

Glacier Hydrology Primer

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Introduction

The role of glaciers in the hydrological cycle: In the context of the hydrological cycle, glaciers can be thought of as water stores with the ability to delay the occurrence of runoff from the time at which precipitation occurs. This delay can range in scale from hours to centuries. Data from glacierized catchments around the world shows that even modest glacier presence in terms of area coverage relative to watershed size can significantly influence hydrological features such as streamflow interannual variability and minimum flows.

Globally, glaciers are not in equilibrium with the current climate. As a consequence, rapid and widespread retreat has been documented, which in turn implies that glacierized catchment hydrology is most likely to remain in non-stationary conditions until glaciers reach a new balance. This future state is uncertain in most cases and may range from total disappearance to a reduced volume and surface area, concentrated in the uppermost reaches of their current location.

Glaciers can be classified in many different ways, according to their geometry, their thermal regime, their location in relation to the local physiography, etc. This document pertains to *mountain glaciers* and does not cover aspects related to ice fields, ice caps, aufeis¹ and other

¹ Sheet-like mass of layered ice formed on top of river flow during subfreezing temperatures.

major glacier categories. Mountain glaciers in turn can be subdivided into many types. For example, cirque, hanging, globe, valley are some of the subtypes that have been proposed. In terms of the hydrological aspects, we will refer in this document to two main types:

- Clean glaciers: consist primarily of solid ice, firn and snow, which are freely exposed to the atmosphere throughout most of the glacier's surface area.
- Debris covered glaciers: a core formed by ice, whose surface is covered with a layer of rock debris. The covering layer may be of variable thickness, from a few centimeters to several meters.

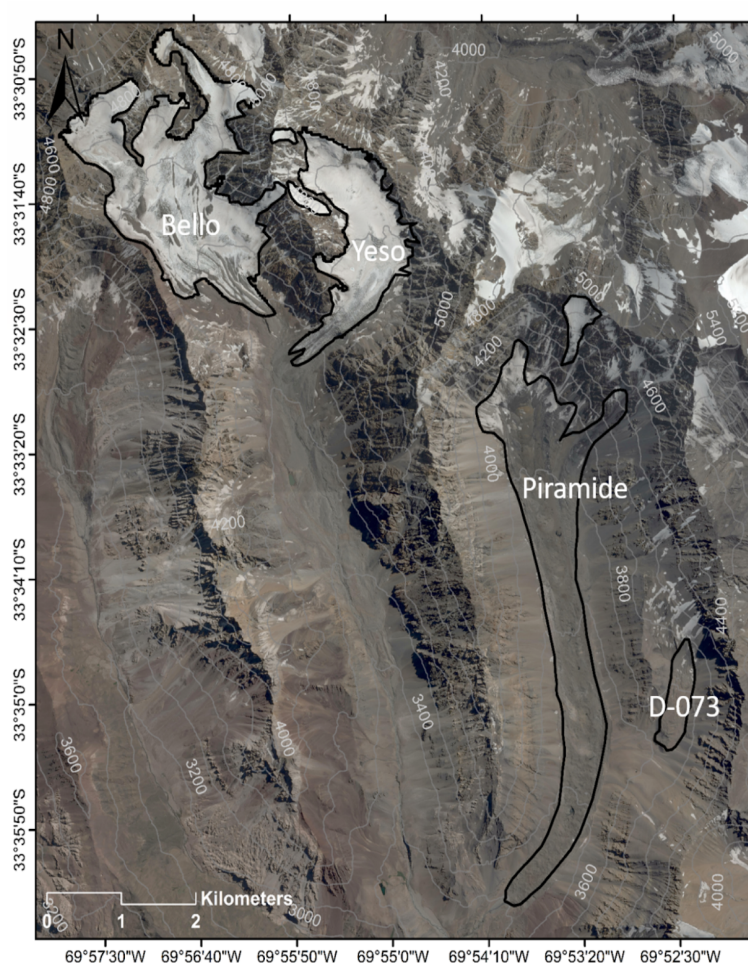


Figure 1. Two clean (Bello and Yeso) and one debris-covered glacier (Piramide) in the Maipo River Basin (Chile). Source: Google Earth.

