Rock Glaciers

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Abstract

Rock glaciers, a key element of alpine mountain geomorphic systems, consist of coarse surface debris that insulates an ice-core or ice-debris mixture. Rates of movement of active rock glaciers vary from 1 to more than 100 cm yr^{-1}. Rock glaciers exist in all major mountain ranges where permafrost occurs but are more common in dryer climates with high talus accumulation rates. New geospatial techniques, high-resolution data sources, and improved technology will contribute to a better understanding of these landforms. This chapter provides an in-depth summary of important research findings pertaining to rock glaciers and offers insight to future research.

Keywords

Climate change, Creep, Distribution, Field monitoring, Flow, Formation, Geospatial, Glacial, Hydrology, Periglacial, Permafrost, Photogrammetry and Remote sensing.

1.1 Introduction

High mountain systems, typically defined as having a large elevation ranges with a belt reaching above timberline that has been sculpted by glaciers and modified by periglacial processes (Troll 1973). Mountains are in the focus of geographic research as they influence weather and climate (Barry 2008), are an important source of fresh water (Immerzeel *et al.* 2020), and many processes are present which shape the mountains (Barsch and Caine 1984). Changes in mean annual air temperature (MAAT), net radiation patterns, snow cover distribution, glacier coverage and other variables will alter the processes modifying mountain systems (Beniston 2000; Beniston 2003; Slaymaker and Embleton-Hamann 2018). Humans, through anthropogenic climate change and recreational usage, threaten the nature and stability of high mountain systems (Barsch 1996; Konrad *et al.* 1999). Development in real estate projects, hiking paths, camping sites, transportation corridors, pipelines, ski lifts, communication towers, and utility towers continues to occur in alpine regions (Giardino and Vick 1987). As a result, mountains and humans have become delicately interconnected (Barsch 1996; Slaymaker and Embleton-Hamann 2018).

Rock glaciers are an important component of high mountain systems, characteristically serving as a visible indicator of mountain permafrost (Barsch 1996; Haeberli 2000) and occur at most mountain ranges on Earth (Table 1) (Figures 1–6) (Jones *et al.* 2018). Rock glaciers are important for the debris transport and storage in mountains (Humlum 2000; Gärtner-Roer 2012) and can also a source of hazards (Kääb *et al.* 2005; Delaloye *et al.* 2013; Schoeneich *et al.* 2015). They also store a significant amount of ice and can be of hydrological importance, especially in arid areas (Bolch and Marchenko 2009; Brighenti *et al.* 2019; Jones *et al.* 2019). Humans have used rock glaciers as a source for construction material, a backdrop for residential areas, dam abutments, drill sites, shaft and tunnel portals, and a water source for urban areas (Giardino and Vick 1987; Burger *et al.* 1999). Studying flow, age, ice content, formation, and climates of rock glaciers can provide a better understanding of geomorphic systems and processes in general and about the hydrological and geomorphological importance of rock glaciers and their response to climate change in specific.

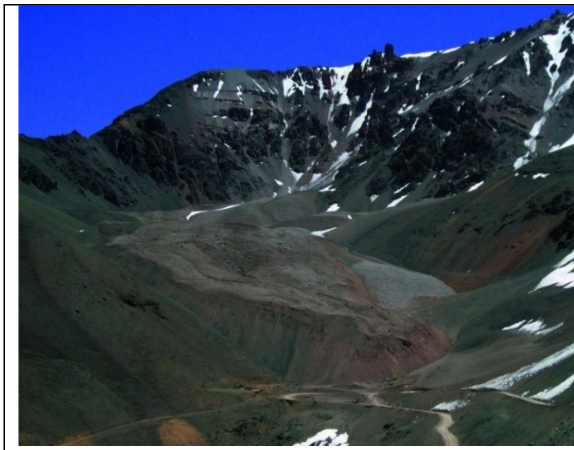


Figure 1 A rock glacier in Paso de Agua Negra, San Juan, dry Andes of Argentina. The road shown in the foreground is approximately 6 m wide.

Figure 2 Rock glaciers nested in multiple cirques in northern Iceland.

Figure 3 Murtél rock glacier at Piz Corvatsch, Upper Engadin, Switzerland. A 58 m borehole was drilled into this rock glacier in 1987. This represents one of the longest (and deepest) temperature measurements through alpine permafrost. (Photo: T. Bolch)

Figure 4 A rock glacier in the volcanic San Juan Mountains, CO, USA (Photo: JR Giardino)

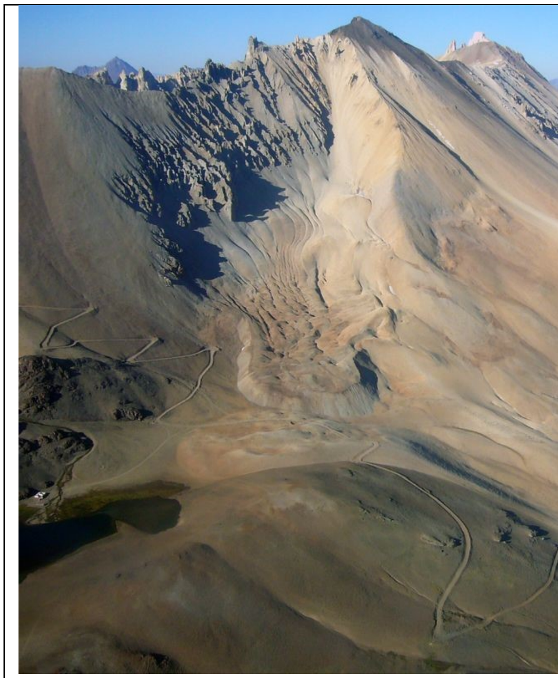


Figure 5 Cryogenic rock glacier near Las Leñas in the Central Andes of Mendoza, Argentina.

Figure 6: Ordzhonikisde rock glacier in Left Talgar Valley, Ile Alatau, Northern Tien Shan, Kazakhstan. Photo: V. Blagovechenskiy

1. 2 Definition and classification

Early studies defined rock glaciers as a 'dead glacier' (Steenstrup 1883), a 'peculiar form of talus' (Spencer 1900), a 'rock stream' (Cross and Howe 1905), or even 'movement of masses in a glacier-like way' which create a 'wrinkled' pattern (Capps Jr 1910). In Russian literature, they were also called "wooden glacier" (Goloskokov 1949). Later studies included new variables, describing rock glaciers as a 'special form of moraine' (Gerhold 1964), 'ablation complexes' (Johnson 1974), and an example of the 'creep of alpine permafrost' (Haeberli 1985). Though scientific inquiry and debate have allowed our comprehension of these unusual alpine landforms to advance, our understanding continues to be refined and improved.

The rock glacier nomenclature has caused much confusion (Hamilton and Whalley 1995; Berthling 2011). The word 'glacier' implies glaciation; however, rock glaciers have been identified in areas that have not been glaciated. Rock glaciers could also be confused with rock streams, rock avalanche deposits, talus, or rubble if identified by surface morphology alone, even though the process of movement is quite different for each (Giardino *et al.* 1992). Hence, the activity and movement of the landform and processes associated with it must be considered. Thompson (1999) suggested that an analysis of fabric could help discriminate among talus, moraines, protalus ramparts, and rock glaciers. Rock glaciers exhibit various transitional features common to the alpine landscape; this makes it difficult to form a general standard definition. However, most rock glaciers demonstrate lobate or tongue-shaped forms resulting from a combination of glacial, nonglacial, and periglacial processes. More specifically, rock glaciers may be characterized according to a combination of the following variables: ice content, debris input (talus or moraines), rates of flow, size, location on the hillslope, microrelief (ridges and furrows), and distribution, which is influenced by lithology, geographic location, topographic and microclimates, and other local environmental variables (Wahrhaftig and Cox 1959; Washburn 1980; Martin and Whalley 1987; Vitek and Giardino 1987; Whalley and Martin 1992; Barsch 1996; Haeberli *et al.* 2006). At present, the widely accepted rock glacier definition is 'the visible expression of cumulative deformation by long-term creep of ice/debris mixtures under permafrost conditions' (Berthling 2011).

Rock glaciers have traditionally been classified as active, inactive, and fossil (relict) (Barsch 1996). Activity status is important since it can provide information about current and past climatic conditions (Sailer and Kerschner 1999; Frauenfelder *et al.* 2001). Active rock glaciers are mainly free of vegetation and have movement rates in the order of few decimeters to meters per year. Sharp crested, cascading front slopes and defined surface features indicate recent movement and are the key structure to identify active rock glaciers. Inactive rock glaciers still contain ice, but their movement has ceased in response to climatic change (the ice is melting or snow input has changed) or they have extended too far away from their debris source (Barsch 1996). As a result, they have less pronounced surface features and front slopes. Rounded crests and stable fronts with a mature lichen cover are related to inactive rock glaciers (Calkin *et al.* 1987) (Figure 7). Inactive rock glaciers typically contain alpine grasses, subalpine dwarf shrubs, or small patches of short trees (Burga *et al.* 2004). Fossil or relict forms are vegetated, no longer contain ice, and display surface collapse features (Barsch 1996).

It is important to note that a clear distinction is often impossible. Many processes operate on active, inactive, and relict rock glaciers other than those that refer to movement or ice content. As a result, descriptions of the processes that are responsible for flow or creep are favored. For example, chemical and physical weathering is still occurring on all rock glaciers, regardless of the rock glacier's active, inactive, or relict flow/creep status. Furthermore, the classification inherently implies that no other environmental or ecological processes are present in some forms;

however, considering other ecological process, rock glaciers could provide an active habitat for animals, e.g., American pika populations (Millar and Westfall, 2010).

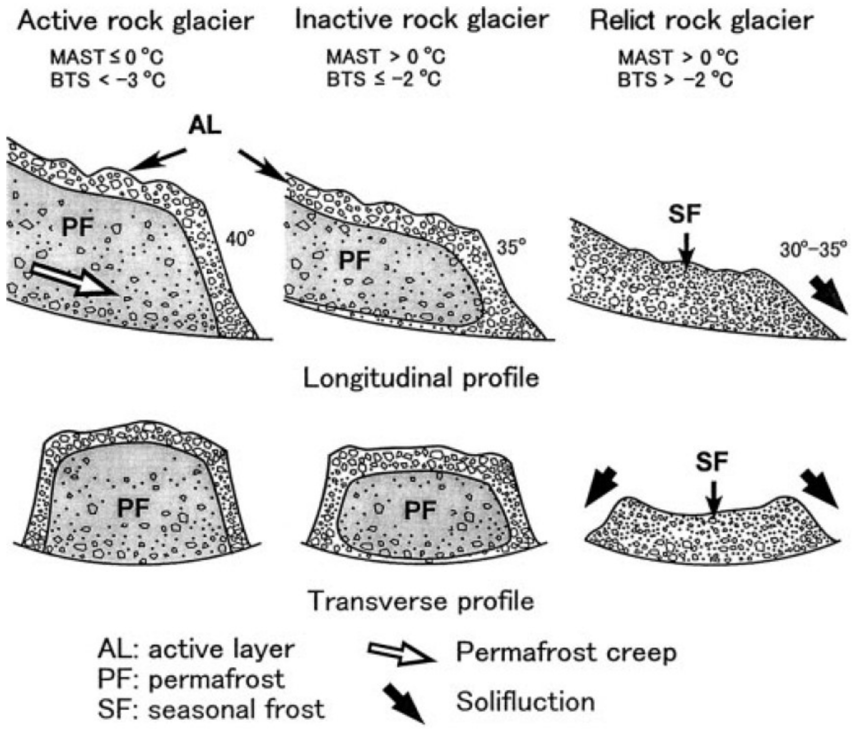


Figure 7: Schematic profiles of active, inactive, and relict rock glaciers. Source: Ikeda and Matsuoka (2002)

1. 3 Objectives

The objectives of this chapter are to (1) address the current state of knowledge applied to rock glaciers, (2) discuss the techniques used by researchers to study the rock glaciers, and (3) provide topics for future rock glacier research.

1. 4 Rock Glaciers as Part of the Mountain System

The surface of Earth, a patchwork of landforms, is shaped by various geologic and geomorphic processes that operate at different spatial and temporal scales. Rock glaciers are commonly a significant part of high mountain systems, consisting of summits and valleys with relief greater than 1000 m, rugged topography, and steep slopes that average 35° to 60° reaching elevations above the treeline. Traditionally, mountain slopes have been classified using distinct processes.

(Barsch and Caine 1984) designed a 4-tiered system with the following categories: (1) glacial, (2) coarse debris, (3) fine clastics, and (4) geochemical. The glacial system occurs at highest elevation and can transport debris from rockfall and erosion at the base of the glacier. The coarse debris system transfers material between cliffs, talus, and the features beneath them, including rock glaciers. Rock glaciers can contain a significant amount of debris/sediment (well above 20%) (Otto *et al.* 2009). In the coarse debris system, movement is largely from mass wasting. The fine sediment system is an open system with input from aeolian sources. The material is exported from the mountain system to the fluvial system. In the geochemical system, solution weathering is important because of its consistency, although solute concentration remains minimal (Krautblatter *et al.* 2012).

The mountain classification system does not imply an always active, cascading system of energy and debris downslope from peaks to valleys. Certain mountains exhibit poorly developed linkages with little coarse debris occurring at lower elevations (Barsch and Caine 1984; Gerrard 1990; Müller *et al.* 2014; Slaymaker and Embleton-Hamann 2018) (Figure 8). This is supported by a rock glacier's slow movement (typically in the range of centimeter to decimeter per year) in comparison to the other types of mass movements such as landslides, debris flows, mudflows, or avalanches. Much of the work done is infrequent and random in its timing, resulting in a pulsed impact (Barsch and Caine 1984). Catastrophic events such as rockfalls and avalanches do much of the work in isolated areas such as on steep slopes, in areas of winter snow accumulation, or near glaciers (Caine 1974; Barsch and Caine 1984; Krautblatter *et al.* 2012). Rock glaciers represent a slow-moving, dynamic system that can maintain equilibrium for thousands of years (Burger *et al.* 1999).

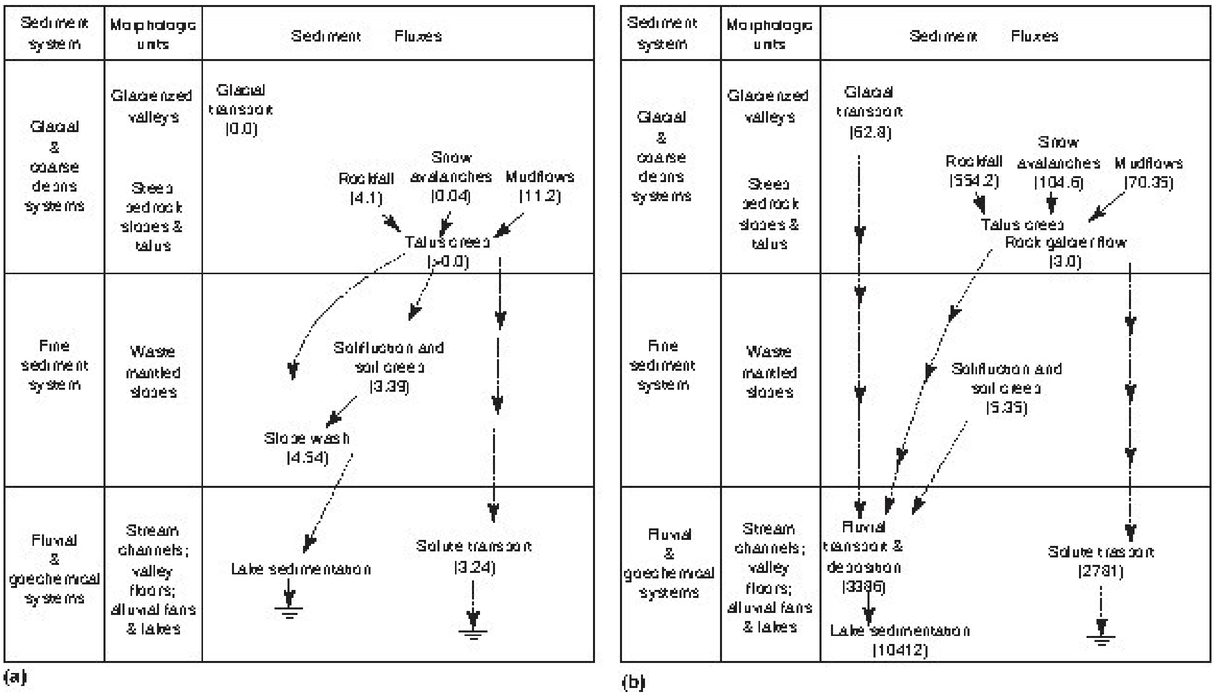


Figure 8 Sediment fluxes in two high mountain areas: (a) Williams Lake Basin, Colorado (Caine 1976) and (b) the Upper Rhine Basin (Jäckli 1956). Note that slow mass wasting accounts for less than 15% of the geomorphic work done. Their relative importance also decreases as the basin size increases. All fluxes are based on vertical transport (10^6 J km^2 yr^{-1}).
Adapted from Barsch, D., Caine, N. (1984).

