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1 **Snow cover and snow albedo changes in the central Andes of Chile and Argentina**
2 **from daily MODIS observations (2000–2016)**

3
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33 **Abstract**

34 The variables of snow cover extent (SCE), snow cover duration (SCD), and snow albedo (SAL)
35 are primary factors determining the surface energy balance and hydrological response of the
36 cryosphere, influencing snow pack and glacier mass-balance, melt, and runoff conditions. This study
37 examines spatiotemporal patterns and trends in SCE, SCD, and SAL (2000–2016; 16 years) for central
38 Chilean and Argentinean Andes using the MODIS MOD10A1 C6 daily snow product. Observed
39 changes in these variables are analyzed in relation to climatic variability by using ground truth
40 observations (meteorological data from the El Yeso Embalse and Valle Nevado weather stations) and
41 the Multivariate El Niño index (MEI) data. We identified significant downward trends in both SCE and
42 SAL, especially during the onset and offset of snow seasons. SCE and SAL showed high inter-annual
43 variability which correlate significantly with MEI applied with a one-month time-lag. SCE and SCD
44 decreased by an average of $\sim 13 \pm 2 \%$ and 43 ± 20 days respectively, over the study period. Analysis of
45 spatial pattern of SCE indicates a slightly greater reduction on the eastern side ($\sim 14 \pm 2 \%$) of the
46 Andes Cordillera compared to the western side ($\sim 12 \pm 3 \%$). The downward SCE, SAL, and SCD
47 trends identified in this study are likely to have adverse impacts on downstream water resource
48 availability to agricultural and densely populated regions in central Chile and Argentina.

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61 **Keywords:** Andes; Argentina; Chile; climate change; ENSO; MOD10A1; MODIS; snow albedo; snow
62 cover extent; time series analysis.

63 **1. Introduction**

64 Snow in the semi-arid mountain regions of the central Andes of Chile and Argentina provides
65 important water resources to more than 10 million people and is of major importance for agriculture in
66 this area (Masiokas et al. 2006). Moreover, snow constitutes a key seasonal component in the surface
67 energy and hydrosphere budgets, reflecting incoming solar shortwave radiation (e.g., Konzelmann and
68 Ohmura 1995). Hydrological balance in the cryosphere is highly influenced by the amount of snow
69 precipitation and the spatiotemporal variability of seasonal snow cover extent (SCE). The combined
70 variability of snow precipitation, SCE, and snow cover duration (SCD) directly influences river-runoff
71 variabilities and glacier surface-mass balance conditions (Ragetti et al. 2016; Wilson et al. 2016).

72 On high mountain glaciers, energy availability for snow and ice melt is regulated by surface
73 albedo which is defined as the ratio of incoming solar radiation reflected by a surface (Cuffey and
74 Paterson 2010). Fresh snow, for example, acts as a near perfect reflector with albedo values of up to
75 0.98. However, snow albedo (SAL) diminishes over time as a result of snow metamorphism,
76 decreasing to as low as 0.46 (Cuffey and Paterson 2010). Rainfall can further enhance this natural
77 lowering of SAL through the addition of latent energy, which can initiate melting (Benn and Evans
78 2010) and cause downwasting and thinning of glaciers (Neckel et al. 2017). Snow and ice albedo can
79 also be reduced by the surface deposition of dust and/or anthropogenic soot (Hansen and Nazarenko
80 2004; Cereceda-Balic et al. 2012). In the central Andes, an additional factor which influences SAL is
81 the seasonal formation of penitents. Often forming in areas of low humidity and high solar elevation,
82 snow penitents can result in significant changes in the surface roughness of snow-covered terrain,
83 which, in turn, influences SAL and sublimation conditions (Corripio and Purves 2006).

84 The overall variability of SAL is influenced by a variety of factors: snow grain size, levels of
85 contamination, solar zenith angle, cloud cover, snow metamorphism, surface roughness, age factor, and
86 liquid water content, amongst others (Warren and Wiscombe 1980; Mernild et al. 2015a). Since SAL is
87 a key parameter determining the amount of energy available for surface melting snow and ice, snow-
88 sublimation, and metamorphosis, spatiotemporal variability in SAL is important when determining
89 snow ablation conditions (Male and Granger 1981; Brock et al. 2000; Hock 2005; Gardner and Sharp
90 2010; Mernild et al. 2016a).

91 Spatiotemporal trends in SCE and SAL interpolated from point measurements often include
92 large errors, especially in remote mountainous regions characterized by limited ground observations,

93 localized climate conditions and complex terrain. In comparison, satellite-based remote sensing and
94 satellite derived snow cover products provide opportune sources of large-scale SCE and SAL
95 measurements and have been successfully used as key inputs in climate, atmospheric and hydrological
96 models (Farr et al. 2007; Mernild et al. 2008; Vuille et al. 2008; Mernild et al. 2015a). Remote sensing
97 systems acquiring data from the visible (VIS) to shortwave infrared (SWIR) spectrum with a high
98 temporal resolution are well suited for monitoring SCE and SAL over large areas, providing good
99 spatial and temporal coverage (Wiscombe and Warren 1980; Dozier and Frew 1981; Dubayah 1992;
100 Knap et al. 1999).

101 Several remote sensing based snow cover products are currently available, most of which apply
102 either the normalized difference snow index (NDSI) (Hall et al. 1995), empirical relationship
103 assumptions or spectral un-mixing models (Klein and Stroeve 2002a). Optical sensor systems,
104 however, are unable to acquire useful information during cloudy conditions (Justice et al. 1998;
105 Marchane et al. 2015). Therefore, frequent satellite observation revisits are essential to study changes
106 in SCE and SAL, since surface conditions can vary rapidly and may change considerably over a few
107 days.

108 To compensate for extensive cloud cover, compromises are often made by conducting satellite
109 analysis based on composite products such as the MODIS (Moderate Resolution Imaging
110 Spectroradiometer) 8 day snow cover MOD10A2 product (Hall et al. 2002), which can mask subtle
111 changes in SCE and SAL over time. In order to avoid this limitation, the MODIS MOD10A1
112 Collection 6 (C6) dataset was used this study. MOD10A1 provides daily SCE and SAL values globally
113 at a spatial resolution of 500 m, making it suitable for evaluating seasonal trends in SCE and SAL (Hall
114 et al. 2002; Liang et al. 2005; Marchane et al. 2015; Hall and Riggs 2016; Saavedra et al. 2016; Li et al.
115 2017; Huang et al. 2017; Dariane et al. 2017), snow cover phenology (Xu et al. 2017) and the relation
116 between SCE and climate (Gurung et al. 2017; Li et al. 2017). Using MOD10A1 data, this study
117 analyses spatiotemporal changes in SCE and SAL in the central Andes of Chile and Argentina by
118 parameterizing a time series of seasonal SCE and SAL metrics at the per-pixel level. Furthermore, this
119 study examines the large-scale influence of ENSO events on SCE and SAL as well as the more
120 localized effect of climatic variability (utilizing meteorological data from the El Yeso Embalse (EYE)
121 and Valle Nevado (VN) automatic weather stations (AWS)) and elevation.

