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Chapter 1

Geological and Hydrogeological Characterization of Springs in a DSGSD Context (Rodoretto Valley – NW Italian Alps)



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Abstract As continuous groundwater monitoring in the upper sector of Rodoretto Valley (Germanasca Valley, Italian Western Alps) is hampered by logistical problem of data collection during winter and spring months, the only tools currently available to derive hydrogeological information are non-continuous and non-long-term dataset of spring discharge (Q), temperature (T) and electrical conductivity (EC).

In order to quantify aquifer groundwater reserve, available Q dataset of a small mountain spring (Spring 1 CB) was investigated by applying the analytical solutions developed by Boussinesq (J Math Pure Appl 10:5–78, 1904) and Maillet (Essais di'hydraulique souterraine et fluviale, vol 1. Herman et Cie, Paris, 1905); T and EC datasets were also used to provide qualitative information about the nature of the aquifer that supplies the spring.

The outcomes of the elaborations highlighted the limits of applicability of these methods in the presence of a non-continuous Q dataset: both Boussinesq (J Math Pure Appl 10:5–78, 1904) and Maillet (Essais di'hydraulique souterraine et fluviale, vol 1. Herman et Cie, Paris, 1905) estimated that discharge values as a function of recession time were found to be consistently lower than the available discharge ones and the estimated groundwater volumes stored over time above the spring level turned out to be underestimated.

Continuous (hourly value) and long-term Q, EC and T values are, therefore, needful to correctly quantify and to make a proper management of groundwater resources in mountain areas.

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1.1 Introduction

The global demand for water has been increasing at a rate of about 1% per year over the past decades as a function of population growth, economic development and changing consumption patterns, and it will continue to grow significantly over the foreseeable future (United Nation 2018).

In Italy, 84.3% of the national clean water derives from groundwater, where 48.0% results from well, 36.3% from spring, 15.6% from surface waters and the remaining 0.1% from marine water: spring represents, therefore, one of the largest and precious sources of water, necessary to meet the water needs of the population (Istat 2017).

The continuous expansion of urban areas has caused a growing interest in finding new potable water sources and led to consider mountain aquifers as an increasingly more strategic resource.

In this contest, the mountain water resource management is a topic that has become increasingly important.

As mountain aquifers can be particularly vulnerable from qualitative and quantitative point of view, they need a high degree of protection: it is important to understand their recharging system, from both geological and hydrogeological perspectives, in order to protect and optimize its present and future management.

A large number of methodologies have been developed over time to derive hydrogeological information about mountains springs recharging systems. In the early 1900s, Boussinesq (1877, 1904) and Maillet (1905) developed analytical models to quantitatively investigate hydrograph and spring recession mechanisms. Such methods were widely applied in the past years as they allowed to determine the possibilities of underground water resource storage and exploitation (Dewandel et al. 2010; Fiorillo et al. 2012; Giacometti et al. 2016; Jakada et al. 2019).

Apart from hydrograph and recession curve analysis, autocorrelation and cross-correlation methods were elaborated and applied to spring monitoring dataset: Desmarais and Rojstaczer (2001), Galleani et al. (2011) and Lo et al. (2014) showed how the spring behavioural model of the drainage can be understood by analysing continuous dataset of discharge (Q), rainfall (P), temperature (T) and electrical conductivity (EC).

Time series analyses of Q, P, T and EC dataset, widely applied to characterize large karst systems (Bonacci 1993; Galleani et al. 2011; Luo et al. 2016; Banzato et al. 2017), have been also recently applied to study mountain springs supplied by porous and shallow aquifers (Lo et al. 2014; Amanzio et al. 2016).

However, in remote mountain settings, continuous groundwater monitoring is often hampered by financial and logistical problem of instrumentation and data collection, especially in winter and spring months.

Alpine areas generally lack easy access and commonly are within sectors designated or managed as wilderness, precluding use of motorized equipment. Traditional techniques for sampling groundwater physical parameters are difficult to use, and modern methods for assessing aquifer properties are often not allowed or strongly discouraged (Clow et al. 2003; Tobin and Schwartz 2016).

The different logistical difficulties just described above were met in the alpine glacial valley of Rodoretto Valley (Germanasca Valley, Italian Western Alps).

In the past years, several authors (Forno et al. 2011; Forno et al. 2012; Piras et al. 2016) have conducted detailed survey, identifying the geological and geomorphological characters and resulting in the production of a new morphological and quaternary geological map of the area. A lot of peculiar morphologies connected to deep-seated gravitational slope deformations (DSGSDs) phenomena as scarps, depressions, transversal trenches and ridges elongated parallel to watersheds have been recognized and described in Forno et al. (2011).

