



Supplement of

Technical note: A physically based correction of WRF temperature using atmospheric lapse rates for high-resolution hydrological simulations in a glacierized catchment

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Contents of the supplementary material

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S1 Introduction

The supplementary material presents additional text, equations, figures, and tables along with the main text of the technical note. The order of the presented materials does correspond to the chronology of the technical note. Cross-references to the supplementary material are marked with an “S” in the technical note.

10 S2 Supplementary Text

S2.1 Thermodynamic correction of the WRF relative humidity

As relative humidity (RH) is not an independent variable and depends non-linearly on air temperature and moisture, we performed a sequential, thermodynamically consistent correction of RH considering the corrected WRF temperature. For this, we applied the empirical August-Roche-Magnus (ARM) formula, which is often used in meteorology to calculate the saturation
15 vapor pressure of water at a given temperature, helping determine RH (Alduchov and Eskridge, 1996; Huang, 2018). First, we define the hourly saturation vapor pressure $e_s(T_{\text{uncor},t})$ [hPa] over water and ice surfaces using the uncorrected WRF $T_{\text{uncor},t}$ in Equation S1. In a second step, we perform the same computation using the corrected WRF $T_{\text{cor},t}$ for deriving hourly saturation vapor pressure $e_s(T_{\text{cor},t})$ [hPa]. Lastly, we recompute the corrected hourly RH $RH_{\text{cor},t}$ [%] by holding vapor pressure constant and considering the corrected temperature with Equation S1:

$$20 \quad RH_{\text{cor},t} = RH_{\text{uncor},t} \frac{e_s(T_{\text{uncor},t})}{e_s(T_{\text{cor},t})}, \quad e_s(T_t) = \begin{cases} 6.1094 \exp\left(\frac{17.625 T_t}{243.04 + T_t}\right), & T_t \geq 0^\circ\text{C} \\ 6.1121 \exp\left(\frac{22.587 T_t}{273.86 + T_t}\right), & T_t < 0^\circ\text{C} \end{cases} \quad (\text{S1})$$

where $T_{\text{uncor},t}$ is the hourly uncorrected and $T_{\text{cor},t}$ is the hourly corrected temperature [$^\circ\text{C}$] of a WRF cell. $RH_{\text{uncor},t}$ represents the hourly uncorrected and $RH_{\text{cor},t}$ the corrected RH [%] of a WRF cell. If the hourly T_t is equal to or larger than 0°C , the ARM formula for the water surface is considered. For temperatures smaller than 0°C , we used the ARM formula for the ice surface. Both ARM coefficients for water and ice are taken from Alduchov and Eskridge (1996), which are commonly used
25 in meteorological data processing and humidity studies (e.g., Huang (2018)). We constrained the corrected RH values to the range 0-100 %, meaning values $< 0\%$ were set to 0% and $> 100\%$ were set to 100% .

Due to the cold bias in the uncorrected WRF temperature, the RH adjustment using the ARM formulas resulted in a downward shift in the corrected mean RH across all elevations, as shown in Figure S4. Since the cold bias is more pronounced in the highest glacirized areas of the WRF temperatures, the RH has the most significant impact in the elevation range 2900-3500
30 m a.s.l.. The mean of the uncorrected RH was about 90 % for this distinct elevation range, and had a mean RH of about 73 % after the correction. In addition, the correction reduces the slope of the RH elevation gradient across the upper Kauner Valley from $2.2\% (100\text{ m})^{-1}$ for the uncorrected to $0.3\% (100\text{ m})^{-1}$ for the corrected RH data. Figure S5 underlines the local effect of the RH correction by exemplarily showing the ECDFs of hourly uncorrected and corrected RH at three station locations. The largest difference between corrected and uncorrected RH is observed at the Vernagt Pegelstation. It gets obvious that the RH
35 correction affects all parts of the ECDF, with the lower quantiles (i.e., $p < 0.25$) showing the strongest reduction in RH.

S2.2 Correction of the WRF precipitation

Climate models, including the WRF model, often overestimate the frequency of light rainy days while simultaneously underestimating the total amounts of extreme observed precipitation (Lazoglou et al., 2024). The model's tendency to overestimate the occurrence of light precipitation events is known as the "drizzle bias" (Lazoglou et al., 2024; Velasquez et al., 2020). We follow the approach of Lazoglou et al. (2024) to adjust the number of rainy days based on the assumption that the relationship between observed and simulated rainy days remains the same in time (thresholding). The correction of the dry-days frequency using thresholding has already been applied in several studies before (e.g., Smitha et al., 2018; Velasquez et al., 2020). Thresholding is a widely used, simple, and efficient method for correcting the frequency of wet days. It involves converting all simulated values below a designated threshold to zero (Lazoglou et al., 2024). We adjust the WRF precipitation frequency by computing wet-day frequencies in both monthly observed and simulated datasets. As the "drizzle bias" has a distinct seasonality (Fig. S8) and is consequently more pronounced in the winter period, we computed mean monthly thresholds based on the observed and simulated counts of rainy days. The first step is to compute the mean monthly precipitation days based on daily station data. We select the meteorological station Vergoetschen (1280 m a.s.l.), operated by HD-Tyrol, as it has a long record of daily precipitation without data gaps from 1971. Observed precipitation days are defined as days with precipitation exceeding 0.01 mm d^{-1} . Observed mean monthly precipitation days from 1971 to 2015 are listed in the second column of Table S2.

We estimate monthly-varying precipitation thresholds to identify WRF time steps whose precipitation is to be set to zero for frequency matching. These thresholds are identified based on the absolute minimum error between observed and WRF precipitation days for hourly resolution (Tab. S2). Figure S7 illustrates the optimization approach. In the second step, the hourly WRF precipitation values below the monthly thresholds (Tab. S2) are set to 0, assuming that the monthly-varying precipitation thresholds are stationary over the period 1971 to 2015, across space, and elevation. The correction of WRF precipitation only leads to a reduction of the annual precipitation over the research site by 4 % compared to the uncorrected, as mainly low precipitation frequencies are set to 0. WRF precipitation intensities (i.e., quantiles) are not evaluated or adjusted due to a lack of reliable precipitation observations from higher elevations ($> 2000 \text{ m a.s.l.}$).

Since WaSiM considers the presence of precipitation in its cloudiness estimate, precipitation frequency can indirectly affect other processes, such as the computation of long-wave incoming radiation (Schulla, 2021). We remove low precipitation events using monthly thresholds (Tab. S2) under the assumption that these are stationary over the period 1971 to 2015. However, this assumption is only partly true as the observed monthly precipitation days at Vergoetschen station show an increasing trend (Fig. S9) for the first ten years (i.e., 1971 to 1981). After 1981, the total number of monthly precipitation days seems to be stationary at this site until 2015. Nevertheless, the interannual variability of monthly precipitation days is well visible (Fig. S9). In principle, we could remove the low precipitation events for each year separately, as the precipitation observations from Vergoetschen station overlap with the period of the longer simulation run (1973 to 2015). However, we have deliberately opted for global monthly thresholds to be independent of the simulation period. From our perspective, seasonal or even sub-seasonal threshold values are needed to remove low precipitation frequencies due to the clear tendency concerning observed monthly precipitation days between summer and winter (Fig. S6).

Apart from the precipitation frequency, we have not further investigated the quality of the WRF precipitation, which is a key driver and source of uncertainty in the hydrological model (Meyer et al., 2019). The evaluation of precipitation timing, magnitude, and intensity is particularly limited by station observations due to the high spatial heterogeneity and, most importantly, by the large measurement uncertainty in high-elevation regions ($> 2000 \text{ m a.s.l.}$). The undercatch of precipitation measurements in cold regions has been demonstrated in several studies (e.g., Kochendorfer et al., 2016; Mair et al., 2016). A potential approach for evaluating WRF precipitation magnitudes would be to evaluate them indirectly by closing the water balance, assuming observed streamflow is available. However, this is only possible when there is no significant contribution from glacier melt and permafrost thawing, which is usually unknown. Due to the many sources of uncertainty, we recommend, in particular, evaluating the seasonal precipitation of RCMs in glaciated high mountain regions with seasonal observation data from glacier mass balance programs, as an evaluation only against streamflow is problematic due to the delayed response and the lack of information on the remaining water balance components, such as evapotranspiration and sublimation. There are only a few glacier observatories in the European Alps with long temporal data availability, including streamflow observations. The Glacier Observatory at Vernagtferner would be a good study site for evaluating seasonal precipitation datasets (WGMS, 2025). Unfortunately, interdisciplinary and long-term observations of cryospheric and hydrological processes are still very limited in

high-elevation catchments. Nevertheless, correcting rainfall intensities remains problematic due to limited observational data and non-stationary conditions (Teutschbein and Seibert, 2012; Meyer et al., 2019).

S2.3 Evaluation of corrected seasonal WRF temperature at station locations

Figure S11 compares observed and corrected WRF hourly temperatures by season. The comparison is done for different data subsets based on the availability of hourly temperature records. The seasonal temperature dynamic is well reproduced by the corrected WRF temperature at all observation sites. When comparing the medians of the four boxplots, we observe a slight seasonal tendency of the corrected WRF to overestimate summer temperature. In addition, the standard deviation is under- or overestimated by the corrected WRF temperature for some months. Nevertheless, we cannot observe a systematic bias in the corrected WRF temperatures across different sites and altitudes.