122

123 **2. Study area**

124 The Andes of central Chile and Argentina (31°S and 40°S) contain some of the highest peaks of
125 the entire Andes Cordillera, reaching altitudes above 6,000 m above sea level (a.s.l.) (Fig. 1). Covering
126 an area of ~1,730 km², the study area chosen is located immediately west of Santiago de Chile
127 (32°50' – 34°50'S; 69°20' – 70°40'W). This study area includes several river basins which supply
128 freshwater to large downstream populations (10+ million people in Chile and 2+ million in Argentina),
129 hydro-power stations, and agricultural lands on both sides of the cordillera (Corripio and Purves 2006).
130 This area of the central Andes also includes the largest glaciated areas in South America outside
131 southern Patagonia (Saavedra et al. 2016). River runoff in this central region originates primarily from
132 snowmelt (Masiokas et al. 2006), with snowfall contributing up to ~85 % of runoff from specific
133 catchments (Mernild et al. 2016b). The availability of snow as a freshwater resource is therefore of
134 vital socio-economic importance in this semi dry region (Peña and Nazarala 1987; Meza et al. 2012;
135 Carey et al. 2017).

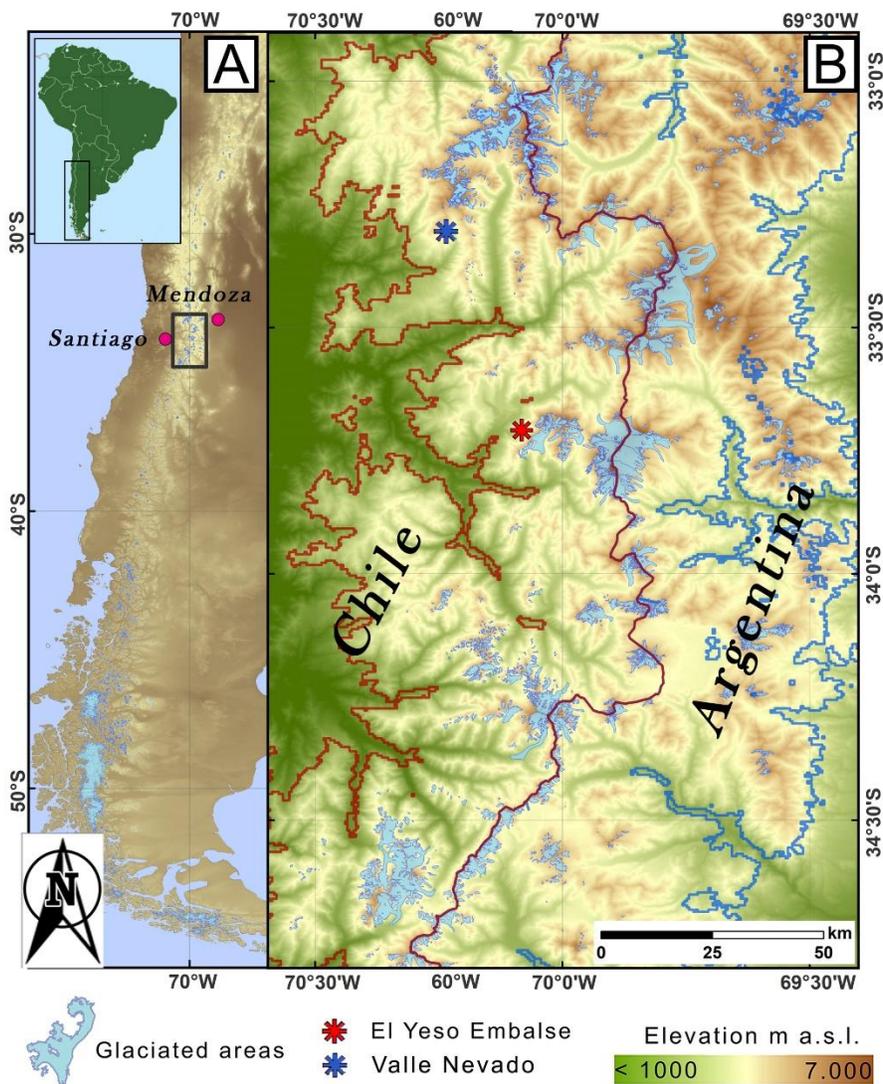
136 The intra-annual variability of precipitation in central Andes is highly influenced by the
137 placement of an atmospheric high-pressure cell over the southeastern Pacific Ocean. This cell normally
138 inhibits precipitation in the Austral summer (December – February) and allows for the passage of
139 westerlies and frontal precipitation during Austral winters (June – August) (Garreaud et al. 2009).
140 Precipitation events are usually concentrated between April and October, providing ~95 % of the mean
141 annual totals, peaking in June or July (Masiokas et al. 2016). The strength of El Niño Southern
142 Oscillation (ENSO) influences inter-annual variability in precipitation, with higher/lower precipitation
143 occurring during El Niño/La Niña events (Rutllant and Fuenzalida 1991; Escobar et al. 1995; Leiva
144 1999; Montecinos and Aceituno 2003; Garreaud et al. 2009). During El Niño events, precipitation
145 increases predominantly during the austral winter (Masiokas et al. 2006; McClung 2013). Whilst El
146 Niño events do influence precipitation amounts, these events shows little or no significant signal in
147 annual mass balance measurements of glaciers located in the central Andes but has been linked to the
148 Pacific Decadal Oscillation (PDO) rather than the ENSO (Mernild et al. 2015a).

149 Along the central Andes, annual accumulation of snow is highest at 4,000 – 5,000 m a.s.l.,
150 where glacier accumulation zones are also present (Cornwell et al. 2016; Mernild et al. 2016b; Mernild
151 et al. 2016c). Precipitation differences observed between the western and eastern sides of the Andes
152 Cordillera occur due to the combination of orographic effects of the mountain relief and the dominating

153 westerly wind direction which results in precipitation amounts and humidity being lower on the eastern
154 Cordillera slopes (Cornwell et al. 2016; Mernild et al. 2016b).

155 For the central Andes, mean surface air temperatures are normally highest between December
156 and March and lowest in July and August (Masiokas et al. 2016) and temperatures in the Andes showed
157 increasing trends from 1975 to 2006 ($\square 0.25^\circ\text{C}/\text{decade}$) (Falvey and Garreaud 2009). The 0°C isotherm
158 for the western side of the cordillera (40 km northeast of Santiago de Chile), was located at 3,385 m
159 a.s.l. between 2009 and 2014 (Mernild et al. 2016c).

160



162 Figure 1: (a) The central Andes of Chile and Argentina; and (b) including the area of interest west and
163 east snow cover regions delineated by red and blue lines, respectively. The divide between the western

164 and eastern Andes also represents the natural border (continental divide) between Chile and Argentina.
165 Blue areas represents glaciers. The red star in the center of the study area represents the location of the
166 El Yeso Embalse (EYE) meteorological station (2475 m a.s.l.) and the blue star the Valle Nevado (VN)
167 meteorological station (3050 m a.s.l.).

