

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

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

1 Introduction

Anthropogenic climate change impacts water availability, particularly in regions dependent on high-altitude water sources within mountain systems (Beniston & Stoffel, 2014, Beniston et al., 2018). Changes in the cryosphere that will occur within the next century include a change from solid to liquid precipitation, declining volumes of cryoforms and snow lines to higher altitudes, and shorter snow seasons. With these changes, the predictions are that water resources in mountain 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., 2019).

Within this climatic context, the Copiapó River exhibits a mixed hydrological regime with very low discharges, typically ranging between ~ 1.49 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 basin 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 Copiapó’s runoff 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 limited amount of rainfall in the Andes, streamflow in the Copiapó River is largely controlled by cryospheric processes. However, the contribution of cryospheric water storage to runoff varies substantially among cryoform 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.

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.

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 reserves, but the potential volumes of water that are present in these cryospheric reserves are highly variable and depend on both the type of cryoform and the spatial distribution. We hypothesize that water is unevenly distributed across the Atacama Desert, depending on the types of cryoforms and their specific locations, which implies that the potential water supply may vary accordingly. Consequently, once the ice bodies begin to melt, significant variability in water runoff is expected throughout the region.

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 reserves and resources within arid watershed contexts.

The main objective of this study is therefore to develop a methodology that is able to identify and integrate the cryospheric components into a quantitative tool that can monitor the potential volumes of water of each sub-watershed. Since there is currently no regulatory framework in Chile on watershed management and conservation, that includes potential water volumes from cryospheric reserves, this study includes a technical workflow on which policies for the regulation of cryospheric watersheds can be based. The hydrological role of these cryospheric watersheds will prove an important factor for sustainable water resource management for the Copiapo watershed, especially when the freshwater reserves currently captured in high altitude cryoforms become available for the Chilean population in the very near future due continuously increasing global temperatures (Flores et al., 2018).

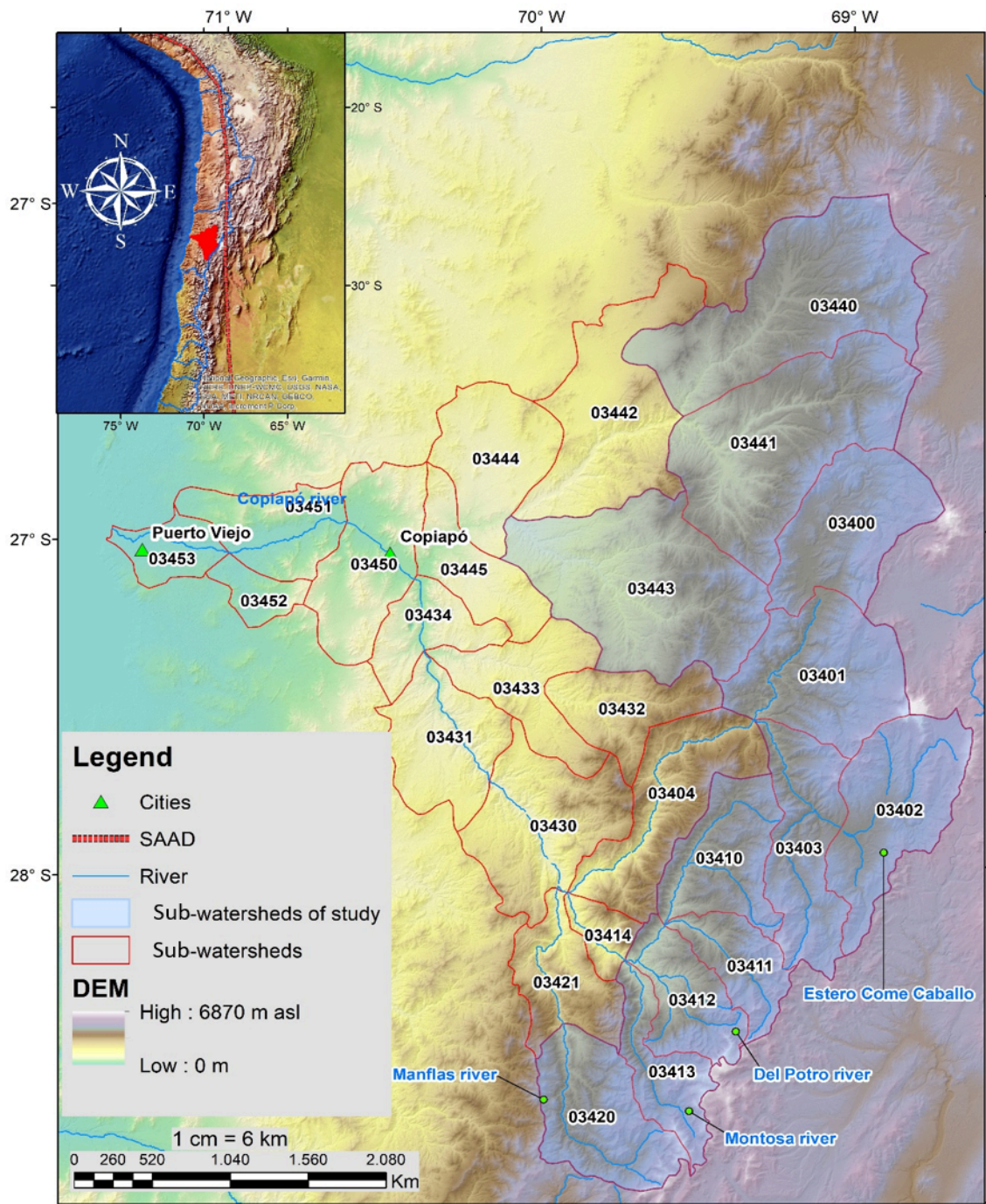


Figure 1. Location map of the Copiapo watershed and the different sub-watersheds (Zone 19S). The codes in each sub-watershed represent the official numbering that the General Water Directorate assigns to each sub-watershed.

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

2 Regional setting

The cryosphere of the study area is composed of both cryospheric reserves and cryospheric resources. In this study, we differentiate between cryospheric reserves and resources, because both reserves and resources contribute differently to the potential volume of water in cryoforms. The cryospheric reserves of the Atacama Desert host year-round stable cryoforms (Fig. 2), such as debris-free glaciers, debris-covered glaciers, rock glaciers, gelifluction taluses, and protalus lobes, as documented by Richmond (1952). Together, these five classes comprise the cryospheric reserves found in the desert regions of northern Chile and Argentina, whose classification and characteristics were clearly defined by García et al. (2017). The cryospheric resources correspond to the snow permanence that produces freshwater by melting in the same year in which it precipitates.

The Copiapo 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 Copiapo from 1795 to the present shows a mean annual value of 22.5 mm (Izquierdo et al., 2024). Among the arid basins in northern Chile, the Copiapo 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 reserves and resources 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 Copiapo, there is a flow of water in the rivers of the Copiapo 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; Azocar and Brenning, 2010; Schaffer et al., 2019).

The glacial and periglacial inventory of the Copiapo 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 Copiapo watershed, the occurrence of snow begins above 3,500 m of altitude. In these extreme environments, studies suggest that 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).

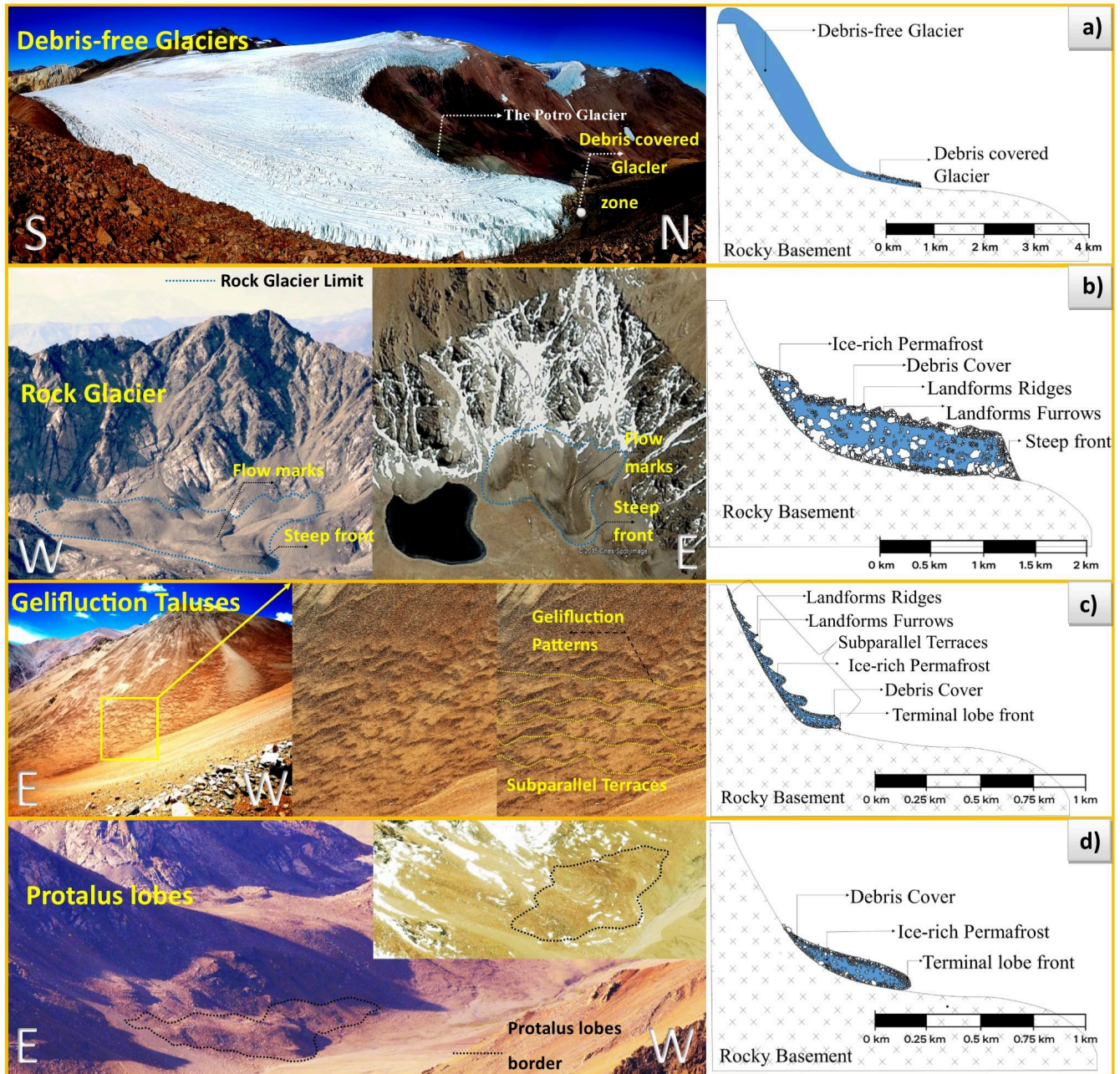


Figure 2. Main cryoforms. Field photographs from the Copiapo watershed on the left, with a schematic interpretation on the right of. a) Debris-free glacier and debris-glaciers, b) Rock glaciers of the Del Potro river sub-watershed. c) Gelifluction taluses of Montosa river sub-watershed, d) Protalus lobes of the La Laguna sub-watershed in Pulido watershed.

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

The hydrology of arid Andean basins 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 Aguas (DGA) hydrometric network show a pronounced summer peak in river flow—when rainfall is nearly zero—together with a high baseflow index, confirming a melt-dominated regime and the central hydrological role of the cryosphere in this region. These patterns are consistent with previous studies across the semi-arid Andes (Favier et al., 2009; Gascoin et al., 2011; Ayala et al., 2016), where snow and frozen water bodies act as short- and long-term reservoirs modulating runoff during dry years (Table 2).

Cryospheric contributions, however, operate over markedly different response times. Seasonal snow responds at sub-seasonal scales, controlling the onset and magnitude of spring–summer discharge. Debris-free and debris-covered glaciers respond at seasonal to multiannual scales, providing sustained meltwater even in years of limited snowfall. Ice-rich permafrost landforms—including rock glaciers, protalus lobes, and gelifluction slopes—contribute at decadal to centennial scales and represent the most climatically buffered frozen-water reservoirs in the basin. Capturing these contrasted response times requires not only quantifying the spatial distribution of cryoforms but also applying a coherent conceptual framework for estimating their ice volumes.

Table 2. Contribution to streamflow (%) indicates the proportion of total river discharge supplied by cryospheric meltwater (glaciers 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	Ragettli & 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—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, 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. (2020), 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 protalus 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., 2020) to all permafrost cryoforms—including rock glaciers, protalus lobes, and gelifluction slopes—supported by DInSAR-derived velocities.

