



Research papers

Annual water balance and hydrological trends in the glacierised Tarfala Catchment, Sweden

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ABSTRACT

Quantifying components of the hydrological cycle in glacierised catchments is important for the assessment of the temporal distribution, quantity and quality of water resources available to downstream regions, especially under a changing climate. However, this assessment requires long time series of observations, which are typically unavailable for remote catchments, such as those in mountainous areas. In this study, we leverage a unique ~40 year time series of hydrological data recorded in the subarctic glacierised Tarfala catchment (Sweden) to explore temporal trends in the components of the catchment water balance (precipitation, runoff, change in storage, and evaporation), and to assess if water balance residuals are associated with specific hydro-climatic conditions. No significant temporal trends were found in precipitation and storage changes of the glacierised area, but significant increases were found in evaporation and summer discharge (in part attributed to glacier volume losses). The annual water balance could not be perfectly closed, and water losses were on average 112 mm y^{-1} larger than the water inputs over the study period. Among the water balance components, discharge contributed most to the total water balance uncertainty, and storage surplus due to antecedent meteorological conditions could explain why water losses in specific years exceeded the uncertainty bounds. It is therefore essential to consider legacy effects from previous years when applying water balance calculations in mountainous and/or glacierised catchments.

1. Introduction

Glacierised catchments play a key role in determining the hydrology of downstream regions that often rely on glacial melt water for human activities and ecological processes (Messerli et al., 2004; Milner et al., 2017; Viviroli et al., 2011). The hydrological regime of glacierised catchments at the middle and high latitudes is characterised by a discharge peak typically more concentrated into the summer months than in non-glacierised catchments due to the glacier melt contribution (e.g., Fountain and Tangborn, 1985). The timing and intensity of the summer streamflow peak and therefore the hydrological regime of the catchment, depend on how glaciers respond to both meteorological and climate variability (e.g., Dahlke et al., 2012). Under warming, the total volume and timing of glacier runoff are expected to be affected due to increased glacier melt (Fountain and Tangborn, 1985). In the short term,

the melt season is expected to lengthen as, due to earlier snowmelt, the ice surface is exposed for a longer period. In the long term, this effect translates into increasing ice mass loss and ice melt runoff contribution to catchment streamflow, until a maximum ('peak water') is reached, after which runoff decreases as glaciers reach an equilibrium with the climatic forcing or disappear (e.g., Huss and Hock, 2018; Jansson et al., 2003).

Both effects (increasing and decreasing trends in annual runoff) have been reported in the Alps and in Canada, whilst the threshold appears not to have been crossed in central Asia or Peru, where only increasing runoff trends are seen (Casassa et al., 2009). In the Alps, catchments with high glacier coverage displayed an increase in the annual runoff associated with glacier mass loss, due to an earlier start of the snow melt season and increasing temperature (Bard et al., 2015; Kormann et al., 2016; Pellicciotti et al., 2010) and lower winter accumulation (Lane and

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Nienow, 2019). In Canada, increasing trends in annual runoff have been observed in glacierised catchments in south-western Yukon, compared with decreasing trends in non-glacierised catchments in the same hydroclimatological setting (Fleming and Clarke, 2003). A study of the effect of glacier retreat on catchment runoff in that region found that runoff changes could be attributed to the observed glacier coverage changes in many, but not all, cases and indicated glacier retreat rate as a key factor for decadal changes in runoff (Chesnokova et al., 2020). Decreasing trends in summer runoff associated with glacier mass and area reductions have been observed in British Columbia, leading to the conclusion that the 'peak water' has been exceeded in this area (Stahl and Moore, 2006). Studies at the global scale have shown that the 'peak water' in large-scale catchments with small glaciers, such as in western Canada or central Europe, has either already been exceeded or is expected to be reached in the next decade, confirming the findings of similar studies at a regional scale (Huss and Hock, 2018). Inferring the effect of glacier retreat on catchment runoff from summer observations remains challenging because factors such as the extent and timing of seasonal snow cover disappearance or the occurrence of precipitation events in late summer influenced by climate variability can mask the contribution of ice melt runoff (Dahlke et al., 2012). These factors are particularly uncertain in high-latitude catchments, which have been studied less than those in the Alps (e.g., Bard et al., 2015; Huss, 2011; Kormann et al., 2016; Pellicciotti et al., 2010) and North America (e.g., Chesnokova et al., 2020; Fleming and Clarke, 2003; Stahl and Moore, 2006). Due to this lack of long-term studies in high-latitude glacierised catchments it remains unclear how variations in precipitation, evapotranspiration and altered glacier mass affect runoff changes in time.

An effective approach to understand catchment hydrological processes is that of quantifying water balance components and then, based on that, detecting changes in the catchment response over time (Dingman, 2015). Not only does the calculation of water balances allow changes in runoff to be attributed to specific processes, it also provides a framework within which to assess the quality of the hydrologic data. An imbalance between the amount of water entering and leaving the catchment (as diagnosed by a water balance that is not closed) can be related not only to observed changes in storage, but also to both measurement errors and processes that were unaccounted for or assumed to be negligible (e.g., storage outside glacier area). Quantifying water balance components and their uncertainty requires long time series of observations that are rarely available in remote areas. To our knowledge, possibly because of this data limitation, there are few studies on water balance (Kane and Yang, 2004). In contrast, several studies quantified the water balance components using modelling experiments to overcome data availability limitations (e.g., Chesnokova et al., 2020; Engelhardt et al., 2014; Finger et al., 2011; Huss, 2011). Modelling estimates, however, are also affected by the uncertainty of observational data used, for model calibration and validation (McMillan et al., 2012) and by process representation and parameter uncertainty (e.g., Fenicia and Kavetski, 2021). These challenges demonstrate the knowledge gap in investigating uncertainties in water balance components and closure in general and especially in high-latitude glacierised catchments. This raises the question of how water balance residuals of catchments in these settings vary in relation to the other components or more in general in relation to climatic conditions, once the uncertainty in hydrologic inputs and outputs is accounted for.

The long time series of hydrological data recorded at the subarctic Tarfala Research Station, northern Sweden, makes it possible to quantify the different components (precipitation, runoff, change in storage, and evaporation) of the annual water balances of the glacierised Tarfala catchment over ~40 years (1969–2008). Distinguishing these different components provides a more complete picture of catchment processes compared to previous work in the same area focusing on precipitation and discharge alone (Dahlke et al., 2012). Storglaciären, the largest glacier in this catchment, is predicted to lose between 50 and 90% of its volume by 2100 (Radić and Hock, 2006), similar to other nearby glaciers

which also display ongoing negative mass balance trends (Taveirne et al., 2021). Therefore, given the sensitivity of the hydrological regime to changes in glacierisation levels, the hydrological behaviour of the Tarfala catchment is expected to undergo significant changes in the coming years. To assess the ongoing effects of climate change, it is important to conduct a detailed evaluation of the hydrological conditions of the past approximately 40 years, during which global warming effects started to become considerable in the Arctic (Bring et al., 2016) and also became apparent in retreating nearby glaciers (Taveirne et al., 2021).

The Tarfala catchment is therefore an ideal case study to address two related questions relevant for high-latitude glacierised catchments in general: (i) how are the components of the hydrologic balance changing through time and, in particular, are changes in runoff related to glacier volume loss and/or altered precipitation? and (ii) how well can the catchment annual water balance be closed and do the residuals vary systematically as hydro-climatic conditions change through time? Accordingly, our aims are to:

- (1) quantify changes in runoff and other hydrologic components (precipitation, evaporation, storage changes) in the Tarfala catchment, considering the entire multi-decadal study period, as well as at decadal time scales.
- (2) assess how the uncertainty of the water balance components affects the magnitude of unexplained residuals and investigate if residuals and their trends are related to any of the components of the hydrologic balance, which would suggest that hereto unquantified hydrological processes might be occurring.

