



UNIVERSITÀ
POLITECNICA
DELLE MARCHE

PhD program:

Civil, Environmental and Building Engineering and Architecture (ICAEA)

PhD curriculum:

Civil, Environmental and Building Engineering and Architecture

XXXIV Cycle (2019-2021)

PhD Thesis title:

Definition of the hydrogeological conceptual model with an integrated approach in earthquakes-induced changing contexts

PhD Candidate:

Davide Fronzi

Advisor:

Prof. Alberto Tazioli

Curriculum Coordinator:

Prof. Francesco Fatone

Making the invisible visible...

Table of Contents

Table of Contents	
Abstract	
Riassunto.....	
Dedication.....	
Acknowledgements.....	
List of figures.....	
List of tables.....	
1 Introduction.....	1
1.1 General outline.....	1
1.2 Synthesis: the state of the art and the state of practice	3
1.2.1 Carbonate, karstic and fissured aquifers	3
1.2.2 Methods of tracing techniques.....	8
1.2.2.1 <i>Artificial tracers</i>	9
1.2.2.2 <i>Isotope hydrogeology</i>	14
1.2.3 Hydrogeological modification due to seismic events	21
1.2.4 Hydrogeological conceptual models	26
1.3 Aims and motivations of this PhD research project	29
1.4 Flow-chart of the research method	30
1.5 Geological and hydrogeological setting of the study area	31
1.5.1 Geo-lithological features and tectonic lineaments.....	31
1.5.2 Seismicity of the area.....	38
1.5.3 Climatic and meteorological features	42
1.5.4 Hydrogeological features	43
2 Material and methods	55
2.1 Site characterization	55
2.1.1 Precipitation and meteo-climatic monitoring system	55
2.1.2 Standard Precipitation Index	56
2.1.3 Hydrological monitoring system	58
2.1.4 Hydrological balance	68
2.1.4.1 <i>Indirect methods (Turc method vs WaterbalANce)</i>	68
2.1.4.2 <i>Calibration of Infiltration Rate through direct method</i>	72

2.1.5	Recession curve analysis.....	76
2.2	Tracer tests.....	78
2.2.1	Selection of injection point and injection method.....	78
2.2.2	Monitoring points.....	81
2.2.3	Tracer tests processing.....	83
2.2.4	Sliding Windows Cross-correlation function vs tracer tests	84
2.3	Isotopic and hydrochemical sampling.....	88
2.3.1	Monitoring points.....	88
2.3.2	Snowfall and snowmelt	90
2.3.3	Laboratory operations.....	92
2.3.4	Isotopic data processing	93
2.3.4.1	Find changes in isotopic signals	93
2.3.4.2	The GIS based- $\delta^{18}\text{O}$ isotope overlapping method	94
2.4	Geological field work and GIS supported analysis (The northern portion of the study area - Upper Nera River basin).....	95
3	Results	97
3.1	Precipitation and meteo-climatic results.....	97
3.2	Direct hydrological balance and long-term discharge variation at regional scale	110
3.3	Spring discharge analysis.....	126
3.4	Artificial tracer tests analysis	135
3.4.1	Regional groundwater flow modification corroborated by artificial tracer tests	153
3.4.2	Comparison between periodic tracer tests and time-series analysis.....	158
3.5	Isotope hydrology outcomes.....	164
4	Discussion.....	182
4.1	Post-seismic hydrogeological conceptual model	204
4.2	Snowpack melting processes	208
5	Conclusions	210
	References	214

Abstract

Understanding the groundwater flow, recharge, and storage processes in carbonate aquifers represents a challenging aspect in hydrogeology. These aquifers are hosted in geological formations characterised by strong heterogeneities and anisotropies affecting the hydrodynamic characteristics of the systems. This comprehension becomes even more complicated when dealing with mountainous aquifers, where the geological and the geo-structural complexity is combined with high topographical heterogeneity. The topographical heterogeneity can significantly influence the rainfall regime, and the higher altitudes, often characterised by the presence of snow cover, make a non-negligible contribution to the aquifers' recharge processes.

In this context, it has been demonstrated that high-magnitude earthquakes can change the hydrodynamic behaviour of springs and rivers showing transient or long-lasting hydrogeological modifications, making the water management a real hard task.

This is the case of a wide Apennines area located in central Italy, the one affected by an intense seismic sequence between August 2016 and January 2017, characterized by nine events of $M_w \geq 5.0$ ($M_{max} = 6.5$) whose epicentre is located within the study area.

This research aims to increase the hydrogeological and hydrodynamic knowledge related to groundwater circulation in carbonate aquifers by investigating possible changes in the qualitative and quantitative characteristics of groundwater flow, during and after a series of strong seismic events.

A combined holistic approach, involving different disciplines and techniques, was used.

The investigation starts with a more traditional hydrogeological approach, based on the application of the hydrological balance and the analysis of springs hydrographs, then validated through the results of several artificial tracer tests, hydrochemical and isotopic analyses.

During the study, a detailed hydro-geo-structural survey was carried out, integrated with GIS-supported analysis, aimed both at identifying the main geological structures responsible for the groundwater flow modifications, and at characterising the hydrogeological connections between the hydrostructures recognised in the study area.

The results of this research confirm the geo-structural complexity of the domain and its influence on the hydrogeological features, highlighting the role of faults, having a regional or local extension, played in modifying the groundwater flow.

The outcomes obtained from each investigation technique had compared each other in order to outline a detailed conceptual hydrogeological model which has been updated step by step following the post-seismic phases to provide to the public and private drinking companies operating in the area, a tool for the correct and sustainable groundwater management operations.

Keywords: carbonate aquifers; earthquake hydrogeology; tracer tests; isotope hydrology; water balance; conceptual models.

Riassunto

La comprensione dei processi di ricarica, di movimento e di immagazzinamento delle acque sotterranee negli acquiferi carbonatici rappresenta una grande sfida per l'idrogeologia. Infatti, questi acquiferi sono ospitati in formazioni geologiche caratterizzate da forti eterogeneità e anisotropie che si possono ripercuotere nelle caratteristiche idrodinamiche del sistema. Questa comprensione diventa tanto più complicata quando si trattano acquiferi ricadenti in aree montuose, ove la complessità geologico-strutturale si unisce a una grande eterogeneità topografica. Quest'ultima può influenzare il regime pluviometrico in maniera significativa, ed inoltre le quote elevate sono spesso caratterizzate dalla presenza del manto nevoso che apporta un contributo non trascurabile ai processi di ricarica degli acquiferi.

In questo contesto, è stato dimostrato che i terremoti ad elevata magnitudo possono cambiare il comportamento idrodinamico delle sorgenti mostrando variazioni transitorie o di lunga durata, rendendo la gestione della risorsa idrica molto impegnativa.

Questo è il caso di una ampia zona Appenninica dell'Italia centrale, quella interessata da una intensa sequenza sismica tra agosto 2016 e gennaio 2017, caratterizzata da nove eventi di $M_w \geq 5.0$ di cui il più forte di $M_w = 6.5$ con epicentro ubicato all'interno dell'area di studio.

Questa ricerca incrementa notevolmente le conoscenze idrogeologiche ed idrodinamiche relative alla circolazione delle acque sotterranee negli acquiferi carbonatici, approfondendo i possibili cambiamenti delle caratteristiche qualitative e quantitative del flusso idrico sotterraneo, durante e dopo una serie di forti eventi sismici. È stato utilizzato un approccio olistico combinato coinvolgente diverse discipline e tecniche, a partire da un'indagine idrogeologica più tradizionale basata sull'applicazione del bilancio idrologico e sull'analisi degli idrogrammi delle portate sorgentizie, validata mediante i risultati di numerosi test con traccianti artificiali, e analisi geochemiche ed isotopiche. Durante le indagini è stato effettuato un rilevamento idrogeostrutturale di dettaglio, integrato con un'analisi basata su sistemi geografici informativi, finalizzato sia alla identificazione delle strutture geologiche principali responsabili del cambiamento del flusso idrico sotterraneo, sia alla caratterizzazione delle varie idrostrutture presenti nell'area in studio e dei rapporti tra esse.

I risultati confermano la complessità geologico-strutturale di questa area e la ripercussione sull'assetto idrogeologico, evidenziando in particolare il ruolo che le faglie, ad estensione sia regionale che locale, giocano nelle modificazioni del flusso di acque sotterranee.

Le evidenze ottenute da ciascuna tecnica di indagine sono state confrontate tra loro per delineare un modello idrogeologico concettuale indispensabile per la corretta e sostenibile gestione della risorsa idrica sotterranea.

Parole chiave: acquiferi carbonatici; idrogeologia dei terremoti; prove di tracciamento; idrologia isotopica; bilancio idrogeologico; modelli concettuali.

Dedication

The three years spent in doctoral school were an important period of transition and both personal and professional growth.

The Covid-19 pandemic has marked severely this journey, but at the same time, the warmth of my family has been felt despite the situation.

This manuscript is a dedication to the family I come from and the one I will have.

To my big and one family.

Acknowledgements

The completion of this research could not have been possible without the expertise of my supervisor Prof. Alberto Tazioli and the support of Prof. Daniela Valigi.

I would also like to give special thanks to Prof. Carlo Cardellini, Prof. Francesco Mirabella, Dr. Costanza Cambi and Dr. Stefano Palpacelli for supporting the scientific discussions and models proposed during the various development steps of this research.

My warmest thanks to the curriculum coordinator Prof. Francesco Fatone, to all the staff of the “Dipartimento di Scienze e Ingegneria della Materia, dell'Ambiente ed Urbanistica” of Marche Polytechnic University, friends, and colleagues.

I would like to acknowledge the Dept. technicians in the person of Mirco Marcellini and Ing. Jonathan Domizi for supporting the field and laboratory operations, as well as Dr. Stefano Caliro of INGV Institute and Dr. Mario Mussi of CNR-IGG Institute for the isotopic analysis.

Many thanks to Paola Maria Vivalda and Torquato Nanni.

This thesis could not have been developed without the data provided by CIIP s.p.a., Tennacola s.p.a, Regione Umbria, Regione Marche Civil Protection Agency, whom I sincerely thank.

Special thanks to Dr. Elisa Mammoliti, for supporting me both personally and professionally.

Finally, I would like to thank the reviewers of this PhD thesis and the committee members.

List of figures

Figure 1.1 Carbonate aquifer distribution in Europe, modified form Goldscheider, 2005.

Figure 1.2 Carbonate aquifers distribution in Italy, modified form Lancia et al., 2020.

Figure 1.3 Reference models for karstic aquifers. a) Mangin, 1975 model b) Drogue, 1980 model A: highly permeable upper zone, B: blocks with low-permeability cracks with slow flow, C: high-permeability karst conduits with rapid flow; c) Doerfliger & Zwahlen, 1955 conceptual model.

Figure 1.4 Main isotopic fractionation processes occurring within the water cycle, modified from Fischer, 2012.

Figure 1.5 Iterative process for the conceptual modelling through the consensus or multi-model approach, from Enemark et al., 2019.

Figure 1.6 Flow-chart of the research method.

Figure 1.7 Stratigraphic column of geological formations belonging to the Umbria-Marche domain, modified from Deiana & Marchegiani, 2002.

Figure 1.8 Tectonic setting of the Umbria-Marche domain and location of Sibillini Mts. Massif, modified from Montone & Mariucci, 2020.

Figure 1.9 Geological map of the Sibillini Mts., modified from Pierantoni et al., 2013.

Figure 1.10 Location of main earthquakes occurred in the area during the 2016-2017 seismic sequence with pattern of primary surface ruptures observed for an overall length of 20-25 km along the Mt. Vettore - Mt. Bove fault system, modified from Emergeo W.G. et al., 2016.

Figure 1.11 Sketch of relationships between the simplified stratigraphy (left) and hydrogeological complexes (right) (not in scale, modified from (Viaroli et al., 2021). Ev: dolomite and evaporite sequence, CM: Calcare Massiccio, Co: Corniola, Ra: Rosso Ammonitico, Pos: Marne a Posidonia, CD: Diaspri, Bu: Bugarone, Mai: Maiolica, MF: Marne a Fucoidi, SB: Scaglia Bianca, SR: Scaglia Rossa, SV: Scaglia Variegata, SC: Scaglia Cinerea, Bi: Bisciaro, Cerr: Marne a Cerroghna, Sch: Schlier, Laga: Flysch della Laga. Hydrogeological complexes: TRI: Dolomitic-Evaporitic aquiclude, BAS: Basal aquifer, JUR Jurassic aquiclude, MAI: Maiolica aquifer, MF: Marne a Fucoidi aquiclude, SCA: Calcareous Scaglia aquifer, SVC: Upper aquiclude, Laga aquitard.

Figure 1.12 Map of conceptual hydrogeological models for each aquifer belonging to the Sibillini Mts., drawn by Boni et al., 2010.

Figure 1.13 Hydrostructural map of Umbria Region, published by Mastrotillo et al., 2009.

Figure 1.14 Hydrogeological map and cross-section of Sibillini Mts. – scale 1:50000, with detail of hydrostructures number (i.e., hydrostructure 1, 2 and 3) modified from Viaroli et al., 2021.

Figure 1.15 Hydrogeological conceptual model identified by Boni & Petitta, 2007, for the BAS aquifer modified by Mastrotillo et al., 2012.

Figure 1.16 Hydrogeological conceptual model for the Upper Nera River basin located in hydrostructure 2, modified from Tarragoni, 2006.

Figure 1.17 Piezometric map of Norcia Plain located in hydrostructure 3, from Petitta, 2011.

Figure 1.18 Main schematic traced flow paths, with hydrogeological and structural sectors, derived from the first tracer test (a), the second tracer test (b) and third artificial tracer tests (c). From Nanni et al., 2020.

Figure 2.1 Location of rainfall stations used to calculate Standard Precipitation Index (SPI).

Figure 2.2 Location of discharge monitoring points.

Figure 2.3 Discharge data for points 1, 2 and 3 located in the Hydrostructure 1. Principal vertical axis refers to point 2 while secondary vertical axis refers to point 1 and 3. The grey window indicates the co-seismic period (24th August 2016-18 January 2017).

Figure 2.4 Discharge data for points 4 and 5 located in the Hydrostructure 2a. The grey window indicates the co-seismic period (24th August 2016-18 January 2017).

Figure 2.5 Discharge data for points 6, 7 and 8 located in the Hydrostructure 2 and belonging to the Upper Nera River Basin up stream to Visso village. The solid lines referred to the continuous data, while the dots refer to discrete discharge measurements. The grey window indicates the co-seismic period (24th August 2016-18 January 2017).

Figure 2.6 Discharge data for points 9, 10 and 11. The points 9 and 10 are located in the Hydrostructure 2 while the point 11 is located in the Hydrostructure 3. All the points belong to the

Nera River Basin downstream to Visso village. The solid lines referred to the continuous data, while the dots refer to discrete discharge measurements.

Figure 2.7 Discharge data for points 12 and 13. The points are located in the Hydrostructure 3 and belong to the Campiano stream. The solid line refers to the continuous data, while the dots refer to discrete discharge measurements. The grey window evidences the co-seismic period (24th August 2016-18 January 2017).

Figure 2.8 Discharge data for points 14, 15, 16, 17 and 18. The points are located in the Hydrostructure 3 and belong to the Sordo River Basin. The solid lines refer to the continuous data, while the dots refer to discrete discharge measurements. The grey window evidences the co-seismic period (24th August 2016-18 January 2017).

Figure 2.9 Discrete discharge data for points 19 and 20. The points are located in the Hydrostructure 3 and belong to the Corno River Basin.

Figure 2.10 Continuous discharge data for point 21. The point is in the Hydrostructure 2 in the Pian Grande Plain (Mèrgani sinkhole).

Figure 2.11 Architecture of WaterbalANce WebApp to perform the hydrological balance computation.

Figure 2.12 Intersection between the slope map (a) obtained by the DTM (20 m x20 m cell size), and the hydrogeological complexes (b).

Figure 2.13 Flowchart of indirect (Turc vs WaterbalANce) and direct hydrological balance.

Figure 2.14 Representation of a spring flow discharge graph with nomenclature referring to the various parts, modified from Martin & White, 2008.

Figure 2.15 Pian Grande Plain with a detail of polje.

Figure 2.16 Historical map of Pian Grande Plain area with typical karstic forms. The Mèrgani sinking stream is evidenced by the red circle. Modified from Giovagnotti, 1979

Figure 2.17 Artificial tracer injection operations conducted during TEST4 on 20th March 2018 in Mèrgani sinking stream.

Figure 2.18 Field fluorometric probes used for artificial tracer tests: (a) Albillia GGUN-FL24 and GGUN-FL30 produced by Albillia Co. (Neuchatel, Switzerland) and (b) PME Cyclops-7 Logger (PME Inc. USA).

Figure 2.19 RF-6000 laboratory spectrofluorometer produced by Shimadzu Corporation (Kyoto, Japan).

Figure 2.20 Calibration curve of Shimadzu RF-6000 laboratory spectrofluorometer.

Figure 2.21 (a) Input data to perform SWCCF (rainfall and snow depth) (b) output data to perform SWCCF (springs discharge data). The grey window refers to the co-seismic period (24th August 2016 – 18th January 2017).

Figure 2.22 Map of hydrochemical-isotopic monitoring points.

Figure 2.23 Hydrochemical-isotopic sampling.

Figure 2.24 Snow cover sampling field operations, performed near Mt. Prata rainfall/snow depth station.

Figure 2.25 Graphical example of geo-isotopic modelling. The graph shows the hydrogeological complexes $\delta^{18}\text{O}$ distribution obtained by the GIS clipping procedure for each analysed spring. The relative percentage to the total can be calculated overlapping the average spring $\delta^{18}\text{O}$ content $\pm \sigma$.

Figure 3.1 SPI-12 obtained for the selected rain gauges representing the different sectors of the Sibillini Mts. massif, having continuous data between 1991 and 2021.

Figure 3.2 SPI-12 obtained for the Montmonaco rain gauge (black line) representing the Adriatic side of the Sibillini Mts. massif, and for the Vallo di Nera rain gauge (blue line) representing the Tyrrhenian side of the Sibillini Mts. massif.

Figure 3.3 SPI-12 obtained for the Arquata del Tronto rain gauge.

Figure 3.4 Linear regression between mean annual temperature and elevation. (a) Time interval 2009-2020 (b) Time interval 2015-2020.

Figure 3.5 Linear regression between mean annual rainfall and elevation for the whole area. (a) Time interval 2009-2020 (b) Time interval 2015-2020.

Figure 3.6 Linear regression between mean annual rainfall and elevation for the Adriatic side. Time interval 2015-2020.

Figure 3.7 Linear regression between mean annual rainfall and elevation for the Northern Tyrrhenian side. Time interval 2015-2020

Figure 3.8 Linear regression between mean annual rainfall and elevation for the Southern Tyrrhenian side. Time interval 2015-2020.

Figure 3.9 Mean annual temperature distribution (2015-2020) over the study area with detail of the hydrostructures (i.e., 1, 2, 2a and 3).

Figure 3.10 Mean annual rainfall distribution (2015-2020) over the study area with detail of the spatialization sides and hydrostructures (i.e., 1, 2, 2a and 3).

Figure 3.11 Linear regression between mean annual water surplus and elevation for the whole area. Time interval 2015-2020.

Figure 3.12 Linear regression between mean annual water surplus and elevation for the Adriatic side. Time interval 2015-2020.

Figure 3.13 Linear regression between mean annual water surplus and elevation for the Northern Tyrrhenian side. Time interval 2015-2020.

Figure 3.14 Linear regression between mean annual water surplus and elevation for the Southern Tyrrhenian side. Time interval 2015-2020.

Figure 3.15 (a) Comparison between PET calculated with Thornthwaite-Mather method and Eto calculated with Hargreaves & Samani method over the analysed time span; (b) Correlation between the variables with Pearson determination coefficient and MAE.

Figure 3.16 Detail of hydrostructure 1 and the discharge monitoring points.

Figure 3.17 Detail of hydrostructure 2 and part of hydrostructure 3 (Upper Nera River basin) with the discharge monitoring points.

Figure 3.18 Detail of hydrostructure 2a and the discharge monitoring points.

Figure 3.19 Detail of hydrostructure 3 and the discharge monitoring points along the Campinao stream, the Sordo and Corno Upper Basins.

Figure 3.20 Discharge data of points 12 (blue dots), 13 (red line - continuous data), 14 (green dots) after the 30th October 2016, earthquake. The dashed blue line represents the possible discharge trend recorded in point 12 just after the seismic sequence, having a similar behaviour with respect to the point 14.

Figure 3.21 Exponential fitting (black dashed line) of the baseflow recession curve (blue line) for point 15 (years 2019-2020). The red dashed line represents the pre-seismic mean annual spring discharge.

Figure 3.22 (a) Discharge increase measured between point 16 and point 17; (b) discharge increase measured between point 17 and point 18; (c) overall discharge increase measured between point 16 and point 18 vs local daily rainfall. The red dashed lines represent the pre-seismic mean annual discharge increase between the monitoring points.

Figure 3.23 On the right schematic map of discharge modifications recorded within each hydrostructure. On the left a picture of a dry spring located 854 m a.s.l. within hydrostructure 3. The dots refer to punctual springs (red: dry, yellow: discharge decreasing, green: discharge increasing), while the green lines represent the increasing of discharge due to the seismic sequence recorded along the linear springs.

Figure 3.24 (a) Continuous discharge data (m³/s) used to investigate any significant hydrogeological modification both in the short- and long-term periods. The grey window represents the co-seismic period (24th August 2016, 18th January 2017). (b) SPI-12 of Arquata del Tronto rain gauge between November 2013 and December 2020.

Figure 3.25 The depletion coefficients, α , for the years 2003–2019. Blue lines and dots indicate pre-earthquakes recession periods, red lines and dots are related to post-earthquake periods.

Figure 3.26 The depletion coefficients, α , for the years 2003–2019. Blue lines and dots indicate pre-earthquakes recession periods, red lines and dots are related to post-earthquake periods.

Figure 3.27 The depletion coefficients, α , for the years 2010, 2011 and 2012 (pre-seismic period) and 2017 to 2020 (post-seismic period). Blue lines and dots indicate pre-earthquakes recession periods, red lines and dots are related to post-earthquake periods.

Figure 3.28 The depletion coefficients, α , for the years 2015 and 2016 (pre-seismic period) and 2017, 2018-2019 (post-seismic period). Blue lines and dots indicate pre-earthquakes recession periods, red lines and dots are related to post-earthquake periods.

Figure 3.29 Tracer breakthrough curve (BTC) for Capotenna spring (hydrostructure 1) and spring discharge data recorded during TEST5. The green line refers to Fluorescein.

Figure 3.30 Tracer breakthrough curves (BTCs) for Foce spring (hydrostructure 1) and spring discharge data recorded during (a) TEST1, TEST2 and TEST3, modified from Nanni et al., 2020 (b) TEST4 (c) TEST5 and (d) TEST6. The green lines refer to Fluorescein, while the light blue lines refer to Tinopal CBS-X.

Figure 3.31 Tracer breakthrough curves (BTCs) for Capodacqua spring (hydrostructure 2a) and spring discharge data recorded during (a) TEST1, TEST2 and TEST3, modified from Nanni et al., 2020 (b) TEST5. The green lines refer to Fluorescein, while the blue lines refer to Tinopal CBS-X.

Figure 3.32 Tracer breakthrough curves (BTCs) for Pescara del Tronto spring (hydrostructure 2a) and spring discharge data recorded during (a) TEST2 and TEST3, modified from Nanni et al., 2020 (b) TEST4 and (c) TEST6. The green lines refer to Fluorescein, while the blue lines refer to Tinopal CBS-X.

Figure 3.33 Tracer breakthrough curves (BTCs) for the monitoring points in the Upper Sordo River basin (i.e., San Martino spring, Torbidone spring and Sordo linear springs- hydrostructure 3) and discharge data recorded during TEST1, TEST2 and TEST3, modified from Nanni et al., 2020. The blue lines and dots refer to Tinopal CBS-X.

Figure 3.34 Tracer breakthrough curves (BTCs) for Sordo linear springs (hydrostructure 3) and spring discharge discrete data recorded during (a) TEST4 and (b) TEST. The green lines refer to Fluorescein.

Figure 3.35 Tracer breakthrough curves (BTCs) for Torbidone spring (hydrostructure 3) and spring discharge data recorded during (a) TEST4 and (b) TEST5. The green lines refer to Fluorescein.

Figure 3.36 Tracer breakthrough curve (BTC) for San Martino spring (hydrostructure 3) and spring discharge data recorded during TEST5. The green line refers to Fluorescein.

Figure 3.37 Tracer breakthrough curves (BTCs) for Campiano spring (hydrostructure 3) and spring discharge discrete data recorded during (a) TEST4 and (b) TEST5. The green lines refer to Fluorescein.

Figure 3.38 Tracer breakthrough curves (BTCs) for Nera i Molini linear springs (hydrostructure 2) and discharge data recorded during (a) TEST1, TEST2 and TEST3, modified from Nanni et al., 2020 (b) TEST4. The green lines and dots refer to Fluorescein, while the blue lines refer to Tinopal CBS-X.

Figure 3.39 Tracer breakthrough curves (BTCs) for Nera at Castelsantangelo-San Chiodo spring (hydrostructure 2) and spring discharge data recorded during (a) TEST1 and TEST3, modified from Nanni et al., 2020 (b) TEST4 (c) TEST5 and (d) TEST6. The green lines refer to Fluorescein, while the light blue lines refer to Tinopal CBS-X.

Figure 3.40 Tracer concentration measured in punctual samples during TEST1, TEST2, TEST4 and TEST5. The principal vertical axis refers to the yellow symbols, while the secondary vertical axis refers to the green ones.

Figure 3.41 (a) Location of the monitoring points, the new mapped fault and main tracer direction from the injection point (Mèrgani sinking stream in Pian Grande Plain) to the analysed basins (Castelsantangelo and Ussita). The green dots refer to the point where the tracer has been recovered (point 8 and 9) while the red crosses refer to the monitoring point where the tracer has not been detected in this study (point 6, 7, 11 and 12). The monitoring point 10, 13 and 14 are not instrumented for tracer detection. (b) Outcrop of the new mapped fault with its kinematics represented in the stereo plot.

Figure 3.42 Illustration of pre-seismic (TEST1 and TEST2) traced groundwater flow directions from the injection point (Mèrgani sinking stream). The monitoring points show the filed instrument to detect tracer arrivals. The empty symbols refer to the pints where tracer has never been recorded. The results are synthetized from the data published by Nanni et al., 2020 and integrated by the results analysed in this study.

Figure 3.43 Illustration of post-seismic (TEST4 and TEST5) traced groundwater flow directions from the injection point (Mèrgani sinking stream). The monitoring points show the filed instrument to detect tracer arrivals. The empty symbols refer to the points where tracer has never been recorded.

Figure 3.44 (a) Maximum tracer velocity vs distance from injection point for TEST4; (b) Maximum tracer velocity vs distance from injection point for TEST5; (c) Mean tracer velocity vs distance from injection point for TEST4; (d) Mean tracer velocity vs distance from injection point for TEST5.

Figure 3.45 Comparison between maximum tracer concentration expressed in ppb vs days after injection in the monitoring points (Pescara del Tronto, Capodacqua and Foce springs) and rainfall recorded at Monte Prata during the five analysed tests. The green area represents the Fluorescein, while the blue one the Tinopal CBS-X. When the fluorometric probe was not installed in a specific spring during a specific test, the corresponding graph is represented in light grey, and the expression “No Probe” is added in the figure.

Figure 3.46 Contour graphs that represent variations in the curve shape of the snow cover/ raw spring discharge CCFs, over the whole period (i.e., among the sliding windows shown in (Table 2.8). (a) Mt. Prata snow depth/Pescara del Tronto spring raw discharge (b) Mt. Prata snow depth /Capodacqua spring raw discharge (c) Mt. Prata snow depth /Foce spring raw discharge. The positive correlation value corresponding to the 95% significance level ($p < 0.05$) is 0.06.

Figure 3.47 Contour graphs that represent variations in the curve shape of the high-elevation rainfall/residual spring discharge CCFs, over the whole period (i.e., among the sliding windows shown in (Table 2.8). (a) Mt. Prata rainfall/Pescara del Tronto spring residual discharge (b) Mt. Prata rainfall /Capodacqua spring residual discharge (c) Mt. Prata rainfall /Foce spring residual discharge. The positive correlation value corresponding to the 95% significance level ($p < 0.05$) is 0.06.

Figure 3.48 Contour graphs that represent variations in the curve shape of the low-elevation rainfall/residual spring discharge CCFs, over the whole period (i.e., among the sliding windows shown in (Table 2.8). (a) Capodacqua rainfall/Pescara del Tronto spring residual discharge (b) Capodacqua rainfall /Capodacqua spring residual discharge (c) Montemonaco rainfall /Foce spring residual discharge. The positive correlation value corresponding to the 95% significance level ($p < 0.05$) is 0.06.

Figure 3.49 (a) Meteoric $\delta^{18}\text{O}$ distribution above the study area: 1- $\delta^{18}\text{O}$ -elevation relation determined in this study, 2- $\delta^{18}\text{O}$ -elevation determined by Giustini et al., 2016, 3- $\delta^{18}\text{O}$ -elevation determined by Tarragoni, 2006. (b) Results of statistic tests on the Meteoric $\delta^{18}\text{O}$ distribution above the study area, group 1 (our relationship) is significantly different from the literature ones (groups 2 and 3). (c) Linear regression between $\delta^{18}\text{O}$ and elevation determined by eight rain gauges' water

collectors and eleven minor springs of the area obtained in this study (red solid line) and linear regression obtained by Tarragoni, 2006 for the same area (black dashed line).

Figure 3.50 Application of the inverse $\delta^{18}\text{O}$ -elevation relation obtained in this study (red line and red dots), the one obtained by Giustini et al., 2016 (green dashed line and green dots) and the one presented by Tarragoni, 2006 (black dashed line and black dots), to the mean pre-seismic $\delta^{18}\text{O}$ value of Sordo (20), Pescara (4), Capodacqua (5), Foce (2) and Nera Castelsantangelo (8).

Figure 3.51 Simplified hydrogeological map with isotopic and hydrochemical sampling points.

Figure 3.52 Streams topographic intersection with the hydrogeological complexes and localization of the monitoring points for the analyzed basins: (a) Castelsantangelo stream (b) Ussita stream.

Figure 3.53 Plot of the chemical composition of all the monitoring points on the Langelier-Ludwig diagram. (b) Detail of the Langelier-Ludwig of referring to the box reported in (a).

Figure 3.54 Results of the isotopic data processing. The blue dots refer to the pre-seismic period, the red dots refer to the post-seismic period. On the right raw springs $\delta^{18}\text{O}$ data with pre-seismic $\delta^{18}\text{O}$ mean value (black solid line) $\pm \sigma$ (black dashed lines). On the right the results of findchangepts performed in MATLAB®.

Figure 3.55 Recharge area for each spring determined by the hydrogeological setting and the mean $\delta^{18}\text{O}$ values.

Figure 3.56 Meteoric $\delta^{18}\text{O}$ distribution for each hydrogeological complex within each spring recharge area.

Figure 3.57 Hydrogeological complexes $\delta^{18}\text{O}$ distribution outcropping in each recharge area for each spring: (a) Capodacqua spring (b) Pescara del Tronto spring (c) Foce spring (d) Nera San Chiodo-Castelsantangelo (e) Sordo spring. For each area the spring $\delta^{18}\text{O}$ content $\pm \sigma$ is plotted in the chart bar (blue dot: pre-seismic period; red dot: post-seismic period).

Figure 3.58 (a) Snow cover isotopic sampling phase. (b) Snow isotopic characterisation performed in the winter 2012-2013. (c) Snow isotopic characterisation performed in the winter 2020-2021.

Figure 3.59 $\delta^{18}\text{O}$ data for the Mèrgani sinking stream (blue dots) vs the snow depth recorded in Mt. Prata station expressed in cm. The black solid line represents the Mèrgani water $\delta^{18}\text{O}$ mean value, while the black dashed lines represent $\pm \sigma$ interval.

Figure 3.60 Snow density recorded at three different snow depth during the melting phase of the winter 2020-2021.

Figure 4.1 Schematic hydrogeological cross-section involving Pian Grande Plain and Pescara del Tronto spring. The hydrodynamic groundwater divide between hydrostructure 2 and 2a is determined by the hinge of the anticline. The colours of the hydrogeological complexes refer to the ones reported in Figure 1.11, modified from Fronzi et al., 2020.

Figure 4.2 Schematic hydrogeological cross-sections with main tectonic lineaments (i.e., Normal faults and Sibillini Mts. Overthrust). The sketch (not in scale) represents the position of the piezometric level hypothesised before and after the seismic sequence with a detail of the shift of hydrodynamic groundwater divide due to earthquakes effects.

Figure 4.3 Hydrogeological cross-section (SSE-NNW oriented) modified from Viaroli et al., 2021. The colours in section refer to the sketch of hydrogeological complexes reported in Figure 1.11.

Figure 4.4 Hydrogeological cross sections with the identification of the new mapped fault and the hypothetic elevation (m a.s.l.) of water table. The colours in section refers to the sketch of hydrogeological complexes reported in Figure 1.11. From Fronzi et al., 2021.

Figure 4.5 Hydrogeological cross-section (SE-NW oriented) modified from Nanni et al., 2020 with the hypothetic elevation of water table and the monitoring points along the Castelsantangelo stream. The vertical scale is exaggerated to better visualize the piezometric gradient, the colours in section refers to the sketch of hydrogeological complexes reported in Figure 1.11.

Figure 4.6 Groundwater sample (black dots) and rainfall samples (red dots) d-excess values on the left, and their distribution on the right.

Figure 4.7 Hydrogeological map for Capodacqua and Pescara del Tronto springs modified from Celico, 2011. The red circle evidences the tectonic contact between BAS (rose color), MAI (lighy grey) and SCA (dark grey), while the red cross highlights the interruption of SW-NE oriented groundwater flow direction after the seismic sequence.

Figure 4.8 3D illustration of hydrostructure 2a (Capodacqua and Pescara del Tronto springs) and Pian Grande Plain. The increase of recharge area elevation for Pescara del Tronto and Capodacqua spring obtained by the water isotopic composition may derive from the decrease of groundwater amount coming from Pian Grande Plain (1300 m a.s.l.). Modified from Dragoni et al., 2003.

Figure 4.9 Schematic conceptual model with hydrogeological cross-sections (not in scale). In legend: (1) tracer injection point (i.e., Mèrgani sinkhole); (2) main traced paths; (3) minor traced paths; (4) Mt. Vettore - Mt.Bove normal fault; (5) co-seismic ruptures; (6) Sibillini Mts. Overthrust; (7) groundwater flow direction; (8) groundwater flow direction (perpendicular to the reader, entering in the plane); (9) spring; (10) rainfall infiltration; (11) piezometric levels; (12) carbonate fissured and fractured rocks; (13) Laga geological Formation (LAG). Modified from Fronzi et al., 2020.

Figure 4.10 Pre- and post-seismic planimetric conceptual models with location of main springs, basal groundwater flowpaths and main tectonic lineaments. The boundaries between the hydrostructures (i.e., Mt. Vettore- Mt. Bove and Nottoria-Preci faults) are represented by the black dashed lines, while the blue dashed lines represent the groundwater divides.

Figure 4.11 Simplified overlook of the pre- and post-seismic conditions schematised as a block diagram. The groundwater exchanges between different hydrostructures are represented by the black arrows. *Discharge data integrated with mean spring discharge values published in Viaroli et al., 2021 and in Valigi, 2021.

Figure 5.1 Simplified scheme illustrating the main pillars involved in this project, following and modified from De la Torre et al., 2020.

List of tables

Table 1.1 Historical strong earthquakes affecting the study area.

Table 2.1 Meteo-climatic monitoring network. Blue colour refers to the Marche Region stations; yellow colour Umbria Region; green colour Lazio Region.

Table 2.2 Classification of drought conditions based on the SPI.

Table 2.3 Characteristics of rainfall stations used to calculate Standard Precipitation Index (SPI).

Table 2.4 Details of discharge monitoring network.

Table 2.5 Mean slope (°) values determined for each hydrogeological complex within each hydrostructure.

Table 2.6 Details of artificial tracer tests performed in the Sibillini Mts. area.

Table 2.7 Details of monitoring points selected to perform the transient time series analysis via SWCCF.

Table 2.8 Details of windows to perform the transient time series analysis via SWCCF.

Table 2.9 Details of meteoric water isotopic sampling network. *Rainfall and snow cover sampling.

Table 3.1 Synthesis of hydrogeological balance for each hydrostructure. Δ (m³/s) represents the discharge difference between pre- and post-seismic periods.

Table 3.2 Pre- and post-seismic mean annual discharge measured in point 1.

Table 3.3 Pre- and post-seismic mean annual discharge measured in point 2.

Table 3.4 Pre- and post-seismic mean annual discharge measured in point 3.

Table 3.5 Pre- and post-seismic mean annual discharge measured in point 6.

Table 3.6 Pre- and post-seismic mean annual discharge measured in point 7.

Table 3.7 Pre- and post-seismic mean annual discharge increase measured between point 9 and point 10.

Table 3.8 Pre- and post-seismic mean annual discharge increase measured between point 10 and point 11.

Table 3.9 Pre- and post-seismic mean annual discharge measured in point 4.

Table 3.10 Pre- and post-seismic mean annual discharge measured in point 5.

Table 3.11 Pre- and post-seismic mean annual discharge increase measured between point 12 and point 13.

Table 3.12 Pre- and post-seismic mean annual discharge measured in point 14.

Table 3.13 Pre- and post-seismic mean annual discharge measured in point 15.

Table 3.14 Synthesis of discharge increase recorded along the Sordo river between point 16 and 17, between point 17 and 18, and the overall discharge increase between point 16 and 18 referred to four different periods.

Table 3.15 Pre- and post-seismic mean annual discharge increase measured between point 19 and point 20.

Table 3.16 Basic statistics calculated for the analysed springs during the pre-, co-, and post-seismic periods.

Table 3.17 Maillet depletion coefficients α and values of Q_0 for the analyzed recession periods. Data under the red line are related to post 30 October 2016 recession periods.

Table 3.18 Maillet depletion coefficients α and values of Q_0 for the analyzed recession periods. Data under the red line are related to post 30 October 2016 recession periods.

Table 3.19 Maillet depletion coefficients α and values of Q_0 for the analyzed recession periods. Data under the red line are related to post 30 October 2016 recession periods.

Table 3.20 Maillet depletion coefficients α and values of Q_0 for the analyzed recession periods. Data under the red line are related to post 30 October 2016 recession periods.

Table 3.21 Maillet depletion coefficients α and values of Q_0 for the analyzed recession periods. Data under the red line are related to post 30 October 2016 recession periods.

Table 3.22 Maillet depletion coefficients α and values of Q_0 for the analyzed recession periods. Data under the red line are related to post 30 October 2016 recession periods.

Table 3.23 Results of TEST4 and TEST5 obtained by Qtracer2, the numbers refer to the ones in plate of Figure 3.42.

Table 3.24 Mean and standard deviation values for the elevation of each basin (Castelsantangelo and Ussita) and for each aquifer outcropping within the basins determined by the GIS analysis.

Table 3.25 Pre- and post-seismic SO_4^{2-} and $\delta^{18}\text{O}$ ($\pm \sigma$) with the elevation determined by the $\delta^{18}\text{O}$ - elevation obtained in this study. The trend column refers to the variation of the aforementioned parameters between the pre- and post-seismic conditions: + sulphates increase in the post seismic, - sulphates decrease in the post-seismic, $\uparrow$ recharge area mean elevation increase, $\downarrow$ recharge area mean elevation decrease, = sulphates content or recharge area mean elevation not varied between the pre- and post-seismic condition.

Table 3.26 Hydrogeological complexes (relative percentage to the total) that can potentially feed the analysed springs, obtained by the overlapping between the complexes $\delta^{18}\text{O}$ distribution and the average springs $\delta^{18}\text{O}$ content $\pm \sigma$.

Table 4.1 Tracer velocity range min-max (m/day) estimated by the free software Qtracer2 (US EPA 2002) for the five tracer tests results coupled with the SWCCF.

1 Introduction

1.1 General outline

Nowadays groundwater is considered one of the most important natural resources for world population. The Water Framework Directive 2000/60/EC establishes the actions in the field of water policy for the European Community. According to that, many water protection and management plans have been developed by the authorities in the last decades with the aim to achieve the quali-quantitative targets for the protection of water resources and water bodies, ensuring a long-term sustainable water supply. From a hydrogeological perspective, it is often necessary to support the authority's decisions, deepening the knowledge on groundwater quantity, movement, and dynamics in specific areas, in order to cope with needs related to agriculture, drinking water use and hygienic-sanitary issues (Asadi et al., 2020). In this context, the mountainous aquifers represent the natural water towers of our planet, providing good quality drinking water to most of worldwide population (Messerli et al., 2004), despite little research had been done on them since most of the hydrogeological information are often available in lowland alluvial areas (de Graaf et al., 2015). In many regions, water supplies are already insufficient to satisfy industrial, urban and agricultural demands (Vaux, 2011) and the water scarcity is one of the major problems that the scientific community have to face today and in the near future (Liu et al., 2017). Groundwater resources management is nowadays one of the most important pillars in our society and the study of regional-scale variations induced by anthropic pressure and climate change involve several research groups and disciplines (Dragoni & Sukhija, 2008; Thornton et al., 2014). Earthquake-induced variations in hydrogeological system, as well as their influences on groundwater quality and availability are worldwide recognised (Ingebritsen & Manga, 2019). However, due to the lack of well-developed hydrogeological monitoring systems in most of the cases and due to the absence of a robust hydrological and hydrochemical dataset prior to seismic sequences, this research field is still characterised by open questions. Many

models have been proposed during the years to explain the hydrogeological changes induced by earthquakes, but each case study seems to have its unique answer and the mechanisms responsible for hydrological modifications are often uncertain (Korup et al., 2021). Considering all these observations, the present PhD research work is developed by inheriting decades of previous knowledge of one of the most studied areas of central Italy from the geological standpoint, that of the central Apennines. The reasons to study in deep this area born afterward the Amatrice-Norcia 2016-2017 seismic sequence. In fact, as an important Latin proverb says “Mater artium necessitas”. After the seismic sequence, a series of detailed studies of tectonics, structural geology, seismicity and hydrogeology begun (Emergeo W.G. et al., 2016; Petitta et al., 2018; Porreca et al., 2018). However, these studies have often different objectives and, due to the high complexity of some areas, are not suitable to be used to build a robust hydrogeological conceptual model.

These studies gave a new impetus on the deepening of the groundwater circulation of a complex carbonate domain located in a wide area which comprises the Sibillini Mts., affected by more than nine $M_w \geq 5.0$ earthquakes in the last decade. Direct hydrogeological investigations conducted between the years 2016-2021 have been analysed and compared with the previous studies in the area, investigating the role of faults before, during and after the seismic period. The overall objectives of this thesis include a definition of a new hydrogeological conceptual model of the area and, understanding the hydrogeological changes that have occurred by integrating several hydrogeological investigation techniques including artificial tracer tests and isotope hydrology. The more specific objectives of this research are reported in section 1.3 of this manuscript.

1.2 Synthesis: the state of the art and the state of practice

In this section will be described existing academic and research methods related to:

1. Carbonate, karstic and fissured aquifers
2. Methods of tracing techniques
3. Hydrological modification due to seismic sequence
4. Hydrological conceptual models

1.2.1 Carbonate, karstic and fissured aquifers

Carbonate rocks cover approximately 10 % of the ice-free continental areas and supply about a quarter of the world population with drinking water (Ford & Williams, 1989). Only in Europe more than 35 % of the land surface is occupied by carbonate terrains (Figure 1.1) (Goldscheider, 2005).



Figure 1.1 Carbonate aquifer distribution in Europe, modified form Goldscheider, 2005.

Some European countries, such as Austria and Slovenia depend on carbonate aquifers for about 50% of their overall drinking water supply (Kovács, 2003). In Italy the main carbonate aquifers are located in the central and southern portion of the Apennines chain (Figure 1.2) and provide good quality groundwater for the population living right up to the coastal areas. However, Preziosi & Piccini, 2000 estimate that the amount of renewable water resources linked to the Italian karst aquifers is 45 km³ and this value is approximately three times the consumption of drinking water in the same country, proving that these aquifers are a very important resource, only partially exploited.



Figure 1.2 Carbonate aquifers distribution in Italy, modified from Lancia et al., 2020.

A large proportion of carbonate rocks underwent karstification processes during geological times. Consequently, together with alluvial plains, karstic rocks may be one of the most important aquifer formations in the world (Bakalowicz, 2005). At the same time, karst aquifers are often highly vulnerable to contamination and are impacted by a wide range of human activities (Drew & Holtz, 1999); for this reason, the investigation and water quality protection are essential for preserving their “natural health”. In this context, some European Action, such as the COST Action 620, dealing with specific vulnerability and risk assessment for carbonate (Karst) aquifers have been developed (Andreo et al., 2009). Community and scientific attention on carbonate aquifers is not merely focused on water quality aspects, but several research groups are trying to explore the hydrodynamic characteristics of these complex systems and quantify the availability of water resources provided by them even under increasing anthropic pressure and global climate change (Younger et al., 2002; Hartmann et al., 2014; Martos-Rosillo et al., 2015; Earman & Dettinger, 2011).

The most important characteristics of karst aquifers, that differentiate them from any other kind of hydrogeological system, is the high flow heterogeneity. Indeed, the water interacting with host rock can react dissolving the carbonates around interconnected voids enlarging fissures and fractures, increasing the primary porosity of the matrix and thus the hydraulic conductivity (Kiràrly et al., 2002).

The dissolving processes and the karstification degree is strictly related to the direction and magnitude of groundwater flux density vector (Kiràrly et al., 1971). As a result, carbonate rocks

acquire a heterogeneous range of voids, of different origin, capable to affect groundwater storage and movement within the aquifers.

During the past decades many models were proposed to simplify and explain the karst aquifers storage and movement from the recharge areas to the springs, and they are all supported by those historically developed by Mangin, 1975; Drogue, 1974, 1980; Király 1975, 2002 and Király et al., 1995.

In Mangin, 1975 the concept of the epikarst was introduced, consisting in a shallow, high permeability karstified zone below the aquifer surface, where water rapidly drains through enlarged vertical shafts, reaching the saturated zone, and contributing at a diffuse recharge. Besides, the main conduit system transmits infiltration water towards a karst spring but is poorly connected to large voids in the adjacent zones, resulting a well-defined horizon where active conduits develops and where most of the flow take place (Figure 1.3a).

Drogue, 1974, 1980 hypothesised that the original rock fracture pattern influences the developing of the karst network. This network consists of fissured blocks with a size in the order of several hundred meters, separated by high-permeability low storage conduits. Every block is dissected by small scale fissures or fracture with considerably larger storage and low bulk permeability. The uppermost part of the aquifer is responsible for enhanced infiltration of meteoric water due to its intensely fissured pattern for decompression phenomena (Figure 1.3b).

Király, 1972, 2002 and Király et al., 1995, combining the previous assumptions, developed a new model, then synthesized by Doerfliger & Zwahlen, 1995. This model involves the epikarst and a hierarchical organization of the conduit networks. It also comprises the hydraulic effect of nested groups of different discontinuities. This means that the aquifer permeability is determined by the fracture intersections, and the effective porosity is determined by fractures plains. Carbonate aquifers can be considered as interactive units of high-conductivity hierarchically organised karst channel network with a low-permeability fissured rock matrix (Figure 1.3c).

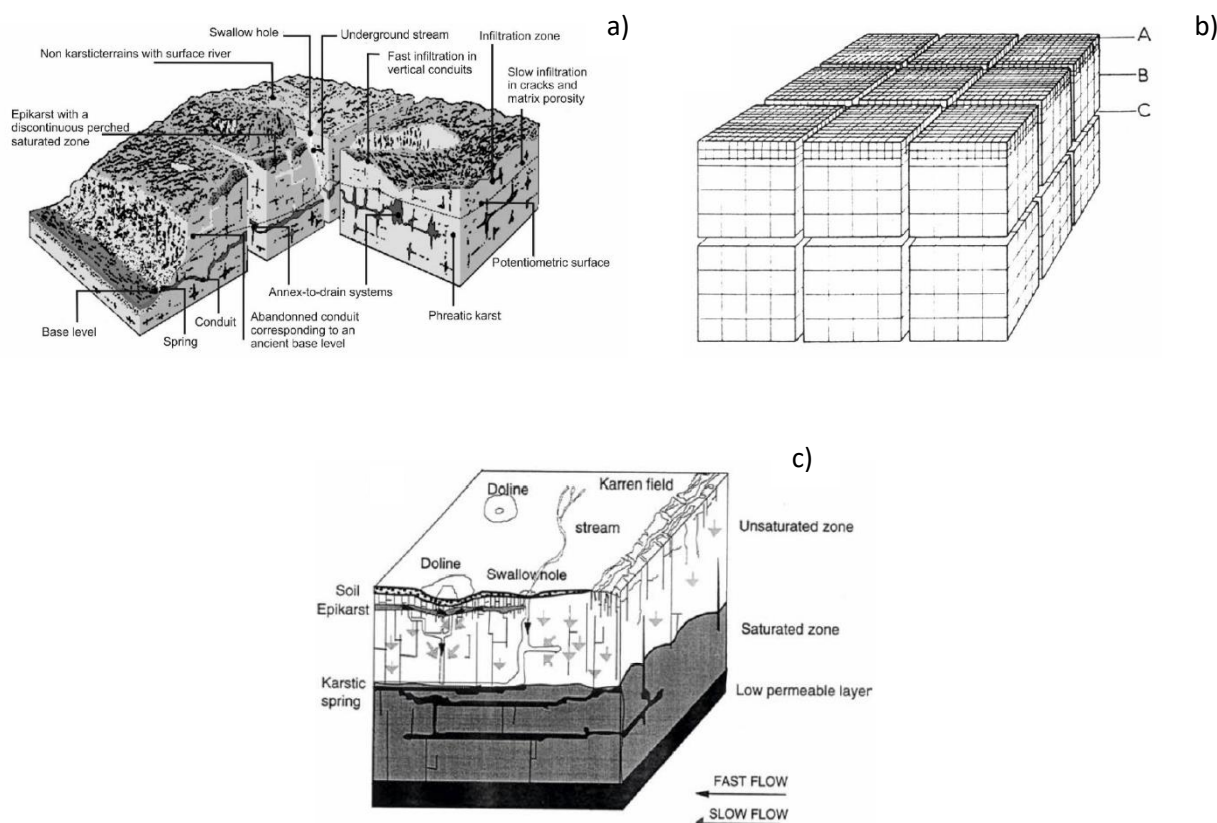


Figure 1.3 Reference models for karstic aquifers. a) Mangin, 1975 model b) Drogue, 1980 model A: highly permeable upper zone, B: blocks with low-permeability cracks with slow flow, C: high-permeability karst conduits with rapid flow; c) Doerfliger & Zwahlen, 1955 conceptual model.

The work recently published by Kovács, 2021 underlain that many karst classifications exist from a geomorphological point of view, while limited and contrasting definitions are referred to the term karst aquifer. Moreover, the existing definitions and classifications, lack of quantitative aspects and are focused only on visible geological and geomorphological aspects.

During the years many efforts were made to provide integral information about karstic systems from a hydrological point of view, based on the assumption that the measurement of discharge or hydraulic head versus time in a spring or monitoring well of a karst aquifer offers important information about the geometry, the hydraulic properties and the hydrodynamic functioning of that system (Burger, 1956; Berkaloﬀ, 1967; Forkasiewicz & Paloc, 1967; Drogue, 1967; Mangin, 1975; Kiraly & Morel, 1976; Kiraly, 2003; Shevenell, 1996; Kresic & Bonacci, 2010; Kovács, 2003; Kovács et al., 2005, 2015; Kovács & Perrochet, 2008, 2014). This assumption led to the application of hydrograph analysis as the most widely used method for defining these characteristics. In this regards, two analytical methods can be recognised (Jeannine & Sauter, 1988): i) Time series analysis

and ii) Single-event methods. The former, based on the investigation of a succession of recharge events, provide only some physical aspects of the functioning of the hydrogeological system. The latter, dealing with the global hydraulic response of a spring to a single recharge event can give important information such as the spring base-flow and quick-flow amount.

Among the time series analysis approaches, univariate and bi-variate methods are recognised. The first able to identify the cyclic variations, the latter very suitable to investigate the relationship between input (recharge) and output (discharge).

Within single event methods, the master recession curve (MRC) is probably one of the most known and used, and it is capable to provide an average characterization of baseflow response (Kresic, 2007).

However, both the methods, if not coupled with reservoir models or analytical solutions, gives only qualitative information about hydrodynamic behaviour. For this reason, since the early 1900s mathematical models of baseflow recession have been developed (Boussinesq, 1904; Maillet, 1905) and further studies improved those simplify models (Forkasiewicz & Paloc, 1967).

In recent decades, thanks to the increasing evolution of computational devices, the hydrogeological modelling has been addressed by complex numerical models that can simulate both the hydrodynamic and hydrogeological behaviour of the systems (Zdechlik, 2016). However, most of them deal in a simplified manner with the heterogeneity of carbonate aquifers, which in many cases appears to play a non-negligible role on local flow and transport condition.

In the mid-nineties Quinlan, J. F. et al., 1996 highlighted the poor performance of numerical models in karstic aquifers resulting from: i) inappropriate emphasis on the local scale flow paths characteristics, ii) the large amount of hydrogeologic data required to model such anisotropic, heterogeneous aquifers and the additional data requirements necessary to establish realistic boundary conditions iii) the necessity to develop models which incorporate triple continuum concepts, iv) the difficulty of properly characterizing and modelling transient events in karst aquifers and, v) the inadequacy of laminar flow equations or Darcy's law to describe the flow in conduits in the aquifer.

In less than twenty years, many efforts were made to improve the numerical models and nowadays numerical simulation may be considered a useful tool to provide temporal information about future recharge rates. (Hartmann et al., 2014).

Besides, the calibration and validation of such models need a large amount of data not deriving only on the hydrodynamic response between input (recharge)-output (discharge) or solely on the mere hydrographs analysis.

For this reason, the geological, geo-structural and traditional hydrogeological data have been coupled with others diagnostic tools such as hydrochemical data, natural and artificial tracers with the aim to obtain robust conceptual models. Physicochemical and hydrochemical parameters are increasingly used to obtain information about aquifer reactivity, vulnerability and water transfer processes, thus contributing to the development of hydrodynamic conceptual models.

1.2.2 Methods of tracing techniques

Tracing techniques can be used to obtain information about water movement and contaminant transport (Käss, 1998). With the term tracers could be identified three main groups: i) hydrochemical tracers, ii) artificial tracers and iii) natural isotopic tracers.

Excluding the anthropogenic influence on the hydrogeological environment, the hydrochemical facies of groundwater mainly result from water-rock interactions phenomena. The groundwater geochemical composition is influenced by aquifer lithology, water circulation conditions and residence time within the system.

Many studies investigate the relations between hydrogeochemical processes and karst physical characteristics (Bakalowicz, 1994; Vaute et al., 1997; Perrin et al., 2003; Barbieri et al., 2005; Terzić et al. 2014) demonstrating that the temporal variations observed in groundwater chemistry are usually related to the aquifer physical features and carbonate permeability, helping to identify hydrodynamic behaviour.

In this context, it must be noticed that the groundwater hydrochemical characterization can be integrated by natural tracing approaches based on the water isotope signature.

The isotopic investigations can be extremely useful in the assessment of groundwater-surface water interactions, water dating, quantification of water-rock interactions and evaluation of water resource

vulnerability (Emblanch et al., 2003; Katz et al., 2009; Batiot et al., 2003; Pulido- Leboeuf et al., 2003; Négrel & Petelet-Giraud, 2005; Adinolfi Falcone et al., 2008; Bouchaou et al., 2008).

The application of environmental tracing techniques often runs parallel to the artificial tracer tests. The latter are often more expensive and time consuming, but they are particularly useful and applicable in karst aquifer systems where transit times are short (thus limiting the required sampling and monitoring periods) and where the sole application of conventional hydrogeological methods, such as piezometric maps and numerical groundwater models, often leads to ambiguous and incomplete results (Worthington et al., 2017).

Theoretical concepts related to artificial tracer methods and isotope hydrology (with particular reference to stable water isotopes) are summarised in the following paragraphs.

1.2.2.1 *Artificial tracers*

Artificial tracers are substances intentionally added within the water to label flowing water (Goldscheider et al., 2008). They can have conservative or non-conservative properties.; the first ones are stable within the groundwater, and they do not show particular interaction or interferences with the host rocks, the latter should interact with the aquifers material, and they are also known as reactive tracers. While the conservative tracers are useful to mimic the water movement, the reactive ones can help to simulate the contaminant transport and its attenuation during times within the aquifers. However, in practice not all conservative tracers exhibit ideal conservative characteristics, but some of them are nearly so. In fact, the ideal tracer should be highly soluble, detectable at extremely low concentrations, behaves conservatively (i.e., no degradation or adsorption), and is nontoxic to humans and the environment (Käss, 1998; Leibundgut et al., 2009).

Another classification of tracer types can be made on the basis of their capability to be water-soluble or not (solutes or particles). The latter are substances that cannot dissolve in water such as particulate tracers. Particulate tracer dimensions range between less than 1 μm to few millimetres, hence they can experience a vast variety of phenomena during their movement such as filtration, sedimentation, exclusion, or other processes (Goldscheider et al., 2008). However, particles are invisible in the water, and they show no analytical interferences with other tracers, so they are recommended when visible colour must be avoided, and soluble tracer-rocks interaction phenomena are expected. Besides, particulate tracers require more experience and an advanced laboratory equipment to detect its

presence and quantify its amount within the water samples. This makes soluble tracers the most widely used in the field of hydrogeology.

The prominent solute tracers are the fluorescent dyes. They are synthetic organic compounds that absorb light at specific wavelengths and emit fluorescence light at longer wavelengths. Different fluorescent dyes are characterised by different detection limit (generally very low 10^{-3} - 10^{-1} µg/L), they are marked by different general or specific problems. For example, Uranine and Eosine (two of the most used tracer types) are sensitive to light and strong oxidants but at the same time they are toxicologically safe. On the contrary tracers such as Rhodamine Wt and Sulphorodamine can show analytical interferences between fluorescent dyes of similar optical properties and are marked by genotoxic and ecotoxic issues respectively. The aforementioned tracers are highly visible in the field of human visible wavelengths (≈ 15 µg/L for Uranine) and they are generally not recommended in areas characterised by catchments of mineral springs. Tinopal CBS-X is recommended when visible effects on water must be avoided and only a slight blue colour may result from the Tinopal CBS-X entering in water.

If the investigated area is not very wide and a short test is expected, salts can provide good results.

Salts are totally colourless and provide a response on water by increasing its electric conductivity (EC) or more generally by increasing its dissolved anion or cation content. Their detectability in terms of increased EC, measured directly in the field, is less than the one provided by individual cations or anions determined by laboratory analytical techniques. They generally require a large amount of mass to be injected if compared to the fluorescent dyes. Cations generally show greater interaction with the host rock while anions are more stable. In some cases, salts can undergo a number of chemical alteration processes forming toxic compounds (e.g. Bromide), in other cases they may be bio-chemically unstable (e.g. Iodide).

Different methods have been developed to design tracer amount to be released into the aquifer during a tracer test. A very simple equation was proposed by Käss, 1998:

$$M = L \cdot k \cdot B$$

where M: tracer quantity, i.e., mass (kg) or number of particles; L: distance (km), k: coefficient for tracer (uranine: 1, other fluorescence dyes: 2–15, salts: 1000–20000, particles: 10¹²–10¹³); B = factor for hydrogeological conditions (0.1–0.9 for karst aquifers).

Further equations were developed during the years (Worthington & Smart, 2003), taking into account the system discharge and target peak concentration or by adding the groundwater velocities (Aley & Fletcher, 1976). Filed 2002, proposed a new design method based on the one-dimensional advection dispersion equation (ADE) named “efficient hydrologic tracer-test design methodology” (EHTD) in which also the sampling frequencies were considered.

In all cases, the tracer mass to be injected should be determined balancing the risk connected by the additional substances introduced into the system and the consequence of failing the test due to the non-achievement of the detection limit (Goldscheider et al., 2008).

Special attention must be paid during the tracer preparation and injection operations avoiding samples contamination and therefore the analytic phases.

The selection of the tracer releasing point is strictly related to the objective of the research. In some cases, the tracer is injected in wells or piezometers (Worthington & Smart, 2016; Tazioli & Tacconi, 1991) directly into the saturated portion of the aquifer, choosing discrete depth or homogenising the tracer in all the water column within the well. In other cases, sinkholes, sinking streams or caves are selected as they transfer the tracer into the aquifer. In this case the tracer travels both in the unsaturated and in the saturated portion of the aquifer following the main active conduits. Generally sinking stream guarantee a large volume of flushing water, avoiding the use of large tanks that must be brought on site to facilitate injection operations. Usually, the tracers are released instantaneously within the system without modifying the natural flow rate.

Different methods are available to observe artificial tracers at relevant monitoring sites (springs, wells, rivers), such as manual or automatic sampling with subsequent laboratory analysis or the use of continuously recording field fluorometers (Schnegg, 2002). In case of preliminary tests (qualitative), for inaccessible sites or numerous monitoring points, the use of the integrative method is recommended when using a fluorometric dye. It consists of the accumulation of tracer over a certain time interval by means of an active carbon trap (i.e., granular activated charcoal). Moreover, the usage of active carbon traps could be helpful as backup for others sampling methods, especially when the expected tracer concentration is low and the automatic samples or the field fluorometers may record a tracer concentration not far from the detection limits.

The sampling of water and the substitution of carbon traps should be faster during the early stages after injection and may slack off in the later period. Before the injection operation a blank sample

must be collected to define the water background value and to calibrate the analytical instruments used to determine the sample tracer concentration.

The main result of a tracer tests is a breakthrough curve (BTC) in which the tracer concentration (discreetly or continuously recorded) is expressed over time. The BTCs provide quantitative information about the tracer transit times and different flow velocities (maximum, mean and peak) if the flowpath length is known, in any case the linear distance between injection and detection point represents a robust approximation providing satisfactory results.

The reliability of a tracer test is evaluated by the tracer recovery (Mr) calculated as follows:

$$Mr = \int_{t=0}^{\infty} (Q \cdot c) dt$$

where Q is discharge, c is tracer concentration, and $(Q \cdot c)$ is flux; it is inherent therefore that the computation of Mr requires the discharge data at the monitored point.

As concerns the velocities, the maximum flow velocity is strictly dependent to the analytical detection limit (i.e., the lower the detection limit, the earlier the first possible detection, hence time of first tracer detection); the peak time (maximum concentration peak) yields the dominant or peak velocity and is the most readily determined.

The mean velocity is the best to be used in analytical models, it is calculated starting from the time of the concentration centroid (time when half of the recovered tracer has passed the sampling site, also known as mean transit time). The mean transit time (t_m) is useful to estimate the system volume considered as the volume of water displaced in the system during the duration of tracer test, and it is calculated as follows:

$$V = \int_{t=0}^{t_m} Q dt$$

where Q is the system discharge.

Several studies show that tracers, as well as water, do not flow at one single advective velocity, but they experience a wide range of velocities. The highest are recorded in the centres of large conduits, while low flow conditions are recorded in the small, interconnected fissures and fractures. For this reason, the analytical solution proposed by Bear, 1979 describing the tracer movement with the advection-dispersion model (ADM) in function of different initial and boundary conditions, has

been modified by an advanced analytical equation such as the multi-dispersion model (MDM) (Maloszewski et al., 1992) by joining several ADMs. However, it has been demonstrated that tracer can remain in dead zones of caves and conduits (pools) and remobilised after rainfall events, making the BTCs characterised by several peaks. This behaviour also occurs in the presence of absorption-desorption phenomena. An analytical solution for tracer transport in a non-equilibrium state (i.e., exchange between mobile and immobile fluid regions) has been proposed by Field & Pinsky, 2000.

The analytical solutions for tracer transport in a non-equilibrium state usually are integrated with spatially distributed or lumped hydrologic models conceptualizing the flowpaths and transit time depicted by artificial tracers, the water residence time obtained by means of natural tracers, yielding the mixing fractions, aquifer characteristics, and diffusion losses to the matrix (Hartmann et al., 2014). In general, the inclusion of tracer test results in groundwater numerical simulations can substantially improve the validity of the models (Worthington, 2009), but at the same time not all the model types provide acceptable results if not characterised by large amount of data. Lumped models generally are characterised by higher adaptability and strength in simulating the hydrogeological behaviour in several different context (Hartmann et al., 2015). However, according to Mudarra et al. 2019, only one research and no efforts were made to addresses prognostic simulations of water resources at the whole catchment scale of carbonate aquifers, by integrating tracer tests data in the lumped models, suggesting that there is clearly room for improvement in karst water resources evaluation and projection. The inclusion of additional information derived from tracer tests can improve the parameter calibration procedure (Mudarra et al. 2019).

In general, it must be highlighted that the use of artificial tracers proves to be decidedly primary in defining the hydraulic connections between different aquifers or within different points belonging to the same groundwater body, or in determining the hydrogeological parameters which are the foundations of conceptual, and consequently numerical models. To the best of my knowledge, there are no long-term (more than 5 years) continuous and periodic tracer tests, involving the same carbonate domain before, during and after a seismic sequence. For this reason, one of the main purposes of this research is to integrate the results of artificial tracer tests with the traditional investigation methods and develop an holistic approach useful for water management purposes to define the hydrogeological characteristics even in a transient condition caused by natural hazards such as big earthquakes.

1.2.2.2 Isotope hydrogeology

Oxygen and hydrogen isotopic measurements have been made for more than 100000 groundwater samples since the 1950s to 1960s (Jasechko, 2019) and some of the interpretations of these isotopic measurements have advanced the hydrological sciences. Nowadays, isotope hydrogeology is considered a subfield of water resources research because it is focused on the definition of processes and reactions occurring during the recharge, storage, and discharge of groundwater.

Isotope hydrogeology typically study stable or radioactive isotope compositions of river, spring, and well water samples, then interpreting isotopic measurements to quantify or conceptualize groundwater flow paths, velocities, and biogeochemical reactions. In addition, the isotopic compositions of stored groundwater can be considered a powerful tool to distinguish flow networks through heterogeneous aquifers (i.e., karst aquifers).

The advantage of investigating the groundwater isotope compositions, lies in the opportunity of tracing movements of molecules, providing information beyond what can be inferred from traditional investigation methods (piezometric, lithologic, meteorologic, and solute concentration measurements). This concept may seem similar to what concerned for artificial tracers, but contrariwise to the artificial tracers, natural tracers, or environmental tracers, such as isotopes, are inherent components of the water cycle. Instead of being added to the water at a well-defined time and location, they are naturally abundant and added spatially and continuously to the hydrological system (Hartmann et al., 2014). In nature, Oxygen has three stable isotopes (^{16}O , ^{17}O , and ^{18}O); Hydrogen has three naturally occurring isotopes (^1H , ^2H , and ^3H); two of them are stable (^1H and ^2H); while ^3H is radioactive.

The stable isotopes ^{16}O (oxygen-16), ^{18}O (oxygen-18), ^1H (protium), and ^2H (deuterium) are the most widely measured in isotope hydrology and hydrogeology. Their abundances are reported in relation to standard ocean water in “delta notation” as follows:

$$\delta^{18}\text{O} = \left[\frac{\left(\frac{^{18}\text{O}}{^{16}\text{O}} \right)_{\text{sample}}}{\left(\frac{^{18}\text{O}}{^{16}\text{O}} \right)_{\text{SMOW}}} - 1 \right] \times 1000$$
$$\delta^2\text{H} = \left[\frac{\left(\frac{^2\text{H}}{^1\text{H}} \right)_{\text{sample}}}{\left(\frac{^2\text{H}}{^1\text{H}} \right)_{\text{SMOW}}} - 1 \right] \times 1000$$

where “sample” refers to the H₂O being measured (e.g., a sample of groundwater), and “SMOW” stands for “standard mean ocean water”. The units of measurement are “per mille” (‰) (Mook & Rozanski, 2000).

The delta notation scale ranges from –1000 to ∞. Positive δ values denote isotopes ratios that exceed SMOW, while negative δ values denote isotopes ratios lower than SMOW. In practice, measurements of δ¹⁸O and δ²H values are frequently made relative to internal laboratory standards, themselves calibrated against international standards (i.e., V-SMOW: Vienna- standard mean ocean water, etc).

¹⁸O, ²H and ³H can be regarded as ideal natural tracers since they are part of the water molecule. Assuming that, the various components of the hydrological cycle can be characterised from an isotopic point of view by sampling rainfall water and snow cover (meteoric recharge), surface water (lakes, rivers), also in solid form (glaciers), water contained in the soil or groundwater by directly springs sampling or groundwater drained from wells or intercepted by piezometers.

The first step in characterising a site or study area is to sample precipitation.

Precipitation δ¹⁸O and δ²H relationships are described by the so called “meteoric water lines” (linear regressions describing precipitation δ²H and δ¹⁸O variations). Precipitation δ¹⁸O and δ²H values often covary.

A “global meteoric water line” (GMWL) was first characterized by Craig, 1961:

$$\delta^2\text{H} = 8 \times \delta^{18}\text{O} + 10$$

A recent analysis conducted by Jasechko, 2019 of n = 68382 precipitation samples collected from stations located around the world yields a nearly identical relationship to the GMWL proposed by Craig, 1961:

$$\delta^2\text{H} = 7.93 \times \delta^{18}\text{O} + 8.99$$

Except for globally developed studies, local meteoric water lines (LMWLs) are required to better investigate the regional isotopic features. It has been demonstrated that LMWLs can have regression coefficients that significantly vary from that of the GMWL, with differences induced by seven globally recognised processes:

i) different δ¹⁸O values of the seawater from which the condensing moisture originates (Araguás-Araguás et al., 2000), ii) physiochemical conditions impacting kinetic and equilibrium fractionation during evaporation at the sea surface where the condensing moisture originated (Merlivat & Jouzel, 1979), iii) mixing of air masses of varying origins and histories as they travel through the atmosphere, iv) temperature-driven rainout that occurs as moisture advection over the continents

(Gat, 1996; Sonntag et al., 1983), v) vapor fluxes that recycle precipitation from Earth's land surfaces back into the atmosphere (Gat et al., 1994; Gat & Matsui, 1991; Kurita et al., 2004; Salati et al., 1979), vi) physiochemical conditions (e.g., temperature) during the condensation of moisture into rain and snow (Merlivat, 1978; Gonfiantini et al., 2001), and finally vii) below-cloud moisture entrainment (Friedman et al., 1962) or partial evaporation (Stewart, 1975) as the precipitation descends toward the land (Craig, 1961; Dansgaard, 1964; Lee et al., 2007; Lee & Fung, 2008; Feng et al., 2009; Bailey et al., 2018).

Consideration of LMWLs is common to many studies employing environmental isotopes because local precipitation isotope compositions exert a first-order control on the isotope composition of local groundwater, so it is important to consider local processes affecting precipitation isotope compositions when interpreting spatial variations in groundwater isotope features.

Precipitation $\delta^{18}\text{O}$ and $\delta^2\text{H}$ values often vary seasonally: higher $\delta^{18}\text{O}$ (and $\delta^2\text{H}$) values during warmer months and lower $\delta^{18}\text{O}$ values during cooler months. The covariation of precipitation $\delta^{18}\text{O}$ and temperature has been termed the “temperature effect” (Dansgaard, 1964).

Correlations between precipitation rates and $\delta^{18}\text{O}$ values are sometimes referred to as “amount effects” (Longinelli & Selmo, 2003).

Even after precipitation occurrence, numerous processes may influence the isotopic composition of groundwater, if the precipitation eventually recharge an aquifer. For example, groundwater $\delta^{18}\text{O}$ and $\delta^2\text{H}$ values may be higher than amount-weighted precipitation if the stored groundwater reaches high temperatures (Giggenbach, 1978; Giggenbach & Stewart, 1982) or if the infiltrating water evaporate partly before recharging the aquifer (Gat & Tzur, 1968; Gat & Dansgaard, 1972). One parameter commonly applied as a preliminary test for the importance of disequilibrium isotope effects (i.e., evaporative isotope effects) is the deuterium excess value (d-excess) (Dansgaard, 1964; Pfahl & Sodemann, 2014) calculated as follow:

$$\text{d-excess} = \delta^2\text{H} - 8 \times \delta^{18}\text{O}$$

Groundwater d-excess values are similar to local precipitation d-excess, if groundwater retain the isotopic composition of recharge, underlying fast infiltration processes (Vogel & Van Urk, 1975; Mathieu & Bariac, 1996; Saxena, 1984).

At a global scale most groundwater samples have d-excess values similar to precipitation (implying partial evaporation of infiltrating water to be of lesser importance) (Jasechko, 2019). Several studies suggest that evaporative isotope effects have minimal impact on isotopic compositions of recharge (Dennis et al., 1997; Mountain et al., 2015), while other works identify groundwater $\delta^{18}\text{O}$ and $\delta^2\text{H}$

values positioned “below” the meteoric water line in $\delta^{18}\text{O}$ - $\delta^2\text{H}$ space, suggesting evaporative isotope effects, those effects may be important especially in certain arid areas (Fontes & Edmunds, 1985).

However, where groundwater isotope compositions remain constant over the seasons and are not altered by evaporation processes, they may differ from local annual precipitation isotope compositions. This differences may arise where i) seasonal differences in recharge/precipitation ratios exist (Brinkmann et al., 1963; Vogel et al., 1975), ii) where groundwater recharge at higher elevations and subsequently flow to a location in an aquifer that underlies lower-elevation lands (Giggenbach et al., 1983), iii) where the groundwater originates as precipitation that fell far from the location it recharged (Darling et al., 1987), iv) where the groundwater derives from precipitation that fell under different climate conditions (Vogel & Ehhalt, 1963; Edmunds & Wright, 1979).

Mixing of groundwater, affected by different origins and histories, may drive spatial variations in their isotope compositions.

Stable isotopes have long been applied to identify recharge sources or areas and many studies have interpreted water stable isotope compositions as indicative of influxes of surface water into aquifer systems or elevations at which groundwater in an aquifer first recharged.

Stable oxygen and hydrogen isotope compositions have also been applied to understand how water exchanges among aquifers and estimate the elevation that groundwater samples recharged (Kattan, 2006; Payne & Yurtsever, 1974).

In mountain areas, groundwater isotopic values are often lower than nearby annual amount-weighted precipitation $\delta^{18}\text{O}$ values (Kattan, 2001; Kebede et al., 2005; Scholl et al., 1996; Weyhenmeyer et al., 2002). Globally, precipitation $\delta^{18}\text{O}$ ($\delta^2\text{H}$) values tend to decrease with elevation, and this aspect is also known as “altitude effect”. (Bowen & Wilkinson, 2002; Poage & Chamberlain, 2001). Inverse relationships between local precipitation $\delta^{18}\text{O}$ and local elevation arise due to the preferential rainout of heavier isotope as vapor in an air mass ascends, cools, and precipitates due to the orographic uplift. Understanding the locations and elevations of groundwater recharge areas is important for developing groundwater quality protection plans.

Stable isotope compositions have been applied to distinguish groundwater i) generated by surface-water sourced recharge (i.e., from lakes, rivers, and wetlands) and ii) high-elevation recharge.

Two steps are required to estimate recharge elevations using groundwater isotope compositions. First, a locally significant empirical relationship quantifying how meteoric water $\delta^{18}\text{O}$ (or $\delta^2\text{H}$) changes with elevation should be developed. Then, recharge elevations can be estimated by applying the inverse $\delta^{18}\text{O}$ (or $\delta^2\text{H}$)-elevation equation to the groundwater isotopic values. Typical

isotopic gradients range from -0.1‰ to -0.5‰ per 100 m of elevation gain for $\delta^{18}\text{O}$ and -0.5‰ to -4‰ per 100 m of elevation gain for $\delta^2\text{H}$ (Jasechko, 2019).

$\delta^{18}\text{O}$ -elevation gradients can vary considerably even over short distances (Scholl et al., 2002), hence, developing local $\delta^{18}\text{O}$ -elevation relationships is important to investigate hydrogeological features.

In mountain regions, groundwater recharged at high elevations likely mix with groundwater recharged at lower elevations prior to sampling (Dafny et al., 2006; Jódar et al., 2016), but isotope hydrologists often implicitly assume recharge takes place at a single elevation. For this reason, mixing take place and no single recharge elevation can characterize all the groundwater in a sample. To this aim, decoupling the potential for seasonally selective recharge (e.g., preferential recharge of snowmelt) versus high-elevation recharge should be done, but it is demonstrated to be a challenge aspect, however each effect may take place and impact groundwater $\delta^{18}\text{O}$ values (Herrera et al., 2016). Untangling high-elevation recharge from seasonal variations in recharge ratios has been proved to be much more challenging (Abbott et al., 2000; Scholl et al., 1996).

Seasonal variations in $\delta^{18}\text{O}$ -elevation relationships exist only in some areas (Gonfiantini et al., 2001). Where the seasonal and interannual recharge is recognised (Jasechko et al., 2014), the potential for temporally variable $\delta^{18}\text{O}$ -elevation relationships introduces a source of uncertainty if these temporal variations are not captured by precipitation isotope monitoring data.

The estimation of recharge area elevations based on isotope compositions rely on the existence of a measurable $\delta^{18}\text{O}$ -elevation gradient in the study site and the methods for estimating recharge elevations using isotope compositions are only applicable where precipitation $\delta^{18}\text{O}$ values correlate significantly with elevation (Jasechko, 2019).

Identifying a specific elevation at which the groundwater sample recharged, implicitly assumes that the entire groundwater sample recharged at a single elevation, which is an oversimplification. Incorporating spatial variability of precipitation isotope compositions into hydrogeological models may improve interpretations of groundwater isotope compositions (i.e., the range of elevations at which the mixture of groundwater in the sample recharged).

What remains unclear is the relative relevance of diffuse versus focused groundwater recharge for high elevation recharge, and how the importance of each mechanism varies spatially (Mercer et al., 2020). Understanding recharge rates at seasonal timescales is also important in order to i) forecast changes to annual recharge rates under changing seasonal climate conditions (Ajami et al., 2012; Earman et al., 2006; Earman & Dettinger, 2011; Niraula et al., 2017), ii) develop more sophisticated conceptual models detailing hydrological processes and conditions conducive to groundwater

replenishment (Alsaaran, 2006; Hammarlund & Edwards, 2008; Pavlovskii et al., 2017; Rathay et al., 2018; Zhang & Wang, 2016).

Groundwater and precipitation isotope compositions have been analysed to quantify proportions of annual recharge derived from winter or summer precipitation (Fritz et al., 1987). The seasonal differences in recharge ratio, where the recharge ratio is the amount of precipitation that becomes recharge (i.e., recharge/precipitation) is defined “seasonal recharge bias” (Cherry et al., 2020). There are a number of potential limitations to consider if applying $\delta^{18}\text{O}$ or $\delta^2\text{H}$ data to evaluate seasonal recharge biases, three of them are i) the changes to snowpack isotope compositions over time, ii) partial evaporation of recharged groundwater, iii) possible influences from spatial variations in precipitation $\delta^{18}\text{O}$.

All the methods presented in literature require an implicit assumption that two seasonal end members approximate the cumulative isotope composition and rate of recharge during two seasons: summer and winter. This is a known oversimplification (Jasechko et al., 2017), as recharge ratios vary within individual seasons.

It remains challenging to decouple the relevance of seasonally selective recharge from other factors that may also lead to low groundwater $\delta^{18}\text{O}$ values relative to annual precipitation $\delta^{18}\text{O}$ values, including recharge at higher elevations where precipitation $\delta^{18}\text{O}$ values are often lower (Weyhenmeyer et al., 2002).

Strong annual rainfall $\delta^{18}\text{O}$ -elevation relationships do not exist in some of the locations where recharge biases have been proposed (Cunningham et al., 1998; Winograd et al., 1998), implying high-elevation recharge cannot explain the observation that the groundwater isotopic values are lower than the annual rainfall isotopic ones in many areas. At other sites, insufficient relief exists to explain groundwater versus precipitation $\delta^{18}\text{O}$ offsets, implying seasonal biases in groundwater recharge likely exist (Penna et al., 2014).

Overall, the possibility that the groundwater isotopic values are lower than the annual rainfall isotopic ones arises due to high-elevation recharge should be tested by i) estimating local precipitation $\delta^{18}\text{O}$ -elevation relationships and ii) evaluating the range of possible elevations at which the local groundwater may have recharged (Jasechko, 2019). In this context it has to be noticed that snowpack isotope compositions can change throughout the snowmelt process, meaning precipitation isotope compositions should be integrated by snowmelt end-members (Ala-Aho et al., 2017; Beria et al., 2018; Dahlke & Lyon, 2013; Feng et al., 2002; Laudon et al., 2002; Lechler & Niemi, 2012; Lee et al., 2010; Zhou et al., 2008).

Earman et al., 2006 evidence that bulk snowpack $\delta^{18}\text{O}$ values tend to increase after deposition stressing that the researches that do not account for changes to snowpack isotope compositions expectedly underestimate true recharge $\delta^{18}\text{O}$ signature.

When dealing with wide areas, far from the evaporative processes occurred in oceans or seas, also the “continental effect” should be comprising in the isotopic studies. The continental effects occur during the transport of vapour masses over long distances moving inland from seacoasts, generating fractionation processes (Zoltán et al., 2020).

A schematic illustration encompassing the most important isotopic fractionation processes (i.e., effects) within the water cycle are reported in Figure 1.4.

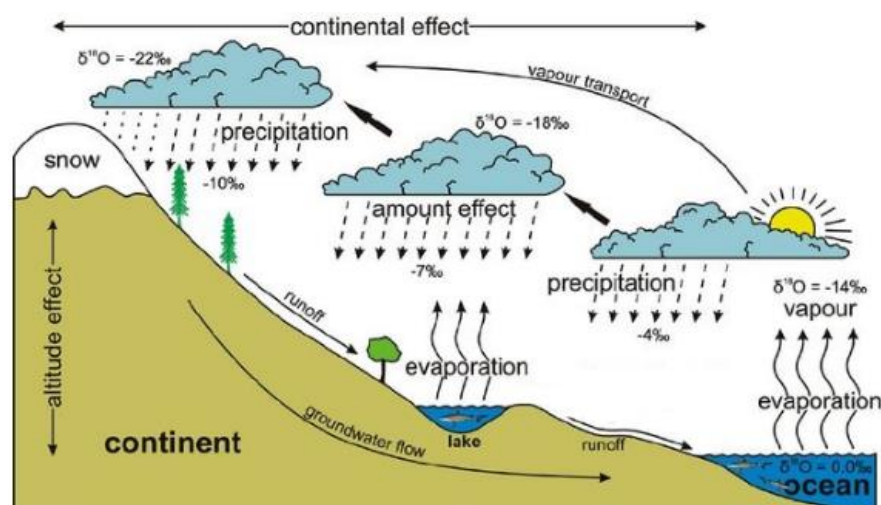


Figure 1.4 Main isotopic fractionation processes occurring within the water cycle, modified from Fischer, 2012.

In general, intercomparison of stable-isotope-based approaches against other measurement-based approaches can further quantify temporal variations in recharge ratios. Combining and comparing multiple approaches is optimal for quantifying groundwater recharge rates (De Vries & Simmers, 2002). Thanks to the isotopic techniques it is possible to develop sophisticated and quantitative conceptual models of groundwater recharge. Isotopic methods can help water scientists identify processes relevant to groundwater quality and sustainability. Isotope hydrology offers most valuable outcomes if coupled with more conventional hydrological and hydrogeological data including meteo-climatic monitoring, piezometric or discharge measurements, geological surveys, and solute concentration analysis.

1.2.3 Hydrogeological modification due to seismic events

The hydraulic conductivity of karst aquifers depends by the geological history of the rocks and by the dynamic evolution of the groundwater flow system (Király, 2003). With this concept taken from Király 2003, Attila Kovács, in the work entitled “Quantitative classification of carbonate aquifers based on hydrodynamic behaviour” published in 2021, referred to the dynamic effects that karst aquifers can undergo due to the opening and widening of fractures, due to the karstification process, which increases void sizes, highlighting that this process represents an important feed-back loop in the evolution of karst systems. In fact, groundwater flux depends on hydraulic conductivity and hydraulic gradient which respectively depends on the aperture of fractures. This concept obviously does not imply modifications linked to seismic events (which were outside of the scope of the work), but it turns a light on important thoughts about the dynamic effects that carbonate and karst systems can undergo over time: some of these are certainly the strong earthquakes.

Earthquakes are historically known to cause hydrogeological modifications, including changes in chemical composition (Cleasson et al., 2004; Skelton et al., 2019), streamflow (Miur-Wood & King, 1993; Rojstaczer & Wolf, 1995; Manga, 2001; Manga et al., 2003; Montgomery & Manga, 2003; Manga & Rowland, 2009; Petitta et al., 2018), and spring hydrodynamic characteristics (Wang & Manga, 2015; Valigi et al., 2019). The hydrogeological modifications can be transitory, or they can display permanent features (Manga & Wang, 2015; Charmoille et al., 2005; Esposito et al., 2009; De Luca et al., 2018; De Luca et al., 2016).

The theoretical explanation for the groundwater system modifications after seismic events should be simple because they reflect the deformations that are caused by seismic waves (La Vigna, 2013). In this regard, earthquakes change both the static stress (i.e., the movement of two portions of the earth's crust along a fault induce a change in the state of crustal stress accumulated prior to the earthquake) and both the dynamic stress (i.e., seismic waves due passage). The stresses increase with increasing seismic motion, but decrease inversely with distance (Manga & Wang, 2007). The hydrological modifications due to earthquakes have been categories into three classes according to the distance from the hypocentre: i) the near-field is the area around the hypocentre at a radius distance of about 1 length from the seismogenic fault , ii) the intermediate-field is the area around the hypocentre at a radius distance > 1 length from the seismogenic fault, but < 10 lengths from the seismogenic fault, and iii) the far-field is the area around the hypocentre at a radius distance more than 10 fault lengths (Wang & Manga, 2010).

In the far-field, the dynamic stress peak is much higher than the one of static stress, while both stresses are comparable in the near-field. In the intermediate-field both stresses are measurable, but the dynamic stress is higher than the static one (Wang & Manga, 2010). Therefore, seismic waves cause spatial stress modifications and, thus, spatial variations in pore pressure; the resulting difference in pore pressure generates pressure gradients and causes flow modifications (Manga & Wang, 2007). These modifications in hydrodynamic behaviour are recognized at regional scales hundreds of kilometres away from the activating fault (Cox et al., 2012; Mohr et al., 2017). Indeed, Barberio et al., 2020 have found a clear correlation between earthquakes distance and magnitude in hydrogeological response of piezometric level fluctuations for 18 worldwide earthquakes. Esposito et al., 2001 described, among others, the consequences of the 1980 Irpinia earthquake, which caused hydrogeological effects as far as 200 km; in particular, a significant increase of the Caposele spring flow was observed, which returned to normal conditions after about one year. On the other hand, Rojstaczer & Wolf, 1992, recognized during the well-known Loma Prieta earthquake, significant hydrological changes in two basins located near-field from San Andreas Fault, attributing the cause of these modifications to the formation of new fractures capable to develop novel continuous flow path or enhance existing continuous flow path.

Changes in pores geometry and fractures due to the variations in interstitial pressure and stress, can be different with regards to the lithologies involved in the seismic deformation (Manga & Wang, 2007). The incoherent deposits, are characterised by sediment grains rearrangement due to the shear deformation, tending to fill in pre-existing pores. This effect reduces the original porosity, resulting in a phenomenon commonly known as consolidation or compaction. Conversely, shear deformation in coherent deposits, causes the sediment grains to roll on top of each other creating new voids. The latter increases the total volume of the sediment and is commonly referred to dilatancy.

The behaviour of lithoid materials with regards to the seismic deformation is significantly different and often exceeds its elastic limit, resulting in the opening of micro-cracks (Brace et al, 1966).

Many authors tried to understand and explain which processes are responsible for the hydrogeological modification after big earthquakes (Lai et al., 2004; Manga & Rowland, 2009; Montgomery, 2003; Adinolfi Falcone et al., 2012). Holzer & Youd, 2007, proposed undrained consolidation (compaction) to explain near-field pore pressure changes, and several authors suggested that seismic waves may increase rock permeability by removing fracture fill; this

phenomenon may lead to redistribution of pore pressure and thus changes in the piezometric features (Brodsky et al., 2003; Matsumoto et al., 2003; Wang & Chia, 2008).

As reported by Mohr et al. 2017, discharge increases after earthquakes, observed mainly on stream gauges, can be explained by three mechanisms: i) changes in static strain determining aquifer geometry modifications, pore pressure changes and consequent variations of hydraulic gradient i (Muir-Wood & King, 1993); ii) dynamic strain effects such as consolidation and shaking of water out of the unsaturated zone (Manga, 2001; Manga et al., 2003; Montgomery, 2003; Manga & Rowland, 2009; Mohr et al., 2017); iii) increased permeability due to dynamic strain (Rojstaczer et al., 1995; Petitta et al., 2018; Valigi et al., 2019; Mohr et al., 2017; Wang et al., 2004; Amoruso et al., 2011; Manga & Wang, 2015; Hosono et al., 2019). Another explanation was reported by Muir-Wood & King, 1993 suggesting that co-seismic and post-seismic hydrogeological responses may be a function of the focal mechanisms of earthquake.

Thus, what appears to be a fairly simple problem from a theoretical point of view, in practice is characterised by extreme complexity depending on the lithologies involved in the seismic events and the distance from the hypocentre. Controls on the magnitude of changes and subsequent evolution of permeability remain poorly documented, and even the mechanisms are uncertain (Ingebritsen & Manga, 2019). Some of the more comprehensive studies have shown that multiple mechanisms must be at play (Hosono et al., 2019; Yan et al., 2014).

Therefore, since the cause of earthquakes is the activation of fault systems, at least in the near-field, the role of faults played before (i.e., inter-seismic), during and after the seismic sequences and their study seems to be the main focus of earthquake hydrogeology.

Nowadays, the role of fault systems in fluid transfer is a scientific challenging aspect, whose interest came into light again especially after the seismic shocks occurred in the last decades (Manga et al., 2003; Charmaillé et al., 2005; Shi et al., 2014). Many authors have focused their attention on the interaction between fault zones and drainage patterns (Goldscheider & Neukum, 2010), while others have tried to develop physical models to simplify and quantify how flow conditions are affected by fault zones (Evans et al., 1997; Fairley et al., 2003; Medici et al., 2021). A whole series of research works focus their attention on the fault behaviour in detail, joining hydraulic, temperature, other natural tracers data and borehole tests results (Bense et al., 2013) to better identify the reservoirs characteristics in close proximity of fault zones. Most of the works deal with fault systems and deep

fluid transports in geo-structural fields for oil or hydrocarbon exploitation purposes while a smaller number of studies are focused on the impact of fault systems on groundwater flow (Scibek et al., 2016). Only a small number of these works, actually, deal with the changes in the role of a certain fault system before and after the seismic period (Xue et al., 2016). However, the latter is really a crucial aspect in hydrogeological studies, especially when they concern groundwater management issues (Fronzi et al., 2020). In fact, it is known that faults behave either as conduits or barriers with respect to groundwater flow depending on their kinematics and fault rock lithology (Caine et al., 1996; Faulkner et al., 2010), but the age-old problem of identifying if they limit or even block the flow of groundwater or if they rather act as a preferential flow path is partially still unsolved (Celico et al., 2006). For instance, Bense et al., 2013, in a review on fault zone hydrogeology suggest that the product of deformation processes accommodating strain can reduce or enhance the permeability within fault zones. In particular, long-lasting spring and stream excess flows are known to accompany major normal faults and strike-slip faults earthquakes (Petitta et al., 2018; Wang & Manga, 2015; Valigi et al., 2019), although discharge increases have also been detected for thrust faults induced earthquakes (Manga, 2001; Mohr et al., 2017). According to Sibson, 2013, for normal faults, fluid movement is favoured along the fault strike (along the σ_2) during the inter-seismic period when the fault surface mainly behaves as a barrier. During the seismic shock, if the earthquake is strong enough to break through the entire seismogenic layer (or even to cause surface ruptures), the fault rock becomes a highly permeable layer, which drains fluids from the uplifting foot-wall to the subsiding hanging-wall.

Increase of spring and streamflow discharge due to extensional tectonics are common effects in the Italian Apennines.

In Italy, some authors suggested different interpretations to explain groundwater modifications due to the main earthquakes occurred in the country (De Luca et al., 2018; Adinolfi Falcone et al., 2012; Amoruso et al., 2010; Esposito et al., 2009).

Amoruso et al., 2011 described the hydrogeological changes that occurred after the 2009 L'Aquila earthquake. These effects included the sudden disappearance of some springs located along the normal fault responsible for the earthquake, an immediate discharge increase of both the drainages along the Gran Sasso highway tunnel and other springs, and, in the following months, an increase of the water table elevation of the Gran Sasso fractured aquifer. According to the authors, these changes were related to the increase of permeability at the aquifer scale, which caused hydraulic

heads raising in the discharge zones and their lowering in recharge areas. Relating to the same earthquake, Adinolfi Falcone et al., 2012 indicated that, in agreement with the observed gradual return to the pre-seismic situation, the observed changes have non-permanent effects because they can be due to a fracture cleaning, rather than to the formation of new micro cracks. Petitta et al., 2018 described the hydrogeological effects of the 2016–2017 seismic sequence in Central Italy, providing preliminary considerations on the causes of the observed phenomena. The authors considered data from 22 sites, located in a wide area, within 100 km from the epicentre. They found that post-seismic discharge from rivers and springs had increased, overall, more than 9 m³/s, with a consequent groundwater release of more than 1 × 10⁸ m³ over 6 months. The authors preliminary conclude that the effects can be due to both an increase of bulk hydraulic conductivity (*K*) at the aquifer scale and, partly, to other mechanisms, such as shaking and/or squeezing effects related to intense subsidence in the core of the affected area or breaching of hydraulic barriers. Valigi et al, 2019 analysed the effects of the 2016-2017 seismic sequence only in the Norcia Plain, close to the epicentre of the October 30th seismic event. They found, by operating only on the springs and streams discharge data, that at least 2.4 m³/s of groundwater was added to the original baseflow amount of the Sordo River with the additional discharge mainly provided by the basal carbonate aquifer.

1.2.4 Hydrogeological conceptual models

The way through out scientists interpret available information to produce a reasonable set of elementary assumptions, to explain a system, is called conceptual modelling (Rushton, 2003; Brassington & Younger, 2010). Models are of fundamental importance in most scientific investigations and have become one of the principal tools of modern applied science (Silvert, 2001). Models varies from physical or imaginary objects to three-dimensional representations, descriptions, equations or combinations of several of these groups. In order to develop any of these models it is necessary to create a theory-based description that synthesize the studied phenomena founded on a set of variables with logical and quantitative relationships. This theory-based description is known as conceptual model. From a hydrogeological point of view, conceptual modelling is often required to demonstrate the sustainability of an action (human or natural) impacting on the hydrogeological system. The objective evaluation, conducted by the hydrogeologists to build a conceptual model are used by developers, public authorities, industries, water management companies and/or scientists to take decisions which may have an impact on the system or to conduct any future studies and research in the same area (Brassington, 2004).

A conceptual model will include a collection of reasonable and simple assumptions which summarize the main characteristics of the real system and thanks to the model, the system behaviour may be more clearly understood. The model should summarise the result of data interpretation, those data can be published information with or without new field and laboratory tests, or if necessary further understanding can be extracted from simple and preliminary mathematical modelling using these data. In most cases, the purpose of developing a conceptual model is to provide a sufficient understanding of the relationships between the principal characteristics of a system so that deductive and/or mathematical methods can be used to evaluate possible outcomes of changes within the system for a range of feasible situations.

Hence, conceptual modelling should always precede any attempt to build a numerical model of a groundwater system (Rushton, 2003). This conceptualization is a crucial step, as numerical groundwater models built on defective conceptual models unavoidably produce inappropriate results (Bredehoeft, 2005; Robins et al., 2005).

Generally, if the solutions provided by the numerical model are not reliable (measured hydraulic heads, discharge rates etc. divergences between the reality and the model output), the conceptual

model (definition of geological units, inclusion of faults, conduits geometry, etc.) must be revised. (D’Affonseca et al., 2020).

In some cases, the conceptual modelling does not necessarily require a mathematical modelling at all after its completion (Brassington, 2006).

The scale of hydrogeological assessments and investigations is very variable. The hydrogeological conceptual models are required for both small private wells (i.e., for irrigation purposes), for regional water management operations or continental evaluation of groundwater availability (Brassington & Younger, 2010). Whatever the scale of the problem, acquiring data prior to modelling is the key point (D’Affonseca et al., 2020).

A number of technical notes and technical reposts, as well as textbooks have been developed with the aim to provide some guidance on hydrogeological conceptualization (Bear & Verruijt, 1987; Rushton, 2003; Younger, 2007; Anon, 2002). However, the activities necessary to complete a hydrogeological investigation and the methodology of developing a conceptual model are not exactly the same. The actions to build up a hydrogeological model includes investigation that are generally focused on collecting information, whereas the emphasis in developing a conceptual model is the interpretation of data as they are collected and identifying additional information needed to complete the conceptual understanding. The hydrogeological investigation can be divided into separate parts, each one builder on the previous adopted method. It will always be necessary to modify the details of the investigation to the needs of each study, and the modeller should be prepared to change the investigation technique based on the problems encountered in applying a particular planned technique (i.e., fieldwork issues that are the common limitation of some hydrogeological methods). Brassington, 2006 schematized the essential phases that would be expected to be included in most investigations in seven steps: i) desk review, ii) field survey, iii) exploration, iv) monitoring programme, v) data management, vi) water balance and vii) completion of the conceptual model.

The assumptions made about each predominant characteristic of a groundwater system are iteratively re-evaluated to further develop understanding as new evidence is acquired and tested against the evolving conceptual mode. This process has been represented by Enemark et al., 2019 as an upward flowing spiral (Figure 1.5).

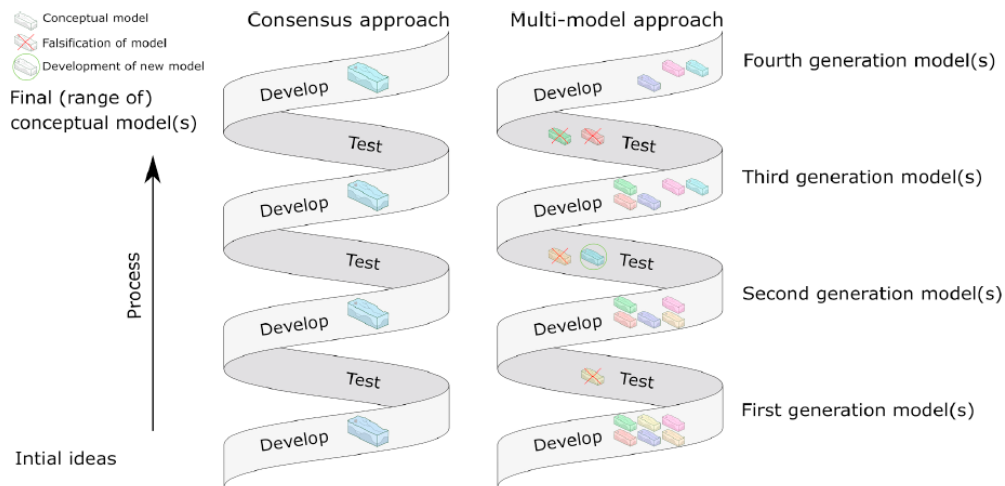


Figure 1.5 Iterative process for the conceptual modelling through the consensus or multi-model approach, from Enemark et al., 2019.

Of course, the computation of the water balance only, to validate a conceptual model is an extremely simplified approach, and the common tendency is to build models by integrating as much information as possible (hydrochemical and isotopic data, groundwater level oscillation, artificial tracer tests etc.). In this context two approaches can be distinguished: the multi-model approach (Neuman et al., 2003) and the consensus approach (Barnett et al., 2012; Izady et al., 2014). The first one considers the results obtained by a single method to construct a conceptual model, and each method provide a model itself, if one model contradicts the results obtained by previous methods, the model is entirely discarded. Contrariwise, the latter, is based on all available observations and data which are iteratively integrated into a single conceptual model, providing a staircase of confidence (Gedeon et al., 2013). In the consensus approach a much more detailed model at each turn of the spiral is achieved by adding, step by step, information obtained from each applied hydrogeological technique (Brassington & Younger, 2010).

