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**Assessing the role of permafrost in the preconditioning and triggering factors of the September 2020 Crête des Grangettes rockfall (southern French Alps)**

*Evaluation du rôle du permafrost dans les facteurs prédisposants et déclenchants de l'éroulement rocheux de la Crête des Grangettes de septembre 2020 (Massif des Écrins)*

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## Résumé

Les parois rocheuses affectées par du permafrost sont très sensibles au changement climatique, entraînant parfois des écroulements rocheux menaçant les populations et infrastructures. Des études ont établi un lien entre l'instabilité des parois et la dégradation du permafrost, mais une documentation plus approfondie de ces événements est nécessaire pour mieux comprendre les mécanismes liés à la dégradation du permafrost. Notre étude porte sur l'écroulement rocheux (~35 800 m<sup>3</sup>) qui s'est déclenché sur la Crête des Grangettes le 17 septembre 2020 (Hautes Alpes, France). Nous analysons les facteurs de prédisposition et de déclenchement de l'événement en combinant des méthodes de suivi des températures de surface du terrain, de modélisations numériques de transfert de chaleur, et de tomographie de résistivité électrique (ERT) associée à une analyse pétrophysique. Les résultats indiquent que le site d'étude est caractérisé par un permafrost tempéré (>-2°C) dont le réchauffement est progressif depuis les années 1990 et s'intensifie depuis 2015 (+0,48±0,12°C par décennie à 22 m de profondeur), ce qui a probablement fragilisé les joints de glace dans les fractures, prédisposant à l'instabilité du versant. Les études ERT confirment la présence de permafrost autour de la niche d'arrachement, avec des secteurs riches en glace (résistivité > 10kΩm). Enfin des précipitation 19 jours avant l'écroulement (91 mm cumulés) ont pu le déclencher par pressions hydrostatiques et/ou érosion de la glace dans les fractures. Cette approche multi-méthode fournis des résultats permettant de mieux évaluer les facteurs qui peuvent prédisposer ou déclencher un écroulement rocheux en contexte de permafrost.

**Mots-clés** : Géomorphologie périglaciaire, Suivi des températures de surface du terrain, Modélisation du permafrost, Tomographie de résistivité électrique, Écroulement rocheux.

## Abstract

Permafrost-affected rockwalls are highly sensitive to climate change, leading to rock slope failures that threaten human lives and activities. Numerous studies have connected rockwall instability with permafrost degradation, but further documentation of mid to large volume events is needed to better understand the specific mechanisms of permafrost degradation and their mechanical implications to develop effective approaches for hazard assessment. Our study investigates the rockfall (c. 35,800 m<sup>3</sup>) that was triggered from the Crête des Grangettes (Southern Alps, France) on September 17, 2020. We decipher the preconditioning and triggering factors of the event by combining ground surface temperature

monitoring, numerical modelling of heat transfer, and geophysical survey using Electrical Resistivity Tomography (ERT) interpreted through a petrophysical analysis. The results show that the study site is characterised by warm permafrost ( $>-2^{\circ}\text{C}$ ) since the 1970s that is progressively warming towards the melting point. This warming has been slow since the 1990s but has become more pronounced since 2015 ( $+0.48\pm 0.12^{\circ}\text{C}$  per decade at 22 m depth). This warming likely weakened ice-filled fractures, evidenced by ice in the rockfall scar, preconditioning the slope for instability. The ERT surveys confirm that the study site is still largely frozen around the scar with possible ice-rich layers (high resistivity values;  $> 10\text{k}\Omega\text{m}$ ). Finally, rainfall occurring 19 days before the rockfall (91 mm cumulative) may have triggered it by increasing hydrostatic pressure and/or eroding ice-filled joints. This multi-method approach provides results that enhance the assessment of thermal factors that may predispose or trigger rockfall in a warming permafrost rockwall.

**Keywords:** Periglacial geomorphology, Ground surface temperature monitoring, Permafrost modelling, Electrical Resistivity Tomography, Rockfall.

### **Résumé allongé en français**

Une série de déstabilisations s'est produite sur la Crête des Grangettes (Le Monêtier-les-Bains), dans le massif des Écrins (Hautes-Alpes), les 17 septembre, 20 décembre 2020, et 23 Juillet 2021 (fig. 1). Pour des raisons d'accessibilité au terrain, cet article étudie essentiellement l'écroulement rocheux ( $\sim 35\,800\text{ m}^3$ ) qui a eu lieu le 17 septembre 2020 à l'amont d'une voie d'alpinisme (la goulotte « Fantomas »), ainsi que d'un chemin de randonnée qui n'a toutefois pas été atteint.

L'écroulement s'est produit au niveau d'une crête rocheuse, au contact entre des grès et des marnes, à proximité immédiate d'une faille indiquée sur la carte géologique (fig. 2A-B). Ces lithologies peu cohésives et la discontinuité au niveau de la niche d'arrachement forment un contexte géologique favorable à des instabilités. Une fracture visible sur une orthophotographie de 2009 (fig. 2C) conditionne également la niche d'arrachement. De la glace et des écoulements d'eau observés dans la niche d'arrachement immédiatement après l'évènement suggèrent la présence de permafrost (fig. 3B), mais sa distribution locale et son rôle dans les processus de déclenchement doivent être vérifiés. D'après la carte de distribution du permafrost issue du *Rock Model* (Boeckli et al., 2012a) et recalculée pour les Alpes françaises avec les données Safran, la niche d'arrachement se situe dans un secteur dans

lequel le permafrost serait présent uniquement dans des conditions très favorables (fig. 4A). En revanche, la carte de susceptibilité de permafrost pour les formations superficielles (Marcer et al., 2017) suggère que le permafrost est présent dans presque toutes les conditions (fig. 4B). Il est donc nécessaire d'approfondir les investigations afin de préciser l'état et la distribution du permafrost dans la zone de départ de l'écroulement.

Le volume de la déstabilisation a été estimé à environ 35 800 m<sup>3</sup> par la comparaison d'un MNT à haute résolution (20 cm) réalisé à partir d'un levé drone effectué le 22 septembre 2021 avec le MNT de l'IGN à 20 cm de résolution de 2019 (fig. 5). Le dépôt de l'écroulement rocheux, très dispersé, a recouvert d'anciens dépôts et est de ce fait difficile à analyser (fig. 3A, C, D).

L'objectif de cette étude est d'évaluer les conditions thermiques qui ont prédisposé et déclenché l'écroulement rocheux. Pour ce faire, nous utilisons une approche qui combine des mesures *in situ* comme celle de la température de la surface du terrain (fig. 6), des études géophysiques (tomographie de résistivité électrique / ERT ; fig. 6) et une modélisation numérique de la distribution du permafrost (fig. 7), afin de détailler les conditions thermiques dans lesquelles l'écroulement s'est produit. En développant une approche globale à plusieurs échelles, nous avons obtenu des informations précieuses pour déchiffrer les processus liés au permafrost qui ont potentiellement conditionné et déclenché l'écroulement de la Crête des Grangettes le 17 septembre 2020.

