

1 Introducing a new normalized cryospheric index (NCI) to categorize 2 sub-watersheds on arid environments

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7 **Abstract.** This study examines sub-watersheds in the arid northern region of Chile (26°41'–28°24'S), situated within the
8 broader Copiapó watershed. The primary water source for this watershed originates from cryospheric landforms. The region
9 exhibits pronounced physiographic and climatic diversity, with its northern sector situated within the South American Arid
10 Diagonal (SAAD), where cryospheric features display greater spatial isolation. The aim of this study is to quantify the water
11 volume contributed by distinct cryospheric landforms to regional watersheds. This study employs a classification methodology
12 to categorize cryospheric landforms within sub-watersheds, integrating an inventory of cryospheric landforms, historical snow
13 permanence derived from satellite imagery, published ice content and depth measurements (ERT and GPR). The Normalized
14 Cryospheric Index (NCI) is calculated for varying hydrological conditions and provides a means to compare potential water
15 volumes across sub-watersheds. The analysis reveals significant spatial variability in cryospheric landforms and their strategic
16 hydrological significance. Under average and low-precipitation conditions, the southern sub-watersheds of the Copiapó
17 watershed show the greatest water storage potential. The Montosa River (NCI = 0.93), Del Potro River (NCI = 0.89), Manflas
18 River (NCI = 0.47), Estero Come Caballos (NCI = 0.34), and Cachitos River (NCI = 0.31) sub-watersheds have been identified
19 as strategic priority areas within the region for sustaining surface runoff and safeguarding water availability. During high-
20 snowfall periods, northern sub-watersheds in the Copiapo watershed, such as Estero Come Caballos, show elevated NCI values
21 despite their limited cryospheric landforms. In contrast, the Montosa, Manflas, and Del Potro sub-watersheds contain the most
22 extensive cryospheric landforms and rank among the top of the sub-watersheds with the highest NCI scores.

23 1 Introduction

24 Anthropogenic climate change impacts water availability, particularly in regions dependent on high-altitude water sources
25 within mountain systems (Beniston & Stoffel, 2014, Beniston et al., 2018). Changes in the cryosphere that will occur within
26 the next century include a change from solid to liquid precipitation, declining volumes of cryospheric landforms, snow lines
27 moving to higher altitudes, and shorter snow seasons. With these changes, the predictions are that water resources in mountain
28 areas will become increasingly scarce. In the arid zones of the Chilean Atacama Desert glaciers continue to shrink (Masiokas,

et al., 2020) and centennial reconstructions of rainfall patterns on the scale of hundreds of years, show a decline in rainfall in the Central Zone and the “Norte Chico” natural regions of Chile (Le Quesne et al., 2006; Minvielle and Garreaud, 2011; Morales et al., 2012; Bozkurt et al., 2017). In addition, the occurrence of mega droughts has intensified since 2010 (Garreaud et al., 2020).

Within this climatic context, the Copiapó River exhibits a mixed hydrological regime with very low discharges, typically ranging between $\sim 1.49 \text{ m}^3 \text{ s}^{-1}$ and $1.82 \text{ m}^3 \text{ s}^{-1}$ at the La Puerta gauging station, and strongly modulated by contributions from its main tributaries, the Manflas, Jorquera and Pulido Rivers, whose headwaters originate in high-altitude cryospheric environments. Long-term analyses indicate that mean annual flows rarely exceed $2 \text{ m}^3 \text{ s}^{-1}$, and the watershed shows recurrent years of negative water balance, leading to progressive aquifer depletion (Suárez et al., 2014). These persistently low flows and the strong dependence on upstream snow and ice melt highlight the structural vulnerability of the runoff of the Copiapó River to changes in the cryosphere.

Because of the high altitude of the Andean cryosphere, precipitation is captured from air masses, predominantly in the form of snow. Generally, snow precipitation occurs during the winter (June–September), although the northern parts also receive snow during summer due to the Tropical Monsoon (i.e. Bolivian winter), which takes place from December to February (Valdivieso et al., 2022). In high-altitude Andean environments, snowfall does not directly translate into liquid runoff but may be preserved over different timescales through distinct cryospheric processes. Under favorable mass-balance conditions, repeated snow accumulation and firnification lead to the formation of debris-free glaciers. Where glacier ice becomes mantled by rockfall or talus material, debris-covered glaciers can develop, reducing ablation rates and enhancing ice preservation. In parallel, a significant fraction of snowfall infiltrates coarse debris slopes, where it refreezes and is preserved as ground ice within ice-rich permafrost landforms such as rock glaciers, protalus lobes, and gelifluction taluses. The melting of these cryospheric systems generates a sustained discharge of water, even during snow-free dry seasons. Given the extremely limited rainfall in the Andes, streamflow in the Copiapó River is strongly dependent on snowmelt and cryospheric water storage. (Masiokas et al., 2019). However, the contribution of cryospheric water storage to runoff varies substantially among cryospheric landforms types and occurs over contrasting temporal scales. Debris-free glaciers represent the most direct and dynamic cryospheric contribution to river discharge, responding seasonally to atmospheric forcing and sustaining spring–summer runoff through glacier melt. Debris-covered glaciers may also contribute meltwater, but their hydrological response is attenuated and delayed due to the insulating effect of supraglacial debris, which reduces melt rates per unit area. The interannual variability of this meltwater contribution is strongly modulated by debris cover thickness and continuity, which can substantially alter glacier hydrological response across seasons and years (Burger et al., 2019).

In contrast, ice-rich permafrost landforms, such as rock glaciers, protalus lobes, and gelifluction taluses, function predominantly as long-term cryospheric water reserves rather than active meltwater sources. Recent studies in the Central Andes indicate that many rock glaciers exhibit near-zero mass loss over recent decades, suggesting that internal ice melt currently contributes only marginally to river runoff (Ferri et al., 2020). Their primary hydrological significance therefore lies in their capacity to store frozen water over decadal to centennial timescales, potentially buffering future water availability

under continued climatic warming. Multi-decadal photogrammetric reconstructions using aerial, satellite, and UAV data have further revealed significant kinematic variability in rock glaciers of the Dry Andes, highlighting their dynamic nature and importance as cryospheric components (Vivero et al., 2021). In addition, future megadrought projections for the Southern Andes suggest that glacier meltwater contributions may decline markedly, underscoring the critical need to understand and protect all cryospheric water reserves, including periglacial landforms (Ayala, 2025).

Seasonal snow represents a distinct cryospheric component with a rapid hydrological response. During wet years, snowmelt dominates river discharge, whereas during dry years runoff becomes increasingly dependent on long-term cryospheric storage at high elevations (Ohlanders et al., 2013). Although snow sublimation losses in the Andes can reach up to ~70% of the seasonal snowpack (Jara et al., 2021), the remaining snowmelt still exerts a first-order control on interannual streamflow variability.

The source of water in the watersheds of the Atacama Desert originate thus predominantly from the volumes of ice that are kept in cryospheric landforms, but the potential volumes of water that are present in these cryospheric landforms are highly variable and depend on both the type and the spatial distribution. We hypothesize that the hydrological relevance of arid sub-watersheds can be quantified by integrating long-term ice storage and seasonal snow permanence into a single normalized index, and that the resulting sub-watershed ranking remains robust under realistic measurement and modelling uncertainty. Testing this hypothesis requires (a) a complete cryospheric inventory at the watershed scale, (b) physics-based ice-volume estimates for both glacial and ice-rich permafrost landforms, and (c) explicit propagation of uncertainty through the index calculation. General methodologies to classify and prioritize watersheds are based on morphometric analysis (Rahaman et al., 2015), watershed hydrology (Wolfe, et al., 2019) and ecological and water quality factors. There are also numerous studies covering watershed health in which protection frameworks have been developed based on ecological and hydrological parameters providing approaches on watershed prioritization (Jaiswal et al., 2015; Ahn & Kim, 2017; Sriyana, 2019; Basuki et al., 2022). Few studies incorporate all cryospheric components into watershed management or conservation strategies. Peng et al. (2022) developed an integrated index to evaluate cryospheric changes across the Northern Hemisphere, emphasizing the influence of cryospheric dynamics on regional hydrology. Nonetheless, there is currently a lack of research addressing cryospheric landforms and snow within arid watershed contexts.

To address this gap, this study delivers three integrated outputs for the Copiapó watershed: (i) an updated, GIS-ready inventory of 1862 cryospheric landforms covering glaciers, debris-covered glaciers, rock glaciers, gelifluction taluses, and protalus lobes; (ii) a unified physics-based framework for estimating ice volume across glacial and ice-rich permafrost landforms, combining the Farinotti et al. (2019) consensus product with viscoplastic creep inversion (Cicoira et al., 2021) constrained by Sentinel-1 DInSAR velocities; and (iii) a Normalized Cryospheric Index (NCI) that ranks sub-watersheds by their combined long-term ice storage and seasonal snow contribution, with Monte Carlo propagation of uncertainty to ensure ranking robustness. We apply this framework to the 12 cryospheric sub-watersheds of the Copiapó watershed and identify those whose protection is critical for sustaining downstream water availability under continued climatic warming.

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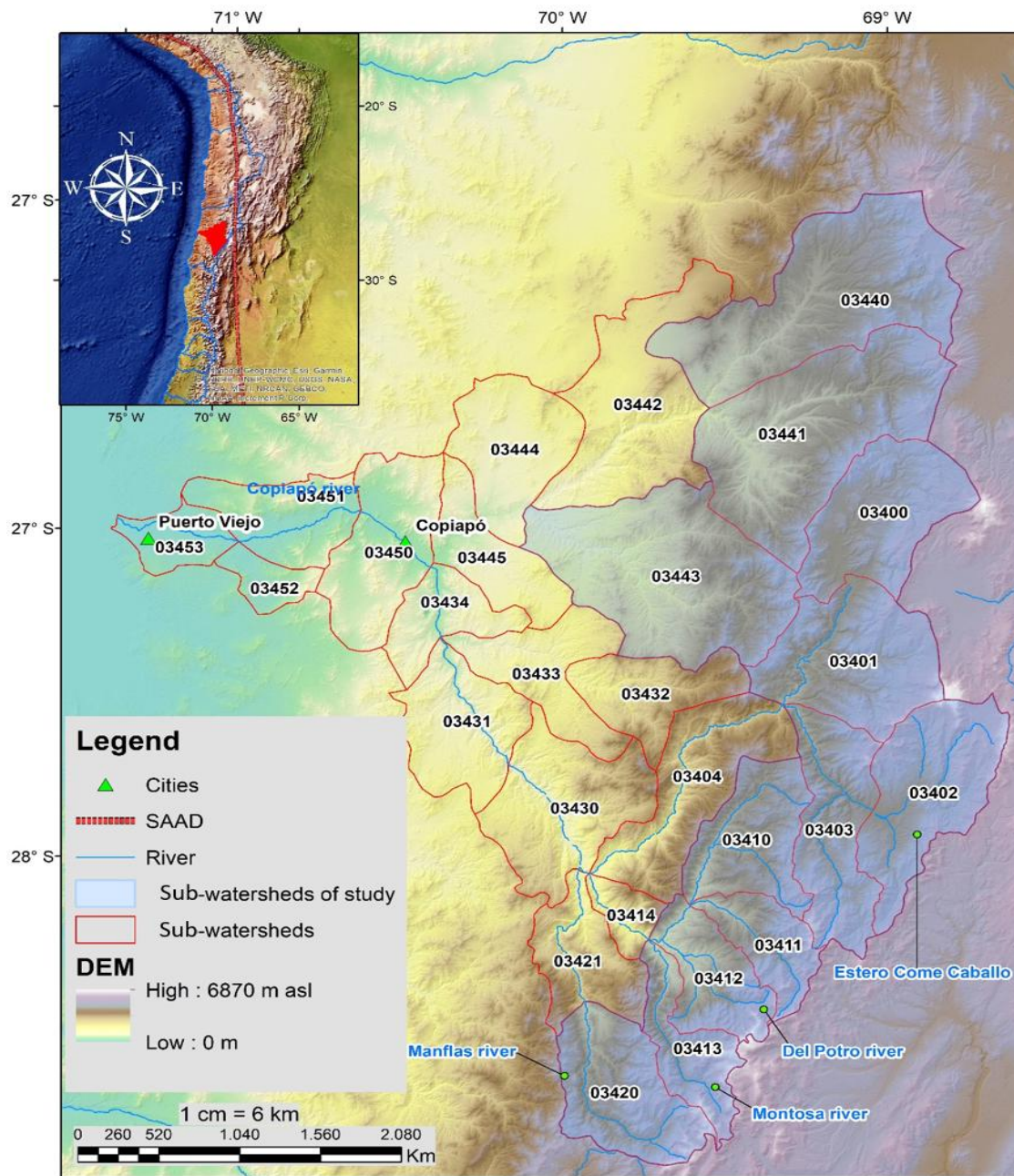


Figure 1. Location map of the Copiapó watershed and the different sub-watersheds (Zone 19S). The codes in each sub-watershed represent the official numbering that the General Water Directorate assigns to each sub-watershed.

103
 104 **Table 1. List of sub-watersheds with cryospheric components in the study area.**

Sub-watershed Code	Sub-watershed name	Sub-watershed area km ²
3400	Quebrada Monardes	803.6
3401	Figueroa River	924.2
3402	Estero Come Caballos	880.6
3403	Cachitos River	734.1
3410	Vizcachas de Pulido river	611.9
3411	Ramadillas River	366.8
3412	Del Potro River	470.6
3413	Montosa River	413.2
3420	Manflas River	728.1
3440	Quebrada San Andres	1476.3
3441	Quebrada Paipote	1493.8
3443	Quebrada Martinez	1513.1

105 **2 Regional setting**

106 **2.1 Study Area**

107 The Copiapó watershed is composed of 27 sub-watersheds, distributed along the high mountain belt that forms the headwaters.
 108 A sub-watershed is defined here as a watershed unit with a lower Strahler stream order than the main watershed. In our research
 109 area, the Copiapó watershed corresponds to the main watershed and contains rivers with a maximum Strahler stream order of
 110 nine. Within this hierarchy, sub-watersheds draining from the headwaters typically initiate with Strahler orders of one to three
 111 and progress downstream until they merge into larger tributaries with stream orders of up to nine. By distinguishing among
 112 sub-watersheds and their associated cryospheric landforms, we can perform both quantitative and qualitative assessments of
 113 the current potential water volumes stored within cryospheric environments.
 114 To classify a sub-watershed as a cryospheric watershed, we applied objective and reproducible criteria based on its cryospheric
 115 components. A sub-watershed is considered cryospheric only when it fulfilled two conditions: (i) the presence of at least one
 116 active cryospheric landforms (debris-free glacier, debris-covered glacier, rock glacier, protalus lobe, or gelifluction slope)
 117 mapped in the regional inventory of glaciers and the periglacial environment of the Atacama Desert (García et al., 2017); and

(ii) measurable seasonal snow, expressed as long-term snow permanence derived from MODIS fractional snow cover (Rittger et al., 2013). Sub-watersheds fulfilling both requirements (≥ 1 cryospheric landforms + snow permanence) were classified as cryospheric.

Applying these criteria resulted in 12 sub-watersheds being identified as cryospheric watersheds (Fig. 1; Table 1). These sub-watersheds are located entirely within high-elevation sectors of the Copiapó watershed (mean elevation 3,617 m a.s.l.), coincide with areas of high permafrost probability in the global permafrost model of Gruber (2012), and contain 97% of the cryospheric landforms mapped in the Copiapó watershed. This classification is therefore geomorphologically justified, reproducible, and consistent with independent permafrost modelling and regional cryospheric inventories.

The Copiapó watershed (Fig. 1) is located in the southern limit of the Atacama Desert, the driest desert of the world (Clarke, 2006), near to the South American Arid Diagonal or SAAD (Zech et al., 2008). The annual precipitation in Copiapó from 1795 to the present has an annual mean of 22.5 mm (Izquierdo et al., 2024). Among the arid watersheds in northern Chile, the Copiapó watershed is regarded as one of the most vulnerable (Suárez et al., 2014). Because of the limited availability of freshwater and the overexploitation of superficial water and groundwater, the deepening of the water table is accelerating and cryospheric landforms and snow are reducing (Casassa et al., 2007; Vuille et al., 2008). Despite this very limited liquid precipitation in the lower areas, e.g. around the city of Copiapó, there is a flow of water in the rivers of the Copiapó watershed year round. This water comes from the Andean cryosphere that is known to host and store large amounts of fresh water (Bolch and Marchenko, 2006; Bórquez et al., 2006; Azócar and Brenning, 2010; Schaffer et al., 2019).

The glacial and periglacial inventory of the Copiapó watershed indicates a transition from areas that are dominated by glacial environments in the south, to areas that are dominated by periglacial environments in the north (García et al., 2017). In the Copiapó watershed, the occurrence of snow begins above 3,500 m of altitude. In these extreme environments, the water budget is significantly influenced by snow sublimation. From 2001 to 2016, approximately 70% of the snow balance was lost to sublimation, effectively reducing the available water volume (Jara et al., 2021). In semi-arid regions of Chile and Argentina, snow accumulation and subsequent melt processes contribute substantially to surface runoff in numerous Andean River Watersheds (Masiokas et al., 2006; Favier et al., 2009).

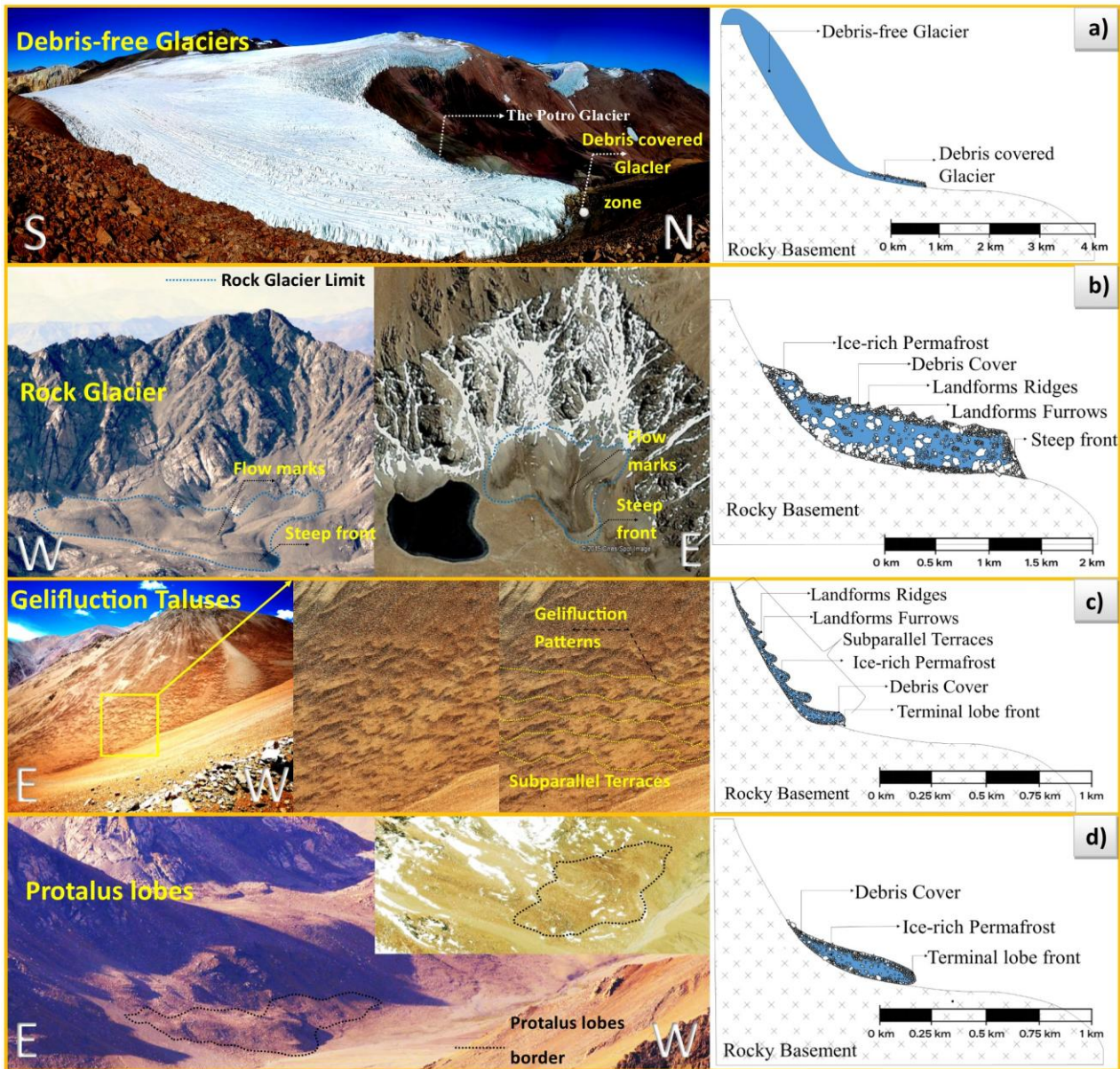


Figure 2. Main cryospheric landforms. Field photographs from the Copiapó Watershed on the left, with a schematic interpretation on the right of. a) Debris-free glacier and debris-glaciers, b) Rock glaciers of the Del Potro River Sub-watershed. c) Gelifluction taluses of Montosa River Sub-watershed, d) Protalus lobes of the La Laguna Sub-watershed in Vizcachas de Pulido Sub-watershed.

2.2 Hydrological background of the cryosphere and conceptual framework for ice-volume estimation

The hydrology of arid Andean Watersheds is dominated by cryospheric processes. In the Copiapó watershed, year-round streamflow persists despite minimal precipitation at mid- to low elevations, indicating that meltwater from snow, glaciers, and ice-rich permafrost constitutes the primary source sustaining discharge. Long-term records from the Dirección General de

152 Aguas (DGA) hydrometric network show a pronounced summer peak in river flow, when rainfall is nearly zero, together with
 153 a high baseflow index, confirming a melt-dominated regime and the central hydrological role of the cryosphere in this region.
 154 These patterns are consistent with previous studies across the semi-arid Andes (Favier et al., 2009; Gascoin et al., 2011; Ayala
 155 et al., 2016), where snow and frozen water bodies act as short- and long-term ice storage modulating runoff during dry years
 156 (Table 2).
 157 Cryospheric contributions, however, operate over markedly different response times. Seasonal snow responds at sub-seasonal
 158 scales, controlling the onset and magnitude of spring–summer discharge. Debris-free and debris-covered glaciers respond at
 159 seasonal to multiannual scales, providing sustained meltwater even in years of limited snowfall. Ice-rich permafrost landforms,
 160 including rock glaciers, protalus lobes, and gelifluction slopes, contribute at decadal to centennial scales and represent the most
 161 climatically buffered ice storage in the watershed. Capturing these contrasted response times requires not only quantifying the
 162 spatial distribution of cryospheric landforms but also applying a coherent conceptual framework for estimating their ice
 163 volumes.
 164 The classification of debris-covered glaciers adopted in this study follows Kirkbride (2011), who defines them as glaciers
 165 whose ablation zone is largely or entirely covered by supraglacial debris across its full width. This distinction is important in
 166 the study area, where debris-covered glaciers occur in close spatial association with rock glaciers but represent fundamentally
 167 different cryospheric processes: debris-covered glaciers are dynamic ice bodies in which the supraglacial covered modulates
 168 ablation, whereas rock glaciers are permafrost landforms whose movement is driven by internal ice deformation in an ice-rich
 169 debris matrix.

