



VENTILATION DYNAMICS AND HEAT EXCHANGE IN CAVES: AN INTEGRATED MONITORING AND MODELING APPROACH

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Ventilation Dynamics and Heat Exchange in Caves: An Integrated Monitoring and Modeling Approach

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Abstract

This thesis investigates ventilation dynamics and heat exchange along ventilated caves. For this work, we prioritised monitoring caves ventilated by a chimney effect, which is due to density/temperature contrasts between the cave and the external atmosphere. Understanding these processes is crucial for various fields, including paleoclimatic reconstructions using speleothems, the conservation of subterranean ecosystems, groundwater quality, carbon cycle assessment and shallow geothermal extraction. The work presents an integrated approach of monitoring and modeling to address gaps in understanding the effects of ventilation on cave climatological conditions and to validate thermal and ventilation numerical models with field data.

In the first part (Part 1), this work focuses on the monitoring methods and strategies adopted to carry out reliable air temperature and air flow measurements. Along ventilated caves, air temperature signals from outside are transferred along cave passages for a distance known as the convection length. Moving away from the entrances where air is drawn into the cave, the heat exchange that occurs along the air convection length attenuates the temperature signals before thermal equilibrium is reached between the air and the rock.

Therefore, airflow, together with air temperatures, is a key parameter to monitor. We adapted and thoroughly tested in the cave environment a low-cost and robust digital flowmeter (Sensirion SFM3003-300-CE) for multi-year airflow monitoring. While the device showed high laboratory accuracy ($\pm 1\%$ for 0.5-5 m/s), *in-situ* measurements carried out by comparing three SFMs in the same cave cross-section exhibited significant uncertainties (20% to $\pm 50\%$). These results are compared against a numerical model, which suggests that these uncertainties are mainly due to non-uniform velocity distribution within complex cave cross-sections, and, thus, instrument positioning is critical. Gauging tests with artificial CO₂ were also used to assess cave airflow. This method is likely the most accurate for one-time airflow assessment, as it does not require velocity integration over a cross-section. In this work, it is also emphasised that measuring airflow with uncertainties lower than 20-30% is not obvious. Therefore, according to the precision required, comparing multiple methods to assess air flow along cave conduits will help to estimate uncertainties affecting airflow measurements. Air temperature measurements also presented challenges near cave entrances due to heat radiation and sensor characteristics. Here, uncertainties on air temperature measurements can reach $\pm 20\text{-}30\%$ of the measured value if instruments are positioned close to the walls. Instead, far from entrances, these uncertainties reduce to $\sim \pm 5\%$ of the measured value.

Investigating airborne matter (including aerosols and gases) dispersion is critical for managing potential degradation in cave environments from external pollutants and visitor disturbances. Gauging tests with artificial CO₂ also revealed that Taylor's theory significantly underestimates the dispersion coefficient in natural conduits due to their complex geometries and roughness. Breakthrough curves carried out from ventilated conduits of one mine and two caves exhibit pronounced "tailing" in natural caves. This indicated that dispersive singularities and chambers along the cave paths slow down the tracer, promoting retardation and high dispersion. Therefore, a 1-D advection-dispersion model is adequate for simple geometries but insufficient for complex caves, requiring more sophisticated numerical simulations (e.g., based on 3-D Navier-Stokes equations).

In the second part of the work (Part 2), monitoring data are used to assess heat exchange along the convection length, to investigate cave ventilation dynamics and to validate numerical models.

A numerical model coupling convective heat transfer in air with conductive heat transfer in the rock mass was used to investigate the propagation of thermal perturbations along the convection length.

This distance was found to be approximately proportional to the amplitude of the annual airflow fluctuations divided by the square root of the cave's mean radius.

Three years of monitoring allowed us to estimate the energy exchange along Longeague cave, which was found to be in the order of $89 \pm 50\%$ GJ. Moreover, the observed convection lengths (thermal anomalies) of upper and lower cave entrances extend up to 150 m, while 90% of the energy exchange between air and rock occurs within the first 50 meters of the convection length. The study suggests that the deep karst zone may not be present in ventilated caves shorter than their convection length. Particular attention was paid to investigate and derive cave aeraulic resistance in order to build a reliable cave ventilation model. In this work, we show that it is possible to simulate ventilation (based on the momentum balance applied to the cave atmosphere) by knowing cave and external temperatures, provided that the cave aeraulic resistance is known. This achievement is a step forward for paleoclimatic studies and cave protection and management. Ventilation modeling helps to simulate plausible past cave ventilation and to forecast effects on ventilation if aeraulic resistance is changed (e.g. by opening or closing cave passages with doors in tourist caves).

In the general discussion, the concept of convection length is also qualitatively extended to hydrologically active systems, where water flows can transport external temperature signals for hundreds or kilometres, depending on water flow, recharge modes and system depth.

The results of this thesis represent a step forward in understanding cave ventilation related to the chimney effect, and provide new, detailed insights into heat transfer mechanisms in karst systems, accounting for air convective heat exchange. We believe that this work, together with the “twin” PhD thesis “The Role of Convective Heat and Mass Transfer in the Thermal Response of Karst Conduits” by Sedaghatkish (2025), will serve as a valuable starting point for researchers interested in cave (paleo)climatology, hydrogeology, geothermal energy extraction, and geosciences in general.

Keywords

Cave ventilation; cave temperature; chimney effect; aeraulic (aerology) resistance; cave monitoring; numerical modeling; convective heat and mass transfer; thermal anomalies; convection length.

Résumé

Cette thèse étudie les dynamiques de ventilation et les échanges thermiques le long de cavités ventilées. Ce travail a priorisé l'observation de grottes ventilées par l'effet cheminée, provoqué par des contrastes de densité et de température entre l'atmosphère externe et l'intérieur de la grotte. Comprendre ces processus est essentiel dans plusieurs domaines, notamment la les reconstitutions paléoclimatiques à l'aide de spéléothèmes, la conservation des écosystèmes souterrains, la qualité des eaux souterraines, l'évaluation du cycle du carbone et l'exploitation géothermique peu profonde. Ce travail propose une approche intégrée de monitoring et de modélisation afin de combler les lacunes concernant les effets de la ventilation sur les conditions climatiques souterraines, et de valider les modèles numériques thermiques et de ventilation par des données de terrain.

Dans la première partie (Partie 1), ce travail se concentre sur les méthodes et stratégies de surveillance adoptées pour effectuer des mesures fiables de température de l'air et de flux d'air. Dans les cavités ventilées, les signaux de température de l'air extérieur sont transférés à l'intérieur sur une distance appelée longueur de convection. En s'éloignant des entrées, là où l'air est aspiré dans la grotte, les échanges thermiques qui se produisent le long de cette longueur de convection atténuent les signaux de température jusqu'à atteindre un équilibre thermique entre l'air et la roche.

Par conséquent, le flux d'air, conjointement à la température de l'air, est un paramètre clé à surveiller. Un débitmètre numérique robuste et à faible coût (Sensirion SFM3003-300-CE) a été adapté et longuement testé en environnement souterrain pour une surveillance du flux d'air sur plusieurs années. Bien que l'appareil ait montré une grande précision en laboratoire (± 1 % pour 0,5–5 m/s), les mesures *in situ*, réalisées en comparant trois SFM sur une même section de la grotte, ont présenté des incertitudes significatives (de 20 % à ± 50 %). Ces résultats sont comparés à un modèle numérique, suggérant que ces incertitudes sont principalement dues à une distribution non uniforme de la vitesse dans des sections transversales complexes, rendant ainsi le positionnement des instruments critique. Des tests de jaugeage avec du CO₂ artificiel ont également été utilisés pour évaluer le flux d'air dans la grotte. Cette méthode est probablement la plus fiable pour une évaluation ponctuelle du flux d'air, car elle ne nécessite pas l'intégration des vitesses sur une section transversale. Ce travail souligne également que mesurer le flux d'air avec des incertitudes inférieures à 20–30 % n'est pas trivial. Ainsi, selon le niveau de précision requis, la comparaison de plusieurs méthodes d'évaluation du flux d'air le long des conduits souterrains permet d'estimer les incertitudes associées aux mesures. Les mesures de température de l'air ont également posé des défis à proximité des entrées en raison du rayonnement thermique et des caractéristiques des capteurs. Dans ces zones, les incertitudes peuvent atteindre ± 20 –30 % de la valeur mesurée si les instruments sont placés près des parois. En revanche, loin des entrées, ces incertitudes se réduisent à environ ± 5 % de la valeur mesurée.

L'étude de la dispersion de la matière en suspension (notamment les aérosols et les gaz) est cruciale pour la gestion des risques de dégradation liés aux polluants extérieurs ou à la fréquentation touristique. Les tests avec du CO₂ artificiel ont également montré que la théorie de Taylor sous-estime fortement le coefficient de dispersion dans les conduits naturels en raison de leur géométrie complexe et de leur rugosité. Les courbes de percée mesurées dans une mine et deux grottes ventilées montrent une traîne importante dans les cavités naturelles. Cela indique que des singularités dispersives et des chambres situées le long des conduits ralentissent le traceur, favorisant la rétention et une dispersion accrue. Un modèle unidimensionnel d'advection-dispersion convient pour des géométries simples, mais est insuffisant pour les grottes complexes, nécessitant alors des simulations numériques plus sophistiquées (par exemple basées sur les équations de Navier-Stokes 3D).

Dans la deuxième partie du travail (Partie 2), les données de surveillance sont utilisées pour évaluer les échanges thermiques le long de la longueur de convection, pour étudier la dynamique de ventilation des grottes et pour valider les modèles numériques.

Un modèle numérique couplant le transfert thermique convectif dans l'air et transfert thermique conducteur dans la roche a été utilisé pour étudier la propagation des perturbations thermiques. Il a été montré que cette distance est approximativement proportionnelle à l'amplitude des fluctuations annuelles du flux d'air divisée par la racine carrée du rayon moyen de la cavité.

Trois années de surveillance ont permis d'estimer les échanges énergétiques de l'air dans la grotte de Longeaigne, évalués à environ 89 ± 50 % GJ. De plus, les longueurs de convection observées (anomalies thermiques) aux entrées supérieure et inférieure de la grotte s'étendent jusqu'à 150 m, alors que 90% de l'échange d'énergie entre l'air et la roche a lieu dans les 50 premiers mètres. L'étude suggère que la zone karstique profonde pourrait être absente dans les grottes ventilées plus courtes que leur longueur de convection. Une attention particulière a été portée à l'analyse et à la détermination de la résistance aéraulique des grottes afin d'établir un modèle de ventilation fiable. Ce travail montre qu'il est possible de simuler la ventilation (basée sur un bilan de quantité de mouvement appliqué à l'atmosphère de la grotte) en connaissant les températures interne et externe, à condition de connaître la résistance aéraulique de la cavité. Il s'agit d'une avancée significative pour les études paléoclimatiques et pour la protection et la gestion des grottes. Cette modélisation pourrait permettre de simuler des régimes de ventilation passés plausibles et de prédire les effets d'une modification de la résistance aéraulique sur la ventilation (par exemple, les modifications dues à l'ouverture ou à la fermeture de passages dans les grottes touristiques).

Dans la discussion générale, le concept de longueur de convection est aussi qualitativement étendu aux systèmes hydrologiquement actifs, dans lesquels les écoulements d'eau peuvent transporter les signaux thermiques externes sur des distances de plusieurs centaines de mètres à plusieurs kilomètres, selon les débits, les modalités de recharge et la profondeur du système.

Les résultats de cette thèse représentent une avancée dans la compréhension de la ventilation des grottes liée à l'effet de cheminée et apporte des détails inédits et significatifs sur les mécanismes de transfert de chaleur dans les systèmes karstiques, en tenant compte des échanges thermiques convectifs dans l'air. Ce travail, ainsi que la thèse de doctorat "jumelle" de Sedaghatkish (2025) intitulée "The Role of Convective Heat and Mass Transfer in the Thermal Response of Karst Conduits", constituent un point de départ précieux pour les chercheurs intéressés par la (paléo)climatologie des grottes, l'hydrogéologie, l'extraction de l'énergie géothermique et les géosciences en général.

Mots-clés

Ventilation et température des grottes; effet cheminée; résistance aéraulique; suivi des grottes; modélisation numérique; transfert convectif de chaleur et transfert de masse; longueur de convection.

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Chapter 1. Introduction

The term “**Karst**” refers to a landscape with unique morphologies and hydrology that hosts caves and extensive underground water systems (Ford and Williams, 2007). The distinctive landscape of the karst is mainly the result arising from the dissolution of soluble rocks (e.g. carbonates and evaporitic rocks) by water. However, here with the term karst we refer exclusively to the “classical karst” in carbonates (Jones and White, 2019; Benac et al., 2013; Knez, 2011). Water carves these terrains via subsurface drainage (e.g. *Karren fields* or *Lapiaz* in French), forming enclosed depressions (dolines, uvala and sinkholes), sinking streams, caves and springs. Aquifers in karst rocks typically exhibit a well-developed secondary porosity (conduits), primarily due to the widening of discontinuities through dissolution. Surface water infiltrates preferentially through fractures, depressions and swallow holes, which are hydrologically connected with the underground networks of conduits (caves) and eventually springs (where the underground water comes back to the surface). The development and features of karst systems are influenced by numerous factors, including lithology, stratigraphy, geological structure, topography, precipitation, temperature, water's chemical composition, and base-level fluctuations. These fluctuations can be driven by processes such as valley incision, sea-level changes, tectonic activity, or diapirism. Therefore, variations of regional or local base levels of erosion influence the long-term evolution of the karst systems (De Waele and Gutiérrez, 2022; Ford and Williams, 2007).

The intersections between the cave passages and the surface represent inlets and outlets of air and water. These can range from large cave entrances to fissures less than a centimetre wide. In the early stages of karstification, water is the most abundant fluid filling small voids associated with rock porosity and minor fractures. As the karstification proceeds and the vadose zone is formed, some cave passages are abandoned by water (Figure 1.1) and air may become the most ubiquitous fluid (Luetscher et al., 2008; Luetscher and Jeannin, 2004a). Indeed, ventilation and ventilated caves are key features of the karst. Consequently, water and air, which abundantly traverse the karst, transport heat (resulting in temperature variations) from the external atmosphere deep into the karst. The result of this is that these fluids control the climatology underground, contributing to fix the temperature and vertical temperature gradients (Badino, 2010a; Luetscher and Jeannin, 2004a) and contributing to karstification and speleogenesis (Badino, 1995; Fairchild and Baker, 2012).

On one side, many works have been focusing on heat flow characteristics within the phreatic zone, e.g. (M D Covington et al., 2011; Liedl et al., 1998; Benderitter et al., 1993), on the other hand, much less can be found about the air advected underground. Yet, there is still debate on the relative importance of the heat transported by air and water; therefore, quantifying the heat exchanged by advected air is becoming crucial to find clues to this question. Researchers tried to quantify the energy exchanged in karst systems, mainly by physical models (e.g. Badino, 2018, 1995; Covington and Perne, 2015; Jeannin et al., 1997; Lismonde, 2002; Wigley and Brown, 1976, 1971) and a few by using field data (e.g., Obleitner and Spötl, 2011; Luetscher et al., 2008). The availability of field data is therefore critical. Notably, monitoring air temperature and air flow rate in caves is the basis for the assessment and quantification of heat exchanges between air and the surroundings. Covington and Perne (2015), Badino (2010) and Luetscher and Jeannin (2004a) underline that there are few field studies on airflow dynamics and the necessity of more field investigations to help in other research topics.

Understanding the ventilation dynamics in karst systems, particularly their impact on underground climatology as well as on energy and mass transfer, is essential for numerous scientific and practical reasons. Karstification and speleothem formations (such as stalactites and stalagmites, for instance)

are influenced by ventilation. Studying ventilation dynamics and temperature variations is crucial for interpreting paleoclimate data and distinguishing external paleoenvironmental signals from local cave conditions. It is known that speleothems are capable of recording important information on external climatic variations over millennial scales and that their isotopic composition is often influenced by air movements (Cuthbert et al., 2014; Banner et al., 2007; Spötl et al., 2005). Condensation and evaporation related to airflow are also thought to play a major role in karstification and speleothem formation (Fairchild and Baker, 2012; Dreybrodt et al., 2005). Therefore, cave ventilation is suggested to control cave temperature, humidity, $p\text{CO}_2$ and isotopic fractionation (e.g., $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$) in speleothem calcite. Ventilation and temperature influence CO_2 degassing from drip water, directly impacting speleothem growth rates. Variations in these factors can either accelerate or slow down growth (Rau et al., 2015; Fairchild and Baker, 2012; A. J. Kowalczyk and Froelich, 2010; Baldini et al., 2008; Spötl et al., 2005). Ventilation patterns are also meaningful for preserving groundwater quality (Jeannin et al., 2017), for evaluating the carbon cycle (Martin-Pozas et al., 2022; Prelovšek et al., 2018; Breecker et al., 2012), as it controls CO_2 concentrations (Kukuljan et al., 2021; Faimon and Lang, 2013; Kowalczyk and Froelich, 2010) in the whole volume of carbonate rocks, and even for heat pumping in geothermal extraction (Borović et al., 2019). As a last but not least note, ventilation is the guiding thread behind explorations held by speleologists in complex cave systems (e.g. Pastore et al., 2019; Cella et al., 2015; Gatti et al., 2015).

From an ecological point of view, cave ecosystems are extremely fragile environments since they are confined places with little spatial and temporal variability (White et al., 2019). Temperature regulation controlled by ventilation has implications for the evolution of habitats (Dredge et al., 2013) of numerous and often endemic species. Exchanges between internal and external atmospheres also introduce and disperse organic matter and nutrients for dwelling fauna, which is adapted to rough cave environments (Medina et al., 2023). Therefore, any disturbance and changes of this equilibrium can provoke their alteration. This is particularly important in show caves (Martin-Sanchez et al., 2015; Mulec, 2014), where the influence of tourism provokes, e.g. temperature and humidity variations that can lead to severe consequences (Martin-Sanchez et al., 2012). The degradation process related to evapo-condensation, heat, aerosols and CO_2 transport driven by air advection and natural convection through caves also impacts the preservation of prehistorical paintings found in several caves used as a shelter (Bourges et al., 2014; Quindos et al., 1987) and adornments in show caves (Addesso et al., 2022b, 2022a; Christoforou et al., 1994; Dredge et al., 2013).

Therefore, tourism impact must be monitored and mitigated to avoid damage to the natural heritage of karst (Addesso et al., 2022a). Moreover, there is increasing interest in airborne matter transport, deposition, distribution and incorporation in speleothems along ventilated caves (e.g. aerosols and pollutants) (Addesso et al., 2022b; Dredge et al., 2013; Faimon et al., 2019; Christoforou et al., 1994; Pashchenko et al., 1993). Particularly, there is concern among show cave managers about the potential degradation of the cave environment caused by aerosol deposition. Degradation can result from external pollutants or disturbances caused by visitors to the delicate cave environment and climate (Dredge et al., 2013) and related mitigation practices have become a significant issue. It is therefore necessary to understand also how airborne matter is dispersed underground. Radon gas is also present and transported along caves (Quindos Poncela et al., 2013), which is often used as a natural tracer for detecting ventilation (Pflitsch and Piasecki, 2003; Przylibski and Piasecki, 1998), but is also a threat to speleologists and show-cave guides, given its potential toxicity with long exposures (Atkinson et al., 1983; Craven and Smit, 2006; Gregoric et al., 2013; Kowalczyk and Froelich, 2010).

Then, understanding cave atmospheres, their variability and the factors controlling them is a matter of priority, e.g. for paleoclimate reconstructions (James et al., 2015; Lyu et al., 2020), cave preservation and management, habitat evolution (Medina et al., 2023; Mammola et al., 2019a; Perry, 2013) and cave thermal response to ongoing climate changes (Obleitner et al., 2024b).

1.1. Zonation of the karst

Although in this work we deal mainly with the airflow, introducing the following morpho-hydrological distinction is necessary since this is the main terminology used by geologists, hydrogeologists and speleologists when referring to karst. Consequently, in this work, we will use this differentiation when we want to identify a karst zone by referring to its position relative to the water table, thus indicating whether we are referring to a water-unsaturated or saturated zone.

Karst systems are preferably divided into (Figure 1.1):

- the vadose zone (above the water table);
- the phreatic zone (below the water table);
- the epiphreatic zone (in between the previous two)

Each hydrogeological zone has unique flow characteristics and plays a crucial role in the karst hydrogeology, speleogenesis (cave formation), heat distribution in the karst and general evolution of karst landscapes (Audra and Palmer, 2013; Ford and Williams, 2007; Luetscher and Jeannin, 2004a).

The vadose zone is the aerated zone of the karst, and it comprises also the soil (if there is any) and the epikarst, and it is along the vadose zone where ventilation takes place. Within the vadose zone, the water primarily flows downward by gravity and voids in the rock are only partially occupied by water, except after heavy rain, when some fill with water. The epikarst or subcutaneous zone is located at the top of the vadose zone. The epikarst is represented by a weathered rock top which is fissured and dissolved; consequently, infiltration is facilitated and quick through this layer (recharge water can be completely absorbed). As epikarst has a complex structure, part of the recharge water can be partially trapped within blind-fissured and poorly connected zones, while part of the water is quickly transferred toward the vertical conduits and shafts of the unsaturated zone below (Perrin et al., 2003).

The vadose zone feeds the phreatic zone through well-developed shafts and sub-vertical conduits, which could alternate with subhorizontal conduits with free-surface streams and morphologies similar to those found on the surface, like meanders, canyons and waterfalls. The downward flow can be interrupted by local water pooling, resulting from geological structures that lead to perched phreatic zones. When descending water reaches the water table, marking the top of the phreatic zone. Here, the flow is mainly controlled by hydraulic gradients. In the phreatic zone, water completely fills closed and rounded conduits and flows through pathways that generally undulate up and down in the direction of the flow. The hydraulic gradient is generally gentle and follows the most efficient paths to nearby valley outlets. The epiphreatic zone, represented by the maximum height reached by rising water during high-water stages, underlines the dynamicity of the water table in karst systems which is highly irregular, fluctuating significantly in response to changes in infiltration rates. This dynamic process profoundly influences cave features. The epiphreatic zone is periodically flooded during high water and drained during low water, exhibiting characteristics of both vadose and phreatic flow (Audra and Palmer, 2013).

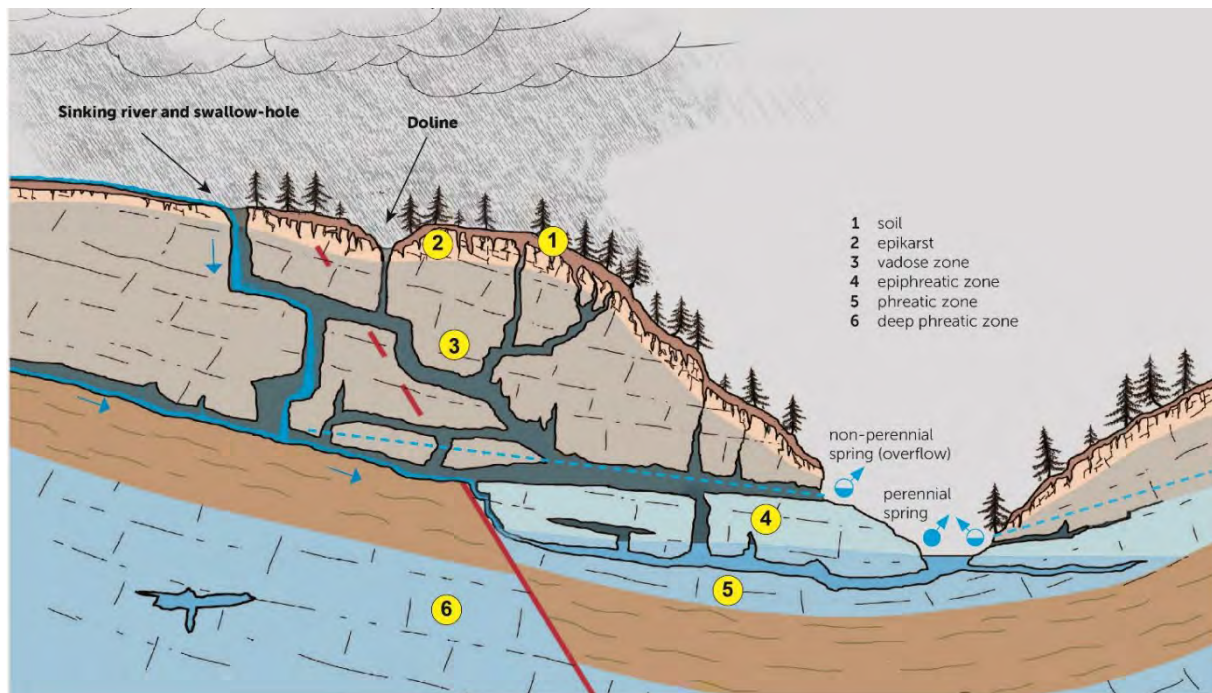


Figure 1.1 – A simplified sketch of the karst zonation (modified after Wenger et al., 2024). The picture is a copyright of SISKa.

1.2. Cave climate and meteorology

Cave climate is known for its constancy in comparison to that on the surface. Moreover, it is frequently asked what the temperature of a cave is, and it is usually answered that it is almost the Annual Mean Temperature (AMT) of the outside at the same elevation. However, temperature shifts of caves' AMT relative to the external AMT are generally observed; thus, it is meaningful to investigate what characteristics (e.g., geographic conditions, temperature of infiltrating fluids, system geometries) cause cave-surface AMT differences. Therefore, the most interesting questions and most relevant unknowns are related to the variability in space and time of cave atmospheres because of their subtle but constant connection with external climatic conditions (Covington and Perne, 2015; Badino, 2010a; Luetscher and Jeannin, 2004a). The physics of cave atmosphere has attracted different researchers, and the two most recent and complete titles that have dealt with the topic are *"Fisica del clima sotterraneo"* (Badino, 1995) and *"Climatologie du monde souterrain"* (Lismonde, 2002). Badino (2010) and Covington and Perne (2015) have stressed the necessary subdivision of cave climatology and cave meteorology. They consider the first as the study of the average cave atmospheric conditions that vary slowly in time, and the second as the study of how the cave conditions fluctuate around this average condition over relatively short timescales. In the same way, Covington and Perne (2015) highlight that a meteorologist often studies relatively dramatic phenomena; instead, a cave meteorologist may study diurnal or seasonal variations on the order of a few tenths of a degree and humidity variations of a few per cent.

In this work, we will focus mainly on the changes induced by airflow variations in caves and their impact on cave temperature oscillations at daily and annual scales.

1.2.1. Cave ventilation

Through the vadose zone, important air movements are observed and different mechanisms can instigate the airflow within caves (Covington and Perne, 2015; Wigley and Brown, 1976; Cigna, 1968).

Cave airflow and atmosphere variability within caves are primarily driven by temperature and barometric differences with the external atmosphere, although other environmental factors such as variations in water discharge may intervene in changes to the cave atmosphere (Covington and Perne, 2015). The chimney effect, controlled by temperature differences, is considered most prevalent at the scale of a karst massif (Badino, 2010a; Covington and Perne, 2015; Kukuljan et al., 2021a; Luetscher and Jeannin, 2004a) and will be detailed in the next Chapter 1.2.1.1. Here we give an overview of other possible processes causing air movements, such as barometric airflow (Gomell and Pflitsch, 2022; Pflitsch et al., 2010), airflow due to water drag (Palmer, 2007; Cigna, 1968), surface wind-driven flow (e.g. Kukuljan et al., 2021a), airflow due to floodwaters changing the volume of air within the system (Gabrovšek et al., 2018; Pivetta et al., 2021) as well as convective cells (e.g. Qaddah et al., 2022; Guerrier et al., 2019). A combination of these phenomena is also possible.

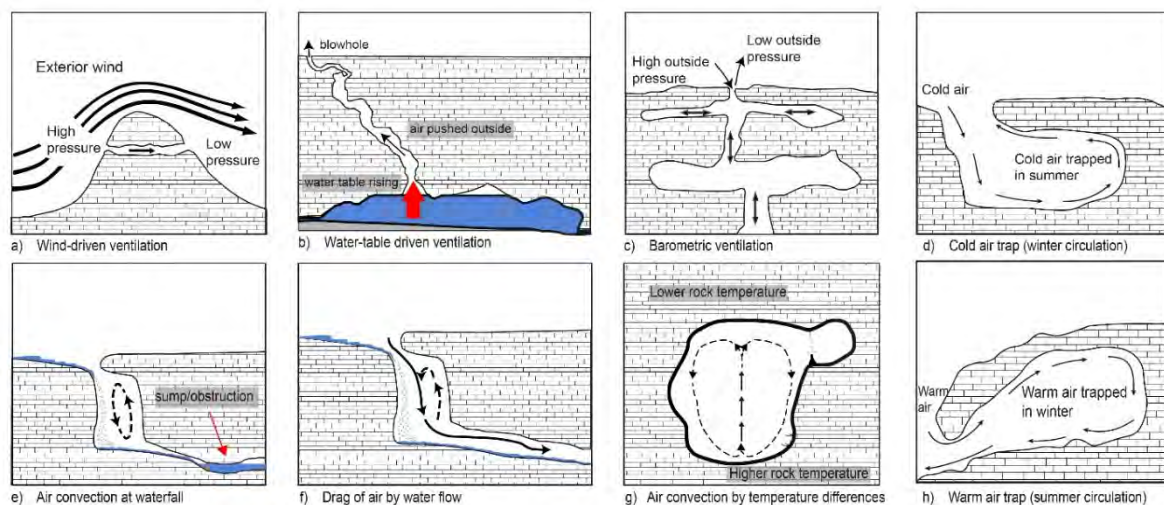


Figure 1.2 – Different processes which can instigate air movements. Black arrows depict airflow. The chimney effect is not included since it is detailed in the next Chapter 1.2.1.1.

External wind can instigate airflow within caves (Figure 1.2.a). Outside wind flowing over a relief causes pressure variation near the ground. If a cave formed through a relief has one entrance exposed on the upwind, this opening will experience a positive pressure, while the entrance exposed on the downwind side will experience a negative pressure. The pressure difference established between the two cave ends generates an airflow (Kukuljan et al., 2021a; Palmer, 2007). This kind of ventilation can sometimes operate together with other mechanisms triggering an airflow (e.g. the chimney effect induced by temperature/density differences between the internal and external atmospheres, cf. next Chapter 1.2.1.1). Kukuljan et al. (2021a) analysed the effects of external wind on Postojna cave ventilation, which is primarily driven by density difference with the exterior (chimney effect), revealing that external winds can increase, decrease, or completely reverse airflow driven solely by the chimney effect. Another study published by Nachshon et al. (2012) has also demonstrated that ground-surface winds can instigate air motion within vadose zone fractures and cracks, favouring gas exchange at the interface between underground and Earth's atmosphere.

During flood events, air masses that fill underground voids are pushed toward the surface by rising groundwater levels (Figure 1.2.b) within the epiphreatic zone (Figure 1.1). This can result in a noticeable airflow, especially if the rising water table pushes the air trapped in large volumes through relatively small passages. If these passages reach the surface as small apertures on the ground, they are known as blowholes (e.g. Gabrovšek et al., 2018). This phenomenon is well known on the Kras/Carso plateau

(Slovenia/Italy), particularly related to the 33 km long Reka-Škocjan-Timavo system (e.g. Pivetta et al., 2021; Gabrovšek et al., 2018; Gabrovšek and Peric, 2006). Except for Škocjan Cave, where the underground river can be followed for several kilometres, most caves in this area that reach the main collector flowing toward the Timavo spring develop vertically. As sumps often block airborne passages, they provide access to only small sections of the river. During flood events, water can rise more than 100 m within the epiphreatic zone, and water squeezes air out from cavities above the base water level, resulting in numerous blowholes on the surface. Most of the caves in this region were discovered through the digging of these narrow blowholes. An example of a vertical cave developing on this plateau (Jama 1 v Kanjaducah) whose airflow is controlled by water table variations is reported in Gabrovšek et al. (2018). A similar phenomenon was studied by Pellet et al. (2024) at Cosquer cave (an Upper Paleolithic decorated cave in southeastern France). This cave is half drowned under the sea level since it develops on coastal karst and is confined by seawater and the surrounding rock. Here, sea level variations due to tides and rough seas provoke air pressure waves, which possibly generate inflow/outflow of the confined cave through karst conduits/fractures inside the surrounding massif.

Water movement can cause local air movements (Figure 1.2.e/f). Air is dragged by water since a shear stress is created between the water flow and the surrounding air. This effect is more pronounced close to waterfalls (Figure 1.2.e), but underground rivers can also generate such air movements. If the cave path is obstructed by, e.g., a sump, the airflow near the stream is counterbalanced by a return circulation of air along the ceiling. However, if the underground stream connects the upper and lower entrances (Figure 1.2.f), it can drive the airflow throughout the cave (Palmer, 2007).

Barometric caves are those caves whose airflow is instigated by external barometric variations (Figure 1.2.c). Air pressure variations in the outer atmosphere are usually transferred to a cave system quite quickly through its openings. However, caves with large underground volumes but small openings limit air exchange with the surface, causing delayed transmission of atmospheric pressure changes. This delay creates pressure gradients that drive airflow: when surface pressure rises, external air flows into the cave, while a drop in surface pressure causes outward airflow. These flows continue until internal and external pressures equalise. Since these air movements are driven by atmospheric pressure variations, barometric airflow responds to the irregular sequences of atmospheric cyclones and anticyclones rather than seasonal patterns (Gomell and Pflitsch, 2022). Barometric airflow occurs in caves with long response times to pressure changes, which depend on and increase with cave volume and decrease with the total cross-sectional area of openings. The number and the position of the openings are not relevant as long as the sum of the cross-sectional areas is sufficiently small in comparison with the cave volume (Gomell and Pflitsch, 2022). Caves that have well-defined barometric patterns are Jewel Cave and Wind Caves in South Dakota (USA), with known lengths of 260.2 km and 339.3 km in 2022 (Gomell and Pflitsch, 2022). In small or medium-sized caves, this phenomenon may also be present, but difficult to measure, since short-term air pressure differences between the outer atmosphere and the cave, as well as fast air pressure changes, are not or hardly measurable in most cave systems. A detailed overview of the processes taking place in barometric caves can be found in (Gomell and Pflitsch, 2022; Badino and Chignola, 2019; Pflitsch et al., 2010; Jernigan and Swift, 2001).

Temperature gradients between the ceiling (colder) and floor (warmer) of a cave can also activate currents related to natural convection (Figure 1.2.g) (e.g. Qaddah et al., 2022; Guerrier et al., 2019). When no other processes causing ventilation intervene, this air movement usually dominates in caves with a single entrance. In caves with small entrances, air exchanges with the outside are often negligible. With inner air humidity near saturation, small temperature differences (on the order of a

few tenths of degrees) between the floor and roof are sufficient to instigate convective cells. However, typical geometries that favour convection cells, allowing air interior-exterior exchanges by the same opening, are those showing large entrances. When the cave slopes downward from the entrance (Figure 1.2.d), these currents are active during cold external temperatures, causing cold, dry air to sink along the floor and warm, moist air to rise and exit along the ceiling. Conversely, if the cave slopes upward from the entrance (Figure 1.2.h), these currents become active during warm external temperatures. In these scenarios, circulating convection works to gradually equalise the temperature difference between the cave's atmosphere and the surrounding rock. When there are air exchanges with the outside, as in the cases shown in (Figure 1.2.d/h), if the outside temperatures are relatively constant, such convective cells will eventually cease when the temperature of the cave rock aligns with the outside temperature. Lack of summer ventilation, or convection cells near entrances, can sometimes create cold air traps (Šebela and Turk, 2018). In sufficiently cold climates, such caves can develop permanent ice deposits, even if average temperatures are above freezing (Luetscher and Jeannin, 2004b). With the cave configuration shown in (Figure 1.2.d), in summer, the cold air in the cave will be trapped underground, being denser than the warmer air outside. Therefore, the external atmosphere will tend to buoy above the inner atmosphere, configuring atmospheric stratification, and summer ventilation will be mainly absent. Indeed, these caves are often known as “cold-trap caves”.

Although all the typologies of airflow can be important for specific cases and detected in many caves, there is a large consensus that the buoyancy-driven airflow, better known as the chimney effect, is the most ubiquitous at the scale of a karst massif (Badino, 2010a; Covington and Perne, 2015; Kukuljan et al., 2021a; Luetscher and Jeannin, 2004a). For well-developed Alpine karst systems, Badino (2010) assesses the minimum specific airflow due to this process to range between 1 and 10 kg/s/km² whereas Luetscher and Jeannin (2004) consider an average rate of between 75 and 150 kg/s/km². This work will focus on caves whose ventilation is instigated by a chimney effect.

1.2.1.1. *The chimney effect*

The chimney effect, or buoyancy effect, is driven by forced convection. It sets in the presence of two or more cave entrances at different elevations when a difference in air density between the underground environment and the external atmosphere produces subterranean airflow between lower and upper entrances (Figure 1.3). Nevertheless, Covington et al. (2013) and Spötl et al. (2005) have highlighted that airflow can occur even if the cave is connected to the atmosphere only through small fractures, without the need for large entrances.

Caves can be seen as an interface between the external atmosphere and the inner atmosphere of the karst massif. These two atmospheres have different physical characteristics, such as temperature, density and pressure, and tend to reach equilibrium. Air density primarily depends on temperature, but it is also influenced by relative humidity, CO₂ content and, to a lesser extent, by barometric pressure (Lismonde, 2002).

Warm air has a lower density than cold air, so a volume of warm air immersed in a cold atmosphere will tend to rise. In winter, the temperature of the air inside a cave is higher than outside; thus, the cave atmosphere acts as a warm volume of air immersed in a colder environment and will rise. As a result, during cold seasons, cold external air enters through lower entrances, while warm cave air exits from upper entrances. To visualise this effect, consider an air column inside the cave of height h and unitary section (Figure 1.3.a). This internal column is lighter than an air column of the same height outside the cave. If we imagine closing the cave's lower entrance with a door (dashed line passing through point A in Figure 1.3.a), the pressure exerted by the air column differs on the two sides. Because the internal air column is lighter, the pressure on the door's inner side is lower than that on the outside (see the two black arrows facing toward point A in Figure 1.3.a). If we open the door, the pressure difference generates a pressure gradient (Driving Pressure, DP) along the cave that, if strong enough to overcome wall friction, induces airflow from the lower entrance toward the upper entrance. Conversely, in summer, the situation reverses (Figure 1.3.b). The external air column becomes warmer and thus lighter than the internal one. This means that the denser cave air tends to sink toward the lower entrance, while warm external air is drawn in through the upper entrances. Namely, the direction of DP reverses (black arrows reversed in Figure 1.3.b), switching the cave airflow accordingly. A quantitative description of the chimney effect and DP is provided in Chapter 1.2.1.2.

Therefore, cave airflow experiences seasonal flow reversal, even though diurnal temperature oscillations may also induce changes in flow direction. During warm periods (e.g., summer or warm daytimes), ventilation is downward, while when external temperature drops below that inside the cave (e.g., winter or cold nights), airflow reverses, and ventilation is upward. The air travelling from one entrance to the other tends to equilibrate with the cave's surrounding rock, which is at a certain temperature, and it does so by exchanging heat. In summer, the air entering from outside is warmer than the cave walls (i.e. in summer), the air cools down, releasing heat to the rock in the upper entrance zone and warming up the cave walls near the upper entrances. In contrast, in winter, the air entering from the lower entrance is colder than the rock, so the air warms up and rises, subtracting heat from the cave walls and lowering the wall temperature near the lower entrances.

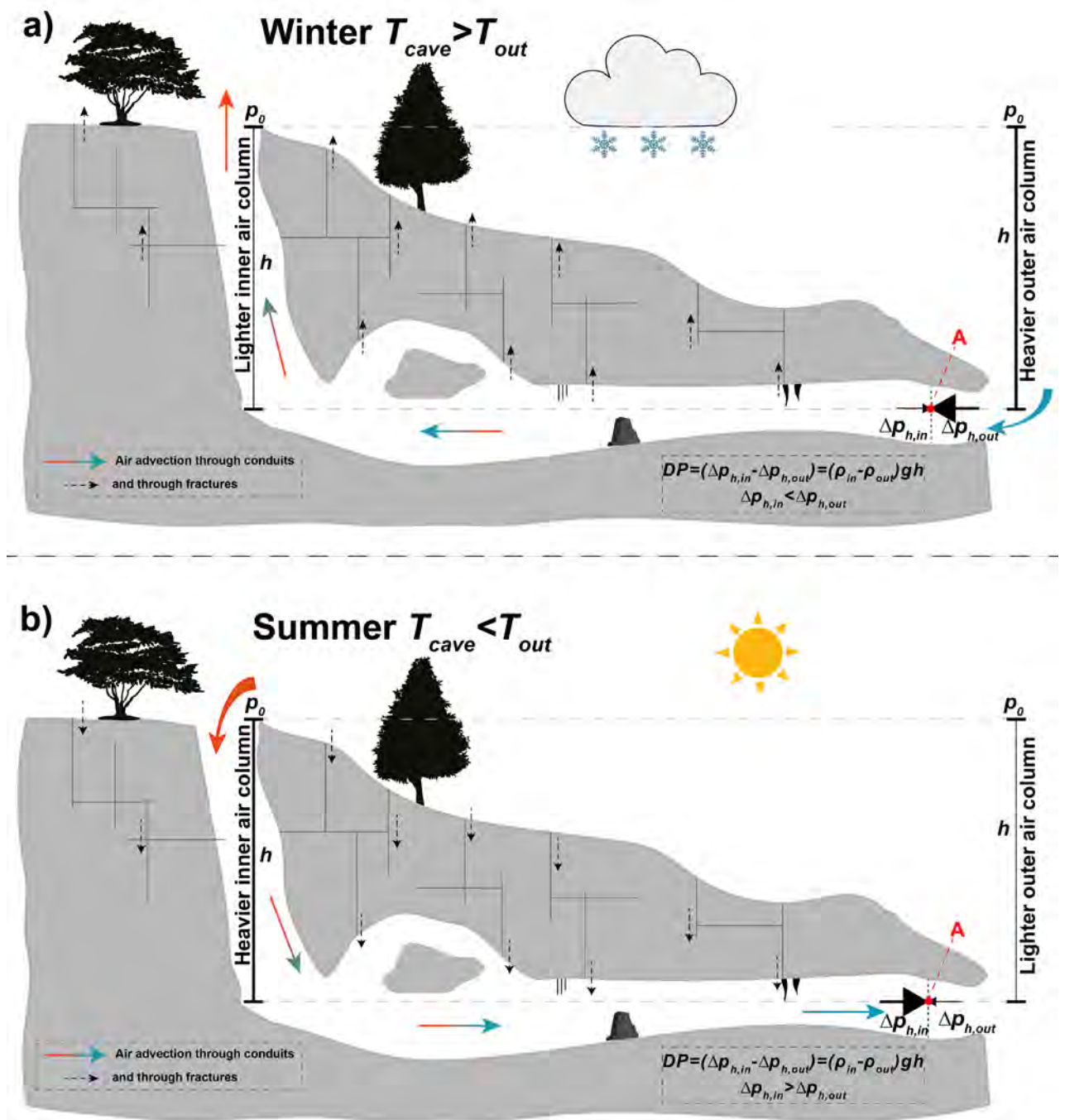


Figure 1.3 – Schematic representation of the chimney effect. The differential pressures ($\Delta p_{h,in}$ and $\Delta p_{h,out}$) between p_0 and point A is determined by the weight of the inner and outer air columns of height h ($g=9.81 \text{ m/s}^2$). The inner and outer columns have different densities (ρ_{in} and ρ_{out} , respectively), due to temperature differences between the internal and external atmospheres. This results in a driving pressure (DP) triggering a chimney effect. A) In winter, the colder external air column of height h is denser and heavier than the internal one, so cold air enters the cave through the lower entrance, displacing the lighter warm air inside the cave, which exits from the upper entrance. Cold air is warmed up to the temperature of the cave on the way to the upper entrance. B) In summer, the inner cave air column is denser and heavier than the external one. The heavier internal air “falls” along the cave’s vertical part and exits through the lower entrance, drawing in warm surface air from the upper entrance. As it moves downward, the incoming warm air gradually cools to the cave temperature. Dry fractures can also be traversed by air and favour air exchanges with the surface.

As a result, upper entrances are always traversed by warm air and lower entrances are always traversed by cold air. These flow reversals, along with the fact that upper and lower entrances are consistently crossed by warm and cold air, respectively, have a significant impact on cave microclimates, particularly on the mean temperature of the entrance zones (Gabrovšek, 2023a; Lismonde, 2002; Wigley and Brown, 1971). This leads to a difference between the external annual mean temperature (AMT_{ext}) and the mean temperature of the entrance zones (Figure 1.4). This difference is negative for lower entrances ($AMT_{ext} > AMT_{Low. entr.}$, Figure 1.4.a) and positive for upper entrances ($AMT_{ext} < AMT_{Up. entr.}$, Figure 1.4.b). Indeed, the air needs a certain distance to reach temperature equilibrium. The length to which the shift from external AMT can penetrate and be observed in a cave depends mainly on the characteristics of the airflow and the cave's average diameter. This length is usually called the thermal or relaxation length; however, in *Chapter 7 – “Modeling heat transfer for assessing the convection length in ventilated caves”*, we will define it as “convection length” since it is due mainly to convection and heat diffusion is negligible.

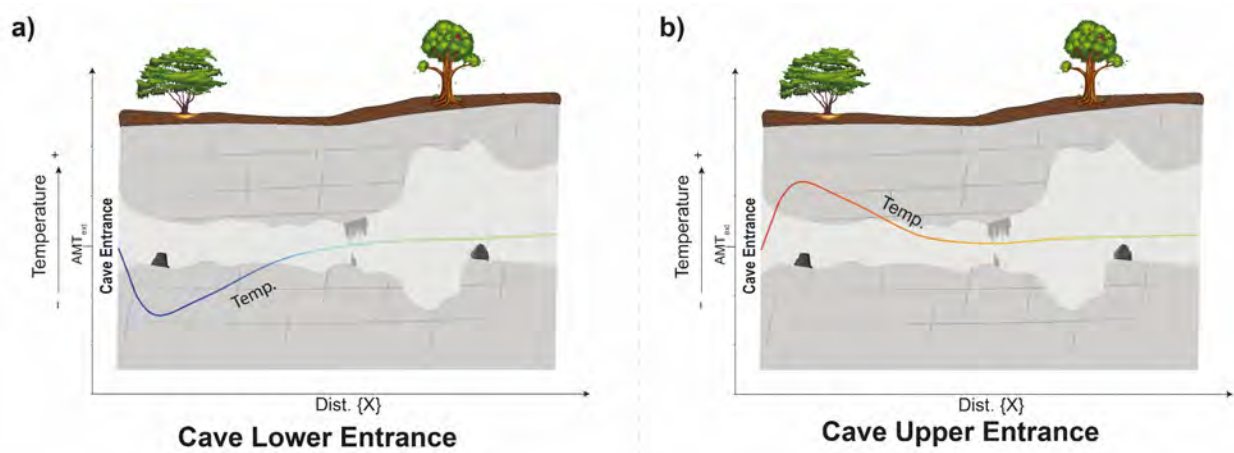


Figure 1.4 – In a), it is shown the temperature profile of a lower cave entrance is shown. In the first part of the lower entrance region, the internal mean temperature is lower than outside AMT_{ext} . In b), the temperature profile along an upper cave entrance. In the first part of the upper entrance region, the internal mean temperature is higher than outside AMT_{ext} . Image modified after Wigley and Brown (1971).

Airflow driven by forced convection can be established in different cave geometries, classified as shown in Figure 1.5 (Covington and Perne, 2015; Lismonde, 2002; Cigna, 1968). L-shaped cave is a classical geometry observed for ventilated caves (e.g. Figure 1.3 and Figure 1.5.a), which develops mainly vertically. Instead, caves that have passages that develop below and above the entrances are respectively defined as U-shaped or reversed U-shaped, then the two entrances show a small difference in altitude. Since the two entrances are at similar altitudes, these caves will exhibit seasonal asymmetry in the airflow, which can be strong in one season (dynamique phase) and very weak or absent in the other (static phase) (Luetscher et al., 2008; Lismonde, 2002). In Figure 1.5.b (U-shaped cave), airflow is stronger in winter because the warm air inside the cave rises through the vertical conduit connecting to the colder external atmosphere, while cold and heavier external air sinks inside the cave. In summer, U-shaped caves can also act as “cold-air traps” (as seen in Figure 1.2.d), since the small altitude difference between entrances results in minimal or absent summer ventilation. An example of this configuration can be found in e.g Ravbar and Kosutnik (2014); Faimon and Lang (2013) and Luetscher et al., (2008). For the same reasons, in a reversed U configuration (Figure 1.5.c), the ventilation will be dominant in summer. Lighter external air entering the cave can rise along the

ascending branch, and colder inner air will fall on the descending side. In the winter season, this cave geometry will be less efficient in ventilation (Covington and Perne, 2015; Lismonde, 2002).

Besides geometries, air composition and characteristics (e.g. seasonal variation in CO₂, moisture content or geothermal heat) can cause air density changes and also result in an asymmetry of the air exchange rate between the cave and surface (Sánchez-Cañete et al., 2013).

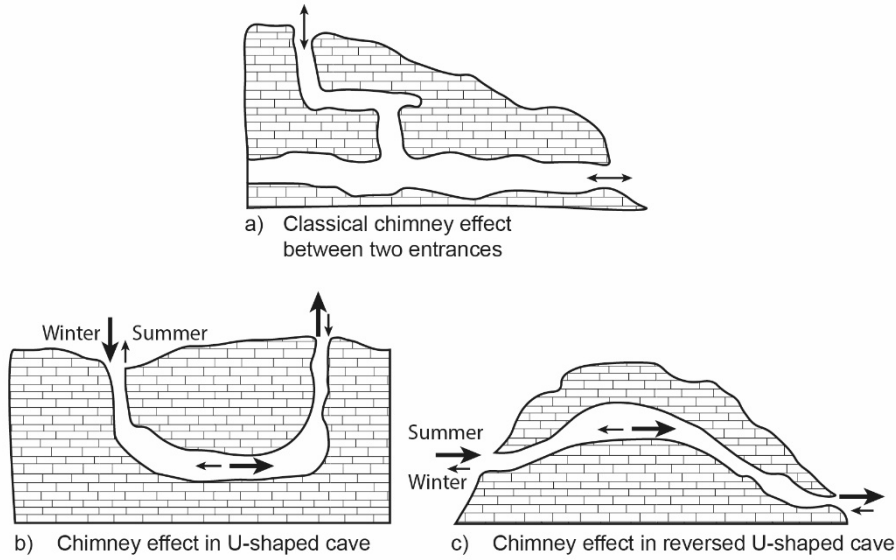


Figure 1.5 – Different settings that produce chimney effect ventilation. Modified after Covington and Perne (2015).

1.2.1.2. Physical mechanisms of a chimney effect: the concepts of the driving pressure and cave aeraulic resistance

The driving pressure DP (also called driving force) refers to the forces that drive airflow within a cave system. It is measured in Pa and it represents the pressure difference that sets a fluid, such as air, in motion within a system, like a cave. In this section, the general concepts of driving force and cave aeraulic resistance, which are necessary to relate air buoyancy, cave geometries and airflow, will be presented (for the sake of consistency of physical quantities, here we use the mass flow rate in kg/s instead of volume flow rate in m³/s).

For ventilated caves under a chimney effect, the main driving pressure is related to the forced convection resulting from density (temperature) differences between the atmosphere in the cave and outside as explained in the previous paragraph (Chapter 1.2.1.1). We can evaluate the pressure at the two sides of a “fictional door” (dashed line passing through point A in Figure 1.3). Specifically, we define the inner differential pressure such as $\Delta p_{h,in}$, and the external differential pressure $\Delta p_{h,out}$.

$\Delta p_{h,in}$ and $\Delta p_{h,out}$ represent the pressure differences between the top of the air columns (p_0) and their bottoms (point A), evaluated separately for the inner and outer air columns of height h (Figure 1.3.a/b). These Δp (in Pa) can be estimated from the following equations (Lismonde, 2002):

$$\Delta p_{in} = \rho_{in} \cdot g \cdot h$$

Eq. 1.1

$$\Delta p_{out} = \rho_{out} \cdot g \cdot h$$

Eq. 1.2

Where ρ_{in} and ρ_{out} are the average air densities of, respectively, the inner and outer air columns (kg/m^3), g is the gravitational acceleration (9.81 m/s^2) and h is the height of the air columns (m). Thus, the resulting driving pressure DP (in Pa), when the “fictional door” in Figure 1.3.a/b is open, is the difference between the two previous terms:

$$DP = (\rho_{in} - \rho_{out}) \cdot g \cdot h$$

Eq. 1.3

In Figure 1.3.a (i.e. winter) $\Delta p_{h,in}$ is smaller than $\Delta p_{h,out}$ therefore the sign of DP will define the upward ventilation, whereas in Figure 1.3.b (i.e. summer) the DP sign will be the opposit. The Eq. 1.3 highlights that DP is proportional to both the difference in density (between the internal and external air) and to the difference in height between the two entrances. In the chimney effect, the difference in elevation between the entrances is the trigger needed to initiate the airflow between the internal and external. In other words, a horizontal conduit ($h=0$), even in the presence of air density differences between inside and outside, is not able to generate airflow due to a chimney effect.

Assuming that the difference in density is due only to temperature differences (i.e. neglecting relative humidity and CO_2 content), DP can be written as (Lismonde, 2002):

$$DP \approx \frac{P_{atm} \cdot g}{R_s \cdot T_{atm}^2} (T_{cave} - T_{atm}) h$$

Eq. 1.4

P_{atm} is the atmospheric pressure (Pa), R_s is the air-specific gas constant ($287 \text{ J/kg}\cdot\text{K}$), T_{atm} is the external temperature ($^{\circ}\text{K}$) and T_{cave} is the cave temperature ($^{\circ}\text{K}$), while h is the difference in elevation between entrances.

The driving pressure DP can also be deduced from the momentum balance applied to the atmosphere inside the cave (Lismonde, 2002):

$$\dot{m} = \pm \sqrt{\frac{DP}{R}}$$

Eq. 1.5

where $\dot{m}$ is the mass flow rate (in kg/s) and R ($\text{kg}^{-1}\text{m}^{-1}$) is the aeraulic resistance (i.e. the term related to the friction along the conduit).

DP and cave aeraulic resistance R are particularly relevant for understanding air movements within caves. R is a physical property of the cave which depends on the friction determined by the system roughness, conduits' diameters and lengths. However, R is not trivial to determine analytically, since system roughness is difficult to quantify.

Theoretically, the aeraulic resistance of a conduit can be inferred from the following relation (Lismonde, 2002; Atkinson et al., 1983):

$$R_T = \frac{8fL}{\pi^2 \rho D^5}$$

Eq. 1.6

Where R_T is the resistance based on the theory, f is the Darcy friction factor, L and D are the length and the average diameter of the conduit, while ρ is the air density in the conduit.

The relation expressed by Eq. 1.6 is a reliable approximation for conduits with a known or predictable friction factor f , typically smooth industrial conduits. Atkinson et al. (1983) attempted the estimation of f in Castelnau Cave using field measurements and found that, despite its 10 km length and wide galleries, the cave's aerodynamic resistance is dominated by just 500 m of crawling passages. They also noted that errors in f arise from the simplified discretisation of the cave (i.e. the number of segments with different diameters composing the main conduits). Since Eq. 1.6 shows that the resistance depends on the average conduit diameter to the fifth power, thus, even small variations in D can significantly affect estimates of f and R_T .

Nevertheless, cave resistance R can be empirically inferred from field data, such as cave temperature, flow rate, and external temperature, thus avoiding the assessment of f . Garagnon et al. (2022) and Luetscher et al. (2008) used field data to extrapolate the aerodynamic resistance R , therefore using it to simulate the airflow by the proportionality given Eq. 1.5 and compare it with the measured airflow. The concepts of DP and R will be applied and extended *Chapter 8 – “Dynamics of chimney-effect ventilation and heat exchange: a model from Longeague Cave”*.

1.2.2. Karst temperatures and thermal distribution

"What's the weather underground?". This question, explored by Badino (2010), highlights the complexity and importance of subterranean climates, especially in karst environments. Although in most geological contexts the temperature increases with depth at an average rate of 2.5–3°C/100 m (Anderson, 2005), the gradient generally observed in karst systems is usually in the order of 0.2 to 0.6°C/100 m, i.e. quite different compared to other geological situations (Luetscher and Jeannin, 2004a). This difference is primarily due to the high permeability of karst. The interior of a karst massif is connected to the external atmosphere through fractures and well-developed conduit networks (caves), allowing meteoric water and air to flow into deeper massif zones. These incoming fluids, initially in thermal equilibrium with external atmospheric conditions, gradually tend to thermally equilibrate with the rock as they move through the massif. This interaction drives heat exchanges between the fluids and the surrounding rock, influencing the subterranean thermal regime. As a result, the average surface temperature exerts a primary control on cave temperatures, making them strongly dependent on the altitude and latitude at which the caves are located.

Within the karst phreatic zone, the geothermal flux, coming from below, is the main source of energy of a karst system (Figure 1.6). However, it is demonstrated that the energy furnished from geothermal heat flux is too low to really warm up the large amount of water that flows through karst systems. It may increase water temperature by some tenth of a degrees during low water conditions. As a consequence, the geothermal heat flux does not reach the vadose zone and is completely drained out by water discharged at springs (Drogue, 1985; Bögli, 1980; Mathey, 1974).

Fluids (water and air) entering the vadose zone from the exterior are in thermal equilibrium with external conditions, i.e. they follow thermal variations at the surface. Energy is transferred from the Earth's surface to the massif through various mechanisms, including heat diffusion through the rock

and the convection of air and water. The heat of water and air is not conservative, so changes in fluid inlet temperature are dampened and retarded as the water and air move through the karst due to heat exchange between fluids and rock. In other words, water and air flowing in contact with each other and with the rock tend to equilibrate their temperature. Within karst, seasonal and short-term fluctuations in temperature might be transmitted over long distances before they are fully damped (Luhmann et al., 2015a). The extent to which subsurface temperature variations are attenuated and out of synch with surface temperature fluctuations depends on the relative significance of these fluxes (Obleitner et al., 2024b), although this relative importance is not yet constrained. The outside atmosphere is thus the primary source of energy (Badino, 2005a) for karst underground environments. In the complexity of the circulation of fluids (water and air) in a karst network, the question “*what’s the weather underground*”, becomes equally complex. And because air and groundwater circulate through these interconnected subsurface channels much more rapidly than in aquifers with intergranular and fracture porosity (De Waele and Gutiérrez, 2022), the influence of atmospheric heat source is more relevant compared to the geothermal heat source (Badino, 2005a; Luetscher and Jeannin, 2004a) in the zone of karst above the phreatic level (Figure 1.1 and Figure 1.6).

The temperature distribution in the subsurface is not only scientifically intriguing but also critically important for various practical applications, including the construction of underground facilities, the utilisation of groundwater resources, and the development and exploitation of geothermal energy (Medici and Rybach, 1995). Since ~20% of exposed continental rocks undergo intense karstic processes, interactions with human activities are common (Ford and Williams, 2007). For instance, during tunnelling along a non-karst bedrock, measured temperature increases with depth, which indicates a compact rock formation. In contrast, encountering a reduction in temperature is often a clear signal of approaching karstified layers. If these are saturated with water, they can be subjected to high pressure (especially if they are located in depth) and potentially high water inrush which can create dangerous situations for workers and facilities (sudden water inrushes of and/or structural collapses due to the discontinuities) (Zini et al., 2015; Badino, 2005a; Luetscher and Jeannin, 2004a). Therefore, a deeper understanding of heat fluxes in karst environments is essential.

In 1969, Pulina introduced the concepts of dynamic and static temperature zones in caves to identify microclimatic regions therein, using temperature fluctuations as a criterion for their distinction. The first zone is represented by a yearly average temperature amplitude $\geq 0.5^{\circ}\text{C}$ and is typical for near-entrance cave parts (usually between 50 and 100 m depth), while the second zone characterises deeper cave parts (usually deeper than 100 m) and is represented by fluctuations $< 0.5^{\circ}\text{C}$. Drogue (1985) adopted this subdivision by renaming the two zones as *heterothermic* (the dynamic zone) and *homothermic* (the static zone). This thermal classification is currently accepted, and based on this subdivision, Luetscher and Jeannin (2004a) developed the conceptual model shown in Figure 1.6 and detailed below.

The heterothermic zone: this zone extends from the karst surface and is estimated to be between 50 and 100 metres deep. In this region, heat fluxes are dominated by advected fluxes of air and water, which fill voids within rock discontinuities, and partially by conductive fluxes through the rock, showing annual temperature fluctuations. Consequently, the temperature vertical gradient in the heterothermic zone can be positive or negative and strictly depends on external variations. For comparison, the depth of temperature variations related to the conduction of heat through the rock (without any advection) is in the order of 7 to 10 meters at a yearly time scale. Due to the high porosity and permeability of the karst, heat is transferred from the ground surface by advection deeper into the massif.

The homothermic zone: this zone is based on temperature stability, then it does not depend on atmospheric temperature oscillations. Badino (1995) demonstrated the major role played by the rock heat capacity in this remarkable stability. Luetscher and Jeannin (2004a), observed thermal gradients in deep caves range between the energy dissipation rate of falling water ($0.234^{\circ}\text{C}/100\text{ m}$) and the adiabatic lapse rate of moist air ($0.5^{\circ}\text{C}/100\text{ m}$). These authors estimate that the energy flux due to air circulation in two caves (Hölloch and La Diau) is 2 to 20 times greater than that due to water, supporting this with evidence that many other observed caves have thermal gradients close to the adiabatic lapse rate of humid air. However, Badino (2010) argues that Luetscher and Jeannin (2004a) overestimate typical air flux and he concludes that water is the dominant factor in most settings. Regardless, observed temperature gradients typically fall between those expected from the dominance of air and water (Covington and Perne, 2015). Furthermore, Luetscher and Jeannin (2004a) observed also that in deep vadose zones ($>400\text{ m}$), the air lapse rate is very often close to that of water. They highlighted that due to the hierarchical structure of karst networks, water tends to concentrate into fewer conduits at greater depths and due to the lower connection with the surface, air circulation entrained by forced convection is progressively reduced. As a result, since water circulation remains significant, its relative effect on heat transfer becomes more pronounced, leading to temperature gradients close to $0.3^{\circ}\text{C}/100\text{ m}$. Within the phreatic zone, included in the homothermic zone, the temperature gradient approaches zero. Below the main conduit system, temperature gradients are predominantly influenced by geothermal heat flux (Figure 1.6).

However, these definitions of *heterothermic* and *homothermic* zones lack a clear physical distinction between them. A more physically-based distinction is provided by the paper Sedaghatkish et al. (2024), which was issued from this project (ThermoKarst Project) and is presented in *Chapter 7 – “Modeling heat transfer for assessing the convection length in ventilated caves”*.

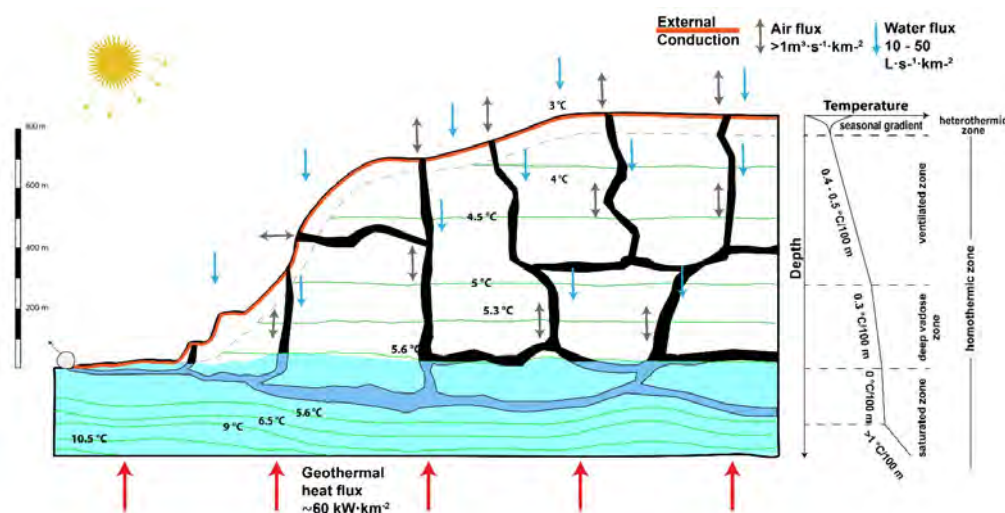


Figure 1.6 – The conceptual model of the temperature distribution in karst presented by Luetscher and Jeannin (2004a), modified.

1.3. State-of-the-art monitoring and modeling of heat and mass transfer in karst

The study of karst temperature distribution and ventilation can be carried out by either models or field monitoring, or preferably both. Making simple models is a way for researchers to explore different complex mechanisms happening in the real world. Although the theoretical physics behind the various mechanisms taking place in karst is quite well described (Sedaghatkish et al., 2024, 2025; Gabrovšek, 2023a; Badino, 2018, 1995; Covington and Perne, 2015; Lismonde, 2002), the applicability of the different available models and their validation with field data is still questionable. In fact, the main

problems stem from the need to couple all processes, but adopting different assumptions and simplifications to reduce calculation time, and the availability of field data to validate a model. For example, some cave heat transfer models included only conduction (Domínguez-Villar et al., 2023a; Salmon et al., 2022; Faimon et al., 2012b; Quindos et al., 1987), others refer only to natural convection in poorly ventilated caves (Guerrier et al., 2019; Qaddah et al., 2023, 2022), and many works focus on water and saturated zone, e.g. Covington et al. (2011) and Luo et al. (2023).

The first steps in modeling ventilated caves were done in 1971 by Wigley and Brown, who pioneered the modeling of airflow and heat flux in karst conduits, using a 1-D model to calculate temperature and moisture profiles based on energy and water mass conservation. They coupled convective and latent heat flux as a function of the distance. These authors also highlighted the presence of thermal anomalies in the entrance zones that extend from the entrance toward the interior of the cave, for a distance equivalent to the so-called relaxation length (or thermal length) of the air temperature, which is the distance at which external temperature fluctuations are completely attenuated along the conduit. Although their model and the further application by De Freitas and Littlejohn (1987) in Glowworm Cave (New Zealand), provided insights into condensation and evaporation processes, assuming convection-limited heat flux, the major model simplification of constant (in space) wall temperature implies that the rock temperature is not influenced by air temperature. Badino (1995) and Lismonde (2002) pointed out that the airflow modifies the temperature of the rock mass, which in turn modifies the air temperature profile along the cave.

Lismonde (2002) developed a model for heat and mass transfer in a straight inclined conduit of constant diameter within a massif. The model coupled the 1-D radial conduction in the rock with convective transfer in the air, using a sinusoidal inlet temperature. The airflow rate was predicted by considering the interaction between the temperature field in the air and the airflow through the buoyancy term in the momentum balance equation of the gas. This model qualitatively reproduced field observations such as flow rate inversions and entrance thermal anomalies. In 2023, Gabrovšek presented a numerical model of a ventilated conduit thermally coupled with the surrounding rock, highlighting the importance of cave passage geometries and cave configurations (e.g. Figure 1.5) in determining airflow patterns. Despite their important contributions, these studies' models were, however, not validated with direct field data.

Cave monitoring is therefore necessary for model validations as well as for direct observation and interpretation of the processes. Continuous monitoring in caves is a quite recent practice thanks to the evolution of technology. In caves, the severe environmental conditions (among which the temperature, condensation and absence of light) and the lack of electricity are the main issues. Speleologists and researchers interested in the climatology of caves had to wait for the technology to step forward to perform consistent cave monitoring. Cigna (2002) makes a review of the state of the art of cave monitoring at that time. He underlined that in the first half of the 20th Century, the study of the physics and mathematics of cave climatology was relatively poor and limited to a few documents (e.g. Bock, 1913; Kyrle, 1923; Crestani and Anelli, 1939). Since the second half of the 20th Century, new studies have been developed and upgraded from qualitative to quantitative (e.g. Cigna, 1968, 1961, 1960, 1958; Wigley, 1967; Wigley and Brown, 1969, 1976). The renewed interest of physicists in the topic led to three of the most complete publications on cave climatology, such as Andrieux (1969), Badino (1995) and Lismonde (2002). The step forward was mainly due to modern technologies such as personal computers and particularly to the appearance on the market of inexpensive and robust data loggers, which can record a great amount of data (Cigna, 2002). Early works contemplating cave monitoring

were already published e.g. in De Freitas et al. (1982) which showed a 12-month cave monitoring proving that the main driving force in Glowworm Cave (New Zealand) is due to air density difference (chimney effect) related to unequal temperatures between internal and external air (already theoretically introduced by e.g. Trombe, 1952, and Cigna, 1968). Other early works about cave monitoring, in particular temperature, were published by Atkinson et al. (1983) and Villar et al. (1984a, 1984b). However, long-term monitoring really started between the late '90s and the beginning of the 21st Century as underlined by Badino (2010). For instance, probably the very first monitoring networks for long-term time series were deployed in Moestroff Cave in Belgium in 1990 (Massen et al., 1997), in Kartchner Caverns (USA) in 1999 and in Cango Caves (South Africa) in 2000 (Cigna, 2002). While the last two caves served as tourist attractions and therefore the availability of electrical supply facilitated the implementation of monitoring networks, for the wild Moestroff Cave was necessary to accurately choose the devices that could be powered by batteries. At that time (1990), the availability of instrumentation was limited and the devices that could be adapted and deployed in a cave were few. Moreover, only 12 V lead acid batteries were available that could feed electric devices for a relatively short time (~2 weeks) before complete depletion. The remote cave access and the weight of 30 kg per battery demanded a great deal of effort from the researchers to keep such a system running for three years. Indeed, many early cave monitoring networks were implemented to measure and record only temperature and humidity for ease of measurement. In fact, the cave monitoring growth was fostered by the miniaturisation of electronic components as well as the improvement of the energy efficiency of batteries. Nonetheless, nowadays most of the data loggers for recording different parameters (e.g. CO₂, O₂, Radon) and not only temperature, have become portable and much more robust, allowing their deployment in remote cave passages.

Another key parameter for cave climatology is the airflow (flow rate and speed), but it has always been challenging to measure (Massen et al., 1997) and was only assessed qualitatively until the late '90s (Cigna, 2002). In the last decades, airflow was mainly measured by using cup (mechanical) or hot-wire anemometers, which, however, had some inefficiency (e.g. mechanical friction in the first case, or lack of direction indication in the second). More recently, ultrasonic anemometers based on the Doppler principle have emerged on the market. These devices have great accuracy and resolution; however, ultrasonic as well as mechanical and hot-wire anemometers, when deployed in underground environments, suffer high degradation of their mechanical and electronic parts, which leads to their rapid uselessness. Due to the high costs of these devices, sometimes researchers have developed their own devices based on the same principles as the commercial ones (e.g. Massen et al., 1997) and others searching for new techniques, as in Borreguero et al. (2009).

Despite these challenges in cave monitoring, significant progress has been made in recent years, largely driven by field studies that have advanced our understanding of underground climatology. As a result, cave monitoring has become increasingly common, boosting numerous studies focusing on the interpretation of cave paleoclimate and climate (e.g. Breitenbach et al., 2015; Hasegawa et al., 2014; Baldini, 2010; Borreguero et al., 2009; Bourges et al., 2006), management and conservation (Addesso et al., 2022a; Bourges et al., 2014; Fernandez-Cortes et al., 2006), but only a few were devoted to investigating and quantifying the energy exchanges within the karst (Obleitner and Spötl, 2011; Luetscher et al., 2008; Ohata et al., 1994a, 1994b).

Luetscher and Jeannin (Luetscher and Jeannin, 2004a) used field data, mostly unpublished, from different caves and deep boreholes to outline a general model of the thermal profile in karst accounting for the presence of the heterothermic and homothermic zones. Luetscher et al. (2008) used data from

monitoring stations in Monlesi Cave (Switzerland) to estimate the energy balance of the system for an annual cycle, and their results demonstrated that forced convection is a driving mechanism for the heat exchange between the cave and the surrounding environment. Notably, in this study, the authors inferred from spot measurements of airflow the cave resistance and the driving force in order to extrapolate a complete airflow time series (e.g. Lismonde, 2002; cf. Chapter 1.2.1.2). The same approach was recently used by Garagnon et al. (2022) in Milandre Cave.

Guidelines on cave and karst protection are reported by the International Union for Conservation of Nature (IUCN; Watson et al.; 1997) and by the international directive “Habitat” n. 92/43/CEE of the European Union Council regarding the protection of cave environments. The latter outlines strategies for the protection and restoration of various habitats and species (both animal and plant) of community interest, as identified in the same document, by establishing a network of protected areas and promoting scientific research (Chiarini et al., 2022; European Commission, 1992).

Postojna Cave is the second longest and the most visited show cave in Slovenia, which also exhibits complex meteorological conditions partially related to the heavy presence of tourists. Therefore, the cave has been the subject of intensive microclimate investigations, among these studies there are e.g. Šebela and Turk (2018), Dominguez-Villar et al. (2015), Šebela et al. (2015), Gregorič et al. (2014a), Šebela and Turk (2014), Gregoric et al. (2013), Šebela et al. (2013), Šebela and Turk (2011). Kukuljan et al. (2021b) used long-term monitoring to interpret the dynamics and distribution of CO₂ in Postojna Cave. They point out that the airflow through the massif is determined by the combined action of the chimney effect and external winds, and that the relationship between the direction of the airflow and the geometry of the airflow pathways explains the observed variations of pCO₂. The same authors, provided a computational model supported by field data to evaluate the relative importance of the chimney effect and external wind influence on Postojna Cave airflow. It is worth also highlighting that Gomell and Pflitsch recently published a work (Gomell and Pflitsch, 2022) based on field monitoring of airflow, pressure and temperatures, shedding light on the aeraulic characteristics of barometric caves.

Monitoring the underground is nowadays recognised as a major tool for the development of research, but it is also seen as a best practice for habitat preservation. The European network Natura 2000 and its directive Habitat (CD n. 92/43/CEE) together with Hemerald Network, specifically its part dedicated to “Terrestrial underground caves, cave systems, passages and waterbodies”, have outlined and planned a comprehensive monitoring of cave environments, with the aim of a European network for the development of best practice in monitoring fragile habitats such as caves. However, Weigand et al. (2022) underline the inconsistency and spreading of the data collected in the different participating countries. It is also underlined the necessity to homogenise the type of collected data and make them comparable, according to the forecasted scope. Cigna (2002) lists the main parameters that should be collected for a complete monitoring of the status of a cave, including temperature, humidity, pCO₂, Radon and airflow.

Dispersion of airborne matter is another key parameter related to air transport (Hart et al., 2013). When a solute is injected in a stream of fluid flowing in a pipe, the interaction between velocity variation over the cross-section and radial diffusion of the solute results in solute dispersion. The solute transport can be described by a 1-D advection-dispersion equation (valid for air and water) in which the dispersion coefficient D (i.e., the effective “diffusion” coefficient in the longitudinal direction) depends on flow characteristics and can be much larger than coefficient related to molecular diffusion. Taylor (1954) derived a theoretical expression of the dispersion coefficient for a turbulent flow in a

rough pipe with a circular cross-section. Taylor's theory states that the dispersion coefficient D is simply proportional to the pipe diameter d and the friction velocity U^* related to the friction in the pipe and flow speed, with $D/d \cdot U^* = 5.05$. This relation is extensively used in mine engineering to infer dust and contaminant transport in time and space along the channels to prevent mine operators' health from the release of mine smoke and fumes (Higgins and Shuttleworth, 1958; Standish, 1988; Thirnonsl and Kissel, 1974; Widiatmojo et al., 2015; Widodo et al., 2008). However, the limits of Taylor's theory have been shown since it underestimates the dispersion in rough and not perfectly circular conduits (Widiatmojo et al., 2015; Widodo et al., 2008). Gauging tests in ventilated conduits, which use gas as a tracer, is a powerful method to infer dispersion and test the applicability of the 1-D advection-dispersion equation in ventilated conduits. However, it was only rarely applied to air dynamics in natural cave systems (e.g. De Freitas et al., 1982), which differ from mines because of their more complex morphologies. Nevertheless, the dust present in mines is analogous to the airborne aerosols present in caves as the latter include external and internal aerosol sources, anthropogenic (visitors) production, hydrological sources and bedrock production (Dredge et al., 2013).

1.4. Main objectives of the work

This thesis is primarily concerned with the monitoring of ventilated caves and the interpretation of field data, but the results of numerical models for which field data have been acquired are also presented. Nevertheless, more details on modeling can be found in the PhD thesis by Sedaghatkish (2025) – *“The Role of Convective Heat and Mass Transfer in the Thermal Response of Karst Conduits”*.

As previously mentioned, significant gaps remain in understanding the effects of air convection on the climatic conditions of ventilated caves. In addition, a systematic comparison with field data is lacking for the validation of thermal and ventilation models.

We are also interested in testing the 1-D advection-dispersion model and Taylor's (1954) theory to assess the dispersion coefficient in the longitudinal direction. Since Taylor's (1954) dispersion theory is derived from artificial pipes, it needs to be tested for dispersion in ventilated natural conduits.

For our project, the main expectations are:

1. to infer the driving force and cave resistance from field measurements and derive the aeraulic model in a ventilated cave;
2. to quantify the heat exchanged at an annual time scale along a cave ventilated by a chimney effect. More specifically, we want to understand where most of this energy is exchanged along the ventilated conduits;
3. to define the concept of convection (or thermal) length, namely the distance from a cave entrance to a location inside the cave where the temperature is influenced by external climatic variations. Field data will be used to correlate conduit geometries and airflow (depending on external fluctuations) with the observed extension of the convection length;
4. to validate by field data and observations several modelled scenarios with varying air and heat-transfer characteristics (e.g. saturated or partially saturated air, finite or infinite heat transfer coefficient) and derive a general thermal model valid for ventilated caves;
5. to test the applicability of Taylor's theory and the 1-D dispersion-advection equation in naturally ventilated conduits;

6. qualitatively extend the concepts of the convection length from ventilated to hydrologically active caves.

To achieve these goals, the following tasks were carried out:

- identifying caves and locations where monitoring systems could be deployed;
- selecting adequate measuring devices and data loggers;
- providing reliable air temperature measurements and related uncertainties;
- testing instruments to measure airflow under harsh cave conditions, validating their functionality, providing robust calibration and estimation of uncertainties related to airflow measurements;
- carrying out gauging tests to infer the dispersion in different naturally ventilated settings.

1.5. Work subdivision

This thesis is paper-based and the four publications are embedded in an outline text, structured as in the following.

- **Chapter 1** is the introduction; it explains the context and the main research questions as well as the structure of the manuscript.

Thereafter, **Part 1 – cave investigations and monitoring** covers monitoring strategies and techniques. It is subdivided as follows:

- **Chapter 2** provides general information on methods and descriptions of study sites;
- **Chapter 3** is a section dedicated to the evaluation of uncertainties in temperature measurements;
- **Chapter 4** is Paper I (first author and published in the International Journal of Speleology), whose title is *“Monitoring air fluxes in caves using digital flow meters”*. It is divided into a technical and a modeling part. The first part deals with testing and calibrating digital airflow meters in caves. The second part presents a modeling section that helps to understand the uncertainties of airflow measurements in caves;
- **Chapter 5** is Paper II (first author and published in the International Journal of Speleology), whose title is *“Dispersion of artificial tracers in ventilated caves”* aims to infer dispersion along naturally ventilated conduits through gauging tests. The results are compared with the dispersion-advection theory and Taylor’s prediction;
- **Chapter 6**, entitled *“Comparing different methods for measuring cave airflow”*, is an extended conference abstract (accepted as an oral presentation) for the 19th International Congress of Speleology (July 20th – 27th 2025, Belo Horizonte, Brazil). This work brings together the results of Chapter 4 and Chapter 5. It aims to compare the different methods to measure and assess cave airflow;

Next, **Part 2** discusses cave ventilation dynamics and heat exchanges within ventilated zones of caves. Models are, therefore, built based on the cave monitoring results illustrated in Part 1. Part 2 is subdivided into:

- **Chapter 7** is Paper III, to which I contributed as the second author. It is entitled “*Modeling heat transfer for assessing the convection length in ventilated caves*” (published in the Journal of Geophysical Research: Earth Surface). It is a modeling paper which aims to give a quantitative definition of the convection length in ventilated conduits (distance from the entrance where internal temperature is no more under the influence of external fluctuations). My work was devoted to the instrumental part and field data management. Field data time series were used to validate the model.
- **Chapter 8** is yet unpublished work entitled “*Dynamics of chimney-effect ventilation and heat exchange: a model from Longeague Cave*”. This chapter presents the main results of three years of monitoring a ventilated cave (Longeague Cave, Switzerland). The main outcomes of this chapter are: 1) a quantitative assessment of heat exchanges related to air advection, 2) the assessment of cave aeraulic resistance through field data and 3) the development of a cave aeraulic model.
- **Chapter 9** is Paper IV, to which I contributed as the second author. It is a modeling paper entitled “*Thermal modeling of caves ventilated by chimney effect*” (published in the International Journal of Thermal Sciences). My work was devoted to the instrumental part and field data management. This study introduces a revised thermal model, based on the framework established in Paper III. The modifications include more realistic assumptions based on field measurements from Longeague Cave, and the model is systematically compared with the field data.
- **Chapter 10** aims to qualitatively extend the concepts of convection length from ventilated to hydrologically active caves. Furthermore, in the general discussion section (Chapter 11.4) a spring classification based on their thermal behaviour is proposed.
- **Chapter 11** is the general discussion section, bringing together the main findings presented in all the papers and focusing on the main questions of the project, followed by a conclusion.

1.6. ThermoKarst Project

This thesis was realised with the support of the Swiss National Science Foundation SNSF through a 4-year project funding. The project was called “*Thermal response of cave and karst systems to climate change*” (short: ThermoKarst, project number n. 200021_188636). The project employed two doctoral students, one for setting up the monitoring in the field and one for developing the modeling part.

The project was completed in June 2024.

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PART 1

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Cave investigations and monitoring

Chapter 2. Methods

2.1. Overview of the monitored caves

In this paragraph, we define the kind of cave settings we have chosen to pursue work objectives (Paragraph 1.4).

Primarily, we wanted to monitor caves ventilated by a chimney effect. The study is expected to bring information about: 1) the physical characteristics of processes controlling ventilation, 2) the impact of ventilation on cave climate and 3) the amount of heat advected by air. The main monitored sites are:

- **Longeaigne Cave**, which lies in Val-de-Travers (Jura mountains, CH) and develops into Jurassic limestone for 1.3 km. It is mainly dry and is crossed by an intense airflow, which complies with the chimney effect due to the presence of two entrances. The elevation difference is 90 m with a distance of c. 300 meters between the two entrances. Since August 2020, we have deployed 9 temperature stations along the cave and one anemometer in the upper entrance conduit. In December 2021, a second anemometer was deployed in the cave's Manivelle passage (station T3). The main results from this cave are shown and discussed in Chapter 8 – “*Dynamics of chimney-effect ventilation and heat exchange: a model from Longeaigne Cave*” as well as the field data from this cave are used for the development of the numerical model presented in Chapter 9 – “*Thermal modeling of caves ventilated by chimney effect*”.

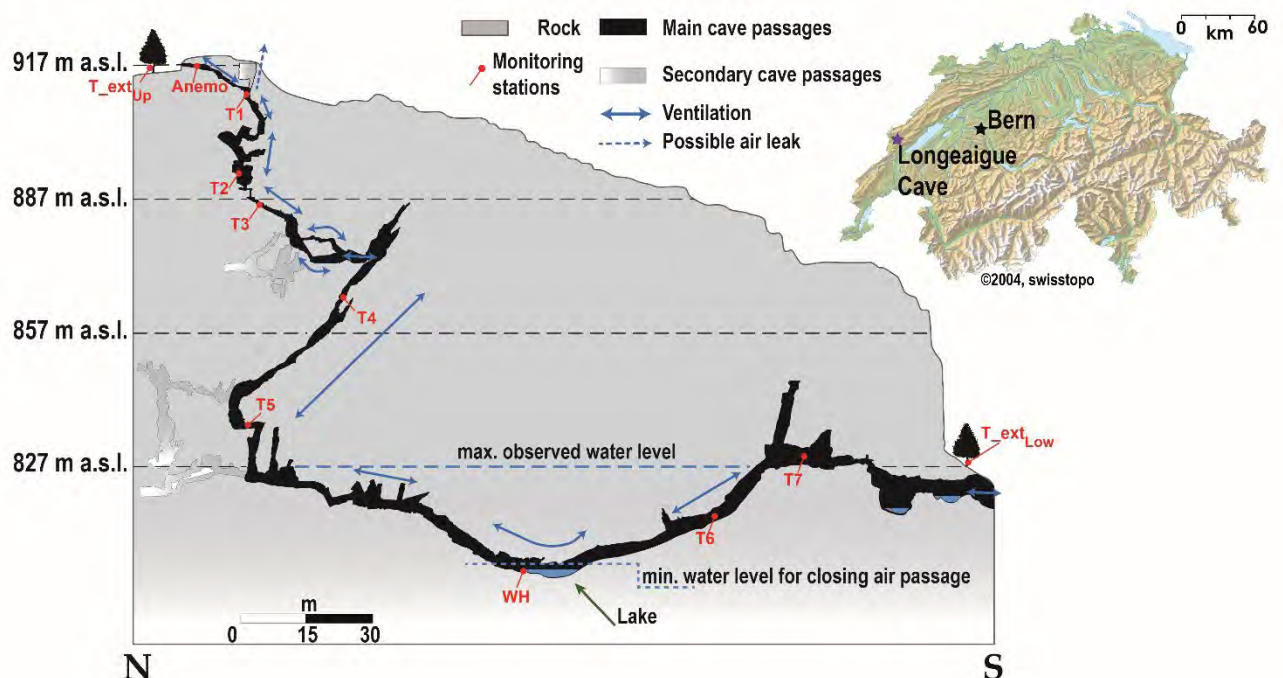


Figure 2.1 – An extract of the Longeaigne Cave topography (modified after Jeannin, 2018).

- **Baulmes mine** is situated in western Switzerland. The lower mine comprises 7 subhorizontal levels with a cumulated length of about 10.7 km. The difference in elevation between the lower and upper entrances reaches 90 m, ensuring a strong chimney effect. The 360 m-long tunnel acts as a lower entrance. Here, we deployed 7 temperature and 1 airflow monitoring stations. Field data from this ventilated tunnel are exploited in Chapter 5 – “*Dispersion of artificial*”

tracers in ventilated caves” and Chapter 7 – “Modeling heat transfer for assessing the convection length in ventilated caves”.

- **D7.1 Cave** is located in the Sieben Hengste karst area (1750 m a.s.l., Switzerland) and is part of a large ventilated cave system with over 160 km of total length, influenced by the chimney effect. Airflow reveals that the cave acts as an upper entrance of the karst system. Field data from this cave are exploited in Chapter 7.

To qualitatively extend the concept of convection length to flowing water, we monitored a cave, which is an active sinkhole. This setting is representative of caves which have an external stream falling into a sinkhole leading to an underground karst system. For this setting, the **Gouffre de la Rouge Eau** cave was chosen, and its results are shown in Chapter 10.

During the project, additional climate data were collected in other ventilated caves, such as the Milandre Cave in Switzerland and Shtares Cave in northern Albania. Although data from these caves will not be analysed and discussed in this thesis, the concepts developed for the other monitored caves are globally compatible with these caves as well.

2.2. Temperature measurements

2.2.1. Hobo and Reefnet probes

During the monitoring campaign, three typologies of thermoprobes were used in the field. These probes had to be reliable at small temperature variations, often close to 0.2°C as an order of magnitude, and resistant to tough environments found in caves.

The used thermo probes are:

- HOBO U23-004 Pro v2; this device is equipped with two temperature sensors, one internal and one wired TMCx-HD sensor. The internal sensor has an accuracy of $\pm 0.25^{\circ}\text{C}$ with a resolution of 0.04°C . The typical response time of 10 minutes in air moving 1 m/s. The external sensor has an accuracy of $\pm 0.21^{\circ}\text{C}$ and a resolution of 0.02°C , with a typical response time of 3 minutes in air moving 1 m/s. The sensor case is made of ASA styrene polymer and consists of screwable parts sealed with o-rings; then, they are waterproof under very low water pressures.
- HOBO U22-001 Water Temperature Pro v2; the device is composed of a watertight case and is equipped with an internal temperature sensor. The declared accuracy is $\pm 0.21^{\circ}\text{C}$, with a resolution of 0.02°C and a typical response time of 12 minutes in air moving at 2 m/s and 5 minutes in water. The case is unscrewable and made of polypropylene. The probe is waterproof to 120 meters.
- ReefNet Sensus-Ultra is a dive recorder for temperature and pressure measurements; it is a robust sensor with a polycarbonate case useful for water pressure and temperature measurements in high-pressure conditions. However, it can also be used for air temperature and pressure. The declared temperature accuracy is $\pm 0.8^{\circ}\text{C}$ and the resolution of 0.01°C , while the pressure sensor has a sensitivity corresponding to 0.3 mH₂O, corresponding to c. ± 30 mBar.

2.2.2. Calibration of temperature probes

To ensure that all thermo probe measurements were comparable to each other, before deploying the devices, the probes were tested and calibrated against a reference Almemo 2290-4 datalogger with a calibrated PT100 (acc. $\pm 0.01^{\circ}\text{C}$). The test was carried out by sinking the probes in a melting ice bath

until reaching the environmental temperature. The test was performed in a 24 L insulated box into which 5 kg of ice and 10 L of distilled water were poured. Agitation of the water was ensured with a magnetic stirrer.

From this calibration procedure, linear correction functions were derived for each probe to adjust their temperature readings. Applying these corrections improves the accuracy of the measurements, ensuring consistency across all sensors. For all sensor types, the coefficient of determination (R^2) of the linear functions was >0.99 .

Therefore, the accuracy of temperature logger types is:

- $\pm 0.10^\circ\text{C}$ (0.02°C resolution) for TMCx-HD external sensor of Hobo U23-004 Pro v2;
- $\pm 0.20^\circ\text{C}$ (0.02°C resolution) for HOBO U23-004 internal sensor and HOBO U22-001 Water Temperature Pro v2 also equipped with an internal sensor;
- $\pm 0.30^\circ\text{C}$ (0.01°C resolution) for ReefNet Sensus-Ultra.

2.3. Airflow measurements with the Sensirion SFM3003-300-CE

The SFM3003-300-CE is a bidirectional mass flowmeter designed for medical and sophisticated gas mass flow measurements. The choice to use the SFM3003-300-CE stems from its compactness and robustness, high resolution (up to 16 bits) and low power requirement (typically 3.3 V to run and ~ 3.8 mA during measurements). This last point is fundamental in a cave, where commonly energy supply is not available if not by using batteries, which in turn should last as long as possible to assure constant data recording. Moreover, the airflow in a cave can be in only two directions according to the chimney effect. The downward and upward cave flow is well recognised by the device, recording the measurements as negative or positive values. The device measures the flow rate through the sensor pipe in Standard Liters Per Minute (SLM), and it needs proper calibration for the conversion into m/s. A detailed description of calibration and airflow measurements with this instrument is given in *Chapter 4 – “Monitoring air fluxes in caves using digital flow metres”*.

In *Chapter 6 – “Comparing different methods for measuring cave airflow”*, we compared different methods to assess cave airflow rates (hand-held devices, fixed devices, gauging tests with CO_2 as a tracer) and to identify weaknesses and strengths of these different approaches. During this project, we also tested other anemometers than SFM device, including an ultrasonic anemometer and a cup anemometer. However, the former proved unreliable under cave’s condensing conditions and was unable to record airflow time series longer than a couple of months. On the other hand, the latter struggled to measure low wind speeds since the airflow in the cave was insufficient to overcome mechanical friction and initiate cup rotation.

2.4. Datalogging systems for integrating the SFM flowmeter

Most commercial solutions for data logging are not suitable for cave environments due to factors such as mud, dust, high air humidity, difficult accessibility and the absence of external power sources. For this project, we needed to build our own dataloggers, used especially for cave airflow measurements.

The dataloggers utilise a TinyPico microcontroller, equipped with a cost-effective yet powerful ESP32 microprocessor. The ESP32 offers superior performance compared to similar microcontrollers like Arduino, thanks to its enhanced memory and higher clock frequency. This microcontroller, appropriately coded, can be used to integrate many sensors for several purposes, e.g. measurements of airflow, CO_2 , temperature and pressure.

We built our dataloggers mainly to integrate the digital flow meter type Sensirion SFM3003-300-CE. These data loggers are powered by a 3.7 V/5500 mAh lithium battery, while the sensor itself requires only 3.3 V to run (~ 3.8 mA during measurements). The battery duration depends on the efficiency of the logging system, but we experienced a battery life of one year with a time step of one hour.

Chapter 3. Effect of wall radiation on cave air temperature monitoring

3.1. Introduction: Which temperature is measured by a sensor?

Air temperature is one of the most important and frequently measured quantities in caves (Peyraube et al., 2025; Cigna, 2002) since it affects thermal energy balance and air movements therein, especially for buoyancy-induced air movements, i.e. chimney effect. Cave air temperature is also a key parameter for cave management and preservation as it influences changes in moisture condensation and water evaporation, potentially altering the cave's original appearance and threatening the resident fauna, which is particularly sensitive to environmental changes (Chiarini et al., 2022). However, as highlighted by Lundström and Mattsson (2020), air temperature measurements in confined environments like caves can be significantly biased, especially under still-air conditions. This is primarily because temperature sensors are subject to radiative heat exchange with surrounding surfaces, such as the cave walls, and in the absence of air movement, the low thermal conductivity of gases hinders effective convective exchange with the sensor, making accurate measurement more challenging. In the presence of air movements, temperature measurement errors can also arise due to the thermal lag of the sensor, which depends on its coating material (e.g., plastic or metal), affecting the sensor's emissivity. Sensor shape can further influence temperature estimation, as heat exchange with the surrounding environment depends on the heat transfer coefficient, which is proportional to air velocity and decreases with sensor size (Lundström and Mattsson, 2020). As pointed out by Lundström and Mattsson (2020), a sensor immersed in a transparent fluid (the air in the present case) receives a convective flux from the fluid and a net radiative flux from the conduit wall. Therefore, its temperature is a weighted average of the air and wall temperatures.

Most of our measurements were carried out from Longeague Cave. The upper vertical part of the cave is dry and narrow, forcing cavers to crawl. In contrast, the cave's lower part is wider and can be flooded during high-water events. These cave characteristics forced us to use water-tight probes and hang them tightly on the conduit walls, a few centimetres from the rock surface. Moreover, these probes have a plastic case with a diameter in the order of 3 cm, increasing probe emissivity. Therefore, wall temperatures may influence air temperature measurements. The following tests aim to clarify the impact of: 1) wall radiations on air temperature measurements and 2) the sensor's size and coating (i.e. sensor type).

3.2. Methods

3.2.1. Radiative experiment

To verify and quantify the error introduced by radiation on sensors with different sheaths (plastic or metal) and at different locations in the cave, we equipped 4 monitoring stations in Longeague Cave. The cave is naturally ventilated by a chimney effect and presents an upper and lower entrance (refer to Figure 2.1 in Chapter 2.1). The monitored stations for these tests are T1 and T2 (cave upper part), as well as T7 (cave lower part). All these stations are in the ventilated cave conduit. Stations T1 and T2 are respectively at 16 (295) m and 42 (265) m from the upper (lower) entrance, whereas T7 is at 20 (294) m from the lower (upper) entrance. The air volume flow rate (m^3/s) and direction (upward or downward ventilation) are measured at Anemo station using a Sensirion SFM-3003 (*cf.* Chapter 2.3 and Chapter 4).

To reduce the error due to the repetibility of the measurements related to different sensors the used thermometers have been calibrated against a reference device (Almemo 2290-4, accuracy $\pm 0.01^\circ\text{C}$; see Chapter 2.2). The used devices are Hobo U23-004 (from now on referred to as U23) with an internal temperature sensor (accuracy $\pm 0.20^\circ\text{C}$) and an external wired temperature sensor (type TMCx, accuracy $\pm 0.10^\circ\text{C}$). From now on we will refer to the U23 internal sensor as U23_{IS}, instead, it will be U23_{TMCx} when referring to the U23 external wired sensor. The U23_{IS} is encased within a polypropylene-made case with a diameter of 30 mm. Instead, the U23_{TMCx} has a metallic cap with a diameter of 5 mm (Figure 3.1). Therefore, the emissivity is higher for U23_{IS} internal sensors than for U23_{TMCx}.

At T7, we deployed two probes type U23 (therefore, two U23_{TMCx} and two U23_{IS}) one next to each other (Figure 3.1). In the first monitoring phase, both TMCx sensors measured in the air. One of the two U23_{TMCx} had the wired sensor shielded into a metallic pipe which allowed the air to pass through and prevented radiation from walls (this sensor will be referred to as U23_{Sh}). Whereas the second U23_{TMCx} sensor was bare (Figure 3.1). In the second monitoring phase, the bare U23_{TMCx} was moved to the wall surface (U23_{rock}) to monitor the rock surface temperature (Figure 3.2). During both monitoring phases, the two U23_{IS} were hung on the wall, close to the rock surface.

At stations T1 and T2, probes U23_{TMCx} have the external sensor measuring as far as possible from the conduit walls (T1 conduit area $\sim 0.90\text{ m}^2$ at T1, T2's $\sim 1.40\text{ m}^2$), whereas U23_{IS} was hung on the wall a few centimetres from the rock surface.