L'analyse des données météorologiques à long terme montre que la chaleur exceptionnelle du printemps 2020, combinée à des températures estivales constamment élevées depuis 2015, ont pu contribuer à préparer l'écroulement rocheux (fig. 9). Par ailleurs, l'analyse des données météorologiques des trois mois qui ont précédé l'événement met en évidence des précipitations importantes 19 jours avant l'écroulement (91 mm en cumulé les 28 et 29 août 2020 ; fig. 10). Cet épisode a pu être un facteur déclenchant de l'écroulement en augmentant les pressions hydrostatiques dans les fractures et/ou en érodant les joints de glace.

À l'échelle locale, nous analysons les fluctuations de la température de surface du terrain grâce à son suivi continu (2021-2022 ; fig. 11). Les relevés montrent des températures annuelles moyennes de la surface des parois allant de -0,26°C (NO) à +1,88°C (O) dans les parois rocheuses autour de la cicatrice et des températures annuelles moyennes de la surface du terrain allant de -2,85°C à +1,18°C dans les pentes de débris. Étant donné que les

températures du terrain ont été mesurées au cours d'une année avec une température de l'air supérieure de 1,45°C à la moyenne 1991-2020, les résultats suggèrent que les conditions sur le site d'étude sont favorables à la présence de permafrost.

À l'échelle pluri-décamétrique, deux profils géophysiques à l'amont de la cicatrice de l'écroulement montrent des valeurs de résistivité élevées ( $>10 \text{ k}\Omega\text{m}$  ; fig.15), correspondant à des matériaux gelés, comme confirmé par les études en laboratoire menées sur deux échantillons prélevés dans les affleurements (fig. 14). Ces mesures révèlent la présence de permafrost autour de la zone de départ de l'écroulement rocheux.

Les simulations de transfert de chaleur suggèrent la présence de permafrost tempéré ( $> -2 \text{ }^\circ\text{C}$ ) à la profondeur de la niche d'arrachement (fig. 12). En effet, les séries temporelles de température proches de la surface montrent également que l'écroulement s'est produit à l'issue d'une période de réchauffement constant depuis environ 1990 (fig. 13).

D'après les observations et l'analyse des données, l'écroulement rocheux de la Crête des Grangettes du 17 septembre 2020 a probablement été déclenché par le réchauffement du permafrost, avec un permafrost tempéré approchant le point de fusion. À plus long terme (décennie), le réchauffement du permafrost a pu résulter de la conduction thermique depuis la surface comme suggéré par les modèles de transfert de chaleur, et d'un affaiblissement mécanique des joints de glace, menant à la déstabilisation de la paroi.

Enfin, notre approche multi-méthodes fournit des informations et des données précieuses pour l'évaluation des risques de déstabilisation de parois et pentes raides, et souligne la complexité des processus pouvant mener à des mouvements de versants à différentes échelles spatiales et temporelles.

## 1. Introduction

High mountain environments are very sensitive to the rise in air temperature, resulting in permafrost degradation. In the European Alps, the temperature of continuous permafrost has increased by 0.4 °C over the past 20 years (Haberkorn et al., 2021). However, ice-poor bedrock has experienced a much higher warming rate, up to three times this average value (Etzelmüller et al., 2020; Haberkorn et al., 2021; Magnin et al., 2024). Nevertheless, these trends remain highly variable depending on the region, and the effect of local ground conditions (microtopography, ground cover such as rock debris, snow, or vegetation, and ice content).

Permafrost degradation could cause slope instabilities (Gruber et Haeberli, 2007; Huggel et al., 2010; Ravanel et Deline, 2011; Ravanel et al., 2017; Tapia Baldis et Trombotto, 2019; Savi et al., 2021; Stoffel et al., 2024). Numerous studies have highlighted the role of permafrost degradation in rockwall destabilisation, for example through the presence of visible water and ice in the scar (Deline et al., 2011; Frauenfelder et al., 2018; Cathala et al., 2024). Other studies have linked the increasing rock slope failure frequency to climate warming (Ravanel et al., 2010; Ravanel et Deline, 2011; Fischer et al., 2012) and to summer heat waves (Temme, 2015; Ravanel et al., 2017; Paranunzio et al., 2019; Legay et al., 2021). In some studies, the use of Electrical Resistivity Tomography (ERT) has revealed a clear relationship between slope instability and permafrost degradation (*e.g.*, Duvillard et al., 2021). Investigations on the link between permafrost dynamics and the mechanical behaviour of rock destabilization explain that various thermo-hydro-mechanical processes could be involved (Gruber et Haeberli, 2007; Krautblatter et al., 2013). Laboratory experiments have showed that ice-infill in fractures significantly mechanically weakens under warming towards the melting point (*e.g.*, Davies et al., 2001; Mamot et al., 2018, 2021), which can be linked to observed rock slope failures in warm permafrost conditions (Deline et al., 2011; Frauenfelder et al., 2018; Legay et al., 2021; Ravanel et al., 2023). Permafrost degradation can also locally be accelerated by heat advection through water flow in fractures, also causing the erosion of the ice-infill and a loss of bonding between rock and ice (Hasler et al., 2011; Krautblatter et al., 2013). But water circulation can also lead to high hydrostatic pressures since the ice infill acts as an aquiclude, that could also trigger rock slope failures (Stoll et al., 2020; Magnin et Josnin, 2021). Despite existing research linking climate warming, permafrost degradation, and rockfalls, we still lack sufficient field data and in-depth analysis of rock failure events to gain

a more advanced understanding of the permafrost-related processes driving rock slope instabilities.

To address these needs, some studies have analysed the preconditioning and triggering factors of specific case studies (Fischer et al., 2010; Deline et al., 2011; Frauenfelder et al., 2018; Etzelmüller et al., 2022; Ravanel et al., 2023; Cathala et al., 2024). Our research aligns with this approach, focusing on the Crête des Grangettes in the Écrins massif (Southern French Alps), which has been affected by slope destabilisations on 17 September, 2020, December 2020, and 23 July 2021. This study focusses on the first rockfall, selected for its convenient site accessibility. Ice coating and liquid water were observed in its scar few hours after the event suggesting the likely presence of permafrost, but its local distribution and its role in predisposing and triggering the event remains to be confirmed.

The aim of this study is thus to assess the thermal conditions that have preconditioned and triggered the rockfall. To do so, we use an approach that combines *in-situ* measurements as ground surface temperature monitoring, geophysical surveys (Electrical Resistivity Tomography; ERT) and numerical modelling of permafrost distribution, to detail the thermal conditions under which the rockfall occurred.

## 2. Study area

### 2.1. Geographical and geological setting

The Crête des Grangettes is located in the Grand Tabuc valley (Écrins massif) within the municipality of Monêtier-les-Bains (Hautes-Alpes) (fig. 1A). The study site is located at 3100 m a.s.l., east of the Monêtier glacier (fig. 1B) whose main body was only 0.7 km<sup>2</sup> in 2018. Three rock slope failures were identified by field observations and aerial images by the RTM (*Restauration des Terrains de Montagne*) on the Crête des Grangettes area on September 17 and December 20, 2020, and on July 23, 2021 (fig. 1B). After these slope destabilisations, the Grand Tabuc valley was closed by a Municipal decree above 2100 m a.s.l. following the September 17 event, and then above 1900 m a.s.l. after the December 20 rockfall. This closure has limited recreational and sporting activities (hiking, mountaineering, ski touring, etc.) in the valley until its reopening on 08 March 2021.