S2.4 Evaluation of elevation-dependent warming in WRF temperature data

In a further analysis, we test whether the increase in WRF temperature shows different trends with elevation in the research area. We first test the significance of the temperature trends using the modified Mann-Kendall test and second, compute the coefficient of a linear model giving the annual temperature at four different altitudes (Tab. S6). The first three elevations correspond to actual meteorological observation stations in Kauner Valley, while the highest (i.e., 3800 m a.s.l.) represents the highest point of the research area. Interestingly, we find stronger warming by 17 % when comparing the lowest and highest locations of the research area, meaning the higher altitudes are experiencing a stronger temperature increase. This observation is also consistent with other studies on elevation-dependent warming in the European Alps (e.g., Rangwala and Miller, 2012; Kuhn and Olefs, 2020).

S2.5 Evaluation of streamflow simulations for longer period (1973-2015)

As almost continuous observed streamflow data are available from 1971 onwards at gauge Gepatschalm, we compare the hourly and daily streamflow over different periods (Fig. S15). The left panels of Figure S15 illustrate the mean of the streamflow in days of the year. The subplots in Figure S15 (a-e) present observed and simulated streamflow for shorter periods of 14 years. Figure S15a shows the earliest period from 1973 to 1987. The blue (T correction) and cyan (T and RH correction) lines of the mean simulated streamflow follow well the seasonality and magnitude of the mean observed streamflow (black line). In general, the annual streamflow regime has a typical pattern of a nival regime. The ECDFs of hourly streamflow (Fig. S15b) indicate a good fit in the mean and high flow parts with both correction products. However, WaSiM tends to underestimate low flows ($p < 0.8$). The mean of hourly streamflow is $2.8 \text{ m}^3 \text{ s}^{-1}$ for the observation and $2.5 \text{ m}^3 \text{ s}^{-1}$ for the simulation with T correction and $2.3 \text{ m}^3 \text{ s}^{-1}$ with T and RH correction.

The period 1987 to 2001, as shown in snapshot Figure S15c, is characterized by slightly higher streamflow magnitudes. The mean observed streamflow is $2.8 \text{ m}^3 \text{ s}^{-1}$ and the simulated mean, considering only T correction, is $2.7 \text{ m}^3 \text{ s}^{-1}$ while the simulation with T and RH correction yields a lower mean streamflow of $2.5 \text{ m}^3 \text{ s}^{-1}$. In general, the summer streamflow period is extending into autumn.

In the following period (2001-2015), the high flow period further extends during the summer months (Fig. S15e). Additionally, the streamflow generation shifts towards spring. As a consequence, the mean observed ($3.1 \text{ m}^3 \text{ s}^{-1}$) and both simulated ($3 \text{ m}^3 \text{ s}^{-1}$ with T correction; $2.9 \text{ m}^3 \text{ s}^{-1}$ with T and RH correction) streamflow increases. The reason for the increased streamflow is the increased temperature (i.e., mean annual temperature is $-3.5 \text{ }^\circ\text{C}$ in 1973-1987 and $-2 \text{ }^\circ\text{C}$ in 2001-2015), which leads to a higher glacier melt contribution (Fig. 8b). Although WaSiM well predicts the observed seasonality, it overestimates the mean streamflow in August and September. However, the overestimation is slightly smaller when using the T and RH correction.

The daily mean of streamflow values computed over the entire simulation period provides a good overview of the general tendency of the WaSiM model for underestimating the low flow period while overestimating streamflow at the end of the ablation period in late summer (Fig. S15g and h). Nevertheless, WaSiM reproduces the seasonal streamflow pattern in the upper Kauner Valley well with both correction products. Over the entire simulation period (1973-2015), the daily KGE is 0.76, percent bias -5.5, and RMSE $2.2 \text{ m}^3 \text{ s}^{-1}$ for the T correction simulation, and the GOFs for the T and RH correction are KGE 0.8, percent bias -11.3, and RMSE $2 \text{ m}^3 \text{ s}^{-1}$. Considering that the model is not calibrated, meaning the model parameters are

not optimized according to the extended simulation period, or for the different corrected forcings, the streamflow results are very satisfying.

130 **S2.6 Comparison of simulated glacier mass with observed geodetic elevation changes**

We performed additional evaluations of the simulated glacier mass balances by comparing them with observed mean mass balances from a geodetic observation product (Hugonnet et al., 2021), which represents mean elevation change in meters over five-year periods (i.e., 2000-2004, 2005-2009, and 2010-2014). We resampled the elevation change product from 100 m to the 25 m grid resolution of the hydrological model and masked it to the same spatial area as the hydrological model. The transformation from elevation change (m) into mm w.e. is done by multiplying the elevation change by a mean bulk density of 850 kg m⁻³ for firn and ice after Huss (2013). Because the geodetic elevation product incorporates several limitations and uncertainties for small areas (< 100 km²), a low sample number of glaciers, and short observation periods (<10 years), plus the mean error for the bulk density assumption of ± 60 kg m⁻³, we assumed a 30 % mean error for the observation. The observation yields mean glacier melt of -1071.5 mm w.e. over 2000-2004, -1218.1 mm w.e. over 2005-2009, and -1263.8 mm w.e. over 2010-2014, indicating a slight acceleration in the annual melt rates over the three consecutive periods (Fig. S14). Both WaSiM simulations overestimate the cumulated melt losses at the end of the simulation period in 2015 by 36 % in the case of temperature corrected and by 22 % in the case of the T and RH correction. Differences due to different glacier outlines between the observation and the model (Fig. 8c) can not be excluded, since the glacier mass balances are given as mean mm w.e. over the entire AoI. Consequently, an overestimation of the glacier cover by WaSiM would lead to an overestimation of glacier melt. Nevertheless, given the fact that the WaSiM model is not explicitly calibrated to glacier mass balances, the results strengthen the model's ability to reliably predict glacier dynamics.

S3 Supplementary Figures

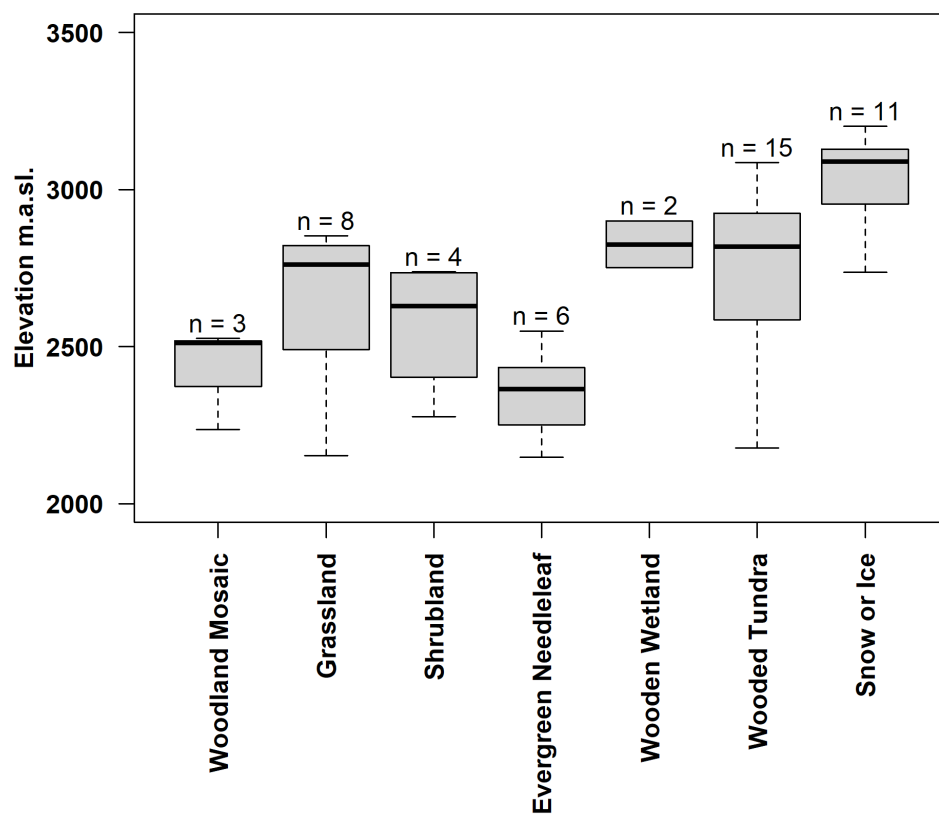


Figure S1. Boxplot of WRF land use categories and WRF elevation in m.a.s.l. of the upper Kauner Valley domain. The numbers above the boxes give the sample sizes (i.e., number of grid cells). The WRF grid cells have a spatial resolution of 2 km.

