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To
Editioral Board HESS
c/o Dr. R. Woods

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Subject Revision statement hess-2015-128

Our reference
WI/20150827

Dear Dr. Woods

I am pleased to submit a revision of our paper "Reconciling high altitude precipitation in the upper Indus basin with glacier mass balances and runoff".

We have addressed all referee comments (2 x) as well as the short comments (3x) submitted during the open discussion. We have attached our pointwise reply below.

We hope the paper is now acceptable for publication in HESS and that it may contribute to the understanding of the complex hydrology of the upper Indus basin.

With kind regards,

Dr. W.W. (Walter) Immerzeel

Anonymous Referee #1 – Referee comment
Received and published: 19 June 2015

The paper estimates indirectly precipitation by discharge, glacier mass balance and actual evapotranspiration for the Upper Indus Basin. Considering the absence of stations at high altitudes, this work is very interesting. The conclusion is that the current precipitation estimates (with land stations and remote sensing) are strongly underestimated.

Thank you.

Although the topic is of prime importance, I have many perplexity that the paper could be published without:

i) re-writing completely the method section. Currently it is too much hermetic. I am not be able to follow exactly what has been done. I have more doubts than answers. Please provide more details in particularly connected with the uncertainty of data. Please separate sections for precipitation, evapotranspiration, mass balance, equations. . . Please provide supplement information file.

It is not entirely clear to us what the reviewer cannot follow specifically, but we have further elaborated on the method section and provide more details in the revised manuscript. We have paid particular attention to explaining our uncertainty analysis and validation in more detail and we have included additional sub-sections for the different datasets used.

ii) re-writing completely the results and discussion section. Even results and discussion are too much condensated. In general I would like to be more convinced by authors about the findings. I strongly suggest to present detailed tables and/or graphs in which the terms of the water balance are presented as estimation and uncertainty. If it was possible this new analysis should be subdivided for main elevation bands and regions. Furthermore previous estimations (by many other authors) need to be presented and discussed (the authors know well the literature) In general it needs to be clear and convincing how/why the present work overcomes the previous ones. In conclusion I suggest an in-depth analysis of the glacier mass storage that is less convincing than the other analysis.

We have partially rewritten and amended these sections. We have included a water balance analysis for the three zones and we have added Turc-Budyko plots (based on catchment observations) to the manuscript as suggested by Dr Andréassian in the open discussion. We believe this is a significant improvement of the validation part of our paper. We have also placed our findings better in the context of previous studies published by our group and others.

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Anonymous Referee #2 – referee comment
Received and published: 12 July 2015

This paper discusses how to obtain a better estimation of distributed precipitation for the well-known upper Indus Basin by inverting its hydrological balance. By using regionally-averaged glacier mass balance data to estimate precipitation gradients in the area, and a distributed hydrological model to compute accumulation and melt, the authors suggest that the precipitation needed to sustain the observed mass balance is higher than the one observed by ground or ridded products. An evaluation is also provided by using corrected precipitation as an input to estimate average annual runoff for sub-basins.

I think that the topic discussed by this paper is relevant. The idea of inverting the local hydrological balance is an interesting approach to solve the problem of gauges deficiencies (i.e., a sparse distribution and/or instrumental under catch), which is a frequent hydrological problem in mountain catchments (see for example results in Fig. 8). Such an approach has been already proposed in the past, to my knowledge, but new applications can help to understand its added value. The application to the UIB is also interesting given the well known reasons that the authors recall in the Introduction (a high demand of water, a growing population in the area etc).

Thank you. Inversions have been conducted in the past based on streamflow. We have included several of those studies in the introduction. We have tested the approach using glacier mass balances in a sub-basin of the upper Indus [*Immerzeel et al.*, 2012]**, however without the rigorous uncertainty analysis and validation of the present study.**

My suggestion here is that the paper would benefit from more details about both methods and results discussion. These could help the reader to understand in a more exhaustive way the implications of the analysis that has been presented. As examples, why and how did you chose to consider up to four different products to estimate ETa? I agree with you that this can help to account for data uncertainty, but I think it would be useful to show why this operation is better than considering just one source. I also think that more details could help when introducing, for example, the geostatistical conditional simulation used to interpolate precipitation gradients. Please consider also to provide additional details about the formulation of the simple model you mention as Eq. 2 and about the hydrological fluxes that are not reported explicitly in the same Equation.

We have elaborated both sections and this was also suggested by reviewer 1.

Evapotranspiration is notoriously difficult to monitor. There are hardly any direct measurements of actual ET in the upper Indus, and even if there are they are not representative for basin wide ET, which varies considerably even at sea level without mountains. We choose to take into account the uncertainty in ET in our stream flow estimates and we choose for products covering re-analysis datasets, a

global hydrological model and an energy balance model. We have added a paragraph on the justification for the ET products to the methods section.

We have provided further details about the geostatistical interpolation. See also our reply to Bettina Sheafli where we explain the use of a standardized semi-variogram and the geostatistical conditional simulation. We have extended the description in the methods section.

We have added a discussion regarding our water balance equation in particular related to groundwater and sublimation to the validation section

As additional (minor) examples:
- Line 26 page 4756: what is currently still poorly understood about UIB hydrology? Please provide examples that could help the reader here;

We have clarified this in the text.

- Line 5 page 4761: please consider including an equation explaining the positive (or negative) lapse in space;

We have added equations to the manuscript.

- Line 13 page 4761: which temperature threshold do you consider to start melting?

At 0ºC

- Section 2.3: why did you choose the log-Gaussian and/or the Gaussian distributions?

We use log-Gaussian distributions for positively-values parameters and Gaussian if parameter values can also be negative. These are maximum entropy priors preferable if no additional information about the actual distributions is available.

- Line 2-4 page 4764: which is the time period when these accumulation measurements were made? Is it consistent with the period considered by the analysis?

These data are older, but the only direct observations available. Our point this that they confirm the vertical gradients we estimate and therefore we think it is good to cite these studies.

- Line 20 page 4766: maybe remove "of"?

OK.

B. Schaefli – Short Comment
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Received and published: 6 May 2015

This interesting manuscript uses an impressive data collection to estimate the amount of precipitation in the upper Indus Basin, as a key for a better quantification of the available water resources. The authors conclude that currently used precipitation estimates yield a gross underestimation of actual precipitation. Given the virtual absence of ground-based precipitation estimates at high altitudes, this work is obviously of prime importance. I am not an expert for this region, but I get from the presented discussion that it is currently not even entirely clear whether runoff in this region is fed by positive net glacier melt or not.

Thanks for the positive feedback and our detailed responses are listed below in bold. Regarding your last comment we note that glacier melt is a key contributor to runoff in the upper Indus [*Immerzeel et al.*, 2013; *Lutz et al.*, 2014]**, however overall there may a stable or slight positive glacier mass balance during the last decade** [*Gardelle et al.*, 2013]**. This does however not imply that glacier melt is negligible; it merely shows that the accumulation is larger than the melt. The glacier melt is not in the water balance equation (Eq. 2) because the melt equates the precipitation when the mass balance is 0 and it is thus accounted for.**

What triggered the present comment on this paper was

i) the overall impression that the used methods and results are presented in such a condensed way that I cannot entirely follow what has been done,

We have extended and clarified these sections.

ii) the absence of a summary of the order of magnitudes and of the uncertainties of the water balance terms,

We will include a water balance analysis including estimated uncertainties to the extent possible in the revised version to substantiate our findings. This is a good suggestion.

iii) the fact that the paper does not mention groundwater. Groundwater is absent from the water balance equation (eq. 2). This might of course be justified for the region / studied period but is nevertheless surprising.

We have assumed that over the observed period from 2003 until 2007 there is no net loss or gain of groundwater in the upper Indus basin. We do acknowledge that groundwater may play an important role in hydrology. A study in the Himalaya in Nepal shows that fractured

> **basement aquifers play an important role. They fill during the monsoon and they purge in the post-monsoon thus causing a natural delay in runoff** [*Andermann et al.*, 2012]**. However this does not imply significant net gains or losses over multiple year periods, which is what we consider. In the revised manuscript we have added this discussion related to role of groundwater and the potential additional uncertainty it may cause.**

The very condensed presentation of the methodology reads well but I would suggest to add some details (or supplementary material).

Point taken and we have elaborated the methodology in the revised manuscript. Some first responses to your queries can be found below.

I do for example not understand how the best precipitation field has been selected among all generated fields (there seems to be some form of optimization, on which criterion?).

There is no optimization within the 10000 realizations. What we present in Figure 5 is the ensemble average precipitation field based on the 10,000 simulations. Each simulation differs due to slightly different parameterization of the key processes in the model simulation (parameter ranges can be found in Table 1). This provides us with the uncertainty of the obtained precipitation field, where the standard deviation (measure of uncertainty) of those 10,000 simulations is shown in panel C.

Also e.g. in the sentence “. By running a multiple regression analysis after optimizing the PGs we quantify the contribution of each parameter to the total uncertainty.” I do not really know what has been done.

We randomly vary 5 parameters (HREF, HMAX, DDFd, DDFdg, TS) 10,000 times. For each glacier and for each parameter combination we optimize the PG such that the total accumulation for a glacier minus the total melt is equal to the mass balance. Once we have 10,000 combinations of parameters and associated PGs we ran a multi-variate linear regression analysis to determine relative contribution of each parameter to the spread in the PG, this is what is shown in Figure 6. The sum of the relative weights per region is 1. We have further clarified this in the method section.

What is the total uncertainty?
We define the total uncertainty as the standard deviation of the 10,000 simulations, which are based on the random variation of 5 key parameters as outlined in paragraph 2.3.

Why do the degree-day factors explain the PGs (“We take into account uncertainty in the following key parameters (HREF, HMAX, DDFd, DDFdg, TS) for the PG”)?
The degree-day factors determine the melt and the total amount of melt determines how much precipitation is required to sustain the observed mass balance. The total amount of precipitation is controlled by the PG.

On what are the PG-fields conditioned in “geostatistical conditional simulation”?
They are conditioned on the PGs determined for each glacier (separately for each of the 10,000 parameter combinations). The continuous PG fields are then obtained by using a variogram to spatially interpolate these PGs, were they are conditioned by the obtained PG at the glaciers. This is clarified in the text.

What do you mean with a standardized semi-variogram.? The same functional type everywhere?
This means that the sill is 1 and that we use the same variogram for the PGs associated with each of the 10,000 parameter combinations but the sill of the variogram is scaled with the variance of the PGs associated with each parameter combination drawn. This is clarified in the text.

Finally: would it be possible to summarize the different estimates of the water balance terms for the different sub-regions, how they are estimated and the order of magnitude of their uncertainty? Are there areas where the uncertainty of precipitation and mass balance is in the order of magnitude of the uncertainty of evaporation, transpiration and groundwater storage change (which would make any inference impossible to my view)? A comment on the order of magnitude of the observed runoff uncertainty could perhaps complete the picture

We have included a water balance analysis to the manuscript and in addition we have assessed the physical plausibility of our results using Budyko-Turc plots as suggested by Dr Andréassian.

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P. Reggiani – Short comment
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P. Reggiani (1), T.H.M. Rientjes (2) and B. Mukhopadhyay (3)

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We would like to thank the authors of the short comment for taking the time to provide this feedback to us. We respectfully disagree with most arguments, but evidently we are pleased to take this opportunity to rebut their points and reply to their feedback.

General

The discussion paper presents an analysis of precipitation estimation by inverse precipitation-stream flow modeling, aimed at proving that a) precipitation gauged by valley stations and b) TRMM remote sensing estimates of precipitation for the Upper Indus Basin (UIB) grossly underestimate actual precipitation.

What we do in the paper is to inversely model high altitude precipitation using the glacier mass balance and we validate our findings by making a first order estimate of streamflow, which we compare to observed records on an annual basis. This is a novel approach and the concept has been successfully tested (and published) at a smaller scale for the Hunza basin [*Immerzeel et al.*, 2012]**. So, we do not perform inverse precipitation-streamflow modelling as suggested.**

As an alternative to TRMM and gauged data, the authors use ERA-interim and MERRA reanalysis products to derive basin-wide mean annual precipitation. The products are artificially corrected, whereby closure of the basin-scale mean annual mass balance equation Q=P-ET+MB serves as a constraint.

This is not correct. In our approach we use the APHRODITES precipitation dataset as a basis and we correct this dataset using precipitation gradients and a presumed elevation of peak precipitation based on published relationships between precipitation and elevation (see caption Table 1). We constrain these precipitation gradients based on the glacier mass balance trends [*Kääb et al.*, 2012a].We validate our finding using streamflow observations after correction with ET. To account for the large uncertainty in ET we use four different ET products (all published) and the MERRA product is one of those.

Losses to groundwater and buffer effects due to longer residence times of water in alluvial deposits (generally composed of silt, sand and gravel) are neither addressed nor mentioned.

We correct precipitation in areas above 2000 m asl, where the terrain is general steep, soils are shallow and the abundance of extended areas with alluvial deposits is limited. We have assumed that over the observed period from 2003 until 2007 there is no net loss or gain of groundwater in the upper Indus basin. We do acknowledge that groundwater may play an important role in the hydrology. A study in the Himalaya in Nepal shows that fractured basement aquifers play an important role. They fill during the monsoon and they purge in the post-monsoon thus causing a natural delay in runoff of a few months [Andermann et al., 2012]. However this does not imply significant net gains or losses over multiple year periods, which is what we consider. Interesting to note here is that the authors themselves also assume a negligible net groundwater flux on an annual timescale ([*Reggiani and Rientjes*, 2014]. We agree it is an important topic and in the revised manuscript we will add a discussion related to role of groundwater and the potential additional uncertainty it may cause.

The discharge Q is the observed long-term mean annual stream flows for various sub-catchments, ET is estimated from reanalysis data or an energy balance model, while glacier mass balance accounting (MB) is based on ICESAT satellite altimetry (25 sqkm resolution).

ICESat is a space-borne laser altimeter and it provides point measurements of surface height in tracks. These data were processed into regional trends in glacier mass balance and the approach is published in Nature [*Kääb et al.*, 2012a]. So it is by no means a gridded dataset at 25 km^2 resolution.

In the inverse model, precipitation P is considered as the dependent variable. The analysis window is 2003-2007. The verification of the mass balance closure is achieved by means of a grid-based distributed hydrological model (PCGLOB) (1 sqkm grid resolution, daily time step), which estimates net precipitation (P-ET) and contains glacier mass balance accounting (MB) with the aim to reproduce observed flows (Q) at a series of observation points.