This study focusses on the rockfall that occurred on September 17, 2020, which occurred upstream the “Fantomas” ice couloir, one of the climbing routes leading to the Dôme de Monêtier (3404 m a.s.l.; fig. 1B), and a hiking path that was not reached.



**Figure 1: Study site.** A: Location map (*source: Esri, Maxar, Earthstar Geographics, and the GIS User Community*). B: Grand Tabuc and Crête des Grangettes map, main instabilities observed in 2020 and 2021, and main hiking and mountaineering paths (*source: IGN 0.2 m resolution DEM*).

1. Hiking or mountaineering route; 2. Stream; 3. Glacier; 4. Release area of the September 17, 2020 rockfall; 5. Release area of the December 20, 2020 rockfall; 6. Release area of the July 23, 2021 rockfall.

**Figure 1 : Site d'étude.** A : Carte de localisation (*source : Esri, Maxar, Earthstar Geographics, and the GIS User Community*). B : Carte du vallon du Grand Tabuc et de la Crête des Grangettes, les principales instabilités observées en 2020 et 2021, et les principaux itinéraires de randonnée et voie d'alpinisme (*source : MNT à 0.2 m de résolution*).

1. Itinéraire de randonnée ou d'alpinisme ; 2. torrent ; 3. glacier ; 4. zone de départ de l'éroulement du 17 septembre 2020 ; 5. zone de départ de l'éroulement du 20 décembre 2020 ; 6. zone de départ de l'éroulement du 23 juillet 2021.

The Crête des Grangettes is included in the oriental part of the Écrins-Pelvoux massif, which is an external crystalline massif of the Western Alps. It appears immediately at the West both of the nummulitic transgression deposits and of the sub-briançonnais overlapping thrust. At this place, several scales are cutting the massif, including the *écaille des Grangettes* (Bravard and Gidon, 1979) which is divided in smaller scales, the upper ones being composed of the Mesozoic cover of the granites.

The rockfall occurred on a ridge at the contact between sandstone and clay (fig. 2). The thrust fault marked on the geological map (fig. 2A, B) indicates the limit between two small scales and allows the Trias series to overlap the Jurassic series. The smaller faults associated with this secondary thrust fault combine with the stratigraphical joints and with the weakly cohesive lithology to create an area of instability. The fault at the release area, that is also evident on a 2009 orthophotograph (fig. 2C), aligning precisely with the September 2020 rockfall scar, is one of these small faults. More details on the geology of the study site are given in Section 3.3.1.

**Figure 2: Geological setting.** A: Geological map of the study site (from BRGM). B: Geology of the release area and the zone instrumented for this study. C: Orthophotograph of the release area (20 cm resolution). White dotted line: structural weakness corresponding to the scar (IGN orthophotograph, 05/08/2009).

1. Fluvial alluvium; 2. Mixed screes, alluvial and/or avalanche fans, sometimes reworked by water runoff; 3. Active to fixed (and  $\pm$  reactivated) recent to modern screes, locally solifluent; 4. Modern to historic glacial moraines or tills; 5. Glaciers and snow patches; 6. Sandstones, microbreccias, sandy limestones, biotrititic limestones, marmoreal limestones; 7. Undifferentiated marls or black shales; 8. Dolomitic shales  $\pm$  dolomites, coarse breccias, spathic limestones; 9. Ophitic dolerite with clustered quartz diorite; 10. Amphibolitic gneiss

and amphibolites; 11. Gneiss, undifferentiated migmatitic gneiss; 12. (Mg-Fe-rich) granites  $\pm$  Turbats porphyroïd; 13. Migmatitic granites, anatexis granites ; 14. (Mg-Fe-rich) fine-grained granites (of Pelvoux-Ailefroide granites); 15. Migmatites, anatexites and gneiss  $\pm$  migmatitic; 16. Undifferentiated migmatites; 17. Visible, observed fault; 18. Supposed fault.

**Figure 2 : Contexte géologique.** A : Carte géologique du site d'étude (d'après BRGM). B : Géologie de la zone de départ et de la zone instrumentée pour cette étude. C : Orthophotographie de la zone de départ (20 cm de résolution). Pointillés blancs : faiblesse structurale correspondant à la niche d'arrachement (orthophotographie IGN, 05/08/2009).

1. Alluvions fluviales ; 2. Cônes mixtes d'éboulis, de déjection et/ou d'avalanches, cônes remaniés par le ruissellement ; 3. Éboulis récents à actuels, actifs à fixés (et  $\pm$  réactivés), localement soliflués ; 4. Moraines glaciaires actuelles à historiques ou tills ; 5. Glaciers et névés ; 6. grès, microbrèches, calcaires gréseux, calcaires biodétritiques, calcaires marmoréens ; 7. marnes ou schistes noirs indifférenciés; 8. Schistes dolomitiques  $\pm$  dolomies, brèches grossières, calcaires spathiques ; 9. Dolérite ophitique à diorite quartzique en amas ; 10. Gneiss amphiboliques et amphibolites ; 11. Gneiss, gneiss migmatitiques indifférenciés ; 12. granites (magnésio-férrifères)  $\pm$  porphyroïdes de Turbat ; 13. granites migmatitiques, granites d'anatexie ; 14. granites (magnésio-férrifères) à grains fins (du granite de Pelvoux-Ailefroide) ; 15. Migmatites, anatexites et gneiss  $\pm$  migmatitiques ; 16. Migmatites indifférenciées ; 17. Faille visible, observée ; 18. Faille supposée.

## 2.2. Periglacial context

The scar is located on a north face. Ice and water coating were observed during an aerial survey of the area a few hours after the event, suggesting the occurrence of permafrost in the release area (fig. 3B).

**Figure 3: The September 17 rockfall, 2020.** A: The Crête des Grangettes and identification of the rockfall scar and deposit (ph.: 03/10/2023) B: Rockfall scar (ph.: PGHM Briançon, 18/09/2020). C: Scar and rockfall propagation area (ph. PGHM Briançon, 18/09/2020). D: Detail of the rockfall deposit (ph.: PGHM Briançon, 18/09/2020).

**Figure 3 : L'écroulement du 17 septembre 2020.** A : La Crête des Grangettes et identification de la niche d'arrachement de l'écroulement (ph. : 03/10/2023) B : Niche d'arrachement (ph. : PGHM Briançon, 18/09/2020). C : Niche d'arrachement et zone de propagation de l'écroulement rocheux (ph. PGHM Briançon, 18/09/2020). D : Détail du dépôt de l'écroulement (ph. : PGHM Briançon, 18/09/2020).

To map permafrost in the steep parts of our study site, we use the *Rock Model* calibrated by Boeckli et al. (2012a) with a sample of measured surface temperature in rockwalls directly coupled with the atmosphere, and that calculates the Mean Annual Rock Surface Temperature (MARST) from the Potential Incoming Solar Radiation and the Mean Annual Air Temperature. We implement this model with the 1981-2010 air temperature using

the S2M SAFRAN air temperature from *Météo France* (Vernay et al., 2022). The calculated MARST is then classified in various permafrost conditions to be interpreted against local ground characteristics (Boeckli et al., 2012b). The map shows that the release area is located in a sector where permafrost would only be present under very favourable conditions (fig. 4A), *i.e.* highly fractured and partially snow cover (Boeckli et al., 2012b). In contrast, the *Permafrost Favorability Index* (PFI; Marcer et al., 2017), that is calibrated with a rock glacier inventory and with climatic and topographic data, and that is more relevant for debris slopes, suggests that permafrost is present in almost all conditions on the study site (fig. 4B). But this PFI map is calibrated with landforms inherited from the Little Ice Age and tends to overestimate permafrost conditions (Cathala et al., 2024). Further investigation is therefore required to clarify the condition and distribution of permafrost in the release area.