2. Study area

The Tarfala catchment (67.911379 N, 18.611312 E, WGS84), northern Sweden, is located on the eastern side of the water divide of the Scandinavian mountains (Fig. 1). The catchment covers an area of approximately 20.1 km² and consists of a valley surrounded by steep slopes, with an elevation range of between 980 and 2097 m a.s.l. of the Kebnekaise peak, the highest point in Sweden. The valley is located within the discontinuous permafrost zone (Christiansen et al., 2010) but permafrost depths of several hundred meters are found at higher elevations in the terrain (Isaksen et al., 2001). The catchment mostly comprises sediments of weathered bedrock and boulders (Andersson et al., 1985; Rafstedt, 1985) and the bedrock on the valley floor and some of the mountain slopes are covered by diamicton. As the catchment is above the treeline (780 m a.s.l. in the area; Wilson and Nilsson, 2009), the vegetation is limited to grass heath on small patches of organic soil in a narrow zone along the valley floor.

The catchment contains six glaciers (Fig. 1), with a total glacierised area of 6.18 km², representing approximately 30% of the study site. Storglaciären is the largest (3 km²) and most extensively studied glacier in the catchment and has the longest mass balance record globally, from 1946 to the present day (Holmlund et al., 2005). Storglaciären is a polythermal glacier with a cold surface layer in its ablation area that has diminished in size in response to climate warming (Gusmeroli et al., 2010; Holmlund and Eriksson, 1989; Pettersson et al., 2003). A stream drains meltwater to the main catchment stream, Tarfalajåkk. Three percent of the catchment area consists of lakes. The largest, Tarfalasjön, is contained within a glacial overdeepened basin that is 52 m deep (Schytt, 1973) and stores ~15 × 10⁶ m³ of water. Other lakes are dammed by moraine ridges and store ~10⁶ m³ of water.

The Tarfala catchment is situated in a transitional zone between a continental climate to the east and a more maritime climate to the west (Holmlund and Schneider, 1997). Precipitation patterns are controlled by orographic mechanisms typical of a high mountain environment, with air masses coming mainly from the north-west, west and south-west (Jansson et al., 2007). In winter, some precipitation is also brought to the eastern side of the valley by south-easterly winds. Strong

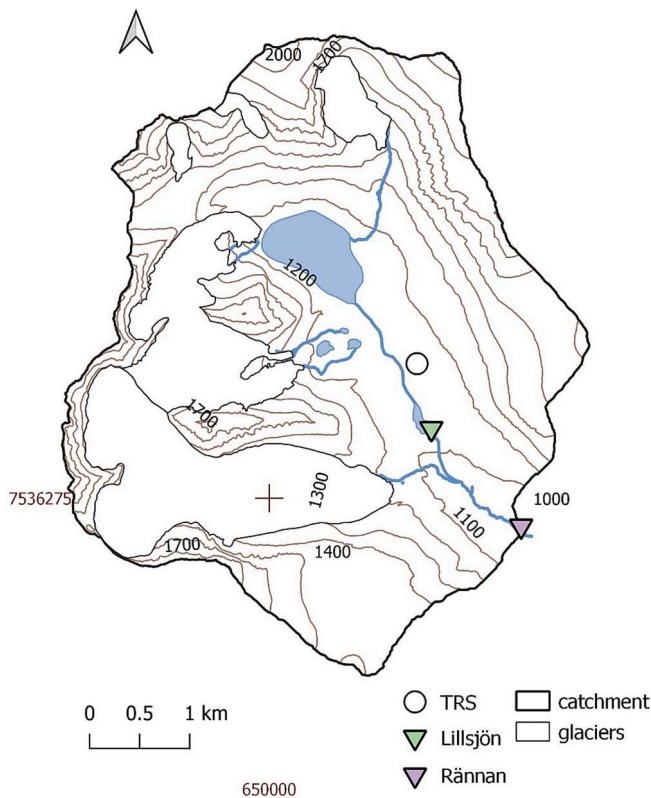


Fig. 1. Map of the Tarfala catchment with the locations of the Tarfala Research Station (TRS) and the Lillsjön and Rännan streamflow gauge stations (northern Sweden). The cross indicates the coordinates in the Swedish reference system (SWEREF 99). The Rännan station is used in the water balance and the Lillsjön for data infilling.

winds along the valley also play a role in determining the winter snow distribution. The observed mean annual air temperature is -3.4 ± 1.0 °C (1965–2014, Tarfala Research Station data) and has been warming linearly at a rate of 0.042 °C yr^{-1} , with the largest increases found in January and October (Jonsell et al., 2013). The summer season is short, with temperatures typically reaching above 0 °C between late May and early September. The hydrological regime is typical of glacierised, high mountain catchments. The beginning of the snowmelt occurs on 11th June on average (1969–2011) (Dahlke and Lyon, 2013). The timing of the relative contributions to runoff of snowmelt, ice melt, and precipitation between July and August depends on snow cover properties and glacier dynamics (Dahlke et al., 2014). As air temperature drops in September, melt ceases and precipitation begins to fall consistently as snow. In this study we used precipitation, air temperature, discharge and glacier mass balance observations available at the Tarfala catchment over the period 1969–2008. After 2008 missing data and the change of the catchment outlet following the destruction of the Rännan gauging station made catchment characteristics change, so that the time series extension to more recent years was not possible.

3. Methods

3.1. Data

3.1.1. Precipitation

Precipitation at Tarfala Research Station (TRS; 1135 m a.s.l.) was initially measured by a standard precipitation gauge from the Swedish Meteorological and Hydrological Institute (SMHI). With the installation of the automatic weather station, precipitation measurements were provided by an unheated tipping bucket system. In this study, precipitation measurements from the standard gauge are used for the period

1969–1988 and from the tipping bucket for the period 1989–2009. Precipitation measurements during winter are considered not reliable as the air temperature is below 0 °C for most of the time and strong winds often occur. Continuous precipitation measurements from the standard gauge and tipping bucket are therefore restricted to rainfall events (using a daily mean air temperature of 0 °C threshold to exclude precipitation that falls as snow) between June and September which define the summer period in this study. The spatial variability of summer precipitation in the catchment is accounted for by estimating the areal precipitation. This is obtained from a Thiessen polygon interpolation of measurements from the period 1965 to 1984 (Quinlan, 2016), when a distributed network of gauges was established for the International Hydrological Decade (Bergman, 1988). A linear regression is then established between the precipitation measurements at Tarfala Research Station and the areal precipitation estimations (Fig. S1), which allows extrapolation of areal summer precipitation (P_a , measured in mm during the summer) estimations to the whole study period (Quinlan, 2016):

$$P_a = 1.2P_{\text{TRS}} - 12.3, \quad (1)$$

where P_{TRS} represents precipitation at Tarfala Research Station in mm during the summer (coefficient of determination, $R^2 = 0.94$).

Winter precipitation is estimated from the total amount of snow accumulated in the catchment area as volumetric water equivalent. This is obtained from manual measurements of snow depth and density carried out in the catchment between the end of May and the beginning of June for the period 1968–1974 (Bronge, 1985; Nilsson and Larsson, 1976). To extrapolate the volumetric water equivalent calculated from these snow surveys to the whole study period, we established a relation between volumetric water equivalent and the winter mass balance (C_w , measured in 10^6 m^3) of Storglaciären (Fig. S2). The resulting linear regression predicts winter precipitation (P_w , measured in 10^6 m^3) as:

$$P_w = 3.5C_w - 1.3, \quad (2)$$

with a coefficient of determination $R^2 = 0.94$. Winter precipitation is then converted to water equivalent (w.e.) depth (mm during the winter) for consistency with other terms in the water balance. The total annual precipitation (P_{tot} in mm yr^{-1}) is obtained by summing winter (P_w) and summer precipitation (P_a).

