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MOUNTAINHYDRO: INTEGRATING GEOLOGY, GEOPHYSICS AND HYDROGEOLOGY IN A MULTIDISCIPLINARY APPROACH

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*'There are more things in heaven and earth, Horatio,
than are dreamt of in your philosophy'
William Shakespeare*

To my father, Vanni

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Abstract

Spring water represents a vital source of drinking water for local communities in mountain areas worldwide. As water demand and pressure on groundwater resources increase in a context of changing climate, there is a growing commitment to assess groundwater availability and to develop sustainable water supply strategies.

Effective groundwater management relies on a comprehensive characterization of aquifer systems. In this regard, karst-fractured terrains pose unique challenges due to their complex and heterogeneous nature. This study focuses on the assessment of aquifer systems in karst-fissured mountain environments, by employing a multidisciplinary approach based on three pillars: geology, geophysics, and hydrogeology.

The **geological** pillar centers on the construction of a 3D geological model derived from a combination of existing geological maps and targeted field observations. The model captures the spatial distribution of geological units and major structural lineaments and serves as a framework for the interpretation of groundwater preferential flow paths.

The **geophysical** pillar involves the use of high resolution DC resistivity surveys and densely spaced transient electromagnetic (EM) methods, both ground-based and airborne. Geophysical surveys play a key role in capturing the subsurface heterogeneity and identifying preferential flow paths within the rock volumes. The integration of the geological model into the geophysical inversion through an iterative process provides insights into the hydrogeological characteristics of the investigated system and enhances reliability of the interpretations. Moreover, the 3D inversion of airborne EM data has proven effective in increasing model accuracy in topographically and geologically complex terrains.

The **hydrogeological** pillar consists of groundwater chemical and stable isotope analyses, as well as hydrodynamic characterization. These analyses support the delineation of recharge areas of the springs and help characterize the behavior of the aquifer system. In addition, numerical simulations of groundwater flow were carried out to test different scenarios of recharge and flow paths in order to validate the conceptual model.

The results demonstrate the strength of an iterative, integrated modelling strategy in improving the understanding of complex karst-fissured aquifers.

Two representative study areas were selected to implement and validate the multidisciplinary approach: one in northern Italy (Val Sabbia, Lombardy), and the other in southern Spain (Sierra de Aracena, Andalusia). These sites exemplify geologically complex mountain terrains, providing robust testing ground for the methodology.

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1. Introduction

1.1 Object of the research

Climate change is one of the main challenges of today's era, with deep effects on natural resources, including water availability. Its impacts are particularly severe in mountain regions, where changes in precipitation patterns, rising temperatures, and altered hydrological cycles threaten the stability of water resources and its accessibility (Beniston, 2012; Chaturvedi et al., 2021; Matiu et al., 2021). In such fragile environments, ensuring a sustainable water supply is crucial, particularly for communities that rely on groundwater as their primary domestic water resource.

To give an example related to the most densely populated region of Italy, in Lombardy mountains cover 40.4% of the territory (ISTAT, 2023). Here, groundwater springs play a significant role in meeting water demand, supplying 17.4% of the region's water use in 2020 (ISTAT, 2020). These springs are fed by aquifers whose characterization is key to sustainable water management. However, despite their strategic importance, groundwater resources remain poorly understood, leading to mismanagement and increased vulnerability to water scarcity (United Nations, 2022).

This study aims to establish a systematic methodology for characterizing natural springs in mountain environments and assessing their potential for sustainable water resource management. A comprehensive understanding of the hydrodynamics of these springs relies on detailed knowledge of the subsurface structures and hydrostratigraphic framework.

To achieve this, the research develops an integrated approach that combines geological, geophysical, and hydrogeological methods for a thorough investigation of groundwater systems in complex, tectonically deformed mountain environments.

By incorporating 3D geological modelling, hydrogeological data analysis, and geophysical imaging, this study will provide critical insights into the spatial distribution of geological formations, fault networks, and karst processes that influence groundwater flow and storage. This multidisciplinary approach will help evaluate their impact on spring recharge, groundwater dynamics, and long-term resource sustainability, offering valuable information for effective water resource management.

To enhance the understanding of groundwater systems, the study employs:

- A 3D geological model, developed using existing geological data and refined with new field observations.

- The analyses of water level and its physicochemical properties to characterize the springs dynamics.
- Ground-based and airborne geophysical surveys, including electrical and electromagnetic methods, to improve subsurface imaging.

The natural world is inherently three-dimensional (3D), yet our understanding of groundwater reservoirs is often fragmented due to limited subsurface data. The structural and stratigraphic setting in which the groundwater resource is hosted determines its availability and sustainable discharge (Bense et al., 2013; Fetter & Kreamer, 2022). Therefore, creating a reliable 3D digital representation of the overall subsurface architecture and its characteristics is essential to support the subsequent processes of developing, managing, and protect the resource itself. In this sense, 3D geological models have become powerful tools for visualizing and understanding the geometry, distribution and connectivity of the rock volumes hosting aquifer systems, promoting access to continuous and homogeneous information. Furthermore, 3D geological models can be used as a foundation for numerical flow simulations, which are useful for the understanding of groundwater flow paths.

However, despite their advantages, 3D geological models rely on sparse and spatially limited geological data, leading to uncertainties in subsurface geometry interpretations. To overcome this limitation, indirect geophysical methods are employed. These techniques are designed to measure physical properties of subsurface materials, allowing for a detailed imaging of geological features geometry and discontinuities at depth (Kirsch, 2009).

One of the key innovations of this study is the use of the 3D geological model as a constraint for geophysical inversions. By integrating a 3D geological model, the inversion process is guided by prior geological knowledge, reducing ambiguity and improving the representation of geological features.

This approach ensures that the resistivity models produced by geophysical inversion align more closely with known geological information, thus reducing the uncertainties in the final model. Moreover, the 3D geological model serves as a bridge between different disciplines (geology, hydrogeology, and geophysics) providing a comprehensive understanding of groundwater flow dynamics and structural controls on groundwater movement.

By constraining geophysical inversion with geological data, this study demonstrates a novel methodology that enhances subsurface characterization, especially in geologically complex mountain environments.

Another key innovation of this study was the application of geophysical methods with broad spatial coverage. In particular, the cutting-edge airborne electromagnetic (AEM) surveying was used among the first applications for mapping groundwater systems in a mountain, karst-fractured environment. The rugged topography and complex subsurface structural setting, which result in strong resistivity contrasts, required the use of 3D inversion of the AEM data. As an alternative to AEM, the Loupe system was also considered in this study. Loupe is an innovative ground-based EM acquisition tool that provides areal coverage, though with a more limited spatial extent and shallower investigation depth compared to AEM. In this study, Loupe was used for the first time in a mountain setting, offering a complementary approach to airborne methods.

To test the effectiveness of this multidisciplinary approach, two study sites were selected for its application and the necessary comparisons between the comparable settings.

The first site is located in the mountain area of Val Sabbia, within the Brescian Prealps, part of the Lombardy region in northern Italy. The area is characterized by carbonate rock formations, that are potentially affected by karst processes. This study area is home to several groundwater springs, which evidence the occurrence of complex groundwater flow systems. Due to their natural abundance and high water quality, these springs are tapped by A2A Ciclo Idrico, the local water network manager, to supply the territory's communities with drinking water. However, recent years have witnessed growing concerns about their sustainability. Seasonal fluctuations, likely exacerbated by climate change, have significantly impacted spring discharge, exposing local communities to water shortages and interruptions.

Here, a 3D geological model was built, springs were monitored, and geophysical surveys were carried out, both ground-based and airborne. However, despite the high density of AEM data collected in Val Sabbia, they have not been made available for analysis. As a consequence, to demonstrate the importance of AEM data for groundwater characterization, the second site was selected to provide the missing dataset in the comparable setting of the Sierra de Aracena region in Spain. This area shares geological and hydrogeological similarities with Val Sabbia, making it a suitable reference site where AEM data are available. By analyzing this dataset, the study aims to showcase the effectiveness of AEM techniques in groundwater resource assessment and highlight their potential application in the study area.

1.2 State of the art

The study of groundwater in mountain environments deploys different approaches to address the complex interactions between geological and hydrogeological factors. These approaches integrate various disciplines to achieve a comprehensive understanding of spring systems.

3D geological modelling has emerged as a critical tool in hydrogeology, offering a detailed and spatially coherent representation of the subsurface that enhances the conceptual understanding of groundwater systems. These models enable the characterization of aquifer geometry, stratigraphic continuity, and structural controls on flow dynamics, which are essential for informed groundwater management and sustainability planning.

The application of 3D modelling allows for a more precise delineation of aquifer extents and hydraulic connectivity, as shown in the work of Hassen et al. (2016), who explored the spatial configuration of an aquifer system in Tunisia. Their study underscores the role of geological continuity in controlling flow pathways and resource distribution, thus supporting quantitative assessments of groundwater availability. Moreover, as emphasized by Thornton et al. (2018), in tectonically complex terrains, the integration of structural geology into 3D models enhances the resolution of hydrogeological interpretations. Similarly, D’Affonseca et al. (2020) employed implicit 3D modelling techniques to conceptualize a karst aquifer influenced by tectonic features in Germany. Their approach facilitated the identification of preferential flow paths and structural barriers, advancing the hydrogeological understanding of fractured and heterogeneous systems.

In the context of resource risk mitigation, Mas et al. (2022) demonstrated how 3D geological modelling reduces uncertainty in subsurface exploration, particularly for deep geothermal applications. The enhanced depiction of subsurface heterogeneity directly informed risk assessment and strategic planning.

Expanding the methodological scope, Stevenazzi et al. (2023) integrated 3D hydrostratigraphic modelling with hydrochemical and hydrological analyses to characterize Alpine aquifer systems. This comprehensive framework enabled the identification of groundwater bodies and recharge mechanisms, demonstrating the benefits of combining geological and hydrogeological data in a spatially explicit model.

By offering a detailed 3D representation of the subsurface, geological models enhance the understanding of groundwater systems. They serve as foundational tools for advancing hydrogeological theory, improving predictive capabilities, and supporting robust, evidence-based decision-making for groundwater governance.

In the context of Italian mountain aquifer systems, particularly those in carbonate rocks, improving the understanding of groundwater systems increasingly relies on integrating hydrodynamic, geochemical, and isotopic analyses. This multidisciplinary approach offers a better understanding of the complexities of groundwater flow. It identifies recharge processes and characterizes the structural and functional behavior of fractured and karst aquifers.

Isotopic tools have demonstrated critical value in delineating flow paths and identifying recharge areas in systems where traditional hydrogeological investigations are constrained by heterogeneity and inaccessibility. Barbieri et al. (2005) highlight the importance of stable isotope monitoring in characterizing groundwater circulation in a karst-fissured aquifer in the Central Apennines, revealing flow patterns that would otherwise remain concealed. Similarly, Doveri et al. (2013) confirm the central role of isotopic analysis in karst systems, where structural complexities and nonlinear hydrodynamics challenge conventional interpretative methods.

Aquifer characterization gains greater resolution when isotopic techniques are integrated with hydrochemical analyses and field-based investigations. This is exemplified by Calligaris et al. (2016), who employ a multidisciplinary approach to delineate the hydrogeological behavior of complex hydrostructure. Their work underscores how such integration is essential for sustainable management and exploitation of groundwater resources. In addition, Vigna & Banzato (2015) investigate the contrasting hydrodynamic and hydrochemical responses of different Alpine aquifer types in a carbonate context. Their findings reflect the diversity of internal aquifer architectures and recharge-discharge relationships, revealing the need for site-specific assessments within broader hydrogeological frameworks. More recently, Corniello et al. (2024) employ a combined hydrochemical and isotopic approach to elucidate the groundwater relationships between two carbonate massifs in the Campania region. Their study quantified groundwater resources and serves as a decision-support tool for integrated and sustainable water management in structurally complex regions.

Together, these studies contribute to a growing body of knowledge that affirms the essential nature of integrating isotopic, hydrochemical, and hydrodynamic methods for accurately characterizing the structure and behavior of mountain aquifer systems. This multidimensional understanding is crucial for making informed assessments of groundwater resources, planning for resilience, and ensuring long-term sustainability in sensitive and often overexploited environments.

The effectiveness of geophysical methods in groundwater studies is widely demonstrated through numerous case studies in scientific literature. Their reliability is based on their ability to generate subsurface images by detecting variations in physical properties such as electrical resistivity. This capability allows for the identification of key hydrogeological features, including lithological boundaries and fault networks, which are essential for understanding groundwater flow dynamics. By providing a non-invasive means of characterizing subsurface structures, geophysical techniques play a crucial role in groundwater exploration.

An overview of contributions made by geophysical methods to karst-system exploration is provided by Chalikakis et al. (2011). In addition, Binley et al. (2015) present how geophysical methods have emerged as valuable tools for investigating shallow subsurface processes and offer a vision for future developments relevant to hydrology. The work of Tarik et al. (2023) provides a demonstration of the application of DC electrical resistivity surveys to identify karst features such as sinkholes in an area of Morocco.

Furthermore, Knight et al. (2018), using the airborne electromagnetic (AEM) method to map the aquifer systems of the Central Valley in California, assess that the acquisition of AEM data could provide critical information about the hydrostratigraphy of the aquifer systems needed for groundwater management.

In geologically complex terrain, the traditional 1D forward modelling approach commonly used to model AEM data may not be sufficient for accurate subsurface imaging. In such environments, the simplifying assumption of horizontal layering fails to account for the structural complexity and topographic effects that significantly influence the EM response. Although 3D modelling of AEM data requires considerably more computational resources, numerous studies have demonstrated its advantages in terms of accuracy and reliability when applied to field data. Most of the existing literature focuses on mineral exploration, where the inherently 3D nature of ore bodies justifies the use of 3D inversion techniques (L. Cox et al., 2024; L. H. Cox et al., 2010; McMillan et al., 2019; Ren et al., 2018; Su et al., 2022; Yang & Oldenburg, 2012). Additional examples exist for environmental applications, such as those using frequency-domain systems (Liu et al., 2018; Zhang et al., 2021b), or layered geological contexts investigated with the HeliTEM multi-pulse system (Liu & Yin, 2016). However, based on a review of the available literature, no documented cases of 3D AEM modelling applied specifically to groundwater exploration in mountain regions have been found. In these settings, steep topography plays a significant role in shaping the EM signal, often making 1D modelling approaches inadequate to resolve the complexity of the subsurface.

Documented examples of integrating geophysical surveys and geological mapping into 3D, catchment-scale hydrogeological models remain limited. One notable case is presented by Vuilleumier et al. (2013), who combined AEM data acquired in Tulum area of Mexico's Yucatan Peninsula from Supper et al. (2009) with a stochastic karst system generator. The resulting network of conduits was incorporated into a finite element mesh flow model as pipe systems, allowing turbulent flow simulation. The approach relies on the expected electrical conductivity contrast between water-filled caves and the surrounding limestone, enabling identification of potential conduits. All simulated conduits were assumed fully saturated and the model was treated as a confined aquifer of constant thickness under steady-state conditions.

Bodin et al. (2022) used a combination of tracer tests, 3D seismic imaging, and borehole flow data for delineating karst conduit networks at the Hydrogeological Experimental Site (HES) in Poitiers, France. The targeted aquifer remains a confined and fully saturated system.

In karst environments, a variety of multidisciplinary approaches have been developed to address different scientific and practical objectives. For example, in the context of the mitigation of the sinkhole risk, several combinations of methods have been tested, often including geophysical techniques and hydrodynamic techniques (Jia et al., 2021; Kaufmann & Romanov, 2016). Multidisciplinary strategies have also proven effective in applied contexts, such as evaluating the feasibility of managed aquifer recharge (MAR) in karst settings (Patekar et al., 2022). In these studies, geophysical methods are employed to characterize subsurface geometry, while hydrochemical and hydrogeological analyses provide insights into the behavior and dynamics of groundwater systems.

1.3 The Multidisciplinary Approach

This study develops a multidisciplinary systematic approach to characterize the aquifer systems of mountain springs using 3D geological modelling, geophysical modelling and hydrogeological characterization.

Here's a breakdown of each step and the rationale behind it.

- I. Definition of the study area and data collection: the first phase of the multidisciplinary approach focuses on understanding the geological, geophysical, and hydrogeological characteristics of the study area. This involves: collecting and harmonizing geological data features, such as lithological features and structural elements; gathering geophysical data to provide spatial coverage of subsurface properties, particularly electrical resistivity; measuring hydrogeological parameters, including discharge rates, chemical and isotopic composition, to understand groundwater origin and the processes

it undergoes. Thus, this step includes the review of existing knowledge to provide background information on the study area. When necessary, it also embraces the collection of new, site-specific geological, geophysical, and hydrogeological data obtained through new, customized acquisition campaigns.

II. Integrated Geological and Geophysical modelling: this phase involves the integration of geological and geophysical models. This process follows an iterative workflow in which geological and geophysical interpretations continuously inform and refine each other, resulting in a more coherent and constrained representation of the subsurface.

a. Geological modelling: the process begins with the development of a geological model involving the synthesis of the collected geological and structural data. This includes 3D mapping of the distribution of rock units and major faults that define the architecture of the study area.

b. Geophysical modelling informed by geology: differences in geophysical parameters provide additional insights into groundwater preferential flow paths and constrain the geology at depth. However, geological units do not have a unique electrical signature, and geophysical data may suffer from equivalence problems. To overcome this challenge, geological data are incorporated into geophysical modelling to guide the inversion process. This is done by introducing constraint breaks at the interfaces between different geological units, as well as imposing to these units compatible resistivity ranges. Moreover, 3D modelling of AEM data is employed to honor the geological and structural complexity of mountain environments.

c. Geological and geophysical model update: the geological model is updated and refined based on the geophysical modelling results. Adjustments are made where inconsistencies arise, requiring reinterpretation of stratigraphic relationships, fault geometries, or the spatial extent of certain rock units; subsequently, new inversions are carried out with the updated constraints. This iterative process continues until a geologically and geophysically consistent model is achieved.

At this point, heterogeneities within geological units or at their interfaces are detected and may be a proxy for the presence of groundwater.

III. Hydrogeological characterization of the aquifer system: this phase aims to understand the hydrogeological behavior of the aquifer system, with a particular focus on identifying recharge areas, characterizing groundwater flow dynamics, and assessing

the potential influence of geological structures on groundwater circulation. A comprehensive investigation is carried out using a combination of hydrochemical, hydrographic, and isotopic analyses.

- IV. Development of the conceptual model of the hydrostructure. Here geology, geophysics and hydrogeology concur in defining the recharge areas, the water circulation and the aquifer structure. In particular: geological modelling defines the volume involved in groundwater circulation; geophysical modelling identifies the preferential flow paths, such as karst-fractured lithologies, and improves the 3D geological understanding; hydrogeology provides information for building the final conceptual model, testing and validating the insights derived from geology and geophysics, also through numerical modelling.

2. Development and validation sites

2.1 Challenges in spring characterizations in mountain areas

Characterizing springs in mountain environments is particularly challenging due to the interplay of multiple factors, including rugged topography, complex geological structures, limited data availability, and harsh climatic conditions. These elements complicate data collection and interpretation.

One of the main obstacles is accessibility. Many springs are located in remote and difficult-to-access areas, posing logistical constraints for field surveys, the installation and maintenance of monitoring equipment, and the collection of long-term data. Additionally, seasonal snow cover further complicates access during winter months, often restricting fieldwork and limiting the availability of observations.

Moreover, in these settings, the structural complexity of the subsurface is often controlled by fractures, faults, and, in some cases, karst systems. This heterogeneity makes it difficult to define recharge areas and flow dynamics.

In this study two examples of mountain environment, one in Italy and the other in Spain, were selected to test the multidisciplinary approach. The two study sites are described in the next sections.

2.2 Val Sabbia, Lombardy (northern Italy)

2.2.1 Physiographic framework

The first site analyzed in this research is the Mt. Ario – Mt. Pezzeda – Corna Blacca mountain group within the Brescian Prealps. It extends in a WSW–ENE direction along the watershed dividing Val Trompia and Val Sabbia (Fig. 2.1). The study area is defined by the following boundaries. To the north it is bordered by the upper Val Trompia. To the south it extends to the Fosso di Marmentino and the “Pertiche” area, approximately following the route of the Strada Provinciale 50 (Tavernole sul Mella/Nozza) and the Strada Provinciale 110 (Forno d’Ono/Livemmo). To the west the area is delimited by the Val Trompia segment between Tavernole sul Mella and Bovegno. To the east it is bounded by the hydrographic left side of the Abbioccolo River, in the homonymous valley. This area is known as the “Piccole Dolomiti Bresciane” because of the carbonate nature of the massif and its morphology, which bears similarities to the better-known neighboring Dolomites.

The topography of the region is dominated by imposing reliefs forming a roughly E-W oriented mountain range, dissected by deep valley incisions, with a mostly N-S and E-W trend. The

highest elevations are found at the Corna Blacca (2005 m), Mt. Pezzeda (1798 m), and Mt. Ario (1756 m) peaks. The major valleys include the E-W oriented Val Trompia (500-1000 m) and Valle d'Irma (500-900 m), as well as the N-S Valle Abbioccolo (400-800 m) and Val Degnone (450-700 m).

The study area is part of the Mella and Chiese river basins, with the Mella River delineating its northern and western boundaries. Although the Chiese River does not flow through the study area directly, its tributaries, including the streams Degnone, Glera, and Abbioccolo, play a significant role in shaping the hydrography. These streams, together with a network of smaller tributaries, form a dendritic drainage pattern, reflecting the structural and lithological characteristics of the region. They exhibit a mix of perennial and seasonal flow regimes.

Key hydrological features include springs, scattered and not well-developed cave systems of karst origin, and sinkholes.

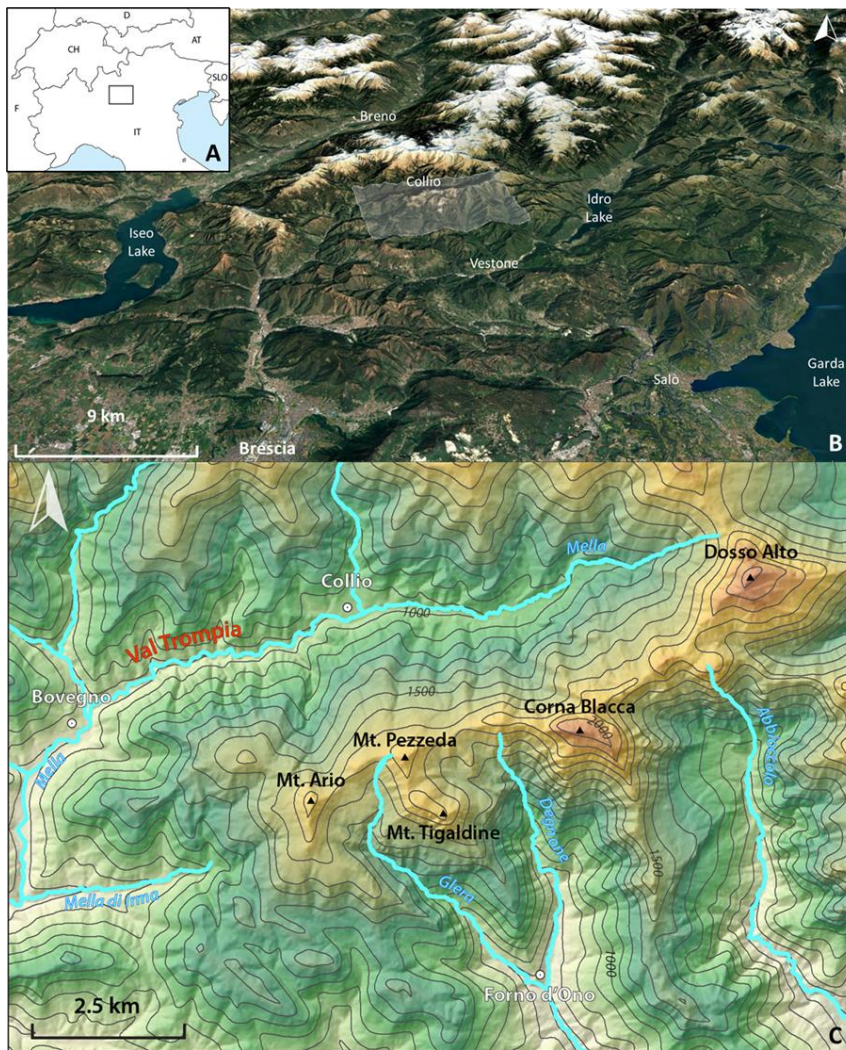


Fig. 2.1 Physiographic setting of the Val Sabbia study area. A: location of the study area in Northern Italy. B: the study area in the Brescian Prealps. C: The Mt. Ario – Mt. Pezzeda – Corna Blacca massif.

2.2.2 Geological and structural framework

The Mt. Ario – Mt. Pezzeda – Corna Blacca massif is part of the Brescian Prealps, a section of the Prealps located in the province of Brescia, Lombardy, in northern Italy. This massif belongs to the broader central Southern Alps domain, a south-verging fold-and-thrust frontal belt (Schönborn, 1992; Schumacher et al., 1995). To the north, it is bounded by the Insubric Line, while to the south, it is partially obscured by Plio-Pleistocene deposits of the Po Plain (Pieri & Groppi, 1981). To the east, the South Giudicarie Line defines the boundary between the central and eastern Southern Alps (Doglioni & Bosellini, 1987) (Fig. 2.2).

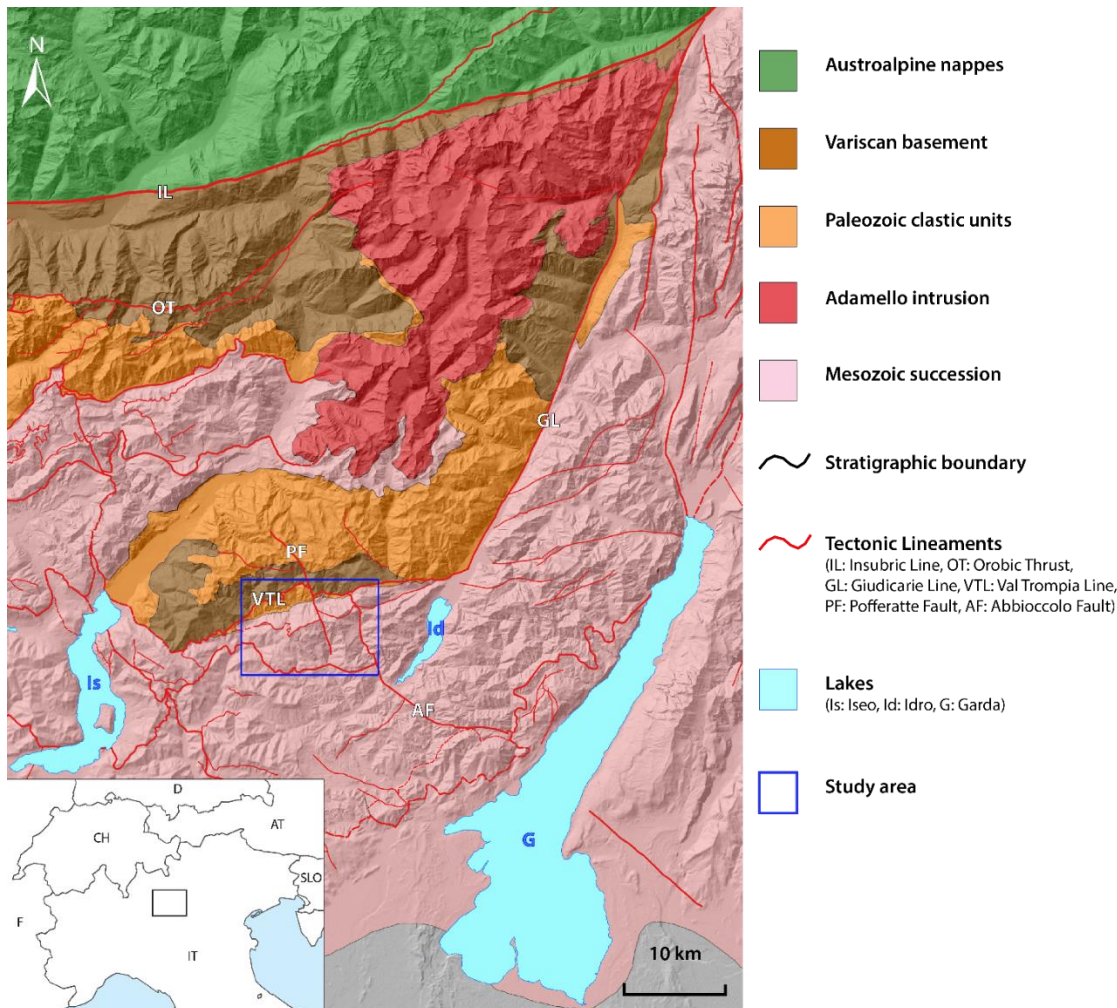


Fig. 2.2 Regional structural sketch of the study area (redrawn after Bigi et al. (1990) and Bosellini et al. (1999)).

The geological studies of the Brescian Prealps date back to the late 19th and early 20th centuries. The region has attracted significant interest due to the exposure of a complete succession, ranging from the crystalline basement of the Southern Alps to Quaternary deposits. The stratigraphy of the area was firstly described in the works of Curioni (1870), Bittner (1881)

and Cacciamali (1914) who recognized and described the Triassic terrains sequence. Moreover, this area sparked interest because of the southernmost outcrop of the Variscan metamorphic basement of the central Southern Alps, known as the Massiccio delle Tre Valli Bresciane (Boni, 1943). This basement is unconformably overlain by the Permian continental clastic rocks of the Verrucano Lombardo (Cassinis et al., 2008). The Permian sedimentary cycle ended in the Early Triassic with a widespread regression, leading to the deposition of the Carniola di Bovegno.

Stratigraphic reviews by De Sitter & De Sitter-Koomans (1949) in the Bergamasc Alps and Assereto & Casati (1965) in Val Camonica indicate that subtidal marine conditions were established in the Middle Triassic, resulting in the deposition of the Angolo Limestone. These conditions were later influenced by terrigenous inputs during the Upper Anisian, leading to the formation of the Prezzo Limestone. Marine sedimentation continued with substantial volcanoclastic contributions, culminating in the deposition of the Buchenstein Formation. During the Upper Ladinian, extensive carbonate platforms developed, represented by the Esino Limestone, which were occasionally interrupted by terrigenous basins, forming the Wengen Formation.

In the Carnian period, sedimentation was characterized by the expansion of the Val Sabbia Sandstone deltaic system, followed by the deposition of peritidal and evaporitic facies in an arid coastal environment, resulting in the San Giovanni Bianco Formation. The deposition of the Norian Dolomia Principale, which began in the Upper Carnian, is considered the earliest evidence of the rifting and spreading processes associated with the Jurassic Tethys Ocean (Jadoul et al., 1992).

The schematic Lower to Upper Triassic stratigraphic succession of the study area is depicted in Fig. 2.3.

The tectonic setting of the Brescian Prealps can be described as a south-verging stacking of tectonic units thrusting along east-west tectonic lineaments (Fig. 2.4). Suess (1869) was the first to identify the presence of a major east-west oriented, north-dipping tectonic structure, that he called the Val Trompia Line. This line separates the schists of the crystalline basement to the north from the Permo-Triassic outcrops to the south. Further investigations by Bittner (1881) revealed a tectonic framework south of the Val Trompia Line characterized predominantly by folding, with the northernmost folds partially thrust.

South of the Val Trompia Line, Tilmann (1907) identified a tectonic lineament east-west oriented, now known as Mt. Ario Line, that he considered as the contact of the southern limb of an Anisian syncline with the southern outcrop of the Middle Triassic sequence. Cacciamali

(1914) interpreted the same structure as marking a southward overturned anticline. Cozzaglio (1923) supported Cacciamali's hypothesis, while Boni (1943), in relation to this lineament, proposed the thrust of two rigid plates. According to more recent interpretations (Boni & Cassinis, 1973; Cassinis et al., 2011), the Mt. Ario Line is considered as an east-west trending subhorizontal thrust, along which the Anisian Angolo Limestone overlies the more recent Triassic units. In a couple of places, between the Angolo Limestone and the underlying Dolomia Principale, there are thrust splays: smaller, subsidiary thrusts that branches off the Mt. Ario Line. One of these comprises a patch of the Val Sabbia Sandstone and is located 200 m south of the summit of Mt. Ario. The other comprises the overturned limb of a syncline in which Esino Limestone, Wengen and Buchenstein Formation crop out, and is located southeast of Mt. Pezzeda, between this and Mt. Frondine, at the head of Valle Glera (Boni, 1943). These thrust splays are remnants of the primitive structure from which the thrust is derived, dragged by the overriding mass.

Further to the south another east-west tectonic lineament crops out. It is named Ombriano-Forno d'Ono Line. According to Tilmann (1907), it represents a subvertical north-dipping fault. In contrast, Cacciamali (1914) interpreted it as subhorizontal thrust. According to more recent interpretations (Bianchi et al., 1971; Boni & Cassinis, 1973), this is a subhorizontal tectonic line that brought the northern Angolo Limestone into contact with the southern Val Sabbia Sandstone.

In the study area, the main tectonic lineaments are displaced by several mainly northwest-southeast trending faults. The most important one is the Pofferatte fault, that causes a lowering of the north-east unit (Bianchi et al., 1971).

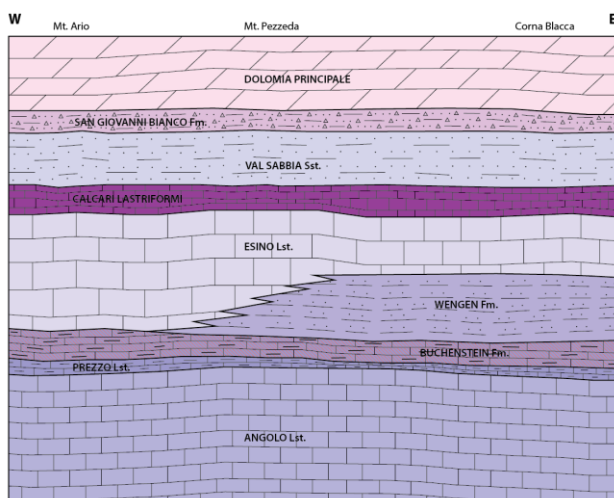


Fig. 2.3 Stratigraphic sketch of the Triassic succession outcropping in the study area (redrawn after Borrromeo (1982)).

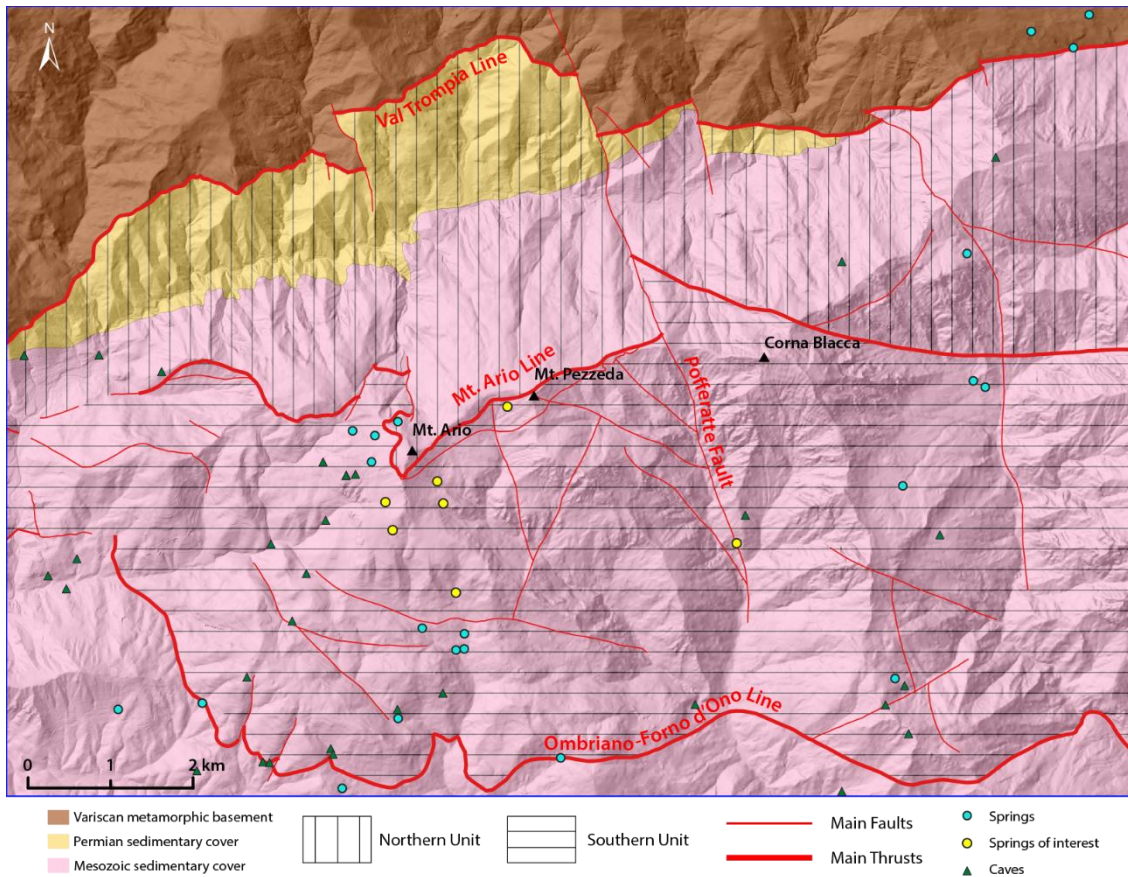


Fig. 2.4 Structural sketch of the study area (Boni & Cassinis, 1973; Servizio Geologico d'Italia, 2011, 2012, 2016) highlighting key features such as springs and caves.

2.2.3 Hydrostratigraphic setting

The Southern Alps domain hosts important groundwater reservoirs, mainly located within the wide and thick Mesozoic sedimentary cover. Here, the carbonate rock outcrops have facilitated the formation of significant karst features (Tognini, 2011).

The identification and characterization of the groundwater bodies in the mountain area of Lombardy has been carried out by Masetti et al. (2021) and Stevenazzi et al. (2023). These works introduce a 3D hydrostratigraphic/hydrostructural model that takes into account the lithological, geometrical, and geomorphological components of the hydrostratigraphic heterogeneity.

Masetti et al. (2021) identified the Brescian Prealps as part of the “Tre Valli Bresciane metamorphic basement and carbonates” groundwater body. This area includes systems associated with both the metamorphic basement of the Southern Alps and the overlying Permo-Triassic carbonate and terrigenous sedimentary sequences.

The hydrostratigraphic classification proposed at a regional scale by Stevenazzi et al. (2023), defines the dominant flow characteristic of each bedrock unit in the Lombardy Alps. According to this classification, the formations outcropping in the study area primarily exhibit karst behavior.

The main aquifer systems in the region include:

- Lower-Middle Triassic carbonates of the Angolo and Prezzo Limestone, where groundwater circulation is classified as karst dispersive/impervious. This is due to the lateral and vertical alternation of karst rocks (mainly limestone and dolostone) with low-permeability rocks (marlstone and shale).
- Middle Triassic massive and layered carbonates of the Esino Limestone and Calcari Lastriformi, which were classified as karst drainage/karst dispersive. These formations function as a major aquifer due to their high karstification potential (Tognini, 2011).
- Massive dolomitic succession of the Dolomia Principale, where the dominant flow type is karst dispersive/karst drainage. It acts as a major aquifer locally, benefiting from extensive fracturing and dissolution features.

In contrast, several aquitards and aquicludes limit groundwater flow in the region, including:

- Middle Triassic terrigenous-carbonate successions of the Buchenstein and Wengen Formation, where groundwater flow is classified as impervious/karst dispersive due to the alternation of shale and carbonate lithotypes.
- Upper Triassic terrigenous and pelitic-evaporitic successions of Val Sabbia Sandstone and San Giovanni Bianco Formation, which exhibit impervious/fissured porous flow characteristics.

Additionally, the region's complex tectonostratigraphic framework plays a crucial role in shaping its hydrogeological dynamics. The presence of tectonic lineaments and faults strongly influences groundwater flow patterns and storage capacity across the area.

The rugged environment that characterizes the study area makes it difficult to locate caves or to access possible endokarst systems. However, the Lombardy Speleological Federation (FSLo) has identified 32 karst cavities within the study area, all of which are recorded in the Lombardy Speleological Register as part of the [Tu.Pa.Ca.](#) project.

These are relatively small cavities, generally with a maximum depth or length of around 10 m. Most of them are shallow caves or recesses located at the edge of the study area, historically

used as shelters. Despite their limited extent, these features offer precious information into the karst processes affecting the area and contribute to the understanding of local hydrogeology. Among them, three caves stand out due to their geological significance, and they are specifically mentioned in the descriptions of the formations where they are found in Section 4.1.3.

The presence of caves is a clear indication of karstification; however, the opposite is not necessarily true. The lack (or non-discovery) of accessible caves does not imply the absence of hydrologically significant karst systems. Even if no humanly reachable passages are found, smaller conduits may still exist, capable of transporting significant volumes of water at high velocities.

2.2.4 Description of the springs

In the study area, groundwater flow is evidenced by the presence of numerous springs (at least 41), which serve as key indicators of the hydrogeological dynamics. These springs emerge at various elevations and locations mainly throughout the Southern Unit (Fig. 2.4), reflecting the complex subsurface flow paths influenced by geological formations, fault and fracture networks, and karst features.

Due to the natural abundance and high quality of the groundwater, A2A Ciclo Idrico, the local water network manager, utilizes these springs as a primary water source for domestic supply. The water is directly collected, treated, and distributed to meet the needs of the local population.

The distribution of the springs across the area highlights the hydrogeological heterogeneity of the area, with different recharge and discharge zones influencing their flow rates and seasonal variability. Some springs exhibit constant and significant discharge, while others show more variable flow.

The presence of these springs not only plays a crucial role in water resource management but also provides information for understanding the hydrogeological setting and aquifer properties within the study area.

Of all the springs issuing in the study area, five have been identified as being of particular interest, four of which are located around Mt. Ario (Fig. 2.5). The fifth, Acqua Bianca, is the

largest and is located at the foot of the Corna Blacca massif. These springs are described in more detail below.

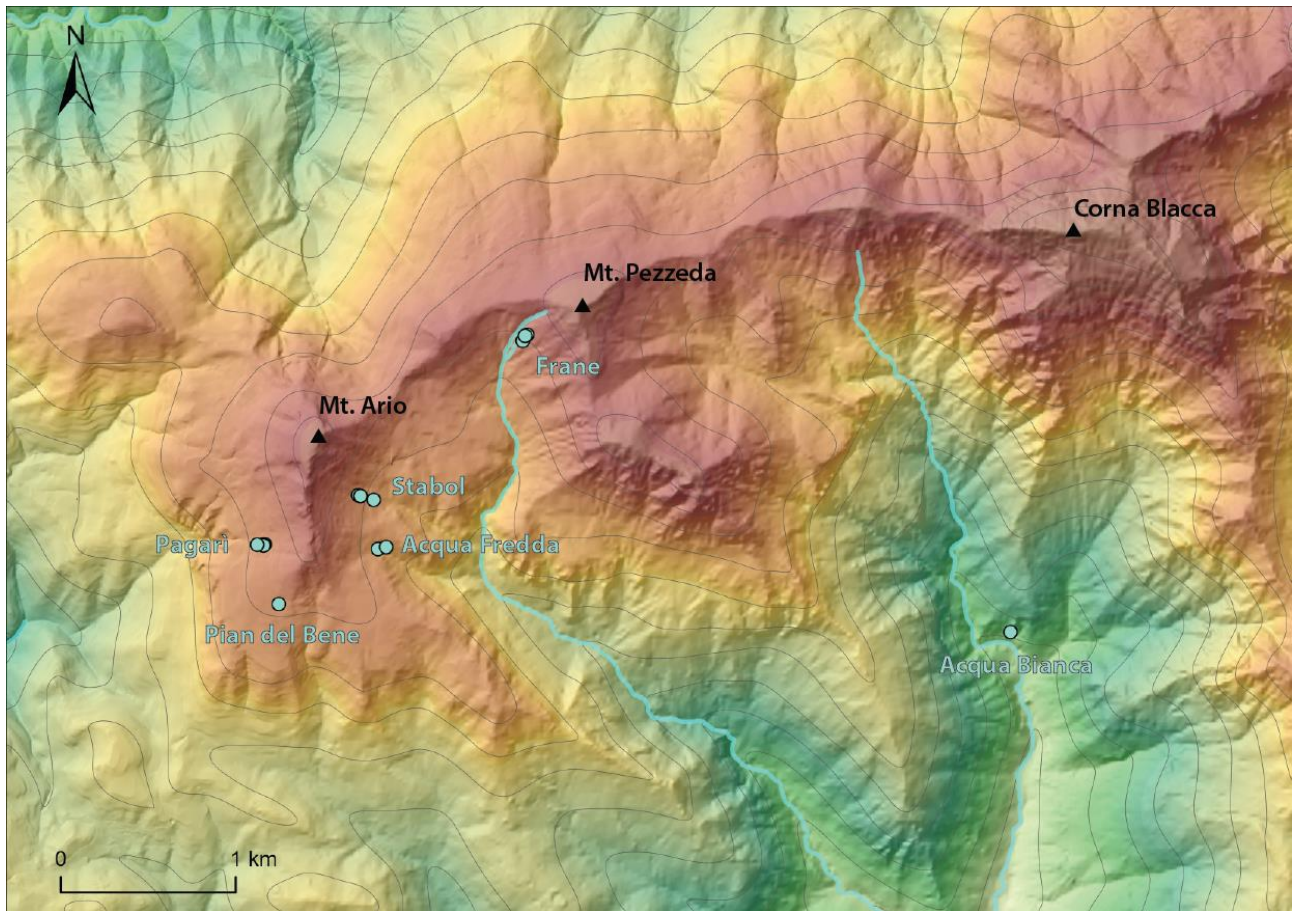


Fig. 2.5 Location of the springs of interest in the study area.

Pian del Bene – Acqua Fredda 1-2 – Stabol 1-3 springs

The Pian del Bene (1509 m), Acqua Fredda (1447 m), and Stabol (1510 m) are three groups of springs located on the southern and eastern side of Mt. Ario (1755 m), and connected with the Pertica Alta (BS) municipal aqueduct (Fig. 2.6).

The water of these springs is collected in a concrete cistern, named Vasca B9, with a Cipolletti thin-plate weir and equipped with Level TROLL 500 sensor pair with VuLink® data logger device (In-Situ) for the continuous measurement of water level, water pressure and temperature.



Fig. 2.6 A: Vasca B9 collector. B: Stabol 3 spring. C: Pian del Bene spring.

Pagari 1-4 springs

The Pagari springs consists of 4 neighboring springs located at an elevation of 1479 m a.s.l. on the western slope of Mt. Ario (1755 m), in the municipality of Marmentino (BS). The spring collector consists of a concrete tank with a triangular thin-plate (Fig. 2.7).

Pagari collector is equipped with Level TROLL 500 sensor pair with VuLink® data logger device (In-Situ) for continuous measurement of water level, water pressure and temperature. The water from these springs is also channelled to Vasca B9.



Fig. 2.7 Pagari 4 spring.

Acqua della Frana 1-2-3 springs

Acqua della Frana (from here on only Frane) refers to a group of three adjacent springs located at the head of the Valle Glera, downstream of Mt. Pezzeda (1798 m) on its southwestern side, at an elevation of 1601 m, in the municipality of Pertica Alta (BS). The water from the three springs is collected in a concrete tank, named Vasca B5, with a triangular thin-plate weir (Fig. 2.8). The spring's discharge ranges from approximately $0.001 \text{ m}^3/\text{s}$ during base flow conditions, reaching up to $0.01 \text{ m}^3/\text{s}$ following intense rainfall events.

The springs' collector is equipped with Aqua TROLL 500 multiparameter sonde pair with VuLink® data logger device (In-Situ). Measured parameters include water level, temperature, pressure, conductivity, pH, Rugged Dissolved Oxygen (RDO®).



Fig. 2.8 Vasca B5, collector of the Frane 1-3 springs.

Acqua Bianca spring

Acqua Bianca is one of the most productive springs in the province of Brescia, with an impressive waterfall about 15 m high. It is located at an altitude of 748 m in the hydrographic left side of the Degnone stream, just above Forno d'Ono, one of the hamlets of the Pertica Bassa municipality (BS). The spring's discharge ranges from approximately $0.03 \text{ m}^3/\text{s}$ during base flow conditions, reaching up to $0.8 \text{ m}^3/\text{s}$ following intense rainfall events.

The geometry of the intake structure is complex (Fig. 2.9): there are two concrete tanks connected, each equipped with a thin-plate weir. The spring is provided of Aqua TROLL 500 multiparameter sonde, and Level TROLL 500 sensor pair with VuLink® data logger device (In-Situ). Measured parameters include water level, temperature, pressure, conductivity, pH, Rugged Dissolved Oxygen (RDO®). The results of the monitoring are presented in Section 4.2.1.



Fig. 2.9 A: Acqua Bianca water collection tank. B-C: water overflow from the rock layers. D: top-down view of the Acqua Bianca waterfall.

2.3 Aracena, Andalusia (southern Spain)

Since the AEM dataset acquired in Val Sabbia was not made available, an alternative dataset was used to assess the impact AEM technique on groundwater exploration. To ensure the validity of the analysis, the selected reference dataset pertain to a site with comparable geological and hydrogeological characteristics to those of the study area in Val Sabbia, as far as possible.

The reference site is located in Andalusia, a region in southwestern Spain, more precisely in the Sierra de Aracena, where complicated structural conditions and topographic roughness provide a suitable analogue for Val Sabbia. In particular, groundwater flow is evident in this area, as demonstrated by the presence of numerous springs, which indicate active underground water circulation. Furthermore, the confirmed presence of a cave provides clear evidence of karst systems within the carbonate rock formations exposed in the area. This cave suggests significant secondary permeability due to fracturing and dissolution processes, which likely influence groundwater flow.

The AEM data acquired in this area provides information for the characterization of the groundwater resource, karst features, and hydrogeological connectivity.

Through the use of this data set, the study will demonstrate how airborne electromagnetic methods can improve subsurface characterization.

2.3.1 Physiographic framework

The Sierra de Aracena is a mountain range located in the province of Huelva, in southwestern Spain (Fig. 2.10). It is part of the Sierra Morena, a mountain range that forms the natural boundary between the Guadalquivir basin to the south and the Meseta Central plateau to the north.

The geomorphology of the Sierra de Aracena is characterized by rolling hills with moderate slopes and small valleys, dominated by Mediterranean forest. Elevations typically range between 600 and 900 m, with Cerro del Castaño (962 m) representing the highest peak of the range. The climate is Mediterranean with Atlantic influences, characterized by mild and humid winters and hot, dry summers, with precipitation mainly concentrated in the autumn and winter months.

The study area is crossed by a well-developed hydrographic network, with marked lithological and structural control over surface drainage. The main axes of the river network exhibit a preferential orientation from NW to SE and from E to W.

The outcrop of carbonate lithologies, particularly limestones and marbles, facilitates the development of karst morphologies. In the Sierra de Aracena, exokarst landforms are widely represented, with the presence of karren fields, dolines, and blind valleys. Simultaneously, the development of endokarst manifests through the presence of complex cave systems, such as the renowned Gruta de las Maravillas, situated in the town, with an estimated volume of 70,000 m³ (Martínez-Moreno et al., 2016). This cave, characterized by an extensive network of galleries, chambers, and underground lakes, in which stalactites and stalagmites are present, represents an emblematic example of hypogene karst in the region.

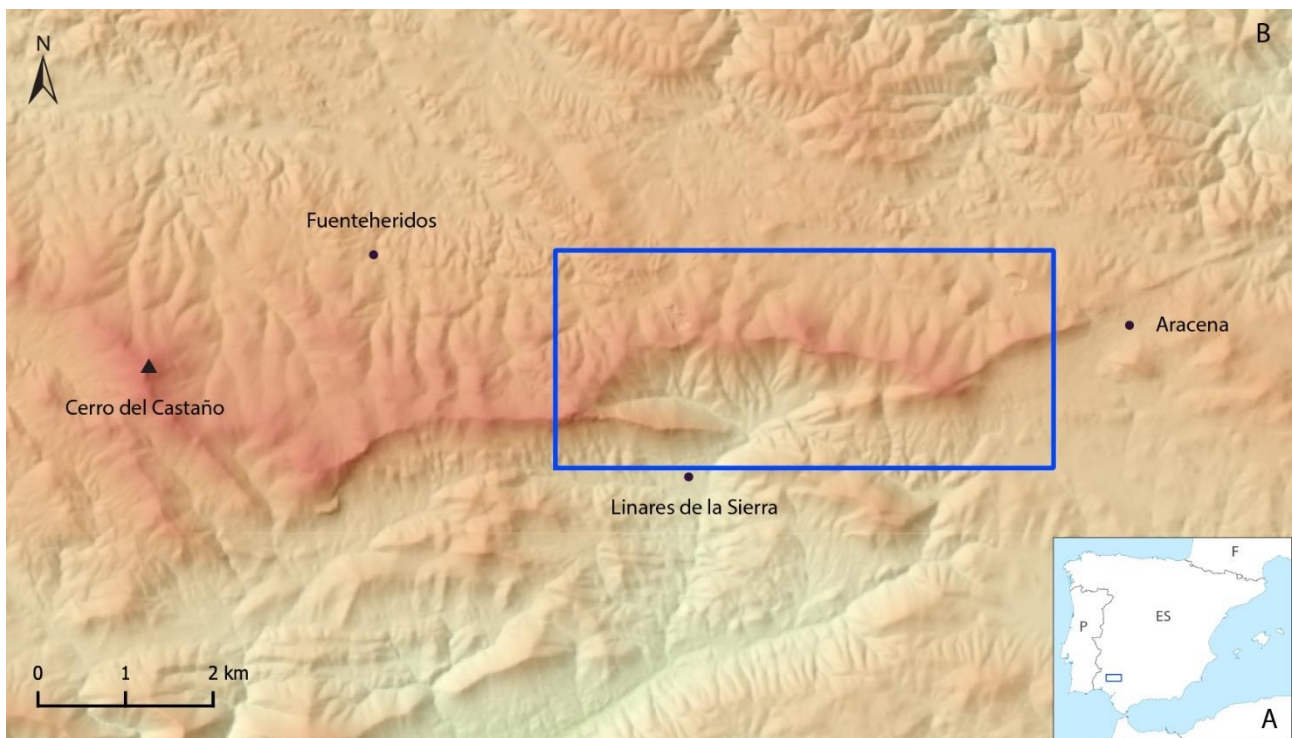


Fig. 2.10 A: the location of the study site in the Iberian Peninsula. B: Aracena physiographic setting.

2.3.2 Geological and structural framework

The study area is located at the boundary between the Ossa-Morena and South Portuguese Zones (OMZ and SPZ respectively), two of the main domains of the Iberian Massif (SW Spain). Specifically, the area lies within the Aracena Metamorphic Belt (AMB), a regionally significant structure characterized by high-grade Variscan metamorphism (Díaz-Azpiroz & Fernández, 2021). The boundaries of the AMB are aligned parallel to the main regional structural trends, highlighting tectonic control over its formation.

The La Umbría Formation is primarily composed of graphitic phyllites, interbedded with black quartzites. The Precambrian age of this formation makes it the oldest in the area. The exact

thickness of the formation remains uncertain due to structural complexity and a lack of continuous outcrops.

The Aracena Formation begins with basal conglomerate layers, which rapidly transition into dolomitic marbles and, to a lesser extent, limestones. This lithological transition indicates a change in sedimentation conditions, shifting from a clastic to a carbonate environment. The presence of dolomitic marbles highlights post-depositional dolomitization processes.

At the top of the Aracena Formation the Volcanosedimentaria Formation crops out. This Precambrian formation is characterized by an alternation of acidic volcanic tuffs and micaceous rocks, with interbedded carbonate layers. These layers exhibit similarities to the dolomitic marbles of the Aracena Formation, suggesting a continuity in carbonate sedimentation processes. Towards the upper part of the formation, marmoreal limestones, sometimes dolomitized, are found, indicating variations in the chemical composition of the carbonate rocks. The ensemble of carbonate layers intercalated between the metavolcanics can exceed 300 m in thickness.

The structure of the Sierra the Aracena range is characterized by the NW-SE oriented SW vergent Fuenteheridos – La Umbría anticlinal overturned fold (Fig. 2.11). At the core of this anticlinal the schists of La Umbría formations crops out.

The geometric continuity of the outcrops and the fold is broken by displacement along two distinct sets of steeply dipping faults. The first set consists of sinistral faults oriented approximately E-W, which are subsequently intersected and offset by the second set. This second fault system is characterized by dextral faults oriented in a NE-SW direction. The interaction between these fault sets creates a complex structural framework, influencing the deformation patterns and spatial distribution of geological units within the study area.

2.3.3 Hydrostratigraphic setting

The study area is characterized by a natural groundwater drainage that produces a large number of springs.

The most promising aquifer-forming formations within the study area, based on their hydrogeological properties and spatial extent, are the carbonates of the Aracena Formation (Martos Rosillo et al., 2008). These consist primarily of dolomitic marbles, with minor Cambrian limestones, which are permeable due to fracturing and/or karstification. Their thickness generally ranges between 250 and 400 m, though in some areas, it can exceed 550 m.

These carbonate rocks rest upon the metapelites of the La Umbría Formation, which serve as a low-permeability substrate (Martos Rosillo et al., 2008). Towards the upper part of the

sequence, there is the alternation of tuffs, porphyries, and micaschists of the Volcanosedimentaria Formation, which behave as aquitards. These layers are interspersed with marbles and limestones, creating a complex hydrogeological setting.

The study area is located in the central sector of the “Sierra de Aracena” aquifer, where the carbonate aquifer materials, primarily from the Aracena Formation, distributed across two geometrically and structurally disconnected outcrops. The northern outcrop consists of the northern limb of the Fuenteheridos – La Umbría anticline, which is repeated due to the effect of a second-phase fold. The southern outcrop is composed of exposures from the southern limb of the anticline. Recharge occurs through the infiltration of rainwater, and in the northern sector the groundwater flows northward toward the Fuente del Castaño spring (FC in Fig. 2.11), while in the southern area, the flow moves southward toward the La Herrería spring (H in Fig. 2.11).

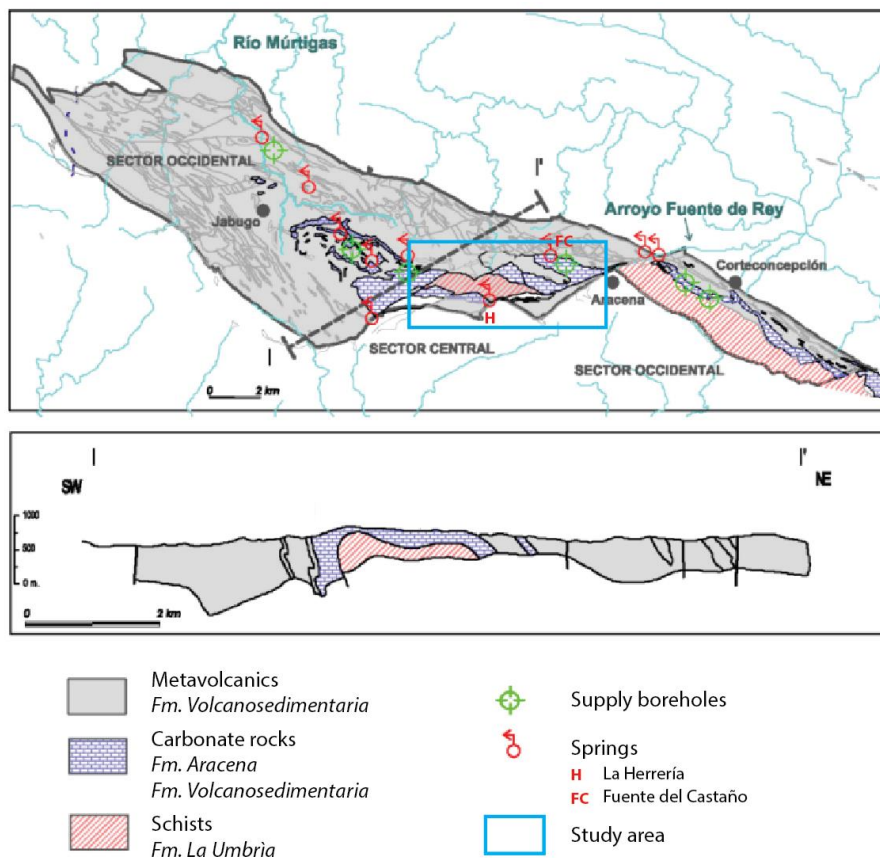


Fig. 2.11 Hydrogeological map and cross section of the carbonate aquifer of the Sierra de Aracena. Redrawn after Martos Rosillo et al. (2008).

3. Methods

3.1 3D geological modelling

3.1.1 Objective and brief review of the 3D geomodelling

Geomodelling consists of a range of mathematical techniques used to create unified and comprehensive representations of geological objects (Mallet, 2002). These models allow spatial visualization and geological interpretation of inaccessible geological conditions from limited information, while systematically incorporating all the associated data to ensure accuracy and consistency (Houlding, 1994). The process often integrates multiple datasets, including field observations, geological and structural maps, cross sections, and geophysical surveys, which are combined to build detailed and realistic representations of the subsurface architecture.

The primary objective of the 3D geomodelling process is to create a geometrically accurate representation of the geological setting within the area of interest. This involves defining stratigraphic boundaries and fault networks, as well as understanding their spatial relationships.

The widespread use of 3D geomodelling in many branches of geology is due to a very common question among geologists: with little scattered data, how can the geometry of geological bodies be interpreted in 3D?

In hydrogeology, 3D geological models have supported groundwater resource estimation for various purposes, such as geothermal energy exploitation and CO₂ storage (Mas et al., 2022). Furthermore, they serve also as a foundation for downstream analyses by enabling the assessment of spatial variability in subsurface geological properties. This is crucial for a range of applications, including groundwater flow simulations (Burs et al., 2016; Hassen et al., 2016). By providing a detailed, continuous and homogeneous representation of the subsurface geology, 3D geomodelling not only aids scientific understanding but also supports practical decision-making.

Different geomodelling methods are implemented in commercial software packages, each offering unique strengths tailored to specific industries or modelling challenges.

One commonly employed method is the explicit modelling, which involves manually delineating geological interfaces using a series of bounding data points (x, y, z) (Ross et al., 2021). The input data points must clearly relate to specific geological interfaces to guide the interpolation process. Interpolation techniques are then used to "fill in" the surface between the known data points, creating a continuous boundary. This process relies heavily on expert interpretation and knowledge, offering precise control over the geometry of the resulting

model. However, it is often labor-intensive and subjective. Moreover, because each surface is constructed independently, ensuring consistency between intersecting surfaces (faults and stratigraphic layers) can be challenging without additional constraints or validation. Explicit modelling is best suited for cases where the available data are detailed and localized, or where expert interpretation is critical to accurately represent complex geological features. The explicit approach is implemented in computer program like GOCAD (Mira GEOSCIENCE).

In contrast, implicit modelling relies on mathematical algorithms to generate geological interfaces by defining them as equipotential contours (iso-surfaces) of a 3D scalar field. This scalar field is generated from scattered data points and takes into account both position and orientation data to represent the subsurface geometry (Lajaunie et al., 1997; McInerney et al., 2005). The process begins with the estimation of a scalar field, which is computed based on input data points known to lie on one or more geological interfaces, as well as additional orientation data that may not necessarily lie directly on the interfaces but provide critical directional constraints for the model (Lajaunie et al., 1997). The scalar field is interpolated across the model domain ensuring that points on the same interface share the same scalar value, while gradients in the scalar field align orthogonally with the orientation data. The resulting scalar field allows for the automatic extraction of geological interfaces as iso-surfaces of constant scalar value using algorithms like "marching cubes" (Lorensen & Cline, 1987). This approach is particularly well-suited for modelling geological complexity, such as faulted and folded terrains, and is highly efficient in managing large and diverse datasets. The software GeoModeller follows such an approach.

The choice of the geomodelling method requires careful consideration of the geological features to be modelled, the data availability and its resolution, as well as the user requirements.

In the context of this project, implicit geomodelling's ability to efficiently generate continuous and consistent representations of subsurface geology, even with sparse data, has made it the preferred method. Moreover, GeoModeller is the ideal software solution for managing the types of data gathered and used in a geologically complex and deformed terrains, such as the study areas.

3.1.2 GeoModeller modelling approach

GeoModeller is an implicit 3D geomodelling software suite developed by BRGM (Bureau de Recherches Géologiques et Minières) and subsequently commercialized by Intrepid Geophysics (Melbourne, Australia).

The 3D interpolator approach employed by GeoModeller is based on potential field method (Calcagno et al., 2008; Chilès et al., 2004; Lajaunie et al., 1997; McNerney et al., 2005).

The general principles on which the potential field method relies are: i) a geological interface serves as the boundary between two distinct geological formations, which may include stratigraphic boundaries, tectonic boundaries, or fault surfaces; ii) orientation data sampled within geological formations, including polarity information, are relevant to model the interfaces separating the formations; iii) the geological interfaces being modelled can be regarded as belonging to an infinite set of surfaces aligned with the orientation field (Calcagno et al., 2008). This method can thus be used when contact data points are available on a geological interface, along with additional orientation data that represent the gradient or derivative of the potential field (Lajaunie et al., 1997). Using co-kriging for interpolation, the method generates a scalar potential field. This field is used to define the geological interfaces as iso-surfaces that align with and honor all available data (Chilès et al., 2004).