**Figure 4: Permafrost maps.** A: Permafrost susceptibility map from the Rock Model of Boeckli et al. (2012). B: Permafrost susceptibility map based on the PFI of Marcer et al. (2017).

1. Glacier; 2. Permafrost in nearly all conditions; 3. Permafrost mostly in cold conditions; 4. Permafrost only in very favourable conditions; 5. Rockfall scar.

*Figure 4 : Cartes du permafrost. A : Carte de susceptibilité de permafrost d'après le Rock Model de Boeckli et al. (2012). B : Carte de susceptibilité de permafrost d'après le PFI de Marcer et al. (2017).*

*1. Glaciers ; 2. Permafrost présent dans presque toutes les conditions ; 3. Permafrost présent principalement dans des conditions froides ; 4. Permafrost présent uniquement dans des conditions très favorables ; 5. Niche d'arrachement.*

### 2.3. Description of the rockfall of September 17, 2020

The rockfall occurred from an outcrop and the released volume is estimated to 35,800 m<sup>3</sup> by comparing a high-resolution (20 cm) Digital Elevation Model (DEM) created from a drone survey conducted on 22 September 2021 with the 20 cm resolution DEM from IGN (*Institut National de l'information Géographique et forestière*) acquired in 2019 (fig. 5A-B). The maximum depth of the scar is c. 49 m, with an average depth of 28 m.

The rockfall propagated into the Grand Tabuc valley. Its trajectory is difficult to reconstruct as the deposit spread along multiple paths (fig. 3A). The rockfall deposit is barely visible due to its dispersion and spreading over sediments from previous rockfall deposits, colluvium, alluvium, and tills (fig. 3A, C, D). However, part of the deposit is visible (fig. 3A,

D), while another part may have reached and/or been remobilised by the Grand Tabuc stream (fig. 3C), although this was not identifiable from satellite images or field observations made on December 3, 2020.

**Figure 5: Study site.** Left: Release area of the rockfall of September 17, 2020 and depth of the scar calculated by comparing a DEM (20 cm resolution) produced from a drone survey on 22 September 2021 with the IGN DEM (20 cm resolution, data acquisition in 2019). Right: Profile of the rockfall scar from the DEMs comparison.

*Figure 5 : Site d'étude. Gauche : Secteur de départ de l'éroulement rocheux du 17 septembre 2020 et profondeur de la niche d'arrachement calculée à partir de la comparaison d'un MNT (20 cm de résolution) réalisé à partir d'un levé drone du 22 septembre 2021 avec le MNT de l'IGN (20 cm de résolution, acquisition des données en 2019). Droite : Profil de la niche d'arrachement à partir des MNTs « avant/après » éroulement.*

### 3. Methods

To assess the thermal conditions that have preconditioned or triggered the rockfall, we combined various methods to analyse both the weather conditions and the thermal characteristics of the ground at various temporal and spatial scales.

#### 3.1. Air temperature and meteorological analysis

The weather conditions preceding the rockfall are analysed through air temperature and precipitation records from the Monétier les Bains Automatic Weather Station (AWS; 1459 m a.s.l.), which provides daily data since 1985. Air temperature records are daily averages calculated from the minimum and maximum temperatures recorded each day while the precipitation records are the sum of rainfall each day. We aggregate these data into seasonal values and calculate anomalies by comparing those seasonal records to the averages of a 30-year reference period (1991 to 2020) for each season.

#### 3.2. Ground temperature measurements

To characterise ground thermal regime on the study site and to assess the permafrost state and evolution around the scar, ground surface temperature (GST) is monitored since 2021. In this paper, we use the temperatures recorded between 22.09.2021 and 20.09.2022. The location of each sensor is shown on Figure 6. The measurements are performed at hourly time steps using 12 temperature sensors combining Mouser® IButton DS1925L-F5# (accuracy  $\pm 0.5$  °C; resolution 0.065 °C) in gentle slopes and Geoprecision® M-Log5W-ROCK (accuracy  $\pm 0.1$  °C; resolution 0.01 °C) on rockwalls, all installed about 10 cm below the ground surface. Four sensors were installed on the rockwalls: three of them around the top

of the scar (between 3139 and 3141 m a.s.l., facing north, northeast, and west, respectively), and the last one upstream of the destabilised area (3234 m a.s.l., facing northeast). Eight sensors were installed between 3148 and 3203 m a.s.l. in debris slope. Sensors Block #1 to Block #6 are located under thin material, while sensors Block #7 and Block #8 are in coarser material where air circulation may be more significant (Delaloye and Lambiel, 2005). Data were collected over the period September 23, 2021 to September 19, 2022. We calculate annual averages and complete three missing days in 2022 by duplicating data from September 17, 2022 to September 19, 2022. Annual averages directly inform whether the conditions are favourable or not to permafrost.

**Figure 6: Instrumentation on the study site.** A: Location map of the temperature sensors installed in the rockwall and in the debris slope, and of the geophysical profile. B-C: Location of temperature sensors and their MAGST recorded in 2021-2022 (ph. 22 Sept. 2021).

1. Temperature sensor located on gentle slope; 2. Temperature sensor located on rockwall; 3. Profile 1 (ERT, 240 m, electrode spacing: 5 m); 4. Profile 2 (ERT, 400 m, electrode spacing: 5 m); 5. Linear section for heat transfer modelling; 6. Rockfall scar

*Figure 6 : Instrumentation du site d'étude. A : Carte de localisation des capteurs de température installés en paroi et dans des formations superficielles, et du profil géophysique. B-C : Emplacement des capteurs de température et leur température annuelle moyenne de surface du terrain enregistrée en 2021-2022 (ph. 22 sept. 2021).*

*1. Capteurs de température installés dans des pentes douces ; 2. Capteurs de température installés sur des parois rocheuses ; 3. Profil 1 (ERT, 240 m, espacement des électrodes : 5 m) ; 4. Profil 2 (ERT, 400 m, espacement des électrodes : 5 m) ; 5. Section linéaire pour la modélisation de transfert de chaleur ; 6. Niche d'arrachement.*

### 3.3. Ground temperature modelling

#### 3.3.1 Model description

In order to estimate the thermal state of the release area when the rockfall occurred and the thermal dynamic prior to this occurrence, we modelled the temperature evolution in the release area using the Feflow® software allowing heat transfer modelling (DHI, version 8.0; Clausnitzer and Mirnyy, 2015; Feflow user guide, 2016). The two-dimensional (2D) domain was defined using the IGN 20 cm resolution DEM (resampled at 1 m resolution), with a linear section-oriented N-S which intersects the rockfall scar (fig. 6A). The domain was extended down to 5 km below the surface to include the geothermal flux in a realistic way (fig. 7A). The upper part of the simulated domain is split into seven parts, to account for geological

layers with distinct thermal properties (thermal conductivity and heat capacity; fig. 7A and tab. 1).

