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Effect of Baseline Snow-Pack Assumptions in the HySIM Model in Predicting Future Hydrological Behavior of a Himalayan Catchment

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Abstract

Glaciers and snow-packs influence stream flow by altering the volume and timing of discharge. Without reliable data on baseline snow and ice volumes, properties andbehaviour, initialising hydrological models for climate impact assessment is challenging.Two contrasting HySIM model builds were calibrated and validated against observed discharge data (2000-2008)assuming that snowmelt of the baseline permanent snow-pack reserves in the high-elevation sub-catchment are either constrained (snow melt is limited to the seasonal snow accumulation) or unconstrained (snow melt is only energy-limited). We then applied both models within a scenario-neutral framework to develop Impact Response Surface of hydrological response to future changes in annual temperature and precipitation.Both models had similar baseline model performance (NSE of 0.69-0.70 in calibration and 0.64-0.66 in validation), but the impact response surfaces differ in the magnitude and (for some combinations) direction of model response to climate change at low (Q10) and high (Q90) daily flows. The implications of historical data inadequacies in snow-pack characterization for assessing the impacts of climate change and the associated timing of hydrological tipping points are discussed.

Keywords: snow pack, uncertainty, TRMM3B42 V7, Impact Response Surface, evapotranspiration, Climate change

1. Introduction:

In mountainous and glacialcatchments, the magnitude and seasonality of river discharge are greatly influenced by snow-pack development and behaviour (Ye et al., 2003). Hydrological modelling in such mountainous regions is, however, significantly constrained with model biases and input data uncertainties regarding snow and ice reserves (Fontaine et al., 2002); thus affecting the reliability of snow/glacier melt- runoff predictions and water resources and flood risk management. For example, snow depth is an important parameter for hydrological simulation (Bell et al., 2016), especially for peak floods caused by snowmelt (Bergeron et al, 2014). Accurate observation and measurement of snow pack and glaciers at the catchment or river basin scale, including data on water-equivalent storage, temporal depth-area relationships, albedo etc., is challenging in such poorly instrumented, remote and often inaccessible environments. Given the importance of these properties for physically-based, spatially-distributed hydrological modelling, modellers may rely on the increasing array of remote sensing datasets that are becoming available, such as the NOAA Geostationary Orbiting Environmental Satellite (GOES) orMODerate resolution Imaging Spectroradiometer (MODIS) on the NASA Terra satellites (Schmugge et al., 2002), along with snow water equivalent (SWE) retrieval methods and algorithms (Dai et al., 2012).

However, practitioners still use various indirect observations, assumptions, expert knowledge and inverse modelling skills for conceptual modelling in ‘melt dominated’ catchments. These include assumptions of linear relationships between terrain elevation and snow depth, estimation of snow water equivalent through evaluation of snow-covered area

(Liston, 1999), use of passive microwave data (Dong et al., 2005) and incorporation of interpolation of ground-based measurements with remote sensing data (Kelly et al 2003, Kelly 2009); but none of these approaches are free from uncertainty in actual snow water equivalent storage.

Despite these challenges in defining snow boundary conditions, conceptual hydrological models with snowmelt components, using either energy budget (e.g. Anderson, 1976), temperature-index or degree-day methods for snow accumulation and melting (e.g. Martinecet al., 1983) are widely used in snow-dominated catchments. Examples include HBV (Akhtar et al., 2008), SRM (Kult et al., 2012), GERM (Farinotti et al., 2012), GSM-SOCONT (Schaefli and Huss, 2011), J-2000 (Nepal et al., 2014), OEZ (Hagg et al., 2007), SNOWMOD (Singh et al., 2006), UBC Watershed (Loukas et al., 2002), WATFLOOD (Comeau et al., 2009) and HySIM (Manley and Water Resource Associates Ltd, 2006).Glacier mass balance is a major climate indicator in glacio-hydrological modelling using index based approaches in conceptual models (e.g.: , TOPKAPI-ETH, SRM-model etc).Previous work in the Alps in Europe has evaluated the influence of snow and ice reserves on simulated hydrological behaviour, demonstrating the importance of priorknowledge on ice-thickness distribution, total ice volume amount and spatial distribution of winter snow and glacier area change in constraining snow and ice melt for runoff projections (e.g. Bavay et al., 2013; Huss et al., 2014), compared to assuming unlimited snow/glacier melt. In this study we have selected HySIM model due to its (i) capability to simulate surface water resources in both data-limited and snow-dominated environments, (ii) ability to simulate daily surface runoff, percolation to groundwater and river flow, (iii) in-built multiple parameter optimization with four objective functions, (iv) inclusion of advanced hydraulic and hydrologic parameters in the model and (v) graphics facility for rapid visualization of inputs and outputs

It is common within climate change studies on glacier dominated catchments for glacier extent to be assumed to be in steady-state (Hagg et al., 2013). However, Hagg et al (2007) considers that treating glaciers as static ice bodies is a major limitation in climate change impact studies.Unsteady state and seasonal snow model setupsmay lead to enhancedhydrological seasonality due to larger direct runoff in wet seasons and lower glacier or snow melt in dry seasons (Juen et al., 2007).

The purpose of this study is to explore the implications of model-setup assumptions regarding the behaviour of historical snow pack reserves for water resources simulation in the Himalayan Beas river basin of north India and to see how the model assumptions on snow and glaciers make an impact on hydrological model simulations under changing climate. The two research questions are (i) How do differing assumptions on baseline snow-pack behavior affectmodel parameter values and performance? (ii) How do the baseline snow-pack assumptions affect simulated future climate change impacts?

Hydrological modelers often ignore inadequacies in the input data or data assumptions which can affect baseline/future model performance, instead focusing more on the quantification of model parameters. Therefore in this study, the consequences of such baseline set-up assumptions for simulated future hydrological response to changed temperature and precipitation under two future snowpack scenarioswere assessed. The paper provides valuable guidance for conceptual hydrological modelling in snow dominated regions, considering the effects of data inadequacies in snow pack characterization and its implications in climate change related hydrological impact studies. However, it must be noted that the purpose of this paper is not to predict the timescale of glacial retreat or loss, recognising the difficulty and uncertainty in estimating rates of loss of glacial area or volume in the Himalayas and Tibetan Plateau (Yao et al., 2012).

2. Materials and Methods

2.1 Beas Basin

This study focuses on the perennial River Beas in thenorth-western Himalayan region of India. The river basin upstream of Pong Dam, bounded by latitude 31°28'- 32°26'N and longitude 75°56'- 77°48'E, has a drainage area of 12,560 km^2 (Figure 1) which varies in elevation from 245 to 6617 meters above sea level. The major land cover classes include forest, glacier and bare rock, with about 65% of the area covered with snow during winter (Singh and Bengtsson, 2003). The Beas basin receives around 70% of the annual rainfall during the summer monsoon between June and September, and the basin is characterized by mean minimum and mean maximum winter temperatures of -1.6^{o}C and 7.7^{o}C, respectively (Singh and Ganju, 2008) although there are significant spatial differences due to the elevation range. Daily gauged inflows to the Pong reservoir are available from January 1998 to December 2008 (11 years) used for this study.

2.2 Meteorological data

This study has used Tropical Rainfall Measuring Mission (TRMM) 3B42V7 daily gridded precipitation data for 1998 to 2008 to represent the precipitation variability of the catchment. The TRMM data, which is widely used inenvironmental and hydrological research, has a spatial resolution of 0.25^{o}x 0.25^o and covers the latitudinal band of 50^o N-S. TRMM data sets have been used in studies of many catchments in theHimalayan region (Bookhagen and Burbank, 2010).Moreover, Xue et al (2013) demonstrated that 3B42V7 provides better basin-scale agreement with observed (2001–2010) monthly and daily rain gauge data and improved rainfall intensity distribution than the earlier 3B42V6.

Daily reference evapotranspiration (ET_0)was calculated using the FAO Penman-Monteith method (Allen et al., 1998)from the 0.312 degree (~38 Km)gridded meteorological variables (daily maximum temperature, daily minimum temperature, daily wind velocity, daily average relative humidity, and daily average solar radiation) from the NCEP Climate Forecast System Reanalysis (CFSR) data for 1998 - 2008.The standard normal probability density functions and standard normal cumulative density functionsof ET_0 for the three sub-basins (Figure 2)demonstrate that the CFSR data adequately representsthe influence of elevation.

2.3 The hydrological model, HySIM

HySIM is a continuous, daily, conceptual rainfall-runoff model that has been extensively used in mountainous catchments including in climate change studies (Wilby, 2005). The HySIM model contains two sub-routines for simulating the river basin hydrology and channel hydraulics. The hydrology in each sub-basin is simulated using seven stores representing snow, vegetation, soil layers, unsaturated and saturated zones,while the channel hydraulic sub-routine uses kinematic routing flows within sub-basins. Details of themodel parameters are given in Pilling and Jones (1999). The empirical degree-day approach is used to calculatedaily snowmelt or accumulation. HySIM uses precipitation, ET_0 and a temperature-based snowmelt modelto simulate stream-flow.

The Beas river basin was sub-divided into three sub-catchments on the basis of the river network and elevation (Figure 1) to appropriately represent the catchment structure / rivernetwork in such a conceptual model. These represent the glacier and permanent snow-covered areas of the basin (Upper), the seasonal snow cover areas (Middle) and the remaining lower elevation parts of the catchment that receive rainfall-only (Lower). The upper, middle and lower basins areas are 5720, 3440 and 3350 km², respectively. Areal averaging was used to transform the precipitation and evapotranspiration data from the different resolution grids.

Soil parameters were initialized based on spatially-weighted values from the Harmonized World Soil Database (HWSD) (FAO/IIASA/ISRIC/ISSCAS/JRC, 2012) although model values for the soil hydraulic parameters were calibrated.