In addition, some water mountain sources, many of which potentially exploitable for drinking purposes, have been identified along the longitudinal trenches mapped in the upper sector of Rodoretto Valley: by mean exclusive geological investigations, it was, therefore, possible to demonstrate how the pattern of the hydrographic network is strongly affected by the recognized gravitational features (Forno et al. 2012).

With the aim to implement such available geological information, it was necessary making hydrogeological studies and a preliminary characterization of mountain springs.

In this work, a selected spring (Spring 1 CB) was investigated by using available non-continuous and non-long-term recorded dataset of discharge (Q), temperature (T) and electrical conductivity (EC) parameters.

Specifically, (1) a critical summary of published works on specific topic of spring monitoring and spring modelling approaches base on analytical (physically based) and statistical methods; (2) recession curve analysis on available dataset of Q in order to quantity aquifer groundwater reserves; and (3) spring monitoring datasets of T and EC analysis to provide qualitative information about the nature of the aquifer that supplies the spring were made.

1.2 Methods

1.2.1 Analysis of Recession Curve – Analytical Methods

The analysis of spring discharge hydrographs is the most suitable tool for the study of mountain springs and for the definition of aquifer characteristics, such as the type and quantity of its groundwater reserves.

The spring hydrograph is the final result of various processes that govern the transformation of precipitation and other water inputs in the drainage area into the single output at the spring (Fig. 1.1).

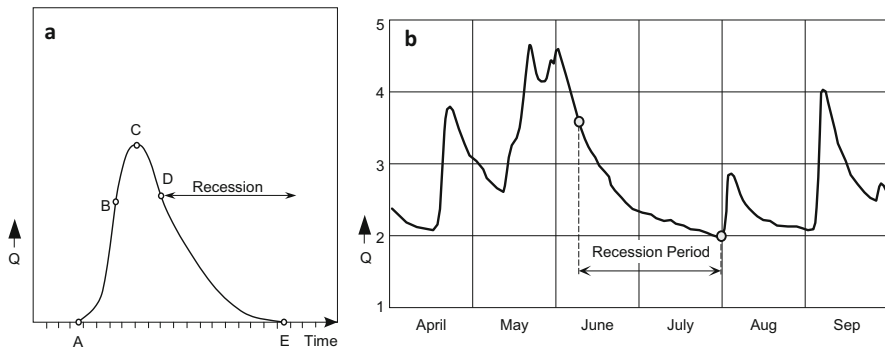


Fig. 1.1 (a) Components of a discharge hydrograph (Modified from Kresic and Bonacci 2010). (b) Example of spring discharge hydrograph (Modified from Kresic 2007)

Its shape depends on the size of the drainage area, as well as the precipitation intensity, and it is constructed according to the discharge values (Q), measured during a hydrogeological year.

It is defined by a rising limb (AB), a crest (BCD) and a falling or recession limb (DE), which corresponds to the recession period (Kresic and Bonacci 2010) (Fig. 1.1a).

Particularly, the recession limb (DE) is that part of the spring hydrograph that extends from a discharge peak to the base of the next rise (Fig. 1.1b). It contains information on storage properties and different types of media, such as porous, fractured and cracked lithology, and provides a function that quantitatively describes the temporal discharge decay and expresses the drained volume between specific time limits (Sayama et al. 2011).

The recession curve analysis has been useful in many areas for estimating the hydrological significance of the discharge function parameters and the hydrological properties of the aquifer (Amit et al. 2002).

Over the decades, many studies were made on recession curve: initial methods for the analysis of hydrograph recession periods, developed in the early 1900s, are based on Boussinesq (1904) and Maillet (1905) equations, which both give the dependence of the flow rate at specified time (Q_t) on the flow at the beginning of recession (Q_0).

