



Climate change impacts on groundwater discharge-dependent streamflow in an alpine headwater catchment

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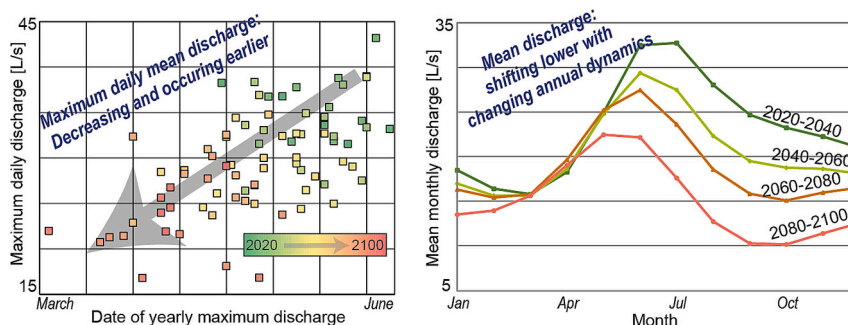
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HIGHLIGHTS

- 3-D groundwater (GW) model developed for a nival-regime alpine headwater catchment
- Annual streamflow dynamics in a GW-dependent stream impacted by climate change
- Minimum flowrates decrease and timing shifts from late winter to early autumn.
- Overall streamflow decreases no-flow periods start to occur by end of century.
- Groundwater has limited capacity to mitigate impact of snow cover loss.

GRAPHICAL ABSTRACT



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ABSTRACT

Climate change will have—and, in much of the world, is already having—a pronounced impact on alpine water resources. A deeper understanding of the future role of groundwater in alpine catchments, including quantification of climate change impacts on groundwater discharge, is vital for understanding the future of alpine water resources as a whole. Here, we develop and couple a geophysics-informed groundwater model with a net recharge model to investigate the impacts of climate change on a nival-regime alpine headwater catchment with significant unconfined Quaternary aquifer coverage. Flow in the groundwater-fed stream at the catchment outlet is analysed to determine changes in its annual dynamics. Comparing the periods 2020–2040 and 2080–2100 under ten RCP-8.5 climate models, we find a 35 % decrease in mean groundwater discharge and an increase in no-flow periods from ~0 % to 4.3 %. We also observe significant changes to the timing of monthly mean discharge maxima and minima, which shift ~1 month and ~5 months earlier, respectively. While groundwater has the potential to dampen the impacts of snow cover loss, currently perennial nival-regime alpine streams could be at risk of becoming intermittent by the end of the century. Our study underscores the increasingly critical role that groundwater will play in alpine catchments and emphasizes the need for quantitative understanding of the limits to its buffering capacity.

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1. Introduction

High mountain regions play a critical role in the water supply of downstream areas (Immerzeel et al., 2020; Viviroli et al., 2020, 2007). They contain large stores of water in the form of snow and ice and receive more precipitation due to the orographic effect on air masses. Climate change disproportionately affects mountain regions. Projected temperature increases and glacial mass depletion in many high mountain ranges threaten future downstream water supply (IPCC, 2021; Vuille et al., 2018). Furthermore, a quarter of the world's population is projected to depend on alpine regions for water resources by 2050 (Viviroli et al., 2020). This growing importance and vulnerability of alpine water resources, combined with studies showing the significant streamflow contribution of groundwater in a diverse range of alpine hydrological systems (e.g., Andermann et al., 2012; Clow et al., 2003; Cochand et al., 2019; Hayashi, 2020; Liu et al., 2004; McClymont et al., 2010; Muir et al., 2011), has led to increasing interest in the role of groundwater in the future of alpine hydrological systems.

Several types of hydrological units with relatively high hydraulic conductivities can be present in alpine catchments. These include talus, moraines, rock glaciers, alpine meadows, and fractured bedrock (Hayashi, 2020; Somers and McKenzie, 2020). They differ in their hydraulic properties, formation processes, extent, and, for Quaternary deposits, granulometry. Talus aquifers are made primarily of large, unsorted rock debris and have correspondingly high hydraulic conductivities. Their hydraulic conductivities may decrease with depth due to an increase in finer sediments with depth (e.g., Kurylyk and Hayashi, 2017). Moraines are formed by glacial action and consist of heterogeneous sediments with varying grain sizes till. Alpine moraines can have complex morphologies and, in the European Alps, commonly date from the various stages in the Würm glaciation or the Holocene (Ivy-Ochs et al., 2009). Because of their ubiquity, hydraulic properties, and large spatial extents, they play an important role in many alpine catchments.

Due to the relative inaccessibility and high heterogeneity of alpine catchments, direct subsurface measurements are often unfeasible. Non-intrusive characterisation and delineation of alpine hydrogeological units can be accomplished via a combination of numerical monitoring, surface observations, and geophysical measurements. The latter includes several methods that have been used in alpine catchments: seismic refraction (Langston et al., 2011; Rogger et al., 2017), ground penetrating radar (GPR) (McClymont et al., 2010), gravimetry (Arnoux et al., 2020a; Halloran, 2022; McClymont et al., 2012), and electrical resistivity tomography (ERT) (Glas et al., 2019; Langston et al., 2011; McClymont et al., 2012, 2010; Muir et al., 2011). ERT, combined with knowledge of the geology of a site (e.g., from surface geology maps), is well-suited to alpine aquifer delineation due to its portability and flexibility. It has been used to investigate the internal structure of moraines (Langston et al., 2011) and rock glaciers (Mewes et al., 2017). Data from geophysical methods can be used to inform finite-element (FE) numerical models that can assimilate more readily-available meteorological and surface hydrological data. In this manner, effective hydraulic properties can be determined. This improves a model's performance and avoids the use of non-site-specific literature values. Even with this restrictive amount of data and many unknown parameters, numerical modelling in the context of alpine hydrogeology has successfully been used to investigate and reproduce groundwater dynamics in various alpine aquifers (Arnoux et al., 2021; Ciruzzi and Lowry, 2017; Evans et al., 2018; Manning and Solomon, 2005; Morán-Tejeda et al., 2015; Rogger et al., 2017; Tague et al., 2008; Voeckler et al., 2014; Welch and Allen, 2014). Not only do numerical models help in the investigation of current groundwater dynamics in alpine aquifers, but they can also be used to predict climate change impacts on groundwater-dependent streams. They have, for example, been used to estimate groundwater discharge variations under glacial loss (e.g., Somers et al., 2019) and soil freezing dynamics (e.g., Evans et al., 2018; Liu et al., 2021).

Here, we present a study of the impact of climate change on an alpine

headwater catchment. Alpine groundwater will play an increasingly important role in sustaining streamflow in perennial alpine streams. By combining ERT measurements, meteorological data, streamflow measurements, FE numerical modelling, climate models and statistical analysis, we examine the impacts of climate change on the annual hydrological dynamics.

2. Study site and field measurements

2.1. The Tsalet catchment

The Tsalet catchment (Valais, Switzerland) is a nival-regime alpine headwater catchment with no glacial cover (Fig. 1). This roughly cone-shaped catchment has a surface area of $\sim 1 \text{ km}^2$ and receives a mean precipitation of $\sim 1100 \text{ mm/year}$. Soils are mostly thin ($\leq 1 \text{ m}$) or absent and the terrain has an average elevation gradient of 70 % from the catchment outlet in the northwest to the south-eastern edge. Elevation ranges from 2260 to 2870 m a.s.l. (above sea level). Because of its high elevation the area has undergone several periods of glaciation and the bedrock (primarily greenschists, gneiss, quartzites, and breccia) is mostly overlain by quaternary sediments (Cochand et al., 2019). The moraine cover can be differentiated between older, more compact moraines dating from throughout the Würm glaciation and younger ones dating from the last glacial maximum (LGM). Talus and colluvium-dominated mixed deposits also account for an important portion of the sedimentary cover in the upper part of the catchment. Arnoux et al. (2020a) established a conceptual model of groundwater dynamics within the catchment by combining various methods, of which the time-lapse gravimetry data underwent subsequent analysis (Halloran, 2022). However, neither of these works developed numerical groundwater models nor did they investigate climate change impacts. Periodic springs are found at the boundary between the relatively lower hydraulic conductivity moraine and the relatively higher conductivity Talus aquifers and, near the catchment outlet, perennial springs ensure continuous streamflow in the Tsalet stream. Several studies describe in greater detail the Tsalet and the wider Réchy catchment in which the Tsalet is located (Arnoux et al., 2020a; Cochand et al., 2019; Halloran, 2022; Mari et al., 2013).

2.2. ERT measurements and interpretation

ERT is an ideal geophysical method for the Tsalet catchment as the predominant lithologies (talus, moraine and quartzite) have contrasting electrical properties. Talus (or scree) generally has been found to have extremely high resistivities of $>20'000 \text{ } \Omega\text{m}$ due to the presence of large voids and poor contact between individual rock fragments, while quartzite bedrock was observed to have resistivities ranging from $5'000\text{--}20'000 \text{ } \Omega\text{m}$ (McClymont et al., 2010). Moraines can have electrical resistivities ranging from $<1000 \text{ } \Omega\text{m}$ to $4000 \text{ } \Omega\text{m}$ depending on water saturation and other factors (Langston et al., 2011).