168

169 **3. Data**

170

171 *3.1 MODIS data*

172 The MOD10A1 C6 (henceforth MOD10A1 unless other version is implied) snow product is
173 derived from daily data acquisitions by the MODIS sensor aboard the Terra spacecraft (Riggs et al.
174 2017). The MODIS global daily snow cover product MOD10A1 (MODIS/Terra Snow Cover Daily L3
175 Global 500m Grid) is derived from cloud free observations and is well suited for regional snow cover
176 and albedo mapping (Hall et al. 2002; Liang et al. 2005; Dozier et al. 2008; Rittger et al. 2013; Fausto
177 et al. 2015; Mernild et al. 2015b). The latest MOD10A1 product was released in the spring of 2016 and
178 includes a range of improvements to the previous version including, amongst others, the removal of
179 Terra sensor degradation issues and improvements in atmospheric calibration (Lyapustin et al. 2014).
180 Importantly, the algorithms used to compile the MOD10A1 snow product are modified to include only
181 the best quality observations from the atmospherically corrected MOD10GA product (Hall et al. 2002).
182 Individual MOD10A1 product parts include NDSI, NDSI snow cover (SCE), SAL and corresponding
183 quality control flags. The MOD10A1 NDSI SCE is produced by using an empirical relationship with
184 NDSI values, where NDSI values are multiplied by a constant (Dozier et al. 2008; Hall and Riggs
185 2016). By using only full snow cover pixels, the accuracy of the MOD10A1 SAL product is improved
186 in terms of ground truth comparisons (Sorman et al. 2007; Mernild et al. 2015b). The overall error of
187 the MOD10A1 SAL product can vary substantially but is in the order of 1–10 % for good observations
188 with low atmospheric disturbances across the Greenland ice sheet (Klein and Stroeve 2002b). The
189 overall error at this location is likely higher due to the complex terrain of the Andes Mountains.
190 However, changes in albedo can still be quantified and here we opted for a per-pixel temporal change
191 analysis that is expected to mitigate the influence of topography to some degree by avoiding direct
192 inter-comparison of pixels influenced by different slope/aspect.

193 MOD10A1 data used in this study was obtained from the NASA Earth Observation System
194 Data and Information System (EOSDIS) Reverb ECHO website (<https://reverb.echo.nasa.gov/>). Out of
195 the 5,844 potential scenes available between March 1 2000 and Feb 29 2016, only 121 (~2 %) were
196 missing in the archive, with a maximum temporal gap of 17 days. The snow cover year (season) was
197 set to start 1 March and end February 28 (29) based on analysis of the data set. Pixels with cloud cover
198 or poor retrievals were omitted as determined from QA flags and only pixels flagged as “best quality”
199 were included in the further analysis (see section 4.1). Out of all pixels in the data series 74.8 %
200 contained “best quality” data (supplementary material S1).

201

202

203 *3.2 Ancillary data*

204 Elevation data were obtained from the Shuttle Radar Topography Mission (SRTM) v.3. SRTM
205 provides elevation data at a spatial resolution of 30 meters with an overall vertical accuracy of ~10 m
206 and geo-position error of ~9 m (Farr et al. 2007). Multivariate El Niño index (MEI) ranks were
207 obtained from the National Oceanic and Atmospheric Administration (NOAA) website
208 (<http://www.esrl.noaa.gov/psd/enso/mei/table.html>). The MEI provides a ranked index of the strength
209 of El Niño and La Nina events. MEI values are normalized for each bimonthly season (Wolter and
210 Timlin 2011) and cover the 16-year study period of snow cover and snow albedo observations.

211 We acquired mean monthly air temperature (MMAT), mean annual air temperature (MAAT),
212 and monthly precipitation sums (2000–2016) from the El Yeso Embalse meteorological station (EYE)
213 supported by available data from the Valle Nevado (VN) meteorological station from 2013 onwards
214 (locations in Fig. 1; data shown in supplementary material S2 and S3) from Dirección General de
215 Aguas (DGA; www.dga.cl) and CryoNET respectively
216 (<http://globalcryospherewatch.org/cryonet/sitepage.php?surveyid=68>). Finally, glacier outlines were
217 obtained from the Randolph glacier inventory v. 5.0 (Pfeffer et al. 2014) in combination with updated
218 glacier shapes from 2013/2014 (Malmros et al. 2016).

219

220 **4. Methods**

221

222 *4.1 Time series preprocessing*

223 The MOD10A1 time series was preprocessed and analyzed using the program TIMESAT
 224 (Jönsson and Eklundh 2002, 2004; Eklundh and Jönsson 2015). TIMESAT originally developed to
 225 analyze vegetation seasonality can be applied to all remote sensing data containing seasonal variability.
 226 We applied a Savitzky-Golay filter within TIMESAT to smooth the MOD10A1 time-series by applying
 227 polynomial fitting to the data points within a moving window of a certain width (Savitzky and Golay
 228 1964; Jönsson and Eklundh 2004). Missing dates were filled with blank scenes before smoothing
 229 (Jönsson and Eklundh 2002) in order to compose a complete time series. Pixels not flagged as “best
 230 quality” from the MOD10A1 QA flags were excluded from the analysis. The width of the moving
 231 window influences the degree of smoothing and the ability of the filter to cope with rapid changes
 232 (parameters used in the TIMESAT preprocessing are shown in Table 1). The polynomial fitting was
 233 iterated and adapted to the upper part of the curve by assigning weights to data points above and below
 234 the result of the previous step (Jönsson and Eklundh 2004). SCD was extracted in TIMESAT for SCE,
 235 and the seasonal snow cover integral (SCI) (defined as the integral under the curve between onset and
 236 end of seasonal snow cover) was extracted to evaluate the accumulated seasonal SCE for each season.
 237 Areas characterized by limited seasonal variability were masked out due to the inability of the
 238 TIMESAT algorithm to estimate SCD for such conditions. The mask was created from the median of
 239 the 16 seasons and excluded areas with less than 16 days in the SCD dataset. Most excluded areas were
 240 located in glacier accumulation zones where constant snow cover prevents seasonal variability in SCE.

241

242 Table 1: Input parameters for use in TIMESAT.

Parameter	NDSI Snow Cover	Snow Albedo
Seasonal parameter	0.7	0.7
Number of envelope iterations	3	3
Adaptation strength	2	2
Savitzky-Golay window size	15	15
Spike Method	Median filter	Median filter
Amplitude season start (%)	70	65
Amplitude season end (%)	20	46

243

244 *4.2 Trend estimation*

245 We conducted linear temporal trend analysis to estimate the magnitude and direction of changes
246 in SCE, SAL, SCD, SCI, MAAT, and annual precipitation. We calculated per-pixel trends by applying
247 a nonparametric linear regression model with time as the independent variable and the abovementioned
248 variables as dependent variables. Since time series of the variables analyzed often do not meet
249 parametric assumptions of normality and homoscedasticity, a median trend (Theil–Sen, TS) procedure
250 was applied using the Theil–Sen slope estimator (median trend) which has proven robust against
251 outliers (Eastman 2009). Uncertainty estimates of trends are calculated from standard deviations of the
252 calculated metrics and are provided as \pm values to all trends reported. The significance of the trends
253 was determined using the nonparametric Mann–Kendall test of significance (Mann 1945; Kendall
254 1975). The Mann–Kendall significance test is commonly used as a trend test for the TS median slope
255 operator (Eastman 2009) and produces outputs of z-scores that allow for the assessment of both the
256 significance and direction of trends. The trends were considered significant at $p < 0.05$ (where p
257 denotes the probability that there is no significant difference between observations over time). Trends
258 in SCE, SCD, and SCI were extracted from the area of maximum values (minimum of 75% at any
259 given time) for the entire study period and similar for SAL.

