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Rock glaciers and mountain hydrology: A review

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Abstract.

In mountainous regions, climate change threatens cryospheric water resources, and understanding all components of the hydrological cycle is necessary for effective water resource management. Rock glaciers are climatically more resilient than glaciers and contain potentially hydrologically valuable ice volumes, and yet have received less attention, even though rock glacier hydrological importance may increase under future climate warming. In synthesising data from a range of global studies, we provide the first comprehensive evaluation of the hydrological role played by rock glaciers. We evaluate hydrological significance over a range of temporal and spatial scales, alongside the complex multiple hydrological processes with which rock glaciers can interact diurnally, seasonally, annually, decadally and both at local and regional extents. We report that although no global-extent, complete inventory for rock glaciers exists currently, recent research efforts have greatly elaborated spatial coverage. Using these research papers, we synthesise information on rock glacier spatial distribution, morphometric characteristics, surface and subsurface features, ice-storage and hydrological flow dynamics, water chemistry, and future resilience, from which we provide the first comprehensive evaluation of their hydrological contribution. We identify and discuss long-, intermediate- and short-term timescales for rock glacier storage, allowing a more balanced assessment of the contrasting perspectives regarding the relative significance of rock glacier-derived hydrological contributions compared to other water sources. We show that further empirical observations are required to gain a deeper hydrological understanding of rock glaciers, in terms of (i) their genesis and geomorphological dynamics (ii) total ice/water volume; (iii) water discharge; and (iv) water quality. Lastly, we hypothesise that at decadal and longer timescales, under future climate warming, degradation of ice within rock glaciers may represent an increasing hydrological contribution to downstream regions, and thus increased hydrological significance while rock glacier water stores persist.

Keywords:
Rock glaciers; Mountain cryosphere; Water storage; Water resources; Climate change; Mountain hydrology

1. Introduction

Glacierised high mountain systems worldwide form natural 'water towers' that constitute a significant freshwater source for downstream regions, particularly in arid and semiarid zones (Messerli et al., 2004; Viviroli et al., 2007). Here, glacial- and snowpack-derived meltwaters buffer hydrological seasonality, contributing to streamflow in otherwise low-flow conditions during drier months (e.g., Kaser et al., 2010). In this context, the mountain cryosphere (snow, ice and permafrost) is important for ecosystem services provision (Grêt-Regamey et al., 2012), supplying multiple societal needs within mountains and the surrounding lowlands – potable water supplies, energy generation (hydropower) and agriculture, for example (Immerzeel et al., 2010; Viviroli et al., 2011). In vulnerable drought-prone regions particularly, glaciers represent an important drought-resilient water source (Bolch, 2017). This has been illustrated for several high-altitude cities located in the Andes (Wouter et al., 2017; Table 1).

Table 1. Relative contribution of glacial melt (%) to the water supply of selected cities under different meteorological conditions. Values in square brackets indicate the uncertainty ranges of the estimates. After Wouter et al. (2017).

	Quito (Ecuador)	La Paz (Bolivia)	Huaraz (Peru)
Annual average (normal year)	2.2 [0.9–5.0]	14.8 [5.9–26.8]	19.0 [7.5–35.4]
Monthly maximum (normal year)	5.3 [2.3–11.1]	61.1 [37.8–77.1]	67.3 [41.9–82.8]
Annual average (drought year)	3.7 [1.5–8.0]	15.9 [6.4–29.4]	27.2 [11.6–46.7]
Monthly maximum (drought year)	15.4 [7.3–27.6]	85.7 [74.1–91.5]	91.1 [78.1–96.0]

The rapid near-global retreat of mountain glaciers, predominantly attributed to anthropogenic causes (Marzeion et al., 2014), has previously been reported (Gardner et al., 2013) and glacial retreat and mass loss is projected to continue throughout the twenty-first century (Marzeion et al., 2012; Radić et al., 2014; Huss and Hock, 2015). Furthermore, under high-end climate change scenarios (RCP8.5) Shannon et al. (2019) project global ensemble mean glacier volume loss to be -64 ± 5% (excluding glaciers situated on the periphery of the Antarctic ice sheet) towards the end of the century, with particular regions experiencing mass losses exceeding 75%, including Central Europe, Caucasus, High Mountain Asia and Southern Andes. Even under conservative projections – limiting global warming to 1.5°C above pre-industrial levels (Paris Agreement) – substantial glacier mass loss occurs; e.g., ~36% reductions in glacier mass by 2100 in High Mountain Asia (Kraaijenbrink et al., 2017). Elevation-dependent warming, i.e. rates of warming are amplified with increasing elevation, suggests high-altitude environments will likely experience comparatively faster warming rates than those at lower elevations, particularly at low-latitudes (Vuille et al., 2008; Mountain Research Initiative EDW Working Group, 2015). Indeed, within certain low-latitude mountain ranges complete loss of glaciers has already occurred (Rabatel et al., 2013). Therefore, long-term glacial stores are finite (Jansson et al., 2003).

In the short-term, annual water release from long-term glacial storage increases; however, beyond 'peak water' enhanced melt rates are overwhelmed by glacier shrinkage, and projected glacier runoff gradually declines. For example, recent global-scale projections encompassing 56 glacierised macroscale (>5000 km^2) basins suggests that to date (2017) peak water has been reached in 45% of the 56 studied basins, increasing to 78% by 2050 (RCP4.5) (Huss and Hock, 2018). Critically, by 2100 Huss and Hock (2018) report glacier runoff reductions exceeding 10% during at least one month of the melt season for one-third of the studied basins; the largest of which occur in Central Asian and Andean basins. These findings are corroborated across a range of spatial scales (e.g., Baraer et al., 2012; Sorg et al., 2014; Frans et al., 2016; Hanzer et al., 2018). Importantly, accompanying glacier mass loss, the potential occurrence of increased rain-to-snow fraction (Berghuijs et al., 2014), reductions in snow cover extent and/or duration (Brown and Mote, 2009; Bavay et al., 2013; Zhou et al., 2017), seasonal runoff maxima shifts towards earlier in the year (Hanzer et al., 2018; Huss and Hock, 2018) and widespread permafrost degradation (Haeberli, 2013; Biskaborn et al., 2019) will further reduce long-term runoff in snowmelt- and icemelt-dominated basins.

The anticipated effects of the future decline of the mountain cryosphere pose far-reaching challenges for effective freshwater resource management. Several studies have previously summarised and discussed such impacts on anthropogenic and ecologic systems (see Beniston, 2003; Barnett et al., 2005; Bolch et al., 2012; Huss et al., 2017; Milner et al., 2017; Beniston et al., 2018), and we refer interested readers to those manuscripts for details. In terms of climate change adaptation strategies, an understanding of all components of the hydrological cycle in high mountain systems is vital (Jones et al., 2018a). However, whilst much has been written on the hydrological role of debris-free glaciers (see Fountain and Walder, 1998; Jansson

53   et al., 2003; Irvine-Fynn et al., 2011) and increasingly on debris-covered glaciers (Fyffe et al., 2019, and
54   references therein), that of rock glaciers has received comparatively less attention (Duguay et al., 2015).
55
56   Rock glaciers are landforms consisting of a continuous, thick seasonally frozen debris layer (known as the
57   *active layer* [AL]) covering ice-supersaturated debris or pure ice (Berthling, 2011; Bonnaventure and
58   Lamoureux, 2013). They are formed by gravity-driven creep as a consequence of internal ice deformation.
59   Intact rock glaciers (i.e. those features within which ice presence is expected beneath the AL [Barsch,
60   1996]) are thought to contain ice volumes of significant value (Azócar and Brenning, 2010; Rangecroft et
61   al., 2015; Jones et al., 2018b; Munroe, 2018). Critically, due to the insulating effect of the AL, internal ther-
62   mal regimes are at least partially decoupled from external micro- and meso-climates in summer (Juliussen
63   and Humlum, 2008; Millar et al., 2013). As a result, rock glaciers are reasonably assumed to have retarded
64   ice melt, which suggests these landforms may prolong long-term water storage in high mountain systems
65   and buffer losses from alternative sources (Millar et al., 2013; Rangecroft et al., 2015; Bosson and Lambiel,
66   2016; Jones et al., 2018b). Furthermore, rock glacier presence and abundance affects the amount and prop-
67   erties of runoff from high mountain watersheds. The potential hydrological value of rock glaciers, and thus
68   their importance in terms of hydrological research, was first noted by Corte (1976). Yet, despite an accel-
69   eration of rock glacier-related research during recent decades – searches within Scopus® using 'Article title,
70   Abstract, Keywords' for all spelling variants of *rock glacier(s)* generated 973 peer-reviewed articles and
71   reviews to date[1], 390 (40%) of which have been published since 2010 – research focusing upon their hy-
72   drological role remains limited. For example, Duguay et al. (2015) report that searches of both ISI Web of
73   Science™ and Geobase™ resulted in just 28 papers with an appropriate discussion on rock glacier hydrol-
74   ogy. Therefore, despite the near-ubiquitous nature of rock glaciers in high mountain systems (Jones et al.,
75   2018a), there remains a need to understand the state-of-knowledge regarding the hydrological role of rock
76   glaciers. The rationale and aim of this manuscript is thus to describe the state of current scientific
77   knowledge about the hydrological role and contribution of rock glaciers in mountain regions and to sign-
78   post towards critical future directions for rock glacier hydrological research.

79   **2. Rock glaciers**

80   *2.1. Rock glacier characteristics*

81   Rock glaciers are lobate or tongue-shaped assemblages of poorly sorted, angular-rock debris and ice (ice-
82   cored or ice-cemented [Section 2.2]) commonly found in high mountain environments, which move as a
83   consequence of the deformation of internal ice (Giardino and Vitek, 1988; Barsch, 1996). Previously, rock
84   glaciers have been classified according to their ice content and dynamic behaviour (Fig. 1); *active rock glac-
85   iers* contain ice and display movement, *inactive rock glaciers* contain ice and no longer display movement
86   and *fossil rock glaciers* do not contain ice and no longer move, and are referred to regularly as *relict*
87   (Haeberli, 1985; Barsch, 1996). Importantly, Kääb (2013) notes that "this classification is a theoretical con-
88   cept. The transition between the three stages is actually continuous".
89
90   *Active rock glaciers* display rates of movement in the order of centimetres to a few decimetres per year (see
91   Table 3 in Janke et al., 2013), although cases have been reported with surface velocities of several metres
92   per year (Gorbunov et al., 1992; Kääb et al., 2003; Delaloye et al., 2013; Sorg et al., 2015; Hartl et al., 2016b;
93   Eriksen et al., 2018). They are characterised by distinctive flow-like morphometric features reflecting their
94   visco-plastic properties; spatially organised longitudinal or transverse ridge-and-furrow assemblages,
95   steep (∼>30–35°) and sharp-crested front and lateral slopes that typically rise 15–70 m above adjacent
96   terrain, light-coloured (i.e. little weathered) frontal slope in contrast to the dark-coloured rock-varnished
97   (i.e. greatly weathered) uppermost surface, a swollen, noticeably longitudinally convex appearance of the
98   rock glacier body, and an absence of vegetation and/or lichen cover (Wahrhaftig and Cox, 1959; Martin and
99   Whalley, 1987; Baroni et al., 2004; Haeberli et al., 2006; Harrison et al., 2008).
100
101   *Inactive rock glaciers* still contain ice but are immobile. Two types of inactivity can be defined: *climatically
102   inactive*, where the ice has melted; and *dynamically inactive* where there is reduced nourishment of talus
103   and/or ice as the rock glacier extends too far from the headwall (e.g., Kellerer-Pirklbauer and Rieckh, 2016),
104   are cited as possible causes for this behaviour (see Barsch, 1996, p. 8–10). In comparison to active rock
105   glaciers, the surface micro-topography of inactive features is relatively subdued. They generally have

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[1] March 2019.

106   gentler, dark-coloured rock-varnished frontal slopes with partial to full vegetation and/or lichen cover
107   (Ikeda and Matsuoka, 2002). As a consequence of the difficulty of differentiating between active and inac-
108   tive forms, particularly through photogeomorphology, active and inactive rock glaciers are often collec-
109   tively termed *intact*.
110
111   *Relict rock glaciers*, defined as former rock glaciers, no longer contain any ice and are characterised by sub-
112   dued surface micro–topography. They often exhibit surface collapse features including thermokarst ponds,
113   i.e. water-filled depressions resulting from melting of stagnant glacial ice, and have gentler ($\sim$<30°) and
114   round-crested front and lateral slopes, a dark-coloured rock-varnished frontal slope, and extensive vege-
115   tation and/or lichen cover (Martin and Whalley, 1987; Barsch, 1996; Baroni et al., 2004; Harrison et al.,
116   2008).
117

Fig. 1. Typical examples of active [a, b], inactive [c, d] and relict [e, f] rock glaciers from around the world: (a) Sourdough rock glacier, Wrangell Mountains, AK, USA (61°23'N, 142°44'W). Note the steep headwall that serves as a source of snow avalanches and rockfall; (b) Caquella rock glacier, Bolivian Andes of South Lipez, Bolivia (21°29'S, 67°55'W); (c) Liapey d'Enfer rock glacier, Hérens valley, Swiss Alps, Switzerland (46°05'N, 7°32'E); (d) rock glaciers in the Niggelingtälli, Turtmann Valley, Swiss Alps, Switzerland (46°13'N, 7°45'E); (e) Hoelltal rock glacier, Niedere Tauern

5

124 Range, Central Eastern Alps, Austria (47°22'N, 14°39'E); and (f) rock glaciers beneath Le Mourin mountain, Valais,
125 Swiss Alps, Switzerland (45°56'N, 7°10'E). On the photographs, dashed lines correspond to the approximate rock glac-
126 ier boundary. Images [a] modified after Anderson et al. (2018) and [b–f] from Google Earth.
127
128 It is important to note that recognition, delineation and classification (in terms of their activity) is not a
129 trivial undertaking both from *in situ* surveys, or from analysis of remote sensing/photographic data, par-
130 ticularly where: (i) the "characteristic" surface morphology is not evident (Whalley et al., 1986), and/or (ii)
131 rock glaciers form multi-lobed complexes with lobes of different activity types superimposed and embed-
132 ded onto/into one another, forming a cascading form with active lobes at higher elevations and relict lobes
133 at lower elevations (Roer and Nyenhuis, 2007). Regardless, identifying and establishing the activity status
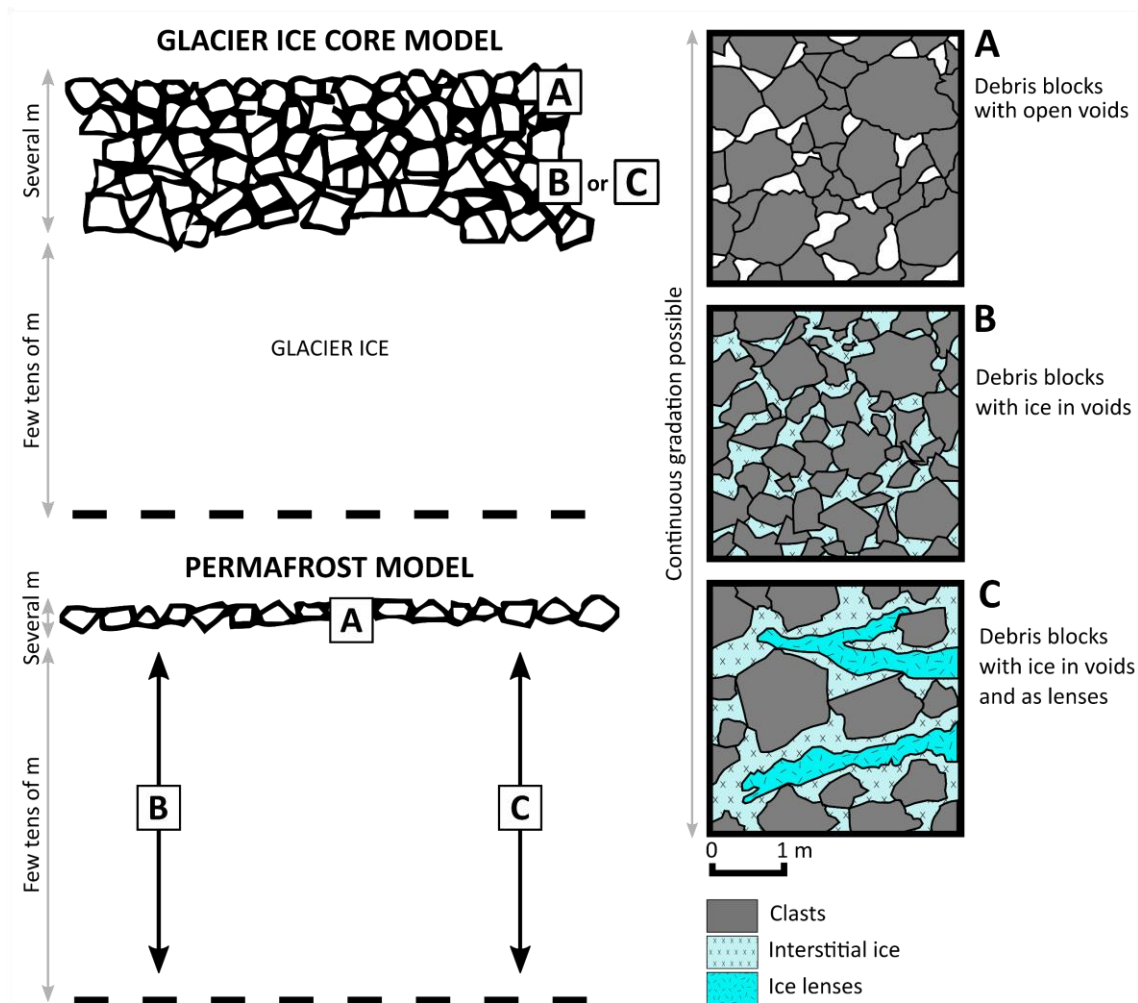134 of rock glaciers represents an important initial step in determining their potential hydrological significance.