In GeoModeller, interfaces of geological formations with sub-parallel behaviors are grouped into *series*¹, allowing a single mathematical interpolation solution to model the entire series (Calcagno et al., 2008; McNerney et al., 2005).

GeoModeller works according to two fundamental rules that govern the construction of geological models. First, it requires a chronologically ordered sequence of geological *formations*², the so-called *geological pile* (Calcagno et al., 2008). This means that all formations included in the model must be organized in a stratigraphic-like sequence that reflects the temporal order in which the formations are processed within the numerical calculation. The boundary between two formations is assigned to the top or the bottom of one of the two formations. Second, the nature of the contact between these formations must be defined to ensure the model reflects the geological processes that have shaped the region. The relationships between formations fall into two main categories, which determine how the formations interact within the geological pile:

- *Erode* relationship: this type of relationship allows younger formations to truncate older ones. In geological terms, this represents erosion processes where a younger layer cuts into and removes parts of the older formations beneath it. This is particularly important for representing unconformities, erosional or tectonic surfaces in the model.

¹ In GeoModeller, the term *series* is used in its conventional English sense, meaning a *sequence* or *set*, and should not be mistaken for its chronostratigraphic definition.

² The term *formation* in GeoModeller refers broadly to any geological unit included in the 3D model, without necessarily adhering to formal stratigraphic classifications.

- *Onlap* relationship: this relationship enables younger formations to fill the available space without altering the geometry of the older ones. Essentially, the younger formations are deposited on top of the existing layers, conforming to the pre-existing structure without significant deformation.

By combining these rules, GeoModeller ensures that the geological model reflects the stratigraphic and structural setting observed in the field.

Handling faults is crucial for constructing a reliable 3D geological model. In GeoModeller faults are treated as geological interfaces and modelled using a potential field, where fault traces serve as contact data points, while fault orientation data define the gradient. Each fault is represented by its own potential field. By convention, the hanging wall and footwall are assigned based on the orientation of the gradient data. Faults can be classified as either infinite, extending across the whole study area, or finite, with a limited spatial influence. For finite faults, their extent must be explicitly defined. The fault extent is determined by the intersection of the potential field iso-surface with a sphere or ellipsoid of a specified radius, centered on the data points that define the fault (Calcagno et al., 2008; Chilès et al., 2004).

Since faults typically affect only specific portions of the stratigraphic succession, they must be explicitly associated with geological formations within the geological pile. Additionally, at least one 3D point must be defined for each geological interface within every fault block. Furthermore, the software requires users to specify the relationships between different faults in the model to ensure accurate structural representation (McInerney et al., 2005).

Despite these capabilities, the accurate computation of discontinuities such as faults and unconformities remains a challenge, requiring careful parameterization and many trial-and-error tests.

3.1.3 3D model export

In GeoModeller it is possible to export the modelling products in multiple formats, depending on the application they are intended for. This option allows for the interoperability of GeoModeller with GIS, CAD, numerical modelling software like FEFLOW, and 3D visualization software like Paraview.

For the 3D geological models, the available export formats are DXF (AutoCAD), TSK (GoCAD), VTK (Paraview). In addition, the 3D geological interfaces can be exported as DXF, TSK, Tsurf, VTP, VTK, GRID (raster formats readable with GIS software).

2D geological cross sections and profiles can be saved as DXF and raster images (PNG, TIFF, JPG, PDF).

3.2 Hydrogeological analyses

3.2.1 Objectives

The characterization of groundwater bodies is a fundamental step in understanding the hydrogeological framework of the study area. This process requires the integration of multiple methods to delineate recharge mechanisms, flow pathways, and the properties of groundwater systems.

Groundwater retains a record of its origin and the path it has travelled before emerging at a spring. This "memory" is preserved in its chemical and isotopic composition, offering information on water's history and the geological formations it has encountered along the way. Chemical composition reflects the minerals dissolved during groundwater flow, influenced by soil-water exchange, rock-water interactions, and potential anthropogenic inputs.

Isotopic composition provides insights into the climatic conditions at the time of infiltration, the altitude of recharge areas, and potential mixing with other water sources.

By analyzing these chemical and isotopic signatures, it is possible to reconstruct groundwater's paths and understand aquifer dynamics, making this information essential for groundwater resource management.

In addition to geochemical methods, spring hydrographs, which represent the temporal variations in groundwater spring discharge, are key tools for assessing aquifer dynamics.

By employing this multidisciplinary approach, it is possible to identify and characterize the aquifer system(s) feeding the springs of interest and assess their role in sustaining spring discharge. The results contribute to a general understanding of the groundwater circulation in the study area.

3.2.2 Hydrochemical features

The chemical quality of spring waters is determined by natural factors and processes as well as human activities. Natural factors include the lithology of the geological formations through which the groundwater flows, mineral dissolution and precipitation, ion exchange, and redox reactions. Human activities can alter the natural hydrochemical composition of groundwater introducing additional contaminants. In this case, since the study area is a sparsely populated mountain location, the only potential anthropic pressure on the quality of the spring comes from seasonal grazing during the summer months, which can be considered negligible.

The major elements analysis determines the hydrochemical signature of groundwater, which represents its unique chemical composition. This signature provides information on the interactions between groundwater and the surrounding geological formations (Freeze &

Cherry, 1979), preserving a "memory" of its flow path and the geochemical processes occurring along the way, helping to understand the hydrogeological behavior of the aquifer system. As groundwater moves through different rock types, it undergoes various geochemical processes, leading to characteristic ion concentrations. This chemical fingerprint aids in tracing groundwater origin, flow dynamics, and aquifer behavior.

The common inorganic solutes in groundwater are calcium (Ca^{2+}), Magnesium (Mg^{2+}), Sodium (Na^+) and Potassium (K^+) among cations; and Bicarbonate (HCO_3^-), Chloride (Cl^-), Sulphate (SO_4^{2-}), Carbonate (CO_3^{2-}) and Fluoride (F^-) among anions (Fitts, 2002).

In addition to chemical composition, the measurement of physical parameters such as electrical conductivity (EC), pH, and temperature (T, expressed in $^{\circ}\text{C}$) provides information on groundwater properties and quality.

Electrical conductivity (expressed in $\mu\text{S}/\text{cm}$) is the ability of a given water to conduct electricity and is directly proportional to the amount of total dissolved ion concentration. The higher the conductivity, the more ions are present, the more mineralized is the water (Younger, 2008). Since it is a temperature-dependent parameter, the measurements are typically corrected to the equivalent value at 25°C to ensure comparability across different sampling conditions.

pH measures the acidity or alkalinity of groundwater, which is largely influenced by the lithology of the aquifer system. Carbonate formations typically lead to alkaline groundwater due to the dissolution of calcium carbonate (CaCO_3), which buffers acidity and increases pH.

Temperature (T, expressed in $^{\circ}\text{C}$) is a key parameter as it provides insights into groundwater circulation depth and potential mixing with deeper or geothermal waters. Shallow groundwater tends to reflect seasonal temperature variations, while deeper aquifers typically maintain more stable thermal conditions.

Graphical methods of data presentation are useful for quickly assessing the results of major ion analyses and identifying overall trends. Here the Piper diagram is used for visualizing the relationships between dominant cations and anions, because it provides the classification and comparison of different water samples based on their ionic composition (Piper, 1944). It consists of two ternary plots at the base, representing the percentage of total milliequivalents per liter (meq/L) of cations in the lower left (with Ca^{2+} , Mg^{2+} , and $\text{Na}^+ + \text{K}^+$ as the three axes) and of anions in the lower right (with $\text{HCO}_3^- + \text{CO}_3^{2-}$, SO_4^{2-} , and Cl^- as three axis). The central diamond integrates projections from both the cation and anion triangles, providing a simultaneous representation of the major ion composition. The position of a water sample within this diagram determines its hydrochemical facies. The hydrochemical facies represent

zones within a groundwater system with different concentrations of ions that are related to the hydrodynamic, hydrochemical and geochemical mechanisms of water-rock interaction, and are therefore conditioned by the type of rock volume through which the water flows (Younger, 2008).

3.2.3 Environmental isotopes analysis

Environmental isotopes are naturally occurring isotopes, both stable and radioactive, that serve as tracers in Earth and environmental sciences (Clark & Fritz, 1997). Among them the most studied are those of hydrogen and oxygen, the fundamental elements of water. The stable isotopes of these elements do not undergo radioactive decay and are integral constituents of water molecules, making them ideal tracers for hydrological investigations (Kendall & McDonnell, 1998).

Meteoric processes modify the stable isotopic composition of water; hence, the recharge waters in a particular environment have a characteristic isotopic signature.

In temperate and humid climates, once rainfall water infiltrates the subsurface, the stable isotopes cease fractionation and preserve the isotopic signature of their source. By analyzing these isotopic variations, it is possible to identify the origin of the groundwater and estimating the approximate elevation of the primary recharge areas of the aquifer system (Longinelli & Selmo, 2003, 2010). The isotopic signature serves as a natural and conservative tracer of waters. Springs primarily fed by direct precipitation will reflect the isotopic composition of local meteoric water.

Hydrogen is omnipresent in terrestrial environments, it has two naturally occurring stable isotopes: ^1H or just H (99.9885%), and ^2H or D, called deuterium (0.0115%) (Hoefs, 2015). Due to the significant mass difference between these isotopes, hydrogen isotopic variations are highly sensitive to phase changes such as evaporation and condensation.

Oxygen is the most abundant element on Earth and has three naturally occurring stable isotopes: ^{16}O (99.757%), ^{17}O (0.038%), and ^{18}O (0.205%) (Hoefs, 2015). In isotope analysis, the $^{18}\text{O}/^{16}\text{O}$ ratio is typically measured because of their higher abundance and larger mass difference, which enhances fractionation effects during phase transitions.

Stable isotope compositions are commonly reported using δ notation and expressed in parts per thousand (denoted as ‰) relative to an international recognized standard (Craig, 1961b). The δ value is calculated using the equation:

$$\delta(\text{‰}) = \left(\frac{R_x}{R_s} - 1 \right) \cdot 1000$$

Where R denotes the ratio of the heavy to light isotope, and R_x and R_s are the isotope ratios in the sample and standard respectively. A positive δ value indicates enrichment in the heavier isotope relative to the standard, while a negative δ value means depletion.

The reference standard for hydrogen and oxygen, the zero point of the δ -scale, is V-SMOW (Vienna Standard Mean Ocean Water), originally defined by Craig (1961b) as the isotope composition of “average ocean water”. Since the D/H ratio in V-SMOW is higher than in most natural samples on Earth, δD values for natural water samples are generally negative.

Craig (1961a) first observed that the isotopic composition of meteoric water behaves in a predictable fashion. He established a linear correlation between the isotopic enrichment of $\delta^{18}O$ and δD expressed as:

$$\delta D = 8\delta^{18}O + 10$$

which describes the relationship between ^{18}O and D in meteoric waters on a global scale. This equation, known as the “Global Meteoric Water Line (GMWL)”, provides a reference for interpreting groundwaters and precipitation sources.

However, local meteoric water lines (LMWLs) often deviate from the global trend due to climatic processes (Hoefs, 2015). Variation in the slope (the numerical coefficient 8) and the deuterium intercept (the constant 10, also referred to as the “deuterium excess” or “d-excess”) depend on the conditions of evaporation, vapor transport and precipitation (Frohlich et al., 2002).

As water moves through the hydrological cycle, fractionation processes alter the relative abundances of the hydrogen and oxygen isotopes. Fractionation occurs due to two main processes: isotopic reactions and kinetic processes. In the former isotopic exchange between different chemical substances, different phase stages, or different molecules prevails, while chemical changes are insignificant. The latter is mainly driven by differences between reaction velocities of isotope molecules. It results in distinct isotopic signatures in different environmental settings.

Several geographical and meteorological factors influence these isotopic compositions. Out of these, the altitude and seasonal factors play significant roles.

The “altitude effect” refers to the systematic decrease in the isotopic composition of precipitation with increasing elevation. As moist air masses ascend, they cool and condense, preferentially removing heavier isotopes from the vapor phase. This process leads to progressively more depleted precipitation at higher elevations (Clark & Fritz, 1997).

The “seasonal effect” arises from temperature-dependent fractionation during precipitation formation (Clark & Fritz, 1997). In colder months, precipitation tends to be more depleted in heavy isotopes due to stronger fractionation at lower temperatures. Conversely, in warmer months, precipitation is enriched in heavy isotopes.

As mentioned at the beginning of this section, the isotopic composition of groundwater can also provide an approximate indication of the elevation of its recharge areas. This application is closely dependent on the knowledge of some local parameters, as the distribution in the area of the vertical isotope gradient ($\Delta\delta^{18}\text{O}/100\text{ m}$) of rainwater. Vertical isotope gradients express a linear change in the value of $\delta^{18}\text{O}$ with elevation.

The remarkable work of Longinelli & Selmo (2003) documented vertical isotope gradients in nine different areas of Italy. These gradients were derived from isotopic measurements of rainfall samples. For the estimation of the mean elevation of the recharge area of the groundwater feeding the springs in the Mt. Ario – Mt. Pezzeda – Corna Blacca area, the vertical isotope gradient calculated at Darfo-Boario and Passo della Presolana was selected as reference value. The former is a village in the Brescian Val Camonica, the latter a site in the Bergamasc Prealps, and are both located approximately 20 km NW of the study area. It is possible to estimate the mean recharge elevation of the springs using the equation:

$$HR_S = \frac{(\delta^{18}\text{O}_R - \delta^{18}\text{O}_S)}{\frac{\Delta\delta^{18}\text{O}_R}{100}} + H_R$$

Where $\delta^{18}\text{O}_R$ and $\delta^{18}\text{O}_S$ correspond to the oxygen composition of the reference area and of the springs under investigation; H_R corresponds to the elevation of the water sampled as a reference; $\Delta\delta^{18}\text{O}_R/100$ is the vertical isotope gradient calculated using the values of the reference area, that is equal to 0.1 (Longinelli & Selmo, 2003).

Given the morphological and atmospheric similarities between these sites and the study area, the application of this isotope gradient beyond its original location is justified. However, the use of this method is associated with potential uncertainties, as the estimated elevations may be either over- or underestimated. To improve accuracy, the results should be checked against the elevations of nearby ridges and peaks, as well as the geomorphological, geological and structural features of the area.

3.2.4 Springs hydrodynamics

To evaluate the characteristics and performance of an aquifer system, it is important to analyze the spring discharges using spring hydrographs. They represent discharge variation over time,

providing useful information on the aquifer's dynamics, including its storage capacity, transmissivity, and the types and volume of its groundwater reserves (Kresic & Bonacci, 2009). In mountain environments, where complex geological structures and lithological heterogeneities influence groundwater flow (Antonellini et al., 2019; Calligaris et al., 2016), hydrographs provide valuable insights into the connectivity between different aquifer units. The comparison of hydrographs with precipitation and temperature reveals how the aquifer responds to droughts and seasonal fluctuations in water availability. Moreover, the amplitude and frequency of discharge variations depicted in the hydrograph pattern help making considerations on the storage capacity of the aquifer, the role of fractures or karst features on groundwater circulation, and identifying the aquifer types (Vigna & Banzato, 2015). This information allows for the assessment of the resilience of groundwater bodies to climate variability and for developing sustainable water resource management strategies. By integrating hydrograph analysis with other hydrogeological methods, a more comprehensive understanding of groundwater behavior can be achieved, enhancing the accuracy of the conceptual models of the aquifer system. Among all the springs issuing in the study area, one spring was selected to be analyzed in detail. The spring is the Acqua Bianca, situated in the eastern portion of the study area in the upper Val Degnone, in the Corna Blacca area, which is the largest spring in the region in terms of discharge. Acqua Bianca is equipped with a probe for continuous water level monitoring. From the water level and knowing the geometry of the spring water collection tank, the spring discharge can be calculated.

3.2.5 Hydrogeological modelling

Mathematical models are fundamental tools for hydrogeological analysis and planning in karst environments. As outlined by Anderson et al. (2015), these models can generally be classified into two broad categories: data-driven models (also known as “black-box” models) and process-based models.

Data-driven models do not explicitly represent the physical processes occurring within the aquifer system, that is treated as a “black-box”. Instead, they rely on empirical relationships derived from observed data (e.g., precipitation, spring discharge) to establish statistical or mathematical correlations between inputs and outputs (Pulido-Bosch, 2021). These models are advantageous in that they require limited geological or hydrogeological data and can be implemented relatively quickly. However, because they lack an explicit representation of

system processes, their results lose their reliability outside the range of conditions they were calibrated for (Hartmann et al., 2014).

Process-based models, in contrast, use processes and principles of physics to represent groundwater flow within the problem domain. Of particular relevance within this category are the distributed models, which can be further subdivided into fully distributed and semi-distributed models (Jeannin et al., 2021).

Fully distributed models provide a detailed representation of both the geometry and physical processes of the system by numerically solving flow equations within a fully discretized spatial domain (Kovács & Sauter, 2014). These models offer a physically accurate description of the system, making them well-suited for detailed, site-specific investigations. Despite requiring large amounts of input data (e.g., topography, geology, hydraulic parameters) and significant computational resources (Hartmann et al., 2014), fully distributed models have been used to incorporate geophysically-driven datasets. There are different strategies to include karst heterogeneity in distributed models. The equivalent porous medium (EPM) approach assumes that hydraulic heterogeneities can be represented by “average” properties (i.e., an equivalent porous medium) (Hartmann et al., 2014). Since the EPM approach does not consider rapid flow in the karst conduits, it becomes unreliable in systems with high karstification. An alternative is the discrete fracture network (DFN) approach, which models only selected sets of fractures as permeable pathways, while assuming the surrounding matrix has negligible permeability (Kovács & Sauter, 2014).

Developing catchment-scale models remains a challenging task (Hartmann et al., 2014), largely due to the difficulty in the identification of preferential flow zones (such as highly fractured areas or conduits) which, in combination with diffuse matrix flow, govern the overall behavior of karst aquifers (Maqueda et al., 2023). Direct observations are often insufficient, as a significant portion of the karst network is inaccessible. Consequently, the resulting spatial statistics tend to be incomplete (Maqueda et al., 2023).

The identification of preferential flow zones in karst systems can be attained through geophysical surveys (Chalikakis et al., 2011). Geophysics gives information that contribute to increase the probability of locating conduits or other preferential flow paths containing groundwater in a karst system (Bakalowicz, 2005). In turn, this information can be used to generate hydrodynamic parameter maps for distributed hydrogeological karst models.

In this study, the primary objective of hydrogeological modelling is to support geological interpretation and refine the conceptual model of the study area. The hydrogeological model

was employed to identify key recharge areas within the catchment that contribute most significantly to spring discharge, an essential step for assessing vulnerability and safeguarding long-term drinking water quality. Additionally, the model was used to explore the potential presence of karst cavities or conduits.

A broader aim is to establish a versatile modelling framework that can be easily integrated with future geophysical data, following the approach of Vuilleumier et al. (2013), where karst features are either directly mapped or inferred and subsequently incorporated into hydrogeological models.

Code selection

For this project FEFLOW 8.0 has been adopted. It is a widely used code with capabilities of simulating flow in complex aquifers (Diersch, 2014), including karst systems (Kavouri & Karatzas, 2016; Neukum et al., 2015; Ohmer et al., 2023; Pereira et al., 2024).

The general equation adopted by FEFLOW to describe isothermal, incompressible flow in a confined aquifer can be written as:

$$S_s \frac{\partial h}{\partial t} = \nabla \cdot (K \nabla h) + W$$

where S_s is the specific storage coefficient, K is the hydraulic conductivity (m/s), h is the hydraulic head (m) and W is the sink-source term that embeds $j_{\psi}^{interface}$, i.e. the exchange occurring at the interface between the two domains, ρ is the fluid density. The general formulation for $j_{\psi}^{interface}$ is:

$$j_{\psi}^{interface} = \frac{1}{\delta S} \int n^{interface} \cdot [j + \rho(w - v)] dS$$

where δS is the projected area of the surface (m²), n is the outward-directed normal unit vector at surface, j is the interface exchange, ρ is the fluid density, w is the velocity vector at the interface and v is the velocity vector of the fluid.

Matrix flow is modelled using the Darcy's laws, according to which the discharge rate Q (m³/s) of a streamtube of cross section A (m²) is proportional to the matrix permeability k (m²), the fluid properties and the gradient. For one-dimensional flow:

$$Q = -KAi = k \frac{\rho g}{\mu} Ai$$

where the minus symbol represents the flow direction, g is the gravity force, μ is the dynamic viscosity and $i = \frac{dh}{dx}$ is the hydraulic gradient.

To simulate flow in discrete features, three different laws can be used: the Darcy's law, as defined above; the Hagen-Poiseuille law, which describes flow in a fracture; the Manning-Strickler law, which described flow in an open channel. While some scholars have adopted the Manning-Strickler law to simulate karst conduits (Ninanya et al., 2018), in this site there is no evidence of "underground rivers" or similar. It is instead more likely that flow occurs as in a highly fractured zones with limited conduit apertures, which may be better conceptualized using the Hagen-Poiseuille law. Using this law, K is proportional to the hydraulic radius (r_{hyd}) of a tube. The discharge rate from a circular tube can be written as

$$Q = -\frac{Wb^3}{12\mu}i$$

where W (m) and b (m) are respectively the width and aperture of the tube, the latter being twice the tube radius.

3.3 Geophysical surveying

3.3.1 Objectives of geophysical investigations

Despite the 3D geological model provides continuous and homogeneous information, uncertainties remain, especially at depth. These uncertainties arise because geological constraints are derived from surface observations and scattered data, making it challenging to extend the information at depth to characterize deeper structures.

To overcome these challenges, geophysical surveys offer reliable and efficient tool to characterize the subsurface architecture at depth (Binley et al., 2015). These methods offer indirect yet valuable insights into the spatial distribution of structural discontinuities and lithological variations and can thus resolve the indetermination of the geological constraints at depth.

The natural heterogeneity of the subsurface strongly influences the overall groundwater behavior. Integrating geophysical data with geological and hydrogeological observations aids in structural characterization of aquifer systems, detecting discontinuities such as faults, fractures, lithological boundaries, and karst features.

In karst-fissured terrains, electrical (E) and electromagnetic (EM) methods are particularly well suited for subsurface investigations (Binley et al., 2015; Chalikakis et al., 2011; Sumanovac & Weisser, 2001) allowing to characterize the electrical resistivity distributions at depth. Since resistivity is highly sensitive to variations in rock type, porosity, connectivity of pores, water content, clay content, and metallic content (Binley et al., 2015; Boyd et al., 2024; Holmes et al., 2022; Revil & Florsch, 2010), these methods have significant potential for hydrogeophysical applications. By refining the structural understanding of the subsurface, geophysical surveys contribute to reduce uncertainties in geological models and improve structural and hydrogeological assessments. Furthermore, geophysical surveying allows to map heterogeneities within geological formations, often linked to the presence of water.

Electrical methods are more sensitive to resistive anomalies, while electromagnetic methods are more sensitive to conductive bodies. Some methods (like EM methods with smaller dipole moments and DCIP methods) provide high resolution at shallow depths, while others penetrate deeper but with lower resolution (like AEM methods). However, for the characterization of springs in mountain areas, the field logistics plays the larger role in defining the most appropriate exploration method. Here only AEM allows to map the subsoil with areal coverage, despite of rugged terrain and dense vegetation, down to hundreds of meters. In addition, new TEM equipment mounted on backpacks (the Loupe system by Loupe geophysics) are able to

survey up to 10 km per day, but with much lower penetration depths. For these reasons, AEM and ground EM in continuous acquisition are the main mapping tools proposed in this study, with classic DC resistivity used for validation.

In the next Sections, an overview of the resistivity values for different geological setting is illustrated, and the different methods deployed in this study are presented.

3.3.2 Electrical properties of geological materials

As previously mentioned, the primary objective of the E and EM surveys is to improve subsurface characterization by analyzing its electrical properties. These methods are most effective in areas where there is a significant contrast in electrical properties between different geological units, allowing them to serve as indirect indicators for subsurface mapping.

Note that these techniques cannot directly determine the precise composition of mineral ores or specific lithological units. This limitation arises because the bulk electromagnetic properties of rocks are complex function of multiple variables, including mineral content, porosity, pore fluid conductivity and saturation. Consequently, these methods provide information but are not prescriptive. Nevertheless, when the correspondence between geological units and geophysical models is achieved, geophysics allows to map heterogeneities within geological units or at their interfaces that may be a proxy for the presence of groundwater.

Fig. 3.1 shows the ranges of electrical resistivity in geological materials, spanning over seven orders of magnitude.

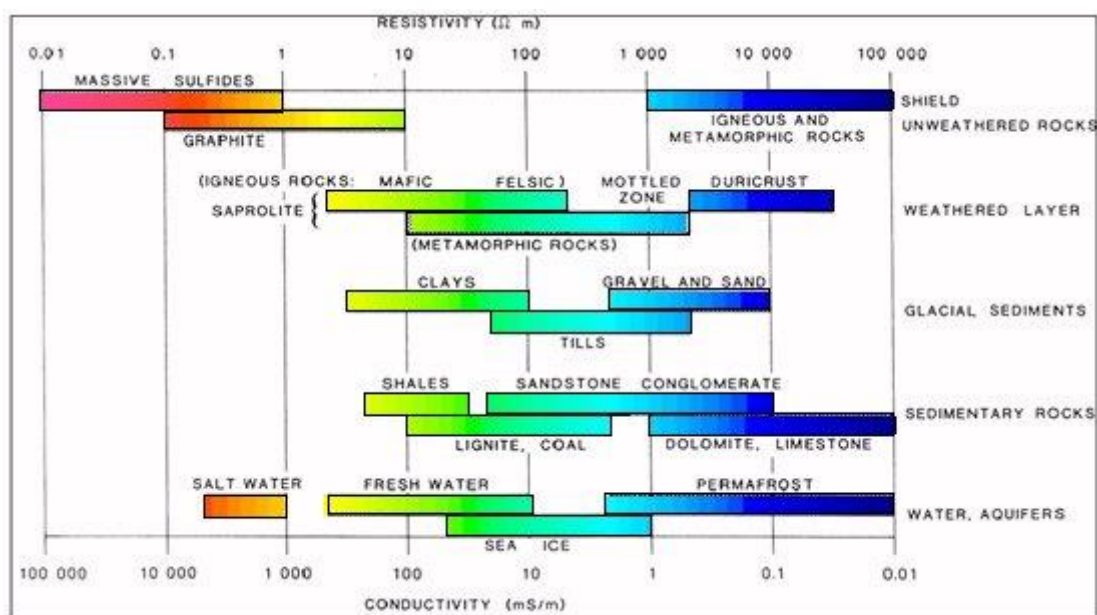


Fig. 3.1 Typical resistivity ranges of the most common lithotypes (Palacky, 1987).

In porous-medium contexts (i.e., in absence of metallic conduction given by the concentration of ore minerals) the ground conductivity is controlled by several geological features including:

- Porosity, which determines the volume of void spaces within the rock.
- Hydraulic permeability, which describes how pores are interconnected.
- Saturation, as the presence of water significantly affects conductivity.
- Concentration of dissolved electrolytes, since ions in the pore fluid contribute to electrical conduction.
- Temperature and phase of pore fluid, which influence ionic mobility.
- Amount and composition of colloids (clay content).

From a physical perspective, these factors control the electrolytic conductivity and the surface conductivity (the one associated with the mineral-fluid interface). Among all, pore space and pore geometry are the most impacting factors in determining bulk conductivity. Porosity in rocks exists mostly in the form of joints, fractures, vugs (dissolution cavities in limestones and dolomites), and intergranular voids in sedimentary rocks. The impact of porosity on the lab-measured resistivity for different rock types is illustrated in Fig. 3.2.

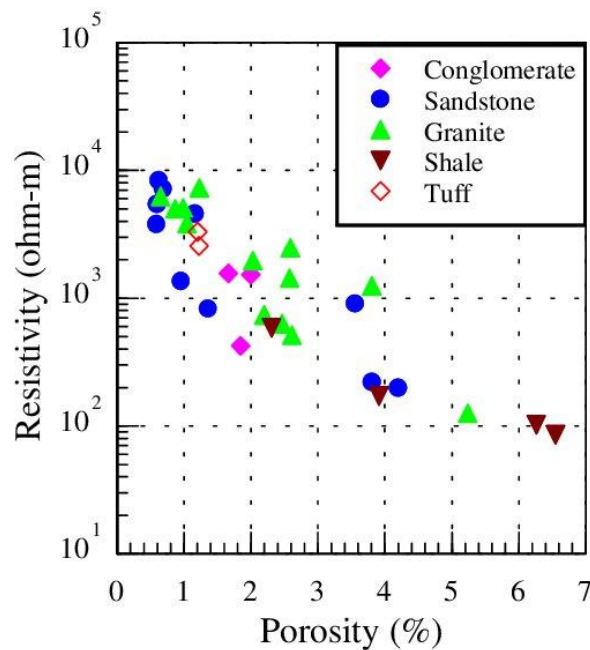


Fig. 3.2 Relationship between electrical resistivity and porosity of different type of rocks (Park et al., 2016).

An empirical relationship between physical properties of porous rocks and resistivity, that is the basics for quantitative interpretation, is provided by well-known Archie's Law (Archie, 1942):

$$\rho_r = F \cdot \rho_w S_w^{-n}$$

In which ρ_r is the rock resistivity with fluids, F is the formation resistivity factor, ρ_w is the pore-fluid resistivity, S_w is the water saturation and n is a parameter experimentally defined which is usually set as 2. In case of full water saturation, the formation factor is related to porosity through the equation:

$$F = \rho_r / \rho_w = a \varphi^{-m}$$

In which a is an empirical constant, m is an empirical constant related to cementation that spans from 1 to 3, and φ is the porosity.

3.3.3 Direct current (DC) resistivity principles and modelling

Direct current (DC) electrical resistivity measurement is a geophysical technique investigating the subsurface electrical resistivity distributions. It is widely used in hydrogeological investigations because of the sensitivity of electrical properties to factors such as lithology, presence of fractures and cavities, water content, and clay content. However, this also represents a potential vulnerability of resistivity measurements: the high risk of misinterpretation or ambiguity in the modelled subsurface structures when supporting information is not available (Binley & Slater, 2020). Electrical Resistivity Tomography (ERT) represents a type of DC resistivity survey, in which resistivity is mapped along 2D profiles or 3D volumes.

The DC resistivity technique involves applying an electric field E to the subsurface, which results in the transport of free charges. The charge carriers can be electrons (e.g., in the presence of high concentrations of ore minerals) or ions (which is the common case for charge transport in fluids in the subsurface).

Resistivity measurements are made using a quadrupole, i.e. a four-electrode configuration. Two electrodes are used to inject electrical current I (units of amperes, A) into the ground the other two measures the resulting potential differences V (units of volts, V). This process is governed by Ohm's law, which states:

$$\vec{J} = -\sigma \cdot \nabla V = \sigma \cdot \vec{E}$$

The current density J is defined as the current per unit cross sectional area, is oriented in the direction of the movement of charges and is generally assumed to be proportional to the electric field strength ($E = -\nabla V$) (Binley & Slater, 2020).

The constant of proportionality between current density and electric field is the material property electrical conductivity σ (units of siemens/meter, S/m) equals the inverse of the electrical resistivity ρ (units of ohm meter, Ωm). For the simple case of a current-carrying wire, the potential difference between the two ends of the wire is the line integral of the electric field along the wire changed of sign:

$$V = - \int \vec{E} \cdot d\vec{l} = - \int \rho \cdot \vec{j} \cdot d\vec{l}$$

The total electrical current I is defined as the flux of the current density $\vec{j}$ thorough a section $\vec{A}$ using the relationship $I = \vec{j} \cdot \vec{A}$.

Thus, assuming a uniform electric field across an electrically homogeneous wire with constant section, the potential difference is:

$$V = \rho \cdot I \cdot \frac{l}{A}$$

By reversing the equation, the resistivity of a wire is obtained:

$$\rho = \frac{V}{I} \cdot \frac{A}{l} = \frac{V}{I} \cdot k$$

where $k = \frac{A}{l}$ is known as the geometrical factor, expressed in m, that depends on the electrode configuration used.

Considering a current passing a point-like electrode on the ground surface ($z=0$). The current flows radially in the ground if it is homogeneous and isotropic. The law of charge conservation states that the net flow of current density through a closed surface (the ground) is equal to the negative time derivative of the total charge contained within the enclosed volume:

$$\nabla \cdot \vec{j} = - \frac{\partial \rho}{\partial t}$$

where ρ represents the electrical resistivity distribution.

The voltage gradient is defined by the Poisson's equation for galvanic data:

$$\nabla \cdot (-\sigma \cdot \nabla V) = - \frac{\partial \rho}{\partial t} = -I \cdot \delta(x, y, z)$$

where the source term is expressed in terms of the current I injected in the position $\delta(x, y, z)$ using the Dirac delta function (which takes on a value of 1 at the position x, y, z and is 0 elsewhere).

In homogeneous and isotropic media, considering that the current density J is the current I divided on the area where it distributes (a hemisphere, therefore $2\pi r^2$) and imposing the boundary condition of no flux out of the ground, using Ohm's law it is possible to calculate the potential at any distance r as:

$$V(r) = \frac{\rho I}{2\pi r}$$

Considering that the configuration of the electrodes in DC methods is composed by two electrodes for injecting current (A and B) and two electrodes to measure the resulting potential (M and N), the potential difference between M and N is expressed as:

$$V_{MN} = V_M - V_N = \frac{\rho I_{AB}}{2\pi} \cdot \left\{ \left[\frac{1}{\overline{AM}} - \frac{1}{\overline{MB}} \right] - \left[\frac{1}{\overline{AN}} - \frac{1}{\overline{NB}} \right] \right\}$$

From which a formulation for ρ is derived:

$$\rho = \frac{V_{MN}}{I_{AB}} \cdot \frac{2\pi}{\left(\frac{1}{\overline{AM}} - \frac{1}{\overline{AN}} - \frac{1}{\overline{BM}} + \frac{1}{\overline{BN}} \right)} = \frac{V_{MN}}{I_{AB}} \cdot k$$

Where $\overline{AM}$, $\overline{MB}$, $\overline{AN}$, and $\overline{NB}$ represent the distances between the electrodes in the quadrupole configuration, V_{MN} is the voltage measured by the potential dipole, I_{AB} is the current injected through the current dipole, and k is the geometrical factor for a quadrupolar measurement on a halfspace.

However, in reality the ground is not a homogeneous halfspace. Therefore, what it will be measured on the field is the apparent resistivity ρ_{app} , which is the resistivity of a homogeneous halfspace that gives the same response to the site investigated (Binley & Kemna, 2005). In other words, it reflects the average resistivity of the ground.

Calculating the apparent resistivity for a given resistivity distribution involves solving the galvanic forward problem.

3.3.4 Transient electromagnetic (TEM) principles and modelling

Electromagnetic (EM) measurements are used to characterize the subsurface electrical properties, similar to galvanic DC methods. However, while both approaches aim to map

resistivity variations, EM techniques rely on different physical principles to probe the underground structures.

The fundamental principles of EM induction are based on two key concepts: i) an electric current flowing through a conductor generates a magnetic field, as described by Ampère's law; ii) a time-varying magnetic field induces an electric current in a conductive medium exposed to the field, in accordance with Faraday's law of induction. Thus, understanding the EM method relies on the equations that govern these two fundamental laws.

In Transient Electromagnetic (TEM) surveys, a continuous electrical current is injected through a transmitter loop, which remains electrically isolated from the ground (i.e., no direct galvanic contact). This current generates a primary magnetic field, which collapses when the current is rapidly turned off. According to Faraday's law of induction, this abrupt change in the magnetic field induces an electromotive force (emf) in the subsurface, proportional to the time-derivative of the flux of the magnetic field changed of sign:

$$\oint_C E \cdot dl = -\frac{d}{dt} \oint_S B \cdot dA$$

Where E is the electric field, B is the magnetic induction and the S and C subscripts of the integral sign represent the surface through which the flux is computed and its boundary, respectively.

Building on the principles, the rapid decay of the primary magnetic field induces secondary electric field in the subsurface. This, in turn, generates eddy currents, which circulate within the conductive earth and gradually dissipate due to ohmic losses. The decay of these currents produces a secondary magnetic field, which serves as the key signal detected in TEM measurements. It is directly recorded by the receiver as a voltage decay curve, known as the "transient". The transient is sampled through a series of acquisition channels, called gates, which open at progressively increasing time intervals, allowing for the investigation of deeper and deeper subsurface layers. Since the signal attenuates with depth, it is necessary to progressively extend the reception windows of each gate (their width increases following a logarithmic progression) in order to enhance the instrument's acquisition capability. The whole process takes only a few milliseconds—after the transmitter current is fully switched off—allowing precise monitoring of the secondary magnetic field evolution over time. The measured transients provide information on the subsurface conductivity distribution, allowing for the detection of geological structures and materials.

Electromagnetic noise in TEM data includes geomagnetic signals, radiation from high-voltage power lines, and disturbances generated by electromagnetic storms (Spies & Frischknecht, 2008). These storms are concentrated in tropical areas, while the activity decreases moving away from the equator. An anomalous EM response can be linked to currents induced in metal conductors such as iron fences or distribution networks and pipelines that are located above the investigated half-space. This generates an electromagnetic coupling that affects the recorded voltage values. In detail, there are two different types of coupling: capacitive and galvanic. The first is easily recognizable on the airborne EM data since it manifests itself as oscillations and sudden changes of sign on the data while the second is more difficult to interpret since a general increase in the signal is observed especially at late times. This is linked to the fact that galvanic coupling is generated by anthropic structures that are not electrically isolated from the ground. EM coupling must be removed from the EM data through accurate manual processing, culling the affected data out; inverting data containing this type of disturbance can lead to high misfit values or can introduce artifacts, making the obtained models unusable.

IP in TEM

When the subsurface resistivity is frequency-dependent, that is, when there is a chargeable medium underground, an induced polarization (IP) effect will occur. When the IP effect is strong, negative responses will be observed in the collected TEM data. At present, significant IP effect data have been found in both ground-based and airborne TEM surveys, which are common in mineral resource exploration, permafrost areas or clay layer areas. In this case, it is necessary to model the IP effect when interpreting TEM data, because ignoring the IP effect can easily lead to erroneous resistivity inversion results. The classic Cole-Cole model of the complex resistivity can be expressed as (Pelton et al., 1978):

$$\rho(\omega) = \rho_0 \left[1 - \frac{m_0}{10^3} \left(1 - \frac{1}{1 + (i\omega\tau_\rho)^C} \right) \right]$$

where ρ_0 is the direct current resistivity ($\Omega \cdot m$), m_0 is the intrinsic chargeability (mV/V), τ_ρ is the relaxation time (s), ω is the angular frequency, and C is the frequency exponent. However, there is a strong correlation between m_0 and C , which increases the complexity of the inversion. To this end, Fiandaca et al. (2018a) proposed a MPA re-parameterization model, which weakens the correlation between polarization parameters and can improve the resolution of the inversion:

$$\mathbf{m}_{MPA} = \{\rho_0, \varphi_{max}, \tau_\varphi, C\}$$

where φ_{max} is the maximum phase angle of the Cole-Cole complex conductivity and τ_φ is the inverse of the frequency at which φ_{max} is reached.

1D EM modelling

In this study, the EEMverter modelling platform (Fiandaca et al., 2024a) was used to simulate TEM data. TEM one-dimensional (1D) modelling and interpretation is currently the most widely used method because of its fast simulation speed and wide range of applications. 1D modelling approximates the earth to a layered geological distribution and calculates the response by setting electrical parameters on each layer of geological structure. In the EEMverter simulation, whether it is a 1D or a 3D modelling, the frequency domain response is first calculated and then converted to the time domain response through the Hankel transform. The magnetic field in frequency domain is computed integrating the contributions of infinitesimal loop pieces along the transmitter:

$$B_{z,FD}(\sigma, h, \omega, x, y, z) = \oint_{Tx\ loop} dB_{z,FD}$$

where $\sigma = \{\sigma_1, \dots, \sigma_N\}$ and $h = \{h_1, \dots, h_{N-1}\}$ represent the electrical conductivity and the layer thicknesses of a N-layered 1D model respectively, ω is the angular frequency, x, y, z indicate the position of the receiver, and $dB_{z,FD}$ is the contribution of the infinitesimal loop piece of length ds , expressed as:

$$dB_{z,FD}(\sigma, h, \omega, x_*, y_*, z_*) = \frac{I \cdot ds}{4\pi} \cdot \frac{y_*}{\rho_*} \oint_0^\infty (1 + r_{TE}) e^{\mu_0 z_*} \cdot \frac{\lambda^2}{\mu_0} \cdot J_1 \cdot (\lambda \cdot \rho_*) \cdot d\lambda$$

Where, in turn, x_*, y_*, z_* are the receiver coordinates in the reference system that sees the infinitesimal loop piece as oriented in the x direction and centered at its origin, J_1 represents the Bessel function of order 1, $\rho_* = \sqrt{x_*^2 + y_*^2}$ is the distance between the receiver and the segment origin, I is the current in the loop, $u_0 = (\lambda^2 - \omega^2 \mu_0 \epsilon_0)^{1/2}$ and r_{TE} represents the reflection coefficient of the electric field transverse to the z-axis, defined as:

$$r_{TE} = \frac{Y_0 - \hat{Y}_1}{Y_0 + \hat{Y}_1}$$

Where Y_0 and $\hat{Y}_1$ are the free space and the surface assumptions respectively, computed as in Ward & Hohmann (1988).

3D EM modelling

1D modelling can quickly image underground resistivity, but the real world is three-dimensional (3D). When geological conditions are complex and TEM data response shows obvious 3D characteristics, using 3D modelling can significantly improve model reliability. The EEMverter modelling platform (Fiandaca et al., 2024b) has adopted a primary-secondary field separation technique for 3D modelling. Unlike the traditional total field computation strategy, this method does not require dense meshing near the source and has a smaller footprint area (Zhang et al., 2021a), conducive to the substantial conservation of computational resources. By separating the primary field E^p and the secondary field E^s , the background conductivity σ^p and the abnormal conductivity σ^s , and applying Dirichlet boundary conditions, the frequency domain EM double-curl partial differential equation is solved based on the vector finite element idea:

$$\int_{\Omega} (\nabla \times \mathbf{N} \cdot \nabla \times \mathbf{E}^s - i\omega\mu(\sigma^s + \sigma^p)\mathbf{N} \cdot \mathbf{E}^s) dV = \int_{\Omega} (i\omega\mu\sigma^s \mathbf{N} \cdot \mathbf{E}^p) dV, \\ E_s|_{\Omega} = 0.$$

Computations are carried out in the frequency domain, and then time-transformed in time domain through a Hankel transform.

3.3.5 Geophysical acquisition systems

Terrameter

In this study, DC resistivity surveying was conducted using an ABEM Terrameter L2 multi-electrode survey system (GuidelineGeo), applying a multiple gradient array configuration (Fig. 3.3). This electrode array was chosen because it is suitable for multichannel data-acquisition system, so that many data points can be recorded simultaneously for each current injection, thus reducing fieldwork time without compromising the data density and providing high-resolution imaging of subsurface resistivity variations (Dahlin & Zhou, 2006). In addition, ERT method with gradient configuration was successfully used in karst terrains (Sumanovac & Weisser, 2001). The DC resistivity method proves particularly effective in such geological settings, as it allows the imaging of electrical resistivity contrasts between high-resistivity massive formations and low-resistivity clayey formations or water-saturated/clay-filled fractures and karst features.

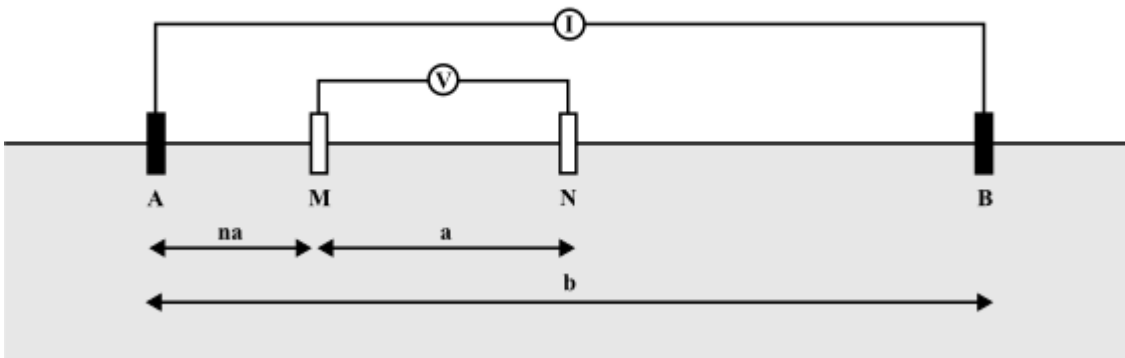


Fig. 3.3 Schematic representation of the multiple gradient array, with A and B are current electrodes, and M and N are potential electrodes.

Loupe system

The Loupe EM profiling system is a portable TEM device designed for high-resolution mapping of electrical resistivity in the near surface (Street et al., 2018). Unlike traditional TEM setups, Loupe is optimized for easy to carry rapid profiling, making it an efficient tool for investigations in rugged mountain terrains.

The system consists of two key components: a transmitter coil and a receiver coil, each assembled on a backpack, allowing it to be carried by two field operators while walking continuously at a slow pace (Fig. 3.4).

The transmitter generates bi-polar square waveforms of current, switching off within 10 microseconds. The receiver captures the resulting secondary EM response, enabling real-time assessment of subsurface conductivity.

Loupe operates in a continuous profiling mode, with a fixed 10 m separation between the transmitter and receiver. The depth of investigation (DOI) varies based on ground conductivity, system geometry, and measurement frequencies. To ensure precise and consistent data collection, the system features real-time communication between the transmitter and receiver, allowing for automatic spacing control and data recording.

An integrated GPS system provides absolute positioning and timing synchronization, enhancing the accuracy of the measurements.

Due to its high mobility and rapid data acquisition, Loupe is particularly well-suited for applications in groundwater exploration in mountain environments, offering a practical and efficient solution for identifying subsurface water-bearing formations and conductivity anomalies.

Since the relative posture of the transmitting coil and the receiving coil is constantly changing during walking, it will bring challenges to data processing and inversion. In addition, the effective depth of investigation for the Loupe system is usually within 50 m, which is suitable for rapid exploration of shallow surface targets.



Fig. 3.4 Two stages of Loupe acquisition in Val Sabbia.

Airborne electromagnetic (AEM) systems

Among EM geophysical methods, airborne EM (AEM) surveys stand out as one of the most versatile and efficient techniques. Their rapid data acquisition capability and ability to collect high-density measurements over large areas make them particularly valuable for mapping subsurface electrical properties in regions where ground-based surveys would be impractical. Like other EM techniques, AEM relies on a changing primary magnetic field generated around a transmitter loop. This induces eddy currents in the subsurface, which in turn produce a secondary magnetic field detected by the receiver coil.

AEM instruments, typically deployed on low-flying helicopters, collect multi-sensor data along a predefined flight path, with sampling rates ranging from a few to several hundred measurements per second.

While AEM systems generally cannot achieve the same lateral and vertical resolution as ground-based instruments, they serve as a complement by extending and interpolating sparse ground-based data across much larger areas (Paine & Minty, 2005).

The data acquisition campaign in Val Sabbia was conducted using the SkyTEM system, whereas the Xcite system (NRG) was employed for data collection in Aracena. These systems differ both in the configuration and in survey depth. Here is a brief presentation of the two.

In SkyTEM helicopter-borne TEM systems (Fig. 3.5A), both the transmitter loop and the receiver coil are mounted on the same rigid structure, known as the frame, with a vertical separation of approximately 2 m. The frame consists of a lightweight octagonal latex structure, suspended about 20 m below the helicopter, minimizing electromagnetic interference from the aircraft.

In SkyTEM system, the primary field is removed by positioning the receiver in the so-called “zero position”, where the primary field is null by definition (Auken et al., 2015). Additionally, SkyTEM is designed for dual-moment operation, it uses both low and high moments (LM and HM respectively), enabling high-resolution near surface imaging while maintaining good resolution at depth (up to 500 m or more). The LM allows the current to be turned off more quickly, enhancing the shallow resolution.

To ensure accurate data collection, the SkyTEM frame is equipped with several monitoring instruments, including: i) a GPS system for precise positioning; ii) three inclinometers to track tilt variations, allowing necessary corrections during data processing; iii) additional sensors to monitor flight altitude and tilt throughout the survey. This instrumentation ensures that positioning errors and frame tilts are accounted for in the early stages of data processing, improving the reliability of the final subsurface resistivity models.

In Xcite helicopter-borne TEM system (Fig. 3.5B), the frame has a 20 m diameter patented inflatable design, and it is suspended about 30 m below the helicopter. The transmitter coil is positioned along the perimeter of the inflatable frame, creating a large loop, within which the receiver coil is placed, near its center, where the primary field is effectively canceled by the bucking coil. It works by generating an opposing magnetic field that nullifies the primary field at the receiver location. This configuration ensures that the receiver primarily detects the secondary field induced in the subsurface, minimizing the influence of the primary field.

Its inflatable frame allows for faster deployment compared to the SkyTEM system; however, it provides less control over the acquisition geometry. Ideally, this geometry should remain rigid and stable to ensure optimal data quality.

This system operates at a single base frequency with an effective penetration depth that varies depending on ground conductivity but is generally suited for shallow to medium depth investigations.

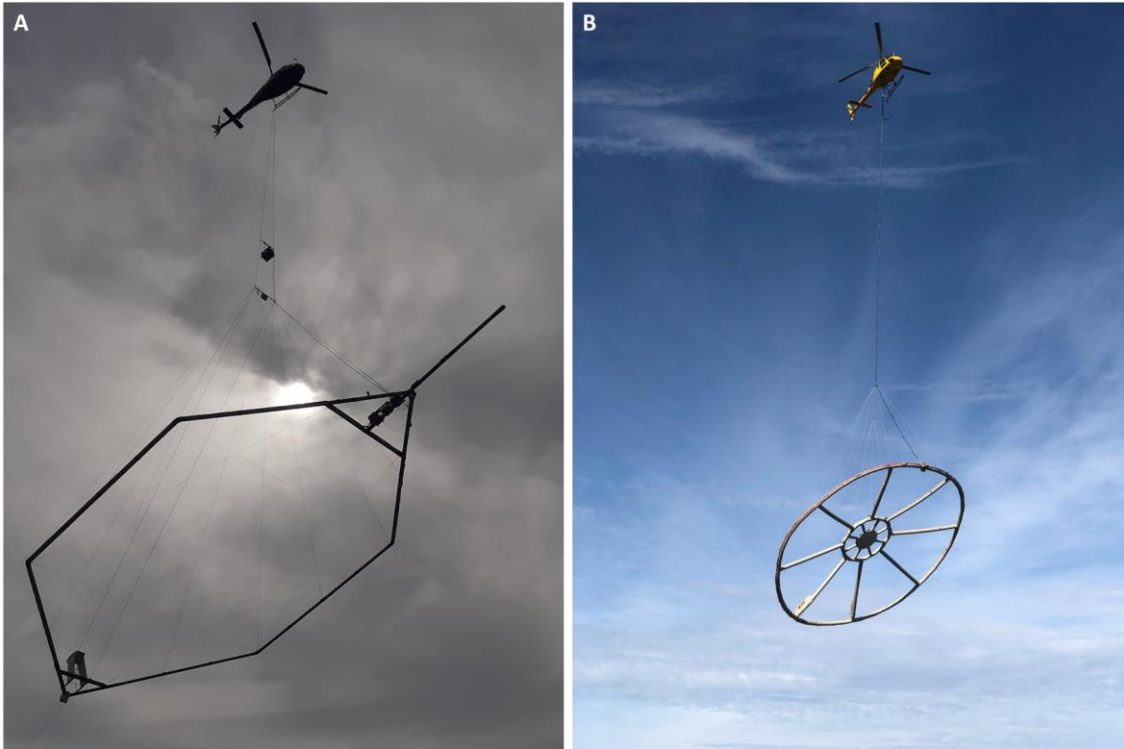


Fig. 3.5 Airborne systems. A: SkyTEM system. B: Xcite system.

3.3.6 Geophysical processing and inversion

Before proceeding with the modelling of the data, it is essential to carefully re-examine all collected datasets to identify and eliminate any sources of error that could compromise the reliability of the inversion results. This step involves checking for inconsistencies, noise, or anomalies that may arise from instrumental errors, environmental interferences, or improper data acquisition. By cleaning and refining the dataset, the modelling process can be conducted with greater accuracy, ensuring that the final resistivity model is as representative of the true subsurface conditions as possible. This step is carried out in EEMstudio (Sullivan et al., 2024), a QGIS plugin for data visualization processing in GIS environment.

In order to translate geophysical measurements into useful information for a better understanding of subsurface systems, two fundamental problems have to be solved: the forward problem and the inverse problem.

Take the forward modelling of DC method as an example, the forward problem consists of predicting the theoretical response of a given resistivity distribution based on the governing physical law, Poisson's equation. This is achieved through forward modelling, which calculates the expected measurements (data) for a known subsurface structure. To represent the variation in resistivity, the subsurface is discretized into layers (in 1D) or cells (in 2D and 3D), each assigned a uniform resistivity value (Binley & Slater, 2020).

However, in geophysical exploration the inverse problem must be solved—meaning that, given a set of measured data (typically apparent resistivity values), the goal is to determine a resistivity model that best explains the observations. This process, known as inversion, is inherently non-linear, requiring sophisticated computational techniques to iteratively refine the model. Note that the inverse problem does not have a unique solution; multiple models can produce similar responses within an acceptable level of uncertainty.

Since inversion relies on forward modelling to compare predicted and observed data, an accurate forward model is crucial for assessing the fit of the solution. By integrating these approaches, E and EM surveys can provide information on the spatial distribution of electrical properties, despite the inherent complexity and non-uniqueness of the inversion process.

In this thesis, the EEMverter inversion software has been used. EEMverter is a forward and inversion modelling software specifically developed to model E and EM data, also in joint and time-lapse inversions. It allows to run modelling of galvanic data, ground- and airborne-based inductive data (Fiandaca et al., 2024a).

In EEMverter the inversion parameters are defined on model meshes which do not coincide with the forward meshes used for data modelling: the link between model and forward meshes is obtained interpolating the model mesh parameters into the forward mesh discretization. The interpolation from the model parameters M defined on the model mesh nodes into the values m at the subdivisions of the forward mesh is expressed through a matrix multiplication. The time-domain Jacobian in the forward mesh is computed as:

$$J_{m,TD} = A \cdot T \cdot J_{m,FD}$$

where the matrix T holds the Hankel coefficients, the matrix A implements the current waveform, gate integration and filters, and the frequency-domain Jacobian $J_{m,FD}$ is calculated in 1D through finite difference and in 2D/3D using the adjoint method and the chain rule as in Fiandaca et al. (2013):

$$J_{m,FD} = J_{\sigma^*} \cdot \frac{\partial \sigma^*}{\partial m}$$

where J_{σ^*} is the Jacobian of the forward mesh with respect to the complex conductivity σ^* and $\frac{\partial \sigma^*}{\partial m}$ is the partial derivative of the complex conductivity versus the model parameters. Finally, the Jacobian of the model space J_M is computed translating the forward-mesh Jacobian through the interpolation Matrix F as:

$$J_M = J_m \cdot F^T$$

The total Jacobian is used for computing the inversion model in a Levenberg-Marquardt linearized approach as follows:

$$M_{n+1,j} = M_{n,j} + [J_{M,j}^T C_d^{-1} J_{M,j} + R^T C_{R,j}^{-1} R_j + \lambda I]^{-1} \cdot [J_{M,j}^T C_d^{-1} \cdot (d - f_{n,j}) + R^T C_{R,j}^{-1} R_j \cdot M_{n,j}]$$

where the subscript j indicates that the inversion process can be split into several inversion cycles: in each cycle j it is possible to change the forward computation for each dataset (e.g. from 1D to 3D), as well as the regularization matrix $C_{R,j}$. In this study both L2 and Sharp regularization schemes are adopted, using the generalized asymmetric minimum gradient support (Fiandaca et al., 2015) for the sharp regularization. In all inversions, the Depth of Investigation is computed as well, following the approach presented in Fiandaca, Christiansen, et al. (2015).

3.4 Methodological development: Integrated Geological and Geophysical modelling

Geophysical inversion often faces challenges related to non-uniqueness and limited resolution issues, meaning that multiple subsurface configurations can explain the same dataset, making it difficult to derive an accurate and unique interpretation.

Integrating 3D geological models (incorporating stratigraphic, structural, and lithological information) can help constrain geophysical inversion, ensuring that the results are consistent with known geological features, and in turn, the geophysical results feed back into and refine the geological model. In this process, geophysics plays a crucial role in guiding the interpretation and modelling of geological interfaces at depth.

The outcomes of geologically constrained inversions can serve as valuable inputs for refining the geometry of deep geological interfaces. By integrating geophysical results into the 3D geological model through an iterative process, it becomes possible to continuously update and improve geological interpretations at depth. This overcomes the issue of sparse geological data. This process of mutual calibration, where the geological model informs the geophysical model and vice versa, highlights the central role of geophysics in guiding the interpretation and modeling of subsurface structures.

This approach offers several key benefits: improved geological consistency, ensuring that the geophysical model reflects realistic subsurface conditions; enhanced resolution of features such as fault networks and lithological boundaries; reduction of inversion artifacts, minimizing distortions caused by data scarcity and increasing the overall reliability of the model.

To implement the integration of the 3D geological model with geophysical models, it was necessary to introduce two fundamental rules into the EEMverter inversion kernel.

The first rule treats the basal boundaries of modelled geological units as resistivity barriers that separate distinct regions within the model. This means that constraints on the electrical resistivity parameter can be broken at these boundaries, allowing for sharper discontinuities in the resistivity profile. The rationale behind this rule is that adjacent geological formations often exhibit significantly different electrical properties due to variations in lithology, porosity, and water content. By allowing resistivity contrasts at geological boundaries, the model can better represent sharp geological transitions, such as contact between highly resistive limestones and more conductive clay-rich or water-bearing units. Constraint-breaking across regions removes limitations in the objective function, without compromising the inversion ability to converge. As such, it represents a relatively safe form of informing the geophysical inversion with geology.

The second rule introduces the ability to define distinct resistivity ranges for each geological region within the model. This approach ensures that the resistivity values assigned to a specific unit remain within geologically plausible limits, reducing the likelihood of unrealistic interpolations or smoothing artifacts. By constraining resistivity variations within predefined ranges, the model can more accurately reflect the electrical properties of different rock units, improving the correlation between geophysical data and geological interpretation. Imposing resistivity ranges is a strong constraint. If the selected range is incorrect, it can prevent the inversion from converging, significantly affecting the final outcome.

These calibration rules help refine the integration of geophysical and geological models, enhancing the reliability of inversion results and supporting more accurate subsurface characterizations. This is particularly beneficial for complex geological settings where faults, karst systems, and stratigraphic variations shape the geological architecture, and consequently the subsurface conductivity distribution.

For the implementation of the first rule in the EEMverter inversion kernel, the initial step involves exporting the modelled geological interfaces (both geological contacts and faults) from GeoModeller in a compatible format (VTP or VTK, see Section 3.1.3).

Once the code successfully identifies the surfaces of the entire geological pile, these interfaces are incorporated into EEMverter to define distinct regions within the inversion model.

The next step is to introduce an option that allows for the break of resistivity constraints across adjacent regions. This means that at the boundaries defined by the geological interfaces, resistivity values are no longer forced to be continuous, enabling the inversion process to better reflect sharp geological transitions and variations in subsurface properties.

For the implementation of the second rule, it is necessary to define a resistivity range for each outcropping geological formation. This range includes a maximum, a minimum, and an average resistivity value, with the latter serving as the starting value for the inversion process.

As the inversion progresses through successive iterations, the resistivity value is allowed to vary within the predefined range. The algorithm adjusts these values dynamically until it converges on the best fit resistivity, optimizing the model to achieve the most accurate representation of the subsurface conditions. This approach helps to maintain geologically realistic constraints while allowing flexibility in the inversion process to accommodate local variations in subsurface properties.

4. Results and discussion on the Val Sabbia study site

4.1 Definition of the 3D geological architecture

4.1.1 Harmonization of geological constraining data

The development of the 3D geological model begins with the input data preparation phase. This process involves gathering and organizing all available datasets, which consist of geological maps and cross sections. Additionally, new data are acquired through targeted field investigations, ensuring that critical geological features are well documented. To ensure spatial accuracy, all collected data are georeferenced within the same coordinate system (EPSG 32632 – WGS 84 UTM Zone 32N). This standardization allows for precise overlay and comparison of different datasets.

A crucial step in this phase is quality assurance, where datasets are reviewed for consistency, accuracy, and reliability. Data standardization is then performed to harmonize different formats and resolutions, facilitating seamless integration within the modelling framework.

Geological maps serve as primary surface constraints for 3D modelling. They contain 2D information on the location of geological boundaries and faults, and the attitude of stratigraphic and tectonic interfaces. However, the geological mapping coverage of the study area is not uniform, with different sources providing various levels of detail and completeness. A summary of all data sources used is given in the Tab. 4.1.

The most recent and authoritative reference maps are produced by the Geological Survey of Italy as part of the CARG Project at a scale of 1:50.000, including the Iseo (099), Breno (078) and Bagolino (079) sheets (Servizio Geologico d'Italia, 2011, 2012, 2016). While these maps offer high-resolution geological data, their coverage is limited to the western and northern portions of the study area, leaving significant areas, particularly in the central and southern portions, without detailed cartography. These maps are available in digital format through the Lombardy Region geoportal (<https://www.geoportale.regione.lombardia.it/>).

To compensate for these gaps, the geological map by Boni & Cassinis (1973), also at a 1:50.000 scale, was incorporated into the study. Although available only in paper format, this map provides complete coverage of the entire study area, ensuring a continuous and comprehensive representation of regional geology. Its inclusion complements the more recent but spatially restricted CARG maps.

Additionally, detailed geological investigations have been conducted in the southeastern sector of the study area as part of academic research, specifically in the thesis by Motta (1953). This

work resulted in high-detail (1:10.000 scale) geological mapping, enhancing the dataset with localized structural and stratigraphic information.

Two lower resolution geological maps were also used to provide supplementary context. These include the 1:100.000-scale geological map by the Servizio Geologico d'Italia (1970), and the 1:200.000-scale regional geological synthesis by Castellarin (1981).

To ensure consistency and integration of the datasets, all paper-based maps were digitized, georeferenced (EPSG 32632 – WGS 84 UTM Zone 32N), and vectorized. This process was carried out within a Geographic Information Systems (GIS) environment, using ArcGIS Desktop. Standardizing the data in this way facilitated its incorporation into the 3D geological modelling workflow, ultimately resulting in a harmonized geological map of the study area (Fig. 4.1).

Name	Author	Scale
CARG Iseo	Servizio Geologico d'Italia (2011)	1:50,000
CARG Breno	Servizio Geologico d'Italia (2012)	1:50,000
CARG Bagolino	Servizio Geologico d'Italia (2016)	1:50,000
Carta geologica delle Prealpi bresciane a sud dell'Adamello	Boni & Cassinis (1973)	1:50,000
Studio geologico sulla zona nord-ovest dell'alta Val Sabbia fino al Lago d'Idro	Motta (1953)	1:10,000
CGI Breno	Servizio Geologico d'Italia (1970)	1:100,000
Carta Tettonica delle Alpi meridionali alla scala 1:200.000	Castellarin (1981)	1:200,000

Tab. 4.1 Geological data sources for the harmonized geological map of the study area.

Since geological maps provide 2D information, a major challenge is extending this data into three-dimensional space to construct a reliable 3D geological model. This process requires integrating surface observations with subsurface constraints, such as cross sections. Cross sections serve as deep constraints for the model, revealing the 3D nature of geometric information from 2D geological maps. Therefore, all available geological cross sections from the literature were gathered, reviewed, and their interpretations assessed to ensure consistency. The geological maps by Servizio Geologico d'Italia (2011, 2012, 2016) are furnished with geological cross sections, three of which are related to the study area, though only marginally. Moreover, the map by Motta (1953) features five detailed geological cross sections. These cross sections are not sufficient to adequately constrain the 3D model at depth. Therefore, a dense network of 18 additional cross sections were developed *ad hoc* based on existing geological maps, so that their distribution would not leave parts of the study area uncovered (Fig. 4.2).

These new cross sections are oriented perpendicularly to the general direction of the main geological structure in the area. The inclusion of this information is essential for the calculation of the 3D model.