Approximately 1.7 km of ERT profiles (Fig. 1) were measured during a field campaign (06–08/09/2021). To optimize coverage, three ERT profiles with a nominal electrode spacing of 5 m were measured using an ABEM Terrameter LS2 Advanced 10/48 in roll-along mode with a multi-channel-optimised Wenner-Schlumberger protocol. Profile A (475 m, 96 electrodes) and profile C (350 m, 72 electrodes) were carried out roughly perpendicular to the slope in the middle and lower part of the catchment, respectively. Profile B (835 m, 168 electrodes) intersects profiles A and C and runs roughly parallel to the slope. The locations of these profiles were chosen as to include talus, both types of moraines and the colluvium/mixed debris aquifers. The bedrock for all three profiles is primarily quartzite, with gneiss present only in the lowest part of profile B. The position of all electrodes was measured to an absolute precision of $<2 \text{ cm}$ using a Leica GNSS surveying system. Inversion of the raw data, including 2-D topography information, to estimate true resistivity cross-sections was carried out using Res2dInv (Geotomo

Software, 2010) with a limit factor of 5 and manual elimination of spurious datapoints. The inverted data had a final combined RMS of $\sim 4\%$ and were graphically analysed using a *python* script (Halloran, 2018) before being manually interpreted using surface geology information (Swisstopo, 2022) and literature values of the above-mentioned lithology (Fig. 2).

The inverted apparent electrical resistivities (Fig. 2) ranged from $\sim 200 \Omega\text{m}$ to $\sim 40'000 \Omega\text{m}$. As expected, the talus aquifers showed the highest resistivities at far above $10'000 \Omega\text{m}$ and the bedrock below this talus aquifer had resistivities of $3000\text{--}7000 \Omega\text{m}$. The area corresponding to the colluvium-dominated mixed debris lithology showed fairly low resistivities which may indicate high clay content and/or high groundwater saturation. The moraine exhibited a two-layer resistivity profile with the top layer having resistivities of $5000\text{--}15,000 \Omega\text{m}$ and the lower layer displaying resistivities of $<3000 \Omega\text{m}$, likely indicating the water saturation boundary. As the resistivities obtained in this study are generally higher than those obtained by Langston et al. (2011), this suggests a lower clay content and correspondingly lower water content in the Tsalet moraines. Similar to the study of Langston et al., 2011, the low-resistivity mixed debris and saturated moraine layers complicate delineation of the bedrock boundary as preferential current flow through the shallow low-resistivity layers leads to a masking effect. However, in our case, this effect was less pronounced and it was thus possible to delineate the moraine-bedrock and mixed debris-bedrock interface. Using this knowledge, the different resistivity values were assigned corresponding lithologies to create geological cross sections showing the thicknesses of the different hydrogeological units (Figs. 2 &

S1) with vertical spatial accuracy on the order of a metre. Overall, interpretation of the ERT data shows that these Quaternary aquifers are relatively shallow with thicknesses of not surpassing $\sim 15\text{ m}$.

3. Numerical modelling

Here, we discuss the models that are developed and used in our study. Firstly, we develop a 3D geological model to provide information on the spatial orientation of hydrogeological units. We then develop a transient hydrogeological model based, in part, on this geological model. We model the transient recharge boundary condition for the hydrogeological model based on meteorological data and topography. The hydrogeological model is calibrated via history-matching of modelled and recorded streamflow at the catchment outlet. Finally, the calibrated model is executed using recharge boundary conditions, modelled in a similar way to those based on recorded meteorological data, based on data from ten established climate models.

3.1. Geological modelling

A 3D geological model was constructed using Geomodeller (Intrepid Geophysics, 2020). The stratigraphic pile was defined as, from oldest to youngest: bedrock, “Würm” moraine, “LGM” moraine, “mixed” debris (colluvium-dominated), and talus (scree) with onlap interfaces. For the topography a resampled 5 m-resolution digital elevation model (DEM) from Swisstopo was used. In areas with no subsurface data, it was assumed, based on the ERT data, that aquifer thicknesses did not exceed

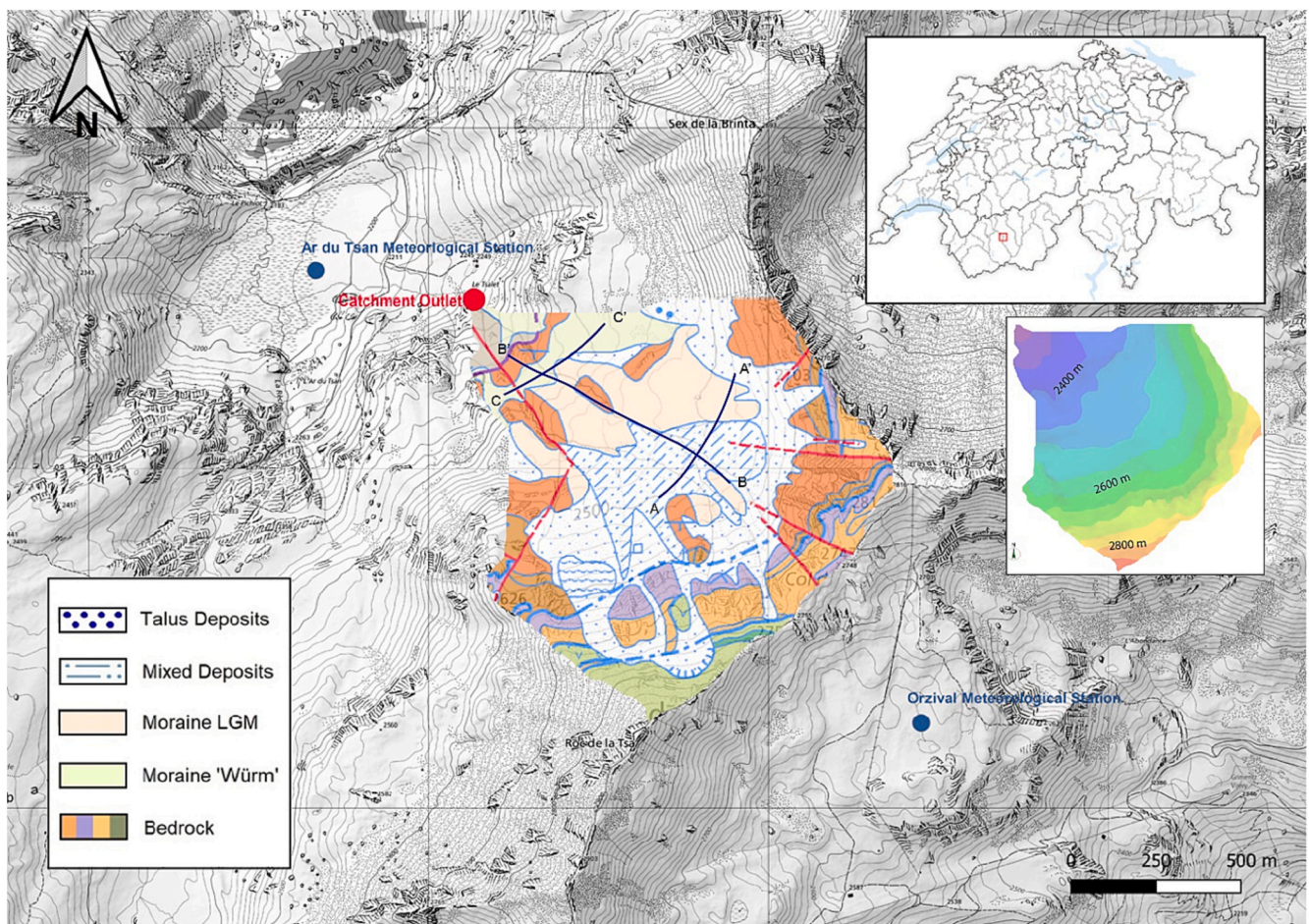


Fig. 1. The Tsalet catchment. Surface geology (Swisstopo, 2022) is shown along with the locations of ERT profiles (see Section 2.2). Locations of weather stations (see Section 3.2.2) and the catchment outlet are also indicated. Basemap from Swisstopo. Insets: Location of the Tsalet catchment in Switzerland and the elevation profile of the catchment (elevation bands of 50 m).