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

3 Methodology

3.1 Study Area

The Copiapó watershed is composed of 27 sub-watersheds, distributed along the high mountain belt that forms the headwaters of the basin. A sub-watershed is defined here as a watershed unit with a lower Strahler stream order than the main watershed. In our research area, the Copiapó watershed corresponds to the main basin and contains rivers with a maximum Strahler stream order of nine. Within this hierarchy, sub-watersheds draining from the headwaters typically initiate with Strahler orders of one to three and progress downstream until they merge into larger tributaries with stream orders of up to nine. By distinguishing among sub-watersheds and their associated cryoforms, we are able to perform both quantitative and qualitative assessments of the current potential water volumes stored within cryospheric environments.

To classify a sub-watershed as a cryospheric watershed, we applied objective and reproducible criteria based on its cryospheric components. A sub-watershed is considered cryospheric only when it fulfilled two conditions: (i) the presence of at least one active cryoform—debris-free glacier, debris-covered glacier, rock glacier, protalus lobe, or gelifluction slope—mapped in the regional inventory of glaciers and the periglacial environment of the Atacama Desert (García et al., 2017); and (ii) measurable seasonal cryospheric resources, expressed as long-term snow permanence derived from MODIS fractional snow cover (Rittger et al., 2013). Sub-watersheds fulfilling both requirements (≥ 1 cryoform + snow permanence) were classified as cryospheric.

Applying these criteria resulted in 12 sub-watersheds being identified as cryospheric watersheds (Fig. 1; Table 1). These units are located entirely within high-elevation sectors of the basin (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 cryoforms mapped in the Copiapó basin. This classification is therefore geomorphologically justified, reproducible, and consistent with independent permafrost modelling and regional cryospheric inventories.

The study area was identified based on the cryospheric components retrieved from the inventory of glaciers and periglacial environment of the Atacama Desert (García et al., 2017). After mapping, we compared the cryoform polygons to the global permafrost model (Gruber, 2012), and 97% of the mapped cryoforms fall within the high-probability permafrost zone. The identification of the 12 cryospheric sub-watersheds of the Copiapó basin was therefore based on both geomorphological evidence (presence of cryoforms) and remote-sensing-derived snow permanence (MODIS), consistent with our methodology for defining cryospheric reserves and resources.

3.2 Survey of cryospheric water reserves

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

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

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

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

3.3 Survey of cryospheric resources: MODIS-derived snow permanence

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

approximately 90% accuracy under ideal conditions (Hall et al., 2006; 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 basin.

In this framework, FSC was used as the primary input for constructing daily snow permanence maps, and snow permanence—computed as the proportion of snow-covered days over the full time series—was incorporated into the Normalized Cryospheric Index (NCI) as the indicator of short-term cryospheric water resources. This approach ensures that the characterization of snow permanence reflects both their spatial extent and their temporal recurrence.

3.4 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 cryoforms 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 basin and contain accessible examples of the main cryoforms present in the region.

During these campaigns, photographic and geomorphological observations were conducted to verify the surface expression of mapped cryoforms, 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 cryoforms in the modelling framework.

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

3.4.1 Geophysical survey on gelifluction slopes and rock glaciers

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



Figure 3. Photography of the fieldwork. a) And b) the Lobulated morphology of studied gelifluction taluses, which are not covered in snow; c) and d) the geoelectric line cable; e) and f) the used equipment, differential GPS and resistivity meter ARES II measuring a resistivity point diagram distribution.

Fieldwork focused on two representative cryoforms within the Montosa sub-watershed: a gelifluction slope and an active rock glacier (Fig. 3 and Fig. 4). 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
Resistivimeter: ARES II	1
30 electrode box	5
10 chaneln multicable II (60 m)	10
Car Bateries	2
Diferential 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 cryoforms in our volume estimations and complementing the GPR-based glacier-thickness validation described in Section 3.4.



Figure 4. 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.

3.5 DInSAR kinematic analysis for glacial and periglacial cryoforms

To characterize surface deformation across glacial and periglacial cryoforms 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:

$$d_{LOS} = \frac{\lambda}{4\pi} \Delta\phi \quad (1)$$

Mean surface velocities were extracted for each cryoform 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 cryoforms and as supplementary kinematic information in glacier-thickness interpretation.

3.5.1 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 $\gamma \approx 0.97$, 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 was applied a factor of $L=10$, the resulting uncertainties are:

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

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

3.6 Calculation of the ice volume and uncertainty

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

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

3.6.2 Debris-covered glaciers

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

3.6.3 Rock glaciers

Rock-glacier thickness was estimated using a viscoplastic creep inversion following Cicoira et al. (2020), where basal shear stress is constrained by observed surface velocity. For each rock glacier, mean slope-parallel surface velocities ($\bar{v}$) and their standard deviation (σ_v) were derived from Sentinel-1 DInSAR measurements (Sect. 3.6.1) and expressed in m yr^{-1} . The basal shear stress τ was inverted using the 1-D viscoplastic formulation with Glen-type exponent $n=3$ and a temperature-dependent rate factor $A(T)$.

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

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

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

Using τ , mean thickness was computed as:

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

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

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

where A is the mapped planimetric area.

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

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

For rock glaciers, C_I was represented as a bounded range consistent with reported values for these type of ice-rich permafrost landforms (Cicoira et al., 2020):

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

Ice mass (in Gt) was then computed as:

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

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 uncertainty in area, ice-content uncertainty, and density uncertainty using quadratic propagation of relative errors (Sect. 3.6.5).

3.6.4 Protalus lobes and gelifluction slopes

Protalus lobes and gelifluction taluses were modelled using the same viscoplastic inversion framework originally developed for rock glaciers (Cicoira et al., 2020), exploiting low but measurable DInSAR-derived surface deformation to constrain basal shear stress and thickness. For each mapped polygon, mean slope-parallel surface velocity and its associated uncertainty (v , σ_v) were used to invert basal shear stress and compute thickness H through Eq. (6). Thermal uncertainty was represented using the same Central Andes temperature prior described in Eq. (5).

However, in contrast to rock glaciers, protalus lobes and gelifluction taluses are characterized by lower ice contents, higher debris fractions, and a dominance of seasonal or shallow permafrost deformation mechanisms. To account for these

rheological differences, we modified the Glen-type flow exponent n used in the inversion. While rock glaciers were modelled assuming $n = 3$, consistent with creep-dominated ice–debris mixtures (Cicoira et al., 2020), protalus lobes and gelifluction taluses were modelled using a reduced exponent ($n = 2$), reflecting more linear, frictionally controlled deformation regimes typical of ice-poor frozen debris slopes. This adjustment reduces the sensitivity of inferred thickness to surface velocity and avoids 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).

3.6.5 Uncertainty propagation framework for ice-rich permafrost cryoforms

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:

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

(13)

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

$$\sigma_{tot} = \sqrt{\sum_i \sigma_{M_i}^2}$$

(14)

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

Table 5. Key references used as input for the calculation of ice volumes of cryospheric reserves.

Cryoform	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., 2020) + DInSAR	40–60%	Cicoira et al. (2020)
Protalus lobes	Adapted viscoplastic model + DInSAR	25–49%	Hilbich et al. (2022)
Gelifluction slopes	Adapted viscoplastic model + DInSAR	25–49%	Hilbich et al. (2022)

3.7 Reserve and resource index and data integration

To develop a classification system that categorizes the capacity to store ice as a strategic water reserve in desert areas of sub-watersheds, we followed a workflow that consists of four phases (Fig. 5). We developed the classification system using cryospheric reserves and resources, integrating both reserves (expressed in gigatons (Gt), equivalent to 1,000 million metric tons of ice) and resources (quantified in km² of snow-cover extent) into the methodological framework. The strategic Normalized Cryospheric Index (NCI) captures the relative significance of each sub-watershed, considering both its ability to store water as ice (cryospheric reserves) and its potential to regenerate water resources through snowfall (cryospheric resources). The NCI needs to be adapted to each specific case. For the Copiapo watershed, the active cryoforms present in cryospheric reserves until 2024 have been used. Those cryospheric reserves include debris-free glaciers, debris-covered glaciers, rock glaciers, gelifluction talus and protalus lobes. Depending on the types of cryoforms available in a specific region, the number of cryoforms included may vary.

The methodological framework for computing the NCI index integrates Monte Carlo simulation with a robustness assessment 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 $N(\bar{V}_i, \sigma_{2v})$ y $N(\bar{S}_i, \sigma_{2s})$, 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:

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

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 cryospheric resources are 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.

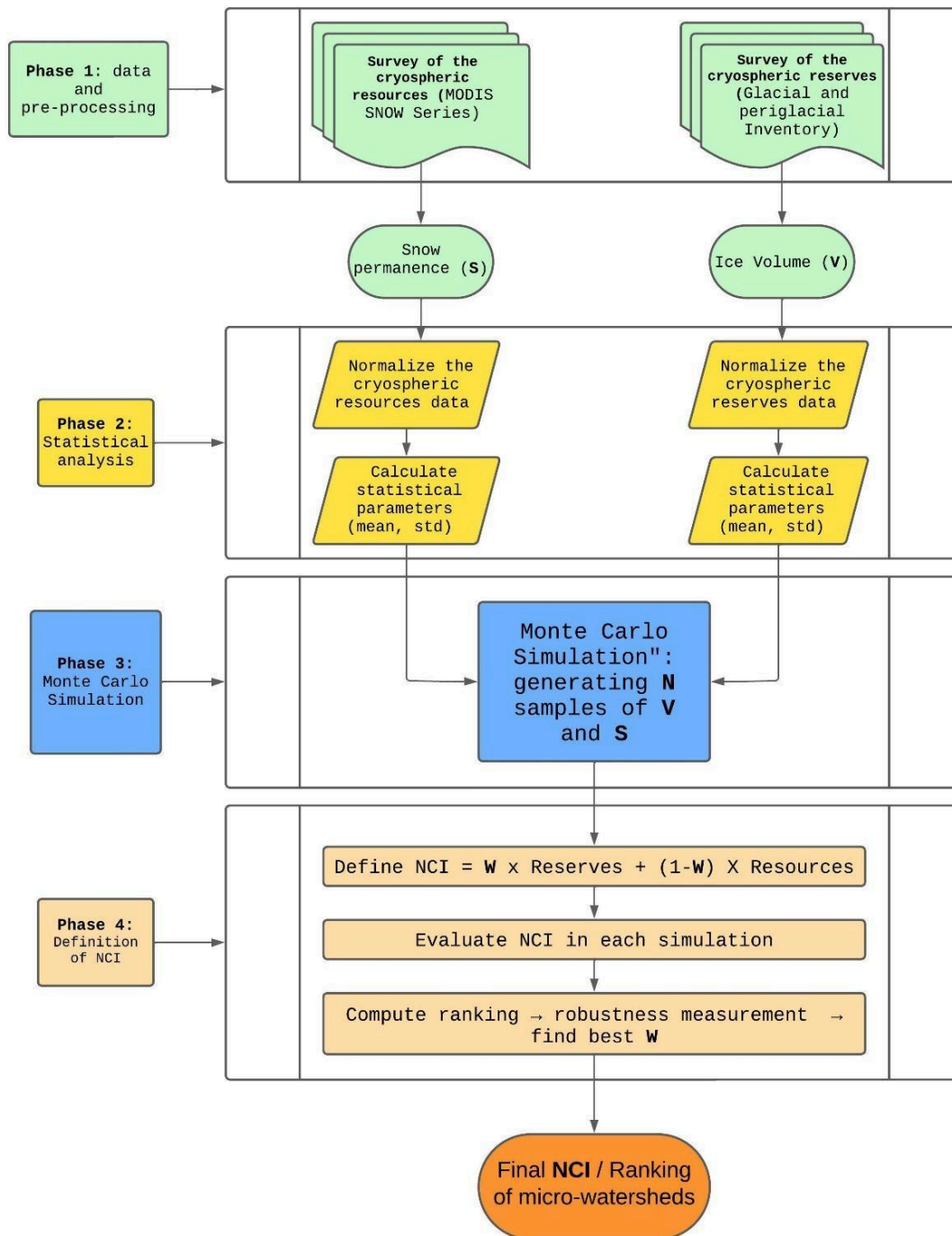


Figure 5. Flow chart of the methodology used to calculate the NCI.