1.3 Aims and motivations of this PhD research project

The southern and central Apennines chain (Italy) is characterised by the presence of powerful carbonate geological formations hosting the main mountain aquifers. The groundwater, drained from these aquifers by the local water management companies, is used for drinking and agricultural purposes and it is distributed both in small villages and coastal cities even located several tens of kilometres away through a well-developed aqueduct system. In mountainous areas, groundwater is also used for hydroelectric power plants and secondarily it provides a livelihood for many fish farms, whose local economy is strictly related on water availability. The geological evolution of this carbonatic domains caused an extremely complex stratigraphic and tectonic configuration reflected in the hydrogeological setting. Moreover, the recent seismic activity, related to the extensional fault systems, can be responsible for hydrogeological modification affecting the groundwater availability and movement within the aquifers.

The Sibillini Mts. Massif was struck by a very intense seismic sequence between August 2016 and January 2017, which caused visible changes in the hydrogeological characteristics of the area. Following the main earthquakes, wide research project and consultancy agreements have been funded by national and local authorities, and private agencies. In this context, various research groups and Universities, coordinated in a single table, have tried to provide specific answers on the hydrogeological modifications induced by the seismic events to improve groundwater management strategies.

This PhD project has been completely integrated within this research framework with the aim to investigate the post-seismic modifications providing an interpretation on the hydrogeological behaviour played by the fault system of the area both during the pre- and post-seismic period.

Besides, using a combined approach based on direct measurements (geological field surveys, discharge data, hydrogeological balance, and hydrochemical sampling) and indirect investigations, such as GIS based analysis, natural tracers data and artificial tracer tests coupled with advanced statistical methods, an attempt was made to deepen the knowledge of mountain carbonate aquifers characterised by such a complex geo-structural setting.

1.4 Flow-chart of the research method

Principles of meteo-climatic analysis, tracing techniques (natural and artificial tracers), GIS based analysis, geological field work and statistics are some of the disciplines involved in this research. Each scientific discipline has been used in workflow at different steps and integrated within the others in an interdisciplinary context (Figure 1.6). Mainly, the thesis is developed by investigating hydrogeological processes at different scales (regional and local). In the research development, the results obtained from the individual methods had integrated each other into an overall view to determine the hydrogeological and hydrodynamic characteristics of the investigated carbonate system. The post-seismic variations were evaluated both in the transient (short-term) and long-term periods. Finally, the field observations, together with the results obtained from the hydrogeological investigation techniques, allowed to develop a hydrogeological conceptual model for the area.

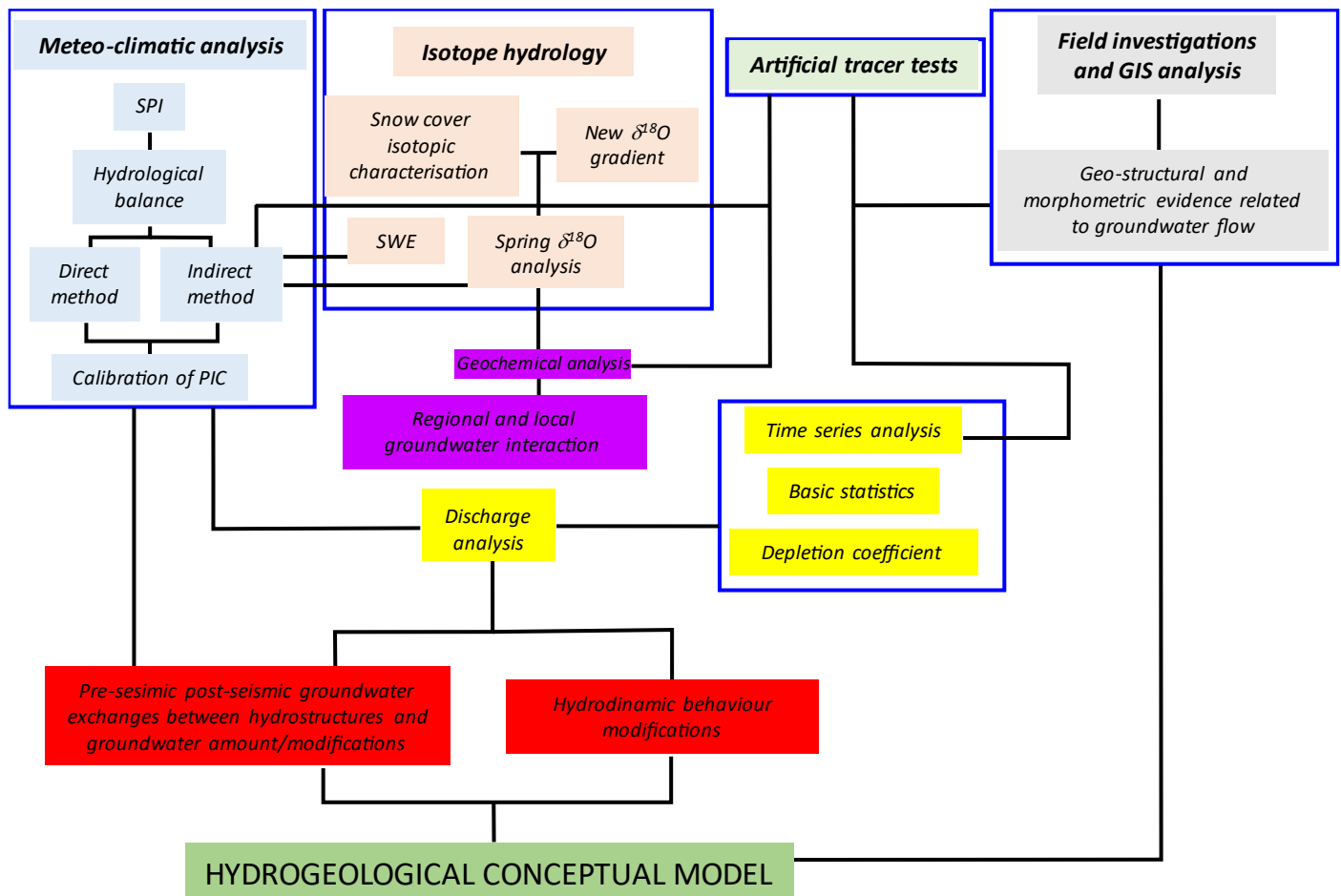


Figure 1.6 Flow-chart of the research method.

1.5 Geological and hydrogeological setting of the study area

1.5.1 Geo-lithological features and tectonic lineaments

From a geological point of view, the study area belongs to the Umbria-Marche stratigraphic succession (U-M). It consists of carbonate multi-layer geological formations (Fms.) (Meso-Cenozoic) and siliciclastic foredeep deposits namely Laga Fm. The succession starts with the deposition of Triassic Evaporites and Dolomites (Anidriti di Burano Fm.) deposited over the Crystalline Basement. Both the Anidriti di Burano and the Crystalline Basement never outcrop in the U-M domain. The sedimentary sequence continues upward with the deposition of the Calcare Massiccio Fm., with a variable thickness of 600-700 m. This is the oldest formation outcropping in the U-M domain and it is mainly composed by platform limestones. During the Lower Jurassic, the transtensional oblique and transversal fault systems (Boccaletti et al. 2005; Carminati & Doglioni, 2012; Centamore & Rossi, 2009) was responsible for the carbonate platform disarticulation in different domains. The Jurassic tectonic displacements give rise to horst and graben structure characterising a different sedimentation respectively defined to as complete, condensed and lacunose, accordingly to the deposition environments. In particular, the deposition of the complete sequence (Lower Jurassic–Upper Jurassic) occurred in the deepest sectors (graben) above the Calcare Massiccio Fm. and is observed in the limestones and marls of the Corniola Fm. (thickness of 200 m), Marne of Mt. Serrone Fm. (thickness up to 150 m), Rosso Ammonitico Fm. (thickness of approximately 40 m), Calcari a Posidonia Fm. (up to 200 m thick) and Calcari Diasprini Fm. (up to 150 m thick). The Bugarone Fm. was deposited in the horst structures above the Calcare Massiccio Fm. and its thickness is approximately 40 m. During the Late Jurassic until the Early Cretaceous period, the sedimentation continued with the pelagic deposition of the Maiolica Fm., a powerful limestone deposit ranging between 150 to 450 m thickness. Above the Maiolica Fm. the stratigraphic sequence is represented by the marls of the Marne a Fucoidi Fm. (50–80 m thick). During the Late Cretaceous the sedimentation continued with the deposition of the Scaglia Bianca Fm. (50- 70 m thickness) and the Scaglia Rossa Fm. (200-400 m) sub divided itself into four members where the limestone, marly limestone and marls percentages varied vertically during the deposition. Starting from the Late Eocene, the marls and marly limestones of the Scaglia Variegata Fm. (50 m thick), the Scaglia Cinerea Fm. (200 m thick), the Bisciaro Fm. (from 50 to 150 m thick) and the Schlier Fm. (from 80 to 200 m thick) were deposited. The Schlier Fm. was followed by the deposition of a powerful succession consisting of siliciclastic turbiditic deposits. The overall thickness of the carbonate sequence above

the Calcare Massiccio Fm. varies due to the presence of complete or condensed Jurassic sequences, reaching a maximum value of 1900 m (Pierantoni et al., 2013). In several wide areas of the U-M Apennines domain, continental Quaternary deposits (river and torrential alluvial deposits, lacustrine, glacial and slope deposits) were deposited (Figure 1.7).

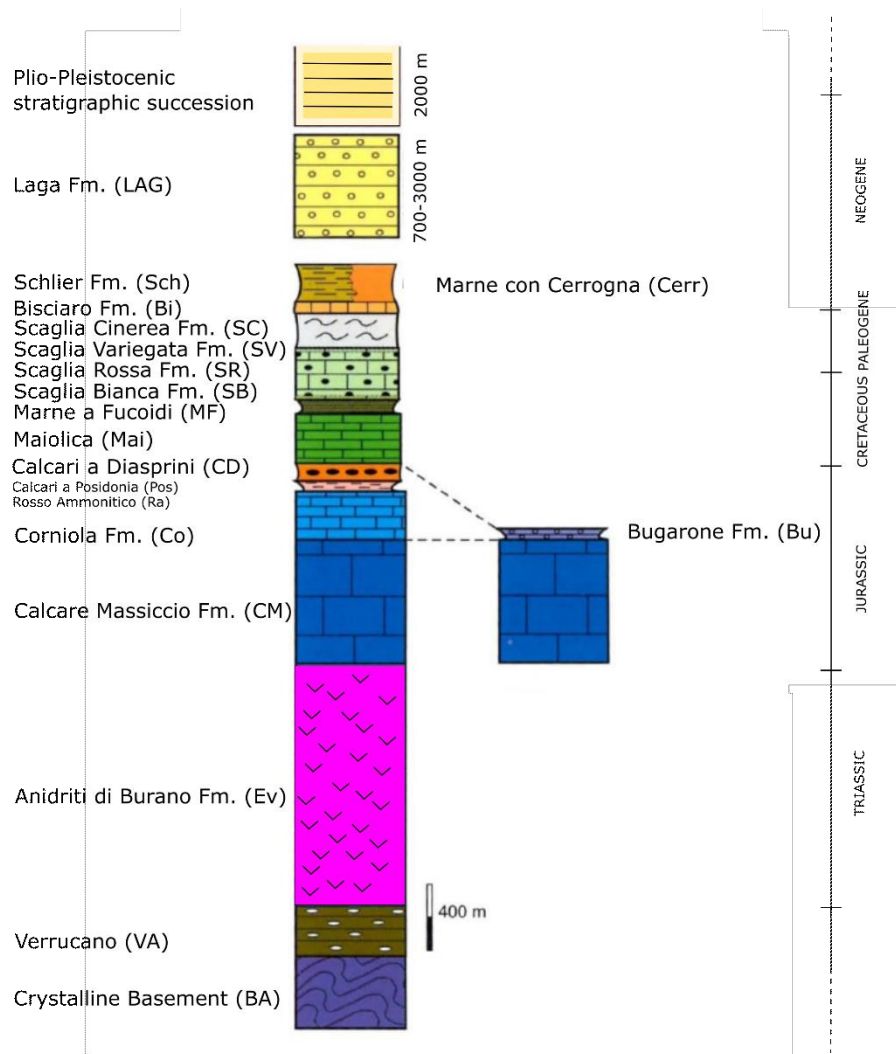


Figure 1.7 Stratigraphic column of geological formations belonging to the Umbria-Marche domain, modified from Deiana & Marchegiani, 2002.

From a geo-structural point of view, the U-M domain is characterized by multiple overthrusts, associated with antiformal and synformal folds, with arcuate shape and ENE vergence. The folds and faults system were originated following the collision between the European (Corsica-Sardinian block) and Adria plate (African affinity) starting from the Oligocene and they developed, in several phases, during the Neogene-Quaternary (Barchi et al., 1986; Boccaletti, 1986; Deiana, & Pialli, 1994; Barchi & Mirabella, 2009; Calamita & Deiana, 1988; Calamita et al., 1994; Cello et al., 1997; Finetti et

al., 2005; Lavecchia et al., 1994, 2003; Mazzoli et al., 2005, 2000; Pizzi et al., 2002). The complex structural setting of the central-northern Apennines is determined also by the tectonic lineaments of the pre-existing paleo geographic domains of the passive margin of the Adriatic plate involved in the formation of the U-M chain. (Carmignani & Kligfield, 1990; Doglioni, 1991; Coward et al., 1999; Boccaletti et al., 2005; Calamita et al., 2009; Di Bucci & Mazzoli, 2002; Finetti, 2005; Malinverno & Ryan, 1986; Mantovani et al., 2009; Scisciani & Calamita, 2009; Scisciani et al., 2002, 2010; Di Domenica et al., 2012) An important role in the development of the actual structural setting of U-M Apennines chain was played by the Jurassic extensional syn-sedimentary tectonic phase responsible for the subdivision of the Triassic carbonate platform into different domains separated by oblique transtensive and transversal fault systems. This phase was followed by the development of ridges and depressions, bordered by normal faults, which conditioned the Cretaceous-Neogene sedimentation of the central-northern Apennines as well as its structural evolution. The current structural setting of the Apennines carbonate ridges is the result of the Jurassic processes of tectonic inversion, both positive and negative, that led to the chain formation, with the reactivation of fault planes (listric faults) and overthrusts (Tavarnelli et al., 2004; Calamita et al., 2003, 2009, 2011; Di Domenica et al., 2012, 2014; Pace & Calamita, 2014).

The U-M carbonatic domain is bordered in the southernmost portion by the Messinian – Early - Pliocene Olevano Romano-Antrudoco-Monti Sibillini thrust fault (named for simplicity Sibillini Mts. overthrust) (Calamita et al., 1990, 2012). This thrust fault is characterized by a NNW-SSE strike direction in the southern portion and is dissected by transtensive faults. Moving northward the Sibillini Mts. overthrust rotates through a N-S strike direction. This is the most important tectonic lineament which separates the carbonate mountain chain from the Mio - Pliocene terrigenous deposits of the Pedepennine Adriatic Foredeep. (Lavecchia, 1985; Calamita & Deiana, 1988; Mazzoli et al., 2005; Pierantoni et al., 2005, 2013; Di Domenica et al., 2012). The U-M Apennines is constituted by two main groups of folds and overthrusts: the Umbria-Marche ridge (Inner Ridge) and the Marche ridge (Outer Ridge), separated by the Acqualagna-Visso tectonic depression in which Miocene deposits outcrop. The latter is marked by the sparse presence of minor carbonate reliefs, and it is characterised by a synclinal and anticlinal pattern (Mirabella et al., 2008). The Outer Ridge is constituted by an anticlinal macroscopic structure, while the Inner Ridge is characterised by a fold system of anticlinal and asymmetrical syncline, with a typical “en-echelon” or overlapped folds

(Galadini et al., 2009). The Inner Ridge and the Outer one converges in the southernmost portion originating the Sibillini Mts. structure (Figure 1.8).

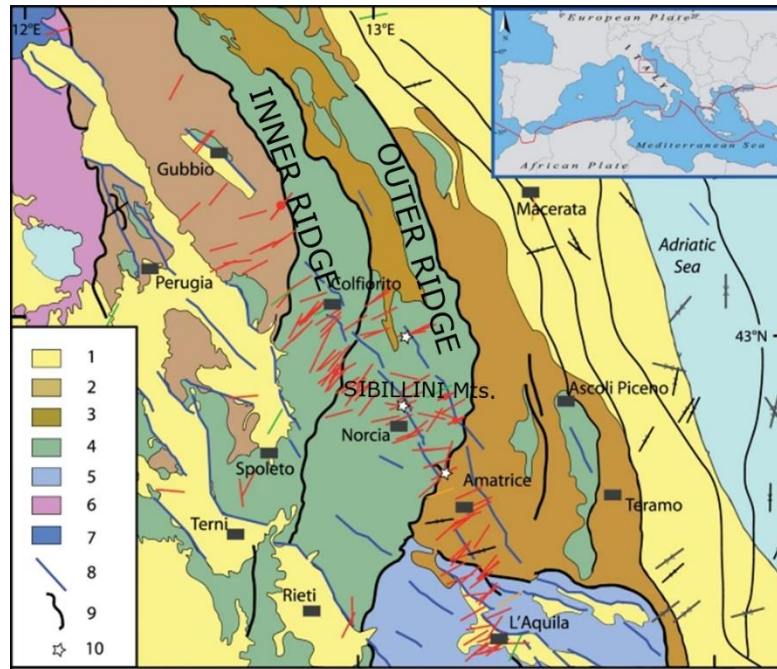


Figure 1.8 Tectonic setting of the Umbria-Marche domain and location of Sibillini Mts. Massif, modified from Montone & Mariucci, 2020. Key: 1) marine and continental Plio-Quaternary sediments; 2) Marnoso Arenacea Formation; 3) Laga; 4) Carbonate sedimentary sequence of the Umbria–Marche domain; 5) Carbonate sedimentary sequence of the Latium–Abruzzo domain; 6) Tuscany nappe units; 7) Liguridi unit; 8) normal fault; 9) thrust fault; 10) Three mainshocks of the 2016 central Italy seismic sequence.

The Quaternary extensional tectonic activity, partly developed from the reactivation of inherited tectonic faults of previous compressional phases, strongly influenced and still influencing the structural evolution of the Apennines chain (Tavarnelli et al., 2004; Butler et al., 2006; Scisciani, 2009; Di Domenica et al., 2012)

The axial zone of the chain (Inner Ridge and Sibillini Mts.) is affected by a Quaternary extensional and transtensional tectonics with high angle fault system and NW-SE strike direction, characterized by faults lengths between 15 and 35 km (Galadini et al., 2009). This system of normal faults was developed and driven by the reactivation of the “en-echelon” fold system.

The activity of these extensional faults, bordering the western slope of the Sibillini Mts., is characterized by hundreds of meters of rejection and is responsible for the formation of intramountain tectonic depressions characterized by powerful Plio-Pleistocene continental deposits (Aringoli et al., 2021). This extensional fault system characterises the seismicity of the central

Apennines as evidenced by the earthquakes that over time have affected the regions of central and northern Italy (Boschi et al., 2000; Barchi & Mirabella, 2009; Emergeo W.G. et al., 2016).

Focusing the attention to the the Sibillini Mts. domain (selected as study area for this PhD research), it is characterized by the outcropping of the carbonate terms of the U-M sedimentary succession, starting from the Jurassic Calcare Massiccio Fm. to the Mio - Pliocenic terrigenous sediments of the Laga Fm. outcropping eastward to the Sibillini Mts. overthrust. The Calcare Massiccio and Corniola Fms. are characterised by a well-developed karstic system. In fact, shallow karst forms such as karren, rillenkarren, kamenitza and sinkholes are recognised in the area (Coltorti and Farabollini 1995); while a lack of knowledge of deep karst systems has to be highlighted due to the absence of direct human inspections (Nanni et al., 2020). The Maiolica Fm. and the Scaglia Fm. are characterised by stratified mycritic limestones and limestones, marly, flinty limestones respectively and the presence of karst forms, such as dolines is observed but they are less developed than the ones in Calcare Massiccio and Corniola Fms. (Coltorti and Farabollini 1995). The Sibillini Mts. area is characterised by an association of folds, overthrust, and faults, such as the entire U-M Apennines domain. The Miocenic compressive structures are dislocated by extensive Quaternary faults (Mazzoli et al., 2005; Scisciani, 2009). These faults are marked by the typical indicators of recent activity (Mazzoli et al., 2005; Di Domenica et al., 2012) and they have a main NNW-SSE strike direction (Aringoli et al., 2016).

The Quaternary faults are responsible for the morpho-structural setting of the Sibillini Mts. area. In particular, three structural sectors have been recognised with blocks lowering and forming steps moving from the Adriatic side (east) to the Norcia Plain to the west. The first block (eastern sector) is comprised between the Sibillini Mts. overthrust and the Mt. Bove – Mt. Vettore normal fault system (VBFS), with a total displacement of about 1500 m (Calamita et al., 1992, Lavecchia et al., 1994; Pierantoni et al., 2013). In this sector the highest reliefs are present, and the elevation frequently exceeds 2000 m above sea level (a.s.l.) the highest point is represented by the Mt. Vettore (2476 m a.s.l.) The central block is comprised between the VBFS to the east and the normal fault that borders the Norcia Plain to the west named Nottoria - Preci fault. This sector is characterised by the presence of a peculiar intramountain tectonic-karstic depression named Pian Grande Plain (approximately 1300 m a.s.l.). The latter is filled by Pleistocene to Holocene fluvial-lacustrine deposits which overlies the bedrock units mainly represented by the limestones of Calcare Massiccio and Maiolica Fms. The western block, in which the Norcia Village is located, is characterised by the presence of powerful quaternary fluvio-lacustrine deposits similar to the ones of Pian Grande Plain and it develops at

about 600-700 m a.s.l. The aforementioned structural sectors show a general NW folds dipping direction towards the Nera River Upper Basin causing the outcropping of the Calcare Massiccio and Maiolica Fms. at higher elevation in the eastern and southern areas (Mt. Vettore and Mt. Bove), on the contrary, moving northward, the same geological formations emerge near the Upper Nera River at lower elevation, between 600-900 m a.s.l. Usually these folds are released by transpressive N-S and E-W oriented faults, which also involves the thrust faults systems (Figure 1.9).

1.5.2 Seismicity of the area

As previously introduced, the Sibillini Mts. area is characterised by several Quaternary active faults. In this context the main active tectonic structure is represented by the Mt. Vettore - Mt. Bove fault systems (VBFS) (Calamita et al., 1992; Calamita & Pizzi, 1992, 1994; Cello et al., 1997; Pizzi et al., 2002, Galadini et al., 2009).

The territory has been historically characterized by multiple strong earthquakes (Table 1.1), some of which have occurred within complexes seismic sequences (Emergeo W.G. et al., 2016). In fact, the seismic history of the area is relatively well known, but the levels of knowledge completeness vary between the different sectors.

More than 228 earthquakes ($M_w \geq 4.0$) have been recorded in the area starting from the 1328 (Rovida et al., 2019). Since the advent of the Italian National Seismic Network, in 1980, more than 163 earthquakes $M_w \geq 3.0$ have been recorded, with a distribution of 15 earthquakes with $M_w \geq 5.0$.

The recent work of Di Matteo et al., 2020 briefly describe the historical seismicity of the area recognising among the oldest earthquakes the one occurred on December 4, 1328, that affected Norcia village and others small centers. In the same area, between the autumn 1599 and January 1600, several earthquakes of similar intensity occurred between Norcia and Cascia, causing the collapsing of tens houses in the Cascia village. A very strong seismic sequence occurred between October 7th and October 14th 1639 near Amatrice village, causing about 500 deaths. The year 1703 is marked by one of the most significant seismic sequence involving a vast area of Umbria and Lazio regions. It occurred between January and February 1703, and nowadays is still recognised as one of the most significant seismic sequence of the last millennium for Central Italy. Then, the western area of the Sibillini Mts. was struck by strong earthquakes in, 1719, 1730 and 1815. In the morning of February 14th 1838, an earthquake located near Norcia generated greatest damages to the structures of the area. Starting from the end of August 1859, the area was stroked by a seismic sequence lasting about a year with the occurrence of two important earthquakes in November 1859 and in May 1860. Norcia and Visso areas have been interested by two earthquakes in February 1879 and in September 1898 respectively. Then, in 1979 a moderate seismic sequence affected some villages in the Valnerina area.

Table 1.1 Historical strong earthquakes affecting the study area. Io (Epicentral intensity - MCS scale), Me (Equivalent Magnitude). Modified from Di Matteo et al., 2020.

Date	Me	Io
4/12/1328	6.4	10
6/11/1599	6.0	9
8/10/1639	6.1	10
15/10/1639	6.2	10
14/01/1703	6.7	11
16/01/1703	6.0	11
27/6/1719	5.5	8
12/5/1730	5.9	9
3/9/1815	5.5	8
14/2/1838	5.3	8
22/8/1859	5.8	9
23/2/1879	5.6	8
12/9/1898	5.3	8
16/11/1916	5.5	8
19/9/1979	5.8	8.5

The recent activity of the VBFS has been studied and demonstrated by several geomorphological and paleoseismological studies (Calamita et al., 1992; Lavecchia et al., 1994; Calamita & Pizzi, 1994; Blumetti & Guerrier, 2007; Coltorti & Farabollini, 1995; Cello et al., 1998; Galadini & Galli, 2003; Galli et al., 2005). The VBFS is the easternmost fault system in the Central Apennines; it is characterized by WSW dipping normal faults, 30 Km long, with “en-echelon” pattern, and in some cases, it is intersected by transfer faults. The VBFS is extended from Mt. Bove to the north until Mt. Vettore to the south, intersecting the Sibillini Mts. overthrust.

Since 24th August 2016, Sibillini Mts. have been interested by multiple earthquakes generating important damages in villages and towns, causing almost 300 fatalities and more than 20000 homeless (Civico et al., 2018; Emergeo W.G. et al., 2016). The seismic sequence (Chiaraluce et al., 2017) started with an Mw 6.0 mainshock at 1 Km west of Accumoli (Lazio Region). Subsequently, on 26th October 2016, a new mainshock of Mw 5.9 occurred at 3 Km NW of Castelsantangelo sul Nera (Marche Region), followed by the largest shock of the seismic sequence (Mw 6.5), on 30th October

2016, located about 5 Km NE of Norcia village (Umbria Region). These have been the strongest seismic event in Italy since the Irpinia earthquake ($M_{\max}$ 6.9) in 1980 (Westaway & Jackson, 1987; Bernard & Zollo, 1989). Additional mainshocks occurred in the southern sector (Abruzzo Region) on 18th January 2017, with a $M_{\max}$ of 5.5. The shocks are confined to the upper crust (9 - 12 Km maximum depth) and follow a roughly NW-SE trend for about 80 Km between the towns of Camerino to the north and Pizzoli to the south (Chiaraluce et al., 2017).

After the most important seismic events, significant surface ruptures were observed along the VBFS. In particular, after the 24th of August 2016, Amatrice earthquake a N155°-trending ruptures, mostly SW dip-slip kinematics were observed and mapped for several kilometres along the southern portion of the VBFS (Emergeo W.G. et al., 2016; Lavecchia et al., 2016). These co-seismic features were interpreted by Livio et al., 2016; Aringoli et al., 2016; Pucci et al., 2017 as the response of primary surface faulting. The earthquake of 26th October 2016 caused sparse and discontinuous co-seismic surface ruptures along the northern portion of the VBFS. Finally, the October 30th Mw 6.5 mainshock close to Norcia produced the widest co-seismic surface effects recognised on an area over 400 Km². The principal effects consisted in primary surface ruptures while the secondary effects, like rock falls and landslides, occurred in the steepest slopes of the Sibillini Mts. massif. The pattern of primary surface ruptures was observed for an overall length of 20-25 km along the whole VBFS (Figure 1.10). The mapped surface rupture typically follows the trace of previously mapped faults (Pierantoni et al., 2013 and references therein) and only in some locations the co-seismic ruptures occurred along fault splays that were not previously recognized. Field observations and remote surveys performed after the October 30th 2016, earthquake revealed that its co-seismic surface ruptures can be considered one of the most complex recorded in Italy and in the Mediterranean area in the past 40 years in terms of number of involved fault splays, in a normal faulting earthquake context (Civico et al. 2018; Emergeo W.G. et al., 2016; Villani et al. 2021).

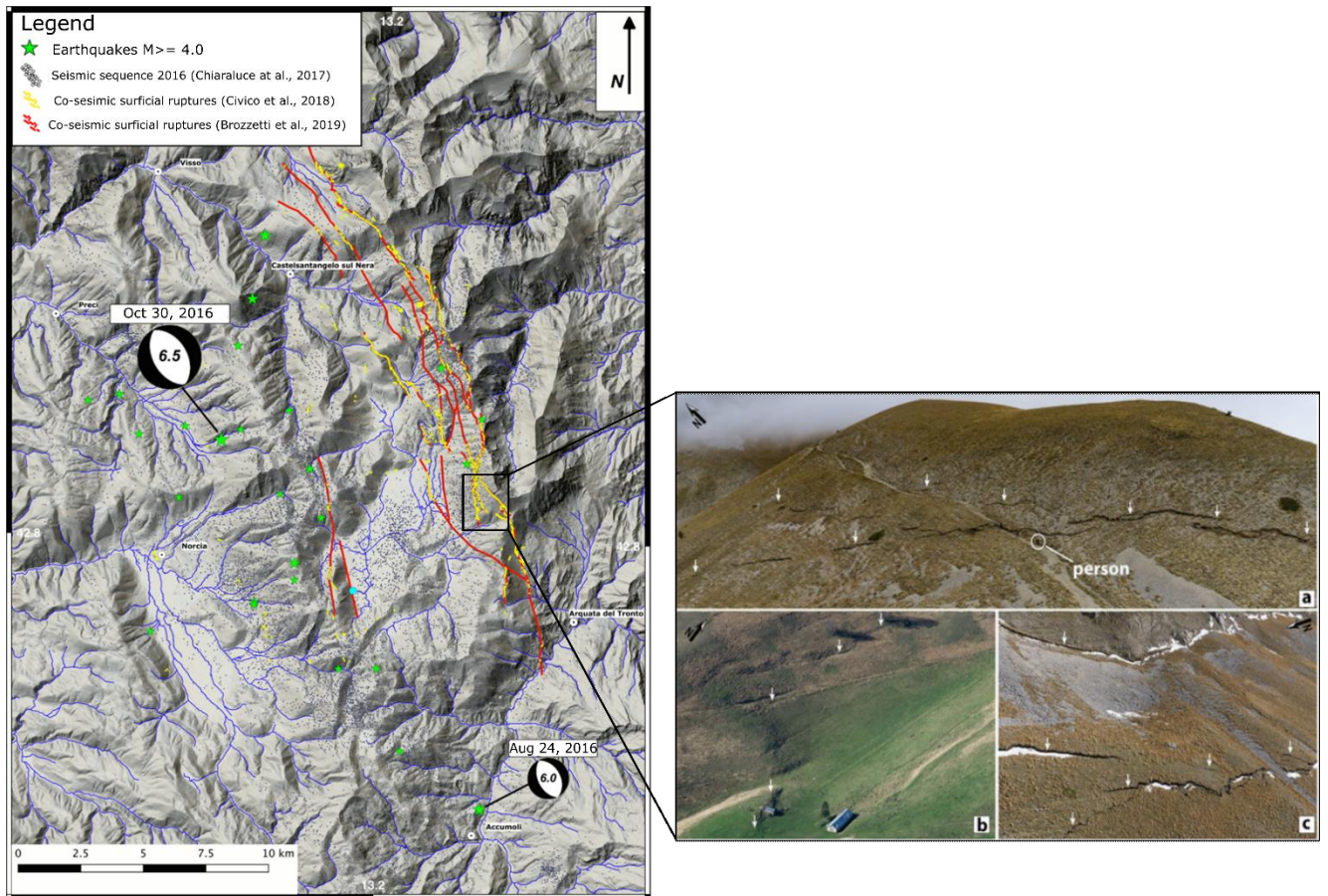


Figure 1.10 Location of main earthquakes occurred in the area during the 2016-2017 seismic sequence with pattern of primary surface ruptures observed for an overall length of 20-25 km along the Mt. Vettore - Mt. Bove fault system, modified from Emergo W.G. et al., 2016.

1.5.3 Climatic and meteorological features

The mountainous region of the Sibillini National Park, despite its important role from the environmental and naturalistic point of view, has not been the site of specific climatic and meteorological studies. In fact, according to the literature of my knowledge, the most complete study in the area is represented by the work performed by Amici & Spina, 2002 and Spina et al., 2002 which, however, is developed on a regional scale and more specifically for the Marche Region.

To define the meteo-climatic characteristics of Marche Region, 102 rainfall stations and 24 temperature stations have been selected by the authors. A statistical study on the trend of temperature and precipitation has been developed, temperature and rainfall trends have been calculated along a period between 1950 and 2000 showing a general decrease in precipitation and an increase in mean annual temperature values. Moreover, in order to study the climatic features of the Marche Region and to identify climatologically similar areas, the method of Thornthwaite, 1957 has been applied using precipitation, temperature, solar radiation and heliophany data for the period 1950-1989, available for a total of 29 stations. The results of the studies conducted by Amici & Spina, 2002 and Spina et al., 2002 highlighted that the Sibillini Mts. is one of the rainiest area in Marche Region with mean annual rainfall of about 1500-1550 mm, this result is in accordance with the one observed by (Boni et al., 1986, 2009; Tarragoni, 2006). In particular, the autumn is the rainiest season (420-450 mm), while the summer season is the driest period (240-255 mm). As concerns the temperature, only the thermometric stations of Montemonaco and Pintura di Bolognola, 987 m and 1380 m a.s.l., respectively, are located in the Sibillini Mts. domain. They show a mean annual temperature of 11.0 °C and 6.8 °C with a minimum value of -8.9 °C recorded by the Pintura di Bolognola thermometric station during the observed period. The maximum temperature in the area has been recorded during July and August, while the minimum temperature has been recorded during winter season (January-February). According to the Thornthwaite climatic classification, the mountainous portion of Marche Region is a humid zone, in particular, Montemonaco station has been classified as humid, second meso-thermal zone with adequate rainfall in all the seasons (B₃B₂'ra').

1.5.4 Hydrogeological features

The main groundwater resources of the central Apennines are located at the boundaries of large carbonate hydrostructures, near important tectonic elements and/or stratigraphic boundaries between formations characterised by different permeability. As far as the U-M domain is concerned, the groundwater circulation is influenced both by the structural characteristics, typical of a chain with folds and overthrusts, and by the stratigraphic setting. The influence of tectonics and litho-structural features on the aquifers characteristics of the U-M domain has been pointed out by Nanni & Vivalda, 2005. They observed that faults and folds create the hydrogeological conditions that favour on one hand the aquifer boundaries, on the other hand permits the groundwater mixing between different aquifers. In particular, the regional Basal aquifer (**BAS**) is hosted by the Calcare Massiccio and Corniola Fms., and is characterised by high infiltration rate and high permeability due to its well-developed karstic system. **BAS** recharges several punctual springs, and is responsible for the discharge increasing along the watercourses due to the intersection between the riverbed and the saturated portion of the aquifer, resulting the so called linear springs characterised by high discharge, widely occurring in the entire Sibillini Mts. area (Fronzi et al., 2021). The **BAS** aquifer is inferiorly confined by the Triassic Anidriti di Burano geological Fm. characterised by low permeability evaporites, acting as a lower regional aquiclude (**TRI**). The Marne of Mt. Serrone, Rosso Ammonitico, Calcari a Posidonia, Calcari Diasprini and the Bugarone Fms., that overlie the **BAS** accordingly to the horst and graben structures, act as an aquiclude complex (**JUR**) at local scale separating **BAS** from the above Maiolica aquifer complex (**MAI**). At regional scale, **BAS** and **MAI** can be stratigraphically in contact, constituting a single regional aquifer. **MAI** is regionally separated from the Scaglia Calcareo aquifer complex (**SCA**) by the aquiclude complex made by Marne a Fucoidi Fm. (**MF**). **SCA** aquifer is hosted by Scaglia Bianca and Scaglia Rossa Fms. and is responsible for local groundwater recharge of linear or punctual springs of the area, such as it is **MAI**, when separated from **BAS**. Both **MAI** and **SCA** are characterised by high to medium-high secondary permeability due to a well-developed fractures and fissures systems linked to the stratification and joints. **SCA** is overlain by the Scaglia Variegata, Scaglia Cinerea, Bisciaro and Schlier Fms. heteropic with Marne con Cerroghna Fm., closing the Umbria-Marche Succession. These formations act as upper aquiclude complex (**SVC**) and are followed by the Flysch della Laga aquitard (**LAG**) (Figure 1.11).

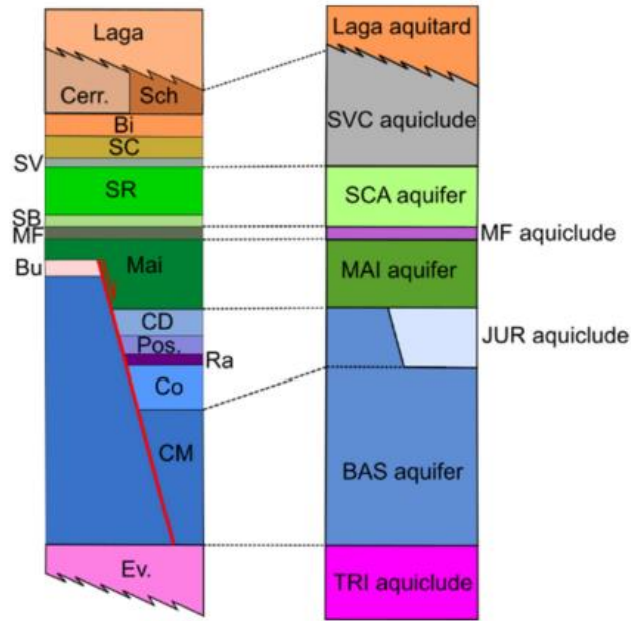


Figure 1.11 Sketch of relationships between the simplified stratigraphy (left) and hydrogeological complexes (right) (not in scale, modified from (Viaroli et al., 2021). Ev: dolomite and evaporite sequence, CM: Calcare Massiccio, Co: Corniola, Ra: Rosso Ammonitico, Pos: Marne a Posidonia, CD: Diaspri, Bu: Bugarone, Mai: Maiolica, MF: Marne a Fucoidi, SB: Scaglia Bianca, SR: Scaglia Rossa, SV: Scaglia Variegata, SC: Scaglia Cinerea, Bi: Bisciaro, Cerr: Marne a Cerrognola, Sch: Schlier, Laga: Flysch della Laga. Hydrogeological complexes: TRI: Dolomitic-Evaporitic aquiclude, BAS: Basal aquifer, JUR Jurassic aquiclude, MAI: Maiolica aquifer, MF: Marne a Fucoidi aquiclude, SCA: Calcareous Scaglia aquifer, SVC: Upper aquiclude, Laga aquitard.

In this stratigraphic context, the groundwater circulation is deeply influenced by the structural setting described in paragraph 1.5.1. The core of anticlines and synclines generally represent recharge areas stratigraphically delimited by low permeability formations in correspondence of which springs often emerge. The compressive structures, such as the thrust faults, generally act as no-flow boundaries, whereas the role of the extensional faults is more debated; they can put in contact different aquifers acting as preferential drainage paths, otherwise isolated from each other for stratigraphic reasons, or they can act as no-flow boundaries (Fronzi et al., 2021).

Many studies were performed in the Sibillini Mts. during the past decades. Starting from the 1991, many research activities have considerably deepened the knowledge in this area and the results obtained from hydrogeological investigations carried out for over twenty years of studies are widely described in numerous works reported in bibliography (Boni et al., 1991; Regione Marche, 1991; Boni & Petitta, 1994; Boni & Preziosi, 1994; Boila & Checcucci, 1995; Mastroirillo, 1996; Mastroirillo, 2001; Boscherini et al., 2005; Tarragoni, 2005, 2006; Boni & Petitta, 2007; Petitta, 2007, Preziosi, 2007; Boni & Petitta, 2008; Boni et al., 2009; Mastroirillo et al., 2009; Mastroirillo et al., 2012; Banzato et al., 2013, Nanni et al., 2013; Banzato, 2014).

Nowadays, this study still remain the factual base to perform hydrogeological studies in the area. In fact, it led to the realization of a hydrogeological map in which the aquifers feeding the main punctual springs and linear spring in the territory have been identified. The study method involves the comparison between the hydro-structural features (which outlines the geometry of the aquifers), the springs distribution and their discharge. The result of this study was a specific conceptual hydrogeological model in which the recharge areas, the main flow directions and the position and nature of hydraulic boundaries between contiguous aquifers were defined. Moreover, the geological formations outcropping in the Sibillini Mts. chain have been classified into hydrogeological complexes based on their capability to infiltrate and store meteoric water.

Boni & Bono, 1982 and Boni et al., 1986 proposed a direct method to assess the effective infiltration, based on the principle that the amount of discharge from a hydraulically closed hydrostructure is equivalent to the average amount of water entering the recharge area. The average effective infiltration, expressed in millimeters/year, is calculated by dividing the average volume of water discharged by a hydrostructure by the extent of its recharge area; the latter is identified based on a detailed reconstruction of the local geological-structural setting (Mastrorillo & Petitta, 2010).

This method encounters some difficulties especially if the investigated hydrogeological units are characterised by deep groundwater circulation and nearby hydrostructures are hydraulically in contact each other. The unknown term is represented by hydraulic exchanges, which are evaluated by comparison between theoretical spring discharge respect to measured spring discharge, evidencing deficit or surplus (Mastrorillo et al., 2009). In this context, Mastrorillo et al., 2009 hypothesize groundwater circulation in compartmentalized hydrostructures with preferential flow directions-oriented N-S, whose geometry follows the morphology of carbonate ridges.

However, as stated by the authors, the hydrogeological model highlights an extremely heterogeneous groundwater circulation and the proposed results must be considered as interpretative hypotheses only partially verified, suggesting that further integrative studies and investigations should be performed. They concluded that in some cases, a more in-depth hydrogeological study is necessary, especially where the geological-structural setting is particularly complex.

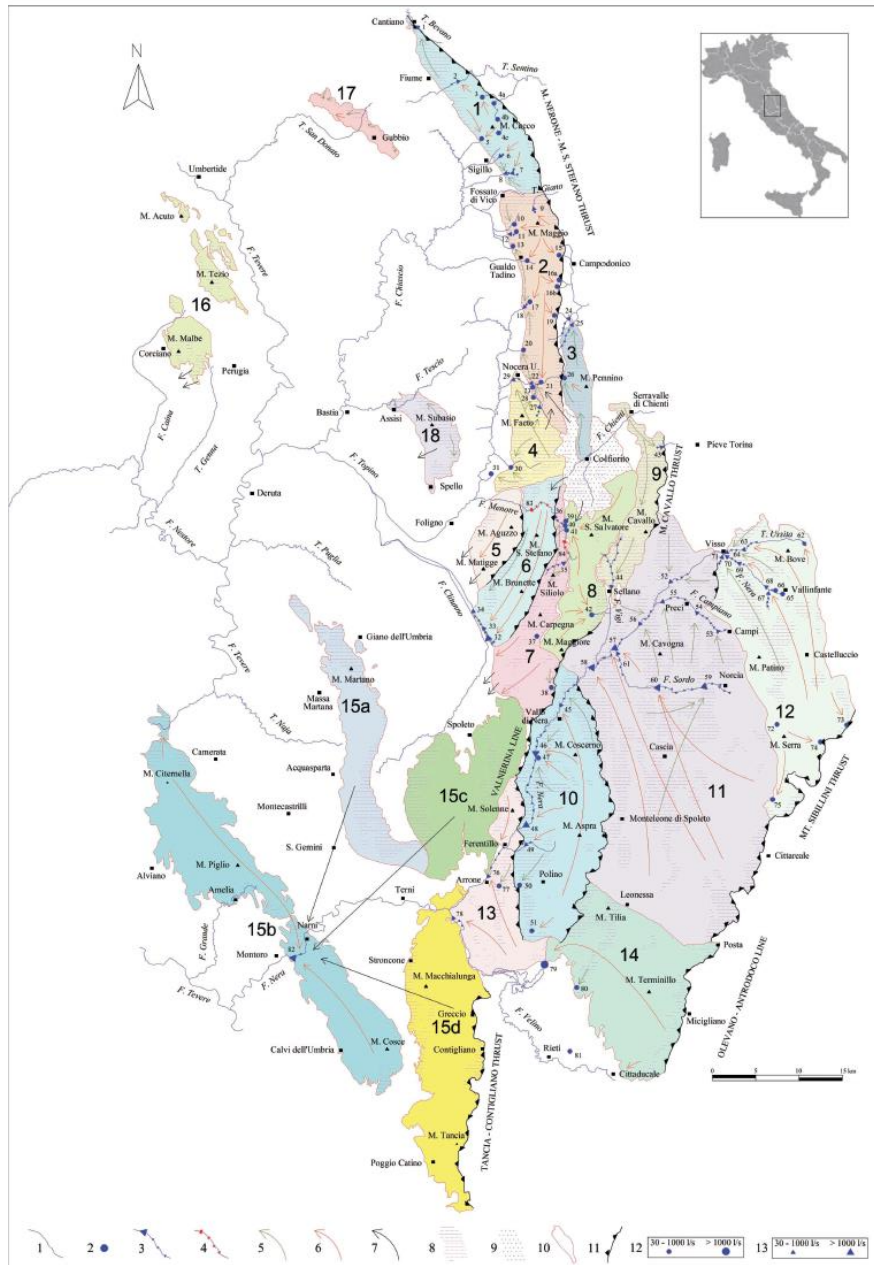


Figure 1.13 Hydrostructural map of Umbria Region, published by Mastrorillo et al., 2009. Legend: 1) Hydrographic network; 2) Perennial spring; 3) Linear spring; 4) River discharge losses; 5) Main groundwater flowpath in the "Scaglia" aquifer; 6) Main groundwater flowpath in the "basal" aquifer; 7) Main seepages between hydrogeological units; 8) Scaglia complex outcropping; 9) Preferential recharge area of Colfiorito Plain; 10) Boundary of hydrogeological units; 11) Regional overthrusts; 12) Discharge ranges for localized springs; 13) Discharge ranges for linear springs.

The historical studies conducted by Roma la Sapienza University have been recently synthesized in a new hydrogeological map (1:50000) of the area (Figure 1.14) describing the pre-seismic conditions (Viaroli et al., 2021).

The motivation behind the work was to provide an updated hydrogeological map by integrating all the data published by the same research group before the seismic sequence. The map was drafted to support the studies aimed at understanding the hydrogeological changes occurred after the 2016-2017 seismic sequence. The authors stated that the complex hydrostructural setting is marked by the presence of the normal faults that dissect the pre-existing thrusts, altering the sealing role of the latter and allowing non-negligible groundwater seepages (not quantified) at local scale.

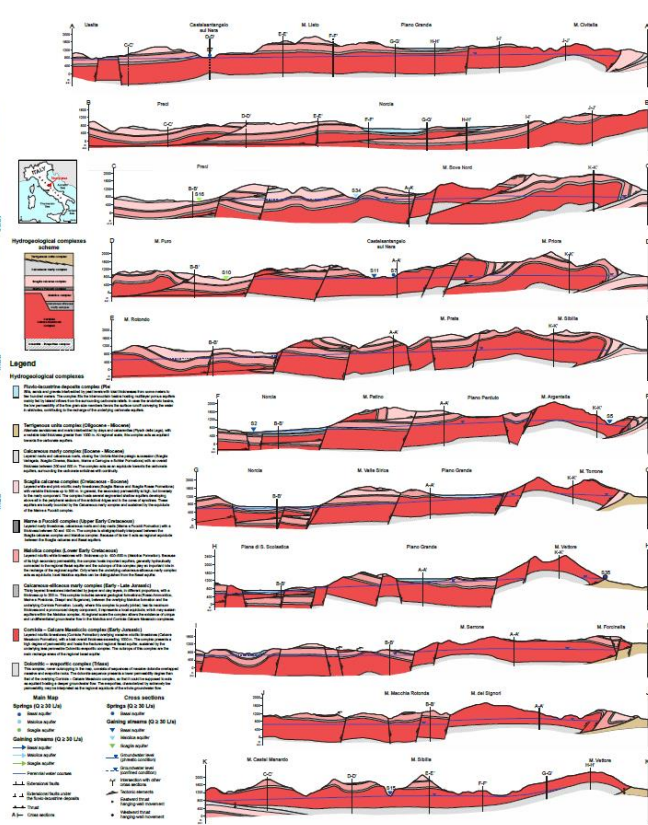


Figure 1.14 Hydrogeological map and cross-section of Sibillini Mts. – scale 1:50000, with detail of hydrostructures number (i.e., hydrostructure 1, 2 and 3) modified from Viaroli et al., 2021.

On the other hand, the map provides important background information on the pre-earthquake hydrogeological condition, highlighting the presence of three hydrostructures separated each other by the main fault systems. The eastmost one is the hydrostructure 1, it is bordered by the Sibillini Mts. overthrust to the east and the Mt. Vettore-Mt. Bove normal fault to the west. This hydrostructure is characterised by the occurrence of the higher reliefs of the area, often exceeding 2000 m a.s.l. (i.e., Mt. Vettore, Mt. Porche, Mt. Sibilla, Mt. Priora). The central hydrostructure is the number 2, and is bordered by the Mt. Vettore- Mt. Bove normal fault to the east and the Nottoria-Preci normal fault to the west. This hydrostructure comprises the tectono-karstic Pian Grande Plain (medium elevation 1300 m a.s.l.). The hydrostructure 3 is the westmost one, it is located at lower elevation with respect to the hydrostructure 1 and 2, and comprises the tectono-karstic Norcia Plain (medium elevation 600 m a.s.l.). The hydrostructure 3 is bordered by the Nottoria-Preci normal fault to the east and the Mt. Coscerno overthrust to the west acting as no-flow boundary. The described hydrogeostructural setting can be represented by a compartmentalized in-series basins system (Celico et al., 2006).

In all the hydrostructure, except for hydrostructure 2, the main regional basal flow belonging to the BAS aquifer is South-North oriented, following the geo-structural setting of the Corniola and Calcare Massiccio geological Fms (in accord to what observed by Mastorillo et al., 2009). The hydrostructure 2 is characterised by the presence of a divergent regional groundwater flow. Part of the basal flow is directed northward towards the Upper Nera River basin, while another part feeds the springs located southward, emerging along the Sibillini Mts. overthrust.

All these considerations have been drawn only on the base of past studies conducted in the area, using discharge data and hydrogeological balance methods, disregarding the presence of actual hydraulic exchanges between the hydrostructures.

The compartmentalized in-series basins system, however, seems to be in contrast to previous studies conducted in the same massif with the aim to identify single springs recharge area or quantify the groundwater resources for single hydrostructure (Petitta, 2011; Mastorillo et al., 2012; Nanni et al., 2013).

Most of the studies have been developed to provide to the local water management companies a tool for a proper management of groundwater resources, and to deepen the knowledge of the natural system of the Sibillini Mts. National Park.

Mastorillo et al., 2012, for example, studied the hydrogeological setting of the eastmost hydrostructure (i.e., hydrostructure 1). The authors espoused the hydrogeological conceptual model identified by Boni & Petitta, 2007, for the **BAS** aquifer (Figure 1.15). In general, the regional flow of

this hydrostructure is oriented from south to north, but the no-flow boundary identified in the southern portion of the hydrostructure (located in the westmost portion of the hydrostructure) seems to disappear moving northward. This trend can be evinced by the piezometric lines drawn for the **BAS** aquifer. In fact, moving further to the north of Mt. Porche a contribution coming from west seems to be present, partially questioning the no-flow boundary between hydrostructure 1 and hydrostructure 2 or the presence of a groundwater divide within hydrostructure 1.

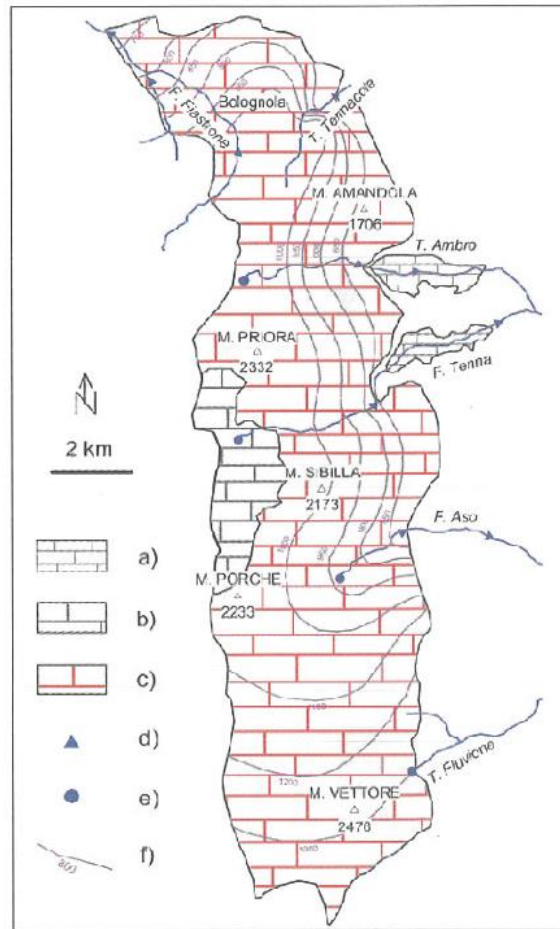


Figure 1.15 Hydrogeological conceptual model identified by Boni & Petitta, 2007, for the BAS aquifer modified by Mastroiillo et al., 2012.

Tarragoni, 2006 deepened the knowledge of hydrostructure 2 by studying exclusively the Upper Nera River basin, (i.e., the one draining the groundwater flow towards the north), leaving aside in its investigation the flow in the opposite direction belonging to the same hydrostructure feeding the springs to the south (Capodacqua and Pescara del Tronto springs). The author uses isotopic techniques based on $\delta^{18}\text{O}$ springs content. A $\delta^{18}\text{O}$ -elevation relationship for the area has been determined by sampling only minor springs, not including rain gauges or snow samples.

The spring fed by **BAS** aquifer show a mean isotopic recharge elevation between 1570 m a.s.l. and 1690 m a.s.l.. The author concludes that the higher elevation (determined by the springs $\delta^{18}\text{O}$ content) with respect to the average elevation of recharge area (1343 m above sea level), is due to the snowmelt component without assuming that part of the flow could come from the runoff involving the western steep slopes of the Vettore Mt. and then infiltrates at its downslope in the Pian Grande Plain (belonging to hydrostructure 2).

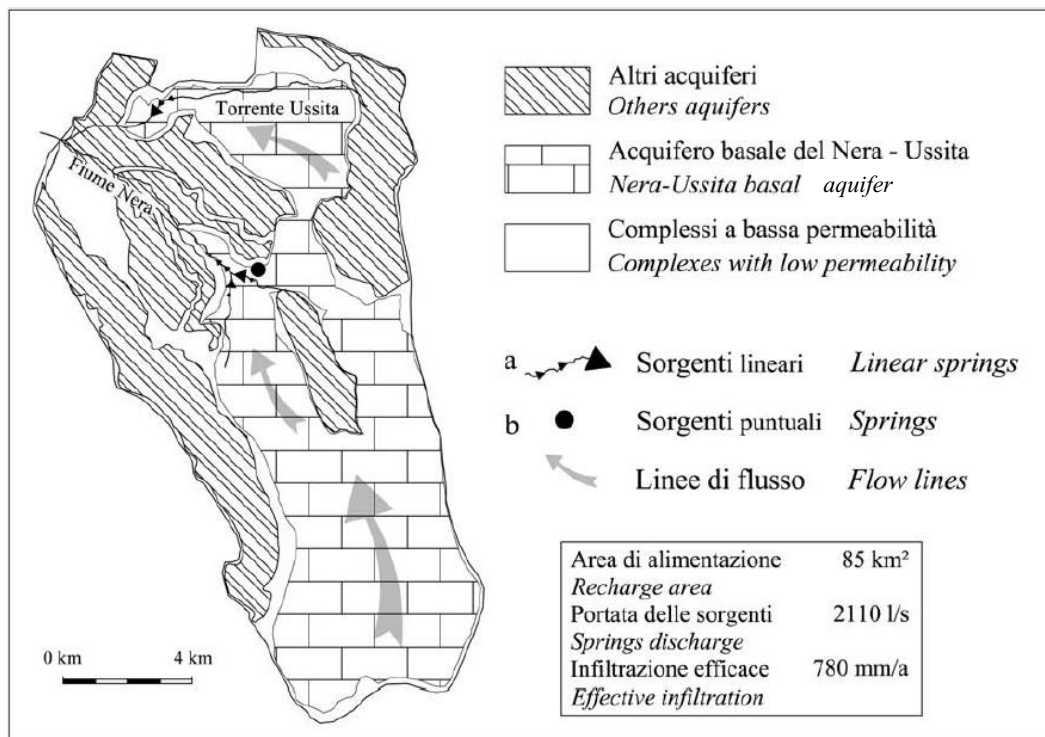


Figure 1.16 Hydrogeological conceptual model for the Upper Nera River basin located in hydrostructure 2, modified from Tarragoni, 2006.

At the same time, there is no evidence of flows in the E-W direction belonging to the **BAS** aquifer (Figure 1.16), (i.e., from hydrostructure 2 to hydrostructure 3). However, the work named “Integrazione della base conoscitiva per la gestione della risorsa idrica sotterranea del Parco Nazionale dei Monti Sibillini” (Petitta, 2011), aimed at studying the hydrogeological features of the Norcia Plain located at about 600 m a.s.l. in hydrostructure 3, shows a clear groundwater contribution coming from the eastern sectors (evidenced by the piezometric map of Figure 1.17). Since the Norcia Plain is located near (immediately to the west) of the Notoria-Preci fault system, a possible groundwater contribution coming from the carbonate geological Fms. of hydrostructure 2, towards hydrostructure 3, may be present.

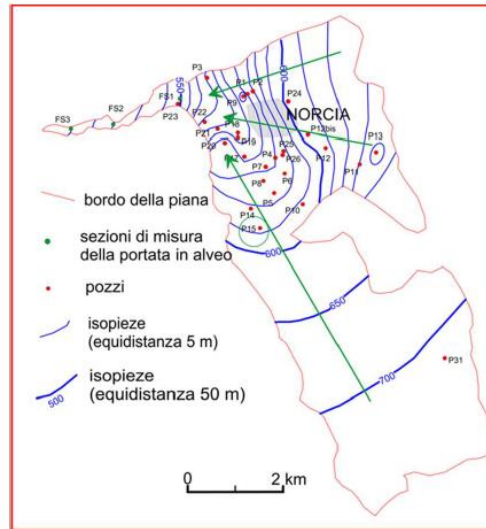


Figure 1.17 Piezometric map of Norcia Plain located in hydrostructure 3, from Petitta, 2011.

Few years later, Nanni & Tazioli deepen the study of Pescara del Tronto and Capodacqua springs (southern-east sector - hydrostructure 2), managed by the groundwater management company (CIIP s.p.a.), identifying the **MAI** and **BAS** aquifers as the principal aquifers feeding these springs. Isotopic investigations have been in progress in the Sibillini Mts. since 2002 (Nanni et al., 2013) while artificial tracer tests have been in progress since 2016. They are conducted by the Università Politecnica delle Marche highlighting the presence of a single hydrostructure, locally compartmentalized, but connected to the sinkhole of the Mèrgani sinking stream located in Pian Grande Plain (Nanni et al., 2020) (Figure 1.18).

Nanni et al., 2020 remarked their quantitative outcomes as preliminary, suggesting more specific investigations to be conducted on single springs or by using integrative methods to be coupled with artificial tracer tests.

The hydrogeological group of the Università Politecnica delle Marche developed its research activities during the years focusing the attention to the hydro-geostructural characterisation of the area performing rock-mass characterisations, isotope hydrology investigations, hydrochemical samplings and artificial tracer tests. The aim was to investigate the aquifers connection, the mixing of groundwater coming from different aquifers and the evaluation of springs recharge areas.

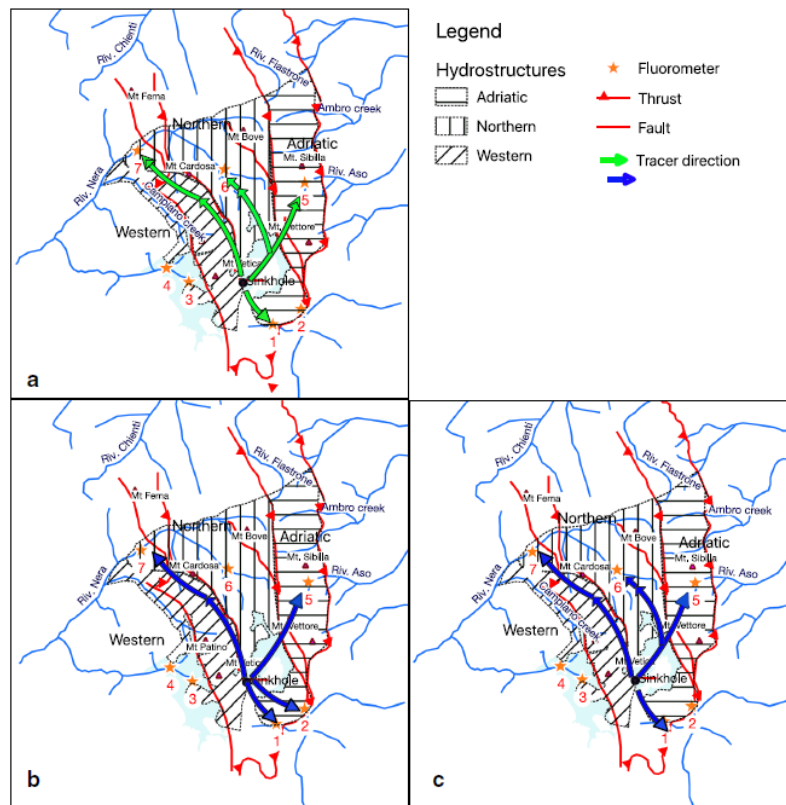


Figure 1.18 Main schematic traced flow paths, with hydrogeological and structural sectors, derived from the first tracer test (a), the second tracer test (b) and third artificial tracer tests (c). From Nanni et al., 2020.

In this context Nanni et al., 2013 highlighted some problems in the application of environmental isotopes to the hydrogeology of high mountain carbonate aquifers attributing them to the effect of orography and climate on isotopic contents, and to the hydrogeological complexity. Moreover, the occurrence of snowpack in the Sibillini Mt. area persisting until late spring, lead the aquifers to be recharged mainly by snowmelt causing possible variations in the isotopic contents of springs.

The peculiarity of this area, although quite limited in extension (1200 km²), is confirmed by the numerous geological, geo-structural and hydrogeological studies carried out over the years.

After the seismic sequence, occurred between 24 August 2016 and 18 January 2017 (nine Mw 5.0–6.5 seismic events), many studies aiming to identify the modification of the hydrogeological setting of Sibillini Mts., were performed. First insights published by Petitta et al., 2018 and Fronzi et al., 2020 showed a general increase in river and springs discharge on the western and eastern slope of the Sibillini ridge, particularly in the systems located in the northern portion, soon after the first seismic shock. After the second and third shocks, the discharge remained high on the western side and suddenly dropped on the eastern side, causing the disappearance of some springs locate at high elevation (Fronzi et al., 2020). The surveys conducted by the various groups operating in the area

were based exclusively on discharge data. In this work, the discharge analysis and the hydrogeological balance are integrated with isotopic, artificial tracer tests and hydrochemical groundwater data to better define the post-seismic groundwater model and deepen the knowledge of this complex carbonate systems.

2 Material and methods

2.1 Site characterization

2.1.1 Precipitation and meteo-climatic monitoring system

Rainfall, snow cover depth and temperature data provided by the authorities of the area involved in this study (Marche, Lazio, Umbria Regions) were freely downloaded from the official hydro-pluviometric databases managed by the regional monitoring centres. As concerns Marche Region, the data have been downloaded from the SIRMIP-online portal (<https://www.regione.marche.it/Regione-Utile/Protezione-Civile/Console-Servizi-Protezione-Civile/SIRMIP-online>) managed by the Regione Marche Civil Protection Agency. The data for the meteo-climatic stations located in Umbria Region have been downloaded from the SIR-RU database (<https://annali.regione.umbria.it/>), while for Lazio Region the data are freely available at <https://protezionecivile.regione.lazio.it/gestione-emergenze/centro-funzionale/rete-idrometreologica>. The rainfall, snow cover depth and temperature data are recorded every 15 minutes for Marche Region by means of a remote-controlled rain/temperature network. Umbria and Lazio Regions provide daily rainfall and thermometric data.

In order to characterize the area from a meteo-climatic point of view and to obtain the precipitation and thermometric regime to perform the hydrological balance, only few stations in the area have been selected. The choice of the meteo-climatic stations network was guided by finding a homogeneous spatial distribution both from a plano-altimetric point of view and by the availability of the most continuous dataset (stations with less than 10% of missing data during the monitoring period). The screening highlighted 15 rainfall stations and 12 temperature stations with the aforementioned characteristics (Table 2.1).

Table 2.1 Meteo-climatic monitoring network. Blue colour refers to the Marche Region stations; yellow colour Umbria Region; green colour Lazio Region.

Name	Elevation (m a.s.l.)	Type
Ussita	749	R, T
Montemonaco	995	R, T
Arquata del Tronto	720	R
Capodacqua	842	R, T
Mt. Prata	1813	R, T, SC
Mt. Bove sud	1917	R, T, SC
Piè del Sasso	711	R
Sellano	608	R, T
Vallo di Nera	310	R, T
Cascia	604	R, T
Norcia	700	R, T
Monteleone di Spoleto	935	R, T
Amatrice	954	R
Cittareale	810	R, T
Mt. Terminillo	1750	R, T

R= rainfall station; T= thermometric station; SC= snow depth thickness.

The daily data have been grouped into monthly and annual data calculating the mean monthly or annual rainfall and the mean monthly or annual temperature for the period 2009-2020. During the same period the mean annual snow cover depth was calculated for Mt. Prata and Mt. Bove sud stations. All data have been used to calculate the indirect hydrological balance to be compared to the direct hydrologic balance (starting from discharge measurements) for the analysed hydrostructures (see paragraph 2.1.4).

2.1.2 Standard Precipitation Index

The Standard Precipitation Index (SPI) (McKee et al., 1993) has been used to detect how the rainfall regime is related to springs discharge variations and to quantify the amount of particularly dry or wet conditions occurred in the area. In this context, few studies explored the correlation between the spring discharge and the SPI (Fiorillo & Guadagno, 2010; Fiorillo & Guadagno, 2012; Viaroli et al., 2018), while the approach of Valigi et al. 2020 and Fronzi et al., 2020 coupled the SPI with the

springs discharge to discriminate rainfall trends with respect to the discharge variation after a seismic sequence.

The SPI is a strong and dynamic index to analyze the occurrence and the magnitude of drought periods (World Meteorological Organization, 2012).

In the SPI computation, cumulated rainfalls over different time scale (1, 2, 3....n months) are fitted to a gamma probability distribution and then transformed into a normal distribution. For a given data time series of precipitation x_i as $x_1; x_2, \dots, x_n$, the SPI is defined as follows:

$$SPI = \frac{x_i - \bar{x}}{\sigma_x}$$

where: $\bar{x}$ is the arithmetic mean of rainfall and σ_x is the standard deviation calculated from the whole monthly time series.

SPI values > 0 indicate a greater than mean precipitation (wet periods), while SPI values < 0 indicate less than mean precipitation (dry periods) (McKee et al., 1993). A drought event occurs any time the SPI is continuously negative and reaches an intensity of at least -1 . The event ends when the SPI becomes positive (Table 2.2).

Table 2.2 Classification of drought conditions based on the SPI.

Condition	Range
Extremely wet	$SPI \geq 2.0$
Very wet	$1.5 \leq SPI < 2.0$
Moderately wet	$1.0 \leq SPI < 1.5$
Near normal	$-1.0 < SPI < 1.0$
Moderately dry	$-1.5 < SPI \leq -1.0$
Severely dry	$-2.0 < SPI \leq -1.5$
Extremely dry	$SPI \leq -2.0$

“n” varying from 1 to 24 months is the best statistical range of application (Guttman, 1998, Guttman, 1999). Therefore the rainfall data were elaborated using a 12 month time scale (SPI-12) since it better represents long-term precipitation patterns and is better tied to groundwater resource dynamics (Fiorillo & Guadagno, 2010; World Meteorological Organization, 2012).

The SPI elaboration was performed for five rainfall stations, which guarantee rainfall data without significant interruptions for at least 30 years in the period Jan-1991 and Sept-2021 (Table 2.3). The stations have been selected to cover the whole study area taking into account all the different sectors

of the Sibillini Mts. (Figure 2.1). In particular, Piè del Sasso rainfall station has been used to characterise the SPI for the Northern area, Montemonaco rainfall station has been used to characterise the climatic conditions of the Eastern side, Amatrice rainfall station has been used to characterise the SPI for the southern side, Vallo di Nera rainfall station has been used to characterise the climatic conditions of the western side of the area, while the Arquata del Tronto rainfall station has been used to characterise the climatic conditions in a barycentric position with respect to the analysed domain. Elevations range from 310 to 995 m s.l.m.

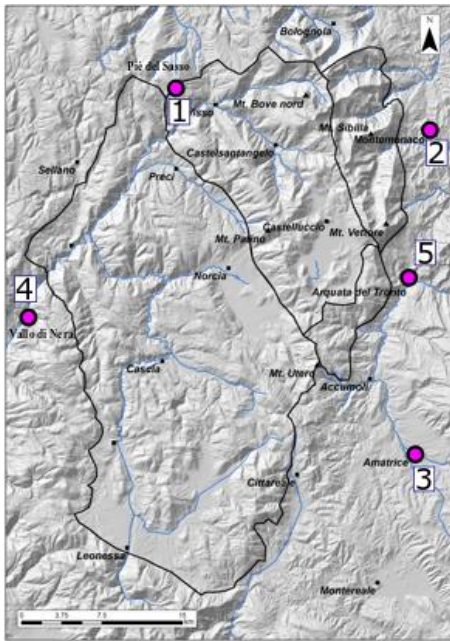


Table 2.3 Characteristics of rainfall stations used to calculate Standard Precipitation Index (SPI).

N° in map	Name	Elevation (m a.s.l.)
1	Piè del Sasso	711
2	Montemonaco	995
3	Amatrice	954
4	Vallo di Nera	310
5	Arquata del Tronto	720

Figure 2.1 Location of rainfall stations used to calculate Standard Precipitation Index (SPI).

2.1.3 Hydrological monitoring system

The discharge measurements were performed in 18 monitoring points using different methods. In this context it must be evidenced that the discrete discharge measurements have been performed cooperating with different research groups, regional authorities and Universities involved in the evaluation of post-seismic groundwater modification of the area, moreover the data have been integrated by the measurements performed during the pre-seismic period and published in (Valigi et al. 2020).

The discharge of punctual springs located in the eastern and south-eastern side of the Sibillini Massif (points 1, 2, 3, 4, 5 Figure 2.2) were provided by the local groundwater management companies. In particular, Capotenna spring (point 1- 1178 m a.s.l.) is managed by Tennacola s.p.a. Servizio idrico Integrato, which provides weekly discharge data recorded in the groundwater drainage tunnel. Foce

spring (944 m a.s.l.), Sassopaccato (1300 m a.s.l.), Pescara del Tronto spring (920 m a.s.l.) and Capodacqua spring (841 m a.s.l.), points 2,3,4,5 respectively, are all managed by the drinking water company CIIP s.p.a. For Foce, Capodacqua and Pescara del Tronto springs the daily discharge data are available and provided by the water company. The discharge has been measured by means of water depth sensor located in a specific weir with rectangular section (in the groundwater drainage tunnel), which have an accuracy of ± 5 % for Pescara del Tronto, ± 1.9 % for Capodacqua and ± 1.5 % for Foce, while Sassospaccato spring discharge is measured monthly. As concerns the northern portion of the study area, in which the Nera River is located, the discharge measurements are available at different scale. Point 6, 7 and 8 are continuously monitored and discharge data are validated by the Regione Marche Civil Protection Agency; the data have been downloaded from the SIRMIP-online portal (<https://www.regione.marche.it/Regione-Utile/Protezione-Civile/Console-Servizi-Protezione-Civile/SIRMIP-online>). However, it must be noticed that during the co-seismic period the rating curve for point 6 was not calibrated, hence nine discharge measurements were conducted and integrated. Point 7 is continuously monitored starting only from October 2017; in the previous period nine discrete discharge measurements were available. In points 9, 10 and 11 discrete discharge monitoring have been performed between 2018 and June 2019 then, starting from August 2019, the point 9 has been equipped with hydrometric pressure transducer (CTD diver Eijkelkamp, accuracy ± 0.5 cmH₂O and resolution 0.2 cmH₂O) compensated by atmospheric pressure, using an acquisition time interval of 10 minutes then grouped at daily scale (mean daily value). The point 10 and 11 have been equipped by continuous hydrometric pressure transducers compensated by atmospheric pressure (OTT mod. Ecolog 800, produced by OTT Corr-tek srl Verona, Italy) since August 2019 and July 2020 respectively. The rating curve for point 9,10 and 11 have been calibrated by performing five discrete discharge measurements for each hydrometric section in different hydrogeological periods during the year 2021.

The western side of the domain is mainly characterised by discrete discharge monitoring.

As the Campiano river concerns, in which point 12 and point 13 are located, 36 discrete discharge measurements have been performed after the seismic sequence and only point 13 have been equipped by continuous daily discharge monitoring (data provided by the fish farming company) since 2018. The continuous discharge data have been compared with 16 discrete measurements in the same point.

The discharge monitoring points located in the Sordo and Corno River basins (point 14, 15, 16, 17, 18, 19, 20) are mainly discrete monitored since the October, 30th 2016 Mw 6.5 earthquake (main shock). In fact, from this event onwards, the appearance of a new spring, namely Torbidone spring, and at the same time its stream, made it possible to carry out 60 discharge measurements a little downstream from the punctual spring (point 14). Forty-seven discrete discharge measurements were performed at the point 15, in which starting from May 2018 a continuous discharge monitoring is operated by continuous hydrometric pressure transducers compensated by atmospheric pressure (OTT mod. Ecolog 800). Also in this case it has deemed necessary to calibrate the rating curve to convert water level into continuous discharge data. A continuous discharge monitoring system is available for point 16 (downstream the confluence between Torbidone stream and point 15). In point 17 and 18, 27 and 23 discrete discharge measurements have been performed respectively after the seismic sequence. Point 18 is characterised by a continuous discharge monitoring system, however starting from February 2019 some technical issues made the system out of operability and only discrete discharge are available.

Points 19 and 20 are characterised by a discrete discharge monitoring starting from February and May 2018 respectively until July 2020 performed by Umbria Region.

Point 21 (Mèrgani sinkhole), located about 1300 m a.s.l., have been instrumented with hydrometric pressure transducer (CTD diver Eijkelpamp, accuracy ± 0.5 cmH₂O and resolution 0.2 cmH₂O) compensated by atmospheric pressure, using an acquisition time interval of 10 minutes then grouped at daily scale (mean daily value). In this point the rating curve has been calibrated by performing five discharge measurements.

The discrete discharge measurements have been performed by using OTT C2 (measure interval 0.025 - 5 m/s; accuracy $\pm 2\%$), OTT MF pro (measure interval: 0 - 6 m/s; accuracy: $\pm 2\%$ between 0 and 3 m/s; accuracy: $\pm 4\%$ between and 5 m/s) current meters (OTT Corr-tek srl Verona, Italy) and FlowTracker Acoustic Doppler Velocimeter (velocity range: ± 0.001 to 4.0 m/s; velocity resolution: 0.0001 m/s; velocity accuracy: $\pm 1\%$ of measured velocity), produced by SonTek / Xylem Inc. (San Diego, California, USA). In case of high flow rate (high velocity and/or water depth) the discharge measurements have been performed by artificial tracer tests by sudden injection method of Fluorescein. The sudden release of dye tracer was conducted up stream of the monitoring point to ensure complete homogenisation of the tracer (for the study area about 200-300 m), and the tracer concentration was continuously recorded by the field fluorometric probe (measure interval 5 s) after

calibration with standard (tracer concentrations: 0 and 100 ppb) (Tazioli, 2011). Specification about the field fluorometric probes used in this study are reported in Chapter 2.2, paragraph 2.2.2).

All the discrete discharge measurements in streams and rivers were carried out away from heavy rainfall periods in order to measure only the base flow due to groundwater component.

All the discharge data have been pre-processed, filtered and, if necessary, only sparse daily missing data have been filled by a moving median method following (Somasundaram & Nedunchezian, 2011).

Discharge data have been used to calculate a direct hydrologic balance to be compared to the indirect hydrological balance (starting from temperature and rainfall) for the hydrostructures (see paragraph 2.1.4). Moreover, continuous discharge data have been used to calculate depletion coefficient (see paragraph 2.1.5), to evaluate the discharge modification due to seismic sequence and quantify the amount of groundwater lost by the seismic induced hydrogeological modifications. The daily discharge data provided by CIIP s.p.a. have been used to investigate the hydrodynamic behaviour of the aquifers by applying a data-driven approach coupled to the results of artificial tracer tests (see paragraph 2.2.4). Where possible, basic statistics have been calculated within three parts of the whole period of time: Pre-seismic (i.e., before the seismic events), Co-seismic (i.e., during the seismic events, when the main shocks took place), and Post-seismic (i.e., after the seismic events) to discern significant springs discharge changes caused by the seismic sequence.

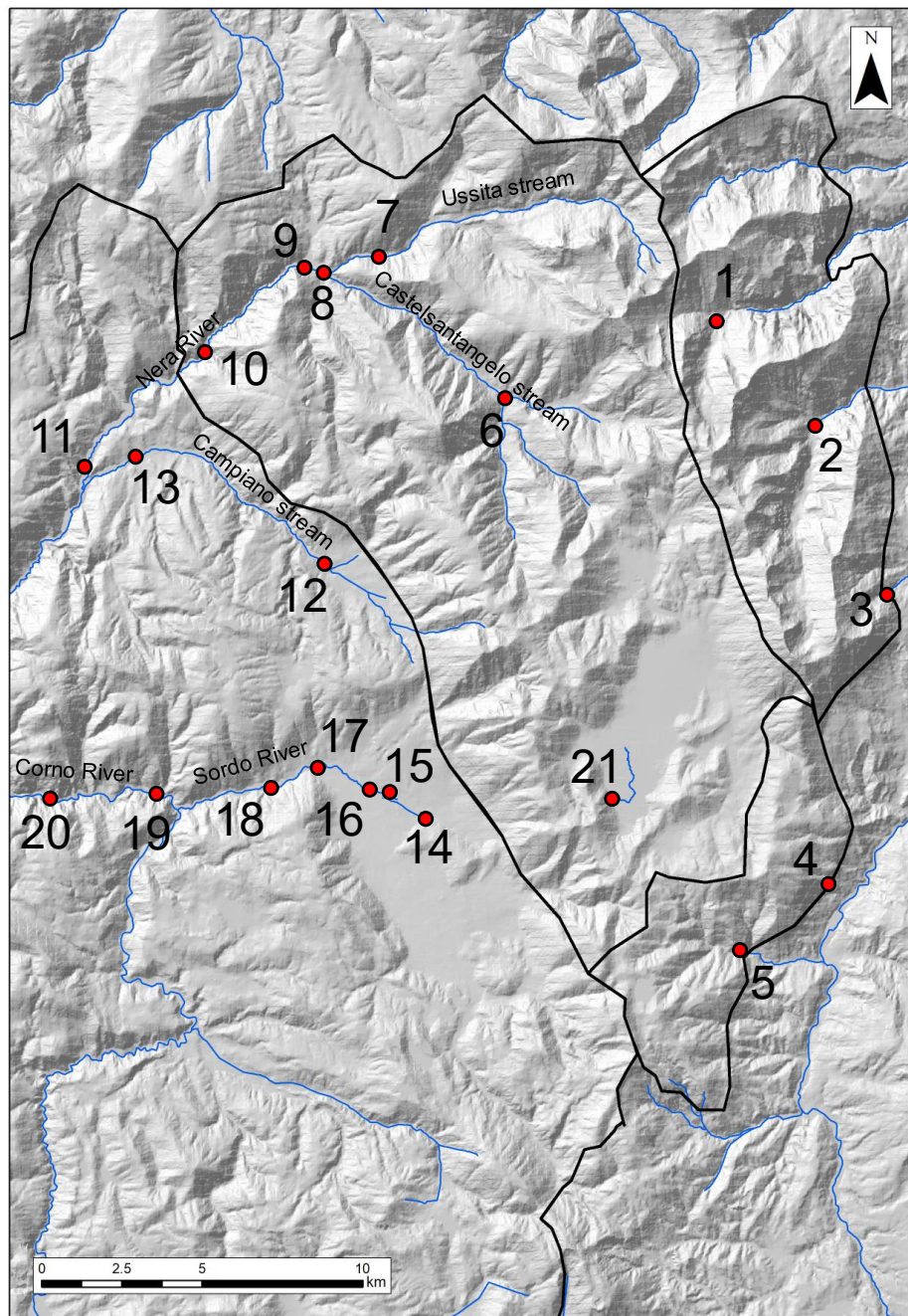


Figure 2.2 Location of discharge monitoring points.

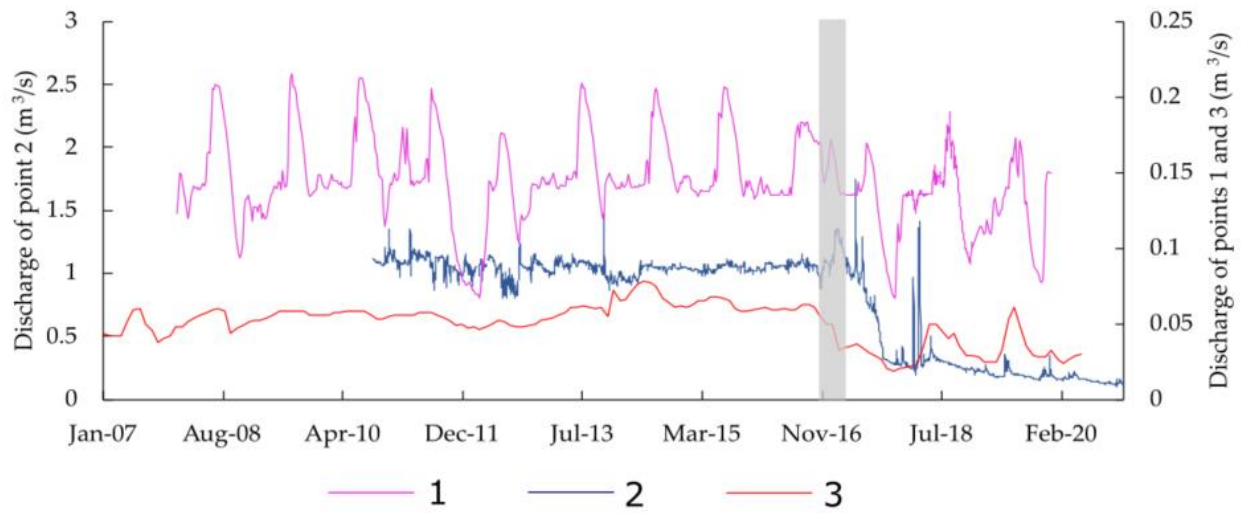


Figure 2.3 Discharge data for points 1, 2 and 3 located in the Hydrostructure 1. Principal vertical axis refers to point 2 while secondary vertical axis refers to point 1 and 3. The grey window indicates the co-seismic period (24th August 2016-18 January 2017).

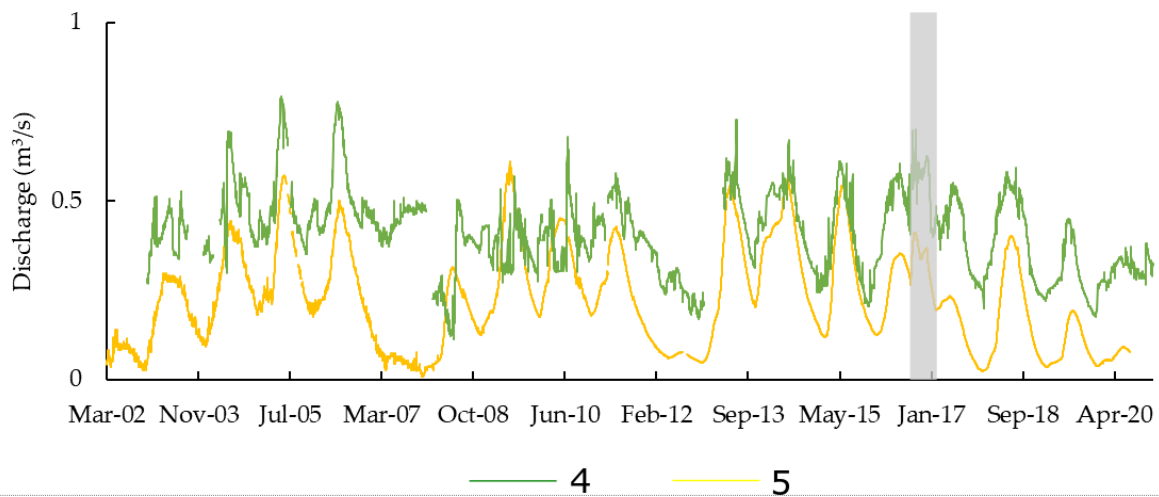


Figure 2.4 Discharge data for points 4 and 5 located in the Hydrostructure 2a. The grey window indicates the co-seismic period (24th August 2016-18 January 2017).

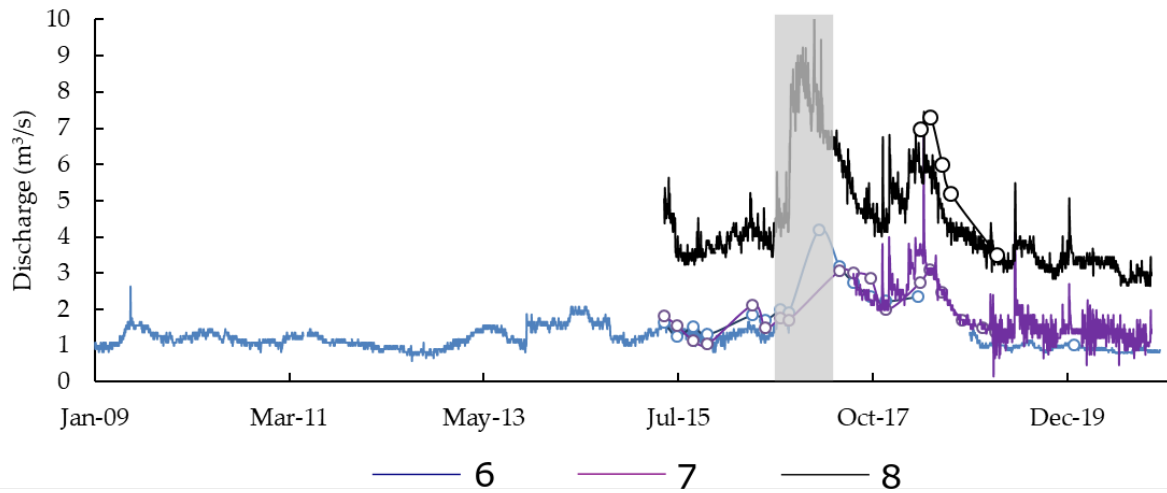


Figure 2.5 Discharge data for points 6, 7 and 8 located in the Hydrostructure 2 and belonging to the Upper Nera River Basin up stream to Visso village. The solid lines referred to the continuous data, while the dots refer to discrete discharge measurements. The grey window indicates the co-seismic period (24th August 2016-18 January 2017).

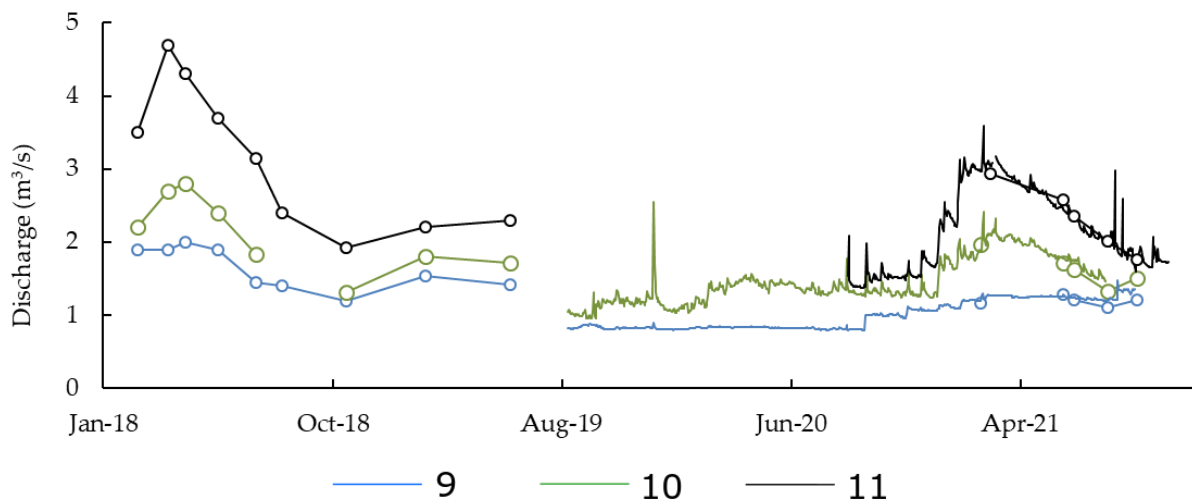


Figure 2.6 Discharge data for points 9, 10 and 11. The points 9 and 10 are located in the Hydrostructure 2 while the point 11 is located in the Hydrostructure 3. All the points belong to the Nera River Basin downstream to Visso village. The solid lines referred to the continuous data, while the dots refer to discrete discharge measurements.

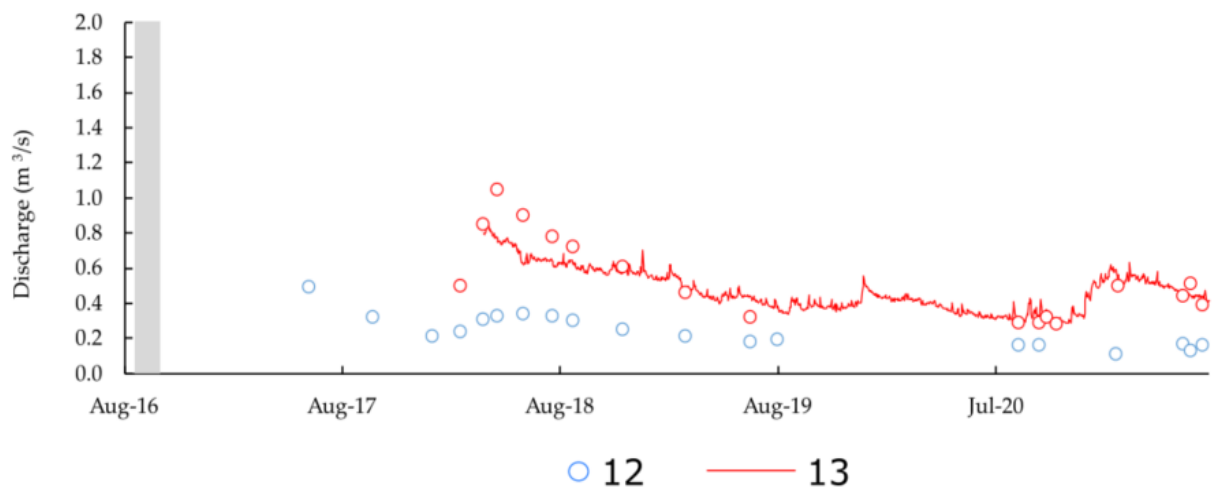


Figure 2.7 Discharge data for points 12 and 13. The points are located in the Hydrostructure 3 and belong to the Campiano stream. The solid line refers to the continuous data, while the dots refer to discrete discharge measurements. The grey window evidences the co-seismic period (24th August 2016-18 January 2017).

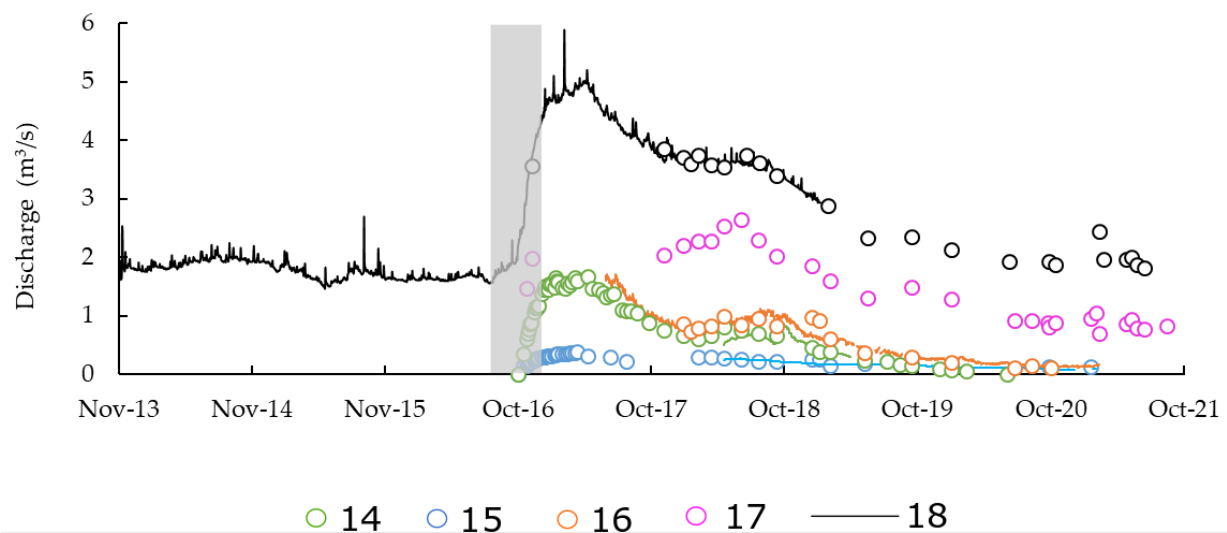


Figure 2.8 Discharge data for points 14, 15, 16, 17 and 18. The points are located in the Hydrostructure 3 and belong to the Sordo River Basin. The solid lines refer to the continuous data, while the dots refer to discrete discharge measurements. The grey window evidences the co-seismic period (24th August 2016-18 January 2017).

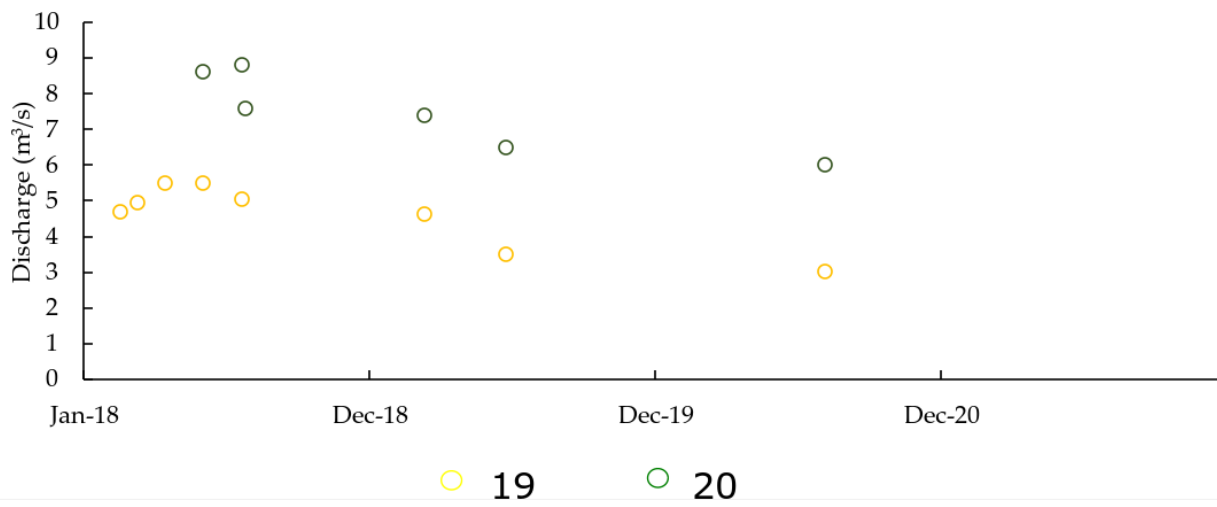


Figure 2.9 Discrete discharge data for points 19 and 20. The points are located in the Hydrostructure 3 and belong to the Corno River Basin.

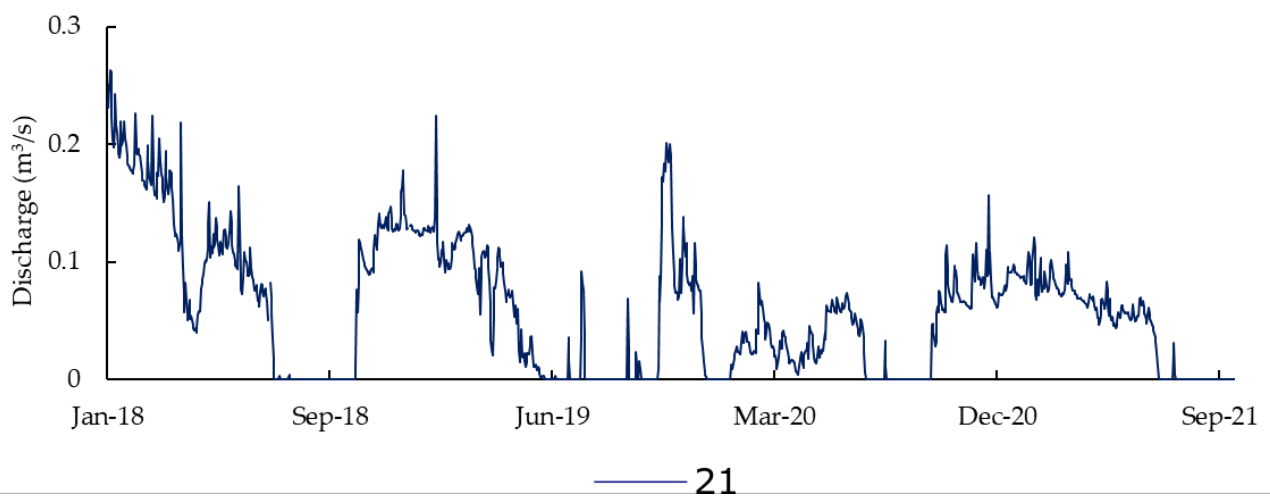


Figure 2.10 Continuous discharge data for point 21. The point is in the Hydrostructure 2 in the Pian Grande Plain (Mèrgani sinkhole).

Table 2.4 Details of discharge monitoring network.

N° in map	Name	Hydrostructure	Start-end monitoring period	Monitoring type	Record time frequency	Rating curve (R ²)
1	Capotenna spring	1	Mar 2008- Dec 2019	Continuous	Weekly	-
2	Foce spring	1	Oct 2010- Dec 2020	Continuous	Daily	-
3	Sassospaccato spring	1	Feb 2007 – Jun 2020	Continuous	Monthly	-
4	Pescara del Tronto spring	2 a	Mar 2002- Jul 2020	Continuous	Daily	-
5	Capodacqua spring	2 a	Mar 2020 – Dec 2020	Continuous	Daily	-
6	Nera Castelsantangelo	2	Jan 2009 – Dec 2020	Continuous/Discrete	Daily	-
7	Ussita Madonna dell'Uccelletto	2	May 2015 – Nov 2020	Continuous/Discrete	Daily	-
8	Nera Visso	2	Jun 2015 – Nov 2020	Continuous/Discrete	Daily	-
9	Nera Visso (ERG)	2	Feb 2018 – Sep 2021	Continuous/Discrete	Daily	0.89
10	Nera I Molini	2	Feb 2018 – Sep 2021	Continuous/Discrete	Daily	0.99
11	Nera Pontechiusita	3	Feb 2018 – Oct 2021	Continuous/Discrete	Daily	0.97
12	Campiano Campi	3	Jun 2017 – Aug 2021	Discrete	-	-
13	Campiano Corone	3	Feb 2018 – Jul 2021	Continuous/Discrete	Daily	0.89
14	Torbidone spring	3	Oct 2016 – Sep 2020	Discrete		-
15	San Martino-Pratarella springs	3	Oct 2016 – Feb- 2021	Continuous/Discrete	Daily	0.94
16	Torbidone + San Martino-Pratarella	3	Jun 2017 – Mar 2021	Continuous/Discrete	Daily	0.97
17	Sordo Trotificio	3	Nov 2016 –Sep 2021	Discrete	-	-
18	Sordo Pontemollo	3	Nov 2013 –Jul 2021	Continuous/Discrete	Daily	0.98
19	Corno a Serrvalle	3	Feb 2018 – Jul 2020	Discrete	-	-
20	Corno a Biselli	3	May 2018– Jul 2020	Discrete	-	-
21	Mèrgani sinkhole	2	Jan 2018 – Sep 2021	Continuous	Daily	0.98

2.1.4 Hydrological balance

The hydrological balance was performed with the aims to: i) observe any discrepancy between the meteoric aquifers recharge amount before and after the seismic sequence, ii) calculate the groundwater surplus lost by the seismic-induced effects, iii) determine the main springs recharge area. The hydrological balance was made by applying both the indirect and the direct methods. The indirect hydrological balance is calculated starting from the rainfall and temperature data while the direct method is calculated starting from the discharge measurements; the methods had compared each other to calibrate the potential infiltration coefficient for each hydrogeological complex.