In 1904, Boussinesq developed an exact analytical solution of the diffusion equation that describes flow through a porous medium, by considering the simplifying assumptions of a porous, free, homogeneous and isotropic aquifer, with the aquifer being limited by an impermeable horizontal layer at the level of the outlet (Eq. 1.1):

$$Q_t = \frac{Q_0}{(1 + \alpha[t - t_0])^2} \quad (1.1)$$

where Q_t (m^3/s) is the flow rate value at $t \neq t_0$; Q_0 is the flow rate at $t = t_0$; α is a constant depending only on the aquifer hydraulic systems (**recession coefficient**) (Eq. 1.2).

$$\alpha = \frac{\sqrt{Q_0} - \sqrt{Q_t}}{\sqrt{Q_t}t} \tag{1.2}$$

Maillet (1905) showed instead that the recession of a spring can be represented by an exponential formula, implying a linear relationship between the hydraulic head and flow rate (Eq. 1.3):

$$Q_t = Q_0 e^{-\alpha(t-t_0)} \tag{1.3}$$

where Q_t (m^3/s) is the flow rate value at $t \neq t_0$; Q_0 is the flow rate at $t = t_0$; α is a constant depending only on the aquifer hydraulic systems (**recession coefficient**) (Eq. 1.4).

$$\alpha = \frac{\log Q_0 - \log Q_t}{e(t - t_0)} \tag{1.4}$$

Both in Boussinesq (1904) and Maillet (1905), the recession coefficient equation was used to determine important hydrogeological parameters as W_0 , the groundwater volume stored above the spring level at the end of spring season (beginning of recession), and W_d , the groundwater volume stored at the end of autumn season (end of recession) (Table 1.1).

Table 1.1 Boussinesq (1904) and Maillet (1905) equations to determine the groundwater volume stored at the end of spring season (W_0) and the groundwater volume stored at the end of autumn season (W_d)

<p>1904</p>	<p>Boussinesq $W_0 = \frac{Q_0}{\alpha(1+\alpha t)^2} \times 86400$ (1.5); $W_d = \left[\frac{Q_0}{\alpha} - \frac{Q_0}{\alpha(1+\alpha t)} \right] \times 86400$ (1.6)</p>	<p>Where: Q_t (m^3/s) is the flow rate value at $t \neq t_0$; Q_0 is the flow rate at $t = t_0$; α is a constant depending only on the aquifer hydraulic systems (recession coefficient)</p>
<p>1905</p>	<p>Maillet $W_0 = \frac{Q_0}{\alpha} \times 86400$ (1.7) $W_d = \frac{(Q_0 - Q_t)}{\alpha} \times 86400$ (1.8)</p>	

About 50 years later, Castany (1967) approximated the connection between the karst spring discharge (Q) during a recession period and the hydraulic head in the aquifer (H) by the following equation (Eq. 1.9):

$$Q = \lambda H^n \quad (1.9)$$

where λ is a coefficient connected to conductivity and effective porosity of the aquifer; $n = 1$ when the discharge is linear; $n > 1$ when the discharge is nonlinear.

A great contribution to recession curve analysis was also given by Schoeller (1967), Drogue (1967) and Mangin (1970). As described in Civita (2008), according to Schoeller (1967) and Drogue (1967) methods, it can be useful to record daily values of discharge on a semi-log scale in order to represent the depletion curve points along a straight line: the intersection of the best fitting line for these points with the ordinate axis gives the value of initial flow at $t = 0$ (Q_0). By inserting this value into Maillet's equation and by calculating the depletion coefficient α , it is indeed possible to obtain the equation representing depletion of the aquifer network reserves over time.

In 1975, Mangin developed an expression that can be used to analyse spring recession curve by including an empirical homographic function that represents the first section of the recession curve in the presence of infiltration phenomena (Eq. 1.10):

$$Q_t = Q_0 e^{-\alpha(t-t_0)} + q_0 + \frac{1 - n't}{\varepsilon't} \quad (1.10)$$

where q_0 represents the difference between the peak flow rate (Q_{\max}) and the initial flow rate (Q_0); η is the infiltration velocity coefficient (d^{-1}) as it is inversely proportional to the duration of infiltration event; ε is the heterogeneity of outflow (d^{-1}) and it identifies attenuation of infiltration rate over time.

Eventually, Dewandel et al. (2003) offered a comprehensive review of approaches for analysing recession by conducting numerical simulations of shallow aquifers with an impermeable floor at the level of the outlet by comparing Boussinesq and Maillet analytical solutions.

In their results, they showed how the Boussinesq formula should be preferred to the Maillet exponential form in such hydrogeological situations: if the Boussinesq method fits the entire recession and could provide correct estimates of the aquifer's parameters, the Maillet solution largely underestimated the dynamic volume of the aquifer.

1.2.2 Analysis of Recession Curve – Time Series Analysis Methods

Assuming that the spring discharge is not affected by a rapid inflow of water into the aquifer, the recession analysis by using analytical relationship provides good insight into the aquifer structure, allowing to predict the volume of groundwater stored at the end of spring and autumn seasons.