171 **Table 2. Contribution to streamflow (%) indicates the proportion of total river discharge supplied by cryospheric meltwater (glaciers**
 172 **and/or seasonal snow), as quantified in each study. Modified from Schaffer et al (2019).**

Location / Watershed	Latitude	Hydrological year	Contribution to streamflow (%)	Cryospheric component evaluated	Reference
Huasco watershed	~29°S	2003/04 and 2007/08	3–23	Glaciers (debris-free + debris-covered)	Gascoin et al. (2011)
Elqui watershed	~30°S	2003	4–9	Glaciers (mainly debris-free)	Favier et al. (2009)
Elqui watershed	~30°S	2011	13	Glaciers (debris-free)	Pourrier et al. (2014)
Juncal River watershed	32–36°S	2005/06 and 2008/09	10 and 31	Glaciers (debris-free), snowmelt	Ragetli & Pellicciotti (2012)
Juncal River watershed	32–36°S	2005/06 and 2008/09	16 and 44	Glaciers (debris-free + debris-covered)	Rodríguez et al. (2016)
Juncal River watershed	32–36°S	2011–2012	2–50	Seasonal snow + glacier melt	Rodríguez et al. (2016)
Yeso River watershed	~33.5°S	2014/15	~42 and ~67	Glaciers + snowmelt (during Megadrought)	Ayala et al. (2016)
Central Chilean Andes	–	1968/69	67	Exposed glaciers (exceptionally dry year)	Peña & Nazarala (1987)

Estimating ice volume in mountain cryospheric systems relies on methodological approaches that vary in complexity, physical basis, and data requirements. Traditionally, regional and global assessments have applied empirical area–volume or area–thickness relationships (e.g., Bahr et al., 1997; Jones et al., 2018). These models provide valuable first-order estimates and remain particularly useful in data-sparse regions. However, by construction, they do not explicitly account for glacier geometry, internal deformation, thermal regime, or spatially variable creep rates, which are processes that exert strong control on ice thickness in both glacial and periglacial landforms.

More recent advances have increasingly emphasized physically based approaches that better represent the mechanics of glacier flow and permafrost creep. For debris-free glaciers, inversion techniques such as the model designed by Farinotti et al. (2017) combine mass conservation, surface slope, and Glen’s flow law to derive spatially distributed ice thicknesses (Fig. 3), and form the basis of the global consensus thickness product of Farinotti et al. (2019). At continental scales, Millan et al. (2022) further demonstrated the value of integrating satellite-derived surface velocities into ice-flow inversion, substantially refining glacier thickness estimates and highlighting the importance of kinematic constraints.

In parallel, physically informed modelling has also progressed for ice-rich permafrost landforms. Viscoplastic formulations, such as those proposed by Cicoira et al. (2021), explicitly link surface velocity, basal shear stress, rheology, and thermal conditions to estimate the thickness of ice-rich layers in rock glaciers. Building on these concepts, Hu et al. (2023) applied deformation-based approaches to regional rock-glacier inventories, illustrating the feasibility of estimating ice content in permafrost landforms using remote sensing–derived kinematics. These developments reflect a broader methodological shift toward deformation-based, physically constrained estimates of ice volume across both glacial and periglacial environments.

In the Andes, the applicability of these physically based approaches is reinforced by the growing availability of satellite-derived deformation data. For example, Villarroel et al. (2018) used Sentinel-1 DInSAR to map surface displacement of more than 2,100 rock glaciers in the Dry Andes, while Vidal-Páez et al. (2024) applied PAZ-SAR SBAS-DInSAR to detect millimetric rock-glacier motion in the Chilean Andes. Together, these studies demonstrate that interferometric and feature-tracking techniques enable the detection of low-magnitude deformation, allowing thickness estimation even in low-creep permafrost landforms such as proglacial lobes and gelifluction slopes.

In this study, we adopt an integrated conceptual framework that applies:

- (1) the physically based Farinotti model to debris-free glaciers;
- (2) a geometry-consistent approach for debris-covered glaciers using thickness gradients from Farinotti et al. (2019); and
- (3) a unified viscoplastic formulation (Cicoira et al., 2021) to all permafrost cryospheric landforms including rock glaciers, proglacial lobes, and gelifluction slopes supported by DInSAR-derived velocities.

203 This unified framework ensures physical consistency across cryospheric landforms with contrasting dynamics and provides
204 the conceptual foundation required for robust estimation of cryospheric ice volumes prior to their integration into the
205 Normalized Cryospheric Index (NCI).

206 **2.3 Glacier and Periglacial Environment**

207 The Copiapó watershed hosts a diverse assemblage of cryospheric landforms (Fig. 2), including debris-free glaciers, debris-
208 covered glaciers, rock glaciers, gelifluction taluses, and protalus lobes, as documented in the regional inventory of García et
209 al. (2017). These landforms exhibit a marked spatial variability across the watershed, reflecting strong controls by elevation
210 and topography.

211 Based on the regional inventory of García et al. (2017), a total of 1862 cryospheric landforms have been mapped within the
212 Copiapó watershed: 62 debris-free glaciers, 16 debris-covered glaciers, 173 rock glaciers, 848 gelifluction taluses, and 763
213 protalus lobes. Gelifluction taluses (46%) and protalus lobes (41%) thus dominate the cryospheric assemblage of the watershed,
214 while debris-free and debris-covered glaciers together account for less than 4% of the inventoried landforms. This numerical
215 distribution reinforces the periglacial character of the Copiapó cryosphere and underscores the relevance of ice-rich permafrost
216 landforms for ice-volume estimations at the watershed scale. This inventory was not originally developed under the Randolph
217 Glacier Inventory (RGI) framework, as it predates the standardization of global glacier datasets in the study region. Instead,
218 the inventory provides a detailed regional mapping of cryospheric landforms, including not only glaciers but also ice-rich
219 permafrost features such as rock glaciers, gelifluction slopes, and protalus lobes. This distinction is particularly relevant
220 because global glacier inventories, such as RGI, are primarily focused on glacier ice bodies and typically do not include
221 periglacial landforms, which are rarely mapped in a systematic manner. As a result, the regional inventory used here captures
222 a broader spectrum of cryospheric components that are especially significant in arid mountain environments, where non-glacial
223 frozen water storage may play an important hydrological role.

224 In general, cryospheric landforms are distributed discontinuously along the high Andes, with notable differences between
225 northern and southern sectors. In areas such as Nevado Jotabeche and Los Tronquitos, landforms are predominantly located
226 on high-elevation ridges and peaks, whereas in the Del Potro region they are more commonly confined to valley bottoms
227 between steep slopes.

228 Gelifluction taluses represent one of the most widespread landforms in the study area and are typically associated with high-
229 elevation environments. In the southern sector, they frequently occur in close proximity to debris-free glaciers, forming
230 coupled glacial–periglacial systems. In contrast, in the northern sector, gelifluction features tend to appear as smaller and more
231 isolated landforms.

232 Rock glaciers are mainly distributed along steep valley flanks at intermediate to high elevations, often occupying topographic
233 depressions or confined valley settings. Debris-free glaciers are generally restricted to the highest elevations and, particularly
234 in the southern sector, are commonly associated with adjacent periglacial landforms.

235 Protalus lobes are especially abundant in some northern areas, where they develop on slopes of varying gradients and exhibit
236 a range of sizes and morphologies. Debris-covered glaciers are comparatively scarce and are typically found in the southern
237 sector, occurring as small bodies located downslope of debris-free glaciers.

238 Overall, the distribution of cryospheric landforms in the Copiapó watershed reflects a transition from glacial-dominated
239 systems in the south to predominantly periglacial environments in the north, consistent with regional climatic and topographic
240 gradients (Fig. 4).

241 The regional inventory of García et al. (2017) predates the RGIK (Rock Glacier Inventories and Kinematics; IPA Action
242 Group) guidelines, but its mapping workflow multi-source satellite photo-interpretation supported by field reconnaissance,
243 identification of active rock glaciers from diagnostic geomorphological criteria (frontal ramp, furrows and ridges), and a
244 GLIMS-compatible attribute database is methodologically consistent with the principles later codified by RGIK. The inventory
245 extends this scope to include gelifluction taluses and protalus lobes, two ice-rich permafrost landforms that dominate the arid
246 Atacama cryosphere but fall outside the RGIK rock-glacier framework.

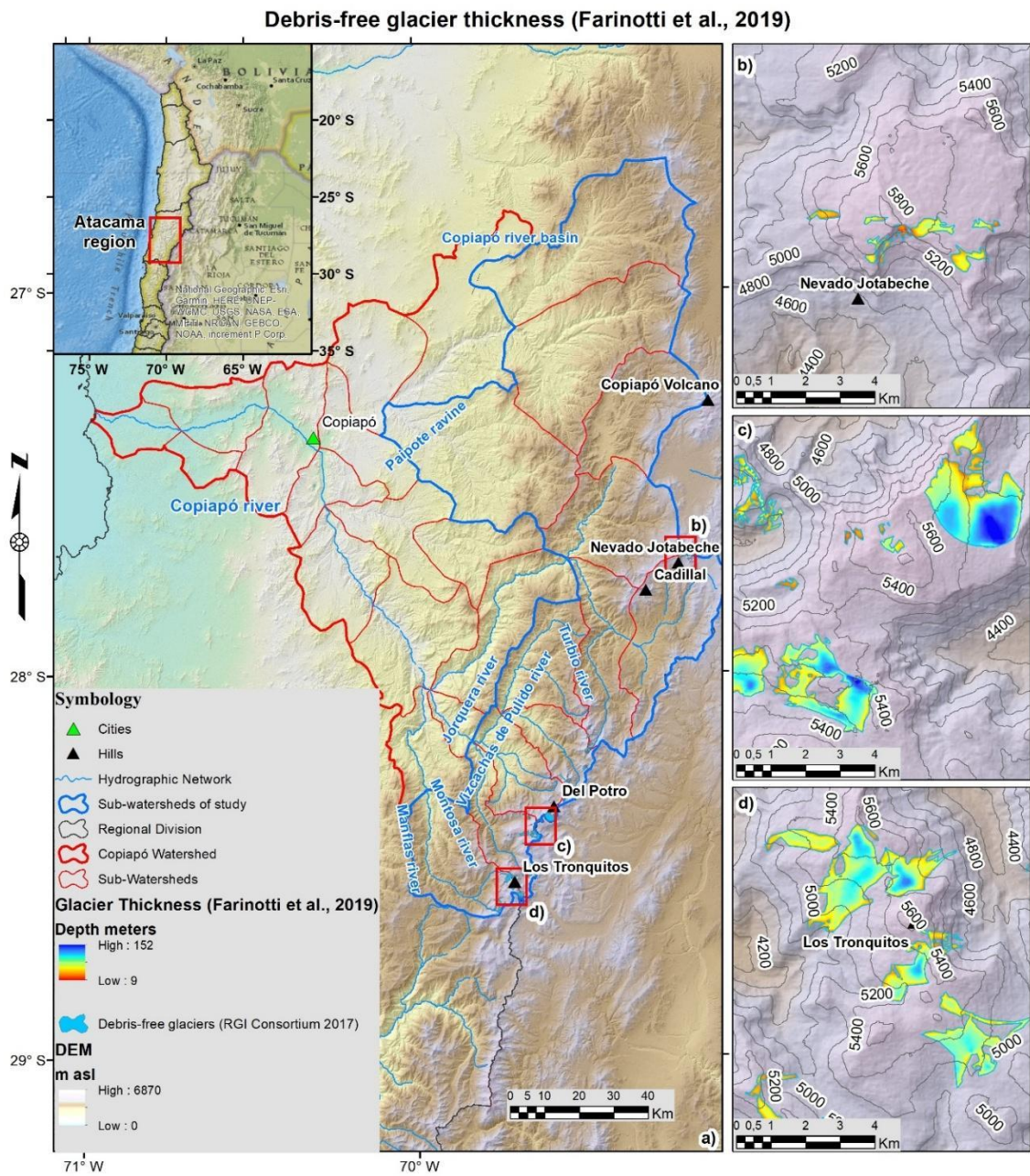


Figure 3. a) Distribution map of the debris-free glacier thickness from Farinotti et al. (2019). b) Jotabeche Volcano cryospheric landforms distribution c) The Del Potro Hill cryospheric landforms distribution d) The Tronquitos Hill cryospheric landforms distribution. Basemap: Esri, Maxar, Earthstar Geographics, and the GIS User Community.

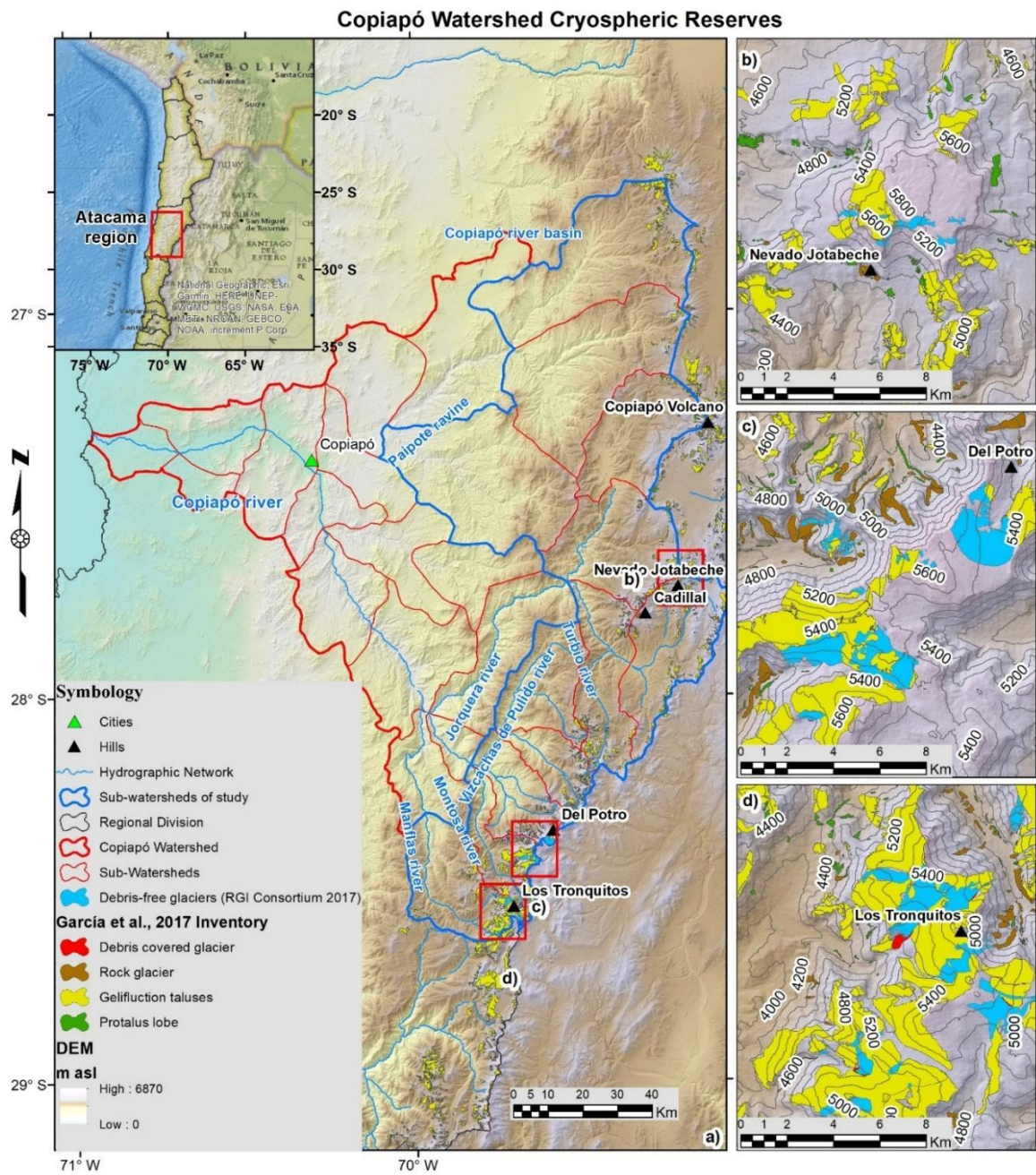


Figure 4. a) Distribution map of the cryospheric landforms inventoried in the Atacama region from García et al. (2017). b) Jotabeche Volcano cryospheric landforms distribution c) The Del Potro Hill cryospheric landforms distribution d) The Tronquitos Hill cryospheric landforms distribution. Basemap: Esri, Maxar, Earthstar Geographics, and the GIS User Community.

258 **3 Methodology**

259 **3.1 Survey of cryospheric landforms**

260 To quantify the ice volume in the Copiapó watershed, we used existing cryospheric inventories without performing new
261 delineations. The regional inventory of glaciers and periglacial landforms of the Atacama Desert (García et al., 2017) was
262 employed directly to extract polygons of debris-covered glaciers, rock glaciers, protalus lobes, and gelifluction slopes. This
263 inventory represents the only comprehensive and validated dataset that explicitly identifies periglacial cryospheric landforms
264 in northern Chile, and its polygons were incorporated without modification into our analysis.

265 Debris-free glaciers were obtained exclusively from the Randolph Glacier Inventory (RGI) v6.0 (RGI Consortium, 2017). For
266 these glaciers, we did not compute ice thickness ourselves. Instead, we used the pre-computed global ice-thickness product of
267 Farinotti et al. (2019), which is physics-based and already calibrated on RGI glacier outlines (Fig. 3). Because the Farinotti
268 model requires RGI geometry as input, the use of RGI v6.0 ensures direct compatibility with the global thickness dataset. Ice
269 volumes for debris-free glaciers were therefore calculated solely by spatially integrating the published Farinotti et al. (2019)
270 thickness values over the RGI glacier polygons within the Copiapó watershed. This approach avoids any reinterpretation of
271 glacier geometry or re-estimation of ice thickness, ensuring methodological transparency and consistency with the global
272 consensus product.

273 The joint use of the inventory developed by García et al. (2017) and RGI v6.0 glacier inventory is thus complementary. RGI
274 provides standardized glacier boundaries and their associated Farinotti-derived thickness fields for debris-free glaciers,
275 whereas García et al. (2017) provides the spatially explicit identification of debris-covered glaciers and periglacial landforms
276 that are not represented in global glacier datasets. By relying directly on these inventories and on the published model outputs
277 of Farinotti et al. (2019), our workflow remains fully reproducible and avoids introducing new mapping or thickness-estimation
278 steps.

279 Mapping uncertainty for all polygons was propagated using the one-pixel boundary buffer method of Paul et al. (2013),
280 ensuring consistent treatment of spatial uncertainty across glacial and periglacial cryospheric landforms.

281 **3.2 Survey of snow cover: MODIS-derived snow permanence**

282 The MOD10A1 Fractional Snow Cover (FSC) product from NASA was used to characterize snow based on long-term patterns
283 of snow permanence. Each MOD10A1 image provides, for every 500×500 m pixel, a value representing the fractional snow
284 cover within that pixel (from 0 to 1), derived through a spectral unmixing algorithm (Rittger et al., 2013). This sub-pixel
285 representation allows a more accurate depiction of discontinuous or patchy snow, which is common in high-elevation arid
286 regions. Instead of relying on isolated scenes, we processed the complete daily MODIS time series (2000–2022) to obtain both
287 the instantaneous snow-covered area and the frequency of snow permanence at the sub-watershed scale. Although cloud
288 contamination can introduce occasional misclassifications, the product maintains approximately 90% accuracy under ideal
289 conditions (Hall et al., 2002; Jiang et al., 2024), and is suitable for long-term characterization of snow permanence.

From the daily FSC series, each MODIS image provides the fractional snow cover for every 500×500 m pixel. Using this long-term record, fractional snow cover (FSC) was used as the core variable to represent the proportion of snow within each pixel in each scene. In addition, we calculated snow permanence as the fraction of days during the 23-year period in which each pixel exhibited snow-covered conditions according to its FSC value. Snow permanence is therefore not an independent MODIS layer, but a temporal aggregation derived directly from the daily FSC maps, summarizing the frequency and recurrence of snow presence across the watershed.

In this framework, FSC was used as the primary input for constructing long-term mean FSC maps. For each pixel, the daily FSC values were averaged over the full 2000–2022 record to obtain a time-mean fractional snow cover. This per-pixel mean FSC was then multiplied by the pixel area (0.25 km^2 for the 500×500 m MODIS grid) and summed across all pixels within each sub-watershed to yield a mean snow-covered area in km^2 . It is this spatially integrated snow-covered area (km^2) that is incorporated into the Normalized Cryospheric Index (NCI) as the indicator of short-term cryospheric water resources, ensuring that both the fractional snow signal at the pixel level and the spatial extent of the sub-watershed are accounted for in the index.

3.3 Field work for ground truthing

Since the mapping workflow is primarily based on remote sensing, satellite imagery, and existing cryospheric inventories, we incorporated targeted field observations and independent geophysical datasets to validate both the identification of cryospheric landforms and the ice-thickness values used in this study. Fieldwork was carried out in the Montosa (03413) and Del Potro (03412) sub-watersheds, which together represent approximately 10% of the Copiapó watershed and contain accessible examples of the main cryospheric landforms present in the region.

During these campaigns, photographic and geomorphological observations were conducted to verify the surface expression of mapped cryospheric landforms, including gelifluction slopes, protalus lobes, and rock glaciers. These observations allowed us to assess the representativeness of the polygons obtained from existing inventories.

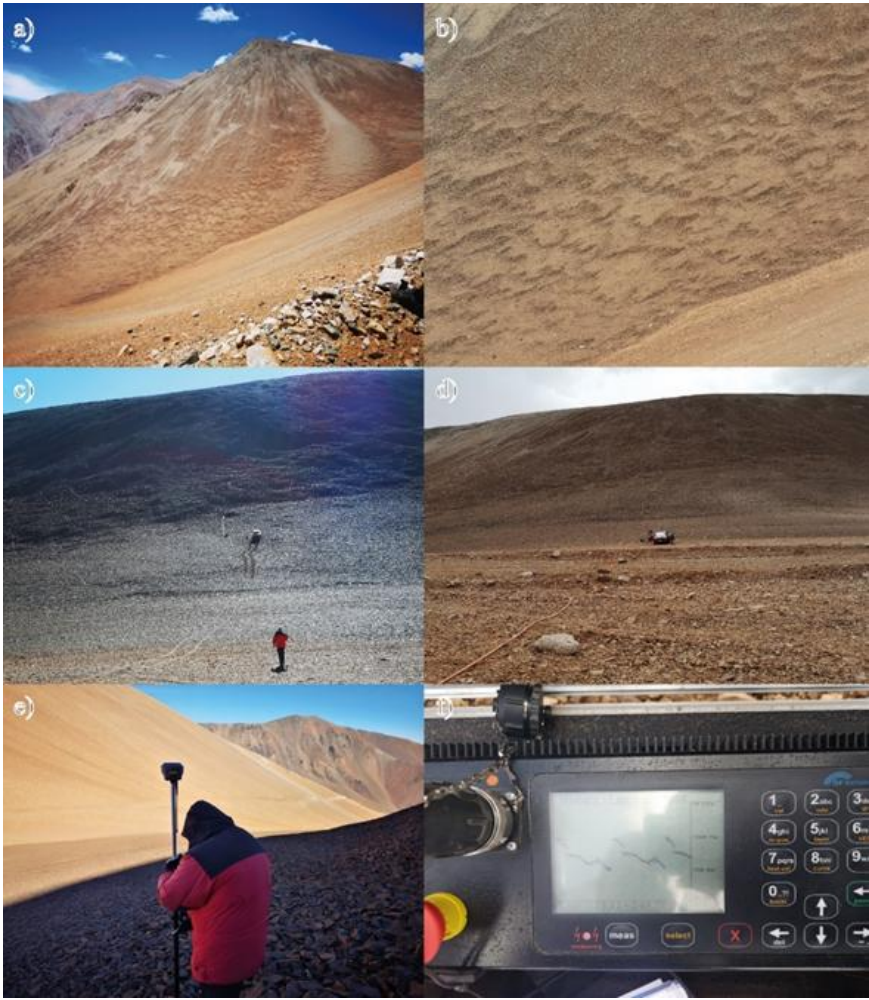
Beyond surface verification, independent geophysical datasets were used to constrain and validate ice-thickness estimates. For debris-free glaciers, ice thickness values from the Farinotti et al. (2019) global product were compared against airborne ground-penetrating radar (GPR) profiles collected by the Dirección General de Aguas (DGA, 2014) over the Del Potro and Tronquitos glaciers.