3.1.2. Air temperature and evaporation

The air temperature data are provided by the Tarfala Research Station. Data gaps were filled using data from secondary sensors at the station and SMHI's Tarfala station, as well as downscaled ERA-40 data and a transfer function from the data of the nearby SMHI Katterjåkk station by Jonsell et al. (2013). Air temperature data are used to estimate the annual potential evaporation rate using the Hamon, (1961) model, which accounts for the effects of temperature on saturation vapor density and thus potential evaporation (we use the term 'evaporation' given the negligible vegetation cover in the catchment). Days with a mean temperature below 0 °C were assumed to have a negligible evaporation rate. We also assumed that in the typically wet conditions of the Tarfala catchment, potential evaporation is a good approximation of actual evaporation. The Hamon method may provide biased estimates in some locations, but it becomes accurate once evaporation rates are adjusted to reflect local conditions (Rosenberry et al., 2004). To this aim, the yearly potential evaporation from Hamon's method in the whole study period is rescaled by the spatially averaged evaporation over the whole Tarfala catchment, previously estimated at 49 mm yr^{-1} . This value is the area-weighted average between 70 mm yr^{-1} of evaporation from the unglacierised areas of the catchment (Schytt, 1973), and 0 mm yr^{-1} of evaporation from the glacierised areas, based on the assumption that evaporation and condensation balance out over the glaciers (Östling and Hooke, 1986).

3.1.3. Discharge

Discharge time series at the catchment outlet are calculated from water stage records at the gauging station Rännan (Fig. 1) through the use of stage-discharge relations (rating curves). These relations are calibrated using stage-discharge measurements (gaugings). The gaugings were largely conducted by salt dilution using the finite mass dilution method (Aastad and Søgne, 1928; Okunishi et al., 1992). Sporadically, dye dilution and current meter methods have also been employed (Bronge, 1985). Some modifications in Rännan's hydraulic cross-section have caused the stage-discharge relation to vary over time. Several rating curves were therefore used depending on their validity period. A summary of the rating curves used until 2002 can be found in Sannebro (2006). For years 2003–2005 the rating curve was updated based on new measurements conducted in 2005 (Snoek, 2005a, 2005b). After 2005, a new rating curve was adopted based on measurements available from 2006 to 2008 (Quinlan, 2016). Sediment transport and a pressure sensor failure made data unreliable in 2003. In addition to the discharge time series at the catchment outlet, stage records are available from the gauging station at Lillsjön (Fig. 1), which has been operational between July and September since the 1960s. This station serves partly to determine the discharge of the Storglaciären sub-catchment as it is located at the outlet of the biggest sub-catchment in the Tarfala valley. In this study we used the mean daily discharge data compiled by Sannebro, (2006) and Dahlke et al. (2012). Similar to the relations established by Schneider (1995) and Sannebro (2006), linear regressions between discharges at Rännan and Lillsjön were determined to fill the gaps where summer discharge data (June–August) were missing. This was the case for the summer discharge in the years 2006 and 2008, when ten days were filled.

For the ten years when reliable winter runoff records are not available, a recession curve was used to reconstruct discharge data. The recession curve follows a relation based on the amount of time after the end of ice melt, which was determined using existing winter discharge records by Schytt (1973) and formalised by Bronge (1985) as:

$$Q = kt^{-\alpha}, \quad (3)$$

where k and α are empirically derived constants and t is the time in days after the end of ice melt, defined as the time at which the daily air temperature decreases below 0 °C.

3.1.4. Glacier mass balance and storage change

The glacier mass balance of Storglaciären is determined by the direct glaciological method (Kaser et al., 2003; Østrem and Stanley, 1969). Snow depth is determined by probing in a 100 × 100 m grid (Jansson, 1999). Density measurements are made in snow pits or by core drilling. Typically, 40–50 ablation stake measurements are used for summer balance calculation. The winter balance is measured in a concentrated period between late April and early May, depending on weather conditions, and the summer balance is measured in mid-September (Holmlund and Jansson, 1999), due to the well-defined and relatively constant timing of transition between melt and accumulation seasons. The accuracy of the mass balance time series was improved by Holmlund et al., (2005) who exploited higher quality topographic maps than were used previously to recalculate the glacier area. This reduced the error, which was largest in early years and typically less than 0.1 m w.e. after 1960. The difference between winter and summer balances results in the net mass balance (Fig. 5).

Errors in the mass balance measurements have been discussed by Jansson (1999) and point at decimeter accuracy in the resulting balance values for the glacier. There are, however, still errors that are not accounted for. The most important of these is internal accumulation (Trabant and Mayo, 1985). Schneider and Jansson (2004) studied the internal accumulation on Storglaciären in two seasons and found that up to 4% (0.05 m w.e.) of the accumulation, as recorded by surface mass balance measurements, could remain in the glacier system as internal

accumulation. This mass accumulates in the firn pack and the total error depends on the size of the firn area. The error due to internal accumulation has not been applied to the mass balance reported for the glacier since a longer time series of the necessary measurements are lacking. In the absence of mass balance observations of the other five glaciers, the net balance of Storglaciären is used to estimate the volume of water stored by all ice bodies in the catchment. While digital elevation models (DEMs) of Storglaciären are available in specific years (Koblet et al., 2010) over the period 1969–2008 and the net mass balance was calculated per elevation band, this is not the case for the other five glaciers. To relate the area-elevation distribution of Storglaciären, with that of the other five glaciers, we used a high resolution (2 m × 2 m) DEM from 2016 from the Swedish topographic authority and extracted the area-elevation distribution for the other five glaciers. The annual Storglaciären net mass balance per elevation band over the period 1969–2008 is then multiplied by the area within the same elevation band for the other glaciers. By summing all the volumes, the total annual amount of water stored by the glacierised area is obtained. The area changes of Storglaciären - the largest in the catchment - between 1970 and 2010 are small (0.14 km²) and mainly occurred at low elevations (Koblet et al., 2010). The extent of the other five glaciers at these elevations is limited. Therefore, the effect of assuming constant an area for the other glaciers on the storage change is small over the study period.

3.2. Data analysis

3.2.1. Water balance estimation

The water balance equation used for the glacierised Tarfala catchment is a modified version of the general equation described by Kane and Yang (2004) in their review of water balance studies at high latitudes and is considered at an annual time scale:

$$P_{\text{tot}} - E - Q - \Delta S = \varepsilon \quad (4)$$

where P_{tot} , E , Q and ΔS are the inputs, representing the total precipitation, the actual evaporation, the runoff and the overall water storage change, respectively, and ε represents the water balance residual. Eq. (4) can easily be recast into a form more typically used in water balance calculations: $\Delta S = P_{\text{tot}} - E - Q - \varepsilon$. The annual water balance components are calculated over the hydrological year. The input, P_{tot} , comprises the summer and winter precipitation (section 3.1.1). Evaporation is estimated using the Hamon method (section 3.1.2) and runoff is calculated using discharge data from the station Rännan (section 3.1.3). The storage change generally accounts for glacier, snow, groundwater and lake storage. The water volume contained in snow patches lasting over one year is expected to be small. Similarly, the contribution of groundwater or soil moisture storage can be considered negligible, as 30% of the catchment area is glacierised, 42% is bedrock or talus and the remaining 25% is till which is partly permanently frozen (Nilsson and Larsson, 1976). Groundwater recharge through boundary exchange is unlikely because the catchment is well contained by mountains with exposed bedrock on all sides. The storage change of Tarfalsjön has been estimated at less than $0.1 \times 10^6 \text{ m}^3$ (<5 mm after converting to a mean water depth) over the hydrological year and the storage change of other, smaller lakes in the valley is estimated to be even less (Schytt, 1973). Lake storage changes are thus considered negligible. Therefore, we assumed that all storage components except glacier storage are negligible at an annual time scale and we tested the validity of this assumption. The storage change is obtained by converting the storage of the glacierised area to a mean water depth over the catchment.

As a negative net mass balance represents a decrease in the volume of water being stored by the glaciers and a positive net mass balance represents an increase in storage by the glaciers, then $\Delta S < 0$ indicates that outputs from the catchment (especially runoff) are greater than inputs. The residuals, ε , indicate the level of combined measurement error, processing error and storage changes that were either unaccounted for

or assumed to be negligible. Positive residuals imply overestimation of the inputs compared with the outputs, or an underestimation of ΔS caused by an increase in unaccounted water storage. Our analysis of correlations between residuals and other hydrologic components elucidates possible over- or under-estimations driven by particular hydroclimatic conditions.