**Figure 7: (A) Model geometry and mesh implemented in Feflow.** Each layer, numbered from 1 to 7, has specific thermal properties, presented in Table 1. (B) Pattern of ground temperature applied at the surface as boundary condition in the model (see text).

*Figure 7 : (A) Géométrie du modèle implémentée dans Feflow. Chaque couche, numérotée de 1 à 7, possède des propriétés thermiques spécifiques, présentées dans le Tableau 1. (B) Profil de température utilisé en conditions aux limites à la surface du modèle.*

**Table 1: Geological layers description and their thermal conductivity and heat capacity values used for the thermal transfer flow.** Thermal conductivity data is from Waples and Waples (2004) and heat capacity data is from Rao et al (2022).

*Tableau 1 : Description des couches géologiques, et leurs valeurs de conductivité et de capacité thermique utilisées pour simuler les transferts thermiques. Les données relatives à la conductivité thermique proviennent de Waples et Waples (2004) et les données relatives à la capacité thermique proviennent de Rao et al (2022).*

The geometry of each layer has been inferred from the geological map and airborne photographs. A triangle mesh gradually coarsening with depth was applied (fig. 7A). The water flow was not taken into account in the simulations. This was achieved by constraining the entire simulated domain to be fully saturated, by applying a constant hydraulic head of 3280 m (slightly higher than the maximum elevation, though) as a boundary condition at all surface nodes. A very low hydraulic conductivity of  $10^{-12} \text{ m s}^{-1}$  is also applied to the entire domain so that no significant flow is allowed. Freeze and thaw effects are accounted for through the piFreeze (version 1.001) extension included in Feflow. The main parameters are shown in Table 2.

**Table 2: Main parameters used in the heat transfer simulations.**

*Tableau 2 : Principaux paramètres utilisés dans les simulations de transfert de chaleur.*

### 3.3.2 Forcing timeseries

We compared the timeseries of daily averaged ground temperature recorded from 23 September 2021 to 19 September 2022 (Sect. 3.2) with air temperature data from the three closest AWS (Fig. 1A): Le Monétier les Bains (hereafter LMLB; 1459 m a.s.l.), Le Monétier

(2185 m), and Ecrins-Nivôse (2970 m). For all sensors, the best correlation is obtained with LMLB AWS (average correlation: 0.92; fig. 8), which we thus used to create forcing timeseries. Temperature records from this AWS started on 1 November 1935, and have 1558 missing data, out of which 1143 correspond to an almost continuous period between 1939 and 1942, and 106 correspond to five non-continuous months between 1985 and 1986. The remaining missing data are sporadic in the dataset. In order to fill these gaps, we proceeded as follows: we first used the data from another AWS (Besse-en-Oisans, 1520 m a.s.l.; 29 km WNW from LMLB; records starting in 1959; correlation coefficient 0.943 between both AWS), after applying a linear offset. For the 1939 – 1942 period, we used the linear fit of the whole dataset. Last, the remaining gaps were filled with a linear interpolation between existing data. In summary, we get a complete daily timeseries of air temperature, from 1 November 1935 onwards at the altitude of LMLB AWS (1459 m a.s.l.). An additional time step corresponding to 1850 conditions is defined as 1°C less than the 1961-1990 temperature average, following Magnin et al. (2017).

**Figure 8: Scatter plots of the ground surface temperature measured by a selection of four sensors from 23 September 2021 to 19 September 2022, versus air temperature measured at Le Monétier les Bains AWS (blue crosses). For each panel, the linear regression is displayed (orange line) with its equation and correlation coefficient.**

*Figure 8 : Diagramme de dispersion entre la température mesurée par quatre capteurs du 23 septembre 2021 au 19 septembre 2022, et la température de l'air mesurée à la station du Monétier les Bains (croix bleues). Pour chaque graphique, la régression linéaire est représentée (ligne orange) avec son équation et son coefficient de régression.*

In a second step, the resulting air temperature timeseries is adjusted with a linear regression to mimic the ground temperature recorded by the sensors. The surface nodes of the modelled domain are split in four distinct parts (fig. 7B). A common lapse rate of 5 °C km<sup>-1</sup> is systematically applied for all parts. The outcrop on the N side is forced with temperature timeseries adjusted on the Mon\_N sensor. Then, the “Block #4” sensor is used as reference for the second part of the profile, while “Block #6” is the reference for the third part. This choice of sensors is based on the correlation against air temperature records, which are the highest amongst the eight sensors on the ground, which is likely explained by a more direct coupling between the ground and the air, with less effect of the snow cover. Finally, the last part of the section is based on “Block #6” sensor with an additional +8 °C offset accounting for the southward aspect (Gruber et al., 2004; Magnin et al., 2015a). Between each part of the profile, transition nodes are linearly interpolated to allow smooth temperature changes (dotted

lines on fig. 7B; reference simulation). To investigate the sensitivity and robustness of the model depending on the choice of sensors, we introduce a modified forcing pattern (warm simulation), using Mon\_W temperature instead of Mon\_N at the middle of the release area (fig. 7B).

### 3.3.3 Model runs and output

A first spinup simulation is performed with constant temperature forcing at the surface corresponding to conditions in 1850 CE, for a sufficiently long time to reach equilibrium. Then, using this spinup simulation to initialise the model, the ground temperatures timeseries are used as input, and the model is run until present. The daily timeseries are resampled to monthly averages until 1989, and weekly averages until present, which is sufficient to capture the dynamics of the temperature evolution at depth, while significantly fastening the model run. Decadal outputs are requested for the entire simulated domain, as long as an additional output for September 2020. A few observations points are also defined on the mesh, with continuous record of the model variables.

## 3.4. Electrical Resistivity Tomography and petrophysical analysis

Two ERT profiles were carried out using an ABEM Terrameter LS2; the location of the profiles is shown in Figure 6. Profile 1 has 48 electrodes while Profile 2 has 96 electrodes with a 5 m electrode spacing between adjacent electrodes. Stainless steel electrodes were used for both current injection and voltage measurements. The measurement protocol is a combination of Wenner- $\alpha$  and Schlumberger quadrupoles, with a multichannel acquisition. The dGPS position of every electrode was measured during the acquisition.

The raw datasets collected in the field were first filtered to eliminate negative resistivity values, as well as outliers in the pseudo-sections. Later on, the data of both profiles were gathered in a same file to carry out a 3D inversion. In total, 1118 data points were used in the inversion. The ResIPY software (Blanchy et al., 2020) was used for the inversion of the recorded data.

In addition to the geophysical surveys, we carried out a petrophysical analysis in laboratory on two rock samples taken along the profiles to measure their resistivity into saturated conditions. The two samples are cut in 4 cm cubes and show highly differentiated lithologies, as indicated on the geological map. Sample GRAN-01 is marl, while GRAN-02 is sandy limestone. They are dried for 24 hours at 58°C. After drying, the samples are saturated



under vacuum conditions with melted snow water to match the ionic content as closely as possible to the conditions during fieldwork. The samples are then left in their pore water solution for 1 month to reach equilibrium. Resistivity is measured using a high-precision impedance analyzer, ZEL-SIP04-V02 (Zimmermann et al., 2008), over a frequency range of 0.01 to 45 kHz. The samples are subsequently fitted with gel carbon electrodes and placed in a thermally controlled bath (Abdulsamad et al., 2019; Coperey et al., 2019). The complex resistivity was measured across a temperature range from +25 °C to -15 °C. Later, the resistivity measured at 1 Hz (the frequency of the field survey) was extracted and used to interpret the field results.