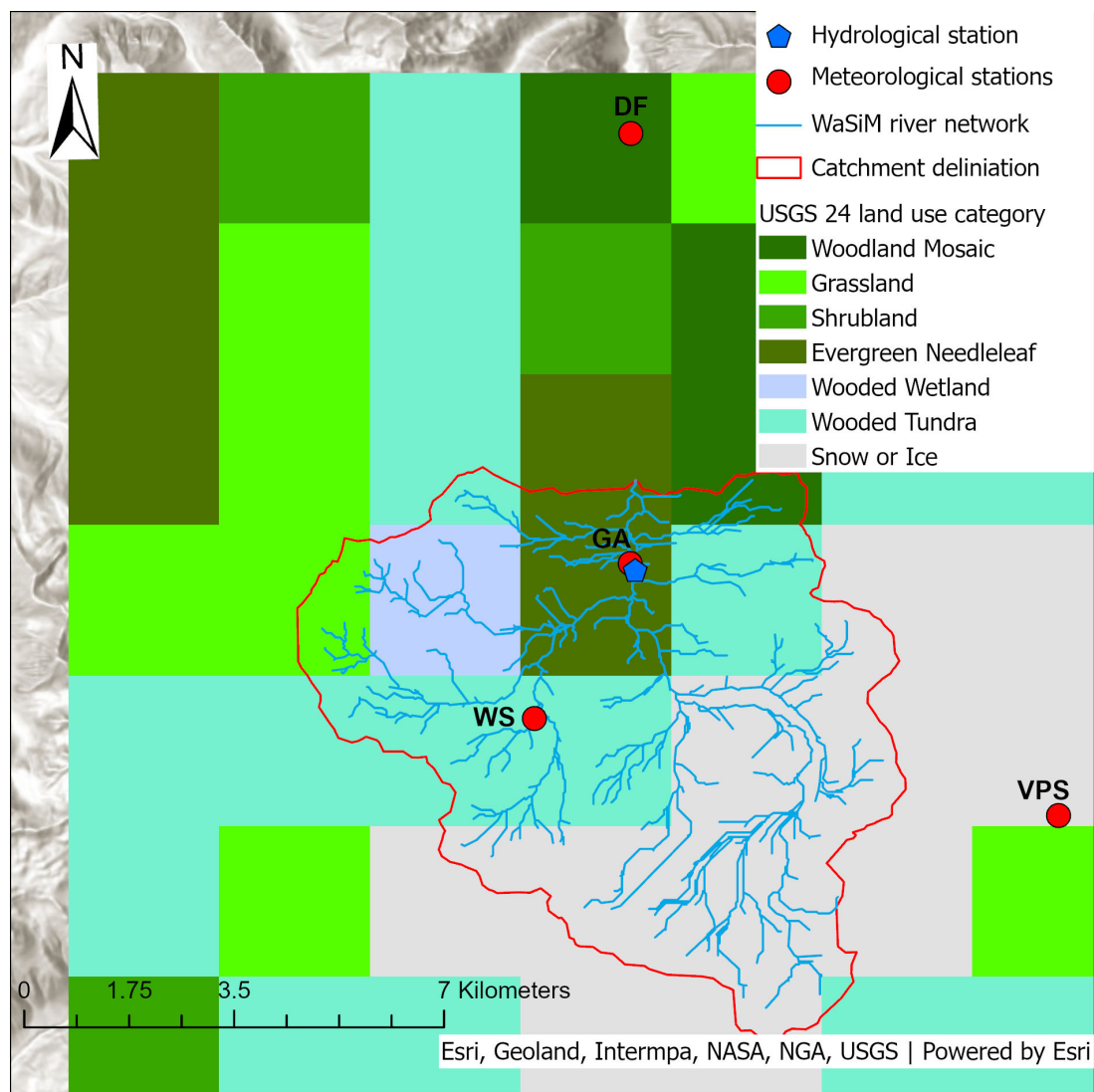


Figure S2. WRF land use categories in 2 km grid resolution and main features (river network and catchment delineation) of the upper Kauner Valley. Meteorological stations are indicated with red circles. Hydrological stations are represented by blue symbols.

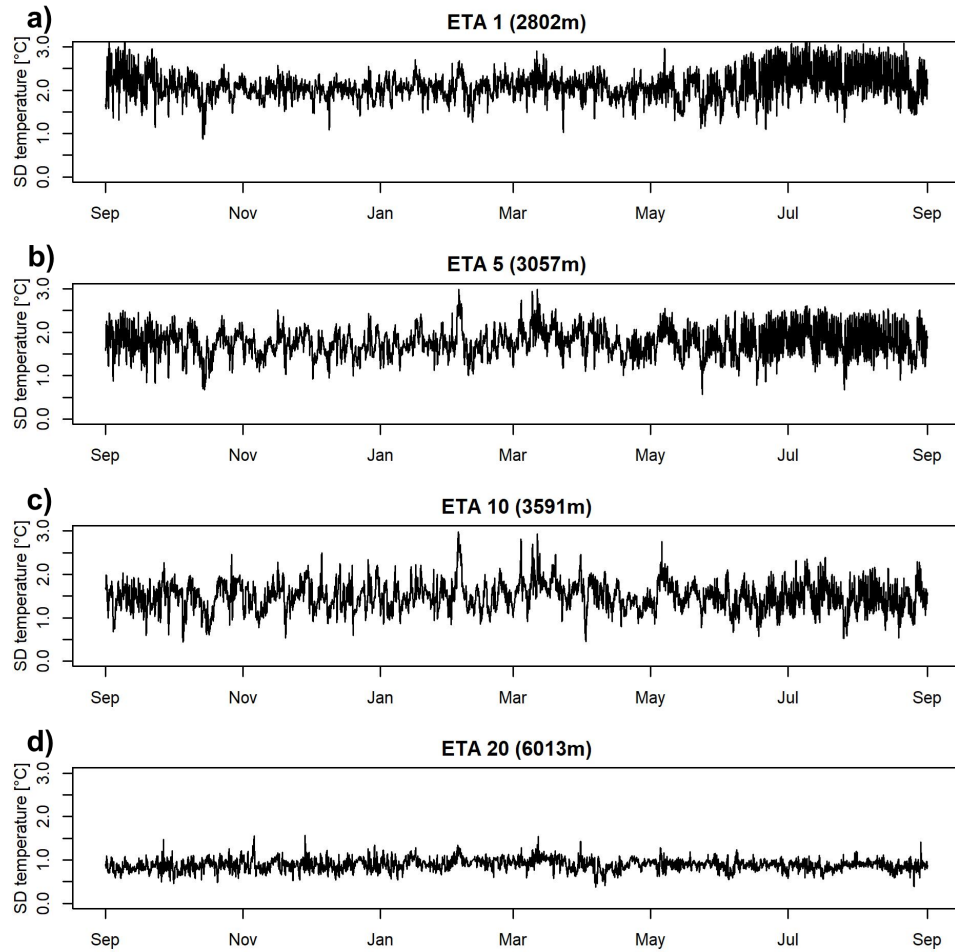


Figure S3. Hourly standard deviation of WRF temperature over the upper Kauner Valley for the period September 1974 to 1975 for eta levels (a) 1, (b) 5, (c) 10, and (d) 20. The altitude in the header gives the mean geopotential height across all WRF grid cells of the upper Kauner Valley domain in m a.s.l..

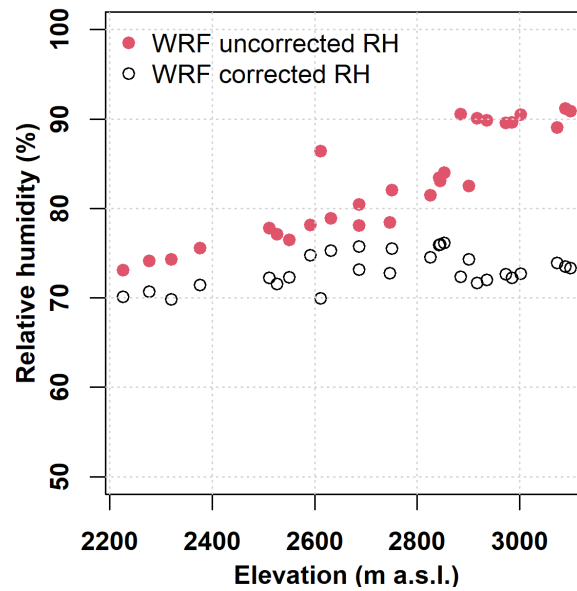


Figure S4. Scatterplot of WRF elevation (m a.s.l.) and mean relative humidity from uncorrected (red dots) and corrected (black circles) over the period 1970-2015.