4 Results

4.1 Glacier and Periglacial Environment Mapping

Mapping the cryospheric reserves inventoried 1862 cryoforms in the Copiapo watershed. Of these cryoforms, 77 cryoforms correspond to debris-free glaciers (Fig. 6), 16 to debris-covered glaciers, 184 to rock glaciers, 369 to gelifluction taluses and 731 to protalus lobes (Fig. 7). There is a gap in cryoforms observed between Nevado Jotabeche and Del Potro (Fig. 6). Around Del Potro (app. at 4,500 m altitude), the cryoforms are found in the valleys in between the steep ridges, whereas around Nevado Jotabeche and Los Tronquitos, the cryoforms cover the highest points (i.e. ridges and peaks) in the region.

Gelifluction taluses account for more than 70% of the mapped cryospheric reserve area in the study region and constitute the most widespread cryospheric landform in the Copiapó watershed, although their volumetric contribution is subject to larger uncertainty than that of glacier ice. The cryoform inventory demonstrates a notable concentration of gelifluction talus features within the southern sector of the study area, proximal to the Argentinian border (Fig. 7b, c). These talus formations occur in close spatial association with debris-free glaciers, exhibiting considerable size variability with surface areas spanning 0.01 to 20.33 km². Their distribution is strongly controlled by elevation, with 89% occurring above 5,000 m a.s.l.. This spatial association between gelifluction taluses and debris-free glaciers is characteristic of the southern, higher-elevation sector of the study area and contrasts with cryoform assemblages observed in the northern sector (Fig. 7a). In the north, gelifluction taluses generally occur as isolated landforms, without adjacent debris-free glaciers, and are restricted to relatively small surface areas ranging from 0.01 to 3.56 km². In contrast, in the southern Del Potro area at elevations around 5,500 m a.s.l., gelifluction taluses commonly occur in close proximity to debris-free glaciers and reach substantially larger surface areas (2.1–20.33 km²), while the associated debris-free glaciers range between 0.18 and 6.96 km².

The second-largest cryospheric landforms in our research area are rock glaciers, making up 25.28% of the cryoforms in the region. Rock glaciers are found predominantly at high altitudes but below 5,000 m, close to steep ridges. Together with the protalus lobes, the rock glaciers are cryoforms that reach the lowest altitudes (4,200 m asl). In our research areas, Del Potro (Fig. 7b) hosts most of the rock glaciers. They appear as elongated bodies varying between 0.02 km² and 1.2 km² that fill the valleys in between the steep ridges.

Debris-free glaciers represent the third most extensive cryospheric component within the study area, exhibiting comparable prevalence to rock glaciers. These features account for 24.58% of all documented cryoforms in the region and are predominantly located at elevations exceeding 5,500 meters above sea level (masl). The debris-free glaciers in the southern part of our research area always appear adjacent to gelifluction taluses, whereas in the north the debris-free glaciers appear as small (< 0.2 km²), isolated bodies at the highest altitudes above 4,500 m. The debris-free glaciers in the south cover surface areas between 0.1 km² and 6.96 km².

In the Nevado Jotabeche region (Fig. 5a), protalus lobes are the dominant cryoform. In this region, protalus lobes manifest in two distinct morphological forms: smaller features (<0.1 km²) positioned along lower slope segments and larger formations

(0.1–0.4 km²) occupying gentler topographic settings. These protalus lobes are typically found at elevations predominantly below 5,000 meters.

Debris-covered glaciers are relatively rare in our research area. They are only found in the south of the Copiapo watershed (Fig. 5). They occur in front of debris-free glaciers and cover small areas of 0.05 km² and 0.1 km². They normally occur at altitudes below 5,000 m.

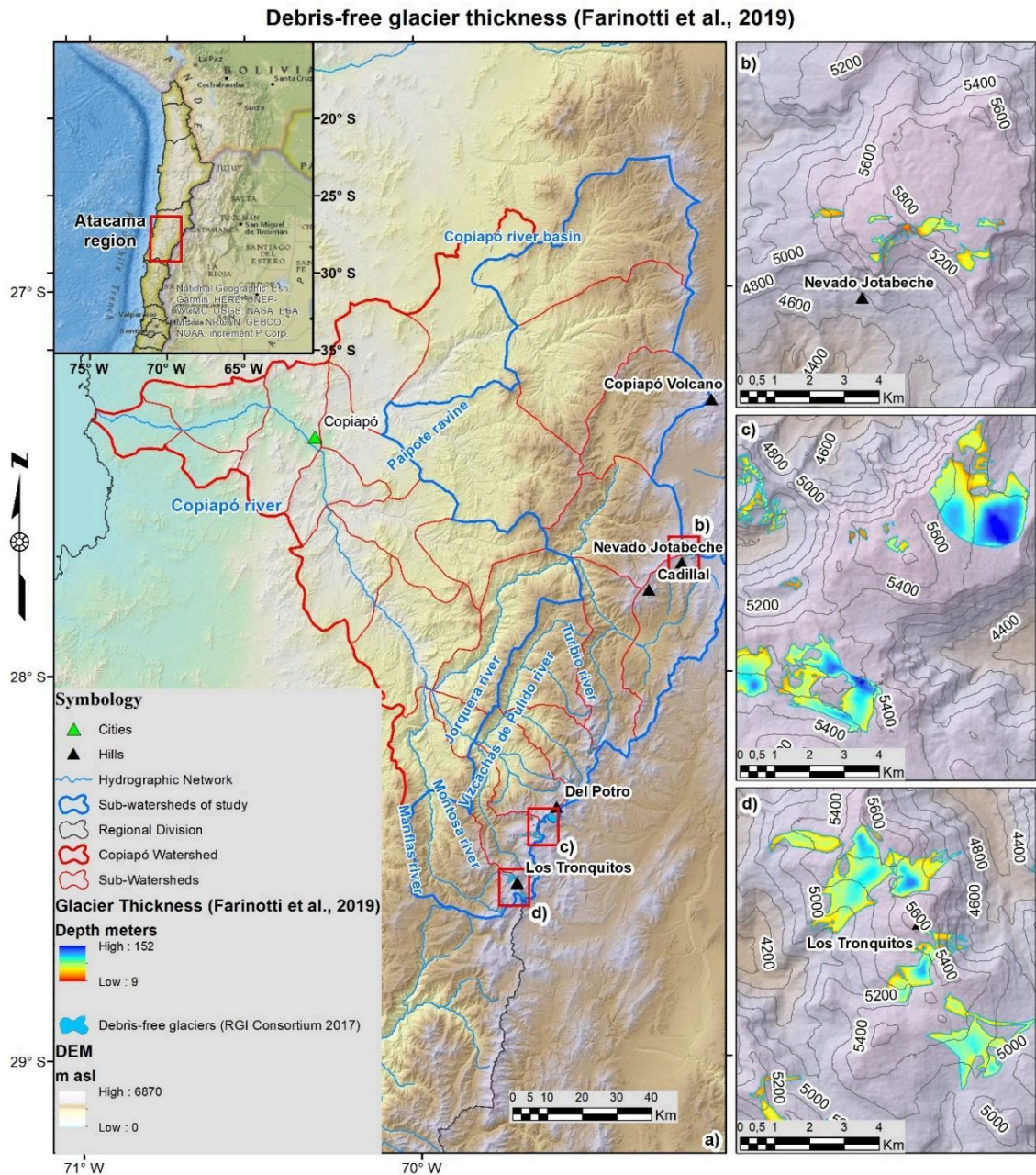


Figure 6. a) Distribution map of the debris-free glacier thickness from Farinotti et al. (2019). b) Jotabeche volcano cryospheric reserves distribution c) The Potro hill cryospheric reserves distribution d) The Tronquitos hill cryospheric reserves distribution.

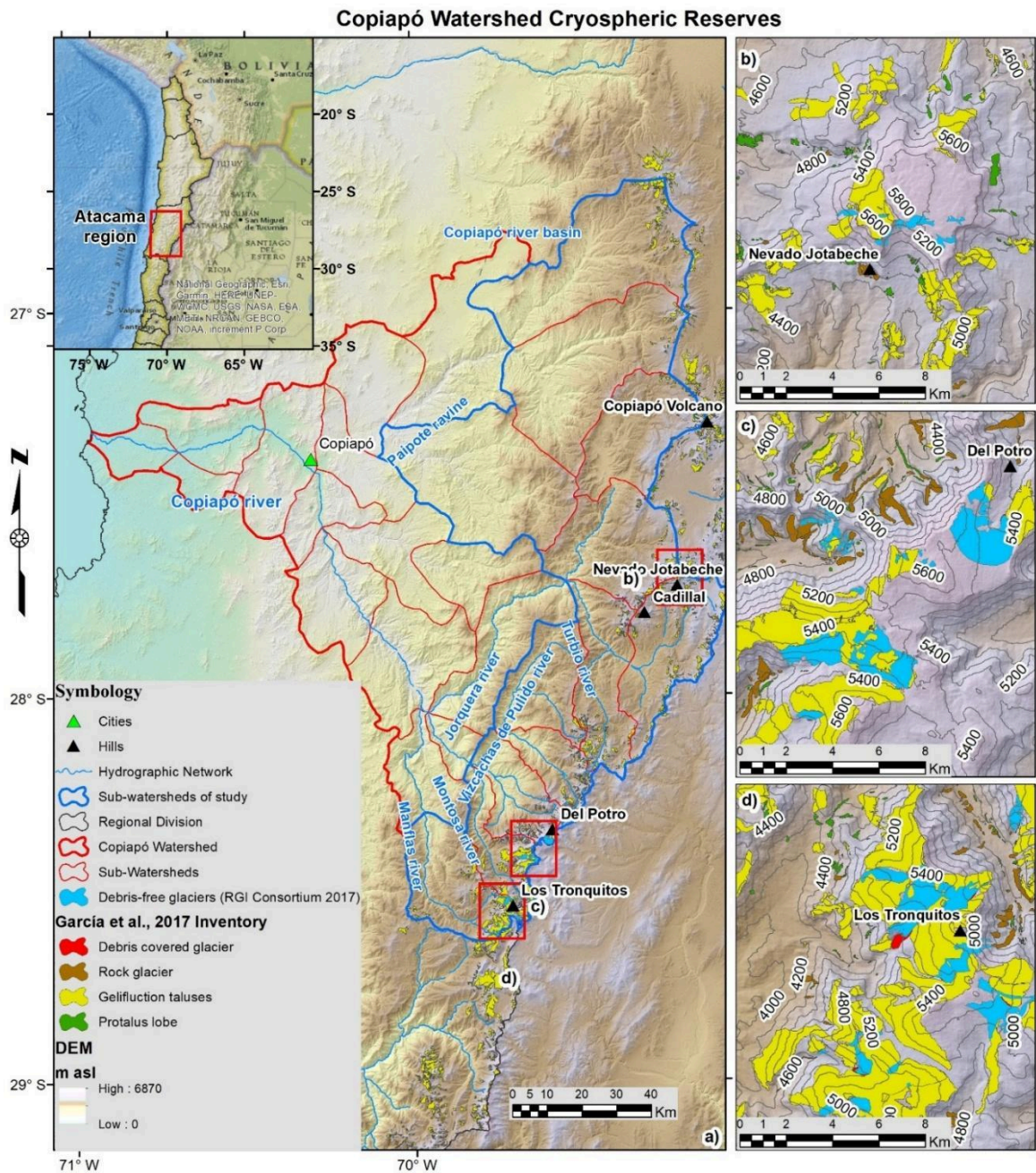
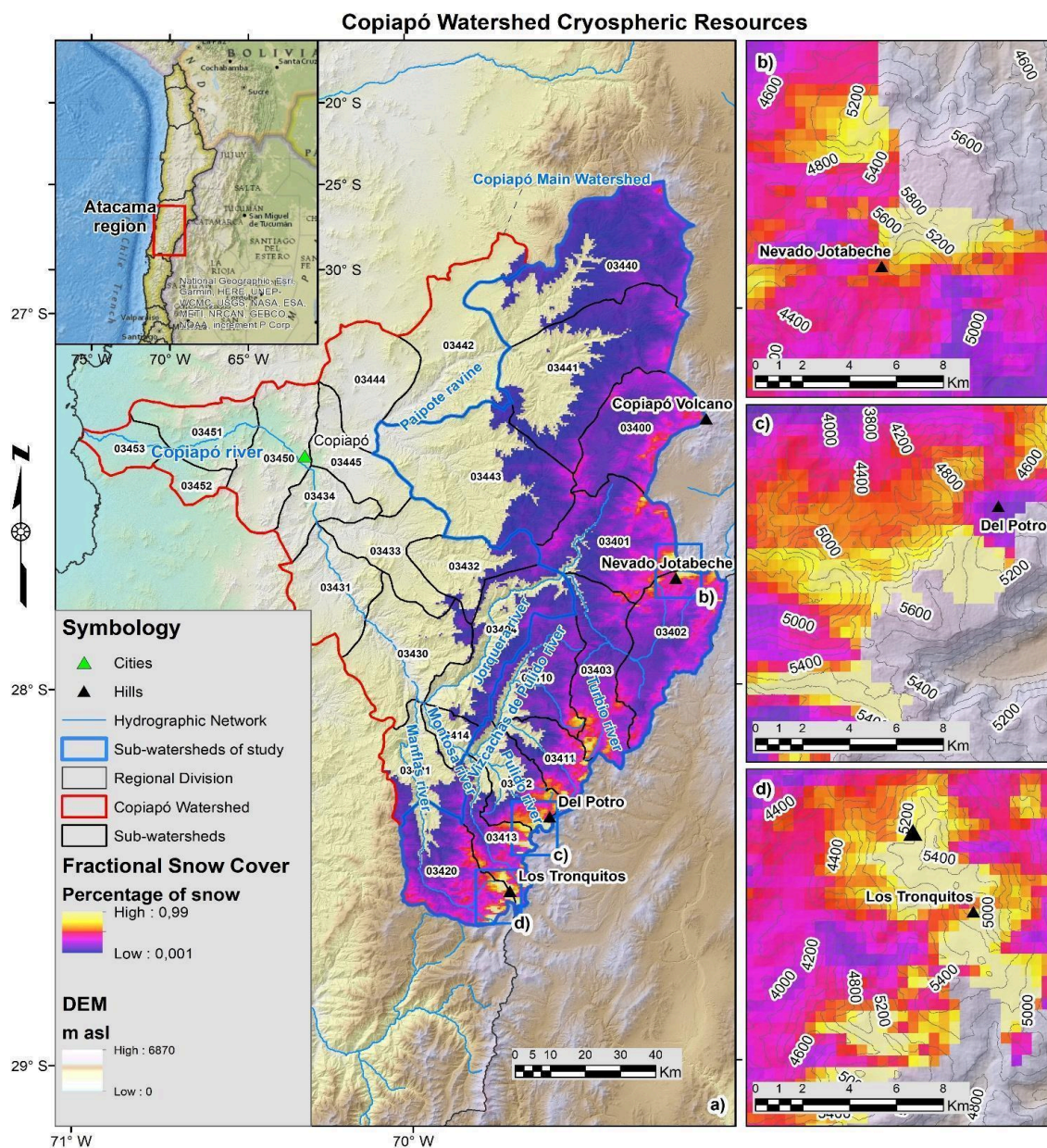