Glaciers in the Hydrological Cycle

Glaciers are storage compartments in the hydrological cycle, much in the same way as the seasonal snowpack, or even the soil column. The main difference to these other forms of storage is that glaciers are concentrated in space, to few very specific locations, and that the *timescales* at which glaciers release water can vary widely (Jansson, Hock, and Schneider 2003). Ice melting at the glacier surface may have been formed decades or centuries before

from snow precipitation. Snow and firn on the glacier surface act as porous media, holding liquid water from precipitation and melt and releasing it after hours, days or months. Channel networks in a glacier's surface, core or base convey runoff to the glacier terminus with time delays of minutes to hours. Appropriately estimating the volume of water stored as ice at any given time, the rate of accumulation and melt and the dynamics of water movement in a glacier are important to properly link glacier hydrology to wider studies of current and future water availability in glacierized watersheds.

Within a glacierized region, it is possible estimate the stored volume in glacier ice from a scaling expression of the type,

$$V = cS^\gamma$$

Where S is surface area, and γ , c are empirical parameters derived from observed data. Despite their empirical origin, there is a theoretical basis to expect that the scale parameter c is a random variable and that γ should be a constant for glaciers belonging to the same geometry class (for example, valley glaciers versus lobe-shaped ice caps). This expression is derived from dimensional analysis and is valid for *groups* of, rather than individual, glaciers. For example, (Bahr, Meier, and Peckham (1997) reported values of $\gamma = 1.36$ for valley glaciers, and $\gamma = 1.25$ for ice-caps based on data from more than one hundred glaciers in three continents. Application to individual glaciers should be made with great care and awareness of its limitations (Bahr, Pfeffer, and Kaser 2015; Bahr, Meier, and Peckham 1997).

Glacier mass balance

Glacier mass balance represents the rate at which a glacier gains or loses ice mass over a given period. For practical purposes, the water year mass balance is usually computed as the sum of winter or accumulation (Figure 2, blue line; usually positive) and summer or ablation (Figure 2, red line; usually negative) mass balances. Accumulation and ablation may occur at any time during the water year, but it is usually accepted that, on average, distinct seasons exist when one process dominates over the other.

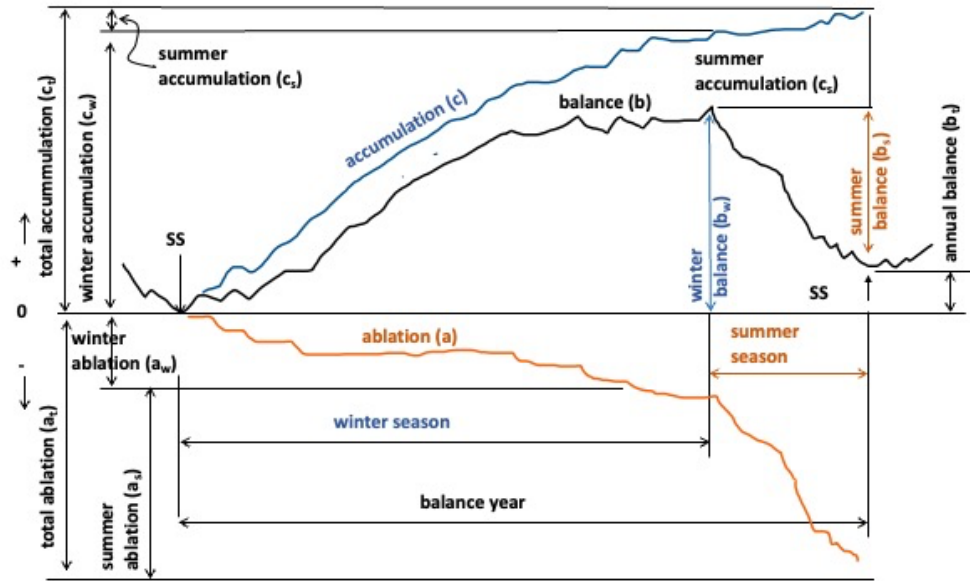


Figure 2. Schematic representation of annual glacier mass balance. Adapted from Cuffey & Paterson (2010)