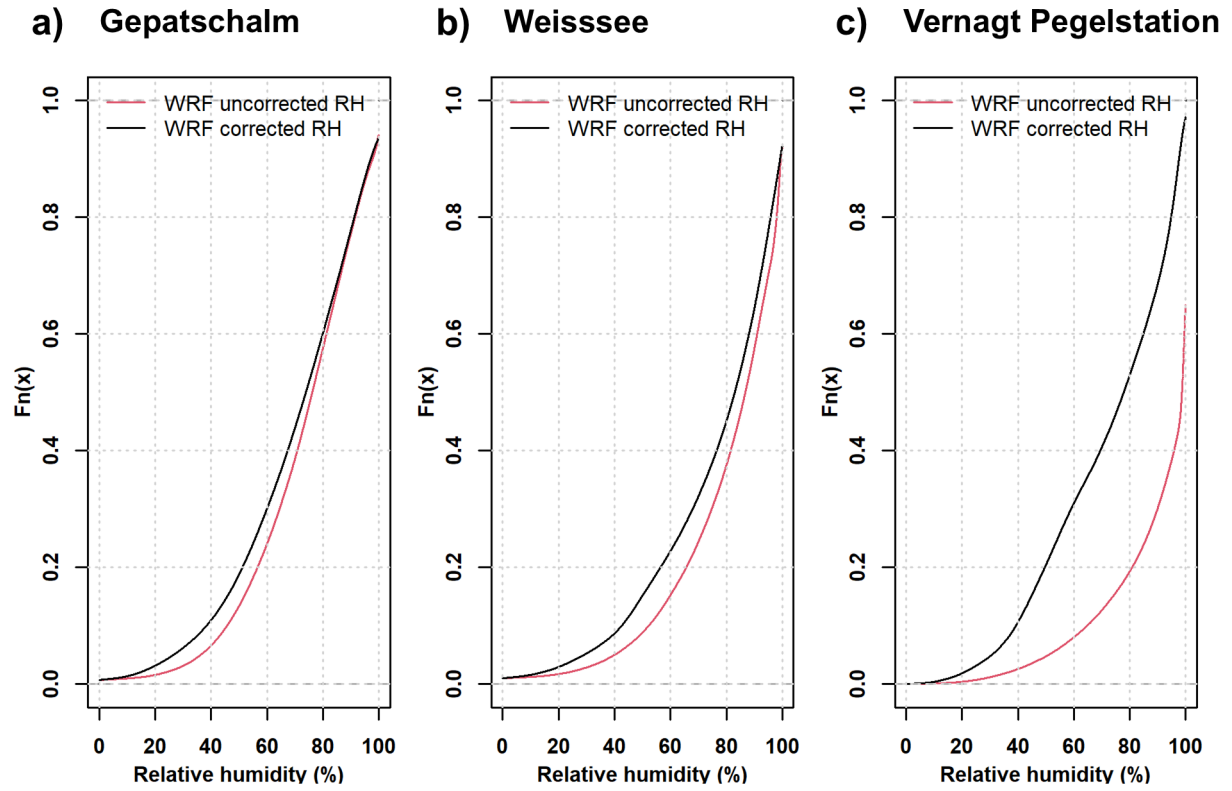


Figure S5. ECDFs of hourly uncorrected (red line) and corrected (black line) relative humidity (%) across the period 1970-2015 at different station locations: a) Gepatschalm, b) Weisssee, and c) Vernagt Pegelstation.

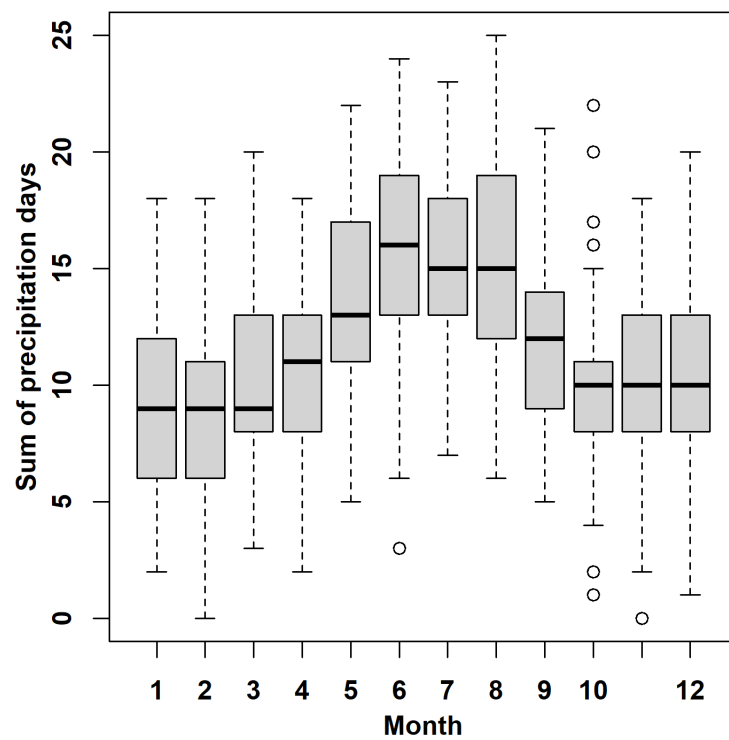


Figure S6. Boxplot of monthly precipitation days at the station Vergoetschen for the period 1971 to 2015.

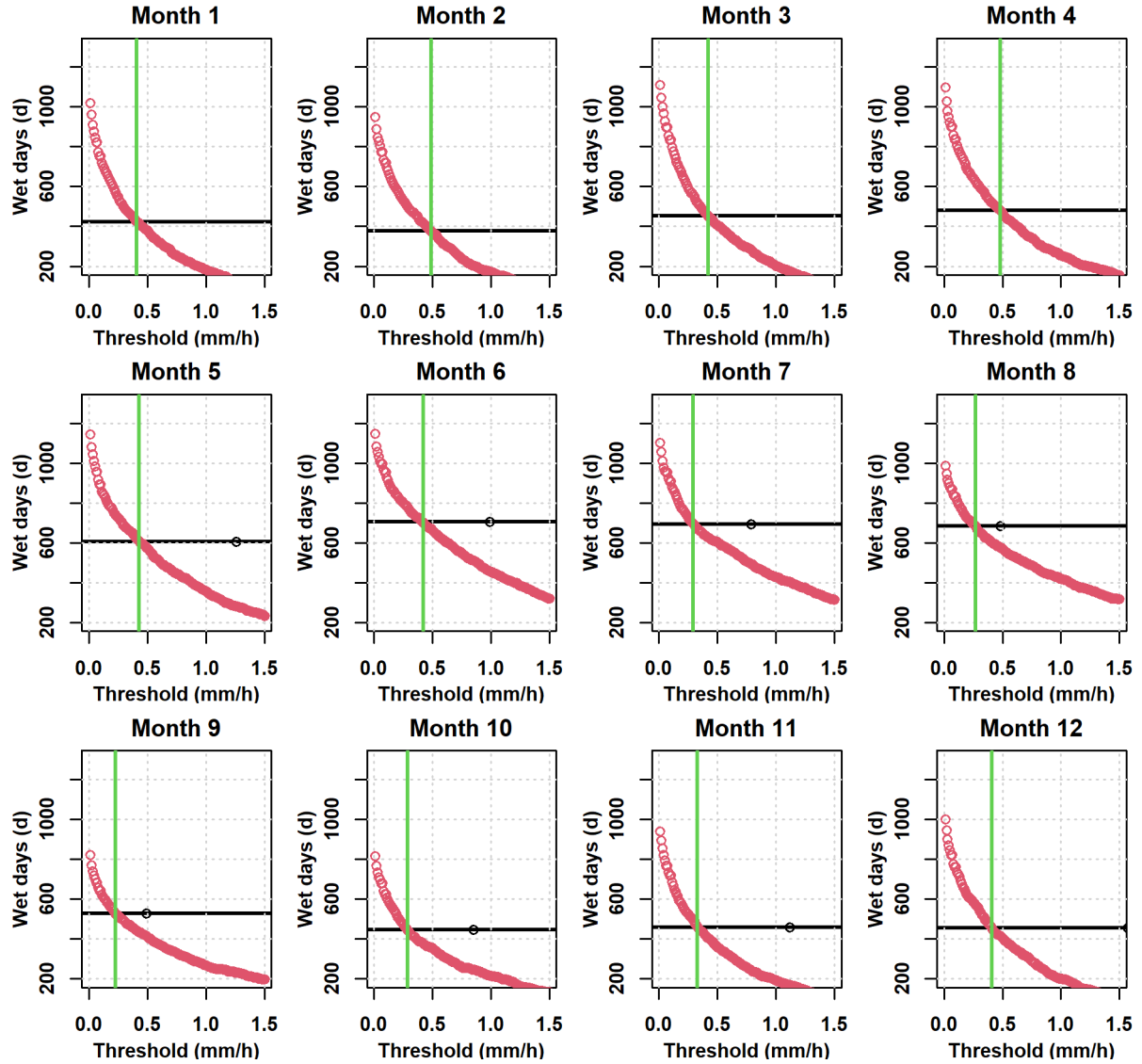


Figure S7. Monthly sum of wet days at station Vergoetschen (black horizontal line) and step-wise increase of the threshold value (mm h^{-1}) (red circles) to identify WRF time steps whose precipitation is to be set to zero for frequency matching. The point of absolute minimum error between observed and WRF precipitation days is indicated with the green vertical line.

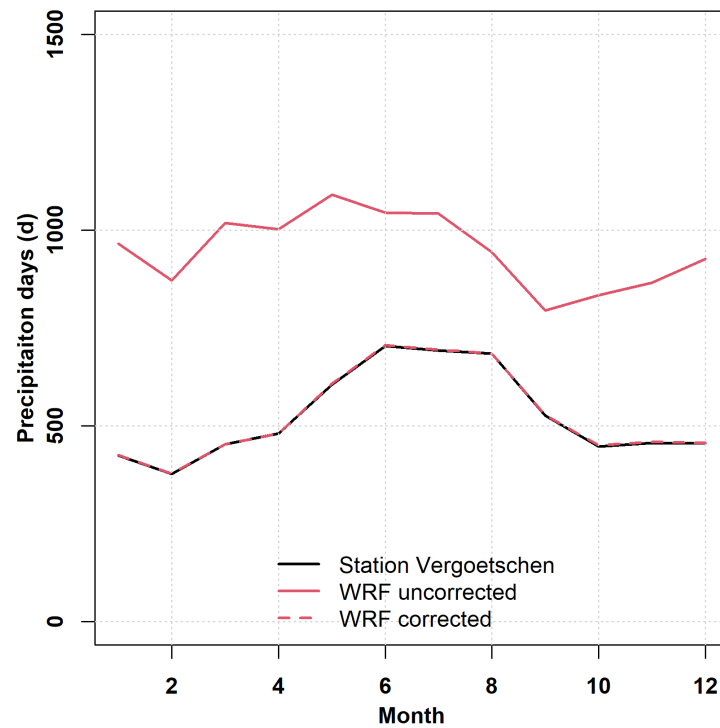


Figure S8. Monthly sum of wet days at station Vergoetschen (black line) and the corresponding WRF grid cell as uncorrected (red line) and after the frequency correction (red dashed line). The meteorological station Vergoetschen is located in the lower Kauner Valley at 1280 m a.s.l. and operated by HD-Tyrol.