Figure 7. a) Distribution map of the cryospheric reserves inventoried in the Atacama region from García et al. (2017). b) Jotabeche volcano cryospheric reserves distribution c) The Potro hill cryospheric reserves distribution d) The Tronquitos hill cryospheric reserves distribution.

4.2 Snow permanence results

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

The calculated probability of the occurrence for the investigated watersheds shows that the snow permanence watershed varies, with higher snow permanence for the southern watersheds compared to the central and northern watersheds. The watershed with the highest resource index is the Montosa river watershed (03413) with 0.18% probability of being covered in snow. The Vizcachas de Pulido river (03410), which is located at the lowest altitude, has the lowest probability of 0.07% to have snow permanence.



4.3 Analysis of runoff

Daily mean streamflow records at the Copiapó River at Pastillo station (Fig. 9) reveal distinct seasonal patterns between wet and dry years. During the wet year (Oct 2017 - Mar 2018), streamflow exhibits a sustained increase from October to December, peaking at 5.7 m³/s in December, followed by a gradual decline through January and a secondary pulse in

February (5.3 m³/s). In contrast, the dry year (Oct 2013 - Mar 2014) shows substantially lower spring flows, with mean streamflow averaging only 1.0 m³/s during October-November, and the most significant increases concentrated in December (1.3 m³/s) and February (2.2 m³/s). Overall, the wet year delivered 3.4 times more water than the dry year, with the greatest relative differences occurring in November (+471%) and December (+343%).

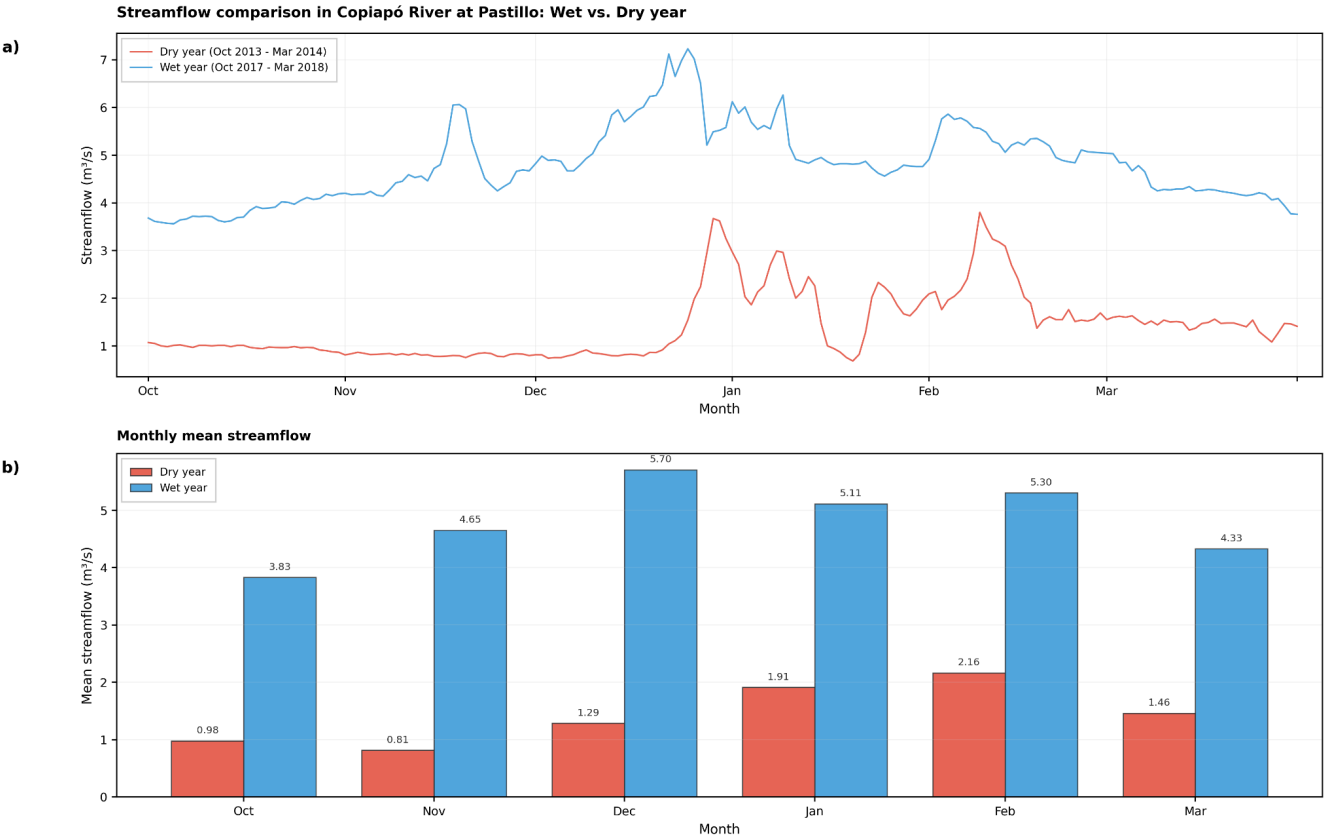


Figure 9. Daily (a) and monthly mean (b) streamflow at Copiapó River at Pastillo for wet (2017-2018, blue) and dry (2013-2014, red) years. Data Source: Dirección General de Aguas, Chile.

These contrasting patterns between individual years reflect broader interannual variability in snowpack accumulation, ice bodies and melt dynamics. To quantify this relationship at the watershed scale, we incorporated a new analysis linking MODIS-derived snow permanence in September—at the onset of the melt season—with mean river discharge observed between October and March (Fig. 6). This relationship shows a robust correlation ($R^2 = 0.58$; $r = 0.76$; $N = 22$), confirming that seasonal snow permanence acts as a dominant short-term cryospheric water resource in the Copiapó watershed. However, the unexplained variance (42%) suggests that other cryospheric components—such as rock glaciers, debris-covered ice, and possibly permafrost—may contribute to streamflow, particularly during low-snow years or late in

the melt season when these longer-residence-time reservoirs become more significant. The integration of these empirical hydrological results into the NCI framework validates that sub-basins with higher NCI values not only exhibit stronger cryospheric signatures in their topography but also display measurable hydrological responses consistent with multiple meltwater sources operating at different temporal scales.

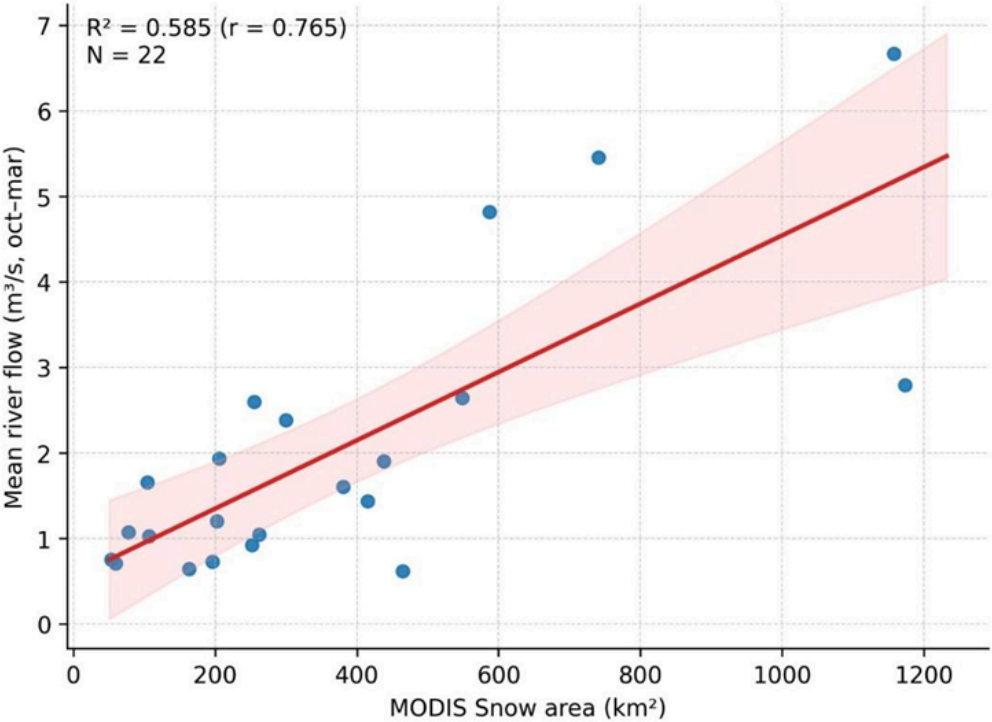


Figure 10. September snow-covered area vs. October-March mean discharge at Copiapó River at Pastillo (2000-2021, N = 22). Linear regression (red line) with 95% confidence interval (shaded). $R^2 = 0.58$; $r = 0.76$; $p < 0.001$. Sources: MODIS MOD10A1 and DGA, Chile.

4.3 Field survey observations of cryospheric reserves.

The fieldwork has yielded observations of cryoforms in the Del Potro and Montosa river sub-watersheds (Fig. 11 and 12). The Del Potro Glacier (Fig. 12d) and the surrounding periglacial environment of the Del Potro River sub-watershed (03412; Fig. 11) were clearly identified during field surveys. In the Montosa River sub-watershed, the Maranceles Glacier (Fig. 12c) was also observed. Additional field observations confirmed the presence of well-developed rock glaciers, gelifluction taluses, and protalus lobes in the Del Potro River basin (03412) (Fig. 11c–g).

Glacial landforms are preferentially distributed on south-facing slopes, as indicated by their aspect, which are generally associated with reduced incoming solar radiation. In contrast, periglacial cryoforms such as protalus lobes and gelifluction

taluses are more frequently found on exposed slopes with steeper gradients and variable aspects, reflecting differences in topographic setting rather than direct measurements of snow accumulation or thermal conditions. Debris-free and debris-covered glaciers are primarily located in topographic depressions and high-elevation headwaters. In contrast, protalus lobes and gelifluction taluses are predominantly distributed on steep slopes and debris-covered sectors. Overall, periglacial cryoforms display a wider spatial distribution across the watershed than glaciers.