For ice-rich permafrost landforms, Electrical Resistivity Tomography (ERT) measurements were conducted on a representative gelifluction slope and on an active rock glacier within the Montosa sub-watershed. These ERT profiles constrain the depth and continuity of ice-rich layers and support the thickness ranges adopted for periglacial cryospheric landforms in the modelling framework.

Together, these field observations and geophysical datasets provide independent validation for both the spatial delineation and the subsurface characteristics of cryospheric landforms within the Copiapó watershed, strengthening the robustness of the ice-volume estimates applied in this study.

323 **3.3.1 Geophysical survey on gelifluction slopes and rock glaciers**

324 Electrical Resistivity Tomography (ERT) was employed to characterize subsurface ice-rich layers in selected periglacial
325 landforms and to provide observational constraints for the ice-thickness values used in the modelling framework. ERT is a
326 multi-electrode resistivity method that generates 2-D vertical sections of the subsurface by injecting electrical current into the
327 ground and measuring the resulting potential differences across electrode arrays. The field configuration used in this study
328 (Table 3) consisted of 36 electrodes operated with Wenner, Schlumberger, and Dipole–Dipole arrays.



329 **Figure 5. Photography of the fieldwork. a) And b) the Lobulated morphology of studied gelifluction taluses, which are not covered**
330 **in snow; c) and d) the geoelectric line cable; e) and f) the used equipment, differential GPS and resistivity meter ARES II measuring**
331 **a resistivity point diagram distribution.**
332

333 Fieldwork focused on two representative cryospheric landforms within the Montosa sub-watershed: a gelifluction slope and
334 an active rock glacier (Fig. 5 and Fig. 6). These sites were selected because they exhibit clear geomorphological indicators of

ice-rich permafrost and provide suitable access for geophysical surveys. The ERT profiles allowed direct identification of high-resistivity zones ($>20\text{--}50\text{ k}\Omega\cdot\text{m}$), characteristic of massive or highly saturated ice-rich permafrost bodies.

Table 3 Used geophysical equipment for fieldwork of this study

Equipment	Amount
Resistivity meter: ARES II	1
30 electrode box	5
10 channels multicable II (60 m)	10
Car Batteries	2
Differential GPS	1

The survey layout required a 600-m profile organized through multicables arranged at 90° angles. A high-precision differential GPS (“South Galaxy”) was used to set the start and end points of each line, ensuring alignment and minimizing installation error. Electrode spacing was fixed at 5 m, with position control verified using a combination of GPS and measured markers at 30-m intervals. All electrodes were tested prior to each acquisition, and any with unacceptable standard deviation values were repositioned.

These ERT measurements provide direct ground validation of subsurface ice for both gelifluction taluses and rock glaciers, supporting the thickness ranges applied to ice-rich permafrost cryospheric landforms in our volume estimations and complementing the GPR-based glacier-thickness validation described in Section 3.3.



Figure 6. Photography of the fieldwork rock-glacier surface. (a) Intelligent multi-electrode cable reel used for deploying the ERT array across the periglacial terrain. (b) Installation of the 600-m geoelectric line. (c) Operation of the ARES II resistivity meter during data acquisition.

Before every measurement, all electrodes were tested to detect electrodes with high standard deviation in resistivity. If the deviation exceeded acceptable limits, the electrodes were repositioned until the deviation fell within an acceptable range to ensure measurement accuracy.

3.4 DInSAR kinematic analysis for periglacial cryospheric landforms

To characterize surface deformation across glacial and periglacial cryospheric landforms and to provide kinematic constraints for ice-thickness modelling, we processed Sentinel-1 IW Single Look Complex (SLC) images using the ESA SNAP v9.0 toolbox. Interferometric pairs were selected based on perpendicular baselines <150 m and temporal baselines <24 days to ensure coherence in high-altitude terrain (Table 4).

Table 4 Sentinel-1 interferometric image pairs and associated temporal and perpendicular baselines tested.

Reference Image Date	Secondary Image Date	Orbit Type	Temporal Baseline (Days)	Perpendicular Baseline	Interferogram status
01-03-2024	13-03-2024	Ascending	12	55.35	Inconsistent
25-03-2024	06-04-2024	Ascending	12	40.12	Inconsistent
03-03-2016	27-03-2016	Descending	24	9.39	Coherent
05-01-2015	29-05-2015	Ascending	24	-121.04	Inconsistent
29-01-2015	22-02-2015	Descending	24	25.41	Inconsistent

SLC scenes were co-registered using the TOPS burst co-registration workflow. Interferograms were generated, filtered using a Goldstein filter ($\alpha = 0.8$), and corrected for topographic phase using the 30-m SRTM DEM. Phase unwrapping was performed with SNAPHU in external mode. Line-of-sight (LOS) displacement was obtained from the unwrapped phase:

$$d_{LOS} = \frac{\lambda}{4\pi} \Delta\phi \tag{1}$$

Mean surface velocities were extracted for each cryospheric landforms polygon (debris-free glaciers, debris-covered glaciers, rock glaciers, protalus lobes, gelifluction slopes). These DInSAR-derived velocities were incorporated as dynamic constraints

for the viscoplastic modelling of permafrost cryospheric landforms and as supplementary kinematic information in glacier-thickness interpretation.

3.4.1 Velocity uncertainty

Uncertainty in DInSAR-derived surface velocity was incorporated by explicitly estimating the phase noise associated with interferometric coherence and temporal baseline. Following Hanssen (2001), the standard deviation of the interferometric phase for a multilooked interferogram is:

$$\sigma_{\phi} \approx \sqrt{\frac{1-\gamma^2}{\gamma^2 L}} \quad (2)$$

Where γ is the interferometric coherence and L the number of equivalent looks. For Sentinel-1 SLC data used in this study, the typical modeled coherence is $\gamma=0.97$, the temporal baseline is $\Delta t=24$ days, and the radar wavelength is $\lambda=0.0555$ m.

Phase noise is converted to line-of-sight displacement noise using:

$$\sigma_{\phi} = \frac{\lambda}{4\pi} \sigma_{\phi} \quad (3)$$

And subsequently velocity uncertainty:

$$\sigma_v = \frac{\sigma_d}{\Delta t} = \frac{\lambda}{4\pi \Delta t} \sqrt{\frac{1-\gamma^2}{\gamma^2 L}} \quad (4)$$

For a representative multilooking a factor of $L=10$ was applied. The resulting uncertainties are:

$$\begin{aligned} \sigma_{\phi} &\approx 0.056 \text{ rad}, \quad \sigma_d \approx 2.5 \times 10^{-4} \text{ m} = 0.25 \text{ mm}, \\ \sigma_d &\approx 1.0 \times 10^{-5} \text{ m d}^{-1} = 0.01 \text{ mm d}^{-1} \approx 3 - 4 \text{ mm a}^{-1}. \end{aligned}$$

This value ($\sigma_d = 0.01 \text{ mm d}^{-1}$) was thus adopted as the 1σ uncertainty of DInSAR velocity measurements.

3.5 Calculation of the ice volume and uncertainty

To quantify the total ice volume stored in glacial and periglacial cryospheric landforms, we applied cryospheric landforms-specific physics-based models according to their mechanics, morphology, and available input data (Table 5). The method applied and associated uncertainties are detailed below.

3.5.1 Debris-free glaciers

Ice thickness for debris-free glaciers was obtained from the global consensus ice-thickness dataset of Farinotti et al. (2019), which integrates multiple physics-based inversion models (Farinotti et al., 2017). Glacier volumes were computed by spatially

408 integrating the consensus ice-thickness rasters over the RGI v6.0 glacier polygons within the Copiapó watershed. A relative
 409 uncertainty of $\pm 26\%$ was applied following the uncertainty assessment reported in Farinotti et al. (2017, 2019).

410 3.5.2 Debris-covered glaciers

411 Mean ice thickness for debris-covered glaciers was derived by sampling the Farinotti et al. (2019) consensus ice-thickness
 412 field in the clean-ice frontal zones immediately adjacent to each debris-covered area. These representative values were then
 413 assigned to the corresponding debris-covered glacier polygons, ensuring consistency with the local glacier geometry and
 414 thickness gradients. Ice volumes were calculated by multiplying the assigned mean thickness by the mapped debris-covered
 415 glacier area. Due to the inherently greater spatial variability and additional uncertainties associated with debris cover (e.g.,
 416 variations in debris thickness, thermal properties, and remote-sensing-derived thickness estimates), published studies of
 417 debris-covered glacier thickness consistently report substantially larger differences between modeled and *in situ* measurements
 418 compared to debris-free glaciers (e.g. $\sim 40\%$ uncertainty for certain debris-covered glaciers; see Pieczonka et al., 2018). Hence,
 419 a conservative uncertainty of $\pm 35\%$ was adopted for debris-covered glacier volume estimates to account for these factors.

420 3.5.3 Rock glaciers

421 Rock-glacier thickness was estimated using a viscoplastic creep inversion following Cicoira et al. (2021), where basal shear
 422 stress is constrained by observed surface velocity. For each rock glacier, mean slope-parallel surface velocities ($\underline{v}$) and their
 423 standard deviation (σ_v) were derived from Sentinel-1 DInSAR measurements (Sect. 3.4.1) and expressed in m yr^{-1} . The basal
 424 shear stress τ was inverted using the 1-D viscoplastic formulation with Glen-type exponent $n = 3$ and a temperature-dependent
 425 rate factor (T).

426
 427 To explicitly represent uncertainty in the thermal state of ice-rich permafrost in the Central Andes, we treated ice temperature
 428 as a normal distribution:

$$430 \quad T \sim N(-3^\circ\text{C}, 3^\circ\text{C}) \quad (5)$$

431
 432 This choice is consistent with reported near-isothermal to moderately cold permafrost conditions in the Central Andes,
 433 including rock-glacier settings, and with the strong temperature sensitivity of creep close to 0°C described in constitutive
 434 studies (e.g. Koenig et al., 2025, Monnier & Kinnard, 2013).

435
 436 Using τ , mean thickness was computed as:

$$437 \quad H = \frac{\tau}{\rho \cdot g \cdot \sin(\alpha)} \quad (6)$$

439 where α is the mean surface slope, ρ is ice density, and g is gravitational acceleration. Total landform volume was computed
440 as:

$$441 \quad V = A \cdot H \quad (7)$$

442 where A is the mapped planimetric area.

443 Ice volume was obtained by applying an ice-content fraction C_I to the landform volume:

$$444 \quad V_{ice} = V \times C_I \quad (8)$$

445 For rock glaciers, C_I was represented as a bounded range consistent with reported values for these type of ice-rich permafrost
446 landforms (Cicoira et al., 2021):

$$447 \quad C_I = 0.5 \pm 0.10 \quad (9)$$

448 Ice mass (in Gt) was then computed as:

$$449 \quad M_{Gt} = \frac{V_{ice} \rho}{10^{12}} \quad (10)$$

450 with $\rho = 900 \pm 20 \text{ kg m}^{-3}$. Uncertainties in H and M_{Gt} were propagated from σ_v , σ_T , slope uncertainty, mapping
451 uncertainty in area, ice-content uncertainty, and density uncertainty using quadratic propagation of relative errors (Sect. 3.5.5).

452 **3.5.4 Protalus lobes and gelifluction slopes**

453 Protalus lobes and gelifluction taluses were modelled using the same viscoplastic inversion framework originally developed
454 for rock glaciers (Cicoira et al., 2021), exploiting low but measurable DInSAR-derived surface deformation to constrain basal
455 shear stress and thickness. For each mapped polygon, mean slope-parallel surface velocity and its associated uncertainty (v ,
456 σ_v) were used to invert basal shear stress and compute thickness H through Eq. (6). Thermal uncertainty was represented using
457 the same Central Andes temperature prior described in Eq. (5).

458 However, in contrast to rock glaciers, protalus lobes and gelifluction taluses are characterized by lower ice contents, higher
459 debris fractions, and a dominance of seasonal or shallow permafrost deformation mechanisms. To account for these rheological
460 differences, we modified the Glen-type flow exponent n used in the inversion. While rock glaciers were modelled assuming n
461 $= 3$, consistent with creep-dominated ice–debris mixtures (Cicoira et al., 2021), protalus lobes and gelifluction taluses were
462 modelled using a reduced exponent ($n = 2$), reflecting more linear, frictionally controlled deformation regimes typical of ice-
463 poor frozen debris slopes. This adjustment reduces the sensitivity of inferred thickness to surface velocity and avoids
464 unrealistically thick solutions for low-velocity landforms.

Because ice contents in these landforms are generally lower and more heterogeneous than in rock glaciers, ice-content fraction was prescribed following Eq. (11) as a bounded prior range rather than a fixed value. This range represents plausible ground-ice fractions for ice-rich permafrost slopes in the Central Andes and is supported by field-calibrated ERT observations showing strong spatial variability in ground-ice content within Andean permafrost landforms (Hilbich et al., 2022), as well as early empirical constraints for frozen debris slopes in arid mountain environments (Schrott, 1996). The ice-content prior was therefore implemented to represent structural uncertainty rather than a deterministic parameter.

$$C_I = 0.37 \pm 0.12 \quad (11)$$

Following thickness estimation, landform volume was computed using Eq. (7), ice volume using Eq. (8), and ice mass using Eq. (10).

3.5.5 Uncertainty propagation framework for ice-rich permafrost cryospheric landforms

For rock glaciers, gelifluction taluses, and protalus lobes, uncertainties in ice thickness and ice mass were propagated deterministically using quadratic propagation of relative errors for multiplicative quantities. First, the uncertainty in mapped area was estimated following Paul et al. (2015), assuming boundary errors occur within ± 0.5 pixels and using:

$$u_{A_i} = \frac{1}{2} P_i \times \lambda \times \sigma \quad (12)$$

where P_i is the polygon perimeter, λ is the spatial resolution of the imagery used for delineation, and $\sigma \approx 0.6872$ is a correction factor corresponding to the standard deviation of a Gaussian distribution with most errors confined within ± 0.5 pixels.

Ice volume was computed using the Eq. 7, and ice mass as $M_{Gt} \propto A \times H \times C_I \times \rho$. Therefore, the relative uncertainty in ice mass was calculated as:

$$\left(\frac{\sigma_M}{M}\right)^2 = \left(\frac{\sigma_A}{A}\right)^2 + \left(\frac{\sigma_H}{H}\right)^2 + \left(\frac{\sigma_{C_I}}{C_I}\right)^2 + \left(\frac{\sigma_\rho}{\rho}\right)^2 \quad (13)$$

and the absolute uncertainty as $\sigma_M = M \times \left(\frac{\sigma_M}{M}\right)$. Watershed totals were obtained by summing masses across polygons, while total uncertainty was aggregated assuming independence using:

$$\sigma_{tot} = \sqrt{\sum_i^n \sigma_{M_i}^2} \quad (14)$$

494 This framework ensures consistent uncertainty propagation across all ice-rich permafrost cryospheric landforms while
495 preserving the physically based dependence of thickness on surface velocity and temperature.

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497 **Table 5. Key references used as input for the calculation of ice volumes of cryospheric landforms.**

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Cryospheric landforms	Method	Ice content	Key references
Debris-free glaciers	Physics-based inversion (Farinotti et al., 2019)	100%	Farinotti et al. (2017, 2019)
Debris-covered glaciers	Thickness assignment from adjacent clean-ice zones (Farinotti 2019)	45–99%	Farinotti et al. (2017, 2019)
Rock glaciers	Viscoplastic model (Cicoira et al., 2021) + DInSAR	40–60%	Cicoira et al. (2021)
Gelifluction taluses and Protalus lobes	Adapted viscoplastic model + DInSAR	25–49%	Hilbich et al. (2022), Schrott, (1996)

514 **3.6 Cryospheric landforms and snow index and data integration**

515 To develop a classification system that categorizes the capacity to store ice as a strategic water reserve in desert areas of sub-
516 watersheds, we followed a workflow that consists of four phases (Fig. 7). We developed the classification system using
517 cryospheric landforms (expressed in gigatons (Gt), equivalent to 1,000 million metric tons of ice) and snow (quantified in km²
518 of snow-cover extent), integrating both into the methodological framework. The strategic Normalized Cryospheric Index (NCI)
519 captures the relative significance of each sub-watershed, considering both its ability to store water as ice (cryospheric
520 landforms) and its potential to regenerate water resources through snowfall. The NCI needs to be adapted to each specific case.
521 For the Copiapó watershed, the active cryospheric landforms present until 2024 have been used. Those cryospheric landforms
522 include debris-free glaciers, debris-covered glaciers, rock glaciers, gelifluction talus and protalus lobes. Depending on the
523 types of cryospheric landforms available in a specific region, the number of cryospheric landforms included may vary.
524 The methodological framework for computing the NCI index integrates Monte Carlo simulation with a robustness assessment
525 of a composite indicator that incorporates ice volume (V) and snow permanence (S), adhering to the uncertainty propagation

principles in simulation-based methodologies outlined by Rubinstein and Kroese (2016). First, each watershed has a mean value (e.g., $\bar{V}_i$ y $\bar{S}_i$) and an estimate of its dispersion (ΔV_i and ΔS_i), usually understood as standard deviation. On the basis that both V and S follow normal distributions $N(\bar{V}_i, \sigma_v)$ y $N(\bar{S}_i, \sigma_s)$, N replicates of each variable are generated for each watershed, thus obtaining “clouds of values” representing the uncertainty of each datum. Subsequently, a normalization is performed (using min-max) in order to avoid that one variable dominates the other only by differences in magnitude.

Once the simulated values are generated, a weighted index is defined whose objective is to combine both variables (Eq. 15):

$$NCI_{i,k}(\omega) = \omega \cdot V_{i,k} + (1 - \omega) \cdot S_{i,k} \quad (15)$$

Where $\omega \in [0,1]$

Where ω is a parameter that determines the relative relevance of ice volume versus snow permanence. For each sample k (among the N simulations), the values of that index in all the watersheds are obtained and ordered from highest to lowest, giving rise to a specific ranking of each run. In this way, each watershed “i” gets a “position” (1 for the watershed with the highest index, 2 for the next one and so on).

The optimization function (Eq. 16) relies on the stability of the ranking under uncertainty, in accordance with the sensitivity analysis framework outlined by Saltelli et al. (2008). Specifically, for each watershed, the standard deviation of its position over the N runs is measured, and this measure of dispersion is summed (or averaged) for all watersheds. A smaller value of this sum indicates greater stability or robustness in the ranking. Formally, if σ_i denotes the standard deviation of the position of watershed under a weight ω , the objective function results:

$$f(\omega) = \sum_{i=1}^n [\sigma_i \cdot (\omega)] \quad (16)$$

Where the idea is to minimize the value of this function. To determine the “optimal” ω , a sweep over [0.1] is performed. The optimal result corresponds to the ω value that maximizes the stability of the watershed ranking system, ensuring minimal sensitivity to errors or noise in the V and S datasets.

Finally, plots are generated to analyze the resulting average ranking and the probability that certain watersheds occupy the first place in various simulations. In this way, a picture is obtained of how uncertainty in ice volume and snow permanence affects the prioritization of watersheds.

Although the NCI has a simple formulation, the variables it combines are affected by substantial uncertainty. Ice volumes integrate multiple sources of error related to thickness modelling, ice content, density, and mapped area, while snow is represented by long-term snow permanence derived from MODIS, which captures both the spatial extent and temporal recurrence of snow. Monte Carlo simulation is therefore applied not to increase index complexity, but to propagate uncertainty

in ice volume and snow permanence into watershed ranking and to identify an objective weighting factor (ω) that maximizes ranking robustness. This stochastic framework avoids subjective parameterization and ensures a stable, transparent, and reproducible classification of cryospheric sub-watersheds.

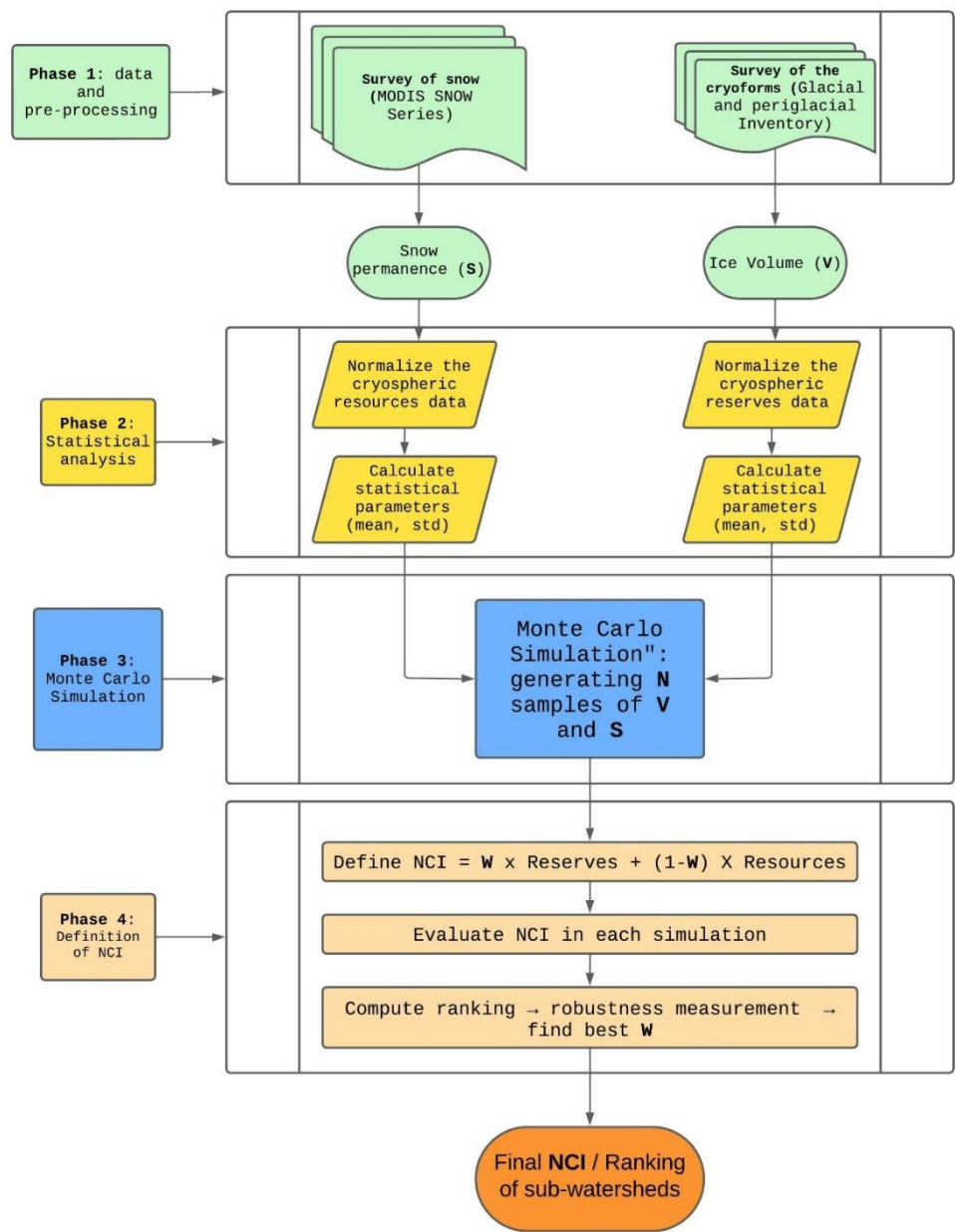


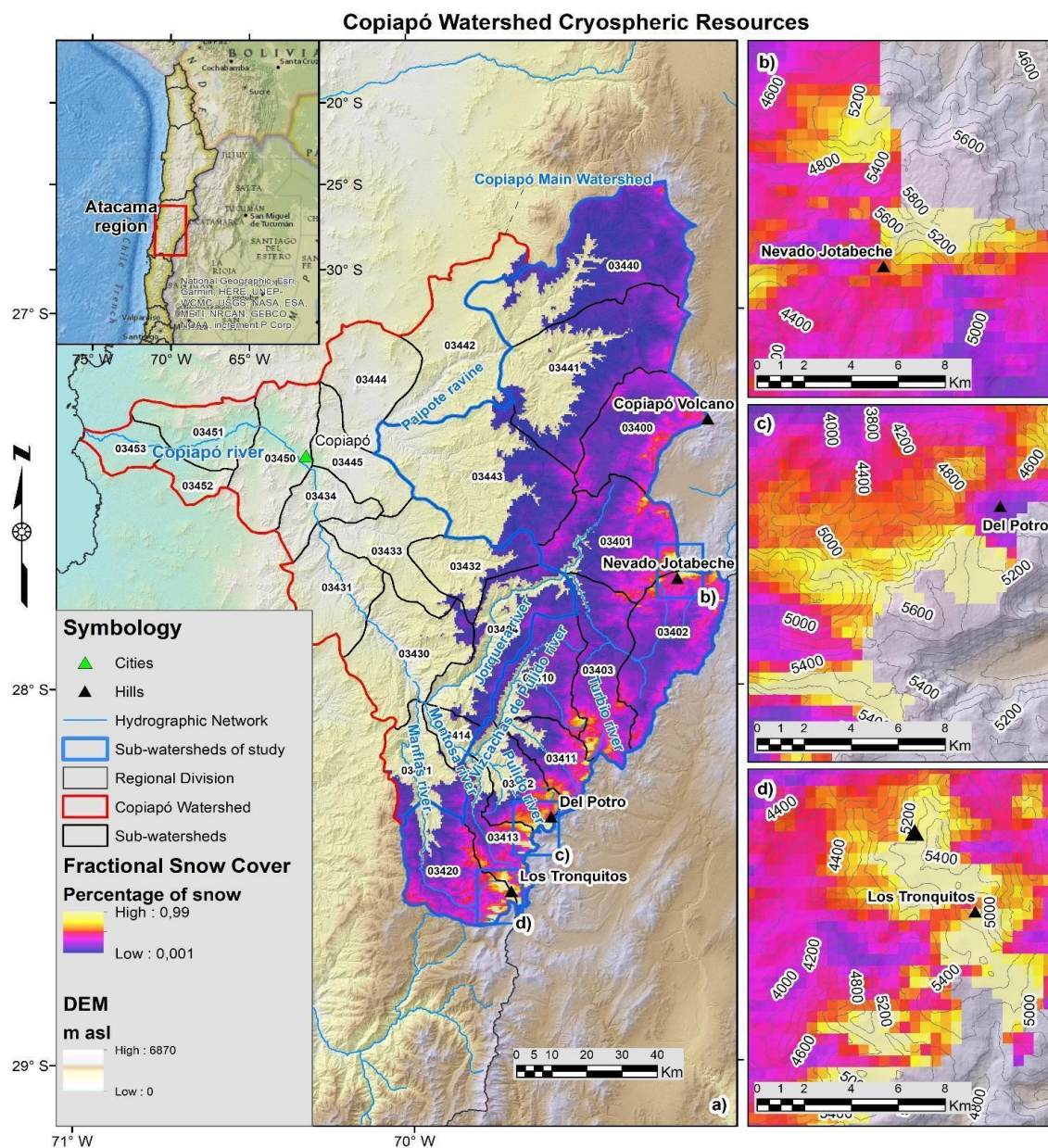
Figure 7. Flow chart of the methodology used to calculate the NCI.

