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Interannual variability in glacier contribution to runoff from a high-elevation Andean catchment: understanding the role of debris cover in glacier hydrology

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Abstract

We present a field-data rich modelling analysis to reconstruct the climatic forcing, glacier response, and runoff generation from a high-elevation catchment in central Chile over the period 2000–2015 to provide insights into the differing contributions of debris-covered and debris-free glaciers under current and future changing climatic conditions. Model simulations with the physically based glacio-hydrological model TOPKAPI-ETH reveal a period of neutral or slightly positive mass balance between 2000 and 2010, followed by a transition to increasingly large annual mass losses, associated with a recent mega drought. Mass losses commence earlier, and are more severe, for a heavily debris-covered glacier, most likely due to its strong dependence on snow avalanche accumulation, which has declined in recent years. Catchment runoff shows a marked decreasing trend over the study period, but with high interannual variability directly linked to winter snow accumulation, and high contribution from ice melt in dry periods and drought conditions. The study demonstrates the importance of incorporating local-scale processes such as snow avalanche accumulation and spatially variable debris thickness, in understanding the responses of different glacier types to climate change. We highlight the increased dependency of runoff from high Andean catchments on the diminishing resource of glacier ice during dry years.

KEYWORDS
debris-covered glaciers, dry Andes of Chile, glacier mass balance, glacier runoff, glacio-hydrological modelling

1 | INTRODUCTION

Seasonal snow and glacier melt in the semiarid Chilean Andes provide water to more than two thirds of Chile's population as well as maintaining key economic activities, ecosystems, and ecosystem services (Favier, Falvey, Rabatel, Praderio, & López, 2009). Central Chile is characterized by warm and dry summers, and humid cold winters, and ice melt provides a key contribution to runoff in dry periods and during late summer and autumn, in a water balance otherwise dominated by snowmelt (Ayala et al., 2016; Ohlanders, Rodriguez, & McPhee, 2013; Ragettli & Pellicciotti, 2012; Ragettli, Pellicciotti, Bordoy, & Immerzeel, 2013; Rodriguez, Ohlanders, Pellicciotti, Williams, & McPhee, 2016). While glaciers in the region have been receding and losing mass over the past few decades (Barcaza et al., 2017; Bown, Rivera, & Acuña, 2008; Malmros, Mernild, Wilson, Yde, & Fensholt, 2016; Mernild et al., 2015; Ragettli, Immerzeel, & Pellicciotti, 2016; Rivera, Acuña, Casassa, & Bown, 2002), the runoff response to climate and glacier changes is still poorly understood. Recent trends of decreasing runoff from high-elevation catchments (Casassa, López, Pouyaud, & Escobar, 2009;Mernild et al., 2016 ; Ragettli et al., 2016) suggest that the peak runoff, corresponding to the maximum contribution from a catchment (Huss & Hock, 2018; Pellicciotti, Bauder, & Parola, 2010), was reached at some time in the past. Results obtained from advanced glacio-hydrological modelling at relatively high resolutions (Mernild et al., 2016; Ragettli et al., 2016) and trend analysis (Casassa et al., 2009) are in agreement with global-scale models that also show declining trends in runoff from central Andean catchments (Bliss, Hock, & Radić, 2014; Huss & Hock, 2018). However, these analyses have been based on either a few intensively studied glaciers (Ragettli et al., 2016) or obtained from large-scale studies with grid resolutions too coarse to capture differences caused by important local-scale processes (Huss & Hock, 2018; Mernild et al., 2016). Multidecadal studies focusing on the processes generating glacier streamflow response to a changing climate are needed to bridge this scale gap.

Recent studies in the region have advanced our understanding of the spatial patterns of ablation and glacier mass balance (Ayala et al., 2016; Ayala, Pellicciotti, & Burlando, 2017; Bravo, Loriaux, Rivera, & Brock, 2017), but they are generally limited to a maximum of a few seasons of data and none has investigated decadal changes of mass balance and runoff. Time series analysis of observational records (Burger, Brock, & Montecinos, 2018; Casassa et al., 2009) is a useful tool to establish general data trends but is of limited use when observations are scarce in space and time and cannot provide insights into which processes drive observed changes. Additionally, while satellite-based glacier inventories (Barcaza et al., 2017; Malmros et al., 2016; Nicholson et al., 2010; Rabatel, Castebrunet, Favier, Nicholson, & Kinnard, 2011; Rivera et al., 2002) aided the establishment of baseline areal changes, they do not generally assess mass balance or volumes change and cannot be used to explain the causes of observed changes. Therefore, there is a need for an integrated approach to understand the midterm and long-term changes in glaciers and glacier runoff in the high-elevation catchments of the central Andes that combines both high-resolution glacio-hydrological modelling and mass balance estimates from remote sensing (Pellicciotti, Ragettli, Carenzo, & McPhee, 2014), which are increasingly used to evaluate model simulations.

Determining glacier mass balance regimes and glacier hydrological contribution in the Andes is further complicated by the presence of debris-covered and rock glaciers, which account for approximately 3,200 km^2 of the 23,700 km^2 of inventoried ice (Barcaza et al., 2017). While the contribution of these glaciers to high-elevation streamflow is poorly understood (Ayala et al., 2016), increasing evidence from other mountain regions shows a very distinct response of debris-covered glaciers to climate change compared with clean ice glaciers (Benn et al., 2012; Rowan, Egholm, Quincey, & Glasser, 2015). In one of the first glacio-hydrological modelling studies to explicitly include debris-covered glaciers, Ayala et al. (2016) showed that the contribution of a debris-covered glacier to total runoff over 2 years was of a similar magnitude to that of two debris-free glaciers in the same catchment.

Here we take advantage of the rare opportunity afforded by a well-instrumented catchment, the Rio del Yeso, to understand the interannual variability of glacier mass balance and glacier contribution to runoff over a 16-year period at the start of the present century (2000–2015). Our main aims are to (1) reconstruct the glacier changes for the period 2000–2015 and (2) estimate the corresponding glacier contribution to runoff for the period. These aims are addressed through application of a physically oriented and fully distributed glacio-hydrological model, in situ data, and the first geodetic mass balance estimates for the region. We use this combination of modelling, field data, and satellite observations to compare the hydrological contributions of a debris-covered glacier and two debris-free glaciers in the study catchment.

2 | STUDY SITE AND DATA

2.1 | Study site

The Rio del Yeso catchment (33.55°S, 69.91°W, 3,000–5,230 m asl, 62 km^2) is located ~70 km east of Santiago in the semiarid Andes of central Chile (Figure 1) and contains three principal glaciers: Bello, Yeso, and Piramide. The former are small, debris-free valley-type glaciers (4.6 and 2.2 km^2, respectively), while the debris-covered Piramide Glacier covers a larger elevation gradient although has a total area similar to that of Bello Glacier (4.7 km^2). Piramide has a typical reverse ablation gradient and an estimated debris thickness ranging from 0.01 to 0.6 m (Ayala et al., 2016; colour scale in Figure 1a). Mixed snow-debris avalanches typically feed the highest elevations of the glacier from local headwalls.

2.2 | Meteorological data

The model was forced with meteorological data (temperature and precipitation) from Yeso Embalse (YE) Automatic Weather station (AWS) from the Chilean Water Directorate (Dirección General de Aguas, DGA) meteorological network (Figure 1b). This AWS recorded daily maximum and minimum temperatures and daily precipitation for the entire simulation period (2000–2015). Additional temperature data from AWSs in the catchment were also used (Table 1). Hourly and daily maximum and minimum temperatures from the Laguna Negra AWS (2,780 m asl; Figure 1b) were used to identify the best disaggregation approach and evaluate the performance of the selected approach to derive hourly data from the temperature time series at YE AWS. Lapse rates used to extrapolate air temperature from YE AWS to the entire catchment were calculated using YE (2,475 m asl), Yeso off-glacier (4,300 m asl), Piramide off-glacier (3,020 m asl), and Piramide on-glacier (3,494 and 3,655 m asl) AWSs (Figure 1c) between 2013 and 2015, using common data periods for each month (Table 2). Air temperature lapse rates were validated against hourly temperatures measured at AWSs installed on Bello and Yeso glaciers. Finally, daily cloud transmissivity coefficients were derived from reanalysis ERA-Interim data by considering constant values during the day.