We do not use the hydrological model PCR-GLOBWB, which we think the authors are referring to, but we use the corrected precipitation, the ensemble average of four evapotranspiration products and the glacier mass balance to make a first order estimate of average annual runoff. We compare this estimate with observation as an independent validation.

From modeling and an uncertainty analysis in which several precipitation correction model parameters are drawn by Monte Carlo analysis, it is concluded, that the mean annual precipitation over the basin must equate 913±323mm/year. This value is approximately a factor three higher than the estimates stated in several earlier publications (Immerzeel et al. 2009, 2010; Bookhagen and Burbank 2010).

That is correct. Understanding the water balance at large of a complex basin such as the upper Indus which lacks direct observations has been a quest of many scientists and slowly but steadily progress has been made including our present study and previous work [*Immerzeel et al.*, 2009, 2010, 2012, 2013, 2014; *Pellicciotti et al.*, 2012; *Ragettli et al.*, 2013; *Lutz et al.*, 2014] and the work of other authors [*Mukhopadhyay and Khan*, 2014a, 2014b; *Reggiani and Rientjes*, 2014]. To our opinion new insights should be allowed in science and that is in fact how progress is made. The comments should be directed to this particular paper being under review and not at papers that have already passed a rigorous process of peer-review by themselves.

Actual evaporation is estimated as an average of four widely disparate products, including ERA Interim evaporation (i), MERRA reanalysis evaporation (ii), an estimate using an energy balance model (iii) and an estimate computed by PCGLOB via soil moisture accounting (iv). The average value and spread between the four products is 359±107mm/yr. In the works by Immerzeel et al. (2009, 2010) and Bookhagen and Burbank (2010) evaporation is neglected.

Yes, one of the things we have learned in previous years is that actual ET (and possibly sublimation at high altitude even more) may play a significant role in the water balance. However, actual ET is notoriously difficult to measure and even if there are point measurements available they are by no means representative to the entire upper Indus. Therefore we have decided to use the ensemble average of four different actual ET products which are all published in peer reviewed journals. We acknowledge these products are subject to uncertainty and the ensemble average ET for the upper Indus is 359 mm/yr and the spread is 107 mm/yr. We take this spread into account in our estimate of annual runoff used to validate our approach.

The paper seems to be another attempt (e.g., Immerzeel et al. 2012a, 2013) to come up with more realistic results than those first published in Immerzeel et al. (2009), where a mass balance analysis of the UIB was performed using basin-average TRMM precipitation estimates of 300 mm/year for the 2001-2005 period to drive the SRM hydrological model (Martinec, 1975). From the modeling results at that time, the authors reached the conclusion that to close the mass balance at Besham Quila gauging station (upstream of the basin outlet at Tarbela Reservoir), where 460 mm/year is the observed long-term mean annual flow, the supplementary discharge required to close the water balance must come from non-renewable glacier wastage at a rate of 1% per year. The authors cited these results in another sequel article (Immerzeel et al., 2010). In Immerzeel et al. (2012b), the Indus basin was labelled as “hot spot” based on the 2010 findings, including the water supply perspective. In Immerzeel et al. (2009) actual evaporation as a forcing term is set to zero. If included, it would lead to a higher (and even more unrealistic) glacier melting rate to close the water balance.

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The authors seem to provide comments here on a paper published in remote sensing of environment of 2009, so there is no immediate need for us to respond to this here, however we are happy to provide some context.

This 2009 paper was the first in a sequence to unravel the Indus water balance and it has been cited 127 times (Scopus) by numerous scientists working in this field. It was also the time when the Karakoram anomaly was still a revolutionary idea postulated by Ken Hewitt [*Hewitt,* 2005], but the idea made it only to the mainstream as late as 2012. It was also the period before the IPCC discussion on the Himalayan glaciers. At that time, we were one of the first to attempt to model the entire upper Indus using a simple hydrological model forced with TRMM precipitation and MODIS snow cover and validated by runoff. We noted indeed a mismatch between the total runoff and the total TRMM precipitation and in the paper we discuss two options in careful terms based on the knowledge available at that point in time: (i) the mismatch is caused by an underestimation of precipitation and (ii) it is caused by a negative glacier mass balance. Now, in 2015, we believe the first reason is the most plausible.

In our view, the discussion paper suffers from a series of conceptual shortcomings:

Firstly, the authors continue to look at a very short time window (2002-2007), ignoring longer, climatic, time scales. For instance, when the 50-year trend of the observed Indus flows at the inlet of Tarbela Reservoir, downstream of Besham Quila, is considered, it should have become outright apparent that flow data exhibit an essentially stable trend from 1961 to date, as indicated by Reggiani and Rientjes (2015) and Mukhopadhyay and Khan (2015a). Moreover, the cumulative reservoir inflow volumes at Tarbela for the 1999-2009 decade were actually 4% below the 1961-2009, 50-year average (see Table 2 in Reggiani and Rientjes, 2015), the same time window for which Kääb et al. (2012) estimated a non-renewable ice mass loss from ICESAT altimetry data equivalent to 231± 46 m3/s of mean annul discharge at Tarbela. This equivalent discharge is 10% higher than the observed long-term mean annual flow and casts doubts on the reliability of the satellite-based ice mass estimates for the UIB. As a result, one should question if the satellite-derived mass balance estimates can be considered and used as an estimator variable for glacier mass balance accounting, and as in this case, to derive inferences about precipitation.

The ICESAT altimetry data are an established means to assess trends in glacier mass balance ([*Kääb et al.*, 2012b, 2015]) and this is not a topic to be debated here. This specific observation seems to re-open a previous discussion. We are well aware of the authors' discussion with Andreas Kääb about the brief communication in the Cryosphere regarding this topic (http://www.the-cryosphere-discuss.net/8/5857/2014/tcd-8-5857-2014-discussion.html**) and we strongly support arguments provided by Andreas Kääb in his reply.**

We constrain our precipitation correction by glacier mass balance observations from ICESat which were only available for the 2002-2007 period. Andreas Kääb has computed the mass balance trends for the three zones we have considered (Himalaya, Hindu-Kush and Karakoram) with a method similar to his Nature paper [*Kääb et al.*, 2012a]. Overall the mass balance trends over this period are slightly negative (see Table 1). The reason is that the Karakoram anomaly does not overlap significantly with the Indus basin boundary [*Kääb et al.*, 2015]. Moreover, we take into account a (considerable) uncertainty in the mass balances to estimate the uncertainty in our precipitation estimates (paragraph 2.1 and Table 1).

The authors base their argument on Tarbela flow which drains only about half of the upper Indus basin we consider in our study. The fact that observed flows (also subject to errors by the way) are stable does not contradict our findings. There are many factors influencing streamflow and a potential change in snow melt regime is the large unknown here.

Finally, even if the glacier mass balances were positive then still precipitation would be significantly underestimated in particular in the north-western part of the basin. The precipitation in the APHRODITES dataset (and other data sets as well) is simply too small to account for the large glacier systems found in the upper parts of the basin. That is our point.

Secondly, different studies have addressed the issue of estimating realistic precipitation and actual evaporation rates. For the Upper Indus Basin (UIB), a large number of gridded rainfall products have been examined. For instance Palazzi et al. (2013) and Reggiani Rientjes (2015) studied several precipitation reanalysis products showing that the basin-average precipitation in the UIB is indeed at least double the rates indicated by the TRMM 3B43 product in Immerzeel et al. (2009; 2010) and in the order of 675± 100 mm/yr, thus significantly higher than those recorded at valley stations (Archer and Fowler, 2004). Several studies with weather stations placed over limited periods at high altitudes have indicated that actual precipitation in the high altitude mountainous areas is significantly higher, reaching up to 2000 -3500 mm and higher of w.e. per year (e.g. Wake 1989, Cramer 2000, Kuhle 2005, Winiger 2005), to then decrease higher up, an already well-known phenomenon (see Fig. 8 in Mukhopadhyay Khan, 2014a).

This is exactly the point of our paper and based on our approach in this paper we estimate the basin precipitation to be 913 ± 323 mm/yr, which is indeed higher than the TRMM 3B43, APHRODITES and most other commonly used gridded products (see our introduction). The authors themselves estimate the basin precipitation to be 675 ± 100 mm/yr, but this is only upstream of Tarbela (about half of the area we consider). Considering the uncertainty margins our estimates do not differ significantly. We have used the work of Matthias Winiger and Ken Hewitt [*Winiger et al.*, 2005; *Hewitt*, 2007, 2011] to estimate values for the elevation of maximum precipitation. Both have decades of field experience in the region. Our final results match well with field observation of high altitude accumulation.

Also, estimates of actual evaporation are provided, which have been presented in literature based on few field experiments at highly glaciated mountain ranges including the Himalayas at large (Buthyani, 1999, Khattak et al., 2011) and valley-based stations (see Fig. 7 in Mukhopadhyay and Khan, 2014a). In particular, Buthyani (1999) indicated a mean annual total evaporation rate in the order of 200 mm/yr for Siachen glacier based on glacier mass balance. In the discussion paper the authors rely on i) gridded estimated actual evaporation with mean values which are at least a factor two higher than observed in glaciated areas in the Himalayas, ii) possibly inconsistent satellite-based glacier mass estimates, iii) and short-term flow records as independent variables to draw inferences about precipitation. The more robust approach would be to rely on evaporation and precipitation estimates and trends to infer on glacier mass balance. In this case, it would become apparent that satellite-derived mass balances are not sufficiently reliable to serve as support in inverse modeling of precipitation.

Most points have been covered already earlier in this reply.

The Khattak paper (2011) does not discuss ET, but only temperature, precipitation and stream flow. Buthyani (1999) provides an ET estimate for a single glacier based on an empirical formula developed to estimate lake evaporation in the US based on air temperature only and in Fig 7. of Mukhopadhyay and Khan (2014a) the Penman-Monteith reference ET is given for selected stations. Reference ET is very different from actual ET and average values plotted here are about 2.5 mm/day, which is about 900 mm / year. This considered we believe that our approach to estimate basin wide ET is much preferred. The points regarding satellite-based glacier mass estimates and stream flow records have been covered above.

Thirdly, the authors chose to ignore long-term observed flow time series. An inverse modeling attempt like the one proposed here, with multiple uncertain independent variables (i.e. ET, MB), cannot replace or serve as a substitute to any sound analysis of observed stream flow data.

We think it does, in particular when uncertainty are considered as rigorous as we do. Relying entirely on streamflow analysis will not solve this puzzle as ET, snow melt, glacier melt, sublimation, rain and groundwater dynamics all have their role in streamflow and isolating specific components from streamflow only is an ill-posed inverse problem that is impossible to solve.

Neither does an inverse steam flow modeling on a time window of half a decade convey a sense of confidence when conclusions need to be drawn on long-term, climate-controlled glacier mass storage.

We do not draw any conclusion on glacier storage, but only on glacier storage changes, which respond directly to the climate. Moreover, we do not apply inverse stream flow modeling as Reggiani et al. suggest. Instead, we use average stream

flow estimates as a means to validate the inverse modeling based on glacier mass balance (with good results).

An analysis of longer flow records in space and time would provide considerably more insights into the mass balance of the basin than numerical modeling alone (in this context we recall that satellite-altimetry derived mass balance in glaciers in extreme topography (Kääb et al., 2012) is essentially an application of reflected electromagnetic wave signal interpretation, which has not undergone any thorough validation for the particular region yet). Rising trends of August flows in the central and eastern Karakoram imply decreasing glacial storage at rates of 0.553 – 0.645 mm/day/year and 0.186 – 0.217 mm/day/year in the Shigar and Shyok watersheds respectively, whereas in the western Karakoram (Hunza watershed) falling trend of August flows implies increasing glacial storage at a rate of 0.552 – 0.644 mm/day/year (Mukhopadhyay Khan, 2014b; Mukhopadhyay et al., 2014; Mukhopadhyay Khan, 2015b). Such rates should be reconciled with the precipitation trends to infer changes in the regionally- averaged glacier mass balance.

This has been covered several times before. In the revised version we have included a water balance estimate for the three regions we consider.

Fourth, the distribution of the various parameters in the uncertainty analysis of precipitation are assumed with a (log-)Gaussian distribution, which the authors have not demonstrated to relate to actual empirical distributions in the region that could in principle be quite different (e.g. bi-modal, skewed, non-Gaussian etc.). The analysis only yields the uncertainty of their precipitation correction model which they have assumed and inserted into the model “a priori” based on values taken from the literature and not necessarily the actual uncertainty of precipitation, which is yet unknown. The precipitation uncertainty analysis pursued in this way is thus akin to a prediction that directly or indirectly causes itself to become true, by the very terms of the prophecy itself (Merton, 1948).

We respectfully disagree. There are 6 parameters which play a key role in our approach (HREF, HMAX, DDFd, DDFdf, TS, MB). The uncertainty in these parameters jointly determine the uncertainty in the precipitation gradient and the resulting precipitation field. We use log-Gaussian distributions for positively-values parameters and Gaussian if parameter values can also be negative. These are maximum entropy priors preferable if no additional information about the actual distributions is available. We base the parameter range on literature values (some collected during our own field campaigns) and we run a rigorous Monte Carlo simulation of 10,000 runs. This forward stochastic approach to uncertainty assessment is a well-accepted approach in science, if no direct information on output uncertainty is available.

The research and results presented in the paper do not provide relevant benefit towards understanding the hydrological balance in UIB. Findings on gridded precipitation and actual evaporation products are significantly higher than those shown in recent publications, whereas long-term streamflow analysis and aspects of glacier mass storage are not analyzed. The underlying assumption that the water balance can be closed by inversely estimating precipitation results in basin-average precipitation estimates that are likely overrated. Given the essentially stable (or statistically insignificant falling) long-term trend in observed stream flows at the basin outlet, the truly important scientific issue is not an estimation of the absolute value of the basin wide mean annual precipitation, which can hardly be achieved in this terrain, but validation of glacier mass loss estimates against the background of a hydrological balance of the basin and spatial patterns and trends in precipitation, as a function of summer and winter seasons. Such an analysis is needed to validate the mass balance of the glaciers and melting rates variously given in Immerzeel et al. (2009), Kääb et al. (2012) and Gardelle et al. (2012, 2013). Consequently, the discussion paper opens more questions than it provides answers, while the methodological approach does not contribute much of value in this respect.

We believe our work is an important step forward in understanding upper Indus hydrology and it provides a new independent estimate of high altitude precipitation using changes in glacier mass balances derived from ICESAT laser altimetry, which is a proven technique published in high quality scientific journals.