To estimate winter mass balance, the following fluxes should be considered:

Direct snowfall on the glacier surface: precipitation rate and precipitation phase are needed to estimate this flux. Usually, no direct observations of snowfall on the glacier surface exist, so indirect estimates are required. Because mountain glaciers are relatively small compared to many global precipitation products, downscaling to local conditions is needed. For studies that aim at estimating historical glacier evolution, climate reanalyses (e.g. ERA5, CFSR) downscaled with dynamical (for example with NWP models) or statistical methods (e.g. bias correction) have shown to provide reasonable estimates of mountain precipitation.

Wind snow redistribution from adjacent areas and within the glacier surface extent: wind velocity and direction are greatly affected by the local topography, so it is required to estimate wind fields from regional estimates by downscaling. Examples of computer routines that accomplish this task include the *micromet* model (Liston and Elder 2006), and *windninja* (Wagenbrenner et al. 2016).

Gravitational redistribution from adjacent slopes: Many mountain glaciers are located within or adjacent to cirque formations or steep slopes from which deposited snow is mobilized by gravity. The amount and opportunity of snow mass input is determined by the surrounding slope and the snow stability conditions. A practical approach to quantifying these fluxes in a modeling setting is that of the *Snowslide* routine (Bernhardt and Schulz 2010).

Summer mass balance is dominated by ablation processes. These include surface melt, surface sublimation and evaporation, and, in some cases, calving of the glacier snout into a water body. Calving is less common in mountain glaciers but can occur when a proglacial lake forms adjacent to the tongue of a retreating glacier.

Energy balance determines ablation processes at the glacier surface. Energy exchange between the ice and the atmosphere results in heat being gained or lost by the ice. Excess energy is released through mass fluxes with melted ice or by evaporation losses to the atmosphere. The energy balance at the glacier surface is given by,

$$M = S + L + H + LE + P + G$$

where M is the energy available for melt, S is net solar shortwave radiation, L is net longwave radiation, H is sensible heat flux, LE is latent heat flux, P is heat transfer due to precipitation and G is conductive heat transfer below the surface. Energy balance at the glacier surface is dominated by local atmospheric conditions, which may vary significantly from regional weather because of local cooling, shading from adjacent slopes, and local wind patterns. Atmospheric variables recorded from automatic weather stations (AWS) located on the glacier surface are thus necessary to conduct detailed studies, and to develop parameterized transfer functions to regional weather conditions so that modeling exercises can be performed even in the absence of local data.

Research shows that individual terms dominating the energy balance can change for the same glacier during an ablation season. For example, Ayala et al. (2017), through field data and numerical modeling propose a conceptual scheme in which sublimation (latent heat flux) is the dominant mass flux term in the early stages of the ablation season for glaciers in dry environments, whereas melt becomes the dominant flux towards the end of summer (Figure 3). Refreezing is a relatively secondary term in the energy partition, with more prominence in the early ablation season than at later dates. The Tapado Glacier, at roughly 4500 masl and 27°S in the Chilean Andes, is a good example of this phenomenon.