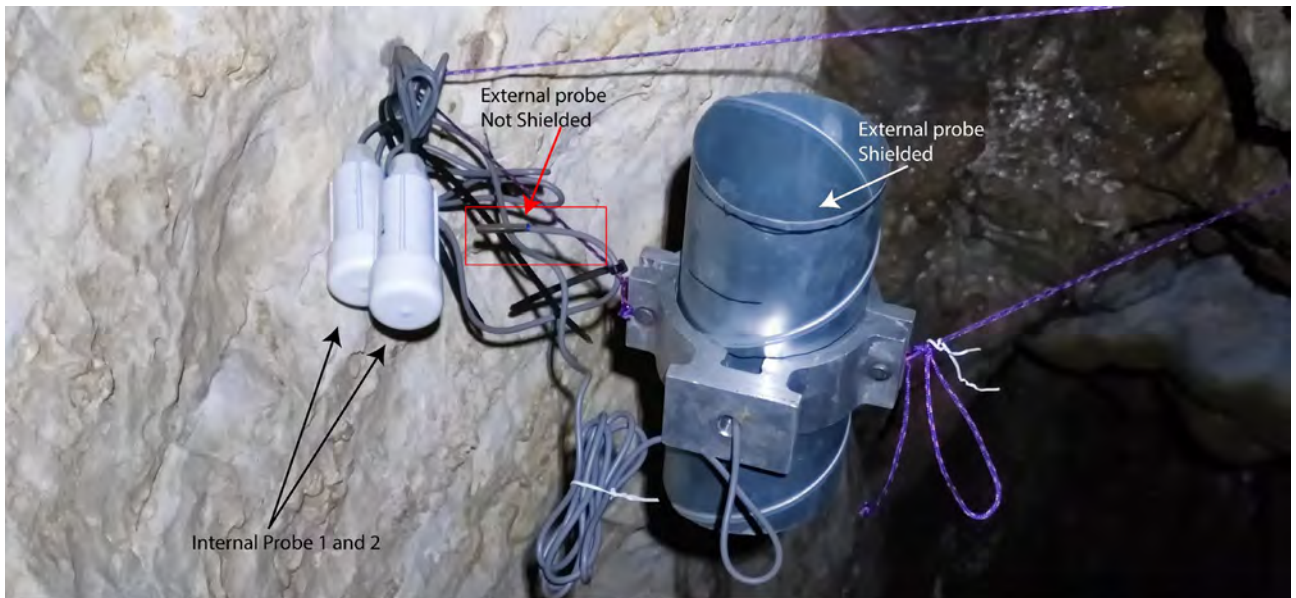


Figure 3.1 – Monitoring setting for the radiative experiment at station T7 in Longeague Cave.



Figure 3.2 – Monitoring setting for the radiative experiment at station T7 in Longeague Cave, with TMCx probes measuring both air and wall surface temperatures.

3.2.1.1. Estimation of the Relative Error of the internal–external temperature difference

Since we are mainly interested in temperature contrasts (ΔT) between external and internal temperatures, we defined the *Rel. Err.* to quantify the deviation of the $U23_{TMCx/IS}$ from a reference $U23_{ref}$ sensor with respect to the temperature contrast between the cave and the external atmosphere T_{ext} .

Therefore, the *Rel. Err.* on the ΔT between the external temperature and cave measurements carried out simultaneously is estimated as:

$$Rel. Err. = \frac{[(T_{U23_{TMCx/IS/rock}} - T_{ext}) - (T_{U23_{ref}} - T_{ext})]}{(T_{U23_{ref}} - T_{ext})} = \frac{T_{U23_{TMCx/IS/rock}} - T_{U23_{ref}}}{T_{U23_{ref}} - T_{ext}}$$

Eq. 3.1

Where T_{ext} is the external temperature, $T_{U23_{TMCx}}$ is the temperature measured with bare $U23_{TMCx}$, $T_{U23_{ref}}$ is the reference air temperature in the cave, measured with $U23_{Sh}$ for *Rel. Err.* due to wall radiation at station T7. For stations, T1 and T2, $U23_{TMCx}$ is used as the reference air temperature. $T_{U23_{IS}}$ is the temperature measured with the $U23_{IS}$ probes hung on the wall (Figure 3.1), T_{rock} is the temperature measured on the wall surface (Figure 3.2).

As stated above, we previously deployed the $U23_{Sh}$ (TMCx shielded sensor at T7) and the bare $U23_{TMCx}$ to estimate relative error (*Rel. Err.*) between these sensors. Here, we assume that $U23_{Sh}$ gives the best estimation of cave air temperature. This will allow us to estimate any influence of radiation on bare $U23_{TMCx}$. If this error is negligible, we can use air temperature measurements carried out with $U23_{TMCx}$ as the reference at T1 and T2 (where no shielded TMCx sensors are deployed). Afterwards, the bare $U23_{TMCx}$ was moved to the wall surface ($U23_{rock}$). Wall surface temperature gives the order of magnitude for the maximum error that can affect $U23_{IS}$, which, in the worst case, could affect the measurement of the wall temperature. Therefore, measurements carried out with $U23_{Sh/TMCx}$ and $U23_{rock}$ are the lower and upper bounds of the *Rel. Err.* in measurements carried out by $U23_{IS}$ hung on the wall.

3.2.2. Infrared images and mesh experiment

To investigate the temperature distribution on cave walls under conditions of strong ventilation, Infrared (IR) images were taken on 17 August 2022 (external temperature $\sim 26^{\circ}\text{C}$). A Flir E40bx infrared camera (accuracy $\pm 2\%$ of reading) was used. At the time, the airflow was downward and strong ($\sim 0.69 \text{ m}^3/\text{s}$), and therefore, the larger impact of warm air on cave walls was expected near the upper entrance. Consequently, IR images were taken along the upper cave conduit (refer to Figure 2.1).

Additionally, to attempt a visualisation of air temperature distribution across the hollow part of a conduit cross-section, we set up an experiment using a wooden mesh. The idea is that the wooden rods composing the mesh are heated differently by incoming external warm air in the radial direction of the conduit (from the conduit centre toward the walls). On the same day (17 August 2022), the wooden mesh was deployed in front of a $\sim 0.5 \text{ m}^2$ cross-section, 10 m far from the upper cave entrance (precise location and experiment set-up are given in Figure 3.6). The mesh had holes large enough to allow air to pass through, while mesh rods (3 cm thick) were impacted by air. The air temperature measured at the mesh location was $\sim 22^{\circ}\text{C}$ (Almemo 2290-4; accuracy $\pm 0.01^{\circ}\text{C}$). IR images were taken approximately half an hour after the mesh was positioned.

3.3. Results

Measurements at T7 (at $\sim 20 \text{ m}$ from the lower entrance) carried out with $U23_{Sh}$, $U23_{TMCx}$ and $U23_{IS}$ were first compared in a period without ventilation (the cave air path was closed by flood events) to test reproducibility between devices. In this condition, cave walls and air should have close temperatures; therefore, with no ventilation, all the sensors should show comparable measurements.

Figure 3.3 shows a time series from station T7 (measurements recorded between 14/12/2022 and 12/02/2023). Figure 3.3.A shows the comparison between $U23_{TMCx}$ and $U23_{Sh}$, while Figure 3.3.B displays the comparison between $U23_{IS}$ and $U23_{Sh}$. The airflow is also depicted in these figures. The ventilation was aborted until 30/01/2023 ($0 \text{ m}^3/\text{s}$), thereafter upward ventilation started (negative airflow values, between -0.6 and $-0.4 \text{ m}^3/\text{s}$). As long as there is no ventilation, the ΔT between the probes is $< 0.1^{\circ}\text{C}$ (blue lines in figures), therefore, all the probes measure very close temperatures. As soon as the ventilation is active, a difference between 0.5°C and 1.2°C appears between the $U23_{IS}$ and $U23_{Sh}$ (Figure 3.3.B). On the other hand, the difference between $U23_{TMCx}$ and $U23_{Sh}$ is as high as 0.2°C (Figure 3.3.A). Therefore, from measurements carried out during non-ventilated periods, it is possible to say that 1) the probes are well calibrated (all probes measured very close values), 2) during ventilated periods the *Rel. Err.* (Eq. 3.1) between the $U23_{Sh}$ and $U23_{TMCx}$ is $< 10\%$. Therefore, the *Rel. Err.* of bare $U23_{TMCx}$, ideally impacted from direct wall radiation, is quite small.

The temperature in the cave is expected to be measured differently by the sensors TMCx and Internal Sensor within the plastic case and the difference is expected to be maximum at high flow rates. It is also expected that the *Rel. Err.* due to wall radiation decreases with distance from the entrances since air and rock temperatures tend to reach close temperatures. The *Rel. Err.* is evaluated for all sensors in the following chapters at stations T7, T2 and T1 during upward (Chapter 3.3.1) and downward (Figure 3.5 in Chapter 3.3.2) ventilation. For stations T2 and T1, where shielded sensors were not deployed, bare TMCx are used as reference air temperature sensors. For monitoring station locations, refer to Figure 2.1.

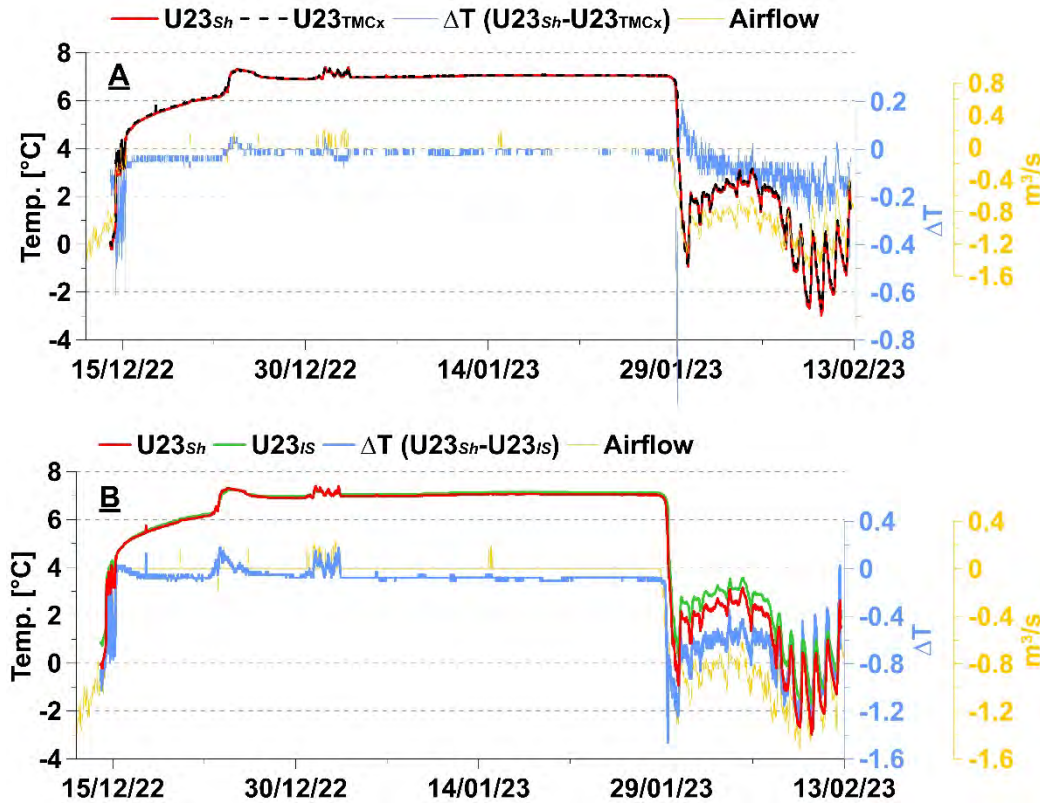


Figure 3.3 – Comparison between sensors at station T7 in Longeigue Cave. Figure A) shows air measurements carried out with U23_{Sh} (shielded sensor) and U23_{TMCx} (bare sensor). Figure B) shows air measurements carried out with U23_{Sh} (shielded sensor) and U23_{Is} (internal sensor hang on the wall surface). The airflow is also depicted (yellow line).

3.3.1. Upward ventilation

Figure 3.4.A/C/E shows the ΔT ($\Delta T = T_{U23} - T_{ext.}$) between the external and internal temperatures during a dominant upward flow, at stations T7 (Figure 3.4.A), T2 (Figure 3.4.C) and T1 (Figure 3.4.E). At T7, internal temperatures were measured with U23_{Sh} (air ref. temp.), U23_{rock} (rock surface temp.) and U23_{Is} (hanging on the wall). At T2 and T1, internal temperatures were measured with U23_{TMCx} (in air) and U23_{Is} (hanging on the wall). The airflow (m^3/s) is measured at Anemo station and depicted in the figures during upward ventilation (negative airflow values) between 02/03/2023 and part of 08/03/2023, and downward ventilation (positive airflow values for the downward ventilation) from the second part of 08/03 to 10/03/2023.

During upward ventilation (02/03/2023 and part of 08/03/2023), the lower entrance is the air inlet. T7 station is 20 m far from the lower entrance and when the upward flow is maximum (negative peaks in Figure 3.4.A), the probe U23_{rock} is constantly the warmest, while U23_{Is} is between the air temperature measured with U23_{Sh} and rock temperature. At Station T2 (which is 265 m far from the lower entrance) all the sensors (U23_{TMCx} and U23_{Is}) measured the same ΔT with external temperatures (Figure 3.4.C). The same results were obtained for station T1 (295 m far from the lower entrance, Figure 3.4.E/F).

The evaluation of the relative error is yielded by Eq. 3.1. Figure 3.4.B shows the *Rel. Err.* of U23_{Is} and U23_{rock} at station T7 (U23_{Sh} as the reference air temperature). The *Rel. Err.* of U23_{rock} gives the order of magnitude of the maximum difference between air and rock temperature, and, ultimately, this is the upper bound of the error for air temperature measurements. During upward flow (air entering from the cave lower entrance at c. 20 m from T7) the air temperature measurements provided by U23_{Is} hanging on the wall surface (Figure 3.2 and Figure 3.4.B) can produce a *Rel. Err.* as high as 30%, while

the *Rel. Err.* of measurements directly on the rock surface is about 60%. For T2 and T1, during upward flow, the *Rel. Err.* between measurements carried out with $U23_{TMCx}$ and $U23_{IS}$ is close to 0%. Therefore, for the latter two stations, there is no difference between measurements carried out with sensors $U23_{TMCx}$ in the air and $U23_{IS}$ hanging on the wall.

Rel. Err. higher than 100% happens when the airflow and temperature suddenly change. This error is likely due to the thermal lag of the $U23_{IS}$ sensor.

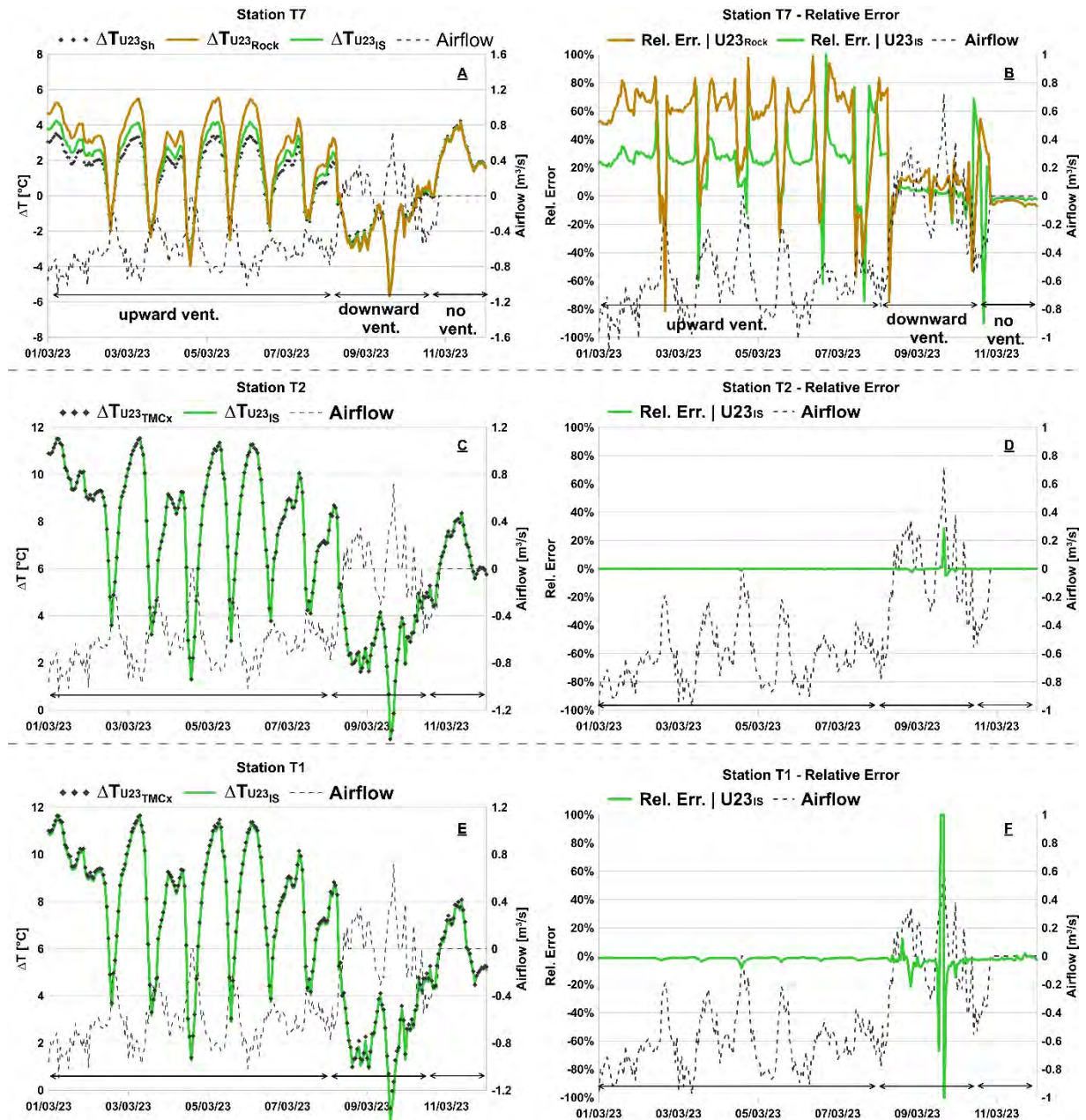


Figure 3.4 – A and B, comparison between sensors at station T7 (lower cave part) during an upward and downward flow. In C, D, E and F, a comparison between sensors at T2 (Figures C and D) and T1 (Figures E and F), which are deployed in the cave's upper part.

3.3.2. Downward ventilation

When the airflow turns downward, the upper entrance is the air inlet. At station T7 (294 m far from the upper entrance), in the period 08/03 to 10/03/2023 (Figure 3.4.A), the ΔT s between external temperature and temperatures measured with $U23_{Sh}$, $U23_{IS}$ and $U23_{rock}$ show very small deviation between them (i.e. air and rock have close temperatures since all the sensors measured the same ΔT between air/rock and external temperature). The *Rel. Err.* is <10% for $U23_{IS}$ and <20% for $U23_{rock}$ (Figure 3.4.B).

The data from the upper part of the cave (T1 and T2 stations, at 16 and 42 m from the upper entrance, respectively) were analysed, focusing on the period from 11/07/2023 to 21/07/2023 during a persistent downward flow (Figure 3.5). At these stations, the comparison is between $U23_{TMCx}$ (air reference temperature) and $U23_{IS}$ (hanging on the wall). No probe was measuring directly the rock surface in the cave's upper part.

Figure 3.5.A shows the ΔT s ($\Delta T = T_{U23} - T_{ext.}$) estimated with both $U23_{IS}$ and $U23_{TMCx}$ at station T1. The deviation between ΔT s is maximum during positive peaks (maximum airflow), while it reduces at lower airflow values. The *Rel. Err.* estimated for $U23_{IS}$ is between 20% and 30% on average (Figure 3.5.B), of the same order of magnitude as the lower cave part (T7) during upward flow.

Whereas, at T2 (Figure 3.5.C), the deviation between ΔT estimated with $U23_{IS}$ and $U23_{TMCx}$ is negligible and the *Rel. Err.* reduces to less than 5% (Figure 3.5.D).

Again, the *Rel. Err.* higher than 100% is likely due to the $U23_{IS}$'s thermal inertia to sudden changes in surrounding conditions (airflow and temperature).

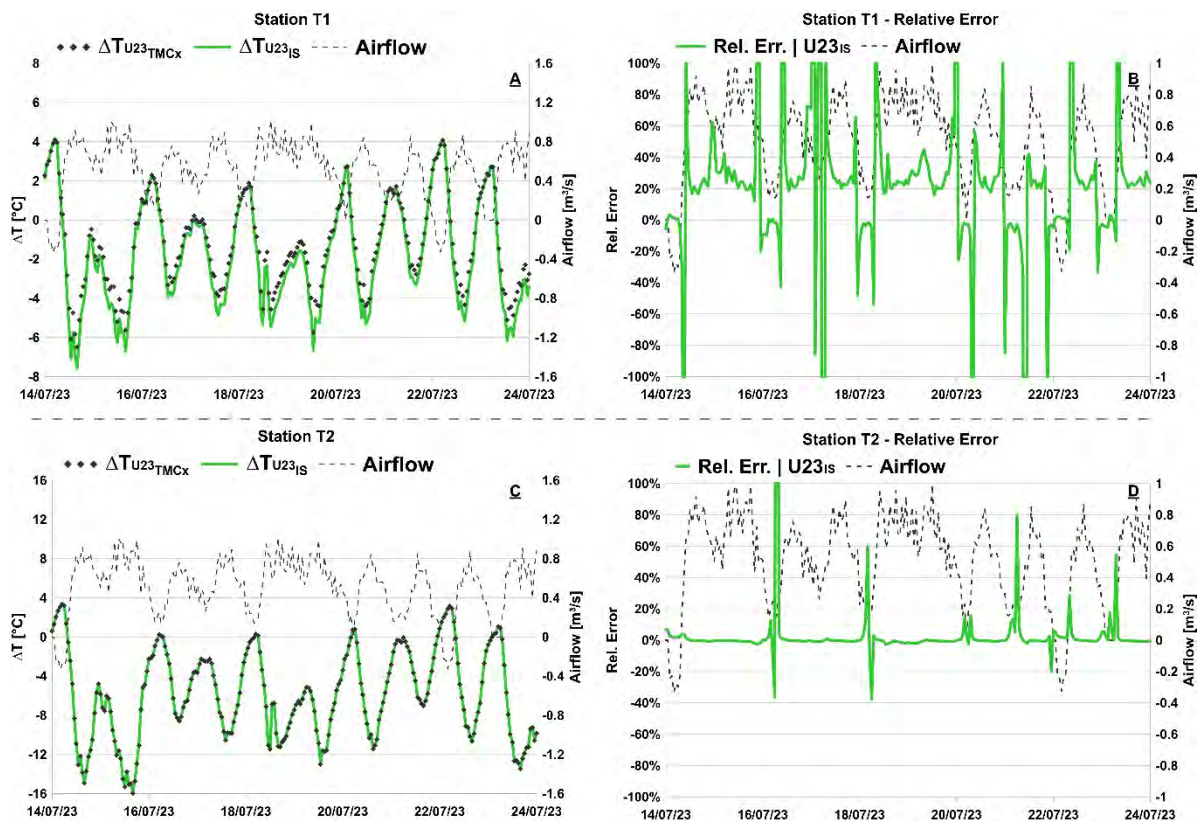


Figure 3.5 – Comparison between sensors at T1 (Figures A and B) and T2 (Figures C and D) deployed in the cave's upper part during a downward flow.

3.3.3. IR images and mesh experiment

To support previous data and visualise the temperature distribution on cave walls and within the hollow part of the cross-section, Infrared (IR) images were captured along the upper entrance conduit (Figure 3.6) during a period of strong downward airflow (summer regime). IR images in Figure 3.6.c have revealed distinct wall temperature variations on the wall approximately 10 m from the upper entrance (ext. temp. $\sim 26^{\circ}\text{C}$; *in situ* air temperature $\sim 22.7^{\circ}\text{C}$; downward airflow $\sim 0.69\text{ m}^3/\text{s}$). IR temperatures of rock (Figure 3.6.c) ranged between 20 and 16°C . Notably, rock chocks exposed to the airflow have shown higher temperatures than recessed zones (Figure 3.6.c).

Figure 3.6.d shows an IR image of the wooden mesh exposed to the airflow. The mesh temperature ranged between 20°C and 22°C , consistent with the air temperature of the side ($\sim 22.7^{\circ}\text{C}$). The mesh temperature was asymmetric concerning its centre, with the warmest area ($\sim 22.7^{\circ}\text{C}$) concentrated on the right side. This asymmetry may relate to the conduit immediately upstream of the mesh (Figure 3.6.b), which could have forced the airflow toward the right side of both the conduit and mesh. Indeed, due to irregularities such as bumps, hitches, and bends, air exchanges heat with surfaces differently depending on their orientation and exposure to the flow.

Rock temperatures were more homogeneous at station T2 (Figure 3.6.g/h), located 42 m from the upper entrance. This image was taken on 17 June 2022, approximately two months before the one shown in Figure 3.6.c-e, when the external temperature was about 29°C . IR images show a rock temperature ranging between 9.3°C (cold zone on the left side of Figure 3.6.g) and 10.8°C . At this location, the sensor T2 displays a slightly higher IR temperature ($\sim 11.4^{\circ}\text{C}$), while the actual measured (by the sensor) temperature was $\sim 13.5^{\circ}\text{C}$, consistent with the air temperature measured at $\sim 13.2^{\circ}\text{C}$.

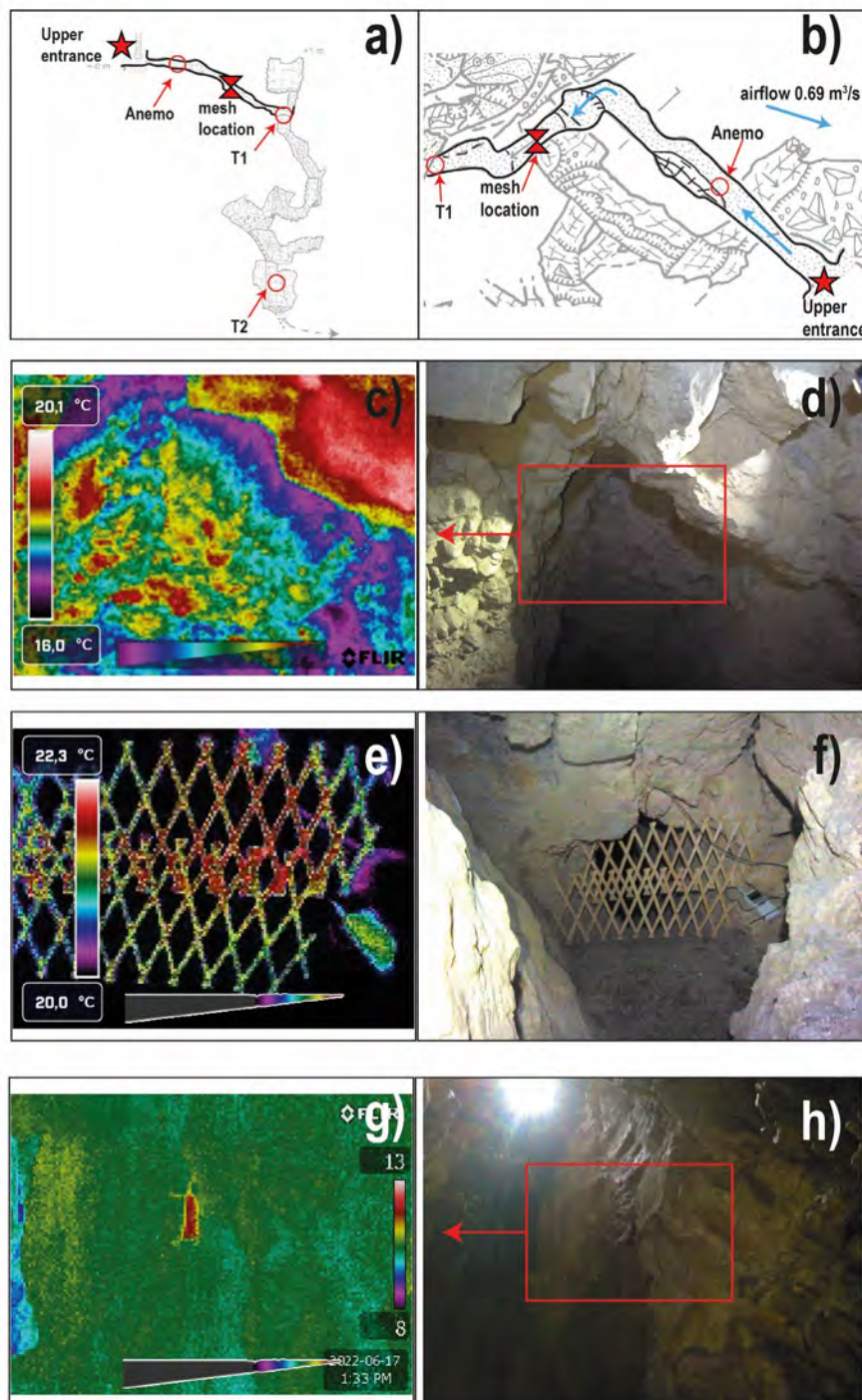


Figure 3.6 – A and B show a shred of Longeigue's survey (section and plan, respectively) displaying the position of the wood mesh during the experiment and of stations T1 and T2. C – D), IR and standard images, respectively, of the ventilated conduit (~10 m from the upper cave entrance) during downward warm ventilation (ext. temp. ~26°C, in situ air temp. ~22°C, vent. ~0.69 m³/s). E – F), IR and standard images, respectively, of the wooden mesh experiment. G – H), IR and standard images, respectively, of station T2 (~42 m from the cave upper entrance).

3.4. Discussion

By measurements carried out at T7 where the wall surface and air temperatures were measured, it is highlighted that the cave wall and air temperature close to the lower entrance (in the region 20 m from the inlet) present a ΔT between air and rock that can be as high as 60% during upward (winter) ventilation. While, at the same station and during the same ventilation, air measurements taken with probes in a plastic case and hanging on the wall (U23_{IS}) are affected by a *Rel. Err.* (against the reference probe U23_{SH}) in the order of 30% to 20% (Figure 3.4.A/B). In contrast, simultaneous measurements at T2 and T1 (at 265 and 295 m from the lower entrance, respectively), U23_{IS} placed showed a <5% *Rel. Err.* (with the reference probe U23_{TMCX}, Figure 3.4.C to F).

These results are consistent with those retrieved during the downward (summer) ventilation. At station T1 (16 m from the upper cave entrance), the measurements carried out with U23_{IS} hanging on the wall showed a *Rel. Err.* in the order of 20 to 30% (Figure 3.5.A/B). While at station T2 (42 m from the upper cave entrance), the *Rel. Err.* reduced to lower than 5% (Figure 3.5.C/D).

IR images confirm these results (Figure 3.6). In fact, rock temperatures have shown inhomogeneous distribution along the first tens of meters of the cave's upper entrance conduit during the downward regime. IR images also highlight that the rock exposed to the flow is sensibly warmer than in the recesses. Also, conduit bends can redirect flow on one conduit side, affecting surface temperatures, as suggested by the experiment made with the wooden mesh (Figure 3.6.g/h). On the other hand, the walls surrounding station T2 have shown a more homogeneous temperature. Applying Eq. 3.1 (*Rel.Err.*) to IR rock temperature (compared with the air reference temperature measured with the Almemo thermometer), this yields a difference between 97% and 40%. Although the absolute temperatures shown by a thermocamera can be affected by several factors such as emissivity variations, reflective surfaces, viewing angle effects (Sass et al., 2023; Yang et al., 2024) and instrumental error (in this case 2% of the measured values), these results suggest that rock exposed to the flow show a lower difference in temperature with airflow (40%) than in area where the heat exchanges are less efficient (97% e.g. in wall recesses). In contrast, at station T2, where the wall temperature is more homogeneous, Eq. 3.1 yields values between 20% and 30%.

Along the first ~20/30 m from the inlet (according to the ventilation direction), both measured temperature and IR images suggest that: 1) air and wall surface temperature can be sensibly different (from 50% to close 100%) and 2) wall temperatures show inhomogeneous temperature distribution. This fact can affect the measurements carried out with probes hanging on the wall and with a plastic case (U23_{IS}), especially in the entrance zones. Indeed, these sensors (T1 and T7 during downward and upward ventilation, respectively) measured a temperature between the wall surface and air temperatures (Figure 3.4.A). The temperature measurements carried out with U23_{IS} sensors can be mainly affected by two factors: 1) the convection acting on the probe and depending on its shape and size, and 2) the thermal inertia of the plastic case U23_{IS} (Lundström and Mattsson, 2020). Moreover, as pointed out by Lundström and Mattsson (2020), radiation from surrounding surfaces impacts sensor temperature. Unless the temperature of the surfaces around the sensor equals the air temperature, the sensor temperature is the average of the air and wall temperatures. Therefore, close to entrances, since wall and air temperatures are very different (about 60% at T7, and up to 97% at the upper entrance conduit), the wall effect radiation on the sensor within a plastic case can play a key role in controlling sensor temperature.

The sensor temperature can be derived from a steady-state energy balance at the sensor surface, assuming heat exchange between the sensor and the air by convection, and with the cave walls by

radiation (Lundström and Mattsson, 2020). The convective heat transfer can be described using Newton's law of cooling, while radiative heat exchange can be expressed through a linearised form of the Stefan–Boltzmann law (valid for small temperature differences). Under these assumptions, the sensor temperature can be expressed as a linear combination of air and wall temperatures, weighted by the corresponding convective and radiative heat transfer coefficients (Lundström and Mattsson, 2020). The model based on energy conservation yields the following equation:

$$T_{sensor} = T_{air} \frac{h_{conv}}{h_{conv} + h_{rad}} + T_{wall} \frac{h_{rad}}{h_{conv} + h_{rad}}$$

Eq. 3.2

where h_{conv} and h_{rad} are the convective and radiative transfer coefficients at the sensor surface. The heat transfer coefficient indicates the efficiency of the heat exchange by radiation or convection.

A sensor mainly measures air temperature when convection is higher than radiation (i.e., when $h_{conv} \gg h_{rad}$). If not, it records a temperature between the air and wall temperatures, or even the wall temperature if radiation prevails ($h_{conv} < h_{rad}$). The convective heat transfer coefficient (h_{conv}) increases with airspeed and decreases as the sensor size increases. The radiative coefficient (h_{rad}) depends on the sensor's surface emissivity. The U23_{IS} sensor is larger than the U23_{TMCx} (30 mm vs 5 mm) and has higher emissivity due to its plastic (polypropylene) casing, compared to the metallic cap of the U23_{TMCx}. Moreover, U23_{IS} sensors were hung on the wall. Here, air velocity is much lower than in the centre of the conduit, thus air temperature is closer to the wall temperature. Because of this, the U23_{IS} sensor tends to give values in between the air and wall temperatures, while the smaller U23_{TMCx}, located more in the centre of the conduit, provides a better estimate of the air temperature.

For higher distances from the air entrance (>30 m), the error related to wall radiation is ≤5%. Here, air temperature carried out with U23_{IS} and U23_{TMCx} matches closely between them since air and wall temperatures are close to each other.

3.5. Conclusions

To summarise, air temperature measurements can be challenging to carry out accurately when one is close to the cave entrances. We have shown that:

- there is no difference between carrying out air temperature measurements with or without a shield around a metallic, small and low-in-emissivity sensor (U23_{Sh} vs bare U23_{TMCx} at T7, Figure 3.3). Therefore, the radiation on small sensors with a metallic cup can be neglected for cave monitoring;
- the roughness of the wall leads to a different heating of the wall surface by the airflow. This fact can introduce an additional uncertainty to temperature measurements, especially close to the entrances;
- air temperature measurements carried out with probes hanging on the wall and/or with a plastic case can show a *Rel. Err.* up to 30% at least in the first tens of meters from the air entrances. This error is mostly related to the vicinity of the wall surface, the size of the instrument and the plastic case, which has high emissivity (Lundström and Mattsson, 2020);
- far away from cave entrances (>30 m), the *Rel. Err.* between U23_{IS} and U23_{TMCx} is as low as 5%.

In the following thesis work, we will refer to these *Rel. Err.s* for air temperature measurements, according to the station and sensor used.

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Chapter 4. Monitoring air fluxes in caves using digital flow metres (published paper)

Pastore, C., Sedaghatkish, A., Weber, E., Schmid, N., Jeannin, P.-Y., Luetscher, M., 2024. Monitoring air fluxes in caves using digital flow meters. *International Journal of Speleology*, 53(1), 63-73. <https://doi.org/10.5038/1827-806X.53.1.2500>

Abstract

Precise measurements of airflow within caves are increasingly demanded to assess heat and mass transfers and their impacts on the karst environment, including subsurface ecosystems, hydrochemistry of karst water and secondary mineral precipitates. In this study, we introduce a new, low-cost and lightweight device adapted to monitoring air fluxes in caves, which addresses the need for reliable measurements, low power consumption, durability and affordability. The device was calibrated in a wind tunnel, showing the high accuracy and precision of the device. Field-related uncertainties were further investigated in a ventilated cave to determine the effect of local airflow conditions on the inferred mass flux. Comparing measured values with a 3-D air velocity distribution modelled on a surveyed cave section suggests that most of the uncertainties in estimating the airflow result from the relative position of the instrument in the streamlines rather than from the accuracy of the device.

4.1. Introduction

Accurate measurements of air fluxes in caves are increasingly demanded to quantify heat and mass transfers for their implication on the karst water hydrochemistry (Jeannin et al., 2017) and secondary calcite precipitates (Spötl et al., 2005) as well as for assessing their impact on subsurface ecosystems (Dominguez-Moñino et al., 2021). Because airflow largely controls the cave climate (Luetscher et al., 2008; Luetscher and Jeannin, 2004), quantitative measurements are needed for paleoclimate reconstruction, for assessing habitat evolution (Medina et al., 2023) of numerous specific and often endemic species, and even for carbon cycle estimates (Martin-Pozas et al., 2022; Prelovšek et al., 2018; Breecker et al., 2012), as they control CO₂ concentrations (Kukuljan et al., 2021; Faimon and Lang, 2013; Kowalczyk and Froelich, 2010) in the whole volume of carbonate rocks. Some authors relied on airflow monitoring to assess the air energy balance in ventilated caves and its impact on underground ice masses (Luetscher et al., 2008).

Different types of anemometers are available on the market, including 1) analogue anemometers, 2) hotwire anemometers and 3) ultrasonic anemometers, offering a wide range of costs and quality of measurements. While each of these instruments is suited to specific monitoring ranges, they either have limited resolution and mechanical inertia (1), lack indication of the airflow direction (2) or require a high power supply (2, 3) (Fan et al., 2018; George et al., 2018), although the available technology is constantly improving (Wang et al., 2021; Fan et al., 2018; Leoni et al., 2018). Moreover, cave air drafts could be associated with low speeds (<1 m/s), in particular when the cross-section is large or with a small driving force (e.g. reduced seasonality, small vertical development and/or natural convection cells) (Gabrovšek, 2023; Kukuljan et al., 2021). There, measurements can be affected by instrumental precision and would require high-resolution monitoring with e.g. ultrasonic devices or hot-wire anemometers, which ensure high precision and accuracy. However, these instruments are sensitive to cave-specific environmental conditions and their prohibitive costs make them difficult to install routinely in remote cave parts. While most of these instruments are designed to be installed outdoors, where climatic conditions are very variable (alternating sunny and rainy days, varying RH and

temperatures), others are designed specifically for indoor applications under steady conditions (e.g. in ventilation ducts or for medical purposes). In cave conduits, climatic conditions are less variable than outside but the condensing environment nonetheless represents an extreme environment for these devices. These climatic conditions can alter the integrity of the collected data during monitoring campaigns, whilst extra heating caps require an additional power supply, possibly influencing the cave climate locally.

We have tested several devices in the past, including a highly sensitive ultrasonic anemometer. However, during long-term monitoring, most of the instruments did not work properly for several reasons: condensing environment, failure of electronic parts, drifting and unreliable measurements.

Here we test a new, low-cost and lightweight device for the automated monitoring of air fluxes in caves. This analysis aims to find out:

- the suitability of the instrument outside the manufacturer's declared working conditions;
- the uncertainties over the range of potential velocities found in caves;
- a reliable conversion curve between the raw standard litre per minute (SLM) measured by the device into a velocity in metres per second (m/s).

Eventually, cave airflow (from now on referred to as volumetric airflow in m^3/s) is derived from the integration of air velocities through a known cross-section. The inferred airflow through a cross-section is, however, subject to uncertainties related to singularities along a cave conduit (e.g. rock chocks and asperities) which can modify the velocity distribution within the conduit (Pastore et al., 2024) and cross-section. This raises the question of the best location to deploy a device to infer a close-to-average airspeed. To address this point, we carried out a numerical simulation of air velocity distributions based on a 3-D lidar scan of a cave conduit. Results were compared with experimental measurements carried out with the digital flowmeter.

4.2. Materials and Methods

4.2.1. Sensirion SFM3003-300-CE

Originally developed for measuring airflow in a closed system for medical applications, the Sensirion SFM3003-300-CE (from now on SFM) is a digital bidirectional mass flowmeter ready to be implemented on a 3.3 V microcontroller based on an I2C protocol. The device is compact and robust, shows a high data resolution (up to 16 bit) and has low power requirements (typically 3.3 V / ~ 3.8 mA during measurements). This last point is fundamental in caves, where electrical energy is often limited to batteries, which should last as long as possible to ensure continuous data recording. Furthermore, the bidirectional cave airflow is well recognised by the device, which records the flow direction as negative or positive values (Figure 4.1).

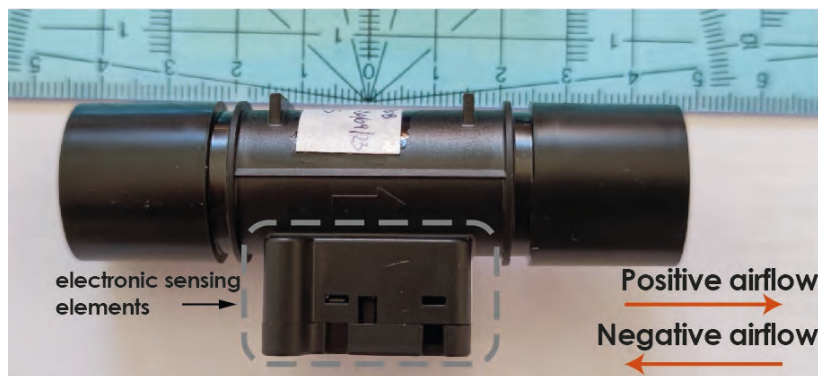


Figure 4.1 – Sensirion SFM 3003-300 CE. Internal pipe diameter: 3.61 cm.

The device measures the flow rate in standard litres per minute (SLM) at standard conditions ($T = 20^{\circ}\text{C}$ and $P = 1013 \text{ mbar}$), with a declared accuracy of $\pm 2\%$ of the measured value. The instrument has a calorimetric element sensor based on a micro-electromechanical system (MEMS). The microelements are on a silicon chip, which hosts a micro-heater and an upstream and downstream temperature sensor. The heater generates a thermal field. In the absence of flow, the temperature of the two sensors is equal and the thermal field is symmetric. In the presence of a draft, the instrument determines the thermal asymmetry (oriented as the flow) by measuring the ΔT at the two extremities which produces a measurable signal that is a function of the flow velocity (Monitaro and Domanski, 2023). To check the SFM precision and accuracy, independent calibration was carried out in a wind tunnel at the Swiss Federal Institute of Metrology (METAS).

4.2.2. Calibration test and instrumental reproducibility

The Swiss Federal Institute of Metrology (METAS) hosts a wind tunnel (flow velocity's BIPM ID: [EURAMET-M-CH-00000CIJ-1](#)) to generate air drafts between 0.10 and 50 m/s at extremely high accuracy (Table 4.1).

Type of instrument	Measuring range	Uncertainty
Anemometers	$v = 0.10 - 1.0 \text{ m/s}$	0.02 m/s for $v < 1 \text{ m/s}$
	$v = 1.0 - 50.0 \text{ m/s}$	2% for $v > 1 \text{ m/s}$

Table 4.1 – METAS' uncertainties related to wind velocity measured in the wind tunnel.

The device SFM_SN0708 was clipped on a rod in an open section of the wind tunnel (Figure 4.2.a/b) and was tested over a range of air velocities between 0.15 and 8 m/s ($n=11$). The instrument was calibrated for both directions (at $\theta=0^{\circ}$ for positive and $\theta=180^{\circ}$ for negative values) and a rotation experiment was carried out at three different velocities (1.2, 1.6, 5.0 m/s) at θ angles of 10° , 20° , 30° and 45° from the vertical axis, respectively (Figure 4.2.c).

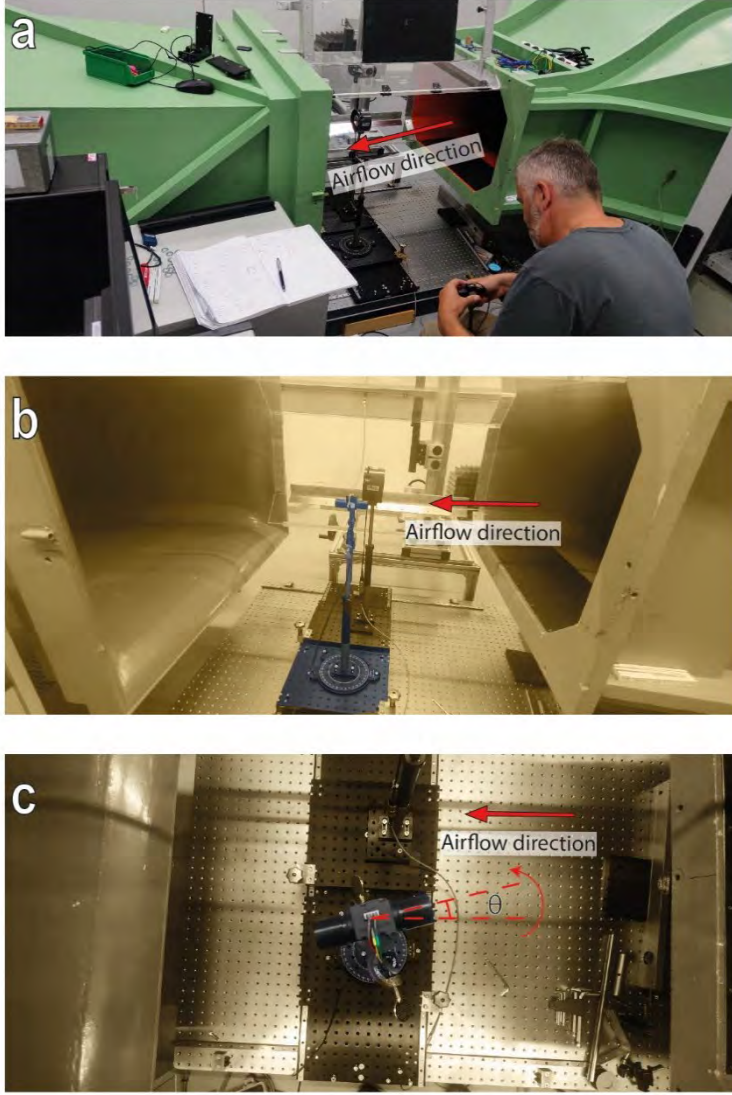


Figure 4.2 – The METAS’ wind tunnel (a), the experimental setup (b) and the rotational experiment (c).

The SFM was connected to the laptop through the SEK-Control Center software v.1.38 (Sensirion®, 2024). The data sampling rate ran at 100 Hz and the flow rate was averaged and logged every 5 measurements. The average (Avg_{SLM}) and the standard deviation (σ_{SLM}) of the measured values ($n=401$) were determined for each reference velocity (V_{ref}).

The SFM’s absolute error (Err_{abs}) is given by $\sqrt{Err_{ref}^2 + Err_{SFM}^2}$, where Err_{ref} is the error on the reference velocity based on the V_{ref} standard deviation and uncertainty given in Table 4.1, and Err_{SFM} is the velocity error on SFM measurements inferred as $V_{ref} \cdot 2\sigma_{SLM}/Avg_{SLM}$.

The reproducibility of the instruments was investigated in the lab at the Suisse Institute for Speleology and Karst Studies (SISKA) by comparing measurements from two different devices, SFM-1 (SN0501) and SFM-2 (SN0382), with the calibrated reference (SFM_SN0708) at different air velocities. The airflow was generated using a fan mounted on a 1.10 m-long pipe and powered with an adjustable supply. The fan velocity was controlled by eight different voltages between 5V and 30V.

Instrumental reproducibility was tested further in the field by placing two other devices (SFM_X2 and SFM_TK.D) in a ventilated passage at Milandre Cave, Switzerland. These two instruments were installed on two opposite corners of a 0.68x0.68 m trapdoor present in the cave and recorded in parallel for 75 days.

4.2.3. Air velocity distributions across a cave cross-section

The uncertainties related to the wall-air boundary layer were investigated in another ventilated cave (Longeague Cave, Val-de-Travers, Switzerland). Longeague Cave ventilation is dominated by a chimney effect (Pastore et al., 2024), with a driving pressure strongly related to the temperature difference between inside and outside the cave (Luetscher and Jeannin, 2004; Lismonde, 2002; Badino, 1995). Inside the cave, an irregular cave cross-section was instrumented using three calibrated SFM devices (SFM-1, SFM-2 and SFM_{wall}, Figure 4.3). These instruments were deployed simultaneously in the 1.2 m² cave passage (1.27 m wide) at a similar height, 0.64 m above ground. One instrument was in direct contact with the cave wall (SFM_{wall}), while the two others (SFM-1 and SFM-2, Figure 4.3) were positioned at distances of 0.31 m and 0.54 m from SFM_{wall}, respectively. The experimental setup was left in the cave for one week (6 October – 13 October 2023) with strong diurnal variations including flow reversals (upward and downward ventilation).

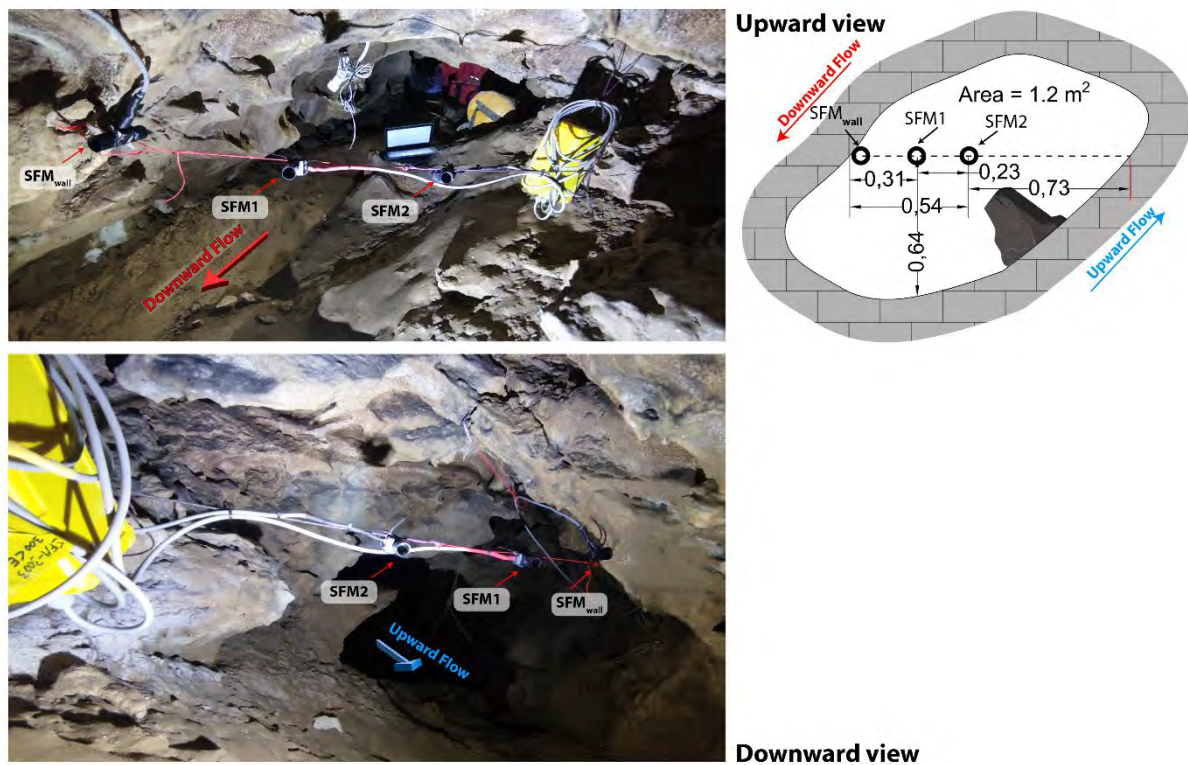


Figure 4.3 – Position of the instruments to test the velocity distribution in Longeague Cave. The dimensions on the right are in meters.

4.2.4. Modeling approach and numerical simulation

A 3-D lidar scan (Leica BLK2GO) was made along the 8 m long conduit hosting the SFMs in Longeague Cave (Figure 4.3). The spatial resolution of the point cloud is up to 5 cm but was reduced to 20 cm during data processing to decrease the computational time of numerical simulations. The conduit is considered far enough from cave entrances such as $T_{\text{wall}} \approx T_{\text{air}}$, therefore the effect of free convection on the wall surface is negligible (Sedaghatkish et al., 2024). Figure 4.4.a/b displays the scanned cave

passage. The conduit has only two openings, which act alternatively as inlets and outlets, depending on the direction of the airflow. When the airflow is downward, the inlet is at the top of the z-axis and the outlet is at the bottom. For upward flow, the inlet and outlet are reversed. The red cross-section in Figure 4.4.b, at the projected distance $y = 5.7$ m, highlights the instrumented cross-section (from now on CS_{5.7m}).

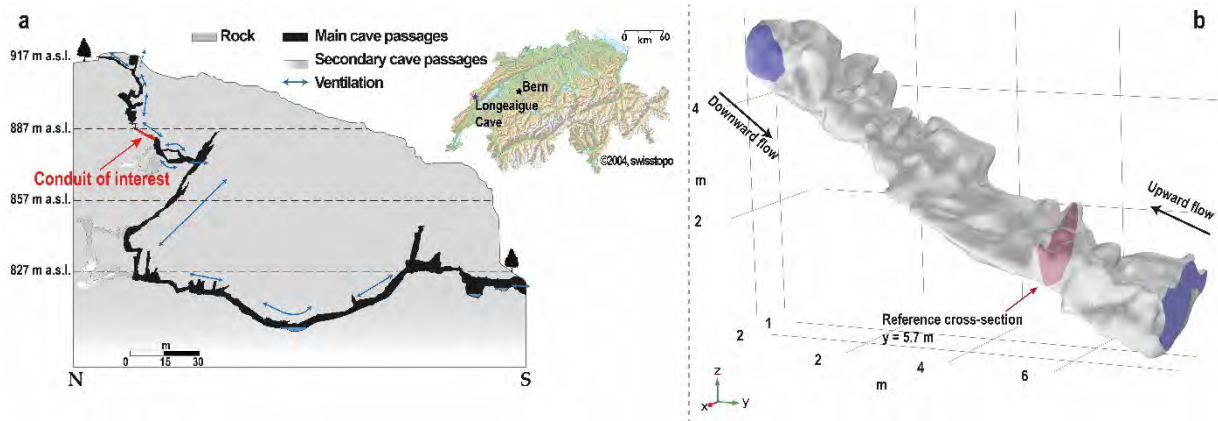


Figure 4.4 – In a) an excerpt of the Longeigue Cave survey (modified after Jeannin, 2018) and the scanned conduit highlighted in red. In b) is the computational domain of the 3-D scanned cave passage. Inlet/outlet boundaries are indicated in blue colour. In red is the reference cross-section CS_{5.7m} ($y=5.7$ m).

We developed a numerical model using COMSOL Multiphysics® v6.2 (COMSOL Multiphysics®, 2023) to calculate the velocity components and pressure. The software works based on Finite Element Analysis (FEA) for the discretization of Partial Differential Equations (PDE) by subdividing the computational domain into smaller parts called finite mesh elements.

The governing equations comprise the continuity-momentum equations as well as the k - ϵ turbulence model based on Reynolds-Averaged Navier-Stokes formulations (RANS). The full form of the equations can be found in Hammoodi et al. (2022) and Sivakumar et al. (2021). The k - ϵ model solves for two variables: k , the turbulence kinetic energy; and ϵ , the rate of dissipation of turbulence kinetic energy. More comprehensive explanations about the k - ϵ turbulence model can be found in Wilcox (1998).

Incompressible and steady-state airflow is considered. A uniform velocity distribution is imposed at the inlet surface. The atmospheric pressure is imposed at the outlet and a no-slip (zero velocity) boundary condition is employed for the cave walls. The air density and dynamic viscosity are assumed equal to 1.24 kg/m^3 and $1.77 \times 10^{-5} \text{ Pa}\cdot\text{s}$, respectively.

Mesh independence was tested for different numbers of grid resolutions to finally select a total of 1.78 million elements. The numerical results do not change with increasing mesh elements.

A turbulent flow is verified when Reynolds number (Re) $> 2 \cdot 10^3$, where $Re = 4Q\rho/\pi d\mu$, Q is the volume flow rate [m^3/s], ρ is the density of humid air (1.24 kg/m^3 at 10°C), d is the aeraulic diameter [m] determined as $4 \cdot \text{area}/\text{perimeter}$, and μ is the dynamic viscosity of air ($1.78 \cdot 10^{-5} \text{ kg/m}\cdot\text{s}$).

The numerical model provides the air velocity distributions under stationary conditions in a turbulent. Although the comparison with field data may be only qualitative because of steady-state assumption, the model provides a better understanding of the relative position of the instruments in the streamlines

and supports the interpretation and quantification of uncertainties derived from complex conduit geometries.

4.3. Results

4.3.1. Device calibration and reproducibility

Figure 4.5 shows the experimental time series measured in the wind tunnel with SFM_SN0708 at 0.5, 1.6, 2.0 and 5.0 m/s. For each velocity, the measured values (in Standard Litre per Minute, SLM) exhibit a standard deviation (σ) in the order of a few per cent ($n=401$). The calibration results are given in Table 4.2 and Figure 4.6.a. The uncertainty is up to 26% for velocities <0.30 m/s, while between 6 and 3% above 0.5 m/s. Table 4.2 confirms that the SFM flow meter measures almost the same value when rotated by 180° within the airflow. This consistency is also evident in Figure 4.6.a, indicating that the measurements are equally reliable in both directions.

V_{ref} (m/s)	Avg _{SLM} at 0° (+SLM)	σ_{SLM} (n=401)	SFM Err _{abs} (m/s)	SFM Err _{rel} (%)
0.15 (± 0.02)	0.19	0.008	0.04	26%
0.30 (± 0.02)	0.61	0.015	0.03	10%
0.50 (± 0.02)	1.54	0.017	0.03	6%
0.80 (± 0.02)	3.33	0.035	0.04	5%
1.20 ($\pm 2\%$)	5.94	0.054	0.04	3%
1.60 ($\pm 2\%$)	8.81	0.077	0.05	3%
2.00 ($\pm 2\%$)	11.66	0.099	0.07	4%
2.50 ($\pm 2\%$)	15.94	0.156	0.08	3%
3.00 ($\pm 2\%$)	20.44	0.165	0.09	3%
5.00 ($\pm 2\%$)	39.28	0.405	0.16	3%
8.00 ($\pm 2\%$)	73.84	1.095	0.30	4%
V_{ref} (m/s)	Avg _{SLM} at 180° (-SLM)	σ_{SLM} (n=401)	SFM Err _{abs} (m/s)	SFM Err _{rel} (%)
0.14 (± 0.02)	-0.15	0.014	0.04	28%
0.30 (± 0.02)	-0.60	0.018	0.04	13%
0.50 (± 0.02)	-1.50	0.020	0.03	6%
0.80 (± 0.02)	-3.37	0.039	0.04	5%
1.60 ($\pm 2\%$)	-9.12	0.087	0.06	4%
2.50 ($\pm 2\%$)	-16.02	0.150	0.08	3%

Table 4.2 – Reference velocities and measured SFM response in standard litre per minute (SLM) at 0° (+values) and 180° (-values).

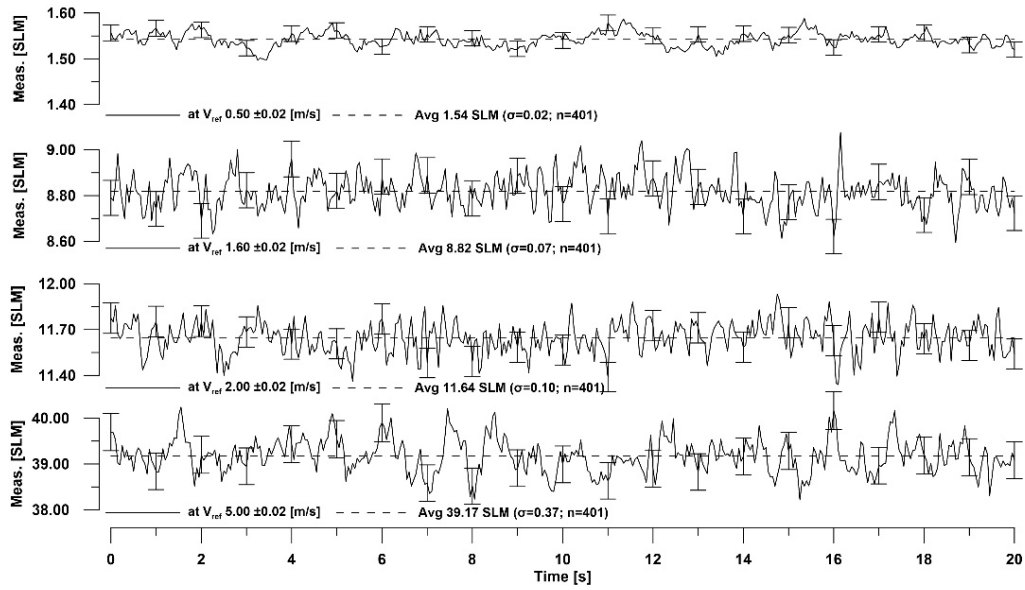


Figure 4.5 – SFM data series (in standard litre per minute) at four reference velocities ($V_{ref}=0.50/1.60/2.00/5.00$ m/s). Results show a low standard deviation (σ ; $n=401$) at constant airflow.

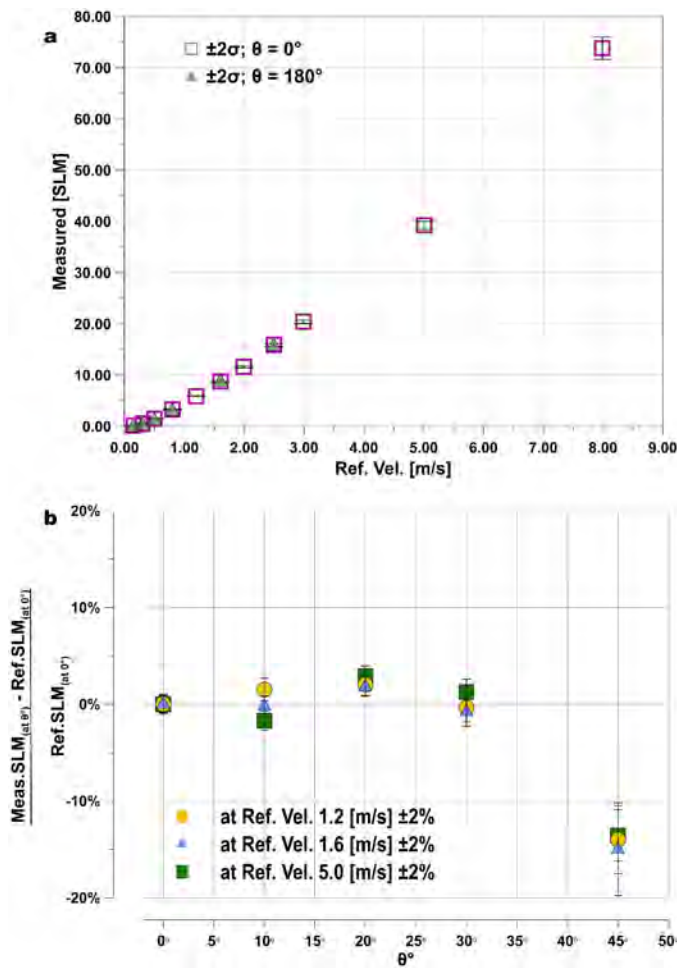


Figure 4.6 – A) Measured SLM (y-axis) against the reference velocities (x-axis). B) Rotation test results at three V_{ref} . For $0^\circ < \theta < 30^\circ$ and $1.2 < V_{ref} < 5.0$ m/s, the velocity deviation from reference is reasonably low (between 0.33% and 3%). For $\theta=45^\circ$, the deviation exceeds 10% for all the velocities.

The rotation experiment (Figure 4.6.b) shows that with angles within 10° to 30° from the flow axis, the deviation of the SFM measurements $[(SLM_{(at \theta^\circ)} - Ref.SLM_{(at 0^\circ)} / Ref.SLM_{(at 0^\circ)}]$ ranges between 0.33% and 3% for V_{ref} values between 1.2 and 5.0 m/s. At 45°, the error exceeds 10% for all the velocity steps. These results indicate that the pressure drops associated with the instrument shape and orientation are not significant as long as the angle stays reasonably low (less than 30°).

The SFM_SN0708 calibrated in the wind tunnel was compared with the other two SFMs (SFM-1_SN0501 and SFM-2_SN0382) tested in the SSKA's lab using the pipe with a fan. Results are shown in Figure 4.7.a. The values are well correlated with a ratio close to 1:1 compared to the reference SFM_SN0708. The high standard deviation evident in Figure 4.7.a is rather due to the flow instability along the pipe than to instrument deviation, showing that measurements should last sufficiently long to compensate for such turbulent fluctuations.

The devices setup in Milandre Cave (SFM_X2 and SFM_TK.D) also confirm the instrumental reproducibility (Figure 4.7.b).

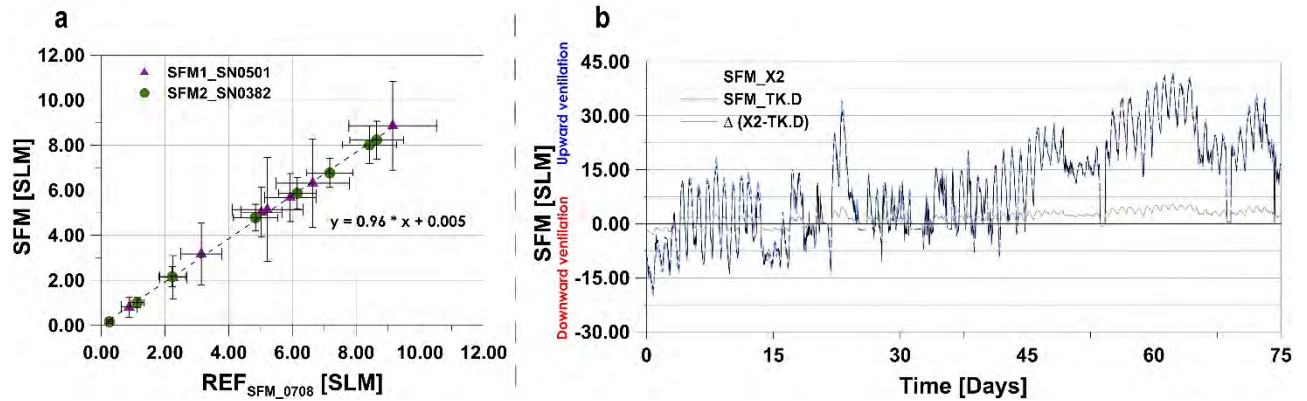


Figure 4.7 – In a), comparison between SFM-1 and SFM-2 using a pipe with a fan. Results highlight the reproducibility of the two instruments. In b), comparison between SFM_X2 and TK.D at the trapdoor in Milandre Cave. These results confirm the reproducibility of the instruments.

4.3.2. Cave monitoring

Figure 4.8 shows the measurements from three different SFMs deployed in the same cross-section in Longeague Cave, close to and far from the cave wall. The hourly time series show clear differences between the instruments (Figure 4.8). SFM-1, located at 0.31 m from one wall and 0.96 m from the other, measured the highest velocities in both ventilation regimes (1.39 ± 0.07 m/s; -1.00 ± 0.04 m/s). SFM_{wall} measured velocities close to SFM-1 during upward flow. Instead, SFM-2, located at 0.54 m and 0.73 m from the walls, logged velocities close to SFM_{wall} during the downward ventilation (+velocities), and comparable to SFM-1 during the upward ventilation (-velocities).

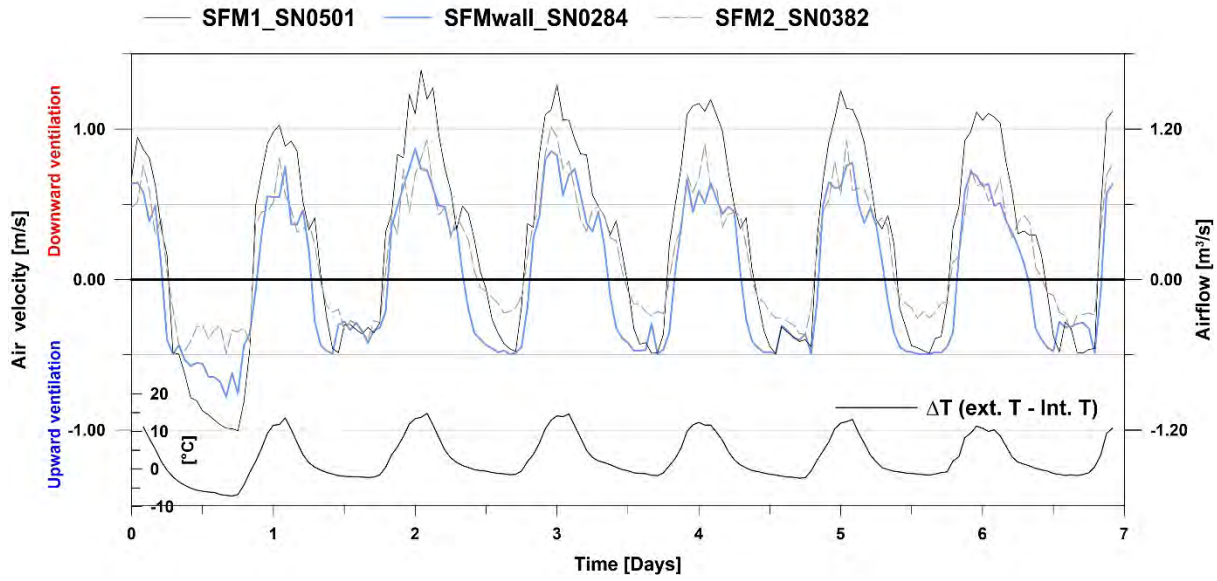


Figure 4.8 – Diurnal ventilation cycles of cave air velocities measured with SFM-1, SFM-2 and SFM_{wall} in the same cross-section at hourly time step. The data were recorded in Longeigue Cave (Switzerland) between 06 October and 13 October 2023. The lower part of the graph shows the ΔT (°C) between the external temperature measured in front of the cave entrance and the cave temperature at the same location as the SFMs.

4.3.3. Modelled air velocities

The simulations were carried out with steady airflows of 0.65, 1.35 and 2.00 m³/s, for both upward and downward flows. These airflow values were chosen based on field observations (Pastore et al., 2024) and cover much of the spectrum of possible airflows in Longeigue Cave. The model results (Figure 4.9) are compared with empirical data measured in a cross-section at 5.7 m (projected distances on the y-axis) from the upper end of the conduit (CS_{5.7m}). The measured velocities confirm a turbulent regime with a Reynolds number ranging between 10⁴ and 10⁵.

The velocity distributions modelled for an airflow of 1.35 m³/s are shown in Figure 4.9 A/B for cross-sections at 2.2 m, 3.5 m (projected distances on the y-axis) and CS_{5.7m}. The model results clearly show that the velocity distribution depends on the 3-D geometries and flow direction: while for downward flow low-speed zones appear to be more prominent near the ceiling in the 3.5 m and CS_{5.7m} cross-sections, such zones are visible on the right wall (from an upward point of view) during upward flow. At the cross-section 2.2 m the presence of an obstacle reduces the speed at the floor for the downward-modelled flow.

Focusing on CS_{5.7m}, the results for the three modelled airflows are shown in Figure 4.9, from C to H. For low airflow (C and F with 0.65 m³/s) we observe a quite homogenous distribution of velocities (between 0 and 1 m/s) compared to higher airflows (0 – 3 m/s). The velocity distribution within the cross-section is also different depending on the airflow direction: a zone of low velocity is more expanded in the top-right corner for all upward models (Figure 4.9.F/H) than for downward models. This low-speed zone is due to the presence of a rock chock visible in the 3-D scan and in Figure 4.3 (downward view). For all airflows, the centre-low of the cross-section always displays the highest velocities (Figure 4.9).

In Figure 4.9, the position of the SFM probes is depicted. Downward velocities from SFM-1, SFM-2 and SFM_{wall} can be compared with velocity distributions predicted for 1.35 m³/s downward airflow (Figure 4.9.D). For this flow rate, modelled velocity distributions show higher values at SFM-1 than at SFM_{wall} as observed *in situ* (Figure 4.8). However, the model predicts a higher velocity at SFM-2 than at SFM-1

and SFM_{wall} , which contrasts with the higher velocities measured at $SFM-1$ during downward ventilation (Figure 4.8).

During upward ventilation, measured velocities (Figure 4.8) compare well with the modelled velocities for $0.65 \text{ m}^3/\text{s}$ (Figure 4.9.F). Both, the measured (dashed line in Figure 4.8) and predicted velocities of $SFM-2$ show lower values than $SFM-1$ and SFM_{wall} .

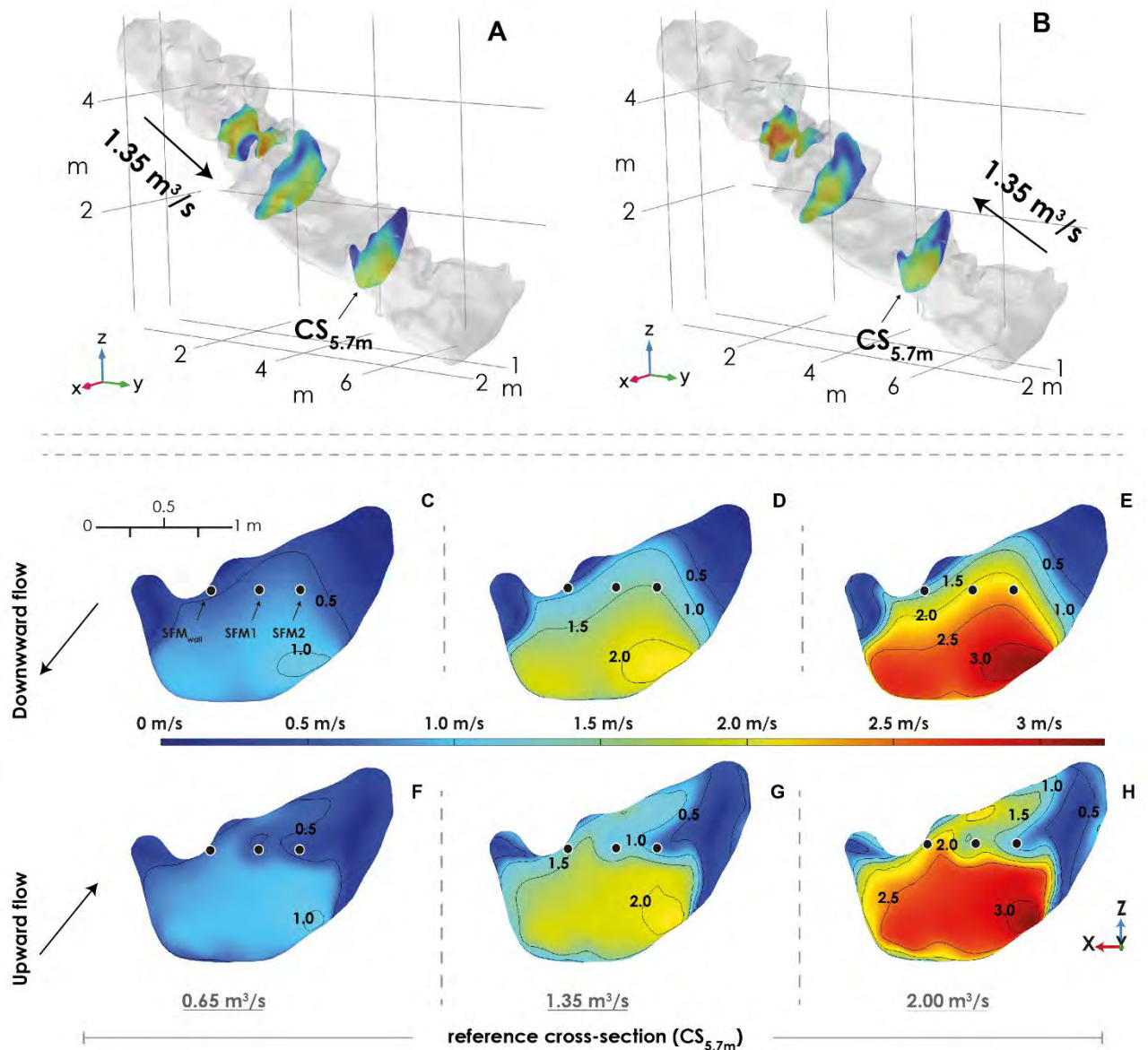


Figure 4.9 – Modelled air velocity distributions along the conduit and at the same cross-section as the cave measurements. $CS_{5.7m}$ velocities refer to three different airflows and opposite flow directions. The three black circles identify the SFM positions as deployed in the cave.

4.4. Discussion

4.4.1. Quality of the measurements and airspeed conversion

Flow measurements under steady conditions in an air tunnel showed a high precision of the SFM ($\pm 1\%$) between 0.5 and 5 m/s (Table 4.2, Figure 4.5 and Figure 4.6.a). The 17 individual measurements (Figure 4.6.a) show an instrumental uncertainty Err_{rel} of $\pm 5\%$ between 0.5 and 8 m/s, and between 26/28% and 10% for $0.15 < V_{ref} < 0.5$ m/s, for both flow directions. Nevertheless, turbulent flow is typical for caves (Lismonde, 2002) where the spatial variability of the cross-section can be significant, leading to an increase in the Reynolds number ($Re \gg 2000$). Thus, in a turbulent field, the stochastic error related to the sampling may impact the measurements (Mauder and Zeeman, 2018; Finkelstein and Sims, 2001).

The data is reproducible (Figure 4.7.a and Figure 4.7.b) and non-sensitive to minor changes in orientation (3% at $\theta = 30^\circ$; Figure 4.6.b) making it suitable for installation in complex cave sections. The device proved to measure correctly in both directions, a requirement that fits well with the dual behaviour of a cave ventilated by a chimney effect.

The relationship between measurements (V_{ref} against SLM) shown in Figure 4.6.a was found to be close to quadratic. Although the relation between mass flow rate and temperature difference in a thermal flowmeter is first-order (i.e., linear), other thermal processes may compete for a non-linear relationship (e.g. thermal exchanges with device walls). In particular, the device is designed to measure in a closed system with a pipe as a flow inlet. Instead, we deploy the instrument immersed in a flow that is free to enter the pipe device, thus probably causing boundary effects of the second order that may disrupt linearity.

The conversion of SLM into m/s can be derived by plotting the points shown in Figure 4.6.a with the two axes swapped (x-axis=SLM and y-axis=m/s). Depending on the interested velocity range, it is possible to fit the equation for the desired V_{ref} range. In the case of Milandre (Figure 4.7.b) and Longeague (Figure 4.8), we aimed to apply the conversion within the range of $0 \leq |SLM| \leq 39.0$ ($0 \leq |V| \leq \sim 5.0$ m/s), ensuring that the intercept of the equation is forced to 0 (i.e., 0 m/s = 0 SLM). Since a single equation would not fit within 2σ for $SLM \leq 6.0$ ($\sim |V| \leq 1.2$ m/s), thereby increasing the relative error at low speeds, two different equations were obtained.

Within the range $6.0 \leq |SLM| \leq 39.0$ ($\sim 1.2 \leq |V| \leq \sim 5.0$ m/s), the conversion is obtained by Eq. 4.1:

- $y = -0.0014 \cdot x^2 + 0.1828 \cdot x$

while for $|SLM| \leq 6.0$ ($|V| \leq \sim 1.2$ m/s) by Eq. 4.2:

- $y = -0.0243 \cdot x^2 + 0.3422 \cdot x$

The absolute velocity value assures the same equation for both flow directions (positive or negative). The original measurement sign (indicating the flow direction) can be restored after the conversion into m/s.

According to the manufacturer, no corrections for either barometric pressure or temperature are needed.

4.4.2. Airspeed distribution through a cross-section

The high reproducibility of the instruments allows for studying the impact of conduit geometries on the measured air velocity (Figure 4.3 and Figure 4.8). The comparison of the values measured by three

different SFM sensors within the same cross-section of a cave passage pointed out differences, which are clearly beyond the range of the instrument's accuracy. This reflects the heterogeneity of the flow field within the cave cross-section. At the test site, SFM-1 measured the highest velocities during the experiment (Figure 4.8). This is likely due to the SFM-1 position in a sort of “channel” free of obstacles, as shown in Figure 4.3. SFM_{wall} did not show the lowest velocities, despite being located on the wall. Instead, the lowest velocities were measured at SFM-2 during the upward regime (negative values) although the instrument was far from the walls (0.54 m and 0.73 m). Therefore, SFM-2 exhibits a different response during downward and upward ventilation, when compared with SFM-1. This is probably due to the presence of an obstacle affecting the flow, causing a different response according to the wind direction. We conclude that the position of the SFM-1 is less affected by the cross-section morphology since it measured the highest velocities. During downward regimes (+val.), the velocity difference between SFM-2 and SFM-1 (taken as a reference) is -42% as well as between SFM-1 and SFM_{wall}. Conversely, in the upward regimes, the velocity difference between SFM-1 and SFM_{wall} reduces to approximately -21%, while staying at -42% for SFM-2.

The airflow modelled along the 3-D conduit geometry confirms the results obtained from field measurements. In fact, the modelled velocities highlight the spatial variability along the conduit (Figure 4.9). These differences are related to the complex geometry of the conduit and the distribution of the obstacles therein (Figure 4.9.A/B). Furthermore, airspeed distributions are less uniform across CS_{5.7m} as airflow increases. Consequently, the differences measured between the three SFMs can sensibly increase at a high volumetric flow rate. The flow distribution changes also according to the flow direction, as shown by the low-speed zone appearing during upward flow. Nevertheless, the model prediction contrasts with SFM-2 field measurements during downward flow, since the model predicted higher velocity than SFM-1. The reduction of the 3-D resolution (from 5 to 20 cm) and the associated loss in spatial resolution could explain a different velocity distribution. Our results show that the measurement of the air velocity in natural settings (e.g. a cave) depends not only on the airflow but also on the relative position of the instrument in the streamlines and flow direction.

4.4.3. Impact on air volume estimation

Long airflow data series in caves are of interest to study the air dynamics (Gomell and Pflitsch, 2022; Faimon and Lang, 2013) and the associated energy exchanged therein (Obleitner and Spötl, 2011; Luetscher et al., 2008). Airflow monitoring usually derives from a punctual airspeed measurement integrated through a specific cross-section, but this may not necessarily represent the average velocity across the entire cross-section. The instrument position determines the measured velocities and these uncertainties impact the calculated airflow (m³/s) distributed over the 1.2 m² CS_{5.7m} (Figure 4.8 and Figure 4.9.C/H). Instruments placed in the same cross-section have shown differences as high as 42% (~0.5 m/s) between them, although they were only a few dozen centimetres away from each other. These experimental data were confirmed by modelled air velocity distribution in a cave cross-section.

The modelled air distributions suggest that there is no clear location to deploy a single anemometer within a cross-section (Figure 4.9). If an anemometer is placed in a high-speed zone, there is a high possibility of overestimating the airflow. Conversely, positioning it close to the wall may result in an underestimation of the flow. The model in Figure 4.9 suggests that the average velocity between SFM-1, SFM-2, and the SFM_{wall} position is likely to provide a value close to the actual airflow rate. For instance, considering velocities of 0.6, 1.3, and 1.5 m/s for SFM_{wall}, SFM-1, and SFM-2, respectively, results in an airflow rate of 1.4 m³/s over the CS_{5.7m} (1.2 m²). This value closely matches the modelled

1.35 m³/s. Therefore, the actual airflow passing through CS_{5.7m} during the monitored period is likely to average the airflows measured at SFM-1, SFM-2 and SFM_{wall}.

The cumulated air volume transiting through Longeague Cave over seven days is shown in Figure 4.10 for the three devices. As expected from the airspeed time series (Figure 4.8), the total air volume integrated with SFM-1 values is the highest with $4.4 \cdot 10^5$ m³ ($\sim 5.4 \cdot 10^5$ kg). It overtakes the air volumes measured by SFM-2 and SFM_{wall} by 39% and 27%, respectively. Since the estimation of the air volume passing through CS_{5.7m} is directly dependent on the inferred airflow, the average air volume estimated from the three instruments should provide a good approximation of the total air volume over the monitored period (Figure 4.10, dashed orange line). Moreover, the three cumulated curves in Figure 4.10 show different slopes, but also vary according to the flow direction (coloured arrows). The latter indicates that the driving pressure related to the ΔT (external – internal temperature, Figure 4.8) did not act symmetrically on the system during the monitored days: the temperature difference was higher during the day (downward ventilation) than at night (upward ventilation). During downward flow, the slopes highlighted by blue arrows in Figure 4.10 are steeper than during upward ventilation (green arrows in Figure 4.10), implying that $Vol_{downward} \gg Vol_{upward}$ for SFM-1 and SFM-2 (Figure 4.10). SFM_{wall}, on the other hand, displayed $Vol_{downward} \approx Vol_{upward}$.

These results show clearly that the positioning of the instrument plays an important role in the measurement uncertainties. In our case, these uncertainties largely dominate the instrument uncertainty ($\pm 5\%$) and those related to the device's orientation (2.5% at 1.2 m/s, $\pm 30^\circ$). Therefore, choosing the appropriate measurement point is not straightforward. Depending on the researched precision, more than one flowmeter should be deployed in a single cross-section to infer an average air velocity to be integrated over the cross-section. However, if only one instrument is to be used, the measuring point could be chosen after carrying out manual measurements with a hand-held anemometer to investigate the velocity distribution (taking care to cover the whole cross-section, and not to disturb the flowfield by the human body within or close to the cross-section). A fixed anemometer should not be deployed in the high-speed zone nor the low-speed zone, but in a velocity field which could represent as closely as possible the average velocity. In the case of one-time measurements to estimate airflow with a hand-held anemometer, one approach could be to measure airflow in three or more different cross-sections along the same conduit. The volumetric airflow derived from measurements with anemometers (hand-held or in place) may also be compared with tracer gauging tests (Pastore et al., 2024). In fact, the latter does not require velocity integration, as it is derived from the mass conservation equation and yields reliable flow rates even at low air speeds.

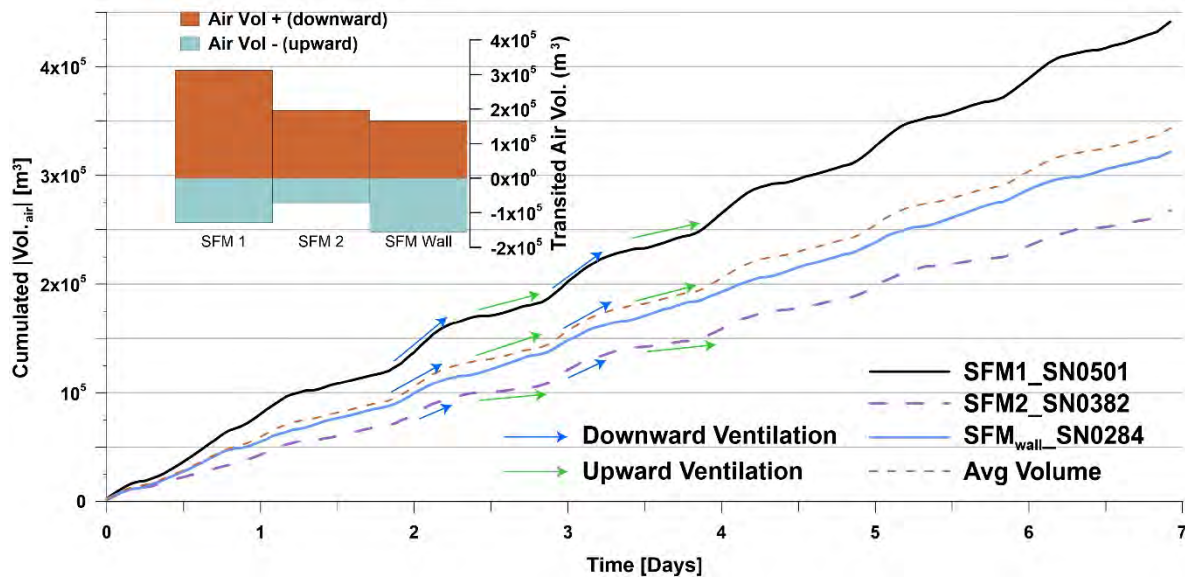


Figure 4.10 – Cumulated air volumes evaluated from the data series shown in Figure 4.8.

4.4.4. Experience from long-term monitoring

The testing of SFM probes in a wind tunnel, in the SSKA's lab with a pipe and in naturally ventilated conduits, revealed these probes to be suitable for cave monitoring, more than most other types of sensors we tested. Several SFMs were installed for at least one year in different caves subject to chimney ventilation. Some of these data have already been used in scientific publications (Sedaghatkish et al., 2024; Garagnon et al., 2022). Furthermore, three years of airflow monitoring data (very few gaps) collected with an SFM device in Longeague Cave are being interpreted to assess air thermal exchanges within the ventilated conduits. Despite environmental conditions including high-condensing environments, freezing conditions, sporadic flooding, and anthropogenic disturbances, the instruments have measured with high reliability, confirming their robustness. In our experience, most of the acquisition gaps did not derive from the sensor, but from the failure of the logger's electronic parts, more sensitive to environmental conditions, or battery failures. Good insulation of electronic parts is, therefore, a priority. The battery duration depends on the efficiency of the logging system since the device itself requires only 3.3 V to run (~ 3.8 mA during measurements). For instance, in D7.1 Cave (Switzerland) we deployed an SFM working with our logger based on a TinyPICO microcontroller appropriately coded and powered by a 3.7 V/5500 mAh lithium battery. This cave is difficult to access (high mountain cave), so we left the instrument working for 1 year with a 1 hour time step and without any maintenance (Sedaghatkish et al., 2024). However, after prolonged exposure, condensation within the SFM pipe could potentially disrupt measurements. Our field experience indicates that installing the device with the electronic sensing part (Figure 4.1) upside down allows moisture to drip away from the sensor. In any case, after drying the sensor (e.g. with a light cloth), the device usually worked again. This positive experience with the SFM contrasts with independent measurements carried out with various other commercial devices, including an ultrasonic anemometer, which could not handle the condensing conditions without extra heating, possibly affecting the cave climate locally, and which would require an additional power supply.

4.5. Conclusions

Measuring airflows in caves is of interest to study the air dynamics and the related energy exchanges in karst systems. Besides reliable measurements, high-performance and robust instruments are key points when investigating challenging environments such as caves. In this article, we tested an instrument that meets our requirements: 1) low power supply, 2) reliable measurements, 3) durability and 4) affordability. The instrument showed a low Err_{rel} , $\pm 5\%$ between 0.5 and 5 m/s, a high resolution (up to 16 bit) and the indication of flow direction. Nevertheless, despite the high resolution and accuracy of the instrument, calculating the volume or mass flow rates from air velocity measurements might be not obvious. This difficulty arises primarily because the velocity field within a single cross-section is non-uniform and fluctuates with the flow rate and direction.

Our results show clearly that the position of a single instrument in a cave conduit plays a major role in the estimation of the airflow. While long-term monitoring with single instruments unlikely records the “real” airflow, determining a plausible range of airflow is key for the calculation of energy balances or the modeling of aerosol transport. Airflow monitoring with several instruments distributed in the same cross-section could help to reduce the measurement uncertainty. Computational models aiming to correct the integration of one-point velocity measurement in a specific cross-section represent another possible approach.

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Authorship statement

CP, ML, and PYJ designed and directed the study. EW and NS developed the electronic devices used for the measurements. CP performed the measurements and analyzed the field data. AS carried out the simulations. CP wrote the paper with input from all authors.

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Chapter 5. Dispersion of artificial tracers in ventilated caves (published paper)

Pastore, C., Weber, E., Doumenc, F., Jeannin, P.-Y., Luetscher, M., 2024. Dispersion of artificial tracers in ventilated caves. *International Journal of Speleology*, 53(1), 51-62. <https://doi.org/10.5038/1827-806X.53.1.2497>

Abstract

Artificial CO₂ was used as a tracer along ventilated karst conduits to infer airflow and investigate tracer dispersion. In the karst vadose zone, cave ventilation is an efficient mode of transport for heat, gases and aerosols and thus drives the spatial distribution of airborne particles. Modeling this airborne transport requires geometrical and physical parameters of the conduit system, including the cross-sectional areas, the airflow and average air speed, as well as the longitudinal dispersion coefficient which describes the spreading of a solute. Four gauging tests were carried out in one mine (artificial conduit) and two ventilated caves (natural conduits). In this paper, we demonstrate that it is possible to gain reliable airflow rates and geometric information of ventilated karst conduits using CO₂ as a tracer. Airflow was gauged along two caves and one mine and compared with punctual measurements made with a hot-wire anemometer. Cross-sectional areas estimated with CO₂ tests were compared with those measured in situ. Moreover, breakthrough curve (BTC) analysis displayed an accentuated tailing along the investigated natural conduits due to the presence of dispersive singularities which possibly enable aerosol deposition. The long tailing observed in Milandre and Longeigue Caves is probably due to cross-section variations. A 1-D advection-dispersion model tested for these sites was unable to fit BTC tailing in natural conduits. In Baulmes artificial conduit, where long tailing is not observed, the dispersion coefficient has been estimated using Chatwin's method, and compared with the prediction of Taylor's theory. Despite the regular geometry of Baulmes Mine, Taylor's correlation significantly underestimates the dispersion coefficient deduced from field data, showing the need for more theoretical work on turbulent dispersion in mines. This paper gives a first insight into air motion and matter dispersion along ventilated karst conduits, preparing for proper aerosol dispersion modeling.

5.1. Introduction

In the karst vadose zone, cave ventilation is an efficient mode of transport for heat, gases and aerosols and thus drives the spatial distribution of airborne particles (Pashchenko et al., 1993; Christoforou et al., 1994; Dredge et al., 2013; Faimon et al., 2019). Modeling this airborne transport requires geometrical and physical parameters of the conduit system, including the cross-sectional areas, the airflow and average air speed, as well as the dispersion processes which describe the spreading of the transported substance (Hart et al., 2013; Adesso et al., 2022). Geometrical parameters and air velocity can sometimes be measured in the field, but inaccuracies derive from spot measurements which may poorly reflect the average air velocity along the conduits. The sizing of cross-sections is also subject to approximations when extracted from a cave survey. Hence, a method that may provide information about the conduit geometry in a safe, inexpensive and reliable way would be welcome (Field, 1999; Hauns, 1999; Palmer et al., 1999; Hauns et al., 2001; Field, 2002; Massei et al., 2006). The gauging of air has been extensively used in mine engineering to infer dust and contaminant transport in time and space along the channels to prevent mine operators' health from the release of mine smoke and fumes (Higgins and Shuttleworth, 1958; Thirnonis and Kissel, 1974; Standish, 1988; Widodo et al., 2008; Widiatmojo et al., 2015; Semin et al., 2022). However, it was only rarely applied to air dynamics in natural cave systems which differ from mines because of their more complex morphologies.

Nevertheless, the dust present in mines is analogous to the airborne aerosols present in caves and includes both, external and internal aerosol sources, anthropogenic (visitors) production, hydrological sources and bedrock production (Dredge et al. 2013). Since the production sources are widely different, the size of particles also varies. The grain dimensions together with the airflow magnitude and flow patterns will control the mode and speed of deposition (Christoforou et al., 1996; Dredge et al., 2013). For instance, the deposition of coarse particles (size $> 1\mu\text{m}$) occurs mainly at the cave entrance, whereas fine particles (size $< 1\mu\text{m}$) may travel further into the cave.

When a solute is injected in a stream of liquid flowing in a pipe, the interaction between velocity variation over the cross-section and radial diffusion of the solute results in solute dispersion. The solute transport can be described by a 1-D advection-dispersion equation where the dispersion coefficient D (i.e., the effective “diffusion” coefficient in the longitudinal direction) depends on flow characteristics and can be much larger than the thermodynamic diffusion coefficient. Experimental values of D can be obtained from the analysis of breakthrough curves (BTC), i.e., time variations of the solute concentration at a certain distance from the injection point. Taylor (1954) used a method based on the frozen-cloud approximation to extract D from BTC. Chatwin (1971) showed the limits of this assumption and proposed an alternative method based on the linear representation of the two BTC limbs. Furthermore, Taylor (1954) derived a theoretical expression of the dispersion coefficient for a turbulent flow in a rough pipe with a circular cross-section. Based on the empirical velocity profile coupled to the Reynolds analogy to obtain the eddy diffusivity, Taylor’s theory states that the dispersion coefficient is simply proportional to the pipe diameter d and the friction velocity U^* , with $D/d \cdot U^* = 5.05$.

Standish (1988) used acetylene (C_2H_2), which has similar physical characteristics (density, molecular weight and diffusivity) to those of air, to infer the dispersion by applying Taylor’s method to different injections into ventilated mine tunnels. More recent applications (Widodo et al., 2008; Sasaki et al., 2009; Widiatmojo et al., 2015;) relied on the injection of SF_6 into ventilation conduits to assess the dispersion coefficient solving the 1-D advection-dispersion equation. Their results have clearly shown that, in their case, the ratio $D/d \cdot U^*$ was between 16.5 and 65.5 times higher than Taylor’s proportionality coefficient of 5.05. Taylor’s approach (Taylor, 1954), which is limited to fully developed flow in a pipe with a circular cross-section, fails to consider singularities as bends, rock blocks and narrow passages. However, these singularities are nonetheless common in natural environments such as caves where their presence needs to be taken into account when trying to assess the dispersion coefficient. The comparison between measured and theoretical BTCs may reveal discrepancies associated with tailing. In hydrogeology and mining aerology, the tailing is interpreted as the effect of singularities (e.g. changes in geometries, presence of cavities or solute absorption) on the contaminant transport caused by particles momentarily trapped and released with retardation in the main flow. Hauns et al. (2001) suggest that retardation and high dispersion provide evidence of an irregular conduit, including either numerous dispersive features or large-scale ones (e.g. chambers in our case). Their experiments have shown that singularities slow down the flow, increase the turbulence and produce eddies. The recirculation of the low-speed flow in the main flow will strongly influence the shape of the resulting breakthrough curve. Widiatmojo et al. (2015) also pointed out that D varied over the years, probably related to the closure of some mine branches (reduction in the mine size) and associated flow rate drops.

