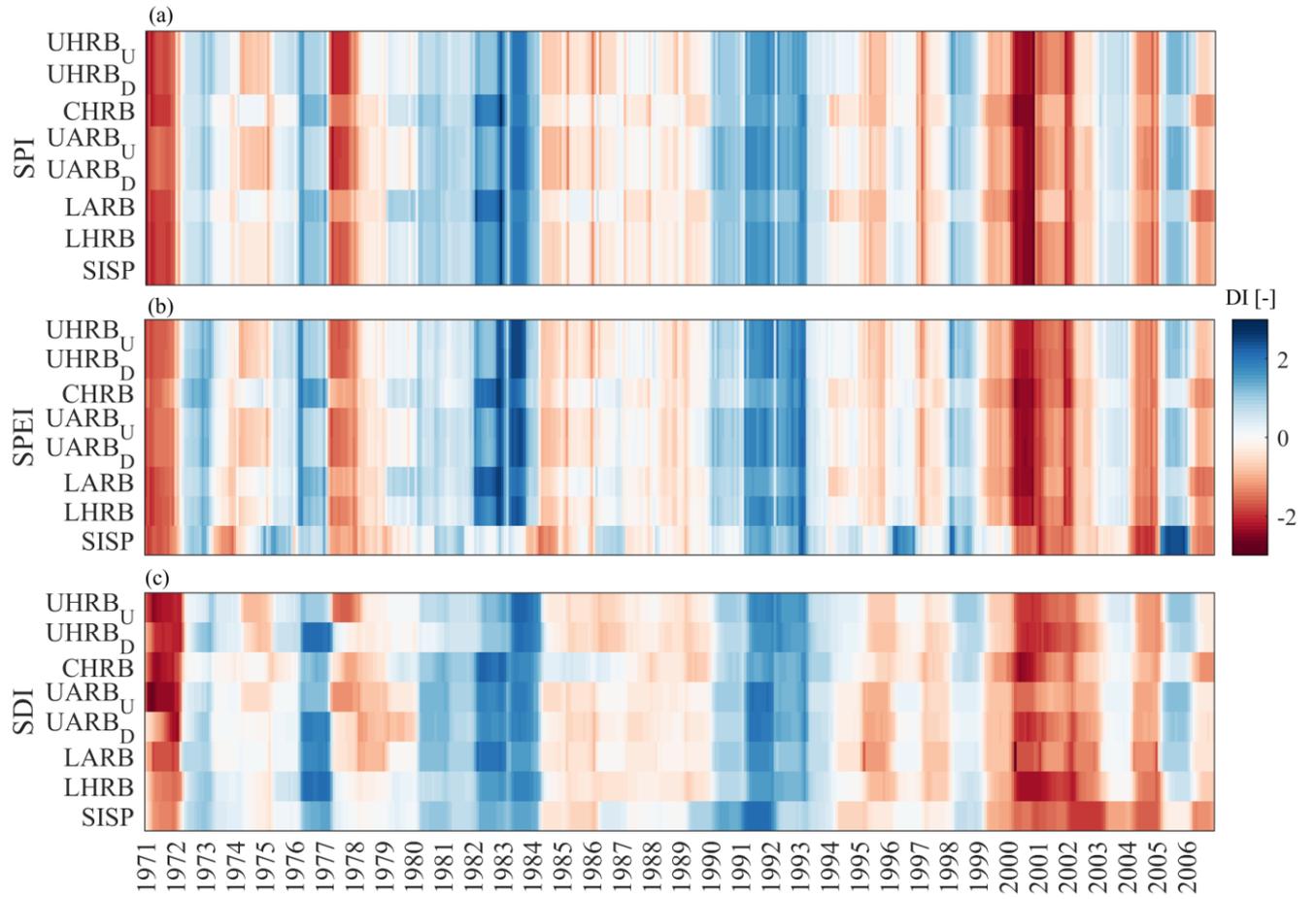
Figure 6. Time series of monthly drought indices (based on 12 months accumulation time) SPI, SPEI and SDI for the sub-basins ID1 – ID8 for the 1970 – 2006 study period
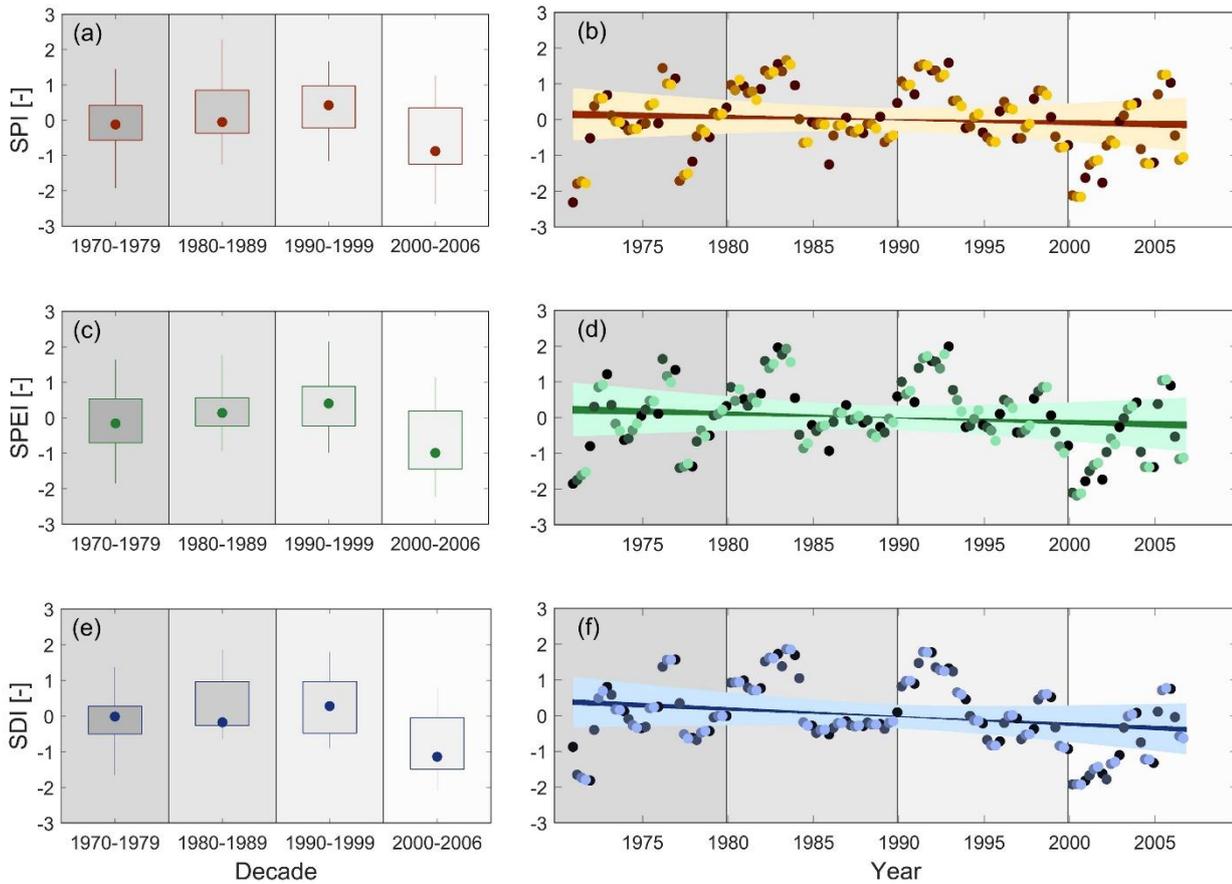
Figure 7. Decadal distributions and time series of mean basin (a)-(b) SPI, (c)-(d) SPEI and (e)-(f) SDI over the study period. The dots in the box plots indicate the median values and the whiskers the 5/95th percentiles. The dark to light shaded dots in the time series plots indicate the monthly drought indices (based on 12 months accumulation time) for all months of January, April, July and October, respectively. The dark shaded areas indicate the envelope of trend lines for the trends estimated based on all months of January, April, July and October, respectively. The light shaded areas show the associated envelope of 5/95th confidence intervals.