In this study, seven parameters were calibrated- Rooting depth (mm) [RD], Permeability - horizon boundary (mm hr^{-1}) [PHB], Permeability - base lower horizon (mmhr^{-1}) [PBLH], Interflow rate within the upper soil layer (mm hr^{-1}) [IU], Interflow rate within the lower soil layer (mm hr^{-1}) [IL], Snow temperature threshold (^{o}C) [ST] and Snow melt rate for each degree of temperature above the threshold (mm $day^{-1}$$^{o}C^{-1}$) [SM]. The permanent Himalayan snow and ice cover within the upper sub-catchment in the model was initialized with an ice/snow depth of 25 m informed by past research (Kulkarni et al., 2005; Kulkarni and Karyakarte, 2014). HySim uses the commonly applied empirical degree-day approach to simulate snow melt and accumulation (Pilling and Jones, 1999). In this model, when the mean air temperature (T) falls below Snow Threshold (ST) parameter, any precipitation will be assumed as snow and added to snow storage. Similarly daily potential snowmelt (M_s, mm/day) will be released from snow storage when the mean air temperature (T) exceeded base Temperature (T_b, assumed as zero ^{0}C), as indicated by Snow melt rate parameter (SM, this is similar to degree-day factor) based on equation [M_s= SM x (T – T_b)]

2.4Methodology Adopted

The methodology adopted in this study is shown in the Figure 3

1. Two separate HySIM models were built using different assumptions as to the baseline behaviour of the glacier and permanent snow-pack in the Upper sub-basin.

a. **Model Assumption 1 –constrained snowmelt [glacier and permanent snow-packare in steady state]**

In this model, the annualsnowmelt in the Upper sub-basin is limited to the seasonal snow accumulation so that there is no net loss of snow and ice reserves over the baseline period.

b. **Model Assumption 2–unconstrained snowmelt [glacier and permanent snow-pack are in an unsteady state]**

In this model, the snowmelt is only energy-limitedso that there can be net loss (or gain) of snow and ice reserves over the baseline period.The model was initialised with a snowpack depth of 25m in the Upper sub-catchment to ensure that annual snow melt is not limited by snowpack availability.

2. Both modelshad a two year warm-up period (1998-1999) and then wereindependently calibrated,by modifying the selected parameter values, against observed discharge data (2000 –2004, inclusive) and validated (2005 –2008, inclusive).The automated optimisation procedure in HySIM, using the reduced error of estimate (REE) technique, was used to generate one hundred parameter sets for each model within the parameter uncertainty range. The hydrological performances were then evaluated by comparing the simulated and measured daily dischargeusing the Nash-Sutcliffe Efficiency criterion (NSE -Nash and Sutcliffe 1970) and the Percent Bias (PBIAS)goodness-of-fit measures,as recommended by Moriasi et al (2007)
3. After identifying the two parameter sets which provided the best calibration/validation performance for the two contrasting models, the parameter values in both models were fixedfor all future simulations.
4. A climate 'scenario neutral framework' was set up that assesses the catchment response to a plausible range of future climate changes, whilst avoiding the

application of time varying GCM/RCM scenarios simulated under particular assumptions of social/economic/environmental policies (Prudhomme et al., 2010; Remesan and Holman, 2015). Ranges of future annual temperature and annual precipitation changes were informed by the regional summary results from 25-39 GCMs given in Christensen et al. (2013). Six temperature change factors between $\Delta T=0°C$ and $\Delta T=+5°C$ (in steps of 1^oC) and seven precipitation change factors from $\Delta P= -10\%$ to $\Delta P= +20\%$ (in steps of 5%) were used. Each absolute temperature change factor was added to the historical NCEP data to provide modified temperature and subsequently ET_0 time series, assuming all other weather variables were unchanged. The relative changes in precipitation were applied to the TRMM historical time series.

5. The two calibrated/validated HySIM models were then initialised with two future contrasting snow/glacier scenarios:
    a. **Medium term scenario(M)**: Permanent snow/ice reserves persistin the Upper sub-basin, with future snowmelt unconstrained by snow and ice reserves;
    b. **Long term scenario(L)**: Insignificantpermanent snow/ice reserves remain in the Upper sub-basin so that future annual snowmelt is limited to the seasonal accumulation of snow

We selected these two scenarios to represent two realistic but distinct behavioral systems for the Himalayas - a nearer-term scenario (which we have termed "Medium") in which the permanent snow/ice reserves persist in the Upper sub-basin and a longer term future scenario in which there are insignificant permanent snow/ice reserves remaining in the Upper sub-basin Given the major uncertainties in the future temporal responses of the Himalayan glaciers to climate change, we have deliberately not ascribed time horizons to our two scenarios. Where appropriate, we have used the notations of 1M, 2M, 1L and 2L to denote the different baseline model assumptions and future scenarios.

6. The four model builds (each combination of calibrated/validated Model 1 and Model 2 with the Medium and Long term future snow-pack scenarios) were then each run for the forty two combinations of changed temperature (6) and precipitation (7), withImpact Response Surfaces (IRS) produced of the changes in daily low flows [Q90] and high flows [Q10] under these plausible future changes in temperature and precipitation,compared to those from 2000-2008 (i.e. with zero temperature/precipitation change).The daily flow that is equalled or exceeded 10 and 90% of the time (annual high flow (Q10) and annual low flow (Q90)), respectively, are commonly used in designing hydropower projects (IITR, 2011).

3. Results and Discussion

This section presents and discusses the effect of the conceptual model structure assumptions ofhistorical snow packbehavior on the calibrated parameter values and baseline hydrological performance of the two models in the Beas river basin.Thentheinfluence of the resultant baseline parameterization on simulated climate change impactsinthe medium and long term snow-pack scenariosare presented and discussed. It is seen that, the glaciers in the Himalayas (excluding the Karakorum) are believed to be more sensitive to future warming than those in the Tibetan Plateau due to seasonal change of snow accumulation (Yao et al., 2012).This is consistent with the estimated current relative rates of loss in the glaciers of Himachal Pradesh (Kulkarni et al., 2007; Yao et al., 2012).

3.1 How do model parameter values and baseline performance change with assumptions on historical snow-pack behaviour?

The time series of observed and modelled flow for both constrained and unconstrained models are given in the Figure 4, which demonstratesthat both models can describe the observed Beas river basin hydrology with the contrasting snow pack assumptions. Thecalibration and validation results of the simulated daily discharge inTable 1 shows that both models give almost identical NSE values for both periods, which fall within the "Very good" model performance range of Henriksen et al. (2003) for the calibration period and on or within 0.01 of the boundary between "Good" and "Very good" model performance for the validation period. There is greater divergence with the results for PBIAS, which are "Excellent" for Model 1 [constrained snow melt] for both periods but "Very Good" (calibration) and on the "Good" boundary (validation) for Model 2 [unconstrained snow melt]. However, the Q-Q plot for 2000-2008 in Figure 5shows thatboth models under-predict the extreme high flow and low flow events in the Beas River, although this is likely to partially represent the limitations of the TRMM precipitation data.The probability of exceedence curves of observed flow, unconstrained modeled flow and constrained modeled flow are given in the Figure 6 which demonstrates that both models reproduce the observed distribution of daily flows reasonably well.

The slight performance disparity between Model 1 and 2 is related to the model assumptions.The differences in the calibrated parameter values between the twomodels (Table 2) indicate that the calibration process has sought to compensate for the hydrological consequences of the differing snow-pack behaviour assumptions between the models, given that input data uncertainties(e.g. in the TRMM daily precipitation data) andmodel structural deficits are constant across the models. However, hydrologicalmodellers often assume that modelling performance issues are substantively caused byinadequacies in the quantification of model parameters themselves, rather than inadequacies inthe input data or data assumptions; but this is not always the case (Mukhopadhyay and Dutta, 2010; Remesan and

Holman, 2015). In such cases, catchment models may have unrecognized limitations in simulating stream flow under scenarios of climate or land-use change that are outside of the calibration conditions, due to what might be thought of as compromised parameter values.

3.2 How do simulated future climate change impacts vary as a consequence of the baseline snow-pack assumptions?

A common assumption in hydrological modelling is that a successfully calibrated and validated hydrological model contains process representationthat is suitable for climate change impact studies. This ignores thepotential expansion of modelling uncertainty associated withthe varying assumptions on initial boundary conditions and resultant parameter values evident in Table 2.

Impact response surfaces have therefore been constructed to understand how the differing parameterization resulting from the contrasting baseline snow-packassumptions translates into uncertainty in the future hydrological response of the river Beas to a plausible range of changes in future annual precipitation and temperature change to 2100 for the region (Christensen et al. 2013) under medium term (continued presence of permanent snow-pack) and long term (loss of permanent snow pack) conditions.

3.2.1 Medium term impacts

Both models simulate increasing average annual discharge with increasing temperature and precipitation, with the temperature increase of +5^{o}C (through its influence on snowmelt) being more able to offset the simulated annual precipitation decrease of -10% (Table 3).The change in average annual discharge is generally larger with Model 2 (unconstrained snow melt), with the difference between the two models ranging from0.6-1.5 x 10^9 m^3/yr (compared to an observed average annual discharge of 7.8 x 10^9 m^3/yr), whilst the percentage change difference between the models is up to 45% of the simulated baseline [with ΔT= 0° C, ΔP= 0 %]. Though it hasnot been considered in this study, it is noted that shrinkage in glacier area and melt duration would impact glacier melt runoff as it can connect to a function of contributing glacier area and melt rate (Bliss et al., 2014).

Figure 7 shows that there is broad similarity in both the Q10 impact response surfaces when the two models are applied to medium term conditions (i.e. 1M and 2M settings) in which future snowmelt is only energy-limited due to a continuing permanent snow-pack. In general Q10 increasesin both models with increasing temperature as there is increased snowmelt of the permanent snow/ice reserves over a longer period, although there is a tipping point in Model 1(constrained snowmelt) at around a temperature increase of 1°C above which increases in snowmelt outweigh the effect of increased actual evapotranspiration. Q10 also tends to increase with increasing precipitation with a similar sensitivity in both models, although the effect decreases with increasing temperature from a range of around 40% of the baseline Q10 for ΔT= +0°C to 30% for ΔT= +5°C. Q10 for the unconstrained snowmelt model (Model 2and 2M setting) is more sensitive to changing temperature and precipitation than the constrained snowmelt model (Model 1and 1M setting),exhibiting generally larger changes (up to an increase of 79% of the baseline Q10, compared to 40% with Model 1) and a larger range of responsesacross the scenario-neutral climate space (of 90% of the baseline Q10, compared to 58%).