135 *2.2. Rock glacier origin and evolution*

136 A long-standing academic debate regarding rock glacier origin pervades the literature (see Barsch, 1977,
137 1987, 1996; Whalley and Martin, 1992; Hamilton and Whalley, 1995; Clark et al., 1998; Whalley and Azizi,
138 2003; Haeberli et al., 2006; Berthling, 2011). Specifically, the rock glacier controversy can be framed as the
139 permafrost model where internal ice is assumed to be of a dominantly periglacial/permafrost origin
140 (Haeberli, 1985) vs. the glacier ice core model where a dominant glacial origin is assumed. In reality, within
141 deglacierising mountains rock glaciers are equifinal inasmuch as they can arise from separate or combined
142 periglacial, glacial and paraglacial (i.e. landscape relaxation) processes (Knight et al., 2019). Indeed, glaci-
143 ological vs. slope vs. climatic controls on the evolution of rock glaciers may vary through the rock glacier
144 life cycle (ibid.). Here, we do not seek to contribute to this debate (this is beyond the scope of this paper) –
145 several papers have already discussed the arguments for and against the different positions (see Whalley
146 and Martin, 1992; Barsch, 1996; Haeberli et al., 2006; Berthling, 2011, for reviews); rather, we adopt the
147 dispassionate view that rock glaciers can be derived through either model (as per Berthling, 2011) and
148 proceed to describe these hypotheses in further detail below.

149
150 The permafrost model for rock glacier genesis follows Wahrhaftig and Cox (1959) where interstitial ice
151 (pore-ice) or segregated ice (ice lenses), primarily derived from either the freezing of rain and meltwater
152 percolating through the rock glacier matrix, the freezing of groundwater, and/or the burial of snow and ice
153 accumulations (Burger et al., 1999; Haeberli, 2000; Haeberli et al., 2006), produces creep (Wahrhaftig and
154 Cox, 1959; Barsch, 1977, 1987, 1988, 1996; Haeberli, 1985). Rock glaciers of periglacial origin are often
155 called *talus-derived rock glaciers* (Humlum, 1996) or *ice-cemented rock glaciers* (Wayne, 1981). Of note is
156 the occurrence of rock glaciers in nonglacial, periglacial environments, therefore past or present glaciation
157 is not a prerequisite for their formation (Haeberli, 1985; Giardino and Vitek, 1988; Hamilton and Whalley,
158 1995). Proponents of this model suggest that little evidence exists to support the glacier ice core model
159 (see Berthling, 2011, for detailed discussion), although they acknowledge the possibility of incorporation
160 of glacier (sedimentary) ice in permafrost (e.g., Haeberli, 1989; Haeberli and Vonder Mühll, 1996; Kääb et
161 al., 1997; Haeberli et al., 2006). In such cases it is argued that the long-term existence of these rock glaciers
162 requires permafrost conditions (Harris and Murton, 2005; Berthling, 2011; Kääb, 2013).

163
164 Based on *in situ* data, other authors have presented evidence that supports a glacigenic origin (i.e. glacier
165 ice core model) for certain rock glaciers (Outcalt and Benedict, 1965; Potter, 1972; Whalley, 1974; White,
166 1976; Whalley et al., 1994; Humlum, 1996; Potter et al., 1998; Ishikawa et al., 2001; Monnier et al., 2013;
167 Petersen et al., 2016; Guglielmin et al., 2018). This model involves the creep of a thin (typically <50 m) ice
168 body, which has been buried and preserved by an insulating weathered rock debris layer (Whalley and
169 Martin, 1992; Whalley and Azizi, 2003). Recently, Anderson et al. (2018) showed that rock glaciers of
170 glacigenic origin represent a plausible end-member response that is captured in numerical models of de-
171 bris-free glaciers. Rock glaciers of this type are also referred to as *moraine-derived rock glaciers* (Harrison
172 et al., 2008)*, ice-cored rock glaciers* (Potter, 1972)*, glacier ice-cored rock glaciers* (Johnson, 1978) or *glacier-
173 derived rock glaciers* (Humlum, 1996). Figure 2 represents a simplified perspective of rock glacier compo-
174 sition for features derived from either the permafrost or glacier ice core model as outlined by Martin and
175 Whalley (1987).
176

6

GLACIER ICE CORE MODEL

Several m

Few tens of m

GLACIER ICE

PERMAFROST MODEL

Several m

Few tens of m

Continuous gradation possible

A
Debris blocks
with open voids

B
Debris blocks
with ice in voids

C
Debris blocks
with ice in voids
and as lenses

0   1 m

Clasts
Interstitial ice
Ice lenses

177
178 **Fig. 2.** Basic representation of rock glacier internal composition associated with the glacier ice-core and permafrost
179 models. Modified after Martin and Whalley (1987).
180
181 Proponents of the two competing concepts of rock glacier formation generally focus on already well-devel-
182 oped landforms (Monnier and Kinnard, 2015b). Yet, there are striking examples of large glacier-rock glacier
183 composite features where the lowermost parts of debris-covered glaciers are *currently developing* into rock
184 glaciers. Cases of glacier-to-rock glacier transition within relatively short timescales have been reported;
185 for example, within ca. 50 years at Sachen Glacier, Nanga Parbat, Pakistan (Shroder et al., 2000) and ca. 60
186 years at Presenteseracae debris-covered glacier, central Andes, Chile (Monnier and Kinnard, 2015b). Nev-
187 ertheless, studies of glacier-rock glacier interactions are rarely orientated towards such examples (Monnier
188 and Kinnard, 2015b), in spite of their potential to elucidate the drivers behind which glaciers will fully
189 transition into rock glaciers and those which will simply downwaste.
190
191 Here, it is important to distinguish between rock glaciers (cf. Section 2.1) and debris-covered glaciers
192 (Hambrey et al., 2008; Benn and Evans, 2010; Cogley et al., 2011; Jiskoot, 2011; Kirkbride, 2011), particu-
193 larly when reviewing the hydrological significance of the former; grouping these features would incorrectly
194 inflate the hydrological significance of rock glaciers. Debris-covered glaciers are glaciers in which the abla-
195 tion zone is partly or wholly covered with thin (typically less than several-decimetres thick) supraglacial
196 debris (Kirkbride, 2011). There are critical differences between rock glaciers and debris-covered glaciers :
197 rock glacier movement is governed by internal deformation, the majority of which occurs in a shear zone
198 at depth within the feature (Arenson et al., 2002; Buchli et al., 2013, 2018; Krainer et al., 2015; Kenner et
199 al., 2017a); debris-covered glacier motion may occur due to internal deformation, basal sliding, and soft
200 bed deformation. Importantly, basal sliding is generally non-occurring or very limited for cold-based de-
201 bris-covered glaciers (i.e. glaciers frozen to their beds), and only debris-covered glaciers underlain by a soft
202 deformable substrate (i.e. unlithified sediments or poorly consolidated sedimentary rocks) exhibit soft bed
203 deformation. The surface of debris-covered glaciers is complex topographically, consisting of a spatially-
204 chaotic mosaic of features such as hummocks, depressions, supraglacial melt ponds and ice cliffs; the

7

surficial instability and discontinuity of the debris-covered glacier debris layer mean that such ice expo-
sures are frequently visible. Contrastingly, rock glacier surfaces are characterised by spatially coherent
flow-like topography (Section 2.1) and surficial ice-exposures are infrequent.

Monnier and Kinnard (2017) identify three types of glacier-rock glacier dynamic relationships within the
literature: (i) the re-advance of glaciers or debris-covered glaciers overriding older permafrost bodies and
coalescing (Lugon et al., 2004; Haeberli, 2005; Ribolini et al., 2007, 2010; Monnier et al., 2013; Dusik et al.,
2015; Bosson and Lambiel, 2016; Kellerer-Pirklbauer and Kaufmann, 2018; Kenner, 2019). Such glacier-
rock glacier interactions are defined by the permafrost model; (ii) the continuous emergence of a rock glac-
ier from a debris-covered glacier, as indicated by the evolution of the surface morphology, together with
the creep of a massive and continuous ice body that has been buried and preserved (Potter, 1972; Potter
et al., 1998; Humlum, 2000; Krainer and Mostler, 2000). These types of glacier-rock glacier interactions are
described by the glacier ice core model; (iii) debris-covered glacier-to-rock glacier transition through the
evolution of the surface morphology and the internal structure, i.e. the development of a perennially frozen
ice-rock mixture resulting from the incorporation of surface-derived debris and periglacial ice and frag-
mentation of the initial massive and continuous ice body (Monnier and Kinnard, 2015b, 2017; Seppi et al.,
2015). Monnier and Kinnard (2015b) describe this as an alternative to the dichotomous debate between a
periglacial or glacial origin. Regarding (ii) and (iii), the emerging rock glacier and debris-covered glacier
morphology are in complete continuity, suggesting a continuum between debris-free glaciers and rock glac-
iers where debris-covered glaciers form an intermediate stage (Giardino and Vitek, 1988; Ackert, 1998).

With continued climate-driven deglaciation, high mountain systems are transitioning from glacial to parag-
lacial dominated process regimes (Harrison, 2009). Ballantyne et al. (2014) note that formerly glacierised
high mountain systems "are characterised by a high spatial density of large-scale post-glacial rock-slope
failures (RSFs) such as major rockfalls, rockslides, rock avalanches and deep-seated gravitational slope de-
formations". RSFs occur in response to debuttressing or deglacial unloading that follows the exposure of
glacially steepened rockwalls by glacier downwastage and retreat (Ballantyne, 2002; Fischer et al., 2006).
In the context of current and future climatic conditions, enhanced paraglacial processes (i.e. RSFs) driven
by continuing deglaciation could increase supraglacial debris accumulation upon glacial forms, and thus
limit ablation of the underlying ice (Section 3) (Lambrecht et al., 2011; Pellicciotti et al., 2014; Lardeux et
al., 2016). Thick supraglacial debris cover (i.e. decimetres to metres) influences glacier dynamics signifi-
cantly, as inefficient sediment evacuation processes encourage glacier-rock glacier interactions (Shroder
et al., 2000). Recent numerical model simulations report that under significant warming (+2.5°C) debris-
covered glacier-to-rock glacier transition can occur rapidly (∼100 years) (Anderson et al., 2018). There-
fore, glacier-rock glacier interactions could enhance the resilience of the mountain cryosphere, preserving
frozen water stores as glaciers transition to rock glacier forms. Importantly, however, these landform as-
semblages are not included in glacio-hydrological modelling, which usually focuses solely on glaciers (Bolch
et al., 2019, and references therein). As such, an improved understanding of glacier-rock glacier dynamic
relationships is critically important in the context of future water resources management. We further argue
that of particular importance are empirical studies that seek to elucidate the drivers behind which glaciers
will fully transition into rock glaciers and those which will simply downwaste.

2.3. Rock glacier water storage

Rock glacier water storage occurs as ice, snow and water at *long-term*, *intermediate-term* and *short-term*
timescales (Jones et al., 2018b; Fig. 3), similar to glacier storage (cf. Jansson et al., 2003). Long-term storage
concerns ice storage below the AL of rock glaciers on multi-annual to centennial and millennial timescales
(e.g., Clark et al., 1996; Krainer et al., 2015). Recent research concludes that rock glaciers constitute non-
negligible long-term water stores, particularly in deglacierising/deglacierised semi-arid and arid regions
(Section 4). In addition, this storage also includes the release of these water stores through degradation of
the internal ice body. Intermediate-term storage encompasses the storage and release of snowmelt and AL
thaw-derived runoff on a seasonal timescale (Section 5). Short-term storage includes diurnal drainage of
en- and sub[rock]glacial water through the rock glacier (Section 5). In this context, intermediate-term and
short-term storage can strongly influence catchment runoff characteristics, especially in catchments where
rock glaciers are more abundant than glaciers. In addition, event storage and releases (i.e. singular storage
releases), which have irregular occurrences and/or irregular intervals, are also types of rock glacier stor-
age; for example, intermittent thermokarst ponds (Giardino et al., 1992; Haeberli et al., 2001; Kääb and
Haeberli, 2001) or rock glacier-dammed lakes formed as a consequence of river channel disruption (Hewitt,
2014; Rosenwinkel, 2018; Blöthe et al., 2019). Detailed knowledge of the timescales and forms of rock

262 glacier water storage represents critical information for the management of such resources but is currently
263 lacking.
264
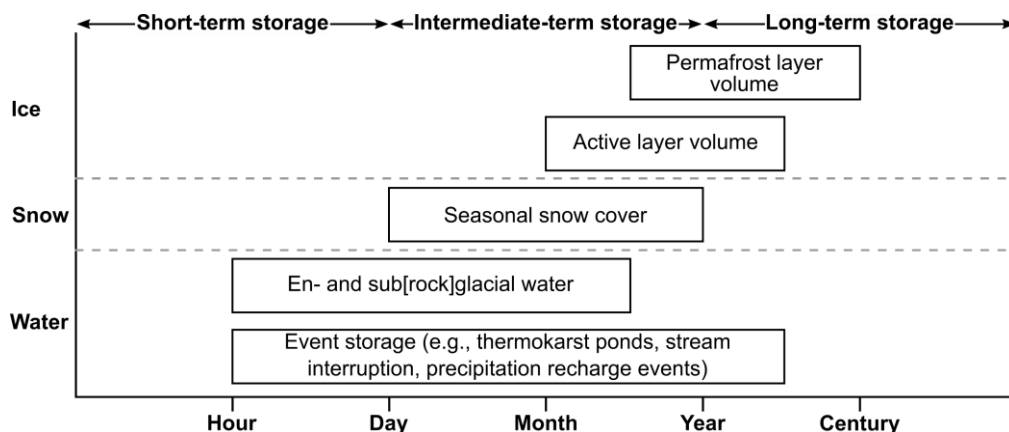
265
266 **Fig. 3.** Schematic diagram showing the different forms of rock glacier storage and their associated timescales. Modified
267 after Jones et al. (2018b).