3.2.2. Statistical analyses

Statistical trends in precipitation, temperature, glacier storage change and discharge time series were analysed using the Mann-Kendall trend test. This is a non-parametric test that tests the null hypothesis H_0 : 'there is no monotonic trend' against the alternative hypothesis, H_1 : 'there is a monotonic trend'. It analyses differences in signs between earlier and later data values by comparing each value with the values preceding it in the time series. The entire study period was tested for the existence of trends using this statistical test at a significance level of 5%. In addition, we investigated differences between decades, comparing the decadal mean precipitation, temperature, and discharge (years 1969–1978, 1979–1988, 1989–1998, and 1999–2008) using the Tukey-Kramer test with a 95% confidence interval.

In addition to the trend analysis, we also compared the flow duration curves of the four different time intervals. Flow duration curves were constructed by calculating the daily discharge, Q , that is exceeded for a given percentage of time in the hydrologic year.

3.2.3. Uncertainty assessment of the water balance estimation

The uncertainty of the water balance for the glacierised Tarfala catchment is assessed by estimating the relative contributions of each uncertainty source to the total uncertainty.

In this paper, we use the 95% uncertainty interval as it is common practice in Hydrometry and recommended by the Hydrometric Uncertainty Guidance (HUG, 2007). Working with uncertainty intervals is not appropriate for investigating the relative contributions of each uncertainty source to the total uncertainty because these intervals are not additive. However, variances of random variables are additive if these variables are independent and follow Gaussian distributions. We therefore assume that errors associated with each source of uncertainty follow a Gaussian distribution centred at zero and with a standard deviation expressed as the quarter-length of the uncertainty interval. This means that when a Gaussian error, ϵ_X , of a random variable X is

specified as $\pm w$, its corresponding standard deviation, σ_X , is $w/2$.

For precipitation measured by a gauge at a single point, systematic errors from undercatch are the main source of uncertainty (McMillan et al., 2012). For areal rainfall estimates, the number and spatial distribution of gauges is a source of uncertainty. This uncertainty was estimated to be 5–10% at a daily scale for 20 or more gauges in a catchment of 135 km² (Villarini et al., 2008). In the Tarfala catchment, the areal estimates are calculated from an average of 20 gauges, uniformly distributed over a smaller area of 20 km². We therefore expect that this error is negligible compared to the undercatch source of uncertainty. Undercatch error has been estimated to typically range between 5 and 16% (McMillan et al., 2012). In this study, the undercatch error affects only summer precipitation (section 3.1.1). We corrected summer precipitation for evaporation from the precipitation gauge, wetting losses and wind losses based on Alexandersson's (2003) approach for monthly precipitation estimates in Sweden. In this approach, evaporation losses from the precipitation gauges are calculated by multiplying the monthly mean temperature by a monthly variable coefficient (equal to 0.14, 0.12, 0.10, 0.10 from June to September). The evaporation correction is applied for months with a mean monthly temperature above 0 °C. Wetting losses (equal to 1.2, 1.5, 1.6, 1.7 mm from June to September) are added to monthly summer precipitation totals. A constant correction factor (equal to 1.12) is applied to correct precipitation for wind losses. This factor is recommended for mountain locations above the tree line. The overall correction of summer precipitation results in a mean increase of 14% for summer over the study period. We therefore assume that systematic errors from undercatch are compensated by the precipitation correction.

After correcting the systematic errors, the random error associated with summer precipitation measurements is assumed equal to $\pm 10\%$ (McMillan et al., 2012). Uncertainties in observations were considered when estimating the winter precipitation component (section 3.1.1). A measurement error of 0.1 m w.e. was assumed for the glacier winter mass balance (Jansson, 1999). This uncertainty was propagated to the winter precipitation estimations (Eq. (2) in section 3.1.1) using Monte Carlo techniques. The winter precipitation estimations with the observation error and the random error on summer precipitation were propagated by Monte Carlo techniques in the computation of the total annual precipitation and its associated standard deviation, σ_P .

The error associated with evaporation estimations, ϵ_E , is assumed

Table 1

Statistics (mean, standard deviation (Std), coefficient of variation (CV)) and Mann-Kendall (MK) trend test of yearly and seasonal precipitation, discharge and air temperature for the Tarfala catchment. Winter and summer are defined by the months October-May and June-September, respectively.

	Mean	Std	CV	MK trend test
Precipitation				
Year	mm y ⁻¹ 1301	mm y ⁻¹ 291	– 0.22	>0.05
Winter	mm per winter 784	mm per winter 207	0.26	>0.05
Summer	mm per summer 517	mm per summer 187	0.36	>0.05
Corrected precipitation				
Year	mm y ⁻¹ 1375	mm y ⁻¹ 311	– 0.2	>0.05
Summer	mm per summer 591	mm per summer 210	0.35	>0.05
Discharge				
Year	m ³ s ⁻¹ (mm y ⁻¹) 0.91 (1446)	m ³ s ⁻¹ (mm y ⁻¹) 0.18 (291)	– 0.20	>0.05
Winter	m ³ s ⁻¹ (mm per winter) 0.10 (104)	m ³ s ⁻¹ (mm per winter) 0.05 (56)	0.55	>0.05
Summer	m ³ s ⁻¹ (mm per summer) 2.6 (1372)	m ³ s ⁻¹ (mm per summer) 0.63 (333)	0.24	<0.05
Air temperature				
	°C	°C	–	
Year	–3.4	0.8	–0.23	<0.05
Winter	–7.5	0.9	–0.12	<0.05
Summer	4.8	1.0	0.21	<0.05

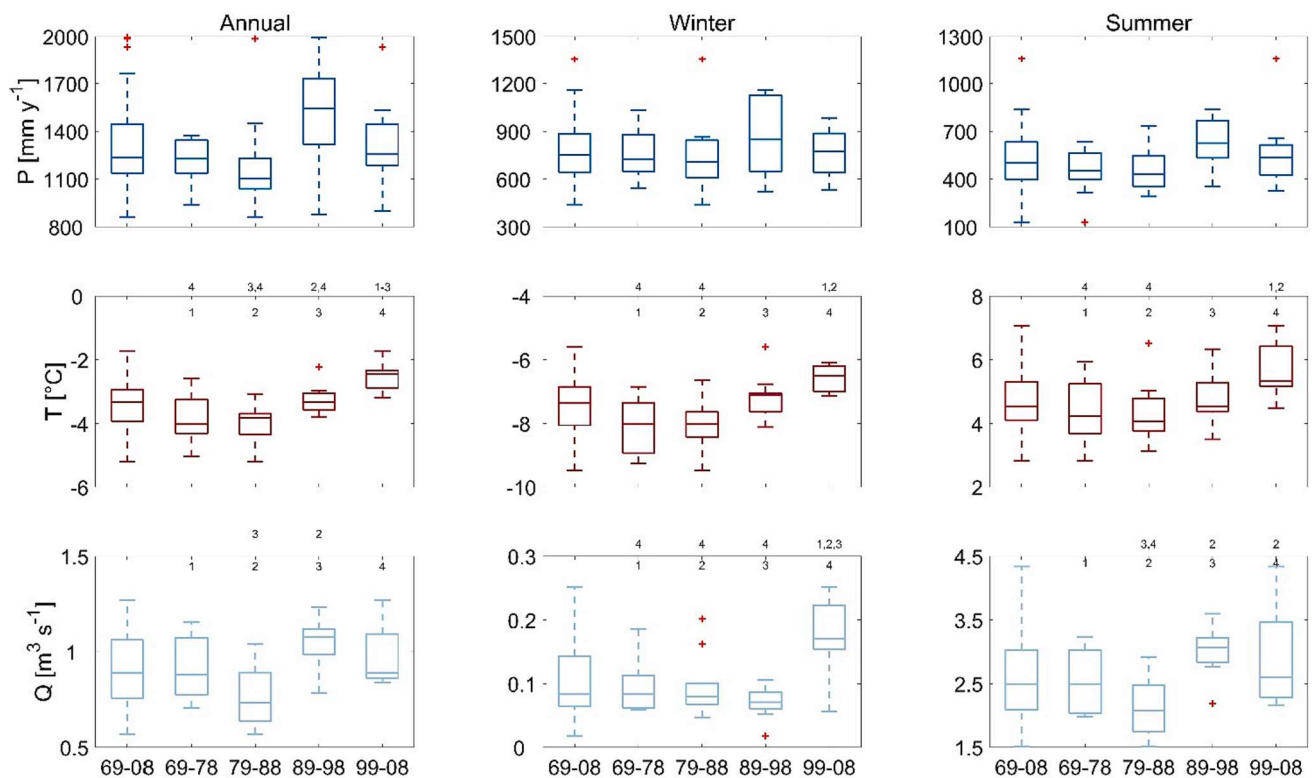


Fig. 2. Annual, winter and summer precipitation (top row), air temperature (middle), and discharge (bottom) distributions for the four decades. Numbers (1–4) within the panels represent the four decades in the considered period; numbers above the panels show when the mean of precipitation, air temperature, and discharge are significantly different between decades. The comparison between the mean for the four decades is performed by the Tukey-Kramer test with a 95% confidence interval.