However, there are very significant differences in the responsesin Q90 between the two models. In Model 2 (unconstrained snowmelt, 2M setting), Q90 increases with increasing temperature and precipitation across the scenario-space by up to 133% of the baseline Q90 due to the increasing duration and magnitude of snow/ice melt, although there is a minor inflection line around ΔT= +1°C. With small temperature increases (<1°C) and a decrease in precipitation, Q90 decreases by up to -12% arising from decreased recharge due to increased actual ET not being offset by increased snow/glacier melt or precipitation.

In contrast, warming leads to general decreases in Q90 in Model 1 (constrained snowmelt, 1M setting) across most of the scenario spaceby up to -73% of the baseline Q90, despite the increase in average annual discharge (Table 3). This arises due to the effect of the calibrated parameterizationon the interplay between a number of hydrological processes within the sub-catchments (Table 2). Firstly, the snow melt rate in the Upper catchment is lower in Model 1 compared to Model 2 (1.0 compared to 1.2 mm/^{o}C/d) leading to smaller increases in dry season snowmelt from this sub-catchment. Secondly, seasonal snow accumulation in the Middle catchment melts sooner and quicker (due to a lower snow melt threshold and higher snowmelt rate) exhausting this seasonal reserve before the dry season, but the potential for increased recharge into the groundwater store to support dry season flows is limited by the smaller lower boundary permeability which leads to increased lateral flow. Finally the Middle catchment in Model 1 has a greater calibrated rooting depth of 2.2m (compared to 1.2m in Model 2),so that more water can be extracted from the soil moisture reservesto meet evapo-transpirative demand leading to reduced recharge and therefore baseflow.

3.2.2 Long term impacts

Across the entire range of temperature and precipitation changes, the long-term future average annual flows are lower than in the medium term for both models [data not shown]. This arises as the volume of snowmelt is limited to the seasonal winter precipitation in the Upper and Middle catchments (as there is no diminishing permanent snow-pack as in the Medium term future) with the result that the increased temperature increases the evapotranspiration and thus decreases rainfall-runoff and recharge. Both models (i.e. 1L and 2L settings) show an increasing average annual discharge with increasing precipitation (compared to [ΔT= 0 °C, ΔP=0%]), the magnitude of which decreases with increasing temperature (in contrast to the medium term response) (Table 3). There is a greater range of changes in average annual discharge across the scenario-neutral space with Model 2 (unconstrained snowmelt), with the difference between the two models in the magnitude of the simulated change from the baseline [ΔT= 0 °C, ΔP=0%] ranging between around 650-1000 x 10^6 m^3/yr, whilst the percentage change difference between the models is only up to 15% of the baseline. Both of these impact uncertainties between the two models were lower than in the medium term future.

There are much more similar hydrological responses in low and high flows for the long term future across the scenario-neutral space between the two models (Figure 8), than for the medium term impact response surfaces (Figure 7). Q10 and Q90 both decrease with increasing temperature and decreasing precipitation, as rainfall and seasonal snow accumulation decrease and evapotranspiration increases. As the snowmelt in both the Upper and Middle catchments are limited to the seasonal snow accumulation, the largest decreases in Q10 of -50% (Model 1, 1L setting) and -62% (Model 2, 2L setting) are associated with the lowest precipitation (ΔP= -10%) and highest temperature increase (ΔT= +5°C). Similarly the largest increases in Q10 of +41% (Model 1) and +51% (Model 2) are associated with the greatest precipitation (ΔP= +20%) and lowest temperature increase (ΔT= +0°C). Under the highest temperature scenario (ΔT= +5°C), changes in Q10 range from the most optimistic changes of -2% (Model1) to -9% (Model 2) for ΔP= +20%, to the most pessimistic changes of -50% (Model1) to -62% (Model 2) for ΔP= -10%.

As would be expected with the loss of the permanent snow/ice cover in the Upper sub-basin, Q90 decreases with increasing temperature, although Q90 can increase under certain conditions of increasing precipitation. For the most pessimistic combination [ΔT= +5°C, ΔP= -10%], Q90 decreases by between -68% (Model1) to -74% (Model 2). This decrease arises from the earlier loss of the seasonal snowmelt contribution (as the duration of the melt period shortens) so that river discharge becomes reliant on the groundwater baseflow. However, the increased evapotranspiration associated with warming leads to decreased recharge in the lower and middle sub-basins and thus to further reduced baseflow. In contrast, the hottest and wettest scenario [$\Delta T= +5^oC$, $\Delta P= +20\%$] leads to increases in Q90 of between 42% (Model1) and 13% (Model 2) as the increased precipitation leads to increased recharge during the monsoon period when soils are at field capacity, outweighing the effects of the increased evapotranspiration during that period.

The long term uncertainty introduced by the two baseline snow-pack behaviour assumptions across the scenario-neutral climate space ranges between -14 to +13% for Q10 and 5 to -31% for Q90. This is about 10% less than the medium term uncertainty in Q10 (of 36% of the baseline Q10) and reflects the limitation imposed on the hydrological response of the two models by the loss of the permanent snow/glacier reserve and the influence of the limited seasonal snow accumulation.

3.3 Implications

The modelling results presented have demonstrated that the long term transition to a Himalayan river whose hydrological response is dominated by rain and the shorter-duration melt of seasonal snow will lead to significant temporal changes in the water balance and flow dynamics. Here, although both of the models show this gross difference between the two future periods, there are important differences in simulated future hydrological response that arise as a consequence of their different baseline parameterization (Table 2). Figure 9 shows how the differences in calibrated parameter values associated with the differing assumptions of baseline snow pack behaviour impact on the change in the flow duration curves for four selected future climates. Also shown is the uncertainty range for the two models for the historical period. The selected scenarios are [ΔP=-10%, ΔT= 0^{o}C] (highest reduction in precipitation and no change in temperature), [ΔP=-10%, ΔT= 5 ^{o}C] (highest reduction in precipitation and highest increasein temperature), [ΔP=+20%, ΔT= 0 ^{o}C] (highest increase in precipitation and no change in temperature), [ΔP= +20%, ΔT= 5 ^{o}C] (highest increase in precipitation and temperature from baseline climate). These show that precipitation changes in the absence of temperature increases tend to lead to the similar percentage changes in discharge throughout the flow duration curve; whereas temperature increases cause the largest increase in discharge around the 25thto50thexceedance probability reflecting increases in pre- and post-monsoon snowmelt. However, the largest differences between the responses of the two models to the same change in climate, arising from the assumptions regarding the historical snow-pack behaviour, occur in the medium term future between about the 60th and 100th exceedance probability reflecting the different contributions of snowmelt to dry season flows.

The assumptions regarding the historical snow pack behaviour lead to an uncertainty between the two baseline models across the historical flow regime that is less than 30% of the observed baseline flow (with the exception of the 97th – 100th exceedance probability). When the future discharge uncertainty across the range of flow exceedance between the two models for each of the two future periods is compared with this, it is apparent that the future uncertainty only exceeds the historical uncertainty for the medium term under the warming scenario space, indicating that the magnitude of the baseline model uncertainty is not conserved but is magnified under such futures. In contrast, in the longer term, there is no general expansion of the uncertainty due to the behavioural constraints imposed by the lack of a permanent snow/glacier. The results show that, giventhe current uncertainty in spatio-temporal dynamics of the glaciers and permanent ice and snowfields, the choices that a modeller makes in the baseline model build can lead to differences in the simulated magnitude of future changesin average annual discharge and hydrological response that are significantly larger than the baseline uncertainty.

The result show an average loss of -0.37m/yr snow water equivalent (unconstrained Model1) and average gain of +0.005 m/yr snow water equivalent (constrained Model 2) during 2000-2008 (negative indicates mass loss and positive values indicates mass gain). The total simulated mass balance loss during 2000-2008 period is -3323mm with uncertainty bounds of [-131 to -691 in mm/year snow water equivalent corresponding to the year 2005 and year 2008 respectively] for the unconstrained model. Similarly the constrained model (with no permanent snow assumption) has shown a total mass change of +46 mm (i.e. seasonal accumulation) during 2000 – 2008 periods with uncertainty bounds of [-37 mm/year to +50 mm/year snow water equivalent for the year 2002 and year 2004 respectively]. The losses are comparable with other studies in the Himalayan region e.g. -0.89m/yr in Shaune Garang basin for 2001-2008 (Kumar et al., 2016), -0.44± 0.09 m/yr for Lahaul and Spiti region for 1999-2011 and the regional review of Pritchard (2017). The stream-flow contribution from snow/ice-melt is shown in the Figure 10 for both constrained and unconstrained model. One can clearly see from this figure that unlike the unconstrained model with permanent snow assumption (i.e. Model 2), the snow/ice contribution from the constrained model is relatively less during low flow periods but compensated with other water balance components. In the case of the Model 2 simulation, the total snow/glacier melt is more than in the Model1 simulation because of our conceptual assumptions.The water balances for theconstrained and unconstrained model simulation arerepresented in the Figure 11 indicating how surface runoff and low AET are compensating for the effect of the lack of a permanent snow packin the constrained model. For better understanding of major water balance gains and losses in Beas Basin with different model assumptions are given in the Figure 12. It is clearly seen that maximum snow pack melting is occurring in the hot monsoonseason (June, July and August months). Variations in soil moisture changes (in the upper and lower layer) and differences in monthly groundwater gains/losses are clearly observed in these figures with compensation effect on AET on both constrained and unconstrained models.

Given the focus of our study on the model response to the differing baseline snow packassumptions alone, it must also be acknowledged that there are many other sources of uncertainty which affect the impact of climate change on the hydrological response of glaciated mountainous river basins. Firstly, the simulated changes in flows resulting from the effect of increased temperatures on snow/ice melt in our two extreme cases donot take into account the change in glacier area which may lead to over-estimation of the snowmelt-temperature response, although our results are consistent with the increases in annual average streamflow of 10% and 18% under warming scenarios of 2^{o}C and 4^{o}C (with no change in precipitation) simulated by Nepal et al. (2014) in the Himalayan DudhKosiriver basin. Secondly,although there are continuing debates regarding the temperature index approach to calculating snow melt and accumulation (e.g. Hock, 2003; Pellicciotti et al., 2012; Mackintosh et al., 2017), it continues to be a widely used method, particularly in data-poor basins.Thirdly, although more sophisticated methods exist, the change factor method used to perturb the historical climatology is still widely used in impact analysis studies (e.g. Anandhi et al., 2011; Fatichi et al., 2015). Finally, the limited duration comparison of9 years of baseline simulation against 9 years of scenario analysis, due to TRMM V7 and flow data availability, is acknowledged.