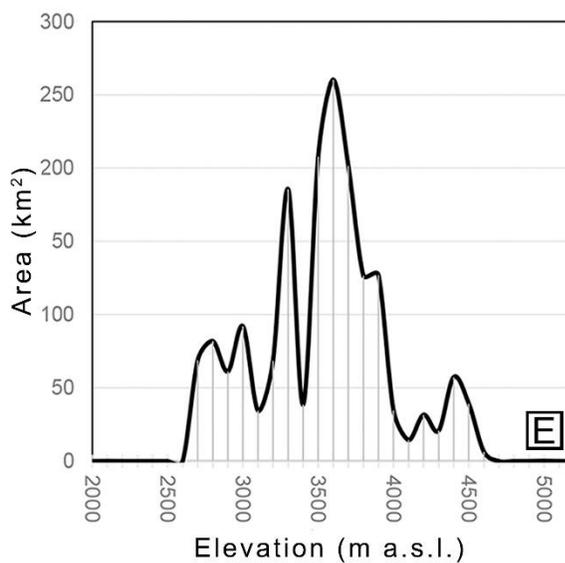
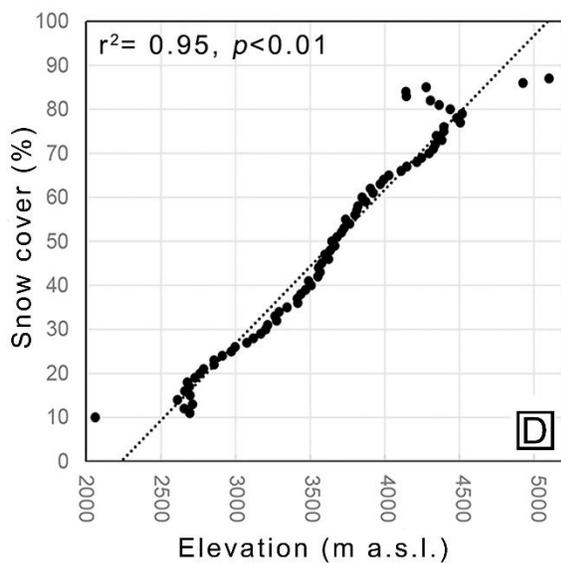
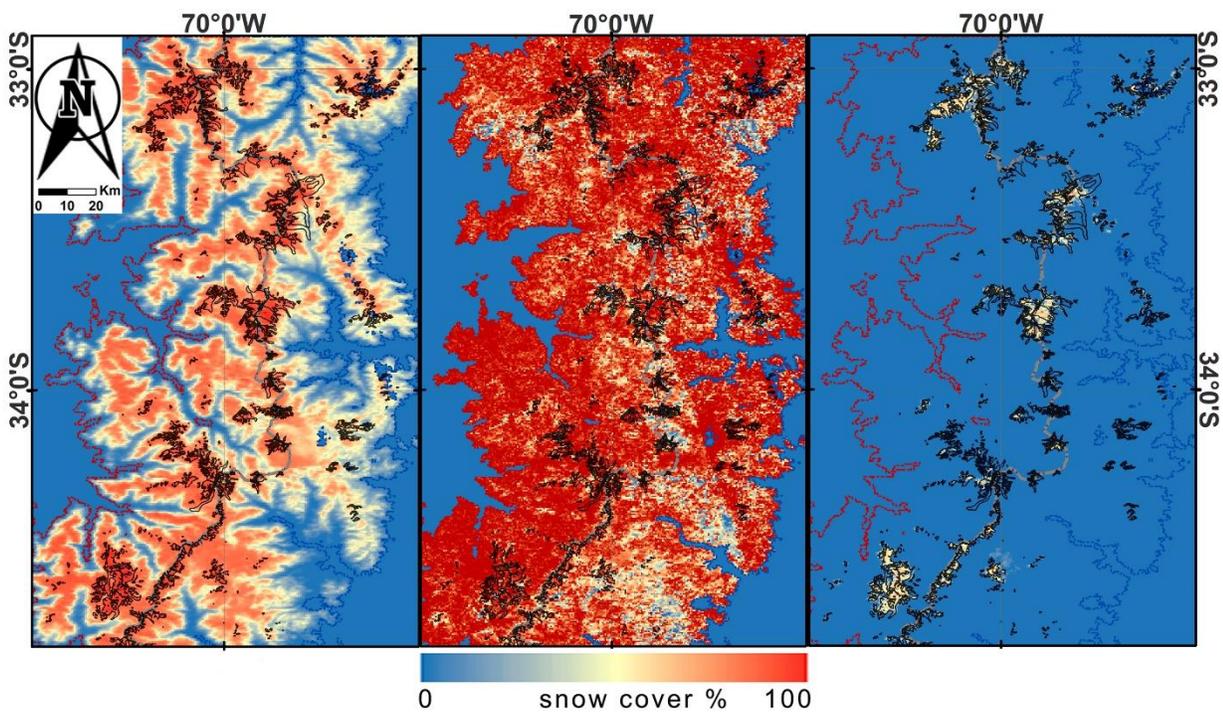
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261 **5. Results**

262 *5.1 Spatio-temporal patterns in snow cover extent*

263 Statistical analysis of the MODIS time series revealed that the minimum, median, and
264 maximum SCE for the entire observation period and area (Fig. 2) was 2 %, 43 %, and 74 %,
265 respectively. SCE showed the presence of a clear linear relationship with elevation (coefficient of
266 determination (r^2) = 0.95, $p < 0.01$; Fig. 2d). However, the relationship was imperfect ($r^2 = 0.34$) above
267 4,000 m a.s.l due to increased scattering in the accumulation zone of the glaciers. At its maximum, SCE
268 covered 1,730 km² (74%) of the study area, only being present at elevations above 1,250 m a.s.l. SCE
269 area was normally distributed with highest concentration of SCE observed between 3,250 and 4,000 m
270 a.s.l. (Fig. 2e). SCE was on average ~11 % less on the eastern side of the Cordillera compared to the
271 west (Fig. 2a-c).

272



273

274 Figure 2: a) Median snow extent; b) maximum snow cover extent; c) minimum snow cover extent for
 275 the 16 snow seasons observed (2000–2016) (glaciers are delineated by black); d) Mean SCE change
 276 with elevation; and e) Snow covered area distribution with elevation.