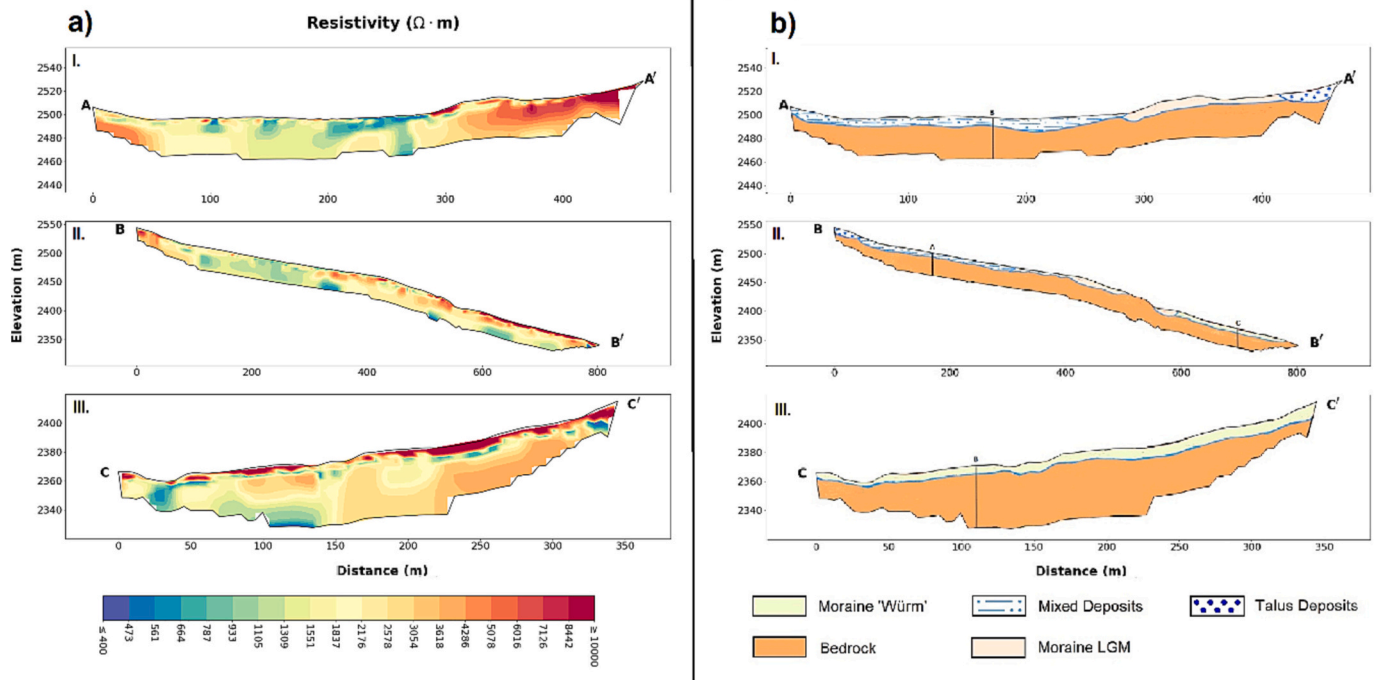


Fig. 2. a) Apparent resistivity (post-inversion) of the three ERT profiles. b) Interpretation of resistivity results and assignment of different lithologies. Locations of profiles appear in Fig. 1.

20 m. Older lithologies were assumed to dip at angles of 20–80°, which is corroborated by the ERT data, most notably for the talus aquifer. Due to the similarities in resistivities between the moraines and mixed debris layer, aquifer-aquifer interfaces could not be delineated with high certainty, although our interpretation is coherent with other studies (McClymont et al., 2012) that have also observed this type of morphology. To improve upon the surface geology data (Swisstopo, 2022), which has spatial precision on the order of several tens of metres, data was manually added to 85 well-distributed profiles ranging from tens to hundreds of metres. Mismatch of the resultant geological model and the input data (DEM, ERT interpretation, and surface geology) was found to be minor, with localised errors limited to <3 m. In order to assess the results of the geological model, multiple hypothetical geological cross sections were constructed from the geological model (Fig. S1). These cross sections included all lithologies and confirmed the geological coherence of the model. In general, aquifer thicknesses are highly correlated to the distance from any outcropping bedrock due to the steep bedrock walls surrounding the talus and to the erosional processes in the moraines. In cases where on the surface geology map different aquifers protrude into other aquifers it can be seen that they sometimes continue underground as expected. These could act as hydrogeological funnels or barriers depending on the difference of the hydraulic conductivity of the different materials.

3.2. Hydrogeological modelling

3.2.1. Subsurface model

A hydrogeological model was built in FEFLOW (DHI, 2021). Triangular mesh generation was constrained by the catchment geometry and a known fault with the minimum element angle set to 30°, resulting in 10,528 2-D elements with a mean side length of ~12 m (Fig. S2). The 3-D model was subsequently constructed by implementing a 5 m-resolution digital elevation model (DEM) and defining a fixed vertical element size of 2 m and depth of 20 m. A recharge-spreading layer (RSL) of 0.2 m thickness was added on top of the first layer (e.g., Beckers and Frind, 2000; Frind et al., 2014; Sousa et al., 2013). The final mesh of the 12-layer model contains 126,336 elements. FEFLOW was executed with

transient and unconfined pseudo-saturation settings and a minimum residual water depth of 1 m to ensure numerical stability (Diersch, 2014). We note that it is expected that the effective hydraulic conductivities obtained by history-matching when using pseudo-saturation approaches are generally overestimates as the complex nuances of unsaturated zone hydraulic conductivities are avoided. Elemental zones were assigned to each hydrogeological unit (Section 3.1, Fig. S3) and were each linked to hydraulic properties. The bottom boundary was set to no-flow and the lateral boundary conditions of the entirety of the model were defined as seepage-faces. Due to the topography and stratigraphy, it is believed that the vast majority of recharge leaves the catchment at its outlet, primarily as surface water (Arnoux et al., 2020a). Net recharge boundary conditions were applied to the top boundary (Section 3.2.2).

Before optimization via history-matching (inverse modelling) (Section 3.2.3), all initial porosities were set to 0.2 and initial hydraulic conductivity (K) values were set to 10^{-9} m/s (bedrock), 10^{-4} m/s (moraines), 5×10^{-4} m/s (mixed deposits), 5×10^{-3} m/s (talus), and 0.5 m/s (RSL), based on literature values (Clow et al., 2003; Kurylyk and Hayashi, 2017; Muir et al., 2011; Rogger et al., 2017). To set initial conditions, the model was first run in steady-state with uniform precipitation of 1,000 mm/year, then, to reach a pseudo-steady state, it was executed in transient mode with recharge defined as repetitions of the first year of the transient model several times. From this, the resultant hydraulic heads were exported into a copy of the FEFLOW model for the history-matching process (Section 3.2.3).

3.2.2. Recharge boundary condition modelling

In order to calculate the net (effective) transient model recharge various data from three different meteorological stations were used. The *Ar du Tsan* (temperature, water equivalent precipitation, mean pressure and snow height) and *Orzival* (temperature, snow height, relative humidity and mean pressure) meteorological stations (Fig. 1) are in close proximity to the catchment and are located at 2,192 and 2,630 m a.s.l., respectively. Because the *Orzival* station had several data gaps, data from a third station in *Evolène* (~9 km SSW of the catchment) were used to aid with gap-filling and also to provide solar irradiation and humidity

data. These data were used in the calculation of snow water equivalent (SWE), snowmelt, and effective evapotranspiration. The overlapping period of these meteorological stations spans 11/4/2014–31/10/2016 and thus model history-matching (“calibration”) was based on this period.

The local temperature gradient, $\nabla T(t)$, required for elevation-based scaling of meteorological data, was calculated based on daily mean temperatures from the *Ar du Tsan* and *Orzival* meteorological stations, with data from *Evolène* used for gap-filling. The daily and seasonal mean temperature gradients showed significant annual variations (Fig. S4) with the winter months experiencing a smaller gradient ($-4.5^\circ\text{C}/\text{km}$) compared to other seasons. Using this, daily mean-, maximum- and minimum temperatures can be calculated for any elevation value. Due to the lack of maximum and minimum temperature for the *Ar du Tsan* meteorological station, maximum and minimum temperatures at different elevations were calculated using the mean temperature gradient. Due to the large variation in elevation within the study area, temperatures between the outlet of the catchment and the highest point can differ by as little as 0°C and as much as 12°C depending on the day and the season. These differences are significant for SWE, snowmelt and effective evapotranspiration (ET). The catchment was divided into 50 m elevation bands (Fig. 1) and the mean elevation used in subsequent calculations. The majority of daily gradients (Fig. S4) lie within the range of dry- and wet-adiabatic temperature gradients ($-1.0^\circ\text{C}/100\text{ m}$ and $-0.6^\circ\text{C}/100\text{ m}$, respectively).

SWE was calculated using the precipitation time-series from the weather station at the *Ar du Tsan* and the elevation-scaled daily mean temperature time-series. To determine liquid and solid precipitation, a binary snow-rain model was assumed:

$$R_{\text{rain}}(z, t) = \begin{cases} P_{\text{liq}}(t), & \bar{T}_z(z, t) > 0 \\ 0, & \text{else} \end{cases} \quad (1)$$

$$SWE_{\text{acc}}(z, t) = \begin{cases} P_{\text{liq}}(t), & \bar{T}_z(z, t) \leq 0 \\ 0, & \text{else} \end{cases} \quad (2)$$

where R_{rain} is recharge from rain; P , recorded precipitation; $\bar{T}_z$, mean elevation-adjusted daily temperature; and SWE_{acc} , snow water equivalent accumulation.