1.5 The Rock Glacier System

As an individual component of the hillslope, rock glaciers can be described as an interconnected, cascading system of debris, ice, and water (Giardino 1979; Müller *et al.* 2014; Jones *et al.* 2019). The rock glacier system is comprised of four interrelated subsystems: cliff and talus, surface, subsurface, and outflow (Figure 9). The output from one subsystem provides input to another subsystem. For example, debris and meltwater from snow or ice are commonly transported from one subsystem to another. Surrounding cliffs provide coarse to fine talus inputs, depending on the patterns and density of headwall fractures and chemical and physical weathering of the surrounding bedrock cliffs (Haeberli *et al.* 1979; Giardino 1983; Müller *et al.* 2014). A rock glacier's continuous contact with the source area of talus helps maintain its form. If a rock glacier flows too quickly, the likelihood of loss of contact with the source area increases, and the subsystems will become disconnected. Snowmelt provides water to refreeze within the coarse debris of the rock glacier. The rock glacier acts as a storage unit for debris and ice but only transfers solutes out of the subsystem through summer melting (Jones *et al.* 2019). Coarse debris is characteristically contained within the system; however, glacial advances can transport debris out of the system. Other aspects of the system are open to external inputs and exhibit cascading characteristics. Aeolian material can be transported from other areas, deposited, and flushed from the system. Rock glacier solute potential has been shown to contribute to the geochemistry of alpine lakes and streams, which could affect other system's water quality (Vitek *et al.* 1981; Williams *et al.* 2006; Baron *et al.* 2009; Jones *et al. 2019)*

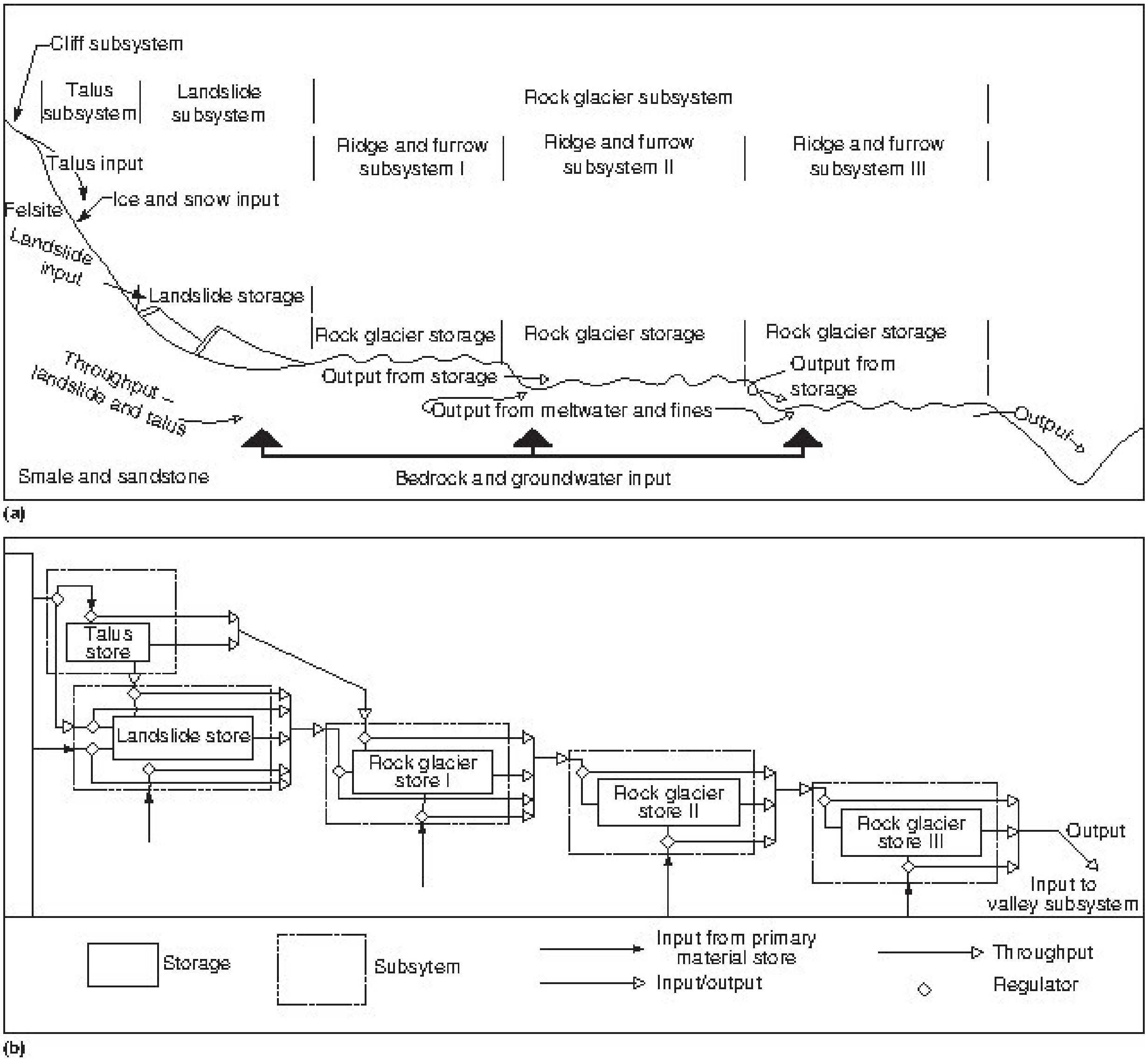


Figure 9 The rock glacier system (a) and storage and subsystems within the rock glacier system (b). Adapted with permission from Vitek, J.D., Giardino, J. (1987).

1.6 Form

From their research in the Alaska Range, Wahrhaftig and Cox (1959) provided a classification framework based on shape. Lobate rock glaciers have single or multiple lobes with a greater width than length, whereas tongue-shaped rock glaciers are longer, extending downslope, usually from a cirque. Spatulate rock glaciers resemble tongue-shaped rock glaciers but display an abrupt widening beyond lateral topographic constraints (resembles a piedmont glacier). Subsequent studies used classifications based on topographical or geographical descriptors such as valley-wall, valley-floor, protalus, debris, talus, or glacial (Outcalt and Benedict 1965; Linder and Marks 1985; Barsch, 1988; Barsch 1996; Humlum 1998). In the Swiss Alps, pebbly rock glaciers contain abundant matrix, but are smaller because of less resistant source rockwall. In contrast, bouldery rock glaciers (clast sizes 15–20 cm) with less matrix and less resistant source walls are smaller (Ikeda and Matsuoka 2006). A single classification scheme has not gained widespread acceptance; however, a subset of Wahrhaftig and Cox's (1959) classification most commonly appears in the literature (Figure 10).

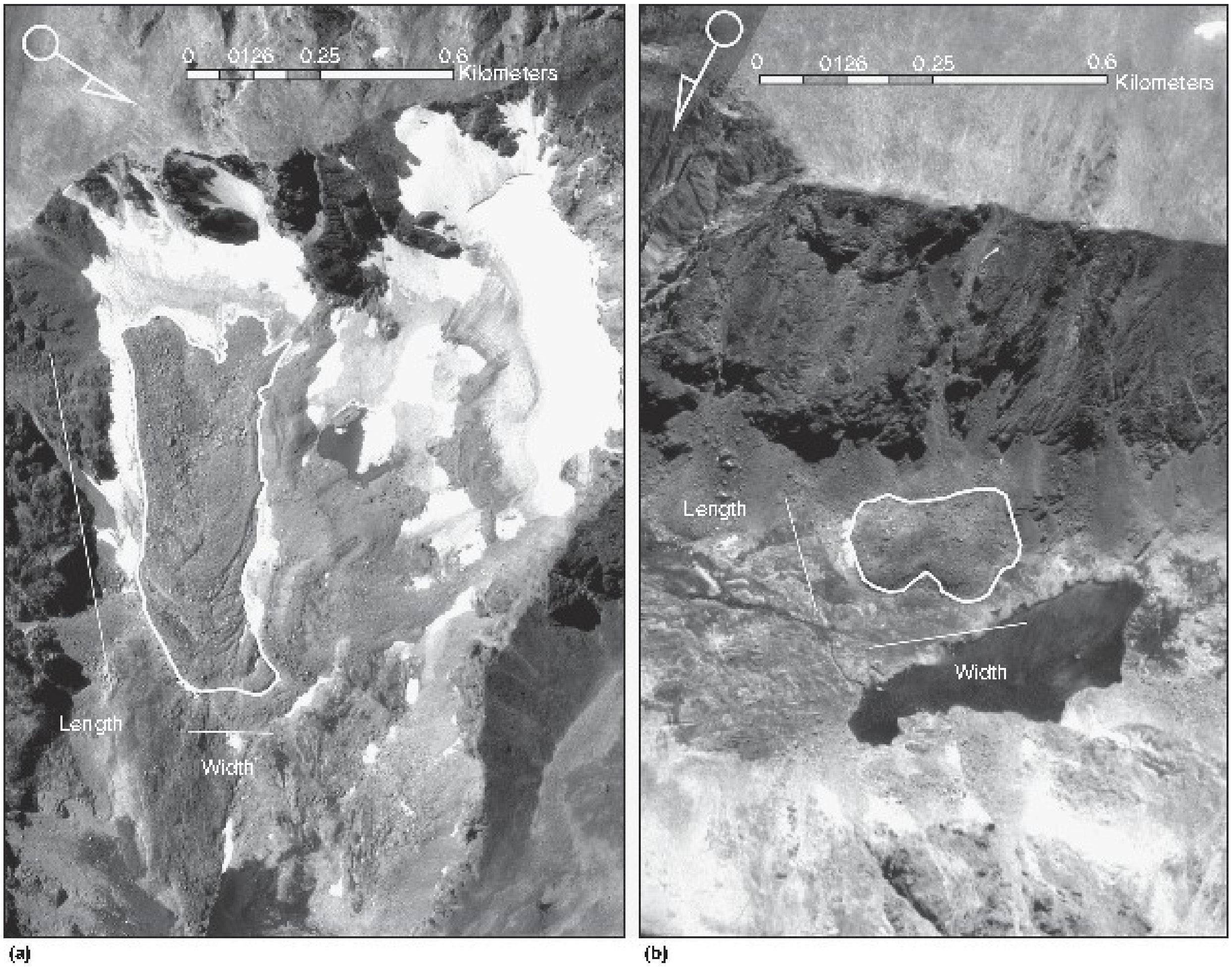


Figure 10 Typical tongue-shaped (a) and lobate (b) rock glacier forms are illustrated by aerial photos of the tongue-shaped Arapaho rock glacier and Green Lake 5 lobate rock glacier in the Front Range, CO.

1.7 Surface Morphology

The surface of a rock glacier consists of mostly mixtures of poorly sorted, angular, blocky rock debris at slopes that are often less than 25°. Rock glaciers that are flowing or creeping exhibit pronounced ridge and furrow complexes, resembling a viscous substance such as pahoehoe lava (Vitek and Giardino 1987; Barsch 1996). Active rock glaciers that are moving are mostly devoid of vegetation and comprised of distinct ridges and furrows. Transverse ridges and furrows form perpendicular to the direction of movement, appear bowed in nature, and originate from over-thrusting of internal shear planes, differential movement of distinct layers, or changes in debris supply (Wahrhaftig and Cox 1959; Haeberli 1985; Burger *et al.* 1999; Kääb and Weber 2004); Janke, 2005d). Longitudinal ridges and furrows form parallel to the principal direction of movement and result from extensional flow, resistance to flow, or are remnants of lateral moraines (Calkin *et al.* 1987; Ackert 1998)).

Rock glaciers exhibit a combination of transverse and longitudinal ridges and furrows. On the edges of the rock glacier, longitudinal ridges may occur, whereas toward the middle, they may be transverse. Ridges and furrows are commonly arched downslope in the lower parts of the rock glacier. In addition, rock glaciers may have a tightly meandering furrow that extends lengthwise down the rock glacier (Wahrhaftig and Cox 1959). Because of thermokarst or ice ablation, the surface morphology of rock glaciers can also appear mottled with spoon-shaped depressions (Washburn 1980). Advancing (active) rock glaciers have a steep front slope near the angle of repose with large boulders present at their toes, but near the head or rooting zone (similar to a bergschrund near the head of a glacier), there is a gradual transition from the source of debris input to the rock glacier (Figure 11) (Vitek and Giardino 1987; Barsch 1996; Müller *et al.* 2014). In general, the curvature is a concave upward profile near the head of the rock glacier and convex toward the toe. Visible surface structures and supporting information about the internal structure can explain the development and movement of rock glaciers. Figures 11–16 illustrate some of the features that characterize rock glaciers.

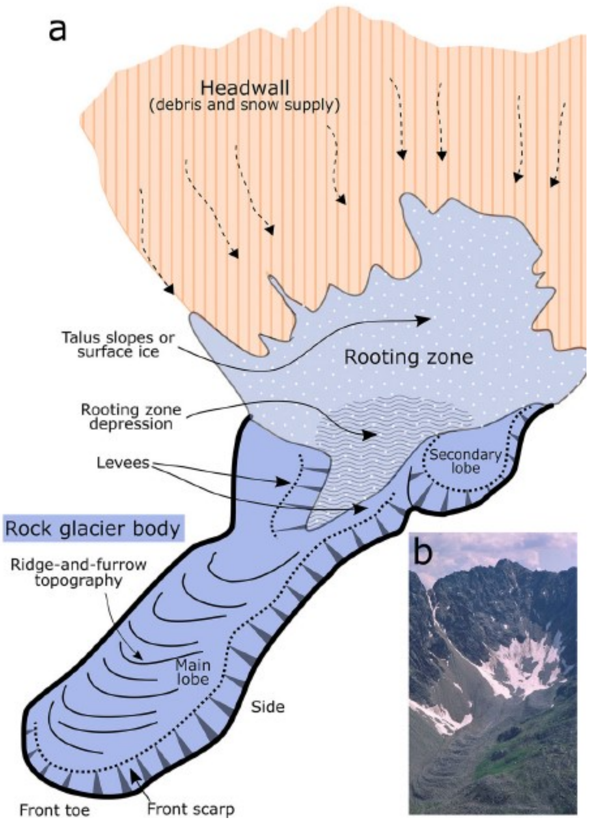


Figure 11 Schematic diagram illustrating the main morphological components of a typical tongue-shaped rock glacier and (b) the corresponding field photograph. Buco del Cacciatore, Valtellina, Central Italian Alps (photo courtesy of Mario Butti). Source: (Brardinoni *et al.* 2019)

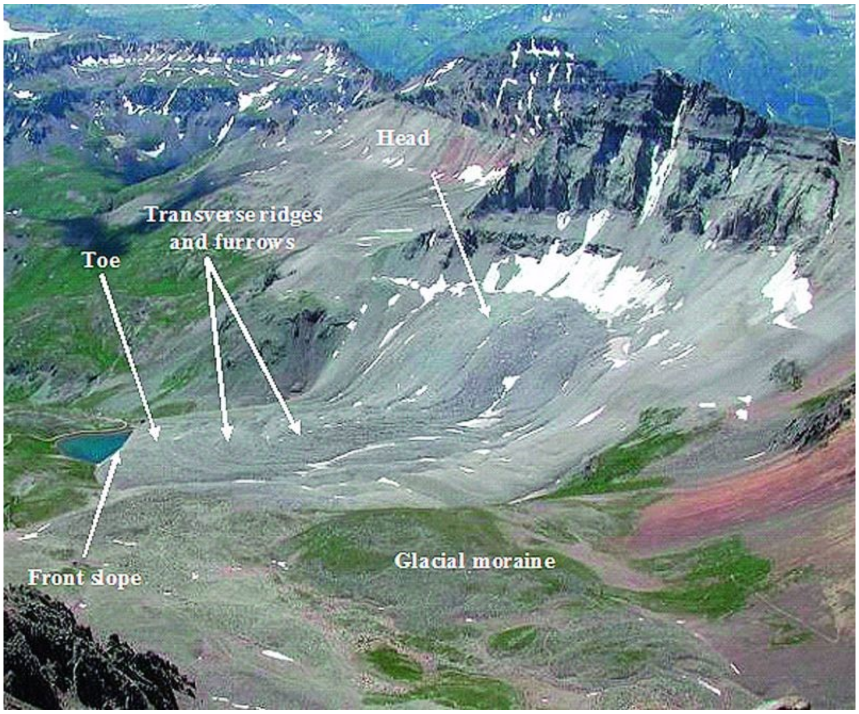
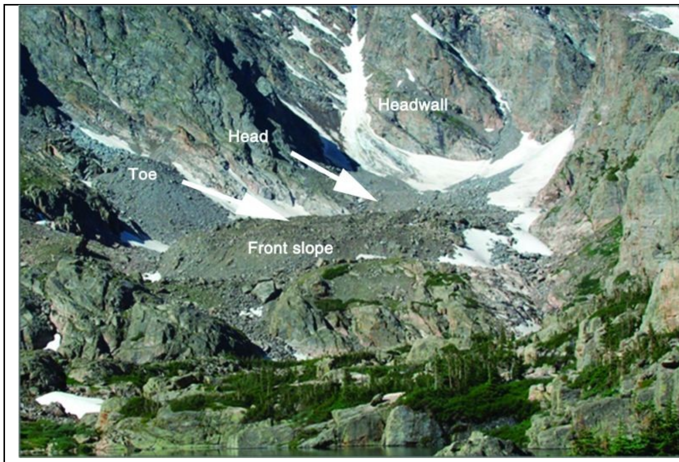
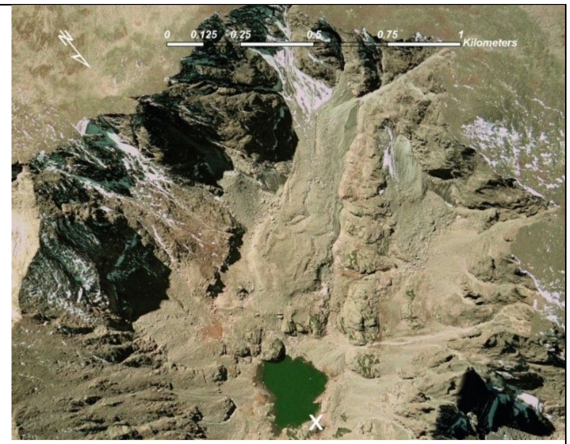


Figure 12 Some surface features of the tongue-shaped Yankee boy basin rock glacier in the San Juan Mountains, CO. Adapted from Degenhardt, J.J. and Giardino, J.R. (2003).

Figure 13 This rock glacier at Larsbreen, Svalbard, exhibits a unique longitudinal ridge and furrow complex. The ridges emanate from convergence of talus cones (not shown), whereas the furrows are associated with slower deposition from divergence between cones.
Adapted from Humlum, O. (1997).

 **Figure 14** A terrestrial photograph of the tongue-shaped Taylor rock glacier in Rocky Mountain National Park, CO. A small lobate rock glacier emanates from the corridor to the left. (Photo: J. Janke)	 **Figure 15** An aerial photograph of Taylor rock glacier. Vantage point of the previous photo is shown with an 'X.'
 **Figure 16** Front slope near the toe of California rock glacier in the Sangre de Cristo Mountains of Colorado (slope ≈>35°; height ≈45 m). (Photo: J. Janke)	

1.8 Rock glacier mapping and inventorying

Exact knowledge about the occurrence of the rock glaciers is important for many reasons such as creating a understanding of the mountain systems, the hydrological importance of rock glaciers, and improve hazard assessments. The distinct characteristics described here allow rock glaciers to be identified and mapped in the field and using remote sensing data. In earlier times, aerial photographs were the main source for rock glacier mapping (e.g., (Wahrhaftig and Cox 1959; Titkov 1988; Gorbunov and Titkov 1989; Guglielmin and Smiraglia 1998), while nowadays satellite images are used. The increasing availability of high-resolution satellite images and GoogleEarth allows even smaller rock glacier to be detected and larger areas to be mapped. Our understanding of rock glaciers has improved in recent years, in particular for the European Alps (Cremonese *et al.* 2011; Kellerer-Pirklbauer *et al.* 2012; Krainer and Ribis 2012; Scotti *et al.* 2013; Colucci *et al.* 2016), Rocky Mountains (Charbonneau and Smith 2018), Andes (Esper Angillieri 2009; Falaschi *et al.* 2014; Rangecroft *et al.* 2014; Falaschi *et al.* 2015; Janke *et al.* 2015; Barcaza *et al.* 2017), Carpathian mountains (Onaca *et al.* 2017), High Mountain Asia including the Tibetan mountains and the Himalaya (Schmid *et al.* 2015; Jones *et al.* 2018; Reinosch *et al.* 2021) and the Tien Shan (Bolch and Gorbunov 2014; Wang *et al.* 2017; Blöthe *et al.* 2019; Bolch *et al.* 2019; Kääb *et al.* 2021). However, although rock glacier inventories exist for many parts of the Earth, larger areas are still missing.

Almost all of the available inventories were generated manually by on-screen digitizing. However, mapping approaches and definitions are varying amongst the studies and inventories. A promising method to automatically and objectively map rock glaciers using satellite data and GIS modelling was proposed by (Janke 2001). (Brenning 2009) used digital elevation data along with satellite imagery and statistical models to detect rock glaciers. Modern machine learning techniques, combined with the availability of high-resolution data, allow further automatization of rock glacier mapping (Robson *et al.* 2020). Although these methods are promising, the uncertainties are still large and require manual correction if an accurate inventory is to be generated. While the rock glacier fronts are usually identifiable, the rock glacier head is often difficult to define due to a smooth transition to the upslope area. An intercomparison exercise revealed large deviation of the rock glacier outlines digitized by different experts especially in the upper regions (Brardinoni *et al.* 2019). Additional information such as differential interferometric synthetic aperture radar (DInSAR) or hillshades derived from high-resolution DEMs supports the mapping of rock glaciers (Barboux *et al.* 2014; Necsoiu *et al.* 2016; Wang *et al.* 2017; Villarroel *et al.* 2018 Kääb *et al.* 2021) (Figure 17).