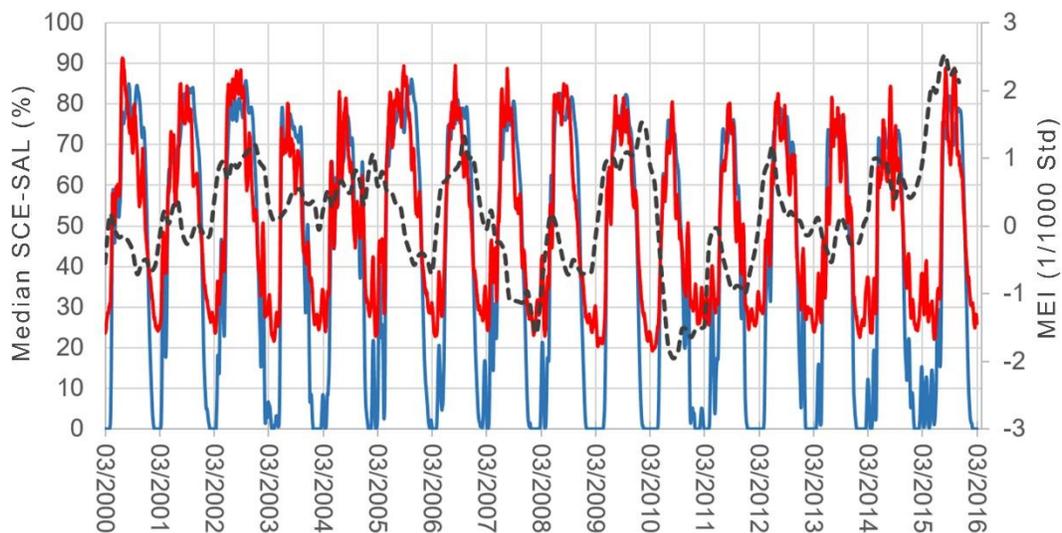
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278 *5.2 Inter- and intra-annual variability in snow cover extent (SCE) and snow albedo (SAL)*

279

280 The mean SCE over the study area between 2000 and 2016 was approximately ~25 % in austral
 281 summer and ~68 % in austral winter, reaching a maximum between July 24th and 25th October and a
 282 minimum in autumn between March 6th and June 6th (Fig. 3). Overall, spatial trends for SCE and SAL
 283 showed widespread downward tendencies (Fig. 4). For SCE, downward trends dominated, with
 284 medium SCE decreasing by 13.4 ± 4.0 % between 2000 and 2016 (0.35 % yr^{-1}). Demonstrating spatial
 285 variabilities, the downward trends in SCE were found to be more pronounced for the the eastern side of
 286 the Cordillera (13.9 ± 4.0 %, 0.9 % yr^{-1}) compared to the west (12.4 ± 3.1 %, 0.8 % yr^{-1}). The eastern side
 287 of the Cordillera also showed lower SCE but higher SCE variability compared to the western side, this
 288 despite the eastern side being more elevated (mean elevation for the western and eastern sides is 3,115
 289 m a.s.l and 3,720 m a.s.l, respectively) (Fig. 2d). Median SAL trends from March 2000 to February
 290 2016 were also downward with mean values decreasing by 7.4 ± 2.2 % (0.5 % yr^{-1}). Downward trends in
 291 SAL were shown to be more widespread in the southern parts of the study area compared to northern
 292 parts. However, we found no significant differences between the eastern and western sides of the
 293 Cordillera.

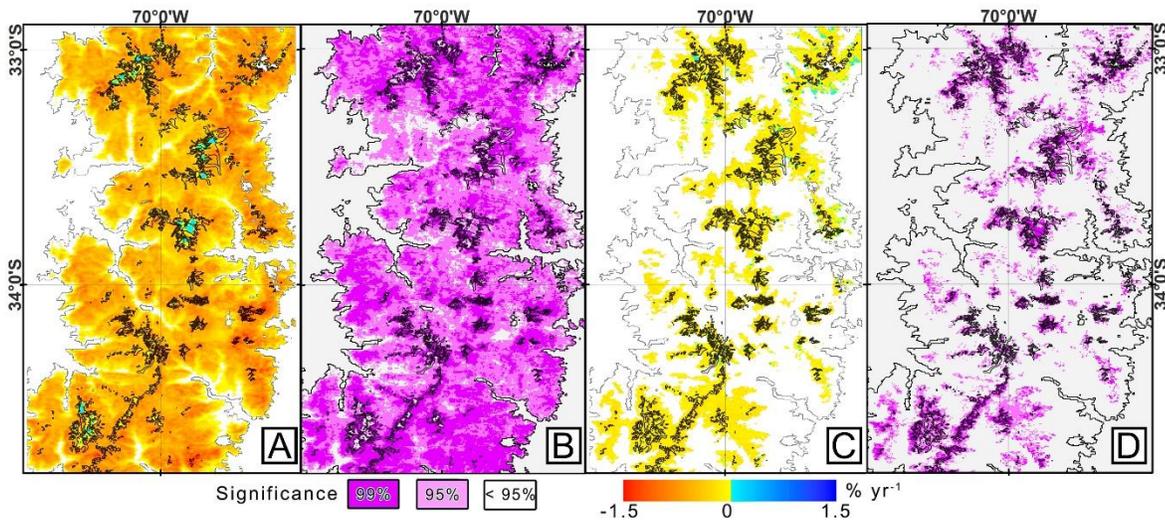
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296 Figure 3: Variation in daily median SCE (blue) and median SAL (red) percentage and the bi-monthly
 297 multivariate ENSO index (MEI) (dark dashed line) in 1/1000 standard variations (Wolter and Timlin
 298 2011).

299



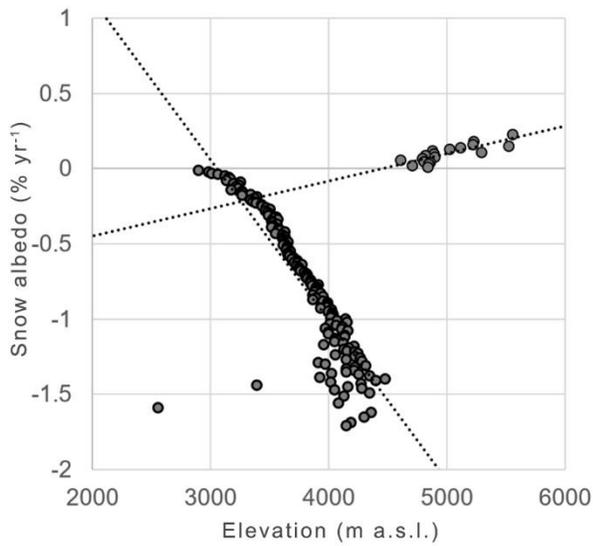
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301 Figure 4: Linear median trend ($p < 0.05$) in: SCE (a) and SAL (c) from Mar 2000 through Feb 2016, and
 302 the corresponding significance level of each trend (b, d).

303

304 Comparisons with SRTM data, the SAL trends observed showed negative correlations below $\sim 4,600$ m
 305 a.s.l. (-1.1 ± 0.8 % yr^{-1} , $r=0.84$, $p < 0.01$). Above $\sim 4,600$ m a.s.l., however, a distinct shift is observed,
 306 with increasing correlations as a function of elevation (0.2 ± 0.01 % yr^{-1} , $r=0.83$, $p < 0.01$) (Fig. 5). In
 307 regards to intra-annual variability, SCE showed pronounced downward trends (< -1 % yr^{-1}) during the
 308 onset (April, May, and June) and offset (October, November, December, and January) of the snowy
 309 season. In comparison to SCE, SAL showed slightly less intra-annual variability with most downward
 310 trends occurring during the onset of the snowy season (Fig. 6).