equal to $\pm 25\%$. This is the approximate error associated with the Hamon method when compared with other empirical and energy balance-based evaporation estimates (Rosenberry et al., 2004). As no uncertainty quantification for discharge in the study catchment was available, the error associated with the discharge estimation, ϵ_Q , is assumed equal to $\pm 10\%$ —a typical value for medium flows (McMillan et al., 2012). The error associated with storage change, $\epsilon_{\Delta S}$, is assumed equal to ± 0.1 m w.e., which is the error associated with the mass balance calculation of Storglaciären (Jansson, 1999).

The uncertainty associated with the water balance estimation is assessed by considering the four sources of uncertainty described above (precipitation, evaporation, discharge and storage changes). Due to the assumption that errors are normally distributed, the total uncertainty associated with the water balance estimation (see Eq. (4) in section 3.2.1) follows a Gaussian distribution centred at zero and with a standard deviation $\sigma_{WB}^2 = \sigma_P^2 + \sigma_E^2 + \sigma_Q^2 + \sigma_{\Delta S}^2$.

The uncertainty budget of the water balance estimation is explored for each year by computing the following four ratios corresponding to each uncertain component: $\left(\frac{\sigma_P^2}{\sigma_{WB}^2}, \frac{\sigma_E^2}{\sigma_{WB}^2}, \frac{\sigma_Q^2}{\sigma_{WB}^2}, \frac{\sigma_{\Delta S}^2}{\sigma_{WB}^2}\right)$. Note that the four components of the uncertainty budget add up to exactly one as $\sigma_{WB}^2 =$

$$\sigma_P^2 + \sigma_E^2 + \sigma_Q^2 + \sigma_{\Delta S}^2.$$

4. Results

4.1. Precipitation

The mean total precipitation was 1301 mm y^{-1} (Table 1). The mean winter (October–May) precipitation was clearly larger ($784 \text{ mm per winter}$) than the mean summer (June–September) precipitation ($517 \text{ mm per summer}$) over this period. The mean total corrected precipitation was 1375 mm y^{-1} . The Mann-Kendall trend test revealed no trends in total, summer or winter precipitation (Table 1). There were no significant differences between the mean total precipitation within each of the four decades (1969–78, 1979–88, 1989–98 and 1999–08; Fig. 2). This was also true for the winter and summer precipitation (Fig. 2).

4.2. Air temperature and evaporation

The mean air temperature during the period 1969–2008 was -3.4°C with mean winter and summer air temperatures of -7.5°C and 4.8°C , respectively (Table 1). Significant trends were detected in the mean

Table 2

Statistics (mean, standard deviation (Std), coefficient of variation (CV)) and Mann-Kendall (MK) trend test of calculated variables for the Tarfala catchment (annual water balance components).

	Mean	Std	CV	MK trend test
Actual Evaporation	mm y^{-1}	mm y^{-1}	–	
Year	49	6	0.1	<0.05
Storage change	mm y^{-1}	mm y^{-1}	–	
Year	–27	104	–3.8	>0.05
Water balance residuals	mm y^{-1}	mm y^{-1}	–	
Year	–112	305	–3.1	>0.05

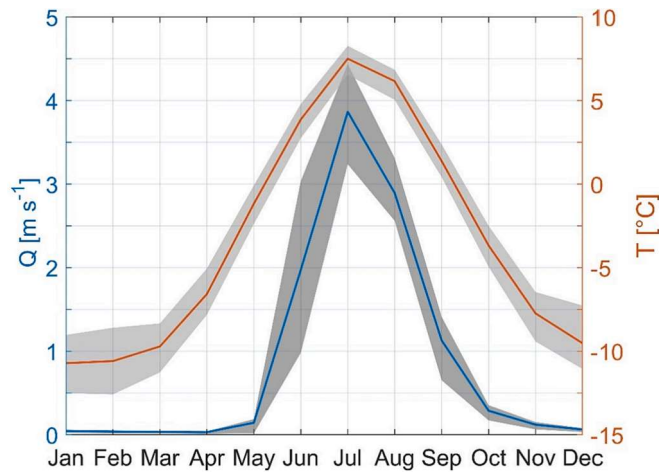


Fig. 3. Mean monthly discharge and air temperature over the period 1969–2008. The shaded areas indicate the 25th and 75th percentiles of the monthly discharge and temperature.

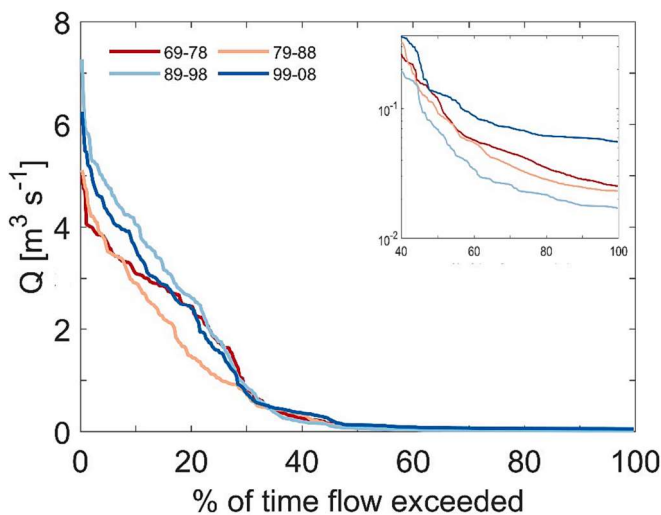


Fig. 4. Annual flow duration curves indicating the discharge, Q that is exceeded for a given percentage of time in the hydrologic year for the four decades (1969–1978, 1979–1988, 1989–1998, 1999–2008).

annual and seasonal air temperature over the whole period (Table 1). Significant differences in mean annual air temperatures were found between the decades 1969–78 and 1999–08, 1979–88 and 1989–98, and 1979–88 and 1999–08 (Fig. 2). Significant differences in mean winter and mean summer air temperatures were also found between the decades 1969–78 and 1999–08 and 1979–88 and 1999–08 (Fig. 2). The mean actual evaporation was 49 mm y^{-1} over the whole considered period (note that this value was used to constrain the estimation from Hamon's approach) and a significant temporal trend was found (Table 2).

4.3. Discharge

The mean annual discharge during the period 1969–2008 was $0.9 \text{ m}^3 \text{ s}^{-1}$ corresponding to 1446 mm y^{-1} (Table 1). The summer discharge between the months of June and September represented the largest proportion of the annual discharge (Fig. 3) with a mean of $2.6 \text{ m}^3 \text{ s}^{-1}$ corresponding to 1372 mm per summer (Table 1). In some years, discharge could not be calculated because of missing data for significant portions of the melt season from both the Rännan and Lillsjön gauging

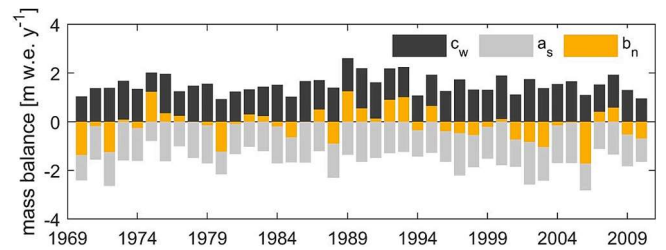


Fig. 5. Storglaciären winter (c_w), summer (a_s) and net mass balance (b_n).