2.1.4.1 *Indirect methods (Turc method vs WaterbalANce)*

Starting from the annual mean rainfall and temperature data obtained from the monitoring network, the elevation-mean annual rainfall and elevation-mean annual temperature relationships were derived (period 2009-2020 and sub-period 2015-2020). Due to the variability of rainfall regime caused by the extremely heterogeneous topographical setting, three different elevation-mean annual rainfall relationships have been obtained (shown in the results section, paragraph 3.1). The Tyrrhenian side was subdivided into Northern and Southern domain by calculating two different relationships, while a single relationship was calculated for the Adriatic side. The snow depth and rainfall data of Monte Bove sud and Monte Prata stations, located almost on the ridge between the Tyrrhenian and the Adriatic side have been taken into consideration both for the Adriatic and the Northern Tyrrhenian sides. The rainfall data of Cascia and Monteleone di Spoleto stations have been taken into consideration to calculate the relationship (elevation-mean annual rainfall) for both the Tyrrhenian domains (Northern and Southern). The mean annual rainfall value for the stations located above 1500 m a.s.l. (Mt. Prata, Mt. Bove sud and Mt. Terminillo), characterised by the persistent snow cover from late autumn to spring, were incremented by the mean annual snow water equivalent (SWE) calculated starting from the procedure described in paragraph 2.3.2, during the snow melt isotopic characterisation of the study area. Hence, the mean maximum annual snow depth recorded in Mt. Prata and Mt. Bove sud stations were converted in SWE (1 cm snow depth $\approx$ 4 mm rainfall) and added to the mean annual rainfall for all the stations located over 1500 m a.s.l.

The atmospheric temperature data, collected from the monitoring network, permitted to calculate a single elevation-mean annual temperature relationship valid for the whole study area.

The spatialization of mean annual temperature and mean annual rainfall above the area has been performed by applying the linear regression equations obtained between elevation- mean annual

rainfall (3 relationships) and between elevation-mean annual temperature to the digital terrain model DTM (20x20 m cell size resolution) achieving two raster files (20x20 m cell size resolution) covering the whole study area.

Starting from the temperature and rainfall raster files, the well-known Turc method (Turc, 1954) to calculate mean annual actual evapotranspiration value (AET) was applied using the Rastercalculator tool in ArcGIS® following the equation:

$$AET = \frac{R}{\sqrt{0.9 + \frac{R^2}{L^2}}}$$

where:

$$L = 300 + 25T + 0.05T^3$$

in which R is the mean annual rainfall value (mm) and T is the mean annual temperature value (°C).

Then, the mean annual water surplus (mm) raster file (WS) was calculated as follows:

$$WS = R - AET$$

A refined approach with respect to the Turc method, based on mean annual (rainfall and temperature) data, were performed by applying the Thornthwaite - Mather method at monthly scale. This method remains one of the most used, especially for hydrogeological purposes (Mammoliti et al., 2021). In fact, in contrast to physically based models, which require many input parameters, the Thornthwaite-Mather method is a simple empirical data-driven procedure in which the associated error to its use is smaller than the one associated with the measure of input data (Di Matteo et al., 2017). Moreover, many authors demonstrated that the method performs well in humid regions where the precipitation and air temperature are the only input data, while it underestimates monthly potential evapotranspiration (PET) under dry and arid climates, because the equation does not consider the saturation vapor deficit of the air (Pelton et al., 1960; Pruitt 1964; Hashemi & Habibian, 1979; Malek, 1987).

The computation was carried out by the WebApp WaterbalANce (Mammoliti et al., 2021) freely available at <https://thornwaterbalance.com/> and appropriately developed within the framework of this research to faster and automate the calculation of the hydrological balance.

WaterbalANce code was written in Python and is driven by a serverless computing approach. It requires the following input parameters: the latitude of the rainfall-thermometric station (LAT , decimal degrees), both the mean monthly temperature (Tm , °C) and both the monthly total precipitation (P , mm) recorded at the station. By adding the soil moisture storage capacity (SM , mm), the rainfall snowfall temperature threshold (SRT , °C) and runoff factor ($beta$, %), it is possible to calculate the hydrological balance output such as water surplus (S , mm). Starting from temperature and latitude, the potential evapotranspiration (PET , mm) is calculated as follows:

$$PET = 16 \times k \times \left(\frac{10 \times Tm}{I} \right)^a$$

where:

$$I = \sum_{n=1}^{12} i \text{ where } i = \left(\frac{Tm}{5} \right)^{1.514}$$

and

$$a = (675 \times 10^{-9} \times I^3) - (771 \times 10^{-7} \times I^2) + (1792 \times 10^{-5} \times I) + 0.49239$$

k is the latitude dependent correction factor accounting the number of days in the month and the actual number of hours of insolation; I is the annual heat index; a is a coefficient strictly proportional to I . For the computation of the actual evapotranspiration (AET , mm) the soil moisture storage capacity must be selected in reason of the soil type and its relation between the presence of plants and roots (Donker, 1987; Thornthwaite & Mather, 1955; Thornthwaite & Mather, 1957). After that, according to the difference between monthly precipitation and PET (mm) the soil moisture content changes to satisfy the water request from the soil system. The water surplus (S) is generated only after the soil moisture conditions are satisfied, and if the temperature is above a rainfall-snowfall threshold (generally temperature > -1 °C). The architecture of the WaterbalANce web app is shown in figure 2.11, while the Python code is reported in Annex A. The applicability of the Thornthwaite-Mather method was tested over the Hargreaves & Samani, 1985 method, by comparing the PET to the reference evapotranspiration (Eto) grouped at monthly scale. However, this comparison was

made possible only for one meteo-climatic station (Montemonaco), having no daily rainfall and temperature missing data for the analysed period.

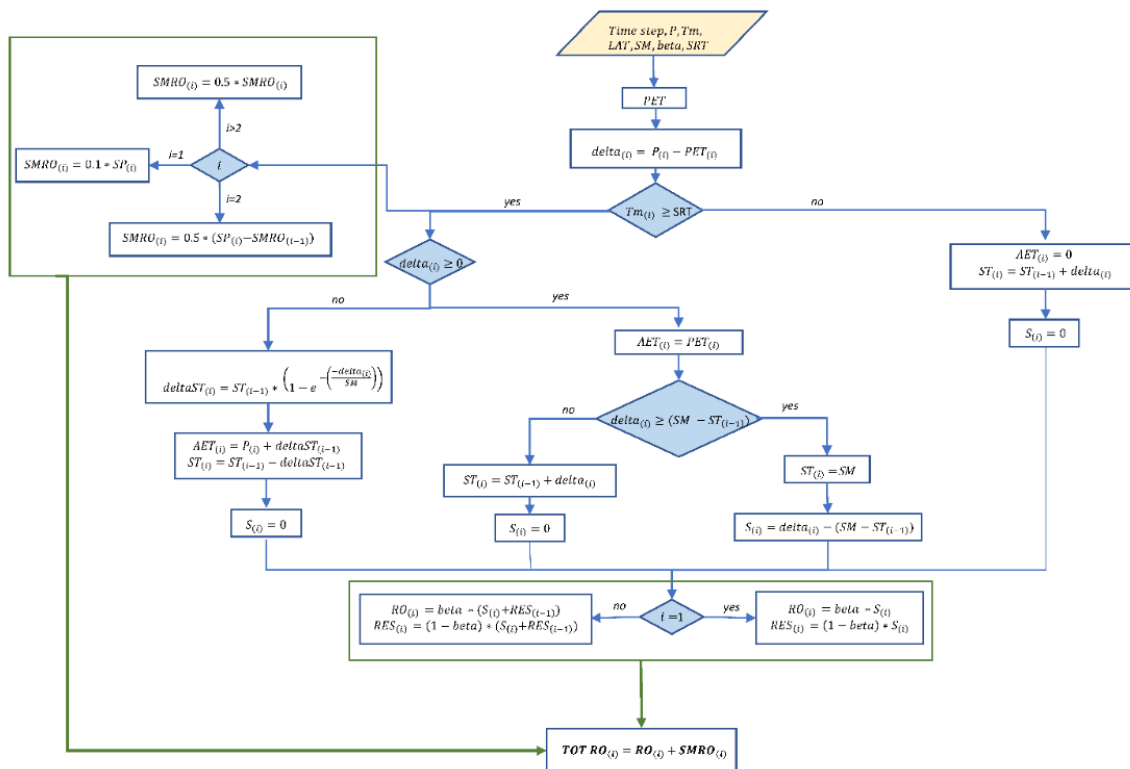


Figure 2.11 Architecture of WaterbalANce WebApp to perform the hydrological balance computation.

In this case the maximum annual snow depth recorded in Mt. Prata and Mt. Bove sud stations were converted in SWE (1 cm snow depth $\approx$ 4 mm rainfall) and added to the mean monthly rainfall for the months in which the snowmelt occurs in all the stations located over 1500 m a.s.l., while the soil moisture storage capacity (SM) was assumed equal to 200 mm following previous studies performed in similar areas of central Italy (Di Matteo et al., 2017), hence specific soil maps and thus soils hydrologic characteristics are not available.

The computation of water surplus was carried out only for those stations in which both rainfall and temperature data were available at daily scale with missing data < 10% in the whole period (2009-2020), then grouped at monthly scale.

The monthly water surplus values, calculated for each termo-pluviometric station, were then grouped at annual scale (total annual water surplus) and for each station a mean annual water surplus was calculated.

The mean annual water surplus value for each station (located at different elevation each other's) was used to obtain the mean annual water surplus-elevation relationships. The strict dependence of the Thornthwaite-Mather method to temperature and rainfall data leads also in this case to calculate three relationships: one for the Adriatic side and two for Tyrrhenian side (northern and southern domains).

The spatialization of mean annual water surplus above the area have been performed by applying the equations obtained between elevation - mean annual water surplus (3 relationships) to the DEM (20x20 m cell size resolution) obtaining the water surplus raster file (20x20 m cell size resolution) covering the whole study area.

2.1.4.2 Calibration of Infiltration Rate through direct method

Starting from the pre-seismic discharge data of Mastrorillo et al., 2009 and Boni et al., 2010, integrated with the ones recently published by Viaroli et al., 2021, a direct hydrological balance method has been applied to calibrate the infiltration rate for each hydrogeological complex outcropping in the area. In fact, excluding external modifications, the mean annual discharge provided by a single spring can be considered as the aliquot of water surplus infiltrating in its recharge area during the same hydrologic year. The mean annual discharge data reported in Mastrorillo et al., 2009, Boni et al., 2010 and Viaroli et al. 2021 were considered as the equilibrium state for the hydrogeological system. For each hydrostructure the Effective Infiltration (EI) was estimated on the outcropping areas of each hydrogeological complexes and for the tectono-karstic plateau occupied by the Quaternary deposits (**DEP**), covering the carbonate geological Fms.

In this context it must be highlighted that some studies performed in the area affirm that the aquifers recharge component is always underestimates and the discharge data (output of the system) is higher than the input (Effective Infiltration) (Boni et al., 2010; Mastrorillo et al., 2021). The reason of this discrepancy has been attributed to the underestimation of precipitation falling in solid form (snow). The SWE for those areas has never been evaluated by field investigations and this amount has never been added into the hydrologic balance equation.

Although in this study the contribution of snow melt amount has been added into the water balance computation, a specific calibration of the infiltration coefficients for each hydrogeological complex became necessary.

Firstly, the mean annual volume of groundwater drained by each hydrostructure (calculated starting from the mean annual springs discharge) was divided by the whole extension of the hydrostructure obtaining a preliminary mean infiltration value expressed in millimetres/year (direct hydrogeological balance) (Allocca et al., 2014).

Then, following Celico, 1988 and Civita et al., 2000 the potential infiltration value (PIC) has been attributed to each hydrogeological complex: PIC for fractured limestone usually ranges between 70 % and 100 % while for marly limestones between 30 % and 70%.

The mean infiltration value calculated by the indirect method for each hydrogeological complex was then average weighted, based on each complex areal extension within each hydrostructure and compared to the mean infiltration value obtained by the direct method.

After several iteration aimed at converging the results of inflow (infiltration) and outflow (springs discharge data) through the indirect method, the PIC values for each hydrogeological complex were modified by taking into account the mean slope angle characterising each hydrogeological complex within each hydrostructure.

Starting from the intersection between each hydrogeological complex and the slope map determined in ArcGIS operating in a DTM 20 x20 cell size, each PIC was directly proportionally decreased starting from the maximum PIC for each complex hypothesised by Celico, 1988 and Civita et al., 2000, as the slope of the complex increases, obtaining a modified potential infiltration coefficient (PICs) as a function of the average hydrogeological complex slope (Annex B).

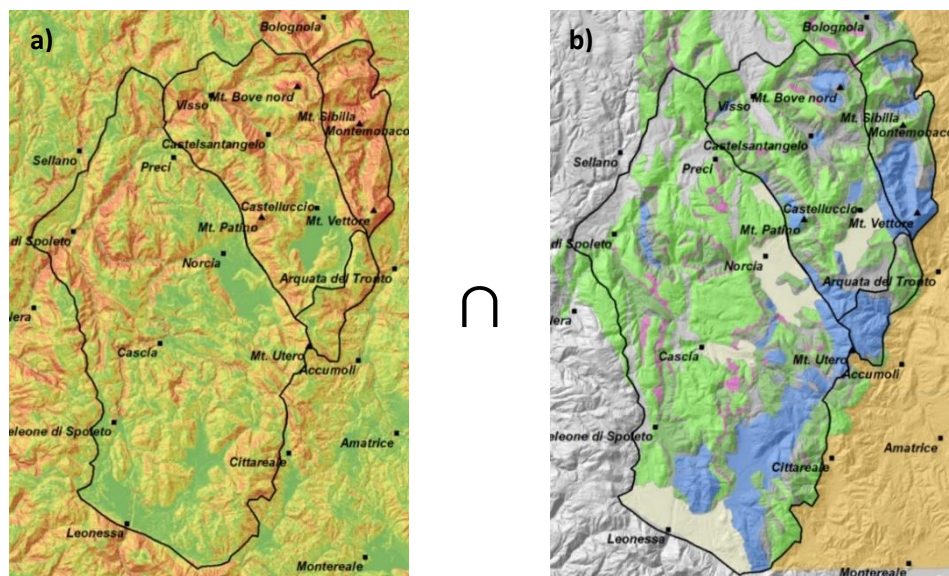


Figure 2.12 Intersection between the slope map (a) obtained by the DTM (20 m x20 m cell size), and the hydrogeological complexes (b).

Table 2.5 Mean slope (°) values determined for each hydrogeological complex within each hydrostructure.

Hydrogeological complex	Hydrostructure 1	Hydrostructure 2	Hydrostructure 2a	Hydrostructure 3
BAS	33.2	26.4	21.4	18.07
JUR	31.8	22.3	24.6	21.2
MAI	32.6	25.8	21.4	21.7
MF	31.6	25.2	23.5	19.8
SCA	31.6	23.3	24.4	19.9
SVC	n.p.	20.2	n.p.	17.1
DEP	n.p.	3.09	n.p.	7.3

n.p.= hydrogeological complex not-present within the hydrostructure.

The value of the mean annual water surplus (WS or S) calculated with both methods (Turc and Thornthwaite- Mather) were used to estimate the mean annual Effective Infiltration as follows:

$$EI = WS \times PICs$$

For the terrigenous complex, located eastward to the Sibillini Mts. overthrust, the PIC and therefore the PICs has not been attributed/calculated because it outcrops only outside the springs recharge areas.

Finally, the aquifer recharge component corresponding to the Effective Infiltration multiplied by the hydrostructures areal extension was directly compared to the pre-seismic discharge data drained by each hydrostructure.

Once the calibration of the modified Potential Infiltration Coefficient was performed, the hydrological balance was applied to the post-seismic data to quantify and discriminate the discharge surplus strictly related to the seismic effects and not attributable only to the recharge processes occurred in the area.

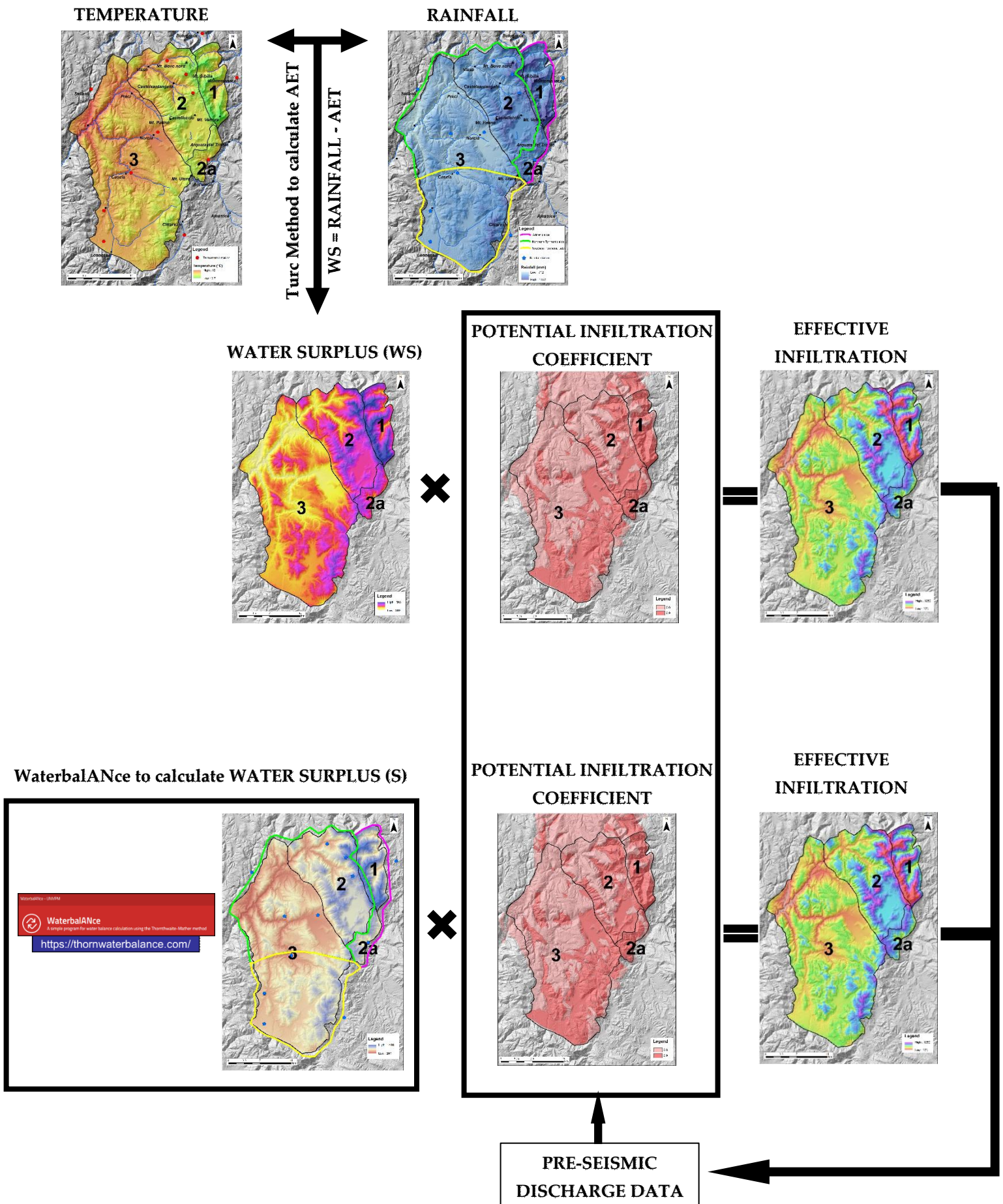


Figure 2.13 Flowchart of indirect (Turc vs WaterbalANce) and direct hydrological balance.

2.1.5 Recession curve analysis

Some studies performed in springs fed by carbonate aquifers of central Italy (Angelini & Dragoni, 1997; Giacometti et al., 2016; Di Matteo et al., 2021) demonstrate that the recession curves can be well approximated by an exponential trend following the Maillet equation (Maillet, 1905):

$$Q_t = Q_0 e^{-\alpha t}$$

in which Q_0 is the discharge at the beginning of the recession period (t_0), Q_t is the discharge at time t (calculated from t_0) and the parameter α is a recession coefficient.

Although Maillet, 1905 developed this equation from catchments characterised by granulated media aquifers, it has been widely used for karst media, i.e., non-homogeneous and anisotropic media with fissures (Bonacci, 1993). This first mathematical characterization of the baseflow recession is based on the drainage of a simple reservoir and presumes that the spring discharge is a function of the volume of water held in storage.

In this research, the discharge data of the monitored springs have been fitted using the Maillet method on every baseflow recession period (not characterised by changes in slope on a single hydrograph recession limb) individuated along the analysed time interval both in the pre-seismic and in the post-seismic periods.

It is worth noting that the recession analysis is performed with the sole scope of assess any change in aquifer parameters, such as hydraulic conductivity or to identify possible modification occurred on aquifer geometry due to seismic sequence. In fact, the baseflow recession (the most stable section of any hydrograph, and also the most characteristic feature of the global response of an aquifer) can be assumed as that segment of a hydrograph least influenced by the temporal and spatial variations of infiltration (Figure 2.14) (Taylor & Greene, 2008). Thus, operating on this section of the hydrograph, any detected change (between pre- and post-seismic) can provide important information about modification within the aquifer operated by external factors such as earthquakes.

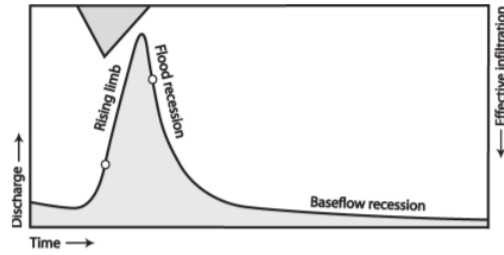


Figure 2.14 Representation of a spring flow discharge graph with nomenclature referring to the various parts, modified from Martin & White, 2008.

The continuous daily discharge data of Pescara and Capodacqua springs have been fitted starting from the baseflow recession period of the year 2003, Foce spring discharge has been fitted starting from 2010, while for the Sordo spring only the years 2015 and 2016 before the seismic period and the years 2017 and 2018-2019 for the post-seismic were available to perform the Maillet method. Capotenna spring discharge were fitted on weekly time interval starting from 2008, while Sassopaccato spring discharge were fitted on monthly time interval starting from the baseflow recession period of 2007.

2.2 Tracer tests

2.2.1 Selection of injection point and injection method

In a central position with respect to the study area there is the presence of a well-developed tectono-karstic basin filled by Holocene lacustrine deposits named Pian Grande Plain (Figure 2.15).



Figure 2.15 Pian Grande Plain with a detail of polje.

The large plateau is located almost 1300 m a.s.l. and is bordered by a normal fault system. The eastern side is bordered by the scarp of the Mt. Vettore and by the higher reliefs of the Umbria-Marche Apennines chain, while the western edge is bordered by the fault at the base of Mt. Castello and is by reliefs at lower elevations. From a geomorphological point of view, the presence of a well-developed exo-karstic system within the plain is highlighted by the occurrence of several dolines, polje and a sinkhole named Mèrgani located on the southern-western side of the plain in which a sinking stream (Mèrgani creek) is located and represents one of the most developed karstic phenomena in the Sibillini area (Figure 2.16). Others minor karst forms such as solution holes, karren, rillenkarrren and kamenitza were mapped (Celso Giovagnotti, 1979; Coltorti & Farabollini, 1995,) and involves primarily the Calcare Massiccio geological Fm. and partially the Maiolica geological Fm. The Scaglia Rossa geological Fm. is characterised only in limited areas by the presence of karstic shapes similar to dolines.

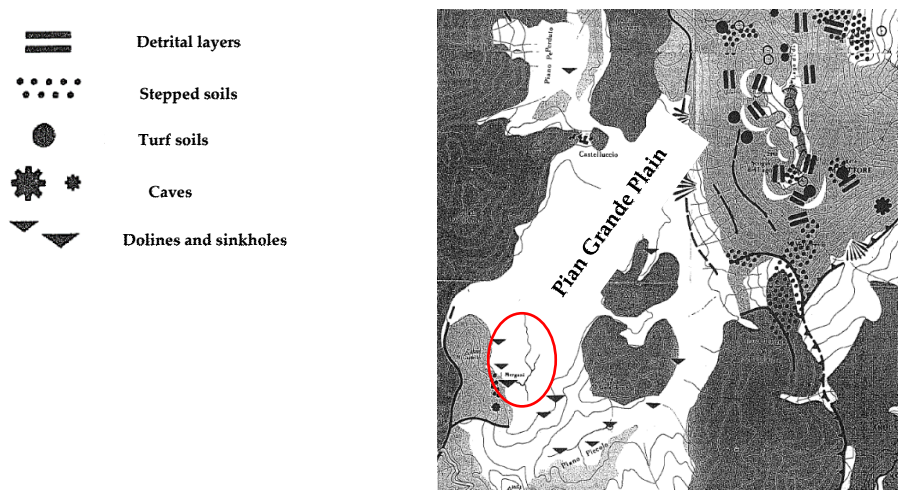


Figure 2.16 Historical map of Pian Grande Plain area with typical karstic forms. The Mèrgani sinking stream is evidenced by the red circle. Modified from Giovagnotti, 1979.

Often the karstic forms are masked by the thick fluvial-lacustrine deposits which can reach a thickness over 400 m at lower elevation. However, due to a lack of direct speleological investigations has meant that knowledge of the karst system, even in the shallow portion is not well known in the entire Sibillini area, unlike other carbonate areas of the central Apennines (Galdenzi & Menichetti, 1995; Tamburini & Menichetti, 2020).

The Mts. Sibillini area has been the setting of artificial tracer tests since long time (Lippi Boncambi, 1950) and some hydrogeological studies of the Pian Grande Plain were performed between 1947-1963 (Lippi Boncambi, 1947; Lippi Boncambi 1963). In this framework, some tracer tests to investigate the hydrogeological connection between Pian Grande Plain and the springs located westward in the Norcia Plain (about 600 m a.s.l.) were performed and the hypothesis of the existence of hydraulic connections between the injection point and the springs were drawn. However, these published outcomes lack quantitative results, probably due to the analytical equipment available at the time.

Recently Nanni et al., 2020 conducted some artificial tracer tests before, during and just after the seismic period. In each test, the Mèrgani sinking stream has been selected as tracer injection point. With the aim of characterising the hydrodynamic response of the springs under different hydrogeological conditions, to characterize the aquifer parameters and to evidence the groundwater flowpaths modifications due to the seismic sequence occurred in 2016-2017, several long time and periodic tracer tests were performed integrating the ones conducted by Nanni et al., 2020.

All the tracer tests have been performed by applying the sudden injection method and two kinds of dye-tracers were selected and used alternately.

During the first tracer test (TEST1), performed on 12 February 2016 (pre-seismic), 2 kg of Na-Fluorescein ($C_{20}H_{10}Na_2O_5$) has been injected into the sinkhole. This is considered a sort of pilot test to observe the main hydraulic connections between the sinkhole and the springs located nearby. The second test (TEST2), performed on 9 June 2016 (pre- and co-seismic), was characterized by the injection of 29 kg of Tinopal CBS-X ($C_{28}H_{20}Na_2O_6S_2$).

After the seismic sequence four tracer tests were performed: the first one on 20 March 2017 (TEST3), the remaining three were conducted specifically within the framework of this research on 20 March 2018 (TEST 4), 8 February 2019 (TEST 5), and 15 June 2020 (TEST6). TEST3 consisted in the injection of 85 kg of Tinopal CBS-X in the Mèrgani sinkhole. TEST4, TEST5 and TEST6 were characterized by the injection of 16 kg, 27 kg of Fluorescein and 80 Kg of Tinopal CBS-X respectively.



Figure 2.17 Artificial tracer injection operations conducted during TEST4 on 20th March 2018 in Mèrgani sinking stream.

During each test, just before the injection of tracer in the sinkhole, a discharge measurement has been performed to quantify the water inflowing in the sinkhole. Starting from January 2018, the Mèrgani creek has been instrumented by a hydrometric station and a continuously daily discharge inflowing in the sinkhole has been recorded.

The details of artificial tracer tests are summarized in table 2.6 with specification about the inflow discharge in the Mèrgani sinkhole during the injection operations.

Table 2.6 Details of artificial tracer tests performed in the Sibillini Mts. area.

N° of test	Date of injection	Type of tracer	Injected mass (kg)	Mèrgani discharge (L/s)	Period	Reference
TEST1	12-Feb-2016	Fluorescein	2	1550	Pre-seismic	Nanni et al., 2020
TEST2	09-Jun-2016	Tinopal CBS-X	29	10	Co-seismic	Nanni et al., 2020
TEST3	20-Mar-2017	Tinopal CBS-X	85	136	Post-seismic	Nanni et al., 2020
TEST4	20-Mar-2018	Fluorescein	16	130		This study
TEST5	08-Feb-2019	Fluorescein	27	70		This study
TEST6	15-Jun-2020	Tinopal CBS-X	80	50		This study

2.2.2 Monitoring points

The tracer tests monitoring points location has been selected in accordance with the hydrogeological setting and they have been radially positioned with respect to the Pian Grande Plain. In general, they are located at the end of the aquifers' outcrops along the streams, in the proximity of the main linear springs or in proximity of the punctual springs used for public water supply by the local water management companies. Two different devices were used to detect the tracer concentration in the monitoring points:

- Field fluorimetric probes
- Portable water samplers

Two kinds of continuous fluorimetric probe were installed: the one produced by Albillia Co. (Neuchatel, Switzerland) and the PME Cyclops-7 Logger (PME Inc. USA) both containing various optics for tracer detection, a standalone power supply and a data logger for the measured data storage.

The Albillia GGUN-FL24 and GGUN-FL30 fluorometers are characterised by a minimum detection limit of 2×10^{-2} ppb, whereas the fluorometer PME Cyclops-7 Logger has 1×10^{-2} ppb as Fluorescein detection limit and 6×10^{-1} ppb as concerns the Tinopal CBS-X detection limit.

The field fluorimetric probes have been calibrated before each tracer test by using a blank sample and one calibration standard at 100 ppb of dye tracer prepared using the same water collected in the monitoring point. Tracer concentration data have been acquired every 10 minutes during the various tests.

The memory data storage of the field fluorometric probe was periodically downloaded to check the correct execution of the tests.

Two kinds of portable automatic water sampler were installed: a model 3700C produced by Teledyne ISCO (Lincoln, Ne, USA) and a WTW produced by Xylem Analytics (Weilheim, Germany). The samplers are sequential collectors containing 24 bottle and for which a specific time interval sampling can be selected by the operator. For the various tests, the automatic sampling time was settled at 12 h. The samplers were periodically emptied once the 24 bottles have been filled, and the water samples were returned to the laboratory.

Moreover, all the monitoring points have been instrumented by an active carbon trap used to fix tracer and every 15 days a water sample was collected manually in each monitoring point during the substitution of fluorometric carbon-active traps. Once collected, both the carbon traps and the water samples were taken to the laboratory.

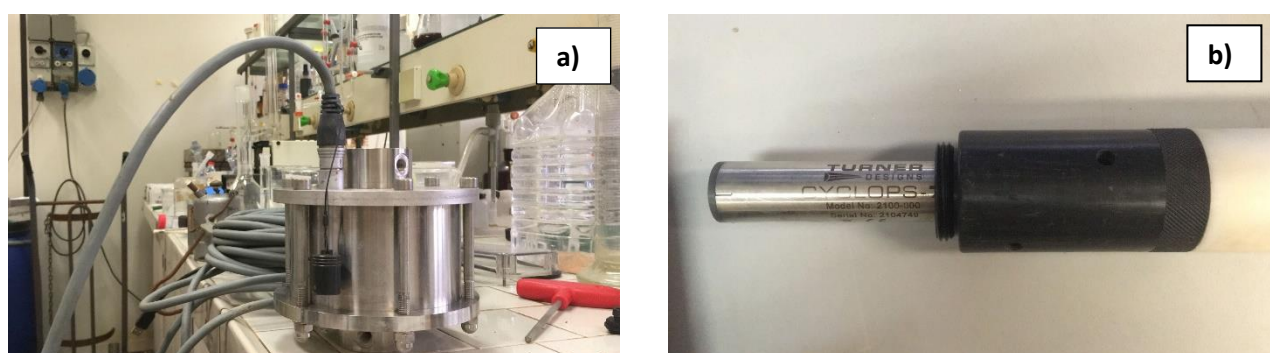


Figure 2.18 Field fluorometric probes used for artificial tracer tests: (a) Albilia GGUN-FL24 and GGUN-FL30 produced by Albilia Co. (Neuchatel, Switzerland) and (b) PME Cyclops-7 Logger (PME Inc. USA).

2.1.1 Laboratory operations

Both Tinopal CBS-X and Fluorescein (if present), were extracted by the carbon-active traps using a potassium hydroxide solution in methanol after being dried in an oven for 24 hours at a temperature below 60°C. Once collected (manually and by means of the automatic samplers), the water samples and the solutions obtained by the extraction (filtered at 45 μm to prevent any impurity entry in the instrument) were analysed by a model RF-6000 laboratory spectrofluorometer (Figure 2.19) produced by Shimadzu Corporation (Kyoto, Japan) having an emission spectrum with a resolution of 1.0 nm, and measures wavelength with a ± 1.0 nm accuracy.

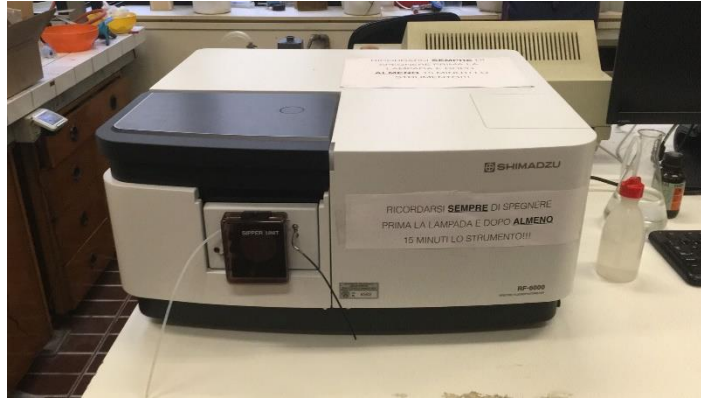


Figure 2.19 RF-6000 laboratory spectrofluorometer produced by Shimadzu Corporation (Kyoto, Japan).

The calibration of the Shimadzu spectrofluorometer was performed by using three concentration standards (10, 20 and 100 ppb) prepared using the same water collected in the field and a blank sample for each monitoring point, sampled before each new tracer injection (Figure 2.20).

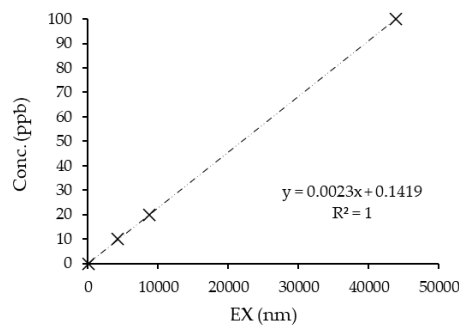


Figure 2.20 Calibration curve of Shimadzu RF-6000 laboratory spectrofluorometer.

The tracer concentration was determined by setting the excitation wavelength (EX) = 490 nm and the emission wavelength (EM) = 520 nm for the Fluorescein while the excitation wavelength (EX) = 380 nm and the emission wavelength (EM) = 430 nm have been selected for the Tinopal CBS-X.

2.2.3 Tracer tests processing

For each monitored point, it was possible to derive the breakthrough curve (BTC), which describes the trend in tracer concentration as a function of time, making it possible to visually grasp how the tracer is released into the aquifer. The graphical data were processed using MS Excel while the free software QTracer2 (EPA 2002, USA) was used for the quantitative evaluation of tracer tests, which was necessary to obtain the aquifer volumes (involved by the tracer), the flow velocities (maximum

and mean), mean residence times and longitudinal dispersivity. Finally, from the mass balance for each monitored point the validity of each test has been evaluated.

Qtracer2 is a program developed to calculate total tracer-mass recovery, hydraulic and geometric parameters from groundwater tracer tests in various hydrologic system, especially in those characterised by karstic and fractured aquifers (EPA 2002, USA). It requires as input parameters the tracer concentration and the discharge data measured in the monitored point. For this reason, only TEST4 and TEST5, for which a more complete monitoring points network and a fairly continuous discharge record was available, have been processed by Qtracer2. In the springs for which the discharge data were available only for discrete interval, the discharge value was considered constant and equal to the previous value until a new discharge measurement was performed. The daily, weekly, or monthly discharge data were downscaled hourly (considered constant) and the tracer concentration value (recorded every 10 minutes by the fluorometric probes and every 12 hours by the automatic water samplers) were grouped on an hourly scale calculating the hourly mean concentration value or by downscaling the data (considered constant). The tracer concentration values obtained by the water samples (manually collected), as well as the tracer presence in the active carbon traps were processed to obtain quali-quantitative information about the hydraulic connections between the injection point and the monitoring points.

2.2.4 Sliding Windows Cross-correlation function vs tracer tests

The springs located in the eastern and southern-east side of the massif (i.e., Foce spring, Capodacqua spring and Pescara del Tronto spring), which displayed a long-term discharge decreasing after the seismic sequence, have been the object of an advanced data-driven investigation coupled with tracer tests results. Those springs are characterised by continuous daily discharge data without significant interruption between 2013 and 2019 (Figure 2.21 b), as well as tracer tests continuous monitoring data almost complete during the same period. These circumstances permitted to develop a transient time series analysis via Sliding Window Cross Correlation Function (SWCCF) (Delbart et al., 2014) to be compared to the tracer tests (from TEST1 to TEST5) results.

Three rain gauges and one snow thickness gauge (above the ones belonging to the Regione Marche Civil Protection Agency network) have been selected to measure the precipitations in the area (Figure 2.21 a). In particular, Mt. Prata rain and snow cover thickness gauge, located at 1813 m a.s.l., has been used to represent and analyse the main meteorological contribution from the highest

portion of the massif, while Montemonaco and Capodacqua rain gauges were used to measure the local meteorological condition, nearby the considered springs (Table 2.7).

Table 2.7 Details of monitoring points selected to perform the transient time series analysis via SWCCF.

Monitoring point	Parameter	Instrument
Pescara spring	Discharge	Water level gauge
Capodacqua spring	Discharge	Water level gauge
Foce spring	Discharge	Water level gauge
Mt. Prata station	Snow cover thickness Rainfall	Snow gauge Rain gauge
Mt. Monaco station	Rainfall	Rain gauge
Capodacqua station	Rainfall	Rain gauge

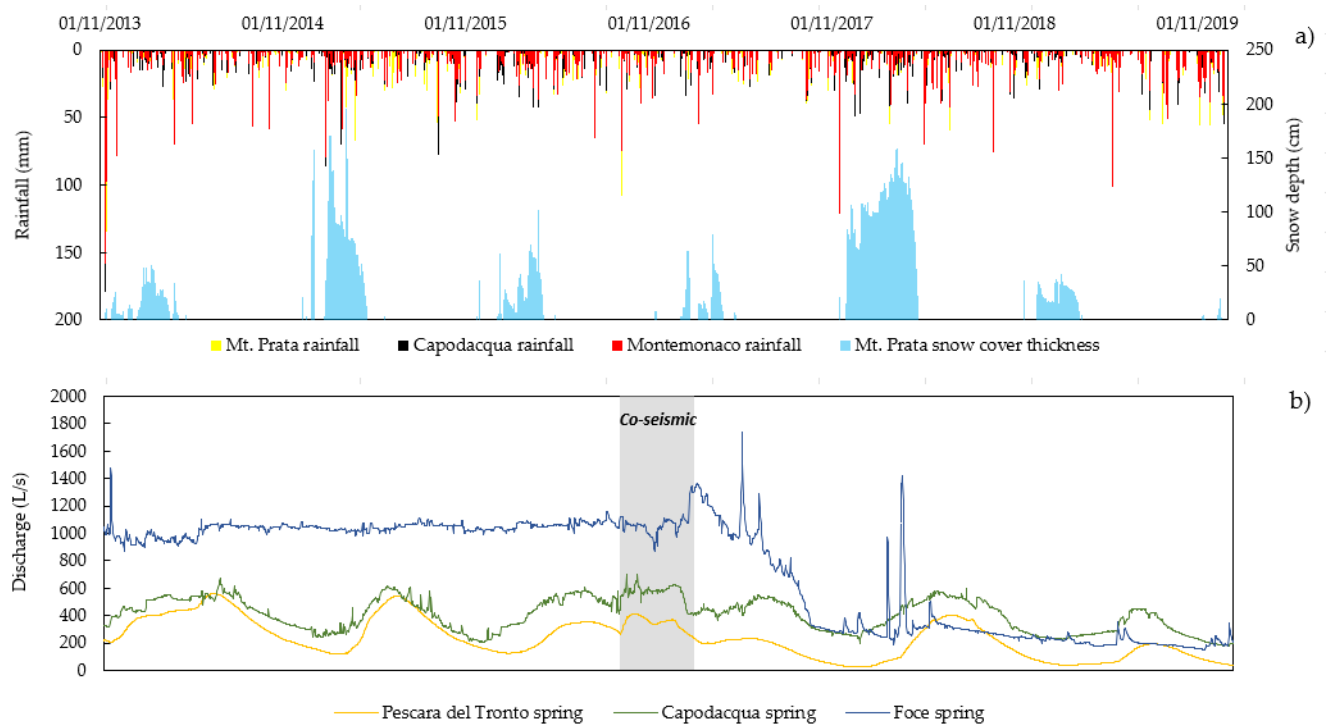


Figure 2.21 (a) Input data to perform SWCCF (rainfall and snow depth) (b) output data to perform SWCCF (springs discharge data). The grey window refers to the co-seismic period (24th August 2016 – 18th January 2017).

To assess mid- and long-term earthquakes-modification of inflows-outflow dynamics in the aquifers feeding those springs, the Sliding-Window Cross-Correlation Function (SWCCF) has been selected. This is an enhanced version of the well-known bivariate technique (Cross-Correlation Function, CCF) (Chiaudani et al., 2017; Viaroli et al. 2019) able to define the output (i.e., spring discharge, or hydraulic head) response time to different inputs (i.e., snowmelt, and rainfall), to identify the different recharge modes affecting the outflow behavior over the time, and to evaluate the actual influence of an inflow on the overall amount of the discharging groundwater.

As for the CCF, a correlation coefficient ($r_{xy}(k)$) is calculated for each time lag (k), by the following equation:

$$r_{xy}(k) = \frac{\frac{1}{N} \sum_{t=1}^{N-k} (x_t - \bar{x})(y_{t+k} - \bar{y})}{\frac{1}{N} \sum_{t=1}^N \sqrt{(x_t - \bar{x})^2 (y_t - \bar{y})^2}} \quad \text{with } k = 0, 1, 2, \dots, K$$

where N is the number of observations in the time-series, (x_t, y_t) are the pairs of data, and $\bar{x}$ and $\bar{y}$ are the mean values of each time-series. Time lags characterized by high positive $r_{xy}(k)$ values correspond to the actual discharge response times to recharge events.

The ground shaking has been proven to cause both reversible and irreversible modifications to groundwater flow, relationships between input and output time-series change as well. This implies that results obtained by a traditional CCF are unreliable. On the other hand, the SWCCF allows performing a transient time-series analysis, that allows the identification of all the changes in the inflow-outflow system related to the seismic sequence. It consists of performing a CCF in time windows with a regular width, partially overlapped. In this research, the selected windows are 2-year wide (i.e., 24 months) and have been shifted by 180 days (i.e., 6 months). This results in nine windows, two interesting the pre-seismic phase, five partially overlapping the co-seismic phase and two interesting the post-seismic phase (Table 2.8).

Table 2.8 Details of windows to perform the transient time series analysis via SWCCF.

Window	N° months	Starting date	Ending date	Period
1	24	01/11/2013	31/10/2015	Pre-seismic
2	24	01/05/2014	30/04/2016	
3	24	01/11/2014	31/10/2016	Co-seismic
4	24	01/05/2015	30/04/2017	
5	24	01/11/2015	31/10/2017	
6	24	01/05/2016	30/04/2018	
7	24	01/11/2016	31/10/2018	
8	24	01/05/2017	30/04/2019	Post-seismic
9	24	01/11/2017	31/10/2019	

In large carbonate aquifers located in mountain region, the rainfall inflow generally represents a limited amount of the whole recharge volume, with respect to snowmelt (Nanni & Rusi, 2003; Fiorillo et al., 2015; Rusi et al., 2018; Chiaudani et al., 2019); hence, it is necessary to remove the

baseflow component of discharge time-series to amplify the rainfall recharge effect of spring behaviour. This aspect was achieved by removing the daily baseflow (D_{BFI}), calculated by a local-minimum method using the BFI+ 3.0 freeware software (Gregor, 2010), from the raw daily discharge time-series (D_{tot}) to obtain the residual daily discharge component (D_{res}):

$$D_{res} = D_{tot} - D_{BFI}$$

Besides, the higher the correlation coefficient value, the more the specific inflow is influent on the discharge behavior over the selected period.

The correlation obtained by the SWCCF have been tested through the Student's t-test (95% of significance level) for $r_{xy}(k)$ values ($p < 0.05$) (Seidenbecher et al., 2003):

$$t = \frac{r_{xy}(k)\sqrt{N-2}}{\sqrt{1-r_{xy}(k)^2}}$$

where t is the t-value, which is equal to 1.645 for the 95% significance level. Positive correlation coefficients below the $r_{xy}(k)$ values that give $t = 1.645$ in the previous equation are considered not significant.

2.3 Isotopic and hydrochemical sampling

2.3.1 Monitoring points

With the aim to characterize the mean recharge elevation of the punctual and linear springs of the Mt. Sibillini area, isotopic analysis of water has been carried out. Water samples were collected in 50 mL high-density polyethylene bottles sealed by plastic inserts to avoid water evaporation both at rainfall sampler, punctual and linear springs (Figure 2.22).

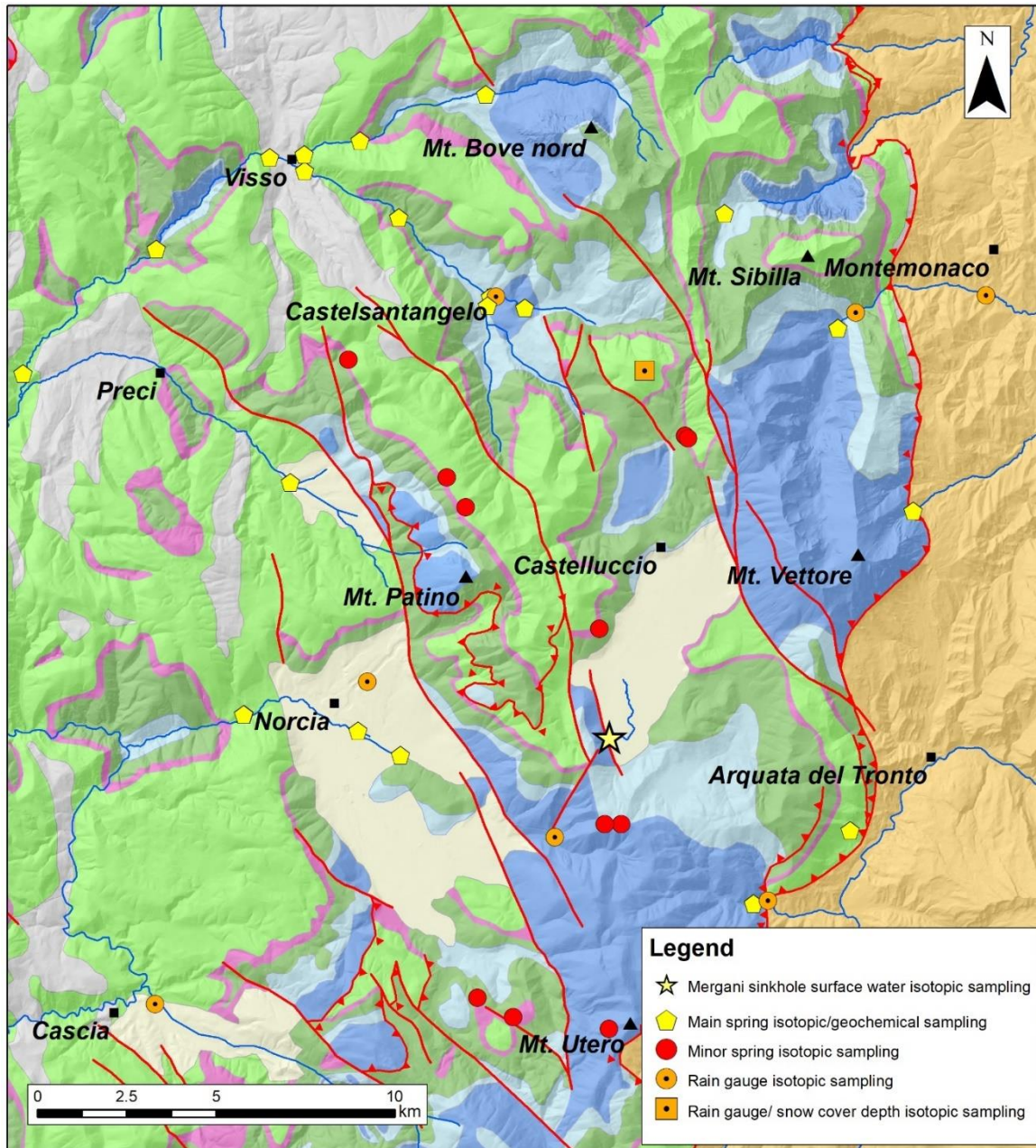


Figure 2.22 Map of hydrochemical-isotopic monitoring points.

A new local correlation between the meteoric $\delta^{18}\text{O}$ and the elevation (i.e., $\delta^{18}\text{O}$ gradient) was determined by eight rainfall samplers mainly located between 600 m and 974 m a.s.l. The higher rainfall sampler was located at 1800 m a.s.l. in the proximity of Mt. Prata rainfall/snow cover thickness station. The meteoric bimonthly sampling was performed cooperating with Perugia University between 2013 and 2019 (Table 2.9) covering the entire hydrologic year. Standard precipitation gauges with funnel diameter of tot cm and 5" gauges (12.72 cm of diameter) were used as rainfall sampler connected, by means of an externally insulated PVC pipe, to a buried totalizers constituted by 20 L volume tank. To avoid evaporation processes, a layer (~1 cm) of pure paraffin oil was inserted in the tank between May and September. Precipitation amount recorded by the rain gauges was compared with the amount of water measured manually for each sampling period. The average $\delta^{18}\text{O}$ values were determined by weighting the amount of the water collected by the rainfall sampler at each sampling respect to the total precipitation of the entire period of observation. To fill the gap between 974 m and 1800 m a.s.l., 11 minor punctual springs in the area have been selected. The selected small springs are characterized by a low discharge value during the entire hydrologic year and accordingly to their limited extensions and elevation of recharge area (not so different from the punctual spring altitude) and their rapid flow response to the rainfall events, they can be used as natural pluviometers (Tazioli et al., 2019; Mussi et al., 2017; Doveri & Mussi, 2014). All the minor springs were monthly sampled during the entire hydrologic year in 2017.

Table 2.9 Details of meteoric water isotopic sampling network. *Rainfall and snow cover sampling.

	Sampling period	Name	Elevation m a.s.l.
Rainfall samplers	2017-2018	Cascia	604
	2017-2019	Norcia	700
	2013-2017	Castelsantangelo sul Nera	737
	2013-2017	Vallefiume	755
	2013-2017	Capodacqua	805
	2013-2017	Foce	930
	2013-2017	San Pellegrino	974
	2013-2019	Mt. Prata*	1800
Minor springs	2017	Prati di San Marco	1093
	2017	Fonte Acqua d'Abeto	1267
	2017	Fonte Secca	1299
	2017	San Lorenzo Basso	1439
	2017	San Lorenzo Alto	1473
	2017	Fonte dell'Asinella	1500
	2017	Fonte Conserva	1520
	2017	Fonte Civitella	1557
	2017	Fonte Utero	1577
	2017	Fonte del Grillo	1598
	2017	Fonte Fausole	1645

The main linear and punctual springs emerging in the area have been sampled to: i) observe any changes in $\delta^{18}\text{O}$ or hydrochemical composition after the seismic sequence, ii) define springs mean elevation recharge area (before and after the seismic sequence) iii) develop an advance isotopic recharge model for the springs for which at least five or more $\delta^{18}\text{O}$ data were collected before the seismic sequence (see paragraph 2.3.4.2). The availability of a robust isotopic and hydrochemical dataset developed by the Università Politecnica delle Marche priorly to the seismic sequence permitted to compare the results. The springs isotopic/hydrochemical sampling and data collection have been performed in collaboration with Perugia University between the years 2016 and 2020 and it was integrated by the data published by (Rosen et al., 2018) for the Upper Nera River basin and by (Petitta et al., 2021) for Capodacqua spring. All the linear springs samples collected along streams and rivers were taken away from heavy rainfall periods to sample only the groundwater contribution.



Figure 2.23 Hydrochemical-isotopic sampling.

2.3.2 Snowfall and snowmelt

The first assessment of the snowfall regime highlights the occurrence of snowpack persisting from late autumn until late spring in the highest topographic area (> 1500 m a.s.l.). The snow depth can reach more than 200 cm near the highest peaks. This evidence led to the necessary evaluation of the snow water equivalent (SWE) and a preliminary snow cover isotopic characterization to better understand the springs recharge processes occurring in the area. To do that a snowpack sampling system was developed in the vicinity of the Monte Prata rainfall/snow depth station, by choosing a site not characterised by snow-drift effects (sheltered from strong wind conditions able to mobilise the snow on the ground) (Adok, 1977; Hawley & Millstein, 2019).

The snow sampling consisted of seven steps, developed as follows:

- i) create a vertical embankment on the snowpack
- ii) measuring the snowpack depth
- iii) sampling horizontally at different depths (on the surface, in half position within the snow depth and at the bottom) by means of a stainless-steel geotechnical die with known volume (0.996 cm^3) avoiding snow compaction.
- iv) empty the contents of the geotechnical die into high-density polyethylene bottles sealed by plastic inserts to avoid water evaporation.
- v) determine the snow density (kg/m^3) and the %V/V ratio once the snow sample is melted (at room temperature), to empirically calculate its SWE.
- vi) an aliquot (50 mL) of the melt water is inserted into high-density polyethylene bottles sealed by plastic inserts to avoid water evaporation and sent to the Stable Isotopes Laboratory of National Research Council (CNR-Pisa, Italy) to perform isotope analysis.

The sampling was carried out during the melting phase of the snowpack in the year 2021 after the isotope fractionation processes occurred. The data were compared with a previous sampling campaign carried out at the same location in 2013 both during the snow cover accumulation and both during the snow melting period.



Figure 2.24 Snow cover sampling field operations, performed near Mt. Prata rainfall/snow depth station.

2.3.3 Laboratory operations

Once collected, the water samples for isotope analyses were filtered with a 0.45 mm Whatman glass microfibre filter and poured into a separator funnel by a glass syringe, then the water flowed into the glass flask below the separator funnel.

Isotopic analyses of oxygen and hydrogen of the water were performed in two different laboratories:

- i) INGV of Naples, using near infrared laser analyzer (L2130i, Picarro, Santa Clara, CA, USA) with the wavelength-scanned cavity ring down spectroscopy technique (analytical error $< 0.08\text{‰}$; $\delta^{18}\text{O}$ and $< 1.0\text{‰}$ for $\delta^2\text{H}$). The data were reported vs Vienna Standard Mean Ocean Water, V-SMOW (Rozanski et al., 1993)
- ii) Stable Isotopes Laboratory of National Research Council (CNR-Pisa, Italy). In this case, the $\delta^{18}\text{O}$ value was determined through the analysis of gaseous CO_2 , previously equilibrated with water at 25°C (Epstein & Mayeda, 1953) by means of a mass spectrometer FINNIGAN MAT252. The mass-spectrometric measurement of the $^{18}\text{O}/^{16}\text{O}$ ratio requires correction because a fractionation between CO_2 and H_2O occurs. Analytical precision on $\delta^{18}\text{O}$ values is better than 0.10‰ . The $\delta^2\text{H}$ value was measured by means of the Europa Scientific GEO 20-20, by reducing the water to elemental hydrogen (Coleman et al., 1982) using magnesium instead of zinc. Because the totality of water is reduced and all hydrogen is converted to hydrogen gas, isotopic fractionation does not occur, and a correction of the mass-spectrometric measurement is not necessary. The analytical error for $\delta^2\text{H}$ is 1.5‰ . The isotopic ratios of CO_2 and H_2 were measured by dual inlet mass spectrometer. Also in this case, isotope results are reported in permille notation vs Vienna Standard Mean Ocean Water, V-SMOW.

The hydrochemical analyses were conducted following different methodologies. In particular, the pre-seismic analyses were performed by Università Politecnica delle Marche - Geochemical laboratory using an ion chromatography system (ICS-1000, Dionex, Waltham, MA, USA) determining major ions concentration, while the post-seismic samplings (completed between September 2016 and June 2020) were performed both by Università Politecnica delle Marche and Perugia University. For these samples, temperature, pH, Eh, electrical conductivity and HCO_3^- were measured in the field. HCO_3^- concentration was determined by acid titration with 0.01 N HCl using methyl orange as indicator. One of the sample aliquots was filtered upon sampling through $0.45\text{ }\mu\text{m}$ membrane filters and then acidified with 1% of 1:1 diluted HCl . Chemical analyses were made at

the Geochemical laboratory of Perugia University. Calcium and Mg concentrations were determined by atomic absorption (AA) flame spectroscopy on the acidified sample while Na and K were determined by atomic emission (AE) flame spectroscopy, using an Instrumentation Laboratory aa/ae spectrophotometer 951. Cl and SO₄ were determined by ion chromatography using a Dionex DX-120 instrument.

2.3.4 Isotopic data processing

The first approach to identify the mean recharge area elevation for a spring, using its $\delta^{18}\text{O}$ content, consists in the application of the $\delta^{18}\text{O}$ gradient, obtained by the linear regression of $\delta^{18}\text{O}$ rainfall content vs elevation, to the spring isotopic data.

To assess if the distribution of the local $\delta^{18}\text{O}$ rainfall content, with respect to the one available in literature, was significantly different, a GIS-based procedure integrated by the one-way ANOVA test and the Kolmogorov-Smirnov test was performed. The computation started by applying the $\delta^{18}\text{O}$ -elevation linear regression (Giustini et al., 2016; Tarragoni, 2006) to the Digital Terrain Model (DTM 20x20m cell size) with the aim to spatialize the $\delta^{18}\text{O}$ rainfall content as a raster file in the whole study area. Secondary, by applying the same methodology used to obtain the $\delta^{18}\text{O}$ raster files (Giustini et al., 2016; Tarragoni, 2006), a $\delta^{18}\text{O}$ rainfall content map for the area was determined by applying the local $\delta^{18}\text{O}$ -elevation linear regression. Finally, a comparison between the $\delta^{18}\text{O}$ rainfall content (Giustini et al., 2016; Tarragoni, 2006) and the spatialized local one was performed by using both the statistical tests and shown as boxplot graphs.

To observe any difference in the mean elevation of springs recharge areas before and after the seismic sequence, the local $\delta^{18}\text{O}$ rainfall content vs elevation relationship was applied to the springs $\delta^{18}\text{O}$ data (pre- and post-seismic). For five springs (characterised by a robust pre-seismic isotopic dataset, for which at least five or more $\delta^{18}\text{O}$ data were available before the seismic sequence) a statistical approach based on finding abruptly signal changes was applied.

2.3.4.1 Find changes in isotopic signals

In order to detect abrupt changes in the springs isotopic signal the `findchangepts` function in MATLAB® was used and applied to the springs $\delta^{18}\text{O}$ content over time. This function consists in finding a sample or time instant at which some statistical property (mean of the signal, its variance,

or a spectral characteristic of a signal) changes abruptly (Trauth, 2017). To find a signal changepoint, the function employs a parametric global method following those steps:

- i) Chooses a point and divides the signal into two sections
- ii) Computes an empirical estimate of the desired statistical property for each section
- iii) At each point within a section, measures how much the property deviates from the empirical estimate. Adds the deviations for all points
- iv) Adds the deviations section-to-section to find the total residual error
- v) Varies the location of the division point until the total residual error attains a minimum

The procedure is clearest when the chosen statistic is the mean. In that case, findchangepts function minimizes the total residual error from the "best" horizontal level for each section.

2.3.4.2 The GIS based- $\delta^{18}\text{O}$ isotope overlapping method

The five springs characterized by the robust isotopic dataset prior to the seismic sequence were analyzed developing a GIS based- $\delta^{18}\text{O}$ isotope overlapping method. In this case for each analyzed spring, the $\delta^{18}\text{O}$ gradients available in literature for central Italy (Giustini et al., 2016; Tarragoni, 2006) were applied and compared to the local one determined in this research. Then, starting from the 1:10.000 Regione Marche and the 1:10.000 Regione Umbria official geological map, a detailed hydrogeological complexes map was obtained for each spring recharge area (determined by the hydrological balance), by combining the geological units into appropriate complexes with similar hydrogeological features (Viaroli et al., 2021). The conceptual aggregation of different geological units into hydrogeological complexes was supported by the available literature (Boni et al., 1991; Regione Marche, 1991; Boni & Petitta, 1994; Boni & Preziosi, 1994; Boila & Checcucci, 1995; Mastrorillo, 1996; Mastrorillo, 2001; Boscherini et al., 2005; Tarragoni, 2005, 2006; Boni & Petitta, 2007; Petitta, 2007; Preziosi, 2007; Boni & Petitta, 2008; Boni et al., 2009; Mastrorillo et al., 2009; Mastrorillo et al., 2012; Banzato et al., 2013; Nanni et al., 2013; Banzato, 2014) and considering specific or detailed in-situ hydrogeological characterization. A GIS clipping procedure was applied over the hydrogeological complexes belonging to the spring recharge area. The recharge areas for the analysed springs were iteratively optimized with the mean $\delta^{18}\text{O}$ value, integrating the results obtained from the hydrological balance during the pre-seismic period (paragraph 3.2). To validate and refine the isotopic results obtained in step two, a third step was introduced. For each recharge area, the $\delta^{18}\text{O}$ meteoric content distribution, clipped over each hydrogeological complex, was plotted in a box plot and chart bar graph to highlight the distribution of $\delta^{18}\text{O}$ for each complex and calibrate

the advanced isotopic recharge model for all the analyzed spring. In fact, subsequently, the average $\delta^{18}\text{O}$ content $\pm \sigma$ of the analyzed springs (pre- and post-seismic) has been overlapped to the $\delta^{18}\text{O}$ hydrological complexes distribution calculating the individual contribution of each complex (expressed as relative percentage to the total) that can potentially feed the analyzed springs (Figure 2.25).

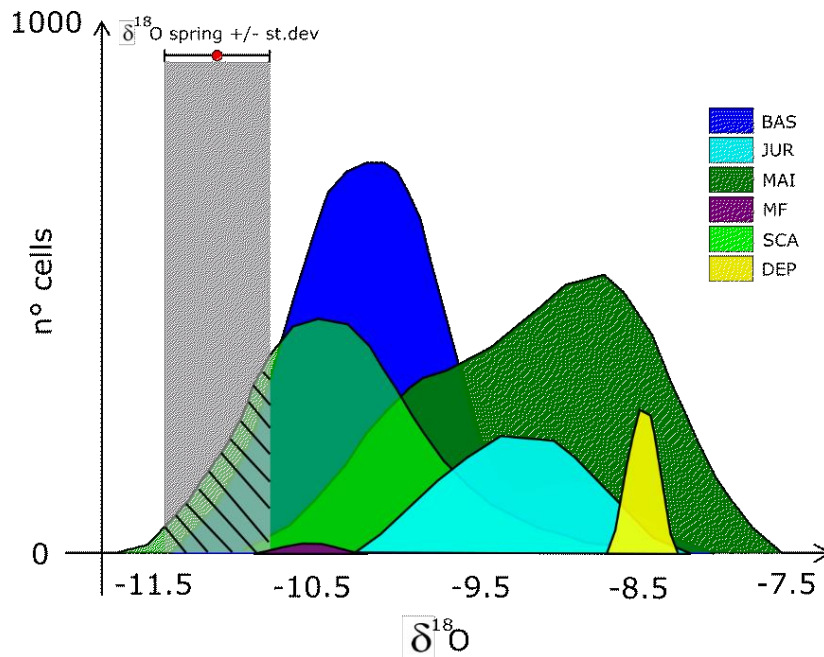


Figure 2.25 Graphical example of geo-isotopic modelling. The graph shows the hydrogeological complexes $\delta^{18}\text{O}$ distribution obtained by the GIS clipping procedure for each analysed spring. The relative percentage to the total can be calculated overlapping the average spring $\delta^{18}\text{O}$ content $\pm \sigma$.

2.4 Geological field work and GIS supported analysis (The northern portion of the study area - Upper Nera River basin)

The Upper Nera River basin, resulting in a complex structural setting, drains most of the groundwater amount towards the northern sectors, leading to a diffuse presence of punctual and linear springs due to the intersection of the main streams and the saturated portion of the aquifers (Fronzi et al., 2021). However, its hydrogeological boundaries, as well as the hydrogeologic connections between the catchment, and possible external recharge areas, have never been demonstrated by using a quantitative approach (Scarpelli et al., 2019).

Hence, in order to better characterise the geological setting, focusing on the evidence of some geo-structural elements connected with the groundwater circulation, detailed field surveys were

performed in specific monitoring points (i.e., linear and punctual springs) and represented along schematic geological cross-sections across the springs. The surveys were conducted using as base maps the 1:10.000 Regione Marche official map, the 1:40.000 Sibillini geological map (Pierantoni et al., 2013) and the hydrogeological map published by Viaroli et al., 2021. In the Upper Nera River basin, characterised by the presence of several linear springs (meant as gaining streams), the structural geology information depicted during the fieldwork were integrated with the existing geological maps and a GIS repository has been developed aiming at identifying and clarifying the groundwater-surface water interaction. Moreover, a GIS procedure has been developed for each sub-catchment characterizing the Upper Nera River basin (i.e., Castelsantangelo stream sub-basin and Ussita stream sub-basin), calculating the mean elevation by the available Digital Terrain Model (DTM 10 meters' cell size) of the area. Subsequently, each hydrogeological complex outcropping in the area has been intersected with the DTM and the mean elevation for each complex for both basins was calculated. At last, a topographic intersection between the outcropping hydrogeological complexes and the streams were performed, to investigate the correlation between streams elevation, drainage and recharge areas. The mean elevation of each hydrogeological complex outcropping in the sub-basins determined by the GIS procedure was then compared to the mean recharge elevation achieved by the isotopic data for the spring draining the same hydrogeological complex.

3 Results

3.1 Precipitation and meteo-climatic results

The SPI-12 calculated from monthly rainfall data over a period of 30 years (1991-2021) shows approximately two circumstances (2002, 2007) in which its value is lower than -2 (extremely dry conditions) in all the analysed stations (Figure 3.1). In the years 2011-2012 and 2017 severely dry conditions ($SPI < -1.5$) are recognised in all the station except for Vallo di Nera rainfall station (Tyrrhenian side) which is marked by extremely dry conditions.

As far as wet periods concerned, the SPI-12 has been higher than 1.5 (very wet conditions) in four intervals (1999-2000, 2009-2010, 2013-2014 and 2018) more visible for the Adriatic and the southern domain of the study area, while the Tyrrhenian side is marked by an extremely wet condition between October 2020 and July 2021.

Before the year 1999 the dry and wet conditions interested the sector of the study area in different periods with a visible delay between Montemonaco (Adriatic side) and the others monitoring stations.

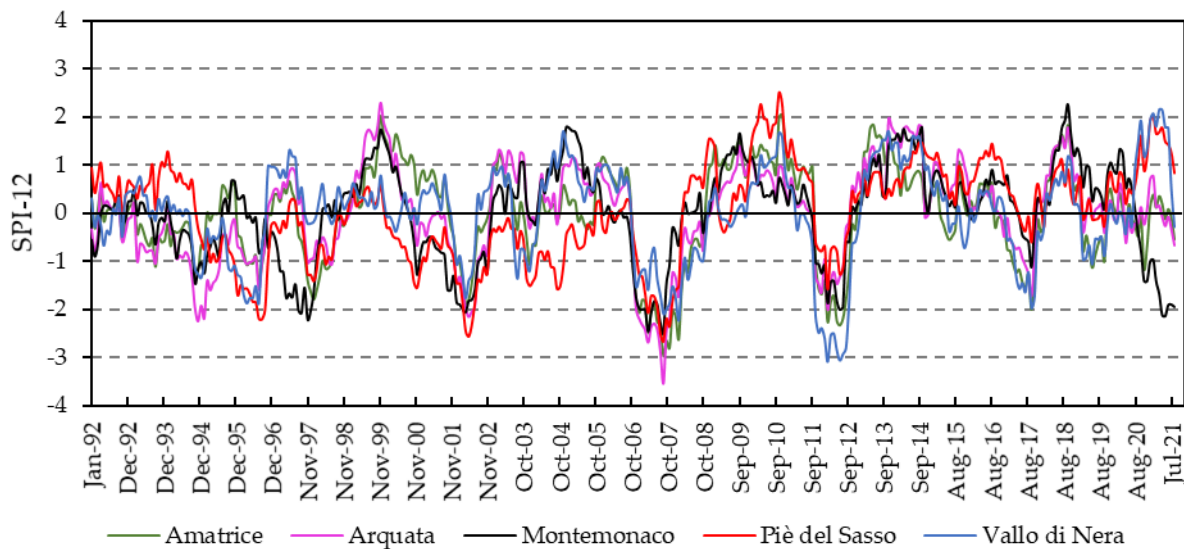


Figure 3.1 SPI-12 obtained for the selected rain gauges representing the different sectors of the Sibillini Mts. massif, having continuous data between 1991 and 2021.

Focusing the SPI-12 values for the Adriatic (Montemonaco) and Tyrrhenian (Vallo di Nera) side (Figure 3.2), it is possible to observe a general correlation between the periods: January 1992 and August 1995 (near normal), June 2004 and February 2009 including a complete oscillation between moderately wet to extremely dry conditions, September 2011 and November 2018 including two dry periods and a very wet period. In all the other intervals, the SPI-12 values show a visible decorrelation much more emphasised in the years (1995-1997 and 2019-2021), in which the SPI-12 for Montemonaco station passes from near normal conditions to extremely dry conditions; Vallo di Nera, on the other hand, is characterised by an opposite trend in the same periods.

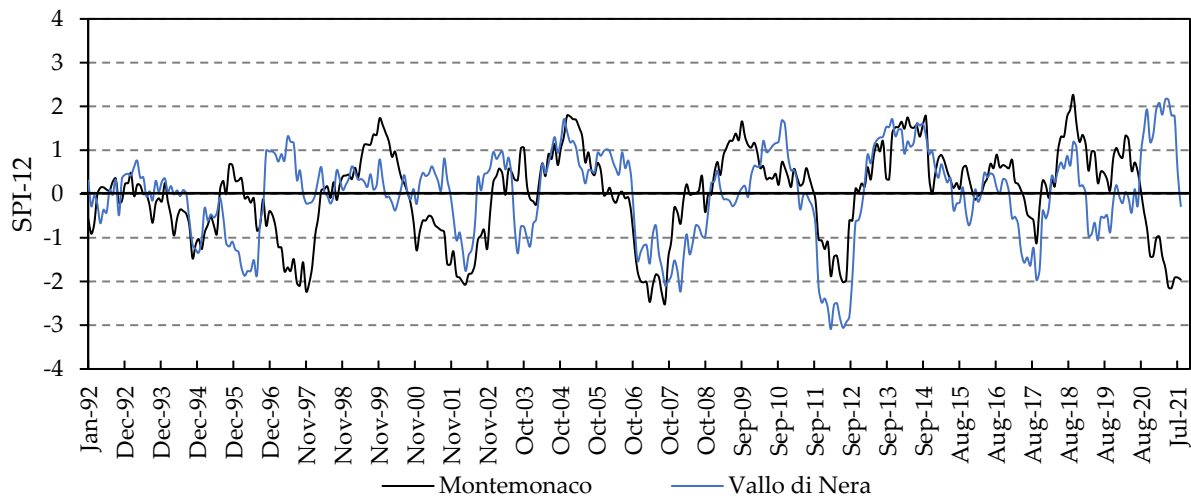


Figure 3.2 SPI-12 obtained for the Montemonaco rain gauge (black line) representing the Adriatic side of the Sibillini Mts. massif, and for the Vallo di Nera rain gauge (blue line) representing the Tyrrhenian side of the Sibillini Mts. massif.

The SPI-12 calculated in the barycentric position of the study area, in this case represented by the Arquata del Tronto rainfall station (Figure 3.3), is characterised by an average condition between the Adriatic and Tyrrhenian side, and according to Valigi et al., 2020 and Fronzi et al., 2020 this is the most appropriate rainfall station to be compared with the springs discharge trends of the Sibillini Mts. especially for those located on the southern side of the massif.

Starting from January 1992 dry conditions with SPI-12 lower than -2, or just slightly higher, lasted particularly long from September 2006 to September 2007. Others extremely dry conditions of less duration have been recorded between October 1994 and February 1995, in May 2002 and March 2012. After the seismic sequence, the SPI-12 varies between -1 and -1.5 (moderately dry conditions) for just a one month (October 2017). The other months are characterised by moderately wet to very wet conditions between May 2018 and September 2018. Starting from October 2018 until September 2021 the SPI-12 varying between 1.0 and -1.0 (near normal conditions).

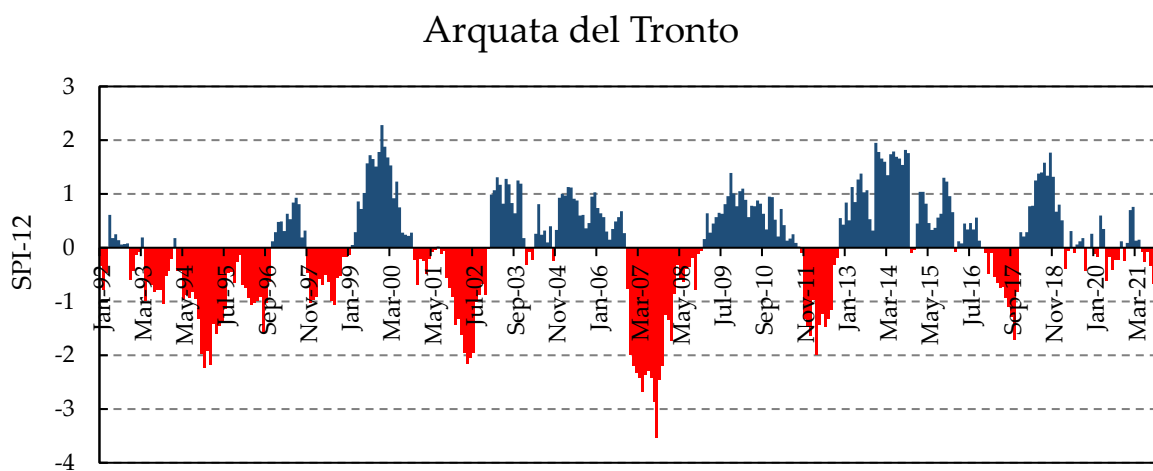


Figure 3.3 SPI-12 obtained for the Arquata del Tronto rain gauge.

The meteo-climatic monitoring system (rainfall and thermometric stations) available in the area permitted to obtain temperature-elevation and rainfall-elevation relations. The best correlation has been obtained by a linear regression between annual data. The regressions have been performed both on 2009-2020 and 2015-2020 temporal intervals.

The temperature gradient calculated for the interval 2009-2020 is equal 4.5°C/100 m (Figure 3.4a) while the data subset (2015-2020) shows a temperature gradient of 4.3 °C/100 m (Figure 3.4b) demonstrating a no significant difference between the thermometric regime calculated after the seismic sequence and the one including the data before the seismic sequence. The strong Pearson's coefficient of determination ($R^2 \geq 0.93$) confirms the good quality of the linear regression for both the time series for the whole area. However, it has to be notice that two elevation gaps not covered from temperature station are located between 918 and 1750 m a.s.l. in which only the Pintura di Bolognola station (Marche Region – 1360 m a.s.l.) is present, as well as a lack of temperature station for the elevation higher than 1917 m a.s.l.

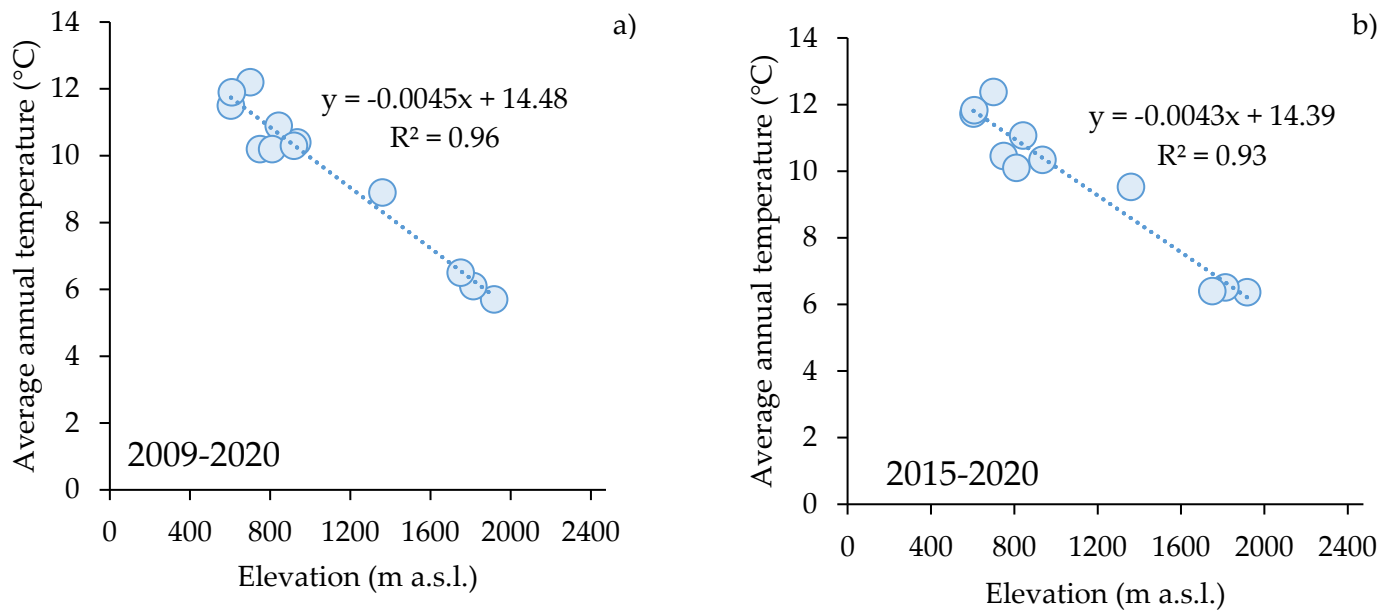


Figure 3.4 Linear regression between mean annual temperature and elevation. (a) Time interval 2009-2020 (b) Time interval 2015-2020.

By applying the linear regression for the mean annual rainfall data (incremented by the SWE – station above 1500 m a.s.l.) in the whole area, both the temporal intervals (2009-2020 and 2015-2020) show a moderate Pearson’s coefficient of determination (Figure 3.5), with no significant differences between the regression calculated after the seismic sequence and the one including the data before the seismic sequence. In this case it has to be notice a lack of data between 995 m a.s.l. and 1750 m a.s.l. and the presence of no rainfall stations over 1917 m a.s.l. as well. The rainfall stations located over 1500 m a.s.l. show a scatter due to a discrete variability of snowfall regime in the area and probably connected to their location influenced by high winds conditions.

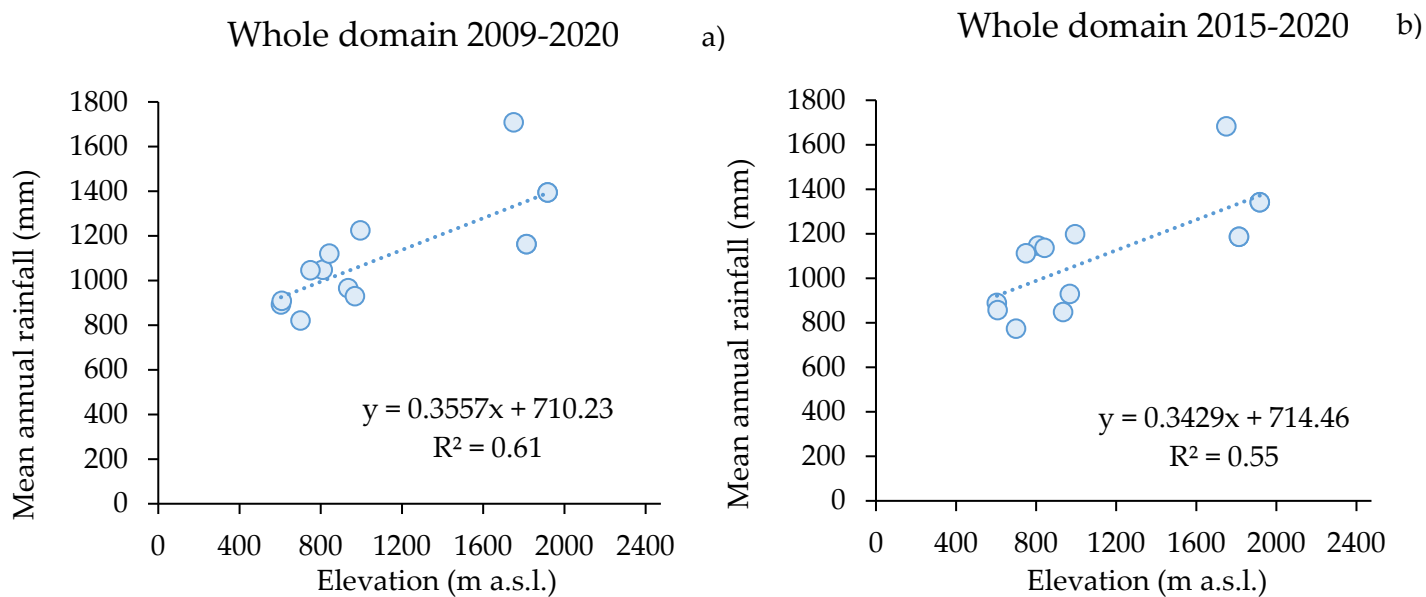


Figure 3.5 Linear regression between mean annual rainfall and elevation for the whole area. (a) Time interval 2009-2020 (b) Time interval 2015-2020.

According to the results obtained by the SPI-12 for the Adriatic and Tyrrhenian sides (Figure 3.2) and the moderate positive coefficients of determination values obtained for the whole area, it has deemed necessary to subdivide the rainfall stations in different sub-domains: the Adriatic side (Figure 3.6), the Northern Tyrrhenian side (Figure 3.7) and the Southern Tyrrhenian side (Figure 3.8). The increasing in the Pearson's coefficient of determinations confirm that the area is characterised by at least three different rainfall-elevation relations.

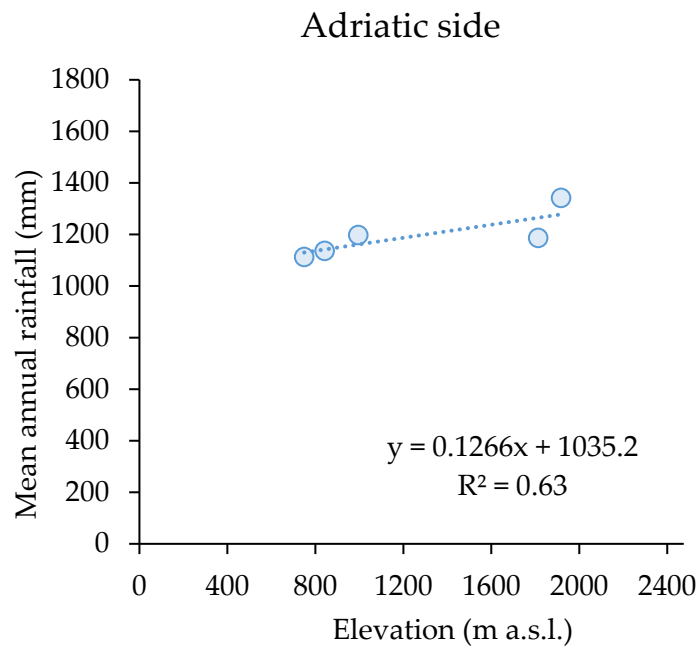


Figure 3.6 Linear regression between mean annual rainfall and elevation for the Adriatic side. Time interval 2015-2020.

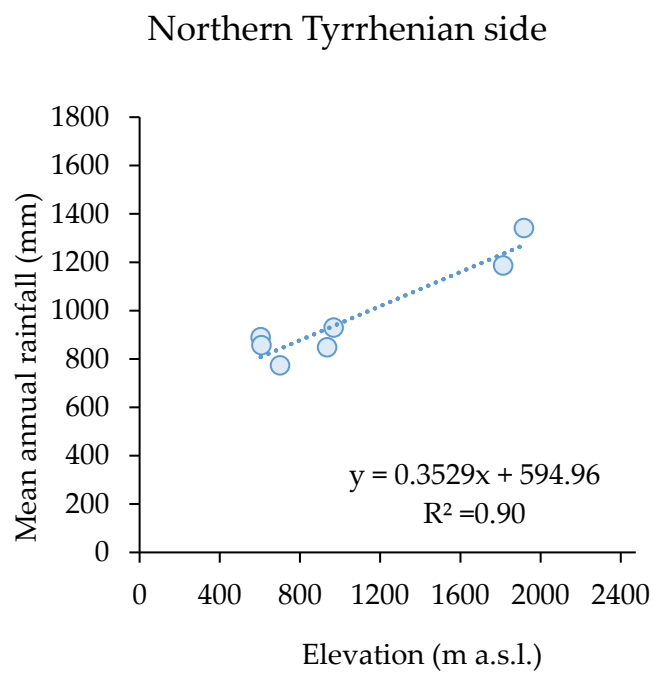


Figure 3.7 Linear regression between mean annual rainfall and elevation for the Northern Tyrrhenian side. Time interval 2015-2020.

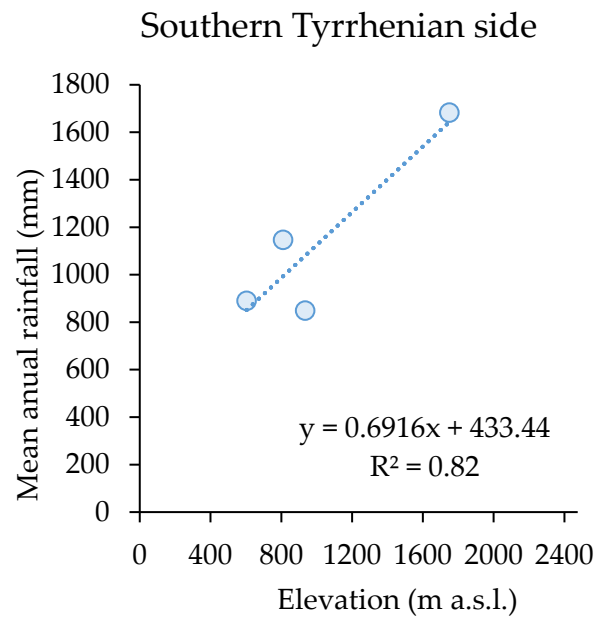


Figure 3.8 Linear regression between mean annual rainfall and elevation for the Southern Tyrrhenian side. Time interval 2015-2020.

The relations have been spatialized in ArcGIS® using the “Rastercalculation ArcToolbox” obtaining the temperature (Figure 3.9) and rainfall (Figure 3.10) distribution maps over the study area, used to calculate the AET by applying the Turc method (Turc, 1961).

The mean annual temperature ranges between 3.7 - 13 °C (Figure 3.9). The hydrostructure 3 is characterised by higher temperature values especially in the northern area while the hydrostructure 1 is characterised by lower temperature values due to its higher elevations. As far the rainfall distribution concerns, their values range between 712 mm and 1452 mm. Figure 3.10 shows the three sub-domain of rainfall ArcGIS® spatialization procedure.

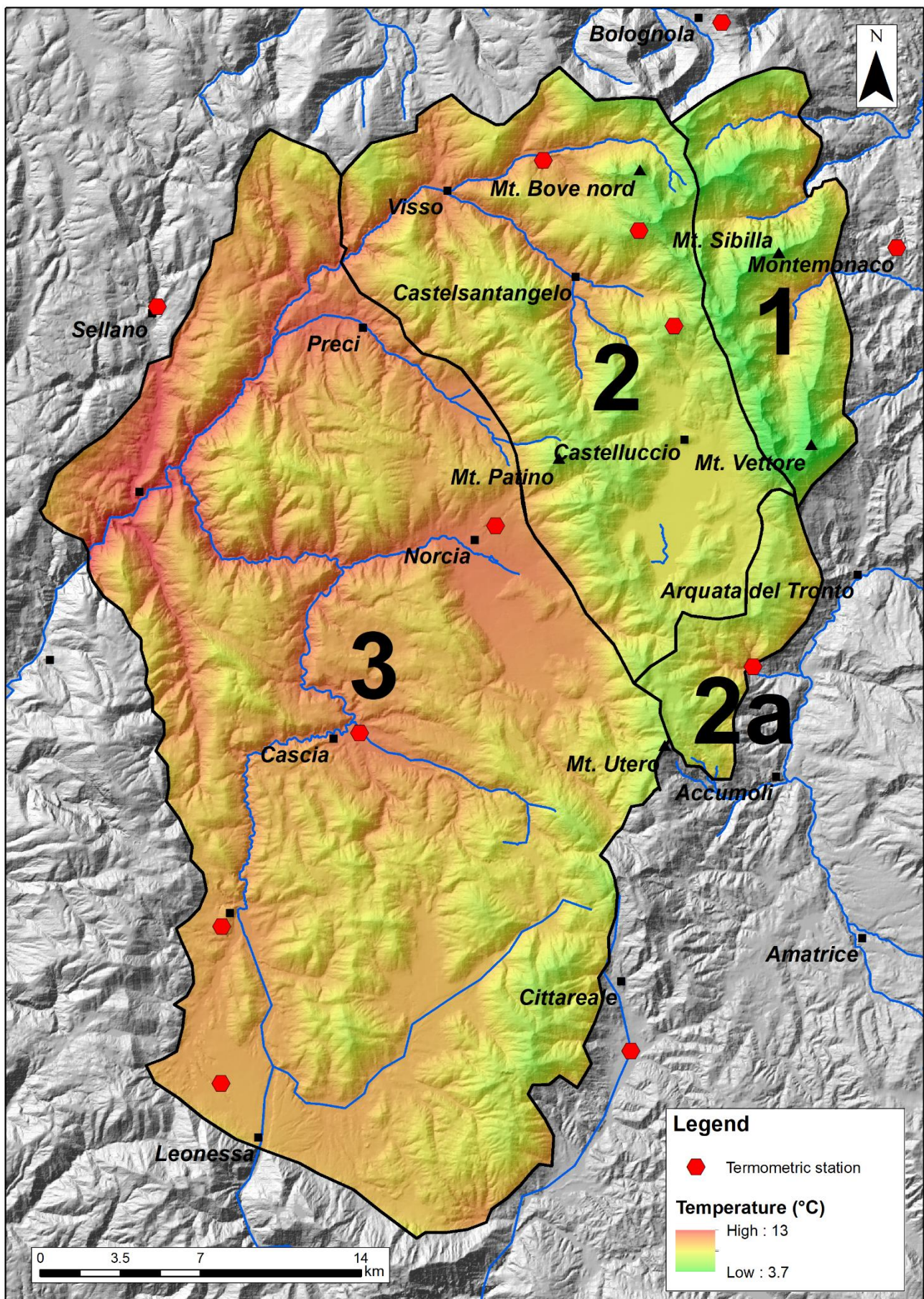


Figure 3.9 Mean annual temperature distribution (2015-2020) over the study area with detail of the hydrostructures (i.e., 1, 2, 2a and 3).

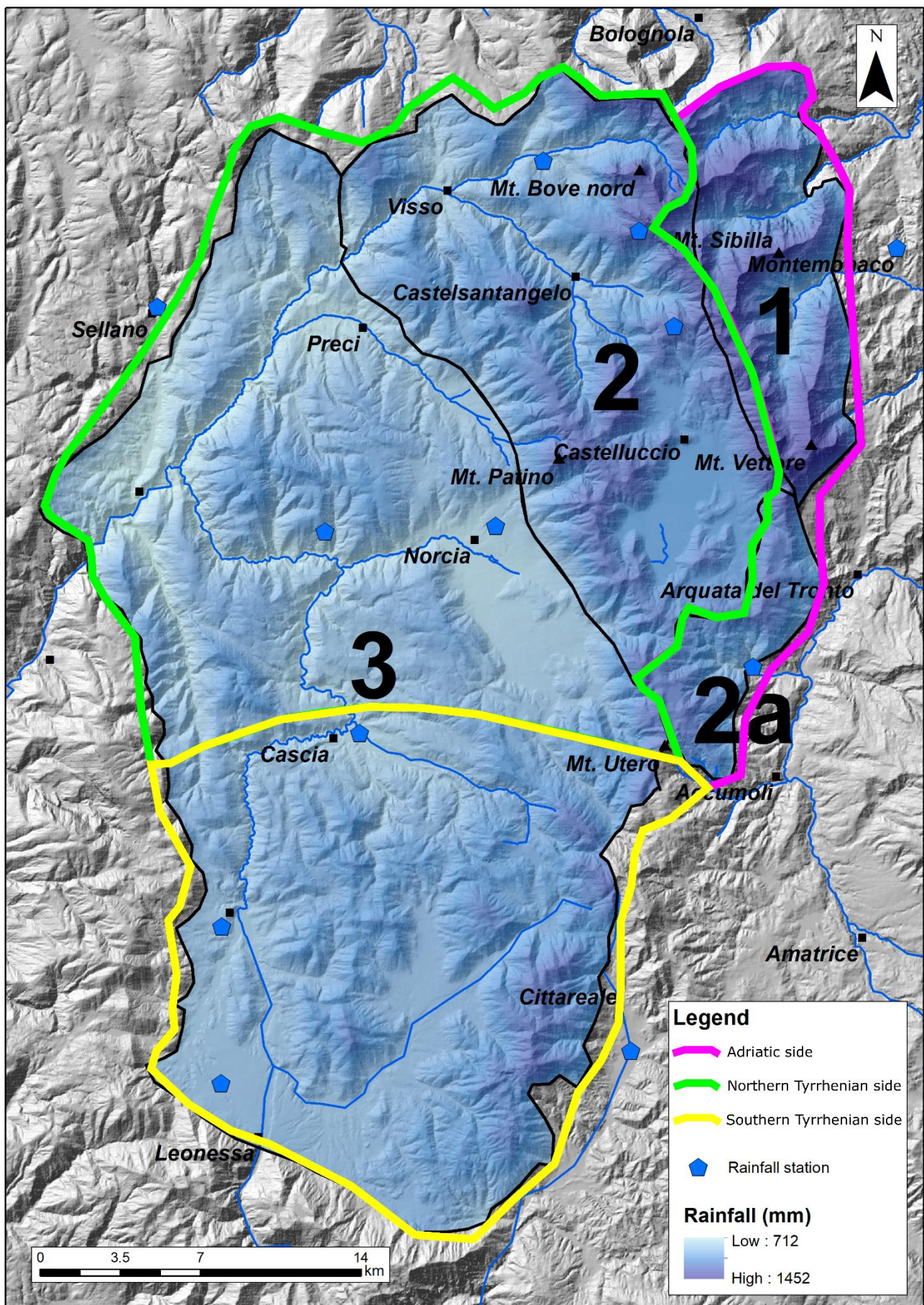


Figure 3.10 Mean annual rainfall distribution (2015-2020) over the study area with detail of the spatialization sides and hydrostructures (i.e., 1, 2, 2a and 3).

The developing of the WaterbalANce Web-app to perform water balance computation by applying the Thornthwaite-Mather method (Mammoliti et al., 2021) permitted to refine the preliminary steps to compute the hydrogeological water balance for the study area. In fact, starting from punctual monthly rainfall and mean monthly temperature data for 12 termo-pluviometric stations (Table 2.1-paragraph 2.1.1), an annual water surplus-elevation relationship (Figure 3.11) has been obtained for the whole area showing a strong positive coefficient of determination ($R^2 > 0.72$).

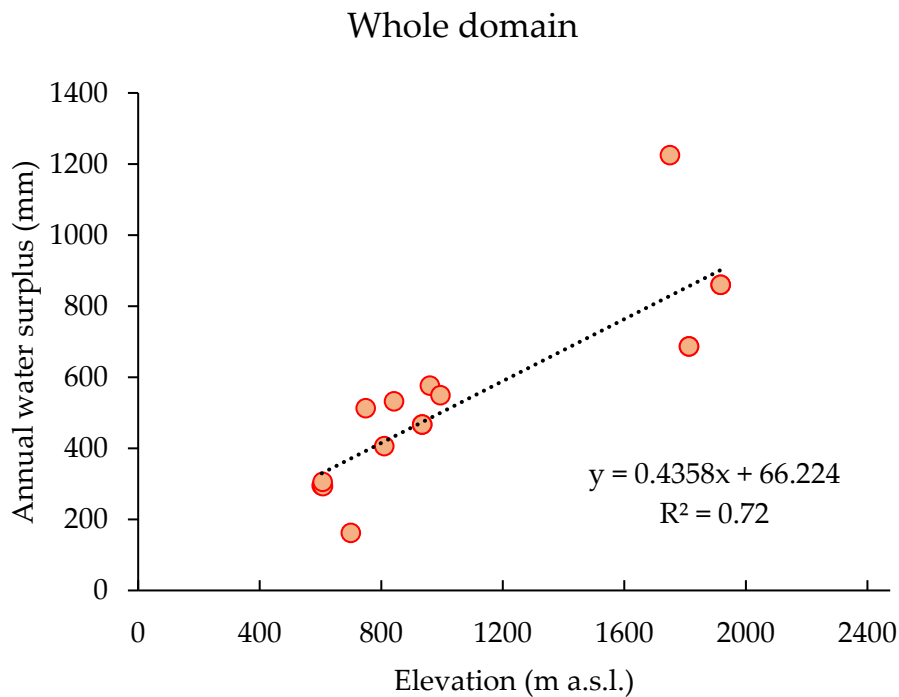


Figure 3.11 Linear regression between mean annual water surplus and elevation for the whole area. Time interval 2015-2020.

However, since the water surplus is strictly related to the rainfall regime over the area, also in this case it has deemed necessary to subdivide the annual water surplus-elevation relationship into three different sectors: the Adriatic side (Figure 3.12), the Northern Tyrrhenian side (Figure 3.13) and the Southern Tyrrhenian side (Figure 3.14). The increasing in the Pearson's coefficient of determinations confirm that the area is characterised by at least three different water surplus-elevation relationships.

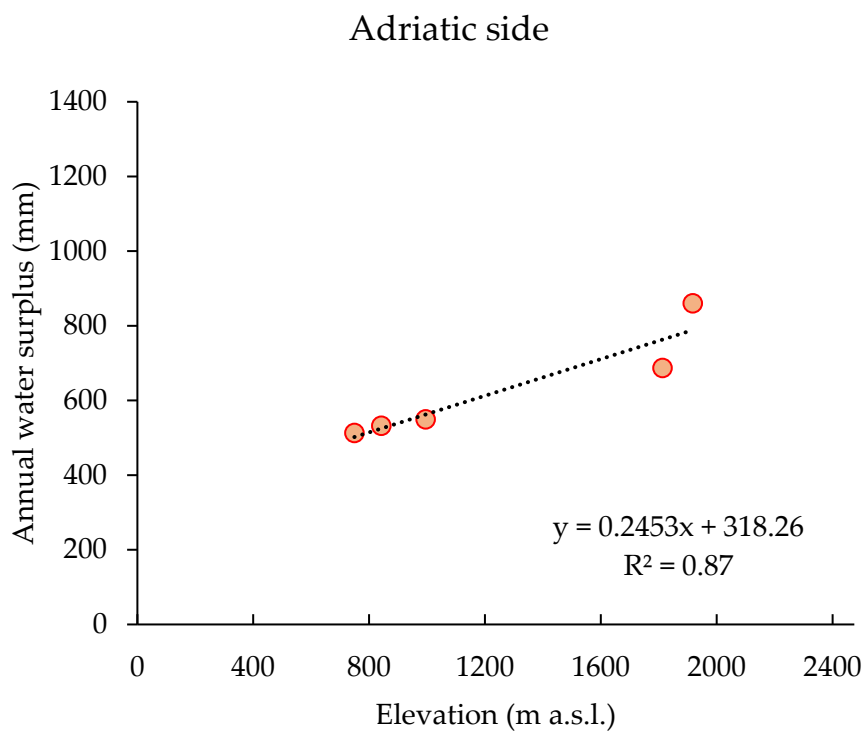


Figure 3.12 Linear regression between mean annual water surplus and elevation for the Adriatic side. Time interval 2015-2020.