311

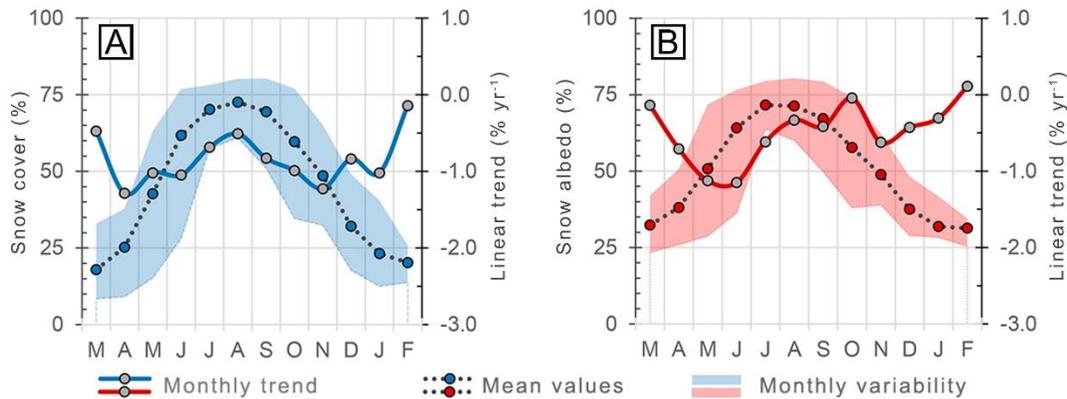


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313 Figure 5: Trends in snow albedo (SAL) with elevation.

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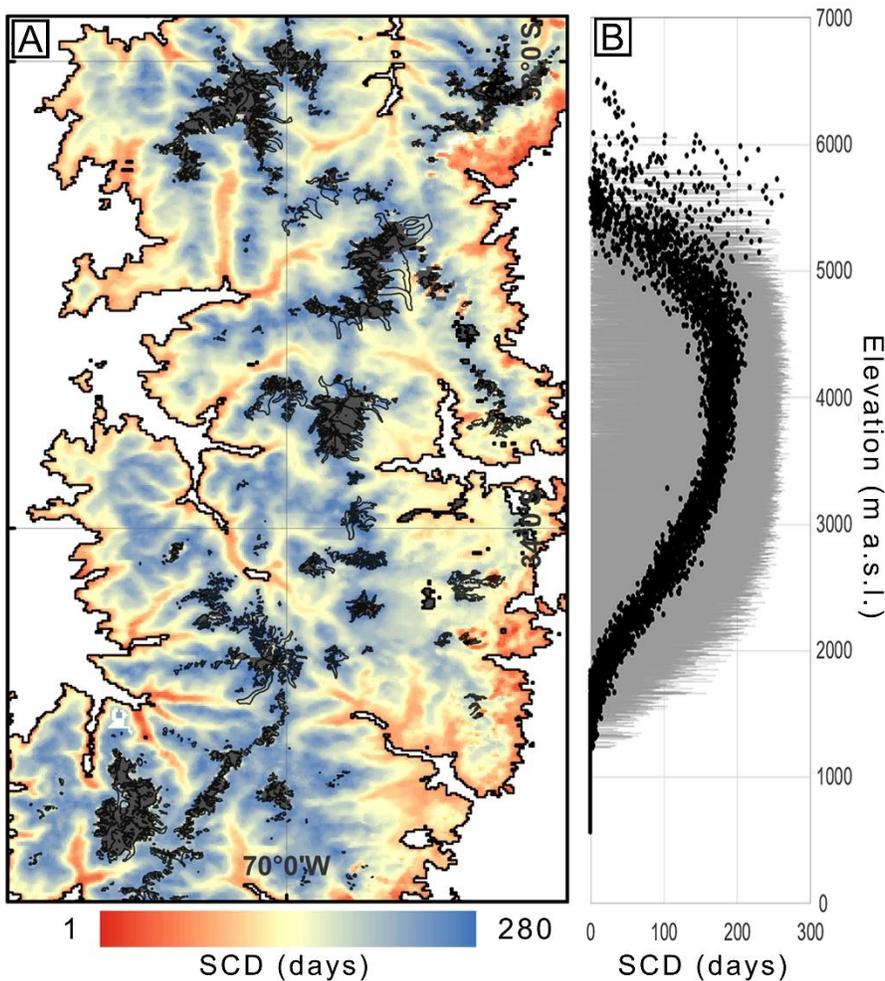
317 Figure 6: Intra-annual linear trends (blue and red lines), monthly mean values (dotted lines) and spread
 318 (colored areas; monthly minimum and maximum values)) in: (a) snow cover extent (SCE); and (b)
 319 snow albedo (SAL) between 2000 and 2016.

320

321 *5.3 Snow cover duration (SCD) and seasonal snow cover integral (SCI)*

322 The seasonal SCD values ranged between 0 and 280 days and the seasonal mean SCD ranged
 323 between 203 days in the 2005–2006 to 130 days in 2011–2012, with an overall median of 173 days
 324 (Fig. 7a). We observed a strong correlation between SCD and elevation (excluding glacier areas). SCD,
 325 for example was shown to increase by an average of ~6 days for every 100-meter increment within the

326 2,000 to 4,600 m a.s.l. elevation range ($r^2=0.80$, $p<0.01$) (Fig. 7a). Overall, the trends in SCD were
 327 downward, showing a mean reduction of 43 ± 20 days for the study period (-2.7 ± 1.3 days yr^{-1}) (Fig.
 328 8a). Trends were especially negative on the eastern side of the Cordillera, with a reduction of 52 ± 36
 329 days (3.3 ± 2.3 days yr^{-1}), whereas on the western side SCD reduced by 35 ± 33 days (2.2 ± 2.0 days yr^{-1}).
 330
 331



332
 333 Figure 7: a) Median snow cover duration (SCD, days) 2000–2016 (dark grey areas represent locations
 334 where the model failed to determine seasonality); and (b) Mean snow cover duration with elevation for
 335 the 2000–2016 period (grey bars show the spread between minimum and maximum values of SCD as a
 336 function of elevation).
 337

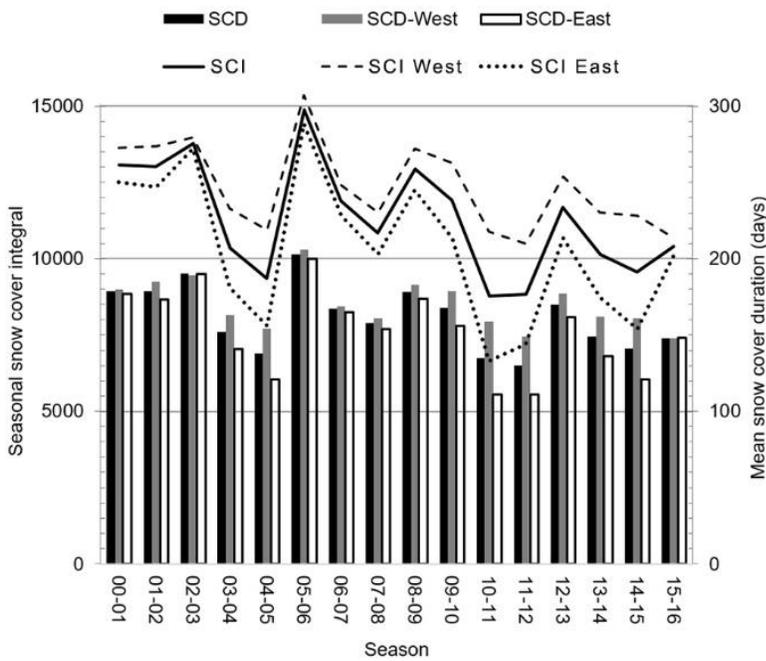
338

339 Figure 8: (a) Per pixel yearly snow cover duration (SCD) trend 2000–2016, and (b) corresponding
340 significance of trend. Dark gray masked areas represent the mask where SCD values are erroneous and
341 black lines represent glaciers.

