

1 **Linking baseflow separation and groundwater storage dynamics in an alpine basin**
2 **(Dammagletscher, Switzerland)**

3
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20

21 **Abstract**

22 This study aims at understanding interactions between stream and aquifer in a glacierized
23 alpine catchment. We specifically focused on a glacier forefield, for which continuous
24 measurements of stream water electrical conductivity, discharge and depth to the water table
25 were available over four consecutive years. Based on this dataset, we developed a two-
26 component mixing model in which the groundwater component was modelled using measured
27 groundwater levels. The aquifer actively contributing to stream flow was assumed to be
28 constituted of two linear storage units. Calibrating the model against measured total discharge
29 yielded reliable sub-hourly estimates of discharge and insights into groundwater storage
30 properties. Our conceptual model suggests that a near-surface aquifer with high hydraulic
31 conductivity overlies a larger reservoir with longer response time.

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35 *Key words:*

36 Mountain hydrology, electrical conductivity, mixing models, linear groundwater reservoir,
37 stream-aquifer interactions

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1. Introduction

Groundwater storage dynamics in alpine catchments are difficult to determine, but could influence the response of mountain hydrology to climate change. A better understanding of stream-aquifer interactions is therefore necessary to predict hydrological flow patterns in the future. Alpine sites put additional constraints on data acquisition, because snow cover, weather conditions, and/or rough terrain limit the available measurements.

In this study, we estimate groundwater storage dynamics in the alpine headwater catchment fed by the Damma glacier in central Switzerland. In previous studies, we focused on local properties of the groundwater flow in specific stream reaches (Magnusson et al. 2014; Kobierska, 2014). The aim is now to use this specific knowledge to upscale our hydrogeological understanding to the whole glacier forefield. We seek to estimate the contribution of groundwater and hyporheic exchange to stream flow during different periods of the year, as well as the volume and response times of groundwater storage.

The topic of contributing storage to stream flow has been covered by many studies. Analytical and numerical formulations of the Boussinesq equation (e.g., Brutsaert and Nieber, 1977; Rupp and Selker, 2006; Rupp et al., 2009) and linear or nonlinear reservoirs (e.g., Wittenberg and Sivapalan, 1999; Hannah and Gurnell, 2001; Majone et al., 2010) have been explored. At our site, traditional recession analysis is challenged by the fact that discharge is dominated by the diurnal dynamics of snow and glacier melt. Pure recession events are therefore very rare.

In alpine sites, mixing models based on natural tracers are a typical avenue for hydrograph separation (i.e. Hinton and Schiff, 1994; Liu et al., 2004; Covino and McGlynn, 2007; Blaen et al., 2013). Dzikowski and Jobard (2011) used electrical conductivity (EC) data to estimate the groundwater contribution to the discharge of an alpine stream. They defined seasonal ranges in the relationship between EC and streamflow rather than predicting groundwater flow and total flow for individual time steps. On the other hand, Covino and McGlynn (2007) presented groundwater table data but did not use them in their mixing model.

We suggest here a different approach to using mixing models with streamwater EC data, which involves a time-varying groundwater input. We implemented a two-component mixing model (glacier melt and groundwater) in which the groundwater exfiltration component is the output of two linear groundwater reservoirs. One reservoir provides a baseflow component. The second reservoir models additional groundwater using five groundwater (GW) stage measurements throughout the forefield. In the following, we refer to infiltration as the flow from the stream into the aquifer (i.e., aquifer recharge) and to exfiltration as the flow of groundwater and hyporheic exchanges back into the stream (i.e., aquifer discharge).

To verify the robustness of the model and to understand the influence of each data input taken separately (EC or GW stage data), we compared our calibrated model to two partial models, each of which held one measured input variable (streamwater electrical conductivity or groundwater level) constant. By further analyzing groundwater interactions (infiltration and exfiltration) with stream water, we: (1) verify that groundwater exfiltration estimates are realistic, (2) provide an upper limit to the volume of the active groundwater reservoir and (3) conclude with a conceptual representation of the forefield's main hydrogeologic features.

89 2. Study site and experimental methods

90 2.1. Site description

91

92 The Damma glacier forefield (Fig. 1) is part of a small (10.7 km²) granitic catchment situated
93 in the central Swiss Alps. It is currently being studied as part of the SoilTrEC project
94 (Bernasconi et al., 2011). The glacier covers 40% of the catchment and has been retreating
95 since the end of the Little Ice Age (LIA). Due to a sharp change in slope gradient, a small
96 piece of the glacier has become detached from the main glacier during its retreat and is
97 referred to as the 'dead ice body'. Large lateral moraines date from approximately 1850 (the
98 end of the LIA) and two terminal moraine bands dating from 1927 and 1992 mark the end of
99 two short periods of re-advance. The elevation of the catchment ranges from 1800 to 3600 m
100 a.s.l. and the entire catchment is covered by snow for approximately six months per year.

101

102 The glacier forefield itself ranges from 1800 to 2000 m a.s.l. and covers an area of
103 approximately 0.5 km². The average annual temperature between November 2008 and
104 November 2012 was 2.2 °C at our automatic weather station (AWS) in the forefield (see
105 Fig. 1). In 2008, annual precipitation and evapotranspiration for the whole catchment were
106 estimated at 2300 mm and 70 mm respectively (Kormann, 2009). With a yearly cumulative
107 discharge of approximately 2700 mm, the water balance of the catchment is clearly positive
108 and corresponds to an average glacier mass loss of about one meter depth per year.

109

110 The basin is characterized by heavy snowfall in winter, making discharge difficult or
111 impossible to measure. Discharge becomes dominated by baseflow as snow and glacier melt
112 gradually cease in late autumn. In late spring (typically end of May), snowmelt leads to a
113 strong increase in discharge and a clear daily cycle is quickly established. In autumn, daily
114 cycles of glacier melt are interrupted by rain events and the recession of a slow-draining
115 aquifer becomes noticeable as melt rates decrease.

116

117 The forefield is encompassed by two steep lateral moraines (Fig. 1). The area in the vicinity
118 of S3 is composed of a relatively impermeable silty surface layer, which leads to surface
119 runoff during storms, as evidenced by scouring of the surface (see Fig. 2 in Kobierska et al.,
120 2014). The area between S5 and S0, where the topography suddenly steepens, is rich in
121 springs, which display seemingly constant flows (in the order of 10 L/s per spring) during the
122 summer season.