268 **3. Ice preservation**

269 Active rock glacier internal structure consists of two basic units: firstly, the AL, which is typically a few
270 decimetres to a few metres in thickness and secondly, underlying the AL, a core of ice-supersaturated de-
271 bris (i.e. nonglacial) or pure ice (i.e. glacial) (Ballantyne, 2018, p. 243). The AL typically comprises coarse
272 blocky clasts ~0.2–5.0 m in length (Humlum, 1997) (Fig. 4a), although boulders exceeding 5 m in length
273 are sometimes present (Fig. 4b).
274

275
276 **Fig. 4.** Ground-based views of the lowermost part of the Chola Glacier (Khumbu Himal, Nepalese Himalaya, May 2017),
277 an ongoing glacier-rock glacier interaction, including: (a) the coarse-blocky openwork debris that is characteristic of
278 rock glacier AL surfaces. Field sampling of the visible clasts suggests mean grain size ~1.9 m (b-axis); and (b) the vari-
279 ability of grain size (i.e. >5 m blocks) visible at rock glacier surfaces. The grain size variability reflects the diversity of

9

280   processes (e.g., rockfalls, snow avalanches etc.) by which debris is transferred to the rock glacier surface (Haeberli et
281   al., 2006). Researcher for scale. Photos: Darren B. Jones.
282
283   Complex internal thermal regimes are generated within the coarse-blocky openwork structure of the AL,
284   such that the AL acts as a thermal filter between the surface energy balance and the subjacent frozen rock
285   glacier core (Humlum, 1997; Hanson and Hoelzle, 2004). Indeed, Gruber and Hoelzle (2008) illustrate that
286   both the temperature beneath the seasonal snow cover and mean annual ground temperatures at greater
287   depth can be considerably reduced due solely to the low thermal conductivity of the openwork blocky de-
288   bris according to a simple and purely conductive model. Consequently, the thermal regime within coarse-
289   blocky openwork debris is different to the surrounding environment leading to the development of cooler
290   ground temperatures when compared to fine-grained or bedrock surfaces in similar settings (Harris and
291   Pedersen, 1998; Juliussen and Humlum, 2008) and mean annual air temperatures (MAATs) (Gorbunov et
292   al., 2004; Millar et al., 2013). This leads to the continued persistence of buried ice at elevations where MAAT
293   exceeds 0°C, i.e. elevations where permafrost persistence outside of rock glaciers and similar landforms is
294   highly improbable (e.g., Baroni et al., 2004; Popescu et al., 2017), occasionally reaching below the tree line
295   (e.g., Sorg et al., 2015; Charbonneau and Smith, 2018) (Fig. 1a).
296
297   It is probable that a number of non-conductive (i.e. convective and advective heat flow) and conductive
298   processes are responsible for the cooling influence of openwork blocky debris in relation to rock glaciers,
299   talus slopes and blockfields (Harris and Pedersen, 1998). These processes include: (i) the Balch effect; (ii)
300   the chimney effect; (iii) continuous air exchange with the atmosphere; (iv) summer-time evaporation or
301   sublimation of water and/or ice; and (v) conductive heat loss via boulder protrusion. At different times of
302   the year all of these mechanisms could feasibly be operating, but under different circumstances governed
303   by the duration, depth and permeability of seasonal snow cover, local topography (i.e. slope length and
304   gradient) and openwork blocky debris thickness (Ballantyne, 2018, p. 51). Each is summarised below.
305

306   (i)   *The Balch effect (or Balch ventilation).* This is defined as the insulating effect of air-filled
307   voids within the openwork blocky debris (Thompson, 1962; Barsch, 1996, p. 238). During
308   periods of low wind speeds and no or little snow cover, density-induced cooling occurs as
309   comparatively cold air masses penetrate the openwork blocky debris and displace
310   warmer air masses. In contrast to this, warmer air masses at the surface are prevented
311   from penetrating into the openwork blocky debris by cold air trapped within inter-clast
312   voids. Consequently, heat transfer is conductive only, thus warming of the openwork
313   blocky debris at depth is sluggish, whereas density-induced non-convective processes
314   (e.g., Balch ventilation) take place much more rapidly, particularly at night (e.g., Humlum,
315   1997). Importantly, wind pumping produced by high winds, however, may upset the
316   above-described pattern and even dynamically displace trapped cold air as comparatively
317   warmer and less dense near-surface air masses are driven into the openwork blocky de-
318   bris by means of forced ventilation (Humlum, 1997). Of note, Ballantyne (2018, p. 50) n
319   that most investigators reject this explanation of negative temperature anomalies within
320   openwork blocky debris as it fails to explain seasonal temperature variations at depth.
321   (ii)   *The chimney effect.* First proposed by von Wakonigg (1996), this theory concerns a sea-
322   sonally switching air circulation mechanism that produces ascending warm air in winter
323   and descending cold air in summer. Consequently, a positive and negative thermal anom-
324   aly develops in the upper and lower parts of the slope, respectively. Specifically, in winter,
325   warmer air masses move upslope through the openwork blocky debris under thick sea-
326   sonal snow cover, exiting through warm funnels in the upper parts of the slope. As a result,
327   the concomitant aspiration of cold air into the lower parts of the slope occurs. This process
328   is most efficient during particularly cold periods leading to the build-up of a cold reservoir
329   in the lower parts of the slope. In summer, this mechanism is reversed, with a gravity dis-
330   charge of cold air through the openwork blocky debris, exiting at the foot of the slope
331   (Harris and Pedersen, 1998; Delaloye and Lambiel, 2005). This process is frequently per-
332   ceptible as a persistent breeze towards the base of rock glacier frontal slopes during the
333   summer (Delaloye and Lambiel, 2005). Importantly, the chimney effect, therefore, facili-
334   tates the preservation of ice in openwork blocky debris on slopes >1000 m below the re-
335   gional limit of discontinuous permafrost, where MAAT is >+5°C (Delaloye and Lambiel,
336   2005). In addition, the results of recent 2-D numerical modelling (cf. Wicky and Huack,
337   2017) successfully replicate the above-described mechanism, placing further emphasis on
338   its importance for the preservation of frozen ground under future climate warming.

    (iii)   *Continuous air exchange with the atmosphere.* Proposed by Harris and Pedersen (1998), this theory is a simple extension of the chimney effect (ii) in areas where continuous snow cover is lacking. In its absence, continuous air exchange with the atmosphere occurs over the entirety of the openwork blocky debris occurs; therefore, in response to changes in air temperature, almost instantaneous warming and cooling of the openwork blocky debris to a considerable depth occurs. This process is most effective in windy situations on steep slopes.

    (iv)   *Summer-time evaporation or sublimation of water and/or ice.* Latent heat absorbed from block surfaces in these processes may cool the blocks in the upper layers of the openwork blocky debris (von Wakonigg, 1996). This process is most effective in regions with relatively dry summer air (Harris and Pedersen, 1998).

    (v)   *Conductive heat loss via boulder protrusion.* Boulders protruding into and through the seasonal snow cover act as "efficient heat bridges", thermally coupling the openwork blocky debris and the surface air in winter and thereby enhancing conduction (Juliussen and Humlum, 2008). In block fields (Elgåhogna and Sølen Mountains, central-eastern Norway), Juliussen and Humlum (2008) recorded mean annual ground temperatures 1.3–2.0°C lower than nearby till and bedrock.
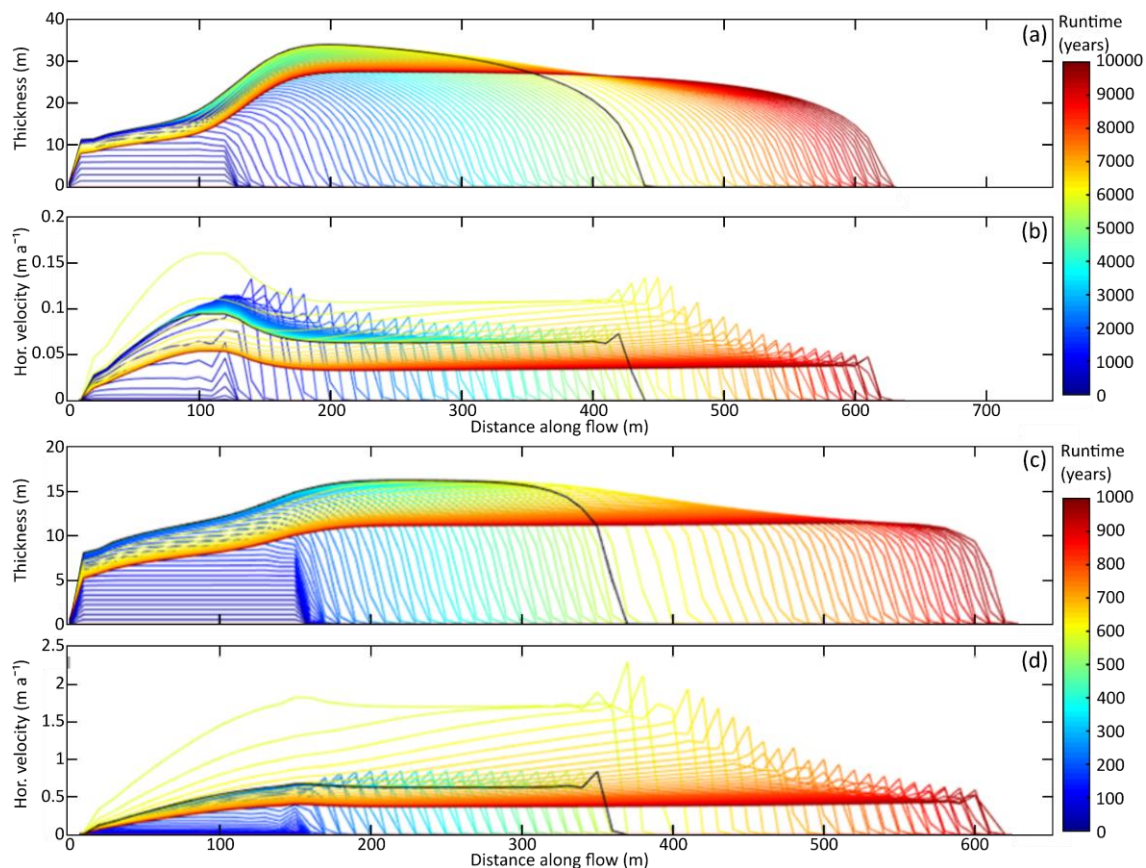
In addition to the above-described processes, ice subjacent to the AL can 'self-preserve' as it increases the heat capacity, thermal conductivity and emissivity of the ground relative to unfrozen ground within which the pore spaces are air-filled. It may act as an effective heat sink, rapidly absorbing, emitting and distributing heat without undergoing high amounts of melt (Kenner et al., 2017b). Critically, therefore, rock glaciers have retarded melt as the ice underlying the AL is thermally decoupled from external macro- and meso-climates. Indeed, long-term ice preservation is feasible within rock glaciers previously considered morphologically inactive or relict (Section 2.1) (e.g., Scapozza et al., 2011; Harrington et al., 2018; Colucci et al., 2019), as the lower ice content potentially increases AL porosity [and potentially void-interconnectedness], which improves internal air circulation (Delaloye and Lambiel, 2005). For example, field investigations at the inactive Helen Creek rock glacier (Banff National Park, Alberta, Canada) indicate the presence of isolated patches of subsurface ice where the AL is most coarse (clasts often >1 m b-axis) (Harrington et al., 2018). As a consequence, the response of rock glaciers to present-day and future climate warming may occur at comparatively longer timescales than glaciers (Giardino et al., 2011; Kenner and Magnusson, 2017; Anderson et al., 2018; Jones et al., 2018a). Indeed, Janke et al. (2013) note that "[u]nlike glaciers that are sensitive to extreme fluctuations on a shorter timescale, a strong climatic signal must exist to produce change in a rock glacier system". On this basis, it is reasonable to conclude that the hydrological role of rock glaciers (i.e. long-term storage) will likely become increasingly important with continued glacier recession (Bolch and Marchenko, 2006; Millar and Westfall, 2008), at least until the rock glaciers themselves become relict.

Of note, however, are emerging observations of increased rock glacier surface velocities potentially in response to recent climate warming at both decadal (Kääb et al., 2007; Kaufmann and Ladstädter, 2007; Delaloye et al., 2008; Bodin et al., 2009; Scapozza et al., 2014; Nickus et al., 2015; Sorg et al., 2015) and seasonal (Kääb et al., 2003; Perruchoud and Delaloye, 2007; Delaloye et al., 2010; Liu et al., 2013; Wirz et al., 2016) timescales. In the European Alps, for example, annual terrestrial geodetic surveys conducted at 16 rock glacier lobes suggest that relative to the respective previous year, mean surface velocity increase was ~+52% during the reporting period (2010–2014) with most rock glaciers reaching new surface velocity maxima in 2014 (PERMOS, 2016). Elsewhere, striking examples of rapid rock glacier movement (i.e. several metres annually) (Delaloye et al., 2013; Hartl et al., 2016b; Eriksen et al., 2018) and partial rock glacier collapse (Bodin et al., 2017) have been explained, at least in part, by climatic factors. Furthermore, the synchronous kinematic behaviour recorded at multiple rock glaciers with considerably different local conditions (e.g., elevation, topographical conditions, lithology), has been attributed to external drivers (i.e. climate warming) (Kääb et al., 2007; Delaloye et al., 2008, 2010; Kellerer-Pirklbauer and Kaufmann, 2012; Sorg et al., 2015; Kellerer-Pirklbauer et al., 2018). In explanation, increased rock glacier surface velocities are potentially due to modification of the rheological behaviour of warming ice (i.e. increased internal deformation) (Kääb et al., 2007) and/or hydrological effects (i.e. increased water content) within the frozen rock glacier core (i.e. higher pore pressure) or at the base (i.e. possibly basal sliding) (Ikeda et al., 2008; Roer et al., 2008; Buchli et al., 2013, 2018; Kenner et al., 2017a).

Rock glaciers with ground temperatures close to 0°C are reported to have a higher thermal responsiveness and creep faster in general than colder features (Kääb et al., 2007; Müller et al., 2016), and thus are likely

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398   to be more vulnerable to future climate warming. Numerical flow modelling of the Huhh1 and Murtèl rock
399   glaciers (Swiss Alps, Switzerland) by Müller et al. (2016) indicate that an immediate 1°C increase in rock
400   glacier temperature considerably affects rock glacier horizontal velocity and surface geometry (Fig. 5) –
401   features with temperatures closest to 0°C are most impacted. Further, Marcer et al. (2019) modelled the
402   destabilisation susceptibility of rock glaciers (see section 'Reaction Typology' in Schoeneich et al., 2015),
403   reporting that occurrence of this phenomenon is more likely in elevations near the 0°C isotherm, north-
404   facing aspects, steep slopes (25–30°) and flat to slightly convex topographies. Indeed, they identify signifi-
405   cant evidence of destabilisation (e.g., crevasses and scarps) at 46 active rock glaciers in the French Alps –
406   10% of all active rock glaciers in this region. Research considering the implications of future climate warm-
407   ing on rock glaciers is therefore much required, yet it remains in its infancy (Rangecroft et al., 2016). In one
408   of the few studies to date, Rangecroft et al. (2016) modelled future MAAT projections at intact rock glacier
409   sites in the Bolivian Andes. Assuming a conservative rock glacier activity threshold of +2°C (MAAT), the
410   authors conclude that >98% of currently intact inventoried landforms will become relict by 2080. Similarly,
411   within the Argentinian Andes >50% of currently active rock glaciers will terminate below the mean annual
412   0 °C isotherm by 2050 under RCP2.6 (i.e. best-case climate scenario) (Drewes et al., 2018). Under RCP8.5
413   (i.e. worst-case climate scenario) the majority will terminate below the mean annual 0 °C isotherm by 2070,
414   e.g., 92.6% (~4120 of 4449 currently active rock glaciers) in the Central Andes sub-region (ibid.). While no
415   allowances were made for a lagged response of rock glaciers to the modelled temperature shifts (see
416   Rangecroft et al., 2016; Drewes et al., 2018), considering the above-described emerging observations, in a
417   changing climate the hydrological significance of rock glaciers as long-term stores is questionable.
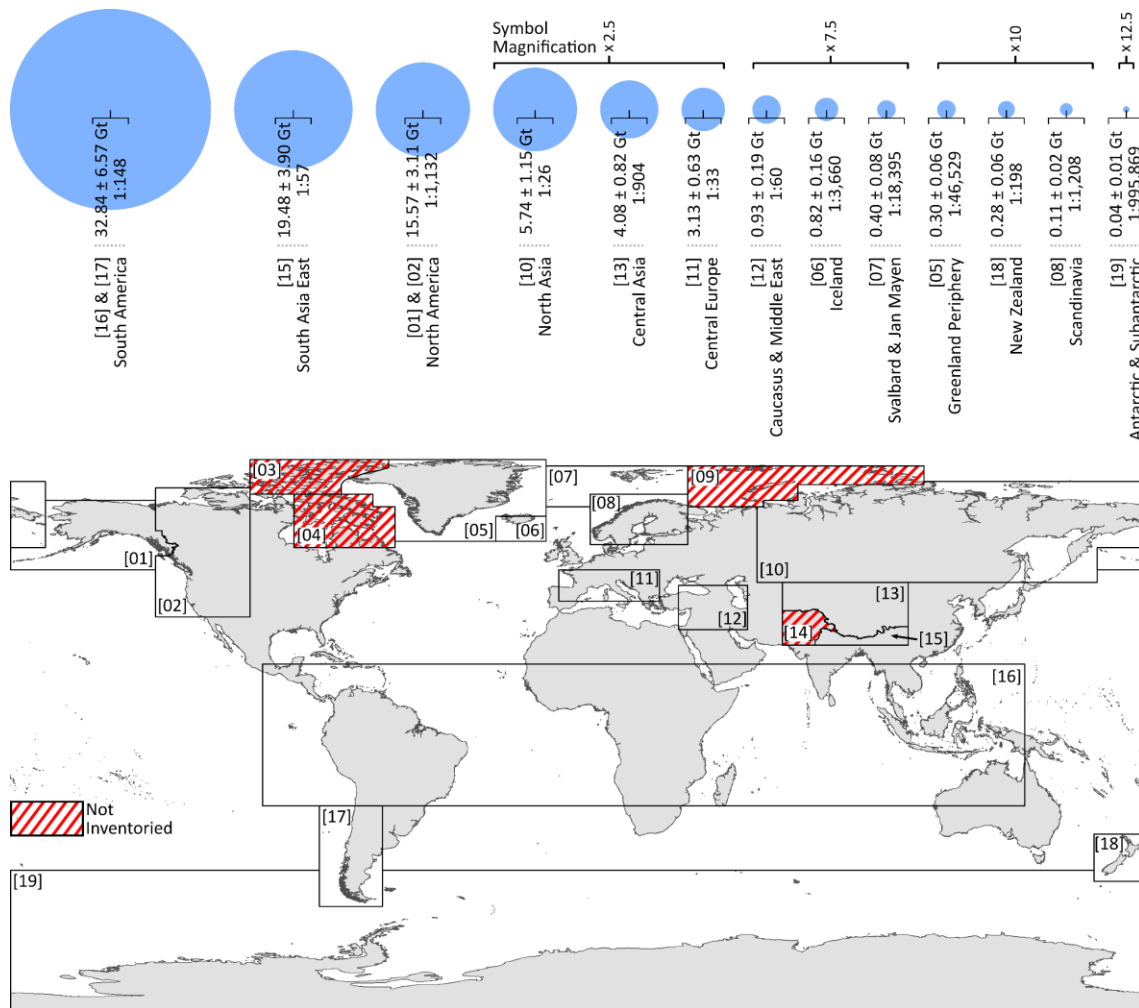418
419   Importantly, however, establishing statistically significant correlations between meteorological variables
420   (e.g., MAAT, snow cover [depth, duration, timing], summer temperature, positive degree days etc.) and rock
421   glacier response mechanisms can prove challenging (Roer et al., 2005; Kellerer-Pirklbauer and Kaufmann,
422   2012; Hartl et al., 2016b; Scotti et al., 2017). As a consequence, it is highly likely that rock glacier kinematic
423   behaviour and surface geometry are determined by a complex combination of geomorphic (e.g., rock glacier
424   internal structure, underlying topography [slope angle, slope convexity]), environmental (e.g., short and
425   long-term temperature change, snow cover – including depth, duration and timing, rainfall events) and
426   paraglacial processes (e.g., landslide and/or rockfall driven debris overload of the rooting zone) (Krainer
427   and He, 2006; Bodin et al., 2009, 2017; Delaloye et al., 2013; Schoeneich et al., 2015; Hartl et al., 2016b;
428   Wirz et al., 2016; Scotti et al., 2017; Buchli et al., 2018; Eriksen et al., 2018). Here, it is important to stress
429   that some rock glaciers yet to exhibit a climatically-forced response have been reported (Potter et al., 1998;
430   Janke, 2005). Nevertheless, in the context of projected climatic warming, clearly, further research is re-
431   quired to critically quantify and assess rock glacier-climate relations and the associated implications for
432   future water resources. To this end, research focused upon the response of rock glaciers to past environ-
433   mental conditions could provide key information for validating projected responses to future climate
434   warming; however, there is currently a paucity of such studies (Sorg et al., 2015).
435