4. Conclusions

The paucity of data and the uncertainty in understanding the snow/ice resources and their spatio-temporal dynamics in poorly instrumented mountainous regions remains a significant challenge for conceptual hydrological modelling in snow and glacier-dominated catchments.

This often necessitates model assumptions to be made regarding initial snow/ice boundary conditions. This study evaluated how contrasting baseline permanent snow pack assumptions in a semi-lumped conceptual model affectedsimulated hydrologicalresponse to changes in future climate. Two conceptual model settings in which the glacier and permanent snow pack are either in steady state (Model 1, in which annual snowmelt is constrained by the seasonal snow accumulation so that there is no net loss over the baseline period) or are not in equilibrium (Model 2, so that snowmelt is only energy-limited and there can be a net loss or gain over the baseline period) were calibrated and validated in the mountainous Beas River catchment in northern India. The findings from the study are that:

1. Despite the contrasting assumptions about historical snow-pack response to baseline climate variability, both models hadgenerally very good and similar baseline hydrological performance. The compensation effect of parameter fitting wasdemonstrated to obscurethe effect of uncertainboundary conditions on process and model behaviour.
2. The effect of differing parameterisation on baseline hydrological process behaviour is hidden, so that models can have unrecognised limitations in simulating stream flow under scenarios that are outside of the baseline conditions. We havedemonstrated that the potential inadequacies of such resultant parameters values only become evident when the models are subjected to changing climatic conditions.
3. Across a scenario-neutral climate change space of changing annual temperature of 1-5^{o}C and changing annual precipitation of -10 to +20%, the model parameterisation under the two baseline snow-pack behaviour assumptions introduced medium term discharge uncertainty across the scenario-neutral climate space of between36% (ΔQ10 of -3 to -39%) for Q10 and 163% (ΔQ90 of -141% to +22%) for Q90.These uncertainty ranges obtained from the difference between Model 1 and Model 2 during

9 year scenario analysis are significantly greater than the modelled baseline uncertainty of around 30%.

4. Under longer term changes associated with the loss of the permanent snow/glacier, the uncertainty across the scenario-neutral climate space amounts was between 28% (ΔQ10 of -15% to +13%) for Q10 and 37% (ΔQ90 of -31 to +6%) for Q90.
5. Although the uncertainty arising from the baseline parameterisation reduces with the transition from a melt-dominated river to a rain-dominated river, associated with the loss of area and volume of permanent snow/ice-pack resources, the significant temporal changes in flow dynamics and magnitudes (Nepal, 2016) will have significant impacts on simulated future dry season water resources for irrigation, hydropower and livelihoods;

6. In this particular case study, although the NSE values are similar for both assumptions, we acknowledge the differences in the PBIAS values (especially in the validation phase). Nevertheless, our study suggests that, where there is considerable uncertainty in historical snow-pack reserves and dynamics, an ensemble of hydrological model-builds calibrated to the different assumptions should be used to inform the understanding of the resultant effect of parameter biases on climate change impact studies.

5. Acknowledgment:

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6. Conflict of Interest

None

7. References:

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Table 1: Statistical indices of hydrological model performance in the Beas basin

Calibration (2000-04)					Validation (2005-08)				
Year	Model 1 [Constrained snowmelt]		Model 2 [Unconstrained snowmelt]		Year	Model 1 [Constrained snowmelt]		Model 2 [Unconstrained snowmelt]	
	PBIAS (%)	NSE	PBIAS (%)	NSE		PBIAS (%)	NSE	PBIAS (%)	NSE
2000	2.4	0.82	16.8	0.75	2005	26.1	0.66	39.9	0.50
2001	7.5	0.70	20.9	0.69	2006	7.9	0.69	31.2	0.70
2002	-0.2	0.52	11.9	0.59	2007	-15.7	0.56	14.6	0.72
2003	-15.8	0.71	1.5	0.77	2008	-19.0	0.70	-2.4	0.74
2004	4.9	0.67	0.5	0.69	**Average**	**-0.2**	**0.65**	**20.8**	**0.66**
Average	**-0.2**	**0.69**	**7.54**	**0.70**					
Standard Deviation	**26.1**	**0.66**	**39.9**	**0.50**	**Standard Deviation**	**21.22**	**0.06**	**18.71**	**0.11**

Table 2: Calibrated parameters in Beas basin for the two models

Parameters	Model 1 [constrained snowmelt]			Model 2 [unconstrained snowmelt]		
	Upper Catchment	Middle Catchment	Lower Catchment	Upper Catchment	Middle Catchment	Lower Catchment
Rooting depth [RD](mm)	#[1]	2242	2750	#	1201	2752
Permeability - horizon boundary [PHB] (mm/hour)	3.3	5.3	5.3	26	8	4
Permeability - base lower horizon [PBLH] (mm/hour)	0.8	1.1	1.09	156	163	17
Interflow - upper [IU](mm/hour)	10	8	8	8	6	6
Interflow – lower [IL] (mm/hour)	4	4	4	330	82	82
Snow Threshold [ST] [^{o}C]	0.3	0.3	#	0.6	0.6	#
Snow Melt [SM] [mm/^{o}C/d]	1.0	1.2	#	1.2	0.8	#

[1] # Not relevant to processes being simulated within sub-catchment

Table 3: Change in average annual discharge for the four model-future combinations under a range of temperature (ΔT) and precipitation (ΔP) changes, expressed as (upper) volume change and (lower) percentage change relative to [ΔT= 0ºC, ΔP=0%], and (shaded) differences between the two models

		Model							
		1 [Constrained] **($x10^6$ m^3/yr)**			2 [Unconstrained] **($x10^6$ m^3/yr)**		Difference between models **($x10^6$ m^3/yr)**		
Future		ΔP=-10%	ΔP=+20%		ΔP=-10%	ΔP=+20%		ΔP=-10%	ΔP=+20%
Medium	ΔT=+5ºC	*2116.1*	*4790.6*	ΔT=+5ºC	*3699.5*	*6015.8*	ΔT=+5ºC	*1583.4*	*1225.2*
	ΔT=+0ºC	*-1370*	*2737.5*	ΔT=+0ºC	*-752.8*	*1816.9*	ΔT=+0ºC	*617.2*	*920.6*
		ΔP=-10%	ΔP=+20%		ΔP=-10%	ΔP=+20%		ΔP=-10%	ΔP=+20%
Long	ΔT=+5ºC	*-3424.6*	*1268.3*	ΔT=+5ºC	*-2559.2*	*293.5*	ΔT=+5ºC	*865.4*	*974.8*
	ΔT=+0ºC	*-1671.8*	*3340.1*	ΔT=+0ºC	*-1028.7*	*2346.6*	ΔT=+0ºC	*643.1*	*993.5*

		Model							
		1 [Constrained] **(% change)**			2 [Unconstrained] **(% change)**		Difference between models **(%)**		
Future		ΔP=-10%	ΔP=+20%		ΔP=-10%	ΔP=+20%		ΔP=-10%	ΔP=+20%
Medium	ΔT=+5ºC	*23*	*52*	ΔT=+5ºC	*59*	*97*	ΔT=+5ºC	*36*	*45*
	ΔT=+0ºC	*-15*	*30*	ΔT=+0ºC	*-12*	*29*	ΔT=+0ºC	*3*	*1*
		ΔP=-10%	ΔP=+20%		ΔP=-10%	ΔP=+20%		ΔP=-10%	ΔP=+20%
Long	ΔT=+5ºC	*-44*	*16*	ΔT=+5ºC	*-58*	*7*	ΔT=+5ºC	*15*	*10*
	ΔT=+0ºC	*-21*	*43*	ΔT=+0ºC	*-24*	*54*	ΔT=+0ºC	*2*	*11*