123

124 Magnusson et al. (2014) studied four groundwater transects (named S1, S3, S5 and S6 in
125 Fig. 1). Each transect was equipped with three pressure transducers: one in the stream and two
126 in piezometric tubes placed on a line perpendicular to the stream. Taking S1 as an example,
127 we adopted the following notation: S1_{stream} for the stream stage measurement, S1_{near} for the
128 piezometer that is closer to the stream, and S1_{far} for the piezometer farther away from the
129 stream. Note that S0 consists of one single piezometer located approximately 50 m from the
130 main stream channel (Fig. 1). Due to difficult field conditions, the piezometers could only be
131 installed to a maximum depth of 1.5 m.

132

133 The water table is driven by stronger gradients along the stream than towards it. This results
134 in strong advection in the direction of stream flow, as shown in Kobierska et al. (2014). The
135 mean gradient between S1 and S7 is 13.5% over a distance of 840 m. Between S0 and S5, the
136 steepest section of the forefield has gradients over 20% for approximately 150 m. Near-stream
137 lateral groundwater gradients are primarily influenced by diurnal stream stage fluctuations,
138 rather than by topography-driven longitudinal gradients (Magnusson et al, 2014).

139

140 This paper focuses on the dynamics of the active groundwater storage, which is the part of the
141 aquifer that can exfiltrate into the stream before it reaches the gauging station. Refraction
142 seismics and electrical resistivity surveys were carried out on four transects of the forefield
143 (Kobierska, 2014). These geophysical studies suggest that the saturated glacial till does not
144 contain permafrost areas over the whole forefield. The sediment layer should also be at least
145 10 m thick in much of the forefield, including the vicinity of the discharge station. This means
146 that an important part of the aquifer in the forefield is ‘non-contributing’, meaning that not all
147 water flowing out of the catchment is measured at the discharge station. The lack of
148 permafrost means that changes in groundwater levels reflect changes in groundwater volumes
149 (rather than a change in the lower boundary due to permafrost melting).

150

151 2.2. Hydrometeorological data

152

153 Groundwater levels were measured with Hobo U20 Water Level Loggers (5-min sampling
154 interval averaged to 30-min values) at S1, S3, S5 and S6 as shown in Fig. 1. The method is
155 described in detail in Magnusson et al. (2014). Stream stage was measured at the catchment
156 outlet (S7 in Fig. 1), using both a cable-supported radar device and a pressure logger installed
157 in a partly perforated tube. The rating curve of discharge as a function of stream level was
158 calibrated with the results of salt and dye tracer dilution tests across a wide range of flows
159 (35 L/s to 4500 L/s, see Magnusson et al. 2012 for further details). According to the
160 manufacturer’s specification, the loggers have 0.14 cm resolution and 0.3 cm accuracy. The
161 absolute pressure readings were adjusted for atmospheric pressure variations (measured at site
162 S7) also using a Hobo U20 pressure sensor.

163

164 Table 1 presents values of the main hydro-meteorological parameters for successive winters
165 and summers (taken from start of June to end of October), as measured by the discharge
166 station and the meteorological station (S7 and AWS in Fig. 1). This highlights the succession
167 of hydro-climatically different years, which presented a good opportunity to test the
168 robustness of the model.

169

170 For example, Table 1 shows large year-to-year variability in snow water equivalent (SWE)
171 and annual rainfall, and also shows that neither water source strongly dominates the water
172 balance. Snow water equivalent (SWE) was estimated from the maximum snow depth of each
173 winter, assuming a density of 0.3. Snow depth was measured at the AWS with a Campbell
174 Scientific SR50 ultrasonic sensor. Cumulated rainfall is calculated from rainfall measured at
175 the AWS. Note that both SWE and rainfall data were measured at the AWS in the forefield
176 and are thus not representative of the water input to the whole catchment which extends 1800
177 m above the forefield. Cumulated discharge also contains a significant ice melt component,
178 which was not estimated in this study.

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180 2.3. Electrical conductivity endmembers

181

182 Streamwater EC and temperature were measured at the main runoff station (S7) with a WTW
183 Tetracon 325 sensor (accuracy 0.5% for EC, 0.5 °C for temperature under 15 °C). The 10-min
184 sampling rate was averaged to 30-min values for this study. Various measurements of
185 groundwater springs were also carried out throughout the forefield with a hand-held WTW
186 Cond 315i device (same accuracy) in order to determine endmember values for use in the
187 mixing model. Continuous EC measurements of groundwater and streamwater are also
188 available for summer 2011 at three transects (S1, S3 and S5) and at some springs between S5
189 and S0. EC was temperature-corrected using a non-linear correction to a reference of 20 °C.

190

191 From those measurements, an electrical conductivity map of the forefield can be sketched
192 (Fig. 1) with 6 geographically distinct areas (all displayed in Fig. 6). Zones L1, L2, H1, H2
193 and H3 serve as a visual representation of low and high EC zones based on 238 single EC
194 measurements (Table 2) and previous work by Tresch (2007) at this site. Only the endmember
195 EC values impact our model and not the extent of those zones. Naturally, the ruggedness of
196 the field site did not allow measuring groundwater and glacier melt electrical conductivities
197 everywhere in the forefield.

198
199 Between 2009 and 2012, EC measured at the main runoff station (S7) varied from 2 to 13.3
200 $\mu\text{S}/\text{cm}$ with an average value of 6.6 $\mu\text{S}/\text{cm}$. The main section of the stream through the
201 forefield is fed by two glacial sub-catchments of low EC (areas L1 and L2, lower end only).
202 Direct measurements of glacier melt on the dead ice body yielded EC values ranging from 1.7
203 to 2.1 $\mu\text{S}/\text{cm}$. We use the lowest EC value measured for melting ice (1.7 $\mu\text{S}/\text{cm}$) as the
204 endmember value for glacier melt. EC can be assumed to be a conservative tracer in open-
205 channel flow because of the short travel time of surface runoff through the forefield (on the
206 order of 10 minutes). This is confirmed by the low EC values (minimum of 2 $\mu\text{S}/\text{cm}$)
207 measured at the discharge station during extreme flow events.