## 4. Results

### 4.1. Meteorological factors

The rockfall occurred in a context of rising seasonal temperature anomalies, particularly in summer (fig. 9A). Since 2015, summer temperature anomalies compared to the reference period have consistently been above +1 °C (+1.3 °C in 2015, +1.5 °C in 2017, +1.0 °C in 2018; +1.7 °C in 2019). In 2020, the average summer temperature was 0.5 °C above the reference period. The average winter temperature in 2020 was 1.8 °C higher than the one of the reference period, making it the fifth warmest winter since 1985, after the winters of 1989, 2007, 2016, and 2017 (+2.0 °C, +1.9 °C, +2.3 °C, and +1.8 °C, respectively). Spring 2020 was the second warmest spring since 1985, with a temperature 1.6 °C higher than the reference period (after spring 2011, which was +1.7 °C).

Figure 9B shows a precipitation anomaly of +46 % (94 mm) in winter 2020, +26 % (50 mm) in spring 2020 and +22 % (40 mm) in summer 2020 compared to the period 1991-2020, which respectively makes them the 7<sup>th</sup>, 6<sup>th</sup>, and 5<sup>th</sup> wettest winter, spring, and summer of the period. Autumn 2020 was drier than the reference period autumns, with cumulative precipitation anomaly of -8.4 % (-24 mm). Therefore, the analysis of seasonal precipitation anomalies reveals no clear factors that could have triggered the rockfall.

**Figure 9: Air temperature and precipitation anomalies recorded at the Monêtier les Bains AWS, calculated in relation to the reference period 1985-2014. A: Air temperature anomalies. B: Precipitation anomalies.**

*Figure 9 : Anomalies de température de l'air et de précipitations enregistrées à la station du Monêtier les Bains, calculées par rapport à la période de référence 1985-2014. A : Anomalies de températures de l'air. B : Anomalies de précipitations.*

In the three months preceding the rockfall, daily average air temperature ranged between +7.7 °C and +21.8 °C at LMLB, which corresponds to a temperature between -0.5 °C and +13.6 °C when extrapolated using an adiabatic lapse rate of 0.5 °C 100 m<sup>-1</sup> (Durand et al., 2009) between the LMLB AWS (1459 m a.s.l.) and the site (3100 m a.s.l.). At LMLB AW, three heat peaks with temperatures above 20 °C (approximately 11.8 °C at the study site) were recorded on July 30, 2020; August 09, 2020; and August 21, 2020. Precipitation records show no exceptional event for the season; however, substantial rainfall occurred 19 days before the event, with a cumulative of 91 mm recorded on 28 and 29 August 2020 (fig. 10).

**Figure 10: Air temperature and precipitation at Le Monétier les Bains AWS (1459 m a.s.l.) from 1 June 2020 to 21 September 2020.**

*Figure 10 : Température de l'air et précipitations à la station du Monétier les Bains (1459 m) du 1<sup>er</sup> juin 2020 au 21 septembre 2020.*

## 4.2. Ground thermal conditions

To assess the thermal state of the ground around the release area, we obtained results from various methods, *in-situ* temperature measurements (Section 4.2.1), heat transfer modelling (Section 4.2.2) and ERT surveys (Section 4.2.3).

### 4.2.1. Ground surface temperature

The sensors located in rockwalls recorded MARST ranging from -0.26 °C on the north-facing slope to +1.88 °C on the west-facing slope (fig. 6B-C; fig. 11A; tab. 3) during the 2021-2022 period. The MAGST recorded with the sensors located in loose debris are ranging from -2.85 °C to +1.18 °C (fig. 6B-C; fig. 11B; tab. 3). The air temperature recorded during the 2021-2022 hydrological year (8.1 °C) was 1.5 °C higher than the 1991-2020 average (6.6 °C), and the temperature at depth is likely lower than the recorded MAGST (2021-2022) due to the effects of past colder decades. As a result, the GST measured by the sensors suggest that conditions across the entire study site are favourable for the presence of permafrost.

**Table 3: Characteristics and data from the temperature sensors installed at the Crête des Grangettes.**

*Tableau 3 : Caractéristiques et données des capteurs de température installés à la Crête des Grangettes.*

The temperature measured by sensor Block #1 showed little variations between 27 September 2021 and 04 July 2022. This sensor is located near a slight depression and was likely covered by a thick snow cover which isolate the ground from air temperature variations. In contrast, Block #2 to #6 sensors exhibit significant temperature fluctuations during winter (fig. 11B), which is consistent with the strong coupling between air and ground temperature shown in Fig. 8. This is likely due to a shallow snow cover resulting from wind erosion, and/or enhanced snow sublimation due to interstitial air circulation between blocks. A similar pattern is seen with the Block #7 and Block #8 sensors, though snow cover smoothed their temperature oscillations from late March. For all sensors in superficial formations, the snowmelt period is evidenced by a nearly constant surface temperature at 0 °C during the melting phase in April, called Zero Curtain Effect (Kelley and Weaver, 1969). This period occurs later for Block #7 and Block #8 sensors (from 19 May to 1 June and 09 June, respectively) and extends until 4 July for Block #1, confirming significant snow cover at this sensor.

**Figure 11: Ground Surface Temperature measurements.** A: Average daily temperature at the surface of the rockwalls (~ 10 cm). B: Average daily temperature at the surface of the debris slopes (~-10 cm).

*Figure 11 : Mesures de la température de surface du terrain.* A : Température moyenne journalière à la surface des parois (~ 10 cm). B : Température moyenne journalière à la surface du dépôt détritique (~-10 cm).

#### 4.2.2. Heat transfer simulations

The results of both simulations are presented in Figure 12 and Figure 13. Overall, these simulations show the presence of permafrost in a major part of the simulated domain, below the N-facing slope. A minimum temperature between -2.0 °C and -3.0 °C is still computed in 2020 in the upper part, but the extent of this cold permafrost domain is vanishing as compared to 2010.

**Figure 12: Simulated ground temperature in reference simulation (left column) and warm simulation (right column), in January 2000 and 2010, and in September 2020.** The scar line is shown (grey line), as well as the observation points 5, 8, and 11 (orange,

magenta, and green dots). Note the temperature intervals for negative temperatures are half those for positive temperatures.

*Figure 12 : Température du terrain simulée dans la simulation de référence (colonne de gauche) et la simulation chaude (colonne de droite), en janvier 2000 et 2010, et en septembre 2020. La ligne de cicatrice est représentée (ligne grise), ainsi que les points d'observation 5, 8 et 11 (points orange, magenta et vert). Notez que les intervalles de température pour les températures négatives sont la moitié de ceux pour les températures positives.*

The acceleration of the warming is highlighted in Figure 13, which shows three distinct phases: until the mid-1990, no significant warming is observed; a marked warming followed by a stabilisation is seen until 2015; finally, a steep temperature rise occurs after 2015. Since 2015, the temperature increase is 0.054, 0.056, and 0.0244 °C a<sup>-1</sup> for observation points 5, 8, and 11, located 20, 22, and 26 m below the surface respectively. This illustrates the effect of the heat capacity of the ground, which limits the temperature rise at depth since the heat energy is gradually consumed while transferring at depth. The warm simulation shows the effect of an additional warming applied on the rockwall, which propagates at depth. The effect of latent heat of freezing, which consumes heat while keeping the temperature constant, is striking for observation points 5 and 8 from 1997 onwards: the seasonal variation of the temperature is much weaker than in the reference simulation, because their temperatures reach -1 °C, which is the onset of ice melting in the model (tab. 2). For the observation point 5 located close to the rockwall (14 m apart, in the horizontal direction), the difference of simulated temperature is less than 1 °C, and the difference is smaller for the observation points 8 and 11 located farther (fig. 13). This gives a good indication of the uncertainty of this method. However, the overall patterns of ground warming are very similar for both simulations, and the rates of warming since 2015 are alike. An important result evidenced by both simulations, is that the temperatures reached the months prior to the rockfall are unprecedented in the previous century.