Many uncertainties remain until the upper Indus hydrology is understood entirely and this will be the challenge for the years ahead; snow melt dynamics, evapotranspiration, sublimation, high altitude atmospheric dynamics, monsoon vs. westerlies, groundwater. Once we have a better understanding of these processes we may be able to unravel trends in observed river flow.

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Dr Andréassian – Short comment
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I agree with the authors that the only possible way to assess unobserved and unobservable precipitation is by inverting the hydrological cycle. I found the study very complete, comparing the major satellite and reanalysis products with catchment water balance.

Thank you for this positive feedback and support for our approach.

I personally would have been interested in seeing Figure 8 complemented by a projection of the points in the Turc-Budyko non-dimensional graphs (Q/P vs P/E0) : an example can be found in Fig. 2 and 4 in Valéry et al. (2010) It would be a way to show how catchments which had a physically unrealistic water balance can be reintegrated into the hydrologically feasible part of the Turc-Budyko plot.

That is a useful suggestion and we have included a Turc-Budyko plot in the revised version of our manuscript.

Last, I also take the immodest liberty to suggest a complement for your literature review, but only because it really deals with the same issue of precipitation gradients : Valéry et al. (2009).

Thanks! We have included the references in our introduction.

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Reconciling high altitude precipitation in the upper Indus basin with glacier mass balances and runoff

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Abstract

Mountain ranges in Asia are important water suppliers, especially if downstream climates are arid, water demands are high and glaciers are abundant. In such basins, the hydrological cycle depends heavily on high altitude precipitation. Yet direct observations of high altitude precipitation are lacking and satellite derived products are of insufficient resolution and quality to capture spatial variation and magnitude of mountain precipitation. Here we use glacier mass balances to inversely infer the high altitude precipitation in the upper Indus basin and show that the amount of precipitation required to sustain the observed mass balances of ~~the ~~large glacier systems is far beyond what is observed at valley stations or estimated by gridded precipitation products. An independent validation with observed river flow confirms that the water balance can indeed only be closed when the high altitude precipitation is up to a factor ten higher than previously thought. We conclude that these findings alter the present understanding of high altitude hydrology and will have an important bearing on climate change impact studies, planning and design of hydropower plants and irrigation reservoirs and the regional geopolitical situation in general.

1 Introduction

Of all Asian basins that find their headwaters in the greater Himalayas, the Indus basin depends most strongly on high altitude water resources (Immerzeel et al., 2010; Lutz et al., 2014) The largest glacier systems outside the polar regions are found in this area and the seasonal snow cover is the most extensive of all Asian basins (Immerzeel et al., 2009). In combination with a semi-arid downstream climate, a high demand for water owing to the largest irrigation scheme in the world and a large and quickly growing population, the importance of the upper Indus basin (UIB) is evident (Immerzeel and Bierkens, 2012).

The hydrology of the UIB ($4.37 \cdot 10^5$ km^2~~) is~~) is, however, poorly understood. The quantification of the water balance in space and time is a major challenge due the lack of measurements and the inaccessibility of the terrain.~~ and t~~ The magnitude and distribution of high altitude precipitation, which is the driver of the hydrological cycle, is one of its largest unknowns (Hewitt, 2005, 2007; Immerzeel et al., 2013; Mishra, 2015; Ragettli and Pellicciotti, 2012; Winiger et al., 2005). Annual precipitation patterns in the UIB result from the intricate interplay between synoptic scale circulation and valley scale topography-atmosphere interaction resulting in orographic precipitation and funnelling of air movement (Barros et al., 2004; Hewitt, 2013). At the synoptic scale, annual precipitation originates from two sources: the south-eastern monsoon during the summer and moisture transported by the westerly jet stream over central Asia (Mölg et al., 2013; Scherler et al., 2011) during winter. The relative contribution of westerly disturbances to the total annual precipitation increases from south-east to north-west, and the anomalous behaviour of Karakoram glaciers ~~are~~ is commonly attributed to changes in winter precipitation (Scherler et al., 2011; Yao et al., 2012).

At smaller scales the complex interaction between the valley topography and the atmosphere dictates the spatial distribution of precipitation (Bookhagen and Burbank, 2006; Immerzeel et al., 2014b). Valley bottoms, where stations are located, are generally dry and precipitation increases up to a certain maximum altitude (HMAX) above which all moisture has been orographically forced out of the air and precipitation decreases again. In westerly dominated rainfall regimes HMAX is generally higher, which is likely related to the higher tropospheric altitude of the westerly airflow (Harper, 2003; Hewitt, 2005, 2007; Scherler et al., 2011; Winiger et al., 2005).

Gridded precipitation products are the de facto standard in hydrological assessments, and they are either based on observations (e.g. APHRODITES (Yatagai et al., 2012)), remote sensing (e.g. the Tropical Rainfall Monitoring Mission (Huffman et al., 2007)) or reanalysis (e.g. ERA-Interim (Dee et al., 2011)) (Fig. 1, panel C to E). In most cases the station data strongly influence the distribution and magnitude of the precipitation in those data products; however the vast majority of the UIB is located at elevations far beyond the average station elevation

(Fig. 1, panel A to B). The few stations that are at elevations above 2000 m are located in dry valleys and we hypothesize that the high altitude precipitation is considerably underestimated (Fig. 1, panel C to E). Moreover, remote sensing based products, such as TRMM, are insufficiently capable of capturing snowfall (Bookhagen and Burbank, 2006; Huffman et al., 2007) and the spatial resolution (25 to 75 km^2) of most rainfall products (and the underlying models) is insufficient to capture topography-atmosphere interaction at the valley scale (Fig. 1, panel C to E). Thus, there is a pressing need to improve the quantification of high altitude precipitation, preferably at large spatial extents and at high resolution.

A possible way to correct mountain precipitation is to inversely close the water balance. Previous studies in Sweden and Switzerland have shown that it is possible to derive vertical precipitation gradients using observed runoff in a physically realistic manner (Valery et al., 2009; Valéry et al., 2010). Earlier work at the small scale in high mountain Asia suggested that the glacier mass balance may be used to reconstruct precipitation in its catchment area (Harper, 2003; Immerzeel et al., 2012a). Fig. 1 (panel A and B) shows that UIB~~the~~ glaciers are located at high elevations that are ~~in higher parts of the UIB, which are not covered~~not represented by station data. Therefore the mass balances of the glaciers may contain important information on high altitude accumulation in an area that is inaccessible~~,~~ and ungauged, but very important from a hydrological point of view. In this study we further elaborate this approach by inversely modelling average annual precipitation from the mass balance of 550 large (> 5 km^2) glacier systems located throughout the UIB. We perform a rigorous uncertainty analysis and we validate our findings using independent observation of river runoff.

2 Methods

We estimate high altitude precipitation by using a glacier mass balance model to simulate observed glacier mass balances. We use a gridded dataset from valley bottom stations as a basis for our precipitation estimate and we compute a vertical precipitation gradient (PG (*% m^{-1}*)) until observed mass balances matchs~~d~~ the simulated mass balance. We repeat this process for the 550 major glacier systems in the UIB, and the resulting PGs are then spatially interpolated to generate a spatial field that represents the altitude dependence of precipitation. We use this field to ~~correct~~update the APHRODITE precipitation and generate a corrected precipitation field that ~~which~~ is able to reproduce the observed glacier mass balances~~ changes~~. We validate the findings independently with a water balance approach. Estimated (annual) runoff, based on the corrected precipitation, actual evapotranspiration and the observed mass balance, is compared with an extensive set of UIB runoff observations. A rigorous uncertainty analysis is also conducted on the six most critical model parameters including potential effects of spatial correlation.

~~We use regionally-averaged glacier mass balance (MB) data from ICESat (Kääb et al., 2012) in combination with daily air temperature fields, a degree day melt model and a detailed elevation model to optimize the precipitation gradient (PG) so as to match the simulated and observed glacier mass balances. We apply geostatistical conditional simulation (Goovaerts, 1997) to spatially interpolate the PG estimates to high resolution PG fields, which are then used to estimate the high altitude precipitation. To account for uncertainties in the estimated precipitation we randomly sample six critical parameters and generate 10,000 equiprobable precipitation fields and we validate our results using observed river runoff data.~~

2.1 Datasets

2.1.1 Glacier mass balance and outlines

Glacier mass balance trends based on ICESat (Kääb et al., 2012a) are recomputed for the period 2003 until 2008 for the three major mountain ranges in the UIB: the Karakoram, the Hindu-Kush and the Himalaya Fig. 1. For each zone the mass balance is computed including a regional uncertainty estimate (Kääb et al., 2012a). From the zonal uncertainty (σ_z) we estimate the standard deviation between glaciers within a zone (σ_g) as

$$\sigma_g = \sigma_z \sqrt{n} \,. \qquad (1)$$

Where n is the number of glaciers within a zone. The σ_g values used in the uncertainty analysis are shown in Table 1.

Glacier outlines area based on the glacier inventory of the International Centre for Integrated Mountain Development (Bajracharya and Shrestha, 2011).

2.1.2 Precipitation and temperature

The daily APHRODITE precipitation (Yatagai et al., 2012) and air temperature datasets (Yasutomi et al., 2011) from 2003 until 2007 are used as reference datasets to ensure maximum temporal overlap with the ICESat based glacier mass balance dataset (Kääb et al., 2012a). The precipitation dataset is resampled from the nominal resolution of 25 km^2 to a resolution of 1 km^2 using the nearest neighbour algorithm. The air temperature dataset is then bias-corrected using monthly linear regressions with independent station data to account for altitudinal and seasonal variations in air temperature lapse rates (Fig. 3).

~~2.1.1~~2.1.3 Runoff and evapotranspiration

We use runoff data, ~~and~~ potential (ETp) and actual evapotranspiration (ETa) data for the validation of our results. For runoff we compiled all available published data, which we complemented with data made available by the Pakistan Meteorological Department and the Pakistan Water and Power Development Authority.

Evapotranspiration is notoriously difficult to monitor and there are few direct measurements of ETa in the upper Indus. In earlier UIB studies, ET was estimated using an empirical formulae based on air temperature but was only applied to the Siachen glacier (Bhutiyani, 1999; Reggiani and Rientjes, 2014). We take into account the uncertainty in ET in our stream flow estimates and develop a blended product based on re-analysis datasets, a global hydrological model and an energy balance model. ~~To account for uncertainty in~~Four gridded ETa and three gridded ETp ~~estimates~~ products ~~we used four different products which~~were ~~we~~ resampled to a 1 km^2 resolution at which we perform all our analys~~i~~es:

- ERA-Interim reanalysis (Dee et al., 2011): ERA-Interim uses the HTESSEL land surface scheme (Dee et al., 2011) to compute actual evapotranspiration (ETa). For transpiration a distinction is made between high and low vegetation in the HTESSEL scheme and these are parameterized from the Global Land Cover Characteristics database at a nominal resolution of 1 km^2.
- MERRA reanalysis (Rienecker et al., 2011): The MERRA reanalysis product of NASA applies the state-of-the-art GEOS-5 data assimilation system that includes many modern observing systems in a climate framework. MERRA uses the GEOS-5 catchment LSM land surface scheme (Koster et al., 2000) to compute actual ET. For the MERRA product ETp is not available.
- ET-Look (Bastiaanssen et al., 2012): The ET-Look remote sensing model infers information on ET from combined optical and passive microwave sensors, which can observe the land-surface even under persistent overcast conditions. A two-layer Penman–Monteith forms the basis of quantifying soil and canopy evaporation. The dataset is available only for the year 2007, but it was scaled to the 2003 – 2007

average using the ratio between the 2003 – 2007 average and the 2007 annual ET based on ERA-INTERIM.

- PCRGLOB-WB (Wada et al., 2014): The global hydrological model PCRGLOB-WB computes actual evapotranspiration using potential evapotranspiration based on Penman-Monteith, which is further reduced based on available soil moisture.

The average annual ETa for the period 2003 – 2007 for each of the four products is shown in Fig. 2. The spatial patterns show good agreement, but the magnitudes differ considerably. The ~~average~~ ensemble mean ETa for the entire upper Indus equals 359 ± 107 *mm y*$^{-1}$.

~~The daily APHRODITE precipitation (Yatagai et al., 2012) and air temperature datasets (Yasutomi et al., 2011) from 2003 until 2007 are used as reference datasets to ensure maximum temporal overlap with the ICESat based glacier mass balance dataset (Kääb et al., 2012). The precipitation dataset is resampled from the nominal resolution of 25 km² to a resolution of 1 km² using the nearest neighbour algorithm. The air temperature dataset is then bias-corrected using monthly linear regressions with independent station data to account for altitudinal and seasonal variations in air temperature lapse rates Fig. 3.~~

2.2 Model description

We use the PC-Raster spatial-temporal modelling environment (Karssenberg et al., 2001) to model the mass balance of the major glaciers in each zone and subsequently estimate precipitation gradients required to sustain the observed mass balance. The model operates at a daily time step from 2003 until 2007 and a spatial resolution of 1 *km*^{2}. For each time step the total accumulation and total melt are aggregated over the entire glacier surface. Only glaciers with a surface area above 5 *km*^{2} are included in the analysis (Karakoram = 232 glaciers, Hindu-Kush = 119, Himalaya = 204 glaciers) as the ICESat measurements do not reflect smaller glaciers~~to avoid scale problems~~. The model is forced by the spatial precipitation and temperature fields. The precipitation fields are corrected using a precipitation gradient (PG, *% m*$^{-1}$). Precipitation is positively lapsed using a PG between a reference elevation (HREF) to an elevation of maximum precipitation (HMAX). At elevations above HMAX the precipitation is negatively lapsed from its maximum at HMAX with the same PG according to:

$$P_{cor} = P_{APHRO} \cdot (1 + (H - HREF)) \cdot PG \cdot 0.01 \qquad \text{Eq. 1}$$

for HREF < H ≤ HMAX.

~~for H ≤ HMAX, and:~~

$$P_{cor} = P_{APHRO} \cdot (1 + (((HMAX - HREF) + (HMAX - H)) \cdot PG \cdot 0.01) \qquad \text{Eq. 2}$$

for H > HMAX, and:

~~for HREF < H < HMAX.~~

HREF and HMAX values are derived from literature (Table 1) and uncertainty is taken into account in the uncertainty analysis. HMAX varies per zone and lies at a lower elevation in the Himalayas than in the other two zones (Table 1). We spatially interpolate HMAX from the average zonal values to cover the entire UIB.