208
209 Three distinct zones are rich in springs (areas H1, H2 and H3) and consistently present
210 conductivities between 13 and 18 $\mu\text{S}/\text{cm}$ (Table 2). Those groundwater exfiltration zones
211 average 15.1 $\mu\text{S}/\text{cm}$ and show very little temporal variability, as witnessed by continuous
212 data-logger measurements for summer 2011 in the upper part of H1. We can therefore
213 confidently attribute an endmember value of approximately 15.1 $\mu\text{S}/\text{cm}$ to groundwater
214 exfiltration in the forefield.

215 3. Models

216 3.1. Two-component mixing model

217

218
219 In the previous section, we found that EC displayed two distinct endmembers: groundwater at
220 15.1 $\mu\text{S}/\text{cm}$, and glacier melt at 1.7 $\mu\text{S}/\text{cm}$. As stream water EC was consistently anti-
221 correlated to runoff, we considered using mixing models to study the relationship between EC
222 and discharge at the basin scale.

223

224 Our modelling approach requires a set of specific assumptions:

- 225 1. The EC measured at the main discharge station is the result of pure mixing between
226 glacier melt and groundwater exfiltration into the stream
- 227 2. Glacier melt has a constant EC of $EC_{gl} = 1.7 \mu\text{S}/\text{cm}$ (the lowest EC value measured for
228 melting ice on the dead ice body)
- 229 3. Exfiltrating groundwater has a constant EC of $EC_{gw} = 15.1 \mu\text{S}/\text{cm}$ (average of all
230 groundwater measurements)

231

232 The first assumption of pure two-component mixing is violated when rain falls. Several rain
233 events affected both discharge and EC signals during the study period. Because quantifying
234 rainfall throughout the forefield and its impact on stream water EC was not the aim of this
235 study, we excluded all periods when more than two millimeters of cumulated rain had fallen
236 in the last five hours. This filter was designed to exclude the direct increase in surface runoff
237 associated with rainfall events, but not the subsequent exfiltration of rainwater that had
238 infiltrated the aquifer. The filter threshold of 2 mm per 5 hours is similar to typical melt rates,
239 and led to removing 10.8% of the data. The deleted time periods can be seen as gaps in the EC
240 data (upper panel of Fig. 3).

241

242 The second assumption is best met in midsummer when melt water runoff is dominated by
 243 glacier melt. The model does not differentiate snowmelt from glacier melt, as the same low
 244 endmember value EC_{gw} is used.

245
 246 Finally the third assumption is justified by continuous EC measurements at several
 247 groundwater springs, which have shown that EC is reasonably constant in time (previous
 248 section).

249 In summary, the three assumptions outlined above lead to the following equations:

$$250 \quad Q(t) = Q_{gw}(t) + Q_{gl}(t) \quad (1)$$

$$251 \quad Q(t) * EC(t) = Q_{gw}(t) * EC_{gw} + Q_{gl}(t) * EC_{gl} \quad (2)$$

252 where $Q(t)$ is total discharge at time t, EC_{gw} and EC_{gl} are respectively the groundwater and
 253 glacier electrical conductivity endmember values, and $Q_{gw}(t)$ is the groundwater exfiltration
 254 flow whose modelling will be presented in the next section. Q_{gl} is not modelled explicitly, but
 255 instead is estimated by end-member mixing analysis. Mathematically, Q_{gl} is eliminated when
 256 we combine Eqs. (1) and (2) to form Eq. (3):

$$257 \quad Q(t) = \frac{(EC_{gw} - EC_{gl}) * Q_{gw}(t)}{EC(t) - EC_{gl}} \quad (3)$$

259

260 3.2. Groundwater exfiltration model

261 3.2.1. Preliminary simulation considerations

262

263 Our preliminary simulations considered groundwater exfiltration as the output of a nonlinear
 264 storage model using two parameters (Wittenberg and Silvapalan, 1999). They were difficult to
 265 optimize due to equifinality problems because multiple parameter combinations led to similar
 266 calibration results, and thus no clear optimum could be found.

267 Other problems arose because the piezometers had to be rather short due to the difficulties of
 268 installation in this environment. Thus most of the piezometers dried up while the stream was
 269 still flowing. A realistic groundwater model for this catchment should therefore account for
 270 slow drainage of the aquifer in winter. In addition, as previously discussed in Magnusson et
 271 al. (2014), the piezometers provided important information on the daily near-surface
 272 interactions with the stream.

273 To avoid equifinality problems and account for both baseflow and shallower groundwater
 274 exchanges, we decided to introduce both a ‘slow’ and a ‘fast’ linear reservoir that could be
 275 calibrated separately. Accordingly, groundwater exfiltration is the sum of each reservoir’s
 276 output, which is a linear function of storage volume as in Eq. (4):

$$277 \quad Q_{gw}(t) = \frac{V_{slow}(t)}{T_{slow}} + \frac{V_{fast}(t)}{T_{fast}} \quad (4)$$

278 where the proportionality factor T is the response time constant of the reservoir (T_{slow} or T_{fast})
 279 and $V(t)$ is the current storage volume in each reservoir (V_{slow} or V_{fast}).

280

281 At the end of winter, the piezometers are empty and snowmelt initially fills the “slow
 282 reservoir”. Our conceptual model considers that the ‘fast’ reservoir only starts filling when the
 283 ‘slow’ reservoir is full. The ‘slow’ reservoir then remains full by receiving a constant inflow
 284 from the ‘fast’ reservoir or the lateral moraines, and in turn providing an equal exfiltration

285 flow (denoted $baseflow_{max}$ in the next section). How the two reservoirs precisely interact is not
286 modelled. At the end of the season, when the ‘fast’ reservoir is empty, the ‘slow’ reservoir no
287 longer receives flow inputs and its storage starts to decrease.

288

289 3.2.2. Slow linear reservoir

290

291 We calibrated T_{slow} with a recession event at the end of 2008, which was the only pure
292 recession event lasting more than two weeks with reasonable discharge amplitude. In all other
293 years, continuous measurements ended too early due to disturbances from snow loads and
294 icing in the river channel. In 2008, the autumn was marked by an early big snow storm after
295 which snow cover persisted into winter. Snow covered the whole forefield but had no effect
296 on the stream geometry, such that the subsequent stage measurements were not affected by
297 snow loads. Pure recession was established because the thick snow cover was efficient in
298 stopping glacier and snow melt, even during some short warm periods that followed. Based
299 on Eq. (4), the recession hydrograph can be fitted using Eq. (5):

$$300 \quad Q_{recession}(t) = Q_{meas}(t_{end}) \cdot \exp\left(\frac{(t_{end} - t)}{T_{slow}}\right) \quad (5)$$

301 Where $Q_{meas}(t_{end})$ is the measured discharge at the end of the recession event and $Q_{recession}(t)$ is
302 the modelled discharge at any time before the end of the measured event (t_{end}). This method
303 has the advantage of not requiring an exact knowledge of when the recession event started.