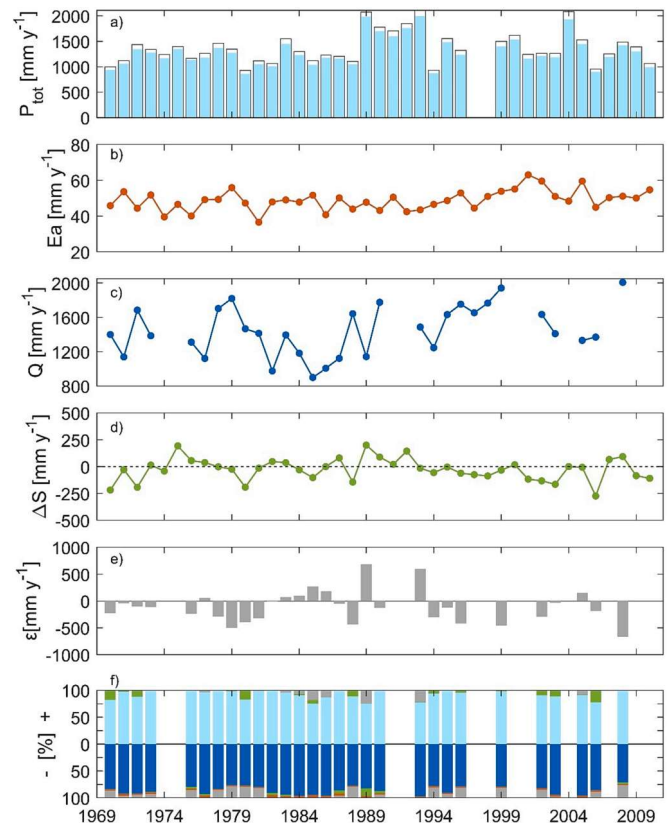


Fig. 6. Annual water balance components: (a) precipitation inputs, P_{tot} (combined summer and winter with empty bar indicating corrected precipitation); (b) actual evaporation rate, E_a ; (c) runoff at the catchment outlet, Q ; (d) change in storage, ΔS (change in the volume of glacierised area); (e) water balance residuals, ϵ . (f) Summary of water balance components, expressed as relative to the total inputs (positive values) or outputs (negative values), and using the same colour scheme as in panels (a)–(e). Note that the y-axis ranges vary among the panels to highlight the variability of each component.

stations. The Mann-Kendall test showed that no significant trend was detected in the total discharge over the study period, while significant increasing trends were found for the summer discharge (Table 1). Significant differences in the total annual discharges were found between the second decade (1979–1988) and the third decade (1989–1998; Fig. 2). Specifically, winter discharges after 1999 were significantly higher than in the previous decades and summer discharges after 1988 were higher than in 1979–1988 (Fig. 2).

Fig. 4 shows the annual flow duration curves calculated for the four decades. In all decades, low flows ($<1 \text{ m}^3 \text{ s}^{-1}$) occurred for more than half of the time, as is typical for a high alpine and highly glacierised catchment. However, a clear change occurred across the four decades, with increased frequency of high flows ($>3 \text{ m}^3 \text{ s}^{-1}$) in the last two decades. The number of days with high flows increased, especially between the second decade (1979–1988) and the last two decades

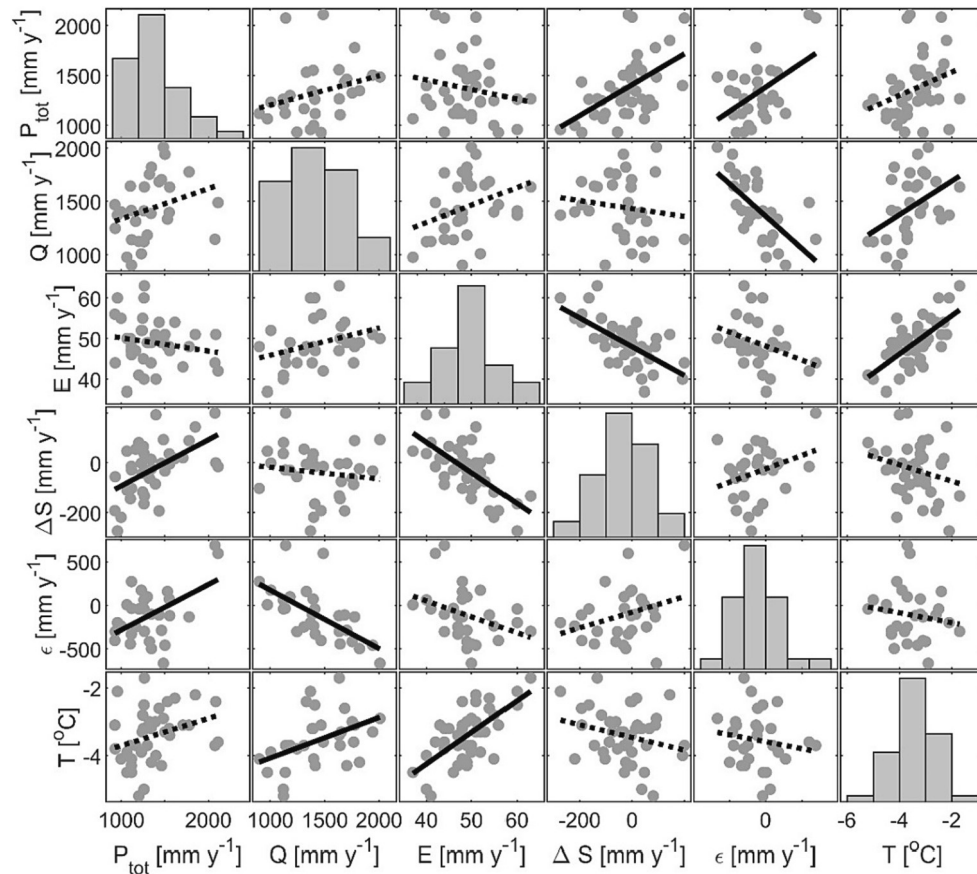


Fig. 7. Scatterplots between pairs of annual water balance components (precipitation, P_{tot} ; discharge, Q ; evaporation, E ; change in glacier storage, ΔS ; water balance residuals, ϵ) and between water balance components and mean annual temperature, T . The histograms on the diagonal represent the frequency plots for each variable. Solid (dashed) lines are least square regressions of the data with slope significantly (not significantly) different from zero.

Table 3

Pearson correlation coefficients between pairs of water balance components (precipitation, P_{tot} ; discharge, Q ; evaporation, E ; storage change, ΔS ; water balance residuals, ϵ) or between water balance components and mean annual temperature, T . Asterisks indicate significant correlations (*, $p < 0.05$; **, $p < 0.01$).

	P_{tot} (mm y ⁻¹)	Q (mm y ⁻¹)	E (mm y ⁻¹)	ΔS (mm y ⁻¹)	ϵ (mm y ⁻¹)	T (°C)
P_{tot} (mm y ⁻¹)	1.00	0.29	-0.18	0.54**	0.50**	0.31
Q (mm y ⁻¹)		1.00	0.33	-0.13	-0.64**	0.44*
E (mm y ⁻¹)			1.00	-0.66**	-0.35	0.66**
ΔS (mm y ⁻¹)				1.00	0.30	-0.25
ϵ (mm y ⁻¹)					1.00	-0.15
T (°C)						1.00

(1989–1998, 1999–2008). The increase in number of days with high flows compared to the second decade was 30 and 20 days, respectively.