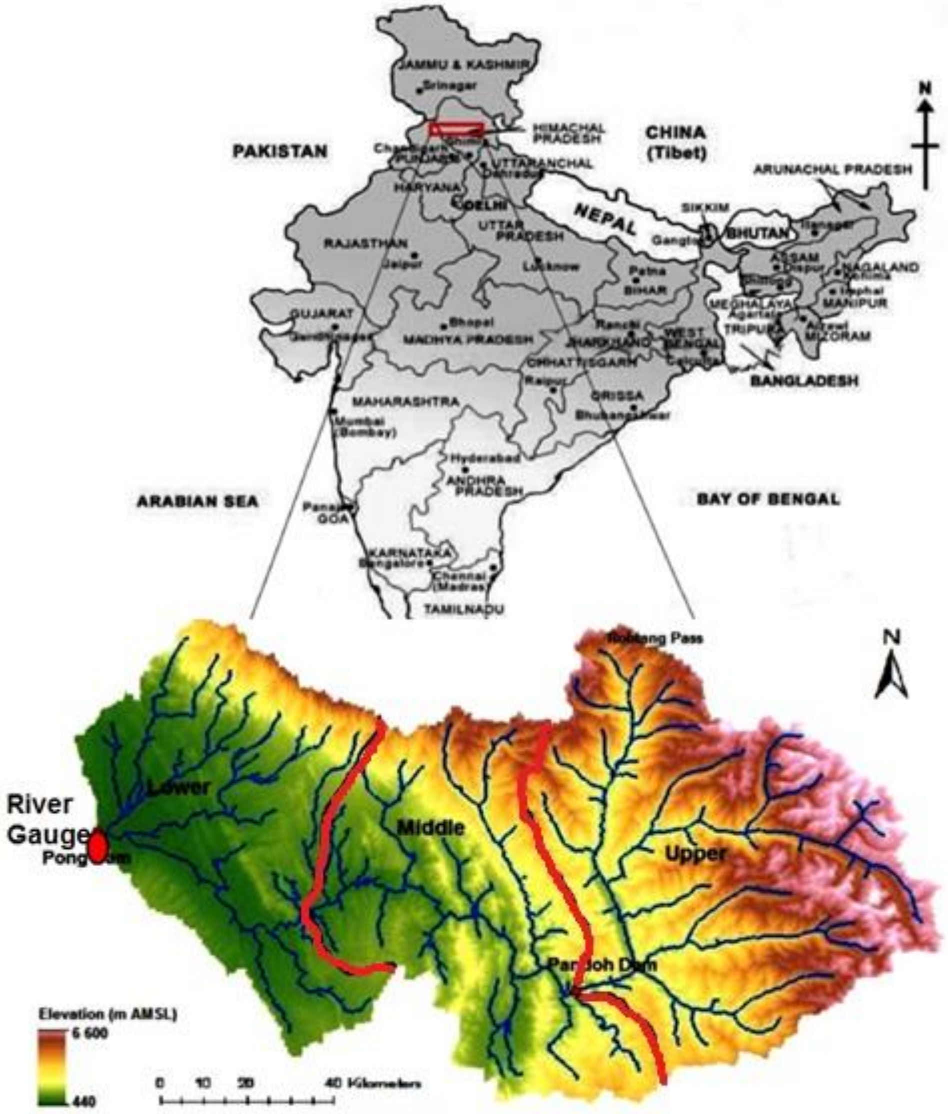

Figure 1. The Beas river basin

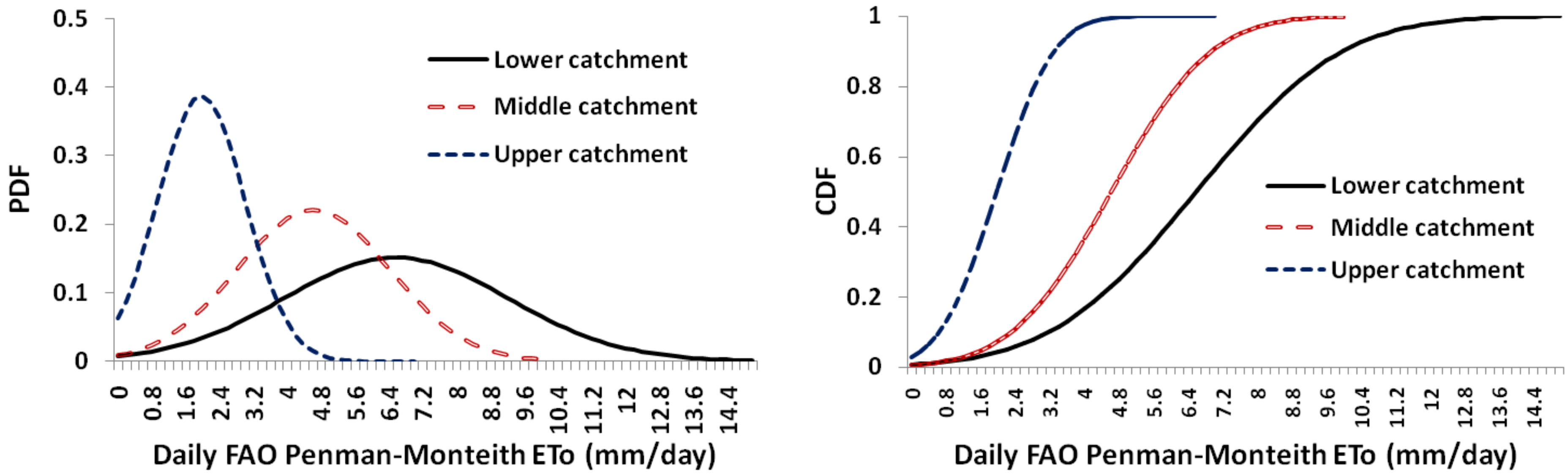

Figure 2: The PDF and CDFs of reference evapotranspiration for three sub-catchments calculated using National Centres for Environmental Protection (NCEP) Climate Forecast System Reanalysis (CFSR) data

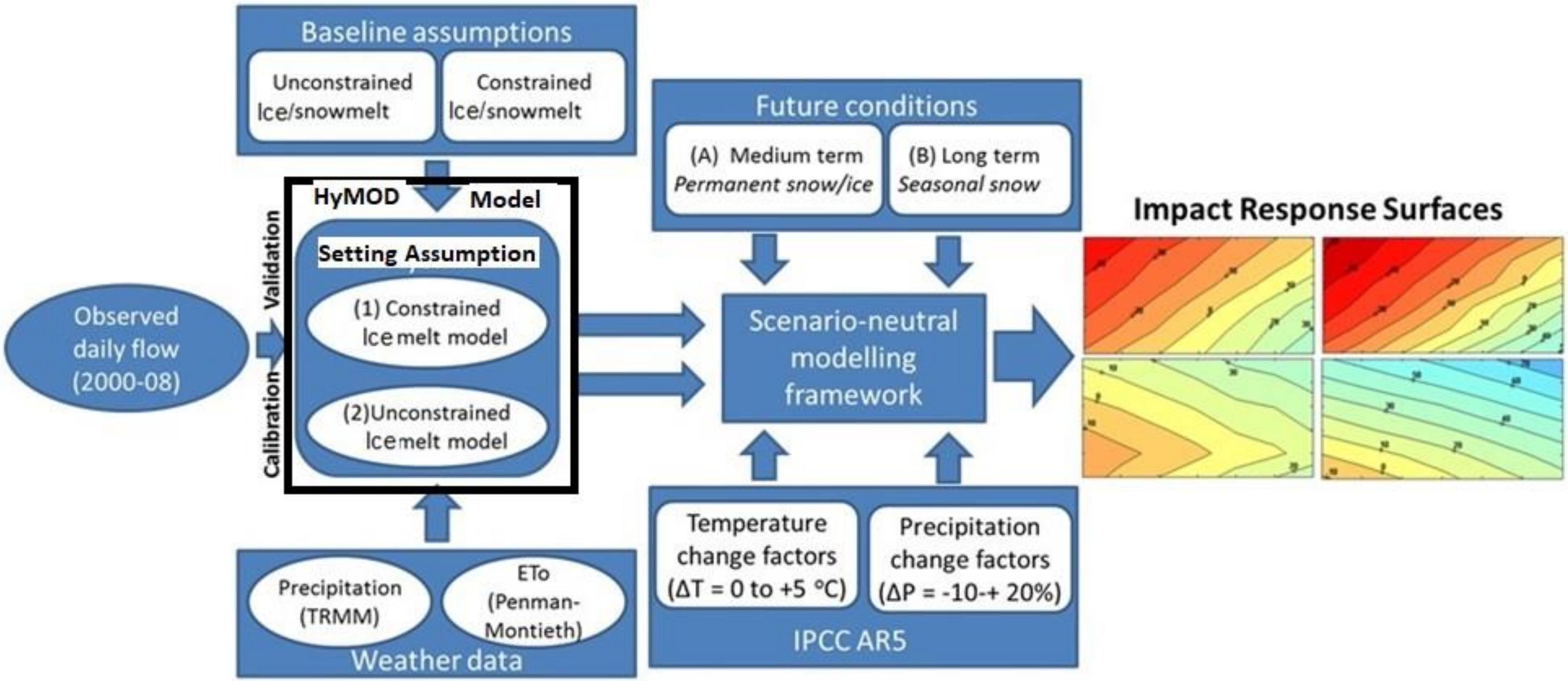

Figure 3: The methodology adopted in this study

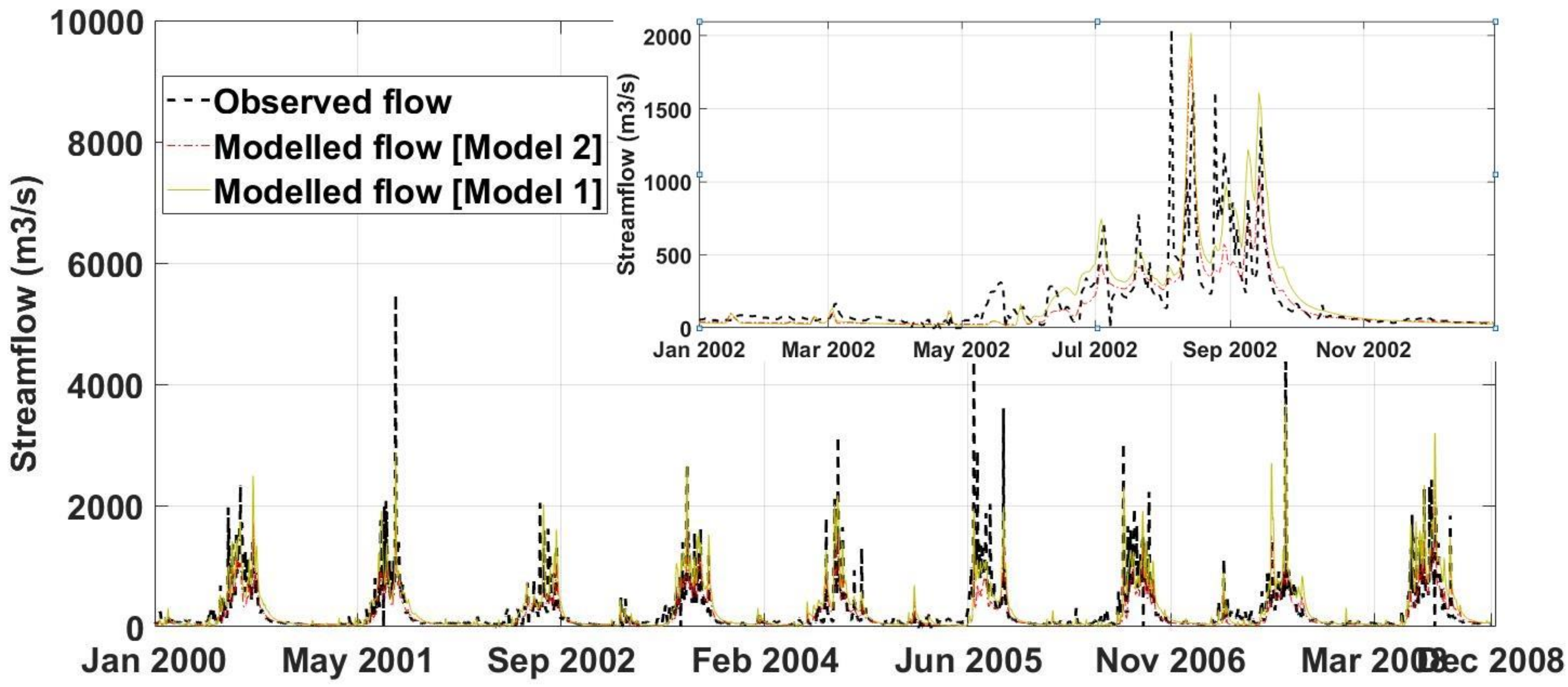

Figure 4: Observed and simulated streamflow using two HySim models with contrasting assumptions regarding baseline snowpack behaviour for 2000-2008 (Year 2002 data is shown as inset)