**Figure 13: Time series of ground temperature at three observation points (5, 8, and 11; see Fig 12 for their location). Full lines: reference simulation. Dotted lines: warm simulation.**

*Figure 13 : Série temporelle de la température du terrain à trois points d'observation (5, 8 et 11 ; voir Fig 12 pour leur localisation). Lignes pleines : simulation de référence. Lignes pointillées : simulation chaude.*

#### 4.2.3. Electrical Resistivity Tomography

The petrophysical analyses show a freeze/no-freeze transition located between 4 and 10 k $\Omega$ m for sample GRAN-01 (marl), while the transition is located between ~1 and 32 k $\Omega$ m for sample GRAN-02 (sandy limestone; fig. 14). The resistivity range for which the freeze/no-freeze transition occurs is therefore smaller for Sample 1. These results show that the contrast between freeze/no-freeze is well pronounced in case of GRAN-02 which facilitate the detection of permafrost in this formation more than in GRAN-01. Furthermore, these results could be fitted with Arrhenius's law or a linear model considering the behaviour of resistivity above/under the freezing temperature (solid line in fig. 14; for more information about this model see Abdulsamad et al., 2019; Coperey et al., 2019).

**Figure 14: Evolution of the electrical resistivity (at 1 Hz) measured in laboratory as a function of the temperature, for the two samples GRAN-01 (marl) and GRAN-02 (sandy limestone) taken along the ERT profile.** The solid lines represent the fit of data with linear model (see Abdulsamad et al., 2019; Coperey et al., 2019).

*Figure 14 : Evolution de la résistivité électrique (à 1 Hz) mesurée en laboratoire en fonction de la température, pour les deux échantillons GRAN-01 (marnes) and GRAN-02 (calcaire gréseux) prélevés le long du profil ERT. Les courbes représentent l'ajustement des données avec un modèle linéaire (voir Abdulsamad et al., 2019, Coperey et al., 2019).*

The results of ERT are presented in Figure 15A-15B. Based on the laboratory measurements (fig. 14), we could delimit the permafrost zone characterized by a resistivity value higher than ~10k $\Omega$ m. From the distribution of the electrical resistivity, we could easily distinguish a zone characterized by a high resistivity value (from 10 to 100 k $\Omega$ m) and a zone with a resistivity two to three order of magnitude lower (from 0.01 to 10 k $\Omega$ m).

Profile 1 (fig. 15) reveals very high resistivity (>10 k $\Omega$ m) at both the upstream and downstream ends, which indicates the presence of permafrost, and likely in the scar area. However, there is also a zone of low resistivity (~3 to 5 k $\Omega$ .m) in the centre of the profile. This zone corresponds to a former permanent snow patch, which may have limited permafrost formation due to its insulating effect. The low resistivities measured here also suggest that the area may be saturated with liquid water resulting from snow melt. Investigating this zone is challenging, as the area of interest (the scar) is located at the profile edge, potentially causing edge effects that could limit the tomogram accuracy.

**Figure 15: Electrical resistivity tomogram taken on September 20, 2022.** Orange line: Profile 1; pink line: Profile 2; white dotted line delineates frozen to non-frozen area.

*Figure 15 : Tomogramme de résistivité électrique réalisé le 20 septembre 2022. Ligne orange : Profil 1 ; ligne rose : Profil 2 ; la ligne blanche en pointillé délimite les secteurs gelé et non gelé.*

Profile 2, with the central electrodes located directly upstream of the scar, confirms the presence of permafrost near the scar area. However, Profile 2 also shows relatively low resistivities ( $<1 \text{ k}\Omega\text{m}$ ) on the western loop of the profile, corresponding to a recently glaciated area (Gardent et al., 2014), which likely explains the low resistivities. At this stage, we could notice a good agreement between the ERT results (*i.e.*, finding the frozen zone or permafrost) and the results of numerical simulations of heat transfer (fig. 12).

## 5. Discussion

### 5.1. Permafrost conditions and ground thermal evolution as predisposing and triggering factors of the Crête des Grangettes rockfall

The combination of *in-situ* observation methods (ground temperature monitoring and geophysical surveys) and heat transfer models enabled the characterisation of the thermal state of the ground and the assessment of permafrost degradation as a preconditioning and triggering factor for the rockfall (Etzemüller et al., 2022; Ravanel et al., 2023; Cathala et al., 2024). Field measurements strongly suggest the presence of permafrost, and ground temperature measurement allows to constrain realistic forcing datasets as input for the thermal model. The model is used to spatialise the permafrost distribution and understand its evolution. Geophysical data provide an independent and indirect dataset that gives another indication regarding permafrost distribution. The petrophysical analysis helps to interpret geophysical surveys on the basis of temperature.

The combination of these methods demonstrates that permafrost warming may have been a preconditioning factor of the rockfall. Indeed, ERT suggests the presence of permafrost at the scar, which is confirmed by heat transfer models. The rockfall likely occurred in a region with warm permafrost that has been warming since at least the 1990s, with an accelerated warming trend since 2012 towards the melting point. This warming trend and the sharp acceleration in ground temperature from 2012 are also highlighted by Cathala et al. (2024) at the Vallon d'Étache site, located 40 km northwest of the Crête des Grangettes. This

warming could have contributed to the rockfall by reducing the ice strength within fractures (Krautblatter et al., 2013). Additionally, the geophysical survey revealed a zone of low resistivity, indicating an area where an old permanent snow patch might have limited permafrost formation due to the insulating effect of the snow.

The significant rainfall that occurred 19 days before the event may have contributed to the rockfall triggering. Water infiltration likely increased hydrostatic pressures and/or facilitated the erosion of ice within fractures (Hasler et al 2011; Magnin et Josnin, 2021). The delay between the rainfall and the rockfall occurrence could be explained by the time required for water to percolate in depth. This phenomenon has been observed for landslides (e.g., Liu et al., 2022), rockslides (e.g., Helmstetter and Garambois, 2010) and rock avalanches (e.g., Cathala et al., 2024), where the response time between rainfall and failure can range from a few days to a month. To further investigate the hypothesis of precipitation as a triggering factor, water mass balance modelling could be employed to estimate the water column available for infiltration at the time of the rockfall (Westermann et al., 2023; Cathala et al., 2024), and used to simulate coupled heat and water transfers (Magnin and Josnin, 2021).