2.3 | Digital Elevation Models

Digital elevation models (DEMs) were used as both a basis for model runs (2,000 SRTM at 30-m resolution), as well as to quantify glacier

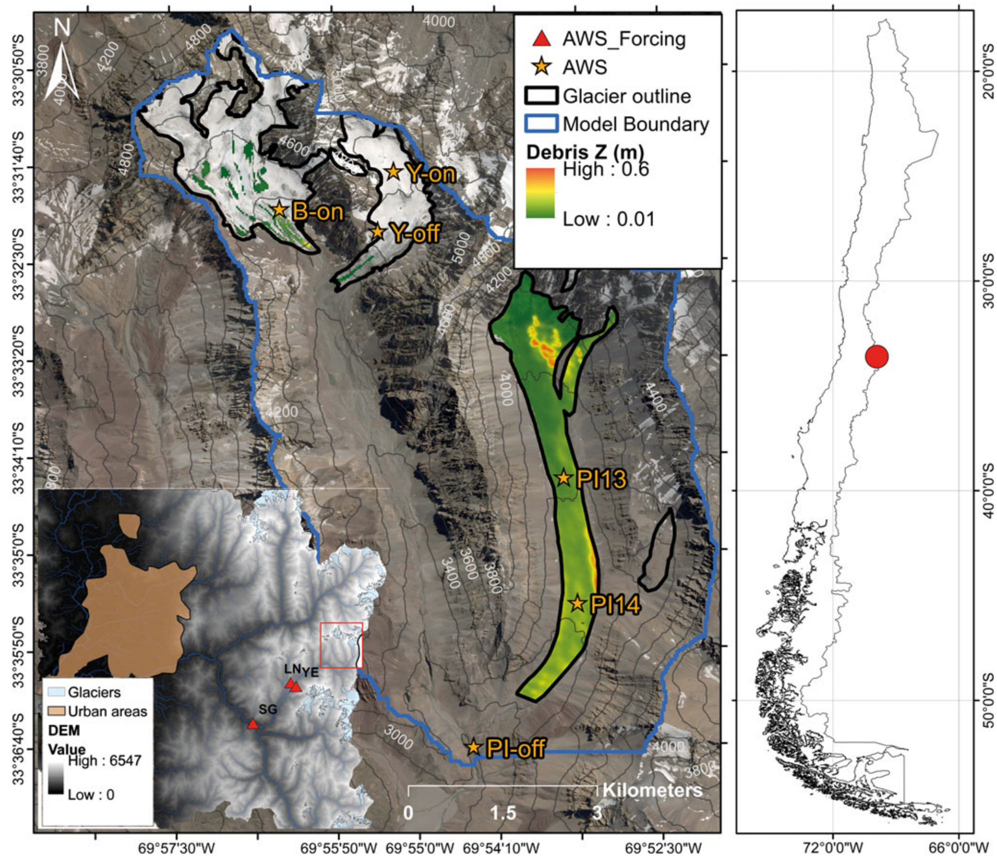

FIGURE 1 (a) Location of the study area in Chile; (b) location of the catchment (red box), near Santiago; and (c) map of the Rio del Yeso catchment including the study glaciers, the local automatic weather stations (AWSs), and the estimated debris thickness map. The locations of Yeso Embalse (YE) and Laguna Negra (LN) weather stations are shown in (b)

thinning through the 16-year simulation period. For the DEM differencing used to validate the model simulations, we used the bistatic TanDEM-X and SRTM-C for the period 2000–2013 and two repeated airborne light detection and ranging (LiDAR) surveys for 2012–2015. TerraSAR/X and TanDEM-X (TDX) correspond to an ongoing satellite constellation launched by the German Aerospace Center (DLR) and Airbus Defense and Space. TDX has a swath width of 30 km with a ground resolution of 0.4 arcsec. The Shuttle Radar Topography Mission (SRTM) was an interferometric synthetic aperture radar (InSAR) mission carried out simultaneously in the C- and X-band frequencies between 11 and 22 February 2000 between 56°S and 60°N (Farr et al., 2007). We used the void-filled Land Processes Distributed Active Archive Center National Aeronautics and Space Administration version of the SRTM DEM at 1-arcsec resolution. The two airborne LiDAR surveys carried out by the DGA have an estimated precision of ±0.30 m with an average of four points per square meter (Dirección General de Aguas, 2012, 2015).

2.4 | Terrestrial photographs

Daily photographs were taken by a time-lapse camera installed in front of Bello Glacier (Figure 1c) on 27 February 2014, covering an area of 0.84 km^2 (Ayala et al., 2016), for which 48 valid photos are available between February to April 2014 and 114 between October 2014 to April 2015. The system included a Canon EOS Rebel T3 camera with a resolution of 12.2 MP and a focal length of 18 mm. The camera was programmed to take photos every day at 13 hr (local time). Photos were georeferenced following Corripio (2004), and albedo was derived from the terrestrial photos as explained in Ayala et al. (2016).

3 | METHODS

3.1 | TOPKAPI ETH model

We used the TOPKAPI-ETH model to simulate glacier mass balance, glacier changes, and runoff generation in the Río del Yeso catchment. TOPKAPI-ETH is an extended version of the original rainfall-runoff model TOPKAPI (TOPographic Kinematic APproximation and Integration; Ciarapica & Todini, 2002 and Liu & Todini, 2002), and it has been applied in glacierized catchments from a few tens to more than 30,000 km^2 in the semiarid Andes (Ayala et al., 2016; Ragettli et al., 2014; Ragettli & Pellicciotti, 2012), the Swiss Alps (Fatichi, Rimkus, Burlando, & Bordoy, 2014; Fatichi, Rimkus, Burlando, Bordoy, & Molnar, 2015; Finger, Heinrich, Gobiet, & Bauder, 2012; Finger, Pellicciotti, Konz, Rimkus, & Burlando, 2011), and the Himalaya (Ragettli et al., 2013, 2016; Ragettli, Cortes, Mcphee, & Pellicciotti, 2014).

TABLE 1 Characteristics of the automatic weather stations (AWS) used in the study together with the variable recorded (T: temperature and P: precipitation) and the period of record

AWS			Elevation		
Map naming	Name	Location	(m asl)	Variable used	Period of record
YE	Yeso Embalse	33.68°S, 70.09°W	2,475	Daily T (Max and Min) P	1999–2015
LN	Laguna Negra	33.66°S, 70.11°W	2,780	T (Daily max and min; hourly)	2013–2015
B-on	Bello on-glacier	33.53°S, 69.93°W	4,134	T (Hourly)	November 2013 to April 2014 October 2014 to June 2015
Y-on	Yeso on-glacier	33.52°S, 69.92°W	4,428	T (Hourly)	November 2013 to April 2014 November 2014 to April 2015
Y-off	Yeso off glacier	33.53°S, 69.92°W	4,300	T (Hourly)	November 2013 to April 2014
PI13	Piramide on-glacier	33.57°S, 69.89°W	3,655	T (Hourly)	November 2013 to April 2014
PI14	Piramide on-glacier	33.59°S, 69,89°W	3,494	T (Hourly)	April 2014 to October 2015
PI-off	Piramide off glacier	33.61°S, 69,91°W	3,020	T (Hourly)	May 2014 to April 2015

TABLE 2 Automatic weather stations (AWSs) and respective years used for the calculation of the air temperature lapse rates

Month	AWSs and year	Valid days per year
January	Yeso Embalse (2014–2015)-Piramide off (2015)-Yeso off (2014)	31 days each year
February	Yeso Embalse (2014–2015)-Piramide off (2015)-Yeso off (2014)	28 days each
March	Yeso Embalse (2014)-Yeso off (2014)	13 days
April	Yeso Embalse (2014)-Piramide on (2014)	12 days
May	Yeso Embalse (2014)-Piramide on (2014)	31 days
June	Yeso Embalse (2014)-Piramide on (2014)	30 days
July	Yeso Embalse (2014–2015)-Piramide on (2014–2015)	31 days each year
August	Yeso Embalse (2014–2015)-Piramide on (2014–2015)	31 days each year
September	Yeso Embalse (2014–2015)-Piramide on (2014–2015)	30 days each year
October	Yeso Embalse (2014–2015)-Piramide on (2014–2015)	31 days
November	Yeso Embalse (2013–2014)-Piramide off (2014)-Yeso off (2013)	25 days for 2014 and 15 days 2013
December	Yeso Embalse (2013)-Yeso off (2013)	31 days

TOPKAPI-ETH offers a compromise between a detailed representation of high-mountain hydrological processes and computational efficiency. The model incorporates physically based parameterizations of most relevant hydrological processes in high-mountain catchments, such as snow and ice melt (Pellicciotti et al., 2005), ice melt under debris (Carenzo, Pellicciotti, Mabillard, Reid, & Brock, 2016), glacier area and elevation changes (Huss, Jouvet, Farinotti, & Bauder, 2010), snow albedo decay (Brock, Willis, & Sharp, 2000), gravitational redistribution of snow (Bernhardt & Schulz, 2010), and englacial storage and release of snow and ice meltwater (Hock & Noetzli, 1997).

In this study, we used the same TOPKAPI-ETH model setup as described in Ayala et al. (2016). Ayala et al. (2016) extensively calibrated and validated the model for the Rio del Yeso catchment using 2 years (2013–2015) of field data that included manual snow depth measurements, ablation stakes, meteorological data from four AWSs, albedo time series from radiation measurements at Bello and Yeso AWSs, distributed fields of daily albedo derived from optical photos, and streamflow measurements at the outlet of Bello and Yeso glaciers. To avoid error compensation and parameter ambiguity, the model was calibrated in a stepwise approach, in which each main parameter set was calibrated individually against specific field observations (see Figure 2 in Ayala et al., 2016). Here we perform an additional calibration step to account for the uncertainty in precipitation over the longer period of record of this study (see Section 3.5 below).