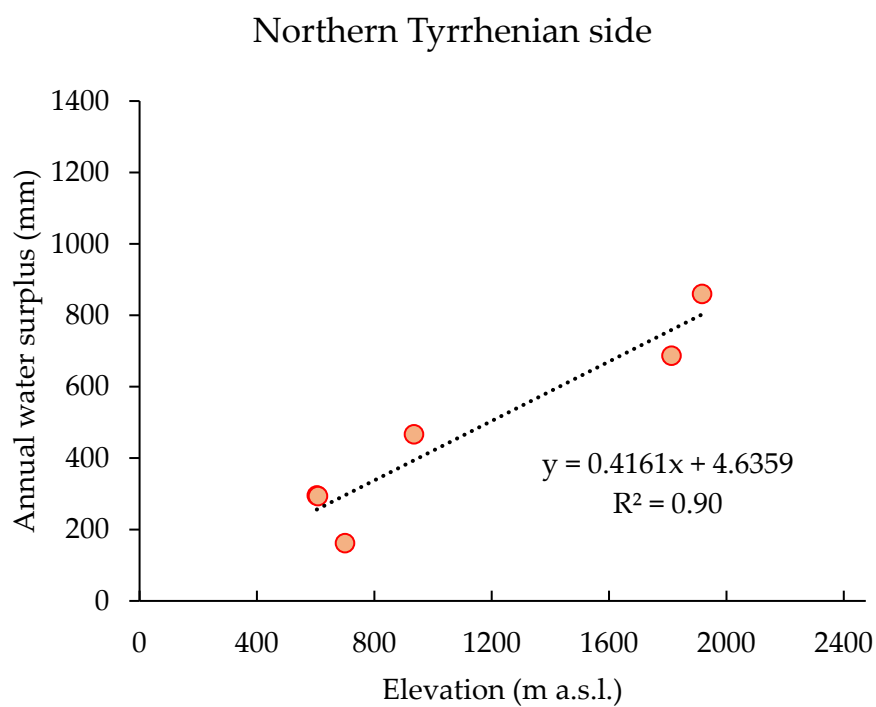


Figure 3.13 Linear regression between mean annual water surplus and elevation for the Northern Tyrrhenian side. Time interval 2015-2020.

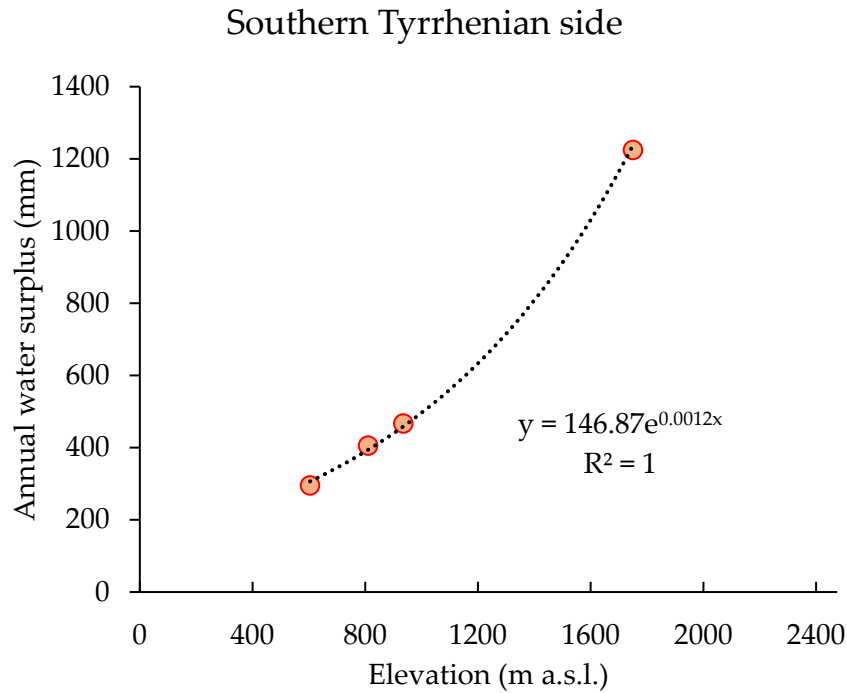


Figure 3.14 Linear regression between mean annual water surplus and elevation for the Southern Tyrrhenian side. Time interval 2015-2020.

The best correlations for existing data have been obtained by linear regression for the Adriatic and the northern Tyrrhenian side, while for the southern Tyrrhenian side the exponential envelop shows a $R^2 = 1$, demonstrating a perfect correlation between the variables. In all the domain it has to be noticed a lack of termo-pluviometric stations between approximately 1000 and 1750 m a.s.l. and over 1920 m a.s.l.

The relations have been spatialized in ArcGIS® using the “Rastercalculation ArcToolbox” obtaining annual water surplus distribution map over the study area, used to perform the “indirect” hydrogeological water balance.

Besides, the comparison between the PET obtained by applying the Thornthwaite-Mather method and the reference evapotranspiration (Eto) obtained by the application of the Hargreaves & Samani method (Hargreaves & Samani, 1985), shows a strong correlation between the variable at monthly scale (Figure 3.15) with a Mean Absolute Error (MAE) of 0.08.

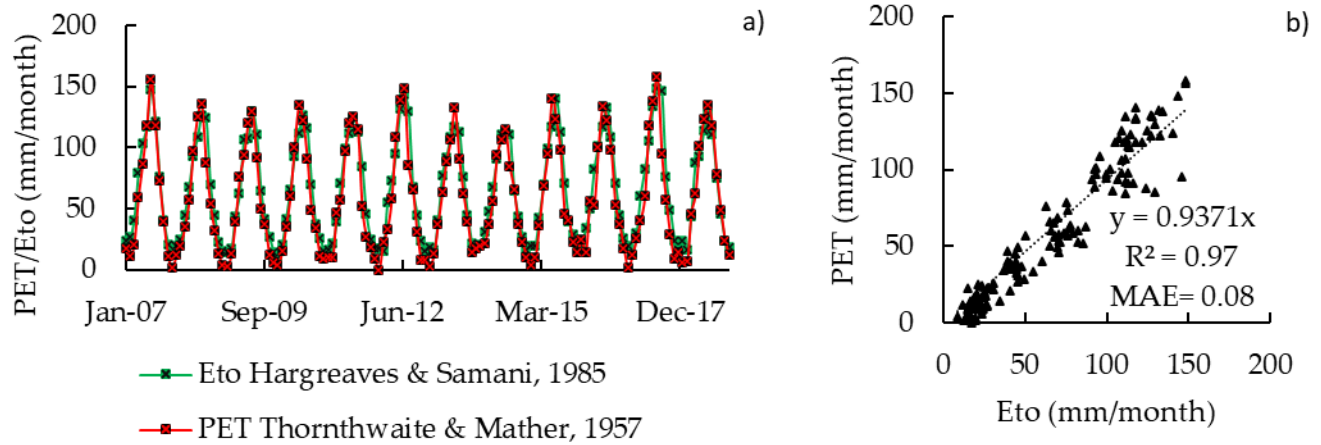


Figure 3.15 (a) Comparison between PET calculated with Thornthwaite-Mather method and Eto calculated with Hargreaves & Samani method over the analysed time span; (b) Correlation between the variables with Pearson determination coefficient and MAE.

3.2 Direct hydrological balance and long-term discharge variation at regional scale

Since a significant rainfall and temperature regime difference between pre and post seismic period was not recorded, the pre seismic discharge data obtained by Mastroiillo et al., 2009; Boni et al., 2010 and Viaroli et al., 2021, permitted to calibrate the potential infiltration coefficient (PIC) for each hydrological complex converging the results of indirect hydrological balance (calculated starting from rainfall and temperature data between 2015-2020) and direct hydrological balance (calculated starting from discharge measurements). Secondary, the post-seismic discharge data have been used to evaluate the groundwater deficit or surplus in each hydrostructure not related to meteoroclimatic conditions, and to quantify groundwater amount modifications at regional scale. A synthesis of the results of hydrological balance is summarised in Table 3.1, while a detailed report of the computation is reported in the Annex B.

Table 3.1 Synthesis of hydrogeological balance for each hydrostructure. Δ (m³/s) represents the discharge difference between pre- and post-seismic periods.

	Hydrogeological complex	Area (km ²)	PICs	Rainfall (mm/year)	EI Turc (mm/year)	Discharge Turc (m ³ /s)	EI Thornthwaite-Mather (mm/year)	Discharge Thornthwaite-Mather (m ³ /s)	Discharge pre-seismic (m ³ /s)	Discharge post-seismic (m ³ /s)	Δ Discharge (m ³ /s) and (%)
Hydrostructure 1	BAS	38	0.93	1245	735	0.89	674	0.81			
	JUR	12	0.61	1222	459	0.17	415	0.16			
	MAI	23	0.82	1241	643	0.47	588	0.43			
	MF	4	0.45	1245	356	0.05	326	0.04			
	SCA	17	0.80	1255	647	0.35	596	0.32			
	Total	94				1.92		1.76	2.01*	1.49*	-0.52 (-25.9%)
Hydrostructure 2	BAS	37.4	0.95	1070	569	0.67	536	0.64			
	JUR	38.3	0.67	1033	372	0.45	521	0.42			
	MAI	69	0.84	1063	496	1.09	556	1.02			
	MF	17.4	0.47	1044	267	0.15	534	0.14			
	SCA	86.4	0.85	1069	508	1.39	563	1.31			
	SVC	17.6	0.45	942	205	0.11	414	0.10			
	DEP	14.5	1.0	1051	577	0.27	542	0.25			
	Total	281				4.13		3.88	3.14*	4.66*	+1.52 (+48.4%)
Hydrostructure 2a	BAS	20.6	0.96	1215	712	0.47	667	0.42			
	JUR	4	0.62	1224	469	0.06	684	0.05			
	MAI	10	0.90	1220	675	0.21	677	0.19			
	MF	1.3	0.5	1209	365	0.02	654	0.01			
	SCA	5	0.81	1202	584	0.09	642	0.25			
	Total	41				0.85		0.76	0.78	0.62	-0.16 (-20.5%)
Hydrostructure 3	BAS	111	1.0	1183	595	2.13	560	2.06			
	JUR	33	0.70	1137	416	0.46	393	0.43			
	MAI	125	0.88	1102	488	2.03	459	1.88			
	MF	38	0.50	1024	260	0.31	244	0.29			
	SCA	285	0.87	1009	466	3.84	437	3.73			
	SVC	73	0.50	882	208	0.41	194	0.42			
	DEP	86	0.90	981	378	1.09	347	0.98			
	Total	753				10.27		9.79	11.21*	14.44	+3.23 (+28.8%)

+ data integrated with mean spring discharge values published in Viaroli et al., 2021; * data integrated with mean spring discharge values published in Valigi, 2021.

A general convergence between the results obtained by indirect and direct hydrological balances is observed during the pre-seismic period for hydrostructure 1 and 2a.

At the same time the hydrostructure 2 is characterised by a water surplus due to effective infiltration of about 1 m³/s with respect to the discharge provided by the same hydrostructure (3.14 m³/s), on the contrary the hydrostructure 3 is characterised by a lack of water due to effective infiltration of about 1 m³/s with respect to the discharge provided by the same hydrostructure (11.21 m³/s). This aspect suggests a groundwater exchange of about 1 m³/s flowing from hydrostructure 2 to hydrostructure 3 during the pre-seismic period. This result is in accordance with the Norcia Plain piezometric map reported by Petitta, 2011 in which two main groundwater directions are visible: the one SE-NW oriented and the one coming from the eastern sector of the plain in which the hydrostructure 2 is located (Figure 1.17).

As evidenced in Table 3.1 the Effective Infiltration calculated with the Turc method at annual scale is always higher than the one calculated by applying the Thornthwaite-Mather method at monthly scale. This difference is much more visible in the discharge values of hydrostructure 3 (0.48 m³/s) characterised by higher areal extension, while it becomes negligible in hydrostructure 2a (0.08 m³/s), characterised by a smaller extension. In general, the discrepancy between the Effective Infiltration calculated by applying the two methods increases as the areal extension increase, evidencing, as expected, that the uncertainty in the application of indirect hydrological balance increases for wider areas, thus suggesting that discharge measurements should always be included in the computation of hydrogeological balance especially in those areas characterised by extremely heterogeneous climatic conditions due to topographic characteristics as well. The effective infiltration values range between 735 and 294 mm/year for the **BAS** and the **SVC** hydrogeological complexes respectively. In fact, the **BAS** is the most permeable complex in the area due to its well-developed karstic features favouring the infiltration, while the **SVC** is the less permeable hydrogeological complex in the area. The effective infiltration ranges between 21 up to 67% of the total precipitation amount occurring in the area.

As the post-seismic condition concerns (after the Mw 6.5 earthquake, 30th October 2016), a discharge decreasing of about 0.5 m³/s and 0.16 m³/s has been recorded in hydrostructures 1 and 2a respectively. At the same time, hydrostructures 2 and 3 show a long-term groundwater discharge increase of about 1.5 m³/s and 3.2 m³/s respectively. In general, a groundwater surplus has been observed in the springs located northward and westward with respect to the study area, while the

springs located eastward and in the southern portion of the Sibillini massif experienced a long-term discharge decrease.

The amount of long-term discharge deficit recorded in hydrostructures 1 and 2a is not equal to the surplus characterising the hydrostructures 2 and 3 for the same period. This aspect may suggest that the seismic effects partially depleted the groundwater reserves.

Following, post-seismic long-term discharge variation in each hydrostructure is highlighted and shown by comparing single spring mean discharge values before and after the main shock.

Hydrostructure 1

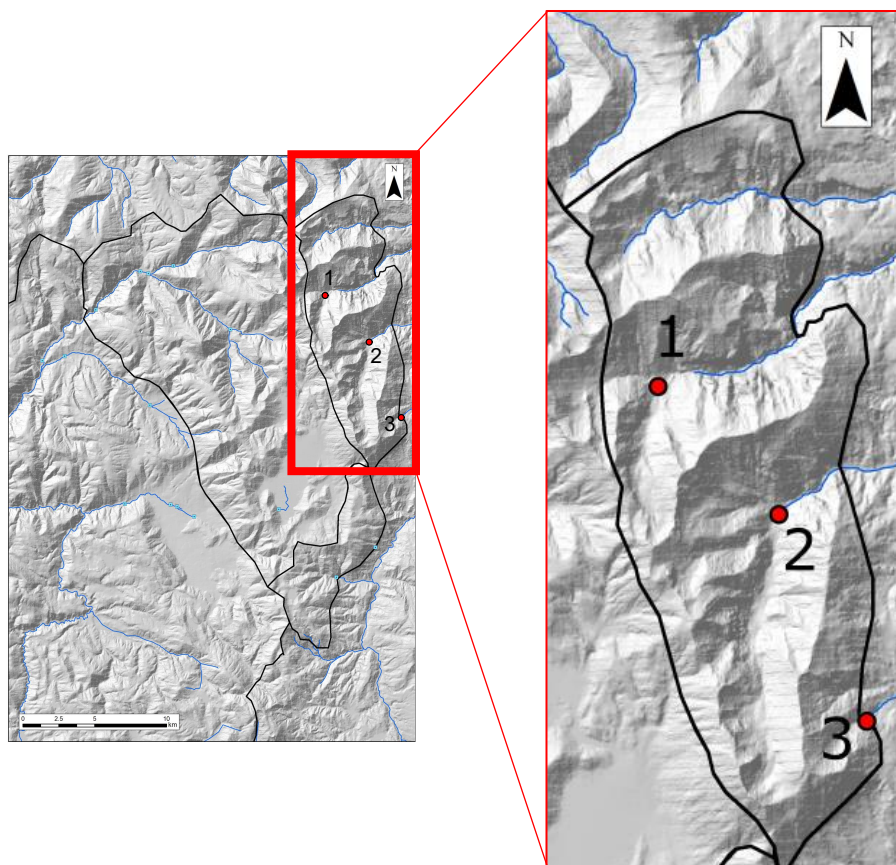


Figure 3.16 Detail of hydrostructure 1 and the discharge monitoring points.

Capotenna spring (point 1) located 1178 m a.s.l. emerging from the **MAI** is characterised by a pre-seismic mean discharge of 0.149 m³/s. The discharge data provided by the local groundwater management company at weekly scale evidence a long-term discharge decreasing of about 0.014 m³/s after the seismic sequence.

Table 3.2 Pre- and post-seismic mean annual discharge measured in point 1.

1	Discharge (m ³ /s)
Pre-seismic mean discharge in point 1	0.149
Post-seismic mean discharge in point 1 November 2016 – December 2019	0.135
Δ (mean) and (%)	-0.014 (-9.4%)

As Foce spring concerns (point 2) located 944 m a.s.l. and emerging from the **BAS**, it is characterised by a pre-seismic mean discharge of 1.05 m³/s. The pre-seismic discharge data have been calculated considering the flow measured inside the drainage tunnel by the local groundwater management company at daily scale and a groundwater contribution emerging just downstream the drainage tunnel not intake for drinking purposes. After the seismic sequence, the groundwater contribution emerging just downstream the drainage tunnel disappeared and the spring shows a long-term discharge decreasing of more than 0.6 m³/s.

Table 3.3 Pre- and post-seismic mean annual discharge measured in point 2.

2	Discharge (m ³ /s)
Pre-seismic mean discharge in point 2	1.05
Post-seismic mean discharge in point 2 November 2016 - December 2020	0.382
Δ (mean) and (%)	-0.667 (-63.3%)

Sassopaccato sping (point 3) emerging from the **BAS** at 1300 m a.s.l. experienced a long-term discharge decreasing of about 60% with respect to the pre-seismic discharge value.

Table 3.4 Pre- and post-seismic mean annual discharge measured in point 3.

3	Discharge (m ³ /s)
Pre-seismic mean discharge in point 3	0.056
Post-seismic mean discharge in point 3 November 2016 – June 2020	0.034
Δ (mean) and (%)	-0.022 (-39.3%)

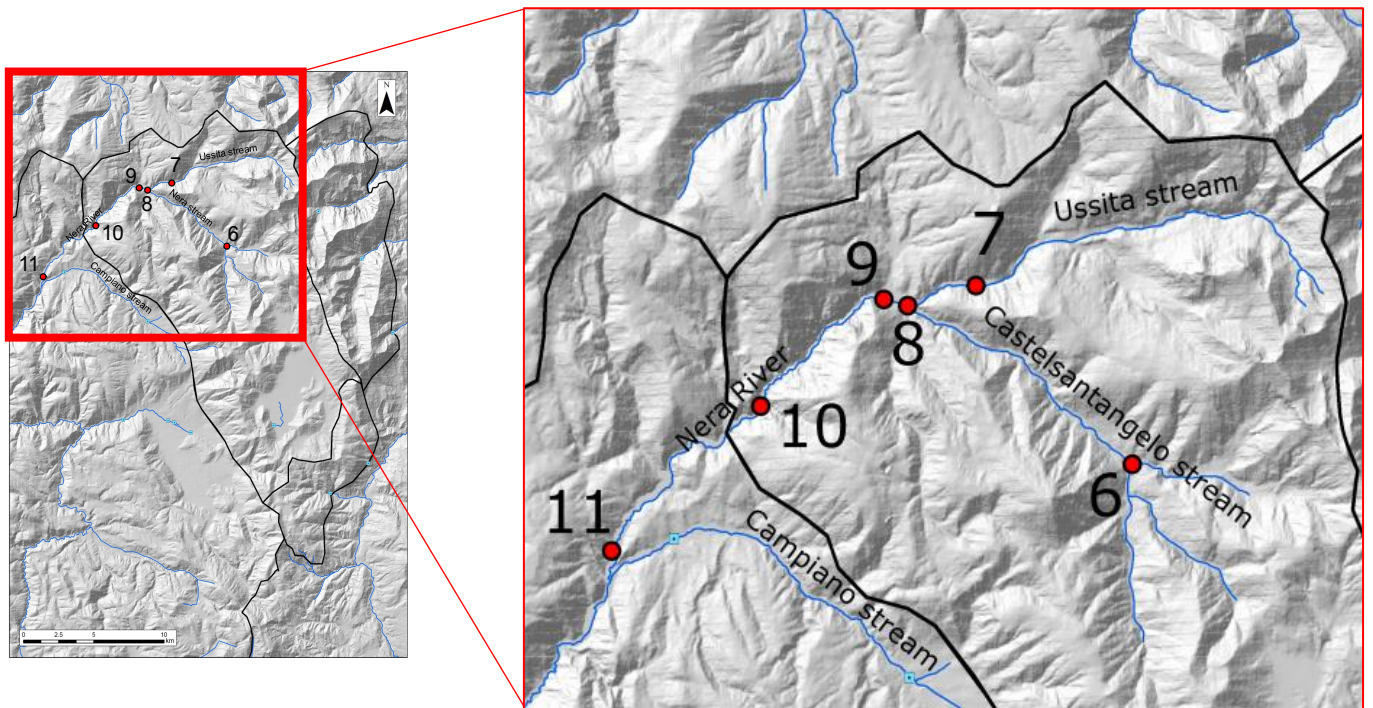


Figure 3.17 Detail of hydrostructure 2 and part of hydrostructure 3 (Upper Nera River basin) with the discharge monitoring points.

Hydrostructure 2 is mainly drained by the punctual and linear springs located in the Upper Nera River basin. The basin extends for about 120 km² and it is divided into two sub-basins, namely Castelsantangelo creek and Ussita creek. The first extends for 80 km² while the latter is 46 km² wide.

The main watercourse flowing in the Castelsantangelo basin is 10 km long while the Ussita basin (Ussita stream) is about 11 km long. Point 6 is in Castelsantangelo basin while point 7 is located in the Ussita stream. The monitoring point 8 is just downstream the confluence of the creeks upstream to the main village of the area named Visso. The Upper Nera River is characterised by the presence of several fish farming and hydropower plants, and it was not possible to effectively determine the water surplus recorded in this area, in particular at point 8, downstream of the aforementioned anthropic activities which may change the spring discharge regime and thus modify the natural hydrographs shape. Downstream to Visso village two linear springs have been recognised, the one located between point 9 and point 10 in which the river drains the **BAS** and the one located between point 10 and 11 in which both the **SCA** is responsible for the discharge increasing along the watercourse.

Nera River upstream to Visso village (points 6, 7 and 8)

Point 6 (Castelsantangelo stream) is characterised by a pre-seismic mean discharge value of about 1.3 m³/s, while during the post-seismic period (until January 2020) an increase in discharge of about 1.23 m³/s has been recorded.

Table 3.5 Pre- and post-seismic mean annual discharge measured in point 6.

6	Discharge (m ³ /s)
Pre-seismic mean discharge in point 6	1.27
Post-seismic mean discharge in point 6 November 2016 - January 2020	2.5
Δ (mean) and (%)	+1.23 (+96.9%)

Point 7 (Ussita stream) is characterised by a post-seismic mean discharge increasing of about 0.3 m³/s.

Table 3.6 Pre- and post-seismic mean annual discharge measured in point 7.

7	Discharge (m ³ /s)
Pre-seismic mean discharge in point 7	1.58
Post-seismic mean discharge in point 7 July 2017 - November 2020	1.87
Δ (mean) and (%)	+0.29 (+18.4%)

Nera River downstream to Visso village (points 9, 10 and 11)

The discharge data of points 9, 10 and 11 permitted to identify and calculate two linear springs along the river (meant as gaining stream). The linear spring located between point 9 and 10 show a discharge increasing of about 0.071 m³/s during the post-seismic period, while the linear spring located between point 10 and point 11 show a discharge increasing of about 0.250 m³/s after the earthquakes.

Table 3.7 Pre- and post-seismic mean annual discharge increase measured between point 9 and point 10.

10-9	Discharge (m³/s)
Pre-seismic mean discharge increasing between point 9 and point 10	0.359
Post-seismic mean discharge increasing between point 9 and point 10 January 2018 – September 2021	0.430
Δ (m ³ /s) and (%)	+0.071 (+19.8%)

Table 3.8 Pre- and post-seismic mean annual discharge increase measured between point 10 and point 11.

11-10	Discharge (m³/s)
Pre-seismic mean discharge increasing between point 10 and point 11	0.710
Post-seismic mean discharge increasing between point 10 and point 11 January 2018 – September 2021	0.960
Δ (m ³ /s) and (%)	+0.250 (+35.2%)

Hydrostructure 2a

This hydrostructure belongs to the hydrostructure 2 (located between the Mt. Vettore-Mt. Bove fault system and Nottoria-Preci fault system). However, a further subdivision has been made to justify the presence of a recharge area for the springs located southward to the Sibillini massif (points 4 and 5), whose groundwater flowpaths are oriented in the opposite direction to the regional basal flow that characterise hydrostructure 2.

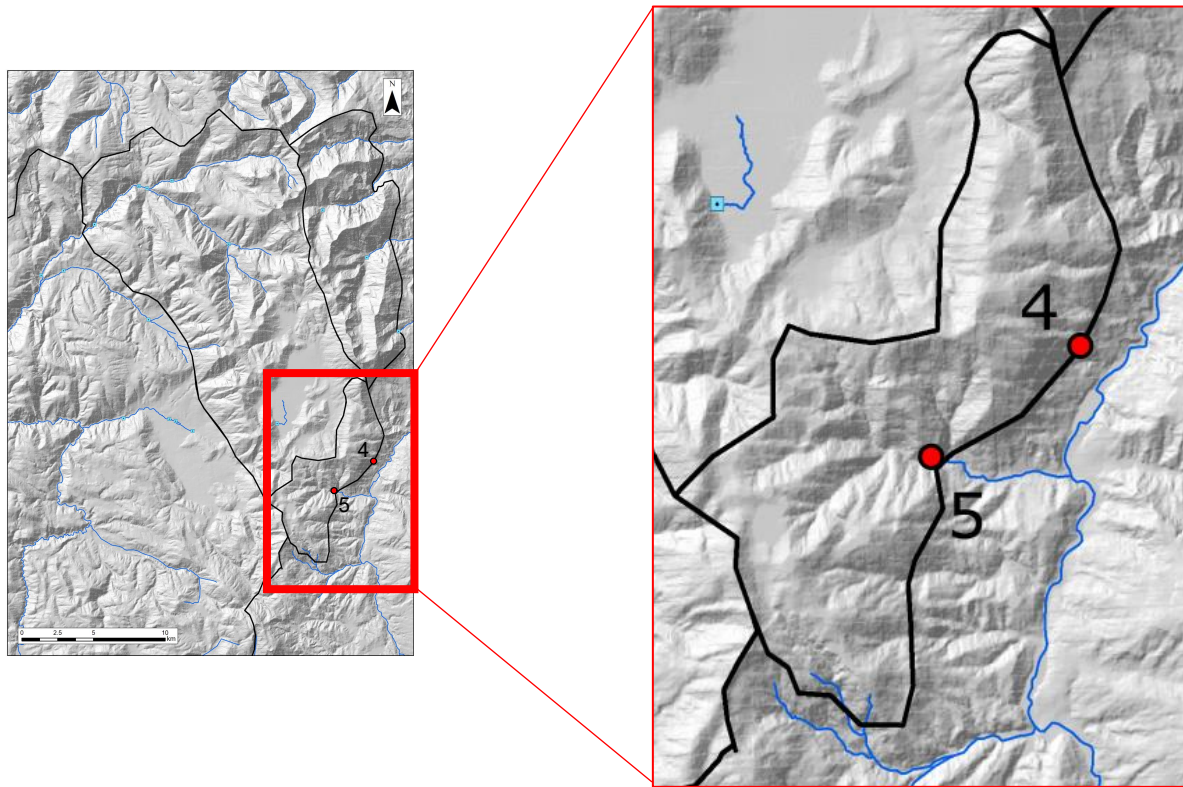


Figure 3.18 Detail of hydrostructure 2a and the discharge monitoring points.

Pescara del Tronto spring (point 4) located 920 m a.s.l. emerging from the SCA is characterised by a pre-seismic mean discharge of 0.245 m³/s. The discharge data provided by the local groundwater management company at daily scale evidence a long-term discharge decreasing of about 0.1 m³/s after the seismic sequence.

Table 3.9 Pre- and post-seismic mean annual discharge measured in point 4.

4	Discharge (m ³ /s)
Pre-seismic mean discharge in point 4	0.245
Post-seismic mean discharge in point 4 November 2016 - July 2020	0.146
Δ (mean) and (%)	-0.99 (-40.4%)

Capodacqua spring (point 5) located 841 m a.s.l. emerging from the **BAS** is characterised by a pre-seismic mean discharge of 0.426 m³/s. The discharge data provided by the local groundwater management company at daily scale evidence a long-term discharge decreasing of about 0.063 m³/s after the seismic sequence.

Table 3.10 Pre- and post-seismic mean annual discharge measured in point 5.

5	Discharge (m ³ /s)
Pre-seismic mean discharge in point 5	0.426
Post-seismic mean discharge in point 5 November 2016 - December 2020	0.363
Δ (mean) and (%)	-0.063 (-14.8%)

Hydrostructure 3

Hydrostructure 3, located westward with respect to the Nottoria-Preci fault system is characterised by the presence of the Campiano stream and the Sordo-Corno River Basin.

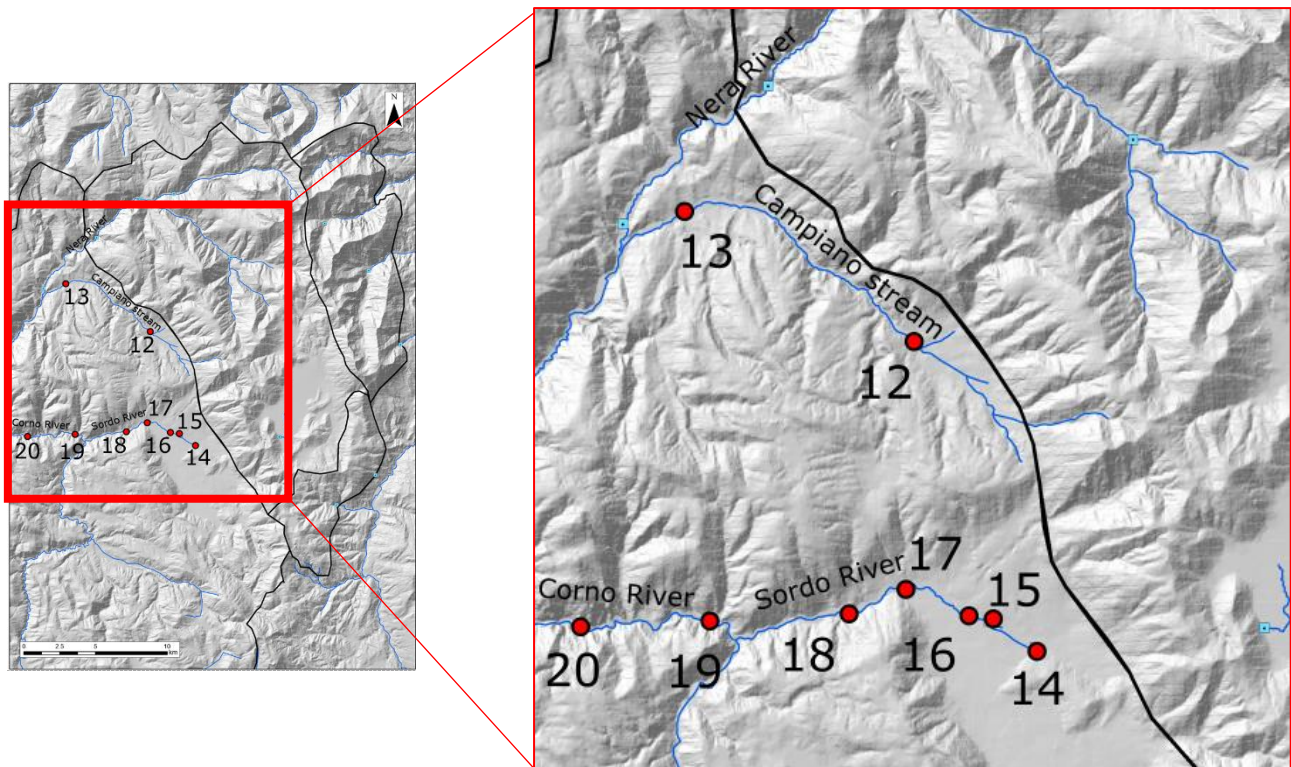


Figure 3.19 Detail of hydrostructure 3 and the discharge monitoring points along the Campiano stream, the Sordo and Corno Upper Basins.

Campiano stream (points 12 and 13)

The discharge data of points 12 and 13 permitted to identify and calculate a discharge increasing along the stream (linear spring). The linear spring shows a slight discharge increasing of just 0.01 m³/s during the post-seismic period. However, it must be noticed that the discharge data for the monitoring points are both available only starting from February 2018, hence a likely discharge increase in the immediate post-seismic condition was not measured.

Table 3.11 Pre- and post-seismic mean annual discharge increase measured between point 12 and point 13.

13-12	Discharge (m ³ /s)
Pre-seismic mean discharge increasing between point 12 and point 13	0.37
Post-seismic mean discharge increasing between point 12 and point 13 February 2018 - July 2021	0.38
Δ (mean) and (%)	+0.01 (+2.7%)

Nonetheless, point 12 show a steady decrease in discharge values following the trend of point 14 since February 2017. This aspect permit to hypotize that point 12 (as well as 14) was characterised by a significant increase in discharge in the immediate post-seismic period (after October 30th 2016) and that the effects have eased at least after the recharge period of the year 2018 (Figure 3.20).

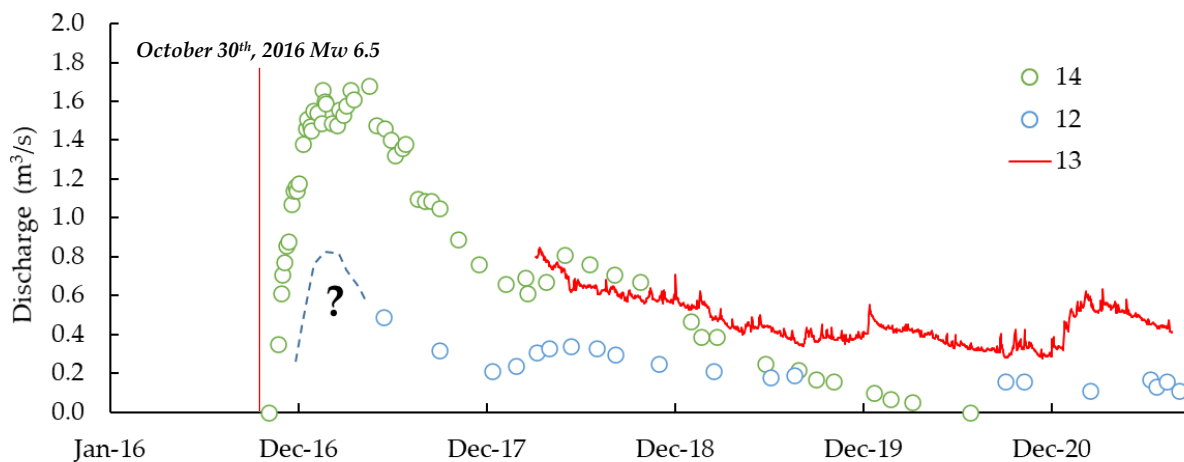


Figure 3.20 Discharge data of points 12 (blue dots), 13 (red line - continuous data), 14 (green dots) after the 30th October 2016, earthquake. The dashed blue line represents the possible discharge trend recorded in point 12 just after the seismic sequence, having a similar behaviour with respect to the point 14.

Sordo River (points 14, 15, 16, 17, 18)

Sordo River Basin is one of the areas that has undergone the greatest changes from a hydrogeological point of view due to the seismic sequence, and those effects are still visible.

Point 14, namely Torbidone spring, appears soon after the October 30th earthquake and its discharge passes from 0 m³/s to about 1.6 m³/s in less than one month. The Torbidone spring post-seismic mean discharge is about 1.04 m³/s, and the post seismic effect have been recorded until July 2020 in which the spring disappear.

Table 3.12 Pre- and post-seismic mean annual discharge measured in point 14.

14	Discharge (m ³ /s)
Pre-seismic mean discharge in point 14	0
Post-seismic mean discharge in point 14 November 2016 - July 2020	1.04
Post-seismic maximum discharge in point 14 November 2016 - July 2020	1.68
Δ (mean) and (%)	+1.04 (+61.5%)

Point 15, is characterised by a pre-seismic mean discharge data of 0.01 m³/s while during the post-seismic period a mean increase of 0.26 m³/s was recorded.

Table 3.13 Pre- and post-seismic mean annual discharge measured in point 15.

15	Discharge (m ³ /s)
Pre-seismic mean discharge in point 15	0.01
Post-seismic mean discharge in point 15 November 2016 – February 2021	0.27
Post-seismic maximum discharge in point 15 November 2016 - February 2021	0.39
Δ (mean) and (%)	+0.26 (+2600%)

The post-seismic groundwater modification affecting point 15 are still visible. The exponential fitting of the baseflow recession curve for the years 2019-2020, obtained by applying the Maillet equation

($R^2 > 0.96$) suggests that the post-seismic discharge increase will return to pre-seismic values (excluding peculiar meteoric recharge conditions) not earlier than the years 2023-2024 (Figure 3.21).

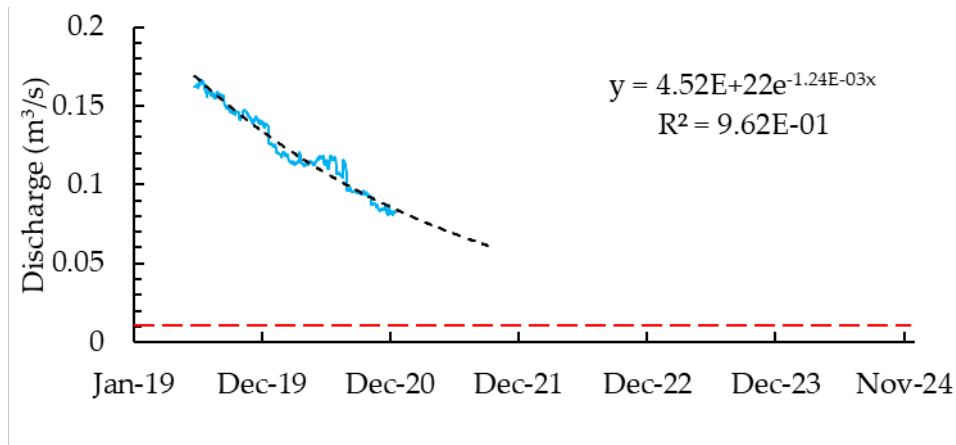


Figure 3.21 Exponential fitting (black dashed line) of the baseflow recession curve (blue line) for point 15 (years 2019-2020). The red dashed line represents the pre-seismic mean annual spring discharge.

The discharge data of points 16, 17 and 18 permitted to identify and calculate two linear springs along the river (meant as gaining stream). The linear spring located between point 16 and 17 show a discharge increasing of about 0.035 m³/s during the whole post-seismic period, while the linear spring located between point 17 and point 18 show a discharge increasing of about 0.18 m³/s after the earthquakes.

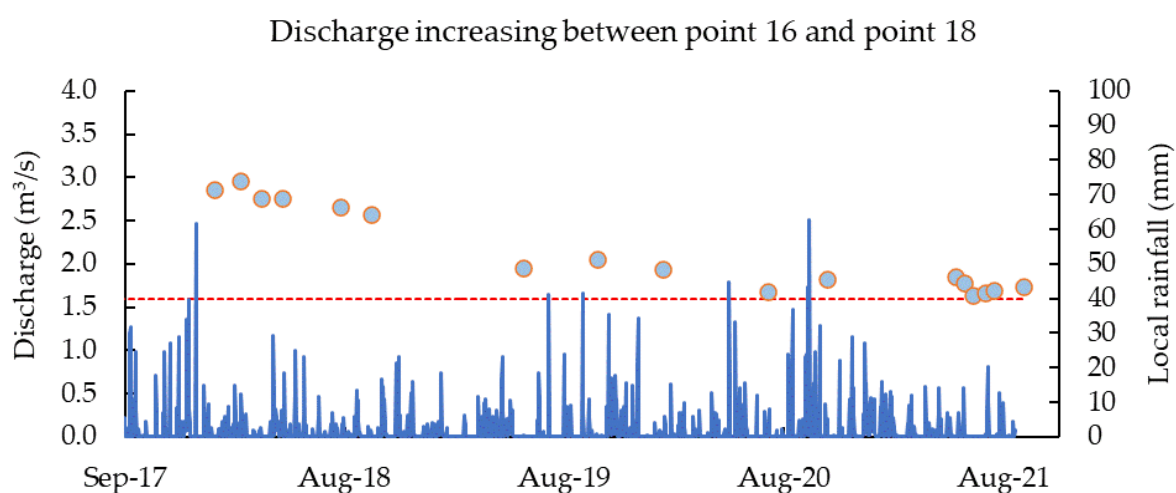
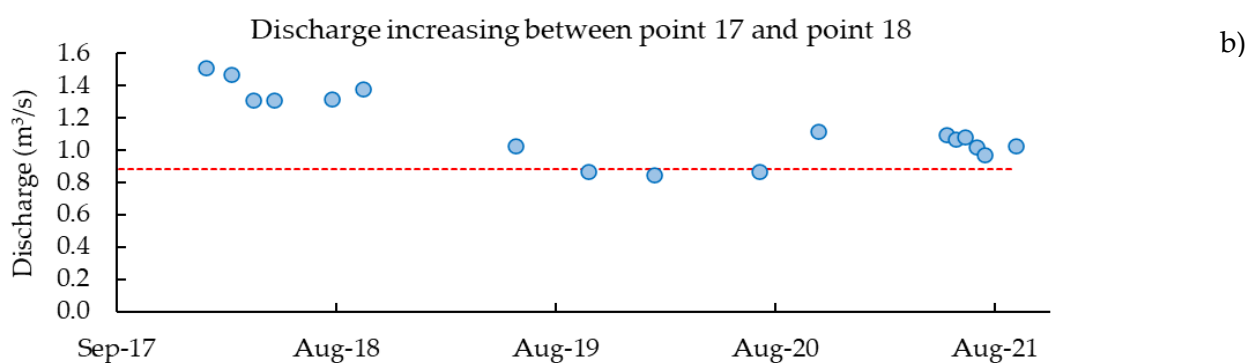
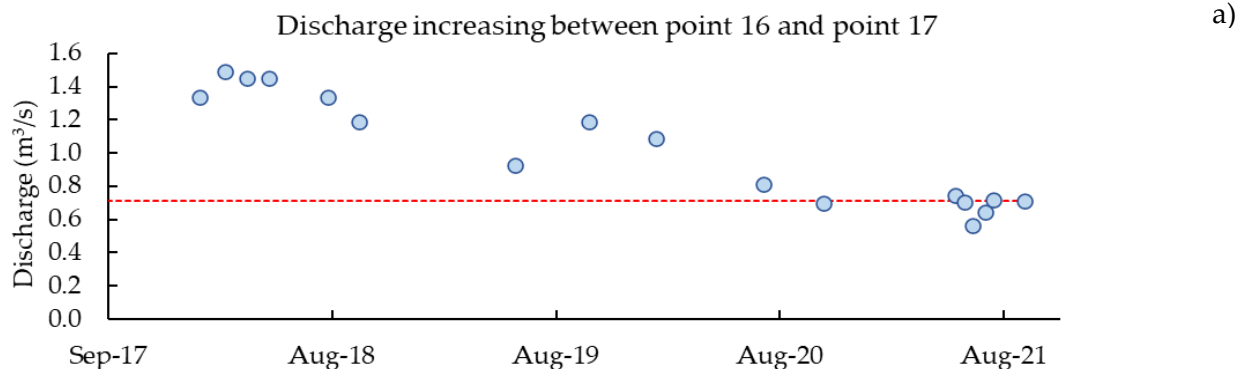


Figure 3.22 (a) Discharge increase measured between point 16 and point 17; (b) discharge increase measured between point 17 and point 18; (c) overall discharge increase measured between point 16 and point 18 vs local daily rainfall. The red dashed lines represent the pre-seismic mean annual discharge increase between the monitoring points.

The discharge data of points 16, 17 and 18 permitted to calculate the post-seismic discharge increasing in different periods.

Table 3.14 shows the discharge increases recorded in the linear springs both in the pre-seismic period [period 1], in the period between January 2018 and June 2019 [period 2], in the whole post-seismic observation period January 2018 - September 2021 [period 3] and during the last year of observation [period 4]. The discharge increase recorded between points 17 and 18 is greater than the one observed between 16 and 17 for the whole post-seismic period. The linear spring located between point 17 and 18 shows a surplus of +1.053 m³/s (January 2018 - June 2019) while it is about half (+0.548 m³/s) in the period January 2018 - September 2021 and only +0.163 m³/s in the period January 2020 - September 2021. Despite the uncertainty in the measurements and the availability of no continuous data in those points, it is clear that a surplus of about 0.160 m³/s in the Sordo River in the last year (January 2020 - September 2021) is still visible. The discharge increase trend between points 16 and 18 shows a slight correlation with the meteoric recharge events. However, a continuous decreasing in discharge values is recorded starting from January 2018.

Table 3.14 Synthesis of discharge increase recorded along the Sordo river between point 16 and 17, between point 17 and 18, and the overall discharge increase between point 16 and 18 referred to four different periods.

18-17-16	Discharge increasing between points 16-17 (m ³ /s)	Discharge increasing between points 17-18 (m ³ /s)	Discharge increasing between points 16-18 (m ³ /s)
Pre-seismic mean discharge (m ³ /s) 2010-2011 [period1]	0.709	0.885	1.594
Post-seismic mean discharge (m ³ /s) January 2018 - June 2019 [period2]	1.313	1.333	2.646
Post-seismic mean discharge (m ³ /s) January 2018 - September 2021 [period3]	1.004	1.136	2.140
Post-seismic mean discharge (m ³ /s) January 2020 - September 2021 [period4]	0.744	1.013	1.757
Δ (mean) [period1- period 2] and (%)	+0.604 (+85.2%)	+0.449 (+50.6%)	+1.053 (+66%)
Δ (mean) [period1- period 3] and (%)	+0.296 (+41.6%)	+0.252 (+28.4%)	+0.548 (+34.3%)
Δ (mean) [period1- period 4] and (%)	+0.035 (+4.9%)	+0.18 (+14.5%)	+0.163 (+10.2%)
Post-seismic maximum discharge (m ³ /s)	1.490	1.513	2.960

Corno River (points 19 and 20)

Between point 19 and 20 an overall discharge surplus of 1 m³/s have been recorded during the post seismic period.

Table 3.15 Pre- and post-seismic mean annual discharge increase measured between point 19 and point 20.

20-19	Discharge (m ³ /s)
Pre-seismic mean discharge increasing between point 19 and point 20	3.5
Post-seismic mean discharge increasing between point 19 and point 20 January 2018 - July 2020	4.5
Δ (m ³ /s) and (%)	+1.0 (+28.6%)

The quantification of water surplus and deficit in linear and punctual springs after the seismic sequence, together with direct field observations, permitted to map the main post-seismic modification at regional scale (Figure 3.23). A decrease in discharge has been observed in the springs located south-eastern and eastern with respect to the study area and persists nowadays as a discharge deficit in hydrostructure 1 and 2a especially in Foce spring (point 2). At the same time, an increase in the discharge values has been observed in the north-western and western sectors of the domain (i.e., Nera River Basin, Campiano stream, Sordo and Corno River Basin). The post-seismic effects, exhausted in the Nera Basin since 2018, as evidenced by Mastrorillo et al. 2020, are still visible as a discharge surplus of about 0.3 m³/s in hydrostructure 3 considering the contribution of point 15 and the linear springs located between points 16 and 18. The field work conducted after the seismic sequence until 2021 highlights a drying up of the springs located over 850 m a.s.l. especially those located in the Tyrrhenian side.

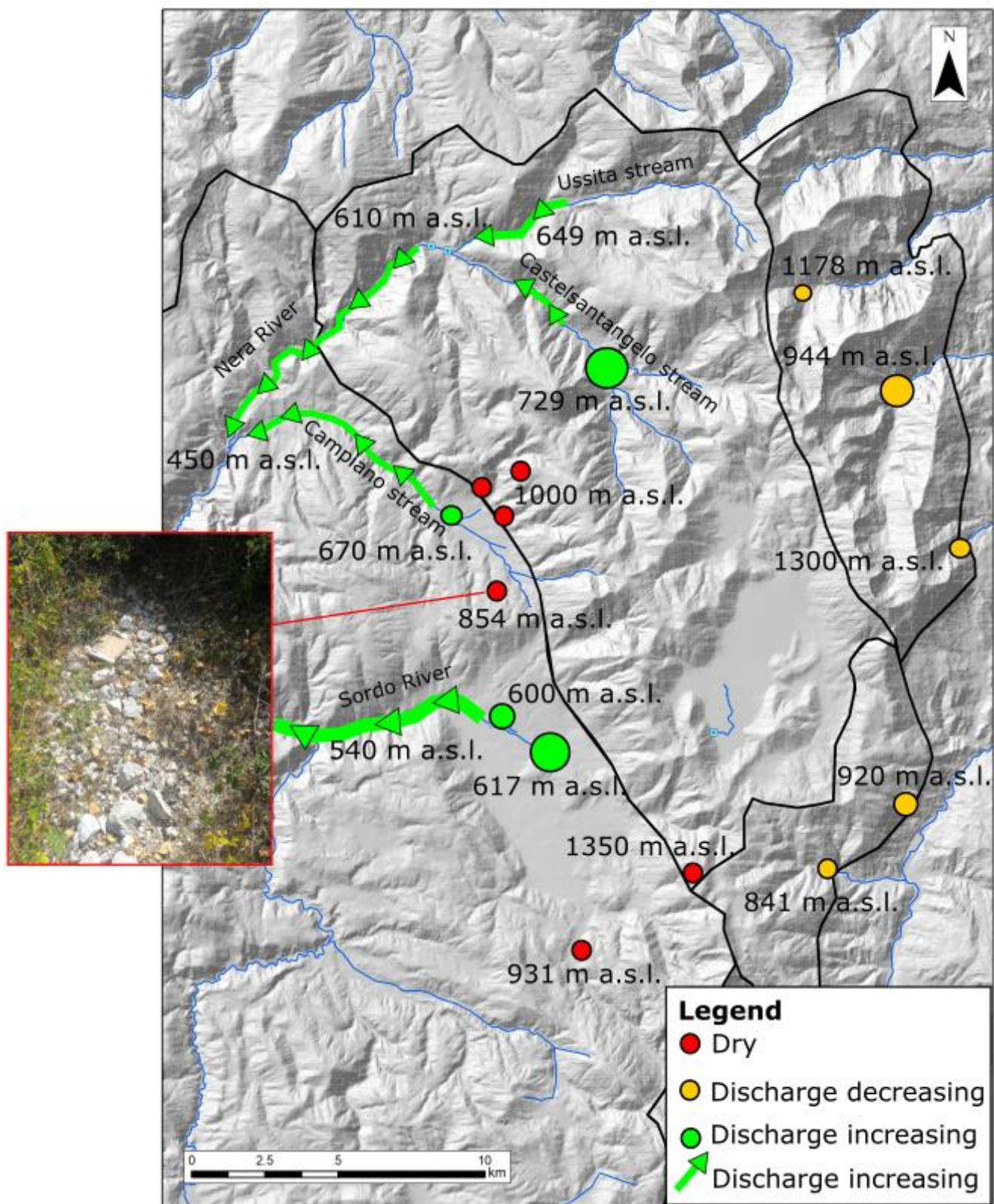


Figure 3.23 On the right schematic map of discharge modifications recorded within each hydrostructure. On the left a picture of a dry spring located 854 m a.s.l. within hydrostructure 3. The dots refer to punctual springs (red: dry, yellow: discharge decreasing, green: discharge increasing), while the green lines represent the increasing of discharge due to the seismic sequence recorded along the linear springs.

3.3 Spring discharge analysis

Basic statistics has been calculated for those springs characterized by continuous discharge data both during the pre- co- and post-seismic periods (Figure 3.24a) to discern long- and short-term significant changes caused by the series of strong seismic events. Moreover, the SPI-12 values of Arquata del Tronto rainfall station permitted to compare the discharge behaviour with local meteorological conditions (Figure 3.24b).

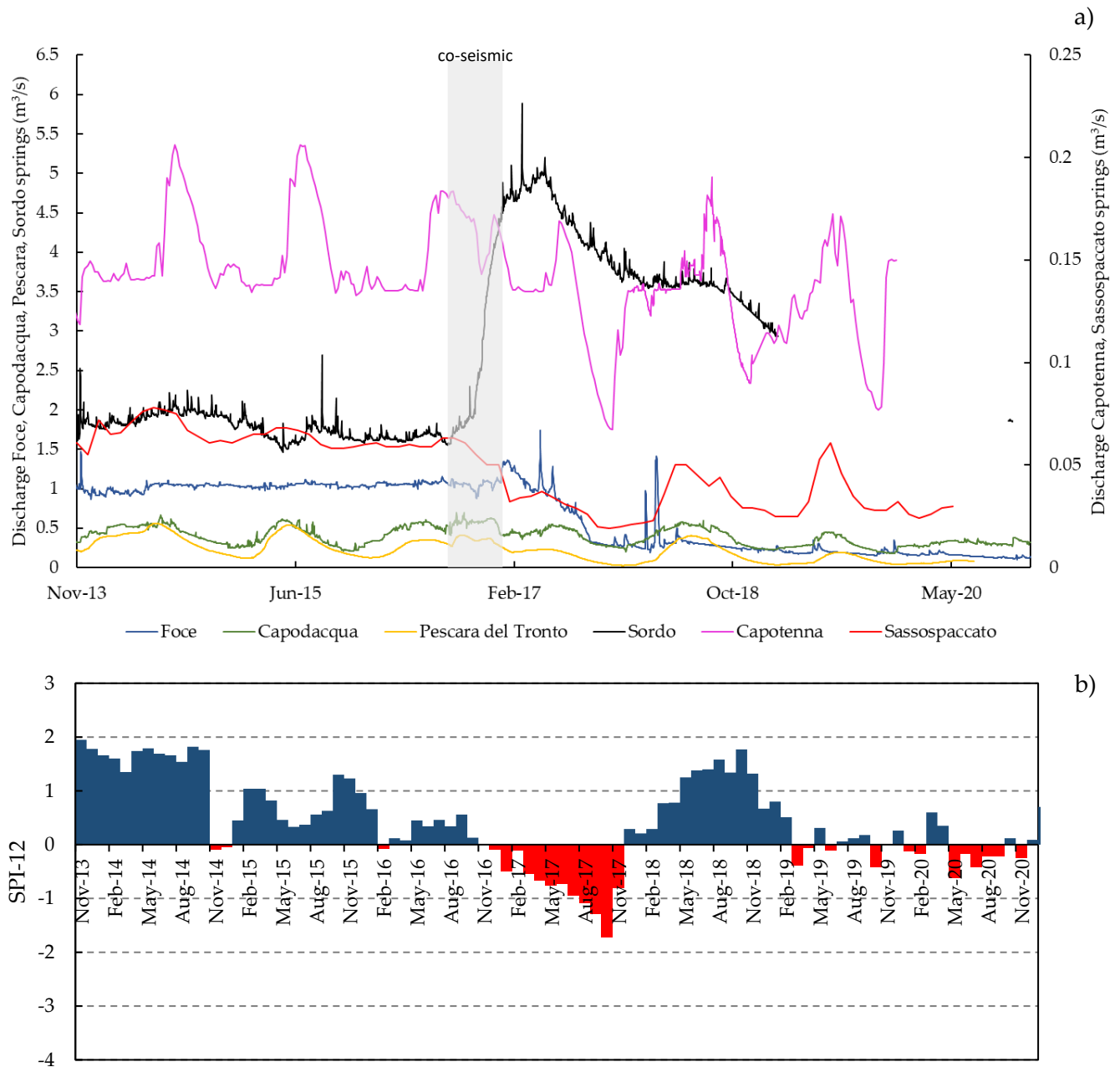


Figure 3.24 (a) Continuous discharge data (m³/s) used to investigate any significant hydrogeological modification both in the short- and long-term periods. The grey window represents the co-seismic period (24th August 2016, 18th January 2017). (b) SPI-12 of Arquata del Tronto rain gauge between November 2013 and December 2020.

As the pre-seismic period concerns, Foce spring is characterised by a very limited seasonal fluctuations (i.e., in the order of tens of litres) to nihil. Capotenna and Sassospaccato springs show a light correlation with seasonal meteoric recharge. The Sordo discharge values are affected by a small run-off component ($\approx 10\%$) since it is monitored along the river, a slight correlation with meteoric recharge is visible in its baseflow. In particular it is possible to identify three periods of baseflow increasing: the year 2014, June 2015 and May 2016, following the moderately wet to extremely wet period according to the SPI-12. Pescara del Tronto and Capodacqua springs are characterised by similar behaviour marked by a strong seasonal variability, as demonstrated by the differences between minimum and maximum values and between 25th and 75th percentiles. The actual nature of the baseflow feeding the different springs accounts for this evidence: Pescara del Tronto, Capodacqua, Capotenna and Sassospaccato springs behaviour depends mostly on the seasonal recharge, while groundwater discharging in Foce and Sordo springs is related to a longer and deeper circulation.

During the co-seismic period, all the analysed springs showed an unusual increase of the discharge value, not linked to clear meteo-climatic changes except Sassospaccato spring for which monthly data cannot permit to clearly identify short-term post seismic discharge increasing.

For all the springs, spotless changes in their behaviour between the pre- and post-seismic period have been detected. The discharges decrease considerably after the co-seismic period in Capotenna, Sassospaccato, Pescara del Tronto and Capodacqua springs. This is clearly indicated by all the statistics shown in Table 3.16 (i.e., mean, minimum, 25th percentile, median, 75th percentile, and maximum values). However, to remove the possible influence of yearly variation in inflow volumes, the minimum value can be considered to point out the long-term decrease in spring discharges. As a result, the springs located in the eastern and southern-east sector of the study area suffered a net discharge loss as already highlighted in the previous chapter. Furthermore, especially for Foce spring, the random discharge peaks are more frequent and characterised by a larger amplitude: from few litres during the pre-seismic period to hundreds or thousands of litres after the co-seismic one. An opposite trend is recorded in Sordo spring, located westward with respect to the study area, for which the minimum discharge value doubled in the post-seismic period and the 75th percentile passes from 1910 L/s during the pre-seismic to 4361 L/s in the post-seismic period.

Table 3.16 Basic statistics calculated for the analysed springs during the pre-, co-, and post-seismic periods.

Pre-seismic period (First available continuous discharge data in November 2013 – 23rd August 2016)								
Spring	Unit	N° of data	Mean	Min	25th	Median	75th	Max
Capotenna*	L/s	147	152.4	118.6	138.2	142.4	163.7	206.4
Sassospaccato⁺	L/s	34	64.3	55	60	63	67	78
Foce	L/s	1027	1034	866.9	1020	1044	1060	1476
Pescara del Tronto	L/s	1027	309.9	119.9	192.8	313.6	412.6	559.5
Capodacqua	L/s	1027	430.3	205.6	326.8	444.6	529.8	671.4
Sordo	L/s	1027	1777	1464	1642	1770	1910	2697
Co-seismic period (24th August 2016 – 18th January 2017)								
Spring	Unit	N° of data	Mean	Min	25th	Median	75th	Max
Capotenna*	L/s	22	165.2	143.1	148.7	168.4	171.3	183.8
Sassospaccato⁺	L/s	5	55.8	50	50	55	61	63
Foce	L/s	148	1066	871.3	1036	1064	1091	1343
Pescara del Tronto	L/s	148	346.6	242.4	320.2	357.0	378.0	412.3
Capodacqua	L/s	148	568.3	410.1	559.2	581.0	602.7	700.5
Sordo	L/s	148	2785	1640	1830	2388	3842	4883
Post-seismic period (19th January 2017 – Last available continuous discharge data)								
Spring	Unit	N° of data	Mean	Min	25th	Median	75th	Max
Capotenna*	L/s	147	128.9	67.2	112.2	134.8	145.2	179.3
Sassospaccato⁺	L/s	41	32.3	19	26	29	35	61
Foce	L/s	1443	343.6	100.4	174.5	232.6	311.8	1740
Pescara del Tronto	L/s	1289	135.5	24.6	54.6	91	199.6	404.0
Capodacqua	L/s	1443	352.1	176.8	274.4	322.9	438.8	593.9
Sordo	L/s	750	3903	2933	3568	3676	4361	5890

* mean weekly data; + mean monthly data

The seismic effects recorded in the springs provided further impetuous to deeper the knowledge at local scale to assess any change in aquifer parameters, such as hydraulic conductivity or to identify possible modification occurred on aquifer geometry.

To do that the depletion coefficient (α) for the aforementioned springs has been calculated and shown in the following tables (Tables 3.17, 3.18, 3.19, 3.20, 3.21, 3.22). The springs for which daily data were available (Foce, Pescara, Capodacqua and Sordo) also shown the graphs of the identified depletion curve in logarithmic scale and the α values for the years before and after the seismic sequence (Figures 3.24, 3.25, 3.26, 3.27). One α value has been obtained for each baseflow recession period, providing a sufficient description of the slope of the recession curve ($R^2 > 0.9$).

Pescara del Tronto spring- hydrostructure 2a (daily data)

Table 3.17 Maillet depletion coefficients α and values of Q_0 for the analyzed recession periods. Data under the red line are related to post 30 October 2016 recession periods.

Year	Recession period	Q_0 (L/s)	α Maillet (d^{-1})
2003	Jul 15 th -Dec 16 th	270	$6.28 \cdot 10^{-3}$
2004	Aug 4 th – Nov 22 nd	402	$5.57 \cdot 10^{-3}$
2005	Aug 2 nd – Nov 10 th	401	$6.88 \cdot 10^{-3}$
2006-2007	Aug 4 th 2006 – Jan 26 th 2007	406	$7.37 \cdot 10^{-3}$
2007	Jun 22 nd – Sep 19 th	59	$6.26 \cdot 10^{-3}$
2008	Aug 22 nd – Nov 23 rd	249	$6.46 \cdot 10^{-3}$
2009	Jun 21 st – Dec 20 th	607	$6.86 \cdot 10^{-3}$
2010	Sep 26 th - Dec 3 rd	272	$5.77 \cdot 10^{-3}$
2011-2012	Dec 9 th 2011–Apr 10 th 2012	128	$5.41 \cdot 10^{-3}$
2012	Aug 23 rd – Nov 30 th	71	$4.08 \cdot 10^{-3}$
2013	Jul 16 th – Oct 26 th	467	$7.48 \cdot 10^{-3}$
2014	Jul 14 th – Dec 25 ^h	487	$7.63 \cdot 10^{-3}$
2015	Jul 17 th – Oct 4 th	453	$8.72 \cdot 10^{-3}$
2016	Jun 27 th –Aug 10 th	351	$4.09 \cdot 10^{-3}$
2016	Oct 2 nd – Oct 29 th	405	$6.80 \cdot 10^{-3}$
2016-2017	Dec 19 th 2016 – Feb 6 th 2017	322	$1.10 \cdot 10^{-2}$
2017	Sep 24 th – Oct 26 th	72	$1.46 \cdot 10^{-2}$
2018	Aug 10 th – Dec 31 st	312	$1.47 \cdot 10^{-2}$
2019	Sept 14 th – Dec 24 th	144	$1.3 \cdot 10^{-2}$

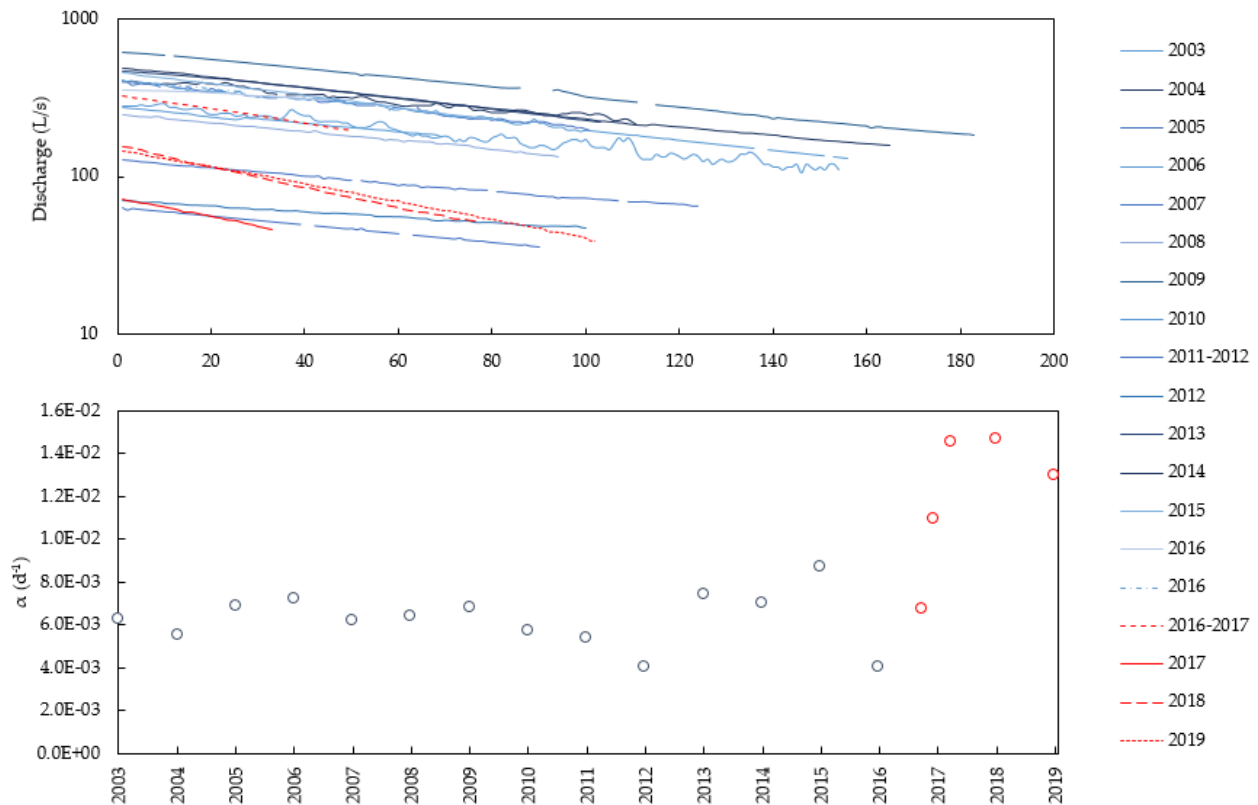


Figure 3.25 The depletion coefficients, α , for the years 2003–2019. Blue lines and dots indicate pre-earthquakes recession periods, red lines and dots are related to post-earthquake periods.

Capodacqua spring - hydrostructure 2a (daily data)

Table 3.18 Maillet depletion coefficients α and values of Q_0 for the analyzed recession periods. Data under the red line are related to post 30 October 2016 recession periods.

Year	Recession period	Q_0 (L/s)	α Maillet (d^{-1})
2003	Aug 4 th -Dec 2 nd	470	$5.66 \cdot 10^{-3}$
2004	Jun 26 th – Aug 16 th	629	$6.34 \cdot 10^{-3}$
2005	Sep 9 th – Oct 29 th	431	$4.21 \cdot 10^{-3}$
2006	Aug 19 th – Oct 17 th	507	$2.72 \cdot 10^{-3}$
2007	-	-	-
2008	Sep 19 th - Nov 30 th	405	$1.48 \cdot 10^{-3}$
2009	Oct 1 st – Dec 6 th	340	$3.81 \cdot 10^{-3}$
2010	Sep 8 th -Nov 20 th	431	$3.25 \cdot 10^{-3}$
2011-2012	Oct 20 th 2011–Feb 2 nd 2012	412	$1.61 \cdot 10^{-3}$
2012	Jun 14 th – Sep 2 nd	307	$3.89 \cdot 10^{-3}$
2013	Jul 16 th – Oct 10 th	552	$5.04 \cdot 10^{-3}$
2014	Aug 15 th – Dec 16 th	456	$3.81 \cdot 10^{-3}$
2015	Sep 8 th – Nov 25 th	323	$6.02 \cdot 10^{-3}$
2016	Jun 30 th –Aug 5 th	558	$4.33 \cdot 10^{-3}$
2016-2017	Dec 23 th 2016 – Jan 15 th 2017	603	$1.92 \cdot 10^{-2}$
2017	Oct 4 th – Dec 11 th	290	$2.29 \cdot 10^{-3}$
2018	Oct 20 th – Dec 18 th	319	$4.92 \cdot 10^{-3}$
2019	Aug 15 th – Dec 14 th	342	$5.1 \cdot 10^{-3}$

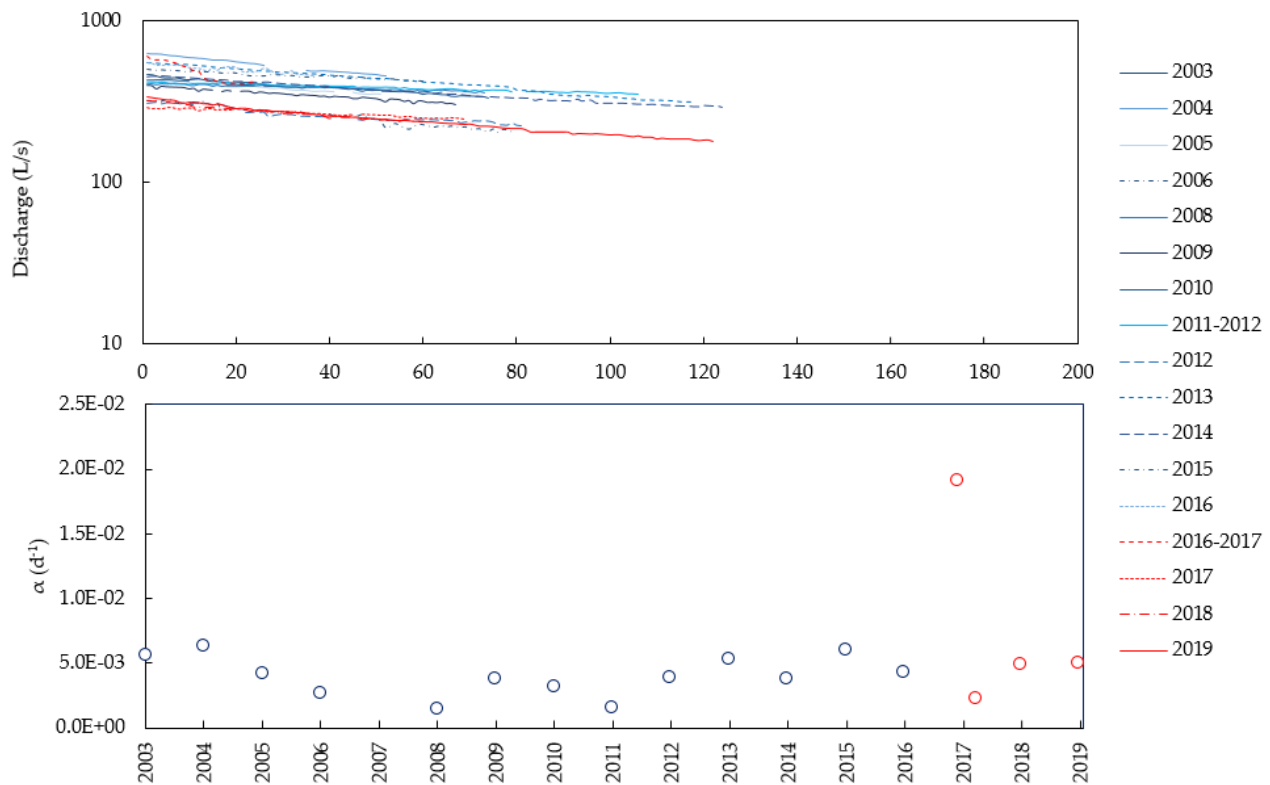


Figure 3.26 The depletion coefficients, α , for the years 2003–2019. Blue lines and dots indicate pre-earthquakes recession periods, red lines and dots are related to post-earthquake periods.

Foce spring - hydrostructure 1 (daily data)

Table 3.19 Maillet depletion coefficients α and values of Q_0 for the analyzed recession periods. Data under the red line are related to post 30 October 2016 recession periods.

Year	Recession period	Q_0 (L/s)	α Maillet (d^{-1})
2010	Sep 14 th -Oct 31 rd	1111	$9.61 \cdot 10^{-4}$
2011	Oct 30 th -Dec 2 nd	1029	$1.77 \cdot 10^{-3}$
2012	Jul 19 th – Sep 11 rd	966	$4.07 \cdot 10^{-3}$
2013	-	-	-
2014	-	-	-
2015	-	-	-
2016	-	-	-
2017	Sep 10 th – Nov 4 th	328	$3.1 \cdot 10^{-3}$
2018	Jun 22 th – Oct 28 th	311	$2.09 \cdot 10^{-3}$
2019	Jul 3 rd – Nov 2 nd	193	$2.09 \cdot 10^{-3}$
2020	Jun 20 th – Nov 19 th	160	$1.87 \cdot 10^{-3}$

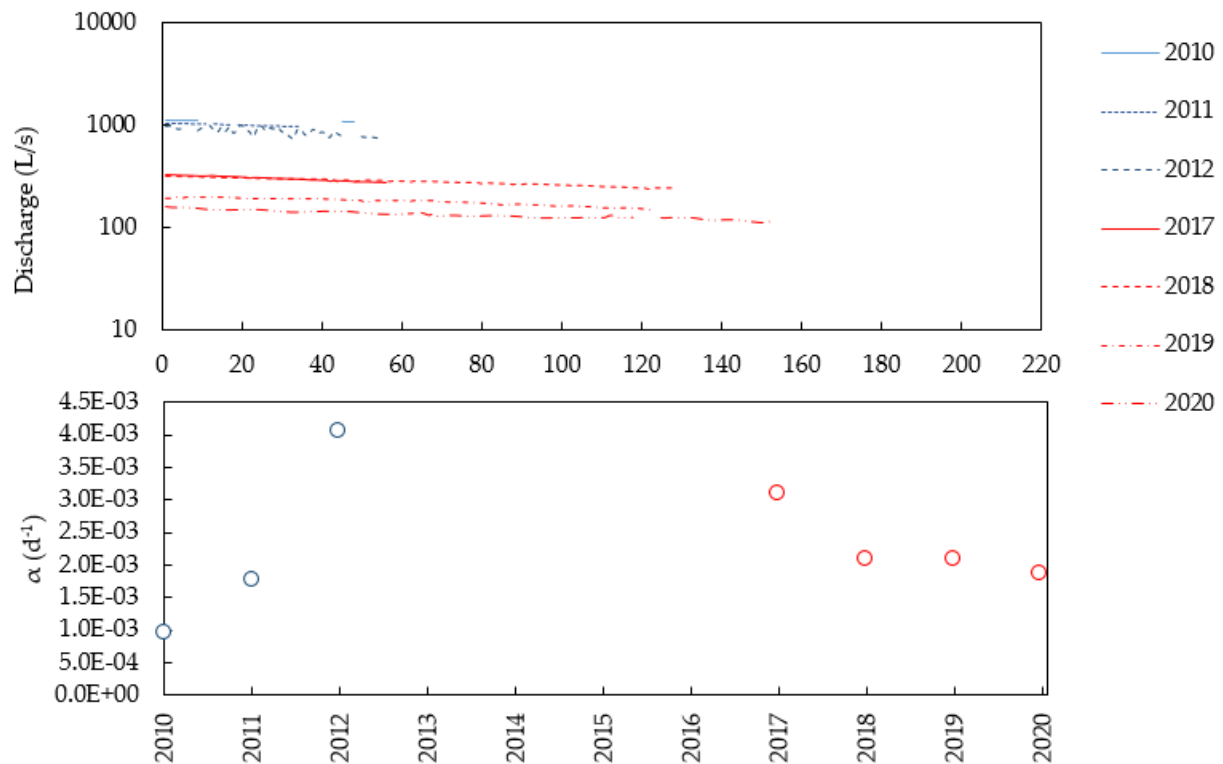


Figure 3.27 The depletion coefficients, α , for the years 2010, 2011 and 2012 (pre-seismic period) and 2017 to 2020 (post-seismic period). Blue lines and dots indicate pre-earthquakes recession periods, red lines and dots are related to post-earthquake periods.

Sordo spring- hydrostructure 3 (daily data)

Table 3.20 Maillet depletion coefficients α and values of Q_0 for the analyzed recession periods. Data under the red line are related to post 30 October 2016 recession periods.

Year	Recession period	Q_0 (L/s)	α Maillet (d^{-1})
2015	Mar 9 th – May 18 th	1764	$2.18 \cdot 10^{-3}$
2016	Apr 9 th – Apr 22 nd	1623	$1.11 \cdot 10^{-3}$
2017	May 16 th – Dec 10 th	4880	$1.24 \cdot 10^{-3}$
2018-2019	Oct 5 th 2018 – Jun 18 th 2019	3472	$1.49 \cdot 10^{-3}$

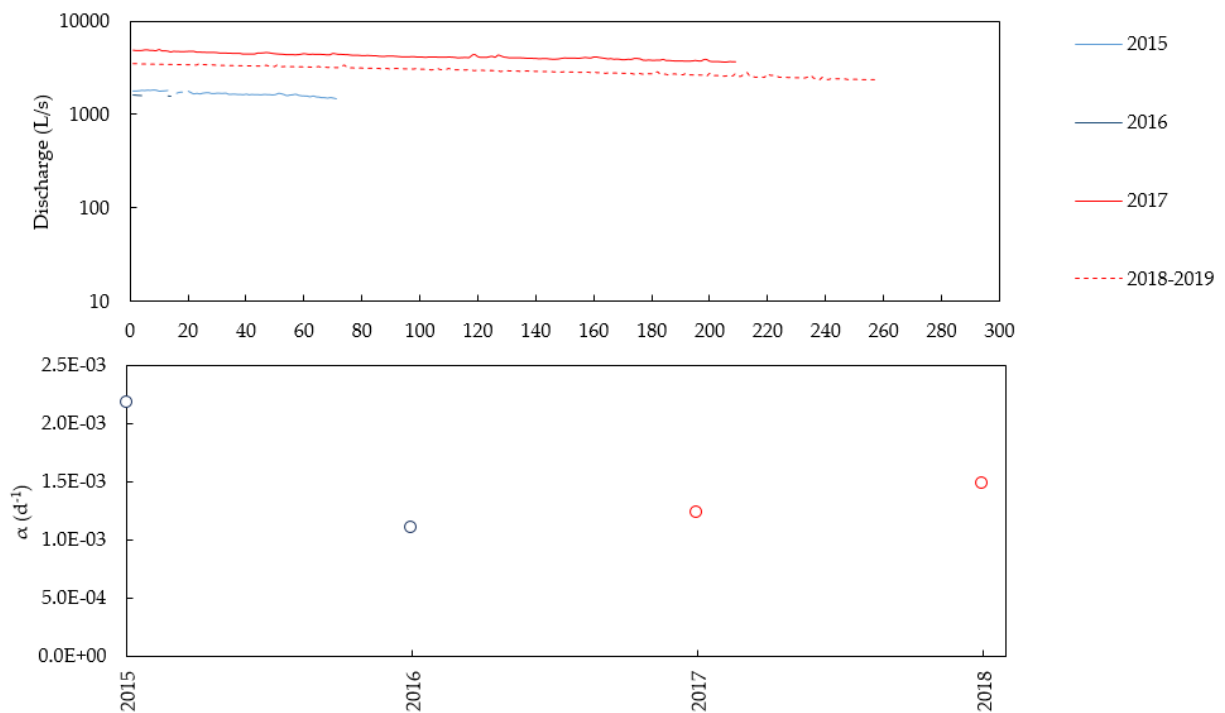


Figure 3.28 The depletion coefficients, α , for the years 2015 and 2016 (pre-seismic period) and 2017, 2018-2019 (post-seismic period). Blue lines and dots indicate pre-earthquakes recession periods, red lines and dots are related to post-earthquake periods.

Capotenna spring - hydrostructure 1 (weekly data)

Table 3.21 Maillet depletion coefficients α and values of Q_0 for the analyzed recession periods. Data under the red line are related to post 30 October 2016 recession periods.

Year	Recession period	Q_0 (L/s)	α Maillet (d^{-1})
2008	Aug 31 rd -Nov 16 th	184	$9.55 \cdot 10^{-3}$
2009	Aug 25 th -Nov 10 th	202	$5.82 \cdot 10^{-3}$
2010	Jul 28 th -Oct 27 th	210	$9.84 \cdot 10^{-3}$
2011	Aug 11 rd -Dec 22 nd	180	$7.33 \cdot 10^{-3}$
2012	Jul 6 th – Sep 14 th	172	$7.92 \cdot 10^{-3}$
2013	Aug 11 rd -Nov 10 th	206	$6.39 \cdot 10^{-3}$
2014	Aug 4 th -Nov 17 th	203	$3.85 \cdot 10^{-3}$
2015	Jul 21 rd -Oct 20 th	206	$4.96 \cdot 10^{-3}$
2016-2017	Dec 28 th 2016 -Feb 10 th 2017	168	$5.18 \cdot 10^{-3}$
2017	Jul 21 rd -Nov 3 rd	154	$8.1 \cdot 10^{-3}$
2018	Sep 15 th – Nov 22 th	151	$7.79 \cdot 10^{-3}$
2019	Sep 1 st – Oct 20 th	131	$9.06 \cdot 10^{-3}$

Sassospaccato spring - hydrostructure 1 (monthly data)

Table 3.22 Maillet depletion coefficients α and values of Q_0 for the analyzed recession periods. Data under the red line are related to post 30 October 2016 recession periods.

Year	Recession period	Q_0 (L/s)	α Maillet (d^{-1})
2007	Jul - Oct	60	$4.81 \cdot 10^{-3}$
2008	Aug - Oct	60	$5.06 \cdot 10^{-3}$
2009	Oct -Nov	58	$1.13 \cdot 10^{-3}$
2010	Aug - Oct	58	$1.48 \cdot 10^{-3}$
2011	Aug - Nov	56	$1.43 \cdot 10^{-3}$
2012	Jul - Sep	51	$9.78 \cdot 10^{-4}$
2013-2014	Dec 2013– Mar 2014	55	$1.5 \cdot 10^{-3}$
2014	Jun- Nov	78	$1.74 \cdot 10^{-3}$
2015	Jun- Oct	68	$1.39 \cdot 10^{-3}$
2016	Sep- Dec	63	$2.6 \cdot 10^{-3}$
2017	May - Nov	37	$3.78 \cdot 10^{-3}$
2018-2019	Jun 2018 – Feb 2019	50	$2.72 \cdot 10^{-3}$
2019	Jul - Sep	61	$6.58 \cdot 10^{-3}$

Where possible the α coefficient of 2016 has been calculated in two or more intervals discriminating the spring depletion curves prior, after the first shock (Mw 6.0 of the Amatrice seismic sequence August 24th 2016) and after the October 30th 2016 mainshock (Mw 6.5 of the Norcia seismic sequence). For Pescara del Tronto spring it can be observed that the average value of the Maillet depletion coefficient α before the October 30th, 2016, mainshock has been $6.35 \cdot 10^{-3} d^{-1}$ (standard deviation = $1.31 \cdot 10^{-3} d^{-1}$). On the contrary, since the first depletion period recorded after the 30th October 2016

earthquake, the α coefficient has significantly increased, with an average value of $1.32 \cdot 10^{-2} \text{ d}^{-1}$ (standard deviation = $2.11 \cdot 10^{-3} \text{ d}^{-1}$).

For Capodacqua spring the average value of Maillet depletion coefficient before the October 30th, 2016, mainshock has been $4.04 \cdot 10^{-3} \text{ d}^{-1}$ (standard deviation = $1.55 \cdot 10^{-3} \text{ d}^{-1}$). No significant increase in its average value has been recorded after the the 30th October 2016 earthquake ($7.88 \cdot 10^{-3} \text{ d}^{-1}$) however it has to be noticed that the first recession interval recorded after the October 30th 2016 mainshock (Dec 23th 2016 – Jan 15th 2017) is characterized by a value of α equal to $1.92 \cdot 10^{-2} \text{ d}^{-1}$, an order of magnitude higher than the pre-seismic period.

The other springs (i.e., Foce, Sordo, Capotenna and Sassospaccato) do not show significant variation of the α coefficient before and after the series of strong earthquakes and its average value is in the order of 10^{-3} d^{-1} for all the springs. In this context it has to be noticed that, unlike in other carbonate Apennines springs (Fiorillo, 2014; Fiorillo et al., 2019), the α values are similar even for highly different initial discharge (Q_0).

3.4 Artificial tracer tests analysis

The outcomes of artificial tracer tests conducted by Nanni et al. 2020 before and just after the seismic sequence (TEST1, TEST2 and TEST3), integrated with the results obtained by the new tracer tests (TEST4, TEST5 and TEST6) specifically designed for this PhD thesis, are described in the following paragraph point by point for each monitored spring with a focus for the Upper Nera River. Unfortunately, the old tracer tests conducted by Lippi Boncambi in 1950 lack quantitative results, and the hydraulic connection between the Mérgani sinking stream and some springs located in the Norcia Plain has been described only from a naturalistic standpoint without a map or a numerical approach. In this chapter, an overall description of the pre- and post-seismic modifications at regional scale is provided and synthesized by means of two graphical maps. Subsequently, the results obtained by the application of Qtracer2 (TEST4 and TEST5) to determine the hydrodynamic characteristics of the monitored springs are displayed. Finally, the outcomes retrieved by the new approach based on the comparison between periodic tracer tests and time series analysis performed by Sliding Windows Cross Correlation Function are shown.

Capotenna spring

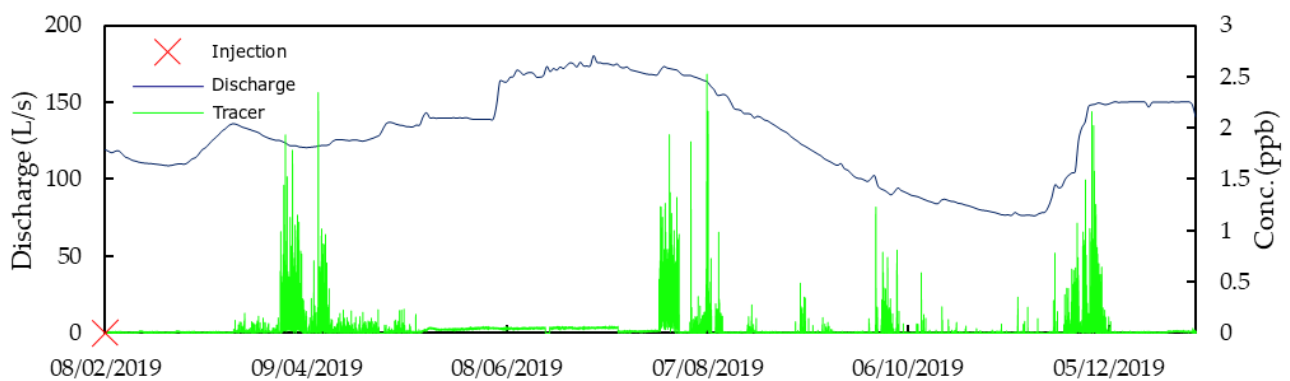
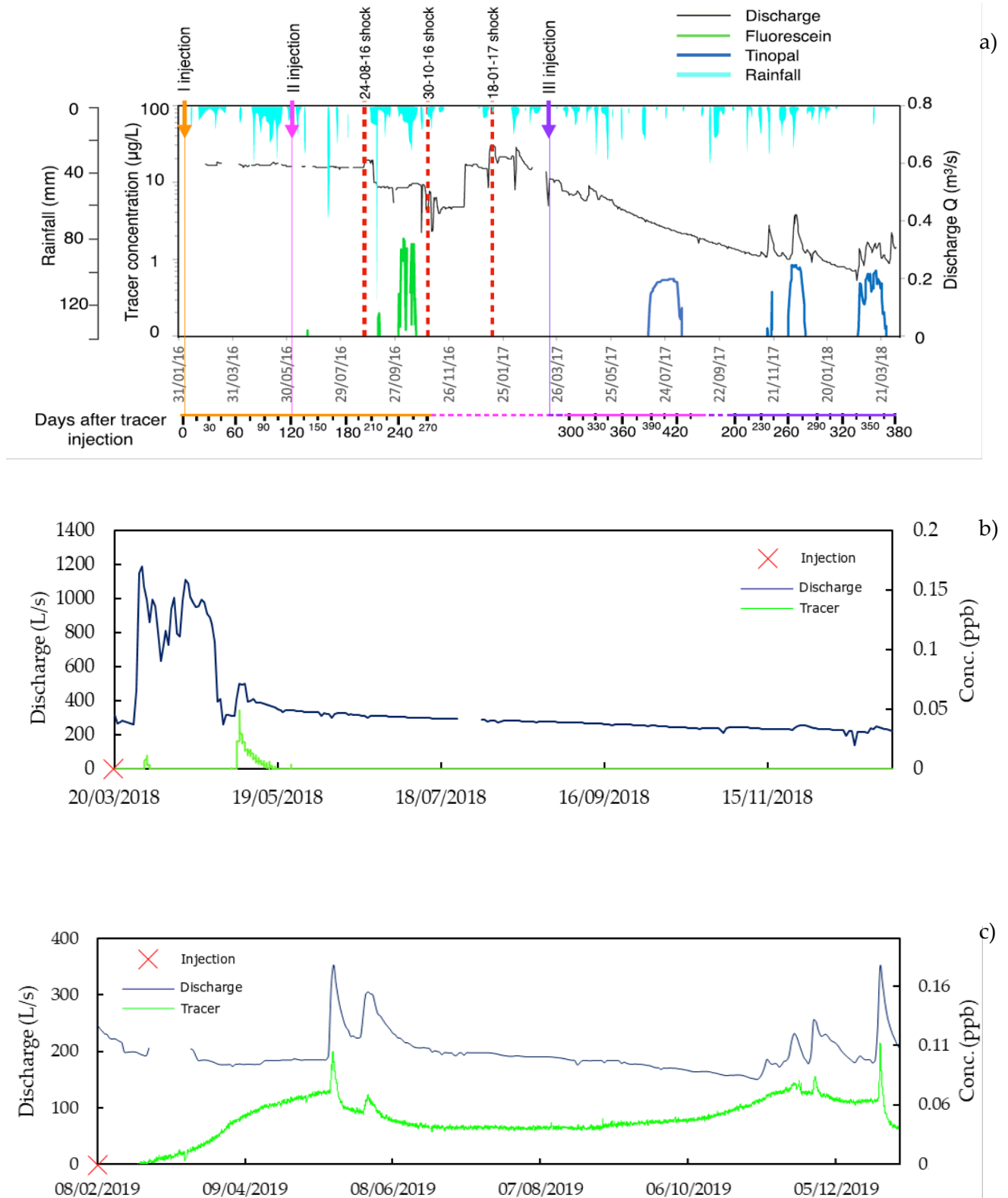


Figure 3.29 Tracer breakthrough curve (BTC) for Capotenna spring (hydrostructure 1) and spring discharge data recorded during TEST5. The green line refers to Fluorescein.

Only during TEST5 Capotenna spring, located about 15.5 km far from the injection point, has been monitored. Figure 3.29 shows the tracer arrivals and the daily discharge data of the spring during TEST5 from the date of the injection. The arrivals of the tracer are characterised by a typical impulsive behaviour, with peaks varying in duration from a few tens of minutes to 1-2 hours. In the graph, four main periods of tracer arrivals can be identified, the first one going from April 1st 2019 (first arrival 52 days after injection) to April 15th 2019, the second one going from July 24th 2019 to August 10th 2019, a third period, characterised by peaks of minor intensity between September 26th

and October 2nd 2019, and a last period going from November 18th to December 5th 2019. The tracer arrivals are not correlated to the discharge regime, except in the last period of arrivals in which an increase in spring discharge is recorded.

Foce spring



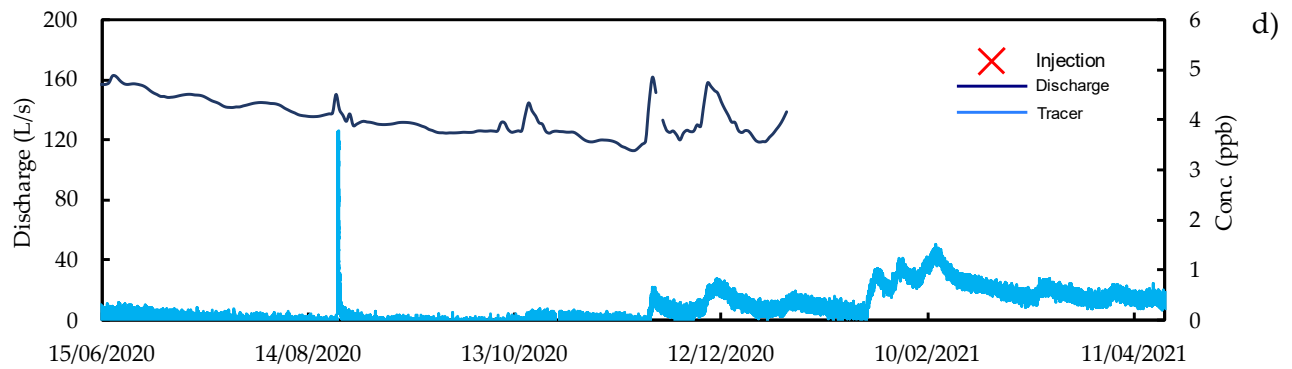


Figure 3.30 Tracer breakthrough curves (BTCs) for Foce spring (hydrostructure 1) and spring discharge data recorded during (a) TEST1, TEST2 and TEST3, modified from Nanni et al., 2020 (b) TEST4 (c) TEST5 and (d) TEST6. The green lines refer to Fluorescein, while the light blue lines refer to Tinopal CBS-X.

Foce spring is located 12.6 km far from the sinkhole. During TEST1 the tracer reached the spring for the first time after approximately 250 days from entering the sinkhole (45 days after the seismic event of August 24th, 2016) and its breakthrough curve is characterised by impulsive peaks. In TEST2, the first tracer arrival was recorded in the month of July, approximately 1 year after being injected, and persists in the spring until August 2017. Even the TEST3 is characterised by a long-lasting tracer arrival, but the peaks are located during the strong increase phases of the spring discharge (Figure 3.30a). TEST4, TEST5 and TEST6 show similar behaviour with respect to the peaks. In TEST4, the tracer was detected just after 45 days from the injection, during a strong increase of the spring discharge which passes from 312 L/s to 498 L/s in about 4 days (Figure 3.30b). During TEST5, a small amount of tracer (however major than the detection limit) was recorded during the whole test; however, the main peaks are located during the strong increases of the spring discharge, like TEST4. During the last test (TEST6), although the spring discharge data is available until December 31st, 2020, the fact that the tracer arrivals are related to the sudden spring discharge increase is confirmed. During all the tracer tests the analysis of fluorescent traps located within the spring drainage tunnel confirmed the arrivals, although characterized by low concentration values.

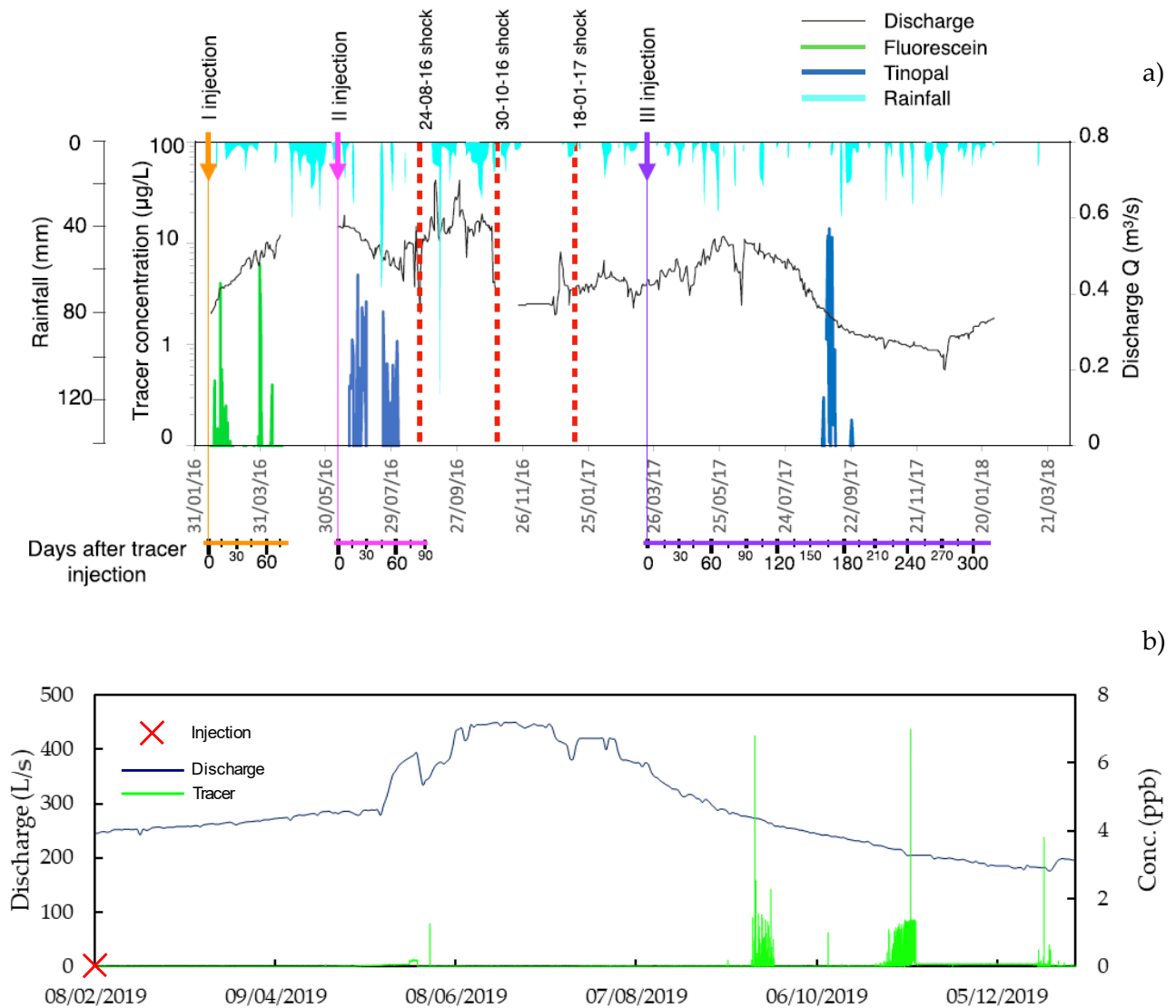


Figure 3.31 Tracer breakthrough curves (BTCs) for Capodacqua spring (hydrostructure 2a) and spring discharge data recorded during (a) TEST1, TEST2 and TEST3, modified from Nanni et al., 2020 (b) TEST5. The green lines refer to Fluorescein, while the blue lines refer to Tinopal CBS-X.

In Capodacqua spring (located 6.3 km from the injection point), the tracer is recovered rather quickly during the discharge increasing and spring recession periods of TEST1 and TEST2 respectively. Both the tests show similar behaviour: the tracer arrives at the spring in about one week after the injection, showing the main peak after 12 days for TEST1 and 20 days for TEST2, respectively. Other peaks are located between 40 and 60 days after the injection. TEST3, conducted after the main seismic events, is characterised by a lower velocity of tracer movement with respect to the others and the tracer

arrives in the spring between 161 and 170 days after the injection with a single marked peak of about 14 ppb related to a series of pulses located in the recession limb of the spring discharge.