304

305 The fit between measured and modeled discharge in Fig. 2 is very good from November 15th
306 to the end of the record. Before this date, analysis of meteorological data suggests that melt
307 inputs contributed to streamflow in addition to groundwater exfiltration. Records shows that
308 substantial snowfall occurred between October 28th and October 31st, bringing snow depth at
309 the meteorological station from 0 to 113 cm. Between the last peak discharge (November 5th)
310 and mid-November, a rain-on-snow event occurred, which prevented total discharge from
311 representing only baseflow. The entire catchment remained covered by snow and on the 11th,
312 as the air temperature sharply dropped below 0°C, the snowpack froze and water percolation
313 through the snowpack stopped. Soil moisture in the upper soil layers subsequently dropped
314 and, from November 15th to the end of the record, the observed flow should represent pure
315 recession from the ‘slow’ reservoir.

316

317 Our conceptual model presented in the previous section considers that streamflow on
318 November 15th 2008 is only constituted of groundwater flow from the ‘slow’ reservoir. At this
319 date, the ‘slow’ reservoir is full and its discharge is 0.07 m³/s (Fig. 2, lower panel).
320 This fixed value denoted $baseflow_{max}$ will represent the contribution of the ‘slow’ reservoir to
321 streamflow for all subsequent periods during which the ‘fast’ reservoir is not empty.

322

323

324 3.2.3. Groundwater level in the fast reservoir

325

326 A total of nine groundwater level sensors and four stream stage sensors were installed in the
327 forefield and could be used to compute a groundwater storage function. In order to represent a
328 balanced spatial average of GW levels in the forefield, we used data from the far piezometer
329 of each transect (S1_{far}, S3_{far}, S5_{far}, S6_{far} and S0). From mid-October onwards, most
330 piezometers were empty except S6_{far} which some years provided stage data until December.
331 For this reason and because there were other periods during which data from some
332 piezometers were missing, we computed a reservoir function every year as an integral of

333 mean stage variations. For each time step, the integral water level in the reservoir $L_{integral}$ was
 334 implemented as follows:

$$335 \begin{cases} L_{integral}(t) = L_{integral}(t - \Delta t) + \sum_{i=1}^n \frac{L_i(t) - L_i(t - \Delta t)}{n} \\ L_{integral}(t_{end}) = L_{residual} \end{cases} \quad (6)$$

336 where the second term on the right of the main equation is the mean variation in groundwater
 337 level between $t - \Delta t$ and t , using all available piezometers (a total of 'n'). This methodology
 338 limits measurement noise and creates a continuous storage function as long as one piezometer
 339 is available. Without the second equation, the computed reservoir would however only offer a
 340 relative value of storage. To correct for this, we assumed that the reservoir drains at the end of
 341 each season (t_{end} ; end of October in this case) to a residual water storage volume $L_{residual}$
 342 which was adjusted for each year. Note that $L_{integral}(t)$ represents the height of saturated material
 343 in the 'fast' reservoir. For this reason, the drainable porosity is introduced in the next section.

344

345 3.2.4. Total groundwater flow

346

347 In our model setup, total groundwater flow is the sum of exfiltration from both the slow and
 348 fast reservoirs. The slow reservoir is always full when the fast reservoir is not empty, that is,
 349 during the main part of the hydrological season (start of June to end of October). It displays a
 350 constant exfiltration rate, denoted $baseflow_{max}$, estimated in section 3.2.2. During this period,
 351 the total groundwater exfiltration flow is obtained by adding the output of both reservoirs
 352 using Eq. (4):

$$353 Q_{gw}(t) = baseflow_{max} + \frac{A_{fast} \times L_{integral}(t) \times \varphi}{T_{fast}} \quad (7)$$

354 where T_{fast} is the time constant of the fast reservoir, A_{fast} is its area and φ is the drainable
 355 porosity. When the fast reservoir is empty (autumn, winter and beginning of spring),
 356 groundwater exfiltration follows Eq. (4).

357

358 The model proposed in this study is obtained by integrating Eq. (7) into Eq. (3) via the
 359 groundwater component Q_{gw} . In the rest of the manuscript, this will be referred to as the
 360 'FULL' model as it uses both electrical conductivity and groundwater data. The complete
 361 modeling framework is schematically summarized in Fig. 3.

362

363 3.3. Model calibration and performance assessment

364 3.3.1. Calibrating against total discharge

365

366 The 'FULL' model and two alternative models (named partial models thereafter) each using
 367 only one type of field measurement (either EC or GW) were calibrated against measured
 368 discharge. The first partial model, denoted P_{EC} , used Eq. (3) with a calibrated constant
 369 groundwater exfiltration rate (Q_{gw}). Weijs et al. (2013) used this model to calibrate a rating
 370 curve using EC rather than stream stage. The second partial model, denoted P_{GW} , had a
 371 variable groundwater inflow as per Eq. (6) but used a constant value for EC (yearly average).
 372 The aim was to determine whether both electrical conductivity and groundwater data used by
 373 the 'FULL' model improved its modelling performance.

374

375 The models were calibrated for each full hydrological year (four years from 2009 to 2012)
 376 and validated with the three remaining years. Calibration started at the beginning of June and
 377 stopped when EC became unavailable, usually mid-October. Relative error was used as a

378 performance measure for calibration. In addition, the Nash-Sutcliffe efficiency (*NSE*) and
 379 benchmark efficiency (*BE*) were evaluated based on Eq. (8):

$$380 \quad Efficiency = 1 - \frac{\sum_t (Q_{meas}(t) - Q_{mod}(t))^2}{\sum_t (Q_{meas}(t) - Q_{bench}(t))^2} \quad (8)$$

381 where Q_{meas} is measured discharge; Q_{mod} is modelled discharge and Q_{bench} is either runoff
 382 predicted by a benchmark model (to compute the *BE*) or by the average of the measured data
 383 (to compute the *NSE*). Our benchmark model uses the discharge value recorded exactly
 384 24 hours earlier, which is a rather stringent test as the signal displays daily fluctuations for
 385 much of the hydrological season. Due to the high-amplitude seasonal discharge record, the
 386 average measured discharge poorly describes the catchment hydrology. For this reason, *BE*
 387 provides a better assessment of model performance than *NSE*, which is bound to be high.