Using the calibrated model setup, we then simulate glacier mass balance, elevation, and areal changes for the period 2000 to 2015. Glacier volume and geometry changes were simulated using the Δh-parametrisation developed by Huss et al. (2010). The Δh-parametrisation is an empirical method that quantifies ice thickness changes as a function of previously observed elevation changes and was developed based on a large datasets of glaciers in the Swiss Alps (Huss et al., 2010). Given the lack of repeated DEMs or ice volume observations for our region, we have used the set of parameters originally calibrated by Huss et al. (2010) for glaciers with an area smaller than 5 km^2. The model was run for 10 years in a spin up mode to produce initial conditions of albedo and snow height and simulations started in 2000 with these initial conditions.

3.2 | Extrapolation of meteorological variables

To drive the glacio-hydrological model, both air temperature and precipitation measurements are required at hourly resolution. While there was relatively good spatial coverage of AWSs in the study catchment in 2013–2015, only a few stations were available over the complete study period (Table 1), and so data were extrapolated in both space and time. Hourly temperature time series were calculated from daily minimum and maximum values recorded at the YE AWS using the method suggested by Wilkerson, Jones, Boote, Ingram, and Mishoe (1983) and adapted by Reicosky, Winkelman, Baker, and Baker (1989) (subroutine WCALC). The approach uses a sinusoidal function to interpolate between extreme values by dividing the day into three time periods: midnight to sunrise plus 2 hr, daylight hours, and sunset to midnight. It assumes that the minimum value occurs 2 hr after sunrise and the maximum halfway between sunrise and sunset time, obtained from the matlab subroutine Sunset, based on Montenbruck and Pfleger (2000).

To evaluate the results of this approach in the study catchment, the method was tested at Laguna Negra AWS (LN, 2,780 m asl; Figure 1b) using the data available for 2013–2015. This site was selected because daily minimum and maximum data as well as independently recorded hourly data were available, and the station was at a similar elevation to the YE AWS (Table 1). The evaluation resulted in a Nash-Sutcliffe model efficiency criterion of 0.93.

The interpolated hourly temperature data were distributed from the YE AWS (Figure 1b) to the rest of the catchment using monthly means of hourly lapse rates, calculated from the available meteorological data in the catchment for each month and year with concurrent data (Table 2). In order to represent the effect of the glacier boundary layer (Brock et al., 2010; Greuell & Böhm, 1998), we use a parameter to decrease air temperature over glacier surfaces (Tmod) by subtracting 1°C for debris-free areas and a $\mathrm{Tmod}_{\mathrm{debris}}$ of 0.3°C to increase temperature for debris cover grid cells, calibrated for glacier-covered areas by Ayala et al. (2016). Figures 2 and 3 show the disaggregated and extrapolated temperatures at the Bello (Figure 2) and Yeso (Figure 3) AWSs. In general, calculated temperatures correspond well to measured values, especially during the daytime in the summer months; however, the disaggregation and extrapolation method performs less well on glacier surfaces before sunrise when there snow was present, and during winter time when air temperatures were below 0° C.

Precipitation is not evenly distributed across the study catchment, and according to previous studies in the region, a logarithmic model can be used to represent precipitation spatial variability at high elevations (Ragettli, Pellicciotti, et al., 2014; Vicuña, Garreaud, & McPhee, 2010). Thus, we extrapolated the hourly precipitation measurements from YE AWS for the period 2000 to 2015 using a logarithmic model as follows (Ragettli, Pellicciotti, et al., 2014):

$$P(z) = P(\text{Yeso} \quad \text{Embalse}) \cdot (0.3866 \log(z) - 2.014) \tag{1}$$

where P (YE) is precipitation measured at YE and $P(z)$ the precipitation at the elevation z. The values of the coefficients in Equation (1) were estimated from the logarithmic fit of the annual mean precipitation at 15 stations in the Maipo catchment for the period 2000–2015. Using a number of stations from the larger Maipo catchment ensures that the large-scale, synoptic variability of the precipitation pattern is reproduced. To represent local effects on precipitation and its accumulation over glaciers, we used a local scaling factor that modifies precipitation over each glacier (Huss, Farinotti, Bauder, and Funk (2008); Magnusson, Farinotti, Jonas, & Bavay, 2011; Farinotti, Usselmann, Huss, Bauder, & Funk, 2012). This factor was calibrated to match the simulated long-term glacier elevation changes with those derived from the geodetic mass balance to account for local processes governing snow accumulation on glaciers. Use of a local scaling factor for each glacier was supported by evidence from field observations of preferential deposition, scouring, and snow removal by wind that cannot be captured by a regional precipitation gradient. The local factor was calibrated against the elevation change of the period 2000–2013 obtained from the DEM differencing. The estimated factors were 0.74 for Bello and Yeso and 1.88 for Piramide. A similar approach has been used by Magnusson et al. (2011) and Farinotti et al. (2012),

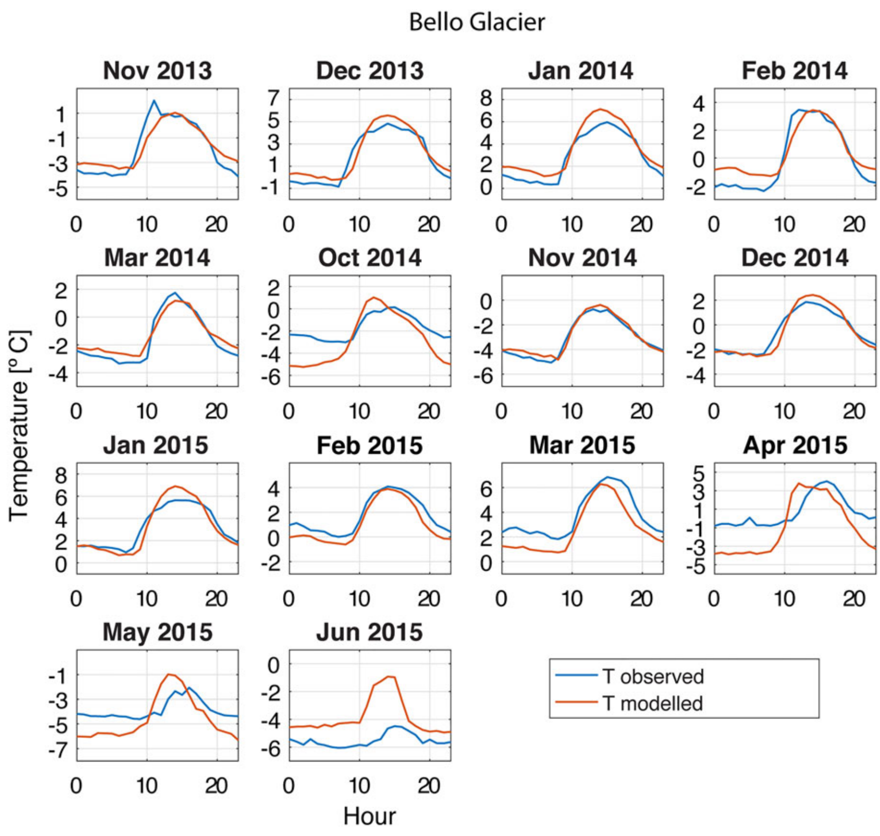

FIGURE 2 Validation of the air temperature extrapolation method on Bello Glacier (Section 3.2): comparison of observed (blue) and modelled (red) average hourly temperature at Bello Glacier AWS for each month, November 2013–June 2015

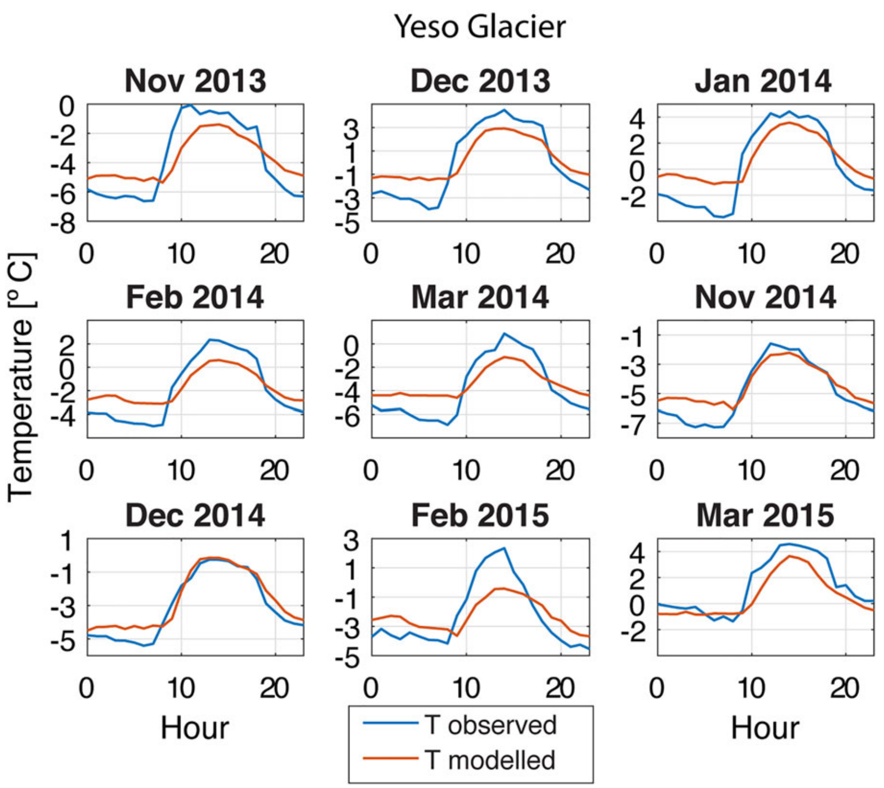

FIGURE 3 Validation of the air temperature extrapolation method on Yeso Glacier: comparison of observed (blue) and modelled (red) average hourly temperature at Yeso Glacier automatic weather station (AWS) for selected months between November 2013 and March 2015. Only months with data available at the Yeso Glacier AWS are shown

and in light of the lack of local observations, it is a plausible way of preventing precipitation uncertainty from dominating the modelling exercise. Finally, precipitation was disaggregated to hourly values by distributing daily values homogeneously during the day.