The melt is modelled over the glacier area using the positive degree day (PDD) method (Hock, 2005), with different degree day factors (DDF) for debris-covered (DDFd) and debris-free (DDFdf) glaciers derived from literature (Table 1). To account for uncertainty in DDF, the DDFd and DDFdf are taken into account separately in the uncertainty analysis. At temperatures below the critical temperature of 2 *°C* (Immerzeel et al., 2013; Singh and Bengtsson, 2004) precipitation falls in the form of snow and contributes to the accumulation. Avalanche nourishment of glaciers is a key contributor for UIB glaciers (Hewitt, 2005, 2011) and to take this process into account, we extend the glacier area with steep areas directly adjacent to the glacier with a slope over an average threshold slope (TS) of 0.2 $m\ m^{-1}$. This average threshold slope is derived by analysing the slopes of all glacieriz~~s~~ed pixels in the basin (Fig.4). To account for uncertainty in TS this parameter is taken into account in the uncertainty analysis.

For each glacier system we execute transient model runs from 2003 until 2007 and we compute the average annual mass balance from the total accumulation and melt over this period. We make two different runs for each glacier system with two different PGs (0.3 *% m*$^{-1}$ and *0.6 % m*$^{-1}$) and we use the simulated mass balances of these two runs and the observed mass balances based on ICESat to optimize the PG per glacier, such that the simulated mass balance matches the observed.

To interpolate the glacier-specific PG-values to PG spatial ~~-~~fields over the entire domain we use geostatistical conditional simulation (Goovaerts, 1997)~~ with a standardized semi-variogram~~. ~~They~~ Simulated spatial fields of PG are thus conditioned on the PGs determined ~~for each glacier (separately for each of the 10,000 parameter combinations). The continuous PG fields are then obtained by using a variogram to spatially interpolate these PGs, were they are conditioned by the obtained PG points~~ at the glaciers centroid. The~~A~~ ~~standardized~~ semi-variogram ~~is~~ has the following parameters: ~~used (~~nugget = 0, the range = 120 *km*, sill = ~~1~~variance of PGs.~~) to interpolate glacier obtained PGs over the study area.~~ ~~The interpolation and construction of the PG is done for each of the~~ ~~In the semi-variogram, the nugget and the range are fixed and the sill is set equal to the variance of the PGs of all glaciers. The spatial PG fields are then used in combination with a digital elevation model to generate the corrected precipitation field.The same variogram for the PGs associated with each of the 10,000 parameter combinations individually (see uncertainty analysis), but the sill of the variogram is scaled with the variance of the PGs associated with the specific parameter combination drawn.~~

2.3 Uncertainty analysis

A rigorous uncertainty analysis is performed to take into account the uncertainty in parameter values and uncertainty in regional patterns. To account for parameter uncertainty we perform a 10,000 member Monte Carlo simulation on the parameters given in Table 1. For each run we randomly sample the parameter space based on the average (μ) and the standard deviation (σ), which are all based on literature values. For the positively-valued parameters we use a log-Gaussian distribution and a Gaussian distribution in case parameter values can be negative. We take into account uncertainty in the following key parameters (HREF, HMAX, DDFd, DDFdg, TS) for the PG as well as uncertainty in the mass balance against which the PG is optimized (MB). ~~Based on the results of the 10,000 simulations we derive the average corrected precipitation field and the associated uncertainty in the estimates~~. We randomly vary the 5 parameters (HREF, HMAX, DDFd, DDFdg, TS) 10,000 times and calculate the PG~~.~~ for each glacier for each random parameter set drawn, thus resulting in 10,000 PG-sets for each glacier considered. For each of the 10,000 PG-sets we then use conditional simulation (see above) to arrive at 10,000 equally probable spatial PG-fields, taking account of parameters uncertainty, mass-balance uncertainty and the interpolation error. Note that for each of the 10,000 sets the variogram is scaled with the variance of the PGs associated with the specific parameter combination drawn. Finally, based on the results of the 10,000 simulations we derive the average corrected precipitation field and the associated uncertainty in the estimates

Using the 10,000 combinations of parameters and associated PGs we ran a multi-variate linear regression analysis to determine relative contribution of each parameter to the spread in the PG to understand which parameter has the largest influence on the PG.

It is possible that certain parameters used in the model are spatially correlated. ~~T~~To account for uncertainty in this spatial correlation and the presence of spatial patterns in the parameters we perform a sensitivity analysis where we consider three cases:

- Fully correlated: we assume the parameters are spatially fully correlated within a zone, e.g. for each of the 10,000 simulations a parameter~~s~~ has the same value within a zone
- Uncorrelated: we assume the parameters are spatially uncorrelated and within each zone each glacier system is assigned a random value
- Intermediate case: we use geostatistical unconditional simulation (Goovaerts, 1997) with a standardized semi-variogram (nugget = 0, sill = variance of parameter, range = 120 *km*) to simulate parameter values for each glacier system.

2.4 Validation

We estimate the average annual runoff (Q) for sub-basins in the UIB from

$$Q = P_{cor} - ET + MB \qquad (~~2~~3)$$

Where P_{cor} is the corrected average precipitation, ET is the average annual evapotranspiration based on the four products described above and MB is the glacier mass balance expressed over the sub-basin area in $mm\ y^{-1}$. We then compare the estimated runoff values to the observed time series (Table 2).

For the three zones (Himalaya, Karakoram and Hindu-Kush) we also perform a water balance analysis to verify whether the use of the corrected precipitation product results in a more realistic closure of the water balance. Finally we test the physical realism of the corrected precipitation product using a non-dimensional Turc-Budyko plot as described in ~~(~~Valéry et al.~~,~~ (2010). This plot is based on two assumptions: (i) the mean annual runoff should not exceed the mean annual precipitation and (ii) the mean annual runoff should be larger than or equal to the difference between precipitation and potential evapotranspiration. By plotting P/ETp versus Q/P on catchment basis it is tested whether the use of corrected precipitation results in more physically-realistic values.

3 Results and discussion

3.1 Corrected precipitation

The average annual precipitation based on 10,000 conditionally simulated fields reveals a striking pattern of high altitude precipitation. The amount of precipitation required to sustain the large glacier systems is much higher than either the station observations or the gridded precipitation products imply. For the entire UIB the uncorrected average annual precipitation (Yatagai et al., 2012) for 2003-2007 is 437 $mm\ y^{-1}$ (191 $km^3\ y^{-1}$), an underestimation of more than 200% compared with our corrected precipitation estimate of 913 ± 323 $mm\ y^{-1}$ (399 ± 141 $km^3\ y^{-1}$ (Fig. 5)). The greatest corrected annual precipitation totals in the UIB (1271 $mm\ y^{-1}$) are observed in the elevation belt between 3750 m to 4250 m (compared to 403 $mm\ y^{-1}$ for the uncorrected case). In absolute terms the main water-producing region is located in the elevation belt between 4250 m and 4750 m where approximately 78 km^3 of rain and snow precipitates annually.

In the most extreme case, precipitation is underestimated by a factor 5 to 10 in the region where the Pamir, Karakorum and Hindu-Kush ranges intersect (Fig. 5). Our inverse modelling shows that the large glacier systems in the region can only be sustained if snowfall in their accumulation areas totals around 2000 $mm\ y^{-1}$ (Hewitt, 2007). This is in sharp contrast to precipitation amounts between 200 and 300 $mm\ y^{-1}$ that are reported by the gridded precipitation products (Fig. 1). Our results match well with the few studies on high-altitude precipitation that~~which~~ are available. Annual accumulation values between 1000 and 3000 *mm* ~~are~~ have been reported for accumulation pits above 4000 m~~eter~~ in the Karakoram (The Batura Glacier Investigation Group, 1979; Wake, 1989; Winiger et al., 2005). Our results show that the highest precipitation amounts are found along the monsoon-influenced southern Himalayan arc with values up to 3000 $mm\ y^{-1}$, while north of the Himalayan range the precipitation decreases quickly towards a vast dry area in the north-eastern part of the UIB (Shyok sub-basin). In the north-western part of the UIB, westerly storm systems are expected to generate considerable amounts of precipitation at high altitude.

Our results reveal a strong relation between elevation and precipitation with a median PG for the entire upper Indus of 0.0989 $\%\ m^{-1}$, but with large~~r~~ regional differences. Median precipitation gradients in the Hindu-Kush and Karakoram ranges (0.260 $\%\ m^{-1}$ and 0.119 $\%\ m^{-1}$ respectively) are significantly larger than those observed in the Himalayan range, e.g 0.044 $\%\ m^{-1}$ (Fig. 6). In the Hindu-Kush, for example, for every 1000 *m* elevation rise, precipitation increases by 260% with respect to APHRODITE, which is based on valley floor precipitation. In combination with a higher HMAX in the Hindu-Kush and the Karakoram (e.g. 5500 m versus 4500 m in the Himalayas, (Hewitt, 2007; Immerzeel et al., 2014a; Putkonen, 2004; Seko, 1987; Winiger et al., 2005)~~)~~ this suggests that westerly airflow indeed has a higher tropospheric altitude and that the interplay between elevation and precipitation is stronger for this type of precipitation. Further research should thus focus on the use of high resolution cloud-resolving weather models (Collier et al., 2014; Mölg et al., 2013) for this region to further resolve seasonal topography-precipitation interaction at both synoptic and valley scales.

The estimated precipitation is considerably higher than what was reported in previous studies. Several studies have used TRMM products to quantify UIB precipitation (Bookhagen and Burbank, 2010; Immerzeel et al., 2009, 2010) and they show average annual precipitation values around 300 *mm*. It was also noted that the water balance was not closing and average annual river runoff at Tarbela exceeded the TRMM precipitation (Immerzeel et al., 2009). Two possible reasons have been suggested to explain this gap: (i) the high altitude precipitation is underestimated, (ii) the glaciers are in a significant negative imbalance (Immerzeel et al., 2009). Since the ICESat study and several other geodetic mass balance studies (Gardelle et al., 2013; Kääb et al., 2012b) it has become clear that the glaciers in this region are~~e~~ not experiencing a significant ice loss and that this cannot be the explanation for the missing water in the water balance. This supports our conclusion that it is the high altitude precipitation that has been underestimated. A study based on long term observations of Tarbela inflow also confirm our results (Reggiani and Rientjes, 2014). In this study the total UIB precipitation above Tarbela is estimated to be 675 ± 100 *mm* y^{-1} and the difference remaining between our results may stem from the fact that the UIB we consider is twice the size of the area above the Tarbela, the different approach used to estimate actual ET, the different period considered and their assumption that ice storage has not changed between 1961 and 2009.

~~3.1~~3.2 Uncertainty

We estimated the uncertainty in the modelled precipitation field with the standard deviation (σ) of the 10,000 realizations (Fig. 5c). The signal-to-noise ratio is satisfactory in the entire domain, e.g. the σ is always considerably smaller than the average precipitation with an average coefficient of variation of 0.35. The largest absolute uncertainty is found along the Himalayan arc and this coincides with the precipitation pattern found here. Strikingly, the region where the underestimation of precipitation is largest, at the intersection of the three mountain ranges in the northern UIB, is also an area where the uncertainty is small even though precipitation gradients are large.

By running a multiple regression analysis after optimizing the PGs we quantify the contribution of each parameter to the total uncertainty. The largest source of uncertainty in our estimate of UIB high altitude precipitation stems from the MB estimates, followed by the DDFdf, DDFd, TS, HREF and HMAX, although regional differences are considerable (Fig. 7~~6~~). The MB constrains the precipitation gradients and thereby exerts a strong control on the corrected precipitation fields, in particular because the intra-zonal variation in MB is relatively large (Table 1). Thus, improved spatial monitoring techniques of the MB are expected to greatly improve precipitation estimates.

Fig. 8~~7~~ shows the result of uncertainty analysis associated to the spatial correlation of the parameters, which reveals that the impact on the average corrected precipitation is limited. Locally there are minor differences in the corrected precipitation amounts, but overall the magnitude and spatial patterns are similar. However, there are considerable differences in the uncertainty. The lowest uncertainty is found for the fully uncorrelated case, the fully correlated case has the highest uncertainty whereas the intermediate case is in between both. For the fully correlated case all glacier systems have the same parameter set for the specific realization and this results in a larger final uncertainty. In the uncorrelated case each glacier system has a different randomly sampled parameter set and this reduces the overall uncertainty as it spatially attenuates the variation in precipitation gradients.

3.3 Validation

The corrected precipitation is validated independently by a comparison to published average annual runoff data of 27 stations (Table 2). Runoff estimates based on the corrected precipitation agree well with the average observed annual runoff (Fig. 9~~8~~, top panel). The runoff estimated for the uncorrected APHRODITE is consistently lower than the observed runoff, and in some occasions even negative. Runoff estimates were also made based on the ERA-INTERIM and TRMM precipitation products. The TRMM results yield a similar underestimation as the uncorrected APHRODITES product, but the runoff estimates based on the ERA-INTERIM precipitation agrees reasonably well with the observations. However the coarse resolution (~70 km^2) (Fig. 1) is problematic and cannot be used to reproduce the mass balance (Fig. 11~~9~~). Averaged over large catchments the precipitation may be applied for hydrological modeling, but at smaller scales there are likely very large biases. As a result, the observed glacier mass balances cannot be reproduced when the ERA-INTERIM dataset is used.

The zonal water balance analysis (Fig. 9, bottom panels) reveals that the water balance is much more realistic when the corrected precipitation is used. Although the uncertainties are considerable, our analysis shows that the Himalaya and Hindu-Kush zones are about twice as wet as the Karakoram zone. For all three zones the glacier mass imbalance only plays a marginal role in the overall water balance and about 60% of the total precipitation runs off while 40% is lost through evapotranspiration. Notable~~y~~ are the values for *Corg*, which represents the water balance gap in case the uncorrected precipitation is used, are approximately 500 $mm\ y^{-1}$ in all three zones. Our validation does not take into account groundwater fluxes and we have assumed that over the observed period from 2003 until 2007 there is no net loss or gain of groundwater in the upper Indus basin. We do acknowledge that groundwater may play an important role in the hydrology. A study in the Nepal Himalaya shows that fractured basement aquifers fill during the monsoon and they purge in the post-monsoon thus causing a natural delay in runoff (Andermann et al., 2012). However this does not imply significant net gains or losses over multiple year periods, which is what we consider. A second component that we have not considered and that may play a role in the high altitude water cycle is sublimation. There are some indications that wind redistribution and sublimation may play a considerable role in the high altitude water balance (Wagnon et al., 2013). However our PGs are constrained on the observed mass balance, hence our precipitation can be considered as a net precipitation and sublimation losses are thus accounted for.