However, ideal recession conditions with a long period of several months without precipitation are infrequent in moderate, humid climates: precipitations can disturb the recession curve and may not be removed unambiguously during analysis (Kresic and Bonacci 2010).

As a consequence, the majority of spring modelling approaches so far to better characterize spring discharge are based on various applications of time series analyses as well as general statistical and probabilistic methods.

Mathematical tools as the auto and cross-correlation functions are frequently used in the past years in hydrogeology for the analysis of a time series of values (Panagopoulos and Lambrakis 2006; Fiorillo and Doglioni 2010; Kresic and Stevanovic 2010; Kresic and Bonacci 2010).

Statistically, the autocorrelation of a random process (e.g. spring discharge) describes the correlation between values of the process (Q) at different points in time (t), as a function of the two times or the time difference; the cross-correlation represents a time-dependent relationship between an output process (e.g. daily spring discharge) and an input process (e.g. daily precipitation).

Primary hydrogeological information about the spring system can be derived by applying such analytical techniques on available spring monitoring dataset of Q , T and EC .

Through hydrograph analysis and calculated correlations between Q , T and EC dataset as a function of infiltration input (P), Galleani et al. (2011) revealed three broad aquifer behavioural categories, based on the drainage network effectiveness (i.e. Replacement, Piston and Homogenization), and developed an index to quantify spring vulnerability (Vulnerability Estimator for Spring Protection Areas).

In Banzato et al. (2011), the VESPA index method, proposed by Galleani et al. (2011), has been examined through a statistical analysis of time series with respect to the following aspects: (1) the time extension of the dataset, (2) the use of averaged data and (3) the number of missing data: the outcomes led to show that the VESPA index is a reliable way to identify higher vulnerability levels, and some uncertainty occurs in the identification of low and medium vulnerability levels.

In Lo et al. (2014), auto and cross-correlation analyses were applied to overall monitoring dataset of Q , P , T and EC of four alpine springs, supplied by porous and shallow aquifers, located in the Aosta Valley. Particularly, by evaluating the auto and cross-correlation coefficient of a time series, they have estimated the spring 'memory effect', a parameter that reflects the duration of system's reaction to input signal (Mangin 1984), understanding how the system reacts to infiltration phenomena induced by precipitation.

1.3 Study Area

The Rodoretto Valley is a left tributary of the Germanasca Valley (Italian Western Alps), located inside the Mesozoic formation of the Greenstone and Schist Complex (Piedmont Zone) (Fig. 1.2).

The geological units have been attributed to the Penninic paleogeographic-structural domain (Sandrone et al. 1993), a multi-fold system represented by units of Mesozoic oceanic crust (Piedmont Zone) and basements in which have been distinguished the Upper Penninic Units (Monte Rosa, Gran Paradiso, Dora Maira Massifs), the Intermediate System (Gran San Bernardo Massif) and the Lower Penninic Units.

The Rodoretto Valley represents a glacial valley with steep rocky mountain sides and a flat valley floor.

The main elements of the landscape are small lateral and frontal moraines and many gravitational elements as minor scarps and counterscarps, linear trenches and landslide niches and landslides: these glacial and gravitational evidences, connected with deep-seated gravitational slope deformations (DSGSDs) (Agliardi et al. 2009), have been recognized on both sides of the valley (Forno et al. 2011, 2013).

The presence of a DSGSD, the widespread presence of open fractures and the consequent high degree of loosening of the bedrock, particularly evident on the upper sector of the right side of the valley (Forno et al. 2011), proved to be very favourable to the storage of large quantities of water (Fig. 1.3).

Several small springs are indeed located at the base of the right side of the Rodoretto Valley: their location appears strongly influenced by the tectonic discontinuities developed in the bedrock and by the morphostructures related to DSGSDs.

1.3.1 Spring 1 CB (Cavallo Bianco)

The spring analysed was named Spring 1 CB (Cavallo Bianco). It is located in the upper sector of Rodoretto Valley at 1954 m a.s.l., elevated about 10 m above the valley floor (Fig. 1.4).

Among the mountain spring located at the base of the slope right side of the Rodoretto Valley it been selected as its location and the outflow mode during summer season made it possible to measure the Q, T and EC parameters with a high degree of reliability and accuracy.

It outflows at the contact between the bedrock and an original slight cover sheet of glacial deposits, representing the aquifer which supply the spring (Fig. 1.5).