In TEST5 as well, the main tracer arrivals are recorded in the depletion phase of the spring discharge. The first tracer arrival (i.e., 112 days after the injection) could be an early arrival of tracer in the spring; however, due to its low concentration, the possibility of being connected to the Fluorescein injection during TEST4 cannot be excluded. Moreover, the fluorimetric probe was not located in the spring during the TEST4 and therefore the first peak of tracer was not recorded. The major peaks of TEST5 are located after 217, 272, 315 days from the injection, suggesting even in this case a lower velocity of tracer with respect to the test performed in the pre-seismic period.

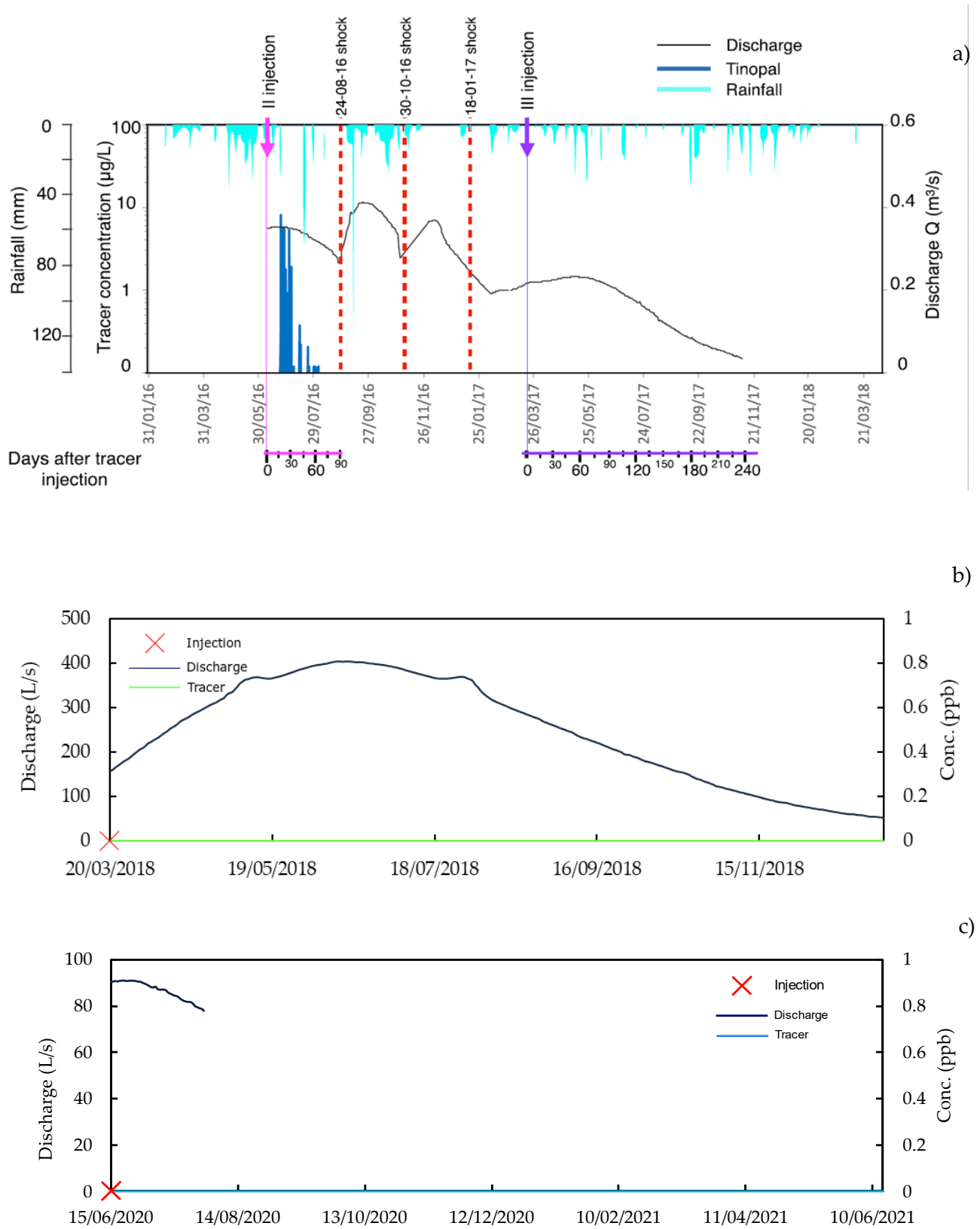


Figure 3.32 Tracer breakthrough curves (BTCs) for Pescara del Tronto spring (hydrostructure 2a) and spring discharge data recorded during (a) TEST2 and TEST3, modified from Nanni et al., 2020 (b) TEST4 and (c) TEST6. The green lines refer to Fluorescein, while the blue lines refer to Tinopal CBS-X.

Pescara del Tronto spring is located approximately 6.8 km from the injection point. The spring has been monitored during TEST2, TEST3, TEST4 and TEST6. Only during TEST2 conducted before the series of strong seismic events on 9 June 2016, by Nanni et al. 2020, the tracer arrives in the spring. The Tinopal CBS-X was recorded by the fluorimetric probe after 13 days from the injection with subsequent peaks at 18, 20, 24, and 26 days. The arrival of the tracer was recorded during the maximum flow rates of the spring (about 0.35 m³/s). The breakthrough curve is pulse-like. During TEST3, TEST4 and TEST6, (after the seismic sequence) the tracer did not arrive in the spring yet.

Upper Sordo River

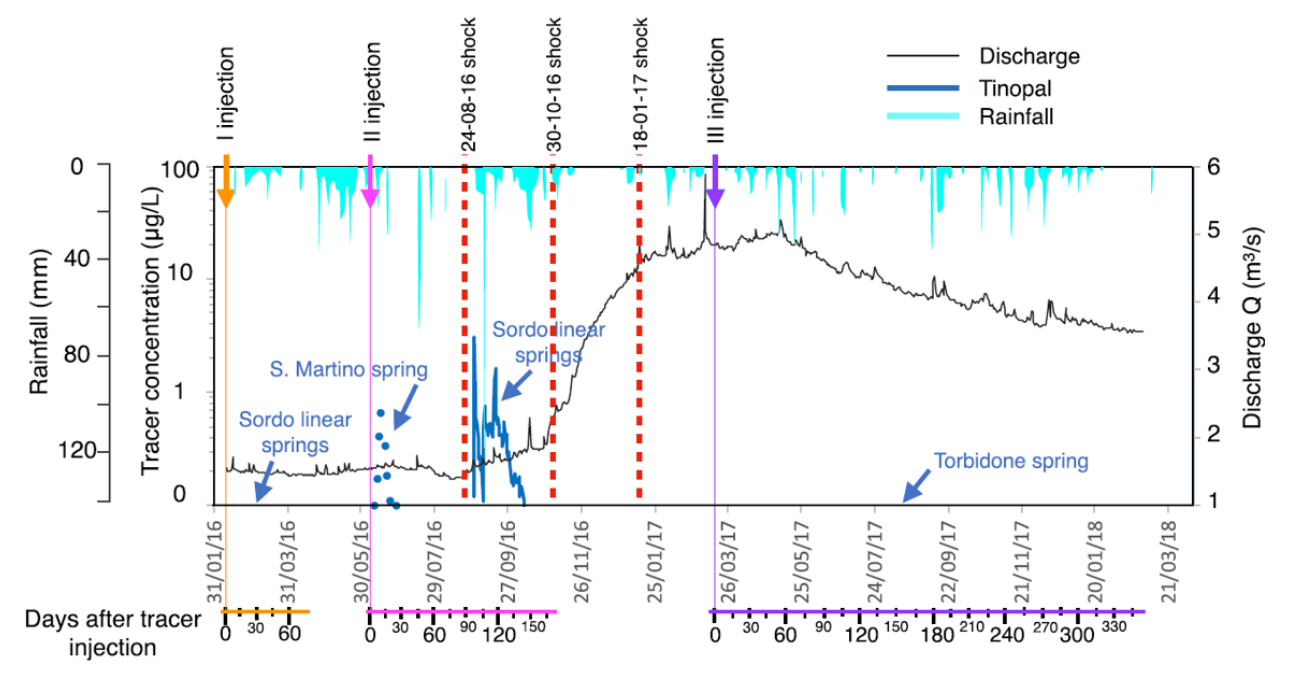


Figure 3.33 Tracer breakthrough curves (BTCs) for the monitoring points in the Upper Sordo River basin (i.e., San Martino spring, Torbidone spring and Sordo linear springs- hydrostructure 3) and discharge data recorded during TEST1, TEST2 and TEST3, modified from Nanni et al., 2020. The blue lines and dots refer to Tinopal CBS-X.

The hydrogeological setting of Sordo River Basin before the seismic sequence was deeply different with respect to the post-seismic period. About 8.8 km far from the injection point the linear springs of the Sordo River, emerges downstream of Norcia village; while upstream of the village, San Martino spring feeds the upper Sordo River basin (7 km from the injection point). During TEST1 no evidence of tracer arrivals were recorded in this area, neither in Sordo linear springs nor in San Martino spring. During TEST2 San Martino spring (periodically and manually sampled), shows a connection with the injection point. The tracer injected during TEST2, in fact, was recorded just 8 days after the injection operations. During the same test, the tracer arrived only 1 week after the 24th August earthquake (approximately 3 months after the tracer injection) in Sordo linear spring when

the river discharge increasing due to the seismic sequence started to be observed. The breakthrough curve is typically hydro-dispersive, and the tracer completely disappeared in approximately 20 days.

After the seismic-sequence the increase in discharge both recorded in the Sordo linear spring and both in San Martino spring was accompanied by the emergence of the new spring namely Torbidone, feeding the upper Sordo River basin upstream to Norcia village (about 6 km from the injection point). During TEST3 only Torbidone spring was monitored, and no evidence of tracer arrivals were recorded. The complex hydrogeological setting and the apparently discordant results observed during the first three tracer tests led to a specific monitoring of each point during the subsequent tracer tests. In particular, during TEST4 the arrivals in the Sordo linear spring and in the Torbidone spring were monitored, while in TEST5 a simultaneous monitoring was carried out in all three points (i.e., Sordo linear spring, Torbidone spring and San Martino spring).

Sordo linear springs

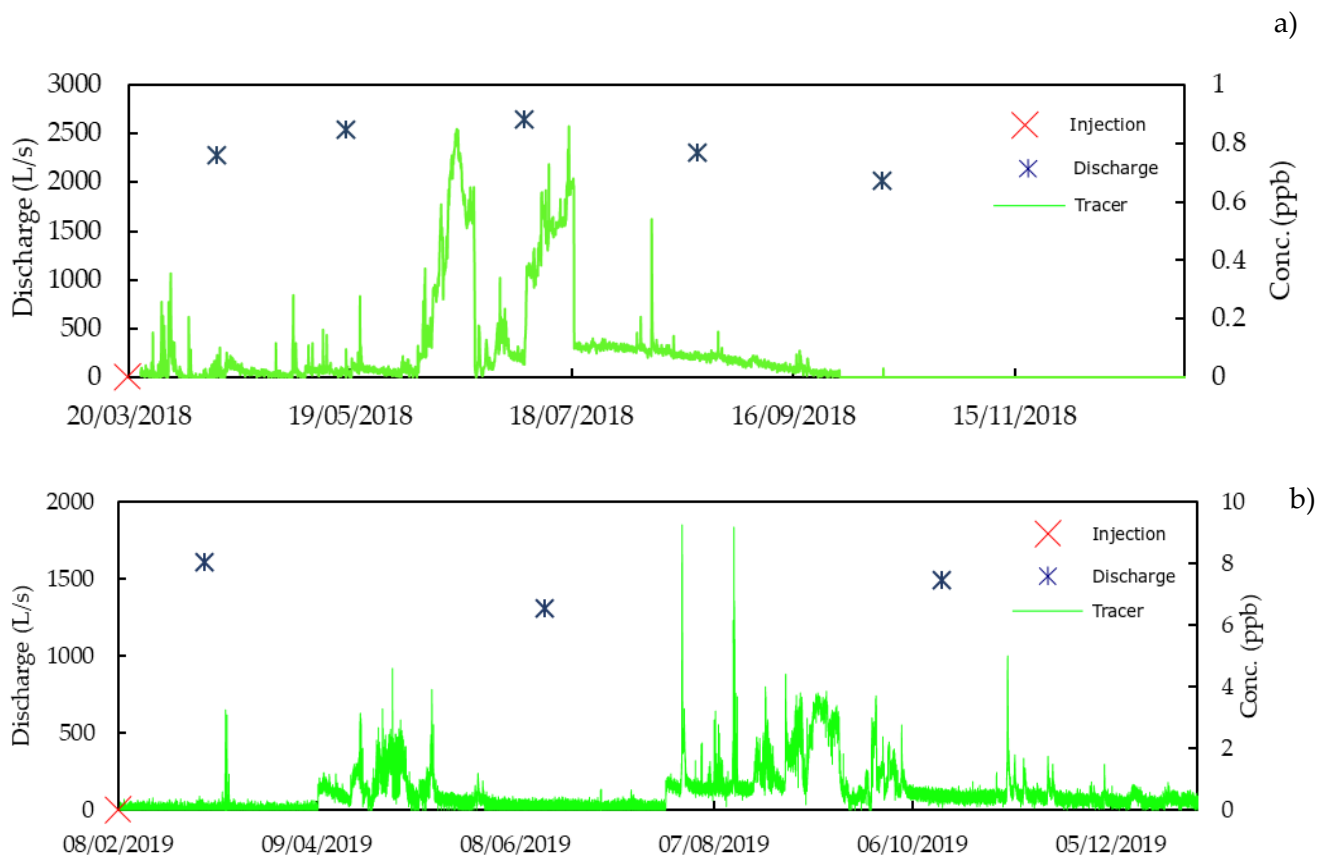


Figure 3.34 Tracer breakthrough curves (BTCs) for Sordo linear springs (hydrostructure 3) and spring discharge discrete data recorded during (a) TEST4 and (b) TEST. The green lines refer to Fluorescein.

During TEST4, the tracer reaches the Sordo linear spring only after a week from the injection. The breakthrough curve shows several impulsive peaks of small intensity repeated in time, and about 80 days after the injection an increase in tracer concentration was observed with a typical bell-shape curve (hydrodispersive behaviour typical of a porous aquifer), lasting about 20 days. A similar bell-shape curve is repeated about a month later with concentration values reaching 0.9 ppb. About 170 days after injection, a final impulsive peak was observed. Most of the tracer was recovered during the “bell-shape periods”, whose breakthrough curve shape depends on the dispersion of the tracer within the alluvial fluvial-lacustrine deposits covering the carbonate geological formation in the Norcia Plain. Unfortunately, the lack of continuous discharge data does not permit to appreciate the possible correlation between the tracer arrivals and the discharge regime.

Figure 3.34b shows the tracer arrivals and the discharge for Sordo linear spring starting from the injection date during TEST5. The breakthrough curve highlights a clear connection between Pian Grande Plain and the monitoring point. The graph shows the superimposition of several peaks pulses-like and others lasting several days; in particular, from April to June 2019 and from August to October 2019, classic bell-shaped curves with total duration up to 11 days were observed, denoting a typically hydro-dispersive character. The first arrival of the tracer at the monitored point is observed 32 days after release. The maximum tracer concentration recorded during TEST5 is one order higher than TEST4, probably due to the higher amount of tracer injected during TEST 5 (27 kg vs 16 kg) or due to the decreasing in discharge values, leading to a higher tracer concentration due to a lower dilution.

Also in this case, the lack of continuous discharge data does not permit to correlate the discharge regime with the tracer arrivals; the discharge values range between a maximum of 1.69 m³/s and a minimum value of 1.15 m³/s, recorded at the end of the test.

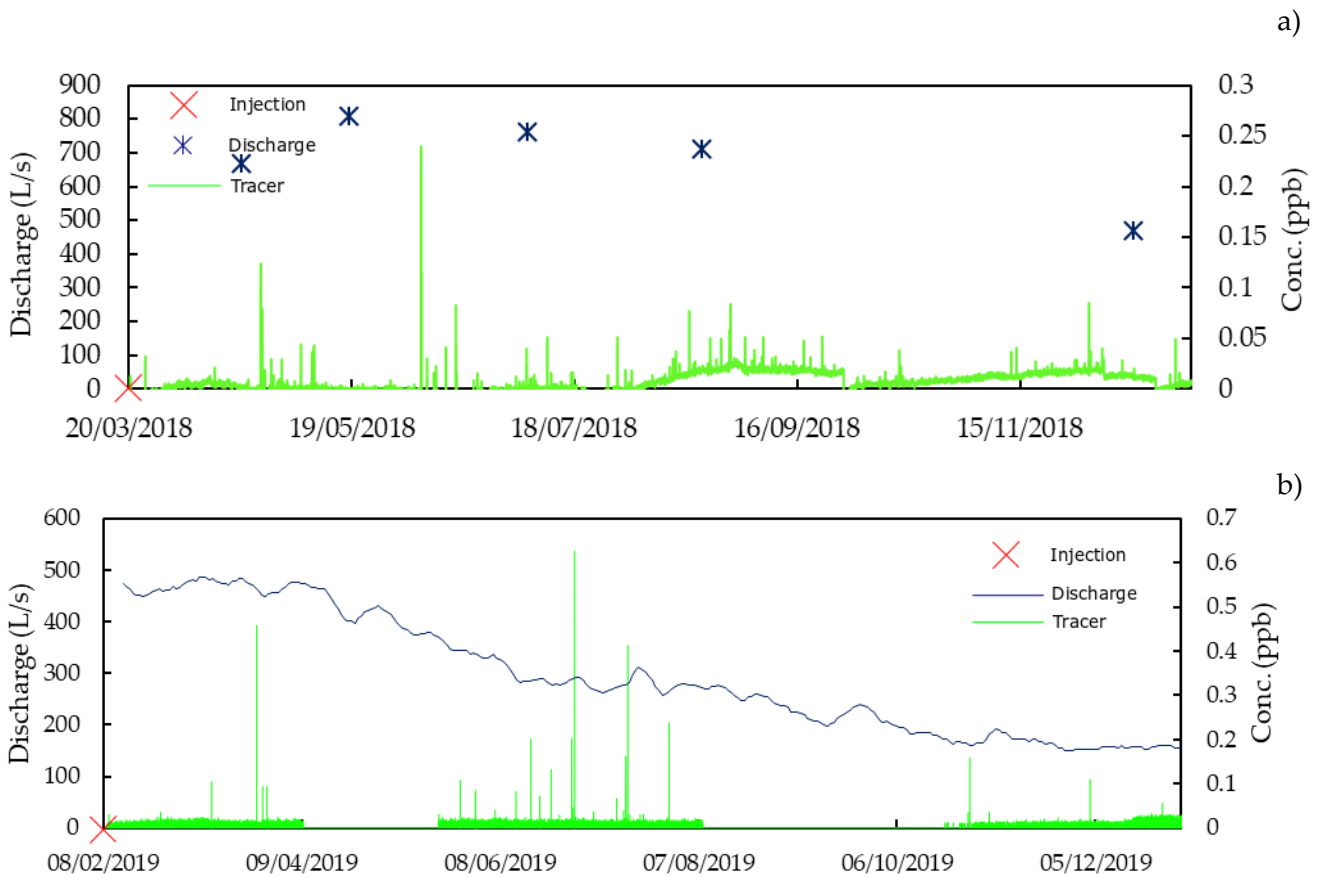


Figure 3.35 Tracer breakthrough curves (BTCs) for Torbidone spring (hydrostructure 3) and spring discharge data recorded during (a) TEST4 and (b) TEST5. The green lines refer to Fluorescein.

The field fluorometer positioned in the Torbidone spring recorded a different tracer arrivals behaviour with respect to the Sordo linear spring, both during TEST4 and TEST5. The tracer arrivals are purely impulsive, with numerous peaks of medium-low intensity repeated from about one month after tracer release. In TEST4 the highest peak is recorded about 75 days after the injection (0.25 ppb). During TEST5 the first tracer arrival is recorded after 46 days from the injection with a concentration value of 0.46 ppb. The maximum concentration peak (0.63 ppb) was recorded 142 days after the injection. Others lower peaks were observed up to approximately 288 days after the tracer release. As can be seen from the graph, the probe showed some operational problems in the periods between April 8th - May 20th, 2019, and between August 8th and October 20th, 2019, periods in which the probe stopped recording. For this reason, the total recovered mass calculated for the monitored point is certainly underestimated. No correlation between the tracer arrivals and the discharge trend, which ranges between a maximum value of about 0.5 m³/s and a minimum of about 0.15 m³/s, with

a general decreasing value, was recorded. Comparing the results of TEST5 to the ones of TEST4, a similar behaviour can be observed with lower concentration peaks during the latter, probably due to the higher quantity of tracer injected during TEST5.

In both the tests the connection of the spring with the water infiltrating in the Pian Grande Plain is clearly visible, and the tracer arrivals are characterized by impulsive peaks lasting from a few minutes to a few tens of minutes. The difference of breakthrough curve with respect to the Sordo linear spring may be because in Torbidone spring the tracer reaches the emerging point directly through the karstified and fissured carbonate geological formation recognised in the emergence zone, without filtering through the fluvial-lacustrine deposits of Norcia Plain.

San Martino spring

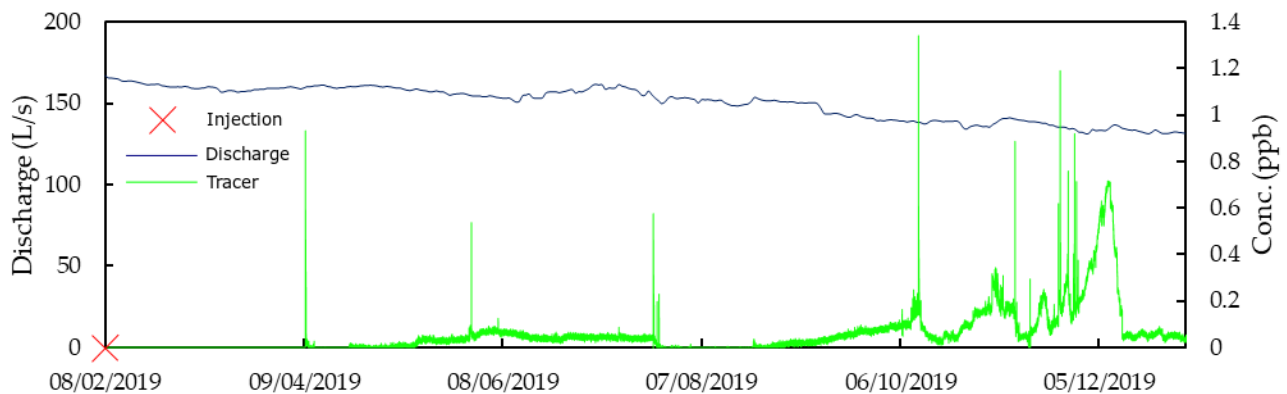


Figure 3.36 Tracer breakthrough curve (BTC) for San Martino spring (hydrostructure 3) and spring discharge data recorded during TEST5. The green line refers to Fluorescein.

Figure 3.36 shows the tracer arrivals and the discharge for San Martino spring starting from the injection date during TEST5. For this monitoring point only TEST5 is available. The connection of the spring with the water infiltrating in the Pian Grande Plain is clearly visible.

The breakthrough curve shows the superposition of pulses-like lasting a few hours suggesting a dominant hydro-dispersive character. In particular, starting from October 15th, several bell-shape curves lasting up to 16 days, are observed. The first tracer arrival is recorded 60 days after tracer release. No correlations between the tracer arrivals and the spring flow are visible, and the discharge values ranges between a maximum value of about 0.16 m³/s and a minimum of about 0.13 m³/s.

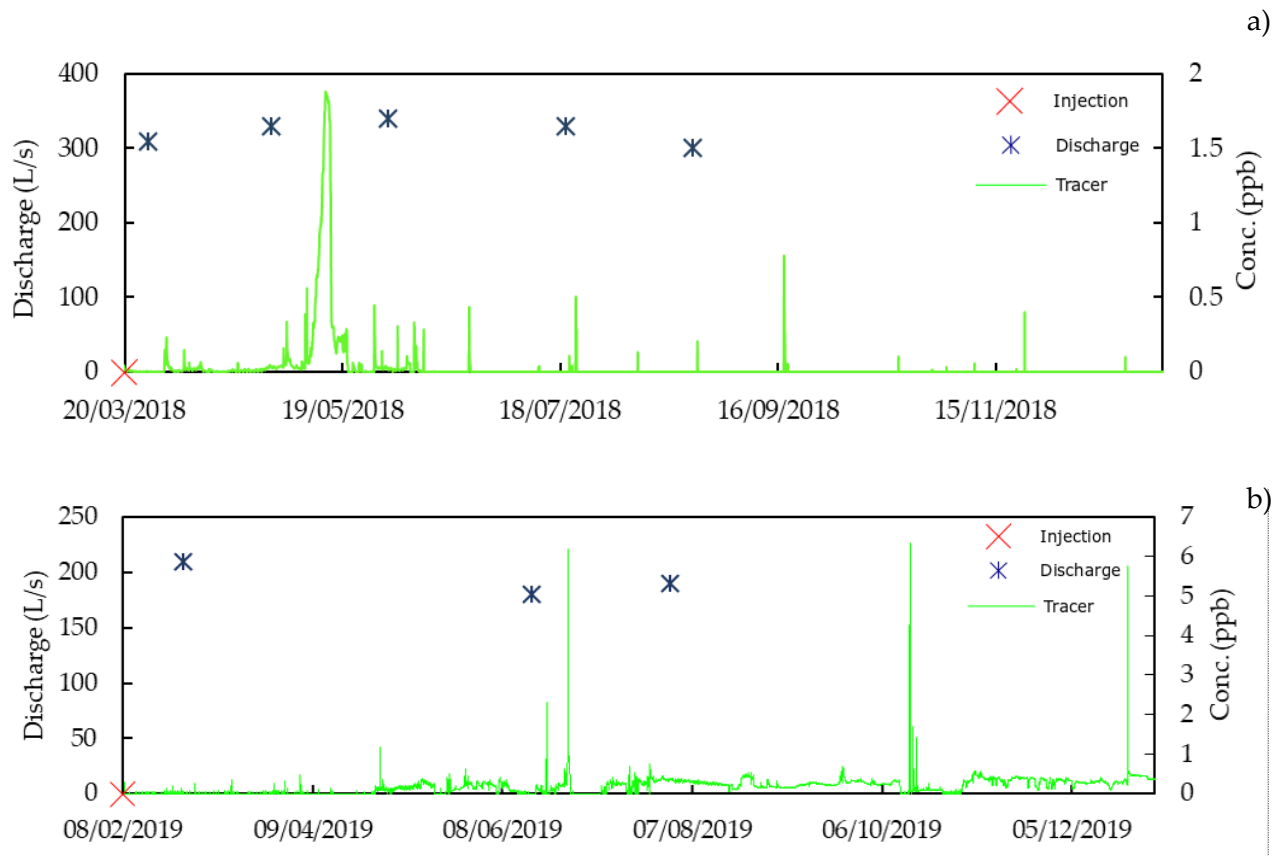


Figure 3.37 Tracer breakthrough curves (BTCs) for Campiano spring (hydrostructure 3) and spring discharge discrete data recorded during (a) TEST4 and (b) TEST5. The green lines refer to Fluorescein.

The monitored point is located along the Campiano stream nearby Campi village (10.6 km from the injection point). The breakthrough curve during TEST4 shows a first arrival after 45 days from the injection, characterised by an impulsive behaviour. Then, between 50 and 60 days from the injection a bell-shaped was observed (probably due to the contribution of one of the two streams of the Campiano stream upstream from the monitoring point, the one flowing on sandy-gravel alluvial deposit). Subsequently, repeated impulsive arrivals were recorded. The most intense after 4 months (1.5 ppb), probably due to the contribution of the other stream, the one emerging from the carbonate outcrops.

Figure 3.37b shows the tracer arrivals and the discharge values recorded during TEST5. The arrivals are characterised by a superposition of pulses lasting several hours with a dominant hydro-dispersive character. In particular, in May, October and December 2019, the presence of very concentrated peaks was recorded (up to 6.3 ppb), superimposed on the classic bell-shape denoting a typically dispersive hydrodynamic feature. The first arrival of the tracer during TEST5 is observed

after 81 days from the injection. Unfortunately, also in this case the lack of continuous discharge data does not permit to correlate the discharge regime with the tracer arrivals; the flow rate ranges between a maximum value of about 0.22 m³/s and a minimum of about 0.14 m³/s, recorded at the end of the test. The tracer tests outcomes evidence a hydraulic connection between the monitoring point and Pian Grande Plain.

Nera i Molini linear springs

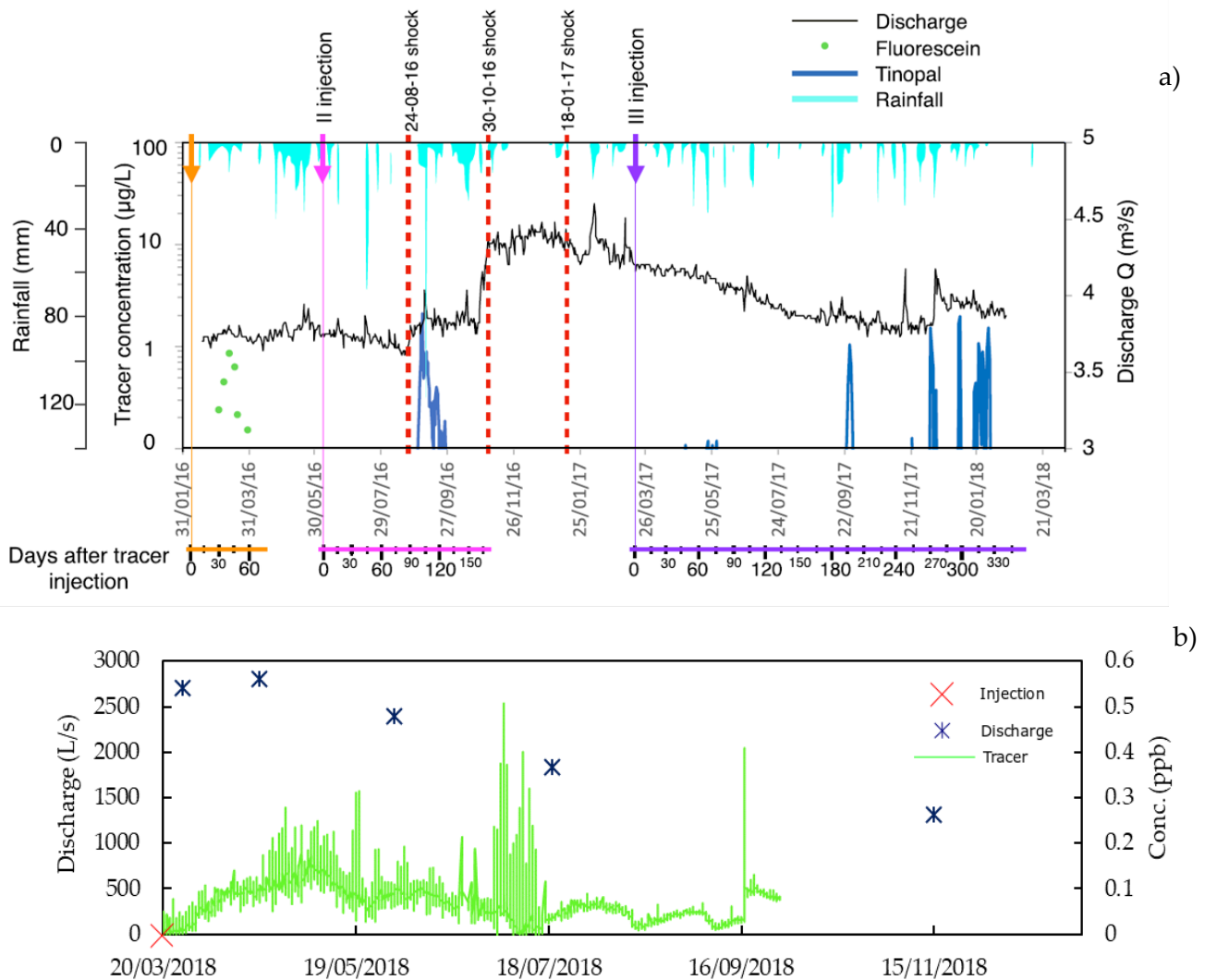


Figure 3.38 Tracer breakthrough curves (BTCs) for Nera i Molini linear springs (hydrostructure 2) and discharge data recorded during (a) TEST1, TEST2 and TEST3, modified from Nanni et al., 2020 (b) TEST4. The green lines and dots refer to Fluorescein, while the blue lines refer to Tinopal CBS-X.

Approximately 3 up to 5 km downstream of Visso village, some linear springs recharge the Nera River in the outcropping zone of the **BAS** (point 10 in figure 2.2), about 18.8 km far from the tracer injection point. During the TEST1, the tracer was detected (through discrete manual sampling) 26 days after its injection, and its arrival lasted approximately 15 days during the increasing of Nera

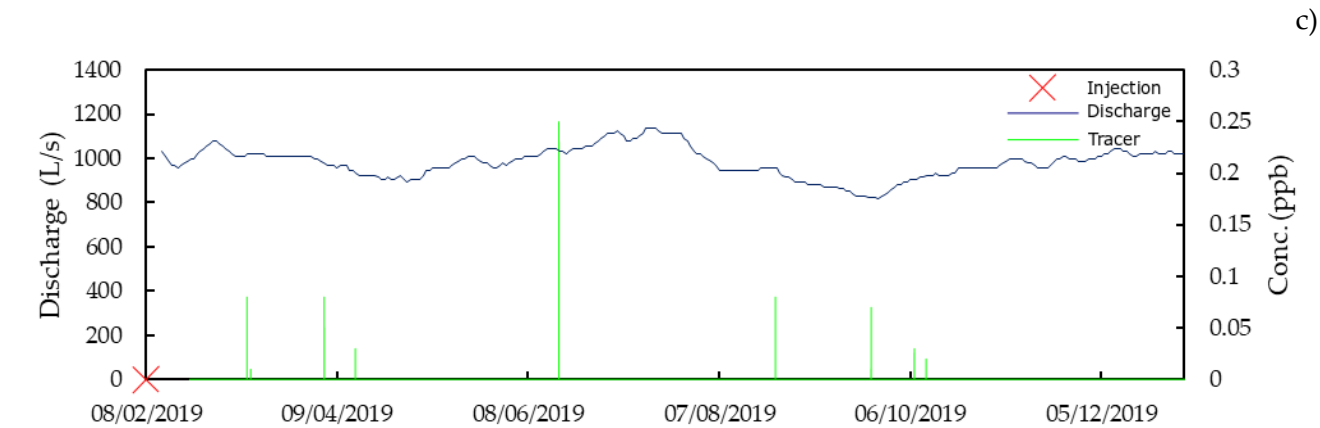
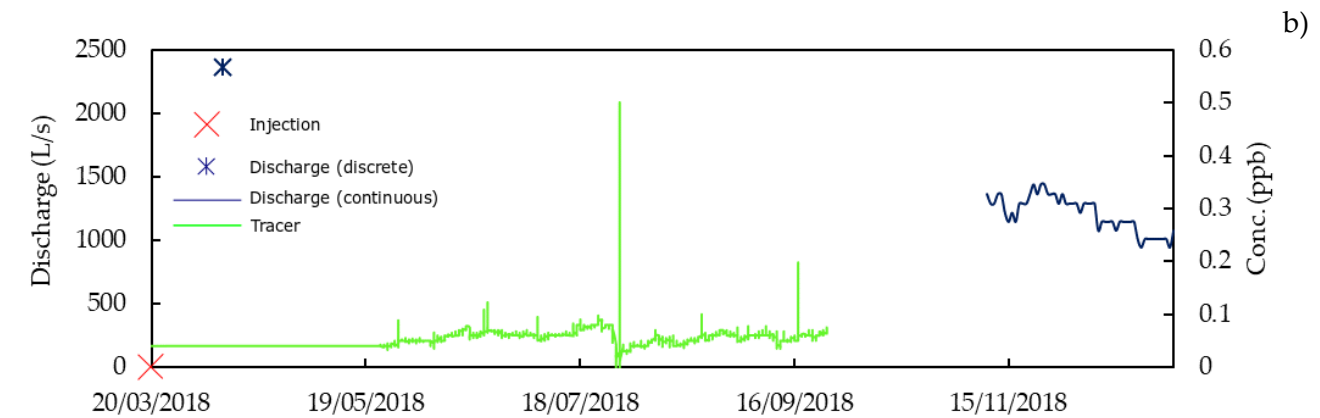
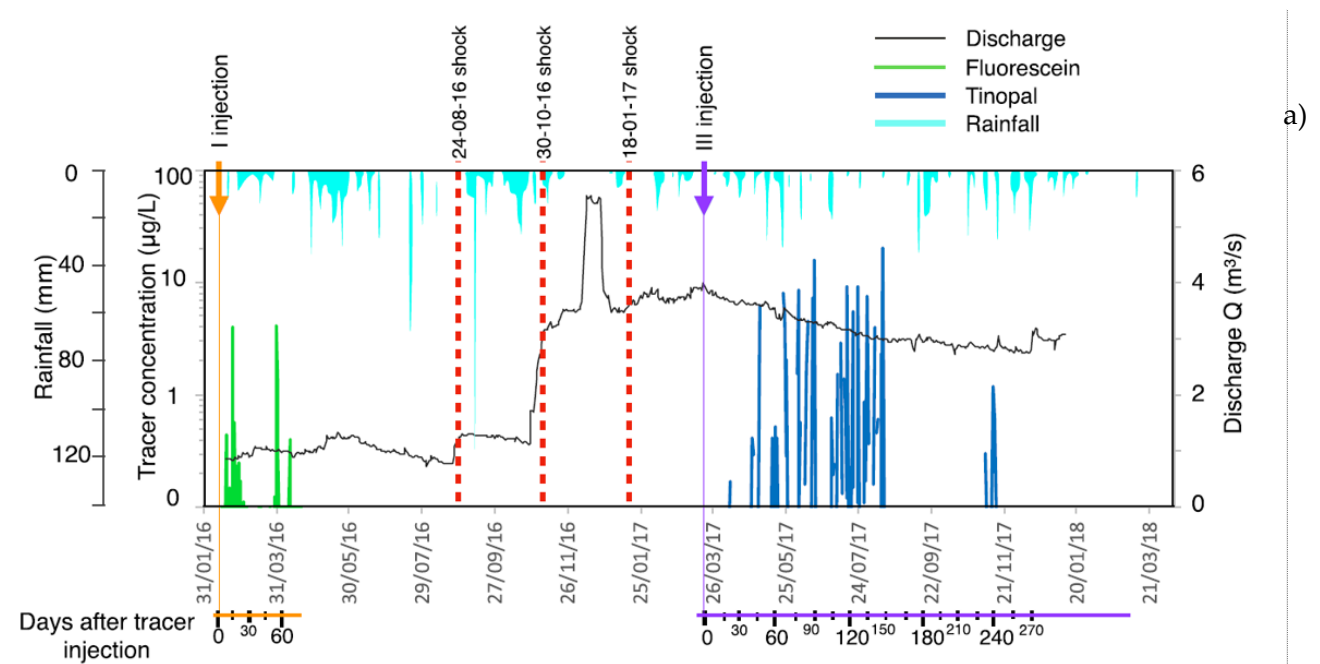
River discharge. The Tinopal CBS-X injected during the recession limb of the discharge curve in TEST2 was detected only 6 days after the August 24th earthquake, probably as a result of a sudden increase in river discharge which passes from 0.9 to 1.2 m³/s, but after 130 days from the injection. Both curves (TEST1 and TEST2) show a sort of dispersion, with some impulsive peaks especially recorded in TEST2. During TEST3, the arrival of the tracer occurred with different peaks separated each other over long-time intervals. The first arrival occurred during the discharge recession period, approximately 120 days after the tracer injection, and the arrivals of the successive peaks coincided with modest increases in discharge values lasting approximately 7 months.

Also in TEST4 the tracer arrives in the linear springs downstream to Visso village; in this case the signal is very disturbed, at least in the initial phase of the test, due to a problem probably caused by an electromagnetic issue, which was later resolved. In addition, some works on a landslide affecting the watercourse just upstream of the measuring point has led to an increase in the turbidity of the water, for this reason it was necessary to raise the detection limit of the tracer, thus losing some sensitivity. The tracer concentration could therefore be slightly underestimated by 0.1-0.2 ppb if compared to the actual concentration. In any case, it is possible to identify some impulsive peaks, starting from a couple of months up to 4 months after the injection, the last peaks were recorded 6 months after tracer release in the sinkhole.

Upper Nera River Basin

The Upper Nera River represents one of the main groundwater drainage area of the Sibillini Mts. towards the northern sector of the massif. The complex hydrogeological setting, characterised by a strong interaction between groundwater and surface water and the fact that the hydrogeological boundaries between different systems have never been clearly defined yet (Scarpelli et al., 2019; Fronzi et al., 2020; Nanni et al., 2020; Fronzi et al., 2021) has prompted a focus on the connection between the two sub-basin (Castelsantangelo stream and Ussita stream) and the Pian Grande Plain. For this reason, 6 monitoring points (6-7-8-9-11-12) have been selected to detect possible tracer arrivals in both the sub-basins (Figure 3.41a).

Nera at Castelsantangelo- San Chiodo spring (point 8 in Figure 3.41a)



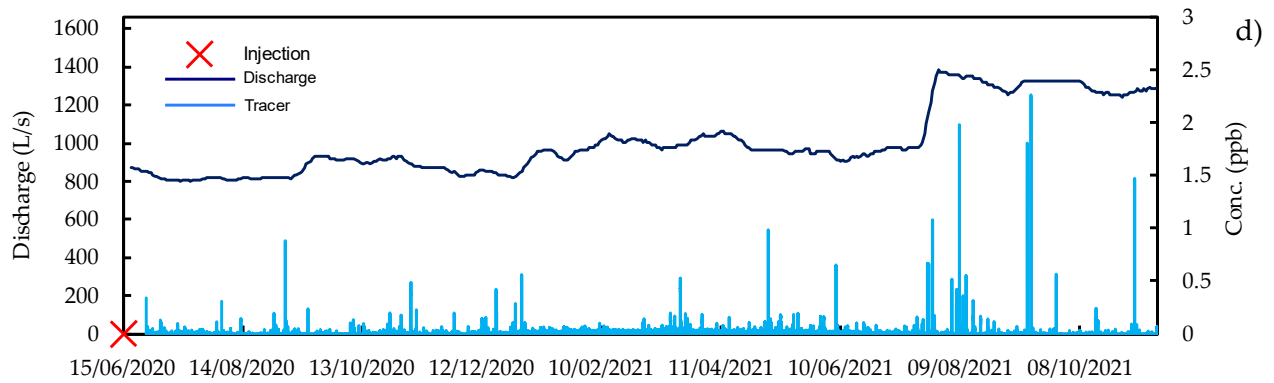


Figure 3.39 Tracer breakthrough curves (BTCs) for Nera at Castelsantangelo-San Chiodo spring (hydrostructure 2) and spring discharge data recorded during (a) TEST1 and TEST3, modified from Nanni et al., 2020 (b) TEST4 (c) TEST5 and (d) TEST6. The green lines refer to Fluorescein, while the light blue lines refer to Tinopal CBS-X.

The San Chiodo spring emerges along the Nera stream valley in the proximity of Castelsantangelo Village (about 12.6 km from the injection point). During TEST1 the fluorescein reached the spring after 19 days from the injection during the rising limb of the spring discharge and a similar quantity of tracer (4 ppb) has been recorded by the fluorometric probe between March 30th and 31st, 2016 (Figure 3.39a). During TEST2 the spring was not monitored. After the earthquake, the Nera River discharge increased about 4 times with respect to the pre-seismic values.

As TEST3 concerns, it was performed during the recession limb of the river hydrograph and the Tinopal CBS-X arrived after 28 days from the injection. The breakthrough curve shows close pulses and high concentrations until 20 ppb, and this arrival pattern was repeated continuously for a period of approximately 4 months. TEST 4, TEST5 and TEST6 show similar breakthrough curves characterised by a pulse pattern. As concerns TEST4, a general decrease in tracer concentration is observed with respect to that detected just after the seismic period during the TEST3. In particular, several peaks of tracer occurred from about two months after the injection. The main peak was recorded by the fluorometric probe on 27 July 2018 (0.5 ppb). TEST5 shows a slight decrement of tracer peak concentration with respect to the previous tracer tests, according to the general decrease of springs and rivers discharge occurred after a first stage of growth. TEST6 show several impulsive tracer arrivals starting from about one month after the injection lasting for the whole monitoring time (until November 2021). TEST6 is characterized by higher concentration value with respect to TEST4 and TEST5, probably due to the higher amount of tracer injected in the sinkhole (80 kg of Tinopal CBS-X).

Castelsantangelo stream linear spring (point 9 in Figure 3.41a)

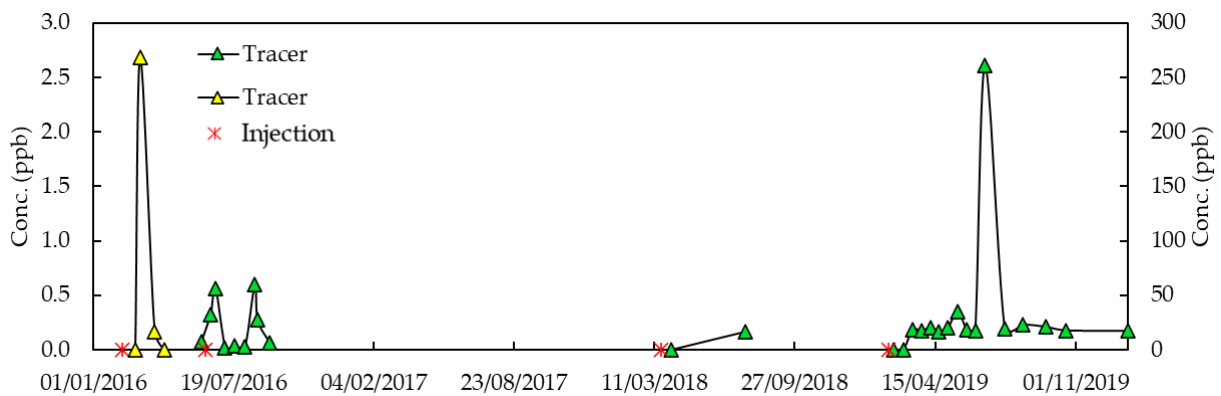


Figure 3.40 Tracer concentration measured in punctual samples during TEST1, TEST2, TEST4 and TEST5. The principal vertical axis refers to the yellow symbols, while the secondary vertical axis refers to the green ones.

Regarding the linear spring emerging downstream of Castelsantangelo village (points 9 in Figure 3.41) it was manually sampled during TEST1, TEST2, TEST4 and TEST5. The spring is located about 14 km far from the Mèrgani sinkhole and it recharges the Nera stream in the outcropping zone of the **MAI**. As concerns TEST1 a water sample collected on March 8th, then analysed in the laboratory spectro-fluorometer, showed a tracer concentration of about 234 ppb. The decrease in tracer concentration following this peak was recorded in the water sample collected on March 29th characterized by 11.5 ppb. During TEST2 a bimodal behaviour in the tracer arrivals is recognised with a maximum tracer concentration of 0.6 ppb. As TEST4 concerns, only two samples were collected in this point, and only the one of July 18th, 2018, is marked by a positive concentration (0.17 ppb). The water samples collected during TEST5 show a main peak on June 25th, 2019 (2.6 ppb).

In all the other monitoring points of the Upper Nera River basin (upstream to Visso village), located both in the Nera and Ussita streams, no tracer was detected neither before neither after the seismic sequence. This outcome is also confirmed by no tracer detection by the active carbons in any point.

At the same time, is important to highlight that the Breakthrough Curves of point 8 and point 9 show a decoupling in time of tracer arrivals and quantity each other, and this aspect is well evidenced in the TEST1 and TEST5. In particular, point 9 is characterised by several peaks showing higher tracer concentration values respect to point 8. In this context the field work conducted during the PhD activities revealed the presence of a previously unmapped fault whose strike roughly corresponds

to the Nera River trace in the area between point 8 and point 9. The fault strikes N170 and dips to the SW at 42, its kinematics is clearly normal and dip-slip as shown by the presence of calcite slicken-lines (Figure 3.41b). This is not a regional scale tectonic structure as it extends for a length in the order of 3–4 km. Nevertheless, the fault, which at surface affects mainly the Maiolica aquifer (MAI) and the Jurassic aquiclude (JUR), extends in the subsurface also dissecting the Basal aquifer (BAS). The graphical results of the tracer tests for the upper Nera River basin, together with the location of the new mapped fault are reported in Figure 3.40a. In particular, the main schematic flow paths from the injection point towards the analysed basins (represented by the green arrows), highlight the hydrogeological connection between the Pian Grande Plain and points 8 and 9, located along the Nera stream.

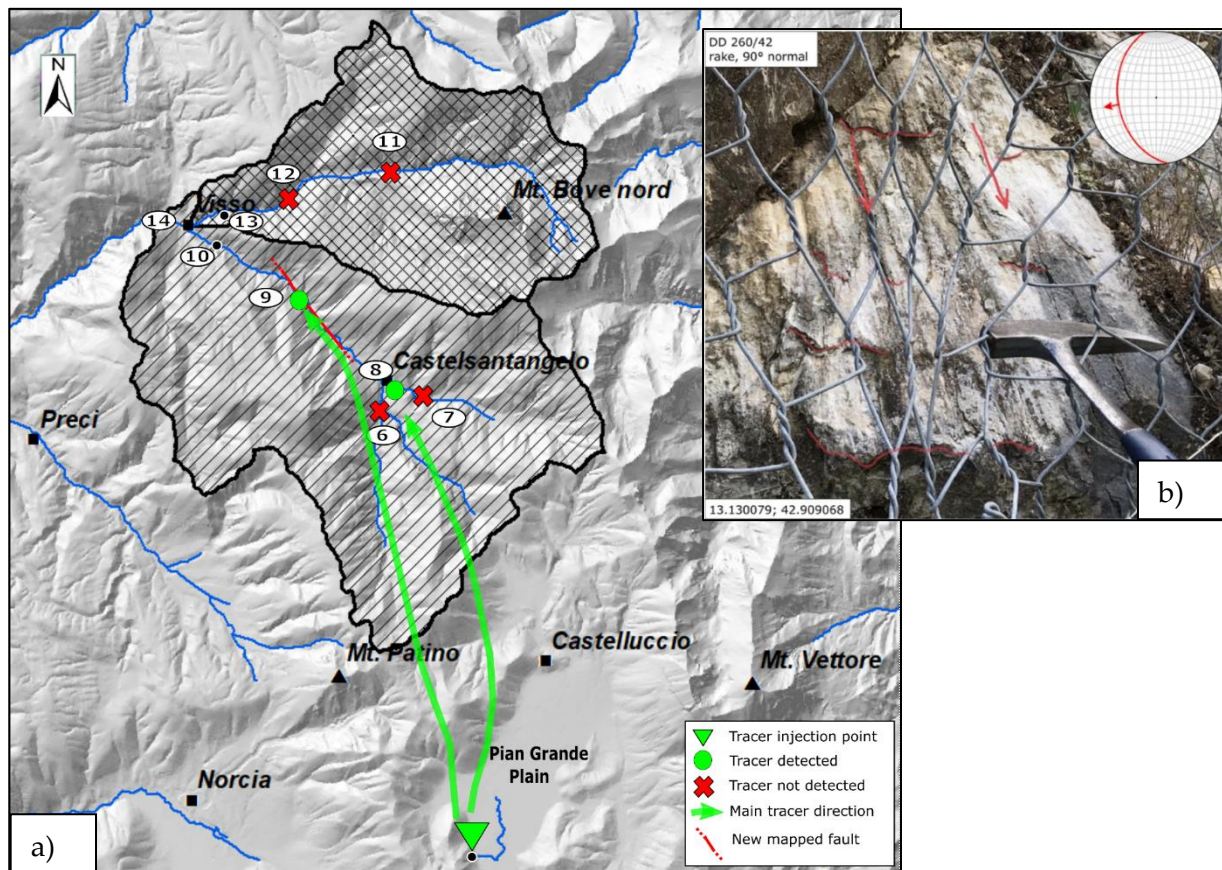


Figure 3.41 (a) Location of the monitoring points, the new mapped fault and main tracer direction from the injection point (Mèrgani sinking stream in Pian Grande Plain) to the analysed basins (Castelsantangelo and Ussita). The green dots refer to the point where the tracer has been recovered (point 8 and 9) while the red crosses refer to the monitoring point where the tracer has not been detected in this study (point 6, 7, 11 and 12). The monitoring point 10, 13 and 14 are not instrumented for tracer detection. (b) Outcrop of the new mapped fault with its kinematics represented in the stereo plot.

3.4.1 Regional groundwater flow modification corroborated by artificial tracer tests

The traced groundwater flow at regional scale, characterising the pre- and post-seismic phases, are synthetized in Figure 3.42 and Figure 3.43 respectively. The pre-seismic condition is the result of the observations obtained from TEST1 and TEST2 conducted by Nanni et al 2020, while the post-seismic condition refers to the TEST4 and TEST5, for which a more complete monitoring network is present.

The general outcomes such as the percentage of tracer recovery mass during the pre- and post-seismic tracer tests, calculated by Qtracer2, is reported in the relative figure. Each monitoring point is marked with a symbol to identify the type of instrument used to determine the arrival or absence of tracer (i.e., fluorimetric probe, water auto-sampler or active carbon trap).

During the pre-seismic phase, groundwater traced flows were predominantly North-South oriented (Apenines flow direction) and mainly compartmentalised in the central hydrostructure (hydrostructure 2-2a) bounded by the Mt. Vettore-Mt. Bove normal fault to the east and by the Nottoria-Preci fault system to the west. Only minor groundwater exchanges in the East-West direction (anti-Apenines flow direction) have been recorded, among these, those going from hydrostructure 2 towards hydrostructure 3 seem to be more significant than those going from hydrostructure 2 towards hydrostructure 1.

After the seismic sequence, the groundwater flowpaths experienced significant modification. The predominantly North-South oriented flow direction remains, but most of the tracer is recovered in the springs of hydrostructure 3 located westward with respect to the Nottoria-Preci normal fault. The hydraulic connection between the Pian Grande Plain and the springs located southward is weakened in favour of those located northward and westward with respect to the injection point. Minor traced water exchanges still remains between hydrostructure 2 and hydrostructure 1.

Details of post-seismic tracer tests (TEST4 and TEST5), such as the time of first tracer arrival and mean tracer transit time together with the percentage of tracer recovery mass for each monitored point is reported in Table 3.23.

Pre-seismic

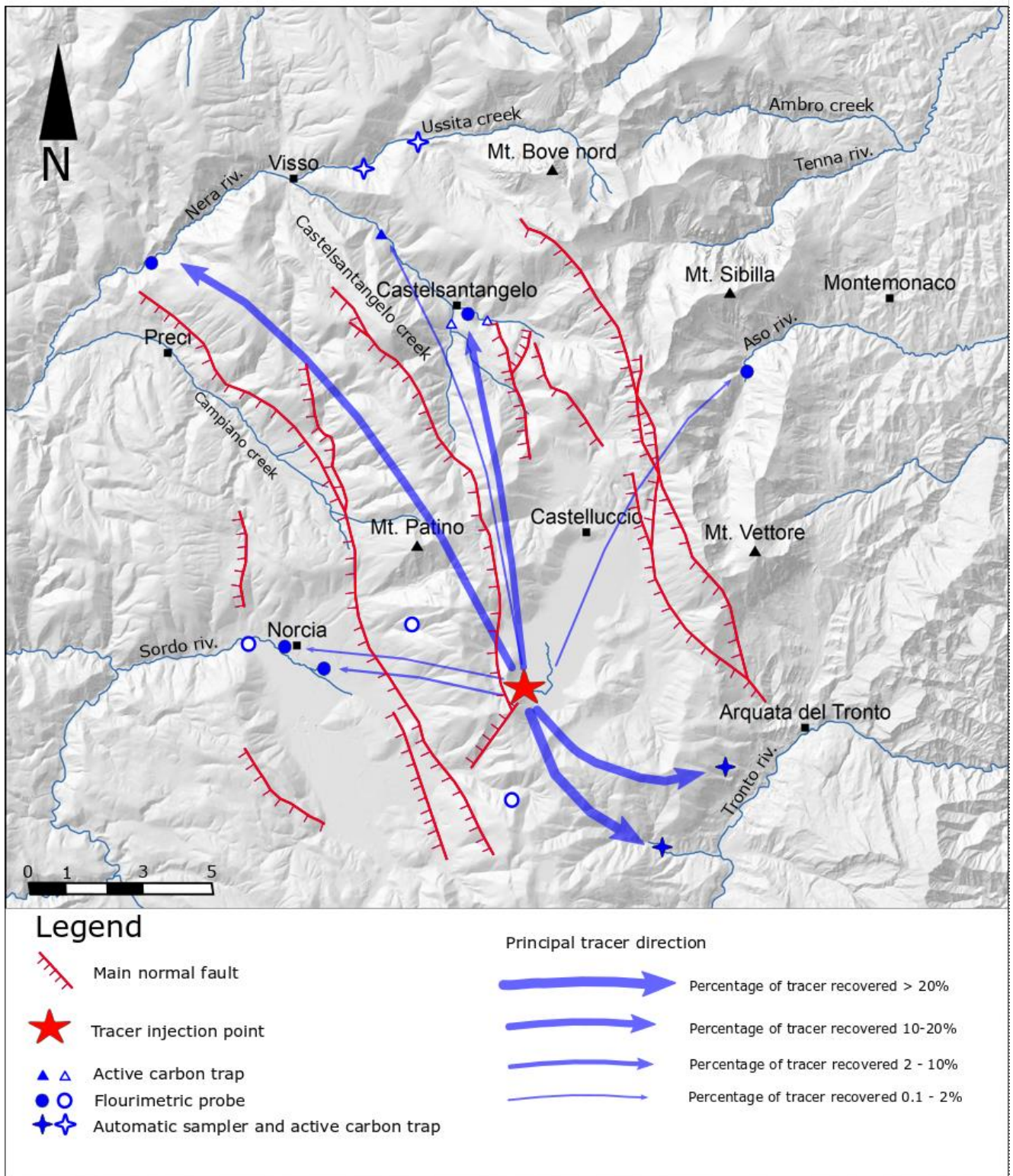


Figure 3.42 Illustration of pre-seismic (TEST1 and TEST2) traced groundwater flow directions from the injection point (Mèrgani sinking stream). The monitoring points show the filed instrument to detect tracer arrivals. The empty symbols refer to the pints where tracer has never been recorded. The results are synthetized from the data published by Nanni et al., 2020 and integrated by the results analysed in this study.

Post-seismic

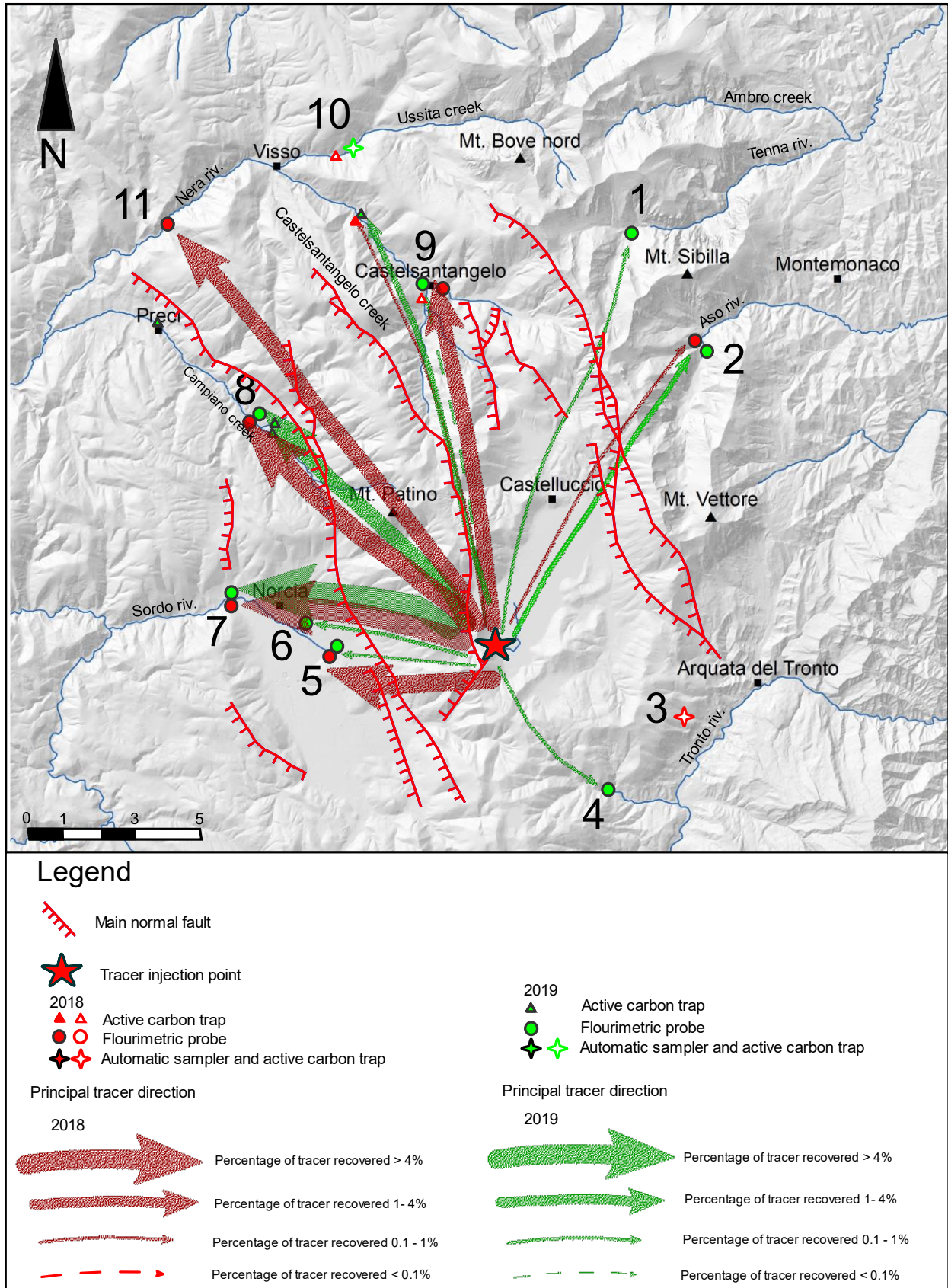


Figure 3.43 Illustration of post-seismic (TEST4 and TEST5) traced groundwater flow directions from the injection point (Mèrgani sinking stream). The monitoring points show the filed instrument to detect tracer arrivals. The empty symbols refer to the points where tracer has never been recorded.

Table 3.23 Results of TEST4 and TEST5 obtained by Qtracer2, the numbers refer to the ones in plate of Figure 3.43.

Name	N° in map	Distance from injection point (m)	T ₁ (d) TEST 5	T ₁ (d) TEST 4	T _g (d) TEST 5	T _g (d) TEST 4	T _{pp} (d) TEST 5	T _{pp} (d) TEST 4	$\bar{v}$ (m/d) TEST 5	$\bar{v}$ (m/d) TEST 4	Recovered mass (%) TEST 5	Recovered mass (%) TEST 4
Capotenna spring	1	1550	52	-	151	-	179	-	102	-	0.61	-
Foce spring	2	12600	94	45	174	49	-	37	72	320	1.21	0.5
Pescara spring	3	6750	-	n.d.	-	n.d.	-	n.d.	-	n.d.	-	0.0
Capodacqua spring	4	6300	112	-	230	-	261	-	27	-	0.9	-
Torbidone spring	5	6000	46	41	116	126	142	83	51	48	0.1	14.7
San Martino spring	6	7000	60	-	174	-	235	-	40	-	0.88	-
Sordo linear spring	7	8800	32	8	193	52	170	118	45	84	92.2	46.9
Campiano at Campi	8	10600	81	45	307	60	140	55	51	178	4.1	6.3
Nera Castelsantangelo	9	12600	31	68	129	124	129	130	98	101	0.001	12.1
Ussita Madonna dell'Uccelletto	10	18000	n.d.	-	n.d.	-	n.d.	-	n.d.	-	0.0	-
Nera I Molini	11	18800	-	38	-	106	-	110	-	175	-	19.5

T₁= time of first arrival; T_g= barycentric time (corresponding to the 50% of tracer recovered mass in the same point); T_{pp}= time of principal peak arrival; $\bar{v}$ = mean tracer velocity (corresponding to barycentric time/distance from injection point); Recovered mass for each point (%) refers to the total recovered mass during the same tracer test for all points; n.d = tracer not detected; - = fluorimetric probe or automatic sampler not in function.

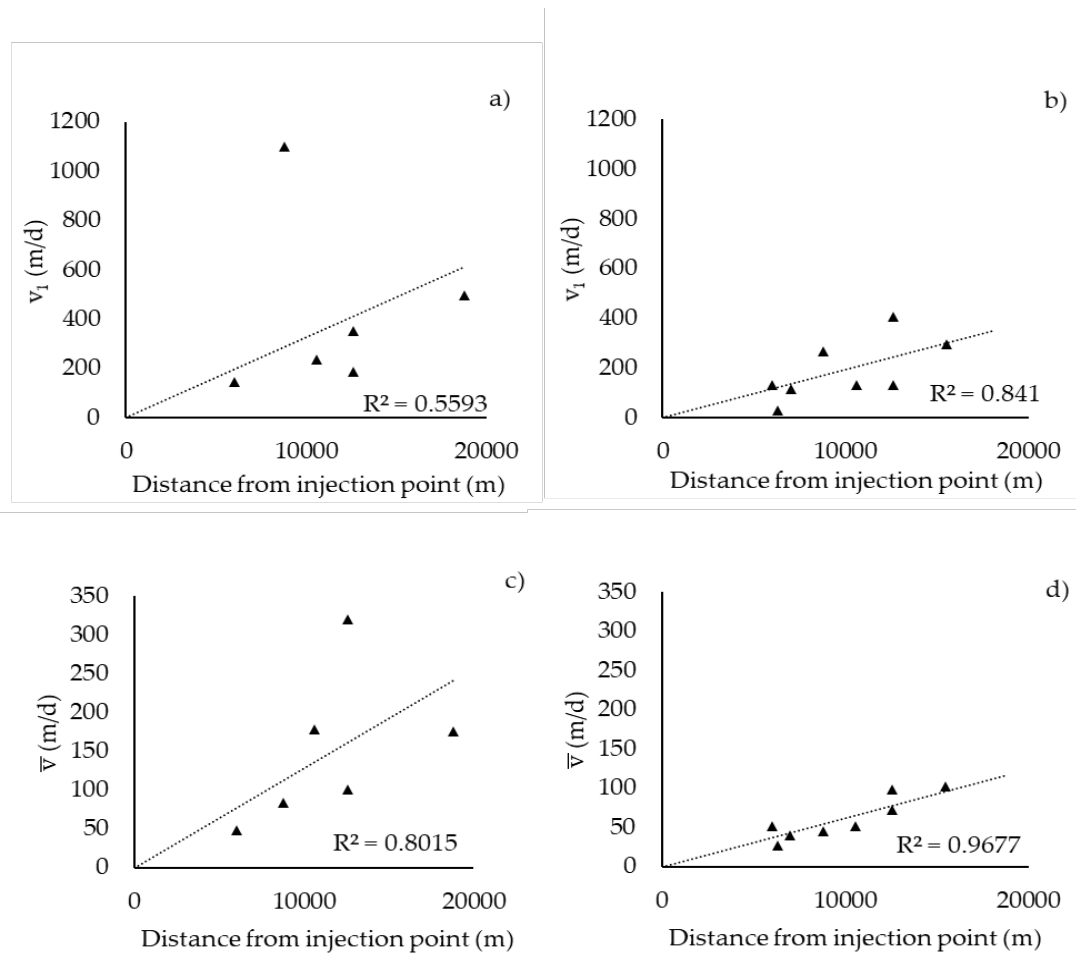


Figure 3.44 (a) Maximum tracer velocity vs distance from injection point for TEST4; (b) Maximum tracer velocity vs distance from injection point for TEST5; (c) Mean tracer velocity vs distance from injection point for TEST4; (d) Mean tracer velocity vs distance from injection point for TEST5.

TEST4 is characterised by a total tracer recovered mass of approximately 4 kg corresponding to 25% of the total mass injected into the Mèrgani sinkhole. This recovery mass is moderately low, but considering a regional long-period tracer tests, involving an area of more than 600 km² it represents a good percentage of tracer recovery mass (Sanford et al., 2006). In this context it must be highlight that, for objective reasons, not all the springs emerging from the Sibillini Mt. domain were monitored due to the unavailability of an adequate number of continuous fluorometric probes and the absent way to physically reach some springs. In addition, manually sampled points were not included in the calculation and nor San Martino spring nor Capodacqua spring were monitored.

TEST5 is characterised by a total tracer recovered mass of approximately 20.31 kg corresponding to 75.2% of the total mass injected. In this case a more complete monitoring network permitted to recover a higher percentage of tracer. Also in this case manually sampled points, such as point 9 in the Upper Nera River basin, were not included in the calculation. For this reason, the recovered mass is definitely underestimated in both the artificial tracer tests.

During TEST4 more than 46% of the recovered mass has been measured in Sordo linear spring and only the monitored springs located western with respect to the Nottoria-Preci fault (Sordo, Torbidone and Campiano) recovered nearly 68% of the total recovered mass. Also during TEST5 the springs located western with respect to the study area (hydrostructure 3) recovered almost the entirety of the injected tracer (97.3%), of which 92.2% was recorded exclusively by the Sordo linear spring.

The graphs in figure 3.44 show the maximum (v_1) and mean ($\bar{v}$) tracer velocities during TEST4 (Figure 3.43 a and c) and TEST5 (Figure 3.44 b and d). As TEST4 concerns, both v_1 and $\bar{v}$ are more scattered than the one calculated for TEST5. As matter of fact, v_1 for TEST4 ranges between 146 m/d (Torbidone spring) to 1100 m/d (Sordo linear spring), while for TEST5 it ranges between 30 m/d (Capodacqua spring) up to 406 m/d (Foce spring). $\bar{v}$ instead, varies between 48 m/d (Torbidone spring) to 320 m/d (Foce spring) during TEST4 and between 27 m/d (Capodacqua spring) to 102 m/d (Capotenna spring).

3.4.2 Comparison between periodic tracer tests and time-series analysis

Specific investigations have been performed at local scale in those springs characterised by a most complete dataset of tracer tests and discharge data (i.e., Pescara del Tronto, Capodacqua and Foce springs).

Figure 3.45 illustrates the daily tracer maximum concentration and the daily rainfall recorded at Mt. Prata rainfall station for each tracer test and spring. This representation provides an innovative alternative to highlight the correlation between tracer arrivals and the rainfall events. Interestingly, this correlation showed a delay between rainfall and tracer arrivals. The delay is interpreted as the travel time of the tracer comprises both the infiltration in the unsaturated zone and the tracer movement in the saturated portion of the aquifers (Mudarra & Andreo, 2011). These graphs provide two different kinds of information: i) the differences in the delay between injection, rainfall events, and tracer arrivals in the springs before (TEST1), during (TEST2), and after the seismic sequence (TEST3, TEST4, and TEST5); ii) the hydrogeological relationships between the examined springs.

Before the main earthquakes, a generally constant delay time in the arrival of the tracer after rainfall for Pescara del Tronto and Capodacqua springs of about 10 and 8 days, respectively, can be highlighted. As Pescara del Tronto spring concerns, this aspect is documented by the tracer arrival located at about 10 days after the rainfall event, which is occurred after 3 days from the injection (TEST2). A subsequent peak is also visible at 23 days, and it is likely correlated to the second intense rainfall event (>20 mm), recorded after 10 days from the injection. Regarding Capodacqua spring, there is a strong similarity in tracer arrivals between TEST1 and TEST2, but with respect to Pescara del Tronto spring, the delay time between rainfall events and tracer arrivals is slightly reduced to 8–9 days. As matter of fact, after the rainfall events occurred 3, 10, and 36 days from the injection (TEST2), the tracer arrived at 12, 19, and 44 days after the injection. This last evidence is clearly visible in TEST2 and confirmed by TEST1. After the seismic events, the tracer did not arrive in Pescara del Tronto at all, while in Capodacqua spring a specific rainfall event seems to be not related to the tracer arrivals during TEST3. Analysing TEST5, a lag of 44–48 days is systematically recognised between intense rainfall events (>50 mm) and the tracer arrivals in the spring.

A completely different behaviour has been observed in Foce spring. Before and during the co-seismic period, a specific rainfall event seems to be not related to the tracer arrivals, while the relation is

much more appreciable after the seismic period especially in TEST4 and TEST5. In these tests the tracer arrives at the springs 12 days after the main rainfall events.

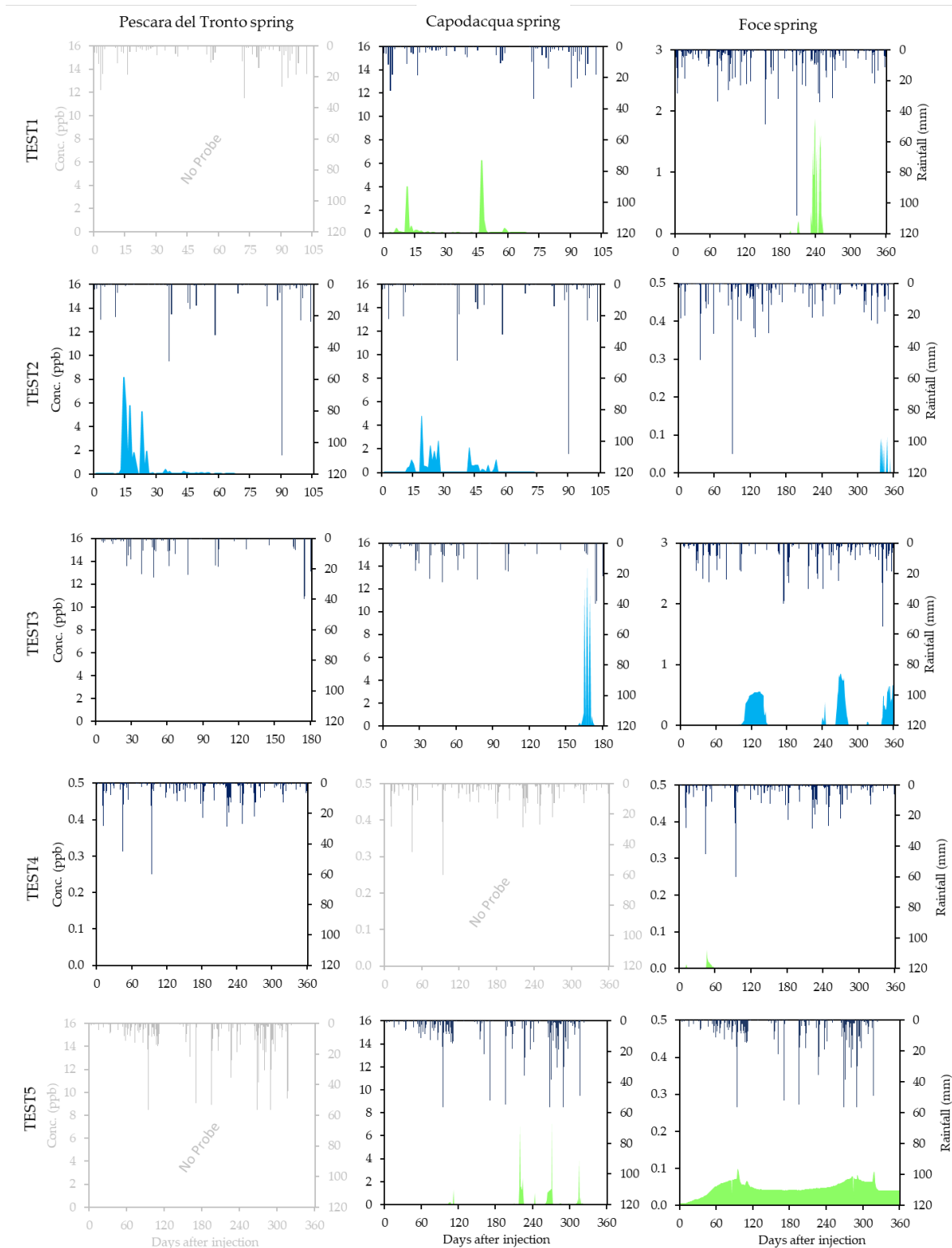


Figure 3.45 Comparison between maximum tracer concentration expressed in ppb vs days after injection in the monitoring points (Pescara del Tronto, Capodacqua and Foce springs) and rainfall recorded at Monte Prata during the five analysed tests. The green area represents the Fluorescein, while the blue one the Tinopal CBS-X. When the fluorometric probe was not installed in a specific spring during a specific test, the corresponding graph is represented in light grey, and the expression “No Probe” is added in the figure.

The Sliding Windows cross correlation function (SWCCF) results related to the Mt. Prata snow depth/rainfall station, considered as the most representative meteo-climatic station for the Pian Grande Plain are shown in Figure 3.46 and Figure 3.47. To avoid misinterpretation (e.g., false response times) of rainfall/residual spring discharge relationships for the main recharge area, the SWCCF analysis has been performed also considering the rain gauges located near each spring (Capodacqua rainfall station for Pescara and Capodacqua springs, and Montemonaco rainfall station for Foce spring). This has allowed a visual filtration of results shown in (Figure 3.47). Furthermore, it is important to highlight that all the discharge responses to the considered inflows represent the piezometric perturbations, due to the recharge arrival in the saturated zone (i.e., pressure transfer); thus, these do not necessarily include the recharge water transport through the saturated zone.

As already seen for other wide mountain carbonate aquifer springs, recharged mainly from snowmelt infiltration (Chiaudani et al., 2019), Pescara del Tronto and Capodacqua springs show a strong response to snowmelt inflow, which is represented by the high correlation values (Figure 3.46). However, this clear response has been pointed out mainly for the pre- (i.e., windows 1 and 2) and post-seismic (i.e., window 8 and 9) periods; even though it is evident also in the first and last windows of the co-seismic period that can be considered as transition periods. In windows that fall in the central part of the co-seismic period (i.e., window 4, 5 and 6), the Pescara del Tronto and Capodacqua discharges seem to totally decorrelate from snowmelt recharge, because of the transient and generalized hydraulic head increase related to the pore pressure propagation following strong seismic events (Amoruso et al., 2010). In this case, the seasonal recharge effect on the piezometric fluctuation seems to be completely hidden by the seismically induced pore pressure propagation perturbation. During the post-seismic period, higher correlation coefficients (i.e., CCF range is 0.4–0.6 before and 0.7–0.9 after the seismic sequence) between the Pescara del Tronto and Capodacqua spring discharges and the snow cover thickness suggest a stronger influence of the seasonal recharge on these spring behaviours. Furthermore, response times increase both for Pescara del Tronto (i.e., from about 110 to about 130 days) and Capodacqua spring (i.e., from about 100 to about 130 days). On the other hand, Foce spring shows a very limited influence of the seasonal snowmelt inflow on the spring behaviour. In windows 1 and 2 of SWCCF between Foce spring discharges and the snow cover thickness, the maximum significant correlation value is in the range 0.3–0.4, with a time lag of about 160 days. During the co- and post-seismic periods, the Foce discharge pattern seems no longer

influenced by the seasonal snowmelt inflow, except for a slightly significant response with an unclear lag ranging between 60 and 90 days.

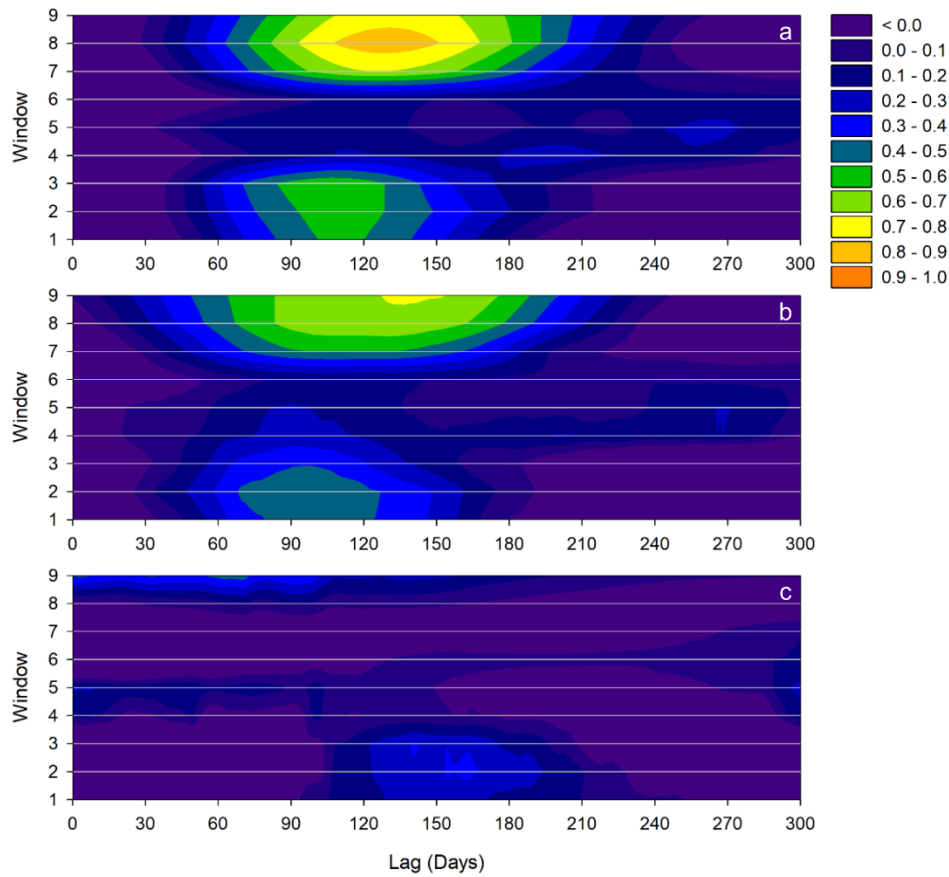


Figure 3.46 Contour graphs that represent variations in the curve shape of the snow cover/raw spring discharge CCFs, over the whole period (i.e., among the sliding windows shown in (Table 2.8). (a) Mt. Prata snow depth/Pescara del Tronto spring raw discharge (b) Mt. Prata snow depth /Capodacqua spring raw discharge (c) Mt. Prata snow depth /Foce spring raw discharge. The positive correlation value corresponding to the 95% significance level ($p < 0.05$) is 0.06.

Comparing the SWCCF results of Mt. Prata rainfall and the residual spring discharge (Figure 3.47) with the local rainfall and the residual spring discharge (Figure 3.48), it appears clear that the quickest responses of residual discharge to rainfall inflow are attributable to the infiltration in the Pian Grande Plain recharge area (Mt. Prata rainfall). This evidence is demonstrated by the high-elevation rainfall/residual spring discharge SWCCFs, that show correlation values higher than the low-elevation rainfall/residual spring discharges' ones. As clearly represented in (Figure 3.47), all of the springs show highly different response patterns to the rainfall inflow. Before the seismic events, Pescara del Tronto residual discharge shows two slightly significant response peaks at about 15 and 33 days, respectively. During the co-seismic period, a strong quick response has been pointed out (i.e., lag $\sim 1-2$ days, with a correlation value > 0.200). At the end of the considered period (i.e., the

post-seismic), only a clear 38-day response time (i.e., correlation value > 0.200) has been detected by the rainfall/residual spring discharge SWCCF. From a hydrodynamic point of view, changes in the Mt. Prata rainfall/Pescara del Tronto residual discharge relationship over the considered period can be related to variations in the overall piezometric level, that have affected the recharge modes.

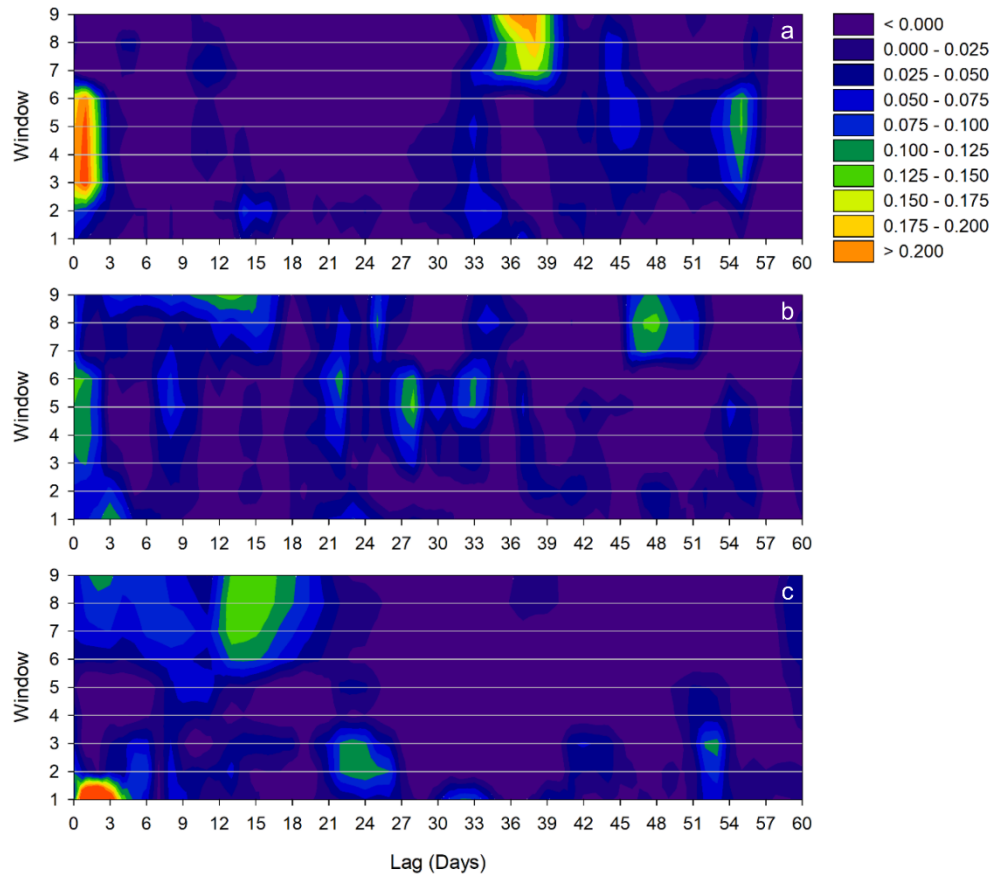


Figure 3.47 Contour graphs that represent variations in the curve shape of the high-elevation rainfall/residual spring discharge CCFs, over the whole period (i.e., among the sliding windows shown in (Table 2.8). (a) Mt. Prata rainfall/Pescara del Tronto spring residual discharge (b) Mt. Prata rainfall /Capodacqua spring residual discharge (c) Mt. Prata rainfall /Foce spring residual discharge. The positive correlation value corresponding to the 95% significance level ($p < 0.05$) is 0.06.

The SWCCF between Mt. Prata rainfall and the Capodacqua residual spring discharge, pointed out a 3-day and a 24-day significant positive response time, during the pre-seismic period. Throughout the co-seismic period, a very quick (i.e., 1 day) and multiple long-period (i.e., 21, 28, and 33 days) response times have been detected by the sliding window analysis. Finally, during the post-seismic period, the response times are at 12 and 45 days.

Finally, the results related to Mt. Prata rainfall and the Foce residual spring discharge sliding window time-series analysis (Figure 3.47) show an evident 2-day response time both in the pre- and

post-seismic period, that disappears in the co-seismic windows (i.e., 3–7). Furthermore, a more retarded significant peak has been detected over the considered time. This has a lag of 24 days in the pre-seismic windows (i.e., 1 and 2), of 10 days in the central window of the co-seismic period (i.e., 5), and 15 days in the last four windows (i.e., 6, 7, 8, and 9).

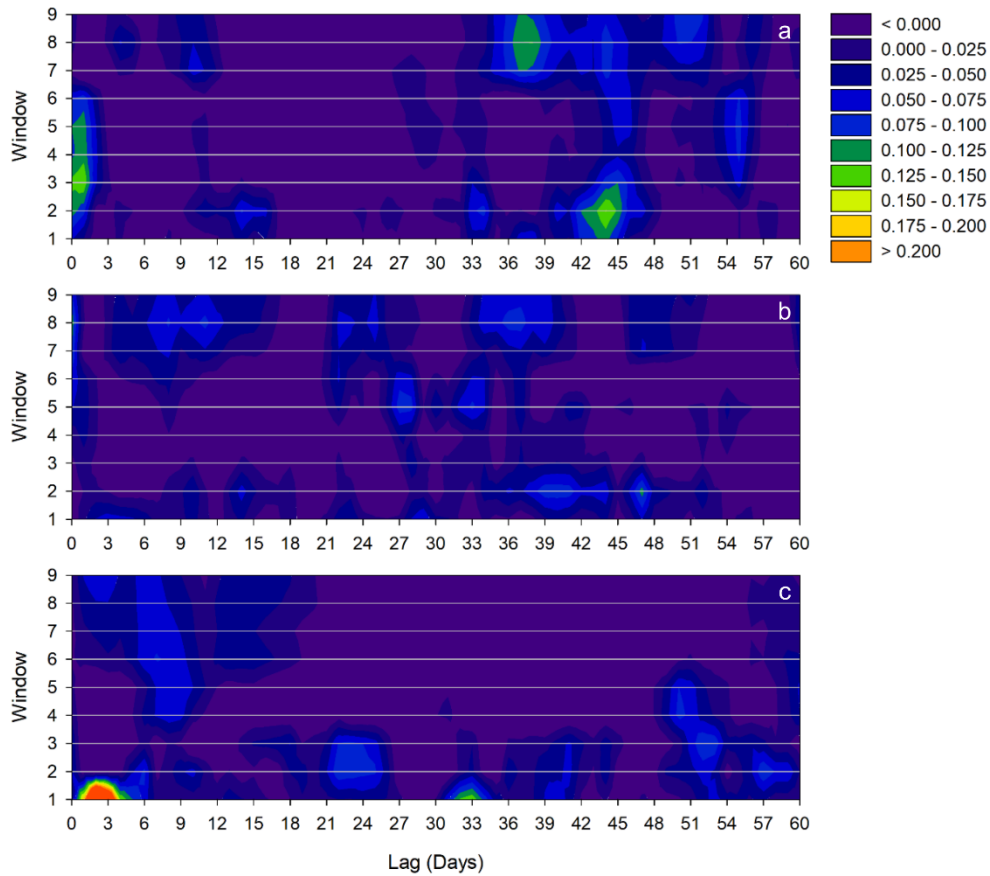


Figure 3.48 Contour graphs that represent variations in the curve shape of the low-elevation rainfall/residual spring discharge CCFs, over the whole period (i.e., among the sliding windows shown in (Table 2.8). (a) Capodacqua rainfall/Pescara del Tronto spring residual discharge (b) Capodacqua rainfall /Capodacqua spring residual discharge (c) Montemonaco rainfall /Foce spring residual discharge. The positive correlation value corresponding to the 95% significance level ($p < 0.05$) is 0.06.

3.5 Isotope hydrology outcomes

Weighing the $\delta^{18}\text{O}$ content obtained for the rainfall samples and for the minor springs water samples with rainfall amount and discharge, the relationship between $\delta^{18}\text{O}$ -elevation has been determined. The isotopic altitude gradient of $0.27\text{‰}/100\text{ m}$ has been calculated by a linear regression of the $\delta^{18}\text{O}$ data vs elevation, showing a strong correlation ($R^2 = 0.81$), as suggested by the Pearson's coefficient of determination. The obtained gradient is slightly major of the one determined by Giustini et al., 2016 for central Italy ($0.20\text{‰}/100\text{ m}$) and by Tarragoni, 2006 for the same area ($0.22\text{‰}/100\text{ m}$). However, both the ANOVA and Kolmogorov-Smirnov tests on the $\delta^{18}\text{O}$ meteoric distribution over the study area demonstrate that the new $\delta^{18}\text{O}$ gradient, actually, is significantly different from the ones available in literature (Figure 3.49a and b), especially to the one valid for central Italy. The comparison between the $\delta^{18}\text{O}$ -elevation relationship obtained in this study and the one obtained by Tarragoni, 2006 is reported in Figure 3.49c.

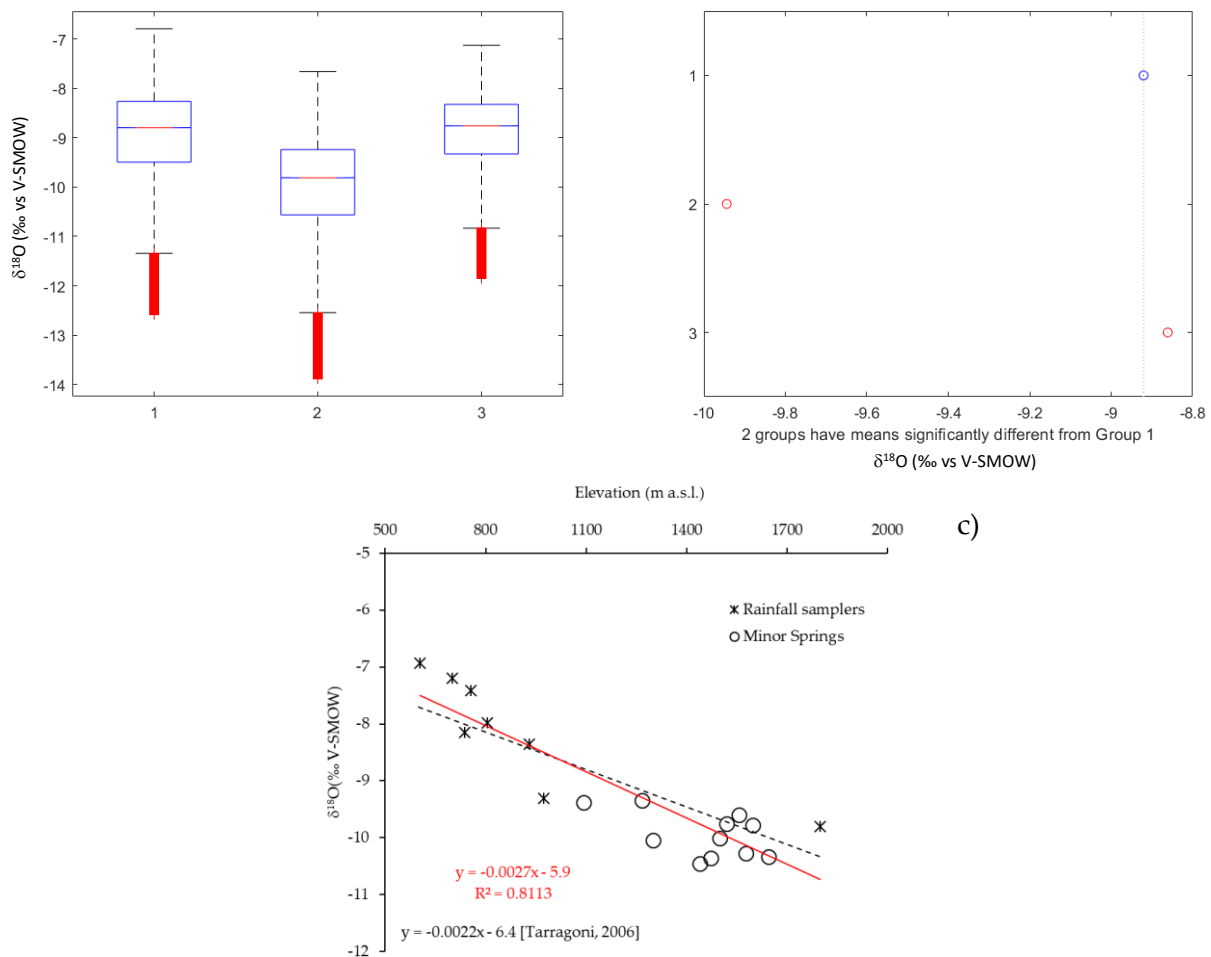


Figure 3.49 (a) Meteoric $\delta^{18}\text{O}$ distribution above the study area: 1- $\delta^{18}\text{O}$ -elevation relation determined in this study, 2- $\delta^{18}\text{O}$ -elevation determined by Giustini et al., 2016, 3- $\delta^{18}\text{O}$ -elevation determined by Tarragoni, 2006. (b) Results of statistic tests on the Meteoric $\delta^{18}\text{O}$ distribution above the study area, group 1 (our relationship) is significantly different from the literature ones (groups 2 and 3). (c) Linear regression between $\delta^{18}\text{O}$ and elevation determined by eight rain gauges' water collectors and eleven minor springs of the area obtained in this study (red solid line) and linear regression obtained by Tarragoni, 2006 for the same area (black dashed line).

The application of the inverse $\delta^{18}\text{O}$ -elevation relation obtained in this study, the one obtained by Giustini et al., 2016 and the one presented by Tarragoni, 2006, to the mean pre-sesimic $\delta^{18}\text{O}$ value of Sordo (20), Pescara (4), Capodacqua (5), Foce (2) and Nera Castelsantangelo (8) springs (n° of data >5) shows significant differences in the springs mean elevation recharge areas (Figure 3.50).

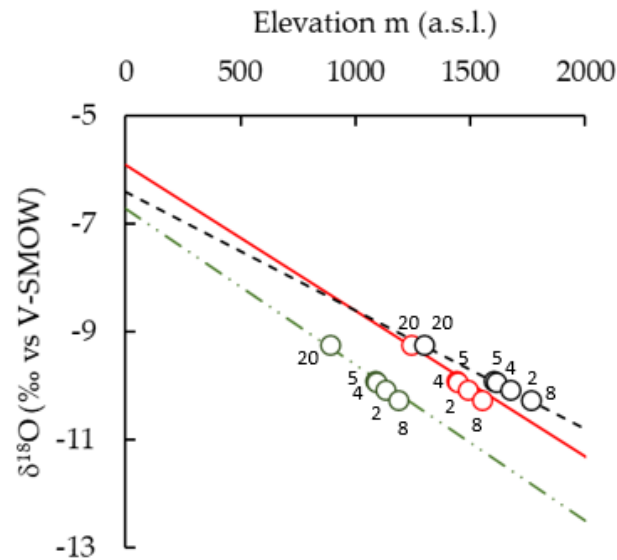


Figure 3.50 Application of the inverse $\delta^{18}\text{O}$ -elevation relation obtained in this study (red line and red dots), the one obtained by Giustini et al., 2016 (green dashed line and green dots) and the one presented by Tarragoni, 2006 (black dashed line and black dots), to the mean pre-sesimic $\delta^{18}\text{O}$ value of Sordo (20), Pescara (4), Capodacqua (5), Foce (2) and Nera Castelsantangelo (8).

This difference is much more significant in the application of the Giustini et al., 2016 relationship, not giving actual indication on the springs mean recharge area elevation. In fact, especially for Pescara and Capodacqua springs, the mean recharge elevation obtained by the $\delta^{18}\text{O}$ -elevation (Giustini et al., 2016 relationship) seems to be coincident with the spring elevation, or slightly higher. This aspect would not be justified by the spring recharge areas topographical elevation upstream of the springs, which is of course, characterised by higher topographic elevation than the spring one.

On the contrary the application of the $\delta^{18}\text{O}$ -elevation relation obtained by Tarragoni, 2006 gives similar results to the application of the relationship obtained from this study.

This result shows that the regional $\delta^{18}\text{O}$ gradient (Giustini et al., 2016) cannot be applied to the areas with complex orography and tectonics (such as mountainous areas with steep slopes having different exposures, or where the springs recharge areas are rather far from the springs itself, and where the geo-structural setting plays a non-negligible role on groundwater paths) but an attempted

evaluation of the local meteoric conditions must be carried out before choosing the location of meteoric sampling points. In addition, it must be stated that the study performed by Giustini et al., 2016 was developed at national scale and did not claim to achieve detailed maps to be used in specific hydrogeological applications.

The $\delta^{18}\text{O}$ gradient obtained by Tarragoni, 2006, on the contrary, show similarities between our results because it was performed at local scale even if limited to the upper Nera River basin.