388 3.3.2. Mass balance verification

389 Our model so far has not taken into account the infiltration of surface water into the aquifer.
 390 Neglecting evapo-transpiration and infiltration from the lateral moraines, the difference
 391 between surface infiltration and groundwater exfiltration represents the change in
 392 groundwater storage in the forefield at every time step. The mass balance equation can be
 393 written to express the instantaneous infiltration rate $Q_{inf}(t)$ as follows:
 394
 395

$$396 \quad Q_{inf}(t) = \frac{dV(t)}{dt} + Q_{gw}(t) \quad (9)$$

397 Our calibration procedure only allowed optimizing the fraction A_{fast} / T_{fast} without considering
 398 the mass balance of the aquifer. Eq. (9) shows that infiltration is dependent on $dV(t)/dt$ so that
 399 rapid variations in modelled groundwater storage could lead to negative and thus unrealistic
 400 infiltration. With the constraint that Q_{inf} may not become negative, Eq. (9) can provide an
 401 upper limit to the total volume of the ‘fast’ reservoir because it is directly related to extreme
 402 negative values of $dV(t)/dt$.

403 4. Results

404 4.1. Model calibration against total discharge

405 The cross-validation results of the four years of data are presented in Table 3. Both partial
 406 models (P_{EC} and P_{GW}) were tested, and displayed worse performance in all cases. This finding
 407 reveals that including both electrical conductivity and groundwater level data benefited the
 408 ‘FULL’ model. Of the two data sources used in the ‘FULL’ model, EC provides better
 409 information for modelling discharge, as model P_{GW} performed much worse than model P_{EC} .
 410
 411

412 The ‘FULL’ model's optimal parameter set for the hydrological season of 2011, however, led
 413 to significantly worse validation results than the other years. This particular year was
 414 characterized by a warm autumn with very late snowfalls. To compensate for year to year
 415 variability in residual water content in the fast reservoir at the end of October, we performed
 416 some adjustments to $L_{residual}$. Table 4 presents the improved validation performance of the
 417 ‘FULL’ model with the addition of a residual water content term. Our model presents high
 418 and reliable performance, which indicates that the main assumptions are coherent with the
 419 physical processes involved. The optimal parameter T_{fast} was 6.5 hours.
 420
 421

422 Fig. 4 shows the model results for 2009. Daily variations in total discharge are appropriately
 423 reproduced, although with some underestimation during most of the early summer (zoom 1).
 424 Discharge recessions following two cold snaps around June 20th and July 10th are however
 425 accurately modeled. The modelling results significantly improve from the beginning of
 426 August onwards, as non-glaciered slopes have become snow free. Zooms 1 and 3 in Fig. 4
 427 focus on periods of underestimation, whereas zoom 2 illustrates slight peakflow
 428 overestimation during intense melt periods. In Section 5.4 we suggest that those deficiencies
 429 are caused by seasonal variations in the “glacier melt” EC endmember (EC_{gl}).

430

431 Those results were obtained with a total volume of the ‘fast’ reservoir based on an area A_{fast} of
 432 1000 m by 100 m (approximate length and width of the forefield). This seems a reasonable
 433 value as infiltration remains positive throughout the season except in very few instances at the
 434 end of the record (Fig. 4). Zooms 1, 2 and 3, show daily cycles in which periods of infiltration
 435 and exfiltration dominance (during day and night, respectively) alternate with one another.
 436 The proposed size of the ‘fast’ reservoir is at the upper limit for a realistic (non-negative)
 437 infiltration. This will be further considered in the discussion.

438

439 Fig. 5 shows the estimated percentage of groundwater exfiltration as a function of total
 440 measured discharge for 2009. Only the best modelled time steps are displayed (less than 10%
 441 absolute error in total discharge). The season starts with medium flow and a high groundwater
 442 contribution (snowmelt-dominated in June), then progresses to high flows with a very low
 443 groundwater contribution (glacier-melt-dominated in August). The end of the season
 444 (September) is characterized by low flows and an increasing groundwater contribution. Those
 445 qualitative results suggest that the model is appropriately describing exfiltration processes.

446

447

448 4.2. Verifying A_{slow} with spring recharge

449

450 The total volume of the ‘slow’ reservoir can be estimated with Eq. (4), using the optimal
 451 parameter T_{slow} ($T_{slow} = 29$ days) and the $baseflow_{max}$ value of $0.07 \text{ m}^3/\text{s}$. For 1000 m of length
 452 and 400 m of width, this yields a maximum depth of 1.73 m. The surface of the aquifer was
 453 assumed based on topographical data (see Fig. 1) and perceptual understanding of the
 454 forefield. Porosity was set to 0.25, the average of all sites mentioned in Smittenberg et
 455 al. (2011).

456

457 The aim of Fig. 6 is to illustrate the recharge of the slow reservoir during spring snow melt.
 458 Using Eq. (4) to relate the volume of water in the reservoir ($V_{slow}(t)$) and its exfiltration rate
 459 ($Q_{gw}(t)$), and adding a recharge term $R(t)$, the storage function can be expressed as follows:

$$460 \quad V_{slow}(t) = V_{slow}(t - \Delta t) + \left(R(t) - \frac{V_{slow}(t - \Delta t)}{T_{slow}} \right) \times \Delta t \quad (10)$$

461 where Δt is the time step (30 minutes in our case).

462 After a 150-day period with little or no recharge (November to end of March), the slow
 463 reservoir would come out of winter with only 10 cm storage remaining. In Fig. 6, the recharge
 464 of the reservoir was simulated using Eq. (10) with a recharge rate of 100 L/s during every
 465 snowmelt period. This rate is equivalent to complete infiltration of 22 mm/day of snowmelt in
 466 the forefield. Even though $S6_{far}$ fills at an early date, for the following interpretation, we retain
 467 the conceptual view that the ‘fast’ reservoir starts filling once the ‘slow’ reservoir is full.