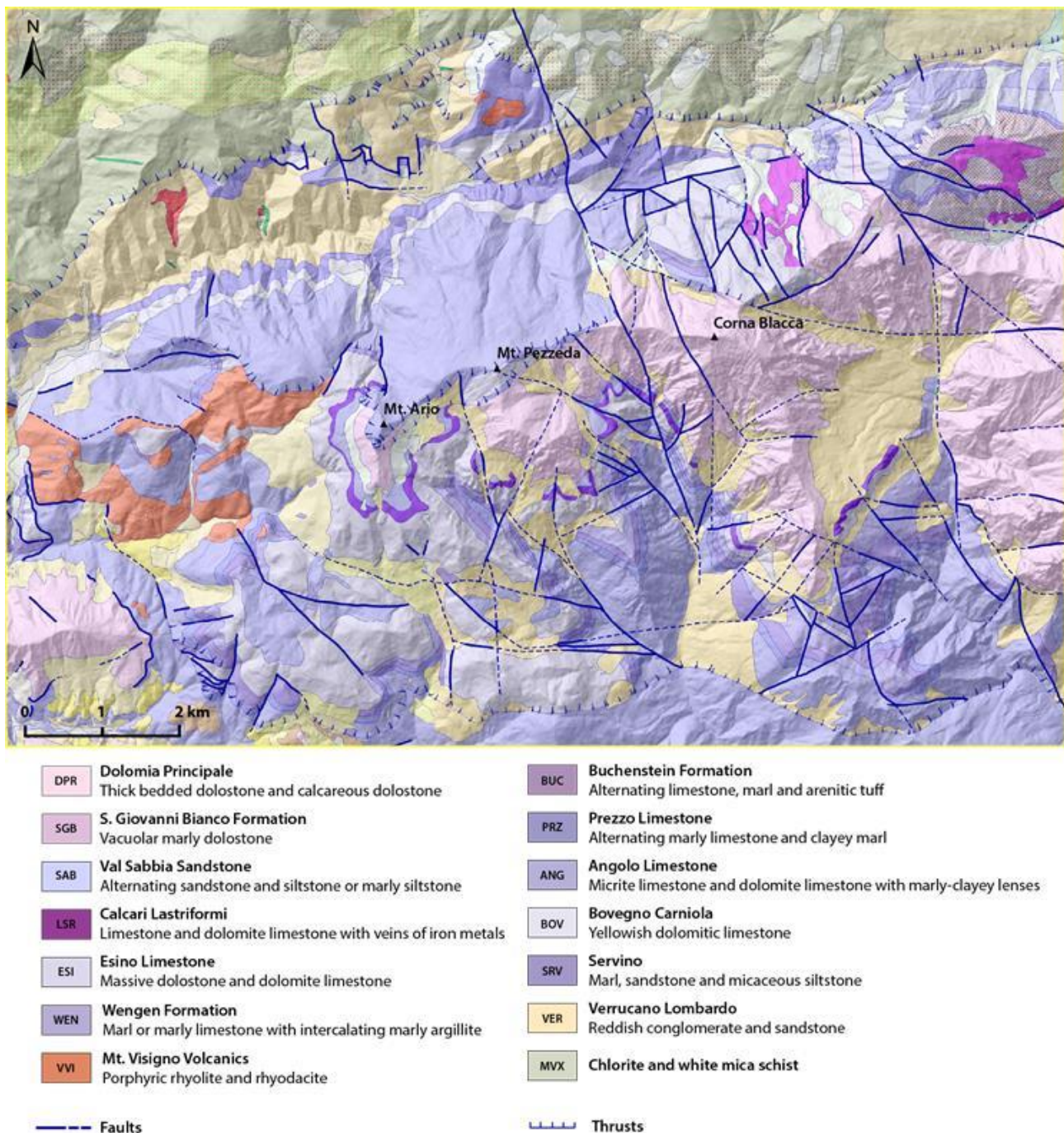


Fig. 4.1 Harmonized geological map of the study area (redrawn after Boni & Cassinis (1973) and Servizio Geologico d'Italia (2011, 2012, 2016)). The legend is normalized to match that of the geological reference maps by Servizio Geologico d'Italia (2011).

As a results, the modelled area is a rectangular prism spanning 13.7 km by 9.6 km by 2.8 km centered on the Mt. Ario – Mt. Pezzeda – Corna Blacca massif (Tab. 4.2).

	Min	Max
X	598057	611836
Y	5066462	5076068
Z	200	3000

Tab. 4.2 3D model extent.

Once properly harmonized and verified, the geological data (maps and cross sections) are imported into GeoModeller, where they serve as superficial and deep constraints to 3D modelling (Fig. 4.3).

At this stage, two important decisions had to be made for the calculation of the model. The first concerned the choice of the reference surface to model, whether the top or the bottom boundary of the geological units. In this case the bottom of each unit was selected as the reference surface. The second decision pertained to the approach for modelling the geological units, whether to assign each unit its own potential field or to group them into a series and represent them using a single potential field. In this case, the modelled units exhibit a lack of concordance, due to the complex geological setting of the study area, preventing their classification as a single series. Consequently, each geological unit is modelled independently using multiple potential fields.

The chronological order of the formations collected in interacting series within a geological pile determines the temporal order of the series within the numerical calculation. Section 4.1.3 describes the geological units that make up the geological pile.

The modelling process incorporates both surface and subsurface constraints.

The surface constraints are derived from mapped geological boundaries and faults, which are directly digitized within the GeoModeller environment. Additionally, some orientation data measured in the field were integrated to enhance model's precision.

Subsurface constraints were established using a dense network of 2D cross sections, where orientation control points were assigned to each geological unit. These cross sections provide critical depth information, ensuring the consistency of the geological framework.

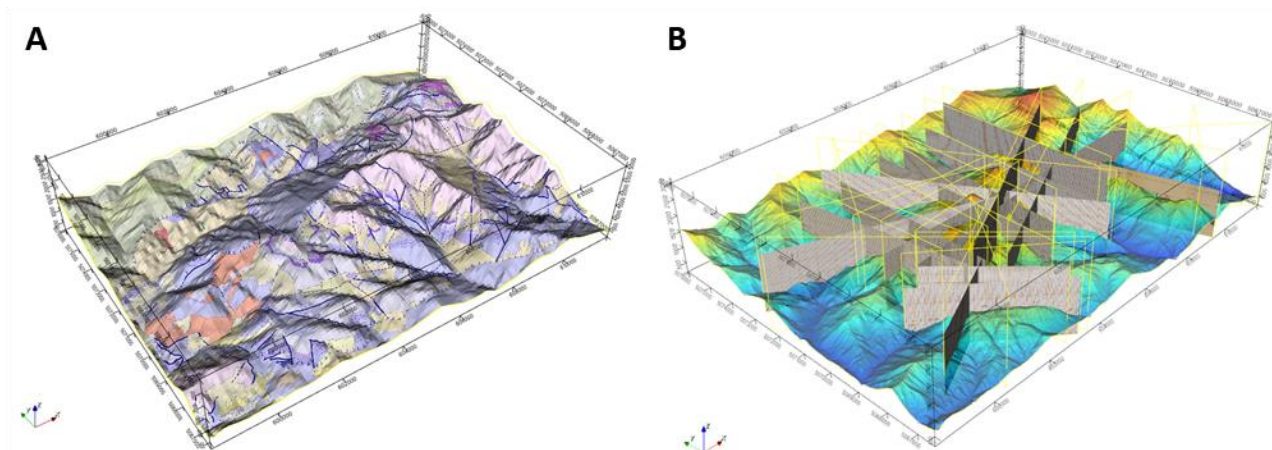


Fig. 4.3 Val Sabbia 3D model setup. A: upload of the DTM and the harmonized geological map. B: upload of the cross sections. Green Y axis in the lower left corner represents the North.

4.1.3 Geological pile definition

As the first step in the modelling process involves defining the rock volumes that constitute the study area (the geological pile using GeoModeller terminology), the understanding of the lithological composition and the geometric relationships between the rock units is essential. Therefore, this Section provides a detailed description of the main lithological characteristics of the outcropping formation used to construct the 3D geological model, as well as their spatial relationships. The outcome of this step is the establishment of the geological pile within the GeoModeller environment.

The lithological and geomorphological characteristics of the formations comprising the Northern and Southern Units (see Fig. 2.4) have been extensively documented in the works of Boni & Cassinis (1973), Rossi (1975), in the unpublished theses of Borromeo (1982) and Maroni (1982), and in the geological reference maps (Servizio Geologico d'Italia, 2011). To further refine the local characterization of each formation, targeted field surveys were conducted. These surveys focused on verifying key geological features, particularly those relevant to hydrostratigraphic and geophysical characterization. The stratigraphic sequence of each structural unit is described below and illustrated in Fig. 2.3.

To simplify the geological model, unconsolidated Quaternary formations were excluded from the representation. This decision was made because, in the study area, these formations have no significant influence on groundwater circulation or storage. Excluding them simplifies the model without compromising its accuracy or the understanding of groundwater dynamics, allowing for a more focused analysis of the formations that directly impact the hydrogeological system. This approach is useful in reducing computational complexity.

The **Northern Unit** consists of Upper Permian to Early Triassic lithotypes. Its southern boundary is marked by the Mt. Ario Line (MAL).

The older terms are represented by the **Verrucano Lombardo** (Upper Permian), clastic rocks, chiefly conglomerates, but also sandstones and siltstones, of a reddish color. This formation crops out in the southern flank of the upper Val Trompia. Its upper boundary is with the Servino. In GeoModeller its base is modelled using the onlap relation.

The **Servino** (Induan – Olenekian) consists of a succession of marls, sandstones, micaceous siltstones with sparse grey limestone or calcareous dolomitic beds. This formation is locally well-known host rock for iron-ore (Cassinis et al., 2007). Actually, Val Trompia has a rich history of mining activities due to the occurrence of strata-bound and vein-type discordant mineralization, which have been exploited since Roman times. In GeoModeller its base is modelled using the onlap relation.

The **Bovegno Carniola** (Lower Anisian) is made up of vacuolar dolomites and yellowish dolomitic limestones; subordinate breccias to dolomitic, calcareous, marly-clayey, calcareous cement elements. It is bounded downwards by Servino and upwards by the Angolo Limestone. In GeoModeller its base is modelled using the onlap relation.

These three formations are not strictly related to the study area, but their inclusion was necessary in order to obtain a reliable 3D model.

The northern unit is capped by the **Angolo Limestone** (Lower – Middle Anisian), which consists of a thick succession of light to dark gray micritic limestones in coarse beds. In its upper part, the formation is characterized by limestones with a wavy bedding surfaces, separated by yellowish silty-sandy terrigenous material (Fig. 4.4A). In the areas of Mt. Ario and the northern slope of Mt. Pezzeda, the Angolo Limestone is intensely fractured and has been thrust southward over the Middle Triassic sequence along the Mt. Ario Line. This thrust coincides with the base of the Northern Unit Angolo Limestone and is modelled in GeoModeller using the erode relation, as it truncates the younger formations.

Angolo Limestone also crops out in the southern unit, at the base of the sequence (Fig. 4.4B). At the watershed between Valle Glera and upper Val Degnone, the Angolo Limestone forms a thick cliff, approximately 600 m high. On this steep slope the Lombardy Speleological Federation (FSLo) documented a cave consisting of a series of interconnected chambers with a total length of 50 m. Due to its limestone composition, the formation could be susceptible to karstification (Tognini, 2011). At the top of the unit, the boundary with the Prezzo Limestone is transitional.

The base of the Southern Unit Angolo Limestone is modelled using the onlap relation in GeoModeller.

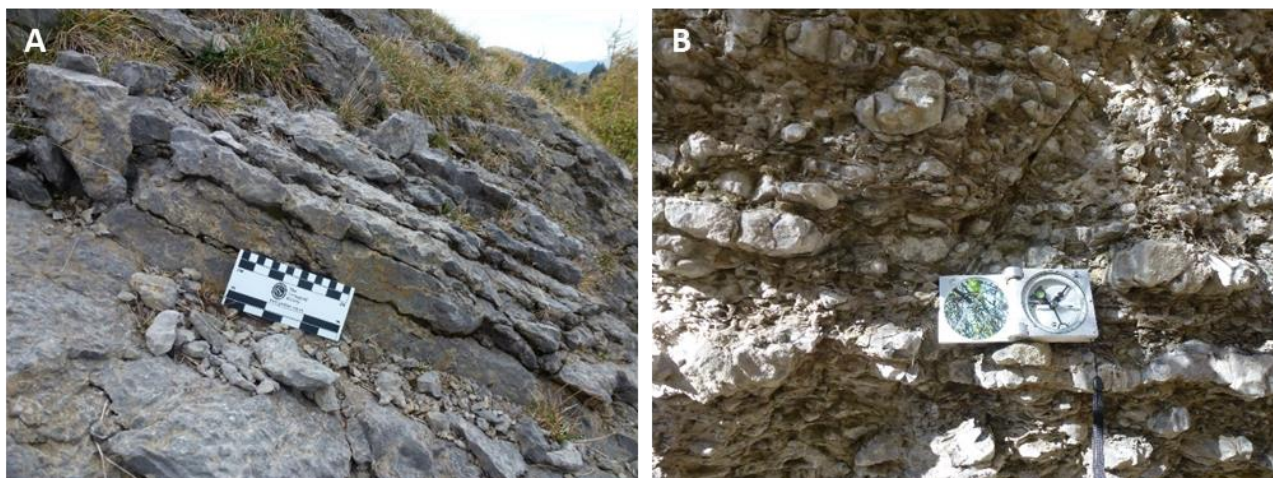


Fig. 4.4 A: Angolo Limestone layers at Mt. Pezzeda. B: Angolo Limestone outcrop in the upper Val Degnone.

The **Southern Unit** includes an extensive Middle-Late Triassic sequence. Its southern boundary is the Ombriano-Forno d'Ono tectonic lineament (OFOL).

The **Dolomia Principale** (Norian) is the youngest unit cropping out in the study area, forming its highest peaks. It consists of light to dark gray dolomites and dolomitic limestones, stratified in thick layers.

In the Mt. Ario area, this formation is reduced due to the influence of the Mt. Ario Line thrust, which descends eastward and causes a progressive thinning of its thickness. At the head of Valle Glera, the formation's continuity is disrupted by the complex fault system of Mt. Pezzeda. Further east, it reappears with increasing thickness, reaching approximately 700 meters at Corna Blacca (Fig. 4.5A).

The carbonate composition of the Dolomia Principale promotes the development of massive, rugged karst pinnacles. These landforms are particularly prominent in the Corna Blacca massif (Fig. 4.5B), but they are also present in the Mt. Ario area.

The formation is characterized by a network of persistent, closely spaced, open fractures oriented E-W, with an inclination of 60° (orientations: 310/62, 300/63, 355/50, 015/63). Additionally, it features steeper fractures with an inclination of 80° , oriented N-S (orientations: 080/80, 065/85, 060/75).

The lower boundary with the San Giovanni Bianco Formation is marked by brecciated layers containing limestone-dolomitic clasts and yellowish calcareous cement, which is sometimes vacuolar and gradually becomes more compact toward the base of the Dolomia Principale.

In GeoModeller its base is modelled using the onlap relation.

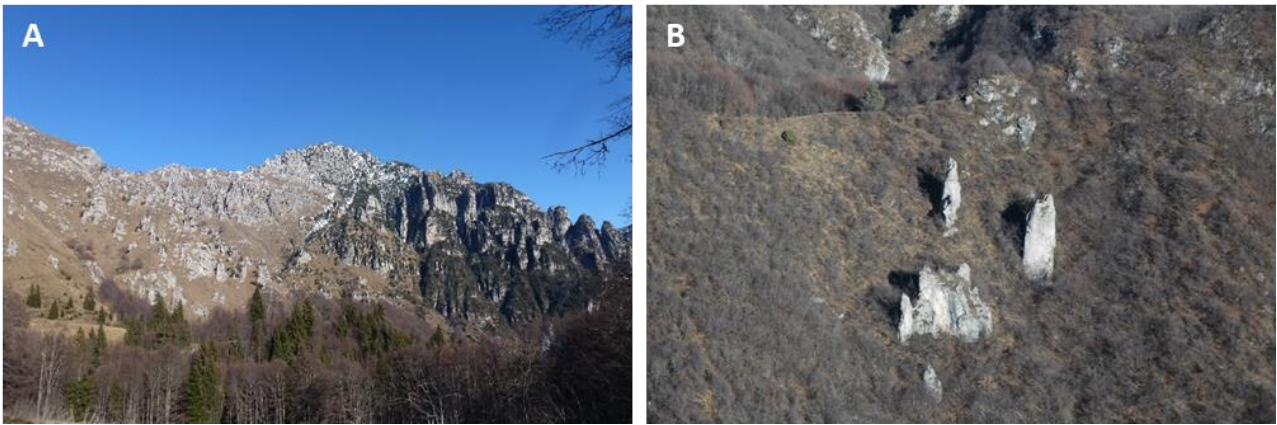


Fig. 4.5 A: Dolomia Principale of the Corna Blacca massif. B: Dolomia Principale karst pinnacles.

The **San Giovanni Bianco Formation** (Upper Carnian) consists predominantly of vacuolar carbonate-sulfate breccias (*“carniole”* Auct.) composed of gray dolomitic clasts and reddish clayey-silty clasts, embedded in a yellowish vacuolar calcareous matrix (Fig. 4.6A). These are interspersed with green and yellowish clayey layers, and in some instances, they exhibit a *fenestrae* structure.

Due to its easily erodible lithology, the outcropping areas of this unit are very limited. The unit forms a thin band surrounding Mt. Ario and extending to the base of Corna Blacca.

The lower boundary with the Val Sabbia Sandstone is gradual, marked by the intercalation of grayish-green clay layers. In contrast, the upper boundary with the Dolomia Principale is defined by a morphological step (Fig. 4.6B). In GeoModeller its base is modelled using the onlap relation.

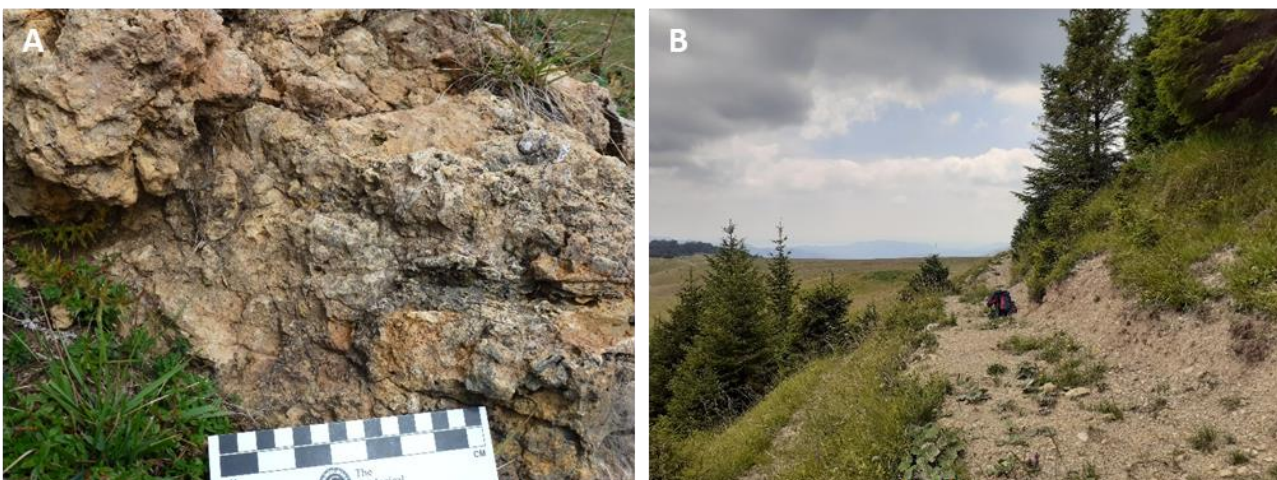


Fig. 4.6 A: a rare San Giovanni Bianco Fm. Outcrop in the area of Mt. Ario. B: the morphological effect of the San Giovanni Bianco Fm around Mt. Ario.

The **Val Sabbia Sandstone** (Lower Carnian) outcrops in several parts of the study area, showing variations in its lithological properties. In general, it consists of layered red and green sandstones, alternating with siltstones and reddish marly siltstones with pervasive fracturing. Around Mt. Ario, this formation has limited outcrops and consists mainly of reddish clayey-silty material highly fractured that contains calcareous nodules joined together to form thin layers (Fig. 4.7A). Here, it forms a gentle, grassy plateau, which is used for summer grazing. This morphology is largely controlled by the high erodibility of the formation and the subhorizontal orientation of its layers, which contribute to the smooth, undulating terrain.

In the Mt. Ario area, sinkholes are present on the outcrop of Val Sabbia Sandstone, ranging in size from a 2 m to approximately 20 m (Fig. 4.7B-C). These landforms likely result from differential dissolution processes, where the soluble Calcari Lastriformi karst bedrock is overlain by the intensely fractured material of the Val Sabbia Sandstone, forming a covered karst system (Waltham et al., 2005).

Moving eastwards towards the Corna Blacca, the sandstones and siltstones are interspersed with more compact conglomerate layers of carbonate composition (Fig. 4.7D).

The lower boundary of the unit is with the Calcari Lastriformi while the upper one is with the San Giovanni Bianco Formation.

In GeoModeller its base is modelled using the onlap relation with two exceptions. As mentioned in Section 2.2.2, the Mt. Ario Line has two splays in which Val Sabbia Sandstone crops out. The basal interfaces of these two portions of the formation, called MAL1 and MAL2, are modelled as Erode.

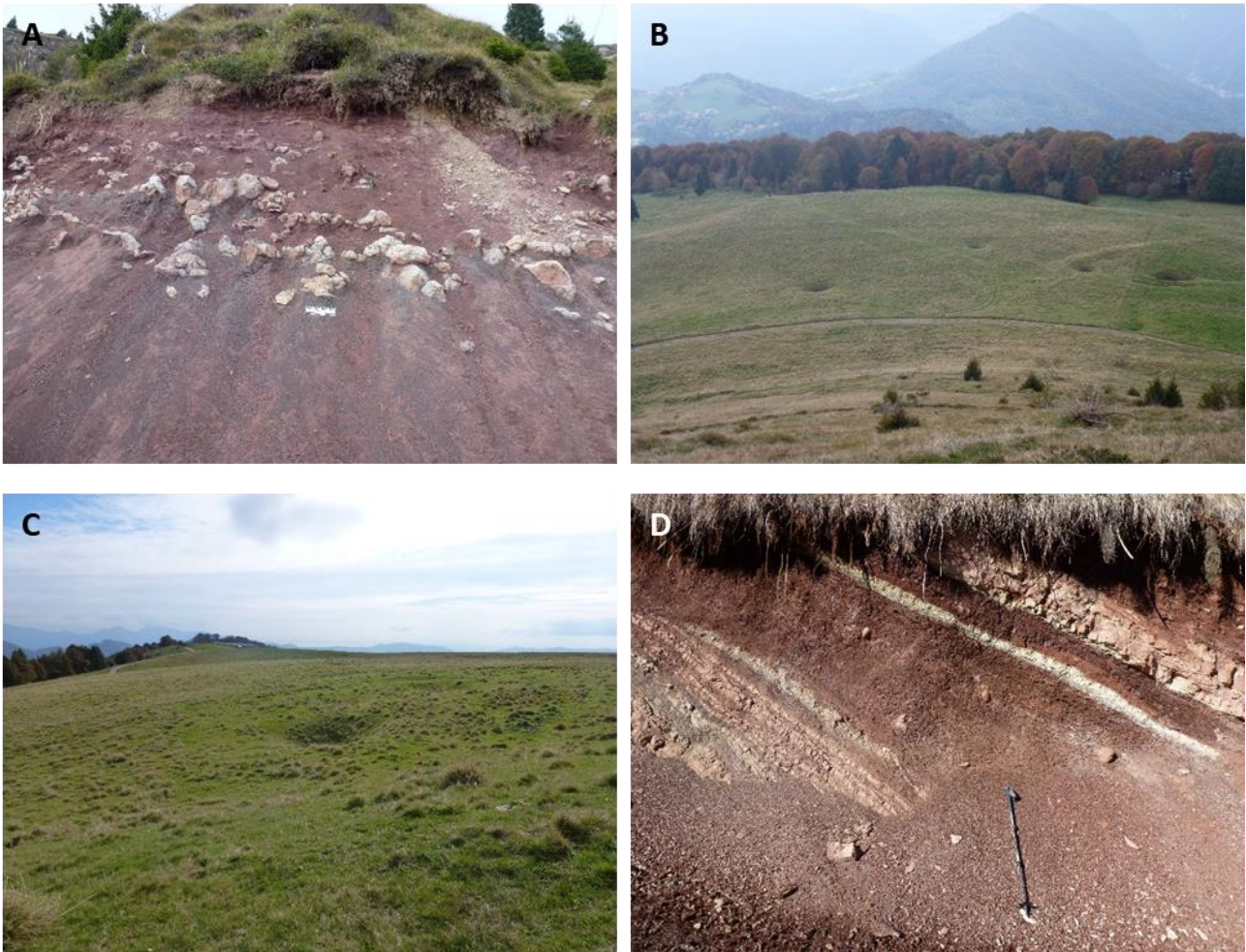


Fig. 4.7 A: Val Sabbia Sandstone outcrop at Mt. Ario. Discontinuous carbonate levels are visible. B-C: sinkholes in the pastures around Mt. Ario. D: alternating layers of different composition in a Val Sabbia Sandstone outcrop in the area of Corna Blacca.

The **Mt. Visigno Volcanics** represents the Carnian magmatism in the Brescian Prealps. These igneous bodies are exposed in the Val d'Irma and consist of porphyric rhyolites and rhyodacites, which were likely emplaced as subaerial flows and subvolcanic bodies (Cassinis et al., 2008; Crisci et al., 1984). Additionally, a conspicuous rhyolite dyke is locally present, intruding the underlying Angolo Limestone.

In GeoModeller its base is modelled using the erode relation, because the emplacement of Mt. Visigno Volcanics crosscut the nearby formations.

The **Calcari Lastriformi** (*"Plattenkalk"* Auct., Lower Carnian) are primarily composed of well-stratified gray to pink limestones, locally mineralized with white calcite and oxidized metallic nodules, with a yellowish weathering appearance (Fig. 4.8A-B).

At Mt. Ario the thin beds of Calcari Lastriformi are intersected by two sets of small but pervasive fractures: one set is subvertical dipping to the W (orientation: 260/85, 280/90, 270/75), while the other dips to the E (orientation: 065/65, 08/62). Moreover, by comparing the geological map and the satellite images, also in this formation sinkholes are recognized.

This formation has good karst potential, due to its limestone composition, but a reduced thickness that limits the possibility of developing large karst cavities (Tognini, 2011).

This formation lies beneath the Val Sabbia Sandstone and overlies the Esino Limestone cliffs. The thickness of the Calcari Lastriformi varies between 30 and 50 m.

In GeoModeller its base is modelled using the onlap relation.

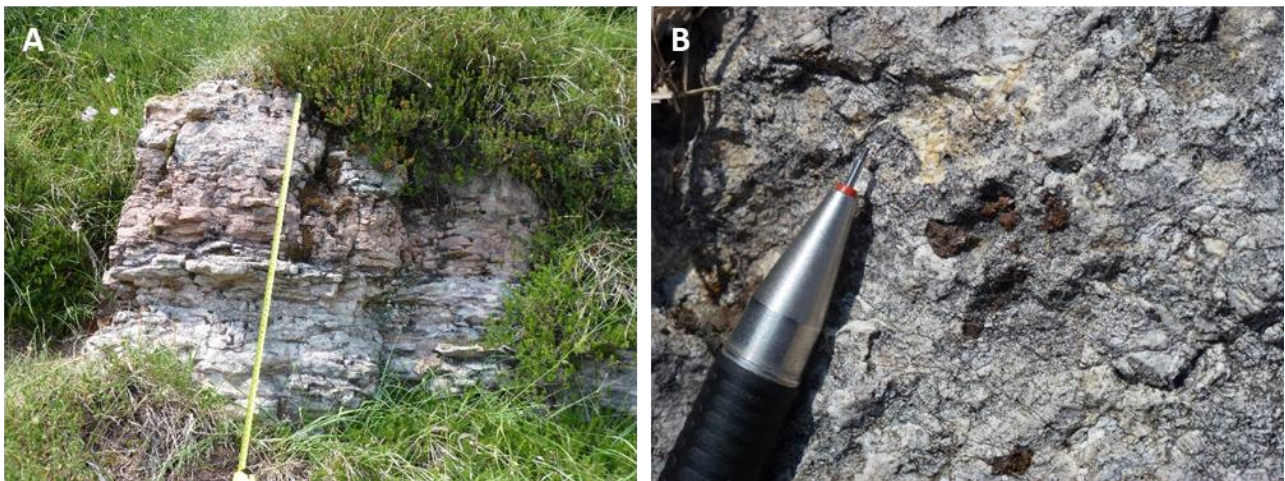


Fig. 4.8 A: pink layers of Calcari Lastriformi outcropping in the Mt. Ario area. B: small metallic mineralization within the Calcari Lastriformi formation.

In the study area, the **Esino Limestone** (Upper Ladinian) consists of dolomites and dolomitic limestones, ranging from whitish to light gray in color, with a massive or faint stratification. They are often coarsely recrystallized with frequent dissolution cavities filled with carbonate or terrigenous materials (Fig. 4.9A).

The Esino Limestone is well known in Lombardy for its high surface and deep karst susceptibility, due to its limestone composition. This characteristic has been extensively documented by the Lombardy Speleological Federation (FSLö) (Tognini, 2011).

Within the study area the Esino Limestone outcrops surround Mt. Ario and partly Corna Blacca. The formation is susceptible to rapid lateral changes in thickness, moving eastward it shrinks considerably until it disappears. This is related to its heteropic relation with the Wengen Formation. In the western part of the study area (around Mt. Ario), the compact and massive structures of the Esino Limestone results in steep reliefs with rugged and imposing slopes (Fig.

4.9B), at the top of which epikarst phenomena occur, with karren and crevasses, the latter also due to gravitational stress release (Fig. 4.9C-D). Moving eastwards, toward Corna Blacca, its thickness decreases and the formation becomes thinner.

The Esino Limestone around Mt. Ario has a massive and fractured structure. The fractures are generally not highly pervasive and can be grouped into two main sets: subvertical fractures dipping to the NE (orientations: 010/80, 020/85, 035/73), fractures dipping to the S (orientations: 165/80, 180/25, 190/20, 205/25).

These differences in thickness, together with the structural characteristics, influence the potential for karst development, affecting groundwater flow and the formation of underground voids. In the absence of fracture permeability, the development of deep karst systems is highly unlikely, as water cannot effectively infiltrate to significant depths. When fracturing is limited, surface karst forms (such as karren) tend to dominate, while the formation of underground cavities remains minimal. Nevertheless, within the Esino Limestone outcropping in the area of Mt. Ario, the Lombardy Speleological Federation reported the presence of “Loèra de Crus” cave, that consists primarily of shafts that reaches a depth of approximately 100 m.

The upper boundary is with the well-stratified Calcari Lastriformi. The lower boundary is either with the Buchenstein Formation in the area of Mt. Ario, and with the Wengen Formation towards Corna Blacca.

In GeoModeller its base is modelled using the onlap relation.



Fig. 4.9 A: detail of a small cavity in the Esino Limestones in the Mt. Ario area. B: Esino Limestones slopes surrounding Mt. Ario. C: karren exokarst formations in the area of Mt. Ario. D: karst crevasses at top of Esino Limestone in the Mt. Ario area.

The **Wengen Formation** (Ladinian) consists of laminated greenish sandstones with volcanic clasts and carbonaceous fragments, alternating with laminated siltstones and green and gray marly siltstones. Upwards, calcarenites are interbedded. Its lateral continuity is interrupted due to its heteropy with the Esino Limestone. Thus, in the area of Mt. Ario this formation is almost absent, while towards the eastern part of the study area, it gradually replaces the Esino Limestone, whose thickness is considerably reduced. The lower boundary of the Wengen Formation is always with the Buchenstein Formation, while the upper boundary is with the Esino Limestone. In GeoModeller its base is modelled using the onlap relation.

The **Buchenstain Formation** (Lower Ladinian) is made of compact dark-gray layered limestones, sometimes with nodules or bands of black chert. It shows local intercalation of sandstone and marls. Characteristic is the presence of pyroclastic horizons of gray-green color.

This formation has a limited thickness. The Lombardy Speleological Federation (FSLo) reported the presence of a cave within this formation: the “Grotta di Cogne” is particularly interesting for its location upstream of the Acqua Bianca spring. This cave is described as a small cavity with a central descending shaft that goes down to a maximum depth of 8 m.

Prezzo Limestone represents the lower boundary of the Buchenstein Formation. The upper boundary can be either with the Wengen Formation or the Esino Limestone. In GeoModeller its base is modelled using the onlap relation.

The **Prezzo Formation** (Upper Anisian) crops out as rhythmical alternation of black marly limestone beds, frequently nodular, with blackish fissile clayey marls. Higher up, the formation is characterized by a calcareous level. Due to the high erodibility of its lithologies, this formation is usually covered by a thick weathering layer. The lower boundary with the Angolo Limestone is gradual and transitional. At the top, the Prezzo Limestone changes to the Buchenstein Formation. In GeoModeller its base is modelled using the onlap relation.

All these formations, each characterized by its specific relationship with the others, collectively form the geological pile in GeoModeller, illustrated in Fig. 4.10A. This pile represents the stratigraphic arrangement of the study area.

Fault systems play a fundamental role in shaping the geological architecture of the study area. The stratigraphic succession in the study area is displaced by both thrusts and faults. As mentioned above, thrusts were treated as erode surfaces in the 3D model. However, modelling faults is more complex, as it requires a specific approach that accounts for the deformation they produce in the affected formations.

Having identified the main faults dislocating the stratigraphic succession in the study area (see structural sketch in Fig. 2.4), it was essential to determine which formations they affected and how (Fig. 4.10B). In addition, their relationships with other faults had to be considered to ensure a consistent structural interpretation (Fig. 4.10C).

To represent their influence, major faults were incorporated into GeoModeller based on their mapped surface traces and inferred orientations from geological maps. Each fault was assigned a defined spatial extent and an influence radius, which determines the zone affected by fault displacement. This approach allows for a more realistic representation of fault-related deformations and their impact on the geometry of geological units.

Although GeoModeller is able to handle this type of information, the number of faults introduced into the model of the study area has been minimized in order to obtain a satisfactory 3D geological model.

Ultimately, the incorporation of 24 fault systems into the 3D geological model resulted in a representation that closely aligns with the available surface constraint data and, to a good approximation, also honors the subsurface constraints.

This refined model captures the geological and structural complexity of the study area, providing a detailed and realistic depiction of stratigraphic relationships and fault interactions.



Fig. 4.10 A: geological pile of the Val Sabbia 3D model (the acronyms of the formations are given in Fig. 4.1). B: a detail of the table that specifies faults and series relationships. C: a detail of the table linking faults with other faults.

4.1.4 Computing the model

As outlined in Section 3.1.2, GeoModeller is based on the potential field method to compute the 3D model. The workflow for constructing the geological model of the study area began with an initial assessment of the series relationships within the geological pile. A key aspect of this process involves testing the hierarchy and interactions of the series to ensure that geological constraints are properly honored.

First modelling iterations focus on the erode interfaces, which in this case correspond to the thrusts and the dykes of the Mt. Visigno Volcanics. Prioritizing these interfaces is crucial for defining their structural hierarchy and geometrical relationships within the model. This step allows to choose their polarity and position within the geological pile.

Once the erode interfaces were successfully tested and validated, the next phase focuses on the onlap series. Initially, these surfaces were assessed individually to verify their spatial distribution within the calculation domain and their fit with geological constraints. Subsequent iterations refined their position in combination with the other series, ensuring they filled the expected spaces reflecting the stratigraphy of the area.

The final step involved incorporating faults into the model. Since all faults in the model are treated as finite faults, to adjust their geometry and influence it is required a trial-and-error tuning of the ellipsoid parameters that define their spatial extent and displacement characteristics. This iterative process results in a more realistic representation of fault-induced deformations within the geological pile.

To further constrain areas with scarce data and higher uncertainty within the domain, 7 additional cross-sections were constructed directly in GeoModeller environment as ancillary constraints, helping to refine the 3D model of the study area's geology at depth.

GeoModeller's implicit modelling approach, which employs the co-kriging interpolation of both interface and orientation data, inherently prevents the unintended overlap of interfaces.

Once the interfaces are calculated, they are rendered using the marching cubes algorithm (Lorensen & Cline, 1987), which enables the visualization of both individual surfaces and full volumetric representations of the subsurface geology.

A best practice in GeoModeller is to introduce new geological data incrementally, following the guideline that "less is better". This approach minimizes the introduction of unnecessary constraints, allowing for a more rapid computation.

Concerning the faults, to simplify the computational process, it is considered good practice to introduce faults one at a time, gradually refining the model until it accurately represents the geological complexity of the area.

The result is a highly detailed and realistic 3D representation of the architectural structure of the rock volumes Fig. 4.12, accurately depicting their spatial distribution, geometry, and relationships. This model provides insights into the subsurface structure, aiding in geological interpretation.

Fig. 4.11 Map of the modelled geology (legend is given in Fig. 4.10A). Red lines represent the modelled faults. Yellow lines are the cross section traces.

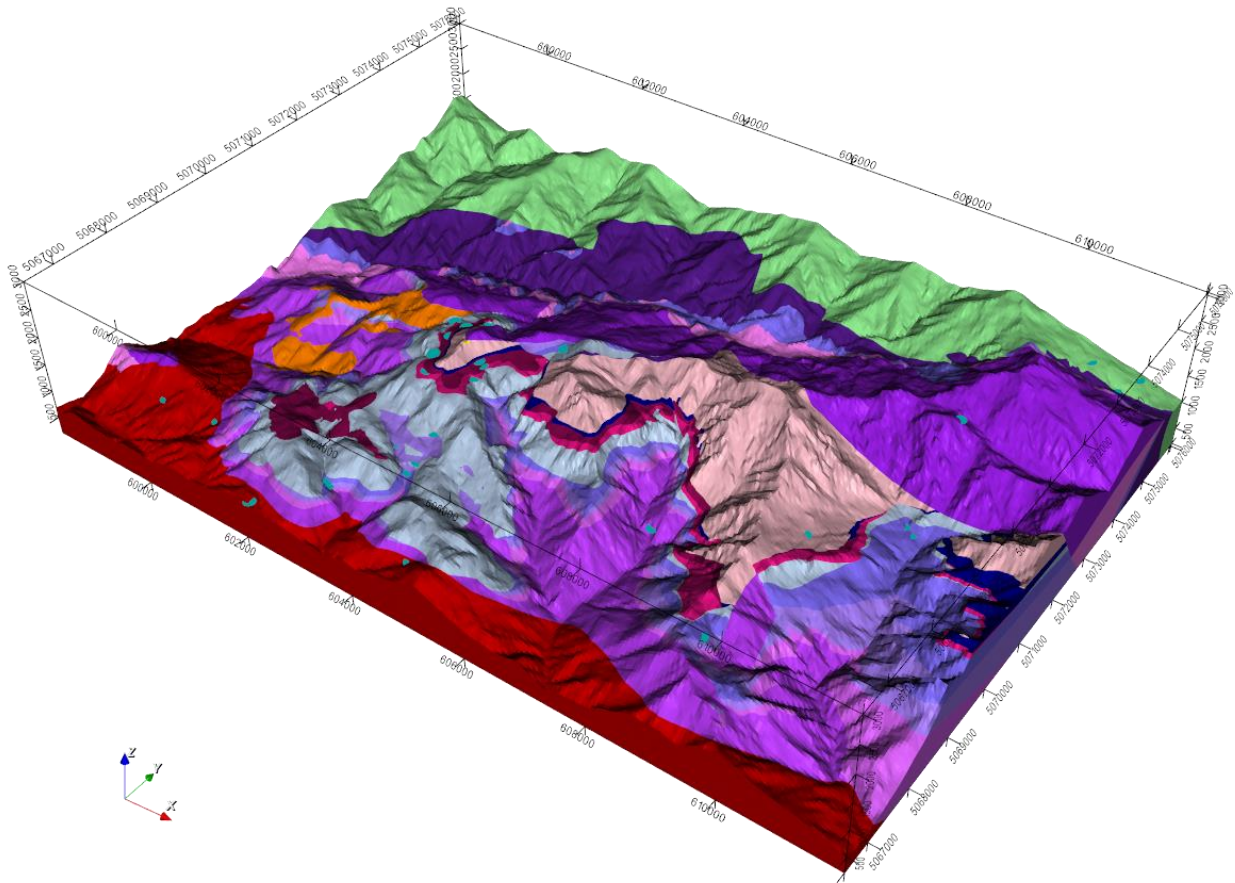


Fig. 4.12 A visualization of the 3D geological model. The blue circles represent the location of all the springs emerging in the study area. The legend is provided in the geological pile (Fig. 4.10A). Y axis is the North.

4.1.5 Uncertainties of the model

The use of GeoModeller for 3D geological-structural modelling in a complex geological setting presents certain challenges and uncertainties.

One of the most significant challenges is the modelling of faults and heteropic formations. Regarding faults, since these discontinuities are not always clearly visible in the field, only surface contact data with a constant dip (using the "dip constant" function in GeoModeller) were used to avoid introducing arbitrary deep constraints unsupported by direct evidence. Additionally, as mentioned in Section 4.1.3, the extent and depth of each fault were determined through a trial-and-error approach, testing different configurations and selecting the one that best reproduces the observed geology. However, this process inherently introduces a degree of uncertainty into the model.

In GeoModeller, handling heteropic relationships between geological formations presents a challenge, as the software does not inherently support lateral facies transitions. To address this

limitation, that in this project regards only the Esino Limestone and the Wengen Formation, a workaround must be implemented to correctly model the pinch-out of formations within the model.

The approach used involves manipulating the orientation constraints assigned to the heteropic formations. By adjusting the order in which these formations are introduced in the geological pile, it is possible to force their termination in a geologically plausible manner. Specifically, the orientation constraints of the heteropic formations can be inverted Fig. 4.13, guiding the model to create a pinch-out geometry where the formations laterally transition into each other.

Iterative adjustments and validation against map data and cross-sections help in refine the model to reflect the expected geological relationships.

Another source of uncertainty in the modelling process is related to the lack of deep geological constraints. Actually, in the model, no deep orientation data were added to prevent excessive reliance on subjective interpretations, which could introduce bias into the results.

The absence of deep constraints means that, in some cases, the model may generate unrealistic geometries at depth. These artifacts arise due to the algorithm's interpolation between limited surface and shallow subsurface data, highlighting the challenge of balancing interpretative freedom with geological plausibility in 3D modelling.

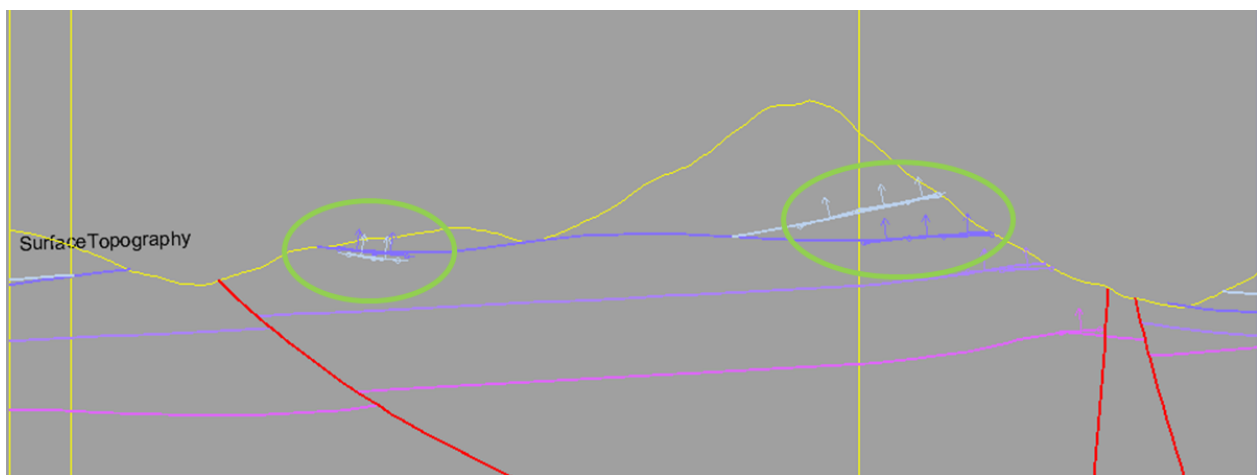


Fig. 4.13 Circled in green are two examples of artefacts used to model heteropic contacts between geological formations in a cross section in GeoModeller. The order of the formation's orientation data is reversed to allow the heteropic formation to pinch-out.

4.2 Hydrogeological characterization

4.2.1 Hydrochemical characterization

Historical springs chemistry data are provided by A2A, the water network manager, for the five springs of interest. Specifically, data were available for the period 2010–2021 for the Acqua Bianca and Frane/Stabol/Acqua Fredda springs, while for the Pian del Bene spring analyses covered the years 2010–2018, and for Pagari springs it spans between 2010–2017. However, for the Frane/Stabol/Acqua Fredda springs, water samples were taken from a shared collection tank. This sampling prevents any distinction between the individual contributions of the three springs, making it impossible to assess potential differences in their chemical composition.

In addition, a water sampling campaign was carried out in July 2023 in cooperation with A2A to perform new analyses of major ions and stable isotopes. During this campaign the water of all the springs were sampled.

The results of the major element analyses are presented in the Piper diagram shown in Fig. 4.14. This diagram provides a classification of the chemical composition of each water sample and enables a direct comparison between all samples. These help to appreciate the similarities and differences in composition among the analyzed springs. Further details on the interpretation of the diagram are provided in Section 3.2.2.

All five spring waters show the same chemical characteristics, with low variability among the years. In the cation ternary plot all the points are shifted toward the Ca-Mg side, with Mg^{2+} percentages around or lower than 50%. In the anions ternary plot, the CO_3^{2-} dominates the water chemistry. Only the Acqua Bianca spring shows a variability in the Ca^{2+} content. All spring waters are classified as calcium-bicarbonate facies.

In terms of hydrogeological implications, this facies suggests that the groundwater has interacted primarily with carbonate rocks, dissolving calcium carbonate ($CaCO_3$) and/or calcium magnesium carbonate ($CaMg(CO_3)_2$) minerals along its pathway.

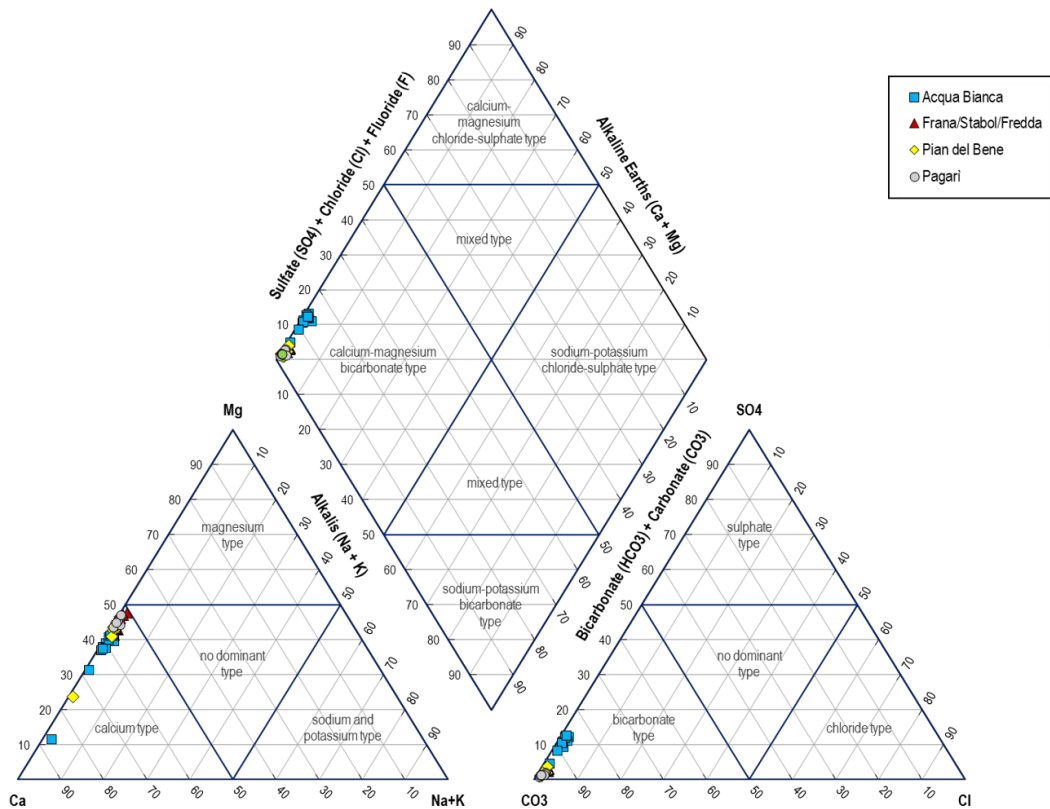


Fig. 4.14 The Piper diagram indicates that the hydrochemical facies of the spring water is of the calcium-bicarbonate type.

To enhance the interpretation of groundwater circulation dynamics, additional parameters, including water temperature, electrical conductivity, pH, and isotopic composition, should be analyzed alongside the geological and structural characteristics of the aquifer system.

Regarding the monitoring of water physical parameters, at the springs equipped with multiparameter probes (i.e., Acqua Bianca and Frane), data collection is carried out continuously at 30-minute intervals throughout the reference period (2022–2024).

In contrast, the Pian del Bene and Pagari springs are not equipped with monitoring probes. However, historical EC and pH data from annual sampling campaigns conducted by A2A in the period 2010–2018 for the former and 2010–2017 for the other spring have been collected. For these two springs no information about the temperature is available. Although these datasets have a lower temporal resolution compared to continuous monitoring, they still contribute to the understanding of the hydrogeological behavior of the spring.

Statistical analyses have been performed on all collected datasets to assess trends, variability, and potential correlations among different parameters, supporting the evaluation of the hydrogeological dynamics of each spring.

The results for Acqua Bianca, Frane, Pian del Bene and Pagari are shown in Tab. 4.3, Tab. 4.4, Tab. 4.5 and Tab. 4.6 respectively.

Parameter	Unit	Arithmetic mean	Min	Max	Median	Standard deviation
T	°C	7,99	7,56	8,28	8,01	0,15
EC	(25°C) μS/cm	299,59	225,88	315,86	303,02	12,71
pH	pH units	8,15	5,68	8,78	8,39	0,84

Tab. 4.3 Acqua Bianca half-hourly data in the period 2022-2024 recorded by the Aqua TROLL probe.

Parameter	Unit	Arithmetic mean	Min	Max	Median	Standard deviation
T	°C	6,94	6,38	8,53	6,91	0,33
EC	(25°C) μS/cm	283,14	253,19	332,40	281,95	4,74
pH	pH units	8,52	7,96	8,90	8,56	0,18

Tab. 4.4 Frane half-hourly data in the period 2022-2024 recorded by the Aqua TROLL probe.

Parameter	Unit	Arithmetic mean	Min	Max	Median	Standard deviation
EC	(25°C) μS/cm	326,44	282,00	365,00	313,00	29,00
pH	pH units	7,83	7,40	8,20	7,80	0,23

Tab. 4.5 Pian del Bene yearly data in the period 2010-2018.

Parameter	Unit	Arithmetic mean	Min	Max	Median	Standard deviation
EC	(25°C) μS/cm	337,63	318,00	364,00	332,00	17,72
pH	pH units	7,61	7,20	7,81	7,70	0,22

Tab. 4.6 Pagari yearly data in the period 2010-2017.

For the Acqua Bianca spring (748 m), the narrow temperature range suggests that groundwater is unaffected by fluctuations in surface air temperature. The range of EC and its standard deviation indicates moderate variations in dissolved ion concentrations, likely influenced by changes in the water flow regime. The average pH reveals neutral to slightly alkaline water.

In contrast, the Frane spring (1601 m) displays a broader temperature range than Acqua Bianca, with a lower minimum temperature, likely due to its higher elevation. The mean EC

reflects a moderate level of dissolved ions, with a relatively small standard deviation indicating stable mineral content. The average pH confirms alkaline conditions. The stability of temperature, EC, and pH supports the presence of a well-established groundwater flow path. For the Pian del Bene spring (1509 m), the mean EC suggests a concentration of dissolved ions comparable with the other springs. The high standard deviation highlights a higher variability in mineral content with respect to the other springs. This elevated EC may indicate prolonged water-rock interaction or extended residence time within the aquifer. The mean pH indicates slightly alkaline conditions.

Finally, for the Pagari spring (1479), the average EC also reflects a relatively high level of dissolved ions, with moderate variability. The mean pH suggests slightly alkaline water, with minimal fluctuations, as confirmed by the low standard deviation.

To conclude, the calcium-bicarbonate facies observed in the springs of the Mt. Ario area (Frane, Pian del Bene and Pagari), combined with their spatial proximity and the similar hydro-structural conditions at their emergence points, strongly suggests that these springs are part of a single groundwater circulation system.

Regarding the Acqua Bianca spring, although it is located further east compared to the Mt. Ario springs, its hydrochemical composition is analogous. This similarity indicates that its groundwater flow path likely occurs within the same aquifer formations.

4.2.2 Environmental isotope characterization

Two spring water sampling campaigns were conducted in the study area, one in September 2022 and the other in July 2023, and isotopic analyses of these water samples were performed by IT²E company using an isotope ratio mass spectrometer (IRMS).

The two sampling campaigns involved the six springs of interest and other seventeen springs issuing in the study area.

In Tab. 4.7 and in Tab. 4.8 the isotopic data measured at springs are reported for the two campaigns. The error on the measurements of $\delta^{18}\text{O}$ is $\pm 0.2\text{‰}$ and the one on $\delta^2\text{H}$ is $\pm 1\text{‰}$.

The $\delta^{18}\text{O}$ values range from -9.96‰ to -8.36‰ , while $\delta^2\text{H}$ values range from -64.4‰ to -54.1‰ . The samples of 2023 exhibit more depleted values, indicating a significant variation in the isotopic signal. This strong fluctuation is attributed to an intense precipitation event (40 mm) that occurred just four days before the 2023 sampling (Fig. 4.15B). In contrast, in 2022, the last major precipitation event before sampling took place seven days earlier, with a total of 34 mm, likely having little to no impact on the isotopic signal (Fig. 4.15A).

Fig. 4.16 reports the local relationship between $\delta^{18}\text{O}$ and $\delta^2\text{H}$ composition of spring water samples and compared them with the Local Meteoric Water Line (LMWL) for Northern Italy by Longinelli & Selmo (2003) and with the Global Meteoric Water Line (GMWL) by Craig (1961a). From the linear distribution of the water samples, the meteoric origin of groundwater is inferred.

Most of the samples fall above the LMWL, indicating more negative isotopic values than those contributing to the theoretical recharge calculated for Northern Italy. This peculiarity accounts for the enrichment in light isotopes typical of recharge areas at higher elevation (the “altitude effect” described in Section 3.2.3). The linear regressions of the samples show a different slope with respect of GMWL and LMWL, indicating that the precipitation that originates the spring waters comes from a likely local recharge, not influenced by global atmospheric circulation, but rather influenced by local climatic and topographic conditions.

SPRINGS NAME	SAMPLING DATE	ELEV	$^{18}\text{O}/^{16}\text{O}$ vs V-SMOW2	$^2\text{H}/^1\text{H}$ vs V-SMOW2
		m	delta x 1000	delta x 1000
Acqua Fredda 2	08/09/2022	1447,6	-8,47	-54,2
Acqua Bianca	12/09/2022	748,6	-9,29	-59,2
Pian del Bene	08/09/2022	1509,1	-9,19	-60
Pagari 4	08/09/2022	1479,5	-9,24	-60,1
Stabol 3	08/09/2022	1510,4	-9,21	-59,7
Frana 1 - 2 - 3	08/09/2022	1601,3	-9,3	-60,2

Tab. 4.7 Results of the isotopic analyses on the water samples collected in 2022.

SPRINGS NAME	SAMPLING DATE	ELEV	$^{18}\text{O}/^{16}\text{O}$ vs V-SMOW2	$^2\text{H}/^1\text{H}$ vs V-SMOW2
		m	delta x 1000	delta x 1000
Acqua Fredda 2	11/07/2023	1447,6	-9,66	-61,8
Acqua Bianca	12/07/2023	748,6	-9,89	-64
Pian del Bene	11/07/2023	1509,1	-9,92	-63,6
Pagari 4	11/07/2023	1479,5	-9,96	-64,3
Stabol 3	11/07/2023	1510,4	-9,86	-63,4
Frana 1 - 2 - 3	11/07/2023	1601,3	-9,93	-64,6

Tab. 4.8 Results of the isotopic analyses on the water samples collected in 2023.

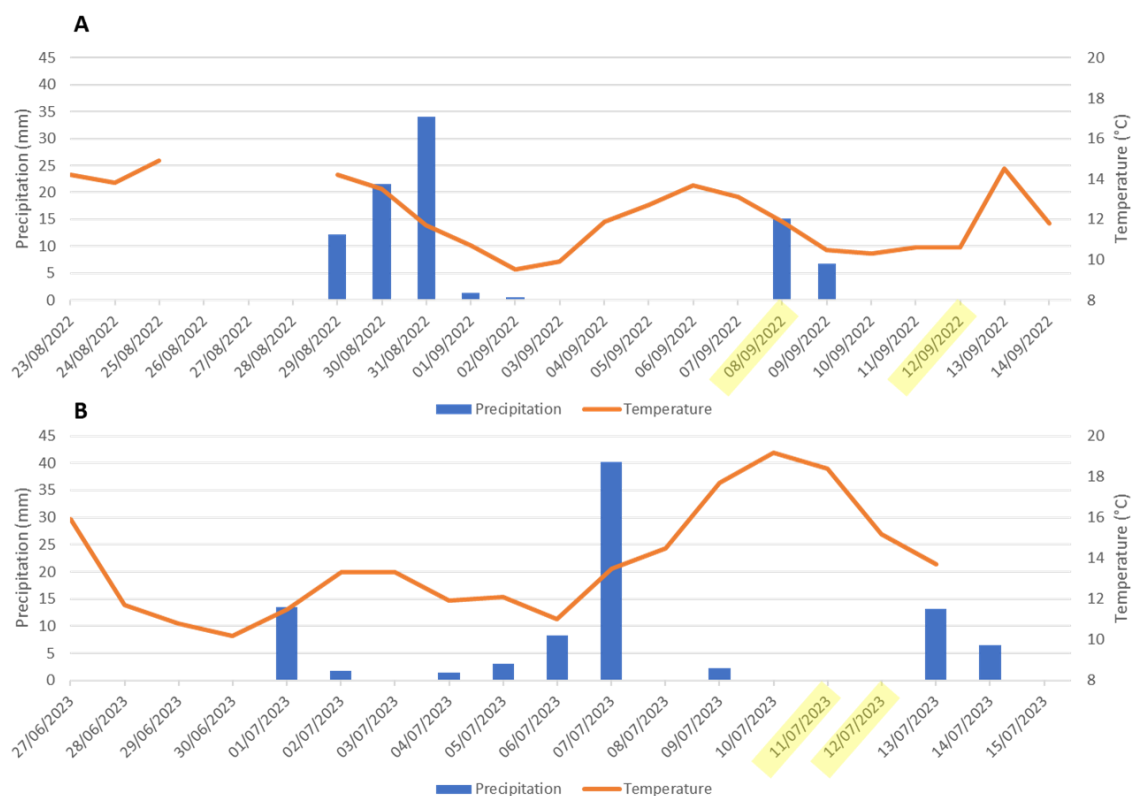


Fig. 4.15 Precipitation and temperature trends 15 days before the samplings in 2022 (A) and 2023 (B). The sampling days are highlighted. During 2022 campaign, temperatures were in the range 10.9 – 11.6 °C, and it rained moderately on the first day of sampling. In contrast, during 2023 campaign, temperatures were higher, 18.4 – 15.2 °C, and all sampling days were dry.

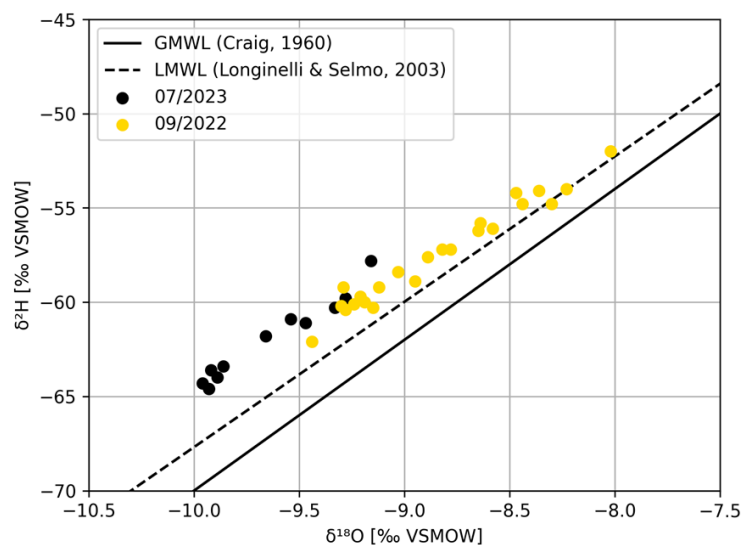


Fig. 4.16 Local relationship between $\delta^{18}\text{O}$ and $\delta^2\text{H}$ of all the 23 sampled springs in the study area plotted with respect of the Global Meteoric Water Line and the Local Meteoric Water Line for Northern Italy.

In addition, for each spring, an approximate recharge elevation range was calculated from a vertical isotope gradient currently available in the literature, as stated in Section 3.2.3. Particularly, in the study area the one determined by Longinelli & Selmo (2003) at Passo della Presolana – Darfo Boario was used, that is -0.1‰/100 m.

To estimate the uncertainties on the recharge elevations, the error on the vertical gradient, the error on the $\delta^{18}\text{O}$ measurement at the reference sites, and the error on the $\delta^{18}\text{O}$ measurement at the investigated springs were considered. The results are reported in Tab. 4.9.

SPRING NAME	SAMPLING DATE	ELEV	RECHARGE ELEV	MIN RECHARGE ELEV	MAX RECHARGE ELEV
			m	m	m
Acqua Fredda 2	08/09/2022	1447,6	1010	802	1218
Acqua Bianca	12/09/2022	748,6	1830	1622	2038
Pian del Bene	08/09/2022	1509,1	1730	1522	1938
Pagari 4	08/09/2022	1479,5	1780	1572	1988
Stabol 3	08/09/2022	1510,4	1750	1542	1958
Frana 1 - 2 - 3	08/09/2022	1601,3	1840	1632	2048

Tab. 4.9 Estimated recharge elevation ranges for the springs of interest.

The estimated recharge elevation aligns with the height of the surrounding ridges and peaks for most of the springs. A detailed examination of the six springs of interest is presented below. The Acqua Bianca mean recharge elevation is 1830 ± 208 m, a range that is compatible with Corna Blacca (2005 m) massif and Mt. Pezzeda (1799 m), Mt. Frondine (1791 m) and Mt. Tigaldine (1765 m) range (Fig. 4.17). The variability in its recharge elevation suggests the possibility of a recharge source from the west, rather than exclusively from the overlying Corna Blacca, which is the more likely recharge area. This hypothesis is not only supported by isotopic analysis but is also reinforced by the presence of a fault system, with the Pofferatte Fault being the most significant. This fault, which outcrops near the spring, may influence groundwater circulation by acting as a drainage pathway that channels groundwater toward the spring.

Furthermore, as described in Section 4.1.3, moving westward, the thickness of the Esino Limestone increases. Since this formation has significant karst potential, where it is thicker it is more likely that dissolution processes play a crucial role in groundwater flow.

On the other hand, the Frane springs emerge within a thrust splay involving the Esino Limestone, Wengen Formation, and Buchenstein Formation, in a tectonically complex area south of Mt. Pezzeda.

The estimated recharge elevation, calculated using the vertical isotope gradient, indicates a mean altitude of 1840 m, which exceeds the elevation of the surrounding topography. However,

considering the uncertainty in the recharge elevation estimate, the reliefs of Dosso Falcone and Mt. Pezzeda remain the most probable recharge areas for this group of springs.

The Pian del Bene, Pagari, and Stabol springs emerge at slightly different elevations around Mt. Ario (1755 m), which represents the most likely recharge area for all three. This hypothesis is supported by the recharge elevations estimated using the vertical isotope gradient, taking into account the error on the estimation: 1730 m for Pian del Bene, 1780 m for Pagari, and 1750 m for Stabol. The consistency of these values suggests that the groundwater feeding the three springs originates from a common recharge area.

Concerning Acqua Fredda spring, although it also emerges around Mt. Ario, the estimated recharge elevation is unreliable (1010 m), as it is lower than the actual elevation of the spring itself. This discrepancy may be due to the poor quality of the water sample analyzed. In fact, the isotopic analysis shows a less negative isotopic values compared to the sample from the nearby Stabol spring. During the sampling campaign, it was reported that the sample collected from Acqua Fredda appeared turbid. Turbidity, which can result from both organic matter and suspended clay particles, may alter the results of the isotopic analysis through various physical and chemical processes. These include the adsorption of isotopes onto suspended particles, fractionation effects, and potential contamination from surface runoff. Nevertheless, it is reasonable to assume that the potential recharge area of the spring is the same as that of the other springs surrounding Mt. Ario.