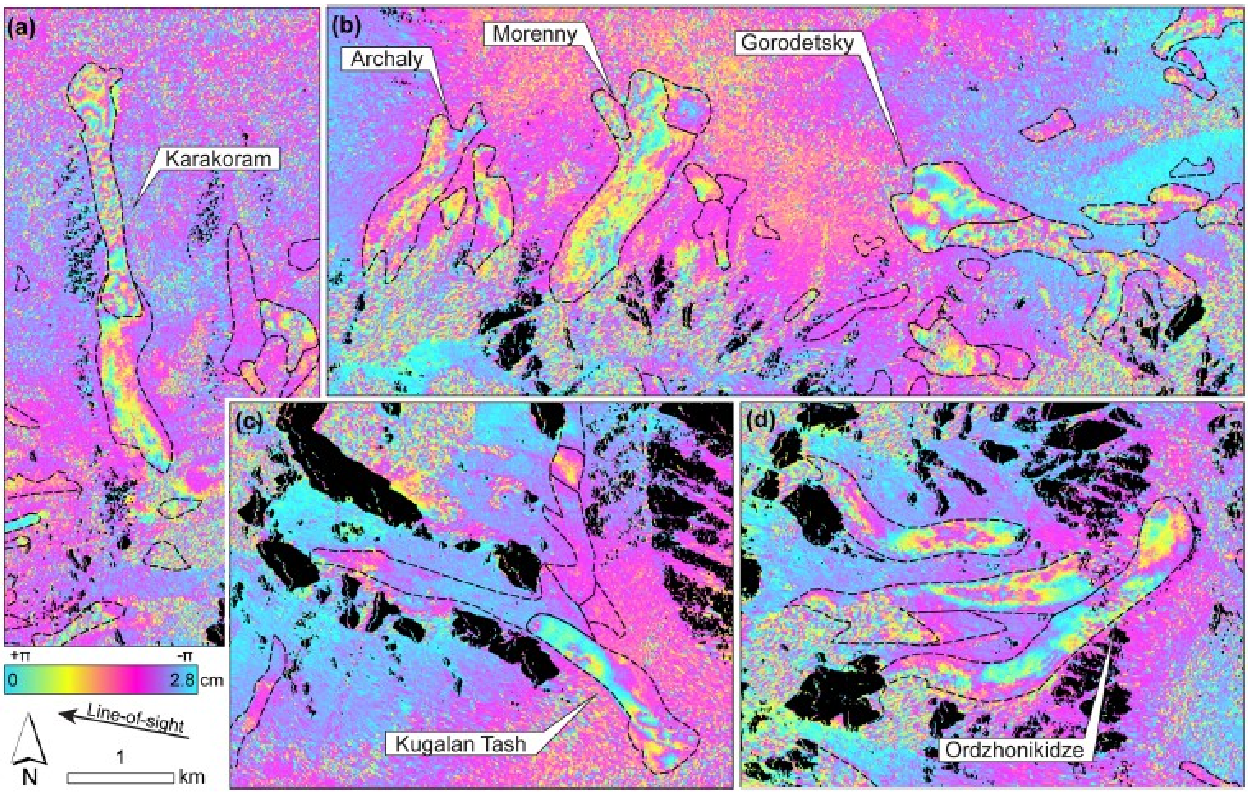


Figure 17 Radar Interferograms to support rock glacier mapping (Example from Northern Tien Shan); Source: Kääb *et al.* 2021

A rock glacier inventory should include the outlines and information about the location and area as well as topographic information which can be obtained from available digital elevation models. Another important variable is the activity of the rock glacier. The activity can be estimated based on the surface morphology (see previous sections) and vegetation cover. Moreover, total plant cover and floristic composition vary between inactive and active forms. Unstable sites are characterized by vascular species tolerating surface instability (Cannone and Gerdol 2003). Inactive rock glaciers typically contain alpine grasses, subalpine dwarf shrubs, or small patches of short trees (Burga *et al.* 2004). However, vegetation might not grow in arid or very cold alpine regions, and hence a clear distinction is often not possible using single images as the only source of information. Activity status is further complicated by lobes of activity superimposed on the surface of previously active rock glaciers or slope deposits (Kellerer-Pirklbauer *et al.* 2008). Therefore the term "intact" can be used for both active and inactive rock glaciers is preferred (Brenning 2005; Kofler *et al.* 2020). Although it is relatively difficult to calculate, the basal shear stress can be used to differentiate active and inactive categories. Giardino and Vick (1987) determined that basal shear stress for active rock glaciers lies between 1.0 and 2.0 bars, and Wahrhaftig and Cox (1959) determined that basal shear stress for inactive rock glaciers is less than 1.0 bars. Remote sensing allows the clearest information about the activity, especially InSAR can detect even a centimeter range displacements (Barboux *et al.* 2014). The working group for rock glacier mapping recommendations is currently working on standardized guidelines and recommends both using high-resolution optical data and InSAR measurements (IPA Action Group 2020).

8.17.9 Origin and Internal Structure

Geophysical investigations (such as ground penetrating radar, seismic refraction, or electrical resistivity tomography) and boreholes from a few rock glaciers provide insights into the internal structure of rock glaciers. It was found that the structure can be generalized as a three-tiered system with a top layer of rock fragments with a thickness of about one to a few meters covering an ice-cemented or ice-cored interior with a thickness in the order of tens of meters, and finally, a layer with low or no ice content that overlies rock deposited and overridden by the top layers (Humlum 2000; Hausmann *et al.* 2007; Springman *et al.* 2012) (Figure 18).

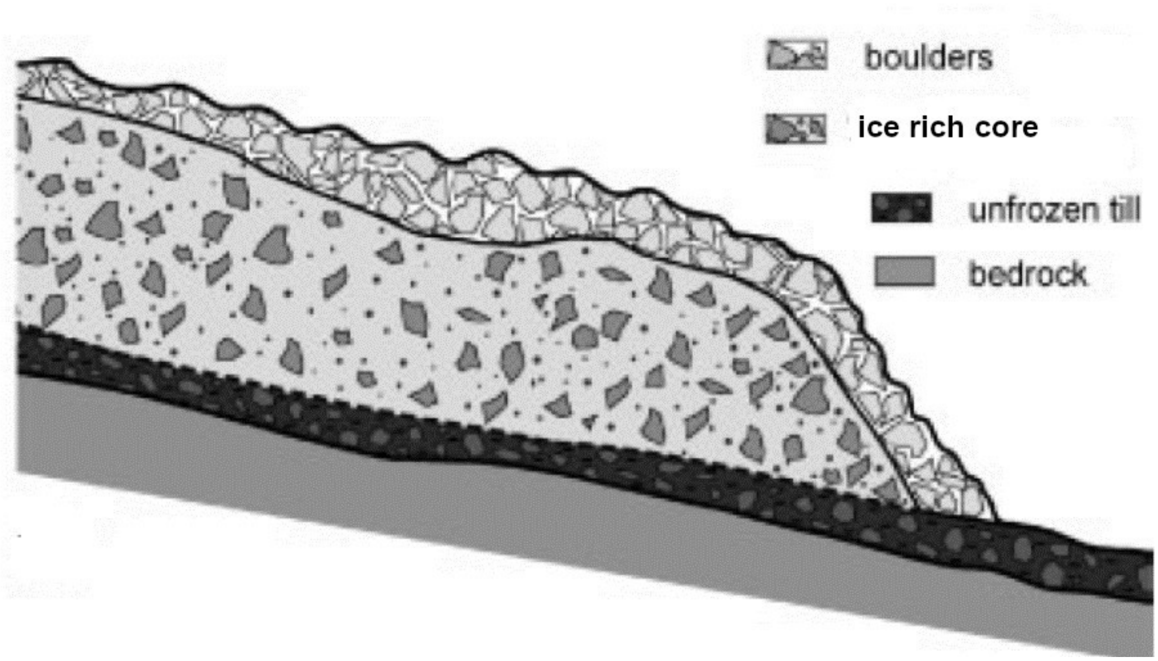


Figure 18 Schematic figure of the structure of a typical rock glacier. Note: The core consists of variable ice and debris content. Source: after Hausmann et al. 2007.

A rock glacier's periglacial or glacial origin is commonly debated, relating to whether or not rock glaciers have an ice-cemented or ice-cored internal structure (Figures 19 and 20). Supporters of the periglacial model believe that cemented ice in the form of interstitial ice (pore-ice) or segregated ice (ice lenses) produces creep (Wahrhaftig and Cox 1959; Barsch 1996; Haeberli *et al.* 2006*)*. Giardino (1979) hypothesized that hydrostatic pressure could help explain the movement of ice-cemented rock glaciers. Supporters of a glacial model advocate an ice-cored structure in which a relict glacier has been buried by debris (Potter 1972; Whalley 1974; Whalley and Azizi 1994; Potter *et al.* 1998). Abundant evidence supports each of the models.

1.9.1 Ice-cored Structure

Evidence for an ice-core has stemmed mainly from the research of rock glaciers in North America, particularly Galena Creek rock glacier in the Absaroka Mountains of Wyoming where the debate began when Potter (1972) observed an ice-core in the rock glacier. Using seismic data, he found evidence of a 1-1.5 m thick debris mantle underlain by an ice core with a thickness of 40-80m. More detailed investigations confirmed these initial findings (Potter *et al.* 1998). Larger variability of the debris mantle had a tendency to thicken down slope. Further evidence was found by Clark *et al.* (1996), who recovered a 9.5 m core of continuous ice. The 2000-year-old ice-core consists of 90% clean, bubbly ice and shows a good correlation with late summer ablation layers for an ice-core obtained from South Cascade glacier in Washington. Peaks in $\delta^{18}O$ correspond closely with the location of the debris bands; this represents warm precipitation from the late summer or early fall and correlates well with records from other temperate glaciers. Recent detailed geophysical investigations using GPR revealed the existence of reflectors that form a network of concave layers dipping upslope (Petersen *et al.* 2020). The authors interpret these as debris bands formed by debris falls and subsequently buried by snow and ice accumulation.

Benedict (1973) also found ablation surfaces on Arapaho rock glacier in the Colorado Front Range. The ablation surfaces were dipped up-valley at angles of 32 to 60°, suggesting a glacial origin. Arapaho rock glaciers could have formed during a catastrophic rockfall event that covered the former glacier, given the steep cirque walls and abundance of massive boulders located near the head (Whalley 1974). Similar processes have been observed in Scotland, where landslide debris is believed to cover and form rock glaciers (Ballantyne and Stone 2004). GPR data in the San Juan Mountains of Colorado suggests a similar formation method for some rock glaciers (Degenhardt 2009). A new core drilled in the rock glacier Murtèl (Swiss Alps) in 2015 also revealed a core with very high ice content (unpublished data).

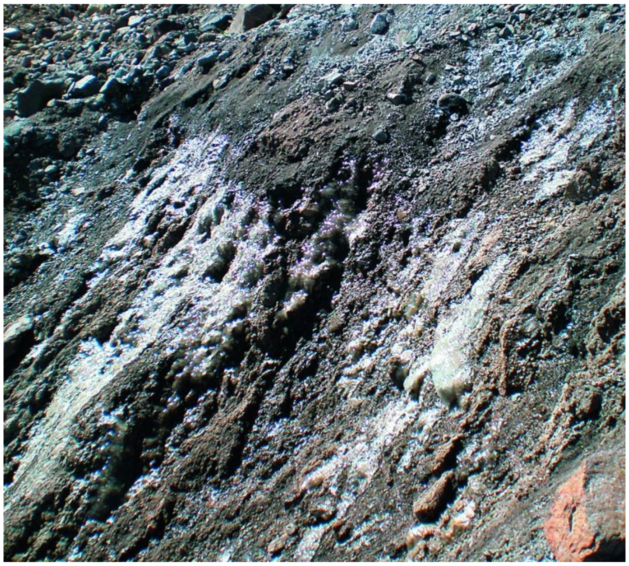


Figure 19 Interstitial ice found in the Elqui Catchment near the Las Tórtolas Mountain, Chile.

Figure 20 Internal ice-core structure on a debris-covered/rock glacier complex in the Pamir (Tajikistan).

1.9.2 Ice-Cemented Structure

Evidence of ice-cemented structures comes from various sources (Wahrhaftig and Cox 1959; Haeberli 1985; Barsch 1987; Barsch 1996; Moore 2014). In their survey of rock glaciers in the

Alaskan Range, Wahrhaftig and Cox (1959) hypothesized that interstitial ice within the rock glacier forms from compacted snow, freezing of water derived from melting snow or percolation of rain, or groundwater that rises beneath the talus and freezes on contact with cold air. Barsch (1977) stated that there is no need for glacier ice if talus production is high and snow avalanches and refrozen meltwater are available. Based on existing data and seismic investigations on Gelena Creek rock glacier, Barsch (1987, 1988, 1996) concluded that the landform is a multiunit rock glacier. The permafrost model is based on the presence of talus. Talus is incapable of much movement unless ice is present. To generate enough ice capable of movement, segregated ice can form where water flows over the permafrost layer. By hydrostatic pressure, water remains in contact with the permafrost, and thus ice lenses can grow (Whalley and Martin 1992). Perhaps the strongest argument relies on the fact that rock glaciers have been identified in areas that have not been glaciated. For example, Barsch and Updike (1971) mapped fossil rock glaciers at Kendrick Peak, AZ, and San Mateo, NM in areas that were not glaciated, thus must be of periglacial origin. Mt. Mesta rock glacier in southern CO is another example of a rock glacier that formed in nonglacial conditions (Giardino and Vitek 1988b). In the Sleza Massif in southwest Poland, relict rock glaciers are likely of periglacial origin (Zurawek 2002).

1.9.3 Continuum

Because abundant evidence exists for each of the models, others have supported the idea of a landscape continuum, a cycle describing the transition among glaciers, rock glaciers, and slope deposits, or a composite model where a combination of processes is responsible for creating rock glaciers (Shroder 1987; Giardino and Vitek 1988a; Corte 1987; Johnson 1987; Janke *et al.* 2015; Monnier and Kinnard 2017; Anderson *et al.* 2018). Capps (1910) alluded to this idea by suggesting that movement started with refreezing and expansion of the glacier, then periglacial processes driven by snow and meltwater formed the interstitial ice. Others have acknowledged the fact that both permafrost and glacially derived rock glaciers can exist. Haeberli and Vonder Mühll (1996) argued that ice of glacial origin would disappear and be replaced by nonglacial ice through time. In the Front Range, Outcalt and Benedict (1965) suggested that rock glaciers resting on the cirque floor are ice-cored, and those that are valley-walled formed from interstitial ice or metamorphism of snow beneath rockfall. Humlum (2000) found that rock glaciers in

Greenland were either talus-derived (periglacial) or glacial-derived (glaciers). He added that notwithstanding this debate, most scientists agree that rock glaciers are the result of a localized, high supply of talus in a permafrost environment. Therefore, it is likely that even if a rock glacier has a glacial ice-core, it is likely that periglacial processes still affect the rock glacier. Steig *et al.* (1998) presented evidence that both glacial and periglacial processes may be acting on the Galena Creek rock glacier. After plotting deuterium (δ D) versus oxygen ($\delta\,^{18}O$), the values fall close to the global meteoric water line. This suggests that the observed isotope stratigraphy largely reflects meltwater infiltration and refreezing rather than runoff derived from glacier ice.

Physical evidence of both permafrost and glacial ice-cores has been noted on several rock glaciers. On a rock glacier in the Hurricane Basin, San Juan Mountains, Brown (1925) found interstitial ice above the glacier core. Resistivity measurements indicate that rock glaciers in the Sainte-Anne cirque have ice-cores near their heads (glacial origin) and ice-cemented materials near the toe (periglacial origin) (Assier et al. 1996). On the Foscagno rock glacier in the Italian central Alps, a lower section of an ice-core between 4 and 7.65 m indicates firn ice, whereas the elongated vertical bubbles in the upper-core between 2.5 and 4 m suggest melting and refreezing (Guglielmin *et al.* 2004; Stenni *et al.* 2007). On the Arapaho rock glacier, extending from the rock glacier mid-section to the toe, drifting snow, rain, and snow meltwater pass downward through the rock debris, freeze, and become interstitial ice. However, the up-valley half consists of glacial ice (White 1971; 1975). In a later paper, White (1976) stated that all ice-core rock glaciers have a saucer-shaped depression between the rock glacier and the headwall, a longitudinal furrow, and pits. Ice-cemented rock glaciers are differentiated from this because they do not have these features but often have longitudinal ridges. In fact, White (1976) stated that lobate rock glaciers occurring on the south side of the Front Range, commonly contain interstitial ice, and move only about 1 to 6 cm yr^{-1}.

Rock glaciers are a transitional form regardless of the processes responsible for its formation (Figure 21) (Giardino and Vitek 1988a). The rock glacier will eventually cease to exist and will become a deposit in the rock record. A clear differentiation between process and form must exist. The term ‘rock glacier’ refers to a form, not a process. Rocks may be in a location before the form is created, and the finality of form as it is expressed in the geomorphic record. Too often, rates of movement are assigned to the form that neglects the processes responsible for producing the motion.

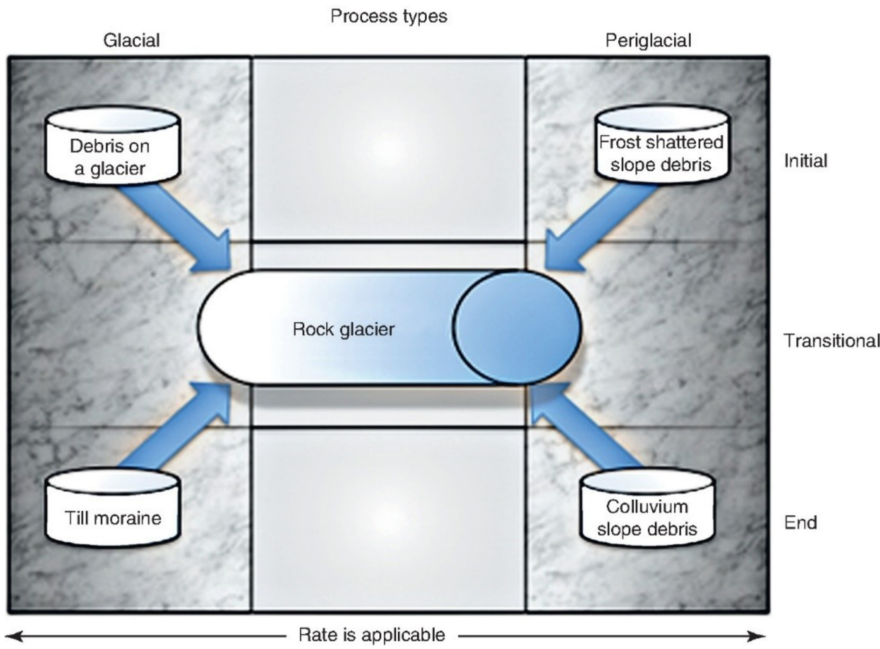


Figure 21 The alpine landscape continuum. Adapted from Giardino, J.R., Vitek, J. (1988a).

Recent research based on modelling and detailed remote sensing investigations (e.g. (Janke *et al.* 2015; Monnier and Kinnard 2017; Anderson *et al.* 2018) provide evidence that debris-covered glaciers can transition to rock glaciers, specifically with strong debris supply and inefficient sediment evacuation. The transition occurs specifically under persistently cold conditions. This is supported by recent work on the sedimentology of rock glaciers and related forms (Knight *et al.* 2018; Jones *et al.* 2019).

