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Rock glacier destabilization in the French Alps: insights from regional and local scale assessments

Marco Marcer

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THÈSE

Pour obtenir le grade de

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GRENOBLE ALPES**

Spécialité : **Géographie**

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préparée au sein des **Laboratoires PACTE** et **EDYTEM**
dans **l'École Doctorale Terre Univers Environnement**

Déstabilisation des glaciers rocheux dans les Alpes françaises: une évaluation à l'échelle régionale et locale

Thèse soutenue publiquement le **19 Décembre 2018**,
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“Homines id quod volunt credunt”

Gaius Iulius Caesar

Abstract

As occurring to several geosystems on our planet, mountain permafrost is threatened by climate change as prolonged warming may compromise the geotechnical properties of the frozen ground. As result, increasing occurrence of rockfall activity, thermokarst formation and rock glacier acceleration was observed in the past decades. Rock glacier destabilization, a process that compromises the structural integrity of these landforms, seems to be linked to atmospheric warming, gaining interest in the past years. The destabilization, which may be triggered by warming permafrost or mechanical stress, is characterized by an anomalous acceleration of the landform and the occurrence of specific features such as cracks and crevasses on its surface. Although the occurrence of these processes is mostly transitory, determining a *crisis* phase of the landform, in exceptional cases it may lead the rock glacier to structural collapse.

This PhD thesis provided an assessment on the occurrence and related processes of rock glacier destabilization in the French Alps. At first, the spatial occurrence of debris permafrost was assessed in order to provide the permafrost distribution map of the French Alps, a tool that was necessary to evaluate permafrost conditions at rock glaciers sites. The second step consisted in an identification of destabilized rock glaciers in the region, which was done by multiple orthoimages interpretation aimed to identify features typically observable on destabilized rock glacier. Once identified the destabilized rock glaciers it was possible to analyse the typical topographical settings in which destabilization occurs and to spot those landforms that are susceptible to experience this phenomenon. After these efforts at the regional scale, the focus was shifted towards local scale investigations at the Lou rock glacier, a partially destabilized landform that, due to frontal failure, in August 2015 triggered a debris flow that caused significant damages to buildings. The analysis aimed to better define the circumstances of this event, focusing on preconditioning, preparatory and triggering factors and their interaction with the destabilization process.

The results provided interesting insights on the issue of destabilizing rock glaciers in the region. Permafrost distribution modeling demonstrated the large extents of the periglacial zone in the region as it can be found in debris slopes above 2300 - 2900 m.a.s.l. depending upon solar exposure and regional precipitation characteristics. Rock glacier destabilization was observed on 46 landforms, i.e. the 12% of the active rock glaciers. Destabilization was found to be more likely to occur in specific local topo-climatic conditions, consisting of north facing, steep and convex slopes at the lower margins of the permafrost zone. A large number of rock glaciers currently not showing destabilization was found to be located in these conditions and suggested to be susceptible to future destabilization. As demonstrated by the Lou rock glacier analysis, destabilization was found to be a relevant phenomena in the context of permafrost hazards. At this site, rock glacier destabilization was linked

to a rapid frontal advance towards a torrential gully. This process seemed to have increased the site predisposition to frontal failure as a mild rainstorm was sufficient to trigger the event.

Despite methodological uncertainties, results indicated that destabilization occurrence is widespread and it may rise the hazard level of a site connected to human infrastructures. Therefore, it is suggested that, where it has been modelled and where stakes may be at risk downslope, rock glacier destabilization deserves to be more carefully investigated. In this sense further efforts should focus towards a better understanding of the destabilization process by site monitoring as well as towards a comprehensive hazard assessment linked to this phenomenon.

Résumé

Le permafrost de montagne est menacé par le réchauffement atmosphérique, une évolution qui s'accompagne de l'augmentation des phénomènes tels que les chutes de pierres, la formation de thermokarsts et l'accélération des glaciers rocheux. La déstabilisation des glaciers rocheux, qui compromet l'intégrité structurelle de ces formes, semble liée au réchauffement atmosphérique, et a suscité un intérêt grandissant au cours des dernières années. Ce phénomène, qui peut être provoqué par le réchauffement du pergélisol ou des contraintes mécaniques externes, est caractérisé par une accélération anormale des glaciers rocheux affectés, et par l'apparition des signes géomorphologiques telles que des fissures et des crevasses à sa surface. Bien que ce processus peut être transitoire, il peut déterminer une phase de crise amenant le glacier rocheux à un effondrement.

Cet étude se préfixe de fournir une première évaluation des phénomènes de déstabilisation de glacier rocheux à l'échelle des Alpes françaises. Dans un premier temps, l'empreinte spatiale du pergélisol a été évaluée afin de produire une carte de répartition du pergélisol régionale, un outil nécessaire pour estimer l'état du permafrost dans les glaciers rocheux. La deuxième étape a consisté à identifier les formes déstabilisées grâce à une observation ponctuelle des images aériennes afin d'identifier les caractéristiques typiquement observables sur les glaciers rocheux déstabilisés. Il est alors possible de comprendre les conditions topoclimatiques typiques dans lesquelles se produit ce phénomène et de repérer les formes susceptibles de subir ce processus. Enfin, les efforts ont été concentrés sur le glacier rocheux du Lou, déstabilisé, qui, du fait d'un détachement de couche active, a conduit à une lave torrentielle en Août 2015. L'analyse a visé à mieux définir les circonstances de cet événement, en mettant l'accent sur les facteurs de préconditionnement, de préparation et de déclenchement et sur leur interaction avec le processus de déstabilisation.

Les résultats ont fourni des informations riches sur la zone périglaciaire de la région. La modélisation de la répartition du pergélisol a mis en évidence les étendues de la zone périglaciaire dans la région qu'on peut trouver sur les pentes de débris au-dessus de 2300 - 2500 m.a.s.l. en fonction de l'exposition solaire et des caractéristiques régionales des précipitations. L'observation des photographies aériennes a permis d'observer 46 formes en cours de déstabilisation, soit 12% des glaciers rocheux actifs des Alpes françaises. Il apparaît que la déstabilisation est plus susceptible de se produire dans certaines conditions topoclimatiques locales spécifiques, en particulier dans des pentes exposées au nord, raides et convexes situées aux marges inférieures de la zone de pergélisol. Un grand nombre de glaciers rocheux ne présentant actuellement aucune déstabilisation sont donc susceptibles d'être affectés par une déstabilisation future. L'analyse du glacier rocheux du Lou a révélé que la déstabilisation est liée à une avancée rapide du front vers un ravin torrentiel. Ce processus semble avoir accru la prédisposition des matériaux détritiques du front à être mobilisés par du ruissellement, des précipitations relativement modérées ayant suffi à

déclencher l'événement. Malgré les incertitudes liées aux méthodes impliquées, les résultats suggèrent que les conditions favorables à la déstabilisation sont fréquentes, et que cette dernière peut augmenter le niveau de risque si le site est connecté à des infrastructures humaines. Des efforts supplémentaires doivent donc être entrepris, afin d'améliorer la compréhension de ces processus, notamment par la surveillance des sites ainsi que par une évaluation locale complète des cascades de processus liés à ce phénomène.

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Well, that's it. These three years have passed so quickly that I am quite glad to have an excuse to take a moment and think back. Moving to Grenoble and working on this project was a remarkable experience, a period that I will gladly remember. It is with great pleasure that I spend these lines to thank all the people that made it possible.

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List of Abbreviations

| | |
|---------------|---|
| AIC | A kaike I nformation C riterion |
| AMOC | A tlantic M eridional O verturning C irculation |
| APIM | A lpine P ermafrost I ndex M ap |
| AUROC | A rea U nder R eceiver O perating C haracteristic |
| CNRS | C entre N ationale de la R echerche S cientifique |
| (d)GPS | (d ifferential) G eographical P ositioning S ystem |
| EDF | E lectricité D e F rance |
| EDYTEM | E nvironnement et D Ynamiques T erritoriales de M ontagne |
| ERT | E lectric R esistivity T omography |
| GAM | G eneralized A dditive M odel |
| GCPs | G round C ontrol P oints |
| GIS | G eographical I nformation S ystems |
| GLM | G eneralized L inear M odel |
| GS | G reenland S tadial |
| HCO | H olocene C limatic O ptimum |
| LIA | L ittle I ce A ge |
| LGM | L ast G lacial M aximum |
| MAAT | M ean A nnual A ir T emperature |
| ONF | O ffice N ational des F ôrets |
| PACTE | P olitiques A ction T erritoir |
| PFI | P ermafrost F avorability I ndex |
| PISR | P otential I ncoming S olar R adiation |
| PTP | P otentially T hawing P ermafrost |
| RAMMS | R apid M ass M ovements |
| SAGE | S ociété A lpine de G éotechnique |
| SfM | S tructure from M otion |
| SRT | S eismic R efraction T omography |
| UAV | U nmanned A erial V ehicle |
| UGA | U niversité G renoble A lpes |
| USMB | U niversité S avoie M ont B lanc |
| WMS | W eb M apping S ervice |

Dedicated to my family.

Chapter 1

Introduction

1.1 General context: climate change and frozen ground

Climate change is a major issue in present day society and one of the greatest challenges of our generation (IPCC, 2007; IPCC, 2013). Although a complete understanding of the 20th century climate change may still be debated by some individuals or governments, the existence of climatic variability in the present and in the past is *certain* (Masson-Delmotte et al., 2013). The main feature of the 20th century climate change is represented by a global mean temperature increase of 0.85°C between 1880 and 2012 (Hartmann et al., 2013), a phenomenon that threatens fundamental aspects of our communities and nations ranging from fresh water availability to agricultural production (Porter et al., 2014). Therefore, it is necessary to take action to identify the effects of climate change on our society in order to adapt our lifestyles, economy and infrastructures.

The present study takes place in the broad issue of the effects that climatic fluctuations have on the frozen water of our planet, i.e. the *cryosphere*. The cryosphere comprehends ice sheets, ice shelves, glaciers, sea ice, fresh water ice, snow cover and permafrost, and covers a large part of the planet (IGOS, 2007). This frozen world is strongly sensitive to climate and responds promptly to its fluctuations. Temperature increase can cause a phase shift from frozen to liquid state, triggering several processes that endanger the existence of the cryosphere and may threaten the populations that live in contact to it. As a result, the cryosphere is one of the Earth system most concerning in the context of climate change (Vaughan et al., 2013). The most popular of these processes is certainly glacial shrinkage, due to its impressive visual impact (Figure 1.1). Since the end of the 19th century alpine glaciers started to retreat, losing 50% of their surface (Zemp et al., 2006). Present glacial extension reached levels as low as during warm periods of the Holocene, several thousands of years before present. However, due to the delayed response time between temperature increase and glacial shrinkage, alpine glaciers are expected to continue losing mass in the near future, even in a scenario of no further increasing temperatures (Vincent et al., 2014). Several small glaciers have already disappeared and model simulations suggest that larger alpine glaciers may encounter the same fate within the end of the next century (Le Meur et al., 2007). The disappearing ice leaves a large

footprint of naked bedrock and debris in the alpine valleys, and, due to the rapidity of the process, the public is often impressed by this phenomenon (Figure 1.1).



FIGURE 1.1: Shrinkage of the [Mer de Glace glacier](#), in the Mont Blanc massif (France), between 1909 and 2017.

Although glaciers are an impressive witness of the struggle of the cryosphere and a powerful way to communicate to the public the sensitivity of the environment to climate fluctuations, they only tell half of the story. Beneath the surface of the cold regions can be found a layer of permanently frozen ground, i.e. the so called *permafrost*. Permafrost occurs at low temperatures, found either at high latitudes or elevations, and its distribution and composition is influenced by the climate (Osterkamp, 2007; Harris et al., 2003). Although frozen ground covers a large extent of the dry lands (9- 14%, Brown et al., 1997, see Figure 1.2), the consequences of climate change are not as striking to observe as in glaciers, due to the fact that permafrost itself is mostly not observable. Increasing temperature causes warming ground and loss of ice content in favour of higher water content, a process generally referred as *permafrost degradation* (Streletskiy et al., 2015). One of the most concerning issue regarding permafrost degradation is due to the fact that permafrost stores almost twice the amount of carbon already present in the atmosphere. Degradation enhances greenhouse gas emissions, possibly triggering a positive feedback of temperature warming (Tarnocai et al., 2009; Schuur et al., 2009; Schaefer et al., 2011; Koven et al., 2011). For this reason, permafrost is a critical component of the cryosphere in the context of global climate change.

Permafrost degradation has also an important role at the local scale. In mountain regions, permafrost degradation may compromise steep ground stability causing mass movements that represent a hazard for the local population (Haeberli et al., 1997; Haeberli et al., 2010; Harris, 2005; Bodin et al., 2015). Loss in stability takes place in ice-rich slopes as permafrost thaws leaving the soil matrix ice free (Davies et al., 2001), leading to an increased susceptibility to rockfalls, landslides and debris flows occurrence. These events may be intense and potentially hazardous if located in proximity to inhabited areas. Among the most impressive cases of failures attributed to permafrost degradation close to urbanized areas, are mentioned here the

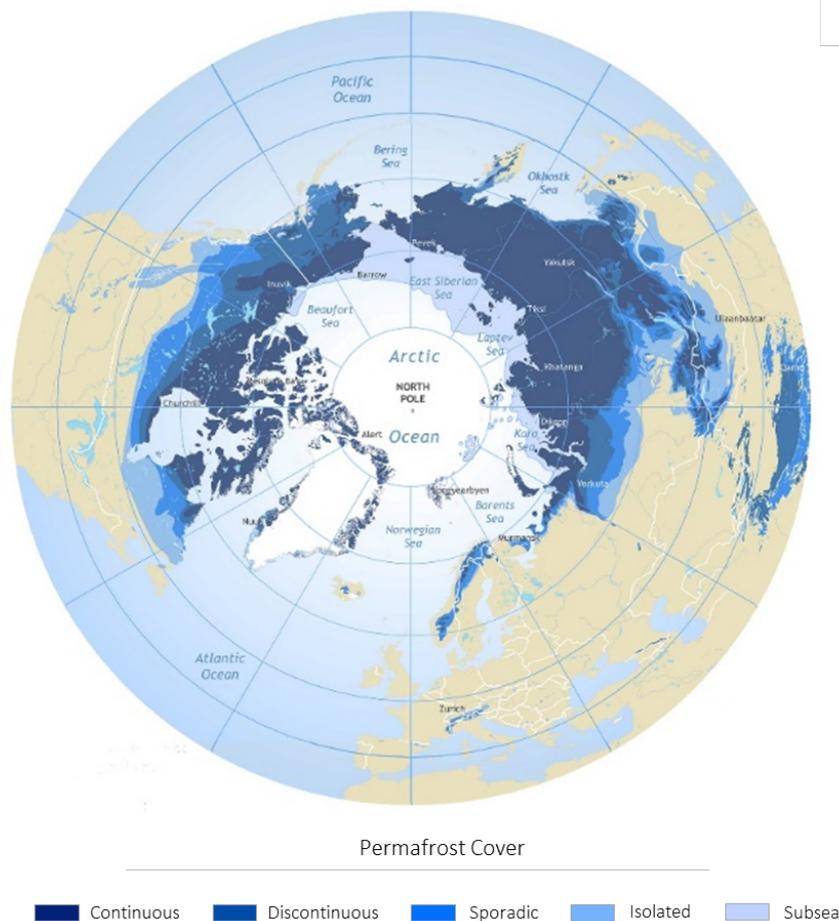


FIGURE 1.2: International Permafrost Association permafrost distribution map in the Northern Hemisphere (Brown et al., 1997).

Dru collapse in 2006 (almost 300 000 m³, Ravanel and Deline, 2008) and the Cengalo rock slide in 2017 (3 000 000 m³).

Although failures in vertical high mountain rock walls produce impressive events, also permafrost in loose debris is significantly subjected to climate change, possibly representing a source of hazards. When loose debris are sufficiently rich in ice, they naturally creeps downslope at few centimetres or decimetres per year, forming peculiar landform called *rock glaciers* (Wahrhaftig and Cox, 1959). Rock glaciers are widely observed to respond to warming with increasing creeping rates (Delaloye et al., 2008b; Kellerer-Pirklbauer et al., 2018), reaching in some cases values of several meters per year. In the case of extremely high acceleration, a series of geomorphological features typical of landslides, as crevasses and cracks, develop on the rock glacier surface (Roer et al., 2008; Delaloye et al., 2013). This process referred as *destabilization* is hypothesized to be caused by a transition between creeping to sliding process of the permafrost body (Roer et al., 2008). Rock glacier destabilization may be triggered either by mechanical shock, e.g. overloading by rockfall, or by permafrost degradation (Roer et al., 2008; Delaloye et al., 2013; Lambiel, 2011; Eriksen

et al., 2018). Destabilization may last few decades and bring the landform to abnormally high velocities, reaching several tens of meters per year (Delaloye et al., 2013; Eriksen et al., 2018; Vivero and Lambiel, 2019).

1.2 The POIA - PERMARISK project

In 2006 a [destabilized rock glacier](#) in the Southern French Alps collapsed, causing a 250 000 m³ landslide (Bodin et al., 2016). Thanks to the remoteness of the area interested, the event did not endangered anthropic areas. Nevertheless, this event was unexpected and highlighted the absence of a proper knowledge on this issue by the French local authorities. As response, the past decade was remarked by a series of efforts involving the local scientific community, in particular the IGA of Grenoble and the EDYTEM laboratory of Bourget le Lac, and the French government, represented by the National Environmental Protection Agency (RTM), aimed to identify and monitor rock glaciers in the French Alps. In 2008 was launched the [PermaFrance](#) project, an effort of permanent monitoring of several permafrost reference sites in the region. In 2015 the first inventory of rock glaciers in the region was completed (Roudnitska et al., 2016). With this database an overview of all the permafrost related landforms in the region was made available, unlocking a huge potential for hazard identification and research questioning.

The project POIA - PERMARISK takes place in the continuity of these efforts, aiming to provide a comprehensive assessment of hazards concerning permafrost-related processes in the French Alps. The project is funded by the ERDF (European Regional Development Fund) through the POIA research program and by the Auvergne-Rhône-Alpes Region through the ARC-3 scholarship. The project is realised in collaboration between the research institutions of PACTE and EDYTEM, part of the UGA and USMB respectively. Since the project involves several actions focused on different processes of the permafrost zone in the region, the project has several partners, involving the RTM, EDF and the two private companies IMSRN and SAGE, giving expertise and support on the different subjects. The project started officially in October 2015 and will terminate in December 2019, aiming primarily to produce a series of tools for the local authorities for the identification of potentially hazardous permafrost sites. Genuine scientific questioning has also a relevant role in the project as the efforts aim to contribute to some relevant questions concerning the permafrost related processes and their interaction with the climate.

1.3 The PhD project – settings and manuscript organisation

The present PhD finds place within the POIA-PERMARISK project by focusing on the specific issue of rock glaciers destabilization. The main aim of the PhD project is to quantify and analyse the phenomena linked to this process in the French Alps

by regional and local scale investigations. Since this goal is broad, the PhD is divided into three research axis that allow more targeted and efficient efforts. The first research axis aims to quantify the extension of the surfaces hosting mountain permafrost in the region by producing a permafrost distribution map. The second research axis assesses the occurrence of destabilized rock glaciers at the regional scale. The third and last research axis investigates a study site where a mass movement triggered on a destabilized rock glacier caused a debris flow, aiming to understand the role of permafrost destabilization in this event. The knowledge achieved by these three research axis is then combined to contribute to the understanding and assessment of destabilized rock glaciers in the French Alps.

This manuscript aims to describe the PhD research development and results. The manuscript consists in a main body and three research articles, covering a research axis each. The main body is divided into nine parts: scientific setting description, three article synopsis, three articles manuscripts and a conclusion. In the scientific setting (Chapter 1) are provided the key concepts to understand the context of the project POIA-PERMARISK, the positioning of this PhD in the project and its relevance to the scientific context. This involves the definition of permafrost (section 2.1), climate change (section 2.2) and its effects on the frozen ground (section 2.3). The three articles composing the scientific corpus of the project are then presented. For each study are outlined the motivations and the aims in order to provide to the reader the continuity of the PhD project. Materials, methodologies and results are only briefly presented, in order to let the reader focus on the significance of the study in the project context. A conclusion of the PhD project is then drawn (Chapter 6). In this chapter the achievements and contributions of the project to the scientific community and local risk managers will be outlined in a continuous perspective with respect to past and future efforts in this scientific setting. Finally, article manuscripts are presented as appendix.

Chapter 2

Scientific Setting

This chapter aims to delineate the scientific context of the PhD project. Here, the reader is provided with the fundamental background necessary to understand the project frame and aims. An overview of permafrost science is given in Section 2.1, with a focus on rock glaciers, which are the main object of this PhD. An overview of post glacial climatic variability is also provided in section 2.2 in order to better understand the present context of climate change in which the permafrost is apprehended in this study. This allows to better frame the impacts to the mountain permafrost zone and related hazards due to the post-industrial temperature rise with a focus on rock glaciers (Section 2.3). Finally, once the reader is made aware of the whole picture, the detailed description of the POIA-PERMARISK project is given in section 2.4, allowing to finally define the relevance and main objectives of the present PhD in section 2.4.2.

2.1 Permafrost: the perennially frozen ground

The perennially frozen ground is commonly called by the acronym *permafrost* (Permanent Frost). The permafrost denomination describes a thermal state of the ground whose temperature has been measured equal or below 0°C for at least two consecutive years (Van Everdingen, 2005). Despite this apparently simple definition, permafrost exists in a wide variety of forms that delineate different processes and conditions involved. Permafrost is defined as *cold permafrost* when its temperature is remarkably lower than 0°C (e.g. -0.5°C), while it is defined as *temperate* when its temperature is close to 0°C (Delaloye, 2005). Permafrost may be *dry* or *ice rich* depending on the presence of frozen water in the soil matrix. In temperate permafrost there may be the coexistence of liquid water which maintains the ground temperature equal to the melting point, due to latent heat. Nevertheless, liquid water may exist also at negative temperature depending on the ground salinity, structure or pressure (Dobinski, 2011).

Permafrost is hidden below the so-called *active layer* (Muller, 1943). This layer is characterized by the seasonal occurrence of freeze and thaw cycles due to the thermal exchanges with the atmosphere. The active layer has a role of heat exchanger that allows, through insulation from solar radiation, to maintain cool the frozen

ground (Harris and Pedersen, 1998). The limit between permafrost and active layer is the depth at which temperature never rises above 0°C despite seasonal fluctuations (Figure 2.1). The depth of the zero annual amplitude, is the point in which seasonal temperature fluctuations are smaller than 0.1°C (Smith and Riseborough, 2002). The vertical extension of the permafrost layer is limited by the existence of the geothermic heat flux that warms the ground (Osterkamp and Burn, 2003). The base of the permafrost layer is the point where ground temperature is zero at the equilibrium between geothermic heat flux and permafrost temperature.

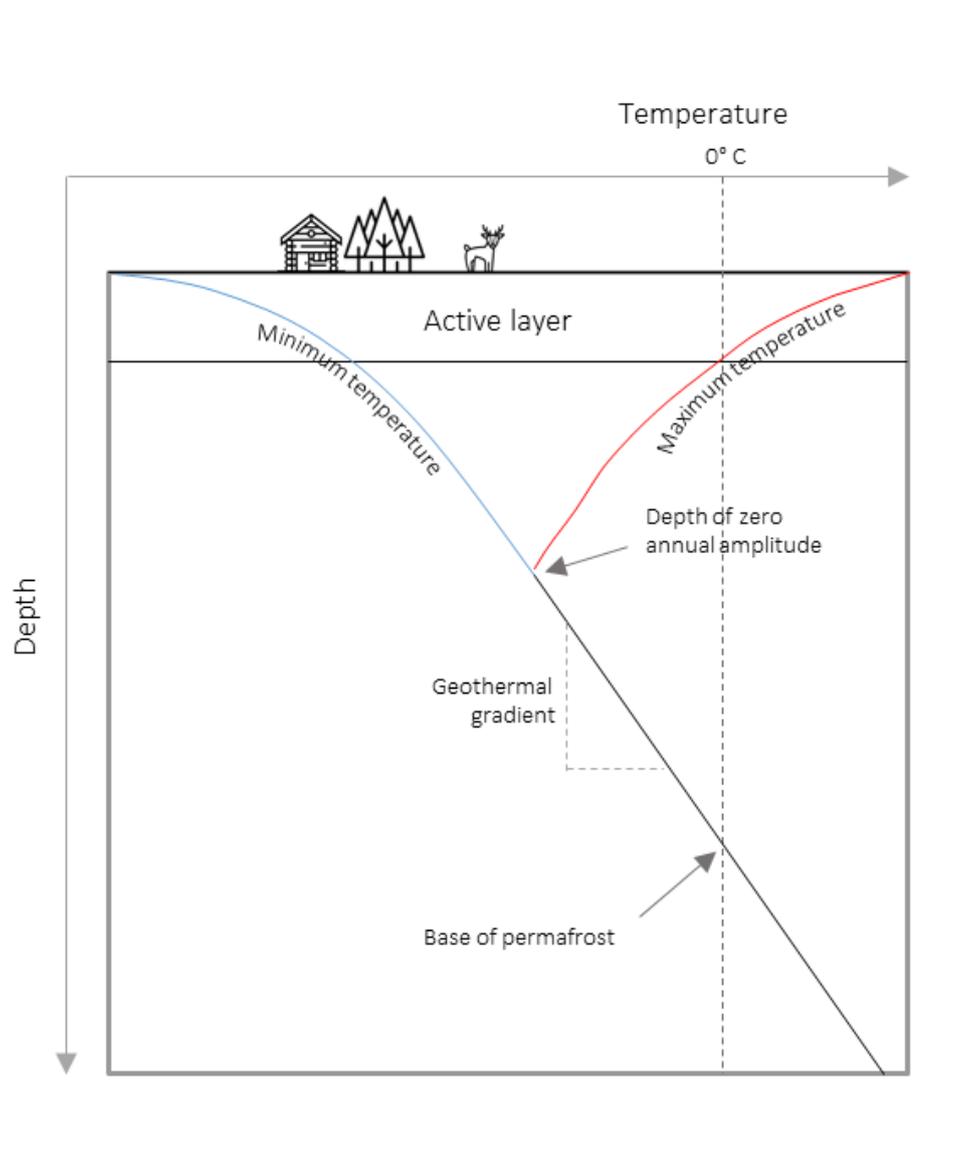


FIGURE 2.1: Conceptual scheme of the permafrost layers and characteristics (adapted from Osterkamp and Burn, 2003).

The attributes that describe the frozen ground, i.e. temperature, layers thickness and water content, depend on several processes and their interactions. Although mean annual air temperature is strongly correlated with permafrost temperature (Harris et al., 2003), its attributes are subjected to strong spatial variability due to

local ground characteristics, solar radiation exposure (Magnin, 2015), water content (Scherler et al., 2013) and snow cover (Apaloo et al., 2012). In all this, permafrost characteristics are still influenced by past climates and ground properties (Majorowicz, 2012). Despite all these elements that characterize permafrost as a spatially variable and complex phenomenon, frozen ground is traditionally grouped into two categories: arctic and mountain permafrost.

2.1.1 Arctic permafrost

Although not relevant in the context of the present PhD, for sake of completeness is briefly presented the Arctic permafrost, i.e. permafrost existing at high latitudes. This permafrost category represents the largest portion of permafrost on the planet, accounting for 76 % of the total permafrost area and mostly located in the Northern hemisphere (Figure 1.2, Brown et al., 1997). Permafrost latitudinal occurrence is often referred as *discontinuous* at lower latitudes to *continuous* at higher latitudes. While the existence of frozen ground in discontinuous permafrost areas is discriminated by local features, continuous Arctic permafrost is widespread on the landscape. Continuous permafrost is several tens of meters thick, although variability is high and depth can reach more than 500 meters (Brown, 1960). The typical active layer is few tens of centimetres thick which, at lower latitudes, hosts the arctic flora. If water content is sufficiently high, permafrost can be ice-rich and ice can be present in the form of massive ice or cemented ice depending on several factors as air temperature and water source. In ice rich conditions, Arctic permafrost may originate several peculiar surface landforms as pingos, palsas and patterned ground (Figure 2.2).



FIGURE 2.2: Images from arctic permafrost sites. Massive ice beneath Arctic Tundra (a). A pingo in the National Pingo Landmark, Canada (b).

Since four million people live in the Arctic, permafrost related issues have a significant interest in these regions (ACIA, 2005). Frozen ground here hosts lakes and rivers and has an influence on the geotechnical properties of infrastructures and communication network. Modifications of the arctic permafrost zone can therefore

affect several aspect of arctic people life. Arctic permafrost has also a global relevance, as emerged in the past years due to the concerns on greenhouse gases storage in frozen ground which may be released in the atmosphere from permafrost thaw (e.g. Tarnocai et al., 2009). Due to the large amount of organic carbon stored in the arctic permafrost, this process may substantially increase greenhouse gas emissions.

2.1.2 Mountain permafrost

Mountain permafrost concerns high mountain ranges, as European Alps, Andes, Himalaya and Pyrenees. It represents 14% of the dry land permafrost and 70% of it is found in the Tibetan plateau (Bockheim and Munroe, 2014). Since air temperature decreases with elevation, permafrost is more likely to be found on higher areas creating a sort of invisible frozen ground belt around the mountains summits. Depending on the local climate, permafrost is found starting from a certain elevation called the *lower limit* of the permafrost zone. This limit may vary significantly across mountain ranges according to the elevation of the 0°C isotherm and precipitation patterns. In the European Alps (45°N) for example, permafrost can be found around 2600 m.a.s.l. (Boeckli et al., 2012b), while, for instance, in the Central Andes (30°S) it can be found above 3900 m.a.s.l. (Azócar et al., 2017).

The typical feature of mountain permafrost is the strong spatial variability caused by the complex topography. Although higher elevations are in general more suitable to permafrost existence, shading plays a relevant role, causing lower limits of permafrost to shift by several hundred meters of elevation within the same mountain summit depending on slope aspect (e.g Haeberli, 1985). Sharp ridges cause a strong discontinuity in permafrost characteristics (Magnin, 2015) and they influence permafrost distribution with three dimensional effects due to lateral heat fluxes (Noetzli et al., 2007).

Complex topography has a direct impact on wind field and therefore snow distribution (Winstral and Marks, 2002), a critical and complex parameter influencing permafrost at the local scale (Delaloye, 2005; Gruber, 2005). Snow cover has a significant insulating power when thicker than 0.6 m (Luetschg et al., 2008), although this value increases in coarser ground (Staub, 2015). Snow appearing in the early winter has a warming effect, protecting the ground from winter cold, while areas kept snow free by wind blow may be significantly cooler (Apaloo et al., 2012). Late disappearing snow can shield the ground from early summer warm temperatures, as it may be the case for avalanche or wind drift deposits (Lerjen et al., 2003). As these snow cover characteristics are influenced by complex interactions between topography and climate, ground temperatures are extremely discontinuous at the local scale and vary up to several degrees both spatially and temporally from one year to another (Gisnås et al., 2014).

Ground surface characteristics have also a critical importance, determining the existence of two different types of permafrost whether the ground is composed of bedrock or loose debris. Bedrock permafrost concerns headwalls and other surfaces

that do not accumulate debris. Its characteristics are affected by several parameters as degree of fracturing and presence of snow accumulation areas. Conduction is the main driver of heat transfer (Kohl, 1999) and massive ice can be found in cracks and joints (Figure 2.3a; for a detailed account of evidences, see Gruber, 2007). Ice content is however generally low, making the active layer thick and the permafrost sensitive to climate variations (Harris et al., 2003; Gruber, 2007; Magnin, 2015).

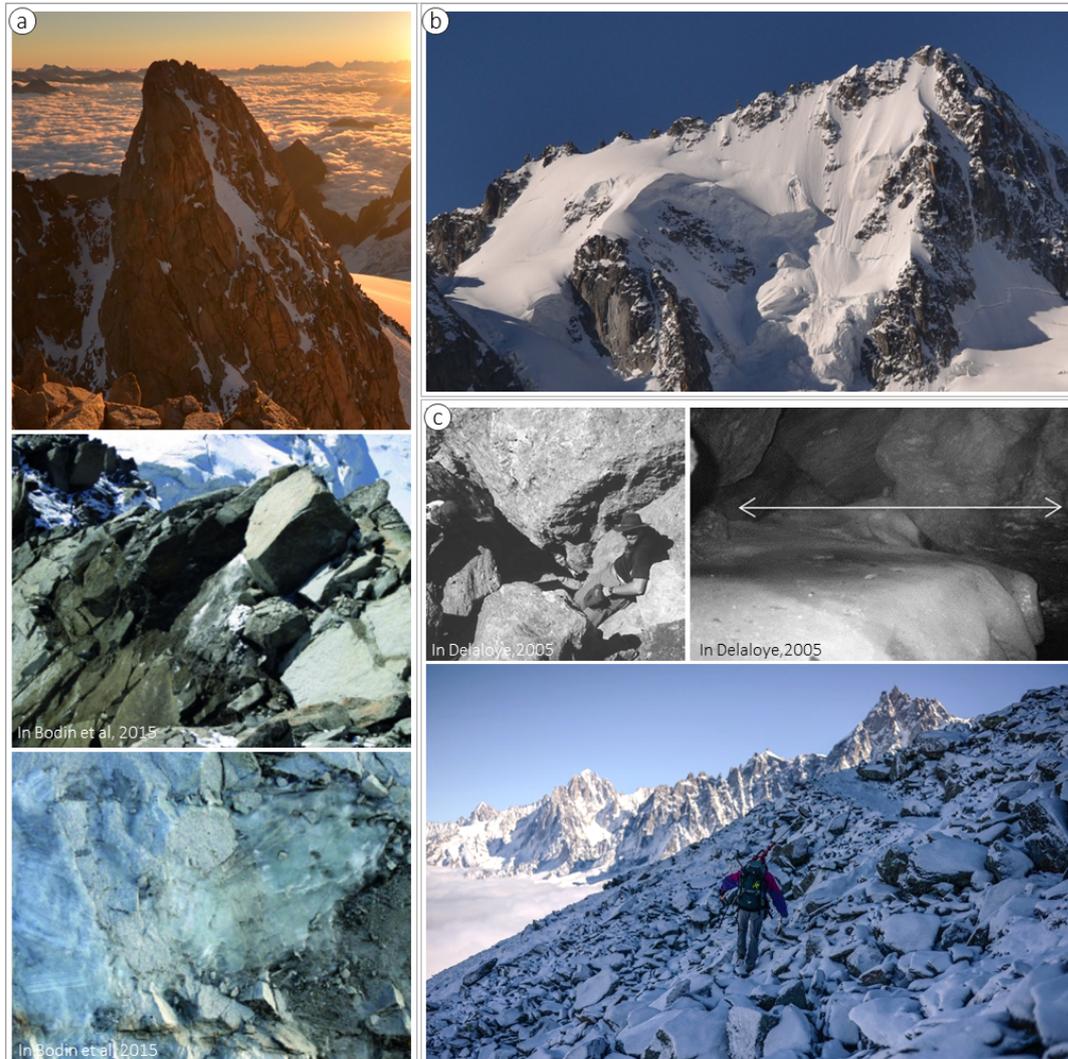


FIGURE 2.3: Examples of mountain permafrost in the European Alps. Bedrock permafrost (a) can be found in high mountain rockwalls and debris-free areas. Frozen water fills the joints of the bedrock. Frozen ground may cause the existence of hanging glaciers (b) by preventing basal sliding of the ice. Coarse debris (c) host permafrost at lower altitudes, thanks to the cooling effect due to ventilation of the active layer. Ice can be found within the ground matrix or in lenses.

Debris permafrost occurs in loose deposits of mountain sediments (Figure 2.3c). The active layer is composed of loose debris, where the heat exchange is dominated by air convection through the Balch effect (Balch, 1900 in Haeberli, 1985). In the pore volumes of the debris, warmer air is displaced by cold air thanks to its higher

density, cooling the ground and creating a thermal offset of the ground with respect to the air temperature (Hoelzle et al., 1999). This cooling effect can be very high, creating thermal offsets up to several degrees, depending on the granulometry of the debris (Hanson et al., 2005; Gubler et al., 2011). For this reason, debris permafrost in the European Alps can be found about 600 m.a.s.l. lower than bedrock permafrost (Boeckli et al., 2012a). Heat conduction becomes relevant at a depth where ground porosity is not sufficient to allow ventilation (Vonder Mühll et al., 2003).

In particular conditions, debris permafrost may be ice-rich, and frozen water can be found either in massive form or mixed with debris. Ice presence in debris permafrost can have two origins: glacial or periglacial. Glacial ice in permafrost generally exists thanks to the high supply of debris from headwalls in the ablation zone of a white glacier located in permafrost conditions. Debris cover may allow the formation of a permafrost thermal regime once it reaches the thickness of the active layer (Monnier and Kinnard, 2015). Glacial retreat may help this process since the interruption of glacial ice accumulation allows the thickening of a debris cover (Monnier and Kinnard, 2017). Glacial ice is typically massive, i.e. ice aggregated in debris-free pockets creating large bodies several meters thick.

The processes that bring to the ice formation in periglacial conditions are essentially two, firnification and congelation, and described by Delaloye (2005). The firnification is a process that involves the compaction of snow where the accumulation of snow and debris is enhanced in permafrost zones (e.g. avalanche cones). If the snow is constantly buried under layers of debris produced during spring freeze-thaw cycles that erode overlying rockwalls or debris transported by avalanches, then periglacial ice may form. This process tends to produce permafrost with high content. Congelation consists in the percolation and refreeze of melt or meteoric water into debris at below freezing temperatures. Congelation origins interstitial ice, i.e. ice mixed with fine-grained debris forming cemented ice. If ice segregation occurs then permafrost can form ice lenses (Figure 2.3c).

Rock glaciers

If the ice content in debris permafrost slopes is above 20-40%, then the ground loses its stiffness and reacts to gravity as a viscous fluid by creeping (Haeberli, 1985). Creeping mountain permafrost is the process at the base of the geomorphological landforms that are at the core of the present thesis: the rock glaciers (for a comprehensive review, see Barsch, 1996; Haeberli et al., 2006). Although creeping permafrost was known to the scientific community since the beginning of the 20th century (Capps, 1910), the first comprehensive study was realized only at the end of the 1950's (Wahrhaftig and Cox, 1959) and systematic interest started two decades later (Barsch, 1978; Haeberli, 1985). At the present date the scientific community has produced a significant effort to understand the features of these landforms. Current knowledge on rock glaciers is exposed here by treating their morphology, creeping characteristics and regional occurrence.

Morphology Rock glaciers are composed by two principal areas: the upslope rooting zone and the downslope creeping lobe (Figure 2.4). The permafrost aggregation and debris accumulation takes place in the rooting zone, a debris talus connected to the debris supplier (Barsch, 1996). If the rooting zone has a glacial origin, e.g. a frontal moraine, then the landform is defined as a *debris rock glacier*. On the other hand, if the rooting zone is connected to a headwall or debris talus, the landform is called a *talus rock glacier* (Humlum, 2000). In the rooting zone, debris are enriched in ice content until downstream creep is triggered. Instead of filling a valley as white glaciers, rock glaciers typically outstand from the surroundings as they have a lateral and frontal thickness of up to several tens of meters. Lateral and frontal talus can be up to 40–45° steep, depending on the creep rate.

Creep characteristics confer to these landforms geomorphological features typical of *lava-stream-like* cohesive flows (Barsch, 1992; Haeberli, 1985). These features exist at different scales and their occurrence vary from case to case (Monnier and Kinnard, 2017). At the meter scale the landform surface forms ripples and undulations. Ridges and furrows are features reaching several tens of meters width caused by the compression forces occurring to the creeping landform. At a larger scale sometimes is observable a superimposition of several lobes creating complex *polymorphic* landforms (Frauenfelder and Käab, 2000). If instead the landform presents a single well defined frontal lobe the rock glacier is defined as *monomorphic*.

The lithology of the debris supply plays a relevant role in the morphology of these landforms, as described by Ikeda and Matsuoka (2006). Densely jointed headwalls, as typically found in schists and platy limestone, produce fine debris that tends to host permafrost with lower ice content. The rock glaciers in these settings are defined as *pebbly* and often terminate in a rounded front located on valley sides slopes, causing an extending creep flow pattern that determines the absence of compressive features as ridges and furrows (Figure 2.5a). On the other hand, crystalline headwalls tend to be more resistant to mechanical weathering and produce coarser debris that feed the so-called *bouldery* rock glaciers (Figure 2.5b). These landforms are characterised by higher ice content, which may be found in massive cores, and by a compressive flow that confers the typical ridges and furrows aspect and high and steep fronts. Although exceptions may exist, pebbly and bouldery rock glaciers often differ also in size. Crystalline headwalls tend to be higher and supply large quantities of debris, allowing a more extensive rock glacier development. On the contrary, pebbly rock glaciers are usually smaller due to the reduced dimensions of the densely jointed headwalls (Matsuoka and Ikeda, 2001; Janke and Frauenfelder, 2007).

Rock glacier activity is also a critical factor shaping the morphology of these landforms, as described by Scapozza (2008). If a rock glacier creeps, then it is considered as active. Active rock glaciers present convex surface and steep front (> 30–35°) with signs of rockfalls. Although vegetation does not cover the rock glacier surface as boulders are constantly moved and tilted, frost resistant pioneer species

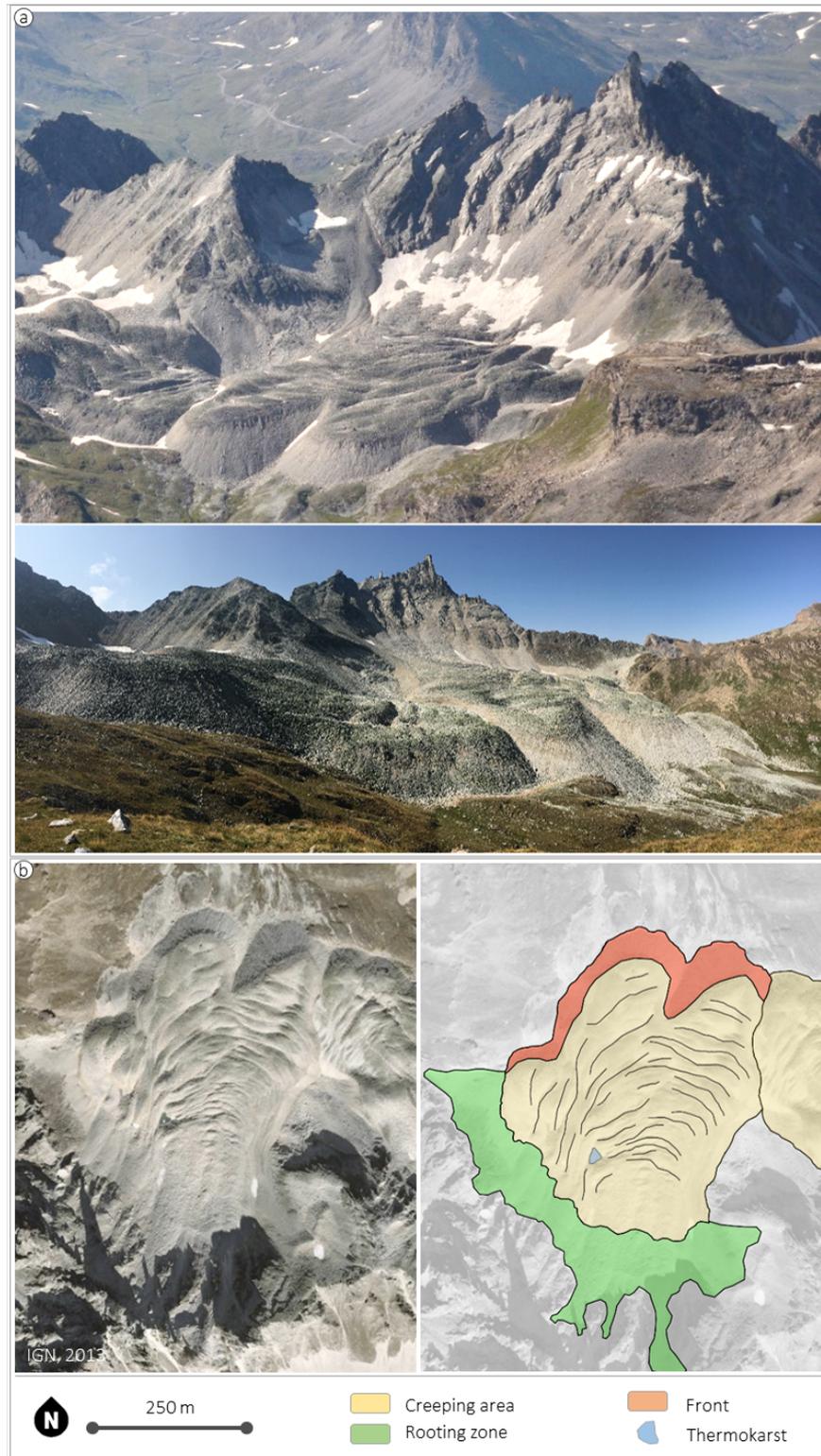


FIGURE 2.4: Overview of an active rock glacier, located in the [Grande Sassiere](#) natural reserve. On top (a) it is shown the appearance of the landform from the field, while on bottom (b) it is shown the appearance on orthoimage and general morphological characteristics.

can be found in fine grained areas. If a rock glacier contains permafrost but does not

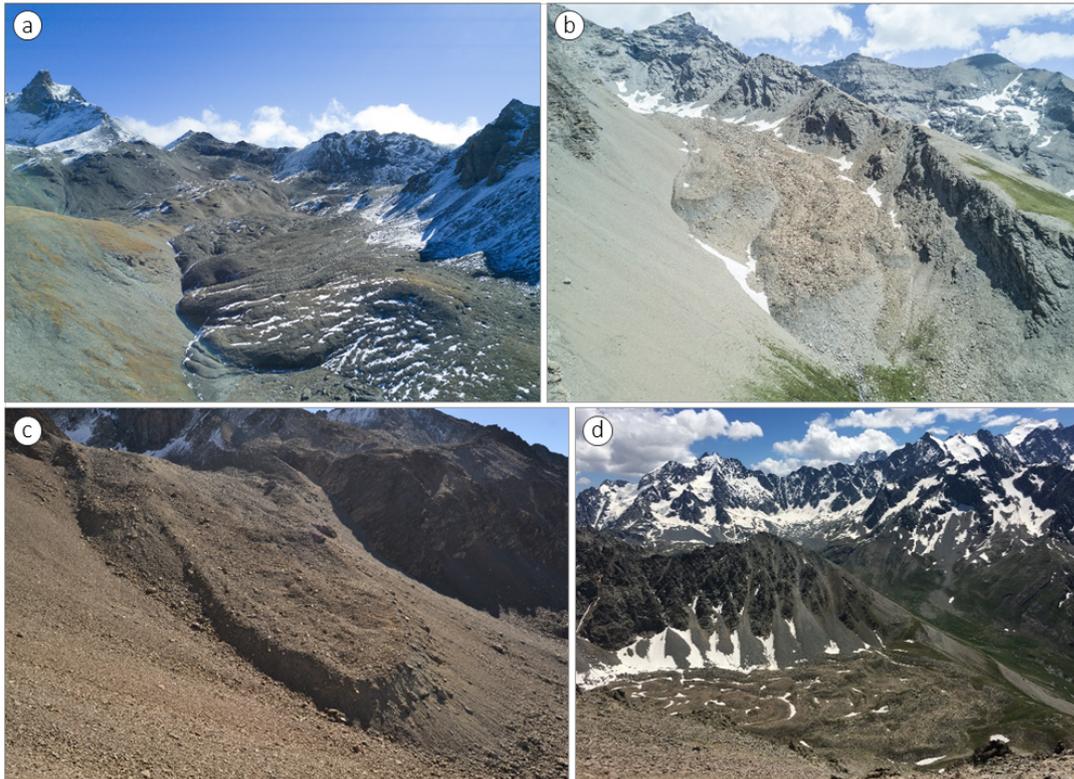


FIGURE 2.5: Different rock glaciers morphologies. **Pebbly** (a) and **bouldery** (b) rock glaciers. **Inactive** (c) and **relict** (d) rock glaciers.

creep, then it is classified as inactive. Inactive rock glaciers present a slightly convex-flat surface and the front is less steep ($30 - 35^\circ$; Figure 2.5c). The surface boulders are more stable and lichens and pioneer plants can be found. Activity is due to the ratio between ice content and slope angle, which if is below a certain threshold characteristic to each rock glacier, may inhibit creep causing inactivation. Inactivation may occur when valley bottom is reached (topographic inactivation) or when atmospheric temperature rises causing ice thaw (climatic inactivation). The extreme end of climatic inactivation is found in rock glacier that lost almost all permafrost content, i.e. relict rock glaciers (Figure 2.5d). These landforms are characterized by concave surface marked by cryokarstic depressions and flat front ($20 - 30^\circ$). Vegetation is widespread and may be in an advanced stage, e.g. trees.

Creep Rock glacier movement is governed by creep, i.e. internal deformation of the ice-rich matrix, that occurs in distinct deep shear horizons where most of the deformation takes place (Arenson et al., 2002). Depending on the activity rate, surface movements are of the order of few centimetres to several meters per year. Although Glen's law (Glen, 1955) can be used to locally describe the creep behaviour, rock glacier rheology is extremely complex, being influenced by several factors as ice content, debris granulometry and permafrost water content and temperature (Arenson and Springman, 2005). Since rock glaciers internal structure is strongly anisotropic (Arenson et al., 2002), strain rates are inhomogeneous and cause a strong spatial

variability of creep rates within the landform surface (e.g. Bodin et al., 2018, Figure 2.6). Although rock glacier creep presents high local variability within a single landform, some regional tendencies could be observed as landforms in colder environment tend to creep at lower rates than those in warmer settings (Kääb et al., 2007).

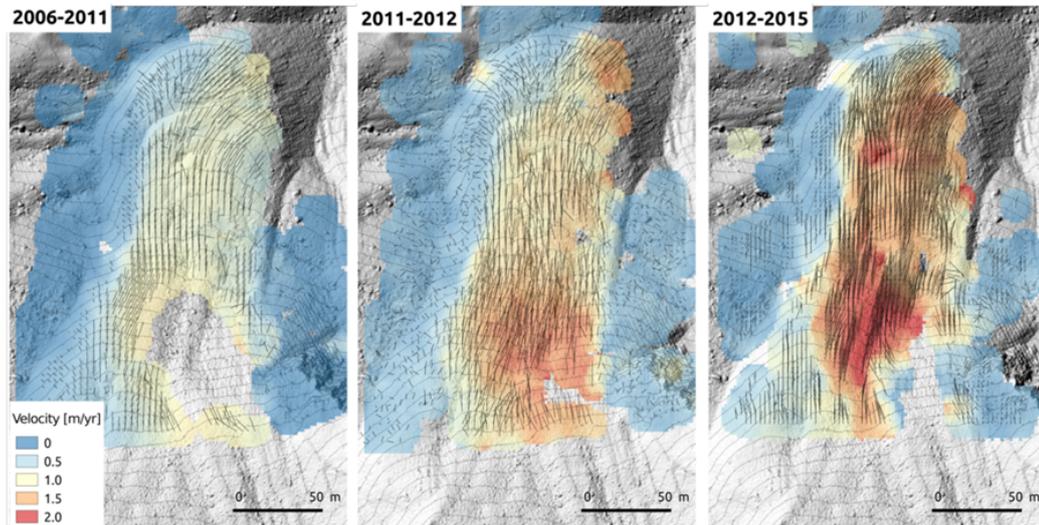


FIGURE 2.6: Spatial and temporal variability of permafrost creep on the Laurichard rock glacier of the Laurichard obtained with high resolution DEMs comparison (Bodin et al., 2018).