3.3 | Debris thickness estimation

We used the debris thickness map derived by Ayala et al. (2016) for the debris-covered areas on the Piramide, Bello, and Yeso glaciers (Figure 1). The map was derived by solving the distributed energy balance of the debris-covered areas at the moment of acquisition of a Landsat 8 thermal image of the area at 90-m spatial resolution. This method was originally presented by Foster, Brock, Cutler, and Diotri (2012) as a physically oriented alternative to empirical relationships between surface temperature and debris thickness (e.g. Mihalcea et al., 2008). Different versions of the method have been subsequently presented (e.g. Rounce & McKinney, 2014), but several uncertainties are still associated with this approach, such as the debris temperature profiles, heat storage rate, and turbulent heat fluxes (Ayala et al., 2016; Schauwecker et al., 2015). Technical details regarding the development of the debris thickness map can be found in Appendix 1 in Ayala et al. (2016).

3.4 | Geodetic elevation change

In recent years, the geodetic method has been widely used to obtain glacier changes over short- or long-term periods (e.g. Bolch, Pieczonka, Mukherjee, & Shea, 2017; Rankl & Braun, 2016). The method, based on the differencing of DEMs, can provide glacier changes over several years for large remote areas.

We followed the TDX processing scheme in Malz et al. (2018) using the GAMMA software. We acquired coregistered single look complex images in HH polarization provided by DLR. All the TanDEM-X scenes along one track were concatenated and the strips processed by differential interferometry (DInSAR; e.g. Malz et al., 2018). We used the SRTM-C DEM as a reference and subtracted it from the bistatic TanDEM-X interferogram. For this, the topographic phase was simulated from the SRTM-C DEM using the TanDEM-X orbit parameters. We unwrapped the interferograms using a Minimum Cost Flow algorithm. In order to remove the phase noise from the differential interferogram we applied a Goldstein filter, and areas with low coherence (coherence < 0.2) were masked out (Goldstein & Werner, 1998; Rankl & Braun, 2016; Vijay & Braun, 2016). The unwrapped differential phase was converted into absolute differential heights. These differential heights were added back to the topographic heights from SRTM-C DEM to generate a TanDEM-X DEM. The resulting TanDEM-X DEMs were geocoded with the SRTM-C DEM to maintain planimetric consistency (e.g. Malz et al., 2018; Vijay & Braun, 2016).

The postprocessing comprises the mosaicking of all raw DEMs resulting from the InSAR strip processing. We used a stable ground mask derived from optical data (Landsat OLI 2013) and corrected vertical biases between the strips applying a polynomial fitting (Malz et al., 2018). TanDEM-X DEMs were iteratively coregistered (vertically and laterally) to the SRTM-C DEM using the approach of Nuth and Kääb (2011).

Uncertainties in the geodetic elevation changes for the period 2000–2013 were estimated by calculating the median absolute deviation for the elevation differences on stable areas. Since the deviation is known to be slope-dependent (Gardelle, Berthier, & Arnaud, 2012), the area of interest was divided in 5° slope-bins and the total median absolute deviation was calculated by weighting the area of each bin (e.g. Malz et al., 2018). We discarded any significant bias associated to the radar signal penetration in snow and ice of SRTM-C and TanDEM-X, as previous studies have shown that summer DEM acquisitions in the Southern Hemisphere (during melting conditions) reveal negligible penetration (Dussaillant, Berthier, & Brun, 2018; Falaschi et al., 2017; Jaber, Floricioiu, & Rott, 2016; Jaber, Floricioiu, Rott, & Eineder, 2013). As the LiDAR surveys included only a small portion of stable areas, we used the precision achieved by the LiDAR and Global Positioning System control points (±0.30 m) as an indication of the error for the period 2013–2015 (Dirección General de Aguas, 2015). Finally, the uncertainties were estimated using the standard principles of error propagation.

3.5 | Calibration and validation

A detailed scheme of the calibration and validation procedure is shown in Figure 4. The geodetic elevation difference for the 2000–2013 period was used to calibrate the precipitation correction factors over glacier surfaces (as described in Section 3.2), while the 2013–2015 dataset (obtained from LiDAR) was used to validate simulated ice elevation differences (Sections 2.3 and 3.4). Model simulations were also validated using albedo and snow-covered area (SCA) obtained from processing terrestrial photos (Section 2.4), as well as albedo observations at AWSs in Bello and Yeso glaciers for the season 2014–2015 and SCA from Moderate Resolution Imaging Spectroradiometer (MODIS). We compared the SCA derived from a daily MODIS MOD10A1 product (Hall, Riggs, Salomonson, Digirolamo, & Bayr, 2002) to the catchment-wide snow cover area derived from the TOPKAPI-ETH on a daily time step. The daily MODIS data were discarded if more than 10% of the total area was covered by cloud, and any remaining cloud covered cells were filtered using a linear interpolation of SCA quantity based upon a temporal search window of 2 days either side of the cloud cover observation at the given cell. The MODIS grids were resampled to a 30-m grid and clipped to the same area as the model domain (see blue line in Figure 1c), and for each day, the catchment average SCA was extracted and compared with that of the TOPKAPI-ETH model simulations. The validation procedure of albedo and SCA using the terrestrial camera is detailed in Ayala et al. (2016) and not repeated here.

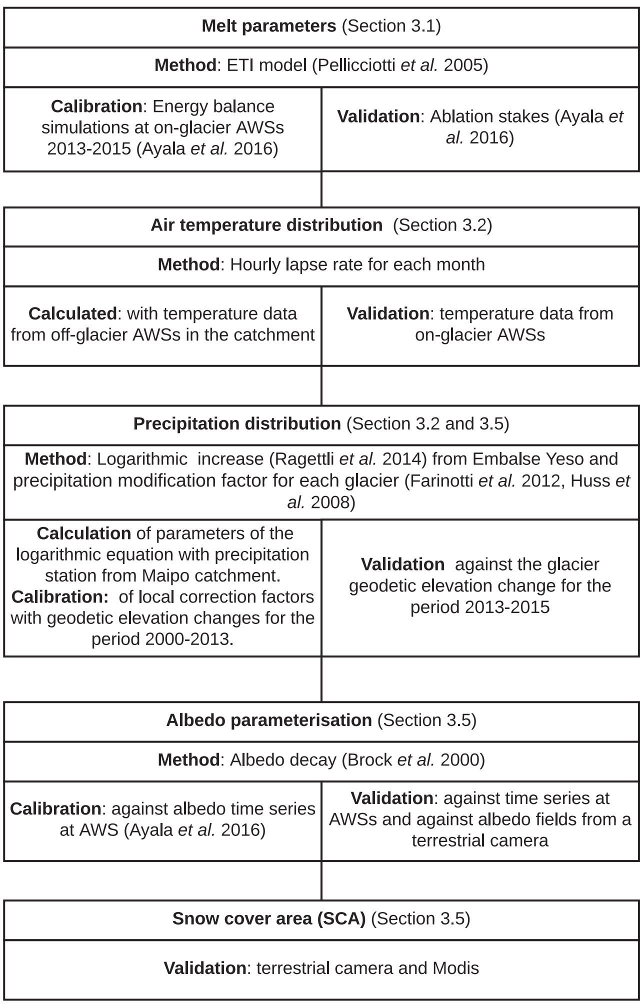

FIGURE 4 Scheme of the calibration and validation approach used in this study together with the datasets used in each step

4 | RESULTS

4.1 | Model validation results

The ground-based and satellite validations of the TOPKAPI-ETH model are given in Figures 5 and 6, respectively. The model captures the variability in albedo measured at Bello and Yeso AWSs and both the albedo and SCA derived from the terrestrial camera (Figure 5). Specifically, modelled albedo at Bello Glacier AWS follows the measured decay rates and albedo increase after spring storm events (Figure 5b), but it overestimates albedo at the Yeso Glacier AWS (Figure 5c), despite capturing the magnitude of the snow albedo decay at the start of 2014. The average albedo and SCA calculated from the camera photos (Figure 5d,e) are replicated over Bello Glacier with a slight tendency towards overestimation in the spring months (October–November). In summer (January 2015), the albedo simulations perform particularly well (RMSE = 0.19).