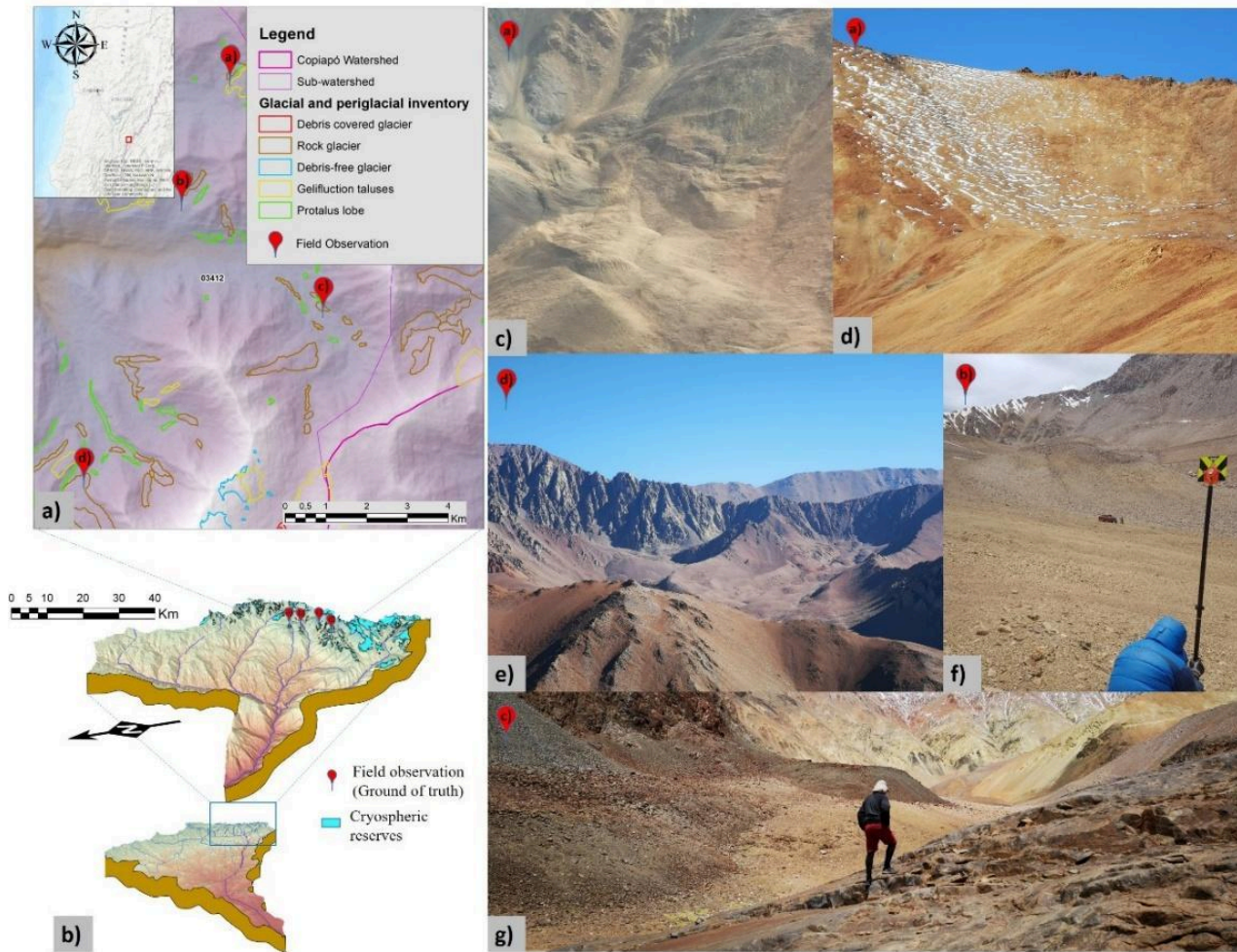


Figure 11. Field observations of cryospheric reserves as the groundtruthing. a) Map of cryospheric reserves distribution, b) 3D view of the watershed, c) Rock glacier of Del Potro river (03412); d) Gelifluction taluses of Del Potro river (03412); e) Rock glacier of Del Potro river (03412); f) Rock glacier of Del Potro river (03412); g) Panorama of different periglacial cryoforms of the Del Potro river (03412).

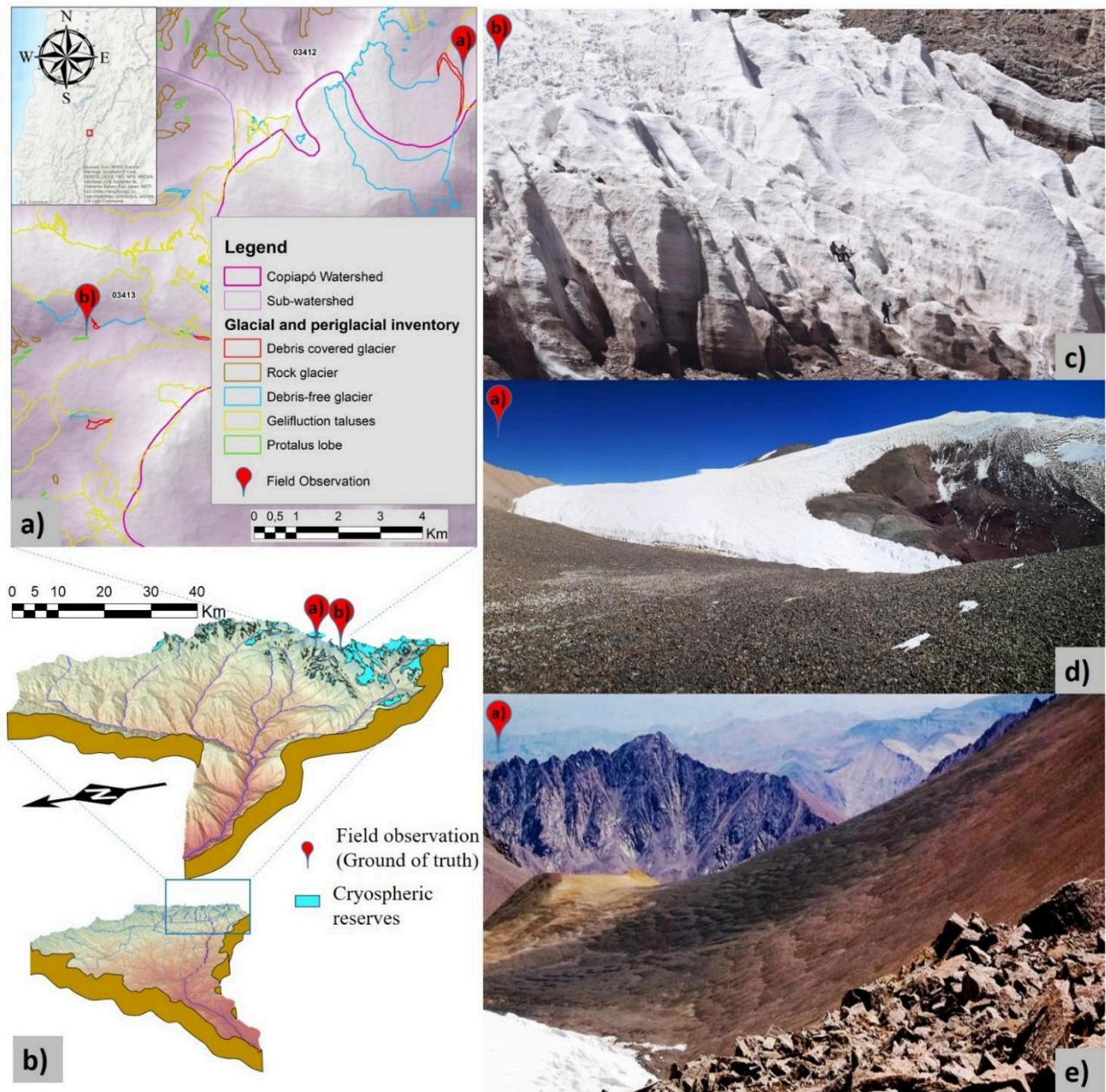


Figure 12. Field observations of cryospheric reserves as the groundtruthing. a) Map of cryospheric reserves distribution, b) 3D view of the watershed, c) Del Potro debris-free glacier of the Del Potro river (03412), d) Maranceles glacier of Montosa river (03413), e) Gelifluction taluses of the periglacial environment of Del Potro river (03412).

4.4 Geophysical results

4.4.1 Gelifluction Taluses

Figure 13, derived from electrical resistivity tomography (ERT), provides the basis for interpreting the internal structure of the gelifluction taluses. The lobate surface morphology characteristic of this cryospheric landform (Figure 3a, b), together with the location and orientation of the tomographic profile (Figure 3c, d), is clearly reflected in the geophysical response. The ERT profile extends over 295 m and crosses a sector of the landform displaying a well-developed pattern of freeze–thaw-related lobulations, starting at 4,430 m a.s.l. in the upstream sector. This geomorphological pattern coincides with a high-resistivity zone (HRZ), with resistivity values ranging from approximately 1,000 to 8,000 Ωm . According to comprehensive reviews and field studies on mountain permafrost, such resistivity values are typical of ice-rich frozen ground, where pore water is largely transformed into electrically resistive ice (Kneisel and Hauck, 2008). Electrical resistivity is known to increase by several orders of magnitude upon freezing, particularly in coarse-grained periglacial materials, making ERT a robust tool for detecting ice-bearing permafrost bodies.

The HRZ delineates an ice-rich permafrost body with an estimated two-dimensional horizontal extension of approximately 20 m² within the tomographic section. Comparable resistivity ranges and internal structures have been documented in ice-rich permafrost landforms investigated with ERT, including rock glaciers, talus slopes and other periglacial slope systems (Hauck and Kneisel, 2008; Hilbich et al., 2009). The association between lobate slope morphology and underlying frozen ground with significant ice content is consistent with the established understanding of gelifluction as a process operating on seasonally thawed active layers overlying permafrost, where ice lenses and frozen substrates exert a first-order control on deformation patterns (Matsuoka, 2001).

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-resistivity zone (LRZ) dominates, with resistivity values between 50 and 200 Ωm . This sharp contrast is interpreted as the transition from frozen ground to unfrozen, water-saturated sediments. Similar resistivity ranges have been reported for unfrozen debris, taliks and shallow groundwater zones underlying permafrost terrain, where electrolytic conduction through liquid water strongly reduces bulk resistivity (Kneisel and Hauck, 2008).

The spatial juxtaposition of the HRZ and LRZ suggests a direct subsurface connection between the ice-rich permafrost body and an underlying groundwater system. Studies combining ERT and seismic methods demonstrate that such resistivity contrasts commonly mark permafrost boundaries and zones of hydrological interaction, where water infiltration and ice loss can be detected geophysically (Hauck et al., 2008; Mewes et al., 2017). In particular, combined geophysical monitoring has shown that increases in liquid water content beneath or adjacent to frozen ground produce pronounced resistivity decreases, even when ice loss is spatially limited. Based on the vertical extent of the HRZ, the ERT data allow for the estimation of an average ice-rich core thickness of approximately 15 m (Figure 11). This thickness falls within the range reported for