In Fig. 10 the Budyko-Turc plot is shown to confirm the physical realism of our results. Those dots located in the hatched part of the graph are physically less ~~un~~realistic. For the uncorrected case almost all dots (open dots) are above the Q/P = 1 line. For the corrected case the Q/P values are much more plausible, however there many catchments which are located slightly to the right side of the theoretical Budyko line, meaning that the Q is smaller than the difference between P and ETp. Possible deviation can potentially be explained by uncertainties in observed flows and ETp estimates, the fact the in glacierized catchments the theoretical Budyko curve may be different because a glacier imbalance can be an additional water balance term that is unaccounted for, a too short time period that is used to construct the water balance and finally that some of the discharge observations do not align in time with the rest of the water balance terms. Overall we conclude though that the use of the corrected precipitation results in physically more realistic results, where the water balance could be closed and no significant amount of precipitation input is missing.

Fig 11 shows how the average simulated zonal glacier mass balance using the corrected, the APHRODITES, the ERA-Interim and the TRMM precipitation datasets. It shows that none of the precipitation products can reproduce the observed mass balance. Mostly the mass balances are underestimated which is consistent with and underestimation of the precipitation. The ERA-Interim dataset overestimates the mass balance in the Himalaya, but underestimates the mass balances in the other two zones as result of the coarse resolution.

~~Our results reveal a strong relation between elevation and precipitation with a median PG for the entire upper Indus of 0.0989 % m^{-1}, but with larger regional differences. Median precipitation gradients in the Hindu Kush and Karakoram ranges (0.260 % m^{-1} and 0.119 % m^{-1} respectively) are significantly larger than those observed in the Himalayan range, e.g 0.044 % m^{-1} (Fig. 10). In the Hindu Kush, for example, for every 1000 m elevation rise, precipitation increases by 260% with respect to APHRODITE, which is based on valley floor~~

~~precipitation. In combination with a higher HMAX in the Hindu-Kush and the Karakoram (e.g. 5500 m versus 4500 m in the Himalayas (Hewitt, 2007; Immerzeel et al., 2014a; Putkonen, 2004; Seko, 1987; Winiger et al., 2005)) this suggests that westerly airflow indeed has a higher tropospheric altitude and that the interplay between elevation and precipitation is stronger for this type of precipitation. Further research should thus focus on the use of high resolution cloud-resolving weather models (Mölg et al., 2013) for this region to further resolve seasonal topography-precipitation interaction at both synoptic and valley scales.~~

4 Conclusions

In this study we inversely model high altitude precipitation in the upper Indus basins from glacier mass balance trends derived by remote sensing. Although there are significant uncertainties, our results unambiguously show that high altitude precipitation in this region is underestimated and that the large glaciers here can only be sustained if high altitude accumulation is much higher than most commonly used gridded data products.

Our results have an important bearing on water resources management studies in the region. The observed gap between precipitation and streamflow (Immerzeel et al., 2009) (with stream flow being larger) cannot be attributed to the observed glacier mass balance (Kääb et al., 2012a), but is most likely the result of an underestimation of precipitation, as also follows from this study. With no apparent decreasing trends in precipitation (Archer and Fowler, 2004) the observed negative trends in stream flow in the glacierised parts of the UIB should thus be primarily attributed to decreased glacier and snow melt (Sharif et al., 2013) and increased glacier storage (Gardelle et al., 2012). In a recent study the notion of ~~of~~ negative trends in UIB runoff was contested and based on a recent analysis (1985 - 2010) it was concluded that runoff of Karakoram rivers is increasing (Mukhopadhyay and Khan, 2014b). The study suggests that increase glacier melt during summer is the underlying reason, which in combination with positive precipitation trends in summer does not contradict the neutral glacier mass balances in the region. From all of these studies it becomes apparent that precipitation is the key to understanding behaviour of glacier and hydrology at large in the UIB. The precipitation we estimate in this study differs considerably, in magnitude and spatial distribution, from datasets that are commonly used in design of reservoirs for hydropower and irrigation and as such it may have a significant impact and improve such planning processes.

The water resources of the Indus River play an important geopolitical role in the region, and our results could contribute to the provision of independent estimates of UIB precipitation.

We conclude that the water resources in the UIB are even more important and abundant than previously thought. Most precipitation at high altitude is now stored in the glaciers, but when global warming persists and the runoff regime becomes more rain dominated, the downstream impacts of our findings will become more evident.

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Table 1. Averages (μ) and standard deviations (σ) of predictors for the precipitation gradient. Values and ranges are based on literature as follows: HREF, HMAX: (Hewitt, 2007, 2011; Immerzeel et al., 2012b, 2014b; Putkonen, 2004; Seko, 1987; Winiger et al., 2005), DDF*d*, DDF*df*: (Azam et al., 2012; Hagg et al., 2008; Immerzeel et al., 2013; Mihalcea et al., 2006; Nicholson and Benn, 2006), MB: (Kääb et al., 2012a)

Variable	**Acronym**	**Distribution**	**μ**	**σ**
Reference elevation (*m*)	HREF	log-Gaussian	2500	500
Maximum elevation (*m*)	HMAX	log-Gaussian		
Himalaya			4500	500
Karakoram			5500	500
Hindu-Kush			5500	500
Degree day factor debris covered glaciers ($mm\ °C^{-1}\ d^{-1}$)	DDFd	log-Gaussian	2	2
Degree day factor debris free glaciers ($mm\ °C^{-1}\ d^{-1}$)	DDFdf	log-Gaussian	7	2
Threshold slope ($m\ m^{-1}$)	TS	log-Gaussian	0.2	0.05
Mass balance ($m\ w.e.\ y^{-1}$)	MB	Gaussian		
Himalaya			-0.49	0.57
Hindu-Kush			-0.21	0.76
Karakoram			-0.07	0.61

Table 2. Runoff stations used for validation. Catchment areas are delineated based on SRTM DEM. * = calculated based on discharge provided by the Pakistan Water and Power Development (WAPDA), ** = based on (Mukhopadhyay and Khan, 2014a), *** = based on (Sharif et al., 2013), **** = based on (Archer, 2003), ***** = based on (Khattak et al., 2011).

Station	**LAT**	**LON**	**Area (km^2)**	**Observed Q ($m^3 s^{-1}$)**	**Period**
Besham Qila*	34.906	72.866	198741	2372.2	2000-2007
Tarbela inflow*	34.329	72.856	203014	2370.3	1998-2007
Mangla inflow*	33.200	73.650	29966	831.8	1998-2007
Marala inflow*	32.670	74.460	29611	956.5	1998-2007
Dainyor bridge*	35.925	74.372	14147	331.8	1998-2004
Skardu - Kachura**	35.435	75.468	146200	1074.2	1970-1997
Partab Bridge**	35.767	74.597	177622	1787.9	1962-1996
Yogo**	35.183	76.100	64240	359.4	1973-1997
Kharmong**	34.933	76.217	70875	452.3	1982-1997
Gilgit**	35.933	74.300	13174	286.7	1960-1998
Doyian**	35.550	74.700	4000	135.7	1974-1997
Chitral**	35.867	71.783	12824	271.9	1964-1996
Kalam**	35.467	72.600	2151	89.6	1961-1997
Naran**	34.900	73.650	1181	48.1	1960-1998
Alam bridge**	35.767	74.600	28201	644	1966-1997
Chakdara**	34.650	72.017	5990	178.9	1961-1997
Karora**	34.900	72.767	586	21.2	1975-1996
Garhi Habibullah**	34.450	73.367	2493	101.8	1960-1998
Muzafferabad***	34.430	73.486	7604	357	1963-1995
Chinari***	34.158	73.831	14248	330	1970-1995
Kohala***	34.095	73.499	25820	828	1965-1995

Kotli***	33.525	73.890	2907	134	1960-1995
Shigar**	35.422	75.732	6681	202.6	1985-1997
Phulra**	34.317	73.083	1106	19.2	1969-1996
Daggar**	34.500	72.467	534	6.9	1969-1996
Warsak****	34.100	71.300	74092	593	1967-2005
Shatial Bridge**	35.533	73.567	189263	2083.2	1983-1997

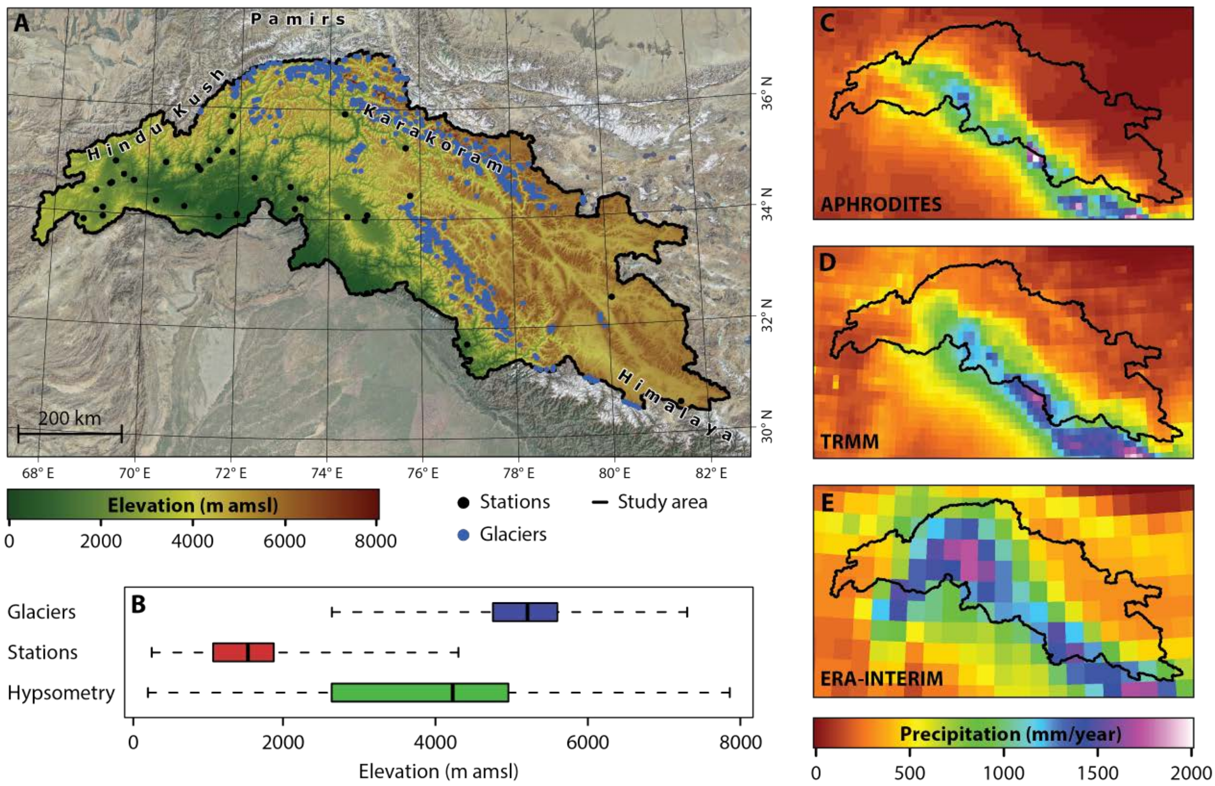


Figure 1. Overview of the UIB, basin hypsometry and three gridded precipitation products. Panel A shows the digital elevation model and the location of the major glacier systems (area > 5 km^2) and the available stations in the Global Summary of the Day (GSOD) of the World Meteorological Organization (WMO). Panel B shows boxplots of the elevation distribution of the basin, the large glacier systems and the GSOD stations. Panel C to E show the average gridded annual precipitation between 1998-2007 for the APHRODITE(Yatagai et al., 2012), TRMM (Huffman et al., 2007) and ERA-INTERIM (Dee et al., 2011) datasets.

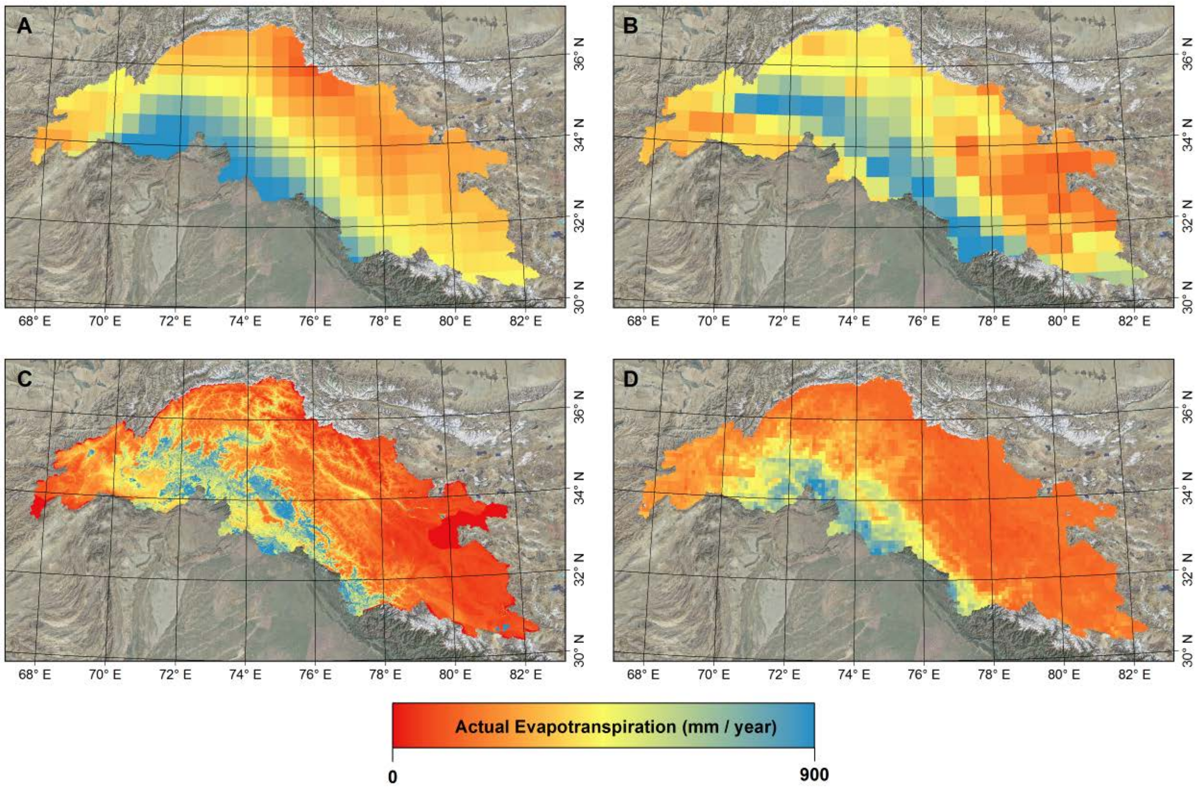


Figure 2. Average annual actual evapotranspiration between 2003 and 2007 for ERA-Interim (A), MERRA (B), ET-Look (C) and PCRGLOB-WB (D).