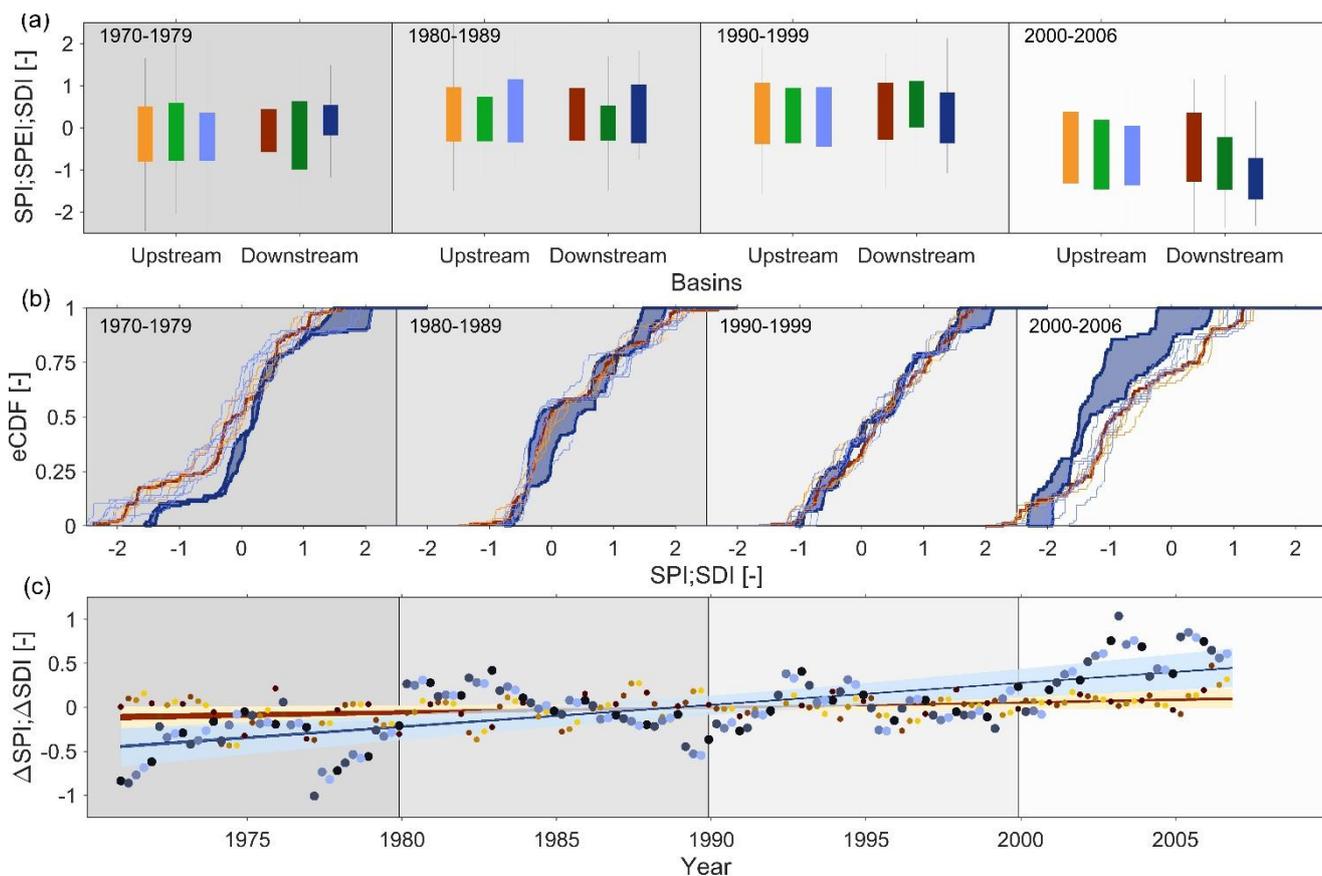
Figure 8. (a) decadal distributions of SPI, SPEI and SDI for the most upstream sub-basins (ID1-ID6) and the downstream sub-basins (ID7-ID8), (b) decadal empirical cumulative distribution functions of SPI (thins red lines upstream basins, bold red lines: downstream basins) and SDI (thin blue lines: upstream basins, bold blue lines: downstream basins). Note that the blue shaded area is added for better visualization of the shifts in downstream basins only and does not have a specific meaning. (c) time series of differences between mean upstream and mean downstream SPI (ΔSPI: yellow and red shades) as well as between mean upstream and mean downstream SDI (ΔSDI: blue shades). The symbols with shades from dark to light indicate the monthly SPI values (based on 12 months accumulation period) for the months January, April, July and October, respectively. The dark shaded areas indicate the envelope of trends in ΔSPI and ΔSDI, respectively, estimated based on all months of January, April, July and October, respectively. The light shaded areas show the associated envelope of 5/95th confidence intervals.