Part of the water, equal to 6.5 l/s, is collected by means of a reinforced concrete structure and reaches, the Rodoretto village (Comune di Prali 2009).

The Spring 1 CB was monitored first during the period between June 15th and October 28th 2016, and second between 11th June and 18th October 2017, with samples taken every 15–20 days approximately.

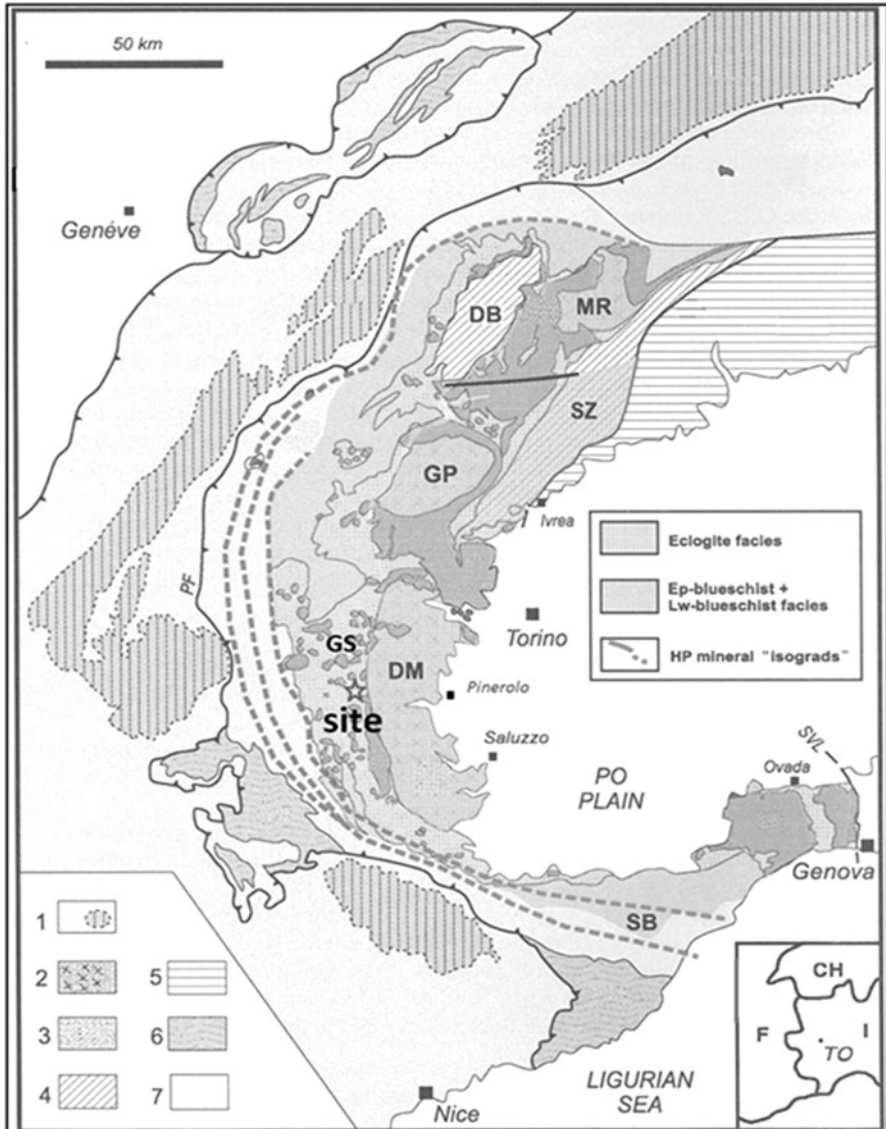


Fig. 1.2 Western Alps sketch map. (Modified from Compagnoni 2003)

1. Jura, Helvetic Domain and external Penninic Domain; 2. Internal Crystalline Massif of the Penninic Domain. Dm: Dora Maira Massif; 3. Piedmont Zone; GS: Greenstone and Schist Complex; 4. Austroalpine Domain; 5. South Alpine Domain; 6. Helminthoid Flysch Nappes; 7. Swiss Molasse

The very steep slopes of the Rodoretto Valley, engraved by a hydrographic network, favour the accumulation of huge snow masses during the winter and spring seasons, made the sector investigated inaccessible in the months from November to



Fig. 1.3 Upper sector of the right side of the Rodoretto Valley (Italian Western Alps)



Fig. 1.4 (a and b) Spring 1 CB

June (Fig. 1.6). Furthermore, traditional method for sampling groundwater physical parameters (Q, T and EC continuous datasets) by using a fixed multi-parameter probe and data logger was not allowed because it was not possible to obtain permission from the competent authority.