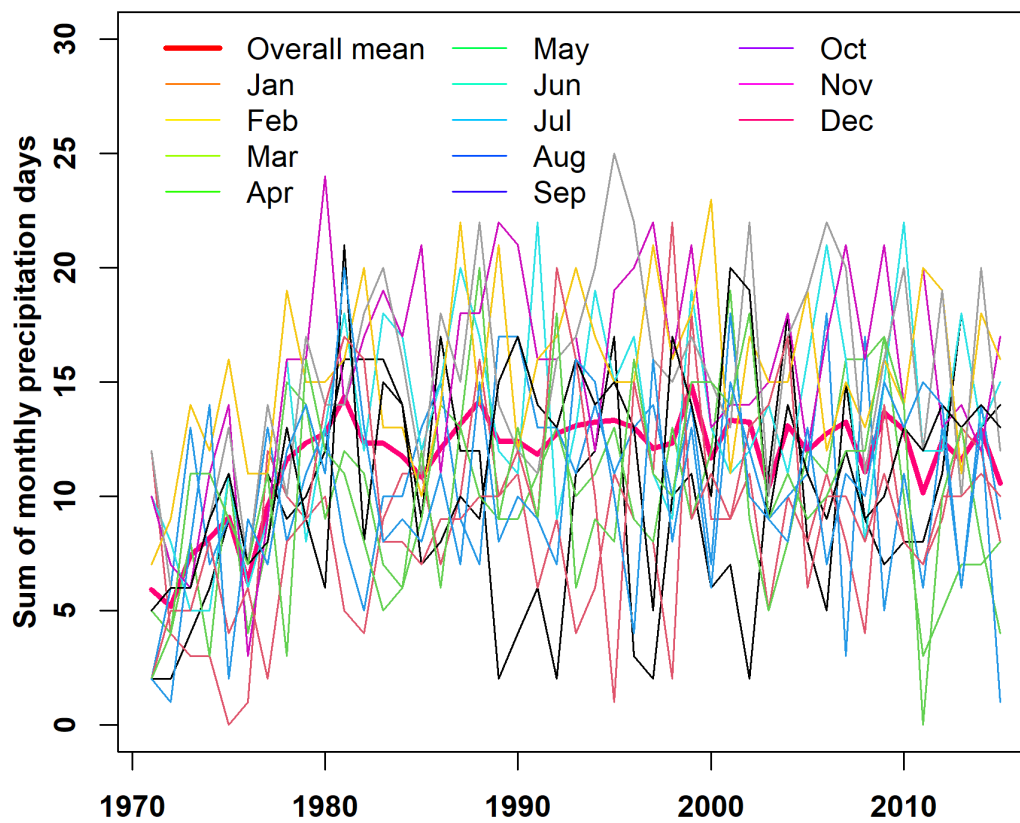


Figure S9. Spaghetti plot of monthly precipitation days at the station Vergoetschen for the individual years over the period 1971 to 2015. The red thick line represents the overall mean of all months.

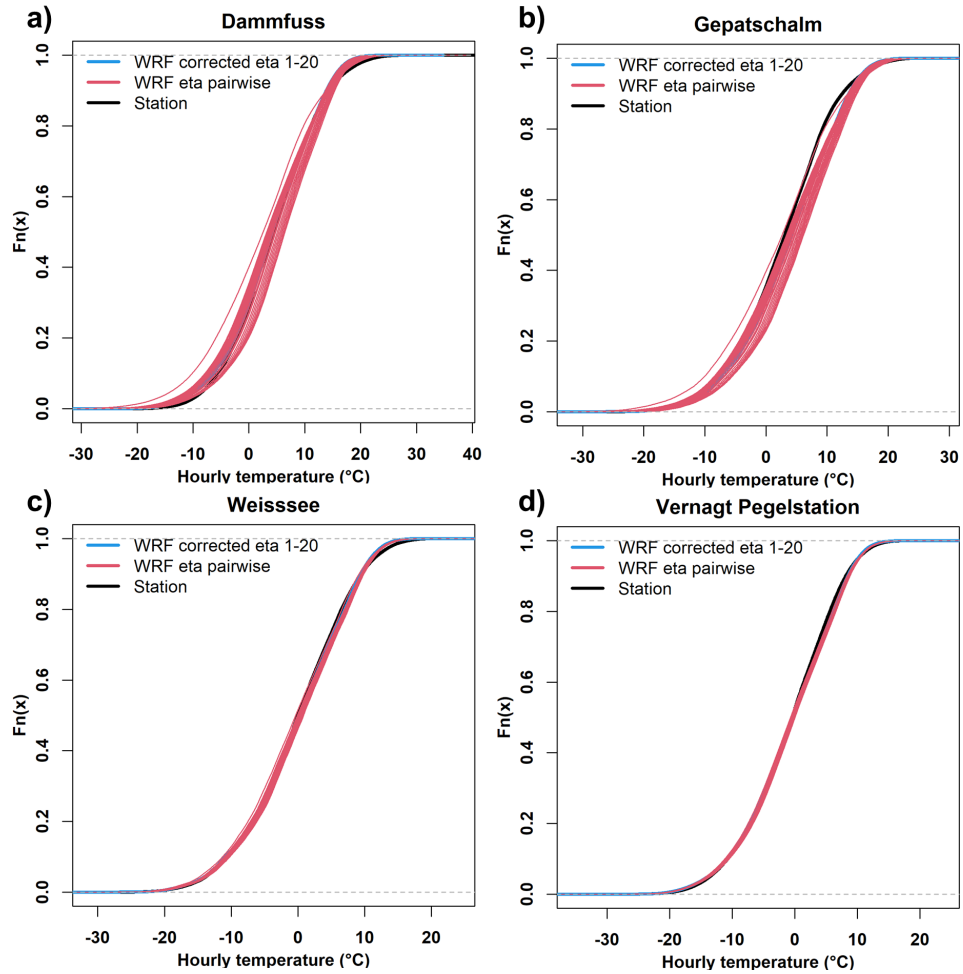


Figure S10. Empirical cumulated distribution functions (ECDF) of observed hourly temperature (black line), corrected WRF temperature (blue line) based on a linear gradient between eta levels 1 and 20, and corrected WRF temperatures (red line) based on adjacent eta level pairs at different stations a) Dammfuss, b) Gepatschalm, c) Weisssee, and d) Vernagt Pegelstation.

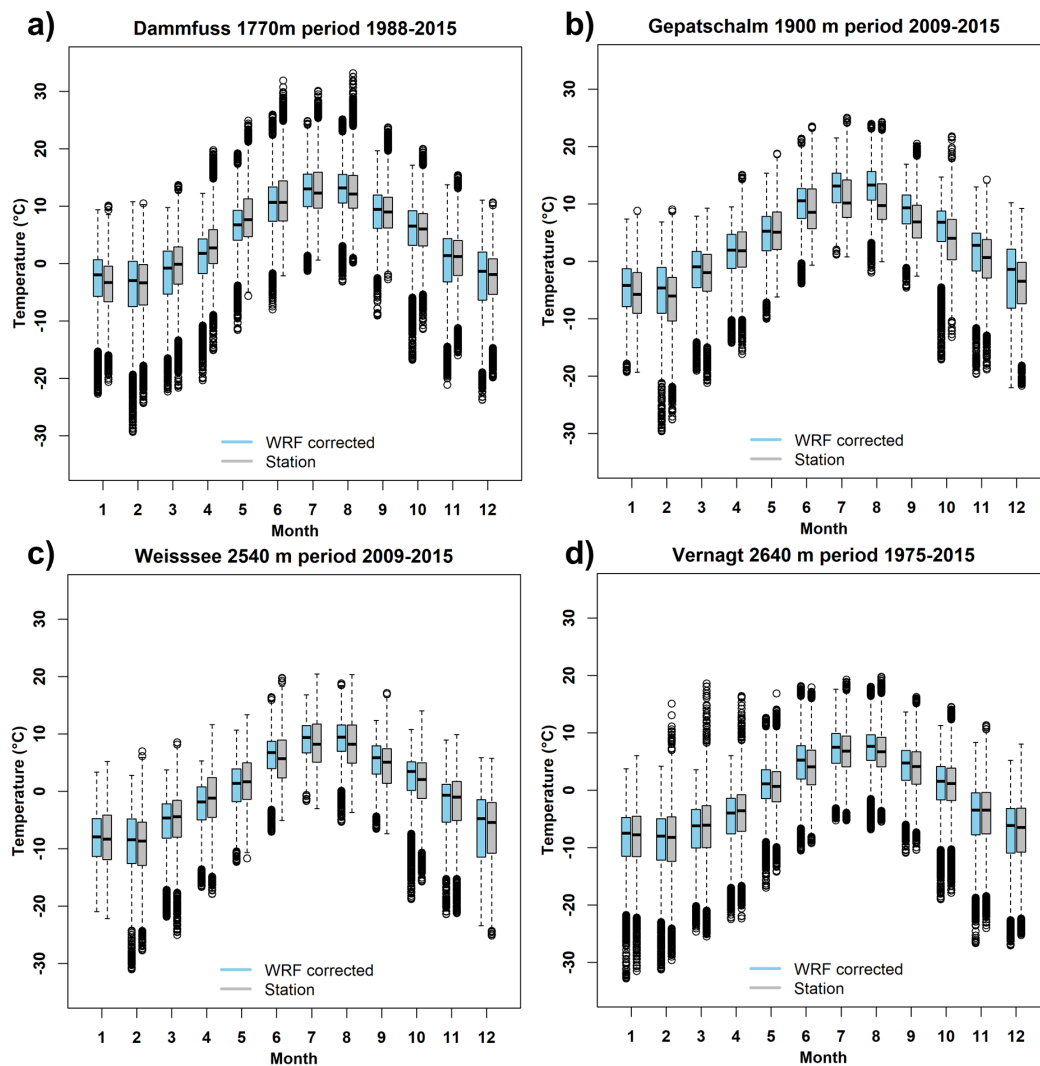


Figure S11. Monthly boxplots of hourly observed temperature (grey) and corrected WRF temperature (blue) at different stations a) Dammfuss, b) Gepatschalm, c) Weisssee, and d) Vernagt Pegelstation. The different temporal coverages are given in the titles of the respective subplots.