443 **4. Rock glacier distribution and storage**

444 A considerable number of rock glacier inventories have been compiled in various mountain ranges (see
445 Jones et al., 2018a). However, as yet no global-scale (complete) inventory has been compiled, despite being
446 described as the "most pressing need" in rock glacier research (Janke et al., 2013). It should be noted that
447 significant recent research efforts have greatly elaborated inventory coverage in the succeeding years, for
448 example in South America (Falaschi et al., 2014; Rangecroft et al., 2014; Falaschi et al., 2015; Falaschi et al.,
449 2016; Azócar et al., 2017; Barcaza et al., 2017; Esper Angillieri, 2017; García et al., 2017; Janke et al., 2017;
450 IANIGLA-CONICET, 2018; Selley et al., 2018), North America (Charbonneau and Smith, 2018; Munroe,
451 2018), Central Europe (Colucci et al., 2016; Kellerer-Pirklbauer et al., 2016; Roudnitska et al., 2016;
452 Salvador-Franch et al., 2016; Triglav-Čekada et al., 2016; Winkler et al., 2016a; Onaca et al., 2017; Palma et
453 al., 2017; Uxa and Mida, 2017; Fernandes et al., 2018; Popescu, 2018), Asia (incorporating Central Asia,
454 South Asia [East], South Asia [West] and North Asia) (Bolch and Gorbunov, 2014; Schmid et al., 2015; Lytkin
455 and Galanin, 2016; Galanin, 2017; Wang et al., 2017; Jones et al., 2018b; Ran and Liu, 2018; Blöthe et al.,
456 2019), Antarctic and Subantarctic (Rudolph et al., 2018) and New Zealand (Sattler et al., 2016). Recently,
457 Jones et al. (2018a) presented a near-global database containing >73,000 rock glaciers (intact = ~39,500,
458 relict = ~33,500). The authors provide an approximate estimate of the water volume equivalent (WVEQ)
459 stored within these rock glaciers as being 83.7 ± 16.7 Gt, equivalent to ~68–102 trillion litres (Fig. 6). Fur-
460 thermore, excluding the Antarctic and Subantarctic and Greenland Periphery Randolph Glacier Inventory
461 (RGI; Pfeffer et al., 2014) regions, along with RGI regions within which no systematic rock glacier invento-
462 ries have been undertaken, the estimated ratio of rock glacier-to-glacier WVEQ is 1:456 globally, implying
463 that glaciers store a volume of water 456 larger than rock glaciers currently, a ratio that will change under
464 a warming climate due to the differential wasting rates of glaciers vs. rock glaciers, as previously discussed.
465

13

Symbol Magnification ×2.5   ×7.5   ×10   ×12.5

[16] & [17] South America — 32.84 ± 6.57 Gt — 1:148
[15] South Asia East — 19.48 ± 3.90 Gt — 1:57
[01] & [02] North America — 15.57 ± 3.11 Gt — 1:1,132
[10] North Asia — 5.74 ± 1.15 Gt — 1:26
[13] Central Asia — 4.08 ± 0.82 Gt — 1:904
[11] Central Europe — 3.13 ± 0.63 Gt — 1:33
[12] Caucasus & Middle East — 0.93 ± 0.19 Gt — 1:60
[06] Iceland — 0.82 ± 0.16 Gt — 1:3,660
[07] Svalbard & Jan Mayen — 0.40 ± 0.08 Gt — 1:18,395
[05] Greenland Periphery — 0.30 ± 0.06 Gt — 1:46,529
[18] New Zealand — 0.28 ± 0.06 Gt — 1:198
[08] Scandinavia — 0.11 ± 0.02 Gt — 1:1,208
[19] Antarctic & Subantarctic — 0.04 ± 0.01 Gt — 1:995,869

Not Inventoried

Fig. 6. Near-global rock glacier WVEQ (Gt) and ratios of rock glacier-to-glacier WVEQ. Rock glacier WVEQs (blue circles) are sized proportionately to the whole. Rock glacier WVEQs reflect 50 ± 10% ice content by volume. The data are organised into the first-order RGI regions, which are reflected in the accompanying world map. No systematic rock glacier inventory studies have been undertaken in RGI regions *03* (Arctic Canada North), *04* (Arctic Canada South), *09* (Russian Arctic), and *14* (South Asia West). Data are derived from Jones et al. (2018a).

Importantly, near-global- and RGI-regional-scale ratios (Fig. 6) suggest that rock glaciers form considerable water stores; however, arguably the abovementioned ratios mask the potential hydrological value of rock glaciers at national and regional level, particularly in semi-arid and arid zones where glacier presence is limited/absent. For example, rock glacier-to-glacier WVEQ ratios of 3:1 in parts of the Chilean Andes (29°–32° S) suggests that rock glaciers collectively contain more water than glaciers (Azócar and Brenning, 2010). Furthermore, Janke et al. (2017) conclude that rock glaciers constitute ~50% of surface water storage in the Aconcagua River Basin, further south in Chile. Elsewhere, studies in Bolivia estimate rock glacier-to-glacier WVEQ ratios to be 1:7 (Sajama region) (Rangecroft et al., 2015) and as high as 1:3 in the West region of Nepal (Jones et al., 2018b). This emphasises the importance of rock glaciers as non-negligible water stores. Indeed, the recently adopted National Glacier Law of Argentina[2] protects glaciers and rock glaciers as strategic water reserves, prohibiting detrimental human activities (e.g., mining and oil and gas activities) in their vicinity, recognising the potential hydrological value of such features. Nonetheless, although a small number of national-scale glacier inventories include rock glaciers, such as the Inventario Nacional de Glaciares of Argentina (Zalazar et al., 2017; IANIGLA-CONICET, 2018), they are omitted from global-scale glacier databases, e.g., the RGI (Pfeffer et al., 2014) or glacier outlines available from GLIMS (www.glims.org) – it is conceivable that in their absence, rock glaciers could be considered as hydrologically insignificant by the wider scientific community and decisionmakers. In addition, while rock glacier inventory density is high in certain areas (e.g., Central Europe [Jones et al., 2018a]) further inventories are required to better understand the hydrological significance of rock glaciers, particularly where future

14

492   climate warming will likely considerably impact freshwater resources in the long-term (e.g., Andes, Hindu
493   Kush etc.). With this in mind, it is important to provide an overview of the techniques to assess both rock
494   glacier distribution (Section 4.1) and subsurface characteristics (Section 4.2).

495   *4.1. Approaches for assessing rock glacier distribution*

496   Whereas debris-free glaciers can feasibly be mapped/monitored through the application of semi-auto-
497   mated and automated remote sensing classification techniques applied to optical satellite data (e.g., Bolch
498   et al., 2010; Guo et al., 2015), the continuous debris-layer that characterises rock glaciers confounds such
499   approaches. In explanation, rock glacier – and debris-covered glacier – surficial debris and debris from the
500   surrounding slopes share a mutual origin, thus their spectral similarity "render[s] them mutually indistin-
501   guishable" (Shukla et al., 2010) using spectral properties alone. Accordingly, combined methodologies have
502   been developed to identify debris-covered glaciers (Paul et al., 2016), for example using multispectral [op-
503   tical and thermal] satellite data (e.g., Shukla et al., 2010); however, such approaches are generally unsuita-
504   ble for mapping rock glaciers or heavily debris-covered glaciers (e.g., Brenning, 2009). Consequently, rock
505   glacier inventory compilation by means of manual-recognition, -delineation and -classification (i.e. dy-
506   namic behaviour) using representative rock glacier features (Section 2.1; Fig. 1) (e.g., Scotti et al., 2013;
507   Schmid et al., 2015) arguably remains the optimum approach. As such, rock glacier inventories are inher-
508   ently subjective (see Scotti et al., 2013; Schmid et al., 2015; Jones et al., 2018b), influenced as they are by
509   the expertise and field knowledge of the mapper. In particular, this is evident in the case of ongoing glacier-
510   rock glacier interactions (cf. Monnier and Kinnard, 2015b; Section 2.2) where defining the divide between
511   rock glaciers and debris-covered glaciers can prove challenging, especially where rock glaciers are not well-
512   developed (i.e. lacking their characteristic surface features) (e.g., Mölg et al., 2018). These ongoing glacier-
513   rock glacier interactions are commonly not included in either glacier, e.g., the RGI (Pfeffer et al., 2014) or
514   GLIMS, or rock glacier inventories in spite of their potential future hydrological value and we argue their
515   inclusion is of critical importance.
516
517   Of further influence on rock glacier inventory accuracy are the resolution and quality of both the terrain
518   (i.e. digital elevation model and maps) and optical (i.e. aerial or satellite imagery) data used. The increased
519   availability of spatial data with high repeat frequency in recent years has enabled existing rock glacier in-
520   ventories to be updated, so the global picture is becoming more complete (e.g., Kellerer-Pirklbauer et al.,
521   2012; Janke et al., 2017). Additionally, open-source and new types of data have provided enhanced oppor-
522   tunities for rock glacier mapping and monitoring, with: (i) fine spatial resolution optical data (e.g., Quick-
523   Bird, WorldView-1 and 2, IKONOS) accessible freely through Google Earth enabling large-scale rock glacier
524   inventories to be compiled (e.g., Rangecroft et al., 2014; Schmid et al., 2015; Jones et al., 2018b); (ii) finer
525   than 1 m resolution LiDAR (light detection and ranging [i.e. airborne laser scanning]) data enabling heavily
526   vegetated relict rock glaciers, which can significantly influence catchment hydrology (Winkler et al., 2016b)
527   (see Section 5.3), to be mapped (e.g., Colucci et al., 2016); and (iii) InSAR (Interferometric Synthetic Aper-
528   ture Radar) data (e.g., ESA Sentinel-1) enabling both rock glacier mapping and accurate classification of
529   activity type through the investigation of surficial kinematics, such as feature displacement (Kenyi and
530   Kaufmann, 2003; Strozzi et al., 2004; Lambiel et al., 2008; Echelard et al., 2013; Liu et al., 2013; Barboux et
531   al., 2014; Wang et al., 2017; Imaizumi et al., 2018; Villarroel et al., 2018).
532
533   We argue that further research addressing the controls on rock glacier distribution and development (see
534   Johnson et al., 2007; Forte et al., 2016; Kenner and Magnusson, 2017) could further exploit the abovemen-
535   tioned data to identify suitable "habitats" for rock glaciers, thereby encouraging more efficient inventory
536   compilation in rock glacier data-deficient regions. Research efforts with respect to this, for instance, using
537   statistical estimation techniques and generalised additive modelling with terrain/optical data (e.g.,
538   Brenning et al., 2007; Azócar and Brenning, 2010; Brenning and Azócar, 2010), have already been under-
539   taken. Further improvements within this research area may enable semi-automated or automated identi-
540   fication and mapping of rock glaciers over larger spatial-scales in the future, although this would demand
541   the use of new datasets, e.g., time-series of SAR interferometry data for distinguishing sedimentary areas
542   experiencing ground deformation (rock glaciers) vs. static features. Critically, better knowledge of rock
543   glacier distribution and more accurate classification of activity type and temporal dynamics will facilitate
544   an informed assessment of future water supplies with respect to rock glaciers. Furthermore, inventories
545   are time-static, and an ongoing concern should be regular monitoring of rock glaciers using up-to-date sat-
546   ellite data, to determine rates of change and wastage over large extents. With new satellite systems in orbit,
547   such as the Sentinel-1 SAR constellation, we are entering a time where routine regular observations can be
548   used to update inventories (e.g., Villarroel et al., 2018), and as Paul et al. (2016) explain this could

15

549 "considerably improve over previous capabilities, thanks to increased spatial resolution and dynamic
550 range, a wider swath width and more frequent coverage". Rock glacier scientists should also explore the
551 capabilities offered by cloud-based image processing on Google Earth Engine, which could enhance the ef-
552 ficiency with which such large volumes of data can be processed in the future (Gorelick et al., 2017).