At the catchment scale, TOPKAPI-ETH simulates the timing of snow cover disappearance of the daily MODIS SCA product (Figure 6). In the period 2001 to 2009, the appearance and disappearance of snow are well reproduced, with a RMSE of ~11–15% for individual years. After 2010, the model performance declines, and the timing of the snow cover disappearance in the spring and summer is not well reproduced (RMSE > 20%). The total summer SCA minimums, however, are in line with the MODIS results, which suggest that in general the model can capture the seasonal variability of snow cover.

4.2 | Glacier elevation changes

Results from the geodetic elevation change show generally positive or stable values for the 13 years from 2000 to 2013 (Figures 7a,b) and then a generally negative mass balance in the period 2013–2015 (Figures 7c,d). The change in the second period is noteworthy, with an ice thinning rate of −1.15 ± 0.15 m/year (−2.31 ± 0.30 m) for Bello Glacier, in contrast with a mean rate of −0.01 ± 0.09 m/year (−0.15 ± 1.23 m) for the period 2000–2013 (Table 3). Yeso Glacier

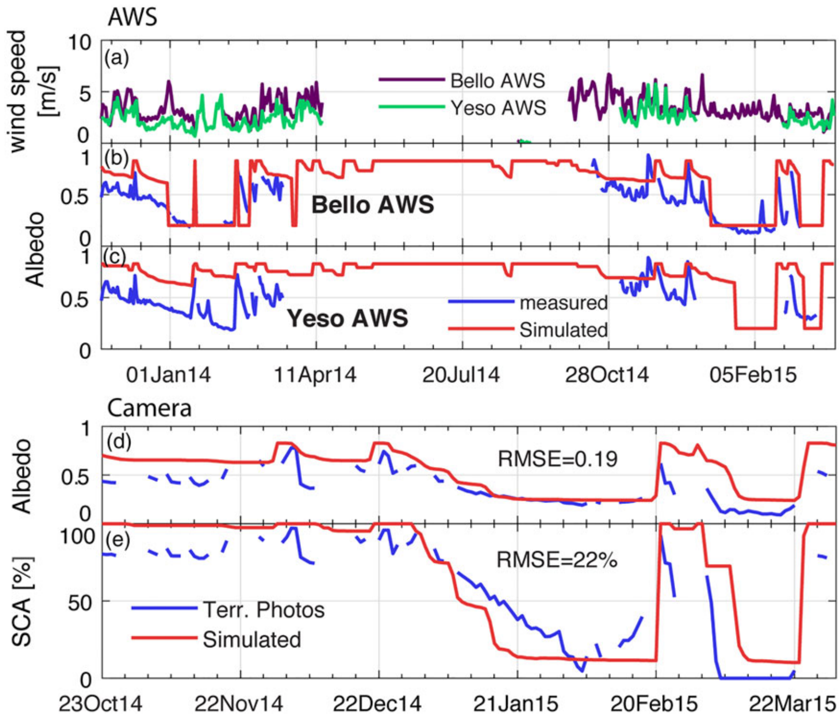

FIGURE 5 (a) Wind speed measured at Bello and Yeso automatic weather stations (AWS) during the period January 2014 to February 2015. (b and c) Comparison of observed and modelled daily albedo at the point-scale of the AWS on Bello and Yeso glaciers during the period January 2014 to February 2015. (d and e) Comparison of observed (from a terrestrial camera) and simulated distributed albedo and snow covered area on a portion of Bello Glacier (see Figure 1 and Section 3.5) covering an area of 0.84 km^2, during the period October 2014 to March 2015. Note the different *x*-axis scale of the lower two panels

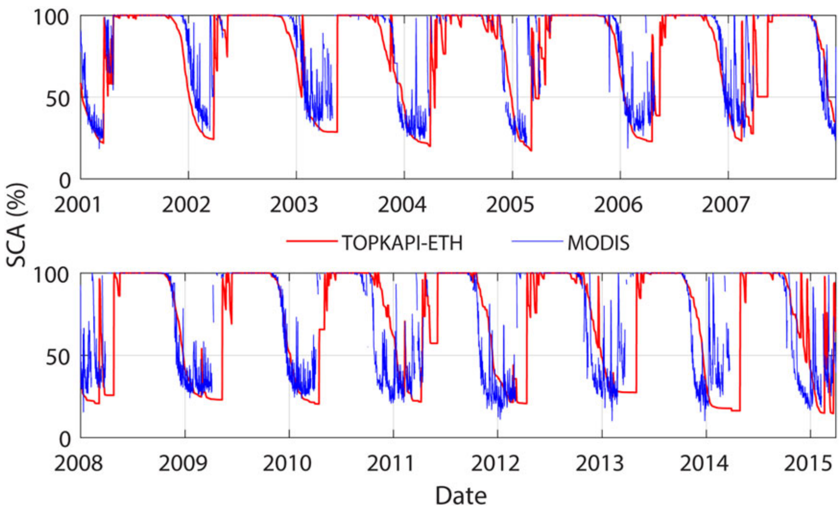

FIGURE 6 Comparison of the modelled TOPKAPI-ETH snow cover area (SCA) for the period 2001–2015 and Moderate Resolution Imaging Spectroradiometer (MODIS) MOD10A1 SCA. The MODIS SCA is provided as a catchment-wide average after filtering of clouds (see text)

has similar thinning rates of −0.03 ± 0.09 m/year (−0.43 ± 1.23 m) for the period 2000–2013 and −1.08 ± 0.15 m/year (−2.17 ± 0.30 m) for the subsequent period.

For 2013–2015, the highest mass losses on Piramide Glacier are observed in the upper section (Figure 7d), where debris is thin (Figure 1a). Comparatively, there is a smaller loss on the lower tongue, and the lowest elevation differences are observed in the central section of the glacier (Figure 7d). Areas of thick debris interspersed with ice cliffs are characterized by heterogeneous surface differences for both periods (Figure 7b,d). For the latter period (2013–2015), the reduction of snow accumulation for the upper Piramide Glacier resulted in greater surface lowering (Figure 7d) where debris is thin and snow is normally supplied by high avalanche loads.

TOPKAPI-ETH simulations show an initial positive glacier elevation change followed by a neutral or slightly negative change for the two debris-free glaciers (Bello and Yeso) for the period 2000–2013, while estimates for the debris-covered Piramide glacier indicate an initial ice thickness increase followed by a decline in 2009 until reaching cumulative negative values of approximately −3.4 m in 2015 (Figure 8). Bello and Yeso lag behind Piramide, showing that strong thickness decreases are the final 2 years and a lower cumulative decrease by 2015 than at Piramide (Figure 8).

There is considerable spatial variability in mass balance, with distinct patterns for each glacier (Figure 9a–c). The surface height changes of Bello and Yeso become more positive with increasing elevation and are particularly negative for thin debris areas on the tongue of Yeso Glacier (between 3,900 and 4,100 m asl). The simulated profile of surface height change on Piramide Glacier peaks in the midglacier area for the period 2000–2013, with a slight increase with elevation for both periods between 3,200 and 3,700 m asl (corresponding to areas of thick debris), and then positive mass balance for the period 2000–2013 between 3,800 and 4,200 m asl, in