Rock glacier creep has not only a marked spatial variability but also a temporal one. Due to the influences of temperature and liquid water, creep rates vary according to seasonality and to interannual climatic trends (Figure 2.7). Some rock glaciers may not be responsive to seasonal variability, possibly when the shear horizon is too deep to be influenced by short term climatic variability (Arenson et al., 2002). When the rock glacier is responsive to seasonality, snow cover thickness, onset and disappearance date determine the timing and magnitude of the response. These landforms are slower in winter and accelerate as melting season starts, peaking in early autumn, with an annual variability up to 80% (Krainer and Wolfram, 2006; Perruchoud and Delaloye, 2007; Delaloye et al., 2010).

Decennial variability is caused by climatic trends. In particular, air temperature variations seem to influence the creeping rates with several months up to a year delay, suggesting that this behaviour is due to the long term warming or cooling of the permafrost body (Bodin et al., 2009). On the contrary to seasonal variability, interannual variability cause synchronous creeping rate variations at the regional scale (Delaloye et al., 2008b). In the European Alps, rock glaciers experienced a sharp increase in creeping rate since the end of the 90s, peaking in 2004. After that velocities decreased until 2006-2008, and acceleration is occurring since then (Kellerer-Pirklbauer et al., 2018).

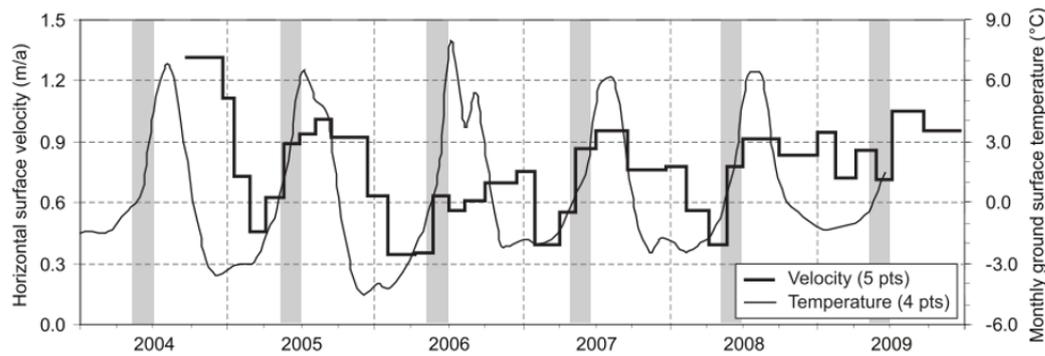


FIGURE 2.7: Temporal variability of the surface velocity at the seasonal scale in relation with ground temperature and snow melt periods (grey shaded areas), measured by Delaloye et al. (2010) at Becs-de-Bosson/Réchy rock glacier.

Occurrence Rock glaciers are common landforms in mountain regions where permafrost is widespread. In the European Alps, rock glaciers count several thousand individuals spread in all ranges (Cremonese et al., 2011). In arid regions where glaciers are rare, rock glaciers are the dominant cryospheric landform and become relevant in water storage and supply (Bolch and Marchenko, 2006; Garcia et al., 2017). These landforms are most likely present also in other cold planets as Mars (Whalley, 2003).

The strongest limit to rock glacier development are white glaciers. Due to their erosive power, white glaciers dominate over rock glaciers and they represent a physical limit to the development of periglacial landforms. The critical feature controlling the occurrence of glaciation versus periglaciation is the topographical suitability to accumulate snow. If the glacial equilibrium line is located in an area where snow cannot (can) accumulate, then a rock glacier (white glacier) is more likely to occur (Humlum, 1998). In this sense, rock glaciers are satellite processes of the glaciation and tend to occupy the recently deglaciated areas (Cossart et al., 2010).

In some cases, despite the widespread glaciation, rock glaciers may still exist at the terminus of the glaciers. This can happen if the glacier ablation zone or frontal moraine is located in permafrost conditions and there is enough debris supply to form a continuous active layer (Monnier and Kinnard, 2015). In these cases there is the formation of a continuous system from white glacier to debris-covered glacier to rock glacier. The system is dynamic and it may shift its landform proportions accordingly to climatic variability (Monnier and Kinnard, 2017). These kinds of systems are more common in the highest mountain ranges of the planet as the Andes and Himalaya.

Apart from glaciation, lithology is a critical factor controlling the existence of these periglacial landforms. High erosion rates are necessary to produce enough debris to stock sufficient ice-rich permafrost and initiate creep (Kenner et al., 2017). Headwall jointing controls the debris size which influences the thermal offset due to the Balch effect, allowing for example blocky rock glaciers to exist at lower elevations

(Janke and Frauenfelder, 2007). These processes cause strong discontinuities in rock glacier distribution at the regional scale (Monnier, 2006).

2.2 Climate change

Climate change in the French Alps defines the frame of the POIA-PERMARISK project. As introduced in the previous chapter, permafrost characteristics are bound to the climate and they may be significantly affected by temperature variations. Nevertheless, climate is a dynamic system and it has always been changing. So, why should we be concerned by the permafrost fate if it most likely already faced climate change? To answer this question, this chapter aims to delineate the characteristics of the climatic variability since the end of the last glaciation to present days, describing also future scenarios. This analysis, based on revision of palaeoclimatology science results, will allow the reader to understand the novel features of the present climate change and, in the next section, the associated effects on permafrost.

2.2.1 Post-glacial climate

Since rock glaciers developed following the last deglaciation, which started in the French Alps about 20 000 y BP, climate after Last Glacial Maximum (LGM) is the subject of interest (Cossart et al., 2010). This period consists in late Pleistocene and Holocene epochs and covers about 15 000 years. On the contrary of glaciations that have multi-millennia scale and are due to extra planetary forcing (Milankovitch orbital cycles), the abrupt climatic fluctuations that occurred in this period have centennial scale and are related to on-earth conditions. Since explaining the mechanisms at the core of these climatic changes exiles from the goals of the present study, this section will only describe magnitude and timing of temperature variations.

The late Pleistocene was characterized by strong climatic instability (Taylor et al., 1993). The LGM, characterized by temperatures about 10°C lower than the current period, was ended by a general warming trend, i.e. the Bølling-Allerød warming (Hoek, 2009). This warming trend was occasionally interrupted by sudden negative anomalies, called stadials. The so-called Greenland Stadial 1 (GS-1, also known as *Younger Dryas*) that took place 12.9 - 11.7 ky BP and lasted about 1200 years (Schwander et al., 2000) is relevant in the context of the present study. The GS-1 involved a temperature negative anomaly of 10- 15°C at the poles and 2.5 – 3°C in the Alps (Carlson, 2013). Greenland ice sheet cores suggest that the end of the Younger Dryas was characterised by an abrupt warming estimated to be of 5 – 10°C and that took place within 30 years (Alley, 2000). Similar signals are observed in the rest of Europe, causing Alpine glaciers to shrink to a size comparable to present days (Ivy-Ochs et al., 2009).

The abrupt end of the Younger Dryas marked the beginning of the current epoch, the Holocene. Holocene climate was characterized by a smaller climatic variability

than the late Pleistocene, as observable temperature anomalies rarely exceeded 2°C. Despite several cold events registered in the Northern Hemisphere during Preboreal at 8.2 ky BP, temperature steadily rose and stabilized between 8 to 6 ky BP. This warm period is often referred to as the Holocene Climatic Optimum (HCO, Wanner et al., 2011). At mid-latitudes mean annual temperatures were comparable to the current record, while warmer climate occurred in the Arctic.

The HCO marked the warmest point in the interglacial stage and was ended by a global cooling trend that lasted until the 20th century called the Neoglacial. Driven by the Milankovitch cycle that progressively reduced summer insolation in the Northern hemisphere, the Neoglacial is interpreted as the entering stage of a new glaciation (Porter and Denton, 1967). The Neoglacial was marked by 5 distinct cold events that occurred at the millennial scale and were interrupted by warmer periods (Mayewski et al., 2004). Timing and magnitude of these phases were different according to the region, suggesting that these oscillations were not global but local. In the context of the present work are noticeable the Goeschenen I and Goeschenen II cold anomalies and relative Iron Age (or Roman) and Medieval warm anomalies (Le Roy et al., 2017).

Through the neoglacial many alpine glaciers reached their Holocene maximum during the coldest and most recent of these events, the so-called *Little Ice Age* (LIA, Matthes, 1939). Although the temporal span of the LIA varies according to local conditions, it is generally agreed that in the French Alps it took place between the early 14th and mid-19th century (Nussbaumer and Zumbühl, 2011). The last period of the LIA is also referred to as the pre-industrial period and coincides with the firsts systematic weather measurements, marking the chronological frontier between palaeoclimatology estimations and historical climatology.

2.2.2 20th century climate change

The end of the LIA was marked by the last Holocene glacial maximum, 1850 circa, and consequent retreat of most European glaciers (Lüthi, 2014). It is noticeable that mean air temperature slightly decreased until about 1920, while glacial retreat was triggered earlier by a decrease in winter precipitation (Steiner et al., 2008; Sigl et al., 2018, Figure 2.8). Overall, between 1880 and 2012 the global temperature rose $0.85 \pm 0.2^\circ\text{C}$ with an increasing warming rate that passed from 0.07 to 0.13°C per decade from the periods 1880-1950 to 1950-2012 (IPCC, 2013). Although the magnitude of this temperature fluctuation is not unprecedented in the Holocene as current temperatures are comparable to the HCO, the climate change rate is alien to this epoch as temperatures shifted from the Holocene minima to the Holocene maxima within 130 years while previous anomalies used to take several centuries to onset.

Locally, climate change characteristics are very variable. The Arctic regions experienced the most dramatic changes, as the surface mean air temperature was up to 3°C higher in the period 2005 – 2009 than in the period 1951 – 2005. In the

European Alps temperature increase is also estimated to be above the global average. Mean temperatures increased by 1.5 (southern Alps) to 2.1°C (northern Alps) since 1950 with a rate of 0.4°C/decade (Einhorn et al., 2015). This rate increased to 0.5°C/decade since 1980. Temperature increase characterized mainly minimum temperature, while more modest increase of maximal temperatures was observed, with the exception of summer 2003. Temperature increase is not uniform across altitudes, as mid-altitudes (1500 – 2000 m.a.s.l.) seem to have warmed at higher rate than high – altitudes (>4000 m.a.s.l., Beniston, 2006).

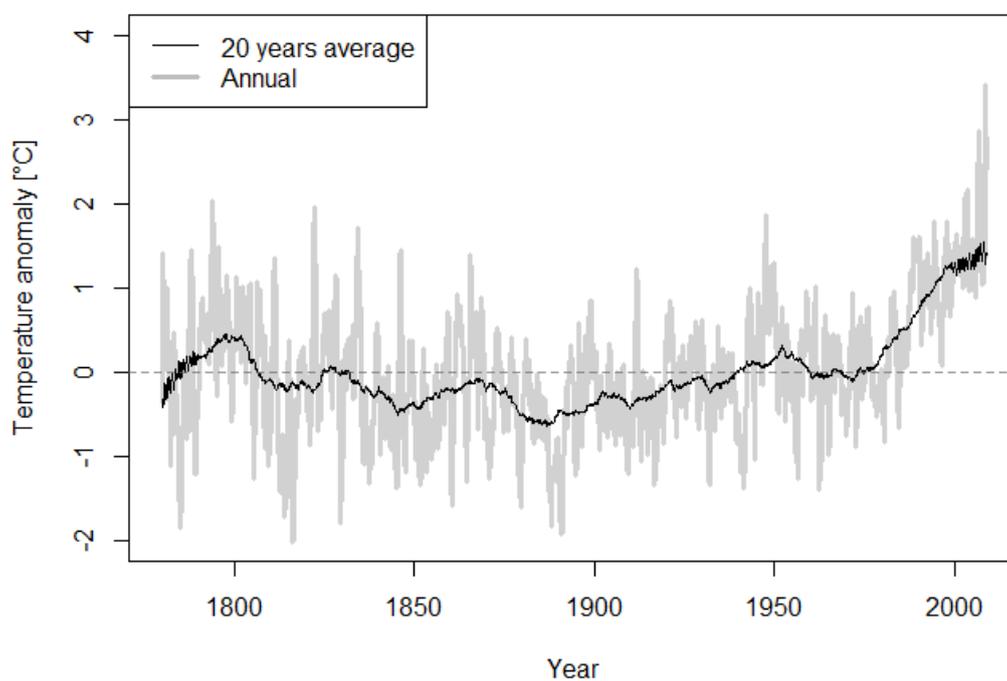


FIGURE 2.8: Mean annual temperature anomaly in the northern French Alps with respect to the period 1850 - 2006, obtained using Histalp data (Auer et al., 2007).

2.2.3 21st century projections

Climate change has a main role in ecology and civilization. Abrupt changes and even milder changes are believed to have caused the decline of several human civilizations (Mayewski et al., 2004). Therefore, as we are living in a context of climate change unprecedented in the Holocene, there is a strong interest in forecasting future trends and considerable efforts are being made in this sense. Accurate climate change forecasting involves understanding the processes at the core of the change itself. After decades of research, the IPCC declared that the 20th century climate change is *very likely* due to anthropogenic greenhouse emissions (Myhre et al., 2013). Consequently, climatic models are based on the hypothesis that anthropic

greenhouse gases emissions are a significant forcing in climate. However, future anthropic emissions are to be forecasted as well, a not straight-forward task given the battle between environmental protection, industrial interests and developing countries. Moreover, several side processes may become more relevant in an evolving climate as, for example, the influence of the Atlantic Meridional Overturning Circulation (AMOC, Chen and Tung, 2018). As a result, climatic projections have large uncertainties and vary greatly from model to model.

Despite uncertainties, global mean temperature is *likely* to exceed 2°C by the end of the 21st century with respect to pre-industrial values (Collins et al., 2013). It is *unlikely* that this increase will exceed 4°C. This suggest the warming trend not only is not expected to be reversed in the near future but also warming rates may keep on increasing. Warming could affect mainly minimal temperatures which will rise in average. Extreme events, as heavy rainfalls and heat waves, are expected to increase in frequency (Kirtman et al., 2013). Temperature warming is very likely to cause important arctic permafrost degradation, mass loss of Greenland ice-sheet and Arctic Ice shelf within a millennium, triggering positive albedo feedbacks and significant variations in the oceanic circulation, processes that may significantly affect the climatic evolution at the global scale.

In the European Alps, despite the large uncertainty due to complex interactions between global climate and local scale phenomena, models unanimously suggest persistency in the warming trend (Heinrich et al., 2013). Models suggest a warming of 1.2 to 1.6°C by 2050 with respect to the period 1961 – 1990. By the end of the century temperature may rise by 2.7 to 3.8°C (Gobiet et al., 2014). Warming is expected to be more pronounced in late summer and winter and amplified at higher altitudes (Kotlarski et al., 2012). Extreme rainfall events, which present a great interest in the context of natural hazards, cannot be represented by numerical modelling due to their reduced spatio-temporal occurrence. Nevertheless, statistical methods suggest that return time of extreme events is likely to decrease, suggesting that heavy exceptional rainfall may become more common in the near-future (Rajczak et al., 2013).

2.3 Mountain permafrost and climate change

Although nowadays temperatures are comparable to warm stages of the Holocene, the 20th century is characterized by the most abrupt climatic variation since the Pre-boreal. The cryosphere is therefore enduring a significant warming in an unusually small amount of time. Permafrost can have a significant thermal inertia and ice can persist underground for long time despite significant atmospheric warming. Nevertheless, frozen grounds may enter in a state of strong disequilibrium with the climate where change of state may cause abrupt phenomena. Partial or total change of state from frozen to liquid water causes in general a loss of the ground stiffness, triggering a series of processes that depend upon the type of permafrost involved and the topographical context.

Bedrock permafrost

Permafrost degradation in rockwalls can severely compromise the geotechnical properties of the material. Laboratory tests suggest that the factor of safety of ice-filled joints is strongly dependent on temperature (Davies et al., 2001) and warming can be responsible for triggering failures from minor localized rockfalls to slope scale rockslides (Krautblatter et al., 2013, Figure 2.9). Water infiltration has also an important role as it can transfer heat more efficiently than conduction, potentially enhancing warming effects on larger portions of the headwalls (Gruber, 2007).

All these processes, firstly theorized and then observed in laboratory experiments, seem to fit the observations in permafrost mountain regions. Warm interstadials through the late glacial were found to correlate with increasing rock falls occurrence in permafrost areas (Gallach et al., 2018). Several studies pointed out that frequency of rockfalls is increasing and peaks in warm periods and extreme precipitation events (Ravanel and Deline, 2010; Huggel et al., 2012; Ravanel et al., 2017). Due to the expected increase of heat waves and heavy rainstorms in the future, these events may become more common (Huggel et al., 2010).

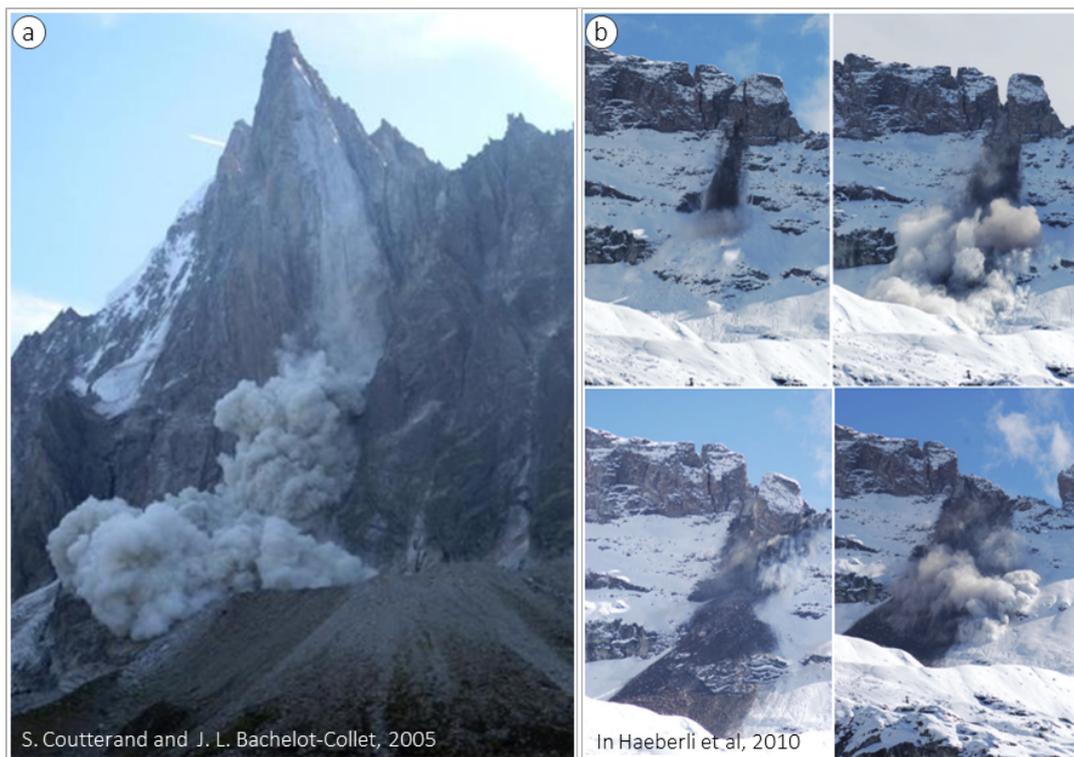


FIGURE 2.9: Rock fall at the Dru (a), and the collapse of a pillar from the Gross Charpf (b), from Haeberli et al. (2010)

Debris permafrost

Permafrost degradation in debris slopes can compromise the ground characteristics. Ice behaves as a cement in loose debris slopes and its warming decreases ground

stiffness (Harris et al., 2003). Due to the loss of the bonding effect of the warming or thawing ice matrix, debris permafrost zones may be subjected major morphological adjustments (Nater et al., 2008). Thawing areas may be more prone to water infiltration which enhances heat transfer and reduces further ground strength. Also, degrading permafrost slopes may be more subjected to erosion as supra permafrost flows can result in active layer failures (Kummert et al., 2017). The magnitude of these processes depends on ground properties and ice content.

Thermokarsts When the topography is flat and the debris have a high ice content, thermokarstic depressions may be generated (Figure 2.10). These phenomena are mostly observable in flat landforms containing massive ice as glaciogenic rock glaciers. Since massive ice is impermeable, thermokarstic depressions (or simply thermokarsts) may become a topographic trap for melt water, meteoric water or runoff water depending on the conditions creating a lake.



FIGURE 2.10: Detail of the thermokarstic lake of the Marinets rock glacier (a) and formation process of a Thermokarstic depression in the Tignes rock glacier (b).

Thermokarstic lakes may be stable or developing. Stable thermokarsts consist of periglacial topographic traps that host small lakes hardly varying in dimensions on the decennial scale. Developing thermokarsts have crater-like shape and can be observed to rapidly increase in size (Kääb and Haeberli, 2001, Figure 2.10b). The water in these depressions, jointed possibly with massive ice on the sides, may enhance permafrost thaw, triggering a positive feedback in the thermokarst growth. This typology is particularly unstable and stored water volume is constantly increasing at the expenses of the ice rich ground damming it.

Rock glaciers The proof of the rock glacier sensitivity to climate change is given by the existence of relict landforms. Although increasing temperatures tend to increase active rock glaciers displacement rates, warming can cause critical ice loss to landform inactivation (Scapozza et al., 2010). Since relict rock glaciers are generally

found at lower elevations than intact rock glaciers, they are most likely witnesses of colder stages of the late Pleistocene. In particular, the typical elevation of relict rock glaciers in the European Alps, i.e. 400 – 600 m.a.s.l. lower than nowadays active rock glaciers, is compatible with late Pleistocene cold stadials as the GS-1 or Younger Dryas (Kerschner, 1978; Haeberli, 1985; Frauenfelder et al., 2001; Dramis et al., 2003; Oliva et al., 2018). Inactive rock glaciers on the other hand are believed to have inactivated during warm stages of the Holocene as the medieval warm period (Scapozza et al., 2010).

Considering active rock glaciers, mean annual temperature modulates to a large extent the landform displacement rates. The creep rate peak of 2004 was the consequence of a decade scale warming trend peaking with the 2003 heat wave (Delaloye et al., 2008b). The high creep rates registered in this decade are the results of the continuously increasing temperatures and heat wave episodes (Kellerer-Pirklbauer et al., 2018). This process, joined with a constant temperature warming probably long lasting in the future, rising questions on the fate of these landforms.

Warming temperatures may increase the predisposition of the rock glacier to the so-called *destabilization* process (Roer et al., 2008). Rock glacier destabilization has a central role in this study, being the object of Chapters 4 and B. Destabilization was described in several ways by different authors, revealing the complexity of the phenomena (Delaloye et al., 2013; Roer et al., 2008; Avian et al., 2005; Bodin et al., 2016; Bodin et al., 2012; Scotti et al., 2016; Ikeda et al., 2008; Eriksen et al., 2018). Destabilized rock glaciers are recognizable because they develop atypical geomorphological features that are linked to strong traction within the landform and marked creep rate increase (Figure 2.11).

Based on the studies previously cited, these features can be grouped in crevasses, scarps and cracks, here called *surface disturbances* (Figure 2.12). Crevasses are deep incisions in the rock glacier surface, cutting the permafrost body. They are normally isolated and can reach several meters depth. Cracks are shallower, possibly affecting only the active layer. Scarps are steep shear planes that dislocate the rock glacier in two different bodies that are completely disconnected. These features are often observable to be stable for decades, while their sudden increase in number and dimension is synchronous to the strong acceleration of the landform and destabilization onset (Delaloye et al., 2013; Eriksen et al., 2018).

The creeping pattern is also peculiar: the rock glacier portion upstream the surface disturbances may creep relatively slowly, while the downstream part moves remarkably faster (Figure 2.13). This difference in velocity within the same landform is believed to cause the traction creating the surface disturbances. This highlights a disconnection of the lower part of the rock glacier, which is driven by a different dynamical process than the upper part. Although still unclear, it was suggested that destabilized rock glaciers show a transition between creep to basal sliding, a process that involves higher velocities and the de-coupling from the climatic forcing (Roer et al., 2008; Schoeneich et al., 2015).

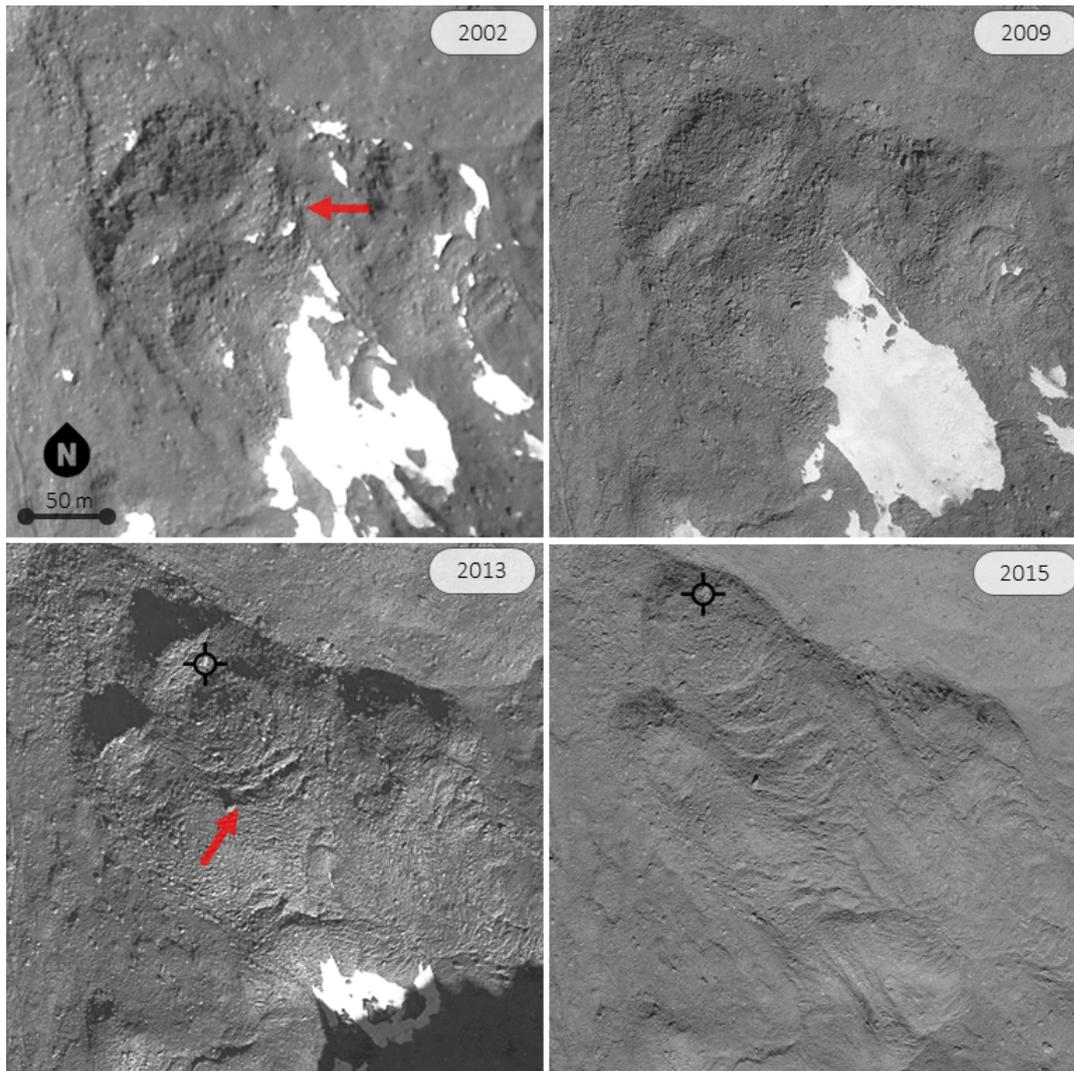


FIGURE 2.11: Example of **rock glacier** destabilization from the southern French Alps. A slightly pronounced scarp and cracks were observable since 2002 (red arrow) and slowly evolving until 2009. In 2013 the morphology two large crevasses were observable (red arrow). At this point the destabilization started, causing a displacement, measured on the black target on these orthoimages, of about 40 meters in two years

Basal sliding is suspected to be caused by the formation of a water saturated layer that reduces shear resistance (Kenner et al., 2017). Water infiltration is enhanced by the surface disturbances (Figure 2.14) causing a positive feedback process in the destabilization, observable by the simultaneous increasing of displacement rates and surface disturbances dimensions (Ikeda et al., 2008).

Destabilization occurrence depends upon several factors. Local topography and rock glacier structure seem to determine the predisposition to the site to destabilization, as destabilized rock glacier tend to be found on steep and convex slopes (Delaloye et al., 2013). Permafrost degradation is suspected to contribute to the destabilization occurrence by increasing rock glacier velocity and water content (Eriksen

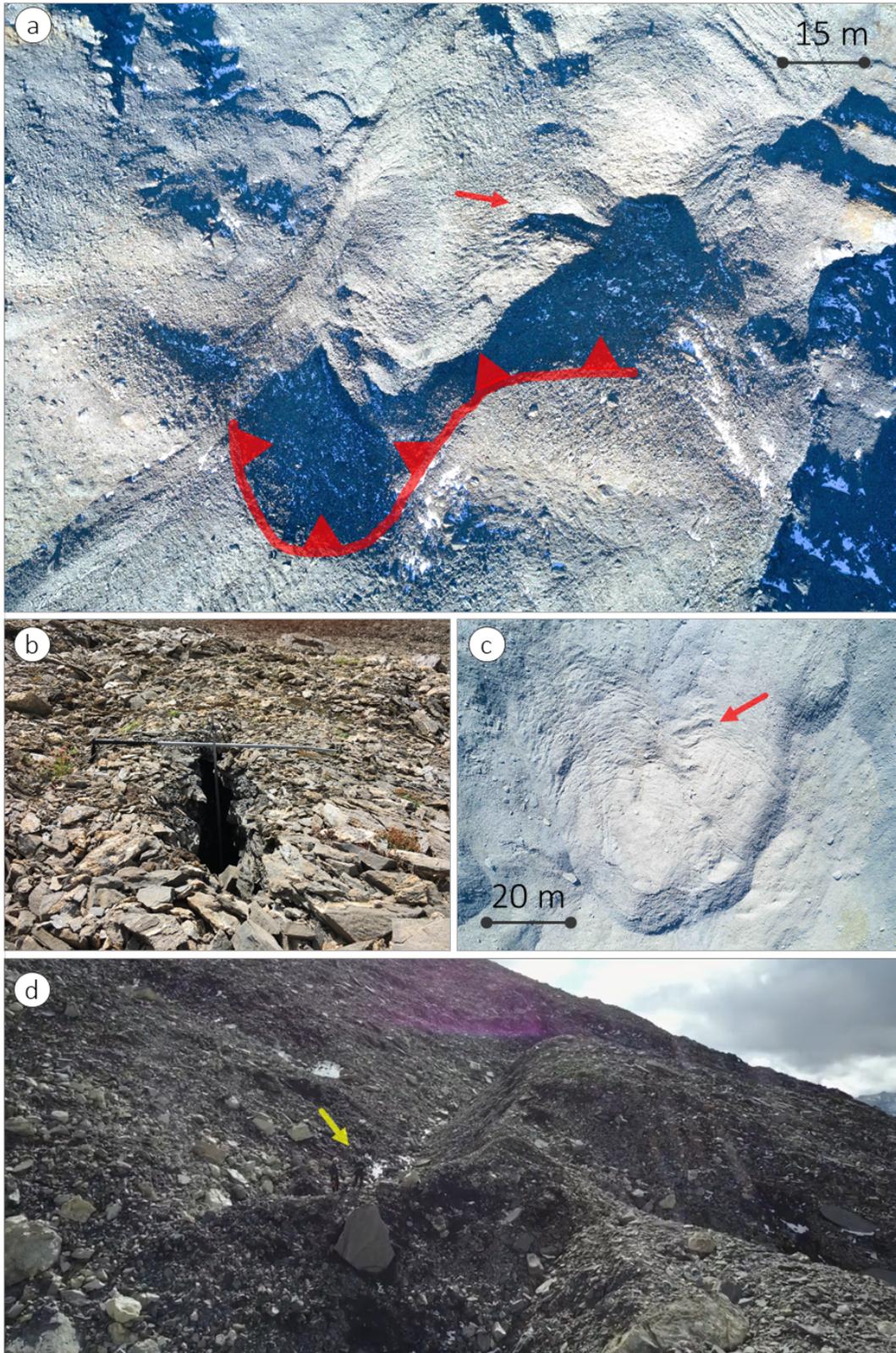


FIGURE 2.12: Some pictures of surface disturbances taken on the field. In (a) a UAV image of a scarp, about 40 m high. Red arrow indicates a small crevasse. In (b) a crack, with hiking sticks as scale. In (c) UAV image of several cracks on a pebbly rock glacier with a more pronounced feature resembling to a shallow crevasse (red arrow). In (d) a deep crevasse with people for scale (yellow arrow).

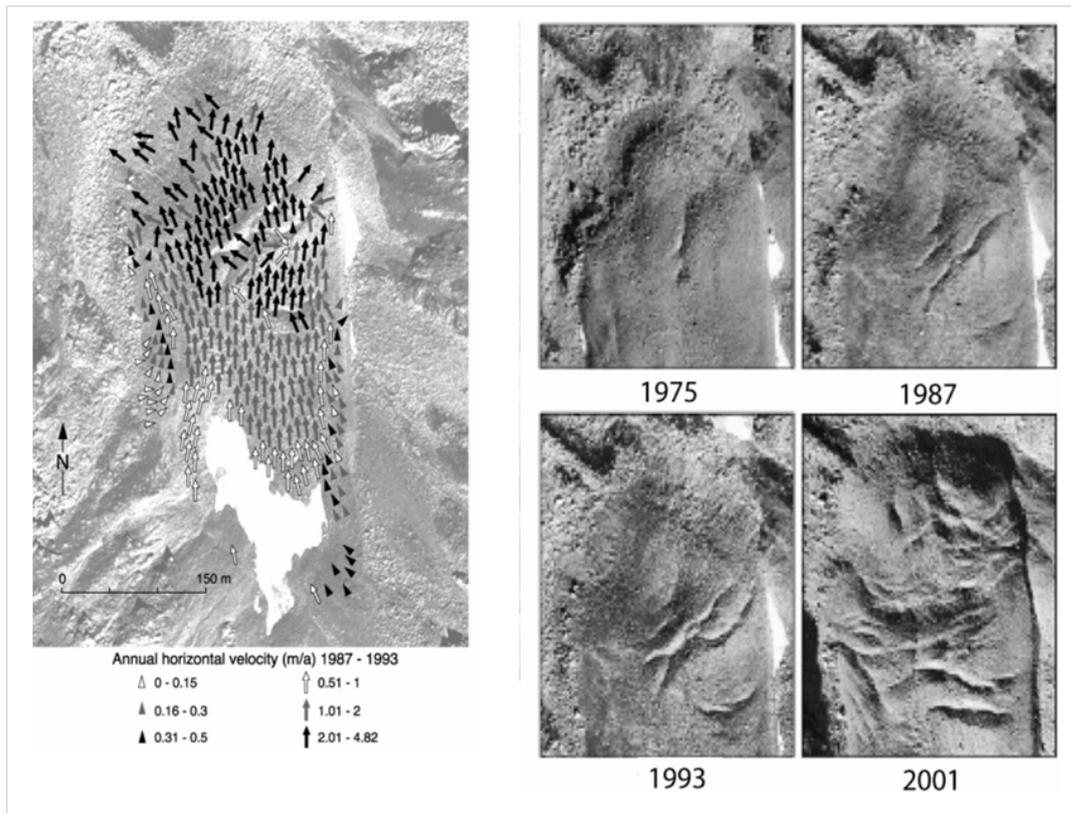


FIGURE 2.13: Creeping pattern of a destabilized rock glacier and evolution of surface disturbances, presented in Roer et al. (2008).

et al., 2018). Destabilization itself can be triggered when a certain threshold of velocity or water content is reached, causing by abnormal acceleration regardless the climatic forcing. Several authors proved that destabilization can be triggered by a mechanical action as a severe load on rock glacier surface caused by glacial advancement or massive rock falls (Roer et al., 2008; Delaloye et al., 2013; Scotti et al., 2016). Delaloye et al. (2013) suggested that this load caused a compressive wave that propagate through the rock glacier in several decades and caused its destabilisation when reaching weak spots of the landform as convex areas.

After a period of increasing creep rate, destabilized rock glaciers tend to slow down and stabilize. In this sense, the destabilization process can be interpreted as a *crisis* phase and may repeat itself periodically with several decades of interval (Delaloye et al., 2013). Nevertheless, when the destabilization is exceptionally severe, the landform may collapse on a sudden landslide. In this catastrophic event, an entire portion of the rock glacier detaches and slides downslope, involving volumes of several hundred thousand cubic meters. Although extremely rare, as to date only two cases, one in the Andes and one in the French Alps have been reported (Bodin et al., 2012; Bodin et al., 2016, Figure 2.15), this kind of event creates concerns in the context of permafrost related hazards. In the French Alps, it was the Bérard rock glacier, an active landform made of schists located in the Parpallon range, that suddenly collapsed in summer 2006 after a rainy event of mild intensity. The collapse



FIGURE 2.14: Crevass filled with old and recent snow and water.
Photo taken in October 2018.

caused a landslide of 250 000 m³. Historical orthoimagery analysis revealed that the landform presented a crevasse cutting transversally the permafrost body since the 60's. A posteriori analysis suggested that a combination of warm summers, high meltwater availability and natural predisposition to failure due to topography and lithology, may have compromised the stability of the landform (Bodin et al., 2016).

Debris flows Since mountain permafrost has a stabilizing role on debris slopes, climate change - related processes may cause a higher supply of debris in torrential channel, potentially increasing the occurrence of debris flows (Haeberli et al., 2010). This process is mostly observed at rock glaciers having a steep front directly connected to steep stream channels (Springman et al., 2013; Kummert and Delaloye, 2018). Rock glacier acceleration may increase the volume contribution of debris to the sediment chain (Kummert and Delaloye, 2018). Also, frontal layer failures can be the initiating cause of debris flows of great magnitude by releasing at once large volumes of recently unconsolidated materials (Lugon and Stoffel, 2010).

Kummert et al. (2017) described in detail the processes that trigger slope failures at the front of fast moving rock glaciers. Failures are more likely in spring and summer, generally triggered by meltwater infiltration and/or intense rainfalls. Among several failures typologies, the *concentrated flows* are remarkable. This phenomenon consists of deep failures caused by supra and intra permafrost water flows

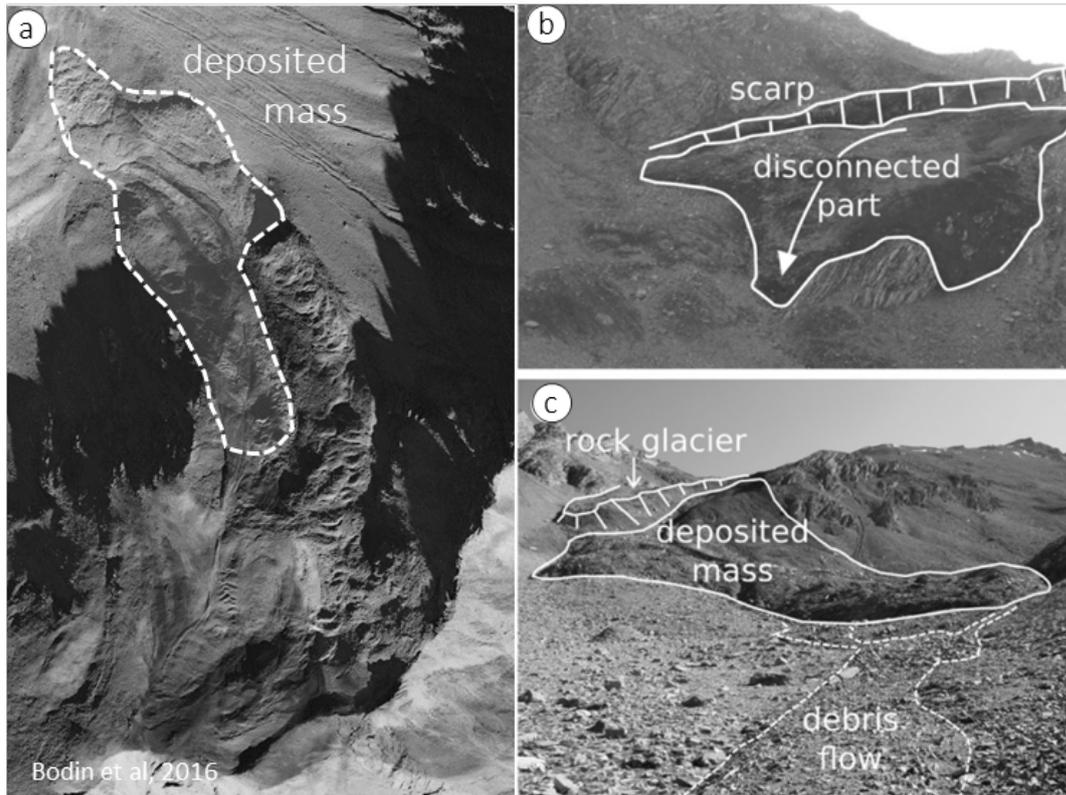


FIGURE 2.15: Bérard rock glacier collapse, from Bodin et al., 2016

that can have a significantly intense discharge. Since concentrated flows occur only in specific conditions of ground saturation not frequently met, they are more rare to observe than others sediment transfer processes. Nevertheless, this process may involve the whole active layer and mobilize consistent volumes triggering consistent debris flows, as analysed in Chapter 5.

2.3.1 Emerging hazards related to permafrost degradation

Several mountain ranges, and the European Alps in particular, are subjected to a heavy and extensive anthropization that extends from valley bottoms (settlements, communication lines, reservoirs) to mountains tops (cable cars, ski slopes, mountaineering activities). This coexistence started centuries ago and alpine communities learned through time to deal with the hazards related to the proximity to this challenging environment. Permafrost degradation-related processes, combined to specific topographic settings, may add to these hazards by compromising high mountains infrastructure stability and preconditioning mass movements (Haeberli et al., 1997; Harris, 2005; Käab et al., 2005; Haeberli et al., 2010; Bodin et al., 2015).

Infrastructure stability

Geotechnical alterations due to permafrost thaw and warming can cause a threat to infrastructures installed on frozen grounds. Infrastructures foundations are designed according to specific geotechnical parameters that may vary in the context of degrading permafrost due to increase in water content and thaw. Alterations can be enhanced by continuous loading and heat transfer to the ground through foundations. Furthermore, the frozen ground is more prone to lose its stiffness due to water infiltration and permafrost thaw, increasing the vulnerability of infrastructures (Fabre et al., 2015).

In the European Alps, the infrastructures concerned are mostly high mountains installations as cable cars and huts. Only in the French Alps, almost 1800 infrastructure elements are built in permafrost areas (Duvillard et al., 2015). Typical damages consist in shifting, tilting, and fracturing of the infrastructure because of ground loss of stiffness, subsidence or unexpected creep (Fabre et al., 2015). Damages are typically slowly developing, allowing to take action before failure. As a result, so far, no sudden unexpected collapse of infrastructures was observed in the region.

Recuperation actions often concerns an adaptation of the foundation to the developing ground consolidation. In some extreme cases, the whole infrastructure is moved to a more stable location. Therefore, these actions may have a significant economic impact on the companies exploiting mountain environments and there is a rising interest in preventing infrastructure instability by performing permafrost-oriented foundations assessments.

Mass movements hazards

Mass movements are a natural hazard very common in alpine environments and involve avalanches, rockfalls, debris flows, glacial collapses and slides. These hazards have been always threatening the alpine community and risk management is performed by two strategies mainly. At first, hazardous areas are determined based on historical data and modelling. In these areas urbanization is limited, as well as transit in specific periods. Second, protection infrastructures, as dams or emergency reservoirs, are built to physically prevent or contain extreme mass movements.

Protection infrastructures are designed on the basis of predicted extreme events in order to be the best compromise between risk avoidance and cost. The magnitude of these extreme events is extrapolated on the statistically expected return time (Mays, 2005). The underlying assumption behind extreme events statistics is that the available data are representative on the long term and that the processes do not vary in time, i.e. a stationary hypothesis. As a result, protection infrastructures are often designed on data representative for the mid-20th century (Level, 2013). Nevertheless, climate change itself is expected to involve more extreme meteorological events as intense rainstorms and heat waves are becoming more frequent (Kirtman et al., 2013). In this context the stationary hypothesis is no longer valid, and protection

infrastructures may not be adapted to the frequency and magnitude of incoming extreme events.

Permafrost degradation related mass movements may involve hazards that were rarely considered in the design process. Increasing erosion rates and debris supply are processes that may add extra volumes to extreme events and, since these phenomena are developing in the past decades, they violate the stationary hypothesis. In some cases, permafrost degradation may be the source of unprecedented mass-movements by causing a failure source by varying the equilibrium of the sediment chain. In these cases, protection infrastructures may be either inadequate or absent, involving a significant risk. The increase of permafrost degradation phenomena therefore is suspected to put in a state of obsolescence the protection strategy of the alpine regions.

2.4 The Project POIA-PERMARISK

The project POIA-PERMARISK takes place within this context of emerging risks linked to permafrost degradation in the French Alps. Emerging risks started to gain interest in the French scientific community since 2006, the year characterised by the Dru rockfall and the Berard rock glacier collapse (Bodin et al., 2016, Figure 2.15). These events triggered complementary efforts by universities, local authorities and private companies to better understand processes linked to bedrock and debris permafrost degradation in the region. The POIA-PERMARISK project aims to assess processes related to debris permafrost degradation in order to identify potential hazards at the scale of the French Alps.

2.4.1 Structure

The POIA – PERMARISK project is oriented towards the hazard identification, i.e. to understand in which situations mass movements may be triggered or infrastructure stability may be compromised. The main aim of the project is to identify debris-permafrost related hazards and to produce a series of tools and analysis that may help to better characterize the risk areas. These tools involve potential hazard maps and sensitive sites identification. Strict collaborations with project partners allow to share knowledge between researchers and risk managers, producing tools oriented to perform risk assessment. Nevertheless, the project also aims to contribute to scientific knowledge concerning the processes involved by permafrost degradation.

In order to achieve these goals, the project involves several masters and PhD students, and researchers from CNRS, USMB and UGA. The broad topic of degrading permafrost in debris terrain is divided into three actions that aim to tackle the issues relative to one specific process. The three actions concern (i) rock glacier characterization, (ii) thermokarst identification and (iii) infrastructure stability on permafrost. Since the three actions cover themes that are not completely independent, sharing

between the parties working on the different topics is ensured by collaboration on mixed sites, i.e. where multiple processes are occurring.

Rock glaciers characterization This action aims to contribute to the knowledge concerning rock glaciers characteristics in the region, with the final goal of assessing hazard related to these landforms. The first step of this process, which constitutes the fulcrum of this PhD, consists in understanding the occurrence of rock glacier destabilization in the region. This issue, which will be described in detail in section 2.4.2, was also tackled by several master students of the IUGA who focused on destabilized rock glaciers characterization at the site scale (Ribeyre, 2016; Nielsen, 2018; Obregon, 2018) and at the regional scale (Serrano, 2017).

The second step will focus on the characterization of the rock glacier connectivity to the sedimentary chain in order to define hazard related to active landforms. This issue was engaged by a study at the local scale by Hovgaard and Eisenbrückner (2017), who evaluated debris flow scenarios related to an active rock glacier connected to a torrential system presenting downslope vulnerabilities. Further development will focus to a regional scale characterization, as described in detail in the conclusion of this PhD, section 6.2.1.

Thermokarsts Thermokarst have been sparsely observed in the region. Nevertheless, regional occurrence is unknown as no inventory of these formations has ever been compiled. The main goal of this action is therefore to map thermokarstic depressions, as well as other supra permafrost trapped waters. As second step, significant study sites are selected in order to better understand the evolution of the thermokarstic depressions in relation with climate change. At the present time, this action is being developed by Villard (2018), who mapped thermokarstic depressions in the region and defined priority study sites for monitoring.

Infrastructures on Permafrost This action aims to assess the stability of infrastructures built on permafrost in the French Alps. Most of these infrastructures involve cable car elements (pylons and arrival stations) and high mountain huts. The analysis is conducted at two scales. On the regional scale, using a GIS approach, infrastructures in a context of permafrost and potentially degrading permafrost are identified. At the local scale, infrastructures built on permafrost that experienced instability are investigated using geophysical and photogrammetric methods. At the present time, this action is being developed by the PhD work of Pierre-Allain Duvillard, at EDYTEM laboratory, in collaboration with the engineering company IMSRN.

2.4.2 The PhD Project

This PhD takes place in the context of the POIA-PERMARISK project by focusing on destabilized rock glaciers. Rock glaciers are a common landform in the French

Alps and destabilization is a phenomena known to occur in the region. The milestone case is the previously cited Bérard rock glacier, (Bodin et al., 2016) started to rise concerns on the potential risks related to the destabilization of these landforms. As first response, the RTM started mapping in 2009 the rock glaciers in the region by orthoimagery interpretation, completing the inventory in 2015 (Figure 2.16). The realization of the inventory was a key step in performing a comprehensive risk assessment of rock glacier related phenomena and a valuable database for investigating issues related to these landforms.