Speleological literature about cave air tracing is poor albeit several attempts have been reported, e.g. with injected CO_2 (Cella et al., 2015a, 2015b; Gatti et al., 2015), SF_6 and CCl_2F_2 (De Freitas et al., 1982) as well as using natural tracers (Przylibski & Piasecki, 1998; Pflitsch & Piasecki, 2003) or artificially

induced airflow (Lucas et al., 2013). Most studies aimed at obtaining information on air inlets/outlets or airflow rates but none of these attempted to derive geometric information about the conduit network nor tracer dispersion along natural conduits. It should be noted that certain tracers employed are currently classified as environmentally hazardous, e.g. SF₆ used in Thirnonsl and Kissel (1974), De Freitas et al. (1982) and Widiatmojo et al. (2015) or dangerous due to their flammability, e.g. C₂H₂ used in Standish (1988).

In this paper, we demonstrate that it is possible to gain reliable air flow rates and geometric information using CO₂ as an air tracer. As for hydro(geo)logical studies (Palmer et al., 1999; Hauns et al., 2001; Field, 2002), flow rates and geometrical parameters of the conduits can be derived from the information yielded by the tracer BTC. Moreover, we demonstrate that in the quite regular geometry of a mine, it is possible to assess the dispersion coefficient D using Chatwin's method (Chatwin, 1971), commonly used in hydrogeology. Finally, we suggest that the tailing observed with CO₂ in caves with complex geometry could be caused by the same singularities that would cause aerosol deposition.

5.2. Methods

5.2.1. Site descriptions

Four CO₂-tracing tests were carried out in one naturally ventilated mine (artificial conduit, Baulmes Mine) and two ventilated caves (natural conduits, Milandre and Longeague Caves). These conduits are subject to a chimney effect (Badino, 1995; Lismonde, 2002; Luetscher and Jeannin, 2004) and three of the four tests were carried out during an upward ventilation regime (with $T_{system} > T_{ext}$), while one was performed during a downward ($T_{system} < T_{ext}$) ventilation regime (Figure 5.2, Figure 5.3, Figure 5.4).

The cross-sectional areas were surveyed with several polar points (distance and inclination) measured along the conduit's perimeter using a modified Leica DistoX310 telemeter that provides tilt, azimuth and distance (± 0.05 m). In Milandre and Longeague, the cross-sections were surveyed every 10 m along the conduits of interest. In Baulmes the sections were surveyed every 10 m for the first 200 m, and every 20 m until 360 m. For each section, between 16 and 25 data points were surveyed. The sections were drawn within CAD software to determine the cross-sectional area with an estimated uncertainty of ± 0.02 m² (Figure 5.1).

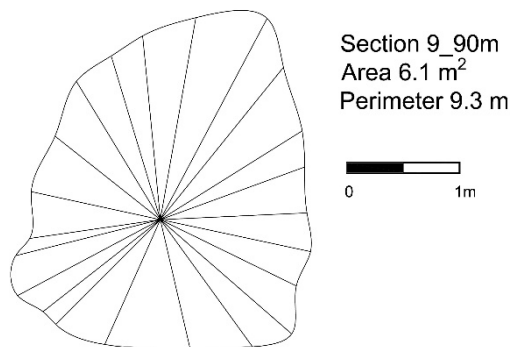


Figure 5.1 – Example of a vertical cross-section surveyed in Baulmes Mine (Section_9 at 90 m from the tunnel entrance).

5.2.1.1. Baulmes Mine

The Baulmes Mine (46°47'33" N 6°31'35" E, 655m a.s.l.) was exploited during the first half of the 20th century for the production of lime and cement. The lower mine comprises 7 subhorizontal levels with a cumulated length of about 10.7 km (Figure 5.2)(Deriaz et al., 2007). The difference in elevation

between the lower and upper entrances reaches 90 m, ensuring a strong chimney effect. Tunnel #10 gives access to the lowest mine level with a 360 m-long sub-horizontal conduit reaching the main network of the mine. The geometry is typical of a mining tunnel: flat ground (except for some local collapses), vertical walls and a barrel-vaulted ceiling. Water flows in an occasionally open culvert from the innermost parts of the mine. No air inlets or outlets are observed along Tunnel #10, except for the main entrance. Three CO₂ data loggers were deployed at different stations (Figure 5.2): for the test during the upward regime at 128 m (St_B3), 280 m (St_B4) and 360 m (St_B5) from the lower entrance; at 232 m (St_B3), at 270 m (St_B2) and 360 m (St_B1) for the test during the downward regime.

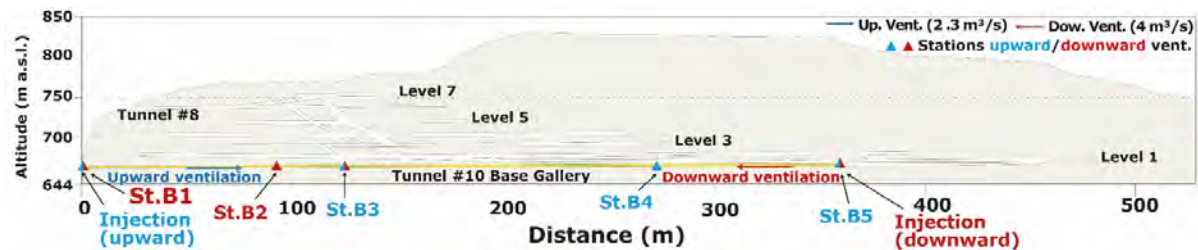


Figure 5.2 – Baulmes Mine ("mine inférieur"; Deriaz et al., 2007, modified). Blue color shows the test stations during upward ventilation, red color refers to the tests during downward ventilation.

5.2.1.2. Milandre Cave

The 10.5 km long Milandre Cave (47°29'6" N 7°0'58" E, 402 m a.s.l.), is located in Northern Switzerland on a limestone plateau at an elevation of about 500 m a.s.l. The main underground river passages cumulate 4.6 km between the upper to the lower entrances (135 m difference in elevation) (Jeannin et al., 2010). The cave complies with the chimney effect, ensuring permanent ventilation (downward or upward, according to the external temperature, Garagnon et al., 2022). The CO₂-gauging test was carried out in a 90 m long transect in the lower part of the cave (Figure 5.3). Two measurement stations were deployed at, respectively, 20 (Mil_20) and 90 m (Mil_90 m) from the CO₂ release point.

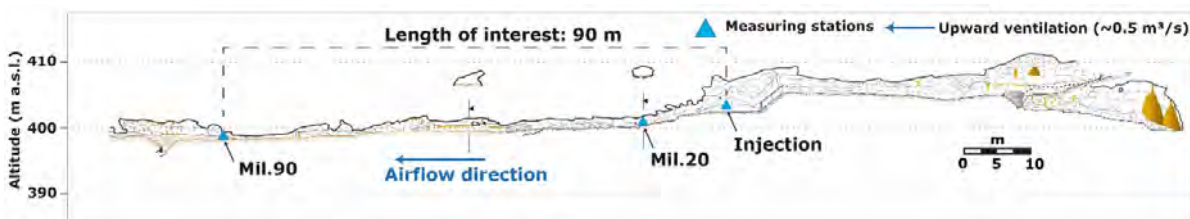


Figure 5.3 – An excerpt of the Milandre survey, showing the transect concerned by the CO₂ gauging test.

5.2.1.3. Longeaigue Cave

Longeaigue Cave lies in Val-de-Travers (Upper entrance: 46°52'22" N 6°31'09" E, 917 m a.s.l.; Lower entrance: 46°52'18" N 6°31'09" E, 820 m a.s.l.) and develops for 1.3 km. It is mainly dry and is crossed by an intense airflow due to the chimney effect developing between the two entrances. The elevation difference is 90 m with a distance of ~300 meters between the two entrances. The CO₂ test was carried out along the main passage, releasing the gas close to the lower entrance and monitoring the gas breakthrough at 3 stations: Lo_239 (239 m from the injection point), Lo_278 (278 m from the injection point) and Lo_294 (294 m from the injection point). The cave is quite large until ~237 m, with extensive phreatic subhorizontal galleries showing some vadose reshaping (shafts) morphologies, and becomes narrower in the upper part where the cave has squeezing conduits. Between the stations Lo_294 and

Lo_278 a secondary cave passage is present (red arrow in Figure 5.4) which can cause airflow drops in this cave section.

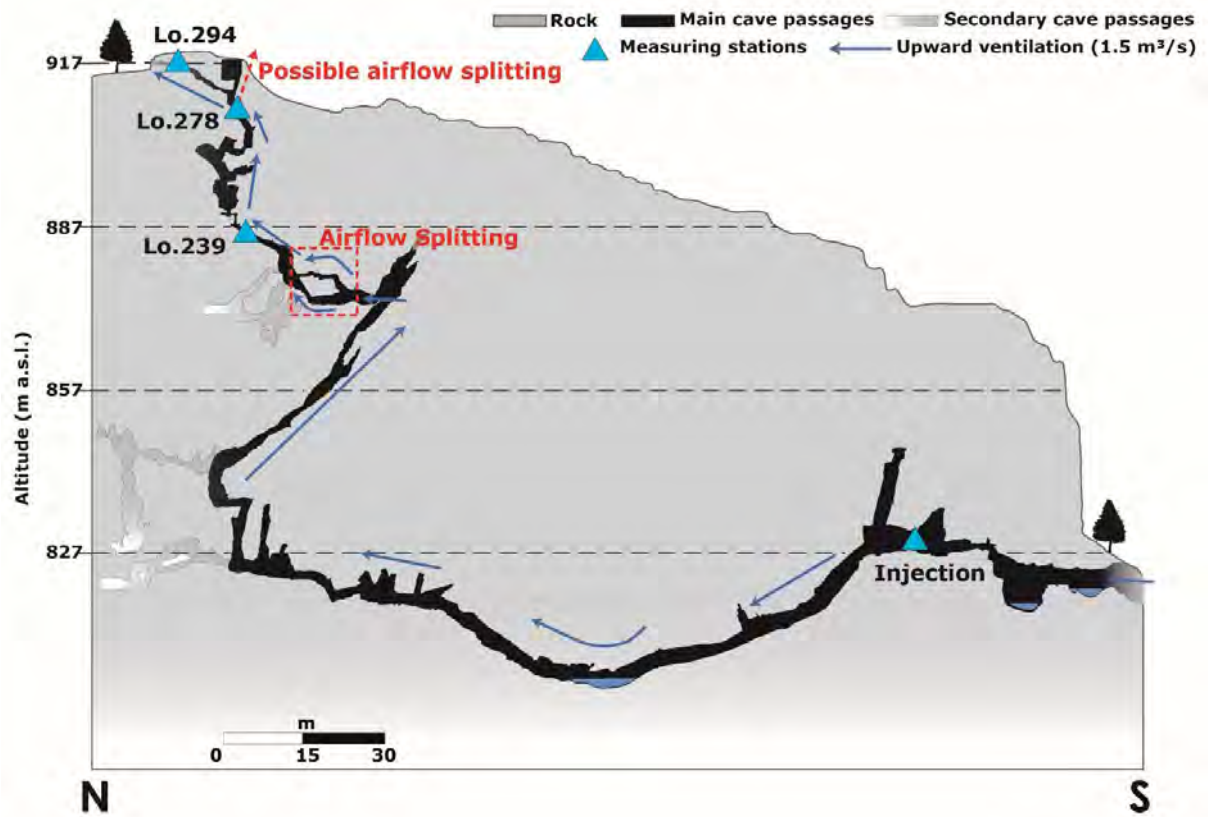


Figure 5.4 – An excerpt of the Longeigue Cave survey, showing the transect concerned by the CO₂ gauging test. Modified after (Jeannin, 2018).

5.2.1.4. Comparison between the three sites

Twenty-eight cross-sections were surveyed in Baulmes, 10 in Milandre, and 30 in Longeigue (Figure 5.5). For all but the first 50 meters, the Baulmes tunnel shape is quite homogeneous (Figure 5.5) showing a low standard deviation. In the first 128 m of the tunnel, the cross-section average is $6.4 \pm 1.8 \text{ m}^2$ ($n=13$) and from the entrance to the second station at 280 m it is $6.3 \pm 1.3 \text{ m}^2$ ($n=29$). Along the entire 360 m long tunnel the average cross-section was $6.6 \pm 1.4 \text{ m}^2$ ($n=37$). In Milandre Cave, the cross-section average of the first 20 meters (Mil_20) is $10.8 \pm 6.6 \text{ m}^2$ ($n=3$), while the whole conduit of 90 m (Mil_90) has a mean cross-section of $6.3 \pm 4.9 \text{ m}^2$ ($n=10$). In Longeigue Cave, the mean surveyed cross-section from the injection station to the first station (Lo_239) of $7.4 \text{ m}^2 \pm 4.2$ ($n=25$), to the second station (Lo_278) of $7.0 \pm 4.3 \text{ m}^2$ ($n=27$) and to the last station (Lo_294) of $6.7 \pm 4.3 \text{ m}^2$ ($n=29$). The standard deviation of averaged cross-sections (presented also in Table 5.2) makes it possible to state that Longeigue and Milandre natural conduits have a less uniform distribution of the cross-sections over a smaller distance than Baulme's anthropic tunnel and that cross-section variations in Baulmes are less pronounced than in Longeigue and Milandre (Figure 5.5). Moreover, the changes in cross-section dimensions between two measurement points along Longeigue are sharper than in Milandre and appear distributed all along the conduit. Only the last 40 meters of Longeigue (Figure 5.5) show a more uniform distribution of small cross-sections.

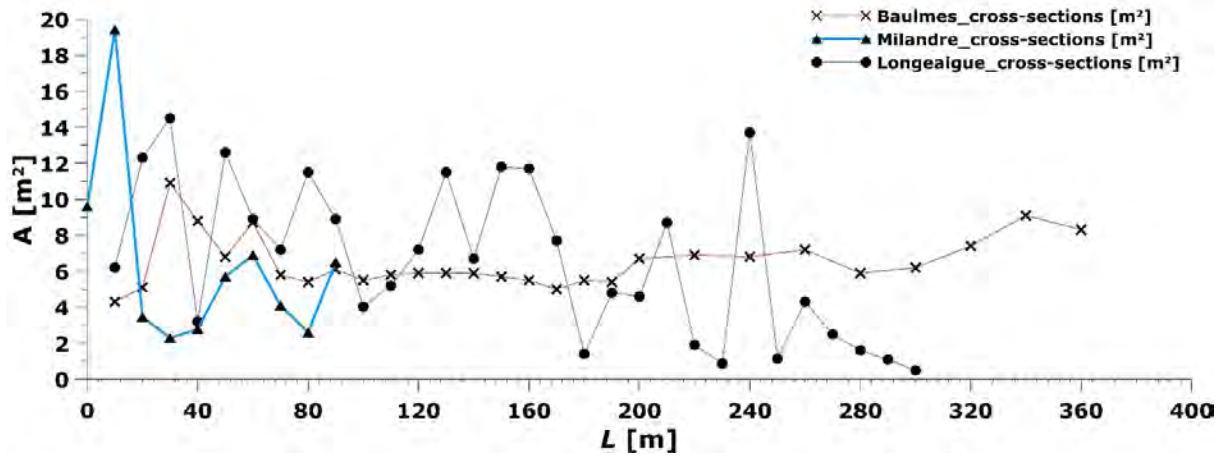


Figure 5.5 – Cross-section surface areas measured along the investigated conduits in Baulmes Mine, Milandre and Longeigue Caves.

5.2.2. Instrumentation

A tank of 1.5 kg CO₂ (2 L, 5.7·10³ kPa) was used to release the tracer. The tank was provided with a faucet valve. Hence, to ensure that $t_{release} < t_{transit}$ it was opened for 1 to 2 minutes (depending on the travel distance). The released CO₂ mass (± 0.05 kg) was obtained by weighing the tank before and after each test. CO₂ was monitored at 1 min in Baulmes and 20 sec time intervals in Milandre and Longeigue, using SCD-41 CO₂ sensors (Adafruit; device accuracy ± 50 ppmv +5% of reading) implemented on a 3.3 V microcontroller based on I2C protocol, functioning as a data logger. The error on measured concentrations is generally between 10% and 15%. The background CO₂ concentration was measured between thirty minutes and an hour before the tracer release, showing no significant variation (<50 ppmv). Therefore, it was assumed that the background CO₂ concentration remained constant during the tracer travel time (ranging from 10 minutes to 1.5 hours, depending on the site) and was subsequently subtracted from the raw data. The impact of anthropogenic exhaled CO₂ can be neglected (only two people, who left the site immediately after the injection). The maximum concentration at the injection point for each site was estimated through mass conservation. It ranged between 2·10³ mg/m³ (at Baulmes Mine) and 20·10³ (at Milandre Cave with the lowest flow rate).

A hand-held hotwire anemometer (Testo 425; device accuracy ± 0.03 m/s +5% of reading) was used to measure the flow in the conduits by integration of 6 measurements averaged over 30 s at each cross-section. Results are compared with airflow inferred from the CO₂ gauging, assumed constant during the tests.

5.2.3. Theoretical approaches

5.2.3.1. Airflow gauging

Tracing experiments are based on the dispersion and mass conservation of an artificial tracer in a fluid flow. The airflow rate can be derived by:

$$Q = \frac{m}{\int c(t)dt} \approx \frac{m}{\sum_{i=0}^k (c_{i(CO_2)} \cdot \Delta t)}$$

Eq. 5.1

where Q is the airflow rate [m³/s], m is the known mass of the injected tracer [g], c is the measured concentration [ppmv] at time t [s].

It is worth noting that Eq. 5.1 provides an accurate estimation of the flow rate Q if mass conservation is satisfied, i.e. if there is no leak between the injection point and the sensor. If this condition is not well-constrained in the field, Eq. 5.1 might yield wrong results.

The mean air speed U and the mean cross-sectional area A can be determined by a quantitative analysis of the BTCs. The mean velocity of the tracer (U_p) is obtained by dividing the travel distance by the time to the BTC peak (max value):

$$U_p = \frac{L}{t_{peak}}$$

Eq. 5.2

The mean residence time $\bar{t}$ [s], namely, the time required for the mass of the tracer to traverse the entire length L [m], for a short-pulse release is given by (Gaspar, 1987a modified by Field, 2002):

$$\bar{t} = \frac{\int_0^\infty t C(t) Q(t) dt}{\int_0^\infty C(t) Q(t) dt}$$

Eq. 5.3

For a constant flow rate Q , the volume V [m³] of air crossing the entire length L during the test is given by:

$$V = Q\bar{t}$$

Eq. 5.4

and the mean cross-section A [m²] along the length L is given by:

$$A = \frac{V}{L}$$

Eq. 5.5

5.2.3.2. Numerical evaluation of the BTCs

At a distance from the injection point much larger than the pipe diameter, the 1-D dispersion-advection equation reads (Taylor, 1954)

$$\frac{\partial C}{\partial t} + U \frac{\partial C}{\partial x} = D \left(\frac{\partial^2 C}{\partial x^2} \right)$$

Eq. 5.6

where C is the solute concentration averaged over the cross-section, U is the mean air velocity, t is the time and x represents the distance from the injection point. In the following, we consider the case of a single airway of constant cross-section A with a probe at a length L from the injection point where a mass m of tracer has been injected at time $t = 0$.

A long-time asymptotic solution of Eq. 5.6 at the probe location is given by (Taylor, 1954; Chatwin, 1971):

$$C(t) = \frac{m}{A\sqrt{4\pi Dt}} \exp \left[-\frac{(L - Ut)^2}{4Dt} \right]$$

Eq. 5.7

A necessary condition of validity of Eq. 5.7 is $t \gg D/U^2$. Eq. 5.7 is adopted independently for water (Field, 1999; Hauns et al., 2001; Field, 2002; Massei et al., 2006) and air (Widodo et al., 2008; Sasaki et al., 2009; Widiatmojo et al., 2015;). Eq. 5.7 depends only on the flow velocity and geometrical characteristics of the conduit. The concentration variations over the cross-section being much smaller than in the longitudinal direction, we assume that the local concentration provided by the probe is a good approximation of the cross-section averaged concentration C included in Eq. 5.6 and Eq. 5.7.

5.2.3.3. The Chatwin geometrical method to assess D

The geometrical method proposed by Chatwin (1971) is often adopted to assess the hydro(geo)logical dispersion coefficient D (Field, 1999; Hauns et al., 2001; Field, 2002; Massei et al., 2006). It consists in rewriting Eq. 5.7 as follows:

$$\sqrt{t_i \ln \left(\frac{C_{peak} \sqrt{t_{peak}}}{C_i \sqrt{t_i}} \right)} = \left| \frac{L}{2\sqrt{D}} - \frac{U}{2\sqrt{D}} t_i \right|$$

Eq. 5.8

where C_i is the concentration at the instant t_i , while C_{peak} is the peak concentration at the corresponding instant t_{peak} . If Eq. 5.7 is valid, plotting the left side of Eq. 5.8 against time yields two straight lines of equation:

$$y = \frac{L}{2\sqrt{D}} - \frac{U}{2\sqrt{D}} t$$

Eq. 5.9

which provides the dispersion coefficient D and the fluid mean velocity U by linear regression. In an ideal case $U = U_p$.

5.3. Results

5.3.1. CO₂ breakthrough curves

Instances of BTCs measured at the last station of each site are shown in Figure 5.6. The time t and the concentration C have been normalized by the coordinates of the concentration peak t_{peak} and C_{peak} , respectively. For the Baulmes test, points are quite symmetrically distributed around $t^* \equiv t/t_{peak} = 1$. Instead, BTCs from caves show a more asymmetrical shape, with a significant part of the distribution included in the time range from $t^* = 1.5$ to 2.5 . Because the external temperature did not vary significantly during the tests, the internal airflow is assumed constant and the extended tails likely result from the more irregular geometry of caves compared to Baulmes' artificial conduit (see Figure 5.5). In the case of Milandre Cave, the background concentration was not fully recovered because the CO₂ was released in a chamber (Figure 5.3) favouring tailing due to air recirculation and diffusion.

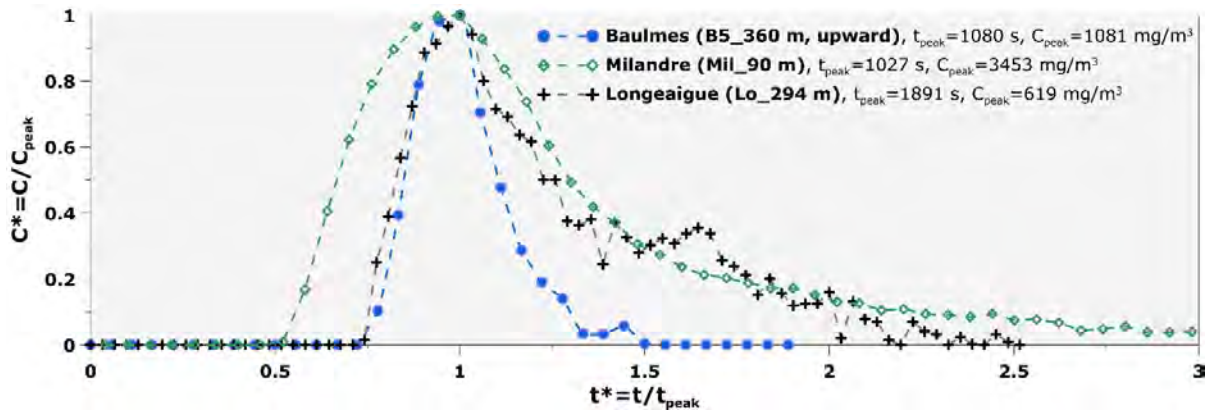


Figure 5.6 – Instances of CO₂ BTCs for the three investigated sites (normalized concentrations and times). All points are shown at a logging rate of 60 seconds.

5.3.2. Airflow rate and conduit geometry

To infer the airflow, tracer mass conservation is assumed for all the tests to solve Eq. 5.1. The results between gauged and directly measured airflow are compared in Table 5.1. The relative deviation between CO₂ gauged airflow and the direct measurements (Table 5.1) consistently stays below 5% in Baulmes. However, this difference expands to around 30% in Milandre and can rise to as much as 47% at the last Longeigue probe, between Lo_294 and Lo_278.

Site	Inj. mass	Station	Q_{CO_2} [m ³ /s]	Q_m [m ³ /s]	$(Q_{CO_2} - Q_m)/Q_{CO_2}$
Baulmes (upward) 15/02/2023 09:30	0.78 kg	St_B3	2.30	2.30	0.00
		St_B4	2.10	-	-
		St_B5	2.30	-	-
Baulmes (downward) 02/05/2023 17:08	1.00 kg	St_B3	3.80	4.00	0.05
		St_B2	3.70	-	-
		St_B1	3.60	-	-
Milandre 08/02/2022 12:20	1.00 kg	Mil_20	0.39	0.52	0.33
		Mil_90	0.35	-	-
Longeigue 13/12/2022 15:14	1.00 kg	Lo_239	1.50	1.76	0.17
		Lo_278	1.50	-	-
		Lo_294	1.50	0.79	-0.47

Table 5.1 – Gauged air flowrate Q_{CO_2} and manually measured air flowrate Q_m in Baulmes, Milandre and Longeigue.

From the analyses of the BTCs, it was possible to infer the average cross-sections (A_{CO_2}) at the gauging stations (Eq. 5.5). The twenty-eight surveyed cross-sections (Figure 5.5) were averaged (A_{avg}) for each station, enabling the comparison with CO₂-gauged cross-sections (A_{CO_2}). The results are presented in Table 5.2 for Baulmes Mine, Milandre and Longeigue Caves.

The differences between A_{avg} and A_{CO_2} at Baulmes Mine were within 1.5% and 3.2%, showing a good agreement between the inferred and surveyed cross-sections for both ventilation regimes. In Milandre Cave the disagreement between the two methods was higher. The average cross-section evaluated by the CO₂ gauging along the 90-meter-long conduit in Milandre (Mil_90) is smaller than the surveyed one by 20%. However, it is within the variability of this cave transect (st.dev. ± 4.9 m²). Lastly, cross-sections gauged in Longeigue show differences of up to 74% (Lo_294) from those obtained by the *in situ* survey (Table 5.2). Although even in Longeigue the relative difference between the two methods is within the variability of the surveyed cross-sections, the agreement between A_{avg} and A_{CO_2} is noticeably worse

compared to that of Milandre. This higher difference can be attributed to the presence of more complex cave morphologies (Figure 5.5 and Table 5.2) generating the long tail evident in the raw data from Longeague (Figure 5.6) which can influence the estimation of conduit volumes (Eq. 5.4) and cross-sections (Eq. 5.5).

Site	Station	$A_{avg} [m^2] \pm 1\sigma$	$A_{CO_2} [m^2]$	$(A_{CO_2} - A_{avg})/A_{avg}$
Baulmes (upward)	St_B3	6.4±1.8 (n=13)	6.5	0.02
	St_B4	6.3±1.4 (n=29)	6.5	0.03
	St_B5	6.6±1.4 (n=37)	6.7	0.02
Baulmes (downward)	St_B3	6.6±1.2 (n=25)	6.4	-0.03
	St_B2	6.5±1.1 (n=29)	6.5	0.00
	St_B1	6.6±1.4 (n=37)	6.5	-0.02
Milandre	Mil_20	10.8±6.6(n=3)	10.5	-0.03
	Mil_90	6.4±4.9(n=10)	5.0	-0.20
Longeague	Lo_239	7.5±4.1(n=25)	11.2	0.49
	Lo_278	7.2±4.1(n=27)	11.3	0.60
	Lo_294	6.8±4.2(n=29)	11.8	0.74

Table 5.2 – Comparison between the surveyed cross-sections together with results inferred from the CO₂ gauging at different sites. A_{avg} : average cross-sections from the injection point to the station; A_{CO_2} : average cross-section inferred from CO₂ tracer velocity.

5.3.3. Data analysis: dispersion coefficient D and curve simulations

U and D were fitted using Eq. 5.9 after plotting the left side of Eq. 5.8 as a function of time (see the insets in Figure 5.7). In Milandre and Longeague Caves, the model cannot reproduce the last part of the data sets where the tailing is accentuated. Therefore, this part of the curves (blue dots in the insets of Figure 5.7) was not considered in the fit. Conversely, the complete data sets of Baulmes were taken into account. In all cases, a high correlation between the selected data and the fitting lines was obtained ($R^2 \geq 0.97$, see Figure 5.7). Since Eq. 5.2 and Eq. 5.9, both give average airspeeds (U_p and U respectively), it was verified that the ratio $(U - U_p)/U_p$ never exceeds $\pm 2\%$, so that $U_p = U$ can be assumed. As expected from the fit of U and D , the simulated BTCs (Figure 5.7) match always well with the empirical data from Baulmes, whereas the simulation does not reproduce the last part of the BTCs in Longeague and Milandre, where the concentration decreases more slowly compared to an exponential model (Eq. 5.7).

Discussing the dispersion coefficients obtained in Milandre and Longeague is not meaningful, since they are estimated using Eq. 5.7, which fails to describe the BTCs. The dispersion coefficient D in Baulmes is of the order of 1 m²/s (see Table 5.3). It is slightly larger for downward flow, which is consistent with the higher air velocity during this test. Surprisingly, D slightly varies with the distance of the sensor from the injection point, for both tests (downward and upward flows).

Site	Station	L [m]	d [m]	U_p [m/s]	D [m ² /s]	D_T [m ² /s]	$\frac{D}{D_T}$
Baulmes (upward)	St_B3	128	2.6	0.36	1.39	[0.40 – 0.51]	[2.7 – 3.5]
	St_B4	300	2.6	0.36	0.83	[0.40 – 0.51]	[1.6 – 2.1]
	St_B5	360	2.6	0.36	0.87	[0.40 – 0.51]	[1.7 – 2.2]
Baulmes (downward)	St_B3	232	2.6	0.64	1.98	[0.64 – 0.82]	[2.4 – 3.1]
	St_B2	270	2.6	0.56	1.71	[0.64 – 0.82]	[2.1 – 2.7]
	St_B1	357	2.6	0.56	1.30	[0.64 – 0.82]	[1.6 – 2.0]
Milandre	Mil_20	20	2.6	0.13	0.64	-	-
	Mil_90	90	1.6	0.12	0.57	-	-
Longeaigne	Lo_239	239	2.6	0.18	1.41	-	-
	Lo_278	278	2.6	0.17	0.75	-	-
	Lo_294	294	2.4	0.16	0.70	-	-

Table 5.3 – Length of the conduit transects L, aeraulic diameters d ($d = 4 \cdot A/\text{perimeter}$, derived from surveyed cross-sections), peak velocity U_p , dispersion coefficients D inferred from field data, dispersion coefficients D_T predicted by the Taylor's theory (Eq. 5.5), the ratio D/D_T . The uncertainty range on D_T results from the uncertainty on the Darcy friction coefficient.

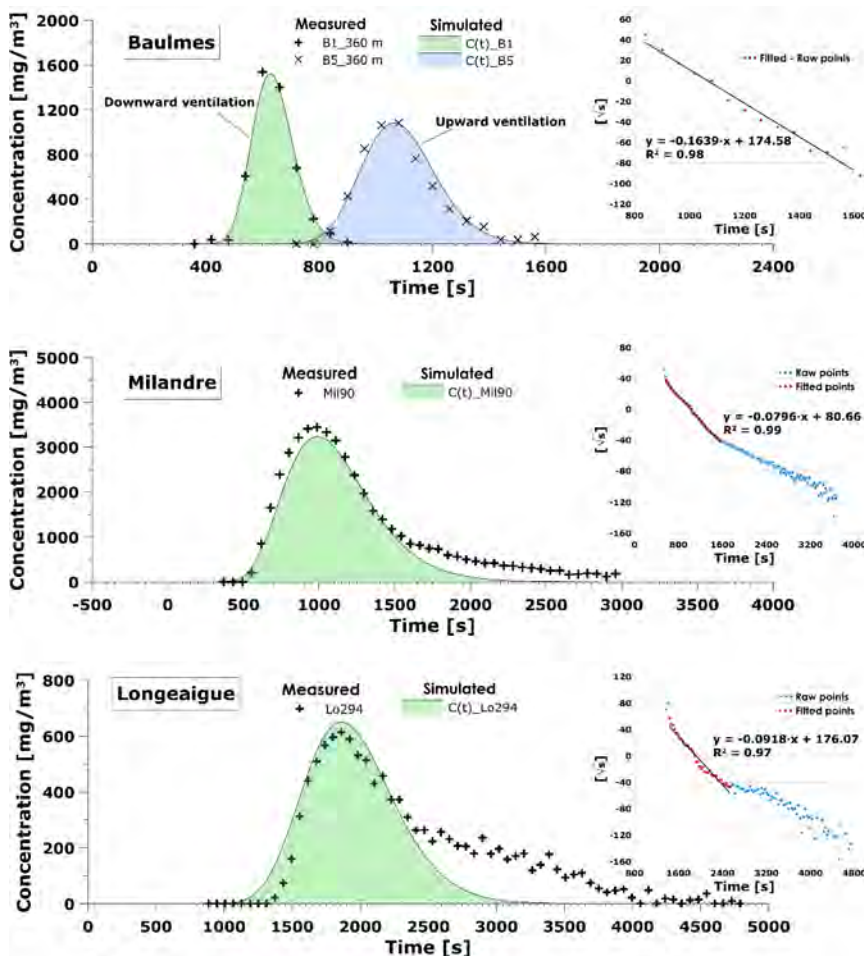


Figure 5.7 – Measured CO_2 concentrations (crosses) and simulated BTCs (Eq. 5.7) at each site. Results obtained at intermediary stations are shown in Table 5.3. Insets: left side of Eq.8 (dots) and fit using Eq.9 (solid line). In orange are the points selected for the fit.

5.4. Discussion

Four CO₂ tests were carried out in one mine (artificial conduit) and two ventilated caves (natural conduits), to investigate the reliability of the method for gauging airflow (Figure 5.6), assessing average cross-sections (Table 5.2) and inferring the dispersion value along ventilated karst conduits (Table 5.3 and Figure 5.7).

The gauging tests were consistent with most of the independent measurements performed manually with an anemometer. The difference between the gauged airflow and the manual measurements generally ranges between 0 and 33%. An exception is station Lo_294 in Longeague, where the difference reached 47% when the airflow was measured at 0.78 m³/s against 1.5 m³/s inferred from CO₂-gauging.

The first possible explanation could be related to the 3-hour difference between the manual measurement and the injection time, which was much higher than for the other sites. However, the temperature difference between the cave and the outside temperature, which drives the airflow inside the cave by a chimney effect, was larger at the moment of the manual measurement than during the CO₂ gauging. Therefore, the flow rate measured manually should have been higher than the flow rate obtained from CO₂ gauging, in contrast to what has been observed. Alternatively, the manually measured airflow may have been impacted by the presence of the operator in the 0.5 m² narrow conduit. Another possible reason for this deviation can be related to the assumption of mass conservation (Eq. 5.1) which could possibly not be correct if the system under investigation is not well-constrained and known. If in Milandre and Baulmes the absence of air divergences is warranted, it is not possible to say the same for Longeague. As shown in Table 5.1, the information of a possible leak (53% of the mass) is given by direct measurements, otherwise not visible only through the gauging test. As observed, the divergence located between Lo_239 and Lo_294 (dashed red arrow in Figure 5.4) could provoke sometimes a secondary airflow reducing the airflow at the upper entrance. More investigation will be necessary to clarify this point. To determine the airflow, the basic assumption of air mass conservation has to be verified, otherwise the gauged airflow may be unreliable. However, the presence of an air leak could be checked with the gauging technique by installing a logger along the possible leakage conduit, to confirm or not the passage of injected CO₂. Therefore, a good knowledge of the studied site is fundamental to make sure that mass conservation is assured.

Air velocity can be referred to as a local velocity or as an average velocity along a conduit. The first can be inferred by spot measurements with e.g. anemometers, while the second requires gauging tests and can be represented by BTC peak velocity (U_p in Table 5.3). In caves, punctual air velocity is usually measured in cross-sections where the air draft is sufficiently strong to be measured, i.e. in narrow passages. Therefore, spot measurements are unlikely to yield an average air velocity over the entire length of the cave. Moreover, high uncertainties are expected at low air velocities, such as when the cross-sections are large or in the presence of a small driving force (e.g. in caves with a small vertical distance between entrances or tiny ΔT between external and internal atmosphere) (Faimon and Lang, 2013; Gabrovšek, 2023). In particular, at low speeds, measurements can be significantly affected by instrumental accuracy (e.g. accuracy of ± 0.03 m/s compared to measurements of 0.15 m/s) and would require high-resolution monitoring with digital devices (Pastore et al., in prep.). Therefore, the CO₂ gauging test is a reliable method for assessing airflow rate under low-velocity conditions.

The regularity of the gallery in Baulmes (Figure 5.5) ensures a consistent result between the CO₂-inferred cross-sectional areas and the surveyed mine sections (Table 5.2), with an instrumental

accuracy on cross-section (A_{CO_2}) assessment between $\pm 5\%$ and $\pm 7\%$. In contrast, in Milandre and Longeague Caves, the difference between A_{avg} and A_{CO_2} reaches 20% and 74%, respectively. Although the surface areas inferred by the CO_2 gauging tests are included in the natural variability of the conduit geometries, the mismatch can be due to 1) the estimation of the mean tracer residence time $\bar{t}$ (Eq. 5.3) or 2) the low spatial resolution (10 m) of the *in situ* survey. In Longeague, sudden and systematic changes in diameter (Figure 5.4 and Figure 5.5), bends and air splitting (dashed rectangle in Figure 5.4) are present. Narrow passages whose cross-sections are between 0.5 m^2 and 1.1 m^2 were surveyed throughout the cave (Figure 5.5), and the largest cross-sections are found between these passages, forcing the air to continuously constrict and expand, causing tailing and lagging (high $\bar{t}$). Larger cross sections will slow down the transport by decreasing the air velocity, increasing the residence time $\bar{t}$. A higher $\bar{t}$ will be reflected in an overestimation of the average conduit volume (Eq. 5.4) and thus an overestimation of the average conduit cross-section (Eq. 5.5). Consequently, dead flow zones and air recirculation causing the tailing likely lead to a better representation of larger cross-sections. The second reason for the discrepancy may be that, in a cave, changes in the cross-sectional area occur at a resolution of less than a few meters and the average surface areas present a large standard deviation. Therefore, the ten-meter resolution surveyed at Milandre and Longeague may not be fully representative. For instance, the 90 m-long Milandre conduit is more variable than the Baulmes tunnel over 360 m, showing standard deviations of $\pm 4.9 \text{ m}^2$ and $\pm 1.8 \text{ m}^2$, respectively. In Longeague Cave, the variability of the cross-section is even more marked than in Milandre. The lower part of Longeague Cave (from the injection point to the station Lo_239) presents morphologies characterized by shafts, rockfalls, and narrow passages alternating with chambers and galleries. In contrast, the upper part of the cave is mainly characterized by rounded narrow passages (from Lo_239 to Lo_294). The number of surveyed cross-sections necessary for a representative average in a cave may be difficult to assess.

Simplifying the characterization of flow and dispersion in complex conduits is not straightforward. Gauging tests proved to yield a first piece of information about average cross-sections and average velocity along the conduit, both necessary for appropriate modeling. Using CO_2 as a tracer across different ventilated systems brings values for average cross-sections and conduit volumes, which can be compared to cave mapping. Cave cross-section errors brought about by the CO_2 tests along complex ventilated cave conduits can be reduced by carrying out CO_2 tests on shorter conduit transects, aiming to minimize the tailing effect. These geometrical parameters can allow the comparison of such values between different caves and could enable the definition of new morphometric indices such as the ratio of a system's volume to its average cross-sectional area and length (Piccini, 2011). These analyses are of particular interest in geoplanetology when comparing terrestrial lava tube morphologies and similar features observed on the Moon and Mars (Sauro et al., 2020; Bell et al., 2022;). Sauro et al. (2013) applied several methods to infer the average cross-section of terrestrial lava tubes, either through 3-D surveys or by extrapolating the mean height and width from speleological surveys. In the presence of airflow, CO_2 -gauged geometries could prove useful in advancing this new field of research.

5.4.1. Comparison with Taylor's theory

For Baulmes tests, it is interesting to compare the dispersion coefficients estimated from the field data with the prediction of Taylor's theory (since Taylor's theory implicitly assumes the validity of Eq. 5.7, comparisons with results from Milandre or Longeague would not make sense). The Reynolds number $Re = 4 \cdot Q_{CO_2} / p \cdot \nu$ (with p the conduit perimeter derived from surveyed cross-sections, and $\nu = 1.43 \times 10^{-5} \text{ m}^2/\text{s}$ the air kinematic viscosity) was of the order of 10^5 during Baulmes tests, clearly indicating turbulent airflow. Taylor's prediction for the turbulent regime reads (Taylor, 1954):

$$D_T = 5.05 d U^*$$

Eq. 5.10

where the friction velocity U^* is related to the mean velocity U_p by the relation (Widiatmojo et al., 2015):

$$U^* = U_p \times \sqrt{\frac{f}{8}}$$

Eq. 5.11

where f is the Darcy friction factor, which depends on Re and the relative roughness $\varepsilon_r = \varepsilon/d$ (with $d=4\cdot A/p$ the hydraulic diameter of the conduit). The wall's absolute roughness ε is not accurately known, but we estimate that it should be in the range of 10 to 30 cm. The Moody chart (Moody, 1944) yields the corresponding friction factors f ranging from 0.06 and 0.1. Entering these values in Eq. 5.11 and Eq. 5.10 provides the uncertainty range of D_T displayed in Table 5.3.

Our results reveal that the dispersion coefficient D estimated from Baulmes BTCs is systematically 1.6 to 3.5 times higher than the prediction from Taylor's theory (see Figure 5.8 and Table 5.3). Instrumental uncertainties, which represent $\pm 15\%$ of D , cannot explain these discrepancies. Moreover, Taylor's theory is known to be accurate for $Re \gtrsim 2 \times 10^4$, a condition that is verified since we obtained $Re \sim 10^5$ during the tests. A first possible explanation for the discrepancy between the theoretical D_T and experimental D could be that Baulmes Tunnel#10 is not strictly circular (see Figure 5.1), as assumed in Taylor's theory. Moreover, the section of Baulmes Mine, although much more regular than caves, significantly varies throughout the conduit (see Figure 5.5). Assessing the consequences of these geometrical "defects" is not possible with the available data. Our results are nonetheless consistent with those obtained by Widiatmojo et al. (2015) (Figure 5.8), which also identified enhanced dispersion along mine conduits compared to Taylor's predictions. These authors suggested that the high dispersion coefficients measured in the Kushiro mine could be related to the presence of cavities along the mine tunnels, which trap the tracer and reintroduce it into the main flow with a delay. The absence of such cavities along the Baulmes Tunnel#10 is consistent with a lower dispersion coefficient, closer to Taylor's theory.

Taylor's theory is relevant neither in caves nor in mines, even though the conduit geometry is more regular in the latter case. However, a 1-D simulation based on Eq. 5.6 is still possible in mines, using empirical dispersion coefficients deduced from field data (see Baulmes results in Figure 5.7). Note that in this site, the inferred D at the last station (B1 or B5 according to the flow direction) is intended to represent the average D value for the entire conduit length ($L=360$ m). However, because D decreases systematically with increasing L , if a D value corresponding to a shorter length than L (e.g. to B3 or B2) is assigned for the simulation, this will result in an overestimation of the peak concentration.

BTC analysis displayed an accentuated tailing along the investigated natural conduits due to the presence of dispersive singularities which possibly enable aerosol deposition. For example, the long tailing observed in Milandre and Longeague Caves is probably due to cross-section variations. Additionally, the 1-D advection-dispersion model was found to be unable to fit BTC tailing. Along natural caves, airways encounter bends, bottlenecks, collapses, variable wall roughness, speleothems, and other structural factors which are absent in the artificial conduit of Baulmes Mine. This is a key point

that will likely impact the air movement and consequently the airborne aerosol distribution and settlement. Therefore, a more complex model is required. Adesso et al. (2022) showed that, in caves, changes in fluid dynamics occur where these singularities are present, resulting in the deposition of airborne particles. Therefore, the tailing observed in CO₂ tests could suggest that, in the case of airborne matter, an increased tailing corresponds to greater sedimentation along the investigated conduit. Hauns et al. (2001) highlighted that retardation could result from the presence of dispersive structures if those have a storage time at the same order of magnitude as the duration of the tracer BTC.

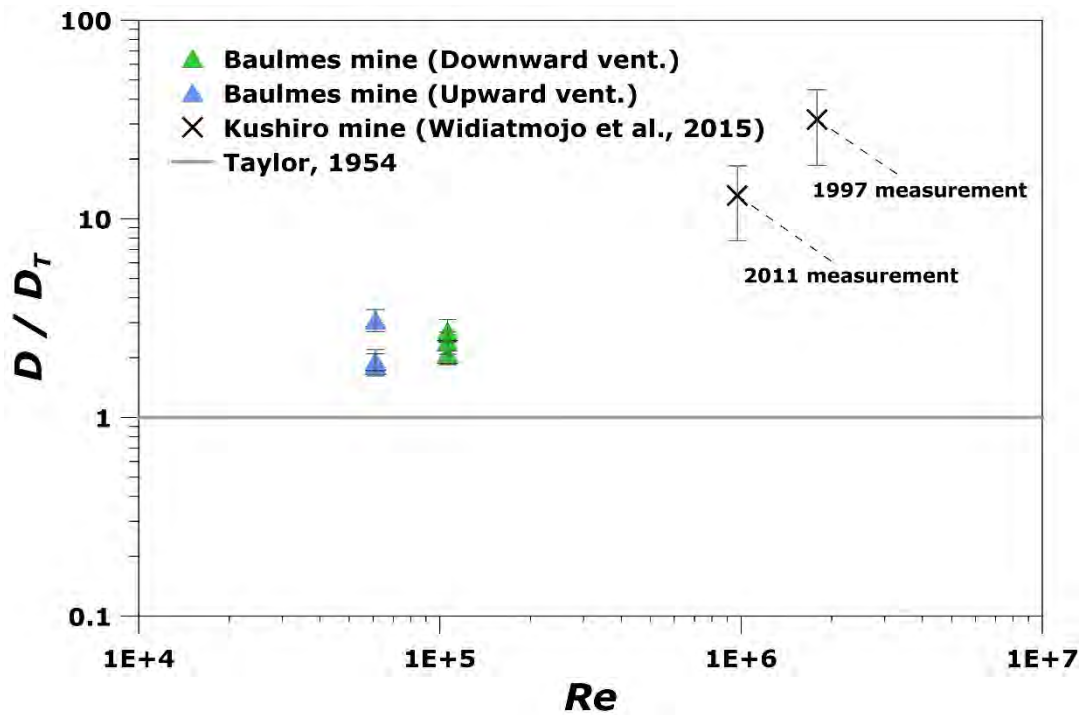


Figure 5.8 – Comparison between the dispersion coefficient obtained from BTCs and Taylor's theory for the Baulmes Mine (this work) and Kushiro mine (Widiatmojo et al., 2015). The vertical error bars correspond to the uncertainty on Darcy's friction factor required to estimate D_T .

Therefore, tailing duration allows an estimate of how long matter can be trapped in low-velocity structures. Stokes' law reads that areas of slow flow (dead flow zones) are favourable to the deposition of aerosols transported in a fluid (e.g. air). Dead flow zones cause the deposition of suspended matter and the quantity of deposited matter is somehow proportional to the tailing duration. Hence, along natural karst conduits, the variable geometry and presence of obstacles lead to dead flow zones. Those dead flow zones (e.g. chambers) retard the transport producing tailing and modify the dispersion with distance. For instance, in the case of Longeague, the tailing suggests that larger chambers with a low air velocity are more prone to deposition than narrower ventilated conduits.

5.5. Conclusions

In the karst vadose zone, cave ventilation is an efficient mode of transport for heat and aerosols and drives the spatial distribution of airborne particles in a cave system. Modeling this airborne transport requires geometrical and physical parameters of the conduit system, including the cross-section areas, the airflow and average air speed, as well as the dispersion coefficient which describes the spread of a solute. The four experiments carried out in natural and artificial conduits, by using CO₂ as a tracer, showed to be a suitable technique to infer the airflow and average conduit cross-sections. The value of

dispersion has also been inferred by applying Chatwin's method. BTC analysis displayed an accentuated tailing along the investigated natural conduits due to the presence of dispersive singularities which possibly enable aerosol deposition. The long tailing observed in Milandre and Longeigue Caves is probably due to cross-section variations, therefore chambers are likely prone to deposition. In contrast with Longeigue and Milandre Caves, the 1-D advection-dispersion model correctly reproduces the shape of the BTCs in the Baulmes Mine. However, Taylor's (1954) theory significantly underestimates the dispersion coefficient deduced from the BTCs. Numerical simulations of 3-D Navier-Stokes equations would be interesting to investigate solute dispersion in caves and mines, in particular, to assess the role of the conduit geometry. It was already attempted for water flow (Hauns et al., 2001). Similar works should be done for airflow.

This paper gives a first insight into air motion and matter dispersion along ventilated karst conduits, preparing for proper aerosol dispersion modeling.

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Authorship statement

CP, ML, and PYJ designed and directed the study. EW developed the electronic devices used for the measurements. CP performed the measurements and analyzed the data. CP and FD carried out the simulations. CP wrote the paper with input from all authors.

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Chapter 6. Comparing different methods for measuring cave airflow (conference paper)

Pastore, C., Luetscher, M., Jeannin, P.-Y., 2025. Comparing different methods for measuring cave airflow, in: Proceedings of the 19th ICS. Presented at the 19th International Congress of Speleology, Sociedade Brasileira de Espeleologia, Belo Horizonte, Minas Gerais, Brazil, pp. 239–244.

Abstract

Precise measurements of airflow within caves are increasingly demanded to assess heat and mass transfers and their impacts on the karst environment, including subsurface ecosystems, hydrochemistry of karst water and secondary mineral precipitates. This study compares different methods for measuring airflow in ventilated caves. Airflow estimates from gauging tests using an artificial tracer are compared with direct measurements from hand-held anemometers and digital airflow meters equipped with dataloggers. The tests generally yield similar results, although the comparison reveals some discrepancies, with differences of up to 50% in small conduit cross-sections. Deployment of three digital flowmeters along a 300 m ventilated cave indicated either a potential air leak or more likely the upper and lower bounds of the estimated airflow. This comparison underlines that the use of different methods for estimating airflow should be a welcome practice when monitoring ventilated caves, to derive a range of plausible airflow rates that accurately reflect actual conditions.

6.1. Introduction

Caves are complex and dynamic environments. They are characterised by airflows that influence the distribution of heat, gases, and particles therein. Precise monitoring of these flows is crucial for understanding the hydrochemistry of karst waters (Jeannin et al., 2017), the formation of secondary minerals (Spötl et al., 2005), the distribution of subterranean ecosystems (Mammola et al., 2019) and paleoenvironmental reconstructions (Fairchild and Baker, 2012; Rau et al., 2015). Cave ventilation, often driven by the chimney effect (Badino, 1995; Lismonde, 2002; Luetscher and Jeannin, 2004; Gabrovšek, 2023), plays a fundamental role in regulating temperature and thermal exchanges, thus long airflow data series in caves are of interest to assess e.g. the associated energy exchanged therein (Luetscher et al., 2008; Obleitner & Spötl, 2011). Various anemometer types are available, including analogue, hotwire, and ultrasonic models, each differing in cost and measurement quality. While suited to specific monitoring ranges, they have limitations: analogue devices suffer from low resolution and mechanical inertia, hotwire anemometers often lack airflow direction detection (except for the so-called X-anemometers and triple-hot-wire anemometers, which can measure velocity vectorial components; Barros da Guia and Marques, 2018; Legg et al., 1984), and both hotwire and ultrasonic models require a high power supply. Nowadays, ultrasonic anemometers are widely used in caves (e.g. Pflitsch and Piasecki, 2003; Obleitner and Spötl, 2011; Gomell and Pflitsch, 2022); however, they could suffer cave climatological conditions (temperature, humidity and lack of sunlight), failing during long-term monitoring. Nevertheless, technological advancements continue to improve their performance (Pastore et al., 2024a; Wang et al., 2021; Fan et al., 2018; Leoni et al., 2018).

Airflow in ventilated caves can be assessed with direct measurements (i.e. with anemometers) or with gauging tests using a tracer gas. Direct measurements with anemometers provide the air velocity at specific locations. To estimate the volume flow rate (m^3/s), these velocities need to be integrated over the cross-section or the cave passage. In Pastore et al. (2024a) it was demonstrated that this procedure may be associated with uncertainties related to the velocity distribution across the cross-section. In a

second work by Pastore et al. (2024b), CO₂ was used as an artificial tracer and it was demonstrated to be a reliable method for assessing airflow rates within ventilated conduits. Therefore, the comparison between different methodologies to assess the airflow rate is essential for understanding the uncertainties associated with each approach and for obtaining accurate estimates of airflow. This work aims to compare results obtained using gauging tests with CO₂ as a tracer (Pastore et al., 2024b), and flow rates estimated with both a hand-held anemometer and a digital flowmeter used for long-term monitoring (Pastore et al., 2024a), adding insights into the comparison between these different methods.

6.2. Materials and Methods

The first method applied was gauging the airflow by injecting CO₂ as a tracer. This method relies on the mass conservation equation: $Q = \frac{m}{\int c(t)dt}$ [Eq.1], where Q is the airflow rate [m³/s], m is the known mass of the injected tracer [g], and c is the measured concentration [ppmv] at time t [s]. More details about the method applied to the airflow are given in Pastore et al. (2024b). To assess the airflow, tracer mass conservation is always assumed to solve Eq. 1 (no gas leakage along the conduit). The second method consisted of manual measurements of flow velocity across the conduit cross-sections, carried out with a Testo 425 hand-held hotwire anemometer (from now on “Testo”; accuracy ± 0.03 m/s +5% of reading). The volume flow rate (m³/s) is derived by the integration from 3 to 6 velocity measurements (based on the cross-sectional surface area) averaged over 30 s per monitored cross-section. As a third method, we installed a digital device such as the Sensirion SFM3003-300-CE (from now on “SFM”) flow meter, which allows recording airflow time series. In Pastore et al. (2024a), we provided a reliable calibration for this device and a rating curve to convert the measured Standard Litre per Minute (SLM) into m/s. The accuracy of this instrument is $\pm 5\%$ of the measured value for velocities above 0.50 m/s and $\pm 26\%$ for velocities < 0.50 m/s. The high instrumental precision ($\pm 1\%$) and reproducibility were verified between different SFM devices (Pastore et al., 2024a). The SFM can be connected to a data logger and deployed to record the flow speed at a specific location in the conduit. Thus, the volumetric flow rate (m³/s) can be derived from the integration of the measured airspeed in the conduit cross-section.

External temperatures were measured with a ReefNet Sensus Ultra device (accuracy ± 0.3 °C, resolution 0.01 °C).

The tests were conducted in two different sites, which are naturally ventilated conduits (chimney effect): 1 old mine and 1 cave (the latter in Figure 6.1), both in the Swiss Jura Mountains (the test dates, cross-section sizes and distances from the CO₂ injection point of each station are given in Table 6.1).

The first test site is Baulmes Mines, where we carried out spot measurements with Testo and two gauging tests with CO₂ (during upward and downward ventilation, see Table 6.1). The measurements were carried out at station B3 (Table 6.1), located within a 360 m-long, sub-horizontal tunnel that provides access to the lowest mine level before connecting to the main mining network. Indeed, this tunnel serves as a lower entrance for the chimney effect. The geometry is typical of a mining tunnel: flat ground (except for some local collapses), vertical walls, a barrel-vaulted ceiling and homogeneous cross-sectional areas. No air inlets or outlets are observed along the selected tunnel, except for the main entrance. The survey and more details of this testing site can be found in Pastore et al. (2024b).

The second test site is Longeague Cave, where we conducted two gauging tests and measurements using both the Testo and SFM devices. We carried out supplementary measurements with Testo and SFM without performing CO₂ tests (see the last two periods in Table 6.1). Longeague Cave is a vertical cave (Figure 6.1). Here, we carried out the gauging tests and measurements at three stations along the

300 m-long conduit that connects the upper and lower entrances. The stations are close to the cave's upper entrance (station Lo_UE), lower entrance (station Lo_LE) and in a cave transect between the previous (station Lo_M) (Figure 6.1 and Table 6.1). Moreover, a perennial lake between the two entrances, whose level is controlled by external precipitations, temporarily stops the airflow when the level rises high enough to close the air passage (mainly in winter).

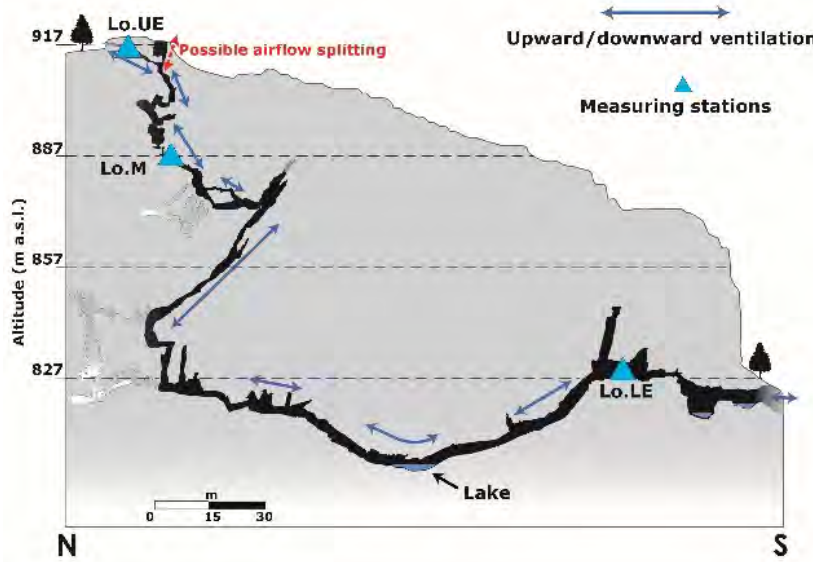


Figure 6.1. An excerpt of the Longeague Cave survey, showing the ventilated conduit (300 m long and 100 m deep). Modified after Jeannin (2018) and Pastore et al. (2024b).

6.3. Results

The SFM measurements are available only for Longeague. The relative deviation between SFM and CO₂ tests is between 27% and 45% (Table 6.1). Between SFM and Testo measurements, the deviation constantly stays below 30% (Table 6.1), showing sometimes higher values for SFM (negative % in Table 6.1) and sometimes for Testo (positive %). Figure 6.2 presents the time series obtained with SFMs at the three monitoring stations in Longeague Cave compared with the flow rates estimated with gauging tests and the Testo device at the same locations, in the periods 16 to 22 of June 2022 (Figure 6.2.a, downward flow) and 10 to 12 of December 2022 (Figure 6.2.b, upward flow). The positive and negative values on the flow rate indicate downward and upward ventilation, respectively. For both ventilation modes, the airflow time series show similar patterns but with different amplitudes between the three stations and also strong daily fluctuations following the external temperatures during the downward flow (Figure 6.2.a, c,d). In Figure 6.2.a, the flow rate at Lo_UE (orange line) shows constantly lower values ($0 < Q < 1.3 \text{ m}^3/\text{s}$) than the other two devices. In contrast, the device positioned at Lo_LE (blue line) generally displays the highest values ($0.3 < Q < 1.5 \text{ m}^3/\text{s}$), while the device placed at Lo_M (green line) measured an airflow generally comprised between the previous two. However, certain flow peaks in the SFM time series and manual measurements at station Lo_M exceed those recorded at station Lo_LE, reaching $1.6 \text{ m}^3/\text{s}$ at Lo_M during the measurement period shown in Figure 6.2.a.

Site	Station	Distance [m]	Cross-section [m ²]	Q _{CO2} [m ³ /s]	Q _{Testo} [m ³ /s]	Q _{SFM} [m ³ /s]	(Q _{Testo} -Q _{CO2})/Q _{CO2}	(Q _{SFM} -Q _{CO2})/Q _{CO2}	(Q _{SFM} -Q _{Testo})/Q _{Testo}
Baulmes (upward) 15/02/2023	B3	128	4.5	2.3	2.3	-	0%	-	-
Baulmes (downward) 02/05/2023	B3	232	4.5	3.8	4	-	5%	-	-
Longeigue (upward) 13/12/2022	Lo_UE	294	0.5	1.5	0.79	1.00	-47%	-33%	27%
	Lo_M	239	1.2	1.5	1.76	-	17%	-	-
	Lo_LE	0	3.1	-	1.63	1.50	-	-	-8%
Longeigue (downward) 17/06/2022	Lo_UE	0	0.5	-	0.89	0.90	-	-	1%
	Lo_M	55	1.2	1.1	1.64	1.60	49%	45%	-2%
	Lo_LE	20	3.1	1.1	1.27	1.40	15%	27%	10%
Longeigue (downward) 09/08/2022	Lo_UE	0	0.5	-	0.64	0.51	-	-	-20%
	Lo_M	55	1.2	-	1.10	1.18	-	-	7%
	Lo_LE	20	3.1	-	1.40	1.29	-	-	-8%
Longeigue (downward) 09/06/2023	Lo_UE	0	0.5	-	0.69	0.84	-	-	22%
	Lo_M	55	1.2	-	0.87	-	-	-	-
	Lo_LE	20	3.1	-	0.97	0.70	-	-	-28%

Table 6.1. The grey columns show the geometrical characteristics of the monitoring stations. Orange columns show the airflow rate estimated with the three methods. Green columns show the ratio between them.

During the upward flow (Figure 6.2.b), periods constantly at 0 m³/s at the beginning and end of the graph are due to the lake closure during external rainfall (Figure 6.1). Therefore, the upward ventilation was active for about 4 days, between Dec. 10th 2022 and Dec. 14th 2022, before the lake closed the air passage again. When the ventilation was active, the airflow rate at Lo_UE stabilised between 1.3 and 1.1 m³/s, while at Lo_LE oscillated between 1.5 and 1.1 m³/s. The flow rate of 1.5 m³/s estimated at Lo_LE with manual measurements fits with SFM time series and with the airflow yielded by CO₂ tests along the cave. The absence of daily fluctuations relates to low fluctuations of external temperatures, which stayed between -11°C and -6 °C (Figure 6.2.b). Figures 2.c and 2.d show two supplementary periods (see the last two periods in Table 6.1) where manual measurements are compared with the SFM time series at the three stations in Longeigue. The measurements confirm that the airflow estimated at Lo_LE is generally the highest, in contrast with the lowest airflow estimated at Lo_UE, while measurements from Lo_M are in between.

6.4. Discussion

The tests conducted aimed to compare different methods for measuring cave airflow. The chosen methods are gauging tests with an artificial tracer (CO₂ in this case), and measurements with both a hand-held anemometer (Testo) and in-place digital flowmeters (SFM). These tests were carried out in one cave (Longeigue) and one mining tunnel (Baulmes), both naturally ventilated by a chimney effect (Pastore et al., 2024b). Gauging tests are a reliable method to assess volume flow rates along ventilated conduits (Pastore et al., 2024b; Widiatmojo et al., 2015; Standish, 1988; Thirnonsl and Kissel, 1974). The method does not require an integration of the velocity over a cross-section, as the airflow is derived from the mass conservation equation [Eq.1] and yields reliable flow rates even at low air speeds. In the following, flow rates assessed with this method, when available, will be used as reference values. The results indicate that the measurements performed manually (Testo) and those obtained with the digital flow meter (SFM) are comparable, with differences between the two that stayed below 30% (last column of Table 6.1). In contrast, the comparison between the gauging tests and most of the direct measurements at Longeigue Cave reveals more significant discrepancies than in Baulmes (Table 6.1). Moreover, time series from SFM and manual measurements (Figure 6.2), seem to indicate an increase in the flow rate from the upper (Lo_UE, Figure 6.1) to the lower station in the cave (Lo_LE, Figure 6.1). Different reasons may determine these differences. The first possible reason for the deviation between

direct measurements and CO₂ tests can be related to an unknown inlet/outlet along the conduit, or in other words that the assumption of air mass conservation along the conduit [Eq.1] is not true. To determine the airflow with gauging tests, this basic assumption has to be verified, otherwise the gauged airflow may be unreliable. If in Baulmes the absence of air divergences is warranted, it is not possible to say the same for Longeigue. The relative difference between direct measurements (SFM and Testo) and CO₂ gauging tests (Table 6.1) shows values that suggest a possible air leak between the upper and lower entrances. Our field observations identified a divergence located between Lo_M and Lo_UE (red arrow in Figure 6.1) which may sometimes provoke a secondary airflow reducing/increasing the airflow between these two stations. This possible leak would remain undetected through the CO₂ tracing alone since mass conservation is assumed. The presence of an air splitting could be checked with the air tracing if a logger is deployed along the possible leakage, to confirm or not the passage of the injected tracer. Further investigation will be necessary to clarify this point.

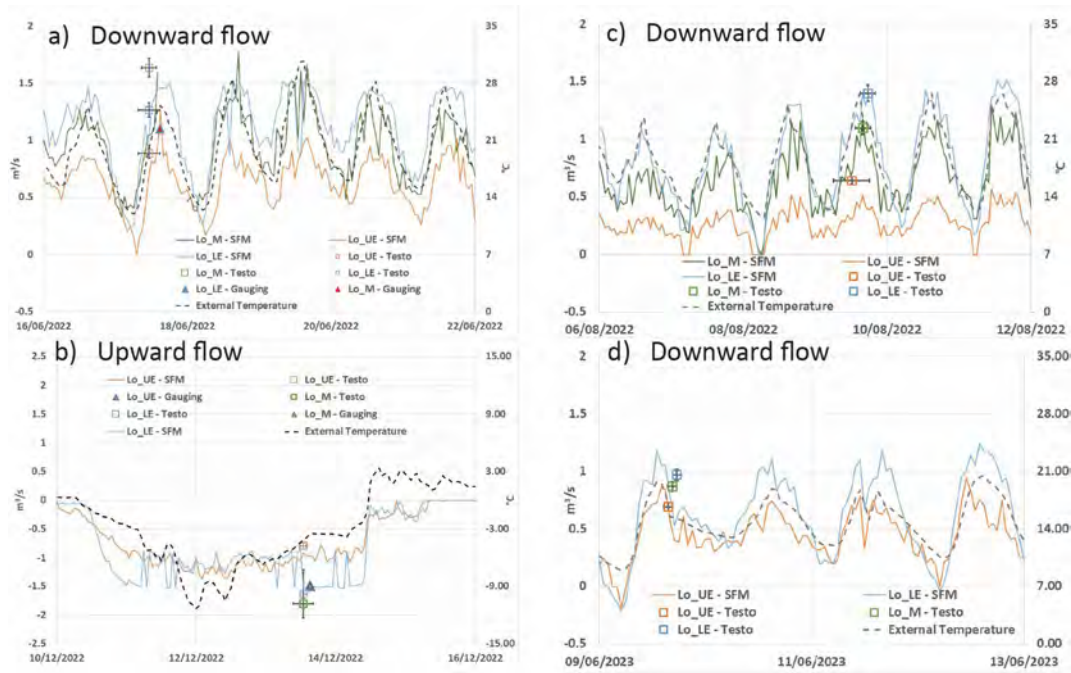


Figure 6.2. Airflow rates estimated with the three methods at the three stations in Longeigue Cave during a) downward and b) upward ventilation. In c) and d) airflow rates were estimated only with Testo and SFM at the three stations during downward ventilation. Dates are dd/mm/yyyy.

When working with anemometers (e.g. Testo) or flowmeters (e.g. SFM), the volume flow rate is calculated based on the estimation of an average air velocity derived from one or several punctual measurements within the cross-section. Therefore, another reason possibly causing the discrepancy between gauging tests and direct measurements involves errors stemming from the estimation of this average. In caves, the spatial variability of the cross-section can be significant and a turbulent flow is typically observed (Lismonde, 2002). Thus, in a turbulent field, the stochastic error related to the sampling may impact the measurements (Finkelstein and Sims, 2001; Mauder and Zeeman, 2018). Moreover, because of the nonuniformity of the cave walls, the distribution of velocity fields varies as the flow rate and its direction change, creating high and low-speed zones (Pastore et al., 2024a). Manual measurements that have shown sometimes higher (49%) and sometimes lower (-47%) values than those from gauging tests were observed in Longeigue Cave at Lo_M and Lo_Ue, respectively. In contrast, estimations made at Lo_LE and Baulmes mines have shown smaller differences (Table 6.1). During measurements with hand-held devices (e.g. Testo), the presence of the operator in narrow conduits may affect the flow velocity within the conduit. The stations Lo_UE and Lo_M, are located within crawling passages with reduced cross-sections (0.5 and 1.2 m², respectively; Table 6.1).

Performing measurements at these stations obliges the operator to lie down on the conduit floor, possibly affecting the measurements, although this is likely to cause a lower flow estimate, as for Lo_UE, rather than higher values, as for Lo_M. Instead, stations Lo_LE and B3 have much larger cross-sections (3.1 and 4.5 m², Table 6.1) and may be less affected by the presence of the operator who can stay outside the main airflow.

Another explanation for differences between CO₂ tests and direct measurements (with SFM and Testo) could be related to the mismatching time between the manual measurements (horizontal error bars in Figure 6.2), the gauging tests and SFM logging times. The temperature difference between the cave and the outside temperature, which drives the airflow inside the cave by a chimney effect (Pastore et al., 2024b), may change during the experiment likely causing a transient flow rate. The low temperature variation outside the cave shown in Figure 6.2.b does, however, not support this possibility for the upward flow (Pastore et al., 2024b). Instead, the external temperature varied and increased during the duration of the measurements (3 to 4 hours) shown in Figure 6.2.a/c/d (downward flow). We always took measurements going from Lo_UE to Lo_M, then to LO_LE. The external temperature variations during the time needed to move from one station to the other may explain the different airflow measured at the three stations. In fact, since these manual measurements were taken at subsequent times, they should be seen as a part of a time series. The increasing airflow on the manual measurements is supported by the actual airflow time series from SFM devices (Figure 6.2), which also show peaks at the same time.

However, there must be another reason explaining the fact that SFM time series from different parts of the cave show different airflow amplitudes at the same time. Comparing time series from the three stations along the cave (Figure 6.2) airflow increases up to 50% from Lo_UE (upper station) and Lo_LE (lower station). This could indicate again the presence of a supplementary inlet/outlet as discussed before. However, it also appears that the airflow rate measured at Lo_M is in between the Lo_UE and Lo-LE flow rates, and does not support the presence of an unknown inlet along the cave. As stated before, stations Lo_UE and Lo_M are crawling passages. Here, SFM devices were positioned at a few centimetres from the conduit walls to facilitate the transit of cavers. Instead, the device deployed at station Lo_LE was positioned in the middle of the 3.1 m² cross-section. This suggests that the instruments close to the wall (Lo_UE and Lo_M) could have measured a lower velocity than the actual average speed of the cross-section, while the SFM positioned at Lo_LE was far from the walls, and could have yielded higher values for the average flow rate. If, with a hand-held device, one can take measurements at several points across the cross-section to average the flow velocity, the location of a single device is crucial for providing a good assessment of the average air velocity.

Long airflow data series in caves are of interest to study e.g. the air dynamics (e.g. Faimon & Lang, 2013; Gomell & Pflitsch, 2022) and the associated energy exchanged therein (Luetscher et al., 2008; Obleitner & Spötl, 2011). When an instrument is installed at a specific point within the cross-section, it is important to note that the device may be measuring an airspeed that does not necessarily represent the average velocity across the entire cross-section (Pastore et al., 2024a). In Pastore et al. (2024a), we deployed simultaneously three SFMs within a 1.2 m² cross-section (the same location as Lo_M). We have shown with field data and a numerical model that the position of the instruments within the same cross-section can produce high uncertainties (up to ±50%) in the estimation of the airflow. Moreover, if an anemometer is placed in a high-speed zone, there is a high possibility of overestimating the airflow. Conversely, positioning it close to the wall may result in an underestimation of the flow.

Therefore, by comparing the results from the different tests, the airflow estimated close to the wall at Lo_UE is likely to represent the lower uncertainty bound, while the airflow estimated in the centre of Lo_LE cross-section represents the higher uncertainty bound of the actual cave airflow.

6.5. Conclusion

The main conclusion of this study is that measurements of airflow quantity in caves are difficult and that uncertainty is usually in the order of ± 30 to $\pm 50\%$. Each of the methodologies applied for measuring an airflow (gauging tests, spot measurements with a hand-held anemometer and airflow time series) offers specific advantages depending on the type of measurement required. Spot measurements can help to identify airflow differences along a conduit due to airflow deviations (i.e. airflow splitting into different flow paths) since they can be carried out all along a ventilated conduit, whereas airflow time series provide insights into airflow dynamics (e.g. airflow variations and direction related to external conditions). Among these, the gauging test with a tracer is likely the most accurate method to assess a one-time airflow rate since it does not require velocity integration over a cross-section. Nevertheless, the assumption of air mass conservation can be a limitation of the method, especially over long distances and in complex systems, where it may be difficult to verify this requirement. Despite a general consistency in measured airflow among the three methods, uncertainties inherent to each technique affect accuracy. Factors such as instrument placement, external variability (e.g., temperature fluctuations), the number of measurement points used to average air velocity, and environmental constraints (e.g. small cross-sections) contribute to these uncertainties. As a result, reaching an uncertainty lower than 20–30%, requires significant and dedicated efforts. For instance, in the case of manual measurements, one approach could be to increase the number of measurement points to assess an average velocity value that is as representative as possible of the velocity fields across the cross-section. It is crucial to cover the entire cross-section and avoid disturbing the airflow with the operator's presence. Moreover, to decrease the stochastic error due to flow turbulences, punctual measurements of air velocity should last long enough to compensate for this error due to turbulence (e.g. at least 30 sec. per measuring point). Another approach could be to measure airflow in three or more different cross-sections along the same conduit. Another key challenge is selecting an appropriate instrument position for long-term monitoring. The field tests and the 3D model of air velocity distributions within a cave cross-section presented by Pastore et al. (2024a) suggest that there is no clear position to deploy a single anemometer within a cross-section. To achieve greater precision, multiple devices should be used in a single cross-section to assess the average air velocity. If only one instrument is available, the measurement point should be selected after conducting manual measurements with a hand-held anemometer to assess the velocity distribution. The fixed anemometer should be placed in a velocity zone that closely represents the average velocity, avoiding both high-speed and low-speed areas. Gauging tests with an artificial tracer, performed under varying airflow conditions (i.e. different directions and flow rates), can provide a rating curve for in-place devices to account for site-specific discrepancies. Long-term monitoring with a single instrument is unlikely to provide a fully accurate airflow rate (Pastore et al., 2024a), so it is crucial to determine a plausible range of airflow uncertainties.

High-precision airflow measurements are challenging to achieve in caves, not only due to methodological uncertainties but also because of the difficulty in precisely determining the instrument's position within the velocity fields of a measuring cross-section. Therefore, comparing different methods can help establish the order of magnitude of the flow and its variations. This approach will allow for the estimation of upper and lower bounds of the flow rates, which are essential for the calculation of e.g. energy balance and aerosol transport modeling.

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PART 2

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Cave ventilation dynamics and heat exchange

Chapter 7. Modeling heat transfer for assessing the convection length in ventilated caves (published paper as second author)

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<https://doi.org/10.1029/2024JF007646>

Author statement: My work for this paper involved: Data curation, Methodology and Investigation in the field

Abstract

The present study focuses on heat transfer in ventilated caves for which the airflow is driven by the temperature contrast between the cave and the external atmosphere. We use a numerical model that couples the convective heat transfer due to the airflow in a single karst conduit with the conductive heat transfer in the rock mass. Assuming dry air and a simplified geometry, we investigate the propagation of thermal perturbations inside the karst massif. We perform a parametric study to identify general trends regarding the effect of the air flowrate and conduit size on the amplitude and spatial extent of thermal perturbations. Numerical results support the partition of a cave into three regions: (1) a short (few meters) diffusive region, where heat mainly propagates from the external atmosphere by conduction in the rock mass; (2) a convective region where heat is mainly transported by the air flow; (3) a deep karst region characterized by quasi-constant temperatures throughout the year. Numerical simulations show that the length of the convective region is approximately proportional to the amplitude of the flowrate annual fluctuations divided by the square root of the cave radius. This result is tested against field data from a mine tunnel and two caves. Our study provides first estimates to identify climate sensitive regions for speleothem science and/or ecosystemic studies.

7.1. Introduction

Understanding heat transfer in karst systems is a key issue for underground biota (Mammola et al., 2019), preservation of cave art (Bourges et al., 2014), speleothem growth rates (Spötl et al., 2005; Banner et al., 2007), or paleoclimate reconstruction (Borsato et al., 2016; Casteel and Banner, 2015; Domínguez-Villar et al., 2021). However, heat transfer in karst results from an intricate coupling between several mechanisms, including heat conduction in the rock mass (Quindos et al., 1987), convection due to water or air flow in caves (Cropley, 1965), or radiative transfer between cave walls (Guerrier et al., 2019). Latent heat exchanges are also present because of evaporation and condensation on cave walls (Dreybrodt et al., 2005) or ice formation and permafrost (Luetscher et al., 2008). A key issue for the understanding of heat transfer in karst is to determine the processes that dominate in a given configuration, and those that can be neglected.

The simplest heat transfer model of a vadose karst assumes 1D conduction heat transfer in a rock mass of infinite depth with periodic temperature fluctuations at the ground surface (Villar et al. 1983; Domínguez-Villar et al. 2023; Salmon et al., 2023). This model predicts that the depth corresponding to 99% attenuation of the temperature fluctuations at the ground surface is a few tens of centimeters for the daily fluctuations and approximately 10 m for the annual fluctuations. Quindos and co-workers (Villar et al., 1983; Quindos et al., 1987) measured the monthly mean temperatures in Altamira Cave during a year. They found that the periodic 1D conduction model correctly predicts the amplitude and

the phase shift of the annual fluctuations at the cave roof whose depth varied from 3.5 m to 17.5 m. In contrast, the floor temperature was much closer to the roof temperature than expected from a 1D conduction model. The authors attributed this effect to the homogenization of the temperature field by radiative transfer between roofs and floors. Guerrier et al. (2019) and Qaddah et al. (2023) confirmed by numerical simulations the significance of radiative transfer inside weakly ventilated shallow caves, and showed that turbulent free convection also had to be considered. Both heat transfer mechanisms contribute to the homogenization of the wall temperature and the deformation of the temperature field in the rock matrix around the cave.

The situation is different when forced convection is present. It has long been known that, in large cave systems, the convective transport of heat by air currents and streams can propagate thermal perturbations over much larger distances than predicted by a pure conduction model. Cropley (1965) found that temperature measurements performed in two caves in West Virginia supported the definition of three zones: zone I immediately adjacent to the cave entrance, where the cave temperature follows the surface temperature fluctuations; zone II where the cave temperature is driven by streams and air currents; zone III where the temperature is approximately constant and close to the annual mean temperature at the surface. Areas where the amplitude of temperature fluctuations was lower than 1°F (ca. 0.56°C) around the annual mean temperature were located at more than 1500 m from the cave entrances. This long distance was mainly due to cold water carried into the caves by winter water recharge. Luetscher and Jeannin (2004) defined the heterothermic vadose zone as the surficial part of a karst massif showing seasonal inversions of the temperature gradient, in contrast with the homothermic vadose zone characterized by a temperature gradient close to the lapse rate of humid air (ca. 0.5°C 100 m⁻¹). These authors estimated the depth of the heterothermic zone at about 100 m in most cases.

Considerable work has been undertaken to simulate heat transfer by forced convection due to water flow in the vadose and phreatic zones of karst (Gong et al., 2019; Long and Gilcrease, 2009; Sinokrot and Stefan, 1993). Covington et al. (2011) showed that the relative significance of different heat fluxes including convection and conduction are timescale dependent. Conduction through the rock surrounding a conduit determines heat flux at times of the order of weeks and longer. Coupling convection in water with conduction in rock is thus necessary to get a realistic model valid at all time scales.

Forced convection due to airflow must be considered when a network of karst conduits has at least two entrances. In this configuration, several mechanisms can contribute to the production of airflow through the cave, resulting in significant convective heat transfer between the cave and the external atmosphere. Potential driving forces are barometric fluctuations (Gomell et al., 2021), dynamic pressure effect due to external winds (Kukuljan et al., 2021), diphasic flow due to water circulation (Atkinson, 1977), or the buoyancy due to the density contrast between the air inside and outside of the cave (Gabrovšek, 2023). The latter mechanism is likely the most significant in most cases. The air density depends on moisture content, CO₂ concentration and temperature. In temperate climates, temperature fluctuations are the main cause of density variations (Gabrovšek, 2023). Therefore, the flow direction is upward when the temperature inside the cave is warmer (or downward when inside the cave is colder) than the external atmosphere. Because of the thermal inertia of the massif, a lower entrance operates as an inlet of cold external air during most of the winter, whereas it acts as an outlet during most of the summer. In the latter case, the air cooled down when crossing the massif before getting out the lower entrance. The lower entrance thus receives in summer an airflow colder than the external temperature. Because of this seasonal asymmetry, the annual mean temperature is shifted to colder values at lower entrances (cold thermal anomaly). The same mechanism produces a shift of the annual mean temperature to hotter values at upper entrances (hot thermal anomaly) (Lismonde,

2002). In alpine ice caves, where this ventilation pattern represents the normal unless the cave is clogged with sediments, these thermal anomalies usually extend over a few hundred meters from the cave entrances (Luetscher et al., 2008).

Wigley and Brown (1971) calculated the temperature and moisture content profiles in the air as a function of the distance from the entrance. They used a 1D model based on the energy and water mass conservation in the airflow. The convective heat transfer coefficient between the air and the wall was estimated using an empirical correlation valid for forced convection in smooth pipes. The model assumes prescribed uniform wall temperature, a major simplification that allows to obtain simple closed-form expressions for the temperature and moisture profiles. However, this approach implicitly assumes that the heat flux through the cave wall is convection-limited, and that the rock mass does not play any role. Lismonde (2002) pointed out that the airflow modifies the rock temperature, which in turn changes the air temperature profile in the cave. This author developed a model to predict heat transfer in a straight inclined duct of constant diameter included in a rock massif. The model coupled 1D radial conduction in the rock mass with convective transfer in the gas, using a sinusoidal function of time as inlet temperature. The airflow rate was predicted by considering the interaction between the temperature field in the gas and the airflow through the buoyancy term in the momentum balance equation of the gas. This model reproduced qualitatively some field observations, e.g. the hysteresis of the flow rate or the thermal anomaly in the entrance areas. Gabrovšek (2023) used a similar model to investigate by numerical simulations the effect of conduit shape on the airflow pattern in ventilated caves.

The present study focuses on the numerical simulation of heat transfer in ventilated caves for which the airflow is driven by the temperature contrast between the cave and the external atmosphere. Our objective is to go beyond the qualitative results obtained from previous numerical studies available in the literature. We aim at assessing orders of magnitudes of the spatial extent of the karst region where the airflow induces significant thermal perturbations. We use for that a numerical model close to that already developed by Lismonde (2002) or Gabrovšek (2023), and we apply it to the simplest possible configuration in order to quantify general trends common to most ventilated caves. The convective heat transfer due to dry airflow in a single horizontal karst conduit of constant diameter is coupled with the conductive heat transfer in the impermeable rock mass.

This article contains a first theoretical part (sections 2-5) dedicated to numerical simulations based on a model including many simplifying assumptions. The relevance of these assumptions to achieve our objective, which consists in the assessment of general orders of magnitude, is evaluated in a second part (section 6) by comparison with the field data obtained from three sites, a mine tunnel and two caves.

7.2. Model definition

7.2.1. Cave geometry and computational domain

We consider the idealized ventilated cave displayed in Figure 7.1, located in a rock massif of infinite extent in both vertical directions. The cave consists of a vertical conduit connected by two horizontal conduits to two entrances at different elevations. The rock mass is assumed impermeable. The horizontal conduits are supposed long enough for the vertical conduit to be fully included in the homothermic area where the temperature gradient reduces to the adiabatic lapse rate. The aerologic and thermal problems are uncoupled in this simplified configuration, since the temperature field in the horizontal conduits have no effect on buoyancy. The air flowrate thus only depends on the temperature contrast between the temperature in the vertical conduit, independent of time, and the temperature of the atmosphere outside the cave, a known function of time.

Another source of simplification is that the thermal problems in the regions of upper and lower entrances are uncoupled. They can thus be treated separately. We arbitrarily focus on the upper entrance, but all the results can be easily translated to a lower entrance (the only difference is that the air flows through the upper entrance inward during summer and outward during winter, and vice-versa through the lower entrance).

The computational domain is a cylinder of length L_{dom} and outer radius R_{dom} displayed in red in Figure 7.1. It contains a conduit of same axis and length and constant radius $R \ll R_{dom}$, inside which the air circulates. The length L_{dom} and outer radius R_{dom} are set so that the computational domain includes the whole thermal perturbation induced in the rock mass by the airflow. Practically, in a given configuration, L_{dom} and R_{dom} are increased until they no longer influence the results.

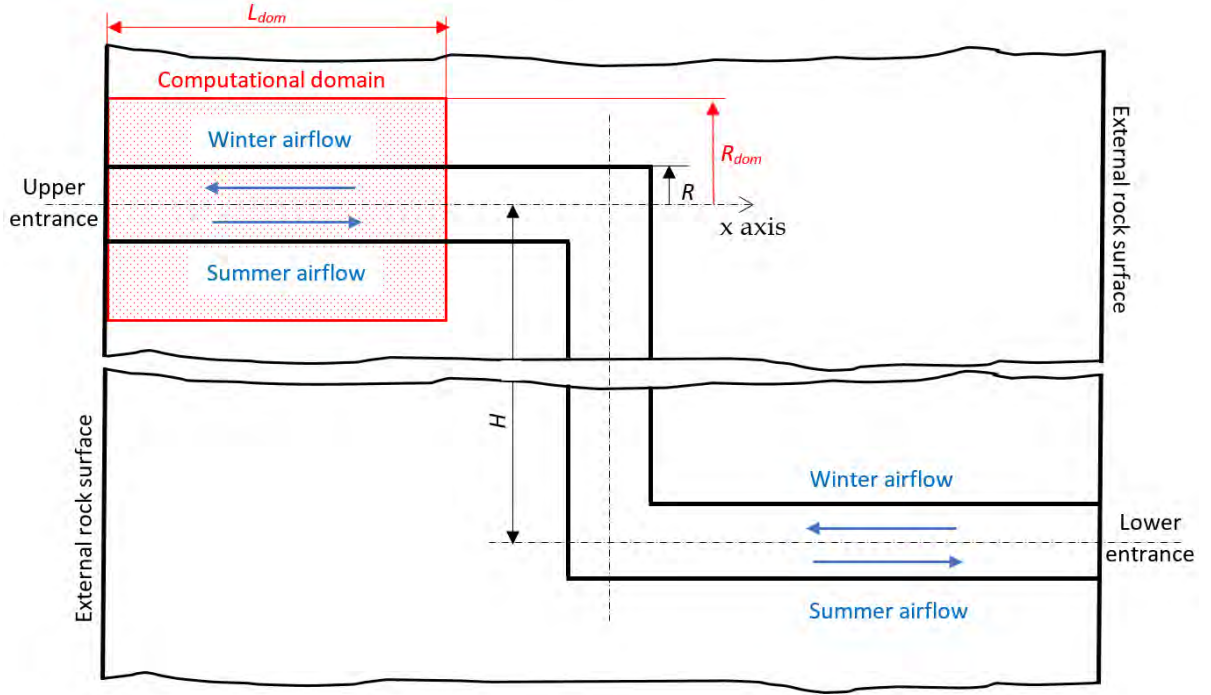


Figure 7.1. Cave geometry (the computational domain is displayed in red). The rock mass is assumed infinite in both vertical directions.

7.2.2. Atmosphere temperature and mass flow rate

For the sake of simplification, only the annual fluctuation of the atmosphere temperature T_{atm} is considered. It is thus assumed that $T_{atm}(t)$ follows the simple law:

$$T_{atm}(t) = T_m + \Delta T_y \sin\left(2\pi \frac{t}{\tau_y}\right) \quad (1)$$

where T_m and ΔT_y are the annual mean temperature (AMT) and the amplitude of the annual temperature fluctuation (ATF), respectively, in the external atmosphere at the elevation of the upper entrance. The corresponding period is $\tau_y = 1$ year.

The air mass flow rate $\dot{m}$ is deduced from the momentum balance applied to the air inside the cave. Assuming negligible inertia, the balance between friction and buoyancy yields (Lismonde, 2002):

$$K \dot{m} |\dot{m}| = -(\bar{\rho}_{atm} - \bar{\rho}_m)gH, \quad (2)$$

where H is the length of the vertical conduit and g the gravitational acceleration. The cave aerodynamic resistance K can be assumed constant in the turbulent regime. $\bar{\rho}_{atm}$ and $\bar{\rho}_m$ are the densities of the air

outside the cave and inside the homothermic zone, respectively, averaged over the cave height. Assuming that: (1) the difference of densities depends only on the temperature drop between the external atmosphere and the homothermic zone, (2) the temperature field in the vertical conduit follows the AMT in the external atmosphere, we get after linearization:

$$\bar{\rho}_{atm}(t) - \bar{\rho}_m = -\frac{M_a P_{atm}}{R_g T_m^2} (T_{atm}(t) - T_m) \quad (3)$$

where P_{atm} is the atmospheric pressure, M_a the molar mass of air and R_g the ideal gas constant. Eqs. (2) and (3) show that the air flowrate is proportional to the square root of the temperature drop ($T_{atm} - T_m$). Injecting Eqs. (1) and (3) in Eq. (2) yields

$$\dot{m}(t) = S \Delta \dot{m} \sqrt{\left| \sin\left(\frac{2\pi}{\tau_y} t\right) \right|} \quad (4)$$

where $S=-1$ for $T_{atm} < T_m$ (wintertime, upward flow) and $S=1$ for $T_{atm} > T_m$ (summertime, downward flow). The positive constant $\Delta \dot{m}$ is the amplitude of the annual fluctuation of the flow rate. It is an input of the model.

7.2.3. Governing equations

The air temperature verifies the energy balance

$$\dot{m} c_{p,a} \frac{\partial T_a}{\partial x} = \varphi_w P \quad (5)$$

where $c_{p,a}$ is the air heat capacity at constant pressure, T_a is the mixing temperature of the air inside the conduit, x the distance from the external rock surface, P the conduit perimeter ($P = 2\pi R$ for a circular cross-section) and φ_w the conductive flux at the conduit wall, positive when going from the rock to the air. Equation (5) is a balance between the energy advected by the air flow (LHS) and the conduction flux at the conduit wall (RHS). The air thermal inertia has been neglected compared to advection (quasi-steady approximation). Equation (5) requires a single boundary condition at the conduit inlet located at $x=0$ for downward flow ($S=1$) or at $x=L_{dom}$ for upward flow ($S=-1$). We thus impose the boundary condition

$$T_a(x = 0, t) = T_{atm}(t) \text{ for } S=1 \quad \text{or} \quad T_a(x = L_{dom}, t) = T_m \text{ for } S=-1 \quad (6)$$

The conductive flux φ_w is required to solve Eq. (5). The conduction equation must therefore be solved in the impermeable rock matrix around the cave. The transient 2D axisymmetric conduction equation reads

$$\frac{1}{r} \frac{\partial}{\partial r} \left(r \frac{\partial T_r}{\partial r} \right) + \frac{\partial^2 T_r}{\partial x^2} = \frac{1}{\alpha_r} \frac{\partial T_r}{\partial t} \quad (7)$$

where $T_r(x, r, t)$ is the rock temperature, α_r the rock thermal diffusivity, and r the distance from the cave axis. The boundary conditions are as follows. The atmosphere temperature is imposed on the external rock surface:

$$T_r(x = 0, r, t) = T_{atm}(t) \text{ for } R < r < R_{dom} \quad (8)$$

The boundary at $r=R_{dom}$ can be assumed adiabatic because the temperature field only depends on x far from the cave (i.e., for a large value of R_{dom}), and the radial component of the temperature gradient is thus close to zero. The boundary condition at $x=L_{dom}$ is also adiabatic, because this boundary is located in the homothermic zone. We get

$$\frac{\partial T_r}{\partial r}(x, R_{dom}, t) = 0 \text{ for } 0 < x < L_{dom} \text{ and } \frac{\partial T_r}{\partial x}(L_{dom}, r, t) = 0 \text{ for } R < r < R_{dom} \quad (9)$$

The last boundary condition is at the conduit wall. It is given by Newton's law of cooling and the heat flux continuity at the conduit wall:

$$\varphi_w = \lambda_r \frac{\partial T_r}{\partial r}(x, R, t) = h_{th}(t)(T_r(x, R, t) - T_a(x, t)) \quad (10)$$

where λ_r is the rock thermal conductivity, $T_r(x, R, t) = T_w(x, t)$ is the temperature of the conduit wall and h_{th} the heat transfer coefficient. The latter is time-dependent since it depends on the air flowrate $\dot{m}(t)$. All the conditions at the boundaries of the rock domain are displayed in Figure 7.2. No initial condition is required because we are looking for the periodic regime, i.e., the solution asymptotically reached by the model at infinite time.

7.2.4. Determination of the heat transfer coefficient

The heat transfer coefficient h_{th} in Eq. (10) is a key parameter. The estimation of h_{th} must therefore be performed with great care to get a reliable model. Assuming forced convection in conduits, Wigley and Brown (1971) and Gabrovšek (2023) used a correlation close to the Colburn correlation (Bergman et al., 2017), valid for fully developed turbulent flow in smooth pipes. Lismonde (2002) pointed out that a cave cannot be considered as a smooth pipe, and multiplied by two the numerical prefactor of the Colburn correlation to account for the effect of wall roughness. Covington et al (2011) used the Gnielinski correlation (Bergman et al., 2017), which explicitly considers the effect of wall roughness.

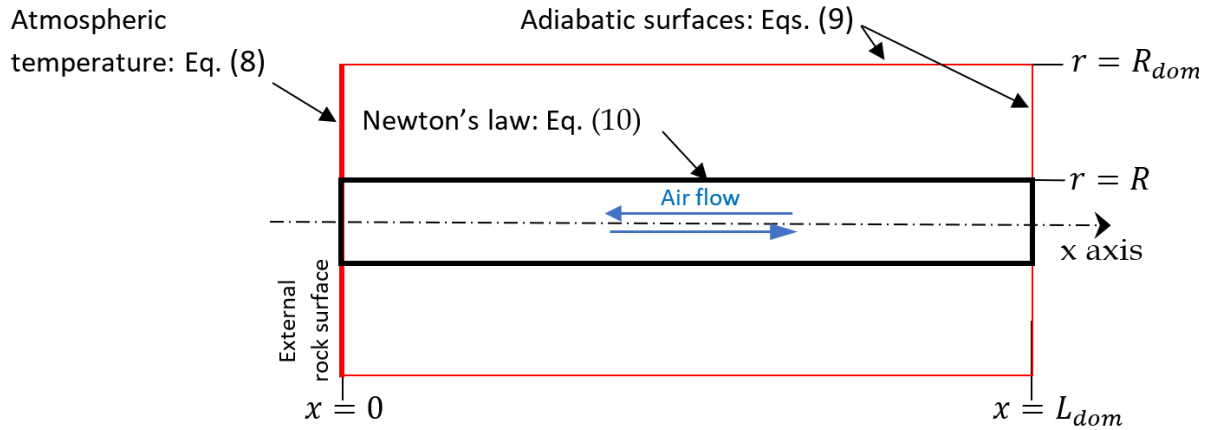


Figure 7.2. Rock domain with the boundary conditions of the heat conduction equation (7).

However, it is unlikely that a cave can be considered as a rough pipe. Indeed, caves can have complicated shapes, ranging from sub-circular conduits with several meters in diameter to complex cave passages associated with the collapse of the host-rock. A succession of bends, conduit contractions or enlargements, or obstacles of any kind are expected to increase the transfer of heat between the air and the wall. Indeed, all of these singularities not only increase the heat exchange surface between the wall and the fluid, but also enhance the heat transfer coefficient by generating secondary flows and by increasing the turbulence level. It is well known that heat exchanger performance can be improved by the insertion of twisted tapes, longitudinal fins or helical ribs inside the tubes (Bergman et al., 2017). For instance, spirally corrugated tubes can increase the heat transfer coefficient by a factor 3 (Promthaisong et al., 2016; Pethkool et al., 2011). However, although empirical correlations exist to predict heat transfer coefficient for a wide range of well-defined geometries used in heat exchangers, no correlation is available for the irregular and tortuous geometries commonly encountered in caves.

To get around this difficulty, two distinct cases will be considered. In the first case (case A in the following), correlations for forced convection and fully developed flow in pipes will be used to estimate the heat transfer coefficient. This approach is expected to give a lower bound of the heat transfer coefficient, and thus a lower bound of the heat flux at the conduit wall. In the second case (case B in the following), a higher bound of the wall heat flux will be obtained assuming infinite heat transfer coefficient. Reality must lie between these two limiting cases.

Case A: forced convection and fully developed flow.