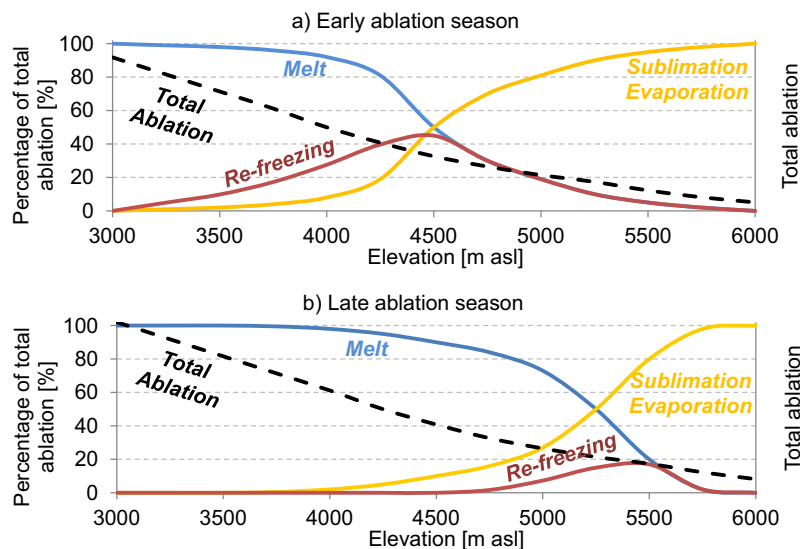


Figure 3. Hypothesized profiles of dominant terms in surface ablation from debris-free mountain glaciers in semi-arid regions. Adapted from Ayala et al. (2017)

An important distinction must be made between clean and debris-covered glaciers in this respect, because for the latter, sensible heat conduction from the surface to the underlying ice

is the only process driving ice melt. Reid & Brock, (2010) proposed an energy balance model for debris covered glaciers, that computes melt as a function of the sensible heat transferred from the debris layer to the ice:

$$G_i(t) = -k_d \left(\frac{dT_d}{dz} \right)_{base}$$

and

$$M = \frac{G_i \Delta t}{\rho_i L_i}$$

Alternative methods include so-called *temperature index* models, which estimate daily melt from empirically derived relations to air temperature (and shortwave radiation, in some cases). These methods are discussed in a previous [ICSIH paper](#).

A major challenge when estimating surface ablation is to estimate the glacier albedo, which has enormous impact on shortwave energy exchange and thus on available energy for melt. Recent evidence from remotely sensed data suggests glacier darkening in recent decades, and that this darkening can be correlated with local climate (Shaw et al. 2021). Another important problem is to appropriately estimate air temperature across the glacier surface because this variable is usually required to diagnose which areas of a glacier are contributing or not to melt (Ragettli and Pellicciotti 2012).

As a glacier loses mass, its area changes as well. Therefore, in any modeling exercise a method for keeping track of area changes is needed. Usually, updating of the estimated glacier area is done once every several time steps of the simulation model, as it is accepted that area changes are much slower than the hourly or daily time step required to run temperature-index or energy balance melt calculations. A good starting point is the Δh -parameterization proposed by Huss et al., (2010), which provides an estimate of glacier thinning due to mass loss. From this starting point, area updates can, at least in theory, be attempted in long-term future glacier evolution projections.

Glacier melt runoff modeling

In order to estimate water contributions from melting glaciers, the first challenge is to estimate the daily melt rate, M , at the glacier surface. Then melt water must be routed through some scheme to the glacier outlet, from where it is further routed to the surface or subsurface flow system depending on local conditions.

It is possible to explicitly compute M by solving the energy balance equation at the ice surface. This is achieved thanks to advances in estimating local meteorological conditions from numerical climate products, given that local measurements rarely are available. Once daily melt rates are computed, they must be aggregated into runoff amounts and routed to the

glacier terminus. Liquid water from ice, firn or snow melt at the glacier surface must travel through the glacier hydrological system before reaching downstream areas. A linear reservoir approach, though simple, has proven very skilled in reproducing observed runoff from glacierized areas. The formulation states that, for a draining reservoir, runoff at time t can be linked to some initial condition through the expression,

$$Q(t) = Q_0 \exp(Kt)$$

where K is the time constant and Q_0 is the initial condition. Typically, parallel reservoirs are set for ice, firn, and snow in hydro-glaciological applications. Each reservoir has a distinct time constant; faster for ice, slower for firn, and somewhat in between for snow (Figure 4).