To calculate daily potential snow water equivalent depletion, the linear degree-day method snowmelt model with a refreezing condition controlled by the daily minimum temperature was used (United States Department of Agriculture, 2004):

$$SWE_{\text{dep}}(z, t) = \begin{cases} 0, & T_{\text{zmin}}(z, t) \leq 0 \\ C_m \times \bar{T}_z(z, t), & \bar{T}_z(z, t) > 0 \end{cases} \quad (3)$$

where SWE_{dep} is potential snow water equivalent depletion; C_m , the degree-day coefficient; $\bar{T}_z$, mean elevation-adjusted daily temperature; and T_{zmin} , minimum elevation-adjusted daily temperature. SWE was calculated by adding snow water accumulation or subtracting potential snowmelt from the value for SWE from the day before:

$$SWE(z, t) = \begin{cases} SWE_{\text{acc}}(z, t), & (SWE(z, t-1) - SWE_{\text{dep}}(z, t)) \leq 0 \\ SWE(z, t-1) + SWE_{\text{acc}}(z, t) - SWE_{\text{dep}}(z, t), & \text{else} \end{cases} \quad (4)$$

The degree day coefficient C_m ($3.9\text{ mm}/^\circ\text{C}$) was calibrated using snow depth data from the *Orzival* meteorological station and the calculated SWE. The results of the SWE calculations were then validated against SWE calculations by Arnoux et al., 2021 (Fig. S5). Using the daily SWE data, recharge from snowmelt (R_{SM}) can be directly calculated by:

$$R_{\text{SM}}(z, t) = \begin{cases} SWE(z, t-1) - SWE(z, t), & SWE(z, t-1) - SWE(z, t) < 0 \\ 0, & \text{else} \end{cases} \quad (5)$$

Lastly, the potential recharge (R) without evapotranspiration can be

calculated:

$$R(z, t) = R_{\text{rain}}(z, t) + R_{\text{SM}}(z, t) \quad (6)$$

The results of SWE and recharge calculations for two elevations (Fig. S6) demonstrate the importance of elevation on daily recharge and recharge timing with the snowmelt period ending ~ 2 months earlier at the lower elevation. For the history-matching period, evapotranspiration (ET) was calculated using the Priestley-Taylor (P-T) formula (FAO, 1998) with the P-T α parameter set to 1 (Muir et al., 2011). The slope of the vapour pressure curve was calculated using the mean elevation-adjusted daily temperature in $^\circ\text{C}$ calculated using the daily temperature gradient data (Fig. S4) and the saturation vapour pressure in kPa:

$$\Delta(z, t) = \frac{4098 \times e_0(z, t)}{(\bar{T}_z(z, t) + 237.3)^2} \quad (7)$$

where e_0 is the saturation vapour pressure [kPa], which was, in turn, calculated by:

$$e_0(z, t) = 0.6108 \exp\left(\frac{17.27\bar{T}_z(z, t)}{\bar{T}_z(z, t) + 237.3}\right) \quad (8)$$

The psychrometric constant was calculated using daily pressure data from the *Ar du Tsan* meteorological station, linearly adjusting pressures by elevation and using the average pressure gradient (∇P) between 2,100 m and 2,700 m of elevation ($0.0095\text{ kPa}/\text{m}$):

$$\gamma(z, t) = (P_1(t) + (z_1 - z) \times \nabla P) \times 0.000665 \quad (9)$$

where P_1 and z_1 are the pressure and elevation at the *Ar du Tsan* meteorological station. Net radiation (R_n) was calculated by:

$$R_n(z, t) = (R_i(t) \times \alpha) - R_{nl}(z, t) \quad (10)$$

where α is albedo (0.25), R_i is the measured incoming short-wave radiation [$\text{MJm}^{-2}\text{d}^{-1}$] at the *Evolène* meteorological station and R_{nl} is the net long-wave radiation, calculated by the modified Stefan-Boltzmann equation (FAO, 1998) using relative humidity values from the *Evolène* station. Evapotranspiration throughout the period was calculated and integrated into a binary model for effective evapotranspiration (ET_{eff}) wherein evapotranspiration was assumed negligible under the presence of snow:

$$ET_{\text{eff}}(z, t) = \begin{cases} ET(z, t), & SWE(z, t) = 0 \\ 0, & \text{else} \end{cases} \quad (11)$$

Overall, effective evapotranspiration ranges from 5 mm/day during summer to 0 in periods with snow cover (Fig. S7). The differences in elevation cause evapotranspiration at the elevation of 2,275 to be generally around 0.2–0.4 mm/day higher during summer. During winter days without snow cover this difference is negligible. A more defining factor between the different elevations was the presence of snow. For higher elevations, periods with snow cover lasted longer and thus the overall yearly effective evaporation rates were lower.

Finally, daily elevation-dependent net recharge (Fig. 3), i.e., total potential recharge minus effective ET, was calculated by:

$$R_{\text{net}}(z, t) = R(z, t) - ET_{\text{eff}}(z, t) \quad (12)$$

Negative R_{net} values correspond to days where evapotranspiration exceeds recharge. For each 50 m elevation band, the corresponding transient net recharge timeseries was imported into FEFLOW and integrated as a boundary condition to the corresponding regions of the model (Section 3.2.1).

3.2.3. Optimisation and validation

To perform parameter optimisation, history-matching was carried out using PEST (Doherty, 2016) to minimise a weighted least-squares objective function (ϕ) based on observed and modelled discharge (Q) at the catchment outlet:

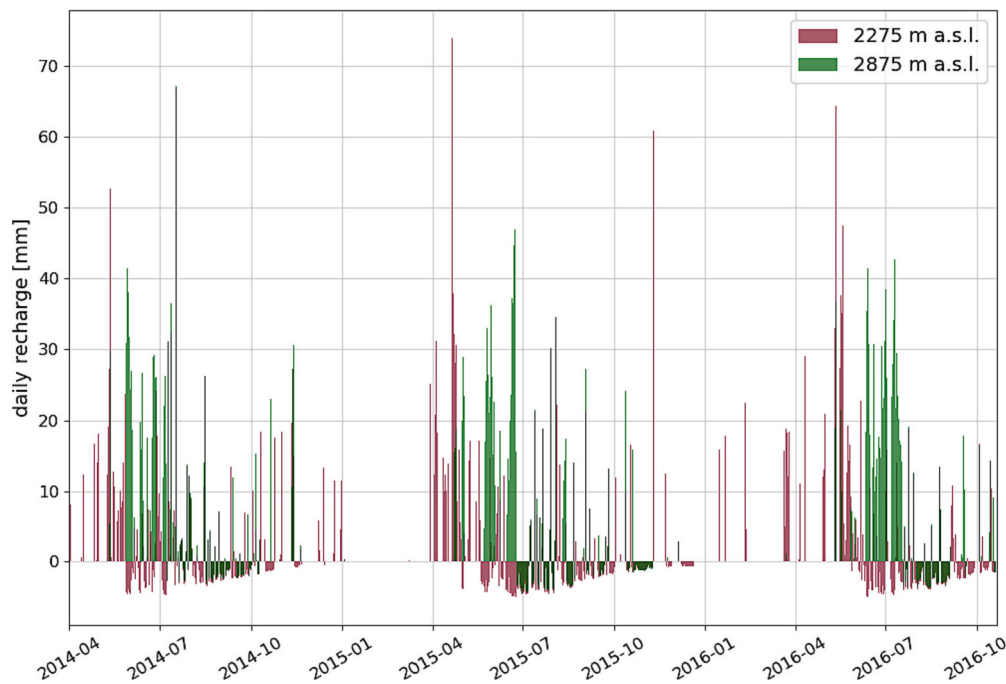


Fig. 3. Calculated net recharge (i.e., total recharge minus evapotranspiration) (Eq. (12)) for the history-matching period near the lower (2,275 m a.s.l.) and upper (2,875 m a.s.l.) limits of the catchment.

$$\phi = \sum_{t=0}^n (w(t)(Q_{obs}(t) - Q_{mod}(t)))^2 \quad (13)$$

where $w(t)$ are weights. Streamflow data (Fig. S8), described in detail in a previous study (Arnoux et al., 2020a), were converted to daily averages and used as calibration data (Q_{obs}). The first year of the time-series was used for initial condition set-up and calibration, while the second year was used for calibration and the third for validation. Hydraulic properties were modelled using effective hydraulic conductivities and porosities for each hydrogeological unit, resulting in twelve parameters (Table S1), following the principle of model parsimony. As the model has a time-step of 1 day and is set up to calculate long-term changes in seasonal groundwater discharge to the stream, weights for rapid surges in discharge due to rain-on-snow runoff events in spring were set to 0 for history-matching. History-matching model runs were also executed with a known fault (Fig. 1) explicitly modelled; however, the presence of this fault was found to have negligible effect on modelled discharge.

Parameters were optimised through history-matching (Table S1), which resulted in a 56 % reduction in the objective function. Overall, hydraulic conductivity (K) values coincide with literature values (Clow et al., 2003; Kurylyk and Hayashi, 2017; Muir et al., 2011; Rogger et al., 2017), with the lower elevation moraine having a higher K value than the upper, likely due in part to higher average saturation levels. Porosities (n) for the quaternary sediments showed no sensitivity to history-matching. Parameter sensitivities were also analysed temporally via the Jacobian matrix (Fig. S9) and the statistical performance was evaluated using RSR, NSE, and PBIAS (Table S2). Following well-established criteria (Van Liew et al., 2003; Moriasi et al., 2007; Santhi et al., 2001), a “satisfactory” rating was achieved for RSR (0.68) and NSE (0.55) and a “good” rating was achieved for PBIAS (19.3 %). Although rapid, short-term variations in discharge are underpredicted, overall, the modelled discharge matches the yearly dynamics and total discharge well (Fig. 4) given the available data.