4.4. Glacier mass balance and storage change of the glacierised area

A negative net mass balance of Storglaciären, which is used in this study to extrapolate the mass balance of the whole glacierised area, characterised 24 out of 40 years (Fig. 5). This translated into a

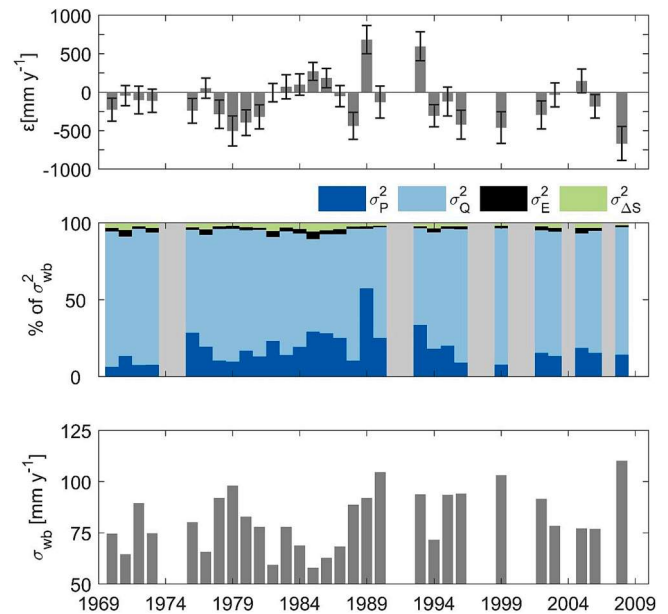


Fig. 8. (a) Annual water balance residuals with error bars indicating all combined errors of the water balance components; (b) summary of the uncertainty of each water balance component (σ_P^2 , σ_E^2 , σ_Q^2 , $\sigma_{\Delta S}^2$), expressed as relative to the total water balance uncertainty (σ_{wb}^2 ; the grey shade indicates missing data); (c) standard deviations of the total error.

corresponding negative storage change at the catchment scale (Fig. 6d) averaging -27 mm y^{-1} for the study period (1969–2008, Table 2).

4.5. Water balance estimation and uncertainty

Fig. 6a–e shows all the components of the annual hydrologic balance. The relative contribution of each component is illustrated in Fig. 6f. The water balance and the residuals were estimated for the 29 years of the study period for which all components could be measured or estimated. The mean of the estimated water balance residuals was -112 mm y^{-1} (Table 2). Large residuals are associated with individual years (Fig. 6e) and no significant trends in the residuals were found over the whole study period (Table 2). Precipitation consistently dominates the water balance inputs and runoff represents the largest output throughout the study period (Fig. 6f). The storage changes and water balance residuals are generally comparable in magnitude, whilst evaporation represents the smallest contribution to the water balance (Fig. 6f).

Some of the water balance components were significantly correlated (Fig. 7; Table 3). In particular, precipitation was positively correlated to change in storage and discharge was positively correlated to mean annual temperature. The water balance residuals were positively correlated with precipitation and negatively with discharge, whereas evaporation was negatively correlated with change in storage and positively with mean annual temperature.

Fig. 8 illustrates the results from the uncertainty assessment of the water balance estimation for the 29 years for which all water balance components could be estimated. The annual residuals from the water balance estimation are first shown with their associated total uncertainty intervals (Fig. 8a). Then, the contribution of the different sources of uncertainty is shown for each year (Fig. 8b). Finally, the standard deviations of the total uncertainty are shown for each year in Fig. 8c. While the error associated with the evaporation estimations was relatively large ($\pm 25\%$), its contribution to the total uncertainty of the water balance estimation was almost negligible each year (always under 2%, Fig. 8b). The relative contributions of discharge and storage changes to the total water balance uncertainty are on average less than 76% and less than 4%, respectively (Fig. 8b). The storage change contributed less than discharge to the water balance (Fig. 6) and its contribution to the total uncertainty is relatively low (Fig. 8b). The contribution of precipitation uncertainty was under 18% (Fig. 8b), lower than that of discharge. The discharge component contributed the most on average to the total water balance uncertainty. Its contributions to the total uncertainty are more important in years when the water balance residuals are large and thus the standard deviation of the total uncertainty are larger in absolute value (Fig. 8c).

5. Discussion

5.1. Trends in the water balance components

A mean total precipitation of 1301 mm y^{-1} and mean summer precipitation of 517 mm per summer were found for uncorrected precipitation over the study period (Table 1). After applying the precipitation correction, these values increased to a mean total precipitation of 1375 mm y^{-1} (white bars in Fig. 6a; Table 1) and a mean summer precipitation of 591 mm per summer (Table 1). Early estimates of total precipitation at the Tarfala Research Station were in the range $900\text{--}1000 \text{ mm y}^{-1}$, of which 350 mm occurred in the summer for the period 1965–87 (Grudd, 1990; Schytt, 1973; used 1 station). Later estimates equalled $1997 \pm 450 \text{ mm y}^{-1}$, of which June–Aug precipitation accounts for $400\text{--}600 \text{ mm}$ for the period 1980–2009 (Dahlke et al., 2012; used 1 station). Our (corrected) precipitation estimates fall between these values. Winter precipitation was the largest precipitation component, which is expected for a northern mountain catchment. The average proportion of total precipitation which falls as snow (59% for uncorrected precipitation and 56% for corrected precipitation) is similar to

that found in Finnish low Arctic catchments with a maritime influence (Young et al., 2006). The total annual and total winter precipitation did not show significant differences between the decades (Fig. 2) and no significant trends were detected (Table 2).

In contrast to the relatively steady precipitation rates, increasing trends in mean annual, winter and summer air temperatures were found (Table 1). Mean winter and mean summer air temperature were significantly different between the first and last decade by about $+1^\circ\text{C}$ (Fig. 2). This warming trend was reflected in the increasing evaporation rate, as evaporation was estimated from air temperature. The mean evaporation rate of 49 mm y^{-1} (imposed to correct bias in Hamon's method) is within the range found for Norwegian glacierised catchments located between maritime and moderate continental climates (Engelhardt et al., 2014) and lower than evaporation estimates for glacierised catchments in Svalbard and Greenland (Killingtveit et al., 2003; Mernild et al., 2008). It is possible that our approach underestimates water vapour losses because it neglects sublimation, but it might overestimate summer evaporation during relatively dry years, because we have assumed that actual and potential evaporation are the same. Any error in the evaporation estimates, however, would not contribute enough to the water balance to explain the residuals given the small value of evaporation rates compared to precipitation, discharge, and storage changes.

Annual mean total discharge in the Tarfala catchment was 1446 mm y^{-1} with 1372 mm y^{-1} occurring during summer (June–Sept) over the study period. No significant trends were found in annual discharge, while significant trends were found in summer discharge over the study period (Table 1). Mean summer discharge was higher during the last two decades, 1989–2009, compared to the second decade, 1979–1988 (Fig. 2). Annual precipitation rates remained relatively stable over time (Fig. 2, upper row), but summer precipitation in the last decade (1999–2008) was lower than the previous decade (1989–1998). This result suggests that the increase of summer discharge can be attributed to increasing air temperatures affecting snow cover and glacier melt (Fig. S3).

This result is consistent with that found by Dahlke et al. (2012) for the same period, who explained the increase of summer discharge with more extensive disappearance of glacier snow and occurrence of earlier snow melt in the catchment due to increasing air temperatures. In that study, the mean summer discharge from the Tarfala catchment was significantly positively correlated to the annual average Atlantic Multidecadal Oscillation (AMO), of which the positive phase brings increased temperatures (particularly in summer) to Scandinavia (Dahlke et al., 2012). Earlier snowmelt from the glaciers translated into a longer exposure of glacier ice (with lower albedo), which in turn produced more meltwater. This hypothesis seems to be supported by the aforementioned temperature increase (Fig. 2, last column), the occurrence of more consecutive years of annual negative storage changes during the same period (Fig. 6c) and an increase in summer glacier storage contributions (Fig. S3). The contribution of annual storage changes is relatively small compared to the annual precipitation ($4 \pm 1\%$ in 1994–2008). However, it can be large compared to summer precipitation ($14 \pm 23\%$ in 1994–2008). Glacier melt contributions can also be enhanced by extreme precipitation events (Van Tiel et al., 2021). Dahlke et al. (2012) reasoned that not only increasing temperatures, but also more frequent extreme events in summers could have caused increased glacier contribution to discharge. The redistribution from the medium to short flow exceedance in the last two decades highlighted by the analysis of the flow duration curves confirms this hypothesis (Fig. 4).