The Nusselt number $Nu = \frac{h_{th} D_h}{k_a}$ used for the estimation of h_{th} is displayed in Figure 7.3 as a function of the Reynolds number $Re_t = \frac{V(t) D_h}{\nu_a}$, for three values of the relative roughness ε (ratio of the wall roughness over D_h). k_a and ν_a are the conductivity and the kinematic viscosity of the air, V is the mean air velocity at time t , $D_h = 4A/P$ is the hydraulic diameter with A and P the cross-sectional area and the perimeter of the conduit. Nu is computed as follows:

$$Nu = \max(Nu_L, Nu_T) \quad (11)$$

where $Nu_L = 3.66$ is the Nusselt number in the laminar regime (independent of the roughness) and Nu_T is given by the Gnielinski correlation (Bergman et al., 2017), which takes into account the effect of the wall roughness expected in the turbulent regime:

$$Nu_T = \frac{\left(\frac{f_d}{8}\right) (Re_t - 1000) Pr}{1 + 12.7 \left(\frac{f_d}{8}\right)^{0.5} (Pr^{2/3} - 1)} \quad (12)$$

where Pr is the air Prandtl number ($Pr=0.71$). f_d is the Darcy friction factor which depends on the Reynolds number and the wall relative roughness ε . f_d was estimated using the Haaland correlation (Haaland, 1983):

$$\frac{1}{\sqrt{f_d}} = -1.8 \log \left[\left(\frac{\varepsilon}{3.7} \right)^{1.11} + \frac{6.9}{Re_t} \right] \quad (13)$$

All the simulations of case A were done using $\varepsilon = 0.01$ as a lower bound of the relative roughness in a cave (red curve in Figure 7.3).

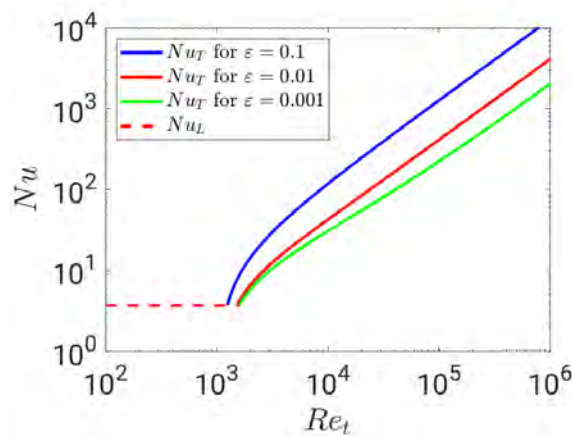


Figure 7.3. Nusselt number as a function of the Reynolds number (Prandtl number: $Pr=0.71$). The red curve was used for the simulations of case A.

Case B: Infinite heat transfer coefficient

In this case, the conduit wall temperature is equal to the bulk air temperature and Eq.(10) thus reduces to

$$\varphi_w = \lambda_r \frac{\partial T_r}{\partial r}(x, R, t) \quad \text{and} \quad T_r(x, R, t) = T_a(x, t) \quad (14)$$

7.2.5. Physical properties

All the physical properties are gathered in Table 7.1. They are assumed constant and estimated at the temperature $T_m = 12^\circ\text{C}$.

Table 7.1 : Thermophysical properties (estimated at temperature $T_m = 285 \text{ K} = 12^\circ\text{C}$ and atmospheric pressure $p_{ref} = 101325 \text{ Pa}$).

properties	Value and unit	Reference
Conduit relative roughness	$\varepsilon = 0.01$	(see section 2.4)
Rock density	$\rho_r = 2320 \frac{\text{kg}}{\text{m}^3}$	(Covington et al., 2011)
Rock heat capacity	$c_{p,r} = 810 \frac{\text{J}}{\text{kgK}}$	(Covington et al., 2011)
Rock thermal conductivity	$k_r = 1.656 \frac{\text{W}}{\text{m.K}}$	(Guerrier et al., 2019)
Rock thermal diffusivity	$\alpha_r = \frac{k_r}{\rho_r c_{p,r}} = 8.81 \times 10^{-7} \frac{\text{m}^2}{\text{s}}$	-
Air dynamic viscosity	$\mu_a = 1.77 \times 10^{-5} \text{ Pa.s}$	(Bergman et al., 2017)
Air density	$\rho_a = 1.23 \frac{\text{kg}}{\text{m}^3}$	(Bergman et al., 2017)
Air kinematic viscosity	$\nu_a = \frac{\mu_a}{\rho_a} = 1.43 \times 10^{-5} \text{ m}^2/\text{s}$	-
Air thermal conductivity	$k_a = 0.0251 \frac{\text{W}}{\text{m.K}}$	(Bergman et al., 2017)
Air Prandtl number	$Pr = \frac{\nu_a}{\alpha_a} = 0.71$	-
Molar mass of air	$M_a = 28.97 \frac{\text{g}}{\text{mol}}$	-
Ideal gas constant	$R_g = 8.314 \frac{\text{J}}{\text{molK}}$	-

7.2.5.1. Dimensionless equations

We define the dimensionless temperatures $\theta_a = (T_a - T_m)/\Delta T_y$ and $\theta_r = (T_r - T_m)/\Delta T_y$ in the air and the rock, respectively. The dimensionless temperature 0 thus corresponds to the external annual mean temperature and 1 is the amplitude of the external annual fluctuations. We take the period $\tau_y = 1$ year as the unit of time and the diffusion length $L_{dif} = \sqrt{\alpha_r \tau_y} \approx 5.3 \text{ m}$ as the unit of length. The dimensionless atmospheric temperature θ_{atm} and air flowrate μ are deduced from Eqs.(1) and (4):

$$\theta_{atm}(\tilde{t}) = \sin(2\pi\tilde{t}) \quad (15)$$

$$\mu(\tilde{t}) = \frac{\dot{m}(t)}{\Delta \dot{m}} = S\sqrt{|\sin(2\pi\tilde{t})|} \quad (16)$$

The dimensionless counterparts of Eqs. (5)-(10) read

$$\mu(\tilde{t}) Re \frac{\partial \theta_a}{\partial \tilde{x}} = \left(4 \frac{k_r}{k_a} Pr^{-1}\right) \tilde{\varphi}_w, \quad (17)$$

$$\theta_a(\tilde{x} = 0, \tilde{t}) = \sin(2\pi \tilde{t}) \text{ for } S=1 \quad \text{or} \quad \theta_a(\tilde{x} = \tilde{L}_{dom}, \tilde{t}) = 0 \text{ for } S=-1, \quad (18)$$

$$\frac{1}{\tilde{r}} \frac{\partial}{\partial \tilde{r}} \left(\tilde{r} \frac{\partial \theta_r}{\partial \tilde{r}} \right) + \frac{\partial^2 \theta_r}{\partial \tilde{x}^2} = \frac{\partial \theta_r}{\partial \tilde{t}}, \quad (19)$$

$$\theta_r(\tilde{x} = 0, \tilde{r}, \tilde{t}) = \sin(2\pi \tilde{t}) \text{ for } \tilde{R} < \tilde{r} < \tilde{R}_{dom}, \quad (20)$$

$$\frac{\partial \theta_r}{\partial \tilde{r}}(\tilde{x}, \tilde{R}_{dom}, \tilde{t}) = 0 \text{ for } 0 < \tilde{x} < \tilde{L}_{dom} \text{ and} \quad (21)$$

$$\frac{\partial \theta_r}{\partial \tilde{x}}(\tilde{L}_{dom}, \tilde{r}, \tilde{t}) = 0 \text{ for } \tilde{R} < \tilde{r} < \tilde{R}_{dom},$$

$$\tilde{\varphi}_w = \frac{\partial \theta_r}{\partial \tilde{r}}(\tilde{x}, \tilde{R}, \tilde{t}) = \eta(\tilde{t}) Bi \left(\theta_r(\tilde{x}, \tilde{R}, \tilde{t}) - \theta_a(\tilde{x}, \tilde{t}) \right), \quad (22)$$

where the tilde ($\sim$) indicates dimensionless time or lengths. $\eta(\tilde{t}) = \frac{h_{th}(\tilde{t})}{H_{th}}$ is the heat transfer coefficient scaled by H_{th} , its maximum value reached for $\dot{m} = \Delta \dot{m}$. The Reynolds number $Re = \frac{4 \Delta \dot{m}}{P \mu_a}$ (with μ_a the air dynamic viscosity) and the Biot number $Bi = \frac{H_{th} \sqrt{\alpha_r \tau_y}}{k_r}$ are based on the maximum values $\Delta \dot{m}$ and H_{th} , respectively. $\mu(\tilde{t}) Re$ and $\eta(\tilde{t}) Bi$ are the Reynolds and Biot numbers at a given time $\tilde{t}$. In a given configuration, $\tilde{R}_{dom}$ and $\tilde{L}_{dom}$ that define the size of the computational domain are increased until they do not modify the results, as stated in section 2.1.

With the constant physical properties defined in Table 4.1, a configuration is well-defined if only two dimensionless numbers are known: the dimensionless conduit radius $\tilde{R}$ and the Reynolds number Re . In case A, i.e., when fully developed forced convection is assumed for the estimation of the heat transfer coefficient, the Biot number in Eq. (22) is related to the Nusselt number by the simple relation $\eta(\tilde{t}) Bi = \frac{Nu}{\tilde{R}} \left(\frac{k_a}{2 k_r} \right)$, where Nu depends on $Re_{\tilde{t}} = \mu(\tilde{t}) Re$ through Eqs.(11)-(12). In case B, the Biot number is infinite, i.e., $\eta(\tilde{t}) Bi \rightarrow \infty$. Eq. (22) thus reduces to

$$\tilde{\varphi}_w = \frac{\partial \theta_r}{\partial \tilde{r}}(\tilde{x}, \tilde{R}, \tilde{t}) \quad \text{and} \quad \theta_r(\tilde{x}, \tilde{R}, \tilde{t}) = \theta_a(\tilde{x}, \tilde{t}) \quad (23)$$

7.3. Method of resolution

We are looking for the periodic solution of the mathematical model defined in section 2.6. The periodic regime could be obtained by time integration starting from an arbitrary initial condition, after simulating a number of cycles large enough to get a good approximation of the solution at infinite time. However, the slow convergence of this method can generate large computational times. We thus prefer to use another method, based on Fourier series. All time-varying variables, including the temperatures in the rock and in the air, are approximated by truncated Fourier series and inserted in the mathematical model of section 2.6. This yields a set of coupled partial differential equations (PDE), which does not include the time. The numerical resolution by the Galerkin method of this set of PDE gives the spatial distribution of the amplitude and phase of the Fourier modes. The temporal evolution of the dependent variables is then recovered from Fourier series. This method is detailed in supplemental material. The validation by comparison with a standard time integration method is presented in the same appendix. All the numerical simulations were performed with the commercial software Comsol Multiphysics, version 6.1.

The method of resolution based on Fourier series used in this work is particularly relevant for the simplified time variation of the external temperature considered in Eq.(1). Considering all the Fourier modes included in real times series of the external temperature would clearly increase both the complexity of the method implementation and the computational resources required for the simulations. In that case, only the most significant modes of the input temperature should be considered. In contrast, using Fourier series to simulate the complex geometry of real caves would not cause any particular problem.

7.4. Characterization of the thermal perturbation

Figure 7.4 displays the wall and air temperatures inside the cave at different distances from the entrance, for case A with $\tilde{R} = 0.189$ and $Re = 1.8 \times 10^5$. Figure 7.4a shows the amplitude of the Fourier modes that stems directly from the numerical simulation. The mode k refers to the dimensionless period $1/k$. The amplitudes of the modes $k=0$ and $k=1$ are thus the annual mean temperature (AMT) and the amplitude of the annual temperature fluctuation (ATF), respectively. The modes $k=2$, $k=3$, and so on, refer to the periods 6 months, 4 months, etc. Although the external temperature imposed at $\tilde{x} = 0$ contains the single mode $k=1$ (since it reduces to Eq. (15)), the coupling between modes due to convection results in the existence of other modes than $k=1$ in cave temperatures (modes $k=0$ to 18 were considered in the simulations, see supplemental material). Figure 7.4b displays the periodic time series built from the Fourier series. The amplitude of the temperature fluctuations decreases when the distance from the entrance increases, as expected. The hot thermal anomaly expected at the vicinity of an upper entrance is also observed. Indeed, for $\tilde{x} = 10$ to 10^3 , the dimensionless temperature is positive throughout the year, which means that it is always higher than the external AMT, even in winter.

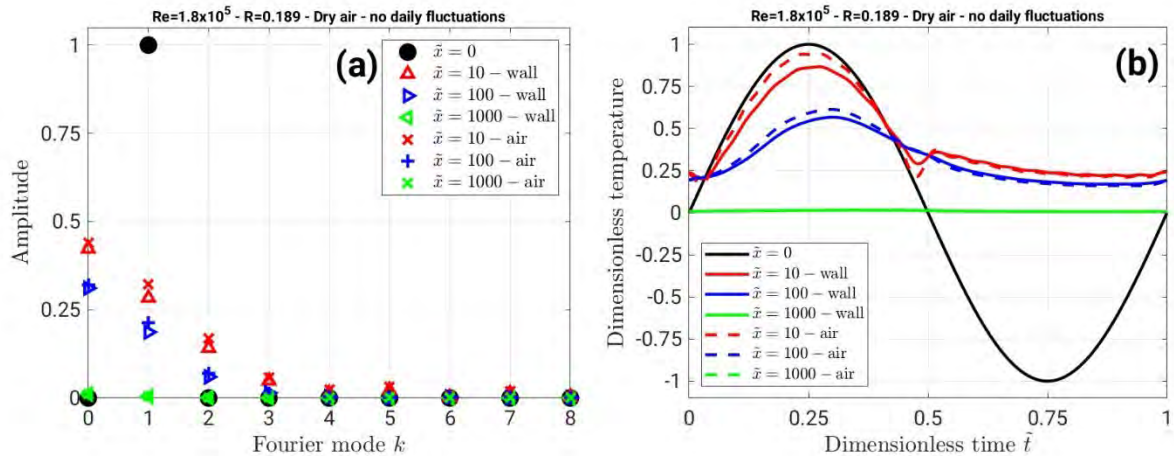


Figure 7.4. Wall and air temperatures at different distances $\tilde{x}$ from the entrance (case A with $\tilde{R}=0.189$ and $Re=1.8 \times 10^5$). (a) Amplitude of the Fourier modes. (b) Time series.

Figure 7.5a displays the temperature profiles in the air and at the cave wall at times $\tilde{t} = 0, 0.25, 0.5$ and 0.75 , corresponding to the external temperatures 0, 1, 0 and -1, respectively. Following an approach similar to that of Cropley (1965), we split the cave into three zones.

- 1- A first zone, from $\tilde{x} = 0$ to $\tilde{x} \approx 1$ (the diffusion length), where the cave temperature follows the fluctuations of the external temperature. More precisely, the sign of the temperature gradient changes in summer and winter (see the inset in Figure 7.5a). This first zone is dominated by the temperature fluctuations of the external wall, propagating inside the rock mass by heat conduction. It will be called diffusive region in the following.

- 2- A second zone characterized by a positive dimensionless temperature and reduced but significant fluctuations. In the specific case of Figure 7.5a, it approximately extends from $\tilde{x} \approx 1$ to 10^3 . In this zone, heat diffusion from the external wall is negligible compared to convection. It will be called convective region in the following. Because of the seasonal flow reversals, the inlet temperature of the air flow is 1 in summer (inward flow) and 0 in winter (outward flow, coming from the deep karst). As a consequence, this zone does not “feel” the negative winter temperatures. This results in: (1) a shift of the AMT to positive values, the so-called thermal anomaly, (2) a reduction of the amplitude of the fluctuations, because the temperature oscillates between 0 and 1 whereas the external temperature fluctuates between -1 and 1.
- 3- A third zone, called deep karst in the following, where the dimensionless temperature is approximately constant and close to zero (the external AMT).

These three regions appear clearly on the amplitude of the Fourier modes plotted in Figure 7.5b as a function of the distance from the entrance $\tilde{x}$. The amplitude of the mode $k=1$ (i.e. the ATF) decreases in two steps. It first drops down from 1 to 0.3 in the diffusive region (from $\tilde{x} = 0$ to $\tilde{x} \approx 1$), then slowly decreases in the convective region, over a distance much longer than 1 (of the order of 10^3 in the specific case of Figure 7.5). The amplitude of the mode $k=0$ (i.e. the AMT) increases in the diffusive region until it reaches a quasi-plateau. Then it slowly decreases in the convective region, over a distance of the same order as for the mode $k=1$. The modes such that $k \geq 2$ follow qualitatively the behavior of mode $k=0$, with a lower amplitude and a shorter attenuation length. In the following, we will focus on both modes with the highest amplitude, $k=0$ and $k=1$, as representative of the AMT (the thermal anomaly) and the ATF, respectively.

In all the simulations we did, the transition between the diffusive and convective regions always take place at a distance from the entrance of the order of 1, as expected. Conversely, the characteristics of the convective domain (amplitude of the thermal perturbation and its spatial extent) depend on the model parameters: the Reynolds number Re (containing the air flow rate) and the cave radius $\tilde{R}$. This dependence will be systematically investigated in the parametric study of the next section. In order to facilitate the interpretation of the results, we define a small number of relevant parameters characterizing the perturbation of the cave temperature field by the air flow. The magnitude of the perturbation is characterized by the AMT and the ATF at the beginning of the convective region, where the impact of convection is maximum, and the effect of thermal conduction from the external wall negligible. We define $\Delta\theta_w$ as the maximum amplitude of the mode $k=1$ in the convective region, for the wall temperature. Practically, it is evaluated at the change of slope noticed on the solid blue line in the inset of Figure 7.5b. Similarly, $\bar{\theta}_w$ is the maximum amplitude reached by the mode $k=0$ for the wall temperature (maximum of the red solid line in the inset of Figure 7.5b). The corresponding amplitudes for the air temperature, $\Delta\theta_a$ and $\bar{\theta}_a$, are estimated at the same location as $\Delta\theta_w$ and $\bar{\theta}_w$, respectively. For example, in the specific case of Figure 7.5, we get $\Delta\theta_w \approx 0.30$ and $\Delta\theta_a \approx 0.34$ (maximum ATF), $\bar{\theta}_w \approx 0.425$ and $\bar{\theta}_a \approx 0.44$ (maximum AMT).

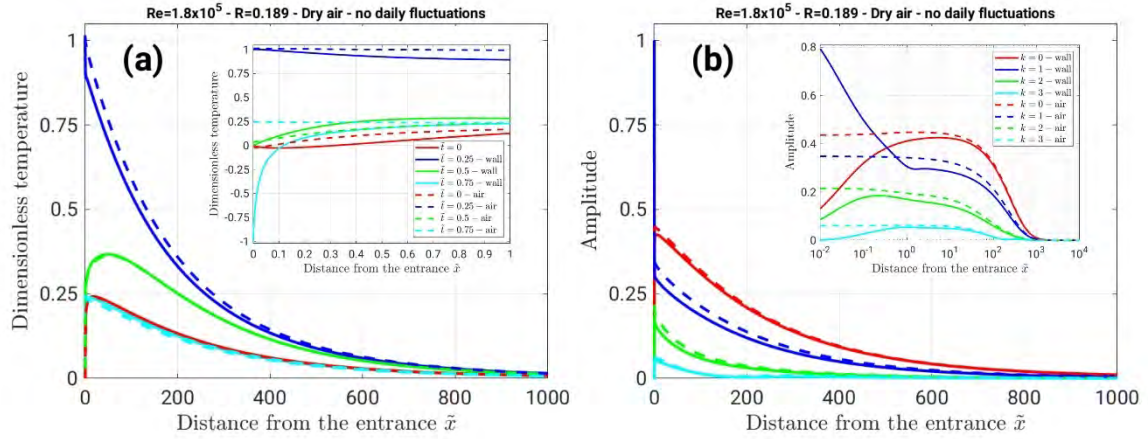


Figure 7.5. Wall (solid lines) and air (dashed lines) temperatures as a function of the distance $\tilde{x}$ from the entrance (case A with $\tilde{R}=0.189$ and $Re=1.8 \times 10^5$). (a) Temperature profiles at various times (Inset: zoom in the diffusive region). (b) Amplitude of the Fourier modes $k=0$ to 3 (Inset: log scale on the $\tilde{x}$ -axis).

The four parameters $\bar{\theta}_w$, $\bar{\theta}_a$, $\Delta\theta_w$ and $\Delta\theta_a$ represent the maximum amplitudes of the perturbation in the convective region. These amplitudes decrease with the distance from the entrance, as the heat transported by the air flow is gradually dissipated by conduction in the rock mass. We arbitrarily assume that a Fourier mode has been damped when its amplitude has been divided by 10. Figure 7.6a shows the thermal perturbation in a rock massif with the black solid lines corresponding to an AMT equal to $\bar{\theta}_w/10 = 0.0425$. The rock domain thus defined can be characterized by the length $\tilde{L}_0$ along the cave axis and the maximum width $\tilde{W}_0$ along the radial direction. Note that this maximum width is observed at a certain distance from the external wall because of the competition between radial diffusion from the cave which transports the thermal anomaly through the massif, and axial diffusion from the external wall which reduces it. Similarly, the black solid lines in Figure 7.6b correspond to an ATF equal to $\Delta\theta_w/10 = 0.030$. The size of the corresponding domain is characterized by its length $\tilde{L}_1$ and its width $\tilde{W}_1$. The latter is taken close to the entrance, but out of the diffusive region. The lengths $\tilde{L}_0$ and $\tilde{L}_1$ can also be estimated from the air temperature (length such that the amplitude of AMT and ATF in the air are equal to $\bar{\theta}_a/10$ and $\Delta\theta_a/10$, respectively).

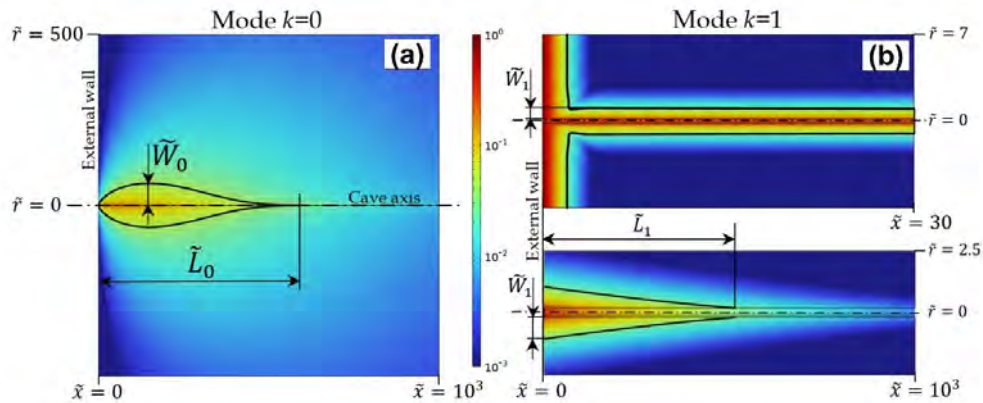


Figure 7.6. Amplitude contour of the thermal perturbation in a symmetry plane of the rock massif (case A with $\tilde{R}=0.189$ and $Re=1.8 \times 10^5$). (a) Mode $k=0$; the black solid lines correspond to the amplitude $\bar{\theta}_w/10 = 0.0425$ (n.b.: at the scale of the figure, the cave reduces to a line). (b) Mode $k=1$; the black solid lines correspond to the amplitude $\Delta\theta_w/10 = 0.030$ (n.b.: the top view focuses on the entrance region; the bottom view has been plotted with different scales on $\tilde{x}$ and $\tilde{r}$ to display the whole perturbed region).

7.5. Parametric study

7.5.1. Amplitude of the thermal perturbation induced by the air flow

Figure 7.7 displays the AMT ($\bar{\theta}_w$ and $\bar{\theta}_a$) and the ATF ($\Delta\theta_w$ and $\Delta\theta_a$) at the beginning of the convective region. Cases A and B yield close values of $\bar{\theta}_a$, which is also little sensitive to the model parameters Re and $\tilde{R}$. Indeed, $\bar{\theta}_a$ varies only from 0.25 to 0.6 when Re and $\tilde{R}$ are increased by three and two orders of magnitude, respectively. With the exception of small conduits ($\tilde{R} = 0.2$) at small Reynolds number ($Re \lesssim 3 \times 10^3$), a similar remark can be made on $\Delta\theta_a$.

The evolution of the wall temperature is more intricate. In case B, the assumption of infinite Biot number imposes $\bar{\theta}_w = \bar{\theta}_a$ and $\Delta\theta_w = \Delta\theta_a$ (since the air and wall temperatures are equal). Assuming that Case A provides a lower bound of the actual Biot number in a cave, Eq.(14) allows to estimate the consequences of an error in the determination of the Biot number on the estimation of the wall temperature at the beginning of the convective region. At high Reynolds number ($Re \gtrsim 3 \times 10^5$), no accurate estimation of the relation of the function $Nu(Re)$ is required because the heat transfer is completely driven by heat conduction in the rock massif. Conversely, the relation $Nu(Re)$ is a key information for convection-limited heat transfer at small Reynolds number ($Re \lesssim 10^4$). At intermediate Re , an accurate estimation of Nu is required to get the ATF of the wall in a cave with a large radius ($\tilde{R} \gtrsim 1$).

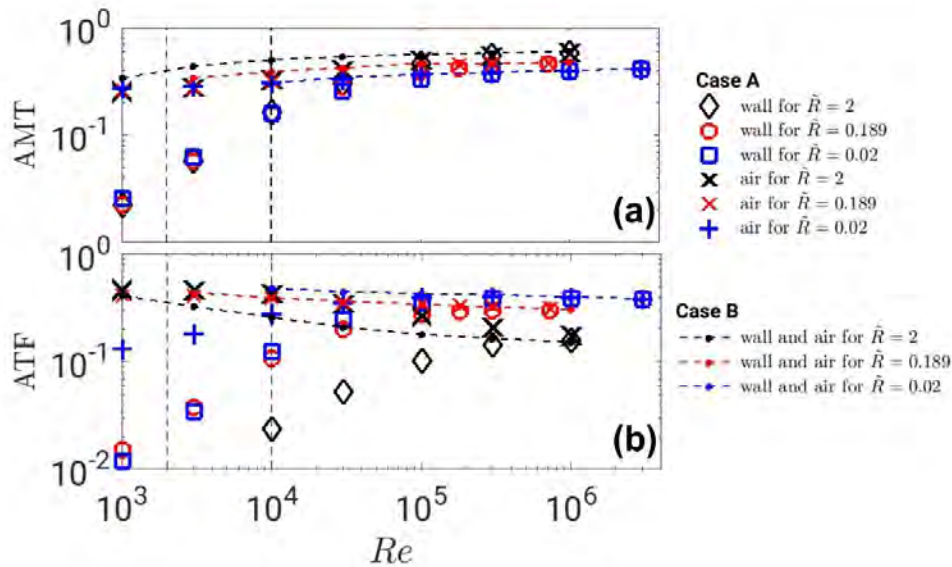


Figure 7.7. Maximum thermal perturbation at the beginning of the convective region for cases A (Biot number estimated from forced convection correlations) and B (infinite Biot number). (a) Maximum AMT ($\bar{\theta}_w$ and $\bar{\theta}_a$ for the wall and the air, resp.). (b) Maximum ATF ($\Delta\theta_w$ and $\Delta\theta_a$ for the wall and the air, resp.). The vertical dashed lines correspond to the limits of the laminar and turbulent regimes.

7.5.2. Spatial extent of the thermal perturbation induced by the air flow

The characteristic lengths $\tilde{L}_0$ and $\tilde{L}_1$ for the damping of the AMT and ATF along the cave axis are displayed in Figure 7.8. The values of $\tilde{L}_0$ computed in case B (infinite Biot number) support the simple relation obtained from curve fitting:

$$\tilde{L}_0 \approx 3.1 \times 10^{-3} Re^{1.06} \tilde{R}^{0.61} \quad (24)$$

In case A, estimating $\tilde{L}_0$ from the air or the wall temperatures give the same results (the lengths required to get $\bar{\theta}_a$ or $\bar{\theta}_w$ divided by 10 are the same, even when $\bar{\theta}_a$ and $\bar{\theta}_w$ are different). At high Reynolds number, $\tilde{L}_0$ estimated from case A follows Eq.(24) derived for case B, as expected when heat transfer is conduction-limited. When Re is decreased, using the forced convection correlations of case A reduces the heat flux at the cave wall compared to the infinite Biot number of case B, leading to larger values of $\tilde{L}_0$ than predicted by Eq. (24).

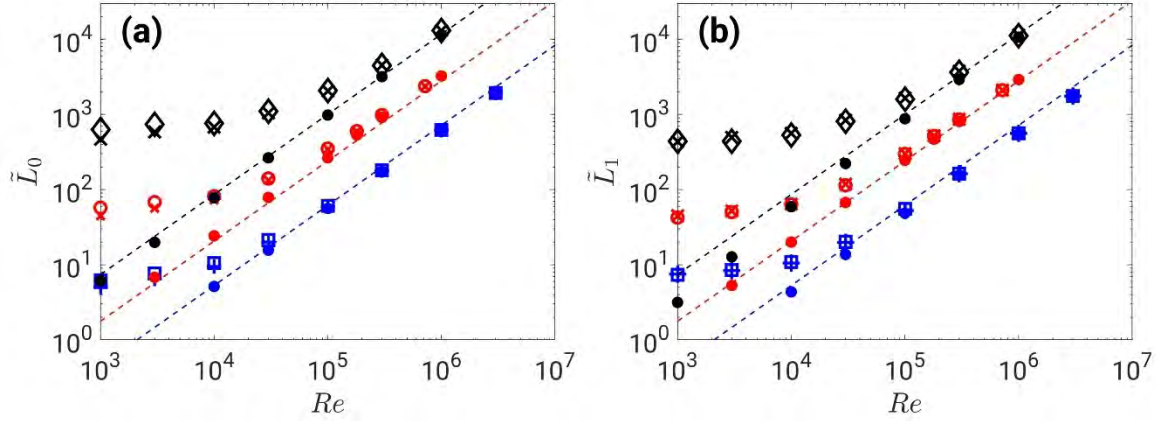


Figure 7.8. Characteristic lengths $\tilde{L}_0$ (a) and $\tilde{L}_1$ (b) as a function of the Reynolds number Re for various radius $\tilde{R}$, computed from wall or gas temperatures. The dashed lines correspond to Eq.(24). Symbols: same legend as in Figure 7.7.

$\tilde{L}_1$, the distance corresponding to the damping of the ATF, is nearly equal to $\tilde{L}_0$, with the exception of slight differences at low Reynolds number for $\tilde{R}=2$ (see Figure 7.8b). The length of the convective region (called convection length in the following) can thus be indifferently determined from the damping of the AMT or ATF, based on the wall or the air temperatures. A higher bound of the convection length is provided by case A, while case B provides a lower bound. We can see in Figure 7.8 that the uncertainty on the convection length increases when the Reynolds number Re is decreased and when the cave diameter $\tilde{R}$ is increased.

The width $\tilde{W}_0$ corresponding to the damping of the AMT in radial directions inside the rock mass is displayed in Figure 7.9. It is approximately proportional to $\tilde{L}_0$ and an order of magnitude smaller.

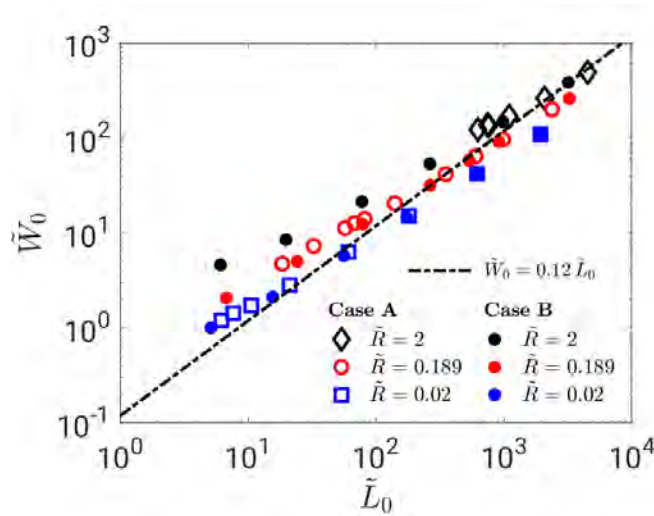


Figure 7.9. Width $\tilde{W}_0$ corresponding to the damping of the AMT in the radial directions inside the rock mass.

7.6. Comparison with field data

7.6.1. Baulmes mine (Switzerland)

The Baulmes mine (46°47'33"N, 6°31'35"E, 655 m a.s.l.) is located in western Switzerland. This mine comprises a network of 11 sub horizontal levels with a cumulated length of about 17 km (Figure 7.10). The difference in elevation between the uppermost and lowest entrances is 90 m, ensuring a strong chimney effect. The sub horizontal conduits at intermediate elevations between the lower and upper entrances, connected by shafts, lead to dead ends in the horizontal directions (for more information about Baulmes mine, see Bourret et al. (2007)). The temperature delivered by Mathod weather station (46°44'13"N, 6°34'4"E, 435 m a.s.l., at 7 km SW from the study site) is used as an estimation of the external atmosphere temperature. A correction has been applied to account for the difference of elevation with the mine entrance, using the vertical thermal gradient $-5.2^{\circ}\text{C}/\text{km}$ (mean value from the network of swiss meteorological stations for the time range 1991-2020). The resulting external temperature is displayed in Figure 7.11 for one year from July 1st 2022 at 0h00 (this date is taken as time $t = 0$ for all the data from Baulmes mine). The sampling period was 1 hour. The Fast Fourier Transform (FFT) of the external atmosphere temperature yielded $T_m = 10.4^{\circ}\text{C}$ (after correction) and $\Delta T_y = 9.4^{\circ}\text{C}$ for the external AMT and ATF, respectively.

The 360 m long tunnel #10 is connected horizontally to the lower entrance with no connection to other branches. This is the zone of interest of study. It is indicated by the yellow shading in Figure 7.10. This conduit has a typical horseshoe cross section. The geometrical characteristics of this conduit, including the mean perimeter and cross-sectional area, their standard deviation as well as its total length are summarized in Table 7.2. We instrumented this tunnel with seven temperature probes at different distances from the entrance. They are listed in Table 7.3. The first six temperature probes are Reefnet (Sensus Ultra) with $\pm 0.3^{\circ}\text{C}$ accuracy. The last one located at 360 m from entrance is HOBO (u22-001) with $\pm 0.2^{\circ}\text{C}$ accuracy. The temperature inside the conduit was recorded by these seven probes from February 3rd 2023 ($t = 0.596$ year) to July 7th 2023 ($t = 1.017$ year). This time range is displayed by the horizontal arrow in Figure 7.11. The sampling period was 1 hour.

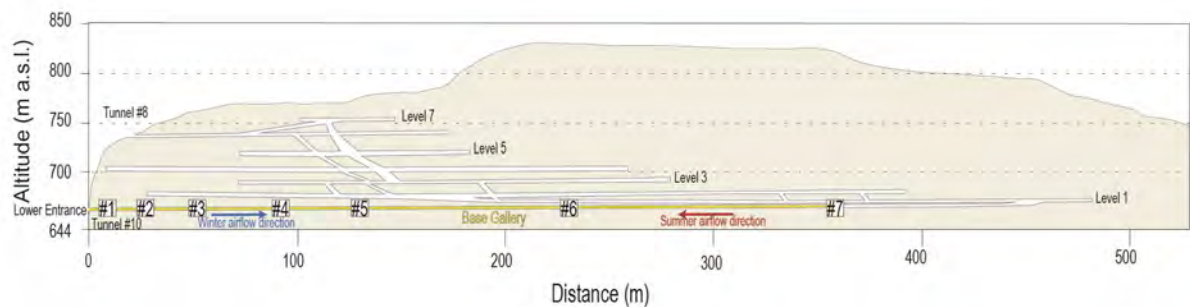


Figure 7.10. Map of Baulmes mine (vertical section).

Table 7.2: Geometrical characteristics of the instrumented conduit of Baulmes mine (length L_c , mean value and standard deviation of the perimeter P and the cross-sectional area A from the survey of 28 cross-sections).

L_c (m)	P (m)		A (m ²)	
	Mean	std	Mean	std
360	9.7	1.1	6.6	1.5

Table 7.3: location of the seven temperature probes inside the instrumented conduit of Baulmes mine. x is the distance from the lower entrance.

Probe	#1	#2	#3	#4	#5	#6	#7
x (m)	10	26	50	90	128	220	360
$\tilde{x} = x/L_{dif}$	1.9	4.9	9.4	17	24	42	68

The air mass flow rate was measured sporadically with a manual hot wire anemometer (Testo425 with an accuracy of $0.03 \text{ m/s} \pm 5\%$ of the measured value). The measurements were carried out at the four dates listed in Table 7.4 and displayed by vertical red bars in Figure 7.11. There are two dates in winter and two in summer. The mean flowrates in winter and summer are 2.4 kg/s and 5.0 kg/s , respectively. In the following, we consider $\Delta \dot{m} \sim (2.4 + 5.0)/2 \sim 4 \text{ kg/s}$ as a rough estimation of the annual amplitude of the air flowrate. With the mean conduit perimeter $P=9.7 \text{ m}$ from Table 7.2 and the air viscosity $\mu_a \approx 1.77 \times 10^{-5} \text{ Pa.s}$ from Table 7.1, we get the order of magnitude of the Reynolds number $Re = \frac{4 \Delta \dot{m}}{P \mu_a} \sim 10^5$. The airflow is thus in the turbulent regime most of the time. The conduit radius reads $R = P/(2\pi) \approx 1.5 \text{ m}$ that yields the scaled value $\tilde{R} = R/L_{dif} \approx 0.3$. The conduit is thus close to the configuration displayed in Figure 7.4 and Figure 7.5 corresponding to $Re = 1.8 \times 10^5$ and $\tilde{R} = 0.189$. In this configuration, cases A and B yield approximately the same results. The heat transfer is thus limited by conduction in the rock rather than convection. Therefore, Eq.(24), for which conduction-limited heat transfer is assumed, is expected to provide a fair estimation of the convection length, and not only a lower bound. We get $\tilde{L}_0 \approx \tilde{L}_1 \sim 3 \times 10^2$. The theory thus predicts a convection length significantly larger than the conduit length $\tilde{L}_c = L_c/L_{dif} \approx 68$.

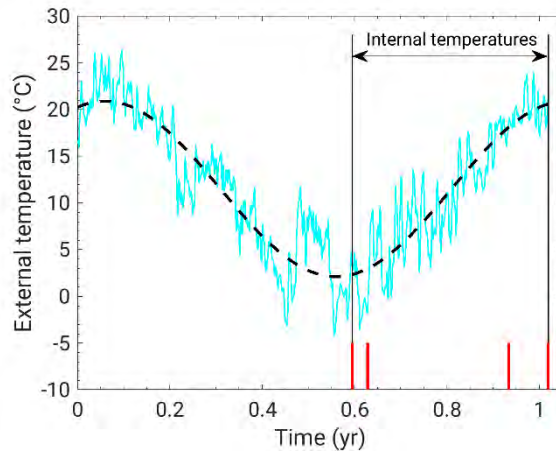


Figure 7.11. External atmosphere temperature estimated at Baulmes mine (daily averaged data). The horizontal arrow displays the time range during which the temperature inside the cave was measured by the sensors listed in Table 7.3. The vertical red bars correspond to the dates of the air flowrate measurements listed in Table 7.4. The black dashed line displays the mode $k=0$ and $k=1$ of the external temperature.

Table 7.4: air mass flow rate measured in winter and summer.

	Winter (upward flow)		Summer (downward flow)	
Date	02/02/2023	02/15/2023	06/06/2023	7/7/2023
$\tilde{t} = t/\tau_y$	0.595	0.628	0.933	1.018
$\dot{m} \text{ (kg/s)}$	1.9	2.8	4.5	5.4

The time series of the external atmosphere temperature and the temperatures inside the conduit have been plotted in Figure 7.12a. The dimensionless temperatures inside the conduit are always negative (i.e., lower than the external AMT), even in summer. This is the cold thermal anomaly expected at a lower entrance.

In the following, we assume that the temperature profiles $\theta_w(\tilde{x})$ at time $\tilde{t} = 0.611$ and $\theta_s(\tilde{x})$ at time $\tilde{t} = 1.016$ give an assessment of the coldest and hottest temperatures reached inside the conduit in winter and summer, respectively. These two times are displayed by vertical bars in Figure 7.12a, and the corresponding profiles are plotted in Figure 7.12b (the temperature at $\tilde{x} = 0$ is the external

atmosphere temperature). In winter, the dimensionless temperature continuously increases from -1.6 at the inlet ($\tilde{x} = 0$) to -0.56 at the end of the conduit ($\tilde{x} = 68$). Spatial temperature variations are thus significant over the entire length of the conduit. This suggests that the convection length should be larger than the conduit length. In summer, the reverse flow direction yields a different temperature profile. The temperature at the end of the conduit ($\tilde{x} = 68$) is -0.06, close to the external AMT. When getting closer to the entrance, the temperature decreases to -0.16 at the first probe ($\tilde{x} = 1.9$). The temperature then suddenly increases within the diffusive region, from -0.16 at the first probe ($\tilde{x} = 1.9$) to the positive value 0.78 at the entrance ($\tilde{x} = 0$).

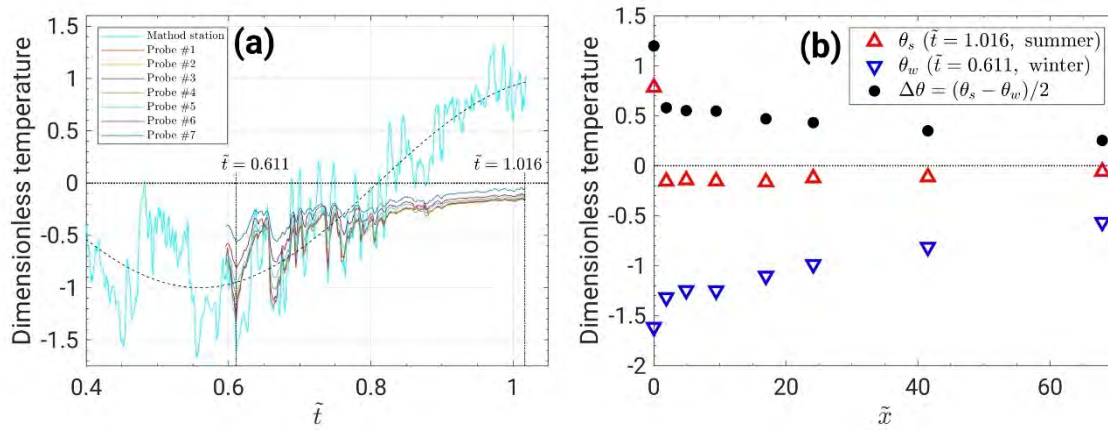


Figure 7.12. Temperatures in Baulmes conduit. (a) Daily averaged temperatures as a function of time (the temperature at $\tilde{x} = 0$ is the external atmosphere temperature measured at Method weather station). (b) Temperature profiles $\theta_s(\tilde{x})$ and $\theta_w(\tilde{x})$ measured respectively in summer and winter, and amplitude of the annual variation $\Delta\theta = (\theta_s - \theta_w)/2$. Dimensionless temperatures defined as $\theta = (T - T_m)/\Delta T_y$.

The difference $\Delta\theta = (\theta_s - \theta_w)/2$ between summer and winter profiles yields an assessment of the amplitude of the ATF in the conduit. $\Delta\theta$ drops from 1.2 to 0.58 between the entrance ($\tilde{x} = 0$) and the first probe ($\tilde{x} = 1.9$). This sudden decrease takes place in the diffusive region. $\Delta\theta$ then smoothly decreases from 0.58 (first probe at $\tilde{x} = 1.9$) to 0.25 at the end of the conduit ($\tilde{x} = 68$). Assuming that the first sensor yields the temperature at the beginning of the convective region, the amplitude of the temperature fluctuation is thus divided by approximately 2.3 in the convective domain of the conduit. This confirms that the conduit length is significantly shorter than the convection length, defined as the distance required to divide by 10 the ATF measured at the beginning of the convective region (see section 4).

7.6.2. D7.1 Cave (Sieben Hengste, Switzerland)

D7.1 cave is located in the Sieben Hengste karst area (46°45'43"N, 7°49'51"E, 1750 m a.s.l). It is part of a large ventilated cave system (>160 km of cumulated length) subject to chimney effect (see the map in Figure 7.13). Field observations reveal a seasonal airflow consistent most of the time with an upper entrance of the cave system (i.e., air inflow during the summer season). The ingress of warm external air during summer is marked by some condensation within the first 20 m of the meander, drained as a streamlet at the base of the meander. Otherwise, the cave passage remains remarkably dry. Unexpectedly, temperature and flow rate measurements show periods of time in winter during which the cave sporadically behaves as a lower entrance, with cold air inflow. The explanation of these unexpected flow reversals is still being searched.

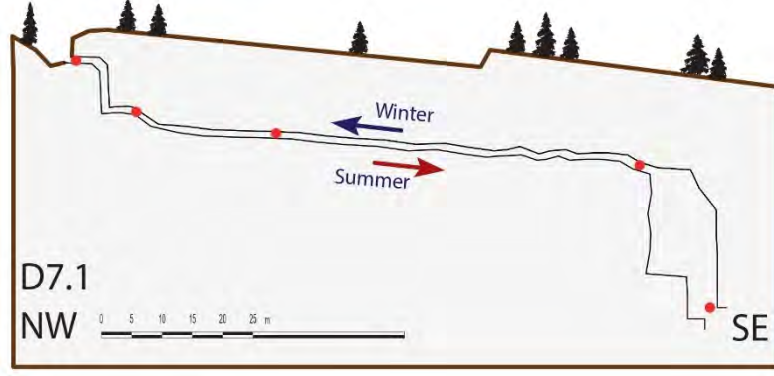


Figure 7.13. Map of D7.1 Cave. The red circles indicate temperature sensors.

External air temperatures displayed in Figure 7.14a are provided by an automatic weather station located at 1850 m elevation (SLF), approximately 2 km SE of our study site (sampling period: 1 hour). The external atmosphere temperature was corrected from the elevation as in section 6.1. The FFT yielded $T_m = 5.0^\circ\text{C}$ (after correction) and $\Delta T_y = 7.1^\circ\text{C}$ for the external AMT and ATF, respectively.

The *ca.* 100 m-long entrance meander of D7.1 Cave has a simple geometry with an average hydraulic diameter of *ca.* 1 m, despite local variations in the cross-sectional area and perimeter. The perimeter $P \approx 3$ m and the scaled radius $\tilde{R} \approx 0.1$ will be considered in the calculations. The meander was instrumented for temperature and airflow measurements between 25/08/2021 and 10/08/2022. Five temperature probes were installed along the main conduit at distances from the cave entrance listed in Table 7.5. Probes #1 to #4 are Reefnets Sensus Ultra with $\pm 0.3^\circ\text{C}$ accuracy (sampling period: 30 min). Probe #5 is a HOBO U22-001 with $\pm 0.2^\circ\text{C}$ accuracy (sampling period: 2 hours). It is located at the base of a 20 m shaft following the entrance meander. Time series from the five probes are displayed in Figure 7.14b.

Table 7.5: location of the five temperature probes inside the instrumented conduit of D7.1 Cave. x is the distance from the entrance.

Probe	#1	#2	#3	#4	#5
x (m)	2	11	33	93	125
$\tilde{x} = x/L_{dif}$	0.38	2.1	6.2	18	24

The cave airflow was measured using a low pressure drop flowmeter (Sensirion SFM 3000 with accuracy of 5% and 20% when velocity is higher and lower than 0.3 m/s, respectively; 50 m from cave entrance) and validated with independent flow measurements (CO_2 gauging) on 10/08/2022. The FFT of the air flow yields the order of magnitude $\Delta \dot{m} \sim 0.1$ kg/s rate (red dashed line in Figure 7.14a), which results in $Re \sim 7 \times 10^3$. The cave is thus most of the time in the transition between laminar and turbulent regimes. Eq.(24) yields the theoretical convection length $\tilde{L}_0 \approx \tilde{L}_1 \sim 10$. It is important to note that this value must be considered as a lower bound of the convection length. Indeed, Eq.(24) stands for case B (infinite Biot number). At $Re \sim 7 \times 10^3$, case A (finite Biot number) yields a convection length significantly larger than case B (see Figure 7.8).

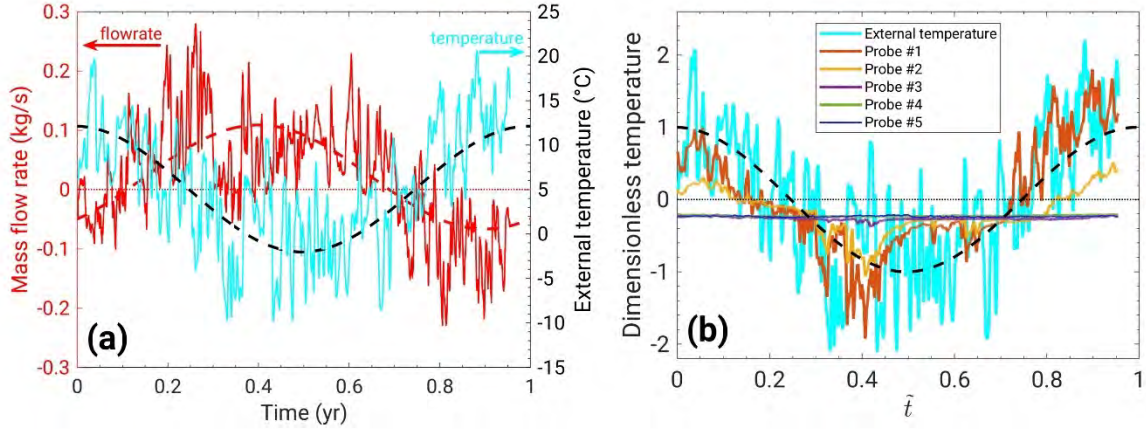


Figure 7.14. Field data from D7.1 cave (all the data are daily averaged). (a) External temperature and mass flow rate; black and red dashed lines represent the modes $k=0$ and $k=1$ computed from the FFT of the external atmosphere temperature and the air flow rate, respectively. (b) Dimensionless temperatures outside and inside the cave ($\theta = (T - T_m)/\Delta T_y$).

The AMT inside D7.1 Cave is displayed in Figure 7.17a. It decreases to reach a negative quasi-plateau in the range from -0.26 to -0.23 in between probes #3 ($\tilde{x} = 6.2$), and #5 ($\tilde{x} = 24$). This is not the behavior expected for an upper entrance, for which the AMT is supposed to be positive in the convective region before decreasing to zero at some distance from the entrance. Sporadic flow reversals in winter can contribute to this negative AMT (an instance of such an event can be observed around time $\tilde{t} \approx 0.33$ in Figure 7.14). Moreover, the cave entrance is in a topographic low, resulting in local cold air trapping during winter. In addition, snow is likely to accumulate near the cave entrance. Both may induce local free convection cells in the entrance zone in winter resulting in the cooling of the cave below the external AMT.

If we assess $\tilde{L}_0$ from the distance over which the AMT significantly varies inside the cave, it would range between $\tilde{x} = 2.1$ (probe #2) and $\tilde{x} = 6.2$ (probe #3). This estimation is confirmed by the variation of the ATF with the distance from the entrance (Figure 7.17b). The ATF at probe #1 ($\tilde{x} = 0.38$), located in the diffusive region, is close to 1. Then it decreases to 0.36 at probe #2 ($\tilde{x} = 2.1$), the first probe in the convective region, and drops down to 0.02 at probe #3 ($\tilde{x} = 6.2$). The ATF being reduced by a factor larger than 10 between probes #2 and #3, $\tilde{L}_1$ should be in between these two probes. It is worthwhile to note that this estimation is consistent with the limit of condensation, which falls in between probes #2 and #3 (20 m corresponds to $\tilde{x} \approx 4$). The convection length of D7.1 Cave is thus in the range from 2.1 to 6.2.

7.6.3. Bärenhöhle Cave (Austria)

The cave is located in Vorarlberg, Austria (47°22'17"N, 9°52'52"E, 901 m asl). The external atmosphere temperature is obtained from the weather station located at Saint Gallen (47°25'31.7094"N, 9°23'54.7008"E, 776 m a.s.l.), Switzerland about 37 km WNW of our study site (the closer Schopernau station (47°19' N, 10°1'E, 835 m a.s.l.; ZAMG 2012) is subject to thermal inversions during the winter season and thus considered as being less representative). The temperature measured at Saint Gallen has been corrected from the elevation as in section 6.1. It is displayed in Figure 7.16a. The FFT yielded $T_m = 8.7^\circ\text{C}$ (after correction) and $\Delta T_y = 7.6^\circ\text{C}$ for the external AMT and ATF, respectively.

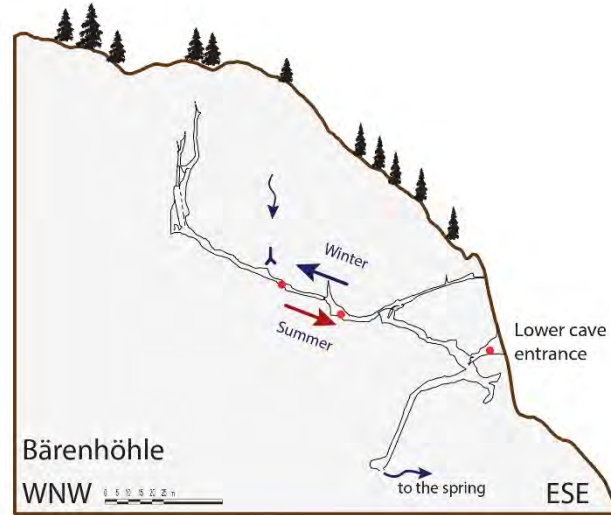


Figure 7.15. Map of Bärenhöhle. The red circles indicate temperature sensors.

Chimney effect drives the airflow in the cave which behaves as a lower entrance (air inflow in winter, see Figure 7.15). The difference in elevation between the lower and upper entrances is *ca.* 130 m for a conduit length along the main air passage approximating 240 m (Perkmann and Luetscher, 2013). The average hydraulic diameter is assessed to about 1 m, so that $P \approx 3$ m and $\tilde{R} \approx 0.1$. Air temperature was recorded at 1h interval using three HOBO (U22-001) temperature probes at distances from the entrance listed in Table 7.6. The corresponding time series are displayed in Figure 7.16b.

Table 7.6: location of the three temperature probes inside Bärenhöhle. x is the distance from the lower entrance.

Probe	#1	#2	#3
x (m)	5	85	120
$\tilde{x} = x/L_{dif}$	0.95	16	23

Air flow was recorded between 16/04/2013 and 16/04/2014 at the lower entrance using a thermo-anemometer calibrated to ± 0.1 m/s. The FFT of the air flow rate (red dashed line in Figure 7.16a) yields the orders of magnitude $\Delta \dot{m} \sim 0.6$ kg/s and $Re \sim 5 \times 10^4$. The cave is thus in the turbulent regime most of the time. In this configuration ($\tilde{R} \approx 0.1$ and $Re \sim 5 \times 10^4$), cases A and B predict approximately the same theoretical convection length (see Figure 7.8). Eq.(24) yields $\tilde{L}_0 \approx \tilde{L}_1 \sim 70$.

The AMT of Bärenhöhle Cave is displayed in Figure 7.17a. Negative values are consistent with a lower entrance. The lowest AMT (-0.27) is measured at probe #2 ($\tilde{x} = 16$). It increases slightly between probe #2 and the last probe #3. The ATF displayed in Figure 7.17b decreases over the whole instrumented conduit, from 1 outside the cave to 0.11 at the last probe ($\tilde{x} = 23$). These two pieces of information suggest that the convection length should be longer than 23, the cave length.

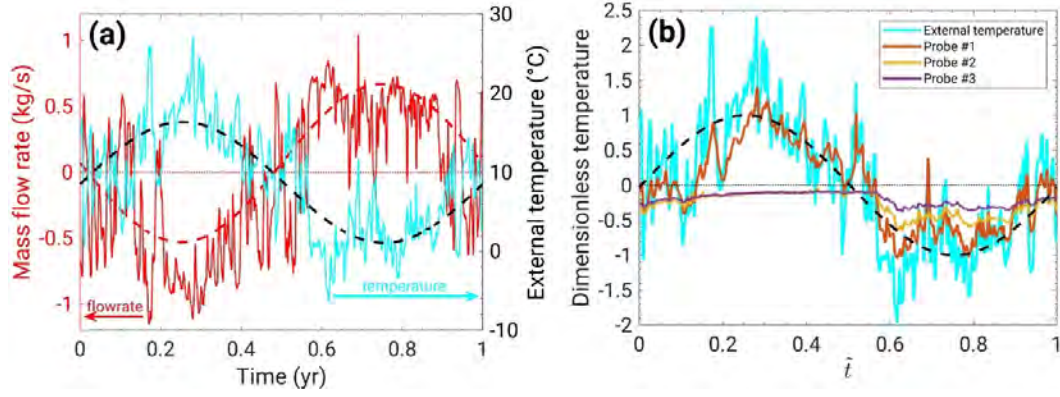


Figure 7.16. Field data of Bärenhöhle Cave (all the data are daily averaged). (a) External temperature and mass flow rate; black and red dashed lines represent the modes $k=0$ and $k=1$ computed from the FFT of the external atmosphere temperature and the air flow rate, respectively. (b) Dimensionless temperatures outside and inside the cave ($\theta = (T - T_m)/\Delta T_y$).

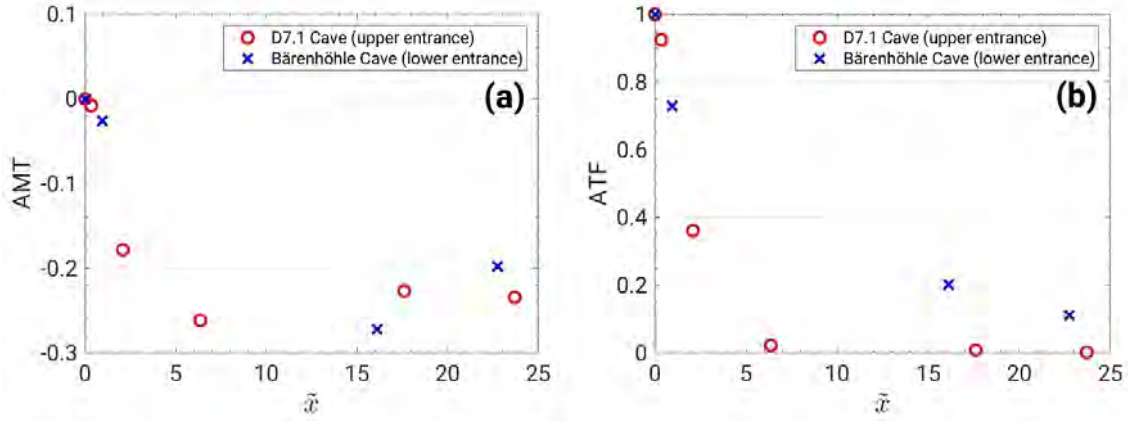


Figure 7.17. AMT (a) and ATF (b) in D7.1 and Bärenhöhle Caves.

7.7. Discussion

The main results from the three investigated sites are summarized in Table 7.7. For two sites (Baulmes mine and Bärenhöhle Cave), we do not observe any contradiction between the model and the field data since both yield a convection length larger than the instrumented conduit (a hundred meters for Bärenhöhle, a few hundreds of meters for Baulmes). In D7.1 Cave, the field data suggest a shorter convection length, in the range from 10 m to 30 m, whereas the theoretical prediction yields a lower bound of 50 m. Although the theoretical order of magnitude is correct, the theory overestimates the convection length. In a cave with a small Reynolds number and a short convection length, local effects in the entrance region, not considered in the model, could overcome the effect of the mean flow generated by the chimney effect. In the case of D7.1 Cave, cold air ingress in winter might cool down the cave passage more than considered in the model, thus producing an actual shorter thermal length.

Table 7.7 : Summary table of the three investigated sites (R : cave radius in meters, $\tilde{R}$: dimensionless cave radius; $\Delta\dot{m}$: amplitude of the air flow rate in kg/s; Re : Reynolds number; $\tilde{L}_0$ and $\tilde{L}_1$: dimensionless convection lengths deduced from the AMT and the ATF, respectively; L_0 and L_1 : same variables in meters.

		Baulmes mine	D7.1 Cave	Bärenhöhle Cave
R		1.5 m	0.5 m	0.5 m
$\tilde{R}$		0.3	0.1	0.1
$\Delta\dot{m}$		4 kg/s	0.1 kg/s	0.6 kg/s
Re		10^5	7×10^3	5×10^4
$\tilde{L}_0 \approx \tilde{L}_1$	Field data	>68	[2.1 , 6.2]	> 23
	Theory	$\sim 3 \times 10^2$	> 10	~ 70
$L_0 \approx L_1$	Field data	> 360 m	[10 m , 30 m]	> 120 m
	Theory	$\sim 2 \times 10^3$ m	> 50 m	$\sim 4 \times 10^2$ m

Using a theoretical approach, we defined the characteristic lengths $\tilde{L}_0$ and $\tilde{L}_1$ from the damping in the convective region of the AMT (mode $k=0$) and ATF (mode $k=1$), respectively. $\tilde{L}_0$ and $\tilde{L}_1$ are close to each other, and similar results are obtained if the estimation of these lengths is based on the wall temperature or that of the air (see Figure 7.8). A single convection length is thus sufficient to characterize the extent of the AMT and the ATF along the cave axis. The case B (infinite Biot number) provides a lower bound of the convection length through the simple expression (24). If the Reynolds number is large enough (typically, if $Re > 3 \times 10^4$ for a conduit with a radius of 1 m, see Figure 7.8), the approximation of infinite Biot number is valid and Eq. (24) is expected to give a right estimation of the convection length. Turning this relation into unscaled variables yields

$$L_0 \approx L_1 \approx 4.0 \times 10^2 \frac{\Delta\dot{m}^{1.06}}{R^{0.45}} \quad (25)$$

where L_0 , L_1 and R are in meters and $\Delta\dot{m}$ in kg/s. Eq.(25) gives the convection length resulting from the balance between heat advection by the air flow and heat conduction in the rock mass, in the conduction-limited regime. In this case, the convection length is approximately proportional to the amplitude of the flowrate annual fluctuations $\Delta\dot{m}$ divided by the square root of the cave radius R . Ventilated conduits with R of the order of 1 m traversed by an airflow with $\Delta\dot{m}$ ranging from 0.1 kg/s to 1 kg/s are commonly encountered in mines and caves (see the case studies of section 6). For such conduits, Eq.(25) yields a convection length along the conduit axis between a few tens and a few hundreds of meters. Thermal perturbations also propagate in the radial direction, i.e., along the directions perpendicular to the conduit. As expected, the amplitude of the annual fluctuations vanishes over the diffusion length, i.e., a few meters. Conversely, the thermal anomaly (defined as the shift of the annual mean temperature), propagates along distances of the order of $L_0/10$ (see Figure 7.9), i.e., over distances of the order of a few meters to a few tens of meters. It is worthwhile to note that in some cases, the extent of the thermal perturbation might be much larger than mentioned above. Luetscher and Jeannin (2004) measured air flow rates around 10 kg/s in Hölloch Cave (Switzerland) and La Diau (France). Recently, airflow rates around 40 kg/s were measured in Shpella Shtares Cave, in Albania (42.31766° N, 19.85403° E, 1427 m a.s.l, see Ref. (Pastore et al., 2019) for more details about this cave). Taking a couple of meters as a rough estimation of the conduit radius in Shpella Shtares, the thermal perturbation should extend over kilometers and hundreds of meters in the axial and radial directions, respectively.

Assessing the thermal penetration length at different time-scales is fundamental to quantify biogeochemical processes in karst systems. Our study demonstrates that seasonal temperature fluctuations propagate from a few tens of meters up to several thousands of meters into a karst system,

depending on ventilation rate and conduit geometry. By way of illustration, let us cite the large ice caves in Austria which host perennial ice formations over several hundreds of meters in the lower entrance zone of ventilated cave systems (e.g. Eisriesenwelt; (Obleitner and Spötl, 2011)). However, it is the advent of speleothems as precisely and accurately dated paleoclimate archives which rises the significance of our study. Stalagmites and flowstones are commonly used to reconstruct climate fluctuations based on their isotope (oxygen and carbon) and trace element concentrations (Fairchild and Baker, 2012). In both cases, the proxy partitioning depends strongly on the cave temperature. While it is commonly assumed that the cave temperature at depth is constant, this is only true beyond the convection length which strongly depends on the time-scale of interest.

Because the temperature-dependent oxygen isotope fractionation between water and calcite ranges between -0.18 and -0.21 ‰/°C (Kim and O'Neil, 1997)(Tremaine et al., 2011), seasonal temperature changes >0.5 °C represent a significant source of uncertainty in the speleothem-proxy interpretation and it might be of interest to sample beyond the convective region. Meanwhile, it can be strategic to focus speleothem research on the convective region associated with decadal to millennial cycles, as this area will respond sensitively to outside climate changes without picking up short-term temperature fluctuations.

7.8. Conclusion

We developed a model coupling buoyancy-induced convection (chimney effect) in a single karst conduit with conduction in the rock mass. We used this model to investigate the propagation of thermal perturbations inside a karst massif. Assuming dry air and a simplified geometry, and reducing the time variations of the external temperature to the annual fluctuations, we performed a parametric study to identify general trends regarding the effect of the air flowrate and conduit size on the amplitude and relaxation length of the thermal perturbations induced by the air flow. The thermal anomaly (defined as a shift in the annual mean temperature) and the annual temperature fluctuations both propagate over the same distances from the entrance. We call this distance the convection length and we show that at high Reynolds number, it is approximately proportional to the air flow rate divided by the square root of the cave radius. The order of magnitude of the theoretical convection length seem to compare satisfactorily with field data obtained from a mine and two caves. Convection lengths of a few tens to a few hundreds of meters should be commonly encountered in ventilated caves. In extreme cases, it could go up to kilometers.

Acknowledgements

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Open Research

Text files including the numerical results plotted in Figures 4-9 are available in <https://doi.org/10.5281/zenodo.10953068> cited as Sedaghatkish and Doumenc (2024). The Comsol Multiphysics file containing the data and results related to the reference case of Figures 4-6 is also available, but a license of Comsol Multiphysics software is necessary to use it. All the field data (including the three instrumented sites) can be found in <https://doi.org/10.5281/zenodo.10514555> associated to Sedaghatkish et al. (2024).

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7.9. Supporting Information

Introduction

Text S1 describes the method based on Fourier series to compute the time-periodic solution of the mathematical model defined in section 2.6. The numerical validation of the method is displayed in Text S2 (including Figs. **S1** to **S3**). A brief comparison between Fourier series (FS) and the resolution of the transient problem by time integration (TI) is presented in Text S3 (including Figs. **S4** to **S6**).

Text S1

The resolution by Fourier series requires a slight modification of the model defined in section 2.6. Eq. (17) resulting from the air energy balance is of order 1 in space, so that only 1 boundary condition must be imposed, the air temperature at the inlet. Since the position of the inlet changes at each flow reversal, the boundary condition has to be applied alternatively on each side of the conduit. This is possible when the problem is solved by time integration, but difficult to implement with Fourier series. This difficulty is overcome by adding a dispersion term in the air energy balance which turns to second order in space, making it possible to apply Dirichlet boundary conditions on both sides of the conduit whatever the flow direction. Eqs. (17-18) turn to:

$$\mu(\tilde{t}) Re \frac{\partial \theta_a}{\partial \tilde{x}} = \left(4 \frac{k_r}{k_a} Pr^{-1} \right) \tilde{\varphi}_w + Re Pe^{-1} \frac{\partial^2 \theta_a}{\partial \tilde{x}^2} \quad (\text{S1})$$

$$\theta_a(\tilde{x} = 0, \tilde{t}) = \sin(2\pi\tilde{t}) \quad \text{and} \quad \theta_a(\tilde{x} = \tilde{L}_{dom}, \tilde{t}) = 0 \quad (\text{S2})$$

It is important to note that the new dispersion term in Eq. (S1) is a numerical trick with no physical sense. If the Peclet number Pe is large enough, the addition of the dispersion term in Eq. (S1) does not significantly modify the model output. An exception is the outlet region, where the dispersion term induces the air temperature to fit the external temperature at the end of the conduit. This change in the air temperature field takes place over a distance from the outlet of the order of Pe^{-1} , which can be arbitrarily small if Pe is large enough. The constant value $Pe = 10^5$ was set in all the simulations. We

checked on a few cases that imposing $Pe = 10^6$ did not change the results. In a second step, all the functions of time are approximated by truncated Fourier series and inserted in the mathematical model. These functions of time include:

- Two model inputs: the reduced air flowrate $\mu(\tilde{t})$ and heat transfer coefficient $\eta(\tilde{t})$.
- Two model outputs: the rock and air temperatures, respectively $\theta_r(\tilde{x}, \tilde{r}, \tilde{t})$ and $\theta_a(\tilde{x}, \tilde{t})$.

We begin with the model inputs. The exact expression of $\mu(\tilde{t})$ and $\eta(\tilde{t})$ are known *a priori* (see section 2.6). Considering that $\mu(\tilde{t})$ is an odd function of time and $\eta(\tilde{t})$ an even function of time, their approximations by truncated Fourier series read:

$$\mu(\tilde{t}) = \sum_{k=1}^{N_\mu} \mu_k \sin(2\pi k \tilde{t}) \quad \text{and} \quad \eta(\tilde{t}) = \eta_0 + \sum_{k=1}^{N_\eta} \eta_k \cos(2\pi k \tilde{t}) \quad (\text{S3})$$

Where μ_k and η_k are real coefficients deduced from the exact expressions of $\mu(\tilde{t})$ and $\eta(\tilde{t})$, respectively. N_μ and N_η are the number of modes taken into account. Increasing N_μ and N_η improves the accuracy of the approximated relations (S3), but requires more computational resources. We now focus on the model outputs. The temperatures fields in the rock $\theta_r(\tilde{x}, \tilde{r}, \tilde{t})$ and in the air $\theta_a(\tilde{x}, \tilde{t})$ are approximated by the truncated Fourier series:

$$\begin{aligned} \theta_r(\tilde{x}, \tilde{r}, \tilde{t}) &= \theta_{r,0}(\tilde{x}, \tilde{r}) + \sum_{k=1}^{N_\theta} \theta_{r,k}(\tilde{x}, \tilde{r}) \cos(2\pi k \tilde{t} + \varphi_{r,k}(\tilde{x}, \tilde{r})) \\ &= \sum_{-N_\theta}^{N_\theta} \Theta_{r,k}(\tilde{x}, \tilde{r}) \exp(2\pi j k \tilde{t}) \end{aligned} \quad (\text{S4})$$

$$\begin{aligned} \theta_a(\tilde{x}, \tilde{t}) &= \theta_{a,0}(\tilde{x}) + \sum_{k=1}^{N_\theta} \theta_{a,k}(\tilde{x}) \cos(2\pi k \tilde{t} + \varphi_{a,k}(\tilde{x})) \\ &= \sum_{-N_\theta}^{N_\theta} \Theta_{a,k}(\tilde{x}) \exp(2\pi j k \tilde{t}) \end{aligned} \quad (\text{S5})$$

Where the same number of modes N_θ is considered for the rock and air temperatures. The complex coefficients $\Theta_{r,k}$ are such that $\Theta_{r,-k}$ is the conjugate of $\Theta_{r,k}$. They are related to the real amplitudes ($\theta_{r,k}$ and $\theta_{a,k}$) and phase shifts ($\varphi_{r,k}$ and $\varphi_{a,k}$) through the standard relations:

$$\theta_{r,0} = \Theta_{r,0} \quad \text{and} \quad \theta_{a,0} = \Theta_{a,0} \quad (\text{S6})$$

$$\theta_{r,k} = 2\sqrt{\Theta_{r,k}\Theta_{r,-k}} \quad \text{and} \quad \theta_{a,k} = 2\sqrt{\Theta_{a,k}\Theta_{a,-k}} \quad \text{for } k > 0 \quad (\text{S7})$$

$$\tan \varphi_{r,k} = -\frac{j(\Theta_{r,k} - \Theta_{r,-k})}{\Theta_{r,k} + \Theta_{r,-k}} \quad \text{and} \quad \tan \varphi_{a,k} = -\frac{j(\Theta_{a,k} - \Theta_{a,-k})}{\Theta_{a,k} + \Theta_{a,-k}} \quad \text{for } k > 0 \quad (\text{S8})$$

Injecting Eqs. (S3-S5) in the mathematical model of section 2.6 yields the equations:

$$\frac{Re}{2j} \sum_{m=1,3,\dots}^{N_\mu} \mu_m \left(\frac{d\Theta_{a,k-m}}{d\tilde{x}} - \frac{d\Theta_{a,k+m}}{d\tilde{x}} \right) = \left(4 \frac{k_r}{k_a} Pr^{-1} \right) \tilde{\Phi}_{w,k} + RePe^{-1} \frac{d^2 \Theta_{a,k}}{d\tilde{x}^2} \quad (\text{S9})$$

$$\Theta_{a,k}(\tilde{x}=0) = \frac{1}{2j} \text{ for } k=1, \quad \Theta_{a,k}(\tilde{x}=0) = 0 \text{ for } k \neq 1 \quad (\text{S10})$$

$$\Theta_{a,k}(\tilde{x} = \tilde{L}_{dom}) = 0, \quad (\text{S11})$$

$$2\pi jk\Theta_{r,k} = \frac{1}{\tilde{r}} \frac{\partial}{\partial \tilde{r}} \left(\tilde{r} \frac{\partial \Theta_{r,k}}{\partial \tilde{r}} \right) + \frac{\partial^2 \Theta_{r,k}}{\partial \tilde{x}^2} \quad (\text{S12})$$

$$\Theta_{r,k}(0, \tilde{r}) = \frac{1}{2j} \text{ for } k=1, \quad \Theta_{r,k}(0, \tilde{r}) = 0 \text{ for } k \neq 1 \quad (\text{S13})$$

$$\frac{\partial \Theta_{r,k}}{\partial \tilde{x}}(\tilde{x} = \tilde{L}_{dom}, \tilde{r}) = 0 \text{ and } \frac{\partial \Theta_{r,k}}{\partial \tilde{r}}(\tilde{x}, \tilde{r} = \tilde{R}_{dom}) = 0 \quad (\text{S14})$$

$$\text{Case A: } \tilde{\Phi}_{w,k} = \frac{\partial \Theta_{r,k}}{\partial \tilde{r}}(\tilde{x}, \tilde{R}) \quad (\text{S15})$$

$$= Bi \sum_{n=0,2,\dots}^{N_\eta} \frac{\eta_n}{2} \left[\Theta_{r,k-n}(\tilde{x}, \tilde{R}) - \Theta_{a,k-n}(\tilde{x}) + \Theta_{r,k+n}(\tilde{x}, \tilde{R}) - \Theta_{a,k+n}(\tilde{x}) \right]$$

$$\text{Case B: } \tilde{\Phi}_{w,k} = \frac{\partial \Theta_{r,k}}{\partial \tilde{r}}(\tilde{x}, \tilde{R}) \text{ and } \Theta_{a,k}(\tilde{x}) = \Theta_{r,k}(\tilde{x}) \quad (\text{S16})$$

Where $j = \sqrt{-1}$ and k is an integer varying from 0 to N_ϑ . Eqs. (S9) and (S12) thus define a set of $2(N_\vartheta + 1)$ coupled partial differential equations (PDE) with boundary conditions (S10-S11, S13-S16). This set of PDE, which does not include the time variable $\tilde{t}$, is solved numerically by finite elements (Galerkin method with quadratic Lagrangian elements, Comsol Multiphysics software). The resolution yields the complex coefficients $\Theta_{r,k}(\tilde{x}, \tilde{r})$ and $\Theta_{a,k}(\tilde{x})$. The real amplitudes and phase shifts are deduced from relations (S7-S8) and the temperatures as a function of time from relations (S4-S5).

Text S2

The choice of N_μ , N_η and N_ϑ results from a compromise between the accuracy of the results on one hand, the difficulty of the implementation and the available computational resources on the other hand. The sensitivity of the simulation results to N_ϑ , N_μ and N_η is investigated below for case A with $\tilde{R} = 0.2$ and $Re = 7.2 \times 10^5$. We first set $N_\eta = 0$ and $N_\mu = 1$ and we investigate the effect of N_ϑ . Figure **S1** displays the effect of N_ϑ on the time series of the wall temperature at $\tilde{x} = 10$ and $\tilde{x} = 100$. The general shape of the curve is little modified for $N_\vartheta \geq 6$, but the description of the rapid oscillations around $\tilde{t} \approx 0.05$ and $\tilde{t} \approx 0.45$ requires a larger number of modes. The same approach is used to investigate the effect of N_η and N_μ . Figure **S2** shows that these parameters have little effect on the results for the selected configuration. In all the simulations, we imposed $N_\vartheta = 18$, $N_\eta = 6$ and $N_\mu = 7$ which happened to be a reasonable compromise between accuracy and complexity. We checked on a few configurations that increasing the number of modes above the selected values did not significantly changed the results. In order to confirm its validity, the numerical solution based on Fourier series was compared with the solution obtained by time integration (Galerkin method with quadratic Lagrangian elements, time discretization using implicit Backward Differentiation Formula, Comsol Multiphysics software). The TI simulation was done over 28 cycles (i.e., from time $\tilde{t} = 0$ to $\tilde{t} = 28$) taking as the initial condition the solution obtained from Fourier series at $\tilde{t} = 0$. The comparison between TI (last cycle, from time $\tilde{t} = 27$ to 28) and FS is displayed in Fig. **S3**. Small discrepancies are observed during the

rapid oscillations around times $\tilde{t} \approx 27.05$ and $\tilde{t} \approx 27.45$ for $\tilde{x}=10$ and $\tilde{x}=100$, but the overall agreement is excellent.

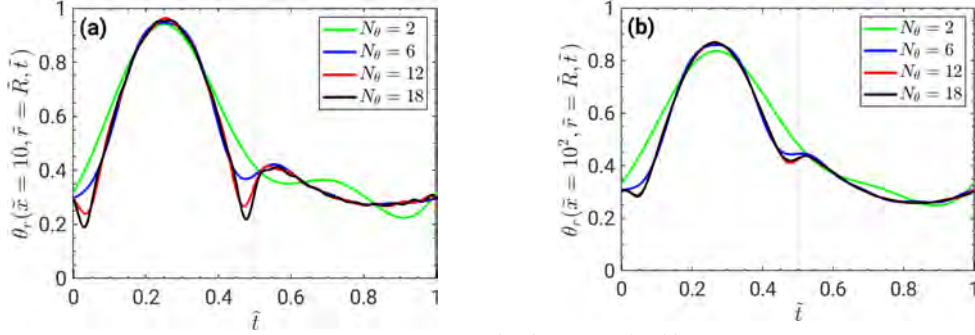


Figure S1. Time evolution of the wall temperature at $\tilde{x}=10$ (a) and $\tilde{x}=100$ (b) for $N_\eta = 0$, $N_\mu = 1$ and different values of N_θ . Case A with $\tilde{R}=0.2$ and $Re = 7.2 \times 10^5$.

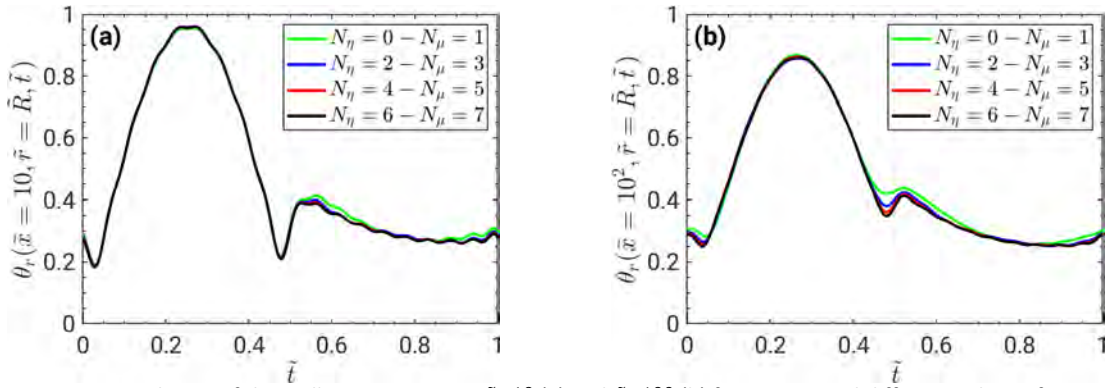


Figure S2. Time evolution of the wall temperature at $\tilde{x}=10$ (a) and $\tilde{x}=100$ (b) for $N_\theta=18$ and different values of N_η and N_μ . Case A with $\tilde{R}=0.2$ and $Re = 7.2 \times 10^5$.

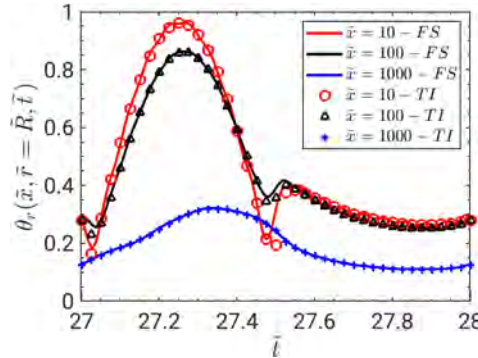


Figure S3. Time evolution of the conduit wall temperature calculated using Fourier series (FS) or time integration (TI), at different distances from the entrance $\tilde{x}$. FS solution at $\tilde{t}=0$ used as the initial condition of TI simulation. Case A with $\tilde{R}=0.2$ and $Re = 7.2 \times 10^5$.

Text S3

A standard approach, not used in this study, would consist in solving the time-dependent problem by TI starting from an arbitrary initial condition (e.g., $\theta_r(\tilde{x}, \tilde{r}, \tilde{t}=0) = \theta_a(\tilde{x}, \tilde{t}=0) = 0$). The time-dependent solution converges to the periodic regime at infinite time. A good approximation can thus be obtained if the simulated time is long enough. Figures (S4-S6) display the parameters defined in Section 4 calculated from TI as a function of time or FS, for case A with $\tilde{R}=0.189$ and $Re = 1.8 \times 10^5$.

With the exception of $\tilde{W}_1$, TI results converge slowly to the periodic regime. Taking FS as a reference, TI yields errors at time $\tilde{t} = 50$ in the range from 10% to 30% for $\bar{\theta}_w$, $\Delta\theta_w$, $\tilde{L}_0$ and $\tilde{L}_1$. These errors are acceptable if rough estimates are required. In contrast, TI underestimates $\tilde{W}_0$ by an order of magnitude. To get a correct estimate of this parameter and improve the accuracy of others, TI would require to simulate a time of the order of the diffusion time $\tilde{W}_0^2 \approx 64^2 \approx 4 \times 10^3$, i.e. thousands of years. Such simulations would require computational times hardly compatible with current computer resources.

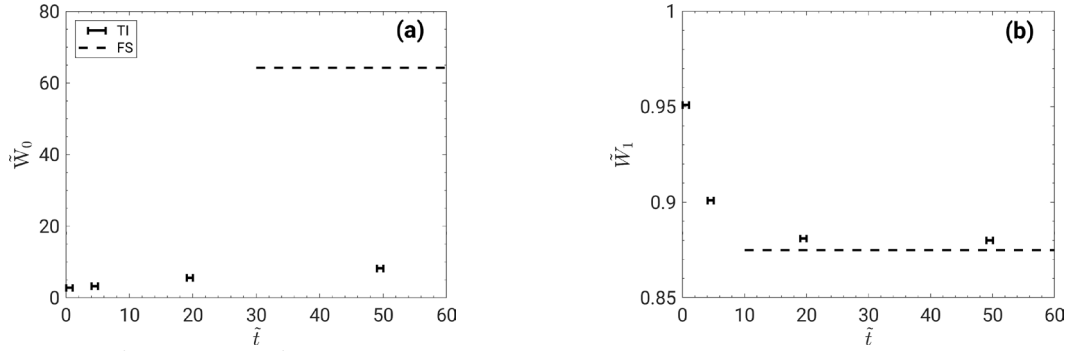


Figure S4. $\tilde{W}_0$ (a) and $\tilde{W}_1$ (b) estimated from Fourier series (FS) or time integration (TI). Initial conditions in TI simulations set to zero. Case A with $\tilde{R} = 0.189$ and $Re = 1.8 \times 10^5$.

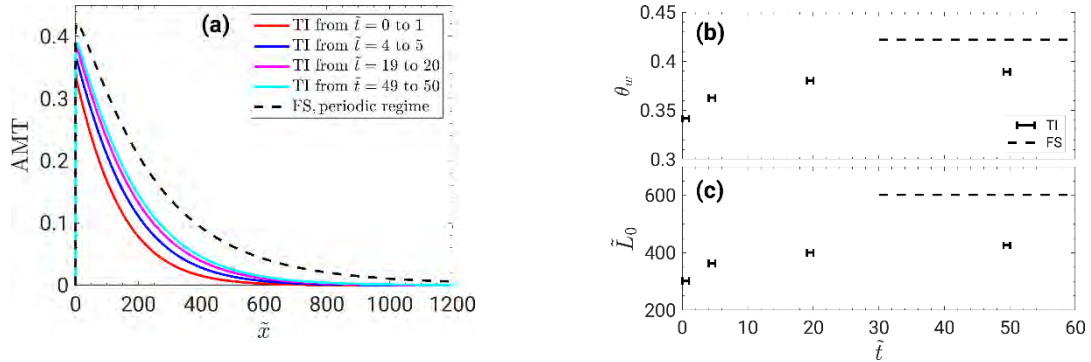


Figure S5. AMT of the conduit wall (a), $\bar{\theta}_w$ (b) and $\tilde{L}_0$ (c) estimated from Fourier series (FS) or time integration (TI). Initial conditions in TI simulations set to zero. Case A with $\tilde{R} = 0.189$ and $Re = 1.8 \times 10^5$.

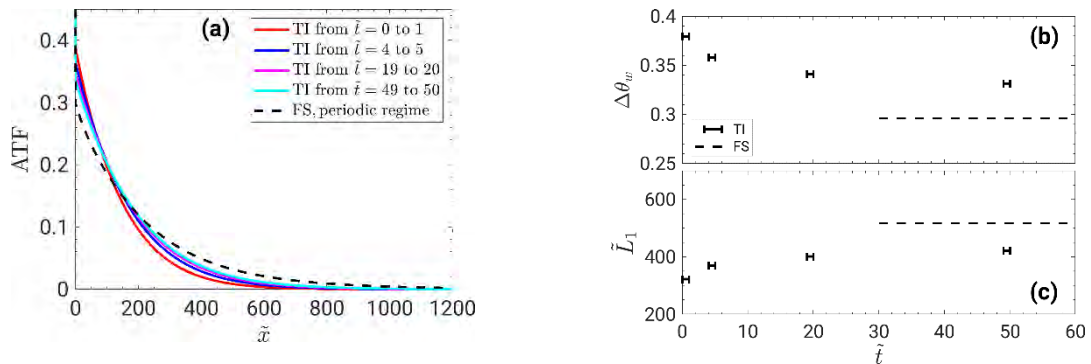


Figure S6. ATF of the conduit wall (a), $\Delta\theta_w$ (b) and $\tilde{L}_1$ (c) estimated from Fourier series (FS) or time integration (TI). Initial conditions in TI simulations set to zero. Case A with $\tilde{R} = 0.189$ and $Re = 1.8 \times 10^5$.

Chapter 8. Dynamics of chimney-effect ventilation and heat exchange: a model from Longeague Cave (unpublished paper)

Abstract

This study presents a model of heat exchange and conduit resistance in Longeague Cave (Swiss Jura Mts.), a cave ventilated by the chimney effect. In the karst vadose zone, heat exchange is mainly controlled by advective flows of water and air, with the latter becoming dominant in conduits abandoned by water. The chimney effect, driven by air density differences between the underground and the external atmospheres and due to multiple entrances at different altitudes, is largely considered the most widespread mechanism instigating ventilation. Caves ventilated by this mechanism show cold and warm zones (compared to the external mean annual temperature) in the lower and upper entrance zones, respectively. The distance to which these anomalies penetrate the cave is called the convection length or heterothermic zone. Through multi-year monitoring of temperatures and airflow in Longeague Cave (a double entrance and L-shaped cave in the Swiss Jura Mts.), we identified the convection length and quantified the energy exchanged at the air/rock interface along the ventilated conduit. The results show that the warm and cold convection lengths of the 300-meter-long cave were found to be comparable to half of its length (150 m), indicating the absence of a deep (homothermic) karst zone in the ventilated section. This study demonstrates that, in a cave ventilated by the chimney effect, most of the energy exchange (approximately 90%) primarily occurs in the entrance zones. Moreover, a reliable aeraulic model was derived from field data, linking flow rate to driving pressure and revealing that the cave behaves similarly to a Tesla Valve, exhibiting different aeraulic resistance for upward and downward flows. Furthermore, changes in the ventilation regime, affected by fluctuations in the level of an internal lake that can block the air passage, can significantly alter heat exchange and the cave's thermal response to external climate variations. This research contributes to a broader understanding of cave ecosystems, climate regulation and sustainable cave management practices. Eventually, the developed aeraulic model allows airflow estimation based on internal and external temperatures.

8.1. Introduction

In the karst vadose zone, heat exchanges and subsequent temperature distribution are mainly controlled by advective fluxes of water and air (Badino, 2010), whilst the latter may become dominant in cave passages abandoned by water (Luetscher et al., 2008). For well-developed Alpine karst systems, Badino (2010) assesses the minimum specific airflow to range between 1 and 10 kg/s/km², whereas Luetscher and Jeannin (2004) suggest an average rate of between 75 and 150 kg/s/km². Cave airflow can be triggered by various mechanisms, among which forced natural convection (mostly known as the chimney effect or also buoyancy-driven airflow) is considered the most prevalent at the scale of a karst massif (Kukuljan et al., 2021a; Covington and Perne, 2015; Badino, 2010; Luetscher and Jeannin, 2004). The chimney effect sets in the presence of multiple entrances at different altitudes when a difference in air density between the underground environment and the external atmosphere leads to pressure gradients driving subterranean airflow. During cold phases (outside temperature lower than the inside) the warm cave air rises toward the upper entrances, drawing in cold air at the lower entrances. Conversely, when the outside temperature is warmer than the inside (warm phases) the colder inner

air falls toward the lower entrance, drawing warm air in through the upper entrances. Diurnal and seasonal temperature fluctuations around the external annual average induce airflow reversals (Sedaghatkish et al., 2024; Lismonde, 2002; Wigley and Brown, 1971). As a consequence of flow reversals, lower inlets show an average annual temperature lower than the external Annual Mean Temperature (AMT), while upper inlets display an average temperature higher than the AMT.

Depending mainly on the geometry of the cave (number and size of entrances, conduit diameters, direct connection to the water table, etc.), several other mechanisms can instigate air flow inside caves (Covington and Perne, 2015; Wigley and Brown, 1976; Cigna, 1968), including fluctuations in atmospheric pressure producing barometric airflow (e.g. Gomell and Pflitsch, 2022; Pflitsch et al., 2010), airflow due to water drag (Palmer, 2007; Cigna, 1968), surface wind-driven flow (e.g. Kukuljan et al., 2021a), airflow due to floodwaters changing the volume of air within the system (Pivetta et al., 2021; Gabrovšek et al., 2018) as well as convective cells (e.g. Qaddah et al., 2022; Guerrier et al., 2019). A combination of these phenomena is also possible. Besides geometries, air composition and characteristics (e.g. seasonal variation in CO₂, moisture content or geothermal heat) can cause air density changes which could affect the cave air flow rate (Sánchez-Cañete et al., 2013).

Airflow largely controls cave climate (Luetscher et al., 2008; Luetscher and Jeannin, 2004) and quantitative airflow and temperature measurements are relevant for the interpretation of paleoclimate proxy-records from cave speleothems and sediments (Cuthbert et al., 2014), for assessing the evolution of the habitats (Medina et al., 2023; Mammola et al., 2019) of numerous specific and often endemic species, for groundwater quality (Jeannin et al., 2017) and even for evaluating the carbon cycle in a karst system (Breecker et al., 2012; Martin-Pozas et al., 2022; Prelovšek et al., 2018) as it controls cave air CO₂ concentrations (Kukuljan et al., 2021a; Kowalczyk and Froelich, 2010; Faimon and Lang, 2013). Ventilation also has a large impact on ice bodies preserved in caves (Obleitner et al., 2024; Spötl et al., 2024; Gabrovšek, 2023; Obleitner and Spötl, 2011; Luetscher, 2005). Airflow measurements in caves are required to quantify heat and mass transfers advected by air and authors such as Luetscher et al. (2008) relied on airflow monitoring to assess the air energy balance in ventilated caves and its impact on underground ice masses.

Pulina (1969) identified what he called a “*dynamic zone*” showing temperature variations $\geq 0.5^{\circ}\text{C}$ (at a yearly time scale) characterizing the first tens of meters of a cave (typically between 50–100 m from the entrance), and a “*static zone*” with annual temperature oscillations $< 0.5^{\circ}\text{C}$ (at a distance usually $> 50/100$ m). Later, in 1985, Drogue renamed them *heterothermic* and *homothermic* zones, respectively. This classification, still accepted, was used by Luetscher and Jeannin (2004) to develop a conceptual model of thermal characteristics in a karst system.

Three thermal sub-zones can be distinguished and characterised by different vertical gradients: 1) the conductive zone, 2) the heterothermic zone, and 3) the homothermic zone (Luetscher and Jeannin, 2004; Jeannin et al., 1997). Conductive heat fluxes related to annual fluctuations of the outside conditions, including external radiation, are known to reach a depth below the ground surface in the order of a few meters ($\sim 5\text{--}10$ m; Sedaghatkish et al., 2024).

Although thermal stability in deep karst systems is related to the heat capacity of the bulk rock (Badino, 1995), which dampens external fluctuations, even small temperature differences between the air and the mass rock allow for energy exchanges to and from the cave walls (Luetscher et al., 2008).

The heterothermic zone, which characterizes the upper part of the vadose zone, extends from the karst surface and is estimated to be between 50 and 100 metres deep. In this region, air and water fluxes from the surface flow through rock voids while transporting and exchanging heat. As it takes a certain distance for the air and water to equilibrate with the rock temperature, the gradient in the heterothermic zone can be positive or negative depending on external conditions. Temperatures in the heterothermic zone are thus controlled by water and air circulation.

The homothermic zone identifies the deeper part of the karst where the temperature oscillations are very low or nil. Some authors defined it as the region where annual temperature variations are lower than 1 or 0.5°C (Mavlyudov, 1997; Jeannin, 1990; Pulina, 1969), but an attenuation of 90% of the amplitude of the outside daily temperatures is probably a better definition (Sedaghatkish et al., 2024). In Alpine karst, this homothermic zone is characterised by vertical thermal gradients ranging between 0.234 and 0.55°C/100 m since it is constrained by the physical gradients of water and humid air, respectively. In the presence of a high water flow rate and low air circulation, the air temperature gradient tends to about 0.3 °C/100 m, close to the energy dissipation rate of water (i.e. 0.234°C/100 m; Luetscher and Jeannin, 2004; Jeannin et al., 1997).

Looking at the temperature variations along a single cave conduit, Wigley and Brown (1971) proposed a 1-D numerical solution of air temperature and moisture content along a unidirectional ventilated pipe, accounting for cold and warm temperature anomalies (compared to the external AMT) at the pipe inlet. Their model was based on mass and energy conservation and has shown a “thermal relaxation length”, the distance to which the temperature of an incoming airflow is damped by a factor $1/e=0.37$ (i.e. where the temperature fluctuation is reduced by 63%). However, their model assumes a constant wall temperature with distance from the inlet, which is a major simplification since the heat transfer due to air convection and rock conduction are not coupled. Therefore, in Wigley and Brown's (1971) model, the bulk-rock temperature does not play any role (Sedaghatkish et al., 2024). Lismonde (2002), Gabrovšek (2023) and Sedaghatkish et al. (2024) clarified that the rock temperature has a major role in determining air temperature profiles along karst conduits and in turn, the airflow temperature modifies the rock temperature. The first two authors (Gabrovšek, 2023; Lismonde, 2002) developed two similar models to estimate cave convection length and ventilation. However, their models lack in comparison with field data. In recent work (Sedaghatkish et al., 2024), we proposed a similar model to investigate quantitatively the convection length along ventilated caves giving orders of magnitude. We concluded that turbulent airflow, very common in ventilated caves, significantly modifies the rock temperature. The results of the model presented in Sedaghatkish et al., (2024) were validated against field measurements. Air/wall temperature distributions allow for the differentiation of three thermic zones (Sedaghatkish et al., 2024): 1) diffusive, 2) convective and 3) deep karst zones are identified according to the dominating heat transfer process. The first zone, between 0 and a few meters (~5 m at yearly time scale) from the cave entrance, is dominated by heat diffusion through the rock from outside to the interior and it is related to the rock diffusivity. After this distance, forced convection dominates the heat exchange. In this region, called the convective zone, still-significant fluctuations follow variations of the external temperatures, although they are buffered with distance. The length of the convective zone, also called “convection length”, is approximately proportional to the amplitude of the flowrate annual fluctuation divided by the square root of the cave's conduit radius (Sedaghatkish et al., 2024). The model presented by Sedaghatkish et al. (2024) estimates that for commonly encountered airflow rates in alpine regions (0.1 – 1.0 kg/s) and cave radius in the order of 1 m, the distance at which the thermal perturbation can propagate along the conduits is from a few tens to

hundreds of meters. While, the thermal perturbation propagates in the radial direction in the order of one-tenth of the convective length, i.e. from a few meters to some tens of meters.

Therefore, the region between the cave entrance and the deep karst zones, namely along the convective zone, is the place where most energy exchange between the air coming from the outside and the rock surrounding the cave takes place. It can thus be interesting to pay greater attention to this part of the karst from an energetic point of view. The cave air energy balance was previously investigated in the context of ice caves (Obleitner and Spötl, 2011; Luetscher et al., 2008; Ohata et al., 1994a, 1994b). In ice caves, where heat exchanges happen at the interface between different water phases (liquid/solid/vapour) and with the surrounding rock, turbulent airflow plays a major role in removing/releasing latent heat from/to the system. Luetscher et al. (2008) quantified to -75 ± 8 GJ the net energy removed by the airflow over an annual cycle in Monlesi Cave (Switzerland), a cave located in the heterothermic zone. Moreover, the same authors used punctual flow rate measurements to infer the cave aeraulic resistance (R) to reconstruct the annual airflow time series. In fact, most authors (Sedaghatkish et al., 2024; Gabrovšek, 2023; Badino, 2010; Lismonde, 2002; Atkinson et al., 1983; Wigley and Brown, 1971) argue that the cave airflow depends on 1) the temperature difference between the internal and external atmospheres (establishment of a driving pressure, DP), 2) the cave geometries (conduit length, diameter, roughness and general cave shape) and that the annual mean temperature of cave entrances (convection length) should show a thermal shift from the external AMT (the so-called thermal anomalies).