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

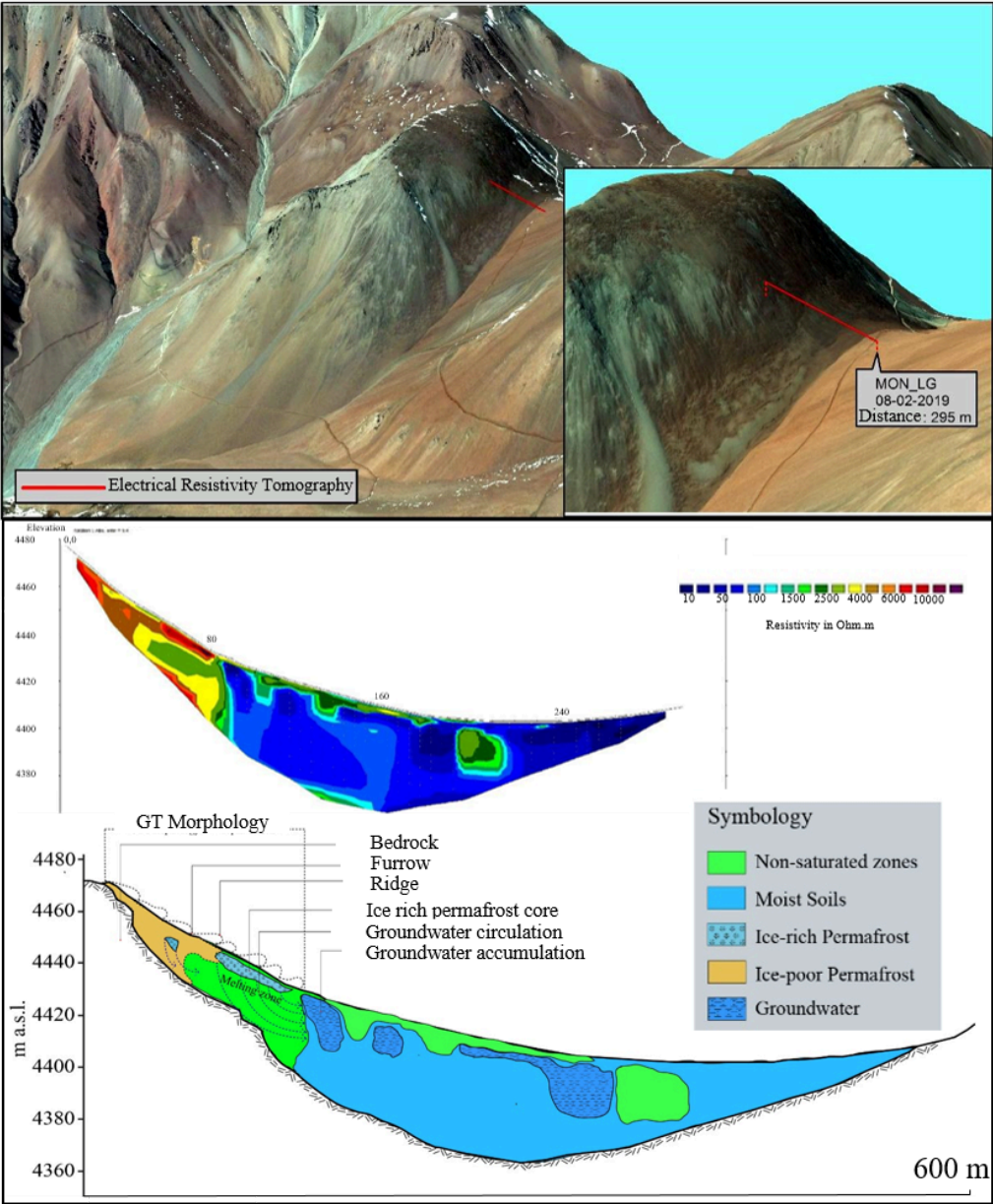


Figure 13. Results of the electrical resistivity tomography (ERT) over the gelifluction taluses cryoforms

4.4.2 Rock Glacier

The Electrical Resistivity Tomography (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 and Hauck, 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 based on DInSAR surface velocities (Cicoira et al., 2020).

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.

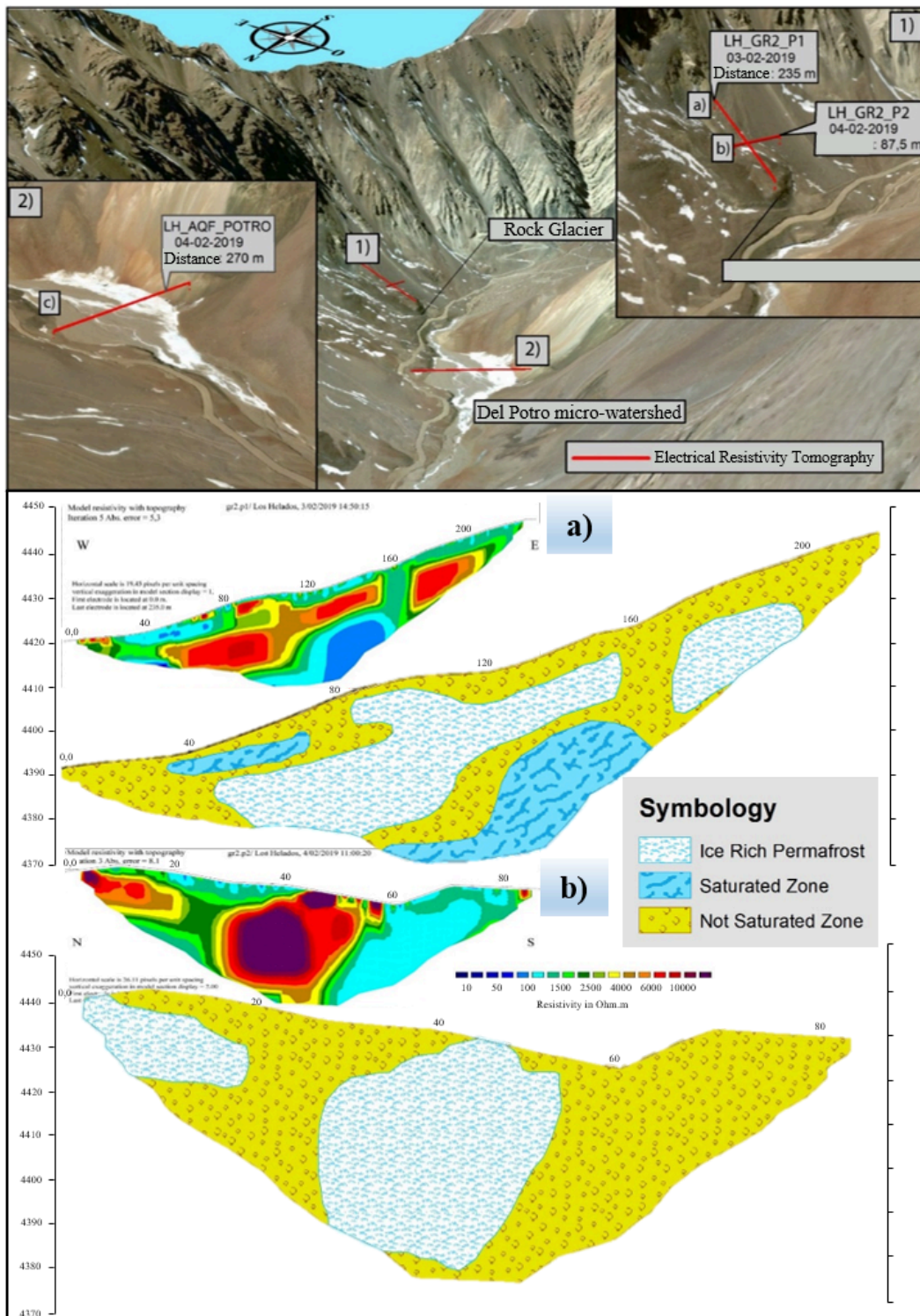


Figure 14. Results of the electrical resistivity tomography (ERT) over rock glacier cryoforms.

4.5 Thickness estimations on periglacial environment

Figure 15 shows the surface displacement of a representative rock glacier in the Copiapó watershed derived from Sentinel-1 DInSAR. Coherent downslope motion is clearly confined to the rock glacier body, while surrounding terrain remains stable, supporting the reliability of both the cryosphere inventory and the DInSAR processing. These surface velocities provide a direct kinematic constraint for the viscoplastic thickness modelling of ice-rich permafrost landforms used in this study.

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

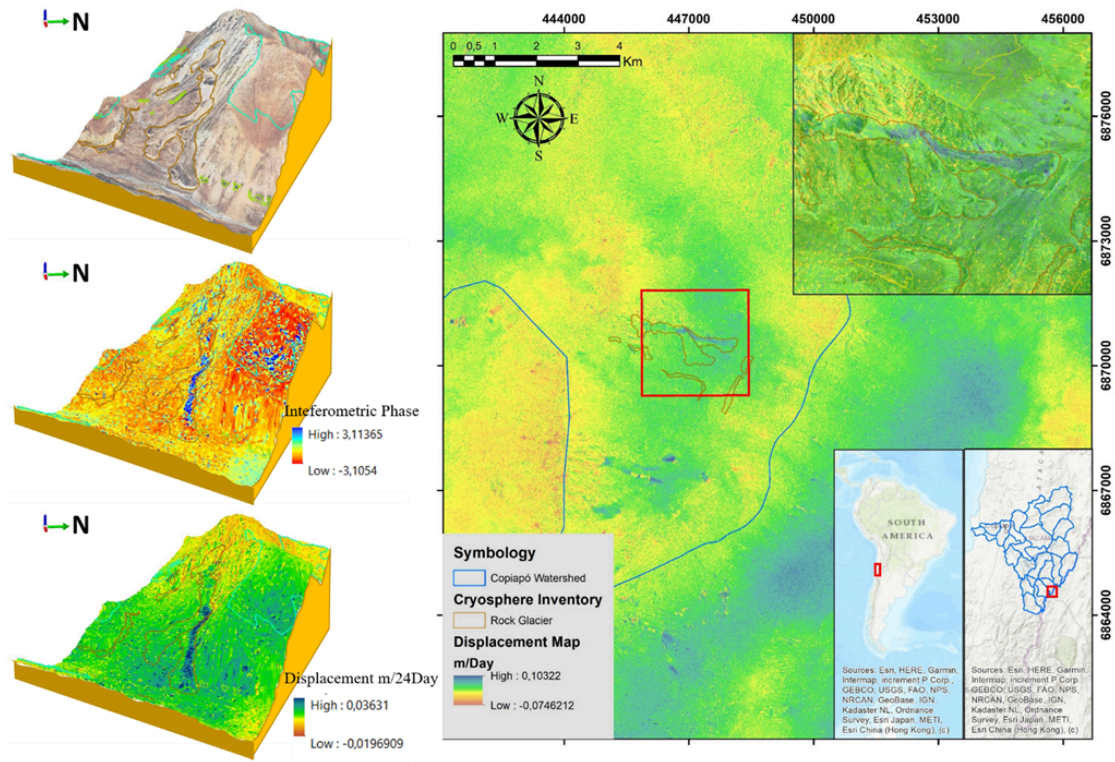


Figure 15. Example of DInSAR-derived surface displacement of a rock glacier in the Copiapó watershed.

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 cryoform groups remain clearly distinguishable in thickness–velocity space, supporting the application of a unified viscoplastic model while preserving cryoform-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.

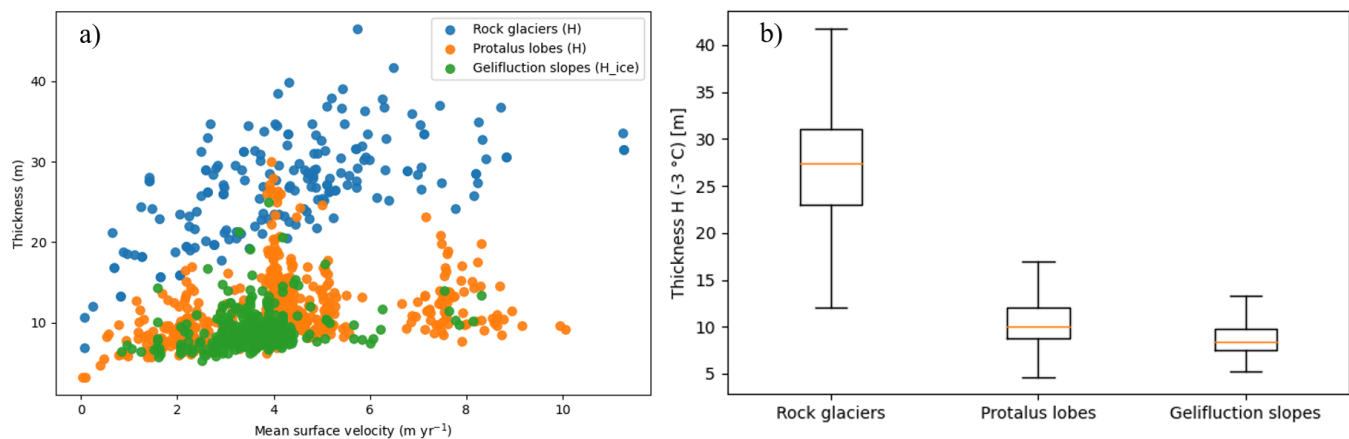


Figure 16. Modeled thickness–velocity relationship for periglacial cryoforms,

4.5 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 reserves of the Copiapo river watershed have been calculated (Table 6). The watershed with the largest ice reserves 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 cryospheric resources, 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 cryospheric resources 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).