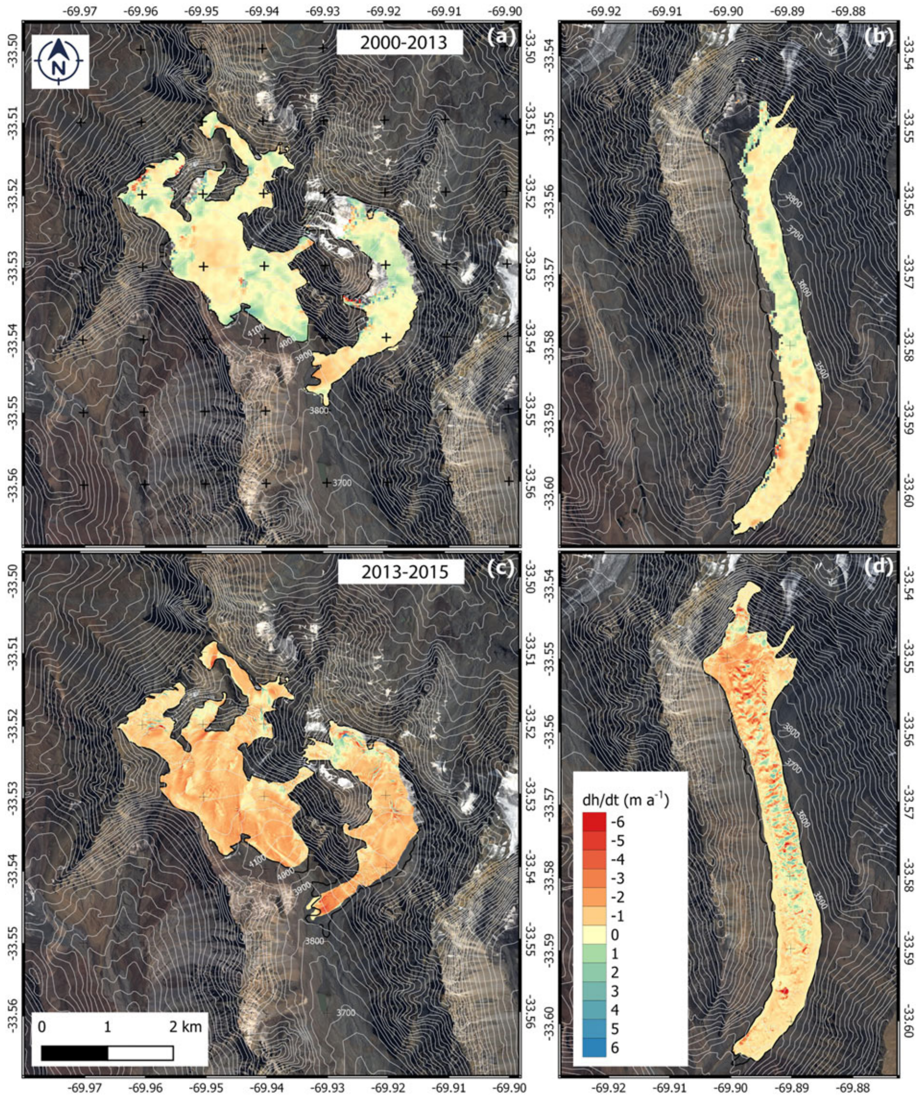

FIGURE 7 Geodetic elevation changes (Section 3.4) for the periods 2000–2013 [from Shuttle Radar Topography Mission and TanDEM-X digital elevation models]) (a and b) and 2013–2015 (c and d) (from the light detection and ranging digital elevation models). For both periods, Bello and Yeso glaciers are shown on the left panels (a and c) and Piramide Glacier on the right (b and d)

TABLE 3 Geodetic surface elevation change rate in the periods 2000–2013 and 2013–2015 for the three study glaciers

Glacier	Δ*h* 2000–2013	Δ*h* 2013–2015
Bello	−0.01 ± 0.09 (m/year) −0.15 ± 1.23 (m)	−1.15 ± 0.15 (m/year) −2.31 ± 0.30 (m)
Yeso	−0.03 ± 0.09 (m/year) −0.43 ± 1.23 (m)	−1.08 ± 0.15 (m/year) −2.17 ± 0.30 (m)
Piramide	−0.14 ± 0.09 (m/year) −1.88 ± 1.23 (m)	−0.75 ± 0.15 (m/year) −1.50 ± 0.30 (m)

correspondence with the areas of highest avalanche mass mobilization (results not shown). The mass balance is also negative between 4,200 m asl and the upper reaches of the glacier at 4,600 m asl, which correspond to areas of thin debris cover (Figure 1a), where an elevation gradient is reestablished (Figure 9c).

The geodetic elevation changes show a similar pattern as simulations over Yeso and Bello for the final 3 years of the study period (Figure 9a,b). However, geodetic values were more negative than the 2000–2013 simulations between 4,200 and 4,800 m asl (Figure 9a,b).

The differences between the debris-free and debris-covered glaciers are also evident in the spatial distribution of the geodetic mass balance, with a profile that is quite smooth for Bello (and slightly less so for Yeso), while the spatial variability at Piramide is very high, a feature also well captured by the model. According to the geodetic mass balance between 2000 and 2013, simulations in Bello and Yeso overestimate melt below ~4,300 m asl and overestimate accumulation above that elevation, while at Piramide the geodetic mass balance pattern for that period is similar to the simulations, with a small melt overestimation between 3,500 and 3,800 m asl

4.3 | Runoff and runoff components

Total runoff and the annual contribution to runoff from snowmelt, icemelt, and precipitation are shown in Figure 10. A clear overall decline in runoff can be observed, which is marked for the period after

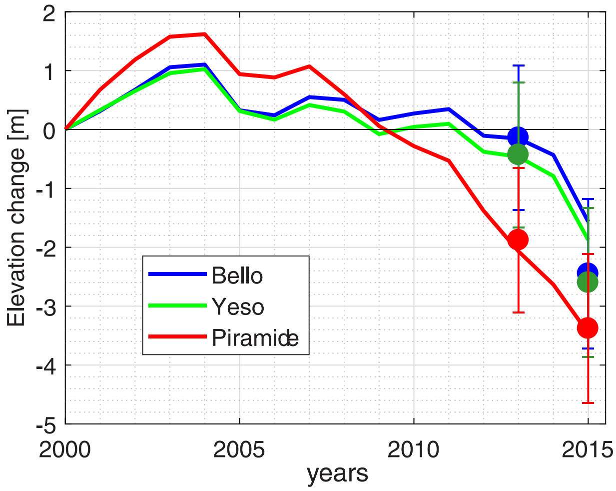

FIGURE 8 Cumulative elevation changes for the three study glaciers as simulated by TOPKAPI-ETH for the period 2000–2015 (colour lines). The dots in 2013 indicate the geodetic elevation change between 2000 and 2013 derived from differencing the SRTM and TanDEM-X DEMs and used for calibration of the precipitation correction factors. The dots in 2015 show the geodetic elevation change for the period 2000–2015 (obtained from the sum of the elevation difference for the first period 2000 to 2013 and that of the second period 2013–2015 from differencing of the light detection and ranging digital elevation models)

2009. The main runoff contribution comes from snowmelt, accounting for 66 to 93% of total runoff per year. Icemelt contribution fluctuates between 3.5 and 32% and is highest in dry years with low total runoff (such as 2014–2015, which have the highest proportional contribution from icemelt). The liquid precipitation contribution is consistently small and does not exceed 6% in any given year. While annually snowmelt represents the main water input to the system, in summer (January and February) icemelt becomes increasingly dominant, contributing equally with snowmelt by March (Figure 11). The contribution of icemelt to total runoff from each of the three glaciers is distinct (Figure 12). Bello and Yeso have the smallest average icemelt runoff with 5.4 and $5.7 \cdot 10^{-5}$ m/hr, respectively, while the highest icemelt contribution is from Piramide, with $8.3 \cdot 10^{-5}$ m/hr. It should be noted that Bello Glacier displays larger interannual variations in streamflow contributions from icemelt than Piramide. Snowmelt contribution to total runoff from Bello and Yeso are 4.6 and $4.3 \cdot 10^{-5}$ m/hr, respectively, and Piramide has the highest snowmelt contribution of $1.59 \cdot 10^{-4}$ m/hr, being the glacier with more snowmelt variability. Since 2011, snowmelt contribution from all three glaciers has reduced compared with the previous years (orange and red lines on the right hand panels in Figure 12). The seasonality is also distinct from 2013 onwards, with snowmelt and icemelt both occurring earlier at Piramide (Figure 12e,f) than on the other two glaciers (Figure 12a–d).

5 | DISCUSSION

5.1 | Mass balance and runoff contribution

Geodetic mass balance estimates show a very distinct behaviour in the two analysed periods (2000–2013 and 2013–2015). While an almost neutral elevation change was obtained for the first period (−0.01 to

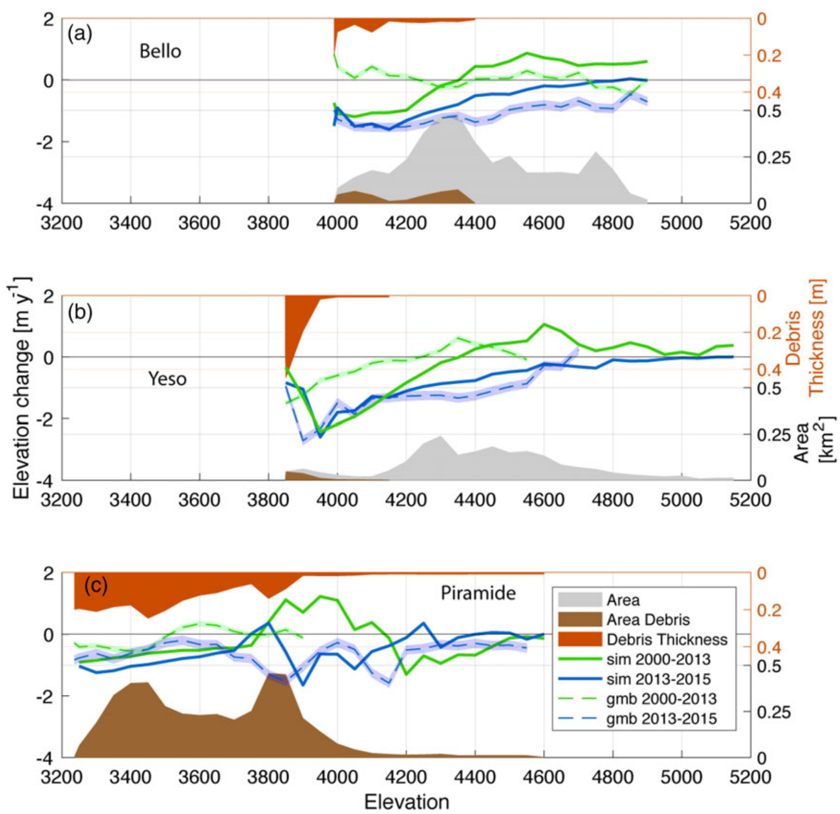

FIGURE 9 Simulated average elevation changes for 100-m elevation bands, and debris free (light grey) and debris-covered (brown) areas on (a) Bello, (b) Yeso, and (c) Piramide glaciers. Average debris thickness for each elevation band is shown on the upper right axis. Simulated elevation changes in the periods 2000–2013 and 2013–2015 are shown by the green and blue continuous lines, respectively. Geodetic elevation differences are shown by the green and blue segmented lines in the period 2000–2013 and 2013–1015, respectively. Incomplete lines for the geodetic elevation change profiles are due to missing pixels in the corresponding elevation band