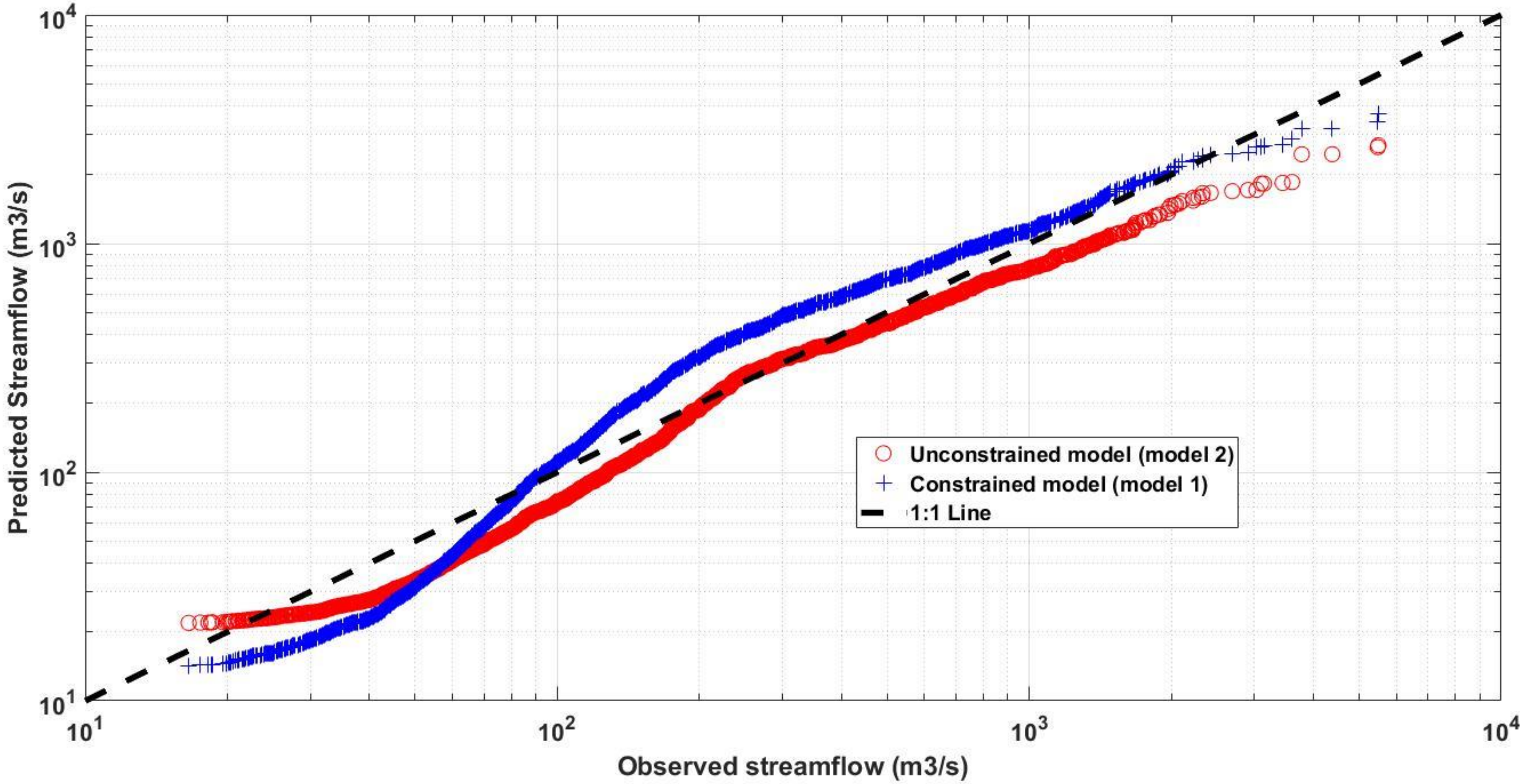

Figure 5: Q-Q plot of the two HySim models with contrasting assumptions regarding baseline snowpack behaviour for 2000-2008

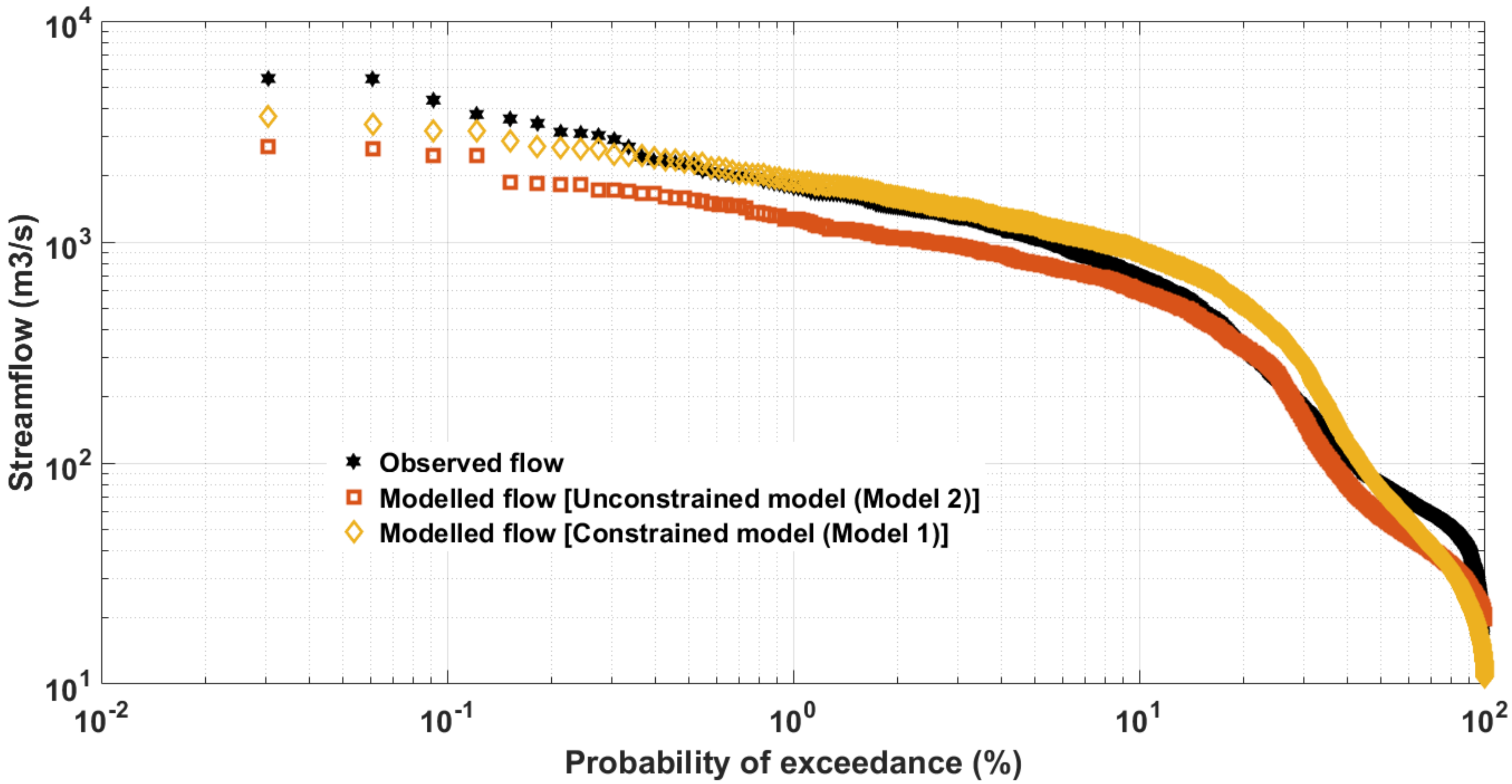

Figure 6: The probability of exceedence curves for the HySim models with contrasting assumptions regarding baseline snowpack behaviour along with observed flow data for total simulation period