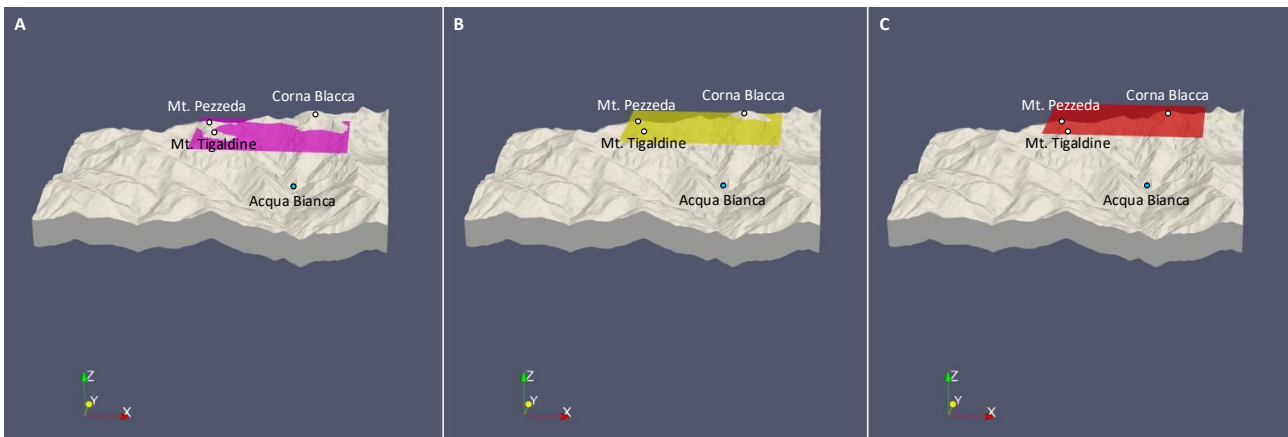


Fig. 4.17 Range of recharge elevation estimated with the vertical isotope gradient for the Acqua Bianca spring. A: Minimum elevation. B: mean elevation. C: maximum elevation.

4.2.3 Hydrodynamic characterization of springs

Since the Acqua Bianca is the most fruitful spring of the study area, devices for continuous recording of water level have been installed by A2A, as detailed in Section 2.2.4.

Unfortunately, this spring is located in an area without cellular network coverage, making real-time data transmission impossible. This limitation has affected the completeness of the dataset, as there have been multiple instances where, upon manually downloading the data, sometimes the probes were found switched off due to battery depletion. As a result, gaps occurred within the time series. To address this issue, the reference period chosen was from 01/10/2023 to 01/10/2024, as it proved to be the most complete in terms of data. For comparison, the hydrograph for the period 01/10/2022-01/10/2023 for the Acqua Bianca spring is also provided, but it lacks a lot of data, both in terms of discharge measurements and precipitation records (Fig. 4.19).

The hydrograph of the springs shows how the mean daily discharge reacts to cumulated daily precipitation over the specified period. In this case it reveals different trends in the measured discharge values. The observed discharge variations are strongly influenced by the local meteorological conditions, with discharge patterns closely following precipitation events.

Precipitation data were downloaded from the monitoring weather station installed by A2A in the Mt. Ario area, where some springs emerge (Fig. 4.18). The rain gauge is able to distinguish the type of precipitation. The analysis of the dataset for the period of interest indicates that rainfall is the main type of precipitation (83% of all the precipitation events). Although snowfall did occur in the area, time constraints did not allow to consider its effect on groundwater recharge in this study.

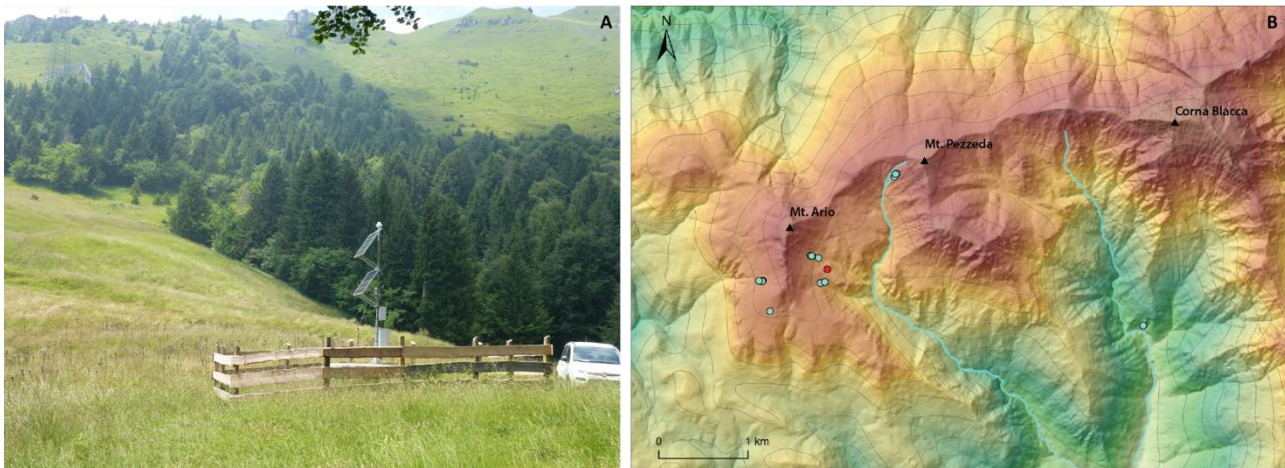


Fig. 4.18 Weather station (A) and its location with respect to the springs (B).

The hydrograph of the Acqua Bianca spring is reported in Fig. 4.19 and Fig. 4.20. The discharge value of the spring is an estimate, as it is calculated based on the water level measured in the collection tank of the Acqua Bianca spring, considering the tank's geometry. However, as shown in Fig. 2.9A, the tank's shape is irregular, requiring some simplifications in the estimation

process. Additionally, the reported discharge corresponds to the water captured by the spring intake structure, which represents only a portion of the total groundwater emerging at the site. In fact, alongside the intake structure, groundwater also flows directly from the rock, forming a natural waterfall (see Fig. 2.9D). Descriptive statistics on the discharge values are presented in Tab. 4.10.

Parameter	Units	Arithmetic mean	Median	Mode	Standard Deviation	Min	Max
Q	m ³ /s	0,043	0,03	0,03	0,038	0,002	0,344

Tab. 4.10 Descriptive statistics of Acqua Bianca spring discharge.

The graphical analysis of the hydrograph displays significant fluctuations in discharge throughout the year. Sharp discharge peaks occur following rainfall events, suggesting that the spring is primarily fed by direct meteoric water infiltration. Rainfall is quickly absorbed into the subsurface and channeled toward the spring due to the presence of a well-developed network of fractures and/or karst conduits within the carbonate lithologies. As a result, once the input from infiltration stops, spring discharge also declines rapidly, returning to the baseline levels observed prior to the precipitation event.

The retention curve after rainfall events shows a smooth, exponential-like decay. The presence of this residual discharge between rainfall events suggests that, in addition to a rapid discharge component, there is also a slower discharge component, likely associated with small, interconnected fractures in the rock matrix.

The variations in discharge over time were compared with changes in the main hydrochemical parameters (e.g., EC, T and pH) (Fig. 4.21).

An inverse correlation between discharge and electrical conductivity (EC) is observed. During peak discharge events, poorly mineralized meteoric water enters the aquifer system diluting the overall ion concentration. In addition, the rapid response of discharge to precipitation events reduces the interaction between recharge waters and surrounding rocks, thus failing to enrich itself with ions, and its electrical conductivity remains lower. This pattern is typical of springs fed by aquifers characterized by high permeability zones.

Temperature data further support this behavior. These data maintain relative stability throughout the reference period but exhibit sharp declines at discharge peaks. This phenomenon is linked to the contribution of colder meteoric water.

Also the pH remains relatively stable over time. Some minor fluctuations could be associated with variations in water chemistry due to rock interaction or recharge events involving slightly more acidic waters.

In conclusion, Acqua Bianca spring is fed by an aquifer with dominant drainage behavior, according to the classification promoted in Vigna & Banzato (2015). This type of aquifer is characterized by high permeability zones, probably related to intense and localized fracturing and/or conduits presence due to karstification, within a rock volume characterized by interconnected discontinuities.

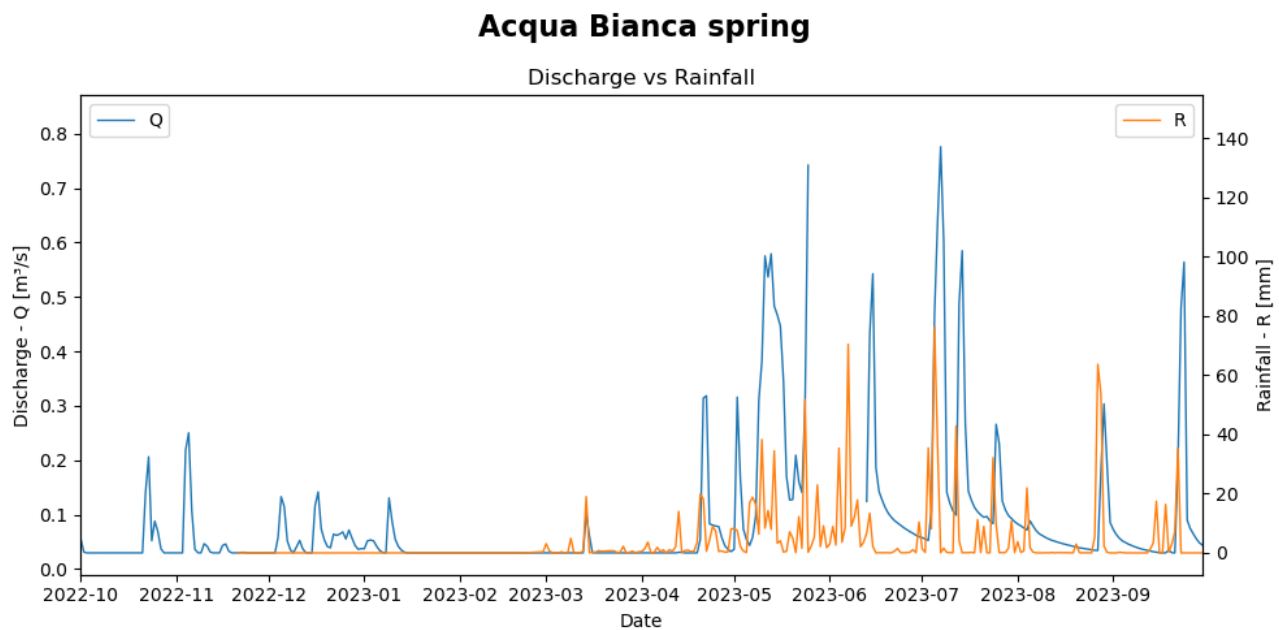


Fig. 4.19 Hydrograph of the Acqua Bianca spring in the period 01/10/2022 – 01/10/2023. In orange is the cumulative daily rainfall measured at Mt. Ario weather station. In blue is the average daily discharge calculated from the water level measured at the spring

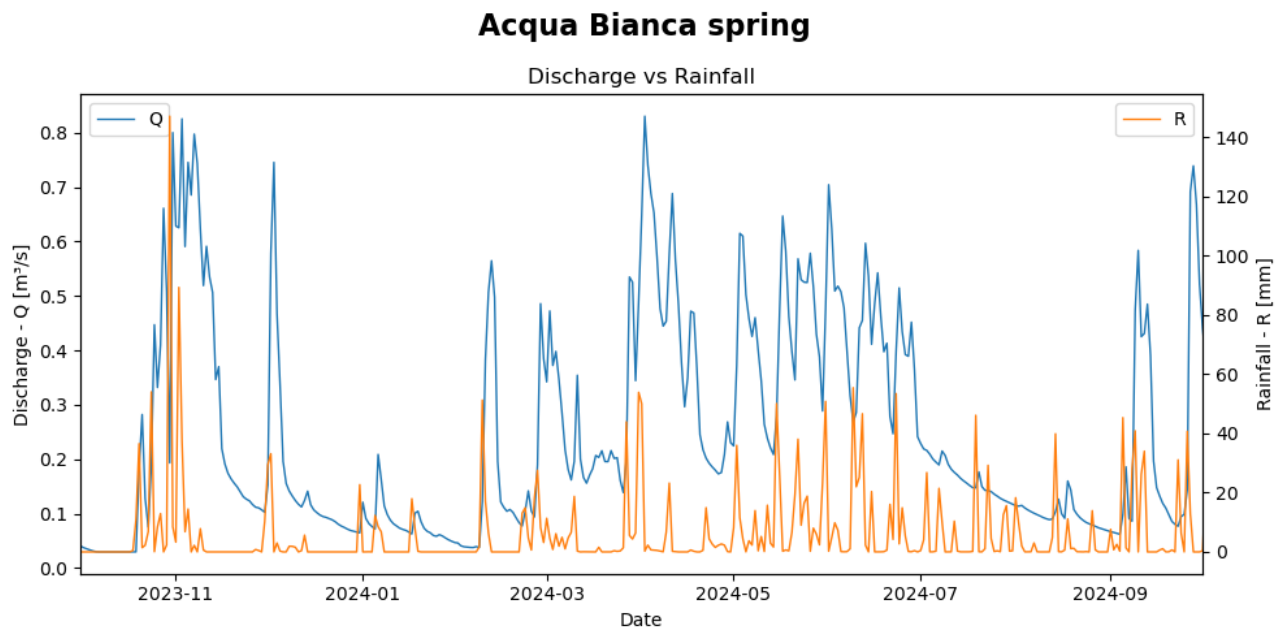


Fig. 4.20 Hydrograph of the Acqua Bianca spring in the period 01/10/2023 – 01/10/2024. In orange is the cumulative daily rainfall measured at Mt. Ario weather station. In blue is the average daily discharge calculated from the water level measured at the spring.

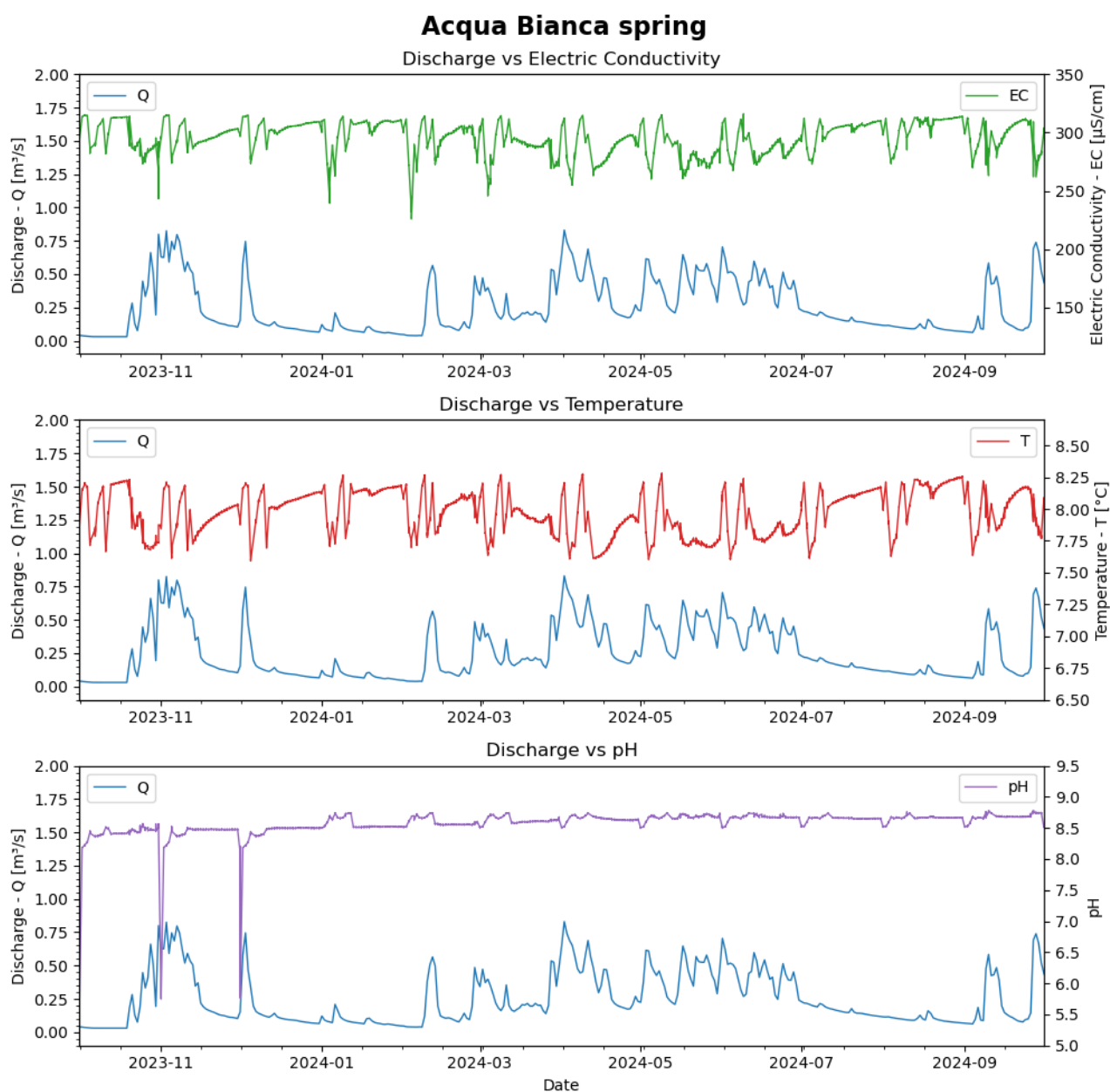


Fig. 4.21 Discharge and electrical conductivity (EC), temperature (T), and pH trends of the Acqua Bianca spring in the period 01/10/2023 – 01/10/2024.

4.3 Geophysical data and modelling results

The geophysical surveying in Val Sabbia was designed to provide an areal coverage. This included both airborne EM surveys and ground-based EM in continuous acquisition using the Loupe instrument. In addition, several DC resistivity profiles were collected along a 4 km transect that intersects the majority of the geological units of the area. The two ground-based geophysical data acquisition campaigns (Loupe and DC) were both carried out by the EEM Team for Hydro and Exploration of the University of Milano.

The first survey campaign, conducted in the summer of 2022, focused on the acquisition of DC resistivity data. The second campaign, carried out in the fall of 2022, involved the collection of EM data using the Loupe system. Both campaigns were limited to the Mt. Ario sector (Fig. 4.22), as it was the only location within the study area where the spring zones were accessible and data acquisition was logistically feasible.

Unfortunately, the AEM data acquired in Val Sabbia have not been made accessible for this thesis. Nevertheless, a synthetic study was conducted using the 3D geological model as a reference framework. This synthetic modelling exercise aimed to simulate how the subsurface would appear in AEM surveys and to evaluate the resolution capabilities of AEM in imaging fractured and karst formations, also using 3D EM modelling. The following sections present the results from the DC resistivity, Loupe, and AEM data and modelling analyses.

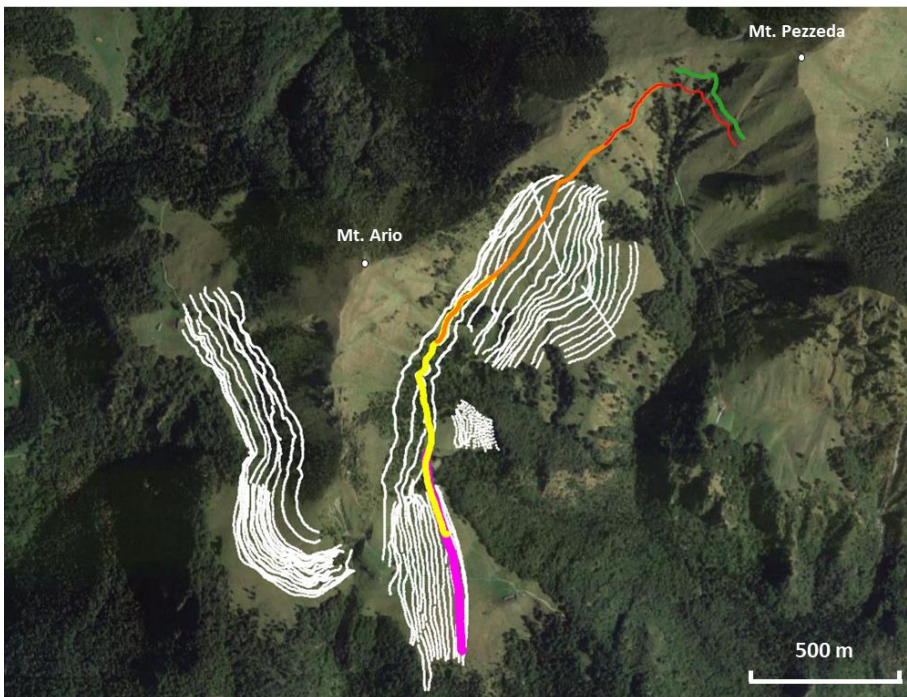


Fig. 4.22 Map of the ground-based geophysical surveys in the Mt. Ario area, showing Loupe acquisition tracks (white lines) and DC resistivity profiles (colored lines).

4.3.1 DC resistivity data

For the electrical acquisition at the study area, 5 profiles of galvanic DC data were collected using ABEM Terrameter LS2 instrument. Each profile was acquired using a four-cable layout with 4x21 (81 electrodes) spread and 5 m electrodes spacing, employing a roll-along protocol and the gradient array. The profile lengths ranged from 0.38 km to 1.15 km, with 300 m overlap between the profiles in the same line (Ario01_A, Ario01_B, Ario01_C and Ario01_D), with a total length of 4 km and 19260 measured quadrupoles across the area. The data acquisition employed a 100% duty cycle energizing scheme (Olsson et al., 2015) with four second on time and 3.75 kHz full-waveform sampling.

Given the rugged topography of the area, the survey layout had to be adapted to topographic constraints, following the natural terrain while maintaining optimal electrode contact (Fig. 4.23). Despite these challenges, the method provided a good distribution of resistivity anomalies, which are indicative of lithological and structural heterogeneities.

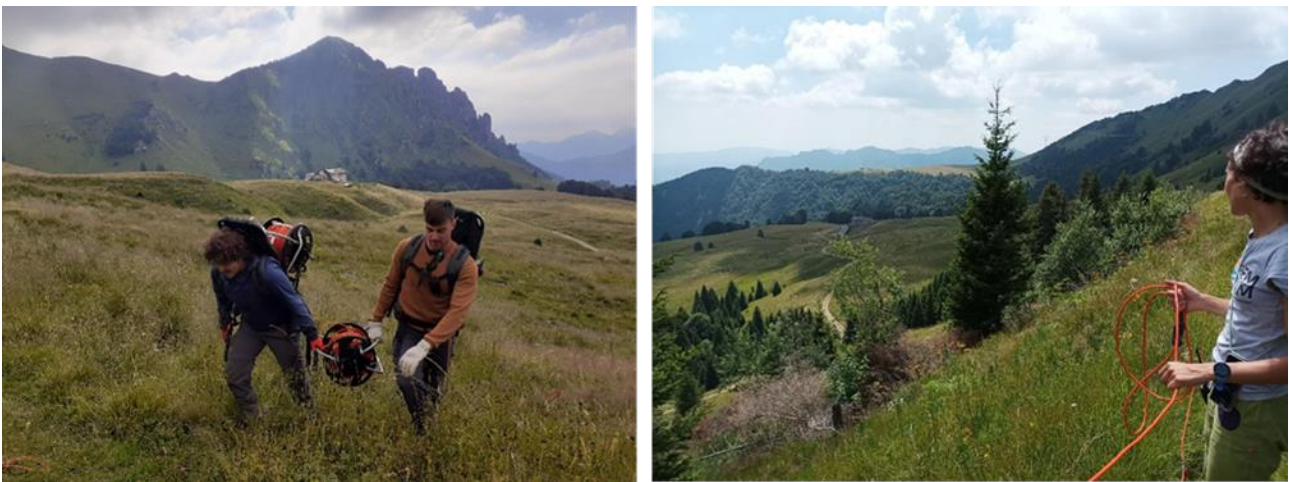


Fig. 4.23 Moments during the DC resistivity data acquisition campaign in Val Sabbia.

The acquired datasets were processed using EEMstudio (Fig. 4.24-Fig. 4.28) to remove the outliers and subsequently modelled in 2D using the EEMverter inversion code. The resulting model consists of five concatenated high-resolution resistivity profiles. This level of resolution makes it possible to assess variations in subsurface properties. Formations with different lithological properties, which also result in distinct electrical responses, are clearly appreciable in the resistivity models. In contrast, lithotypes with comparable characteristics are more challenging to perceive.

The modelling approach follows a two-step strategy. Firstly, the standard inversion of the electrical data was run, which provides an initial resistivity distribution based solely on the

geophysical measurements. Subsequently, the geologically constrained inversion was run (as detailed in Section 3.4), where the modelled geological interfaces are incorporated to refine the results by breaking the vertical constraints on the resistivity data and imposing resistivity ranges for each geological unit. Tab. 4.11 presents the resistivity ranges associated with the formations used in the inversions. The decision to assign wide resistivity ranges was made to account for the significant heterogeneity that can exist within a single geological unit and aligns with the ranges reported in literature for similar formations (Palacky, 1987). These internal variations can lead to resistivity differences spanning several orders of magnitude (for example, in the case of SAB, which exhibits highly diverse lithological characteristics, as reported in Section 4.1.3). In contrast, for formations known to be more homogeneous (e.g. ESI), it was considered reasonable to apply narrower resistivity ranges. Several trial-and-error attempts were made, initially using more conservative values. These ranges were then progressively adjusted based on their impact on the inversion misfit, aiming to strike a balance between geological plausibility and data fit.

The comparison of the results obtained from the two inversion strategies is presented for all lines in Fig. 4.29, where the inversion models are also cropped with the DOI, which reached approximately 120 m of depth. Furthermore, line-by-line results are presented in Fig. 4.30 – Fig. 4.34. In general, strong resistivity contrasts, both vertical and lateral, are found in models produced by both inversion strategies. The inversion guided by the modelled geological interfaces, in some parts of the arrays (especially the central profiles, as shown in Fig. 4.32 and Fig. 4.33), leads to a smoother resistivity distribution.

Tab. 4.12 compares the misfit values calculated for the standard and geologically-constrained inversions. The L2 and Sharp measures of misfit are reported for both strategies. This comparison allows for an assessment of how geological constraints and different inversion approaches influence the quality of the model's fit to the measured data. For profiles Ario01_D and Frane02, the high misfit is mainly due to significant topographic effects and to the curvature of the survey lines. Moreover, in these cases, the geologically-constrained inversion shows a higher misfit than the standard inversion. This is primarily because the assigned resistivity range for the WEN formation is not suitable. Specifically, the minimum resistivity value set for this formation (200 Ωm , see Tab. 4.11) fails to account for the conductive anomaly observed near the Frane springs.

The results of the DC resistivity data inversions reveal a marked heterogeneity in the distribution of resistivity values even within single geological units. This variability highlights

how different portions of the same formation may exhibit contrasting electrical properties, which can be attributed to differences in fracturing or the presence of water-filled voids. Although based on a limited number of profiles, these findings underscore the effectiveness of DC resistivity surveys in improving the characterization of subsurface conditions, as they allow for the identification of localized anomalies that may correspond to preferential groundwater flow paths.

Furthermore, incorporating geological constraints enhances the resolution of deeper structures and reduces ambiguities.

Unfortunately, the models are significantly influenced by 3D effects induced by the complex and rugged topography, which introduce geometric distortions in the resistivity distribution. These distortions can cause artificial patterns in the inversion results, such as the "drop-shaped" and "smile-shaped" anomalies observed in the shallow conductive regions. Since these shapes are a byproduct of the modelling process rather than true subsurface features, they should not be considered geologically meaningful. However, despite these distortions, the resistivity values themselves remain reliable, as they still provide information about subsurface conductivity variations.

Region	Min resistivity value (Ωm)	Max resistivity value (Ωm)
ANG	50	5000
PRZ	50	20000
BUC	50	5000
WEN	200	20000
ESI	200	20000
LSR	0.1	100000
SAB	0.1	100000
SGB	0.1	100000
DPR	50	5000
BOV	50	5000
SRV	10	1000
MAL	10	1000
MAL1	10	1000
MAL2	10	1000
VVI	10	1000
OFOL	10	1000
MVX	10	1000

Tab. 4.11 Resistivity ranges for each modelled geological unit that identifies a region in the inversion (for the acronym see Fig. 4.1).

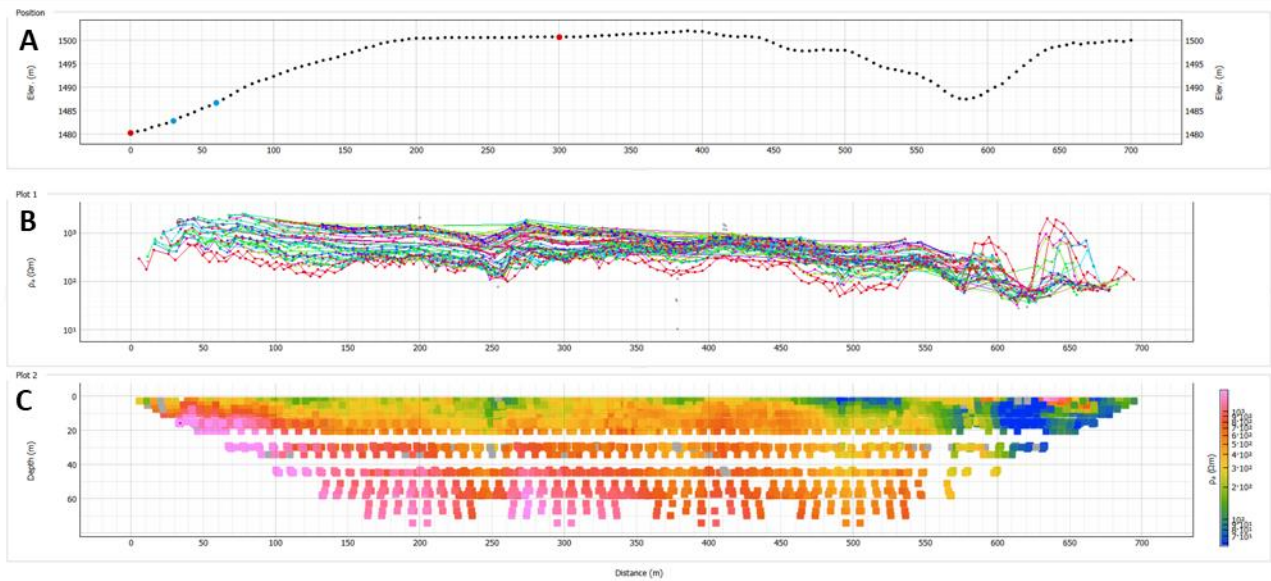


Fig. 4.24 A visualization of the DC resistivity data in EEMstudio along the acquisition profile Ario01_A. A: electrodes and topography. B: apparent resistivity as a function of lateral focus. C: pseudosection. Grey dots in B and C represent data culled out in the processing phase.

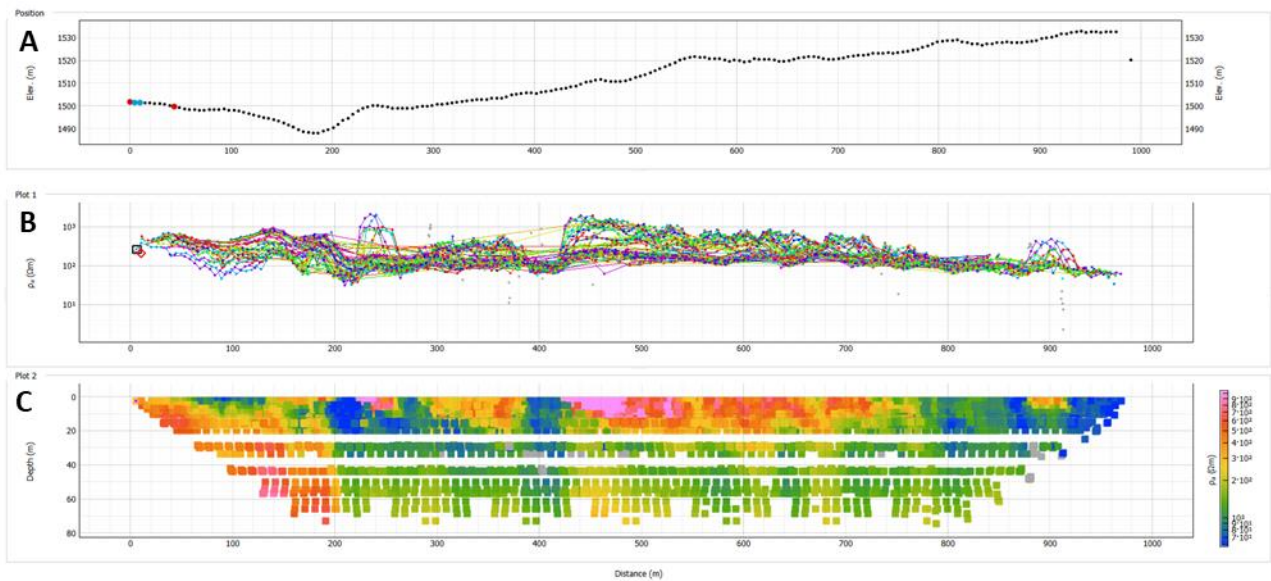


Fig. 4.25 A visualization of the DC resistivity data in EEMstudio along the acquisition profile Ario01_B. A: electrodes and topography. B: apparent resistivity as a function of lateral focus. C: pseudosection. Grey dots in B and C represent data culled out in the processing phase.

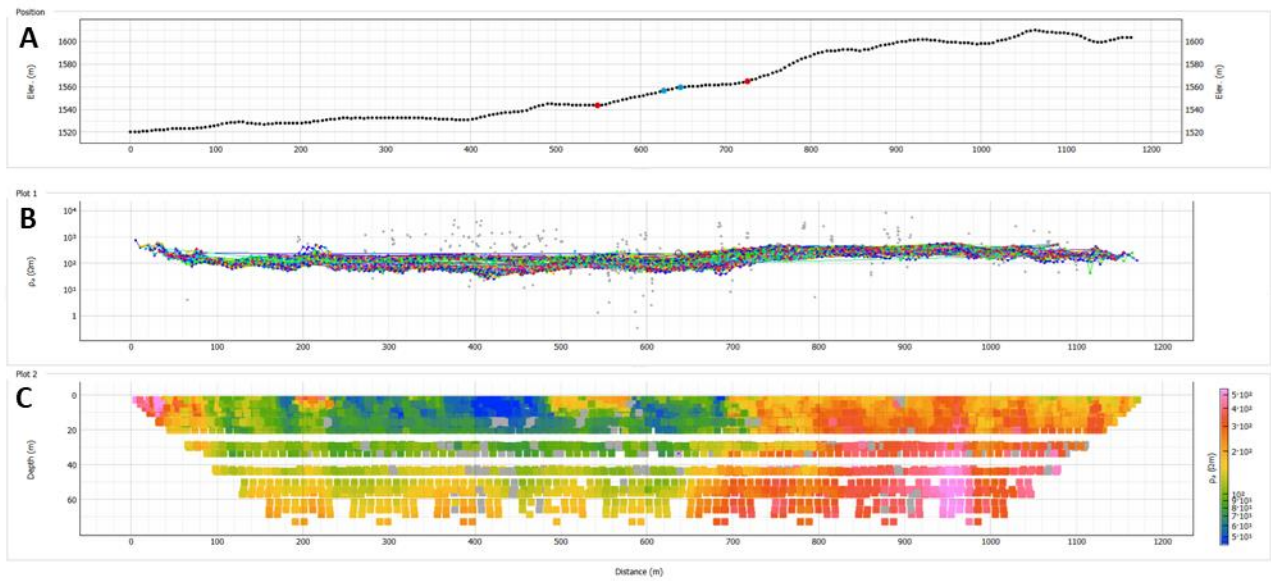


Fig. 4.26 A visualization of the DC resistivity data in EEMstudio along the acquisition profile Ario01_C. A: electrodes and topography. B: apparent resistivity as a function of lateral focus. C: pseudosection. Grey dots in B and C represent data culled out in the processing phase.

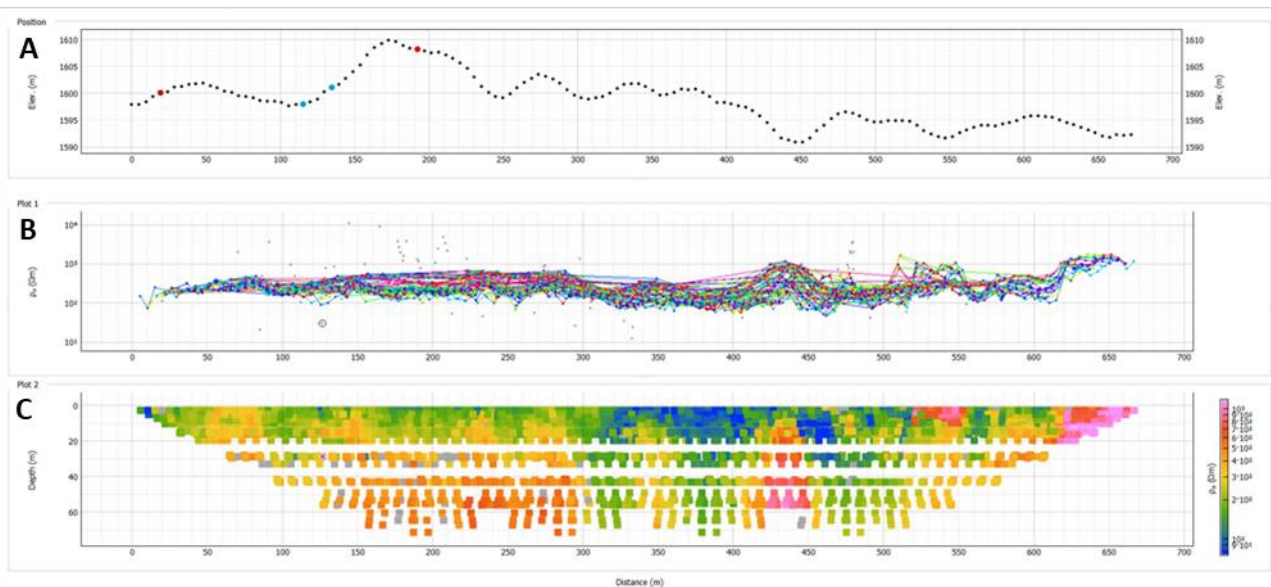


Fig. 4.27 A visualization of the DC resistivity data in EEMstudio along the acquisition profile Ario01_D. A: electrodes and topography. B: apparent resistivity as a function of lateral focus. C: pseudosection. Grey dots in B and C represent data culled out in the processing phase.

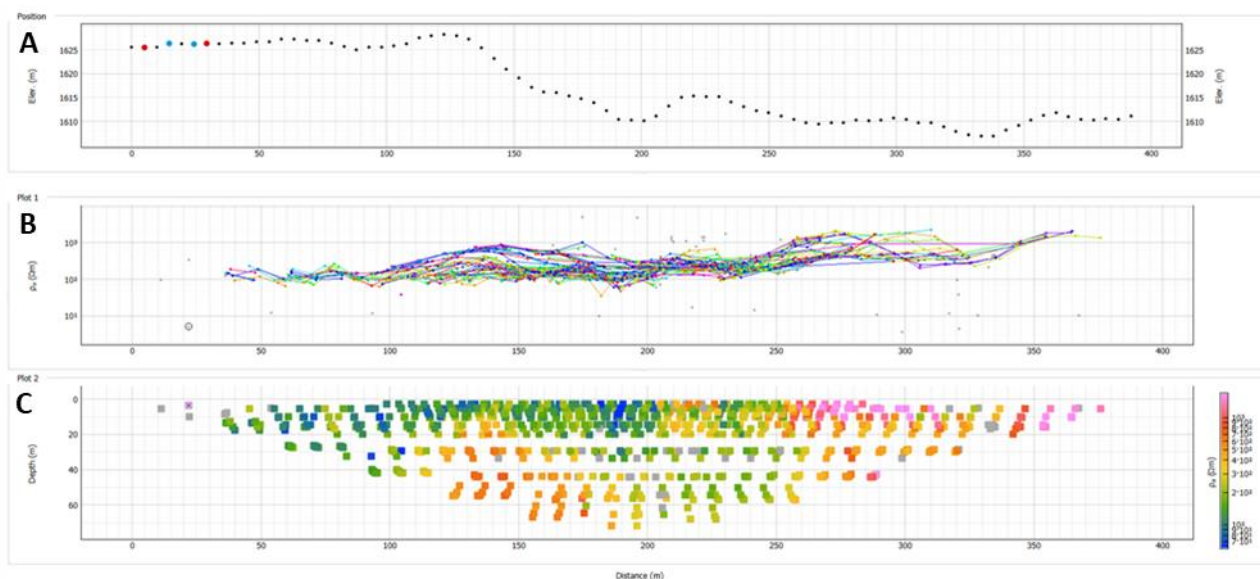


Fig. 4.28 A visualization of the DC resistivity data in EEMstudio along the acquisition profile Frana02. A: electrodes and topography. B: apparent resistivity as a function of lateral focus. C: pseudosection. Grey dots in B and C represent data culled out in the processing phase.

According to the DC resistivity models (Fig. 4.29), the conductive anomalies ($20 - 300 \Omega\text{m}$) at shallow depth are localized within the Carnian-aged formations (San Giovanni Bianco Fm. – SGB, Val Sabbia Sandstone – SAB, Calcarei Lastriformi – LSR). As described in Section 4.1.3, these formations are heterogeneous in composition: SGB and SAB are composed primarily of dolostone and marly siltstone, with interbedded clayey and carbonate-rich layers. In contrast, LSR consists of fractured layered limestone. Despite their different lithologies, these formations display similar electrical behavior, though for different reasons. In SGB and SAB, the conductive response is primarily due to their clayey content and the presence of water within microfractures. Over time, this water can dissolve the carbonate portions, further altering their conductivity. While in LSR, the conductive anomaly is associated with the presence of water within its fracture systems. Since limestone itself is generally resistive, the fractures act as pathways for water infiltration, increasing overall conductivity.

Here a brief discussion of the main evidences is given line-by-line.

- Ario01_A (Fig. 4.30). A highly resistive body (ranging from 2,000 to 10,000 Ωm) is observed cropping out in the southern section of the first profile, extending northward with approximately 30° dipping angle. This resistive anomaly closely corresponds to the Esino Limestone (ESI) outcrop in this area. Typically, ESI carbonate formation is prone to karstification due to its relatively pure limestone composition, which usually leads to

the development of cavities and underground drainage systems. However, in the Mt. Ario area, ESI exhibits a massive and compact structure, which makes it less susceptible to karstification. This structural characteristic explains its high resistivity, as fewer fractures and dissolution features mean less water infiltration, reducing the conductive pathways that would otherwise lower its resistivity. Therefore, in the study area ESI behaves as a relatively impermeable formation.

The LSR formation, which is prone to karstification, displays significant lateral heterogeneities with conductive anomalies in the range of 70 – 150 Ωm , in the vicinity of the sinkholes evident on the surface (see Section 4.1.3). This part of the profile is particularly straight, with gentle topography variations: the conductive heterogeneities are not related to effects due to 3D topography and/or profile bends. The conductive anomalies observed within the LSR formation are interpreted as water-bearing zones associated with the presence of fractures. These fractures are believed to have been enlarged by dissolution processes, which enhance permeability and allow for increased groundwater circulation, resulting in lower resistivity values in those areas.

The SAB formation exhibits conductive behavior upstream of the springs Acqua Fredda 1 and 2, suggesting increased moisture content or the presence of saturated zones.

- Ario01_B (Fig. 4.31). This profile reveals pronounced heterogeneities within the SAB formation, and to a lesser extent within the LSR formation. Significant conductive anomalies are observed upstream of the Acqua Fredda and Stabol springs, likely indicating zones of groundwater accumulation. However, some of the apparent heterogeneities coincide with areas of strong profile bending and are likely artifacts caused by 3D effects in the inversion process.
- Ario01_C (Fig. 4.32). This profile displays heterogeneities within the SAB formation, similar to those observed in Ario01_B, while the LSR formation appears significantly more homogeneous. A change in resistivity distribution is evident along the profile at the location of the MAL thrust splay. However, a conductive artifact is present near the area of pronounced profile bending, likely resulting from 3D inversion effects. Overall, the ANG formation in this profile is more resistive compared to both the LSR and ESI formations.
- Ario01_D (Fig. 4.33). This profile exhibits the most pronounced curvature among all the survey lines and was therefore split into two segments (Ario01_D01 and Ario01_D02) during the inversion process to reduce the impact of bending-related distortions.

Although some 3D inversion artifacts are still present, the resulting resistivity distribution at depth appears more uniform compared to the other profiles.

- Frana02 (Fig. 4.34). This profile captures similar features to those observed in Ario01_D, but it shows a decrease in resistivity in the area corresponding to the Frana springs.

PROFILE	DC INVERSIONS	MISFIT
Ario01_A	L2 standard	3.03
	L2 geologically-constrained	2.99
	Sharp geologically-constrained	2.41
	Sharp standard	2.92
Ario01_B	L2 standard	4.08
	L2 geologically-constrained	3.97
	Sharp standard	3.80
	Sharp geologically-constrained	3.53
Ario01_C	L2 standard	5.61
	L2 geologically-constrained	6.79
	Sharp standard	5.49
	Sharp geologically-constrained	6.05
Ario01_D01	L2 standard	5.59
	L2 geologically-constrained	6.42
	Sharp standard	5.18
	Sharp geologically-constrained	5.80
Ario01_D02	L2 standard	5.94
	L2 geologically-constrained	7.98
	Sharp standard	5.24
	Sharp geologically-constrained	7.67
Frana02	L2 standard	10.12
	L2 geologically-constrained	14.06
	Sharp standard	9.86
	Sharp geologically-constrained	13.62

Tab. 4.12 Misfit comparison of the two different inversion strategies for DC resistivity data.

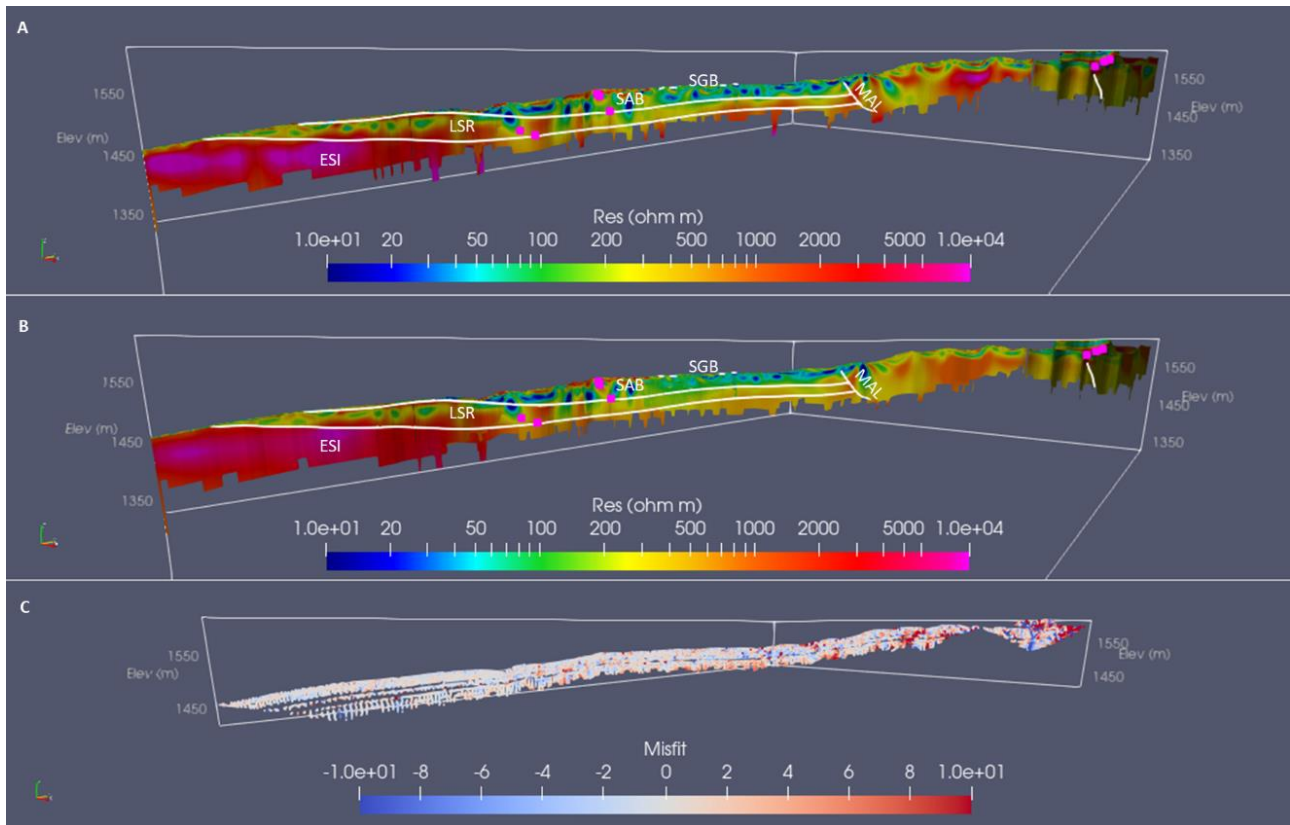


Fig. 4.29 Inversion of the DC resistivity data along all profiles cropped by DOI. A: standard inversion. B: geologically-constrained inversion. C: DC resistivity misfit. Purple dots in A and B are the springs of interest. White lines represent the modelled geological interfaces.

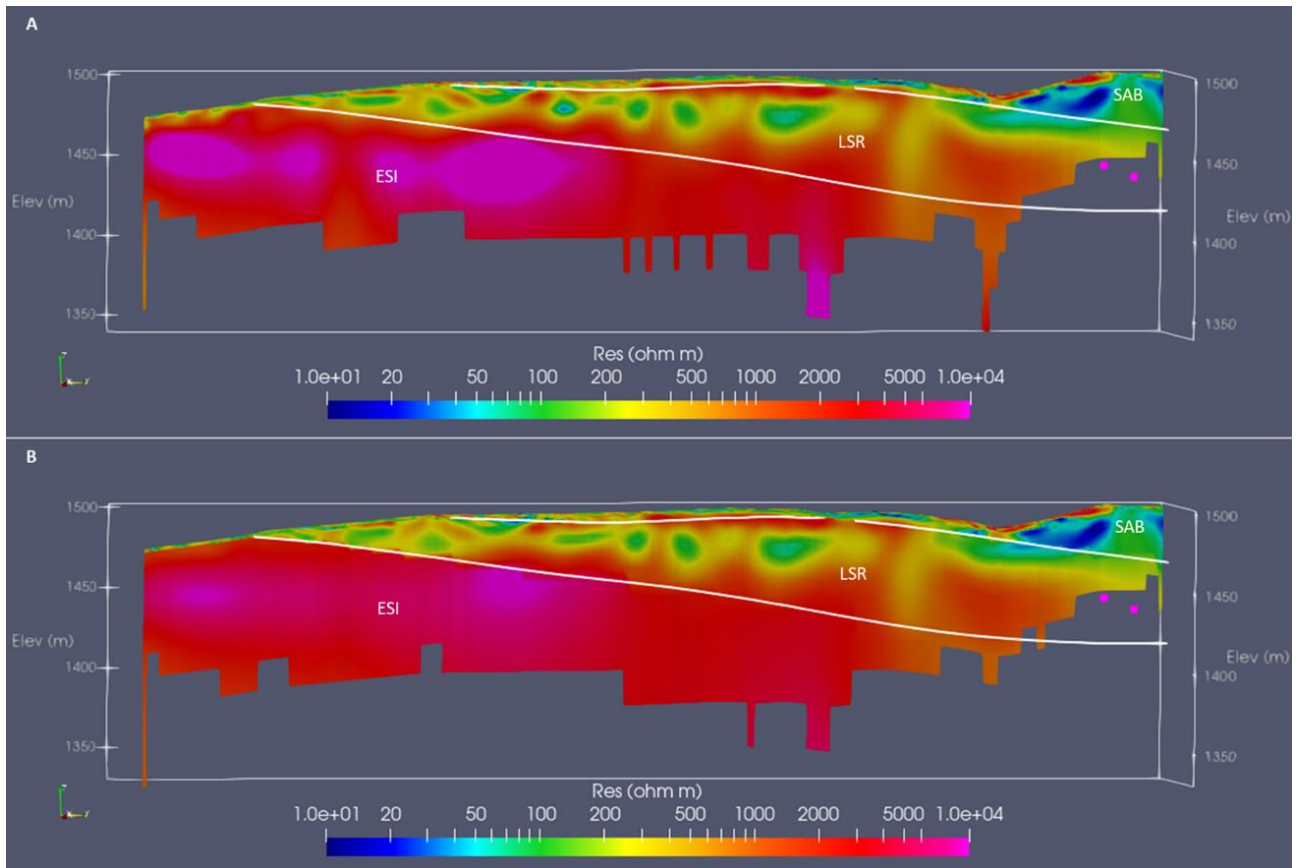


Fig. 4.30 Inversion of the DC resistivity data of the first profile, Ario01_A, cropped by DOI. A: standard inversion. B: geologically-constrained inversion. Purple dots are the springs of interest. White lines represent the modelled geological interfaces.

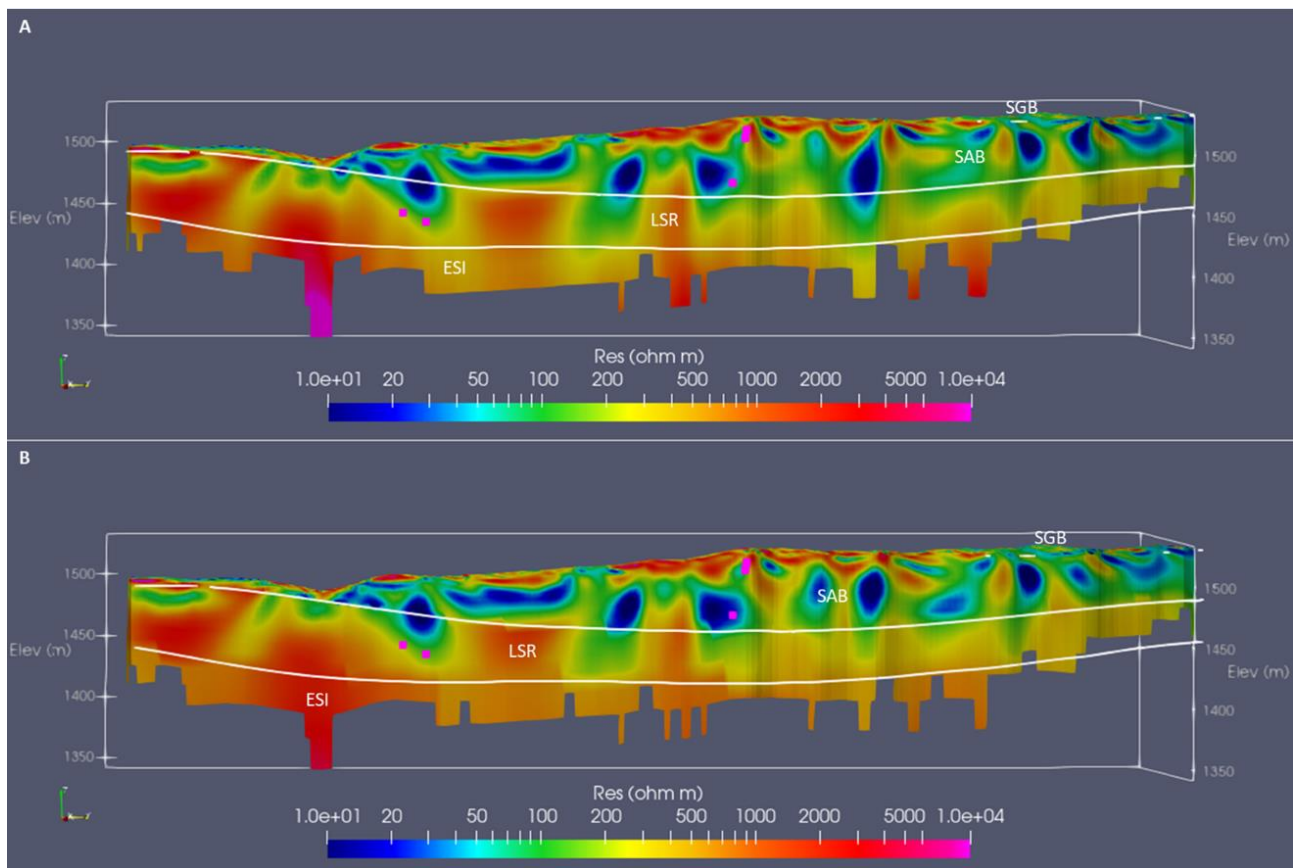


Fig. 4.31 Inversion of the DC resistivity data of the second profile, Ario01_B, cropped by DOI. A: standard inversion. B: geologically-constrained inversion. Purple dots are the springs of interest. White lines represent the modelled geological interfaces.

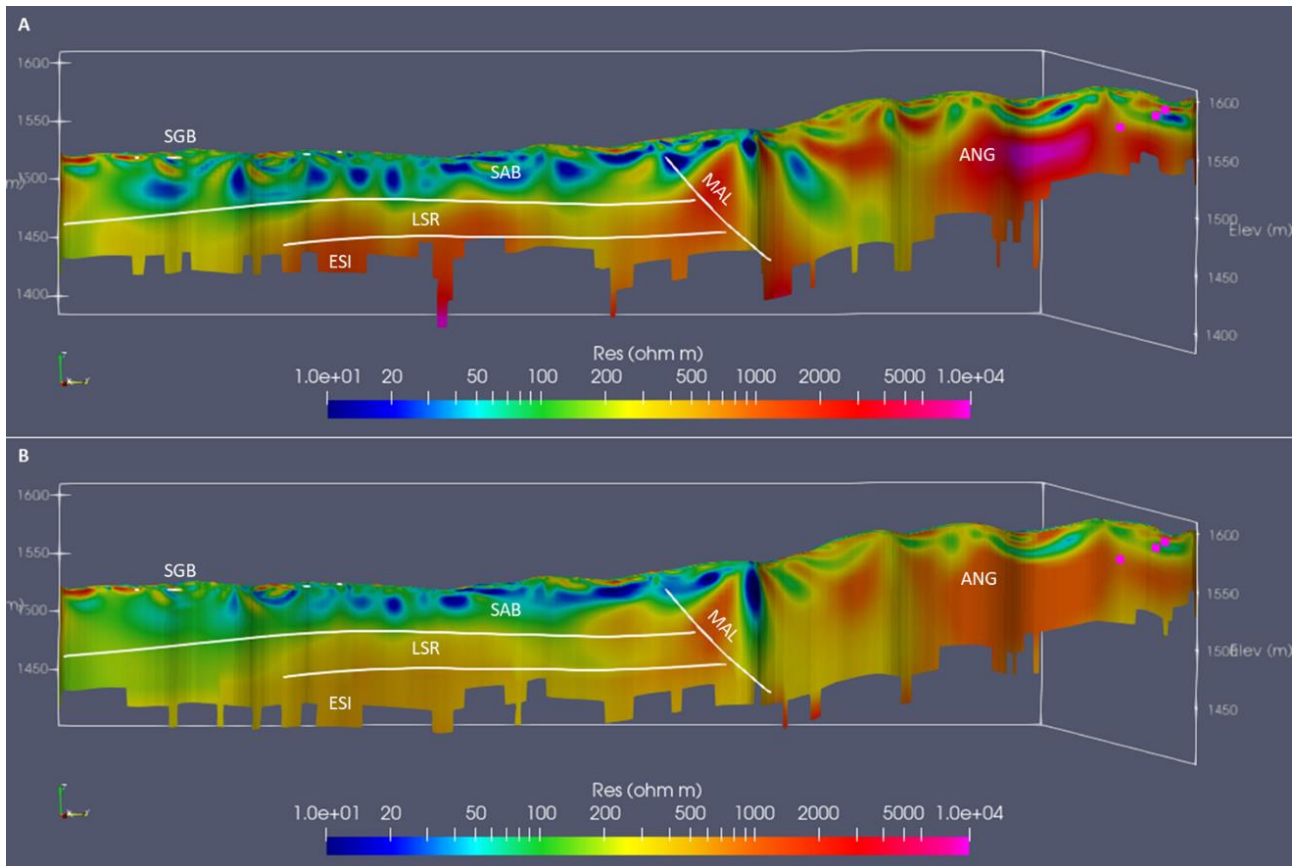


Fig. 4.32 Inversion of the DC resistivity data of the third profile, Ario01_C, cropped by DOI. A: standard inversion. B: geologically-constrained inversion. Purple dots are the springs of interest. White lines represent the modelled geological interfaces.

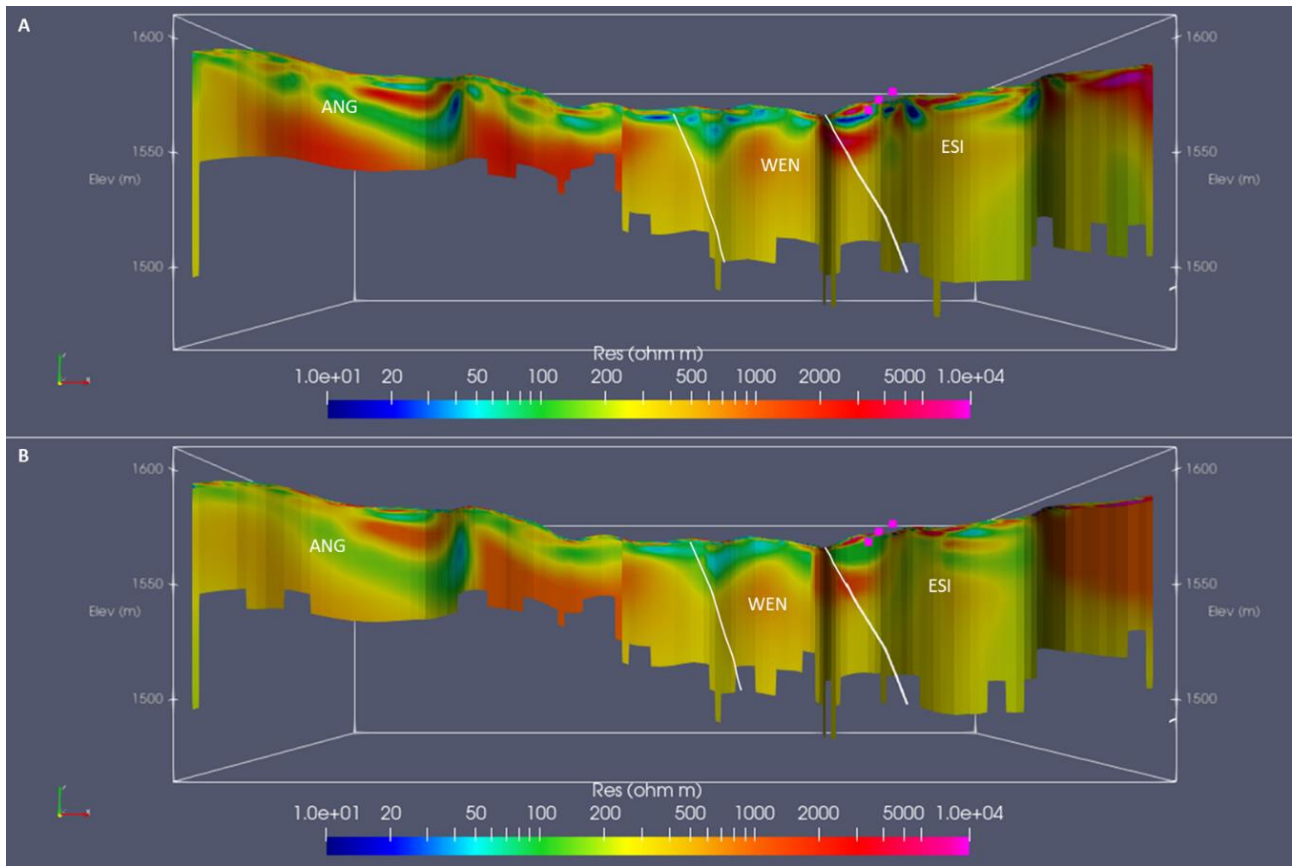


Fig. 4.33 Inversion of the DC resistivity data of the fourth profile, Ario01_D, cropped by DOI. A: standard inversion. B: geologically-constrained inversion. Purple dots are the springs of interest. White lines represent the modelled geological interfaces.