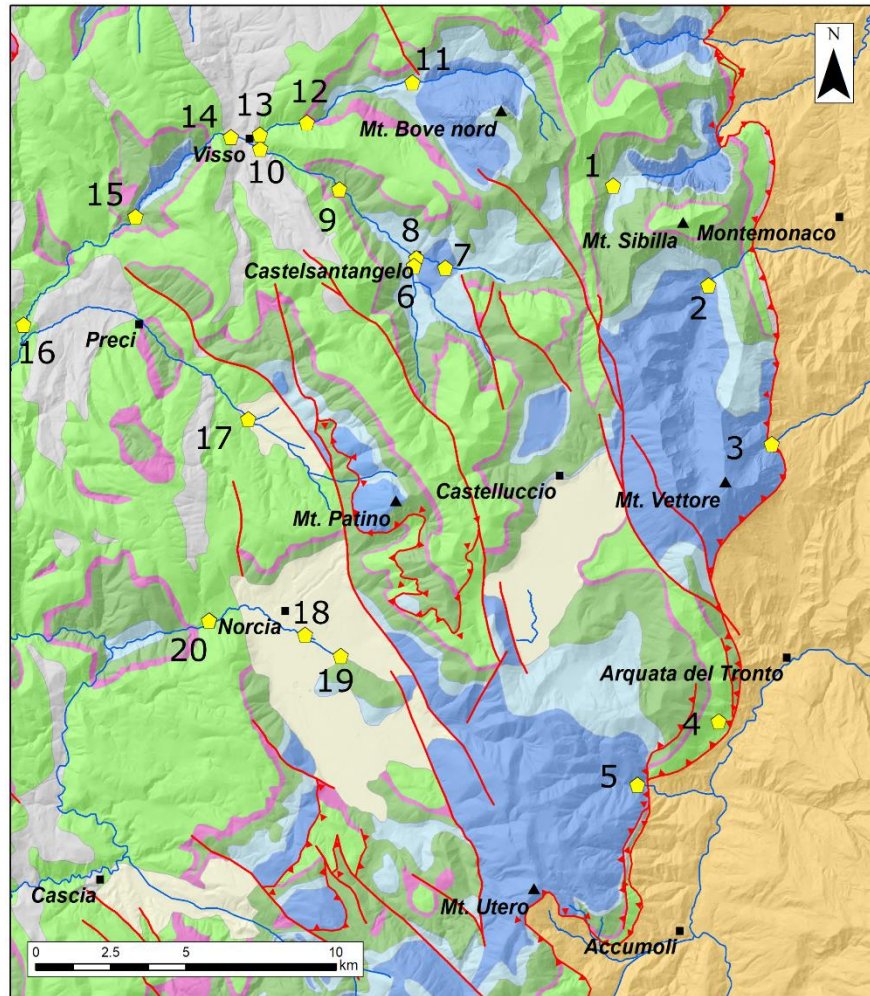


Figure 3.51 Simplified hydrogeological map with isotopic and hydrochemical sampling

The results obtained from the application of the inverse $\delta^{18}\text{O}$ -elevation relation to the $\delta^{18}\text{O}$ springs data, provide further interesting evidence. For example, the Castelsantangelo spring (8), belonging to the Castelsantangelo stream (Upper Nera River), emerging from the **BAS**, appears to have a much higher isotopic recharge area elevation than the mean elevation of the **BAS** outcropping upstream to the same monitoring point. These inconsistencies between hydrogeological features and isotopic elevations were investigated in detail for the Upper Nera Basin both for the Ussita creek and Castelsantangelo creek through a GIS-based analysis.

The latter highlights that the Ussita creek mean elevation is higher respect to the Castelsantangelo creek and so is the overall mean elevation of geological Fms. hosting the main aquifers (**BAS**, **MAI** and **SCA**) formations (Table 3.24). Nonetheless, the three hydrogeological complexes hosting the main aquifers in the U-M Succession (**SCA**, **MAI** and **BAS**) are distributed differently in the two sub-catchments in terms of mean outcropping elevation. In particular, as concerns the Ussita creek, the hydrogeological complex outcropping at the highest elevation (1440 m a.s.l.) is **MAI**, the **BAS** mean elevation is intermediate (1328 m a.s.l.), while the **SCA** mean elevation is the lowest, (1213 m a.s.l.). In the Castelsantangelo creek, the **BAS** mean elevation is the lowest (862 m a.s.l.), the **MAI** mean elevation is intermediate (1278 m a.s.l.) and **SCA** outcrops at the highest mean elevation (1357 m a.s.l.). The higher elevation of the Ussita basin with respect to the Castelsantangelo one, is reflected also on the topographic cross section along the two streams (Figure 3.52). Ussita stream extends from 1225 m to 625 m a.s.l. while the Castelsantangelo stream extends from 1060 m to 625 m a.s.l. As concerns the **BAS**, the Ussita stream intersects it from 1200 m to 790 m a.s.l., at a higher elevation than the Castelsantangelo stream, which intersects the same aquifer from 830 m to 730 m a.s.l. As concerns the **MAI**, the Ussita stream crosses it from 720 to 700 m a.s.l. and from 685 m to 660 m a.s.l. (with a short interruption caused by the stream intersection with the **JUR** aquiclude) while the Castelsantangelo stream intersects it between 711 m and 670 m a.s.l. Therefore, the mean outcrops elevations are quite similar among each other. The **SCA** outcrops along the Ussita stream between 660 m and 620 m a.s.l. while in the Castelsantangelo stream the same hydrogeological complex outcrops from 665 m to 650 m a.s.l., so that the complex mean elevation is slightly higher in the Castelsantangelo catchment, despite its lower mean elevation. Just 500 m before their confluence near the Visso village, both streams intersect the **SVC** aquiclude.

Table 3.24 Mean and standard deviation values for the elevation of each basin (Castelsantangelo and Ussita) and for each aquifer outcropping within the basins determined by the GIS analysis.

	Mean elevation (m a.s.l.)	Std. dev. (m a.s.l.)
Ussita creek		
DTM _{tot}	1291	377
SCA	1213	321
MAI	1440	451
BAS	1328	291
AQUIFERS _{mean}	1327	354
Castelsantangelo creek		
DTM _{tot}	1245	301
SCA	1357	256
MAI	1278	331
BAS	862	71
AQUIFERS _{mean}	1165	219

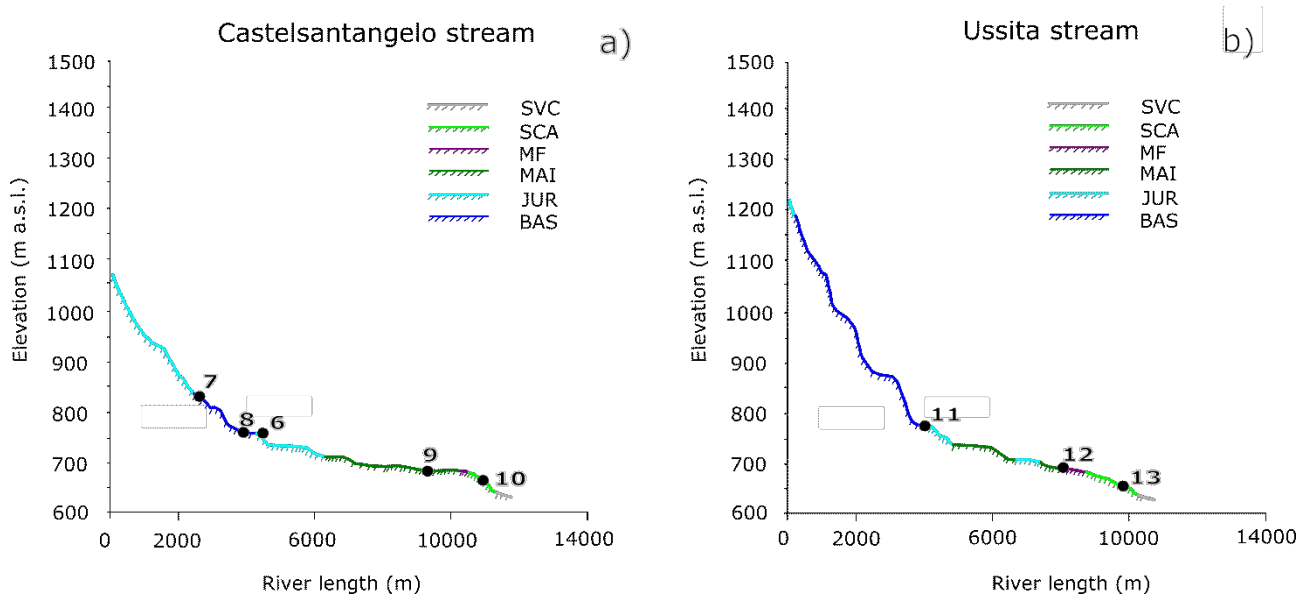


Figure 3.52 Streams topographic intersection with the hydrogeological complexes and localization of the monitoring points for the analyzed basins: (a) Castelsantangelo stream (b) Ussita stream.

In general, before the seismic period, a different $\delta^{18}\text{O}$ content is observed between point 6 and 7, similar among each other, and point 8 even if they all emerge from the **BAS** aquifer complex. In particular, the point 6 and 7 show a more likely local recharge area with a mean elevation of 1356–1379 m a.s.l. with respect to the point 8 which is characterized by a higher recharge area (1545 m a.s.l.). The mean **BAS** elevation outcropping upstream the point 8 in the Castelsantangelo basin, determined by the GIS analysis, is equal to 862 m a.s.l. and this value, if compared with the one determined by the isotopic $\delta^{18}\text{O}$ content, suggests a contribution of groundwater external from its

hydrologic basin. The monitoring point 12 shows a mean $\delta^{18}\text{O}$ content of 10.19‰ suggesting a mean elevation recharge area of about 1526 m a.s.l. During the post-seismic period a more complete dataset allowed for the mean elevation recharge area evaluation for all the monitoring points (Table 3.25). Points 6 and 7 located in the upper part of the Castelsantangelo creek, both emerging from the **BAS**, are characterized by a quite similar $\delta^{18}\text{O}$ content, thus indicating a mean elevation recharge area between 1330 and 1400 m a.s.l., similar to the pre-seismic period. The point 8 is characterized, also during the post-seismic period, by a less enriched $\delta^{18}\text{O}$ content with respect to point 6 and 7, and also in this case a contribution of groundwater from the external area can be supposed. Moving downstream from point 8 to point 10 along the Castelsantangelo creek, the mean elevation recharge areas seem to decrease, being approximately near 1500 m a.s.l., even much higher than the hydrogeological complexes outcropping in the basin. As concerns the Ussita creek, a slight increase in mean recharge elevation area is observed between point 11 and 12, and this result is in accordance with the GIS based analysis showing that the **BAS** mean elevation, upstream to the point 11, outcrops at 1328 m a.s.l. while the **MAI** mean elevation, upstream to the point 12, outcrops at 1440 m a.s.l. Moving downstream from point 12 to point 13 a decreasing in $\delta^{18}\text{O}$ content shows a lower mean elevation recharge area. Also in this case the isotopic result is in accordance with the GIS based analysis (Table 3.24).

Table 3.25 Pre- and post-seismic SO₄ 2- and $\delta^{18}\text{O}$ ($\pm \sigma$) with the elevation determined by the $\delta^{18}\text{O}$ -elevation obtained in this study. The trend column refers to the variation of the aforementioned parameters between the pre- and post-seismic conditions: + sulphates increase in the post seismic, - sulphates decrease in the post-seismic, $\uparrow$ recharge area mean elevation increase, $\downarrow$ recharge area mean elevation decrease, = sulphates content or recharge area mean elevation not varied between the pre- and post-seismic condition.

No. on Map	Pre-seismic period			Post-seismic period			Trend	
	SO ₄ ²⁻ $\pm \sigma$ (mg/L)	$\delta^{18}\text{O} \pm \sigma$ (‰ V-SMOW)	Elevation (m a.s.l.)	SO ₄ ²⁻ $\pm \sigma$ (mg/L)	$\delta^{18}\text{O} \pm \sigma$ (‰ V-SMOW)	Elevation (m a.s.l.)	SO ₄ ²⁻	Elevation
1	3.58 $\pm$ 0.15	-10.09 $\pm$ 0.18	1495	-	-9.73 $\pm$ n.a.	1388		$\downarrow$
2	15.8 $\pm$ 8.73	-10.06 $\pm$ 0.95	1487	21.4 $\pm$ 6.24	-10.2 $\pm$ 0.11	1529	+	$\uparrow$
3	3.40 $\pm$ n.a.	-10.19 $\pm$ 0.31	1526	-	-	-		
4	1.72 $\pm$ 0.24	-9.92 $\pm$ 0.39	1445	1.70 $\pm$ 0.21	-10.21 $\pm$ 0.13	1533	=	$\uparrow$
5	4.93 $\pm$ 1.03	-9.9 $\pm$ 0.66	1440	5.34 $\pm$ 1.38	-10.34 $\pm$ 0.27	1571	+	$\uparrow$
6	1.43 $\pm$ 0.06	-9.62 $\pm$ 0.07	1356	-	-9.57 $\pm$ 0.05	1341		=
7	1.69 $\pm$ 0.18	-9.65 $\pm$ 0.07	1379	1.59 $\pm$ 0.15	-9.63 $\pm$ 0.15	1361	-	$\downarrow$
8	23.1 $\pm$ 3.73	-10.26 $\pm$ 0.15	1545	24.8 $\pm$ 3.26	-10.35 $\pm$ 0.23	1575	+	$\uparrow$
9	12.6 $\pm$ 0.17	-	-	24.4 $\pm$ 2.33	-10.17 $\pm$ 0.25	1519	+	
10	-	-	-	16.87 $\pm$ n.a.	-10.04 $\pm$ 0.33	1482		
11	3.34 $\pm$ 0.06	-	-	2.73 $\pm$ n.a.	-10.01 $\pm$ n.a.	1473	-	
12	11.2 $\pm$ 0.23	-10.19 $\pm$ 0.11	1526	24.2 $\pm$ 3.1	-10.07 $\pm$ 0.33	1490	+	$\downarrow$
13	-	-	-	22.1 $\pm$ 1.32	-9.95 $\pm$ 0.18	1455		
14	-	-	-	17.9 $\pm$ 0.75	-9.99 $\pm$ 0.03	1469		
15	17.9 $\pm$ 0.75	-9.91 $\pm$ 0.13	1442	19.15 $\pm$ 0.92	-9.83 $\pm$ 0.26	1420	+	=
16	-	-9.90 $\pm$ 0.04	1438	18 $\pm$ 0.18	-9.87 $\pm$ 0.02	1431		=
17	1.25 $\pm$ 0.04	-		3.61 $\pm$ 0.36	-9.82 $\pm$ 0.18	1416	+	
18	120.7 $\pm$ 10.8	-9.35 $\pm$ 0.04	1275	124.1 $\pm$ 20.2	-9.71 $\pm$ 0.16	1383	+	$\uparrow$
19	-	-		80.9 $\pm$ 29.6	-9.82 $\pm$ 0.18	1415		
20	-	-9.23 $\pm$ 0.25	1240	86.65 $\pm$ 2.53	-9.51 $\pm$ 0.16	1324		$\uparrow$

The application of the inverse $\delta^{18}\text{O}$ -elevation relation obtained in this study to the mean $\delta^{18}\text{O}$ springs data, permitted to compare, where presents, pre- and post-seismic recharge areas mean elevation determined by the isotopic data.

The springs located in hydrstructure 1 show the higher recharge areas mean elevation (points 1, 2, 3) with respect to the other springs, except for point 8 and 12 belonging to the Upper Nera River basin. This aspect is easily correlated with the geological setting of the springs recharge area (located between the Mt. Vettore- Mt. Bove normal fault to the west and the Sibillini overthrust to the east) characterized by a topographic elevation comparable with the average recharge area determined by the mean $\delta^{18}\text{O}$ springs content. The Capotenna spring (point 1) shows a decrease in its mean recharge area elevation during the post-seismic phase, passing from 1495 m a.s.l. to 1388 m a.s.l.

Foce spring (point 2) instead, show a slight increase in its mean recharge area elevation (however not significant, see paragraph below).

Both the springs located in hydrostructure 2a (points 4 and 5) show a slight increase in their post-seismic recharge area mean elevation determined by the isotopic data. In particular, Pescara del Tronto spring (point 4) passes from 1445 m a.s.l. to 1533 m a.s.l. and Capodacqua spring (point 5) passes from 1440 m a.s.l. to 1571 m a.s.l. during the post-seismic phase.

The springs located along the Nera River downstream of the Visso village (points 14, 15 and 16) do not show significant variations in their isotopic content with an average recharge area elevation around 1435 m a.s.l.

The hydrogeological system characterized by Sordo River- San Martino spring and Torbidone spring (isotopic data not available during the pre-seismic period for the latter) belonging to the hydrostructure 3 (points 18, 19 and 20), shows a mean recharge area elevation much higher than the mean main aquifers elevation outcropping within its hydrological basin upstream of the same springs, suggesting an external groundwater supply, probably coming from hydrostructure 2 (E-W oriented). The same springs show a sharp increase in their mean recharge area elevation of more than one hundred meters during the post-seismic phase.

The Campiano stream, monitored in point 17 (hydrostructure 3), for which no pre-seismic isotopic data are presents, suggests during the post-seismic period a mean recharge area elevation higher than the average outcrop elevation of the main aquifers within its hydrological basin upstream of the spring. Also in this case an external groundwater supply, probably coming from hydrostructure 2, can be assumed.

The decrease in the mean $\delta^{18}\text{O}$ isotopic and thus an apparent increase in the springs mean recharge area elevation is accompanied by an increase in the SO_4^{2-} content (monitoring points 2, 5, 8, 18) except for point 12 where the increase in SO_4^{2-} is related to a decrease in the spring recharge area mean elevation.

In one case the decrease of mean recharge area elevation is accompanied by a decrease in SO_4^{2-} content (point 7). Point 4 is characterised by an increase in recharge area mean elevation with no significant variation in the SO_4^{2-} content during the post-seismic, contrariwise, in point 15 the increase in SO_4^{2-} content is not accompanied by an increase in recharge area mean elevation after the seismic sequence.

The general hydro-chemical composition is bicarbonate-calcium (Figure 3.53) however, three main sub-groups can be identified: (i) the first group with a low sulphate content (points 1, 3, 4, 5, 6, 7, 11

and 17) i.e., the springs having the higher elevation on the Mt. Sibillini domain, or those having their mean recharge area elevation (calculated by their $\delta^{18}\text{O}$ content) not far from their elevation and the springs belonging to hydrostructure 2a characterised by short time between recharge events and discharge response (see SWCCF results) (ii) the second group with slightly higher average sulphates content (points 2, 8, 9, 10, 12, 13, 14, 15 and 16) are those emerging from the **BAS** or others hydrogeological complexes having an interaction with the **BAS** for tectonic or stratigraphic reasons. The latter are likely characterised by long groundwater flow paths, wider recharge areas (such as the spring belonging to the Upper Nera River basin or Foce spring). The third group (iii) to which the springs of the Tyrrhenian side of the Sordo River basin characterized by a higher sulphate content than the others (points 18 during the pre-seismic) and 18, 19 and 20 in the post-seismic phase. The presence of a sulphate rich component can be ascribable to the interaction of groundwater with the upper portion of the Triassic dolomites/evaporites (**TRI**), acting as an aquiclude, underlying the regional basal aquifer (**BAS**) and therefore indicating a deeper groundwater flowpath. In fact, according to the Corine Land Cover 2018 (CLC, 2018) there are no agricultural activities in the area and the presence of some fish farms is not responsible for the increase of the SO_4 concentration as reported in (Arpa U., 2002) so, other sources of sulphates like anthropogenic pollutants can be neglected.

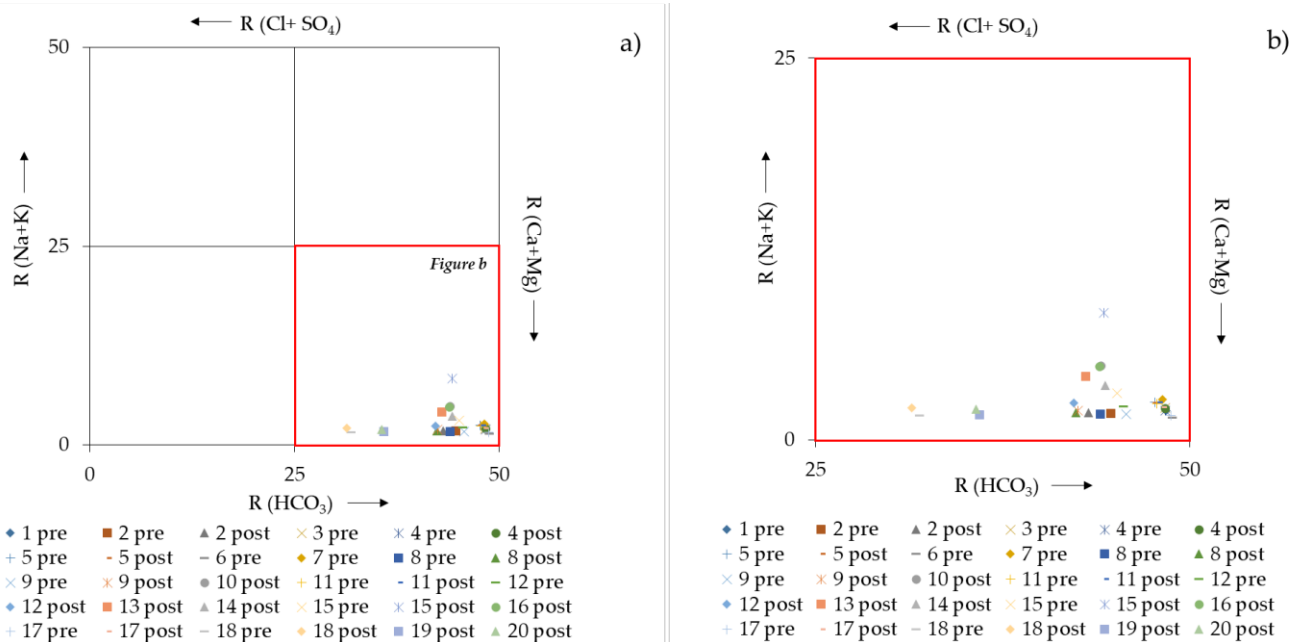


Figure 3.53 Plot of the chemical composition of all the monitoring points on the Langelier-Ludwig diagram. (b) Detail of the Langelier-Ludwig of referring to the box reported in (a).

The pre- and post-seismic $\delta^{18}\text{O}$ mean values shown in Table 3.25 provide a general overlook for the isotopic variations that the springs underwent because of the seismic sequence. A more rigorous method based on the identification of significant abrupt changes in the isotopic signal has been performed in selected springs. Those springs (points 2, 4, 5, 8 and 20) are characterised by a significant $\delta^{18}\text{O}$ number of data both before and after the seismic events and for which water sampling was carried out during several hydrological years following both the recharge and recession periods to avoid sampling bias. The results of the isotopic data processing are shown in Figure 3.54.

Foce spring (point 3) is characterised by 12.9 % (7 over 54) of pre-seismic $\delta^{18}\text{O}$ data outside from the pre-seismic mean $\delta^{18}\text{O} \pm \sigma$ limits, three of them having an enriched isotopic content. The post-seismic $\delta^{18}\text{O}$ data are all comprises between pre-seismic mean $\delta^{18}\text{O} \pm \sigma$ value. As Capodacqua spring concerns (point 5) 8 over 53 pre-seismic $\delta^{18}\text{O}$ data (15 %) exceeds the standard deviation bounds, in particular two of them are less enriched while six are characterized by a more enriched isotopic content. Eight over 15 post-seismic isotopic data (53.3 %) are characterized by a $\delta^{18}\text{O}$ value lower than the lower σ confidence limit. Pescara del Tronto spring (point 4) display 10 over 54 $\delta^{18}\text{O}$ pre-seismic values exceeding the standard deviation bounds, only two of them characterized by a $\delta^{18}\text{O}$ value lower than the lower σ confidence limit. During the post-seismic phase, the 28.6% of $\delta^{18}\text{O}$ spring water samples are characterised by a less enriched isotopic value with respect to the lower σ confidence limit. Castelsantangelo spring (point 8) is marked by 2 over 5 pre-seismic $\delta^{18}\text{O}$ water samples value outside from the pre-seismic mean $\delta^{18}\text{O} \pm \sigma$ boundaries. During the post-seismic phase 6 over 14 data exceed the $\delta^{18}\text{O}$ confidence limits, four of them (28.6 %) are lower than the lower σ confidence limit. As far Sordo spring concerns (point 20) 1 over 5 pre-seismic $\delta^{18}\text{O}$ water samples value is outside from the pre-seismic mean $\delta^{18}\text{O}$ confidence limit (the upper limit), while during the post-seismic period 5 over 7 (71.4 %) of their samples show $\delta^{18}\text{O}$ value lower than the lower σ confidence limit. In few cases, the water samples (those collected simultaneously in Foce, Pescara and Capodacqua springs between September and October 2004) show a peculiar isotopic enrichment probably linked to particular meteoric recharge conditions.

For the same springs, a slight seasonal isotopic oscillation is observed during the pre-seismic period. The less enriched isotopic content is recorded in late winter, then starting from the spring until late summer-early autumn (July-August-September) the $\delta^{18}\text{O}$ values become less negative, after which a downward trend in $\delta^{18}\text{O}$ values is recorded.

The isotopic data analysis carried out with findchangepts function in MATLAB® (Trauth, 2017), identifies for Foce, Capodacqua and Pescara springs a period of abrupt change in the isotopic signal in the pre-seismic period (between September and October 2004) when the isotopic enrichment was recorded. By adding the post-seismic isotopic data Foce does not identify any other abrupt change in the signal, while Capodacqua and Pescara del Tronto springs show a significant change (decrease in the $\delta^{18}\text{O}$ values). The same behaviour is evidenced by the Nera San Chiodo-Castelsantangelo and

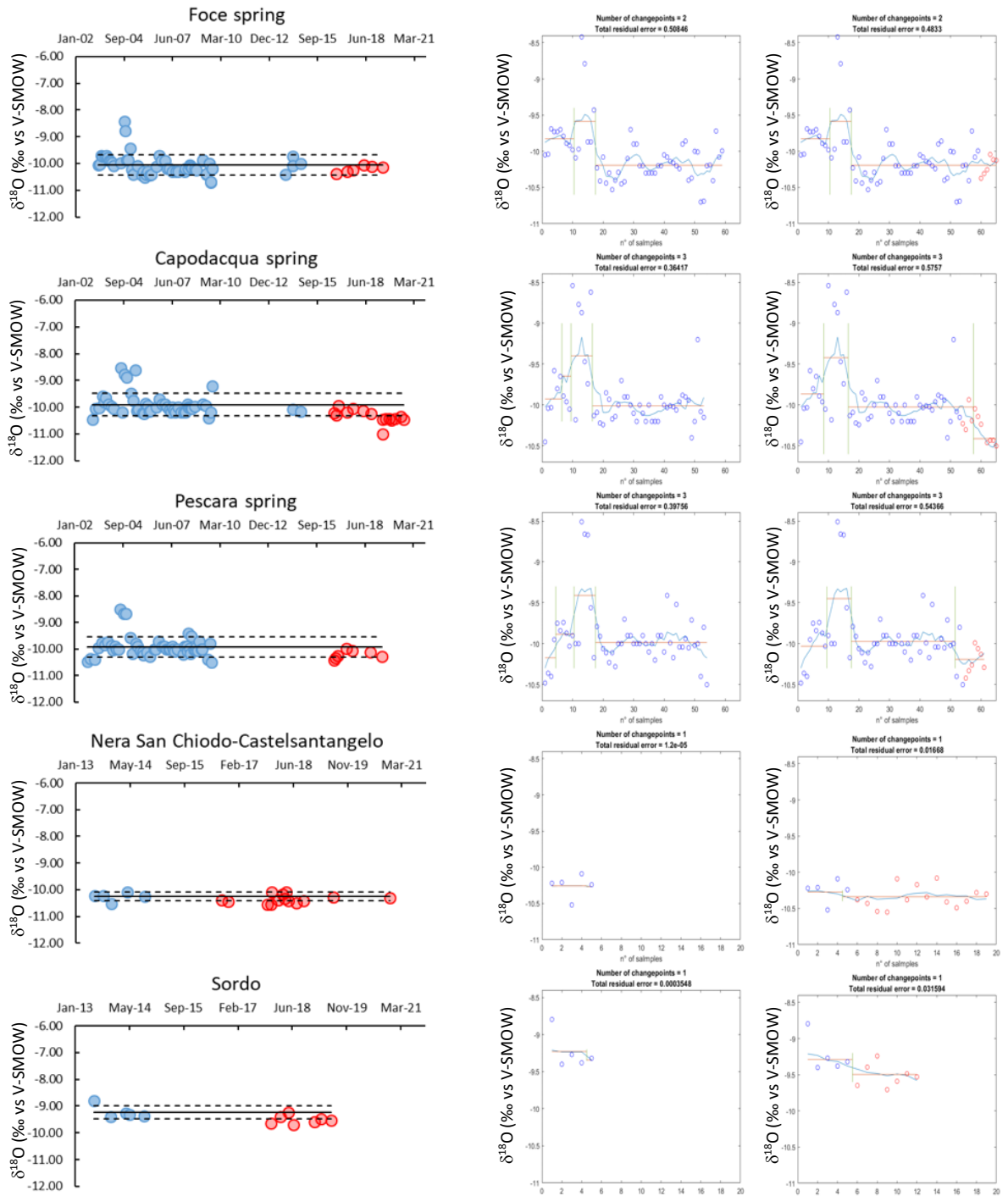


Figure 3.54 Results of the isotopic data processing. The blue dots refer to the pre-seismic period, the red dots refer to the post-seismic period. On the right row springs $\delta^{18}\text{O}$ data with pre-seismic $\delta^{18}\text{O}$ mean value (black solid line) $\pm \sigma$ (black dashed lines). On the right the results of findchangepts performed in MATLAB®.

Sordo springs. The latter is characterised by a moderate variation in the post-seismic isotopic content (becoming more negative). The complete dataset of hydrochemical and isotopic analysis used in this research are attached in the Annex C.

In light of the aforementioned outcomes, an advanced investigation to define the interaction between water and the outcropping hydrogeological complexes in the spring recharge areas, using isotopic data, was carried out by a GIS-based overlapping method. The recharge area for each spring determined by the hydrogeological setting and the mean $\delta^{18}\text{O}$ values is shown in Figure 3.55 while the meteoric $\delta^{18}\text{O}$ distribution for each hydrogeological complex within each spring recharge area is presented in the boxplots of Figure 3.56. Capodacqua spring recharge area is characterised by the outcropping of **BAS**, **JUR**, **MAI**, **MF** and **SCA**. **BAS** and **SCA** are located at lower and higher elevation respectively as shown by their meteoric $\delta^{18}\text{O}$ distribution. Pescara del Tronto spring recharge area involves **JUR**, **MAI**, **MF** and **SCA** complexes. As Foce spring concerns, only **BAS** and **JUR** hydrogeological complexes outcrops in its recharge area. Both Castelsantangelo spring and Sordo spring are characterised by the outcropping of **BAS**, **JUR**, **MAI**, **MF**, **SCA** and **DEP** in their recharge areas. The fluvial-lacustrine deposits (**DEP**) outcropping in Castelsantangelo spring recharge area are those covering the Pian Grande Plain, while the ones outcropping in Sordo spring are mainly located in the Norcia Plain (lower elevation) and only partially involves the Pian Grande Plain.

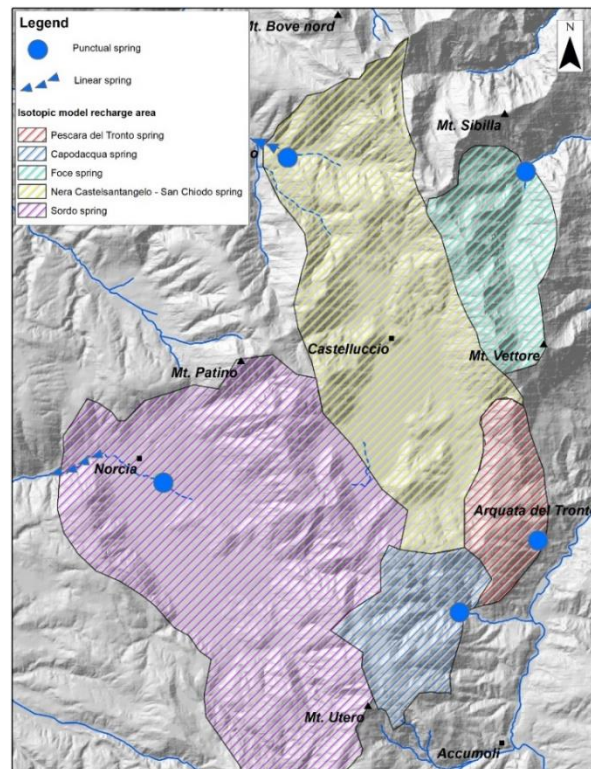


Figure 3.55 Recharge area for each spring determined by the hydrogeological setting and the mean $\delta^{18}\text{O}$ values.

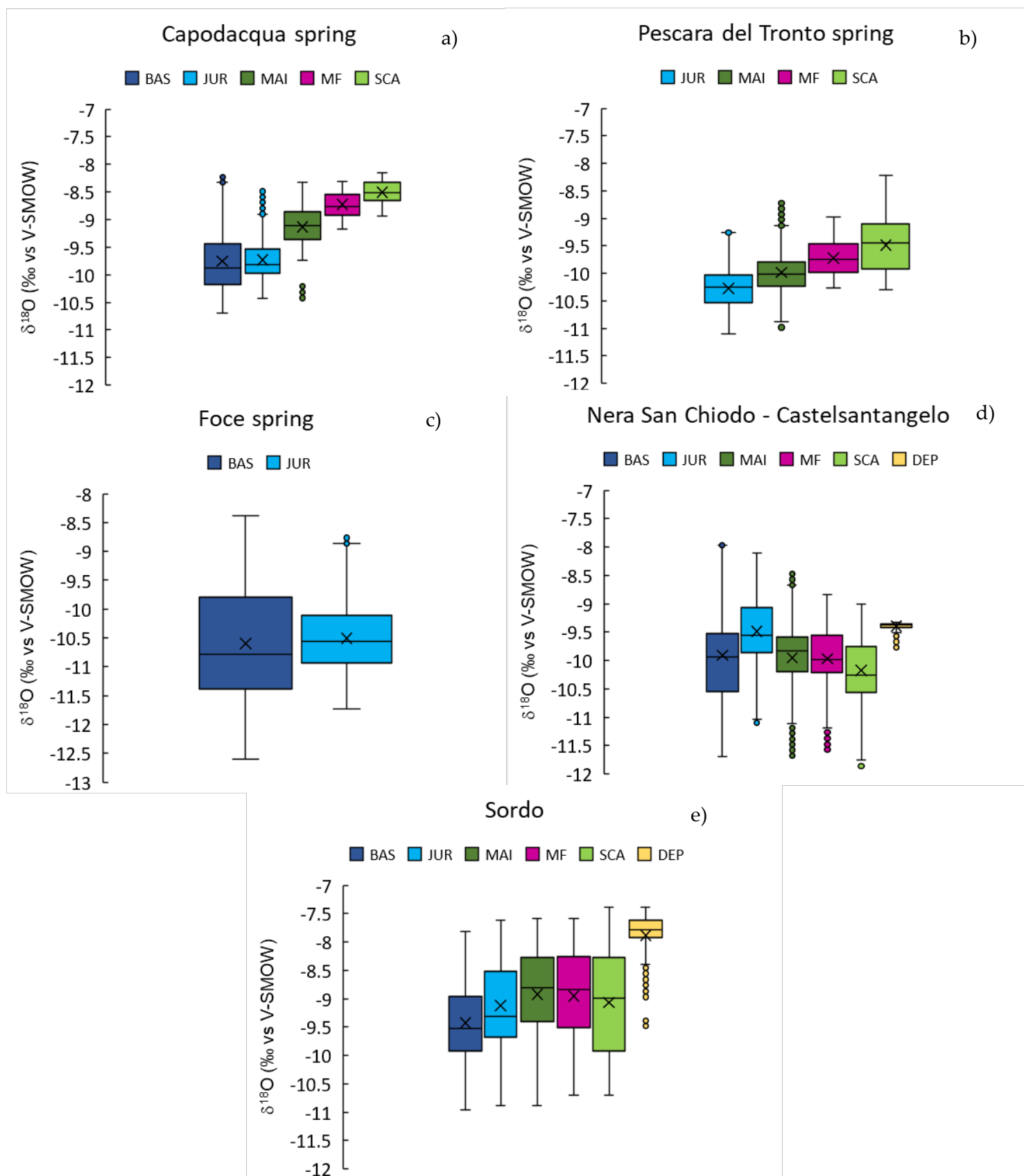


Figure 3.56 Meteoric $\delta^{18}\text{O}$ distribution for each hydrogeological complex within each spring recharge area.

The overlapping of average $\delta^{18}\text{O}$ content $\pm \sigma$ of the analyzed springs to the $\delta^{18}\text{O}$ hydrological complexes distribution (Figure 3.57) permitted to calculate (both in pre- and post-seismic conditions) the individual contribution of each complex (expressed as relative percentage to the total) that can potentially feed the analyzed springs (Table 3.26).

During the pre-seismic period Capodacqua spring is mainly feed by **BAS** (88.6 %) and subordinately by **JUR** (11.1%), while the **MAI** contribution can be considered negligible (< 1%). After the seismic sequence, the interaction with the **BAS** complex seems to be more evident (95.4 %). Pescara del Tronto spring is mainly feed by the **MAI** both during the pre- and post-seismic periods and subordinately by **SCA**. Foce spring is mainly recharged by the meteoric water infiltrating in the **BAS** complex (> 70 %) both during the pre- and post-seismic phases. A more complex hydrogeological setting is highlighted by Castelsantangelo and Sordo springs. The Castelsantangelo spring is fed, according to the isotope overlapping method, mainly by the **SCA**, secondary by the **MAI** and then by the **BAS**. The percentage ratios between the various hydrogeological complexes do not change significantly before and after the seismic sequence. However, the fluvial-lacustrine deposits (**DEP**) covering the Pian Grande Plain, although largely belonging to the spring recharge area, are not identified by isotopic overlapping method. This result may derive from the fact that rainfall involving the western slope of Mt. Vettore (characterised by steep slopes), located at higher elevations with respect to the Pian Grande Plain, contributes to recharge the fluvial-lacustrine deposits (**DEP**) of Pian Grande area and thus the Castelsantangelo spring.

As far as the Sordo spring is concerned, it is characterised both in the pre- and post-seismic phases by a water supply mainly due to **BAS** which counts for itself more than 40 % with respect to the other hydrogeological complexes. However, during the post-seismic phase this contribution from the **BAS** seems to slightly decrease in favour to the fluvio-lacustrine deposits of the Pian Grande Plain. Also in this case, this result can be interpreted as follows: the meteoric recharge, involving the western slope of Mt. Vettore, contributes to feed the fluvial-lacustrine deposits (**DEP**) of Pian Grande area and thus partly recharging the Sordo spring.

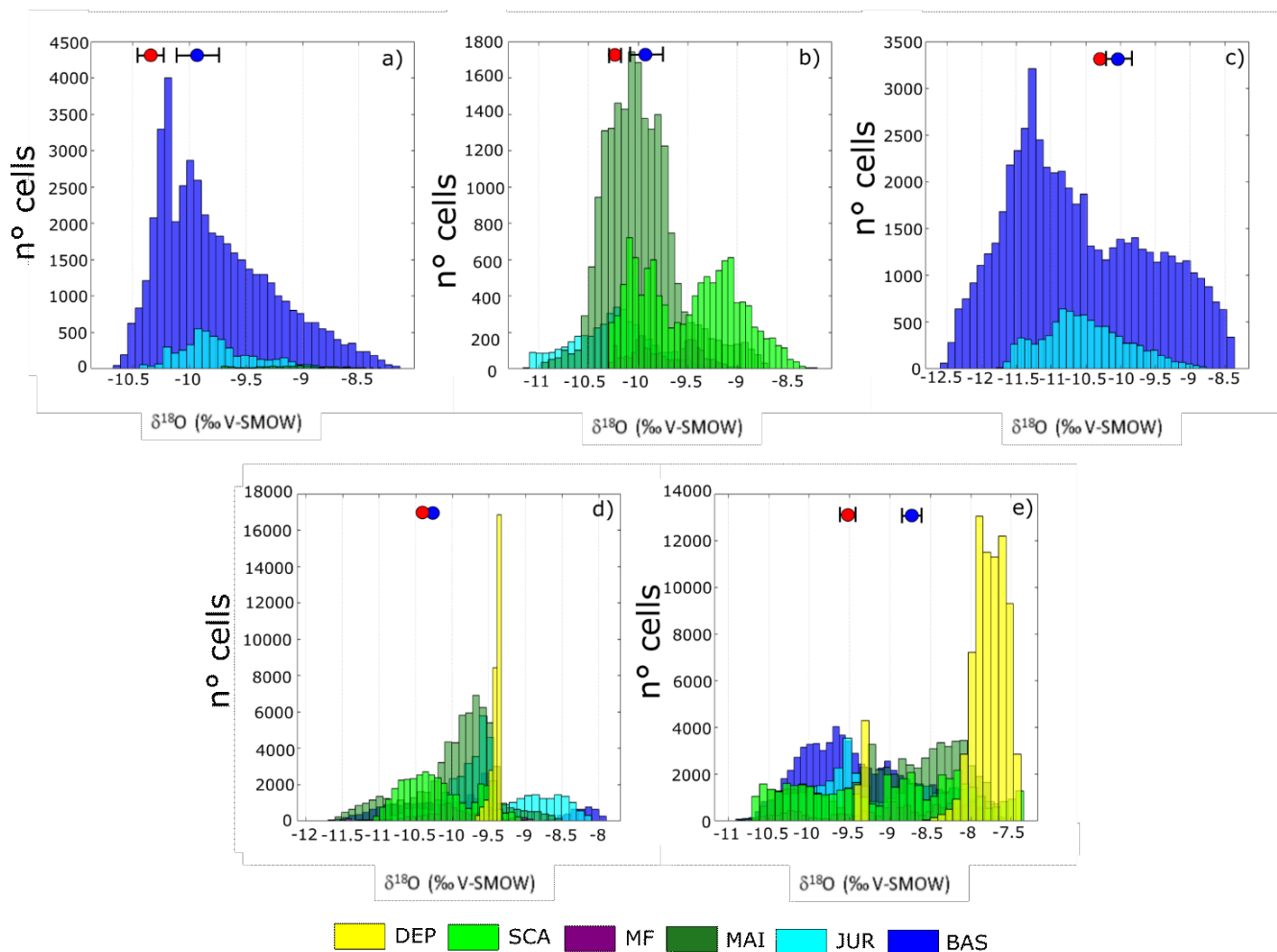


Figure 3.57 Hydrogeological complexes $\delta^{18}\text{O}$ distribution outcropping in each recharge area for each spring: (a) Capodacqua spring (b) Pescara del Tronto spring (c) Foce spring (d) Nera San Chiodo- Castelsantangelo (e) Sordo spring. For each area the spring $\delta^{18}\text{O}$ content $\pm \sigma$ is plotted in the chart bar (blue dot: pre-seismic period; red dot: post-seismic period).

Table 3.26 Hydrogeological complexes (relative percentage to the total) that can potentially feed the analysed springs, obtained by the overlapping between the complexes $\delta^{18}\text{O}$ distribution and the average springs $\delta^{18}\text{O}$ content $\pm \sigma$.

Period	Spring name (n° hydrostructure)	%BAS	%JUR	%MAI	%MF	%SCA	%DEP
Pre-seismic	Capodacqua (2a)	88.6	11.1	0.3	-	-	-
	Pescara del Tronto (2a)	-	9.3	60.8	7.3	22.6	-
	Foce (1)	77.7	22.3	-	-	-	-
	Nera San Chiodo-Castelsantangelo (2)	13.7	11.6	30.5	7.9	36	-
	Sordo (3)	48.6	21.8	12.5	2.6	14.6	0.01
Post-seismic	Capodacqua (2a)	95.4	4.6	0.1	-	-	-
	Pescara del Tronto (2a)	-	13.9	64.8	4.1	17.3	-
	Foce (1)	73.7	26.3	-	-	-	-
	Nera San Chiodo-Castelsantangelo (2)	14.1	10	27.1	5.9	42.9	-
	Sordo (3)	41.5	23.4	13.1	2.9	13.3	5.8

3.5.1 Snowmelt processes and isotopic fractionation

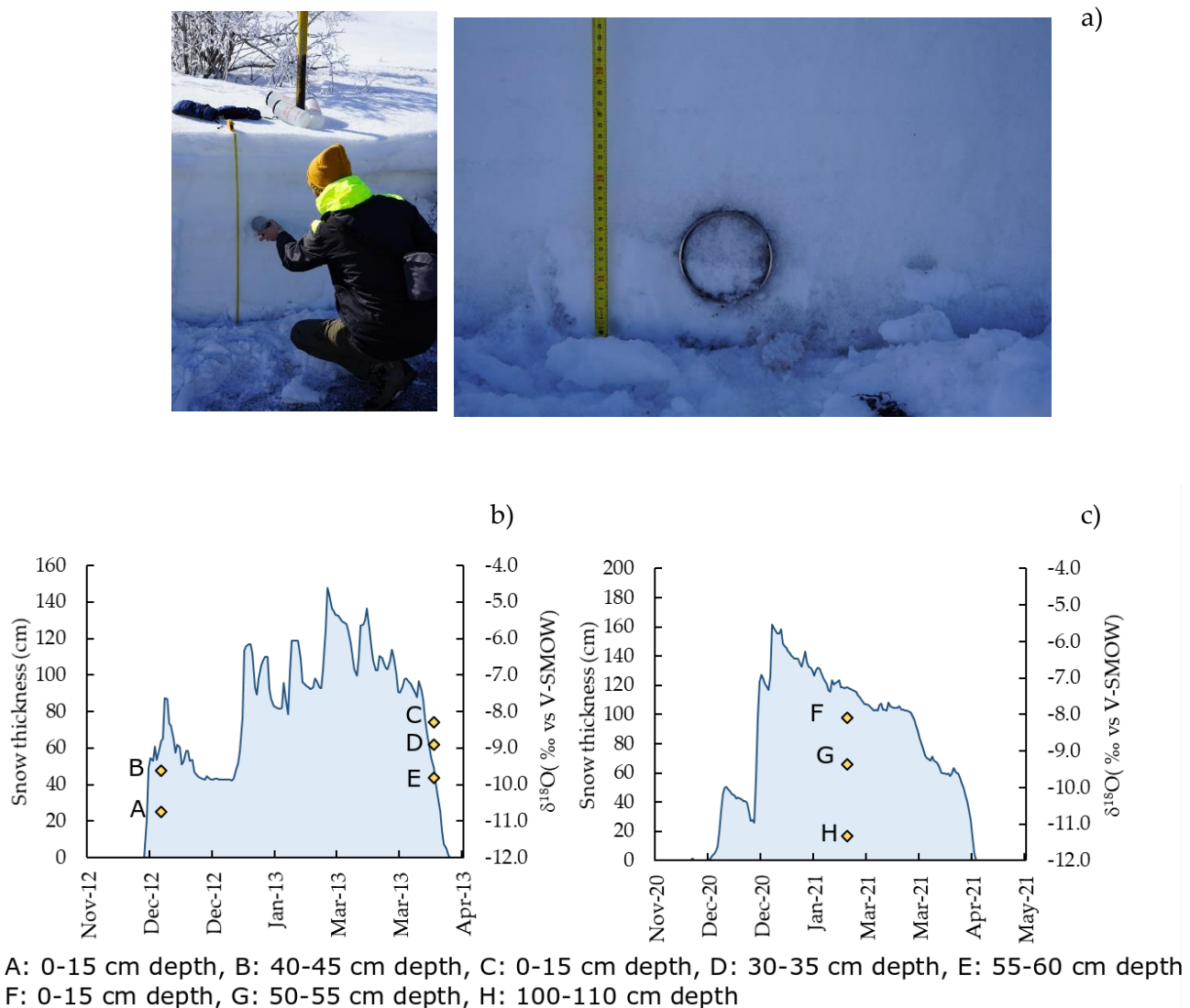


Figure 3.58 (a) Snow cover isotopic sampling phase. (b) Snow isotopic characterisation performed in the winter 2012-2013. (c) Snow isotopic characterisation performed in the winter 2020-2021.

The snow cover isotopic sampling, performed during the snowpack accumulation on December 2012, shows a more enriched isotopic composition in the deeper sample (B) collected 40-45 cm from the snow surface, with respect to the one collected between 0 and 15 cm below the snow cover surface (A). An opposite trend is observed during the snowmelt period. The shallow sample (C) is characterized by a more enriched isotopic composition with respect to the one collected in the mid-portion of the snow cover thickness (D) and the one collected in its deep portion (E) respectively (Figure 3.58b). The same trend is observed during the snowmelt period of the year 2021. The shallow

sample (F) is characterized by a more enriched isotopic composition with respect to the one collected in the mid- (G) and in the deep portion (H) of the snowpack respectively (Figure 3.58c).

The preliminary results show that the snowpack undergoes a series of isotopic fractionation processes during the winter, leading to an isotopic stratification due to diurnal melt/refreeze processes within the pack (Taylor et al., 2001; Koeniger et al., 2008; Eskelinen et al., 2016).

However, the snowpack ripening during the winter and its melting does not show a clear isotopic signal within groundwater or surface water.

As matter of fact, the snow depth recorded in Mt. Prata station compared to the surface water $\delta^{18}\text{O}$ values of the Mèrgani sinkhole (having a seasonal activation mainly due to snowmelt processes involving the Pian Grande Plain and its surrounding slopes), not shows a clear correlation each other (Figure 3.59).

The snowpack can be considered isotopically homogenized during ripening due to exchange of water percolating through it. In addition, this hypothesis is supported by the fact that snowmelt occurs in few days during the spring, leading to a homogenisation of its isotopic content before infiltrating into the unsaturated portion of the aquifers or generate run-off. (Ala-aho et al., 2017).

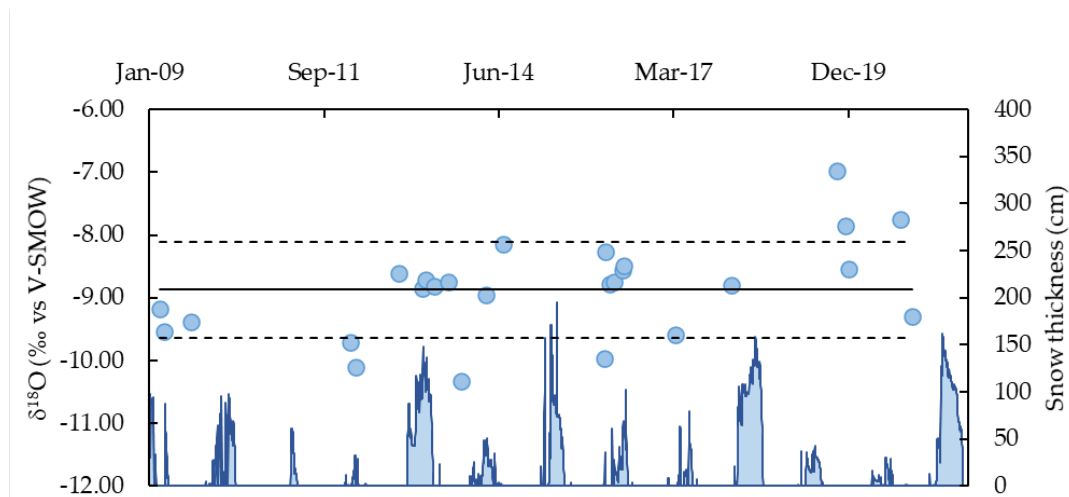


Figure 3.59 $\delta^{18}\text{O}$ data for the Mèrgani sinking stream (blue dots) vs the snow depth recorded in Mt. Prata station expressed in cm. The black solid line represents the Mèrgani water $\delta^{18}\text{O}$ mean value, while the black dashed lines represent $\pm \sigma$ interval.

Hence, the snowpack that rapidly melts and infiltrates into the aquifers (characterised by a big amount of groundwater volume) does not give a different isotopic signal with respect to the rainwater infiltrating at the same elevation. In fact, the mean $\delta^{18}\text{O}$ content of the snowpack

(measured at different depth) is approximately the same as that of rainfall in the same point, especially those measured in the deeper portion of the snowpack, which is, among other, the densest part of the snowpack itself (therefore provides more water available for infiltration).

The last aspect is evidenced by the snowpack sampling campaign for which both a different equivalent density along its thickness, both an increase in density during the melting phase (as evidenced by the comparison between the samples taken in February 2021 and April 2021) is recorded (Figure 3.60). In this context, the deeper samples are the one marked by a greater increase in density probably due to the combining effects of i) melting/freezing processes ii) compaction due to its own weight.

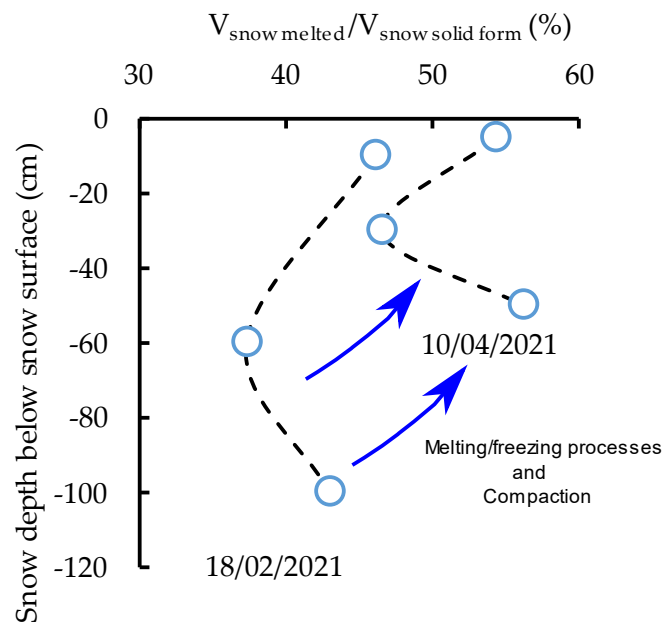


Figure 3.60 Snow density recorded at three different snow depth during the melting phase of the winter 2020-2021.

However, as this is a preliminary snow cover isotopic characterization for the study area as well as the first attempt in the Apennines chain (especially for hydrogeological purposes), it is necessary to go deeper in this aspect to assess the influence of snowmelt/isotopic fractionation processes on the spring's isotopic variations during the recharge periods.

4 Discussion

From a meteo-climatic point of view, the rainfall regime of the area has been pointed out by calculating the SPI-12 for five rainfall stations covering different sectors of the Sibillini Massif (Figure 3.1). Extremely dry conditions have been recorded by all the stations in two circumstances, both belonging to the pre-seismic period (2002 and 2007). As the post-seismic period concerns, only October 2017 is characterised by a severely dry condition in the whole area. The rainfall station of Montemonaco located in the eastern side of the domain (i.e., Adriatic side) shows a different rainfall regime with respect to the one characterising the western sector (i.e., Tyrrhenian side) represented by the Vallo di Nera rainfall station (Figure 3.2). A visible decorrelation between the SPI-12 for those stations has been observed during the years 1995-1997 and during the years 2019-2021. This aspect suggests that the main drought periods, recognised at national scale (Bordi et al., 2001; Romano et al., 2013) and at European scale in the Mediterranean areas (Caloiero et al., 2018), are well evidenced by all the rainfall stations and affect in the same way the whole domain. On the contrary, local dry or wet conditions may affect the various sectors of the Sibillini Mts. at different temporal intervals lasting up to several months.

The different rainfall regime characterising the area was confirmed by the relationship observed between the rain gauge elevations and the mean annual rainfall calculated for fifteen rainfall stations (Figures 3.6, 3.7, 3.8). In particular, the rainfall stations belonging to the Adriatic side at lowest elevation are characterised by higher rainfall values with respect to the ones located at the same elevation on the Tyrrhenian side. Moreover, the Tyrrhenian side is characterised by two different rainfall-elevation relationships suggesting that the southernmost portion is marked by higher precipitation values.

The mean annual rainfall of the rain gauges located over 1500 m a.s.l. has been incremented by the snowfall contribution thanks to the determination of the SWE performed during the snowpack sampling in the year 2021. Contrariwise to other European contexts, in which the standard relationship 1 cm snow depth = 1 mm rainfall has been used in a simplistic manner (DeWalle & Rango, 2008), in this area the peculiar geographic position (located between the Adriatic and the Tyrrhenian seas) leads to a probable arrival of moisture-laden masses resulting in wetter snow on the ground (1 cm $\approx$ 4 mm).

Although the area is characterised by different rainfall regime due to orographic effects, a strong positive correlation between elevation and mean annual rainfall has been identified in all the domains. Previous hydrogeological studies carried out in the area neglect this aspect and most of them adopted traditional methods of rainfall spatialization such as the topohyet and isohyet methods (Boni et al. 2010). However, as the authors suggested, those methods are poorly suitable to describe the actual distribution of average meteoric inflows into the Sibillini Mts. chain, as they not reflect the local topographical effects.

On the other hand, the average annual temperatures do not show significant differences between the Adriatic and Tyrrhenian sides and a single relationship between average annual temperature and elevation has been identified for the whole Sibillini area showing a strong correlation (Figure 3.4).

Both for rain gauges and for temperature stations a lack of monitoring points is observed at the highest elevations; above 1917 m a.s.l. and 1813 m a.s.l. for temperature and rainfall stations respectively.

This aspect may deeply influence the input in the water budget calculation, generally underestimating the recharge amount with respect to the outflow (discharge).

The developing of a new Web-app, namely WaterbalANce, through a Python code (Annex A) and freely available for academic or scientific purposes (Mammoliti et al., 2021) permitted to accelerate the preliminary steps to perform the hydrogeological balance and starting from mean monthly temperature and rainfall, the mean monthly water surplus has been calculated by applying the Thornthwaite-Mather method, then grouped at annual scale. This procedure highlights a strong positive correlation between mean annual water surplus and elevation, and three different relationships have been identified for the studied area (Figures 3.12, 3.13, 3.14).

The mean annual rainfall and temperature, spatialized above the study area, permitted to calculate the AET by applying the Turc method and then perform the “indirect water balance”. At the same time, the spatialized water surplus directly calculated by the WaterbalANce Web-app by applying the Thornthwaite-Mather method, allowed to perform the “indirect water balance” and compare the results obtained by the previous method.

Although the Thornthwaite -Mather method is based on monthly data, providing a more detailed information about the recharge amount, it is deeply influenced by the soil moisture storage capacity

selected for the calculation of the actual evapotranspiration (AET) and it must be chosen in reason of the soil and crop type (Donker, 1987; Thornthwaite & Mather, 1955; Thornthwaite & Mather, 1957). In this research the soil moisture storage capacity (SM) has been attributed equal to 200 mm based on previous studies conducted in similar mountain areas of central Italy (Di Matteo et al., 2017; Mammoliti et al., 2021).

At this stage, the selection of the potential infiltration coefficient for each hydrogeological complex represents a crucial phase. As matter of fact, the PIC cannot be attributed univocally as a function of the lithology outcropping in the area, but the stratigraphic and tectonic characteristics, as well as the topographic setting and the average slope values for each complex, must be considered (Andreo et al., 2008; Mussi et al., 2017; Viaroli et al., 2019).

In this research, the PIC has been modified considering the mean slope value characterising each hydrogeological complex within each hydrostructure, by proportionally decreasing the PIC value from its maximum, but not exceeding its lower limit following Celico, 1988 and Civita et al., 2000.

The procedure was driven by the convergences between the weighted average infiltration rate (mm/years) calculated starting from rainfall and temperature (indirect method) and the one obtained directly by the mean annual springs discharge values.

However, this approach is only a raw attempt to decrease the uncertainty on the water balances linked to the great variability of PIC attribution. The intersection between the literature PIC values for the lithologies outcropping in the area and their mean slope values reveals the limitations of the method, not considering the fracturing grade of the rock substratum over the study area.

In literature many works have been developed to specifically solve the issues linked to the calibration of infiltration rate both at regional and local scale (Andreo et al., 2008; Mudarra et al., 2019; Allocca et al., 2014; Martos-Rosillo et al., 2015; Imanian et al., 2020; Rossi & Donnini, 2018). According to Mussi et al., 2017 at local scale, the combined effects of tectonic setting, fracturing grade and slopes are more relevant in the calculation of the water balance, but at regional scale they can be considered negligible if related to the error linked to the determination of the water balance inputs (i.e., rainfall and snowfall) due to the non-homogeneous distribution of the thermo-pluviometric stations (Di Matteo et al., 2017). According to Di Matteo et al., 2017, the errors linked to the evaluation of precipitation amount, frequently exceed the error committed due to the inaccurate attribution of the PIC or to that related to the chosen water balance method (i.e., simple data driven

methods such as Turc or Thornthwaite-Mather vs physically based approach such as Penman-Monteith etc.).

In any case, for each hydrostructure, the effective infiltration estimated with the Turc method is slightly higher than the one calculated with the Thornthwaite-Mather method (using the same PICs), suggesting that $SM = 200$ mm univocally selected for the whole area leads to an overestimation of the AET and therefore to a lower availability of water to generate runoff or infiltration.

The mechanisms implied in the recharge processes occurring in the area are not the main focus of this research, therefore future studies aimed to characterise the actual infiltration amount by evaluating specific soils hydraulic properties and vegetation characteristics, as well as the rock fracturing grade characterisation, should be developed.

In general, the effective infiltration for the **BAS** varying between 500 and 700 mm/year, for the **MAI** ranges between 360 and 670 mm/year, while for **SCA** the effective infiltration is comprised between 500 and 650 mm/year. These results are in accord to what observed by Boni et al., 1986, for which a mean value of about 550 mm/year has been determined operating in all the Umbria-Marche Apennine carbonate ridge.

The comparison between the direct and the indirect water balance methods permit to conclude that in areas characterised by a complex hydrogeological setting and high altitude reliefs, the merely water balance based on rainfall and temperature data do not provide exhaustive information about the recharge amount, and the direct discharge measurements should be always performed in the main punctual or linear springs draining the principal aquifers, at least to better calibrate the potential infiltration coefficient (Wendland et al., 2007).

In this context the “indirect balance” was calculated to observe whether any discrepancies in the pre- and post-seismic recharge conditions occurred, and to compare the effective infiltration values, expressed as groundwater discharge, with those actually drained by the individual hydrostructures during the pre- and post-seismic periods to better identify where the main hydrogeological modifications took place and quantify the groundwater exchanges between hydrostructures.

As the pre-seismic condition concerns, a general convergence between the indirect and direct water balance has been observed for hydrostructure 1 and 2a, those characterised by more defined hydrogeological boundaries: i) the Sibillini Mts. overthrust to the east and south-east, acting as a no-flow boundary (Mastrorillo et al., 2012), ii) the Mt. Vettore-Mt. Bove normal fault to the west of

hydrostructure 1 and iii) the hydrodynamic groundwater divide determined by the hinge of the anticline to the northern portion of hydrostructure 2a (Figure 4.1). The latter is also responsible for the opposite groundwater directions within hydrostructure 2. The ones directed northward towards the Upper Nera River basin and the one directed southward towards Capodacqua and Pescara del Tronto springs (hydrostructure 2a).

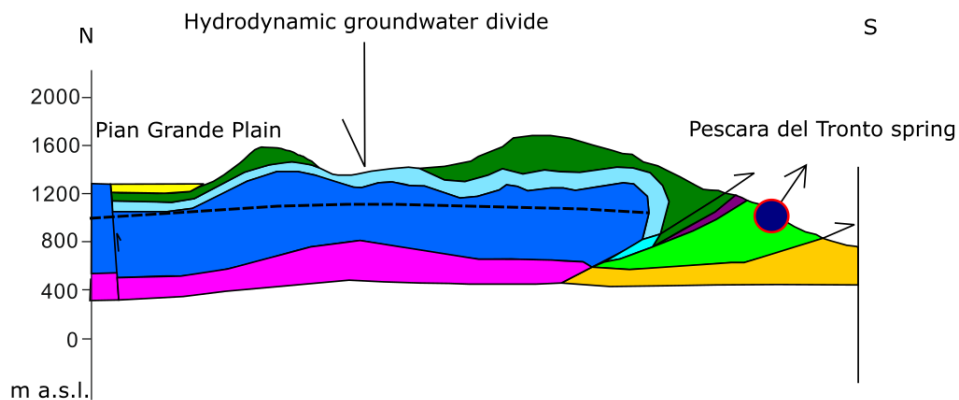


Figure 4.1 Schematic hydrogeological cross-section involving Pian Grande Plain and Pescara del Tronto spring. The hydrodynamic groundwater divide between hydrostructure 2 and 2a is determined by the hinge of the anticline. The colours of the hydrogeological complexes refer to the ones reported in Figure 1.11, modified from Fronzi et al., 2020.

The hydrostructure 2 and 3 are characterised by a higher discrepancy between the effective infiltration calculated by applying the indirect methods with respect to the groundwater drained from the same hydrostructures (direct method). In particular, a lack of about $1 \text{ m}^3/\text{s}$ has been recorded starting from discharge data of hydrostructure 3 with respect to that calculated by the indirect method (both Turc and Thornthwaite-Mather). On the contrary the hydrostructure 2 is characterised by a surplus of about $1 \text{ m}^3/\text{s}$ due to effective infiltration with respect to the actual groundwater amount drained by the same hydrostructure during the pre-seismic phase. This outcome highlights a groundwater exchange between the aforementioned hydrostructures from the hydrostructure 2 towards hydrostructure 3 of about $1 \text{ m}^3/\text{s}$ during the inter-seismic (no significant earthquakes recorded).

This preliminary observation permitted to have an overview on the role of the main fault systems of the area during the inter-seismic period. The Sibillini Mts. overthrust, bordering the U-M carbonate domain to the East act as a no-flow barrier as previously reported by (Mastrorillo et al., 2012), while the main normal fault systems (i.e., Mt. Vettore – Mt. Bove and Nottoria-Preci) are characterised by a different behaviour with respect to the groundwater flows. The Mt. Vettore – Mt. Bove fault system

acts mainly as a no-flow barrier and only minor groundwater exchanges are observed between hydrostructure 2 and hydrostructure 1, while the Nottoria-Preci fault system is responsible for local groundwater exchanges E-W oriented. This aspect is also evidenced by the pre-seismic tracer tests results (Figure 3.42) for which a hydraulic connection between the tracer injection point located in hydrostructure 2 (i.e., Mèrgani sinkhole in Pian Grande Plain) and the springs located in the hydrostructure 3 (i.e., Sordo River basin) has been observed.

Moreover, the pre-seismic artificial tracer tests (TEST1 and TEST2) demonstrate that the main groundwater flows were predominantly N-S oriented and mainly compartmentalised in the hydrostructure 2. This evidence supports the idea that the regional normal faults direct groundwater flows along their strike during the interseismic period and only a minor groundwater amount can cross the fault core in a perpendicular direction. Also TEST1 and TEST2, emphasise a divergent groundwater flow within hydrostructure 2. In fact, the traced groundwater was directed both towards the springs located in hydrostructure 2a (i.e., Pescara del Tronto and Capodacqua springs) to the South and both to some punctual and linear springs located in the Upper Nera River basin to the North.

The post-seismic discharge data for each hydrostructure substantially differ from the effective infiltration calculated by applying the indirect method, suggesting that the changes in the hydrogeological regime of the springs is not influenced by peculiar differences in the meteo-climatic conditions. Indeed, the SPI-12 of Arquata del Tronto rainfall station (barycentric position) highlights a severely-dry condition on October 2017 and this is also evident in the Vallo di Nera station (western side) that is even drier than that located on the eastern side (Montemonaco), yet to the west a discharge increase is recorded, while in the eastern side the springs show a long-term discharge decrease.

Hence, both increases and decreases in the spring discharge were recorded in the whole domain. Hydrostructure 1 is characterised by an overall decreasing discharge of about 30% with respect to the pre-seismic one, while hydrostructure 2a passes from a mean discharge value of 0.78 m³/s to 0.62 m³/s. Hydrostructure 2 is marked by an increase in mean discharge values of 1.5 m³/s, while hydrostructure 3 recorded an increase in mean discharge values of more than 3 m³/s.

The springs that show the most important long-term discharge modifications are Foce spring emerging from the **BAS** of hydrostructure 1 and the springs located in the Upper Sordo River Basin

(hydrostructure 3). Foce spring displayed a mean discharge decreasing of more than 60% with respect to the pre-seismic period, while the springs located in the Sordo River basin experienced a long-term discharge increasing (still visible) and the appearance of a new spring, named Torbidone spring, characterised by a mean discharge of 1.04 m³/s during the post-seismic phase. Nowadays the groundwater surplus due to seismic effects are still recorded in the Sordo linear springs emerging downstream to Norcia village and in San Martino spring (mean discharge value ten times higher than the pre-seismic discharge values) for which the seismic effects will return to pre-seismic values not earlier than 2023-2024 (Figure 3.21).

The regional groundwater flow modifications are evidenced by the post seismic tracer-tests (TEST4 and TEST5) results, demonstrating that 60 up to 90% of the injected tracer mass was recovered in the monitoring points of hydrostructure 3. This aspect, combined with the long-term spring discharge increasing in hydrostructure 3, suggests a groundwater transversal exchange E-W oriented towards the Nottoria-Preci fault system coming from hydrostructure 2. At the same time, the discharge increase recorded in the Upper Nera River Basin (hydrostructure 2) for which the groundwater post-seismic effects exhausted in 2018 (Mastrorillo et al., 2020), can be explained by the groundwater transversal exchange E-W oriented towards the Mt. Vettore- Mt. Bove fault system coming from hydrostructure 1. In this context, the main normal faults, responsible for the subdivision into well-defined hydrostructures, through which only minor groundwater exchanges before the earthquakes were recognised, experienced an increase in permeability perpendicular to the strike direction, due to new co-seismic fractures formation (Brozzetti et al., 2019). The enhanced permeability along the fault strike allowed for groundwater crossing the fault core perpendicularly (Figure 4.2).

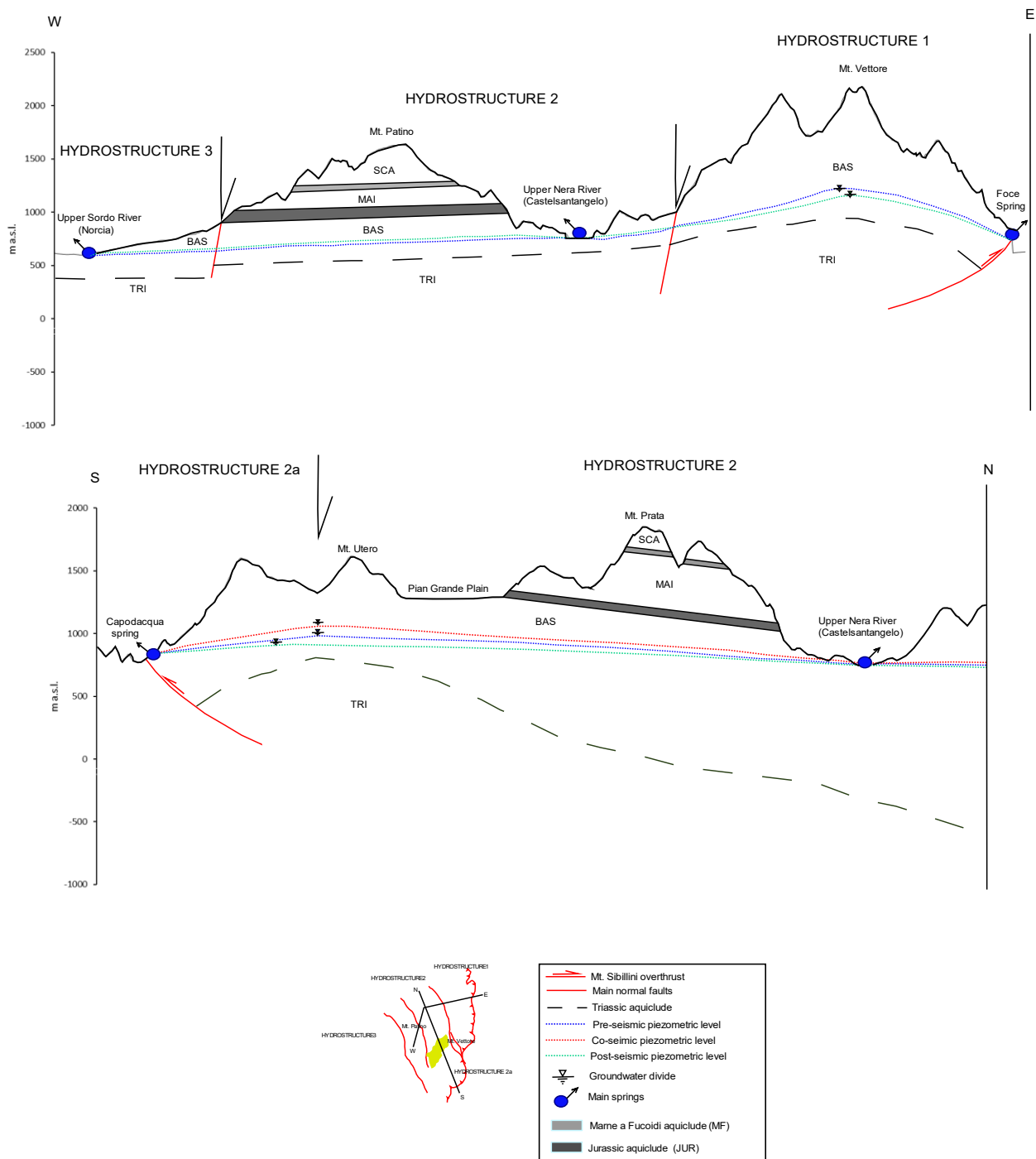


Figure 4.2 Schematic hydrogeological cross-sections with main tectonic lineaments (i.e., Normal faults and Sibillini Mts. Overthrust). The sketch (not in scale) represents the position of the piezometric level hypothesised before and after the seismic sequence with a detail of the shift of hydrodynamic groundwater divide due to earthquakes effects.

The $\delta^{18}\text{O}$ data analysis of the springs support this hypothesis. In fact, both the springs located in the Upper Nera River Basin (hydrostructure 2), characterised by a direct hydrological connection with Pian Grande Plain (point 8 and point 9 of Figures 3.40, 3.50, 3.51), experienced a mean recharge area elevation increase. The last aspect is much more evidenced by the springs located in the Sordo River

Basin (hydrostructure 3) for which the mean recharge elevation determined by the mean $\delta^{18}\text{O}$ values displays an increase of more than 100 m a.s.l. after the seismic sequence (point 18 and point 20 of Table 3.25).

As far the pre-seismic period concerns, the springs $\delta^{18}\text{O}$ data describe a complex hydrogeological setting for which in some cases the springs are fed by the aquifers outcropping just upstream to the spring (local groundwater circulation), showing mean recharge area elevation not far from the spring elevation and in other cases the springs recharge areas are characterised by higher values with respect to the aquifers outcropping upstream to them, within its hydrological basin. This aspect has been evidenced in the Upper Nera River basin for which the coupled use of hydrogeological and geo-structural approaches, supported by GIS-analysis, allowed for a detailed conceptual model.

In particular, a difference between the two analyzed basins (Castelsantangelo stream and Ussita stream) is observed with respect to the geo-structural features which deeply influence the hydrogeological response and behaviour of the streams. In the Ussita creek catchment, the Basal aquifer (**BAS**) has higher geo-structural elevation (horst) with respect to the regional **BAS**, as noticeable in the hydrogeological section of Figure 4.3. For this reason, the recharge of the Ussita stream's springs seems to be not strictly connected to the regional basal flow characterising the hydrostructure 2 (S-N oriented).

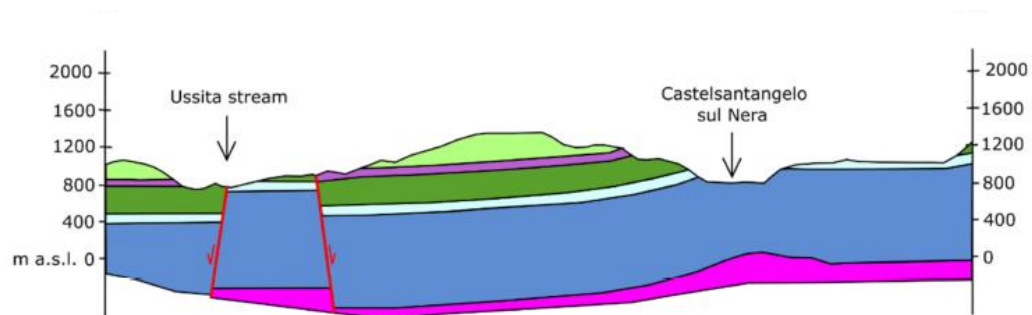


Figure 4.3 Hydrogeological cross-section (SSE-NNW oriented) modified from Viaroli et al., 2021. The colours in section refer to the sketch of hydrogeological complexes reported in Figure 1.11.

Moreover, accordingly to the GIS analysis a general higher topographic elevation is observed in the Ussita sub-basin with respect to the Castelsantangelo one (Table 3.24). This is apparent also in the linear springs of the Ussita creek, which are located at higher elevation than those emerging along the Castelsantangelo stream (Figure 3.52). The tracer tests analysis supports this hydrogeological

configuration. In fact, tracers have never been detected in the Ussita sub basin neither before nor after the seismic sequence indicating that the groundwater recharge area of the Ussita stream is comparable to its catchment. On the contrary tracer has been detected both in point 8 and 9 (Figure 3.41) indicating that there is a connection between the Castelsantangelo stream and areas outside its catchment both before and after the seismic period.

The new mapped fault seems to have a remarkable role on the groundwater path feeding the Castelsantangelo creek. In fact, TEST1 and TEST5 highlights a decoupled behaviour in quantity and time of tracer arrivals in points 8 and 9 (Figure 3.40), thus suggesting that the new mapped fault can act as preferential flow path along its strike, and groundwater coming from the Pian Grande Plain (tracer injection point) is directed towards the point 9 located near the fault (Figure 4.4).

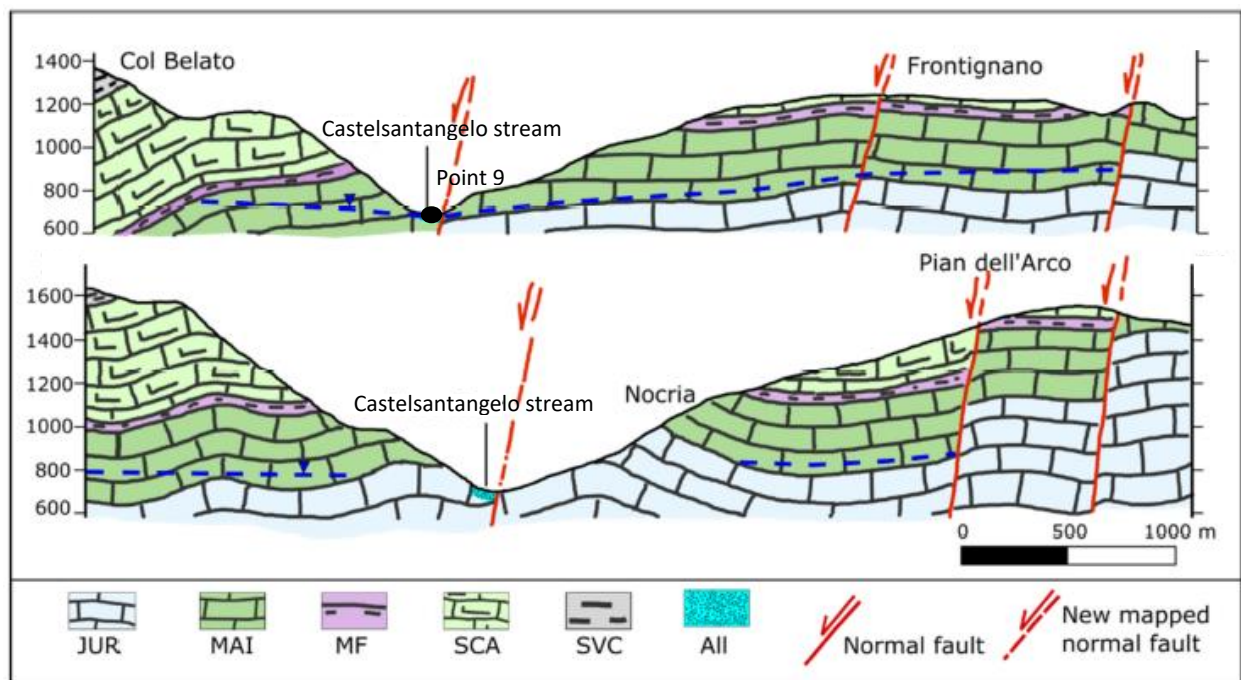


Figure 4.4 Hydrogeological cross sections with the identification of the new mapped fault and the hypothetical elevation (m a.s.l.) of water table. The colours in section refers to the sketch of hydrogeological complexes reported in Figure 1.11. From Fronzi et al., 2021.

On the contrary, the connection between the tracer injection point (Mèrgani sinking stream) and point 9 (Castelsantangelo-San Chiodo spring), evidenced also during the last tracer test (TEST6), seems to be limited to the **BAS** aquifer complex as suggested by the cross-section SE-NW directed in Figure 4.5.

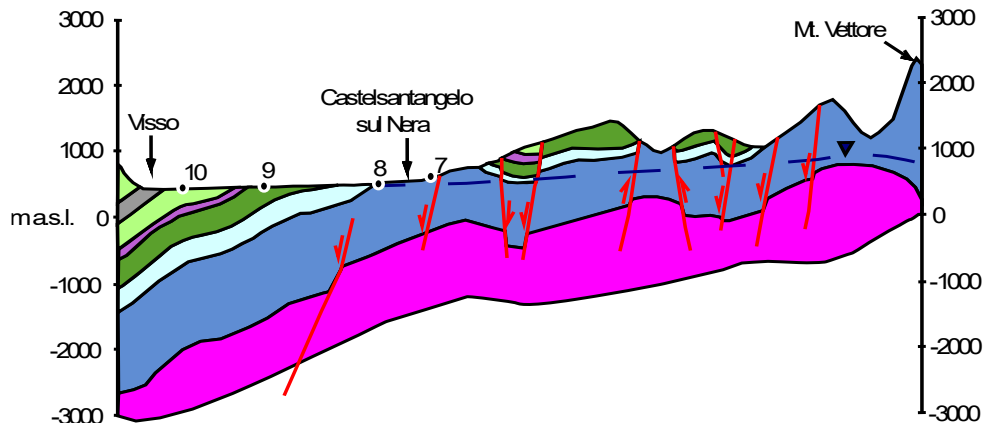


Figure 4.5 Hydrogeological cross-section (SE-NW oriented) modified from Nanni et al., 2020 with the hypothetical elevation of water table and the monitoring points along the Castelsantangelo stream. The vertical scale is exaggerated to better visualize the piezometric gradient, the colours in section refers to the sketch of hydrogeological complexes reported in Figure 1.11.

This hypothesis is in accordance with the lower tracer concentration recorded in point 8 justified by the higher dilution of tracer within the regional **BAS** aquifer. The new mapped fault can be considered as a minor tectonic feature in relation to the regional ones having the same bedding orientation, but a local influence on groundwater circulation has been observed and it represents a non-negligible geo-structural element in relation to local groundwater flow. The low value of SO_4 recorded before the seismic sequence at points 11, 12 (Ussita creek) with respect to the point 8 in the Castelsantangelo creek, confirms that the water emerging in the Ussita creek is not involved in the deep groundwater circulation interacting with the **TRI** aquiclude below the Mt. Vettore–Pian Grande Plain, characterized by high SO_4 concentration due to the leaching of the Triassic evaporites (Madonna et al., 2005).

The increase of SO_4 concentration recorded both in point 9 and 12 after the seismic sequence seems to be relate to different mechanisms.

The new mapped fault favoured a connection between the **BAS** aquifer of Mt. Vettore– Pian Grande Plain area (rich in SO_4) and point 9 as it acted as a preferential flow path. In contrast, the SO_4 increase observed in point 12 is likely related to the tectonic displacement of Mt. Vettore–Mt. Bove fault system during the seismic sequence. This fault system, acting as a flow-barrier during the pre-seismic period as observed at regional scale, became responsible for the connection between the eastern sector of the Mt. Sibillini domain (hydrostructure 1) and the western portion (hydrostructures 2 and 3), and in the high structural of Mt. Bove in which the Ussita basin develops, it caused an increase in SO_4 at point 12 not connected to tracer arrivals. The isotopic data for the

monitoring spring of the area, coupled with the tracer tests and the hydrochemical data, permitted to observe a deep difference between the analyzed basins as well.

The $\delta^{18}\text{O}$ water content measured in point 8 and 11, respectively in the **BAS** aquifer of Castelsantangelo and Ussita, indicate that the recharge area elevation of Castelsantangelo is around 1575 m a.s.l. while that of Ussita is about 1473 m a.s.l. This occurs despite the outcropping of **BAS** is higher in the Ussita catchment than in the Castelsantangelo one, where the average elevation of **BAS** is only 862 m a.s.l. This confirms that a high elevation area external to the hydrologic basin (Pian Grande Plain) feeds the creek around point 8.

Although it is known that in the epikarst of the carbonate areas, infiltration water may stagnate for a long time in micro fissures, resulting in increasing mineralization and isotope differentiation (Williams, 2008; Zuber & Motyka, 1998; Motyka, 1998), this does not seem the case of the investigated area. In fact, the research area is characterised by a well-developed surficial epikastic system only in the proximity of the Pian Grande Plain (Coltorti & Farabollini, 1995) and the infiltrating water, which can remain stagnant in the deepest fissures and micro fissures of the rock and therefore not affected by the evapotranspiration processes, is quickly remobilized when a rainfall event takes place. After all, the study area is characterized by a well-developed karst and fissured system of both saturated and unsaturated zone, which guarantees a rapid infiltration process down to the saturated zone. Williams, 2008 indicates that the stagnant water in the upper most part of the vadose zone is observed only where a high porosity contrast exists throughout the underlying bedrock. According to Nanni et al., 2020, the infiltration processes of meteoric water in this area are as fast as the water movement within the aquifers. This aspect can be evidenced by the d-excess calculated for all the groundwater samples (Figure 4.6). In fact, the d-excess value is strictly related to several environmental factors: temperature, humidity, and the evaporation processes (Dansgaard, 1964). Generally, high d-excess values indicate that more evaporated moisture has been added to the atmosphere (Cappa et al., 2003; Gat & Matsui, 1991), while low values are associated with samples fractionated by evaporation. Sappa et al., 2018 reported that the western part of the Mediterranean basin has a rainfall d-excess of +14‰, whereas the eastern part shows an excess of +22‰ (Cox et al., 2011), and this variation reflects a mixture between the Mediterranean and the Atlantic air masses (Gat et al., 2003; Bortolami et al., 1978). In the Sibillini area the precipitation samples are characterised by a mean d-excess of +9.72 ‰, while the groundwater samples present d-excess values in the range of +11.20‰ (25th percentile) to +13.59‰ (75th percentile) with an average value of +12.48 ‰ (Figure 4.6), indicating the significance of the eastern Mediterranean area as a moisture

source for this ridge. At the same time, the lower value of rainfall d-excess with respect to the springs one, as well as no significant seasonal trends can be observed in the latter, confirming that the rainfall rapidly infiltrates into the unsaturated portion of the rocks avoiding the evaporation processes occurring in the soil and sub-soil.

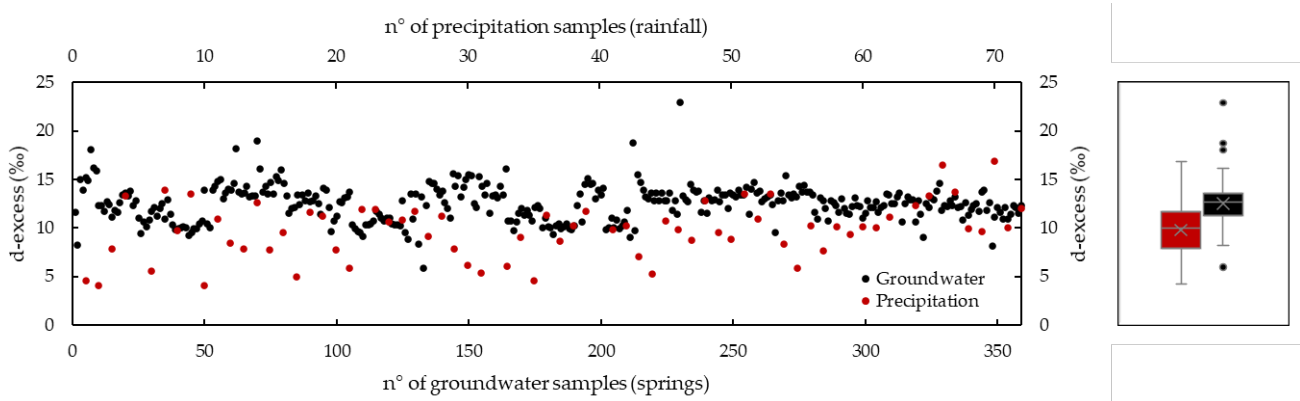


Figure 4.6 Groundwater sample (black dots) and rainfall samples (red dots) d-excess values on the left, and their distribution on the right.

Moreover, the results obtained by the sliding windows cross-correlation (SWCCF) between rainfall, and residual spring discharge, show quick time response between the precipitation events and the spring discharge increasing (Figure 3.47).

The Sliding Windows Cross Correlation analysis, coupled with the results of the periodic and long-time tracer tests, clarifies important issue related to the water circulation within the unsaturated and saturated portions of the aquifers both during the pre-, co- and post-seismic conditions, and permitted to refine the post-seismic conceptual model highlighting transitory and long-term hydrogeological modifications (Fronzi et al., 2020).

The discharge time-series of Pescara del Tronto and Capodacqua springs (hydrostructure2a) show a clear decrease in the long-term discharge values and the larger response times to the inflow recorded after the earthquakes (windows 8 and 9) can be attributed to a post-seismic piezometric level drop in the recharge area, that makes unsaturated a larger portion of the previously saturated zone and indeed increases the travel times of recharge water, as demonstrated by the results of TEST4 and TEST5 (Figure 3.45). Moreover, the fact that the tracer was not recorded in Pescara del Tronto spring after the seismic sequence may justify the piezometric level drawdown.

In fact, the tracer recorded in the Capodacqua spring before the seismic sequence (TEST2) reached the Pescara spring approximately 2 days later, probably following the flowpath SW-NE oriented, identified by Celico, 2011 during the pre-seismic condition (Figure 4.7). This specific flowpath direction was not found at all in the post-seismic phase, when the fault system putting in contact the

BAS (upstream to Capodacqua spring) with the **MAI** and **SCA** aquifers (upstream to Pescara del Tronto spring) can no longer be overcome by the piezometric level (Figure 4.7). This evidence agrees with the dry up of the minor spring named Forca Canepine located at 1350 m a.s.l. near the Capodacqua spring recharge area (Figure 3.23).

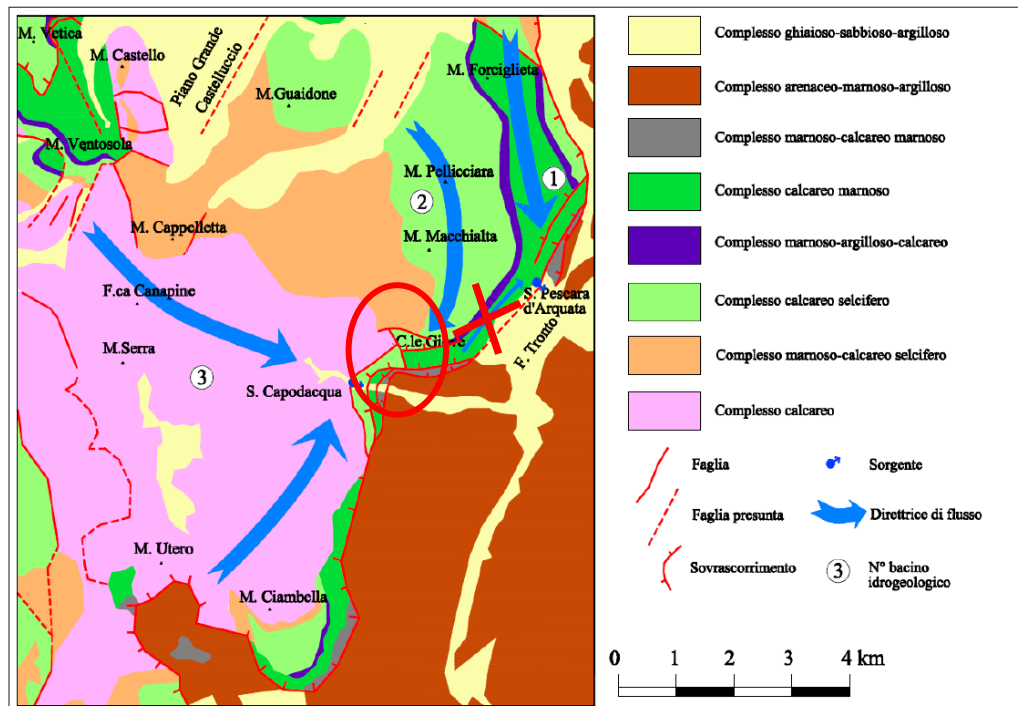


Figure 4.7 Hydrogeological map for Capodacqua and Pescara del Tronto springs modified from Celico, 2011. The red circle evidences the tectonic contact between **BAS** (rose color), **MAI** (light grey) and **SCA** (dark grey), while the red cross highlights the interruption of SW-NE oriented groundwater flow direction after the seismic sequence.

In addition, the isotopic and hydrochemical results for the Pescara del Tronto spring (Table 3.25), suggest an increasing in mean elevation recharge area (compatible with the reliefs located northward to the spring) during the post-seismic phase, while a groundwater contribution coming from lower elevations (compatible with the Pian Grande Plain elevation 1300 m a.s.l.) can be excluded in the post-seismic phase (Figure 4.8).

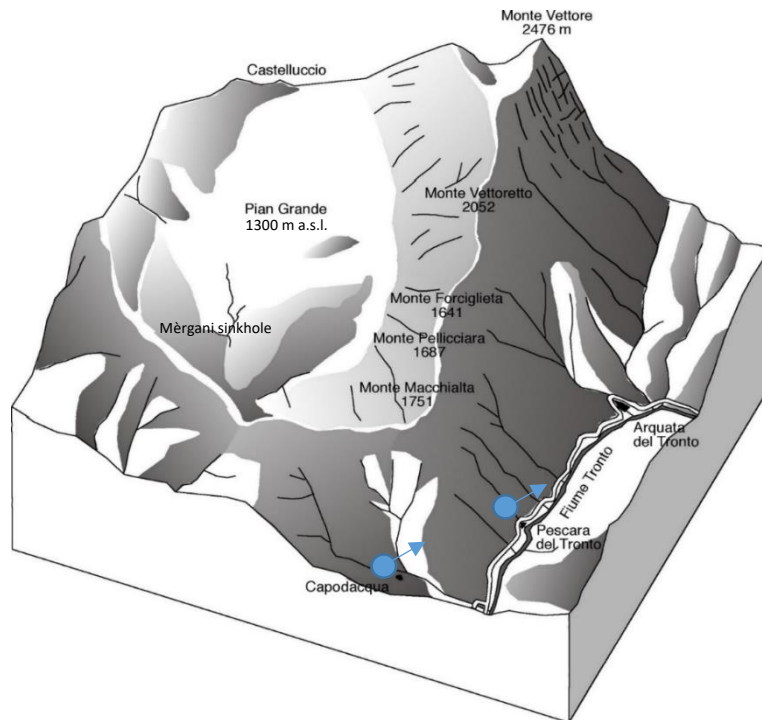


Figure 4.8 3D illustration of hydrostructure 2a (Capodacqua and Pescara del Tronto springs) and Pian Grande Plain. The increase of recharge area elevation for Pescara del Tronto and Capodacqua spring obtained by the water isotopic composition may derive from the decrease of groundwater amount coming from Pian Grande Plain (1300 m a.s.l.). Modified from Dragoni et al., 2003.

Moreover, the increase in the recharge area mean elevation is not accompanied by a significant increase in SO_4 content, excluding an extra groundwater contribution coming from the **BAS** after the earthquake.

Concerning the Pescara del Tronto spring, main groundwater flow feeding the spring is roughly parallel to the Forca di Presta syncline axe and to the Mt. Vettore - Mt. Gorzano fault system, activated during the 2016-2017 seismic sequence (Valigi et al., 2020). Surface faulting on this fault segment, which is part of the Mt. Vettore- Mt. Bove fault system, occurred during both the 24th of August and the 30th of October mainshocks accompanied by co-seismic ruptures propagated along the spring recharge area. A permeability enhancement of the aquifer feeding the spring is consistent with the observed discharge increase and is supported by the occurrence of co-seismic fracturing in the spring recharge area, along a fault in the same directions as the main groundwater flow feeding the spring. An enhancement in hydraulic conductivity is also supported by the increase of the calculated α depletion coefficients after the seismic sequence since α is known to be directly proportional to hydraulic conductivity ($\alpha \propto K$).

Hydraulic conductivity changes could be permanent or temporary. For the Pescara del Tronto spring, besides the immediate discharge increase, it was observed that the spring discharges four

years after the earthquakes remained lower than expected based on the recharge amount. These low discharge values can be justified by a transient increase of hydraulic conductivity, which has led to an exceptional and excessive emptying of the aquifer groundwater reserve. The aquifer emptying triggering the extra flow led a progressive drop of the water table; in fact, if the hydraulic conductivity (K) increase had been permanent, the system would have evolved towards a new equilibrium until the $K \times i$ product had become the same as it was before the earthquake. The discharge would have, therefore, progressively lowered until reaching values consistent with the water surplus available for recharge, and the relationship between water surplus and discharge would have been similar to those observed before the earthquake. Nonetheless, this kind of mechanism cannot explain the very low discharge values observed in 2018, 2019 and 2020, not accompanied by severe drought conditions. In fact, the progressive discharge lowering should have arrested after the new equilibrium was reached, without getting down to the very low values recorded in 2018 - 2020.

According to the findings of this work, the aquifer emptying occurred during the five months following the 24th of October earthquake reduced the hydraulic gradient, which, therefore, is still presently too low to guarantee “normal” discharge, considering that permeability, after an initial and sudden increase, is decreasing. It can be supposed that the system will go back to the previous equilibrium only when the water table, and consequently the hydraulic gradient, will rise to pre-earthquake values. This will be accompanied by a progressive reduction of α coefficient, consistent with the gradual hydraulic conductivity reduction.

During the post-seismic period, the 33-day response time characterising Pescara del Tronto spring raises to 38 days (see SWCCF results figure 3.47a), due to the water table drop that likely increases the portion of unsaturated flow path toward the saturated in the diffuse fracture matrix.

As for Pescara del Tronto spring, and Capodacqua spring, the rise of the water table, caused by the pore pressure propagation during the co-seismic period, has shortened the unsaturated path in the matrix of fractures, in favour of the karst one, and activated new long and/or slow paths. The following decrease of the water table has provoked a generalized increase of the response times and the deactivation of unsaturated paths, temporarily activated during the co-seismic period. As the 45-day response time does not correspond to any of the pre-seismic peaks, the formation of a new permanent unsaturated flow path, caused by the Mt. Vettore-Mt. Bove fault system activation, cannot be excluded.

As far Foce spring concerns (hydrostructure 1), the mid- and long-term changes in the relationship between the discharge and snow cover thickness observed by the SWCCF can be explained by a combination of (i) pore pressure propagation due to the seismic perturbation, that causes a general piezometric uprising, and (ii) non-homogeneous hydraulic conductivity variation within the aquifer, that has been proven to change hydrodynamics at different scales.

It was also noticed that during the TEST5 a low amount of Fluorescein was recorded during the entire monitoring period for Foce spring. This aspect could be linked to the decrease of spring discharge (Table 3.16 and Figure 3.24a). This reduction of water volume flowing toward Foce spring, attributable to the change in baseflow directions, caused a limitation of dilution and allowed a larger concentration of tracer to be detected by the fluorometric probe.

Since the post-seismic period is characterised by a generalised drop of the water table, the possibility that the tracer remained stuck more easily and more often in karst conduits and fractures within a widened unsaturated zone seems to be the most significant hypothesis.

The sharp hydraulic head increase, due to the pore pressure propagation in the co-seismic period, has likely caused the saturation of previously unsaturated conduits; hence, intense rainfall events reach the water table by karst circulation and not through combined karst and diffuse fracture matrix flow paths.

The quickest response time in Foce spring disappears probably because it is related to an unsaturated karst flow path, that actually brings a smaller amount of recharge water toward the saturated zone; thus, the generalized pore pressure propagation throughout the Sibillini aquifer following the repeated seismic solicitations can have masked the response of Foce residual discharge. Regarding the more retarded peak, although the pore pressure propagation can account for the decrease of the response time during the co-seismic period, its lag during the post-seismic one has been reduced to 15 days likely because co-seismic deformation could have created new fractures that have increased the vertical hydraulic conductivity in the unsaturated portion of the aquifer.

This framework permit to point out the hydrogeological characteristics and modifications occurred both at a local scale and regional scale for the eastern side of Sibillini aquifers.

During the pre-seismic period (Figure 4.9a), the baseflow fed all the springs located on the eastern side of the domain; however, the SWCCF between the snow cover thickness and the raw spring discharge highlights that the snowmelt recharge influenced considerably only the Pescara del

Toronto and Capodacqua discharge, while the Foce spring regime did not show any seasonal variation.