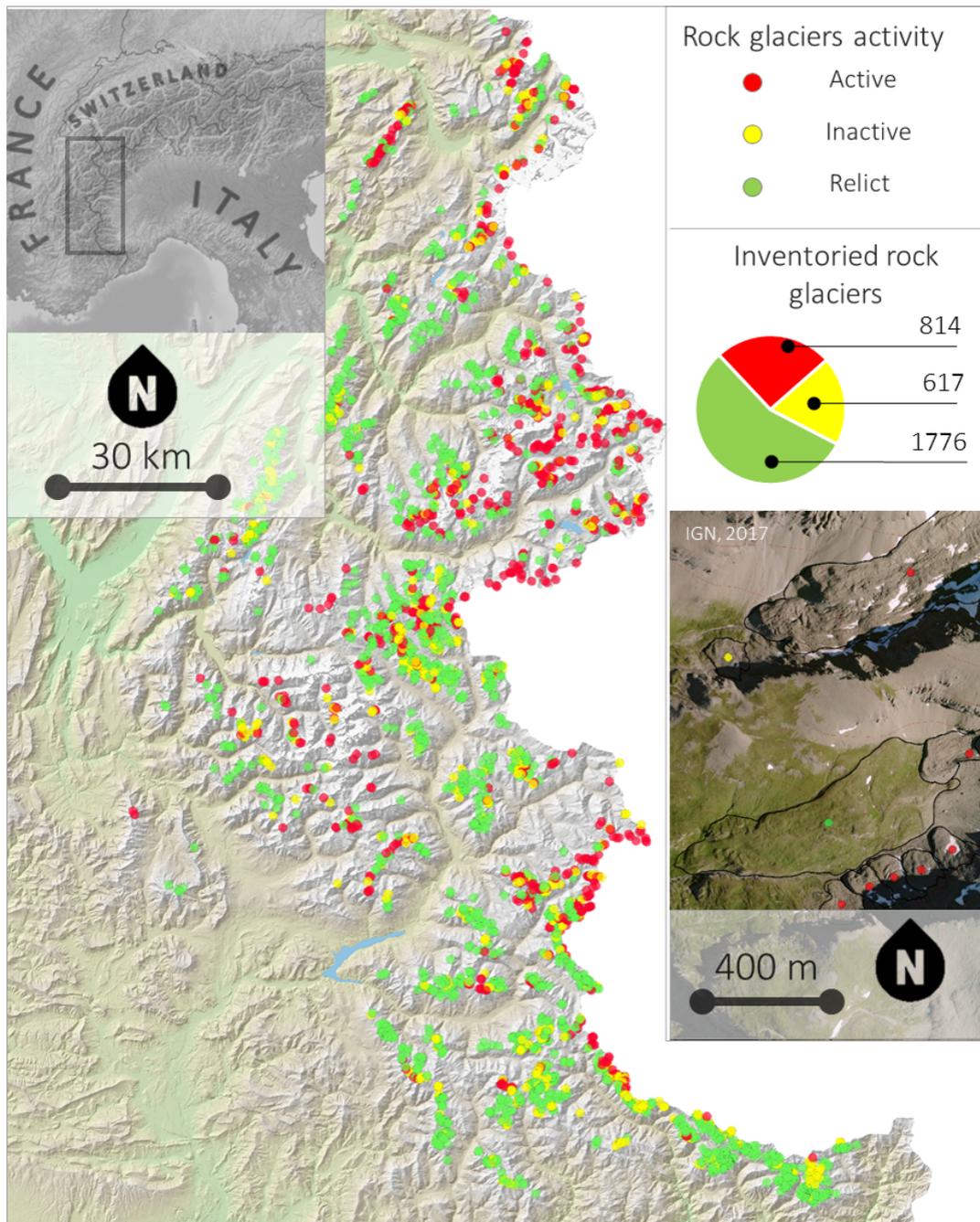


FIGURE 2.16: Rock glacier inventory as provided by the RTM

In the past five years, new cases of destabilisation were identified in the region.

By using InSAR analysis, Echelard (2014) spotted a fast moving rock glacier in the Vanoise range. The rock glacier, called *Pierre Brune*, was moving at several meters per year and presenting a sequence of scarps and crevasses on its surface. The first cracks appeared on the rock glacier surface in the 50's and slowly evolved since the 90's when destabilization occurred causing sudden acceleration of the frontal lobe and new crevasses formation.

On 14th August 2015, a debris flow flooded the town of Lanslevillard in Haute Savoie, causing damages for several hundred thousand euros (Schoeneich et al., 2017). The initiation points of the debris flow were found to be located on the front of the active rock glacier of *Lou* and caused a mass movement of 10 000 - 15 000 m³. Initiation points consisted in concentrated flows that eroded the active layer reaching the permafrost table. Further field inspection revealed that part of the rock glacier was most probably experiencing destabilization, as suggested by multiple cracks on the landform surface.

Serrano (2017) investigated the occurrence of rock glacier destabilization in the Haute Maurienne and Ubaye areas. The study highlighted that several rock glacier were showing evidence of destabilization observable from aerial imagery. Four of these sites were investigated by field survey and historical orthoimages analysis, proving the occurrence of the destabilization process (Obregon, 2018).

These cases, together with several observations in the neighbouring regions (e.g. Delaloye et al., 2013; Roer et al., 2008), suggest that rock glacier destabilisation is not an exceptional phenomenon. This requires systematic assessment of these landforms and related phenomena in order to be included in the periglacial risk assessment.

PhD strategy

The PhD aims to contribute to the knowledge of the phenomena related to destabilized rock glaciers in the French Alps. The main research question is to quantify and understand the occurrence of the destabilization phenomena in the region. The PhD also aims to contribute to the POIA-PERMARISK project by providing decision-support tools that may be helpful to identify potential hazards related to rock glacier destabilisation at the regional scale of the French Alps. This is done by following a rigorous strategy that follows three research axis (Figure 2.17). Each axis targets a specific issue of the problem.

Axis 1: Modeling permafrost spatial distribution In order to characterize permafrost related processes, it is important knowing permafrost distribution. Rock glaciers are landforms that fit into the regional permafrost zone and knowing its extents is fundamental to better understand the occurrence and characteristics of these landforms. Although permafrost distribution in the region was already investigated (Bodin et al., 2008; Boeckli et al., 2012b), the studies did not have access to extensive dataset covering the region, rising some questions concerning the validity of the results of these researches.

Therefore, the first step of the PhD project is to produce a permafrost distribution map specifically designed for the French Alps using. This was done by using the RTM rock glacier inventory as database of permafrost evidence and a statistical approach to model the relation between permafrost occurrence and terrain attributes. This allowed to produce a permafrost distribution map of the region, a tool also used in the others two actions of the POIA-PERMARISK projects to characterize permafrost conditions at infrastructures and thermokarst locations. Therefore, this step is fundamental for the whole project. This issue is described in Chapter 3 and is the object of Article I, presented in Chapter A.

Axis 2: Rock glacier destabilization assessment Thanks to the inventorying effort of the RTM, it is now known that rock glaciers are numerous in the French Alps, reaching several hundreds of active landforms. So far, only few landforms were identified as destabilized. As there is not a comprehensive study that aims to identify all the destabilized rock glaciers, this axis aims primarily to identify destabilized rock glaciers at the regional scale (Section 4.1). Destabilizing rock glaciers have been rarely estimated at the regional scale, as most of the studies focus on site-scale approaches. Therefore, in this axis a novel methodology to spot destabilization is proposed.

A second point investigated in this axis is the relation between terrain parameters and rock glacier destabilization occurrence in order to understand if there are typical topographical and climatic settings that precondition destabilization (Section 4.2). Understanding this relation may help to predict destabilization occurrence of active rock glaciers at the regional scale. This issue has not yet been tackled by the scientific community and therefore a novel methodology is proposed here. Overall, this axis represents not only the core but also the main challenge of the PhD project. This axis is described in Chapter 4 and is the object of Article II, presented in Chapter B.

Axis 3: Study case: rock glacier failure and realized risk Although hazards related to creeping permafrost destabilization have been theorized (Schoeneich et al., 2015), confirmed cases are scarce in the French Alps. Therefore, there is a lack of knowledge on the mechanisms that may lead to failure. Knowing these processes is relevant in order to understand the alert indicators prior to failure. Also, there is not a solid knowledge on which volumes can be involved in catastrophic events. These estimations are important to allow an assessment of the suitability of protection infrastructures facing these events.

Therefore, the last research axis of the PhD is to define and study one site presenting both destabilization and hazardous mass movements, aiming to understand the link between these two phenomena. The study site of the project is the Lou rock glacier, a destabilized rock glacier that triggered a consistent debris flow in August 2015 that caused damages to the town of Lanslevillard, Savoie. This issue, presented in Chapter 5 is the subject of Article III, presented in Chapter C.

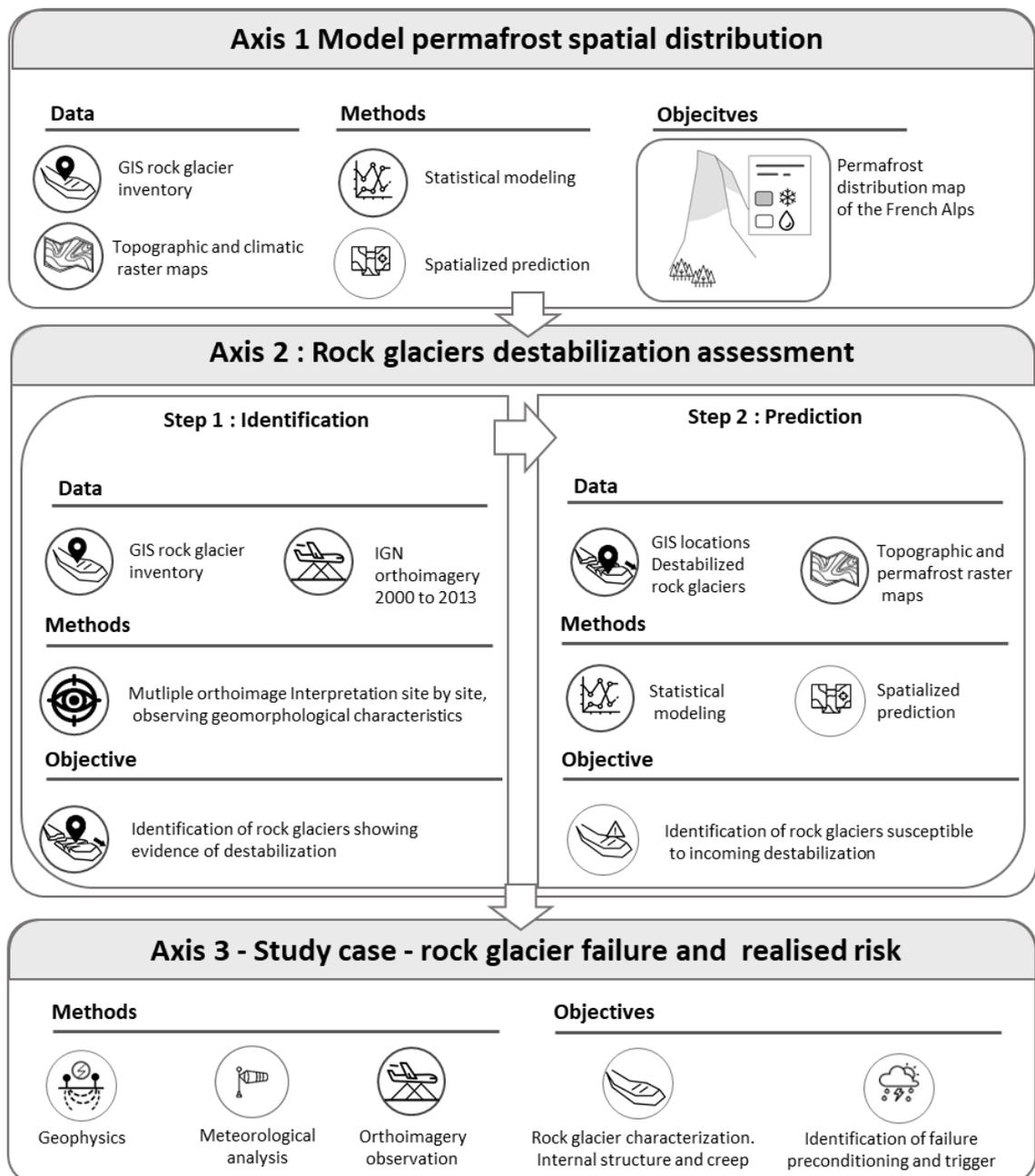


FIGURE 2.17: General workflow of the PhD project based on the three research axis, with specified objectives and methodologies.

Chapter 3

Modeling permafrost spatial distribution

Knowing the spatial distribution of permafrost is an important step in this study as it allows to better characterize permafrost conditions at sites of interest (Haeberli et al., 2010). In the French Alps, the spatial distribution of permafrost was described by two regional studies, one specifically designed for the region (Bodin et al., 2008) and one at the European Alps scale (APIM, Boeckli et al., 2012b). These studies shared the common weak point that the region was poorly covered by the database used, as permafrost evidences used to calibrate the models covered only partially the region. Therefore, there is still some uncertainty concerning the quality of these maps as their performance is unknown in areas not used to calibrate the models (e.g. Bonnaventure and Lewkowicz, 2008). The aim of the present chapter is to provide a synopsis of the first comprehensive permafrost distribution model in the French Alps, object of the Article I in chapter A.

3.1 Objectives and methodology

Permafrost is an invisible phenomenon that manifests itself on the surface only in localized landforms, as rock glaciers in mountain environments for instance. Knowing its extents beyond these landforms therefore requires a modelling approach which can be either physically based or statistical. Physically-based approaches reproduce ground temperature by modelling the interactions between atmosphere, snow cover and soil characteristics (Westermann et al., 2013). Statistical methods on the other hand are based on modelling correlations between observed permafrost evidences and typical topo-climatic patterns (e.g. Keller, 1992). Statistical models are therefore empirical and more simple to implement than numerical models, as they don't require the knowledge of all the physical processes behind ground temperature regimes. Due to their flexibility, these methods are the ones traditionally used to produce regional scale distribution maps.

Statistical permafrost distribution modelling is a subject widely covered by previous studies. First efforts date to the beginning of the 90s, when Keller (1992) implemented on a GIS a set of empirical rules-of-thumb deduced from observations on

sites where permafrost existence was observed. Since then, a number of studies with a similar approach took place, mainly covering Swiss mountain ranges (Imhof, 1996; Lambiel and Reynard, 2001; Frauenfelder, 2004). Rock glaciers are often used as permafrost evidence since they are observable on orthoimages, allowing to gather large datasets at the regional scale. Using rock glaciers to model permafrost distribution was done under the assumption that active and inactive landforms host permafrost and the local topo-climatic conditions at their location are suitable for the existence of frozen ground.

The paper by Boeckli et al. (2012a) is the study that substantially upgraded the methodology of statistically - based permafrost distribution modeling. Boeckli et al. (2012a) proposed the introduction of the logistic regression as statistical modelling tool. This method, made popular in rock glacier occurrence modelling by Brenning (2005), has the advantage to evaluate the probability of a binary variable, e.g. presence/absence of permafrost, in a continuous range of probability. This probability can be interpreted as an index describing the suitability of local topo-climatic conditions to the realization of the dependent variable. This feature is a great advantage in the production of a map, which consists in smooth transitions between areas with expected presence/absence of permafrost, while previous models resulted in an unnatural discrete transition between permafrost presence and absence. After Boeckli et al. (2012a), later studies, as well as the work proposed in this PhD project, adopted the methodology proposed with only minor variations (Sattler et al., 2016; Azócar et al., 2017).

Following this methodology, this research axis aimed to produce the first permafrost distribution map of the French Alps using a database covering the whole region, i.e. the RTM rock glaciers inventory. Active rock glaciers were used as evidences of permafrost existence, while relict rock glaciers were used as proof of permafrost absence. Using a GIS approach, several topographical and climatic attributes were sampled at rock glaciers locations in order to model the occurrence of permafrost presence/absence in relationship to these terrain attributes. Terrain attributes were chosen as proxies of processes determining the existence of permafrost and involved mean annual air temperature, potential incoming solar radiation and precipitation patterns (Boeckli et al., 2012a). This allowed to get insights on the relationship between climatic patterns and rock glacier permafrost existence in the region and to predict permafrost occurrence at the regional scale (for a general workflow example, see Figure 3.1).

In addition to the standard procedure to produce the permafrost distribution map, the present study proposed some variations with respect to the classical modelling approach. These methodologies were inspired by previous studies in fields of permafrost and landslide spatial modelling and aimed to propose a more solid statistical process. Proposed improvements are here described.

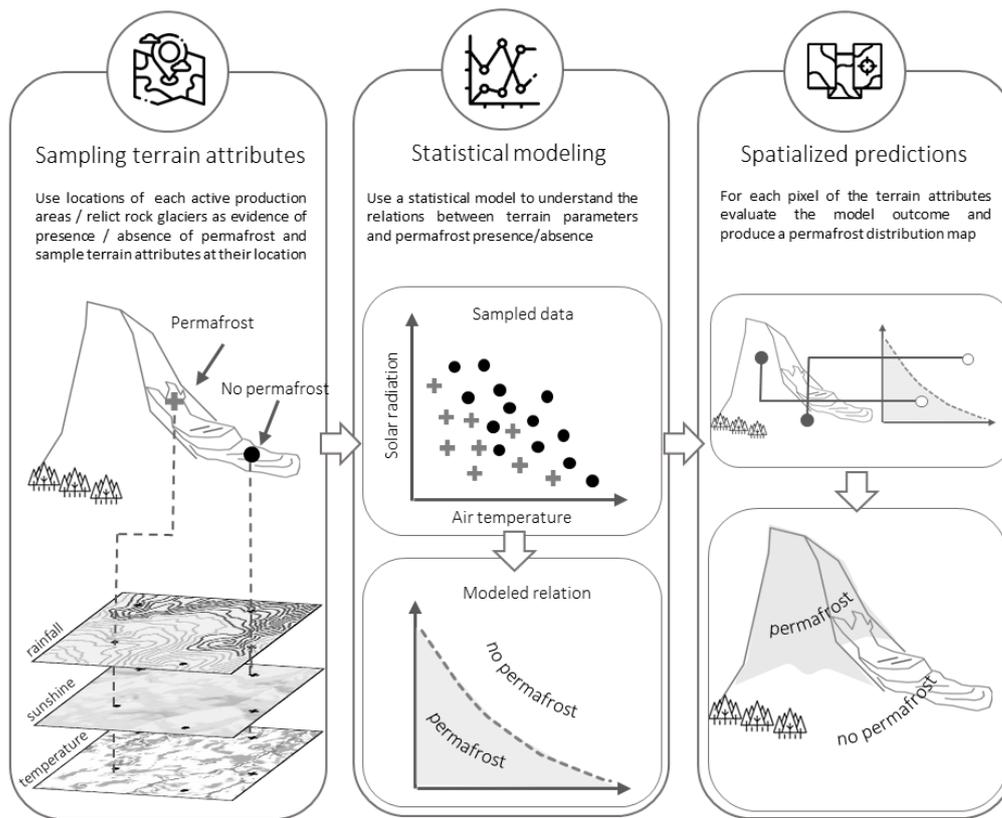


FIGURE 3.1: General workflow of statistical modeling applied to permafrost distribution prediction.

Significance of rock glaciers in statistical modeling Boeckli et al. (2012a) hypothesized that intact rock glaciers are evidence of the existence of permafrost. Centroids of the landforms were taken as sampling point for evaluating local topo-climatic conditions at rock glaciers sites and build the database for statistical modelling. However, several authors also pointed out that rock glacier dimensions depends on topographical and geological settings which influence rock glacier dynamics as well (Matsuoka and Ikeda, 2001; Frauenfelder et al., 2003; Brenning, 2005; Janke and Frauenfelder, 2007). Faster or larger rock glaciers may flow into non-permafrost areas before encountering climatic inactivation. This suggests that the location of the landform centroid is not solely dependent on climatic conditions, but also on the dynamics and geometry of the rock glaciers. Therefore, the hypothesis that rock glacier locations are an evidence of climatic suitability is not completely correct, creating a bias in the model.

In this sense, rock glaciers can be seen as a micro-environment where permafrost can exist in otherwise un-favourable conditions. Coarse blocks provide ventilation that cools the ground (Balch effect, Balch 1900 in Haeberli, 1985) and insulate the permafrost body. Several studies (e.g. Harris and Pedersen, 1998) tried to estimate this thermal offset between rock glacier and nearby ice-free ground, which may vary

between 2 and 7°C. In order to compute the APIM, Boeckli et al. (2012b) aware of this issue compensated this effect by integrated in the permafrost model a thermal offset of 2°C. However, this method was described in the study as the greatest source of uncertainty concerning the APIM. Therefore, further efforts may target this issue.

A solution to the issue was proposed by (Sattler et al., 2016), which considered that the production area (rooting zones) is a more consistent evidence of the location where topo-climatic conditions are favourable to permafrost existence. Without a production area suitable to permafrost formation the rock glacier would not exist. This hypothesis, which was already suggested by Bodin et al. (2008), was supported by the present study and used in the modelling framework. Therefore, for each active rock glacier in the RTM inventory, production areas were digitized using orthoimages interpretation. Rock glaciers with ambiguous production areas were discarded. Furthermore, rock glaciers with clear glacial origin were discarded, since, in this case, the formation of the production area originates from processes different from periglacial permafrost aggregation.

Model validation method In order to understand which is the goodness of the model in predicting permafrost distribution occurrence, models are typically validated using a cross validation method. Cross validation consists in leaving out a part of the database, e.g. 10%, and training the model with the remaining 90%. The left out part, called validation set, is used then to evaluate the performance of the model by comparing model prediction and actual data (Hand, 1997). This framework is repeated several times by randomly subsetting the database in validation and training sets, giving an insight of the model in predicting permafrost occurrence in areas where data is missing. Although this method is well established in statistics and used by several studies concerning permafrost (Boeckli et al., 2012a, e.g.), it is known that cross validation is a poor validation method in spatialized models. Indeed, leaving out randomly a portion of the data-set results in training and validation points that are spatially close to each-others, making impossible to detect whether the model poorly performs in specific areas of the study site (Ruß and Brenning, 2010).

To solve this issue, in landslide science regular cross validation is often replaced by spatial cross validation methods (Goetz et al., 2011). Spatial cross validation consists in selecting a spatially grouped cluster of data from the database as validation set. In this way it is possible to estimate model performance in an area where data was not used for calibration. This is a more consistent method as it gives a proper quantification of the model predictive power in areas where data is missing. In the present study, spatial cross validation was used as validation method. Spatial clusters were selected by dividing the study region into 5 sub-regions with respect to lithological, topographical and climatic conditions that may influence permafrost distribution. For each subregion a model was trained and tested on the rest of the French Alps.

Uncertainty estimation Statistical modelling based on inventory involves a propagation of inventory errors, e.g. bias related to operator's interpretation of rock glacier activity and shape, on the modelling results (Steger et al., 2016). To date, no study investigated the effects of this source of uncertainty in the permafrost distribution modelling framework. The present study proposed an estimation of the effects of inventory uncertainties on the modelling performance. To do so, two main sources of uncertainty were investigated.

The first source of uncertainty consisted in the shape of the rock glacier, which may vary according to the operator in charge (Schmid et al., 2015). Different rock glaciers shapes cause a variation of the centroid position which, at its turn, causes a variability in the sampled terrain attributes and therefore on the model outcome. The effects of this source of uncertainty were assessed by having different operators digitizing the same 10 rock glaciers. The shapes were compared and the variability in the resulting position of the sampling point allowed to estimate the variability in the sampled predictor variables related to this source of uncertainty. This variability was then used to perform a Monte Carlo simulation consisting of 1000 model runs in which the values of the predictor variables were randomly perturbed by the estimated variability. By comparing the resulting models it was possible to assess the error propagation of this inventory uncertainty in the final models.

The second source of uncertainty consisted in the attribution of the activity based on geomorphological indicators, which may also be dependent on the subjectivity of the operator. This source of uncertainty is well known in rock glaciers inventories and it is due to the fact that interpretation of the geomorphological feature characterizing the activity is often not sharp between classes. As results inactive rock glaciers may be classified as active and vice-versa, violating the statistical hypothesis behind the model. To assess this source of uncertainty, active rock glaciers creep was investigated by observing orthoimages taken at different dates. Several orthoimages covering the French territory from 2000 to 2013 are made available by the national geographic institute (IGN) either online on [Géoportail](#), or as WMS (Web Mapping Service) for institutions (IGN, 2017). These images allowed to observe rock glaciers movements and better characterize rock glacier activity. Therefore each rock glacier was inspected for movements over a time span of one decade and assigned an index describing if movements were observable. Considering time between orthoimages and resolution, only rock glaciers moving more than 0.3 m/y could be observed as moving. To estimate the influence of activity attribution uncertainty on the final two models were computed, one with the allegedly active rock glacier in the original RTM inventory and one with only the rock glaciers observed to be creeping in the IGN orthoimages. The two models were then compared in order to observe the permafrost distribution difference in relation to this source of uncertainty.

3.2 Main results

Permafrost distribution was expressed by the Permafrost Favourability Index (PFI, Figure 3.2), a probability scale ranging from 0 (maximal probability of permafrost absence) to 1 (maximal probability of permafrost existence). Although the term *permafrost distribution* is used across the study, it is emphasized that the PFI does not describe the actual permafrost distribution. Due to the statistical hypothesis behind the model, the PFI describes the topo-climatic suitability of any given location in the map to the existence of an active rock glacier production area. In other words, when PFI is close to 1, then the local topo-climatic conditions are suitable to the existence active production areas. On the other hand, when the PFI is close to zero, then the local topo-climatic conditions are usually suitable to the existence of relict rock glaciers. Local permafrost patches due to processes that are not relevant in rock glacier existence at the regional scale, as snow cover or wind patterns, cannot be represented by this approach. Also, considering that active rock glaciers are representative of the climate of cold epochs of the Holocene (e.g. Lambiel and Reynard, 2001; Cossart et al., 2010), this map describes the permafrost favourability in a no longer valid climatic context as rock glaciers may still be active despite warmer temperatures thanks to their thermal inertia.

Overall, permafrost occurrence was found to be favourable at a Mean Annual Air Temperature (MAAT) lower than 1.8°C on north facing slopes and lower than -1.8°C on south facing slopes on the Northern Alps. These values are 1°C higher in the Southern Alps. The permafrost zone covers about 770 km^2 in the region, excluding bedrock permafrost not accounted by the model. Although the PFI and APIM (Boeckli et al., 2012b) are based on dissimilar statistical hypothesis and comparison should be cautious, the two maps are very similar and differences seemed to be minimal. It is therefore suggested that the APIM map is valid for the French Alp region.

Permafrost existence was found, coherently to previous studies, to be mostly determined by elevation and solar radiation. It was also found that the latitude had a significant role in permafrost distribution. Latitude is a proxy for precipitation regional patterns that vary thanks to the different influences of Atlantic and Mediterranean circulations. Analysing regional precipitation data from Gottardi (2009), it was found that seasonal precipitation distribution may play a relevant role in this pattern. In the Southern Alps, fall months are drier than in the Northern Alps and winter precipitation is more abundant in late winter (Durand et al., 2009). This is suggested to influence ground temperatures by allowing more efficient early winter frost in the southern Alps, explaining permafrost at higher MAATs in this region.

Lithology was also suggested to play a significant role in permafrost distribution. Permafrost existence was found to be more favourable in crystalline lithology as the lower limits of permafrost are about 100 m lower compared to sedimentary lithology. Crystalline lithology tends to produce blocky rock glaciers (Matsuoka and Ikeda, 2001) having coarser surface that can enhance the ventilation effect. It is suggested

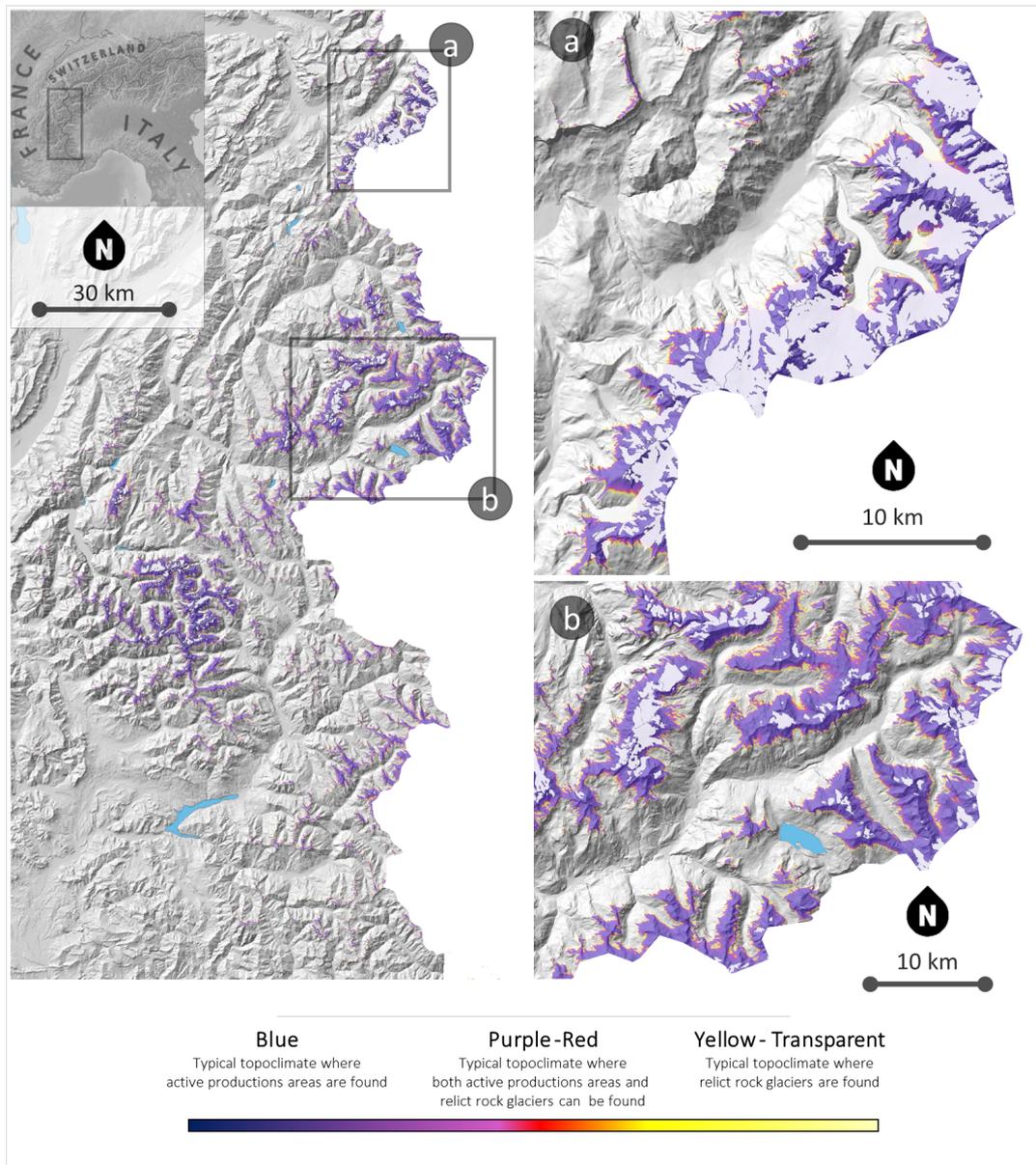


FIGURE 3.2: The Permafrost Favourability Index map (PFI) in the French alps with explanatory detail of (a) Mont Blanc and (b) Vanoise ranges.

therefore that, thanks to this difference in granulometry between blocky and pebbly rock glaciers, permafrost distribution may follow regional scale lithological patterns (Janke and Frauenfelder, 2007).

Model performance and uncertainties The statistical model achieved an *outstanding* performance (Hosmer and Lemeshow, 2000), confirming the suitability of this method to describe permafrost occurrence at the regional scale suggested by previous studies (Boeckli et al., 2012a; Sattler et al., 2016). Spatial cross validation suggested that the model had a high predictive power. Nevertheless, it was found that using localized inventory to extrapolate permafrost distribution at a larger scale may

introduce severe biases.

Modeling outcome was significantly sensitive to inventory uncertainty. Digitalisation uncertainties may produce errors up to 0.2 PFI points. Activity attribution uncertainty was even more compromising as errors reached 0.35 PFI points in the permafrost probability. This highlighted the importance of having a high quality database in modelling permafrost spatial distribution with this approach.

3.3 Conclusions

This study allowed to understand and define the spatial pattern of the debris permafrost zone in the French Alps. Permafrost was confirmed to be widespread in the region, covering a relevant portion of the French Alps, i.e. 770 km². For comparison, white glaciers cover 280 km² (Gardent, 2014). This suggested that there is a conspicuous amount of this hidden cryosphere in the region. Permafrost was found at high altitudes, above 2300 – 2700 m.a.s.l. in the Northern Alps and 2500 – 2900 m.a.s.l. in the Southern Alps. Differences between Northern and Southern Alps were suggested to be due to the latitudinal variability of precipitation patterns determined by the Atlantic and Mediterranean influences. This involved that climate change may have effects on permafrost distribution at the regional scale not only by air temperature variations but also by changing precipitation patterns (Harris et al., 2003). Lithology was also found to have a significant effect on permafrost distribution, possibly due to debris size influencing ventilation effect.

The present PFI map has a strong relevance in the context of the PhD and POIA-PERMARISK projects as it identifies the permafrost areas in equilibrium with a LIA climate. Permafrost located at lower altitudes may possibly be subjected to more intense thaw as conditions for permafrost existence are not met anymore due to the present temperature warming and permafrost may still exist thanks to thermal inertia of the ground (Hoelzle and Haeberli, 1995). This feature will be of fundamental importance for the Chapter 4, i.e. estimating the spatial occurrence of rock glacier destabilisation. Estimating the extent of this area will give an insight on the relation between thaw occurrence and rock glacier destabilization.

3.3.1 Suggested improvements

The present study follows the classical statistical approach to produce a regional scale permafrost distribution map, achieving comparable results to previous studies conducted with similar approaches but different databases (Boeckli et al., 2012a). It is therefore not recommended to try ameliorate the map using the same statistical approach. The major improvement needed to complete the map is to integrate a bedrock permafrost distribution model in order to cover all the mountain permafrost distribution as proposed by Boeckli et al. (2012b). Currently, in France, bedrock permafrost is described in the Mont Blanc range by Magnin (2015) using a statistical approach based on a temperature loggers network. Nevertheless, bedrock permafrost

is most likely present, at least, in the highest rock faces of the ranges of Ecrins (peaking at 4102 m.a.s.l.) and Vanoise (peaking at 3855 m.a.s.l.). In these ranges there are no systematic temperature measurements oriented to detect permafrost. Therefore, considering that bedrock permafrost characteristics may strongly vary from massif to massif, a monitoring effort in these ranges is necessary to have a suitable database for modelling.

In the context of permafrost distribution modeling, a more detailed map can be produced using a physically-based approach proposed by Westermann et al. (2013). This method allows to create a ground temperature map based on the thermal exchange between ground, snow cover and atmosphere. The resulting map reproduce local permafrost patterns not achievable by the statistical approach used here. Also, the map is not stationary, i.e. it does not represent a single climatic condition, as it is the case for the PFI map which is significant only for the LIA conditions, but represents the ground thermal evolution in relation to the climatic time series. This allows to integrate ground temperature and permafrost thaw in future climatic scenarios, an interesting feature in the context of climate change (Westermann et al., 2015; Westermann et al., 2016). Nevertheless, it is emphasized that such modeling approach requires a large amount of data as spatialized climatic time series and ground characterization that may be challenging to acquire. If future efforts will engage this issue, it is recommended to set up strong collaborations with the institutions that are specialized in producing or obtaining such data.

Chapter 4

Rock glacier destabilization assessment

Although in the French Alps several studies reported the occurrence of rock glacier destabilization (Echelard, 2014; Bodin et al., 2016; Ribeyre, 2016; Serrano, 2017; Obregon, 2018), there is not a comprehensive assessment of the incidence of this phenomenon. Quantifying the incidence of this phenomenon at the regional scale has a double interest. First, there is a need to identify where destabilized rock glaciers are located in a hazard assessment perspective. Destabilized rock glaciers involve sudden acceleration that, in topographic settings of high connectivity between the front and the torrential gully, may be linked to increasing frequency and magnitude of mass movements. Second, the destabilization process is still not completely understood and regional scale investigation may contribute to its understanding. As the French Alps host a large number of rock glaciers, destabilization may be a common feature, allowing to understand its general characteristics. According to these two interests, the present study aimed to perform a regional-scale assessment of rock glacier destabilization occurrence. The issue was tackled by following a two steps methodology:

1. *Identification of destabilized rock glaciers* The first step was to identify all the rock glaciers that present evidence of destabilization. The result was an inventory of destabilized rock glaciers, a tool that can give an insight of regional occurrence of the phenomenon as well as the identification of the most interesting sites.
2. *Identification of rock glaciers susceptible to destabilization* The second step aimed to understand and model the typical terrain attributes in which destabilization occurs. In this way it was possible to evaluate which terrain attributes are significant in preconditioning the occurrence of the destabilization. This also allowed to map areas where there is a susceptibility to destabilization occurrence, highlighting areas that present potential development of destabilization.

In the present section, these two steps are described into two independent subsections. At the end of the chapter it will be drawn a general conclusion (in section

4.3), highlighting the usefulness of the results and their limitations in the POIA-PERMARISK and research scenario.

4.1 Identification of destabilized rock glaciers

Identifying destabilized rock glaciers at the regional scale represents a significant challenge. Destabilized rock glaciers have been traditionally studied using a site-by-site fieldwork approach, involving detailed analysis of historical orthoimagery, GPS measurements and ERT surveys (e.g. Bodin et al., 2016; Scotti et al., 2016; Eriksen et al., 2018). This approach allows to gather important knowledge on the landform and the processes involved into the destabilization. Nevertheless, it requires an amount of effort not applicable for a regional scale assessment.

Recognizing abnormal creeping rates at the regional scale can be done by using an InSAR approach (Delaloye et al., 2008a; Barboux et al., 2012; Echelard, 2014). Despite some methodological challenges, this method can be very precise and useful, allowing to obtain precise quantifications of creeping rates and identify destabilized landforms, e.g. the Pierre Brune rock glacier (Echelard, 2014). Nevertheless, destabilization may not necessarily involve abnormally high creep rates, as observed on the Lou rock glacier, creeping at *only* 3 m/y (Schoeneich et al., 2017). Also, some destabilized rock glaciers were observed to present low creeping rates before sudden failure (e.g. Bodin et al., 2016). This method may therefore not spot *slow-moving* destabilized rock glaciers and those landforms at the edge of the failure.

To overcome these issues, the present study proposed a novel methodology based on multiple orthoimagery interpretation of geomorphological characteristics of the active rock glaciers. The IGN provides three orthomosaics at high resolution (50 cm x 50 cm/pixel to 1 x 1 m/pixel depending on the area and year) from 2000 to 2013 covering the entire national territory (IGN, 2017). By comparing orthomosaics of different years, it was possible to observe rock glacier activity and their geomorphological evolution in the past decade. In order to spot destabilization, it was necessary to define the typical characteristics of the destabilized rock glacier that could be observed on multiple orthoimagery. This was done by a combination of fieldwork investigation, literature analysis and historical imagery observation, aimed to understand the recurrent patterns occurring in these landforms. Overall, reported cases of known ongoing destabilization present (i) surface disturbances that (ii) increase in time by number and/or dimension and (iii) determine a discontinuity in the creeping pattern of the landform (Figures 2.12 2.11 and 2.13). These characteristics seem to be recurrent in every case of known destabilization and were therefore used in our recognition process.

Nevertheless, active landforms presented a wide variety of these characteristics and a binary classification in *stable* and *destabilized* was problematic due to many cases presenting intermediary characteristics between the two states. Therefore, two

intermediate classes of destabilization, i.e. *unlikely destabilized* and *suspected of destabilization*, were used. These two classes express the fact that the rock glacier presents some typical characteristics of destabilization but they are not as pronounced as in typical cases of destabilization. In particular, likely stable rock glaciers present surface disturbances slightly pronounced and non-developing in time, while landforms suspected of destabilization have more pronounced surface disturbances developing in time that however are not associated with typical discontinuous creeping pattern observed in destabilized landforms. These different degrees of destabilization were called here *destabilization ratings*, varying from zero (stable rock glaciers) to three (potentially destabilized rock glaciers).

An additional issue concerned the definition of potentially destabilized rock glaciers as this definition was based on known cases of destabilization, which are mostly concerned by blocky rock glaciers, while in the French Alps there is a large incidence of pebbly rock glaciers. While destabilization on blocky rock glaciers was often described to occur in relation to deep surface disturbances as crevasses and scarps, in fine-grained landforms was observed a large incidence of shallow cracks as only typology of surface disturbances. This large incidence of cracks on pebbly rock glacier rose the question whether this surface disturbance was a significant indicator of destabilization or rather a natural consequence of the extensive creep pattern in a context of a generalized creeping acceleration in the past decades. On the other hand, some studies presented destabilized rock glaciers showing shallow cracks, suggesting that this feature is a significant evidence of destabilization (Serrano, 2017; Schoeneich et al., 2017). Overall, at the current stage, there is not enough knowledge to provide a convincing answer to this question, which represents a source of relevant uncertainty in the present study. To acknowledge this source of uncertainty, potentially destabilized rock glaciers were divided into two categories depending on the type of surface disturbances observed, i.e. cracks versus crevasses and scarps.

Once defined this methodology to consistently classify each active landform, it was performed a detailed orthoimages interpretation. Each active rock glacier was inspected using the available orthoimagery and comparing different dates, aiming to observe creep pattern, surface disturbances occurrence and evolution. destabilization rating was assigned to each landform according to its observed characteristics (Figure 4.1). This was done by three different operators working independently and results were then compared, in order to limit personal biases in the interpretation. As the knowledge and skills of the operators evolved thorough the process thanks to the discussions about dubious cases, several comebacks on the rating were performed.

4.1.1 Results

The study identified destabilization features on 259 rock glaciers, i.e. more than 50% of the active landforms in the region. Most of these landforms were rated as unlikely destabilized (127) and suspected of destabilization (86). In total were identified 46

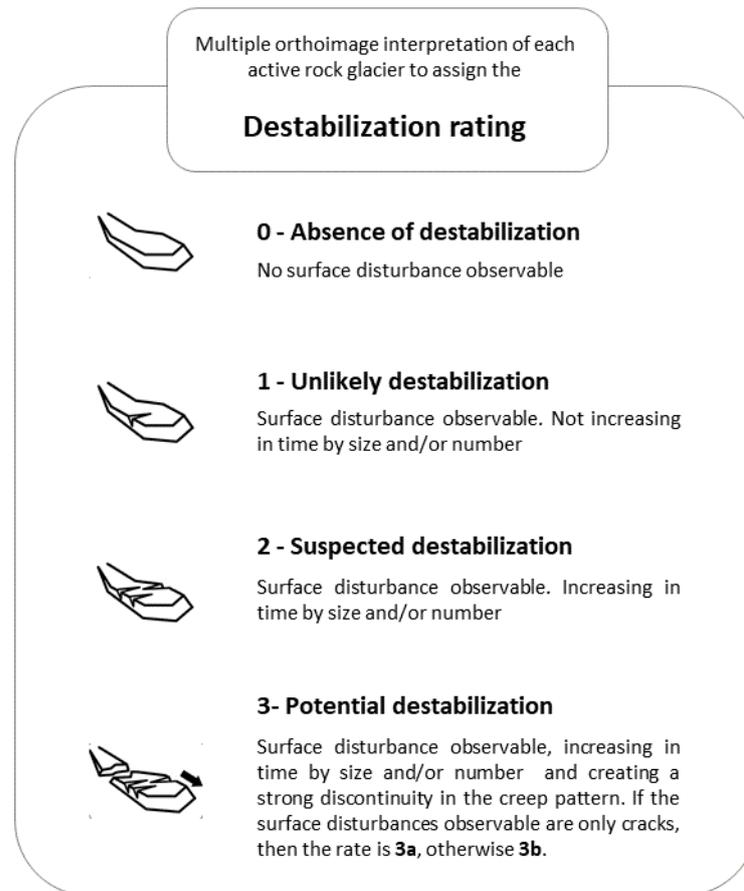


FIGURE 4.1: Destabilization rating, attributed by multiple orthoimages observations of the occurrence and evolution of the surface disturbances and creep pattern.

potentially destabilized rock glaciers, i.e. the 11.7% of the active landforms in the region. Of these, 33 were related to shallow cracks occurrence, while potentially destabilized rock glaciers associated to deep surface disturbances presented a wide variety of morphologies, from a single crevasses cutting the whole landform body to multiple surface disturbances disturbing the entire surface of the rock glacier. In general, destabilization features were mainly observed in the Vanoise and Ubaye regions, two areas characterised by densely jointed lithology and pebbly rock glaciers (Figure 4.2). Complementary to this, cracks were the most observed surface disturbance (on 167 landforms), while crevasses and scarps were more rare (on 40 and 27 landforms respectively).

The main limitation of the method was found to be the subjectivity of the classification. During the process there was the feeling of being subjected to the *prevalence induced concept change* (Levari et al., 2018), i.e. that bias that varies human judgement according to the amount of a certain feature in a group rather than their actual characteristics. In other words, classification tended to get milder (stricter) if a group

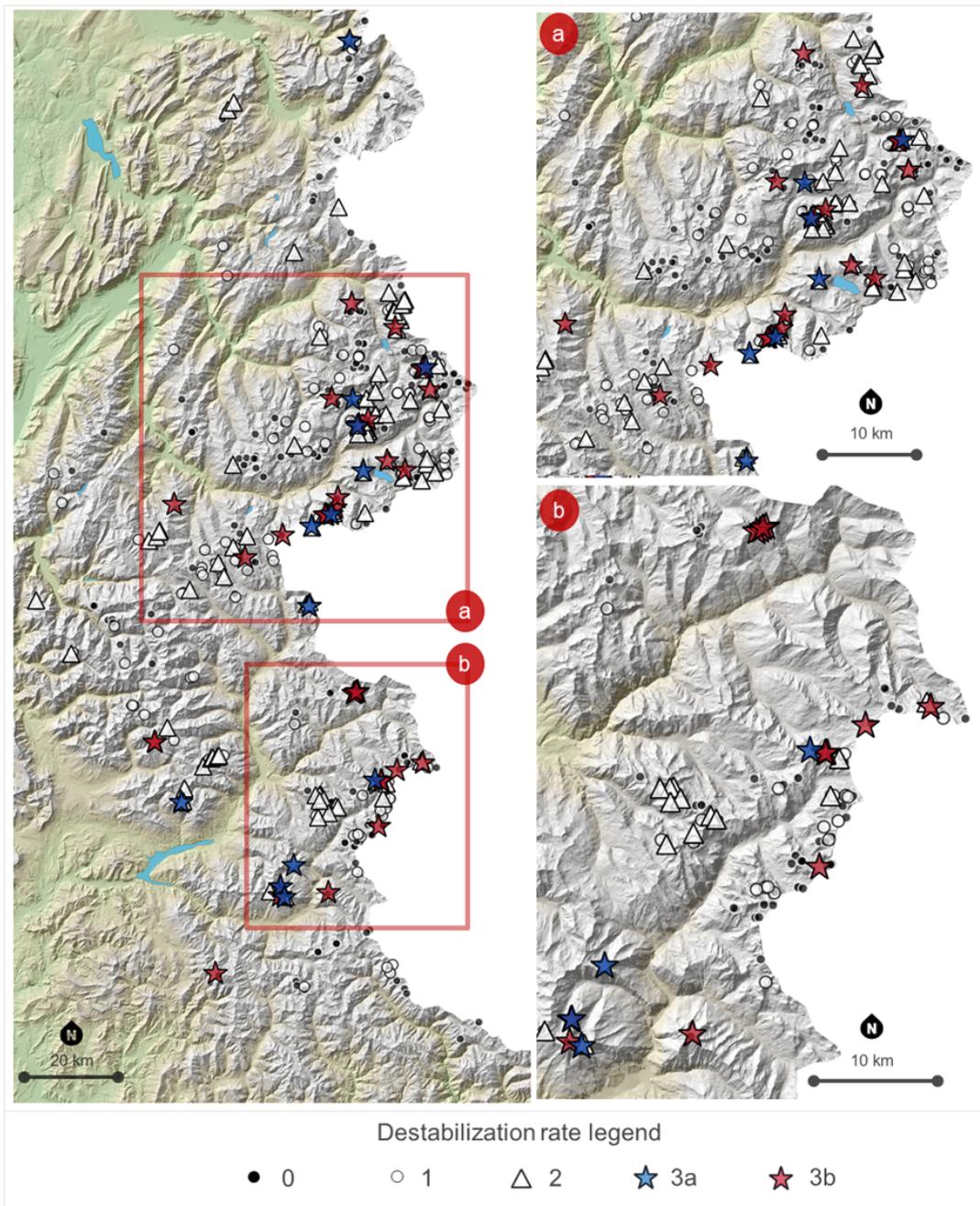


FIGURE 4.2: Active rock glaciers by destabilization rating in the French Alps, with focus on the Maurienne (a) and Ubaye (b) areas.

of stable (unstable) rock glaciers were encountered consecutively. Despite the use of several operators, experience sharing and comebacks on the process, some landforms were subject of long discussion and it was difficult to find agreement between the operators.

4.2 Identification of rock glaciers susceptible to destabilization

The first step of this study allowed to identify several potentially destabilized rock glaciers in the French Alps. This database offered the possibility to investigate the occurrence of destabilization at the regional scale by aiming to address two main issues. First, it was investigated if there are recurrent terrain features that increase the local susceptibility to rock glacier destabilization. Understanding these relations between destabilization and terrain characteristics would allow to have an interesting insight on the destabilization phenomena. The second issues was to define the rock glacier areas that are susceptible to destabilization, i.e. those rock glaciers that may present low destabilization rating but are located in topographic conditions where destabilization typically takes place. This was investigated by extrapolating the typical conditions in which destabilization occurs to the areas occupied by active rock glaciers. In this sense, it was possible to produce a *susceptibility map* to rock glacier destabilization a tool that identify areas where terrain attributes are suitable for destabilization occurrence. A rock glacier located in this area, although not showing destabilization features, may be susceptible to incoming destabilization if subjected to proper trigger.

Understanding and mapping rock glacier destabilization susceptibility was done following a classical statistical modeling approach in landslide science (Goetz et al., 2011). This method was conceptually not dissimilar from the one presented to model permafrost distribution in the previous issue (see Figure 4.3). Stable and potentially destabilized rock glaciers were treated as a binary variable indicating rock glacier stability and instability respectively. Stability and instability locations were used to sample terrain attributes as elevation and slope, creating a large database for modelling. Relationships between stability and instability occurrence and the terrain attributes were modelled using the statistical model Generalized Additive Model (GAM) with logistic link function. The GAM model allowed to understand the conditions in which destabilization occurs, e.g. typical slope angle and altitudes, and predict destabilization susceptibility at rock glacier sites.

Terrain attributes were chosen to describe the local susceptibility to destabilization based on in-situ observations on known cases of destabilization reported by previous studies. Destabilization was observed to occur on moderately steep slopes and convex profile curvatures. This was suggested to be due to the extensive flow patterns in convex and steep profiles that may compromise the cohesion of the permafrost body by increasing internal stress (Roer et al., 2008; Delaloye et al., 2013). Also shading was included in the analysis as it controls snow cover characteristics as north facing slopes host snow patches longer through the summer season, a process that may be relevant in water supply during summer.

The relation between permafrost degradation and rock glacier destabilization

was also investigated. Here, permafrost degradation was described by the Potentially Thawing Permafrost (PTP), i.e. the *melting area* in Lambiel and Reynard (2001). The PTP consisted in the lower margins of the permafrost zones and described an area in which climatic conditions are no longer suitable to permafrost existence due to atmospheric temperature increase. The PTP therefore does not directly quantify the degrading permafrost, but rather represent a proxy of degradation, as frozen ground is expected to thaw and/or increase water content in these sensitive areas (Hoelzle and Haeberli, 1995; Lambiel and Reynard, 2001).

All the terrain attributes described above were sampled at potentially destabilized and stable rock glacier locations and used to model destabilization occurrence. Once the model was computed, its predictive power was assessed using a spatial cross validation approach, as proposed for the PFI model in the previous section (see section 3.1). The model was then used to produce a destabilization susceptibility map, i.e. a map that predicts the susceptibility to destabilization in any given location. Rock glacier destabilization susceptibility was classified from *very low* to *very high* according to its modelled susceptibility value. It is emphasized that the map has value only within active rock glaciers.