468

469 The main feature of Fig. 6 is the successive appearance of permanent water in the different
470 piezometers (plotted GW levels are the depth of water in each piezometer). S6_{far} is located at
471 the lower end of the forefield and is quickly filled by permanent water. S1_{far}, on the other
472 hand, displays daily peaks for approximately three weeks before water permanently rises on
473 May 10th. We suggest that snowmelt regularly fills the piezometers but infiltrates deep into
474 the aquifer through an unsaturated zone because the ‘slow’ reservoir is not yet full. The ‘slow’
475 reservoir depth is about 1.3 m when S1_{far} permanently fills, whereas its maximum depth was
476 earlier estimated at 1.73m. This is reasonable because S1_{far} is not quite at the highest point of
477 the forefield, and the reservoir may still keep filling under the ‘dead ice body’. This result
478 suggests that if the recharge rate was well estimated, then the reservoir volume too was
479 correctly estimated, providing an independent method to verify the T_{slow} parameter derived
480 from the 2008 recession analysis.

481

482 5. Discussion

483 5.1. Constraining the fast reservoir volume

484

485 Previous sections presented satisfying total discharge modeling results. Complementary
486 verifications also suggested that both groundwater exfiltration rates and the total volume of
487 the ‘slow’ reservoir had been well estimated. Finally, infiltration analysis provided an upper
488 limit on the possible volume of the ‘fast’ reservoir, under the constraint that infiltration may
489 not become negative.

490 This limit is however directly dependent on the choices made to compute the integrated
491 groundwater level $L_{integral}$. It is possible that $L_{integral}$ displayed excessively large daily
492 fluctuations requiring a small ‘fast’ reservoir for infiltration to remain positive. Magnusson et
493 al. (2014) showed that the damping of daily stream stage fluctuations into the aquifer is a
494 significant process influencing groundwater storage. We used the piezometers that were
495 farthest away from the stream for the computation of the reservoir function. However, those
496 piezometers may have been too close to the stream to accurately describe the average storage
497 fluctuations of the aquifer.

498

499 We suggest that the absolute depth of this ‘fast’ reservoir is on the order of one meter for the
500 following reasons: (i) the maximum value attained by $L_{integral}$ is 0.9 meters (i.e., the
501 piezometers are on average approximately one meter deep), (ii) most of them are nearly
502 empty by the end of the season (end of October) when the ‘fast’ reservoir has depleted. Based
503 on this depth, the simulations in Fig. 4 were carried out with a ‘fast’ reservoir area (A_{fast}) of
504 1000 by 100 meters. This corresponds to the length of the forefield by twice the distance from
505 the stream to S0, and is also roughly the average width of the braided river system over the
506 forefield (slightly smaller than the green zone in Fig. 7).

507

508 Based on those geometrical aspects, we suggest that the ‘fast’ aquifer is characterized by high
509 hydraulic conductivities, spans the riparian and hyporheic zone of the braided stream network
510 and is on the order of one meter deep.

511

512 5.2. Conceptual hydrogeological model of the forefield

513

514 The aim of this section is to propose a conceptual overview of the site’s hydrogeology, based
515 on modelling insights and previous results. This is illustrated by Fig. 7.

516

517 The modelling chain presented in this study yielded robust simulation of total discharge as a
518 function of groundwater levels in the forefield and stream EC at the discharge station. The
519 model then enabled the estimation of an active groundwater reservoir in the forefield. Based

520 on the initial hypothesis of a combination of two linear reservoirs, we found that the deeper
521 reservoir empties slowly and has a volume equivalent to the area of the forefield (1000 by 400
522 meters) with a depth of 1.7 m if porosity is assumed constant at 0.25. A shallower aquifer fills
523 on top of the base aquifer during summer and responds rapidly to daily fluctuations in stream
524 stage.

525

526 Geophysical campaigns have however shown that depth to bedrock is likely to be at least
527 10 m in most of the forefield (Kobierska, 2014). We can therefore expect part of the saturated
528 sediment volume to act as a non-contributing aquifer, flowing below the discharge station.
529 How much this hidden groundwater flow component affects the yearly water balance would
530 be difficult to assess as total sediment depth and its hydraulic properties are technically
531 challenging to measure. Note that at the beginning of spring snowmelt recharge in 2011 (Fig.
532 6), modelling shows that the slow 'active' reservoir had not completely emptied over the
533 winter before recharge by snowmelt started.

534

535 5.3. Limitations and uncertainties

536

537 One key problem with the use of mixing models in such an environment is the limited range
538 of variation in EC. Also, the recorded values are at the lower end of what can be measured by
539 typical instrumentation. However, the use of four years of data defined by strong and
540 consistent daily fluctuations allowed for interesting findings. Brown (2002) highlights that
541 mixing models are not as well adapted to glacial environments as previously thought. In our
542 case, however, the length and high temporal resolution of the time series make the technique
543 worth testing.

544

545 Considering hydrology in the forefield as the mixing of only two water sources is clearly a
546 simplification. We can list a total of four components: snowmelt, glacier melt, groundwater
547 exfiltration, and rainwater. Rainfall is hard to quantify due to strong elevation gradients. Had
548 rainfall been known, a three-component mixing model with a rain endmember of $6.05 \mu\text{S}/\text{cm}$
549 (Table 2) would not have had a significant impact, since the average measured EC at the
550 discharge station was $6.6 \mu\text{S}/\text{cm}$. For this reason, as well as the quick routing of rainwater
551 through the catchment due to steep topography, the model performance did not significantly
552 improve with further filtering of rainfall (see the modelling assumptions in section 3.1).

553

554 Modelled groundwater exfiltration does not solely describe localized groundwater resurgence
555 via springs. Quick hyporheic exchange must lead to some increase in stream water EC as
556 water flows through the forefield. Those processes are considered as groundwater exfiltration
557 by the model and may represent a significant fraction of groundwater flow in the forefield.
558 Brutsaert (2005) stressed that characterizing a basin as a single lumped unit with basin-scale
559 parameters is a useful concept but has limitations. The heterogeneity between different
560 sections of the aquifer is not taken into account, since the model considers the aquifer as a
561 homogenous body. It is nonetheless noteworthy that our simple model, consisting of only two
562 linear reservoirs and considering only two water sources, reliably reproduced discharge. This
563 suggests that despite its simplicity, the modelling approach provided an adequate description
564 of the catchment's hydrogeology. The rugged topography and heterogeneous soils should lead
565 to non-linear behaviors at a smaller scale. However, as pointed out by Fenicia et al. (2006),
566 groundwater reservoirs at the catchment scale tend to show relatively simple behavior.