565 **4 Results**

566 **4.1 Snow permanence results**

567 By integrating all MODIS fractional snow cover images and computing their long-term mean over the period 2000–2022, we
568 derived a snow permanence metric, defined as the probability of snow presence at each pixel. The resulting raster represents
569 the spatial distribution of snow permanence across the Copiapó watershed (Fig. 8). At the regional scale, snow permanence
570 shows limited spatial variability at high elevations, remaining consistently high above approximately 5,500 m a.s.l. These
571 regions are located close to the Argentinean border. In the central area between Caserones Hill and Cadillal Hill, no cryospheric
572 landforms were identified, which coincides with the lowest snow permanence values in the study area, with snow present only
573 around 30% of the time. In contrast, the highest snow permanence values are observed in the southern sector of the study area.
574 In the central area of the Copiapó watershed the snow permanence is concentrated around the Nevado Jotabeche Hill area. The
575 northern part of the watershed does not show concentrated major snowfall anomalies. Snow permanence has persisted
576 approximately 30–50% of the time exclusively at high-elevation zones (>4,500 meters above sea level) near the Argentinian
577 border, contrasting sharply with lower altitudes starting at 2,000 m a.s.l.

578 The calculated probability of the occurrence for the investigated watersheds shows that the snow permanence watershed varies,
579 with higher snow permanence for the southern watersheds compared to the central and northern watersheds. The watershed
580 with the highest resource index is the Montosa River watershed (03413) with 0.18 probability of being covered in snow. The
581 Vizcachas de Pulido River (03410), which is located at the lowest altitude, has the lowest probability of 0.07 to have snow
582 permanence. These per-pixel snow permanence values, when multiplied by the pixel area (0.25 km²) and integrated across
583 each sub-watershed, yield the mean snow-covered area in km² reported in Table 6 and used as the snow variable S in the NCI
584 calculation.

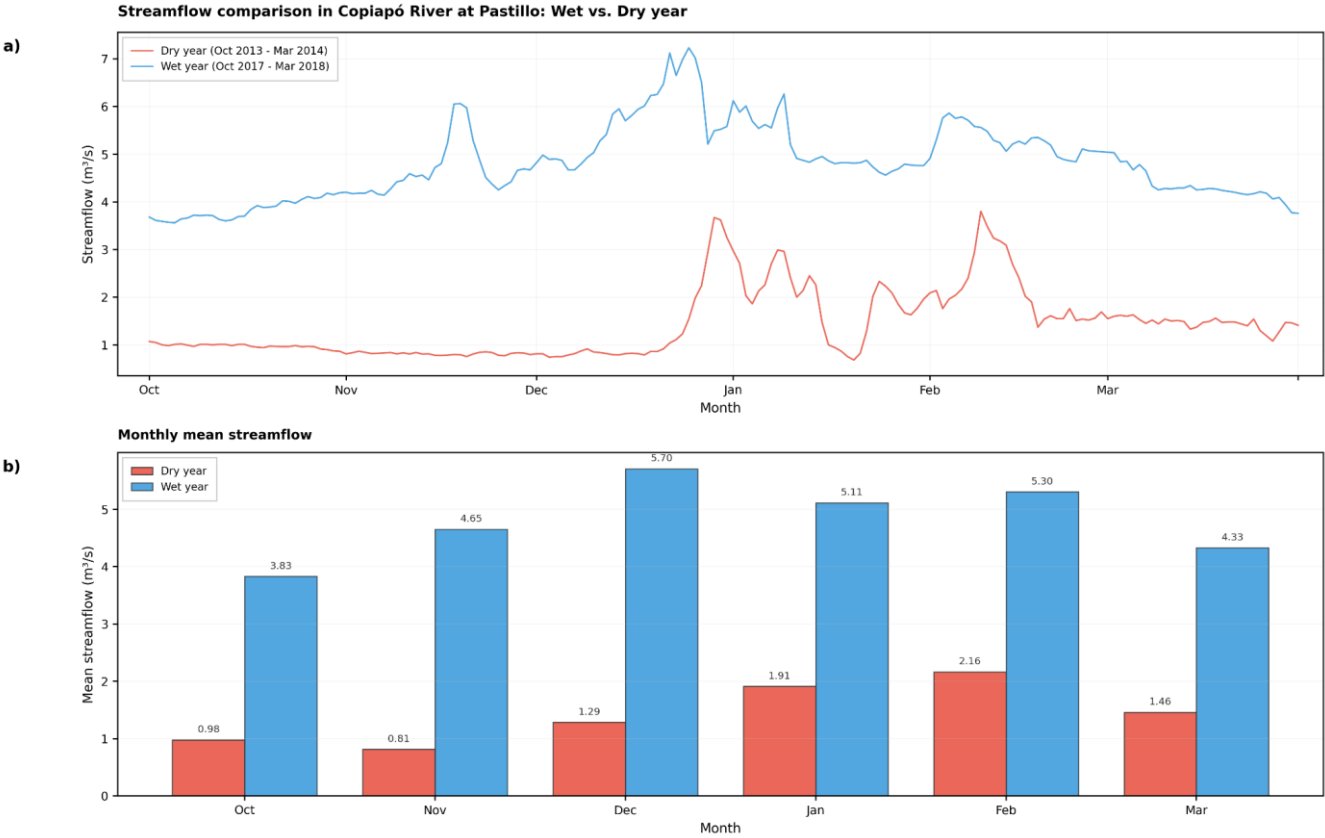


4.2 Analysis of runoff

Before incorporating snow permanence as a core variable of the NCI, we tested empirically whether MODIS-derived snow cover translates into a hydrological signal at the watershed scale. To do this, we contrasted streamflow behaviour during representative wet and dry years and quantified the statistical relationship between September snow extent and Oct–Mar mean

593 discharge at the Copiapó River at Pastillo gauging station. Daily mean streamflow records at the Copiapó River at Pastillo
 594 station (Fig. 9) reveal distinct seasonal patterns between wet and dry years. During the wet year (Oct 2017 - Mar 2018),
 595 streamflow exhibits a sustained increase from October to December, peaking at 5.7 m³/s in December, followed by a gradual
 596 decline through January and a secondary pulse in February (5.3 m³/s). In contrast, the dry year (Oct 2013 - Mar 2014) shows
 597 substantially lower spring flows, with mean streamflow averaging only 1.0 m³/s during October-November, and the most
 598 significant increases concentrated in December (1.3 m³/s) and February (2.2 m³/s). Overall, the wet year delivered 3.4 times
 599 more water than the dry year, with the greatest relative differences occurring in November (+471%) and December (+343%).

600



601

602 **Figure 9. Daily (a) and monthly mean (b) streamflow at Copiapó River at Pastillo for wet (2017-2018, blue) and dry (2013-2014, red)**
 603 **years. Data Source: Dirección General de Aguas, Chile.**

604 These contrasting patterns between individual years reflect broader interannual variability in snowpack accumulation, ice
 605 bodies and melt dynamics. To quantify this relationship at the watershed scale, we incorporated a new analysis linking MODIS-
 606 derived snow permanence in September, at the onset of the melt season, with mean river discharge observed between October
 607 and March (Fig. 10). This relationship shows a robust correlation ($R^2 = 0.58$; $r = 0.76$; $N = 22$), confirming that seasonal snow
 608 permanence acts as a dominant short-term cryospheric water resource in the Copiapó watershed. However, the unexplained

609 variance (42%) suggests that other cryospheric components, such as rock glaciers, debris-covered ice, and possibly permafrost,
 610 may contribute to streamflow, particularly during low-snow years or late in the melt season when these longer-residence-time
 611 ice storage become more significant. The integration of these empirical hydrological results into the NCI framework validates
 612 that sub-watersheds with higher NCI values not only exhibit stronger cryospheric signatures in their topography but also
 613 display measurable hydrological responses consistent with multiple meltwater sources operating at different temporal scales.
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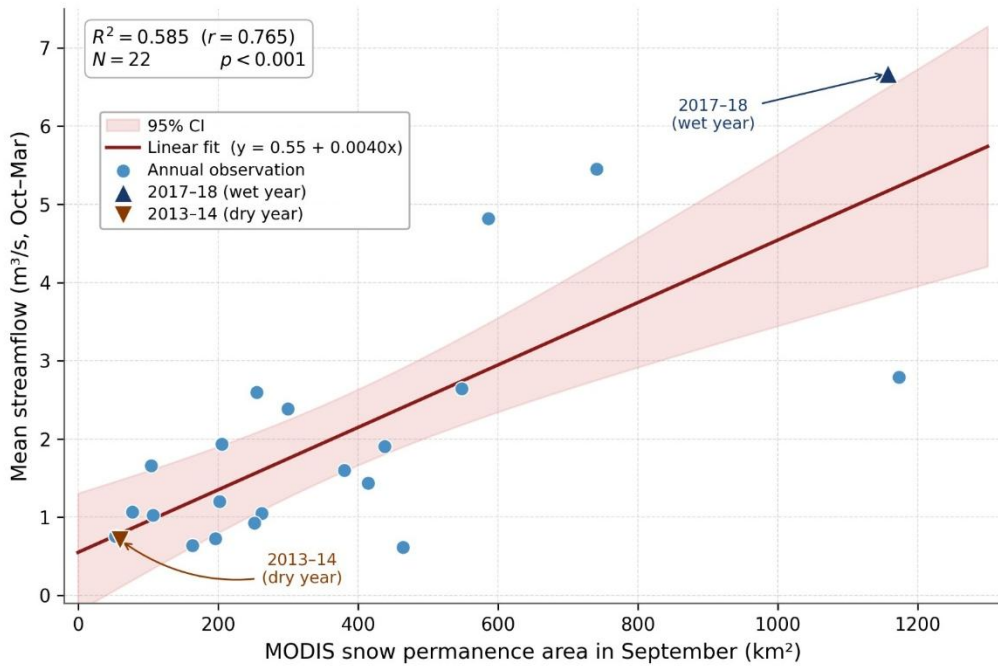
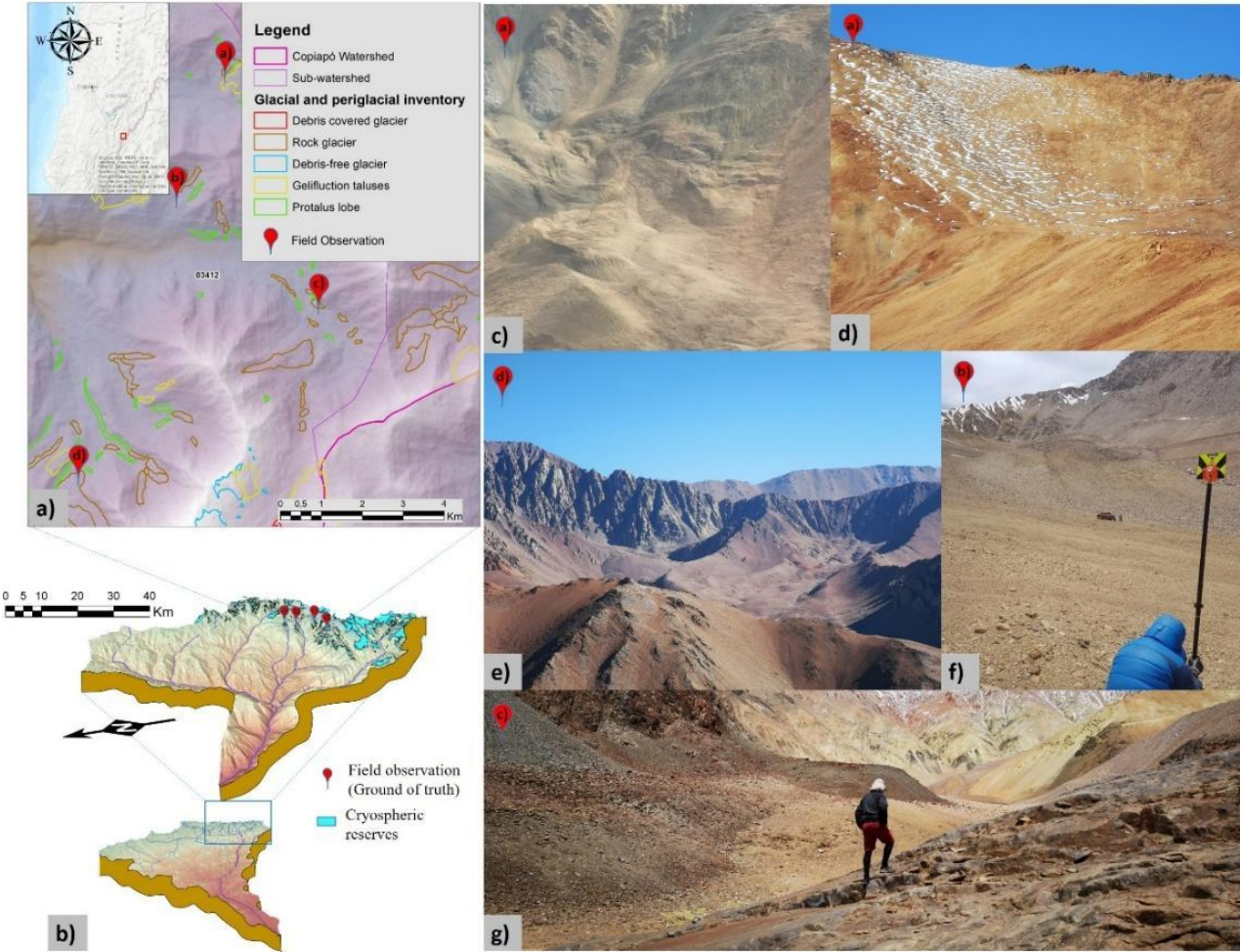


Figure 10. September snow-covered area vs. October-March mean discharge at Copiapó River at Pastillo (2000-2021, N = 22). Linear regression (red line) with 95% confidence interval (shaded). $R^2 = 0.58$; $r = 0.76$; $p < 0.001$. Sources: MODIS MOD10A1 and DGA, Chile.

620 4.3 Field survey observations of cryospheric landforms.

621 The fieldwork has yielded observations of cryospheric landforms in the Del Potro and Montosa River sub-watersheds (Fig. 11
 622 and 12). The Del Potro Glacier (Fig. 12d) and the surrounding periglacial environment of the Del Potro river sub-watershed
 623 (03412; Fig. 11) were clearly identified during field surveys. In the Montosa River sub-watershed, the Maranceles Glacier
 624 (Fig. 12c) was also observed. Additional field observations confirmed the presence of well-developed rock glaciers,
 625 gelifluction taluses, and protalus lobes in the Del Potro river watershed (03412) (Fig. 11c–g).
 626 Glacial landforms are preferentially distributed on south-facing slopes, as indicated by their aspect, which are generally
 627 associated with reduced incoming solar radiation. In contrast, periglacial cryospheric landforms such as protalus lobes and
 628 gelifluction taluses are more frequently found on exposed slopes with steeper gradients and variable aspects, reflecting

629 differences in topographic setting rather than direct measurements of snow accumulation or thermal conditions. Debris-free
 630 and debris-covered glaciers are primarily located in topographic depressions and high-elevation headwaters. In contrast,
 631 protalus lobes and gelifluction taluses are predominantly distributed on steep slopes and debris-covered sectors. Overall,
 632 periglacial cryospheric landforms display a wider spatial distribution across the watershed than glaciers.



633
 634 **Figure 11. Field observations of cryospheric landforms as the ground truthing. a) Map of cryospheric landforms distribution, b) 3D**
 635 **view of the watershed, c) Rock glacier of the Del Potro River (03412); d) Gelifluction taluses of the Del Potro River (03412); e) Rock**
 636 **glacier of the Del Potro River (03412); f) Rock glacier of the Del Potro River (03412); g) Panorama of different periglacial cryospheric**
 637 **landforms of the Del Potro River (03412). Basemap: Esri, Maxar, Earthstar Geographics, and the GIS User Community.**

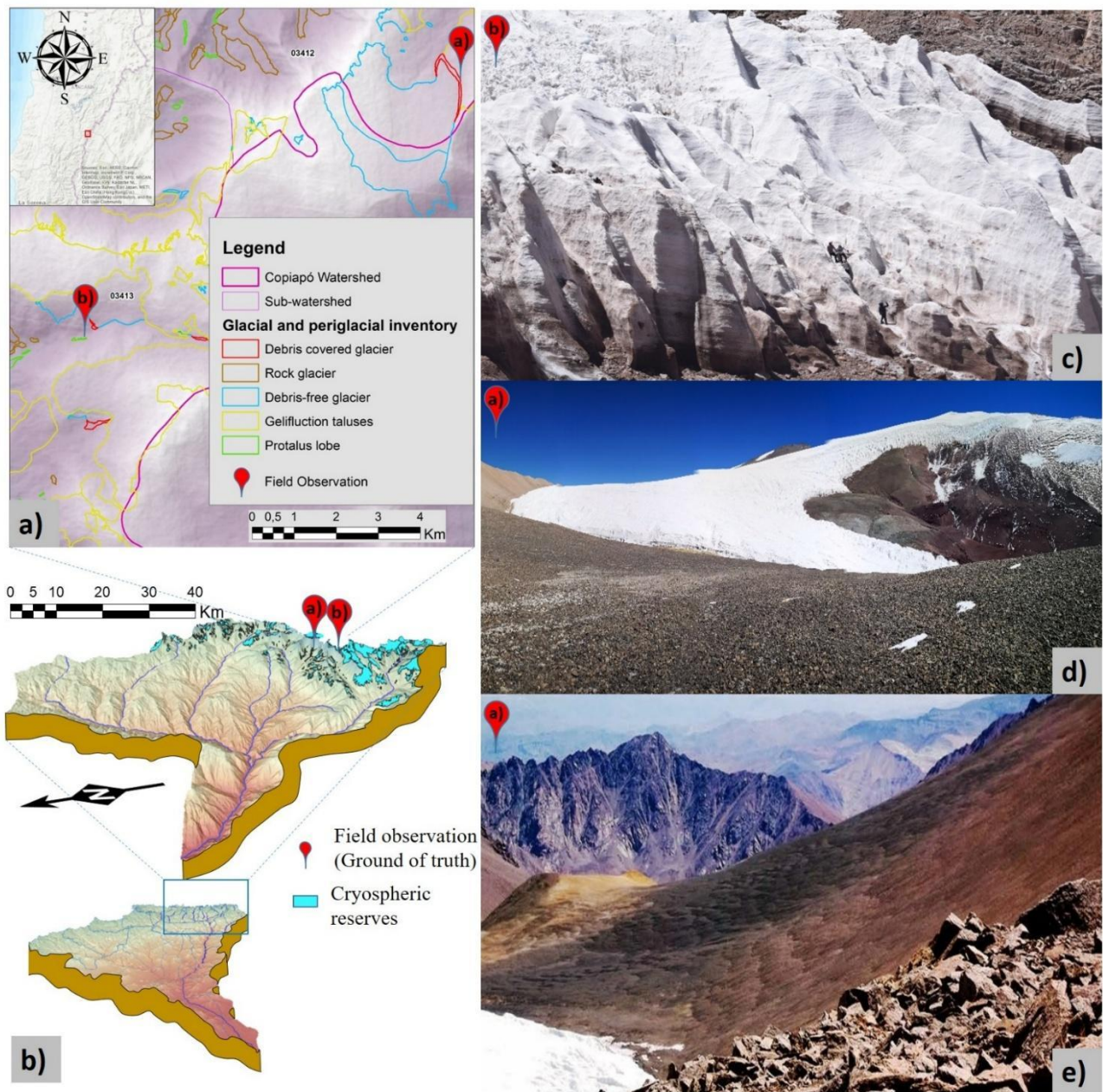


Figure 12. Field observations of cryospheric landforms as the ground truthing. a) Map of cryospheric landforms distribution, b) 3D view of the watershed, c) Del Potro debris-free glacier of the Del Potro River (03412), d) Maranceles glacier of the Montosa River (03413), e) Gelifluction taluses of the periglacial environment of the Del Potro River (03412). Basemap: Esri, Maxar, Earthstar Geographics, and the GIS User Community.