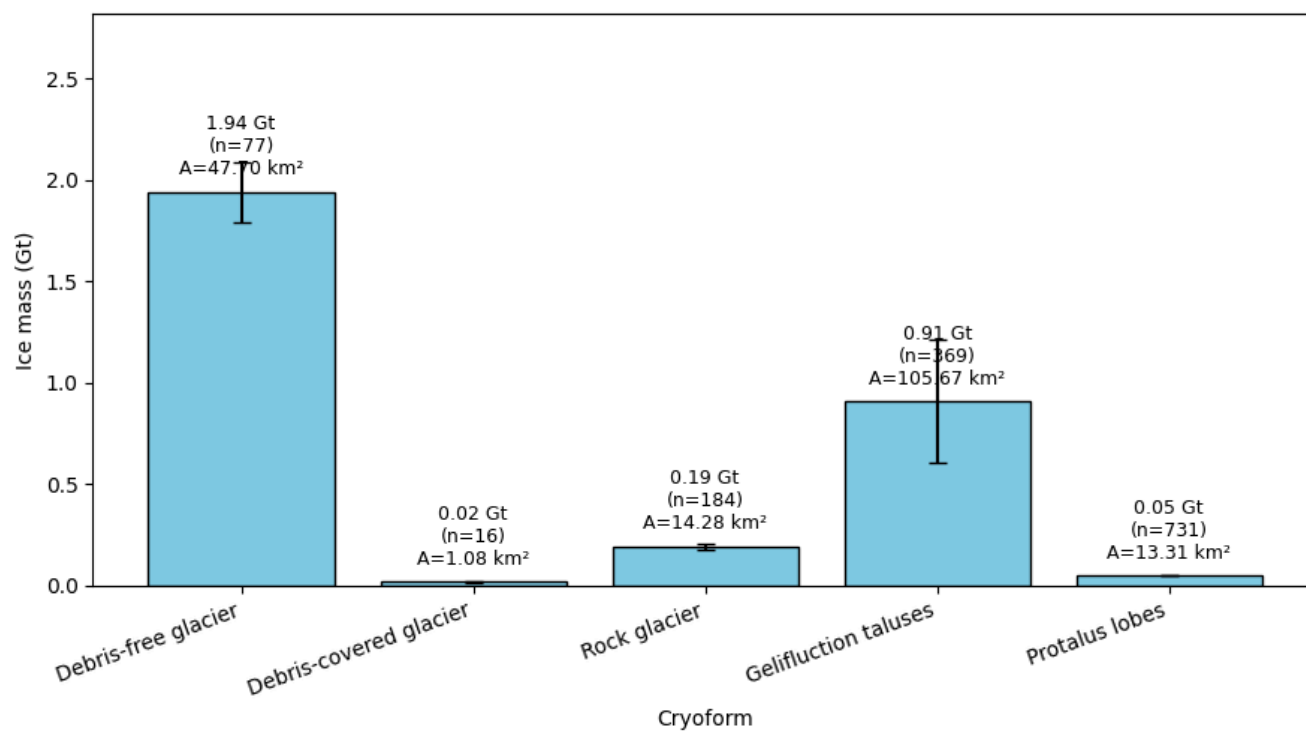


Figure 17. Total ice volumes of the Copiapo river watershed.

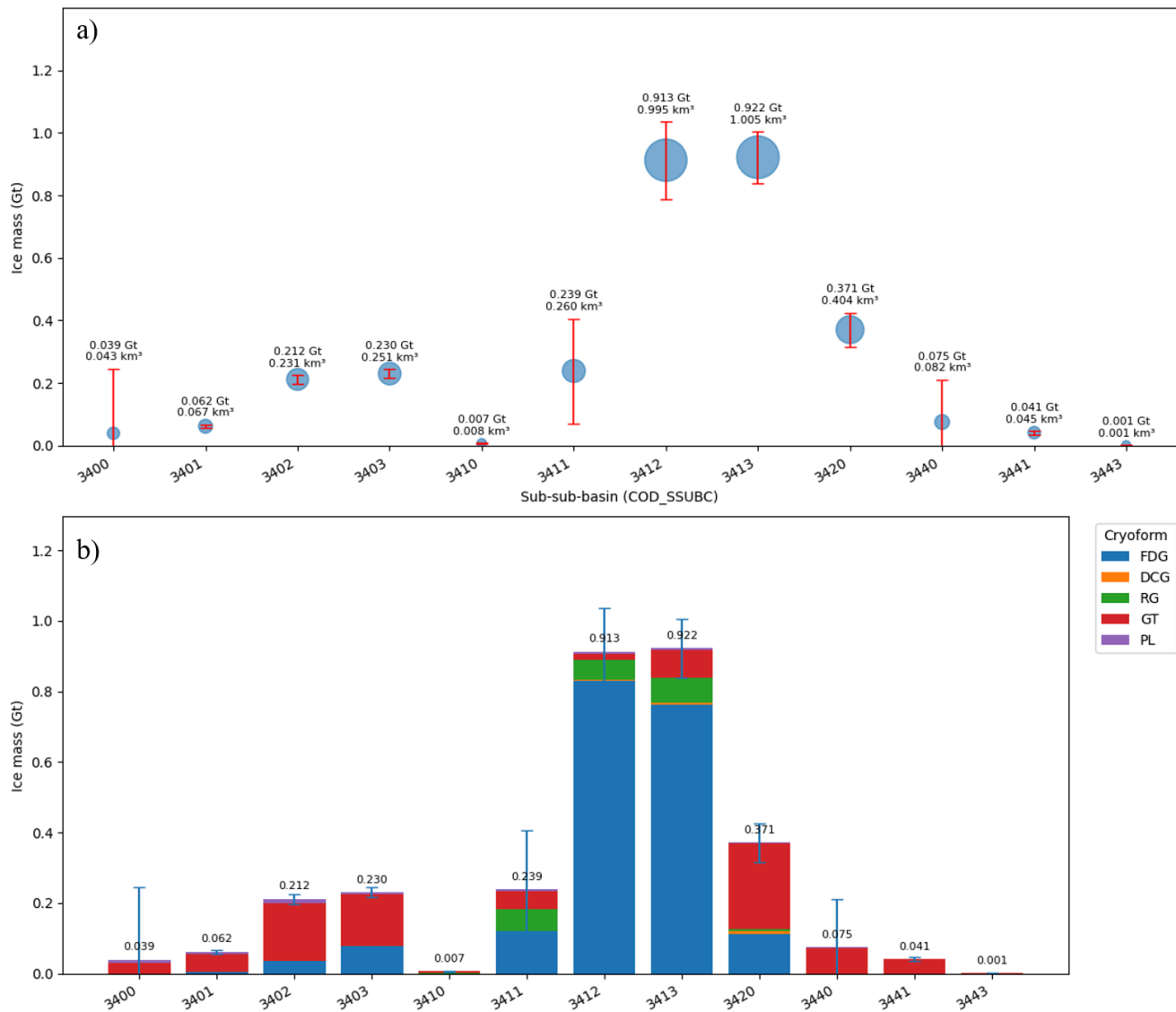


Figure 18. Total ice volumes of the Copiapó river watershed.

Figure 18 illustrates the distribution of ice mass across cryospheric sub-sub-basins of the Copiapó watershed, expressed both as total ice mass with associated uncertainty (panel a) and as stacked contributions by cryoform (panel b). The results reveal a highly heterogeneous spatial distribution of cryospheric reserves, with a small number of sub-sub-basins 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-sub-basins, each storing approximately ~0.9 Gt of ice. These basins clearly dominate the cryospheric reserves 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-sub-basins exhibit ice masses below 0.1 Gt, reflecting either limited cryoform 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 reservoirs, whereas ice-rich permafrost cryoforms—particularly rock glaciers and gelifluction taluses—represent a substantial fraction of the total ice mass in intermediate and smaller sub-sub-basins. Protalus lobes contribute minor but non-negligible volumes in selected basins, emphasizing their cumulative importance at the watershed scale.

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

Overall, these results demonstrate that cryospheric ice mass is strongly concentrated in a limited number of sub-sub-basins, while many others contain comparatively small but non-negligible ice reserves distributed among different cryoforms. 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 reserves (ice mass) and cryospheric resources (snow persistence) into a unified metric, introduced in the following section as the Normalized Cryospheric Index (NCI).

Figure 19 displays four panels that provide a comprehensive depiction of 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. Overall, the results show how the choice of weight (w) influences the mean value of the NCI and, ultimately, the final ranking of the sub-watersheds.

Panel (a) shows the evolution of the mean NCI value of each sub-watershed as w varies from zero to one. 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.

Panel (b) shows the position in the ranking that each subwatershed occupies for each w value. 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.

Panel (c) presents 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. This value quantifies the overall stability of the ranking for each w . A marked minimum is seen around $w=0.3$, indicating that, at that point, the ranking of watersheds is less sensitive to the

uncertainty of the data of V and S. In other words, assigning that weight in the combination of variables leads to a more consistent ranking in terms of robustness.

Finally, panel (d) shows the mean NCI (with its error bars) for each watershed when $w=0.3$, the optimal value detected with the optimization function. The vertical axis reflects the average magnitude of the index, while the error bars indicate the standard deviation derived from the Monte Carlo simulations. Thus, the visualization offers a clear identification of the watersheds with the highest NCI values, while simultaneously exposing the degree of uncertainty surrounding these estimates. This result confirms the relevance of choosing $w=0.3$ and provides a detailed view of the dispersion of each watershed under such a scenario.

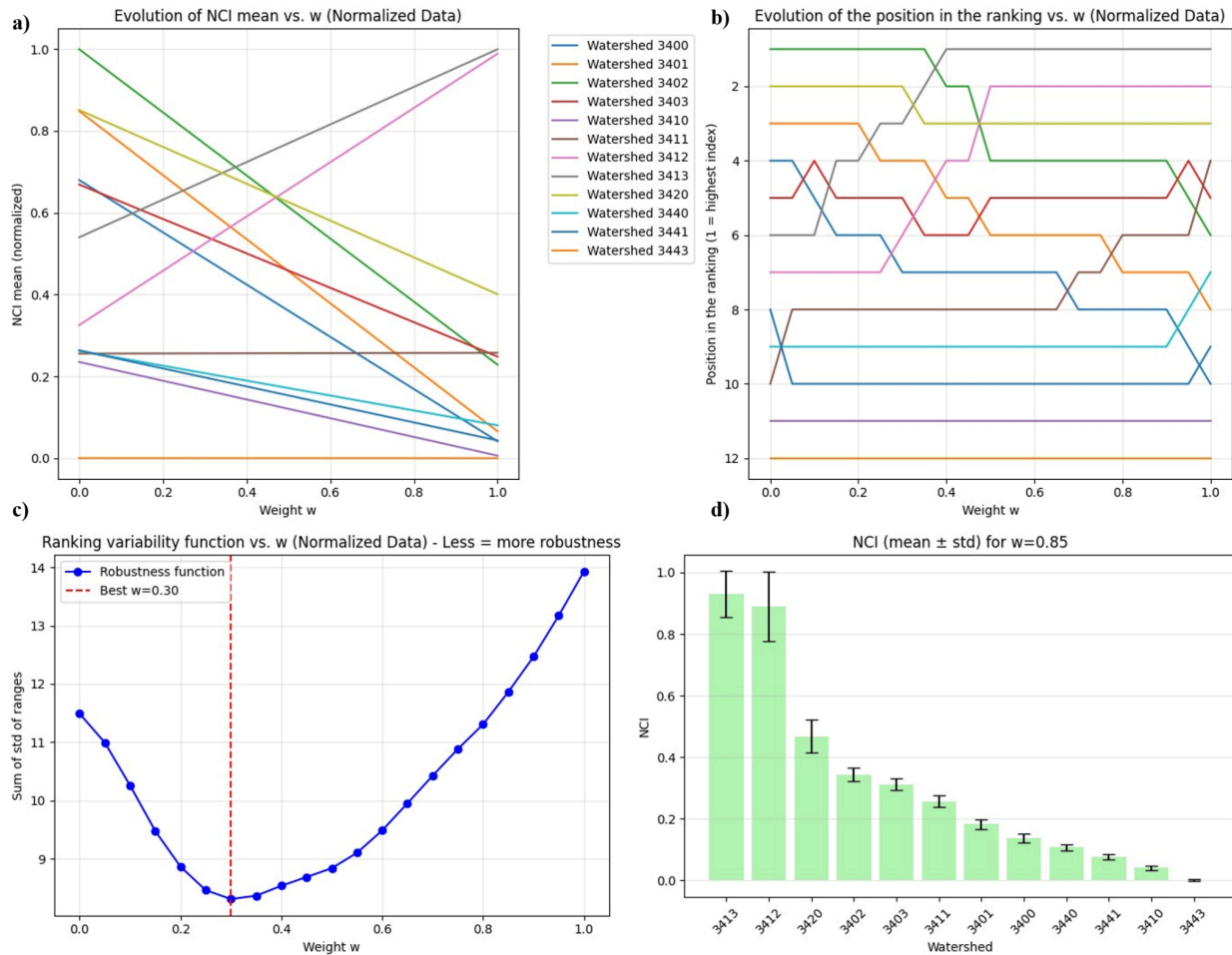


Figure 19. Analysis of NCI and robustness of the ranking as a function of w weighting (a) Evolution of the mean NCI for each sub-watershed as w varies between 0 and 1 (b) Position in the ranking of each sub-watershed as a function of w (c) Ranking variability function (sum of the standard deviations of the positions), where the minimum at $w \approx 0.3$ indicates the maximum stability of the ranking (d) Mean NCI $\pm$ standard deviation for $w=0.3$, evidencing the magnitude and dispersion of the index in each sub-watershed under the most robust condition.