1.10 Distribution and Climate

Rock glaciers play an important role in the elevational distribution of landforms in glacial or periglacial environments (Caine 1984). Traditionally, rock glaciers were thought to exist in mainly low precipitation, continental climates where frost weathering is dominant and temperatures are cool enough to maintain ice (Barsch 1977; Haeberli 1983). In the Alps, Haeberli (1983) indicated rock glaciers exist in dry, continental climates at altitudes below the equilibrium line of glaciers but above the lower permafrost limit. Humlum (1998) and Janke

(2007) found that rock glaciers and glaciers locations are driven by topoclimates that favor high talus production rates, not regional climates (Figure 22). Temperatures over the surface of a rock glacier appear to vary greatly over short distances. A horizontal distance of less than 535 m and a 100 m change in elevation exist among the sites. Temperatures do not decrease with higher elevation on the California rock glacier in Colorado. In fact, site 3 is the warmest in the winter but coldest in the summer (Figure 22). A 1.6 °C mean difference exists during the winter, and a 2.8 °C mean difference exists in the summer. Ground temperature measurements at different depth at Murtèl rock glacier show that temperatures near the surface exhibit high annual variability with temperatures up to 15°C in 0.25 cm depth but temperatures remain below 0°C in depths greater than about 5 m (Figure 23)

When the distribution is mapped, rock glaciers provide insight into Holocene and Pleistocene climates (Barsch and Updike 1971; Harris 1994; Steig *et al.* 1998; Frauenfelder and Kääb 2000; Moran *et al.* 2016). In Arizona and New Mexico, Barsch and Updike (1971) concluded that permafrost once extended as low as 2600 m during Wisconsin times. In the Tyrolean Alps, Kerschner (1985) found that permafrost extended from 500 to 600 m lower than today with a 3.0–4.0 °C lower MAAT. In the Swiss Alps, Frauenfelder and Kääb (2000) found that temperatures were once 3.0–4.0 °C cooler, depressing the permafrost limit by 500–600 m. In Greece, relict rock glacier slope deposits indicate that MAATs were 6.3–9 °C cooler, in agreement with paleoclimatic reconstructions (Hughes *et al.* 2003; 2006). In Colorado, Janke (2005c) found that the climate was once some 3.0–4.0 °C cooler, with permafrost extending some 600–700 m lower than today.

Present climates affect rock glaciers by preserving or ablating an internal ice structure (Martin and Whalley 1987). Ordinarily, temperate alpine glaciers do not have clear ^{18}O and ^{16}O ratios for more than a few hundred years because of ablation, melting, and refreezing. However, a thick rock glacier debris cover slows these processes providing a longer interval of untapped climatic information (Barsch 1988; Haeberli *et al.* 1988; Haeberli 1990; Haeberli 1992; Clark *et al.* 1996; Steig *et al.* 1998; Konrad *et al.* 1999).

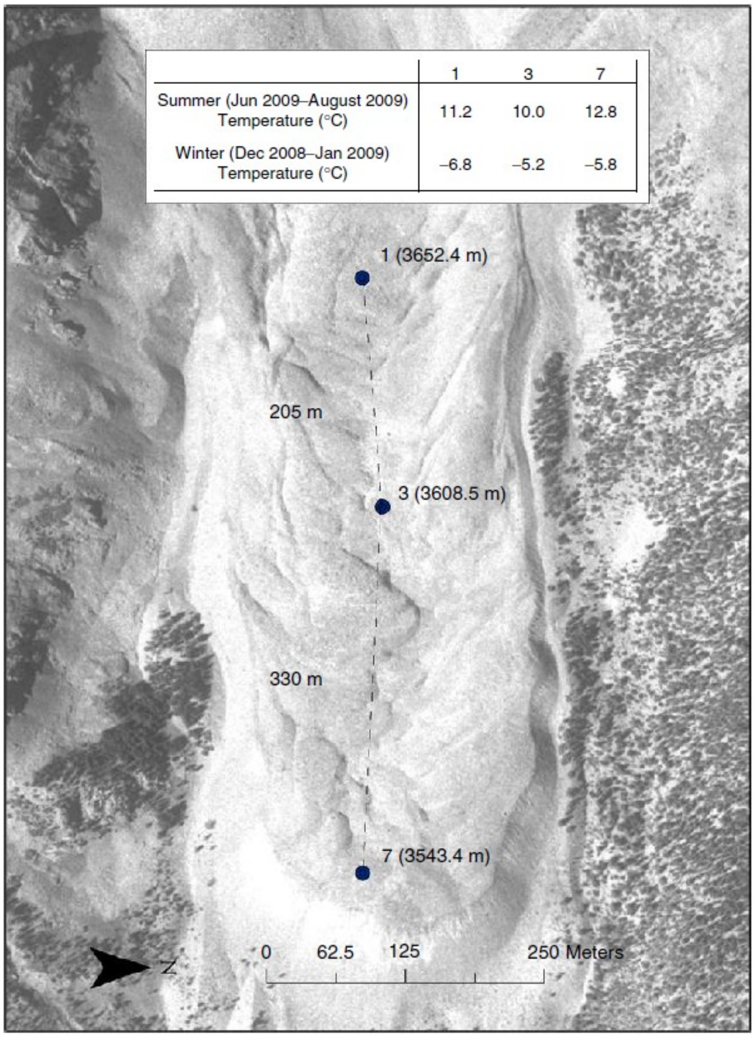


Figure 22 An example of the effects of microclimate on the California rock glacier, Huerfano valley, CO. Site 3 is the warmest in the winter but coldest in the summer which illustrates the

effect that rock glacier topography and surrounding cirque wall geometry have on microclimate formation.

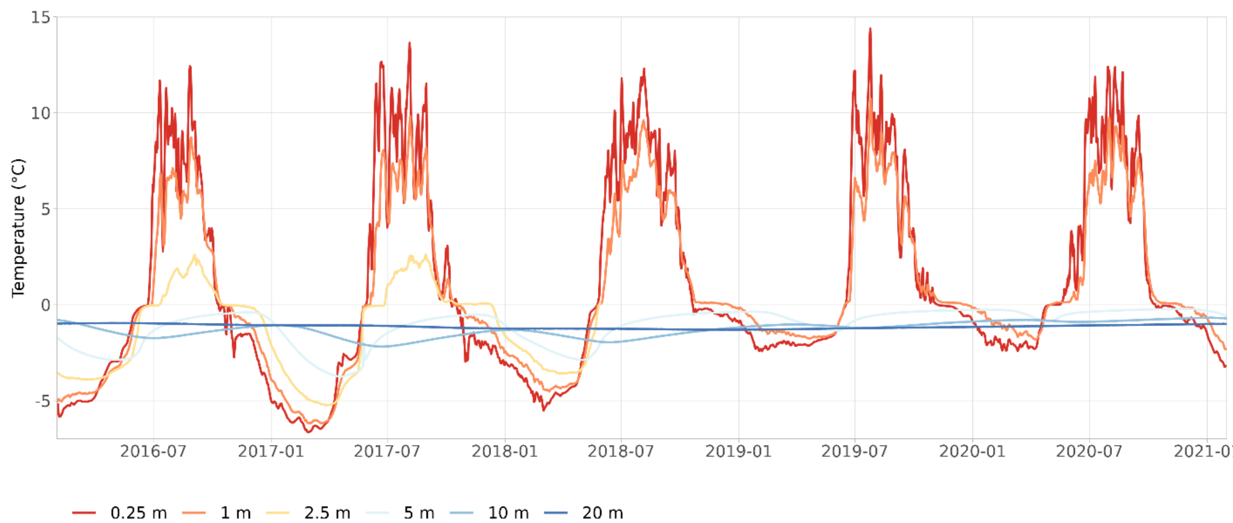


Figure 23 Temperature measurements in different depths from Murtèl rock glacier in the Swiss Alps from February 2016 until February 2021. Data source: Swiss Permafrost Monitoring Network (PERMOS).

1.11 Rock Glacier Age

The age of rock glaciers can be inferred from various techniques that include geomorphological climatic reconstructions, lichenometry, C-14 dating of organic remains trapped englacially, or weathering of surface clasts. Care must be taken when using these methods: it is important to determine whether the age refers to time of formation, time of rock deposition, age of the ice-cemented or ice-cored structure, or age of recent advance. Lichenometry has been used to date rock glacier surfaces based on lichen thallus size. Two different approaches, direct and indirect, are used. In the direct approach, growth rates are determined by monitoring individual thalli over several years, typically of abundant yellow green specimens (subgenus *Rhizocarpon geographicum agg*.). Growth rate curves can then be extrapolated to estimate age. The indirect approach uses surfaces of known ages to calibrate growth rates empirically. The accuracy of lichenometry dating is ±15 to 20% (Andrews and Miller 1972; Calkin and Ellis 1980; Bickerton and Matthews 1992; Solomina and Calkin 2003). Rates of lichen growth are affected by environmental factors such as snow, which slows growth or kills lichens, and talus, which disintegrates lichens. Even over short distances, *R. geographicum* growth rates can be quite different. This indicates that microclimates influence growth rates, which makes it difficult to extend growth rates over several decades (Bradwell and Armstrong 2007).

The relative age of a rock glacier can be determined by measuring the hardness of surface clasts using the Schmidt hammer test: weaker rocks exposed to more weathering are likely older (Haeberli *et al.* 2003). This method assumes rock hardness changes in a unidirectional manner; therefore, it may not correlate with other age indicators for a given locality (Burger *et al.* 1999). Radiocarbon dating of fossil wood remains encapsulated in rock glaciers can also provide an age estimate (Scapozza *et al.* 2010).

Most measurements indicate formation within the Holocene (last 10 000 years), although the formation of some recent rock glaciers are believed to be associated with events during the Little Ice Age. On the Murtél-Corvatsch rock glacier in the Swiss Alps, subfossil stems of different bryophyte species were found at depths 6 m below the surface and about 3 m below the permafrost table, with ages ranging 800 BC to 0 AD. The formation is believed to begin with the onset of the Holocene when glaciers were absent (Haeberli et al. 1999). Diameter of *R. geographicum* thalli revealed ages between 5000 and 6000 years which indicate the age of deposition of the boulders of Murtél rock glacier (Burga *et al.* 2004). Wohlfarth *et al.* (1994) suggested rock glaciers developed around 10 700 BP in southwestern Switzerland. (Krainer *et al.* 2015) dated plant macrofossil in a core of Lazaun I rock glacier and found ages of 8960 cal yr BP at the bottom of the rock glacier in almost 24 m depth. Lewis and Hanvey (1993) believed that rock glaciers in South Africa were flowing some 21 000 years BP.

In the Central Italian Alps, Calderoni *et al.* (1998) found peaty soils buried by rock glaciers that dated at 1120 years BP for active rock glaciers. Dates near the front of inactive rock glaciers were about 2700 years BP. C-14 dating of a buried paleosoil on the Foscagno rock glacier in the Italian central Alps indicates an age younger than 2200±60 years BP (Guglielmin *et al.* 2004). In the Schiantala Valley of the Maritime Alps, Italy, rock glaciers formed after a phase of glacier expansion occurring about 2550±50 years BP. Calibrated C-14 ages of the tree remains found in a rock glacier front in the Russian Altai Mountains were 293±21 years BP and 548±21 years BP; given the time of emergence, the rock glacier developed between 3800 and 2600 and 550 years BP (Fukui *et al.* 2007). In the British Isles, rock glaciers likely developed during the Younger Dryas (Harrison *et al.* 2008). Age of plant remains from the basal sediments of a pond suggest that rock glaciers in northern Maine formed during the Younger Dryas (11 320 years BP). Ballantyne *et al.* (2009) identify relict rock glaciers in the Cairngorm Mountains of Scotland and date them between 12 100±600 years to 15 400±800 years. However, Jarman et al (2013) re-investigated the identified landforms based on field surveys and aerial photo interpretation and question whether these lands forms are relict rock glaciers. Radiocarbon dates from the Lazaun I rock glacier in the southern Otztal Alps indicated an ice age at the base of nearly 10 300 years old (Krainer *et al.* 2015).

At the Faeroe Island in the North Atlantic Ocean, a Younger Dryas age was suggested (Humlum 1998). Rock glaciers in Svalbard began forming about 3500 years BP (Andre 1994). Another estimate in Svalbard suggests that two tongue-shaped rock glaciers developed with the onset of the Holocene (Isaksen *et al.* 2000).

Organic materials removed from debris layers within Galena Creek rock glacier reveal a mid-Holocene or earlier age (Konrad *et al.* 1999). Lobate rock glaciers advanced in the Front Range 3340 years BP (Benedict 2005). *Rhizocarpon geographicum* on 48 lobes of 18 rock glaciers in the Elk Mountains and the Sawatch Range of central Colorado indicate ages of 3080 years BP (Refsnider and Brugger 2007). Hetu and Gray (2000) indicated a Younger Dryas to Early Holocene formation of lobate rock glaciers in the Chic-Choc Mountains in Quebec. Schmidt hammer measurements and weathering-rind thickness indicated that rock glaciers began forming around the Late Glacial or onset of the Holocene in the Japanese Alps (Aoyama 2005). Tropical Andes rock glaciers originated in the early Holocene (Francou *et al.* 1999; Fabre *et al.* 2001).

Rock glaciers in other parts of the world suggest that formation is associated with the Little Ice Age. Lichenometry suggests that rock glaciers in northern Iceland formed immediately after the Little Ice Age (Hamilton and Whalley 1995; Kirkbride and Dugmore 2001). Diameter of the largest lichens on the Mendel rock glacier in the Sierra Nevada, California indicate formation before the Little Ice Age (≈700 years BP) (Konrad and Clark 1998). Humlum (1996) suggested that rock glaciers in Greenland formed during the Little Ice Age, about 550 years ago, when air temperatures were some 2 to 4 °C cooler than present (Humlum 1999). The Thabor-Cheyal Blanc rock glacier in the northern French Alps was not visible on ancient maps or pictures, so it must have formed during the last 200 years, during the end of Little Ice Age (Monnier 2007).

1.12 Geophysical Methods Applied to Rock Glaciers

Permafrost surface and subsurface features are highly variable; geophysical techniques have shown promise when detecting this heterogeneity (Kneisel *et al.* 2008). Since geophysical properties, such as electrical resistivity (and conductivity), permittivity, and seismic wave velocity, vary greatly from a phase change of water to ice, geophysical techniques have been used to investigate the subsurface properties of rock glaciers, map active layers, and monitor change (degradation or aggradation) of permafrost over time.

Direct current electromagnetic techniques, such as capacitive coupled electrical resistivity tomography (CCERT), measure permafrost distribution based on electrical resistivity or electrical conductivity. Conductive water is easily distinguished from nonconductive ice, although soil texture, porosity, water content, and salinity can alter measurements (Table 2). This is the most common method for determining properties of frozen ground. In a CCERT survey, resistivity data are obtained by pulling transmitters and receivers across the ground along a profile, which will result in a 2D dataset (Kneisel 2003; Hauck and Vonder Muhll 2003). A constant current is supplied by electrodes and the resulting voltage is measured at potential electrodes. From these values, resistivity can be calculated. Advantages of this method are its fast data acquisition, rapid deployment, and the fact that galvanic contact with the ground is not required, making it easier to obtain data on frozen ground.

Other electromagnetic techniques such as ground penetrating radar (GPR) are sensitive to change in dielectric permittivity, which shows a marked difference for frozen and unfrozen materials. Permittivity is affected by attenuation or the loss of signal intensity and is determined by the speed by which GPR waves move through different structures (Maurer and Hauck 2007). Structural change affects permittivity at contact zones where energy is reflected back to the surface or the wave propagates further into the ground. GPR uses two antennas, one to transmit short radio waves between 10 and 1000 MHz, and the other receive reflected energy. When antennas are moved across a survey line, a high-resolution image of subsurface ground properties can be obtained.

Additionally, gravimetry can be used to detect the presence and the thickness of ground ice. Data must be corrected for latitude, elevation, Earth tides, and topography of surrounding terrain. A model is then created to determine shape and density of body producing the gravity anomaly. The method has many limitations; intense modeling is required to correct the data, and resolution can be limited from the density variation of materials. Additional data in the form of gamma-gamma logs, boreholes, or refraction data are generally necessary to support data modeling. Problems may also arise when ice content increases continuously along a profile (Vonder Mühll and Klingelé 1994).

Radio echo soundings can also be used to get a rough estimate of the permafrost thickness and ice form; however, the depth of penetration is limited and unfrozen water causes absorption and scattering effects (King *et al.* 1987). Seismic refraction can be used to interpret the internal structure, but the method is limited because the seismic velocity contrasts between layers cannot always be detected because of layer thickness or rock type (Wagner 1996).

A combination of two or more geophysical techniques is generally beneficial because misinterpretation can result from using a single electromagnetic property (Table 2) (Schrott and Sass 2008). There are many limitations of these techniques when studying rock glaciers. The coarse surface of rock glaciers may require additional salt water-soaked sponges to achieve ground contact. CCERT allows mapping of spatial changes of permafrost bodies, but it does not represent stratigraphy well. On the other hand, GPR provides important stratigraphic information, but does not reveal changes in permafrost at reflectors. As a result, integration of these two data sources can provide valuable crosscheck information that validates interpretation.

Frozen ground and active layer thickness have been measured in both alpine and Arctic regions using electrical resistivity. Ice masses on rock glaciers and talus slopes have been assessed using electrical resistivity measurements (Isaksen *et al.* 2000; Hauck and Vonder Muhll 2003; Kneisel *et al.* 2008). P-wave velocities can be used to differentiate between active and inactive rock glaciers, whereas DC resistivities reflect structural differences such as ice-cemented or ice-rich rather than frozen and unfrozen differences (Ikeda 2006). Cui and Cheng (1988) detected two rock glacier layers using DC resistivity soundings near the Urumqi River, Tianshan Mountains. They observed a 1.5 m thick active layer associated with 2.5 kΩ soundings. The active layer is underlain by frozen sands and gravels with ice as indicated by soundings ranging from 32 to 37 kΩ. Francou and Reynaud (1992) determined an active layer of 4 to 5 m thickness underlain by a 30 to 40 m thick (800 kΩ) permafrost layer in Laurichard, French Alps using the geoelectrical sounding method. Geoelectrical tomographies revealed massive ice lenses at the apex and front of rock glaciers in the Central and Western Alps (Ribolini *et al.* 2010). In Svalbard, DC resistivity values indicate an ice-rich layer at 20 to 35 m lying on ice-saturated sediments (Isaksen *et al.* 2000). Electrical resistivity on rock glaciers in the Grizzly Creek region, southwest Yukon indicate no massive ice, even on rock glaciers that were once thought to have ice-cores (Evin *et al.* 1997). Geoelectrical soundings have been used to detect super-saturated sediment with ice that is tens of meters thick on the Portette rock glacier in Italy (Ribolini and Fabre 2007). Changes in the active layer revealed that the rock glacier is strongly influenced by latent heat, snowmelt infiltration, and air circulation. In the Andes of Argentina, geoelectric geophysical methods detected a wet layer underneath an 18.5 m frozen permafrost structure at the base of the El Paso rock glacier (Croce and Milana 2002). On the Murtèl rock glacier, DC resistivity tomography revealed a steeply dipping boundary near the toe, and high resistivities indicate high ice content (Hauck *et al.* 2003).