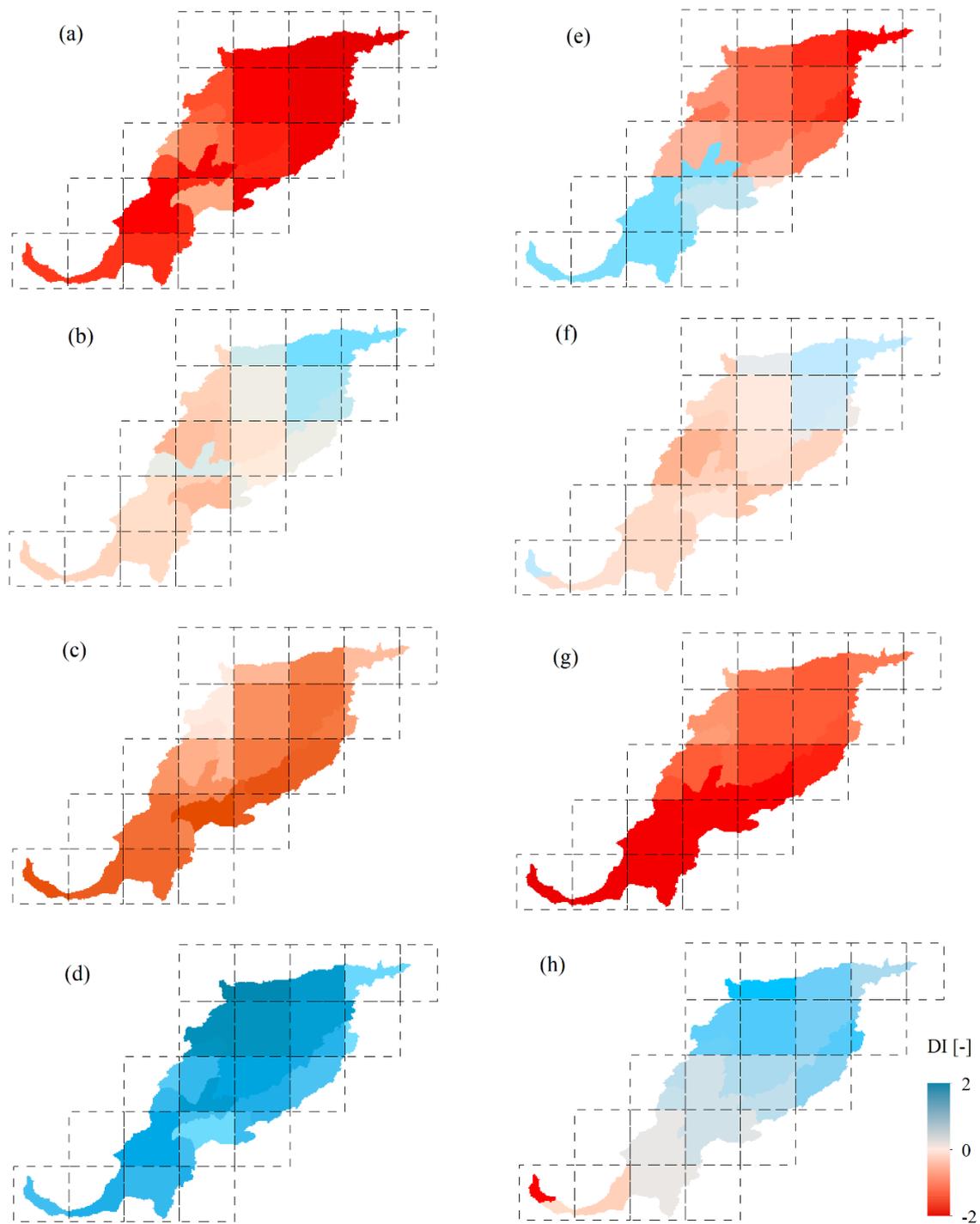
Figure 9. Spatial distribution of (a)-(d) SPI and (e)-(h) SDI for the years 1977, 1987, 2002 and 2003, based on the grid cells of the model application.