8.1.1. Aeraulic resistance and its role in chimney effect ventilation

In a chimney effect, the cave mass flow rate ($\dot{m}$, in kg/s) is related to the balance between the friction and buoyancy terms of the momentum balance (Lismonde, 2002; Badino, 1995; Atkinson et al., 1983). Thus, neglecting inertia, the driving pressure DP can be deduced from the momentum balance applied to the atmosphere inside the cave (Lismonde, 2002):

$$\dot{m}(t) = \pm \sqrt{\frac{DP(t)}{R}}$$

Eq. 8.1

with DP (Pa) = $(\rho_{cave} - \rho_{ext})g\Delta z$, being approximated as $DP \approx \rho_{cave} \frac{g}{T_{atm}} \Delta T \Delta z$ (where $\Delta T = T_{cave} - T_{atm}$ in °K, Δz is the difference in altitude between cave entrances in meters and $\rho_{cave/ext}$ are air densities inside and outside the cave, as well as t is the time in seconds. The $\pm$ symbol is necessary to distinguish between upward (-) and downward (+) flow directions). The driving pressure is therefore dependent on density differences between external and internal atmospheres due to the existence of a ΔT . The cave aeraulic resistance (R), although not directly related to thermal exchange processes, is a critical parameter to define a cave aeraulic model since it puts in relation air mass flow rates with DP . R is a physical property of the cave which depends on the friction determined by conduits' roughness (Darcy friction factor, f), length and sinuosity, but it mainly depends on conduits' diameters. R is shown to be inversely proportional to the fifth power of the diameter (Lismonde, 2002; Atkinson et al., 1983), i.e. a small passage determines greater friction than a large one. Although geometrical parameters (conduit lengths and diameters) might be measured in the field, assessing the friction factor (f) for natural conduits is not straightforward. Caves under forced convection typically exhibit a turbulent flow (Reynolds number $Re \gg 2 \times 10^3$; Lismonde, 2002; Badino, 1995), then f depends weakly on Re and mainly on the relative roughness of the conduit (the ratio between the size of the asperities and

diameter), with the latter generally unknown. The friction factor for natural conduits can be approximated through the Moody chart (Moody, 1944) or by other empirical correlations (Haaland, 1983). For caves, f values reported in the literature are estimated to range from 0.04 to almost 1 (Albright and Springer, 2022; Springer, 2004; Jeannin, 2001; Atkinson et al., 1983) but it could be potentially larger (Zhou et al., 2016). Indeed, cave aeratic resistance R is difficult to estimate analytically for a specific investigated site. Therefore this parameter has to be inferred empirically from field data and can be approximated by plotting $\dot{m}^2 = \pm \frac{\Delta P}{R}$, where R^{-1} is the slope of the straight line and the $\pm$ indicates the direction of the flow (upward or downward). Different authors verified for some caves under a chimney effect the consistency with this relation (Garagnon et al., 2022; Kukuljan et al., 2021b; Kašing and Lenart, 2020; Covington and Perne, 2015; Faimon and Lang, 2013; Luetscher et al., 2008; De Freitas et al., 1982), but they also observed an asymmetry between upward and downward flow (i.e. $R_{upward} \neq R_{downward}$). This asymmetry could result from different conduit friction according to flow direction ($f_{upward} \neq f_{downward}$) but also from the ΔP approximation. Faimon and Lang (2013) and Gabrovšek (2023) explained that a same $|\Delta T|$ could produce different $|\Delta P|$ whether it happens in winter or summer, for different reasons including air density having non-linear variation with temperature, relative humidity, thermal effect of cave walls, cave shape and CO_2 concentration (Sánchez-Cañete et al., 2013). After Luetscher et al. (2008), Garagnon et al. (2022) used field data to extrapolate the aeratic resistance R for the upward and downward ventilation and built a ventilation model for Milandre Cave. Garagnon and coworkers (2022) showed that the model based on the proportionality given in Eq. 8.1 explains approximately 95% of the measured airflow, confirming the major role played by temperature in driving the air dynamics.

8.1.2. Work objectives

In this work, we aim to quantify the energy exchanged at the air/rock interface along the cave and to demonstrate that in a cave ventilated by a chimney effect, the heat advected by air is primarily exchanged in the entrance zones of the cave, accounting for cold and warm anomalies (Gabrovšek, 2023; Lismonde, 2002; Wigley and Brown, 1971). Furthermore, we will show that the convection length of a cave can be close to half of its length (i.e., the central part of the cave is not within the homothermic or deep karst zone) and that changes in the ventilation regime can influence heat exchange and the cave thermal response to external climate changes. Additionally, a reliable cave aeratic model is derived from field measurements. Once the model is constituted for a specific cave, the airflow may be inferred directly by knowing the internal and external temperatures. We present the results of airflow and temperature monitoring in Longeigue Cave (Switzerland), a cave ventilated by a chimney effect and favourable for the study of heat exchange along the convective length. The cave ventilation is directly related to the external temperature variations, which control both, its direction and intensity (Pastore et al., 2024a, 2024b). When the outside temperature is greater than inside, the cave airflow is descending. Vice versa, when the inner temperature is higher than the outside air, the air is drawn in at the lower cave entrance. A perennial lake in the lower cave section acts as a gate for airflow and closes the air passage during floods if the water level rises by ~ 1 m above the base level. We hypothesize that this particular cave feature may impact cave ventilation and climate.

By shedding light on the intricate processes governing cave ventilation and heat exchanges, this research contributes to our broader understanding of cave ecosystems, climate regulation and sustainable cave management practices.

8.2. Material and methods

8.2.1. Study Site

The “Baume de Longeaigne” (Longeaigne Cave from now on) opens in the Swiss Jura Mountains (Val-de-Travers; Upper entrance: 46°52'22" N 6°31'09" E, 917 m a.s.l.; Lower entrance: 46°52'18" N 6°31'09" E, 820 m a.s.l.). The cave (Figure 8.1) consists of one main branch with a variable diameter and a length of 0.3 km over a 97 m elevation difference between the lower and upper entrances. The cave shows larger passages along the subhorizontal lower part (~2.5 m as average diameter), with extensive phreatic subhorizontal galleries showing some vadose reshaping (shafts) morphologies, and becomes narrower in the upper part where the cave has squeezing conduits (~1.7 m as average diameter, with the smallest diameter of ~0.6 m²). In total, the explored length of the cave is ~1.3 km, including a series of sub-horizontal and sub-vertical secondary branches leading to depths of between -89 and -132 m from the upper entrance. These branches are not or only partially traversed by an airflow.

The cave is mainly dry and is crossed by an intense airflow between the two entrances. The lower entrance is a temporary spring acting as an overflow of the “Raies” hydrogeological system, whose catchment area is estimated to be between 60 and 80 km². The maximum discharge is assessed to be between 3 and 5 m³/s (Jeannin, 2018). During floods, a consequent rise of water, up to ~20 meters in elevation, is observed inside the cave (Figure 8.1). A perennial lake is present in the lower cave part and the shape of the cave below 827 m a.s.l. implies that this part of the cave can be completely or partially submerged. During high water conditions, the airflow is aborted.

Between the stations Anemo and T1, a side cave passage is present (dashed blue arrow in Figure 8.1) which may sometimes cause a local convection cell (Pastore et al., 2024b).

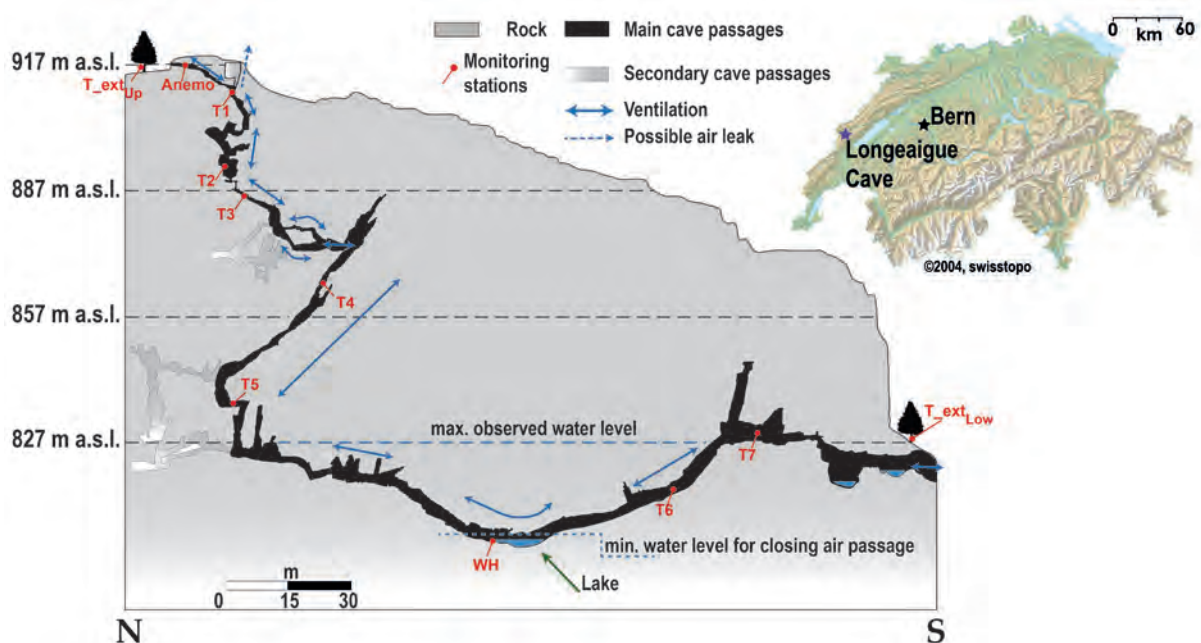


Figure 8.1 – An excerpt of Longeaigne Cave topography (modified after Jeannin, 2018). The red points indicate the monitoring stations and the arrows the ventilation.

8.2.2. Temperature measurements

From August 2020 to August 2023, we installed 8 stations along the ventilated cave conduit for monitoring the air temperature (T-stations and Anemo_Temp. in Figure 8.1 and Table 8.1). The stations are deployed at intervals of $\sim 10 - 30$ m along the main airflow pathway, except between T5 and T6 where the distance is ~ 100 m. External temperature is monitored at two external stations close to the upper (T_{extUp}) and lower (T_{extLow}) entrances.

All the sensors at T-stations, listed in Table 8.1, were tested against a reference thermometer (Almemo 2290-4; accuracy $\pm 0.01^\circ\text{C}$) in an ice bath in an isothermic box and a linear correction was obtained. This allowed us to have a homogeneous calibration and reduce the relative error between the sensors (accuracy column in Table 8.1). In contrast, the Anemo station (which hosts the airflow meter, see next Chapter 8.2.3) is equipped with a temperature sensor type MS8607 (Adafruit) not calibrated against the reference thermometer, therefore we refer to the manufacturer accuracy of $\pm 1^\circ\text{C}$ and resolution of 0.01°C .

STATION NAME	DISTANCES FROM THE UPPER (AND LOWER) ENTRANCE	ELEVATION	PROBE TYPE	ACCURACY	RESOLUTION
UPPER EXT. ST.	-	917 m a.s.l.	Reefnet	$\pm 0.30^\circ\text{C}$	0.01°C
ANEMO	4 m (310 m)	917 m a.s.l.	SFM (airflow)	$\pm 5\%$, $0.5\text{ m/s} < V$ $\pm 26\%$, $0.5\text{ m/s} > V$	Up to 16 bits
ANEMO_TEMP	4 m (310 m)	917 m a.s.l.	Adafruit MS8607	$\pm 1^\circ\text{C}$	0.01°C
T1	16 m (298 m)	911 m a.s.l.	Hobo U22/U23	$\pm 0.20/\pm 0.10^\circ\text{C}$	$0.04^\circ/0.02^\circ\text{C}$
T2	42 m (272 m)	892 m a.s.l.	Hobo U22/U23	$\pm 0.20/\pm 0.10^\circ\text{C}$	$0.04^\circ/0.02^\circ\text{C}$
T3	55 m (259 m)	884 m a.s.l.	Hobo U22/U23	$\pm 0.20/\pm 0.10^\circ\text{C}$	$0.04^\circ/0.02^\circ\text{C}$
T4	97 m (218 m)	865 m a.s.l.	Hobo U22	$\pm 0.20^\circ\text{C}$	0.04°C
T5	140 m (174 m)	837 m a.s.l.	Reefnet	$\pm 0.30^\circ\text{C}$	0.01°C
WH (water height)	217 m (97 m)	803 m a.s.l.	Reefnet	-	-
T6	269 m (45 m)	816 m a.s.l.	Hobo U22	$\pm 0.20^\circ\text{C}$	0.04°C
T7	294 m (20 m)	820 m a.s.l.	Reefnet	$\pm 0.30^\circ\text{C}$	0.01°C
T8	outside the ventilated conduit	808 m a.s.l.	Reefnet	$\pm 0.30^\circ\text{C}$	0.01°C
LOWER EXT. ST.	-	820 m a.s.l.	Reefnet	$\pm 0.30^\circ\text{C}$	0.01°C

Table 8.1 – Used devices, distances from the upper (and lower in brackets) entrance and elevations of the monitoring stations in Longaigue Cave.

Five temperature stations were equipped with Hobo U22-01 Water Temperature Pro v2 (Table 8.1). Reefnet Sensus Ultra temperature and pressure devices were deployed at three internal and two external stations (Table 8.1). The sensors measured at an hourly time step.

The upper stations (from T1 to T5) are along the narrower, vertical and dry cave part. In contrast, the probes below 827 m a.s.l. (T6 and T7, Figure 8.1) can be flooded during high-water events. These

conditions forced us to use water-proof probes (Hobo U22 and Reefnets) and hang them tightly on the conduit walls, a few centimetres from the rock surface. Therefore, wall temperatures may influence air temperature measurements. To evaluate the error of temperature measurements due to the rock vicinity, in August 2022 we added U23-04 Water Temperature Pro v2 probes (from now on Hobo U23) at stations T1, T2, T3 and T7. These Hobo loggers are equipped with a wired external temperature probe type TMC1-HD and were installed in cross sections to measure as far as possible from the conduit walls. We estimated the error due to close-to-wall measurements is $\pm 20\%$ for stations in the first 30 m from entrances (T1 and T7) and $\pm 5\%$ for more internal stations. During high water conditions, when the lower part of the cave is traversed by water, sensors T6 and T7 are influenced by water temperature.

An additional sensor was placed at station T8 located in the non-ventilated part of the cave (outside the topography excerpt in Figure 8.1). This zone is generally dry when the water table is low, while it is submerged when the water table rises. Temperature measurements from this sensor will be used only in the discussion, for a short comparison with the ventilated conduit.

Although some of the probes failed and data periods are missing (i.e. 16/10/20 – 29/05/21; 17/06/22 – 10/08/22), meaningful considerations can be made on cave temperature dynamics.

8.2.3. Airflow measurements

The air velocity is measured with a Sensirion SFM3003-300-CE (hereafter referred to as SFM, Figure 8.2.a). The device is a digital bidirectional (positive and negative values) mass flowmeter, compact and sturdy, with a high data resolution (up to 16 bits). In Pastore et al. (2024a), we provided a reliable calibration and conversion function (from Standard Litre per Minute to m/s) for this device. The accuracy of the instrument is $\pm 5\%$ of the measured value for velocities above 0.50 m/s and $\pm 26\%$ for velocities between $Vel. < 0.50$ m/s. The air velocity is measured every hour in the cave upper entrance conduit (station Anemo in Figure 8.1) at around 5 m from the inlet in a 0.5 m^2 cross-section. The instrument is installed close to the wall (Figure 8.2.b) to keep the passage clear and avoid accidental damage by cavers because the cross-sectional area is rather small (crawling passage in 0.5 m^2 cross-section). Since the SFM is positioned close to the wall, we determined through in-place tests that the measured airflow likely represents the error lower bound of airflow measurements. Consequently, we estimate the error upper bound to be up to 50% higher than the SFM measurements. For a short period (from January to June 2022) a second SFM was deployed at station T3, with its data reported only for the discussion in Chapter 8.2.3.

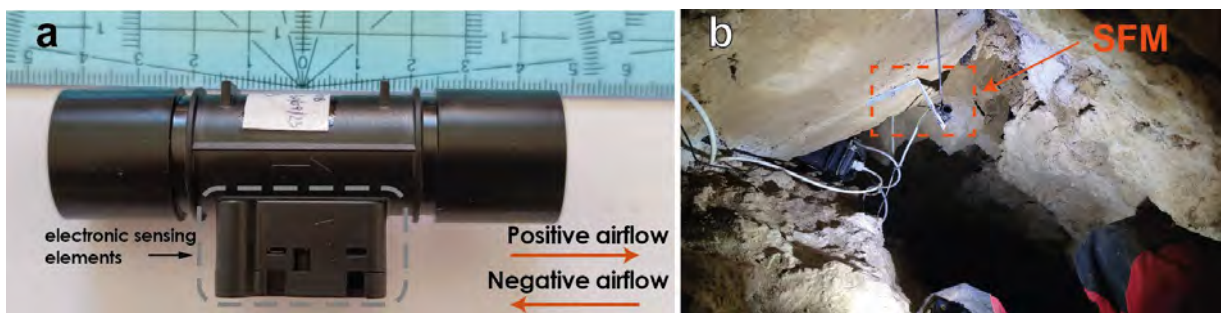


Figure 8.2 – Sensirion SFM 3003-300 CE.

8.2.4. Water height monitoring

At the WH station (Figure 8.1 and Table 8.1), a Reefnet sensor (press. resol. 1 mbar, accuracy ± 3 mbar) records variations in the water column that cause airflow to stop. The cave lake is located at the base of a sort of gooseneck and the sensor was installed at a height equal to the minimum level the water

has to reach to fill the cross-section and close the air passage (~ 1 m above the low water condition, Figure 8.1).

The pressure of the water column is measured at an hourly time step and corrected for air pressure measured at station T5 (~ 33 m above WH) considering $1 \text{ mbar} = 0.0102 \text{ m of water column (m}_{\text{H}_2\text{O}})$.

8.2.5. Assessment of the cave aeraulic resistance

In the case of a natural forced convection, the airflow through the cave can be estimated from the momentum balance equation (Eq. 8.1).

In Eq. 8.1, the driving pressure DP represents the pressure difference at the two extremities of the cave determined by the density difference between internal and external air columns (because of different air temperatures). The assumption to take a single conduit to determine DP , with constant slope and temperature, does not well apply to the complexity of Longeague Cave. Thus, we discretized the cave into eight different conduits (Figure 8.3 and Table 8.2), and the driving pressure DP is defined as follows:

$$DP(t) = \frac{P_{atm} \cdot g}{R_s \cdot T(t)^2_{atm}} \left| \sum_{i=1}^{N=8} [T_i(t) - T_{atm(z_i)}(t)] \cos(\theta_i) L_i \right|$$

Eq. 8.2

where P_{atm} is the atmospheric pressure (Pa), g is the gravitational acceleration (9.81 m/s^2), R_s is the air-specific gas constant (287 J/kg/K), T_{atm} is the external temperature ($^{\circ}\text{K}$), whereas T_i is the temperature of the i^{th} cave transect and t is the time (s). Then the buoyancy term (the term in the summation of Eq. 8.2) is determined for each i^{th} cave transect, for which the angle with the gravity vector (θ_i) and the length (L_i) are shown in Table 8.2. Whereas, the conduit temperature is considered homogeneous for its length and is retrieved from the T_i station positioned at the nod between two transects, according to Table 8.2. Other methods to average the temperatures along the cave (e.g. by using an average temperature between two nods) were tested leading to small differences in the assessed DP .

The outside temperature (T_{atm}) is corrected as a function of the elevation (z_i) of each station (Table 8.1) by the linear relation:

$$T_{atm(z_i)} = \frac{(T_{atm_UE} - T_{atm_LE})}{\Delta H} \Delta z_i + T_{atm_LE}, \quad P_{atm} = 1[\text{atm}]$$

Eq. 8.3

where $T_{atm(z_i)}$ is the external temperature corrected by the difference in altitude (Δz_i) between the lower entrance (taken as the reference point at 0 m) and the i^{th} conduit altitude. T_{atm_UE} and T_{atm_LE} are the external temperatures measured at the upper and lower entrances, and finally ΔH is the difference in altitude between them (90 m).

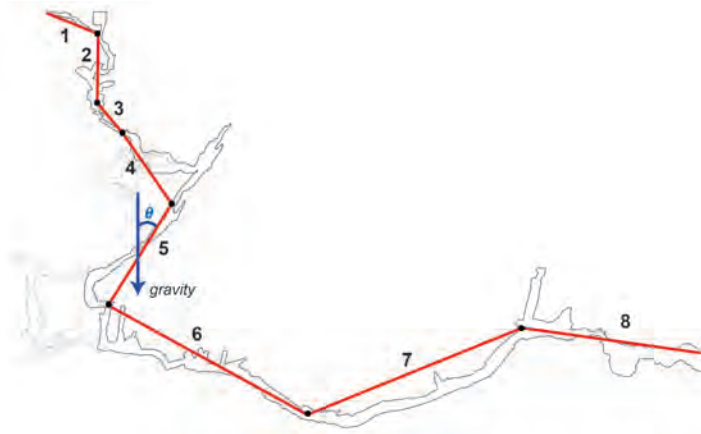


Figure 8.3 – Cave discretization into different transects i^{th} of length L_i and slope θ_i (see Table 8.2).

Section index i (from upper to lower entrance)	Angle with g vector (θ_i)	Air temperature for T_i (at station i)	Conduit length L_i (m)	Conduit average diameter D_i (m)
1	64°	T1	11.7	0.91
2	0°	T2	17.6	1.40
3	28°	T3	10.7	1.28
4	50°	T4	24.8	1.67
5	37°	T5	40.7	2.24
6	63°	T5	71.5	2.70
7	112°	T6	66.4	2.55
8	81°	T7	42.3	2.55

Table 8.2 –Transect geometrical properties and corresponding monitoring stations for transect temperature T_i .

The cave aeraulic resistance R can be inferred by plotting $\dot{m}$, derived from cave measurements, against DP (Eq. 8.2). In the result Section 8.3.3 (“Aeraulic model”) we show that the cave airflow follows Eq. 8.1 and that the cave resistance can be estimated. The validation of the aeraulic model is obtained by solving back Eq. 8.1 using the inferred R and the known DP for assessing $\dot{m}(t)$. The estimated airflow $\dot{m}(t)$ is then compared with the measured one.

8.2.6. Air-rock heat exchanges along the ventilated conduit

To assess the energy exchanged between air and cave walls, we impose the following system boundary conditions.

The boundary limits are defined by the convection length (L_c) as the distance at which the amplitude of the Annual Temperature Fluctuation (ATF) is reduced by 90%. We also define the convection width (W_c) as $0.1 \times L_c$ (Sedaghatkish et al., 2024). The deep karst (D_k) is the zone where the temperature fluctuations are buffered by >90%, and rock and air temperatures are close to equilibrium. We evaluate the amount of energy exchanged between the air and the rock on a three-year time scale, therefore the system boundary conditions (Figure 8.4) are defined 1) from the limit of the seasonal diffusive zone (L_f in Figure 8.4, ~ 10 m thick), 2) at a distance W_c from the conduit, the rock temperature is constant (W_c depends on the considered time scale). Therefore, we assume that all the heat is exchanged with the advected air and within the convection length (L_c) and is stored in the surrounding rock within the convection width (W_c). During wet cave conditions, when the airflow is stopped by rising water in the

1. Radiation
2. Advection
3. Conduction
4. Diffusion

L_c : convection length
 w_c : convection width $\approx 0.1 L_c$
 L_r : diffusion length

D_K : Deep karst
AMT : Annual Mean Temperature
ATF : Annual Temperature Fluctuation

Therefore, energy exchanges over three years can be estimated as:

$$\Delta E_{air} = - \int_0^{3yrs} |\dot{m}| \cdot c_{air} \cdot \Delta T \, dt \approx - \sum |\dot{m}| \cdot c_{air} \cdot \Delta T \cdot \Delta t = -\Delta E_{rock}$$

where $\Delta E_{air/rock}$ (J) is the air/rock system enthalpy, $\dot{m}$ is the mass flow rate of dry air (kg/s), ΔT is the difference in temperature ($^{\circ}\text{C}$) ($\Delta T = T_{T7} - T_{Ext}$ for a downward flow; $\Delta T = T_{T1} - T_{Ext}$ for an upward flow); c_{air} is the specific heat of dry air ($=1002 \text{ J/kg}\cdot^{\circ}\text{C}$) and $\Delta t=3600 \text{ s}$ is the logging interval between two measurements. The stations T1 and T7 are the last stations along the cave for the upward and downward regimes (Table 8.1), respectively. Therefore, those temperatures represent the temperature reached by the air along the cave before exiting from the outlet. We consider that during downward airflow ($\Delta T < 0$, $\dot{m} > 0$), the rock gains energy from the air, thus $\Delta E_{rock} > 0$ and $\Delta E_{air} < 0$. Conversely, when $\Delta T > 0$ and $\dot{m} < 0$ (upward ventilation) the air removes energy from the system, therefore $\Delta E_{rock} < 0$ and $\Delta E_{air} > 0$.

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Therefore, Eq. 8.4 can be rewritten as follows:

$$\Delta E / L_{Input \rightarrow Output} = - \frac{\int_0^t \dot{m} \cdot c_{air} \cdot \Delta T_{(T_{Output} \leftarrow T_{Input})} dt}{L_{Input \rightarrow Output}}$$

Eq. 8.5

$\Delta E / L_{Input \rightarrow Output}$ (J/m) is the energy exchanged between two consecutive stations normalised for the distance $L_{Input \rightarrow Output}$ (m) between them, $\Delta T_{(T_{Output} \leftarrow T_{Input})}$ is the difference between the temperature measured at an input station T_{Input} and that measured at an output station T_{Output} . For example, $T_{Input} = T_{ext}$ and $T_{Output} = T_{T1}$ for the first station inside the cave during a downward flow, then $T_{Input} = T_{T1}$ and $T_{Output} = T_{T2}$ for the second station and so on. The same for an upward flow ($T_{Input} = T_{ext}$ and $T_{Output} = T_{T7}$).

Owing to the measurement error on the airflow ($\pm 50\%$), the main error in the energy assessment $\Delta E_{air/rock}$ arises from this. Therefore, the amount of energy is evaluated for a minimum and a maximum value.

8.3. Results

For assessing the heat exchanges and the cave aeraulic resistance, we measured the airflow (in m^3/s) and the air temperatures along the cave during the period between 13 August 2020 and 13 August 2023 (Figure 8.5.a and b).

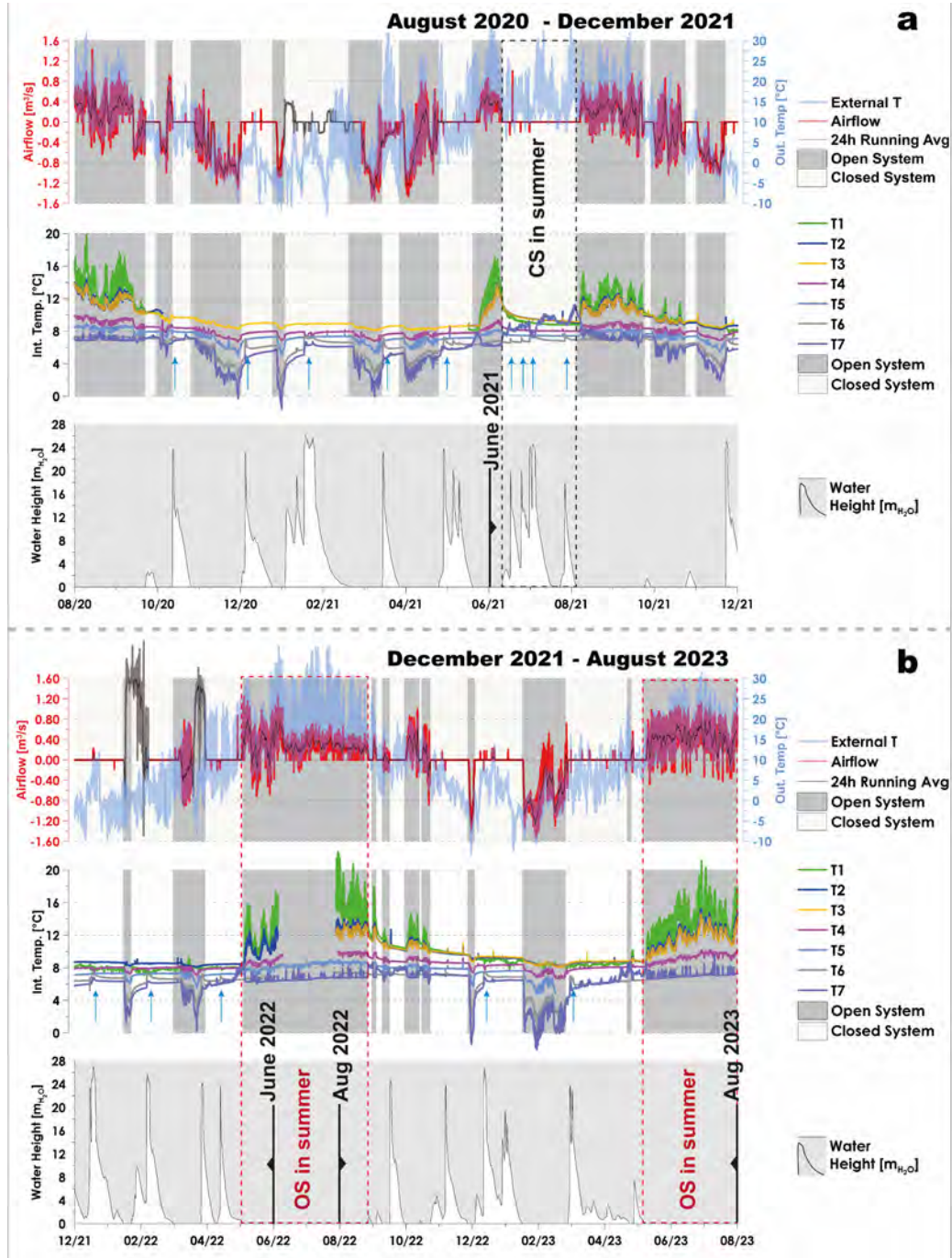


Figure 8.5 – Airflow, external and internal temperatures are compared. The internal temperatures are shown for each station. The water height is measured at the lake (WH station in Figure 8.1). **Open System** = active airflow, **Closed System** = aborted airflow. Graph a) shows the period from 13th August 2020 until half of December 2021, while graph b) shows the period from half of December 2021 until 13th August 2023. Airflow depicted with a grey line are problematic airflow measurements. See in the text. Blue arrows indicate temperature peaks at T6 and T7 due to high water stages (**CS** periods) The black bars delimit the periods taken into account for evaluations in the text (June 2021 – June 2022; August 2022 – August 2023).

8.3.1. Airflow

A parallel evolution between cave ventilation and external temperatures is confirmed by the field data (Figure 8.5.a and b). The airflow measurements (positive=downward regime; negative=upward regime) follow the general trend of the external temperature throughout the investigated period (Figure 8.5).

In Longeague Cave, ventilated phases (**OS**, Open System, dark grey periods in Figure 8.5) alternate with periods of no ventilation (**CS**, Closed System, 0 m³/s periods light grey in Figure 8.5), corresponding to high water levels in the lake indicating that the air path is shut off when lake water rises high enough to fill the entire lake cross-section.

During **OS** periods, the downward airflow is generally between $\sim 0.30 \pm 0.15$ and $\sim 0.80 \pm 0.40$ m³/s, while the upward flow is between $\sim -0.40 \pm 0.20$ and $\sim -1.50 \pm 0.75$ m³/s. Airflows lower than 0.25 m³/s (0.50 m/s or ~ 0.30 kg/s) are considered as null and mainly correspond to a **CS** (Figure 8.5).

The airflow reversals happen with an external temperature of $\sim 10 \pm 1^\circ\text{C}$, meaning that ventilation is downward when the external temperature is higher than this value, while it is upward when it is lower.

During two periods (27/01/2022 – 19/02/2022 and 01/04 – 19/04/2022, airflow depicted with a grey line in Figure 8.5), the airflow exceeded 1.5 m³/s with a downward regime and partially overlaps periods where the WH probe indicates **CS**. However, the temperatures inside and outside the cave were typical of an upward regime before the cave turned into **CS**. The same kind of signal but less intense, happened between January and March 2021, again during the transition from an upward flow to a **CS**. These events, which happened sporadically in winter, could be due to 1) free convection cells which may bring cold air inside the cave causing this contradictory signal, 2) an “intermittent” air leak (Pastore et al., 2024b) in the upper cave part (dashed arrow in Figure 8.1), 3) the influence of the external wind or 4) a failure of the monitoring system. Therefore, in the following, these periods will be treated as downward **OS** periods but without an estimated air flow rate when the water level measured at the WH station indicates **OS**. When the system turns into **CS**, possibly measured airflow is neglected in further calculations.

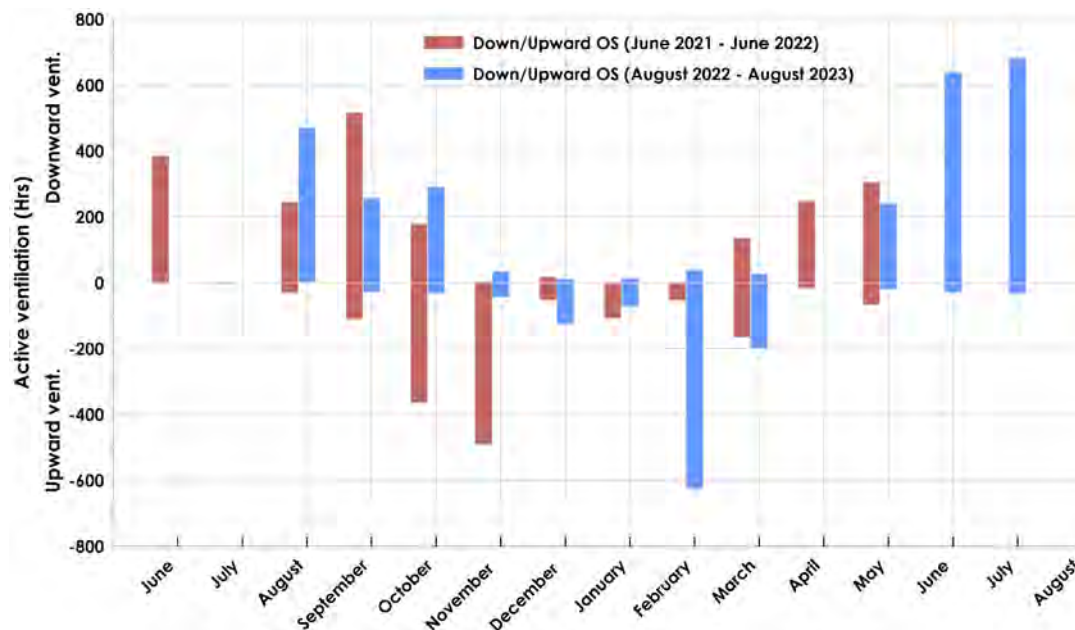


Figure 8.6 – Monthly distribution of cumulated hours of downward (positive) and upward (negative) Open System (**OS**) ventilation phases (Open System, **OS**) from June 2021 to July 2023.

Figure 8.6 shows the number of hours of **OS** grouped by month. We consider the years from 1st of June 2021 to 31st of May 2022 (June 2021/2022 in the following) and from 13th of August 2022 to 12th of August 2023 (August 2022/2023 in the following) for comparison, since they do not have lacunae in the data set if not the inconsistent airflow period stated above. Therefore, the two compared periods are shifted by about 2 months. The total duration of **OS** (both downward vent. and upward vent.) is very similar between June 2021/2022 and August 2022/2023, since in the first case $n_{2021-2022}=4303$ hours (~ 179 days) and in the second case $n_{2022-2023}=4788$ hours (~ 200 days).

Figure 8.6 displays, however, a different distribution of **OS** periods between the two years. While the ventilation was active only for 395 hours (~ 17 days) during the summer period from June to July 2021, in the period from June to July 2023, the **OS** cumulated time was 1387 hours (~ 58 days). A contrasted **OS** duration is also noticeable when comparing August 2021 and August 2022 (Figure 8.6). Conversely, during 2021, ventilation was mainly active from September to November, totalling 1670 hours. In September 2021 the ventilation was mainly downward, whereas, from October to November 2021 the system ventilation reversed upward (Figure 8.6). In 2023, the upward ventilation took place mainly in February and March, totalling 890 hours.

8.3.2. Air temperatures

Using the recorded temperature datasets, meaningful considerations can be made on cave temperature dynamics. As a general trend over the three monitored years (from August 2020 to August 2023, Figure 8.5.a and b), during **OS** phases, all temperatures follow the same evolution, with an increase in summer and a decrease in winter, mimicking the external atmospheric variations. While the internal temperatures show marked oscillations at stations near the upper entrance (T1, T2 and T3) in the downward regime, on the other hand, during the upward regime, the greatest oscillations are observed at T7, which is located near the lower entrance. In contrast, more internal temperatures (T4, T5 and T6) show a strong attenuation of the fluctuations with increasing distance from the cave entrance. The larger amplitudes of temperature oscillations happen with higher flow rates; conversely, they are completely flattened when the airflow is aborted, namely during **CS** periods.

As for the airflow and due to the lack of temperature data from June to August 2022 (Figure 8.5.b), we compare the two complete years, June 2021/2022 and August 2022/2023. It is interesting to take a look at these two periods, as the outside meteorological conditions were very different between them. From June to August 2021, 421 mm of rain fell, keeping the airpath mostly closed throughout that summer. This contrasts with the following year, when only 222 mm of rain fell (rainfall data source MeteoSwiss, Couvet station, 13 km from Longeague Cave) between June and August 2022, and 146 mm between June and August 2023, allowing the cave to be ventilated during those months. These two summers with contrasting ventilation phases (**OS** and **CS**) are highlighted in Figure 8.5 with dashed rectangles: red for **OS** and black for **CS**.

Consequently, the cave's average temperatures and ventilation may also vary between the two considered periods (June 2021/2022 and August 2022/2023), providing important information on the effects on the internal climate as external meteorological and hydrological conditions change. The presence of the lake that influences the ventilation status (**OS** or **CS**) could help characterise the variability of *Ventilated Cold* and *Warm Zones* (namely the so-called cold and warm anomalies) to external changes. In the following, we show that the cave's Annual Mean Temperatures (AMTs) have varied over the years, depending on whether the system was predominantly open or closed in winter

or summer, resulting in a different thermal shift of the ventilated cold and warm zones from the external AMT.

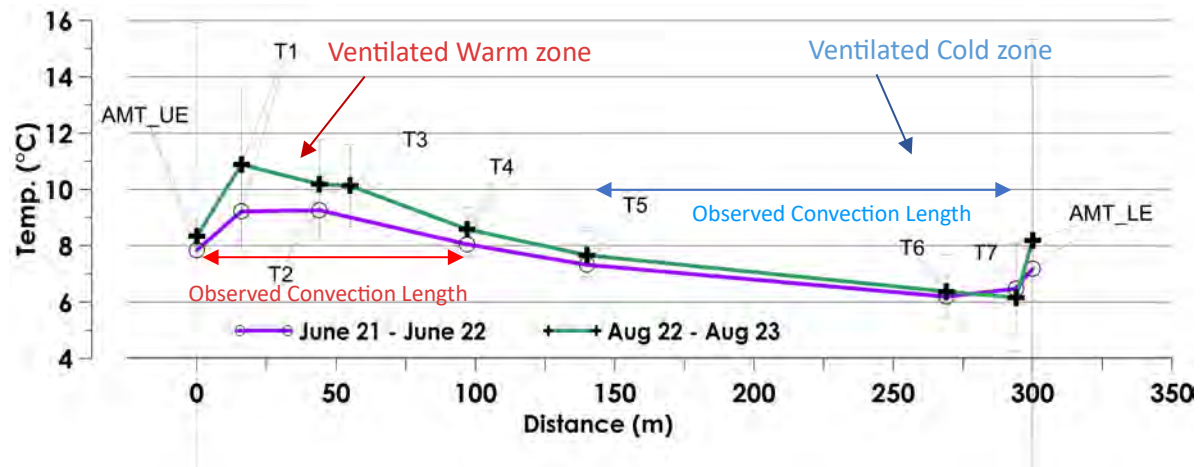


Figure 8.7 – Annual mean temperatures along Longeague Cave during the periods June 2021/June 2022 (purple curve) and August 2022/August 2023 (green curve), including periods of Open System (OS) and Closed System (CS).

	1	2	3	4
	Avg Temp. (°C±σ)	Downward OS Temp. (°C±σ)	Upward OS Temp. (°C±σ)	CS Temp. (°C±σ)
June 2021 – June 2022				
AMT_UE	7.92±7.87	11.24±7.89	4.40±4.83	7.13±7.95
T1	9.22±1.75	10.73±2.39	9.01±0.89	8.52±0.84
T2	9.22±1.06	10.03±1.49	8.93±0.67	8.97±0.60
T3	-	-	-	-
T4	8.05±0.40	8.33±0.53	7.71±0.42	8.01±0.17
T5	7.43±0.43	7.53±0.57	6.94±0.51	7.33±0.17
T6	6.19±0.73	6.36±0.22	5.90±0.29	6.70±0.25
T7	6.48±1.58	6.04±1.49	5.21±1.46	7.06±1.36
AMT_LE	7.29±7.27	10.50±7.08	3.68±3.44	6.54±7.53
August 2022 – August 2023				
AMT_UE	8.34±7.59	16.23±4.56	0.24±4.93	8.34±5.04
T1	10.89±2.74	13.89±2.44	9.08±1.33	9.51±1.29
T2	10.18±1.58	11.85±1.33	8.89±0.87	9.49±0.87
T3	10.13±1.46	11.63±0.84	8.81±0.84	9.56±0.84
T4	8.59±0.78	9.34±0.53	7.55±0.69	8.40±0.42
T5	7.66±0.87	8.43±0.43	6.22±0.84	7.82±0.28
T6	6.37±1.32	7.04±0.54	3.90±1.61	6.68±0.58
T7	6.16±1.91	6.85±0.54	2.71±2.69	6.75±0.85
AMT_LE	7.19±7.27	10.50±7.08	3.68±3.44	6.54±7.53

Table 8.3 – External Annual Mean Temperatures (AMTs) at lower and upper entrances are compared with cave AMTs for each station. In column 1, the averages are evaluated for the whole dataset. Column 2 and column 3 refer to averages calculated from temperature measured during Open System periods for downward and upward flow, respectively. Column 4 refers to Closed System average temperatures.

The averaged temperatures for each station over these periods are compared in Figure 8.7 and Table 8.3 (column 1). The external AMT (n=8785, number of averaged measurements for each year) close to the upper entrance (AMT_UE) in the period June 2021/2022 is 0.4°C lower than the AMT_UE of the

period August 2022/2023 (Table 8.3). Instead, AMT at the lower entrance (AMT_LE) for June 2021/2022 is $\sim 1^\circ\text{C}$ lower than August 2022/2023 ($n=8785$, both years).

The *Ventilated Warm* and *Cold Zones* are visible in Figure 8.7 and their characteristics are resumed in Table 8.4. The AMT ($n=8785$) at T1 (16 m from the upper entrance) represents the local maximum at 10.89°C and 9.22°C for August 2022-2023 and June 2021-2022, respectively. Whereas, the AMT ($n=8785$) at T7 (20 m from the lower entrance) is a local minimum at 6.48°C and 6.16°C , respectively for August 2022-2023 and June 2021-2022 (Table 8.3).

The differences between T1 and T7 with external AMT (ΔT_{warm} and ΔT_{cold} , respectively) are shown in Table 8.4. The *Ventilated Warm Zone* is more pronounced than the cold one ($\Delta T_{\text{warm}} > \Delta T_{\text{cold}}$), then, the external AMT is only approached in the cave at T4 (for the downward flow) and T5 (for the upward flow). These stations are located at 97 m and 140 m from the upper entrance and 197 m and 154 m from the lower entrance, respectively. At these internal stations, the external fluctuations are attenuated by $\sim 90\%$ ($\sigma_{T4/T5}$ vs $\sigma_{\text{AMT_UE/LE}}$, Table 8.3). Thus, we can state that their distances from cave entrances are the observed convection lengths (Figure 8.7 and Table 8.4). For both years, during the upward flow, the cave AMTs decrease with distance from the upper entrance, while during the upward ventilation, the mean cave temperatures increase from the lower one toward the inside. The absence of a plateau in the temperature profile (Figure 8.7) denotes also the absence of a deep karst zone in the middle part of Longeigue Cave (D_k in Figure 8.4), at least in its ventilated part (Figure 8.1).

	Ventilated Warm Zone	Ventilated Cold Zone
Distance from the entrance of the max./min of AMT	16 m (T1 from UE)	20 m (T7 from LE)
$\Delta T_{\text{warm/cold}}$ with external AMT (June 21-22)	1.30°C	-0.81°C
$\Delta T_{\text{warm/cold}}$ with external AMT (Aug. 22-23)	2.55°C	-1.03°C
Observed convection length	~ 100 m	~ 150 m

Table 8.4 – Characteristics of the Ventilated Warm and Cold Zones. ΔT_{warm} and ΔT_{cold} represent the differences between the maximum and minimum AMT measured in the cave (at T1 and T7, respectively, column 1 in Table 8.3) and the external AMT. The observed convection length is the distance at which $\text{AMT}_{\text{cave}} \approx \text{AMT}_{\text{ext}}$.

Moreover, internal AMTs differ between the considered years (Figure 8.7 and Table 8.3). The internal average temperatures observed at stations close to the upper entrance (T1, T2 and T4) are lower for June 2021/2022 by 1.67°C (T1), 0.92°C (T2) and 0.49°C (T4) than for August 2022/2023. These internal temperature differences between the two years have values comparable, if not larger, than those observed for the external AMT in the same periods.

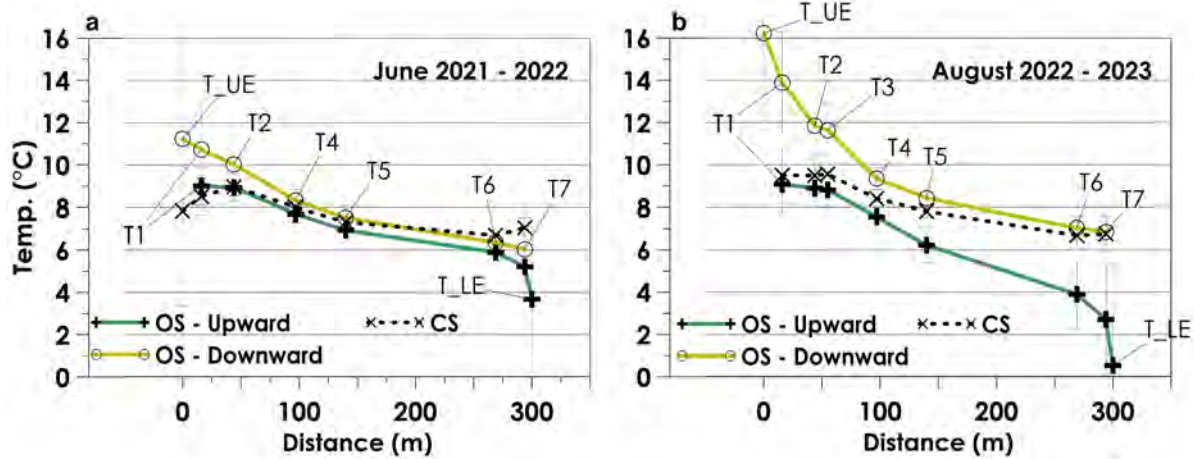


Figure 8.8 – Continuous lines represent the air average temperature for each station evaluated for Open System (OS) periods, and according to the flow direction (green and yellow colours). The dotted lines are for air temperature averages evaluated for Closed System (CS) periods.

In Figure 8.8.a/b, the AMT for each station in the cave has been calculated by distinguishing between **OS** (solid green lines for the upward flow and solid yellow lines for the downward flow) and **CS** (dashed lines). This distinction highlights how the yearly average temperatures of the cave for the periods June 2021/2022 and August 2022/2023 are affected by the status of the ventilation (**OS** and **CS**) and its direction (downward and upward). In fact, the external AMT in Figure 8.8 (T_{UE} and T_{LE}) represents the mean temperature of the air effectively entering the cave.

During **OS**, the data point count (n) for temperature averaging reflects the number of hours with active airflow (upwards or downwards). The average T_{UE} and T_{LE} evaluated only during periods when the system was open (**OS**) varied significantly between June 2021/2022 and August 2022/2023 (Figure 8.8.a/b, Table 8.3 columns 2 and 3). During **OS** and downward ventilation, outside the cave's upper entrance, the T_{UE} was 11.23°C ($n=2451$, $\sigma=7.89$) for June 2021/2022 and 16.23°C ($n=2711$, $\sigma=4.55$) for August 2022/2023, showing a difference of 5°C between these two years. During **OS** and upward ventilation, at the lower entrance, this difference was 3.14°C ($\sigma=5.88$) since T_{LE} was 3.68°C ($n=1219$, $\sigma=3.44$) and 0.54°C ($n=1412$, $\sigma=4.77$), for June 2021/2022 and August 2022/2023, respectively. These differences are reflected by temperature averages in the cave (Figure 8.8.a/b, Table 8.3 columns 2 and 3). Furthermore, Eq. 8.1 reads that the flow rate is proportional to $\sqrt{\Delta T}$ (ΔT between internal and external temperatures). As a result, if **OS** periods take place when the external air temperatures have a reduced contrast with internal cave temperatures, this could also result in a reduced airflow rate in the cave.

Conversely, during **CS** periods, air temperatures are in between those recorded during **OS** periods and no temperature fluctuations are observed (Figure 8.5). Moreover, it is interesting to note in Figure 8.8.a/b that for both years, the **CS** average temperatures at the stations near the upper entrance ($T1$ and $T2$) are close to those recorded during **OS** upward ventilation, and those recorded close to the lower entrance ($T6$ and $T7$) are close to those recorded during the downward ventilation.

The temperature peaks observed at stations $T6$ and $T7$ (blue arrows in Figure 8.5) during the **CS** periods relate to the lake level. During floods, $T6$ and $T7$ sensors, both located in the lower part of the cave, are submerged or very close to the water flow, causing the observed peaks (i.e., the sensors measured water temperature). During rainfall, water flows from the warmer upper part of the cave toward the

cave's colder zones near the lower entrance. The fact that the lower entrance has an average temperature lower than the external AMT, and that water temperature peaks during flood events are always positive, indicates that intruding water is always warmer than the surrounding rock in the lower cave regions (Sedaghatkish et al., 2025).

From June 2021 to about 15 August 2021 a sensible increase in T7 temperature appears during a **CS** period and without airflow (in Figure 8.9 as a zoom of Figure 8.5). This happened after a period of strong downward ventilation followed by intense rainfall that kept the lake closed, impeding air circulation. This condition could be explained by a free convection cell.

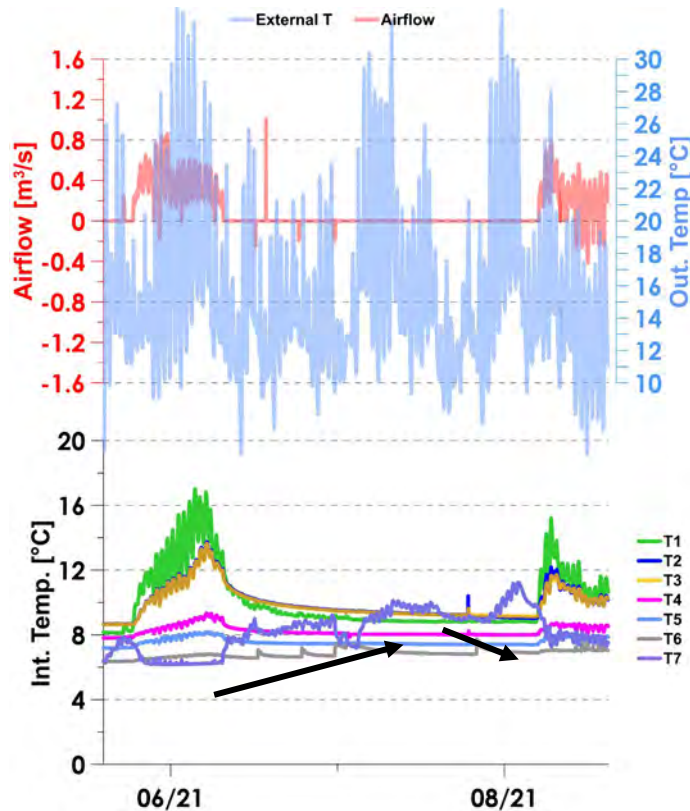


Figure 8.9 – Temperature increase at station T7 (lower entrance). The temperature increased (black arrows) during a **CS** phase (no airflow).

8.3.3. Aeraulic model

Three years of airflow and temperature data allow us to plot (Figure 8.10) the mass flow rate ($\dot{m}$, derived from the measured volumetric airflow times the air density $\rho = 1.25 \text{ kg/m}^3$) against DP (Eq. 8.2). Figure 8.10 shows that Longeigue's ventilation follows Eq. 8.1. The points lacking in the central part of the graph correspond to a mass flow rate lower than $\pm 0.3 \text{ kg/s}$ ($\sim 0.25 \text{ m}^3/\text{s}$), which corresponds to the measurements with large uncertainty when the cave is switching to **CS** periods (i.e. disconnecting from the outside temperature). The coefficients of the fitted curves in Figure 8.10 represent $1/\sqrt{R}$ of Eq. 8.1, then the aeraulic resistance R is estimated at $-26.14 \text{ kg}^{-1}\times\text{m}^{-1}$ during the upward flow and $63.90 \text{ kg}^{-1}\times\text{m}^{-1}$ during the downward flow.

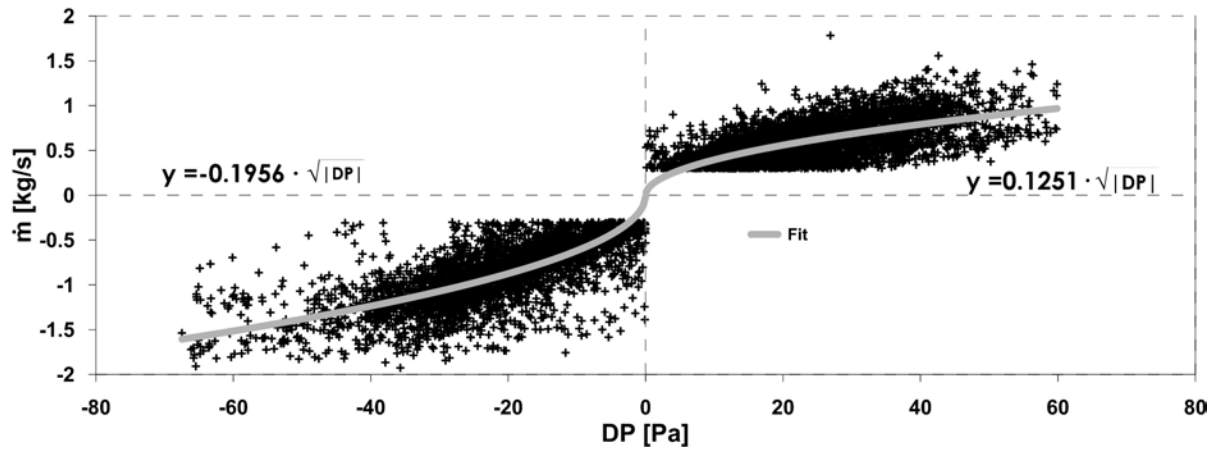


Figure 8.10 – Quadratic relation between the measured mass flow rate and the driving force DP for the period August 2020–2023, obtained after the cave discretization (Figure 8.3). Negative values refer to the upward flow, while positive values refer to the downward flow.

With the derived cave resistance R for both regimes (Figure 8.10), Eq. 8.1 is used to back-calculate the mass flowrate $\dot{m}$, which is then compared with the measured one (Figure 8.11). Using water height data from station WH, we determined periods of **OS** and **CS** to include lake fluctuations in the model.

The model correctly predicts daily fluctuations (boxes a and b in Figure 8.11), but it shows discrepancies with fitting minimum and maximum measured values. The model relative error is estimated as $[(\dot{m}_{mod} - \dot{m}_{meas})/\dot{m}_{meas}]$. Figure 8.12.a shows that most of the predicted points have an error of $\pm 30\%$, which increases at low $\dot{m}$ values (Figure 8.12.b).

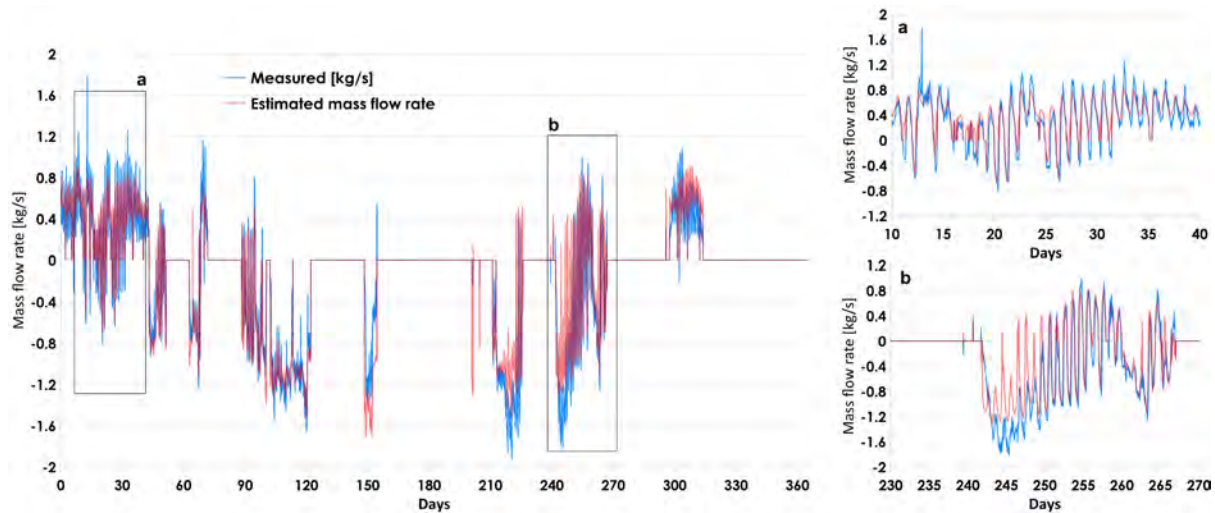


Figure 8.11 – Measured mass flow rate (in blue) against the modelled mass flow rate (in red). The a and b boxes on the right side are zooms on the periods in the black boxes on the left side of the figure.

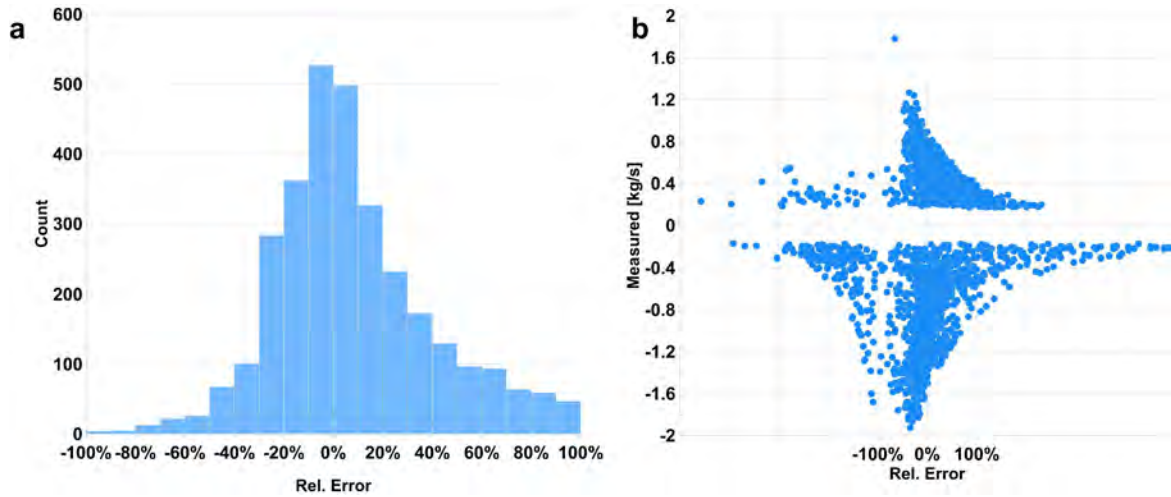


Figure 8.12 – The picture displays the model's relative error $[(\dot{m}_{mod} - \dot{m}_{meas})/\dot{m}_{meas}]$. a) the model relative error is generally lower than 30%; b) the relative error of the model against the measured mass flowrate $\dot{m}$. The highest error is related to low mass flow rates (i.e. transition between **OS** and **CS** phases).

8.3.4. Heat transfer

To evaluate the heat transfer between air and rock for three annual cycles (from August 2020 to August 2023), Eq. 8.4 was solved for the whole cave length (from outside to T7 for downward flows and from outside to T1 for upward flows). Figure 8.13 shows the result as a time series (dark green and orange lines) and as cumulated curves (purple and red lines), distinguishing between the energy released toward and removed from the rock. Estimation of heat exchanges is derived when ventilation is active (**OS**, both upward and downward regimes, dark grey areas in Figure 8.13). Positive ΔE_{rock} (purple and dark/light green lines in Figure 8.13) indicates that heat is released from the air to the rock during downward flows, therefore ΔE_{air} in Eq. 8.4 is negative. Instead, a negative ΔE_{rock} (orange/yellow and red lines in Figure 8.13) indicates heat removed from the rock by the air during upward flows (positive ΔE_{air} in Eq. 8.4). When ventilation is aborted (**CS**, light grey areas in Figure 8.13) $\Delta E_{air/rock} = 0$, no heat exchanges occur due to air convection. The black line in Figure 8.13 indicates the sum of the orange and green lines. Main uncertainties on the energy transfer derive from airflow measurements, for which the relative error is estimated at $\pm 50\%$ of the measured value. Light green and yellow lines, as well as error bars in Figure 8.13 display these uncertainties.

The amount of energy released over 3 years from the air towards the rock during all the downward regimes is estimated at $162 \pm 50\%$ GJ (red line in Figure 8.13), whereas the energy removed by the air from the rock during all the upward regimes is estimated at $-80 \pm 50\%$ GJ (red line in Figure 8.13).

The time series in Figure 8.13 shows that the higher intensity of the thermal exchanges (higher positive peaks) happened during downward regimes which took place mainly in summer. However, from November 2020 until May 2022, the energy exchange between air and rock happened mostly during upward regimes (negative values of the cumulated black curve). This observation agrees with the time distribution of ventilation phases (Figure 8.6). In fact, during 2021, the ventilation was observed mainly upward, since the cave was in the **CS** phase from June until August 2021. Therefore, between November 2020 and May 2022, most of the heat exchanges happened in the lower part of the cave. This is also confirmed by the less pronounced ventilated warm zone in the upper part of the cave in 2021 (Figure 8.7 and Figure 8.8).

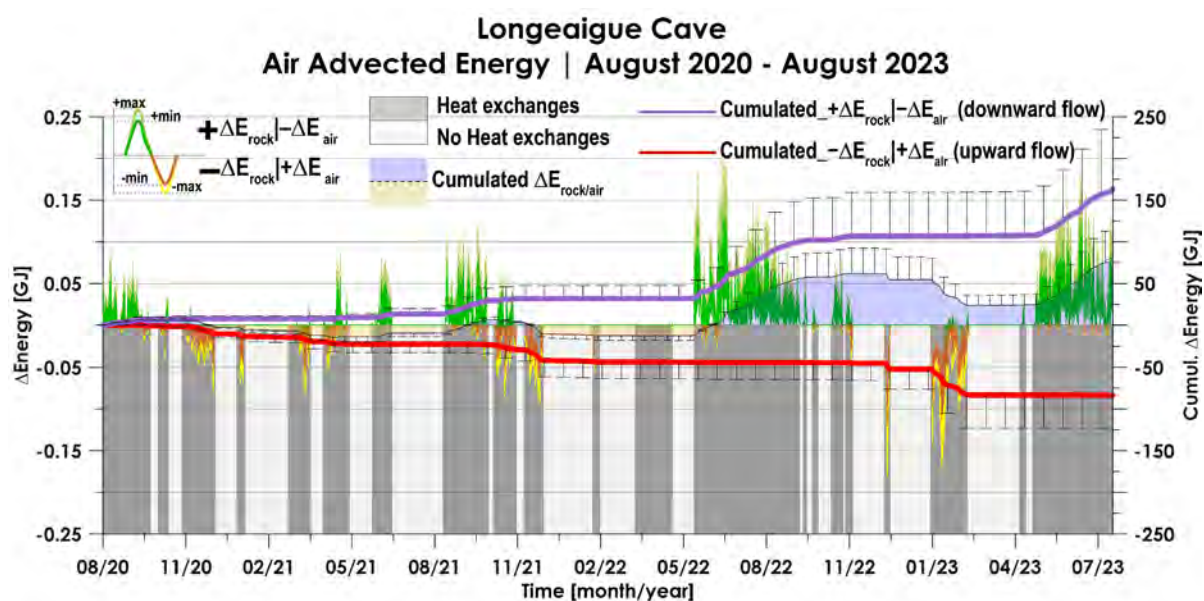


Figure 8.13 – Heat exchanged (ΔE) as time series and as cumulated curves evaluated from the cave data shown in Figure 8.3. The estimation of the heat transfer is derived for the whole cave length (from outside to T7 or T1 for downward and upward flows, respectively).

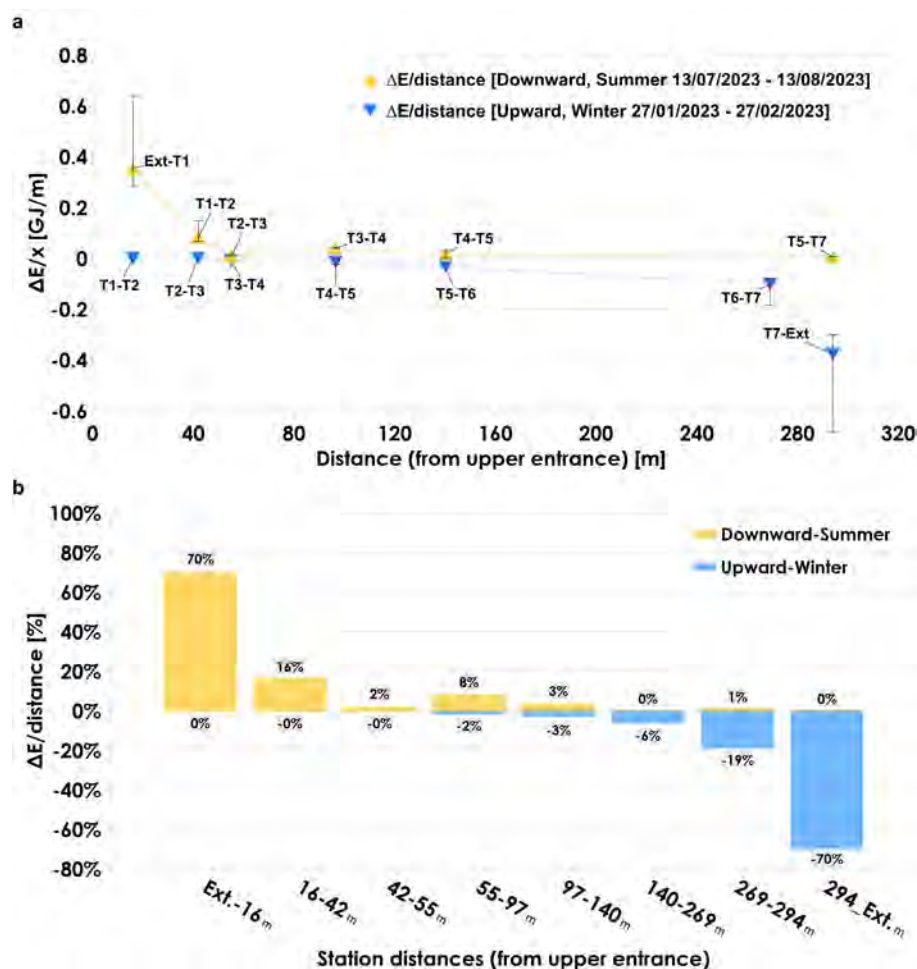


Figure 8.14 – Energy exchanged between the walls and air at each monitoring station during one month of winter and summer regimes.

The arrangement of monitoring stations inside the cave allows the estimation of heat exchanges along the conduits (Eq. 8.5). Figure 8.14.a illustrates the distribution of heat exchanges (GJ/m) between stations during one month of upward and one month of downward air circulation (**OS**). These results show that most of the energy ($\sim 90\%$, Figure 8.14.b) is exchanged between the air and the cave walls in the entrance regions, within the first 50 m of the cave. This distance is shorter than the observed convection length (between 100 and 150 m in Figure 8.7). Larger temperature fluctuations take place near cave entrances, while they are damped along the convection length. Fluctuation buffering along the conduit (Figure 8.5) relates to the energy exchange between air and rock. As a result, most of the heat transfer (90%) takes place within the first 50 meters of the convection length, where the temperature difference between air and rock is greatest, while the whole convection length is required for the complete damping of the fluctuations.

8.4. Discussion

Longeague Cave's airflow proved to be linked with external temperature fluctuations and confirmed that forced convection (chimney effect) drives the ventilation. The cave airflow is between 0.80 and -1.5 m³/s, which is an intermediate weak airflow, as many caves have shown higher values (e.g. Badino et al., 2018; Badino, 2010; Luetscher and Jeannin, 2004; Dueñas et al., 1999) as well as many others have shown weaker air drafts (e.g. 0.08 m³/s measured in Císařská Cave, Czech Republic; Faimon and Lang, 2013).

Internal temperature oscillations are synchronous with airflow fluctuation, which is in turn controlled by the external temperature variations. The direct link between these three parameters is noticeable in Figure 8.5. Airflow is downward when the temperature outside the cave is higher than $\sim 10 \pm 1^\circ\text{C}$, whereas it is upward when the external temperature is lower than this value. Owing to the presence of the perennial lake (Figure 8.1) whose level is controlled by external precipitations, the cave ventilation is quite complex and differently distributed over the years (Figure 8.6) according to the external weather conditions. The data in Figure 8.5 show that over yearly cycles, the cave alternates periods of Open System (**OS**), with upward or downward ventilation, and periods of Closed System (**CS**) when the passage of air is closed by water.

Looking at temperatures in the cave (Figure 8.5), we distinguish three different trends of temperature fluctuations according to the ventilation phase (**OS** or **CS**). During **OS** periods of downward ventilation (mainly during summer), the temperatures along the cave decrease with distance from the upper entrance. Conversely, during **OS** periods of upward ventilation (mainly in winter), the air temperatures increase with distance from the lower entrance. Whereas, in **CS** periods, the internal temperature oscillations are flattened and stabilised at a certain value. These periods represent time intervals during which there is no air exchanged by forced natural convection between the internal and external atmospheres. Since there is no advected heat, this temperature represents the local rock temperature, with which the air equilibrates by conduction.

The comparison of the cave AMTs between two different years (Figure 8.7) reveals differences. This suggests that the contrasted distribution of **OS** and **CS** over the years (Figure 8.6) is driven by lake fluctuations (and consequently by the presence or absence of forced air convection), and that along with the transited air masses and the temperature at which air enters the cave, likely plays a major role in determining cave AMTs.

The air volumes transited per year relate to $\sim 5.4 \times 10^6 \text{ m}^3$ (for both upward and downward) and $\sim 7.72 \times 10^6 \text{ m}^3$ ($\sim 4.7 \times 10^6 \text{ m}^3$ downward, $\sim 3.0 \times 10^6 \text{ m}^3$ upward), respectively, for the periods June 2021/2022 and August 2022/2023. Between June 2021/2022 and August 2022/2023, the total time of **OS** differs by only 21 days. Although this is not a long time, it accounts for 30% of the air volume transited in the period August 2022/2023. The volumetric airflow (in m^3/s) derives from the integration of a punctual airspeed through the cross-section and, in a natural setting such as a cave, the air velocity measured by the instrument can be influenced by its position in the cross-section (e.g. distance from the wall, presence of obstacles and bends) thus impacting airflow estimation given a non-uniform velocity distribution (Pastore et al., 2024a). The error on the volumetric estimation is, therefore, assessed at $\pm 50\%$. It is also highlighted that the volume estimation for June 2021/2022 lacks the air amount transited in the periods between 27/01/2022 – 14/02/2022 and 01/04 – 19/04/2022 since airflow data are untrustable (airflow grey line in Figure 8.5.b, cf. 8.3.1). The annual volumes of air transited in Longeague compare with the $13.6 \times 10^6 \text{ m}^3$ of air volume estimated by Luetscher et al. (2008) at Monlesi Cave, ventilated only during winter with an average airflow of $\sim 8 \text{ m}^3/\text{s}$.

In June 2020-2021 (Figure 8.7) cave AMTs were lower than in August 2022-2023, especially in the upper entrance zone ($\sim 100 \text{ m}$ from the entrance) between T1 and T4 (Figure 8.7 and Table 8.3). This is consistent with the lake closure during the summer of 2021 (Figure 8.6) which prevented warm air from entering the cave. In 2021 the ventilation was aborted (**CS**) for ~ 80 days between June and August (dashed black rectangle in Figure 8.5.a) when the ventilation was expected to be downward (summer). This correlates with 421 mm of rainfall concentrated in summer (over a total precipitation of 1171 mm), keeping the airpath closed. In comparison, in the year August 2022/2023, the total rainfall amount was 1092 mm, but only 146 mm fell between June and August 2023 and the cave was ventilated with warm air (dashed red rectangles in Figure 8.5.b). Therefore, the time of year when **CS** periods occur is crucial for determining the cave AMTs. This is also confirmed in Figure 8.8.a and Figure 8.8.b. These figures show the average air temperatures distinguishing between **OS** (upward and downward) and **CS** periods. With light green curves, it is displayed that the temperature of the external air (T_{UE}) entering the cave in June 2021/2022 during downward **OS** ventilation (i.e. warm air entering from the upper entrance) had an average temperature (11.24°C , $\sigma=7.89^\circ\text{C}$) $\sim 5^\circ\text{C}$ lower than during downward ventilation in August 2022/2023 (16.23°C , $\sigma=4.56$). This resulted in a lower mean temperature along the cave in the period June 2021/2022 (Figure 8.8.a) than during August 2022/2023 (Figure 8.8.b). This also relates to the lower air volumes that transited during the downward regimes in June 2021/2022 compared to August 2022/2023. Indeed, during June 2021/2022, the cave was primarily in a downward regime ($T_{\text{cave}} < T_{\text{ext}}$) from late August to September 2021 for ~ 32 days (763 hours), while during 2022/2023, the downward ventilation occurred between June and early August 2023 for ~ 63 days (1506 hours). In the first period, this resulted in the intake of $\sim 1.1 \times 10^6 \pm 50\% \text{ m}^3$ of air through the upper entrance, compared to $\sim 3.0 \times 10^6 \pm 50\% \text{ m}^3$ of air during the second period. Therefore, during the downward regime in August 2022/2023, more and warmer air passed through the cave, leading to higher AMTs compared to downward regimes that took place in June 2021/2022, when less and colder air transited.

In contrast, during upward **OS** (i.e. cold air entering from the lower entrance) the cave was colder in 2022/2023 than in 2021/2022 (Figure 8.8.a/b). Although it is not possible to quantify exactly the air volume transited for June 2021/2022 during the upward ventilation (because of the untrustable airflow measurements, airflow depicted with a grey line in Figure 8.5.b, cf. 8.3.1) it can be seen from Figure 8.6 that most of the upward ventilation occurred from the end of September until October (2021), alternating with downward ventilation, and then from November to the beginning of December (2021)

in only upward ventilation. The temperature of the external air entering the cave during upward **OS** for June 2021/2022 (T_{LE} in Figure 8.8.a) had an average temperature of 3.68°C ($\sigma=5.7$). Probably, as cave temperatures suggest, also during the periods of problematic airflow measurements (airflow grey line in Figure 8.5.b, cf. 8.3.1) the ventilation was upward. If we consider these periods, the T_{LE} reduces at 3.05°C . Thus, since in the period June 2021/2022, most of the upward ventilation occurred during autumn (October-December 2021, Figure 8.6) and at lower outside temperatures only for a relatively short period of January and part of March 2022 (Figure 8.6), the air drawn in the cave was not very cold. Instead, during August 2022/2023, the upward ventilation occurred mainly from February to March (Figure 8.6), therefore the entering air was much colder (-0.51°C , $\sigma=3.2$).

These results show that although the total duration and total transited air mass of the downward and upward ventilation regimes are comparable between June 2021/2022 and August 2022/2023, differing by only a few days, the internal AMT difference between the two years is related to the temperatures at which air enters the cave. This, in turn, is linked to the distribution of ventilation phases (**OS** and **CS**) throughout the year (Figure 8.6), which are driven mainly by external precipitations.

This transient system in Longeigue is comparable to other caves which could be opened and closed through different mechanisms. In show caves, the presence of doors allows to “open” and “close” the system. Doors are opened and closed to permit tourist transit and this could trigger modification of ventilation, temperature, relative humidity and CO_2 content (Li et al., 2024; Dredge et al., 2013; Spötl et al., 2005; De Freitas and Schmokal, 2003; Pflitsch and Piasecki, 2003) and therefore may impact the delicate microclimatic cave equilibrium (Mammola et al., 2019). Cycles of evapo-condensation and $p\text{CO}_2$ variations due to airflow changes may induce e.g. corrosion of carbonate speleothems. On the other side, with a general climate warming, underground glaciers and ice caves are threatened (Wind et al., 2022; Colucci and Guglielmin, 2019; Spötl et al., 2018). In the ice cave Eisriesenwelt (a show cave in Austria) ice formations are preserved by keeping the entrance door open only during wintertime to allow cold external air to enter (the cave is a lower entrance under a chimney effect). In summer, the cave's lower entrance is sealed to prevent downward and warm ventilation (from the inner parts of the system) that could melt the ice (Spötl et al., 2024; Obleitner and Spötl, 2011). Also, snow and ice accumulations at cave entrances and external glaciers (e.g. Luetscher et al., 2011) could represent transient obstructions interrupting cave ventilation. The Cenote Abyss, opening at 2940 m a.s.l. in Italian dolomites, is an example that is being studied by researchers to monitor the internal ice body volume (Santagata et al., 2017). The cave entrance is in a sinkhole and snow, ice and rock boulders have sealed the bottom of the doline allowing the formation of a perennial lake. In 1994 an abrupt emptying of the lake took place, unveiling the cave but also triggering an airflow. Since this event, the cave has been sealed by ice following an irregular pattern over the years, and the temperature therein could underline open and closed system cycles. Santagata et al. (2017) suggest a likely irreversible process of warming of the cave and melting of the ice therein due to the opening of the passage.

Other caves may have passages lying in the epiphreatic zone (i.e., parts of the karst that can be submerged during floods; e.g., Jeannin, 2001; Vuilleumier et al., 2021), as in the lower part of Longeigue. However, part of the air exchange may also occur through the soil and fractured rock surface (e.g., Gregorič et al., 2014). In both cases (epiphreatic passages and fractured surficial zones), airflow can stop during rainy periods. Additionally, natural geomorphological events, but also human activities such as mining, may block cave entrances permanently or at different times (e.g., the Obir cave, Spötl et al., 2005), preventing ventilation and altering the cave's internal climate.

Badino (2010), remarked that even small variations in conduit cross sections (caused e.g. collapses, flooding or anthropogenic modifications) can significantly affect ventilation and the cave's energy balance. Accordingly, cave closure can occur over different time scales and may be related to a variety of processes. As shown in Eq. 8.1, the flow rate depends on the driving pressure DP and cave aeraulic resistance R . Notably, R is related to the conduit's geometry, therefore any cross-sectional change modifies R , resulting in different airflow under the same pressure gradient. We explore the cave resistance in the following chapter.

8.4.1. Cave ventilation

The aeraulic model of a cave under a chimney effect (Figure 8.10) is meaningful to visualize the cave's aeraulic behaviour and eventually to estimate the flow rate from the driving pressure.

Unlike other aeraulic models (Gabrovšek, 2023; Garagnon et al., 2022; Faimon and Lang, 2013; Luetscher et al., 2008; Lismonde, 2002) which have assumed one single conduit or simple geometries, in this study, we discretised the cave passages into different segments (Figure 8.3) with measured geometrical parameters (length and slope, Table 8.2) and temperature to estimate the driving force DP (Eq. 8.2). We have also shown the validity of Eq. 8.1 (Figure 8.10) for Longeague Cave and we inferred R . Thus, the predicted mass flow rate ($\dot{m}$) was compared against the measured airflow (Figure 8.11). Our model has shown to be able to well reproduce the measured airflow fluctuations and to be in good agreement with measured values (Figure 8.11). The plausible relative error of the model is in the order of $\pm 30\%$ (Figure 8.12.a) or higher for very low mass flow rates (Figure 8.12.b). This error arises mainly from the following reasons: 1) from air temperature measurements that could be partially affected by rock temperature (Sedaghatkish et al., 2025) and sensor's size and shape (Lundström and Mattsson, 2020), 2) from the lake fluctuations which make the cave diameters change when the airflow is still active (transition from **OS** to **CS** and vice versa), 3) from the number of discretised conduits (Figure 8.3) composing DP (Eq. 8.2) and 4) from the significant uncertainty affecting the assessment of the real mass flow rate (derived from air velocity measurements).

The relation between the (mass) flow rate and the driving pressure (DP) shown in Figure 8.10 is not symmetrical since it shows different coefficients for the square root of DP , $(-)0.1956$ for the upward and 0.1251 for the downward regimes. This means that a given pressure DP corresponds to different flow rates depending on the direction of the ventilation. This characteristic is not rare and it has already been observed and highlighted in different caves by other authors (De Freitas et al., 1982; Faimon and Lang, 2013; Covington and Perne, 2015; Kašing and Lenart, 2020; Kukuljan et al., 2021b; Garagnon et al., 2022). The coefficients of the curves in Figure 8.10 represent $1/\sqrt{R}$ in Eq. 8.1, where R is the cave aeraulic resistance. Table 8.5 shows cave resistances approximated from aeraulic models found in the literature for other caves. Since we express R in $\text{kg}^{-1}\times\text{m}^{-1}$, cave resistances in Table 8.5 have been corrected in these unities for consistency with Longeague's results considering an air density of 1.25 kg/m^3 , 101.325 kPa as air pressure, as well as cross-sections and differences in altitude between entrances estimated from the literature.

It is noticeable that all the caves in Table 8.5 show an asymmetry between the upward and downward (when present) ventilation. All these cave resistances (Table 8.5) derive from field data fitting as in this work (Figure 8.10) except for that of Castelpard Cave for which R is inferred from Eq. 8.6.

Theoretically, the aeraulic resistance of a conduit composed of N segments can be inferred from the following relation (Atkinson et al., 1983; Lismonde, 2002):

$$R_T = \sum_{i=1}^N \left(\frac{8fL_i}{\pi^2 \rho_i D_i^5} \right)$$

Eq. 8.6

Where R_T is the resistance based on the theory, f is the Darcy friction factor (usually considered constant between different segments), while L_i , D_i and ρ_i are the length, the diameter and the density of air of the i^{th} conduit segment.

This relation is a good approximation for conduits whose friction factor is known or predictable, usually in the case of smooth industrial ducts. Atkinson et al. (1983) attempted to estimate f in Castelnard Cave from field measurements. They have shown that in a cave of 10 km with quite wide galleries, crawling passages with a total length of 500 m are responsible for the entire cave's aeraulic resistance. They have also highlighted that errors in the estimation of f are related to the simplified cave discretization (number of segments with different diameters). Indeed, Eq. 8.6 shows the strong dependency of the aeraulic resistance on cave diameters as they are raised to the power of 5. Thus, small variations in D_i (i.e. number of considered segments with varying D) have large consequences on the estimation of f and R_T . For example, the estimated R_T for Longeague Cave, using friction factors from the literature between 0.04 and 1 (Atkinson et al., 1983; Springer, 2004; Jeannin, 2001; Albright and Springer, 2022) and data from Table 8.2 for D_i and L_i , would be in the range of 0.75 and 19 kg⁻¹×m⁻¹. These values are much smaller than the results from field data (Table 8.5 and Figure 8.10). Thus, in our case, Eq. 8.6 seems to underestimate cave aeraulic resistance under natural and complex conditions. Furthermore, the smallest section of the cave corresponding to the narrow passage above T3 (area of 0.7 m², perimeter of 3.4 m and length of 3 m, Figure 8.1) is part of segment no. 3 in Table 8.2. The average diameters of the passages reported in Table 8.2 are derived from the survey of the cross-sections measured every 10 m along the passage, and this narrow passage may not fall within the average if it is outside this distance.

By evaluating the R_T for this small passage alone, Eq. 8.6 yields a resistance of 4.5 kg⁻¹×m⁻¹, assuming a friction factor (f) of 0.04. This value is significantly higher than the R_T estimated for the entire cave (0.75 kg⁻¹×m⁻¹, $f=0.04$) based on the average D_i values reported in Table 8.2. Since the total aeraulic resistance of a cave is the sum of the resistance of each cave segment, either a high value of friction factor or a single narrow passage may significantly increase the total value of R_T . Therefore, using Eq. 8.6 requires a thorough distribution of surveyed cross-sections along the cave, and small cave passages must be included in the average diameter estimation. Furthermore, although the relative roughness for standard pipes is considered always less than one (Moody, 1944), recently, Zhou et al. (2016) have proposed a new model to estimate the Darcy friction factor as a function of Reynold number. They have shown that in fractured contexts where the relative roughness can be very high (e.g. from 5 to 500), f can be much higher than 1.

Using field measurements allows us to estimate R , bypassing the assessment of f . In Table 8.5, while Longeague, Milandre, Postojna and Velká caves have R values comparable to each other, Císařská shows resistances two orders of magnitude higher, and Monlesi Cave shows an R value two orders of magnitude lower. The Císařská Cave presented by Faimon and Lang (2013) could display such high R values since the cave is closed at both sides with doors having 0.20x0.20 m windows that allow air

passage. The airflow of this cave was measured through these small windows (0.04 m² each) (Faimon and Lang, 2013). Since R is inversely proportional to the fifth power of the conduit diameters (Eq. 8.6) the major role in determining the resistance of Císařská might be played by these very small cross-sections compared to the rest of the cave. Therefore, the resistance derived from this cave might be an artefact. Moreover, such small passages could determine punctual head-losses (sudden expansion and sudden contraction; Maréchal, 1994; Lismonde, 2002) in the order of ~ 1 to ~ 2 Pa, comparable with the driving forces estimated by the authors for this cave. Concerning Monlesi Cave (Luetscher et al., 2008), the low R value ($\sim 3 \times 10^{-2}$ kg⁻¹m⁻¹) is consistent with the large cross-section (7.5 m²) of the entrance shaft where the airflow was measured. Such a small R value explains the large airflow measured in the shaft ($1 < \dot{m} < 17.5$ kg/s).

In Longeague Cave, the asymmetry of R could be in part related to the different friction due to geometries (e.g. bending, narrow passages/cave diameter) and roughness (f) in one direction or the other. We can state that a cave somehow acts as a Tesla Valve (Tesla, 1920), which allows fluids to flow preferentially in one direction rather than in the other, namely the valve has an anisotropic behaviour. Nguyen et al. (2021) highlight the electronic-hydraulic analogy between an electrical diode and a Tesla Valve. In fact, one of the main properties of a diode is to allow the current to freely flow in one direction while being cut off in the opposite direction. Therefore, for a Tesla Valve, this anisotropy is called diodicity (D_y in the following) and is defined as the ratio between $|R_{downward}/R_{upward}|$ (Nguyen et al., 2021). If this ratio is $\neq 1$, the valve has a diodic behaviour. This analogy could help in interpreting observations carried out from ventilated caves since for the electrical context the resistance is given by Ohm's law ($R = \Delta V/I$, where ΔV is the potential difference and I is the current) while for a fluid $R = \Delta P/Q$ (where Q is the flow rate). The electronic-hydraulic analogy was already used for ventilated caves for example by Badino (1995) and Lismonde (2002). Nguyen and coworkers (2021) highlight that the “valve anisotropic resistance derives from its asymmetric internal geometry interacting with high-inertia flows” (i.e. with high Re) which is what seems to happen for caves. In Longeague Cave, the anisotropic behaviour is identified by $D_y \simeq 2.7$, which can be compared to values reported from a few other caves (Table 8.5). These values seem to agree with the tests conducted by Nguyen et al. (2021), where a typical diodicity of ~ 2 is observed for flow conditions showing Reynold's number between 10^2 and 10^4 . Furthermore, for a Tesla Valve, plotting the flow rate against the ΔP for both directions will result in a very similar curve shape (e.g. Nguyen et al., 2021, pp. 8) to that shown in Figure 8.10, which allows us to state that a ventilated cave under a chimney effect acts as a Tesla Valve. Interestingly, in the same work, they pointed out that for very rough conduits, such as a Tesla conduit as well as cave conduits (e.g. Pastore et al., 2024b), friction factors (f) can be significantly greater (i.e. $\varepsilon/D \sim 1$ or even higher, as shown in Zhou et al. 2016, where ε is the effective roughness and D is the conduit diameter) than those documented in the literature at comparable Re (e.g. for caves, Albright and Springer, 2022; Springer, 2004; Jeannin, 2001; Atkinson et al., 1983). This fact reflects the extreme degree of roughness of natural channels. The complex geometry of natural conduits (with bends, bottlenecks, collapses, variable wall roughness, speleothems, and other structural factors) introduces additional disturbance to the flow, further increasing the friction factor. Indeed, Nguyen et al. (2021) conclude: “a better understanding of the hydraulics of very rough channels generally would benefit from further work that varies Re over a wide range”.

Moreover, the fluctuations of the perennial lake in Longeague might also play an important role in determining cave resistance. Figure 8.10 shows point scattering, particularly in winter, indicating a difference of about ± 1.5 kg/s for the same DP during upward ventilation (negative DP in Figure 8.10).

This suggests that for a given flow direction and the same driving pressure, the flow rate could vary according to the aperture of the lake cross-section, and in winter, the water level fluctuates much more frequently than in summer (Figure 8.5). Therefore, the cave resistance is much higher when the lake cross-sectional area becomes smaller due to the incoming water (until R reaches an “infinity” value, i.e. closed lake).

Cave	R (upward) [kg ⁻¹ m ⁻¹]	R (downward) [kg ⁻¹ m ⁻¹]	Mean Flow rate range [kg/s]	D_y	Reference
Milandre	~5	~11	-3.5 < $\dot{m}$ < 2.5	~2	Garagnon et al. (2022)
Monlesi	~3x10 ⁻²	-	~1 < $\dot{m}$ < 17.5 Only upward	-	Luetscher et al., (2008)
Císařská	~1.3x10 ³	~3.3x10 ³	-0.1 < $\dot{m}$ < 0.1	~2.5	Faimon and Lang (2013)
Castelguard	~816 – 2x10 ³ (from Eq. 8.6)	-	-2.4 < $\dot{m}$ < -1.7 Only upward	-	Atkinson (1983)
Postojna	~57 – ~95	~11 – ~19	~-1 < $\dot{m}$ < 1.9	~5.5 – 5	Kukuljan et al. (2021b)
Velká Ondrášova	~200			-	Kašing and Lenart (2020)
Longeaigne	26	64	-1.8 < $\dot{m}$ < 1	~2.7	This study

Table 8.5 – Approximated cave resistances R derived from aeraulic models and cave geometries found in the literature compared with the results of this study.

On the other hand, considering equal friction for both flow directions (i.e. $D_y=1$), the maximum DP (or the highest airflow) that can be reached in one season rather than the other could be also related to the non-linear relationship between air density and temperature. Indeed, different DP can be produced for the same $|\Delta T|$ (between external and internal temperature). Faimon and Lang (2013) demonstrated that this could sensibly affect the driving pressure. DP is generated primarily by the density differences ($\Delta\rho$) between the internal and external air column, therefore, given a certain $|\Delta T|$ between the cave and the exterior, e.g. $|\Delta T| = 20^\circ\text{C}$, the resulting DP will be greater if this difference results from $T_{\text{cave}}=7^\circ\text{C}$ and $T_{\text{ext}}=-13^\circ\text{C}$ ($\Delta\rho=0.096\text{ kg/m}^3$) than if it is $T_{\text{ext}}=27^\circ\text{C}$ ($\Delta\rho=0.084\text{ kg/m}^3$), favouring larger airflow during cold seasons (Faimon and Lang, 2013).

The shape of the Longeaigne Cave should also play a primary role in generating airflow in winter than in summer. Models suggest (Gabrovšek, 2023; Lismonde, 2002) that in L-shaped caves, the convection length plays a major role in determining the DP built along the conduit. Both authors have shown that caves with an L-shaped geometry (such as Longeaigne, which has a lower entrance connected to a long horizontal segment and an upper entrance linked to a vertical shaft) tend to exhibit stronger airflow in winter (dominant season) compared to summer (non-dominant season). During downward flow (non-dominant season), warm air heats the walls in the vertical part, allowing heat to be stored in the rock. During the upward and dominant flow, the cold air is gradually heated along the horizontal part, promoting thermal equilibrium between air and rock. Then, the warmer temperature of the rock in the vertical part keeps the rising air warmer and lighter because of the thermal inertia of the walls. On the other hand, in summer (non-dominant season), the warm air has to cool down as it enters the cave along the vertical section, but it does so less efficiently since air “falls” along the warmest part of the

cave; therefore, forced convection towards the lower entrance is counteracted by the thermal inertia of the wall. The increase in air pressure due to the increase in depth could also play a role in determining DP (adiabatic lapse rate; Badino, 2017).

At this stage, it is not possible to distinguish between the role played by these different factors in building the asymmetry of the aerodynamic cave behaviour (larger airflow in winter than in summer). However, for Longeigue Cave it seems that both R , which varies with lake fluctuations and probably has no equal friction f for both directions, and the L-shaped cave morphology have the largest impact in determining the highest volume flow rate in winter rather than in summer. Despite this, we have shown in the previous paragraph that the greatest air volume transited in summer and not in winter.

8.4.2. Cave convection length

The entrance zones of the cave show the presence of regions that are warmer and colder than the external AMT (Figure 8.7). These ventilated thermal zones are usually identified as warm and cold anomalies (Gabrovšek, 2023; Lismonde, 2002; Wigley and Brown, 1971) because their AMTs show a thermal shift from the external AMT. The *Ventilated Warm* and *Cold Zones* penetrate inside the cave for a certain distance. The warm zone is characterised by a higher AMT than the external atmosphere (Table 8.4) and the cave mean temperatures approach the external AMT with increasing distance from the upper entrance. The *Cold Zone* is at a lower AMT than the outside atmosphere (Table 8.4) and the average temperature increases with distance from the lower entrance. The distance at which the external temperature fluctuations are reduced by 90% of their amplitude is the convection length (also called penetration, thermal or relaxation length; Sedaghatkish et al., 2024). Field data from Longeigue Cave reveal convection lengths in the order of ~ 100 m for the *Warm Zone* and ~ 150 m for the *Cold Zone* (Table 8.4 and Figure 8.4).

The convection length (L_c) corresponds to both the extent of the thermal zone and the distance of propagation of the annual temperature fluctuations. L_c is expected from theoretical models presented by e.g. Wigley and Brown (1971), Lismonde (2002) and Gabrovšek, (2023). Although these authors have explained that the presence of these thermal zones is due to heat exchanges through air convection, their models lack a comparison with direct cave measurements to link them to real environments and provide plausible estimates of these distances. In recent work (Sedaghatkish et al., 2024), we presented a model that couples convective heat transfer with conductive heat transfer in the rock mass. This model aims to predict the order of magnitude of L_c along conduits ventilated by a chimney effect and was compared against field data. Sedaghatkish et al. (2024) model reads that in the presence of a turbulent airflow with $Re \geq 10^4$ (for Longeigue it is expected to be between $10^4 < Re < 10^5$; Sedaghatkish et al., 2025), at an annual time scale L_c is approximately proportional to the amplitude of the flowrate annual fluctuations divided by the square root of the cave radius (Eq. 8.7). Numerical simulations showed that in ventilated caves with a typical radius of 1 m and air flow rates ranging between 0.1 and 1.0 kg/s the convection length should be in the order of a few tens to a few hundreds of meters, but probably longer for extreme cases (Sedaghatkish et al., 2024).

The order of magnitude of the convection length can be assessed using the equation we proposed in Sedaghatkish et al. (2024):

$$L_c \approx 4.0 \times 10^2 \frac{\Delta \dot{m}^{1.06}}{r^{0.45}}$$

Eq. 8.7

where L_c is the convection length (m), $\Delta\dot{m}$ is the Annual Flowrate Fluctuation (AFF) of mass flow rate (kg/s) and r is the average cave radius (m). This formula was thus tested on Longeigue data.

To obtain the $\Delta\dot{m}$ for Longeigue Cave, a Fast Fourier Transform (FFT) was applied to the measured mass flow rate for June 2021/2022 and August 2022/2023 (Figure 8.15.a/b, respectively). The FFT analysis revealed the $\Delta\dot{m}_{2021-2022}=0.19$ kg/s (~ 0.24 m³/s, Figure 8.15.a) and $\Delta\dot{m}_{2022-2023}=0.43$ kg/s (~ 0.53 m³/s, Figure 8.15.b). In a pure chimney effect where $\dot{m} \propto \sqrt{DP}$ (Eq. 8.1) as observed in the caves studied e.g. by Sedaghatkish et al. (2024), AFF would be expected to dominate the airflow signal, followed by the Daily Flowrate Fluctuations (DFF) as it is for external temperatures (dashed lines in Figure 8.15).