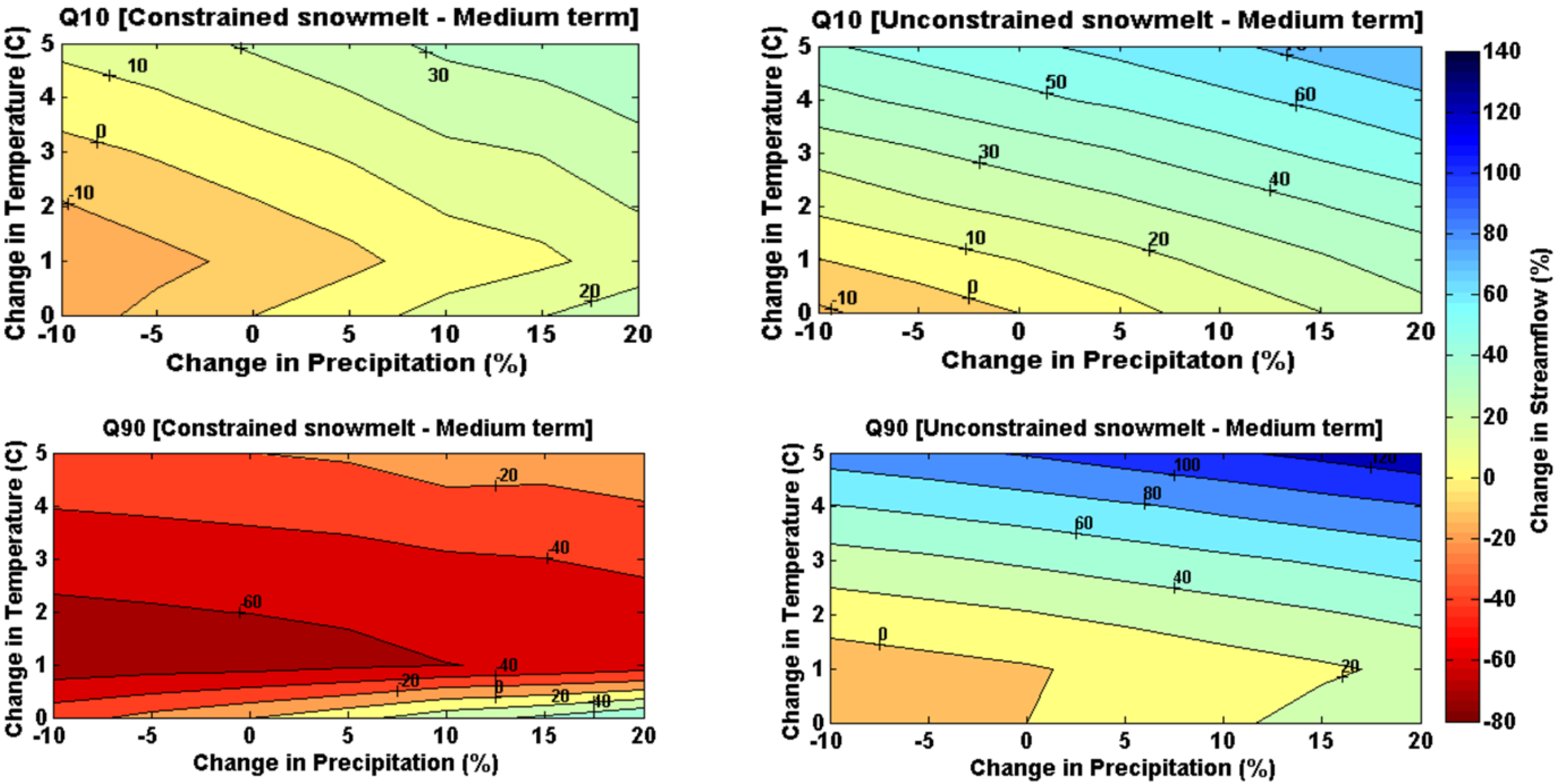

Figure 7: Impact Response Surfaces of the change in simulated (upper) Q10 and (lower) Q90 daily discharge under changed annual temperature and precipitation for medium term future conditions with permanent snow/icepack for models with contrasting baseline snow/ice pack behaviour assumption

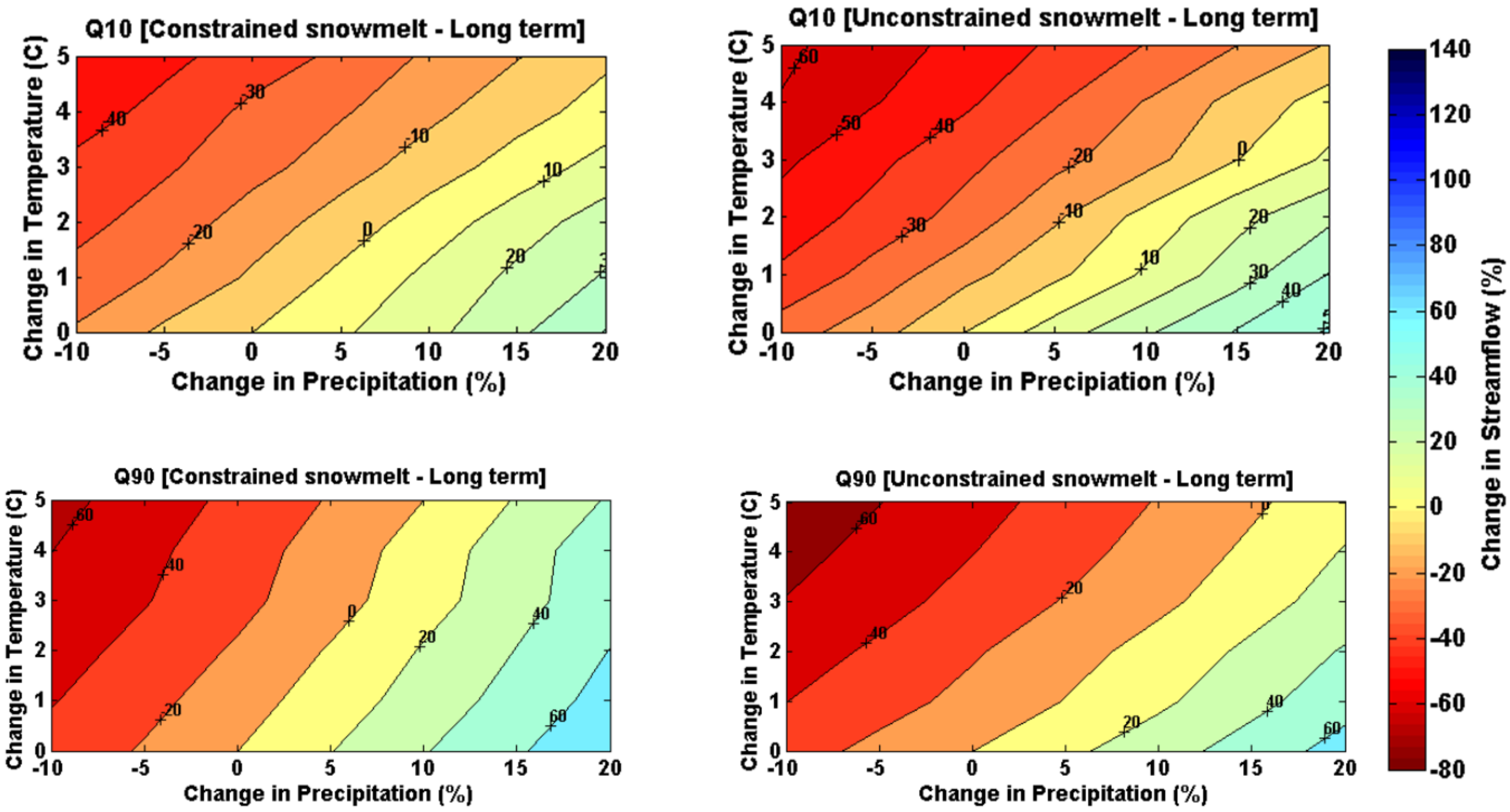

Figure 8: Impact Response Surfaces of the change in simulated (upper) Q10 and (lower) Q90 daily discharge under changed annual temperature and precipitation for long term future conditions with no permanent snow/icepack for models with contrasting baseline snow/ice pack behaviour assumption

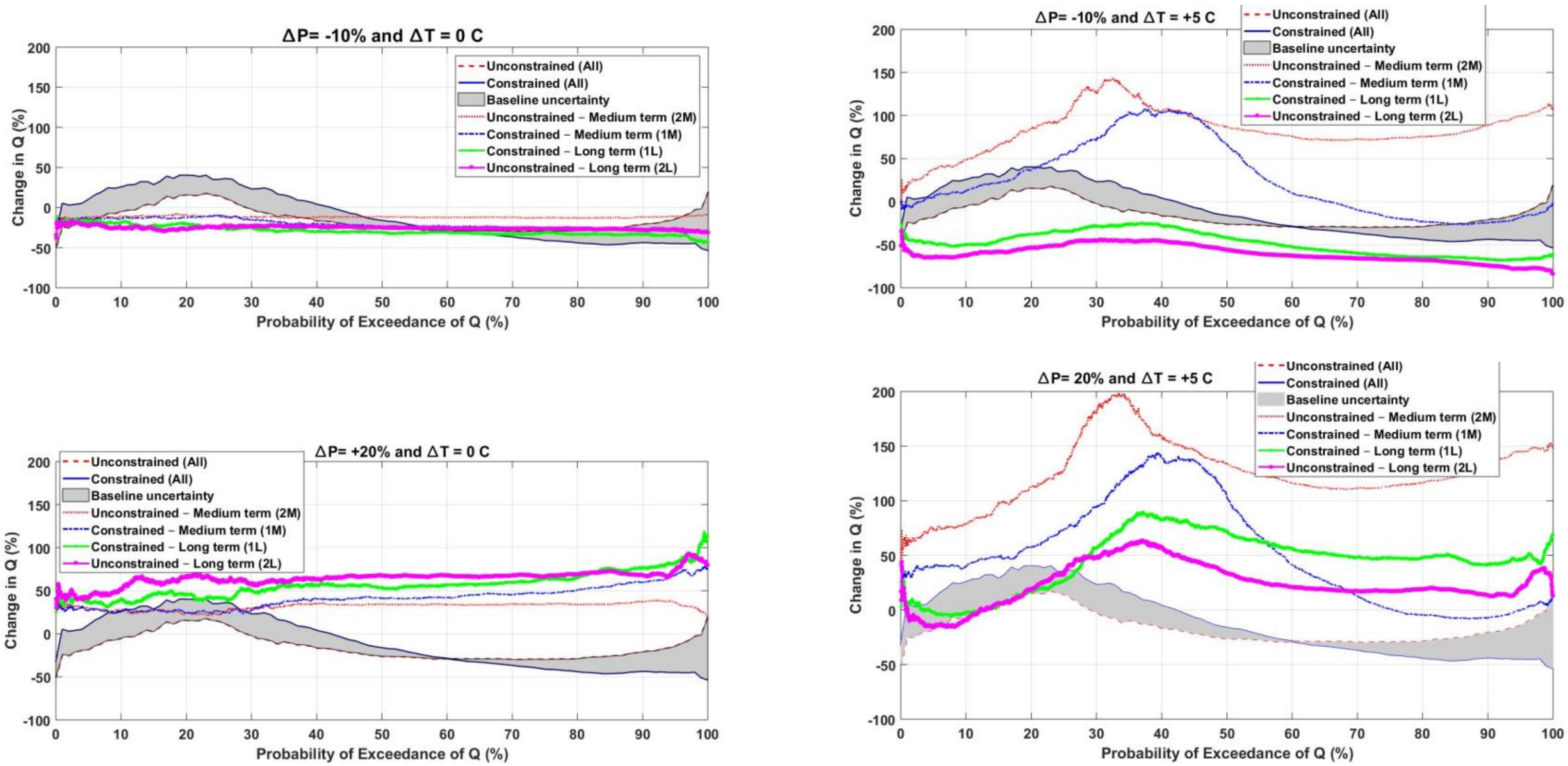

Figure 9: Comparison of medium and long term future impact uncertainty (as given by the percentage change in daily discharge [compared to ΔT= 0°C, ΔP=0%] across the flow exceedance probability for four selected change factor scenarios for the two models) and their baseline model uncertainty. N:B. We have used notations like 1M, 2M, 1L and 2L etc for denoting different models with medium term and long term assumptions.