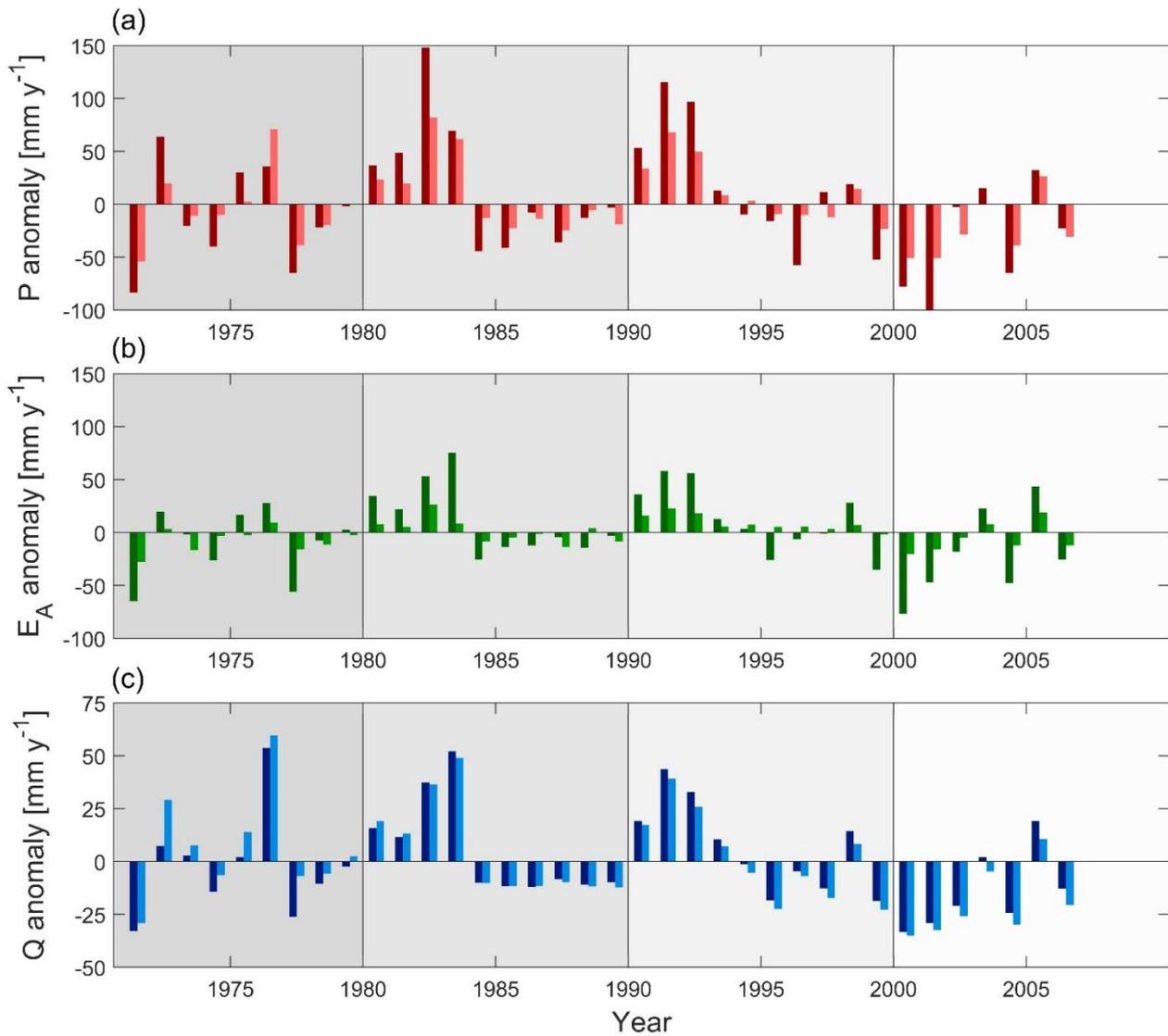
Figure 10. (a) Precipitation anomalies, (b) actual evaporation anomalies (here: $E_A = E_I + E_T$) and (c) streamflow anomalies over the study period. All anomalies are calculated based on the 1970-2006 mean values. The dark shaded bars indicate the combined flows to/from the upper basin (ID1-ID6), the light shaded bars show the flows to/from the lower basin (ID7).