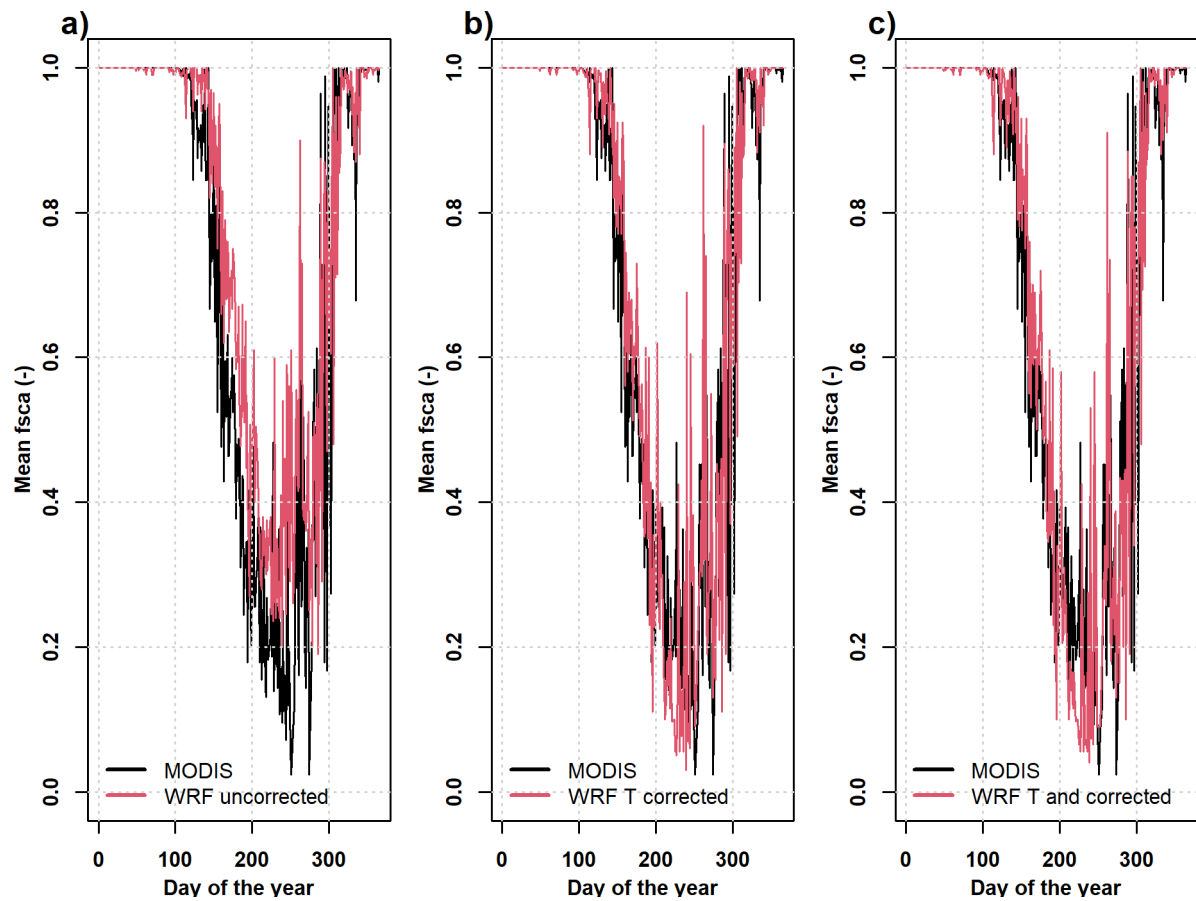


Figure S12. Day of the year mean fractional snow covered area (fsca) of MODIS (black line) and simulated fsca with a) uncorrected WRF forcing, b) corrected WRF temperature, and c) corrected WRF temperature and RH for the period 2007 to 2015.

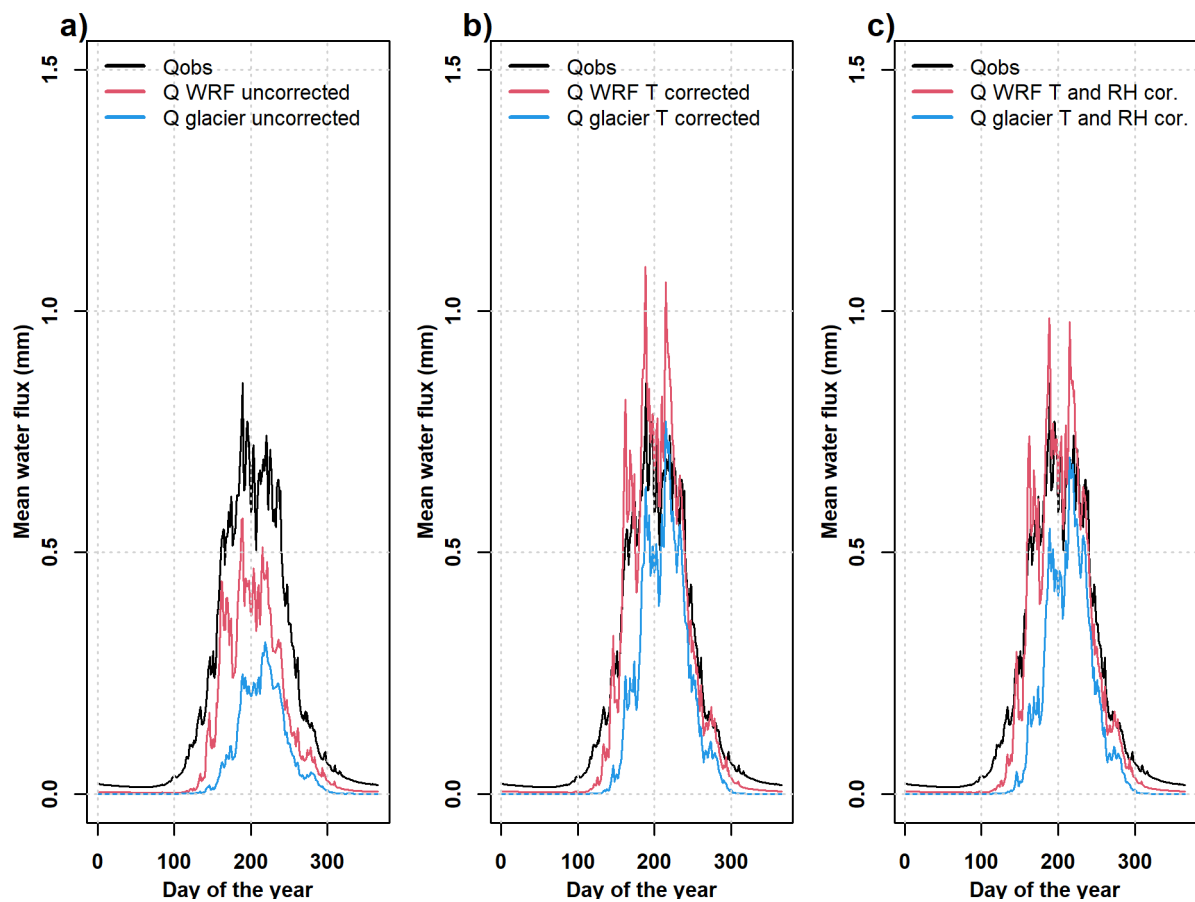


Figure S13. Observed (black line) and simulated (red line) mean water flux (mm/d) at streamgauge Gepatschalm and the simulated contribution from the glacier melt (blue line) with a) uncorrected, b) corrected WRF temperature, and c) corrected WRF temperature and RH for the period 2007 to 2015.