Degenhardt and Giardino (2003) used GPR to study the internal structure of a rock glacier in the San Juan Mountains, CO. Based on the GPR results, they found that the interior of the landform is composed of a layered permafrost matrix of ice, sediment, and ice lenses that comprise thicker depositional units formed through high-magnitude debris falls. Folds are observed in the uppermost layers, resembling the surface expression of ridges and furrows, which suggests that compressive stresses originating in the accumulation zone are transmitted downslope through the rock glacier. On two tongue-shaped rock glaciers in Svalbard, GPR revealed that longitudinal layers slant downward in relation to the rock glacier surface until an upward slant is reached near the toe. The layers are interpreted as snowdrifts that have been covered with debris from slides (Isaksen *et al.* 2000). GPR has also indicated layering structures on other rock glaciers that may correspond to mass movements of higher magnitude that covered snow patches or possibly the active layer above supersaturated permafrost (Berthling *et al.*

2000). On a polar rock glacier on James Ross Island, Antarctica, longitudinal GPR profiles show upward dipping surfaces; transverse GPR profiles indicate a syncline structure inclined toward the central part of the rock glacier (Fukui *et al.* 2008). Radar measurements have even been transmitted through boreholes to identify cross-sections of ice-rich zones, underlying ice-free zones, and unconsolidated glacial sediments with air-filled voids (Musil *et al.* 2006). Monnier and Kinnard (2013) used GPR and boreholes to interpret the internal structure and composition of rock glaciers in Chile, which revealed degrading conditions.

Other geophysical techniques have also been used to investigate rock glaciers. Transient electromagnetic (TEM) identified a shear plane some 18–28 m deep on the Fireweed rock glacier, Alaska (Bucki *et al.* 2004). In Mullins Valley, Antarctica, seismic surveys indicated a thick glacier core that ranges from 90 to 95 m near the valley head (Shean *et al.* 2007). On the Murtél-Corvatsch rock glacier, Vonder Mühll and Klingelé (1994) found four layers according to drillings and gravimetric methods: an active layer (3 m); massive ice (3 to 20 m thick); an ice–silt–sand layer (20 to 30 m); and a layer with ice-saturated blocks (30 to 50 m) (Fig. 24D)

Ideally, a combination of geophysical methods and other field methods should be used to interpret the internal structure of rock glaciers (Otto and Sass 2006). For example, Vonder Mühll *et al.* (2000) used a combination of a 58 m borehole and geophysical methods to determine the internal structure of the Murtél-Corvatsch rock glacier, which allowed mapping of the permafrost table and lateral extent of a deep shear horizon. Later (Maurer and Hauck 2007) applied geoelectric, seismic and georadar surveys at the same rock glacier which facilitated the interpretation of the internal structure (Fig. 24).

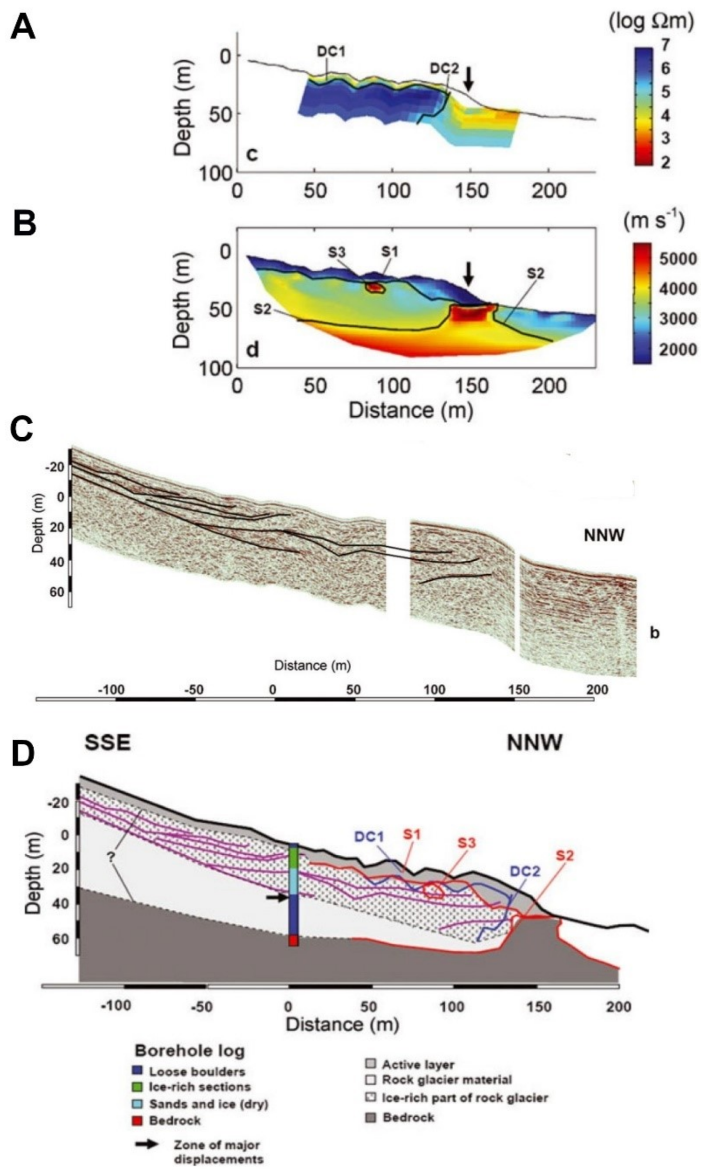


Figure 24 Results from the (A) geoelectric, (B) seismic and (C) georadar surveys conducted on the Murtèl rock glacier. Interpretations superimposed on the tomograms of (a) and (b), respectively. (D) Integrated interpretation of boundaries observed. The borehole log is also displayed. DC1 and S1 denote the active layer, DC2 is the boundary between frozen and

unfrozen material, S2 is the bedrock interface and S3 is probably a huge boulder, Source: (Maurer and Hauck 2007).

1.13 Characteristics of Rock Glacier Flow and Surface Elevation Changes

Rates of surface motion rock glaciers vary from a few centimeters to several meters per year. According to Haeberli *et al.* (2006), the following variables must be taken into consideration to understand a rock glacier flow/creep: (1) rock weathering, snow avalanches, rockfall, and associated rock sizes; (2) freezing processes and ice formation in a combination of fine and coarse-grained rocks; (3) the thermohydromechanics of creep and failure processes in frozen debris; (4) kinematics of nonisotropic, heterogeneous and layered permafrost; and (5) down valley spatial and temporal variation in motion as evident in ridges and furrows.

The surface displacement can either be measured using field surveying techniques such as repeated Global Navigation Satellite System (GNSS) measurements or triangulation, near-range remote sensing such as terrestrial laser scanning or drone survey or air- and spaceborne remote sensing using feature tracking or DInSAR techniques (Table 3). Manual photogrammetric point-to-point methods with stereocomparators or analog plotters were first used to measure the horizontal and vertical displacement of large surficial rocks (Figures 25 and 26).

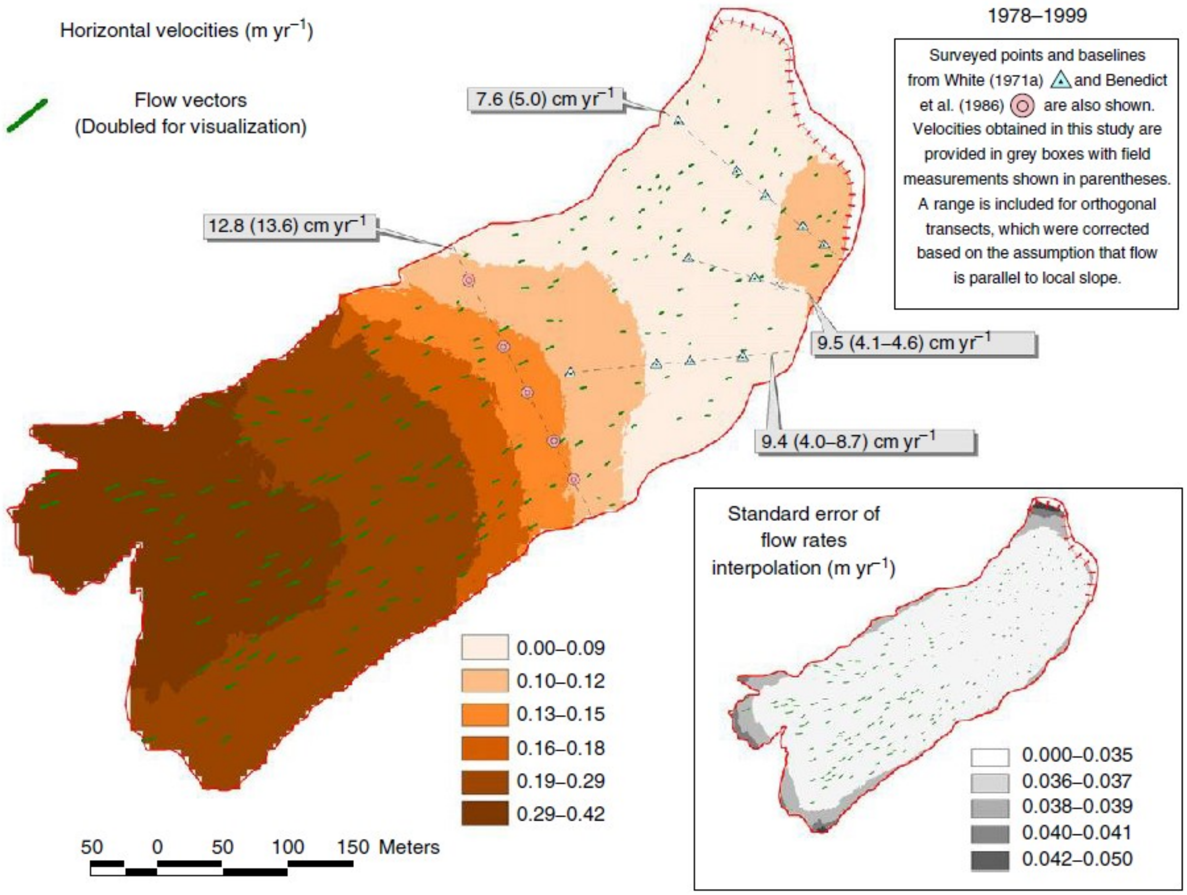


Figure 25 Photogrammetrically derived flow vectors and field measurements of flow on the Arapaho rock glacier in Colorado (Janke 2016).

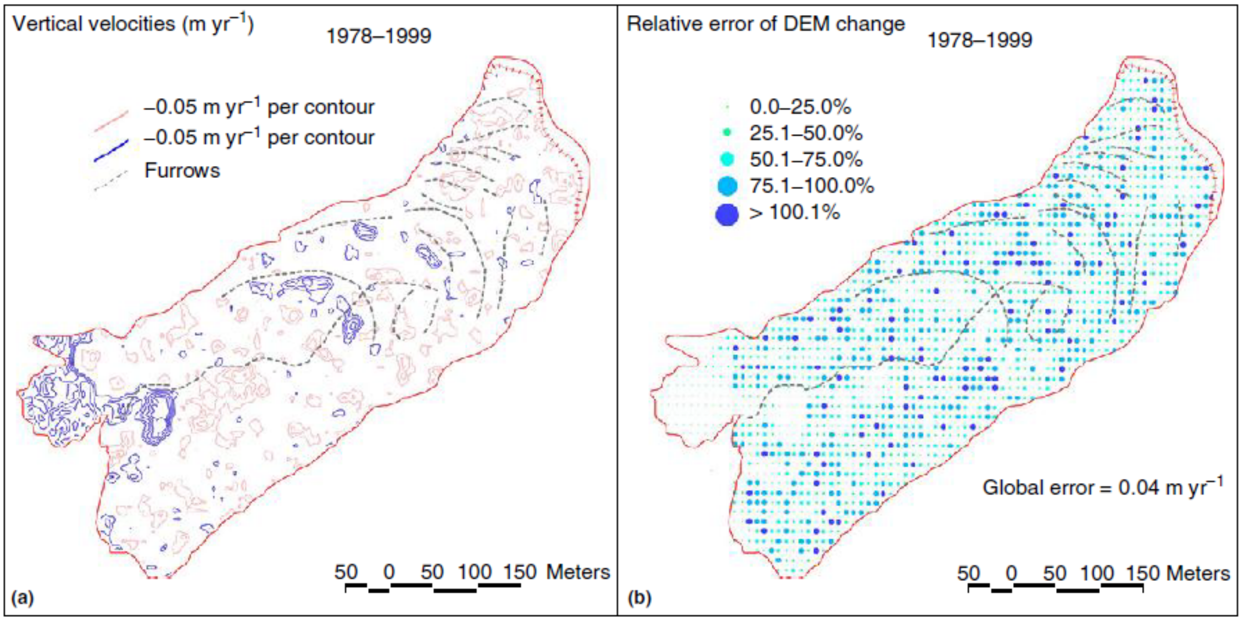


Figure 25 Vertical change on the Arapaho rock glacier (a) and relative error of measurements (b) (Janke 2016).

Advances both in remote sensing data and methods allow the surface displacements to be measured with subpixel accuracy, allowing even displacements in the centimeter range to be detected over the entire rock glacier surface (Figures 27 and 28)

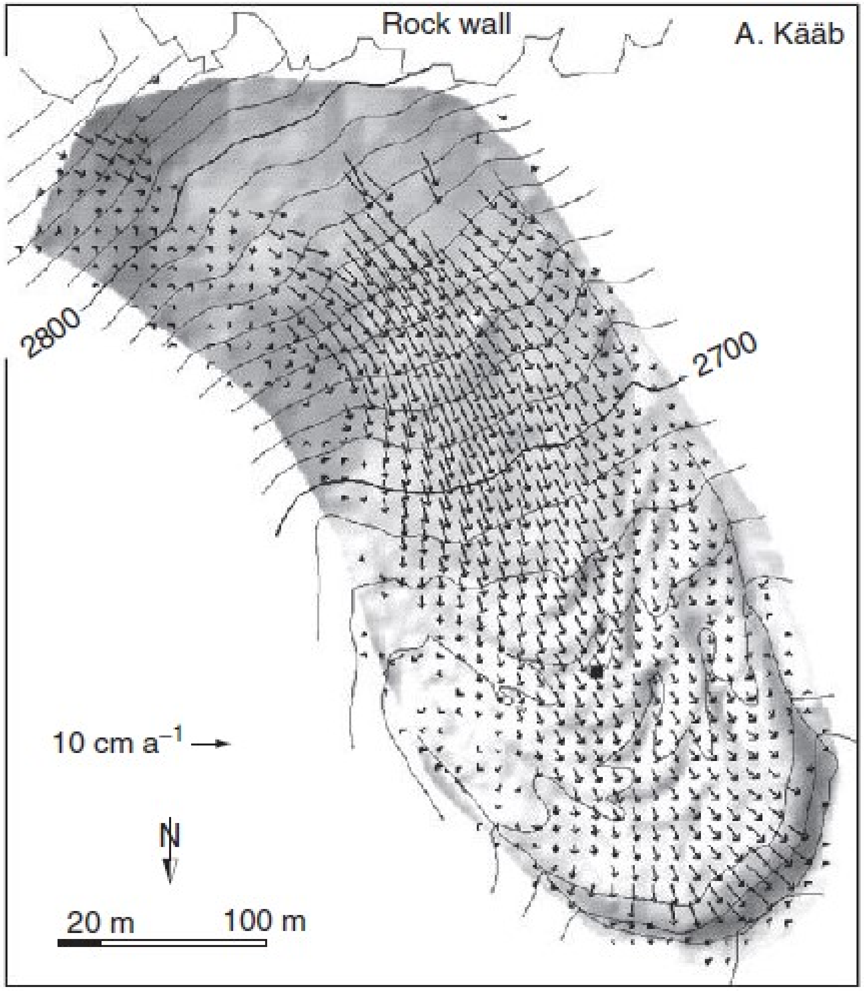


Figure 27 Horizontal displacements 1987 – 1996 on the Murtél rock glacier, Upper Engadine, Switzerland. Source: (Kääb *et al.* 1998)

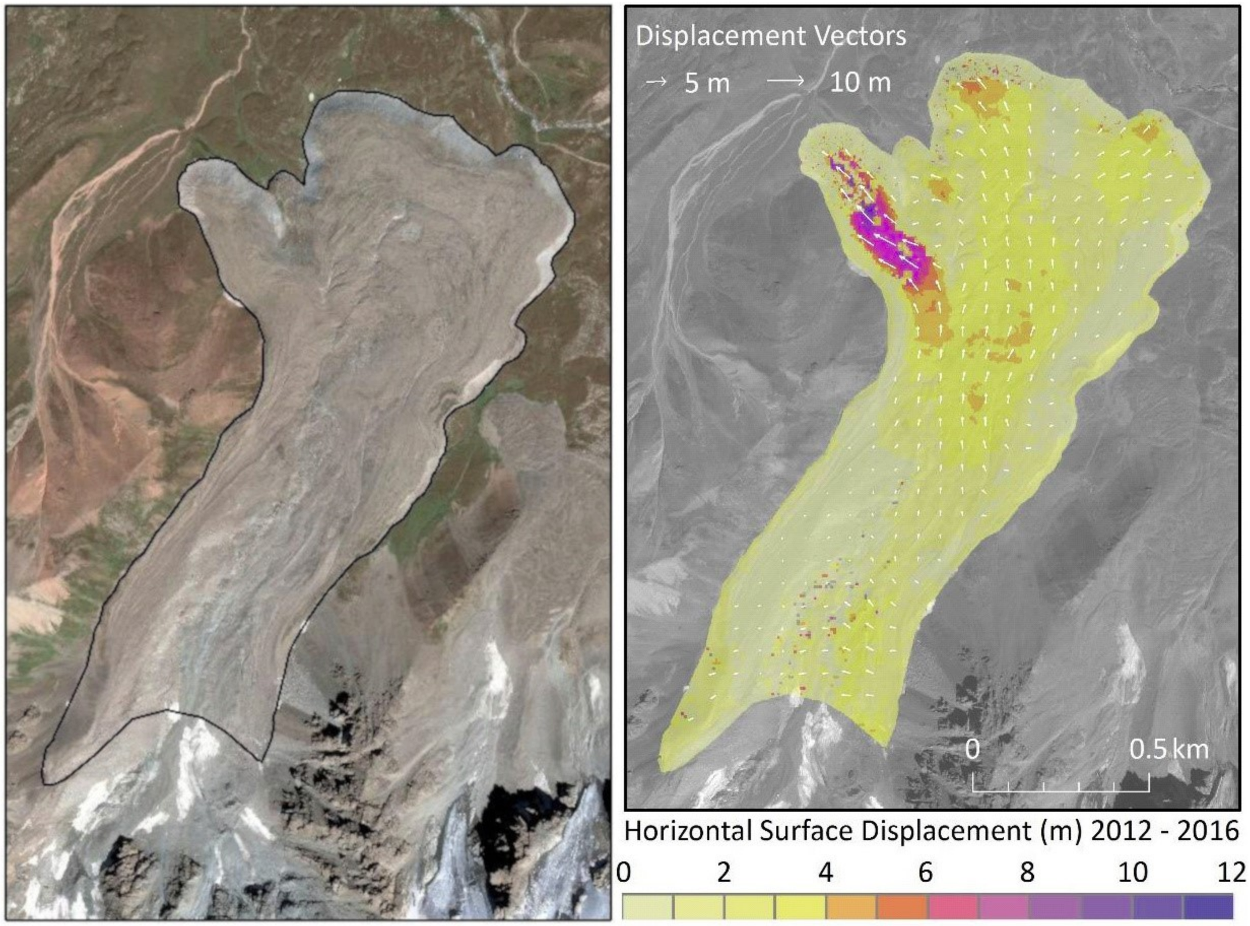


Figure 28 Morreiny rock glacier, Northern Tien Shan; Left: 2012 GeoEye image, Right: Surface displacements measured using the feature tracking method and high-resolution 2012 GeoEye and 2016 Pléiades data. Figure generated by A. Strel and T. Bolch. Pleiades data © 2016 CNES and Airbus Defence and Space.