The tracer injected in the Mèrgani sinkhole to simulate the recharge process arrived in the south-eastern sector (i.e., Capodacqua spring) in about 10–20 days, while reaches Foce spring in more than 200 days. Intense rainfall events infiltrating during the pre-seismic period affected the discharge patterns by reaching the saturated zone with travel times of few days for Capodacqua and Foce springs and tens of days for all the spring on the eastern portion of Sibillini aquifer. Therefore, these rainfall events mobilised the tracer stack in the unsaturated zone, bringing it to the spring with an additional delay of few days.

This complex but defined framework changed considerably during the co-seismic period (Figure 4.9b) because subsequent strong seismic events caused a pore pressure propagation throughout the entire aquifer as demonstrated by the basic statistics performed on three springs located in hydrostructure 1, two springs located in hydrostructure 2a and one spring located in hydrostructure 3. This phenomenon accounts for the complete disappearance of the seasonal response in the SWCCF between the snow cover thickness and all the raw discharge time-series. Besides, another result of the pore pressure propagation was the generalised transient increase of the hydraulic head, which can explain the quicker response of the discharge of all the springs to intense rainfall events. The water table rise provoked the saturation of a shallower portion of the aquifer that is expected to have porosity more attributable to karst features than to fracture ones. For the same reason, other additional flow paths previously unsaturated were activated, as highlighted especially in Capodacqua by the SWCCF between rainfall and residual spring discharge time-series. Although changes in the relationships between the rainfall recharge and spring discharge were observed from a hydrodynamic standpoint, no significant variations were detected in the tracer travel times after intense rainfall events (Figure 3.45).

In the post-seismic period, the Mt. Vettore-Mt. Bove fault system activation, together with multiple ground shakings, caused consistent variations in the hydraulic conductivity distribution throughout the aquifer. The fault system activation produced new co-seismic ruptures and fractures, while the ground shaking caused the already existing void clean up. This heterogeneous hydraulic conductivity increases within the aquifer provoked a general lowering of the piezometric level, especially in the recharge area, and a shift on the groundwater divide with a subsequent hydraulic gradient variation. As a result, the regional groundwater flow direction changed considerably,

becoming E-W (Figure 3.9c). This variation in the regional groundwater flow accounts for the clear discharge decrease in all the springs belonging to the eastern and southern-east sector. Besides, both snowmelt (Figure 3.46) and rainfall (Figure 3.47) travel times increased, especially in Pescara del Tronto and Capodacqua springs, as pointed out by the SWCCF results. Comparing the tracer velocity variations between the pre- and post-seismic periods (Table 4.1), additional insights about changes due to the 2016–2017 period on the distribution of hydraulic conductivity can be obtained.

Table 4.1 Tracer velocity range min-max (m/day) estimated by the free software Qtracer2 (US EPA 2002) for the five tracer tests results coupled with the SWCCF.

Spring	Distance from injection point (m)	TEST1	TEST2	TEST3	TEST4	TEST5
Pescara del Tronto	6750	-	165-561	n.d.	n.d.	-
Capodacqua	6300	108 -572	114-525	37-39	-	20-29
Foce	12600	50-59	35-37	35-116	210-273	39-132

*n.d.: tracer not detected, - fluorometric probe not in function.

In detail, Foce spring shows a higher tracer velocity, that is attributable to the hydraulic conductivity increase due to the co-seismic ruptures. Besides, this new fracture network created a more permeable zone in the area between the Mèrgani sinkhole and Foce spring favouring a more rapid infiltration of intense rainfall events. In Foce spring, the simultaneous arrivals of the tracer and the abrupt and consistent discharge increase (TEST4, TEST5, TEST6) confirmed the crucial role of the co-seismic ruptures in changing this spring hydrodynamics. Contrariwise, a decrease in the overall tracer velocity has been observed in Pescara and Capodacqua springs confirming the occurrence of water table drop and the subsequent widening of the unsaturated zone increasing the probability for the tracer to remain stack in karst conduits and fractures.

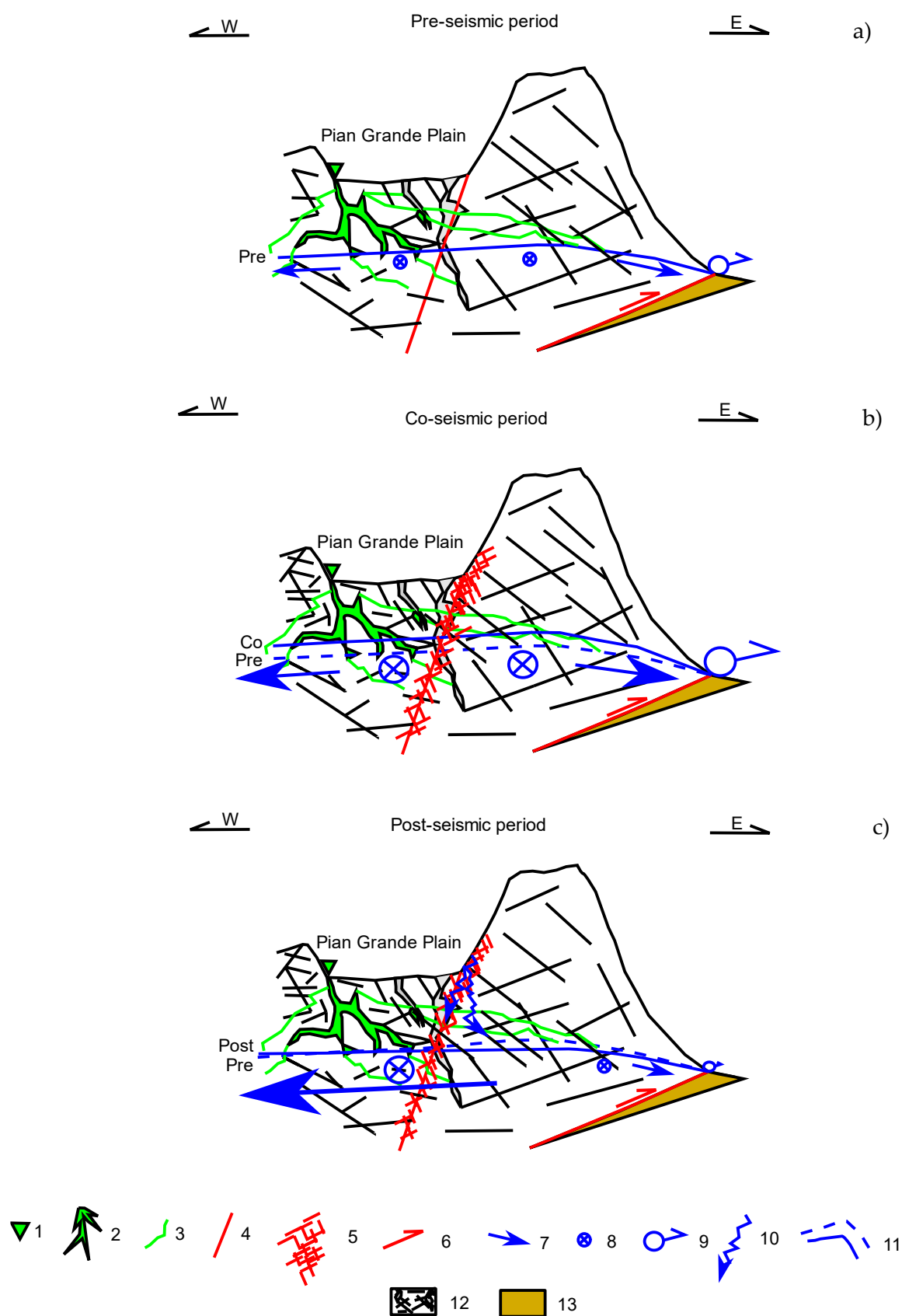


Figure 4.9 Schematic conceptual model with hydrogeological cross-sections (not in scale). In legend: (1) tracer injection point (i.e., Mèrgani sinkhole); (2) main traced paths; (3) minor traced paths; (4) Mt. Vettore - Mt. Bove normal fault; (5) co-seismic ruptures; (6) Sibillini Mts. Overthrust; (7) groundwater flow direction; (8) groundwater flow direction (perpendicular to the reader, entering in the plane); (9) spring; (10) rainfall infiltration; (11) piezometric levels; (12) carbonate fissured and fractured rocks; (13) Laga geological formation (LAG). Modified from Fronzi et al., 2020.

In this context different mechanisms influencing the springs hydrodynamic can be hypothesised. The springs emerging from the **BAS** in the Eastern sector (hydrostructure1 i.e., Foce spring), showing a medium-high sulphate content due to the interaction between groundwater and **TRI** (Table 3.25- point 2), are fed, during the pre-seismic period, by a sustained regional groundwater flow.

After the earthquakes, the co-seismic fracturing mainly affected the unsaturated rocks portion and the new fractures aperture developed along the Mt. Vettore – Mt. Bove fault system did not alter the permeability conditions of the saturated portion of the **BAS** without significant variations in the α values (Figure 3.27 and Tables 3.19, 3.21).

Hence, in Foce spring, the sudden increase in discharge rates due to intense rainfall events (daily rainfall > 30 mm) is linked to the decrease of base regional groundwater flow amount intercepted by the spring's drainage tunnel, and therefore single rainfall events are highlighted as sharp peaks in the spring hydrograph (Figure 3.30).

This phenomenon is in line with the greater co-seismic fractures opening in the proximity of the faults towards the topographic surface, which gradually closes at greater depth due to the increasing of lithostatic loading (Stemberk et al., 2019).

The springs emerging from the **BAS**, but with a more limited non-regional groundwater flowpath, (see tracer arrival velocity even in the pre-seismic phase and lower discharge values such as in Capodacqua spring) show only a transitory α increasing effect (Figure 3.26 and Table 3.18) provoked by the piezometric level increase involving the well-developed epikarstic zone underlying the Pian Grande Plain. In fact, in the long-term period the α values returned to the pre-seismic condition due to the general piezometric level decreasing.

Pescara del Tronto spring, fed by a more local suspended aquifers developed on **SCA** and **MAI**, (showing a small contribution of groundwater coming from the **BAS** due to tectonic reasons during the pre-seismic period), is characterised by several co-seismic topographic fractures affecting its recharge area and therefore, due to its aquifer shallow depth also its saturated area suffered for a permeability increase with consequent α increasing.

The GIS based $\delta^{18}\text{O}$ – overlapping method, specifically developed to investigate the interaction between water and the outcropping hydrogeological complexes in the spring recharge areas, generally provide a useful result, but some precautions are necessary.

This approach works well where the infiltrating water does not undergo fractionation processes which make the $\delta^{18}\text{O}$ -elevation relationship not applicable to the recharge areas. If the last aspect occurs, the fractionation processes should be characterised during the entire recharge period by isotopic sampling along the soil profile and along the unsaturated portion of the aquifers.

On the other hand, if a spring is fed by regional deep groundwater flows, if different aquifers show similar outcropping elevation within the same spring recharge area, or if the operator selects a model area not comparable to the real spring recharge area this approach suffers of reliability, but it can still provide useful information.

As matter of fact, in this research the application of the method provided satisfactory results for the springs located in hydrostructure 1 (i.e., Foce spring) and 2a (i.e., Capodacqua and Pescara del Tronto spring) for which the hydrogeological boundaries of their recharge areas are better defined. On the contrary, the Nera San Chiodo - Castelsantangelo spring (hydrostructure 2) fed by the aquifer with an area almost 4 times larger than its upstream hydrological catchment, involving largely of Pian Grande Plain, together with the similar outcrop elevation of each hydrogeological complexes within the recharge area, provides a general no-agreement with the real hydrogeological setting well-defined coupling GIS-analysis, tracer tests, isotope hydrology and geochemistry (Fronzi et al.,2021).

Regarding the Sordo spring (hydrostructure 3), important outcomes concerning the pre- and post-seismic modification were obtained through the application of the GIS based $\delta^{18}\text{O}$ – overlapping method. In particular, a percentage increase of **DEP** passing from 0.01 % to 5.8 % is observed. A wider extension of **DEP** of Pian Grande Plain located eastward of the Nottoria-Preci fault (hydrostructure 2) was included in the spring recharge area during the post-seismic period, confirming the increase of E-W oriented groundwater flow perpendicularly to the fault strike.

The importance of this model lies in the fact that when an operator performs a hydrological investigation starting from isotopic data only, can notice step by step if there are inconsistencies between purely modelled results and the real hydrogeological setting.

4.1 Post-seismic hydrogeological conceptual model

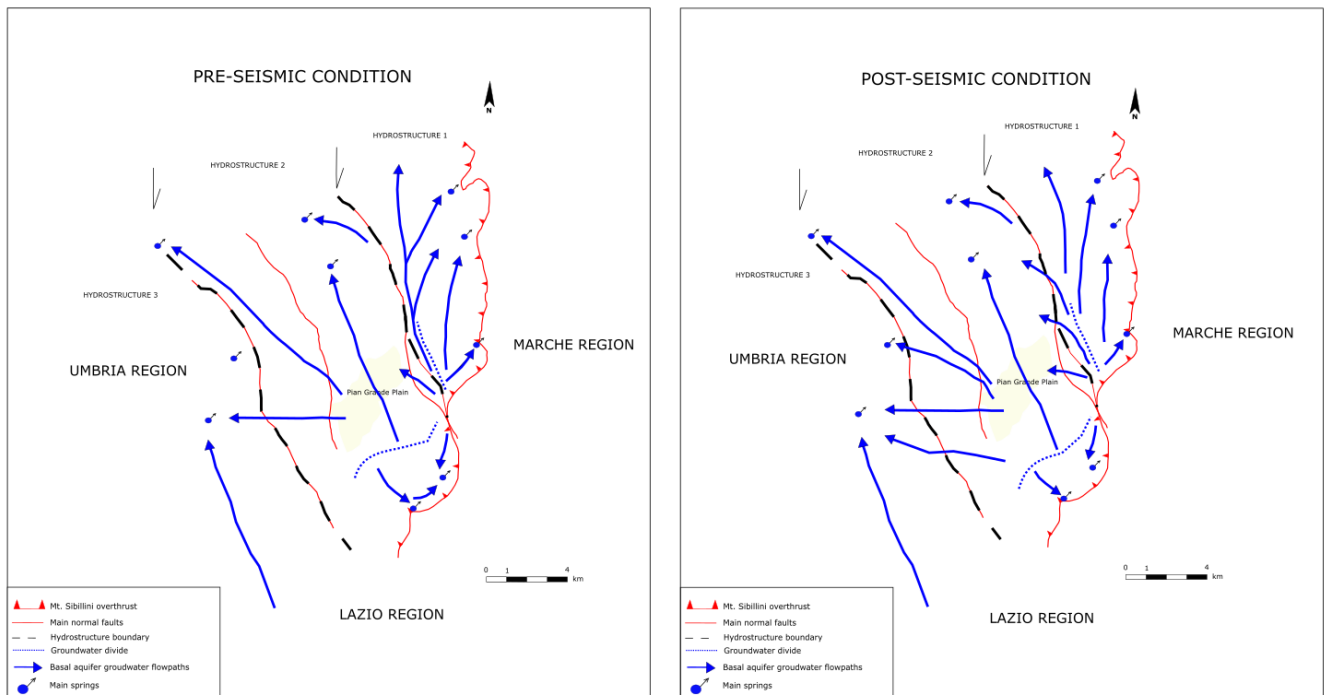


Figure 4.10 Pre- and post-seismic planimetric conceptual models with location of main springs, basal groundwater flowpaths and main tectonic lineaments. The boundaries between the hydrostructures (i.e., Mt. Vettore- Mt. Bove and Nottoria-Preci faults) are represented by the black dashed lines, while the blue dashed lines represent the groundwater divides.

In the previous paragraph, partial conceptual models have been presented for small areas of the studied domain. In Figure 4.10 a 2D conceptual model developed at regional scale has been defined relating to groundwater flows pertaining to the BAS.

The sketches report the location of the main springs and tectonic lineaments, as well as the Pian Grande Plain which represents a barycentric area with respect to the model.

As can be seen, the eastern and south-eastern springs emerge near the Sibillini Mts. Overthrust which acts as a no-flow boundary separating the carbonate domain (to the west) and the Miocene terrigenous complex (to the east) as previously observed by Banzato, 2009 and Mastrorillo et al., 2012.

During the pre-seismic phase, the main flows are S-N oriented with a divergent path in hydrostructure 2 towards the south. The groundwater flowpaths of hydrostructure 1 follow the Mt. Vettore -Mt. Bove fault system along its strike direction, which acts westward, as a no-flow boundary. The contribution of groundwater flowing between the dynamic groundwater divide and the fault feeds the springs belonging to the same hydrostructure (to the east) moving northward.

In the post-seismic phase, the right-shift of groundwater divide of hydrostructure 1, due to the Mt. Vettore - Mt. Bove fault displacement, bring to a “domino effect of discharge increase” both in hydrostructure 2 and hydrostructure 3, to the west. The groundwater, located between the dynamic groundwater divide of hydrostructure 1 and the Mt.Vettore- Mt-Bove fault, crosses perpendicularly the fault itself and then the Nottoria-Preci fault system (characterised by a smaller groundwater exchange visible during the pre-seismic phase).

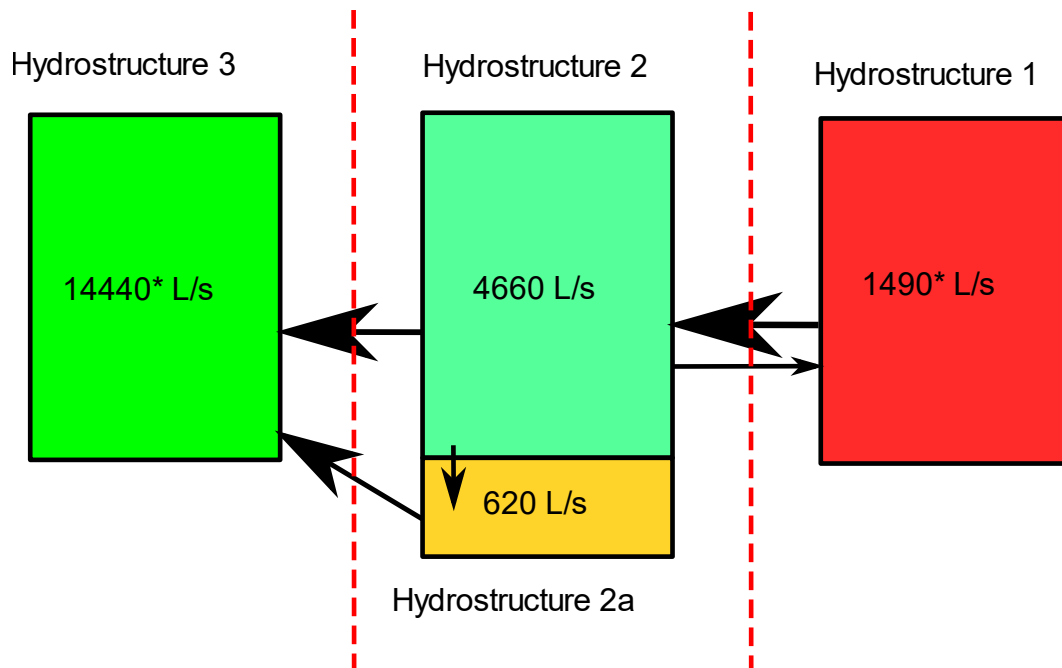
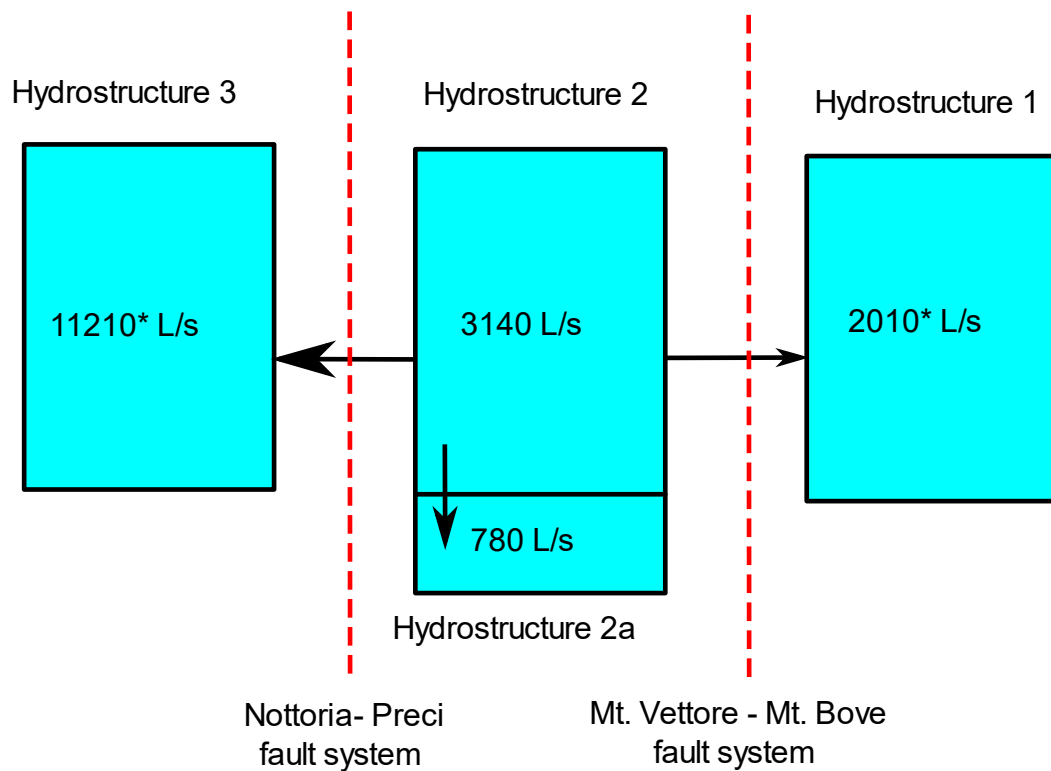
The hydrogeological disequilibrium generated by the seismic sequence, that led to draining more groundwater to the springs and linear springs located western and north-western of the Sibillini Mts. domain, provoked at the same time a shift of the dynamic groundwater divide of hydrostructure 1 and hydrostructure 2a towards east and south-east respectively, with a concomitant decrease of the springs discharge emerging to the south and eastern side of the study area. Similar mechanisms have been drawn for hydrostructure 1 by Mastroiillo et al., 2020, operating only on spring discharge data.

Furthermore, possible flow paths may have been interrupted due to piezometric lowering towards the east and others may have been activated in the co-seismic and post-seismic phases, especially towards the western sectors as shown by the large amount of tracer recovered in the springs located in hydrostructure 3.

The hydrogeological modifications observed in the post-seismic phase in the short-term is slowly returning to pre-seismic conditions. This aspect is demonstrated by the post-seismic effects exhausted in the Nera River Basin since 2018 (Mastroiillo et al. 2020), by the disappearance of Torbidone spring since June 2020 (Valigi et al.,2020) and by the continuous discharge decreasing recorded in San Martino spring (hydrostructure 3). Moreover, the very heterogeneous mean and maximum tracer velocities recorded in TEST4 are gradually linearising with the distance from injection point in the TEST5 (Figure 3.44) However, at the moment the persistence of near-normal conditions according to the standard precipitation index (not moderately, very- or extremely-wet periods recorded since May 2019) suggests that the lost groundwater amount will not be recovered in the short-term, and the term earthquake-induced groundwater overexploitation can be effectively coined. According to Mastroiillo et al., 2021 in the last two hydrologic years, the co-occurrence of climatic and post-seismic events are responsible for the depleting of groundwater resources of the Sibillini Mts. aquifers.

The conceptual models, usually developed prior to the numerical ones, with the aim of achieving a comprehensive understanding of groundwater flow circulation, are a necessary condition for rigorous modelling (Beven, 2001). The understanding of the geological and geomorphological framework, from which aquifers boundaries and geometry can be inferred, is a key point (Goldscheider & Drew, 2014). Combining the surface (geological formations outcrops and topographic setting) information with time series obtained from as many investigative approaches as possible (i.e., continuous precipitation and springs discharge monitoring, isotopes, geochemistry, artificial tracer tests) enables to link single results and to label all the components involved in the flow and storage patterns of the aquifers (Mudarra et al. 2019). Such knowledge plays an essential role for groundwater management operations and water protection plans. For these reasons, a simplified overlook on the pre- and post-seismic conditions, has been summarized with a block diagram that highlights the exchanges between different hydrostructures through the main fault systems (Figure 4.11).

PRE- SEISMIC CONDITION



POST- SEISMIC CONDITION

Figure 4.11 Simplified overlook of the pre- and post-seismic conditions schematised as a block diagram. The groundwater exchanges between different hydrostructures are represented by the black arrows. *Discharge data integrated with mean spring discharge values published in Viaroli et al., 2021 and in Valigi, 2021.

4.2 Snowpack melting processes

The importance of the snow contribution (i.e., snowfall and snowmelt) from a hydrogeological point of view on this area has been suggested by various studies (Nanni et al., 2013, Boni et al., 2010; Mastorillo et al., 2021). However, none of them have investigated this aspect using quantitative approaches. The snow melting processes have never evaluated, neither from the aquifers recharge standpoint, nor from the isotopic signal point of view. In literature is recognised that the snowmelt sequence show isotopes fractionation processes and they may be modelled and used in catchments tracing studies (Unnikrisna et al. 2002).

In this research, the Sliding Windows Cross Correlation Function performed between the snowpack depth and spring discharge allowed to quantify the time lag between the melting processes and the seasonal increase of spring discharge. A significant correlation among these phenomena is visible between 100 and 180 days (3-5 months). Since the maximum snow level is recorded on January-February, snowmelt recharge effects can be observed in the spring discharge regime between May, June, and July. In the same months, the spring discharge reach their maximum annual values confirming that the seasonal discharge regime is strongly influenced by snowmelt processes.

The snow water equivalent was empirically calculated (unfortunately for one sampling year only) and added to the mean annual precipitation. This approach, although preliminary, allowed for a more reasonable quantification of recharge amount involving the aquifers and the convergence between the direct and indirect hydrogeological balance confirm its efficiency.

Regarding the snow isotopic characterisation, the sampling campaign carried out in winter 2020-2021 was compared with that carried out in winter 2012-2013, providing comparable results, in line with the main scientific literature. In fact, all the snowpacks displayed variations of isotopic composition with depth (Friedman et al., 1992). The snowmelt had low $\delta^{18}\text{O}$ values early in the season and became progressively enriched in $\delta^{18}\text{O}$ as the pack continued to melt. The progressive enrichment in $\delta^{18}\text{O}$ of snowmelt and the dependence of $\delta^{18}\text{O}$ on melt rates can be explained by isotopic exchange between liquid water and ice (Taylor et al., 2001). Hence, melting and refreezing, altering the snowpack density, were clearly identified as processes that caused isotope fractionation (Hürkamp et al., 2019).

In the past, Moser & Stichler, 1974 affirmed that isotopic changes due to different snow processes (snowmelt, sublimation, metamorphism, etc.) are limited to the top layer of a snowpack and that the

isotopic compositions of deeper layers remain unaltered. However, further studies demonstrated that snow metamorphism can also homogenize the whole snowpack (Winograd et al., 1998).

The complex mechanism of the snow isotopic signature modification during its ripening can be explained as follows: the so-called pervasive flow (water percolation within the snowpack), from snowmelt at the surface, induces a downward translation of the isotopic composition of the snowpack along the direction of the moving water particles (Beria et al., 2018).

Niewodnizański et al., 1981 observed that altitude effect for snow precipitation is almost conserved in the snowpack.

In this study it is recognised that the snowpack ripening during the winter and its melting does not show a clear isotopic signal within groundwater or surface water and it does not give a different isotopic signal with respect to the rainwater infiltrating at the same elevation.

However, it must be evidenced that this investigation is a preliminary approach (first attempt in the Apennines chain) and specific studies carried out with heated rain gauges and a well-developed lysimeters network should be integrated with field snow sampling operations to characterize its processes and its influence on groundwater dynamics.

According to Ala-aho et al. 2017, in snow-influenced catchments, snowmelt produces a traceable isotopic signal, which differs from original snowfall isotopic composition because of isotopic fractionation in the snowpack.

To this aim, five snowmelt sampling stations, and a lysimeter have been installed at the end of this PhD programme. A research programme was specifically designed in collaboration with the Italian National Research Council-IRPI Institute and other Universities, to investigate the recharge processes and the groundwater-surface water interaction in complex aquifers through isotopic methods, selecting the Upper Nera River Basin as one of the study sites. The kick-off of this research project was in January 2022, and the sampling operation are ongoing.

5 Conclusions

The main objective of this research project was to get a deeper insight into the effects that a series of strong seismic events can have on the hydrodynamic of carbonate aquifers and to the discharge regime of its springs, both at mid- and long-term. At the same time, a hydrogeological characterisation of a complex mountain carbonate domain intended to be refined for the correct and sustainable management of groundwater resources.

The role of faults and their interaction with the groundwater flow, especially in seismic active areas, is a challenge aspect in hydrogeology and the application of a single methodology can be misleading (Fronzi et al., 2021).

The integrate approach permits to overcome the limits of a single method and join the outcomes of each approach for a better interpretation of the connection between tectonics and hydrogeology (De la Torre et al., 2020). In fact, in tectonically complex areas, the connection between the geological framework knowledge and the success of hydrogeological investigations is recognised and the two disciplines (structural geology and hydrogeology) can benefit each other. The hydrogeological functioning of the geological structures can be inferred by traditional or innovative methods, such as hydrochemical modelling or artificial/natural tracing techniques (Terzić et al., 2022).

Tracing techniques are one of the most effective tools to clearly confirm groundwater connections and transport dynamics to define the springs recharge areas, or to better understand surface water-groundwater interactions (Mudarra et al., 2019).

By integrating results and observations from individual methods, it is possible to simply obtain confident information to build up a robust hydrogeological model, such as aquifers boundaries and geometry or flowpaths and flow dynamics (De la Torre et al., 2020).

In this work, the availability of meteo-climatic data, the springs discharge data, the isotopic and hydrochemical samples collected over more than ten-year time interval that comprises the 2016–2017 seismic period, and high-quality tracer tests, performed at different times along the same period, has enabled to define a detailed framework of the modifications to the aquifer features and to the recharge system that affected the Sibillini Massif springs.

A multi-technique approach, following most of the pillars presented by De la Torre et al., 2020, has been implied and this allowed to validate the hydrogeological conceptual model for the area.

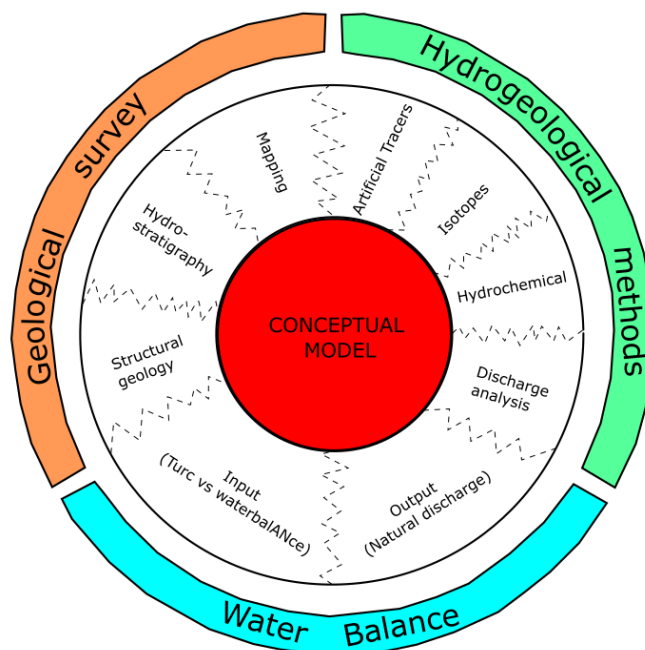


Figure 5.1 Simplified scheme illustrating the main pillars involved in this project, following and modified from De la Torre et al., 2020.

From a hydrogeological point of view, the seismic sequence triggered local transversal groundwater exchanges between aquifers belonging to different hydrostructures, separated from each other by the main extensional normal fault systems of the area. Before the seismic sequence groundwater circulation was mainly north-northwestern directed and parallel to these fault systems (Mastrorillo et al., 2012). In this area, the hydraulic conductivity increase is due to the bulk fracture cleaning triggered by the pore pressure propagation (Adinolfi Falcone et al., 2012; Amoroso et al. 2011) and to the formation of new co-seismic ruptures along with the Mt. Vettore-Mt. Bove fault system (Evans et al., 1997; Porreca et al., 2018; Villani et al.; 2018; Puliti et al., 2020).

It has been demonstrated that in shallow aquifers (i.e., **MAI** and **SCA**), the amount and timing of discharge variations following moderate to strong earthquakes can be different from those related to deeper and wider regional aquifer (**BAS**), the discharge of which can be influenced both by permeability modifications and groundwater divide migration (Mastrorillo et al., 2020). Understanding the details of the mechanisms leading to these different behaviours and their mutual interaction is fundamental to comprehend the overall response of carbonate water systems to earthquakes.

In particular, the geo-structural features (also the minor tectonic lineaments) can play a non-negligible role that should not be underestimated in the hydrogeological setting and in the analysis

of groundwater flow observed to be responsible for a different hydrodynamic response among the monitoring points in the analysed domain.

These findings confirm that, even in the frame of a well-known geo-structural scheme of a specific area (Central Apennines in this case), it is important to define the local geological setting of any strategic groundwater system, to better assess what it could be its evolution due to perturbations related to external factors (Di Matteo et al., 2020) such as earthquake sequences.

The role of long-time tracer tests (the only case in Europe with more than 5 years of continuous monitoring) was essential to go more in depth this aspect.

Besides, the new $\delta^{18}\text{O}$ -elevation relation has been determined for this area and the hydrochemical data, coupled with the isotopic ones, permitted to better characterize some of the hydrogeological changes recorded after the seismic period.

This work highlights the importance of performing an integrated approach to refine and validate hydrogeological conceptual model, especially stressing the hydrogeological modifications due to faults properties and behaviour, in particular before, after and during a seismic period.

Part of this research has tried to compare two techniques aimed at understanding underground hydrodynamics, which usually provides results at different time scales: periodic tracer tests and transient time-series analysis. In the first case, single and localized inflow events and their spatio-temporal influence on the recharge system are simulated; hence, the strong heterogeneity and anisotropy of the system, as well as the meteo-climatic features of the specific year, have a major influence on the results, although the tracer behaviour reflects the actual flow of recharge water. In the second case, the time-series analysis techniques allow identifying the inflow-outflow relationships along with large time scales and independently from the number and intensity of the recharge events. Although this is the first attempt to join time-series analysis and tracer tests and limitations can be found in both cases, the outstanding results obtained by the proposed combined approach gives the chance to statistically explain the tracer path within the aquifer, especially when the tracer gets stuck in the highly heterogeneous unsaturated zone and are mobilized by consecutive recharge events.

The GIS based- $\delta^{18}\text{O}$ isotope overlapping method can be very useful to highlight a series of warnings and suggestions that, if adopted, can provide important information about the springs recharge area

and therefore it can be useful to develop programs of groundwater protection and management of drinking water resources.

Each numerical model must be preceded by the development of a reliable qualitative or quantitative conceptual model. The use of different kind of information, investigative techniques, and modelling approaches, helps to reduce uncertainty in the calibration of the numerical models (Mudarra et al., 2019).

Since models represent the most flexible and suitable tool for all the “water operators” to address the issues and the challenges related to water management, their development with integrated techniques is strongly recommended also to others complex geological contexts.

The strength of hydrogeological models lies in “making the invisible visible” and this can be achieved by developing even detailed, but at the same time, simple 2D or 3D models (Malard et al., 2012; Jeannin et al., 2013; Jeannin & Malard, 2018), future perspective of this research.

Further developments include the snowpack isotopic investigation and its relationship with the groundwater system.

Subsequent artificial tracer tests (by identifying alternative injection points) could be performed in the area to verify further groundwater paths and to develop water resource protection plans for individual springs.

References

- Abbott, M. B., Wolfe, B. B., Aravena, R., Wolfe, A. P., & Seltzer, G. O. (2000). Holocene hydrological reconstructions from stable isotopes and paleolimnology, Cordillera Real, Bolivia. *Quaternary Science Reviews*, 19(17-18), 1801-1820.
- Adinolfi Falcone, R., Carucci, V., Falgiani, A., Manetta, M., Parisse, B., Petitta, M., ... & Tallini, M. (2012). Changes on groundwater flow and hydrochemistry of the Gran Sasso carbonate aquifer after 2009 L'Aquila earthquake. *Italian Journal of Geosciences*, 131(3), 459-474.
- Adok, U. (1977). Snow drift. *Journal of Glaciology*, 19(81), 123-139.
- Ajami, H., Meixner, T., Dominguez, F., Hogan, J., & Maddock III, T. (2012). Seasonalizing mountain system recharge in semi-arid basins-climate change impacts. *Groundwater*, 50(4), 585-597.
- Ala-aho, P., Tetzlaff, D., McNamara, J. P., Laudon, H., Kormos, P., & Soulsby, C. (2017). Modeling the isotopic evolution of snowpack and snowmelt: Testing a spatially distributed parsimonious approach. *Water Resources Research*, 53(7), 5813-5830.
- Aley, T., & Fletcher, M. W. (1976). The water tracer's cookbook. *Missouri Speleology*, 16(3), 1-32.
- Allocca, V., Manna, F., & De Vita, P. (2014). Estimating annual groundwater recharge coefficient for karst aquifers of the southern Apennines (Italy). *Hydrology and Earth System Sciences*, 18(2), 803-817.
- Alsaaran, N. A. (2006). Using environmental isotopes for estimating the relative contributions of groundwater recharge mechanisms in an arid basin, central Saudi Arabia. *Arabian Journal for Science and Engineering*, 31(1), 3-14.
- Amici, M., Spina, R. (2002). Campo medio della precipitazione annuale e stagionale sulle Marche per il periodo 1950–2000. Macerata, IT, Centro di Ecologia e Climatologia - Osservatorio Geofisico Sperimentale
- Amoruso, A., Crescentini, L., Petitta, M., Rusi, S., & Tallini, M. (2011). Impact of the 6 April 2009 L'Aquila earthquake on groundwater flow in the Gran Sasso carbonate aquifer, Central Italy. *Hydrological Processes*, 25(11), 1754-1764.
- Andreo, B., Ravbar, N., & Vías, J. M. (2009). Source vulnerability mapping in carbonate (karst) aquifers by extension of the COP method: application to pilot sites. *Hydrogeology Journal*, 17(3), 749-758.
- Andreo, B., Vías, J., Durán, J. J., Jiménez, P., López-Geta, J. A., & Carrasco, F. (2008). Methodology for groundwater recharge assessment in carbonate aquifers: application to pilot sites in southern Spain. *Hydrogeology Journal*, 16(5), 911-925.
- Angelini, P., & Dragoni, W. (1997). The problem of modeling limestone springs: the case of Bagnara (North Apennines, Italy). *Groundwater*, 35(4), 612-618.
- Anon. (2002) *Geology in Environmental Impact Statements*. Institute of Geologists of Ireland, Dublin.
- Araguás-Araguás, L., Froehlich, K., & Rozanski, K. (2000). Deuterium and oxygen-18 isotope composition of precipitation and atmospheric moisture. *Hydrological processes*, 14(8), 1341-1355.
- Aringoli, D., Farabollini, P., Giacometti, M., Materazzi, M., Paggi, S., Pambianchi, G., ... & Tondi, E. (2016). The August 24th 2016 Accumoli earthquake: Surface faulting and deep-seated gravitational slope deformation (DSGSD) in the Monte Vettore area. *Annals of Geophysics*, 59.

- Aringoli, D., Farabollini, P., Pambianchi, G., Materazzi, M., Bufalini, M., Fuffa, E., ... & Scalella, G. (2021). Geomorphological Hazard in Active Tectonics Area: Study Cases from Sibillini Mountains Thrust System (Central Apennines). *Land*, 10(5), 510.
- ARPA Umbria (2002). Monitoraggio Dell'impatto Ambientale Degli Impianti di Troiticoltura in Valnerina — Monitoring of the Environmental Impact of Trout Farms in Valnerina; ARPA Umbria: Terni, Italy.
- Asadi, E., Isazadeh, M., Samadianfard, S., Ramli, M. F., Mosavi, A., Nabipour, N., ... & Chau, K. W. (2020). Groundwater quality assessment for sustainable drinking and irrigation. *Sustainability*, 12(1), 177.
- Bailey, A., Posmentier, E., & Feng, X. (2018). Patterns of evaporation and precipitation drive global isotopic changes in atmospheric moisture. *Geophysical Research Letters*, 45(14), 7093-7101.
- Bakalowicz, M. (1994). *Water Geochemistry: Water Quality and Dynamics*
- Bakalowicz, M. (2005). Karst groundwater: a challenge for new resources. *Hydrogeology journal*, 13(1), 148-160.
- Banzato F. (2014). Caratterizzazione del flusso in acquiferi carbonatici fratturati dell'Italia centrale per la gestione e la protezione delle risorse idriche sotterranee. Tesi di Dottorato. Università degli Studi di Roma "La Sapienza".
- Banzato F., Mastrorillo L., Nanni T., Palpacelli S. Petitta M. & Vivalda P. (2013). L'acquifero carbonatico fratturato delle sorgenti del fiume Aso (Parco Nazionale dei Monti Sibillini): valutazioni sulla risorsa rinnovabile e sull'area di alimentazione. Atti del Convegno Nazionale "La ricerca carsologica in Italia", 22-23 giugno 2013, Laboratorio carsologico sotterraneo d Bossea, Frabosa Soprana (Pubbl. 1 giugno 2016): 137-145.
- Barberio, M. D., Gori, F., Barbieri, M., Billi, A., Caracausi, A., De Luca, G., ... & Doglioni, C. (2020). New observations in Central Italy of groundwater responses to the worldwide seismicity. *Scientific reports*, 10(1), 1-10.
- Barbieri, M., Boschetti, T., Petitta, M., & Tallini, M. (2005). Stable isotope (2H , 18O and $87\text{Sr}/86\text{Sr}$) and hydrochemistry monitoring for groundwater hydrodynamics analysis in a karst aquifer (Gran Sasso, Central Italy). *Applied Geochemistry*, 20(11), 2063-2081.
- Barchi M.R. & Lemmi M. (1996) Geologia dell'area del M. Coscerno – M. di Civitella (Umbria sud-orientale). *Boll. Soc. Geol. It.*, 115, 601-624.
- Barchi M.R. (2010). The Neogene-Quaternary evolution of the Northern Apennines: crustal structure, style of deformation and seismicity, in: Beltrando M., Peccerillo A., Mattei M., Conticelli S., Doglioni C. (Eds.), *The Geology of Italy*. *Journal of Virtual Explorer*. <https://doi.org/10.3809/jvirtex.2009.00220>
- Barchi, M. R., & Mirabella, F. (2009). The 1997–98 Umbria–Marche earthquake sequence: "Geological" vs. "seismological" faults. *Tectonophysics*, 476(1-2), 170-179.
- Barchi, M., & Lavecchia, G. (1986). Structural style of the Umbrie-Marches fold belt: buckle folding or thrust ramping?. In *Geologia dell'Italia centrale*. Congresso nazionale. 73 (pp. 79-80).
- Barnett, B., Townley, L.R., Post, V., Evans, R.E., Hunt, R.J., Peeters, L., Richardson, S., Werner, A.D., Knapton, A., Boronkay, A. (2012). *Australian Groundwater Modelling Guidelines*. Waterlines Report Series. National Water Commission, Canberra.
- Batiot, C., Liñán, C., Andreo, B., Emblanch, C., Carrasco, F., & Blavoux, B. (2003). Use of Total Organic Carbon (TOC) as tracer of diffuse infiltration in a dolomitic karstic system: The Nerja Cave (Andalusia, southern Spain). *Geophysical research letters*, 30(22).
- Bear, J. (1979). *Hydraulics of Groundwater*. McGraw-Hill, New York, 567 p.

- Bear, J., & Verruijt, A. (1987). Modeling Three-Dimensional Flow. In *Modeling Groundwater Flow and Pollution* (pp. 53-84). Springer, Dordrecht.
- Bense, V. F., Gleeson, T., Loveless, S. E., Bour, O., & Scibek, J. (2013). Fault zone hydrogeology. *Earth-Science Reviews*, 127, 171-192.
- Beria, H., Larsen, J. R., Ceperley, N. C., Michelon, A., Vennemann, T., & Schaefli, B. (2018). Understanding snow hydrological processes through the lens of stable water isotopes. *Wiley Interdisciplinary Reviews: Water*, 5(6), e1311.
- Berkaloff, E. (1967). Limite de validité des formules courantes de tarissement de débit [Limit of validity of current flow recession formulations]. *Chron Hydrogéol* 10:31-41
- Bernard, P., & Zollo, A. (1989). The Irpinia (Italy) 1980 earthquake: detailed analysis of a complex normal faulting. *Journal of Geophysical Research: Solid Earth*, 94(B2), 1631-1647.
- Beven, K. (2001). How far can we go in distributed hydrological modelling? *Hydrology and Earth System Sciences*, 5(1), 1-12.
- Blumetti, A. M., & Guerrieri, L. (2007). Fault-generated mountain fronts and the identification of fault segments: implications for seismic hazard assessment. *BOLLETTINO-SOCIETA GEOLOGICA ITALIANA*, 126(2), 307.
- Boccaletti, M., Calamita, F., & Viandante, M. G. (2005). The lithospheric Apennine Neo-Chain developing since the Lower Pliocene as a result of the Africa-Europe convergence. *Bollettino della Societa Geologica Italiana*, 124(1), 87-105.
- Boccaletti, M., Tortorici, L., & Ferrini, G. L. (1986). The Calabrian Arc in the frame of the evolution of the Tyrrhenian Basin. *Giornale di Geologia*, 48(1-2), 113-120.
- Boila P. & Checcucci R. (1995). Relazioni tra dinamica di deflusso delle sorgenti e suscettibilità all'inquinamento di idrostrutture carbonatiche nell'Appennino centrale. *Atti del 2° Convegno Nazionale sulla Protezione e Gestione delle Acque Sotterranee: metodologie, tecnologie e obiettivi*. Nonantola (Modena), 17-18-19 maggio 1995. Quaderni di Geologia Applicata.
- Bonacci, O. (1993). Karst springs hydrographs as indicators of karst aquifers. *Hydrological sciences journal*, 38(1), 51-62.
- Boni C. & Petitta M. (1994). Sorgenti lineari e valutazione dell'infiltrazione efficace in alcuni bacini dell'Italia centrale. *Quaderni di Geologia Applicata*, 1, Bologna.
- Boni C. & Petitta M. (2007). Studio idrogeologico per l'identificazione e la caratterizzazione degli acquiferi che alimentano le sorgenti dei corsi d'acqua perenni dei Monti Sibillini, esteso all'intera area del Parco Nazionale. Contratto di studio e ricerca. Dipartimento Scienze della Terra Università di Roma "La Sapienza", Autorità di Bacino del fiume Tevere, Parco Nazionale Monti Sibillini. Relazione inedita.
- Boni C. & Petitta M. (2008). Redazione informatizzata della cartografia idrogeologica tematica del territorio della Regione Umbria. Rapporto definitivo. Contratto di ricerca Regione Umbria - Giunta Regionale Direzione Regionale Ambiente, Territorio e Infrastrutture Servizi tecnici Regionali-Dipartimento di Scienze della Terra Università di Roma "La Sapienza".
- Boni C. & Preziosi E. (1994). Le sorgenti lineari nell'alto bacino del fiume Nera (Appennino Umbro-Marchigiano, Italia centrale). *Atti del "1st International Meeting for young Researchers in Applied Geology"*. Lausanne, April 21th 1994, Swiss Federal Institute of Technology. GEOLOP-DGC/EPFL.

- Boni C. & Ruisi M. (2008). Le grandi sorgenti che alimentano il corso del Fiume Nera-Velino: una importante risorsa strategica nell'economia dell'Italia centrale. *Atti dei Convegni Lincei-Accademia Nazionale dei Lincei*, 232, 141.
- Boni C., Baldoni T., Banzato F., Cascone D. & Petitta M. (2010). Hydrogeological study for identification, characterization and management of groundwater resources in the Sibillini Mountains National Park (Central Italy). *Ital. J. Eng. Geol. Environ.*, 2, 21-39.
- Boni C., Bono P. & Capelli G. (1986). Schema idrogeologico dell'Italia Centrale. *Mem. Soc. Geol. It.*, 35 (2), 991-1012.
- Boni C., Falcone M., Giaquinto S., Martini E. & Zoppis L. (1991). Risorse idriche sotterranee dei massicci carbonatici umbri. Le acque sotterranee in Umbria. *Protagon Publ. n. 413 del GNDICI-CNR*: 49-64.
- Boni C., Tarragoni C., Martarelli L. & Pierdominici S. (2009). Studio idrogeologico del settore nord-occidentale dei Monti Sibillini: un contributo alla cartografia idrogeologica ufficiale. *Italian Journal of Engineering Geology and Environment*, 1, 25-40.
- Bordi, I., Frigio, S., Parenti, P., Speranza, A., & Sutera, A. (2001). The analysis of the Standardized Precipitation Index in the Mediterranean area: regional patterns. *Annals of Geophysics*, 44(5-6).
- Bortolami, G. C., Ricci, B., Susella, G. F., & Zuppi, G. M. (1979). Isotope hydrology of the Val Corsaglia, Maritime Alps, Piedmont, Italy. In *Isotope hydrology 1978*.
- Boscherini A., Checcucci R., Natale G. & Natali N. (2005). Carta idrogeologica della Regione Umbria a scala 1:100.000. Regione Umbria Direzione Politiche Territoriali, Ambientali e Infrastrutture. Servizio Geologico. *Giornale di Geologia Applicata* 2.
- Boschi, E., Guidoboni, E., Ferrari, G., Mariotti, D., Valensise, G., & Gasperini, P. (2000). Catalogue of Strong Italian Earthquakes from 461 BC to 1997 (Appendix to volume 43 N° 4, 2000). *Annals of Geophysics*, 43(4).
- Bouchaou, L., Michelot, J. L., Vengosh, A., Hsissou, Y., Qurtobi, M., Gaye, C. B., ... & Zuppi, G. M. (2008). Application of multiple isotopic and geochemical tracers for investigation of recharge, salinization, and residence time of water in the Souss–Massa aquifer, southwest of Morocco. *Journal of Hydrology*, 352(3-4), 267-287.
- Boussinesq, J. (1904). Recherches théoriques sur l'écoulement des nappes d'eau infiltrées dans le sol et sur le débit des sources [Theoretical research on groundwater recharge and spring discharge]. *J Math Pures Appl* 10:5–78
- Bowen, G. J., & Wilkinson, B. (2002). Spatial distribution of $\delta^{18}\text{O}$ in meteoric precipitation. *Geology*, 30(4), 315-318.
- Brace, W. F., Paulding Jr, B. W., & Scholz, C. H. (1966). Dilatancy in the fracture of crystalline rocks. *Journal of Geophysical Research*, 71(16), 3939-3953.
- Brassington, F. C. (2004). Developments in UK hydrogeology since 1974. Geological Society, London, Special Publications, 225(1), 363-385.
- Brassington, F. C., & Younger, P. L. (2010). A proposed framework for hydrogeological conceptual modelling. *Water and Environment Journal*, 24(4), 261-273.
- Brassington, R. (2006) *Field Hydrogeology* (3rd edn). JohnWiley and Sons, Chichester.
- Bredehoeft, J. (2005). The conceptualization model problem—surprise. *Hydrogeology journal*, 13(1), 37-46.

- Brinkman, G. A., Aten Jr, A. H. W., & Veenboer, J. T. (1963). Absolute standardization with a NaI (Tl) crystal — I: calibration by means of a single nuclide. *The International journal of applied radiation and isotopes*, 14(3), 153-157.
- Brodsky, E. E., Roeloffs, E., Woodcock, D., Gall, I., & Manga, M. (2003). A mechanism for sustained groundwater pressure changes induced by distant earthquakes. *Journal of Geophysical Research: Solid Earth*, 108(B8).
- Brozzetti F. & Lavecchia G. (1994). Seismicity and related extensional stress field: the case of the Norcia seismic zone (Central Italy). *Annales Tectonicae* 8, 38–57.
- Brozzetti F., Boncio P., Cirillo D., Ferrarini, F., de Nardis R., Testa A., Liberi F. & Lavecchia G. (2019). High-resolution field mapping and analysis of the August–October 2016 coseismic surface faulting (central Italy earthquakes): Slip distribution, parameterization, and comparison with global earthquakes. *Tectonics* 2019, 38, 417–439.
- Burger, A. (1956). Interprétation mathématique de la courbe de décroissance du débit de l'Areuse, Jura neuchâtelois (Suisse) [Mathematical interpretation of the hydrograph of the Areuse spring, Jura of Neuchatel, Switzerland]. *Bull Soc Neuch Sci Nat* 79:49–54
- Butler, R. W., Tavarnelli, E., & Grasso, M. (2006). Structural inheritance in mountain belts: an Alpine–Apennine perspective. *Journal of structural geology*, 28(11), 1893-1908.
- Caine, J. S., Evans, J. P., & Forster, C. B. (1996). Fault zone architecture and permeability structure. *Geology*, 24(11), 1025-1028.
- Calamita F. & Pizzi A. (1992). Tettonica quaternaria nella dorsale appenninica umbro – marchigiana e bacini intrappenninici associati. *Studi Geol. Camerti*, Vol. Spec. (1992/1), 17 - 25.
- Calamita F. (1990). Thrusts and fold-related structures in the Umbria-Marche Apennines (Central Italy). *Annales Tectonicae*, 4: 83-117.
- Calamita F., Coltorti M., Deiana G., Dramis F. & Pambianchi G. (1982). Neotectonic evolution and geomorphology of the Cascia and Norcia depression (Umbria-Marche Apennine). *Geogr. Fis. Dinam. Quat.*, 5.
- Calamita F., Coltorti M., Farabollini P. & Pizzi A. (1994). Le faglie normali quaternarie nella Dorsale appenninica umbro-marchigiana: proposta di un modello di tettonica di inversione. *Studi Geol. Camerti*, Vol. Spec., CROP 18: 211-226.
- Calamita, F., & Deiana, G. (1988). The arcuate shape of the Umbria-Marche-Sabina Apennines (central Italy). *Tectonophysics*, 146(1-4), 139-147.
- Calamita, F., Coltorti, M., Farabollini, P., & Pizzi, A. (1994). Le faglie normali quaternarie nella dorsale appenninica umbro-marchigiana: proposta di un modello di tettonica di inversione.
- Calamita, F., Eserete, P., Paltrinieri, W., Scisciani, V., & Tavarnelli, E. (2009). Structural inheritance of pre- and syn-orogenic normal faults on the arcuate geometry of Pliocene-Quaternary thrusts: Examples from the Central and Southern Apennine Chain. *Bollettino della Società Geologica Italiana*, 128(2), 381-394.
- Calamita, F., Paltrinieri, W., Pelorosso, M., Scisciani, V., & Tavarnelli, E. (2003). Inherited mesozoic architecture of the Adria continental palaeomargin in the Neogene central Apennines orogenic system, Italy. *BOLLETTINO-SOCIETA GEOLOGICA ITALIANA*, 122(2), 307-318.
- Calamita, F., Pizzi, A., & Roscioni, M. (1992). I "fasci" di faglie recenti ed attive di M. Vettore-M. Bove e di M. Castello-M. Cardosa (Appennino umbro-marchigiano).

- Calamita, F., Satolli, S., & Turtù, A. (2012). Analysis of thrust shear zones in curve-shaped belts: Deformation mode and timing of the Olevano-Antrodoco-Sibillini thrust (Central/Northern Apennines of Italy). *Journal of Structural Geology*, 44, 179-187.
- Caloiero, T., Veltri, S., Caloiero, P., & Frustaci, F. (2018). Drought analysis in Europe and in the Mediterranean basin using the standardized precipitation index. *Water*, 10(8), 1043.
- Cappa, C. D., Hendricks, M. B., DePaolo, D. J., & Cohen, R. C. (2003). Isotopic fractionation of water during evaporation. *Journal of Geophysical Research: Atmospheres*, 108(D16).
- Carmignani, L., & Kligfield, R. (1990). Crustal extension in the Northern Apennines: the transition from compression to extension in the Alpi Apuane core complex. *Tectonics*, 9(6), 1275-1303.
- Carminati, E., & Doglioni, C. (2012). Alps vs. Apennines: the paradigm of a tectonically asymmetric Earth. *Earth-Science Reviews*, 112(1-2), 67-96.
- Celico, F., Petrella, E., & Celico, P. (2006). Hydrogeological behaviour of some fault zones in a carbonate aquifer of Southern Italy: an experimentally based model. *Terra Nova*, 18(5), 308-313.
- Celico, P. (1988). *Prospezioni Idrogeologiche II*, Liguori Ed.
- Celico, P. (2011). *Studio Idrogeologico per la Redazione dello S.I.A. Necessario per il Rinnovo della Concessione di Derivazione della Sorgente Pescara d'Arquata (Arquata del Tronto—AP); C.I.I.P.: Ascoli Piceno, Italy.*
- Cello, G., Mazzoli, S., Tondi, E., & Turco, E. (1997). Active tectonics in the central Apennines and possible implications for seismic hazard analysis in peninsular Italy. *Tectonophysics*, 272(1), 43-68.
- Cencetti C., Dragoni W. & Nejad Massoum M. (1989). Contributo alle conoscenze delle caratteristiche idrogeologiche del Fiume Nera (Appennino centro-settentrionale). *Geol. Appl. e Idrogeol.*, 24: 191-210, Bari.
- Centamore, E., & Rossi, D. (2009). Neogene-Quaternary tectonics and sedimentation in the Central Apennines. *Bollettino della Società geologica italiana*, 128(1), 73-88.
- Charmoille, A., Fabbri, O., Mudry, J., Guglielmi, Y., & Bertrand, C. (2005). Post-seismic permeability change in a shallow fractured aquifer following a ML 5.1 earthquake (Fourbanne karst aquifer, Jura outermost thrust unit, eastern France). *Geophysical Research Letters*, 32(18).
- Checucci R., Mastrorillo L. & Valigi D. (2017). Acque sotterranee e terremoti: alcune considerazioni sugli effetti della sismicità sulla disponibilità della risorsa idrica in Valnerina. *Acque Sotterranee-Italian Journal of Groundwater*, 6.
- Cherry, M., Gilmore, T., Mittelstet, A., Gastmans, D., Santos, V., & Gates, J. B. (2020). Recharge seasonality based on stable isotopes: Nongrowing season bias altered by irrigation in Nebraska. *Hydrological Processes*, 34(7), 1575-1586.
- Chiaraluce, L., Di Stefano, R., Tinti, E., Scognamiglio, L., Michele, M., Casarotti, E., ... & Marzorati, S. (2017). The 2016 central Italy seismic sequence: A first look at the mainshocks, aftershocks, and source models. *Seismological Research Letters*, 88(3), 757-771.
- Chiaudani, A., Di Curzio, D., & Rusi, S. (2019). The snow and rainfall impact on the Verde spring behavior: A statistical approach on hydrodynamic and hydrochemical daily time-series. *Science of The Total Environment*, 689, 481-493.
- Chiaudani, A., Di Curzio, D., Palmucci, W., Pasculli, A., Polemio, M., & Rusi, S. (2017). Statistical and fractal approaches on long time-series to surface-water/groundwater relationship assessment: A central Italy alluvial plain case study. *Water*, 9(11), 850.

- Civico R., Pucci S., Villani F., Pizzimenti L., De Martini P.M., Nappi R. & Open EMERGEIO Working Group (2018). Surface ruptures following the 30 October 2016 Mw 6.5 Norcia earthquake, Central Italy. *Journal of Maps*. doi: 10.1080/17445647.2018.1441756
- Civita, M., De Maio, M., & Ubertini, L. (2000). Valutazione e cartografia automatica della vulnerabilità degli acquiferi all'inquinamento con il sistema parametrico Sintacs R5: a new parametric system for the assessment and automatic mapping of ground water vulnerability to contamination. Pitagora.
- Claesson, L., Skelton, A., Graham, C., Dietl, C., Mörrth, M., Torssander, P., & Kockum, I. (2004). Hydrogeochemical changes before and after a major earthquake. *Geology*, 32(8), 641-644.
- Coleman, M. L., Shepherd, T. J., Durham, J. J., Rouse, J. E., & Moore, G. R. (1982). Reduction of water with zinc for hydrogen isotope analysis. *Analytical chemistry*, 54(6), 993-995.
- Coltorti, M., & Farabollini, P. (1995). Quaternary evolution of the Castelluccio di Norcia basin (Umbro-Marchean Apennines, central Italy). *Il Quaternario*, 8(1), 149-166.
- Console F., Motti A. & Pantaloni M. (2017). L'intermittenza delle sorgenti del Torbidone nella Piana di Norcia: analisi delle fonti storiche a partire dal XIV secolo. *Rend. Online Soc. Geol. It.* (43), 35-56.
- Coward, M. P., De Donatis, M., Mazzoli, S., Paltrinieri, W., & Wezel, F. C. (1999). Frontal part of the northern Apennines fold and thrust belt in the Romagna-Marche area (Italy): Shallow and deep structural styles. *Tectonics*, 18(3), 559-574.
- Cox, K. A., Rohling, E. J., Schmidt, G. A., Schiebel, R., Bacon, S., Winter, D. A., ... & Spero, H. J. (2011). New constraints on the Eastern Mediterranean δ 18 O: δ D relationship. *Ocean Science Discussions*, 8(1), 39-53.
- Cox, S. C., Rutter, H. K., Sims, A., Manga, M., Weir, J. J., Ezzy, T., ... & Scott, D. (2012). Hydrological effects of the MW 7.1 Darfield (Canterbury) earthquake, 4 September 2010, New Zealand. *New Zealand Journal of Geology and Geophysics*, 55(3), 231-247.
- Craig, H. (1961). Isotopic variations in meteoric waters. *Science*, 133(3465), 1702-1703.
- Cunningham, E. E., Long, A., Eastoe, C., & Bassett, R. L. (1998). Migration of recharge waters downgradient from the Santa Catalina Mountains into the Tucson basin aquifer, Arizona, USA. *Hydrogeology Journal*, 6(1), 94-103.
- D'Affonseca, F. M., Finkel, M., & Cirpka, O. A. (2020). Combining implicit geological modeling, field surveys, and hydrogeological modeling to describe groundwater flow in a karst aquifer. *Hydrogeology Journal*, 28(8), 2779-2802.
- Dafny, E., Burg, A., & Gvirtzman, H. (2006). Deduction of groundwater flow regime in a basaltic aquifer using geochemical and isotopic data: The Golan Heights, Israel case study. *Journal of Hydrology*, 330(3-4), 506-524.
- Dahlke, H. E., & Lyon, S. W. (2013). Early melt season snowpack isotopic evolution in the Tarfala valley, northern Sweden. *Annals of Glaciology*, 54(62), 149-156.
- Dansgaard, W. (1964). Stable isotopes in precipitation. *tellus*, 16(4), 436-468.
- Darling, W. G., Edmunds, W. M., Kinniburgh, D. G., & Kotoub, S. (1987). Sources of recharge to the basal Nubian sandstone aquifer, Butana region, Sudan. IAEA.
- de Graaf, I. D., Sutanudjaja, E. H., Van Beek, L. P. H., & Bierkens, M. F. P. (2015). A high-resolution global-scale groundwater model. *Hydrology and Earth System Sciences*, 19(2), 823-837.
- De Luca, G., Di Carlo, G., & Tallini, M. (2016). Hydraulic pressure variations of groundwater in the Gran Sasso underground laboratory during Amatrice earthquake of August 24th, 2016. *Annals of Geophysics*.

- De Luca, G., Di Carlo, G., & Tallini, M. (2018). A record of changes in the Gran Sasso groundwater before, during and after the 2016 Amatrice earthquake, central Italy. *Scientific reports*, 8(1), 1-16.
- De Vries, J. J., & Simmers, I. (2002). Groundwater recharge: an overview of processes and challenges. *Hydrogeology Journal*, 10(1), 5-17.
- Deiana, G. & Piali, G. (1994) The structural provinces of the Umbro Marchean Apennines. *Mem. Soc. Geol. It.*, 48, 473-484.
- De la Torre, B., Mudarra, M., & Andreo, B. (2020). Investigating karst aquifers in tectonically complex alpine areas coupling geological and hydrogeological methods. *Journal of Hydrology* X, 6, 100047.
- Delbart, C., Valdes, D., Barbecot, F., Tognelli, A., Richon, P., & Couchoux, L. (2014). Temporal variability of karst aquifer response time established by the sliding-windows cross-correlation method. *Journal of Hydrology*, 511, 580-588.
- Dennis, F., Andrews, J. N., Parker, A., Poole, J., & Wolf, M. (1997). Isotopic and noble gas study of Chalk groundwater in the London Basin, England. *Applied Geochemistry*, 12(6), 763-773.
- DeWalle, D. R., & Rango, A. (2008). *Principles of snow hydrology*. Cambridge University Press.
- Di Bucci, D., & Mazzoli, S. (2002). Active tectonics of the Northern Apennines and Adria geodynamics: new data and a discussion. *Journal of Geodynamics*, 34(5), 687-707.
- Di Domenica, A., Bonini, L., Calamita, F., Toscani, G., Galuppo, C., & Seno, S. (2014). Analogue modeling of positive inversion tectonics along differently oriented pre-thrusting normal faults: an application to the Central-Northern Apennines of Italy. *Bulletin*, 126(7-8), 943-955.
- Di Domenica, A., Turtù, A., Satolli, S., & Calamita, F. (2012). Relationships between thrusts and normal faults in curved belts: new insight in the inversion tectonics of the Central-Northern Apennines (Italy). *Journal of Structural Geology*, 42, 104-117.
- Di Matteo L., Capoccioni A., Porreca M., & Pauselli C. (2021). Groundwater-Surface Water Interaction in the Nera River Basin (Central Italy): New Insights after the 2016 Seismic Sequence. *Hydrology*, 8(3), 97.
- Di Matteo L., Dragoni W., Azzaro S., Pauselli C., Porreca M., Bellina G. & Cardaci W. (2020). Effects of earthquakes on the discharge of groundwater systems: The case of the 2016 seismic sequence in the Central Apennines, Italy. *Journal of Hydrology*, 583, 124509.
- Di Matteo, L., Dragoni, W., Maccari, D., & Piacentini, S. M. (2017). Climate change, water supply and environmental problems of headwaters: The paradigmatic case of the Tiber, Savio and Marecchia rivers (Central Italy). *Science of the Total Environment*, 598, 733-748.
- Doerfliger, N., Zwahlen, F. (1995). Action COST 65 – Swiss National Report. *Bulletin d’Hydrogéologie de l’Université de Neuchâtel*, 14, 3-33.
- Dogliani, C. (1991). A proposal for the kinematic modelling of W-dipping subductions-possible applications to the Tyrrhenian-Apennines system. *Terra Nova*, 3(4), 423-434.
- Donker, N. H. W. (1987). WTRBLN: a computer program to calculate water balance. *Computers & Geosciences*, 13(2), 95-122.
- Doveri, M., & Mussi, M. (2014). Water isotopes as environmental tracers for conceptual understanding of groundwater flow: An application for fractured aquifer systems in the “Scansano-Magliano in Toscana” area (Southern Tuscany, Italy). *Water*, 6(8), 2255-2277.
- Dragoni, W., Speranza, G., & Valigi, D. (2003). Impatto delle variazioni climatiche sui sistemi idrogeologici: Il caso della sorgente Pescara di Arquata (Appennino umbro-marchigiano, Italia). *Geol. Tec. Ambient*, 3, 27-35.

- Dragoni, W., & Sukhija, B. S. (2008). Climate change and groundwater: a short review. Geological Society, London, Special Publications, 288(1), 1-12.
- Drew, D., Hötzl, H., (1999). Karst Hydrology and Human Activities. International Contributions to
- Drogue, C. (1974). Structure de certains aquifères karstiques d'après les résultats de travaux de forage. (Structure of certain karst aquifers from drilling data). CR Acad Sci Paris, série III (278):2621–2624.
- Drogue, C. (1980). Essai d'identification d'un type de structure de magasins carbonatés fissurés: application à l'interprétation de certains aspects du fonctionnement hydrogéologique (On the structural identification of carbonate storage fissures: application to certain aspects of hydrological functioning). Mém Hors Sér Soc Géol Fr 11:101–108.
- Earman, S., & Dettinger, M. (2011). Potential impacts of climate change on groundwater resources—a global review. Journal of water and climate change, 2(4), 213-229.
- Earman, S., Campbell, A. R., Phillips, F. M., & Newman, B. D. (2006). Isotopic exchange between snow and atmospheric water vapor: Estimation of the snowmelt component of groundwater recharge in the southwestern United States. Journal of Geophysical Research: Atmospheres, 111(D9).
- Edmunds, W. M., & Wright, E. P. (1979). Groundwater recharge and palaeoclimate in the Sirte and Kufra basins, Libya. Journal of Hydrology, 40(3-4), 215-241.
- Emblanch, C., Zuppi, G. M., Mudry, J., Blavoux, B., & Batiot, C. (2003). Carbon 13 of TDIC to quantify the role of the unsaturated zone: the example of the Vaucluse karst systems (Southeastern France). Journal of hydrology, 279(1-4), 262-274.
- Emergeo, W. G., Pucci, S., De Martini, P. M., Civico, R., Nappi, R., Ricci, T., ... & Lancia, M. (2016). Coseismic effects of the 2016 Amatrice seismic sequence: First geological results. Annals of geophysics.
- Enemark, T., Peeters, L. J., Mallants, D., & Batelaan, O. (2019). Hydrogeological conceptual model building and testing: A review. Journal of Hydrology, 569, 310-329.
- Epstein, S., & Mayeda, T. (1953). Variation of O18 content of waters from natural sources. Geochimica et cosmochimica acta, 4(5), 213-224.
- Eskelinen, R., Ronkanen, A. K., Marttila, H., Isokangas, E., & Kløve, B. (2016). Effects of soil frost on snowmelt runoff generation and surface water quality in drained peatlands.
- Esposito, E., Pece, R., Porfido, S., & Tranfaglia, G. (2001). Hydrological anomalies connected to earthquakes in southern Apennines (Italy). Natural Hazards and Earth System Sciences, 1(3), 137-144.
- Esposito, E., Pece, R., Porfido, S., & Tranfaglia, G. (2009). Ground effects and hydrological changes in the Southern Apennines (Italy) in response to the 23 July 1930 earthquake (M S= 6.7). Natural Hazards and Earth System Sciences, 9(2), 539-550.
- Evans, J. P., Forster, C. B., & Goddard, J. V. (1997). Permeability of fault-related rocks, and implications for hydraulic structure of fault zones. Journal of structural Geology, 19(11), 1393-1404.
- Fairley, J., Heffner, J., & Hinds, J. (2003). Geostatistical evaluation of permeability in an active fault zone. Geophysical Research Letters, 30(18).
- Falcone, R. A., Falgiani, A., Parisse, B., Petitta, M., Spizzico, M., & Tallini, M. (2008). Chemical and isotopic ($\delta^{18}\text{O}\%$, $\delta^2\text{H}\%$, $\delta^{13}\text{C}\%$, ^{222}Rn) multi-tracing for groundwater conceptual model of carbonate aquifer (Gran Sasso INFN underground laboratory—central Italy). Journal of Hydrology, 357(3-4), 368-388.

- Faulkner, D. R., Jackson, C. A. L., Lunn, R. J., Schlische, R. W., Shipton, Z. K., Wibberley, C. A. J., & Withjack, M. O. (2010). A review of recent developments concerning the structure, mechanics and fluid flow properties of fault zones. *Journal of Structural Geology*, 32(11), 1557-1575.
- Feng, X., Faiia, A. M., & Posmentier, E. S. (2009). Seasonality of isotopes in precipitation: A global perspective. *Journal of Geophysical Research: Atmospheres*, 114(D8).
- Feng, X., Taylor, S., Renshaw, C. E., & Kirchner, J. W. (2002). Isotopic evolution of snowmelt 1. A physically based one-dimensional model. *Water Resources Research*, 38(10), 35-1.
- Field, M. S. (2002). The QTRACER2 program for tracer-breakthrough curve analysis for tracer tests in karstic aquifers and other hydrologic systems. National Center for Environmental Assessment--Washington Office, Office of Research and Development, US Environmental Protection Agency.
- Field, M. S., & Pinsky, P. F. (2000). A two-region nonequilibrium model for solute transport in solution conduits in karstic aquifers. *Journal of Contaminant Hydrology*, 44(3-4), 329-351.
- Field, M.S. (2003). A review of some tracer-test design equations for tracer-mass estimation and sample collection frequency. *Environmental Geology*, 43(8): 867-881.
- Finetti, I. R., Calamita, F., Crescenti, U., Del Ben, A., Forlin, E., Pipan, M., ... & Scisciani, V. (2005). Crustal geological section across central Italy from the Corsica Basin to the Adriatic Sea based on geological and CROP seismic data. CROP Project: deep seismic exploration of the Central Mediterranean and Italy, 159-197.
- Fiorillo, F. (2014). The recession of spring hydrographs, focused on karst aquifers. *Water resources management*, 28(7), 1781-1805.
- Fiorillo, F., & Guadagno, F. M. (2010). Karst spring discharges analysis in relation to drought periods, using the SPI. *Water resources management*, 24(9), 1867-1884.
- Fiorillo, F., & Guadagno, F. M. (2012). Long karst spring discharge time series and droughts occurrence in Southern Italy. *Environmental Earth Sciences*, 65(8), 2273-2283.
- Fiorillo, F., Pagnozzi, M., Stevanović, Z., & Ventafridda, G. (2019). Main hydrological features and recharge analysis of the Caposele Spring catchment, southern Italy. *Acta Carsologica*, 48(1).
- Fiorillo, F., Petitta, M., Preziosi, E., Rusi, S., Esposito, L., & Tallini, M. (2015). Long-term trend and fluctuations of karst spring discharge in a Mediterranean area (central-southern Italy). *Environmental Earth Sciences*, 74(1), 153-172.
- Fischer, D. G. J. (2012). Palaeoecology, migration behavior, and reproductive pattern of Palaeozoic to Mesozoic freshwater sharks revealed by stable isotopes (Doctoral dissertation, Universität Wien).
- Fontes, J. C., & Edmunds, W. M. (1985). The use of environmental isotope techniques in arid zone hydrology: A critical review.
- Ford, D. C., & Williams, P. W. (1989). *Karst geomorphology and hydrology* (Vol. 601, No. 1). London: Unwin Hyman.
- Forkasiewicz, J., Paloc, H. (1967). Le régime de tarissement de la Foux-de-la-Vis [The recession regime of the Foux-de-la-Vis]. *Chron Hydrogéol* 3(10):61-73.
- Friedman, I., Machta, L., & Soller, R. (1962). Water-vapor exchange between a water droplet and its environment. *Journal of Geophysical Research*, 67(7), 2761-2766.

- Friedman, I., Smith, G. I., Gleason, J. D., Warden, A., & Harris, J. M. (1992). Stable isotope composition of waters in southeastern California 1. Modern precipitation. *Journal of Geophysical Research: Atmospheres*, 97(D5), 5795-5812.
- Fritz, P., Drimmie, R. J., Frape, S. K., & O'Shea, K. (1987). The isotopic composition of precipitation and groundwater in Canada. In *Isotope techniques in water resources development* (pp. 539–550). Vienna, Austria: International Atomic Energy Agency.
- Fronzi D., Banzato F., Caliro S., Cambi C., Cardellini C., Checcucci R., Mastroiillo L., Mirabella F., Petitta M., Valigi D. & Tazioli A. (2020). Preliminary results on the response of some springs of the Sibillini Mountains area to the 2016–2017 seismic sequence. *Acque Sotter. Ital. J. Groundw.*, 9(1), 19-25.
- Fronzi D., Di Curzio D., Rusi S., Valigi D. & Tazioli A. (2020). Comparison between periodic tracer tests and time-series analysis to assess mid- and long-term recharge model changes due to multiple strong seismic events in carbonate aquifers. *Water*, 12(11), 3073.
- Fronzi D., Mirabella F., Cardellini, C., Caliro S., Palpacelli S., Cambi C., Valigi D. & Tazioli A. (2021). The Role of Faults in Groundwater Circulation before and after Seismic Events: Insights from Tracers, Water Isotopes and Geochemistry. *Water*, 13(11), 1499.
- Galadini, F., & Galli, P. (2003). Paleoseismology of silent faults in the Central Apennines (Italy): the Mt. Vettore and Laga Mts. faults. *Annals of Geophysics*.
- Galdenzi, S., & Menichetti, M. (1995). Occurrence of hypogenic caves in a karst region: examples from central Italy. *Environmental Geology*, 26(1), 39-47.
- Galdini, F., Pantosti, D., Galli, P., Messina, P., Montone, P., Pizzi, A. & Salvi, S. (2009). Il terremoto del 6 aprile e le conoscenze sulle faglie attive dell'Appennino centrale, *Progettazione Sismica N.3*.
- Galli, P., Galadini, F., & Calzoni, F. (2005). Surface faulting in Norcia (central Italy): a “paleoseismological perspective”. *Tectonophysics*, 403(1-4), 117-130.
- Gat, J. R. (1996). Oxygen and hydrogen isotopes in the hydrologic cycle. *Annual Review of Earth and Planetary Sciences*, 24(1), 225-262.
- Gat, J. R., & Dansgaard, W. (1972). Stable isotope survey of the fresh water occurrences in Israel and the northern Jordan Rift Valley. *Journal of Hydrology*, 16(3), 177-211.
- Gat, J. R., & Matsui, E. (1991). Atmospheric water balance in the Amazon Basin: an isotopic evapotranspiration model. *Journal of Geophysical Research: Atmospheres*, 96(D7), 13179-13188.
- Gat, J. R., & Tzur, Y. (1968). Modification of the isotopic composition of rainwater by processes which occur before groundwater recharge. Weizmann Inst. of Science, Rehovoth, Israel.
- Gat, J. R., Bowser, C. J., & Kendall, C. (1994). The contribution of evaporation from the Great Lakes to the continental atmosphere: estimate based on stable isotope data. *Geophysical Research Letters*, 21(7), 557-560.
- Gat, J. R., Klein, B., Kushnir, Y., Roether, W., Wernli, H., Yam, R., & Shemesh, A. (2011). Isotope composition of air moisture over the Mediterranean Sea: an index of the air-sea interaction pattern. *Tellus B: Chemical and Physical Meteorology*, 55(5), 953-965.
- Gedeon, M., Mallants, D., Rogiers, B., 2013. Building a staircase of confidence in groundwater modeling: a summary of ten years data collection and model development. *Modflow and More Conference. Translating Science into Practice*, Golden, CO.

- Giacopetti, M., Aringoli, D., Materazzi, M., Pambianchi, G., & Posavec, K. (2016). Groundwater recharge estimation using spring hydrographs: The case of the Tennacola carbonate aquifer (central Apennine, Italy). *RENDICONTI ONLINE SOCIETA GEOLOGICA ITALIANA*, 41, 61-64.
- Giggenbach, W. F. (1978). The isotopic composition of waters from the El Tatio geothermal field, Northern Chile. *Geochimica et Cosmochimica Acta*, 42(7), 979-988.
- Giggenbach, W. F., & Stewart, M. K. (1982). Processes controlling the isotopic composition of steam and water discharges from steam vents and steam-heated pools in geothermal areas. *Geothermics*, 11(2), 71-80.
- Giggenbach, W. F., Gonfiantini, R., Jangi, B. L., & Truesdell, A. H. (1983). Isotopic and chemical composition of Parbati valley geothermal discharges, north-west Himalaya, India. *Geothermics*, 12(2-3), 199-222.
- Giovagnotti, C. (1979). Lineamenti paleogeografici e geomorfologici dei Monti Sibillini. *Biogeographia—The Journal of Integrative Biogeography*, 6(1).
- Giustini, F., Brilli, M., & Patera, A. (2016). Mapping oxygen stable isotopes of precipitation in Italy. *Journal of Hydrology: Regional Studies*, 8, 162-181.
- Goldscheider, N. (2005). Karst groundwater vulnerability mapping: application of a new method in the Swabian Alb, Germany. *Hydrogeology Journal*, 13(4), 555-564.
- Goldscheider, N., & Drew, D. (Eds.). (2014). *Methods in Karst Hydrogeology: IAH: International Contributions to Hydrogeology*, 26. Crc Press.
- Goldscheider, N., & Neukum, C. (2010). Fold and fault control on the drainage pattern of a double-karst-aquifer system, Winterstaude, Austrian Alps. *Acta carsologica*, 39(2).
- Goldscheider, N., Meiman, J., Pronk, M., & Smart, C. (2008). Tracer tests in karst hydrogeology and speleology. *International Journal of speleology*, 37(1), 27-40.
- Gonfiantini, R., Roche, M. A., Olivry, J. C., Fontes, J. C., & Zuppi, G. M. (2001). The altitude effect on the isotopic composition of tropical rains. *Chemical Geology*, 181(1-4), 147-167.
- Gregor, B. M. (2010). *Bfi+ 3.0. User's Manual*. Department of Hydrogeology, Faculty of Natural Science, Comenius University, Bratislava, Slovakia. 21p.
- Guttman, N. B. (1998). Comparing the palmer drought index and the standardized precipitation index 1. *JAWRA Journal of the American Water Resources Association*, 34(1), 113-121.
- Guttman, N. B. (1999). Accepting the standardized precipitation index: a calculation algorithm 1. *JAWRA Journal of the American Water Resources Association*, 35(2), 311-322.
- Hammarlund, D., & Edwards, T. W. (2008). Stable isotope variations in stalagmites from northwestern Sweden document changes in temperature and vegetation during the early Holocene: a comment on Sundqvist et al. 2007a. *The Holocene*, 18(6), 1007-1008.
- Hargreaves, G. H., & Samani, Z. A. (1985). Reference crop evapotranspiration from temperature. *Applied engineering in agriculture*, 1(2), 96-99.
- Hartmann, A., Goldscheider, N., Wagener, T., Lange, J., & Weiler, M. (2014). Karst water resources in a changing world: Review of hydrological modeling approaches. *Reviews of Geophysics*, 52(3), 218-242.
- Hartmann, A., Mudarra, M., Andreo, B., Marín, A., Wagener, T., & Lange, J. (2014). Modeling spatiotemporal impacts of hydroclimatic extremes on groundwater recharge at a Mediterranean karst aquifer. *Water Resources Research*, 50(8), 6507-6521.

- Hashemi, F., & Habibian, M. T. (1979). Limitations of temperature-based methods in estimating crop evapotranspiration in arid-zone agricultural development projects. *Agricultural Meteorology*, 20(3), 237-247.
- Hawley, R. L., & Millstein, J. D. (2019). Quantifying snow drift on Arctic structures: A case study at Summit, Greenland, using UAV-based structure-from-motion photogrammetry. *Cold Regions Science and Technology*, 157, 163-170.
- Herrera, C., Custodio, E., Chong, G., Lambán, L. J., Riquelme, R., Wilke, H., ... & Lictevout, E. (2016). Groundwater flow in a closed basin with a saline shallow lake in a volcanic area: Laguna Tuyajto, northern Chilean Altiplano of the Andes. *Science of The Total Environment*, 541, 303-318.
- Holzer, T. L., & Youd, T. L. (2007). Liquefaction, ground oscillation, and soil deformation at the Wildlife Array, California. *Bulletin of the Seismological society of America*, 97(3), 961-976.
- Hosono, T., Yamada, C., Shibata, T., Tawara, Y., Wang, C. Y., Manga, M., ... & Shimada, J. (2019). Coseismic groundwater drawdown along crustal ruptures during the 2016 Mw 7.0 Kumamoto earthquake. *Water Resources Research*, 55(7), 5891-5903.
- Hürkamp, K., Zentner, N., Reckerth, A., Weishaupt, S., Wetzel, K. F., Tschiersch, J., & Stumpp, C. (2019). Spatial and temporal variability of snow isotopic composition on Mt. Zugspitze, Bavarian Alps, Germany. *Journal of Hydrology and Hydromechanics*, 67(1), 49-58.
- Imanian, A., Sheykhi, V., Zare, K., & Zare, M. (2020). Introducing the recharge–discharge relationship to evaluate the recharge coefficient of karstic aquifers: thematic and indexing approaches. *Hydrological Sciences Journal*, 65(8), 1385-1398.
- Ingebritsen, S. E., & Manga, M. (2019). Earthquake hydrogeology. *Water Resources Research*, 55(7), 5212-5216.
- Izady, A., Davary, K., Alizadeh, A., Ziaei, A.N., Alipoor, A., Joodavi, A., Brusseau, M.L. (2014). A framework toward developing a groundwater conceptual model. *Arab. Journal Geosci.* 7, 3611–3631. <https://doi.org/10.1007/s12517-013-0971-9>.
- Jasechko, S. (2019). Global isotope hydrogeology—Review. *Reviews of Geophysics*, 57(3), 835-965.
- Jasechko, S., Gibson, J. J., & Edwards, T. W. (2014). Stable isotope mass balance of the Laurentian Great Lakes. *Journal of Great Lakes Research*, 40(2), 336-346.
- Jasechko, S., Wassenaar, L. I., & Mayer, B. (2017). Isotopic evidence for widespread cold-season-biased groundwater recharge and young streamflow across central Canada. *Hydrological Processes*, 31(12), 2196-2209.
- Jeannin, P. Y., Eichenberger, U., Sinreich, M., Vouillamoz, J., Malard, A., & Weber, E. (2013). KARSYS: a pragmatic approach to karst hydrogeological system conceptualisation. Assessment of groundwater reserves and resources in Switzerland. *Environmental Earth Sciences*, 69(3), 999-1013.
- Jeannin, P. Y., & Malard, A. (2018). A Way to Predict Natural Hazards in Karst.
- Jeannin, P.-Y., Sauter, M. (1998). Analysis of karst hydrodynamic behavior using global approaches: a review. *Bull Hydrogéol Univ Neuchâtel* 16:31–48
- Jódar, J., Custodio, E., Lambán, L. J., Martos-Rosillo, S., Herrera-Lameli, C., & Sapriza-Azuri, G. (2016). Vertical variation in the amplitude of the seasonal isotopic content of rainfall as a tool to jointly estimate the groundwater recharge zone and transit times in the Ordesa and Monte Perdido National Park aquifer system, north-eastern Spain. *Science of the total environment*, 573, 505-517.
- Käss, W. (1998), *Tracing technique in geohydrology*. Balkema, Rotterdam

- Kattan, Z. (2001). Use of hydrochemistry and environmental isotopes for evaluation of groundwater in the Paleogene limestone aquifer of the Ras Al-Ain area (Syrian Jezireh). *Environmental Geology*, 41(1-2), 128-144.
- Kattan, Z. (2006). Characterization of surface water and groundwater in the Damascus Ghotia basin: hydrochemical and environmental isotopes approaches. *Environmental Geology*, 51(2), 173-201.
- Katz, B. G., Sepulveda, A. A., & Verdi, R. J. (2009). Estimating nitrogen loading to ground water and assessing vulnerability to nitrate contamination in a large karstic springs basin, Florida 1. *JAWRA Journal of the American Water Resources Association*, 45(3), 607-627.
- Kebede, S., Travi, Y., Alemayehu, T., & Ayenew, T. (2005). Groundwater recharge, circulation and geochemical evolution in the source region of the Blue Nile River, Ethiopia. *Applied Geochemistry*, 20(9), 1658-1676.
- Kern, Z., Hatvani, I. G., Czuppon, G., Fórizs, I., Erdélyi, D., Kanduč, T., ... & Vreča, P. (2020). Isotopic 'altitude' and 'continental' effects in modern precipitation across the Adriatic-Pannonian Region. *Water*, 12(6), 1797.
- Kiraly, L. (1975). Rapport sur l'état actuel des connaissances dans le domaines des caractères physiques des roches karstiques, edited by: Burger, A. and Dubertret, L., *Hydrogeology of karstic terrains*, Int. Union of Geol. Sci. B, 3, 53-67.
- Király, L. (2002). Karstification and Groundwater Flow. In: *Proceedings of the Conference on Evolution of Karst: From Prekarst to Cessation*. Postojna-Ljubljana. 155-190.
- Kiraly, L., & Morel, G. (1976). Etude de régularisation de l'Areuse par modèle mathématique [Regulation of the Areuse spring by mathematical modelling]. *Bull Hydrogéol Univ Neuchâtel* 1:19-36
- Király, L., Matthey, B., Tripet, J.P. (1971). Fissuration et orientation des cavités souterraines. Région de la Grotte Milandre. *Bull. Soc. Neuchât. Sci. Nat.* 94, 99-114.
- Kiraly, L., Perrochet, P., Rossier, Y. (1995). Effect of the epikarst on the hydrograph of karst springs: a numerical approach. *Bull. Centre d'Hydrogéol.* 14, 199-220.
- Koeniger, P., Hubbart, J. A., Link, T., & Marshall, J. D. (2008). Isotopic variation of snow cover and streamflow in response to changes in canopy structure in a snow-dominated mountain catchment. *Hydrological Processes: An International Journal*, 22(4), 557-566.
- Korup, O., Mohr, C. H., & Manga, M. M. (2021) Bayesian Detection of Streamflow Response to Earthquakes. *Water Resources Research*, e2020WR028874.
- Kovács, A. (2003). Geometry and hydraulic parameters of karst aquifers: a hydrodynamic modeling approach (Doctoral dissertation, Université de Neuchâtel).
- Kovács, A. (2021). Quantitative classification of carbonate aquifers based on hydrodynamic behaviour. *Hydrogeology Journal*, 29(1), 33-52.
- Kovács, A., Perrochet, P. (2008). A quantitative approach to spring hydrograph decomposition. *J Hydrol* 352(1-2):16-29
- Kovács, A., Perrochet, P. (2014.) Well hydrograph analysis for the estimation of hydraulic and geometric parameters of karst aquifers. In: *H2Karst research in limestone hydrogeology: environmental earth sciences*. Springer, Heidelberg, Germany, pp 97-114
- Kovács, A., Perrochet, P., Darabos, E., Lénárt, L., Szűcs, P. (2015). Well hydrograph analysis for the characterisation of flow dynamics and conduit network geometry in a karstic aquifer, Bükk Mountains, Hungary. *J Hydrol* 530:484-499

- Kovács, A., Perrochet, P., Kiraly, L., Jeannin, P.-Y. (2005). A quantitative method for the characterisation of karst aquifers based on spring hydrograph analysis. *J Hydrol* 303(1–4):152–164
- Kresic, N. (2007). *Hydrogeology and groundwater modeling*, 2nd edn. CRC, Boca Raton, FL
- Kresic, N., Bonacci, O. (2010). Spring discharge hydrograph. In: Kresic N, Stevanovic Z (eds) *Groundwater hydrology of springs: engineering, theory, management, and sustainability*. Elsevier, Amsterdam, 561 pp
- Kurita, N., Yoshida, N., Inoue, G., & Chayanova, E. A. (2004). Modern isotope climatology of Russia: A first assessment. *Journal of Geophysical Research: Atmospheres*, 109(D3).
- La Vigna, F. (2013). Earthquake hydrology. *Acque Sotterranee-Italian Journal of Groundwater*, 2(2).
- Lai, W. C., Koizumi, N., Matsumoto, N., Kitagawa, Y., Lin, C. W., Shieh, C. L., & Lee, Y. P. (2004). Effects of seismic ground motion and geological setting on the coseismic groundwater level changes caused by the 1999 Chi-Chi earthquake, Taiwan. *Earth, planets and space*, 56(9), 873-880.
- Lancia, M., Petitta, M., Zheng, C., & Saroli, M. (2020). Hydrogeological insights and modelling for sustainable use of a stressed carbonate aquifer in the Mediterranean area: From passive withdrawals to active management. *Journal of Hydrology: Regional Studies*, 32, 100749.
- Laudon, H., Hemond, H. F., Krouse, R., & Bishop, K. H. (2002). Oxygen 18 fractionation during snowmelt: Implications for spring flood hydrograph separation. *Water Resources Research*, 38(11), 40-1.
- Lavecchia, G. (1985). Il sovrascorrimento dei Monti Sibillini; analisi cinematica e strutturale. *Bollettino della Società Geologica Italiana*, 104(1), 161-194.
- Lavecchia, G., Boncio, P., Creati, N., & Brozzetti, F. (2003). Some aspects of the Italian geology not fitting with a subduction scenario. *J. Virtual Explorer*, 10, 1-14.
- Lavecchia, G., Brozzetti, F., Barchi, M., Menichetti, M., & Keller, J. V. (1994). Seismotectonic zoning in east-central Italy deduced from an analysis of the Neogene to present deformations and related stress fields. *Geological Society of America Bulletin*, 106(9), 1107-1120.
- Lavecchia, G., Castaldo, R., De Nardis, R., De Novellis, V., Ferrarini, F., Pepe, S., ... & Tizzani, P. (2016). Ground deformation and source geometry of the 24 August 2016 Amatrice earthquake (Central Italy) investigated through analytical and numerical modeling of DInSAR measurements and structural-geological data. *Geophysical Research Letters*, 43(24), 12-389.
- Lechler, A. R., & Niemi, N. A. (2012). The influence of snow sublimation on the isotopic composition of spring and surface waters in the southwestern United States: Implications for stable isotope-based paleoaltimetry and hydrologic studies. *Bulletin*, 124(3-4), 318-334.
- Lee, J. E., & Fung, I. (2008). "Amount effect" of water isotopes and quantitative analysis of post-condensation processes. *Hydrological Processes: An International Journal*, 22(1), 1-8.
- Lee, J. E., Fung, I., DePaolo, D. J., & Henning, C. C. (2007). Analysis of the global distribution of water isotopes using the NCAR atmospheric general circulation model. *Journal of Geophysical Research: Atmospheres*, 112(D16).
- Lee, J., Feng, X., Faiia, A. M., Posmentier, E. S., Kirchner, J. W., Osterhuber, R., & Taylor, S. (2010). Isotopic evolution of a seasonal snowcover and its melt by isotopic exchange between liquid water and ice. *Chemical geology*, 270(1-4), 126-134.
- Leibundgut, C., Malozewski, P., Külls, C. 2009. *Tracers in Hydrology*, John Wiley & Sons: Chichester.

- Lippi Boncambi C. (1947). Idrologia sotterranea dell'Altipiano di Castelluccio. *Annali Istituto di Mineralogia e Geologia dell'Università degli Studi di Perugia*, 104-118.
- Lippi Boncambi C. (1948). Monti Sibillini. *Tipografia Mareggiani*, Bologna, 77 pp.
- Lippi Boncambi C. (1963). Indagini idrogeologiche sull'altipiano di Castelluccio di Norcia. In *Atti del IX Congresso Nazionale di Speleologia*, Trieste.
- Liu, J., Yang, H., Gosling, S. N., Kumm, M., Flörke, M., Pfister, S., ... & Oki, T. (2017). Water scarcity assessments in the past, present, and future. *Earth's future*, 5(6), 545-559.
- Livio, F. A., Michetti, A. M., Vittori, E., Gregory, L., Wedmore, L., Piccardi, L., ... & Wilkinson, M. (2016). Surface faulting during the August 24, 2016, Central Italy earthquake (Mw 6.0): preliminary results. *Annals of geophysics*, 59(5).
- Longinelli, A., & Selmo, E. (2003). Isotopic composition of precipitation in Italy: a first overall map. *Journal of Hydrology*, 270(1-2), 75-88.
- Madonna, R., Signanini, P., Crema, G., Di Sabatino, B., Rainone, M. L., & Di Nunzio, A. (2005). The geothermal area of Acquasanta Terme (Central Italy): Main characteristics and an attempt of field evaluation. In *Proceedings of the World Geothermal Congress* (pp. 24-29).
- Maillet, E. T. (1905). *Essais d'hydraulique souterraine & fluviale*. A. Hermann.
- Malard, A., Vouillamoz, J., Jeannin, P. Y., Weber, E., & Eichenberger, U. (2012, April). SWISSKARST Project-how to document the karst aquifers in Switzerland using the KARSYS approach. In *EGU General Assembly Conference Abstracts* (p. 2637).
- Malek, E. (1987). Comparison of alternative methods for estimating ETp and evaluation of advection in the Bajgah area, Iran. *Agricultural and forest meteorology*, 39(2-3), 185-192.
- Malinverno, A., & Ryan, W. B. (1986). Extension in the Tyrrhenian Sea and shortening in the Apennines as result of arc migration driven by sinking of the lithosphere. *Tectonics*, 5(2), 227-245.
- Maloszewski, P., Harum, T., & Benischke, R. (1992). Mathematical modelling of tracer experiments in the karst of Lurbach system. *Steierische Beiträge zur Hydrogeologie*, 43, 116-136.
- Mammoliti, E., Fronzi, D., Mancini, A., Valigi, D., & Tazioli, A. (2021). WaterbalANce, a WebApp for Thornthwaite-Mather Water Balance Computation: Comparison of Applications in Two European Watersheds. *Hydrology*, 8(1), 34.
- Manga, M. (2001). Origin of postseismic streamflow changes inferred from baseflow recession and magnitude-distance relations. *Geophysical Research Letters*, 28(10), 2133-2136.
- Manga, M., & Rowland, J. C. (2009). Response of Alum Rock springs to the October 30, 2007 Alum Rock earthquake and implications for the origin of increased discharge after earthquakes. *Geofluids*, 9(3), 237-250.
- Manga, M., & Wang, C. Y. (2007). Pressurized oceans and the eruption of liquid water on Europa and Enceladus. *Geophysical Research Letters*, 34(7).
- Manga, M., Brodsky, E. E., & Boone, M. (2003). Response of streamflow to multiple earthquakes. *Geophysical Research Letters*, 30(5).
- Mangin, A. (1975). Contribution à l'étude hydrodynamique des aquifères karstiques. PhD Thèse, Université de Dijon, p. 124.

- Mantovani, E., Babbucci, D., Tamburelli, C., & Viti, M. (2009). A review on the driving mechanism of the Tyrrhenian–Apennines system: Implications for the present seismotectonic setting in the Central-Northern Apennines. *Tectonophysics*, 476(1-2), 22-40.
- Martin, J. B., & White, W. B. (2008). *Frontiers of karst research*, special publication 13. Lessburg, VA: Karst Warer Institute.
- Martos-Rosillo, S., González-Ramón, A., Jiménez-Gavilán, P., Andreo, B., Durán, J. J., & Mancera, E. (2015). Review on groundwater recharge in carbonate aquifers from SW Mediterranean (Betic Cordillera, S Spain). *Environmental Earth Sciences*, 74(12), 7571-7581.
- Mastorillo L. (1996). Contributo alla valutazione delle risorse idriche sotterranee dell'Appennino carbonatico marchigiano. *Quaderni Di Geologia Applicata*. 1/1996: 25-35.
- Mastorillo L. (2001). Elementi strutturali e caratteristiche idrogeologiche della dorsale carbonatica Umbro - Marchigiana interna. *Mem. Soc. Geol. It.* 56: 219-226.
- Mastorillo L., Baldoni T., Banzato F., Boscherini A., Cascone D., Checcucci R., ... & Boni C. (2009). Analisi idrogeologica quantitativa del dominio carbonatico umbro. *Italian Journal of Engineering Geology and Environment*, 1, 137-155.
- Mastorillo L., Saroli M., Viaroli S., Banzato F., Valigi D. & Petitta M. (2020). Sustained post-seismic effects on groundwater flow in fractured carbonate aquifers in Central Italy. *Hydrol. Process.*, 34(5), 1167-1181.
- Mastorillo, L., Viaroli, S., Banzato, F., Bolzonella, E., Bufalini, M., Ianni, C., Materazzi, M., Petitta, M. & Tonelli M. (2021). Co-occurrence of climatic and post-seismic events depleting groundwater resources of fractured carbonate aquifers of Sibillini Mts. (Central Italy). *Flowpath 2021 – National Meeting on Hydrogeology – Naples, Italy*.
- Mathieu, R., & Bariac, T. (1996). An isotopic study (2H and 18O) of water movements in clayey soils under a semiarid climate. *Water Resources Research*, 32(4), 779-789.
- Matsumoto, N., Kitagawa, G., & Roeloffs, E. A. (2003). Hydrological response to earthquakes in the Haibara well, central Japan–I. Groundwater level changes revealed using state space decomposition of atmospheric pressure, rainfall and tidal responses. *Geophysical Journal International*, 155(3), 885-898.
- Mazzoli, S., Cello, G., Deiana, G., Galdenzi, S., Gambini, R., Mancinelli, A., ... & Tondi, E. (2000). Modes of foreland deformation ahead of the Apennine thrust front. *Journal of the Czech Geological Society*, 45(3-4), 246.
- Mazzoli, S., Pierantoni, P. P., Borraccini, F., Paltrinieri, W., & Deiana, G. (2005). Geometry, segmentation pattern and displacement variations along a major Apennine thrust zone, central Italy. *Journal of Structural Geology*, 27(11), 1940-1953.
- McKee, T. B., Doesken, N. J., & Kleist, J. (1993, January). The relationship of drought frequency and duration to time scales. In *Proceedings of the 8th Conference on Applied Climatology* (Vol. 17, No. 22, pp. 179-183).
- Medici, G., Smeraglia, L., Torabi, A., & Botter, C. (2021). Review of modeling approaches to groundwater flow in deformed carbonate aquifers. *Groundwater*, 59(3), 334-351.
- Mercer, J. J., Liefert, D. T., & Williams, D. G. (2020). Atmospheric vapour and precipitation are not in isotopic equilibrium in a continental mountain environment. *Hydrological Processes*, 34(14), 3078-3101.
- Merlivat, L. (1978). The dependence of bulk evaporation coefficients on air-water interfacial conditions as determined by the isotopic method. *Journal of Geophysical Research: Oceans*, 83(C6), 2977-2980.