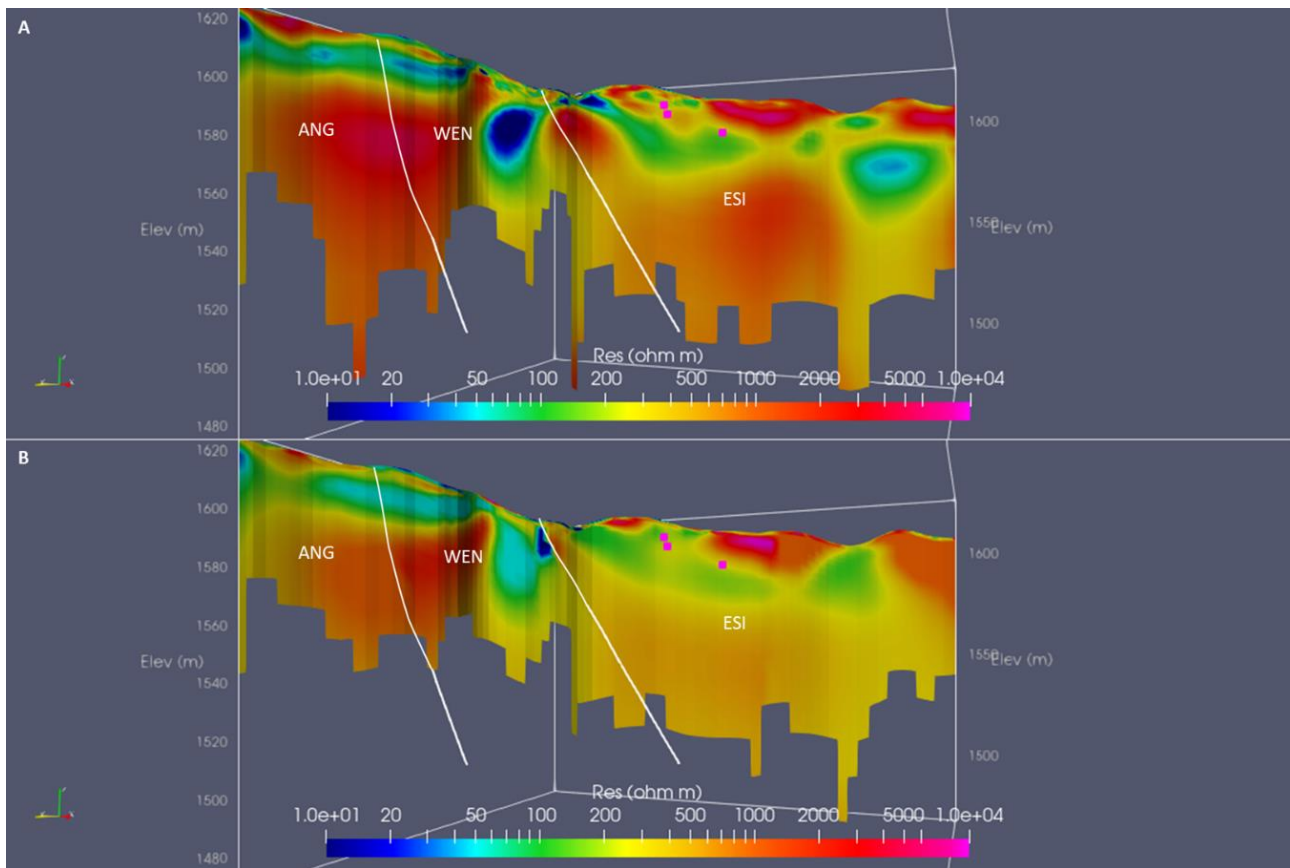


Fig. 4.34 Inversion of the DC resistivity data of the last profile, Frane02, cropped by DOI. A: standard inversion. B: geologically-constrained inversion. Purple dots are the springs of interest. White lines represent the modelled geological interfaces.

4.3.2 Loupe resistivity model

In the study area, a total of 43.44 km of TEM data were collected using the Loupe instrument in continuous acquisition mode, with repetition frequency set to 75 Hz and 10 m Tx-Rx separation. The data collection focused on the area surrounding Mt. Ario, ensuring an areal coverage of key geological features and of the springs. However, the design of acquisition lines was significantly constrained by logistical challenges, related to the inaccessibility of steep and rugged slopes, as well as vegetation. In particular, it has proven very challenging to cover areas with trees and bushes, due to the cable connection between the Tx and Rx Loupe units.

Fig. 4.35 presents a representative data example, along with the number of gates retained after processing the entire survey. Overall, this site posed significant challenges in terms of signal-to-noise ratio, due to the high resistivity associated with some geological units (e.g. ESI).

The area's rugged topography had a considerable impact on the acquisition geometry, often causing the Tx and Rx units to have differing orientations in terms of bearing, pitch, and roll.

During processing, data with significant deviation from the ideal acquisition setup were carefully removed, and only the Z component of the field has been modelled. These factors, together with the low signal-to-noise ratio, resulted in a high final data misfit in both the standard and geologically-constrained inversions, as reported in Tab. 4.13. The misfit is particularly elevated in areas where the ESI formation outcrops (Fig. 4.36), due to its high resistivity (ranging from 500 to 1000 Ωm), which increases susceptibility to noise. Moreover, the presence of buildings and potential underground utilities in these areas may further contribute to signal distortion. Nevertheless, the models are consistent with the geology of the area and the DC results, as shown in the following.

Fig. 4.37 to Fig. 4.43 present the Loupe results as XY depth slices, both as standard inversions (A panels) and geologically-informed inversions (B panels), from -1 m to -50 m depth below the ground. Overall, the resistivity models are generally more conductive than those obtained from DC resistivity surveys, although the resistivity distribution is consistent. The main features highlighted by the XY slices are:

- in the Loupe resistivity models, most superficial layers are characterized by low resistivities due to the presence of non-consolidated deposits (Fig. 4.37);
- moving deeper, in the southernmost and the eastern portion of the acquired lines, where the Esino Limestone (ESI) crops out, a volume characterized by a high resistive signature appears (500 – 1000 Ωm) (Fig. 4.38 – Fig. 4.41);
- in contrast, the overlying Calcari Lastriformi (LSR), Val Sabbia Sandstone (SAB), and San Giovanni Bianco Formation (SGB) exhibit conductive values (10 – 100 Ωm), not uniformly distributed within the volume (Fig. 4.38 – Fig. 4.43).

Applying constraint-breaking across the regions defined by the 3D geological model, and assigning specific resistivity ranges to each region during the inversion of Loupe data, led to a smoother distribution of resistivity values within each geological unit. This distribution more accurately reflects the geological stratigraphy and better represents the physical properties of each formation, as shown by the comparison between the standard and geologically-constrained inversions (panels A and B in Fig. 4.37 to Fig. 4.43). The advantages of the geologically-informed inversion become even more evident when comparing Fig. 4.44 and Fig. 4.45, which show two geological cross sections overlaid with the results of the Loupe inversion using the standard and geologically-informed approaches, respectively. In Fig. 4.45 the improvement achieved through the geologically-informed inversion is clearly visible: the resistivity values are more consistently distributed within each geological unit, and the data

misfit is reduced (see Tab. 4.13). This approach also allows for the unambiguous identification of significant heterogeneity within the SAB formation. Fig. 4.46 further illustrates this in 3D, showing the heterogeneous resistivity distribution obtained from the geologically-constrained inversion of Loupe data within the SAB formation, as defined by the 3D geological model.

Fig. 4.47 offers a direct comparison between the DC and Loupe resistivity inversion results, with the latter extracted along the same profiles as the DC data. The comparison reveals a strong agreement between the two datasets, both in terms of resistivity contrasts and the geometry of subsurface features. The Loupe model (panel A) exhibits a smoother distribution of resistivity values, while the DC model (panel B) show more pronounced heterogeneities, likely due to the higher sensitivity of DC methods to near-surface variations.

A detailed comparison of conductive anomalies (resistivity $< 50 \Omega \cdot \text{m}$) from Loupe model with the DC model highlights a notable spatial consistency: anomalies identified in the Loupe model closely align with those detected in the DC inversions. This correspondence supports the reliability of Loupe data acquisition in complex mountain terrains for identifying geological structures and potential water-bearing zones within stratigraphic units.

However, it is important to note that the depth of investigation for the Loupe system is relatively limited, reaching approximately 50 m. For deeper subsurface imaging with regional or areal coverage in rugged mountain settings, Airborne EM methods remain the most suitable and practical solution.

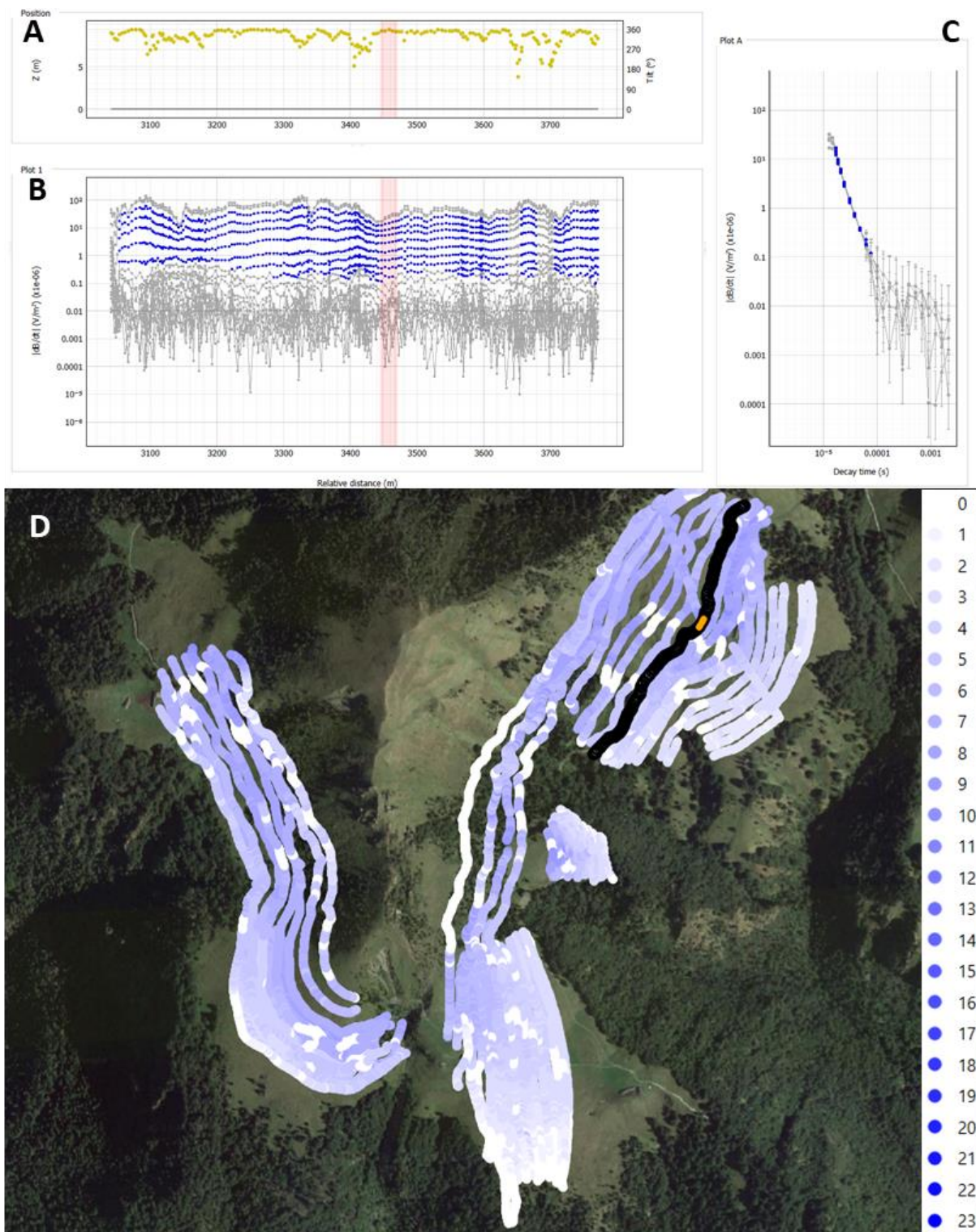


Fig. 4.35. A: example of Loupe data along an acquisition profile showing Tx-Rx distance (yellow points), with decays highlighted in red (see C panel); B and C: data used in the inversion are shown in blue, while grey indicates data discarded during processing. D: map of the number of gates retained after data processing (white to blue markers), along with the position of the data shown in A (black markers) and C (yellow markers).

SECTOR	LOUPE INVERSIONS	MISFIT
West	L2 standard	3.12
	L2 geologically-constrained	2.79
	Sharp standard	2.54
	Sharp geologically-constrained	2.34
East	L2 standard	3.50
	L2 geologically-constrained	2.89
	Sharp standard	2.64
	Sharp geologically-constrained	2.38

Tab. 4.13 Misfit comparison of the two different inversion strategies for Loupe resistivity data.

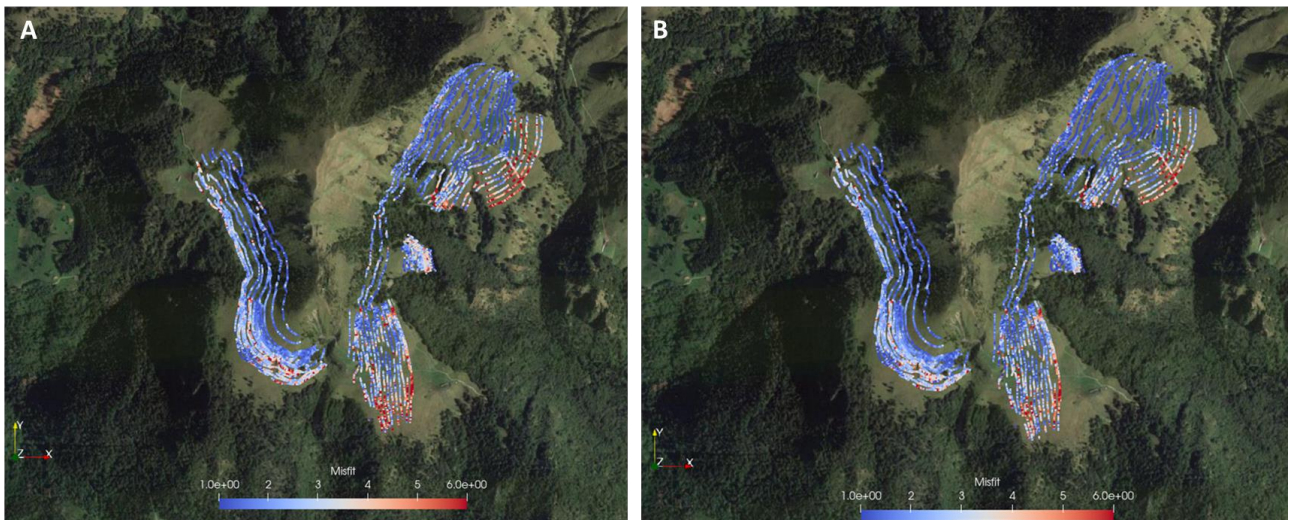


Fig. 4.36 A: Map of the Loupe data misfit for sharp standard inversion; B: Map of the Loupe data misfit for sharp geologically-constrained inversion.

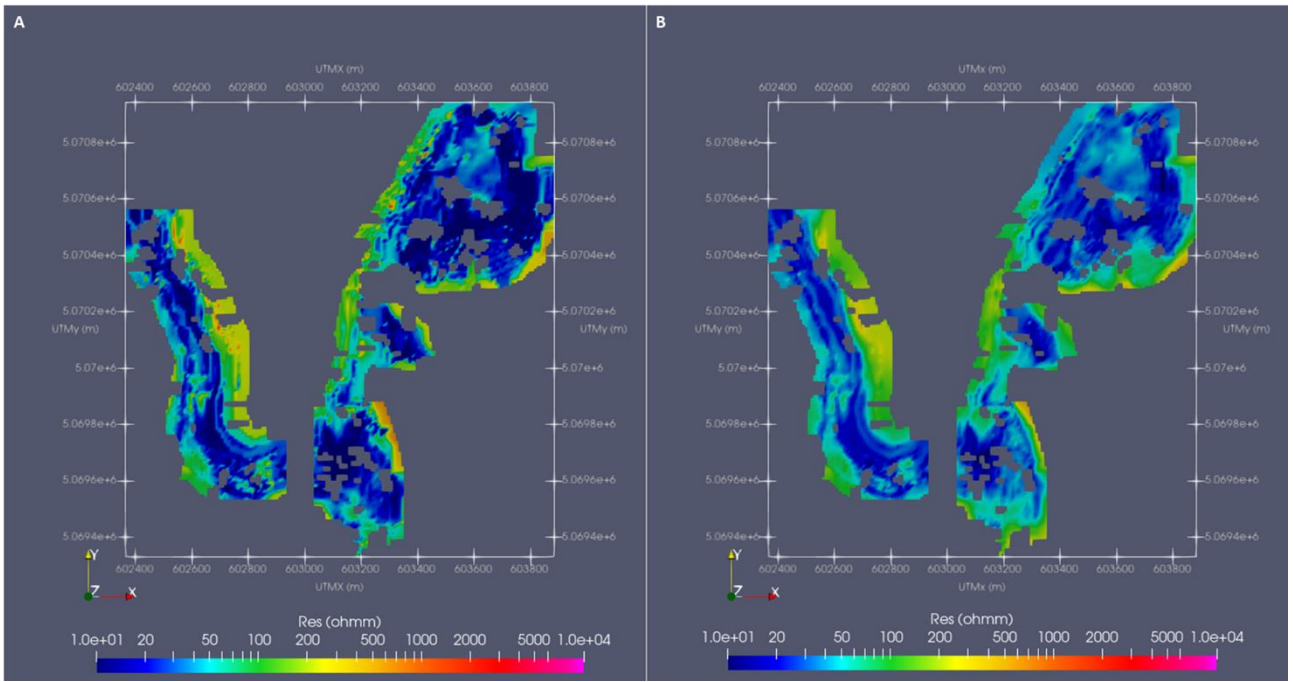


Fig. 4.37 Depth slice at $z = -1$ m from the Loupe data inversion cropped with DOI. A: standard inversion. B: geologically-informed inversion.

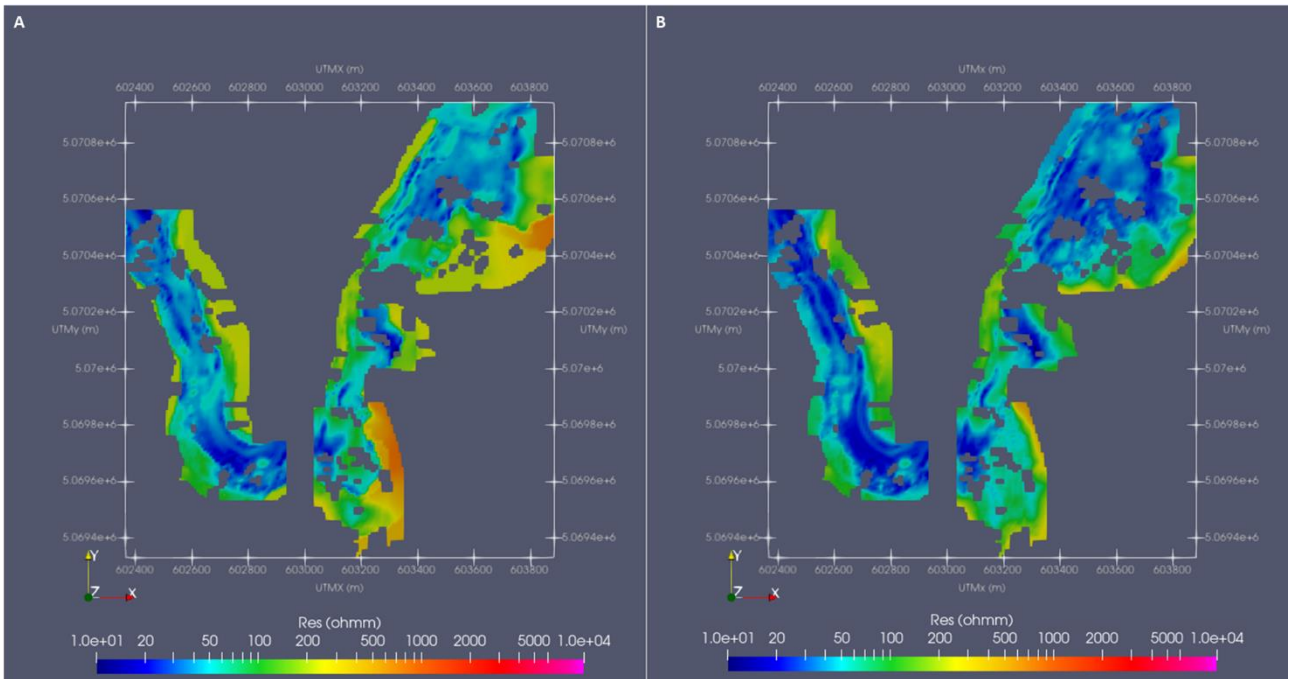


Fig. 4.38 Depth slice at $z = -5$ m from the Loupe data inversion cropped with DOI. A: standard inversion. B: geologically-informed inversion.

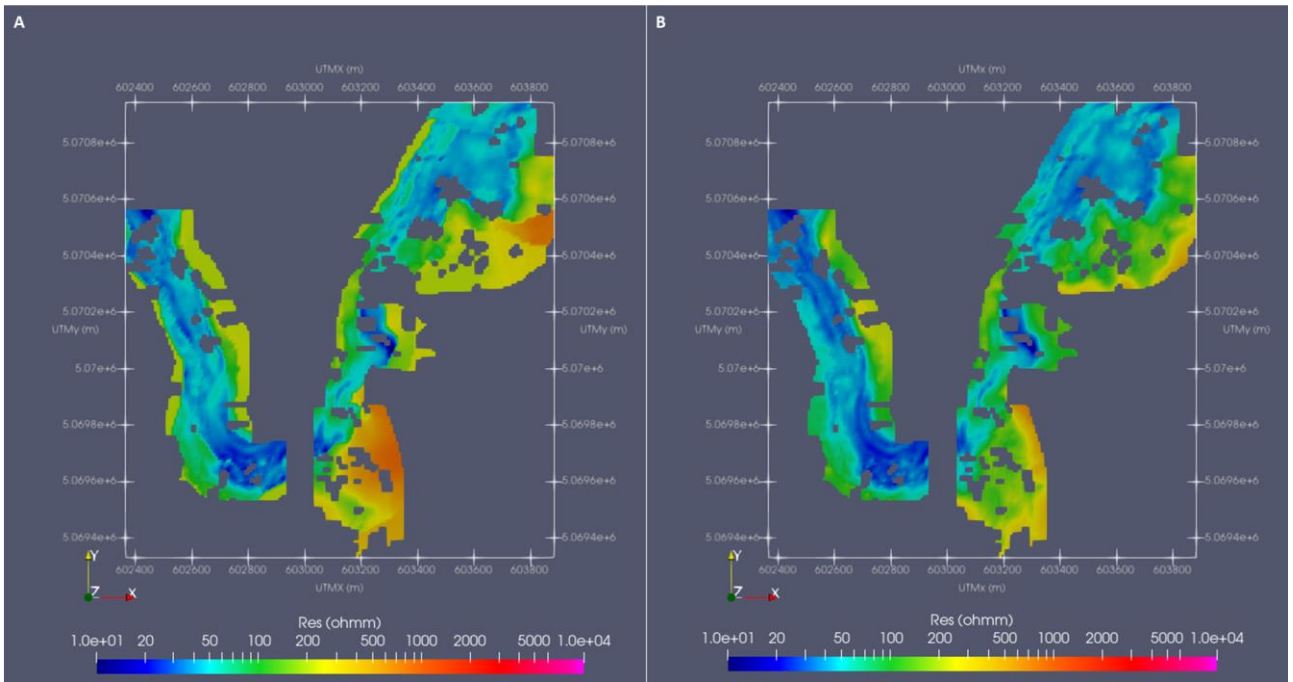


Fig. 4.39 Depth slice at $z = -10$ m from the Loupe data inversion cropped with DOI. A: standard inversion. B: geologically-informed inversion.

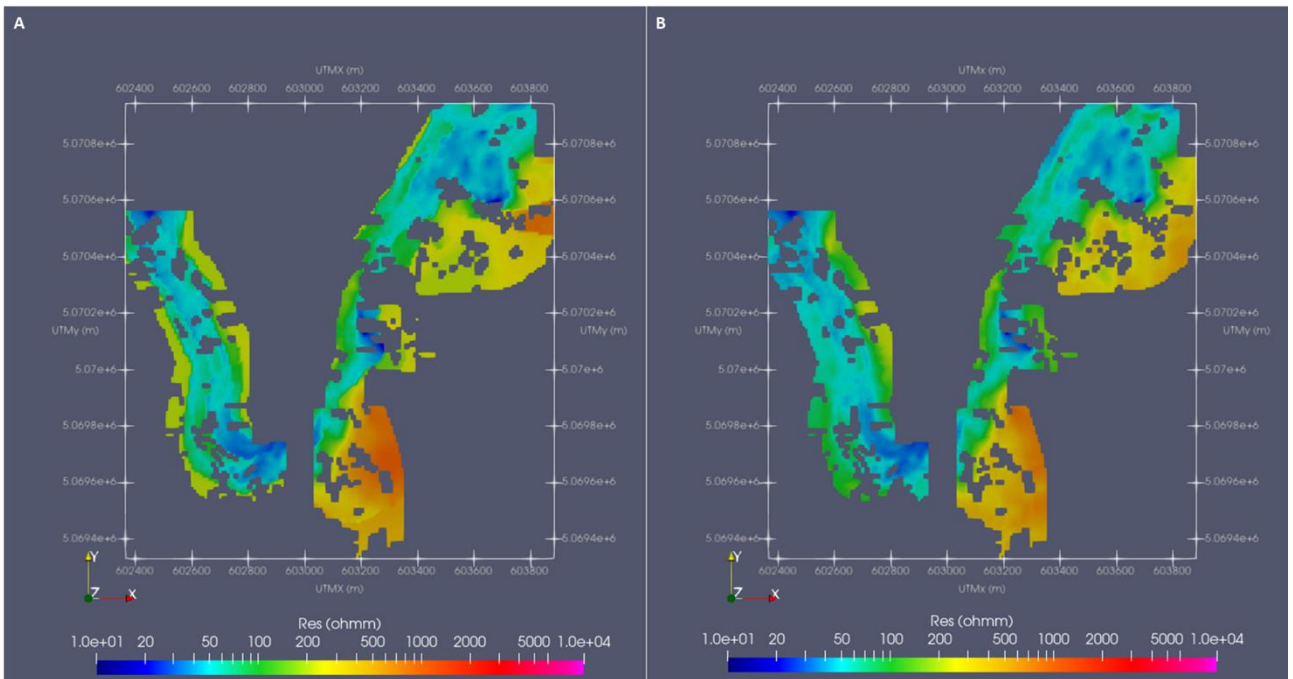


Fig. 4.40 Depth slice at $z = -20$ m from the Loupe data inversion cropped with DOI. A: standard inversion. B: geologically-informed inversion.

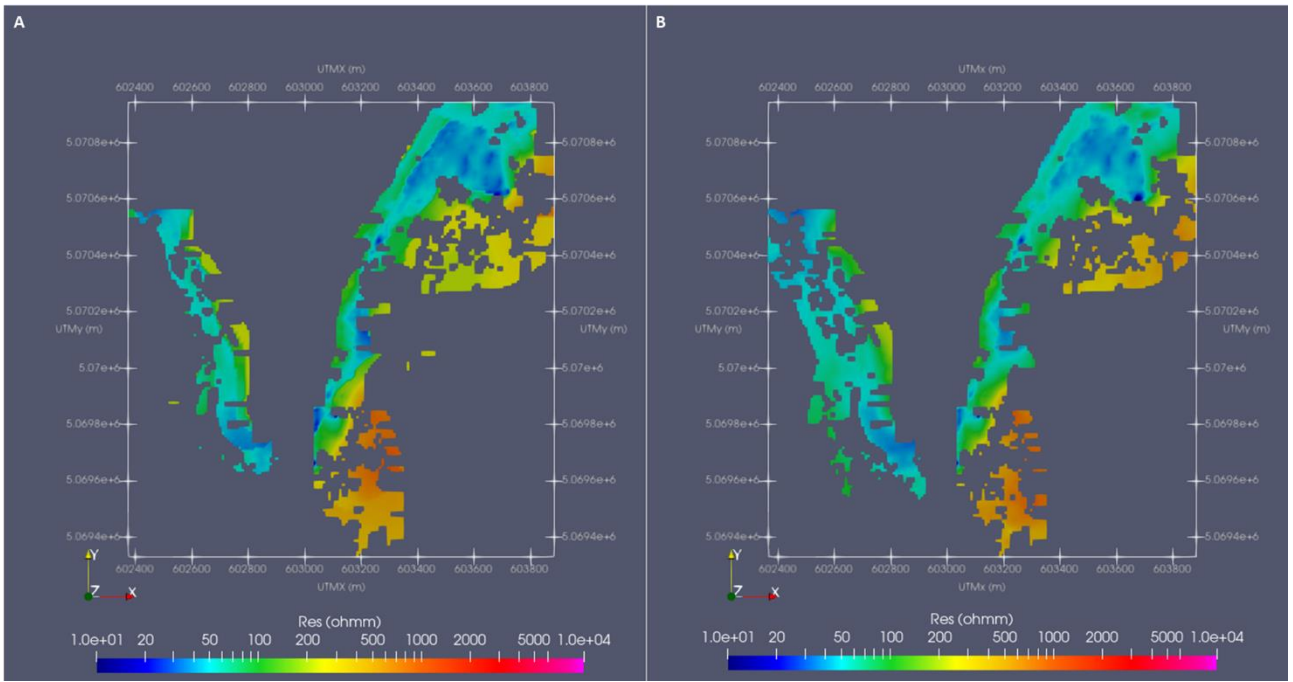


Fig. 4.41 Depth slice at $z = -30$ m from the Loupe data inversion cropped with DOI. A: standard inversion. B: geologically-informed inversion.

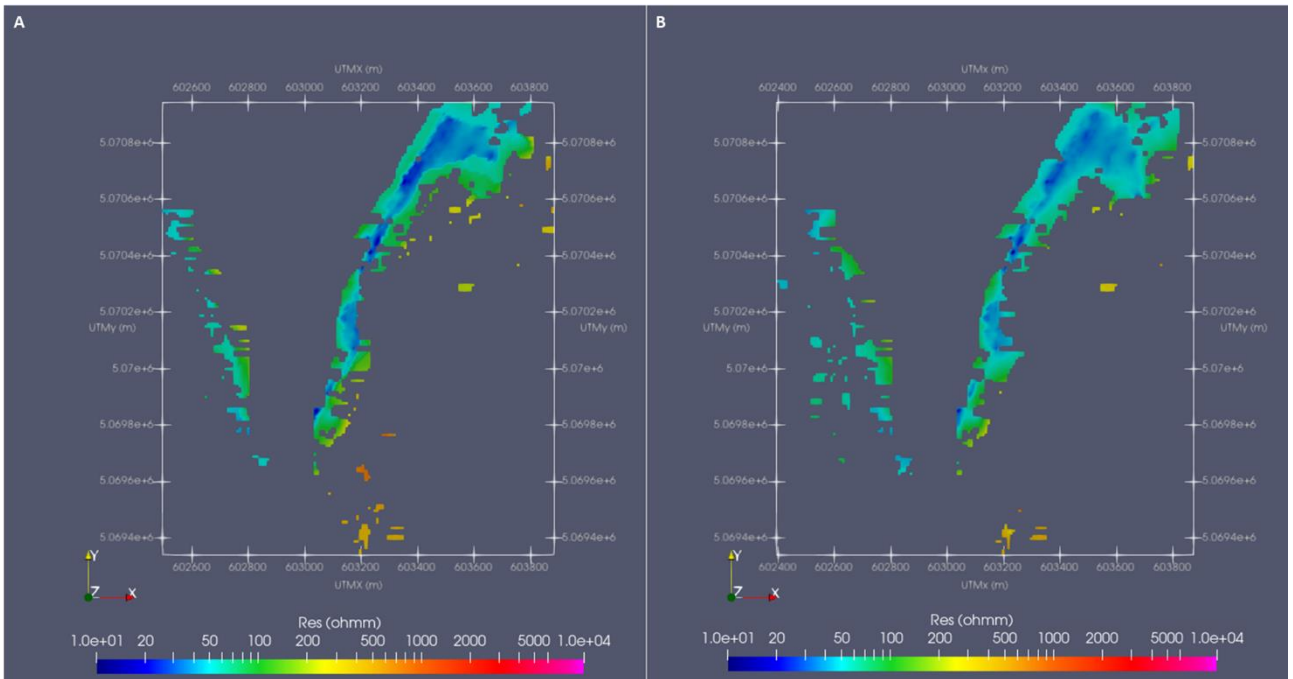


Fig. 4.42 Depth slice at $z = -40$ m from the Loupe data inversion cropped with DOI. A: standard inversion. B: geologically-informed inversion.

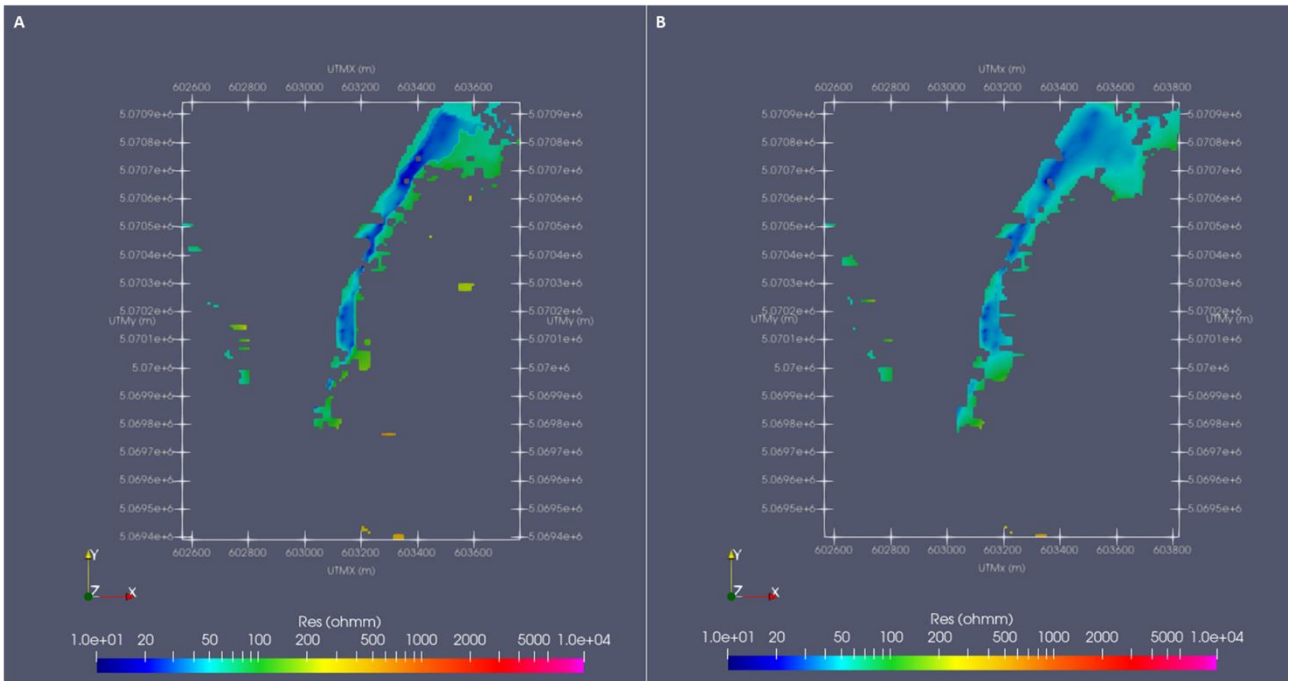


Fig. 4.43 Depth slice at $z = -50$ m from the Loue data inversion cropped with DOI. A: standard inversion. B: geologically-informed inversion.

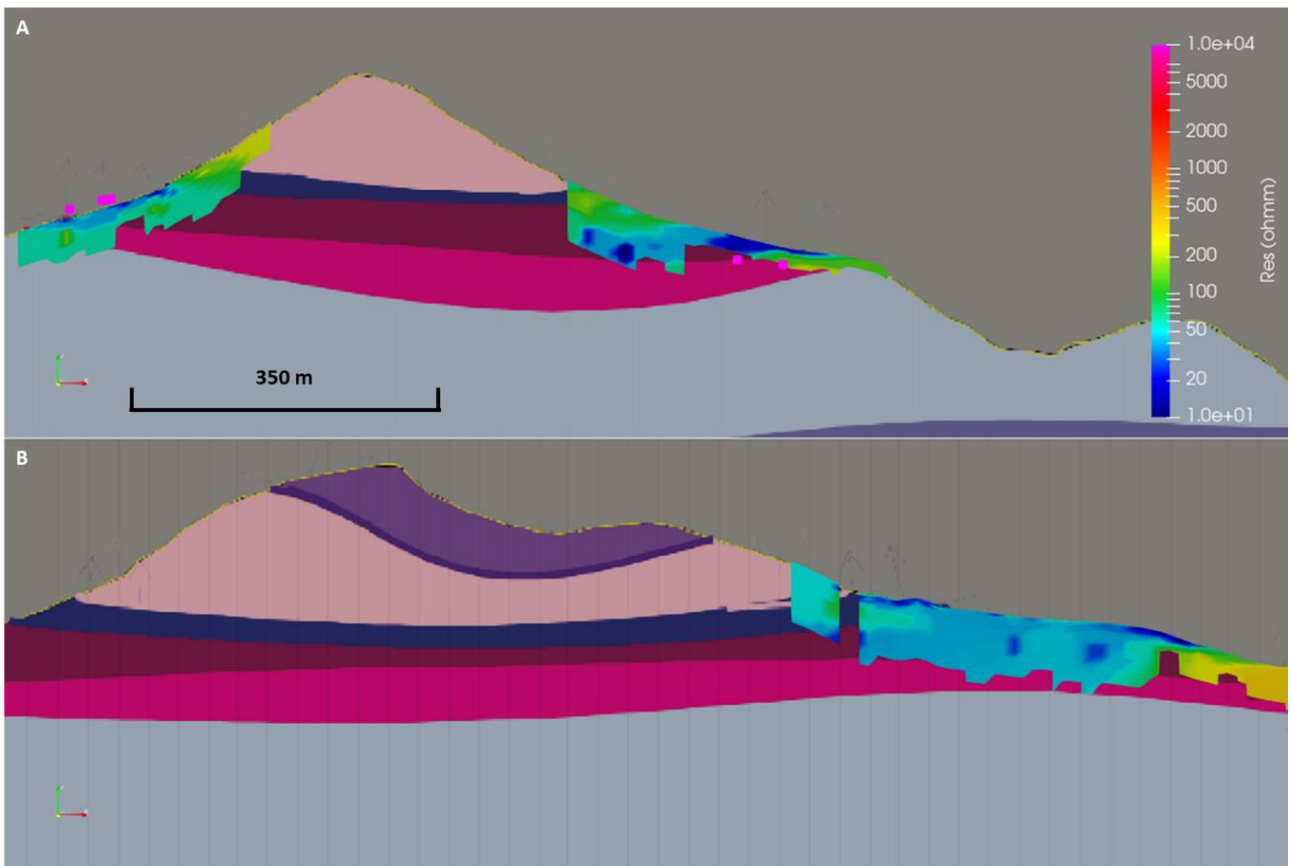


Fig. 4.44 Two cross sections of the modelled geology intersecting Mt. Ario (A: E_I, B: E_II according to the Fig. 4.2), overlaid with the Loupe resistivity model obtained using standard Sharp inversion cropped with DOI. Purple dots are the springs of interest.

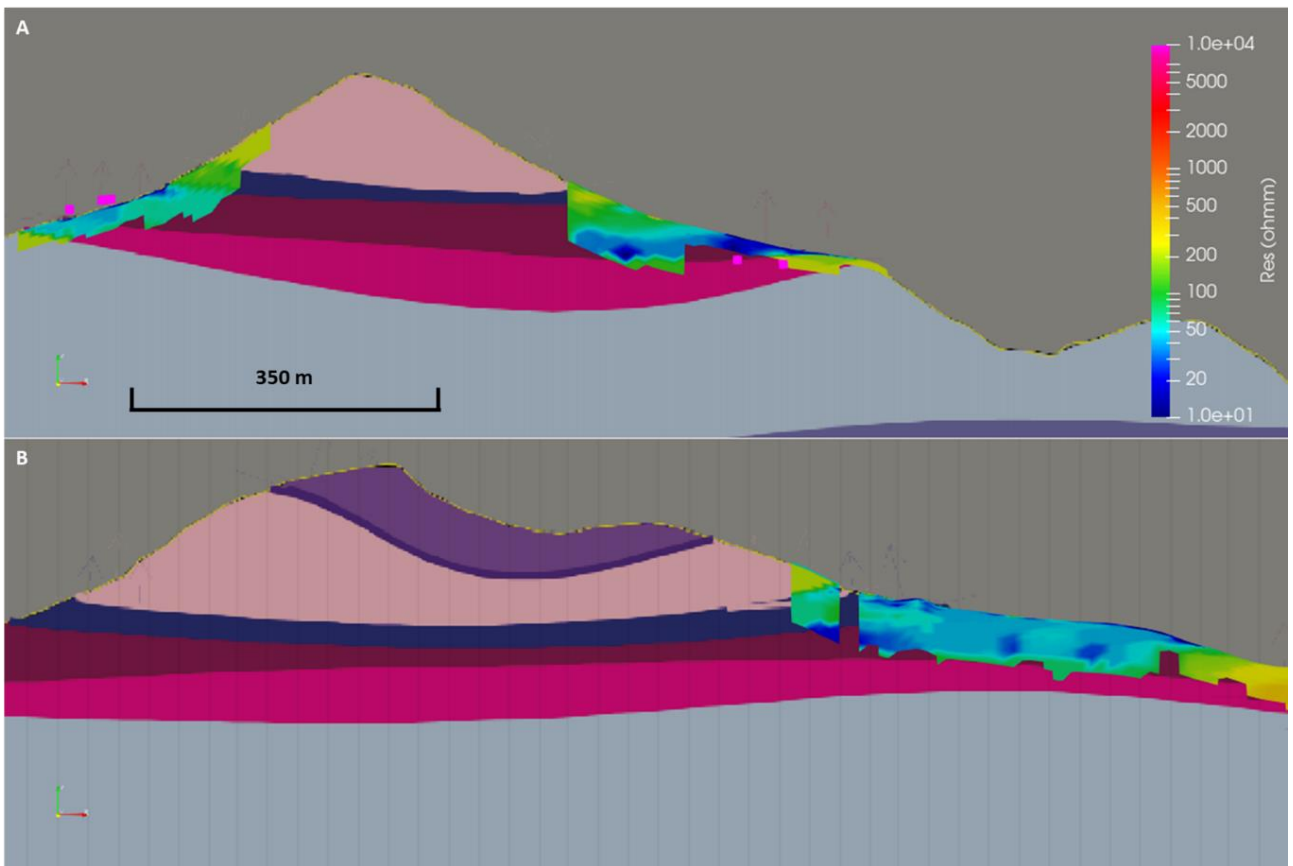


Fig. 4.45 Two cross sections of the modelled geology intersecting Mt. Ario (A: E_I, B: E_II according to the Fig. 4.2), overlaid with the Loupe resistivity model performed using geologically-constrained Sharp inversion cropped with DOI. Purple dots are the springs of interest.

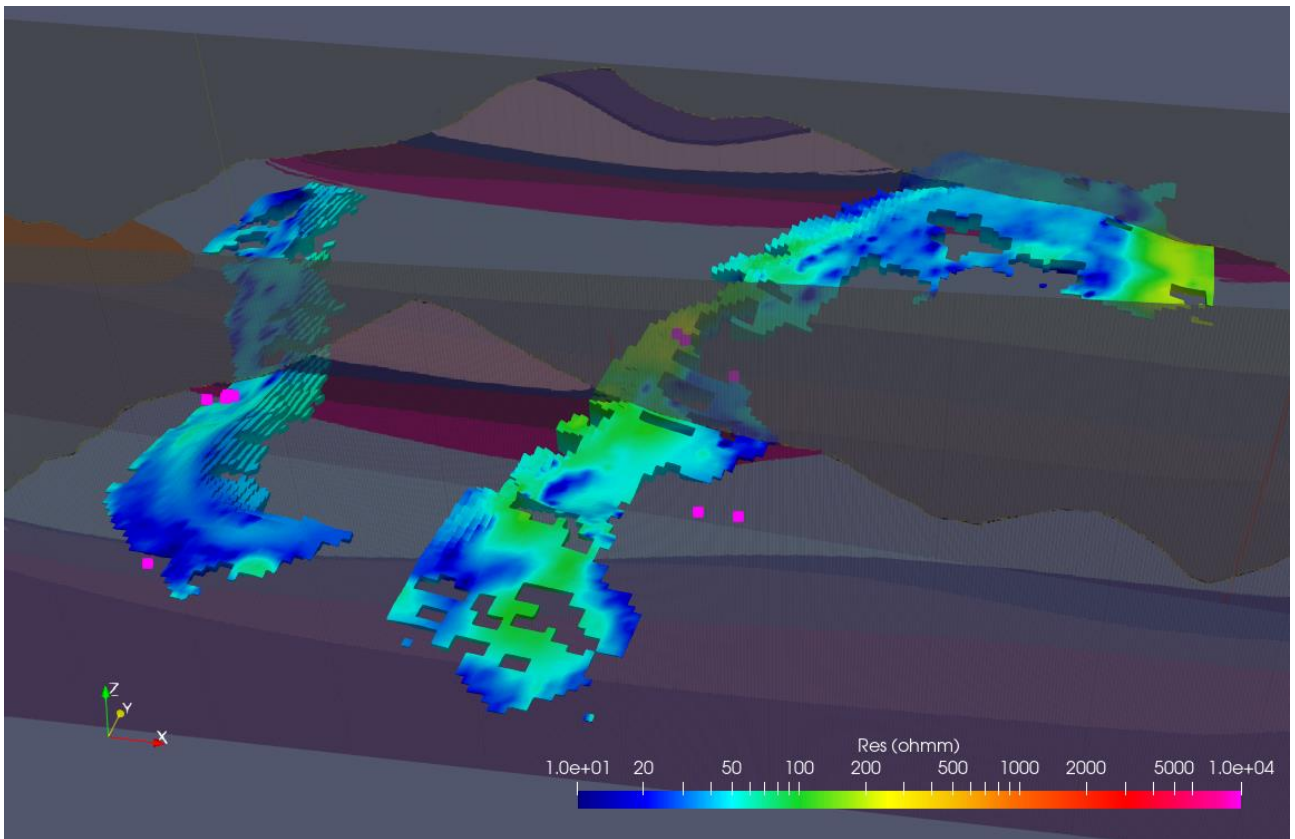


Fig. 4.46 3D Loupe resistivity model of the SAB formation, illustrating its internal heterogeneity. Two semi-transparent cross sections of the modelled geology (E_I in the foreground and E_II in the background, from Fig. 4.2) are included for reference. Purple dots are the springs of interest. Y axis is the north.

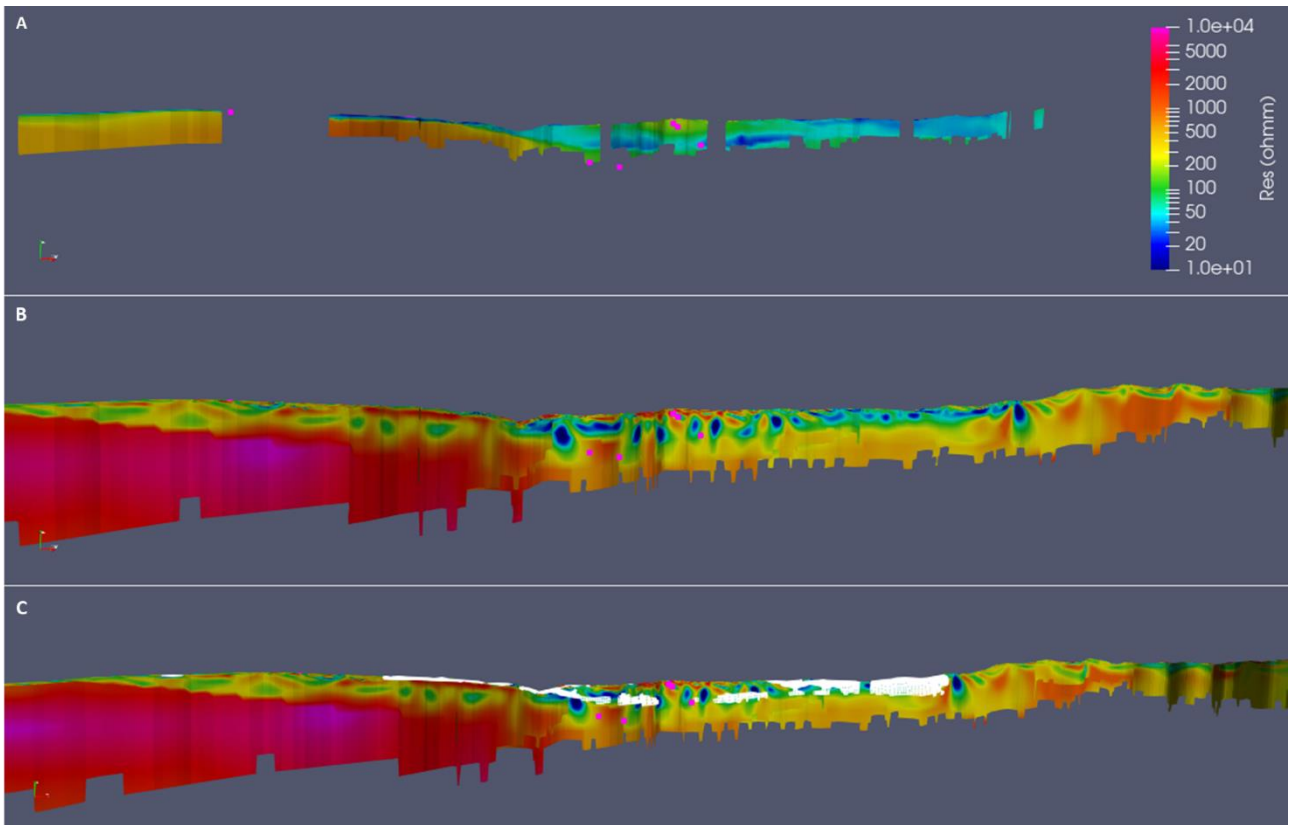


Fig. 4.47 Comparison between DC and Loupe resistivity models, both cropped with DOI. A: Loupe resistivity model sliced along the DC profiles. B: DC resistivity model. C: DC resistivity model with overlaid conductive zones ($< 50 \Omega m$) from the Loupe model shown in white. Purple dots are the springs of interest.

4.3.3 Synthetic AEM models

Since AEM data for the Val Sabbia study area were not made available, a synthetic modelling approach was developed to investigate the potential geophysical response of the subsurface using the 3D geological model as structural reference (Fig. 4.48). This synthetic model was designed to simulate how the subsurface in the Acqua Bianca spring sector of the study area would appear in AEM surveys. To achieve this, each geological unit represented in the 3D model was assigned a resistivity value (Tab. 4.14) selected based on typical ranges found in the literature for lithologies with similar characteristics (Fig. 4.49). To simulate features of particular hydrogeological interest, three discrete conductive elements were introduced into the model to represent possible karst conduits (Fig. 4.50). These features are considered plausible, also supported by the hydrograph of the Acqua Bianca spring (as explained in Section 4.2.3). They were placed within the Esino Limestone formation (ESI), which is known to be prone to karstification, and aligned along the main structural discontinuity in the area, the Pofferratte Fault. A buffer zone was included around each conduit to account for the likely presence of weathered or altered rock surrounding the main conduit, thereby better approximating the conductive signature such features would produce in an AEM survey

Region	RES (Ωm)
ANG	500
PRZ	900
BUC	900
WEN	5000
ESI	5000
LSR	100
SAB	100
SGB	1000
DPR	1000
BOV	500
SRV	900
MAL	500
MAL1	100
MAL2	100
VVI	900
OFOL	100
MVX	5000

Tab. 4.14 Resistivity values for each geological unit modelled (formation acronym from Fig. 4.1).

This approach provides a realistic geophysical scenario that could be used to evaluate the potential response of the AEM method in a complex, karst-fractured mountain environment. The synthetic AEM data, generated from a hypothetical regular grid of flight lines (Fig. 4.51), provided insights into how different geological features such as faults, karst conduits, and interfaces between geological units with different electrical behavior might be represented in resistivity sections. Two different forward modellings were carried out: a standard 1D modelling and a full 3D modelling, both considering a constant flight altitude of 100 m, a typical value in mountain areas where a higher ground clearance is required for safety concerns. Three main reasons motivated the decision to run the 3D modelling: i) the significant topographical variability of the area, exacerbated by the high ground clearance and the large footprint of AEM systems; ii) the strong resistivity contrasts among geological formations, as mapped by the DC and Loupe data collected in the adjacent Mt. Ario sector; iii) the relatively small dimension of the modelled karst conduits, compared to the system footprint.

Fig. 4.52 shows the comparison of 3D and 1D forward responses for line 27, placed above the modelled karst conduits. Very strong differences are already evident in the data, with much weaker signal in 3D in correspondence to the karst conduits. Both 1D and 3D forward data were contaminated with gaussian noise before inversion, with an error model composed by 3% uniform noise plus a $t^{-\frac{1}{2}}$ time-diminishing absolute noise with value of 1 nV/m² at 1 ms. Data with total standard deviation above 30% were culled out before inversion, as shown in Fig. 4.52D for 3D forward data.

Both 1D and 3D forward data were inverted in 1D, with a standard inversion approach as well as with the geologically-constrained inversion, with constraint-breaking across geological interfaces, but not applying formation-specific resistivity ranges as with the ground field data. Furthermore, both L2 and Sharp regularizations were used for the vertical/horizontal constraints, to highlight the effectiveness of the geologically-constrained inversion also in comparison to standard inversion with regularizations that allows for sharp resistivity changes across formations.

All the inversion results are summarized in twelve figures, as described in the following.

- Fig. 4.53 and Fig. 4.54 present the sharp geologically-constrained inversion result in 3D view, for the inversions corresponding to 1D and 3D forward, respectively. While the 1D inversion of 1D forward data shows an excellent correspondence with the true model, the 1D inversion of 3D forward data shows evident artifacts: mountain tops and crests show overestimated resistivity values; conductive surficial formations show

underestimated resistivity values in the valleys, with evident striping in correspondence to the acquisition lines.

- Fig. 4.55 to Fig. 4.59 show the 1D inversions of 1D forward responses along profiles 1, 18, 26, 27 and 29. In all these figures the quality of the reconstruction can be ranked, from the lowest to the highest, as: L2 standard inversion; Sharp standard inversion; L2 geologically-informed inversion; Sharp geologically-informed inversion. Overall, the geologically-informed inversion presents a better reconstruction of deep formations, as well as formation boundaries. The simulated karst conduits are very well reconstructed, as well as the surrounding buffer areas simulating the weathered zone.
- Fig. 4.60 to Fig. 4.64 show the 1D inversions of 3D forward responses along profiles 1, 18, 26, 27 and 29 (the same shown for 1D inversions of 1D forward responses). These inversions, despite giving a decent overall understanding of the complex modelled geology, present significant artifacts that might lead to risky misinterpretations. The main artifacts, similarly to what shown in Fig. 4.54, can be summarized as follows: i) resistivity values at mountain tops and ridges are overestimated because they also include the contribution of the surrounding air, which is highly resistive; ii) in contrast, valleys appear much more conductive, as the surrounding environment consists of rocks rather than air; iii) thicknesses of conductive formations are often underestimated; iv) resistive mountain tops or ridges trigger significant artifacts in the form of deeper conductive anomalies, which act as compensating artifacts for the artificially high resistivity values caused by the surrounding air; v) small conductive conduits are poorly imaged, without a proper geometry recognition but with a weak decrease of resistivity. Nevertheless, the lack of correspondence between the imaged geometries and the true resistivity distribution in the 1D inversions of the 3D forward responses shows that the geologically-informed inversions do not introduce artifacts at the formation boundaries when the data do not support resistivity change across these boundaries.
- Fig. 4.65 to Fig. 4.69 show a comparison between different 1D geologically-constrained inversion results of 3D forward responses. Panel A presents the inversion results using the full dataset, with the corresponding misfit shown in Panel B. Panel C displays the inversion results obtained after excluding data points with a misfit greater than 2, and Panel D shows the resulting misfit. By comparing Panel A and Panel B, it is evident that the data fit is poorer at mountain tops and valley bottoms. Where a conductive layer does not continue but instead terminates into air (such as at mountain tops) high data

misfits are observed. When data points with a high misfit are removed, the resulting model (Panel C) shows no significant changes and it still retains most of the artifacts; however, the 1D inversion achieves a better overall data fit (Panel D). This outcome demonstrates that removing high-misfit data does not suffice in avoiding 3D artefacts, highlighting the necessity of performing 3D inversions for more reliable and accurate subsurface characterization in areas with high topography variations. In Fig. 4.70 a map of the data fit is reported for the L2 and Sharp 1D inversion of the 3D forward model of the entire dataset (A and B respectively) and for the dataset without the high misfit data (C and D respectively).

Tab. 4.15 provides a comparison of the overall misfit values obtained for each inversion approach. Both L2 and Sharp misfit measures are reported for all strategies.

All these findings highlight the importance of 3D AEM modelling in mountain areas in presence of significant resistivity contrasts, and might explain the lack of success in identifying karst conduits with AEM reported in literature: the 1D inversion is not able to retrieve geometry of conductive conduits, but only an overall decrease in resistivity in their surroundings. The effectiveness of the geologically-informed 3D inversion of AEM data is proved for field data in the next section, where the results of the Aracena study site are presented.

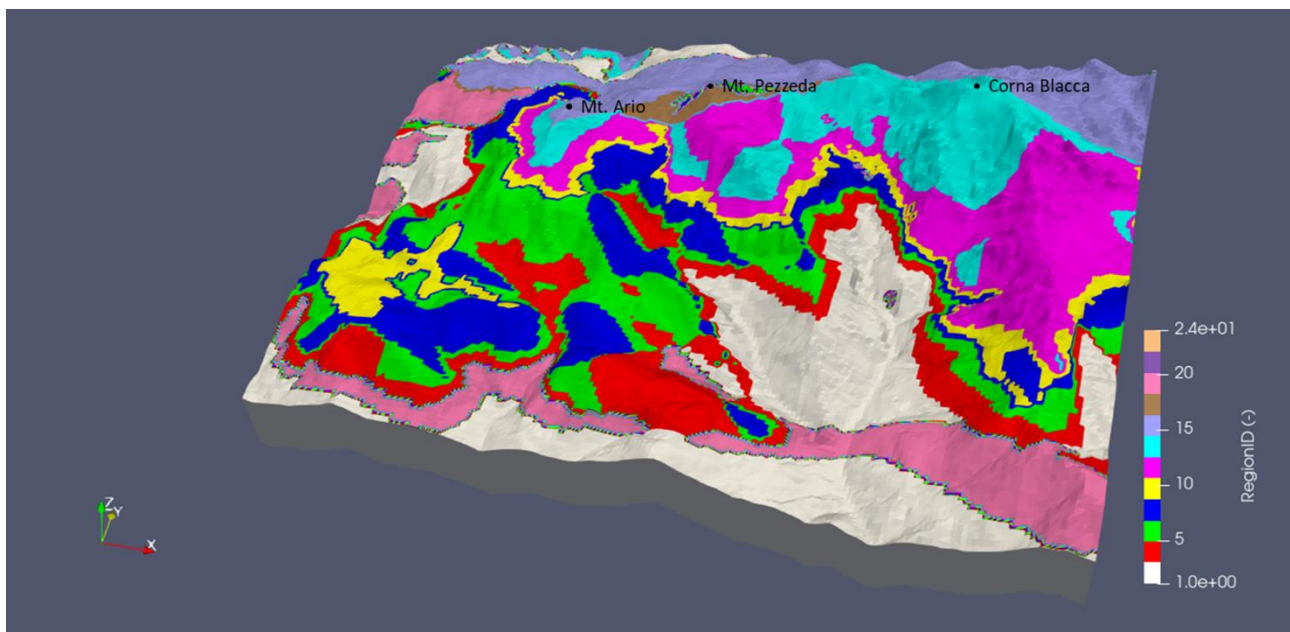


Fig. 4.48 3D geological model. All the modelled geological units are represented with different colors. Although the colors differ, this model can be compared with the one in Fig. 4.12 and its corresponding legend in Fig. 4.10 to understand which colors correspond to which formations. Y axis is the north.

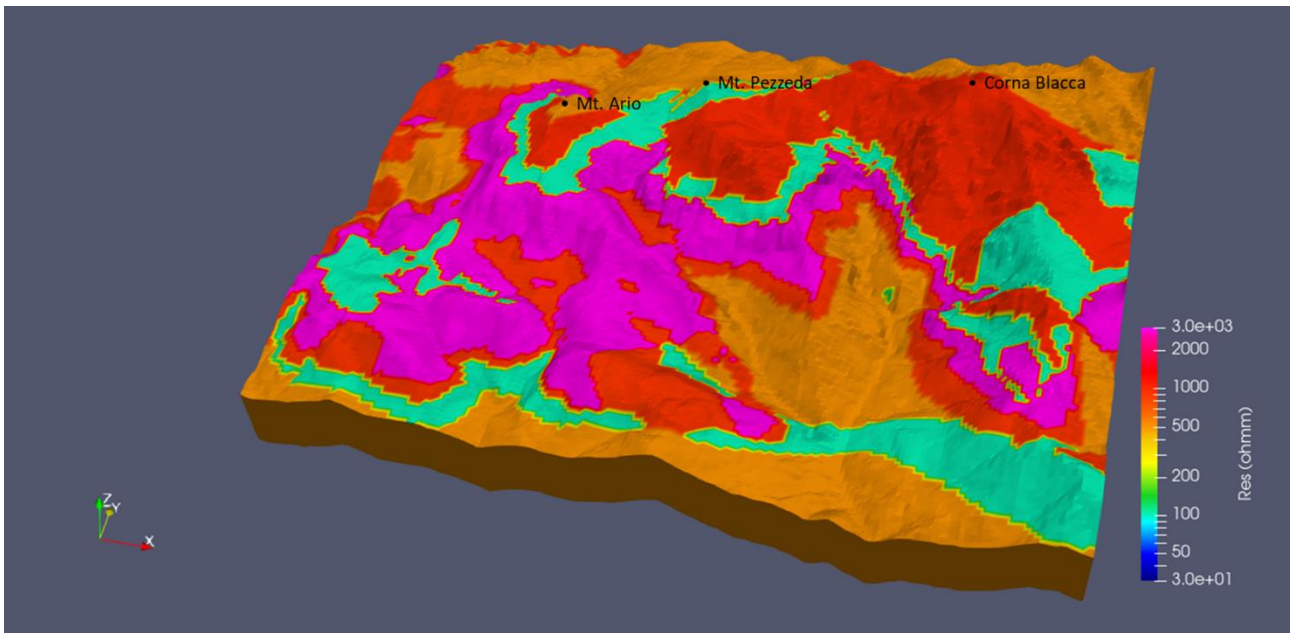


Fig. 4.49 3D resistivity distribution model based on the 3D geological model. Y axis is the north.

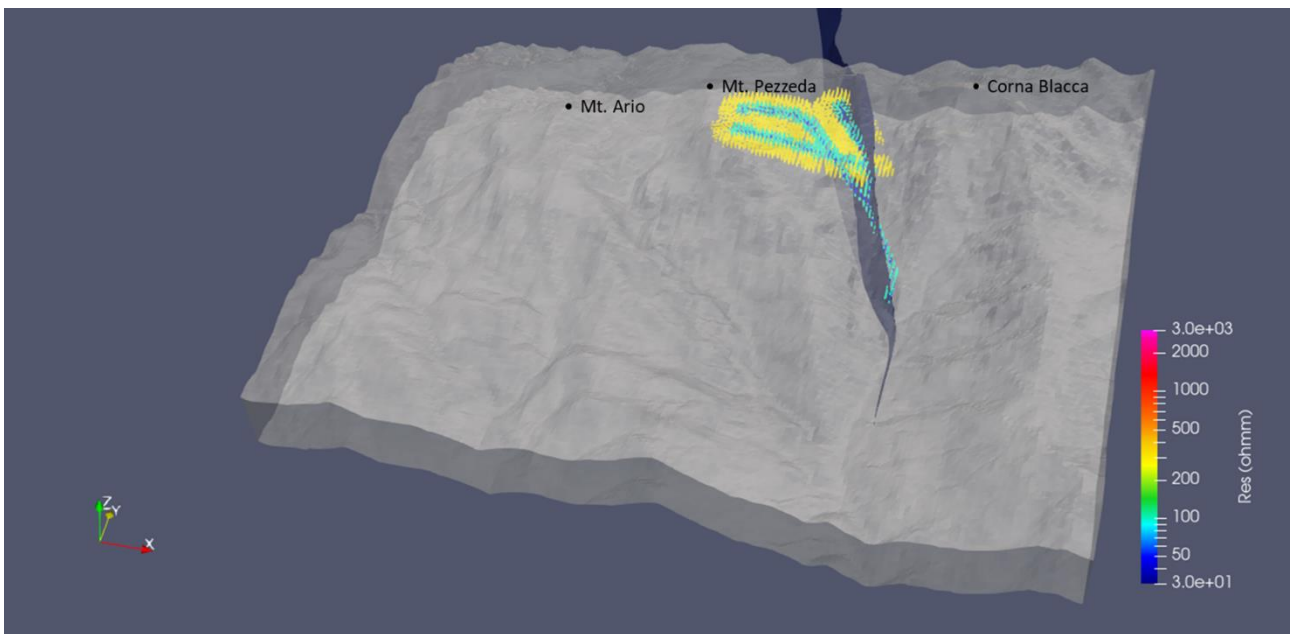


Fig. 4.50 3D model of the conduits represented together with the modelled Pofferatte fault. Y axis is the north.