553 *4.2. Approaches for assessing rock glacier subsurface characteristics*

554 *4.2.1. Internal structure*
555 Information concerning the number, spatial distribution and morphometric characteristics of rock glaciers
556 is useful; nevertheless, complementary subsurface data (i.e. internal structure: origin, thickness and pro-
557 portional ice content, hydrological flowpaths) are required to better understand the hydrological role of
558 rock glaciers. Knowledge of the subsurface properties of rock glaciers is derived predominantly from geo-
559 physical techniques such as ground-penetrating radar (GPR), electrical resistivity tomography (ERT) and
560 seismic refraction tomography (for reviews see: Scott et al., 1990; Maurer and Huack, 2007; Kneisel et al.,
561 2008; Hauck, 2013). Geophysical investigations suggest that rock glacier internal structure comprises four
562 general layers: (i) a debris mantle (i.e. AL) with a thickness in the order of meters; (ii) an ice-rich (i.e. high
563 ice content [vol.%]) frozen layer a few tens of metres thick; an ice-poor (i.e. low ice content) or ice-free
564 sediments underlying the ice-rich layer; and (iv) the surface of the bedrock. For example, the Ölgrube and
565 Kaiserberg rock glaciers (Ötztal Alps, Austria) each comprise a 4–6 m thick AL that is underlain by ice-rich
566 permafrost 20–30 m in thickness, then 10–15 m of ice-free sediments (Hausmann et al., 2012). The above-
567 described subsurface composition has been confirmed through geophysical investigations at a number of
568 other rock glaciers (e.g., Isaksen et al., 2000; Fabrot et al., 2005; Hausmann et al., 2007; Bodin et al., 2009).
569 Furthermore, information from boreholes drilled through the well-studied Murtèl-Corvatsch and Muragl
570 rock glaciers (Haeberli et al., 1998; Arenson et al., 2002) also support the general rock glacier internal
571 structure as earlier described (Kääb, 2013). Importantly, rock glacier subsurface characteristics (e.g., AL
572 thickness, multiple ice origins [i.e. freezing of sub- and supra-groundwater, seasonal snow accumulation,
573 avalanching and buried glacial ice], volumetric ice content, permafrost table topography, rock debris abun-
574 dance, etc.) within and between rock glaciers can be highly heterogeneous (Haeberli et al., 2006) (Fig. 7).
575 Two cores drilled on the Lazaun rock glacier in the southern Ötztal Alps, Italy, further demonstrate the
576 horizontal and vertical heterogeneity of rock glacier subsurface and ice content (Krainer et al., 2015; Fig.
577 8). This spatial heterogeneity is also reported at the Las Liebres rock glacier (Chilean Andes, Chile) where
578 ice content markedly varies along a longitudinal GPR profile (22–83%) (Monnier and Kinnard, 2015a).
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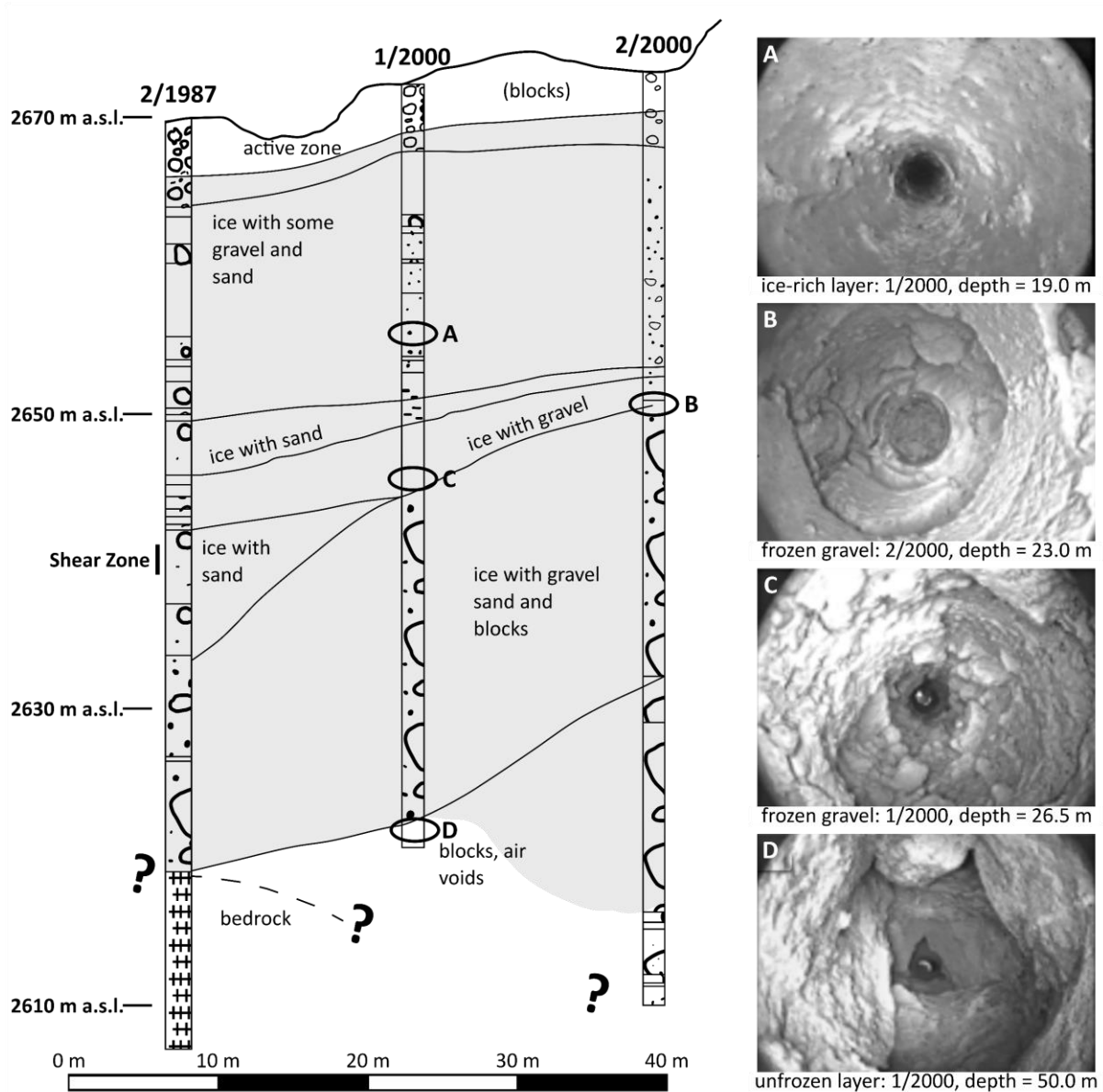
Fig. 7. Detailed stratigraphy of the Murtèl-Corvatsch rock glacier through three boreholes (BH 2/1987, BH 1/2000 and BH 2/2000) (Arenson et al., 2002). Photographs A-D from a borehole camera reflect the stratigraphy of the rock glacier at 19.0 m, 23.0 m, 26.5 m and 50.0 m depths. These subsurface investigations demonstrate the heterogeneity of rock glaciers, with zones of massive and interstitial ice present. Modified after Springman et al. (2012).
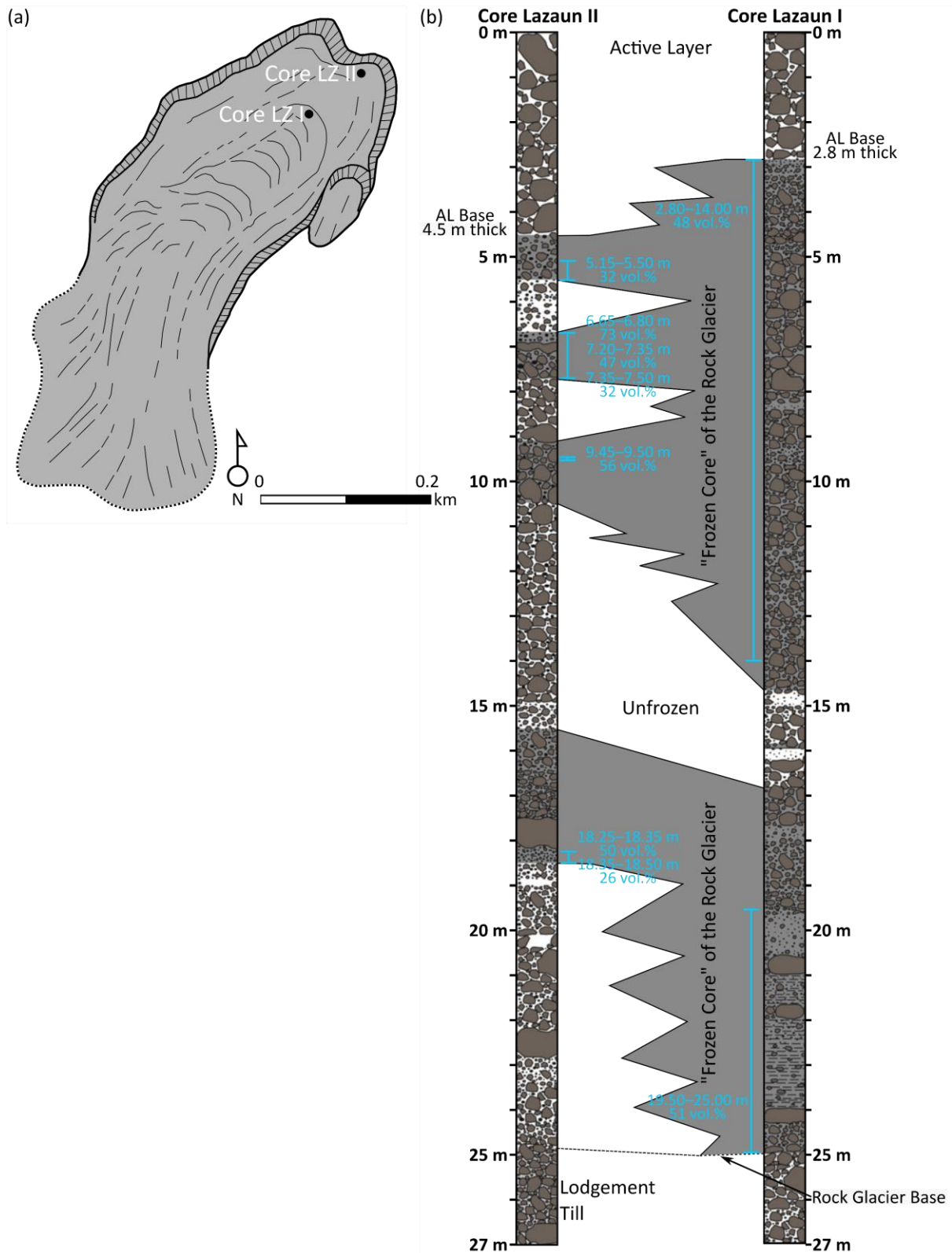
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Fig. 8. (a) Map of the Lazaun rock glacier including the locations of core Lazaun I and II; and (b) the rock glacier internal structure from core Lazaun I and II. The average ice content of core Lazaun I is 43 vol.%, varying between 0–98 vol.% and core Lazaun II is 22 vol.%, varying between <2–73%. Examples of higher ice content within the cores are shown. Note the presence of intra-permafrost taliks (particularly in core Lazaun II), the formation of which has been attributed to advective and conductive heating by infiltrating water and circulating air (Luethi et al., 2017). Modified after Krainer et al. (2015).

While significant progress has been made regarding scientific knowledge of the interior characteristics of rock glaciers, such understanding is derived from a relatively small number of features due to the

18

596 formidable logistical challenges of fieldwork in such environments (e.g., government restrictions on sam-
597 pling and/or equipment use, field location accessibility, presence of large boulders and megaclasts [i.e. b-
598 axis >4.1 m] at the surface [Fig. 4b]) that would require considerable expense and time to overcome, par-
599 ticularly drilling (Degenhardt Jr and Giardino, 2003; Janke et al., 2013). Indeed, very few rock glaciers have
600 been drilled (see Table 3 in Haeberli et al., 2006). Therefore, although non- or minimally invasive geophysi-
601 cal techniques can rapidly provide 2-D subsurface data over extended survey areas, in contrast to 1-D in-
602 formation provided by drilling (Haeberli et al., 2010), geophysical investigations have rarely been validated
603 by direct measurements (Degenhardt Jr and Giardino, 2003; Hauck and Kneisel, 2008). As a consequence,
604 where detailed information on the density and composition of rock glaciers are unavailable and geophysi-
605 cal data quality is variable, significant uncertainties in the interpretation of geophysical data can occur due
606 to ambiguities in relating subsurface data to material properties within the rock glacier (e.g., differentiating
607 between rock and ice) (Hauck et al., 2011; Hauck, 2013).

609 In response, innovative, alternative methodologies have been developed; for example, at the Äußeres
610 Hochebenkar rock glacier (Ötztal Alps, Austria), feature thickness was calculated using a simple creep
611 model based on high-resolution surface displacement and slope data derived from multitemporal digital
612 elevation models. Comparison of model results and GPR data proved helpful in fine-tuning the analysis of
613 the latter (Hartl et al., 2016a). Also, where multitemporal high-resolution digital elevation models are avail-
614 able, rock glacier thickness and volume could be estimated with this method (Hartl et al., 2016a). As the
615 modelling approach of Hartl et al. (2016a) accounts for individual rock glacier-specificities (i.e. rock glacier
616 dynamics and local topography), estimated rock glacier thickness and volume will be more accurate than
617 those generated using more general approaches, such as empirical power-law relations (Section 4.2.2).

619 Of further note, improved geophysical techniques will provide opportunities to improve understanding of
620 rock glacier internal structure and ice content. For example, helicopter-mounted quasi-3-D GPR was re-
621 cently successfully used to investigate the internal structure of Furggwanghorn rock glacier, Swiss Alps
622 (Merz et al., 2015a; Merz et al., 2015b), overcoming some of the logistical challenges associated with de-
623 tailed *in situ* geophysical surveys. Emmert and Kneisel (2017) found that quasi-3-D electrical resistivity
624 tomography enables the mapping and monitoring of spatial variations within the rock glacier internal
625 structure, with considerable potential to deliver new insights into surface and subsurface process-linkages.
626 Additionally, the geophysically based 4-phase model (see: Hauck et al., 2011) combines electrical resistivity
627 tomography and refraction seismic tomography measurements to obtain volumetric proportions and char-
628 acterise the lateral and vertical distribution of the rock glacier constituents: air, water, ice, and rock. The
629 model has already been successfully applied to rock glaciers for this purpose (e.g., Hauck et al., 2011;
630 Schneider et al., 2013). Furthermore, Mewes et al. (2017) demonstrate the suitability of the 4-phase model
631 for detecting interannual and seasonal phase change between liquid water and ice in rock glaciers; there-
632 fore, this model is a valuable tool for monitoring degradation of the internal ice body (Section 3). These
633 new techniques will provide a platform to better understand the internal structure, dynamics and response
634 of rock glaciers to future warming. In particular, more detailed geophysical investigations may enable vol-
635 ume and WVEQ estimation for glacier-rock glacier transitional forms (Section 2.2), for which simplified
636 geomorphometric approaches (Section 4.2.2) are not suitable due to the internal complexity of such fea-
637 tures (Bolch et al., 2019).

638 *4.2.2. Volume and WVEQ*
639 Since detailed subsurface information is available for only a limited number of rock glaciers, 2-D-area-re-
640 lated statistics have enabled first-order approximations of thickness and volume of unmeasured rock glac-
641 iers. In particular, *empirical thickness-area (H-S) relations* have been applied to predict rock glacier thick-
642 nesses and derive rock glacier volume (e.g., Brenning, 2005a; Azócar and Brenning, 2010; Bodin et al., 2010;
643 Perucca and Esper Angillieri, 2011; Rangecroft et al., 2015; Janke et al., 2017; Jones et al., 2018a, 2018b).
644 Empirical H-S relations can be expressed as $\bar{h} = c \cdot S^{\beta}$, where mean rock glacier thickness $\bar{h}$ (m) is calcu-
645 lated as a function of surface area S (km^2) and a scaling parameter c (50) and scaling exponent β (0.2)
646 (Brenning, 2005a). Rock glacier volumes were determined by $V = \bar{h} \cdot S$. Rock glacier WVEQ was subse-
647 quently estimated through the multiplication of V and estimated ice content (% by vol.) and assuming an
648 ice density conversion factor of 900 kg m^{-3} (Paterson, 1994).

650 To our knowledge, no study has evaluated empirical power-law relations (i.e. H-S and V-S [volume-scaling]
651 scaling relations) for their suitability to approximate rock glacier thickness and volume. Empirical power-
652 law relations have been evaluated in relation to glaciers (see Frey et al., 2014; Bahr et al., 2015), upon which
653 they have more commonly been used to predict regional- and global-scale glacier thickness and volume

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(e.g., Chen and Ohmura, 1990; Bahr et al., 1997; Radić and Hock, 2010; Grinsted, 2013; Andreassen et al., 2015). Frey et al. (2014) demonstrate that considerable uncertainty is associated with the application of power-law relationships for this purpose. It is reasonable to assume that shortcomings experienced when applying these approaches to glaciers and rock glaciers are similar. Regarding V-A scaling relations (i.e. $V = c \cdot S^\gamma$ [where γ is $\beta+1$]) for glaciers, the scaling exponent γ is considered a constant (Bahr et al., 1997; Bahr et al., 2015) whereas the scaling parameter c should be considered as a variable determined by several parameters that vary between glaciers (e.g., basal topography, sliding parameters, ice-flow parameters) (Bahr et al., 2015). Further, several researchers have suggested different, region-specific values for both c and γ from regression of the available data (see Grinsted, 2013). Brenning's (2005a) empirical power-law relation was developed from morphometric field measurements at 19 rock glaciers in the Andes of Santiago, Chile (Bodin et al., 2010). As such, this approach cannot account for regional specificities of rock glaciers around the world (e.g., dominant ice origin, local climatic parameters, local topography, etc.) (Jones et al., 2018a) that could reasonably be expected to influence the value of scaling parameter c, as described above. To date, a constant scaling parameter c and scaling exponent β have thus been used in rock glacier studies applying Brenning's (2005a) empirical power-law relation (e.g., Brenning, 2005a; Azócar and Brenning, 2010; Bodin et al., 2010; Perucca and Esper Angillieri, 2011; Rangecroft et al., 2015; Janke et al., 2017; Jones et al., 2018a, 2018b). Therefore, use of such methods with respect to rock glaciers requires careful and critical evaluation. To improve the effectiveness of empirical power-law relations for rock glaciers requires that (i) the physics (i.e. dynamics) of rock glaciers are better understood, (ii) the sample size used to constrain the scaling parameter c and choose the value for β needs to be increased, and (iii) the scaling parameter c is localised.

Previous studies predicting rock glacier WVEQ have assumed a 'typical' volumetric rock glacier ice content of 50% (Brenning, 2005b; Azócar and Brenning, 2010; Perucca and Esper Angillieri, 2011) – WVEQ values for 40–60% vol. (i.e. lower [40%], mean [50%] and upper [60%] estimates) are sometimes provided to account for uncertainty (Brenning, 2005a; Bodin et al., 2010; Rangecroft et al., 2015; Jones et al., 2018a; Jones et al., 2018b) – consistent with *in situ* data derived from different climatic regions worldwide (e.g., Elconin and LaChapelle, 1997: >50%; Arenson et al., 2002: 40–70%; Croce and Milana, 2002: ~55%; Hausmann et al., 2007: 45-60%; Hausmann et al., 2012: 40–60%). Importantly, studies that have used Brenning's (2005a) empirical power-law relation – excluding Janke et al. (2017) – have considered intact rock glaciers (i.e. active and inactive rock glaciers treated collectively). Yet, Arenson and Jakob (2010) suggest that the application of the abovementioned statistical model should better differentiate between the volumetric ice content of active and inactive rock glaciers. Indeed, a classification system for debris-covered glaciers and rock glaciers divides the latter into three subclasses: (i) "Class 4: Proper" (i.e. active), ice content = 25–45%; (ii) "Class 5: Transitional" (i.e. inactive), ice content = 10–25%; and (iii) "Class 6: Glacier of rock" (i.e. relict), ice content = <10% (Janke et al., 2015). As a consequence, large uncertainties will be introduced to rock glacier WVEQ predictions by the typically lower volumetric ice content of inactive rock glaciers (Brenning, 2010). In addition, existing rock glacier WVEQ estimations have not considered volumetric air content, reported to be up to 25% in intact rock glaciers (Arenson and Springman, 2005), that would further reduce predicted WVEQ (Arenson and Jakob, 2010). Nonetheless, existing rock glacier WVEQ estimations provide a much-needed first approximation that should stimulate further research and gain the attention of policymakers.