The calculated NCI, demonstrated that the sub-watershed with the highest NCI was the by the Montosa river watershed (03413) (NCI = 0.93), followed by the Del Potro river watershed (03412) (NCI = 0.89), Manflas river watershed (03420) (NCI = 0.47), Estero Come Caballos watershed (03402) (NCI = 0.34) and Cachitos river watershed (03403) (NCI = 0.31). These are the sub-watershed with the largest cryospheric reserves. In the northern region, the Estero Come Caballos sub-watershed, it was highlighted by the index primarily attributed to its persistent snow cover.

Table 6 Summary of results by sub-watershed of reserves in Gt, resources in km² and indexes in different scenarios.

Code of the sub-watershed	Reserves gigatons	Reserves uncertainty	Resources 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 Cryoforms and snow permanence in the Copiapo watershed

From the cryoform and snow occurrence mapping exercises (Fig. 6 and 8), we observed that the cryospheric reserves of the Copiapo 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 cryoforms 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 cryoforms occur in smaller bodies (Fig. 6). This resulted in smaller, separated cryosphere's covering and surrounding the highest point (i.e. field observations in Figure 6 a, b and c). They host smaller cryospheric reserves, for example near the Cadillal and the Nevado Jotabeche, compared to the southern part of our study area. Because those northern cryospheric reserves are located in or in proximity of the SAAD (Abraham et al., 2020), snowfall and permanent snow occurrences are limited, leading to isolated cryoforms (Fig. 5). The potential volume of water that is captured in cryospheric reserves in the sub-watersheds of the Copiapo watershed (Fig. 1) is thus strongly influenced by the South American Arid Diagonal (SAAD) that controls the distribution of the different types of cryospheric reserves (García et al., 2017). In our case, the proximity of the Copiapo 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 cryoforms 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 cryoforms 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 cryoforms in the southern regions in our research area. A last negative spatial correlation between the snowfall and the cryoforms is the large snowfall anomaly in the northern area. The permanent snow occurrence is high, while not many cryospheric reserves 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 cryoforms.

From the different cryoforms, we see that gelifluction taluses cover the largest surface area in the Copiapo watershed compared to other cryoforms in our research areas. Field observations also confirm that gelifluction taluses extend across significantly broader areas compared to other cryoforms (Fig. 7). This is consistent with regional observations that gelifluction taluses cover most surfaces (Garcia et al., 2017). 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. (2015).

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. 10). In the north of the Copiapo 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 reserves. 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. 7). 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 Del Potro, the modeled volume is ~24% larger than the GPR-based estimate; at Tronquitos, 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 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.

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 (km³)	DGA 2014 GPR volume (km³)
Del Potro	1.715.087	0.362	0.293
Tronquitos	1.715.038	0.157	0.092

The watershed with the best balance between reserves and resources is the Montosa river watershed (03413) (Fig. 14), which contains high quantities of both cryospheric reservoirs and resources. The field observations of geomorphological features that are indicative of active cryoforms, such as the subparallel terraces (Fig. 10e), fit well with the mapped cryoforms from our workflow. The mapped surface area and thickness are supported by the observed surface area and thickness in the field. Lastly, debris-covered glaciers occur predominantly in areas with a relatively continuous cryosphere (García et al., 2017). This is consistent with our findings that debris-covered glaciers are exclusively located in the southern region of the study area, a zone characterized by a more extensive and contiguous cryospheric distribution relative to the fragmented northern sectors.

5.2 Classifying and prioritizing sub-watersheds in the Copiapo watershed

Unlike rainfall-dominated catchments, the upper Copiapó watershed is strongly influenced by cryospheric processes, with its headwaters characterized by the widespread presence of glaciers, rock glaciers, and ice-rich permafrost landforms. Cryospheric elements are concentrated in the high-altitude sub-watersheds, where water storage and seasonal runoff generation are controlled primarily by snow accumulation, snow persistence, and long-term ice storage rather than by direct rainfall inputs.

The Normalized Cryospheric Index (NCI) identifies three main sub-watersheds with a dominant cryospheric influence on the hydrological functioning of the Copiapó system, based on the combined contribution of ice mass and long-term snow persistence. The spatial distribution of cryospheric reserves shows a 97% overlap with high-probability permafrost zones predicted by the global permafrost model of Gruber (2012), providing independent support for the relevance of ice-rich permafrost as a key component of water storage in the basin. Together, these observations justify classifying the upper Copiapó as a cryosphere-influenced watershed, in which cryospheric processes exert a first-order control on water availability. The critical sub-watersheds include: the Montosa River (03413), Del Potro River (03412) and Manflas River (03420) (Table 6). Their spatial distribution corresponds to zones of maximum snow permanence and the most extensive cryospheric continuity within the Copiapó watershed (Figs. 6 and 8). These high indexes thus directly correlate with the cryoforms 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 cryospheric resources (snow persistence) and reserves (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). 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 Copiapo 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 third 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 Copiapo river.

Given the observed relationship between snowfall and cryoform 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 reserves 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 four sub-watersheds constitute the basis for continuous water production in the Copiapo watershed during dry years, because they host 83.44 % of the cryospheric reserves of the Copiapo watershed. From these four sub-watersheds, the Montosa river watershed (03413) maintains the best balance between resources and reserves (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 cryospheric resources (i.e. snowfall and precipitation) but are less enriched in cryospheric reserves (in the form of cryoforms). 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 reserves 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 Copiapo watershed in terms of cryospheric resources (Fig. 20).

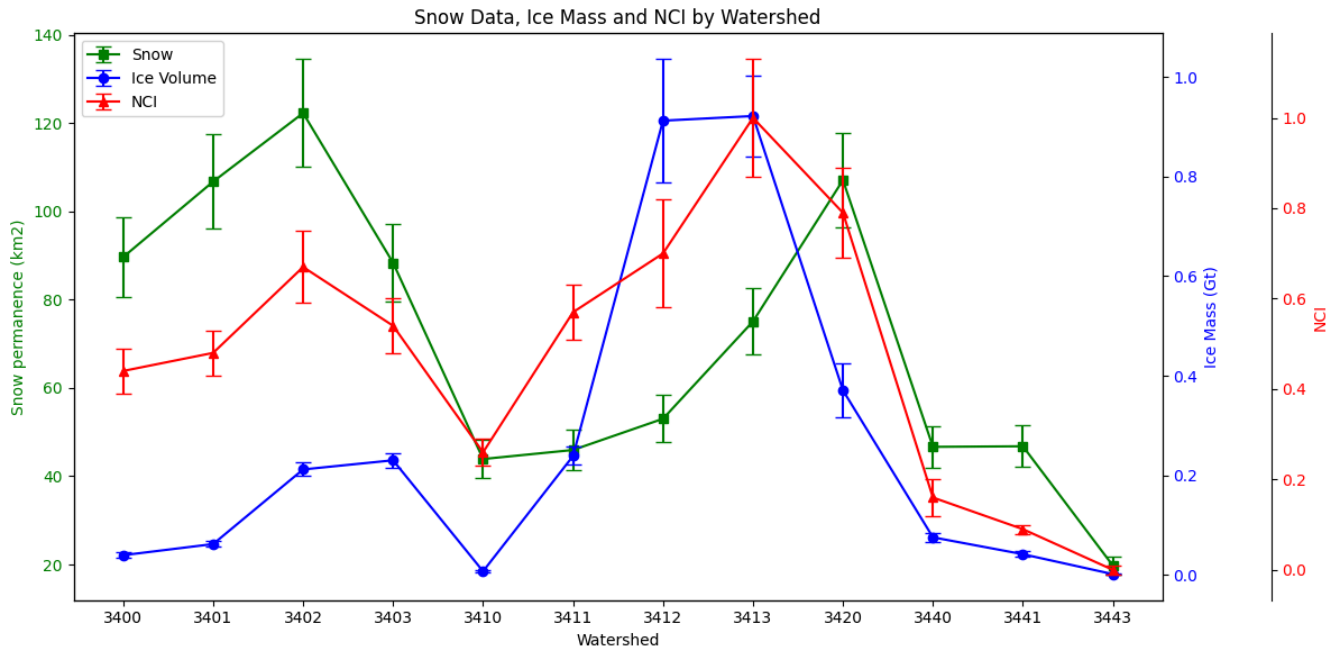


Figure 20. Summary graph of the normalized indexes for resources, reserves and the cryosphere for each sub-watershed of the Copiapo river watershed.

5.3 Impact on watershed management

We have analyzed 12 cryospheric sub-watersheds in the Copiapo 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 cryoforms 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 cryoforms.

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

6. Conclusions

6.1 Main findings of the paper

We have developed a new workflow to identify cryospheric reserves and resources and applied this method to the Copiapo watershed. The workflow consists of mapping cryoforms and snow occurrences and the calculation of a normalized cryospheric index (NCI) for sub-watersheds.

Our cryoform and snow occurrence mapping exercise resulted in the identification of cryospheric reserves in more isolated bodies north of Cadillal Hill and more continuous reserves south of Caserones Hill.

Gelifluction taluses constitute a critical strategic cryospheric reserve for safeguarding water security in northern Chile's watersheds. These landforms represent the periglacial cryoform type most likely to contain the largest ice volumes within the region's cryosphere, a hypothesis that requires further validation through targeted ERT (Electrical Resistivity Tomography) geophysical surveys of such cryospheric features.

The NCI allowed us to identify four main targets for watershed management. The four identified main targets are the Montosa river (03413), Del Potro river (03412), and Manflas river watershed (03420) with NCI of 0.93, 0.89 and 0.47 respectively. We propose that these sub-watersheds are the most important because that can sustain the runoff in the

watershed in times of drought. In years with snowfall the northern sub-watersheds as the Estero Come Caballos sub-watershed (03402) have the potential to be highly productive hydraulically by snow melt. This ranking allows the prioritization of these sub-watersheds.

We conclude that the Montosa river (03413), Del Potro river (03412) and Manflas river (03420) are the sub-watersheds that should be prioritized in the national water resources strategy because of the high NCI.

6.2 Limitations of this work and future research

The workflow developed in this study enables the assessment of the role of a specific cryospheric sub-watershed within a broader drainage system, providing a basis for establishing protection frameworks in water resource management. This tool is transferable to other arid regions that are sourced through cryospheric reserves and resources. However, the method is calibrated for the Copiapo river watershed, and complete resource and reserve data should be available for the NCI method to be used on other watersheds.

To ensure more accurate error propagation in the NCI definition, it is essential to compute the uncertainty associated with the fractional snow cover product. We believe that assuming a 10% error under ideal conditions is one of the main limitations of the study of the cryospheric resources of the watershed. Our goal is to obtain a code that allows us to understand the uncertainties associated with the detection of snow in the mountainous areas of the watershed in order to improve the NCI and take it to the next level of research.

6.3 Broader impacts

The tool provides technical and objective data that water control agencies and those in charge of integrated watershed management, such as the DGA (General Directorate of Water), oversight boards and groundwater communities, should consider when implementing policy on the watershed management of the Copiapo watershed.

This study could help other arid regions of the planet with cryospheric reserves and resources to be able to manage the territory with a strategic plan to address the management of these cryospheric components.

7. Open science statement

The data generated for this project will become available upon publication. The cryospheric reserves data can be downloaded via this link: [<https://doi.org/10.5281/zenodo.15633086>]. The glacial and periglacial inventory for the Atacama region can be downloaded via this link: [<https://doi.org/10.5281/zenodo.14921499>]. The raster image for historical snow data can be downloaded via this link [<https://doi.org/10.5281/zenodo.14921552>].

8. Acknowledgments

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9. References

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