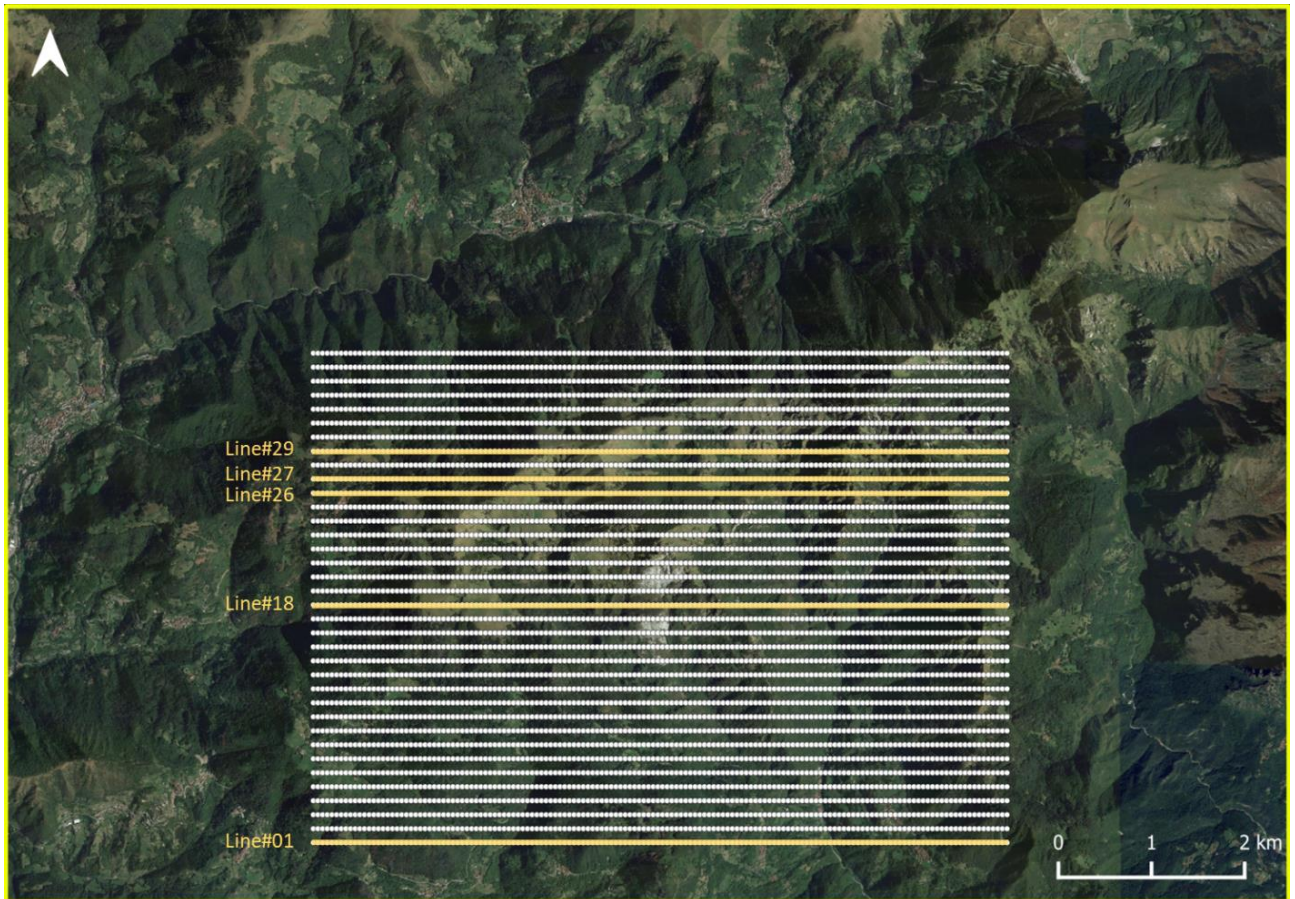


Fig. 4.51 Map of the AEM synthetic survey lines within the study area frame. The lines presented in the next figures are highlighted.

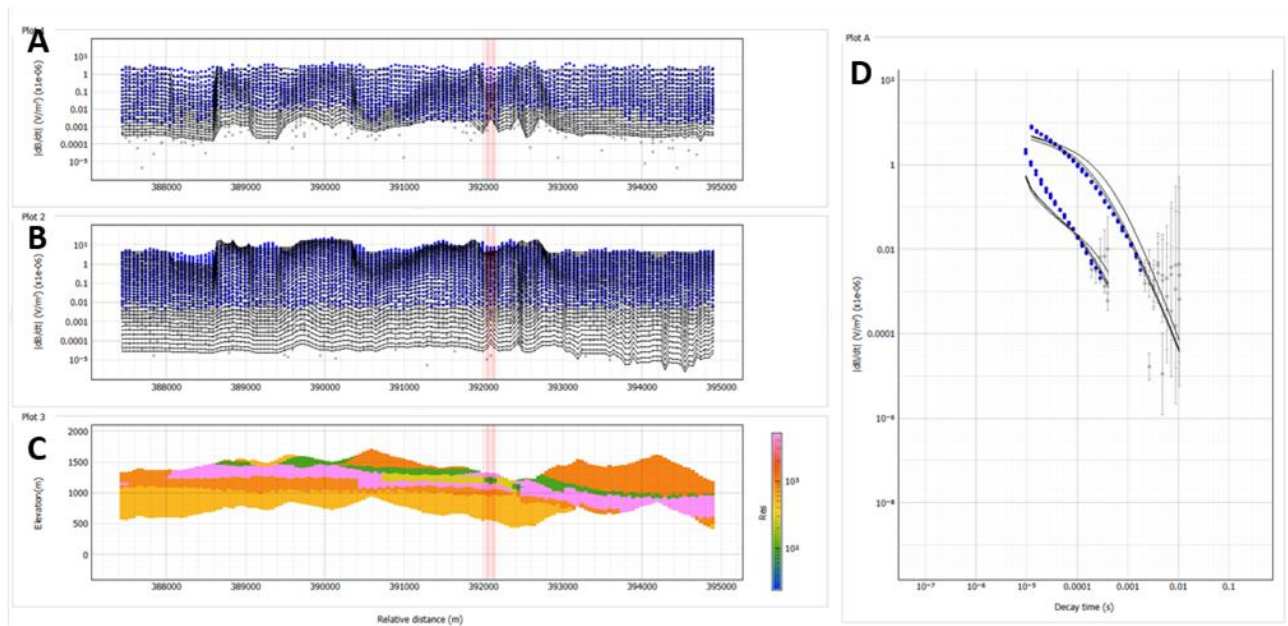


Fig. 4.52 Comparison of 3D and 1D forward responses along line 27, for low and high moments (A and B, respectively), and corresponding model (C). D: soundings highlighted in red in the data stripes. 3D data

(blue markers) are contaminated by gaussian noise, and data with standard deviation above 30% (grey markers) have been culled out. Black lines represent the 1D forward response.

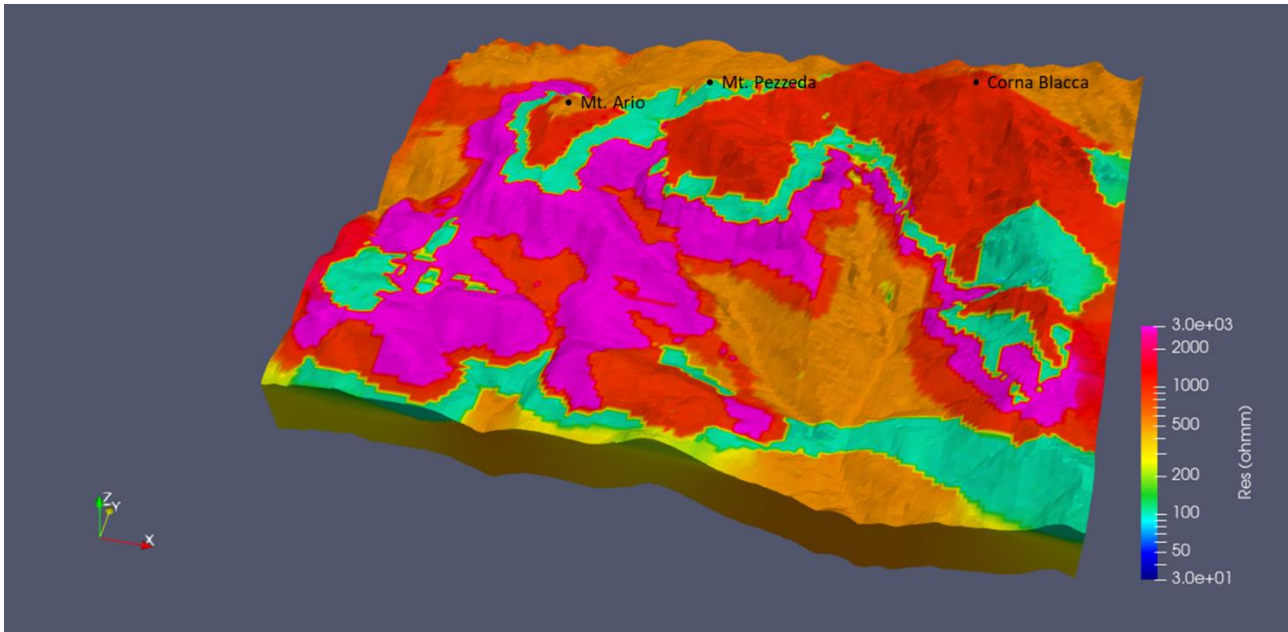


Fig. 4.53 1D Sharp inversion of 1D forward model. Y axis is the north.

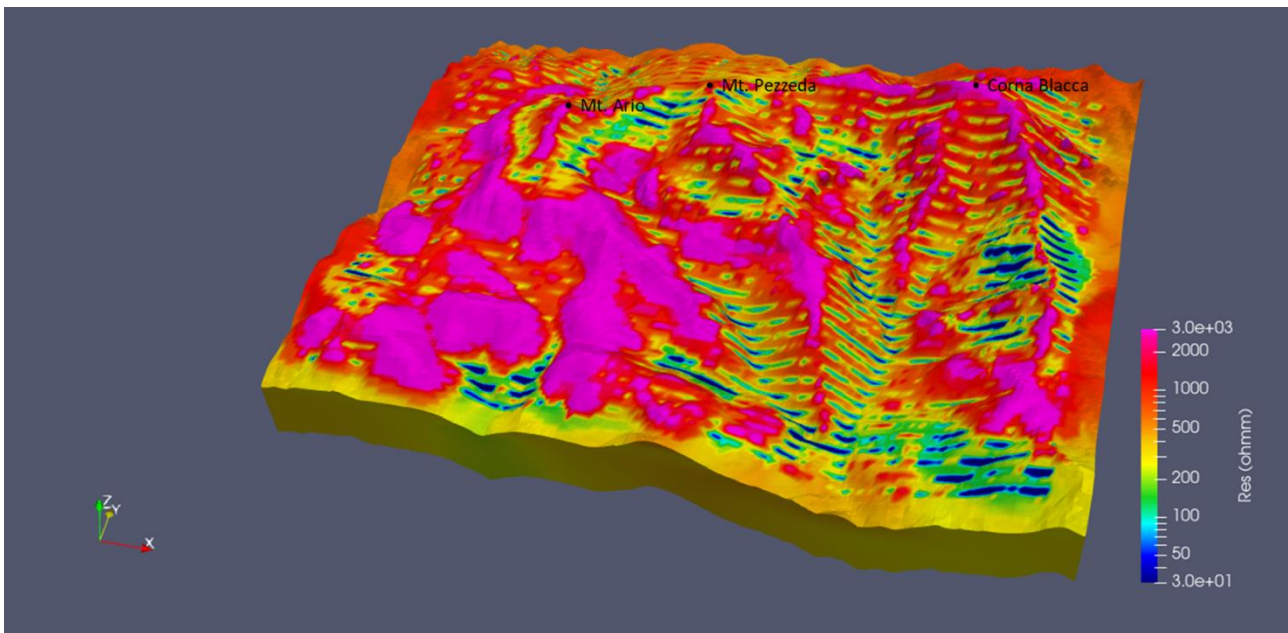


Fig. 4.54 1D Sharp inversion of 3D forward model. Y axis is the north.

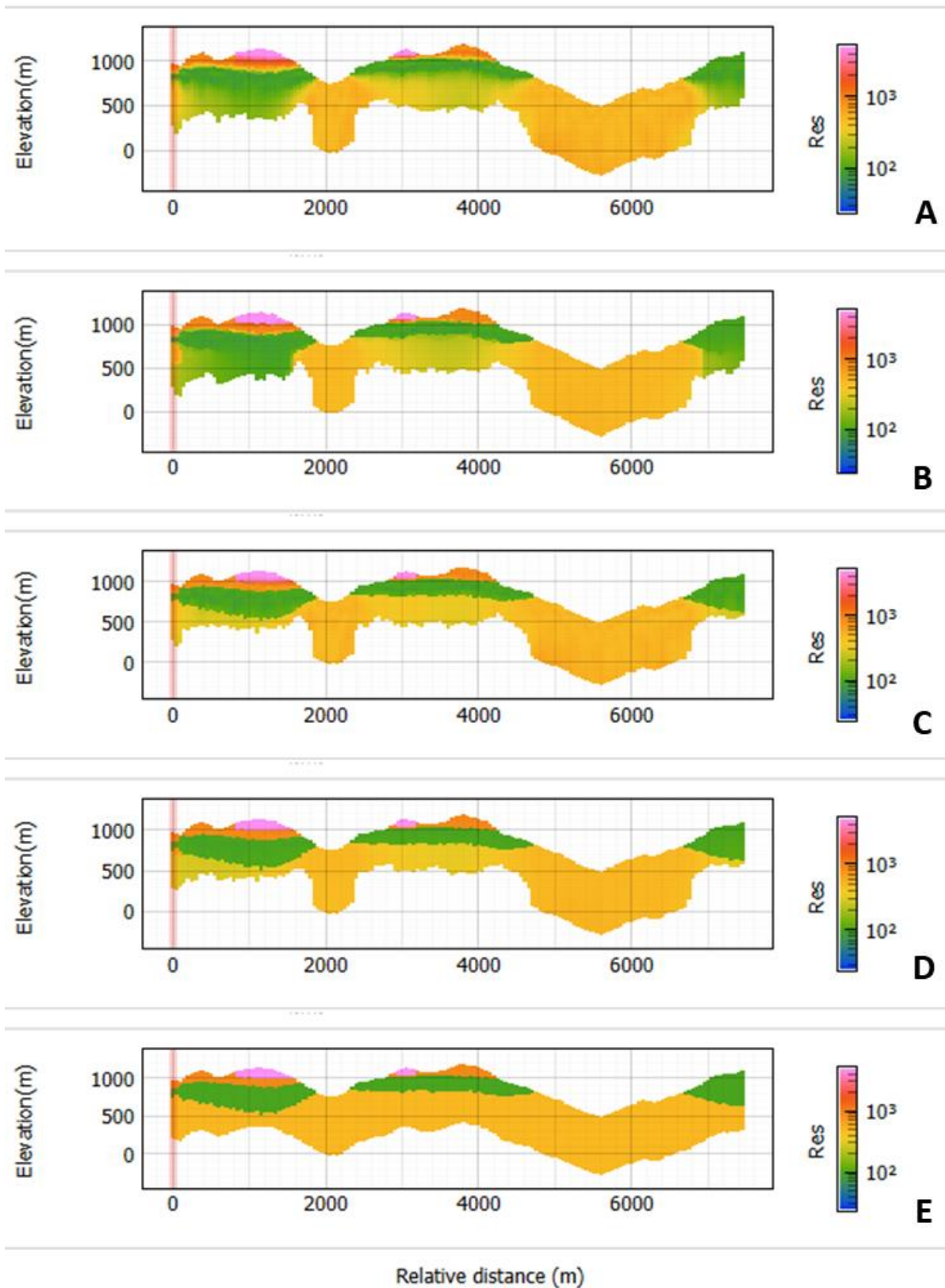


Fig. 4.55 Line 1. 1D inversion of 1D forward model. A: L2 standard inversion. B: Sharp standard inversion. C: L2 geologically-constrained inversion. D: Sharp geologically-constrained inversion. E: true model. All inversion models are cropped with the DOI.

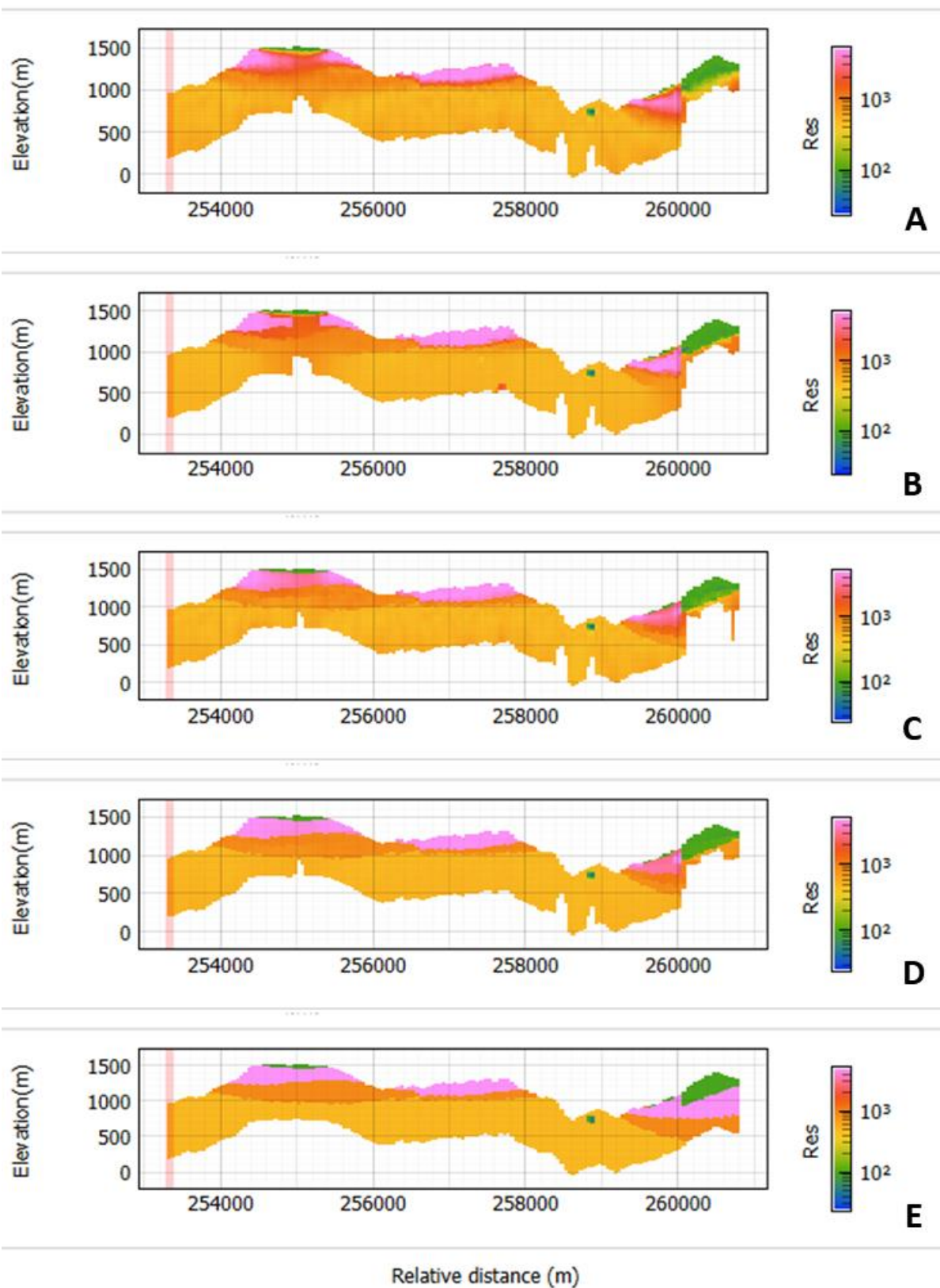


Fig. 4.56 Line 18. 1D inversion of 1D forward model. A: L2 standard inversion. B: Sharp standard inversion. C: L2 geologically-constrained inversion. D: Sharp geologically-constrained inversion. E: true model. All inversion models are cropped with the DOI.

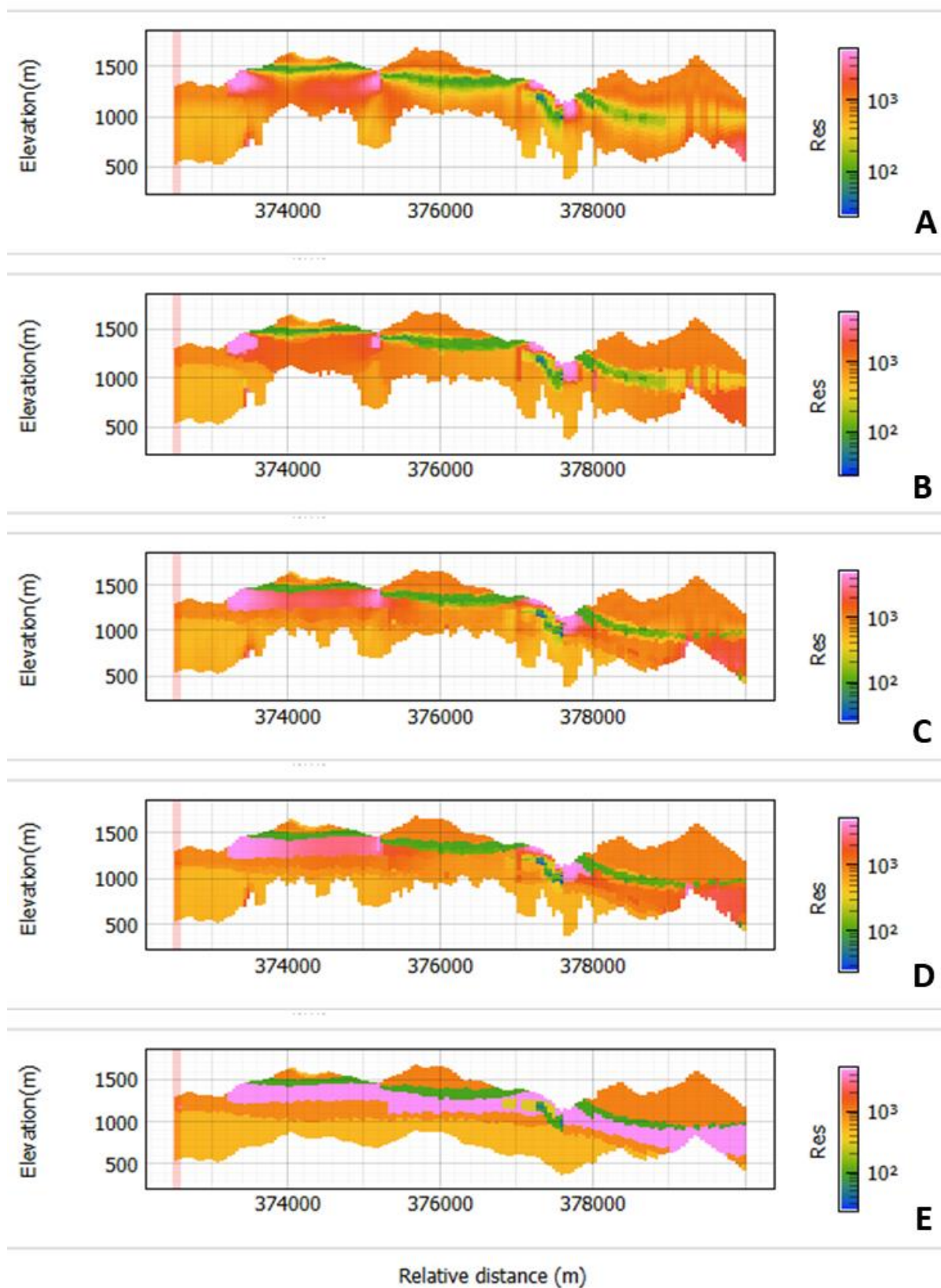


Fig. 4.57 Line 26. 1D inversion of 1D forward model. A: L2 standard inversion. B: Sharp standard inversion. C: L2 geologically-constrained inversion. D: Sharp geologically-constrained inversion. E: true model. All inversion models are cropped with the DOI.

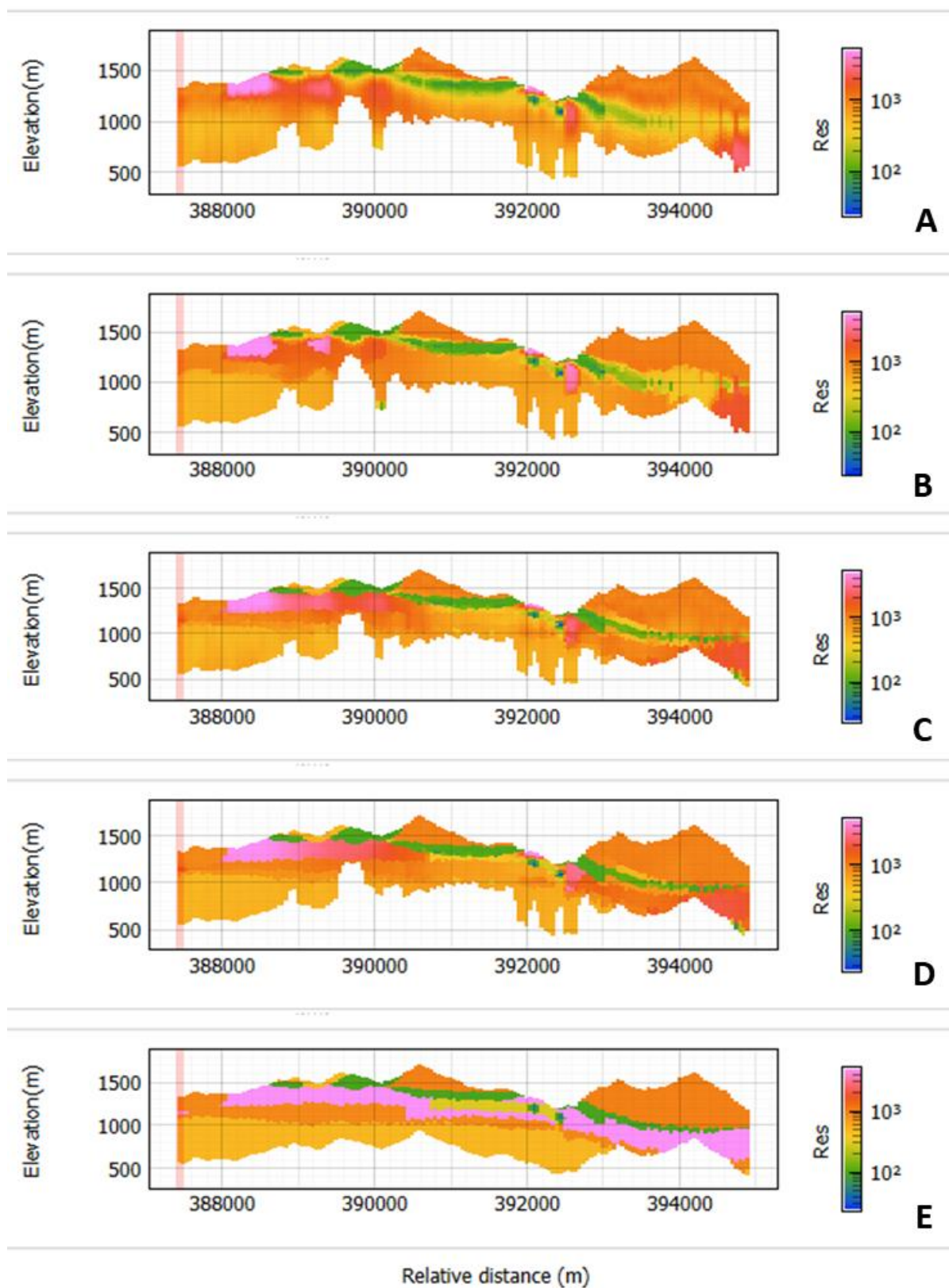


Fig. 4.58 Line 27. 1D inversion of 1D forward model. A: L2 standard inversion. B: Sharp standard inversion. C: L2 geologically-constrained inversion. D: Sharp geologically-constrained inversion. E: true model. All inversion models are cropped with the DOI.

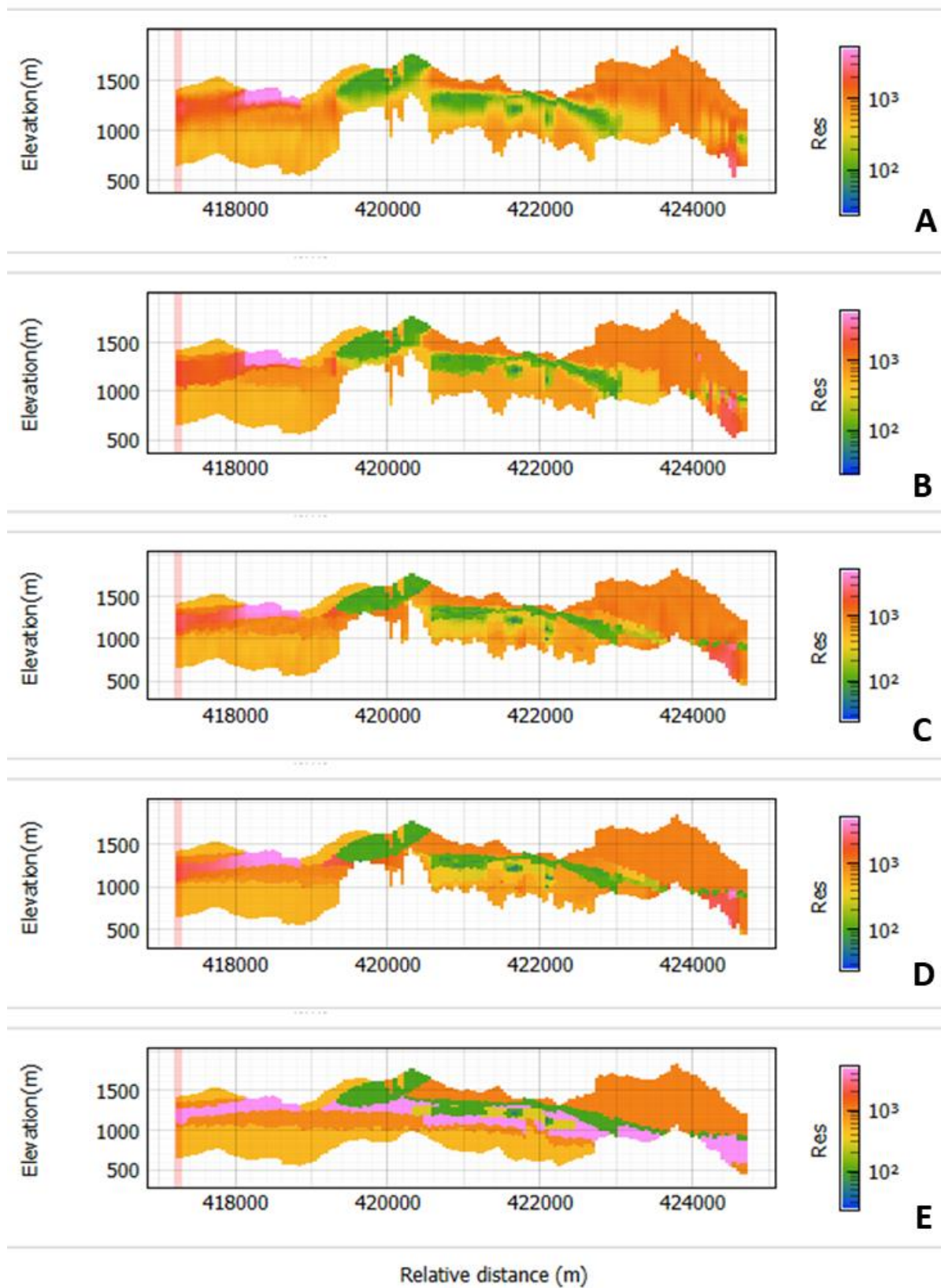


Fig. 4.59 Line 29. 1D inversion of 1D forward model. A: L2 standard inversion. B: Sharp standard inversion. C: L2 geologically-constrained inversion. D: Sharp geologically-constrained inversion. E: true model. All inversion models are cropped with the DOI.

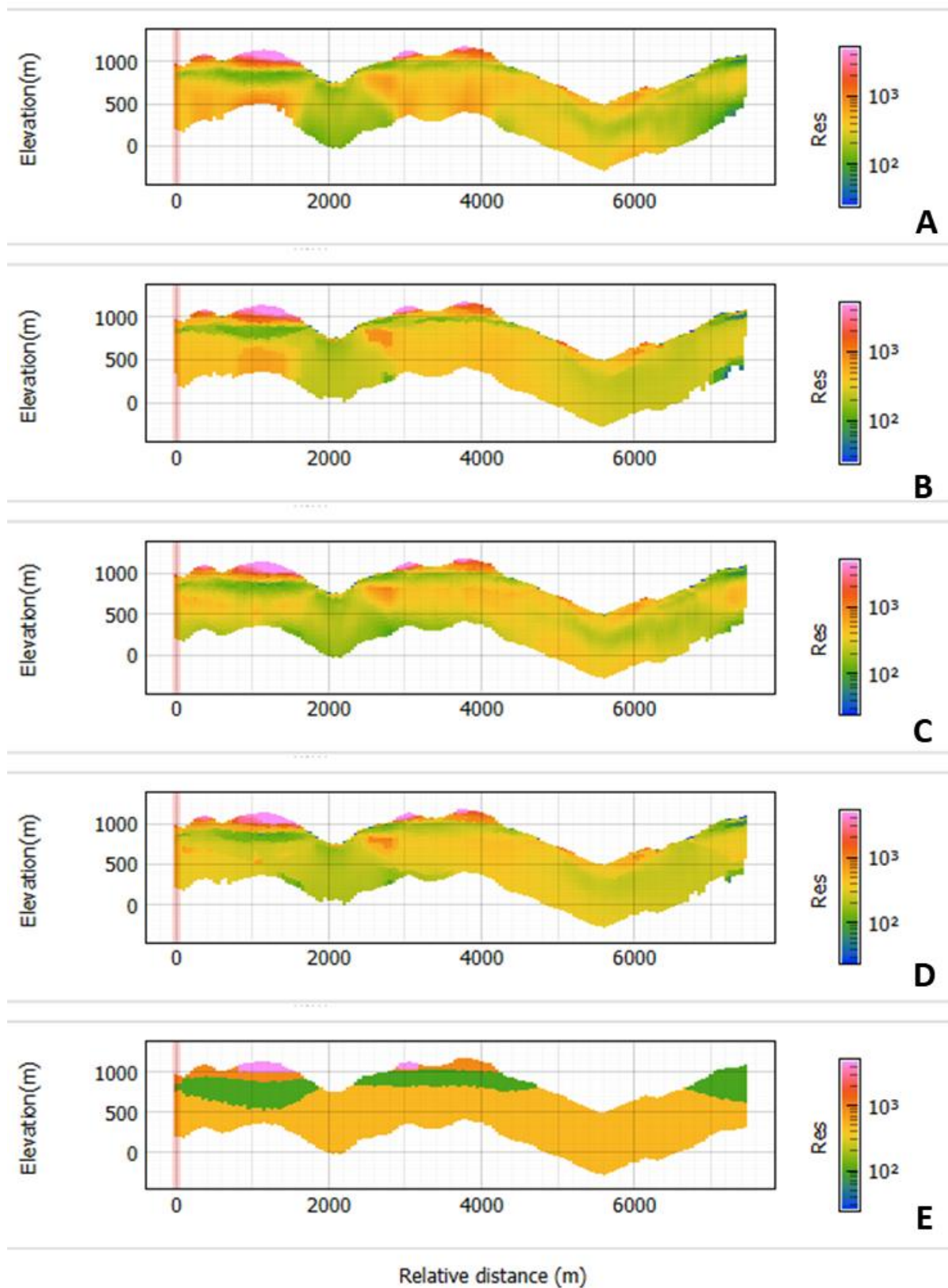


Fig. 4.60 Line 1. 1D inversion of 3D forward model. A: L2 standard inversion. B: Sharp standard inversion. C: L2 geologically-constrained inversion. D: Sharp geologically-constrained inversion. E: true model. All inversion models are cropped with the DOI.

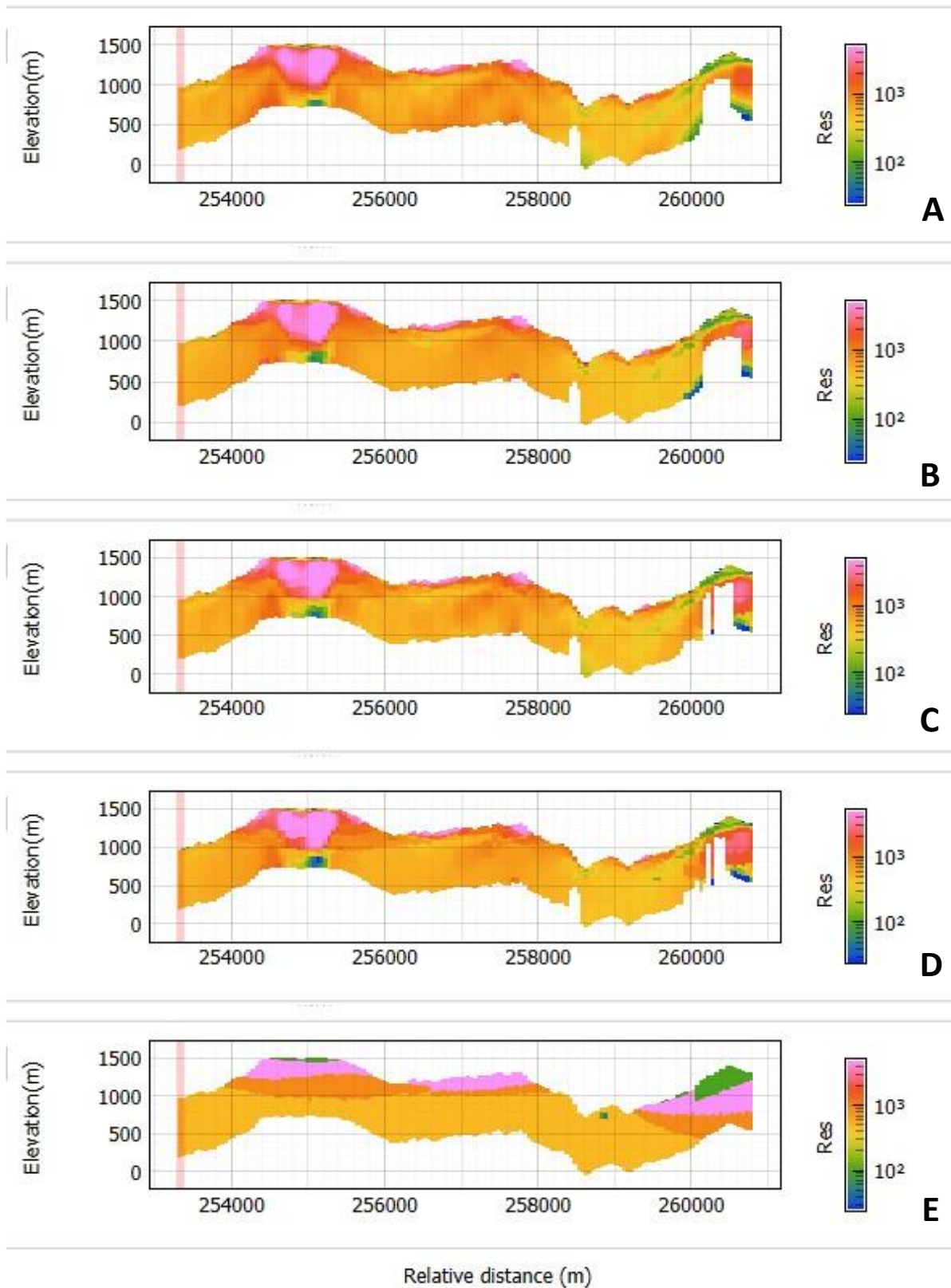


Fig. 4.61 Line 18. 1D inversion of 3D forward model. A: L2 standard inversion. B: Sharp standard inversion. C: L2 geologically-constrained inversion. D: Sharp geologically-constrained inversion. E: true model. All inversion models are cropped with the DOI.

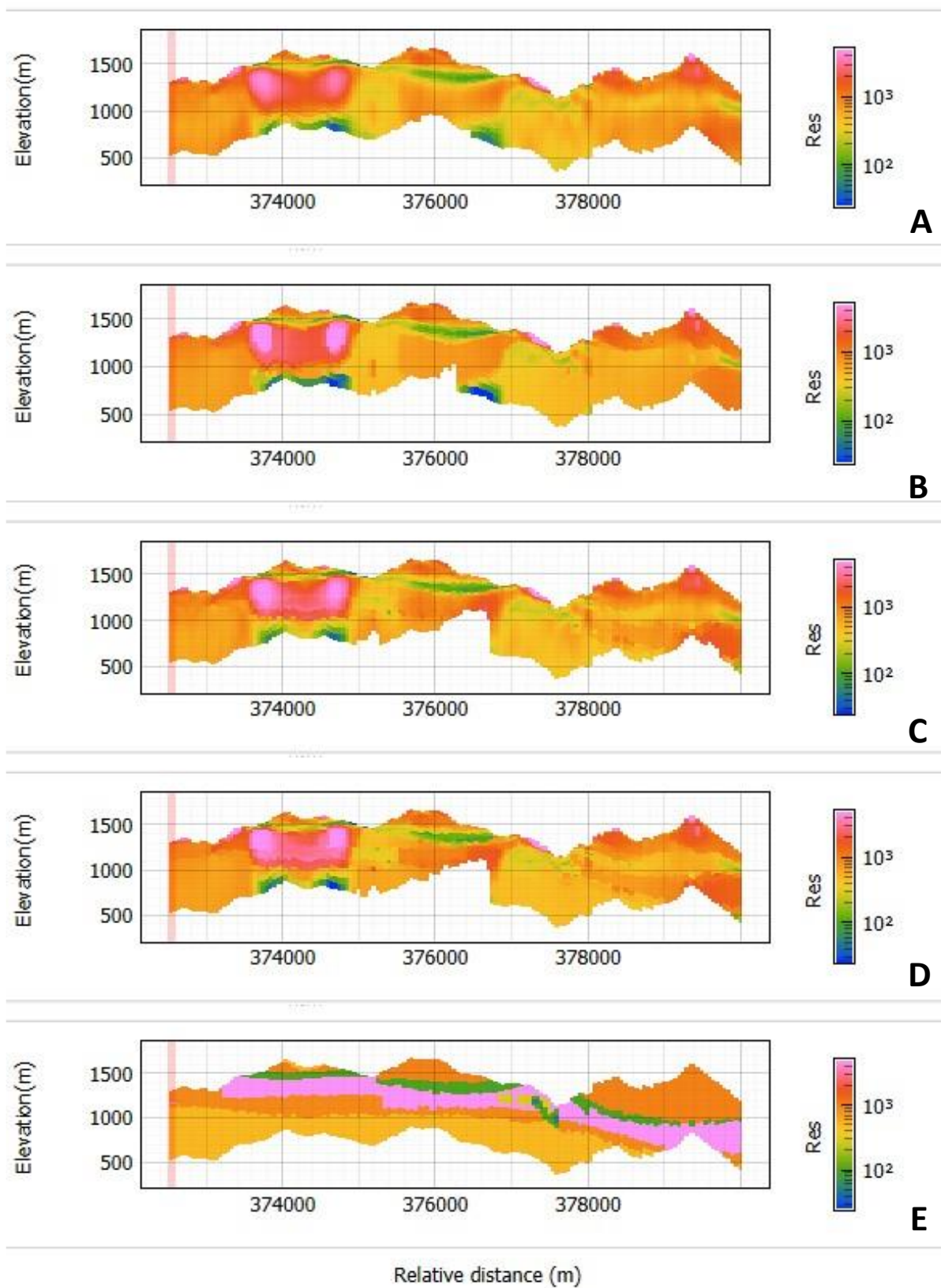


Fig. 4.62 Line 26. 1D inversion of 3D forward model. A: L2 standard inversion. B: Sharp standard inversion. C: L2 geologically-constrained inversion. D: Sharp geologically-constrained inversion. E: true model. All inversion models are cropped with the DOI.

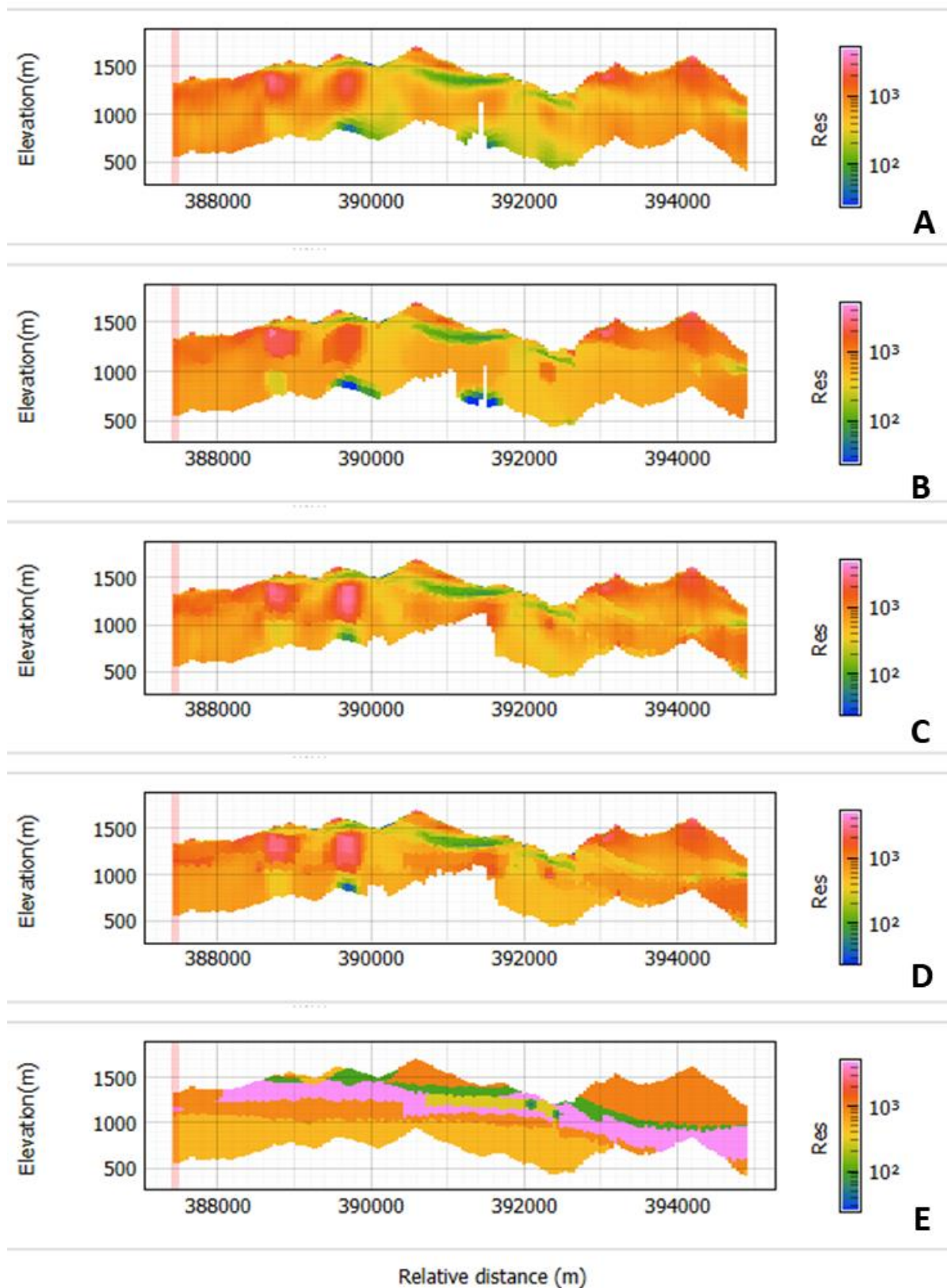


Fig. 4.63 Line 27. 1D inversion of 3D forward model. A: L2 standard inversion. B: Sharp standard inversion. C: L2 geologically-constrained inversion. D: Sharp geologically-constrained inversion. E: true model. All inversion models are cropped with the DOI.

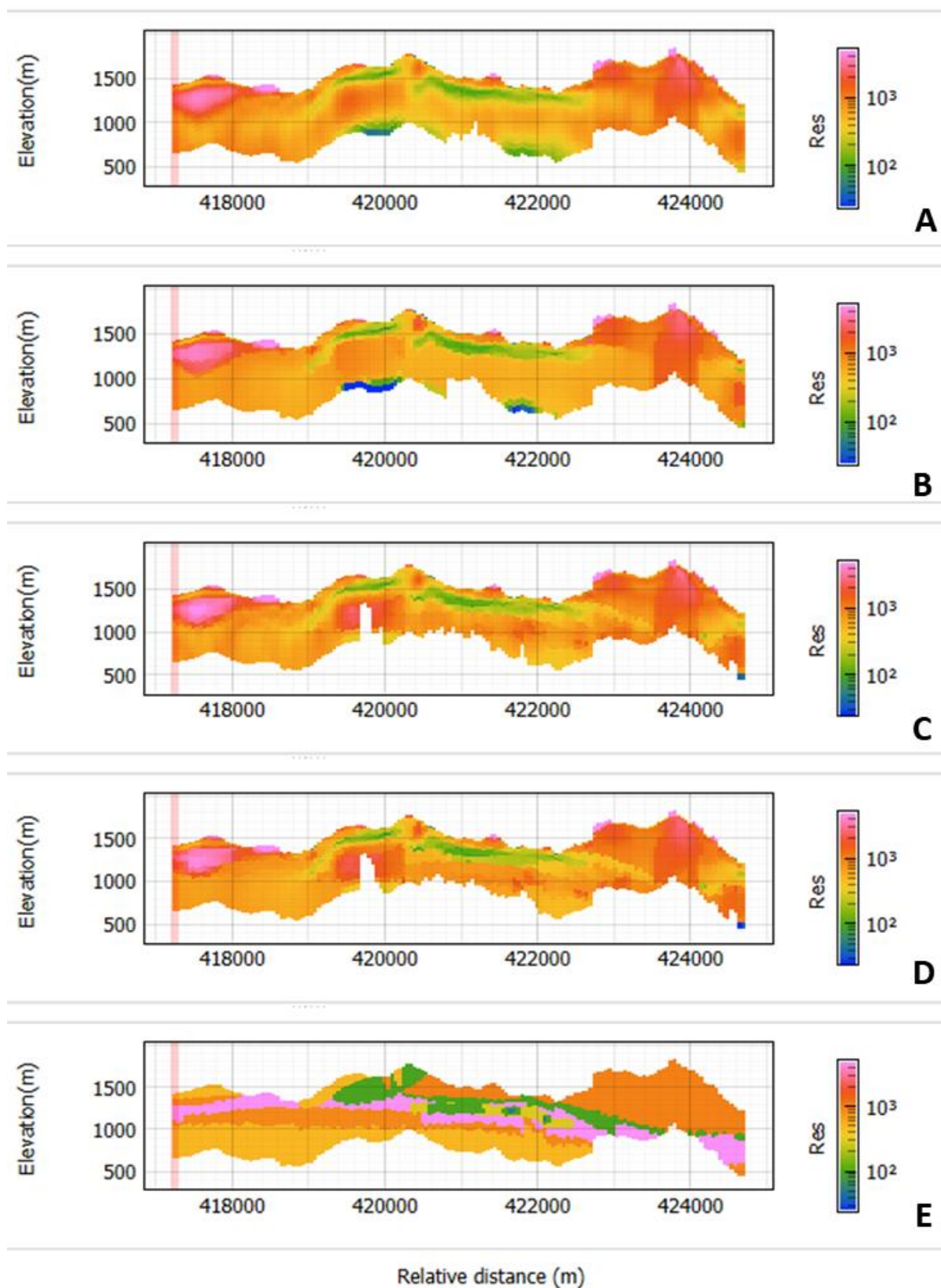


Fig. 4.64 Line 29. 1D inversion of 3D forward model. A: L2 standard inversion. B: Sharp standard inversion. C: L2 geologically-constrained inversion. D: Sharp geologically-constrained inversion. E: true model. All inversion models are cropped with the DOI.

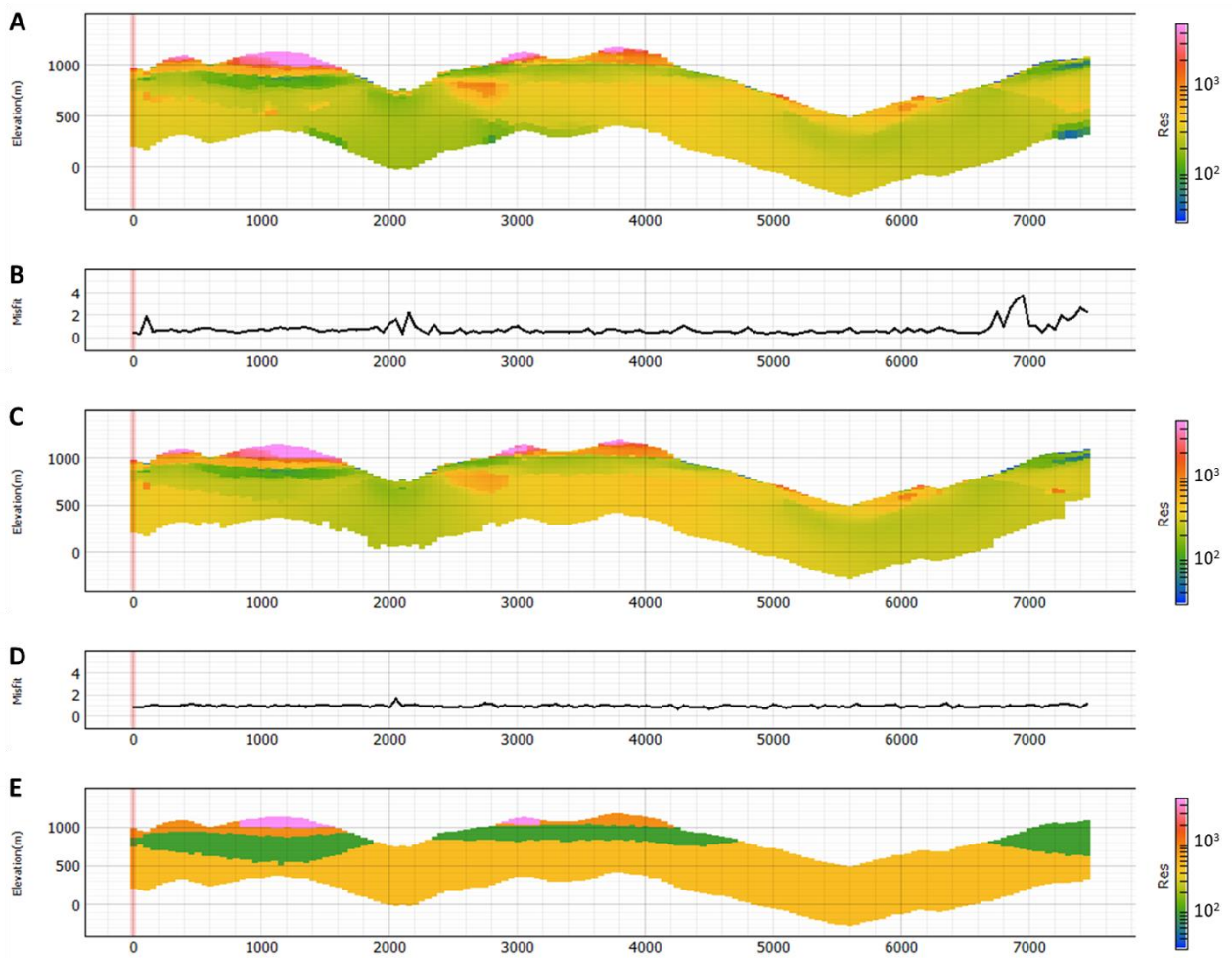


Fig. 4.65 Line 1. A: 1D Sharp geologically-constrained inversion of 3D forward model. B: misfit of A. C: 1D Sharp geologically-constrained inversion of 3D forward model excluding high (>2) misfit data. D: Misfit of C. E: true model.

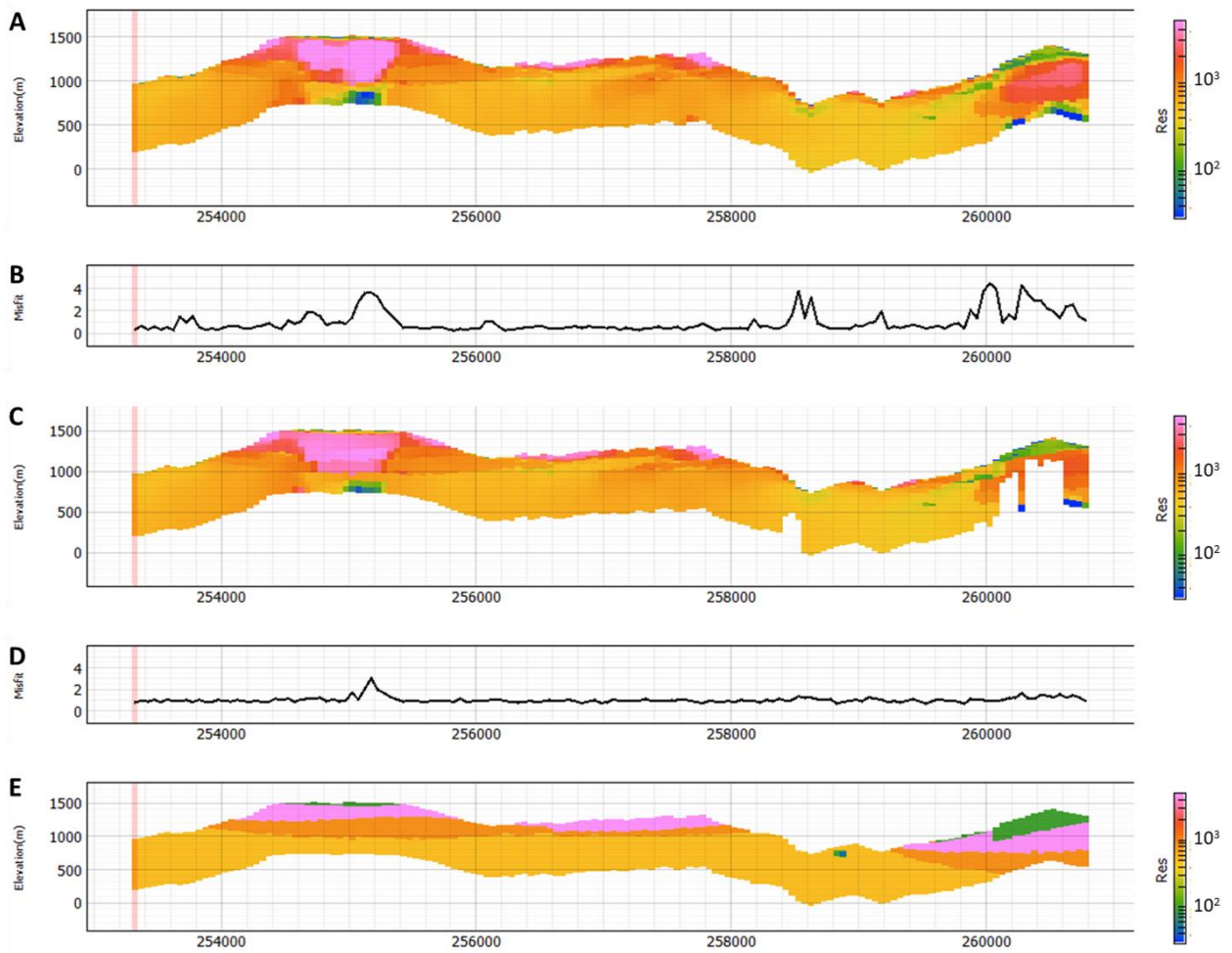


Fig. 4.66 Line 18. A: 1D Sharp geologically-constrained inversion of 3D forward model. B: misfit of A. C: 1D Sharp geologically-constrained inversion of 3D forward model excluding high (>2) misfit data. D: Misfit of C. E: true model.

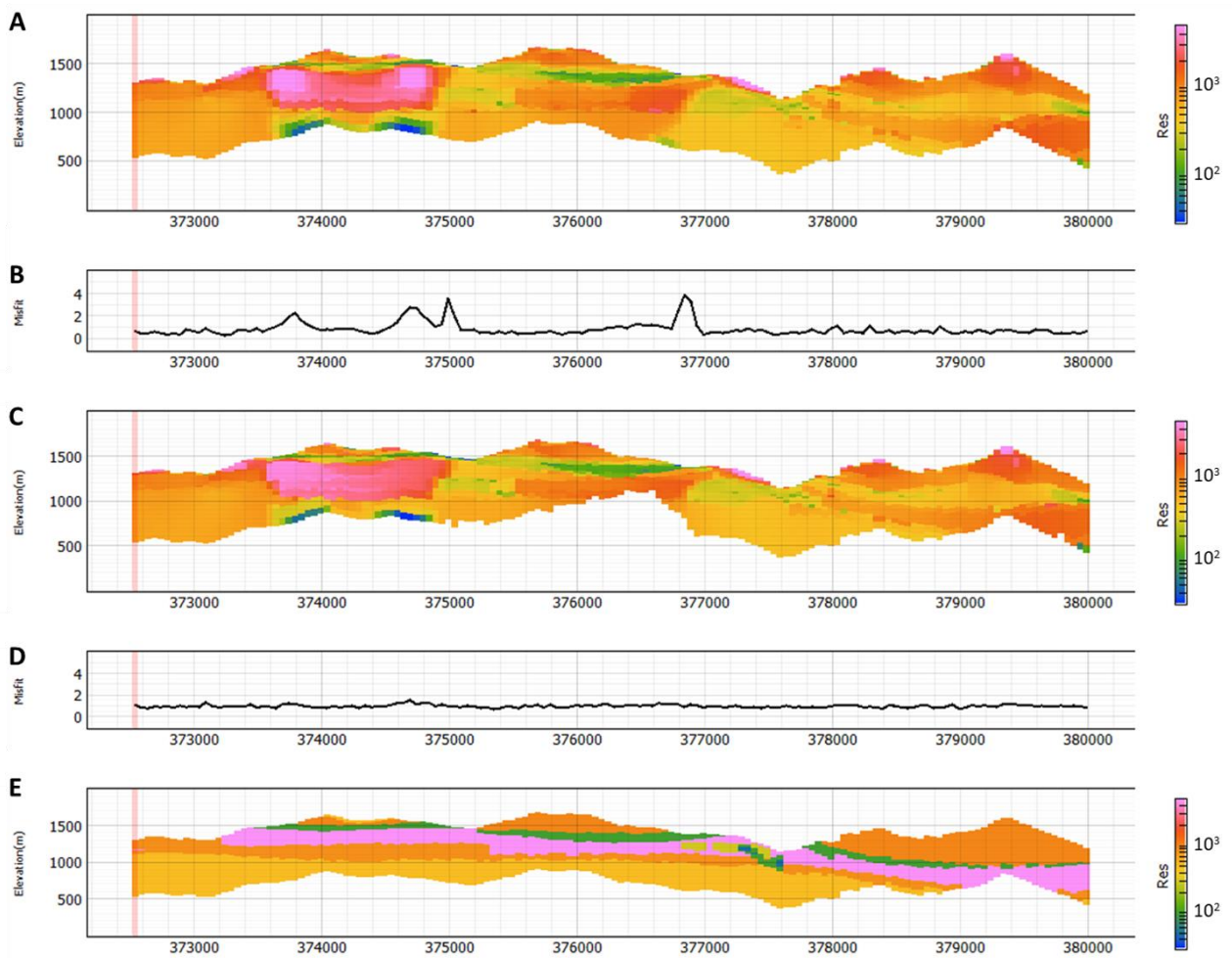


Fig. 4.67 Line 26. A: 1D Sharp geologically-constrained inversion of 3D forward model. B: misfit of A. C: 1D Sharp geologically-constrained inversion of 3D forward model excluding high (>2) misfit data. D: Misfit of C. E: true model.