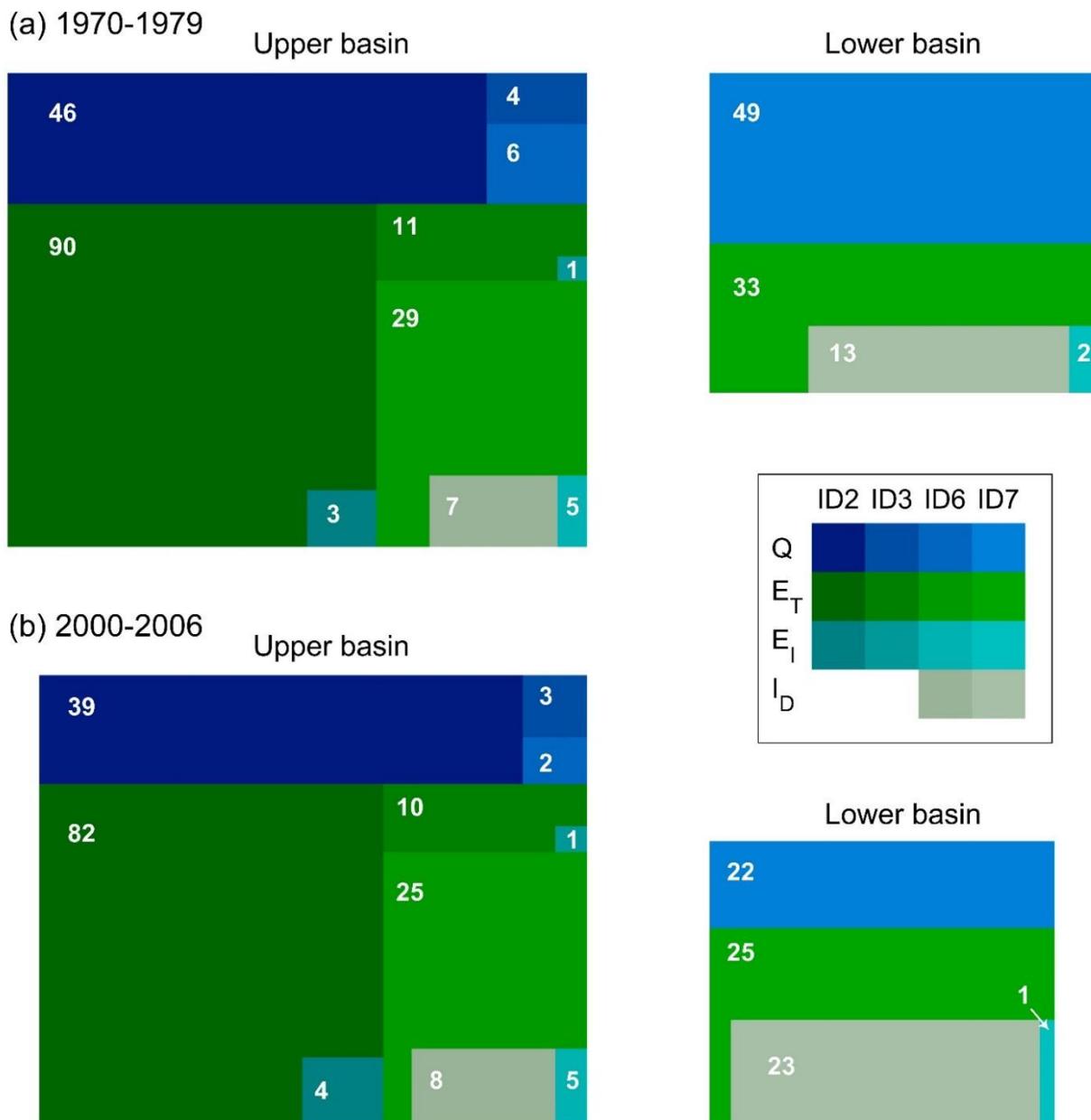
Figure 11. Water balances of the Upper and the Lower basins, respectively for (a) the 1970-79 and (b) the 2000-2006 periods. The size of the outer squares is equivalent to the total water available, i.e. for the upper basin precipitation P, for the lower basin precipitation P plus the combined inflow Q from the upper basin. The size of the internal rectangles of each flux (Q: streamflow, E_T: transpiration, E_I: interception evaporation and I_D: irrigation demand) in each sub-basin is equivalent to its fraction of the total available water in the upper and lower basin, respectively. The fluxes represent the decadal mean values and are shown in [mm y^{-1}].