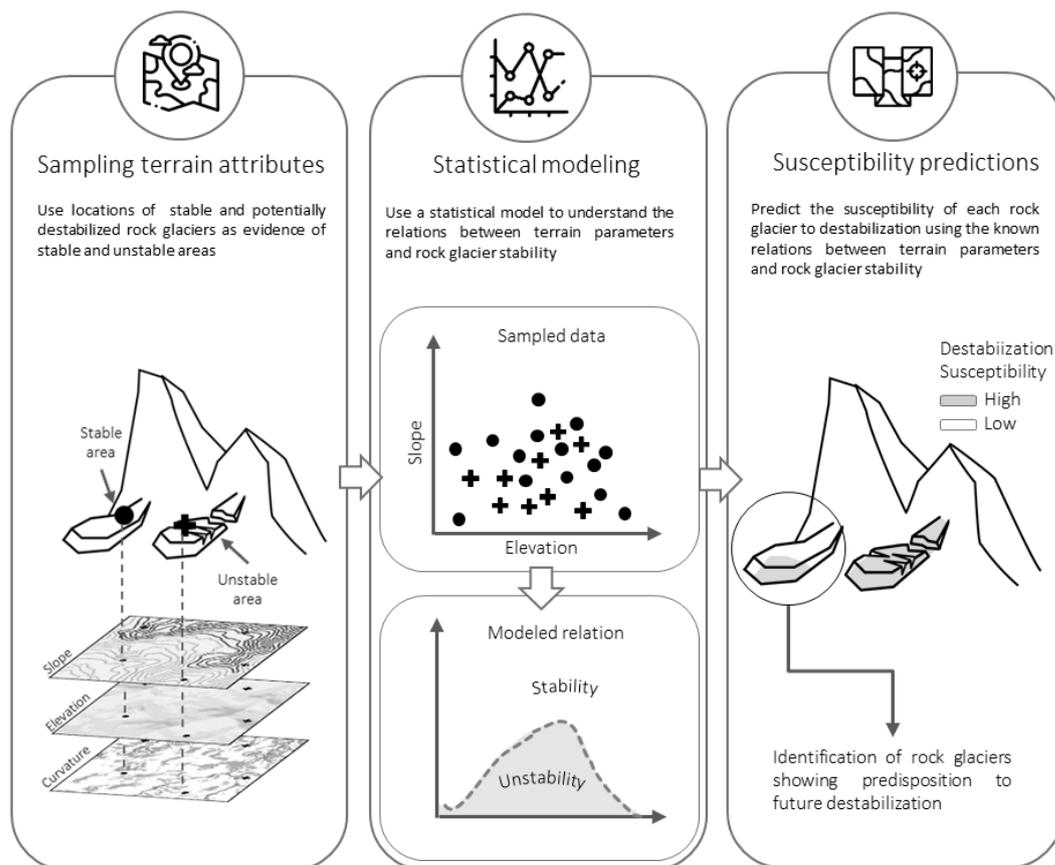


FIGURE 4.3: Conceptual scheme used to understand and predict rock glacier destabilization susceptibility.

4.2.1 Results

Destabilization occurrence was found to be significantly correlated with terrain attributes, suggesting that typical settings that precondition creeping permafrost destabilization exist. In particular, steep and convex slopes were found suitable for this phenomenon, in agreement with previous studies that identified in higher internal stress in steep soils and in extensive flows due to convexity (e.g. Delaloye et al., 2013). The study also pointed out that destabilization is more likely in shady areas, as most of the potentially destabilized rock glaciers were located on North-facing slopes. Although with the current elements it is difficult to provide a convincing explanation for this phenomenon, it is suggested to investigate the relationship between snow cover characteristics, meltwater supply and destabilization mechanisms in order to have an insight to this issue. It was also found that rock glacier destabilization was more likely to occur at the lower margins of the permafrost zone, i.e. where the PTP was highly expected. This suggests that rock glaciers in these areas are more susceptible to destabilization, possibly due to a stronger impact of atmospheric warming on the frozen ground.

The GAM model was found to have a *good* predictive power in the extrapolation of the susceptibility (Hosmer and Lemeshow, 2000), suggesting that this framework was significant in modelling rock glacier destabilization occurrence (Figure 4.4). The computed map indicated that about 75% of the creeping permafrost had a low or very low susceptibility to destabilization. High susceptibility covered 10% of the active rock glaciers, for a total surface of 2.8 km². A considerable amount, i.e. more than 50%, of this area was located in rock glaciers with lower classes of destabilization, suggesting that there is a high potential for incoming destabilization.

4.3 Conclusions

The issue of the occurrence of destabilizing rock glaciers in the region was tackled by novel methodologies operating at the regional scale, achieving interesting results. The results showed that destabilization is a common phenomenon as several rock glaciers showed evidence of the occurrence of this process. The results also indicated the relations between terrain attributes and destabilization occurrence, suggesting interesting patterns. This allowed to identify those rock glaciers that may be susceptible to future destabilization and to observe that there is an high potential for further development of this process. Overall, it is suggested that rock glacier destabilization is a relevant process to be investigated and to be monitored in the future.

In this sense, this study axis provided two main tools to be used in this context: the rock glaciers destabilization rating and the rock glacier destabilization susceptibility map. These products may be useful to define monitoring and survey priorities. The destabilized rock glaciers rating identifies rock glaciers that already presented typical characteristics found in known cases of destabilization and that therefore

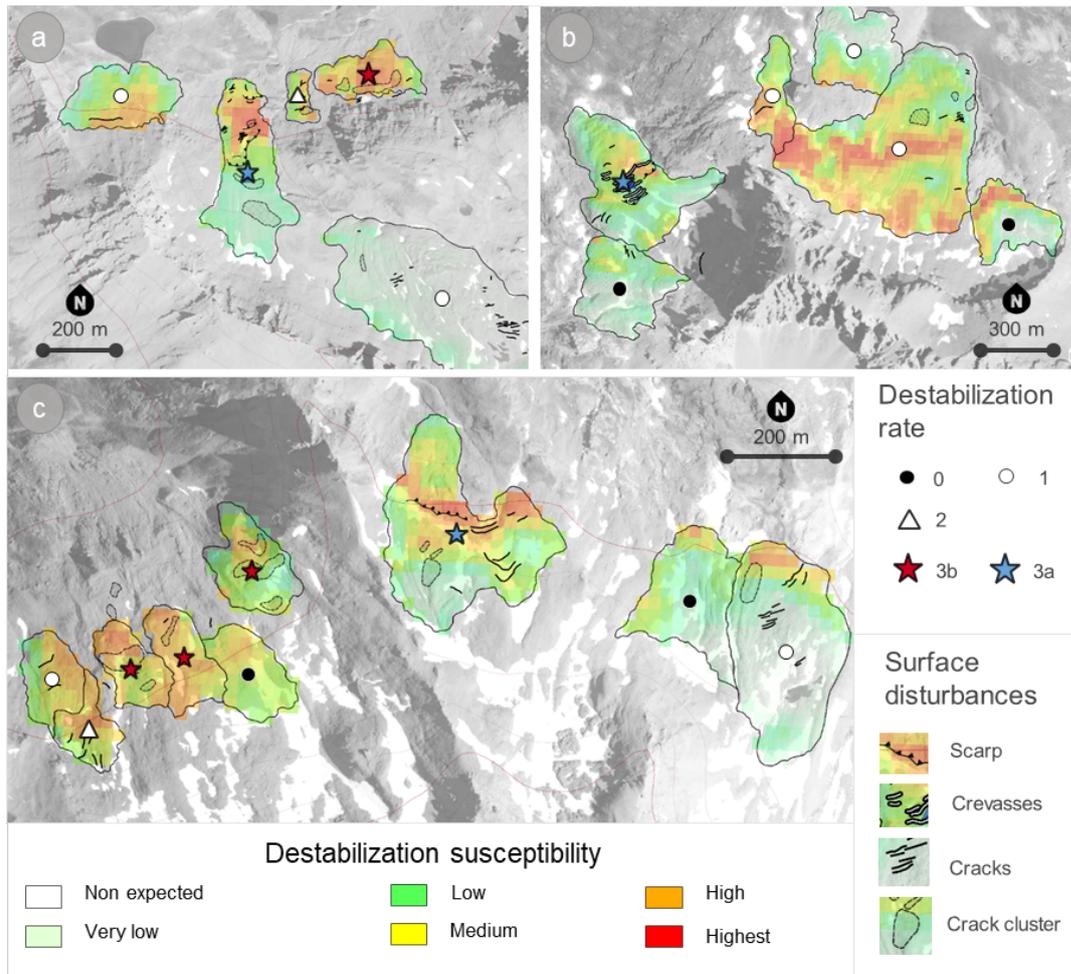


FIGURE 4.4: Examples of destabilization susceptibility map.

can register strong accelerations. Landforms presenting lower rates of destabilization should be considered as rock glaciers that did not present convincing destabilization evidence in the period 2000 – 2013. Nevertheless, it is possible that these rock glaciers may develop destabilization later, as it has been observed that surface disturbances may be present on rock glaciers for decades before destabilization is triggered. In these landforms the susceptibility map becomes handy as it identifies areas susceptible to destabilization. Rock glaciers with high susceptibility to destabilization should be regarded as possible spots of incoming destabilization and their state should be kept under some degree of monitoring. Monitoring effort may be limited to remote sensing optical images interpretation in order to spot geomorphological evolution towards instability.

These products may be useful the context of the periglacial hazard assessment of the project POIA-PERMARISK. It is reminded that both the destabilization rating and susceptibility map target only the sources of potential mass movements, while it is the connectivity between rock glacier and watershed to be a key point discriminating the occurrence of hazardous mass movements (Kummert and Delaloye, 2018). In case of strong connectivity, potentially destabilized rock glaciers

may present sudden acceleration and furnish abnormal quantities of sediments to the downstream torrential channel, increasing the watershed susceptibility to debris flows. In these areas, the suitability of protection infrastructures should be assessed taking into account the modifications to the sediment budget caused by the landform destabilization. Another process linked to destabilization may be rock glacier collapse. At the present state, evidences seem to suggest that this eventuality is remote, as the Bérard collapse was the only one case was registered in the European Alps (Bodin et al., 2016) and that others rock glaciers believed to be on the edge of total failure eventually stabilized (Delaloye et al., 2013). Nevertheless, it is suggested that stakeholders should keep in mind that this is an eventuality that may generate mass movements 1 to 2 orders of magnitude higher than debris flows and protection infrastructures are most likely inadequate to bear these kind of events.

4.3.1 Further work

Further efforts should target the sources of uncertainty related to the approaches used. The first source of uncertainty was due to the significance of cracks and cracks clusters in pebbly rock glacier destabilization. These landforms are still sparsely documented, while most of the previous studies described deep crevasses and scarps as source and evidence of the failure process. Therefore, there is still some uncertainty on the destabilization mechanisms of pebbly rock glaciers and the significance of surface cracks. It is therefore highly recommended to better investigate, by a study site approach, these landforms, their morphology and their evolution in relation to the destabilization process. This knowledge is needed to better understand the relevance of the destabilization ratings and susceptibility map produced by this study. In this sense, the study case presented in chapter 5, contributes to this issue as the Lou rock glacier has characteristics typical of these landforms to be yet fully understood.

Another source of uncertainty was due to the personal biases in attributing the destabilization rating. Personal metrics in the rating attribution may vary as it was remarked that different operators did not agree on several cases. Integrating new orthomosaics could allow to better validate or disprove the rating proposed in this study. Orthomosaics are periodically updated by the IGN with the most recent photogrammetric campaigns. It is therefore recommended to periodically update the destabilization rating using the newest data available.

Finally, further efforts should address the validity of the susceptibility map. This tool indicates the best candidates to experience this process under the hypothesis that in the future the occurrence of this process may increase. This hypothesis should be the target of future work. In this sense, it is recommended to monitor the future evolution of rock glaciers showing lower rates of destabilization but presenting high susceptibility to destabilization. Destabilization development or persistent stability of these areas may validate or disprove the utility of the map.

Chapter 5

Study case: rock glacier failure and realized risk

In this last issue it will be illustrated a case of mass movement generated on a destabilized rock glacier in the French Alps that caused damages to human infrastructures. The main goal is to provide a reference study in the region on this issue by investigating the interaction between rock glacier, destabilization occurrence and mass movements. Also, it is investigated the role of the protection infrastructures in the event trying to understand how they failed in securing the town from the mass movement.

5.1 Context and objectives

The chosen site is the Lou rock glacier, an active landform located in the Mont Cenis range, in Savoie (Figure 5.1). The Lou rock glacier is part of the Arcelle Neuve watershed, a tributary of the Arc river that springs on the North face of the *Grand Signal du Mont Cenis*. The larger section of the rock glacier consist of a flat plateau at 2800 m.a.s.l. and is supplied in sediments by nearby densely jointed schist rock-walls. The front of the rock glacier, which outflows from the plateau, can be subdivided in three main lobes, surrounding the North slopes of the torrential channel of the Arcelle Neuve stream. These lobes, called here western, central and eastern, are characterized by a high steepness and a direct connection to the torrential gully of the Arcelle Neuve stream. The western slope is characterized by destabilization evidence, i.e. intense fracturing of its surface.

The 14th August 2015 two slope failures at the rock glacier front were triggered after several days of rainstorm. The failures were located on the sides of eastern and western lobes and oriented towards the torrential channel. The failures consisted on a slide of the active layer on the permafrost surface, creating a smooth shear plane with cemented permafrost at its bottom. A consistent amount of water was observed to spring out of the head of the failure right after the event. Once triggered, the mass movement sled to the Arcelle Neuve stream and, thanks to the large availability of loose sediments in the torrential bed, developed a debris flow that reached the town of Lanslevillard, located at the confluence with the Arc river (Figure 5.2). Although

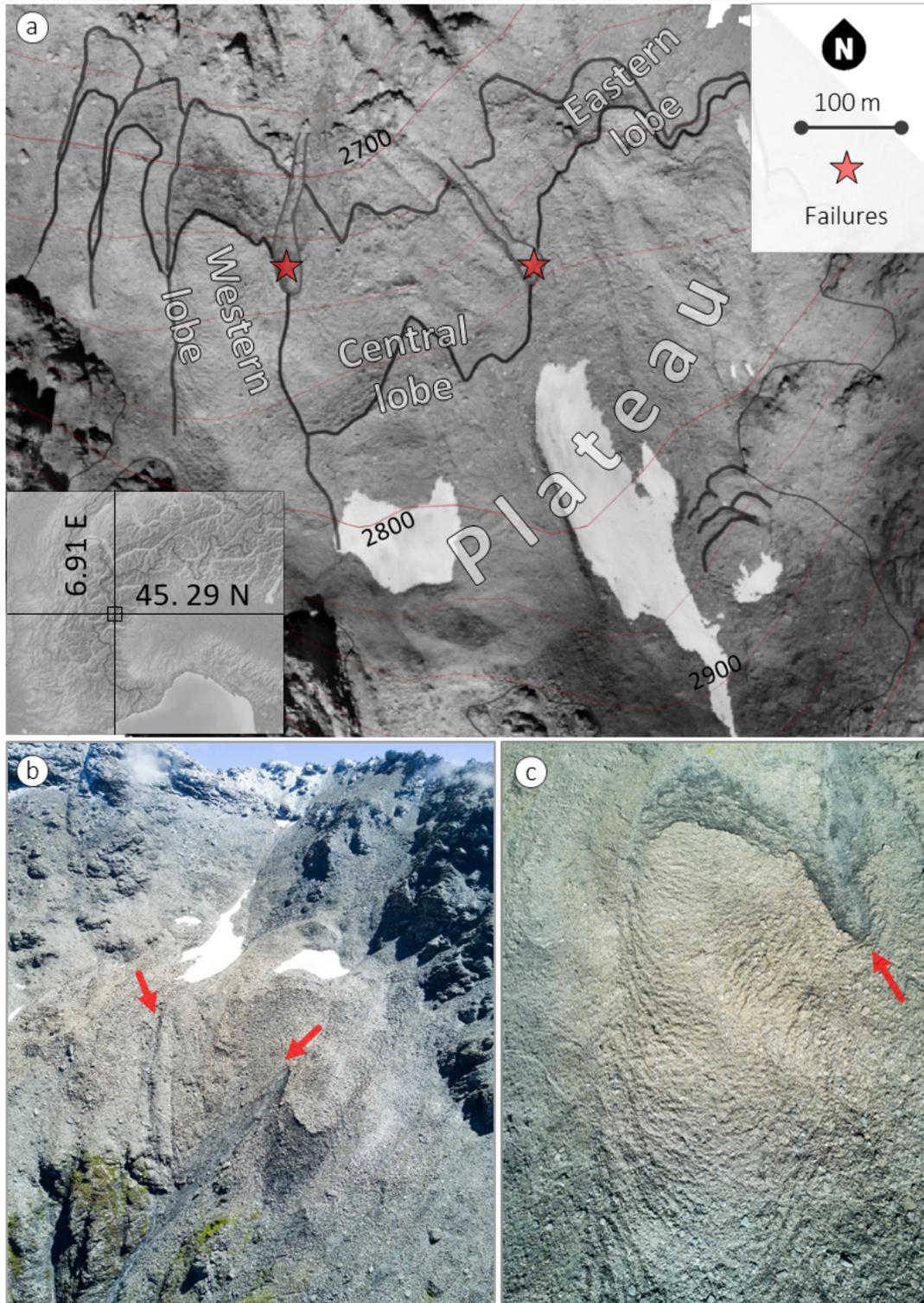


FIGURE 5.1: Presentation of the Lou rock glacier. In (a) overview of the landform on orthoimage. On bottom aerial images of the entire landform from South (b) and detail of the cracks on the western lobe (c). Red arrows indicate locations of the failures involved in the 2015 event.

the Arcelle Neuve watershed was known to produce debris flows, the river infrastructures resulted inadequate to contain this debris flow, causing the flooding of a

part of the town. The volume that reached the town was estimated to be 15 000 m³, causing several thousand hundred euros worth of damages.



FIGURE 5.2: Overview of the debris flow stages. In (a) is presented the Arcelle Neuve stream, connecting the Lou rock glacier (1) and Lanslevillard (2). In (b) is illustrated the flooded area. In (c) is shown an excavator clearing the area from debris. In (d) is shown a pipe that channels the stream below the town before the confluence.

This event presents relevant characteristics in the context of the PERMARISK project as (i) the debris flow trigger was located on a rock glacier showing destabilization evidence and (ii) the infrastructure was inefficient to face the mass movement. This event therefore poses two main research questions. The first question is to understand what is the role of the rock glacier destabilization in the event. Although the failures were triggered by a meteorological event, it is investigated if the destabilization contributed to the failure. The second question concerns the river infrastructures that failed in containing the debris flow. Since channelization was designed acknowledging the predisposition of the Arcelle Neuve stream to debris

flows occurrence, it is questioned in which way the infrastructures were inadequate to this specific debris flow.

This study embeds the efforts of several members of the POIA – PERMARISK project. Ribeyre (2016) evaluated the historical movements of the landforms by orthoimagery interpretation. RTM (2016) analysed the hydraulic aspects of the debris flow, focusing on the failure of the channelization system. Schoeneich et al. (2017) presented a synthetic report of the rock glacier characteristics based on field measures made after debris flow, involving geomorphological interpretation, dGPS and ERT. Nielsen (2018) contributed to the understanding of the geophysical aspects of the landform. The present study gathered the results of these works, as well as providing new investigations, into one comprehensive diagnostic of the 2015 event.

5.2 Methodology

Although the Lou rock glacier was already part of the RTM rock glacier inventory and known to the authorities, it was not under monitoring at the date of the event. The first survey was performed after the event in late August 2015 and the landform is periodically surveyed ever since. Therefore, there are no site specific data available before the slope failures. The analysis was performed by combining historical data not specific for the site and in situ data acquired since 2015. The analysis was based on the hypothesis that the mass movement was caused by a combination of meteorological trigger and site predisposition to failure, which relate to topographical settings, creeping characteristics and internal structure of the site. These elements were investigated individually using morphological analysis, dynamical investigations and meteorological data assessment. Ultimately, the adequateness of the river protection infrastructures was investigated by analysing the hydraulic assessment of the watershed realized by RTM (2016).

Rock glacier and failures morphology Investigating the rock glacier morphology can give an insight on the natural predisposition of the site to failure. The analysis was carried on different levels of complexity. At first, in situ topographical, lithological and geomorphological observations were performed aiming to understand general characteristics of the landform as front steepness. Particular attention was paid to analyse the two slope failures and their characteristics. The observations were also supported by the use of a high resolution DEM obtained by UAV photogrammetric survey. Aerial images, acquired in September 2016 by a DJI Phantom II, were used to compute the 3-dimensional structure of the site using the traditional Structure from Motion approach (Smith et al., 2016) in the software Agisoft Photoscan. This allowed to obtain a DEM of centimetric resolution, which allowed detailed observations and morphometric measurements of the rock glacier characteristics.

In a second time, several geophysical investigations were carried, involving ground resistivity measurements by ERT and p-wave velocity measurements by SRT (Nielsen,

2018). These methods allowed to obtain an important insight on the internal structure of the landform, involving permafrost distribution and ice content. Measurements were conducted in September 2016 and October 2018. The 2016 campaign focused on the rock glacier plateau and involved two ERT profiles. The 2018 focused on the western lobe were two ERT profiles and one SRT profile were performed. ERT data were inverted using the Res2Dinv software, while SRT data were inverted using ReflexW.

Creep characteristics The creep characteristics of the rock glacier were investigated by analyzing (i) the historical evolution of the landform and (ii) its current creep patterns and rates. Knowing the historical evolution of the landform velocity in the past decades is useful to characterize the rock glacier response to the atmospheric warming. The historical evolution of the creeping rates were evaluated by observing surface movements on four orthoimages from 1970 to 2017. Orthoimages were computed using aerial images triangulation in PCI Geomatica and Agisoft Photoscan (Ribeyre, 2016, following the method proposed by Kääb and Vollmer, 2000). Boulders on the rock glacier surface identifiable on the orthoimages were used as remarkable features and tracked in the different frames to compute an averaged velocity between the two frame's dates (Scotti et al., 2016).

The current creeping pattern of the landform was investigated by UAV imagery acquired in August 2017 and August 2018 using a DJI Mavic Pro. Images were treated in order to produce high resolution orthoimages following the SfM approach in Agisoft Photoscan (Smith et al., 2016; Dall'Asta et al., 2017). High precision dGPS coordinates of Ground Control Points were acquired prior the UAV surveys to reduce the distortions in the models (Smith et al., 2016). By repeating the measure in the same period year and at the same time of the day at one year distance, shadows generated by terrain features on orthoimages were similar. This allowed to perform an automated feature tracking of movements on the orthoimages using the SAGA GIS module IMCORR. The result was a map of creeping rates, allowing to observe the spatial distribution of the creep pattern.

Meteo-climate analysis Meteorological events were reported by several studies as the trigger of permafrost failures (Lugon and Stoffel, 2010; Springman et al., 2012; Kummert et al., 2017). Kummert et al. (2017) suggested that failures may not be triggered by an extremely intense rainy event by rather by a particular meteorological sequence that is specific to each landform. It is therefore important to understand what meteorological sequence caused the frontal failures at the Lou rock glacier. Using the MeteoFrance weather station network in proximity of the site, snow cover, temperature and precipitation data are analysed. The data from the 2014-2015 season were compared to the available record, which started in 1992, in order to understand the characteristics of meteorological events that triggered the failures.

River infrastructure analysis The Arcelle Neuve channelization system was entirely described by RTM (2016) using field survey and analysing original projects of infrastructures. The analysis focused on finding the weak points of the system as well as proposing a new infrastructure to protect the town from future events. The document provided by the RTM was here inspected in order to understand mainly two points. At first it was investigated which infrastructure failed in preventing the flooding of the town. Second, it was investigated how this infrastructure was designed, in order to understand what caused the infrastructure failure. In particular, it was questioned if the volumes involved in the 2015 event were unexpected or if the infrastructure was poorly designed and doomed to failure in any debris flow event.

5.3 Results

Rock glacier morphology and creep The geomorphological observations of the rock glacier highlighted the natural predisposition of the site to failure. The frontal slopes reach 40° , directly connected to the the Arcelle Neuve stream. The rock glacier lithology consists of black schist and, although some large blocks are present, the debris diameter generally ranges between from few centimetres to decimetres. These fine debris are unstable and small slides could be triggered just by walking on the rock glacier front.

The rock glacier structure was found to be complex, due to glacial (on the East) and periglacial (on the West) processes, characterised by substantial variability in permafrost ice content distribution. Periglacial areas had low ice content, possibly saturated in water. The western lobe presented evidences of ongoing destabilisation, as the lobe's surface is covered by several shallows tension cracks every 2-3 meters. Historical imagery revealed a strong acceleration of this area, as it were creeping at few dm/y between 1970 and 1996 and reached 3.7 m/y in the past two years (Figure 5.3). This acceleration occurred in parallel to a significant front advance of about 20 meters, observable between 2006 and 2012 (Figure 5.4). After 2012, the front edge location stabilized until nowadays.

Areas influenced by glacial presence, possibly until the LIA, presented a concave morphology separated by the rest of the landform by a lateral moraine. Here geophysical investigations suggested high ice content. This part of the landform is not destabilized as surface disturbances are absent and creep rates and accelerations are milder than the western lobe. The eastern lobe was found to creep at about 1.5 m/y between 2017 and 2018 and historical orthoimages suggested that also on this sector there was significant acceleration although as not intense as on the western lobe (Figure 5.3).

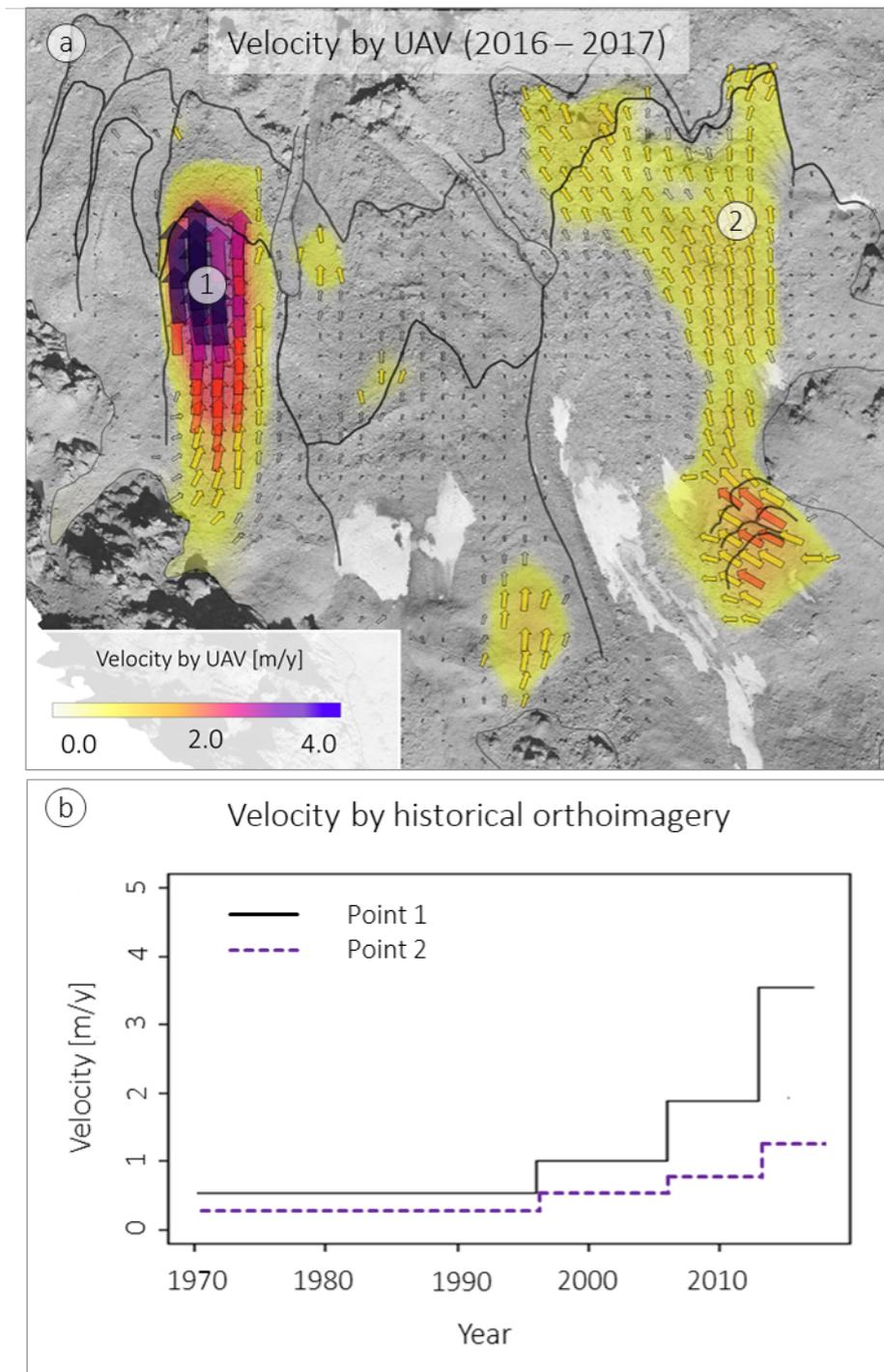


FIGURE 5.3: Creep pattern (a) between 2017 and 2018 and velocity evolution of the western and eastern lobes since 1970 (b)

Failures morphology The two failures were located in two areas of the rock glacier presenting different characteristics. The eastern failure, located on the lobe influenced by glacial processes, could be observed on historical images since the 50s, suggesting that the event occurred in 2015 was not novel. The western failure, located on the edge of the destabilized west lobe and influenced by periglacial processes, on the other hand could not be observed in historical imagery indicating that

the 2015 event was a first time failure. This failure was located on an area that was occupied only recently by the rock glacier due to the front advance between 2006 and 2012 (Figure 5.4).

Both of the failures were observed to be pouring water from the top of the permafrost table right after the event. The shape and the depth of the slides indicated that the failure affected only the active layer of the front. The failures therefore were most probably caused by the supra – permafrost water flow that eroded the unfrozen debris of the active layer. The water flow was probably due to an excess in soil saturation and water collected by the rock glacier catchment during a rainy sequence. Supra-permafrost flow may have been particularly intense in the spots where the failures were triggered, probably due to sub-surface hydrological paths. This process of concentrated flow (Kummert et al., 2017), was observed in several rock glaciers in Switzerland and identified as trigger of debris flows when the rock glacier front is connected to a steep torrential gully.

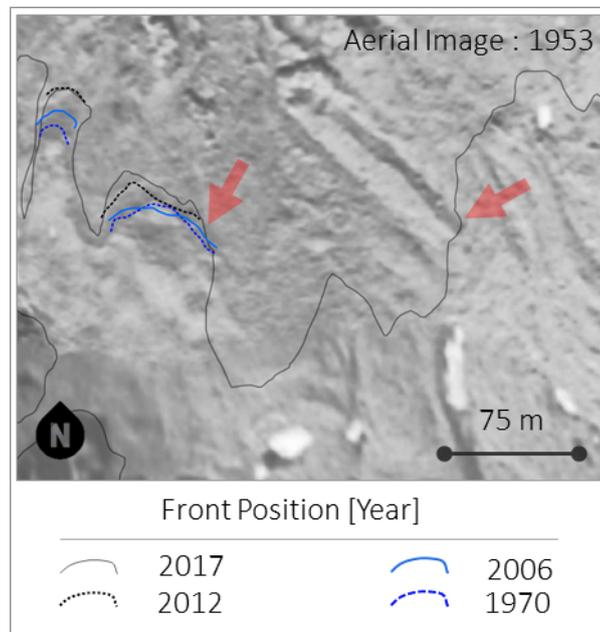


FIGURE 5.4: Evolution of the frontal position since 1970. Red arrows indicate the location of the western and eastern failures. It can be observed that the eastern failures was active already in 1953, while the western failure occurred on a spot only recently occupied by the rock glacier front.

Meteorological preparation and trigger Precipitation data indicated that the winter 2014 -2015 was particularly dry, with a snow cover about 0.5 m thinner than the average. Snow data suggested that the melt season ended one month before than usual. This indicated that the rock glacier was snow free by August, as observed in situ few days after the event. The summer 2015 was remarkable for a strong heat wave that lasted from mid-June to mid-July, reaching temperatures 7 °C higher than

seasonal means and setting the record for warmest period on record. After a complete absence of precipitations during the heat wave, the late July to August period was characterized by several precipitations events. In this period of three weeks were registered at least seven relevant precipitation events with an intensity ranging from 10 to 30 mm/day, for a cumulative rainfall of about 200 mm. Compared to previous years, this period was the wettest three-weeks summer period in the record. The failures were ultimately triggered by a rainfall of about 30 mm/day, a relatively non-intense value as precipitations above 60 mm/day were commonly recorded.

Inadequate infrastructures The weak point of the channelization of the Arcelle Neuve was identified in a pipe of 2 m diameter that channels the stream below the town for 130 m and allows discharge directly into the Arc river (Figure 5.2d). The diameter of the pipe was considered adequate by RTM (2016) for a 100 y return period event. The failure was due to a grid located at the entrance of the pipe, possibly to avoid gradual pipe jamming below the town where large debris removal may have been difficult. This grid was obstructed by the solid material transported by the 2015 debris flow, resulting in the failure of the pipe purpose and consequent flood. The failure of the pipe system is therefore most probably due the presence of this grid rather than the volume of the debris flow.

5.3.1 Diagnostic

The study indicated that the 2015 event was caused by several factors. The Lou rock glaciers had a natural predisposition to generate mass movements, due to its topographical and lithological characteristics. The east lobe was likely already subjected in the past to erosion processes similar to the 2015 failure, as the east slide could be observed to be active in historical imagery. The west slide on the other hand, was a novel process. The dynamical and geomorphological observations revealed that the western lobe is experiencing destabilisation, observable by strong acceleration, surface fracturing and frontal advancement observed between 2006 and 2012. Considering that the west slide occurred in a zone that the rock glacier only recently occupied, it is suggested that rock glacier destabilization caused significant geomorphological changes on the west lobe that increased its susceptibility to failure.

The failures occurred after a period of prolonged heat wave between mid-June and mid-July and gradual meteoric water loading since the end of July. Water collected by the rock glacier catchment flowed downstream above the permafrost table and caused the two failures in two spots where the flow was more concentrated, i.e. concentrated flow process (Kummert et al., 2017). The fact that the failures were ultimately triggered by a mild rainstorm, suggested that the meteorological sequence prior to the event was a fundamental preparatory factor.

Although the debris flow was initiated by the rock glacier failures, the Arcelle Neuve stream added a large amount of debris to the flow, possible up to ten times

the initial slides volume. This suggests that there was a large availability of loose materials in the channels and/or that the flow had a strong erosion power. Loose materials are believed to be in part due to the highly jointed lithology of the bedrock surroundings the channels below the rock glacier. Nevertheless, the rock glacier possibly has a relevant role in charging the channels by conveying sediments downslope thanks to its high dynamics and steep front connected to the torrential gully.

Ultimately, the flood of the town has to be imputed to a pipe that channels the Arcelle Neuve stream under the town into the Arc river. The grid protecting the pipe was inadequate with respect to the solid transport of the debris flow causing the jamming and consequently the overflow. In this sense, the pipe is suggested as inadequate not only specifically with respect to the 2015 event but also to any debris flow involving consistent solid transport. It is unclear the reason why such infrastructure was built in a sensitive area where debris flow were not unexpected. The presence of a grid protecting the pipe entrance acknowledged the possibility of solid transport as grid obstruction would allow an easier fixing of the infrastructure preventing large debris or trees penetrate deep into the pipe. It is possible that this practice was motivated by an underestimation of the eventuality of complete obstruction and consequent outflow.

5.4 Conclusions

The investigations at the Lou rock glacier revealed some important elements that caused the frontal failures of the landform. The failures were triggered only by a mild rainy event, suggesting that climate prior to the event had a significant, if not determinant, preparatory role. Several climatic anomalies were detected in the months prior to the event and consisting of the strongest heat wave since the 90s, a particularly dry winter and a gradual and continuous loading by mild precipitations during two weeks prior to the failures. To which extent these meteorological characteristic influenced the failure occurrence is still unclear and deserves further attention.

The site had a natural predisposition to slides, due to its frontal steepness, schist lithology and connection to a steep torrential gully. As proof, it was observed on historical aerial imagery that the eastern failure was not a novel event. Nevertheless, the western failure was indeed a novel event and suggested to be linked to the western lobe destabilization. The destabilisation process is suggested to have caused significant morphological modifications on the lobe, as frontal advance, surface fracturing favorising water infiltration and sediment deconsolidation. These processes may have significantly preconditioned the failure occurrence. In this sense, it is suggested that rock glacier destabilization is a phenomenon that may increase the site susceptibility to failure.

The town flooding was ultimately the result of underestimation of the natural hazard represented by the Arcelle Neuve stream. Although the event itself was not

major, improper river structure design caused the flooding and damages. In order to ensure the protection of the town, RTM (2016) proposed the construction of a channelling infrastructure through the town. The storage volume of the channel was evaluated on the basis of the 2015 event (15 000 m³) plus a volume due to exceptional contribution of the erodible areas of the watershed (10 000 m³ assuming a 100 y return time). The total volume estimated was 25 000 m³. The adequacy of this volume will depend upon the future occurrence of frontal failures of the Lou rock glacier. This parameter represents a great source of uncertainty as it is unknown if the 2015 event was exceptional or the warning of an incoming major crisis of the landform. Nevertheless, it is highlighted that the rock glacier volumetric contribution to the 2015 debris flow was about 7 - 10%, suggesting that the magnitude of the debris flow mainly depends upon the available sediments in the Arcelle Neuve stream.

5.4.1 Further work

The 2015 debris flow at Lanslevillard represents a symbolic event in the project POIA-PERMARISK. The Lou rock glacier is an interesting permafrost site with natural predisposition to mass movements and ongoing destabilization that should become a reference site in the PermaFrance network for similar processes. It is therefore suggested to perform yearly surveys aimed to monitor the evolution of the displacement rates, surface morphology and thermal properties of the rock glacier. While a combination of UAV and dGPS methods seem to be adequate to monitor the rock glacier surface evolution, thermal characteristics can be investigated using permanent temperature loggers. At this regard, in 2015 four GST temperature loggers were installed to register continuously ground surface temperature in different sectors of the landform. These data may become relevant to understand the role of climatic preparation in a case of a new failure.

In order to ensure continuous monitoring of the frontal activity of the rock glacier (Kummert et al., 2017), it is also recommended to install a permanent camera. Permanent cameras can be useful to better understand the processes of torrential gully recharge, frontal erosion and location of water outflows (Kummert et al., 2017). Permanent cameras can be also connected to a wireless data stream for real time monitoring. This system allows to set up a basic alert system in case of upcoming failure. The major challenge of this action may be to find an appropriate view point to install the system. It has to be considered that the site is avalanche prone and several gazex are installed on the headwalls surrounding the rock glacier. Therefore the site is exposed to avalanches in winter that may damage the camera installation.

Finally, it is suggested to better investigate destabilization mechanisms of the western lobe. The Lou rock glacier is a case of destabilized pebbly rock glacier were

surface tension cracks are the geomorphological evidence of the destabilization process. As explained in section 4.1, there is still some uncertainty concerning the significance of these geomorphological features in the context of rock glacier destabilization. As similar features have been identified by the present PhD in several active landforms across the region, a better understanding of these landforms is therefore required, as specified in section 4.3.1. In this context, the Lou rock glacier can represent a relevant study site where further efforts should focus on understanding the destabilization process. In particular, it is suggested to investigate the motion mechanism of the destabilized lobe by performing borehole inclinometry (Arenson et al., 2002). This operation may allow to understand whether the west lobe is experiencing basal sliding or enhanced creep. Borehole excavation can also give an insight on the internal structure of the landform, revealing the presence or absence of water layers that may be connected to the destabilization occurrence (Eriksen et al., 2018).

Chapter 6

Conclusion

In this last chapter are drawn the conclusion of this PhD work. At first, it is proposed a summary of the main findings that emerged from this study and what is their relevance in the context of the project POIA-PERMARISK. Then, it is proposed a description of the main questions that rose from this work as well as future efforts needed to ameliorate our knowledge on these subjects.

6.1 Summary of the results

The present work aimed to obtain an insight on the permafrost occurrence and rock glacier destabilization in the French Alps, issues that were still to be fully developed. The contribution of study to the knowledge of these issues can be summarized into five main findings:

1. Permafrost is a common feature in the French Alps

In the French Alps the periglacial environment is developed and rock glaciers are common landforms. More than 500 rock glaciers are active. Permafrost distribution modelling highlighted the widespread occurrence of mountain permafrost in the region, covering about 770 km². Permafrost distribution was observed to be related to regional scale variability, possibly due to different characteristics of the precipitation patterns and their seasonal intensity.

2. Rock glacier destabilization is a common phenomenon

Creeping permafrost destabilization was observed to be common phenomena in the region. More than half of active landforms presented geomorphological features that are commonly observed in destabilization processes. In total, 46 landforms presented evidences of potential destabilization, i.e. presenting evolving surface disturbances that seemed to be related to an abnormal creeping pattern. Nevertheless, most of these landforms were characterized by the presence of cracks only, a geomorphological feature that characterizes pebbly rock glaciers. Although already observed on some destabilized landforms, these features were found to be exceptionally abundant in the region and therefore it questioned their significance in the context of observing ongoing destabilization.

3. Terrain attributes precondition destabilization

Destabilization was observed to occur in particular topographic settings as steep, convex and north facing slopes. Rock glaciers located in sedimentary lithology were the only observed to experience destabilization. Destabilization was also observed to be more likely at the lower margins of the permafrost zone. As permafrost in this area is hypothesized to be more impacted by climate warming, it is suggested that rock glacier destabilization susceptibility may be enhanced by increasing mean annual temperatures.

4. High potential for future destabilization

Susceptibility modelling revealed that almost 2 km² of creeping permafrost have a high predisposition to destabilization but currently showing low destabilization rating. These sites should be regarded as possibly experiencing future destabilization if subjected to a proper trigger. This suggest that there is a significant amount of rock glaciers that may encounter destabilization in the future.

5. Rock glacier destabilization may precondition new mass movements

As emerged from the Lou rock glacier investigation, destabilization may cause morphological changes on the landform that can increase the predisposition of the site to mass movements. In case of strong connectivity with torrential channels, destabilized sites may trigger significant debris flows and river infrastructures may not be adequate to sustain the entity of these phenomena.

Overall, despite the uncertainties due to the novel methodologies applied, the results of the present study seem to indicate that rock glacier destabilisation is a relevant phenomenon in the region. Destabilization was identified on a large amount of rock glaciers and its occurrence is susceptible to increase. This process may raise the hazard level by causing significant morphological changes on the rock glaciers structure that may increase the susceptibility to failure. It is therefore concluded that rock glacier destabilization is an important issue that should deserve future investigations.

6.2 Open questions and future work

Through this PhD were tackled a wide variety of subjects in the contexts of permafrost distribution, rock glacier destabilization and related hazards. As often occurs in the scientific process, some questions were addressed but many more arose. This last section aims to point out the main voids left from this PhD and how it may be possible to fill them, a legacy addressed to the researchers that tomorrow will be eager to commit to this subject.

6.2.1 Periglacial risk assessment at the regional scale

The future effort with the highest priority in this context has to be considered the finalization of the rock glacier hazard assessment. As mentioned in section 2.4.2, this PhD should be acknowledged as a single step of the long process of characterizing rock glacier hazards. This study proposed an assessment of rock glacier destabilization, which may be useful in a second time to define the upslope potential hazards linked to these landforms. Nevertheless, the existence of a real hazard is controlled by the connectivity of the rock glacier with the sedimentary chain. Therefore, further work should focus in evaluating the connectivity of active rock glaciers fronts following the scheme proposed by Kummert and Delaloye (2018). It is also reminded here that in such topographic conditions all active rock glaciers may represent a source of hazard as they charge the torrential channel increasing debris flow susceptibility.

The tools developed in this study may ease the hazard assessment challenges related to the large extent of the region involved as analysis should be conducted at the regional scale. Nevertheless, it is important to understand the real meaning and limitations behind each tools before using it in this context. In this sense, it is strongly recommended constantly question the validity of the destabilization rating and susceptibility map by integrating in situ observations and new data (orthoimages, DEMs) in order to spot judgement errors or biases.

To complete the rock glacier hazard assessment, it is reminded to focus on the lower part of the sediment chain as well. Vulnerable areas, as roads and towns close to streams subjected to this periglacial activity should be inventoried. The suitability of these infrastructures to bear significant mass movements in the selected areas should be assessed in detail. In this sense, it may be relevant to estimate the propagation of these mass movements and their impact on the existing infrastructures. This can be done using GIS-based numerical modeling, i.e. the RAMMS model, to predict the temporal propagation of an hypothetical debris flow generated by an active layer failure on the front of a connected rock glacier (Hovgaard and Eisenbrückner, 2017). This method allows to propose flooding scenarios based on debris flow volumes. Nevertheless, this approach requires the knowledge of several mechanical and rheological properties of the debris flow which are scarcely reported in the periglacial context. If future efforts will commit to this method, a crucial point to tackle is therefore the definition of these properties, possibly by calibrating the model on previous known events.

6.2.2 Methodological advancements

The proposed models are based on strictly empirical methodologies. This causes limitations in the predictive power in future contexts of climate changes, as the models are stationary while climate is not. The introduction of numerical simulations of physical processes in the modeling phase could help to predict ground response to

climatic variations (e.g. Westermann et al., 2013). Physically-based models can be based on thermal simulation of the ground temperature, as well as of dynamical behaviour of creeping permafrost. Nevertheless, it has to be taken into account that at the present state of the art models require large amounts of site specific data. It is therefore recommended to carefully evaluate the data availability before committing to this modelling approach.

6.2.3 Open questions on rock glacier destabilization

Understanding the processes that determine the occurrence of mass movements in the periglacial area is suggested to be a central point to be further developed in the future. This study revealed several questions that should be addressed by future efforts in this field.

Interactions between short-term climatic events and rock glacier failures The first question concerns the role of the seasonal climate on the stability of the landform. The influence of rock glacier stability of smaller scale events, as heat waves, was observed but still unclear. The Lou frontal failures for example occurred in a context of scarce snow cover, abnormally warm weather and persistent precipitation. It is not yet clear whether one of these processes was more relevant than the others or if the concomitance of all them caused a suitable condition to trigger the failure. It is also unclear in which way these processes may have prepared the site to failure. In this sense, it is recommended to persist monitoring efforts on this site and integrate it in the PermaFrance network.

Role of climate warming trend in destabilization triggering The second question concerns the main destabilization triggering. Rock glacier destabilization is a process that may be triggered either by mechanical or climatic forcing (Delaloye et al., 2013). The relevance of the climatic forcing leads to the question whether a further temperature rise may cause generalized destabilization triggering at the regional scale. This question is quite interesting as it concerns the future occurrence of rock glacier destabilization. To give an insight to this issue it is suggested to investigate the rock glacier creeping rate evolution in the past decades in relation to the destabilization rate in order to observe general pattern in destabilization onset timing. This can be done using the IGN orthoimagery covering the French Alps which, at the present state, counts 4 frames per site (missions from 1950 to 1965, mission from 2000 to 2004, mission from 2005 to 2009 and mission from 2012 to 2013). In some departments are available more frames and the number should increase in the future. By manually feature tracking boulders moving on the landforms surfaces, it is possible to estimate creeping rate evolution in the past decades.

This approach was used in an exploratory analysis on a reduced sample of rock glaciers, using the most recent orthoimages. The results, briefly presented in Figure 6.1, show a generalized marked increase in displacement rate in the late 2000's,

in agreement with general knowledge on rock glacier creeping behaviour (e.g. Delaloye et al., 2008b). It is interesting to notice that many potentially destabilized rock glaciers seem to have suffered a sharper increase in their creeping rate in the first decade of the 2000s with respect to landforms with lower destabilization rates. This exploratory result suggests that there is indeed an interesting pattern between creeping rate temporal variability and destabilization occurrence that may be worth investigating more into details.

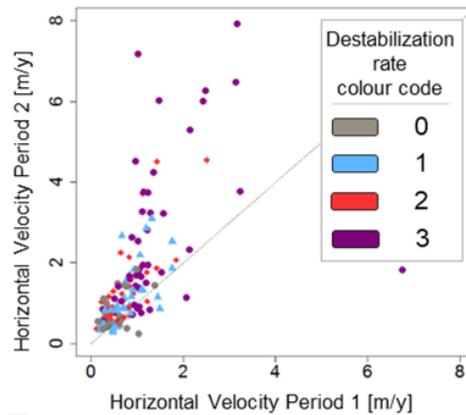


FIGURE 6.1: Fastest observable boulder velocity per rock glacier according to their destabilization rating in in the period 1 (2000 – 2004 to 2008 – 2009) and period 2 (2008 -2009 to 2012 -2013).

Significance of different surface disturbances on destabilization evolution The last question concerns the characteristics of the destabilization process in rock glaciers presenting different typologies of surface disturbances. As explained in section 4.2, surface cracks were found to be so common to raise question about their significance in the context of the destabilization process. The Lou rock glacier investigations highlighted that these features can be linked to important morphological changes on the landform, suggesting their relevance in this context. Nevertheless, only few of the destabilized rock glaciers showing these features were investigated by a field approach and uncertainties hold. It is therefore suggested to focus further efforts on this issue.

In this context, an action has already been launched as four destabilized rock glaciers are monitored by UAV survey since 2017 (Figure 6.2). These sites, located on the flanks of the [Grand Signal de l’Iseran](#), [Roc Noir](#), [Pointe du Grand Vallon](#) and [Tete du Longet](#) summits, were selected according to their different morphologies, representing different surface disturbances. The aim of this monitoring effort is to understand past and recent geomorphological and deformation pattern evolution in the past decades, combining observations based on historical imagery and yearly UAV surveys.

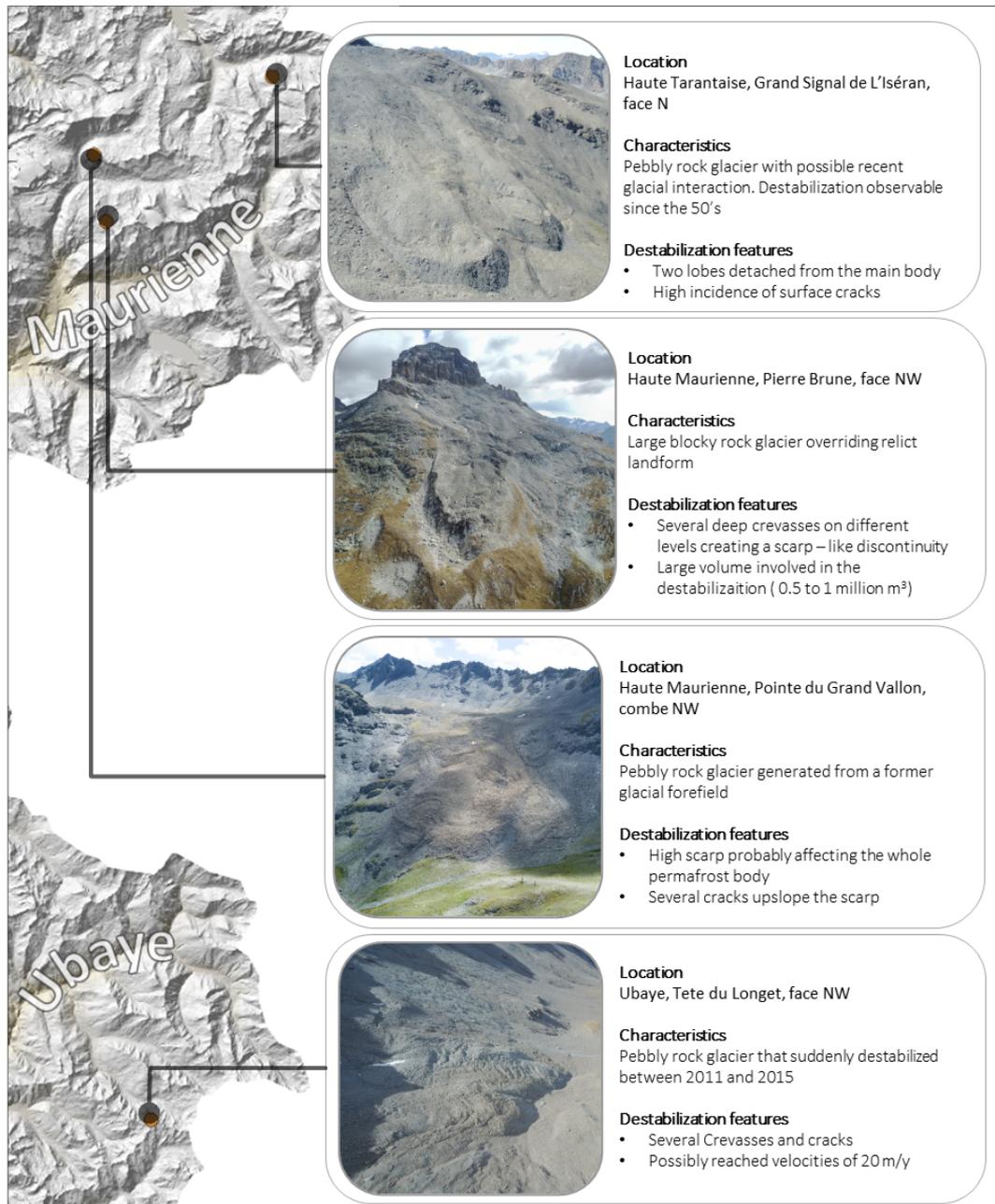


FIGURE 6.2: Localization and main characteristics of the destabilized rock glaciers that started to be monitored by yearly UAV surveys in summer 2017.