3.3. Recharge boundary condition modelling based on climate models

In order to assess the impact of climate change on catchment and discharge thus the availability of water, transient recharge boundary conditions extending to the year 2100 were calculated. Temperature and precipitation time-series data (NCCS, 2018) consisted of daily gridded maps (2×2 km resolution), with each cell containing RCP-8.5 temperature or precipitation timeseries from 1982 to 2099. This emissions scenario was chosen in order to adequately test the limits of groundwater's ability to maintain perennial streamflow in alpine catchments that are currently nival-regime. Data for every available climate model with higher resolution was extracted, totalling 10 different climate models (EUR-11 ensemble): CLMCOM-CCLM4-ECEARTH, CLMCOM-CCLM4-HADGEM, CLMCOM-CCLM4-MPIESM, DMI-HIRHAM-ECEARTH, MPICSC-REMO1-MPIESM, MPICSC-REMO2-MPIESM, SMHI-RCA-ECEARTH, SMHI-RCA-HADGEM, SMHI-RCA-IPSL and SMHI-RCA-MPIESM. Data from these climate models with a reference elevation of 2,730 m were scaled to the catchment following a similar procedure to that used in the model calibration phase (Section 3.2.2), but with the simpler Hargreaves equation (FAO, 1998) used to calculate evapotranspiration due to the lack of relative humidity data. Elevation-dependent temperature was modelled using the same gradients as in the calibration phase of modelling (Fig. S4) and was integrated into the SWE calculations. Using elevation-adapted mean temperature, minimum temperature and precipitation timeseries, elevation-dependent net recharge is calculated.

During the measurement period, snowmelt events and the start of SWE accumulation occurred relatively late compared to climate data for the 2014–2020 period (Fig. 5a). The above-listed climate models predict that the maximum SWE for the 2080–2100 period will decrease by a factor of 2.5 for this elevation band and will reach its maximum ~1 month earlier compared to the early 21st century. Correspondingly, maximum mean monthly total recharge will decrease by 30 % and shift from May to April, although total recharge is predicted to change very little due to dryer summers being compensated by wetter (and warmer) winters (Fig. 5b).

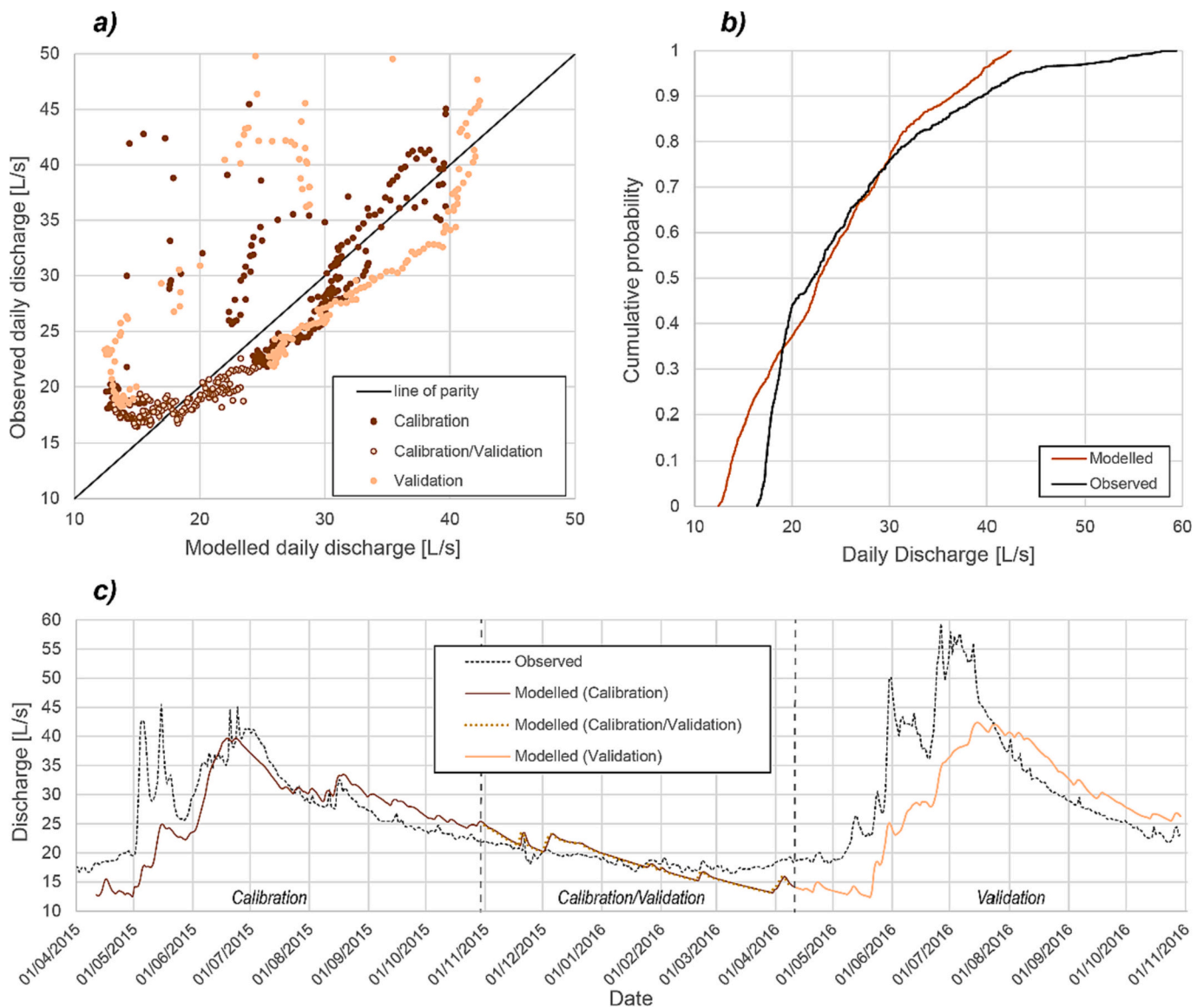


Fig. 4. (a) Observed vs. modelled outlet discharges differentiating between calibration- and validation periods. (b) Cumulative probability curve of outlet discharges modelled and observed of both the calibration and validation periods. (c) Observed vs. modelled time-series differentiating between calibration and validation periods.

Daily net recharge for each elevation band was calculated by subtracting effective ET from daily total potential recharge (Fig. 5c). As the length of the snowmelt accumulation period decreases, net recharge peaks shift 1–1.5 months earlier with ~50 % lower magnitudes. Maximum evapotranspiration timings differ from maximum recharge timings for both the study period and climate model monthly averages. Monthly net recharge rates decrease rapidly after snowmelt, particularly in the early 21st century. The mean yearly net recharge is expected to decrease by ~33 % from 674 mm/year to 450 mm/year by the end of the century. This change of net recharge is expected to directly affect discharge, albeit with complex subsurface transport affecting the temporal dynamics. The modified elevation dependent daily net recharge was imported into FEFLOW as the new recharge boundary conditions for each separate climate model and run from the end of the validation period to the year 2099.

4. Results and discussion

As outlined, the calibrated groundwater model (Section 3.2) was executed with ten RCP-8.5 climate scenarios (Section 3.3) using a daily

timestep. Discharge at the outlet of the model corresponds to streamflow in the Tsalet stream (Section 2.1), which is currently perennial. Here, we present and discuss the modelled climate change impacts to yearly streamflow dynamics and compare these results with the findings of other studies (Arnoux et al., 2021; Arnoux et al., 2020b; Evans et al., 2018; Somers et al., 2019).

4.1. Discharge time-series based on climate models

As expected, the resulting discharge rates for the different climate models (Fig. 6) display significant variations on short time-scales, but present a coherent long-term trend. Yearly mean discharge falls by ~0.6 % per year (~23 L/s to ~15 L/s over 60 years) between the periods 2020–2040 and 2080–2100. As these are multi-decadal averages and the residence time of groundwater in the quaternary aquifers is not expected to extend into the decadal range (Arnoux et al., 2020a), this decrease is assumed to be primarily due to the decrease in recharge. Mean daily discharge of all models shows a yearly periodicity with high values during and right after snowmelt and low values during late recession just before the snowmelt period. The lowest mean daily discharge of all