Remote sensing and GNSS techniques show much promise to detect subtle seasonal and long-term rates of flow at unprecedented spatial precision and accuracy (Strozzi *et al.* 2004). In the Beacon Valley, Antarctica, a precision of mm yr^{-1} can be obtained using synthetic aperture radar (SAR) data; peak velocities were about 4 cm yr^{-1} on rock glaciers (Rignot *et al.* 2002). Differential SAR interferometry (D-InSAR) was used to detect an average displacement of 0.6 cm per 35 days on some rock glaciers. The head of the rock glacier is deforming at 1.8 cm per 35 days, and the toe is moving at about 1.0 cm per 35 days (Kenyi and Kaufmann 2003). (Liu *et al.*

2013) found surface displacements between 14 and 87 cm/a in the Sierra Nevada, USA, based on late summer ALOS Palsar data.

Terrestrial laserscanning (LiDAR) has been used to monitor the front of Hinteres Langtalkar rock glacier from 2000 to 2008 in the Austrian Alps. A positive surface elevation change over the entire analysis period was evident, but decreasing surface lifting became apparent over the last 3 years (Avian et al. 2009). On the Besiberris rock glacier in Spain, a 10-year (1993–2003) geodetic survey revealed that velocities increase from the head to the toe; velocities are greatest in the interior or axis of the rock glacier, and vertical lowering has also been detected (Chueca and Julian 2005). As cameras and sensors continue to improve, high-resolution stereo cameras (e.g., HRSC-A) or stereo satellite images (e.g., Pléiades, GeoEye, Worldview) that records digital multispectral and panchromatic stereo bands could be used to construct high-resolution DEMs and detect change (Otto *et al.* 2007; Bolch *et al.* 2019). The best accuracy can be achieved with a UAV/Drone survey or terrestrial laser scanning (Abermann *et al.* 2010; Vivero and Lambiel 2019). Time series can be extended back using historical aerial images or high-resolution reconnaissance satellite imagery (Bolch and Strel 2018; Bolch *et al.* 2019). The surface elevation changes of active rock glaciers are typically characterized by surface lowering below the rooting zone, variable changes in the middle part (especially due to the displacement of the ridges and furrows, and a surface elevation gain at the front due to the forward movement of the rock glacier (e.g. (Abermann *et al.* 2010; Müller *et al.* 2016), Figures. 29, 30)

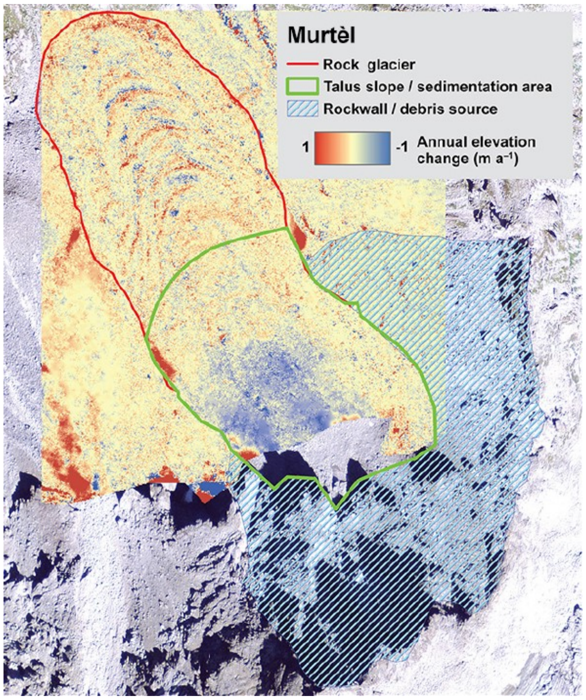


Figure 29 Surface elevation change of Murtel rock glacier (Swiss Alps) between 1997 and 2006 based on stereo aerial images (Müller *et al.* 2016).

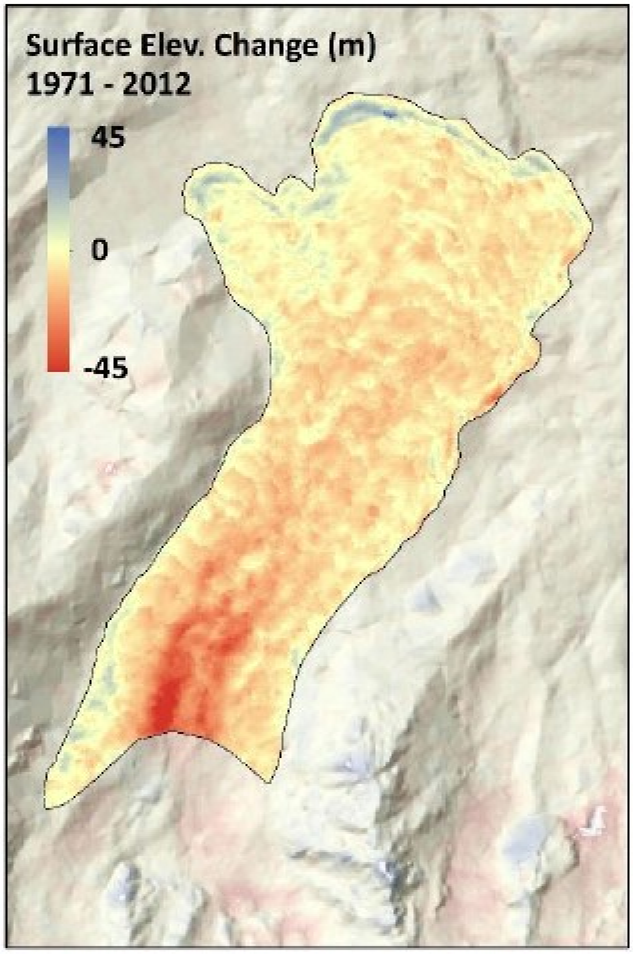


Figure 30 Surface elevation change of Morainny rock glacier in Ile Alatau, Tien Shan, Kazakhstan between 1971 and 2012 based on stereo Corona and GeoEye images. Source: A. Strel, T. Bolch.

MAAT has the strongest correlation with mean annual velocity, but it is also important to examine the time scale of the response to climate change (Janke and Frauenfeldeler 2008; Scapozza *et al.* 2014). Warm ice can deform more quickly than cold ice, or meltwater can act as a lubricant and reduce friction between internal shear planes (Bucki and Echelmeyer 2004; Jansen and Hergarten 2006; Krainer and Mostler,2006; Ikeda *et al.* 2008). Recent investigations showed that the rates of flow are mainly impacted by meltwater (Bodin *et al.* 2009; Cicoira *et al.* 2019). GNSS real-time kinematics (RTK) provides precise data to detect horizontal and vertical motion to simulate 3D rotational movements of boulders over short time scales (Lambiel and

Delaloye 2004). Continuous measurements at selected rock glaciers in the Swiss Alps showed similar intra-annual variability with rapid acceleration in spring and a smooth decreasing velocity in late autumn. Short-term velocity peaks can be related to strong water input from snowmelt or precipitation (Wirz *et al.* 2016; Buchli *et al.* 2018; Cicoira *et al.* 2019).

The Büz North rock glacier on the lower limit of the permafrost belt in the Swiss Alps deforms rapidly during snowmelt periods but decelerates below a dry snow cover during the winter (Ikeda *et al.* 2008). At the Ölgrube rock glacier, flow velocities vary seasonally with higher rates during the melt season as evident by several springs near the front slope of the rock glacier (Krainer and Mostler 2006).

Deformation in boreholes revealed distinct shear zones where horizontal and vertical differential movements are concentrated (Arenson et al. 2002) **(**Figure 31**)**. Musil *et al.* (2002) found that shallow seismic profiles and four boreholes drilled up to 70 m indicate two distinct velocity regimes superimposed on a general increase in velocity with depth. Conveyor belt advance mechanisms have also been observed on the Gruben and Suvretta rock glaciers. The toe of Suvretta rock glacier has a linear velocity profile over the entire thickness, whereas the

Gruben rock glacier has horizontal mass transport concentrated in a 5–10 m thick surface layer (Kääb and Reichmuth 2005).

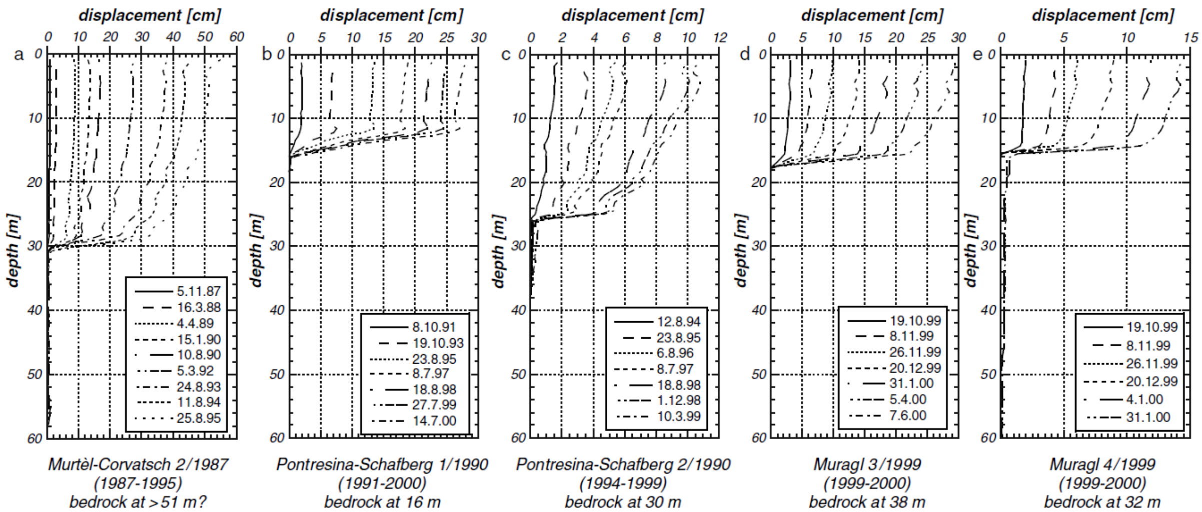


Figure 31 Displacement measurements at selected boreholes in rock glaciers in the Swiss Alps. All measurements show a distinct shear horizon (Arenson *et al.* 2002).

Dendrochronology can provide information on long-term changes of the rock glacier front (Shroder and Giardino 1987). On the Hilda rock glacier, overridden trees indicate an advancement rate of 1.6 cm yr^{-1} since the late 1790s (Bachrach *et al.* 2004; Carter *et al.* 1999). Giardino *et al.* (1984) studied the motion of a rock glacier for more than 400-years based on the tree rings. Shroder and Giardino (1987) studied rock glaciers on Table Cliff Plateau, Utah and Mt. Mestas, CO, and dated the movement of these two rock glaciers for two centuries.

1.14 Hydrology

Water movement through rock glaciers is controlled by local weather, the seasons, the thermal properties of the debris layer, and the physical mechanisms that control meltwater (Geiger *et al.* 2014; Pourrier *et al.* 2014; Harrington *et al.* 2018). Although the general discharge pattern is the same, the average yearly mean specific discharge from active rock glaciers is significantly lower than glaciers (Krainer and Mostler 2002). Giardino *et al.* (1992) described the interior of rock glacier as similar to an aquifer. The water input to the rock glacier can come directly from precipitation, melt from snow patches on the rock glacier surface, runoff from adjacent slopes, or from groundwater. Rock glaciers provide a continuous meltwater supply during the summer months, analogous to an alpine aquifer (Burger *et al.* 1999). Rock glacier melt is of specific importance in arid regions with no or low glacier coverage (Rangecroft *et al.* 2013). Water can flow near the surface, above the active layer, or can flow in the deep subsurface beneath permafrost (Figure 32). The permafrost layer acts as an impermeable boundary between these two layers. During late summer/early fall, falling temperatures create freezing fronts that move upward from the perennial permafrost and downward from the rock glacier surface, squeezing together the remaining water in the central portion of the active layer (Giardino *et al.* 1992; Burger *et al.* 1999).

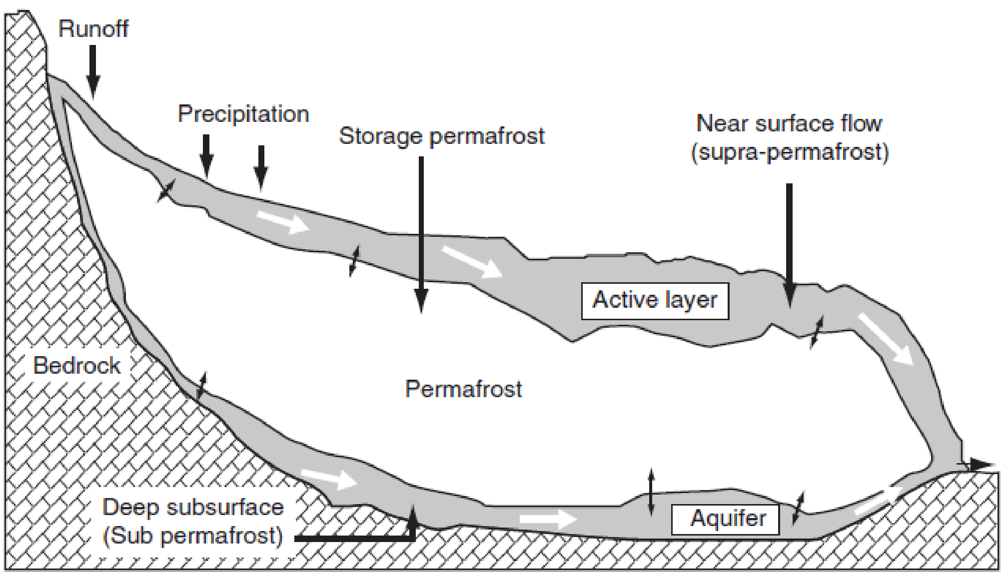


Figure 32 A hypothetical model of water flow through a rock glacier (Giardino, *et al.* 1992).

In addition to the aforementioned inputs, discharged water can also come from melting ice-rich permafrost or an internal ice structure. Baltensperger *et al.* (1990) suggested that a pH>6 indicates groundwater rather than meteoric water, according to data from the Múrtel rock glacier. Water temperature is typically below 1 °C during the melt season. Electrical conductivity is low during high discharge and high during cold weather (Krainer and Mostler 2002).

In the Andes or other areas where precipitation is scarce, rock glaciers appear to be a more significant source of water storage compared to temperate glaciers (Fabre *et al.* 2001; Azocar and Brenning 2010). In the Bolivian and Argentinean Andes, rock glaciers act as a water storage unit in a low precipitation areas (Francou *et al.* 1999; Perucca and Angillieri 2008; Rangecroft *et al.* 2015). Rock glaciers are estimated to have 0.3 km^3 of stored water per 1000 km^2 in the mountainous area in the Andes of Santiago, Chile – an order of magnitude higher than the Alps (Brenning 2005). In Northern Tien Shan in Central Asia, rock glaciers store a significant amount of ice which are an important water source during summer (Bolch and Marchenko 2009). Also, in many other mountain regions of the globe, rock glaciers contain large ice reserves (Jones *et al.* 2019). Rock glaciers could represent a significant water source in a future warmer climate, such as in the Sierra Nevada or other mountain ranges (Millar and Westfall 2008; Harrison *et al.* 2021).

Discharge from rock glaciers can alter the geochemistry of alpine streams, rivers, and lakes by altering temperature, pH, major ions, trace elements, diatoms, dissolved organic matter, bacterial richness, etc. (Thies *et al* 2013; Ilyashuk *et al.* 2014; Fegel *et al.* 2016; Colombo *et al.* 2018). Williams *et al.* (2007) found that 7 Rocky Mountain rock glaciers emit higher nitrate concentrations (69 μmoles l^{-1}) compared to snow (7 μmoles l^{-1}) and rain (25 μmoles l^{-1}). In the Green Lakes Valley, Colorado, nitrate concentrations from rock glaciers increased seasonally, reaching a maximum in October – the highest concentration in the valley. Seasonal variation has also been observed for other geochemical solutes (Williams *et al.* 2006). At Green Lakes 5 rock glacier, Mg^{2+} concentrations increased to more than 4000 μeql^{-1} in September; Ca^{2+} was greater than 4000 μeql^{-1}; and SO_4^{2-}reached 7000 μeql^{-1}. The concentrations were significantly higher than those observed from June to August. At the Rasass See in the European Alps, electrical conductivity has risen 18-fold and nickel concentrations have exceeded drinking water limits by an order of magnitude from increased melt from rock glaciers (Thies *et al.* 2007). In the Loch Vale Watershed in Colorado, it is believed that thicker active layers of rock glaciers are increasing stream nitrate concentrations, which are 50% greater since 2000 compared to the 1991–99 period (Baron *et al.* 2009). Nitrate, calcium, and sulfate are also higher in the rock glacier meltwater from 2000 to 2006. In the Andes of Argentina, tropical glaciers deliver a HCO_3^- and Ca^{2+} solution, whereas rock glaciers emit an SO_4^{2-}, HCO_3^-, and Ca^{2+} solution (Lecomte *et al.* 2008). Gypsum dissolution through sulfide oxidation is the most important geochemical mechanism delivering the oxides. Silicate weathering is more important when meltwater has a longer residence time, and calcite and gypsum dissolution is more noticeable in recently melted water (Lecomte *et al.* 2008).

1.15 Topoclimatic Conditions

Remote sensing has improved our understanding of rock glacier distribution, their characteristics, topoclimates that facilitate rock glacier formation. Topoclimates have been shown to be important for rock glacier formation, more so, compared to regional climates, even where rates of insolation are high.(Vitek and Giardino 1987; Brenning and Trombotto 2006). In the subtropical Andes, local and catchment slope and potential incoming solar radiation characterize the locations of rock glaciers (Bolch and Schröder 2001; Brenning and Azócar 2010). High elevations, southern or eastern-facing aspects, areas with low solar radiation, and slopes less than 20° favor rock glacier occurrence in the Andes of Argentina (Angillieri 2010).

In Greenland, dry, continental areas are ideal for rock glacier occurrence. In the Southern Alps, New Zealand, ice-rich forms of glacial origin exist in the humid north, and debris-rich forms of periglacial origin exist in the more arid south (Brazier *et al.* 1998). In Greenland, active rock glaciers tend to plot close to ELA for modern glaciers. Rock glaciers, however, require high retreat and weathering rates in the source area (Humlum 2000). Ideal sites also favor the high production of talus in the Southern Alps (Kirkbride and Brazier, 1995). Surface snow near the head of the rock glacier freezes and acts as a slideway to deliver rocks detached from the headwall (Pancza 1998). In Trollaskagi, northern Iceland, rock glaciers form at the foot of steep northern slopes where snow accumulates from avalanches and drifting from southerly winds (Farbrot *et al.* 2007). Steady-state rockfall accumulation, however, may not be sufficient to initiate rock glacier formation; a rock-slope failure is needed to supply enough debris (Sandeman and Ballantyne 1996). In Nanga Parbat Himalaya, Pakistan, inefficient sediment transfer processes, mainly from slow ice movement, allow rock glaciers to form (Shroder *et al.* 2000). The size of weathered rocks supplied can also influence the ability to support a matrix used to induce creep; pebbly rock glaciers that can support matrix debris have a greater length compared to bouldery rock glaciers (Matsuoka *et al.* 2005).