At the annual time scale, on the one hand, positive storage changes were associated with high precipitation, but not with low temperature (Fig. 7). On the other hand, discharge was positively correlated with air temperature, but not with precipitation and negative storage changes. These results can be explained by the hydrological response, in part to climate forcing at the interdecadal scale and in part to interannual meteorological variability within the decades. The decade 1979–1988 was characterised by low temperatures, favouring conditions of reduced

snow and glacier melt (Fig. 2 and Fig. 6c). In addition, the summer precipitation variability suggests that precipitation can also generate increased discharge in specific years, thereby weakening discharge-precipitation correlations. The effect of increasing air temperature on annual catchment runoff due to glacier retreat is therefore not yet evident in the Tarfala catchment over the study period, unlike the glacierised catchments in central Europe (e.g., Bard et al., 2015) and in some regions in Canada (e.g., Fleming and Clarke, 2003).

Some of the correlations illustrated in Fig. 7 are expected and depend on how hydrologic fluxes were calculated. The negative correlation between evaporation and change in storage is partly explained by evaporation being a function of monthly temperatures when above zero degrees; warm summers cause both higher glacier volume losses and higher evaporation. Moreover, it is not surprising that evaporation increases significantly with annual temperature because annual temperature is correlated with the mean temperature of the months where it is above zero degrees, which is used to calculate evaporation.

The positive and negative correlations of the water balance residuals with precipitation and discharge, respectively, might be expected because the residuals are calculated as the difference between precipitation and water outflows, including discharge. These two fluxes are the largest in the water balance, and they do not balance each other at the annual time scale. Therefore, the annual residuals increase when the precipitation increases are not met by similarly higher discharge. We discuss how legacy effects from previous years might explain this result in the following section.

5.2. Uncertainty in the water balance components and residuals

The mean of the water balance residuals was equal to -112 mm y^{-1} with a standard deviation of 305 mm y^{-1} (Table 2). This result indicates that our water balance could not be perfectly closed. The mean value is similar to those found by Kane and Yang (2004) in their review of high latitude catchments and by Killington et al., (2003). Large residuals were present in a few individual years when measurement errors or data reconstruction could have affected water balance estimations (Fig. 6e and f).

In this regard, the assessment of the water balance uncertainty showed that even if the uncertainty associated with discharge could be considered low (error of $\pm 10\%$ of discharge for each year), its contribution to the total uncertainty is important (on average 76%) compared

with other components (Fig. 8b). For example, we assumed a larger error for evaporation ($\pm 25\%$) but because evaporation contributes the least to the water balance (Fig. 6f), its contribution to the overall uncertainty was also negligible compared to the other components. This suggests that a proper quantification of the uncertainty associated with the discharge component would enhance the uncertainty assessment of the whole water balance. Precipitation (the other major water balance component) was corrected for undercatch in its summer component. This was considered the main source of uncertainty, since the uncertainty related to areal rainfall estimates was considered negligible (section 3.1.1). The remaining random errors of the summer precipitation measurements and the winter precipitation uncertainty contributed less than uncertainties in discharge. In general, errors in the water balance components were set up with constant standard deviations in this paper. Whilst this approach allows assessment of the uncertainty in the absence of more specific information, a time-invariant uncertainty may not be equally appropriate for all components. For example, the stage-discharge relations have indeed shifted over time at the Rännan station, which indicates the need for a rating curve uncertainty quantification and suggests that discharges computed from the rating curves may have uncertainties that vary through time.

In some years, the water balance residuals lay within the total water balance uncertainty (Fig. 8a) and thus they could be explained by the error associated with the water balance components. In other years, the error associated with the water balance components could only explain part of the water balance residual (Fig. 8a). However, the majority of these larger residuals were connected to specific climatic conditions, as shown in Fig. 9. The most negative of the residual values could be mostly explained by differences in precipitation and temperature between the year when the water balance was calculated and in the antecedent hydrological year. Specifically, high (i.e., above the 29-year median) winter or summer precipitation in cold antecedent years could cause large parts of the catchment to remain snow covered or could result in snow at the higher elevations, even in the summer. This represents an extra storage component that was not measured, but that could contribute to high catchment discharge in years with warm summers that followed a cold year (Fig. 9 and Fig. S4).

Some negative water balance residuals could not be explained by these conditions, but they could be explained by more than one consecutive year of high winter and summer precipitation. In more recent decades, increasing air temperature caused a reduction of the seasonal snow cover in the catchment. In such years, it is likely that negative residuals could be associated with a different storage contribution, e.g., water stored in the active layer between the land surface and permafrost. Under warming, changes in controlling factors such as temperature, precipitation phase and snow properties are expected to affect the dynamics of seasonally frozen ground, including areas of discontinuous permafrost (Ala-aho et al., 2021), with a resulting effect on their hydrological regime. The impact of this phenomenon on the water balance is, however, difficult to establish in this study, given the gaps in the annual discharge data and consequently residual estimations in the last decade of the study period, and the lack of detailed data on soil and groundwater storage. Work by Bodeving (2020) showed that hillslope storage and runoff generation processes in particular are still little understood in the Tarfala catchment and future work is needed to investigate their role in the seasonal and annual water balance.

6. Conclusions

We analysed the different components (precipitation, runoff, change in storage, and evaporation) of the seasonal and annual water balances of the glacierised Tarfala catchment, Sweden, over ~ 40 years (1969–2008). No significant trends were found in precipitation over the study period, while summer discharge increased. A positive correlation was found between the annual mean air temperature and discharge, suggesting that the ongoing warming trend was causing an increase in

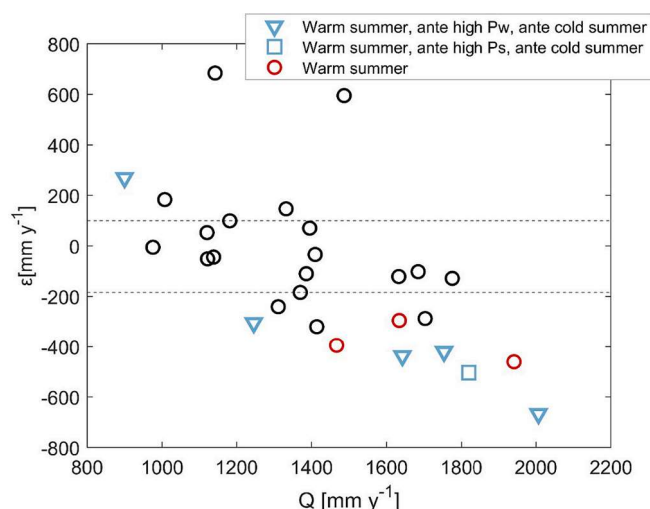


Fig. 9. Relation between annual discharge and water balance residuals. Dashed lines indicate the limits of the combined error of the water balance components ($-175; 85$). Inverted triangles and squares indicate years with warm summers, a cold antecedent summer, and high winter (Pw) and summer precipitation (Ps), respectively; red circles indicate years with warm summers.

discharge in the first decade of the millennium.

We also found negative water balance residuals, indicating that the water balance could not be perfectly closed because water inputs, e.g., precipitation and water added from storage changes (negative glacier mass balances, or soil, lake, groundwater and multi-year snowpack contributions) were underestimated, or the water balance components were too uncertain. Specifically, we identified uncertainties in discharge as particularly important. Therefore, future work should be devoted to better quantifying uncertainty in this component (e.g., estimating rating curve uncertainty over time). Finally, some (negative) residuals were outside the uncertainty bounds. These water balance residuals could, in principle, be caused by precipitation underestimations, discharge overestimations or surplus storage from antecedent years, but could also have a physical basis. In fact, we found that years with large negative residuals were characterised by warm summers and preceded by years with high winter precipitation and cold summers. In those warm years, the snow storage accumulated from the preceding years was released, causing high summer discharge and explaining the non-closure of the yearly water balance. This shows that to estimate the water balance in glacierised catchments, it is necessary to consider, in addition to measurement uncertainty, its components across hydrological years to understand ongoing and future changes.

Declaration of Competing Interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

Data availability

Data will be made available on request.

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Appendix A. Supplementary data

Supplementary data to this article can be found online at <https://doi.org/10.1016/j.jhydrol.2023.130028>.

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