5. Rock glacier water discharge

5.1. Characteristics of rock glacier discharge

To date, only very few studies have investigated the hydrological aspects of rock glacier water discharge, in spite of recent research reporting that rock glaciers constitute non-negligible long-term water stores (Section 4). Besides the above-stated formidable logistical challenges of rock glacier-related fieldwork (Section 4.2.1), this is because water discharge measurements are challenging or virtually impossible to conduct for rock glaciers (i) with multiple, often inaccessible springs (Krainer et al., 2012), (ii) with no spring(s), i.e. the water drains within the debris (Krainer and Mostler, 2002), (iii) that grade into downslope landforms, and (iv) that terminate in lakes or ponds (Colombo et al., 2018c). Based on the few available measurements, the discharge from springs of intact rock glaciers is estimated to range from <1 to >1000 L s^{-1} during the melt season (Table 2).

Table 2. Discharge from intact rock glaciers during the melt season.

20

Rock Glacier Name	Location	Discharge (L s^{-1})	Reference
Gruben	Swiss Alps, Switzerland	2–50	(Haeberli, 1985)
Hilda	Rocky Mts., Alberta, Canada	90–270*	(Gardner and Bajewsky, 1987)
Dos Lenguas	Central Andes, Argentina	5–8	(Schrott, 1991, 1996)
East Slims River	St. Elias Mts., Yukon, Canada	0.7–3.7	(Harris et al., 1994)
Morenas Coloradas	Central Andes, Argentina	230–1000	(Trombotto et al., 1997)
Reichenkar	Stubai Alps, Austria	20–375	(Krainer and Mostler, 2002)
Gößnitz	Hohe Tauern Mts., Austria	5–310	(Krainer and Mostler, 2002)
Innere Ölgrube	Ötztal Alps, Austria	30–1000	(Berger et al., 2004)
Kaiserberg	Ötztal Alps, Austria	25–600	(Krainer et al., 2007)
Reichenkar	Stubai Alps, Austria	20–450	(Krainer et al., 2007)
Mt. Tukuhnikivatz	La Sal Mts., Utah, USA	0.1–25.2	(Geiger et al., 2014)
Lazaun	Southern Ötztal Alps, Italy	9–140	(Krainer et al., 2015)
Äußeres Hochebenkar	Ötztal Alps, Austria	10–300	(Nickus et al., 2015)

* The Hilda rock glacier has been classified as an inactive rock glacier (Luckman and Crockett, 1978). Furthermore, (Harris et al., 1994) note that the rock glacier spring "is actually a groundwater spring mantled by the rock glacier", and thus the results may not reflect only rock glacier discharge.

Intact rock glacier discharge patterns are characterised by strong seasonal and diurnal variability, primarily determined by local weather conditions, the thermal conditions within the AL, and the physical mechanisms that control meltwater flow through the rock glacier (Krainer and Mostler, 2002; Berger et al., 2004; Krainer et al., 2007; Krainer et al., 2015). The water released from intact rock glaciers is derived from (i) melting of the winter snowpack, (ii) melting of glacier ice, (iii) melting of rock glacier ice, (iv) rainfall, particularly during summer thunderstorms with heavy rainfall, and (v) groundwater (Krainer and Mostler, 2002). The contribution to intact rock glacier discharge of the abovementioned water sources changes considerably during the melt season. Typically, discharge rates are highest during the spring/early summer snowmelt and gradually decline through summer and autumn to low or zero flow in the winter months. In addition, rock glacier discharge fluctuates strongly in response to rainfall events and periods of colder weather with snowfall (Johnson, 1981; Blumstengel and Harris, 1988; Krainer and Mostler, 2002; Berger et al., 2004; Krainer et al., 2007) (Fig. 9). Further, intense melting of the winter snowpack during periods of warm weather produce marked diurnal runoff cycles, with low discharge at noon and peak discharge towards late evening (Krainer and Mostler, 2002). In this respect, generally, rock glacier discharge patterns mimic those of glaciers, although at considerably lower magnitude (Krainer and Mostler, 2002; Geiger et al., 2014). Additionally, various authors suggest that rock glacier discharge fluctuates less than glaciers (Potter, 1972; Corte, 1976, 1987; Gardner and Bajewsky, 1987), at least on diurnal timescales (Haeberli, 1985; Pourrier et al., 2014); however, few have demonstrated this empirically.

Isotopic analyses of rock glacier springs, alongside discharge and electrical conductivity sampling (EC) data, support the above-described seasonal evolution of intact rock glacier discharge composition (Krainer and Mostler, 2002; Berger et al., 2004; Williams et al., 2006; Krainer et al., 2007). For example, Krainer et al. (2007) report that rock glacier discharge in the Stubai and Ötztal Alps, Austria, exhibits: (i) intermediate δ^{18}O values at the beginning of the melt season, reflecting the mixed origin of runoff from re-frozen meltwater from the preceding autumn (high δ^{18}O) and snowmelt (low δ^{18}O); (ii) significantly reduced δ^{18}O and EC values following the onset of spring discharge, with runoff predominantly derived from snowmelt; and (iii) gradually increasing δ^{18}O and EC values during the melt season, which reflect the continued depletion of the winter snowpack and increased icemelt (i.e. melting of the frozen rock glacier core) and groundwater contributions. Intermittent sharp peaks in rock glacier discharge, δ^{18}O values and concomitant depressions of EC suggest that outflows temporarily derive predominantly from summer rainfall events (which have higher δ^{18}O values and are EC-depleted) (Krainer et al., 2007; Fig. 9).

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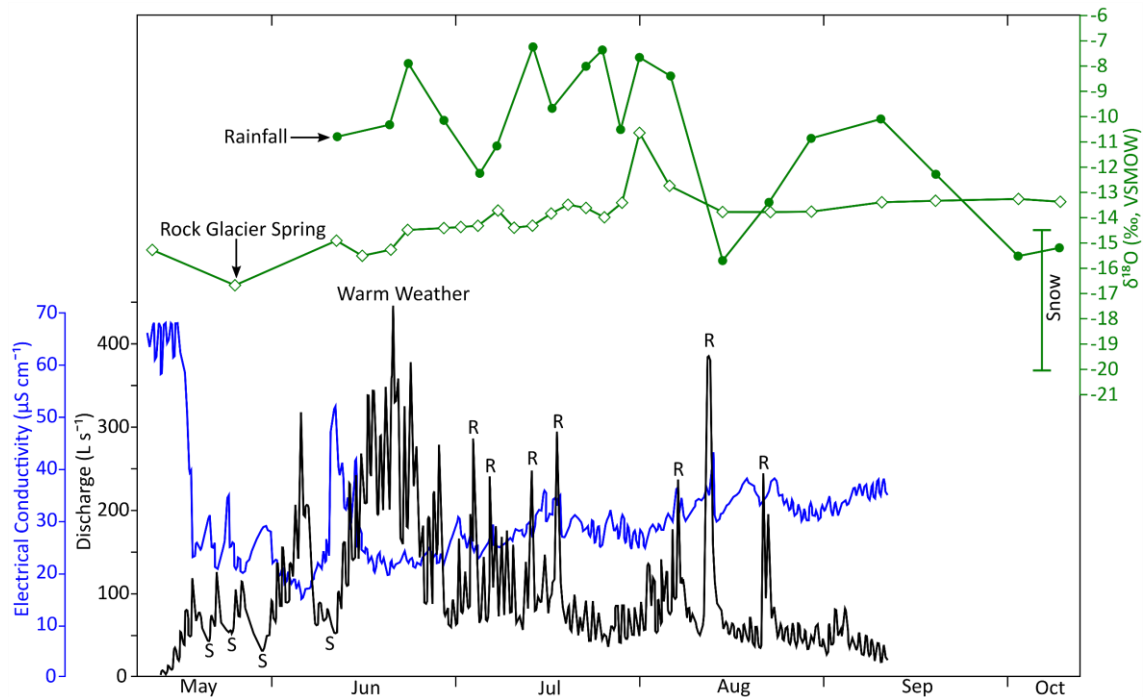
Fig. 9. Hydrograph of the meltwater stream at the Reichenkar rock glacier and EC between mid-May and mid-September 2002. Major snowfall (S) and rainfall (R) events are indicated. Oxygen isotope composition ($\delta^{18}O_{water}$) of the rock glacier spring and rain precipitation is also depicted for the period from mid-May to mid-October 2002. The snow profile was sampled in April 2002. Of note, $\delta^{18}O$ values of the frozen rock glacier core (not depicted) are similar to those of the rock glacier spring. Modified after Krainer et al. (2007).

Of note, previous studies have primarily investigated the temporal evolution of $\delta^{18}O$ in conjunction with discharge and EC data to ascertain the origins of rock glacier outflows. Yet hitherto very few attempts have been made to determine the proportional contribution to rock glacier discharge of different sources. Williams et al. (2006) analysed the outflow components of the Green Lake 5 rock glacier (Colorado Front Range, USA) through geochemical and stable isotopic analyses in combination with end-member mixing analysis: 30% was snowmelt, 32% was soil water, and 38% was baseflow – the latter includes icemelt. Notably, studies have not successfully isolated the contribution of icemelt to rock glacier discharge (Krainer et al., 2007); however, seasonal discharge, $\delta^{18}O$ and EC data, as prior discussed, suggest water released from rock glaciers principally derives from snowmelt, glacial meltwater and intermittent rainfall events, with negligible or non-measurable contributions derived from icemelt (Croce and Milana, 2002; Krainer and Mostler, 2002; Krainer et al., 2007). Minimal icemelt contributions are predominantly due to the insulative effects of the AL, which prevents substantial melt rates (Section 3). For instance, Krainer et al. (2015) estimate that melting of the frozen rock glacier core contributed ~0.6 L s^{-1} or ~2.3% of the total annual mean discharge of ~26 L s^{-1} during the melt season at the Lazaun rock glacier. At the Helen Creek rock glacier (Banff National Park, Alberta, Canada) – an inactive feature – minimal icemelt contributions (3–5% [July-August]) support these findings (Harrington et al., 2018). Elsewhere, using radioisotope analysis, Cecil et al. (1998) compared tritium concentrations in rock glacier outflows (9.2–13.2 TU [Tritium Units]) to those in an ice core at the same location (-1.3–0.2 TU). They conclude that frozen rock glacier core formation pre-dates the peak of above-ground nuclear weapons testing (1950s/1960s), whereas rock glacier outflows primarily derive from modern precipitation (i.e. minor contributions from icemelt).

Notably, rock glacier-catchment hydrology interactions remain poorly understood and opinions regarding the hydrological contributions of rock glaciers diverge. Indeed, a number of studies have suggested that rock glacier hydrological contributions to downstream runoff are significant; however, the majority of these conclusions are based upon non-quantitative data (for summary see: Duguay et al., 2015). It is important to note that investigations focused upon rock glacier hydrology generally consider present as opposed to future rock glacier hydrological contributions. Therefore, thus far the hydrological significance of rock glaciers has been defined according to a restricted timescale. Critically, while rock glaciers form long-term water stores and thus may not constitute a readily available water resource (Duguay et al., 2015; Rangecroft et al., 2015), at decadal and longer timescales, under climate warming, degradation of the frozen

22

782   rock glacier core may represent an increasing hydrological contribution to streamflow (Thies et al., 2013;
783   Geiger et al., 2014).

784   *5.2. Characterisation of rock glacier hydrological flowpaths*

785   Importantly, as well as rock glacier hydrological contributions, rock glacier-catchment hydrology interac-
786   tions should be considered with respect to discharge timing. In addition to the seasonal availability of dif-
787   ferent water sources (Section 5.1), rock glacier discharge timing is determined by spatially- and tempo-
788   rally-variable, convoluted subsurface flowpaths (Burger et al., 1999). Based upon the following infor-
789   mation, a model of the hydrological flowpaths within rock glaciers of periglacial and glacial origin was con-
790   structed (Fig. 10).
791
792   Conceptual models of rock glacier hydrology suggest that subsurface water movement occurs along two
793   flowpath types: *supra-permafrost flow* and *sub-permafrost flow* atop and below an impermeable frozen rock
794   glacier core, respectively (Giardino et al., 1992). This is consistent with the conclusions of Krainer and
795   Mostler (2002) who identified two storage reservoirs, the "quickflow reservoir" (herein *quickflow*) and
796   "baseflow reservoir" (herein *baseflow*), which correspond to supra-permafrost flow and sub-permafrost
797   flow, respectively. Rock glacier hydrographs, dye tracer tests and EC values demonstrate that quickflow
798   and baseflow have significantly different characteristics of water storage and release (Krainer and Mostler,
799   2002). These characteristics are likely determined by the AL thickness (i.e. depth of the 0 °C isotherm),
800   hydraulic connectivity, sedimentological characteristics, and the degree of rock glacier activity (i.e. volu-
801   metric ice content [% by vol.]) (Tenthorey, 1992; Harris et al., 1994; Krainer and Mostler, 2002; Williams
802   et al., 2006; Buchli et al., 2013).
803
804   Dye tracer test results reported in Tenthorey (1992) and Krainer and Mostler (2002) indicate flow veloci-
805   ties of 54–327 m h^{-1}, demonstrating that the coarse-blocky openwork structure of the AL promotes
806   quickflow. Elsewhere, the rapid temporal evolution of rock glacier discharge in conjunction with δ^{18}O and
807   EC values suggests that waters derived from high-intensity weather events (i.e. intense snowmelt or sum-
808   mer rainfall events) are generally transmitted through active rock glaciers as quickflow with a lag time of
809   a few hours (Krainer et al., 2007; Fig. 9). These results also show that quickflow occurs predominantly as
810   channelised flow along a network of conduits eroded into the frozen rock glacier core (Krainer and Mostler,
811   2002). Occasionally, quickflow is visible and/or audible in the AL of intact rock glaciers (Tenthorey, 1992;
812   Krainer and Mostler, 2002; Berger et al., 2004; Rogger et al., 2017).
813
814   Baseflow contributions are likely to form a relatively small proportion of water discharge and are charac-
815   terised by high EC values, indicating that baseflow predominantly derives from slow diffuse groundwater
816   flow transiting through unfrozen, fine-grained material at the base of the rock glacier (e.g., Krainer and
817   Mostler, 2002). Investigations of the internal structure of active rock glaciers confirm the presence of ice-
818   free, fine-grained sediments at the base of rock glaciers (Section 4.2.1), supporting the presence of the
819   baseflow reservoir. Water tracer tests indicate that baseflow exhibits highly variable residence times, rang-
820   ing from several days to several months (e.g., Tenthorey, 1992; Harris et al., 1994). Following depletion of
821   the snow cover, rock glacier hydrographs and EC data demonstrate that baseflow contributions increase
822   (e.g., Krainer and Mostler, 2002).
823
824   A further subsurface flowpath in rock glaciers with low ice content, *intra-flow* (i.e. slow internal flow), was
825   identified through dye-tracing experiments (Tenthorey, 1992, 1994). Intra-flow also includes water trans-
826   iting through active rock glaciers by means of unfrozen drainage networks, i.e. *intra-permafrost flow*
827   (Tenthorey, 1992, 1994; Krainer and Mostler, 2002). Intra-permafrost flow has been reported at depth in
828   the frontal portions of active rock glaciers (e.g., Arenson et al., 2010). In explanation, Ikeda et al. (2008)
829   suggested that networks of air voids and fractures develop within the frozen rock glacier core as a result of
830   fast deformation rates, based upon data from the Büz North rock glacier (Swiss Alps), and thus increased
831   intra-permafrost flow throughout the landform. This has significant implications for rock glacier creep ve-
832   locities and stability, including via the reduction of effective stress by infiltrated water (Krainer and
833   Mostler, 2006; Perruchoud and Delaloye, 2007; Ikeda et al., 2008); however, these implications are not
834   discussed here. Typically, intra-flow is released from the rock glacier as baseflow.
835
836   Importantly, geophysical methodologies, such as the 4-phase model (Section 4.2.1), provide opportunities
837   to estimate the liquid water content within the frozen rock glacier core and to identify preferential

23

838   subsurface flowpaths (Mewes et al., 2017, and references therein). Limitations and required improvements
839   within the 4-phase model are detailed in Hauck et al. (2011) and Mewes et al. (2017).
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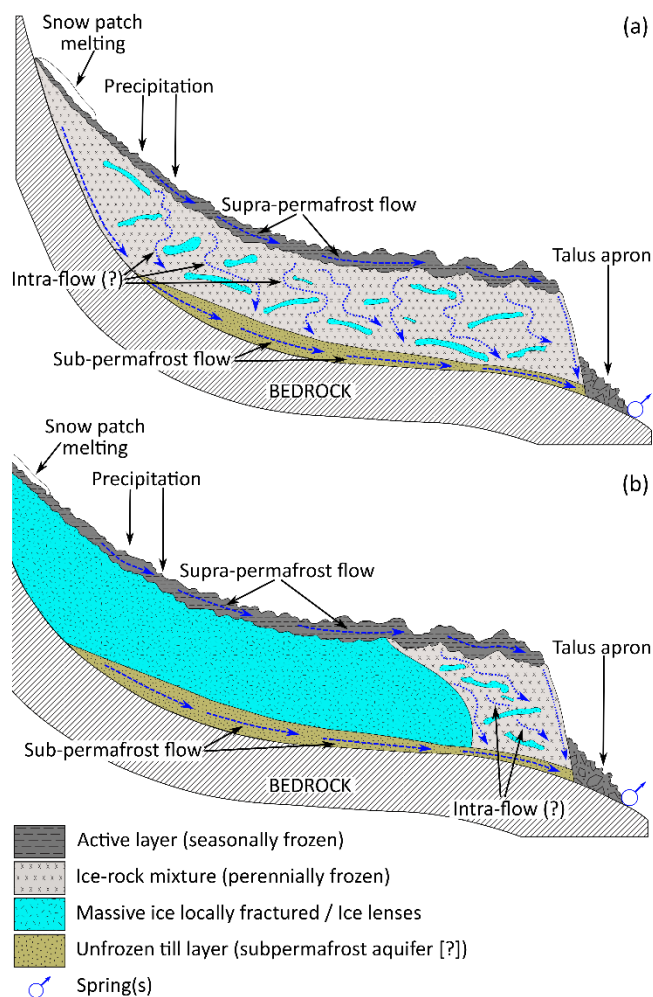
841
842   **Fig. 10.** A simplified hypothetical model of hydrological flowpaths through an active rock glacier of: (a) periglacial
843   origin; and (b) glacial origin. N.B. hydrological flowpath terminology was maintained for [a] and [b] for consistency (i.e.
844   supra-*permafrost*, intra-*permafrost*, and sub-*permafrost* flow), regardless of the dominant ice origin. Arrows depict the
845   flowpath direction.