Coarse surface layers act as a thermal filter that protects the frozen rock glacier core when snow cover is absent or thin (Humlum 1997). Ingelrest *et al.* (2010) deployed a monitoring system on a rock glacier in Switzerland and discovered microclimatic characteristics that lead to cold air release. Advective air movement within blocky debris has much less influence on the thermal regime than the vertical displacement of air masses (Hanson and Hoelzle 2004).

GIS has been used to examine the distribution and characterize topoclimatic conditions of rock glaciers. In the Adam-Presanella Massif of the Italian Alps, active rock glaciers lie below the −1 °C MAAT isotherm, suggesting that rock glaciers are not in equilibrium with the current climate, or rock glaciers may have extended to lower elevations (Baroni *et al.* 2004). Bolch and Gorbunov (2014) modelled permafrost distribution based on field evidence, temperature, and solar radiation in Northern Tien Shan and showed that the lower limit of many large rock glaciers is located lower than the zone where permafrost is likely. Headwall height, size of the contributing area, and slope explained the occurrence of rock glaciers. In the Lemhi Range in central Idaho, rock glacier locations are determined by topographic shading, lithology, relief, aspect, and elevation (Johnson *et al.* 2007). In the San Juan Mountains of Colorado, nonlinear factors of slopes, contributing area slope, local curvature, and contributing area size exhibit topographic controls on rock glacier occurrence (Brenning *et al.* 2007). In the Front Range of Colorado, tongue-shaped rock glaciers occur at higher elevations, northern aspects, and have gentler slopes compared to lobate forms. Active tongue-shaped rock glaciers had similar topographic variables compared to temperate glaciers, but lobate forms showed a significant difference (Janke 2007). In the Sierra Nevada, CA, rock glaciers exist at mainly on NNW to NNE aspects at a range of elevations from 2225 to 3932 m (Millar and Westfall 2008). In the Swiss Alps, Frauenfelder *et al.* (2008) simulated spatio-temporal rock glacier processes at a regional scale and found that the model is highly dependent on input parameters, illustrating the importance of local variables on the distribution of rock glaciers.

Because rock glaciers contain ice, they have been used to represent regional permafrost distribution. In the eastern Himalayas, rock glacier distribution indicates discontinuous permafrost at 4800 m on the north facing slope and at 5300 m on south to east facing slopes.

These estimates are considerably higher than the western Himalayas (Ishikawa *et al.* 2001). (Schmid *et al.* 2015) assessed permafrost distribution modelling using rock glacier occurrence in the whole Hindu-Kush-Himalaya region. Imhof (1996) used altitude, slope, aspect, and ground cover types associated with permafrost to develop a PERM model. Rock glacier locations across the Bernese Alps showed a good correlation with the PERM results. A regional map of permafrost (Iceland) based on meteorological data was validated based on rock glacier inventories (Etzelmüller *et al.* 2007). Rock glaciers show a good correlation with regional permafrost models in Switzerland, classifying about 70% of active rock glaciers correctly (Nyenhuis *et al.* 2005). Janke (2005b) used DEM variables (elevation and aspect) from rock glaciers that contain ice and land cover data to estimate permafrost occurrence in the Front Range of Colorado. Azocar *et al.* (2017) utilized rock glacier activity status and temperature data to model distribution in the semi-arid Chilean Andes. Mercer *et al.* (2017) developed a permafrost favorability index for the French Alps based on a rock glacier inventory. A combination of geophysical, remote sensing, and field detection methods are best to verify the presence of ice-rich permafrost in glacier forefields, rock glaciers, or other kinds of ice-debris landforms (Kneisel and Kääb 2007; Bolch *et al.* 2019). Permafrost mapping could be improved by monitoring flow through photogrammetric techniques to verify motion that is driven by ice (Strozzi *et al.* 2004; Kääb and Kneisel 2006; Roer and Nyenhuis 2007).

1.16 Response of Rock Glaciers to Climate Change

The climatic response of rock glaciers is different compared to glaciers. Their typical bouldery layer (1–3 m thick in some instances) acts as an insulator and protects internal ice core or ice debris mixture, and smooths their response to climate (Barsch 1996; Konrad *et al.*,1999). This debris filters short-term climate anomalies; therefore, a strong climatic signal must exist to produce a change in a rock glacier system. Degradation of rock glaciers is generally measured from slumping surface morphology, frontal activity, amount of internal ice, variation in downslope movement, or temperature of the frozen material (Francou *et al.* 1999; Bosson and Lambiel 2016). A variety of geomorphic (internal structure, underlying topography, slope, curvature, etc.) and environmental (snowfall totals, snowmelt timing, temperature change, etc.) variables must be taken into account to properly evaluate the response of a rock glacier to climatic fluctuations. In case the active layer thickness exceeds the thickness of the bouldery layer, the internal ice core starts to melt, and the rock glacier commonly shows an overall surface thinning because of warming. On the Iberian Peninsula, geodetic surveys and GPS measurements on the Veleta rock glacier indicate dominant vertical subsidence compared to horizontal flow (de Sanjose-Blasco *et al.* 2007). On the Veleta rock glacier, Sierra Nevada, subsidence of dead ice has been consistent since 2001 (Ortiz *et al.*, 2008). Some small-scale ridge advection or heaving may also be observed; advancing ice may cause compression or overthrusting near the toe (Kääb and Weber 2004; Chueca and Julian 2005; Janke, 2005 a, d; Kellerer-Pirklbauer *et al.* 2008). Although degradation of rock glaciers is much slower than of glaciers, they show a direct climate response, most evident in their movements (see section 1.13).

However, some rock glaciers have not shown a clear response to climate change. Field measurements for the Galena Creek (Wyoming) during the mid-1960s until 1995 and for Arapaho, Taylor, and Fair (Front Range) rock glaciers during 1961-2002 suggest that the velocities of these rock glaciers have not significantly changed in the measurement periods (Potter *et al.* 1998; Janke 2005d). During the warmer 1980s and 1990s, some rock glaciers in the European Alps have shown increasing subsidence rates by two to three times, decreasing horizontal velocities, and increasing borehole temperatures (Haeberli 1994; Barsch 1996; Kääb *et al.* 1997; Kaufmann and Ladstädter 2002).

However, an acceleration of the rock glacier flow during the last decades in response to increasing temperature has been found by both in-situ and remote sensing derived measurements for many rock glaciers in the Alps and Tien Shan (Kääb *et al.* 2007; Kääb *et al.* 2021; Marcer *et al.* 2021). Argualas rock glacier, Central Pyrenees, was surveyed between 1991 and 2000 using a total station; displacements showed temporal variations related to temperature change (Serrano *et al.* 2006). High surface flow rates (up to 3 m yr^{-1}) were measured near the front of 3 rock glaciers in the Austrian Alps using differential GPS. Analysis of the Reichenkar rock glacier in the Stubai Alps from 1954 to 1990 indicates a mean velocity of 6 cm yr^{-1}, whereas GPS measurements since 2001 indicate an increase of up to 30 cm yr^{-1} (Hausmann *et al.* 2007). The

Laurichard rock glacier in France has shown an increase in surface velocity and decreased in internal resistivity between 1980 and 2000s (Bodin *et al.* 2009).

1.17 Hazards related to Rock glaciers

The most common hazards related to rock glaciers are outburst of lakes that form on or are dammed by rock glaciers and rock avalanches originating from rock glacier fronts (Haeberli *et al.* 2001; Schoeneich *et al.* 2015) (Figures 33, 34). Degradation of permafrost near the front slope can increase the amount of erodible debris and reduce the mechanical stability of the rock glacier (Kneisel *et al.* 2007). Specifically, if the frontal part of the rock glaciers starts to accelerate strongly, crevasses or scarps can open, and the rock glaciers destabilize or even collapse (Roer *et al.* 2008; Bodin *et al.* 2017). In the case of rock glaciers located on steep slopes, this can result in a rockfall or a debris flow endangering the downstream area (Figure 34). Fast-flowing rock glaciers can also be the source of rockfall or trigger debris flows, as reported from rock glaciers in the Mattertal/Swiss Alps (Graf *et al.* 2013; Faillettaz *et al.* 2019).

In addition, the surface movements of rock glaciers themselves can cause some damage if structures like chair lifts or roads are built on a rock glacier. Special construction techniques were used to build a tramway station across a rock glacier located near Lone Mountain, MT (Figure 35). The rock glacier moves about 10 to 20 cm yr^{-1}, which is comparable to the stretch of the cable. To stabilize the station, gravel was placed on top of the rock glacier, covered by 20 cm thick 60-PSI blue board. This was covered by a 61 cm monolithic pad with massive amounts of rebar to build a 'boat' that floated on the surface of the rock glacier in 1995. The structure was weighted down (2600 t) in order to resist the force of the tram, which would cause it to slide across the blue board. The structure settled in a corner by 38 cm in 15 years because of ice ablation. Recently, the subsidence seems to have slowed (David Hamre personal communication).

Figure 33 A rock glacier blocks a valley and forms Lake Sarez near Irkht in Tajikistan.

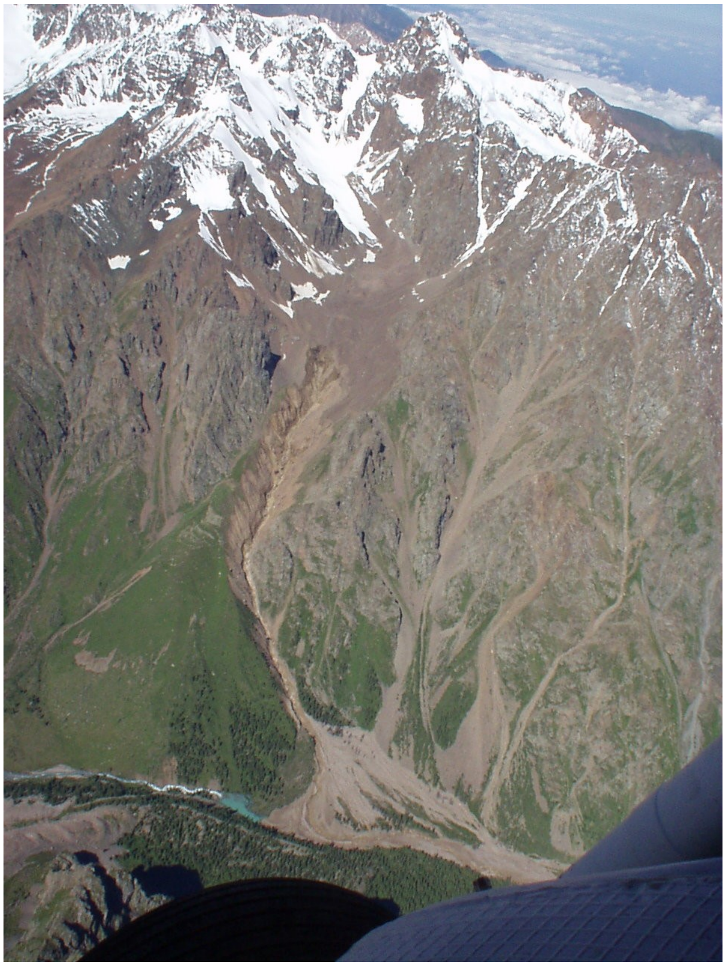


Figure 34 The 2003 debris flow in Left Talgar Valley (Ile Alatau, Tien Shan, Kazakhstan) originated from a degrading rock glacier front. Photo: Institute of Geography, Almaty.

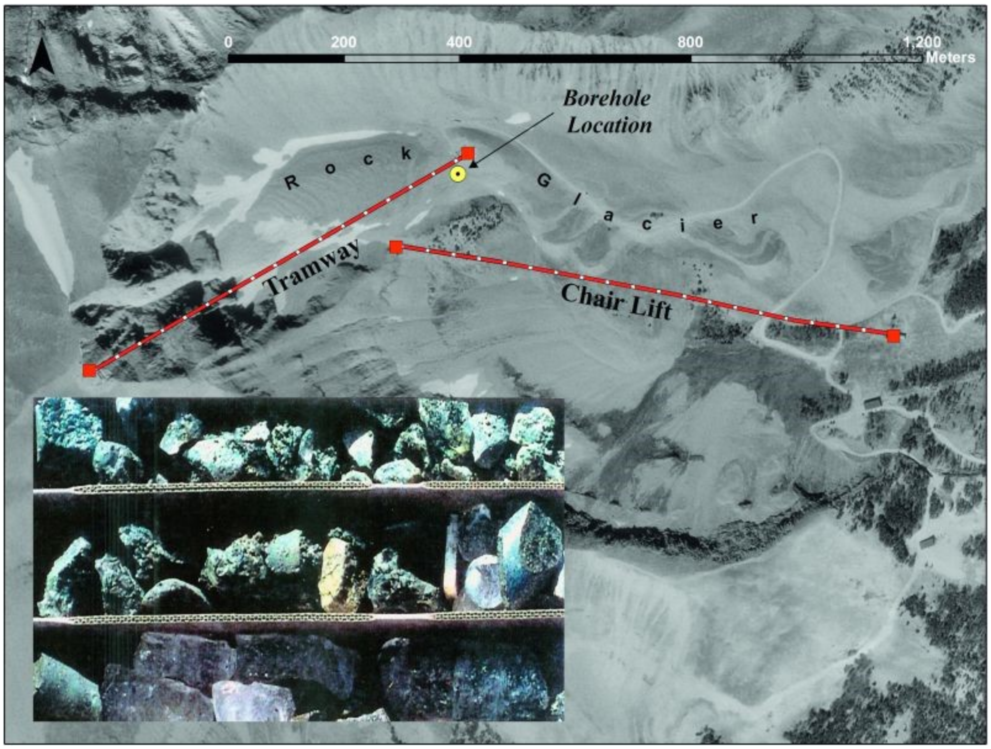


Figure 35 A rock glacier near Lone Mountain, MT, in the Big Sky ski resort is shown with a photo of a core sample ranging from 19.8 to 23.5 m depth. The ice-core shows both an ice-rock aggregate as well as pure ice.

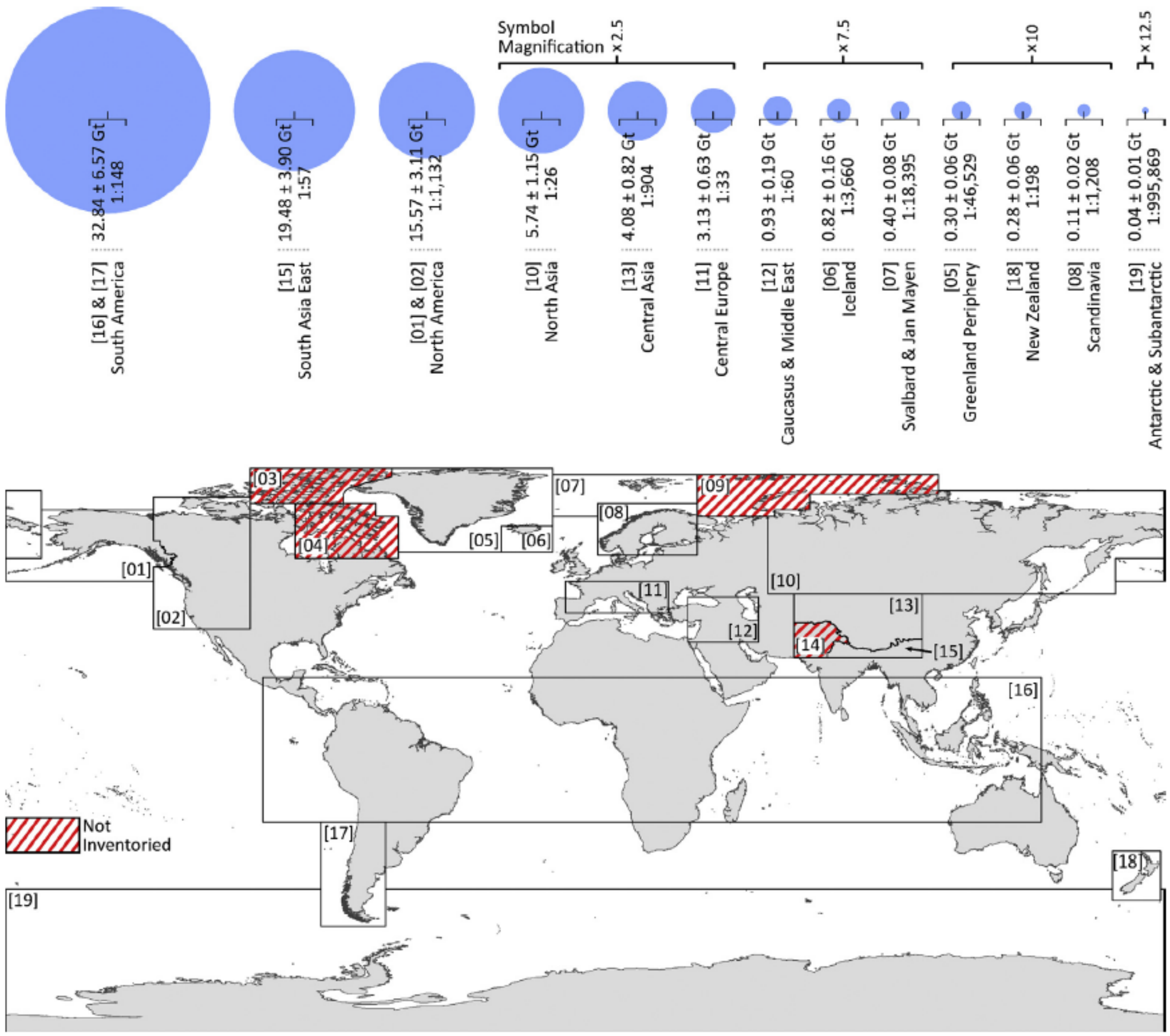


Figure 36 Estimated water volume equivalent stored in rock glaciers and glacier to rock glacier ratios based on existing inventories (Jones *et al.* 2019). Note: In many regions, no complete rock glacier inventory exists; some regions have no information about rock glacier occurrence.

1.18 Future Research

Even though the knowledge about rock glacier occurrence, origin, characteristics and response to climate has much improved, many knowledge gaps still exist. It is important to generate a complete and consistent world inventory of rock glaciers containing basic information such as location, area, length, width, elevation, aspect, slope and activity, and flow rates. Existing information need to be synthesized, and rock glaciers in missing areas should be mapped. Better information about distinguishing debris-covered glaciers from rock glaciers or investigating the glacier-rock glacier continuum under permafrost conditions is needed. The IPA action group of rock glacier kinematics and inventories was established in 2018 and is working on standardized guidelines and rock glacier mapping (IPA Action Group 2020).

Rock glaciers store a significant amount of water in alpine systems (Figure 36). These catchments provide drinking water for nearby communities. As the ice in rock glaciers and other periglacial ice-debris landforms such as ice-covered moraines (I-DLs) are more resilient to the atmospheric warming due to the insulating debris layer, they can somewhat counterbalance the expected declining glacier runoff (Figure 37). The effect of decaying ice that was previously stored in periglacial ice-debris landforms, including rock glaciers, needs to be assessed at the watershed scale and included in glacio-hydrological models. At present, no glacio-hydrological model includes the melt from rock glaciers for larger regions. Currently, local measurements from a few rock glaciers do not adequately represent additional periglacial inputs contained within a watershed. As ion selective probes and affordable laboratory work become more readily available, a better understanding of impacts on water quality will result.