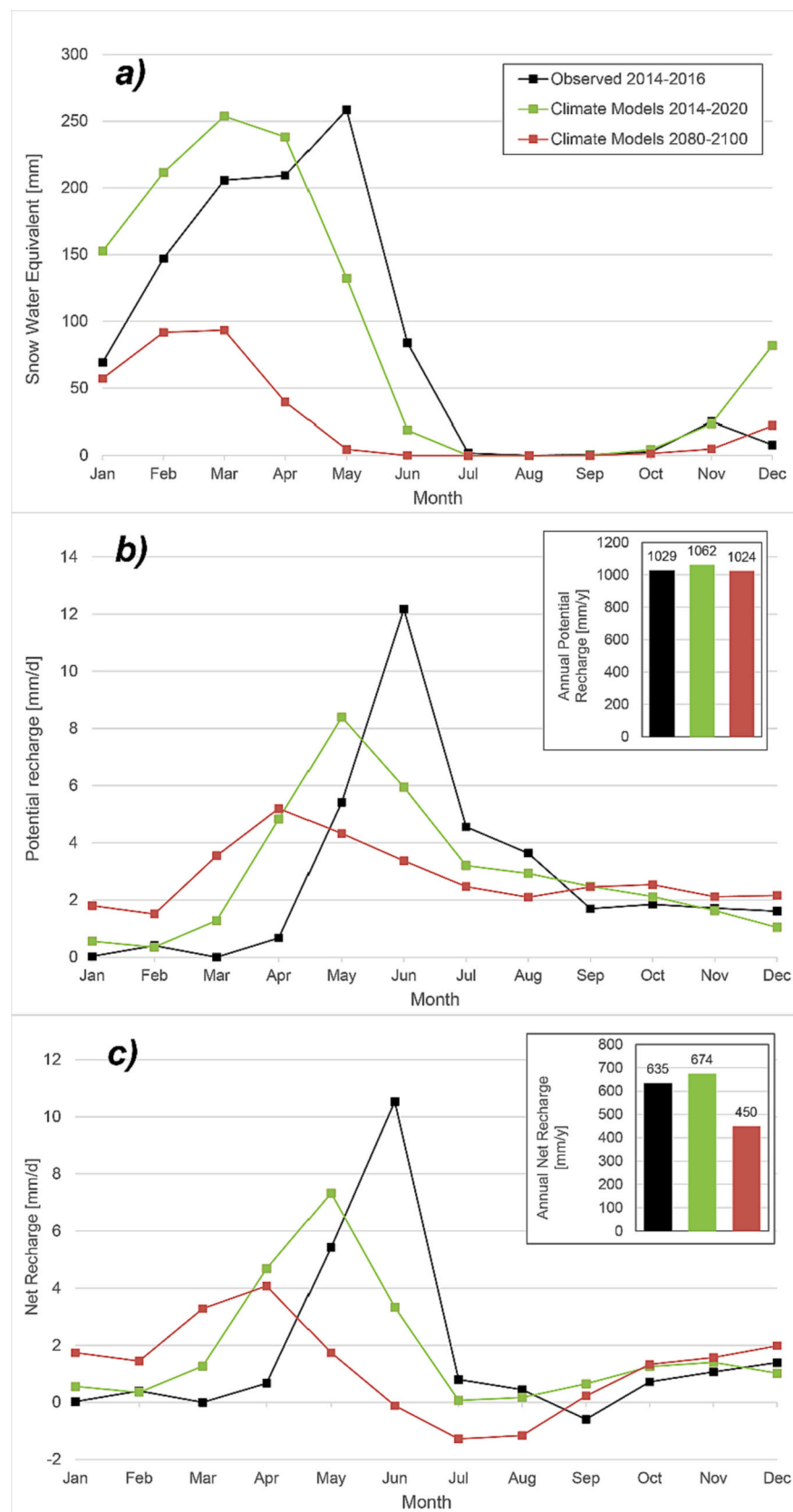


Fig. 5. SWE, potential recharge, and net recharge based on the measurement period (2014-2016) and the climate model ensembles (2014-2020 and 2080-2100). a) Mean monthly SWE at an elevation of 2,525 m calculated using the real available meteorological data and the average of the monthly mean of SWE at the same elevation for all climate models SWE calculations for the start and the end of this century. b) Mean monthly total potential recharge at an elevation of 2,525 m calculated using the real available meteorological data and using all climate models. b inset) Yearly mean totals. c) Mean monthly total recharge at an elevation of 2,525 m calculated using the real available meteorological data and using all climate models. c inset) Yearly mean totals.

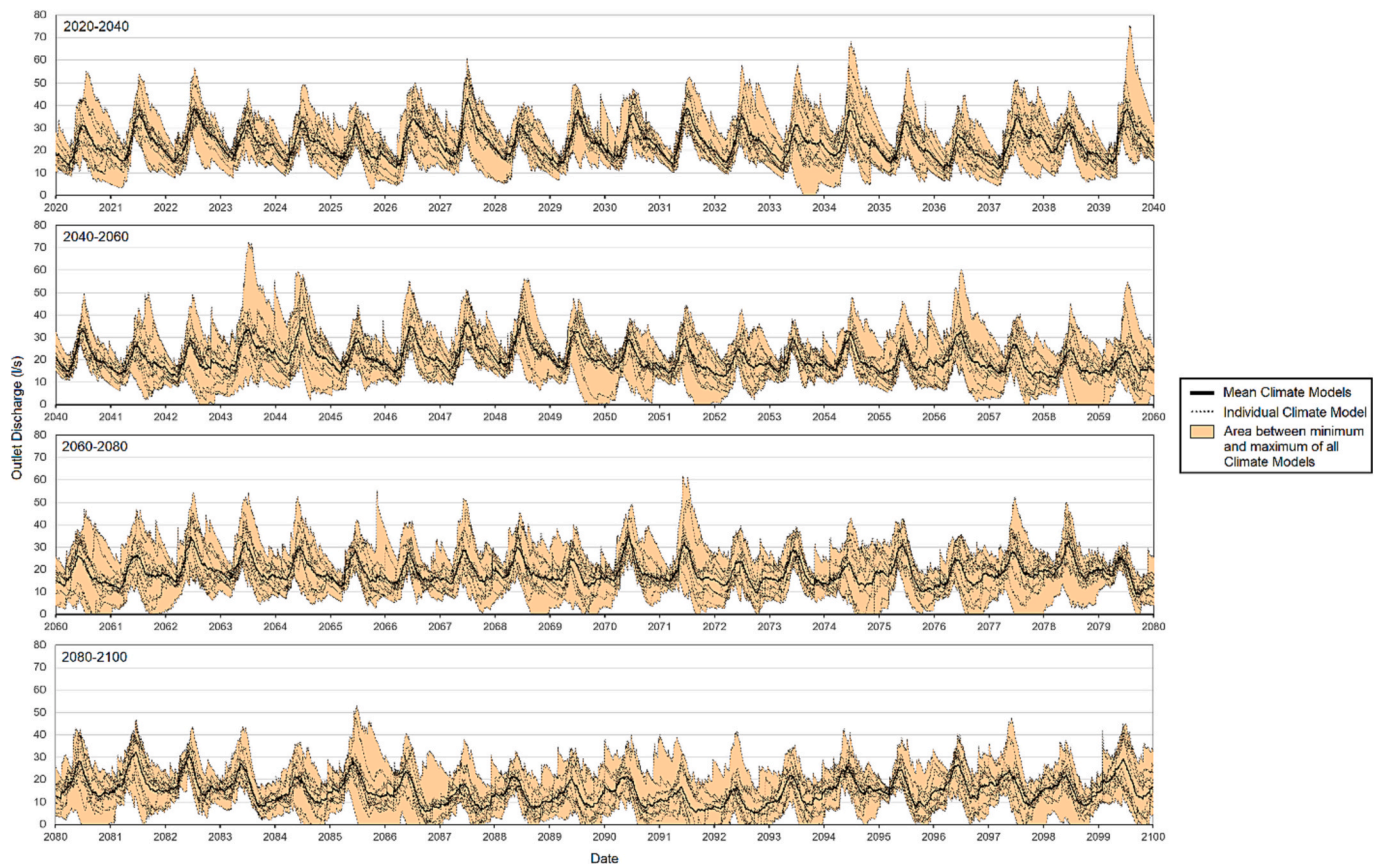


Fig. 6. Mean outlet discharge for all climate model recharge boundaries over the period of 2020–2099 with single outlet discharge rates for every separate climate model recharge boundary condition shown and highlighting the area/variation between maximum and minimum daily recharge for all climate models.

models was observed in late summer of 2092 with a mean minimum discharge rate of ~ 6 L/s. The greatest minimum discharge rate in the model ensemble for this specific year was ~ 17 L/s.

4.2. Increase in the occurrence of no-flow conditions

A notable phenomenon is the occurrence of zero-discharge periods (Fig. 6) wherein the currently perennial *Tsalet* stream ceases to flow. While there is no period in which the ensemble average is 0 L/s, the frequency with which individual models predict a dry stream increases from with time. Cumulative probability curves (Fig. 7a) clearly show the trends in discharge, including the above-discussed increased probability of zero flow conditions. Between 2020–2040 and 2080–2100 the fraction of time during which discharge is zero increases by a factor of 86 from 0.05 % to 4.30 %. While no-flow conditions are the extreme and can have disastrous impacts on riparian and wetland ecosystems, such as the wetland into which the *Tsalet* stream flows immediately downstream of the catchment (Arnoux et al., 2020a; Cochand et al., 2019), increased low-flow conditions can also impact ecological functions and human water supplies (Garssen et al., 2014; Politti et al., 2014). The fraction of time where the discharge rate is ≤ 10 L/s increases by a factor to 6.4, from 4.87 % to 30.98 %, between 2020–2040 and 2080–2100. The maximum outlet discharge value of decreases from 75 to 53 L/s over these same periods.

4.3. Maximum discharge rates and timings

Maximum outlet discharge values and timings aid in predicting the peak water availability from the catchment and determining changes in its expected timeframe. A clear trend is observed in the mean maximum discharges for each year (Fig. 7b,c). Overall, as global warming

progresses, maximum outlet discharge rates are predicted to decrease by 1.9 L/s per decade and to occur ~ 7 days earlier per decade. The results for individual models (Fig. 7c) illustrate the year-to-year variability in addition to the general trend of lower maximum flow rates that occur earlier in the year. Maximum discharge timings are likely to become more variable throughout time as snow accumulation and melt dynamics change significantly. The cumulative probability curves for the maximum catchment discharge rate (Fig. 8a) and its date of occurrence (Fig. 8b) show an increase from 0.33 % to 2.5 % in <10 L/s max flow rate and from 7 % to 23 % for <20 L/s when comparing the entire 2020–2280 period to the 2080–2100 period. These lower yearly discharge maxima will likely have impacts on sediment transport (Maruffi et al., 2022; Rainato et al., 2018) that may have implications for riverine and riparian ecosystems. For the maximum outlet discharge timing (Fig. 8b) during the period of 2080–2100, 24 % of the time maximum outlet discharge occurred before May, while for the period 2020–2080 this occurred only 7 % of the time. Similarly, the maximum quintile date (Fig. 8b), i.e., when 80 % of cases have already experienced their maximum annual discharge, shifts from July 4th (2020–2080) to June 13th (2080–2100).