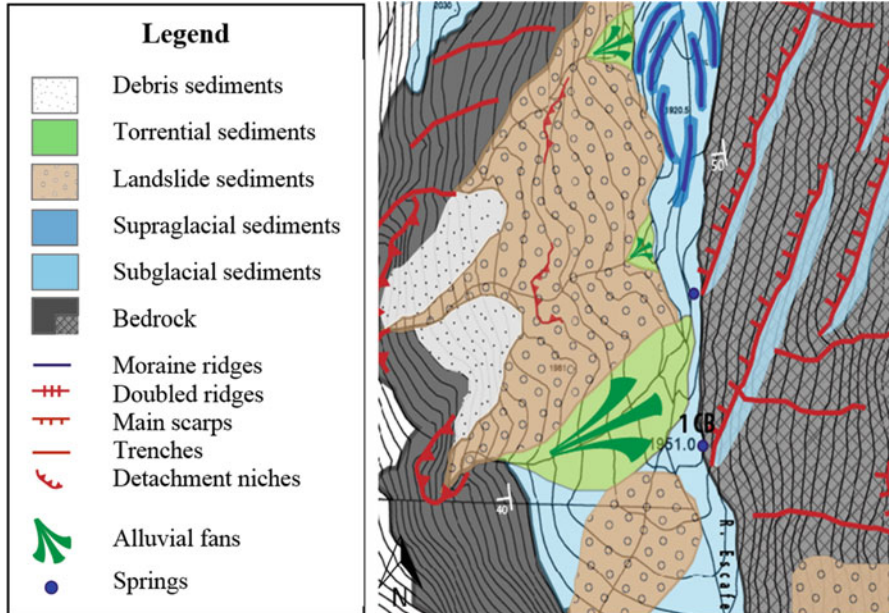


Fig. 1.5 Geological map of the Spring 1 CB area (scale 1:10000)



Fig. 1.6 (a and b) Accumulation of snow masses in the upper sector of Rodoretto Valley (11/06/2017 to 15/07/2017)

1.4 Results and Discussion

The majority of methods of spring characterization are based on various applications of time series analysis as well as general statistical methods described above in the methods paragraph. Such methods require to be applied on multi-years' time series of data on aquifer recharge (P), spring discharge (Q), temperature (T) and electrical

Table 1.2 Discharge values of Spring 1 CB during the monitoring period 2016–2017. Discharge values of spring calculated by Boussinesq (1904) and Maillet (1905) solutions

	Date	Days	Q (m ³ /s)	Boussinesq (m ³ /s)	Maillet (m ³ /s)
2016					
1	29/06/2016	–	0.0125	–	–
2	13/07/2016	0	0.0118	0.0118	0.0118
3	28/07/2016	15	0.0109	0.0097	0.0098
4	21/08/2016	39	0.0087	0.0064	0.0066
5	03/09/2016	52	0.0075	0.0051	0.0051
6	25/09/2016	74	0.0069	0.0040	0.0040
7	28/10/2016	–	0.0070	–	–
2017					
1	30/06/2017	–	0.0130	–	–
2	15/07/2017	0	0.0124	0.0124	0.0124
3	28/07/2017	13	0.0119	0.0105	0.0107
4	23/08/2017	39	0.0088	0.0062	0.0063
5	12/09/2017	59	0.0067	0.0040	0.0041
6	28/09/2017	75	0.0066	0.0035	0.0035
7	18/10/2017	–	0.0067	–	–

conductivity (EC): this represents, therefore, a key limiting factor for many practical hydrogeological spring investigations with short execution times (Kresic 2007).

As in Rodoretto Valley, remote mountain settings generally lack easy access and continuous groundwater monitoring data collection with suitable probes or data logger is prevented by logistical problem, especially in winter and spring months.

Consequently, in many cases, the only tools available to hydrogeologists to derive hydrogeological information about mountains springs recharging systems are non-continuous and non-long-term dataset of discharge (Q), temperature (T) and electrical conductivity (EC) parameters.

Considering (1) the frequency at which the spring 1 CB discharge (Q) could be measured (see Table 1.2); (2) the non-kars condition of analysed spring 1 CB; (3) the absence of exceptional precipitation or new water rapid inflow events during the monitored period (summer and autumn seasons 2016/2017) which affect the summer spring discharge (Fig. 1.8), it has been applied Boussinesq (1904) and Maillet (1905) approach to analyse the depletion curves (un-influenced stage) with a modest degree of reliability.