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Appendix A

Article I

Permafrost Favorability Index: Spatial Modeling in the French Alps Using a Rock Glacier Inventory

Citation :

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Main Findings

- Debris permafrost conditions can be found over 770 km² in the French Alps, above 2300 m.a.s.l. in the northern Alps and 2500 m.a.s.l. in the southern Alps.
- The regional North-South trend of the lower limits in permafrost distribution is suggested to be linked to precipitation patterns.
- Statistical approach is a consistent method for predicting permafrost distribution. Nevertheless, uncertainties in the modeling database used may significantly propagate in the results.



Permafrost Favorability Index: Spatial Modeling in the French Alps Using a Rock Glacier Inventory

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In the present study we used the first rock glacier inventory for the entire French Alps to model spatial permafrost distribution in the region. Climatic and topographic data evaluated at the rock glacier locations were used as predictor variables in a Generalized Linear Model. Model performances are strong, suggesting that, in agreement with several previous studies, this methodology is able to model accurately rock glacier distribution. A methodology to estimate model uncertainties is proposed, revealing that the subjectivity in the interpretation of rock glacier activity and contours may substantially bias the model. The model highlights a North-South trend in the regional pattern of permafrost distribution which is attributed to the climatic influences of the Atlantic and Mediterranean climates. Further analysis suggest that lower amounts of precipitation in the early winter and a thinner snow cover, as typically found in the Mediterranean area, could contribute to the existence of permafrost at higher temperatures compared to the Northern Alps. A comparison with the Alpine Permafrost Index Map (APIM) shows no major differences with our model, highlighting the very good predictive power of the APIM despite its tendency to slightly overestimate permafrost extension with respect to our database. The use of rock glaciers as indicators of permafrost existence despite their time response to climate change is discussed and an interpretation key is proposed in order to ensure the proper use of the model for research as well as for operational purposes.

Keywords: permafrost modeling, French Alps, mountain permafrost, rock glaciers, statistical modeling

INTRODUCTION

In the European Alps widespread evidence of ice-rich mountain permafrost existence has been observed, mostly intact rock glaciers (in the sense of Barsch, 1996, i.e., all the rock glaciers for which the presence of ice is expected) that can be found in almost all massives above 2,500 m (Baroni et al., 2004; Monnier, 2006; Ribolini and Fabre, 2006; Bodin, 2007; Cremonese et al., 2011). In addition, observations of accelerating and destabilizing rock glaciers are interpreted as potential signals of ice-rich permafrost degradation (Roer et al., 2008; Delaloye and Morard, 2011; Ramelli et al., 2011; Delaloye et al., 2013; Scotti et al., 2017). In the French Alps two recent cases of geomorphological phenomena linked to permafrost degradation received particular attention: the

collapse of the Bérard rock glacier in 2006 (Bodin et al., 2016), and the active layer detachment of the Lou rock glacier that resulted in a destructive debris flow, damaging the town of Lanslevillard in August 2015 (Schoeneich et al., 2017). In the present context of warming climate, such hazardous situations are expected to become more common as a consequence of the degradation of ice-rich permafrost (Haeberli et al., 1993; Bodin et al., 2015). For operational needs, knowing the spatial extent and the status of permafrost is therefore an extremely important issue (Haeberli et al., 2010).

Statistical models of permafrost distribution have been used for more than two decades in Switzerland (Keller, 1992; Imhof, 1996; Lambiel and Reynard, 2001; Frauenfelder, 2004). In France, a first model was built using a rock glacier inventory covering the massif du Combeynot (30 km², around 45°N) and locally validated using two independent datasets in the Vanoise and Mercantour massives (Bodin et al., 2008). More recently, as an international effort, the APIM (Alpine Permafrost Index Map) was conceived to estimate permafrost extent over the entire Alpine range (Boeckli et al., 2012a,b). The APIM was however calibrated using only a limited number of rock glaciers in the French Alps (only the Combeynot inventory, Cremonese et al., 2011) and, considering that many authors pointed out that permafrost distribution models tend to be site-specific and weak when transferred to others sites (Frauenfelder et al., 1998; Lambiel and Reynard, 2001; Baroni et al., 2004; Bonnaventure and Lewkowicz, 2008), its significance in this region still remains unknown. Therefore, a permafrost modeling effort focused on this region is necessary.

In France, Restauration des Terrains en Montagne (RTM), a division of the French National Forest Office (ONF) that is responsible for natural hazard management in mountainous regions, has become concerned about the potentially emerging hazards related to ice-rich permafrost degradation and rock glacier destabilization. A Geographical Information System (GIS) inventory of all rock glaciers of the French Alps was therefore created between 2009 and 2016, compiling information for more than 3000 landforms (Roudnitska et al., 2016). It represents a unique database for permafrost modeling in the region.

The present work aims to exploit this inventory to produce and evaluate a permafrost distribution model specifically designed for the French Alps. This is done following the two main assumptions recurrent in the permafrost distribution modeling literature (Boeckli et al., 2012a; Sattler et al., 2016; Azócar et al., 2017). First, the rock glaciers indicate the presence (absence) of permafrost if they are active (fossil). Second, the activity of the rock glacier is controlled by a set of topo-climatic proxies of the local climate at the rock glacier location. These two assumptions allow us to develop a statistical model, described in section Statistical Modeling, which describes the spatial distribution of permafrost, represented by the rock glaciers activity, using the spatial distribution of the topo-climatic proxy variables, which are described in section Data. The assessment of the modeling performance is described in section Uncertainty Estimation of the RGLs inventory. Results and results discussion are presented in sections Results and Discussion, including an interpretation

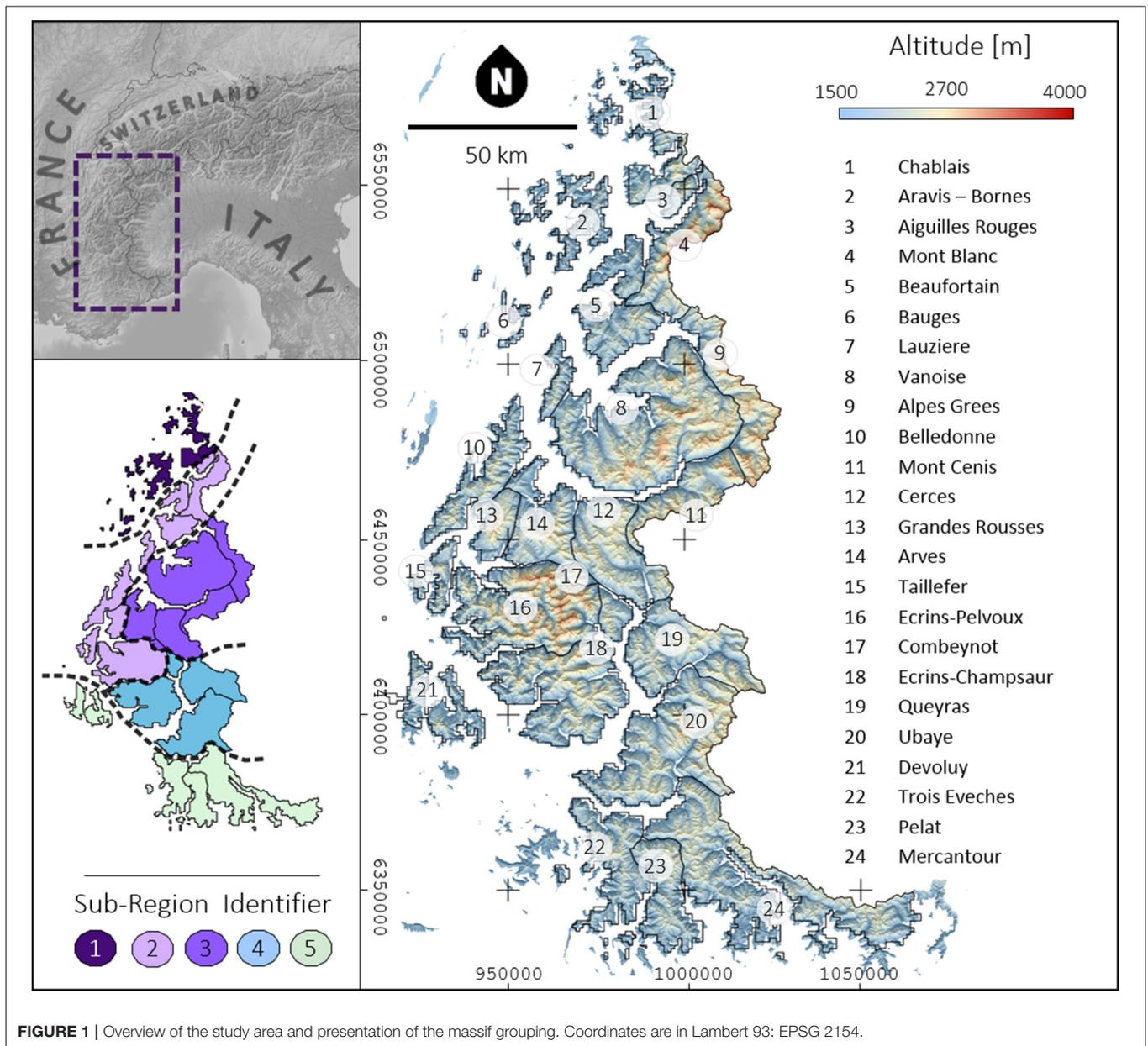
key for the model and its comparison with previous studies (section Model Interpretation).

STUDY AREA

In the present study, the French Alps are defined as the portion of the Alpine range within the French national borders. This mountain range forms a north-south oriented arc about 250 km long and 50–75 km wide (between 44 and 46°N and from 5.7 to 7.7°E) that hosts the highest summits of the entire Alpine Range. The climate is complex, being influenced by the Atlantic Ocean from northwest and by the Mediterranean Sea from south (Beniston, 2006). The climate of the French Alps is usually divided into southern and northern Alps at the geographical limit of the Col du Lautaret (45°N), where a climatic breakpoint is noticeable (Bénévent, 1926; Gottardi, 2009). The zero °C isotherm increases from ~2,200 m in the Chablais up to ~2,700 m in the Mercantour (Durand et al., 2009a). Precipitation patterns are even more variable than temperature, being influenced by atmospheric circulation patterns and topographic sheltering effects. The north-western Alps are remarkably wetter than the rest of the French Alpine range, being directly exposed to the Atlantic perturbations, while the Mediterranean influence determines a drier climate in the south. A topographic sheltering in the interior valleys is also observable, due to the high elevations of the massives directly facing the oceanic air masses (Durand et al., 2009b).

In order to analyse this vast area with its heterogeneous climatic, geological and geomorphological settings, we divided it into five sub-regions (Figures 1, 2 and Table 1) that group the areas that present similar characteristics with respect to the periglacial environment (Kenner and Magnusson, 2017). Sectors 1, 2, and 3 comprise the Western and Eastern side of the Northern Alps. In sub-region 2, where the highest peaks of the French Alps are located, there is a significant degree of glacierization (due to both high elevation and high precipitation), which leaves little space for the development of rock glaciers. Sub-region 1 is also exposed to this climate but, due to the substantially lower elevations hardly reaching 2,500 m a.s.l., there are only small glaciers. Sub-region 3 is sheltered from this influence, resulting in a drier and more continental climate, where periglacial activity flourishes (Guodong and Dramis, 1992; Harris and Corte, 1992; Monnier, 2006). These three sub-regions present also a different lithology. The sub-region 2 is the external crystalline range, dominated by granitic and metamorphic rocks. Sub-region 3 is characterized by an inhomogeneous geological structure dominated by schist, micashists, and ophiolites and Sub-region 1 is composed by low altitude limestones and sandstones.

The southern French Alps present lower elevations and under a strong Mediterranean influence (Gottardi, 2009), resulting in a drier and warmer climate with a low degree of glacierization. Similar to the northern Alps, the southern Alps can be divided into northwest and southeast sub-regions. The eastern sub-region (number 5) hardly reaches 3,000 m a.s.l. and limestone dominates the lithology, with the notable exception of the significantly higher, crystalline massif du Mercantour. The

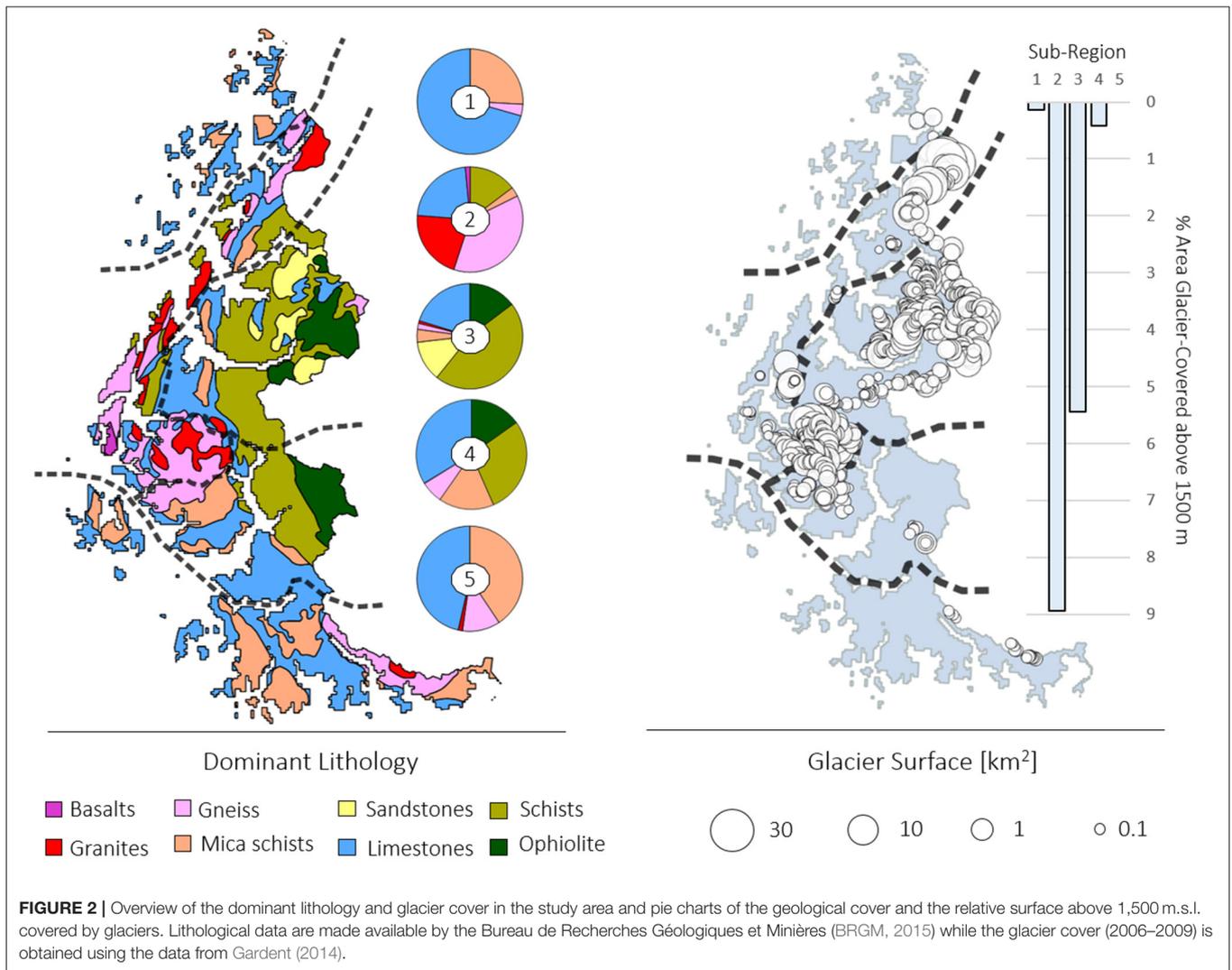


western sub-region (number 4) reaches 3,500 m a.s.l. and it is dominated by micashists and ophiolites.

METHODS AND DATA

Permafrost distribution was modeled using the statistical approach proposed already by several authors (Keller, 1992; Imhof, 1996; Frauenfelder et al., 1998; Lambiel and Reynard, 2001; Bodin et al., 2008; Deluigi and Lambiel, 2011; Boeckli et al., 2012a; Sattler et al., 2016; Azócar et al., 2017). Using the French Alps rock glaciers inventory (Roudnitska et al., 2016), we first hypothesized that the rock glaciers, here treated as dependent variable, are indirect indicators of the presence and absence of permafrost according to their activity. The second

hypothesis is that the conditions for permafrost existence can be discriminated by a set of local topoclimatic controls, here treated as predictor variables, such as present-day air temperature, incoming solar radiation and precipitation patterns that were sampled from raster maps at each rock glacier’s location. The authors acknowledge that permafrost presents a delayed signal with respect to the climate and it is assumed that the present day climatic patterns are representative of climatic patterns of the past several decades. A statistical model was then trained to evaluate the correlation between the permafrost existence indicators, called the response variable, and the topoclimatic controls, called predictor variables. In using a model of rock glacier activity status to predict permafrost in areas outside rock glaciers, it is assumed that the same relationships that apply



to permafrost presence / absence within rock glaciers hold in surrounding terrain.

It is emphasized that fitting a statistical model based on rock glacier activity status does not model the actual permafrost distribution, but rather the likelihood of the occurrence of the complex processes associated to the rock glaciers existence and activity, which are an expression of permafrost presence. Therefore here we will use the term “Permafrost Favorability Index” (PFI) instead of permafrost distribution to express this likelihood (Azócar et al., 2017).

Statistical Modeling

In the present study statistical modeling was performed using a generalized linear model (GLM) with logistic link function. It has been demonstrated that the GLM is suitable for modeling permafrost distribution (Boeckli et al., 2012a; Sattler et al., 2016), periglacial landform existence (Marmion et al., 2008) and landslide occurrence (Goetz et al., 2015) as long as relationships between predictors and response are approximately linear and the sample size is larger than few hundred individuals (Hjort

and Marmion, 2008). This parametric model offers a good transparency to the user, which allows direct interpretation of the predictor variables role by examining the model coefficients. More complex models, such as random forests or support vector machines, may achieve higher performances, but their black box nature could limit the interpretation of results (Deluigi and Lambiel, 2011), and their flexibility may result in overfitting. Generalized additive models are sometimes used as a compromise between interpretability and flexibility when nonlinearities justify their use (Goetz et al., 2015; Azócar et al., 2017). Since, at regional scale, the impact of some climatic forcing on permafrost distribution is still unclear, e.g., precipitation and seasonality, it was considered important here to produce a model transparent to interpretation, and we therefore used a GLM, or logistic regression.

PFI was modeled as a categorical response variable, presence or absence of permafrost, using continuous predictor variables, e.g., elevation. Fitting a linear model directly on the categorical response variable would lead to values of probability that may exceed the interval [0;1]. The logistic link function transforms

TABLE 1 | Sub-Regions (SR) repartition of the French Alps according to main characteristics (lithology and glaciation) for rock glacier existence.

| | Massives | Predominant Lithology | Glacierization | Rock glaciers |
|------|---|---|---|--|
| SR 1 | Chablais, Bornes, Aravis | Limestones, marls | Extremely Sparse | Low density, strong interaction with glaciers |
| SR 2 | Mont Blanc, Aiguilles Rouges, Lauziere, Beaufortain, Belledonne, Ecrins, Taillefer, Grandes Rousses | Crystalline rocks, | Large cirque glaciers, some ice caps | Low density in glacierized areas. |
| SR 3 | Vanoise, Alpes Grees, Mont Cenis | Schists and ophiolites | Some cirque glaciers, isolated ice-caps | High density. Predominance of intact forms |
| SR 4 | Cerces, Arves, Queyras, Ubaye, Escreins, Chambeyron, Champsaour | Limestone, ophiolites, and mica schists | Sparse | Medium density, predominance of fossil form |
| SR 5 | Devoluy, Trois Evêchés, Pelat, Mercantour | Limestone and mica schists Isolated Crystalline Batholith | Only on the Italian Side. | High density. Predominance of polymorphic fossil forms |

the probability of permafrost occurrence to the logits, continuous and unbounded variables that can be linearly modeled (Brenning, 2005). This was done by considering the permafrost occurrence ($Y = 1$) as a probability conditioned by a set of predictors (\mathbf{x}):

$$\pi(\mathbf{x}) = P(Y = 1|\mathbf{x})$$

The conditional probabilities, also called odds, were transformed to logits by the logit link function

$$\text{logit}(\mathbf{x}) = \ln\left(\frac{\pi(\mathbf{x})}{1 - \pi(\mathbf{x})}\right)$$

which are continuous and unbounded, and the linear model consisting of the intercept α and the predictor variables coefficients β could therefore be fitted upon them:

$$\text{logit}(\mathbf{x}) = \beta\mathbf{x} + \alpha$$

The model was fitted by maximum likelihood, and it could be used to predict the probability of permafrost occurrence for any set of predictors. The model was computed using the function `glm` in the R software, while the data handling was done using the packages `RSaga` (Brenning, 2008) and `rgdal` (Bivand et al., 2015).

Model performance was evaluated using the Akaike Information Criterion (AIC) and the Area Under the Receiver Operating Characteristic (AUROC; Hosmer and Lemeshow, 2000) estimated by cross-validation (Hand, 1997). The AIC was used for model selection, measuring the goodness of fit of different models trained on the same dataset. Since the AIC includes a penalty term that depends on model complexity, less complex models are preferred to larger models that fit the data equally well. Smaller AIC values indicate better goodness of fit. In this study, the AIC is used to select a subset of predictor variables by stepwise forward selection.

The AUROC is a tool to evaluate the quality of the model classification. Varying from 0.5 (random classification) to 1 (perfect classification), it quantifies the probability that the model will properly classify a dependent variable for a set of predictors variables.

Cross-validation is a statistical estimation procedure that can be used to assess the predictive performance and determine the

degree of overfitting of a model. In general, it is performed by randomly dividing the dataset into two subsets used for training the model and testing its predictive capabilities, respectively. The process is repeated until all the data have been used at least once as a test set. All test-set predictions are combined to obtain a cross-validation estimate of the performance measure, in this study the AUROC. A model performing on average substantially better on the training set compared to the cross-validation results is said to overfit.

Here, cross-validation was performed using the sub-regions as partitioning class: the model was trained on 4 sub-regions, while the model performance is evaluated on the left-out sector. The hypothesis behind this validation method was that climate, lithology and landscape history within a sector are relatively homogeneous compared to the other sectors. By building models on a group of massives and testing them in the remaining sector, it was possible to assess the transferability of the model to different environments within our study region. Model performances were evaluated using the R package `sperrorest` (Brenning, 2012).

Data

In the present study the observational units were the fossil rock glaciers (absence of permafrost) and the production areas of active rock glaciers (presence of permafrost).

Response Variable: Rock Glacier Inventory

The response variable was obtained using the first exhaustive rock glacier inventory of the French Alps, compiled and supplied by the RTM (Roudnitska et al., 2016). The rock glaciers were inventoried using Bing satellite imagery (yearly updated and accessible via QGIS), IGN aerial orthophotography (available for 2013 and at 50 × 50 cm resolution) (IGN, 2013). The digitized outlines of the landforms included their rooting zones, i.e., the slopes that furnish the debris to the rock glaciers and where the ice-debris mixture is supposed to originate (Humlum, 1998). The activity status was judged according to morphologic indicators (Barsch, 1996; Scapozza, 2008).

Since the inventory was compiled by different operators, the landforms' geometry and activity status were cross-revised by the authors in order to ensure the quality and conformity of the digitization method and activity attribution. Rock glaciers that were suspected to interact with glacial processes were

discarded because of their ambiguous rooting zone. Also, only monomorphic forms (Frauenfelder and Käab, 2000) with fully identifiable outlines were selected. Polymorphic forms are witnesses of a complex interaction between climate and local geomorphology (Frauenfelder and Käab, 2000; Bodin, 2013). Since we study rock glaciers in correlation with climatic proxies, these complex interactions may not be trivial to be modeled statistically and may induce biases.

Historical orthophotos were used to determine if an allegedly active rock glacier was actually moving. Three sets of historical orthophotos (from 2000 to 2006, 2006 to 2010, and 2013 to 2015) are made available to the public by the National Institute of Geography (IGN, 2011b, 2013). Image resolution ranges from 1×1 m for the older data to 0.5×0.5 m for the 2013–2015 data. Displacement was assessed by visual inspection. In general, orthophotos present different degree of distortions, often higher in steep slopes. Rock glacier movement was confirmed when consistent with slope orientation and greater than any geometric distortions observed in the neighborhood of the landform.

In the present study the production area, or rooting zone, only of active rock glaciers was considered as evidence of permafrost existence (Bodin et al., 2008; Sattler et al., 2016). The rooting zone is the area where the mixture of ice and debris is created, and due to an over-saturation with ice, permafrost creep starts generating the rock glacier (Barsch, 1996; Humlum, 1998). The hypothesis behind this was that the climatic conditions favorable for mountain permafrost existence are found primarily in the rooting zone of active rock glaciers. In this way the model was considered to avoid bias by including the rock glaciers' dynamical processes that export permafrost toward lower altitudes and that depend not only on the local climate but also on the morphometric characteristics of the rock glaciers' catchment (Humlum, 1998; Matsuoka and Ikeda, 2001; Frauenfelder et al., 2003; Brenning, 2005; Janke and Frauenfelder, 2007; Cossart et al., 2010). The break between the talus and the typical morphology of an active rock glacier, dominated by ridges and furrows, was used as a morphological criterion to identify the lower limit of production areas. Landforms whose production area was not clearly identifiable were discarded. In agreement with previous studies (Boeckli et al., 2012a; Sattler et al., 2016; Azócar et al., 2017), fossil rock glaciers were used as evidence of permafrost absence. While in active rock glaciers the rooting zone is often clearly identifiable, this is not the case for the fossil forms because of vegetation cover and re-arrangement of the surface after the ground thaw. For this reason, we chose to use the centroids of fossil rock glaciers instead of fossil production areas as evidence of permafrost absence.

Predictor Variables: Topoclimatic Data

The predictor variables aim to describe the processes that determine the realization of the response variable.

Altitude and potential incoming solar radiation

A DTM at a 25×25 m pixel size (IGN, 2011a) was used to calculate the Potential Incoming Solar Radiation (PISR) with the *Terrain analysis* toolbox in SAGA (Wilson and Gallant, 2000). The PISR is the sum of the computed direct and diffusive

components of the solar radiation using the DTM. Calculation is made assuming clear sky conditions, i.e., transmittance at 70%, and absence of snow cover. The PISR describes the process of ground warming caused by the direct exposure to the sunlight, and it also controls snow cover duration.

Air temperature

Gridded temperature data at a 1×1 km resolution belongs to EDF (Gottardi, 2009) and are the result of an inverse distance weighting interpolation corrected for elevation of data from a network of 216 weather stations in the French Alps during the 1975–2005 period. In order to take into account strong climatic gradients due to topography, the method is based on a “ridge crossing” distance, which penalizes the interaction between two weather stations separated by a topographic barrier. Using a cross-validation method, the RMSE of the interpolation was assessed to be 1.07°C for maximum temperatures (TX, mean over the warmest month) and 0.77 for minimum temperatures (TN, mean over the coldest month).

Mean Annual Air Temperature (MAAT) was calculated using the same method for TN and TX on the same weather station network. TX and TN were considered as proxies of winter and summer temperatures, respectively. A seasonal variation of temperature (TSEASON) was also calculated as the difference between TX and TN, as proposed by Boeckli et al. (2012a). Since water bodies tend to reduce seasonal variation, TSEASON was also a measure of continentality, which is suspected to play a significant role in rock glacier distribution (Sattler et al., 2016).

In order to capture the local variability of air temperature due to the complex terrain, the data were resampled using the 25×25 m DTM. Since lapse rates are known to vary from the northern to the southern Alps, using a constant lapse rate, e.g., $-0.65^\circ\text{C}/100$ m, would be incorrect (Gottardi, 2009). Therefore, lapse rates were locally obtained by computing the linear relationship between neighboring temperatures and elevation data for each grid cell at 1 km resolution. Lapse rates were then resampled at 25 m resolution using a bilinear interpolation method. Finally, the temperatures were adjusted by calculating the differences in elevation between the 25 m and the 1,000 m DTMs and multiplying them by the resampled lapse rate.

Precipitation patterns

The precipitation dataset belongs to EDF (Gottardi, 2009). Gottardi (2009) classified the precipitation in the French Alps during the 1948–2003 period into eight regional-scale perturbation systems, called Weather Types (WTs). Gottardi (2009) evaluated the spatial pattern of the precipitation intensity using the same statistical approach used to estimate the air temperatures, and finally obtained raster maps at 1×1 km resolution of mean precipitation amount for each WT. Interpolation was done including also high-altitude nivometers, making the estimation more reliable for the mountainous environments.

Using the time series of the daily WT, it was possible to create precipitation maps for different time periods. The monthly frequency of each weather type was evaluated by counting their mean occurrences for each month along the whole time

series. Frequencies are then used to weight the sum of the WT maps to create monthly precipitation maps that were further aggregated to produce a Mean Annual Precipitation (MAP) map and seasonal precipitation maps. The latter were created by aggregating the months when solid and liquid precipitations are expected to fall in the high mountain environment. Solid precipitation is expected to fall in winter (October to May), while liquid precipitation is expected to fall in summer (June to September). Finally, the Early Winter Precipitation Ratio (EWPR), i.e., the ratio of the total winter precipitation falling before February was also computed.

Uncertainty Estimation of the RGLs Inventory

The rock glacier activity attribution using the classical geomorphological approach involves a certain degree of subjectivity. To assess the influence of this source of uncertainty on the statistical modeling results, a model trained on the full active production areas dataset was compared to a model trained on the confirmed production areas, i.e., production areas feeding a rock glacier that were found to be moving by inspecting the historical orthophotos. Uncertain active production areas were assigned an Activity Uncertainty Index (AUI) of 2 and confirmed active production areas were assigned an AUI equal to 1. By this approach, we aimed at quantifying uncertainties in the model arising from uncertain interpretation of rock glacier movement.

The subjectivity of the rock glacier digitization process is also a known source of uncertainty (Schmid et al., 2014). Here, we assessed this uncertainty by having three operators digitalize 10 randomly selected rock glaciers and comparing their boundaries and activities. The influence of the digitization uncertainty on the model was then assessed by a Monte Carlo approach. For each landform in the database the values of MAAT and PISR were randomly varied using a normal distribution of standard deviation equal to the values found by the digitization uncertainty assessment process. The randomization of the database was performed 1,000 times and for each run a model was fit on the database. The models were then compared by evaluating the difference of the PFI in the predictor variable space MAAT-PISR.

RESULTS

Presentation of the Rock Glacier Inventory

The complete inventory (Figure 3) included 3,261 rock glaciers, 814 of which were classified as active, 671 as inactive and 1,776 as fossil. By visual inspection of the rock glacier inventory, it can be observed at first that the landform distribution within the different sub-regions is uneven by density and activity status. While sub-regions areas were comparable (with the exception of sub-region 1), rock glacier numbers ranged from about 600 in sub-region 2 to more than 1000 in sub-region 3. Also the distribution of rock glacier activity within each sub-region was uneven, ranging from 60% of fossil rock glaciers in sub-region 4 to more than 50% of active ones in sub-region 1.

Almost half of the rock glaciers were not isolated and coexist sharing lobes or being superimposed upon each other in polymorphic structures (Frauenfelder and Käab, 2000).

Polymorphic structures represented a great source of uncertainty since the separation of different lobes and flow paths using only aerial imagery could be subjective. The number of rock glaciers given above is therefore not an absolute one. A considerable number of rock glaciers was found to be interacting with present and former glaciers, especially in sub-regions 1 and 2.

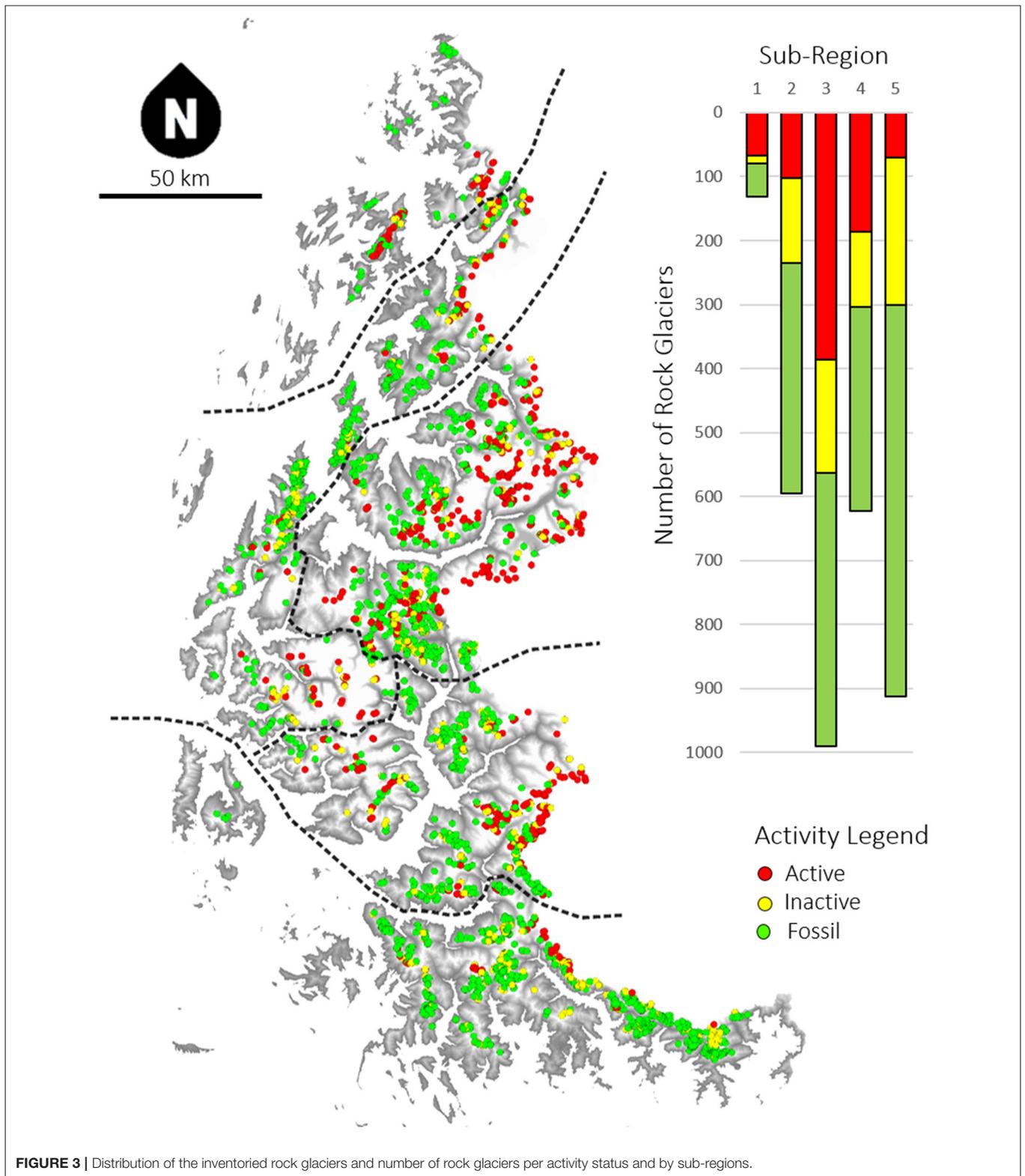
The revision of the inventory allowed to discard landforms not suitable for modeling. The sampling used for the modeling stage comprised 541 fossil rock glaciers and 515 active production areas (Figure 4). Of the 515 active production areas, 373 were feeding rock glaciers whose movement was clearly identifiable on historical imagery. The centroids of active production areas and fossil rock glaciers are available as spreadsheets files in the Supplementary Materials.

Statistical Modeling

A summary of the modeling results is shown in Table 2. In general, the logistic regression model was found to describe the rock glacier distribution extremely well, which is in agreement with previous studies (Boeckli et al., 2012a; Sattler et al., 2016). The AUROC values can be classified as “outstanding” (Hosmer and Lemeshow, 2000). The cross-validation analysis showed that the models had a strong predictive power, since training and test performances were comparable. Nevertheless, the high AUROC values were at least partly due to the elimination of inactive rock glaciers from the data set.

As found in previous work, MAAT (or its topographical proxy, the elevation) and PISR were the most significant predictors. The stepwise variable selection showed that MAP and seasonality are not significant predictors. EWPR was the only predictor linked to precipitation to have high statistical significance. The north-south trend described above was also highly significant, indicating that in the southern Alps the permafrost belt is located at higher elevations but it can exist at higher MAATs (cf. Figure 8). The relationships with respect to the MAAT and PISR and EWPR (Figure 4) were inspected. It could be observed that the behavior of rock glaciers with respect to the local climate varies throughout the region roughly from north to south. Active production areas could be found at positive MAATs commonly in the Southern Alps, while fossil rock glaciers were also found at negative MAATs in sub-region 3. This regional trend was found to be coherent with EWPR, which decreases toward the southern Alps. The role of PISR in separating presence/absence of permafrost was also variable, being more relevant in sub-regions 1, 4, and 5.

A topographic model, i.e., using only terrain attributes (PISR, Elevation and Northing), and a climatic model, i.e., using climatic data (PISR, MAAT, and EWPR), were fitted to the database of confirmed active and fossil landforms (AUI = 1). The topographic model fitted substantially better than the climatic model (AIC = 132 vs. AIC = 261). The PFI map was therefore computed using the topographic model and the database of confirmed active production areas (AUI = 1). The PFI map is available as geotiff image in the Supplementary Materials (Figure S1). The difference between the predicted PFI and the actual PFI at the rock glacier locations is presented in Figure 5. When the residual was close to 1 then the model failed to predict



the PFI, since the active production area (respectively fossil rock glacier) was in a low (high) PFI. Overall there is a slight regional trend in the residuals: in Mercantour (Sector 5), Ecrins and

Belledonne (2) some active production areas can be found in low PFI values, while in Eastern Vanoise (3) and Ubaye (4) fossil rock glaciers can be found within the modeled permafrost belt.

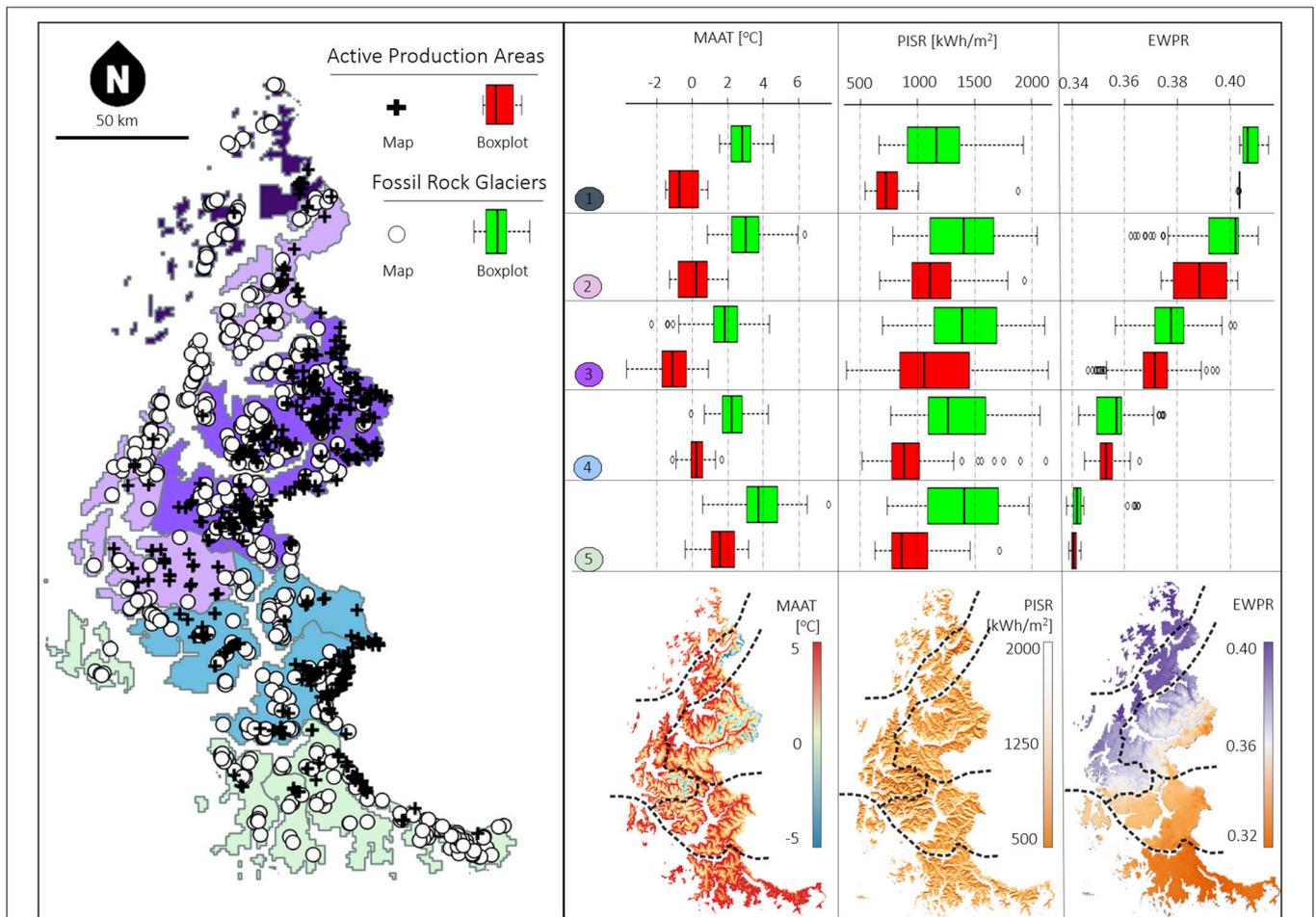
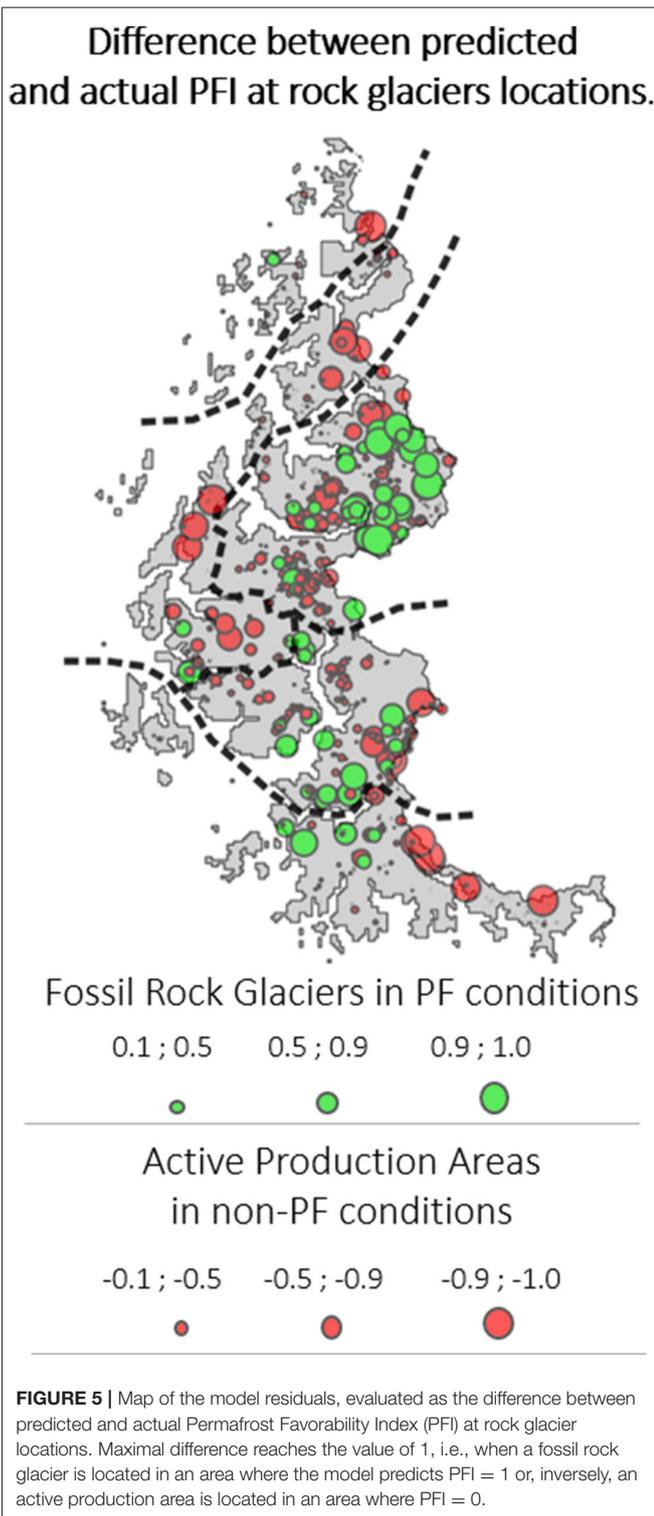


FIGURE 4 | Predictor-variable boxplots for the sampling of rock glaciers used in the modeling. Green boxes refer to P₀ (active production areas of rock glaciers), while red boxes refer to P₁ (fossil rock glaciers). Predictor-variables boxplots are presented by sub regions. The data P₀ and P₁ are available as spreadsheet files in the Supplementary Materials.

TABLE 2 | Model coefficients and standard error in brackets, followed by model performances.

| | AUI = 1 | | AUI = 1 & 2 | |
|----------------|---------------------------|---------------------------|---------------------------|--|
| | Topographic model | Climatic model | Climatic model | |
| Intercept | -1.78 e+02 (3.98 e+01)*** | 2.02 e+02 (3.93 e+00)*** | 2.14 e+02 (3.10 e+00)*** | |
| PISR | -9.14 e-03 (2.69 e-03)*** | -5.85 e-03 (6.24 e-04)*** | -4.29 e-03 (4.33 e-04)*** | |
| Elevation | 2.45 e-02 (2.69 e-03)*** | - | - | |
| Northing | 1.99 e-05 (5.22 e-06)*** | - | - | |
| MAAT | - | -2.54 e+00 (1.99 e-01)*** | -2.20 e+00 (1.46 e-01)*** | |
| EWPR | - | -3.00 e+01 (1.02 e+01)** | -3.77 e+01 (7.90 e+00)*** | |
| CV Train AUROC | 0.99 | 0.98 | 0.98 | |
| CV Test AUROC | 0.98 | 0.97 | 0.96 | |
| AIC | 132 | 261 | 404 | |

AUROC values are given with respect to the cross validation (CV) method.
 p-values significance: ***p < 0.001, **p < 0.005.



Uncertainty Assessment

In **Table 2** are reported the predictor coefficients and performances of the models trained with confirmed production areas and with all the production areas. While directly comparing the model performances is not meaningful due to the different database used, we could observe a substantial difference between

the two models in the PISR coefficient. Introducing the uncertain production areas in the modeling dataset generated a model that gives less importance to the role of the PISR. The difference in the two models, shown in **Figure 6**, could reach a difference in the PFI of 0.35 in the MAAT-PISR space.

Comparing the digitisation of 10 landforms by three different operators revealed results comparable to previous knowledge (Schmid et al., 2014). Most of the disagreement among the operators concerned the upper boundaries of the landforms and areas with dense vegetation (**Figure 7**). Differences in surface areas were high up to 16%. The resulting uncertainty on the predictor variables was 0.10°C for the MAAT and 61 kWh/m² for the PISR, values used as standard deviation in the Monte Carlo perturbation of the inventory data. The range of the resulting models is presented in the space of the predictor variables MAAT and PISR in **Figure 6**. It could be observed that the digitization uncertainty may introduce biases of up to about 10%, which are maximized (15%) for high solar radiation (~2,000 kWh/m²) and near-zero temperatures.

DISCUSSION

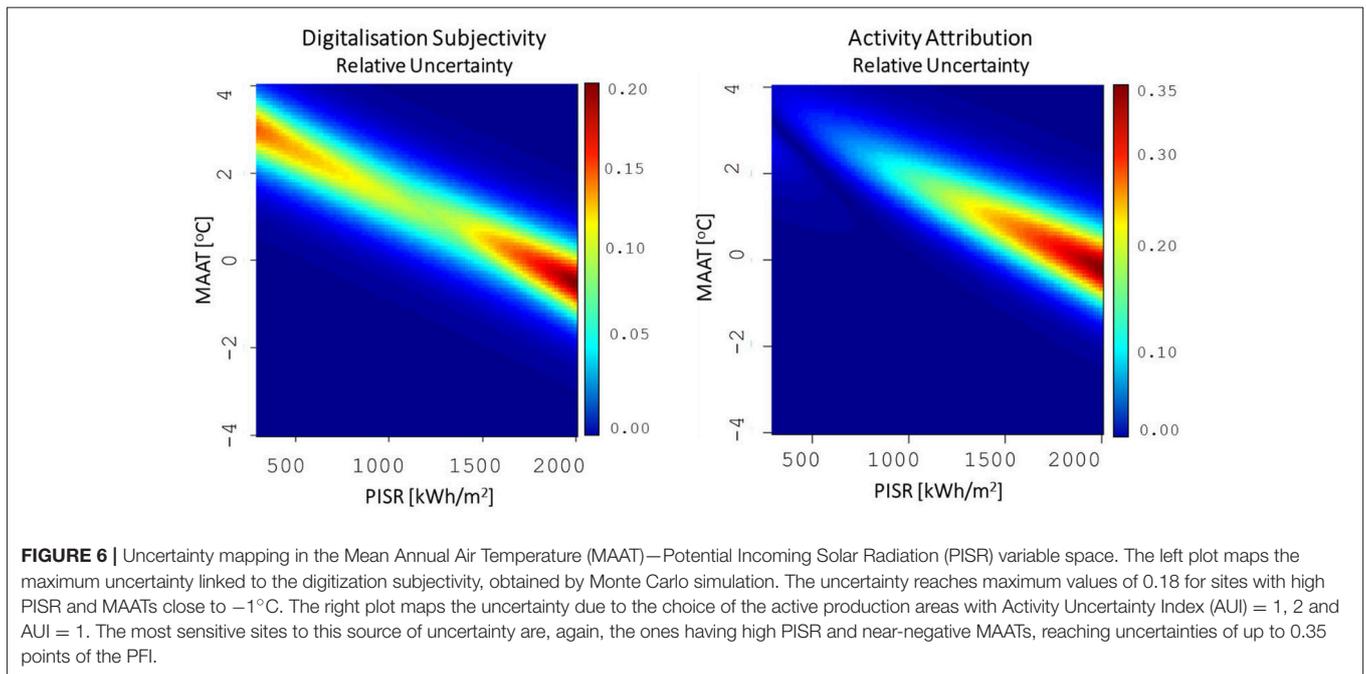
Database

Rock Glacier Inventory

The observed uneven distribution of the rock glaciers is in agreement with the findings of previous studies (Guodong and Dramis, 1992; Harris and Corte, 1992; Kenner and Magnusson, 2017). In glacierized areas, i.e., Sector 2, rock glaciers are sparse. In sector 3, in contrast, glaciers are rather limited due to a drier climate, more space is left for periglacial landforms, and active rock glaciers are therefore widespread. In the Southern Alps rock glaciers are abundant, thanks to the non-existent glaciation and a densely jointed lithology prone to rock-weathering and talus production (Matsuoka and Ikeda, 2001). The predominance of fossil forms suggests favorable conditions for widespread permafrost in the late-glacial cold events, i.e., in a drier and colder climate (Frauenfelder and Kääb, 2000).