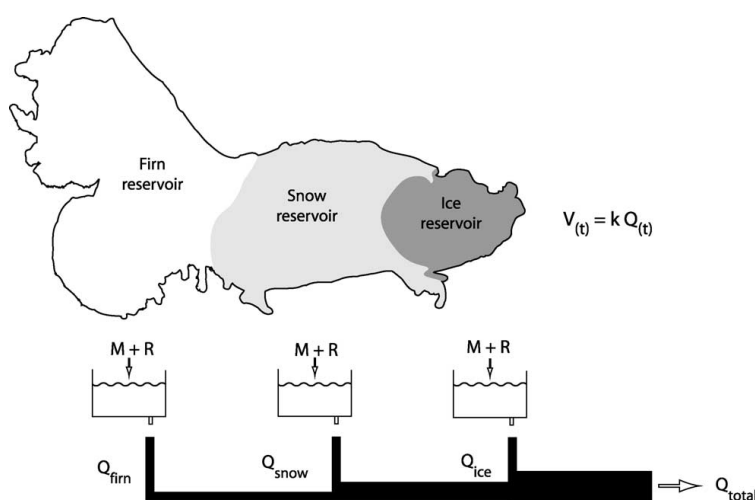


Figure 4. Illustration of three-reservoir concept based on a hypothetical glacier outline. Each reservoir is fed from melt (M) and rainfall (R). Runoff (Q) is proportional to reservoir volume (V) through a empirical constant, k. Surface extents of firn, snow and ice are illustrative only.
Adapted from Jansson et al. (2003)

New developments and technologies

The vast majority of glaciers in the world are non-monitored, which means that knowledge gained from direct observations gathered in a few controlled glaciers in mountainous regions around the world must be transferred in space (and time) if one is to estimate the evolution and hydrological contribution of glaciers regionally. Remotely sensed properties offer the opportunity to complement this transfer problem by providing estimates of relevant processes and properties.

Glacier extent (surface area) and surface albedo can be mapped with optical sensors mounted on a variety of platforms (e.g. [MODIS](#), [Landsat](#), [Pleiades](#)). Ice flow velocity requires more detailed information, such as can be obtained from radar data ([Sentinel-1](#), CryoSat, [AMSR2](#)). Repeat surface topography (DEM) estimates can serve to approximate glacier mass balance, and can be obtained from altimetry missions, radar, or structure from motion from space (Pleiades) or UAVs. Lidar is also a good source of surface elevations, but to date can mostly

be deployed from terrestrial platforms (TLS) or aerial manned or unmanned vehicles. More on UAV-based mapping can be found in this [ICSIH paper](#), and more extensive reviews on remote sensing of the cryosphere can be found in Tedesco (2014).

Conclusions

Estimating the role of glaciers in the hydrological cycle at the watershed scale requires proper representation of the amount of ice stored, and the rate at which the ice melts and contributes to the runoff cycle. Because most glaciers are located in remote areas and are un-monitored, indirect estimation methods are essential, which introduces several sources of uncertainty to this exercise. The first step is to properly characterize local atmospheric conditions, which are themselves modified by the presence of glaciers. Downscaling of reanalysis or NWP data can prove useful. Ablation estimation requires knowledge of the energy exchange characteristics between the atmosphere and the glacier surface, and its representation through modeling tools that vary in complexity. Regardless of the approach employed, key steps that must be taken when approaching the problem of hydro-glaciological estimation include,

- Obtaining meteorological inputs at the appropriate spatial scale.
- Obtain a diverse dataset for model calibration and evaluation, including glacier runoff, snow cover, glacier topography, surface ablation and energy balance (AWS), etc.
- Choose an appropriate glacier spatial representation and modeling scheme.
- Evaluate parameter sensitivity and assess transferability across time and space.

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