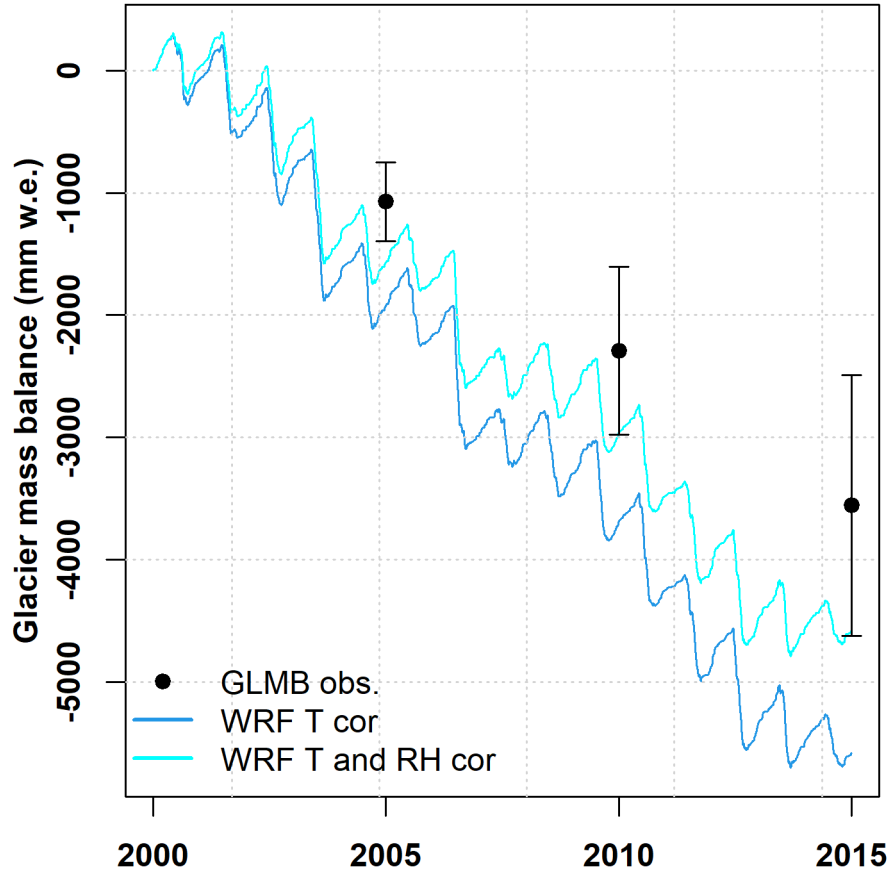


Figure S14. Observed and simulated mean glacier mass balances across the upper Kauner Valley over the period 2000-2015. The years are given as calendar years. The observed glacier mass balance is based on geodetic elevation changes from Hugonnet et al. (2021) and represents a mean over a five-year period (i.e., 2000-2004, 2005-2009, and 2010-2014). The transformation from elevation change (m) into mm w.e. is done by multiplying the elevation change by a mean bulk density of 850 kg m^{-3} after Huss (2013). The error bar for the observation assumes a 30 % mean error, incorporating the uncertainty of the geodetic elevation product and of the density assumption. The cumulated WaSiM simulations are showing the long simulation run (blue line) with T correction, and the simulation run (cyan line) with T and RH correction.

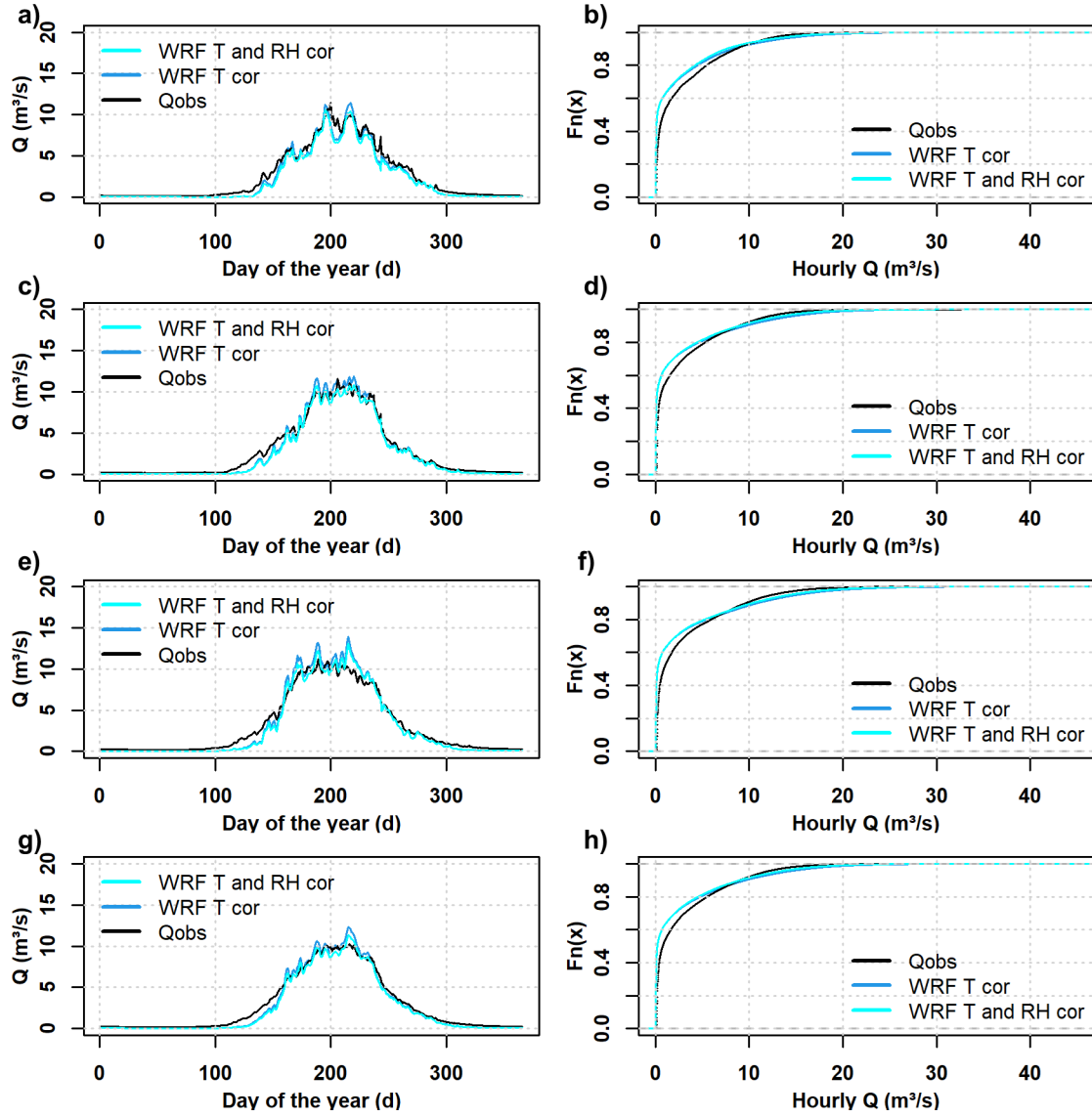


Figure S15. Left panels show mean day of the year of observed (black line) and simulated streamflow with corrected T (blue line) and corrected T and RH (cyan line). The right panels show hourly empirical cumulative distribution functions over four different periods, i.e., 1973-1987 (a, b), 1987-2001 (c, d), 2001-2015 (e, f), and as the overall mean (g, h).

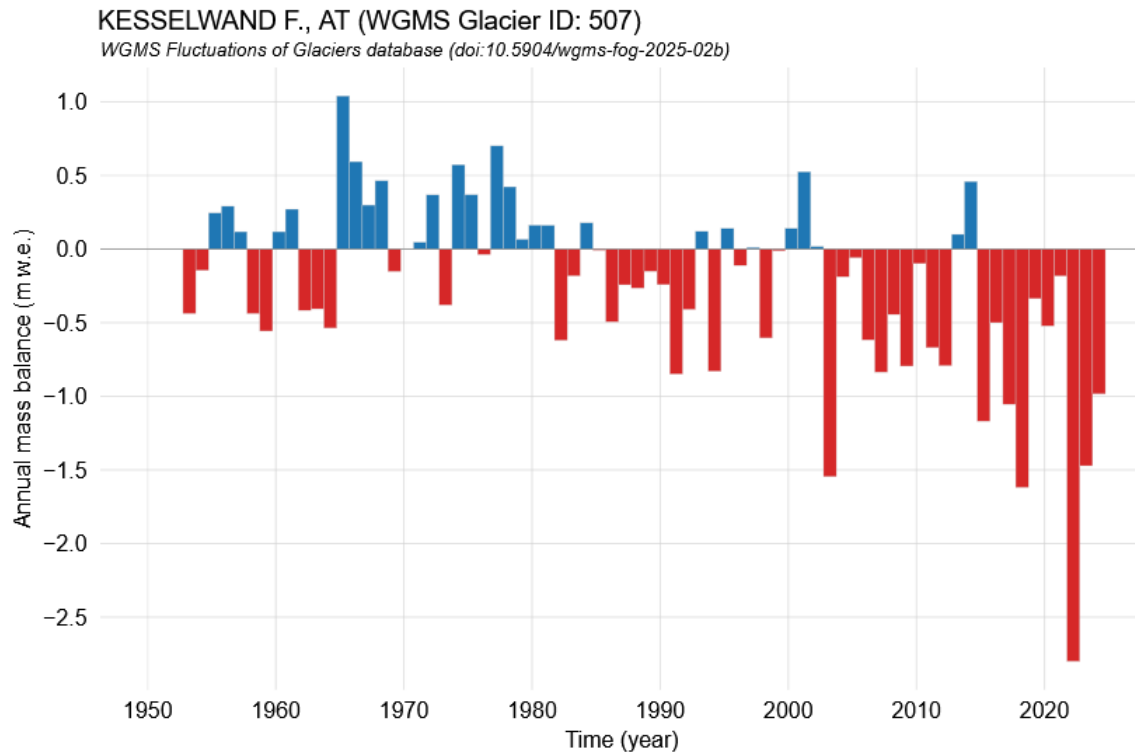


Figure S16. Annual observed glacier mass balances at the Kesselwand glacier in the Ötztal mountains (World Glacier Monitoring Service (WGMS), 2025). Red bars indicate years with negative mass balance, and blue bars indicate positive mass balance. The years are given as glaciological years. The Gepatsch and Kesselwand glaciers share parts of the accumulation area.