342

343 The SCI for the entire study area declined by $1.5 \pm 0.5 \text{ \% yr}^{-1}$ (median trend), corresponding to 25 ± 8
344 % during the full period of 16 snow seasons. The SCI also shows substantial inter-annual variation
345 (Fig. 9), especially on the eastern side of the Cordillera, where SCI is on average 17 % smaller and
346 trends are substantially more downward ($1.8 \pm 0.5 \text{ \% yr}^{-1}$) than for the western side ($1.2 \pm 0.4 \text{ \% yr}^{-1}$).

347



348

349 Figure 9: Mean snow cover duration (SCD, bars) and seasonal snow cover integral (SCI, lines) for the
350 entire study area and for the western (SCD-W) and eastern (SCD-E) sides of the cordillera separately.

351

352 5.4 Impact of changes in temperature, precipitation, and El Niño Southern Oscillation (ENSO) on snow 353 cover duration and snow albedo

354

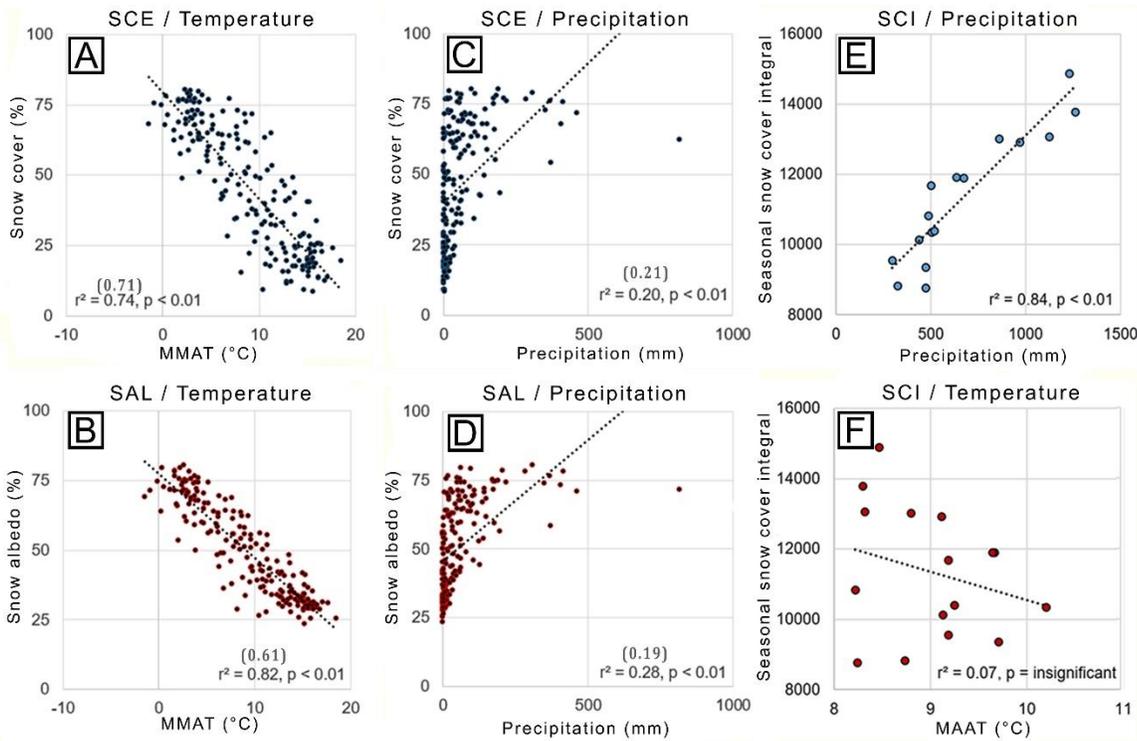
355 Analysis of the meteorological data available from the EYE AWS (Figure 1b) revealed a

significant downward trend in precipitation of -4 mm yr^{-1} and an insignificant trend in MAAT of

356 0.05°C yr⁻¹ between 2000 and 2016. Between 2000 and 2009, nine extreme precipitation events (>200
357 mm month⁻¹) were identified, with the 2003-2004 season (March-February) receiving a maximum of
358 1,259 mm. These 'extreme' events occurred mostly during austral winters and account for large
359 differences in inter-annual precipitation amounts, SCE and SCD sums. Post 2009, no extreme
360 precipitation events occurred, however a minimum of 317 mm was observed for the 2014-2015 season.
361 The mean annual precipitation sum for the 2000 to 2016 observation period was 677 mm. MAAT for
362 this observation period was 9.0°C, with maximum and minimum values of 10.3°C and 7.3°C measured
363 for the 2003-2004 and 2011-2012 seasons, respectively.

364 Statistical comparisons between the MODIS-derived snow data and the observed EYE
365 meteorological data revealed that monthly SCE and SAL averaged over the study area and for the pixel
366 where EYE is located (in brackets) correlated strongly with MMAT (r^2 -values of 0.74 (0.71) and 0.82
367 (0.61), respectively) (Fig. 10a-b). For the VN met-station, the corresponding comparison between SCE
368 and MMAT also show a significant correlation for the 2013 to 2016 period (r^2 -values of 0.62). Mean
369 monthly SCE and SAL also correlate with monthly precipitation sums, but to a lesser extent (r^2 -values
370 of 0.20 (0.21) and 0.28 (0.19)) (Fig. 10c-d). In comparison, SCI correlated more strongly with annual
371 precipitation sums (r^2 -value of 0.84) (Fig. 10e).

372



373

374 Figure 10: Relationships between monthly meteorological data (EYE station; Fig. 1) and monthly SCE,
 375 monthly SAL, and SCI averaged over the study area. a) Monthly SCE against MMAT; b) monthly SAL
 376 against MMAT, c) Monthly SCE against monthly precipitation sums, d) Monthly SCE against monthly
 377 precipitation sums, e) SCI against annual precipitation sums, and f) SCI against MAAT. r^2 -values
 378 between the EYE station data and the pixel where it is located is shown in brackets.