Predictor Variables

The climatic data used are specific for the region and the altitudes of interest and therefore considered appropriate for the present study. However, they cover only the second half of the past century, while the onset of active production areas is a phenomenon that took place at the scale of several centuries to millennia (Kerschner, 1978, 1985; Frauenfelder and Kääb, 2000; Frauenfelder et al., 2001; Cossart et al., 2010). Temperature and precipitation are known to vary considerably during the Holocene (Ilyashuk et al., 2011; Fohlmeister et al., 2013), therefore the data for each rock glacier derived from the climate dataset should be interpreted with respect to regional trends and not as the absolute values of the climatic variables. Furthermore, the predictor variables are only proxies for processes affecting permafrost existence. Each predictor variable represents the combined effect of several processes and results therefore have to be interpreted carefully since the assumed predominant process, described by a predictor variable may not be trivial. For example, air temperature, being strongly correlated to the altitude, has an influence on ground cooling, snow cover persistence and also



on cloud cover. Also, rock glacier activity status is more likely related to thermal conditions decades to centuries ago (Lambiel and Reynard, 2001). While any overall shift in temperature would be absorbed by the model’s intercept term, we must still assume that spatial patterns of present-day MAAT are coherent with spatial patterns in temperature during the time period relevant for controlling rock glacier activity status and permafrost presence/absence.

Statistical Modeling Model Performance

By examining the out-of-sample AUROC values of each sub-region it can be observed that the simple model combining MAAT and PISR performs very well (Table 3). By adding precipitations as a third predictor to the model, the AUROC reaches lower performances. This is attributed to the fact that precipitation varies at a regional scale and the model may tend to overfit the response variable when trained at the local scale. This indicates that a complex model performs well where the training points cover evenly one area but performs poorly when transferred to other areas. A simple model is therefore preferable to predict permafrost occurrence outside the dataset’s spatial domain.

Model performances in the entire region generally vary around AUROC = 0.98, which can be classified as “outstanding” (Hosmer and Lemeshow, 2000). Although it is complex to compare model performances of different studies, the results obtained by Boeckli et al. (2012a) (AUROC = 0.91) and Sattler et al. (2016) (AUROC = 0.98) using the same statistical models, are similar. The lower performance obtained by Boeckli et al. (2012a) can be explained by the fact that we used active production areas and fossil rock glaciers to train the model, while

the APIM is based on intact and fossil rock glaciers, which are “climatically” in between because of the dynamic behavior of these landforms that moves permafrost toward lower elevations.

The topographic model performs marginally better than the climatic model, suggesting that the elevation is a better predictor than MAAT. However, the elevation, which is a proxy of MAAT, does not capture the climatic patterns influenced by the Atlantic Ocean and the Mediterranean Sea. For this reason we focus in section Regional patterns in the spatial distribution of the rock glacier activity on the interpretation of the climatic model.

Model Uncertainties

While the Monte Carlo simulation highlights that the model is reasonably robust to random digitization error in the rock glacier inventory, the uncertainty concerning rock glacier activity attribution by morphometric parameters (Scapozza, 2008) is not negligible. The use of historical imagery of the rock glaciers reduces this uncertainty, ensuring that the active production areas are actually evidence of permafrost since creep could be observed. However, it must be noticed that in a certain number of cases the movements of the rock glaciers could not be assessed because of distortions and poor georeferencing in the orthophoto or presence of snow cover. Also, due to the relatively short time span of 8–15 years covered by the aerial imagery, the movement of rock glaciers creeping at small velocities ($\sim 0.1\text{--}0.2\text{ m/y}$) may have remained undetected.

Regional Patterns in the Spatial Distribution of the Rock Glacier Activity Climatic Influence

The influence of the Atlantic and Mediterranean weather systems seems to play a significant role for rock glacier activity. It is

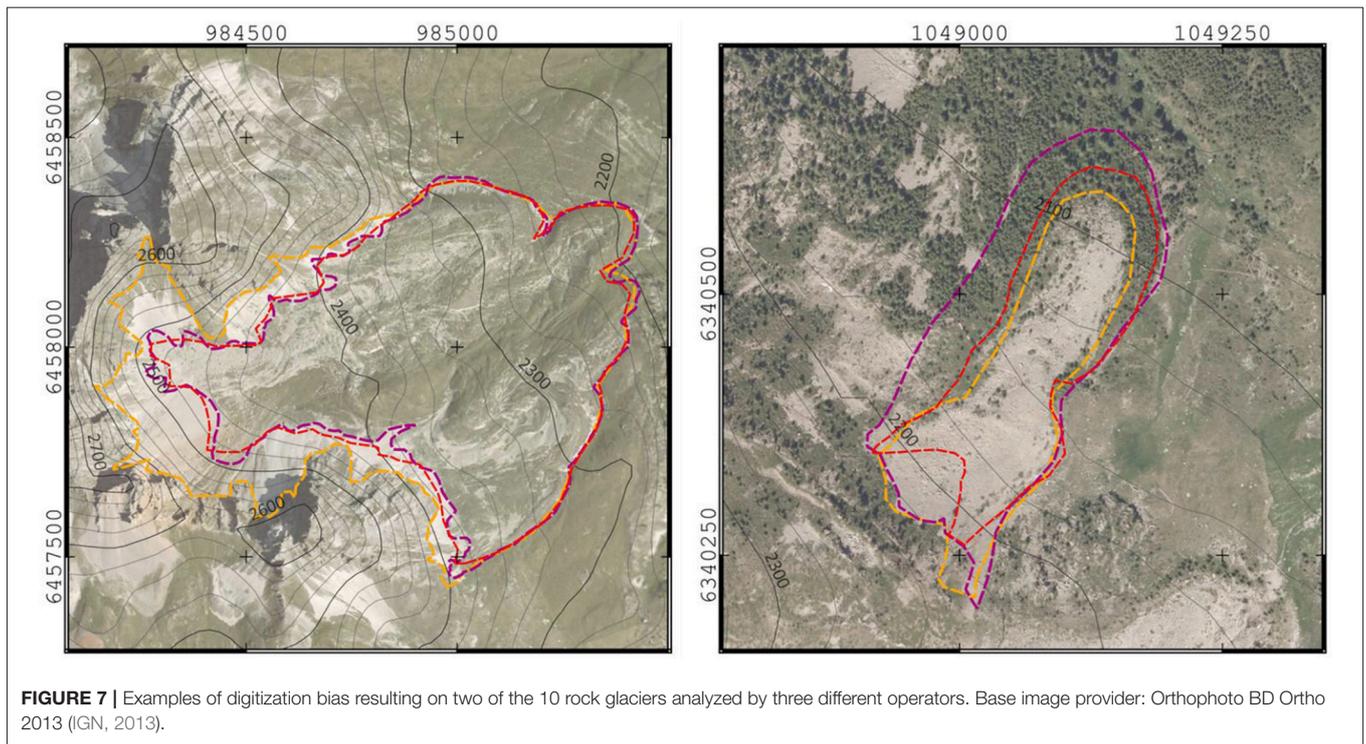


TABLE 3 | AUROC values for two different models.

| Training sub-regions | Model: MAAT + PISR | Model: MAAT + PISR + EWPR |
|----------------------|--------------------|---------------------------|
| Sub-region 1 | 0.94 | 0.74 |
| Sub-region 2 | 0.97 | 0.96 |
| Sub-region 3 | 0.96 | 0.94 |
| Sub-region 4 | 0.97 | 0.97 |
| Sub-region 5 | 0.98 | 0.60 |

The models are trained using a single sub-region as training set and tested on the 4 sub-regions left out. The process is repeated for each sub-region.

interesting to notice that, while PFI limits increase in elevation toward the South, they are found at higher temperatures (Figure 8). Previous studies managed to explain regional trends by analyzing the mean annual precipitation or the continentality (Ridefelt et al., 2008; Boeckli et al., 2012a; Sattler et al., 2016; Azócar et al., 2017). However, in the present study these predictor variables were found not significant in the statistical model. EWPR is found to be the predictor better explaining this behavior, suggesting that precipitation patterns may play a more important role in the distribution of rock glaciers activity.

Crossing our findings and the data in Durand et al. (2009b), it appears that the snow cover presents varying characteristics from North to South. While previous studies highlighted the relevance of the snow cover characteristics for permafrost existence on a local scale (Lerjen et al., 2003; Delaloye, 2005; Bodin, 2007; Luetschg et al., 2008; Gubler et al., 2011; Apaloo et al., 2012), we suggest that this factor could be relevant on a regional

scale, influencing rock glacier activity distribution by altering the ground thermal regime. An insulating and long-lasting snow cover appearing in early winter characterizes the northern sectors impeding an efficient ground cooling during the first part of the cold season. On the other hand, a thinner cover more prone to ground cooling characterizes the southern Alps, allowing active production areas to exist in conditions otherwise unsuitable for permafrost.

Lithological Influence

Lithological influences on rock glacier distribution have been identified by many authors (e.g., Johnson et al., 2007). Matsuoka and Ikeda (2001) observed that crystalline rockwalls tend to produce coarser debris, resulting in “blocky” rock glaciers, while shales rockwalls produce finer debris, resulting in “pebbly” rock glaciers. A coarser matrix may enhance the Balch thermal ventilation, potentially allowing blocky rock glaciers to exist in higher MAAT (Johnson et al., 2007). These observations seem to be consistent with the distribution of residuals shown in Figure 5 and dominant lithology (BRGM, 2015) in Figure 2. Indeed, it can be noticed that active production areas in low PFIs are more likely found in granites and gneiss, while fossil rock glaciers in high PFIs are more likely found in schists, ophiolites, and limestones. In order to test the statistical validity of this observation, a GLM is computed adding the lithology as categorical predictor variable. Lithology is divided into two groups, i.e., granites and gneiss vs. the others types of rocks. The resulting model has an AIC of 120 (vs. AIC = 132 for the topographic model) and the lithology predictor has a *p*-value of 0.0018, confirming that the lithological patterns have a

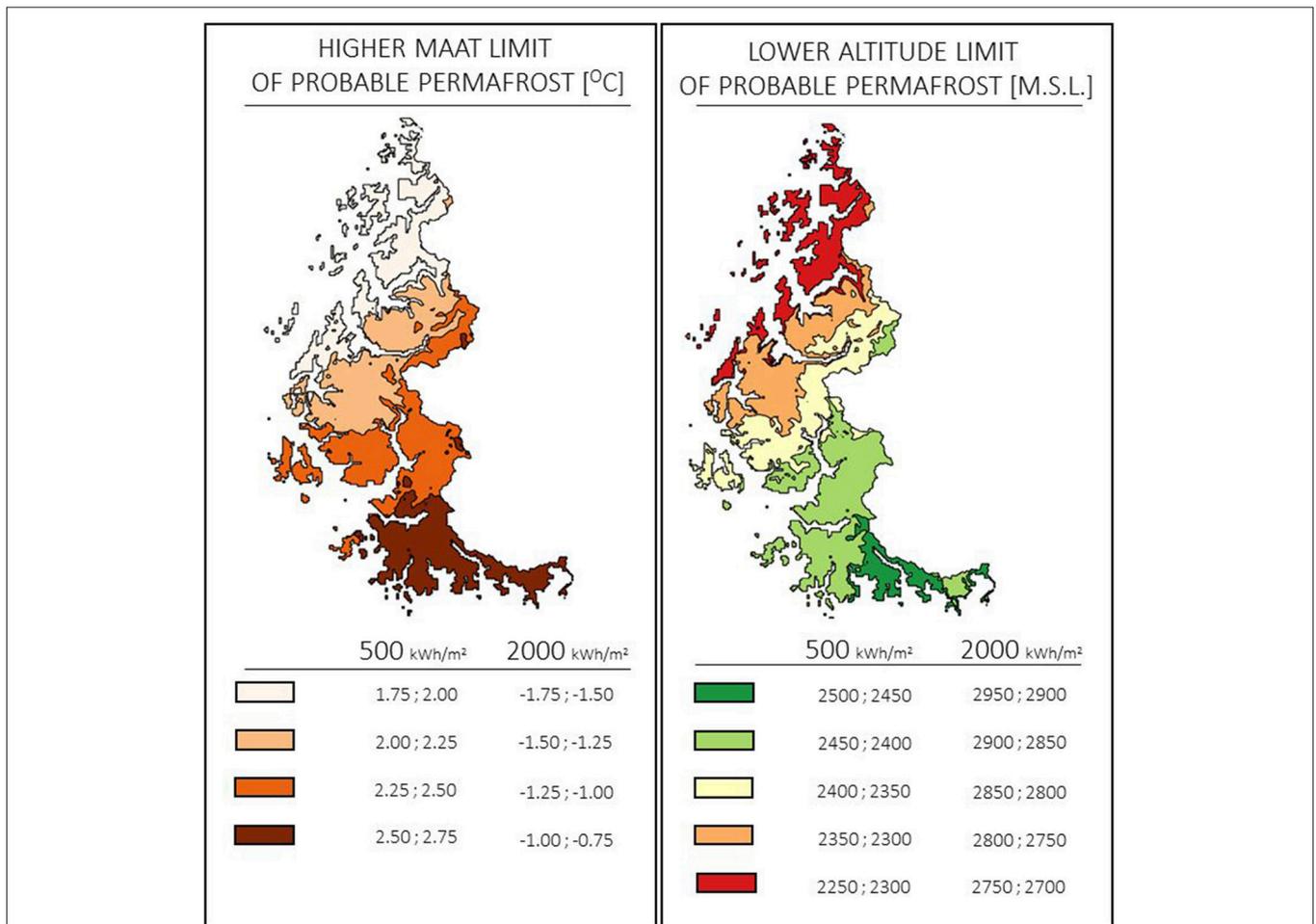


FIGURE 8 | Lower limit of probable permafrost [Permafrost Favorability Index (PFI) > 0.6] at the regional scale with respect to Mean Annual Air Temperature (MAAT) and elevation. Data are presented for two values of solar radiation: 500 kWh/m², corresponding to a typical small north-facing cirque, and 2,000 kWh/m², corresponding to an open south face with moderate steepness (~30°). It can be observed that permafrost limits rise in elevation from the northern toward the Southern Alps, due to the warmer climate. However, probable permafrost is found at higher MAAT in the Southern Alps, thanks to the favorable precipitation patterns.

statistically significant role in rock glacier activity distribution. Specifically, the odds ratio of a rock glacier being active (vs. fossil) with respect to the lithology is 18.5. This corresponds to a shift downwards by 111 m of the permafrost belt in crystalline lithology.

Model Interpretation Comparison with the APIM

Since both APIM and the present PFI cover the French Alps, a comparison between the two models is proposed. However, we wish to stress that, since the two models are based on a different database and different statistical approaches, any comparison is limited. We therefore propose a merely qualitative comparison, presented in **Figure 9**, without the intention to prove the superiority of one or the other model. Considering that the French Alps are the only region that was not extensively used to calibrate the APIM, our test may give an insight into the predictive power of APIM. In general, APIM is found to be suitable for the French Alps, as permafrost distribution

is comparable to the PFI. The distribution of the residuals, presented in **Figure 10**, shows that APIM has the tendency to slightly overestimate permafrost probability with respect to our database, as fossil rock glaciers are more likely to have an APIM values between 0.1 and 0.2 rather than 0. This can be explained by the omission of inactive rock glaciers from the present model fit, which creates a sharper transition between the zones of expected presence and absence of permafrost, witnessed by a higher AUROC.

Use of the Model

Process-related assumptions at the core of the PFI and the APIM models are substantially different, resulting in different interpretations of the distribution maps. While the APIM uses intact rock glaciers as indicators of permafrost existence and successively applies a thermal offset to account for the dynamic and thermal bias induced by the creep of ice-rich coarse debris, the present model predicts the probability, based on the climate, of the existence of an active production area vs. the presence of

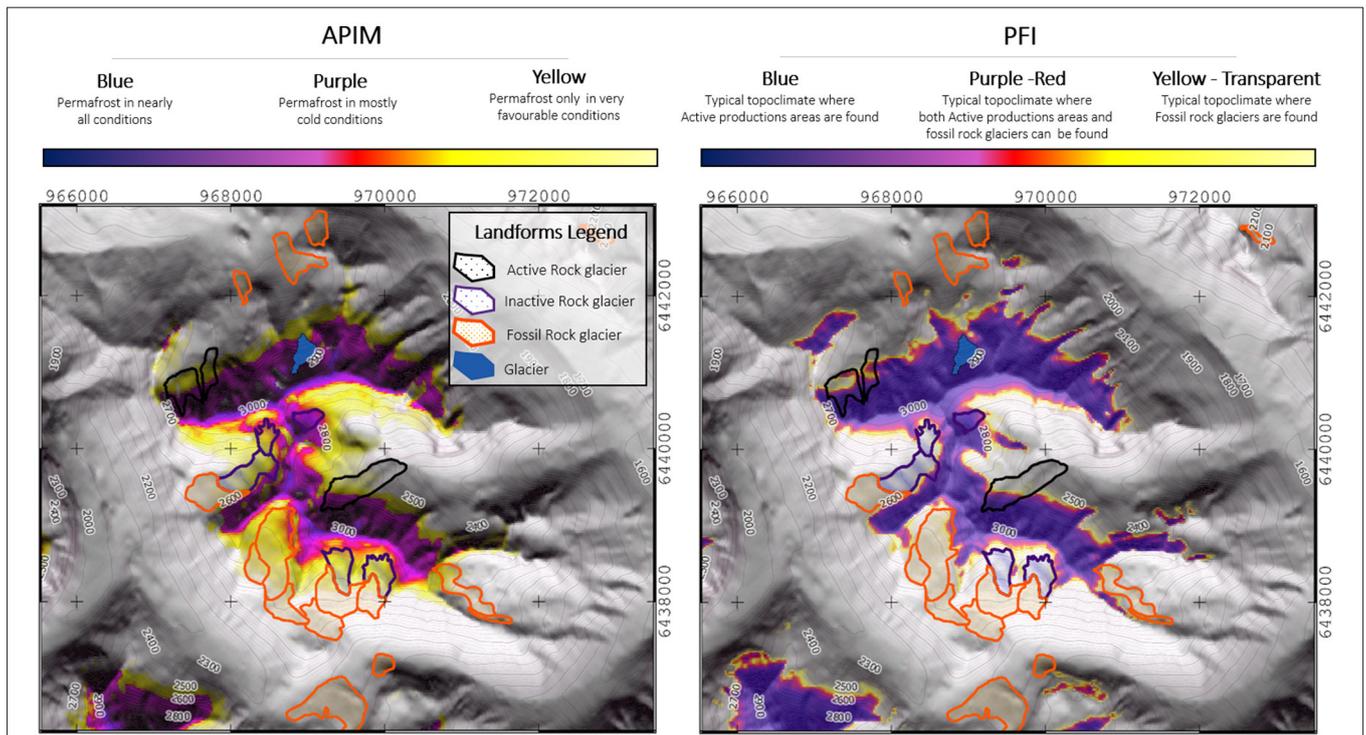


FIGURE 9 | Comparison between Alpine Permafrost Index Map (APIM) and Permafrost Favorability Index (PFI) in Massif du Combeynot. The PFI map is available as geotiff image in the Supplementary Materials.

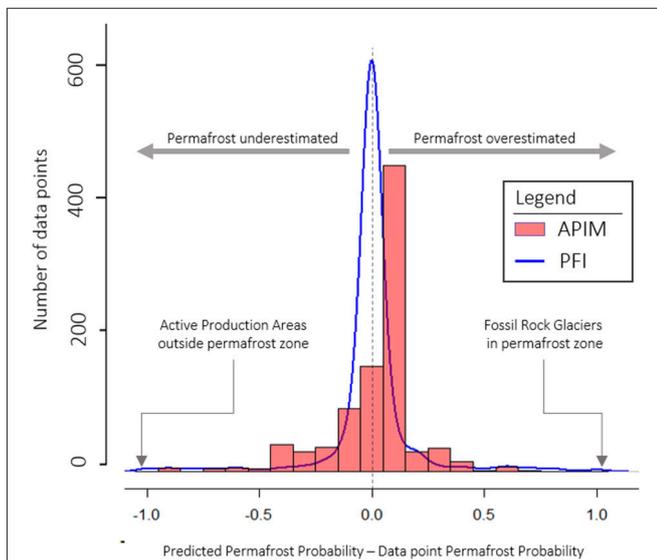


FIGURE 10 | Comparison between the residuals on our inventory with respect to Alpine Permafrost Index Map (APIM) and Permafrost Favorability Index (PFI). Since the PFI is calibrated on this inventory, residual distribution is not surprisingly centered to zero.

a fossil rock glacier. PFI probabilities should be interpreted as climatic suitability for the existence of permafrost in relation to rock glaciers. An active rock glacier which is located in a low-PFI

site can be interpreted as a rock glacier which is in a climate where fossil forms are statistically more likely to occur; such a rock glacier could either currently be degrading, or it might be either explained by processes not described by the model, e.g., unusual permafrost preservation.

Mountain permafrost degradation and the climatic inactivation of rock glaciers have been observed in recent climatic episodes such as the Medieval Warm Period (Scapozza et al., 2010; Ramelli et al., 2011; Luetscher et al., 2013). Therefore, the currently active rock glaciers are believed to be representative for the climatic conditions of the colder periods of recent centuries, e.g., the Little Ice Age (LIA) (Lambiel and Reynard, 2001; Cossart et al., 2010). Although rock glaciers have been demonstrated to be sensitive to climate change, the response time is estimated to be of several decades to centuries (Scapozza et al., 2010; Scotti et al., 2017). As consequence, the present model likely reflects the permafrost favorability in climatic conditions that are no longer valid in the present climate, and even less so in warmer future climate.

CONCLUSIONS

The study presents the first permafrost distribution model of the French Alps calibrated on a rock glacier inventory covering the region. Rock glacier activity distribution is best modeled using a set of three predictors: Elevation, PISR and Northing. The statistical significance of the Northing predictor shows

a regional trend that is suggested to be correlated with the properties of snow cover insulating the ground. The relevance of the dominant lithology is also highlighted, showing that active production areas fed by crystalline rockwalls tend to exist in topoclimatic conditions unfavorable to permafrost existence while the opposite may occur to fossil rock glaciers generated by rockwalls dominated by schists, limestones, and ophiolites. While previous studies indicate the importance of these processes for thermal state of permafrost at local scale, here we suggest that snow cover characteristics and lithology may control rock glacier activity at the regional scale. Several new procedures are proposed to improve the established modeling process, such as inventory revision using historical imagery, digitization uncertainty assessment and spatial cross-validation. The use of historical imagery to verify rock glacier activity was found to substantially reduce the model uncertainty. The model was found to be fairly robust to the subjectivity of the digitization process. However, it is suggested that performing the inventory revision is necessary when the study area is spatially partitioned among different operators, which may introduce personal biases on spatial clusters of rock glaciers. The spatial cross-validation suggests that the model has a good predictive power which, however, decreases significantly when training the model on a reduced dataset and using a larger number of predictors.

Although comparing models designed on different databases and statistical inferences is problematic, a qualitative comparison with APIM data is proposed. Considering that the APIM was not calibrated using this dataset, the present study suggests that this model has a very good predictive power. Differences between the two models are mainly due to the fact that the PFI has a sharper transition between permafrost and non-permafrost areas. As a consequence, the APIM generally tends to slightly over-estimate the permafrost spatial distribution with respect to our database. The differences between PFI and APIM are however at the core of the statistical approaches used: the PFI has to be interpreted as the

probability that the local climate during the LIA was suitable for the existence of an active production area. The model is therefore a useful tool to detect rock glaciers that are in disequilibrium with the local climate and may present degradation issues. However, the model is lagging in time since it does not account for the present disequilibrium that already affects Alpine permafrost. Therefore, future work will focus on integrating degradation data in the model in order to understand the spatial extent of the processes behind these phenomena.

AUTHOR CONTRIBUTIONS

PS, RC, XB, and MM conceived the project. RC and FG supplied and pre-processed fundamental data, which were revised by MM and XB. AB and MM designed the statistical modeling approach. MM wrote the paper and produced the figures. AB, XB, and PS contributed to the writing providing feedback, revision, and editing of the paper.

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SUPPLEMENTARY MATERIAL

The Supplementary Material for this article can be found online at: <https://www.frontiersin.org/articles/10.3389/feart.2017.00105/full#supplementary-material>

Figure S1 | Permafrost Favorability Index map in format geotiff. The reference system is Lambert 93 (EPSG:2154).

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Conflict of Interest Statement: The rock glaciers inventory used for the modeling process was realized under the financial support of the RTM. Therefore, this database is property of the French Minister of the Environment, which is engaged to publish it free of charge to the public by the end of 2018.

The authors declare that the research was conducted in the absence of any commercial or financial relationships that could be construed as a potential conflict of interest.

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Appendix B

Article II

Evaluating the destabilization susceptibility of active rock glaciers in the French Alps

Citation :

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Main Findings

- Rock glacier is a common phenomena on the region as 11.7% of the active landforms showed evidence of potential destabilization.
- Destabilization is more likely in specific terrain settings as north facing steep and convex slopes. Destabilization seems to occur the most at the lower limits of the permafrost zone.
- A large number of rock glaciers not showing evidence of destabilization are susceptible to encounter this process.



Evaluating the destabilization susceptibility of active rock glaciers in the French Alps

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Abstract. In this study, we propose a methodology to estimate the spatial distribution of destabilizing rock glaciers, with a focus on the French Alps. We mapped geomorphological features that can be typically found in cases of rock glacier destabilization (e.g. crevasses and scarps) using orthoimages taken from 2000 to 2013. A destabilization rating was assigned by taking into account the evolution of these mapped destabilization geomorphological features and by observing the surface deformation patterns of the rock glacier, also using the available orthoimages. This destabilization rating then served as input to model the occurrence of rock glacier destabilization in relation to terrain attributes and to spatially predict the susceptibility to destabilization at a regional scale. Significant evidence of destabilization could be observed in 46 rock glaciers, i.e. 10 % of the total active rock glaciers in the region. Based on our susceptibility model of destabilization occurrence, it was found that this phenomenon is more likely to occur in elevations around the 0 °C isotherm (2700–2900 m a.s.l.), on north-facing slopes, steep terrain (25 to 30°) and flat to slightly convex topographies. Model performance was good (AUROC = 0.76), and the susceptibility map also performed well at reproducing observable patterns of destabilization. About 3 km² of creeping permafrost, or 10 % of the surface occupied by active rock glaciers, had a high susceptibility to destabilization. Considering we observed that only half of these areas of creep are currently showing destabilization evidence, we suspect there is a high potential for future rock glacier destabilization within the French Alps.

1 Introduction

Warmer mean annual air temperatures (IPCC, 2013) are linked to a general trend in increasing permafrost temperature (e.g. Harris et al., 2003) and its water content (e.g. Ikeda et al., 2008), causing permafrost degradation, a phenomenon widely observed in the European Alps (Haeberli et al., 1993, 2010; Springman et al., 2013; Bodin et al., 2015). The occurrence of permafrost degradation is dependent on the ground properties, snow cover, and permafrost ice content (Scherler et al., 2013) and is therefore an heterogeneous phenomenon. Permafrost grounds affected by degradation experience a loss in strength due to the increasing ice ductility and reduced internal friction caused by the warmer ice and increasing water content (Davies et al., 2001; Haeberli et al., 1997; Harris and Davies, 2001; Nater et al., 2008; Huggel et al., 2010). Abnormal rockfall activity at high elevations (e.g. Ravanel and Deline, 2010) and increasing rock glacier displacement rates (Delaloye et al., 2008) are often assumed to be indicators of this change of state in the mountain permafrost. These processes may trigger mass movements that, in specific topographic conditions, may represent a hazard to alpine communities. Therefore, there is a growing need to understand the occurrence of these phenomena at a regional scale to allow for a targeted risk assessment and land use planning (Haeberli et al., 2010).

In this context, rock glaciers experiencing destabilization have recently become of interest. While active rock glaciers commonly present moderate interannual velocity variations that correlate with the ground temperature (Delaloye et al., 2008; Kellerer-Pirklbauer and Kaufmann, 2012; Bodin et al.,

2009), destabilized rock glaciers are characterized by a significant acceleration that can bring the landform, or a part of it, to abnormally high velocities (Delaloye et al., 2013; Roer et al., 2008; Scotti et al., 2016; Lambiel, 2011; Eriksen et al., 2018). During this acceleration phase, morphological features typical of sliding processes, such as crevasses and scarps, appear and grow on the rock glacier surface. This suggests that the destabilization occurrence is caused by a basal sliding process over the normal creep movement of rock glaciers (Roer et al., 2008; Schoeneich et al., 2015). In this sense, crevasses and scarps are interpreted as the possible transition between creep-driven sections and sliding sections of the landform (Roer et al., 2008). This destabilization phase, also referred as a “surge” (Schoeneich et al., 2015) or a “crisis” (Delaloye et al., 2013), may last decades and it usually results in a deceleration or inactivation of the landform. In very rare circumstances, destabilized rock glaciers may reach complete failure and collapse in a landslide (Bodin et al., 2016).

The destabilization process can be triggered by either mechanical forces or changes in climate. An overload on the glacier surface caused by a landslide or glacio-isostatic uplift can cause a compressive wave that propagates through the landform, increasing its displacement rates and leading to destabilization (Delaloye et al., 2013; Roer et al., 2008). A warmer climate may also trigger a destabilization crisis as increasing temperatures may cause permafrost degradation of the rock glacier. This process may result in the onset of water-saturated shear layers in which sliding occurs, triggering the crisis (Lambiel, 2011; Schoeneich et al., 2015; Eriksen et al., 2018). The onset of crevasses and scarps can also increase the predisposition of the landform to trap water percolating into the permafrost body, causing a positive feedback process of destabilization (Ikeda et al., 2008). Although triggers are necessary to the destabilization occurrence, not all rock glaciers subjected to these external forces destabilize. For example, permafrost degradation in rock glaciers mainly causes permafrost thaw and results in inactivation (Scapozza et al., 2010). Destabilization can be triggered only if there is a local topographical predisposition of the rock glacier to this process, such as steep slopes (Roer et al., 2008; Delaloye et al., 2013). Therefore, the terrain attributes of the rock glaciers to the onset of a destabilization phase are a critical parameter in the process occurrence.

The purpose of this study was to obtain regional-scale insights into the issue of destabilizing rock glaciers in the French Alps. Destabilization has been observed by several studies in the region (Echelard, 2014; Bodin et al., 2016; Serrano, 2017; Schoeneich et al., 2017); however, there has not yet been a comprehensive assessment of this phenomenon. This was carried out by (i) identifying the rock glaciers showing evidence of destabilization in order to provide an assessment of destabilized landforms, and by (ii) modelling the occurrence of this phenomenon in order to spot rock glaciers susceptible to incoming destabilization. Destabilized rock glacier identification was performed with multi-temporal aerial image interpretation based on expert field knowledge (Sect. 2.2). The geomorphological features typically occurring on destabilized landforms such as scarps and crevasses, here called “surface disturbances”, were mapped and used to assign a destabilization rating ranging from 0 to 3 to each active rock glacier (Sect. 2.2). Rock glaciers attributed with a higher destabilization rating have typical geomorphological characteristics reported in known cases of destabilization, including pronounced surface disturbances that increased by number and size in the past decades. These rock glaciers were suggested to be potentially destabilized while rock glaciers not presenting surface disturbances were classified with lower ratings of destabilization (i.e. stable rock glaciers).

The following step, i.e. modelling the destabilization occurrence, was performed by using a statistical approach that has been used for mapping landslide susceptibility (Goetz et al., 2011; Sect. 2.3). Potentially destabilized rock glaciers were used as destabilization evidence and their relation with terrain attributes (e.g. slope angle and elevation) was modelled using a generalized additive model (GAM). This model can be applied to better understand the relation between destabilization occurrence and terrain predisposition and to compute a destabilization susceptibility map, which provides an overview of potentially destabilizing landforms at a regional scale (Sect. 2.3.1). Strengths and limitations of the methodology, as well as the contribution of the study to enhancing our knowledge rock glacier destabilization, are widely discussed in the paper.

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2 Methods

2.1 Study area and rock glacier inventory

The French Alps cover an area approximately 50–75 km wide and 250 km long, located between 44 and 46° N and 5.7 and 7.7° W (Fig. 1). Apart from the noticeably high Mont Blanc massif (peaking at 4810 m a.s.l.), mountain ranges commonly peak between 3000 and 4000 m a.s.l. The lithology is heterogeneous across the region. The northern French Alps can be roughly divided into the west side, dominated by granite and gneiss (ranges of Mont Blanc, Belledonne, Écrins and Grandes Rousses), and east side, where ophiolites and schists are more common (ranges of Vanoise, Tabor and Mont Cenis). In the southern French Alps ophiolites, limestone and mica schists are the most common lithology (ranges of the Ubaye), while the crystalline range of Mercantour can be found at the southernmost end of the region. Dominant geology is described in the BRGM (2015) at 1/1 000 000 scale, and the vectorial version of this map is used in this study to observe destabilization occurrence in relation to lithology.

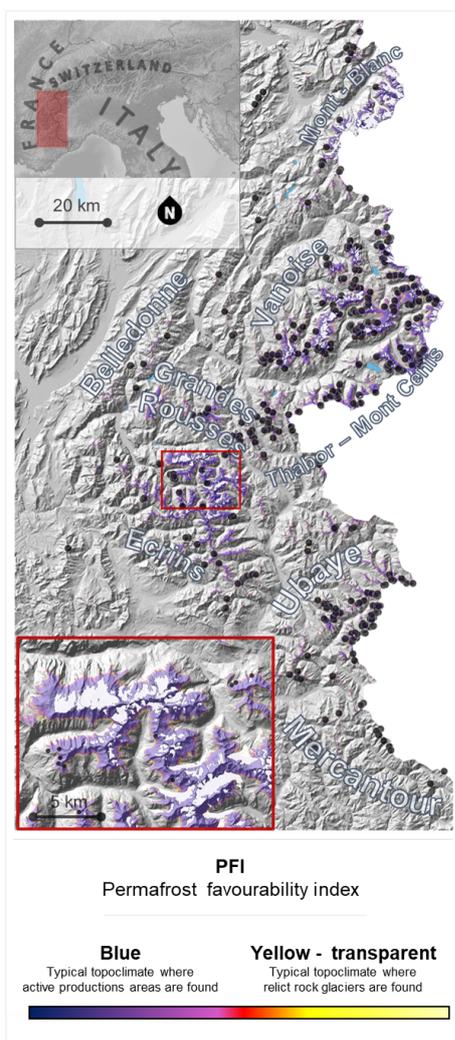


Figure 1. Identification of the study area in the European Alps and overview of the periglacial environment. Permafrost distribution is represented by the PFI map (Marcer et al., 2017). Black dots identify active rock glacier locations (Marcer et al., 2017).

In this region permafrost was estimated to cover up to 770 km² (Boeckli et al., 2012; Marcer et al., 2017). The 0 °C annual isotherm at the end of the 20th century ranged from 2500 m a.s.l. in the south to 2300 m a.s.l. in the north (Gotardi, 2009). The periglacial landforms of the region were inventoried by the “Office national des forêts” (ONF: the National Forest Office) (Roudnitska et al., 2016) and revealed the high presence of active rock glacier in the region (i.e. 493 mapped rock glaciers). This inventory was compiled between the years 2009 and 2016 by inspecting aerial imagery and revised by Marcer et al. (2017). This inventory was used in the present study to identify active rock glacier locations and to investigate the occurrence of destabilization.

According to Auer et al. (2007), mean annual air temperature increased by up to 1.4 °C in the French Alps during

the 20th century, and this rate has been increasing in recent decades. This climate warming is suspected to have caused some noticeable effects on the permafrost characteristics in the region. The only deep permafrost borehole in the region, located in the Écrins massif in temperate permafrost (−1.3 °C) with low ice content, showed a temperature increase rate of 0.04 °C decade^{−1} between 2010 and 2014 (Schoeneich et al., 2012), similar to other sites in Switzerland where data series are longer (PERMOS, 2016). Increasing air temperature was also addressed to be responsible for the acceleration since the late 1990s of the active Laurichard rock glacier located in the Combeynot massif of the French Alps (Bodin et al., 2009). Several cases of rock glacier destabilization, such as the collapsed Bérard rock glacier (Bodin et al., 2016) and the Pierre Brune rock glacier (Echelard, 2014), were also observed in the region. Serrano (2017) mapped destabilized rock glaciers in the Maurienne valley, Vanoise National Park and Ubaye valley, highlighting the high incidence of destabilized rock glaciers in these areas.

2.2 Mapping rock glacier destabilization

The first step to identify destabilized rock glaciers was mapping surface disturbances on rock glaciers. Previous studies that described destabilized rock glaciers showed that these landforms present a wide variety of geomorphological features (e.g. Roer et al., 2008). Here, we followed a methodology similar to that of Serrano (2017), which consisted of defining a catalogue of typical surface disturbances that can be found on destabilized rock glaciers. Surface disturbances on rock glaciers were classified in three distinct categories, depending on their morphology: cracks, crevasses and scarps. Surface disturbances are described in detail in Table 1 and illustrated in Fig. 2.

In this study, surface disturbances were mapped for the inventoried rock glaciers based on interpretation of a set of multi-temporal high-resolution aerial imagery for the French Alps. This orthoimagery collection was obtained from the Institut géographique national (IGN, National Institute of Geography), which is freely available from the official website (<https://www.geoportail.gouv.fr/>, last access: 10 December 2018) or can be accessed as a Web Map Service (IGN, 2011, 2013). The IGN orthoimagery collection consists of orthomosaics covering all of France for three different collection periods. The first orthomosaic is composed of images taken from 2000 to 2004, the second from 2008 to 2009 and the third from 2012 to 2013. All images are of high resolution: 50 cm × 50 cm for the most recent mosaic and slightly lower values (1 m × 1 m at its lowest) for the older mosaics, depending on the location. This resolution was sufficient to identify the smallest features to be mapped, i.e. the surface cracks (Fig. 2a). Nevertheless, several limitations during the mapping process, such as image distortion or illumination, were encountered and will be discussed in Sect. 4.4.1.

Table 1. Description of surface disturbance features that could be observed in the field or from orthoimagery to identify signs of rock glacier destabilization.

| Feature | Description |
|----------|--|
| Cracks | These are shallow linear incisions in the surface of an active rock glacier where a strain is applied (called “scars” in Roer et al., 2008). Cracks can be several tens of metres long and occur either individually or in a great number, being spaced out by only a few metres. In this case we define the feature as a “crack cluster” (translated from Serrano, 2017). Their proximity and shallowness led to the assumption that they affect only the active layer of the landform. Nevertheless, this feature was found to be largely predominant on the Lou (Schoeneich et al., 2017), Signal de l’Iseran (Serrano, 2017) and Tsaté-Moiryl (Roer et al., 2008; Lambiel, 2011) rock glaciers and therefore considered of interest in the context of the study. |
| Crevasse | These deep transverse incisions on the rock glacier surface can range in length from several metres to the entire landform width (Avian et al., 2005; Delaloye et al., 2008; Roer et al., 2008). Their depth is substantially larger than the active layer thickness, suggesting the presence of a shear plane sectioning the frozen body. Crevasse may be isolated or grouped. Spectacular crevasse can be found on Pierre Brune rock glacier (Fig. 1), where they are up to 7 m deep and 10 m wide, cutting across the entire landform (about 150 m). Similar dimensions are reported in the Furggwanghorn rock glacier (Roer et al., 2008). |
| Scarps | Scarps are described by Scotti et al. (2016) and Delaloye et al. (2008) as steep slopes (30 to 40°) several metres high, transversally cutting the entire rock glacier. Scarps are associated with deep shear planes that disconnect the rock glacier into two bodies that creep at different rates. Their activation is associated with a sudden acceleration of the downstream portion of the landform. One of the biggest scarps observable in the region is the one on Roc Noir rock glacier (Serrano, 2017). This S-shaped scarp, 20–30 m high and 40–45° steep, transversally cuts the whole landform (120 m) and the downstream lobe creeps about twice as fast as the upper part. |

Using a single orthoimage to map surface disturbances can lead to misinterpretations in the case of poor illumination of the terrain and snow patches covering the ground (Serrano, 2017). Indeed, as the surface morphology of a rock glacier is naturally shaped according to spatially varying creep patterns, it is easy to mistake actual surface disturbances related to compression features, such as furrows, depending on image quality. Therefore, surface disturbances, i.e. those morphological features not related to the creeping of the ice-rich permafrost, were mapped using all three available orthoimages in order to check that actual strain occurred where surface disturbances were located and to overcome limitations related to poor quality of an individual image.

Rating the degree of destabilization

After the rock glacier surface disturbances were mapped, a rating of the degree of destabilization was assigned to each rock glacier. This rating was given not only to provide some insight into the observed levels of destabilization in the French Alps, but also to provide a confidence rating to describe a rock glacier as stable or unstable for the spatial distribution modelling of rock glacier destabilization.

Assigning a rating to quantify the degree of destabilization of a rock glacier required the definition of the characteristics of the “typical” destabilized rock glacier that can be observed in multiple orthoimages. To do so, we investigated the features of destabilized rock glaciers reported in the literature that could be observed by orthoimagery interpretation. At first, it was observed that the presence of surface distur-

bances was a necessary but not sufficient condition to the occurrence of destabilization, as rock glaciers may present surface disturbances but be stable for decades. For example, in the Pierre Brune, Roc Noir and Hinteres Langtalkar rock glaciers, although crevasse could be observed in aerial imagery since the 1940s to the 1960s, destabilization occurred only in the late 1990s (Echelard, 2014; Serrano, 2017; Roer et al., 2008). Second, the destabilization process can be linked to an increase in surface disturbance occurrence (see Fig. 3). Also, surface disturbances on destabilized landforms were observed to create a discontinuity in the creep pattern. For example, the Plator, Grosse Grabe and Gänder rock glaciers have gone through a sharp transition from displacement speeds on the order of 0.1–0.9 m yr⁻¹ to displacement speeds of the order of several metres per year (Scotti et al., 2016; Delaloye et al., 2008). Finally, a high displacement rate may not be a necessary feature, as some destabilized rock glaciers, e.g. Lou and Furggwanghorn, moved at a “normal” rate of around 2 m yr⁻¹ (Schoeneich et al., 2017; Roer et al., 2008).

These observations suggest that destabilization may be spotted in orthoimages if the landform has surface disturbances increasing over time by frequency and/or magnitude, as well as if disturbances also create a strong discontinuity in the deformation pattern of the landform. Nevertheless, rock glaciers were observed to show a wide variety and combination of these features, making it unrealistic to construct a binary classification of stable versus destabilized landforms. In order to acknowledge this, we proposed a rock glacier destabilization rating based on four rates that

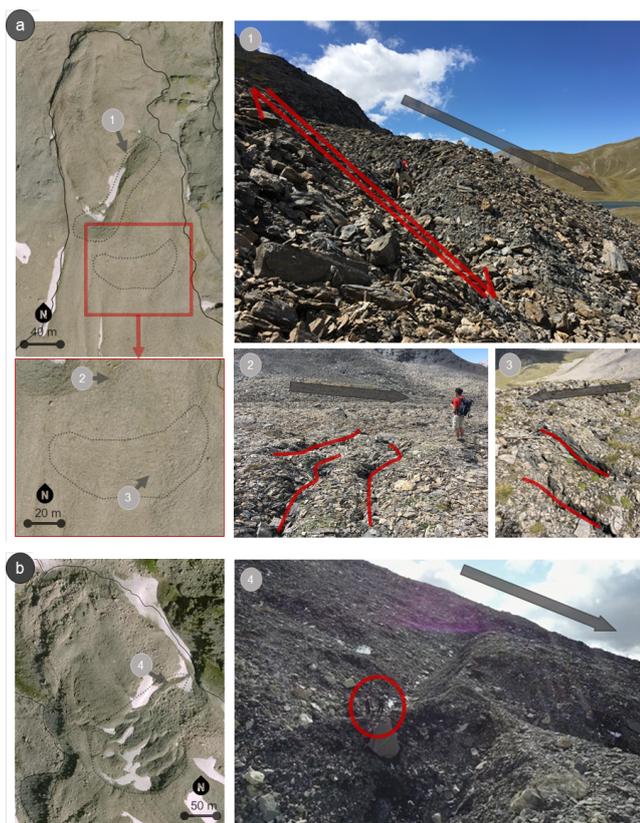


Figure 2. Examples of surface disturbances observable in the available orthoimages of 2013 in comparison to field observations on (a) Roc Noir (Serrano, 2017) and (b) Pierre Brune (Echelard, 2014) destabilized rock glaciers. The black arrows indicate the rock glacier displacement direction. A scarp (1) and cracks (2, 3) have been observed on the Roc Noir rock glacier. Large crevasses (4) can be seen on the Pierre Brune rock glacier. The dotted black lines indicate how the surface disturbances were mapped on these orthoimages.

varied from 0 (stable rock glaciers) to 3 (rock glaciers potentially destabilized), which is explained in more detail in Table 2. For each active rock glacier, a rating of the degree of destabilization was assigned by observing the combination of surface disturbances and a qualitative assessment of recent deformation patterns. This rating was applied using a standardized workflow (Fig. 4). A comparison of the available IGN multi-year orthoimagery was used to observe the temporal evolution of the surface disturbances and surface deformation patterns.

Potentially destabilized rock glaciers were then classified into two different categories according to the type of surface disturbances observed. Most of the destabilization cases observed by previous studies described rock glaciers characterized by surface disturbances that may reach several metres of depth, i.e. crevasses and scarps, and therefore suggested splitting the permafrost body. These surface disturbances

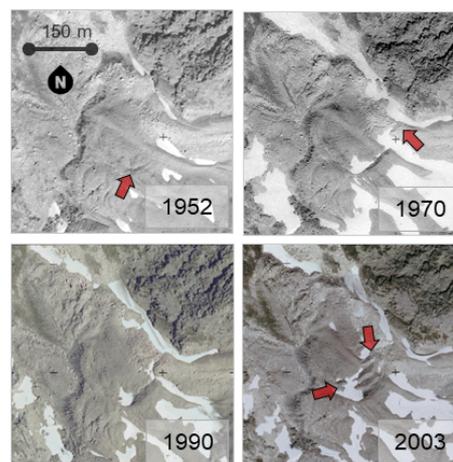


Figure 3. The evolution of the destabilization of the Pierre Brune rock glacier. The destabilization evidence, in this case a crack observable since 1952, evolved to a crevasse, observable in 1970. Afterwards, the landform was stable for 20 years as destabilization evidence did not further evolve. Between 1990 and 2003 the rock glacier experienced severe destabilization with the formation of new crevasses and a scarp at the location of the 1952 crack.

were mostly observed in coarsely grained (i.e. blocky; sensu Ikeda and Matsuoka, 2006) rock glaciers. Nevertheless, in the French Alps many active rock glaciers are finely grained, and some destabilization cases, e.g. the Lou (Schoeneich et al., 2017) and Iseran (Serrano, 2017) rock glaciers, were observed to be characterized by the presence of cracks only. These surface disturbances are shallower than crevasses and scarps and are therefore suggested to affect only the upper layer of the rock glacier. As these observations were relatively recent, at present there is still not enough knowledge concerning the significance of these shallow cracks in the context of rock glacier destabilization. We therefore decided to separate rock glaciers showing shallow surface disturbances from rock glaciers showing deep surface disturbances. This distinction was made to make the reader aware of this gap in knowledge.

2.3 Modelling rock glacier stability

Modelling the rock glacier stability aims to identify the terrain attributes that may precondition rock glacier destabilization. The modelling followed a statistical approach similar to previous studies on landslides (Goetz et al., 2011) and arctic permafrost slope failures (Rudy et al., 2017) that used the GAM with logistic link function (R package “mgcv”). The GAM was selected because of its flexibility in modelling non-linear interactions between dependent and predictor variables. The logistic link function allows us to model the occurrence of a categorical response variable as a function of continuous variables (predictor variables). All numeric predictors were represented using spline-based

Table 2. Rating classes used to describe rock glacier destabilization.

| Rating | Label | Description |
|--------|---|--|
| 3 | Potential destabilization, potentially destabilized rock glaciers | Surface disturbances are well recognizable and evolve in time, increasing in number and/or size. The deformation pattern of the rock glacier is discontinuous and some sectors move significantly faster than others. The source of the discontinuity may be located at the rock glacier’s root and the whole landform may be affected by destabilization. Deformation pattern discontinuities are sharp and coincide with the presence of surface disturbances. Sectors moving appreciably faster may also present a series of surface disturbances. If the dominant surface disturbances are deep (i.e. crevasses and scarps), then it is attributed the rating <i>3a</i> . If the dominant surface disturbances are shallow (i.e. crack and crack clusters) then it attributed the rating <i>3b</i> |
| 2 | Suspected destabilization | In these landforms the surface disturbances are well recognizable and evolve in time, by increasing in number and/or size. The velocity field is continuous, i.e. there are no abrupt spatial differences in the velocity field. If there are sectors moving faster than others, their transition is smooth |
| 1 | Unlikely destabilization | In these landforms surface disturbances do not appear to evolve in time. The rock glacier presents a continuous deformation pattern, with no sectors moving substantially faster than others. |
| 0 | Non-observable destabilization | Active rock glaciers not presenting surface disturbances are considered as stable. |

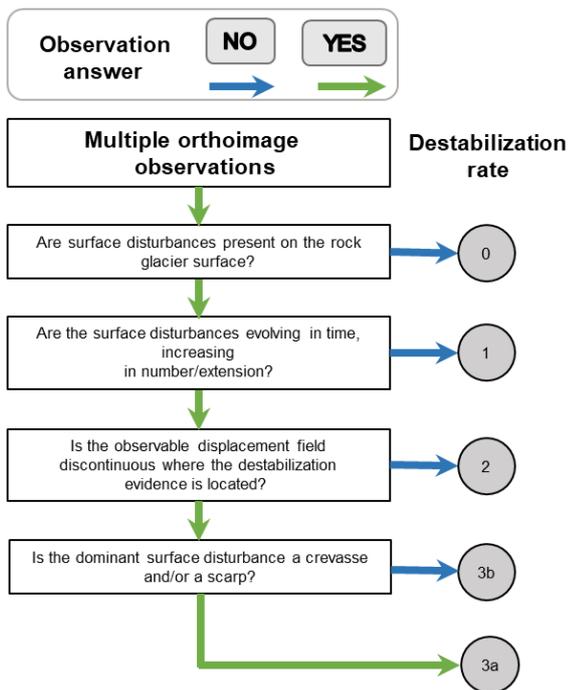


Figure 4. General pipeline used to rate rock glacier destabilization by observing surface disturbances and the qualitative displacement field. Higher destabilization ratings indicate potentially unstable rock glaciers, while lower ratings indicate stable rock glaciers.

smoothing, for which we chose a maximum basis dimension of 4 in order to limit their flexibility and reduce overfitting. The actual degree of smoothness of the splines was deter-

mined using a generalized cross-validation procedure (Wood, 2017).