4.4. Non-uniform impacts on seasonal streamflow and groundwater export

Climate change impacts to yearly streamflow rates are not uniform throughout the year. The largest reduction in average streamflow occurs in the late summer/early autumn period (Fig. 9) when snow cover is absent and dependence on groundwater discharge to ensure streamflow is high. Over a period of 60 years (2020–2040 to 2080–2100), the maximum mean monthly outlet discharge decreases by 31 % (from 32.5 to 22.5 L/s) and shifts from late June to early May. Over this same

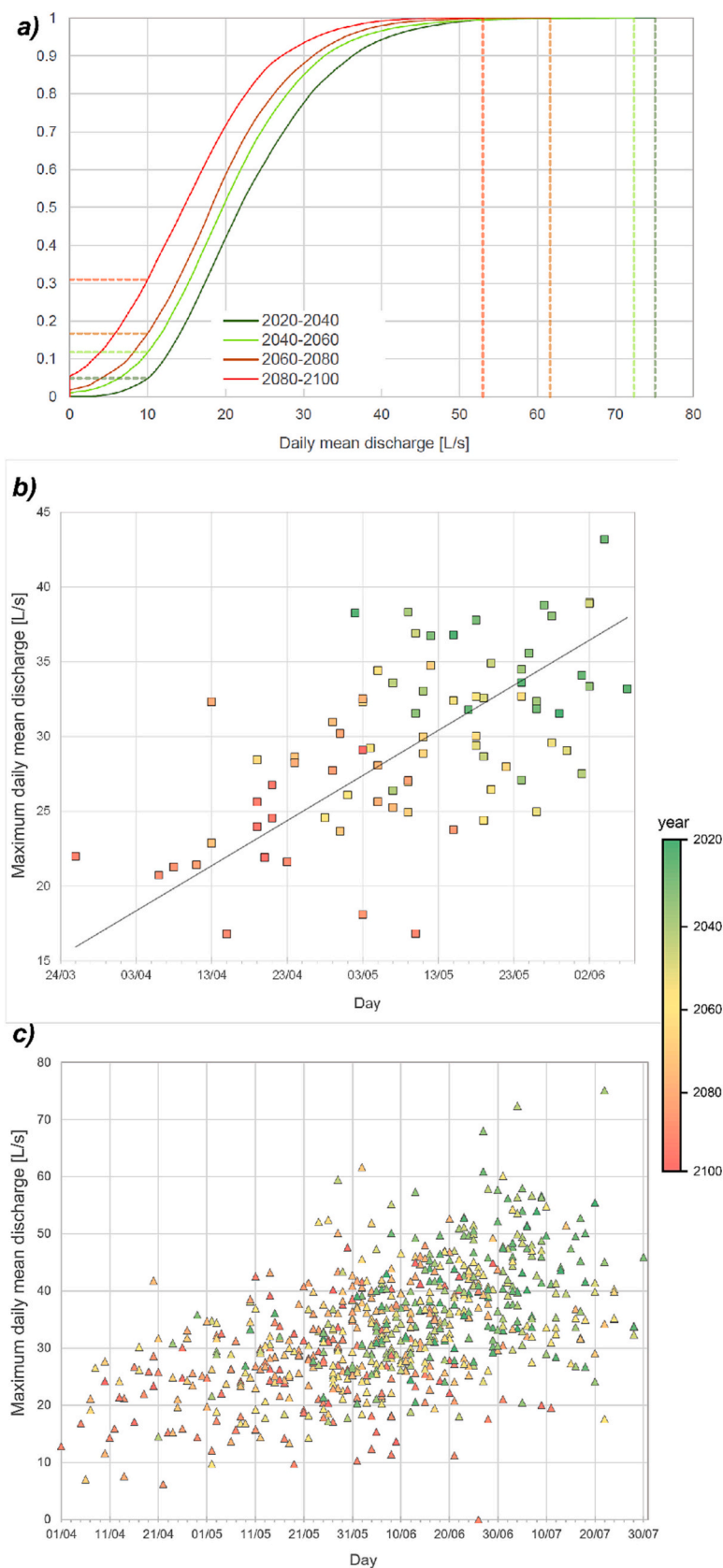


Fig. 7. a) Cumulative probability curves for catchment discharge (see Fig. 6) for 4 twenty-year periods. The horizontal dashed lines indicate the cumulative probability of a flow rate of 10 L/s or less, while the vertical dashed lines indicate the maximum values. b) Yearly evolution of mean yearly maximum outlet discharge values and timings as calculated from the model ensemble. The line of best fit (linear least-squares) is also shown. c) Yearly evolution of yearly maximum outlet discharge values and timings for each model run with a different climate model recharge boundary condition.

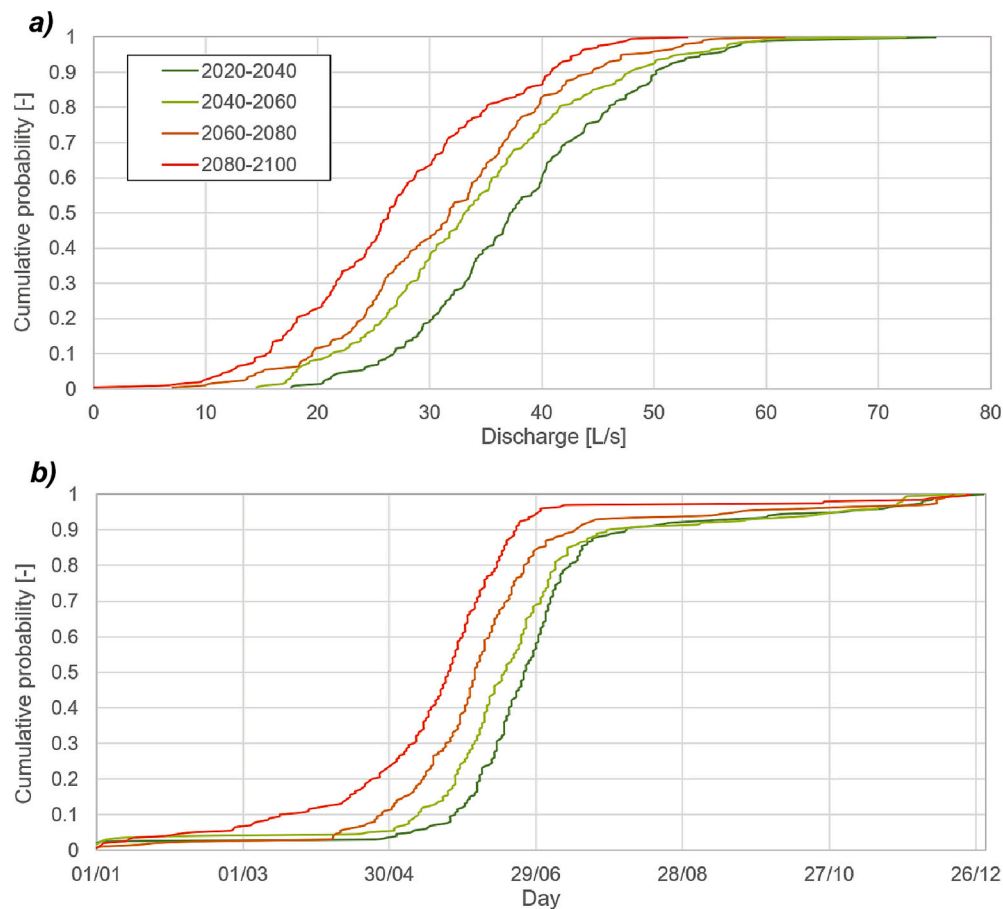


Fig. 8. a) Maximum outlet discharge cumulative probability curve for different periods using all climate model boundary conditions. b) Maximum discharge date cumulative probability curve for different periods using all climate model boundary conditions.

period, the maximum monthly net recharge as calculated mid-catchment (Fig. 5c) also shifted from mid-May to early April and decreased by 45 %. While in both cases the absolute shift between the two periods of ~1.5 months was observed, the delay between peak net recharge in the mid-catchment vs peak catchment discharge is slightly greater than 1 month. This, coupled with the different forms of the yearly recharge and discharge curves illustrates the vital and complex buffering role that alpine aquifers play. Even more striking is the shift in the timing of the minimum mean monthly discharge (Fig. 9a), which shifts from late winter (March) to late summer or early autumn (September/October). In addition to increased evaporation rates and decreased precipitation during summer, higher temperatures during winter cause liquid precipitation instead of snow accumulation in lower elevations which causes outlet discharges to start increasing again during winter. The largest decreases of mean monthly outlet discharge occur in late summer (58 %). This is especially troublesome as summer droughts are predicted to increase in both severity and duration (NCCS, 2018). In many regions of the world, including central Europe, during droughts down-stream water users have primarily relied on water resources from alpine regions to guarantee water availability. While hydroelectric dams may provide a partial solution to these challenges in some regions, their futures under climate change are also uncertain (Gaudard et al., 2014; Schaeffli et al., 2007).

Finally, the model was also used to investigate groundwater export from the catchment – that is, water exiting the catchment in the subsurface (e.g., Schaller and Fan, 2009). Water exits the catchment through the bottom moraine aquifer and via the bedrock itself, although this subsurface export is an order of magnitude smaller than the direct discharge to the stream at the catchment outlet. Yearly dynamics in

groundwater export (Fig. S10) are similar to those of discharge (Fig. 9) although during low-flow periods (later summer/early autumn), discharge rates for the outlet are predicted to decrease by 58 %, while groundwater export is predicted to only decrease by 28 %. The relative importance of groundwater export and the role of lower permeability units in controlling seasonal and long-term groundwater discharge and export dynamics can thus be expected to increase in the future.