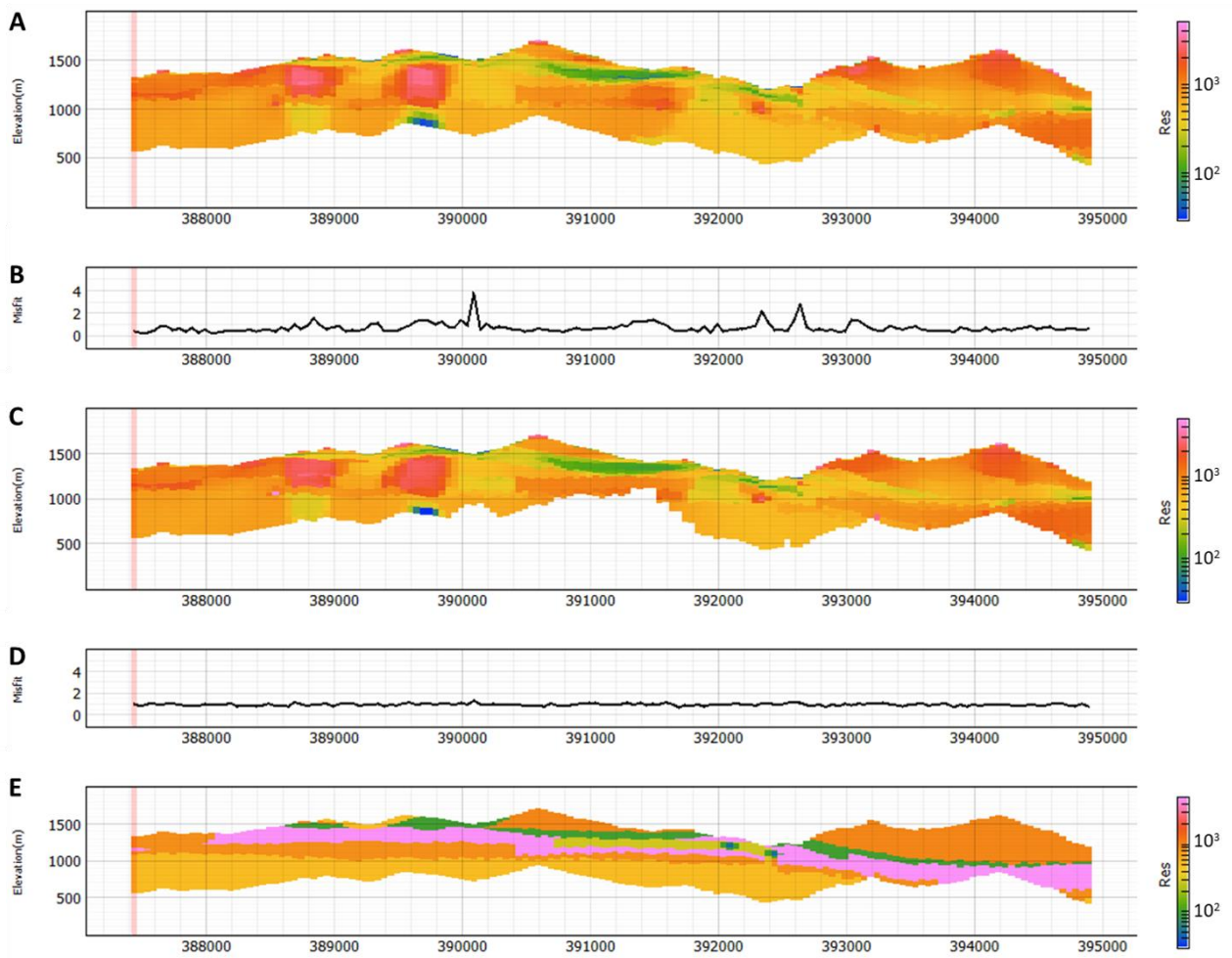


Fig. 4.68 Line 27. A: 1D Sharp geologically-constrained inversion of 3D forward model. B: misfit of A. C: 1D Sharp geologically-constrained inversion of 3D forward model excluding high (>2) misfit data. D: Misfit of C. E: true model.

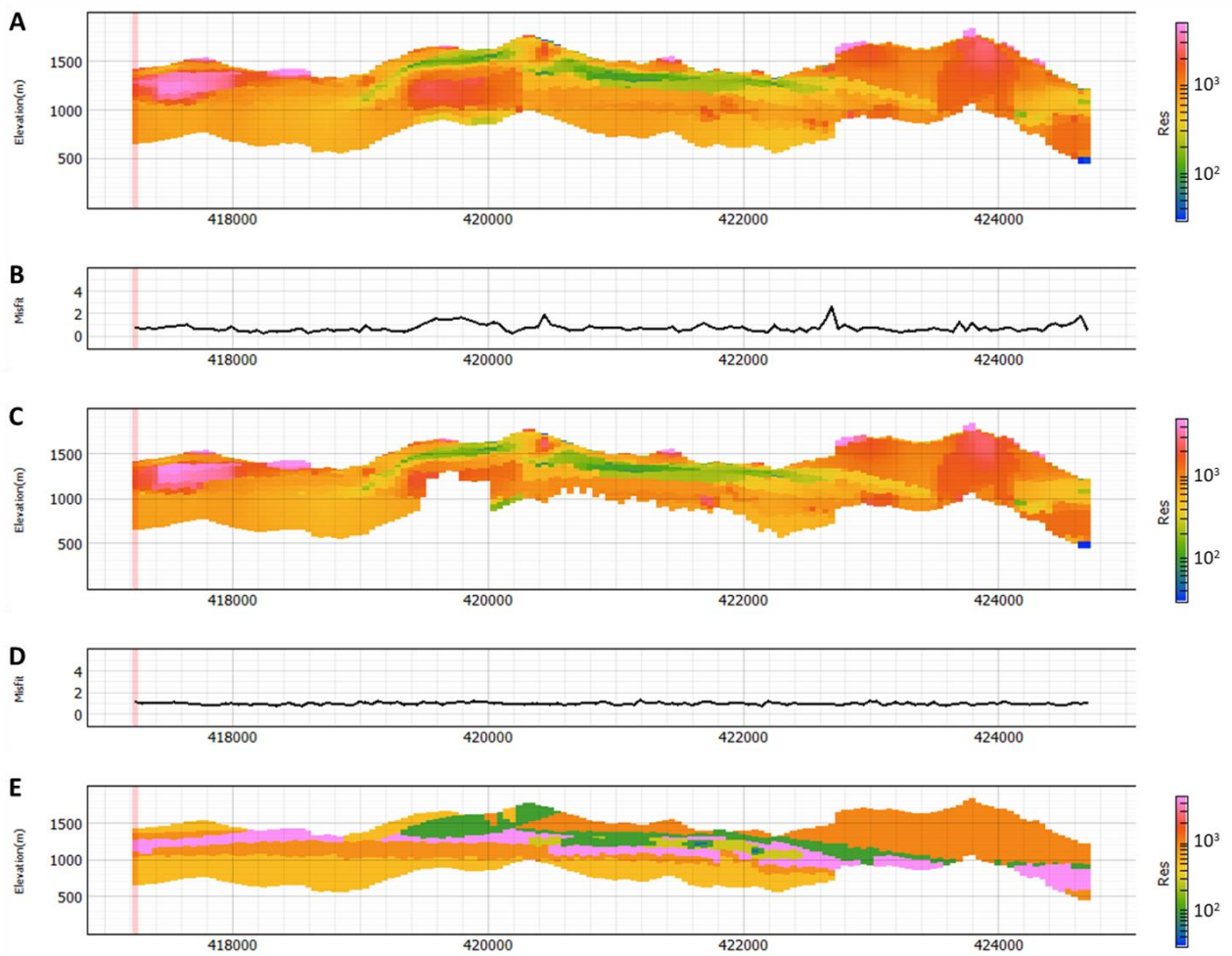


Fig. 4.69 Line 29. A: 1D Sharp geologically-constrained inversion of 3D forward model. B: misfit of A. C: 1D Sharp geologically-constrained inversion of 3D forward model excluding high (>2) misfit data. D: Misfit of C. E: true model.

APPROACH	AEM 1D INVERSIONS	MISFIT
1D Forward	L2 standard	1.03
	L2 geologically-constrained	0.98
	Sharp standard	1.03
	Sharp geologically-constrained	1.01
3D Forward	L2 standard	1.57
	L2 geologically-constrained	1.50
	Sharp standard	1.50
	Sharp geologically-constrained	1.50
3D Forward without high misfit data (>2)	L2 geologically-constrained	0.94
	Sharp geologically-constrained	0.97

Tab. 4.15 Comparison of data misfit from the inversion of three different synthetic AEM resistivity models.

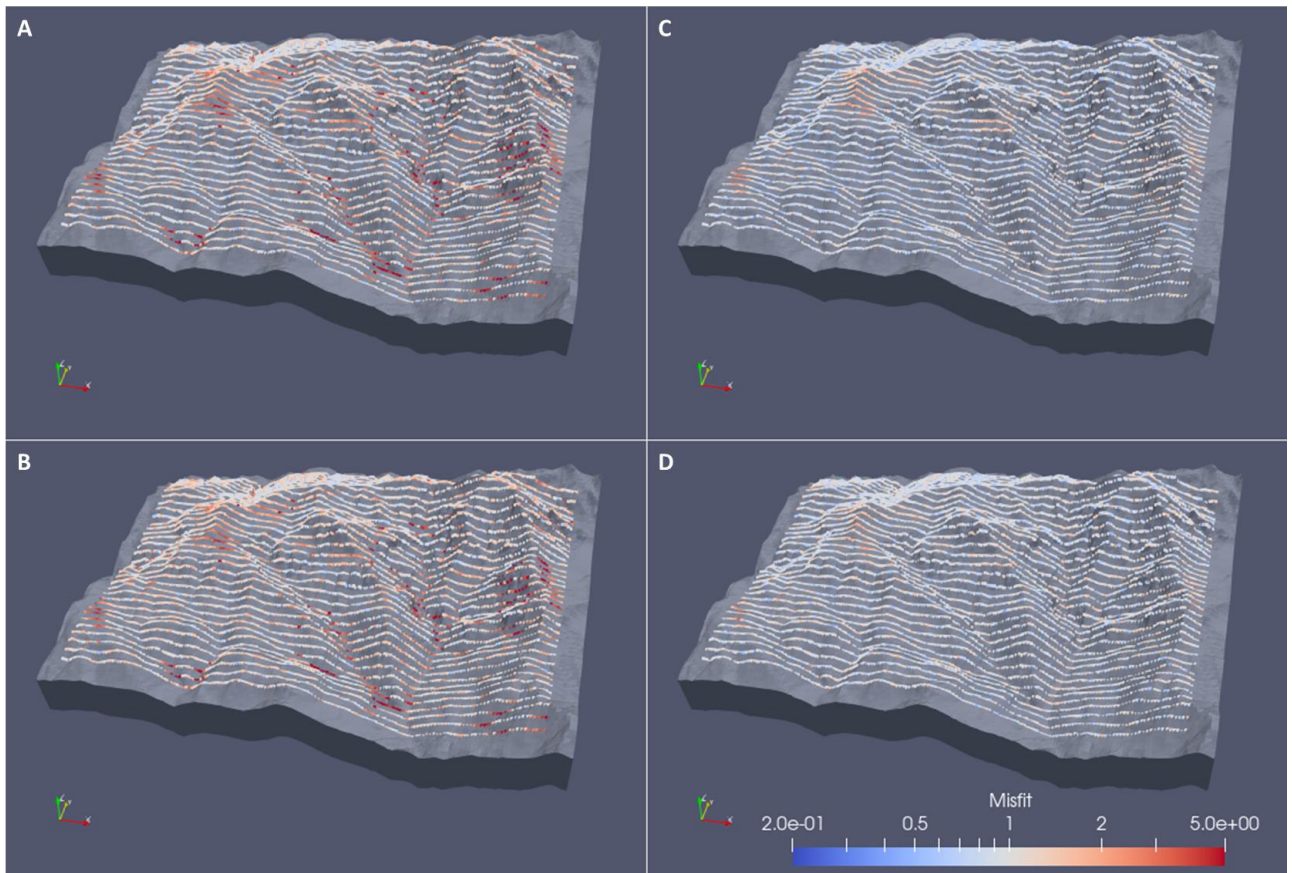


Fig. 4.70 Map of data misfit of 1D inversions of 3D forward response. A: 1D L2 geologically-constrained inversion of 3D forward model. B: 1D Sharp geologically-constrained inversion of 3D forward model. C: 1D L2 geologically-constrained inversion of 3D forward model re-run after removing high misfit data (removal with data misfit > 2). D: 1D Sharp geologically-constrained inversion of 3D forward model re-run after removing high misfit data (removal with data misfit > 2).

4.4 Testing conceptual model of groundwater flow through numerical modelling

4.4.1 Model setup

To test the conceptual model of groundwater flow, the Acqua Bianca catchment area was selected (Fig. 4.71), as it represents the most important spring in term of discharge in the study area. The numerical model was set up using a 3D layered model, consisting of 8 slices and 7 layers. The elevation of the first slice corresponds to ground surface and was computed using the digital terrain model (DTM) provided by Regione Lombardia (<https://www.geoportale.regione.lombardia.it/>) with a resolution of 5 m.

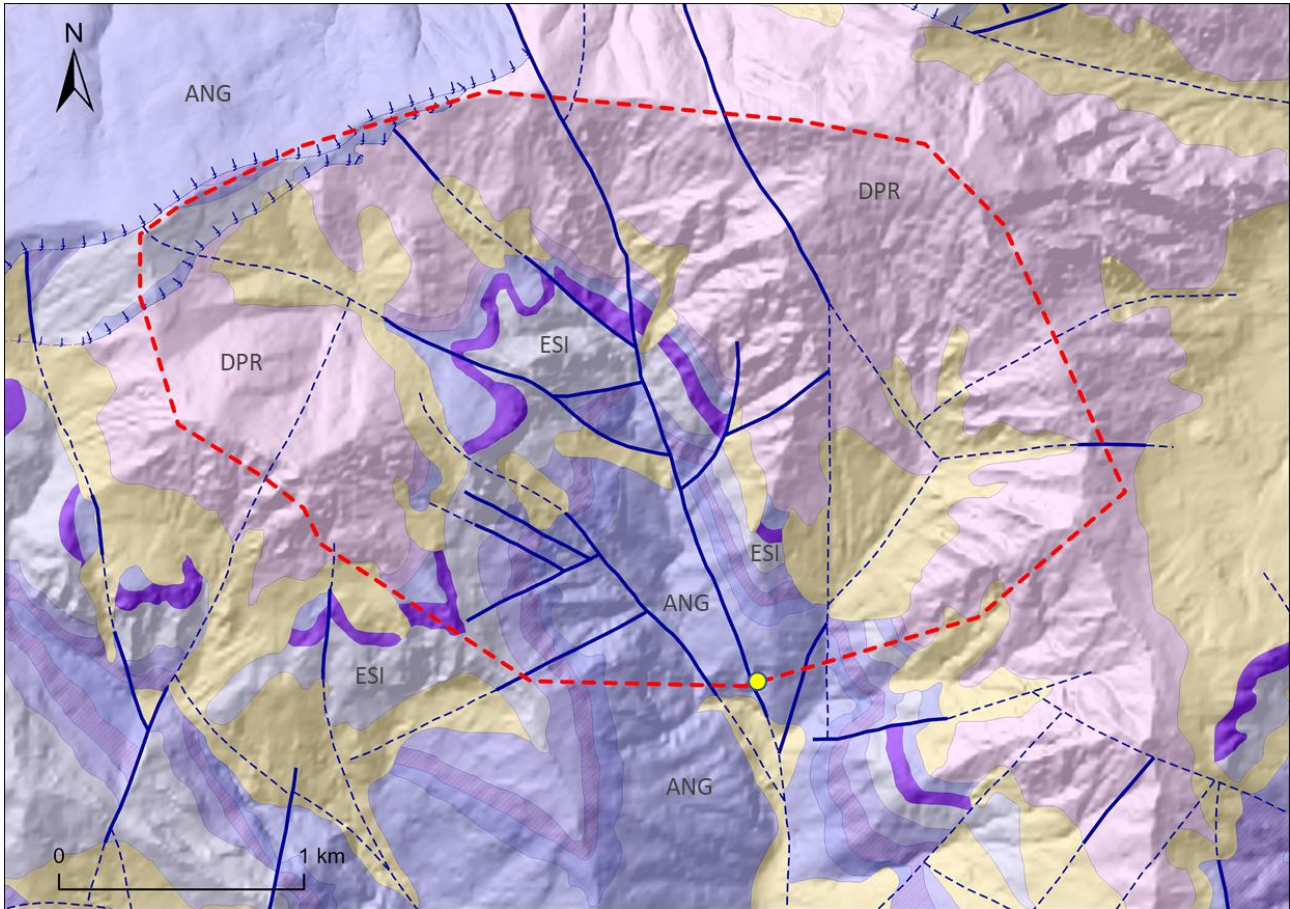


Fig. 4.71 Red polygon represents the model area, the Acqua Bianca spring is marked by the yellow point. The basemap is a detail of the harmonized geological map – DPR: Dolomia Principale, ESI: Esino Limestone, ANG: Angolo Limestone (for more details compare the legend in Fig. 4.1).

The simulation was run in transient state to reproduce the 365 daily observations between 01/10/2022 and 30/09/2023, which were the first time series initially available for the Acqua Bianca spring (discharge measured after 01/10/2023 will be used as calibration or validating datasets in the follow-up). A preliminary run simulated steady state conditions prior to the beginning of the transient simulation. Such conditions represented the “dry” (i.e., base flow)

summer condition in the immediate stages before the “wet” autumn, with consequently increase in spring discharges.

The Acqua Bianca spring (748 m) was assumed to be the main and sole groundwater outlet of the catchment. This assumption is justified by the lack of other relevant springs in the area. First kind (Dirichlet-type) boundary condition was used to reproduce the Acqua Bianca spring. Initially, third-kind (Cauchy-type) boundary conditions were tested, in line with a commonly used approach for simulating springs. However, this configuration led to numerical instability, characterized by significant oscillations. Consequently, a more stable first-kind boundary condition was adopted.

Recharge was mainly ascribed to rainfall, as no significant snow cover was observed in the reference period. Rainfall data were obtained from the regional environmental protection agency (ARPA) weather station located on Mt. Guglielmo (Tavernole sul Mella). These data were selected because records from the closer Pian del Bene weather station became available only after the simulation had already begun. Mt. Guglielmo station is situated approximately 16 km west of the study area as the crow flies and is located in a similar climatic and topographic setting.

Most of the rainfall was assumed to be quickly transferred to the aquifer, without limiting effects by epikarst and soil.

It was assumed that infiltration was concentrated only over the Dolomia Principale outcrops, as it represents the uppermost, fractured and with karst potential outcropping formation, characterized by little covering soils that may limit the infiltration. Moreover, the outcrops of this formation are located at elevations consistent with the recharge altitudes estimated from isotope analysis for the Acqua Bianca spring (see Fig. 4.17).

In addition, this area is affected by the presence of the Pofferatte fault, which could play an important role in draining water from the surface to the spring. Particularly, it was considered recharge area only the portion of Dolomia Principale outcropping west of the fault, to test the hypothesis of a western recharge, as suggested by the recharge elevation estimated with the isotopes (see Section 4.2.2) (Fig. 4.72). A simple water balance that compared the cumulative volume of rainfall occurring on top of the Dolomia Principale outcrop western to fault and the cumulative volume of water measured at the spring supported these assumptions.

The aquifer parameters, largely unknown, were defined based on the conceptual geological model outlined in previous sections. The system was conceptualized as a dual-domain aquifer,

with matrix flow assumed to occur throughout the domain and preferential flow primarily concentrated in the western sector, associated with highly fractured and karstified rocks. Initial high hydraulic conductivities were assigned to all formations underlying the outcropping Dolomia Principale, reflecting the expectation of a well-developed vadose zone. This zone is thought to allow vertical connectivity between the surface and the saturated aquifer through preferential flow paths, particularly activated during recharge events (Fig. 4.73). Although literature reports hydraulic conductivities as high as 10,000 m/d for similar fractured and karst lithologies, an initial value of $K = 1000$ m/d was adopted. This value was subsequently adjusted during model calibration to match the observed spring discharge. As caves and conduits are expected in the western sector of the model domain, discrete features were introduced as illustrated in Fig. 4.72. In line with the approach used for the synthetic AEM models, these elements were placed within the Esino Limestone, which is the formation with the highest karst potential in the area. Each element was assigned a uniform width of 1 m.

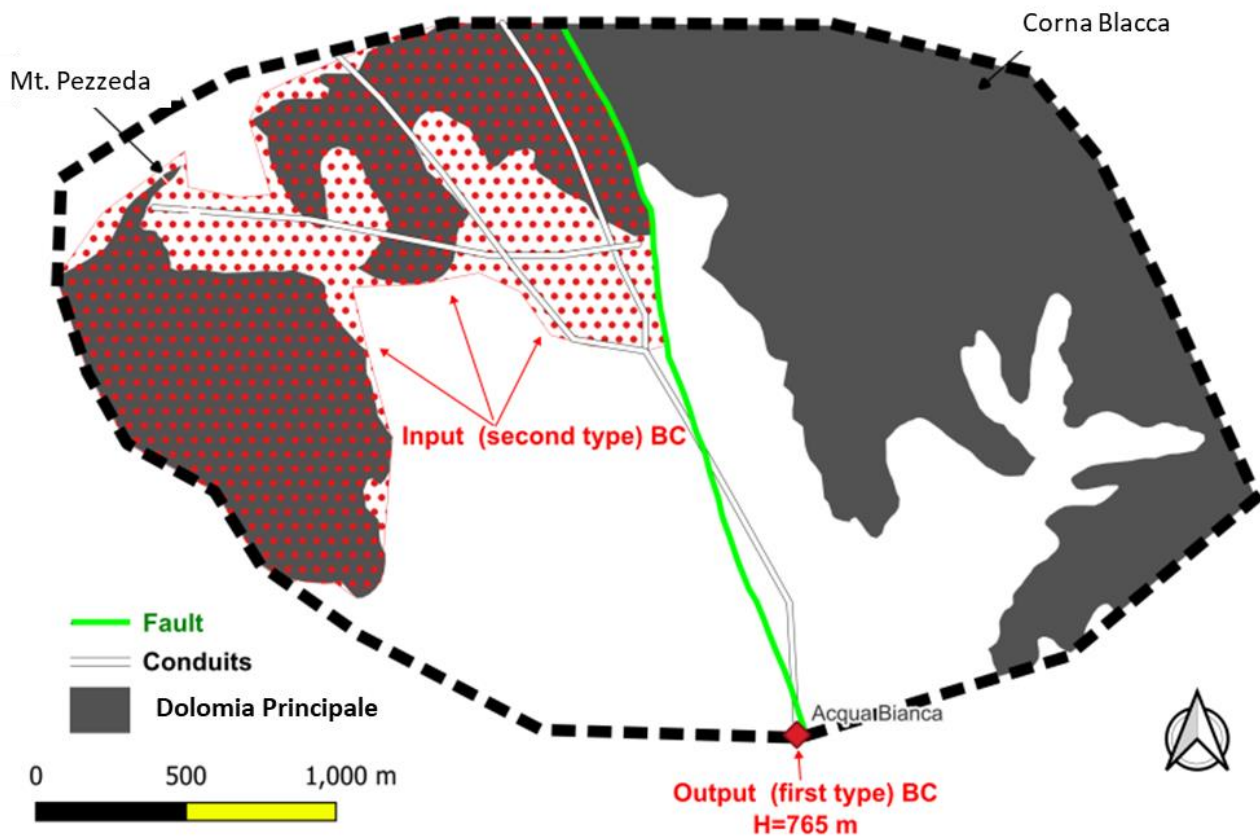


Fig. 4.72 A map with the main elements for numerical simulations, i.e. the discrete elements and the boundary conditions.

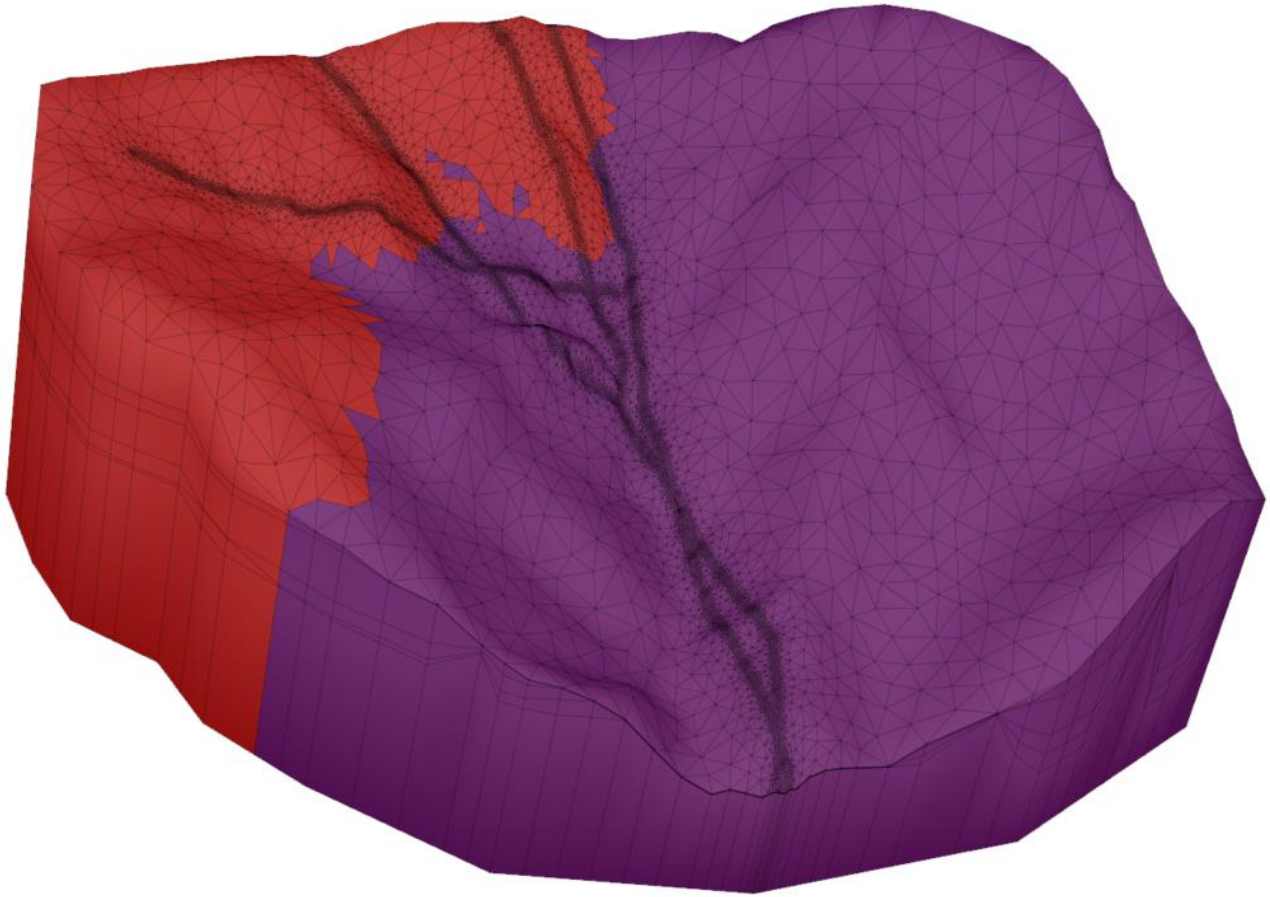


Fig. 4.73 FEFLOW mesh with refinements in proximity of the discrete features representing the fast flow karst conduits. The purple elements represent a less fractured aquifer, with hydraulic conductivity $K=1$ m/d in all simulations. The red elements indicate zones of increased fracturing, potentially associated with diffuse karst. The K value assigned to this zone varies according to the simulation.

In contrast, for the eastern portion of the domain, where the rock mass is assumed to be less fractured, a lower hydraulic conductivity ($K_E = 10^{-5}$ m/d) was applied. This value reflects the reduced permeability due to fewer and less connected fractures, resulting in significantly lower flow rates compared to the western sector.

A sensitivity analysis was conducted on two key parameters: the hydraulic conductivity of the western matrix and the aperture (b) of the conduits. This analysis involved a manual trial-and-error calibration by varying the parameters within a physically plausible range. Although the modelling software FEFLOW supports automated calibration tools such as PEST (Doherty, 1994), a manual approach was deemed suitable for testing the consistency and plausibility of the conceptual hydrogeological model of the study area.

To evaluate the performance of the different simulation scenarios, the root mean square error (RMSE) was used as the fitting metric, calculated as:

$$RMSE = \sqrt{\frac{\sum(Q_{obs} - Q_{model})^2}{N}}$$

where Q_{obs} and Q_{model} represent respectively the observed and modelled discharge and N is the number of observations.

4.4.2 Impact of diffused karst permeability

The first set of simulations aimed to assess the influence of diffuse karst permeability on groundwater flow dynamics. To isolate the effect of this parameter, the hydraulic conductivity of the western sector of the domain (hereafter referred to as K_W), the one considered to be the recharge area of the Acqua Bianca spring, was varied while maintaining a fixed conduit aperture of $b = 1$ m. Specifically, K_W was increased from an initial value of 10 m/d to 1000 m/d.

The initial simulation, using $K_W = 10$ m/d, represents a scenario in which the matrix permeability in the western portion is only ten times higher than that in the eastern part. In Fig. 4.74A, simulated spring discharge (blue line) is compared with observed discharge data (red line with dotted markers), while daily precipitation (in mm/d) is shown as black dashed lines. Fig. 4.74B displays the cumulative discharge curves. Due to missing observations (e.g., around day 250), the cumulative values were calculated only for the time steps with available measured discharge data.

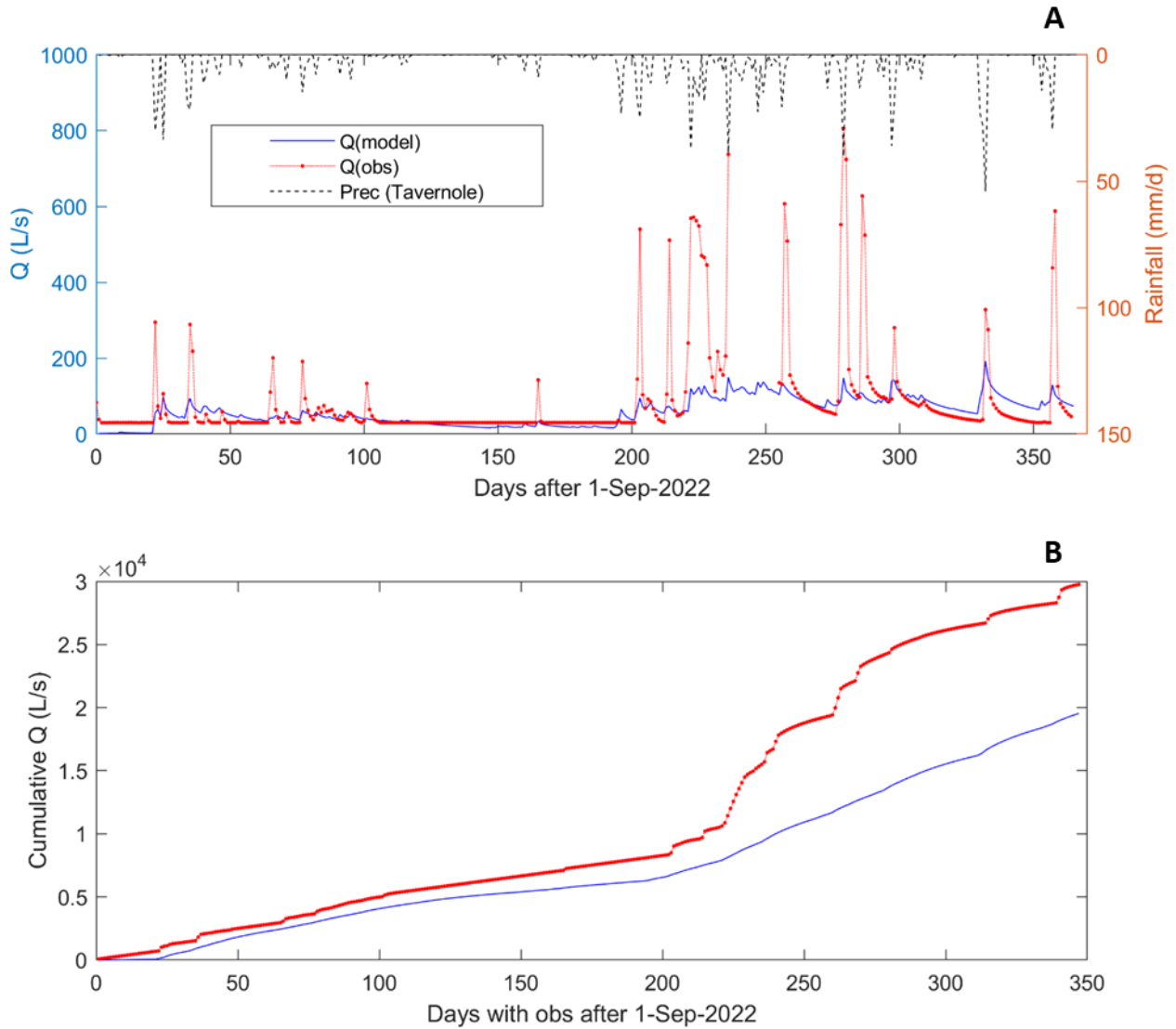


Fig. 4.74 Simulation results with $K_w = 10 \text{ m/d}$ and a conduit aperture of $b = 1 \text{ m}$ for the discrete features representing karst conduits. A: simulated spring discharge (blue line) vs. observed discharge data (red line). B: cumulative discharge curves.

From Fig. 4.74A it can be observed that the model adequately reproduces the baseflow and most of the recession limbs of the hydrograph, particularly after simulation day 100. Given that, the initial conditions are not well constrained, the early part of the simulation is considered a model warm-up period and is therefore less suitable for interpreting flow processes. The model, however, fails to replicate the observed discharge peaks at the spring, resulting in a notable divergence between observed and simulated cumulative discharge after simulation day 200 (Fig. 4.74). This discrepancy is reflected in a relatively high root mean square error ($RMSE = 65.95$) and leads to a mismatch in the overall water budget. At the end of the simulation, the cumulative observed discharge ($\sum Q_{\text{obs}} = 29.763 \text{ l/s}$) is significantly higher than the simulated

discharge ($\sum Q_{\text{sim}} = 19.538 \text{ l/s}$), with the model underestimating the total outflow by approximately 33%.

As the hydraulic conductivity of the matrix is progressively increased to $K_W = 50 \text{ m/d}$ (Fig. 4.75) and $K_W = 100 \text{ m/d}$ (Fig. 4.76), the model begins to generate spring discharge peaks that closely resemble the observed data. The corresponding cumulative simulated discharge ($\sum Q_{\text{sim}} = 24.113 \text{ l/s}$ for $K_W = 100 \text{ m/d}$) suggests that increasing the diffuse matrix permeability in the western portion of the domain improves the agreement with observed discharge volumes. Nevertheless, discrepancies remain. The model is still unable to reproduce discharge peaks exceeding 300 l/s, and although the observational dataset contains gaps (e.g., around day 190), the overall water budget remains underestimated.

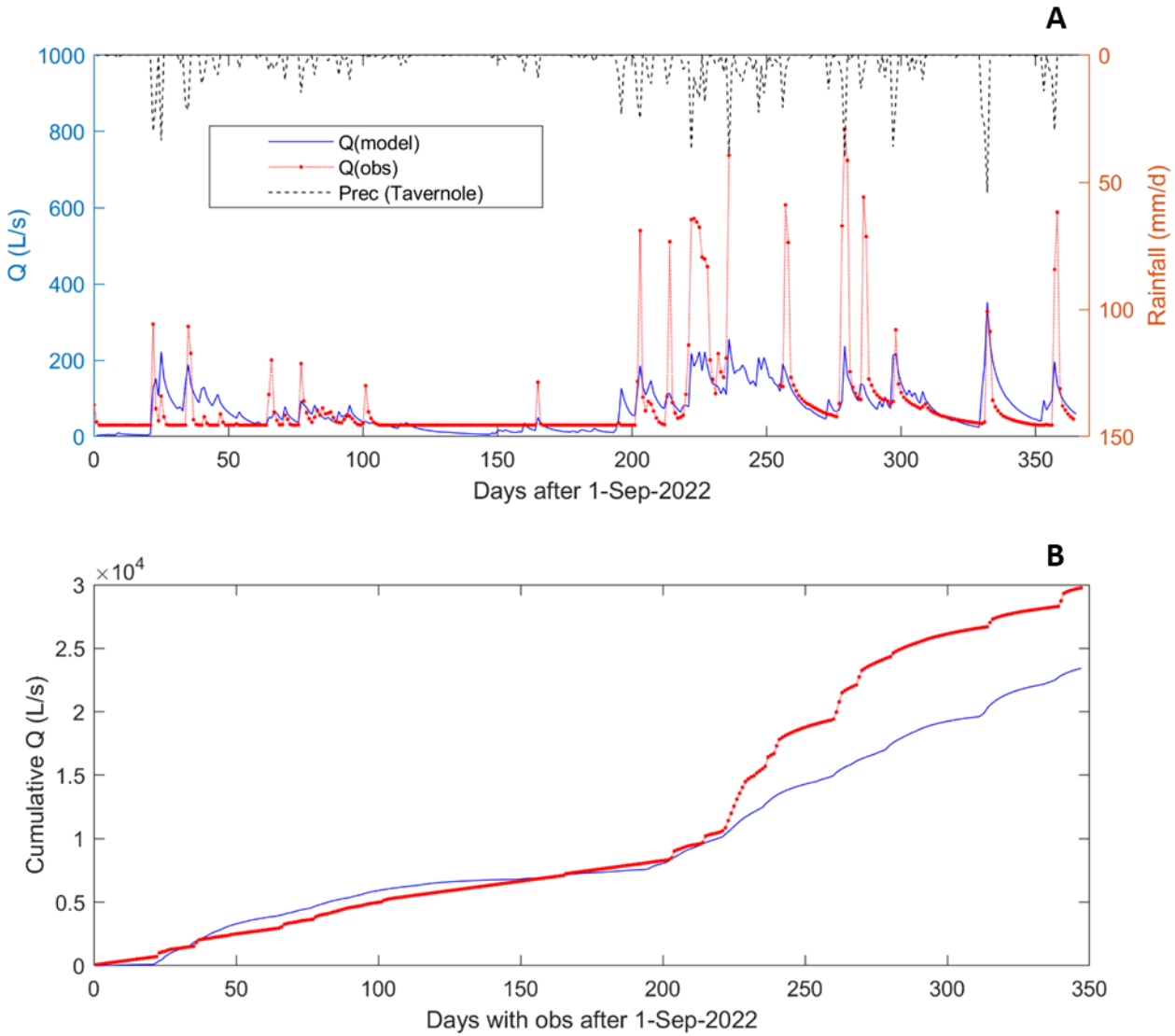


Fig. 4.75 Simulation results with $K_W = 50 \text{ m/d}$ and a conduit aperture of $b = 1 \text{ m}$ and cross section $A = 1 \text{ m}^2$ for the discrete features representing karst conduits. A: simulated spring discharge (blue line) vs. observed discharge data (red line). B: cumulative discharge curves.

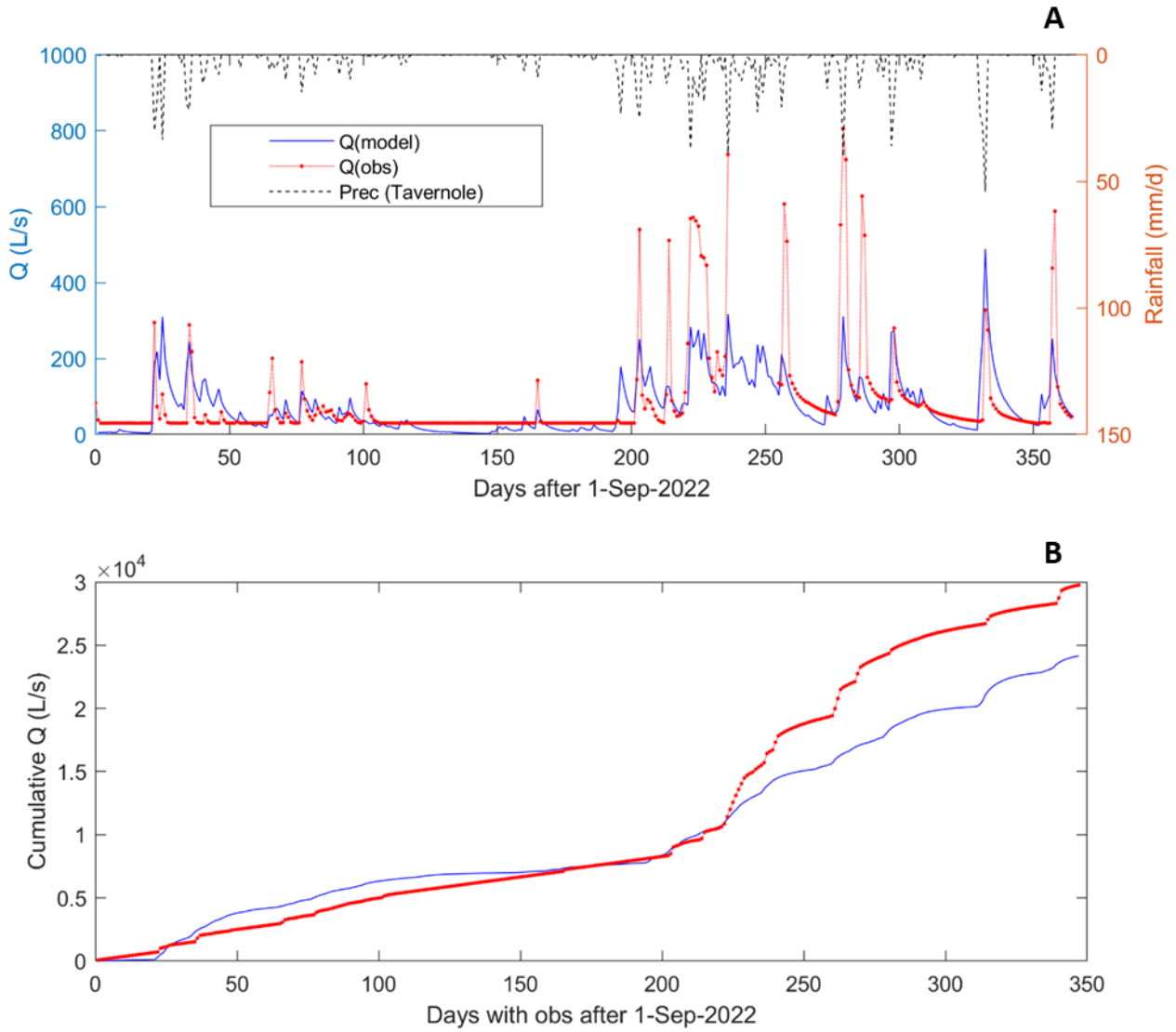


Fig. 4.76 Simulation results with $K_W = 100$ m/d and a conduit aperture of $b = 1$ m and cross section $A = 1$ m² for the discrete features representing karst conduits. A: simulated spring discharge (blue line) vs. observed discharge data (red line). B: cumulative discharge curves.

When the hydraulic conductivity of the western matrix is set to $K_W = 500$ m/d (Fig. 4.77) and $K_W = 1000$ m/d (Fig. 4.78), the calculated water budget ($\sum Q_{\text{sim}} = 25.125$ l/s for $K_W = 1000$ m/d) improves, bringing it closer to the observed values. However, the model fails to accurately reproduce the baseflow, recession curves, and discharge peaks, leading to a higher $RMSE = 150.78$ for $K_W = 1000$ m/d compared to the lower K_W scenarios.

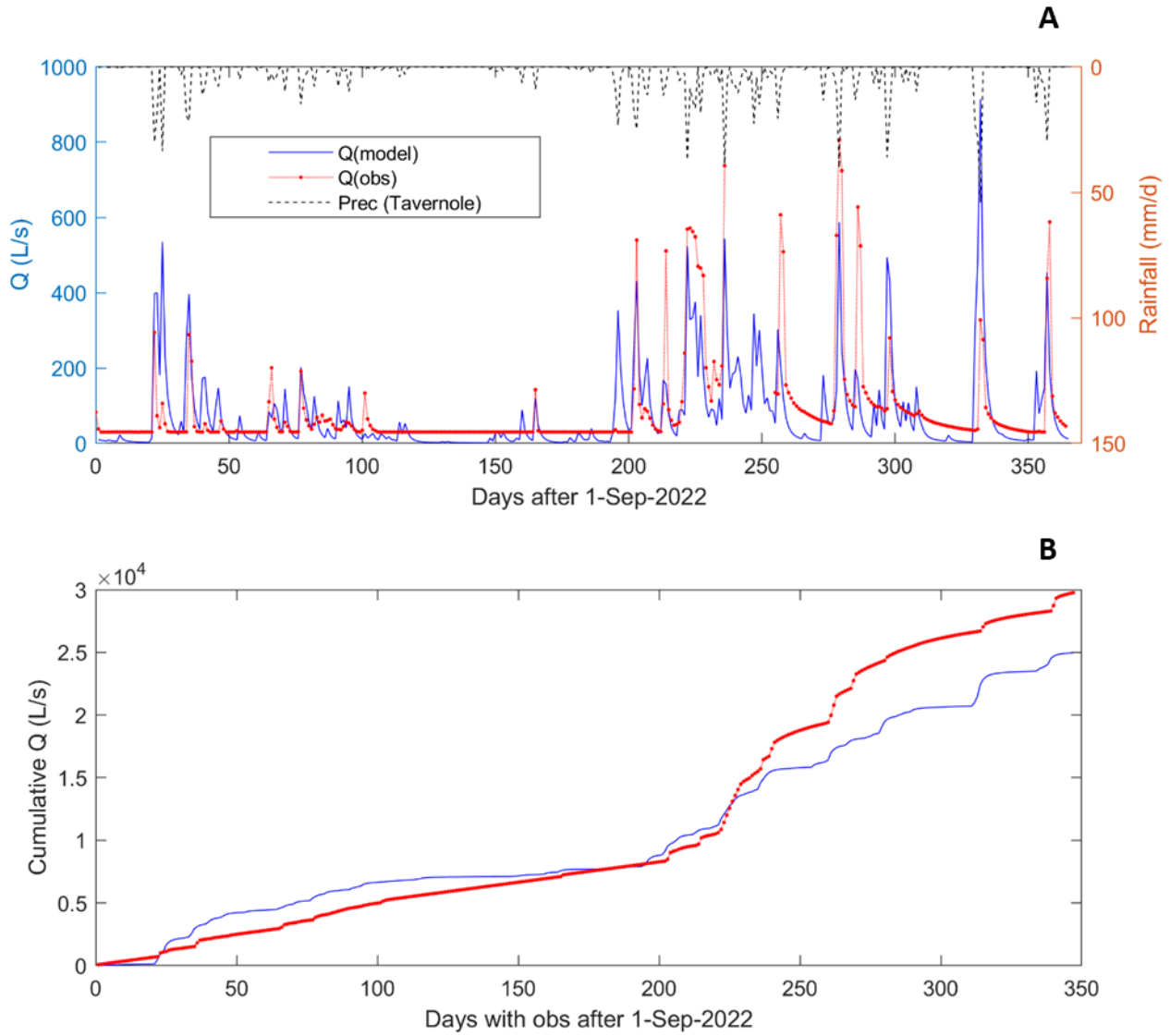


Fig. 4.77 Simulation results with $K_w = 500 \text{ m/d}$ and a conduit aperture of $b = 1 \text{ m}$ and cross section $A = 1 \text{ m}^2$ for the discrete features representing karst conduits. A: simulated spring discharge (blue line) vs. observed discharge data (red line). B: cumulative discharge curves.

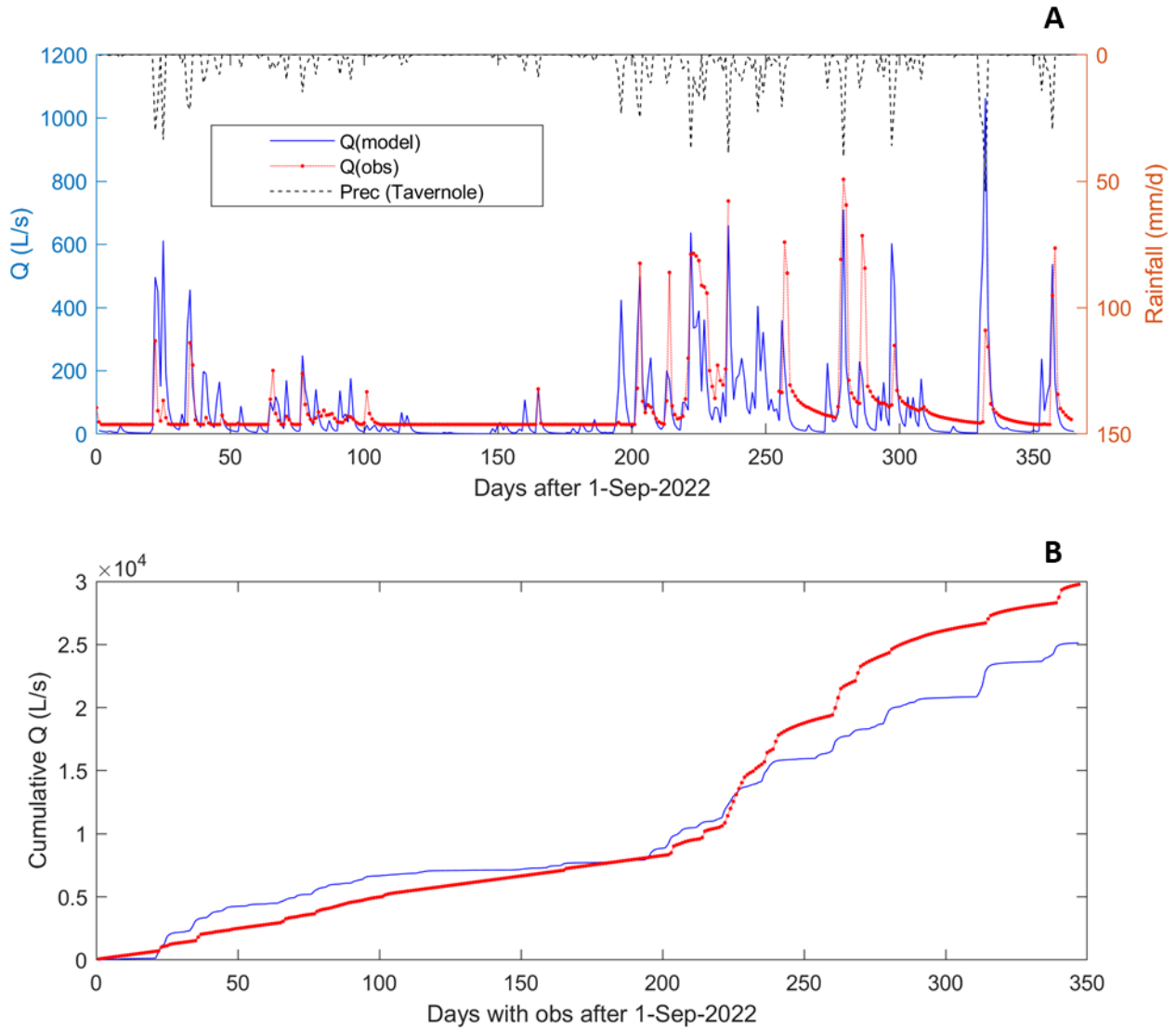


Fig. 4.78 Simulation results with $K_w = 1000$ m/d and a conduit aperture of $b = 1$ m and cross section $A = 1$ m² for the discrete features representing karst conduits. A: simulated spring discharge (blue line) vs. observed discharge data (red line). B: cumulative discharge curves.

4.4.3 Impact of discrete feature aperture

The impact of fast flow discrete features (i.e., conduits) on spring outflow was then evaluated. Since the aperture (b) controls the flow rate along the conduits, based on the Hagen-Poiseuille law, some of the previously presented simulations (all with $b = 1$ m) were re-run using different aperture values. Specifically, it was hypothesized that a lower aperture ($b < 1$ m) could improve the model fitting for simulations with $K_w = 500$ and 1000 m/d, which had previously produced discharge peaks larger than those observed with $b = 1$ m. Conversely, it was expected that a higher aperture ($b > 1$ m) would yield better results for simulations with $K_w = 100$ m/d or lower. One representative simulation was selected for each case and analyzed.

In the first case, using a very low aperture ($b = 0.001$ m) combined with a very high hydraulic conductivity for the matrix ($K_W = 1000$ m/d) did not result in an increase in spring discharge (Fig. 4.79). Neither the baseflow nor the peaks were accurately captured, suggesting that diffuse karst alone may not account for the flow rates without a substantial contribution from the conduits.

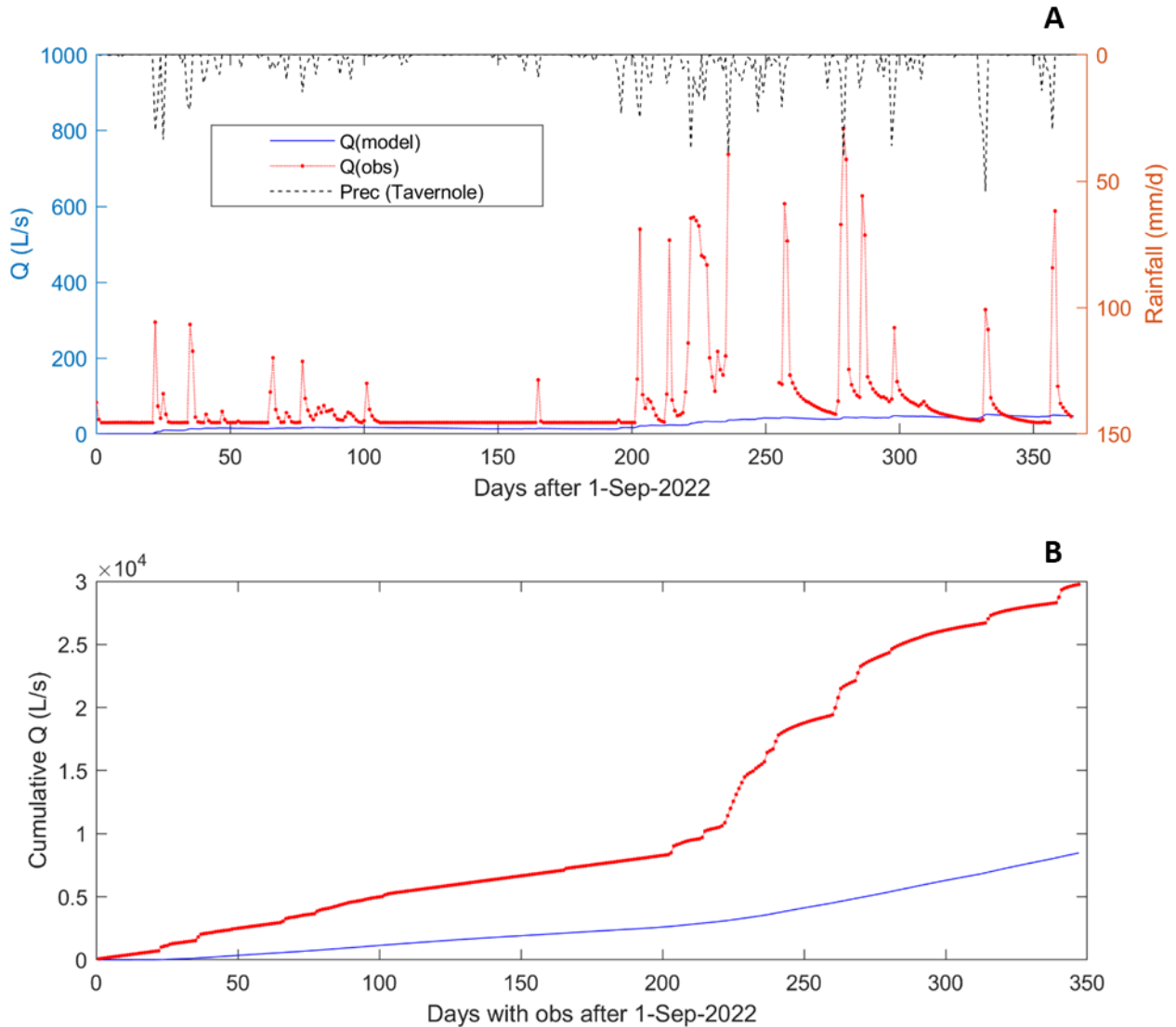


Fig. 4.79 Simulation results with $K_W = 1000$ m/d and a conduit aperture of $b = 0.001$ m and cross section A = 1 m^2 for the discrete features representing karst conduits. A: simulated spring discharge (blue line) vs. observed discharge data (red line). B: cumulative discharge curves.

In the second case, a very high aperture ($b = 20$ m) combined with a lower matrix flow ($K_W = 50$ m/d) results in an increased basal flow and the generation of spring discharge peaks (Fig. 4.80). This confirms that diffuse karst combined with fast flow along the conduits may provide a reasonable conceptualization of the hydrodynamics occurring at Acqua Bianca spring.

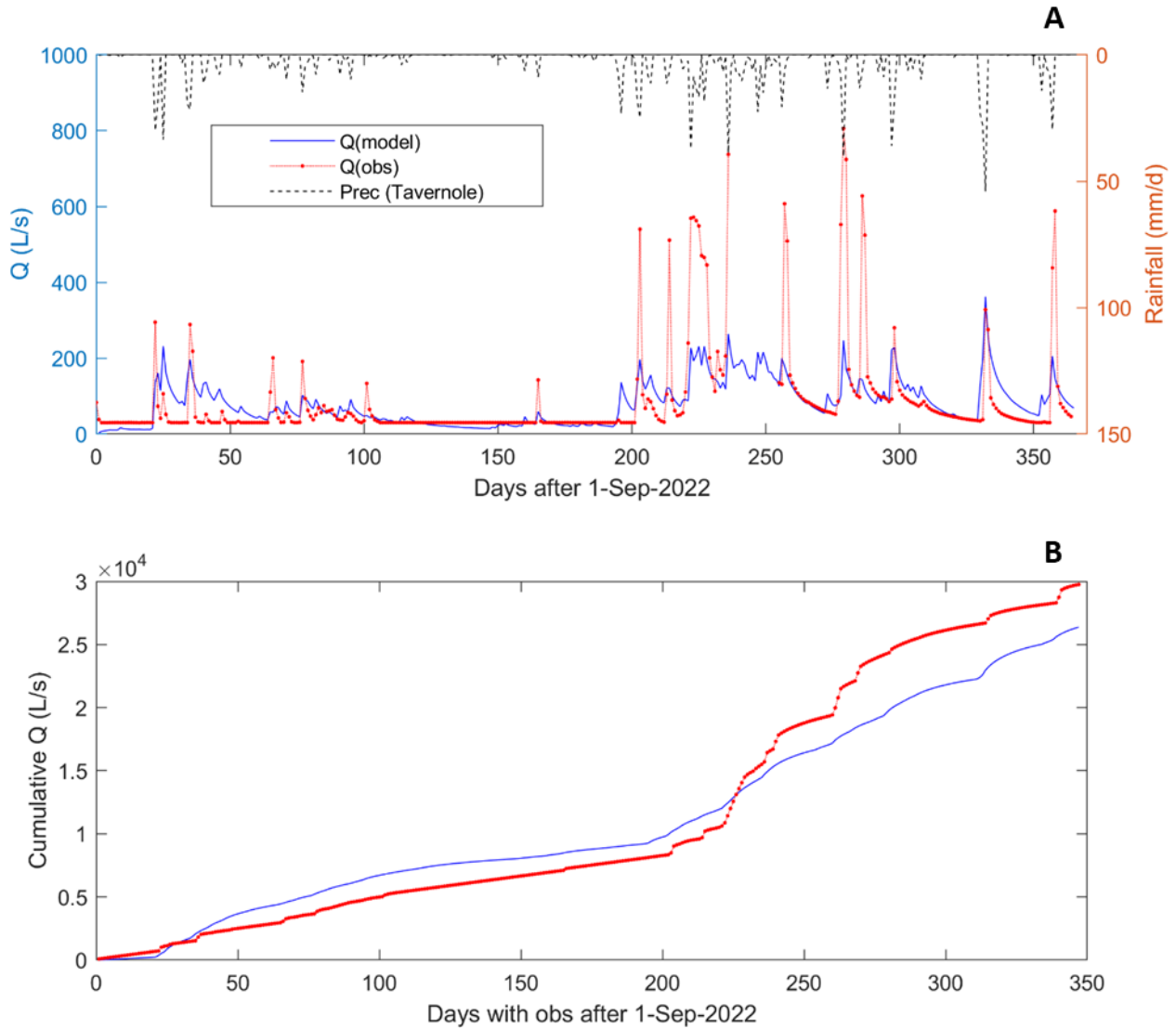


Fig. 4.80 Simulation results with $KW = 50 \text{ m/d}$ and a conduit aperture of $b = 20 \text{ m}$ and cross section $A = 1 \text{ m}^2$ for the discrete features representing karst conduits. A: simulated spring discharge (blue line) vs. observed discharge data (red line). B: cumulative discharge curves.

These simulations support the hypothesis that diffuse karst may coexist with conduits in the western part of the study area, and that recharge occurring on this area is sufficient to produce the entire water volume outflowing at the Acqua Bianca spring. This initial study provides the basis for future follow-up investigations.

The hydrogeological model was prepared to facilitate the integration of future geophysical-derived information. Although not available at present for this site, it is expected that such an additional dataset will improve the physical interpretation of the hydrodynamic mechanisms in the aquifer. The integration of geophysical data into hydrogeological models in karst environments faces key challenges, particularly in linking petrophysical properties to karst

features like conduit topology and hydraulic conductivity. Resistivity maps can serve as proxies for identifying water-bearing zones.

In this work, we developed a layer-based model. While several studies have adopted a similar methodology, the geometry of the analyzed karst systems may be better reproduced using an unstructured grid approach (Panday et al., 2013), which should reduce the computational cost with respect to use of a regular uniform grid (Panday et al., 2013). Unstructured vertical discretization represented directly by hydrogeological layers can replicate complex geological features, including karst conduits (Duran & Gill, 2021). A potential benefit of this approach is that a direct coupling could be established between 3D geological and hydrogeological models. Indeed, GeoModeller is already linked with FEFLOW, and unstructured meshes created using the former (geological) code can be exported to the latter (hydrogeological) code. Unfortunately, our attempts to couple the two codes have always failed, probably due to the strong complexity of the system analyzed. Therefore, the use of unstructured meshes is left open for future development.

5. Results and discussion on the Aracena study site

5.1 Definition of the 3D geological architecture

As for the Val Sabbia model, the extent of the 3D Aracena project is established before modelling begins. The bounding box of the investigated domain was designed to include the area of the AEM survey campaign. Once the domain was selected, the DTM was imported to provide the topographic limit of the model. It consists of a raster dataset with a horizontal resolution of 20 m. It was not possible to find more resolute open access DTM. As the base of the model, a planar surface set at 0 m asl was chosen, which was considered to be sufficiently deep with respect to the morphological characteristics of the study area.

As a result, the modelled area is a rectangular prism spanning 8 km by 3.4 km by 0.9 km west of the town of Aracena (Tab. 5.1).

In this case, the only available geological map of the area is the “Aracena” sheet of the geological map of Spain (Mapa Geológico de España) at 1:50,000 scale (Instituto Geológico y Minero de España, 1995) (Fig. 5.1). It is imported in GeoModeller environment, where it serves as a constraint for the 3D geological modelling process.

The next step in the workflow involved selecting the reference geological surface to model. Consistently with the approach used in the Val Sabbia model, the bottom boundary of each geological unit was chosen as the reference surface for modelling.

Unlike the Val Sabbia model, where each geological formation was modeled as a separate independent unit, the Aracena model was designed with a different approach. In this case, two of the three outcropping formations were grouped together in a single series, resulting in the definition of two series: one series corresponding to the schists of the La Umbría Formation, and a second series combining the dolomites of the Aracena Formation and the metavolcanic rocks of the Volcanosedimentaria Formation.

This approach was chosen to achieve better control over fault-induced deformation within the lithostratigraphic succession, while avoiding the introduction of deep geological constraints that are not supported by the available data.

	Min	Max
X	706313	714355
Y	4194683	4198066
Z	0	900

Tab. 5.1 Aracena 3D model extent.

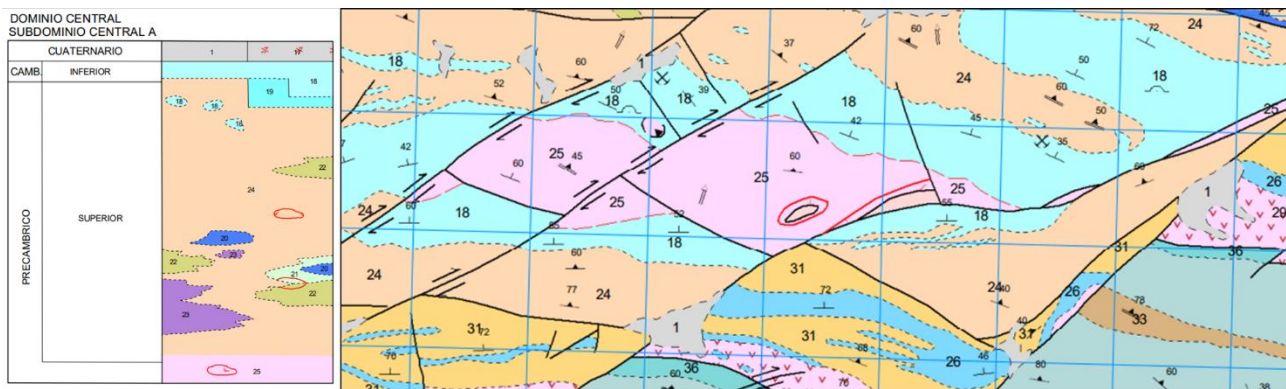


Fig. 5.1 Aracena study site structural sketch on the left and geological map on the right. The code 25 identifies the La Umbría Formation, 18 the carbonates of the Aracena Formation, and 24 the metavolcanics of the Volcanosedimentaria Formation. Black lines are the faults. (Instituto Geológico y Minero de España, 1995)

As previously mentioned, the geological pile of the Aracena 3D model is composed of two series: the La Umbría schists series, and a combined series including the Aracena Formation (dolomites) and the Volcanosedimentaria Formation (metavolcanic rocks).

The chronological order of these series within the geological pile dictates their temporal arrangement in the numerical modelling process, ensuring that the stratigraphic relationships are properly represented.