567

568 Distributed physically-based models could potentially yield better results, but they require
569 reliable soil data at high spatial resolution. It is typically difficult in alpine catchments to
570 gather such type of data. Obtaining adequate snowmelt and glacier melt data already presents

571 important modelling challenges, as described in previous works at this catchment (Magnusson
572 et al., 2011; Kobierska et al., 2013).

573

574 5.4. Effect of snowmelt and sub-glacial glacier melt

575

576 Every year, the model tended to overestimate total discharge in mid-summer and
577 underestimate it at both ends of the season. At the beginning of the 2009 season (Fig. 4,
578 zoom 1), for instance, discharge is clearly underestimated during high flow spells, whereas
579 mid-August is more correctly modelled (Fig. 4, zoom 2). As mentioned in the results section,
580 this deficiency is likely due to variations in the “glacier melt” EC endmember. In early
581 summer, this component is actually mainly snowmelt. The difficulty is that snowmelt has a
582 relatively slow release rate and is in direct contact with saturated ground. Snowmelt is thus
583 more likely to infiltrate into soils than glacier melt, leading to intermediate EC values. This
584 had been evidenced in earlier studies such as Sueker et al. (2000).

585

586 Peak discharge values were also under-estimated in autumn (Fig. 4, zoom 3). The glacier melt
587 endmember may again have been slightly too small. According to Hindshaw et al. (2011), at
588 the end of each season the formation of a thin snow cover on the glacier leads to sub-glacial
589 routing of residual glacier melt, in contrast to the fast-flowing melt channels observed during
590 summer (see Fig. 9 in Hindshaw et al., 2011). This implies longer residence times under the
591 glacier, thereby increasing electrical conductivity and leading to underestimation of total
592 discharge.

593

594 Hindshaw et al. (2011) had focused on seasonal variations of the contribution of a “sub-
595 glacial component”. They, however, termed stream water - groundwater exchanges under the
596 ‘dead ice body’ as sub-glacial, which in our context may be misleading. As mentioned above,
597 we agree with their conclusions that the sub-glacial component changes behavior during the
598 hydrological season. In our opinion, stream water - groundwater exchanges under the ‘dead
599 ice body’ affect stream water EC more than sub-glacial flow under the bulk of the glacier,
600 which appears to lack a substantial underlying sediment layer.

601

602

603 5.5. Suggestions for future studies

604

605 Year-round availability of reliable groundwater level data would have been very useful to
606 further test the robustness of this approach. One difficulty in defining the fast reservoir
607 function was indeed the shallowness of most of our piezometers. We lacked an absolute
608 measure of storage during winter and the integral storage function had to be shifted so that the
609 upper fast reservoir emptied to a low residual value at the end October. This is typically the
610 end of the main hydrological season in high Alpine catchments. As mentioned in the methods,
611 a residual groundwater storage was added to the reservoir for the years that did not display
612 pure recession from the slow reservoir at the end of the calibration period. Groundwater data
613 usually became less reliable in November because potential icing or snow cover affected the
614 atmospheric pressure compensation of the signal. For this reason, the storage calculation was
615 stopped every year at the end of October.

616

617 In contrast to EC measurements, isotope ratios have the advantage of being fully conservative.
618 But since the hydrological signal displays daily fluctuations, mixing models would require
619 automated isotope sampling, infrastructure that is both fragile and expensive to install in such a
620 site. Manual oxygen isotope samples (see $\delta^{18}\text{O}$ values in Hindshaw et al., 2011) also showed
621 that groundwater mainly consisted of glacier meltwater. Only in localized areas did heavier

622 isotopes suggest some mixing with rainwater. However, at sites presenting stronger contrasts
623 between rainwater, groundwater and stream water, high-resolution isotope sampling could be
624 of great interest to complement the methodology presented in this study.

625

626 To better understand if the assumption of EC as a conservative tracer had an impact on the
627 mixing model, we attempted to measure EC at different locations along the stream. The
628 contrasts were too small to provide reliable insights into the progressive ionic enrichment of
629 stream water. Such an approach could be interesting in calcareous sites where EC contrasts
630 are usually stronger.

631 **6. Conclusions**

632

633 The main aim of this study was to estimate the contribution of groundwater exfiltration and
634 hyporheic exchange to streamflow at different times of the year in the partly glacierized
635 Damma glacier catchment in the central Swiss Alps. This site presented experimental
636 challenges specific to alpine areas, making it difficult to collect high-quality data throughout
637 the year. With this study, we improved our understanding of streamwater and groundwater
638 interactions during the main hydrological season as well as during winter and early spring.

639

640 Our approach builds on previous work which used two-component mixing models but did not
641 allow groundwater inflow to vary. We assumed that groundwater exfiltration was produced by
642 the combination of two linear storages. A 'slow' reservoir with a response time constant of 29
643 days was calibrated against a recession event in November 2008. It was overlain by a 'fast'
644 reservoir, which was modelled using groundwater level data from five locations in the
645 forefield. Groundwater exfiltration from both reservoirs fed a two-component mixing model
646 whose output was calibrated against measured discharge. The mixing model assumed that
647 stream water was composed of glacier melt and groundwater exfiltration end-members, which
648 displayed distinct and constant electrical conductivity values.

649

650 The model also yielded a realistic volume for groundwater storage actively contributing to
651 streamflow. Our results suggest that the 'slow' reservoir spans most of the forefield with an
652 average depth of approximately 1.7 m. The volume of the 'fast' aquifer was difficult to
653 estimate but is likely smaller. The 'fast' aquifer had a response time constant of 6.5 hours,
654 suggesting that it is highly hydraulically conductive and contributes to daily riparian and
655 hyporheic exchanges with the stream.

656

657 Modelling assumptions limiting water sources to two endmembers proved consistent with
658 field processes, as the model yielded reliable and reasonable estimates of streamflow. The set
659 of calibrated parameters worked for successive hydrological years marked by climatic
660 variability. In addition, total reservoir volumes and emptying rates were in agreement with
661 previous experimental work carried out at the forefield. This approach provided valuable
662 insights in a difficult alpine catchment and we believe it would be of interest at other sites to
663 infer essential properties of groundwater storage.