In contrast, the FFT applied at the Longeigue mass flow rate (continuous line in Figure 8.15) highlighted that other frequencies with amplitudes comparable to or even higher than the AFF or DFF compose the signal. These modes with intermediate frequencies originate from the intermittent lake closures. These frequencies, between AFF and DFF in Figure 8.15, are related to the distribution of ventilation (**CS** and **OS**) throughout the year and not directly to external temperature fluctuations (Figure 17). This could lead to an underestimation of $\Delta\dot{m}$ (AFF in Figure 8.15). It is unclear to what extent these frequencies affect the estimation of L_c . Their impact can be significant, since the comparison between the measured mass flow rate and its annual fluctuation (Figure 8.16) shows a notable difference of approximately ~ 1 kg/s, indicating that the AFF component on the signal is significantly attenuated.

Moreover, the Annual Mean Flowrate (AMF, i.e. the average mass flow rate over the year at the arbitrary zero frequency in Figure 8.15) in August 2022-2023 was 0.05 kg/s, which indicates a seasonal asymmetry in the flow rate. Indeed, this result is consistent with the stronger airflow observed in winter (asymmetry in Figure 8.10). On the other side, in June 2021-2022, the AMF is close to 0 kg/s, suggesting that upward and downward flow are symmetric. However, this result can be influenced by the untrustable data removed from the airflow time series in the periods 27/01/2022 – 13/02/2022 and 30/03 – 13/04/2022.

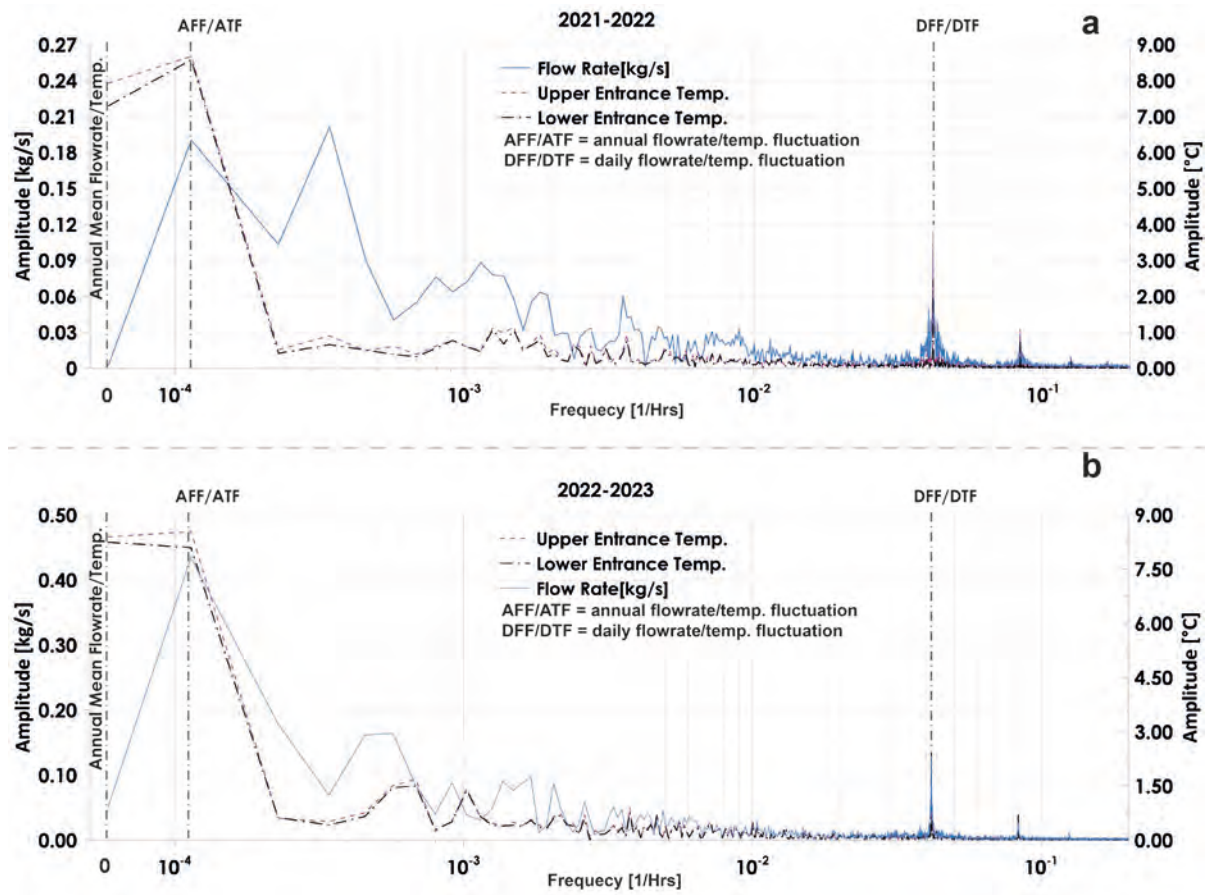


Figure 8.15 – FFT analysis applied to the measured mass flow rate in the cave compared with external temperatures measured at the upper and lower entrances, a) for the annual cycle June 2021-June 2022, while b) is for the annual cycle August 2022-August 2023. Due to the logarithmic scale the annual mean flowrate and AMT, which correspond to zero frequency, is located at an arbitrary origin.

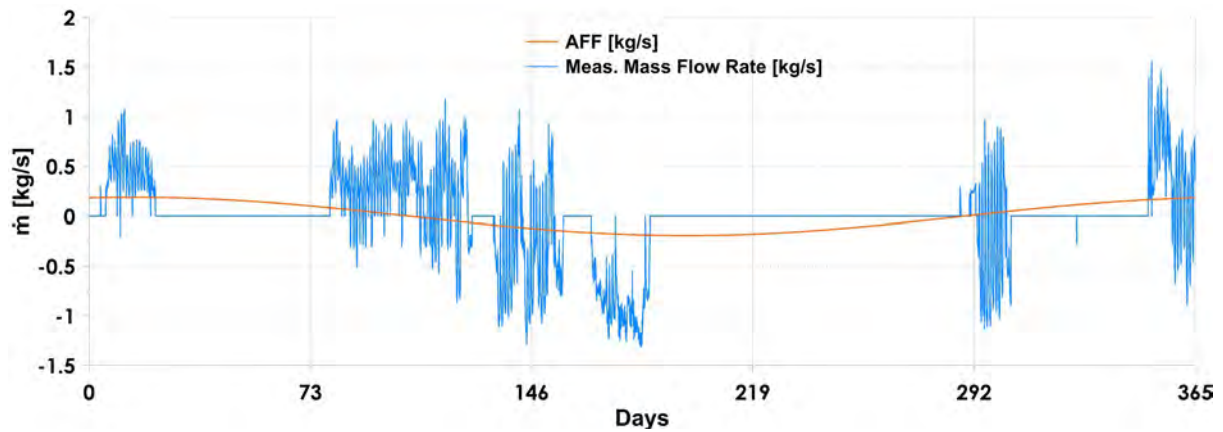


Figure 8.16 – Annual mass flow rate fluctuation derived for the FFT (June 2021-June 2022) compared with the measured mass flow rate.

Although Sedaghatkish et al.'s (2024) model only accounts for the AFF and does not consider the DFF or other frequencies and variations in the radius along the conduit (Table 8.2), the estimation of Longeauge L_c attempted from Eq. 8.7 (cave $r \approx 1.0$ m, Table 8.2) indicates an order of magnitude of $\sim 10^2$ m (Figure 8.7), consistent with the observed convection lengths (Table 8.4). These results are also

comparable with L_c values given by Sedaghatkish et al. (2024) for other caves (Table 8.6). However, a major difference between Longeague and the caves described in Sedaghatkish et al. (2024) is that in the latter case, no thermal interference between entrances is assumed due to the presence of deep karst. The deep karst zone is defined as the zone of the karst where temperatures are stable and do not depend on seasonal variations in the outside atmosphere (Badino, 2005; Luetscher and Jeannin, 2004) and this does not happen in Longeague Cave. Indeed, the absence of a plateau in the cave AMT profile (Figure 8.7) and that the observed convection length is longer than/close to half of the cave length, indicate that Longeague does not show a deep karst zone (D_k in Figure 8.4). The relatively short length of the cave (~ 300 m) and consequently the proximity between the entrances impede the presence of a deep karst zone since the warm and cold L_c affect each other. As a result, the cave lies within the heterothermic zone for its entire length since the cave temperature varies during the year according to the ventilation direction. On the other hand, the lake fluctuations, which determine the distribution of **OS** and **CS** periods over a year (Figure 8.6), affect the difference between outside and cave AMT along the convection length as shown in Figure 8.7, Figure 8.8 and Table 8.4.

		Longeague Cave (this work)	Baulmes mine (Sedaghatkish et al., 2024)	Sieben Hengste Cave (Sedaghatkish et al., 2024)	Bärenhöhle Cave (Sedaghatkish et al., 2024)
Average radius		1 m	1.5 m	0.5 m	0.5 m
$\Delta \dot{m}$		[0.2-0.4] kg/s	4 kg/s	0.1 kg/s	0.6 kg/s
L_c	Field data	100/150 m (warm/cold)	>360 m	[10 m, 30 m]	>120 m
	Theory	$\sim 10^2$ m	$\sim 2 \times 10^3$ m	> 50 m	$\sim 4 \times 10^2$ m

Table 8.6 – Comparison between estimated and observed convection lengths of Longeague Cave and other caves presented in Sedaghatkish et al. (2024).

As the convection length is longer than/close to half of the cave and as the lake induces a disequilibrium between winter and summer conditions we can hypothesize that the temperature in the middle of the cave may differ from the temperature of the massif surrounding the cave. It could be warmer if the lake is open longer in summer and colder if it is open longer in winter. As part of the answer, we can look at the temperature of the cave passages located beside the ventilated one (T8 station in Table 8.1). In Figure 8.17 we show temperature measurements from the non-ventilated part of the cave compared to station T4 in the ventilated part (which is located about halfway along the cave). The non-ventilated parts of the cave are dry and human-accessible during standard **OS** periods, but they are flooded during wet **CS** periods. In the image below (Figure 8.17), one can see that the station out of the ventilated conduit shows a stable temperature when the ventilation is active (**OS**), while it is influenced by water temperature during **CS** periods. In fact, although T4 in Figure 8.5 shows damped fluctuations compared to stations closer to the cave entrances, these fluctuations are clearly visible when this station is compared to the non-ventilated conduit.

When the ventilation is downward (summer) or the system is closed, the ventilated conduit is warmer than the non-ventilated part, as well as it could also be slightly colder during the downward regime. This is a partial answer as we do not know whether at a larger time scale than one year the influence of the ventilated conduit can reach the deeper part of the system, which is under the control of

conduction (between air and wall) during **OS** periods and under the control of water advection during **CS** periods.

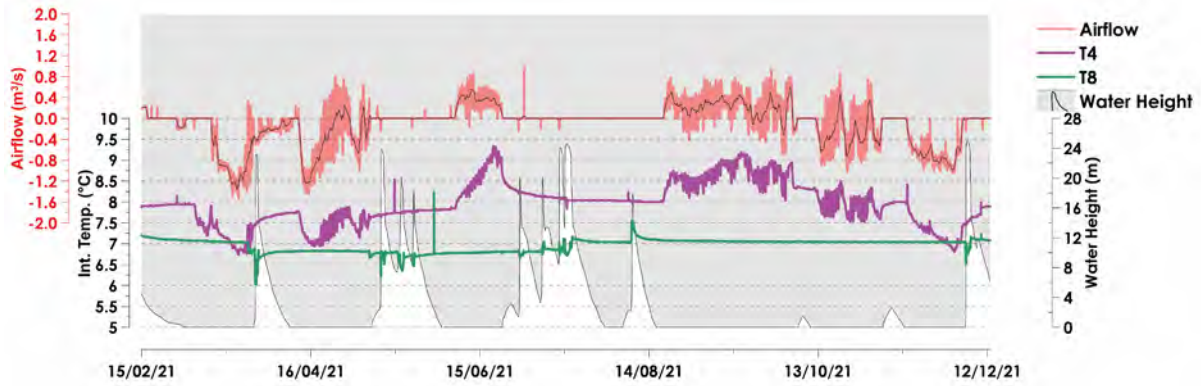


Figure 8.17 – Comparison between temperature measurement at stations T4 (ventilated conduit) and T8 (non-ventilated conduit). Lake fluctuation and airflow are also given.

8.4.3. Quantification of heat advected by air

Temperature buffering along cave conduits indicates that the airflow releases or gains heat to/from the rock along the conduit.

The instantaneous heat exchanges are displayed in Figure 8.13. The amount of heat exchanged depends mainly on the air mass flow rate, which in turn depends on the outside temperature. In Longeague Cave, it may be strongly reduced when the lake is about to close or completely stopped when it is closed. Figure 8.14 shows that most of the heat advected by air is exchanged within the first ~50 m from the inlets. This is in agreement with the observed *Ventilated Warm* and *Cold Zones* shown in Figure 8.7.

We estimate that the energy exchanged during the downward flow is $\Delta E_{rock} \approx 162 \pm 50\%$ GJ/3 years, whilst that exchanged during upward flow is $\Delta E_{rock} \approx -80 \pm 50\%$ GJ/3 years, which results in ~ 82 GJ/3 years as net energy exchanged between rock and air. This paper brings an order of magnitude of the amount of heat which can be exchanged between a cave and the surrounding massif. However, a closer look also implies considering the energy brought to the cave by conduction through the rock. A limitation of our evaluation is that we cannot consider it as an energy balance since we do not know the conductive heat flux through the rock toward the conduit. In a new work (Sedaghatkish et al., 2025), we have shown models that strongly suggest that the heat fluxes from the wall can sensibly affect the temperature in the conduit and that the initial condition of external ground temperature is a key parameter to be thoroughly assessed. Therefore, to close the cave energy balance (within the convection lengths L_c in Figure 5), the term related to the heat through the rock and the associated heat transfer coefficient has to be considered. To infer them, extensive rock surface temperature measurements are necessary since part of the energy can be lost toward or gained from the diffusive zone and the radial direction around the conduit (beyond the convection width, W_c in Figure 8.4). Modeling can be an alternative to estimate the heat transfer coefficient (Sedaghatkish et al., 2025). Therefore our assumption that $|\Delta E_{air}| = |\Delta E_{rock}|$ (i.e. only convective heat exchange) could be partially wrong since there could be a conductive heat flux from the cave walls and ground toward the

conduit. Since we did not directly measure the temperature gradients within the rock we cannot assess if this is really the case and how much this affects the calculation of energy exchanges.

In the same paper (Sedaghatkish et al., 2025) we also explained that the vicinity of the sensor to the conduit walls and the shape of the Hobo U22 can influence the air measurements. Sensors close to the wall can measure a temperature in between the air and rock temperature, and this is particularly true for the sensors deployed close to the cave entrance (T1 for the downward flow and T7 for the upward flow). In addition, the shape of the sensor, its size, and the material of the casing can affect the temperature measured by the sensor (Lundström and Mattsson, 2020). The temperature measured by a sensor depends on the competition between the air convection and the radiative flux from the walls (which is proportional to the sensor characteristics and emissivity coefficient) (Sedaghatkish et al., 2025; Lundström and Mattsson, 2020). Since Hobo U22 sensors are deployed close to the wall and have a large radius (3 cm) as well as a high emissivity due to their polypropylene-made case, we wanted to estimate the error brought by this issue. We added at some monitoring stations (*cf.* §8.2.2) Hobo U23 sensors with a wired small sensor (0.5 cm) equipped with a metallic cap (lower emissivity than the U22) in order to measure far away from the conduit walls. By comparing measurements from U22 (plastic case) and U23 (wired metallic sensor) we estimated an error on air temperature measurements of $\pm 20\%$ of the measured value for stations in the first 30 m from entrances (T1 and T7) and $\pm 5\%$ of the m.v. for all the others monitoring stations. These errors are included in the estimation of heat exchanges shown in Figure 8.14.

Our calculation has also the limitation that the heat term related to water flow is not assessed. During floods, the average cave water discharge is estimated between 1.5 and 2 m³/s, which in terms of mass flow rate means three orders of magnitude higher than air. The energy transported by water dissipates through the massif. However, water flows massively only during **CS** periods in the lower part of the cave (Figure 8.1) and relates to lake levels in the time series (Figure 8.5). As stated previously, none of the stations is directly affected by the water temperature when the cave overflows, except for T6, which is submerged when the water table rises, and station T7, where water can approach very close to the sensor in high water conditions (blue arrows during **CS** in Figure 8.5 showing temperature peaks related to water flows). The phreatic part of the system is fed by water infiltrating through diffuse recharge in the catchment area (Jeannin, 2018) allowing for a low water velocity in the upper part of the massif. This results in negligible heat advected by water compared to conduction through the rock along the upper part of the massif (Luhmann et al., 2011). Field data shown in Figure 8.5 indicate that during **CS** periods the climate of dry cave parts is controlled by heat conduction between air and walls resulting in flat temperatures along the cave, while the effect of water flow is only perceptible in the lower part of the conduit (T6 and T7 in Figure 8.5 and T8 in Figure 8.17). Whereas, during **OS** periods cave climate is controlled by convection in the conduit coupled with heat conduction through the hosting rock. The competition between air and water in heat transfer in a cave remains still open since water would play a major role in the case of concentrated recharge (e.g. stream sinking in a *ponor*). In this case, the water would flow along the whole conduit at a higher speed then advecting more heat, and influencing also the climate in the upper part of a system (Luo et al., 2023; Luhmann et al., 2012; Covington et al., 2011; Luhmann et al., 2011).

However, the computed energy due to advected air is in agreement with the order of magnitude quantified by Luetscher et al. (2008) at Monlesi Cave for air-advected heat (~ 75 GJ removed from the system). These results suggest that quantifying an order of magnitude is possible despite uncertainties remaining on the individual components of the energy estimation.

8.4.4. Free convection cells and unexplained airflow behaviour

With the aid of a thermal camera (Flir E40bx infrared camera, accuracy $\pm 2\%$ of reading), at the lower entrance and during the downward ventilation in summer (17/06/2022; ext. temp. $\sim 29^\circ\text{C}$), we observed temperature differences between the ceiling and floor of the lower entrance chamber (Figure 8.18). This temperature contrast can relate to air stratification that can likely be instigated by a local convection cell established in the lower entrance chamber (Figure 8.18.a). Infrared (IR) images (Figure 8.18.b/c/d/e) suggest that during summer, the large size of the lower entrance chamber (i.e. a large portal measuring $\sim 5 \times 5$ m with a rising ceiling towards station T7; Figure 8.1) favours warm air from the outside to enter the cave, flowing close to the chamber ceiling (ceiling warmer than the floor, Figure 8.18.b/d/e). The entering air cools down inside the cave as it reaches close to station T7 (approximately 20 m from the entrance, with homogeneous wall temperatures as shown in Figure 8.18.c), mixes with cold air from the inner cave parts, and moves outward due to the chimney effect (Figure 8.18.d/e). Moreover, the pothole present at the lower entrance (~ 5 m deep) can work as a cold air trap (Figure 8.18.e), blocking part of the cold air descending along the shaft (Figure 8.18.d). During summer and **OS** periods, T7 temperature (Figure 8.5) shows small peaks that can relate to this local convection cell at the lower entrance.

A convection cell can also justify the temperature increase at T7 (Figure 8.9) during a **CS** period in the summer of 2021 (between June and August). In this period, the lake was closed due to the persistent precipitation, as we have shown and discussed previously. In Figure 8.9 (a focus of Figure 8.5 for the period June to 15 August 2021), it can be observed that only at station T7 (close to the lower entrance), temperature increases sensibly without any measured airflow (**CS**). This was likely induced by natural convection between the cold air from the cave's lower part and the warm external atmosphere (rainy summer). Given the shape of the lower entrance of the cave and since the lake was closed (**CS**, no chimney effect), warm air seems to have reached station T7. Eventually, the void above station T7 may act as a warm air trap and keep warm air during the winter season. A similar situation was documented in the nearby Grotte de la Cascade (Jeannin, 2021). Throughout the monitored period at Longeague (2020 – 2023), the instance shown in Figure 8.9, i.e. constant increase in air temperature during a **CS** period, is the only one we observed during **CS** periods, as the lake is usually open in summer, and then the chimney ventilation is active.

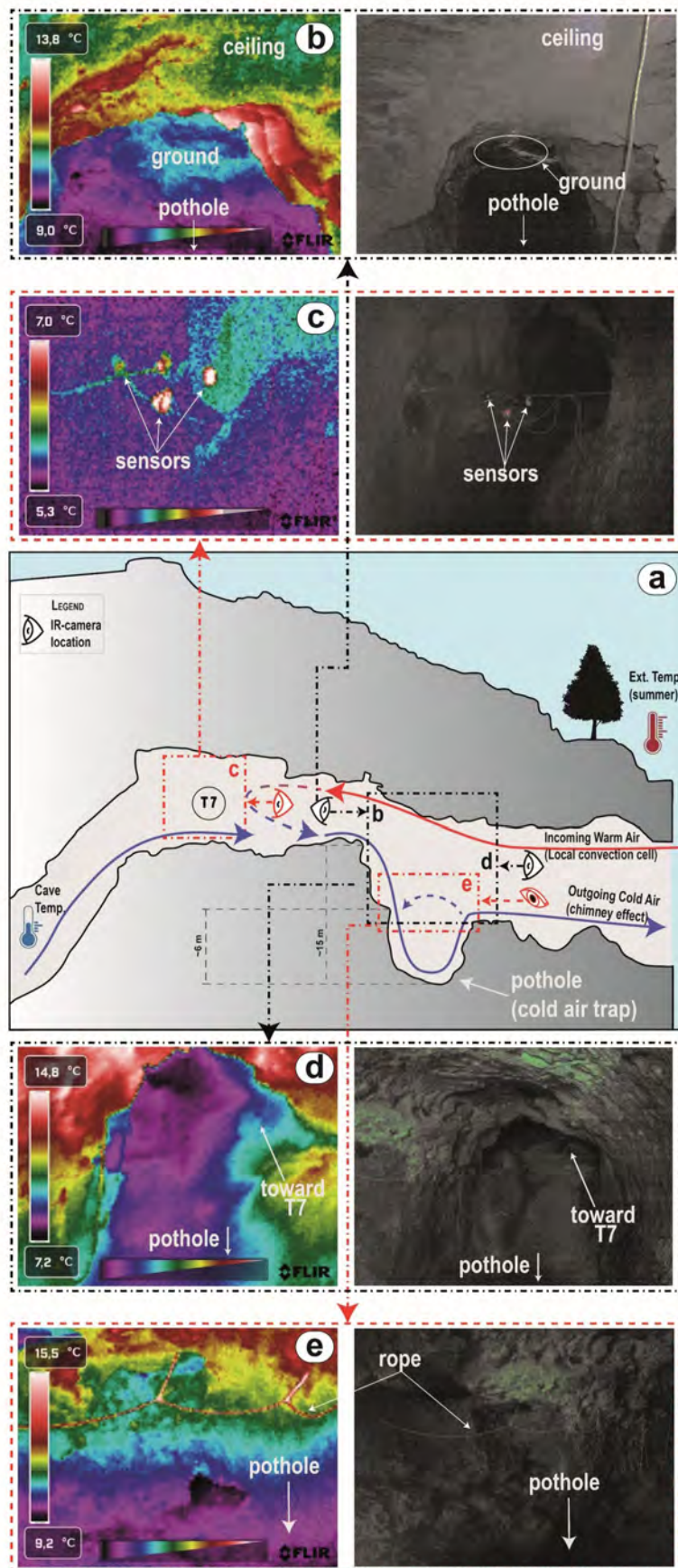


Figure 8.18 – A) Schematic representation of a local convection cell at the cave's lower entrance, possibly forming during downward airflow in summer (chimney effect). B) IR image looking toward the shaft that descends to the lower entrance chamber. C) IR image framing station T7. D) IR image from the lower entrance chamber toward the shaft leading to station T7. E) IR image taken in the lower entrance chamber and showing the wall above the pothole acting as a cold air trap. Walls around station T7 (c) show quite homogeneous wall temperature (5.5 to 6.0°C; contact with the sensors caused higher IR temperature readings), while figures b), d) and e) highlight the stratification of wall temperature, from the ceiling (warm) to the ground (cold). Cold air from the inner cave parts flows outward (chimney effect) close to the ground, while warm air from the outside (local convection cell) flows inward on the ceiling. In c), a noteworthy feature is the cold air "cascade" that descends along the shaft and becomes trapped in the pothole (purple colour in D). Eye icons in a) show the approximate IR-camera location; figures on the right sides of b), c), d) and e) show the cave passages.

A similar phenomenon might have occurred during winter at the upper entrance of Longeague Cave. A localised and sporadic free convection cell at this entrance could have triggered an unexpected response from the airflow meter at Anemo station, as shown by the airflow depicted by instances with a grey line in Figure 8.5.a/b. This behaviour remains unexplained by the chimney effect model. The most evident event occurred between 27/01/2022 and 13/04/2022 (inconsistent airflow in Figure 8.19), when downward ventilation (positive values) exceeded 1.5 m³/s twice, overlapping between **OS** and **CS** phases. Something similar was recorded between January and March 2021 but with a smaller amplitude (airflow depicted with a grey line in Figure 8.5.a). For a short period (from January to June 2022), a second SFM flow meter was deployed at station T3 (displayed by a green line in Figure 8.19). The airflow at T3 station was clearly upward during **OS** phases, and negligible throughout the **CS** periods. This contrasts with the ventilation assessed from Anemo station, although between 21 and 31 March 2022, the two datasets seem to match. Due to these discrepancies, the airflow data from the Anemo station in the period shown in Figure 8.19 are considered inconsistent with a chimney effect. A comparable occurrence was observed in D7.1 Cave (Sedaghatkish et al., 2024). In this cave, during a period when an upward flow was expected, the measured one was subject to a flow reversal. D7.1 cave lies in a topographic low, a condition that could lead to local cold air trapping during winter and probable snow accumulation near the cave's upper entrance. This instance may also trigger local free convection cells in the entrance zones.

Notably, these unexplained events occurred in winter when the outside temperature was lower than that of the cave (Figure 8.19), inducing upward ventilation during **OS** periods. Initially, both stations (Anemo and T3) recorded an upward airflow. However, for an unknown reason, station Anemo abruptly shifted to a strong downward airflow, while station T3 continued to detect an upward flow. When the lake closes, flow at T3 steps down to zero, but it remains high at Anemo station. Temperature measurements at the Anemo and T1 stations are consistent with the ventilation measured at T3. The inconsistent airflow measured at Anemo could have various reasons, i.e.:

- 1) The most likely explanation is a temporary failure of the airflow monitoring system at the Anemo station.
- 2) Another potential explanation is the influence of external wind-driven drafts, but a comparison with meteorological data did not establish a clear correlation with these events.
- 3) Free convection may have brought cold air along the floor of the cave passage, compensated by the exit of warm air flowing along the ceiling of the cave. In this case, the flowmeter device could have measured only the inflow component. This hypothesis does not fit well with the position of the anemo, which is closer to the ceiling than to the floor, nor to the measured temperature at the Anemo station.
- 4) An alternative hypothesis involves an “intermittent” air leak (dashed arrow in Figure 8.1) located between Anemo and T1 stations, leading to a local air circulation between the upper cave entrance and this “third entrance” (Pastore et al., 2024b). Since these inconsistent airflow events at Anemo were observed three times in winter (airflow depicted with a grey line in Figure 8.5 and highlighted as inconsistent in Figure 8.19), it could be speculated that this supplementary cave inlet remains closed for most of the winter, possibly due to e.g. water, snow or ice. However, the mechanism by which it occasionally opens in winter remains unclear. Again, the temperature fluctuations at Anemo and T1 seem more plausibly explained by an outflowing of cave air at Anemo station rather than by the inflow of cold outside air.

Hypothesis 1) seems the most probable.

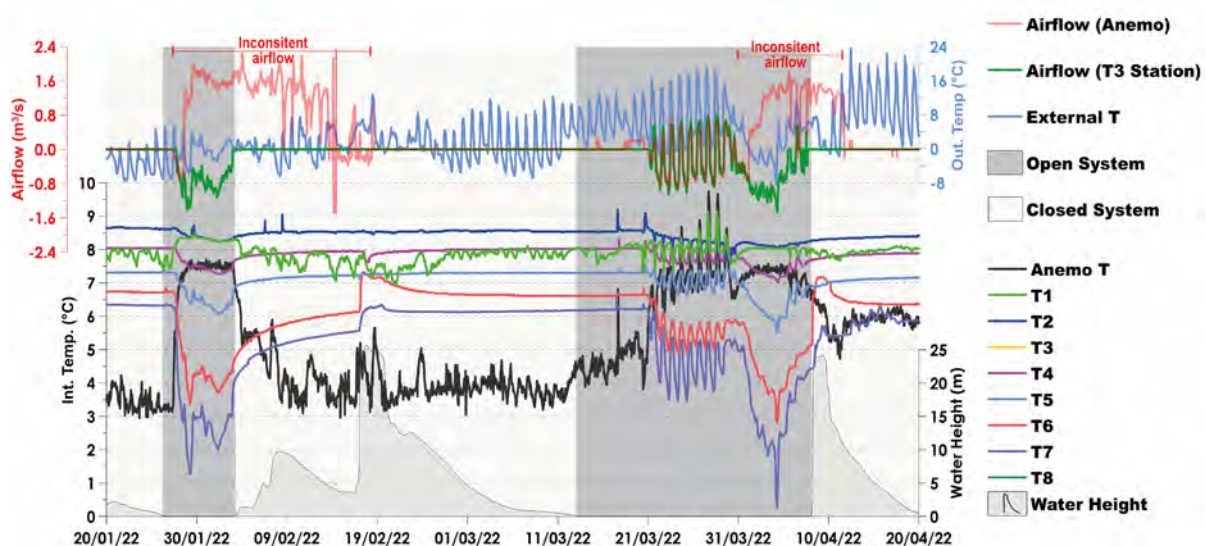


Figure 8.19 – Airflow and temperature time series in the period 27/01/2022 and 13/04/2022. The comparison between stations in the cave shows temperatures typical of an upward flow during OS periods and stable temperatures typical of CS periods. This contrasts with the measured airflow, typical of a downward regime.

8.5. Conclusions

The relationship between external and cave climates is crucial for a better understanding of speleogenetic processes (Badino, 1995; Fairchild and Baker, 2012), for the preservation and conservation of cave environments and dwelling life (Addesso et al., 2022; Medina et al., 2023) as well as for speleothems-based paleoenvironmental reconstruction (Fairchild et al., 2006). After Luetscher et al., 2008, for the first time, we have quantified the heat exchanges taking place along a ventilated cave (Figure 8.13 and Figure 8.14). In such a cave, so-called *Ventilated Warm* and *Cold Zones* are observed (Figure 8.7) corresponding to the heterothermic zone (or convection length) related to both upper and lower entrances, respectively. Identifying these two zones along ventilated caves allows for a better characterization of heat exchanges.

In this work, we demonstrate that ventilation exerts a primary influence on the temperature of a ventilated cave and the hosting rock. Along the heterothermic zone (or convective zone) external fluctuations are transferred by convection inside the cave much faster than by considering only conduction from the surface (Sedaghatkish et al., 2024). Cave ventilation significantly increases the depth of thermal fluctuation and decreases the time of propagation (Sedaghatkish et al., 2025). Field data have shown that little changes in ventilation regime distribution over the years (Figure 8.6) lead to significant changes over a yearly time scale on the AMT of the *Ventilated Warm* and *Cold Zones* in the cave (Figure 8.7 and Figure 8.8). The length of the heterothermic zone in caves corresponds to the convection length as defined by Sedaghatkish et al. (2024). This distance depends mainly on the cave aerodynamic resistance, which in turn depends much on the size of the ventilated conduits (Sedaghatkish et al., 2024; Gabrovšek, 2023). It is usually in the order of ~50 m for metric conduits (Table 8.6 and Eq. 8.7) but can reach kilometres for very big, smooth (typically tunnel) and strongly ventilated conduits. It is reduced to the diffusion length (5-10 meters) for small conduits. We have shown that the deep karst zone (homothermic) may not be present if the ventilated cave is shorter than or close to the convection length. This is the case in Longeigue Cave where the amplitude of the annual temperature variations in the middle of the cave is in the order of 1 to 2°C (Table 8.4).

The assessment of heat exchanged between the air and the rock along the cave shows that ~90% of the heat transfer between air and rock takes place along the convection length (or heterothermic zone). In ventilated caves, air advection is more efficient in transporting heat through karst systems than pure conduction through the rock, although the host rock plays a major role in dulling the external fluctuations with distance along the conduit (Badino, 1995; Luetscher et al., 2008). However, for small caves (centimetric) as well as for very shallow and poorly ventilated caves the conduction of heat from the ground surface may be larger than the heat transferred by the airflow (Qaddah et al., 2023; Guerrier et al., 2019; Quindos et al., 1987; Villar et al., 1984). Beyond the length of the heterothermic zone, Luetscher and Jeannin (2004) suggested that the cave temperature in the homothermic zone (deep karst) is also controlled by cave ventilation. These authors demonstrated that in many caves the temperature gradient along the homothermic zone is close to that of moist air (0.5°C/100 m). However, the numerical models we presented in Sedaghatkish et al. (2025) suggest that, while the gradient is controlled by the thermal gradient of moist air, the mean cave temperature itself could be regulated by the rock temperature, which in turn depends on the surface rock temperature, while cave ventilation significantly increases the depth reached by daily and annual thermal oscillations. In contrast, in very deep vadose zones (>500 m), due to the lower connection with the outer surface and the hierarchical structure of karst networks that will be concentrated in fewer conduits at greater depths, air circulation driven by forced convection will be progressively reduced (Luetscher and Jeannin, 2004). This results in less effect of air on the climate of very deep caves, which may instead be dominated by water circulation (Luetscher and Jeannin, 2004).

We have also demonstrated that cave monitoring is fundamental for assessing cave aeraulic resistance (R) and Driving Pressure (DP) for reconstructing a cave aeraulic model. For paleoclimatic studies, this could potentially allow estimation of the order of magnitude of the airflow rate in the past, as long as the cave resistance did not change during the time due to natural processes (Santagata et al., 2017; Luetscher et al., 2011) or due to anthropic management (see e.g. Spötl et al., 2005). It also should be considered that cave resistance is strongly dependent on cross-section diameters, and any (anthropic) modification of this geometrical parameter can have a large impact on the cave ventilation and climate (Badino, 2010) and, therefore, also on the habitats therein. This is a concern that should be taken under consideration when a cave is going to be modified (by widening a passage or closing it by a door) e.g. to permit tourist transit, especially in show caves.

In conclusion, the thermal perturbations induced by any change in the airflow will have consequences on the extension of the convection length and width (L_c and W_c), i.e. of the *Ventilated Warm and Cold Zones* observed at entrances of ventilated caves. This suggests that at larger time scales (hundreds to thousands of years) the influence of convection length (L_c) and (W_c) may reach the deep karst zone, which by definition is not directly affected by external (daily and annual) fluctuations.

Our results add another important piece of information for developing karst studies focusing on the impact of ongoing climate changes but also for paleoenvironmental studies.

Acknowledgements

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Open Research

The dataset of the cave temperatures and air flow plotted in Figure 8.5, Figure 8.10 and Figure 8.11 are available at the link <https://doi.org/10.5281/zenodo.18166879> cited as Pastore et al. (2026).

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Chapter 9. Thermal modeling of caves ventilated by chimney effect (published paper as second author)

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Author statement: My work for this paper involved: Data curation, Methodology and Investigation in the field.

Abstract

Cave ventilation significantly increases the depth of natural thermal oscillations and decreases the time of propagation compared to heat conduction in the rock mass. This makes it necessary to develop and test thermal models for the prediction of temperature fields in ventilated karst massifs. Here, we develop a thermal model of a single conduit ventilated by chimney effect. The model is based on the diffusion equation in the rock mass coupled to the conservation of energy and water vapor mass in the airflow. The effect of the latent heat of evaporation and condensation is considered. In parallel, the main conduit of a ventilated cave has been equipped with a flowmeter and several temperature sensors. The model is tested against field data collected during a complete year. The relevance of the model assumptions (geometry simplification, initial and boundary conditions, use of transfer coefficients to couple the air and the conduit wall) is thoroughly analyzed. The model correctly predicts the temperature fluctuations at daily and yearly scale, but underestimates the mean annual temperatures inside the cave. A biased assessment of the ground temperature seems to explain this discrepancy. The effect of condensation and evaporation on the cave climate turns out to be low on cave temperature, but significant on air humidity with consequences for ecology or paleoclimatology. This study is a first step towards the elaboration and validation of models providing a quantitative assessment of caves' thermal response at any location and time scale.

9.1. Introduction

Extensive cave networks in karst massifs are traversed by airflows that considerably enhance heat and mass transfer between the bedrock and the external environment (Lismonde, 2002). The impact of airflow must be considered in a wide range of areas including paleoclimatology and ecological issues. Indeed, isotope fractionation that occurs during calcite precipitation is temperature dependent. Airflow also controls evaporation and condensation of water, the transport of carbon dioxide, and their consequences on precipitation and dissolution rates of calcite (Kukuljan et al., 2021a; Kowalczyk and Froelich, 2010). A good understanding of heat and mass transfer associated with airflow is thus requested for paleoclimate reconstructions (Cuthbert et al., 2014; Rau et al., 2015). It is also relevant for the study of subterranean fauna and flora (Culver and Pipan, 2019). Indeed, caves are home to many temperature sensitive species, including bats, beetles, or thysanurans (Braack, 1989). These subterranean ecosystems and their biodiversity are particularly vulnerable to climate change (Castaño-Sánchez et al., 2020). For instance, Rizzo et al. (2015) observed that the survival rate of Pyrenean beetles over 7 days was 100% at 6-20°C, and collapsed to 0% at 23-25°C. This example illustrates the need for reliable thermal model for predicting the consequences of global warming on the temperature field in karst massifs.

The fluctuations of the external pressure can play a role in cave ventilation (Gomell et al., 2021). Large caves with a single narrow entrance are ventilated by barometric effect (Borsato et al., 2024), and wind pressure may be the dominant driving force in caves with multiple entrances at similar heights (Kukuljan et al., 2021b). In caves with multiple entrances at different heights, the most significant cause of ventilation is the chimney effect resulting from the density contrast between the air inside and outside the cave (Lismonde, 2002). In temperate climates, this density contrast is mainly due to temperature (Gabrovšek, 2023). Because of the high thermal inertia of the massif, the airflow direction is upward during most of the winter, when the temperature inside the cave is warmer than the atmospheric temperature. Conversely, the air flows downward during most of the summer. A noticeable consequence of this seasonal inversion of the airflow is the thermal anomalies observed in the entrance regions (Lismonde, 2002). In lower entrance areas, the annual mean temperature is shifted to a lower value than outside because this part of the cave receives cold air at the atmospheric temperature in winter and air cooled down after circulating in the massif in summer. The same mechanism produces a warm anomaly in upper entrance areas.

Airflow modifies the rock temperature in the cave, which in turn changes the air temperature profile inside the cave. Therefore, both the air and the rock mass must be considered in a model aiming at predicting the climate of a ventilated cave. Lismonde (2002) and Gabrovšek (2023) developed such models based on the 1D energy balance in the airflow and the heat conduction equation in the rock mass. Both equations were coupled by the Newton's law of cooling, which states that the local heat flux through the cave wall is proportional to the difference between the wall temperature and the mixing temperature of the air. The coefficient of proportionality, the so-called heat transfer coefficient, depends on the conduit geometry, the fluid properties, the wall roughness and the fluid velocity. It was assessed using standard correlations for fully developed forced convection in pipes (see for instance Bergman et al., 2017). This approach allowed Lismonde (2002) and Gabrovšek (2023) to reproduce qualitatively in numerical simulations some field observations as for instance the seasonal hysteresis of the flow rate or the thermal anomaly in the entrance areas.

Recently, we used similar models to investigate the propagation of thermal perturbations by the airflow inside a karst massif in a quantitative way (Sedaghatkish et al., 2024). We considered the simple geometry of a straight horizontal conduit of constant circular cross-section. Numerical simulations showed that turbulent airflow, very common in ventilated caves, significantly modifies the rock temperature. Numerical simulations supported the partition of the cave into three regions as first proposed by Cropley (1965): (a) a short diffusive region, where heat mainly propagates from the atmosphere by conduction in the rock mass; (b) a convective region where heat is mainly transported by the air flow; (c) a deep karst region characterized by quasi-constant temperatures throughout the year. The length of the diffusive region is of the order of a few meters, whatever the values of the air flowrate and conduit diameter. The length of the convective region, defined as the convection length, corresponds to both the extent of the thermal anomaly and the distance of propagation of the annual temperature fluctuations. Numerical simulations showed that the convection length can commonly reach a few tens to a few hundreds of meters. It is approximately proportional to the amplitude of the flowrate annual fluctuations divided by the square root of the cave radius (Sedaghatkish et al., 2024). The orders of magnitude predicted by the model were consistent with field data from a mine tunnel and two caves.

In the present paper, we want to go beyond orders of magnitude by carrying out detailed and systematic comparisons between numerical simulations and field data. Indeed, numerical simulation of ventilated cave climate is still a challenge. To our knowledge, direct and quantitative comparisons between models and field data are still lacking. The present article provides such a comparison, with

the aim of testing the validity of the physical assumptions included in the model. We focused on the following issues.

- A cave is a succession of bends, conduit contractions or enlargements, and obstacles of any kind. The first of the problems posed by this complex geometry is the level of accuracy required for its description. More specifically, is it necessary to consider a full 3D problem, or can we define an equivalent 2D geometry to save computational resources?
- The irregular geometry of caves also has important consequences on the assessment of the heat transfer coefficient. Correlations commonly used in thermal models are valid for straight rough pipes of uniform cross-section (Bergman et al., 2017). Geometrical singularities are expected to increase the turbulence level, which in turn can significantly enhance the heat transfer coefficient. Some corrections exist to take into account the effect of the well-defined singularities commonly found in heat exchangers (Rohsenow et al., 1998). However, they can hardly be used for the tortuous geometries typically encountered in caves.
- Numerical simulations are based on mathematical models that require initial and boundary conditions. Ideally, the temperature field at any point of the massif should be known at the initial time, which is never the case. Another issue is the boundary condition at the external surface (i.e., the interface between the massif and the atmosphere). The simplest choice is to assume that the external surface temperature is equal to the atmospheric temperature (Gabrovšek, 2023; Sedaghatkish et al., 2024). However, the earth surface does not only interact with the surrounding air. Its temperature results from an energy balance including convection with the air, solar irradiation (short wavelength), radiative exchange with the surrounding environment and the sky (long wavelength), latent heat due to evaporation or condensation, and conduction inside the rock (Salmon et al., 2023). This raises the question of the most relevant boundary condition at the external surface.
- Condensation and evaporation of water are commonly observed on the walls of ventilated caves (De Freitas and Schmokal, 2005; Fernandez-Cortes et al., 2013; Gázquez et al., 2022). Since the latent heat of evaporation or condensation induces cooling or warming, its contribution to cave climate must be evaluated.

With the aim of addressing these issues, flowrate and temperatures in the ventilated cave of Longeague (Switzerland) have been recorded during a complete year. In parallel, we developed a thermal model of the cave, for which results have been tested against the field data collected in the cave.

9.2. Field data

9.2.1. Study site

The “Baume de Longeague” displayed in Figure 9.1 is located in the Swiss Jura Mountains (Val-de-Travers; Upper entrance: 46°52′22″ N 6°31′09″ E, 917 m a.s.l.; Lower entrance: 46°52′18″ N 6°31′09″ E, 820 m a.s.l.). The difference of elevation between both entrances is 97 m. This induces an intense buoyancy-driven airflow inside the conduit that connects both entrances. A secondary conduit (blue arrow on Figure 9.1), likely connecting the cave to the atmosphere through a network of fractures, intersects the main conduit at approximately 16 m from the upper entrance. The possibility of a leak through this secondary conduit cannot be excluded.

The total length of the main conduit of Longeague Cave is 311 m. Its shape and size are highly variable. The cross-sectional area A and perimeter P are displayed in Figure 9.2 as a function of x , the distance from the upper entrance following the conduit axis. Both values were obtained using a laser-scanner

with a 10^{-2} m resolution every 10 m along the cave. Figure 9.3 displays the angle $\theta(x)$ between the upward vertical direction and the velocity vector of the air when flowing from the upper to the lower entrance.

The lower entrance is a temporary spring acting as an overflow of the “Raies” hydrogeological system, whose catchment area is approximately 60 to 80 km². The maximum water discharge-rate at Longeague entrance does probably reach approximately 5 m³/s (Jeannin, 2018). A perennial lake has formed in the lower part of the main conduit (see Figure 9.1). During flood events, the rise of the water level is sufficient to close the conduit, interrupting the airflow. This is an important feature of Longeague Cave that must be considered in the thermal model.

Since August 2020, several stations monitor the cave temperature as well as the airflow along the main conduit of the cave (red symbols in Figure 9.1).

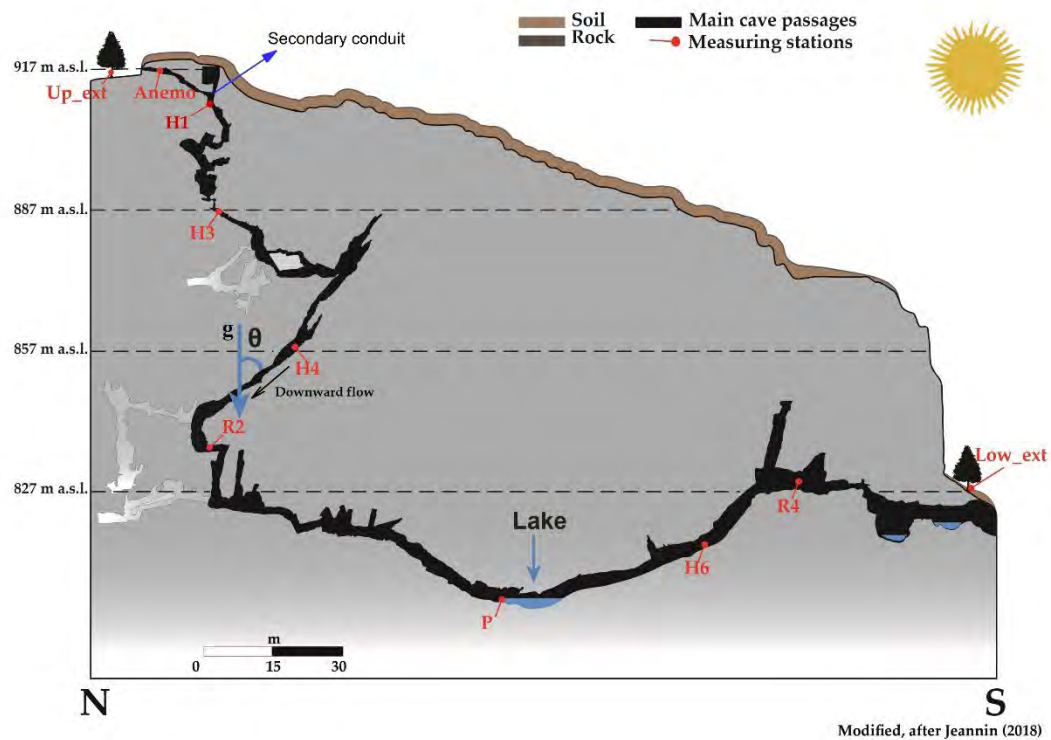


Figure 9.1. Longeague Cave topography (developed vertical profile; main conduit in black, sensor locations in red).

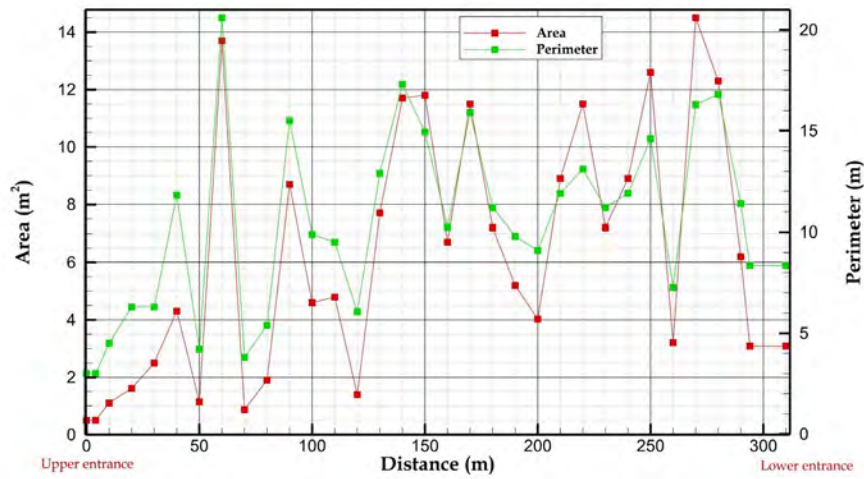


Figure 9.2. Variation of the cross-sectional area A and perimeter P as a function of the distance x from the upper entrance.

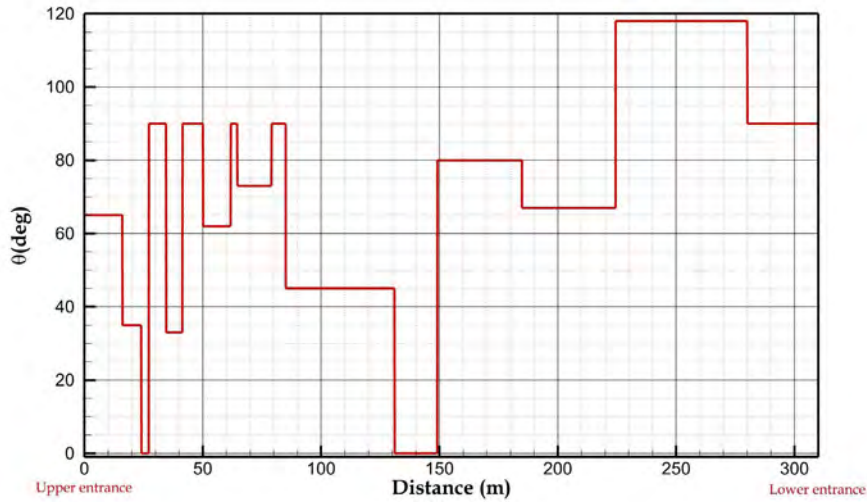


Figure 9.3. Angle θ between the gravity vector and the velocity vector of the air when flowing from the upper to lower entrance as a function of the distance x from the upper entrance (θ is larger than 90° in the conduit segment between P and $R4$ stations, see Figure 9.1).

9.2.2. Temperature measurements

The temperature data logging stations indicated by H in Figure 9.1 were equipped with Hobo Water Temperature Pro v2 (sensor type U22-01, accuracy $\pm 0.21^\circ\text{C}$, resolution 0.02°C). Reefnet Sensus Ultra temperature (accuracy $\pm 0.3^\circ\text{C}$, resolution 0.01°C) were deployed at the stations indicated by R. Two additional Reefnet sensors were used to monitor the atmospheric temperature close to the upper and lower entrances. Another temperature sensor was installed in Anemo station close to the upper entrance (sensor type MS8607 (Adafruit), accuracy $\pm 1^\circ\text{C}$, resolution 0.01°C). The distance of each station from the upper (and lower) entrance are listed in Table 9.1. To make the conduit open enough for the commuting of cavers, all the sensors were located close to the wall, at a distance ranging from 2 to 4 cm.

With the exception of H1, the temperatures used in this paper were recorded from 13/08/2020 to 13/08/2021, which is exactly one year. The Hobo water probe installed in H1 at a few centimeters from the wall was complemented with a Pt100 sensor (TMC1-HD from Hobo Company, resolution: 0.02°C ,

accuracy: 0.1) suspended approximately at the center of the cross-section. The temperature of these two probes were recorded for approximately one month, from 11/10/2022 to 06/11/2022 (sampling rate for all temperature measurements: one measuring point per hour).

Table 9.1. Location of measuring stations

Station	Distance from the upper (lower) entrance
Up_ext	Outside, near the upper entrance
Anemo	4 m (307 m)
H1	16 m (295 m)
H3	55 m (256 m)
H4	97 m (214 m)
R2	140 m (171 m)
P	216 m (95 m)
H6	269 m (42 m)
R4	294 m (17 m)
Low_ext	Outside, near the lower entrance

External upper and lower temperatures are displayed in Figure 9.4a. The dashed lines represent the function:

$$T_{atm}(t) = T_m - \Delta T \sin \left[\frac{2\pi}{\tau} (t - t_0) \right] \quad (1)$$

where $\tau = 1$ year, t_0 is a parameter that accounts for the origin of time, T_m is the annual mean temperature (AMT) and ΔT the amplitude of the annual temperature fluctuations (ATF). t_0 , T_m and ΔT are deduced from the Fast Fourier Transform (FFT) of the atmospheric temperatures displayed in Figure 9.4b, which also provides the amplitude of all other modes, including that of the daily temperature fluctuations (DTF). As expected, the modes showing the highest amplitudes are the ATF and the DTF. The ATF amplitudes at the upper and lower entrances (respectively 8.34 and 8.28°C) are very close to each other. The amplitudes of the DTF differ a little more (3.4°C and 2.4°C at the upper and lower entrances, respectively). The AMT is higher at the upper entrance (7.38°C) than at the lower entrance (6.75°C). The mean vertical thermal gradient between both entrances is thus 6.5 °C/km.

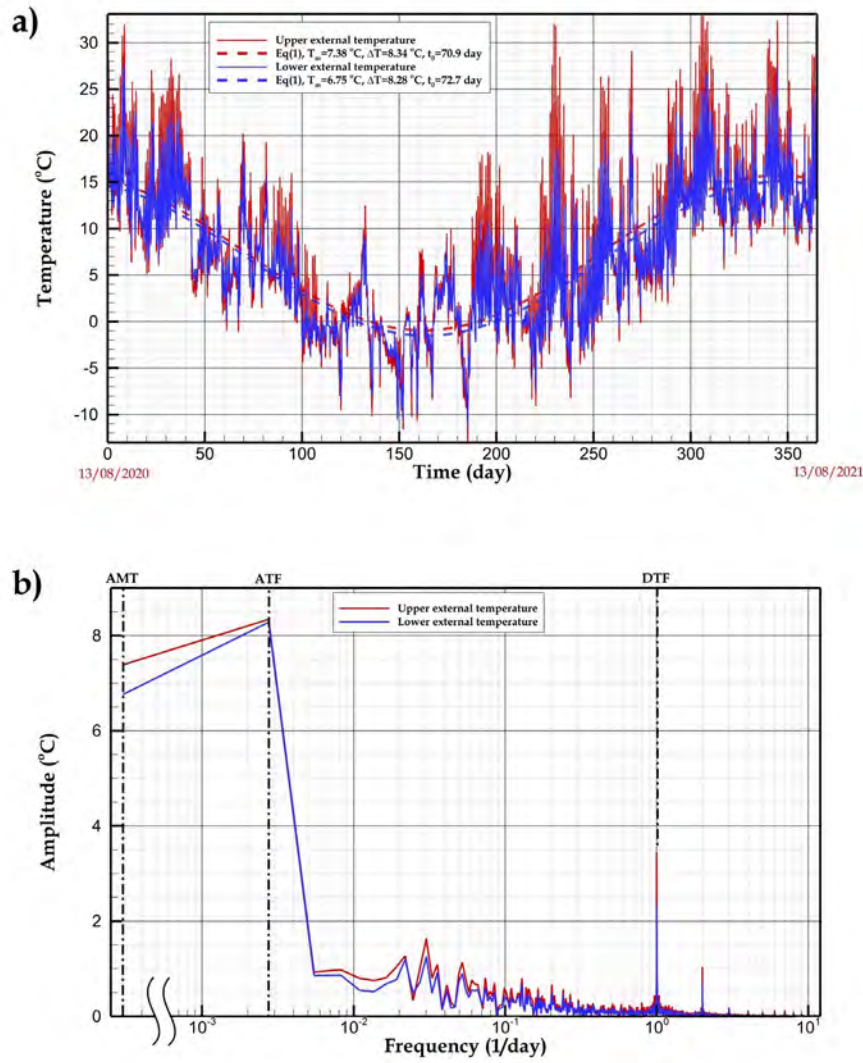


Figure 9.4. Atmospheric temperature close to the upper and lower entrances. a) Time series; b) Amplitude as a function of frequency deduced from FFT (due to the logarithmic scale the AMT, which corresponds to zero frequency, is located at an arbitrary origin).

9.2.3. Airflow measurements

The Sensirion SFM3003-300-CE (from now on SFM) is a digital bidirectional mass flowmeter installed at Anemo station, at 4 m from the upper entrance (Figure 9.1). The device is compact and robust. It shows a high data resolution (up to 16 bit) and low power requirement (typically 3.3 V and 3.8 mA). Furthermore, the bidirectional cave airflow is well recognized by the device which records the flow direction as negative or positive values. The device measures the flowrate that passes through a tube with an inner diameter of 20 mm. This flowrate is then converted in local air velocity through appropriate calibration with an accuracy of $\pm 5\%$ for speeds higher than 0.30 m/s and $\pm 26\%$ for speeds in the range from 0.15 to 0.30 m/s (Pastore et al., 2024). The flowrate through the conduit is obtained by multiplying this local velocity by the cross-sectional area. The main cause of uncertainty is the non-uniform velocity field over the cross-section. Complementary measurements performed by other means (including manual measurements with an anemometer, CO₂ gauging, comparison with other SFM) suggest that: (a) the flowrate measured at Anemo station is certainly underestimated, (b) the ratio of the real flowrate over the measured flowrate cannot be larger than 1.5.

The air mass flowrate was recorded simultaneously with the temperatures using the same sampling rate (see Figure 9.5a). It varies approximately in the range from -1.5 kg.s^{-1} in winter to 1 kg.s^{-1} in summer (positive values correspond to airflow direction from the upper to the lower entrance). The sporadic interruptions of airflow when the lake completely fills the lower part of the conduit are clearly observed in Figure 9.5a. A pressure sensor has been installed in point P (see Figure 9.1 and Table 9.1) to measure the lake level. It provides the time ranges during which the main conduit is blocked by the lake (called “cave closures” in the following). Figure 9.5a shows that the flowrate measured at Anemo station during these cave closures is not always zero. This can be due to the existence of a leakage rate through the secondary conduit (see Section. 2.1 and Figure 9.1), or to sporadic free convection cells developing from the upper entrance, inducing local velocities recorded by the SFM. Conversely, it can happen that no airflow is measured whereas the pressure sensor indicates that the cave is open (e.g., for time ranging from 196 and 213 days). This happens when the flowrate is too small to be detected by the SFM, or because of an uncertainty on the lake level corresponding to cave closures. During the monitoring year, the cave was closed for 203 days (56% of the time). The airflow was directed from the upper to the lower entrance for 66 days and in the reverse direction for 95 days (respectively 18% and 26% of the time).

The FFT of the mass flowrate is displayed in Figure 9.5b. In contrast with the temperature spectrum that was dominated by the annual and daily fluctuations, many other modes emerge with amplitudes comparable or even higher than the annual flowrate fluctuations (AFF) or daily flowrate fluctuations (DFF). These modes with intermediate frequencies originate from the intermittent lake closures. Moreover, the annual mean flowrate (AMF) is strictly positive, and equal to half the AFF. This seasonal asymmetry is due to the highest flowrates reached in winter. The dashed lines in Figure 9.5a represent the same sine function as in Eq. (1) after substituting temperatures for mass flowrates.

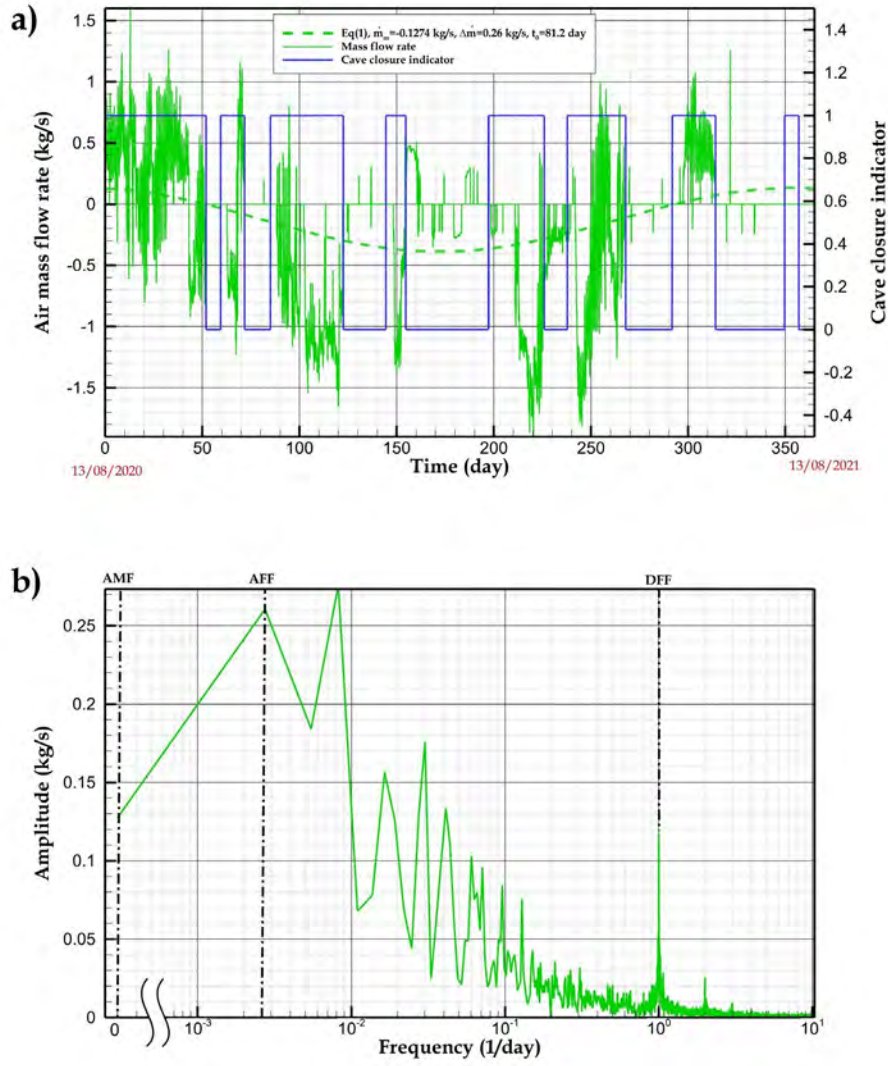


Figure 9.5. Air mass flowrate measured at Anemo station; a) Time series (dashed line corresponds to Eq. (1) where T_m and ΔT are replaced by the AMF $\dot{m}_m$ and the AFF $\Delta \dot{m}$, respectively). The cave closure periods are shown by blue solid line determining if the cave is open (one) or closed (zero). b) Amplitudes versus frequency deduced from FFT (due to the logarithmic scale, the AMF, which corresponds to zero frequency, is located at an arbitrary origin).

9.3. Numerical simulations

9.3.1. Simplifying assumptions

Only the main conduit connecting the upper to the lower entrance is considered in the numerical simulations (see Figure 9.1). For the sake of simplification, it has been “unfolded” to yield a rectilinear conduit of the same length L_{cave} as the real cave (see Figure 9.6). A circular cross-section of variable diameter $D_p(x)$ is assumed. $D_p(x)$ is set to get the same perimeter $P(x)$ as the true conduit (i.e., $D_p(x) = P(x)/\pi$, where the perimeter $P(x)$ is known from the cave survey, see Figure 9.2). The sensitivity study presented in Appendix A suggests that this choice, which preserves the exchange surface between the air and the rock, is a good approximation in most cases. The conduit is located in an impermeable rock domain of outer radius R_{dom} where conduction is the only heat transfer process. The outer radius must be large enough to ensure that the corresponding domain boundary is not

significantly influenced by the airflow in the conduit and is thus approximately adiabatic. These simplifications allow to consider a 2D axisymmetric conduction problem in the rock mass, resulting in a significant reduction in the computational resources compared to the real 3D geometry. Moreover, we assume that the temperature of the external surface is equal to the measured atmospheric temperature. This boundary condition offers the advantage of simplicity, but neglects the effect of radiative transfer and evaporation or condensation on the temperature of the external surface.

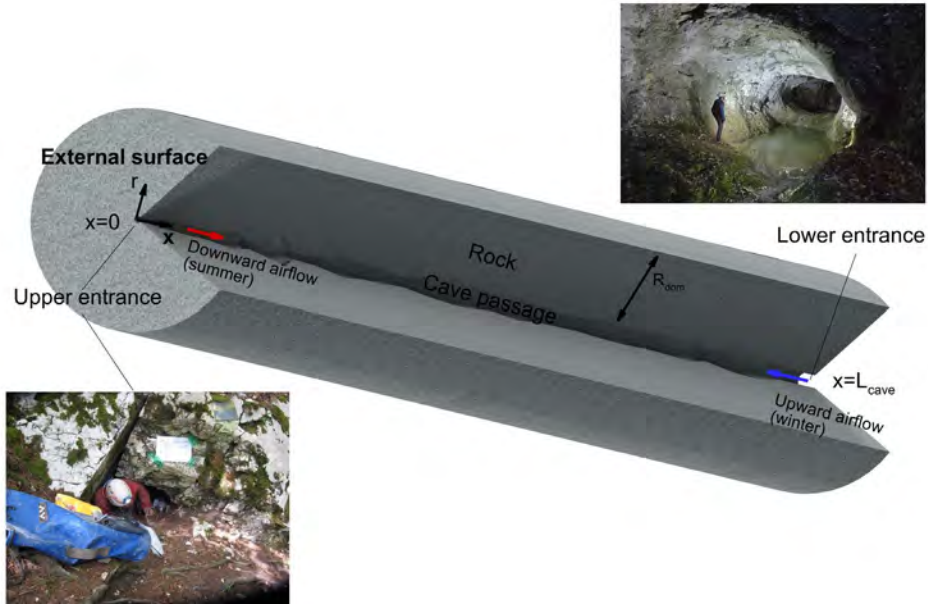


Figure 9.6. The simplified geometry of the ventilated cave with a single conduit and two entrances at different elevations

The need for an aerodynamic model is avoided by using the air flowrate measured by the SFM. During cave closure periods, we assume that the air only circulates in the part of the conduit between the upper entrance and the junction with the secondary conduit (at 16 m from the upper entrance). We impose zero flowrate in the rest of the conduit (i.e., between the junction with the secondary conduit and the lower entrance). Although the conduit is assumed rectilinear, the variation of potential energy with the altitude will be considered in the air energy balance using the conduit tilt angle $\theta(x)$ displayed in Figure 9.3.

The effect of the latent heat of evaporation or condensation inside the cave is an important issue that must be evaluated. A difficulty is that the cave walls might be covered with a water film or completely dry, depending on the season and the weathering conditions. To avoid the complicated task of defining a model predicting the amount of water lying on the walls, we define two simple limiting cases representing a lower and an upper bound of the latent heat effects.

- Case 1: evaporation and condensation are disregarded (no latent heat effect).
- Case 2: a permanent thin film of water covers all the cave walls, so that evaporation cannot be interrupted by the lack of water on the walls. In addition, the water film is assumed thin enough to neglect its thermal resistance compared to that due to conduction in the rock or convection. Condensation in the gas phase is disregarded. We assume that it only takes place on the conduit wall (same assumptions as in Qaddah et al., (2023)).

The advection of heat by water flowing down the conduit is disregarded. This assumption seems reasonable when the water flow reduces to a thin liquid film due to condensation or percolation

through the rock porosity. In contrast, it is questionable during cave flooding induced by intense water recharge. This point is discussed in Section 4.2.

The heat flux through the conduit wall, which couples the energy balance in the air to the conduction equation in the rock mass, is estimated using the Newton's law of cooling based on a heat transfer coefficient. When condensation and evaporation are considered, the mass balance of the water vapor transported by the humid air is also included in the model. This equation contains a mass transfer coefficient for the assessment of the evaporation/condensation fluxes at the cave wall. The Lewis analogy (Cussler and Cussler, 2009) states that the heat and mass transfer coefficients follow the same physical laws. To overcome the difficulty that arises from the complex geometry of a cave, we adopt the same method as in our previous work (Sedaghatkish et al., 2024). We perform two distinct set of simulations to assess a lower bound and a higher bound of the heat and vapor fluxes through the cave wall. In a first set, the transfer coefficients are estimated from standard correlations valid for forced convection and fully developed flow in pipes, which is expected to yield a lower bound. In a second set of simulations, an upper bound of these fluxes is obtained assuming infinite transfer coefficients. Combining dry or humid air along with finite or infinite transfer coefficients yields the four limiting cases listed in Table 9.2.

Table 9.2. Abbreviated names of the four cases considered in the numerical simulations. The estimation of the transfer coefficients required for DA and HA models is detailed in Appendix B.

	Transfer coefficients	
	Finite	Infinite
1- Dry air	DA	DA_∞
2- Humid air	HA	HA_∞

9.3.2. Governing equations

We first detail the complete set of equations for the **HA** case (humid air and finite transfer coefficients), which is the most general. Then we give the simplifications for the simulation of dry air or infinite transfer coefficients.

9.3.2.1. Rock mass

The temperature field in the impermeable rock mass follows the heat conduction equation in the cylindrical coordinates system:

$$\frac{1}{r} \frac{\partial}{\partial r} \left(r \frac{\partial T_r}{\partial r} \right) + \frac{\partial^2 T_r}{\partial x^2} = \frac{1}{\alpha_r} \frac{\partial T_r}{\partial t} \quad (2)$$

$T_r(x, r)$ is the rock temperature at distances x from the upper entrance and r from the conduit axis. α_r is the thermal diffusivity of the rock.

Defining right boundary condition at the external ground surface is challenging. For the sake of simplicity, the atmospheric temperature is imposed at the external surfaces located at $x = 0$ and $x = L_{cave}$:

$$T_r(0, r, t) = T_{atmU}(t) \quad \text{for } R_p(0) \leq r \leq R_{dom}, \quad (3)$$

$$T_r(L_{cave}, r, t) = T_{atmL}(t) \quad \text{for } R_p(L_{cave}) \leq r \leq R_{dom}, \quad (4)$$

where $R_p(x) = D_p(x)/2$ is the cave radius that yields the perimeter obtained from the cave survey (Figure 9.2). T_{atmL} and T_{atmU} are the atmospheric temperatures measured in the vicinity of the lower

and upper entrances, respectively (Figure 9.4). The impact of the boundary conditions defined in Eqs. (3,4) will be analyzed in detail in section 5.1.

Assuming an adiabatic boundary at $r = R_{dom}$ yields

$$\frac{\partial T_r}{\partial r}(x, R_{dom}, t) = 0 \text{ for } 0 \leq x \leq L_{cave} \quad (5)$$

The energy conservation at the conduit wall reads:

$$k_r \frac{\partial T_r}{\partial r}(x, R_p, t) = \varphi_c + L_v J_w \quad (6)$$

where k_r is the thermal conductivity of the rock, φ_c is the convective heat flux at the conduit wall (positive when directed from the rock to the air), L_v is the molar latent heat of evaporation (J.mol^{-1}) and J_w is the molar flux at the conduit wall ($\text{mol.m}^{-2}.\text{s}^{-1}$), positive for evaporation, negative for condensation. Eq. (6) states that the sum of the convective and latent heat fluxes is equal to the conductive heat flux leaving the rock mass. This equation couples the rock mass with the airflow.

9.3.2.2. Airflow

Since the water vapor is always dilute (its mass fraction in the humid air never exceeds 2%), the heat capacity of the water vapor can be neglected against that of the air. With this assumption, the energy balance in the air reduces to:

$$\rho_a c_{v,a} A(x) \frac{\partial \bar{T}_a}{\partial t} + \dot{m} \left(c_{p,a} \frac{\partial \bar{T}_a}{\partial x} - g \cos(\theta(x)) \right) = P(x) \varphi_c, \quad (7)$$

where $\bar{T}_a(x)$ is the mixing temperature of the air, i.e., the air temperature averaged over the conduit cross-section (Bergman et al., 2017):

$$\bar{T}_a(x) = \frac{1}{\dot{m}} \iint T_a(x, r) \rho_a u(x, r) dA \quad (8)$$

where $T_a(x, r)$ is the local air temperature and $u(x, r)$ the axial component of the local velocity vector. $A(x)$ and $P(x)$ are the cross-sectional area and perimeter obtained from the cave survey and displayed in Figure 9.2. $\dot{m}$ is the air mass flowrate measured by the SFM-device and displayed in Figure 9.5 (positive when the air enters through the upper entrance). ρ_a , $c_{v,a}$ and $c_{p,a}$ are the air density, specific heat at constant volume and specific heat at constant pressure.

The first term in the left-hand side (LHS) of Eq. (7) is an approximate expression of the rate of variation of the air internal energy. This term is necessary to regularize the solution when rapid fluctuations occur during flow reversals. Most of the time, it is negligible compared to the other terms. Therefore, Eq. (7) mainly expresses a balance between the energy transported by the fluid (thermal and gravitational potential energy in the second term of the LHS) and the heat transferred to the conduit wall, in the right-hand side (RHS).

The inlet temperature is equal to the atmospheric temperature T_{atmL} or T_{atmU} according to the direction of the flow:

$$\bar{T}_a(0, t) = T_{atmU}(t) \quad \text{for } \dot{m} > 0 \quad (9)$$

$$\bar{T}_a(L_{cave}, t) = T_{atmL}(t) \quad \text{for } \dot{m} < 0 \quad (10)$$

The water vapor balance in the humid air and its associated boundary conditions are similar to Eqs. (7-10):

$$A(x) \frac{\partial \bar{c}_v}{\partial t} + \frac{\dot{m}}{\rho_a} \frac{\partial \bar{c}_v}{\partial x} = P(x) J_w \quad (11)$$

$$\bar{c}_v(0, t) = \phi_{ext} c_{s,atmU} \text{ for } \dot{m} > 0 \quad (12)$$

$$\bar{c}_v(L_{cave}, t) = \phi_{ext} c_{s,atmL} \text{ for } \dot{m} < 0 \quad (13)$$

where $\bar{c}_v$ is the water vapor molar concentration (mol.m^{-3}) averaged over the cross-section similarly to the temperature in Eq. (8). Index “s” indicates saturated condition defined as:

$$c_s = \frac{p_{sat}(T)}{RT} \quad (14)$$

where p_{sat} is the saturated water vapor pressure at temperature T (Lide, 2004). R is the ideal gas constant. Thus, $c_{s,atmU}$ and $c_{s,atmL}$ are the saturated water vapor concentrations based on the atmospheric temperature close to the upper and lower entrances, respectively. ϕ_{ext} is the atmospheric relative humidity. The same constant value of 75% is assumed at the lower and upper entrances.

9.3.2.3. Heat and vapor fluxes at the conduit wall

The Newton’s law of cooling is a general law that applies to any convection process (Bergman et al., 2017). This law states that the heat flux transferred from the wall to a flowing fluid is proportional to the temperature difference between the wall and the fluid:

$$\varphi_c(x, t) = h_t (T_w(x, t) - \bar{T}_a(x, t)) \text{ for } 0 \leq x \leq L_{cave}, \quad (15)$$

where $T_w(x, t) = T_r(x, R_p, t)$ is the wall temperature and $\bar{T}_a(x, t)$ the air mixing temperature defined in Eq.(8). The heat and mass transfer analogy provide the same kind of relation for the water vapor flux:

$$J_w(x, t) = h_m (c_{sw}(x, t) - \bar{c}_v(x, t)) \text{ for } 0 \leq x \leq L_{cave}, \quad (16)$$

where c_{sw} is the saturated water vapor concentration obtained by substituting T for T_w in Eq. (14). h_t and h_m are the heat and mass transfer coefficients, respectively. In a pipe flow, they depend on the physical properties of air (Table 9.3) and the conduit geometry defined by the cross-section $A(x)$ and the perimeter $P(x)$ (Figure 9.2). In the turbulent regime, they also depend on the air flowrate $\dot{m}$ (Figure 9.5) and the wall relative roughness ε , defined as the ratio of the roughness over the hydraulic diameter $D_h(x) = 4A(x)/P(x)$. We set $\varepsilon = 0.01$ in all simulations.

Assessing the heat and mass transfer coefficients requires the evaluation of the Reynolds number Re that characterizes the ratio of inertia to viscous friction in the airflow:

$$Re(x) = \frac{\rho_a |\bar{u}(x)| D_h(x)}{\mu_a} = \frac{4 |\dot{m}|}{P(x) \mu_a} \quad (17)$$

where $\bar{u} = \frac{\dot{m}}{(\rho_a A)}$ is the mean velocity over the cross-section and μ_a the dynamic viscosity of the air. At fixed air flowrate, the Reynolds number is inversely proportional to the conduit perimeter $P(x)$. At the maximum mass flowrate (of the order of 1 kg.s^{-1} , see Figure 9.5) Re ranges from 10^4 (in the largest parts of the conduit) to 10^5 (in the narrowest parts) denoting turbulent flow throughout the cave. The method used to estimate the transfer coefficients from standard correlations for fully developed flow in pipes in the laminar and turbulent regimes is detailed in Appendix B.

9.3.2.4. Infinite transfer coefficient and/or dry air

Eqs. (2-16) assume humid air and finite transfer coefficients (**HA**). In the limiting case of infinite heat transfer coefficients (**HA_∞**), the conditions at the conduit wall Eqs. (15-16) are replaced by

$$T_a(x, t) = T_w(x, t) \text{ for } 0 \leq x \leq L_{cave}, \quad (18)$$

$$c_v(x, t) = c_{sw}(x, t) \text{ for } 0 \leq x \leq L_{cave}. \quad (19)$$

Injecting Eqs. (18-19) in Eqs. (7) and (11) provides the relations for φ_w and J_w required by Eq. (6). A noticeable consequence of Eqs. (18-19) is that the humid air inside the cave is saturated at the wall temperature all over the conduit.

In the case of dry air with finite transfer coefficients (**DA**), Eqs. (11-14) and (16) are ignored and we impose $J_w = 0$ in Eq. (6). When, in addition, the heat transfer coefficient is infinite (**DA_∞**), Eq. (15) is replaced by Eq. (18).

9.3.2.5. Initial condition

The initial condition required for solving Eq. (2) is the temperature field at any point of the rock mass at the beginning of the cave monitoring (i.e., on August 13, 2020). Obviously, this information is not directly available, since the temperature is only known at a few locations where sensors have been installed. Because of the thermal inertia of the massif, the initial temperature field depends in a non-trivial way on the history of the atmospheric temperature and the air flowrate.

Interestingly enough, the upper entrance of Longeague was enlarged by cavers in 1982. This artificial modification likely induced a significant increase of the air flowrate through the main conduit. Although it is unlikely that the air flowrate before the opening was strictly equal to zero, it was certainly much lower than after the opening. We thus do the approximation that the main conduit was closed before this date, and the air flowrate insignificant. Considering this specific feature of Longeague Cave, an approximate initial temperature field was assessed assuming a simplified history consisting of two steps. We first assume no airflow in the conduit, resulting in an adiabatic conduit wall. More precisely, we solved the pure conduction problem defined by Eqs. (2-6), with $\varphi_c = 0$ and $J_w = 0$ in Eq. (6). The atmospheric temperatures in Eqs. (3,4) were approximated by Eq. (1) (displayed by the dashed lines in Figure 9.4a), which only considers the annual fluctuations. Starting from an arbitrary uniform initial temperature field, the simulations of this 1D conduction problem was carried out over 200 years, a time long enough to reach the periodic regime. In a second step, we used this result as the initial temperature field to simulate the 38 years between the opening of the conduit in 1982 and the beginning of the monitoring in 2020. The full **HA** model defined in Section 3.2 was used with the simplified atmospheric temperatures of Eq. (1) as in the first step, along with the same kind of simplified function for the air flowrate (dashed line in Figure 9.5a). The resulting temperature field was used as the initial condition for the numerical simulation of the year during which the cave was monitored.

9.3.3. Physical properties

The physical properties used in the numerical simulations are assumed constant. They are listed in Table 9.3. The saturated vapor pressure of water as a function of temperature was interpolated from data in (Lide, 2004).

Table 9.3. Physical properties used in the numerical simulations (reference temperature: 12°C)

Properties	Value and unit	Reference
Rock density	$\rho_r = 2325 \frac{kg}{m^3}$	(Tiwari et al., 2020)
Rock heat capacity	$c_{p,r} = 841.09 \frac{J}{kgK}$	(Tiwari et al., 2020)
Rock thermal conductivity	$k_r = 2.302 \frac{W}{m.K}$	(Tiwari et al., 2020)
Rock thermal diffusivity	$\alpha_r = 1.177 \times 10^{-6} \frac{m^2}{s}$	(Tiwari et al., 2020)
Air dynamic viscosity	$\mu_a = 1.77 \times 10^{-5} Pa.s$	(Bergman et al., 2017)
Air density	$\rho_a = 1.23 \frac{kg}{m^3}$	(Bergman et al., 2017)
Air kinematic viscosity	$\nu = \frac{\mu_a}{\rho_a} = 1.44 \times 10^{-5} \frac{m^2}{s}$	-
Air heat capacity at constant pressure	$c_{p,a} = 1007 \frac{J}{kgK}$	(Bergman et al., 2017)
Air heat capacity at constant volume	$c_{v,a} = 719.29 \frac{J}{kgK}$	-
Air thermal conductivity	$k_a = 0.0251 \frac{W}{m.K}$	(Bergman et al., 2017)
Air thermal diffusivity	$\alpha_a = 2.02 \times 10^{-5} \frac{m^2}{s}$	-
Ideal gas constant	$R = 8.314 \frac{J}{molK}$	-
Molar mass of water vapor	$M_w = 18.015 \frac{g}{mol}$	-
Diffusion coefficient of water vapor in air	$D_w = 2.43 \times 10^{-5} \frac{m^2}{s}$	(Cussler and Cussler, 2009)
Molar specific vaporization latent heat	$L_v = 44.55 \frac{kJ}{mol}$	(Riddick et al., 1986)
Prandtl number	$Pr = \frac{\nu}{\alpha_a} = 0.71$	-
Schmidt number	$Sc = \frac{\nu}{D_w} = 0.59$	-

9.3.4. Numerical methods

All the numerical simulations were performed with the commercial software Comsol Multiphysics, version 6.1. This software solves partial differential equations by finite elements (Galerkin method). Time discretization was based on implicit backward differentiation formula. The mesh was made of Lagrangian quadratic elements, including 12600 2D elements in the rock mass (mapped mesh) and 200 1D elements in the air. The hyperbolic Eqs. (7) and (11) required the implementation of the stabilization techniques detailed in Appendix C.

9.4. Results

9.4.1. Field data

The AMT, ATF and DTF deduced from the FFT of the field data are displayed by black symbols in Figs. 7a to 7c. Thermal anomalies are clearly visible in Figure 9.7a. Indeed, the AMT passes through a local maximum at $\sim 9.5^{\circ}\text{C}$ at H3 station located at 55 m from the upper entrance (warm anomaly). Conversely, a local minimum at $\sim 5.7^{\circ}\text{C}$ is observed at R4 station at 20 m from the lower entrance (cold anomaly). The warm anomaly is more pronounced than the cold one (approximately $+2^{\circ}\text{C}$ and -1°C relative to the outside temperatures at the upper and lower entrances, respectively). Between these two extrema (i.e., between H3 and R4 stations), the AMT continuously decreases with increasing x . The absence of plateau in the temperature profile suggests that the convection length is larger than half the cave length, and that there is no deep karst region in Longeague Cave.

Figure 9.7b displays the ATF. A sharp decrease is observed close to the entrances, as expected in diffusive regions (Sedaghatkish et al., 2024). The ATF then gradually declines in the convective region to reach a minimum nearby the middle of the conduit.

The DTF is displayed in Figure 9.7c. A sharp decrease is observed close to the entrances, similarly to the ATF. However, in the convective region, the DTF decreases over a much shorter distance than the ATF. It becomes hardly measurable beyond a few tens of meters from the entrances. This suggests that the distance of propagation of a thermal perturbation advected by the airflow should depend on the frequency.

The time series of field data from three selected stations are displayed by solid black lines in Figure 9.8. Station H3 (Figure 9.8a) is close to the upper entrance, station R2 (Figure 9.8b) is approximately in the middle of the cave, and station R4 (Figure 9.8c) is close to the lower entrance. Comparisons with the atmospheric temperatures displayed in Figure 9.4a confirms the strong damping of the daily temperature fluctuations. In contrast, the impact of cave closures on the temperatures is clearly observed at all stations displayed in Figure 9.8 (see for instance the time range from 120 to 196 days).

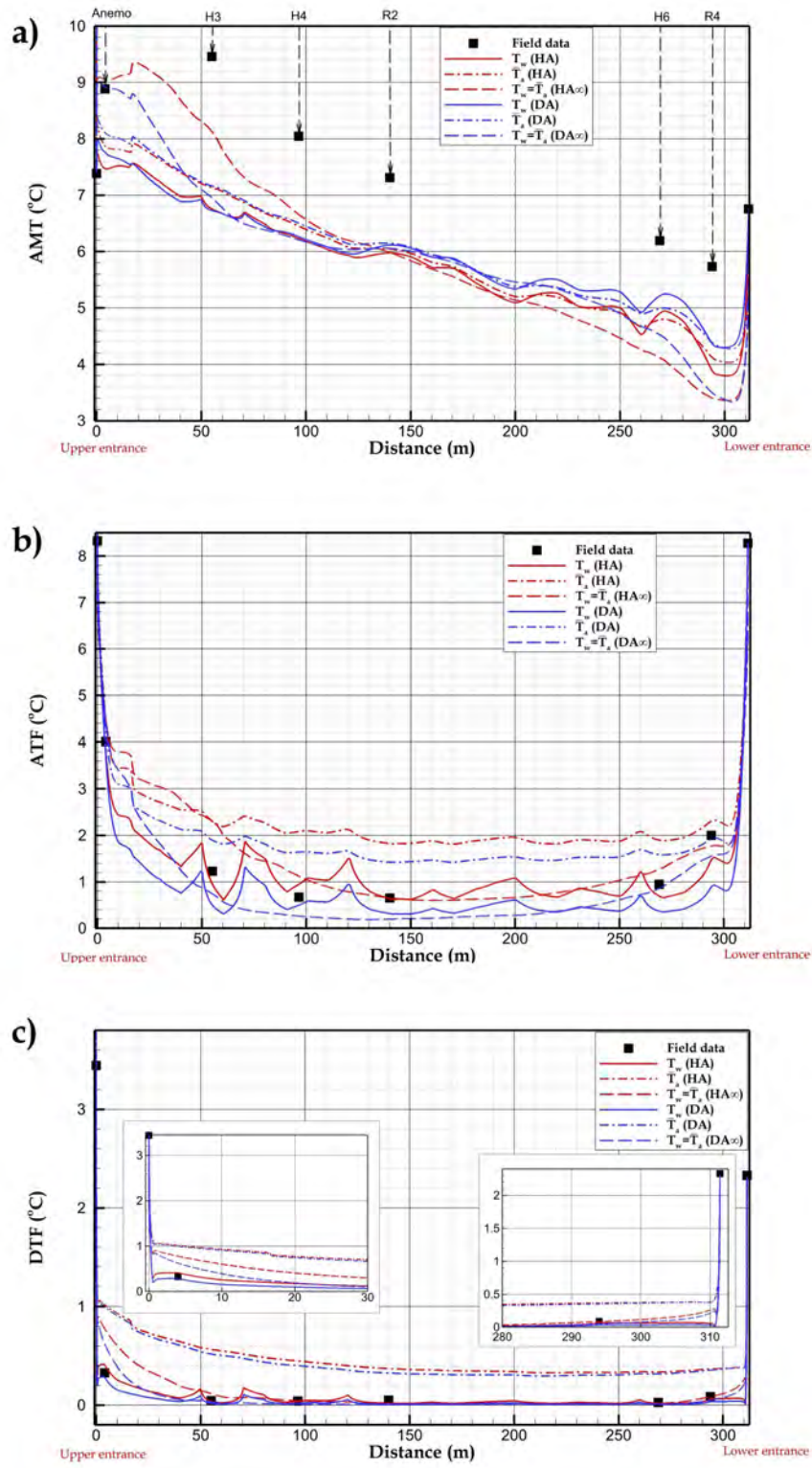


Figure 9.7. AMT(a), ATF (b) and DTF (c) from the field data and the numerical simulations. Data at $x = 0$ and $x = L_{cave} = 311$ m correspond to the atmospheric temperature close to the upper and the lower entrances, respectively.

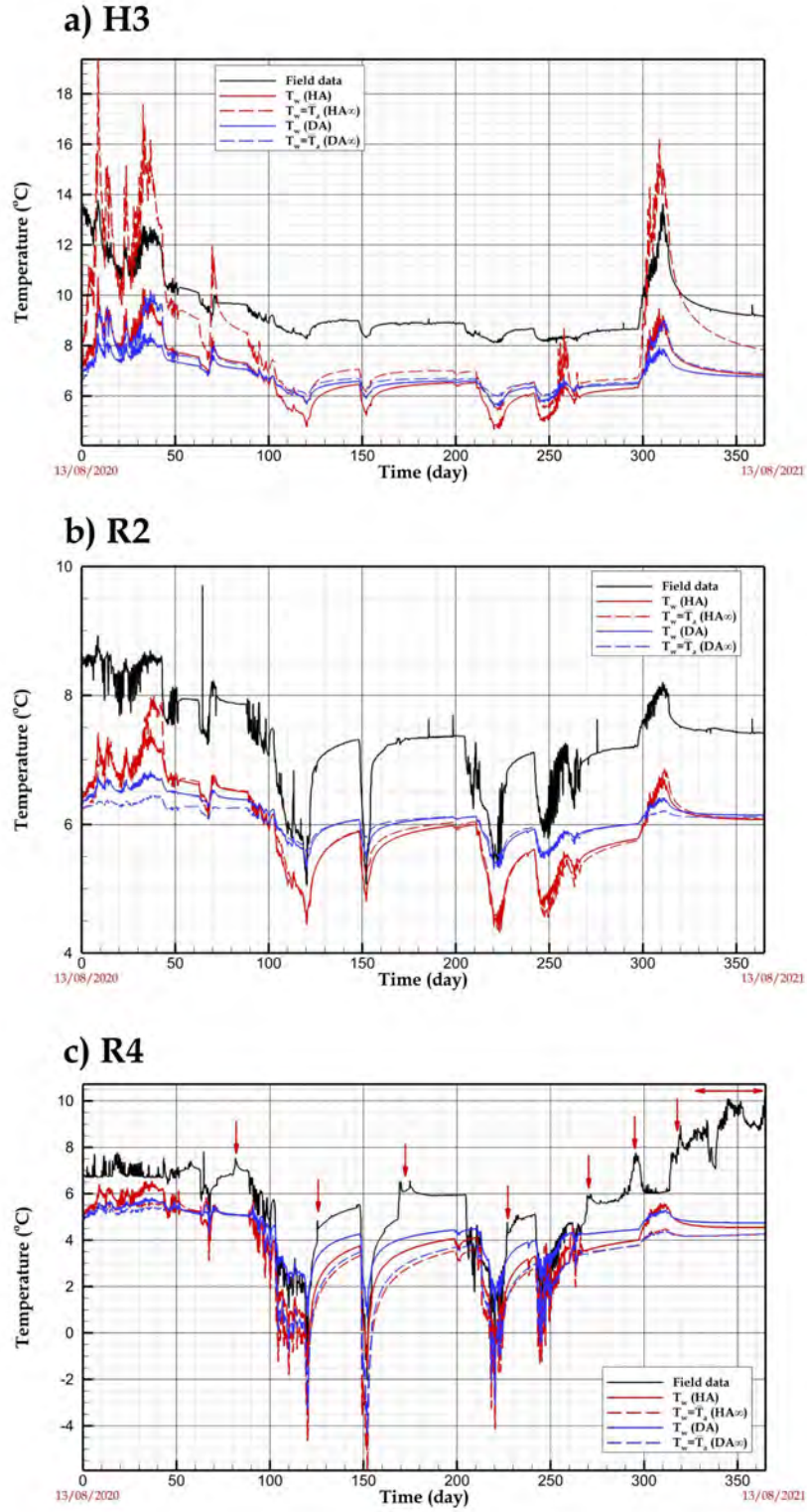


Figure 9.8. Temperature time series in three stations inside the cave a) H3 ($x=55$ m); b) R2 ($x=140$ m); c) R4 ($x=294$ m), red arrows indicates temperature peaks in the field data not predicted by the model.

9.4.2. Numerical simulations

Figure 9.7a displays the AMT of the conduit wall and the air as a function of the distance from the upper entrance, for the four models defined in Table 9.2 (notice that in $\mathbf{DA}_\infty$ and $\mathbf{HA}_\infty$ models, wall and air temperatures are equal). All models significantly underestimate the AMT computed from the field data (approximately by 1 or 2°C). $\mathbf{DA}_\infty$ and $\mathbf{HA}_\infty$ predict larger thermal anomalies compared to $\mathbf{DA}$ and $\mathbf{HA}$. $\mathbf{HA}_\infty$ seems to better reproduce the shape of the field data close to the upper entrance.

Regarding the ATF (Figure 9.7b), the four models reproduce qualitatively the main trends observed in the field data. All of them predict a sharp decrease over the first few meters from the entrances (i.e., in the diffusive regions) followed by a more gradual decline. $\mathbf{DA}_\infty$ globally underestimates the field ATF. $\mathbf{HA}_\infty$ is in good agreement with the field data in the lower half of the cave, but overestimates the ATF at $x = 55$ m (by approximately 1°C). The air and wall temperatures predicted by $\mathbf{DA}$ or $\mathbf{HA}$ models significantly differ from each other. The simulated air temperatures of both models overestimate the field data by approximately 1°C all over the cave. A significantly better agreement is obtained with the simulated wall temperatures. $\mathbf{HA}$ model shows a maximum error of approximately half a degree at H4 station, at 20 m from the lower entrance.

Similar comments apply to the DTF displayed in Figure 9.7c. $\mathbf{DA}_\infty$ and $\mathbf{HA}_\infty$ slightly overestimate the field data close to the upper entrance. The best fit is obtained for the wall temperature with $\mathbf{DA}$ and $\mathbf{HA}$, whereas the air DTF predicted by these models is significantly above the field data. This is confirmed by the time series at R4 station, at 20 m from the lower entrance (Figure 9.9). This station is close enough to the lower entrance to get a significant DTF when the lower entrance operates as an inlet. Compared to the field data, the wall temperature predicted by the $\mathbf{HA}$ model is shifted to lower values, with comparable (slightly lower) amplitudes. In contrast, the model predicts that the air temperature fluctuates with a much larger amplitude than the field temperatures.

The comparison between the simulated wall temperature and the field temperatures is completed by the time series displayed in Figs. 8a to 8c. At H3 ($x = 55$ m) and R2 ($x = 140$ m) stations, the simulated wall temperatures follow most of the time the field temperatures with a quasi-constant shift to lower values, in agreement with the difference between the simulated and field AMT observed in Figure 9.7a. An exception is the first 30 days of the monitoring period. During this time range, the field temperatures globally decrease, whereas all models predict increasing temperatures. This discrepancy is due to the approximated initial condition used in the simulations. The actual history of the atmospheric temperature before the initial time, different from the periodic function assumed in the simulations (see Section 3.2.5), results in significant errors during the first month. This time lapse is short compared to the duration of the monitoring (1 year).

The behavior observed in Figure 9.8c at R4 station ($x = 294$ m), located in the lower part of the cave, is different from H3 and R2. The evolution of simulated and measured temperatures not only differs during the first month of the simulated time range, but the field temperature also shows several peaks not predicted by the simulations. They are indicated by red arrows in Figure 9.8c. The most significant event is not localized at a specific time, but extends over approximately two months, from time $t = 310$ days to the end of the monitoring period (see Figure 9.8c). The same behavior has been observed at station H6 (figure not shown). These events are correlated with the lake level. During floods, H6 and R4 stations, both located in the lower part of the cave, are submerged, or very close to the water stream which modifies the temperature field in the cave. Because of thermal anomalies, a downward water flow induces heat transfer from the warmer upper part of the cave to the colder lower part (compare for instance the time series of the field temperatures at stations H3 ($x=55$ m) in Figure 9.8a and R4 ($x=294$ m) in Figure 9.8c). This explains that, in the lower part of the cave, floods always result in

temperature increases. In Longeague Cave, the rise of the atmospheric temperature with the altitude (see Section 2.2) might also play a role.

As stated in Section 3.1, heat advection by water flow is not considered in our models. Taking into account this effect would require the development of a hydraulic model, an intricate task beyond the scope of this article. However, comparisons between field and simulated temperatures reveal that the effect of water flow is only perceptible in the lower part of the conduit. At higher elevation, the climate is controlled by the airflow coupled to heat conduction in the rock mass. This is likely a general pattern in case of diffuse recharge through the epikarst. At the top of the massif, water infiltration is distributed all over the catchment area inducing low water velocity and thus negligible advection compared to heat conduction in the rock mass. When flowing downward, water concentrates in specific conduits until it reaches the spring where the water velocity, and thus the advection of heat, are maximum. The situation would be different in case of concentrated recharge (swallowing stream), as described by (Covington et al., 2011; Luhmann et al., 2012; Birk et al., 2006). In this case, the effect of water advection on the temperature field would be significant throughout the conduit, including its upper part.