643 4.4 Geophysical results

644 4.4.1 Gelifluction Taluses

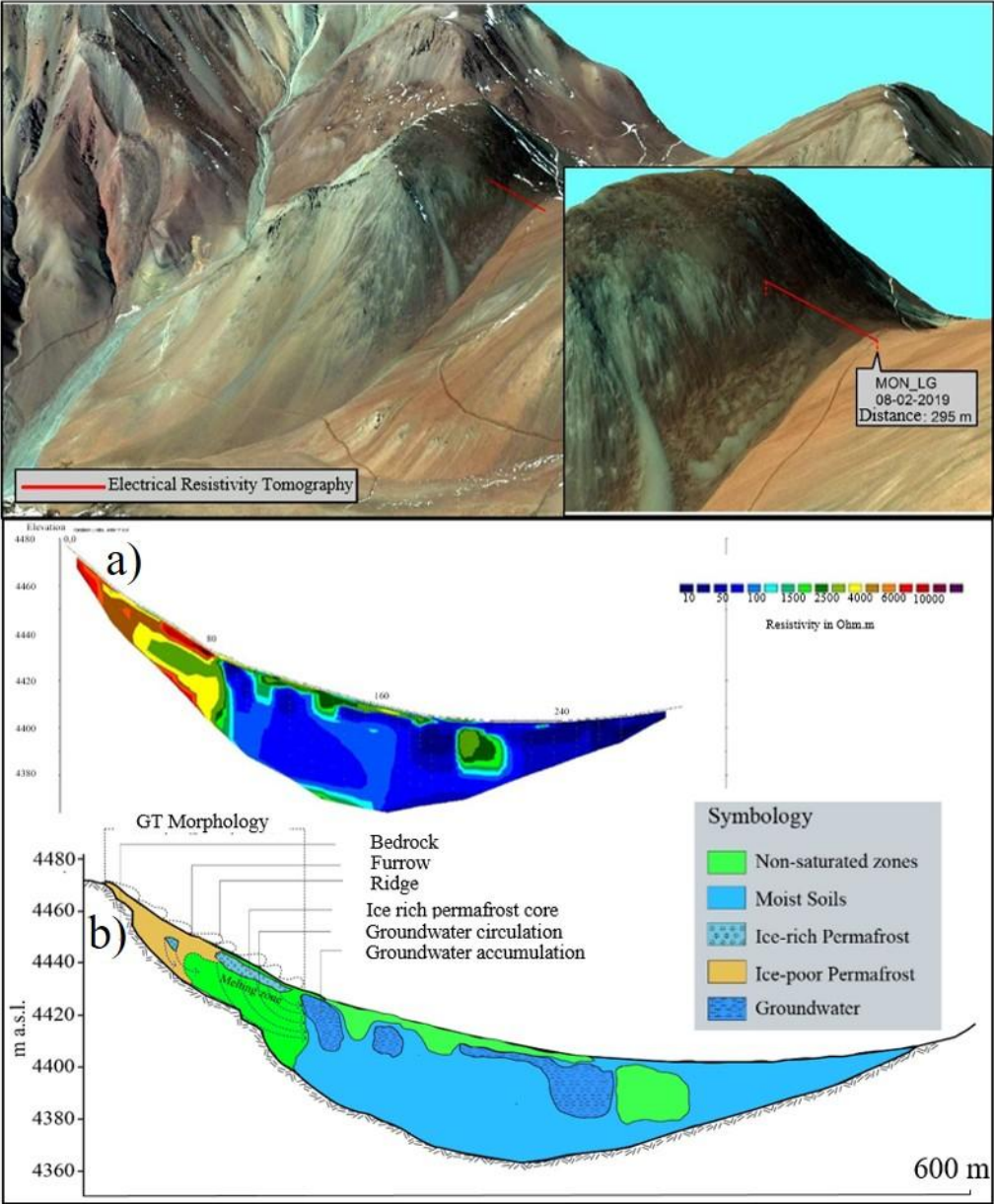
645 Electrical resistivity tomography (ERT) results (Fig. 13) provides the basis for interpreting the internal structure of the
646 gelifluction taluses. The lobate surface morphology characteristic of this cryospheric landform (Fig. 5 a, b), together with the
647 location and orientation of the tomographic profile (Figure 5 c, d), is clearly reflected in the geophysical response. The ERT
648 profile extends over 295 m and crosses a sector of the landform displaying a well-developed pattern of freeze–thaw-related
649 lobulations, starting at 4,430 m a.s.l. in the upstream sector of the Montosa sub-watershed. This geomorphological pattern
650 coincides with a high-resistivity zone (HRZ), with resistivity values ranging from approximately 1,000 to 8,000 Ωm .
651 According to comprehensive reviews and field studies on mountain permafrost, such resistivity values are typical of ice-rich
652 frozen ground, where pore water is largely transformed into electrically resistive ice (Hauck and Kneisel, 2008). Electrical
653 resistivity is known to increase by several orders of magnitude upon freezing, particularly in coarse-grained periglacial
654 materials, making ERT a robust tool for detecting ice-bearing permafrost bodies.

655 The HRZ delineates an ice-rich permafrost body with an estimated two-dimensional horizontal extension of approximately 20
656 m^2 within the tomographic section. Comparable resistivity ranges and internal structures have been documented in ice-rich
657 permafrost landforms investigated with ERT, including rock glaciers, talus slopes and other periglacial slope systems (Kneisel
658 et al., 2008; Hilbich et al., 2009). The association between lobate slope morphology and underlying frozen ground with
659 significant ice content is consistent with the established understanding of gelifluction as a process operating on seasonally
660 thawed active layers overlying permafrost, where ice lenses and frozen substrates exert a first-order control on deformation
661 patterns (Matsuoka, 2001).

662 Downstream of 4,430 m a.s.l., the HRZ terminates abruptly at the frontal sector of the last lobulations. From this point, a low-
663 resistivity zone (LRZ) dominates, with resistivity values between 50 and 200 Ωm . This sharp contrast is interpreted as the
664 transition from frozen ground to unfrozen, water-saturated sediments. Similar resistivity ranges have been reported for
665 unfrozen debris, taliks and shallow groundwater zones underlying permafrost terrain, where electrolytic conduction through
666 liquid water strongly reduces bulk resistivity (Hauck and Kneisel, 2008).

667 The spatial juxtaposition of the HRZ and LRZ suggests a direct subsurface connection between the ice-rich permafrost body
668 and an underlying groundwater system. Studies combining ERT and seismic methods demonstrate that such resistivity
669 contrasts commonly mark permafrost boundaries and zones of hydrological interaction, where water infiltration and ice loss
670 can be detected geophysically (Hauck et al., 2008; Mewes et al., 2017). In particular, combined geophysical monitoring has
671 shown that increases in liquid water content beneath or adjacent to frozen ground produce a pronounced decrease in resistivity,
672 even when ice loss is spatially limited. Based on the vertical extent of the HRZ, the ERT data allow for the estimation of an
673 average ice-rich core thickness of approximately 15 m (Fig. 13). This thickness falls within the range reported for ice-bearing

674 periglacial slope landforms investigated using geophysical methods, where frozen layers commonly extend several tens of
675 meters below the surface depending on material properties and thermal conditions (Kneisel et al., 2008; Hauck and Kneisel,
676 2008).



677
678 **Figure 13. Results of the electrical resistivity tomography (ERT) over the gelifluction taluses cryospheric landforms in the Montosa**
679 **Sub-watershed. a) Electrical resistivity tomography, b) Subsurface interpretation of resistivity test. Basemap: Esri, Maxar,**
680 **Earthstar Geographics, and the GIS User Community.**

4.4.2 Rock Glacier

A second ERT survey was conducted across a representative active rock glacier in the Montosa sub-watershed (Fig. 14). The tomographic sections reveal a complex internal structure characterized by laterally continuous high-resistivity zones (HRZ), which are interpreted as ice-rich permafrost bodies embedded within a coarse debris matrix, a configuration commonly reported for active rock glaciers worldwide (Kneisel et al., 2008).

Resistivity values exceeding approximately 20–50 $\text{k}\Omega\cdot\text{m}$ dominate the central and frontal sectors of the landform, indicating the presence of massive or highly ice-saturated material. Comparable resistivity ranges have been documented in ice-rich rock glaciers in the Andes and other high-mountain regions, where electrical resistivity contrasts are primarily controlled by ground-ice content rather than lithological variability (Kneisel et al., 2008; Pavoni et al., 2021). These high-resistivity anomalies are spatially coherent with the surface morphology of the rock glacier, including its lobate structure and well-defined frontal slope, supporting a close correspondence between geomorphological expression and subsurface ice distribution.

Beneath the HRZ, lower-resistivity zones (LRZ; $<1\text{--}5 \text{ k}\Omega\cdot\text{m}$) are observed, which are interpreted as unfrozen or partially saturated debris layers and/or underlying bedrock. The relatively sharp transition between high- and low-resistivity domains suggests a distinct boundary between ice-rich permafrost and unfrozen substrate, a feature widely reported in ERT studies of active rock glaciers (Kneisel et al., 2008).

The ERT-derived geometry indicates an average ice-rich layer thickness on the order of 20–30 m, with local maxima exceeding 35 m in the central part of the rock glacier body. These thicknesses fall within the range reported for active rock glaciers in the dry Andes based on geophysical surveys and regional assessments (Brenning, 2005; Azócar and Brenning, 2010; Halla et al., 2021). The observed thickness values provide an independent observational constraint for the thickness estimates derived from the viscoplastic creep inversion (Cicoira et al., 2021) based on DInSAR surface velocities.

Importantly, the spatial extent of the HRZ aligns closely with the mapped rock glacier polygon, reinforcing the validity of the inventory-based delineation and confirming that the geophysically detected ice body corresponds to the geomorphological unit considered in the volume calculations. The ERT results therefore serve as ground-based validation for both the presence of ice-rich permafrost and the order of magnitude of the modeled thicknesses applied to rock glaciers within the Copiapó watershed, consistent with previous multi-method approaches combining geomorphology, geophysics, and kinematic observations.

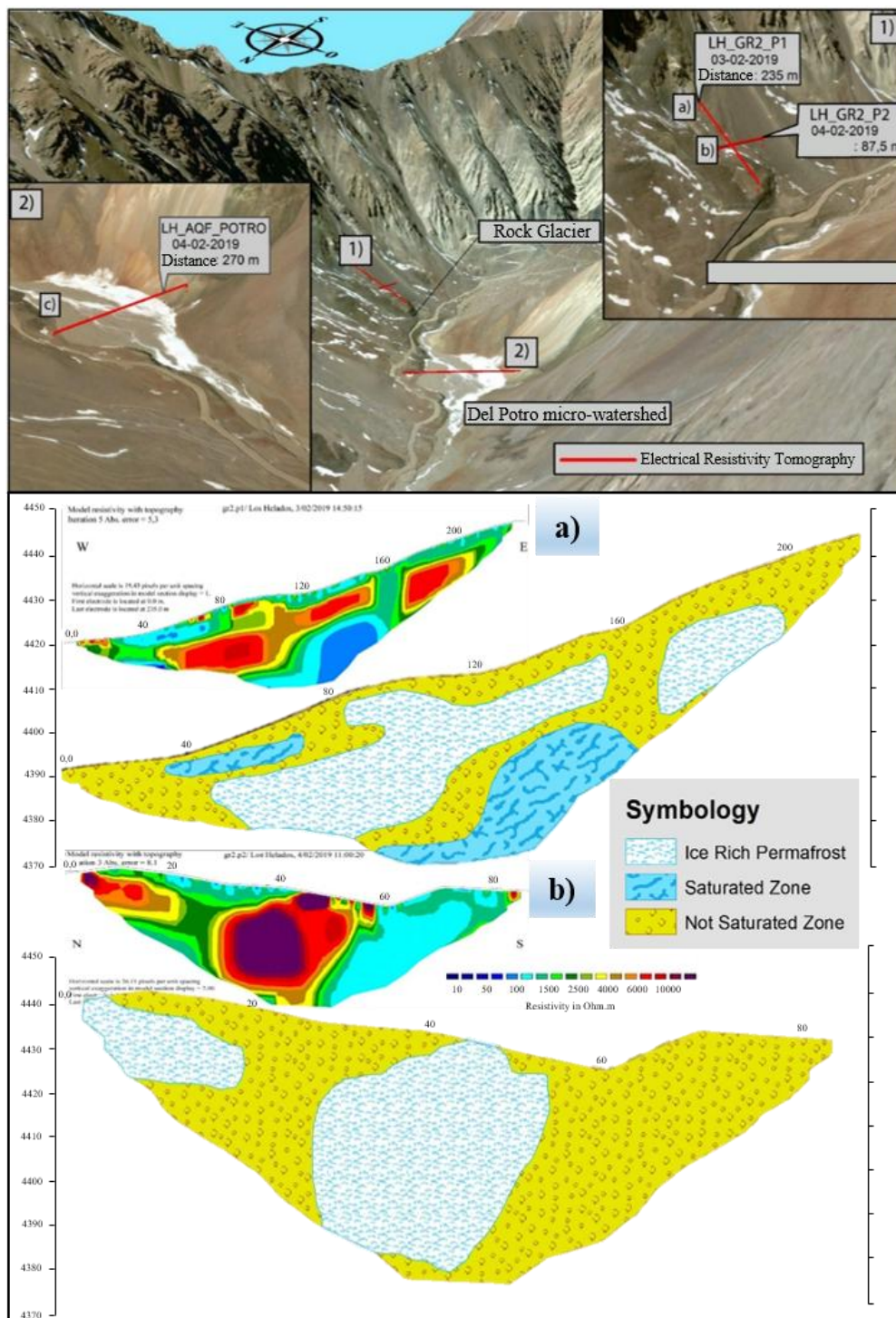


Figure 14. Results of the electrical resistivity tomography (ERT) over rock glacier cryospheric landforms in the Montosa Sub-watershed. Basemap: Esri, Maxar, Earthstar Geographics, and the GIS User Community. a) Longitudinal tomography, b) Cross-sectional tomography.

711 **4.5 Thickness estimations on periglacial landforms**

712 Surface displacement maps (Fig. 15) of a representative rock glacier in the Copiapó watershed show a coherent downslope
713 motion that is clearly confined to the rock glacier body, while surrounding terrain remains stable, supporting the reliability of
714 both the cryosphere inventory and the DInSAR processing. These surface velocities provide a direct kinematic constraint for
715 the viscoplastic thickness modelling of ice-rich permafrost landforms used in this study.

716 The modeled ice-rich layer thicknesses for the main periglacial cryospheric landforms (rock glaciers, protalus lobes, and
717 gelifluction slopes) and their relationship with DInSAR-derived surface velocities (Fig. 16) show that rock glaciers exhibit the
718 largest modeled thicknesses, with median values on the order of ~25–30 m and a wide interquartile range, reflecting both their
719 higher surface velocities and the greater mechanical efficiency of creep in these landforms. In contrast, protalus lobes show
720 intermediate thicknesses, typically below ~15 m, while gelifluction slopes are characterized by consistently thinner ice-rich
721 layers, generally <10 m. The scatter plot highlights a positive relationship between mean surface velocity and modeled
722 thickness across all periglacial cryospheric landforms, consistent with the viscoplastic framework in which higher deformation
723 rates require larger ice thicknesses to sustain basal shear stress under comparable slope conditions.

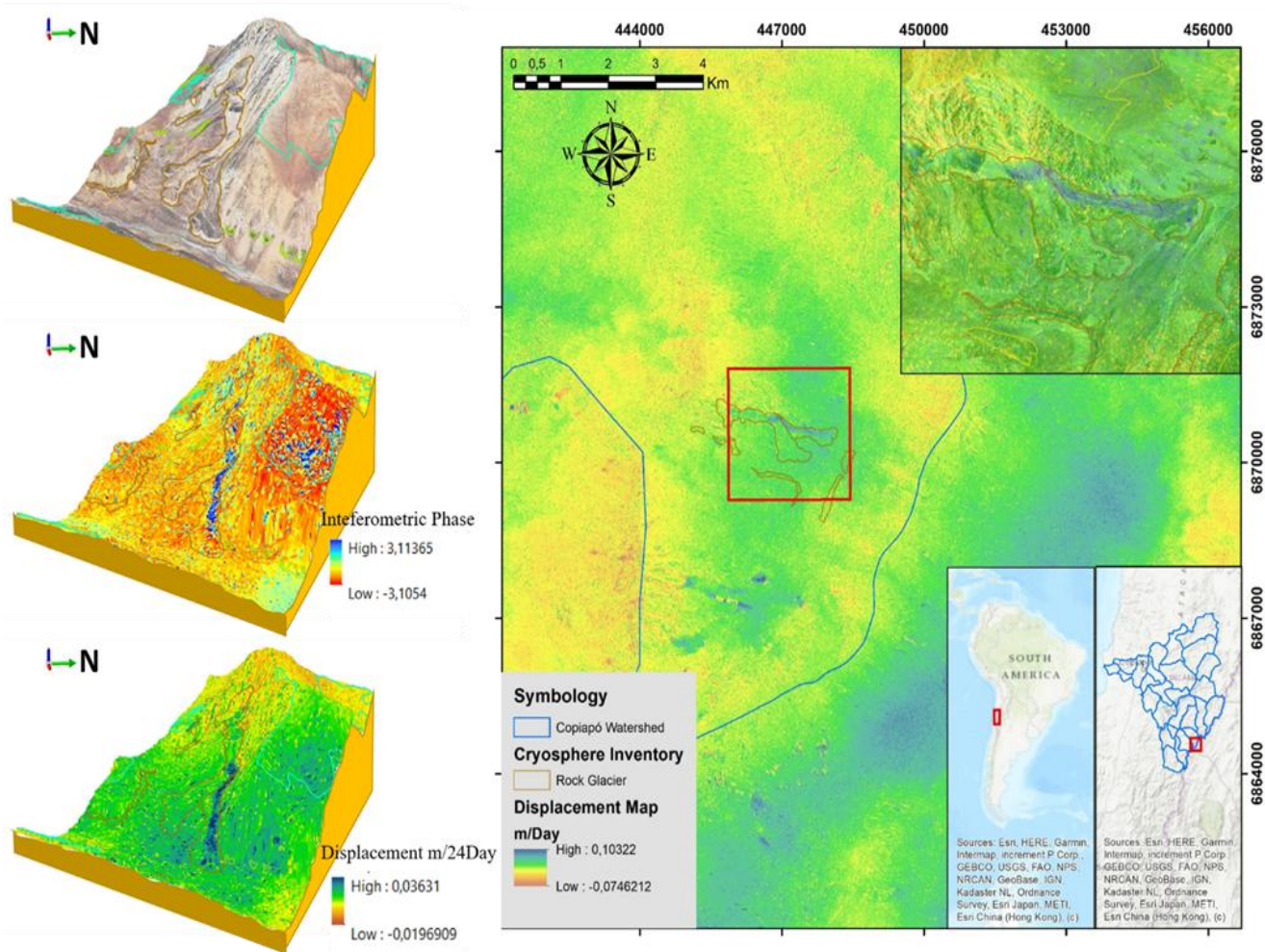


Figure 15. Example of DInSAR-derived surface displacement of a rock glacier in the Copiapó Watershed. Basemap: Esri, Maxar, Earthstar Geographics, and the GIS User Community.

Rock glaciers cluster at higher velocities and thicknesses, whereas protalus lobes and gelifluction slopes occupy lower-velocity regimes, with substantial overlap reflecting variability in slope, thermal state, and internal ice content. Despite this overlap, the three cryospheric landforms groups remain clearly distinguishable in thickness–velocity space, supporting the application of a unified viscoplastic model while preserving cryospheric landforms-specific structural differences. These results confirm that surface kinematics derived from DInSAR provide a physically meaningful constraint on subsurface ice thickness and justify their use for estimating ice volumes in ice-rich permafrost landforms across the Copiapó watershed. These thickness estimates are further contextualized by multi-decadal kinematic records from the Dry Andes showing that rock glaciers exhibit marked interannual velocity variability (Vivero et al., 2021), and by debris-covered glacier studies demonstrating that supraglacial debris exerts a strong control on the hydrological behavior of ice bodies in high-elevation Andean catchments

736 (Burger et al., 2019). The long-term hydrological relevance of these landforms is reinforced by projections indicating reduced
737 glacier meltwater contributions under future megadrought scenarios (Ayala, 2025), which elevate the importance of accurately
738 quantifying periglacial ice reserves such as those estimated here.

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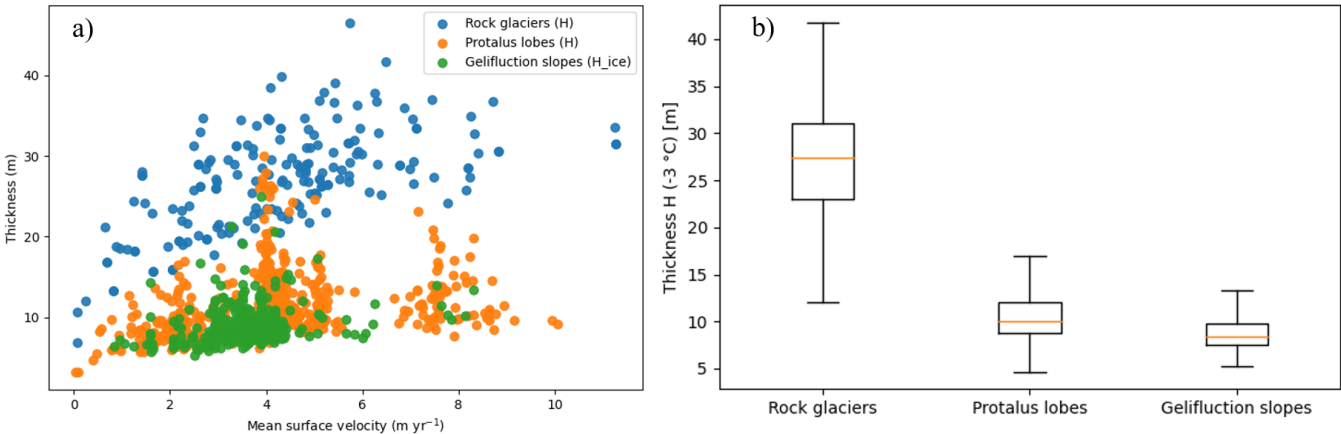


Figure 16. Modeled thickness–velocity relationship for periglacial cryospheric landforms of Copiapó Watershed.

4.6 Ice volumes, snow area and NCI for sub-watershed

The ice volumes, snow area and the NCI for each of the sub-watersheds that constitute the cryospheric landforms of the Copiapó River watershed have been calculated (Table 6). The watershed with the largest ice masses is the Montosa River watershed (03413), with 0.922 Gt, followed by the Del Potro River watershed (03412) with 0.913 Gt and thirdly the southernmost watershed, the Manflas River watershed (03420) with 0.371 Gt of ice. In terms of volume, the largest ice volumes in the entire watershed are found in debris-free glaciers, gelifluction taluses and rock glaciers (Fig. 17).

Following the mapping of snow, we observe that the surface area distribution of snow cover is uneven, with a higher density in the southern part and the northern end of the watershed. In the south, the watersheds with the largest snow permanence are the Manflas River (03420), with 107.06 km² of mean snow cover (2000-2024), the Montosa River (03413) with 75.06 km² and the Del Potro River (03412) with 53 km². In the north, the most important rivers are the Estero Come Caballos (03402) with 122.24 km², Figueroa River (03401) with 106.72 km², Quebrada Monardes (03400) with 89.63 km² and the Cachitos River (03403) with 88.34 km² (Table 6).