1.4.1 Hydrograph Analysis

Available data related to discharge (Q) for the investigated springs are shown in Table 1.2.

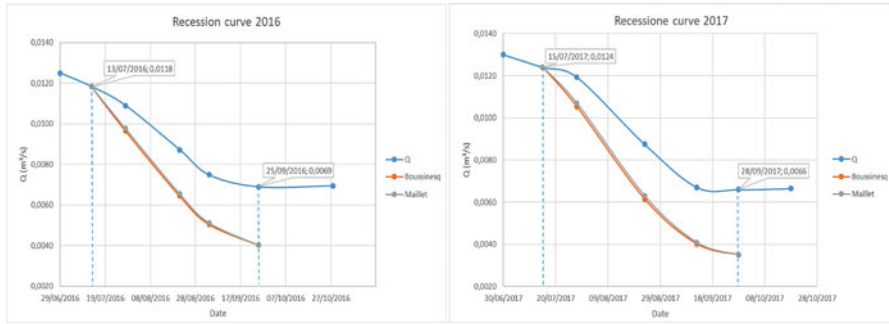


Fig. 1.7 Comparison between recession curve characteristics for Spring 1 CB (2016–2017) and estimated recession curves by Boussinesq (1904) and Maillet (1905) solutions

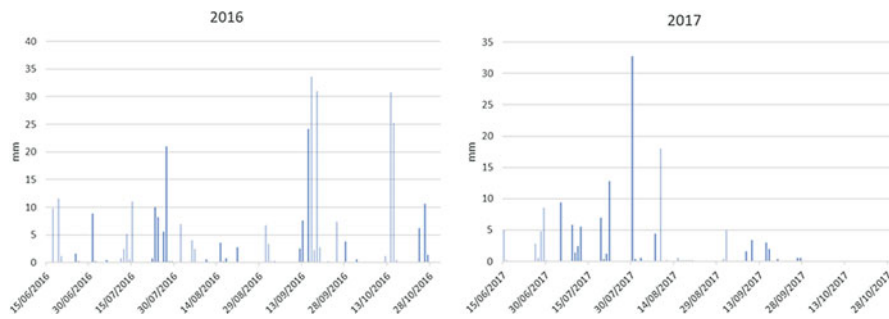


Fig. 1.8 Available rainflow data (Villa di Prali meteorological station, Arpa Piemonte 2016–2017)

The investigated periods were between (1) 13/07/2016 and 25/09/2016 and (2) between 15/07/2017 and 28/09/2017. During these periods, the recession curves do not show quick flow variations or pronounced discharge fluctuations.

The spring 1 CB shows a relatively low discharge rate with a discharge peak in early summer: the maximum value of discharge recorded in 2016 and 2017 is 0.013 m³/s (29/06/2016 to 30/06/2017). The minimum value of discharge in both 2016 and 2017, recorded at the end of September, is 0.007 m³/s (25/09/2016 to 28/09/2017).

For the test site, the Boussinesq (Eq. 1.1) and Maillet (Eq. 1.3) solutions were applied and the calculated discharge data were found to be consistently lower than the available discharge ones (Table 1.2; Fig. 1.7).

Due to the measurement uncertainties, both Boussinesq (Eq. 1.1) and Maillet (Eq. 1.3) solutions will cause errors in estimating the recession coefficient (α).

Nevertheless, the estimated recession coefficients allowed to quantify two important hydrogeological parameters, useful for the quantification of the resources available during the recession period: W_0 is the estimated groundwater volume stored above the spring level at the end of spring season (beginning of recession)

Table 1.3 Estimated recession coefficient (α), groundwater volume stored at the end of spring season (W_0), groundwater volume stored at the end of summer season (W_d)

	α	W_0 (m ³)	W_d (m ³)
2016			
Boussinesq (1904)	0.004	186,707,23	57,764,68
Maillet (1905)	0.007	140,295,09	58,466,17
2017			
Boussinesq (1904)	0.005	158,142,40	58,621,62
Maillet (1905)	0.008	127,414,77	59,597,23