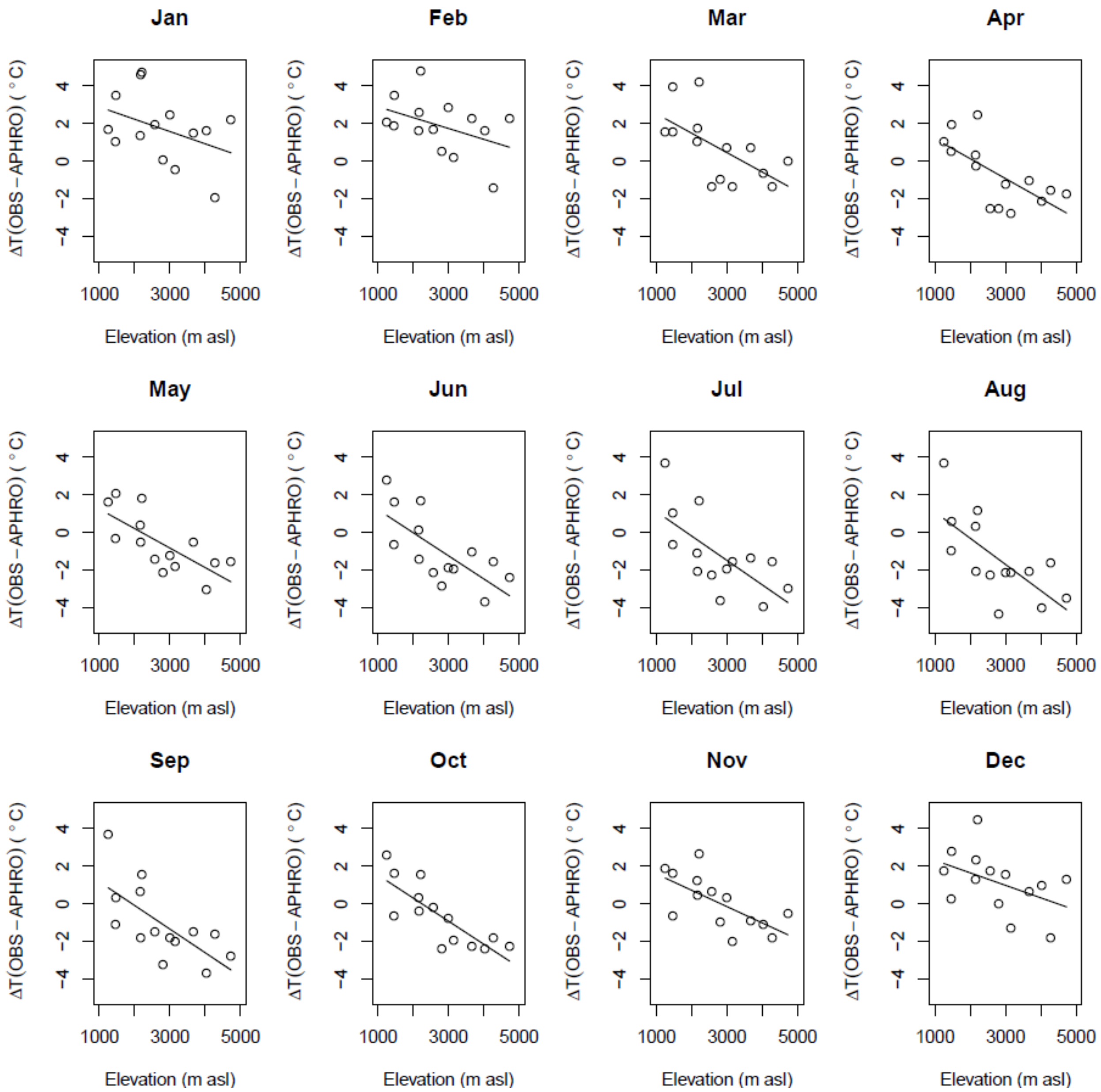


Figure 3. Monthly relation between observed temperatures at meteorological stations (OBS) and the APHRODITE temperature fields (APHRO) (Yasutomi et al., 2011).

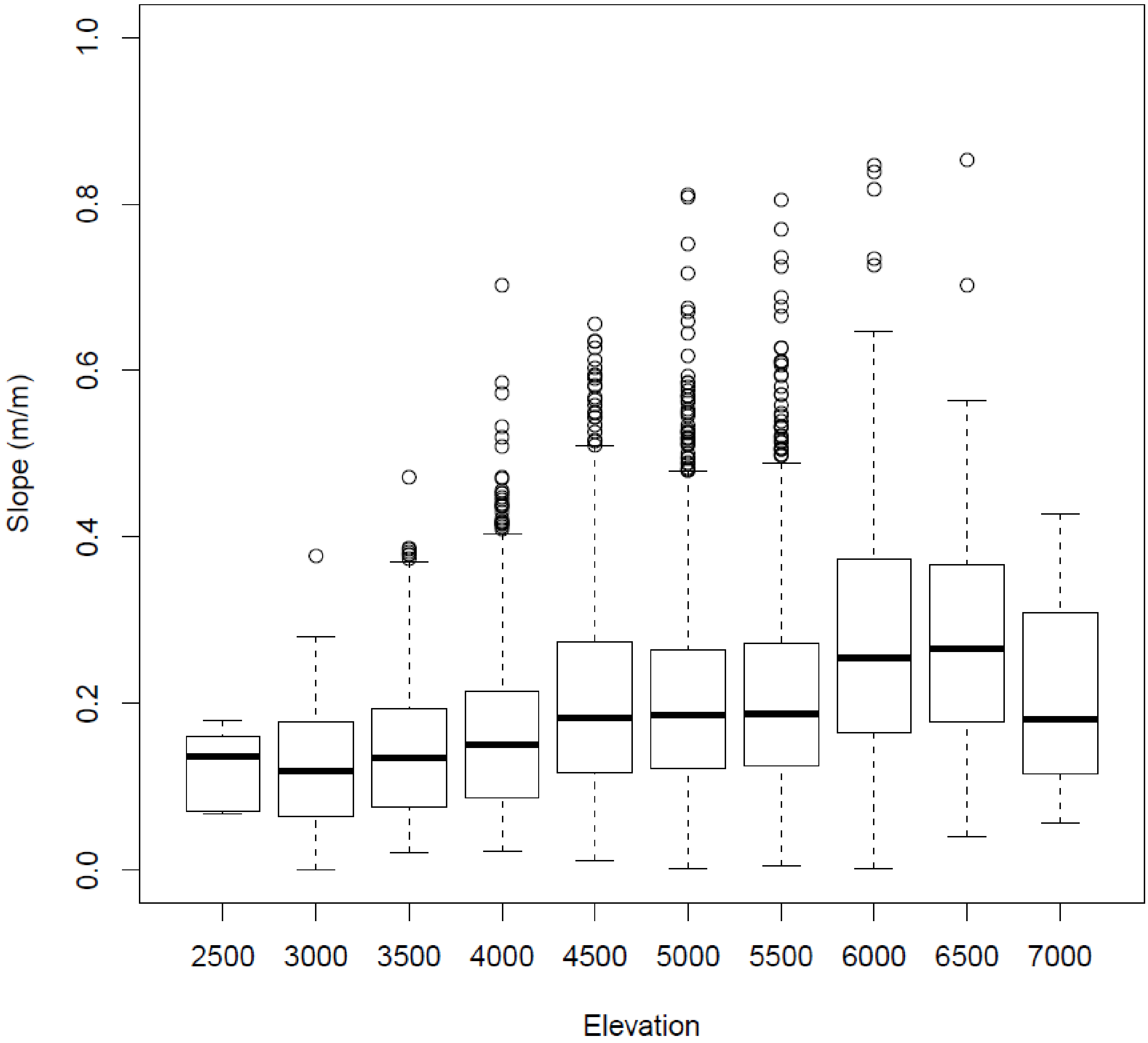


Figure 4. Boxplots of slopes of glacierised areas per elevation bin.

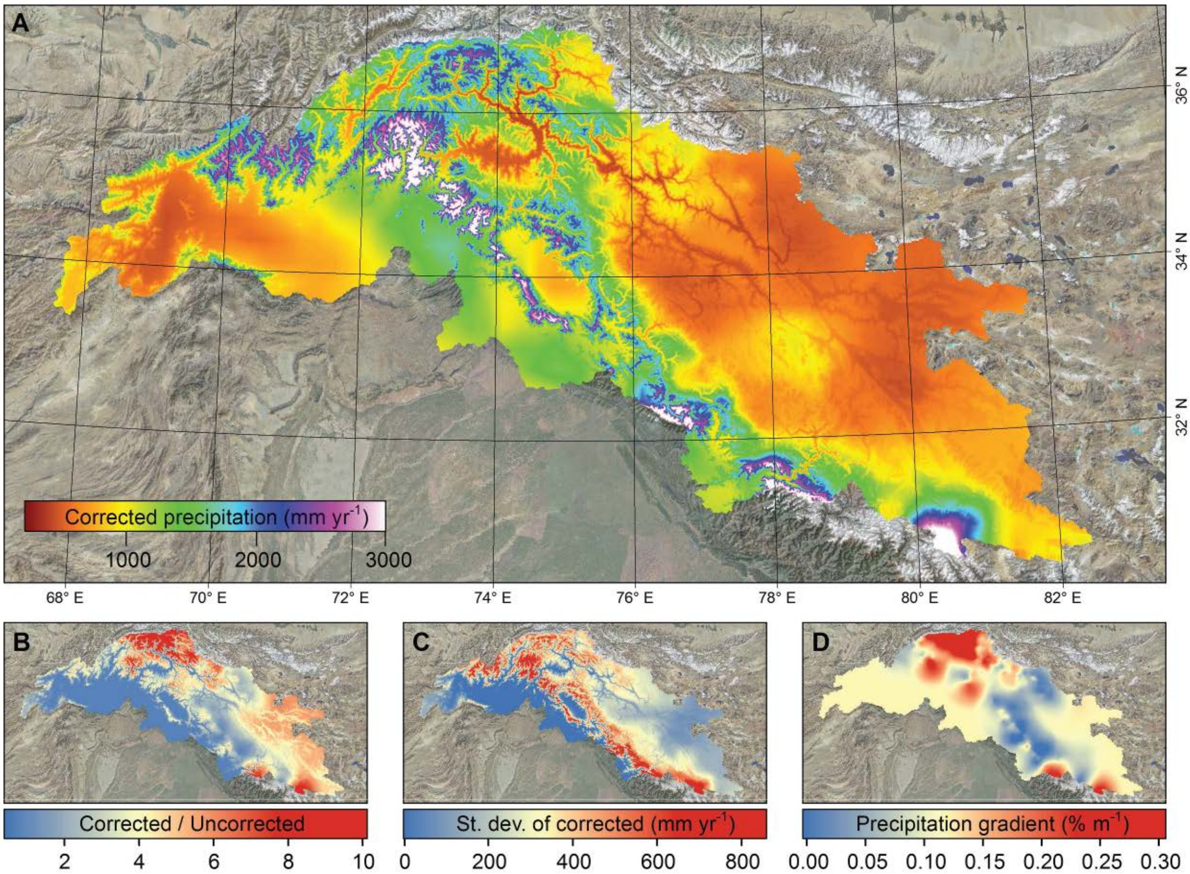


Figure 5. Corrected precipitation and estimated uncertainty for the UIB for the case with intermediate spatial correlation between model parameters. Panel A shows the average modelled precipitation field based on 10,000 simulations for the period 2003-2007, Panel B shows the ratio of corrected precipitation to the uncorrected APHRODITE precipitation for the same period, Panel C shows the standard deviation of the 10,000 simulations and panel D shows the average precipitation gradient.

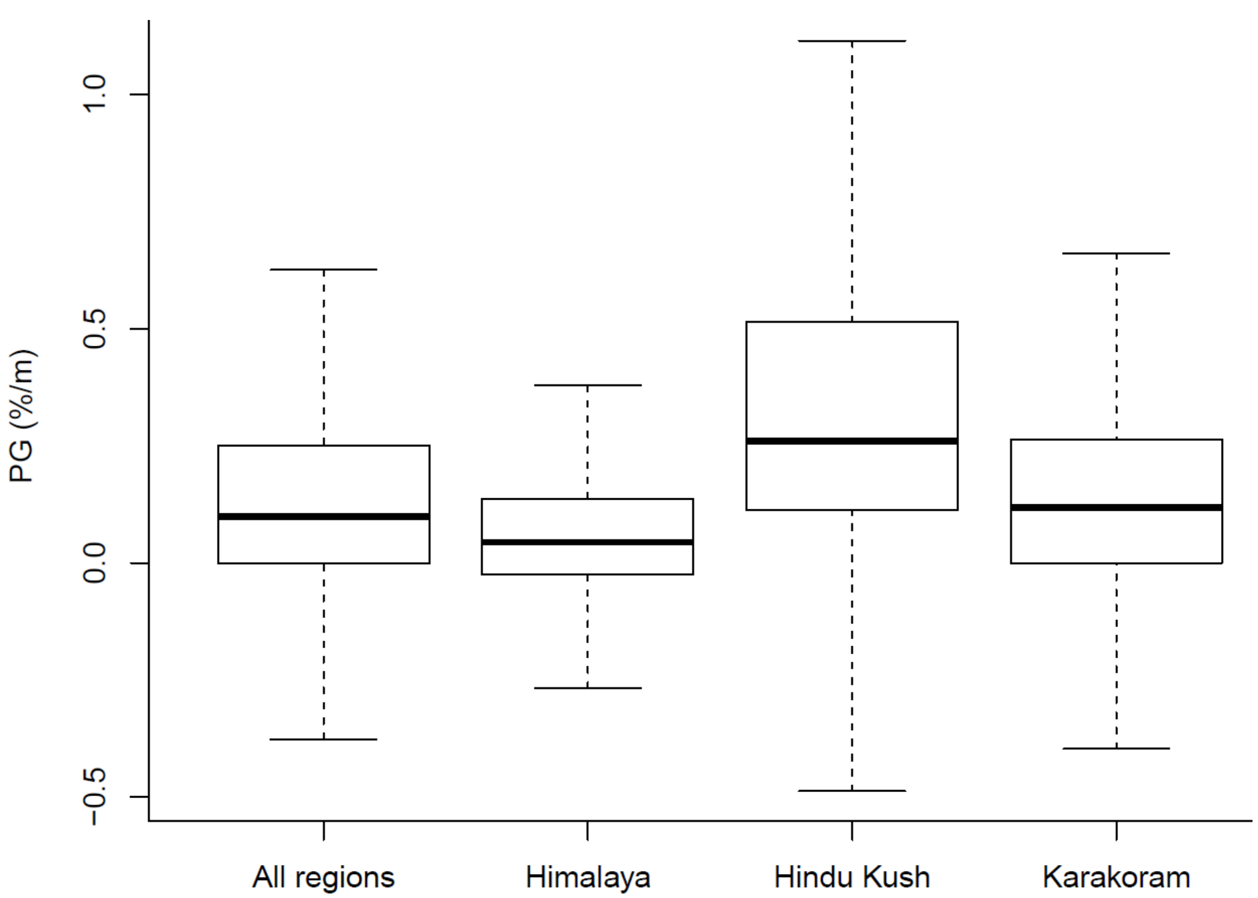


Figure 6. Box plots of precipitation gradients for the entire UIB and for the three regions separately.