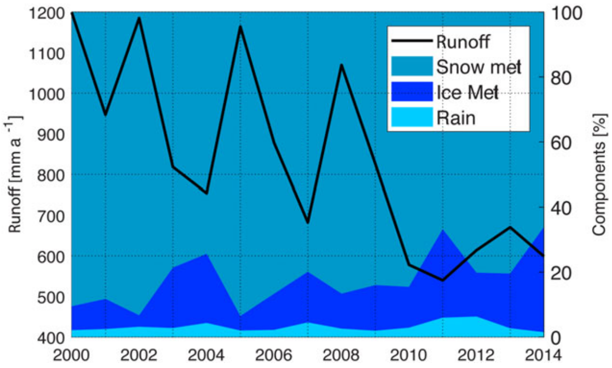

FIGURE 10 Annual average catchment runoff (left axis) and relative contribution (right axis) from snowmelt, ice melt, and rain

−0.14 m/year), a high ice thinning rate is evident for the second period (−0.75 to −1.16 m/year). An important asset of model simulations is that they allow the identification of different mass balance trends between the acquisition times of the DEMs used in the geodetic change detection. In this case, results from TOPKAPI-ETH suggest that the neutral mass balance of the period 2000–2013 is a result of a moderate positive trend in 2000–2009 and a strong negative trend in 2010–2013, which continued in 2013–2015. These results agree with the mass balance measurements on Echaurren Norte Glacier, the only long-term glacier mass balance monitoring program in the Andes of central Chile, which shows similar patterns from the year 2000 (Masiokas et al., 2016), having 4 years of positive mass balance between 2000 and 2009 and no positive mass balance years since then (Masiokas et al., 2016; WGMS, 2017). The strong negative trend in glacier mass balance observed from 2010 is clearly related to the severe drought observed in the Chilean central regions during recent years (Garreaud et al., 2017), termed the *Mega-drought* (Boisier, Rondanelli, Garreaud, & Muñoz, 2016), which has been characterized by historically low precipitation levels, shallow seasonal snowpacks (Cornwell, Molotch, & McPhee, 2016; Cortés & Margulis, 2017), and high temperatures (Garreaud et al., 2017), especially in spring and autumn (Burger et al., 2018).

There is a good agreement between the simulated and geodetic mass balance at Piramide Glacier (Figure 8). The model does not fully reproduce the values from the geodetic mass balance at Bello and Yeso glaciers in 2015, though they are within the range of uncertainties of the geodetic measurements. Simulations do show a decreasing trend starting in 2010, but the geodetic mass loss in the period 2010–2015 is larger than that from the model. Despite these differences, a more detailed analysis of the patterns of glacier surface change shows that the model is able to reproduce some of the elevation-dependent differences evident in the geodetic elevation changes (Figure 9). Importantly, the model is able to resolve much of the spatial variability of glacier surface change in relation to the differences in debris thicknesses, as well as those related to the avalanches that feed the debris-covered Piramide glacier.

The glaciers analysed in this study provide an excellent example of the different processes that affect the long-term evolution of debris-

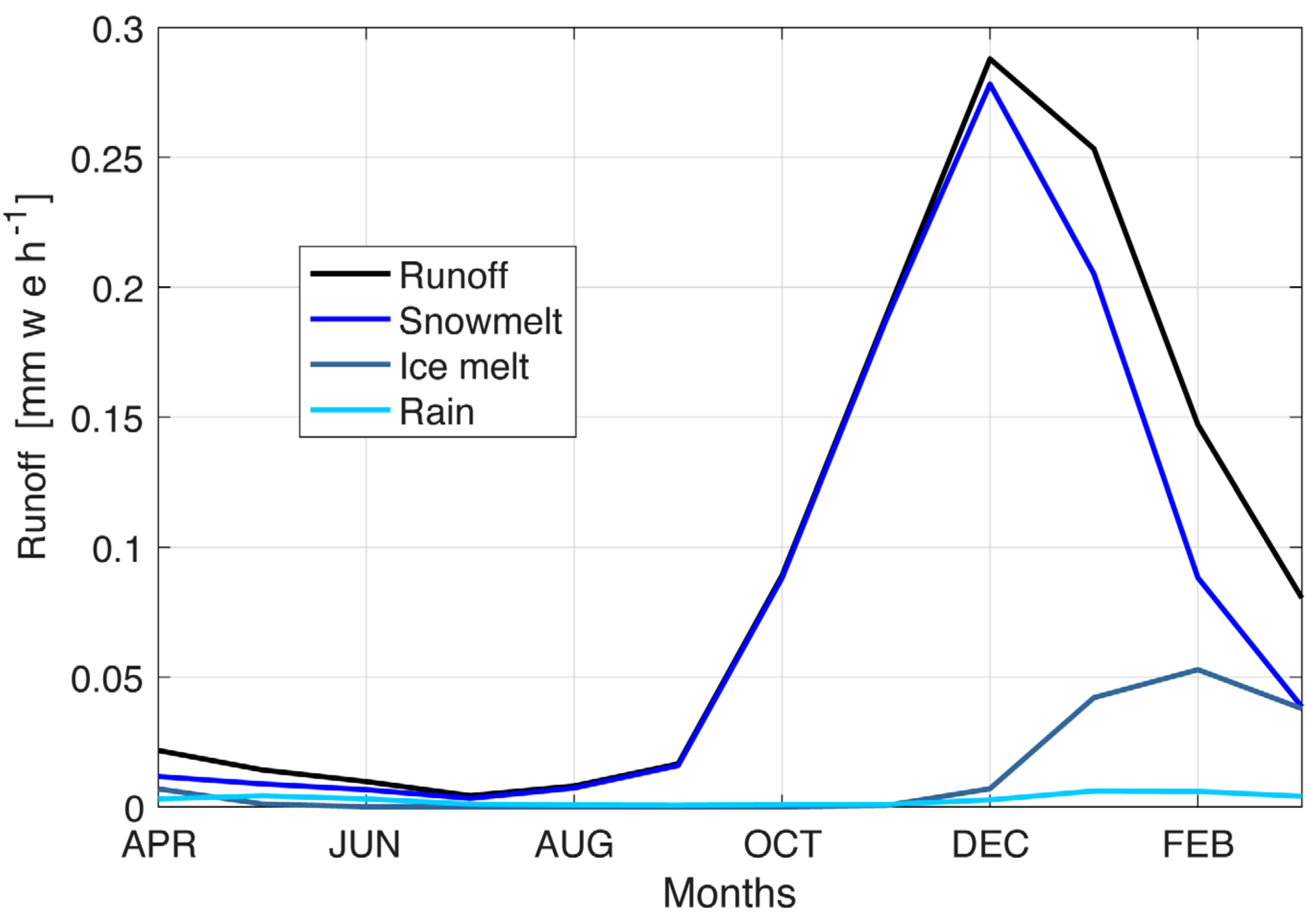

FIGURE 11 Monthly averages of simulated total runoff and runoff components (snowmelt, icemelt, and rain) over the study period (2000–2015)