846   *5.3. Rock glacier-catchment hydrology interactions*

847   Very few systematic studies of catchment-scale geomorphic drivers of streamflow regimes (i.e. timing, mag-
848   nitude and duration of discharge) have conducted comparative assessments of the hydrological response
849   between alpine catchments. Physical catchment parameters, including slope, elevation range, drainage
850   area, and bedrock geology, have previously been used assess inter- and intra-catchment variations in the
851   hydrological response. Contrastingly, the hydrological influence of discrete debris accumulations (DDAs)
852   and/or ice-debris landforms (I-DLs), e.g., talus slopes, protalus ramparts, protalus lobes, and rock glaciers,
853   are largely neglected (Weekes et al., 2015). N.B. ground ice exists within I-DLs. Yet, recent research demon-
854   strates that DDAs and I-DLs can modulate the hydrologic response (e.g., Clow et al., 2003; Geiger et al.,
855   2014; Weekes et al., 2015; Rogger et al., 2017).
856
857   In the La Sal Mountains, Utah, a comparative assessment of an [active] rock glaciated vs. non-rock glaciated
858   catchment indicates that the presence of active rock glaciers had a pronounced influence upon catchment
859   hydrology (Geiger et al., 2014). Hydrographs from the rock glaciated (i) and non-rock glaciated catchments
860   (ii) are distinctly different. (i) In this catchment, flood peaks are delayed following the onset of precipita-
861   tion, flood peaks are higher, and the proportion of precipitation leaving the catchment as stormflow is high.
862   (ii) Contrastingly, in this catchment, flood peaks occur much more quickly and are short in duration, and
863   the proportion of precipitation leaving the catchment as stormflow is significantly lower than in the rock

24

glaciated catchment (see Table 2 in Geiger et al., 2014). This indicates that in non-rock glaciated catchments, significant losses of precipitation to deep groundwater storage may occur. Importantly, rock glacier hydrographs have shown that rock glacier stormflow constitutes a significant proportion of total catchment runoff (15–30%). It appears that the proportion of rock glacier stormflow to catchment stormflow is determined by precipitation intensity rather than precipitation volume. For example, 37.6 mm of precipitation (1.21 mm h^{-1}) and 7.2 mm of precipitation (2.40 mm h^{-1}) derived from two different storms resulted in a 15% and 30% contribution to total catchment runoff, respectively. The above-described observations indicate that active rock glaciers act as impervious surfaces following high-intensity weather events, with the net effect of increasing runoff generation within rock glaciated catchments (Geiger et al., 2014). In the Krummgampen catchment, western Ötztal Alps, Austria, model simulations indicate that complete disappearance of permafrost will reduce flood peaks by ~17%, supporting the above-described findings (Rogger et al., 2017).

Notably, the characteristics of rock glacier water storage and release (i.e. storage behaviour) differ considerably between active, inactive, and relict features. Indeed, lessening of the degree of activity through degradation of the frozen rock glacier core may potentially increase feature porosity, and thus increase the storage capacity. In turn, this will cause changes in the rock glacier discharge pattern, and consequently influence runoff generation in alpine catchments (Rogger et al., 2017). In response, several scientific studies have considered rock glacier-catchment hydrology interactions in relation to relict features (e.g., Wagner et al., 2016; Winkler et al., 2016b; Rogger et al., 2017). Globally, relict rock glaciers are numerous (Section 4) and under future climate warming intact rock glaciers will transition towards relict activity status, and thus from quickflow to baseflow dominated landforms. In explanation, the coarse-blocky open-work structure of rock glaciers promotes rapid infiltration and transmission of water inputs to the groundwater system – i.e. relict rock glaciers constitute unconfined aquifers (Geiger et al., 2014). For example, research in the Niedere Tauern Range (Eastern Alps, Austria) has shown that subsequent to precipitation events, relict rock glaciers rapidly (within hours) contribute ~20% precipitation volume to discharge; the remaining ~80% is delayed considerably with a mean residence time of ~7 months. This large storage component facilitates large baseflow rates long after precipitation events (Winkler et al., 2016b). As a consequence, relict rock glacier water stores can maintain streamflow during summer baseflow periods (e.g., Wagner et al., 2016). In addition, a large proportion of the terrain >2000 m a.s.l. drains through rock glaciers (e.g., Seckauer Tauern Range = 51% [Winkler et al., 2016a]); therefore, it is reasonable to assume that relict rock glaciers constitute significant water stores at intermediate-term timescales. Notably, to date studies assessing rock glacier WVEQ (Section 4.2.2) have not included relict rock glacier water stores; however, including these stores would introduce large uncertainties into WVEQ estimations, likely due to the difficulty of estimating such transient stores. Nevertheless, it is clear that this rock glacier type strongly influences both the water storage capabilities and discharge behaviour of alpine catchments.

Investigations of rock glacier-catchment hydrology interactions that consider large glacier-rock glacier composite features are sparse. At the Tapado catchment (Coquimbo region, Chile) – an assemblage of cascading cryospheric landforms (debris-free glacier, debris-covered glacier, rock glacier and moraine complexes) – Pourrier et al. (2014) found that the storage capacity and transmissive function of rock glaciers is considerably different to debris-covered glaciers. Pourrier et al. (2014) present a conceptual model that synthesises the hypothesised hydrological functioning of this glacier-rock glacier composite feature (Fig. 11). Analysis of the flow dynamics shows rapid and concentrated or slow and diffuse hydrological transfers for the debris-covered glacier and rock glacier, respectively. In this study, the debris-covered glacier forms a weakly capacitive but highly transmissive medium, whereas the rock glacier forms a highly capacitive but weakly transmissive medium. The hydrological data suggest that rock glaciers exhibit: (i) a strong buffering effect on the daily-to-monthly variability of transferring glacier meltwater to downstream areas; and (ii) a high storage capacity that partially delays glacier meltwater transfer to downstream areas (Pourrier et al., 2014). Therefore, catchment hydrology will be significantly influenced by glacier-rock glacier interactions, particularly where debris-covered glaciers are *currently developing* into rock glaciers. As a consequence of the transition from glacial to paraglacial process regimes, glacier-to-rock glacier transition in high mountain systems is increasingly likely (Section 2.2); therefore, it is important to better understand the catchment hydrology implications of glacier-rock glacier interactions.

25
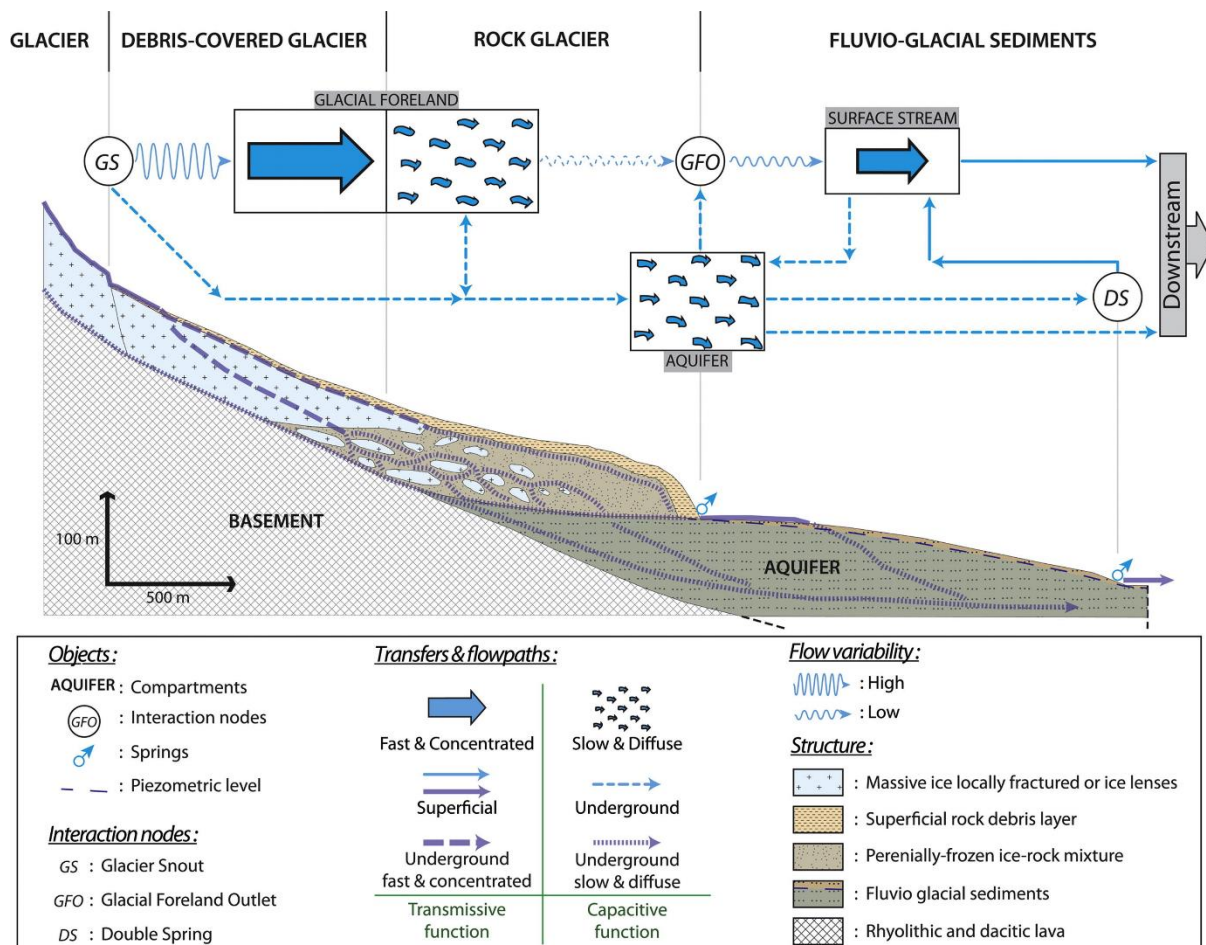
Fig. 11. Schematic transect through the Tapado catchment synthesising the hypothesised hydrological interactions between the catchment components (i.e. debris-free glacier, debris-covered glacier, rock glacier, and fluvioglacial sediments). Figure after Pourrier et al. (2014).

6. Rock glacier water hydrochemistry

Rock glacier water hydrochemistry is the focus of very few scientific investigations (see Table 1 in Colombo et al., 2018b). It is known that rock glaciers lose ice volume at slower rates than glaciers (Section 3), therefore the former potentially affect water hydrochemistry over longer timescales (Fegel et al., 2016). Indeed, work has shown that rock glacier thaw modifies the inorganic chemistry of both water bodies located downstream (Thies et al., 2007; Baron et al., 2009; Ilyashuk et al., 2014; Colombo et al., 2018a; Ilyashuk et al., 2018) and streams (Williams et al., 2006; Thies et al., 2013; Nickus et al., 2015; Fegel et al., 2016; Munroe, 2018). As a consequence, the water quality characteristics of rock glacier outflows should be a major focus of research, given their potential as potable water resources (Burger et al., 1999).

Previously, rock glacier outflow has been described as *clear* (i.e. predominantly sediment free), containing comparatively lower *suspended sediment concentrations* (SSC) and higher *total dissolved solids* (TDS) relative to glacier-fed outflow (Gardner and Bajewsky, 1987). In addition, *total load* (i.e. SSC + TDS [excluding bedload transport]) within the latter are at least one order of magnitude greater than that within rock glacier outflow (Gardner and Bajewsky, 1987). It is important to note that these conclusions are based on a small sample size, with just a small number of studies evidencing this. For example, SSC measurements from the Hilda rock glacier (Canadian Rocky Mountains, Canada) were generally low, ranging between 1–3 mg L^{-1} and reaching 20 mg L^{-1} only in response to precipitation events. As expected, increased rock glacier-derived discharge causes increases in SSC; however, a threshold value of ~150 L s^{-1} was reported after which SSC declines (Gardner and Bajewsky, 1987). SSC measurements reported for the Boundary glacier, situated ~1 km north of the Hilda rock glacier, ranged between 70–4000 mg L^{-1} ($\bar{X}$ = 600–800 mg L^{-1}) during the same precipitation event (Gardner and Bajewsky, 1987), illustrating the potential filtering effect of rock glaciers on fine-grained sediment throughput. SSC measurements from the Reichenkar rock glacier