- Merlivat, L., & Jouzel, J. (1979). Global climatic interpretation of the deuterium-oxygen 18 relationship for precipitation. *Journal of Geophysical Research: Oceans*, 84(C8), 5029-5033.
- Messerli, B., Viviroli, D., & Weingartner, R. (2004). Mountains of the world: vulnerable water towers for the 21st century. *Ambio*, 29-34.
- Mirabella, F., Barchi, M., Lupattelli, A., Stucchi, E., & Ciaccio, M. G. (2008). Insights on the seismogenic layer thickness from the upper crust structure of the Umbria-Marche Apennines (central Italy). *Tectonics*, 27(1).
- Miur-Wood, R. & King, G.C.P. (1993). Hydrological signatures of earthquake strain, *J. Geophys. Res.*, 98, No B12, 20.035-20.068.
- Mohr, C. H., Manga, M., Wang, C. Y., & Korup, O. (2017). Regional changes in streamflow after a megathrust earthquake. *Earth and Planetary Science Letters*, 458, 418-428.
- Montgomery, D. R., & Manga, M. (2003). Streamflow and water well responses to earthquakes. *Science*, 300(5628), 2047-2049.
- Montone, P., & Mariucci, M. T. (2020). Constraints on the structure of the shallow crust in central Italy from geophysical log data. *Scientific reports*, 10(1), 1-12.
- Mook, W., & Rozanski, K. (2000). Environmental isotopes in the hydrological cycle. IAEA Publish, 39.
- Moser, H., & Stichler, W. (1974). Deuterium and oxygen-18 contents as an index of the properties of snow covers. *International association of hydrological sciences publication*, 114, 122-135.
- Motyka, J. (1998). A conceptual model of hydraulic networks in carbonate rocks, illustrated by examples from Poland. *Hydrogeology Journal*, 6(4), 469-482.
- Mountain, N., James, A. L., & Chutko, K. (2015). Groundwater and surface water influences on streamflow in a mesoscale Precambrian Shield catchment. *Hydrological Processes*, 29(18), 3941-3953.
- Mudarra, M., & Andreo, B. (2011). Relative importance of the saturated and the unsaturated zones in the hydrogeological functioning of karst aquifers: The case of Alta Cadena (Southern Spain). *Journal of hydrology*, 397(3-4), 263-280.
- Mudarra, M., Hartmann, A., & Andreo, B. (2019). Combining experimental methods and modeling to quantify the complex recharge behavior of karst aquifers. *Water Resources Research*, 55(2), 1384-1404.
- Mussi, M., Nanni, T., Tazioli, A., & Vivalda, P. M. (2017). The Mt Conero limestone ridge: The contribution of stable isotopes to the identification of the recharge area of aquifers. *Italian Journal of Geosciences*, 136(2), 186-197.
- Nanni T. & Petitta M. (2012). Studio del bacino idrogeologico carbonatico (Monti Sibillini) del fiume Aso finalizzato alla caratterizzazione delle condizioni di alimentazione e di circolazione idrica, alla valutazione del bacino idrogeologico e analisi della sorgente di Foce. Contratto di studio e ricerca. Dipartimento SIMAU Università Politecnica delle Marche, CIIP Spa e Dipartimento Scienze della Terra Università di Roma "La Sapienza".
- Nanni T. & Vivalda P. (1991). Caratteri idrogeologici delle sorgenti carbonatiche dell'Appennino marchigiano. *Atti Conv. "Ricerca e protezione delle risorse idriche sotterranee delle aree montuose"*, Fondazione Bresciana Ric. Scient., 42 (1), 269-331, 24-25 ottobre 1991, Brescia.
- Nanni T. & Vivalda P. (2006). Spring monitoring and analysis of groundwater circulation in the Sibillini Mountains aquifers. (Adriatic side of central Apennine, Italy). Conference: 8th Conference on limestone hydrogeology 2006, 197-200, Neuchatel, Switzerland, September 21-23. ISBN 2-84867-143-2 978-2-84867-143-2.

- Nanni T., Vivalda P.M., Palpacelli S., Marcellini M., Tazioli, A. (2020). Groundwater circulation and earthquake related changes in hydrogeological karst environments: a case study of the Sibillini Mountains (central Italy) involving artificial tracers. *Hydrogeol. J.* 1–20, doi:10.1007/s10040-020-02207-w
- Nanni, T. & Vivalda, M.P. (2005). The aquifers of the Umbria-Marche Adriatic region: relationships between structural setting and groundwater chemistry. *BOLLETTINO-SOCIETA GEOLOGICA ITALIANA*, 124(3), 523.
- Nanni, T., & Rusi, S. (2003). Idrogeologia del massiccio carbonatico della montagna della Majella (Appennino centrale). *Bollettino-Società Geologica Italiana*, 122(2), 173-202.
- Nanni, T., Tazioli, A., & Vivalda, P. M. (2013). Problems in the application of environmental isotopes to the hydrogeology of high mountain aquifers. *AquaMUNDI*, 4, 55-66.
- Négrel, P., & Petelet-Giraud, E. (2005). Strontium isotopes as tracers of groundwater-induced floods: the Somme case study (France). *Journal of Hydrology*, 305(1-4), 99-119.
- Neuman, S. P., Wierenga, P. J., & Nicholson, T. J. (2003). A comprehensive strategy of hydrogeologic modeling and uncertainty analysis for nuclear facilities and sites. Division of Systems Analysis and Regulatory Effectiveness, Office of Nuclear Regulatory Research, US Nuclear Regulatory Commission.
- Niewodnizański, J., Grabczak, J., Barański, L., & Rzepka, J. (1981). The altitude effect on the isotopic composition of snow in high mountains. *Journal of Glaciology*, 27(95), 99-111.
- Niraula, R., Meixner, T., Ajami, H., Rodell, M., Gochis, D., & Castro, C. L. (2017). Comparing potential recharge estimates from three Land Surface Models across the western US. *Journal of hydrology*, 545, 410-423.
- Pace, P., & Calamita, F. (2014). Push-up inversion structures v. fault-bend reactivation anticlines along oblique thrust ramps: examples from the Apennines fold-and-thrust belt (Italy). *Journal of the Geological Society*, 171(2), 227-238.
- Pavlovskii, I., Hayashi, M., & Lennon, M. R. (2018). Transformation of snow isotopic signature along groundwater recharge pathways in the Canadian Prairies. *Journal of Hydrology*, 563, 1147-1160.
- Payne, B. R., & Yurtsever, Y. (1974). Environmental isotopes as a hydrogeological tool in Nicaragua. In *Isotope techniques in groundwater hydrology 1974*, Vol. I. Proceedings of a symposium.
- Pelton, W. L., King, K. M., & Tanner, C. B. (1960). An evaluation of the Thornthwaite and mean temperature methods for determining potential evapotranspiration 1. *Agronomy Journal*, 52(7), 387-395
- Penna, D., Ahmad, M., Birks, S. J., Bouchaou, L., Brenčič, M., Butt, S., ... & Vreča, P. (2014). A new method of snowmelt sampling for water stable isotopes. *Hydrological Processes*, 28(22), 5637-5644.
- Perrin, J., Jeannin, P. Y., & Zwahlen, F. (2003). Implications of the spatial variability of infiltration-water chemistry for the investigation of a karst aquifer: a field study at Milandre test site, Swiss Jura. *Hydrogeology Journal*, 11(6), 673-686.
- Perrone E. (1908). Carta idrografica d'Italia. Il Tevere, 22.
- Petitta M. (2011). Integrazione della base conoscitiva per la gestione della risorsa idrica sotterranea del Parco nazionale dei Monti Sibillini. Contratto di studio e ricerca. Dipartimento Scienze della Terra Università di Roma "La Sapienza", Autorità di Bacino del fiume Tevere, Parco Nazionale Monti Sibillini. Relazione inedita; p. 101.
- Petitta M., Mastroiillo L., Preziosi E., Banzato F., Barberio M. D., Billi A., ... & Di Salvo C. (2018). Water-table and discharge changes associated with the 2016–2017 seismic sequence in central Italy: hydrogeological data and a conceptual model for fractured carbonate aquifers. *Hydrogeology Journal*, 1-18.

- Petitta, M., Banzato, F. & Barberio, M. D. (2021). Studio idrogeologico della sorgente di Capodacqua del Tronto. Relazione Finale.
- Petitta, M., Barberio, D. M., Barbieri, M., Billi, A., Doglioni, C., Passaretti, S., & Franchini, S. (2020). Groundwater monitoring in regional discharge areas selected as “Hydrosensitive” to seismic activity in Central Italy. In *Advances in Natural Hazards and Hydrological Risks: Meeting the Challenge* (pp. 21-25). Springer, Cham.
- Pfahl, S., & Sodemann, H. (2014). What controls deuterium excess in global precipitation?. *Climate of the Past*, 10(2), 771-781.
- Pierantoni P., Deiana G. & Galdenzi S. (2013). Stratigraphic and structural features of the Sibillini Mountains (Umbria-Marche Apennines, Italy). *Ital. J. Geosci.* 132, 497–520. <https://doi.org/10.3301/IJG.2013.08>
- Pierantoni, P. P., Deiana, G., Romano, A., Paltrinieri, W., Borraccini, F., & Mazzoli, S. (2005). Geometrie strutturali lungo la thrust zone del fronte montuoso umbro-marchigiano-sabino. *Bollettino della Società geologica italiana*, 124(2), 395-411.
- Pizzi, A., Calamita, F., Coltorti, M., & Pieruccini, P. (2002). Quaternary normal faults, intramontane basins and seismicity in the Umbria-Marche-Abruzzi Apennine Ridge (Italy): contribution of neotectonic analysis to seismic hazard assessment. *Boll. Soc. Geol. It.*, 1, 923-929.
- Poage, M. A., & Chamberlain, C. P. (2001). Empirical relationships between elevation and the stable isotope composition of precipitation and surface waters: considerations for studies of paleoelevation change. *American Journal of Science*, 301(1), 1-15.
- Porreca M., Minelli G., Ercoli M., Brobia A., Mancinelli P., Cruciani F., Giorgetti C., Carboni F., Mirabella F., Cavinato, G. et al. (2018). Seismic Reflection Profiles and Subsurface Geology of the Area Interested by the 2016-2017 Earthquake Sequence (Central Italy). *Tectonics*, 37, 1116–1137.
- Preziosi E. & Romano E (2009). Dall'analisi idrostrutturale allo studio modellistico di acquiferi regionali. From a hydrostructural analysis to the mathematical modelling of regional aquifers (central Italy). *Italian Journal of Engineering Geology and Environment*, 1, 183-198.
- Preziosi E. (2007). Simulazioni numeriche di acquiferi carbonatici in aree corrugate: applicazioni al sistema idrogeologico della Valnerina (Italia centrale). *Quaderno 125. Istituto di Ricerca sulle Acque - CNR*.
- Preziosi, E., & Piccini, L. (2000). Idrogeologia carsica: le indagini speleologiche con strumento di prospezione e ricerca. *Geologia dell'Ambiente*, 3: 33-38
- Principi P. (1911). *Idrologia sotterranea della Pianura di Norcia*. Boll. Soc. Geol. It., Vol. XXX.
- Pruitt, W. O. (1964). Evapotranspiration—A guide to irrigation. *Calif Turfgrass Cult*, 14, 27-32.
- Pucci, S., De Martini, P. M., Civico, R., Villani, F., Nappi, R., Ricci, T., ... & Pantosti, D. (2017). Coseismic ruptures of the 24 August 2016, Mw 6.0 Amatrice earthquake (central Italy). *Geophysical Research Letters*, 44(5), 2138-2147.
- Pulido-Leboeuf, P., Pulido-Bosch, A., Calvache, M. L., Vallejos, Á., & Andreu, J. M. (2003). Strontium, SO₄²⁻/Cl⁻ and Mg²⁺/Ca²⁺ ratios as tracers for the evolution of seawater into coastal aquifers: the example of Castell de Ferro aquifer (SE Spain). *Comptes Rendus Geoscience*, 335(14), 1039-1048.
- Puliti, I., Pizzi, A., Benedetti, L., Di Domenica, A., & Fleury, J. (2020). Comparing slip distribution of an active fault system at various timescales: Insights for the evolution of the Mt. Vettore-Mt. Bove fault system in Central Apennines. *Tectonics*, 39(9).

- Quinlan, J. F., Davies, G. J., Jones, S. W., & Huntoon, P. W. (1996). The applicability of numerical models to adequately characterize ground-water flow in karstic and other triple-porosity aquifers. In *Subsurface Fluid-Flow (Ground-Water and Vadose Zone) Modeling*. ASTM International.
- Mastrorillo L. & Mazza R. (2021). Rapporto Tecnico Università Roma Tre (Restart) (2021). Revisione dei modelli concettuali dei corpi idrici sotterranei e valutazione dei relativi bilanci idrici a seguito degli effetti indotti dagli eventi sismici del 2016-2017. Redatto da
- Rathay, S. Y., Allen, D. M., & Kirste, D. (2018). Response of a fractured bedrock aquifer to recharge from heavy rainfall events. *Journal of Hydrology*, 561, 1048-1062.
- Renghi D. (2018). Analisi idrostrutturale della sorgente Torbidone a seguito del terremoto del 30 ottobre 2016 (M_w 6.5). Tesi di Laurea Università degli Studi di Perugia.
- Robins, N. S., Rutter, H. K., Dumpleton, S., & Peach, D. W. (2005). The role of 3D visualisation as an analytical tool preparatory to numerical modelling. *Journal of Hydrology*, 301(1-4), 287-295.
- Rojstaczer, S., & Wolf, S. (1992). Permeability changes associated with large earthquakes: An example from Loma Prieta, California. *Geology*, 20(3), 211-214.
- Rojstaczer, S., Wolf, S., & Michel, R. (1995). Permeability enhancement in the shallow crust as a cause of earthquake-induced hydrological changes. *Nature*, 373(6511), 237-239.
- Romano, E., Del Bon, A., Petrangeli, A. B., & Preziosi, E. (2013). Generating synthetic time series of springs discharge in relation to standardized precipitation indices. Case study in Central Italy. *Journal of hydrology*, 507, 86-99.
- Rosati A. (2017). Monitoraggio di sorgenti e captazioni minori dopo il terremoto di Amatrice-Norcia del 2016: variazioni indotte dagli eventi sismici e problematiche di gestione della risorsa. Tesi di Laurea Università La Sapienza di Roma.
- Rosen, M. R., Binda, G., Archer, C., Pozzi, A., Michetti, A. M., & Noble, P. J. (2018). Mechanisms of earthquake-induced chemical and fluid transport to carbonate groundwater springs after earthquakes. *Water Resources Research*, 54(8), 5225-5244.
- Rossi, M., & Donnini, M. (2018). Estimation of regional scale effective infiltration using an open-source hydrogeological balance model and free/open data. *Environmental Modelling & Software*, 104, 153-170.
- Rovida, A. N., Locati, M., Camassi, R. D., Lolli, B., & Gasperini, P. (2019). Catalogo Parametrico dei Terremoti Italiani CPTI15, versione 2.0.
- Rozanski, K., Araguás-Araguás, L., & Gonfiantini, R. (1993). Isotopic patterns in modern global precipitation. *Geophysical Monograph-American Geophysical Union*, 78, 1-1.
- Rushton, K. R. (2003). Groundwater hydrology. *Civil Engineering*, 38, 04-105.
- Rusi, S., Di Curzio, D., Palmucci, W., & Petaccia, R. (2018). Detection of the natural origin hydrocarbon contamination in carbonate aquifers (central Apennine, Italy). *Environmental Science and Pollution Research*, 25(16), 15577-15596.
- Salati, E., Dall'Olio, A., Matsui, E., & Gat, J. R. (1979). Recycling of water in the Amazon basin: an isotopic study. *Water resources research*, 15(5), 1250-1258.
- Sanford, W. E., Cook, P. G., Robinson, N. I., & Weatherill, D. (2006). Tracer mass recovery in fractured aquifers estimated from multiple well tests. *Groundwater*, 44(4), 564-573.
- Sappa, G., Ergul, S., & Ferranti, F. (2014). Water quality assessment of carbonate aquifers in southern Latium region, Central Italy: a case study for irrigation and drinking purposes. *Applied Water Science*, 4(2), 115-128.

- Sappa, G., Vitale, S., & Ferranti, F. (2018). Identifying karst aquifer recharge areas using environmental isotopes: A case study in Central Italy. *Geosciences*, 8(9), 351.
- Saxena, R. K. (1984). Seasonal variations of oxygen-18 in soil moisture and estimation of recharge in Esker and Moraine formations: Paper presented at the Nordic Hydrological Conference (Nyborg, Denmark, August-1984). *Hydrology Research*, 15(4-5), 235-242.
- Scarpelli, G., Bellezza, I., Sante, M. D., Fratalocchi, E., Fruzzetti, V. M. E., Mazzieri, F., ... & Vivalda, P. (2019). Environmental Sustainability of Engineering Works: Geological and Geotechnical Aspects. In *The First Outstanding 50 Years of "Università Politecnica delle Marche"* (pp. 259-278). Springer, Cham.
- Schnegg, P. A. (2002). An inexpensive field fluorometer for hydrogeological tracer tests with three tracers and turbidity measurement. In *Articles of the Geomagnetism Group at the University of Neuchâtel, Groundwater and Human Development* (pp. 1484-1488).
- Scholl, M. A., Gingerich, S. B., & Tribble, G. W. (2002). The influence of microclimates and fog on stable isotope signatures used in interpretation of regional hydrology: East Maui, Hawaii. *Journal of Hydrology*, 264(1-4), 170-184.
- Scholl, M. A., Ingebritsen, S. E., Janik, C. J., & Kauahikaua, J. P. (1996). Use of precipitation and groundwater isotopes to interpret regional hydrology on a tropical volcanic island: Kilauea volcano area, Hawaii. *Water Resources Research*, 32(12), 3525-3537.
- Scibek, J., Gleeson, T., & McKenzie, J. M. (2016). The biases and trends in fault zone hydrogeology conceptual models: global compilation and categorical data analysis. *Geofluids*, 16(4), 782-798.
- Scisciani, V. (2009). Styles of positive inversion tectonics in the Central Apennines and in the Adriatic foreland: Implications for the evolution of the Apennine chain (Italy). *Journal of Structural Geology*, 31(11), 1276-1294.
- Scisciani, V., & Calamita, F. (2009). Active intraplate deformation within Adria: Examples from the Adriatic region. *Tectonophysics*, 476(1-2), 57-72.
- Scisciani, V., Agostini, S., Calamita, F., Cilli, A., Giori, I., Pace, P., & Paltrinieri, W. (2010). The influence of pre-existing extensional structures on the Neogene evolution of the Northern Apennines foreland fold-and-thrust belt. *Rendiconti Online Società Geologica Italiana*, 10, 125-128.
- Scisciani, V., Tavarnelli, E., & Calamita, F. (2002). The interaction of extensional and contractional deformations in the outer zones of the Central Apennines, Italy. *Journal of Structural Geology*, 24(10), 1647-1658.
- Seidenbecher, T., Laxmi, T. R., Stork, O., & Pape, H. C. (2003). Amygdalar and hippocampal theta rhythm synchronization during fear memory retrieval. *Science*, 301(5634), 846-850.
- Shevenell, L. (1996). Analysis of well hydrographs in a karst aquifer: estimates of specific yields and continuum transmissivities. *J Hydrol* 174:331-355
- Shi, Z., Wang, G., Wang, C. Y., Manga, M., & Liu, C. (2014). Comparison of hydrological responses to the Wenchuan and Lushan earthquakes. *Earth and Planetary Science Letters*, 391, 193-200.
- Sibson, R. H. (2013). Stress switching in subduction forearcs: Implications for overpressure containment and strength cycling on megathrusts. *Tectonophysics*, 600, 142-152.
- Silvert, W. (2001). Modelling as a discipline. *International Journal of General System*, 30(3), 261-282.
- Skelton, A., Liljedahl-Claesson, L., Wästeby, N., Andrén, M., Stockmann, G., Sturkell, E., ... & Kockum, I. (2019). Hydrochemical changes before and after earthquakes based on long-term measurements of multiple

- parameters at two sites in northern Iceland—a review. *Journal of Geophysical Research: Solid Earth*, 124(3), 2702-2720.
- Somasundaram, R. S., & Nedunchezian, R. (2011). Evaluation of three simple imputation methods for enhancing preprocessing of data with missing values. *International Journal of Computer Applications*, 21(10), 14-19.
- Sonntag, C., Münnich, K. O., Jacob, H., & Rozanski, K. (1983). Variations of deuterium and oxygen-18 in continental precipitation and groundwater, and their causes. In *Variations in the global water budget* (pp. 107-124). Springer, Dordrecht.
- Spina, R., Stortini, S., Fusari, R., Scuterini, C. & Di Marino, M. (2002). Caratterizzazione climatologica delle Marche: campo medio della temperatura per il periodo 1950-200. Centro di Ecologia e Climatologia - Osservatorio Geofisico Sperimentale, Macerata, IT
- Stemberk, J., Dal Moro, G., Stemberk, J., Blahůt, J., Coubal, M., Košťák, B., ... & Tondi, E. (2019). Strain monitoring of active faults in the central Apennines (Italy) during the period 2002–2017. *Tectonophysics*, 750, 22-35.
- Stewart, M. K. (1975). Stable isotope fractionation due to evaporation and isotopic exchange of falling waterdrops: Applications to atmospheric processes and evaporation of lakes. *Journal of Geophysical Research*, 80(9), 1133-1146.
- Tamburini, A., & Menichetti, M. (2020). Groundwater circulation in fractured and karstic aquifers of the Umbria-Marche Apennine. *Water*, 12(4), 1039.
- Tarragoni C. (2005). Definizione dell'assetto idrogeologico dei Monti Sibillini e sua verifica col contributo della simulazione numerica delle condizioni di flusso e dell'idrologia isotopica. Tesi di Dottorato. Università degli Studi di Roma "La Sapienza".
- Tarragoni C. (2006). Determinazione della "quota isotopica" del bacino di alimentazione delle principali sorgenti dell'alta Valnerina. *Geologica Romana*, 39, 55-62.
- Tavarnelli, E., Butler, R. W. H., Decandia, F. A., Calamita, F., Grasso, M., Alvarez, W., ... & D'offizi, S. (2004). Implications of fault reactivation and structural inheritance in the Cenozoic tectonic evolution of Italy. *The Geology of Italy, Special*, 1, 209-222.
- Taylor, C. J., & Greene, E. A. (2008). Hydrogeologic characterization and methods used in the investigation of karst hydrology. Field techniques for estimating water fluxes between surface water and ground water, edited by: Rosenberry, DO and LaBaugh, JW, US Geological Survey, Reston, Virginia (EUA), 71-114.
- Taylor, S., Feng, X., Kirchner, J. W., Osterhuber, R., Klaue, B., & Renshaw, C. E. (2001). Isotopic evolution of a seasonal snowpack and its melt. *Water Resources Research*, 37(3), 759-769.
- Tazioli, A. (2011). Experimental methods for river discharge measurements: comparison among tracers and current meter. *Hydrological Sciences Journal*, 56(7), 1314-1324.
- Tazioli, A., Cervi, F., Doveri, M., Mussi, M., Deiana, M., & Ronchetti, F. (2019). Estimating the isotopic altitude gradient for hydrogeological studies in mountainous areas: are the low-yield springs suitable? insights from the Northern Apennines of Italy. *Water*, 11(9), 1764.
- Tazioli, G. S., & Tacconi, P. (1991). Application of artificial tracer in pollution studies. Use of artificial tracers in hydrology, 7.
- Terzić, J., Frangen, T., Borović, S., Reberski, J. L., & Patekar, M. (2022). Hydrogeological Assessment and Modified Conceptual Model of a Dinaric Karst Island Aquifer. *Water*, 14(3), 404.

- Terzić, J., Marković, T., & Reberski, J. L. (2014). Hydrogeological properties of a complex Dinaric karst catchment: Miljacka Spring case study. *Environmental earth sciences*, 72(4), 1129-1142.
- Thornthwaite, C. W., & Mather, J. R. (1955). *The Water Balance*; Johns Hopkins Univ., Laboratory in Climatology: Baltimore, MD, USA; Volume 8, pp. 1-104.
- Thornthwaite, C. W., & Mather, J. R. (1957). Instructions and tables for computing potential evapotranspiration and the water balance (No. RESEARCH). Centerton.
- Thornton, P. K., Ericksen, P. J., Herrero, M., & Challinor, A. J. (2014). Climate variability and vulnerability to climate change: a review. *Global change biology*, 20(11), 3313-3328.
- Trauth, M. H. *Detecting Change Points in Time Series with MATLAB*. (2017)
- Turc, L. (1954) Calcul du bilan de l'eau: evaluation en fonction des precipitations et des temperatures. *Int. Assoc. Sci. Hydrol.* 38(3), 188-202
- Turc, L. (1961). Evaluation des besoins en eau d'irrigation, évapotranspiration potentielle. *Ann. Agron.*, 12, 13-49.
- Unnikrishna, P. V., McDonnell, J. J., & Kendall, C. (2002). Isotope variations in a Sierra Nevada snowpack and their relation to meltwater. *Journal of Hydrology*, 260(1-4), 38-57.
- Valigi D., Banzato F., Beddini B., Cambi C., Cardellini C., Checcucci R., De Giorgi F., Mirabella F., Toccaceli B. & Stelluti M. (2019b). Gestione delle variazioni dell'assetto idrogeologico indotte dalla sequenza sismica del 2016 nell'area nursina e dei Monti Sibillini. Atti del IV Convegno Interdipartimentale "Gestire l'emergenza. Il contributo della ricerca interdisciplinare" 14-15 febbraio 2019. Università degli Studi di Perugia. ISBN/EAN: 978-88-9392-138-I.
- Valigi D., Cardellini C., Mirabella F., Tazioli A., Petitta M., Caliro S. (Responsabili Scientifici), Cambi C., Banzato F., Beddini G., Fronzi D., Lacchini A., Mastrorillo L., Palpacelli S., Sbarbati C., Viaroli S. (Collaboratori) (2020a). Modello concettuale. Contributo alla ricerca: "Caratterizzazione dei sistemi idrogeologici del territorio umbro influenzato dagli eventi sismici del 26-30 ottobre 2016 e valutazione degli effetti del sisma sull'approvvigionamento idrico". Referente Regione Umbria: Dr. Geol. Roberto Checcucci.
- Valigi D., Fronzi D., Cambi C., Beddini, G., Cardellini C., Checcucci, R., Mastrorillo L., Mirabella F. & Tazioli A. (2020b). Earthquake-induced spring discharge modifications: The Pescara di Arquata spring reaction to the August-October 2016 central Italy Earthquakes. *Water (Switzerland)*, 12 (3), 1-18.
- Valigi D., Mastrorillo L., Cardellini C., Checcucci R., Di Matteo L., Frondini F., Mirabella F., Viaroli S. & Vispi I. (2019a). Springs discharge variations induced by strong earthquakes: the Mw 6.5 Norcia event (Italy, October 30th 2016). *ROL* 2019, 47, 141-146.
- Vaute, L., Drogue, C., Garrelly, L., & Ghelfenstein, M. (1997). Relations between the structure of storage and the transport of chemical compounds in karstic aquifers. *Journal of Hydrology*, 199(3-4), 221-238.
- Vaux, H. (2011). Groundwater under stress: the importance of management. *Environmental Earth sciences*, 62(1), 19-23.
- Viaroli, S., Di Curzio, D., Lepore, D., & Mazza, R. (2019). Multiparameter daily time-series analysis to groundwater recharge assessment in a caldera aquifer: Roccamonfina Volcano, Italy. *Science of the total environment*, 676, 501-513.
- Viaroli, S., Mastrorillo, L., Lotti, F., Paolucci, V., & Mazza, R. (2018). The groundwater budget: a tool for preliminary estimation of the hydraulic connection between neighboring aquifers. *Journal of hydrology*, 556, 72-86.

- Viaroli, S., Mirabella, F., Mastorillo, L., Angelini, S., & Valigi, D. (2021). Fractured carbonate aquifers of Sibillini Mts.(Central Italy). *Journal of Maps*, 17(2), 140-149.
- Villani, F., Maraio, S., Bruno, P. P., Improta, L., Wood, K., Pucci, S., ... & Pantosti, D. (2021). High-Resolution Seismic Profiling in the Hanging Wall of the Southern Fault Section Ruptured During the 2016 Mw 6.5 Central Italy Earthquake. *Tectonics*, 40(9), e2021TC006786.
- Vogel, J. C., & Ehhalt, D. (1963). The use of the carbon isotopes in groundwater studies. In *Radioisotopes in Hydrology. Proceedings of a Symposium*.
- Vogel, J. C., & Van Urk, H. (1975). Isotopic composition of groundwater in semi-arid regions of southern Africa. *Journal of Hydrology*, 25(1-2), 23-36.
- Wang, C. Y., & Chia, Y. (2008). Mechanism of water level changes during earthquakes: Near field versus intermediate field. *Geophysical Research Letters*, 35(12).
- Wang, C. Y., & Manga, M. (2010). Hydrologic responses to earthquakes and a general metric. *Geofluids*, 10(1-2), 206-216.
- Wang, C. Y., & Manga, M. (2015). New streams and springs after the 2014 Mw6. 0 South Napa earthquake. *Nature Communications*, 6(1), 1-6.
- Wendland, E., Barreto, C. E. A. G., & Gomes, L. H. (2007). Water balance in the Guarani Aquifer outcrop zone based on hydrogeologic monitoring. *Journal of Hydrology*, 342(3-4), 261-269.
- Westaway, R., & Jackson, J. (1987). The earthquake of 1980 November 23 in Campania—Basilicata (southern Italy). *Geophysical Journal International*, 90(2), 375-443.
- Weyhenmeyer, C. E., Burns, S. J., Waber, H. N., Macumber, P. G., & Matter, A. (2002). Isotope study of moisture sources, recharge areas, and groundwater flow paths within the eastern Batinah coastal plain, Sultanate of Oman. *Water Resources Research*, 38(10), 2-1.
- Williams, P. W. (2008). The role of the epikarst in karst and cave hydrogeology: a review. *International Journal of Speleology*, 37(1), 1.
- Winograd, I. J., Riggs, A. C., & Coplen, T. B. (1998). The relative contributions of summer and cool-season precipitation to groundwater recharge, Spring Mountains, Nevada, USA. *Hydrogeology Journal*, 6(1), 77-93.
- World Meteorological Organization, (2012). *Standardized Precipitation Index User Guide* (M. Sodova, M. Hayes and D. Wood). WMO – No 1090, Geneva, 18 pp.
- Worthington, P. F. (2017). A scaled perspective on permeability: How to get the best from what we have. *Journal of Petroleum Science and Engineering*, 156, 212-219.
- Worthington, S. R. (2009). Diagnostic hydrogeologic characteristics of a karst aquifer (Kentucky, USA). *Hydrogeology Journal*, 17(7), 1665-1678.
- Worthington, S. R., & Smart, C. C. (2003). Empirical determination of tracer mass for sink to spring tests in karst. In *Sinkholes and the engineering and environmental impacts of karst* (pp. 287-298).
- Worthington, S. R., & Smart, C. C. (2016). Determination of tracer mass for effective groundwater tracer tests. *Carbonates and Evaporites*, 31(4), 349-356.
- Xue, L., Brodsky, E. E., Erskine, J., Fulton, P. M., & Carter, R. (2016). A permeability and compliance contrast measured hydrogeologically on the San Andreas Fault. *Geochemistry, Geophysics, Geosystems*, 17(3), 858-871.

- Yan, R., Woith, H., & Wang, R. (2014). Groundwater level changes induced by the 2011 Tohoku earthquake in China mainland. *Geophysical Journal International*, 199(1), 533-548.
- Younger, P. L., Teutsch, G., Custodio, E., Elliot, T., Manzano, M., & Sauter, M. (2002). Assessments of the sensitivity to climate change of flow and natural water quality in four major carbonate aquifers of Europe. Geological Society, London, Special Publications, 193(1), 303-323.
- Younger, P.L. (2007) *Groundwater in the Environment: An Introduction*. Blackwell Publishing, Oxford.
- Zdechlik, R. (2016). A review of applications for numerical groundwater flow modeling. *International Multidisciplinary Scientific GeoConference: SGEM*, 3, 11-18.
- Zhang, M., & Wang, S. (2016). A review of precipitation isotope studies in China: Basic pattern and hydrological process. *Journal of Geographical Sciences*, 26(7), 921-938.
- Zhou, S., Nakawo, M., Hashimoto, S., & Sakai, A. (2008). The effect of refreezing on the isotopic composition of melting snowpack. *Hydrological Processes: An International Journal*, 22(6), 873-882.
- Zuber, A., & Motyka, J. (1998). Hydraulic parameters and solute velocities in triple-porosity karstic-fissured-porous carbonate aquifers: case studies in southern Poland. *Environmental Geology*, 34(2), 243-250.

Annex A

```
import os

import numpy as np

import pandas as pd

from pandas import ExcelWriter

from numpy import *

def thornMater(data,LAT,SM,SRT,beta):

    filename="K_latN_S.csv"

    k = np.genfromtxt(filename, delimiter = ';', filling_values = 0)

    k = k[:, k[0] == LAT]

    Tm = data[:,2]

    im = (Tm/5)** 1.514

    k1 = np.unique(data[:,0])

    for j in range(len(k1)):

        index = [data[:,0] == k1[j]]

        I = im[index]

        I = np.nansum(I)

        break

    II = [I] * len(data)

    Iyear1 = [k1, II];

    n = len(data)

    npp = np.resize(k[1:], n)

    II = np.array(II)

    npp = np.array(npp)

    Tm = np.array(Tm)

    a = (675 * 10**-9 * (II**3)) - (771 * 10**-7 * (II**2)) + (1792 * 10**-5 * II) + 0.49239

    a = np.round(a, 2)

    PET = 16 * npp * (((10 * Tm)/ II)** a);

    PET = np.array(PET)

    PET = np.round(PET, 1)

    where_are_NaNs = isnan(PET)

    PET[where_are_NaNs] = 0

    threshold = 0

    substitute = 0
```

```

PET[PET <= threshold] = substitute
P = data[:,3];
delta = P - PET;
delta = np.array(delta)
P1 = [None] * len(P);
for ih in range(0, len(P)):
    P1[ih] = P[ih];
P1 = np.array(P1)
P1[Tm > SRT]= 0
SP = np.convolve(P1,np.ones(10,dtype=float),'full')
SP = SP[0:len(P1)]
SP1 = np.zeros((len(SP),))
for d in range(len(SP)):
    if d == 0:
        SP1[d] = SP[d]
    else:
        if SP[d - 1] < SP[d]:
            SP1[d - 1] = 0
            SP1[d] = SP[d]
        else:
            SP1[d] = SP[d]
for di in range(len(SP1)):
    if SP1[di] != 0:
        if di ==0:
            SP1[di] = SP1[di]
        else:
            if SP1[di] < SP1[di-1]:
                SP1[di] = 0
            else:
                SP1[di] = SP1[di]
    else:
        SP1[di] = SP1[di]

SMRO = np.zeros((len(SP1) + 1,))
ii = 0

```

```

for h in range(len(SP1)):
    if SP1[h] > 0:
        if ii == 0:
            SMRO[h] = 0
            ii = ii+1
            if ii==1:
                SMRO[h + 1] = SP1[h]*0.10
            elif ii==2:
                SMRO[h + 1] = (SP1[h]-SMRO[h - 1])/2
            else:
                SMRO[h + 1]= SMRO[h]/2
        else:
            ii = 0;
            SMRO[h + 1]= SP1[h]
SMRO = SMRO[0:len(P1)]
SMRO = np.round(SMRO, 1)
S = np.zeros((len(SP),))
S[0] = P[0] - PET[0]
AET = np.zeros((len(SP),))
ST = np.zeros((len(SP),))
ST.astype(float)
AET.astype(float)
PET.astype(float)
P.astype(float)
for h in range(len(Tm)):
    if Tm [h] >= SRT:
        if delta[h] < 0:
            S[h]=0
            if h == 0:
                ST[h] = SM - delta[h]
                AET[h]= PET[h]
            else:
                ST[h]= ST[h - 1] - (ST[h - 1] * (1 - np.exp(-(PET[h]- P[h])/SM)))
                AET[h]= P[h] + (ST[h - 1] * (1- np.exp(-(PET[h]- P[h])/SM)))
            else:

```

```

    if h == 0:
        ST[h]= SM
        AET [h] = PET[h]
    else:
        if delta[h] < (SM - ST[h - 1]):
            ST[h]= ST[h - 1] + delta[h]
            S[h]=0
            AET [h] = PET[h]
        else:
            ST[h]= float(SM)
            AET [h] = PET [h]
            if ST[h - 1] > SM:
                S[h] = delta[h]
            else:
                S[h]= delta [h]-(SM-ST[h-1])
else:
    if h == 0:
        ST[h] = SM
        AET[h] = 0
        S[h] = 0
    else:
        ST[h] = ST[h-1] + delta[h]
        AET[h] = 0
        S[h] = 0
RO = np.zeros((len(SP),))
RES = np.zeros((len(SP),))
for h in range(len(S)):
    if h == 0:
        RO[h] = beta * (0 + S[h])
        RES[h] = (1 - beta)*(0 + S[h])
    elif h==1:
        RO[h] = (S[h] + RES[h-1]) * beta
        RES[h] = (1 - beta) * (RES[h-1] + S[h])
    else:
        RO[h] = (RES[h-1] + S[h]) * beta

```

```

    RES[h] = (1 - beta) * (RES[h-1] + S[h])
RO = np.round(RO, 1)
tot_RO = np.zeros((len(SP),))
for f in range(len(S)):
    tot_RO[f] = SMRO[f]+RO[f]
out1 = [data[:,0], data[:,1], data[:,2], data[:,3], a, PET, delta, AET, ST, S, RO, RES, SMRO, tot_RO]
book = xlwt.Workbook()
sheet1 = book.add_sheet('sheet1')
col = ['year','month','Tm','P','k','T','a','PET','delta','AET','ST','S','RO','RES','SMRO','TOT_RO']
for row, array in enumerate(out1):
    for col, value in enumerate(array):
        sheet1.write(col, row, value)
name = "output_waterbalANce.xls"
book.save(name)
book.save(TemporaryFile())
#plot data in subplots
precip = data[:,3]
datetime = np.arange(0, len(Tm), 1)
fig, axs = plt.subplots(5, 2)
axs[0,0].plot(np.float64(datetime),np.float64(Tm),'--',color='black')
axs[0,0].set_ylim([min(Tm)-10,max(Tm)+10])
axs[0,0].set_ylabel("Temp. [ $^{\circ}$ C]")
axs[0,1].plot(np.float64(datetime),np.float64(precip),'--',color='black')
axs[0,1].set_ylim([min(precip),max(precip)+20])
axs[0,1].set_ylabel("Rainfall [mm]")
axs[1,0].plot(np.float64(datetime),np.float64(PET),color='magenta')
axs[1,0].set_ylim([min(PET),max(PET)+10])
axs[1,0].set_ylabel("PET [mm]")
axs[1,1].plot(np.float64(datetime),np.float64(delta),color='cyan')
axs[1,1].set_ylim([min(delta),max(delta)+10])
axs[1,1].set_ylabel("P-PET [mm]")
axs[2,0].plot(np.float64(datetime),np.float64(AET),color='orange')
axs[2,0].set_ylim([min(AET),max(AET)+10])
axs[2,0].set_ylabel("AET [mm]")
axs[2,1].plot(np.float64(datetime),np.float64(ST),color='red')

```

```

axs[2,1].set_ylim([min(ST),max(ST)+10])
axs[2,1].set_ylabel('ST [mm]')
axs[3,0].plot(np.float64(datetime),np.float64(S),color='yellow')
axs[3,0].set_ylim([min(S),max(S)])
axs[3,0].set_ylabel('S [mm]')
axs[3,1].plot(np.float64(datetime),np.float64(RO),color='green')
axs[3,1].set_ylim([min(RO),max(RO)])
axs[3,1].set_ylabel('RO [mm]')
axs[4,0].plot(np.float64(datetime),np.float64(SMRO),color='blue')
axs[4,0].set_ylim([min(SMRO),max(SMRO)])
axs[4,0].set_ylabel('SMRO [mm]')
axs[4,1].plot(np.float64(datetime),np.float64(tot_RO),color='black')
axs[4,1].set_ylim([min(tot_RO),max(tot_RO)])
axs[4,1].set_ylabel('tot RO [mm]')
for ax in axs.flat:
    ax.set(xlabel="Time [months]")
    ax.set_xticks(np.arange(min(datetime), max(datetime)+1, 12))
plt.rcParams.update({'font.size': 15})
figure = plt.gcf()
figure.set_size_inches(32, 18)
plt.savefig('output_waterbalANce.png', bbox_inches='tight',dpi=300)
plt.show()

```

Annex B

								Turc Method						Thornthwaite-Mather Method			
Hydrostructure	Hydrogeological complex	PICs	Area (km²)	Mean Elevation (m a.s.l.)	Rainfall		Temperature °C	AET		WS		EI		S		EI	
					mm/year	m³/s		mm/year	m³/s	mm/year	m³/s	mm/year	m³/s	mm/year	m³/s	mm/year	m³/s
1	BAS	0.93	38.0	1655	1245	1.50	7.27	454	0.55	791	0.95	735	0.89	724	0.87	674	0.81
	JUR	0.61	12.0	1474	1222	0.46	8.05	469	0.18	753	0.29	459	0.17	680	0.26	415	0.16
	MAI	0.82	23.0	1624	1241	0.90	7.41	457	0.33	784	0.57	643	0.47	717	0.52	588	0.43
	MF	0.45	4.0	1655	1245	0.16	7.27	454	0.06	791	0.10	356	0.05	724	0.09	326	0.04
	SCA	0.80	17.0	1740	1255	0.68	6.91	447	0.24	809	0.44	647	0.35	745	0.40	596	0.32
	TOTAL		94										1.92				1.76
2	BAS	0.95	37.4	1346	1070	1.27	8.60	471	0.56	599	0.71	569	0.67	565	0.67	536	0.64
	JUR	0.67	38.3	1240	1033	1.25	9.06	477	0.58	556	0.68	372	0.45	521	0.63	349	0.42
	MAI	0.84	69.0	1325	1063	2.32	8.69	472	1.03	590	1.29	496	1.09	556	1.22	467	1.02
	MF	0.47	17.4	1272	1044	0.58	8.92	475	0.26	569	0.31	267	0.15	534	0.29	251	0.14
	SCA	0.85	86.4	1342	1069	2.93	8.62	471	1.29	597	1.64	508	1.39	563	1.54	479	1.31
	SVC	0.45	17.6	984	942	0.53	10.16	487	0.27	455	0.25	205	0.11	414	0.23	186	0.10
	DEP	1.00	14.5	1291	1051	0.48	8.84	474	0.22	577	0.27	577	0.27	542	0.25	542	0.25
	TOTAL		281										4.13				3.88
2a	BAS	0.96	20.6	1422	1215	0.79	8.28	474	0.31	742	0.48	712	0.47	667	0.44	640	0.42
	JUR	0.62	4.0	1490	1224	0.16	7.98	468	0.06	756	0.10	469	0.06	684	0.09	424	0.05
	MAI	0.90	10.0	1462	1220	0.39	8.10	470	0.15	750	0.24	675	0.21	677	0.21	609	0.19
	MF	0.50	1.3	1370	1209	0.05	8.50	478	0.02	731	0.03	365	0.02	654	0.03	327	0.01
	SCA	0.81	5.0	1321	1202	0.19	8.71	482	0.08	721	0.11	584	0.09	642	0.10	520	0.08
	TOTAL		41										0.85				0.76
3 North	BAS	1.00	22	1122	991	0.68	9.56	504	0.35	487	0.34	487	0.34	522	0.36	522	0.36
	JUR	0.70	14	925	922	0.42	10.41	517	0.24	404	0.18	283	0.13	431	0.20	302	0.14
	MAI	0.88	53	888	908	1.52	10.57	519	0.87	389	0.65	342	0.57	414	0.69	364	0.61
	MF	0.50	22	923	921	0.65	10.42	518	0.37	402	0.29	201	0.14	430	0.31	215	0.15
	SCA	0.87	190	893	910	5.49	10.55	519	3.13	391	2.36	340	2.05	417	2.51	363	2.18
	SVC	0.50	66	762	864	1.79	11.11	525	1.09	339	0.70	170	0.35	356	0.74	178	0.37
	DEP	0.90	33	695	840	0.88	11.40	528	0.55	312	0.33	281	0.29	325	0.34	293	0.31
	TOTAL		400	887									3.87				4.12
3 South	BAS	1.00	90	1150	1229	3.50	9.45	527	1.50	702	2.00	702	1.80	597	1.70	597	1.70
	JUR	0.70	19	1254	1300	0.78	9.00	516	0.31	784	0.47	548	0.33	691	0.42	484	0.29
	MAI	0.88	72	1171	1243	2.85	9.36	522	1.20	721	1.65	634	1.46	629	1.44	554	1.27
	MF	0.50	16	1063	1168	0.60	9.82	530	0.27	638	0.33	319	0.16	546	0.28	273	0.14
	SCA	0.87	95	1119	1206	3.65	9.58	526	1.59	680	2.06	592	1.79	588	1.78	512	1.55
	SVC	0.50	8	872	1037	0.25	10.64	543	0.13	494	0.12	247	0.06	421	0.10	210	0.05
	DEP	0.90	53	918	1069	1.80	10.44	541	0.91	528	0.89	475	0.80	446	0.75	402	0.67
	TOTAL		353										6.40				5.67

Annex C

Supplementary material containing physio-chemical measurements, major ions concentrations, δ ¹⁸ O and δ ² H for the monitoring points																	
No. On Map (Figure 3.48)	Sampling time	Type (Spring=S Linear Spring=LS River=R)	X (UTM_WGS84-32)	Y (UTM_WGS84-32)	T °C	PH	Conduc. (µS/cm)	Ca++ mg/L	Mg++ mg/L	Na+ mg/L	K+ mg/L	HCO3- mg/L	Cl- mg/L	SO4-- mg/L	δ ¹⁸ O ‰ vs V-SMOW	δ ² H ‰ vs V-SMOW	Ref.
1	10/04/2003	S	355904	4752585	n.d.	n.d.	n.d.	71.10	0.66	2.07	0.22	201.30	2.78	3.75	n.d.	n.d.	a
1	10/04/2003	S	355904	4752585	n.d.	n.d.	n.d.	63.60	0.65	2.58	0.31	181.60	3.57	3.51	n.d.	n.d.	a
1	11/06/2003	S	355904	4752585	n.d.	n.d.	n.d.	61.20	0.64	2.60	0.28	n.d.	3.48	3.42	n.d.	n.d.	a
1	11/06/2003	S	355904	4752585	n.d.	n.d.	n.d.	63.20	0.64	2.55	0.31	n.d.	3.49	3.45	n.d.	n.d.	a
1	21/11/2003	S	355904	4752585	n.d.	n.d.	n.d.	72.30	0.65	2.14	0.15	n.d.	n.d.	n.d.	n.d.	n.d.	a
1	21/11/2003	S	355904	4752585	n.d.	n.d.	n.d.	64.20	0.68	2.67	0.22	n.d.	3.90	3.78	n.d.	n.d.	a
1	04/03/2004	S	355904	4752585	n.d.	n.d.	n.d.	75.90	0.64	2.31	0.16	196.90	2.23	3.66	n.d.	n.d.	a
1	04/03/2004	S	355904	4752585	n.d.	n.d.	n.d.	67.60	0.70	2.61	0.23	188.60	2.62	3.49	n.d.	n.d.	a
1	15/04/2014	S	355904	4752585	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.21	-66.96	a
1	26/09/2014	S	355904	4752585	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.96	-65.74	a
1	09/05/2018	S	355904	4752585	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.73	-64.64	a
2	21/03/2003	S	358873	4749161	6.70	n.d.	263.00	n.d.	n.d.	1.09	0.29	138.60	1.74	23.05	n.d.	n.d.	a
2	10/04/2003	S	358873	4749161	6.80	8.12	264.00	35.63	11.72	1.13	0.35	138.80	1.88	23.16	-10.05	-61.39	a
2	09/05/2003	S	358873	4749161	6.80	7.24	259.00	34.43	11.15	1.20	0.36	136.20	1.89	21.28	-10.04	-64.22	a
2	11/06/2003	S	358873	4749161	6.80	8.22	262.00	34.28	11.65	1.14	0.38	137.40	1.79	22.72	-9.69	-63.79	a
2	15/07/2003	S	358873	4749161	6.80	8.14	262.00	36.82	12.04	1.17	0.38	138.70	1.85	24.75	-9.73	-63.51	a
2	20/08/2003	S	358873	4749161	6.70	8.13	n.d.	36.24	12.04	1.13	0.36	137.10	1.79	25.21	-9.73	-64.36	a
2	22/09/2003	S	358873	4749161	6.80	8.21	264.00	36.59	12.11	1.13	0.35	134.70	1.72	25.79	-9.70	-62.93	a
2	21/10/2003	S	358873	4749161	6.70	7.99	264.00	36.09	11.99	1.13	0.36	136.90	1.68	25.08	-9.79	-64.84	a
2	21/11/2003	S	358873	4749161	6.80	8.01	264.00	35.53	12.01	1.13	0.30	137.90	1.86	25.36	-9.89	-63.87	a
2	22/12/2003	S	358873	4749161	n.d.	7.93	264.00	35.40	12.06	1.16	0.31	138.10	1.89	26.22	-9.92	-64.51	a
2	22/01/2004	S	358873	4749161	6.80	8.15	266.00	33.10	11.79	1.17	0.31	137.40	1.91	25.19	-9.98	-63.85	a
2	23/02/2004	S	358873	4749161	6.70	7.96	264.00	35.58	11.96	1.14	0.28	136.90	1.21	24.40	-10.09	-66.17	a
2	25/03/2004	S	358873	4749161	6.80	7.89	262.00	32.30	11.64	1.07	0.29	n.d.	n.d.	n.d.	n.d.	n.d.	a
2	22/04/2004	S	358873	4749161	6.70	7.91	266.00	32.80	11.05	1.08	0.29	n.d.	1.68	21.50	n.d.	n.d.	a
2	26/05/2004	S	358873	4749161	n.d.	n.d.	n.d.	32.40	10.90	n.d.	n.d.	n.d.	1.58	21.47	n.d.	n.d.	a
2	18/06/2004	S	358873	4749161	n.d.	n.d.	n.d.	32.80	11.31	n.d.	n.d.	n.d.	1.64	22.74	n.d.	n.d.	a
2	20/07/2004	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.97	-66.43	a
2	27/09/2004	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-8.42	-55.79	a
2	22/10/2004	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-8.79	-58.34	a
2	15/11/2004	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.87	-66.89	a
2	15/12/2004	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.87	-65.58	a
2	24/01/2005	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.43	-63.00	a
2	25/02/2005	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.23	-68.39	a
2	25/03/2005	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.41	-70.51	a
2	24/06/2005	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.08	-67.04	a
2	23/09/2005	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.44	-70.82	a
2	28/10/2005	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.30	-69.62	a
2	25/11/2005	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.53	-70.96	a
2	23/12/2005	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.40	-70.65	a
2	31/01/2006	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.29	-70.88	a
2	08/03/2006	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.45	-69.53	a
2	07/04/2006	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.42	-69.42	a
2	26/07/2006	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.10	-68.70	a
2	29/09/2006	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.70	-68.00	a
2	14/11/2006	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.90	-68.50	a
2	26/01/2007	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.90	-68.00	a
2	13/03/2007	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.20	-68.90	a
2	24/04/2007	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.20	-69.00	a
2	24/06/2007	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.30	-69.30	a
2	29/08/2007	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.30	-69.20	a
2	19/10/2007	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.30	-68.70	a
2	25/02/2008	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.30	-72.10	a
2	17/03/2008	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.20	-71.70	a
2	19/05/2008	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.20	-72.00	a
2	10/06/2008	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.05	-70.85	a
2	02/07/2008	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.08	-71.52	a
2	21/07/2008	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.14	-70.74	a
2	04/08/2008	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.16	-70.99	a
2	29/09/2008	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.20	-71.15	a
2	27/10/2008	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.24	-71.23	a
2	16/03/2009	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.90	-67.50	a
2	16/03/2009	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.86	-67.45	a
2	25/05/2009	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.40	-72.20	a
2	25/05/2009	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.37	-72.21	a

2	21/07/2009	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.00	-69.00	a
2	21/07/2009	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.01	-69.01	a
2	27/08/2009	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.70	-75.20	a
2	27/08/2009	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.69	-75.20	a
2	29/09/2009	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.20	-71.30	a
2	29/09/2009	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.19	-71.30	a
2	26/11/2012	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-8.27	-53.28	a
2	05/03/2013	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-7.38	-49.50	a
2	23/05/2013	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-6.85	-46.00	a
2	21/06/2013	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-7.80	-48.86	a
2	11/09/2013	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-6.07	-37.63	a
2	22/11/2013	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.41	-69.71	a
2	09/04/2014	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.72	-69.45	a
2	15/04/2014	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.08	-67.45	a
2	17/07/2014	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-6.45	-45.76	a
2	26/09/2014	S	358873	4749161	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.00	-67.60	a
2	28/01/2016	S	358873	4749161	5.60	n.d.	245.00	39.93	6.82	1.33	0.46	151.80	1.88	7.98	n.d.	n.d.	a
2	28/01/2016	S	358873	4749161	7.30	n.d.	244.00	43.66	3.74	1.53	0.44	151.30	2.35	5.49	n.d.	n.d.	a
2	09/02/2016	S	358873	4749161	n.d.	n.d.	n.d.	45.85	3.97	1.47	0.51	150.80	1.94	6.04	n.d.	n.d.	a
2	09/02/2016	S	358873	4749161	n.d.	n.d.	n.d.	43.85	3.75	1.51	0.44	143.00	2.32	5.56	n.d.	n.d.	a
2	18/02/2016	S	358873	4749161	n.d.	n.d.	n.d.	36.36	8.74	1.17	0.35	150.50	1.87	8.03	n.d.	n.d.	a
2	18/02/2016	S	358873	4749161	n.d.	n.d.	n.d.	43.73	3.91	1.44	0.36	149.30	2.35	5.12	n.d.	n.d.	a
2	10/03/2016	S	358873	4749161	n.d.	n.d.	n.d.	38.93	7.50	1.25	0.42	150.10	1.98	8.38	n.d.	n.d.	a
2	10/03/2016	S	358873	4749161	n.d.	n.d.	n.d.	44.12	3.62	1.49	0.40	147.60	2.48	5.41	n.d.	n.d.	a
2	18/03/2016	S	358873	4749161	6.20	7.68	241.00	37.90	8.48	1.24	0.41	149.60	2.13	8.29	n.d.	n.d.	a
2	18/03/2016	S	358873	4749161	7.60	7.52	243.00	44.97	3.30	1.59	0.42	149.30	2.85	4.87	n.d.	n.d.	a
2	29/03/2016	S	358873	4749161	n.d.	7.31	n.d.	38.54	8.27	1.22	0.41	149.80	2.12	8.42	n.d.	n.d.	a
2	29/03/2016	S	358873	4749161	n.d.	7.79	n.d.	45.08	3.25	1.59	0.42	149.10	2.89	4.52	n.d.	n.d.	a

4	30/06/2004	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.00	-67.50	a
4	30/07/2004	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-8.51	-55.96	a
4	05/10/2004	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-8.66	-54.00	a
4	23/11/2004	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-8.67	-55.07	a
4	18/02/2005	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.56	-63.05	a
4	12/04/2005	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.17	-66.78	a
4	29/06/2005	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.79	-66.84	a
4	20/07/2005	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.91	-65.94	a
4	23/09/2005	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.06	-67.11	a
4	28/10/2005	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.11	-67.79	a
4	25/11/2005	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.23	-67.57	a
4	31/01/2006	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.11	-67.47	a
4	07/04/2006	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.28	-66.17	a
4	26/06/2006	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.00	-69.30	a
4	26/07/2006	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.00	-69.30	a
4	29/09/2006	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.70	-67.90	a
4	06/12/2006	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.90	-68.60	a
4	26/01/2007	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.90	-67.70	a
4	13/03/2007	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.00	-68.00	a
4	24/04/2007	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.00	-68.70	a
4	24/05/2007	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.00	-68.20	a
4	05/07/2007	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.90	-67.90	a
4	26/07/2007	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.00	-69.30	a
4	29/08/2007	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.10	-68.70	a
4	19/10/2007	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.00	-67.70	a
4	18/02/2008	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.20	-69.60	a
4	17/03/2008	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.90	-69.20	a
4	08/05/2008	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.90	-68.20	a
4	19/05/2008	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.10	-70.70	a
4	10/06/2008	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.41	-65.23	a
4	02/07/2008	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.99	-70.54	a
4	21/07/2008	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.17	-71.17	a
4	06/08/2008	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.52	-65.74	a
4	30/09/2008	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.04	-70.42	a
4	29/10/2008	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.04	-70.22	a
4	17/11/2008	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.94	-69.10	a
4	15/12/2008	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.05	-70.42	a
4	19/01/2009	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.71	-67.80	a
4	24/02/2009	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.02	-69.91	a
4	16/03/2009	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.00	-67.70	a
4	21/07/2009	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.40	-69.70	a
4	27/08/2009	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.80	-67.80	a
4	29/09/2009	S	358200	4735049	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.50	-69.50	a
4	28/01/2016	S	358200	4735049	8.60	n.d.	254.00	50.57	1.09	1.75	0.42	159.80	3.56	1.54	n.d.	n.d.	a
4	09/02/2016	S	358200	4735049	n.d.	n.d.	n.d.	50.27	0.77	1.74	0.40	158.40	3.67	1.37	n.d.	n.d.	a
4	09/02/2016	S	358200	4735049	n.d.	n.d.	n.d.	50.66	1.20	1.72	0.41	147.00	3.50	1.66	n.d.	n.d.	a
4	18/02/2016	S	358200	4735049	n.d.	n.d.	n.d.	50.91	1.03	1.77	0.34	157.60	3.93	1.79	n.d.	n.d.	a
4	18/02/2016	S	358200	4735049	n.d.	n.d.	n.d.	48.18	1.16	1.67	0.35	148.40	3.54	1.82	n.d.	n.d.	a
4	08/03/2016	S	358200	4735049	n.d.	n.d.	n.d.	49.51	0.73	1.78	0.33	153.50	4.02	1.71	n.d.	n.d.	a
4	18/03/2016	S	358200	4735049	8.70	7.59	248.00	47.71	0.65	1.94	0.29	153.50	5.43	2.22	n.d.	n.d.	a
4	29/03/2016	S	358200	4735049	n.d.	7.84	n.d.	50.49	0.76	1.94	0.40	154.20	4.89	1.98	n.d.	n.d.	a
4	12/04/2016	S	358200	4735049	n.d.	7.33	n.d.	51.38	1.00	1.71	0.40	159.10	3.45	1.32	n.d.	n.d.	a
4	30/08/2016	S	358200	4735049	8.10	7.84	247.00	51.10	1.12	1.97	0.82	166.07	3.60	1.59	-10.42	-68.25	a
4	07/10/2016	S	358200	4735049	7.90	7.73	264.00	49.80	0.89	1.61	0.34	168.06	4.13	1.92	-10.33	-68.10	a
4	15/11/2016	S	358200	4735049	8.10	7.59	260.00	54.60	1.12	1.95	0.03	171.72	3.31	1.58	-10.26	-67.48	a
4	06/12/2016	S	358200	4735049	n.d.	7.54	n.d.	47.97	1.13	1.61	0.41	161.30	3.49	1.74	n.d.	n.d.	a
4	12/05/2017	S	358200	4735049	9.10	7.83	260.00	49.88	1.03	1.91	0.29	161.04	3.63	1.90	-9.99	-66.95	a
4	12/05/2017	S	358200	4735049	n.d.	n.d.	n.d.	54.31	1.02	1.77	0.63	n.d.	4.35	2.04	n.d.	n.d.	a
4	15/09/2017	S	358200	4735049	8.30	7.71	263.00	48.20	1.17	2.14	0.12	153.42	4.82	1.45	-10.06	-66.58	a
4	02/10/2018	S	358200	4735049	8.20	7.90	256.00	47.20	1.32	1.88	0.12	171.11	3.57	1.53	-10.13	-67.65	a
4	23/05/2019	S	358200	4735049	8.40	7.99	265.00	48.40	1.42	1.97	0.47	138.47	3.66	1.69	-10.29	-68.23	a
5	20/12/2002	S	355349	47332016	8.40	7.76	287.00	39.40	6.80	n.d.	0.46	150.10	3.56	3.21	-10.45	-71.99	a
5	20/02/2003	S	355349	47332016	8.40	7.85	294.00	49.03	7.43	n.d.	0.50	181.80	3.27	4.34	-10.04	-72.12	a
5	09/05/2003	S	355349	47332016	8.40	7.88	295.00	51.20	7.67	n.d.	0.44	179.80	3.23	4.40	-10.03	-65.23	a
5	16/07/2003	S	355349	47332016	8.40	7.89	294.00	49.09	7.89	n.d.	0.49	183.70	3.09	4.18	-9.58	-62.71	a
5	20/08/2003	S	355349	47332016	8.40	8.02	292.00	52.00	7.90	n.d.	0.50	182.80	3.10	4.14	-9.80	-63.18	a
5	22/09/2003	S	355349	47332016	8.40	7.90	293.70	49.70	7.78	n.d.	0.43	178.40	3.24	4.26	-9.65	-62.34	a
5	10/11/2003	S	355349	47332016	n.d.	n.d.	n.d.	51.50	7.75	n.d.	0.41	180.80	3.60	4.73	-9.89	-61.04	a
5	22/01/2004	S	355349	47332016	8.40	7.88	293.00	51.30	7.51	n.d.	0.42	179.90	3.61	4.69	-9.96	-63.49	a
5	23/02/2004	S	355349	47332016	8.30	7.88	294.00	51.10	7.56	n.d.	0.42	n.d.	n.d.	4.36	-10.05	-64.54	a
5	09/04/2004	S	355349	47332016	8.30	7.88	294.00	46.30	7.70	n.d.	0.42	n.d.	n.d.	n.d.	n.d.	n.d.	a
5	26/05/2004	S	355349	47332016	n.d.	n.d.	n.d.	46.00	7.46	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	a

5	28/06/2004	S	355349	47332016	n.d.	n.d.	n.d.	45.40	7.50	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	a	
5	30/07/2004	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-8.54	-56.05	a
5	27/08/2004	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.18	-69.15	a
5	05/10/2004	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-8.77	-58.43	a
5	23/11/2004	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-8.87	-58.21	a
5	18/02/2005	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.47	-63.37	a
5	12/04/2005	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.74	-66.84	a
5	24/05/2005	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-8.62	-57.16	a
5	29/06/2005	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.13	-69.37	a
5	20/07/2005	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.09	-68.11	a
5	28/10/2005	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.22	-68.35	a
5	25/11/2005	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.24	-68.29	a
5	23/12/2005	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.85	-65.45	a
5	31/01/2006	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.93	-65.63	a
5	08/03/2006	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.17	-69.06	a
5	07/04/2006	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.14	-68.35	a
5	26/07/2006	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.00	-69.00	a
5	29/09/2006	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.70	-68.20	a
5	06/12/2006	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.90	-68.60	a
5	26/01/2007	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.90	-69.10	a
5	13/03/2007	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.00	-69.20	a
5	24/04/2007	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.10	-69.10	a
5	24/05/2007	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.20	-69.50	a
5	05/07/2007	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.00	-68.80	a
5	26/07/2007	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.10	-68.80	a
5	29/08/2007	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.20	-69.10	a
5	19/10/2007	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.00	-69.10	a
5	20/12/2007	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.20	-68.70	a
5	18/02/2008	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.20	-70.20	a
5	17/03/2008	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.20	-71.30	a
5	08/05/2008	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.90	-69.40	a
5	19/05/2008	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.00	-70.10	a
5	10/06/2008	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.03	-70.22	a
5	02/07/2008	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.01	-69.98	a
5	21/07/2008	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.02	-70.15	a
5	06/08/2008	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.06	-71.20	a
5	29/09/2008	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.96	-70.14	a
5	29/10/2008	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-6.06	-38.54	a
5	16/03/2009	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.88	-69.11	a
5	25/03/2009	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.90	-68.80	a
5	25/05/2009	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.94	-68.77	a
5	21/07/2009	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.40	-69.30	a
5	27/08/2009	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.20	-71.20	a
5	29/09/2009	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.20	-63.60	a
5	15/04/2014	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.08	-66.71	a
5	26/09/2014	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.15	-66.89	a
5	28/01/2016	S	355349	47332016	7.30	n.d.	314.00	54.82	6.31	2.57	0.81	177.40	4.47	6.55	n.d.	n.d.	a	
5	09/02/2016	S	355349	47332016	n.d.	n.d.	n.d.	55.49	6.70	2.56	0.79	189.80	4.25	6.21	n.d.	n.d.	a	
5	18/02/2016	S	355349	47332016	n.d.	n.d.	n.d.	58.39	5.98	2.74	0.85	202.80	5.68	6.68	n.d.	n.d.	a	
5	08/03/2016	S	355349	47332016	n.d.	n.d.	n.d.	57.48	6.02	2.58	0.75	197.60	5.18	6.49	n.d.	n.d.	a	
5	18/03/2016	S	355349	47332016	7.70	7.86	327.00	57.60	5.50	2.54	0.80	199.60	6.16	5.70	n.d.	n.d.	a	
5	18/03/2016	S	355349	47332016	n.d.	7.92	n.d.	48.90	6.40	2.03	0.53	178.60	4.13	3.78	n.d.	n.d.	a	
5	18/03/2016	S	355349	47332016	n.d.	7.89	n.d.	49.26	6.41	1.99	0.45	178.60	4.17	3.84	n.d.	n.d.	a	
5	29/03/2016	S	355349	47332016	n.d.	8.06	n.d.	57.56	5.72	2.59	0.78	196.90	5.26	5.76	n.d.	n.d.	a	
5	12/04/2016	S	355349	47332016	n.d.	7.67	n.d.	53.87	6.13	2.51	0.74	187.10	4.27	5.57	n.d.	n.d.	a	
5	30/08/2016	S	355349	47332016	9.00	7.72	287.00	47.90	6.03	2.20	0.64	195.65	3.83	4.35	-10.23	-67.03	a	
5	07/10/2016	S	355349	47332016	9.00	7.74	297.00	46.40	6.72	1.98	0.46	190.63	4.03	5.37	-10.30	-67.40	a	
5	15/11/2016	S	355349	47332016	8.50	7.45	279.00	50.30	3.44	2.60	0.54	180.36	4.05	2.72	-9.94	-66.46	a	
5	18/11/2016	S	355349	47332016	n.d.	7.22	n.d.	41.42	6.53	1.73	0.42	187.30	4.37	6.55	n.d.	n.d.	a	
5	06/12/2016	S	355349	47332016	n.d.	7.33	n.d.	45.50	6.98	1.96	0.48	185.40	3.58	5.08	n.d.	n.d.	a	

5	27/05/2020	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.34	-69.40	b
5	27/07/2020	S	355349	47332016	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.45	-70.30	b
6	18/03/2016	LS/R	349167	4750614	8.70	7.66	297.00	50.50	0.49	1.24	0.31	156.00	2.26	1.48	-9.57	n.d.	a
6	29/03/2016	LS/R	349167	4750614	n.d.	8.05	n.d.	48.80	0.47	1.15	0.29	150.00	2.19	1.39	-9.67	n.d.	a
6	27/11/2020	LS/R	349167	4750614	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.57	-63.59	a
7	28/01/2016	S	350778	4750224	n.d.	n.d.	n.d.	57.40	0.71	1.42	0.40	184.00	4.22	1.75	n.d.	n.d.	a
7	09/02/2016	S	350778	4750224	n.d.	n.d.	n.d.	52.90	0.71	1.48	0.41	154.00	4.42	1.83	n.d.	n.d.	a
7	18/02/2016	S	350778	4750224	n.d.	n.d.	n.d.	57.30	0.64	1.58	0.40	171.00	4.71	1.67	n.d.	n.d.	a
7	08/03/2016	S	350778	4750224	n.d.	n.d.	n.d.	56.70	0.71	1.53	0.40	178.00	4.80	1.96	n.d.	n.d.	a
7	18/03/2016	S	350778	4750224	7.80	7.32	284.00	57.90	0.65	1.38	0.38	175.00	5.17	1.75	n.d.	n.d.	a
7	29/03/2016	S	350778	4750224	n.d.	7.67	n.d.	58.90	0.67	1.45	0.37	177.00	4.86	1.81	n.d.	n.d.	a
7	12/04/2016	S	350778	4750224	n.d.	7.41	n.d.	58.20	0.67	1.42	0.37	174.00	4.89	1.78	n.d.	n.d.	a
7	06/06/2016	S	350778	4750224	n.d.	6.81	296.00	62.60	1.06	2.65	1.67	n.d.	6.84	1.71	-9.70	-64.00	c
7	06/09/2016	S	350778	4750224	n.d.	7.91	288.00	59.10	1.30	2.50	2.15	n.d.	6.74	1.64	-9.60	-64.00	c
7	27/09/2016	S	350778	4750224	9.50	7.52	314.00	48.80	0.93	3.19	2.16	n.d.	5.94	1.33	n.d.	n.d.	c
7	26/10/2016	S	350778	4750224	n.d.	6.62	278.00	50.70	1.15	2.83	2.05	n.d.	5.92	1.79	n.d.	n.d.	c
7	28/10/2016	S	350778	4750224	n.d.	6.63	271.00	42.60	1.18	3.02	1.99	n.d.	5.75	1.34	n.d.	n.d.	c
7	02/11/2016	S	350778	4750224	n.d.	6.72	288.00	61.40	1.29	2.69	2.10	n.d.	6.36	1.67	-9.70	-64.00	c
7	04/11/2016	S	350778	4750224	n.d.	7.04	290.00	61.90	1.21	2.75	1.64	n.d.	5.70	1.34	n.d.	n.d.	c
7	07/11/2016	S	350778	4750224	n.d.	7.76	287.00	59.70	1.11	2.77	2.10	n.d.	6.18	1.67	n.d.	n.d.	c
7	07/11/2016	S	350778	4750224	n.d.	7.72	286.00	60.30	1.08	2.73	2.08	n.d.	6.30	1.64	n.d.	n.d.	c
7	10/11/2016	S	350778	4750224	n.d.	7.02	285.00	59.60	1.20	2.32	1.95	n.d.	5.73	1.38	-9.60	-64.00	c
7	15/11/2016	S	350778	4750224	n.d.	6.99	287.00	56.70	1.01	2.60	1.70	n.d.	6.23	1.60	-9.70	-64.00	c
7	15/11/2016	S	350778	4750224	n.d.	6.94	287.00	59.20	1.01	2.51	1.80	n.d.	6.27	1.62	n.d.	n.d.	c
7	18/11/2016	S	350778	4750224	8.70	8.44	254.00	55.20	1.02	2.57	1.80	n.d.	6.23	1.63	n.d.	n.d.	c
7	23/11/2016	S	350778	4750224	n.d.	7.67	247.00	46.00	1.14	2.76	1.77	n.d.	5.75	1.35	-9.60	-64.00	c
7	06/12/2016	S	350778	4750224	n.d.	7.55	248.00	61.20	1.14	2.81	1.96	n.d.	6.28	1.61	n.d.	n.d.	c
7	09/01/2017	S	350778	4750224	n.d.	7.40	243.00	47.50	0.98	2.76	1.77	n.d.	5.72	1.35	n.d.	n.d.	c
7	01/02/2017	S	350778	4750224	n.d.	7.63	256.00	59.20	1.05	2.58	1.71	n.d.	6.38	1.62	-9.60	-64.00	c
7	01/03/2017	S	350778	4750224	n.d.	7.49	248.00	53.40	1.04	2.38	1.98	n.d.	6.52	1.70	-9.70	-64.00	c
7	20/03/2017	S	350778	4750224	n.d.	7.70	293.00	55.70	1.16	2.78	1.77	n.d.	6.05	1.52	n.d.	n.d.	c
7	20/03/2017	S	350778	4750224	n.d.	7.67	292.00	53.90	1.27	2.79	1.78	n.d.	5.99	1.51	n.d.	n.d.	c
7	03/04/2017	S	350778	4750224	n.d.	7.70	303.00	60.40	1.17	2.63	1.94	n.d.	6.61	1.77	n.d.	n.d.	c
7	10/04/2017	S	350778	4750224	n.d.	7.80	283.00	57.60	1.01	2.69	2.36	n.d.	6.18	1.69	n.d.	n.d.	c
7	10/05/2017	S	350778	4750224	n.d.	7.65	305.00	57.30	1.06	2.51	1.85	n.d.	7.78	1.83	-9.60	-65.00	c
7	10/05/2017	S	350778	4750224	n.d.	7.71	304.00	56.30	0.97	2.49	1.84	n.d.	7.60	1.82	-9.60	-64.00	c
8	21/06/2013	S	349834	4750561	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.22	-70.30	a
8	11/09/2013	S	349834	4750561	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.21	-58.70	a
8	22/11/2013	S	349834	4750561	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.52	-70.87	a
8	16/04/2014	S	349834	4750561	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.09	-67.80	a
8	26/09/2014	S	349834	4750561	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.24	-69.19	a
8	09/02/2016	S	349834	4750561	n.d.	n.d.	n.d.	45.10	5.65	1.21	0.43	141.00	2.26	n.d.	n.d.	n.d.	a
8	18/02/2016	S	349834	4750561	n.d.	n.d.	n.d.	47.40	5.96	1.43	0.51	143.00	2.43	21.22	n.d.	n.d.	a
8	08/03/2016	S	349834	4750561	n.d.	n.d.	n.d.	47.40	5.83	1.33	0.43	151.00	2.29	21.00	n.d.	n.d.	a
8	10/03/2016	S	349834	4750561	n.d.	n.d.	n.d.	47.00	5.94	1.28	0.43	150.00	2.24	21.56	n.d.	n.d.	a
8	06/09/2016	S	349834	4750561	8.70	7.95	305.00	48.50	6.99	1.46	0.70	154.00	2.49	28.70	-10.38	-68.53	a
8	04/11/2016	S	349834	4750561	8.20	7.74	305.00	55.30	5.95	1.78	0.81	160.00	2.27	24.26	-10.43	-69.58	a
8	18/05/2017	LS/R	349122	4750765	n.d.	n.d.	n.d.	44.50	6.85	1.38	0.60	135.00	2.36	29.03	n.d.	n.d.	a
8	31/10/2017	S	349834	4750561	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.54	-70.59	a
8	28/11/2017	S	349834	4750561	7.90	7.74	293.00	44.50	5.77	1.39	0.41	158.00	2.52	21.82	-10.55	-70.54	a
8	07/12/2017	S	349834	4750561	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.09	-69.07	a
8	03/02/2018	S	349834	4750561	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.38	-70.21	a
8	20/03/2018	S	349834	4750561	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.17	-69.85	a
8	04/04/2018	S	349834	4750561	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.34	-69.63	a
8	20/04/2018	S	349834	4750561	8.30	8.00	291.00	45.20	6.62	1.66	0.32	152.00	2.88	26.01	-10.08	-67.87	a
8	08/05/2018	S	349834	4750561	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.41	-70.37	a
8	20/07/2018	LS/R	349122	4750765	9.10	8.27	299.00	46.60	5.73	1.69	0.52	161.00	7.29	21.13	-10.49	-70.02	a
8	28/09/2018	S	349834	4750561	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.40	-70.48	a
8	26/06/2019	LS/R	349122	4750765	10.70	8.63	326.00	48.00	6.12	1.27	0.47	173.00	2.51	23.73	-10.28	-68.87	a
8	14/01/2020	LS/R	349122	4750765	n.d.	n.d.	n.d.	53.10	9.65	1.51	0.60	154.00	2.57	22.52	n.d.	n.d.	a
8	26/06/2020	LS/R	349122	4750765	n.d.	n.d.	n.d.	52.20	8.38	1.52	0.56	156.00	2.33	29.93	n.d.	n.d.	a
8	27/11/2020	LS/R	349122	4750765	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.30	-69.00	a
9	08/03/2016	LS/R	346232	4753613	n.d.	n.d.	n.d.	54.00	4.76	1.59	0.49	175.00	3.43	12.78	n.d.	n.d.	a
9	18/03/2016	LS/R	346232	4753613	7.10	7.58	240.00	54.10	4.73	1.59	0.51	173.00	3.48	12.38	n.d.	n.d.	a
9	29/03/2016	LS/R	346232	4753613	n.d.	7.90	n.d.	52.90	4.78	1.58	0.54	171.00	3.39	12.68	n.d.	n.d.	a
9	16/04/2016	LS/R	346232	4753613	n.d.	7.47	n.d.	53.10	4.84	1.47	0.45	170.00	3.37	12.57	n.d.	n.d.	a
9	04/04/2018	LS/R	346232	4753613	n.d.	n.d.	n.d.	57.40	6.40	1.94	0.77	148.00	3.34	19.32	n.d.	n.d.	a
9	18/07/2018	LS/R	346232	4753613	n.d.	n.d.	n.d.	38.40	6.95	1.61	0.66	119.00	2.66	21.67	n.d.	n.d.	a
9	15/02/2019	LS/R	346232	4753613	n.d.	n.d.	n.d.	54.20	7.80	1.66	0.78	161.00	2.58	25.64	n.d.	n.d.	a
9	16/05/2019	LS/R	346232	4753613	n.d.	n.d.	n.d.	54.30	7.72	1.58	0.65	163.00	2.69	25.71	n.d.	n.d.	a
9	30/05/2019	LS/R	346232	4753613	n.d.	n.d.	n.d.	52.30	7.54	1.61	0.65	162.00	3.19	24.96	n.d.	n.d.	a
9	11/06/2019	LS/R	346232	4753613	n.d.	n.d.	n.d.	54.90	7.39	1.56	0.68	163.00	2.92	25.02	n.d.	n.d.	a

9	25/06/2019	LS/R	346232	4753613	n.d.	n.d.	n.d.	53.10	7.55	1.84	0.90	164.00	3.11	25.34	n.d.	n.d.	a
9	23/07/2019	LS/R	346232	4753613	n.d.	n.d.	n.d.	54.80	7.87	1.50	0.67	162.00	2.43	26.24	n.d.	n.d.	a
9	14/01/2020	LS/R	346232	4753613	n.d.	n.d.	n.d.	54.90	7.72	1.60	0.64	158.00	2.71	25.84	-10.00	-67.95	a
9	27/11/2020	LS/R	346232	4753613	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.33	-69.05	a
10	20/07/2018	LS/R	343960	4754785	11.10	8.14	309.00	52.40	5.20	1.84	4.35	177.00	7.00	16.87	-9.86	-65.51	a
10	27/11/2020	LS/R	343960	4754785	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.22	-68.56	a
11	08/03/2016	S	349404	4756421	n.d.	n.d.	n.d.	49.90	2.02	2.10	0.50	155.00	3.85	3.40	n.d.	n.d.	a
11	18/03/2016	S	349404	4756421	7.20	7.59	260.00	50.20	1.98	2.08	0.51	160.00	4.16	3.26	n.d.	n.d.	a
11	29/03/2016	S	349404	4756421	n.d.	8.06	n.d.	49.90	2.00	2.08	0.50	159.00	4.04	3.38	n.d.	n.d.	a
11	12/04/2016	S	349404	4756421	n.d.	7.80	n.d.	49.40	2.00	1.96	0.49	158.00	3.50	3.32	n.d.	n.d.	a
11	18/07/2018	LS/R	349404	4756421	n.d.	n.d.	n.d.	39.20	2.20	1.61	0.55	120.00	2.53	2.73	n.d.	n.d.	a
11	27/11/2020	LS/R	349404	4756421	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.01	-66.20	a
12	16/04/2014	LS/R	346494	4755324	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.27	-68.75	a
12	26/09/2014	LS/R	346494	4755324	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.11	-67.43	a
12	08/03/2016	LS/R	345593	4755302	n.d.	n.d.	n.d.	50.50	4.17	2.05	0.52	162.00	4.40	11.50	n.d.	n.d.	a
12	18/03/2016	LS/R	345593	4755302	8.70	7.69	283.00	51.10	4.11	2.04	0.54	163.00	4.65	11.23	n.d.	n.d.	a
12	29/03/2016	LS/R	345593	4755302	n.d.	7.04	n.d.	51.30	4.01	1.91	0.52	159.00	4.59	11.02	n.d.	n.d.	a
12	16/04/2016	LS/R	345593	4755302	n.d.	7.11	n.d.	50.10	4.03	1.84	0.50	157.00	4.20	11.03	n.d.	n.d.	a
12	28/11/2017	LS/R	346494	4755324	8.90	7.63	315.00	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.38	-68.86	a
12	18/07/2018	LS/R	345593	4755302	n.d.	n.d.	n.d.	40.10	6.27	2.08	0.76	117.00	3.78	20.97	n.d.	n.d.	a
12	15/02/2019	LS/R	345593	4755302	n.d.	n.d.	n.d.	56.80	7.55	2.28	0.68	157.00	4.17	26.78	n.d.	n.d.	a
12	14/03/2019	LS/R	345593	4755302	n.d.	n.d.	n.d.	56.70	7.33	2.53	0.96	174.00	4.53	26.62	n.d.	n.d.	a
12	30/05/2019	LS/R	345593	4755302	n.d.	n.d.	n.d.	54.30	5.77	1.95	0.66	157.00	3.87	19.62	n.d.	n.d.	a
12	25/06/2019	LS/R	345593	4755302	n.d.	n.d.	n.d.	55.40	6.67	2.13	0.74	162.00	4.43	25.37	n.d.	n.d.	a
12	14/01/2020	LS/R	345593	4755302	n.d.	n.d.	n.d.	57.50	7.38	2.39	0.87	164.00	4.48	25.88	-9.69	-66.12	a
12	27/11/2020	LS/R	345593	4755302	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.14	-67.10	a
13	20/07/2018	R	344032	4754843	10.50	8.12	321.00	51.70	5.36	2.25	4.68	177.00	8.17	21.19	-9.99	-65.22	a
13	26/06/2019	R	344032	4754843	12.90	8.51	360.00	52.50	5.67	3.01	0.55	176.00	4.59	23.06	-9.76	-64.45	a
13	27/11/2020	R	344032	4754843	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.11	-67.05	a
14	15/07/2018	R	343799	4754920	11.00	8.01	314.00	51.70	5.33	1.84	2.75	186.00	5.94	17.37	-10.00	n.d.	a
14	20/07/2018	R	343351	4755051	11.80	8.00	310.00	51.90	4.94	2.10	2.07	181.00	5.15	18.43	-10.02	-67.49	a
14	27/11/2020	R	343799	4754920	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.97	-66.75	a
15	16/04/2014	LS/R	339995	4752650	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.82	-65.33	a
15	26/09/2014	LS/R	339995	4752650	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.00	-66.64	a
15	28/01/2016	LS/R	339995	4752650	9.30		335.00	60.46	4.29	3.27	0.80	203.70	6.05	14.73	n.d.	n.d.	a
15	09/02/2016	LS/R	339995	4752650	n.d.	n.d.	n.d.	61.33	4.38	3.38	0.79	178.10	5.58	14.58	n.d.	n.d.	a
15	18/02/2016	LS/R	339995	4752650	n.d.	n.d.	n.d.	65.79	3.99	4.33	0.76	205.80	7.44	15.32	n.d.	n.d.	a
15	08/03/2016	LS/R	339995	4752650	n.d.	n.d.	n.d.	64.30	4.00	3.79	0.75	193.00	6.62	14.46	n.d.	n.d.	a
15	18/03/2016	LS/R	339995	4752650	10.20	7.43	334.00	61.53	3.91	3.55	0.74	193.00	6.74	13.86	n.d.	n.d.	a
15	29/03/2016	LS/R	339995	4752650	n.d.	8.18	n.d.	61.03	3.85	3.62	0.73	187.40	6.67	14.11	n.d.	n.d.	a
15	12/04/2016	LS/R	339995	4752650	n.d.	7.10	n.d.	59.93	3.86	3.04	0.65	188.40	5.86	13.69	n.d.	n.d.	a
15	31/10/2017	LS/R	339995	4752650	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.91	-66.91	a
15	07/12/2017	LS/R	339995	4752650	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.28	-64.71	a
15	08/05/2018	LS/R	339995	4752650	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.73	-65.00	a
15	20/07/2018	LS/R	339995	4752650	12.60	8.28	324.00	55.10	4.46	2.79	1.89	188.80	5.78	18.51	-10.02	-66.52	a
15	28/09/2018	LS/R	339995	4752650	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.96	-66.82	a
15	26/06/2019	LS/R	339995	4752650	14.60	8.60	408.00	56.80	4.47	2.75	18.78	188.19	4.83	19.80	-9.99	-64.53	a
15	27/11/2020	LS/R	339995	4752650	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.93	-66.11	a
16	16/04/2014	LS/R	336062	4749507	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.87	-65.83	a
16	26/09/2014	LS/R	336062	4749507	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.93	-65.93	a
16	20/07/2018	LS/R	336062	4749507	12.20	8.14	333.00	56.50	4.13	3.07	2.03	184.22	6.43	17.88	-9.85	-65.64	a
16	26/06/2019	LS/R	336062	4749507	13.70	9.02	348.00	58.20	4.01	2.90	5.07	192.15	8.96	18.13	-9.89	-64.83	a
17	08/03/2016	LS/R	343290	4745922	n.d.	n.d.	n.d.	59.30	0.78	1.52	0.41	181.80	3.28	1.27	n.d.	n.d.	a
17	18/03/2016	LS/R	343290	4745922	8.50	7.52	291.00	58.51	0.71	1.52	0.43	182.00	3.56	1.21	n.d.	n.d.	a
17	29/03/2016	LS/R	343290	4745922	n.d.	7.63	n.d.	59.68	0.80	1.57	0.44	181.30	3.79	1.29	n.d.	n.d.	a
17	12/04/2016	LS/R	343290	4745922	n.d.	7.26	n.d.	59.23	0.78	1.56	0.44	182.00	3.46	1.22	n.d.	n.d.	a
17	20/06/2017	LS/R	343290	4745922	11.10	7.54	280.00	56.20	0.79	2.23	0.57	180.56	3.20	0.97	-9.80	-64.72	a
17	19/07/2017	LS/R	343290	4745922	9.00	7.65	267.00	53.90	0.74	1.82	0.41	174.46	3.06	0.92	-10.03	-65.81	a
17	21/09/2017	LS/R	343290	4745922	8.90	7.72	272.00	53.90	0.77	1.71	0.25	166.53	3.29	1.15	-9.92	-65.70	a
17	20/07/2018	LS/R	343290	4745922	11.70	7.90	341.00	62.50	2.49	2.38	1.02	205.57	4.38	7.72	-9.56	-62.79	a
17	26/06/2019	LS/R	343290	4745922	13.50	8.39	346.00	63.10	2.46	1.98	1.07	196.42	4.12	7.29	-9.78	-64.87	a
18	20/05/2014	S	344707	4738842	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.32	-62.98	a
18	26/09/2014	S	344707	4738842	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.38	-64.13	a
18	28/01/2016	S	344707	4738842	11.50	n.d.	564.00	87.76	18.55	2.84	0.81	217.90	4.78	103.80	n.d.	n.d.	a
18	09/02/2016	S	344707	4738842	n.d.	n.d.	n.d.	90.41	19.09	2.86	0.79	220.30	5.32	115.50	n.d.	n.d.	a
18	18/02/2016	S	344707	4738842	n.d.	n.d.	n.d.	88.71	18.73	2.81	0.68	219.40	5.53	116.80	n.d.	n.d.	a
18	08/03/2016	S	344707	4738842	n.d.	n.d.	n.d.	88.19	18.60	2.77	0.72	217.20	5.31	115.00	n.d.	n.d.	a
18	18/03/2016	S	344707	4738842	11.40	7.28	561.00	87.40	18.19	2.62	0.74	220.10	5.44	129.60	n.d.	n.d.	a
18	29/03/2016	S	344707	4738842	n.d.	7.23	n.d.	88.82	18.52	2.75	0.73	219.80	5.38	130.30	n.d.	n.d.	a
18	12/04/2016	S	344707	4738842	n.d.	7.36	n.d.	88.70	18.60	2.76	0.73	240.30	5.67	133.60	n.d.	n.d.	a
18	06/12/2016	S	344707	4738842	n.d.	7.20	n.d.	89.37	16.92	2.76	0.78	227.90	6.21	114.90	n.d.	n.d.	a
18	01/02/2017	S	344707	4738842	11.70	7.30	589.00	92.70	10.94	3.55	2.78	248.27	4.03	85.19	-9.72	-64.63	a

18	01/02/2017	S	344707	4738842	11.20	7.39	572.00	82.85	16.46	3.31	0.72	229.67	7.11	85.82	-9.57	-63.81	a
18	13/03/2017	S	344707	4738842	11.50	7.36	590.00	98.25	10.63	3.18	2.08	235.00	3.86	96.89	-9.63	-64.99	a
18	13/03/2017	S	344707	4738842	11.20	7.28	569.00	84.01	15.44	3.22	0.66	221.13	8.55	88.21	-9.36	-64.21	a
18	20/06/2017	S	344707	4738842	12.20	7.32	605.00	94.95	12.62	3.15	2.57	216.25	3.23	118.66	-9.96	-66.86	a
18	20/06/2017	S	344707	4738842	11.20	7.49	570.00	84.35	16.26	3.58	0.87	216.25	7.97	89.84	-9.55	-63.74	a
18	19/07/2017	S	344707	4738842	11.80	7.69	586.00	97.75	12.50	3.16	1.75	198.25	3.87	125.47	-9.77	-65.86	a
18	19/07/2017	S	344707	4738842	10.90	7.48	556.00	84.90	16.54	3.46	1.12	215.03	7.99	100.20	-9.46	-64.02	a
18	21/09/2017	S	344707	4738842	11.20	7.50	606.00	96.70	13.46	2.87	1.27	188.32	3.26	138.87	-9.86	-65.85	a
18	22/11/2017	S	344707	4738842	11.10	7.59	610.00	84.70	17.36	3.78	1.00	222.35	6.06	111.20	-9.73	-65.81	a
18	26/01/2018	S	344707	4738842	10.70	7.50	649.00	97.40	15.58	2.69	1.03	216.25	3.22	141.78	-9.73	-66.32	a
18	26/01/2018	S	344707	4738842	10.90	7.66	616.00	85.60	17.92	3.28	0.66	207.71	5.27	116.92	-9.66	-65.86	a
18	09/03/2018	S	344707	4738842	11.30	7.66	610.00	96.40	15.62	3.45	1.06	216.55	4.61	128.81	-9.78	-66.00	a
18	16/03/2018	S	344707	4738842	10.90	7.45	623.00	100.30	15.44	2.95	1.30	208.32	3.42	142.75	-9.91	-66.23	a
18	16/03/2018	S	344707	4738842	12.30	7.59	596.00	88.30	16.72	3.76	1.22	205.57	5.44	120.66	-9.70	-65.24	a
18	10/04/2018	S	344707	4738842	11.40	7.50	621.00	99.00	14.04	2.67	1.47	211.98	3.44	142.66	-9.82	-66.35	a
18	10/04/2018	S	344707	4738842	11.60	7.49	602.00	92.20	14.88	3.71	1.48	207.10	5.33	122.09	-9.37	-63.95	a
18	18/05/2018	S	344707	4738842	12.50	7.53	625.00	100.00	15.28	2.59	1.52	209.54	3.38	142.89	-9.49	-64.40	a
18	18/05/2018	S	344707	4738842	11.90	7.46	619.00	97.20	14.52	3.50	3.13	209.23	6.71	130.73	-9.70	-64.85	a
18	05/07/2018	S	344707	4738842	12.40	7.59	618.00	96.00	15.24	3.31	9.41	227.53	11.18	131.25	-9.90	-67.06	a
18	10/10/2018	S	344707	4738842	11.20	7.73	628.00	93.40	16.16	3.35	3.28	245.53	7.55	139.19	-9.76	-66.02	a
18	14/11/2018	S	344707	4738842	10.80	7.48	633.00	95.20	16.94	2.48	0.80	204.96	3.13	155.22	-9.84	-66.08	a
18	14/11/2018	S	344707	4738842	10.80	7.42	604.00	88.80	17.04	3.17	0.61	244.61	5.75	127.27	-9.62	-65.39	a
18	17/01/2019	S	344707	4738842	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.59	-64.64	a
18	14/03/2019	S	344707	4738842	11.20	7.04	624.00	96.00	16.94	2.92	1.14	243.39	4.17	139.48	-9.84	-66.36	a
18	14/06/2019	S	344707	4738842	11.20	8.04	653.00	95.10	16.46	2.88	1.30	206.79	4.14	140.70	-9.79	-64.82	a
18	14/06/2019	S	344707	4738842	10.80	7.90	638.00	91.20	16.88	3.24	0.75	211.67	4.70	133.73	-9.91	-65.84	a
18	02/09/2019	S	344707	4738842	11.00	7.50	660.00	96.80	16.62	2.07	0.99	206.79	3.33	148.56	-9.82	-66.01	a
18	02/09/2019	S	344707	4738842	10.80	7.55	643.00	93.00	16.86	2.71	0.82	208.93	4.27	137.19	-9.75	-65.46	a
19	06/12/2016	S	345832	4738085	n.d.	7.36	n.d.	75.10	13.36	1.97	0.55	206.20	3.29	88.76	n.d.	n.d.	a
19	16/12/2016	S	345832	4738085	9.60	n.d.	461.00	70.90	10.94	2.26	0.50	190.69	3.52	60.73	-9.92	-66.08	a
19	25/01/2017	S	345832	4738085	9.70	7.54	435.00	68.20	10.91	2.48	0.39	192.87	3.41	52.27	n.d.	n.d.	a
19	01/02/2017	S	345832	4738085	10.00	7.50	449.00	67.80	10.14	2.45	0.35	213.81	3.36	52.53	-9.96	-66.01	a
19	10/02/2017	S	345832	4738085	n.d.	7.50	424.00	65.10	8.87	2.30	0.41	275.72	4.20	51.16	-9.42	-64.71	a
19	02/03/2017	S	345832	4738085	n.d.	n.d.	n.d.	83.67	18.53	1.98	0.60	n.d.	3.78	145.60	n.d.	n.d.	a
19	13/03/2017	S	345832	4738085	9.90	7.56	449.00	69.45	10.15	2.19	0.50	200.69	3.28	54.84	-9.83	-66.13	a
19	28/03/2017	S	345832	4738085	n.d.	n.d.	n.d.	79.89	17.38	2.01	0.61	n.d.	3.73	135.40	n.d.	n.d.	a
19	13/04/2017	S	345832	4738085	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	3.68	130.10	n.d.	n.d.	a
19	13/04/2017	S	345832	4738085	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	3.80	133.00	n.d.	n.d.	a
19	13/04/2017	S	345832	4738085	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	4.01	126.40	n.d.	n.d.	a
19	13/04/2017	S	345832	4738085	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	3.64	146.20	n.d.	n.d.	a
19	20/06/2017	S	345832	4738085	9.90	7.49	447.00	71.35	11.16	2.22	0.31	204.05	3.20	51.94	-9.88	-65.79	a
19	19/07/2017	S	345832	4738085	10.00	7.75	428.00	68.65	11.52	2.33	0.42	204.35	3.20	53.82	-9.81	-65.60	a
19	31/07/2017	S	345832	4738085	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-10.03	-67.28	a
19	05/09/2017	S	345832	4738085	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.89	-68.50	a
19	21/09/2017	S	345832	4738085	10.40	7.60	546.00	81.20	17.02	2.44	0.68	190.46	3.19	110.70	-9.88	-66.29	a
19	31/10/2017	S	345832	4738085	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.99	-68.54	a
19	07/12/2017	S	345832	4738085	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.68	-68.36	a
19	26/01/2018	S	345832	4738085	10.10	7.66	544.00	75.30	16.72	2.22	0.52	212.89	3.60	95.60	-9.91	-66.76	a
19	01/02/2018	S	345832	4738085	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.84	-66.52	a
19	01/02/2018	S	345832	4738085	10.00	7.79	457.00	65.90	11.52	2.12	0.50	195.81	3.30	56.40	-9.88	-65.88	a
19	09/03/2018	S	345832	4738085	10.90	7.72	479.00	72.80	14.14	2.39	0.57	218.38	3.40	74.46	-9.93	-66.57	a
19	16/03/2018	S	345832	4738085	10.60	7.61	495.00	75.40	15.62	2.25	0.65	204.96	3.44	89.54	-9.98	-66.02	a
19	16/03/2018	S	345832	4738085	10.10	7.57	434.00	69.20	10.84	2.45	0.60	205.57	3.50	56.38	-10.16	-66.66	a
19	10/04/2018	S	345832	4738085	10.60	7.63	494.00	75.90	14.24	2.21	0.97	208.62	3.31	88.72	-9.78	-66.40	a
19	10/04/2018	S	345832	4738085	10.30	7.54	435.00	66.40	10.42	2.89	0.89	205.57	3.36	56.06	-9.77	-65.90	a
19	08/05/2018	S	345832	4738085	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.85	-66.55	a
19	18/05/2018	S	345832	4738085	10.40	7.59	506.00	73.90	15.06	2.19	0.98	203.13	3.15	89.85	-9.66	-64.47	a
19	18/05/2018	S	345832	4738085	9.90	7.57	432.00	66.50	10.80	2.11	0.99	207.71	3.48	56.99	-9.44	-63.23	a
19	18/05/2018	S	345832	4738085	11.40	7.77	480.00	73.20	13.38	2.24	0.88	210.45	3.35	75.82	-9.81	-65.28	a
19	05/07/2018	S	345832	4738085	10.80	7.77	480.00	71.40	13.04	2.27	0.38	207.71	3.39	76.87	-9.93	-67.25	a
19	28/09/2018	S	345832	4738085	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-9.99	-67.71	a
19	10/10/2018	S	345832	4738085	10.50	7.77	481.00	70.30	18.58	2.28	0.37	244.00	3.21	77.85	-9.63	-66.21	a
19	14/11/2018	S	345832	4738085	10.10	7.55	519.00	72.90	16.34	2.18	0.31	212.28	3.39	97.00	-9.81	-65.92	a
19	14/11/2018	S	345832	4738085	9.80	7.48	438.00	65.60	11.64	2.22	0.25	209.84	3.30	57.38	-9.64	-65.85	a
19	17/01/2019	S	345832	4738085	7.10	7.81	482.00	72.20	13.70	2.45	0.74	n.d.	3.74	79.16	-9.79	-66.33	a
19	14/03/2019	S	345832	4738085	10.40	7.66	456.00	75.60	10.54	2.50	0.75	209.84	4.38	52.79	-9.76	-65.74	a
19	14/03/2019	S	345832	4738085	10.40	7.86	478.00	71.00	14.16	2.18	0.78	230.58	3.30	76.13	-9.91	-66.71	a
19	14/06/2019	S	345832	4738085	10.50	7.69	476.00	72.70	11.36	2.47	0.54	201.91	3.77	54.61	-9.57	-64.91	a
19	14/06/2019	S	345832	4738085	10.10	8.36	525.00	70.50	16.50	2.14	0.60	202.83	3.12	87.51	-10.15	-67.46	a
19	14/06/2019	S	345832	4738085	12.00	8.20	496.00	69.60	14.32	2.21	0.85	206.49	3.44	73.24	-9.68	-63.54	a
19	02/09/2019	S	345832	4738085	11.00	7.69	484.00	77.00	8.31	2.49	0.62	234.55	4.31	43.92	-9.39	-63.32	a
19	02/09/2019	S	345832	4738085	10.00	7.65	517.00	71.10	15.14	1.92	0.59	201.00	3.20	83.80	-9.89	-66.47	a