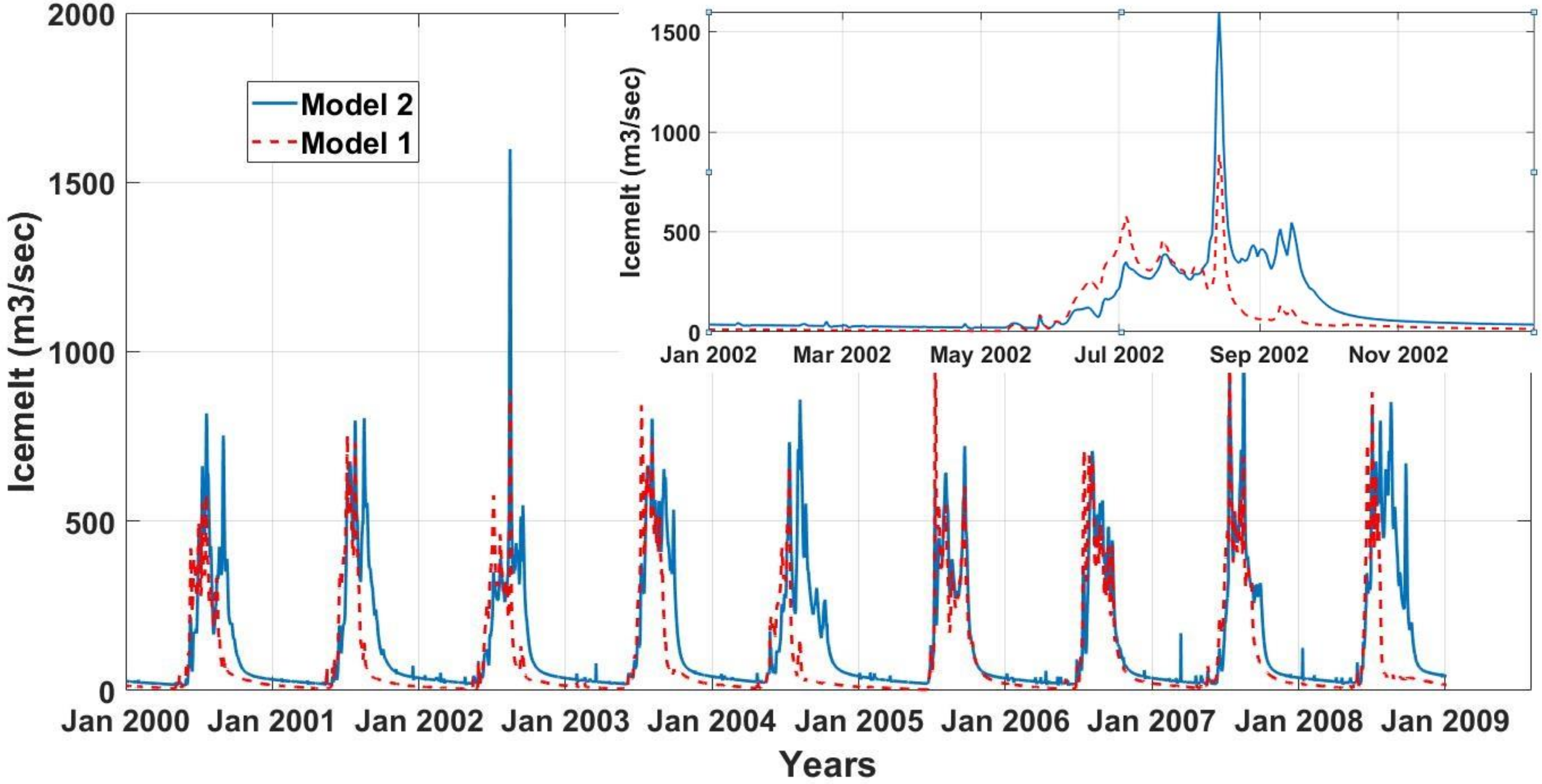

Figure 10: Comparison of snow/ice melt contributions to streamflow in both constrained and unconstrained models in Beas Basin from 2000 to 2008 (Year 2002 data is shown as inset)

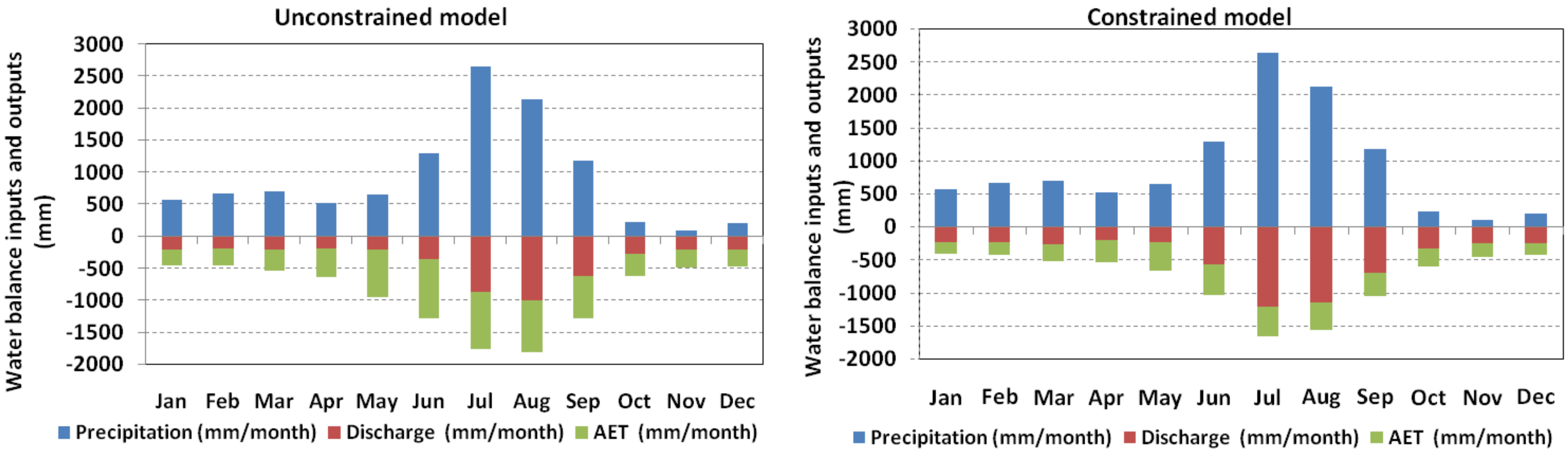

Figure 11: Comparison of major water balance inputs and outputs in Beas Basin for both unconstrained and constrained model simulations (2000 -2008)

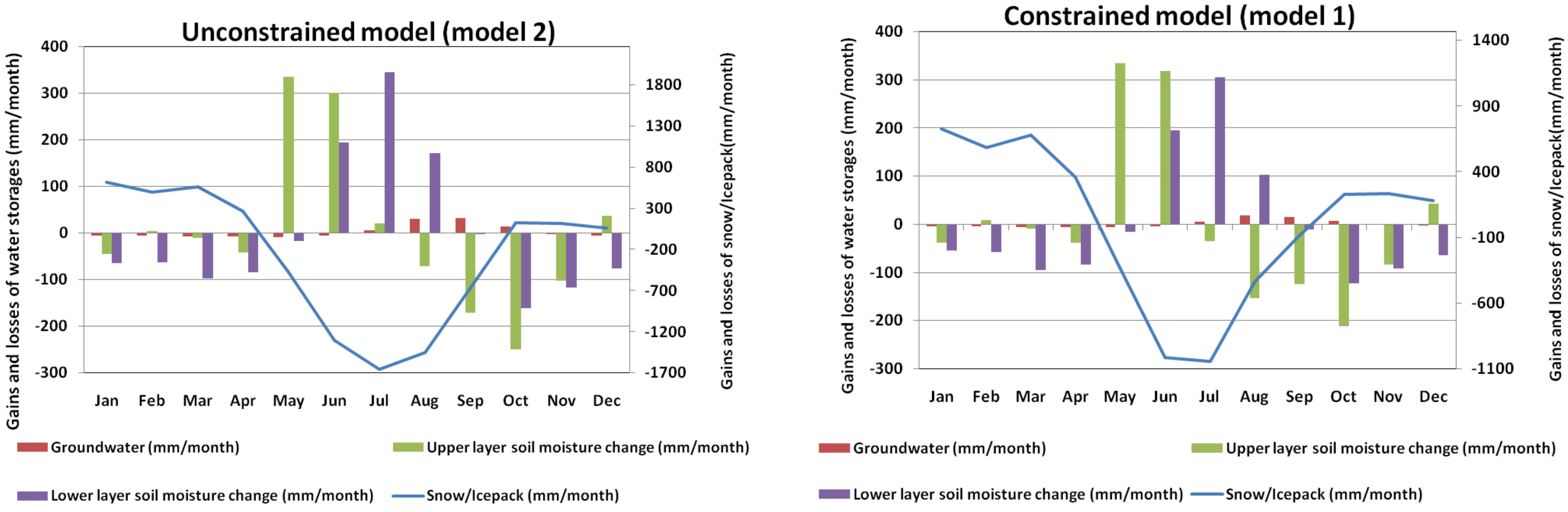

Figure 12: Comparison of major water balance gains and losses in Beas Basin for both unconstrained and constrained model simulations (2000 -2008)