In this study, rock glacier stability was hypothesized to be preconditioned by a series of local terrain attributes. In particular, rock glacier destabilization grouped by either presence or absence was used as the response variable, while terrain attributes describing local topography and climate were used as predictor variables. Multiple-variable models were computed using different combinations of predictor variables. Different models were compared using the Akaike information criterion (AIC), which is a measure of goodness of fit that penalizes more complex models. The best multiple variable model was selected by iterating a backward-and-forward stepwise variable selection, aimed at identifying which combination of predictors was better at describing the response variable by means of a lower AIC. Finally, the best model performance was estimated using the area under the receiver operating characteristic (AUROC) (Hosmer and Lemeshow, 2000). The AUROC estimates the ability of the model to discriminate stable and unstable areas.

The predictive power of the model was estimated with spatial cross-validation (R package *sperrorest*). The method selected was the *k*-means clustering, which consisted of dividing the mapped data in *k* spatially contiguous clusters (Ruß and Brenning, 2010a). All but one cluster were used to train the model, while the remaining cluster was used to test the predictive power of the model. This process was repeated until each cluster was used at least once in both training and test sets. Here, we divided the database into *k* = 5 clusters of equal size per run and used 100 repetitions. Performance indicators were evaluated for the respective test sets, and the

overall model performance was evaluated using the average and standard deviation over all partitioning clusters.

The variable importance was assessed using permutation-based variable importance embedded in the spatial cross-validation (Ruß and Brenning, 2010b). This method consisted of permutating the values of each predictor variable one at a time and calculating the reduction in model performance caused by the permutations. A total of 1000 permutations were performed for each spatial cross-validation repetition. Predictor variables causing higher deviations while permuted were considered the most important ones in the model.

2.3.1 Response variable

Surface disturbances of potentially destabilized rock glaciers were used as evidence of creeping permafrost destabilization. This was performed under the hypothesis that surface disturbances were the geomorphological expression of rock glacier destabilization. Although many surface disturbances could be observed on rock glaciers that were classified as unlikely destabilized or as suspected of destabilization, potentially destabilized rock glaciers could be observed to increase surface disturbances over time by number and size, creating a discontinuity in the deformation pattern, which provided stronger evidence of destabilization. Therefore, only surface disturbances located in potentially destabilized rock glaciers were considered to be solid evidence of rock glacier destabilization.

As surface disturbances were digitized as linear features, they were buffered and merged into an “unstable areas” polygon database. A buffer distance of 30 m was chosen. The model was found to be insensitive to changes in buffer size up to 90 m. All remaining areas within the polygons of stable and likely stable rock glaciers were used as “stable areas”. Polygons of both unstable and stable areas were sampled using a 25 m × 25 m point grid in order to assign the response variable to the modelling database. The point values were then used as binary response variables with values of 0 for stable areas of (likely) stable rock glaciers, while 1 was assigned for unstable areas of potentially destabilized rock glaciers in the modelling stage.

Since the rock glacier inventory counted a relatively small number of potentially destabilized cases (46 individuals), selecting only one point per rock glacier would have caused large uncertainty in the model outcome. Therefore, a simple exploratory analysis was performed to identify a suitable number of points per rock glacier to be used for modelling. Multiple points from one to 10 were randomly selected within each rock glacier perimeter and used to compute a model. This was repeated 10 times per point sample size to measure the variability in the model performance in relation to the point sample size. Since the model performances were found to stabilize for more than five points selected per rock glacier, the number of points randomly ex-

tracted per rock glacier used for modelling was five. Overall, the model was computed using 225 points with evidence of instability and 1785 points with evidence of stability.

2.3.2 Predictor variables

Terrain attributes used in modelling needed to be selected to act as proxies for processes that precondition destabilization. Although destabilization is found to occur in different conditions, some topographical features seem to be recurrent. Destabilization has been observed to occur on steep slopes, as high slope angles tend to increase the internal shear stress (Delaloye et al., 2013). Surface disturbances are often located in convex-shaped bedrock surfaces, which causes an extensive flow pattern and a thinning of the permafrost body (Delaloye et al., 2013). Solar exposure may also be significant in the destabilization occurrence since all known cases of destabilized rock glaciers in the French Alps are north facing. Solar exposure can also be a proxy of the snow cover duration, as north-facing slopes are more prone to conserve longer snow patches through the summer, making meltwater available through the summer. Elevation and mean annual air temperature can also be proxies of snow cover duration that have the possibility of affecting permafrost characteristics. Considering this, slope angle, profile curvature, potential incoming solar radiation (PISR) and elevation were tested as predictor variables.

Terrain attributes were derived from the BD ALTI DEM, 25 m × 25 m spatial resolution (IGN, 2011). Slope angle and downslope curvature (Freeman, 1991) were evaluated using the Morphometry Toolbox in SAGA GIS (version 2.2.2). Negative values of curvature indicate concave topography, while positive values indicate convex topography. Also, PISR was calculated using the Terrain analysis toolbox in SAGA as the sum of the computed direct and diffusive components of the radiation (Wilson and Gallant, 2000). Clear-sky conditions, a transmittance of 70 % and absence of a snow cover were assumed in the calculation of the annual total PISR. Finally, it was decided to evaluate the relation between rock glacier destabilization and the spatial distribution of degrading permafrost in order to give insight into the significance of the warming climate with respect to the destabilization phenomena. The spatial distribution of degrading permafrost was evaluated following the method already presented by other studies (Hoelzle and Haeberli, 1995; Lambiel and Reynard, 2001; Damm and Felder, 2013), which consisted of artificially shifting a permafrost map proportionally to the estimated climate warming occurring between the period of validity of the map and the current climate. Here, we used a permafrost favourability index (PFI) map (Marcer et al., 2017) to act as a permafrost distribution map for the region. The PFI map was calibrated using active rock glaciers as evidence of permafrost occurrence, and it represents the permafrost conditions during the cold episodes of the Holocene, e.g. Little Ice Age (LIA). The climate warming between the

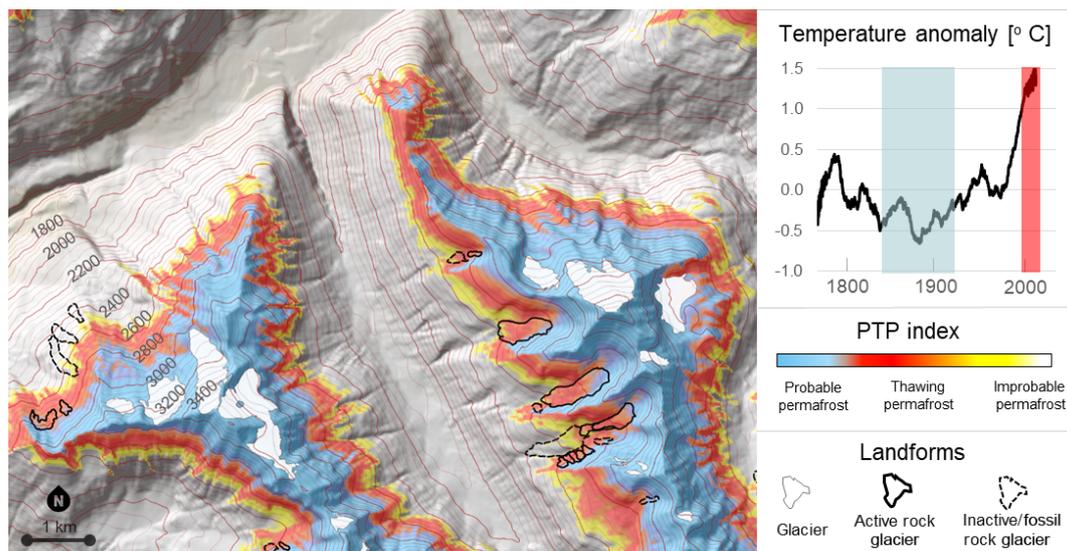


Figure 5. Map of potential thawing permafrost (PTP) distribution in the Mont Cenis range, indicating the extent of the permafrost zone not in equilibrium with the present climate (red coloured areas). Temperature warming to compute the map is evaluated using HISTALP data (Auer et al., 2007) between the end of the Little Ice Age (light blue shaded period in the temperature anomaly plot) and the current climate (red shaded period).

years 1850–1920 and 1995–2005 was determined using the HISTALP database (Auer et al., 2007) over the region. A permafrost distribution map was then recomputed taking into account these temperature variations and represented the theoretical permafrost distribution in equilibrium with the current climate. By comparing this theoretical permafrost distribution and the PFI, a map of the potential thawing permafrost zone (PTP, i.e. the so-called “melting area” in Lambiel and Reynard, 2001) was obtained. In order to use the PTP as a predictor variable, it was represented by an index ranging between 0, i.e. no thaw expected, and 1, i.e. potential thaw.

It should be emphasized that PTP is only a proxy of permafrost degradation, which occurs at all the elevations, while the PTP zone consists of a belt of 250 to 300 m in elevation that affects about 50 % of the lower margins of the permafrost zone (Fig. 5). PTP is used under the hypothesis that degradation is more intense at the lower margins of the permafrost zone where permafrost conditions may be more temperate, richer in water and more sensitive to climate variations.

2.3.3 Susceptibility modelling

The model of rock glacier stability was also used to predict the occurrence of degrading permafrost over the French Alps by producing a susceptibility map (e.g. Goetz et al., 2011). This was carried out using the R package RSAGA and the raster images of the predictor variable maps, which allowed extrapolation of the relationships between rock glacier stability and terrain attributes at the landscape scale. We would like to highlight that since the model is constructed using data on destabilized rock glaciers, the susceptibility map ap-

plies mainly for processes relative to destabilization of ice-rich debris slopes. Therefore, in areas where creeping permafrost does not exist, the extrapolated susceptibility may have high uncertainty. The model predicted a DEFROST index, which was classified into five susceptibility zones using the 50, 75, 90 and 95 percentiles (Rudy et al., 2017; Goetz et al., 2011). These zones described very low (< 50), low (50–75), medium (75–90), high (90–95) and very high (> 95) susceptibility to permafrost destabilization.

3 Results

3.1 Destabilized rock glacier inventory

More than 1300 surface disturbances were digitized, involving 259 active rock glaciers (Fig. 6). Overall, more than the 50 % of the active rock glaciers may be affected by some degree of destabilization as 46 rock glaciers (9.7 %) showed potential destabilization, 86 (17.0 %) were suspected of destabilization and 127 (25.7 %) were unlikely destabilized. Only 13 potentially destabilized rock glaciers presented deep surface disturbances. Location and destabilization rate of each active rock glacier in the region is provided as a shapefile in the Supplement.

Potentially destabilized rock glaciers were mainly located in the Vanoise National Park and in the Queyras and Ubaye mountain ranges. In these areas, densely jointed lithologies (i.e. ophiolites and schists) dominate. Rock glaciers in crystalline lithologies (i.e. gneiss and granite) were found to have low destabilization ratings. That is, only two rock glaciers

Table 3. Number of rock glaciers per dominant lithology in relation to destabilization rate.

| Destabilization rate | Ophiolites | Schist | Sandstone | Mica-schist | Gneiss | Granite | Limestone | Totals |
|----------------------|------------|--------|-----------|-------------|--------|---------|-----------|--------|
| 0 | 47 | 88 | 21 | 11 | 31 | 3 | 32 | 233 |
| 1 | 39 | 37 | 11 | 3 | 13 | 2 | 22 | 127 |
| 2 | 33 | 28 | 5 | 0 | 1 | 1 | 18 | 86 |
| 3a | 5 | 2 | 1 | 0 | 0 | 0 | 5 | 13 |
| 3b | 18 | 7 | 1 | 3 | 0 | 0 | 4 | 33 |

Table 4. Number of rock glaciers per destabilization rating showing a specific surface disturbance.

| Destabilization rating | Cracks | Crack clusters | Crevasses | Scarps |
|------------------------|--------|----------------|-----------|--------|
| 1 | 86 | 54 | 13 | 8 |
| 2 | 52 | 51 | 15 | 11 |
| 3a | 10 | 9 | 10 | 8 |
| 3b | 23 | 29 | 0 | 0 |
| Totals | 187 | 152 | 40 | 27 |

were rated as possibly destabilized over a population of 55 (Table 3).

The predominant surface disturbance observed was cracks, which were present in 187 of the active rock glaciers (Table 4). Crack clusters also had a high number of observed cases (152), while the deep surface disturbances occurred in about 15 % of all the examined rock glaciers. In general, the occurrences of surface disturbances were dependent on the destabilization rating. Scarps and crevasses were found in about 10 % of unlikely destabilized landforms. The observation of each surface disturbance was highest for potentially destabilized rock glaciers with deep surface disturbances, indicating that in these landforms multiple surface disturbances coexist.

3.2 Modelling

Following a stepwise backward and forward selection, the chosen model included PISR, slope angle, elevation and curvature as predictors. The mean cross-validated AUROC was 0.76 on the test set, indicating a good performance (Hosmer and Lemeshow, 2000). The predictors having the most influence on the response variable were the PISR (AUROC change = 0.162), curvature (AUROC change = 0.068), slope angle (AUROC change = 0.031) and elevation (AUROC change = 0.018).

The model transformation functions revealed the relations between terrain attributes and rock glacier stability (Fig. 7). Higher predisposition to destabilization was more likely to occur in an altitudinal range between 2700 and 2900 m a.s.l. and slope angles ranging between 25 and 30°. Slightly nega-

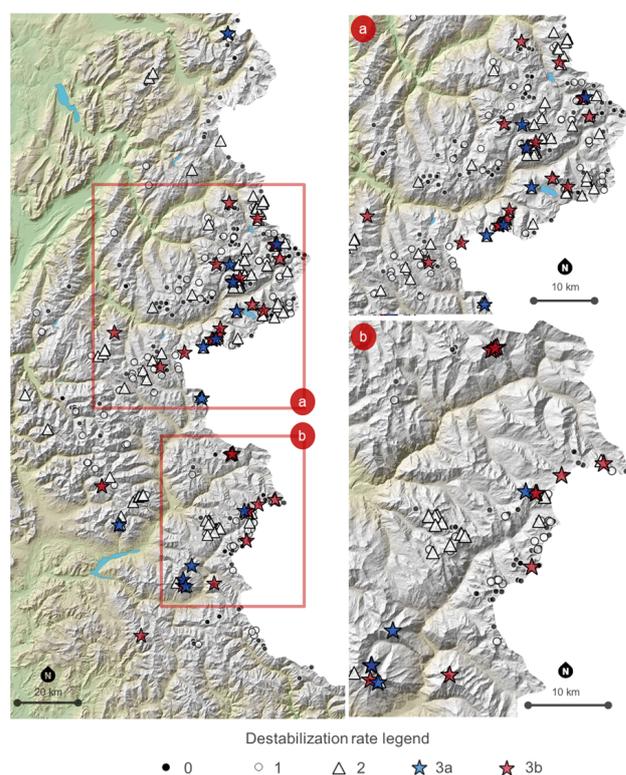


Figure 6. Map of active rock glaciers in France by rock glacier destabilization rating, with focus on the (a) Vaonise–Mont Cenis and (b) Ubaye ranges as most of potentially destabilized landforms were observed in these areas.

tive to positive curvature was also favourable to destabilization. PISR was negatively correlated with the destabilization probability, indicating that rock glacier destabilization was more likely to occur on north-facing slopes. The relation between PTP and destabilization was also explored by including this predictor variable in the model instead of elevation. Although the PTP caused lower model performance, it could be observed that the PTP was positively correlated with the destabilization.

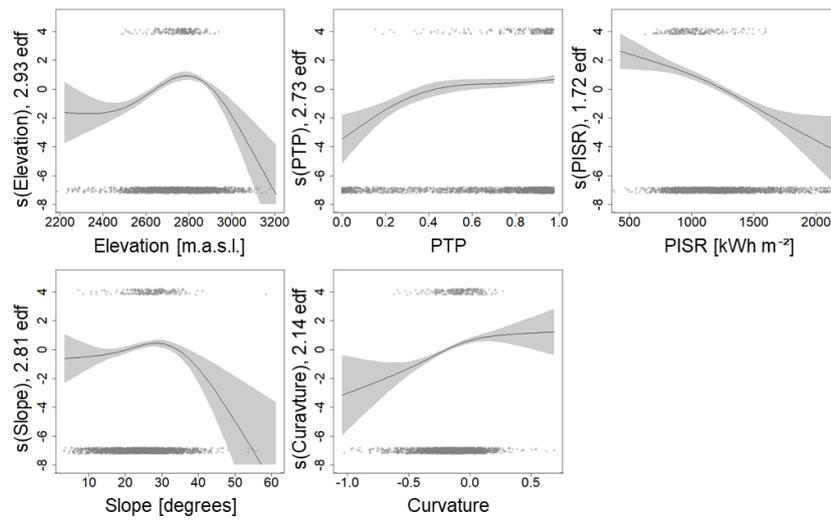


Figure 7. Transformation function plots of the GAM model showing the relationship between each predictor variable and destabilization occurrence. The data distribution with respect to predictor variables is indicated with dots on top (destabilization evidence) and on the bottom (stability evidence) of the plots. The y axis represents the transformation of the predictor variable by the GAM’s spline, indicated here by “s(predictor)”. The effective degrees of freedom are also reported. The PTP is presented here for explanatory purposes, although it was not included in the final model.

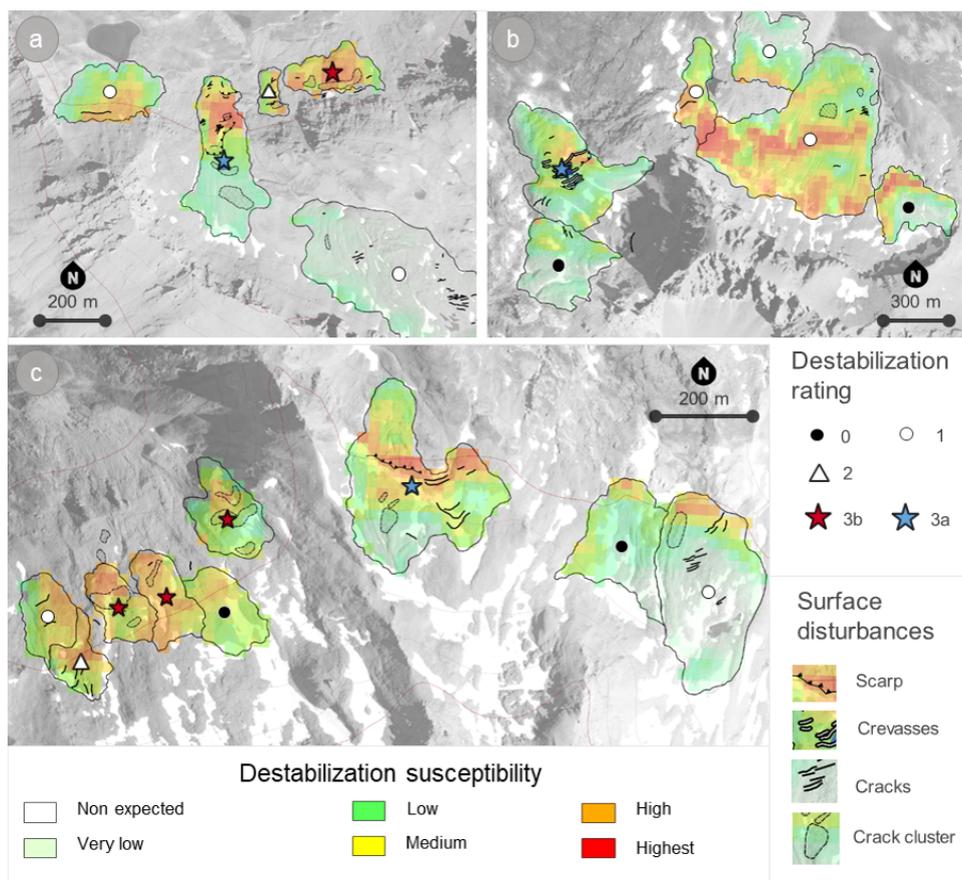


Figure 8. Examples of the susceptibility map in (a) Roc Noir, (b) Pierre Brune, and (c) Iseran and neighbouring rock glaciers. The susceptibility map successfully identifies instabilities observed on the potentially destabilized rock glaciers. Nevertheless, some predicted instabilities were observed in areas that appear stable by observing the orthomosaics.

3.3 Susceptibility map

The susceptibility map highlights creeping permafrost areas susceptible to destabilization based on regional-scale model predictions (examples shown in Fig. 8, and the full map is available in the Supplement). The susceptibility map reproduced the previously known cases of destabilization well. The destabilized areas of Iseran, Roc Noir and Pierre Brune were predicted to have a high susceptibility to destabilization, which matches field observations. In some cases, the susceptibility map predicted high destabilization susceptibility in areas belonging to stable rock glaciers.

Rock glacier surfaces were investigated with respect to each susceptibility class (Table 5). About 75 % of the creeping permafrost was found at low or very low susceptibility to destabilization. Creeping permafrost at high and very high susceptibility to destabilization accounted for 10 % of the total creeping permafrost surface, i.e. 2.9 km². While about one-third of this surface was located in potentially destabilized rock glaciers, more than 1.4 km² of stable and unlikely destabilized rock glaciers was found at high and very high destabilization susceptibility.

4 Discussion

4.1 Rating rock glacier destabilization

The present study provided the first comprehensive assessment of rock glacier destabilization for the French Alps and indicates the potentially high prevalence of this phenomenon. Destabilized rock glaciers were more likely located in the Vanoise, Queyras and Ubaye ranges. In these areas the densely jointed lithology was suspected to generate mainly pebbly rock glaciers (Matsouka and Ikeda, 2001; Ikeda and Matsuoka, 2006). This indicates that destabilization may be more likely to develop in pebbly rock glaciers, as observed in the Bérard, Roc Noir and Lou rock glaciers. Also, rock glaciers in crystalline lithology did not show signs of potential destabilization. However, recognizing surface disturbances on pebbly rock glaciers may be easier than in “blocky” rock glaciers, as smaller cracks are more evident. This may create a bias, which should be studied in more detail by investigating geomorphological features of destabilization occurring on blocky rock glaciers.

The majority of rock glaciers showing potential destabilization were characterized by shallow cracks (33 cases versus 13). Although this is suggested to be partially due to the high incidence of rock glaciers located in densely jointed lithology, there are a number of questions that still need to be answered in this context. At present, we are unsure about the significance of these surface disturbances in the context of destabilization. Cracks may be either “mild” evidence of destabilization as they affect only the upper layer of the landform, or a typical surface disturbance occurring on destabi-

Table 5. Active rock glacier area per class of destabilization susceptibility.

| Destabilization rating | Surface per susceptibility class (km ²) | | | | |
|------------------------|---|------|--------|------|-----------|
| | Very low | Low | Medium | High | Very high |
| 0 | 8.09 | 3.21 | 1.70 | 0.43 | 0.37 |
| 1 | 4.03 | 2.16 | 1.29 | 0.42 | 0.38 |
| 2 | 2.18 | 1.50 | 0.93 | 0.34 | 0.30 |
| 3a | 0.17 | 0.27 | 0.17 | 0.05 | 0.05 |
| 3b | 0.07 | 0.19 | 0.31 | 0.24 | 0.38 |
| Cumulative surface | 14.54 | 7.33 | 4.41 | 1.47 | 1.48 |

lized pebbly rock glaciers. In the first case, using cracks as destabilization evidence could lead to an over-interpretation of the destabilization severity of the landform. Conversely, it was observed that destabilization may occur when only these type of surface disturbances occurred (Schoeneich et al., 2017; Serrano, 2017). Concerning this issue, this study suggested that these landforms deserve more attention due to their high incidence in the regional territory.

Overall, rock glacier destabilization rating can be a relevant tool for the local authorities to focus monitoring efforts related to periglacial risk assessment, as we identified all rock glaciers presenting signs of destabilization in the region. The destabilization rating, if combined with an assessment of displacement rates and landform connectivity, could indicate the severity of the potential hazard and be used to help identify actions that should be undertaken to deal with the problem. In general rock glaciers with a low destabilization rating are currently evolving slowly or are stable, and consequently monitoring based on remote sensing may be sufficient. Suspected or potentially destabilized rock glaciers require more caution and in situ monitoring is recommended.

Uncertainties in rating rock glacier destabilization

A potential source of uncertainty in this study was the subjectivity that can occur while mapping surface disturbances and rating the degree of destabilization. These activities were based on expert knowledge; however, it is possible that mapping and rating results vary depending on the operator. For example, the operators in charge of the digitization process were requested to interpret surface features that in many cases have small dimensions with respect to the resolution of the orthoimages, making the identification challenging. Orthoimages can have varying illumination from one year to another, causing surface disturbances to change their appearance. Orthoimages may also be distorted, creating unrealistic deformation patterns of the rock glacier surface. Also, although surface disturbances were inventoried into the catalogue in an attempt to standardize the classification, destabilized rock glacier morphology is complex, and its identifica-

tion requires intense training. In many cases the boundaries between the different typologies proposed were not sharp. Personal knowledge of the process evolved through the inventory compilation, requiring various iterations to review the work.

Another issue was that the operator's metrics of judgment were subjected to the "prevalence-induced concept change" (Levari et al., 2018), as the classification might become stricter (or looser) when the operator deals with a series of destabilized (or stable) rock glaciers. The ratings were compiled and revised by different operators in an attempt to mitigate these effects. Some cases were the subject of debate, highlighting significant individual biases. These biases can influence the resulting susceptibility model (Steger et al., 2016). It is therefore strongly recommended to integrate the inventory with in situ observations when possible and to maintain a critical attitude towards the data. Currently, France does not have a lidar-based high-resolution DEM covering the study region. Such data could be used to revise the inventory in the future in order to reduce errors due to poor quality of the orthophotos. In particular, having a high-resolution DEM could allow us to avoid issues related to the differentiation between isolated crack and crevasse, as the judgment based on orthoimages may vary depending on the lighting.

Although observing aerial orthoimagery or high-resolution DEMs could not replace the relevance of a proper in situ survey, it provides us with data and resulting insights that would normally not be possible with in situ surveys alone, a characteristic that fitted with the aim of the study. Additionally, the use of orthoimagery has been proven to be a useful approach for mapping rock glacier surface disturbances by Serrano (2017), who compared the results of field observation to observations from orthoimagery. Although Serrano (2017) investigated a limited number of sites, those results were encouraging, showing that the method was relevant. The use of multiple orthoimages was believed to successfully reduce issues related to subjectivity and poor image quality in most of the cases. Observing the movements of the landforms was a valuable decision support tool, as surface disturbances could be related or not to discontinuities in a pronounced displacement field. Also, the use of multiples orthoimages reduced potential errors due to bad lighting that may enhance features that may be unrelated to destabilization processes (Serrano, 2017).

4.2 Modelling the predisposition to rock glacier destabilization

Despite the various limitations of the data, the results were encouraging. The spatially cross-validated model had a good performance, suggesting that the method is valuable in the context of modelling rock glacier stability. The relationships with predictor variables were found to be consistent with topographic settings observed in known cases of destabiliza-

tion. High slope angles are suggested to increase internal shear, making the landform more susceptible to destabilization (Schoeneich et al., 2015). Convex slopes cause an extensive flow pattern as creep velocity is higher downslope from the convexity (Delaloye et al., 2013). This suggests that a thinning of the permafrost body and the generation of traction forces may intensify the occurrence of surface disturbances.

PISR had the most importance in the model, suggesting that rock glacier destabilization was primarily more likely to occur on north-facing slopes. We cannot offer a convincing explanation of this phenomenon since, at the present state of the art, there is no systematic study comparing rock glacier characteristics in relation to their solar exposure. Nevertheless, we suggest that a possible explanation resides in the variability in meltwater input of the rock glaciers with respect to solar exposure. Ikeda et al. (2008) suggest that high water input can boost destabilization by reducing internal friction. Considering that snow patches tend to last longer on north-facing slopes, meltwater inputs may be more significant than on south-facing slopes.

Modelling rock glacier destabilization using PTP instead of elevation revealed that an increasing potential in permafrost thaw was linked to an increase in susceptibility to destabilization, indicating that destabilization was more likely to occur where the permafrost zone was expected to be thawing. This seems to be consistent with the relationship between destabilization and elevation, as potentially destabilized rock glaciers are more often located around 2800 m a.s.l., which roughly coincides with the lower margins of the regional permafrost zone.

4.3 Susceptibility map

Overall, permafrost destabilization was adequately described, as indicated by the cross-validated performance, in most of the observed cases of destabilization. Although cases of potential destabilization were inventoried, rock glaciers that have a low rating of destabilization and are located in areas with high susceptibility should be identified as having a high potential of future destabilization. Results indicated that these rock glaciers had a large area of high predisposition to destabilization, suggesting that there is a high potential for future destabilization in the region. The map may therefore be used to spot rock glaciers that present a predisposition to develop destabilization. In particular, the Laurichard rock glacier is a site currently under monitoring and was found to present a low to medium susceptibility to destabilization in this study (Bodin et al., 2008). The comparison of the future evolution of this landform with respect to the susceptibility map is therefore recommended.

5 Conclusions

The present study aimed to give insights into the extent of destabilizing rock glaciers in the French Alps. Mapping and modelling rock glacier destabilization in this region was conducted using an orthoimagery collection, a 25 m × 25 m resolution DEM and statistical modelling. This methodology carried several limitations, due to subjectivity and modelling issues. Therefore, absolute model performance and the appearance of the susceptibility map may not be exact, and further work is strongly encouraged. Integrating the observations with a high-resolution lidar DEM and with new field-observations could spot possible systematic biases in the destabilization rating attribution and significantly reduce uncertainty.

Despite the limitations of this methodology, the study contributes to the knowledge related to permafrost degradation in the French Alps. Rock glacier destabilization potentially involves 46 active landforms, uniquely located in non-crystalline lithologies, which are typically densely jointed as ophiolites and schist. Shallow surface disturbances (i.e. cracks) had the highest incidence in potentially destabilized rock glaciers. At present, there are several questions concerning the destabilization of pebbly rock glaciers presenting these shallow surface disturbances, as only a few studies tackled the subject. Therefore, considering the high incidence of these landforms in the region, it is suggested to dedicate more attention to these issues in the future.

The destabilization of creeping permafrost was found to be a widespread phenomenon that involves more than 10 % of the total surface of active rock glaciers, i.e. 3 km². Only half of this surface was attributed to rock glaciers currently showing a relevant degree of destabilization, suggesting that several stable rock glaciers have a significant degree of susceptibility to experience destabilization in the future. Rock glacier destabilization was found to more likely occur at the lower margins of the permafrost zone, i.e. where permafrost thaw due to climate warming is expected to be more intense. This suggests that climate warming may have increased the predisposition of creeping permafrost to slope failure. In this context, the present study contributes by having mapped potentially destabilized rock glaciers and areas considered susceptible to destabilization, allowing us to focus future monitoring efforts. In this sense, we suggest that the modelling framework proposed is relevant and further efforts to better acknowledge the phenomena are strongly encouraged.

Code and data availability. Shape files of destabilization ratings at rock glacier locations and destabilization susceptibility maps are available in the Supplement (.tiff format). Data are referenced in EPSG: 2154.

Supplement. The supplement related to this article is available online at: <https://doi.org/10.5194/tc-13-141-2019-supplement>.

Author contributions. MM, CS, XB and PS conceived the project and collected and interpreted data. MM, AB and JG designed the modelling approach. MM wrote the paper. All authors contributed to discussion and editing. AB and JG provided feedback on the writing.

Competing interests. The authors declare that they have no conflict of interest.

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Appendix C

Article III

Investigating the permafrost slope failures at the Lou rock glacier front, French Alps.

Citation :

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Main Findings

- The frontal active layer of the Lou rock experienced two localized failures due to supra-permafrost concentrated flow. One of the failures was associated to a novel event and located on a destabilized area of the rock glacier.
- The failures were triggered after a peculiar sequence of a strong heat wave followed by a rainy period characterized by mild but continuous precipitations.
- The debris flow that stroke the town of Lanslevillard was due to poor management of the river infrastructures and underestimation of the solid transport in the stream.

Investigating the permafrost slope failures at the Lou rock glacier front, French Alps

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ABSTRACT

On August 14th 2015 a large debris flow initiated by the occurrence of two slope failures at the front of the Lou rock glacier flooded part of the town of Lanslevillard, France. The present study provides a back analysis with the aim of understanding the climatic and geomorphological context that led to these failures. Investigations were conducted by combining data on climate, surface movements and geophysical transects. The analysis indicates that the Lou rock glacier is directly connected to an active torrential channel and has a natural predisposition to frontal failure due to the steepness of its front. The slope failures were triggered after a heat wave followed by a three-week period of almost continuous rainfall. Water flowing on top of the permafrost table was observed in the two head scarps, suggesting that regressive erosion consecutive to this concentrated subsurface water flow triggered the failures. For one of the slides, traces of previous failures were observable on historical aerial imagery dating back to the 1950's, while the second slide corresponded to a novel event and developed on the frontal slope of a fast moving and destabilized rock glacier lobe. We also discuss the increase in local predisposition to failure related to the remarkable morphological modifications such as frontal advance and development of surface cracks associated with the lobe destabilization.

1 Introduction

The town of Lanslevillard was hit by two consecutive debris flows on August 14th 2015, at 15h30 and 16h45 (Figure 1). The debris volume that reached the town was estimated to be around 15 000 m³, causing more than 100,000 of damage. The debris flow was triggered in the headwaters of the Arcelle Neuve stream which reaches the Arc river valley in the western part of

28 Lanslevillard. Before reaching the Arc river, the waters of the Arcelle Neuve stream are channelled into a pipe which passes
29 below the local road RD 1106 (Figure 1c). During the event, the large amount of material carried by the flow rapidly clogged
30 the pipe, causing flooding. The overflow damaged vehicles and equipment located in the Vieux Moulin cable car station and
31 buried a restaurant cellar under two meters of debris. At that point the flow followed the local road RD 1106 towards Les
32 Champs, an inhabited hamlet. Using an excavator, an emergency dam was quickly created and the flow was deviated into the
33 Arc River.

34 On August 15th 2015, a helicopter survey carried out by the National Environmental Protection Agency (RTM) located
35 the initiation points of the debris flows on the front of the Lou rock glacier, an active landform located in the upper part of
36 the Arcelle Neuve watershed and which was previously unknown to the local authorities (Figures 2, 3). The initiation points
37 consisted of two slope failures whose upper limits were defined by pronounced head scarps located near the front of the rock
38 glacier. Water springs were observed in both head scarps, a few meters below the surface of the rock glacier (Figure 3b). On
39 August 25th 2015 an in-situ survey took place in order to observe the nature of the initiation points and the characteristics of the
40 rock glacier. The eastern initiation point was characterized by an active water spring despite the dry period after the event. The
41 head scarp was up to 15 meters wide and 20 meters long. The scarp was divided into two minor gullies, reaching a depth of 3.5
42 meters (Figure 3). This failure therefore presented characteristics similar to those observed in the Dirru, Gugla and Tsarmine
43 rock glaciers (Swiss Alps¹), and which are described as resulting from large “concentrated flow”, i.e. linear regressive erosion
44 of the sediments lying on the frontal slope due to saturation from permanent groundwater and water flow from springs. The
45 western initiation point was characterized by a marked “U” shape with a large flat bottom surface where debris cemented by
46 hard ice were observed. The flat surface was 2.5 meters below the rock glacier original surface, extending 15 meters in width
47 and 30 meters in length (Figure 3c, d). These characteristics seemed to indicate a slightly different failure development at
48 the western initiation point, resembling the so-called “active layer detachments” (ALD) that are commonly observed in arctic
49 permafrost²⁻⁴.

50 Since the debris flow was triggered on a rock glacier, this event takes place in the context of periglacial hazards. In the
51 European Alps, periglacial hazards most commonly involve high altitude rock falls and increased sediment availability in
52 Alpine watersheds^{5,6}. The aim of the present study is to propose an interpretation, from a geomorphological perspective,
53 of the drivers that led to the frontal failures of the Lou rock glaciers on August 14th 2015. The objective is also to provide
54 original documentation on the case of a debris flow originating from the front of a rock glacier. Since the rock glacier was not
55 being monitored at that time, this study presents a back analysis mostly based on data that was collected after the event. The
56 underlying hypothesis is that the occurrence of slope failures was controlled by a combination of local predispositions such
57 as the topographical setting, the internal structure and the dynamic of the landform, and a preparatory sequence leading to a
58 trigger that can be linked to the climatic and meteorological events prior to the earth movement event.

59 These hypotheses are based on previous studies evaluating mass movements in permafrost terrain. For instance, the local
60 topographical setting can be conducive to the development and propagation of debris flows due to the presence of uninterrupted

61 steep slopes between the front of the rock glacier and the main torrential channel^{7,8}. Rock glacier dynamics and stability,
62 which are also related to local topography^{9–11}, can further increase the susceptibility to frontal instabilities^{12,13}. In addition,
63 meteorological conditions can represent important preparatory factors and directly trigger the failure. It has been observed that
64 high temperatures reduce shear strength¹⁴ and fracture toughness¹⁵ of frozen materials and that heat waves have often been
65 linked to significant mass movements in permafrost terrains^{16,17}. Meteoric or meltwater infiltration has also been identified as
66 an important preparatory and triggering factor for frontal instabilities^{1,12}.

67 Under the assumption that these external (weather related) and internal (internal structure and dynamic) factors controlled
68 the failure occurrence, we investigated the event by analysing data describing these parameters. Meteorological and nivological
69 data were analysed initially, in order to understand the climatic context prior to and during the event (methods in section 3.1).
70 Decadal-scale evolution of the surface velocity as well as the changes in the surface morphology were investigated using aerial
71 orthomosaics, which were retrieved by historical aerial imagery and UAV surveys after the event (methods described in section
72 3.2). The topographical settings and geomorphological characteristics were observed in the field as well as by UAV surveys
73 (study site described in section 2). The internal structure of the landform was inferred from the electrical resistivity tomography
74 and seismic refraction campaign conducted in 2016 and 2017 (methods described in section 3.3). A comprehensive analysis of
75 the data is then proposed, resulting in diagnosis of the 2015 failures (section 5.3). Finally, the legacy of this event is presented
76 including the risk management plan adopted by the local authorities and the recommendations for future efforts on this site
77 (section 5.4).

78 **2 Study site**

79 **2.1 General setting**

80 The Lou rock glacier is located in the Mont Cenis range, i.e. the orographic barrier between the Maurienne (France) and Susa
81 (Italy) valleys. The Pointe de Ronce (3612 m a.s.l.) is one of the highest peaks in the massif (Figure 2a). Its west ridge descends
82 to two secondary summits, respectively the Signal du Grand Mont Cenis (3356 m a.s.l.) and the Pointe de la Nunda (3023 m
83 a.s.l.) located further to the west. The north face of the saddle separating these two minor summits (Col du Lou, 3040 m a.s.l.)
84 contains the zone of origin of the Lou rock glacier, which develops on a ledge in the topography at 2700 m a.s.l..

85 The Mont Cenis range is part of the Piemont-Liguria Penninic nappe, consisting of cretaceous Bunder-schists ophiolites¹⁸.
86 In the north face of the Signal du Gran Mont Cenis where the rock glacier is located, two stratigraphic series can be identified.
87 From the valley bottom to 2600 m a.s.l. calcschists and marbles are found, while the mountain summit is made up of phyllite
88 marbles. In addition, black schists can be observed above 3400 m a.s.l. on the highest summits of the range. All these series are
89 densely jointed and prone to mechanical weathering, as can be seen in the many scree slopes and late Würm glacial deposits in
90 the area¹⁸.

91 Climatically speaking, the area presents the typical traits of the dry climate that characterizes the eastern side of the Northern
92 French Alps¹⁹. The 0°C isotherm (second half of the 20th century) is located at 2600 m a.s.l. while mean annual precipitation

93 at the valley bottom is of the order of 800 mm/y^{19,20}. Glaciers can be found only in the form of small ice aprons perched on the
94 north faces of the highest summits and cover in total about 2.5 km², extending from 3000 to 3500 m a.s.l. (data from 2009²¹).
95 Conditions favourable to the existence of permafrost are expected to be found from 2500 m a.s.l. on north faces and from 2900
96 m a.s.l. on south slopes²². HISTALP data²³ suggest a mean temperature shift of +1.4°C between the late Little Ice Age (LIA)
97 and recent times (1998 – 2008), which is expected to cause the altitude of conditions favourable to the existence of permafrost
98 in sedimentary materials to rise by an elevation of about 300 meters on all orientational aspects¹¹. The observed widespread
99 glacial retreat also bears witness to this temperature shift²¹.

100 Surface sediment transport is mainly dominated by gravitational (rock falls) and torrential processes (gullying, debris
101 flows), and numerous gullies and couloirs can be observed on steep slopes on all aspects. Although periglacial environment is
102 extensive in the region^{22,24}, the Signal du Gran Mont Cenis area contains only a few rock glaciers due to the rareness of flat
103 surfaces where rock debris can accumulate. One of the few ledges is the site of the Lou rock glacier, which has a surface of
104 about 0.2 km² and is characterized by the presence of several fronts overhanging the steep gullies (>40°), which correspond to
105 the headwaters of the Arcelle Neuve stream where the 2015 debris flow took place.

106 **2.2 The Lou rock glacier**

107 The Lou rock glacier can be defined as "pebbly"²⁵ as the largest clast size observed at the surface is of the order of a few
108 decimeters. Instead of the typical single tongue and single front shape, the Lou rock glacier has five frontal lobes surrounding
109 the gullies in cirque-type configuration with steep frontal slopes ranging between 38°- 42°(Figure 2b). The landform can thus
110 be defined also as a polymorphic rock glacier²⁶. Due to this complex morphology, the specific toponymy presented in Figure 2c
111 will be used throughout the paper.

112 Two main frontal lobes separated by an embankment that bends eastwards characterize the eastern and central part of the
113 rock glacier. These eastern and central lobes form outwards from a plateau which receives a supply of debris from the scree
114 slope leading to the Col du Lou pass. While the central part of this plateau is relatively flat, the eastern side presents a clear
115 depression, suggesting the past presence of a small glacier. The embankment that separates the central and eastern part of the
116 plateau can be interpreted as the lateral moraine of this glacier. The easternmost slope failure related to the occurrence of the
117 2015 debris flows, i.e. the eastern slide, developed where this lateral moraine reaches the front line and leans over the gully.

118 The western side of the rock glacier complex is characterized by three lobes which are supplied directly by the surrounding
119 headwalls. While the two westernmost lobes, i.e. inactive lobes in Figure 2c), are partially vegetated and present a rounded
120 topography, the third lobe shows clear signs of intense activity. This lobe is characterized by a dense network of cracks that
121 scar the surface every 2 – 5 meters. The cracks are probably shallow, measuring up to 20 meters long and creating a rugged
122 micro-topography at the surface that features “steps”, measuring one to two meters in height and width (Figure 4). None of the
123 cracks appear to cut the lobe through its entire thickness, but rather disconnect it from the other rock glacier lobes (inactive
124 lobes and central lobe). These characteristics are typical of pebbly rock glaciers showing potential destabilization¹¹. The
125 western slide associated with the 2015 debris flow events occurred on the orographic right-hand side of this western lobe’s front.

126 **3 Methods**

127 **3.1 Meteorological data**

128 Climatic conditions prior to the event were investigated by analysing meteorological and nivological data including air
129 temperature, rainfall and snow cover data relative to the 2014-2015 winter season and the summer of 2015. Climatic anomalies
130 were also investigated by comparing the data from 2014-2015 with the average and typical variability of the available time
131 series.

132 The main meteorological data (air temperature and precipitation) was obtained from the Mont Cenis weather station located
133 at 2035 m a.s.l., 3 km away from the site (station id: 73144001, Figure 2a). The station is property of Météo-France and has
134 recorded daily mean air temperature and precipitation since 1992. In addition, data from the Bessans weather station located at
135 1710 m a.s.l. and 7 km north-east from the site (station id: 73040005) was used to fill the gaps in the data series from the Mont
136 Cenis station. The procedure to fill these gaps was carried out by looking at the correlation between data for periods during
137 which both stations provided continuous measurements and applying the linear regression model to predict missing data at
138 Mont-Cenis.

139 The Bessans weather station also provided snow height data series for a period spanning from winter 2011-2012 to 2017.
140 Since snow height values were acquired only every 1 to 4 days, the resulting database is discontinuous and only suitable for
141 qualitative analysis.

142 **3.2 Dynamical behaviour**

143 The spatio-temporal evolution of permafrost creep velocities that characterize the rock glacier was investigated in order to
144 understand its past and current dynamics. First, the observation of time series of orthoimages allowed the evolution of the rock
145 glacier surface velocities over the past decades to be reconstructed at coarse temporal resolution, i.e. velocity averaged over
146 several years^{27,28}. Differential GPS (dGPS) surveys have then been performed at least once a year since 2015 to quantify annual
147 displacements of a few points to a high degree of precision²⁹. Finally, UAV imagery was acquired in 2016, 2017 and 2018 to
148 compute and compare high-resolution orthophotos and Digital Elevation Models (DEMs) to understand the deformation pattern
149 and the spatial characteristics of erosion-deposition on the destabilized lobe^{13,30}. UAV surveys were preferred to terrestrial
150 radar and laser scanning^{12,31} because of the lack of suitable viewpoints to set up instruments near the Lou rock glacier.

151 **3.2.1 Historical aerial imagery**

152 Historical aerial imagery was used to reconstruct the dynamic behaviour of the rock glacier in the second half of the twentieth
153 century. For that purpose, several orthoimages acquired at different dates were computed by triangulating aerial images using
154 the Ortho Engine module in PCI Geomatica²⁷. The raw aerial images used in this study were directly obtained from the Institut
155 Géographique National (IGN) website where they are freely available. Distortion coefficients and focal lengths were provided
156 by the IGN support team for the requested missions. Images from 1970 and 1996 were selected for the analysis because of the
157 absence of features such as snow patches and shadows that may significantly reduce the precision of the velocity estimation. In

158 addition, already orthorectified aerial images from 2006 and 2013 (resolutions of 1 and 0.5 meters respectively) were made
159 available by the IGN³² and could also be used in this study.

160 In order to infer surface velocity values, the selected orthoimages were compared with each other on the QGIS software.
161 Noticeable boulders were tracked between pairs of orthoimages and their displacement rate was computed by dividing the
162 measured displacement by the time elapsed between the two images. The uncertainty caused by image distortions and errors
163 was quantified by tracking apparent movements on non-moving areas (i.e. areas of vegetation and bedrock).

164 **3.2.2 DGPS surveys**

165 DGPS surveys were conducted using a rapid-static approach, which consists of post treating data acquired using a mobile
166 receiver jointly with data acquired using a base station of known coordinates. This method allowed high precision (2 cm) to
167 be reached with low acquisition time (30 seconds to 5 minutes per point, depending on satellite configuration). A Trimble
168 Geo7x antenna combined with a TopCon (GB 1000) receiver was used as a basis. Firstly, four targets were marked and their
169 positions measured on August 25th 2015. The targets consisted of crosses made of red and green paint on several large visible
170 boulders located on the surface of the western destabilized rock glacier lobe. Coordinates of the targets were measured again on
171 19th July 2016 and 14th September 2016 with a mobile receiver and new targets (17 in total) were added to increase spatial
172 resolution. Surveys were then repeated on August 12th 2017 and August 22nd 2018. Data were treated using the software
173 Trimble Pathfinder.

174 **3.2.3 UAV surveys**

175 UAV surveys were used to acquire close-range high-resolution aerial images of the destabilized lobe. The UAV-borne
176 photogrammetric surveys were systematically performed on the same dates as the dGPS surveys (14th September 2016, 12th
177 August 2017 and 22nd August 2018). The 2016 mission was performed using a DJI Phantom 3, flying in manual mode. Pictures
178 were captured at different heights, ranging from 20m to 90m above the surface. Eight targets were measured by dGPS and used
179 as Ground Control Points (GCPs). The 2017 and 2018 surveys were performed using a DJI Mavic Pro. Surveys were carried
180 out using an automatic flight planner, i.e. the Drone Map software. Pictures were taken at 70 to 150 m altitude and 9 (in 2017)
181 to 12 (in 2018) GCPs were also acquired by dGPS for these campaigns.

182 UAV images were processed using Agisoft Photoscan following a Structure from Motion approach in order to obtain
183 DEMs and Orthoimages³³. Firstly, point clouds were computed using the UAV on-board GPS data to facilitate the structure
184 reconstruction. In a second stage, GCPs were used to georeference the models. The locations of GCPs in the aerial images was
185 manually adjusted in order to optimize the model until errors were below the pixel size³³. Point clouds were then densified,
186 meshed and exported as Orthoimages and DEMs.

187 The resulting orthoimages were downsampled at a resolution of 25 cm x 25 cm and used to estimate displacements using
188 the module of automatic feature tracking IMCORR³⁴. Unfortunately, the orthoimage obtained from the 2016 UAV survey
189 had to be discarded due to an insufficient coverage of the Eastern rock glacier lobe. As a consequence, only the orthoimages
190 from 2017 and 2018 were selected to perform the feature tracking analysis. Results were manually cleaned to avoid single

191 points of both high speed and movements characterized by unrealistic direction in comparison to the general slope aspect.
192 Correlation points were interpolated using a kriging algorithm to produce a heatmap of displacement over the rock glacier
193 surface for the 2017-2018 time interval. The accuracy of the result was evaluated by (i) comparing the computed displacement
194 with movements obtained from dGPS surveys and (ii) by comparing apparent movements on stable areas.

195 Finally, DEMs were also compared to observe surface elevation changes on the destabilized western lobe. In order to have a
196 more precise evaluation of the surface height variations, the models were co-registered relative to the 2016 model by matching
197 stable areas in the best fit ICP algorithm³⁵ in the CloudCompare software³⁶.

198 **3.3 Geophysical surveys**

199 Electrical Resistivity Tomography (ERT) and seismic refraction tomography (SRT) campaigns were performed in 2016 and
200 2017 to investigate the internal structure of the landform. These methods are largely used in mountain permafrost investigations
201 to visualize the vertical transition from unfrozen to frozen ice-rich sediments using ERT^{37,38} and the thickness of the different
202 horizontal layers using SRT³⁹.