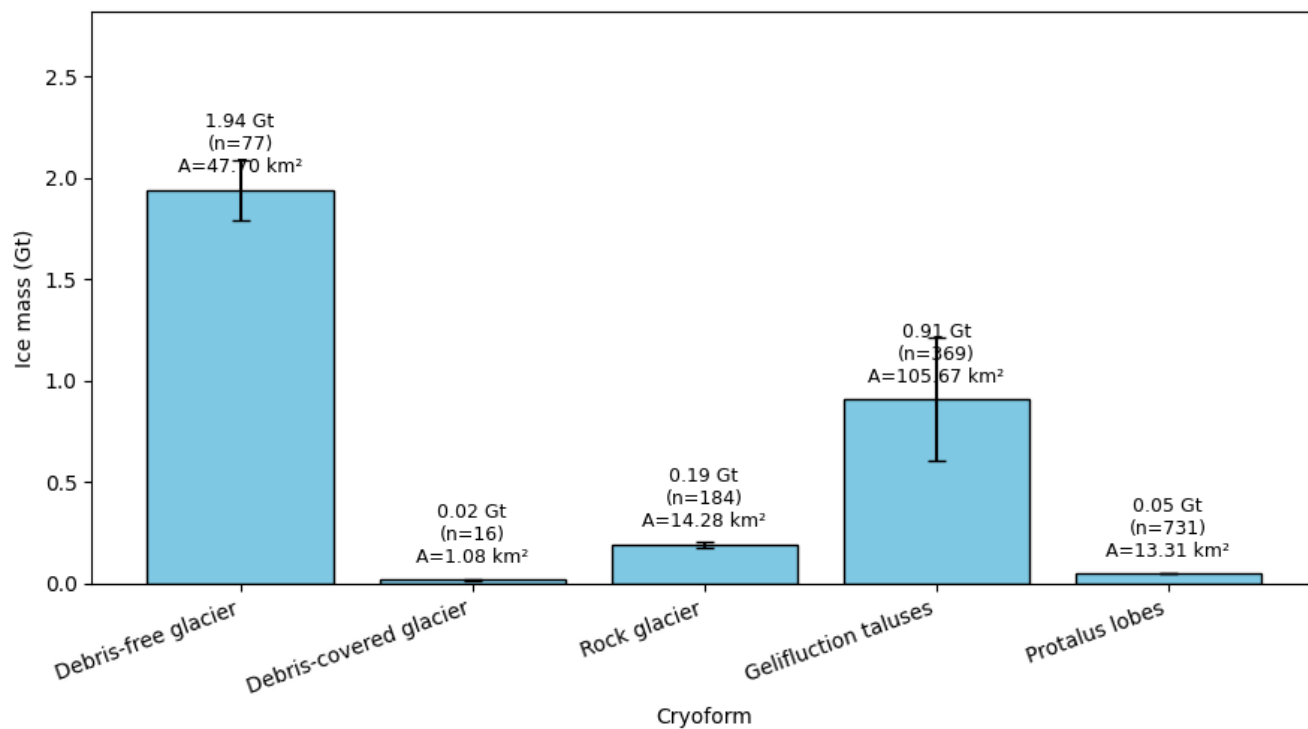


Figure 17. Total ice mass of the Copiapó Watershed.

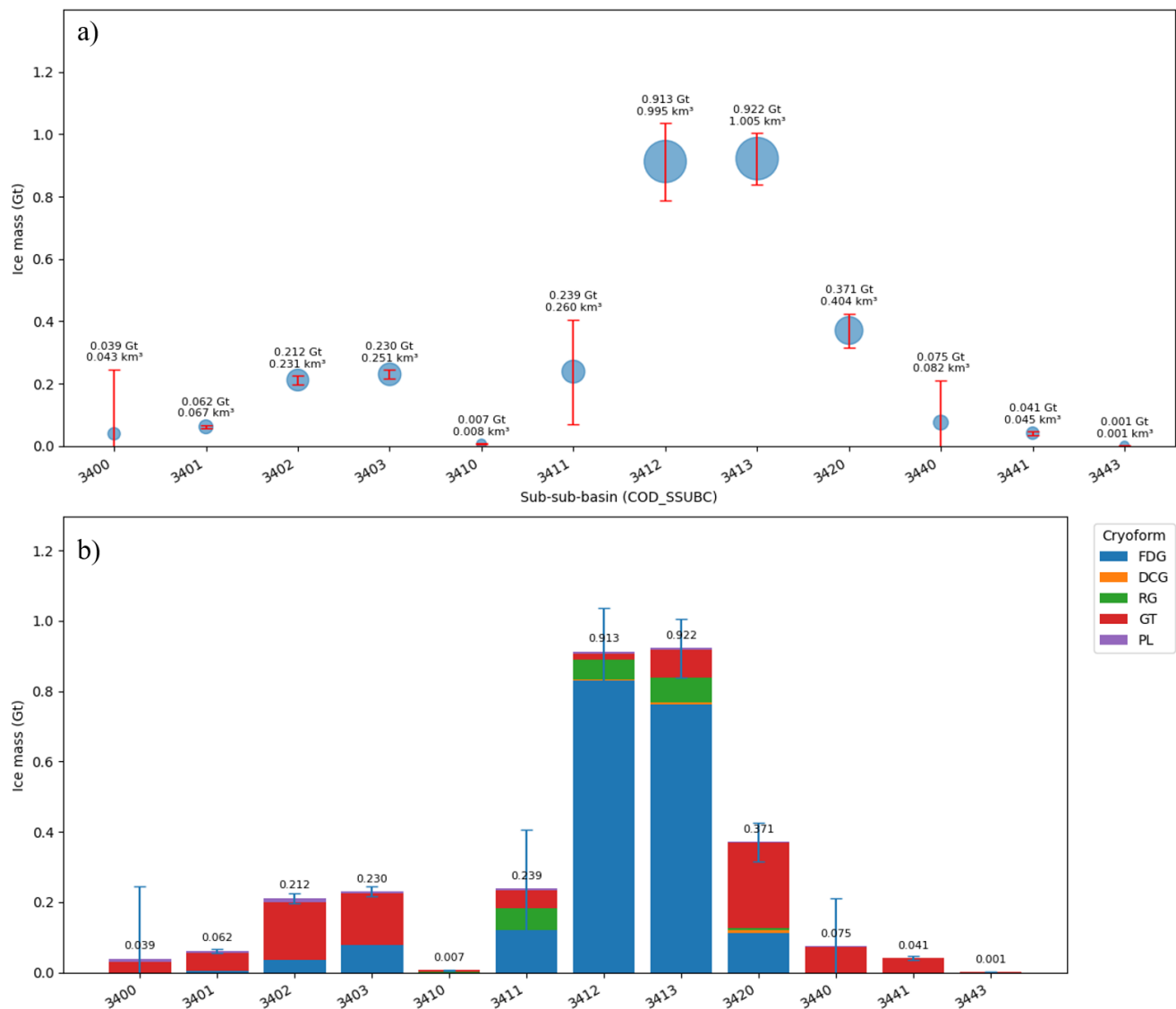


Figure 18. Total ice volumes of the Copiapó Watershed expressed both as a) total ice mass with associated uncertainty and b) as stacked contributions by cryospheric landforms.

The distribution of ice mass across cryospheric sub-watersheds of the Copiapó watershed (Fig. 18), reveal a highly heterogeneous spatial distribution of cryospheric landforms, with a small number of sub-watersheds concentrating the majority of the total ice mass.

The largest ice masses are found in the Montosa (COD_SSUBC 3413) and Del Potro (3412) sub-watersheds, each storing approximately ~0.9 Gt of ice. These watersheds clearly dominate the cryospheric landforms of the watershed and are characterized by the combined presence of debris-free glaciers, rock glaciers, and extensive ice-rich permafrost landforms. In

contrast, several sub-watersheds exhibit ice masses below 0.1 Gt, reflecting either limited cryospheric landforms extent or the dominance of thin periglacial landforms.

The stacked representation highlights that debris-free glaciers constitute the principal contribution to ice mass in the largest ice storage, whereas ice-rich permafrost cryospheric landforms, particularly rock glaciers and gelifluction taluses, represent a substantial fraction of the total ice mass in intermediate and smaller sub-watersheds. Protalus lobes contribute minor but non-negligible volumes in selected sub-watersheds, emphasizing their cumulative importance at the watershed scale.

Uncertainty ranges are largest in sub-watersheds dominated by periglacial cryospheric landforms, reflecting the combined effects of variability in surface velocity, thermal state, ice content, and mapped area. Nevertheless, the relative ranking of sub-watersheds by total ice mass remains robust, with Montosa and Del Potro consistently emerging as the primary cryospheric landforms within the Copiapó watershed.

Overall, these results demonstrate that cryospheric ice mass is strongly concentrated in a limited number of sub-watersheds, while many others contain comparatively small but non-negligible ice volumes distributed among different cryospheric landforms. However, ice mass alone does not fully capture the hydrological relevance of a watershed, as it does not account for the capacity to regenerate water resources through snowfall. This motivates the integration of cryospheric landforms (ice mass) and snow (Permanence) into a unified metric, introduced in the following section as the Normalized Cryospheric Index (NCI).

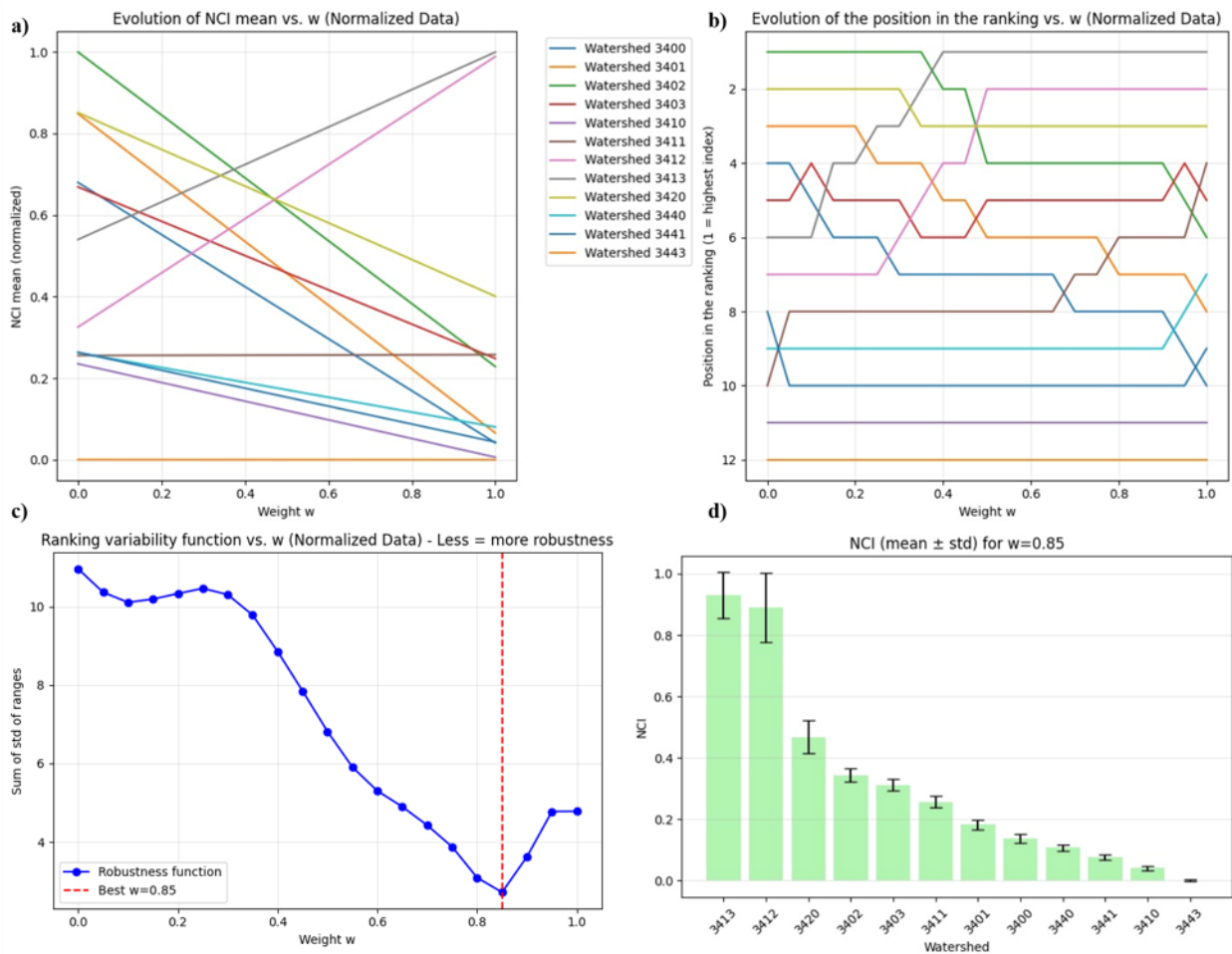
The Normalized Cryospheric Index (NCI) behavior in relation to the relative weighting of ice volume (V) and snow occurrence (S), along with the stability of sub-watershed rankings under uncertainty (Fig. 19) shows how the choice of weight (w) influences the mean value of the NCI and, ultimately, the final ranking of the sub-watersheds.

The evolution of the mean NCI value of each sub-watershed as w varies from zero to one (Fig. 19a). Each line represents a different sub-watershed, so that the slopes and crosses between lines allow us to appreciate which sub-watershed are favored by a greater relevance of ice volume V (when w tends to 1) or snow cover S (when w tends to 0). This preliminary behavior shows, on the one hand, the sensitivity of each subwatershed to weighting and, on the other hand, the possible dominance of one variable over the other in certain w zones.

The position in the ranking that each subwatershed occupies for each w value (Fig. 19b) shows that the horizontal lines extend from the first position (1) to the last. It highlights consistently privileged-ranked subwatersheds regardless of w , in contrast to unstably ranked subwatersheds. This early insight into the stability of the ranking compellingly underscores the urgency of undertaking a rigorous robustness analysis.

The ranking variability function, defined as the sum (or average) of the standard deviations of the position of each sub-watershed in the different simulations (Fig. 19c) quantifies the overall stability of the ranking for each w . A marked minimum is seen around $w=0.85$, indicating that, at that point, the ranking of watersheds is less sensitive to the uncertainty of the data

796 of V and S. In other words, assigning that weight in the combination of variables leads to a more consistent ranking in terms
 797 of robustness.
 798 The mean NCI (with its error bars) for each watershed when $w=0.85$, the optimal value detected with the optimization function
 799 (Fig. 19d), shows that the vertical axis reflects the average magnitude of the index, while the error bars indicate the standard
 800 deviation derived from the Monte Carlo simulations. Thus, the visualization offers a clear identification of the watersheds with
 801 the highest NCI values, while simultaneously exposing the degree of uncertainty surrounding these estimates. This result
 802 confirms the relevance of choosing $w=0.85$ and provides a detailed view of the dispersion of each watershed under such a
 803 scenario.
 804



805
 806 **Figure 19. Analysis of NCI and robustness of the ranking as a function of w weighting (a) Evolution of the mean NCI for each sub-**
 807 **watershed as w varies between 0 and 1 (b) Position in the ranking of each sub-watershed as a function of w (c) Ranking variability**
 808 **function (sum of the standard deviations of the positions), where the minimum at $w \approx 0.85$ indicates the maximum stability of the**
 809 **ranking (d) Mean NCI $\pm$ standard deviation for $w=0.85$, evidencing the magnitude and dispersion of the index in each sub-watershed**
 810 **under the most robust condition.**

811

812 The calculated NCI, demonstrated that the sub-watershed with the highest NCI was the by the Montosa River watershed

813 (03413) (NCI = 0.93), followed by the Del Potro River watershed (03412) (NCI = 0.89), Manflas River watershed (03420)

814 (NCI = 0.47), Estero Come Caballos watershed (03402) (NCI = 0.34) and Cachitos River watershed (03403) (NCI = 0.31).

815 These are the sub-watershed with the largest cryospheric landforms. In the northern region, the Estero Come Caballos sub-

816 watershed, it was highlighted by the index primarily attributed to its persistent snow cover.

817

818 **Table 6 Summary of results by sub-watershed of cryospheric landforms in Gt, snow in km² and NCI index.**

Code of the sub-watershed	Cryospheric landforms gigatons	Cryospheric landforms uncertainty	Snow in sq. km	NCI	NCI uncertainty
3400	0.039	0.006	89.63	0.14	0.01
3401	0.062	0.005	106.72	0.18	0.02
3402	0.212	0.014	122.24	0.34	0.02
3403	0.230	0.015	88.34	0.31	0.02
3410	0.007	0.001	43.90	0.04	0.01
3411	0.239	0.018	45.95	0.26	0.02
3412	0.913	0.124	53.02	0.89	0.11
3413	0.922	0.082	75.06	0.93	0.08
3420	0.371	0.055	107.06	0.47	0.05
3440	0.075	0.009	46.66	0.11	0.01
3441	0.041	0.006	46.79	0.08	0.01
3443	0.001	0.0005	19.76	0.00	0.00

5. Discussion

5.1 Cryospheric landforms and snow permanence in the Copiapó watershed

From the cryospheric landforms and snow occurrence mapping exercises (Figs. 4 and 8), we observed that the cryospheric landforms of the Copiapó watershed are predominantly located at altitudes above 4,500 m. This compares well with other regions in which snowfall has been identified to be controlled by altitude (Saydi and Ding, 2020). We also observed that cryospheric landforms occur in smaller, isolated regions ($< 0.25 \text{ km}^2$ in width) in the northern part of our research area. This corresponds with our own field observations, where most cryospheric landforms occur in smaller bodies (Fig. 4). This resulted in smaller, separated cryospheric landforms and surrounding the highest point (Figure 4 b and c). They host smaller cryospheric landforms, for example near the Cadillal and the Nevado Jotabeche, compared to the southern part of our study area. Because those northern cryospheric landforms are located in or in proximity of the SAAD (Abraham et al., 2020), snowfall and permanent snow occurrences are limited, leading to isolated cryospheric landforms (Fig. 4). The potential volume of water that is captured in cryospheric landforms in the sub-watersheds of the Copiapó watershed (Fig. 1) is thus strongly influenced by the South American Arid Diagonal (SAAD) that controls the distribution of the different types of cryospheric landforms (García et al., 2017). In our case, the proximity of the Copiapó watershed to the SAAD limits the occurrence of debris-free glaciers in the watershed north of 28.1° S latitude and limits the continuity of the cryosphere that is mainly attributed to cryospheric landforms associated with ice-rich permafrost, such as small rock glaciers, protalus lobes and predominantly gelifluction taluses.

In the more central domain in between Cadillal and Caserones no cryospheric landforms occur. This can be explained by the mechanical forcing of the South American low-level jet and the topographic blocking of the Andes (Insel et al., 2009). In the central sector, Mount Pissis exhibits greater elevation on the Argentinian side. Consequently, the prevailing northeasterly winds transporting moisture from the Pacific Ocean are obstructed by the Argentine portion of the mountain. This orographic interception results in localized snowfall on its Argentinian slopes, rather than allowing precipitation to be deflected by the comparatively lower-elevation Chilean topography. The topography north of Cadillal and south of Caserones is high enough to block these winds on the Chilean side, leading to snowfall and eventual cryospheric landforms in the southern regions in our research area. A last negative spatial correlation between the snowfall and the cryospheric landforms is the large snowfall anomaly in the northern area. The permanent snow occurrence is high, while not many cryospheric landforms are produced. The reason for this is the location of this region in the SAAD, resulting in melting or sublimation of the snow that does arrive in this area (Réveillet et al., 2020), meaning that the snow does not end up in cryospheric landforms.

From the different cryospheric landforms, we see that gelifluction taluses cover the largest surface area in the Copiapó watershed compared to other cryospheric landforms in our research areas (Garcia et al., 2017). Field observations also confirm that gelifluction taluses extend across significantly broader areas compared to other cryospheric landforms (Fig. 12). The limited understanding of gelifluction taluses has led to their exclusion from global cryospheric reserve inventories. Nevertheless, in assessing watershed ice volume, these taluses emerge as the most critical features for ice storage, surpassing

other formations in significance (Fig. 17). Furthermore, the ERT-derived thickness measurements across the gelifluction lobe reveal a minimum detectable thickness of 15 meters, which markedly contrasts with the range of 1 to 5 meters reported by Hilbich et al. (2022).

On a more regional scale, we observe that in the south of the watershed, debris-free glaciers are always in contact with gelifluction taluses. This is confirmed in the field (Fig. 12). In the north of the Copiapó watershed, gelifluction taluses occur more as isolated bodies. The topography of the northern area is controlled by Tertiary volcanoes (Kay et al., 1994). In general, these volcanoes exhibit gentler slopes than those of the Principal Cordillera (Harrington, 1961) and largely shape the topography of our southern research area. Given that gelifluction taluses require steep slopes to form (Fig. 2), they are more likely to develop in the southern sector, where the terrain is dominated by the steep relief of the Principal Cordillera unlike in the north, where this feature is absent. Moreover, in the Central Cordillera glacial cirques are abundantly present (Evans, 2006). Glacial cirques provide sheltered environments that shield snow from extreme weather conditions, facilitating its recrystallization into exposed glaciers and other cryospheric landforms. The distribution of rock glaciers and protalus lobes is likewise influenced by the presence of glacial cirques (Palacios et al., 2020), resulting in their predominant occurrence in the southern portion of our study area. In contrast, no exposed glaciers are found north of Copiapó Hill (Fig. 4). Instead, gelifluction taluses and protalus lobes allow the continuity of the cryospheric fringe of the watershed. The absence of debris-free glaciers in this region is directly related to the South American low-level jet and the topographic blocking of the Andes (Insel et al., 2009).

These DGA (2014) volumes were used to provide a magnitude check. At the Del Potro Glacier, the modeled volume is ~24% larger than the GPR-based estimate; at the Tronquitos Glacier, the difference is larger (~70%). This spread is within the expected uncertainty range for individual glacier inversion products in Farinotti et al. (2019), particularly for small glaciers with limited thickness constraints. The DInSAR-derived surface velocities of rock glaciers in the Copiapó watershed are systematically lower than those reported for rock glaciers at comparable elevations further south in the Dry Andes. Vivero et al. (2021) documented mean surface velocities of 0.26 m yr⁻¹ during 1956-1978 and up to 1.25-1.80 m yr⁻¹ during 2014-2019 for the Tapado, Las Tolas, and El Cachito rock glaciers at 30°S in the Coquimbo Region. In contrast, the rock glaciers of the Copiapó watershed (26°41'-28°24'S), as detected by Sentinel-1 DInSAR, exhibit velocities on the order of a few millimetres per year, one to two orders of magnitude slower. Two complementary factors explain this systematic difference. First, the Copiapó watershed lies within or immediately adjacent to the South American Arid Diagonal (SAAD), where extremely limited precipitation results in drier, colder permafrost conditions and reduced ground-ice temperatures relative to sites further south. Lower permafrost temperatures translate directly into higher ice viscosity and reduced creep rates under equivalent basal shear stresses, consistent with the strong temperature dependence of the Glen-type flow law applied in this study (Cicoira et al., 2021; Koenig et al., 2025). Second, the cryospheric landform genesis differs fundamentally between the two sites. The Tapado complex studied by Vivero et al. (2021) comprises glacial rock glaciers, landforms that originated from or remain connected to glacier ice and push moraines, and which therefore tend to be ice-rich and more dynamically active. The rock glaciers inventoried in the Copiapó watershed are predominantly cryogenic in origin, having developed through progressive

freezing of talus and debris-slope material under permafrost conditions, without a direct glacial ice source. Cryogenic rock glaciers characteristically exhibit lower ice contents, more heterogeneous debris-ice matrices, and slower creep rates than glacial forms (Monnier and Kinnard, 2015). The slower kinematics observed in the Copiapó watershed are therefore physically consistent with both the arid thermal regime imposed by proximity to the SAAD and the cryogenic origin of the inventoried rock glaciers, and do not reflect anomalous model behaviour in the viscoplastic thickness inversion.

The DGA (2014) GPR-derived ice volume estimates were used as an independent benchmark to evaluate the magnitude of the modeled glacier volumes (Table 7). For Del Potro Glacier, the modeled volume (0.362 km³) exceeds the GPR-based estimate (0.293 km³) by approximately 24%. In the case of Tronquitos Glacier, the discrepancy is larger, with the modeled volume (0.157 km³) being about 70% higher than the GPR-derived value (0.092 km³).

Despite these differences, the observed spread is consistent with the uncertainty ranges reported for individual glacier ice-thickness inversion products in Farinotti et al. (2019). This is particularly relevant for small glaciers, where limited field-based thickness constraints and complex geometries tend to amplify relative errors in volume estimates. Therefore, while the discrepancies, especially for Tronquitos, are non-negligible, they remain within the expected variability for regional-scale glacier volume modeling.

A relevant consideration regarding the ice volume estimates for debris-free glaciers in the Copiapó watershed concerns the difference in glacier area between the two inventories used. The regional inventory of García et al. (2017) maps a total of 1,907 ha of uncovered glaciers within the Copiapó watershed, whereas the Randolph Glacier Inventory (RGI) v6.0, which forms the basis for the Farinotti et al. (2019) thickness product used in this study, encompasses a larger glacier area for the same watershed. Because the Farinotti et al. (2019) volumes are computed over the RGI glacier polygons, which are more extensive than those delineated by García et al. (2017), the total ice volume attributed to debris-free glaciers in this study may overestimate the volume that would be obtained if the García et al. (2017) glacier boundaries were used instead. This discrepancy arises from methodological differences in glacier delineation, inventory date, and mapping criteria between the two datasets, and represents an inherent source of uncertainty in the volume estimates reported here. Future work should aim to reconcile these inventories and assess the sensitivity of the modeled ice volumes to the choice of glacier outlines, particularly for small and fragmented glacier bodies in the northern sector of the watershed.