The modelling workflow incorporates geological boundary contact data and structural orientation data of each formation at the surface as constraints.

Additionally, the study area is characterized by a complex fault system that dislocates the outcropping succession (see Section 2.3.2). To accurately represent fault-related displacements in the 3D geological model, five major faults (F1-F5) were introduced into GeoModeller environment.

The surface constraints used in the model, including mapped geological boundaries and fault traces, were directly digitized and integrated into GeoModeller. Given the availability of sufficient surface orientation data, no deep constraints were added to the model to avoid introducing additional complexity in calculating the potential field for each geological series.

In conclusion, this modelling phase has produced a reliable and detailed 3D geological model that accurately captures the complexity of the area (Fig. 5.2). The model successfully represents all the structural features, including the overturned anticline with the schists of La Umbría Formation at its core (Fig. 5.3), and the displacement of the two nearly perpendicular fault sets.

This comprehensive representation of the study area serves as a strong foundation for guiding and enhancing geophysical modelling efforts.

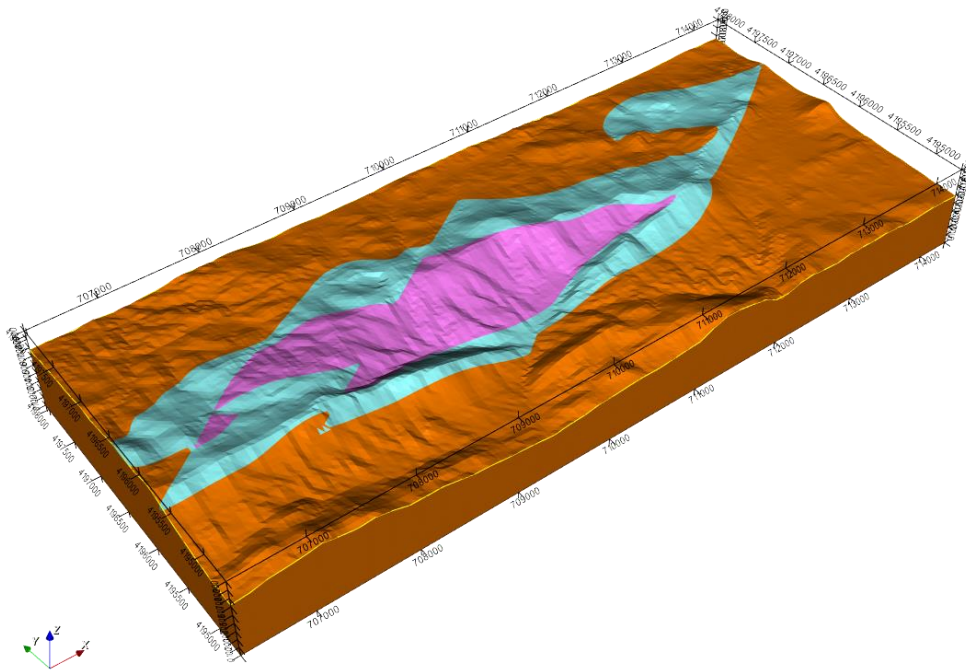


Fig. 5.2 3D geological model of the Aracena study area. The Y axis is the North.

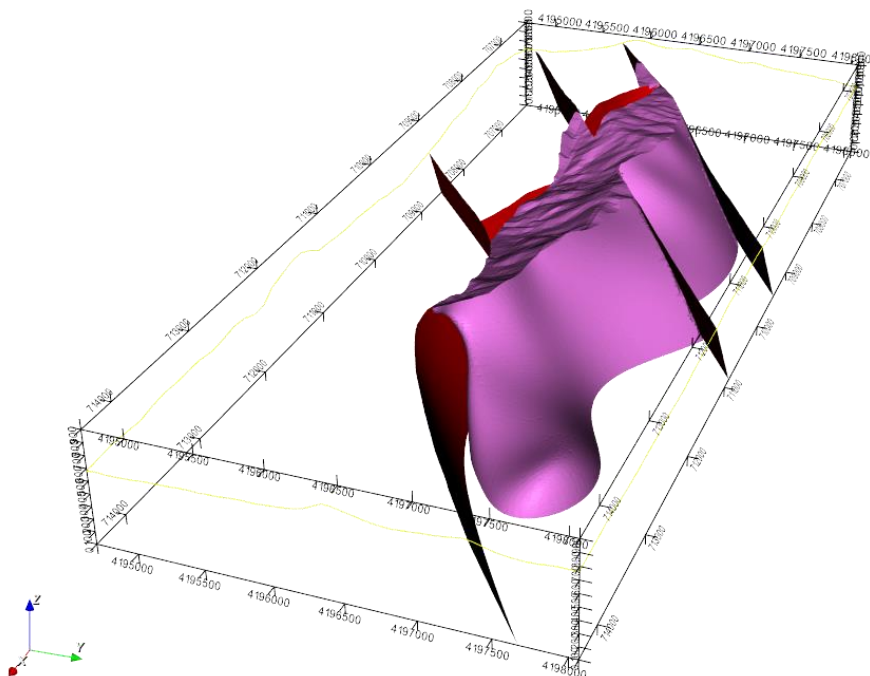


Fig. 5.3 The volume of the schists of La Umbría Formation displaced by the faults network. The Y axis is the North.

5.2 Geophysical modelling results

5.2.1 AEM 1D and 3D resistivity models

In the Aracena study site 35 line-km of AEM data were acquired in 2022 by NRG Europa (Fig. 5.4). The AEM data were first processed in EEMstudio (Fig. 5.5) and then inverted in EEMverter using three different approaches:

- a standard 1D inversion with L2 roughness constraints;
- a geologically-informed 1D inversion with L2 roughness constraints;
- a geologically-informed 3D inversion with L2 roughness constraints.

Considering that the data exhibit significant IP signature with change of sign at late times (Fig. 5.5), all inversions were carried out using the Maximum Phase Angle (MPA) re-parameterization of the Cole-Cole model presented by Fiandaca et al. (2018b).

Tab. 5.2 provides a comparison of the overall misfit values associated with the three inversion strategies mentioned above.

Fig. 5.6 shows the misfit along typical survey line exhibiting IP effects, while Fig. 5.7 presents the comparison between the parameter ρ_0 of the three inversions as 3D volumes, highlighting the correspondence between the resistivity distributions and the surface geology shown in Fig. 5.2. Instead, Fig. 5.8 shows the locations of 5 sections along which results are presented in detail, as explained below.

- The three W-E oriented sections 'a', 'b' and 'c' are shown in Fig. 5.9, Fig. 5.10 and Fig. 5.11, respectively, in terms of parameter ρ_0 . Section 'a' is relatively similar with all inversion approaches, but the 3D inversion in sections 'b' and 'c' is significantly different. In particular, section 'b' shows a much more conductive anomaly between UTMx coordinates 709500 m and 711000 m, and section 'c' between UTMx coordinates 710500 m and 711000 m. Furthermore, in all sections the 3D geologically-informed inversion exhibits a net resistivity contrast at the interface with the schists, which are more conductive in the north section 'a'. On the contrary, the 1D inversions locate the more conductive anomalies in the dolomites in the East portions of the profiles. In all the three sections the continuity of the resistivity features is abruptly broken in correspondence to faults.
- Some portions of the schists and of the shists-dolomites contact are also chargeable, as evident in the chargeability φ_{max} sections 'a', 'b' and 'c' shown in Fig. 5.12, Fig. 5.13 and Fig. 5.14.

- Fig. 5.15 presents the resistivity parameter ρ_0 for the N-S sections 'd' and 'e', while Fig. 5.16 shows the chargeability parameter φ_{max} for the same sections. These N-S profiles allows to understand better the geometry of the resistivity and chargeability contrasts: it is evident that the northern limb of the Fuenteheridos – La Umbría anticline is conductive and chargeable, both within the schists of the La Umbría Formation and at the interface between these schists and the overlying dolomites of the Aracena Formation. In contrast, the schists on the southern limb of the fold exhibit higher resistivity values, are not chargeable. In both sections the 3D geologically-informed inversion presents much clearer resistivity contrasts across formations, with a very evident conductive and chargeable northern contact between schists and dolomites.
- Finally, Fig. 5.17 and Fig. 5.18 highlight the geometry of the conductive and chargeable northern limb of the Fuenteheridos – La Umbría anticline in a 3D view, where again the 3D geologically-informed inversion presents the better correspondence between geology and geophysical anomalies.

These findings suggest the development of a weathered zone at the contact between the dolomites and schists, which might be largely influenced by groundwater circulation. In particular, groundwater might flow through the permeable dolomites on the northern limb of the anticline and, where structural discontinuities are present, it might infiltrate and interact with the underlying schists. This fluid-rock interaction might facilitate weathering processes at the interface, leading to mineral alteration and enhanced permeability. Consequently, this altered zone exhibits higher electrical conductivity compared to the relatively unaltered schists on the southern limb. The extent and intensity of the weathering process are primarily controlled by groundwater flow dynamics, but also by the heterogeneous mineralogical composition of the schists.

These results highlight the coupling between hydrogeological processes and the geophysical signatures observed in the study area.

AEM INVERSIONS	APPROACH	MISFIT
1D Inversion	L2 standard	1.22
	L2 geologically-constrained	2.38
3D Inversion	L2 geologically-constrained	1.90

Tab. 5.2 Comparison of data misfit across the three inversion strategies applied to the Aracena AEM resistivity dataset.

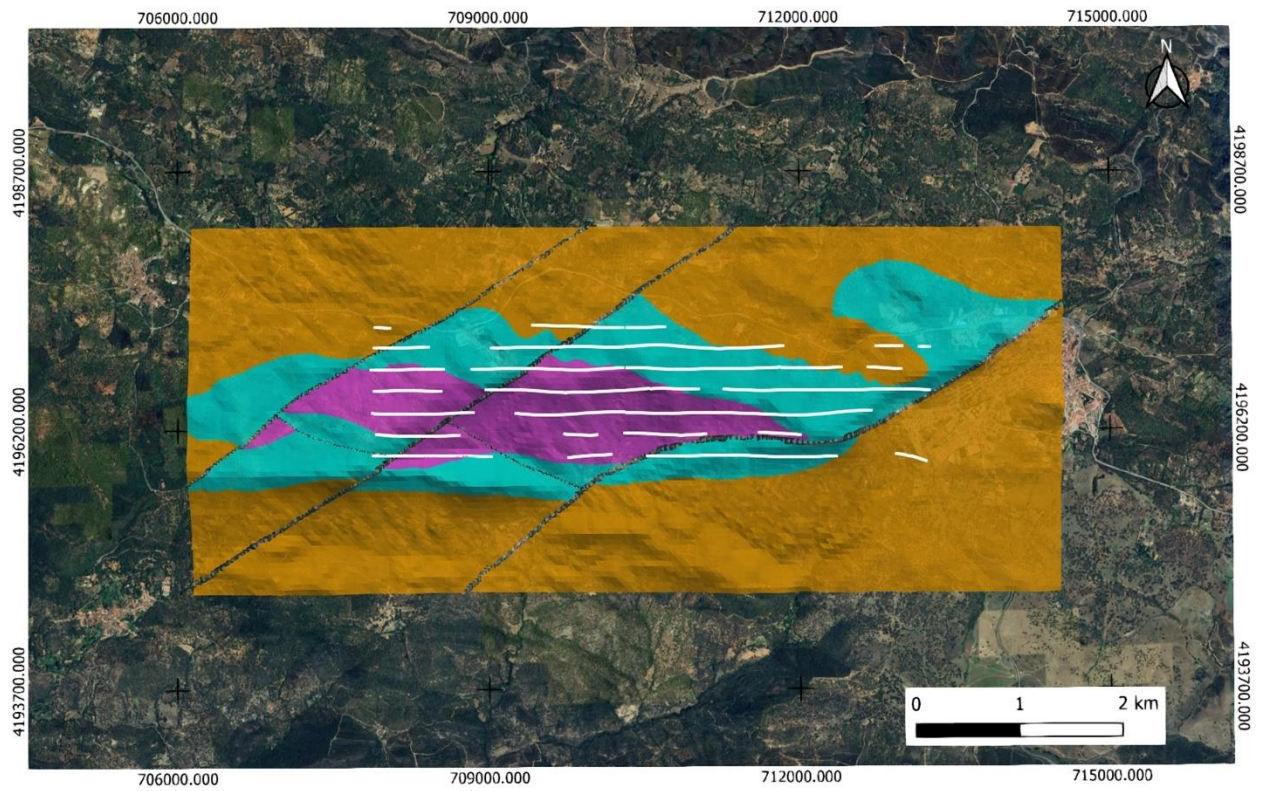


Fig. 5.4 AEM survey lines with respect to the geological map obtained from the 3D modelling in the Aracena study area.

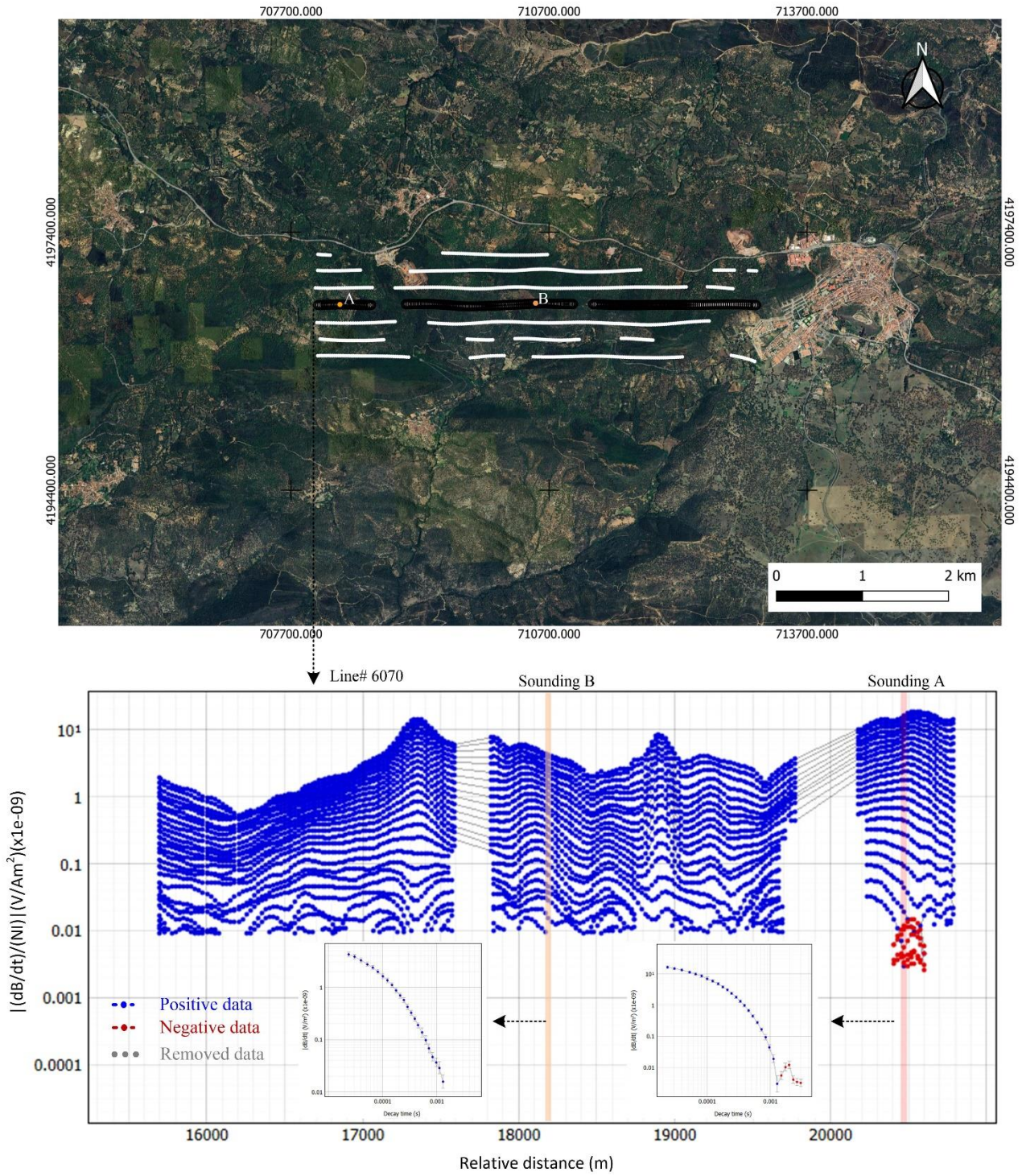


Fig. 5.5 AEM data visualization along a survey line (indicated by black markers in the map on the top). Blue represents the data used in the inversion, red represents IP data, and grey represents the data removed during processing.

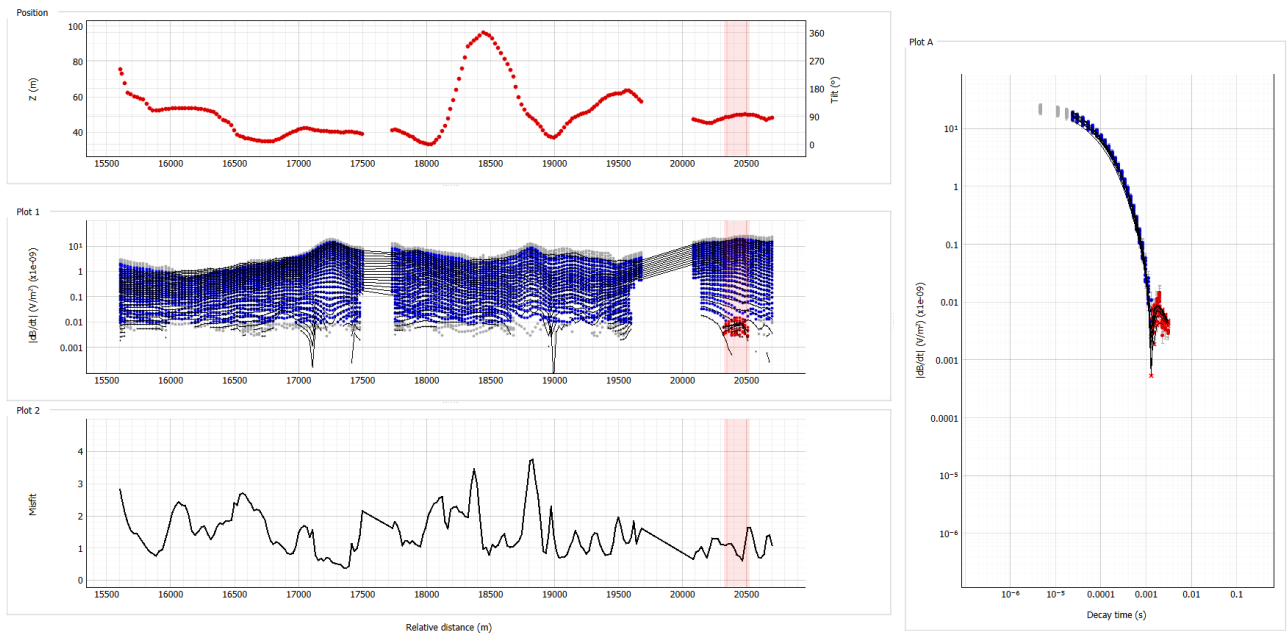


Fig. 5.6 Data fit along the same survey line of Fig. 5.5. Top-left panel – flight altitude. Mid-left panel – data used in the inversion (blue markers), data culled out (grey markers) and forward response (black lines). Bottom-left panel – data misfit. Right panel – data and fit for the soundings highlighted in red in the left panels.

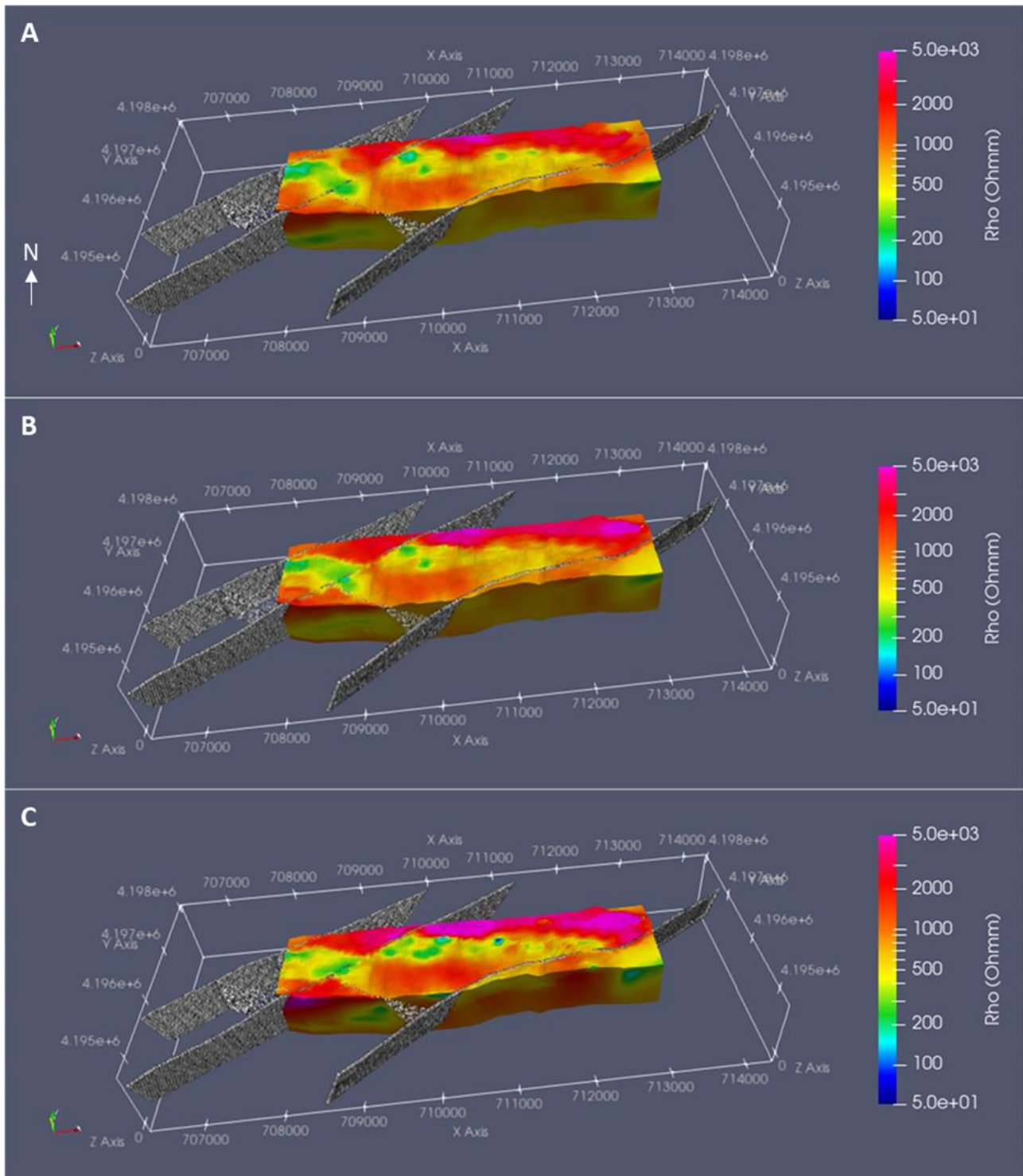


Fig. 5.7 AEM resistivity data model. Grey planes represent the faults. A: 1D standard inversion. B: 1D inversion with constraint-breaking across regions defined by the 3D geological model. C: 3D inversion with constraint-breaking across regions defined by the 3D geological model.

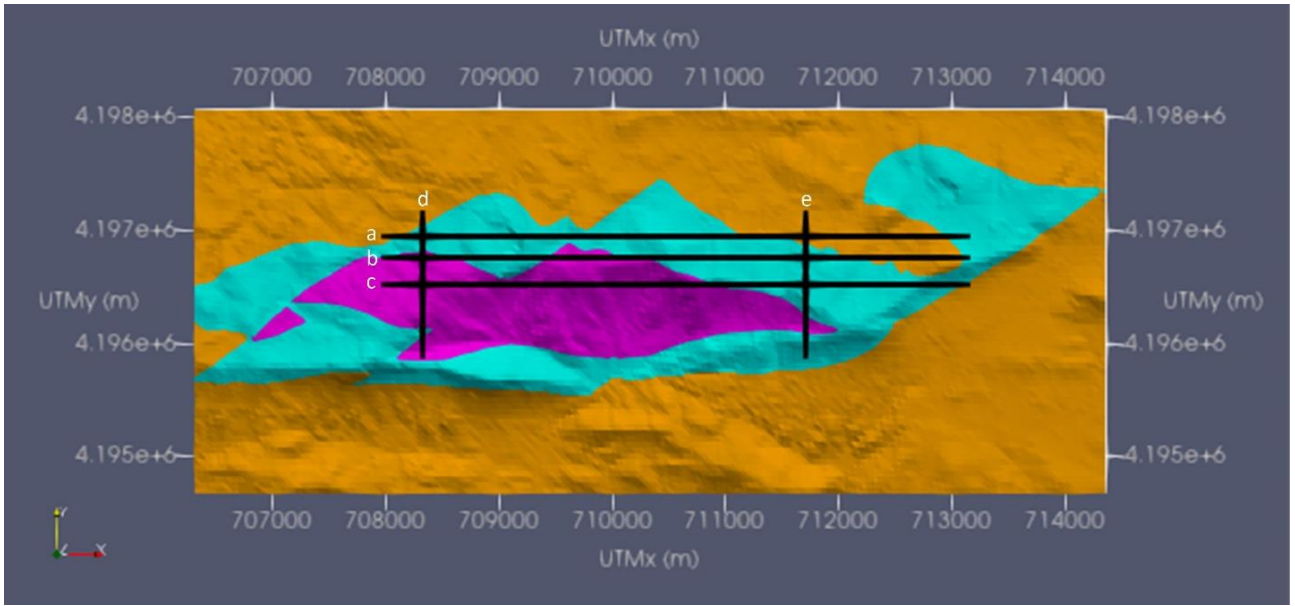


Fig. 5.8 Tracks of the 2D resistivity cross sections (a-e) projected on the geological model of Aracena study area.

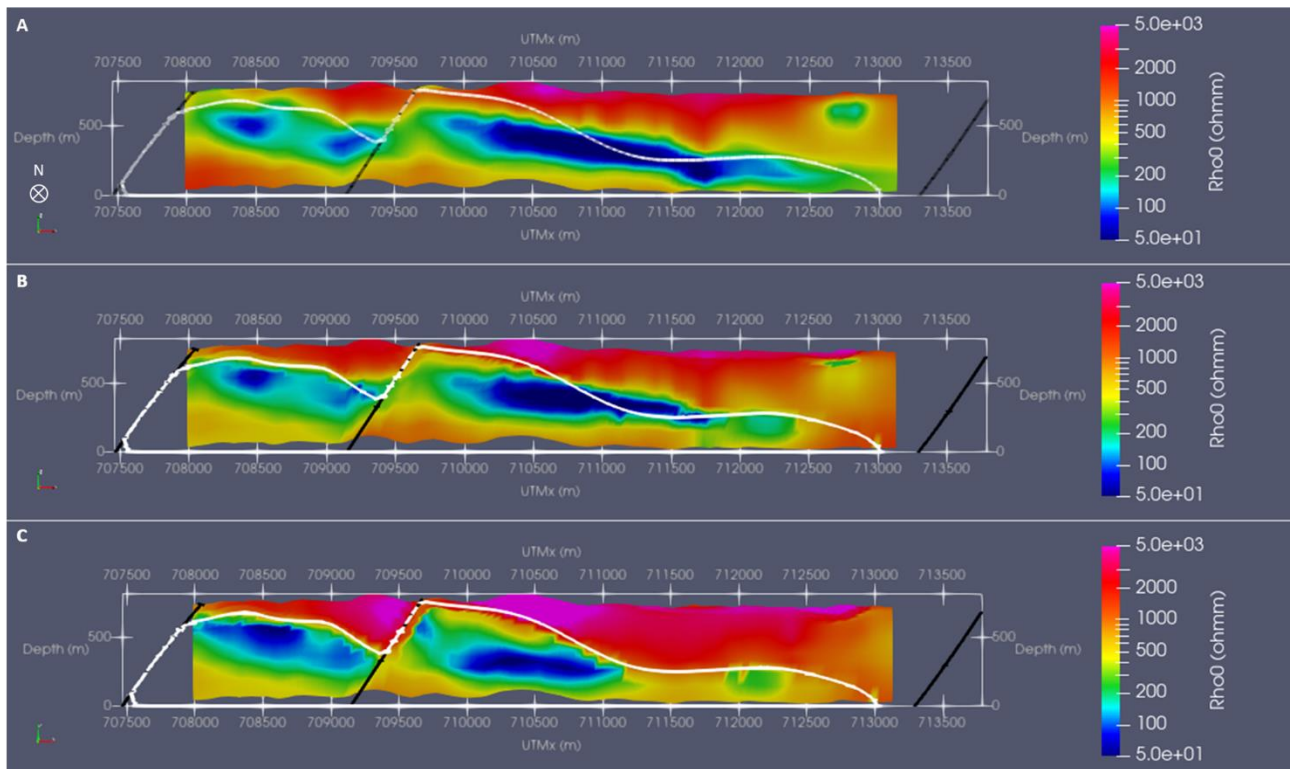


Fig. 5.9 AEM resistivity data model cross section 'a'. A: 1D standard inversion. B: 1D inversion with constraint-breaking across regions defined by the 3D geological model. C: 3D inversion with constraint-breaking across regions defined by the 3D geological model. White line represents the interface of the schists, black lines represent the faults. Aspect ratio 1.

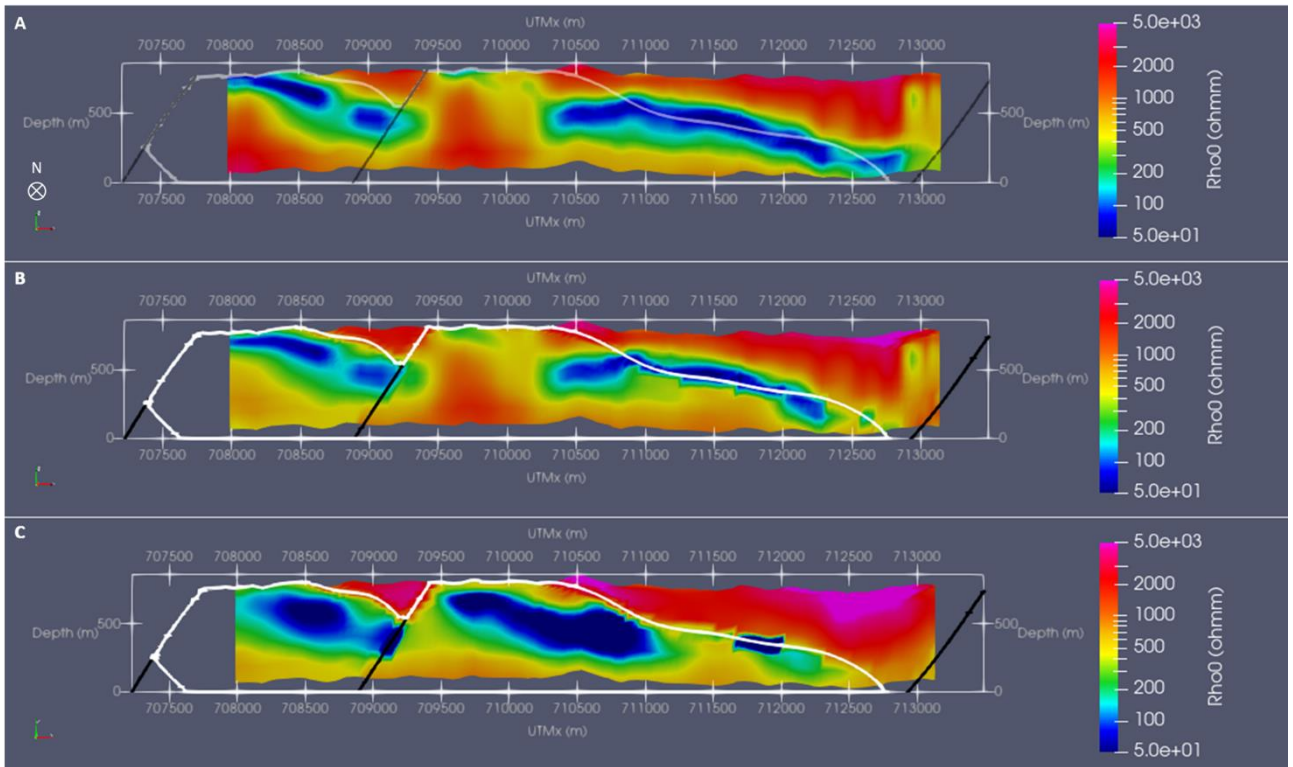


Fig. 5.10 AEM resistivity data model cross section 'b'. A: 1D standard inversion. B: 1D inversion with constraint-breaking across regions defined by the 3D geological model. C: 3D inversion with constraint-breaking across regions defined by the 3D geological model. White line represents the interface of the schists, black lines represent the faults. Aspect ratio 1.

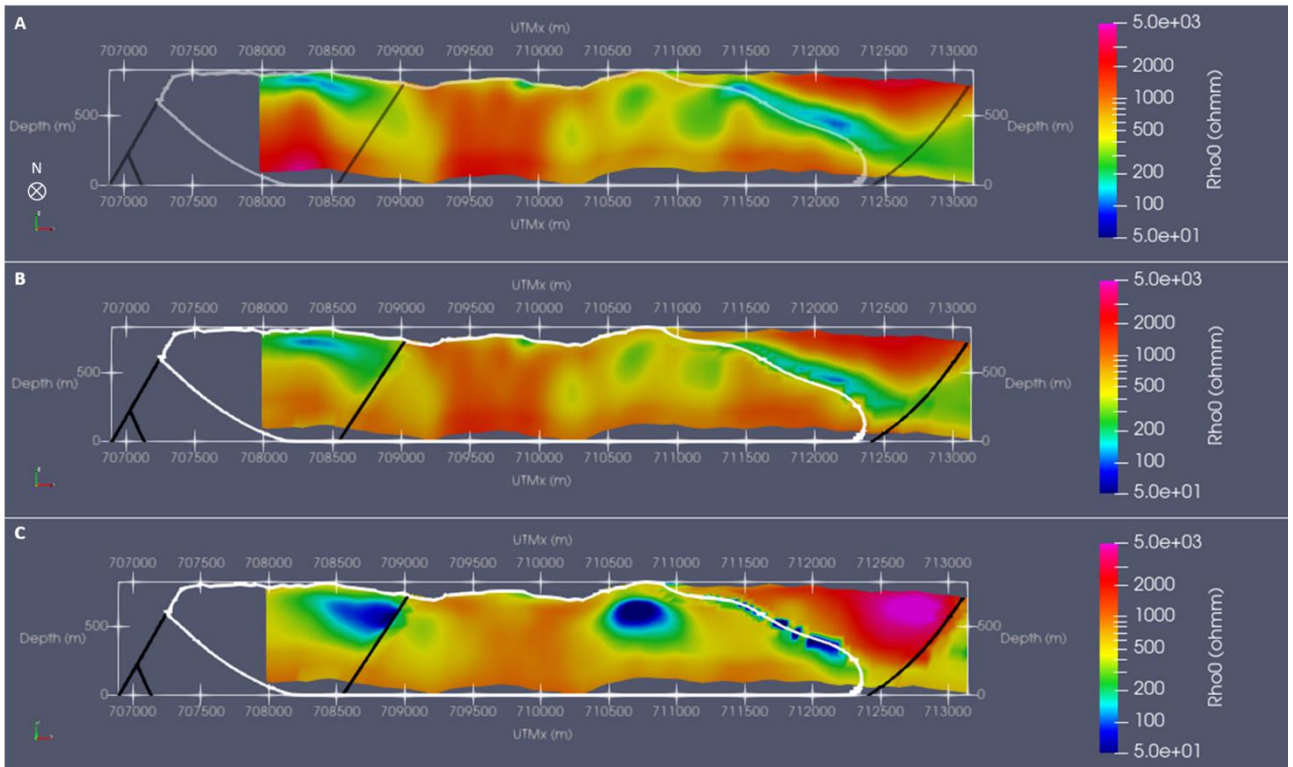


Fig. 5.11 AEM resistivity data model cross section 'c'. A: 1D standard inversion. B: 1D inversion with constraint-breaking across regions defined by the 3D geological model. C: 3D inversion with constraint-breaking across regions defined by the 3D geological model. White line represents the interface of the schists, black lines represent the faults. Aspect ratio 1.

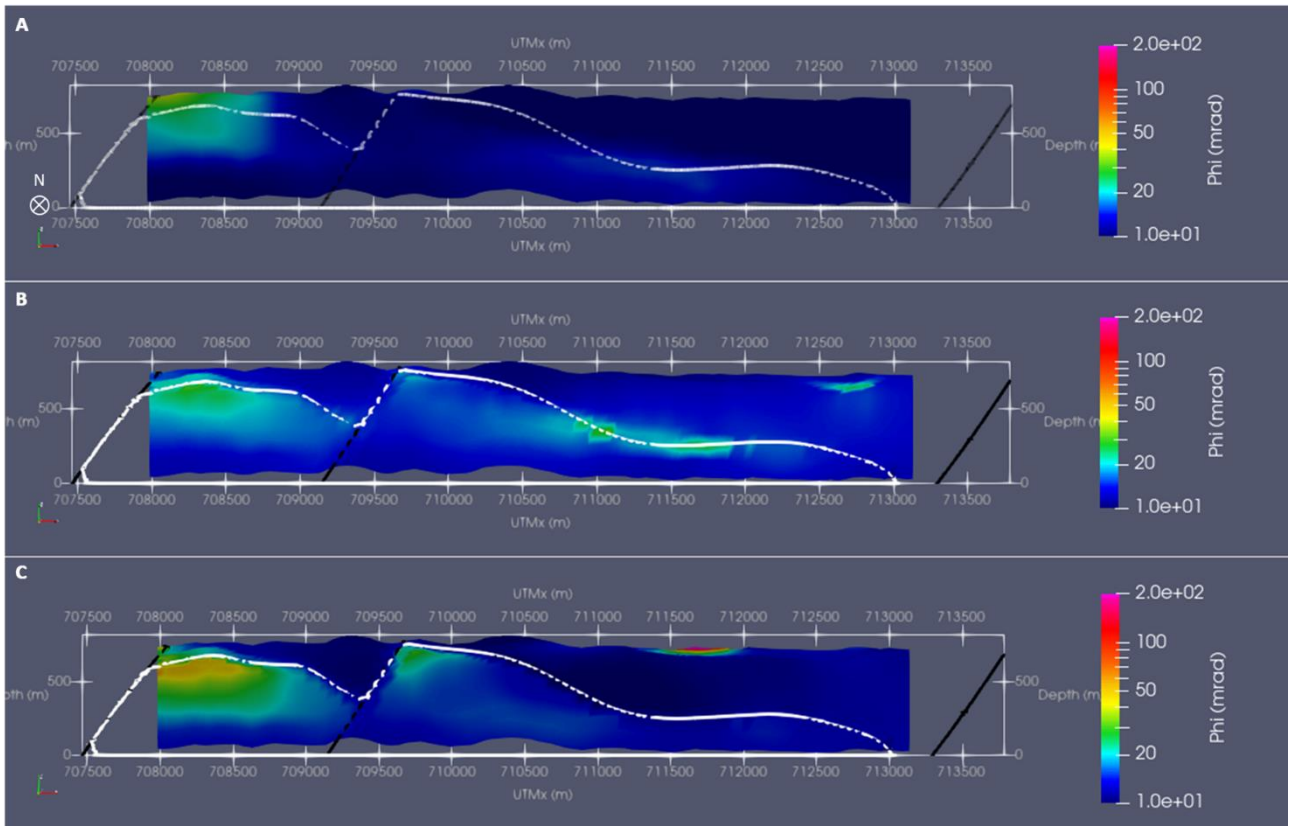


Fig. 5.12 AEM IP data model cross section 'a'. A: 1D standard inversion. B: 1D inversion with constraint-breaking across regions defined by the 3D geological model. C: 3D inversion with constraint-breaking across regions defined by the 3D geological model. White line represents the interface of the schists, black lines represent the faults. Aspect ratio 1.

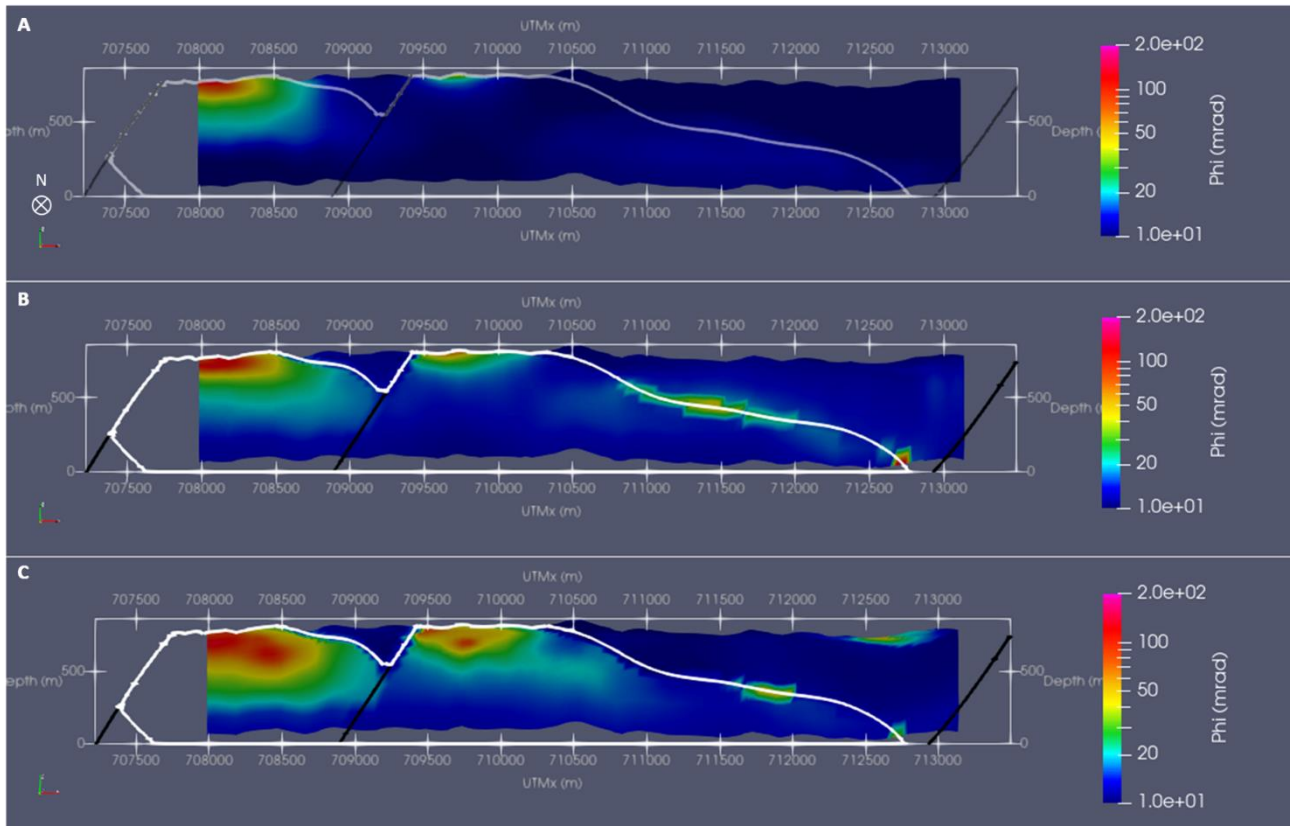


Fig. 5.13 AEM IP data model cross section 'b'. A: 1D standard inversion. B: 1D inversion with constraint-breaking across regions defined by the 3D geological model. C: 3D inversion with constraint-breaking across regions defined by the 3D geological model. White line represents the interface of the schists, black lines represent the faults. Aspect ratio 1.

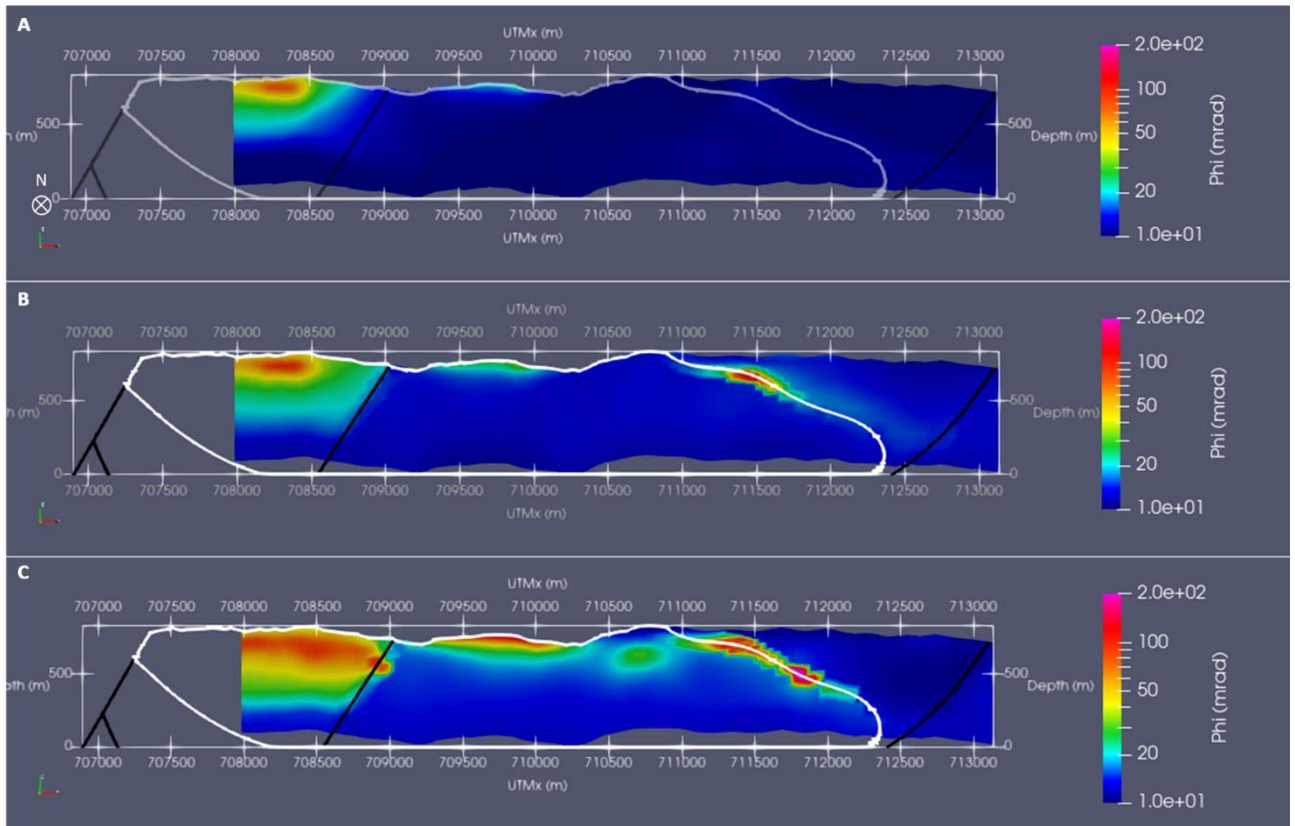


Fig. 5.14 AEM IP data model cross section 'c'. A: 1D standard inversion. B: 1D inversion with constraint-breaking across regions defined by the 3D geological model. C: 3D inversion with constraint-breaking across regions defined by the 3D geological model. White line represents the interface of the schists, black lines represent the faults. Aspect ratio 1.

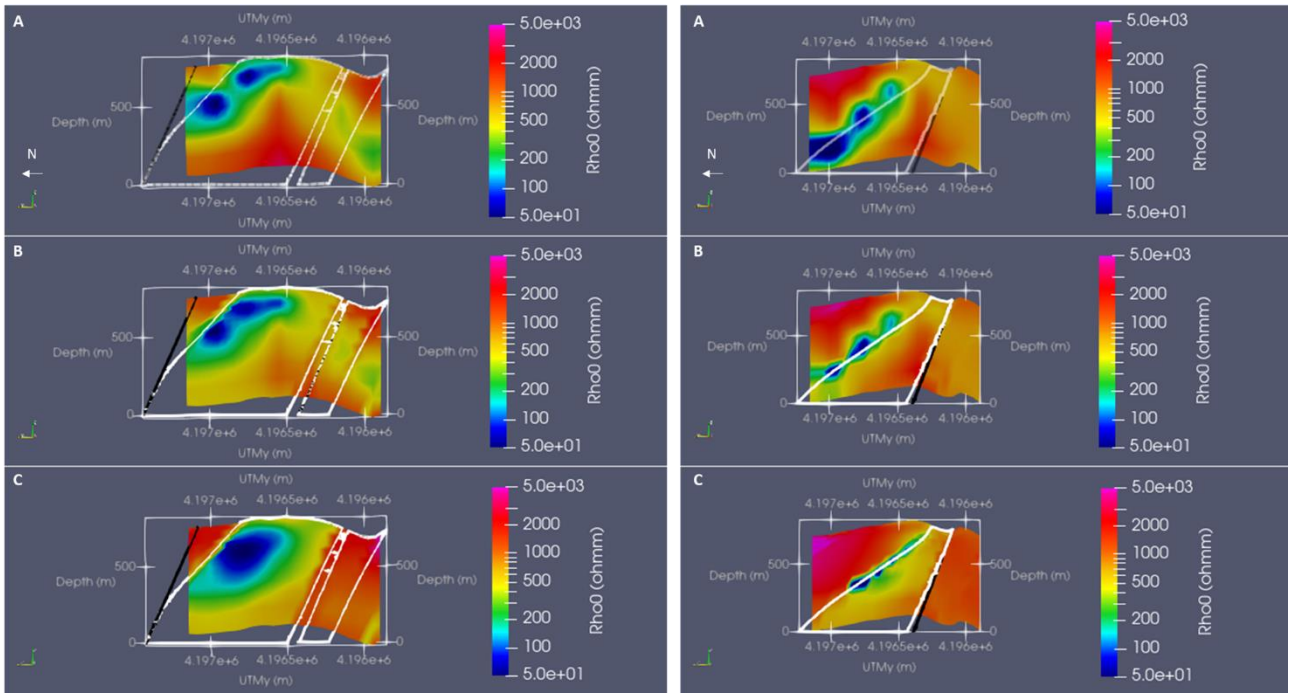


Fig. 5.15 AEM resistivity data model cross section 'd' on the left and 'e' on the right. A: 1D standard inversion. B: 1D inversion with constraint-breaking across regions defined by the 3D geological model. C: 3D inversion with constraint-breaking across regions defined by the 3D geological model. White line represents the interface of the schists, black lines represent the faults. Aspect ratio 1.

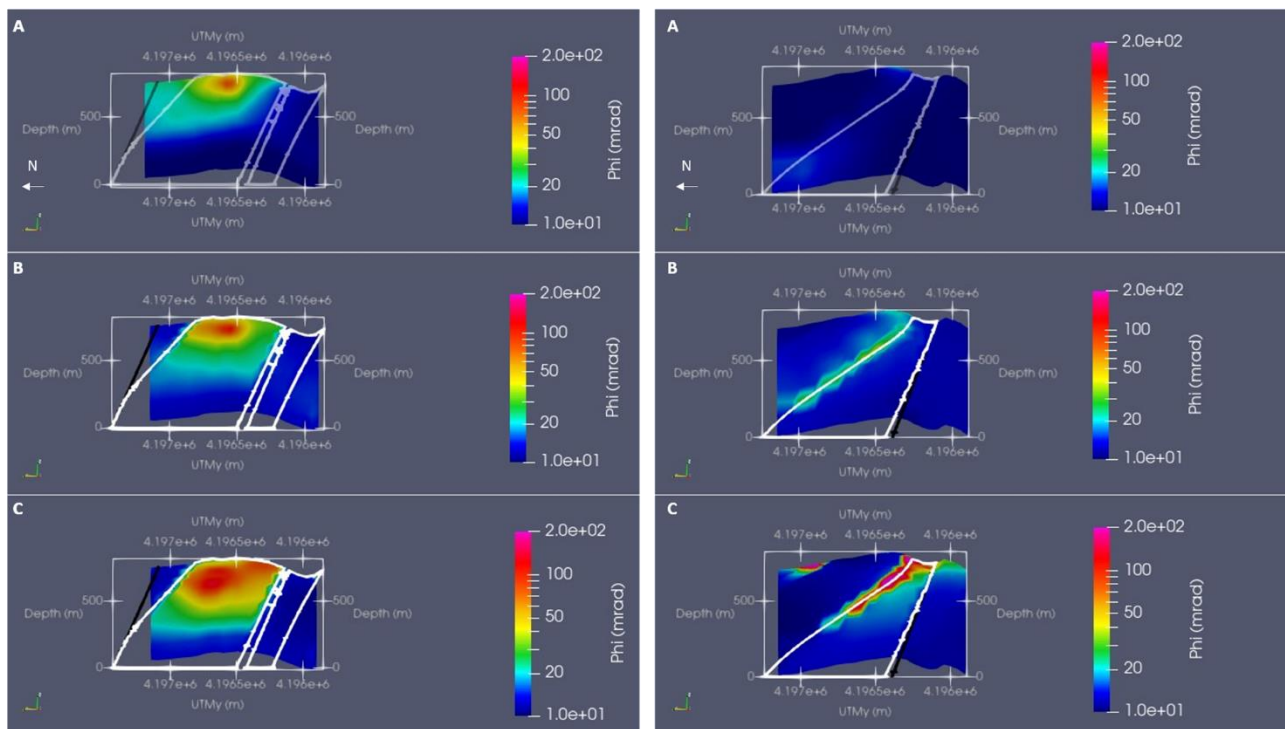


Fig. 5.16 AEM IP data model cross section 'd' on the left and 'e' on the right. A: 1D standard inversion. B: 1D inversion with constraint-breaking across regions defined by the 3D geological model. C: 3D inversion with constraint-breaking across regions defined by the 3D geological model. White line represents the interface of the schists, black lines represent the faults. Aspect ratio 1.

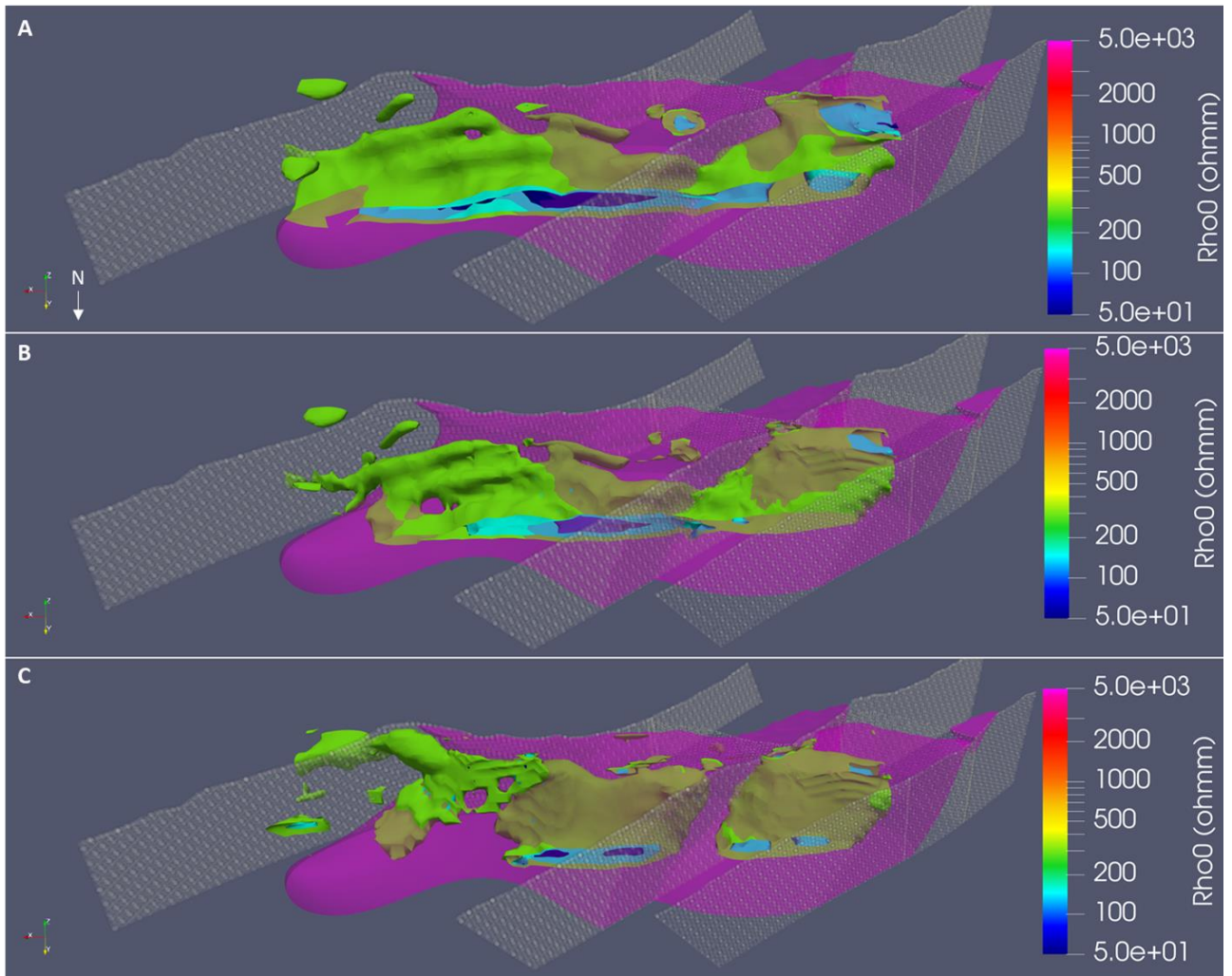


Fig. 5.17 Conductive bodies within the schists (pink volume) and at the top of their interface with dolomites clipped based on resistivity values (green: 300 Ωm , light blue: 150 Ωm , blue: 50 Ωm). Grey planes represent the faults. A: 1D standard inversion. B: 1D inversion with constraint-breaking across regions defined by the 3D geological model. C: 3D inversion with constraint-breaking across regions defined by the 3D geological model.

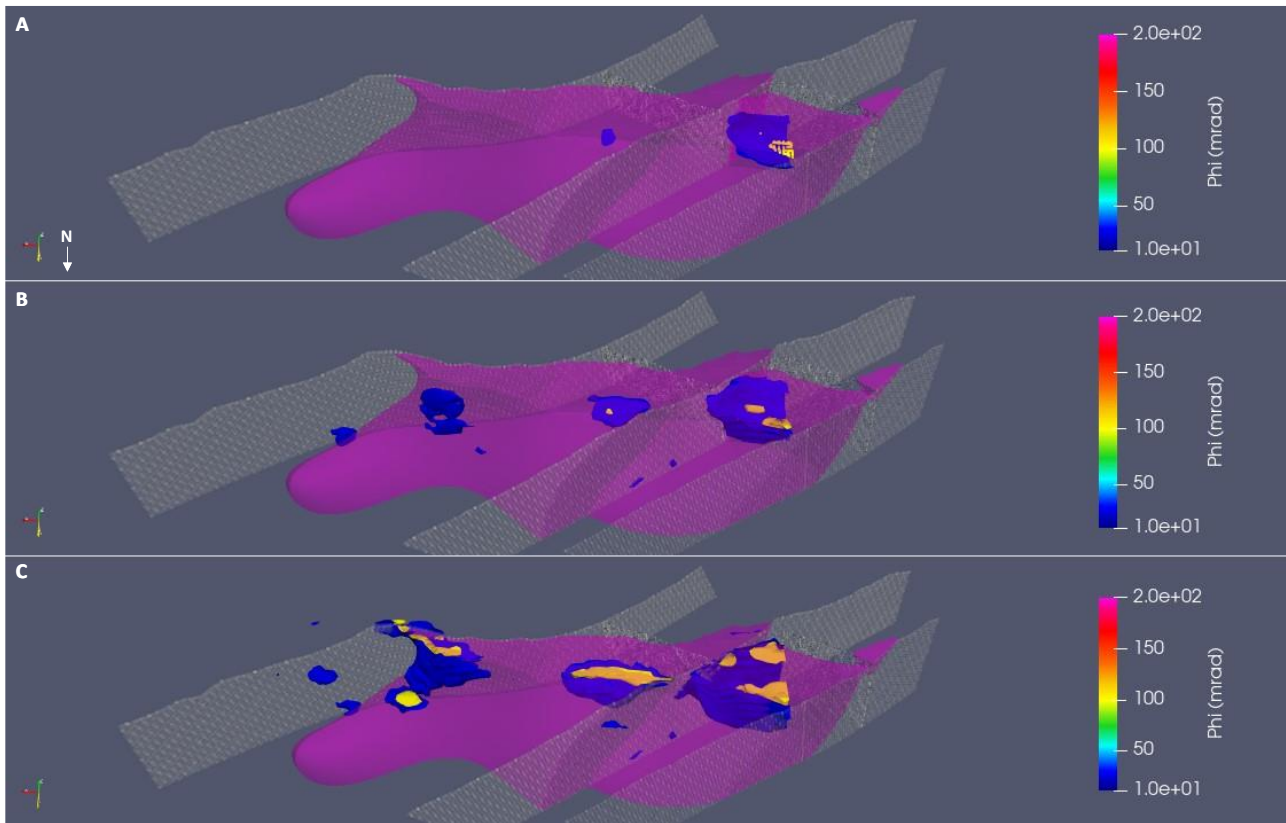


Fig. 5.18 Chargeable bodies within the schists (pink volume) and at the top of their interface with overlying dolomites clipped based on phi values (yellow: 100 mrad, green blue: 30 mrad, blue: 6mrad). Grey planes represent the faults. A: 1D standard inversion. B: 1D inversion with constraint-breaking across regions defined by the 3D geological model. C: 3D inversion with constraint-breaking across regions defined by the 3D geological model.

6. Conclusions

A multidisciplinary approach was developed for the characterization of springs in mountain environments, based on the integration of the three key pillars required to understand subsurface water flow systems: geology, geophysics, and hydrogeology. These pillars are needed simultaneously because the information they provide is complementary and require a joint analysis.

In particular, an integrated and iterative modelling strategy was developed to jointly use geological and geophysical data for the identification of potential recharge zones and preferential groundwater flow paths in karst-fractured mountain environments. This approach relies on two key components. First, detailed surface geology, including the mapping of lithological units and major structural features, is used to build a consistent 3D geological model of the study area. Second, high resolution geophysical data with broad spatial coverage are incorporated to refine and constrain the model. These geophysical datasets are acquired through both ground-based methods (such as Loupe) and airborne surveys (AEM), allowing for subsurface characterization in remote or inaccessible terrains. The significant topographical variations and the strong resistivity contrasts typical of mountain areas require a 3D modelling of the AEM data, as proven by both synthetic and field data analysis.

In addition, numerical hydrogeological modelling was employed to test and validate (or falsify) the conceptual model of groundwater flow dynamics feeding a spring. This modelling effort aimed to assess the plausibility of the hypothesized recharge mechanisms, flow pathways, and discharge processes within a complex hydrogeological setting. By simulating different scenarios and adjusting key parameters (hydraulic conductivity, conduit aperture, and recharge distribution) the numerical model allowed for a quantitative evaluation of the conceptual understanding of the system. In this way, it provides critical insights into how groundwater flows through fractured and/or karst zones and eventually contributes to spring discharge.

Two study sites were used for the development and validation of the proposed methodology. Val Sabbia site served as a testing ground where extensive field-based geophysical acquisitions and detailed geological modelling were carried out. This site allowed for the development and refinement of our integrated approach, particularly in terms of combining 3D geological modelling with geophysical inversions and hydrogeological characterization. However, the final step, i.e. integrating AEM data into the geophysical model informed by geology, could not be completed because the AEM data acquired in the area were not made available for this study.

However, a synthetic study informed by the geological model was carried out, proving the effectiveness of the geologically-informed inversion developed in this study and the necessity of 3D AEM modelling in mountain areas for spring characterization. The AEM synthetic modelling approach highlights critical limitation of 1D inversion: it can fit the data well even when the underlying model contains unrealistic features. In other words, 1D inversion can provide a misleading sense of reliability. This underscores the importance of using 3D inversion approaches, especially in complex geological settings. Moreover, the common practice of discarding high-misfit data as a quality control step in inversion processing proves insufficient on its own. Even after this filtering, there remains a substantial risk of misinterpreting artifacts as real geological features. Furthermore, numerical modelling was carried out for building the conceptual model of the groundwater flow.

The Aracena case study offered the opportunity to test on field data the effectiveness of AEM data for characterizing groundwater resources in mountainous terrain and integrating them with a 3D geological model. The availability of high-coverage AEM data in this area enabled the identification of subsurface features potentially associated with water-bearing zones and supported the refinement of the hydrogeological conceptual model in a structurally complex setting.

As evidenced by the synthetic models developed for the Val Sabbia area, the geologically-constrained 3D AEM inversion in Aracena, which accounted also for induced polarization, produced results that aligned more closely with the expected geological framework and facilitated a more straightforward interpretation of subsurface structures.

The systematic multidisciplinary methodology developed in this study will contribute to the comprehensive understanding of the hydrodynamic of natural springs in mountain environments and to the assessment of their potential for sustainable water resource management, with a holistic interpretation based on geology, geophysics, and hydrogeology.

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