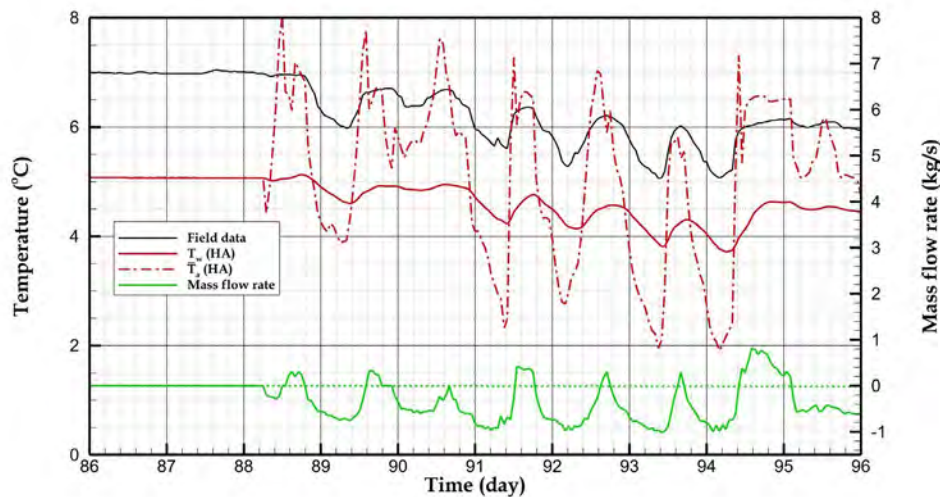


Figure 9.9. Temperature time series from field data and HA model at R4 station, at 17 meters from the lower entrance. Positive airflow indicates that the air enters through the upper entrance (origin of time: 13/08/2020).

The effect of the uncertainty on the airflow rate has been investigated by recalculating the outputs of **DA** and **HA** models after multiplying the airflow rate by 1.5 (see Appendix D). Although non-negligible, the impact on the results during the monitoring year is limited, and does not modify the observations drawn from the comparisons between the simulated and field temperatures. However, these observations raise issues requiring in-depth analyses developed in the next section:

- 1) Models systematically underestimate the annual mean temperatures obtained from field data;
- 2) Although the temperature sensors are located in the air, the amplitude measured in the field are much closer to temperature simulated for the cave walls than for the air;

Models with finite or infinite transfer coefficients yields significantly different results. What is the most relevant assumption?

9.5. Discussion

9.5.1. Annual mean temperatures

As pointed out in Section 4.2, the numerical simulations underestimate by approximately 1 or 2°C the AMT obtained from the field data (see Figure 9.7a). The predominant cause of this discrepancy, which is the main weakness of the model, must be sought in the boundary conditions at the external surfaces (i.e., the interfaces between the rock mass and the atmosphere, see Figure 9.6). Indeed, Eqs. (3-4) impose that the external surface temperature is equal to the atmospheric temperature. However, several effects may induce a significant gap between these temperatures. The energy balance at an external surface reads (Salmon et al., 2023):

$$(-k_r \cdot \vec{\nabla} T_r) \cdot \vec{n} = \varphi_{conv} + \varphi_{rad,lw} - \varphi_{rad,sw} + \varphi_{evap}. \quad (20)$$

This equation states that the conduction flux leaving the rock (LHS) is the sum of all the thermal fluxes transferred from the external surface to the external environment (RHS). $\vec{n}$ is the normal unit vector pointing to the atmosphere, φ_{conv} the convective heat flux from the rock to the atmosphere, $\varphi_{rad,lw}$ the net radiative flux lost by the rock in the long wave-length range, $\varphi_{rad,sw}$ the sun irradiation (short wave-length range) and φ_{evap} the latent heat flux (positive for evaporation). Some of these terms tend to increase the soil temperature (e.g., the sun irradiation), others to decrease it (e.g., the latent heat of evaporation). Molnar (2022) analyzed the atmosphere and soil AMT from 212 sites throughout the world. He concluded that the soil temperature is generally warmer than the atmosphere. The difference between the AMT of the soil and the atmosphere mainly depends on the land surface cover. It approximately ranges from 1°C in wetlands and forests to 3-5°C or more in arid or cold regions. Figure 9.10a displays the values measured at different weather stations located in Switzerland. Most data are included in the range from 0 to 2°C, with significant variations from year to year.

The sensitivity of the conduit temperature to the temperature of the external surfaces was assessed from additional numerical simulations. Figure 9.10b displays the AMT of the conduit wall obtained from the **HA** model in two cases: (a) equal atmospheric and wall temperatures, as specified by Eqs. (3-4) (blue curves), (b) external surface temperatures increased by $\Delta T = 2^\circ\text{C}$ compared to the atmospheric temperatures (red curves). In the latter case, the boundary conditions

$$T_r(0, r, t) = T_{atmU}(t) + \Delta T \quad \text{for } R_p(0) \leq r \leq R_{dom}, \quad (21)$$

$$T_r(L_{cave}, r, t) = T_{atmL}(t) + \Delta T \quad \text{for } R_p(L_{cave}) \leq r \leq R_{dom}, \quad (22)$$

have been used instead of Eqs. (3-4). Setting $\Delta T = 2^\circ\text{C}$ is a rough approximation since the actual difference between the atmospheric and external surface temperatures is unknown and may vary with time and space. However, this is relevant for a sensitivity analysis. To simulate a large time range (200 years) with a reasonable computational time, we used the simplified atmospheric temperatures of Eq. (1) and the same kind of simplified function for the air flowrate (dashed line in Figure 9.5a). The initial condition was determined using the method detailed in Section 3.2.5. Figure 9.10b shows that increasing the external surface temperatures by 2°C results in a similar temperature raise throughout the conduit wall, over the entire time range considered in the simulations (200 years). This result strongly suggests that the main source of discrepancy between the field and simulated AMT at the conduit wall may be the boundary conditions at the external surfaces.

Improving the accuracy of the model thus necessitates to replace Eqs. (3-4) with more realistic boundary conditions, based for instance on the energy balance Eq. (20). This would require appropriate models for assessing the various terms included in this equation (see for instance Larwa, 2018), and thus the monitoring of additional input data (e.g., the wind velocity for the assessment of φ_{conv} or the sky temperature for $\varphi_{rad,lw}$). An alternative could be to disseminate temperature sensors at a few tens

of centimeters below the external surface and use the measured temperatures in Eqs. (3-4) instead of the atmospheric temperature.

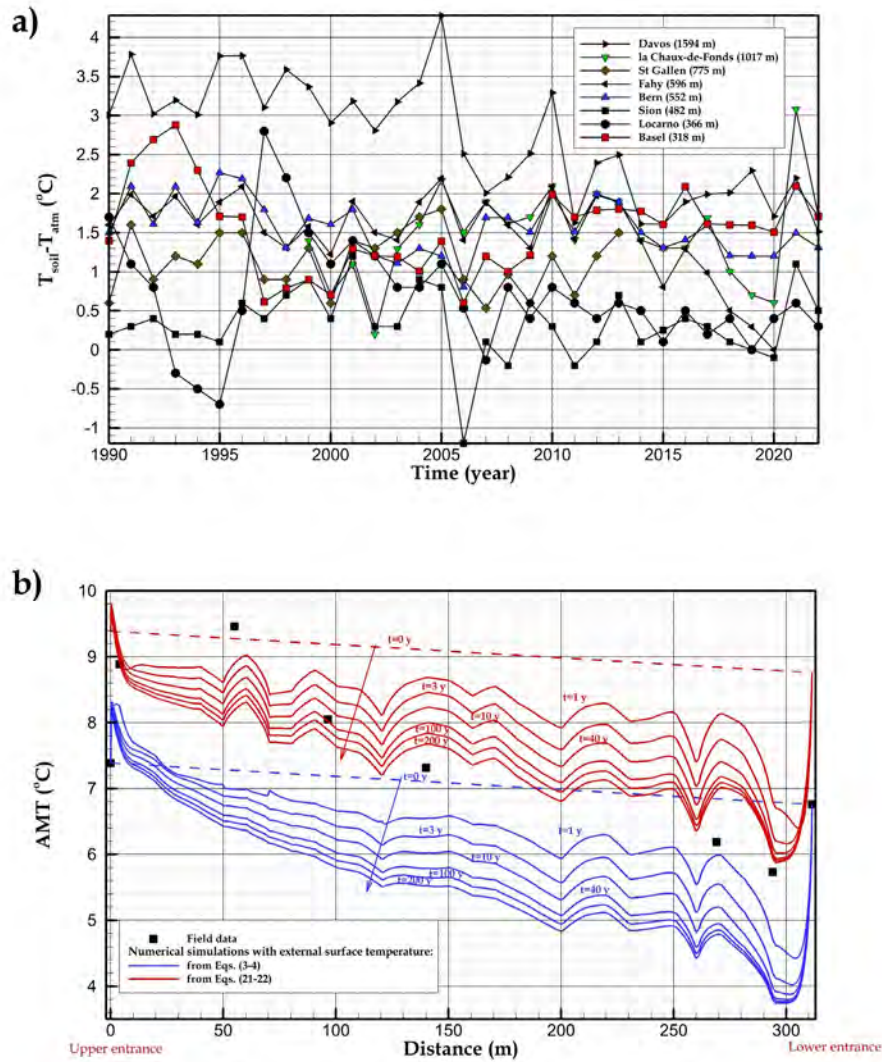


Figure 9.10 a) Mean annual temperature difference between the soil and the atmosphere ($T_{\text{soil}} - T_{\text{air}}$) for some different weather stations in Switzerland (from Federal Office for Meteorology and Climatology of Switzerland (2024)). The soil temperature is measured at 5 cm depth, the atmospheric temperature at 2 m above the ground. b) Effect of the external surface temperatures on the AMT of the conduit wall (numerical simulations with simplified time functions for the air flowrate and the atmospheric temperature, HA model). Blue curves: external surface temperature equal to the atmospheric temperature (Eqs. (3-4)). Red curves: external surface temperatures increased by $\Delta T = 2^{\circ}\text{C}$ (Eqs. (21-22)). Dashed lines: initial conditions obtained by solving the diffusion problem. The arrows point to increasing times.

9.5.2. Temperature fluctuations

The ATF and DTF of the air predicted by **DA** and **HA** models show significant discrepancies with field data (see Figs. 7b and 7c). **DA**_∞ and **HA**_∞ better fit the measured values, but are slightly less accurate than the wall temperatures yielded by **DA** and **HA** models. This is unexpected since the temperature sensors are located in the airflow, not in the rock. There are two possible explanations for this: (1) **DA** and **HA** models fail in predicting the air temperature, **DA**_∞ and **HA**_∞ models based on infinite transfer coefficients and equal temperatures in the air and the wall are closer to reality; (2) The wall temperature from **DA** and **HA** models fit well the field temperatures because the temperature sensors measure the wall temperature rather than the air temperature. Although counterintuitive, the latter explanation is the most likely. A first point is that the local air temperature T_a varies within the cross-

section, and reaches the rock temperature at the conduit wall. Incidentally, the mixing temperature of the air $\bar{T}_a$ provided by the numerical simulations is a cross-sectional averaged temperature. Considering its definition in Eq. (8), we expect $\bar{T}_a$ to be quite close to the local temperature in the center of the conduit, where the local velocity is maximum. A second point to be considered is that a temperature sensor always measures its own temperature (Lundström and Mattsson, 2020), which might differ from the temperature of the surrounding fluid.

Figure 9.11 displays the temperatures measured in the air at H1 station (at 16 m from the upper entrance) during approximately a month. Two temperature sensors were installed, a Hobo Water Pro v2 at a few centimeters from the wall (similar to the other probes used in this study) and a Pt100 sensor suspended approximately in the center of the conduit. Figure 9.11 shows that the amplitude of the daily oscillations measured by the Pt100 is two or three times larger than that measured by the Hobo Water sensor.

As pointed out by Lundström and Mattsson (2020), a sensor immersed in a transparent fluid (the air in the present case) receives a convective flux from the fluid and a net radiative flux from the conduit wall. Therefore, its temperature is a weighted average of the air and wall temperatures. A simple model based on the energy conservation yields the following expression for the sensor temperature (Lundström and Mattsson, 2020):

$$T_{sensor} = \frac{h_{conv}}{h_{conv} + h_{rad}} T_a + \frac{h_{rad}}{h_{conv} + h_{rad}} T_w \quad (21)$$

where h_{conv} and h_{rad} are the convective and radiative transfer coefficients at the sensor surface. Eq. (21) qualitatively explains the results displayed in Figure 9.11. A temperature sensor measures the air temperature if convection prevails over radiation (i.e., if $h_{conv} \gg h_{rad}$). Otherwise, it measures a temperature in between the wall and air temperatures, or even the wall temperature if $h_{conv} \ll h_{rad}$. h_{conv} is an increasing function of the air velocity and a decreasing function of the sensor size. h_{rad} is proportional to the emissivity of the sensor surface. Compared to the Pt100 sensor, the Hobo Water sensor is larger (30 mm against 5 mm) and has a higher emissivity (polypropylene sheath against metallic sheath). In addition, the Hobo Water sensor is located closer to the wall, where the air flows with a smaller local velocity and a local temperature T_a closer to the wall temperature T_w . All of these effects bring the Hobo Water sensor nearer the wall temperature. This sensor thus yields a better estimate of the wall temperature T_w than the air mixing temperature $\bar{T}_a$. In contrast, the temperature of the Pt100 sensor yields a better estimate of the air temperature in the center of the conduit. This interpretation, consistent with the results displayed in Figure 9.11, explains the good agreement between the field data and the wall temperature predicted by **DA** and **HA** models, and the discrepancy with the air temperature. Indeed, the configuration of all sensors used in this work is similar to that of the Hobo Water sensor used in the comparison displayed in Figure 9.11.

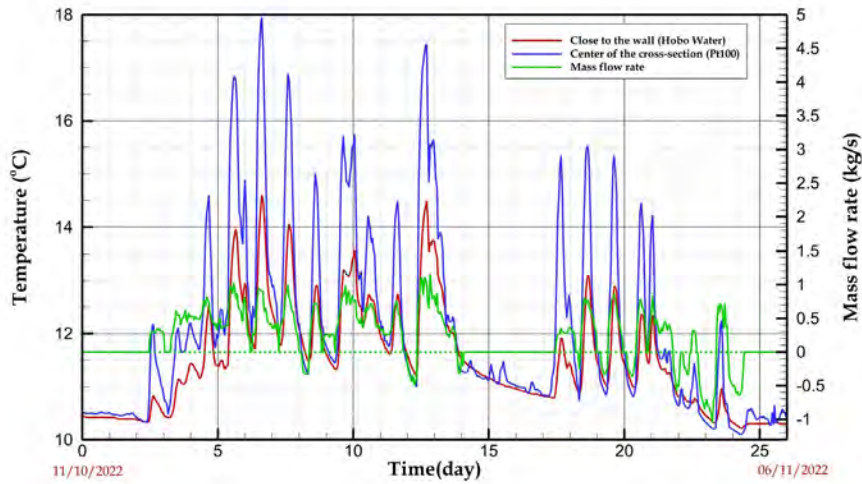


Figure 9.11. Temperature measured by a Hobo Water sensor at a few centimeters from the conduit wall and a Pt100 sensor in the center of the conduit. Test section H1 located at 16 m from the upper entrance. Positive flowrate indicates airflow entering through the upper entrance.

9.5.3. Finite versus infinite transfer coefficients

As pointed out in the introduction, assessing relevant values of the transfer coefficients in the complex geometry of a cave is a difficult task. However, the impact of transfer coefficients on the prediction of cave climate can be more or less significant depending on whether heat transfer is convection-limited or conduction-limited. In a previous work (Sedaghatkish et al., 2024), we showed from numerical simulations that no accurate estimation of the heat transfer coefficient is required when the Reynolds number corresponding to the maximum mass flowrate is larger than 3×10^5 (conduction-limited regime). Conversely, the heat transfer coefficient is a key information when the maximum Reynolds number is lower than 10^4 (convection-limited regime).

The maximum Reynolds number in Longeague Cave ranges from 10^4 to 10^5 , depending on the cave section (see Section 3.2.3). Longeague Cave is thus in the intermediate regime, in which both convection in the air and conduction in the rock mass should be considered. Consequently, the temperature fields predicted by **DA** and **HA** (finite transfer coefficients) or by **DA_∞** and **HA_∞** (infinite transfer coefficients) are significantly different. If we assume that:

- (1) the temperature sensors better estimate the wall temperatures than the air temperatures (see Section 5.2),
- (2) the general underestimation of the AMT by all models is mainly due to the initial and boundary conditions (see Section 5.1),

then **DA** and **HA** models yield more accurate results compared to **DA_∞** and **HA_∞** (see Figure 9.7). With this in mind, the fact that **DA_∞** and **HA_∞** better reproduce the measured AMT in the upper part of the cave (see Figure 9.7a) is likely a coincidence resulting from error compensation. In line with this, Figure 9.7b shows that, in the same region of the cave, the ATF is underestimated by **DA_∞** and overestimated by **HA_∞**.

As pointed out in Section 3.1, **DA** and **HA** models are two limiting cases (no latent heat effect in the former case, permanent water film on the cave walls in the latter case). Comparisons between the outputs of both models suggest that the effect of the latent heat on the cave climate is rather weak (see Figs. 7 and 8). Considering the uncertainties on the airflow rate and the temperature

measurements, the accuracy of **DA** or **HA** models for the prediction of the temperature fields are approximately equivalent.

9.6. Condensation and evaporation

9.6.1. Vapor transfer rate and consequences on paleoclimatology

Figure 9.12a displays the mass fluxes of water vapor at the conduit wall (positive and negative for evaporation and condensation, respectively) at two selected times corresponding to high air flowrates (**HA** model). Both mass fluxes are close to the maximum values reached during the year, one in summer ($t = 32.6$ days, $\dot{m} = 1.26 \text{ kg.s}^{-1}$, air intake from the upper entrance), the other in winter ($t = 220.3$ days, $\dot{m} = -1.92 \text{ kg.s}^{-1}$, air intake from the lower entrance).

In summer, evaporation is only observed in areas very close to the entrances: less than a meter from the upper entrance, a few meters from the lower entrance. Evaporation at the upper entrance arises from the need to increase the relative humidity from 75% (the atmospheric humidity) to 100% before condensation starts. At the lower entrance, evaporation results from the increase in the wall temperature in the diffusive region. With the exception of these two short regions, condensation takes place everywhere over the conduit wall. As expected, the condensation rate is maximum close to the inlet (the upper entrance), where the temperature contrast between the air and the wall is greatest. However, the condensation rate does not decrease monotonically with increasing distance from the upper entrance. Oscillations are correlated with conduit size variations (compare Figure 9.12a with the conduit perimeter displayed in Figure 9.2). We demonstrate in Appendix B that, in the turbulent regime and for a given air flowrate, the transfer coefficients are inversely proportional to the product of the conduit perimeter by the hydraulic diameter. This quadratic dependence results in a strong effect of the conduit size on the vapor transfer rate, with highest values in the narrowest parts of the conduit.

In winter, condensation is limited to a few meters from the upper entrance (the air outlet in winter). Condensation in this short region results from the decrease in the conduit wall temperature in the diffusive region. Evaporation takes place anywhere else. As expected, the evaporation rate is maximum close to the inlet (the lower entrance). Despite the higher air flowrate in winter, the evaporation rate is lower than the condensation rate in summer, for two reasons: (a) lower temperatures in winter (especially near the inlets) resulting in lower saturated vapor pressure and thus lower vapor concentration in the air; (b) larger conduit size in the lower part of the cave compared to the upper part (see Figure 9.2), resulting in lower transfer coefficients. The total amount of condensation and evaporation is assessed and discussed in Section 6.2.

A quantitative estimate of a cave's thermal response to the outside environment is essential for the interpretation of paleoclimate records from speleothems (Lyu et al., 2024). Although variations in the drip water composition largely control the proxy partitioning in speleothems (Fairchild and Baker, 2012), trace elements and oxygen isotope fractionation during CaCO_3 precipitation are also temperature-dependent (Tremaine et al., 2011; Wassenburg et al., 2020) and become significant for temperature amplitudes exceeding $\pm 0.5^\circ\text{C}$. The model developed in this study is capable of calculating this amplitude on daily and yearly scale at any location within the cave and illustrates the temperature sensitivity in the convection zone. Moreover, Figure 9.12a (red line) reveals that, in winter, evaporation is present all along the cave at significant rates but, in particular, close to the lower cave entrance. There, the evaporation flux may reach $2 \times 10^{-3} \text{ g.m}^{-2}.\text{s}^{-1}$ during intense upward air flow ($t = 220.3$ days; $\dot{m} = -1.92 \text{ kg.s}^{-1}$). Whilst this rate is still significantly lower than the potential evaporation rate considered in geochemical models (Dreybrodt and Deininger, 2014), it may become locally important at very low drip rates feeding speleothems. Results from the highly ventilated Longeague Cave thus suggest that the role of evaporation on the isotope proxy record can often be neglected in hydrologically active caves but must be considered in semi-arid environments with only sporadic drips

9.6.2. Role of condensation in water production

Evaporation and condensation produce and consume water inside the cave. Figure 9.12b displays the evaporation rate integrated throughout the conduit wall as a function of time (HA model, negative values refer to condensation). Condensation takes place during approximately two months in summer. The total amounts of condensed and evaporated water during the year were -8.8×10^3 and 16.7×10^3 kg, respectively. The mass of evaporated water throughout the year is twice the mass of condensed water. These values are low ($<0.1\%$) compared to the water flow-through in the groundwater catchment of Raies/Longeague system. Nonetheless, our model also shows that condensation may happen, preferentially at the upper entrance, during downward ventilation regimes. Whilst the total amount of condensed water during the 2020-21 annual cycle reached 8800 kg, it is still negligible with respect to the hydrological mass balance of Longeague Cave. In arid environments, this amount may nonetheless be sufficient to maintain a moist atmosphere in the upper entrance zone of a cave system and thus support the local ecosystem. Because cave-adapted species may be sensitive to even small changes in the subsurface environment (Medina et al., 2023; Jegla and Poulson, 1970), prediction of the amplitude and frequency of temperature changes might be crucial for cave conservation issues.

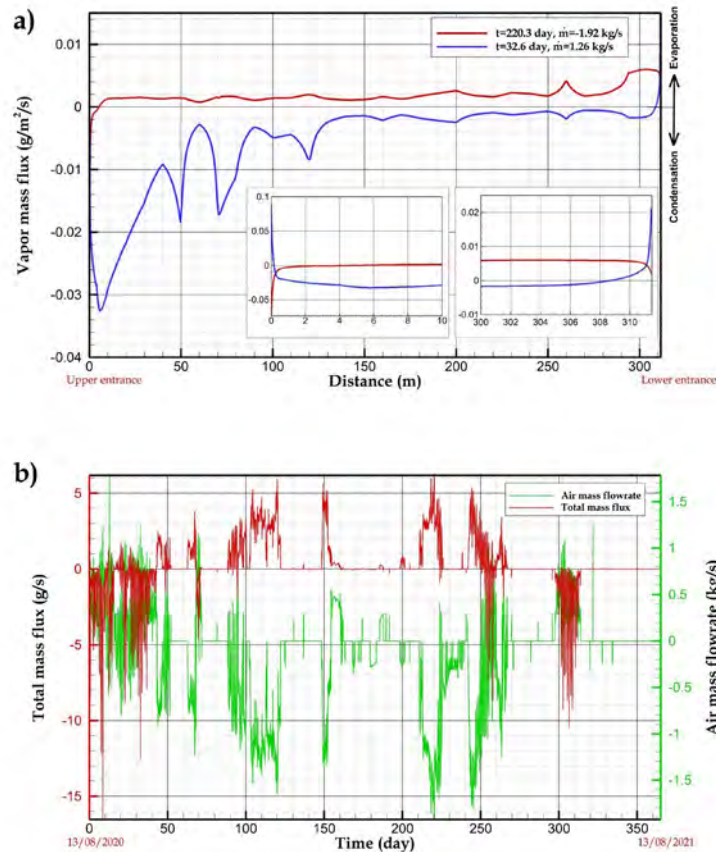


Figure 9.12. a) Vapor mass flux at the conduit wall corresponding to high air flowrate events in summer ($t = 32.6$ days) and winter ($t = 220.3$ days); insets: zooms in the first 10 meters from the entrances b) Evaporation flux integrated over the conduit wall as a function of time. In both figures, positive and negative values denote evaporation and condensation, respectively (HA model).

9.7. Conclusion

The main conduit of Longeague Cave is ventilated by a strong airflow driven by chimney effect. In order to predict the temperature field in this conduit, we developed a thermal model based on the diffusion equation in the rock mass along with the conservation of energy and vapor mass in the airflow. The conduit wall and the air are coupled by transfer coefficients assessed from standard correlations for fully developed forced convection in pipes. The model has been tested against the field data collected in the cave during a complete year.

The complex geometry of a cave is a serious difficulty for thermal modeling. A major simplification was done by assuming a rectilinear conduit of circular cross-section with variable diameter. In this context, the conduction problem in the rock mass is 2D axisymmetric, and only two independent parameters are required to characterize the conduit cross-section: the effective diameter D_p based on the conduit perimeter and the hydraulic diameter D_h . Using D_p in the conservation equations preserves the exchange surface between the air and the rock, while D_h is the relevant geometrical characteristic to be used in the estimation of the transfer coefficients.

The numerical results underestimate by 1 or 2°C the mean annual temperatures obtained from field data. This is the most significant weakness of the model. This discrepancy is mainly due to the assumption of equal rock and atmospheric temperatures at the external surface of the massif. More realistic boundary conditions are proposed.

In contrast, the model accurately predicts the temperature fluctuations from daily to yearly time scales, which includes in the specific case of Longeague Cave the temperature fluctuations generated by intermittent cave closures. The impact of the initial condition assessed from a simplified history is limited to the first month after the beginning of the simulation. Moreover, comparisons between field and simulated temperatures reveal that the effect of water flow due to diffuse water recharge is only perceptible in the lower part of the cave. At higher elevation, the temperature field is controlled by the airflow coupled to heat conduction in the rock mass. The latent heat effect of evaporation and condensation seems to play a minor role on the cave climate. However, predicting the order of magnitude of condensation or evaporation rate is valuable for applications related to ecology or paleoclimatology. In arid environments, maintaining moist atmosphere in the upper part of caves might be crucial for the ecosystems. The order of magnitude of evaporation rates also show that, in temperate European caves, the role of evaporation on the isotope proxy record can often be neglected.

Compared to heat diffusion in a rock mass, cave ventilation significantly increases the depth of natural thermal oscillations and decreases the time of propagation. This makes it necessary to develop and test thermal models for the prediction of temperature fields in ventilated karst massifs. We provided and discussed simplifying assumptions that allow the accurate prediction of the temperature field in a ventilated cave from daily to yearly time scale. However, extrapolating the results of this study to larger time horizons (centuries or more) must be undertaken with great caution. The long-term impact of geometry simplifications, or initial and boundary conditions, will require further research. This study is a first step towards the elaboration and validation of models capable of tackling these issues. Our model opens the way for a quantitative assessment of the cave's thermal response at any location, providing a well-known cave geometry.

The thermal model presented in this article is applied to caves naturally ventilated by chimney effect. However, some issues raised in this article are relevant to engineering applications with artificial ventilation, as mines (Yu et al., 2013; Parra et al., 2006) and tunnels (Lv et al., 2020). Defining the right boundary conditions or convection model also matters in these configurations. More specifically, there is a growing interest for the recovery of geothermal energy from tunnels (Dornberger et al., 2022).

Taking into account the advection of heat by the airflow and the coupling between convection and conduction in the rock mass should improve performance predictions for these complex geotechnical structures.

Appendix A. Effect of the cross-sectional shape on wall temperature and heat flux

The numerical simulation of 3D geometries requires a large amount of computational resources. The objective of this appendix is to test whether a 2D (axisymmetric) circular cross-section can be substituted for the 3D geometry of a real conduit with an acceptable loss of accuracy. To address this issue, we computed the temperature field in the rock mass surrounding the conduits displayed in Fig. A1. Three cases are considered: (a) the circular shape, (b) the triangular shape as an instance of simple 3D geometry, (c) a more complicated shape obtained from a survey in a real cave. The conduit perimeter P is expected to be an important parameter since it imposes the exchange surface between the rock and the air (see Eqs. 7 and 11). Therefore, all the comparisons between the different shapes displayed in Fig. A1 will be done between conduits of equal perimeter P .

The transient conduction equation is solved in the cross-sectional plane assuming uniform temperature in the direction parallel to the conduit. The temperature of the dry air inside the conduit follows a sinusoidal function of time. The heat flux at the conduit wall is deduced from the Newton's law of cooling, assuming uniform heat transfer coefficient all along the conduit circumference.

The problem variables are the coordinates x and y , the time t , the temperature in the rock mass T_r , and the temperature of the air $\bar{T}_a$. The corresponding dimensionless variables read:

$$X = \frac{x}{L_d}, Y = \frac{y}{L_d}, t^* = \frac{t}{\tau}, \theta_r = \frac{T_r - T_m}{\Delta T}, \theta_a = \frac{\bar{T}_a - T_m}{\Delta T} \quad (A1)$$

where T_m is the mean temperature of the air and ΔT the amplitude of the temperature fluctuations of period τ . $L_d = \sqrt{\alpha_r \tau}$ is the diffusion length in the rock mass. $P^* = \frac{P}{L_d}$ and $D_p^* = \frac{D_p}{L_d} = \frac{P}{\pi L_d}$ are the dimensionless perimeter and perimeter-based diameter of the conduit, respectively.

The diffusion equation reads:

$$\frac{\partial^2 \theta_r}{\partial X^2} + \frac{\partial^2 \theta_r}{\partial Y^2} = \frac{\partial \theta_r}{\partial t^*} \quad (A2)$$

The Newton's law of cooling yields the boundary condition at the conduit wall:

$$(-\vec{\nabla} \theta_r) \cdot \vec{n} = Bi(\theta_r - \theta_a) \quad \text{with } \theta_a = \sin(2\pi t^*) \quad (A3)$$

where the Biot number $Bi = \frac{h_t L_d}{k_r}$ is the dimensionless heat transfer coefficient. $\vec{n}$ is the outward unit normal vector. From air mass flowrate measurements and correlations in Appendix B, the maximum heat transfer coefficient in Longeague is of the order of $20 \text{ W.m}^{-2}.\text{K}^{-1}$. Therefore, Bi can be as high as 3 for the daily fluctuations ($\tau=1$ day and $L_d = 0.32$ m) and 50 for the annual fluctuations ($\tau=1$ year and $L_d = 6.1$ m). The computational domain is sufficiently large to make the external boundaries adiabatic. The simulation time is increased until the periodic regime is reached.

Fig. A2 shows the amplitude of the local temperature fluctuation along the conduit circumference. The 3D shape of the triangular and real cross sections results in non-uniform fluctuations, lower in the corners and larger in the tips. However, the amplitudes averaged over the conduit circumference are close from each other (less than 4% difference). Therefore, if we just need the mean temperatures, the circular cross-section (a) with the same perimeter as the 3D shapes (b) and (c) yields satisfying results.

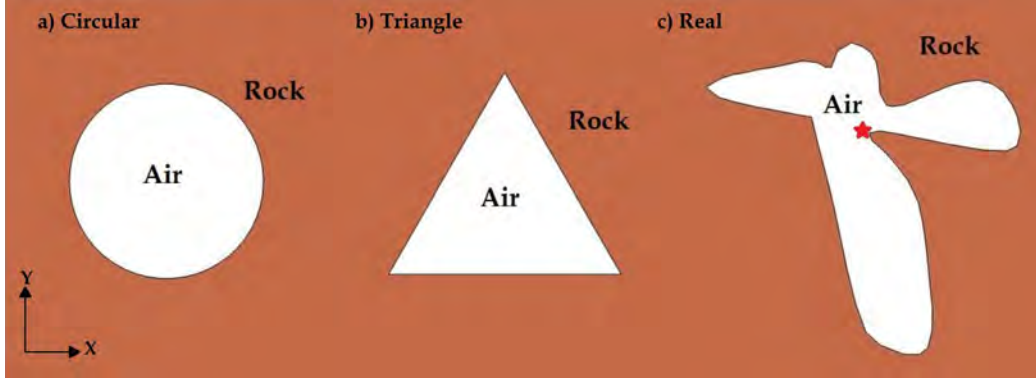


Fig A1: Three different conduit shapes (the figures are not scaled). Cross-section (c) was surveyed at 47.7 m from the entrance of D7.1 Cave located in Sieben Hengste, Switzerland (Sedaghatkish et al., 2024).

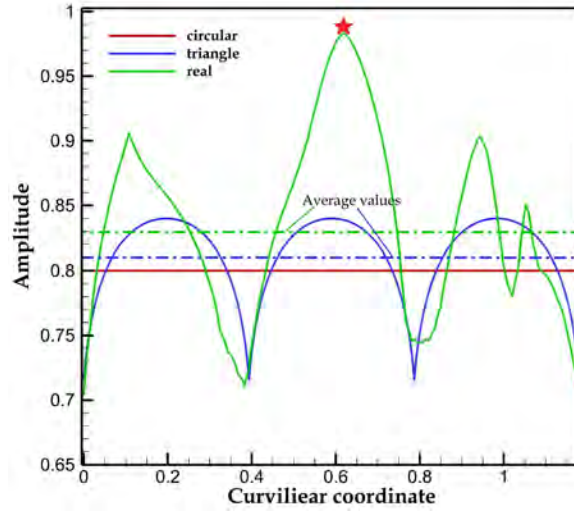


Fig A2. Amplitude of the temperature fluctuations along the conduit circumference (solid lines) and corresponding averaged values (dashed-dotted lines), for $Bi = 15.9$ and $D_p^* = P^*/\pi = 0.378$. For case (b), the origin of the curvilinear coordinate is taken in a corner of the triangle. For case (c), the position of the maximum is indicated by a star in Fig. A1c.

We performed a systematic parametric study varying the Biot number Bi and the perimeter-based diameter D_p^* . The amplitude of the wall temperature variations is displayed in Fig. A3. The maximum difference between circular and 3D cross sections is less than 20 %. We are also interested in the heat flux through the conduit wall integrated over the conduit perimeter $\Phi = \int_p Bi(\theta_r - \theta_a) d\hat{l}$ (where $\hat{l}$ is the curvilinear coordinate over the conduit circumference). Fig. A4 compares the values obtained for the three cross-sectional shapes. The difference between the circular cross section and the other ones increases with Biot number. It never exceeds 25%.

This simple study suggests that substituting a conduit with a circular cross-sectional shape to the real 3D geometry results in a reasonable loss of accuracy, if we content ourselves with the temperatures and heat fluxes averaged over the conduit circumference.

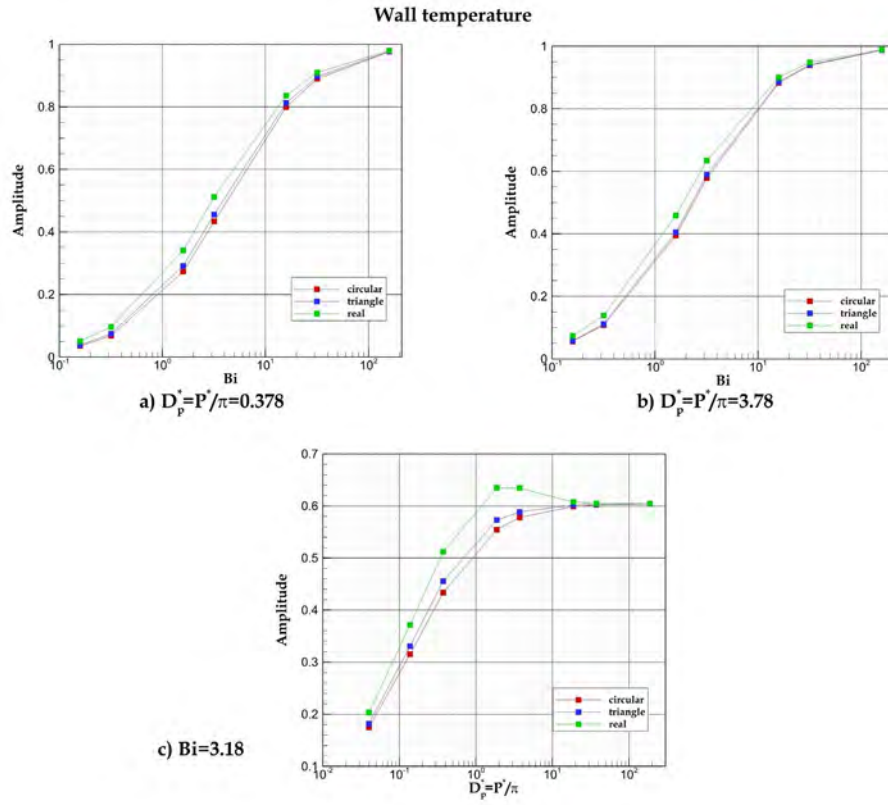


Fig A3. Amplitude of the temperature fluctuations averaged over the conduit perimeter, for different values of the perimeter-based diameter $D_p^* = P^*/\pi$ and the Biot number Bi .

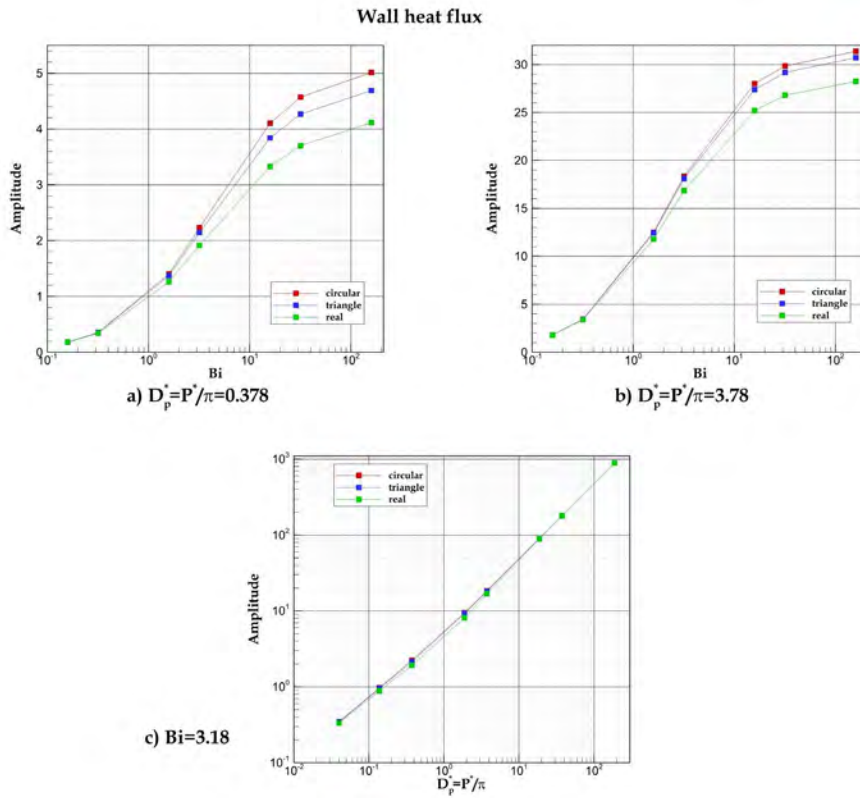


Fig A4. Fluctuation amplitude of the wall heat flux integrated over the conduit circumference, for different values of the perimeter-based diameter $D_p^* = P^*/\pi$ and the Biot number Bi .

Appendix B. Estimation of heat and mass transfer coefficients at the conduit wall

In the turbulent regime, the heat transfer coefficient is obtained from correlations between the Reynolds (Eq. 17), Prandtl and Nusselt numbers, respectively defined as:

$$Pr = \frac{c_{p,a}\mu_a}{k_a} \quad \text{and} \quad Nu = \frac{h_t D_h}{k_a} \quad (B1)$$

Nu and Re are based on the hydraulic diameter D_h to account for the specific shape of the conduit. The Gnielinski correlation (Gnielinski, 1975) has the advantage to take into account the wall roughness. This correlation reads:

$$Nu = \frac{\left(\frac{f_d}{8}\right)(Re - 1000)Pr}{1 + 12.7\left(\frac{f_d}{8}\right)^{0.5}(Pr^{2/3} - 1)} \quad \text{for } Re \geq 4000 \quad (B2)$$

The Darcy friction factor f_d , which depends on the wall relative roughness ε and the Reynolds number, was estimated from the Haaland correlation (Haaland, 1983):

$$\frac{1}{\sqrt{f_d}} = -1.8 \log \left[\left(\frac{\varepsilon}{3.7} \right)^{1.11} + \frac{6.9}{Re} \right] \quad (B3)$$

In the laminar regime, the Nusselt number is equal to a constant that depends on the conduit shape. For a circular cross-section, we get:

$$Nu = 3.66 \quad Re \leq 2000 \quad (B4)$$

We did not try to adapt the value of the constant to the actual shape of the conduit. Indeed, the heat transfer coefficient in the laminar regime is small. It is thus not expected to have significant effect on the results.

In all simulations, the Prandtl number was set to $Pr = 0.71$ (from Table 9.3), and the wall relative roughness to $\varepsilon = 0.01$. The resulting function $Nu = f(Re)$ is displayed in Fig. B1. A “smooth” transition between the laminar and turbulent regimes was achieved using the step function of Comsol Multiphysics for $2000 < Re < 4000$.

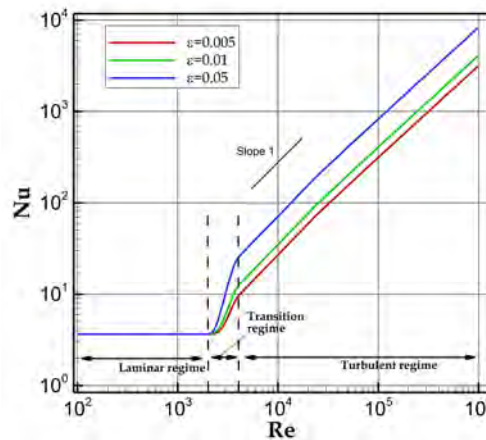


Fig B1. Nusselt number as a function of the Reynolds number for three different relative wall roughness and $Pr = 0.71$.

Using the Lewis analogy (Cussler and Cussler, 2009), the mass transfer coefficient h_m is obtained by substituting the Sherwood number $Sh = \frac{h_m D_h}{D_w}$ and the Schmidt number $Sc = \frac{\nu}{D_w}$ for Nu and Pr in Eqs. (B2) and (B4). It is worth noting that gases verify $Sc \approx Pr$ (from Table 9.3, $Sc \approx 0.59$ and $Pr \approx 0.71$). The Sherwood number Sh used in the simulations is thus close to the Nusselt number Nu displayed in Fig. B1.

A simple relation between transfer coefficients on one hand and the air flowrate and the conduit geometry on the other hand can be easily deduced from Fig. B1. This figure shows that, in the turbulent regime (i.e., for $Re \geq 4000$) and for a rough tube, the Nusselt number is approximately proportional to the Reynolds number (these results can also be deduced from Eqs. (B2-3)). The definitions of the Nusselt and Reynolds numbers yields:

$$h_t \propto \frac{\dot{m}}{P D_h} \quad (\text{B.5})$$

The same relation can be inferred for h_m using the Lewis analogy. The heat and mass transfer coefficients are thus proportional to the air flowrate and inversely proportional to the product of the perimeter by the hydraulic diameter.

Appendix C. Solver Stabilization

The Galerkin method used in Comsol Multiphysics is numerically unstable for the hyperbolic partial differential Eqs. (7) and (11). Classically, stabilization is achieved by adding in the equations a minor amount of artificial diffusion.

The general 1D advection-diffusion equation for the arbitrary variable S reads:

$$\frac{\partial S}{\partial t} + u \frac{\partial S}{\partial x} = c \frac{\partial^2 S}{\partial x^2} + F \quad (\text{C1})$$

u and c are the velocity and diffusivity, respectively. F is a source term. It has been mathematically proven that the Galerkin method is numerically stable if the cell Peclet number Pe is lower than a critical value Pe_c (Kuzmin and Hämäläinen, 2015):

$$Pe = \frac{|u|h}{c} < Pe_c \quad (\text{C2})$$

where h is the mesh element size and $Pe_c = 2$. Obviously, Eqs. (7) and (11) cannot satisfy Eq. (C2) since, in these equations, $c = 0$ results in $Pe \rightarrow \infty$.

This problem is solved by adding in Eqs. (7) and (11) an artificial diffusion term (equivalent to the first term in the RHS of Eq. C1), with a variable diffusivity set to satisfy the condition in Eq. (C2):

$$c = \max \left(c_{min}, \frac{u h}{Pe_c} \right) \quad (\text{C3})$$

where c_{min} is a lower bound of the virtual diffusivity, set to the air thermal diffusivity in Eq. (7) and the water vapor diffusivity in Eq. (11).

Eq. (C3) insures the numerical stability of the Galerkin method, but we must check that the artificial diffusion introduced in the equations does not significantly affect the results. The virtual diffusivity c is a decreasing function of the mesh size. The final test for the regularization validity consists in reducing the mesh size until it has no significant effect on the results.

Appendix D. Impact of the air flowrate uncertainty on the model outputs

Figs. D1a to D1c display the impact of the air flowrate uncertainty on the conduit wall and air temperatures obtained from HA and DA models. We assume that the actual air flowrate is in the range from the measured value to this value multiplied by 1.5.

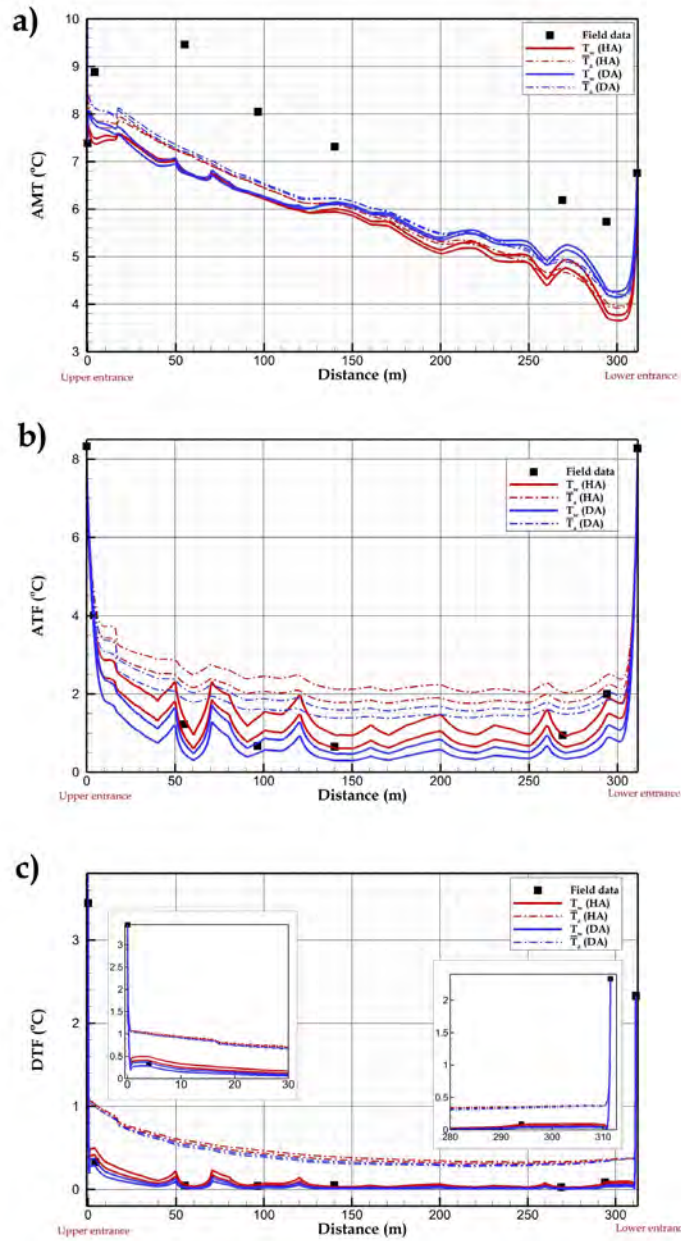


Fig D1. Impact of the air flowrate uncertainty on the model outputs, (a) AMT, (b) ATF, (c) DTF. The solid and dash-dotted lines delineate the uncertainty ranges for the conduit wall and the air temperatures, respectively.

CRediT authorship contribution statement

Amir Sedaghatkish: Conceptualization, Investigation, Methodology, Software, Visualization, Writing – original draft. **Claudio Pastore:** Data curation, Methodology, Investigation. **Frédéric Doumenc:** Conceptualization, Methodology, Writing – review & editing, Validation, Supervision. **Pierre-Yves Jeannin:** Project administration, Supervision, Writing – review & editing. **Marc Luetscher:** Conceptualization, Supervision, Writing – review & editing.

Declaration of competing interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

Data availability

The Comsol Multiphysics file for HA scenario (humid air with finite heat transfer) is available in <https://doi.org/10.5281/zenodo.14535340> associated with (Sedaghatkish, 2024) but a license of the software is necessary to use it.

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Chapter 10. Thermal length in hydrologically active caves: insights from the Rouge Eau swallow hole (unpublished paper)

In this chapter, we aim to qualitatively extend the concepts of convection length to two hydrologically active caves. Therefore, the discussion of the results aims to give the orders of magnitude of water convection lengths for two caves with punctual infiltration (sinkholes), which also show inflows of water with different orders of magnitude.

10.1. Introduction

The concept of convection length applies not only to air but also to water flowing through vadose and phreatic karst pathways (Covington et al., 2012) before reaching springs. Throughout the year, underground rivers and karst springs usually show a relatively stable water temperature (when compared to external rivers) close to the air AMT at the spring's elevation (Covington et al., 2011). However, short-term variations occur and are often linked to flood events (Luo et al., 2023). As for the heat advected by air, forced water convection plays a critical role in the water's energy balance (Covington et al., 2011; Luhmann et al., 2011), and, since air and water have different thermal properties, the thermal lengths of these two fluids also differ. Looking at the thermal behaviour and temperature fluctuations of karst springs (and underground rivers) can be a significant indicator of the climatic conditions of the karst systems. Indeed, beyond the convection length (i.e. in the homothermic zone or deep karst) and owing to the greater heat capacity of water than that of air, it has been observed that the temperature of cave air is close to that of underground rivers (Luetscher and Jeannin, 2004; Badino, 2004), with air temperature following water temperature fluctuations (e.g. Skoglund et al., 2023). Therefore, in hydrologically active caves (i.e. with flowing water), the climatic conditions therein result from the thermal interaction between air, water and rock (Badino, 2004).

Manga (2001) states that water temperature is a valuable tool for gaining information on the karst system properties, such as conduit geometry, recharge mode, flow velocity, and residence time, provided that convective heat exchange remains significant. Heat is regarded as a non-conservative tracer (Luhmann, 2011; Anderson, 2005) and reveals system characteristics through temperature peaks and their attenuation. Understanding how thermal pulses move through the karst and the processes governing heat exchange is, therefore, essential for inferring system properties (Gabrovšek and Peric, 2006; Genthon et al., 2005; Birk et al., 2004; Liedl et al., 1998).

We call thermographs the graphs showing variations of the water temperature measured at a spring or along a cave stream. Thermographs reflect the combined influence of heat exchange processes, depending on local conditions, as outlined in the existing literature (Luo et al., 2023; Skoglund et al., 2023; Luhmann et al., 2015; Covington et al., 2012, 2011). The primary cause of temperature changes (i.e. temperature peaks) at springs during flood events is the increased water flow through the system, which modifies the convection length, i.e. the distance necessary for a thermal signal to reach 90% attenuation. If this distance is larger than the conduit length, variations of input temperature will be visible at the spring. During flood events, the convection length increases due to a higher water flow rate, leading to visible temperature variations at the spring. In fact, factors driving changes in the convection length include variations in the water flow rate and residence time, temperature difference between newly infiltrating water and the system, distinct pathways with different geometries during low and high water periods and conduit diameters. In other words, between low and high-flow periods,

there can be a variation in heat transfer efficiency along the system (Luhmann et al., 2011) that is depicted by temperature peaks (positive or negative) in thermographs.

Additionally, how the water inrush infiltrates into the karst is a key factor in determining the convection length (Luo et al., 2023b; Luhmann et al., 2011; O'Driscoll and DeWalle, 2006). Diffuse (or distributed) recharge passes through less permeable components of the aquifer, such as soil layers, epikarst, rock matrix, and narrow fractures, before reaching the conduit network. Diffuse recharge allows a more efficient heat exchange between water and cave systems and tends to dampen temperature peaks. In contrast, localised recharge (i.e. swallow holes) tends to flow directly into the conduit network, which has limited thermal efficiency in exchanging heat (Luhmann et al., 2011). For instance, a stream entering a ponor at 18°C in a region with an AMT of 10°C may take hundreds or even thousands of meters underground to cool down to ~10°C. In contrast, rainwater infiltrating through soil and epikarst achieves equilibrium much faster, often between a few to tens of meters (see Chapter 5 in Sedaghatkish, 2025). Therefore, the balance between localised and distributed recharge plays a crucial role in determining the nature of the thermal length.

For the case of a concentrated recharge, we monitored an active swallow hole (Gouffre de la Rouge-Eau) to give insight into temperature variations along a hydrologically active cave showing a low to medium water flow rate. Our results are qualitatively compared with the Škocjan-Kačna Cave system (Covington et al., 2012; Gabrovšek and Peric, 2006), which, in contrast, has a much higher flow rate (~8 m³/s on average, >300m³/s as maximum measured flow rate). These springs are used as examples to show how the convection length varies according to the water flow conditions.

10.2. Site description

The *Gouffre de la Rouge-Eau* is located in the municipality of Saicourt (Bern), in the Swiss region of the Jura Bernois. This cave is an active swallow hole, which is fed by a perennial sinking creek, whose water flow rate is comprised between ~0.3 and 1 m³/s (Piquerez, 2014). This creek drains water from the nearby valley (*plaine de Bellelay*), which is mainly filled with peat. It enters the karst through the Rouge Eau swallet, at the contact between the impermeable sediments and limestone. The cave drains toward the Blanche Fontaine karst spring, and is mainly vertical, with a depth of -134 m and a total development of ~1.2 km. An excerpt of the topography from Maire et al. (2022) is given in Figure 10.1. Cave ventilation was not directly investigated; however, we observed that air movements are clearly perceptible along the cave, especially at the base of the 30 m entrance shaft (Figure 10.1) where a waterfall is present. Ventilation is not related to a chimney effect (only one entrance). Instead, air movements are related to air convection due to water fall and to water drag (*cfr.* Figure 1.2.e and Figure 1.2.f in Chapter 1.2.1).

10.3. Monitoring stations

The monitoring lasted from 13/05/2022 to 25/05/2023, and water and air temperatures in the cave were measured at the same locations (Figure 10.1). The water temperature was measured with Reefnet Sensus Ultra devices (accuracy ±0.3 °C, resolution 0.01°C; Chapter 2.2.2). The cave air temperature was measured with the Hobo U23-004 Pro v2 devices equipped with a wired TMCx-HD sensor (accuracy of ±0.1°C, resolution of 0.02°C; Chapter 2.2.2). The data were recorded at an hourly time step. The water sensors were placed in the cave, under the water flow at -30 m (T₋₃₀) and -60 m (T₋₆₀) depth from the entrance, while air temperature was measured at the same stations about ten meters away from the flowing water. All water probes were installed within a pierced pipe as protection from accidental damage. External temperature was measured with a Reefnet Sensus Ultra about 100 m from the cave entrance (T_{ext}). The last Reefnet sensor was placed immersed in the water of the external creek (T_{in}) shortly before it was swallowed into the cave.

A direct measurement of the discharge rate of the external creek is not available. Therefore, the Sorne River, a nearby external river partially fed by the karst aquifers to which the Rouge Eau belongs (Weber and Linder, 2004), is here used as an indicator of discharge variations.

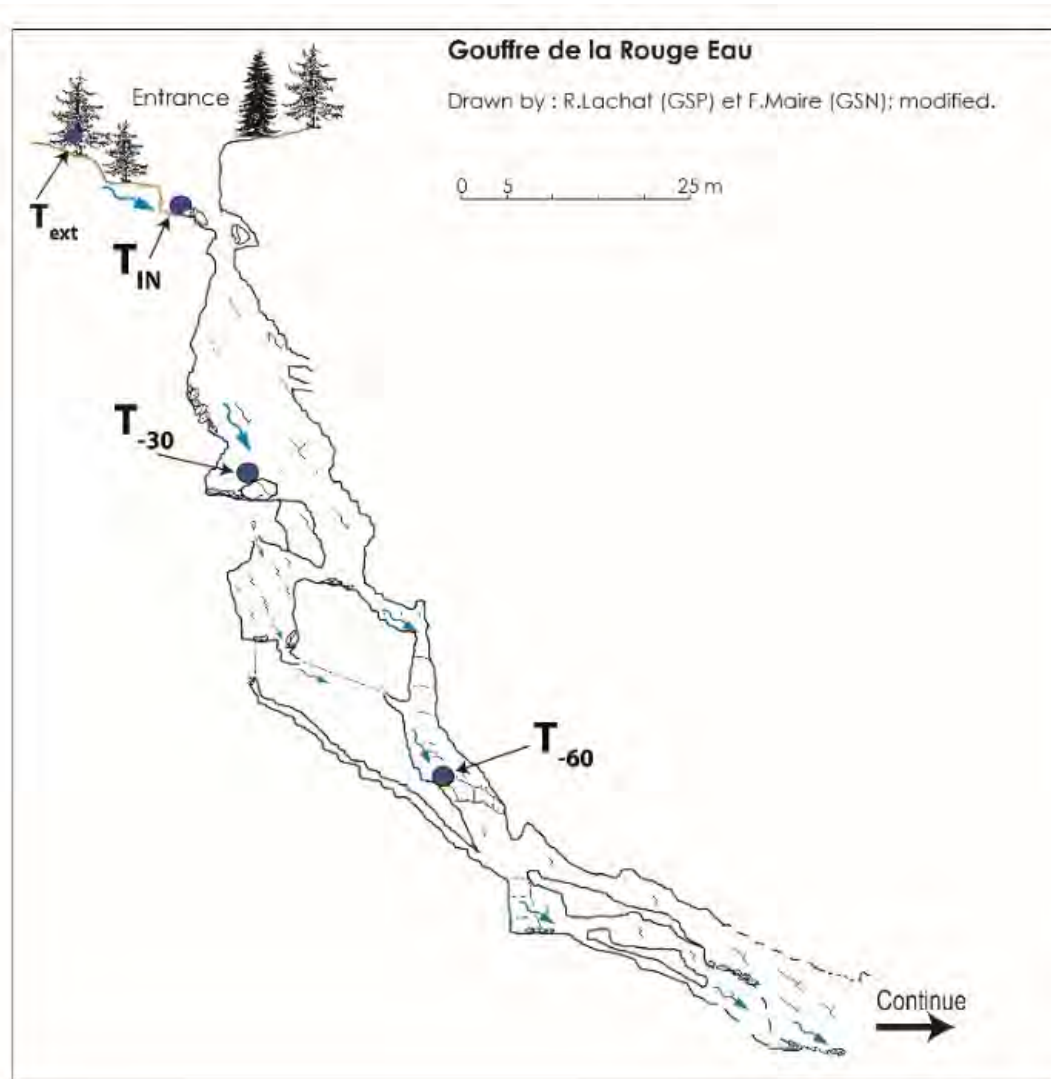


Figure 10.1 – An excerpt from the Rouge Eau cave survey (modified after Maire et al., 2022) with the position of the probes.

10.4. Results

Figure 10.2.a shows the temperature time series recorded at the monitoring stations (including the atmospheric external temperature close to the cave entrance, Figure 10.2.b) as well as the Sorne River discharge (dashed line in Figure 10.2.a).

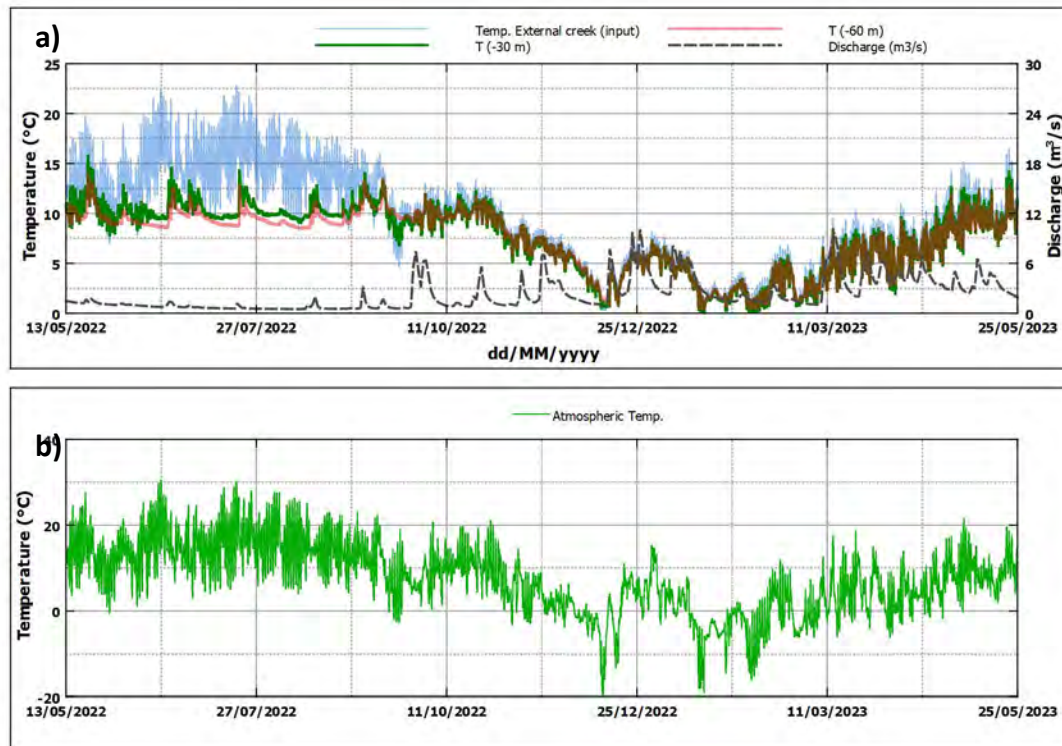


Figure 10.2 – a) Water temperature of the external creek (input temp., blue), at -30 m (green) and -60 m depth (red) in the cave. The discharge of the Sorne River is also depicted. b) Atmospheric temperature recorded near the cave entrance.

From May 2022, the discharge of the Sorne River indicates the beginning of a low water stage that lasted until the beginning of September 2022. During the summer of 2022, the river discharge stayed low ($<1 \text{ m}^3/\text{s}$). For this period, small discharge peaks were visible, likely related to short storm events, which caused an increase in the Sorne's discharge not larger than $3 \text{ m}^3/\text{s}$. From mid-September 2022, the Sorne's discharge increased, and clear flow peaks are visible. Until the end of the monitoring period (May 2023), the Sorne's discharge stayed between 2 and $3 \text{ m}^3/\text{s}$, and flow peaks reached up to $\sim 12 \text{ m}^3/\text{s}$. Accordingly, the Rouge Eau sinking creek also followed the same fluctuations, with low water flow from May to mid-September 2022 and high water flow for the following months.

Two different evolutions of water temperature fluctuations along the monitored cave transect can be observed in Figure 10.2.a:

- In May 2022 (Figure 10.3), the daily fluctuations of water temperature recorded in the external creek were transmitted at the first station in the cave (T_{-30}), although they were reduced in amplitude. In the same period, at T_{-60} the daily fluctuations showed a complete damping.

From June to mid-September 2022 (Figure 10.3), the daily fluctuations were largely damped for both T_{-30} and T_{-60} . Yet, during the period of low water stages, temperature peaks at T_{-30} and T_{-60} are well visible in the graphs and correlate with small flood events revealed by peaks in the water discharge. Temperature peaks indicate a rise in water temperature of up to 5°C at T_{-30} and 4°C at T_{-60} .

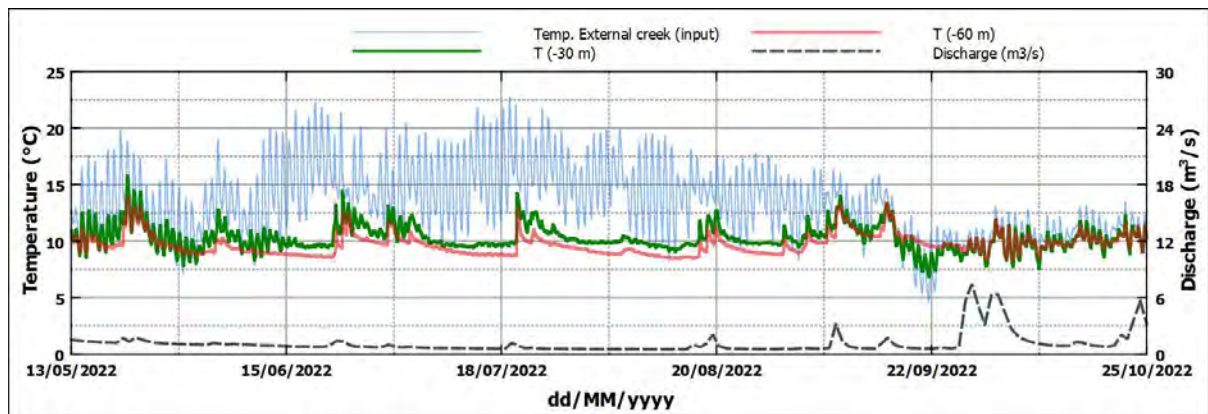


Figure 10.3 – A focus of Figure 10.2.a in the period 13 May to 25 October 2022. The discharge of the Sorne river is also depicted.

- During wet periods and high water stages (Figure 10.4, from the end of September 2022 until May 2023), the daily fluctuations of the water temperature recorded in the external river are transmitted at both T₋₃₀ and T₋₆₀ stations. The external fluctuations of this period (cold season) are smaller than those of the previous one (warm season); however, these did not seem to be dampened. The cave water temperature still shows daily fluctuations, and internal and external temperatures are close to each other. Moreover, although the flood events are larger in this period, temperature peaks related to these events are neither visible outside nor inside the cave.

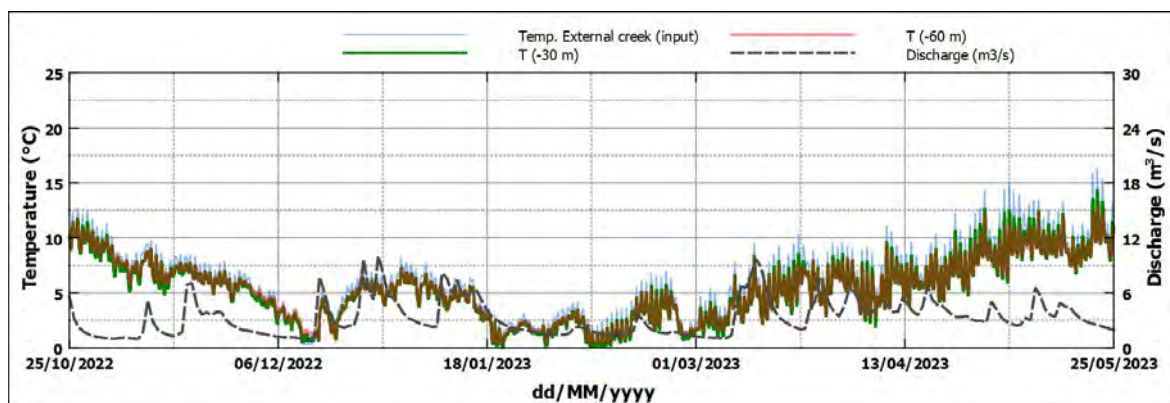


Figure 10.4 – A focus of Figure 10.2.a in the period from 25 October 2022 to May 2023. The discharge of the Sorne river is also depicted.

Resuming, the data show that during spring and summer with low flow (Figure 10.3), there is a strong attenuation of daily temperature fluctuations with increasing distance into the cave. Flood events during summer determine marked temperature peaks, which are transmitted deep in the cave. In contrast, during autumn and winter (Figure 10.4), which show reduced fluctuations of input water temperature and an increased flow rate, daily fluctuations are recorded at both stations in the cave, showing almost no attenuation. Yet, temperature peaks related to flood events are hidden by daily fluctuations.

10.4.1. Comparison between air and water temperatures

In the graphs shown in Figure 10.5.a/b, water and air temperatures at the same monitoring locations are compared (T₋₃₀ in Figure 10.5.a and T₋₆₀ in Figure 10.5.b). Both graphs show that air temperature mimics water temperature. In the graph of station T₋₃₀ (Figure 10.5.a), the difference between air and water temperatures ($\Delta T = \text{air temp.} - \text{water temp.}$) oscillates mostly in the range of $\pm 2.5^\circ\text{C}$, with larger ΔT peaks (max $\Delta T = 3.8^\circ\text{C}$, 11/12/2022) taking place during low water periods (e.g. dashed boxes in

Figure 10.5.a/b) or during small floods in summer. On the other hand, at station T₋₆₀ (Figure 10.5.b), the ΔT between water and air decreases close to 0°C at the end of spring and summer 2022 (from May to September 2022). At this station, large ΔT peaks (max $\Delta T=2.6^{\circ}\text{C}$, 11/12/2022) are always visible during small floods in summer and during low water in winter (dashed boxes in Figure 10.5.b). Finally, at both stations during autumn 2022 and spring 2023, water shows a high flow rate and large daily fluctuations of temperature, and the ΔT oscillates with positive and negative values.

At both stations, during summer and low water periods (from May to September 2022), air temperature is constantly colder than the water ($-\Delta T$), whereas in winter and low water discharge (December 2022 and February 2023, dashed boxes in Figure 10.5.a/b), water temperature is lower than that of air ($+\Delta T$). The two dashed boxes in Figure 10.5.a/b highlight periods when the ΔT is largely positive ($>1.5^{\circ}\text{C}$ with air temp. $>$ water temp., water slightly warmer than 0°C). The maximum ΔT between water and air was reached in December 2022, for both stations, at the apex of winter's continuous decrease in external temperature (lowest ext. temp. $\sim -20^{\circ}\text{C}$ on 12/12/2022, Figure 10.2.a) from November to December 2022 (dashed left boxes in Figure 10.5.a/b), simultaneous with decreasing water discharge. This ΔT peak is followed by a period with a $\Delta T < 1^{\circ}\text{C}$, while water (and external) temperature, as well as the discharge, increased again. Afterwards, another temperature and discharge drop led to a $\Delta T > 1.5^{\circ}\text{C}$ (dashed right boxes in Figure 10.5.a/b)

These results suggest that air cave temperature is controlled by water flow rates and temperature.

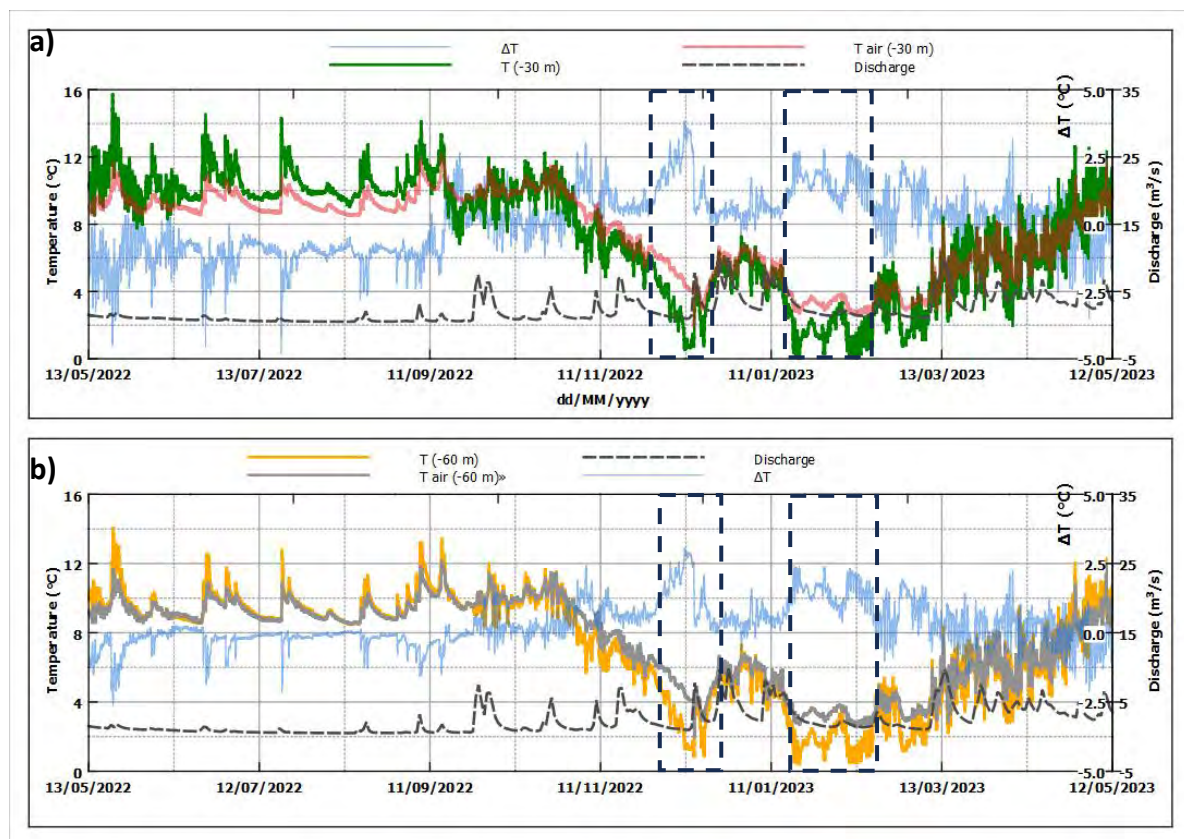


Figure 10.5 – Comparison between air and water temperatures at monitoring stations T₋₃₀ at T₋₆₀. ΔT is air Temp. – water Temp. The discharge of the Sorne river is also depicted. Dashed boxes highlight periods of maximum ΔT (air temp. $>$ water temp.).

10.5. Discussion

10.5.1. Water convection length

We show thermographs recorded along the Rouge Eau cave (Figure 10.6.a) and the Škocjan-Kačna cave system (Figure 10.6.b; Covington et al., 2012; Gabrovšek and Peric, 2006). These caves are active swallow holes with a concentrated recharge related to perennial sinking streams, but with large differences in terms of water discharge. The Rouge Eau Cave is fed by a perennial creek with a discharge between a few litres per second and $\sim 0.3 \text{ m}^3/\text{s}$, and occasionally up to $1 \text{ m}^3/\text{s}$ during the snow melt (Piquerez, 2014). In Figure 10.6.a, T_{in} , T_{-30} and T_{-60} are the temperatures measured in the external river before entering the cave (input temperature), at -30 and -60 m inside the cave below the water entrance, respectively. The Reka River is the main allogenic input for the aquifer in between the Škocjan and Kačna caves, before passing through other caves and reaching the Timavo spring (average discharge $\sim 8.3 \text{ m}^3/\text{s}$, maximum measured discharge $>300 \text{ m}^3/\text{s}$; Gabrovšek and Peric, 2006). In Figure 10.6.b, the Škocjan and Kačna temperatures represent, respectively, the upstream and downstream groundwater temperatures, about 2 km apart. Given that the convection length depends on the flow rate and that discharges of the two systems (Rouge Eau and Škocjan-Kačna) differ roughly by an order of magnitude, we can consider them as two end members for a concentrated recharge.

Since rivers represent the main source of water for both systems, the contribution of diffuse recharge can be neglected along the parts of the systems under consideration. For both Škocjan-Kačna and Rouge Eau caves, the input water temperatures are, therefore, strictly related to the sinking river temperatures which are controlled by the external atmospheric conditions (Covington et al., 2012).

The graph in Figure 10.6.b shows that during low water periods (from 14 to about 24 March 2005), temperature fluctuations are well visible for Škocjan Cave while they are damped when reaching Kačna Cave. Conversely, the temperature fluctuations are well visible on both temperature time series during high water periods (after 24 March 2005). Covington et al., (2012) and Gabrovšek and Peric (2006) suggest that, depending on the flow conditions in the Reka River, diurnal temperature pulses are either transmitted or strongly damped as they pass from Škocjan Cave to Kačna Cave (Figure 10.6.b).

During low flows (identified by the low stage of the dotted grey line), the diurnal oscillations are damped by the heat exchange between water and rock. During storms and high input discharge (high stage peak in Figure 10.6.b), the diurnal variations pass through the system with a very low, or even negligible, damping. According to the authors, the water travel time varies with the discharge rate. They suggest that, when the flow rate is low, the duration of the temperature oscillations is less than the travel time through the system and signals will be strongly damped. In contrast, during storms, the travel time drops to less than a day and the temperature signals are transmitted far in the system, indicating an extension of the convection length. For this system, the convection length can be reasonably estimated in the order of some hundred meters during low-flow conditions, extending to kilometres during flood events.

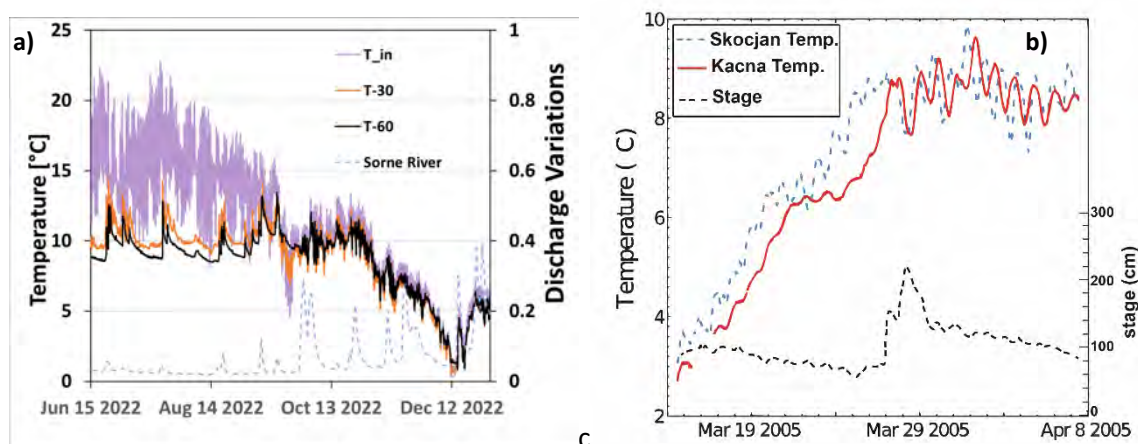


Figure 10.6 – a) Thermographs recorded in the external creek feeding the Rouge Eau cave, as well as at 30 m (T_{-30}) and 60 m (T_{-60}) below the cave entrance. Relative discharge variations (Q/Q_{max} , where Q is the discharge) of the Sorne River. It is a nearby external river used as an indicator of discharge variations of the creek swallowed by the Rouge-Eau cave. b) Water temperatures measured at Škocjan and Kačna caves as well as the water level measured in the Reka River (Covington et al., 2012 after Gabrovšek and Peric, 2006; modified).

The same behaviour can be observed at the Rouge Eau. During summer (from June to mid-September 2022, Figure 10.6.a). The outside water temperature was much higher than the cave temperature and daily temperature fluctuations in the river were large (5 to 10°C). The flow rate of the external creek was low as indicated by the Sorne's discharge. The cave water temperatures did not show daily fluctuations, although the difference between the water and cave temperature was high, but temperature peaks are well marked when even small flood events took place (identified by the small peaks in the discharge of the Sorne River). When the flow rate is low, water has time to equilibrate with the rock and temperatures in the cave are stable. The convection length is therefore short, in the order of a few tens of meters (~30 m) from the cave entrance. In contrast, during flood events in summer and high external temperatures, the increased flow rate transports a significant amount of water with a very different temperature from the system. Therefore, temperature signals are not damped (temperature peaks) in the monitored region of the cave. The increased flow rate prevents the warm water from reaching thermal equilibrium with the rock within the first tens of meters, indicating that the convection length expanded.

During autumn and winter (from mid-September to December 2022, Figure 10.6.b), although the daily input temperature fluctuations are smaller than in summer, they remain clearly visible inside the Rouge Eau cave, with minimal damping from outside to inside the cave. The presence of daily fluctuations in the cave is likely due to 1) input water temperature close to that of the conduit walls (Covington et al., 2011), which does not promote heat exchanges, and 2) the increased water discharge of the sinking creek in the wet seasons, such as winter and autumn. The first condition, such as a reduced heat exchange, implies larger temperature oscillations within the conduits, while the second condition promotes an extension of the convection length. Additionally, in winter, temperature peaks related to the flood events are “hidden” by the daily fluctuations, because during these events, the entering water has a temperature close to that of the system. In general, the water convection length of the Rouge Eau cave can be estimated in the order of some tens of metres for low-flow conditions, while it extends to several hundreds of metres, or even to a few kilometres, during high water periods.

Therefore, the two considered systems have similar behaviour in damping thermal fluctuations during low and high flow periods, but we remark that the input water discharge rates differ by one to two orders of magnitude. Accordingly, the convection lengths between them also differ by the same orders of magnitude.

10.5.2. Air and water temperatures

The comparison between air and water temperatures is meaningful when one wants to investigate microclimatic conditions in caves presenting turbulent water flow, such as active swallow holes and underground rivers. Previous studies have shown that along the homothermic zone, water and air are almost in equilibrium. Indeed, Luetscher and Jeannin (2004) observed that, in the homothermic zone of vertical systems with free-flowing water, the measured air and temperature differ by an order of magnitude of a few tenths of degrees, with water usually slightly colder than air (Luetscher and Jeannin, 2004). Other authors have reported consistent observations, including Badino (2004; 2018) and Skoglund et al. (2023). In the homothermic zone, observed vertical temperature gradients of air are included between water ($0.234^{\circ}\text{C}/100\text{ m}$) and humid air ($0.55^{\circ}\text{C}/100\text{ m}$) lapse rates, being closer to that of water if water flow > air flow (Luetscher and Jeannin, 2004). This suggests that, based on the relative importance of circulating turbulent volumes of water and air, and thanks to the greater thermal inertia of water compared to that of air (Badino 2004), water controls air temperature and cave climatic conditions due to the heat exchanges between air and water (and rock). However, previous authors (Luetscher and Jeannin, 2004) were referring to the homothermic zones, while in the previous paragraphs, we observed and described that the monitoring stations along Rouge Eau are within the water convection (heterothermic) zone. Nonetheless, our results are in agreement with observations reported in the bibliography (e.g. Badino, 2004, 2018; Luetscher and Jeannin, 2004; Skoglund et al., 2023) since water and air temperatures of Rouge Eau cave appear linked. We recall that, although ventilation of this cave has not directly been investigated, we observed that air movements are mainly related to air convection due to water fall and to water drag (no chimney effect; *cfr.* Figure 1.2.e and Figure 1.2.f in Chapter 1.2.1). Therefore, air ventilation is clearly perceptible, but it might be localised to convective cells (for example, close to the waterfall at station T₋₃₀ in Figure 10.1). Although stations T₋₃₀ and T₋₆₀ are reasonably within the water convection length (visible daily and event-related fluctuations), the difference in temperature between air and water is in the order of a few degrees at the -30 m depth and closely matches at the second station at -60 m. At -30 m, in summer, daily fluctuations of air temperature are less visible than those of water temperature, while temperature peaks related to very small flood events are reproduced by air temperature with a smaller amplitude. Instead, always in summer and at -60 m, the two temperatures are very close to each other (ΔT almost 0°C), with ΔT peaks that are still visible during small flood events. During dry periods, the low water flow rate (low heat exchange between air and water) and the likely limited air exchange between the cave and the external atmosphere (no airflow) allow air to show damped fluctuations. Water temperature peaks in summer also produce peaks in air temperature. These peaks relate to transient and limited flooding, which, however, can bring (in the cave) enough water to perturb the air temperature. In winter, larger temperature fluctuations are visible at both stations along the cave. As previously discussed, this is likely related to the limited heat exchange between the water and the rock surrounding the cave, since in winter they may have a closer temperature than in summer. As indicated by the ΔT in Figure 10.5.a/b, during cold seasons, cave air is generally warmer than infiltrating water.

Late in spring and summer, despite the high external and water temperatures, the reduced water flow (low discharge) allowed the water to release much of the heat accumulated outside to the surroundings (air and rock) before reaching station T₋₃₀ (resulting in reduced temperature fluctuations at this station). Nevertheless, at this station, the entering water is still warmer than the air in the cave (large ΔT). Then, the two fluids are not equilibrated, and water continues to transfer heat to the air and the surroundings as it flows further into the cave. Whereas, always in summer but at -60 m, because of the limited flow velocity and larger contact time between water and both rock and air, water has already released part of the external heat (to rock and air), therefore, air and water are closer to equilibrium (small ΔT). Transient and small flood events in summer are able to perturb air temperature, but these events do not last long enough to allow air to equilibrate with water (ΔT peaks). Whereas, during winter, ΔT is

strictly positive when the external temperature suddenly drops and water flow is low (dashed rectangles in Figure 10.5.a/b). We can suppose that, although cave air temperature partially follows the water temperature drop, the simultaneous low flow rate limits the amount of heat exchanged per unit time, so that the largest ΔT occurs when water temperature is at its minimum and flow rate is low, because the reduced water volume cannot immediately equilibrate the air temperature. In contrast, but supporting this interpretation, during the period between the two dashed boxes in Figure 10.5.a/b, the external temperature rises as well as the water flow rate, along with the temperature difference between cave air and water decreases ($\Delta T \approx 0^\circ\text{C}$), indicating that the air temperature has equilibrated with the water. In this case, the high flow rate favours a larger amount of heat exchanged per unit time (i.e., more energy is transferred per second). During spring 2023, large external temperature fluctuations are transported into the cave by the increased water flow rate. Despite this, the ΔT between cave air and water oscillates around 0°C , within approximately $\pm 1^\circ\text{C}$, indicating that air and water are close to thermal equilibrium. However, the presence of residual fluctuations (in the cave) suggests that the higher water velocity reduces the time available for the water to fully transfer heat to the surrounding rock, and may also be influenced by relatively close temperatures between entering water (during spring) and the cave system, which can limit further heat exchange.

Therefore, we can conclude that:

- Cave air temperature follows water temperature oscillations in the presence of turbulent water flow.
- The temperature difference (ΔT) between air and water is mainly negative (air temp. < water temp.) during summer and low water discharge.
- The ΔT is mainly positive (air temp. > water temp.) during low water discharge in winter.
- The ΔT oscillates around negative and positive values during high discharge, especially in autumn and spring.
- Thus, the ΔT between cave air and water is maximum at low water flow rates, because the reduced water flow limits the rate of heat transfer per unit time (although overall water has more time to equilibrate with the surroundings). Instead, ΔT is closer to 0°C at high flow rates, indicating that air and water temperatures equilibrate more quickly.

It is worth highlighting that throughout the discussion about air and water temperatures, the role of the rock has only been touched upon. However, as discussed in the previous chapter, it should always be taken into account that the rock plays a major role in buffering temperature oscillations of both water and air. Therefore, temperature and related oscillations of water, air and rock measured in a cave are always the result of the interaction between these three components and their respective thermal properties. Further work should focus on this key point.

Therefore, for the reasons above, water and air temperatures are intimately linked and studying water temperature variation along underground rivers and at karst springs may reveal key information on climatic conditions over the vadose zone where water infiltrates.

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Chapter 11. General Discussion

This work aimed to develop an integrated monitoring and modeling approach for a better understanding of ventilation dynamics and heat exchange in caves. The main objective was to validate new theoretical models, which couple conduction and convection processes, by comparing them with direct cave measurements, mainly of airflow and temperatures. Previous studies mainly lack in a systematic comparison between numerical models and measured airflow, particular attention was given to provide reliable ventilation measurements, a parameter that is difficult to monitor in a cave, especially when long-term time series are required. To address this, Part 1 of the thesis focuses on the technical challenges of in-cave data acquisition and the solutions adopted to overcome them. In Part 2, the measured parameters are compared with numerical models and used to evaluate heat exchange processes along ventilated conduits.

In this discussion, we emphasise the importance of cave monitoring for environmental studies (Chapter 11.1) and, based on our results, we propose (Chapter 11.2) a revised sketch of the thermal conceptual model proposed by Luetscher and Jeannin (2004) (Chapter 1.2.2, Figure 1.6). We also discuss (Chapter 11.3) the role played by ventilation and convection length on paleoclimatic archives, cave thermal response to climatic changes and subterranean ecosystems. The discussion concludes (Chapter 11.4) by extending the concept of convection length to hydrologically active systems, proposing a classification based on the thermal behaviour of karst springs in relation to recharge modes and system depths.

11.1. The significance of cave monitoring

Although model elaborations are increasingly demanded to explain ventilation processes and the theory on which they are based is being investigated (e.g. Badino, 2018, 1995; Covington and Perne, 2015; Lismonde, 2002; Gabrovšek, 2023), field data are still necessary for model validation and calibration. The main discrepancies between models and real scenarios stem mostly from the assumptions and simplifications taken to build a model (Badino, 2010). This author (Badino, 2010) highlighted the gap between theoretical work and real-world scenarios, emphasising the need for dedicated experiments and cave data to either confirm or deny proposed theoretical models before applying them to real (in-cave) situations. Luetscher and Jeannin (2004) also pointed out the lack of field data to quantify the relative importance of water- and air-related heat fluxes along the karst.

The lack of systematic comparison between models and field data in early theoretical studies was primarily due to the absence of adequate technology for extensive cave monitoring (Cigna, 2002). Technologies available since the end of the 20th/beginning of the 21st Century have made it easier to obtain cave environmental data, such as long-term temperature and airflow time series, which can be used to test numerical models. Indeed, there is a lack of systematic comparisons between airflow measurements and ventilation models in the literature. In addition, some physical parameters used in theoretical models (e.g., friction factor, longitudinal dispersion, or aeraulic resistance) are inferred from theoretical equivalences or estimates derived from other fields of research rather than from the cave environment. Technological advances can now help to conduct field tests to evaluate these parameters specifically for caves.

In previous chapters (Chapter 5 – “*Dispersion of artificial tracers in ventilated caves*” and Chapter 8 – “*Dynamics of chimney-effect ventilation and heat exchange: a model from Longeigue Cave*”), we highlighted that theoretical relationships for assessing longitudinal dispersion and cave aeraulic

resistance are inadequate for cave environments. The theoretical approach significantly underestimates these physical parameters, necessitating *in situ* investigations through direct measurements. This is because both theoretical relationships rely on the Darcy friction factor (or relative roughness), which can hardly be measured or assessed for complex and natural environments.

In Chapter 5, we have demonstrated that the longitudinal dispersion coefficient can be estimated by doing gauging tests and that its assessment based on Taylor's (1954) theory is unsuitable for natural conduits or mine tunnels given their variability in geometry (changes in cross sections along the conduits) and roughness along their length. The dispersion coefficients can be, instead, inferred using Chatwin's method (Field, 1999; Chatwin, 1971) based on the analysis of the breakthrough curves, resulting from gauging tests. Overall, we have demonstrated that the 1-D dispersion model is suitable for conduits with simple geometries and relatively constant diameters, such as mine tunnels. However, it proves inadequate for more complex situations, such as caves, where more complex dispersion models are required.

If the driving pressure and cave aerodynamic resistance are known, ventilation can be modelled (Gabrovšek, 2023; Lismonde, 2002) even if airflow measurements are not available. However, in most cases, the theoretical relationship for cave resistance (R) given by Eq. 8.6 in Chapter 8 leads to unsatisfactory estimations of R due to uncertainties related to conduit/cave geometries and relative roughness. Conversely, we demonstrated that a reliable cave ventilation model can be achieved by inferring cave resistance using direct measurements of airflow and driving pressure (Chapter 8, Figure 8.10). Additionally, we showed that a better ventilation model is obtained if the cave is discretized into distinct segments, with defined temperature and geometry (including segment length, slope, and radius), rather than relying on a simplified geometry with an average temperature for the entire cave.

Conducting experiments and long-term monitoring of cave aerology present significant challenges. Many instruments are not designed for underground environments, making it necessary to adapt sensors (e.g., the SFM we used for airflow monitoring; see Chapter 4 – *“Monitor air fluxes in caves using digital flow meters”*) to the specific constraints found in caves (e.g. low temperature, condensation and absence of light). Using instruments in environments other than those for which they were designed requires thorough calibration and verification in the field to ensure their proper operation in a cave. In Chapter 4, we investigated the use of a digital airflow meter (Sensirion SFM3003-300-CE) after calibrating the device beyond the manufacturer's specified working conditions as well as assessing the uncertainties associated with airflow measurements. Beyond the technical limitations of the instruments, site characteristics are also challenging. For example, airflow is typically measured using a single instrument placed at one point within a cross-section. However, we have demonstrated that this approach can result in under- or overestimation of the airflow rate due to the instrument's position relative to streamlines with varying airspeeds, rather than the device's accuracy. This discrepancy arises primarily from uneven and variable velocity fields within a cross-section, which fluctuate with both flow rate and direction. Our findings emphasize that the placement of a single instrument within a cave conduit significantly influences the accuracy of airflow estimates. While long-term monitoring with a single device is unlikely to capture the precise “true” airflow value, it is more appropriate to determine a reasonable range of airflow values. Moreover, in Chapter 6 – *“Comparing different methods for measuring airflow”* – we compared direct airflow measurements made with digital flow meters, hand-held hotwire anemometers and CO₂ gauging tests. The comparison has shown that there are no striking differences between the results delivered by these different methods. However, we have also highlighted that it is difficult to estimate a cave airflow rate with uncertainties

lower than 20 to 30%. If a fixed anemometer is deployed close to the conduit wall, uncertainties can be as high as $\pm 50\%$. It is important to derive cave airflow values with different methods in order to assess a range of plausible airflow rates (upper and lower uncertainty bounds) that reflect actual conditions. Even temperature measurements are challenging: in cave passages with temperature variations, heat radiation makes it unclear if a temperature probe measures air or wall temperature (see Chapter 3 and Chapter 9).

Thereafter, in our papers devoted to thermal modeling (Sedaghatkish et al., 2024a in Chapter 7 and Sedaghatkish et al., 2025, in Chapter 9), data acquired from cave monitoring (airflow and temperatures) were used either as an input for the models or in comparison with model outputs. The systematic comparison between field data and model outputs allowed us to validate the numerical models we developed and identify their weaknesses.

11.2. Thermal overview of ventilated karst systems

The results of this work allow us to slightly modify the conceptual model proposed by Luetscher and Jeannin (2004) (Chapter 1.2.2, Figure 1.6). These modifications are proposed in Figure 11.1.

Main modifications of the previous global thermal model of karst involve the representation of *Ventilated Warm* and *Cold Zones* (also known as thermal anomalies; Lismonde, 2002; Wigley and Brown, 1971) along the convection zone (or heterothermic zone, i.e. where external daily or yearly fluctuations are still detectable). The distinction between the convection and deep karst zones (or homothermic zone, i.e. where external fluctuations are no more visible) is also illustrated. Figure 11.1.a schematically depicts temperature time series typically observed at the upper and lower entrances as well as in the homothermic zone. At upper entrances, daily variations can be observed during downward ventilation (when $T_{cave} < T_{ext}$), while they are damped during upward ventilation ($T_{cave} > T_{ext}$). Vice versa, during upward ventilation, daily fluctuations can be observed at lower entrances, while they are dulled when ventilation reverses. The thermally perturbed zones (*Ventilated Warm and Cold Zones*, in red and blue in Figure 11.1.a/b, respectively) extend within a cave for a distance equals to the *convection length* (Chapter 7; Sedaghatkish et al., 2024a) and within the encasing rock for a distance around the conduit named as *convection width* (typically 1/10 of the convection length; Chapter 7, Sedaghatkish et al., 2024a). Beyond the annual *convection length* is the homothermic zone (or deep karst; Figure 11.1.a) where temperature fluctuations are smaller than 10% of the external annual fluctuations. The residual seasonal fluctuation reaches deep zones with a time lag which depends on the distance from the entrances and the cave depth, as well as the annual fluctuation of airflow. The extension of the convection lengths, at an annual time scale, mainly depends on air mass flow rates, conduit sizes and singularities present along the air paths. Indeed, the cave geometries determine the cave aerodynamic resistance (R). Changes in the cave resistance over time (in the case of Longeigue Cave due to water table fluctuations, but possibly also induced by other reasons e.g. obstructions of cave entrances by rock falls, cyclical presence of ice/snow occupying part of the cross-section) would affect the amplitude of the airflow fluctuations which may modify the AMT along the conduit, as shown by Figure 8.7 in Chapter 8. Figure 11.1.b depicts vertical gradients along the vadose zone of a schematic cave under a chimney effect. It also displays the average vertical gradient of a standard external atmosphere ($0.65^{\circ}\text{C}/100\text{ m}$; ICAO, 1993). The internal gradients seasonally vary at the entrances (along the heterothermic zones) according to the ventilation regime (downward or upward). Along the homothermic zone, thermal gradients range between 0.234 and $0.55^{\circ}\text{C}/100\text{ m}$ since they are constrained by the physical gradients of water and humid air, respectively. Measured gradients usually range between ~ 0.50 and $0.30^{\circ}\text{C}/100\text{ m}$ (Luetscher and Jeannin, 2004). From previous studies (e.g.

Luetscher and Jeannin, 2004; Figure 1.6) the heterothermic zone hardly overcomes 100 m depth. However, since the convection length depends strongly on mass flow rates, which in turn depend on the cave aeraulic resistance (thus it depends on conduit sizes and singularities present therein), tight passages reduce the thermal penetration distance of external fluctuations. Yet, large cave passages with reduced aeraulic resistance, and consequently potentially larger mass flow rates, may allow the propagation of external thermal perturbations for higher depths/distances. This last statement, however, should be verified with further observation within karst systems showing these characteristics. An overview of the meaning and characteristics of the convection length is given in the next Chapter 11.3.