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767

768 **Table 1.** Key hydro-meteorological parameters of the catchment measured at the automatic
 769 weather station (AWS) and the main discharge station (S7, discharge only). Values in
 770 millimeters were calculated using a catchment area of 10.7 km². The start date for winter
 771 marks the establishment of a persistent snowpack at the AWS.

Winter	Start date	Peak SWE date	Max SWE (mm)	Summer (1 June to 1 Nov.)	Average temperature (°C)	Cumulated rainfall at AWS (mm)	Cumulated discharge at S7 (mm)
2008/2009	29 Oct.	29 Apr.	828	2009	8.1	544	2157
2009/2010	3 Nov.	5 Apr.	525	2010	7.6	598	1995
2010/2011	8 Nov.	20 Mar.	408	2011	8.1	674	2344
2011/2012	5 Dec.	25 Apr.	657	2012	8.7	764	2269

772
773

774 **Table 2.** Results of EC measurements ($\mu\text{S}/\text{cm}$) in the forefield. The zones refer to those
775 drawn in Figs. 1 and 6.

Zone	Average	Min	Max	Standard deviation	N° of samples
H1	15.5	12.8	18.3	1.2	89
H2	12.7	10.9	14	1.2	15
H3	15.4	13.6	17.8	1.3	13
H1+H2+H3	15.1	10.9	18.3	1.5	117
Rain	6.1	4.3	7.5	1.3	4

776

777

778 **Table 3.** Calibration of the full model (*FULL*) and both partial models (P_{EC} and P_{GW}) for four
779 years of data. P_{EC} uses only electrical conductivity data, whereas P_{GW} uses only groundwater
780 level data. For each calibration year, validation is performed on all remaining years.

	Nash Sutcliffe								
	Relative error (%)			efficiency			Benchmark efficiency		
	<i>FULL</i>	P_{EC}	P_{GW}	<i>FULL</i>	P_{EC}	P_{GW}	<i>FULL</i>	P_{EC}	P_{GW}
Calib 2009	13.3	30.6	37.3	0.78	0.44	-0.18	0.49	-0.30	-1.73
Valid 2010, 2011, 2012	27.4	37.3	48.8	0.58	0.29	-0.39	0.09	-0.55	-2.06
Calib 2010	13.5	31.9	41.9	0.90	0.63	-0.17	0.76	0.11	-1.80
Valid 2009, 2011, 2012	25.7	36.3	48.3	0.60	0.30	-0.53	0.14	-0.53	-2.35
Calib 2011	19.1	33.2	42.8	0.86	0.75	-0.23	0.70	0.46	-1.68
Valid 2009, 2010, 2012	42.0	54.2	51.9	0.25	0.29	0.21	-0.73	-0.60	-0.77
Calib 2012	25.7	36.3	52.9	0.64	0.21	0.01	0.25	-0.64	-1.05
Valid 2009, 2010, 2011	22.5	34.7	45.7	0.72	0.48	0.04	0.36	-0.17	-1.18

781 **Table 4.** Calibration and validation results for the '*FULL*' model after adjusting for $L_{residual}$.
782

	Relative error (%)	Nash Sutcliffe efficiency	Benchmark efficiency	$L_{residual}$ (mm)	Optimal T_{fast} (h)
Calib 2009	13.3	0.78	0.49	0	6.5
Valid 2010, 2011, 2012	19.5	0.8	0.57	-	
Calib 2010	13.5	0.90	0.76	26	
Valid 2009, 2011, 2012	19.4	0.76	0.48	-	
Calib 2011	19.1	0.86	0.70	156	
Valid 2009, 2010, 2012	17.5	0.78	0.51	-	
Calib 2012	25.7	0.64	0.25	86	
Valid 2009, 2010, 2011	15.3	0.85	0.51	-	

783
784

785
786

787 **Figure 1.** The Damma glacier forefield: at sites S1, S3, S5 and S6 (solid circles), stream and
788 groundwater levels are recorded. At site S7 (solid square), stream stage is measured for total
789 discharge. At site S8 (solid triangle), only one piezometer is installed. Color patches indicate
790 zones of high (H1, H2 and H3) and low (L1, L2) electrical conductivity. An automatic
791 weather station (AWS) is located in the middle of the forefield. Lateral moraines are indicated
792 with dashed black lines and terrain elevation is shown by 10 m contour intervals. (Figure
793 adapted from Magnusson et al., 2014).

794

795 **Figure 2.** Baseflow recession during November 2008. The lower panel shows measured and
796 modelled discharge at S7 on a logarithmic scale. The dotted black line illustrates how the
797 modelled baseflow recession diverges from measured discharge before November 15th. Note
798 that melting periods (non-negative temperature of snow surface) are indicated as grey shaded
799 bars. The upper panel plots successive rain events.

800

801 **Figure 3.** Schematic flow chart summarizing the functioning of the '*FULL*' model.

802

803 **Figure 4.** The upper section presents model results for the entire 2009 season. The lower
804 section presents zooms on three specific weeks. For each group of three graphs, the bottom
805 panel displays both measured and modelled discharge (m^3/s) at S7. The middle panel presents
806 infiltration and exfiltration (m^3/s). Electrical conductivity ($\mu\text{S}/\text{cm}$) and rainfall (mm/h) are
807 plotted in the upper panel. Time periods that were filtered out can be seen as gaps in the EC
808 data (e.g., in zoom 1).

809

810 **Figure 5.** Modelled ratio of groundwater exfiltration to total modelled discharge (in %) as a
811 function of total measured discharge for 2009. Only time steps with less than 10% relative
812 error against measured discharge were plotted.

813

814 **Figure 6.** Recharge of the 'slow' reservoir from snowmelt in spring 2011. Snow melt periods
815 (non-negative snow surface temperature) are indicated as grey shading. Groundwater levels,
816 displayed in blue, represent the depth of water in each piezometer. Total discharge and
817 reservoir depth are plotted in black. Reservoir level is computed using Eq. 10 based on a
818 surface area of 400 by 1000 meters. The corresponding level of the full reservoir is indicated
819 (dotted line).

820

821 **Figure 7.** Conceptual summary of the forefield's hydrogeology. The stream network is drawn,
822 as well as the main groundwater springs where electrical conductivity was measured and all
823 electrical conductivity zones. A tentative outline of the active reservoir ('slow' + 'fast') is
824 proposed. The discharge station S7 is not displayed as it is slightly outside of the side-cut. The
825 lateral moraines are shown by red dotted lines.

826