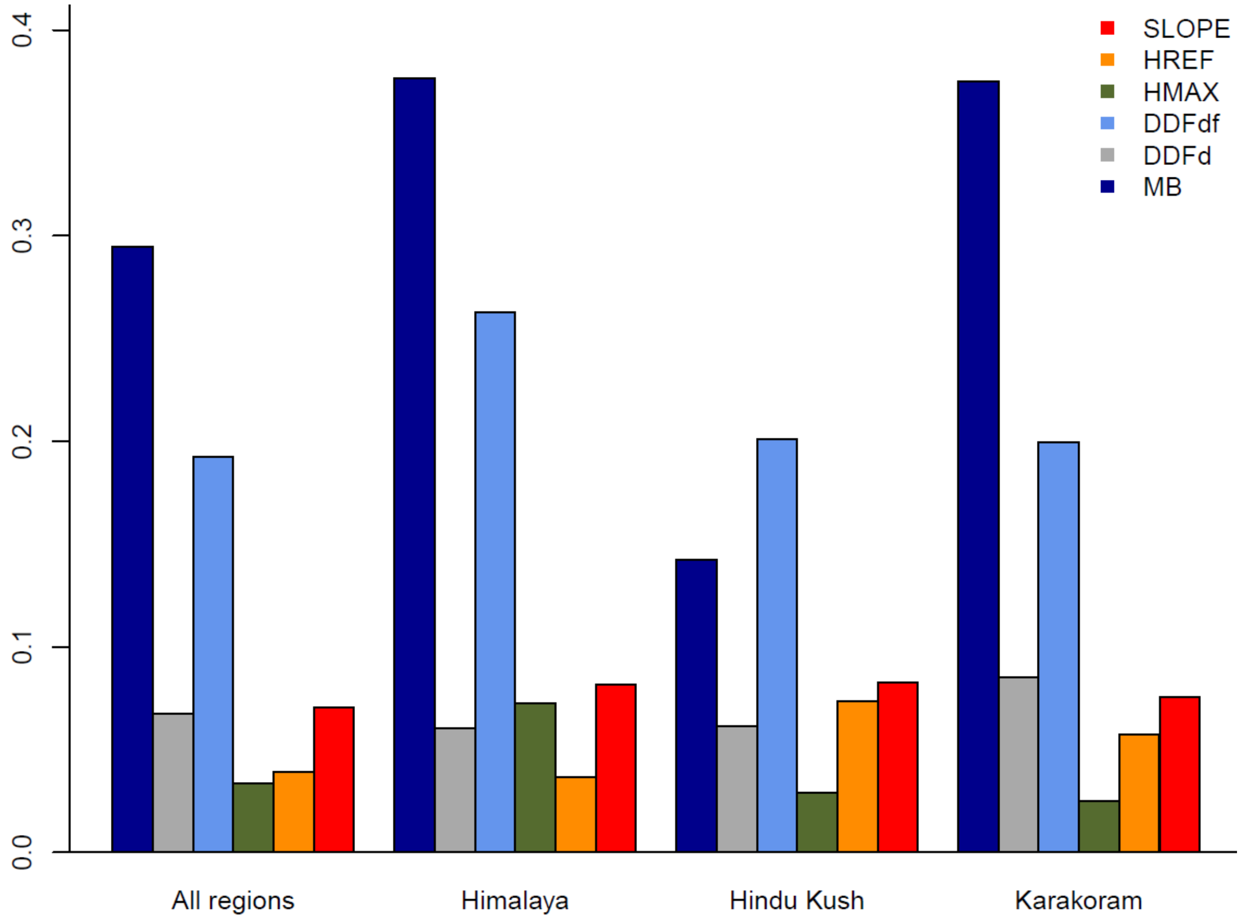


Figure 7~~6~~. Normalized weights of multiple regression of the precipitation gradients by the predictors slope (slope threshold for avalanching to contribute to accumulation), HREF (base elevation from which lapsing starts), HMAX (elevation with peak precipitation), DDFd (degree day factor for debris covered glaciers), DDFdf (degree day factor for debris free glaciers) and the MB (mass balance of the glacier).

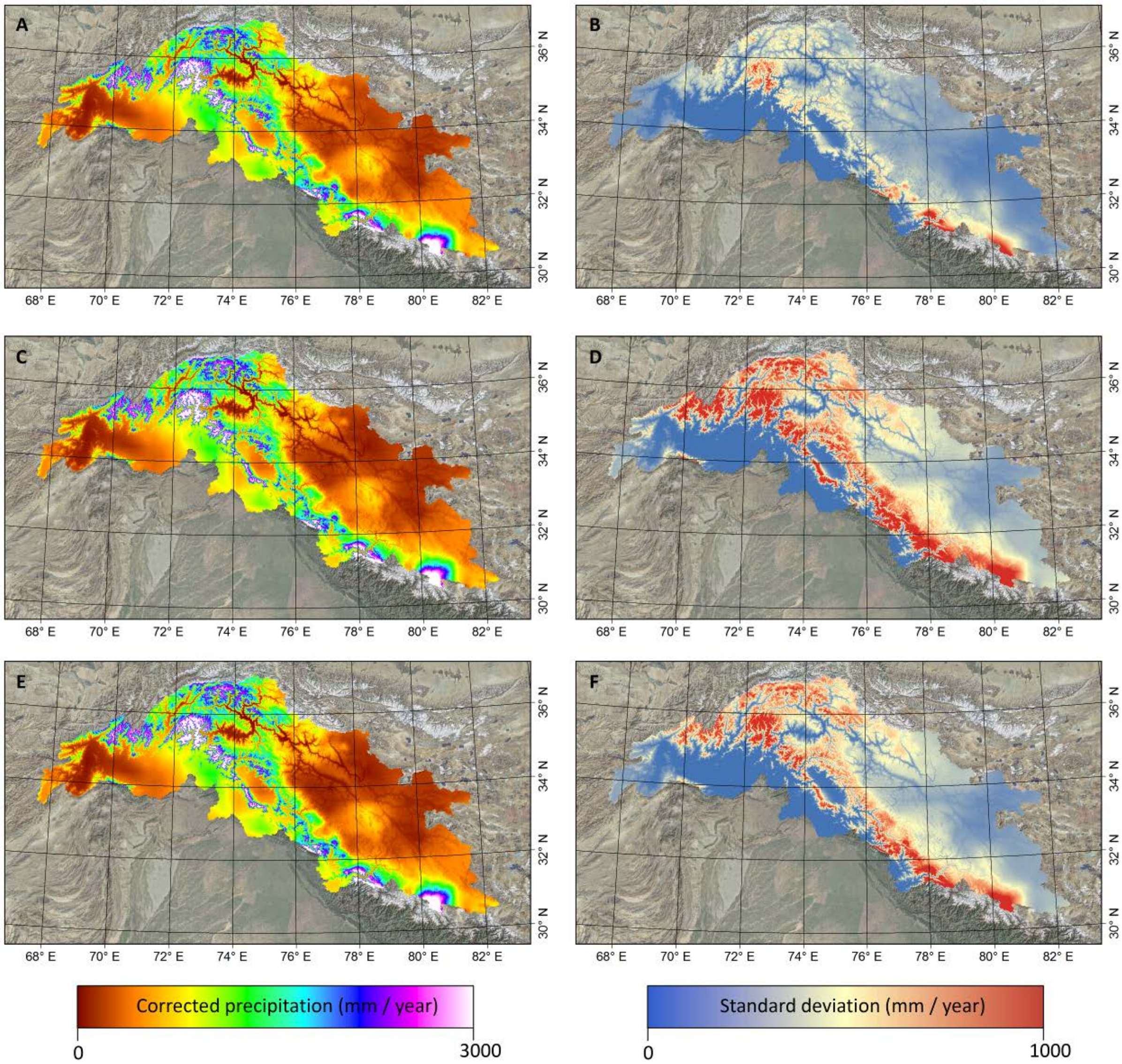
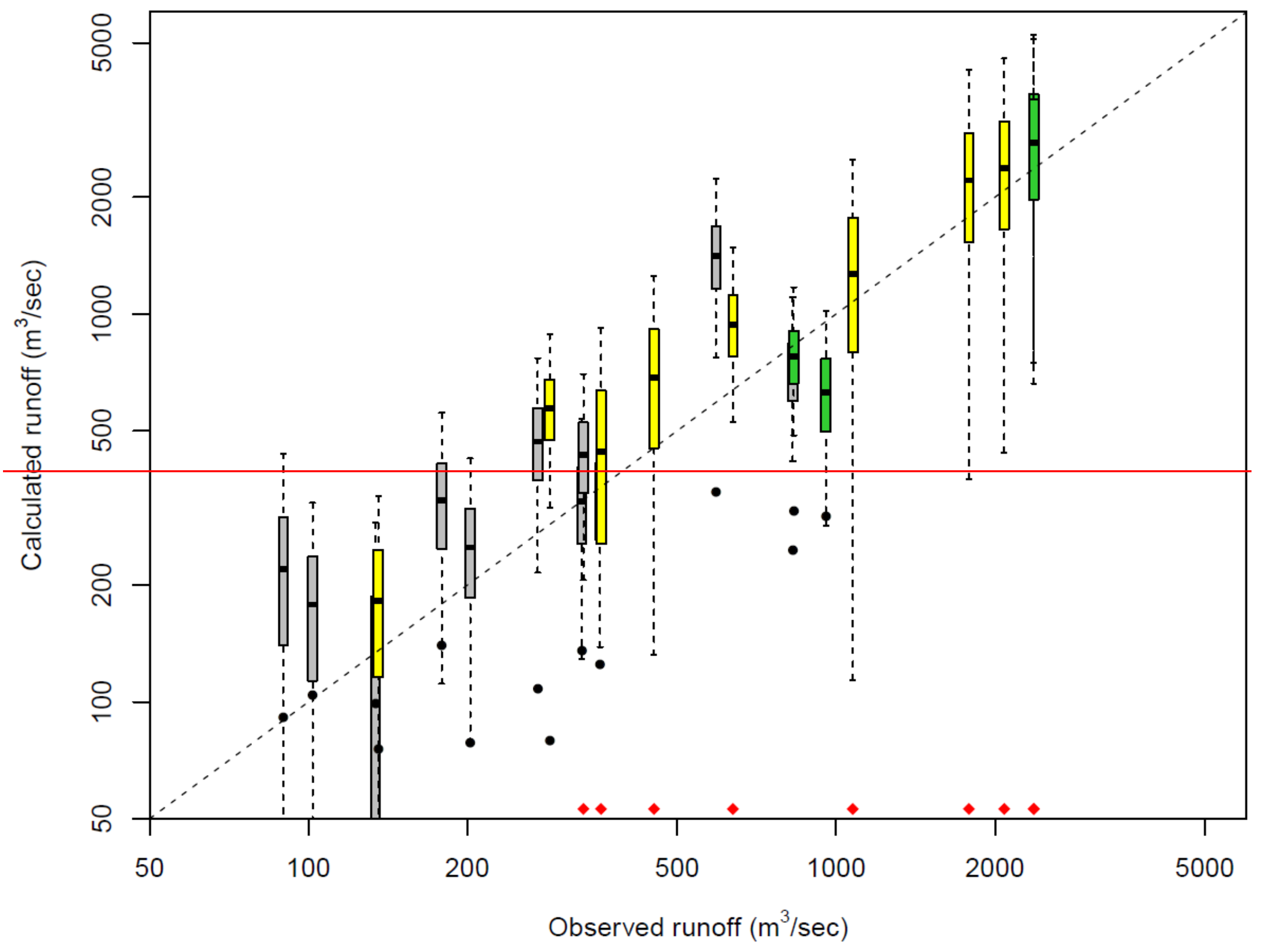


Figure ~~7~~8. Impact of spatial correlation of parameters on the corrected precipitation field and associated uncertainty. The top panels show the corrected precipitation field (panel A) and uncertainty (panel B) for the fully uncorrelated case. The middle panels (D,E) for the fully correlated case and the bottom panels (E,F) for the intermediate case.