203 **3.3.1 Electrical Resistivity Tomography**

204 Two profiles of 64 electrodes (2.5 meters spacing) were acquired on September 15th 2016 on the main plateau (medial and
205 eastern plateau in Figure 2c) of the rock glacier (PE1 and PE2, Table 1). On August 10th 2017, two profiles consisting of 64
206 electrodes (5 meters spacing) were acquired longitudinally (PE5) and transversally (PE6) on the destabilized lobe, from the foot
207 of the front up to the rooting zone and between the two debris flow slides respectively. For all the profiles, the acquisitions
208 were performed using the Wenner array and the resulting data was inverted using the Res2Dinv software⁴⁰. The position of the
209 electrodes was measured by dGPS Trimble Go7x and used to compute the topography of profiles.

210 **3.3.2 Seismic Refraction**

211 Seismic refraction was conducted using a Sesitronix DaqLink III and 24 geophone cables with 5 meters spacing. Real time
212 connection between the Vscope software and the seismograph was ensured by Ethernet cable. Wave source was triggered by
213 hammer hits and for each geophone 2 to 5 triggers were shot in order to achieve a clean signal. Two offset triggers were shot at
214 a distance of 5 and 10 meters from the beginning and end of each profile. The survey took place on the western destabilized
215 lobe on October 18th 2017. In order to cover the whole length of the lobe, three seismic profiles were measured longitudinally,
216 each of them overlapping the previous one by 6 geophones to ensure consistent transition between the different profiles.

217 Data was then processed using the ReflexW software. First arrivals were selected manually and checked using the automatic
218 travel time analysis and the inversion was performed on an initial model consisting of a surface velocity of 400 m/s increasing
219 with depth at a rate of 300 m/s/m. Surface topography was retrieved from in-situ dGPS measurements.

4 Results

4.1 Meteorological conditions prior to the event

During the hydrological year 2014-2015, the meteorological data from the Mont Cenis weather stations presented two main gaps, respectively between November and January, and between April and June (Figure 5a,b). Data issued from the Bessans weather station were found to correlate well with Mont Cenis temperature data ($R^2=0.96$) allowing a robust gap filling, while correlation was weaker for precipitation data ($R^2=0.61$).

Overall, the winter of 2014-2015 was characterized by notably low amounts of snow accumulation (8th lowest amount since 1959). In particular, November, April and May received about 80% less snow than the average since 1959⁴¹. December was also characterized by very little snow precipitation and very thin snow cover (Figure 5c). Temperatures oscillated from abnormally warm at the end of November (3 to 5°C than average on record) to abnormally cold in late December (-5 to -7°C than average on record). Consistent snowfalls arrived in mid-January, as the measured snow height increased from 0.2 m to 0.7 m in the Bessans weather station. However, the snow cover remained relatively thin throughout the whole winter and early spring compared to previous and following years, with snow cover approximately 0.5 m thinner than average at the peak of the accumulation season (late March).

Spring 2015 was characterized by warm temperatures in April – May, occasionally 3-5 °C higher than average. The more intense precipitation event recorded in 2015 reached 50 mm/day and was registered in mid-May at the Mont Cenis station. Given the low temperatures measured at 2000 m on this occasion, this event resulted possibly in solid precipitations at the Lou rock glacier. Temperatures increased steadily through June and then July, with an exception for an episode of rain occurring in mid-June associated with a temporary temperature drop.

Summer 2015 was characterized by a severe heatwave and drought that lasted about a month between late-June and mid-July. Temperatures were 5 to 7 °C higher than average for two consecutive weeks, making this period the warmest on record. Between the end of July and the debris flow event (August 14th), a series of rainfall events with intensities ranging from 10 to 35 mm/day were recorded, for a total amount of 220 mm of precipitation. Compared to previous years, the three-week period previous to the 2015 debris flow event represented the wettest on record to occur in summer.

In particular, the last days before the debris flow events were characterized by a series of relatively intense precipitations. The second-strongest precipitation event of the year (35 mm/day) occurred on August 8th, followed the next day by a smaller event (15 mm/day). After three days without precipitation, 10 mm of rain were recorded on August 13th at the Mont Cenis station and finally an event of precipitation of about 30 mm/day characterized the day of the events (on August 14th). The precipitation events mentioned, including the one occurring on August 14th, were however not exceptional in terms of intensity as rainy events above 60 mm/day and up to 90 mm/day were regularly recorded at the Mont Cenis station since 1992.

4.2 Landform dynamics

4.2.1 Historical aerial imagery

The Lou rock glacier has experienced a generalized acceleration over the past four decades (Figure 6). The western lobe systematically showed the highest velocities of the whole landform, with velocities increasing from 0.5 m/y (1970 – 1996) to 1.0 m/y (1996 -2006), then to 1.9 m/y (2006 – 2013) to finally reach 3.5 m/y (2013 – 2017). The eastern lobe had a similar behaviour with velocity gradually increasing from 0.15 m/y (1970 – 1996) to 1.3 m/y (2013 -2017), while the central lobe apparently encountered a less pronounced acceleration (0.3 m/y in 1996 – 2006 to 0.6 in 2013 – 2017). The inactive lobes on the western side appeared to have experienced a strong reactivation between 2006-2013, as the velocity increased from 0.3 m/y (1996 -2006) to 1.3 m/y (2006 – 2013) with substantial observable frontal advance (Figure 7). Since 2013, these lobes have remained almost inactive, as velocity dropped down to about 0.2 m/y (2013 – 2017). A significant frontal advance of about 20 meters could also be observed for the western lobe between 2006 and 2017 (Figure 7). Depending on image pairs, the uncertainty estimated by observing stable areas was evaluated to be of the order of 0.1 to 0.3 m/y, and thus in most cases significantly lower than the surface velocity values. The calculated displacement rates and their variability can therefore be considered reliable.

4.2.2 dGPS

DGPS measurements were conducted only on the western lobe and only two targets could be measured continuously from 2015 to 2018. Displacement data indicated a slight deceleration in 2016-2017 compared to 2015-2016 and an acceleration between 2016-2017 and 2017-2018, reaching displacement rates about 10% higher than the period 2015-2016. Measurements indicated that the frontal part moved faster (about 3.5 m/y) than the rest of the lobe, with the lowest values of surface velocity (few dm/y) measured at the limit between the western and the central lobes.

4.2.3 UAV photogrammetry

Velocity values extracted from UAV photogrammetry were considered significant above 0.53 m/y because below this threshold, it was possible to observe apparent movements on stable areas. The UAV-derived orthoimages comparison allowed surface displacements at a high spatial resolution to be identified and highlighted the complex morphology and heterogeneous creeping patterns of the rock glacier complex. The western lobe moved significantly faster than the other lobes between 2017 and 2018 (Figure 8). The highest velocities (<3.5 m/y) were observed close to the edge of the front and in the western part of the lobe (Figure 8). Comparison between displacement values obtained from the UAV surveys and from dGPS in this area for the same period showed good correlation (Figure 8). The eastern lobe was found to creep at about 1 to 1.5 m/y, showing also maximum velocities at its front. This lobe seems to diverge into two minor lobes moving in two different directions, possibly due to the underlying bedrock topography. On the slope above the eastern plateau, it was possible to see a system of small fast-moving lobes creeping at rates of 2.5 m/y. A similar pattern is observable upslope of the central plateau, where a minor lobe is creeping downslope at a velocity of 1.5 m/y while the rest of the area shows only small movements (below the limit of detection). Finally, it was possible to identify two small surface slides on the front of the central lobe.

283 DEM subtraction showed a confidence interval, i.e. the double of the standard deviation of elevation changes on stable
284 areas, of 0.48 m. This threshold was used as a lower limit of detection and surface variations below this value were discarded.
285 The comparison of the DEMs obtained for 2016 and 2018 showed a significant debris accumulation on the lower part of the
286 western lobe front, reaching 1.6 m thickness and a total volume of about 1000 m³ (Figure 9). The edge of the front does not
287 seem to have significantly moved forward or backward between 2016 and 2018. The western slope failure did not evolve
288 significantly since the event of 2015, with only minor accumulations observed below the main scarp.

289 **4.3 Geophysical Investigations**

290 **4.3.1 ERT investigations**

291 The PE1 profile revealed a high resistive body located in the centre of the eastern plateau (Figure 10a). Resistivity reached 3
292 000 k Ω in the depression, at about 20 meters deep. This resistive body was surrounded by more conductive terrains, ranging
293 around 10 – 50 k Ω . The upper part of the central plateau showed the presence of a resistive body reaching 500 k Ω at 10 – 15
294 meters depth and located in the middle of the central lobe. It was possible to observe lower resistivity (20 – 40 k Ω) in the zone
295 of origin of the western lobe (Profile PE2).

296 On the western sector resistivity values were generally lower. The longitudinal profile on the western lobe (PE5) ranged
297 from 0.5 k Ω downslope up to 500 k Ω upslope. Starting upward from the front line, a resistive body could be observed at 20
298 meters depth, with resistivity values ranging from about 10 – 30 k Ω in the downslope part and up to 500 k Ω on the upslope part
299 of the profile. Below the resistive body, values dropped quickly to 2 – 0.5 k Ω , reaching similar values to those that can be
300 observed downstream from the front in areas believed to be free from permafrost.

301 The transversal profile across the debris flow slope failures (PE6) showed resistivity values generally ranging from 0.5 to
302 10 k Ω , except for a small resistive body identified at about 20 m deep in the middle of the profile and corresponding to the
303 terminal part of the western lobe. Lower resistivity values ranging from 1 to 5 k Ω were found all long the profile at 20 to 30
304 meters depth.

305 **4.3.2 Seismic refraction investigations**

306 The seismic profile PS3 reached a depth of 17 m in the downslope half and 30 meters in the upslope half (Figure 10b). Wave
307 velocities ranged from 300 to 5000 m/s. Higher velocities could be found upslope, near the limit between the western and the
308 central lobes. Lower velocities were observed in isolated surface patches and down to about 2 – 5 m deep. In general, wave
309 velocity increased linearly with depth. In the downslope part of the profile, wave velocity increased from 500 to 4000 m/s
310 within 15 m depth. In the central section the velocity increase rate was gentler, as velocity ranged from 1000 to 3000 m/s at 30
311 meters depth. Upslope, the velocity strongly increased within 10-15 meters from 300 to 5000 m/s.

312 **5 Discussion**

313 **5.1 Historical and Recent dynamics**

314 **5.1.1 Results Interpretation**

315 Consistent with several other rock glaciers in the European Alps^{42,43}, a marked acceleration was found to characterize the Lour
316 rock glacier during the past decades. On the western lobe, the rate of the acceleration over the past four decades (from 0.8
317 in the period 1970-1996 to 3.6 m/y in the period 2013-2017, i.e. + 400%) can be compared to other cases of rapidly moving
318 rock glaciers^{9,10,12,44,45} and suggests a partial destabilization of the landform. UAV and dGPS surveys confirmed the high
319 current velocity of the western lobe, especially in its frontal zone where velocity rates reached about 3.7 m/y between 2017
320 and 2018. The highest values were measured in the area where the clustered tension cracks have been observed, while the
321 zone surrounding the western slope failure moved at a remarkably slower rate. According to the crack development process
322 described by previous studies^{12,15,46}, pebbly rock glaciers characterized by temperate permafrost conditions and extensive
323 flow patterns are highly susceptible to surface cracking. Cracks generate preferential water flow paths in the landform that
324 encourage reduction of the shear stress resistance and increase the heat transfer by advection. As a consequence, positive
325 feedback processes can be triggered and lead to a further increase in permafrost creep rates, permafrost temperatures and even
326 encourage development of new cracks.

327 **5.1.2 Limitations of the method**

328 Most of the uncertainties associated with the assessment of the rock glacier dynamics were related to the historical analysis.
329 The rock glacier is relatively small in size and its location on steep slopes may induce critical distortions in the computation
330 of orthomosaics from historical images. Stable areas used to assess the errors were often located in zones characterized by
331 flatter slopes in comparison to the rock glacier surface and the calculated uncertainties may be under-estimated in steeper areas.
332 These errors can be illustrated by the change in direction of displacement vectors between consecutive time intervals, which
333 sometimes exceeded 40 degrees of aspect change. Despite limitations, GPS and UAV data confirmed the order of magnitude of
334 the movements observed on the last image pair (2013-2017) and seem to highlight the validity of the values obtained from the
335 analysis of old aerial images.

336 **5.2 Internal structure**

337 **5.2.1 ERT profiles**

338 The highly resistive body measured at the eastern plateau (PE1) indicates the presence of massive ice, which is also supported
339 by the probably glaciogenic formation of this lobe. The depression observed in this area may thus be caused by the partial
340 thaw of this sedimentary ice body. The geoelectric profile suggested a massive ice thickness of about 15 meters, with some
341 uncertainty due to the limited depth of the investigation rendering the identification of a clear transition between the highly
342 resistive massive ice and a potentially underlying more conductive body impossible (bedrock or unfrozen sediments).

343 Lower values of resistivity were measured on the western lobe. They are typical for temperate permafrost with low ice

344 content and fine-grained sedimentary material⁴⁷ and suggest a periglacial genesis of this area of the landform⁴⁸. The resistive
345 body observed in the profiles PE1 and PE2 appeared to be 20 meters deep and about 50 meters wide, showing increasing values
346 of resistivity upslope. The presence of temperate permafrost on the western lobe is in agreement with the development of high
347 displacement rates and surface cracking^{12,15}.

348 **5.2.2 SRT profiles**

349 The seismic profile provided relevant information on the internal structure of the western lobe. Penetration of the waves was
350 limited in depth to around 20 – 30 meters. This depth correlated with the transition between high and low resistivity values
351 observed in the ERT. This transition probably corresponds to a reduction in ice content, and indicates that the western lobe
352 creeps above a layer of unfrozen sediments or densely jointed bedrock. However, a better characterization of the material
353 present under the permafrost body would require further investigations including for instance a deeper sounding.

354 In addition, the lower seismic velocity values observed near the surface suggested an active layer depth of 2-3 meters on
355 the western lobe. An apparent increase in active layer thickness can be identified in the medial part of the lobe on the seismic
356 profile, reaching up to 5 meters in depth. In this area, the velocity anomalies seem to be correlated with the presence of cracks
357 and may be induced by higher air content at shallow depth. The presence of these disturbances up to a depth of 5 m, i.e. deeper
358 than the estimated active layer thickness (2-3 m) indicates that the cracks propagate in the upper part of the permafrost body.

359 **5.3 A comprehensive diagnosis of the frontal failures**

360 Both the geomorphological characteristics and the presence of water springing out of the head scarps indicate that the frontal
361 slope failures resulted from intense linear regressive erosion, previously described as concentrated flow processes from
362 observations conducted at Dirru, Gugla and Tsarmine rock glaciers in the western Swiss Alps¹. In order to describe and explain
363 in detail the occurrence of this process in the case of the Lou rock glacier, we subdivided the discussion into three parts: the
364 local predispositions, the climatic conditions prior to the event and then the triggering factors.

365 **5.3.1 Predisposition : Topographical and geomorphological characteristics of the rock glacier**

366 As mentioned, the Lou rock glacier developed on a flat ledge located in an otherwise very steep topographical setting. The front
367 of the rock glacier is very steep (up to 42°) and directly connected to torrential gullies. The general topographical setting of the
368 area thus encourages the occurrence and propagation of erosion and sediment transfer processes by gravity¹.

369 The eastern slope failure occurred at the limit between the central and the eastern fronts where traces of previous events
370 (head scarps and gullies) were in evidence on historical imagery dating back to 1953 (Figure 7). Therefore, the concentrated
371 flow process causing the 2015 eastern failure had most likely already occurred in the past decades. In addition to the steep
372 front directly connected to the gully system, the presence of a topographical depression located upslope from the head scarp
373 and characterized by the presence of massive ice only a few decimetres below the surface is expected to encourage water
374 concentration and circulation on top of the ice body and lead to frequent release of water onto the frontal slope where the failure
375 was observed. As a result, concentrated water flow can be generated and lead to intense regressive erosion phases, similar to

376 those observed in the western Swiss Alps¹.

377 The western failure, on the other hand, represented a novel event as there were no indications of previous substantial erosion
378 on the historical aerial images. Therefore, one can assume that the occurrence of the 2015 erosional event was either related
379 to recent changes in the internal structure and morphological characteristics of the landform, or due to exceptional weather
380 conditions, or both. The investigations conducted on the western lobe revealed relevant geomorphological modifications in the
381 past decades that were related to the significant increase in velocity for the landform. These geomorphological modifications
382 consisted of a significant advance of the front position and the development of cracks on the surface of the lobe. The observed
383 progression of the front-line appears to have enhanced the connectivity between the rock glacier and the torrential system,
384 since the 2015 failure occurred in an area occupied by the rock glacier only after 2006 (Figure 7). In addition, surface
385 cracking is suspected to have reduced material shear resistance and increased the susceptibility to water infiltration upslope
386 from the front-line¹⁵, possibly enhancing the occurrence of supra-permafrost flow. In this sense, we suggest that the high
387 displacement rates of the rock glacier and consequent destabilization process represent relevant changes that tended to increase
388 the predisposition of the western lobe to failure. Permafrost thermal characteristics and internal structure also played a role
389 in the occurrence of this phenomenon as rock glacier cracking and destabilization is known to be more likely to occur in
390 temperate permafrost conditions^{12, 15}. In addition, the exceptionally wet conditions in the weeks prior to the event must also be
391 acknowledged as an important factor in explaining the occurrence of the 2015 failures in an area of the front that remained
392 previously unaffected by this type of intense erosion process.

393 **5.3.2 Preparation: climatic conditions prior to the event**

394 The climatic sequence that preceded the failure events was characterized by several anomalies consisting of (i) scarce snow
395 cover with an early snow disappearance at the end of the winter, (ii) the warmest heat wave on record between mid-June and
396 mid-July and (iii) the wettest three-week period recorded in summer since 1992. All these factors are known to influence
397 permafrost characteristics and slope stability but, due to the lack of in-situ data, the respective effects of these climatic anomalies
398 on the occurrence of the failures cannot be demonstrated by evidence. Nevertheless, we propose here a brief description of the
399 processes that may have contributed to facilitation occurrence of the failures.

400 Heat waves may lead to significant ground warming resulting in permafrost degradation and active layer thickening. The
401 loss of permafrost ice content due to thawing and the increase of liquid water content associated with permafrost degradation⁴⁹
402 are often addressed as factors favouring the occurrence of mass movements in mountain environments due to the decrease of
403 the frozen ground stiffness and resistance to shear stress¹⁴. This effect can be amplified with an early disappearance of the snow
404 cover which leads to a longer period of ground exposure to solar radiation. In the case of the Lou rock glacier, the absence of
405 data characterizing the thermal state of permafrost and the evolution of the active layer thickness prior to the event does not
406 allow the role of these drivers to be identified as preparatory factors to the debris flow event.

407 On the other hand, repeated rainfalls are known to be an important factor influencing sedimentary slope stability by reducing
408 the shear stress resistance and causing local instabilities¹². Progressive increase of the water content was for instance observed

409 to be a relevant preparatory factor in the occurrence of frontal slope failures at the Dirru, Gugla and Tsarminé rock glaciers in
410 the western Swiss Alps¹. We can thus assume that the high total precipitation recorded during the weeks preceding the events
411 had a large role to play in triggering the Lou slope failures.

412 **5.3.3 Trigger: concentrated flow consecutive to repeated rainfalls**

413 The three weeks prior to the debris flows were characterized by a repetition of rainfall events of small to medium intensity.
414 The frontal failures were ultimately triggered during a minor precipitation event. It appears that the significant infiltration of
415 meteoric water resulting from the frequent rainy episodes led to saturation of the active layer and generated supra-permafrost
416 flow. Water running on top of the permafrost table was still observable the day after the event at the head of the two failures and
417 confirms the (almost) absence of infiltration within or below the permafrost body. Both western and eastern slope failures were
418 then triggered by concentrated flow (¹) that generated regressive erosion from two water springs located respectively at the
419 outflow of the central plateau and the flat eastern plateau, which collects large amounts of water from the nearby slopes.

420 The peculiar morphology of the western slide indicates that parts of the erosion occurred in one single sliding event. Such
421 an event could have been triggered by the removal of sediment further downslope from linear flow erosion, diminishing the
422 support for the materials lying at the front. A large-scale debris slide of this type could be the signature of large concentrated
423 flow events on steep pebbly rock glacier fronts for which the higher ratio of finer grained sediments in comparison to blocky
424 rock glaciers favours higher sediment cohesion and thus aids the occurrence of single slides instead of gradual rock falls.

425 **5.4 Recommendations**

426 Although the Lou frontal failures consisted of a volume only slightly superior to 1000 m³, the debris flow generated was able to
427 mobilize up to 15 000 m³ of debris stored in the Arcelle Neuve stream or on the sides of the torrential channel. As demonstrated
428 in the flooding of the town of Lanslevillard, river infrastructures in this alpine catchment were inadequate to face this event. In
429 order to mitigate the risk linked to possible future debris flow in the Arcelle Neuve stream, the RTM built a new infrastructure
430 at the junction with the Arc River⁵⁰. This infrastructure consists of an open channel which substitutes the former underground
431 pipe in order to allow debris flows to continue and reach the Arc River where sediments will be evacuated by the main river
432 flow. This channel was designed to be operational for debris flows up to 25 000 m³. The volume was evaluated assuming future
433 frontal failures of the same order of magnitude as the 2015 event in terms of mobilized materials and is consistent with other
434 studies documenting similar debris flow mechanisms^{1,8}.

435 Due to its particularities, i.e. its direct connection with torrential channels and displacement patterns, we recommend that
436 the Lou rock glacier should become a future reference site for permafrost research in the region. Annual monitoring involving
437 repeated UAV photogrammetric surveys have been proven to be efficient to monitor both the evolution of the surface velocities
438 and morphological changes characterizing the fronts. Further efforts should focus on analysing the erosion processes at the front
439 of the landform and the sediment transfer rate between the front and the gullies, in order to quantify the role of the rock glacier
440 in the sediment chain. We propose an approach based on fixed camera monitoring^{1,51} and high-resolution DEM difference of
441 the upper and lower gully¹³. Furthermore, borehole investigations could bring significant information about the current state

442 and evolution of the active layer depth, as well as the sedimentary properties of the western lobe. Finally, further efforts should
443 focus on the characterization of other sediment sources and channel recharge rates in the lower sectors of the Arcelle Neuve
444 torrent to better establish debris flow scenarios.

445 **6 Conclusions**

446 In this study we provided a back analysis of the frontal slope failures at the Lou rock glacier. While characteristics of the eastern
447 slide are characteristics of linear regressive erosion from concentrated water flow, the western slide presented geomorphological
448 characteristics that resemble an active layer detachment (ALD), i.e. a slide of a portion of the active layer, mainly observed
449 in arctic permafrost sites. The development of this type of failure, and at such a scale, is probably due to the high ratio of
450 fine-grained material characterizing the Lou rock glacier and which is more conducive to the occurrence of debris slides than
451 individual or cascading rock falls that are commonly observed on boulder rock glacier fronts. In addition, the erosion observed
452 during the 2015 events at the fronts of the Lou rock glacier only mobilized sediments within the active layer, corresponding to
453 similar events observed at other rock glacier sites.

454 Both local topography (steep slopes) and the morphological settings (steep fronts overhanging gullies) of the landform
455 are considered to be important factors preconditioning the occurrence of the slope failures. It has been possible to see the
456 eastern failure in orthoimages since the 1950's, indicating that similar events occurred in the past. On the other hand, the
457 western slide corresponds to the first event recorded in that sector since at least 1953 (oldest useable aerial image). In addition
458 to the exceptionally wet conditions in the few weeks previous to the event, the recent increase in creep velocity leading to
459 destabilization of the western lobe represents the most relevant changes that potentially explain the sudden occurrence of
460 intense erosion in this sector. The destabilization of the western lobe is believed to have facilitated development of a frontal
461 slope failure (i) by reducing the distance between the frontal slope and the torrential gully by means of the significant advance
462 of the front line, enhancing the rock glacier – torrent connectivity and (ii) by leading to the formation of tension cracks on the
463 surface of the lobe, which favour the infiltration of water into the ground and lead to the occurrence of the concentrated flow
464 process that triggered the debris flow events. In this sense, we suggest that rock glacier destabilization may have a significant
465 impact on on-site predisposition to failure. However, further investigations would be needed in order to provide better evidence
466 for the effects of such destabilizations on the sediment transfer activity between rock glacier fronts and torrents.

467 Moreover, we highlighted here the anomalous 2014-2015 climatic conditions, characterized by scarce snow cover in late
468 winter, the occurrence of a strong heat wave between late June and mid-July and an exceptionally wet period in the few weeks
469 preceding the failures. The debris flows occurred after a series of meteoric water inputs, pointing out the important role of water
470 availability on triggering such intense erosion events. In addition, it is emphasized that rainy events were not exceptionally
471 intense but occurred regularly over a three-week period, suggesting that the frequency of precipitation events was critical in
472 triggering the failure, as opposed to their intensity. It has been mentioned that the western slope failure was a novel event and
473 had probably been preconditioned by the recent destabilization of the corresponding lobe. It is also important to acknowledge

474 the exceptional character of the precipitation patterns prior to the event as another explanatory factor for the occurrence of such
475 unprecedented events. If the direct impact of the weather conditions on the landform could not be analyzed in detail due to data
476 scarcity, we point out that the correlation between climatic anomaly and failure events would merit further research effort in the
477 future in order to improve understanding of the link between climate and the thermal state of the rock glacier.

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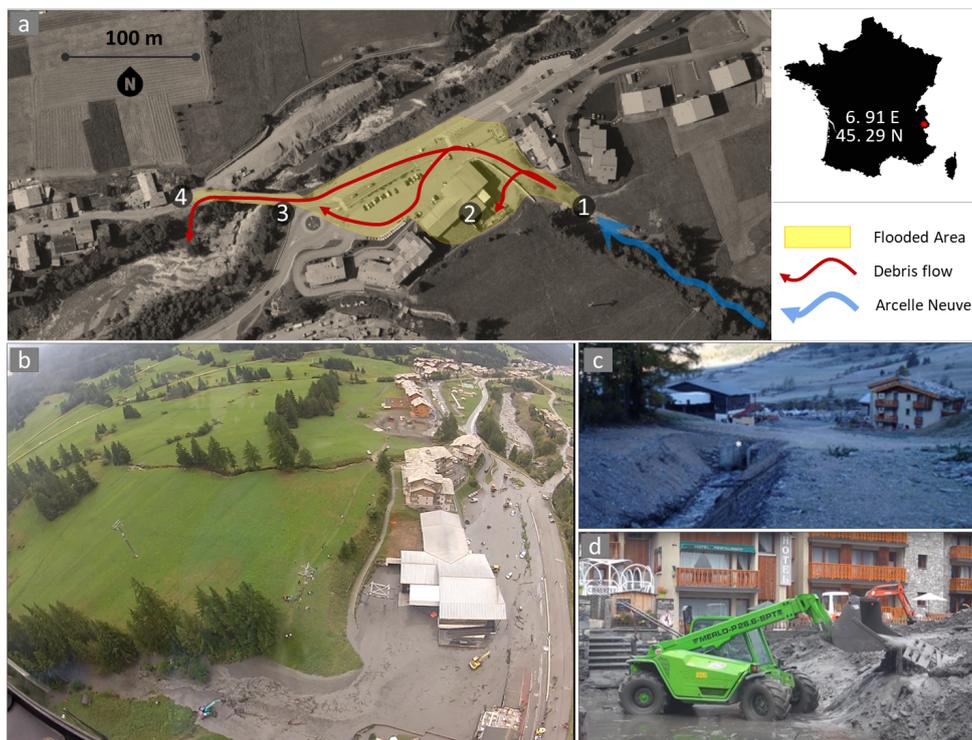


Figure 1. Overview of the debris flow that flooded the town of Lanslevillard the 14th August 2015. On top left (panel a) it is shown (1) the debris flow transiting in the Arcelle Neuve stream, which passes through a narrow pipe under the town (panel c). (2) The pipe got jammed and the debris started to flood the departmental road D902 and buildings nearby (panel b and d). (3) The flow took a secondary road towards the inhabited centre of “Les Champs”. (4) The flow was finally deviated into the Arc river by setting up an emergency embankment

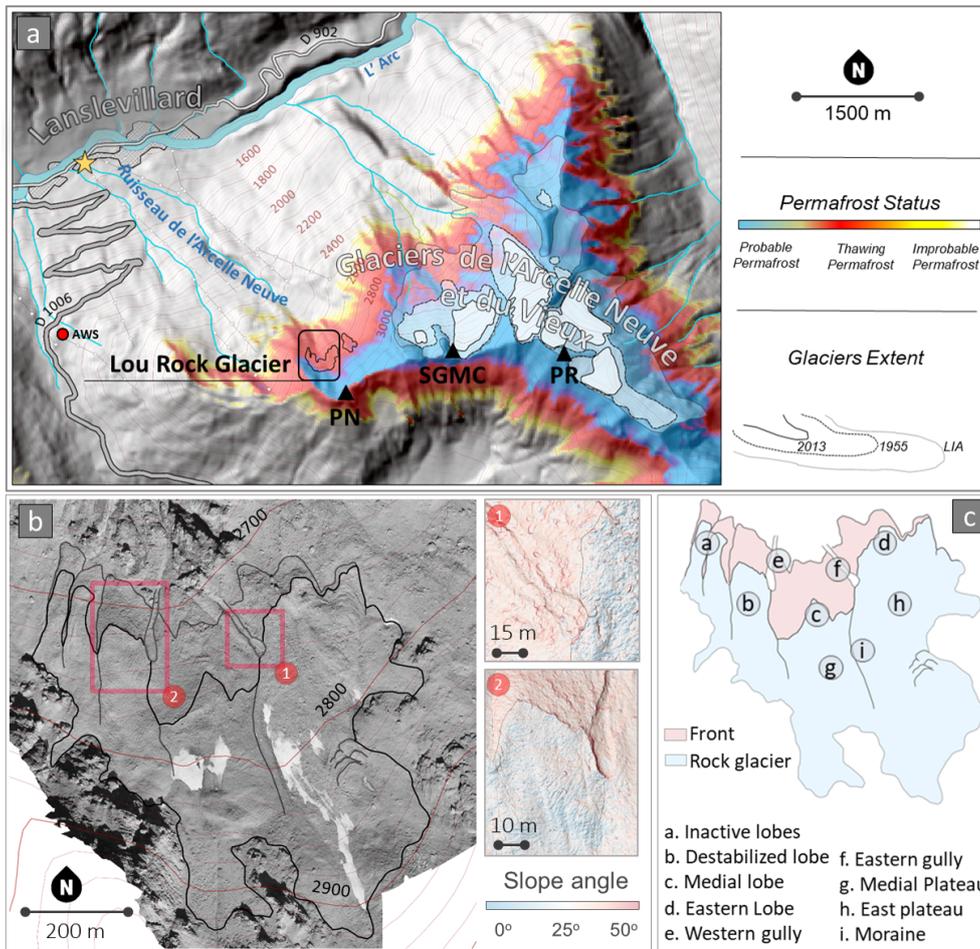


Figure 2. Overview of the Lou rock glacier. In panel a is presented the cryosphere status in the Mont Cenis Range. Glacier data²¹ and permafrost status¹¹. Acronyms: PN: Pointe de la Nunda; PR: Pointe de Ronce; SGMC: Signal du Grand Mont Cenis; AWS: Automatic Weather Station of Mont Cenis. In panel b is presented a high resolution orthoimage of the rock glacier obtained by UAV photogrammetry the 14th September 2016. In panel c is presented toponymy used in the present study to address the structure of the landform.

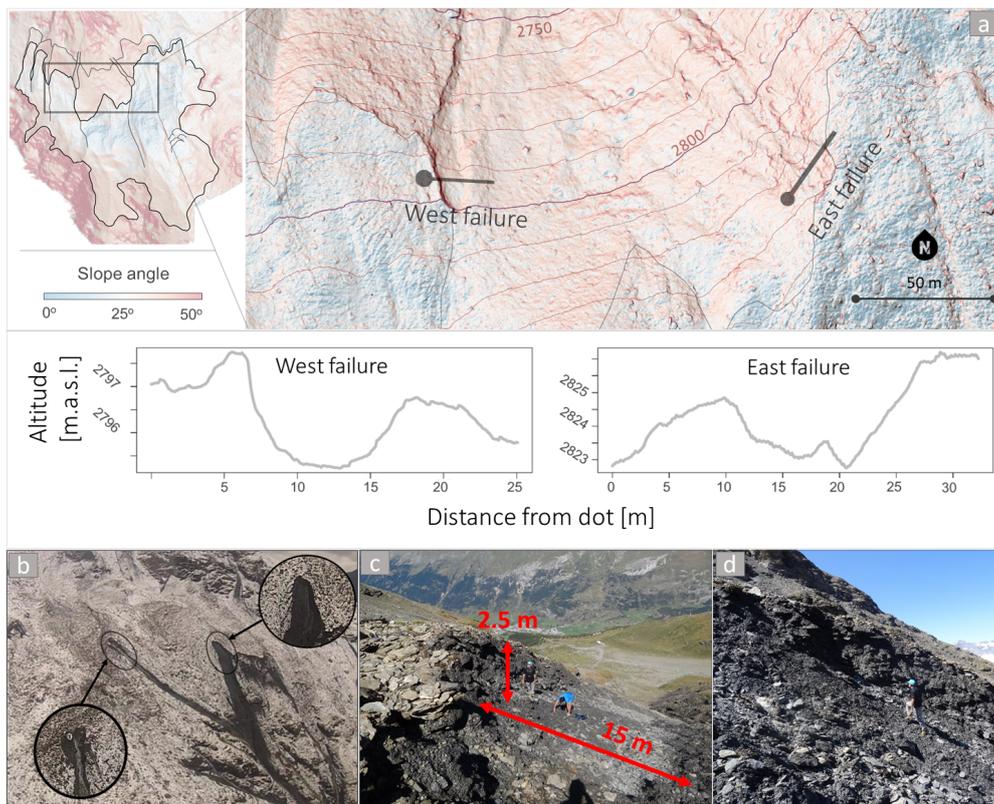


Figure 3. Overview of the frontal slopes failures. On top (panel a) are presented the morphometric characteristics of the two gullies. High resolution altitude, slope and hillshade data were obtained by UAV photogrammetry on September 14th 2016. On bottom left (panel b) is shown an aerial view of the failures taken the day after the event. On centre and right (panels c and d) is shown the west gully, characterizes by flat smooth surface and steep lateral embankments.

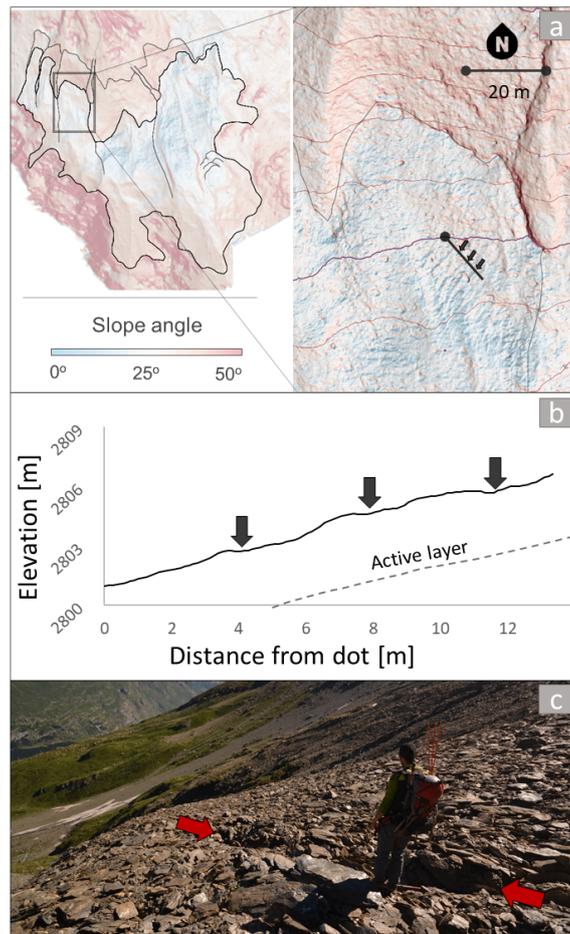


Figure 4. Example of cracks on the western lobe. The cracks are identified by the black arrows in the map (panel a) and on the topographical profile (b). Active layer depth is retrieved from seismic refraction, see section 4.3.2. On panel c is presented the appearance of one crack as it can be observed on the field.

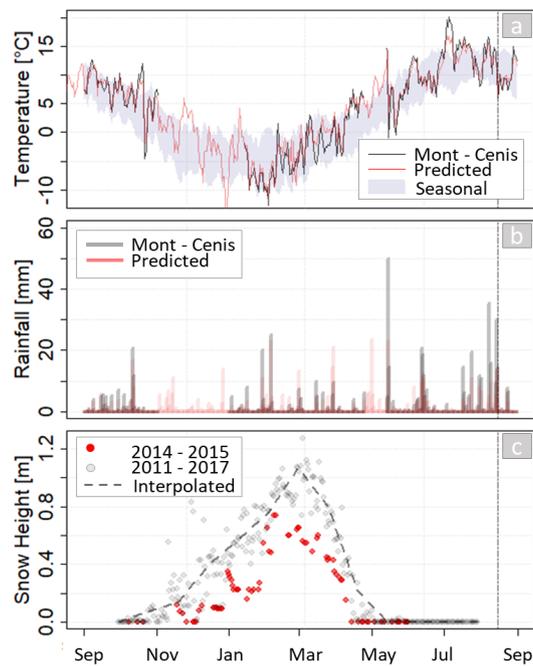


Figure 5. Overview of meteorological and nivological conditions in 2014-2015 prior to the failure event. Temperature and precipitation data (panels a and b) belong to Mont-Cenis weather station. Data from Bessans weather station were used to fill gaps in the data set. Nivological data (panel c) belong to Bessans weather station and are compared to data between 2011 and 2017 to observe the 2014-2015 winter anomaly. Data are not continuous as manually observed every 1 to 5 days. Interpolation between 2011-2017 was performed only for visual support.

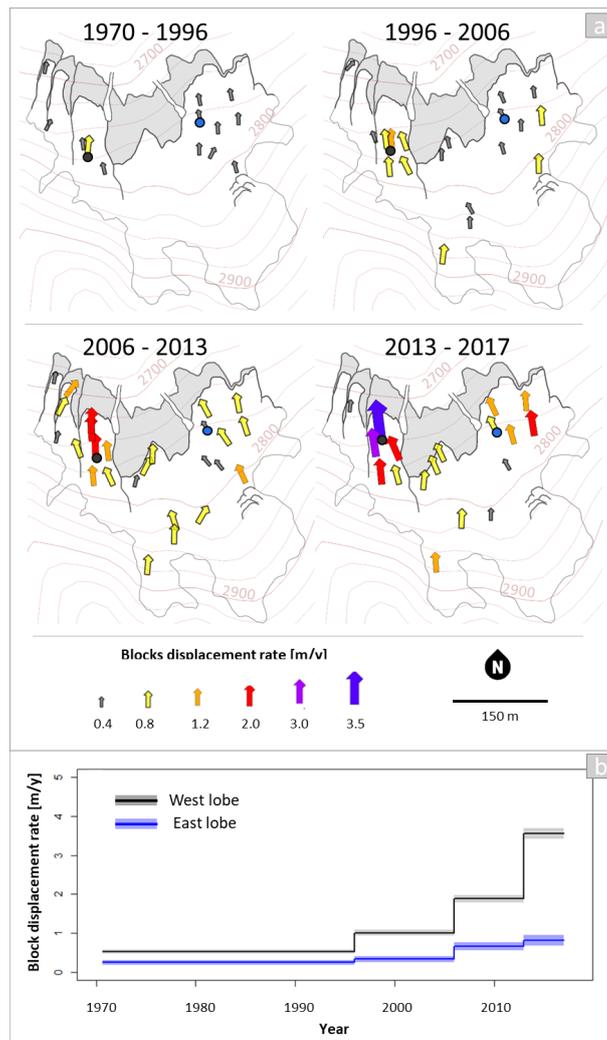


Figure 6. Overview of the historical dynamics of the Lou rock glacier presented as evolution of surface displacement rates retrieved by manual feature tracking on historical aerial imagery. In panel b is presented the graph of the evolution of the displacement rate on two boulders, one for the western lobe and one for the eastern lobe. Shaded areas size represent the mean error evaluated by apparent movements on stable areas. Boulder locations is identified by a dot in the maps in panel a.

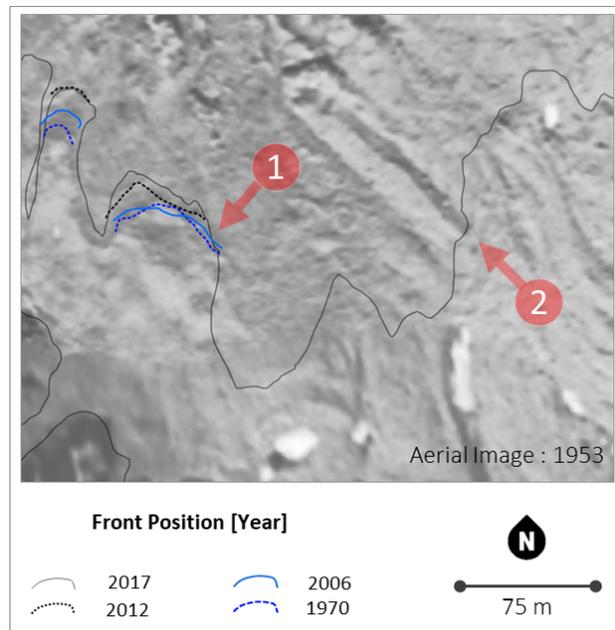


Figure 7. Overview of the advancement of the destabilized lobe front retrieved observable on historical aerial imagery.

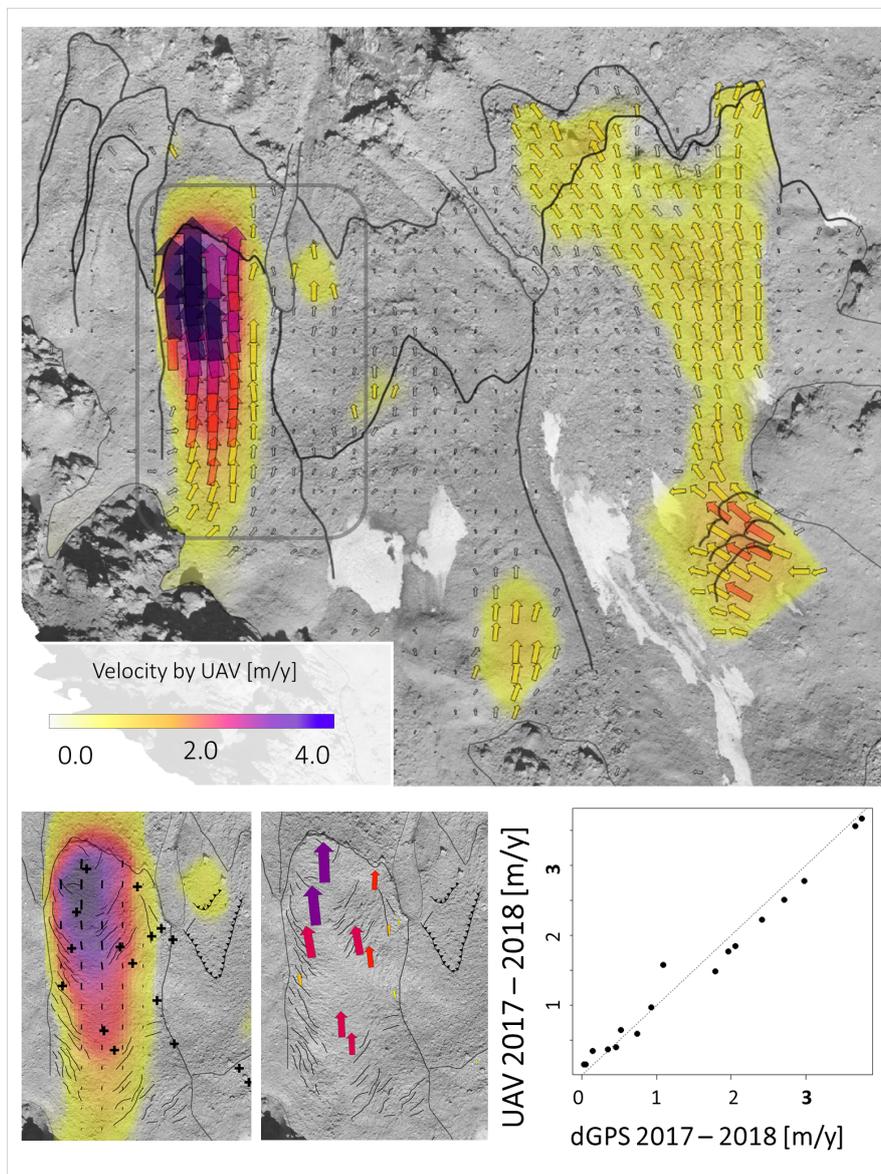


Figure 8. Overview of the dynamics of the rock glacier between 2017 and 2018. On top, measurements by UAV orthoimages comparison using automatic feature tracking in IMCORR. On bottom, comparison between UAV measurements and dGPS punctual measurements.

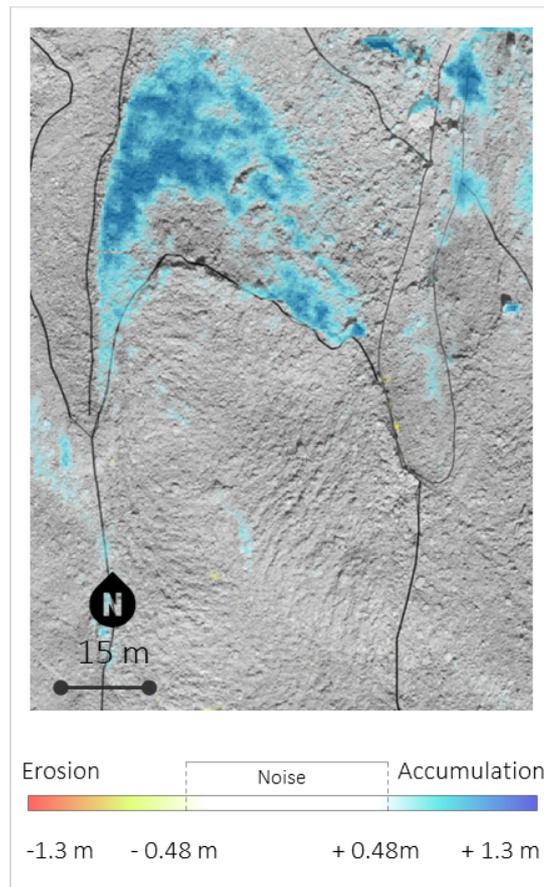


Figure 9. Surface variations at the front of the western lobe between 2016 and 2018 obtained by comparing high resolution UAV DEMs.

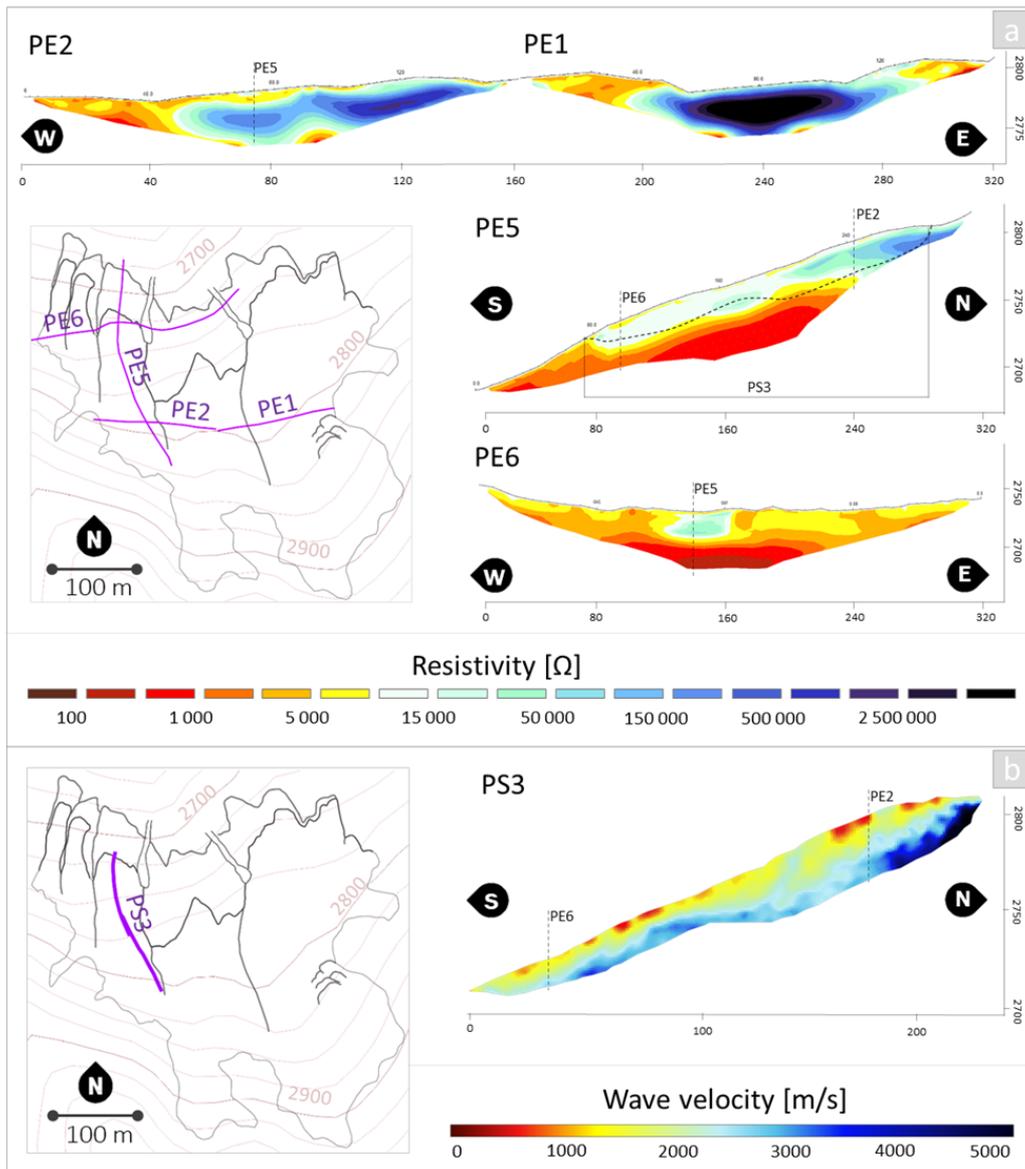


Figure 10. Overview of geophysical investigations at Lou rock glacier. On top (a), ERT transects acquired in 2016 (PE1 and PE2) and 2017 (PE5 and PE6). On bottom (b), SRT transect acquired in 2017.

Table 1. Summary of ERT transects.

| | PE1 | PE2 | PE5 | PE6 |
|----------------------------------|------------|------------|------------|------------|
| Dates of survey | 15/09/2016 | 15/09/2016 | 10/08/2017 | 10/08/2017 |
| Electrode array type | Wenner | Wenner | Wenner | Wenner |
| Electrode spacing | 2.5 m | 2.5 m | 5 m | 5 m |
| Number of electrodes | 64 | 64 | 64 | 64 |
| Number of data-points | 651 | 650 | 456 | 472 |
| Number of inverted points | 646 | 647 | 452 | 459 |
| Iterations | 4 | 4 | 4 | 4 |
| Absolute Error | 5.4 % | 6.3 % | 5% | 6% |