Table 7. Comparison between modeled glacier ice volumes from Farinotti et al. (2019) and GPR-derived ice volumes reported by DGA (2014) for Del Potro and Tronquitos glaciers.

Glacier	RGI ID	Farinotti model volume (km ³)	DGA 2014 GPR volume (km ³)
Del Potro	1.715.087	0.362	0.293

931	Tronquitos	1.715.038	0.157	0.092
932				
933				
934				
935				
936				

937 The watershed with the best balance between cryospheric landforms and snow permanence is the Montosa River watershed
938 (03413) (Fig. 20), which contains high quantities of both cryospheric landforms and snow permanence. The field observations
939 of geomorphological features that are indicative of active cryospheric landforms, such as the subparallel terraces (Fig. 12e),
940 fit well with the mapped cryospheric landforms from our workflow. The mapped surface area and thickness are supported by
941 the observed surface area and thickness in the field. Lastly, debris-covered glaciers occur predominantly in areas with a
942 relatively continuous cryosphere (García et al., 2017). This is consistent with our findings that debris-covered glaciers are
943 exclusively located in the southern region of the study area, a zone characterized by a more extensive and contiguous
944 cryospheric distribution relative to the fragmented northern sectors. The interannual variability in glacier meltwater
945 contribution in high-elevation Andean catchments is strongly modulated by the insulating effect of supraglacial debris, which
946 attenuates ablation rates and delays the hydrological response of debris-covered glaciers relative to their clean-ice counterparts
947 (Burger et al., 2019). Complementing this, multi-decadal monitoring using aerial and satellite photogrammetry and UAV
948 surveys has demonstrated that rock glaciers in the Dry Andes exhibit significant kinematic variability, underscoring their
949 dynamic role in the cryospheric system of the region (Vivero et al., 2021).

950 **5.2 Classifying and prioritizing sub-watersheds in the Copiapó watershed**

951 Unlike rainfall-dominated catchments, the upper Copiapó watershed is strongly influenced by cryospheric processes, with its
952 headwaters characterized by the widespread presence of glaciers, rock glaciers, and ice-rich permafrost landforms. Cryospheric
953 elements are concentrated in the high-altitude sub-watersheds, where water storage and seasonal runoff generation are
954 controlled primarily by snow accumulation, snow persistence, and long-term ice storage rather than by direct rainfall inputs.
955 The Normalized Cryospheric Index (NCI) identifies three main sub-watersheds with a dominant cryospheric influence on the
956 hydrological functioning of the Copiapó system, based on the combined contribution of ice mass and long-term snow
957 persistence. The spatial distribution of cryospheric landforms shows a 97% overlap with high-probability permafrost zones
958 predicted by the global permafrost model of Gruber (2012), providing independent support for the relevance of ice-rich
959 permafrost as a key component of water storage in the watershed. Together, these observations justify classifying the upper
960 Copiapó as a cryosphere-influenced watershed, in which cryospheric processes exert a first-order control on water availability.
961 The critical sub-watersheds include: the Montosa River (03413), Del Potro River (03412) and Manflas River (03420) (Table
962 6). Their spatial distribution corresponds to zones of maximum snow permanence and the most extensive cryospheric
963 continuity within the Copiapó watershed (Figs. 6 and 8). These high indexes thus directly correlate with the cryospheric

landforms and snowfall occurrences. For example, enhanced snowfall is concentrated in the Cerro del Potro sub-watershed (03412), where the largest debris-free glacier of the Copiapó watershed is located (Fig. 8c). This area hosts El Potro Glacier, the most extensive binational glacier in northern Chile (García et al., 2017), indicating a strong coupling between snow accumulation, glacier development, and long-term ice storage. As a result, the Cerro del Potro sub-watershed ranks among the highest NCI values, reflecting the combined importance of snow (permanence) and cryospheric landforms (ice mass). This combination has the potential to control the presence of water resources downstream when global temperatures continue to rise (Flores et al., 2018). Notably, recent projections for the Southern Andes indicate that glacier meltwater contributions may decline substantially during future megadrought conditions (Ayala, 2025), reinforcing the urgency of identifying and protecting cryospheric sub-watersheds such as those ranked highest by the NCI. With this study, we detected the sub-watersheds that should be prioritized. In wet years, snow controls the watershed hydrology in the northern part of the Copiapó watershed. This means that in these locations infrastructure needs to be put in place to capture the larger amounts of run-off that will arrive with increasing warmer temperatures, such as the Estero Come Caballos sub-watershed (03402), with the fourth place of the NCI. In times of water scarcity, we identified that the sub-watersheds of the Montosa River (03413), Del Potro River (03412), and Manflas River watershed (03420) will contribute most of the flow to the Copiapó River.

Given the observed relationship between snowfall and cryospheric landforms distribution, prioritizing the Copiapó watershed is essential to safeguard its water resources. We propose a systematic framework to evaluate and rank sub-watersheds according to preliminary assessments of cryospheric landforms and resource availability. From this ranking (Table 6), the main target, with the highest NCI is the sub-watershed of Montosa River (03413). The second target is located towards the south, the Del Potro River (03412). The third target is the Manflas River watershed (03420) and finally the last targets are Estero come Caballos (03402) and Cachitos River (03403) watershed (Table 6). These five sub-watersheds constitute the basis for continuous water production in the Copiapó watershed during dry years, because they host 85.09 % of the cryospheric landforms ice volume of the Copiapó watershed. From these four sub-watersheds, the Montosa River watershed (03413) maintains the best balance between snow and cryospheric landforms (Fig. 20). We have shown that with this hydrological approach, the NCI allows us to infer that the sub-watershed with the highest hydrological importance is the Montosa River watershed (03413). This index also allowed us to observe that the watersheds of the northern region are high in snow cover but are less enriched in cryospheric landforms. During years of substantial snowfall, management efforts should prioritize northern sub-watersheds, such as the Estero Come Caballos Sub-watershed (03402). This suggests that northern watersheds exhibit a limited ability to develop cryogenic niches, as snow in these areas is primarily lost through melting or sublimation. A potential explanation for the lack of cryospheric landforms in northern watersheds lies in their geomorphological characteristics, which may hinder the accumulation and preservation of ice.

Based on the calculation of the NCI, we have been able to identify that the Estero Come Caballos watershed (03402) is the most strategic for snow accumulation, even surpassing the watersheds in the southern part of the Copiapó watershed in terms of snow permanence (Fig. 20).

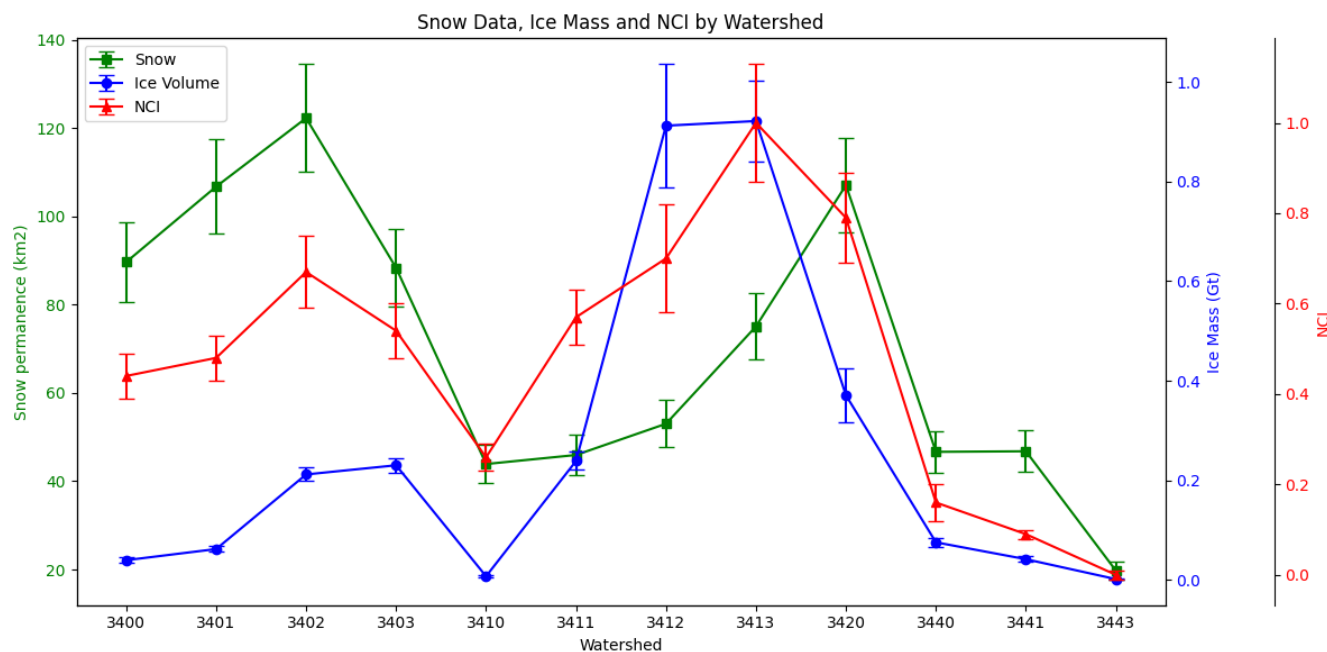


Figure 20. Summary graph of the normalized indexes for snow, cryospheric landforms and the cryosphere for each sub-watershed of the Copiapó River watershed.

5.3 Impact on watershed management

We have analyzed 12 cryospheric sub-watersheds in the Copiapó watershed and showed that there are some that should be prioritized in watershed management strategy. The methodological framework developed and presented in this study can be extended to assess the comprehensive hydrological potential of arid watersheds in high-elevation Andean regions of Chile and other globally distributed arid zones with a focus on cryospheric contributions. In addition, this novel sub-watershed indexing method is fundamental to identify those watersheds with an important snow, glacial and periglacial component that are suitable for inclusion in the watershed protection frameworks that are implemented in the water management strategies of each watershed in Chile. Given the persistent progression of climate change, it is imperative to integrate water reserves sourced from cryospheric watersheds into comprehensive policy-making frameworks governing hydrological regulation.

Implementing and integrating such cryosphere inventories will also allow other regions to preserve their water resources. Currently, the Chilean national cryospheric inventories distinguish between glacier and periglacial landforms. Debris-free and debris-covered glaciers are comprehensively included in the National Glacier Inventory of Chile, whereas within the periglacial environment, only rock glaciers are systematically inventoried. As a result, ice volumes potentially stored in other ice-rich periglacial landforms, such as protalus lobes and gelifluction taluses, are not explicitly accounted for in existing national inventories, despite being addressed in this study. Increasing the variety of cryospheric landforms would facilitate more accurate estimations of the volumes of ice in each sub-watershed, especially in northern Chile where the gelifluction taluses and protalus lobes dominate the cryospheric landforms.

1017 Identifying the strategic position of these sub-watersheds does not imply guaranteeing constant runoff, but rather highlights
1018 those sub-watersheds with a higher potential to sustain streamflow under low-snowfall or dry-year conditions. The runoff
1019 analysis shows that, during dry years, discharge is strongly reduced but not suppressed, and remains seasonally coupled to
1020 melt processes originating in high-elevation cryospheric environments. Sub-watersheds ranked as strategic by the NCI
1021 concentrate larger cryospheric landforms and higher snow permanence, which together contribute to buffering runoff
1022 variability and maintaining baseflow during periods of limited snowfall. The extraction of precious metals has consumed large
1023 quantities of water that was stored in cryospheric landforms (Brenning, 2008). A lack of comprehensive understanding
1024 regarding the spatial distribution and volumetric potential of cryospheric water reserves has contributed to the degradation of
1025 cryospheric landforms critical for supplying freshwater to downstream populations residing at lower elevations. These sub-
1026 watersheds are predominantly cryospheric rather than pluvial in character, indicating that their hydrological regime is mainly
1027 influenced by meltwater processes associated with glaciers, ice-rich periglacial landforms, and seasonal snow permanence,
1028 particularly during the melt season. When the potential volume of water in each sub-watershed is understood, policy and
1029 infrastructure can be adapted to maintain those cryospheric components. Prioritizing the extraction of water from cryospheric
1030 landforms and implementing snow capture techniques is essential to ensure the sustainable development and enhanced
1031 productivity of the examined region, as well as the arid northern regions of Chile. These sub-watersheds may therefore play
1032 an important role in the prevention of water disasters greater than those already known in Chile. The methodological basis
1033 developed in this study can be a key-tool for developing a law to protect Chile's cryosphere and its associated water resources
1034 when adopted in future water management strategy plans.

1035 **6. Conclusions**

1036 **6.1 Main findings of the paper**

1037 We developed a reproducible workflow that maps glacial and periglacial cryospheric landforms and seasonal snow, estimates
1038 their ice volumes with a unified physics-based framework, and integrates both into a Normalized Cryospheric Index (NCI)
1039 that ranks sub-watersheds with explicit Monte Carlo uncertainty propagation. Applied to the Copiapó watershed, the workflow
1040 classified 12 of 27 sub-watersheds as cryospheric; these contain 97 % of the inventoried landforms and overlap 97 % with
1041 high-probability permafrost predicted by Gruber (2012), providing independent support for the classification.
1042 The empirical link between September snow extent and October–March discharge at Copiapó River at Pastillo ($R^2 = 0.58$; $r =$
1043 0.76 ; $N = 22$) confirms that snow permanence exerts a first-order control on seasonal runoff. The 42 % of unexplained variance
1044 is consistent with longer-residence cryospheric storage (rock glaciers, debris-covered ice, ice-rich permafrost) sustaining
1045 baseflow during low-snow years, which is precisely the contribution the NCI is designed to capture.
1046 Ice mass is strongly concentrated: the Montosa (03413; 0.92 Gt) and Del Potro (03412; 0.91 Gt) sub-watersheds alone dominate
1047 the basin's cryospheric storage, with Manflas (03420; 0.37 Gt) third. Debris-free glaciers are the principal contributor in these

top reservoirs, whereas ice-rich permafrost landforms—particularly gelifluction taluses and rock glaciers—account for a substantial and, at the basin scale, cumulatively important fraction, especially in intermediate and smaller sub-watersheds. The NCI consistently ranks Montosa (0.93), Del Potro (0.89) and Manflas (0.47) as the primary priority sub-watersheds, which together concentrate ~71 % of the ice mass inventoried across the 12 cryospheric sub-watersheds (Table 6). These three are the basins most likely to sustain runoff during droughts and should be prioritized in the national water-resources strategy. Estero Come Caballos (03402; NCI = 0.34) and Cachitos (03403; NCI = 0.31) emerge as secondary, snow-driven basins: in high-snowfall years their hydrological productivity rises, so management there should focus on capturing increased runoff rather than on protecting long-term ice. Adding these two raises the prioritized share of inventoried ice mass to ~85 %. This dual behaviour—southern, ice-rich basins safeguarding dry-year baseflow versus northern, snow-rich basins driving wet-year runoff—is the central management insight of the index.

6.2 Relevant topics detected

Three findings have implications beyond the Copiapó basin. First, gelifluction taluses and protalus lobes—which dominate the arid Atacama cryosphere by extent and store non-negligible ice—are currently excluded from Chile's national periglacial inventory, which systematizes only rock glaciers. Their omission means national ice-storage assessments likely underestimate the periglacial reserve in northern Chile. Second, our ERT surveys returned ice-rich cores ~15 m thick in a gelifluction talus and 20–35 m in an active rock glacier, an order of magnitude greater than thicknesses assumed for analogous landforms in parts of the literature, reinforcing that periglacial ice is being undercounted and warranting targeted geophysical campaigns. Third, the very low DInSAR surface velocities (mm yr^{-1}), one to two orders of magnitude slower than rock glaciers further south in the Dry Andes, are physically consistent with the colder, drier permafrost imposed by proximity to the South American Arid Diagonal and with the predominantly cryogenic (rather than glacigenic) origin of these landforms; this constrains how kinematic models should be transferred across the Andes.

6.3 Limitations and outlook

The NCI is calibrated for the Copiapó watershed and requires complete resource and reserve data to be transferred elsewhere. The main quantified uncertainty in the snow term is the assumed ~10 % accuracy of the MODIS fractional-snow-cover product under ideal conditions; deriving a spatially explicit FSC uncertainty for high-mountain terrain is the priority next step to refine the index. Given projections of reduced glacier meltwater under future Southern Andes megadroughts (Ayala, 2025), identifying and protecting high-NCI cryospheric sub-watersheds is increasingly urgent. The framework provides technical, objective input for the DGA, water-monitoring boards and groundwater communities, and offers a methodological basis for integrating glacial and periglacial reserves into watershed-protection policy—and potentially into a legal instrument to protect Chile's cryosphere and its associated water resources.

1078 **7. Open science statement**

1079 The data generated for this project will become available upon publication. The ice volume calculation data can be downloaded
1080 via this link: [<https://doi.org/10.5281/zenodo.21628142>]. The glacial and periglacial inventory for the Atacama region can be
1081 downloaded via this link: [<https://doi.org/10.5281/zenodo.21628105>]. The raster image for historical snow data can be
1082 downloaded via this link [<https://doi.org/10.5281/zenodo.21628032>].

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1091 **9. References**

1092 Abraham, E. M., Rodríguez, M. G. L., Rubio, M. C., Guida-Johnson, B., Gomez, L. V., and Rubio, C.: Disentangling the
1093 concept of "South American Arid Diagonal", *J. Arid Environ.*, 175, 104089, <https://doi.org/10.1016/j.jaridenv.2019.104089>,
1094 2020.

1095 Ahn, S. R. and Kim, S. J.: Assessment of integrated watershed health based on the natural environment, hydrology, water
1096 quality, and aquatic ecology, *Hydrol. Earth Syst. Sci.*, 21, 5583–5602, <https://doi.org/10.5194/hess-21-5583-2017>, 2017.

1097 Ayala, A., Pellicciotti, F., Macdonell, S., McPhee, J., Vivero, S., Campos, C., and Egli, P.: Modelling the hydrological response
1098 of debris-free and debris-covered glaciers to present climatic conditions in the semiarid Andes of central Chile, *Hydrol.*
1099 *Process.*, 30, 4036–4058, <https://doi.org/10.1002/hyp.10971>, 2016.

1100 Ayala, Á., Farías-Barahona, D., Huss, M., Pellicciotti, F., McPhee, J., and Farinotti, D.: Glacier runoff variations since 1955
1101 in the Maipo River Basin, semiarid Andes of central Chile, *The Cryosphere*, 14, 2005–2027, [https://doi.org/10.5194/tc-14-](https://doi.org/10.5194/tc-14-2005-2020)
1102 [2005-2020](https://doi.org/10.5194/tc-14-2005-2020), 2020.

1103 Ayala, Á., Muñoz-Castro, E., Farinotti, D., Farías-Barahona, D., Mendoza, P. A., MacDonell, S., McPhee, J., Vargas, X., and
1104 Pellicciotti, F.: Less water from glaciers during future megadroughts in the Southern Andes, *Commun. Earth Environ.*, 6, 860,
1105 <https://doi.org/10.1038/s43247-025-02845-6>, 2025.

1106 Azócar, G. F. and Brenning, A.: Hydrological and geomorphological significance of rock glaciers in the dry Andes, Chile
1107 (27°–33°S), *Permafrost Periglac. Process.*, 21, 42–53, <https://doi.org/10.1002/ppp.669>, 2010.

1108 Bahr, D. B., Meier, M. F., and Peckham, S. D.: The physical basis of glacier volume-area scaling, *J. Geophys. Res. Solid Earth*,
1109 102, 20355–20362, <https://doi.org/10.1029/97JB01696>, 1997.

1110 Basuki, T. M., Nugroho, H. Y. S. H., Indrajaya, Y., Pramono, I. B., Nugroho, N. P., Supangat, A. B., Indrawati, D. R., Savitri,
1111 E., Wahyuningrum, N., Purwanto, Cahyono, S. A., Putra, P. B., Adi, R. N., Nugroho, A. W., Auliyani, D., Wuryanta, A.,
1112 Riyanto, H. D., Harjadi, B., Yudilastyantoro, C., and Simarmata, D. P.: Improvement of integrated watershed management in
1113 Indonesia for mitigation and adaptation to climate change: a review, *Sustainability*, 14, 9997,
1114 <https://doi.org/10.3390/su14169997>, 2022.

1115 Beniston, M. and Stoffel, M.: Assessing the impacts of climatic change on mountain water resources, *Sci. Total Environ.*, 493,
1116 1129–1137, <https://doi.org/10.1016/j.scitotenv.2013.11.122>, 2014.

1117 Beniston, M., Farinotti, D., Stoffel, M., Andreassen, L. M., Coppola, E., Eckert, N., Fantini, A., Giacona, F., Hauck, C., Huss,
1118 M., Huwald, H., Lehning, M., López-Moreno, J. I., Magnusson, J., Marty, C., Morán-Tejeda, E., Morin, S., Naaïm, M.,
1119 Provenzale, A., and Vincent, C.: The European mountain cryosphere: a review of its current state, trends, and future challenges,
1120 *The Cryosphere*, 12, 759–794, <https://doi.org/10.5194/tc-12-759-2018>, 2018.

1121 Bolch, T. and Marchenko, S.: Significance of glaciers, rock glaciers, and ice-rich permafrost in the Northern Tien Shan as
1122 water towers under climate change conditions, in: *Proceedings of the Workshop Assessment of Snow-Glacier and Water*
1123 *Resources in Asia*, 28–30 November 2006, Almaty, Kazakhstan, 199–211, 2006.

1124 Bórquez, R., Larraín, S., Polanco, R., and Urquidi, J.: Glaciares chilenos. Reservas estratégicas de agua dulce para la sociedad,
1125 los ecosistemas y la economía, Chile Sustentable, LOM ediciones, Santiago, Chile, 124 pp., 2006.

1126 Bozkurt, D., Rojas, M., Boisier, J. P., and Valdivieso, J.: Climate change impacts on hydroclimatic regimes and extremes over
1127 Andean basins in central Chile, *Hydrol. Earth Syst. Sci. Discuss.*, <https://doi.org/10.5194/hess-2016-690>, 2017.

1128 Brenning, A.: Climatic and geomorphological controls of rock glaciers in the Andes of central Chile: combining statistical
1129 modelling and field mapping, PhD Thesis, Humboldt University, Berlin, Germany, 2005.

1130 Brenning, A.: The impact of mining on rock glaciers and glaciers: examples from central Chile, in: *Darkening Peaks: Glacier*
1131 *Retreat, Science, and Society*, edited by: Orlove, B. S., Wiegandt, E., and Luckman, B., University of California Press,
1132 Berkeley, USA, 196–205, 2008.

1133 Burger, F., Ayala, Á., Farías, D., Shaw, T. E., MacDonell, S., Brock, B., McPhee, J., and Pellicciotti, F.: Interannual variability
1134 in glacier contribution to runoff from a high-elevation Andean catchment: understanding the role of debris cover in glacier
1135 hydrology, *Hydrol. Process.*, 33, 214–229, <https://doi.org/10.1002/hyp.13354>, 2019.

1136 Casassa, G., Haeberli, W., Jones, G., Kaser, G., Ribstein, P., Rivera, A., and Schneider, C.: Current status of Andean glaciers,
1137 *Global Planet. Change*, 59, 1–9, <https://doi.org/10.1016/j.gloplacha.2006.11.013>, 2007.

1138 Clarke, J.: Antiquity of aridity in the Chilean Atacama Desert, *Geomorphology*, 73, 101–114,
1139 <https://doi.org/10.1016/j.geomorph.2005.06.008>, 2006.