Calculated runoff (m3/sec)
Observed runoff (m3/sec)
50
100
200
500
1000
2000
5000

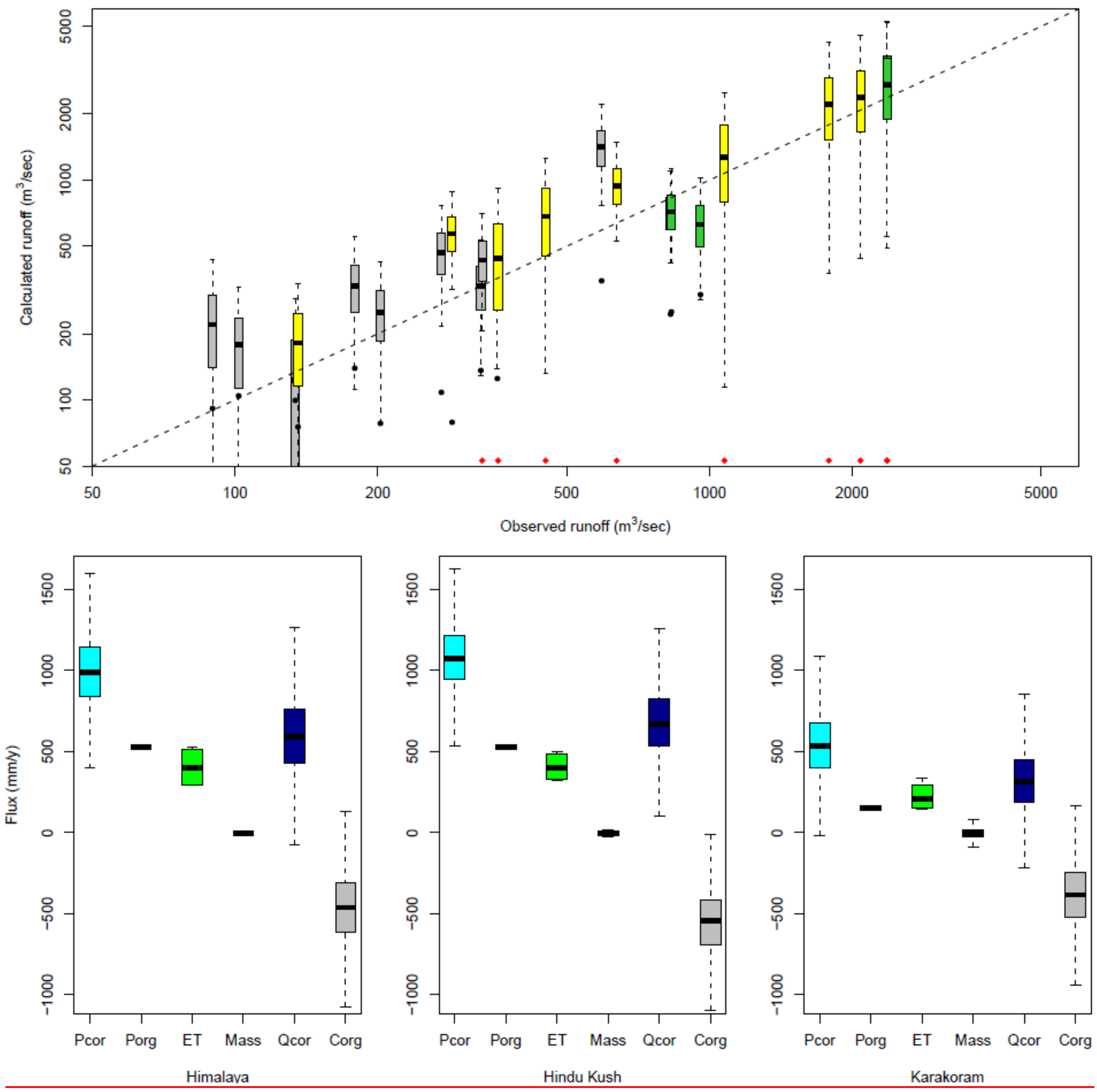


Figure ~~8~~9. Validation of the precipitation correction using observed discharge (Table 2). Top panel: The box plots are based on the runoff estimate based on 10,000 corrected precipitation fields (grey: stations for which the observed record does not coincide with the 2003-2007 period, yellow: stations for which the 2003 – 2007 period is part of the observational record, green: stations for which the observations are based precisely on the 2003 – 2007 period. The black dots and red diamonds (estimated runoff below 50 m^3s^{-1}) show the estimated runoff based on the uncorrected precipitation. Bottom panels: Water balance components of each zone (Pcor = corrected precipitation, Porg = uncorrected APHRODITES precipitation, ET = actual evapotranspiration, Mass = glacier mass balance, Qcor = estimated runoff, Corg = water balance gap in case the Porg is used).

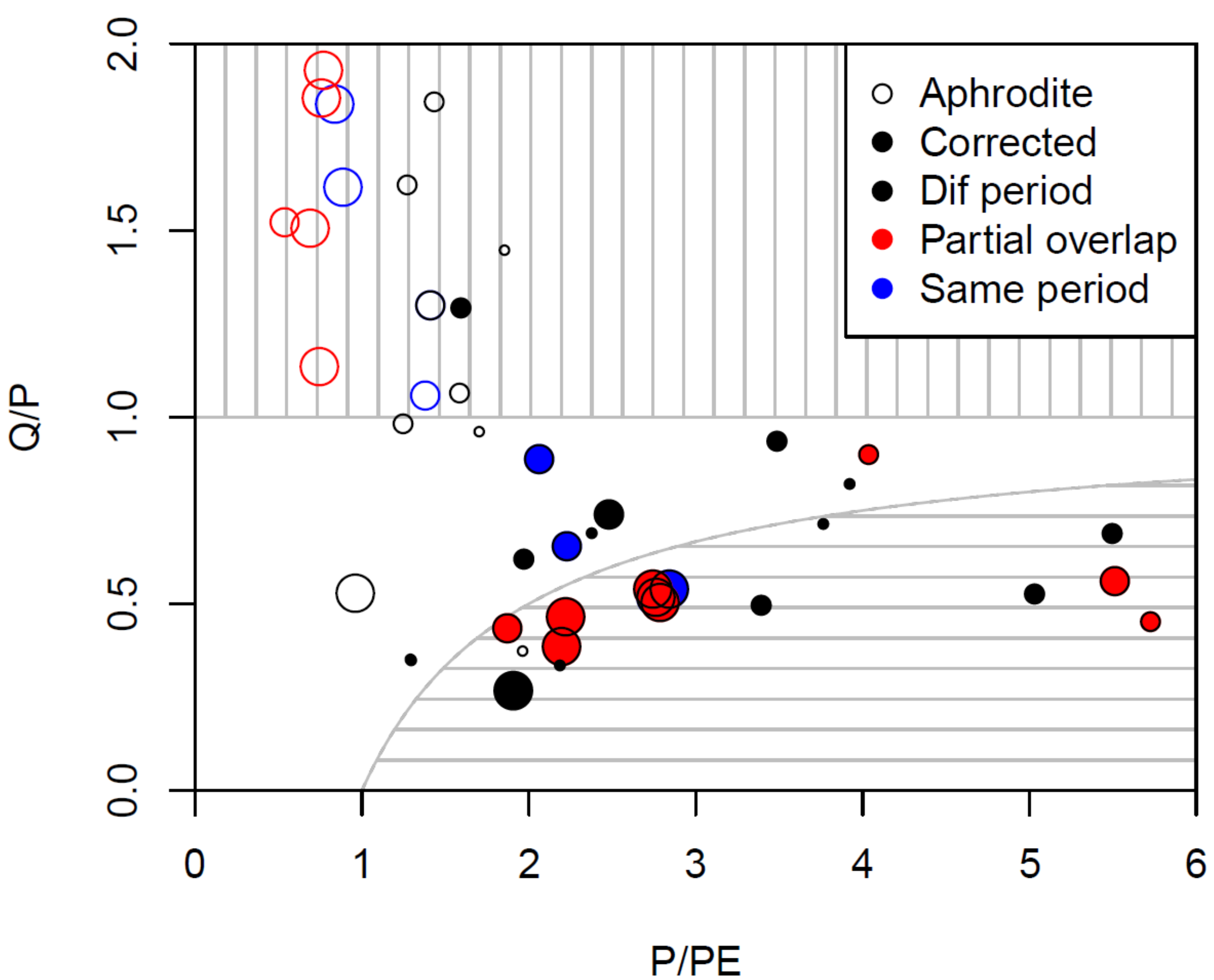


Figure 10. Non-dimensional graphical representation of catchments using their mean runoff, Q, precipitation, P, and potential evapotranspiration, PE. The grey line in the empty centre area represents the theoretical Budyko relationship in the non-dimensional graph. The size of the dots is scaled to the catchment area.

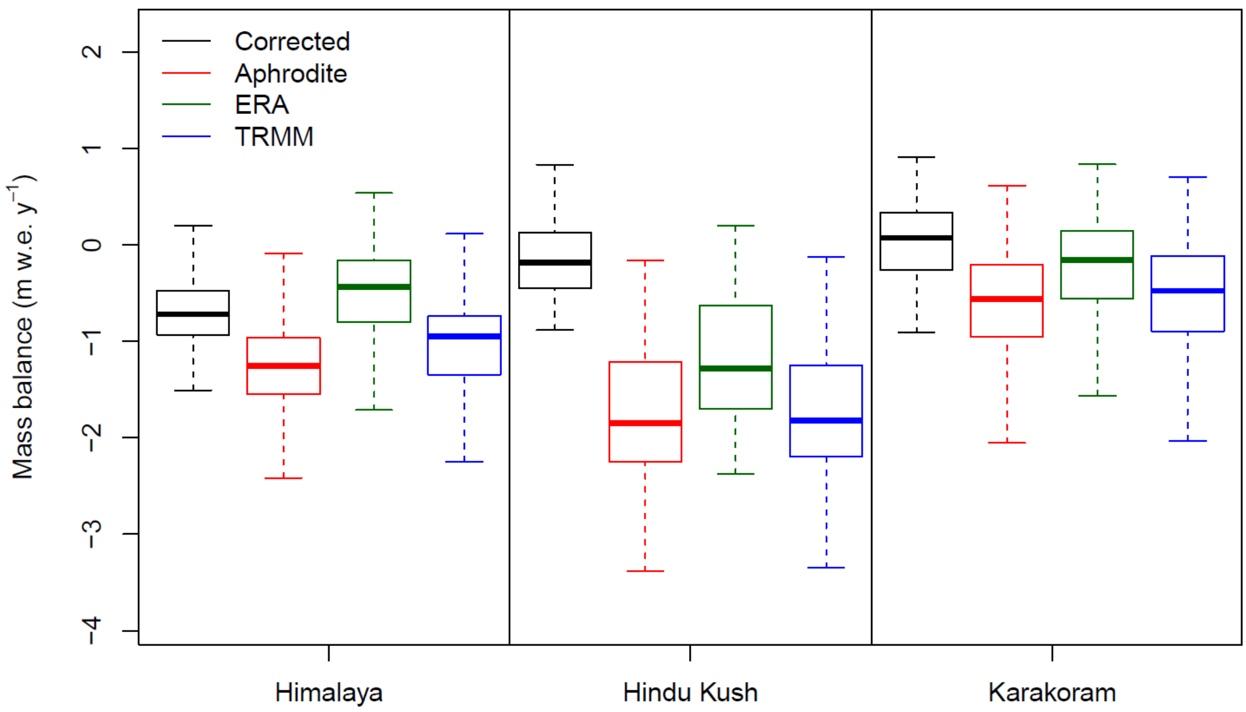
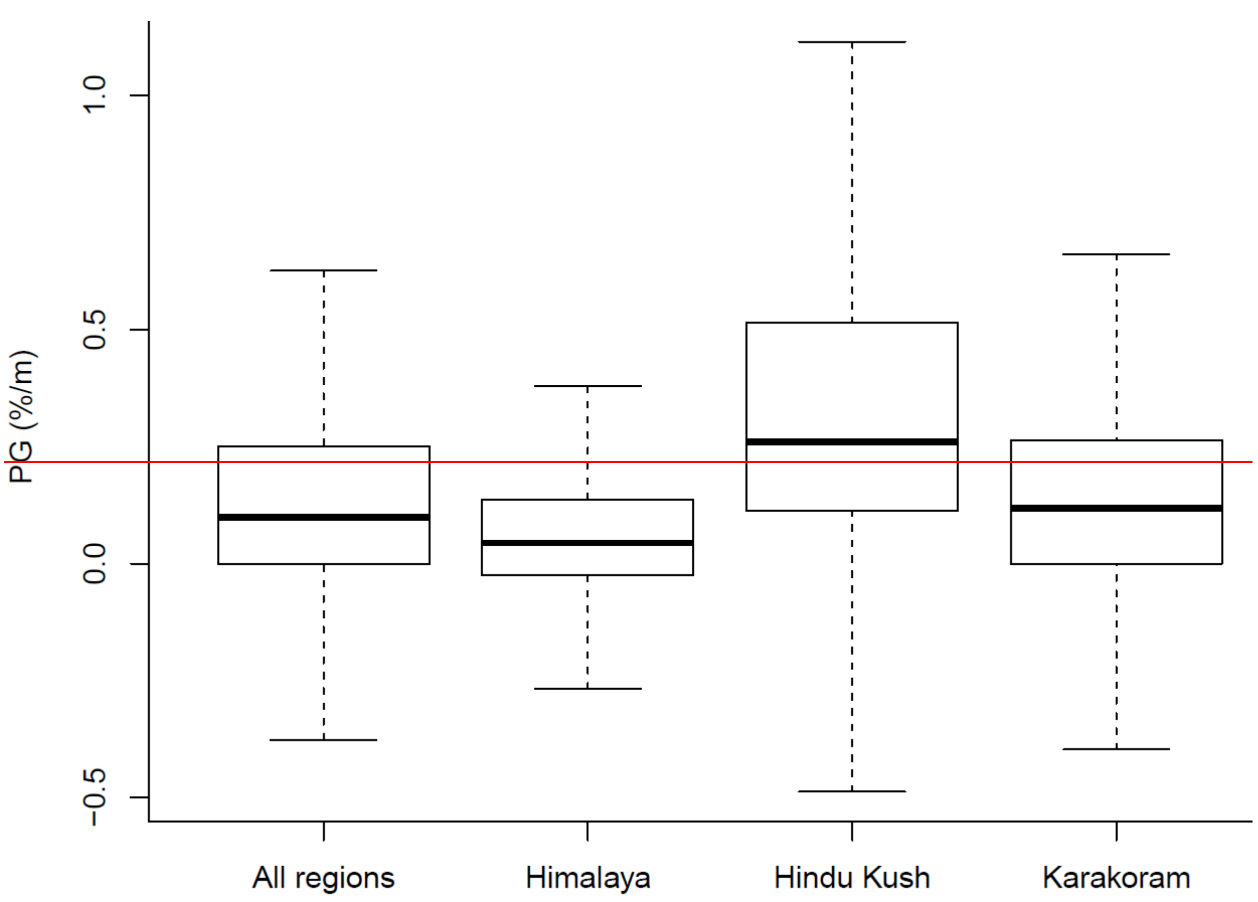


Figure ~~9~~11. Reconstructed mass balances based on the corrected, APHRODITE, ERA-INTERIM and TRMM datasets. The black horizontal dotted line shows the observed mass balance for each zone.



~~Figure 10. Box plots of precipitation gradients for the entire UIB and for the three regions separately.~~