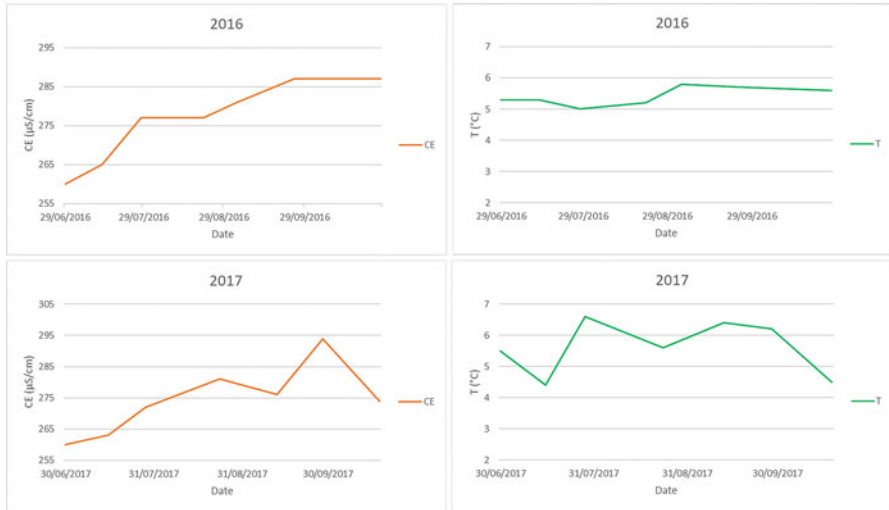


Fig. 1.9 Temporal variations of electrical conductivity (EC) and temperature (T) of the Spring 1 CB (2016–2017)

and W_d represents the estimated groundwater volume stored at the end of summer season (end of recession) (see Table 1.3).

1.4.2 Thermograph and EC Graph

In addition, thermograph and EC graph were plotted using available data, recorded during the monitoring period between June and October 2016 and 2017 (Fig. 1.9).

EC and Q parameter recorded very weak variations during 2016–2017 summer seasons: in the first summer months, the EC appears to feature slightly lower values (260 $\mu\text{S/cm}$) due to the dilution effect of the ionic species, favoured by the supply of glacial melt and the consequent higher values of Q.

Moreover, analysing the values relating to the water temperature, we note that these remain overall constant over the period of time between June and October 2016–2017 (average value of about 5.0 $^{\circ}\text{C}$).

As described above, Spring 1 CB outflows at the base of a sliding scarp that displaces an original slight cover sheet of glacial deposits, representing the aquifer which supplies the spring.

The constancy in the values of CE and T parameters is in accordance with (1) the absence of freshly water recharge events induced by rainflow or snow-melting phenomena that affect in short times the spring summer discharge and physical parameters and (2) the presence of a porous glacial aquifer that supplies the spring of water with same physical characteristics.

1.5 Conclusions

As continuous groundwater monitoring in the upper sector of Rodoretto Valley (Germanasca Valley, Italian Western Alps) is hampered by logistical problem of instrumentation and data collection during winter and spring months, the only tools available to derive hydrogeological information were non-continuous and non-long-term dataset of discharge (Q), temperature (T) and electrical conductivity (EC) parameters.

A small selected mountain spring (Spring 1 CB) was investigated by applying on non-continuous monitoring discharge (Q) dataset the analytical solutions developed by Boussinesq (1904) and Maillet (1905) on spring recession curve (2016–2017).

The outcomes of the elaborations highlighted the limits of applicability of these methods in the presence of a non-continuous discharge dataset: both Boussinesq (1904) and Maillet (1905) gave a good fit of the shape of measured data-recession curve, but the estimated discharge values as a function of recession time were found to be consistently lower than the available discharge ones.

As a consequence, the estimated groundwater volume stored above the spring level at the end of spring season (W_0) and the groundwater volume stored at the end of summer season (W_d) are underestimated and can give only hydrogeological-qualitative information.

Continuous (hourly value) and long-term (complete hydrogeological year) Q and P datasets are indeed needful not only in applications of time series analyses as well as statistical and probabilistic methods but also for the application of Boussinesq (1904) and Maillet (1905) depletion curves analysis.

Finally, the absence of a hydrodynamic impulsion response to the infiltrative processes and the recorded very weak variations of T and EC parameters suggest the presence of a low effective drainage system and an extended saturated zone which allow freshly infiltrated water-resident circulating groundwater homogenization phenomena.

If continuous (hourly value) Q, EC and T monitoring datasets have been available for a significant time interval, they would have allowed to understand in detail to what extent the presence of a DSGSD influences the position, discharge, the chemical–physical characteristics but also the infiltration dynamics and the vulnerability of analysed spring.

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