Beyond thermal and hydrological factors, the rockfall occurred in a lithological and structural context prone to instability, with poorly cohesive rocks and the presence of a fault in the detachment zone. This fault is visible in a 2009 orthophotograph (fig. 2C), corresponding exactly with the scar from September 2020 rockfall. In the field, we also identified another fracture which could be an indicator to further instability (fig. 16). Monitoring the progression of this fracture in the coming years would be valuable to anticipate a possible future rockfall (e.g., Di Matteo et al., 2017).

**Figure 16: Fracture visible approximately 10 m from the scar, parallel to the latter.** The fracture is approximately 5 cm wide. The photography was taken on the Profile 1, between sensors Block #3 and Block #4 (ph: 20/09/2022).

*Figure 16 : Fracture visible à environ 10 m de la niche d'arrachement, parallèle à cette dernière. La largeur de la fracture est d'environ 5 cm. La photographie est prise le long du Profil 1, entre les capteurs Block #3 et Block #4 (ph. : 20/09/2022).*

## 5.2. Limitations of methods and outlooks for investigating permafrost-affected rockwalls

The Crête des Grangettes presents a complex case study due to the challenging topography. The rockfall originated from a rock outcrop, where thermal flow is multidirectional. Temperature differences between the north, east, and west slopes can reach several degrees Celsius, creating a 3D heat flow across the spur (Noetzli et al., 2007). The 2D heat transfer modelling neglects the heat flow from the sides, leading to a potential bias in simulated temperatures. However, the simulations that we performed show that accounting for an additional heat flow changes the temperature, but has almost no effect on the overall evolution of the temperature, leading to robust indications regarding permafrost degradation. Both simulations consistently show that the temperature along the scar reached unprecedentedly high values (warmer than  $-1$  °C, but still negative) during the months prior to the rockfall, which is in good qualitative agreement with the observation of the presence of ice in the scar, that confirms the existence of permafrost, and the persistence of ice within fractures (Magnin et al., 2024).

In addition, the heat transfer model does not account for snow cover, which can significantly modify the heat balance at the surface (Cathala et al., 2024). Yet, the effect of snow cover at this site is probably limited, as evidenced by the marked temperature variations over the whole snow season, for seven out of eight sensors placed in the debris slope, in which air circulation can strongly affect subsurface temperature (e.g., Delaloye and Lambiel 2005; Wicky and Hauck, 2017). Using an energy balance model including snow would be a good solution to account for the effect of snow cover on surface temperature, which could then be used as input in the heat transfer model (Cathala et al., 2024). In the Crête des Grangettes site, regular and local measurements of the snow cover would be required to validate the modelled snow of such an energy balance model, but such measurements are not available yet for this study site.

The site's configuration posed challenges for geophysical surveying because the area of interest, specifically the scar, is located on the rock wall. As a result, Profile 1 was conducted upstream of the release area, with the extremity of the ERT cable placed close to the scar. This configuration induces a low coverage and accuracy at both ends of ERT profile. This setup complicates result interpretation, as the scar is barely visible on the tomogram, and the information derived from ERT in the release area is an extrapolation, potentially leading to edge effects. To address this limitation, Profile 2 was aligned parallel to the scar, and from



a loop close to the scar allowing more coverage of the underground in this case (*i.e.*, more measurements in this zone). Finally, the 3D inversion allows to use all data to get an imagery 3D of the site. However, it would have been valuable to make a third profile placed within the rockwall (specifically in the scar) to obtain a deeper resistivity picture of the rockwall (Magnin et al., 2015b). Unfortunately, this approach is challenging to implement due to the ongoing instability of the area.

Furthermore, the heterogeneous lithology of the study area, consisting of rocks with varying petrophysical properties (porosity, saturation, water content, air content), makes it difficult to distinguish between frozen and unfrozen areas. Petrophysical analyses reveal a freeze/no-freeze transition ranging between 4 and 10 k $\Omega$ m for sample GRAN-01, while for sample GRAN-02, the transition occurs between approximately 1 and 32 k $\Omega$ m (fig. 14). The resistivity of the two samples varies significantly depending on the sample, that make interpretation of the geophysical surveys difficult. Considering the very distinct lithologies along the profiles and the variations in resistivity measured in the laboratory as a function of temperature, it would not be consistent to average the two curves to convert the profiles into temperatures, like the interpretation developed by Duvillard et al. (2021) whose study focuses on a homogeneous site with a unique formation of granite. Another way of interpreting the results in a heterogeneous environment such as the one studied would be to distinguish the layers corresponding to each lithology along the profiles in order to apply the temperature models to each geology (clustering method; Ward et al., 2014; Wang et al., 2021; Isunza Manrique et al., 2024).

Finally, this rockfall presents a highly complex case study for understanding propagation processes. The deposit spread across existing debris slopes and screes, which may have been removed by the Grand Tabuc torrent, makes difficult the identification and mapping of the runout. These factors significantly limit the retro-analysis of the rockfall propagation.

## **6. Conclusions**

By developing a comprehensive approach at several scales, we have obtained valuable information to decipher the permafrost-related processes that potentially preconditioned and triggered the Crête des Grangettes rockfall in September 17, 2020. The main findings of this study are summarized as follows:

- The long-term meteorological data analysis shows that the exceptional warmth of spring 2020, combined with consistently high summer temperatures since 2015, may have contributed to preconditioning the rockfall.
- The analysis of meteorological data over three months before the rockfall shows substantial (though not exceptional) rainfall 19 days prior to the event. This episode may have been a triggering factor by increasing hydrostatic pressures within fractures and/or eroding ice-filled joints.
- At the point scale, we analyse GST fluctuations through continuous monitoring (2021-2022). The records show MARST ranging from  $-0.26^{\circ}\text{C}$  (NW) to  $+1.88^{\circ}\text{C}$  (W) in the rockwalls around the scar, and MAGST ranging from  $-2.85^{\circ}\text{C}$  to  $+1.18^{\circ}\text{C}$  in debris slopes. Considering that GST was measured during a year with air temperatures  $1.45^{\circ}\text{C}$  above the 1991-2020 average, the GST suggests that conditions across the study site are favorable to the presence of permafrost.
- At the pluri-decametric scale, two geophysical profiles around the rockfall scar show high resistivity values ( $>10\text{ k}\Omega\text{m}$ ), that correspond to frozen materials, as confirmed by laboratory study carried out on two samples from outcrops. These measurements show the presence of permafrost around the rockfall scar.
- Heat transfer simulations suggest warm permafrost ( $> -2^{\circ}\text{C}$ ) at the failure depth, with unprecedentedly high temperatures reached around the scar line before the rockfall occurred. The modelled warming rate at 22 m depth is  $(0.48\pm 0.12)^{\circ}\text{C}$  per decade since 2015.

From the observations and data analysis, it results that the Crête des Grangettes rockfall on September 17, 2020 was likely triggered by permafrost warming with warm permafrost approaching  $0^{\circ}\text{C}$ . Over the longer term (decade), the permafrost warming may have resulted from the heat conduction from the surface, as suggested by the heat transfer models. This process, along with the mechanical weakening of ice-filled joints, likely contributed to the destabilisation of the rock slope.

Finally, our multi-method approach provides valuable insights and data for scientific communities focused on permafrost and rock slope failure hazard assessment. It emphasizes the complexity of the processes that can lead to rock slope failures across various spatial and temporal scales, highlighting the need for further research on both comprehensive assessments of specific events and detailed analyses of individual processes.

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