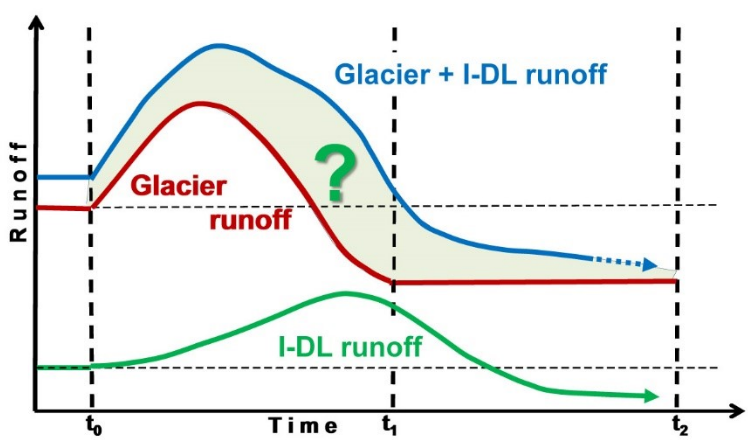


Figure 37 Conceptual figure about the potential future runoff from rock glaciers and other periglacial ice-debris landforms (I-DL). Figure: T. Bolch; the glacier runoff is adapted from (Huss and Hock 2018).

Much information remains to be discovered about the magnitude and timing of runoff from rock glaciers. A better estimate of the total ice content and dynamics of the active layers may help our understanding of water movement and storage. The characteristics and variability of the internal structure is another piece of critical information. Additional cores extracted would provide valuable insights but are only feasible at few sites because of the difficulty of getting coring equipment to remote sites and the associated high costs. Advanced geophysical methods allow more detailed information to be obtained (Hauck *et al.* 2011; Emmert and Kneisel 2017) but due to limited field access and difficulty to work on the boundary surface of the rock glaciers, the feasibility of geophysical investigations are also limited. Helicopter-borne measurements might allow more extensive coverage, although with lower accuracy (Merz *et al.* 2015).

Holistic studies which combine the information from various remote sensing methods and in-situ information obtained from geophysics and other measurements have the potential to provide new detailed insights into the characteristics and climate response of rock glaciers and other periglacial ice-debris landforms (Capt *et al.* 2016; Bolch *et al.* 2019; Buckel *et al.* 2021; Halla *et al.* 2021).

An even more perplexing issue concerns the mechanism of the flow of rock glaciers. Few displacements measurements from boreholes provide important insights and show that in contrast to glaciers that most of the displacements take place at a shear horizon, unlike in glaciers. However, these are point measurements and more information is needed about the deformation of the ice or ice-debris mixture or the potential presence of sliding, especially of rock glaciers that that might have transitioned from debris-covered glaciers. More measurements and especially modelling studies (e.g., of the flow of rock glaciers) that combine a series of different techniques and data sources from different spatial and temporal scales are needed (Müller *et al.* 2016; Wirz *et al.* 2016; Kenner *et al.* 2017; Cicoira *et al.* 2019).

New geospatial techniques to monitor change show promise for future projects. Terrestrial LiDAR, very high-resolution stereo data obtained from very high-resolution satellite images or UAV, unmanned aerial systems, structure from motion through dense image mapping methods, or long-range terrestrial laser scanning could be used to measure detailed rock glacier kinematics (Abermann et al.et al. 2010; Kenner et al. 2014; Gomez-Gutierrez et al. 2014; Piermattei et al. 2016; Dall' Asta et al. 2017; Fey and Wichmann 2017; Vivero and Lambiel 2019; Halla et al. 2021). Deformation of ridges and furrow complexes or advancing or slumping front slopes could be monitored in sensitive areas at an unprecedented spatial scale. The availability of high-resolution SAR sensors such as TerraSAR-X or SAR satellites with high repeat rates (e.g., Sentinel-1) allows detailed insight into the surface displacements and regular monitoring of rock glaciers (Barboux *et al.* 2014; Strozzi *et al.* 2020). Other very high-resolution space-borne hyperthermal sensors could be used to detect changing thermal patterns on the surface of rock glaciers. High-resolution GNSS measurements, coupled with wireless weather stations, could be used to measure real-time displacements or possibly even daily or seasonal movement. It would also be important in understanding the debris transport systems associated with rock glaciers (Müller *et al.* 2014) and to assess and monitor the hazard potential of rock glaciers (Faillettaz *et al.* 2019).

Rock glaciers are abundantly available in continental climates covering large areas and will continue to affect the alpine environment once glaciers have retreated. However, rock glaciers are vastly understudied compared to glaciers. The research community must inspire others to contribute their expertise to thoroughly understand and assess future impacts of an important alpine landform: rock glaciers.

Animations 1–8.

The online version of this chapter contains a Video with additional crystallographic data of compound **1**–**8**. The online version can be found at doi:10.1016/B978-0-12-374739-6.00211-6

See Animation 1–8 which contains additional crystallographic data of compound **1**–**8**.

Animation 1 Terrestrial video of the Arapaho rock glacier, Front Range, CO.

Animation 2 Fly-by of the Arapaho Valley depicting the Arapaho glacier and rock glacier, Front Range, CO.

Animation 3 Fly-by of Huerfano Valley and the California rock glacier located in southern Colorado.

Animation 4 Terrestrial video of the Fair rock glacier, Front Range, CO.

Animation 5 Fly-by of Taylor rock glacier in Rocky Mountain National Park, CO (IKONOS © imagery in the backdrop).

Animation 6 Temporal aerial photographs which depict motion on the Middle St. Vrain rock glacier, Front Range, CO.

Animation 7 Motion on the Arapaho rock glacier, Front Range, CO.

Animation 8 Motion on the Taylor rock glacier, Front Range, CO.

Tables

Table 1 Attributes of some rock glaciers found in the literature

Name/Location	**Position**	**Form**	**Bedrock**	**Status**	**Length (m)**	**Width (m)**	**Elevation (m)**	**Latitude**	**Reference**
Prins Karls Forland, Western Svalbard	Valley wall	Lobate	Quartzite	Active		500		70°50′ N	Berthling *et al.* (1998)
Mullemfjord, Disko Island, Greenland	Cirque	Tongue	Basalt	Active	1700	500	100–200	69°15′N	Humlum (1997)
Central Brooks Range, Alaska, USA	Valley wall	Lobate	Sedimentary	Active and inactive	10–210		900–2000	67°37′–68° N	Calkin *et al.* (1987)
	Cirque	Tongue			<3500		1100–1800	67°37′–68°N	
Kigluaik Mountains, Alaska, USA	Valley floor	Tongue	Metamorphic Granite	Active	350–800		500–930	65°N	Calkin *et al.* (1998)
Alaska Range, USA	Valley wall	Lobate	Metamorphic volcanic	Active and inactive	60–1080	90–3085	865–1850	63°15′–64°N	Wahrhaftig and Cox (1959)
	Cirque	Tongue			150–1540	60–775			
Grizzly Creek, Yukon Territory, Canada	Cirque	Lobate	Metamorphosed sediments; intrusive volcanics	Active and inactive	430–1430	200–1000	1650–2000	61°N	Johnson (1987)

Name/Location	Position	Form	Bedrock	Status	Length (m)	Width (m)	Elevation (m)	Latitude	Reference
Valais Region, Switzerland	Valley floor	Lobate		Active	600		2650–300	46°N	Pancza (1998)
Murtèl/Corvatsch, Grisons, Swiss Alps	Cirque	Tongue	Granodiorite	Active	400	200	2620–2850		Haeberli *et al.* (1988), Vonder Mühll and Klingelé (1994)
Laurichard, French Alps		Tongue	Granite	Active	400	40–200	2500	45°N	Francou and Reynaud (1992)
Galena Creek, Absaroka Mountains, USA	Cirque	Tongue	Volcanic	Active	1600	240–300	2680–3110	44°38′30″N	Potter (1972)
Lombardy, Italian Alps	45% Cirque	Lobate	76% Metamorphic	Active and inactive	535	326	2110–2540		Guglielmin *et al.* (2001)
Mount Emilius, Valle d'Aosta, Italy	Valley wall	Lobate	Gneiss, calc – schist, greenstones	Active	190	200	2815		Guglielmin *et al.* (1994)
				Inactive	140	150	2550		
	Cirque	Lobate		Inactive	470	240	2310		
				Active	470	270	3000		

Name/Location	Position	Form	Bedrock	Status	Length (m)	Width (m)	Elevation (m)	Latitude	Reference
	Valley floor	Tongue		Active	1800		2600		
SW Alps, France and Italy		Lobate	Metamorphic limestone	Active and inactive	200–300		1500–2850	44°N	Evin (1987)
Urumqi River, Tianshan Mountains	Valley fide	Lobate	Gneiss	Active	30–60	100–150	3900	43°04′–43°08′ N	Cui and Cheng (1988)
Northern Tian Shan and Djungar Ala Tau, Kazakhstan	River basin	Tongue	Granite	Active	750–2000	350–980	2100–3500	42°30′–45°30′ N	Gorbunov and Titkov (1992)
Albanian Alps		Tongue	Carbonates, Flysch, magmatic	Inactive	200–1000	75–325	1690–2200	42°27′–42°33′ N	Palmentola *et al.* (1995)
		Lobate			120–200	100–130			
Colorado Front Range, USA	Cirque	Lobate	Granite	Active	405–640	170–215	3320–3710	40°01′–40°16′ N	White (1971), White (1987); Benedict *et al.* (1986)
Mosquito Range, Colorado, USA	Valley wall	Lobate	Granodiorite	Inactive	90–275	180–600	3330–3975	39°10′–39°17′ N	Vick (1987)

Name/Location	Position	Form	Bedrock	Status	Length (m)	Width (m)	Elevation (m)	Latitude	Reference
La Sal Mountain, Utah, USA	Valley wall, Floor cirque	Lobate	Igneous, sedimentary	Active and inactive	400–1000	150–800	2200–3550	37°42′–38°15′ N	Shroder (1987), Nicholas and Butler (1996); Nicholas and Garcia (1997)
	Cirque	Tongue							
Blanca Massif Colorado, USA	Valley wall	Lobate	Granitic, metamorphic		95–295	195–1100	3600–3885	37°37′–37°37′30″ N	Morris (1981), Parson (1987)
	Cirque	Tongue			400–480	110–160	3700–3750	37°37′–37°37′30" N	
California rock glacier, Blanca Massif, USA	Cirque	Tongue		Active	1460	290	3625	37º37′30′ N	
Mount Mestas, Colorado USA	Valley wall	Tongue	Intrusive igneous	Active	130–1930	50–645	2650–3315	37º35′	Giardino (1979)

Name/Location	**Position**	**Form**	**Bedrock**	**Status**	**Length (m)**	**Width (m)**	**Elevation (m)**	**Latitude**	**Reference**
South Fork Pass, Sierra Nevada, USA	Valley floor	Tongue	Granodiorite, quartz monzonite	Active	1600	200–300	3410–3740	36°25′–37°25′ N	Clark and Clark (1994)
Pokalde Massif, Khumbu Himalaya	Cirque	Tongue		Active	410–1100	100–250	4990–5400	27°55′–28° N	Barsch and Jakob (1998)
Lenana, Mount Kenya	Valley side	Lobate	Nephaline syenite	Inactive	300	340	4750	0º	Grab (1996)
Mendoza, Argentina, Central Andes		Tongue		Active and inactive	2100	500	3200–4800	31°–36° S	Corte (1987)
Kanchanjunga Himal, Eastern Nepal		Lobate		Active	280	140	4800–5300	27° N	Ishikawa *et al.* (2001)
Cordillera Principal, San Juan, Argentina	Cirque	Tongue	Andesite	Active	1250–1900		4000–4900	30° S	Schrott (1991)

Source: Reproduced from Burger, K.C., Degenhardt, J.J., Giardino, J.R., 1999. Engineering geomorphology of rock glaciers. Geomorphology 31, 93–132.

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Table 2 Electromagnetic properties of common earth materials

Material	**Permittivity** (ε)	**Electrical Conductivity** (σ , mS m^{-1})[a]	**Propagation Velocity (v, m ns^{-1})**	**Attenuation** (α , DB m^{-1})
Fresh Water	81	0.5	0.03	0.1
Clay	5–40	2–1,00 000	0.06	1–300
Silt	5–30	1–100	0.07	1–100
Shale	5–15	1–100	0.09	1–100
Limestone	4–8	0.5–2	0.12	0.4–1
Granite	4–6	0.1–1	0.13	0.01–1
Dry Sand	3–5	0.01	0.15	0.01
Ice	3–4	0.01	0.16	0.01
Wet Sand	20–30	0.1–1	0.6	0.03–0.3

a Reciprocal of electrical resistivity.

Table 3 Summary of selected rates of flow studies

Location	**Method**	**Years**	**Motion**	**Rates/flow patterns**	**References**
Galena Creek, Absaroka Mountains, WY	Field surveying	1960s	Horizontal	Upper: 80 cm yr^{-1}; middle: 6 cm yr^{-1}; lower: 14 cm yr^{-1}	Potter (1972)
Galena Creek, Absaroka Mountains, WY	Field surveying	1960s–1990s (30-year interval)	Horizontal	Within 10% of above	Potter *et al.* (1998)
Clear Creek, Wrangel Mountains, AK	Field surveying	1949–1957	Horizontal	Upper: 64 cm yr^{-1}; lower: 57 cm yr^{-1}	Wahrhaftig and Cox (1959)
Arapaho, Front Range, CO	Field surveying	1960–1985 (25-year interval)	Horizontal	Center: 19.3 cm yr^{-1}; edges: 6.2 cm yr^{-1}	Benedict *et al.* (1986)
Maroon and Pyramid, Elk Mountains, CO	Field surveying	1964–1968	Horizontal	Average: 63 cm yr^{-1}	Bryant (1971)
King's Throne, Canadian Rocky Mountains	Field surveying	1988–1989; 1988–1996	Horizontal/ Vertical	Horizontal average: 5.35 cm yr^{-1}; vertical average: 2.49 cm yr^{-1}	Koning and Smith (1999)
Selwyn Mountains, Yukon and NW Territories	Field surveying	1983–1995	Horizontal	Horizontal average: 18 cm yr^{-1}±12 cm	Sloan and Dyke (1998)
Arapaho, Taylor, and Fair, Front Range, CO	Field surveying	1961–1966	Horizontal	Arapaho: 5.0 cm yr^{-1}; Taylor: 6.6 cm yr^{-1}; Fair: 9.7 cm yr^{-1}	White (1971)
Arapaho, Taylor, and Fair, Front Range, CO	Field surveying	1961–2002	Horizontal	Arapaho: 7.3 cm yr^{-1}; Taylor: 6.3 cm yr^{-1}; Fair: 9.5 cm yr^{-1}	Janke (2005d)

Location	**Method**	**Years**	**Motion**	**Rates/flow patterns**	**References**
Hiorthfjellet rock glacier, Svalbard	Field surveying	1994–2002	Horizontal	9.5 to 10.8 cm yr^{-1}	Odergard *et al.* (2003)
Macun I, Lower Engadine, Swiss Alps	Field surveying	1967–1988	Horizontal	14.2±0.4 cm yr^{-1}; unlikely to be sliding	Barsch and Zick (1991)
Nautardalur, North Iceland	Field surveying	1977–1994	Horizontal	Center line velocities average 25 cm yr^{-1}; no sliding is evident	Whalley *et al.* (1995)
Reichenkar, Stubai Alps	Field surveying		Horizontal	Rates greater than 200 cm yr^{-1}; basal sliding present	Krainer and Mostler (2000)
Weissmies, South Wallis	Aerial photography: point-to-point	1958–1964	Horizontal/vertical	Horizontal: maximum in center: 60 cm yr^{-1}; vertical: toe: −30 cm yr^{-1}	Messerli and Zurbuchen (1968) from Barsch (1996)
Grosses Gufer, North Wallis	Aerial photography: point-to-point	1950–1962	Horizontal/vertical	Horizontal: lower section: 50 cm yr^{-1}; vertical: toe: −20 cm yr^{-1}	Messerli and Zurbuchen (1968) from Barsch (1996)

Location	Method	Years	Motion	Rates/flow patterns	References
Murtèl I, Swiss Alps	Aerial photography: point-to-point	1932–1971	Horizontal/ vertical	Horizontal: 5.1±3 cm yr^{-1}; velocities slowed by some 46% over 2nd half of study vertical: −2.8±1 cm yr^{-1}; surface lowering during first part, rising during second half	Barsch and Hell (1975)
Six rock glaciers, Northern Tien Shan and Djungar Ala Tau, Kazakhstan	Aerial photography: point-to-point and Field surveys	1969–1984	Horizontal	Rates vary: 20–146 cm yr^{-1}	Gorbunov and Titkov (1992)
Gruben, Swiss Alps	Aerial photography: 50 by 50 m grid	1970–1975	Horizontal	Greatest near the rock glacier head and toe due to extensional flow, compression near the interior	Haeberli *et al.* (1979)
Doesen, Austrian Alps	Aerial photography: computer point-to-point	1954–1993	Horizontal/vertical	Horizontal: 14.3 cm yr^{-1}; vertical: −1.9 cm yr^{-1}	Kaufmann (1998)

Location	Method	Years	Motion	Rates/flow patterns	References
Gruben, Swiss Alps	Aerial photography: computer 50 by 50 m grid	1970–1979	Horizontal/vertical	Horizontal: most total displacements range from 0 to 400 cm; vertical: Permafrost region: 50 to 150 cm, glacier region: 150 cm	Haeberli and Schmid (1988)
Gruben, Swiss Alps	Aerial photography: computer 25 by 25 m grid	1970–1995	Horizontal/vertical	Horizontal: reached values of 100 cm yr^{-1} with some showing 20% change over a few years; vertical: Permafrost region: average of −5 cm yr^{-1}, glacier region: average of −20 cm yr^{-1}	Kääb *et al.* (1997)
Murtèl, Swiss Alps	Aerial photography: computer 10 by 10 m grid	1987–1996	Horizontal/vertical	Horizontal: head: 15 cm yr^{-1}; toe: 5 cm yr^{-1}; vertical: −4 cm yr^{-1}	Kääb et al. (1998)
Muragl, Swiss Alps	Digital aerial photography: cross correlation with a 3 by 3 m grid	1981–1994	Horizontal/vertical	Horizontal: up to 50 cm yr^{-1}; vertical: most in the range of ±10 cm yr^{-1}	Kääb and Vollmer (2000); Kääb (2002)

Location	Method	Years	Motion	Rates/flow patterns	References
Outer Hochebenkar, Oetztal Alps	Digital aerial photography: cross-correlation with random point-to-point matching	1953–1997	Horizontal/vertical	Horizontal: maximum of 1.8 m yr^{-1} with slower rates in the 1980s; vertical: maximum lowering of 1,800 cm over 44 years	Kaufmann and Ladstädter (2002)
Inner Hochebenkar, Oetztal Alps	Digital aerial photography: cross-correlation with random point-to-point matching	1954–1997	Horizontal/vertical	Horizontal: two active sections seperated by an inactive section; vertical: maximum lowering of 80 cm over 44 years	Kaufmann and Ladstädter (2002)
Various rock glaciers, Front Range, CO	Digital aerial photography: point-to-point	1978–1999	Horizontal/vertical	14 to 20 cm yr^{-1}	Janke (2005a)

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