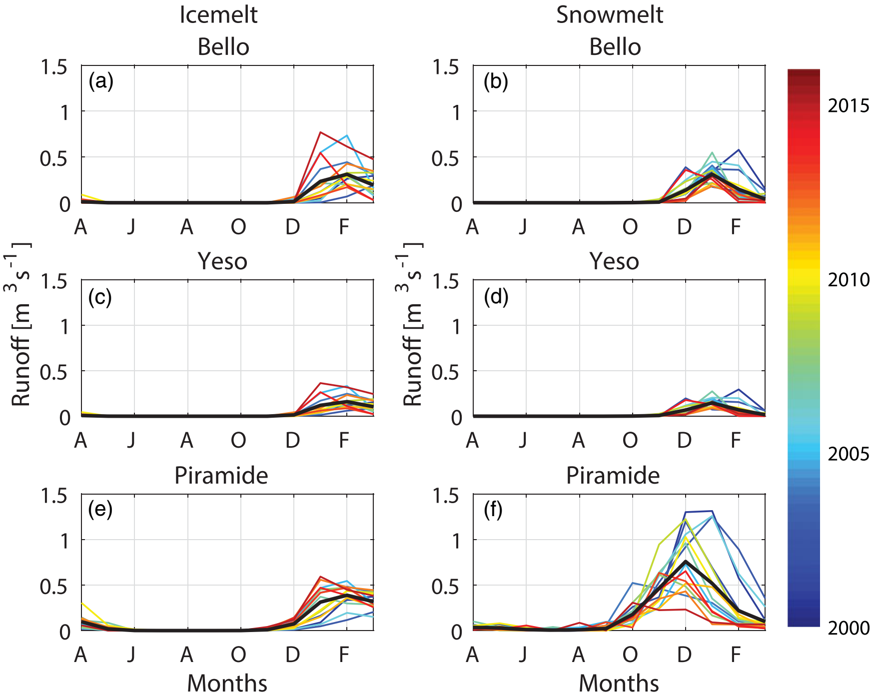

FIGURE 12 Monthly mean runoff generated by ice melt (left panels) and snowmelt (right panels) from (a and b) Bello, (c and d) Yeso, and (e and f) Piramide glaciers. Each thin colour line represents an individual year (indicated by the colour bar). The average for the entire period (2000–2015) is shown as a bold black line

free and debris-covered glaciers within one catchment. While the mass balance of Bello and Yeso is more strongly controlled by temperature gradients affecting the precipitation phase and ablation components of the model (Ayala et al., 2016) and snow removal on the midglacier (evidence from unpublished data on Bello), the mass balance of Piramide Glacier is controlled principally by debris thickness and snow accumulation from avalanches (Figure 9). A sensitivity analysis revealed that ignoring the contribution of avalanching in the model simulations produces a total cumulative elevation change more than 3 times more negative for Piramide compared with our reference model (Figure 13). Ignoring this process has a greater impact on the total absolute glacier elevation change than artificially providing 10 centimetres more surface debris for the glacier (red triangles in Figure 13). The strong control that debris thickness exerts over the glacier mass balance has been extensively demonstrated in the literature (e.g. Ragettli et al., 2016; Vincent et al., 2016) and is shown by the results from our study to be important for long term modelling of debris-covered glaciers mass balance.

The removal of snow from the central sections of Bello Glacier would contradict a hypothesis that glacier areas act as net sinks of snow during the accumulation season (Dadic, Mott, Lehning, & Burlando, 2010; Gascoin, Lhermitte, Kinnard, Bortels, & Liston, 2013). However, the mass losses that seem to occur on Bello and Yeso glaciers might be compensated by the contribution of avalanches from the surrounding upper slopes, which is still shown to be an important process for these glaciers (Figure 13). Evidence of negative surface changes in the central sections of Bello (Figure 7a) reveals a potential local process of wind effects on snow, which agrees with field observations, though modelling snow redistribution on high-elevation glaciers remains an important area for future studies, even if outside of the scope of this paper.

Runoff generation from debris-free and debris-covered glaciers also exhibit a distinct behaviour. Particularly, interesting is the different interannual spread in the runoff generated by icemelt from Bello and Piramide glaciers (Figure 12). Icemelt from Bello Glacier shows a large interannual variability, which most likely indicates a sensitivity to climatic variability of precipitation and temperature. In turn, icemelt from Piramide Glacier is largely insulated from climatic changes due to its supraglacial debris. Interestingly, the opposite pattern is evident for snowmelt, that is, large interannual variability on Piramide Glacier in comparison with Bello Glacier (Figure 12f). In this case, the predominant low elevation of Piramide Glacier makes the

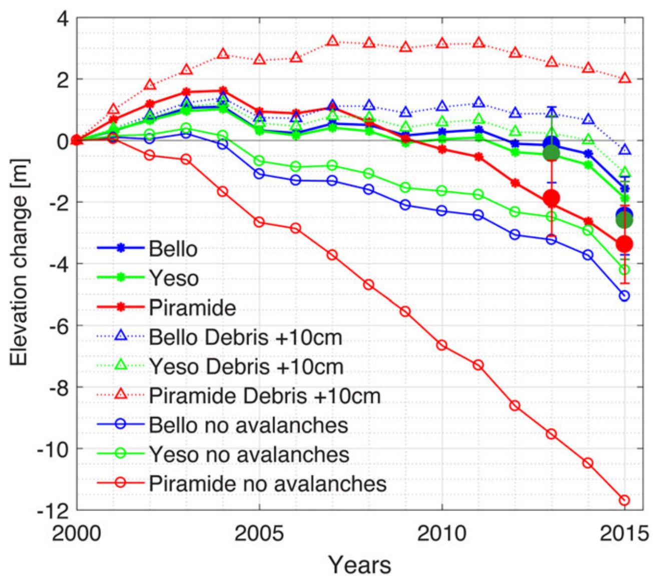

FIGURE 13 Comparison of modelled and observed (geodetic) elevation changes for additional model runs from a sensitivity analyses for the three glaciers: (a) considering an additional 10 cm of supraglacial debris (triangles) and (b) ignoring the avalanching in the TOPKAPI simulations (hollow circles). The reference model (filled markers) and geodetic surface changes (large filled circles and error bars) are shown for comparison

snowpack sensitive to air temperature and the proportional amounts of solid winter precipitation. Further still, the mass loss characteristics of a debris-covered glacier such Piramide Glacier is strongly governed by the presence of ice cliffs (Buri, Pellicciotti, Steiner, Miles, & Immerzeel, 2016; Steiner et al., 2015), which are currently not considered in TOPKAPI-ETH. Future modelling studies of debris-covered glaciers may therefore benefit from physically based or parameterised representations of these processes.

As demonstrated by previous studies for individual years, snowmelt is the largest contributor to runoff in Andean catchments of central Chile with an outlet point around 3,000 m asl (Ohlanders et al., 2013; Ragettli & Pellicciotti, 2012). In this study, we showed that this result holds for a long time period, although with important interannual variability (Figure 10). Particularly, important is the evidence that icemelt contribution becomes more relevant during drought periods. We observed an increase in the relative contribution to runoff of ice melt from 11.6% in the period 2000–2009 to 20.5% in the period 2010–2015. Quantifying the shift in relative streamflow contributions for a drier climate is highly relevant to the socio-economy of central Chile, and our study provides a new insight into the longer term evolution of these contributions, and most significantly a shift to increasing dependency on the declining resource of glacier ice.

5.2 | Sources of uncertainty

As suggested by Ayala et al. (2016), the spatial distribution of forcing variables and the debris thickness are relevant controls on the glacier mass balance in this catchment. Particularly, challenging is the modelling of snow accumulation on debris-free glaciers and wind-exposed locations. As our simulations show, the cumulative elevation change of Bello and Yeso glaciers was consistently overestimated when using regional precipitation gradients that provide a good representation of annual water balances in the region. More precise regional estimates of precipitation and the simulation or parameterization of snow transport would likely improve the simulation of glacier mass balance. On the other hand, more accurate estimates of the current debris thickness distribution and its time evolution will become necessary for future studies on long-term glacier mass balances (Rowan et al., 2015). However, changes in debris thickness over a period of 16 years are likely minor or restricted to specific areas.

6 | CONCLUSION

In this paper, we have used a combination of distributed glacio-hydrological modelling, new estimates of geodetic mass balance for the region, and a large amount of field data over two seasons to investigate the mass balance and runoff contribution of the glaciers of the Rio del Yeso catchment. Our main conclusions are the following:

- The glaciers of the study catchment experienced a period of positive or near-neutral surface elevation change for much of the early 21st century (until 2008 and 2012 for Piramide and the debris free glaciers, respectively), suggesting that they have been in equilibrium with climate over that period. The period of neutral or positive elevation change was followed by a negative trend coinciding with a severe "mega-drought" that has affected central Chile since 2010. The positive elevation changes observed in the first 9 years of our period are consistent with years of positive mass balance in Echaurren glacier between 2000 and 2009. It is only the scarce precipitation associated with the mega-drought that re-established the conditions for strong ablation and thus important mass losses.
- The spatial distribution of the mass balance over the debris-covered glacier is distinct to that of the debris-free glaciers, and its elevation profile is related most strongly to the debris thickness variability and avalanches contribution. There is also a contrast in behaviour in terms of runoff contribution between the debris-covered and debris-free glaciers. The interannual variability of snowmelt contributions from Piramide Glacier is much larger and more sensitive to the 0°C isotherm than the debris-free Bello and Yeso glaciers. By contrast, there is a less variable interannual subdebris icemelt on Piramide Glacier that is decoupled from high frequency climate variability.
- We witness a clear decrease in runoff over the period of study, with very low runoff during the years of the mega-drought, but a decline which is evident from the start of the study period in 2000. Our period of record is too short to confirm a general long-term trend, such as those that have been modelled or suggested by other studies in the region. Dry years show an increased dependency of runoff on the declining resource of glacier ice.

Given this result and those of the global studies, there is a clear need to extend model simulations and reconstruction of geodetic elevation changes for debris-free and debris-covered glaciers to longer time periods, in order to establish whether the peak water has been reached already and what the contribution of distinct types of glaciers is.

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