946 (western Stubai Alps, Austria) reflect clear outflow associated with discharges <100 L s^{-1}, but SSC of 1000
947 mg L^{-1} recorded at peak discharges of >300 L s^{-1} (Krainer and Mostler, 2002); SSC-discharge relationships
948 are therefore unclear. TDS measurements of rock glacier outflow are generally low, although they have
949 been shown to be significantly greater than those of glacier and snowmelt derived outflow. For instance,
950 TDS values of 60–75 mg L^{-1} and 30–40 mg L^{-1} were reported for Hilda rock glacier and Boundary glacier
951 outflow, respectively (Gardner and Bajewsky, 1987).
952
953 Rock glaciers characteristically have a higher debris fraction versus glaciers. Consequently, it is hypothe-
954 sised that significant mineral surface area-ice contact promotes chemical weathering and leads to solute-
955 enrichment of water contained within rock glaciers (Ilyashuk et al., 2014, 2018). This hypothesis is con-
956 sistent with the results from Giardino et al. (1992) who reported that rock glacier outputs in the San Juan
957 Mountains (Colorado, USA) had significantly higher TDS than rock glacier inputs. Rock glacier input pH
958 levels (6.4–6.9) and output pH levels (7.3–8.4), further illustrate the solute-concentrating effect of rock
959 glaciers (Ilyashuk et al., 2014). Giardino et al. (1992) recorded similar pH values for rock glacier inputs
960 (6.4–7.0) and outputs (7.2–8.5) at three features in the Blanca Massif area of the Sangre de Cristo Mountain
961 Range, Colorado, USA. Given that rock glacier thaw resulting from climate change could drive the export of
962 enriched-solute fluxes, the analysis of EC, major ions (e.g., Ca^{2+}, Mg^{2+}, SO_4^{2-}, NO_3^-) and trace elements (e.g.,
963 Ni, Mn, Al, Hg, Pb) in rock glacier outflow has received increasing attention (for review see: Colombo et al.,
964 2018b).
965
966 In the American West (i.e. Cascade Mountains, Rocky Mountains, Sierra Nevada), Fegel et al. (2016) report
967 that rock glacier outflow samples exhibit higher pH and EC values and are also enriched in *weathering*
968 *products* (i.e. SiO_2, Ca^{2+}, K^+, Mg^{2+} and Sr^{2+}) relative to glaciers. Similarly, Crespo et al. (2017) report higher
969 EC values for rock glacier-derived waters ($\bar{x}$ = 1201 µS cm^{-1}) compared to waters from debris-free glaciers
970 ($\bar{x}$ = 32 µS cm^{-1}), snow-dominated streams ($\bar{x}$ = 159 µS cm^{-1}) and debris-covered glaciers ($\bar{x}$ = 805 µS cm^{-1})
971 in the Upper Mendoza River basin (Argentinian Andes). In addition, clear differences in the $\delta^{18}O$ signatures
972 of glacier- (debris-free and debris-covered) vs. rock glacier-derived waters are reported; depleted values
973 (i.e. lower amount of heavy isotopes) and enriched values (i.e. higher abundance of heavy isotopes) are
974 found in glacier- and rock glacier-derived waters, respectively (ibid.). To date, research directly comparing
975 rock glacier- and glacier-derived meltwaters through hydrochemical and stable isotope analysis are few in
976 number; most consider the hydrochemistry of rock glacier outflows over various timescales or relative to
977 surface waters uninfluenced by rock glaciers (not explicitly of glacial origin). Long-term (1985–2005) in-
978 creases in EC (by ~19-fold), Ca^{2+} (by ~13-fold), SO_4^{2-} (by ~26-fold) and Mg^{2+} (by ~68-fold) have been re-
979 ported for the Rasass See alpine lake located beneath an active rock glacier in the Central Eastern Alps, Italy
980 (Thies et al., 2007; Fig. 12). Increased enriched-solute fluxes related to rock glacier thaw are reflected in
981 the strong seasonal variations found in outflow hydrochemistry. For instance, strong seasonal increases in
982 SO_4^{2-} (by ~175-fold), Mg^{2+} (by ~30-fold), Ca^{2+} (by ~20-fold) and Na^+ (by ~4-fold) were recorded for Green
983 Lake 5 rock glacier (Colorado Front Range, Rocky Mountains, USA) outflows in late summer/autumn com-
984 pared with other surface waters during mid-summer (Williams et al., 2006). Such findings have been at-
985 tributed to enhanced rock glacier thaw due to climate warming (Thies et al., 2007; Thies et al., 2013).
986
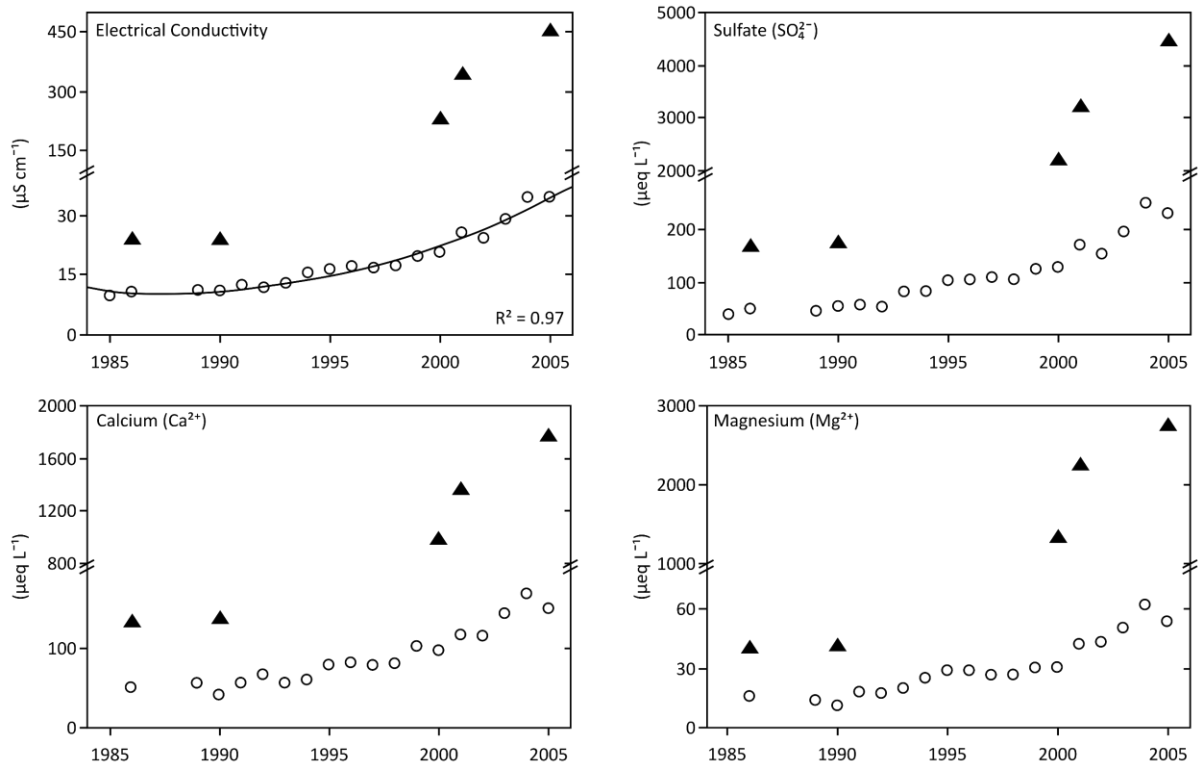
987
988 **Fig. 12.** Long-term EC, sulfate (SO_4^{2-}), Calcium (Ca^{2+}) and magnesium (Mg^{2+}) concentrations in Rasass See (black trian-
989 gles) and Schwarzsee ob Sölden (open circles) alpine lakes (1985–2005). Values for Rasass See and Schwarzsee ob
990 Sölden represent mean values of four to seven discrete samples along the lake vertical profile taken during holomixis
991 (i.e. compete mixing of the lake). Variability among single values is <5%. Diagonal lines on the y-axis depict scale breaks.
992 Modified after Thies et al. (2007).

993
994 Elevated concentrations of trace elements in high-altitude alpine lakes, also attributed to the influence of
995 rock glacier thaw (Thies et al., 2007), have been reported. For instance, natural *acid rock drainage* (ARD)
996 can arise in mineralised catchments with sulfide-bearing lithologies following rock glacier thaw, as the ex-
997 posure of sulfide-rich rocks to air and oxygenated water results in increased oxidation of sulfide minerals
998 (Ilyashuk et al., 2014). ARD typically produces acid-sulfate waters enriched in metals; for example, high
999 concentrations of Mn, Ni and Al in excess of EU maximum permissible levels in drinking water were rec-
1000 orded at Rasass See. Contrastingly, negligible concentrations were reported for the adjacent water body
1001 without a rock glacier in its catchment (Ilyashuk et al., 2014). In mineralised catchments with sulfide-bear-
1002 ing lithologies, under future climate warming, it is reasonable to expect enhanced ARD in response to in-
1003 creased AL thickness which permits atmospheric oxygen to penetrate to greater depths (Colombo et al.,
1004 2018b). Additionally, high-altitude mining operations may adversely affect rock glacier water quality as
1005 mining tailings – a source of ARD (Bellisario et al., 2013) – are piled onto rock glacier surfaces (Brenning,
1006 2008; Bellisario et al., 2013). Of note, the origin(s) of high concentrations of trace elements in rock glacier
1007 ice often remains the subject of considerable uncertainty (Colombo et al., 2018b) and requires further
1008 study. For instance, high heavy metal concentrations (Ni, Zn, Co, Cu, Fe and Mn) were determined in ice-
1009 cores drilled at Lazaun rock glacier, yet lithological analyses indicate the origin of these heavy metals is not
1010 the rocks of the catchment area of the rock glacier (Krainer, 2014; Krainer et al., 2015).

1011
1012 As described above, localised releases of enriched-solute fluxes following rock glacier thaw can alter sig-
1013 nificantly the inorganic chemistry of water surfaces downstream. Further, increased export of enriched-
1014 solute fluxes can be expected under future climate warming. Therefore, an improved understanding of the
1015 meteorological drivers (i.e. air temperature, snowmelt and rainfall) responsible for the export of solute-
1016 enriched waters from rock glaciers is important for (i) determining causality for the hydrochemical re-
1017 sponse of water surfaces downstream and (ii) anticipating, monitoring and planning for future changes in
1018 downstream hydrochemistry (Colombo et al., 2018a). Three hypotheses are identified (Colombo et al.,
1019 2018b): (i) warmer air temperatures enhance rock glacier thaw, releasing solute-enriched waters (e.g.,
1020 Krainer and Mostler, 2002; Berger et al., 2004; Williams et al., 2006; Krainer et al., 2007; Thies et al., 2013;
1021 Nickus et al., 2015; Colombo et al., 2018a); (ii) long-lasting snow cover delays rock glacier thaw thus soluble
1022 materials remain sequestered in rock glacier ice, while snowmelt also dilutes solute-enriched waters

28

1023 (Williams et al., 2006). Similarly, rainfall percolation through rock glaciers lowers solute concentrations in
1024 outflows (e.g., Krainer and Mostler, 2002; Berger et al., 2004; Krainer et al., 2007; Nickus et al., 2015); and
1025 (iii) rainwater percolating through rock glaciers flushes solute-enriched waters, particularly following
1026 snow cover depletion (e.g., Colombo et al., 2018a), in contrast to the effects described in hypothesis (ii).
1027 This lack of agreement likely stems from the limited availability of studies focused on meteorological vari-
1028 ables, hydrologic processes and chemical characteristics of water surfaces downstream (Colombo et al.,
1029 2018a). Further, existing studies have predominantly been undertaken in the European Alps or North
1030 American mountain ranges (Colombo et al., 2018b). As such, it is evident that further scientific investigation
1031 is required, particularly an extension of research evidencing the suitability of water surfaces downstream
1032 of rock glaciers for use as safe, potable water sources to other regions worldwide.

1033 **7. Conclusions**

1034 In this contribution, we synthesise the available literature and present the first comprehensive evaluation
1035 of the hydrological role of rock glaciers over a range of spatial and temporal scales, considering globally-
1036 distributed published studies. Of note, rock glacier-related research has advanced considerably post-1970s,
1037 yet important gaps remain regarding the scientific knowledge and understanding of rock glacier hydrology.
1038 A general lack of consensus concerning the present and future hydrological significance of rock glaciers
1039 pervades the literature.
1040
1041 Importantly, water storage within rock glaciers occurs over a range of timescales; yet, to date we identify
1042 that investigations focused upon rock glacier hydrological significance generally consider present as op-
1043 posed to potential future rock glacier icemelt contributions – rock glacier hydrological significance has
1044 been defined according to a restricted timescale. A near-global first-order approximation of intact rock
1045 glacier WVEQ (water volume equivalent) concludes that intact rock glaciers potentially constitute hydro-
1046 logically valuable long-term stores (see Section 4). Given that rock glaciers are climatically more resilient
1047 than glaciers (see Section 3), we hypothesise that the relative importance of rock glacier hydrological con-
1048 tributions will increase as climate warming proceeds through the twenty-first century, particularly in semi-
1049 arid and arid regions. Furthermore, with continued climate-driven deglaciation and thus a transition from
1050 glacial- to paraglacial-dominated process regimes, glacier-to-rock glacier transition is likely to become in-
1051 creasingly common (see Section 2). This may potentially enhance the resilience of the mountain cryosphere
1052 to future climate warming.
1053
1054 Thus far, rock glacier-related research has primarily defined the hydrological significance of rock glaciers
1055 according to the ice content/WVEQ and/or the proportional contribution to rock glacier discharge of
1056 icemelt. However, for the first time, we have synthesised published work to consider rock glacier "hydro-
1057 logical significance" as encompassing: (i) rock glacier-catchment hydrology interactions (i.e. total discharge
1058 volume, variability, and timing); and (ii) rock glacier effects upon the physical characteristics (i.e. hydro-
1059 chemistry, temperature, etc.) of water inputs. Regarding (i), hydrographs from [active] rock glaciated and
1060 non-rock glaciated catchments have been reported to be distinctly different, and relict rock glaciers – reg-
1061 ularly overlooked in rock glacier hydrology research – can also strongly influence alpine catchment hydrol-
1062 ogy (see Section 5). In regard to (ii), hydrochemical analysis of rock glacier outflows indicates that rock
1063 glaciers may adversely affect the inorganic chemistry of water bodies located downstream and streams
1064 (see Section 6).
1065
1066 In undertaking this detailed synthesis, we have identified candidate areas for future rock glacier hydrology
1067 research, which are currently absent or require further development in the literature. In general, there
1068 exists a relative paucity of quantitative data that continues to restrict scientific assessment of rock glacier
1069 hydrological significance, presumably owing to the formidable logistical challenges of rock glacier-related
1070 fieldwork globally (see Section 4). Importantly, we identify that quantitative data are needed from a range
1071 of locations and timescales. In particular, we would suggest that there is an urgent requirement for studies
1072 to deliver data describing the long-term characteristics of rock glaciers, capturing the full diversity of mor-
1073 phometry and environmental settings of these features. This will require efforts within the research com-
1074 munity to tackle sites that are less easily accessible since much of the existing research has been conducted
1075 to date on the most accessible sites (e.g., within the European Alps), which may provide a limited viewpoint.
1076 We call for further research into the following:
1077

29

- *Rock glacier distribution:* Considerable research efforts have significantly elaborated rock glacier inventory coverage; analysis of inventory data, including rock glacier number, distribution, and morphometric characteristics, forms the first step in understanding their hydrological role. However, rock glacier inventories are absent in many climatically vulnerable regions (e.g., Andes, Hindu Kush, etc.), and relatively few studies share accessible open-access databases (Jones et al., 2018a). Therefore, to enable improved inventory coverage we suggest: (i) existing inventories are made accessible; (ii) habitat suitability models are further developed, thereby encouraging more efficient inventory compilation in rock glacier data-deficient regions; (iii) use of open-access platforms, such as Google Earth; (iv) updated remote sensing data (see Section 4.1) is used to re-evaluate existing inventories. The recently established IPA Action Group (2018–2020), the primary aims of which are "to sustain the first steps toward the organization and the management of a network dedicated to rock glacier mapping (inventorying) and monitoring all around the world and the definition of the necessary standards" support these suggestions (Delaloye et al., 2018).

- *Rock glacier evolution:* It is hypothesised that glacier-to-rock glacier transitions are critically important in the context of future water resources management. Further work is needed to elucidate the drivers behind which glaciers will fully transition into rock glaciers and those which will simply downwaste. This requires a future research focus on *currently developing* features. Furthermore, *currently developing* features are commonly not included in either glacier or rock glacier inventories. To assess their potential future hydrological value efforts to inventory glacier-rock glacier interactions (e.g., Bolch et al., 2019) need to be undertaken.

- *Rock glacier ice content/WVEQ:* Currently, an empirical power-law relationship ($\bar{h} = c \cdot S^{\beta}$) is predominantly used to evaluate rock glacier ice content and thus WVEQ, but there are few empirical studies against which to test the rigour of this relationship (see Section 4). We call for new experiments that: (i) enable the physics (i.e. dynamics) of rock glaciers to be better understood, (ii) increase the sample size used to constrain the scaling parameter c and choose the value for β, and (iii) localise the scaling parameter c. We also call for ground-truthing data that will deliver the evidence-based science to test and review the suitability of the current empirical power-law relationship. Furthermore, an approach to estimate the ice content/WVEQ of glacier-rock glacier interactions, for which simple empirical power-law relationships are likely inappropriate (Bolch et al., 2019), needs to be developed.

- *Rock glacier storage and release:* There is some understanding about the short-term volume, variability, and timing of rock glacier outflows, derived from studies at a few, largely intact sites. As climate change progresses, rock glaciers will evolve, and there is a need for data describing how the transition towards relict activity status will impact discharge. This requires a focus on *space for time substitution* experiments across the intact-to-relict transition. In addition, geophysical methodologies, such as the 4-phase model (see Section 4), provide opportunities to better understand the internal structure of rock glaciers, and thus boost knowledge about subsurface hydrological processes (i.e. preferential subsurface flowpaths). We recognise that delivering empirical data describing rock glacier internal structure is difficult but encourage rock glacier scientists to engage with new methodological/technological approaches to sample rock glacier internal structure.

- *Rock glacier hydrochemistry:* Based upon a limited number of studies, there is evidence that intact rock glaciers can adversely change the inorganic chemistry of water bodies and streams, downstream of outflows. Further scientific investigation of this is required, to extend research evidencing the suitability of water originating from rock glaciers for use as safe, potable water resources. Importantly, future studies should assess waters originating from relict rock glaciers, as hydrological storage capacity and residence time is greater than that of intact features (Colombo et al., 2018b). We note that to include *hydrochemical sampling* in the experimental design of future rock glacier studies would advance this research field.

- *Rock glacier climatic resilience:* A number of studies conclude that complex internal thermal regimes are generated within the coarse-blocky openwork structure of the AL, amplifying the insulation effect of the debris mantle. However, in a changing climate intact rock glaciers could experience drastic change. Indeed, emerging observations of increased rock glacier surface velocities potentially in response to recent climate warming has been reported (see Section 3). In spite of this, research considering the effects of future climate warming on rock glaciers is in its infancy. Improved modelling of the thermal regime (i.e. heat transport) to study the future evolution of rock glaciers under different climate change scenarios (e.g., Pruessner et al., 2018) is critically needed. Further, this will better inform the long-term hydrological role of rock glaciers.

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