Table S1. Metadata of the meteorological observation stations and the respective WRF elevation and land use category (LUC) at the station location.

Station	East	North	Station elevation (m a.s.l.)	WRF elevation (m a.s.l.)	WRF LUC
Dammfuss	632260.0676	5202225.106	1770	2295	Woodland Mosaic
Gepatschalm	632251.1494	5195058.38	1900	2320	Evergreen Needleleaf
Weisssee	630648.4005	5192468.238	2540	2550	Wooded Tundra
Vernagt	639386.08	5190853.37	2640	2830	Snow or Ice

Table S2. Monthly mean and standard deviation of observed precipitation days at station Vergoetschen and derived threshold values for correcting the WRF precipitation frequency from 1971 to 2015.

Month	Mean prec days	SD prec days	Threshold (mm h ⁻¹)
1	9	4.83	0.41
2	8	3.81	0.49
3	10	4.06	0.42
4	11	3.93	0.48
5	13	4.37	0.43
6	16	4.52	0.42
7	15	3.6	0.29
8	15	4.58	0.27
9	12	3.96	0.23
10	10	4.43	0.29
11	10	4.17	0.33
12	10	4.32	0.41

Table S3. Goodness-of-fit criteria for WRF temperature computed based on different eta level combinations and the respective observed station temperature. Adjacent eta levels were used, e.g. from eta levels 1 and 2. Eta 1 to 20 refers to the corrected WRF temperature computed over the eta level range 1 to 20. RMSE stands for Root Mean Square Error, MAE for Mean Absolute Error, and KGE for Kling–Gupta Efficiency.

ETA	Dammfuss			Gepatschalm			Weissee			Vernagt		
	RMSE (°C)	MAE (°C)	KGE (–)	RMSE (°C)	MAE (°C)	KGE (–)	RMSE (°C)	MAE (°C)	KGE (–)	RMSE (°C)	MAE (°C)	KGE (–)
24	4.19	3.37	0.64	4.82	3.85	0.00	3.11	2.41	–1.36	3.08	2.35	0.44
23	4.08	3.27	0.67	4.71	3.76	0.05	3.10	2.40	–1.20	3.06	2.35	0.47
22	4.00	3.19	0.71	4.62	3.67	0.11	3.09	2.39	–1.01	3.06	2.34	0.50
21	3.93	3.12	0.75	4.53	3.59	0.17	3.08	2.38	–0.81	3.06	2.34	0.54
20	3.88	3.03	0.79	4.45	3.52	0.24	3.07	2.37	–0.59	3.05	2.34	0.58
19	3.86	3.05	0.85	4.39	3.47	0.31	3.07	2.37	–0.34	3.05	2.34	0.62
18	3.87	3.06	0.85	4.34	3.43	0.39	3.07	2.37	–0.10	3.05	2.34	0.67
17	3.93	3.11	0.87	4.31	3.42	0.48	3.07	2.38	0.20	3.06	2.34	0.72
16	3.97	3.14	0.86	4.31	3.42	0.51	3.07	2.38	0.33	3.06	2.34	0.74
15	4.04	3.19	0.84	4.31	3.42	0.58	3.09	2.39	0.55	3.06	2.34	0.78
14	4.09	3.22	0.81	4.31	3.42	0.64	3.10	2.39	0.75	3.07	2.35	0.82
13	4.12	3.24	0.79	4.30	3.42	0.68	3.10	2.40	0.83	3.07	2.35	0.84
12	4.13	3.23	0.77	4.29	3.41	0.71	3.10	2.39	0.92	3.07	2.35	0.85
11	4.11	3.23	0.76	4.26	3.38	0.72	3.09	2.39	0.90	3.06	2.35	0.86
10	4.07	3.19	0.73	4.22	3.35	0.73	3.08	2.38	0.88	3.06	2.34	0.86
9	4.03	3.15	0.76	4.18	3.32	0.74	3.07	2.37	0.88	3.05	2.34	0.87
8	3.99	3.12	0.75	4.15	3.29	0.74	3.06	2.36	0.86	3.05	2.33	0.87
7	3.96	3.09	0.75	4.12	3.27	0.74	3.06	2.35	0.86	3.04	2.32	0.87
6	3.93	3.06	0.75	4.10	3.25	0.75	3.05	2.34	0.83	3.03	2.32	0.87
5	3.91	3.03	0.74	4.09	3.23	0.76	3.04	2.34	0.79	3.02	2.31	0.86
4	3.89	3.01	0.73	4.06	3.19	0.77	3.03	2.33	0.74	3.01	2.30	0.88
3	3.90	3.01	0.70	3.99	3.12	0.81	3.00	2.30	0.58	2.99	2.28	0.90
2	3.97	3.05	0.70	3.98	3.10	0.82	2.98	2.28	0.50	2.97	2.26	0.90
1	5.34	4.16	0.40	4.53	3.49	0.64	2.98	2.29	–0.74	2.93	2.23	0.79
1 to 20	3.80	3.00	0.84	4.09	3.25	0.63	3.05	2.35	0.92	3.04	2.32	0.88

Table S4. Statistics of the Kolmogorov–Smirnov (KS) test between observed, uncorrected, and corrected WRF temperature at four stations. The null hypothesis of the KS test is that the two samples have the same distribution. The D-statistic gives the maximum absolute as well as the maximum vertical difference between the ECDFs.

Site	T uncor same distribution	T uncor D-statistic (–)	T uncor max vertical difference (°C)	T cor same distribution	T cor D-statistic (–)	T cor max vertical difference (°C)
Dammfuss	Yes	0.08	–3.06	Yes	0.03	17.46
Gepatschalm	Yes	0.08	9.66	Yes	0.07	9.63
Weisssee	Yes	0.05	–3.8	Yes	0.03	6.45
Vernagt	Yes	0.13	6	Yes	0.03	5.6

Table S5. Water balance components from simulation results without (uncor), with WRF temperature correction (T cor), and with WRF temperature and RH correction (RH cor) in mm. Obs stands for observation, ET for evapotranspiration, and GLMB for glacier mass balance.

Component	2008	2009	2010	2011	2012	2013	2014	2015	Mean
Prec RH cor	1589	1398.6	1353.86	1166.07	1466.79	1813.02	1488.37	1624.12	1487.48
Prec T cor	1589	1398.6	1353.86	1166.07	1466.79	1813.02	1488.37	1624.12	1487.48
Prec uncor	1589	1398.6	1353.86	1166.07	1466.79	1813.02	1488.37	1624.12	1487.48
Snow RH cor	998.8	966.09	816.99	585.3	962.9	1204.76	909.02	983.18	928.38
Snow T cor	998.8	966.09	816.99	585.3	962.9	1204.76	909.02	983.18	928.38
Snow uncor	1088.52	1022.72	884.04	676.34	1024.41	1291.66	989.67	1076.11	1006.68
Rain RH cor	493.31	333.42	457.65	518.93	403.36	485.78	487.81	542.94	465.4
Rain T cor	493.31	333.42	457.65	518.93	403.36	485.78	487.81	542.94	465.4
Rain uncor	386.6	265.94	377.95	412.4	329.99	384.22	392.98	434.61	373.09
Q obs	1798.11	1769.95	1613.95	2226.59	1984.5	1627.43	1598.45	1754.66	1796.71
Q RH cor	1674.06	1526.4	1759.65	1579.49	1784.67	1670.38	1277.48	2134.78	1675.86
Q T cor	1807.1	1659.78	1856.06	1657	1904.33	1777.37	1407.46	2230.33	1787.43
Q uncor	900.64	876.64	1009.37	879.13	1053.32	959.91	795.66	1294.54	971.15
ET RH cor	80.08	97.77	70.12	119.25	93.48	60.93	66.75	85.87	84.28
ET T cor	65.24	83.23	55.65	102.81	75.81	43.16	44.25	71.26	67.68
ET uncor	54.82	72.62	54.04	84.47	74.73	39.16	44.69	68.42	61.62
GLMB RH cor	-282.02	-301.3	-514.83	-577.41	-530.19	-77.33	69.83	-665.52	-359.85
GLMB T cor	-396	-419.05	-594.76	-638.63	-628.03	-161.18	-54.28	-737.19	-453.64
GLMB uncor	389.17	449.13	269.4	153.11	213.83	665.22	590.35	199.17	366.17

Table S6. Statistics on temperature evolution with respect to elevation, p-value of modified Mann-Kendall test, Sen's slope, and mean annual temperature trend from 1973 to 2015.

Elevation (m a.s.l.)	mMK p-value	Sen's slope ($^{\circ}\text{C a}^{-1}$)	Trend ($^{\circ}\text{C a}^{-1}$)
1770	0.0010	0.0031	0.0356
1900	0.0008	0.0031	0.0361
2540	0.0004	0.0032	0.0384
3800	0.0001	0.0035	0.0429

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