4.5. Limitations and comparison with other studies on alpine groundwater

Modelling of subsurface hydrological processes is doubly challenging in alpine catchments due to subsurface heterogeneity, lack of direct groundwater measurements, large gradients, and dynamic snow-melt processes. Our hydrogeological model satisfies established calibration criteria (Table S2) and, importantly, the analysis we perform with it is based on the same data type we use to calibrate it (streamflow at the catchment outlet). Nonetheless, there remain several unavoidable uncertainties in the modelling approach. As is the case with all subsurface models, the delineation of boundaries between hydrological units is inexact, although the ERT measurements allow us to define the depths of the moraine and colluvium/mixed deposit aquifers, which likely play the most critical role in maintaining streamflow. We note that bedrock heterogeneity, which could not be integrated in our model given the available data, may play an important role in many alpine hydrogeological systems and merits further research. The modelled recharge is based on established practices, but is limited by the number of meteorological stations in the study area. Given the limitations in data availability, the relatively low level of parameterisation in the hydrogeological model is appropriate. We also note that while future

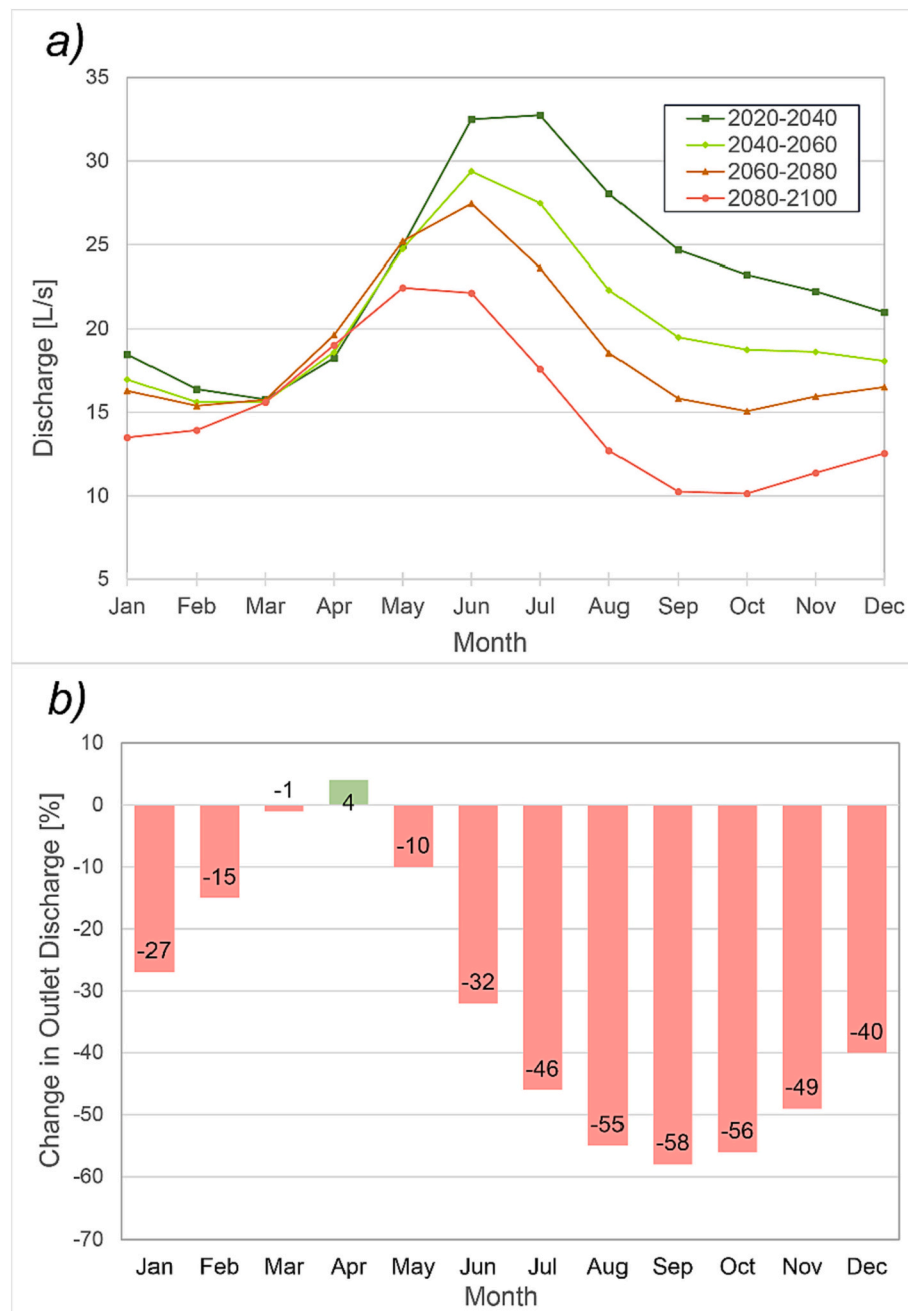


Fig. 9. a) Mean monthly outlet discharge rates for different periods using the monthly mean of the climate model ensemble. b) The change in mean monthly outlet discharge between the periods of 2020–2040 and 2080–2100.

land use/land cover (LULC) changes are not considered, the *Tsalet* catchment is in a nature reserve and is nearly entirely above the treeline, we would not expect this to be relevant over the investigated timescale. In other alpine catchments, however, this may not be the case and LULC changes may be an important consideration.

Our results explicitly demonstrate that while groundwater will help dampen the impacts of decreased snow accumulation, it does have its limits. Similar projections to those presented here have also been made for other catchments in the European Alps (Arnoux et al., 2021), Rockies (Evans et al., 2018), and Andes (Somers et al., 2019). Additionally, multiple studies have demonstrated the current importance of groundwater in sustaining flow in alpine streams (e.g., Cowie et al., 2017; Liu et al., 2004; Yao et al., 2021). Given the findings of this study and others, the relative contribution of groundwater to alpine streamflow globally can reasonably be expected to increase under climate change. Arnoux

et al., 2020b used a physical process-based model with two RCP 8.5 scenarios to predict that the mean yearly discharge in the *Réchy* catchment (of which the *Tsalet* is a subcatchment) will decrease by ~16 % by the end of the century. They also found that minimum discharge rates will shift to late August or September with a total decrease of 38 % to 45 % during August. For other similar alpine catchments in their study, the minimum value also shifted from just before snowmelt to sometime in late summer. Arnoux et al., 2021 presented a conceptual hydrological model under three RCP 8.5 scenarios for several Swiss alpine catchments and found that by the end of the 21st century yearly catchment discharge would decrease between 20 % and 30 % (~20 % for *Réchy*). These results confirm the modelling analysis presented here for the steeper *Tsalet* catchment (70 % mean slope vs. 28 % for *Réchy*) in which we have included detailed delineation of hydrogeological units, as well as finer resolution mesh and net recharge/ET boundary conditions.

5. Conclusions

We performed a detailed study of the impacts of climate change on a typical nival-regime alpine catchment with significant moraine and talus cover using an ERT-informed 3-D hydrogeological model coupled with a climate model-driven recharge model. The results clearly illustrate several key points relevant for alpine streams facing climate change impacts:

- The annual dynamics of alpine streamflow will change. Average flow will decrease, while the timing of maximum discharge will shift earlier in the year and the duration of low-flow periods will increase.
- Groundwater will play an increasingly important role in ensuring flow in alpine streams. As snow-free periods increase in duration and droughts become more common reliance on alpine groundwater will only increase, but this resource has limits.
- Currently perennial nival-regime alpine streams are at risk of becoming intermittent over the course of the current century.

While our study focused on the *Tsalet* catchment, these general observations are applicable to many catchments and streams in the European Alps and in other mountain ranges where anthropogenic climate forcings are expected to have similar impacts on temperature and snow accumulation dynamics (e.g., Beniston and Stoffel, 2014; Blahušáková et al., 2020; Fang and Pomeroy, 2020). While much recent attention on alpine water resources has been focused on conspicuous loss of glacial cover (Castellazzi et al., 2019; Gao et al., 2021; Müller et al., 2022; Zemp et al., 2015), it is important that policy makers and scientists do not neglect the hidden impacts of climate change on alpine groundwater. Indeed, in many alpine catchments, and in many current and future climate scenarios, groundwater may be the single most important water resource. Further research is undoubtedly needed to understand the limits of the buffering capacity of alpine groundwater. Frameworks for evaluating the resilience or vulnerability of alpine water resources absolutely must include groundwater as a key component of the hydrological system. A comprehensive, process-based understanding of alpine groundwater and its role in alpine hydro-systems is essential for environmental policy, water resources management, and climate change adaptation.

CRedit authorship contribution statement

Landon J.S. Halloran: Conceptualization, Methodology, Validation, Formal analysis, Investigation, Resources, Writing – original draft, Writing – review & editing, Supervision, Project administration, Funding acquisition. **Jeremy Millwater:** Methodology, Software, Validation, Formal analysis, Investigation, Data curation, Writing – original draft, Writing – review & editing, Visualization. **Daniel Hunkeler:** Validation, Resources. **Marie Arnoux:** Validation, Resources, Data curation, Writing – review & editing.

Declaration of competing interest

The authors declare no competing interests.

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.scitotenv.2023.166009>.

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