1140 Cicoira, A., Marcer, M., Gärtner-Roer, I., Bodin, X., Arenson, L. U., and Vieli, A.: A general theory of rock glacier creep
 1141 based on in-situ and remote sensing observations, *Permafrost Periglac. Process.*, 32, 139–153,
 1142 <https://doi.org/10.1002/ppp.2090>, 2021.

1143 Dirección General de Aguas (DGA): Estimación de volúmenes de hielo mediante sondeos de radar en zonas Norte, Central y
 1144 Sur, Informe Final, S.I.T. No. 338, Ministerio de Obras Públicas, Santiago, Chile, available at:
 1145 <https://bibliotecadigital.ciren.cl/handle/20.500.13082/32675> (last access: 1 June 2025), 2014.

1146 Evans, I. S.: Local aspect asymmetry of mountain glaciation: a global survey of consistency of favoured directions for glacier
 1147 numbers and altitudes, *Geomorphology*, 73, 166–184, <https://doi.org/10.1016/j.geomorph.2005.07.009>, 2006.

1148 Farinotti, D., Brinkerhoff, D. J., Clarke, G. K. C., Fürst, J. J., Frey, H., Gantayat, P., Gillet-Chaulet, F., Girard, C., Huss, M.,
 1149 Leclercq, P. W., Linsbauer, A., Machguth, H., Martin, C., Maussion, F., Morlighem, M., Mosbeux, C., Pandit, A., Portmann,
 1150 A., Rabatel, A., and Andreassen, L. M.: How accurate are estimates of glacier ice thickness? Results from ITMIX, the Ice
 1151 Thickness Models Intercomparison eXperiment, *The Cryosphere*, 11, 949–970, <https://doi.org/10.5194/tc-11-949-2017>, 2017.

1152 Farinotti, D., Huss, M., Fürst, J. J., Landmann, J., Machguth, H., Maussion, F., and Pandit, A.: A consensus estimate for the
 1153 ice thickness distribution of all glaciers on Earth, *Nat. Geosci.*, 12, 168–173, <https://doi.org/10.1038/s41561-019-0300-3>, 2019.

1154 Favier, V., Falvey, M., Rabatel, A., Praderio, E., and Lopez, D. H.: Interpreting discrepancies between discharge and
 1155 precipitation in high-altitude area of Chile's Norte Chico region (26–32°S), *Water Resour. Res.*, 45, W02424,
 1156 <https://doi.org/10.1029/2008WR006802>, 2009.

1157 Ferri, L., Dussaillant, I., Zalazar, L., Masiokas, M. H., Ruiz, L., Pitte, P., Gargantini, H., Castro, M., Berthier, E., and Villalba,
 1158 R.: Ice mass loss in the central Andes of Argentina between 2000 and 2018 derived from a new glacier inventory and satellite
 1159 stereo-imagery, *Front. Earth Sci.*, 8, 530997, <https://doi.org/10.3389/feart.2020.530997>, 2020.

1160 Flores, B., García, A., and Ulloa, C.: Evolución espacial y temporal de glaciares descubiertos en la Región de Atacama, Chile,
 1161 XV Congreso Geológico Chileno, 742–745, 2018.

1162 García, A., Ulloa, C., Amigo, G., Milana, J. P., and Medina, C.: An inventory of cryospheric landforms in the arid diagonal of
 1163 South America (High Central Andes, Atacama Region, Chile), *Quaternary Int.*, 438, 4–19,
 1164 <https://doi.org/10.1016/j.quaint.2017.04.033>, 2017.

1165 Garreaud, R., Boisier, J. P., Rondanelli, R., Montecinos, A., Sepúlveda, H. H., and Veloso-Águila, D.: The central Chile mega
 1166 drought (2010–2018): a climate dynamics perspective, *Int. J. Climatol.*, 40, 421–439, <https://doi.org/10.1002/joc.6219>, 2020.

1167 Gascoin, S., Kinnard, C., Ponce, R., Lhermitte, S., Macdonell, S., and Rabatel, A.: Glacier contribution to streamflow in two
 1168 headwaters of the Huasco River, Dry Andes of Chile, *The Cryosphere*, 5, 1099–1113, <https://doi.org/10.5194/tc-5-1099-2011>,
 1169 2011.

1170 Gruber, S.: Derivation and analysis of a high-resolution estimate of global permafrost zonation, *The Cryosphere*, 6, 221–233,
 1171 <https://doi.org/10.5194/tc-6-221-2012>, 2012.

1172 Hall, D. K., Riggs, G. A., Salomonson, V. V., DiGirolamo, N. E., and Bayr, K. J.: MODIS snow-cover products, *Remote Sens.*
 1173 *Environ.*, 83, 181–194, [https://doi.org/10.1016/S0034-4257\(02\)00095-0](https://doi.org/10.1016/S0034-4257(02)00095-0), 2002.

1174 Halla, C., Blöthe, J. H., Tapia, C., Trombotto, D., Hilbich, C., Hauck, C., and Schrott, L.: Ice content and interannual water
 1175 storage changes of an active rock glacier in the dry Andes of Argentina, *The Cryosphere*, 15, 1187–1213,
 1176 <https://doi.org/10.5194/tc-15-1187-2021>, 2021.

1177 Hanssen, R. F.: *Radar Interferometry: Data Interpretation and Error Analysis*, Springer, Dordrecht, the Netherlands,
 1178 <https://doi.org/10.1007/0-306-47633-9>, 2001.

1179 Harrington, H. J.: Geology of parts of Antofagasta and Atacama provinces, northern Chile, *AAPG Bull.*, 45, 169–197,
 1180 <https://doi.org/10.1306/0BDA6332-16BD-11D7-8645000102C1865D>, 1961.

1181 Hauck, C., Bach, M., and Hilbich, C.: A four-phase model to quantify subsurface ice and water content in permafrost regions
 1182 based on geophysical data sets, in: *Proceedings of the Ninth International Conference on Permafrost*, edited by: Hinkel, K. M.
 1183 and Kane, D. L., Institute of Northern Engineering, University of Alaska Fairbanks, USA, Vol. 1, 675–680, 2008.

1184 Hilbich, C., Hauck, C., Mollaret, C., Wainstein, P., and Arenson, L. U.: Towards accurate quantification of ice content in
 1185 permafrost of the Central Andes – Part 1: geophysics-based estimates from three different regions, *The Cryosphere*, 16, 1845–
 1186 1872, <https://doi.org/10.5194/tc-16-1845-2022>, 2022.

1187 Hilbich, C., Marescot, L., Hauck, C., Loke, M. H., and Mäusbacher, R.: Applicability of electrical resistivity tomography
 1188 monitoring to coarse blocky and ice-rich permafrost landforms, *Permafrost Periglac. Process.*, 20, 269–284,
 1189 <https://doi.org/10.1002/ppp.641>, 2009.

1190 Hu, Y., Liu, L., Huang, L., Zhao, L., Wu, T., Wang, X., and Cai, J.: Mapping and characterizing rock glaciers in the arid
 1191 Western Kunlun Mountains supported by InSAR and deep learning, *J. Geophys. Res. Earth Surf.*, 128, e2023JF007206,
 1192 <https://doi.org/10.1029/2023JF007206>, 2023.

1193 Insel, N., Poulsen, C. J., and Ehlers, T. A.: Influence of the Andes Mountains on South American moisture transport,
 1194 convection, and precipitation, *Clim. Dynam.*, 35, 1477–1492, <https://doi.org/10.1007/s00382-009-0637-1>, 2009.

1195 Izquierdo, T., Rivera, A.-L., Galeano, Á., Gallardo, D., Salas, V., Aparicio, O., Buylaert, J.-P., Ruiz, F., and Abad, M.:
 1196 Historical catastrophic floods at the southern edge of the Atacama Desert: a multi-archive reconstruction of the Copiapó River
 1197 extreme events, *Global Planet. Change*, 236, 104411, <https://doi.org/10.1016/j.gloplacha.2024.104411>, 2024.

1198 Jaiswal, R. K., Ghosh, N. C., Galkate, R. V., and Thomas, T.: Multi criteria decision analysis (MCDA) for watershed
 1199 prioritization, *Aquat. Procedia*, 4, 1553–1560, <https://doi.org/10.1016/j.aqpro.2015.02.201>, 2015.

1200 Jara, F. M., Lagos-Zúñiga, M., Fuster, R., Mattar, C., and McPhee, J.: Snow processes and climate sensitivity in an arid
 1201 mountain region, northern Chile, *Atmosphere*, 12, 520, <https://doi.org/10.3390/atmos12040520>, 2021.

1202 Jones, D., Harrison, S., Anderson, K., and Betts, R.: Mountain rock glaciers contain globally significant water stores, *Sci. Rep.*,
 1203 8, 2834, <https://doi.org/10.1038/s41598-018-21244-w>, 2018.

1204 Pan, F., Jiang, L., Wang, G., Pan, J., and Huang, J.: MODIS daily cloud-gap-filled fractional snow cover dataset of the Asian
 1205 Water Tower from 2000 to 2022, *Earth Syst. Sci. Data*, 16, 2501–2523, <https://doi.org/10.5194/essd-16-2501-2024>, 2024.

1206 Kay, S. M., Mpodozis, C., Tittler, A., and Cornejo, P.: Tertiary magmatic evolution of the Maricunga mineral belt in Chile,
 1207 *Int. Geol. Rev.*, 36, 1079–1112, <https://doi.org/10.1080/00206819409465506>, 1994.

1208 Le Quesne, C., Stahle, D. W., Cleaveland, M. K., Therrell, M. D., Aravena, J. C., and Barichivich, J.: Ancient Austrocedrus
 1209 tree-ring chronologies used to reconstruct central Chile precipitation variability from A.D. 1200 to 2000, *J. Climate*, 19, 5731–
 1210 5744, <https://doi.org/10.1175/JCLI3935.1>, 2006.

1211 Kirkbride, M. P.: Debris-covered glaciers, in: *Encyclopedia of Snow, Ice and Glaciers*, edited by: Singh, V. P., Singh, P., and
 1212 Haritashya, U. K., Springer, Dordrecht, the Netherlands, 190–199, https://doi.org/10.1007/978-90-481-2642-2_622, 2011.

1213 Kneisel, C., Hauck, C., Fortier, D., and Moorman, B.: Advances in geophysical methods for permafrost investigations,
 1214 *Permafrost Periglac. Process.*, 19, 157–178, <https://doi.org/10.1002/ppp.616>, 2008.

1215 Hauck, C. and Kneisel, C. (Eds.): *Applied Geophysics in Periglacial Environments*, Cambridge University Press, Cambridge,
 1216 UK, 240 pp., ISBN 978-0-521-88966-7, 2008.

1217 Koenig, C. E. M., Hilbich, C., Hauck, C., Arenson, L. U., and Wainstein, P.: Thermal state of permafrost in the Central Andes
 1218 (27–34° S), *The Cryosphere*, 19, 2653–2676, <https://doi.org/10.5194/tc-19-2653-2025>, 2025.

1219 Masiokas, M., Rabatel, A., Rivera, A., Ruiz, L., Pitte, P., Ceballos, J. L., Barcaza, G., Soruco, Á., Bown, F., Berthier, É.,
 1220 Dussaillant, I., and MacDonell, S.: A review of the current state and recent changes of the Andean cryosphere, *Front. Earth*
 1221 *Sci.*, 8, 99, <https://doi.org/10.3389/feart.2020.00099>, 2020.

1222 Masiokas, M., Villalba, R., Luckman, B. H., Le Quesne, C., and Aravena, J. C.: Snowpack variations in the central Andes of
 1223 Argentina and Chile, 1951–2005: large-scale atmospheric influences and implications for water resources in the region, *J.*
 1224 *Climate*, 19, 6334–6352, <https://doi.org/10.1175/JCLI3969.1>, 2006.

1225 Masiokas, M. H., Cara, L., Villalba, R., Pitte, P., Luckman, B. H., Toum, E., Christie, D. A., Le Quesne, C., and Mauget, S.:
 1226 Streamflow variations across the Andes (18°–55°S) during the instrumental era, *Sci. Rep.*, 9, 17879,
 1227 <https://doi.org/10.1038/s41598-019-53981-x>, 2019.

1228 Matsuoka, N.: Solifluction rates, processes and landforms: a global review, *Earth-Sci. Rev.*, 55, 107–134,
 1229 [https://doi.org/10.1016/S0012-8252\(01\)00057-5](https://doi.org/10.1016/S0012-8252(01)00057-5), 2001.

1230 Mewes, B., Hilbich, C., Delaloye, R., and Hauck, C.: Resolution capacity of geophysical monitoring regarding permafrost
 1231 degradation induced by hydrological processes, *The Cryosphere*, 11, 2957–2974, <https://doi.org/10.5194/tc-11-2957-2017>,
 1232 2017.

1233 Millan, R., Mouginot, J., Rabatel, A., and Morlighem, M.: Ice velocity and thickness of the world's glaciers, *Nat. Geosci.*, 15,
 1234 124–129, <https://doi.org/10.1038/s41561-021-00885-z>, 2022.

1235 Minvielle, M. and Garreaud, R. D.: Projecting rainfall changes over the South American Altiplano, *J. Climate*, 24, 4577–4583,
 1236 <https://doi.org/10.1175/JCLI-D-11-00051.1>, 2011.

1237 Monnier, S. and Kinnard, C.: Internal structure and composition of a rock glacier in the Andes (upper Choapa Valley, Chile)
 1238 using borehole information and ground-penetrating radar, *Ann. Glaciol.*, 54, 61–72, <https://doi.org/10.3189/2013AoG64A107>,
 1239 2013.

1240 Monnier, S. and Kinnard, C.: Reconsidering the glacier to rock glacier transformation problem: new insights from the central
 1241 Andes of Chile, *Geomorphology*, 238, 47–55, <https://doi.org/10.1016/j.geomorph.2015.02.025>, 2015.

1242 Morales, M. S., Christie, D. A., Villalba, R., Argollo, J., Pacajes, J., Silva, J. S., Alvarez, A., Llancabure, J. C., and Soliz
 1243 Gamboa, C. C.: Precipitation changes in the South American Altiplano since 1300 AD reconstructed by tree rings, *Clim. Past*,
 1244 8, 653–666, <https://doi.org/10.5194/cp-8-653-2012>, 2012.
 1245 Ohlanders, N., Rodríguez, M., and McPhee, J.: Stable water isotope variation in a central Andean watershed dominated by
 1246 glacier and snowmelt, *Hydrol. Earth Syst. Sci.*, 17, 1035–1050, <https://doi.org/10.5194/hess-17-1035-2013>, 2013.
 1247 Palacios, D., Oliva, M., Gómez-Ortiz, A., Andrés, N., Fernández-Fernández, J. M., Schimmelpfennig, I., Léanni, L., and Team,
 1248 A.: Climate sensitivity and geomorphological response of cirque glaciers from the late glacial to the Holocene, Sierra Nevada,
 1249 Spain, *Quaternary Sci. Rev.*, 248, 106617, <https://doi.org/10.1016/j.quascirev.2020.106617>, 2020.
 1250 Paul, F., Barrand, N. E., Baumann, S., Berthier, E., Bolch, T., Casey, K., Frey, H., Janes, S. P., Krajci, P., Molg, N., Möller,
 1251 M., Nuth, C., Radić, V., Rastner, P., Raup, B., Schar, C., and Winsvold, S.: On the accuracy of glacier outlines derived from
 1252 remote sensing data, *Ann. Glaciol.*, 54, 171–182, <https://doi.org/10.3189/2013AoG63A296>, 2013.
 1253 Paul, F., Winsvold, S. H., Kääb, A., Nagler, T., and Schwaizer, G.: Glacier remote sensing using Sentinel-2. Part II: mapping
 1254 glacier extents and surface facies, and comparison to Landsat 8, *Remote Sens.*, 8, 575, <https://doi.org/10.3390/rs8070575>,
 1255 2016.
 1256 Pavoni, M., Sirch, F., and Boaga, J.: Electrical and electromagnetic geophysical prospecting for the monitoring of rock glaciers
 1257 in the Dolomites, Northeast Italy, *Sensors*, 21, 1294, <https://doi.org/10.3390/s21041294>, 2021.
 1258 Peng, X., Zhang, T., Frauenfeld, O. W., Wang, K., and Mu, C.: An integrated index of cryospheric change in the Northern
 1259 Hemisphere, *Global Planet. Change*, 218, 103984, <https://doi.org/10.1016/j.gloplacha.2022.103984>, 2022.
 1260 Peña, H. and Nazarala, B.: Snowmelt-runoff simulation model of a central Chile Andean basin with relevant orographic effects,
 1261 *IAHS Publ.*, 166, 161–172, 1987.
 1262 Pieczonka, T., Bolch, T., Kröhnert, M., Peters, J., and Liu, S.: Glacier branch lines and glacier ice thickness estimation for
 1263 debris-covered glaciers in the Central Tien Shan, *J. Glaciol.*, 64, 835–849, <https://doi.org/10.1017/jog.2018.75>, 2018.
 1264 Pourrier, J., Jourde, H., Kinnard, C., Gascoin, S., and Monnier, S.: Glacier meltwater flow paths and storage in a
 1265 geomorphologically complex glacial foreland: the case of the Tapado glacier, dry Andes of Chile (30°S), *J. Hydrol.*, 519,
 1266 1068–1083, <https://doi.org/10.1016/j.jhydrol.2014.08.023>, 2014.
 1267 Ragetti, S. and Pellicciotti, F.: Calibration of a physically based, spatially distributed hydrological model in a glacierized
 1268 basin: on the use of knowledge from glaciometeorological processes to constrain model parameters, *Water Resour. Res.*, 48,
 1269 W03509, <https://doi.org/10.1029/2011WR010559>, 2012.
 1270 Rahaman, S. A., Ajeez, S. A., Aruchamy, S., and Jegankumar, R.: Prioritization of sub-watershed based on morphometric
 1271 characteristics using fuzzy analytical hierarchy process and geographical information system – a study of Kallar Watershed,
 1272 Tamil Nadu, *Aquat. Procedia*, 4, 1322–1330, <https://doi.org/10.1016/j.aqpro.2015.02.172>, 2015.
 1273 Réveillet, M., MacDonell, S., Gascoin, S., Kinnard, C., Lhermitte, S., and Schaffer, N.: Impact of forcing on sublimation
 1274 simulations for a high mountain catchment in the semiarid Andes, *The Cryosphere*, 14, 147–163, [https://doi.org/10.5194/tc-](https://doi.org/10.5194/tc-14-147-2020)
 1275 14-147-2020, 2020.

1276 RGI Consortium: Randolph Glacier Inventory – a dataset of global glacier outlines, version 6.0, Global Land Ice Measurements
 1277 from Space, Boulder, Colorado, USA, <https://doi.org/10.7265/N5-RGI-60>, 2017.

1278 Rittger, K., Painter, T. H., and Dozier, J.: Assessment of methods for mapping snow cover from MODIS, *Adv. Water Resour.*,
 1279 51, 367–380, <https://doi.org/10.1016/j.advwatres.2012.03.002>, 2013.

1280 Rodríguez, M., Ohlanders, N., Pellicciotti, F., Williams, M. W., and McPhee, J.: Estimating runoff from a glacierized
 1281 catchment using natural tracers in the semi-arid Andes cordillera, *Hydrol. Process.*, 30, 3609–3626,
 1282 <https://doi.org/10.1002/hyp.10973>, 2016.

1283 Rubinstein, R. Y. and Kroese, D. P.: *Simulation and the Monte Carlo Method*, 3rd edn., Wiley, Hoboken, NJ, USA, 2016.

1284 Saltelli, A., Ratto, M., Andres, T., Campolongo, F., Cariboni, J., Gatelli, D., Saisana, M., and Tarantola, S.: *Global Sensitivity
 1285 Analysis: The Primer*, Wiley, Chichester, UK, 2008.

1286 Saydi, M. and Ding, J.: Impacts of topographic factors on regional snow cover characteristics, *Water Sci. Eng.*, 13, 171–180,
 1287 <https://doi.org/10.1016/j.wse.2020.09.002>, 2020.

1288 Schaffer, N., MacDonell, S., Réveillet, M., Yáñez, E., and Valois, R.: Rock glaciers as a water resource in a changing climate
 1289 in the semiarid Chilean Andes, *Reg. Environ. Change*, 19, 1263–1279, <https://doi.org/10.1007/s10113-018-01459-3>, 2019.

1290 Schrott, L.: Some geomorphological-hydrological aspects of rock glaciers in the Andes (San Juan, Argentina), *Z. Geomorphol.*,
 1291 104, 161–173, 1996.

1292 Sriyana, I.: Developed watershed classification index determining management priority level based on watershed carrying
 1293 capacity, *MATEC Web Conf.*, 270, 04004, <https://doi.org/10.1051/mateconf/201927004004>, 2019.

1294 Suárez, F., Muñoz, J. F., Fernandez, B., Dorsaz, J.-M., Hunter, C. K., Karavitis, C., and Gironás, J.: Integrated water resource
 1295 management and energy requirements for water supply in the Copiapó River Basin, Chile, *Water*, 6, 2590–2613,
 1296 <https://doi.org/10.3390/w6092590>, 2014.

1297 Trombotto, D. L., Sileo, N., and Dapeña, C.: Periglacial water paths within a rock glacier-dominated catchment in the Stepanek
 1298 area, Central Andes, Mendoza, Argentina, *Permafrost Periglac. Process.*, 31, 311–323, <https://doi.org/10.1002/ppp.2043>, 2020.

1299 Valdivieso, S., Hassanzadeh, A., Vázquez-Suñé, E., Custodio, E., and Criollo, R.: Spatial distribution of meteorological factors
 1300 controlling stable isotopes in precipitation in northern Chile, *J. Hydrol.*, 605, 127380,
 1301 <https://doi.org/10.1016/j.jhydrol.2021.127380>, 2022.

1302 Vidal-Páez, P., Fernández-Sarría, A., González-Bonilla, M. J., Derauw, D., Pérez-Martínez, W., Azócar, G., and Ortega, J. H.:
 1303 Monitoring surface velocity changes of rock glaciers in the Chilean Andes using DInSAR technique with PAZ imagery,
 1304 *Procedia Comput. Sci.*, 239, 456–463, <https://doi.org/10.1016/j.procs.2024.06.070>, 2024.

1305 Villarroel, C. D., Tamburini Beliveau, G., Forte, A. P., Monserrat, O., and Morvillo, M. C.: DInSAR for a regional inventory
 1306 of active rock glaciers in the Dry Andes Mountains of Argentina and Chile with Sentinel-1 data, *Remote Sens.*, 10, 1588,
 1307 <https://doi.org/10.3390/rs10101588>, 2018.

1308 Vivero, S., Bodin, X., Farías-Barahona, D., MacDonell, S., Schaffer, N., Robson, B. A., and Lambiel, C.: Combination of
 1309 aerial, satellite, and UAV photogrammetry for quantifying rock glacier kinematics in the Dry Andes of Chile (30° S) since the
 1310 1950s, *Front. Remote Sens.*, 2, 784015, <https://doi.org/10.3389/frsen.2021.784015>, 2021.
 1311 Vuille, M., Francou, B., Wagnon, P., Juen, I., Kaser, G., Mark, B. G., and Bradley, R. S.: Climate change and tropical Andean
 1312 glaciers: past, present and future, *Earth-Sci. Rev.*, 89, 79–96, <https://doi.org/10.1016/j.earscirev.2008.04.002>, 2008.
 1313 Wolfe, J. D., Shook, K. R., Spence, C., and Whitfield, C. J.: A watershed classification approach that looks beyond hydrology:
 1314 application to a semi-arid, agricultural region in Canada, *Hydrol. Earth Syst. Sci.*, 23, 3945–3967, [https://doi.org/10.5194/hess-](https://doi.org/10.5194/hess-23-3945-2019)
 1315 23-3945-2019, 2019.
 1316 Zech, R., May, J. H., Kull, C., Ilgner, J., Kubik, P. W., and Veit, H.: Timing of the late Quaternary glaciation in the Andes
 1317 from ~15 to 40°S, *J. Quaternary Sci.*, 23, 635–647, <https://doi.org/10.1002/jqs.1200>, 2008.