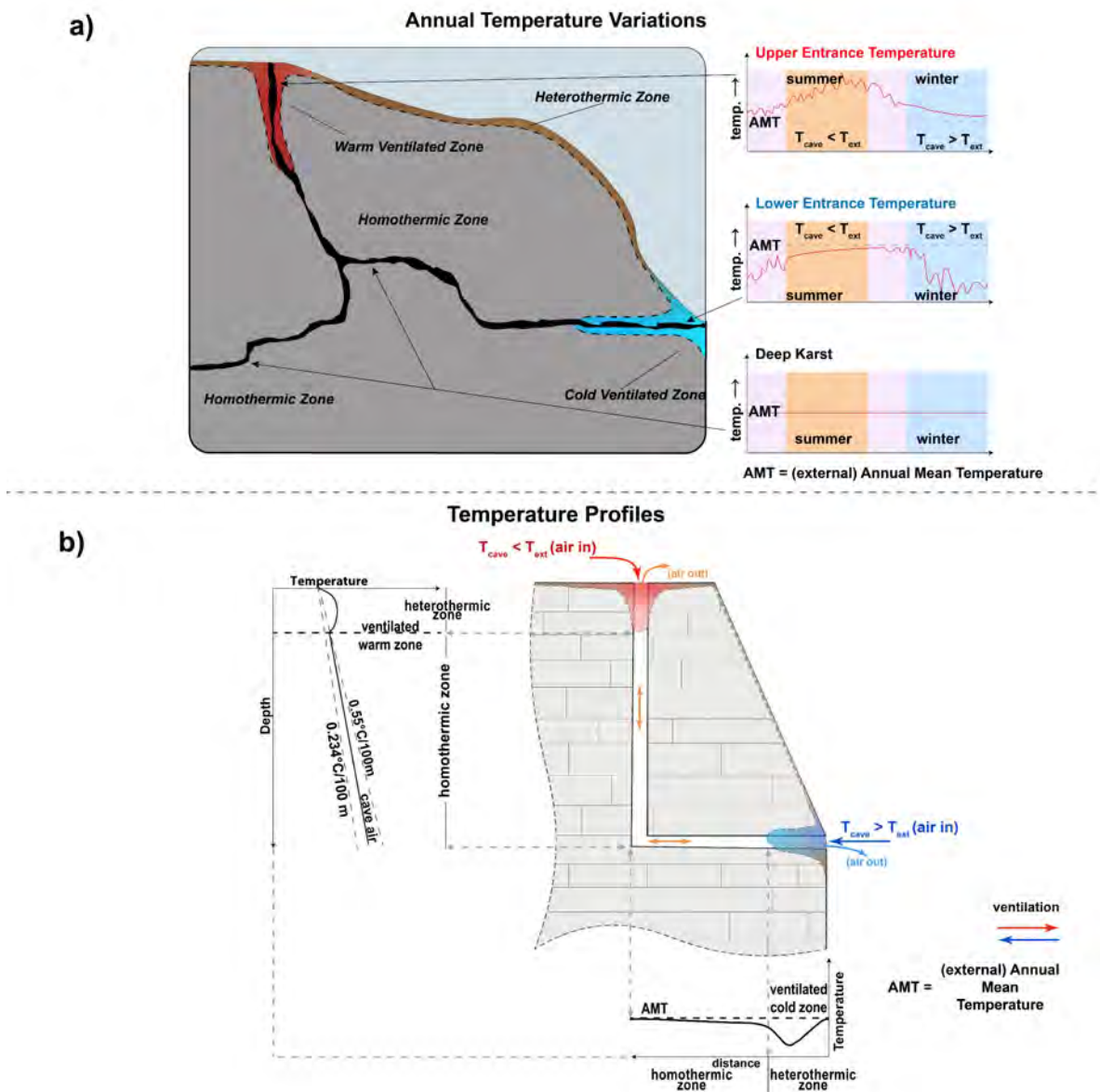


Figure 11.1 – in a), a simplified sketch of a cave under a chimney effect showing typically observed temperature time series at both entrances (heterothermic zones) and in the homothermic zone. In b) schematic representation of a vertical cave under a chimney effect with typically observed vertical thermal gradients (between physical gradients of water and humid air) and horizontal thermal profile (modified after Luetscher and Jeannin, 2004).

11.3. The convection lengths in karst systems

The chimney effect (or forced convection or buoyancy-driven ventilation) is recognised by several authors as the main mechanism driving the ventilation through the karst vadose zone (Kukuljan et al., 2021; Covington and Perne, 2015; Badino, 2010; Luetscher and Jeannin, 2004), while cave ventilation is an efficient mode of transport for heat (Dredge et al., 2013) as we have shown in Chapter 8. Ventilation allows external temperature fluctuations to penetrate the cave until being attenuated with distance.

As long as the chimney effect is the main driving force, the convection length represents the distance necessary to reduce the fluctuation of an advected thermal signal by 90% of the initial amplitude (Sedaghatkish et al., 2024), namely, it is the distance necessary for an input signal to reach equilibrium with the surrounding medium (Covington et al., 2012). Numerical simulations in Chapter 7 show that the convection length is approximately proportional to the amplitude of the flowrate annual fluctuations divided by the square root of the cave radius (Sedaghatkish et al., 2024). Moreover, in Chapters 7 and 8, for the first time, we compared convection lengths estimated through field data and assessed by theoretical models for different caves. The convection length can be in the order of a few tens to a hundred meters within the karst system, and it is much larger than the diffusion length, which typically extends for a few meters (typically between 5 and 10 m) at annual time scales (see Chapter 7). This convective zone is usually followed by a deep karst zone, where temperature is almost constant over the year. In contrast, results from Longeague Cave (Chapter 8) have shown that if the convection length is half of the cave length, the convection lengths related to lower and upper entrances (red and blue zones shown in Figure 11.1) are in contact. As a result, the deep karst zone (or homothermic zone) is not reached along Longeague's ventilated conduit, and the upper and lower convection lengths may affect each other. This 300 m cave has also shown that the upper and lower convection lengths (i.e. related to upper and lower entrances) are characterised by the *Ventilated Warm* and *Cold Zones*, respectively. The observed convection lengths (Figure 8.7) are ~ 150 m, while Eq. 8.7 yields an order of magnitude of 10^2 m.

Furthermore, Eq. 8.7 in Chapter 8 (from Sedaghatkish et al., 2024) applies also to U- and reversed U-shaped caves (Figure 1.5.b and Figure 1.5.c in 0), which comply with a chimney effect. However, it cannot be applied to caves ventilated by mechanisms other than the chimney effect (Figure 1.2). The relation given by Eq. 8.7 is not applicable for caves dominated by a barometric effect (Figure 1.2.c), i.e. caves whose ventilation is driven by the irregular succession of atmospheric cyclones and anticyclones (Gomell and Pflitsch, 202) and to daily and sub-daily variations in Surface Air Pressure (SAP) (Borsato et al. 2024), since the thermal and aerologic problems are uncoupled (i.e., the airflow rate is not controlled by temperature contrasts between the cave and the outside). We recall that barometric caves are characterised by large underground voids connected with the external surface by narrow entrances which limit fast and dynamic air exchanges related to e.g. daily and seasonal temperature variations. Therefore, caves with these characteristics have ventilation often controlled by surface air pressure (SAP) variations (Borsato et al., 2024). Borsato et al. (2024) have studied the barometric ventilation process taking place in Lamalunga cave (Altamura, Southern Italy) and its influence on cave temperatures. This cave is mostly known to host an excellently preserved specimen of a Neanderthal skeleton (Pesce Delfino and Vacca, 1993). The importance of the site fostered DNA and U-Th dating (Lari et al 2015) as well as investigations on (paleo)climatological research (Borsato et al., 2024). In fact, the skeleton is encrusted by myriad calcite coralloids, which allowed U-Th dating (crust age between 130.1 ± 1.9 ka and 172 ± 15 ka; Lari et al 2015). The connections between microclimatic parameters

and speleothem growth are therefore key to understanding the paleoclimatic conditions during which calcite was deposited. In this cave, the barometric ventilation process is favourable since the horizontal underground gallery is connected with the external atmosphere through a narrow and vertical shaft (Borsato et al., 2024). Ventilation is weak but active, especially in the entrance shaft, and follows barometric oscillations (diurnal and sub-diurnal SAP variations), allowing for smooth and continuous air exchanges between the cave and the exterior. Borsato and coworkers (2024) have shown that daily temperature variations are mostly limited to the entrance shaft (a few meters below the surface) when the SAP is higher than the inside pressure. In summer, temperature fluctuations were detected in the inner cave parts, although their amplitude was very small ($\sim 0.0035^{\circ}\text{C}$ at 28 m from the entrance vs $\sim 0.52^{\circ}\text{C}$ in the entrance shaft; Borsato et al., 2024). The authors suggest that these temperature fluctuations are associated with the weak summer ventilation, when the SAP is high and warm air is pumped into the cave, while most of the daily fluctuations are damped along the walls of the entrance shaft. Instead, in winter, the air advection related to density (temperature) differences is confined to the entrance shaft, and no fluctuations related to this process are visible in the cave. The same authors explain that, although in winter cold and dense external air can potentially sink inside the cave through the vertical shaft, allowing for a cold thermal perturbation (Borsato et al., 2024), the narrow diameter of the entrance limits the air exchanges and hinders temperature transmission inside the main gallery of the cave by convection. Therefore, the authors concluded that the influence of forced convection related to density (temperature) contrasts in most parts of the cave is minimal, and its contribution to the energy balance of the cave is limited to an order of magnitude of hundredths of a degree Celsius. Thus, given the limited contribution of convective air exchanges, the detection of a small barometric effect on cave temperatures is allowed throughout the year in all the cave galleries of Lamalunga cave (Borsato et al., 2024). Although barometric ventilation ensures air exchanges between the cave and the external atmosphere, results from Borsato et al. (2024) also suggest that the thermal convection length could not be present in these kinds of ventilated caves, or it could be limited to a few meters from the entrance or even reduced to the diffusion length. In contrast, Bourges et al. (2006) presented details about the particular microclimate of L'Aven D'Ornagac, a cave where these authors showed the coexistence of both density-driven and pressure-driven ventilations. The cave is characterised by a large and vertical opening that leads to the initial chambers, and does not present any other entrances. The large and open volumes close to the entrance allow the establishment of a winter regime dominated by a density-driven ventilation, in which cold, dense air from outside can flow into the cave. Therefore, cave temperatures in these cave areas are strongly connected with the outside. Viceversa, in summer, warm and light external air is not able to penetrate the cave atmosphere, and cave temperatures in the entrance zone are stable. Thus, this initial part of the cave acts as a cold-air trap. In contrast, the far regions of the cave (such as the Salle Plane 500 m far from the entrance; Bourges et al., 2006) the air temperature is extremely homogeneous and is characterised by a half-daily signal of amplitude less than 0.03°C . Bourges and coworkers (2006) showed that the temperature is completely independent of external conditions and, instead, is correlated with variations in atmospheric pressure (relationship between cave temperature and dP/dt), which induce micro-thermal changes linked to compression and relaxation processes in the air volume. Therefore, this cave presented by Bourges et al. (2006) can show a heterothermic zone related to the thermo-convective ventilation, while the inner cave part, which might be considered in the homothermic zone, shows minor thermal fluctuations related to the pressure-induced temperature variations (Bourges et al. 2006).

The relation for estimating the convection length cannot be directly used either for caves with one entrance and acting as a cold air trap (Figure 1.2.d) given that cold air is entering and warm air is

escaping the cave through the same entrance, although a penetration length exists, as it is shown by e.g. Šebela and Turk, (2018). Šebela and Turk (2018), Edenborn et al. (2012) and Luetscher et al. (2008) have shown that, in cold air traps, due to their geometry, ventilation is active when external air temperature is lower than the inner temperature (active periods), while it is stopped when external temperature is higher than the inside (closed periods). Therefore, the aerologic and thermal problems are in this case coupled. However, if cold and warm air transits from the same entrance as sketched in Figure 1.2.d, air stratification could affect temperature distributions between the ground and ceiling. This may affect the convection length but also the convection width, which could be different between the rock surrounding the ceiling and that surrounding the ground, possibly being larger in the second case since cold air is denser and flows preferentially close to the soil. The same reasoning applies to warm air traps (Figure 1.2.h). Further research and models are therefore needed in order to adjust Eq. 8.7 and make it suitable for estimating the convection length of external fluctuations in these geometries.

The heat exchange that occurs along the convection length also propagates into the surrounding rock far from the conduit for a thickness equivalent to the convection width, which is typically one-tenth of the convection length for AMT (i.e. referring to an annual time scale; Sedaghatkish et al., 2024).

Wall and air temperature are determined by the interaction between convection and conduction (Gabrovšek, 2023; Lismonde, 2002), allowing for a faster propagation of external temperature signals inside the conduit than only pure conduction (e.g. Domínguez-Villar et al., 2023). Therefore, the effect of air convection and conduction must be coupled when one wants to investigate the heat exchanges in ventilated conduits, as long as the airflow is turbulent. Conversely, when airflow reduces temperature difference between air and walls becomes smaller, and heat exchange reduces as well, such as in the case of Longeigue Cave during periods of no ventilation (closed system periods in Chapter 8).

Our findings are particularly meaningful for:

- interpretation of paleoclimatic proxies from caves;
- thermal cave responses to external climate changes;
- conditions controlling ecological niches in caves.

11.3.1. Paleoclimatic outlooks

Caves are widely regarded as important “archives” of proxies used for (paleo)environmental reconstructions, while the extensive geographic distribution of caves allows researchers to make correlations between cave proxies at a (nearly) global scale (see e.g. SiSALv2 database [Comas-Bru et al., 2020] and SiSalv3 database projects [Kaushal et al., 2024]). Primary cave archives are speleothems as well as sediments and perennial ice bodies. Paleoenvironmental models aim to reconstruct past external conditions and paleoenvironmental changes at a regional scale and can be tested against the “information” recorded from cave proxies. For example, information on past external conditions (e.g., wet or dry weather periods, ice ages, temperature) is carried into a cave by seepage water that “fixes” the information during the growth of speleothems. Those cave objects generally offer a high temporal resolution and a quite large temporal range for paleoenvironmental reconstructions (up to ~500 ka for speleothems; Fairchild and Baker, 2012; Richards and Dorale, 2003) as long as they are protected in a stable environment, which can mitigate the alteration of a proxy and preserve its paleoenvironmental prints. Regional signals from outside recorded by cave proxies can be hidden by in-cave environmental

conditions (e.g. cave temperature, humidity and ventilation) (Fairchild and Baker, 2012). Therefore, for caves with, for instance, intense airflow, deciphering paleoclimatic signatures can be more complicated.

Ventilation largely controls cave temperature, humidity, $p\text{CO}_2$ and isotopic fractionation of e.g. carbon (^{13}C) and oxygen (^{18}O) preserved in the calcite lattice forming a speleothem (Frisia et al., 2011; Luetscher et al., 2011; Baldini et al., 2008; Spötl et al., 2005). CaCO_3 isotopic contents such as $\delta^{18}\text{O}$ (ratio between ^{18}O and ^{16}O) and $\delta^{13}\text{C}$ (ratio between ^{13}C and ^{12}C) are widely used to reconstruct e.g. changes in external climatic and hydrological conditions in the past (Luetscher et al., 2021; Cuthbert et al., 2014; Fairchild and Baker, 2012; Spötl et al., 2005). The isotopic fractionation of these isotopes is mainly controlled by the *amount effect* (i.e. the isotopic composition is inversely related to the amount of rainfall), temperature, CO_2 degassing and flow rate of the dripping water (Lyu et al., 2020; Rau et al., 2015; Fairchild and Baker, 2012; Baldini et al., 2008; Spötl et al., 2005). Rau et al. (2015), Baldini et al. (2008), and Spötl et al. (2005) pointed out the role of ventilation on the isotopic enrichment of dripping water and that water cooling related to evaporation induced by air drafts could modify the forming temperature of speleothems. Therefore, investigating cave ventilation dynamics and the related temperature variations helps to interpret paleoclimatic data retrieved from cave proxies, especially for speleothems, as highlighted by researchers (Faimon et al., 2012; Kowalczyk and Froelich, 2010; Spötl et al., 2005), and to untangle pieces of information related to external paleoenvironmental conditions from those related to local cave conditions.

Ventilated Warm Zones close to the upper entrances and *Ventilated Cold Zones* close to the lower entrances (Figure 11.1) are characteristic of ventilated caves. The most evident effect is ice formation at lower entrances and in cold air traps in caves opening in high mountain regions, e.g. Eisriesenwelt ice cave (Perşoiu, 2018), Monlesi cave (Luetscher et al., 2008) or Grotte de Pingouin (Borreguero et al., 2009). We have shown in Chapter 8 that 90% of the heat exchanges happen along the first tens of meters (50 m) of the convection length (for Longeague Cave, $\sim 100/150$ m from the upper/lower cave entrance). The presence of these thermal anomalies (*Ventilated Warm* and *Cold Zones*) needs to be taken into account when sampling speleothems since, for example, their isotopic fractionation could be affected by the temperature fluctuations within cold/warm zones, thus speleothem-based paleoclimatic interpretations could be affected by a temperature bias due to this shift. Therefore, paleoclimate studies based on speleothems primarily concentrate on samples collected from the homothermic sections of caves (Obleitner et al., 2024) to assess e.g. average paleotemperatures (e.g. using clumped isotope thermometry Δ_{47} , Rau et al., 2015; Kluge et al., 2014; Affek et al., 2008). Here, the cave air and rock have close temperatures while their annual and daily fluctuations are almost nil, although fluctuations at lower frequencies (decennial or larger) might be important. Indeed, Rau et al. (2015) suggest that the best place to sample speleothems is where the relative humidity is constantly close to saturation and there is no airflow, to minimise the effects related to evaporative cooling. Yet, knowledge about e.g. active/not active ventilation phases, due to e.g. glacier evolutions clogging cave entrances (e.g. Luetscher et al., 2011), can be retrieved from speleothems collected along the heterothermic zone. In Chapter 7 (Sedaghatkish et al., 2024) and Chapter 8, we furnished an equation (Eq.25 in Chapter 7 and Eq. 8.7 in Chapter 8) to assess the order of magnitude of the convection length for annual fluctuations in caves ventilated by a chimney effect. Moreover, in cases where caves are composed of very large galleries and the airflow is in the order of several m^3/s , e.g. the Albanian cave Shpella Shtares (42.31766°N , 19.85403°E , 1,427 m a.s.l.) where the airflow was assessed at $\sim 30 \text{ m}^3/\text{s}$ and conduit diameters usually larger than 2 m (Pastore et al., 2019), the convection length could extend over kilometres and hundreds of meters in the axial and radial conduit directions, respectively. This will

extend the heterothermic zone and therefore speleothems far from the entrances may still be affected by external signals. The evaluation of the convection length may prove useful in identifying places where annual ventilation-induced temperature fluctuations are dulled and likely poorly interfere with speleothem growth.

There is limited knowledge about how rapidly climate changes in the external atmosphere penetrate caves (Obleitner et al., 2024). Our aeraulic model proposed for Longeigue Cave (Chapter 8), although based on an annual time scale (high fluctuation frequencies), can be used as a basis for investigating smaller frequency scales (decadal, centennial and millennial) to infer the rate at which air-advected thermal perturbations penetrate a karst system.

We have shown (Chapter 8) that it is possible to set up a reliable cave aeraulic model to determine air mass flow rate as long as ΔT (i.e. the driving pressure, ΔP) and aeraulic resistance R (determined by cave geomorphological characteristics) are known. Both these parameters (ΔP and R) can be retrieved from field measurements. We underline this point since a theoretical approach to quantify R leads in most cases to a wrong estimation (see paragraph 8.4.1 in Chapter 8). Moreover, cyclic modifications of cave aeraulic resistance (in the case of Longeigue due to water table fluctuations, but possibly also induced by other reasons e.g. obstructions of cave entrances by rock falls, cyclical presence of ice/snow occupying part of the cross-section) would affect both the cave air flow rate and the distance that temperature fluctuations can reach from the entrances, thereby varying the convection length. This has important consequences for the cave's thermal response and the temperature recorded by speleothems (Spötl et al., 2005). Furthermore, variations in the convection length may cause areas of the cave that were previously, for example, in the homothermic zone to become part of the heterothermic zone and vice versa. It is noteworthy that Longeigue's upper entrance was made human accessible in 1982 by digging the upper entrance passage (Jeannin, 2018). This modification, made by speleologists, induced a significant increase in the air flow rate through the cave (Jeannin, 2018). Although the airflow was probably not nil before the opening, changes in the cave geometry have certainly modified the airflow by increasing it. This fact may likely have modified the cave's inner climate and penetration length. Moreover, in Longeigue Cave, the alternate active/non-active ventilation due to the water table fluctuations strongly modifies the cave thermal response by changing the air mass flow rate.

Cave ventilation drives the degassing speed of CO_2 (which is controlled by CO_2 partial pressure of the cave atmosphere) from dripping water forming a speleothem (Rau et al., 2015; Fairchild and Baker, 2012; Kowalczyk and Froelich, 2010; Baldini et al., 2008; Spötl et al., 2005). CO_2 degassing enhances calcite precipitation, leading to larger calcite laminae forming, for example, a stalagmite. Changes in ventilation can modify CO_2 concentrations along a cave, which drives speleothem growth by accelerating or reducing it. Therefore, larger laminae in a speleothem are not always due to an increase in the dripping that shapes a speleothem but are also linked to the alternation between ventilated/non-ventilated phases that can lead to annual growth cycles (e.g. Boch et al., 2010, 2009). This fact is of less importance for studies that focus on long time periods, e.g. decadal or larger time scales. Our studies are related to annual time scales, providing a benefit to high-resolution investigations (e.g., micromill or laser ablation; Baldini et al., 2008; Spötl and Matthey, 2006). These investigations are necessary to identify and interpret climate imprints related to seasonal cycles. For high-resolution studies using fast-growing speleothems (Obleitner et al., 2024), our model for reconstructing airflow based on internal/external temperatures and cave resistance is quite relevant. If the aeraulic model is based on plausible ranges of paleo-driving pressures (derived from ΔT), it could be used to estimate past cave

flow rates while accounting for the thermal delay between external and internal cave conditions during climate change periods. This delay during a climate change period is due to the time required for a thermal signal to propagate to distant cave areas, as it is buffered by the rock's thermal inertia (Mammola et al., 2019), and would likely increase temperature contrasts between T between inner and outer temperatures, thereby influencing ventilation regimes in both duration and intensity (Mammola et al., 2019).

Such insights would help in the interpretation of paleoclimatic proxies, helping to distinguish local effects due to ventilation from those driven by external climatic conditions.

Therefore, when studying cave proxies, preliminary field observations should determine when it is most appropriate to apply a convective or conduction model, i.e. whether a cave is (or was) ventilated. A cave ventilation model and the assessment of the cave convection length can help to 1) better choose sampling places, 2) interpret the paleoclimatic signals recorded by proxies.

11.3.2. Cave thermal response to external climatic changes

If cave proxies record external paleoclimatic conditions, it is because climatic changes can penetrate a karst massif. Therefore, it is fair to ask to what extent the ongoing climate changes (general global warming, Gulev et al., 2021) are affecting cave environmental conditions and, eventually, how fast the external variations penetrate caves.

It has been demonstrated that external temperatures are rising by about 0.3°C per decade in the European Alps regions (IPCC, 2022). In a recent study (Obleitner et al., 2024), the authors have shown that several caves lying in the European Alps are experiencing a temperature increase of about 0.2°C per decade, while in the areas surrounding the caves they studied, the increase reached 0.5°C per decade. These authors have monitored four caves for 20 years, which have shown a temperature increase with a delay of about 15 years. They also observed that this temperature increase is related to air convection (at least for three of the four caves) due to a chimney effect, and that the temperature variations indicate that the monitored cave parts relate to remote parts of the convection length (or heterothermic zone) since annual fluctuations are still detectable in the order of a few tenths of degrees.

The authors Obleitner et al. (2024) have shown that the decadal warming is stronger near the entrance(s) than in the deeper cave sections since the air temperature fluctuations are well expressed, in contrast to far cave sections where the fluctuations are attenuated. Moreover, Obleitner et al. (2024) highlighted differences in temperature increase between periods when a cave system is open (ventilated) and closed (no ventilation). They observed that the warming trends are higher during open periods than during closed periods. Although our results are at a yearly time scale, whereas those from Obleitner et al. (2024) are at a decadal time scale, their observations match ours from Longeague Cave (Chapter 8). There, we have quantified that 90% of the heat exchanges (between air and rock) took place in the first tenth of meters from the entrances and that the cave's temperature averages, variations, and overall climatic conditions are tightly linked to ventilation regimes. Indeed, the net amount of the energy exchanged depends on the duration of open (active ventilation) and closed (no ventilation) periods. Field data (Chapter 8, Figure 8.6) have shown that the yearly distribution of open and closed periods led to significant changes in the annual mean temperatures of the *Ventilated Warm* and *Cold Zones* in the cave, observed close to the upper and lower entrances, respectively. We have also demonstrated that, in the case of Longeague Cave, open and closed periods are controlled by external rainfall. In this specific case, external precipitations indirectly control cave climate and thermal

exchange processes (i.e. by convection during open periods or conduction during closed periods) since the water level of the perennial lake determines whether ventilation is active or not. However, it is not possible to generalise and globally assess which is the impact of cave sumps that occupy and close cave passages. It is known that global warming also affects the intensity, duration and yearly distribution of precipitation events (Trenberth, 2011). For example, snowfall occurring in winter can clog cave entrances, particularly at higher elevations. Reduced snow accumulation may result in upper cave entrances remaining open for longer periods throughout the year, thereby enhancing the chimney effect. Counterintuitively, the most significant impacts can occur during summer: without snow obstructions in spring and summer, warmer external air can easily enter the cave through upper entrances, causing cave air temperatures to increase. Also, variations in rainfall rates may affect caves in which ventilated passages could be shut by the rise of the water table. Or vice versa, long periods of drought may open passages by lowering the water table and starting ventilation.

This will lead to variations in the ventilation regime, also affecting the thermal regime of the cave. Disequilibrium between winter and summer ventilation may cause differences between the cave temperature and the surrounding massif. Therefore, a cave could be warmer if the system is open longer in summer and colder if it is open longer in winter. Moreover, temperature increases are projected to be greater during the winter season than during the summer season (IPCC, 2023). This could reduce the temperature contrast between the cave and the exterior during cold periods, thereby also decreasing ventilation strength in winter. As a result, the cave would be exposed to prolonged summer ventilation, eventually leading to an increase in average cave temperatures.

Although our study focused on the heterothermic zone and that of Obleitner et al. (2024) on its remote parts, it can be anticipated that also the homothermic parts will be affected by external changes with a certain delay. Yet, less is known about how quickly an external warming (or cooling) can propagate far inside a cave. This clearly depends on ventilation, cave geometries and dimensions and also on the role played by water infiltration (Obleitner et al., 2024). This propagation velocity and the delay necessary for an external perturbation to reach very deep karst areas are key questions for speleotherm-based paleoclimatic studies since they aim to date external climate conditions, especially research looking at high-resolution timescales. Long-term monitoring (e.g. Obleitner et al., 2024) and advanced modeling (e.g. our works Sedaghatkish et al., 2025, 2024) will help to have quantitative insights in this field.

11.3.3. Impact on ecological niches

Caves, along with fissure systems (Mammola et al., 2019, 2016; Tobin et al., 2013), are recognised as unique natural laboratories for the study of multiple ecological and evolutionary processes, due to their geographical isolation and generally recognised climatic stability. This isolation has led to a high specialisation of cave organisms and diverse biodiversity from place to place (Mammola et al., 2019; Yao et al., 2016; Lunghi et al., 2014). In order to reduce the number of variables at play, scientists have often studied isolated ecological niches and dwelling organisms to investigate the role of climate change on ecological response. Confined habitats, such as mountain summits, oceanic islands, lakes, as well as caves and other underground environments, have provided valuable insights into climate impacts on biodiversity, serving as models that are then upscaled to a broader range of systems and organisms (Mammola et al., 2019).

The environmental characteristics of caves have allowed subterranean species to evolve in almost stable environments, in particular: stable temperatures and certainly a lack of sunlight. However,

according to our results and also from previous studies (Gabrovšek, 2023; Covington et al., 2012; Wigley and Brown, 1971), in ventilated caves, the temperature might be considered stable only beyond the convection length, in contrast with the entrance zones. Cave entrances represent the ecological interface between the surface and underground world (Tobin et al., 2013), experiencing greater (although smaller than the outside) fluctuations in temperature, humidity and light than the deep cave parts. Yet, Yao et al. (2016) and Tobin et al. (2013) demonstrated that cave entrances are also vital ecological niches, supporting as many endemic species as the deeper karst zones. Medina et al. (2023) also suggest that daily cycles of ventilation (and flow reversals) related to daily temperature fluctuations may impact the circadian rhythms of underground organisms, which in turn regulate different processes in organisms (Medina et al., 2023; Friedrich, 2013; Yerushalmi and Green, 2009). Yet, subterranean species, often endemic, show low tolerance to climatic perturbations and, due to their limited dispersal capacity (namely, the capacity to change their habitat), are more prone to local extinction than surface species (Mammola et al., 2019).

Our quantitative results presented in Chapter 8, alongside prior qualitative research (e.g. Badino, 2004; Mammola et al., 2016), pointed out that the air-rock energy transfer mostly takes place in the close-to-entrance sections of caves, i.e. in the heterothermic zone. In Longeigue Cave, most (90%) of the energy flux from the external atmosphere is dissipated close to the entrances. Therefore, we can anticipate, in agreement also with observations reported by Obleitner et al. (2024), that climate changes will primarily affect climatic conditions along the convective zone. However, this could also be followed by dramatic effects on biological components (e.g. temperature and humidity) of deeper hypogean ecosystems, which may induce species extinctions given the reduced tolerance to temperature changes and low dispersal capacity of underground fauna (Mammola et al., 2018).

In the homothermic zone, cave thermal inertia is expected to produce the observed time lag between the beginning of external climate change and its impact on the internal environment (Tobin et al., 2013). Authors mainly carried out quantitative studies that assumed pure conductive models (e.g. Domínguez-Villar et al., 2015; Rau et al., 2015), focusing on shallow caves with limited air circulation. Through such models, authors could assess an upper bound of the time duration necessary for climate change to propagate in a cave. Depending on depth, the time lag inferred from conductive models may range from decades to centuries. For instance, Domínguez-Villar et al. (2015) estimated that transferring an external temperature fluctuation to a depth of approximately 40 meters by conduction alone could take as long as 20 years. From an ecological standpoint, this suggests that deep underground species have not yet evolved to accommodate ongoing external changes (Mammola et al., 2018). Although studies focused only on conduction through the host rock and heat fluxes advected by water (e.g. Luhmann et al., 2015; Covington et al., 2011; Long and Gilcrease, 2009; Birk et al., 2004) as the main mechanisms controlling cave temperature, we have shown that cave ventilation significantly increases the depth reached by thermal oscillations, promoting the formation of *Ventilated Warm* and *Cold Zones* and controls their length (convection length). Although conduction is still an important component of a cave energy balance, nevertheless, our findings, supported by the observations made in Alpine caves by Obleitner et al. (2024), suggest that air convection may transport external perturbations much faster and deeper into a cave than conduction solely, particularly within the convection length and in caves such as Longeigue lying entirely within this convective zone. We presumably suppose that if the driving pressure (ΔP) increases because of varied AMT contrasts between the cave and outside caused by climate change, the annual airflow fluctuation will vary too, and accordingly, the convection length will also vary. We can hypothesise that if the driving pressure

(ΔP) increases due to greater contrasts in annual mean temperature (AMT) between the cave and the outside environment caused by climate change, then the amplitude of annual airflow fluctuations will also increase (Eq. 1.4 and Eq. 1.5). Consequently, the convection length within the cave system will extend as well. However, this statement should be investigated through numerical models at larger than annual time scales and coupling both conductive and convective processes.

Tobin et al. (2013) differentiate cave habitats into three macro-zones, based on environmental conditions and variability: 1) an entrance zone that is climatically variable and strongly affected by external changes in the surface environment; 2) a deep karst zone, which is characterized by relatively stable temperature and humidity (or near saturation); 3) a transition zone that seasonally shows characteristics of either the entrance or deep karst zone. Tobin and coworkers (2013) also underline that the spatial extents of these zones vary between different caves and are primarily determined by cave characteristics such as passage morphologies and airflow patterns.

While the convection length is often only briefly mentioned in ecological studies (Moldovan et al., 2018) and called as transition zone (Tobin et al., 2013), estimation of its length could also be considered a critical biological indicator of the transition from surface to more stable deep cave conditions. Taxa distribution within caves is strongly affected by cave ecological characteristics such as resource availability, microclimate variability, other species invasion history and species interactions, showing a certain gradient of distribution from entrances to deep parts (Tobin et al., 2013), and from “deep, dark and humid caves, to dry caves with wider entrances and extended photic areas” (Lunghi et al., 2014). We recall that the convection length increases with increasing flow rates and larger diameters (since the aerologic resistance decreases for larger conduits). Modification of the thermal penetration length may impact the distribution of taxa along an underground ecosystem. Given the importance of ventilation in regulating cave temperature and humidity, factors like cave resistance and diameters could prove useful parameters to consider when building evolutionary models of underground species, given that they correlate cave ventilation with convection length. Changes in cave geometries over time (e.g. variation in diameters due to natural or anthropogenic occurrences, passage clogs, etc.) affect ventilation patterns, which in turn influence the temperature and very likely characteristics of ecological niches.

Longeigue Cave represents a potentially interesting site to investigate underground fauna, given its climatic characteristics. Specifically, variation in ventilation regimes, related to open and closed system periods, could be interesting to study for their potential impact on circadian rhythms of dwelling fauna (e.g. Medina et al., 2023). Special attention could be given to the fact that the cave's upper entrance was opened around 40 years ago (in 1982), which triggered a chimney effect (Jeannin, 2018). The cave has been observed to host large communities (Gigon, 1976), especially in the warmer upper entrance, such as e.g. *Areneina*, *Opiliones*, *Copepoda*, *Diplura*, *Coleoptera* (Gigon, 1976), as well as *Oxychilus sp.* and *Dolichopoda sp.* (observed by the author during this project). Most of these species were observed and described before cave opening and ventilation started. Therefore, by making new observations on cave fauna, Longeigue can prove useful to investigate whether, and how, this modification may have impacted underground life.

Airborne nutrient and microbial material dispersion is also critical for species distribution within caves. For example, airborne fungi and bacteria monitored in Altamira Cave by Dredge et al. (2013) exhibit distributions influenced by ventilation, with fungi following the same pattern as suspended aerosols, while the aerial distribution of suspended bacteria differs from the observations of bacterial

colonisation, which is dominant near the cave entrance. Dredge and coworkers (2013) suggest that the connection between bioaerosols (bacteria and fungi) and cave inner-surface colonisation is likely influenced by the distribution of inorganic aerosols, which supply nutrients and create favourable conditions for microbial growth. Our studies about dispersion (Chapter 5) suggest that Taylor's dispersion model and, in general, a 1-D model may be insufficient to explain the transport of airborne nutrients in naturally ventilated passages. Thus, more advanced models, supported by field data, may be needed for accurate assessments.

11.4. Classification of springs based on their thermal behaviour

In this chapter, we want to follow up on the discussion begun in Chapter 10, which qualitatively extends the concepts of convection length to hydrologically active caves. Although in this work we mainly focused on almost dry and ventilated caves, it is worthwhile to have a look at the thermal behaviour of flowing water in caves and at karst springs, as water and cave air temperatures are intimately linked along the vadose zone (Luetscher and Jeannin, 2004; Badino, 2004).

In Chapter 10 (Figure 10.5), we have shown that air and flowing-water temperatures (recorded in Rouge Eau Cave) are linked to each other, with air temperature closely mimicking water temperature. Results reported in Chapter 10 from Rouge Eau suggest that, owing to heat exchange between air, water, and rock, and to the greater thermal inertia of water compared to air (Badino, 2004), cave air temperature tends to approach water temperature with increasing distance from the surface. Therefore, we can reasonably suppose that, when there is abundant and turbulent flowing water along the convection length (or heterothermic zone) and the homothermic zone, water may control air temperatures (especially when airflow is reduced), while both fluid fluctuations are dulled by rock thermal inertia.

Other factors can influence climatic conditions (air and water temperature and related thermal processes) within hydrologically active caves; for example, Luetscher and Jeannin (2004) highlight that water control on cave air temperature is facilitated by a reduced airflow, which (a reduced airflow) can derive from geometrical factors, like system depths and aeraulic resistance, as we have also shown in previous chapters. These authors (Luetscher and Jeannin, 2004) explain that, for example, the airflow can be strongly reduced where there is a small difference in altitude (i.e., vadose zone thinner than 100 m, e.g. lowland karst plateaus) and/or when the cave aeraulic resistance is high (small driving pressure ΔP and/or high resistance in Eq. 1.4). Also climate conditions with low temperature contrasts between cave and external temperatures and low seasonal oscillations (e.g. coastal and equatorial regions) will cause reduced cave ventilation. Underground rivers flowing in caves lying in these regions, for example, Puerto Princesa Underground River (Philippines), described in Badino (2018) and Muruk Cave (New Guinea), described in Audra (2001), show cave air temperature controlled by water temperature (which in turn is controlled by external conditions; Badino, 2018).

Therefore, for the reasons above, water temperature and related fluctuations at karst springs or along underground rivers may reveal information on climatic conditions over the vadose zone where water flows. In the following chapters, we will introduce previous works that aimed at classifying thermal patterns observed at karst springs, and then we will propose an alternative way of classifying thermal patterns at springs.

11.4.1. Previous classifications

In Chapter 10, we have qualitatively shown that the water convection length of hydrologically active caves varies according to the water flow rate. Flood events cause temperature peaks visible on the thermographs and, in the case of concentrated recharge, these peaks are in the order of several degrees that can be transmitted with very low damping for several hundreds of meters or even

kilometres along the cave system (e.g. at Rouge Eau caves and Škocjan-Kačna, respectively in Figure 10.6.a/b, Chapter 10).

In this chapter, we will have an outlook on the range of thermographs identified at springs based on the existing literature (Luo et al., 2023; Luhmann et al., 2011) and we will further propose a possible classification based on the recharge modes and depth of the system.

Although some authors attempted to construct models to explore the heat transport in underground water flow (e.g. Luhmann et al., 2015; Covington et al., 2011; Long and Gilcrease, 2009; Birk et al., 2004), only a few works (Luo et al., 2023; Luhmann et al., 2011) attempted to define categories of springs' thermographs and to provide conceptual interpretations. Luhmann et al. (2011) proposed to quantify the efficiency of heat transfer between water and the surrounding rock based on the Stanton number (St), which is a dimensionless number that represents the ratio between the heat flux at the wall of the conduit and the heat transported along the conduit. A high St ($St \gg 1$) indicates efficient (or effective) heat transfer between water and conduit walls, while a low St ($St \ll 1$) suggests inefficient (or ineffective) heat transfer. The thermal response observed at a spring, or along an underground river, largely depends on the proportion of flow through thermally ineffective and effective pathways (Luhmann et al., 2011).

Luhmann et al. (2011) analysed thermographs from 25 springs and recognised four different patterns (Patterns 1, 2, 3 and 4; Figure 11.2). They classified Pattern 1 and 2 as “ineffective” while Pattern 3 and 4 as “effective” in exchanging heat.

The four patterns in Figure 11.2.e are described and interpreted as follows:

- Pattern 1: This pattern (Figure 11.2.a) results from transient and punctual inflow during snowmelt or rain events. Water flows rapidly through the system at a different temperature from that of the aquifer and traverses the system before the thermal equilibrium is reached. The related thermographs show peaks (warm and cold) retained from the water input temperature signal. The thermograph given as an example is from Goliath's Cave (Luhmann et al., 2011). The monitoring station of Goliath's Cave is approximately 26 m below the surface in a stream that runs through the cave, which is mainly fed by a swallow hole 3 km away (Alexander Jr. et al., 2015; Luhmann et al., 2011). During flood events, to which Goliath's cave reacts rapidly in discharge, large temperature variations ($\gg 1^\circ\text{C}$), both positive and negative, are portrayed in the thermograph. In contrast, during the dry season, water temperature is flat and stable. Given the stable temperature observed during dry seasons, i.e. characterized by the absence of short (daily) and periodic (seasonal) fluctuations, such as in Figure 11.2.b and c (see Patterns 2 and 3 below), and the presence of temperature peaks caused by concentrated recharge and increased flow rates during wet seasons, the behaviour of Goliath's Cave (Figure 8.2.a) suggests the following: under low-flow conditions, the convection length is limited to the system length, resulting in no thermal signal at the spring. In contrast, during high-flow periods, the convection length extends beyond the system, allowing temperature peaks to appear at the spring. Specifically, this spring reveals characteristics of an intermediate deep/long aquifer.
- Pattern 2: Luhmann et al. (2011) relate this pattern (Figure 11.2.b) to springs fed by shallow aquifers primarily recharged by perennial sinking streams. In such systems, surface water (directly influenced by external temperatures) infiltrates underground and moves rapidly through the aquifer. The thermograph given as an example is from Vreeman Spring (Luhmann

et al., 2011). Vreeman Spring (Figure 11.2.b) displays a sinusoidal temperature signal with an amplitude of $\sim 4^{\circ}\text{C}$ and a phase lag of about 40 days relative to surface temperatures (Figure S0.1 in the Supplements). This behaviour reflects the influence of a shallow aquifer receiving concentrated recharge, where water travels from the swallow hole to the spring within a few days. Superimposed on this periodic signal are temperature peaks associated with faster, transient recharge events. The increased flow velocity, corresponding to a shorter travel time, yields an extension of the convection length (namely, it becomes longer than the system).

- Patterns 3 and 4: According to Luhmann et al. (2011), these two patterns (Figure 11.2.c and d for Patterns 3 and 4, respectively) occur when recharge water moves relatively slowly through the aquifer and has enough time to equilibrate with the temperature of the aquifer. For systems exhibiting these patterns, the average residence time of water is likely longer than that of flows producing Patterns 1 and 2. Krage Spring (Figure 11.2.c) is fed by a shallow system with a diffuse recharge through fractures (Luhmann et al., 2011). The thermograph from Krage Spring does not show any temperature peaks related to recharge events but shows a periodic (sinusoidal) signal with an amplitude slightly lower than 1°C and about 5 months out of phase with the surface temperature (the outside maximum temperature was observed in July, while at the spring it was observed in the following January; reference air temperature in Figure S0.1 in the Supplements; Luhmann et al., 2011). On the other hand, the diffuse recharge completely attenuates daily and event-scale temperature fluctuations. Although the thermograph in Figure 11.2.c is similar to that of Vreeman Spring (Figure 11.2.b) which is related to the heat retained by the sinking river water, for shallow systems with a diffuse recharge (such as Krage Spring in Figure 11.2.c), the periodic (sinusoidal) signal visible in the thermographs relates to conduction through rocks from which the water infiltrates (Luhmann et al., 2011). Since conduction solely from the surface cannot allow the temperature signal to penetrate further than a few metres into the rock ($\sim 3\text{ m}$), water convection within the fractures should also play a role in determining aquifer temperature (Luhmann et al., 2011).

In contrast, the graph displayed in Figure 11.2.d (Highway 76 MNDNR Spring) shows long-term temperature stability, over timescales of weeks to years. Luhmann et al. (2011) suggest that these differences between the two springs (Figure 11.2.c and Figure 11.2.d) stem from aquifer depth and rock temperature. If the aquifer is shallow enough that the surrounding rock temperature changes seasonally due to heat conduction from the surface, this may give rise to Pattern 3. Whereas, if the aquifer is deep and the external variations are completely attenuated by the rock (i.e., a constant temperature of the surrounding rock), the associated thermograph will result in Pattern 4.

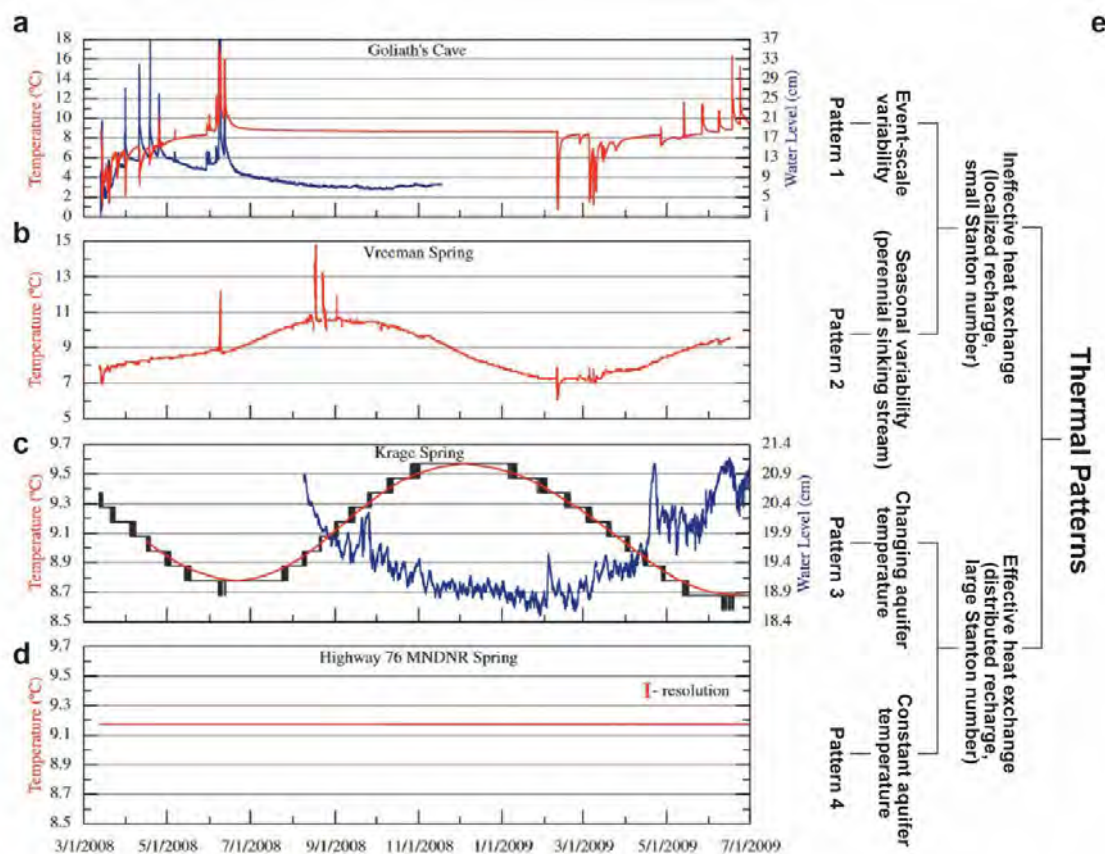


Figure 11.2 – From a) to d): spring thermographs presented by Luhmann et al. (2011) carried out from different springs in Minnesota (USA). The temperature responses during flood events have been summarised by the authors (Luhmann et al., 2011) in the flow chart in e), which groups them into different thermal patterns. The reference air temperature is given in Figure S0.1 in the Supplements. Copyright © 2010 The Author(s). Journal compilation © 2010 National Ground Water Association.

Luhmann et al. (2011) explained that Patterns 1 and 2 are classified as “thermally ineffective patterns” since they yielded $St \ll 1$, while Patterns 3 and 4 are “thermally effective patterns” with $St \gg 1$. Thermally ineffective patterns are usually considered as showing temperature variations at the event scale since they are produced by rainfall or snowmelt events carrying water into the system (such as flood events occurring at Škocjan-Kačna and Rouge Eau caves, Figure 10.6.a/b).

Therefore, the balance between localised and distributed recharge plays a crucial role in determining the nature of the thermal length. Luhmann et al. (2011) also underline that karst aquifers could include both thermally effective and ineffective flow. For instance, Figure 11.2.a (Crage spring) shows a constant temperature during a certain period of the year, and the same applies to the spring shown in Figure 11.2.d throughout the entire year. Therefore, the authors (Luhmann et al., 2011) conclude that the spring in Figure 11.2.a may comply either with Pattern 1 during flood events or with Pattern 4 during low-water periods, indicating a variation in the convection length.

In Figure 11.3 we show another thermograph which could comply with either Pattern 2 or 3. The water temperature was recorded in Milandre Cave (Jura Canton, Switzerland) at 40 m depth (510 m a.s.l.) beneath the surface, from a water tributary into the cave which further flows along a 5 km underground river (data series are made available online by the Swiss Federal Office of the Environment; FOEN, 2025; Boncourt - Milandre Amont station). Water attains this underground tributary after infiltrating through a catchment area characterised by a diffuse recharge and relatively deep unsaturated zone (between

40 and 80 m deep; Perrin et al., 2003). The thermograph from this cave (Figure 8.3) seems to show a yearly sinusoidal signal dumped at an amplitude in the order of 0.2°C and shifted by ~ 4 months with respect to the external temperature (Figure S1.2 in the Supplements). Above the periodic signal, temperature shows evident warm or cold peaks with temperature variations in the range of 0.1 to 0.4°C . Cold peaks took place between January and April 2022 and between March and May 2023, while warm peaks occurred from May 2022 until February 2023. Therefore, temperature peaks are slightly displaced compared with the external temperature, since cold peaks occurred mainly in winter and spring, while warm peaks took place mainly from spring until midwinter. This could probably be due to the residual thermal signal retained by the rock in the infiltration zone during warmer months of the year and advected by the water with a delay. Since the event scale temperature fluctuations as well as the periodic signal are in the order of a few tenths of degrees, the water convection length might be close to the system length. Yet, the classification of this underground spring is not straightforward concerning that proposed by Luhmann et al. (2011), since it is not clear if the system is either thermally effective or ineffective.

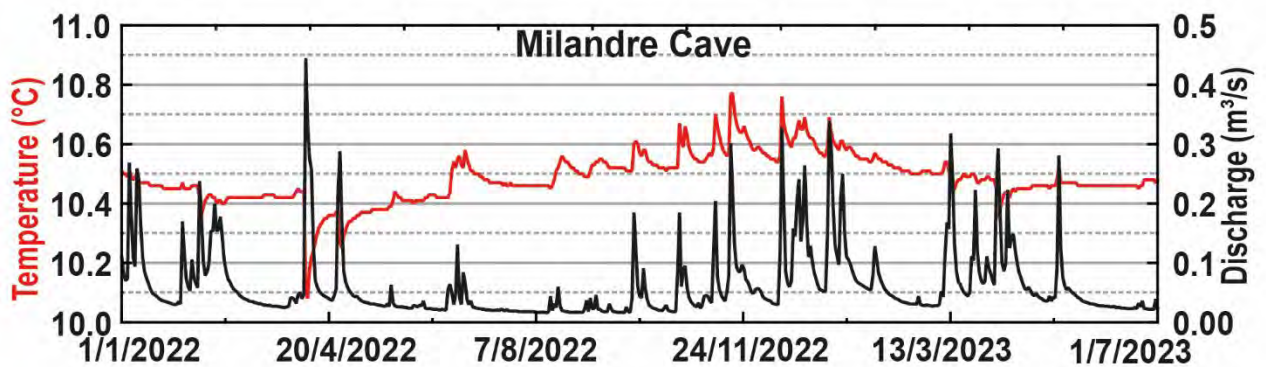


Figure 11.3 – Thermograph (in red) and discharge rate (in black) of Milandre Cave (Jura Canton, Switzerland). Temperature peaks are both negative and positive, which are in the order of some tenths of degrees. The reference air temperature is given in the Figure S0.2 in the Supplements. The data series is made available online by the Swiss Federal Office of the Environment (FOEN, 2025).

Furthermore, not all plausible natural situations seem to be included in Luhmann et al.'s (2011) classification. Other problematic scenarios to classify could be those springs that are recharged by systems with very large height differences (>500 m) between the catchment area and spring (deep systems). The case shown in Figure 11.4 is a thermograph from Heilongquan Spring (HLQ; South China) presented by Luo et al., (2023). According to Luo et al. (2023), the water recharge is mostly concentrated through swallow holes present in the catchment area (concentrated recharge), about 800 m above the spring.

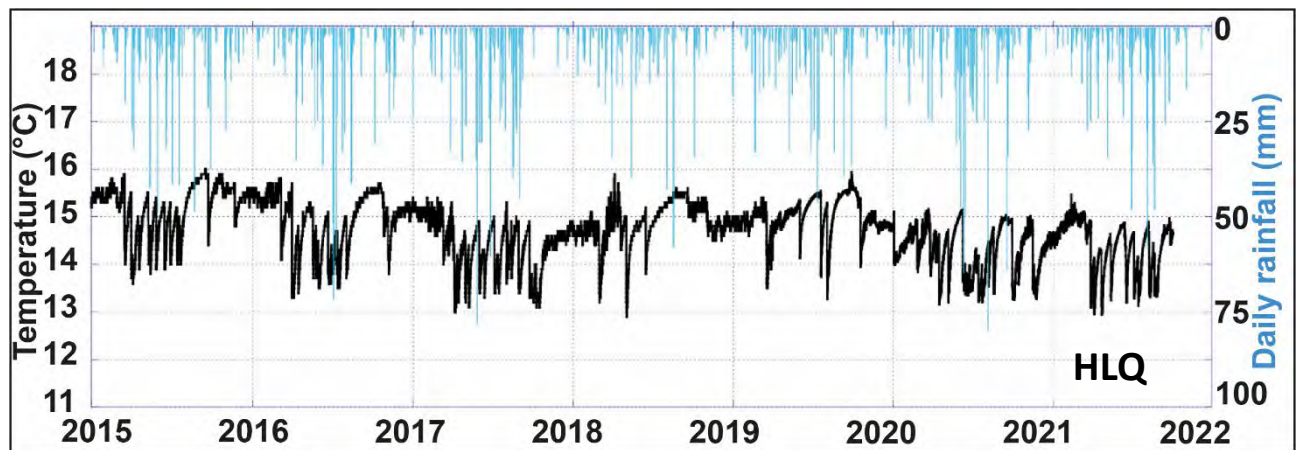


Figure 11.4 – Thermal response curve of Heilongquan (HLQ) spring from 2015 to 2021, presented in Luo et al. (2023). The reference air temperature is given in Figure S0.3 in the Supplements. Reprinted from Luo et al. (2023). Copyright (2023), with permission from Elsevier.

Figure 11.4 shows that, although temperature variations during flood events take place, these always produce negative peaks (decreasing temperature) independently of the season during which they happened. The distance between the catchment area and the spring should be large enough to allow water to equilibrate with the surrounding rock for daily and transient-event signals. Luo et al. (2023) explain that, during flood events with increased recharge and flow velocity, the infiltrating water has no time to equilibrate its temperature throughout the system, and it reaches the spring with a lower temperature than expected at the spring's altitude. But it does so independently if the initial temperature pulse is due to cold or warm water. This suggested to these authors that somehow a certain equilibrium with the rock is reached, although the water temperature matches that of the rock ~300 m above the spring.

We can reasonably suppose that the same happens in Schlichenden Brönnen Spring (SBS, Figure 11.5), another deep system but with a diffuse recharge, that opens in the Schwyz Canton (Switzerland). The karst system feeding the SBS spring (640 m a.s.l.) has a maximum elevation difference of about ~1700 m with the recharge diffused over a 32 km² catchment area (Jeannin, 2016). Independently of the season, the thermograph of SBS (Figure 11.5; data series are made available online by the Swiss Federal Office of the Environment; FOEN, 2025; Schlichenden Brönnen station) shows negative peaks generally included between ~0.2 and ~0.4 °C and matching with increased flow rates. On the other hand, the largest temperature decrease took place between April and June (for both 2021 and 2022) during the snowmelt and when the water infiltration is very high, causing water temperature to drop by about 0.8 °C.

The HLQ shows temperature variations $\geq 1^{\circ}\text{C}$, which are much larger than those observed at SBS ($< 1^{\circ}\text{C}$). The diffuse recharge of the SBS system ensures that the temperature variations stay in the order of some tenths of a degree, in contrast with HLQ whose recharge is concentrated and temperature variations are larger. In both systems, the vertical distance between the infiltration zones and the springs is expected to be longer than their respective convection lengths. However, an increased flow rate and velocity during floods result in negative peaks on thermographs. In vertical systems with rapid water circulation, these peaks have an amplitude which depends on the water velocity and then on the flow rate. In very deep systems, external temperature signals are fully buffered during both low and high flow periods, due to the system's length being sufficient to attenuate daily temperature variations.

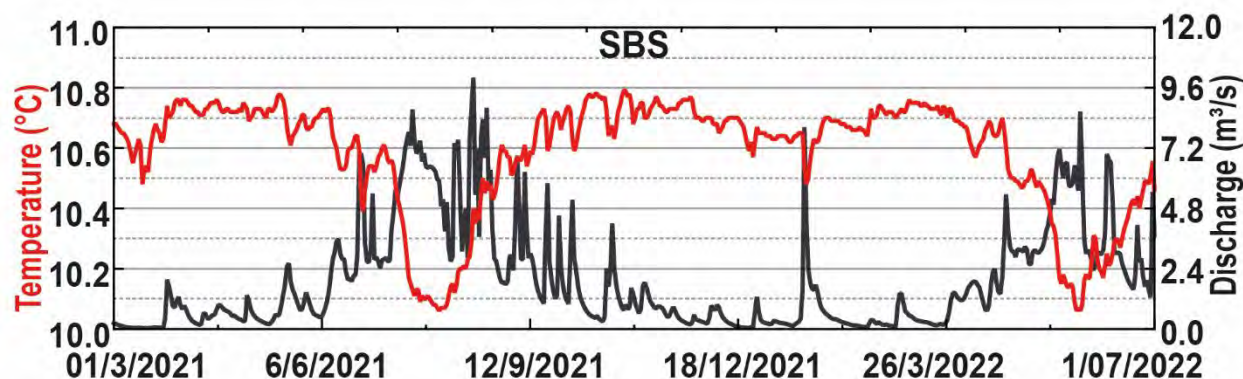


Figure 11.5 – Thermograph (in red) and discharge rate (in black) of Schliechenden Brünnen Spring (SBS; Schwyz Canton, Switzerland). Note that the temperature peaks are always negative and in the order of some tenths of degrees. The reference air temperature is given in Figure S0.2 in the Supplements. The data series is made available online by the Swiss Federal Office of the Environment (FOEN, 2025).

Nevertheless, during periods of high flow, their depth is a distance not sufficient to achieve complete water/rock equilibrium throughout the system. This results in transmission to the spring of temperature peaks related to the rock temperature situated at higher elevations, and then colder than expected at the spring. As a result, temperature peaks may reach the spring, reflecting the rock temperature at higher elevations, which is colder than the expected temperature at the spring.

Therefore, are these springs related to an effective or ineffective system? What should be their classification according to Luhmann et al. (2011)? In our opinion, Luhmann et al.'s (2011) classification of thermal responses of karst springs can be rearranged to give a more direct indication of the recharge mode and depth of the system. Our classification (Figure 11.6) is based on two criteria: 1) the recharge mode (x-axis, from diffuse to concentrated) and 2) the system depth (y-axis, subcutaneous, shallow, intermediate and deep).

11.4.2. Proposed classification

The recharge mode directly influences the convection length: a diffuse recharge tends to shorten the convection length; conversely, a concentrated recharge tends to extend the convection length.

Therefore, in the proposed classification (Figure 11.6):

- on the right side of the x-axis are classified systems with a significant part of concentrated recharge, i.e. for which the convection length is longer than the distance between the catchment and the spring, and temperature variations are significant ($>1^{\circ}\text{C}$). On the left are arranged systems with diffuse recharge, for which the convection length is shorter than the distance between the catchment and the spring, i.e. where temperature variations are small ($<1^{\circ}\text{C}$);
- the vertical axis represents the effect of the thickness of the vadose zone: 1) subcutaneous systems show an annual cycle which is shifted compared to seasons. As the top of the phreatic zone is close to the ground surface, the aquifer is reached by the annual fluctuation due to heat conduction through the rock. 2) Shallow systems have the unsaturated zone ranging between ~ 15 and ~ 100 m, without any clear annual cycles due to conduction, but with small (in the order of few tenths of degrees) positive and negative peaks according to whether the input is

of warm or cold water. 3) systems between 100 and 500 m depths and that show close to nil temperature fluctuations. 4) Deeper systems (≥ 500 m) display only negative variations with a rather low amplitude. The given system depths are in order of magnitude; thus, they may largely vary for specific situations.

The classification proposed in Figure 11.6 is based on the thermographs of: 1) four springs published by Luhmann et al. (2011) (Goliath's Cave, Vreeman, Krage and Highway 76 MNDR springs located in Minnesota), 2) Heilongquan Spring (HLQ, South China) from Luo et al. (2023), 3) Pollaccia and Equi Terme springs (Apuan Alps, Tuscany; IT) from Nannoni and Piccini (2022) and Piccini et al. (2023), respectively, 4) Milandre Cave (Jura Canton, CH), 5) Schliechenden Brünnen Spring (SBS; Schwyz Canton, CH). These springs lie in different world regions (Minnesota, Switzerland, Italy and South China) and climate conditions. However, the comparison between thermographs in Figure 11.6 is allowed since air temperatures shown in the supplementary images (Figure S0.1, Figure S0.2, Figure S0.3 in the Supplements) display a climate situation with four seasons for all considered cases. For all the climate situations, the yearly fluctuation oscillates between cold and warm temperatures, with a shift above the 0°C for South China (Figure S0.3). The precipitation amounts are also comparable between the different sites, being ~900 mm/yr for Minnesota, between ~1000 and ~2000 mm/yr in Switzerland (in the vicinity of Milandre cave and SBS, respectively; Jeannin, 2016; Perrin et al., 2003), from 1900 to 2800 mm/yr in the area of the Apuan Alps in Italy (Piccini et al., 2023) and from 900 to 1200 mm/yr for South China (Luo et al., 2023). In contrast, extreme climates showing very low temperature variability during the year (e.g. tropical or coastal climates) are not considered in this first classification attempt. We anticipate that water temperature at these springs would also show limited variability during the year, regardless of the recharge modality (diffuse or concentrated).

The vertical axes of graphs in Figure 11.6 (from a to h) display the ΔT , which indicates the temperature peaks due to an increased water flow. For the graphs Figure 11.6.a/c/e/g (diffuse recharge), the ΔT scale is in the order of 0.1°C, whilst for Figure 11.6.b/d/f/h (concentrated recharge) it is in the order of 1°C.

It is worth enough to note that thermal behaviours similar to those presented here may occur, but due to factors not strictly related to climatic conditions but rather to geological situations, such as mixing with a thermal water component or the effect of geothermal flow. In general, exceptions to the proposed classification may exist and should be assessed on a case-by-case basis. Additionally, intermediate or mixed cases between the outlined categories are also possible. In the following, we detail the categories depicted in Figure 11.6.

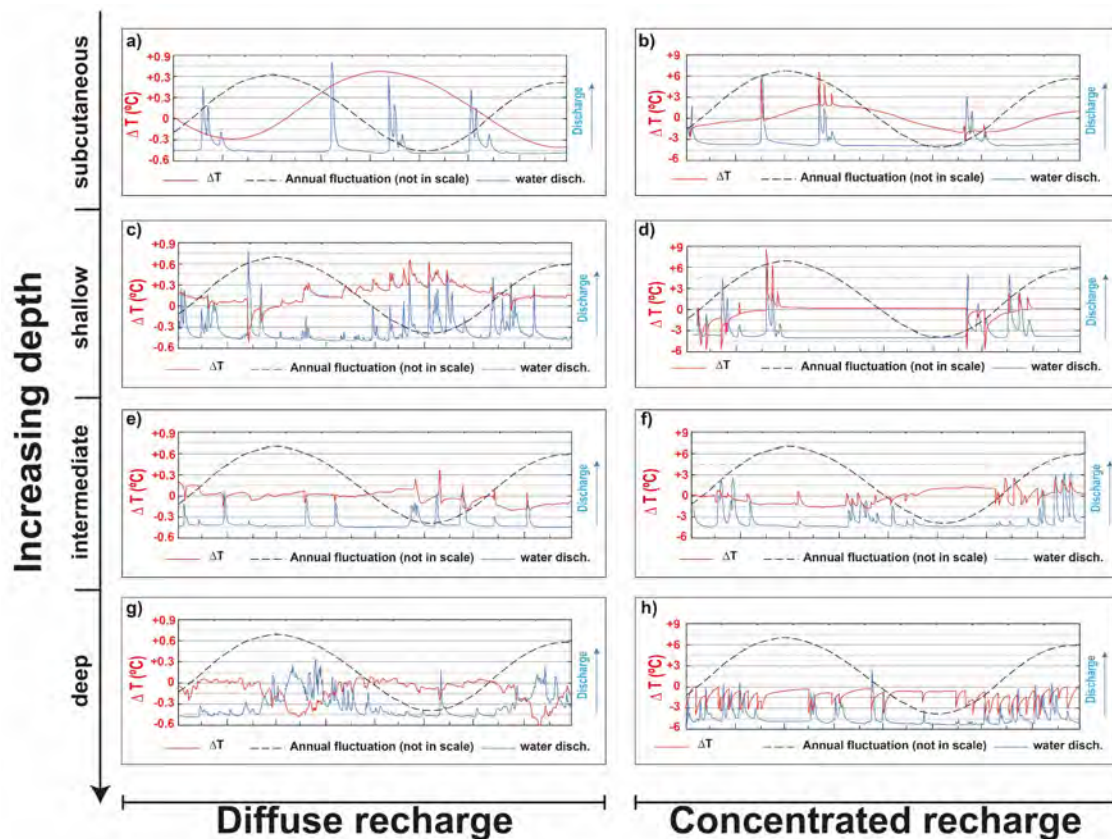


Figure 11.6 – Suggested classification of spring thermographs. The criteria considered are 1) the recharge mode (horizontal axis, diffuse and concentrated) and 2) the system depth (vertical axis, from subcutaneous to deep). In red, ΔT represents temperature variations. Black dashed line, outside annual air fluctuation, *not in scale*. In blue, water discharge.

- **SUBCUTANEOUS SYSTEMS:**

We include in this category systems with a depth in the order of ~ 10 m or lower. The thermographs are shown in Figure 11.6.a and Figure 11.6.b and they are representative of subcutaneous systems with diffuse or concentrated recharge modes, respectively.

Subcutaneous systems with diffuse recharge (Figure 11.6.a) show thermographs with a periodic (sinusoidal) signal with an amplitude lower or close to 1°C and a time delay of several months (typically from 5 to 6) out of phase with the surface temperature (i.e. the outside maximum temperature summer, while at the source spring it was observed in the following winter). The diffuse recharge allows the attenuation of daily and event-scale temperature fluctuations (no temperature peaks). For subcutaneous systems, the periodic (sinusoidal) signal visible in the thermographs is, therefore, related to the seasonally shifted temperature of the rock surrounding the aquifer. Seasonal temperature signals at the surface are transmitted into the subsurface as attenuated, out-of-phase fluctuations, primarily through conduction and, to a lesser extent, through water convection in fractured media. The water infiltrating via fractures and porous flows slowly enough to reach equilibrium with the rock, but it should be turbulent enough to allow the rock temperature signal to reach the spring (Luhmann et al., 2011).

Subcutaneous systems with concentrated recharge (Figure 11.6.b) show a sinusoidal signal but with an amplitude larger than 1°C with superimposed peaks. Since they are fed by a perennial sinking stream (concentrated recharge) which flows until the spring within days, the sinusoidal

signal, although similar to that of subcutaneous systems with a diffuse recharge which is due to the conduction through the rock, is instead related to the input water temperature and its short travel time (days) within the system (Luhmann et al., 2011). Therefore, spring temperature is slightly out of phase (in the order of 1 or 2 months) with the surface temperature (in contrast with the several months typical of those systems with a diffuse recharge). The water temperature of the perennial sinking river is fully dampened for daily fluctuations, provided that the flow velocity is low enough to ensure a travel time sufficient to attenuate short-term variations. However, the depth (length) of the system is probably too short to attenuate completely the yearly fluctuation. In contrast, during flood events, the concentrated recharge allows a sudden surge of the inrush flow rate during precipitation, causing an increase in the flow velocity (Luhmann et al., 2011). The increased velocity, corresponding to a shorter travel time than days, yields an extension of the convection length (namely, it becomes longer than the system), and temperature cold/warm peaks, according to the season and related to flood events (event scale signals), are superimposed on the periodic signal.

- **SHALLOW SYSTEMS:**

We include in this category systems with a depth between ~15 and ~100 m. The thermographs depicted in Figure 11.6.c and Figure 11.6.d are representative of shallow systems with diffuse or concentrated recharge modes, respectively.

Systems at shallow depth with a diffuse recharge (Figure 11.6.c) are characterised by temperature warm or cold peaks with temperature variations in the range of a few tenths of degrees (e.g. in Milandre cave, peaks are between 0.1 and 0.4°C, Figure 11.3). Temperature peaks could be displaced compared with the external temperature (cold peaks could occur at the end of winter and in spring, while warm peaks could take place at the end of summer and in winter). This behaviour is likely due to the delayed advection of the external thermal signal stored in the rock of the infiltration zone. Moreover, for this setting, there could also be a residual shifted (4 to 6 months) periodic (sinusoidal) signal transmitted through the rock by conduction, showing an amplitude $<<1^{\circ}\text{C}$, much more dampened than in the case of subcutaneous systems (Figure 11.6.a). Since the event scale temperature fluctuations as well as the periodic signal are in the order of a few tenths of degrees, the water convection length might be close to the system length.

Shallow systems with a concentrated recharge (Figure 11.6.d) are mainly fed by swallow holes. The springs react rapidly to temperature changes and discharge variations during flood events (rainstorms or snowmelt). During these events, large temperature variations ($>>1^{\circ}\text{C}$ and up to $\sim 10^{\circ}\text{C}$ in some cases, e.g. Figure 11.2.a), both positive and negative, are portrayed in the thermographs. Shallow systems with concentrated recharge could not exhibit periodic fluctuations, provided that the distance between the recharge and discharge points is sufficient to dampen these thermal signals. Specifically, the convection length is reached under low flow conditions (stable temperature), while during high flow rates, it extends beyond the system, resulting in temperature peaks at the spring.

- **INTERMEDIATE-DEPTH SYSTEMS:**

In this category, we include springs fed by systems with depths ranging between 100 and 500 m (Figure 11.6.e/f). We use as reference thermographs those derived from Equi Terme and Pollaccia springs (Italy; Nannoni and Piccini, 2022; Piccini et al., 2023), for diffuse and concentrated recharge,

respectively. These springs are fed by catchment areas ~ 500 m above the spring locations or slightly larger.

Systems within this depth range may have lengths comparable to their convection lengths, which in turn depend on the flow rate. Spring temperatures are expected to be relatively stable, and thermal peaks should exhibit smaller amplitudes compared to those observed in shallow systems or deep systems. Thermographs from these springs may also exhibit mixed characteristics between shallow-system (positive and negative low-amplitude peaks) and deep systems (only negative peaks), according to depth and flow rate. Systems fed with a diffuse recharge of this category should not show seasonal fluctuation (or a really small one) since they should be deep enough to completely attenuate this oscillation (Figure 11.6.e). While in the case of concentrated recharge, seasonal fluctuation might still be visible ($<1^{\circ}\text{C}$; Figure 11.6.f). For both cases, short-term surface events produce temperature peaks (both positive and negative), in the order of a few tenths of degrees ($\sim <0.2^{\circ}\text{C}$) for diffuse recharge and $\lesssim 0.5^{\circ}\text{C}$ for concentrated recharge.

- **DEEP SYSTEMS:**

This category includes systems with a depth of ≥ 500 m. The thermographs depicted in Figure 11.6.g and Figure 11.6.h are representative of deep systems with diffuse or concentrated recharge modes, respectively.

Springs fed by very deep systems (≥ 500 m) show thermographs with negative peaks regardless of the season (e.g. HLQ and SBS in Figure 11.4 and Figure 11.5, respectively), matching with an increased flow.

During flood events, deep systems fed by a diffuse recharge (Figure 11.6.g) have temperature variations in the order of a few tenths of degrees ($<<1^{\circ}\text{C}$, usually between 0.2 and 0.4°C). On the other hand, deep systems, fed by a concentrated recharge (Figure 11.6.h) with high flow velocity, always show decreasing temperature during rainfall, independently of the season, but with temperature variations $\geq 1^{\circ}\text{C}$.

Under low flow conditions, the convection length for daily temperature is reasonably shorter than the system, as daily signals from the input are dampened along the flow path. Although the circulation distance between the infiltration zone and the spring is long enough to allow full damping of daily fluctuations, high flow rates (and high water velocity) can prevent the water from fully reaching the rock temperature in the deeper parts of the system. It tends instead to show temperatures closer to those of the rock higher up in the system. This leads to always negative peaks in the thermograph, since cave water temperature has lower temperatures than expected for spring's altitude, regardless of whether the input signal was warm or cold. This suggests that, somehow, a certain equilibrium with the rock is attained, although this temperature is that of the system zones at higher altitudes than the spring.

11.4.3. Discussion on the proposed classification

The proposed classification is meaningful since:

- It is a practical way to visualise the thermal response of springs together with the depth/length and the recharge mode of the feeding system;

- It reflects whether the system is shorter or longer than the convection length during periods of low and high water flow.

At first glance, the classification in Figure 11.6 reveals that temperature variations on the left side of the graph (diffuse recharge) are on the order of tenths to one degree, while on the right side (concentrated recharge), they reach several degrees. It also appears that thermographs have similar shapes for systems of the same depth (vertical axis). Moreover, this classification respects the subdivision between effective and ineffective systems suggested by Luhmann et al. (2011), which is based on the Stanton number (St). As aforementioned, this dimensionless number describes the efficiency of heat exchanges between a fluid and the media in which it flows. The highest values of St ($>>1$) are typical of a porous matrix in which the fluid flows slowly, while fractures show intermediate St values, whereas conduits have the lowest St ($<<1$) (Luhmann et al., 2011). Therefore, St decreases with an increasing flow velocity and dimensions of the flow paths (from matrix to conduits). In our classification (Figure 11.6) the St tends to decrease from the left to the right side of the x-axis, namely from systems with diffuse recharge (more effective) toward systems with concentrated recharge (less effective). In fact, the flat thermograph associated with Pattern 4 (Figure 11.2.e) represents an end-member system characterised by very diffuse recharge (rather fissured than karstified) and an intermediate deep aquifer (Luhmann et al., 2011).

This classification is a first attempt and should be further refined. Moreover, the detailed characteristics of the systems used as examples are not sufficiently well-known, and other thermographs should be compared with this classification. Numerical modeling of examples in the respective categories can help to validate the proposed hypotheses.

Moreover, the proposed classification can be expanded to include other situations more complex or also rebuilt with thermographs from other climatic regions. In fact, the main weakness of this classification is that the thermographs chosen from the literature are not representative of all the possible thermal situations and are specific cases. We are aware, and agree with Luhmann et al. (2011), that more thermal patterns can be observed in other karst climates. Especially in extreme cases where the climate presents very low-temperature variability during the year (e.g. tropical climates), the water temperature at springs could also show limited variability during the year, regardless of whether the recharge is diffuse or concentrated.

Although this classification could be limited to specific situations, it is believed to provide a new key to understanding the thermal patterns of karst springs. Further discussions will certainly help to improve this proposition.

Chapter 12. Conclusions

This thesis, together with that by Sedaghatkish (2025), represents a significant step forward in the understanding of the heat transfer within karst systems. Measuring temperatures and airflow in caves is of interest to study the air dynamics and the related energy exchanges. Indeed, in this thesis, we mainly focused on the thermal effects of ventilation through the karst vadose zone, although we also introduced a qualitative discussion on the effects of water flows in Chapter 10 and Chapter 11.4.

This work evolved from simplified theoretical models to an *integrated monitoring and modeling approach* that coupled fluid (air) dynamics with rock heat transfer. In this work, we could improve sensors and data loggers to carry out reliable cave data, putting particular focus on the assessment of air flow rates. Recorded data of airflow and temperatures allowed us to validate sophisticated numerical models. Pioneering models of heat and mass transfer along ventilated conduits, such as that of Wigley and Brown (1971), ignored the influence of air flow on the rock temperature. Instead, the research we presented in Sedaghatkish (2024), in agreement with Lismonde (2002) and Gabrovšek (2023), has shown that air flow affects the temperature of the rock, which in turn changes the air temperature profile in the cave. A lack of data was the main gap to validate previous models, so cave monitoring revealed essential information. Good cave data and numerical modeling presented in this work made it possible to show the existence of the so-called “thermal anomalies”, namely the *Ventilated Warm and Cold Zones* (highlighted in Figure 11.1), to assess the convection length and to show that 90% of the energy exchange between air and rock occurs in the areas near the entrances.

Whereas “classic” thermal zonation (Luetscher and Jeannin, 2004; Jeannin, 1990; Drogue, 1985; Pulina, 1969) qualitatively distinguished between *heterothermic* (or *dynamic*) and *homothermic* (or *static*) zones based on depth and on temperature variations $<0.5\text{--}1^{\circ}\text{C}$ (cfr. Chapter 1.2.2), this work introduces the physical parameter of (annual) *convection length*, namely the length of the heterothermic zone and where convective heat exchanges between air and rock dominates heat transfer. This distance defines how far external temperature signals transported by the air flow penetrate a ventilated cave (under a chimney effect) before being attenuated of 90%. It has been shown that this length is *approximately proportional to the amplitude of annual airflow fluctuations divided by the square root of the mean radius of the cave* (Chapter 7; Sedaghatkish, 2024). Since the chimney effect typically develops in caves with multiple entrances at different elevations, each entrance is characterised by its own *convection length* (warm or cold, depending on whether it is an upper or lower entrance). Therefore, in short ventilated caves, such as Longeague (Chapter 8), where the two cave entrances are “only” 300 m from each other, the deep homothermal zone may be completely absent if the length of the cave is comparable with its *convection lengths* (between 100 and 150 m for Longeague). In contrast, the caves we monitored to develop the model presented in Chapter 7 have entrances far enough to allow the presence of a *deep karst zone* (or homothermic zone).

In Chapter 1.2 – “*Cave climate and meteorology*”, it was distinguished between climatology and meteorology. According to Badino (2010) and Covington and Perne (2015), climatology deals with “*average atmospheric conditions*” (such as temperature, humidity, pressure, rainfall, etc.) that vary slowly over time, while meteorology studies “*how these conditions fluctuate on short time scales around this average*”, i.e. meteorology deals with daily or seasonal weather. In practice, for the scientific community, the climate of a specific place is the average of the atmospheric conditions over at least 30 years (IPCC, 2023). Therefore, the difference between climatology and meteorology, as the sciences studying the climate and the weather, respectively, relates exclusively to the observation time.

But this definition could be inadequate for the cave environment. Given the minimal fluctuation in atmospheric variables in caves compared to external variations (e.g. tens vs tenths of degrees for outside and inside atmospheres), a period of 30 years may not really be meaningful to make this distinction. On the other hand, cave meteorology could be limited to the annual *convection length*, i.e. where fluctuations can still be measured on annual time scales. However, distinguishing between cave meteorology and climatology is probably less relevant than recognizing the annual *convection length* in caves, which is relevant for paleoclimatic studies, for instance.

We highlighted in Chapter 7 that ventilation acts as a “rapid vector” since external thermal perturbations penetrate much faster and deeper by air convection than they would by only conduction through the rock. This may have an impact on cave deep part temperatures on time scales in the order of decades or more, then relating more to climate than meteorological changes. Obleitner et al. (2024) highlighted that changing the observation time, i.e. moving from annual to multi-decadal temperature data series, temperature variations can also be observed in cave parts which might be considered remote and with stable temperature. Obleitner and coworkers observed that average temperatures in remote parts of ventilated caves in the Alps increased by about 0.2 °C over twenty years of monitoring. This increase of a few tenths of degrees is assumed to be driven by external warming. The ~15 years delay compared to external warming cannot be triggered by conduction through the rock only, and is rather assumed to correspond to the *convection length* for one or two decades. According to Obleitner et al. (2024) and the results shown in this work, this propagation depends on cave geometry as well as on the interaction between ventilation and water infiltration. Paleoclimate studies based on speleothems usually focus on samples from the inner parts of caves because of stable temperatures, while much less is known about how rapidly external atmospheric warming or cooling is transmitted into caves (Obleitner et al., 2024). Such delays are not critical for studies on long timescales (e.g. orbital scales). Nevertheless, they become important for high-resolution analyses using, e.g. fast-growing or annually laminated speleothems (Obleitner et al., 2024). Although it is beyond the aim of this work to provide a definition of cave climate and meteorology, we suggest keeping in mind that cave ventilation has a clear effect on how external climatic variations are transferred into caves in terms of distance, amplitude, and time scale. Assuming a heat transfer from the surface through the rock solely by conduction is a crude simplification.

For understanding heat exchanges between outside air flowing through cave systems, we used a cave ventilation model. The ventilation model representative of a ventilated cave (under a chimney effect) needs the assessment of the driving pressure (**DP**) and the cave aeraulic resistance (**R**). Multi-year temperature monitoring allowed the estimation of **DP** (Pa), which is derived from ΔT (between the inner and outer atmospheres) and cave geometries (length and slope of conduits). Additionally, we used airflow monitoring to infer the aeraulic resistance **R**, a geometric parameter that does not depend on the time of observation and that can prove useful in cave climate studies. Indeed, this work proposes to infer resistance **R** directly from field data (temperature and flow rate), allowing for the development of more reliable ventilation models (Chapter 8), e.g to create models with different climate scenarios (varying **DP**). Future work may focus on the **R** parameter to better understand whether and how it varies along a ventilated conduit, and, eventually, if it represents punctual head losses or the average of head losses along the cave. Atkinson et al. (1983) highlighted that the resistance of an entire ventilated gallery appears to be determined by the narrow and short passages (compared to the total length) that can be found along a cave. This suggests that this parameter expresses the higher value of **R** in a ventilated conduit. However, it has also emerged in this work (Chapter 8.4.1) that ventilated

caves can behave like a “Tesla Valve”, exhibiting different resistances depending on whether the flow is downward or upward. This rather suggests that R is an average value of the cave resistance. For a better understanding of this asymmetric behaviour, a more detailed monitoring and modeling would be necessary. Right now, we can just keep in mind this fact and determine one aeratic resistance for each flow direction. In our opinion, to simplify models of future 3D numerical modeling requiring aeratic resistance, it is worth exploring this analogy with a Tesla valve.

Average cross sections and average diameters are also necessary parameters when one wants to model realistic ventilated conduits. In Chapter 5, we have shown that CO₂ gauging tests revealed a good method to assess the average areas and diameters of cross-sections along ventilated conduits, air flow rates and average velocities, as well as to infer the dispersion coefficient (necessary for 1D advection-dispersion model). The latter parameter is necessary when one wants to use the 1D advection-dispersion model. The study of dispersion along ventilated conduits is of interest for the transport of gases (e.g. CO₂ and Radon) and aerosols (e.g. pollutants, smoke, fumes, fine dust, bacterial and fungal spores, etc.). Overall, dangerous gases, fumes and smoke can threaten the safety of workers operating underground (e.g. miners and guides in tourist caves), while the transport of e.g. aerosols, pollutants and bacteria can damage fragile ecosystems or cave paintings, especially in tourist caves (Dredge et al, 2013). Through the injection and gauging of artificial CO₂ in ventilated caves, we have shown that 1D advection-dispersion is not sufficient to explain the tailing characterising tracer breakthrough curves (BTC). Then, for ventilated caves, more complex models need to be developed (e.g. numerical simulations of 3D Navier-Stokes equations) to investigate solute dispersion. On the other hand, for more regular conduits (i.e. constant cross-sections), like naturally ventilated mines, the 1D relation is still valid. Despite this, the theoretical relation proposed by Taylor (1954) largely underestimates the dispersion coefficient along naturally ventilated conduits, whereas Chatwin's (1971) geometrical method proved to be more reliable.

Finally, based on the knowledge gained by studying thermal interaction between air and walls along ventilated caves, we qualitatively transferred the concepts of *convection length* to hydrologically active caves. Water and cave air temperatures are intimately linked along the “active” vadose zone (Luetscher and Jeannin, 2004; Badino, 2004). While air dominates in dry (without flowing water) parts of the vadose zone, water may become the main heat vector in active systems, transporting external temperature signals over much greater distances (hundreds of metres or kilometres; Covington et al. 2012) due to its high heat capacity (four times higher than that of air). As we have shown in Chapter 10, along the *convection length* of caves with significant water flow (e.g. swallow holes), the air temperature was found to be directly controlled by water temperature, as already highlighted by Jeannin (1990). Although water and air may show close temperatures after a relatively short distance in the cave, water temperature oscillations (e.g. temperature peaks during floods) due to external variations can sometimes be detected at karst springs. When this happens, it indicates that the water *convection length* is longer than the distance between the infiltration point and the spring. Therefore, estimating the water *convection length* also helps in understanding where and how infiltrating water exchanges heat along the massif. Moreover, since temperature peaks serve as a natural non-conservative tracer (Luhmann, 2015; Anderson, 2005), the interpretation of spring thermographs allows us to study thermal signal propagation in water and to infer important system properties (e.g., recharge mode and depth). Although the classification of thermal response at karst springs proposed in Chapter 11.4 might still be incomplete and needs data for more comprehensive case studies, we consider it a good way to highlight the link between water *convection length*, system recharge mode

(diffuse or concentrated) and depth. Indeed, in agreement with Luhman et al. (2011), duration, order of magnitudes (degrees or tenths of degrees), and sign (positive or negative) of water temperature peaks during floods provide information about these characteristics of the system and on how the *convection length* expands during flood events. In fact, an increased water flow rate brings external heat much deeper into the massif than during periods of low water flow. Therefore, in karst environments, rock, water and air are thermally woven and open questions remain on quantifying the relative importance of heat transported by air compared to that transported by water and by conduction through the walls. Although our analysis of water *convection length* is only qualitative and should be completed using quantitative data from multiple springs in different geological settings and climates, as well as verified against numerical modeling coupling heat transfer in water, air, and rock, it provides a final piece to the ThermoKarst framework, which has yielded new insights into the “thermal structure” of karst massifs.

In conclusion, ventilation is a key factor in creating and maintaining the specific environmental conditions that characterise caves, which are fragile ecosystems. Understanding cave ventilation is key for paleoclimatic models and interpretations, as it controls the deposition and chemistry of paleoclimatic proxies (e.g. speleothems). Moreover, ventilation dynamics are fundamental for predicting cave response to climate change. Ventilation also influences underground habitats, as it controls environmental factors that are critical to subterranean organisms (e.g. temperature and humidity) and, finally, it carries airborne nutrients supporting the underground trophic chain. Climate change as well as structural cave changes (e.g. collapses changing the aeraulic resistance) may alter ventilation characteristics, with potential consequences for the biodiversity and functionality of these fragile environments. Future research should focus on understanding the complex interactions between ventilation, climate, and cave biology to develop suitable evolution models.

12.1.1. Summary of the main conclusions

To punctually answer the questions that motivated the research and are listed in Chapter 1.4, the main conclusions can be summarised as follows:

- Field measurements and cave monitoring proved to be essential for a quantitative assessment of air heat exchanges, for inferring physical parameters of caves (e.g. cave aeraulic resistance, dispersion, geometries) and for testing the reliability of numerical models.
- In Chapter 4 (“*Monitoring air fluxes in caves using digital flow metres*”), Chapter 5 (“*Dispersion of artificial tracers in ventilated caves*”) and Chapter 6 (“*Comparing different methods for measuring airflow*”), we used and compared different methods for measuring airflow within caves. We have shown that it is not easy to measure a cave airflow with an accuracy better than $\pm 20\%$ to $\pm 30\%$ (more likely this uncertainty is in the order of $\pm 50\%$). In caves, these large uncertainties partly relate to the measurement methods and the devices used. But, they also depend on the instrument position relative to the distribution of wind velocity fields within the cross-section. Our comparison between different methods for measuring airflow (Chapter 5 and Chapter 6) has shown that the gauging method with a tracer gas is probably the most accurate way to obtain a one-time assessment of an airflow rate since it does not require velocity integration over a passage cross-section. However, the assumption of air mass conservation can be a limitation of the method, especially over long distances and in complex systems. Nevertheless, comparing the results from the different measuring methods allows one to assess the upper and lower bounds of uncertainties to fix the order of magnitude of the

real airflow. In future works, carrying out several gauging tests under different airflow conditions could help in yielding a rating curve to calibrate direct and punctual measurements carried out with hand-held or in-place anemometers.

- In the karst vadose zone, cave ventilation efficiently transports heat and aerosols, influencing particle distribution (Dredge et al., 2013). Modeling this transport requires parameters such as cross-sectional areas, airflow, and dispersion coefficients. In Chapter 5, experiments using CO₂ as a tracer proved effective for estimating conduit cross-sections and inferring dispersion values. Taylor's (1954) theory significantly underestimates the dispersion coefficient deduced from the breakthrough curves (BTC), whereas applying Chatwin's, (1971) method seems more appropriate. BTC analysis revealed tailing in natural conduits (caves), likely due to cross-section variations, and the presence of dispersive singularities and chambers. Such enlargements promote aerosol deposition due to low velocity. The 1-D advection-dispersion model worked well in quite regular conduits like mine tunnels, while it failed in cases with less regular geometries, such as caves. Therefore, more complex numerical simulations (e.g. 3-D Navier-Stokes equations) will be necessary to investigate solute dispersion, transport and deposition in caves. It was attempted for water flow (Hauns et al., 2001), but similar work could be done for airflow.
- In Chapter 8 (*"Dynamics of chimney-effect ventilation and heat exchange: a model from Longeigue Cave"*), we demonstrate that ventilation exerts a primary influence on the temperature of a ventilated cave and the surrounding rock. Cave ventilation significantly increases the depth of thermal fluctuations and decreases the time of propagation (Sedaghatkish et al., 2024). In caves dominated by a chimney effect, the extension of ventilated warm and cold zones (close to upper and lower entrances, respectively) relates to the *convection length* (Chapter 7), which is approximately proportional to the amplitude of the annual flow rate fluctuations over the square root of the cave's average radius. The convection length is often in the order of ~100 m for metric conduits (Table 8.6 in Chapter 8) but can reach kilometres for very big, smooth (typically tunnel) and strongly ventilated conduits, and reduce to the diffusion length (5-10 meters) for small conduits. The assessment of heat exchanged between the air and the rock along a ventilated cave (Chapter 8) shows that ~90% of the heat transfer between air and rock takes place along the the first tens of meters (50 m for Longeigue Cave) of the convection length, while the host rock plays a major role in dulling the external fluctuations with distance along the conduit. Moreover, in Chapter 8, we have also shown that the deep karst zone may not always be present in ventilated caves shorter than or close to the convection length, such as Longeigue Cave.
- Intermittent ventilation behaviour (open and closed period, **OS** and **CS**, respectively) affects air heat exchange processes. When a system is open and ventilated, the convective heat exchanges dominate, whereas during closed periods, without ventilation, convective heat exchanges reduce. In Longeigue Cave, **OS** and **CS** relate to temporary closures of passages due to water table fluctuations linked with external precipitations. Therefore, external precipitation indirectly controls cave ventilation. Our results have shown that little changes in **OS** and **CS** distribution over the years (i.e. depending on whether closures happen mainly in summer or winter) led to significant changes on annual heat balance of the *Ventilated Warm and Cold Zones* in the cave (Chapter 8).

- In Chapter 8, we have demonstrated that an aeraulic model can be developed to estimate cave airflow in the absence of continuous data, provided that aeraulic resistance (R) and internal–external temperature differences (which determine ΔP) are known. These parameters can be obtained through cave monitoring (perhaps a short period might be sufficient; e.g., a few weeks during both ventilation regimes). In fact, R should be derived from field measurements, rather than theoretical estimates, due to high uncertainty in cave roughness (friction factor). It should also be considered to discretise the ventilated conduit into smaller segments, if the cave geometries are complex, since the cave R is the sum of the resistance of the single segments. Ultimately, R depends on diameters. The relationship between R , ΔP , and airflow may allow inference of geometric and aeraulic properties of the conduit (e.g. the ratio R/D). Since convection length also depends on both conduit diameter and airflow, future works may investigate how to combine the cave convection length with R and ΔP to better characterise cave aeraulics.
- In Chapter 10 (Thermal length in hydrologically active caves), we qualitatively extended the concept of convection length to water flows at karst springs. Due to the different thermal characteristics between water and air, in the case of water fluxes, the convection length can be in the order of kilometres with large water discharge. Consequently, the thermal response at springs and the convection lengths largely depend on the recharge mode (whether it is diffused or concentrated) and the depth of the system. Then, in Chapter 11.4, we proposed a classification of spring thermographs based on these two characteristics, which depict the proportion existing between the system depth and its convection length during periods of low and high water flow. However, this classification is only qualitative and should be verified against quantitative data from multiple springs from different geological settings and climates, as well as with modeling work.

Open Research

The datasets of the cave temperatures and air flow are free to access.

Chapter 7 – Text files including the numerical results plotted in Figures 7.4-7.9 are available in <https://doi.org/10.5281/zenodo.10953068> cited as Sedaghatkish and Doumenc (2024). The Comsol Multiphysics file containing the data and results related to the reference case of Figures 7.4-7.6 is also available, but a license of Comsol Multiphysics software is necessary to use it. All the field data (including the three instrumented sites) can be found in <https://doi.org/10.5281/zenodo.10514555> associated to (Sedaghatkish et al., 2024b).

Chapter 8 – data plotted in Figure 8.5, Figure 8.10 and Figure 8.11 are available at the link <https://doi.org/10.5281/zenodo.18166879> cited as (Pastore et al., 2026).

Chapter 9 – The Comsol Multiphysics file for HA scenario (humid air with finite heat transfer) is available in <https://doi.org/10.5281/zenodo.14535340> associated with Sedaghatkish (2024) but a license of the software is necessary to use it.

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Supplements

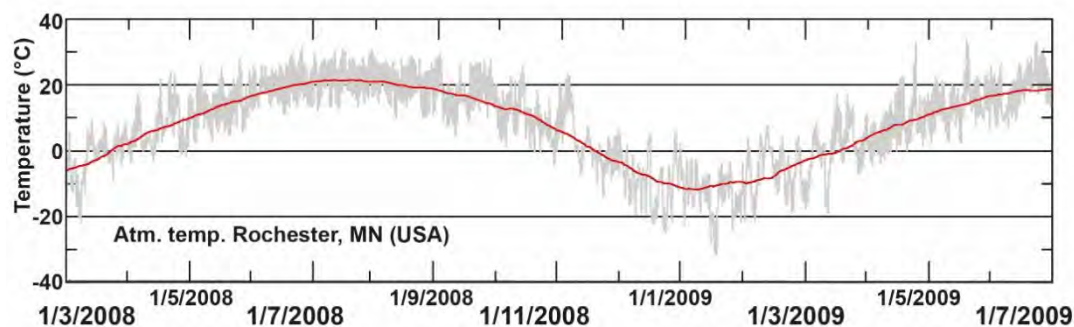


Figure S0.1 – Surface temperature at Rochester, Minnesota (USA). This external temperature is the reference air temperature for springs in Figure 11.2 and Figure 11.6.a/f/e (Luhmann et al., 2011).

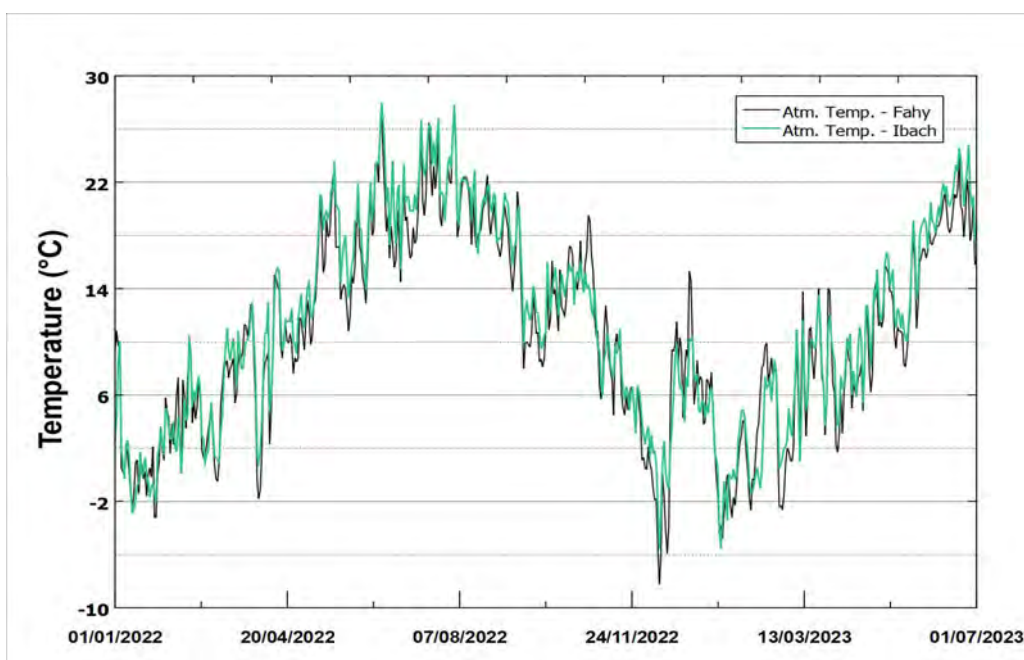


Figure S0.2 – Daily atmospheric temperatures at Fahy and Ibach weather stations (Switzerland). These weather stations give air reference temperatures for Milandre cave (Figure 11.6.b) and SBS spring (Figure 11.6.c), respectively.

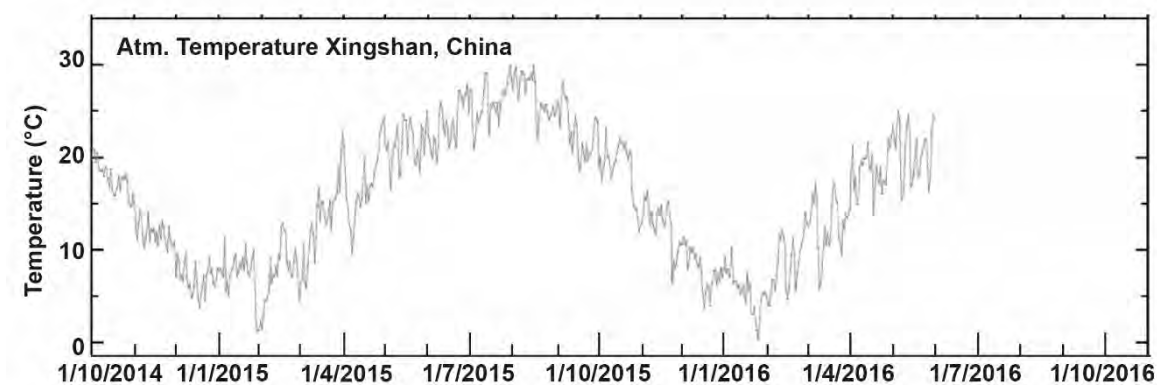


Figure S0.3 – Average air temperature from Xingshan weather station (China). This weather station gives air reference temperature for HLQ spring (Luo et al., 2023). Modified after (Luo et al., 2018); Copyright (2018), reproduced with permission from Springer Nature.

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