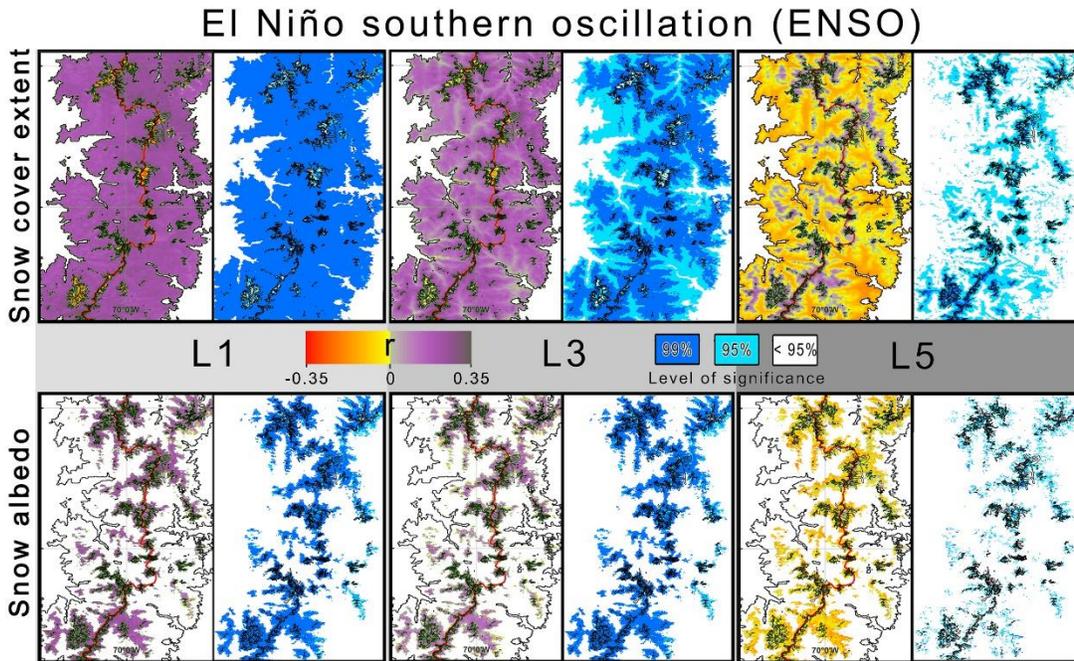
379

380 There can be large uncertainties introduced by comparing a single station with observations
 381 from a larger study area. However, currently it remains the only long-term continuous dataset available
 382 in the area, and trends from the local pixel and the station confirms the relationship with trends for the
 383 area in general (Fig. 10). Furthermore, EYE and NV temperature data significantly correlated (r^2 of
 384 0.89) showing the temperature trends at EYE and NV being similar at least for that part of the western
 385 side of the cordillera in the 2013-2016 period.

386

387 ENSO events (MEI) (plotted in Fig. 3) show significant correlations with mean SCE and SAL
 388 values for the study area (Fig. 11). Mean correlation coefficient values (r) between MEI and SCE/SAL
 389 were strongest when a lag of one month (L1) was applied to the SCE/SAL time series (mean r -values

390 of 0.21 for the study area). Generally, glaciated areas where snow cover is already present are
 391 characterized by negative SCE and ENSO correlations for shorter lag periods (L1), which is in contrast
 392 to the surrounding snow covered areas. This pattern is reversed for longer lag periods (e.g., five
 393 months), when glaciated and higher elevated areas significantly correlate with ENSO, whereas
 394 surrounding lower altitude areas are characterized by negative correlations.
 395



396
 397 Figure 11: Per-pixel correlations and significance of SCE/SAL with the ENSO index (MEI) including
 398 one-, three- and five-month lag time (L1, L3, and L5). Red line demarks the east west divide.
 399

400 6. Discussion

401 402 6.1 Monitoring and assessment of snow cover and snow albedo

403 Spatio-temporal analysis of MODIS-derived SCE, SAL, SCD, and SCI data revealed significant
 404 changes during the observation period of this study. Key to the interpretation of these results is the
 405 quantification of sensor related errors and their influence on the trends observed (2000–2016).
 406 Unfortunately, only a few validation studies of the MOD10A1 C6 products have been published to
 407 date. Recent studies using MOD10A1 collection C6 for the Greenland ice sheet however found that

408 v.C6 corrects for the C5 temporal trend bias in dry snow areas and that albedo retrieval accuracy in C6
409 is substantially improved over C5 (Box et al. 2017; Casey et al. 2017). Therefore, the accuracy of this
410 product is expected to be better than or at least as good as the C5 product, which has been evaluated in
411 several previous studies (Tekeli et al. 2005; Hall and Riggs 2007; Gao et al. 2010; Arsenault et al.
412 2014; Marchane et al. 2015). These studies estimate an overall detection error ranging from ~5 % to
413 ~48 %, depending on locational properties and type of ‘ground truth’ observations used. In general,
414 spatially homogeneous locations with flat terrain produces less error than in complex terrain with
415 mixed surface as being the case for this study (Wang et al. 2014; Burakowski et al. 2015; Moustafa et
416 al. 2017). By applying a per-pixel temporal change analysis approach in the current study, thereby
417 avoiding direct inter-comparison of pixels characterized by different slope/aspect, the influence of
418 topography is expected to be reduced to some degree. This should also be seen in the context of the
419 latitudinal location of the study area (32°50’– 34°50’ S), characterized by an annual range in solar
420 zenith angles of 28-60 for MODIS overpass times. This makes the region less prone to influences from
421 mountain shadowing as compared to complex terrain of higher latitudes of the northern/southern
422 hemisphere. Snow detection in v.6 is expected to show improvements in comparison to previous
423 versions especially above 1.300 m a.s.l., where the surface temperature screen used in the product
424 algorithm (which has previously caused false negatives) has been rolled back leaving fewer gaps in the
425 data (Hall and Riggs 2016).

426 The smoothed and gap-filled MOD10A1 C6 dataset produced with TIMESAT is assumed to
427 correctly represent the seasonal snow distribution for the area. Here, only the best quality MOD10A1
428 C6 observations (by including information available from the QA flags; 0=“best quality”) were used
429 for the Savitzky-Golay function fitting in TIMESAT. By adjusting the data fitting to the upper
430 envelope of the daily observations, we ensured that the SCE and SAL values follow rapid changes,
431 which can occur in snow cover extent and albedo. The TIMESAT model uses local function fitting,
432 where values before and after in the time series are considered. This local function fitting reduces the
433 chance of error occurrence in the MOD10A1 observed snow cover (Tekeli et al. 2005). For the upper
434 ablation zones of glaciers, characterized by limited seasonal variability or year round snow SCE, it is
435 not possible to accurately assess seasonality variables and thereby SCD. In this case, a glacier mask
436 was applied based on SCE seasonal variability, in doing so, restricting the SCD analysis to non-

437 glaciated areas (Fig. 7a). Out of 709029144 individual pixel records in the time series 24.2 % were
438 filled with modeled TIMESAT data.

439 Snow albedo detection in mountainous environments from remotely sensed imagery can contain
440 large errors when measured on terrain with steep slopes. Validation of satellite or aerial imagery based
441 data using stationary point albedometers can also be challenging because of pronounced mixed pixel
442 and geolocation issues (Liang et al. 2005; Sorman et al. 2007; Mernild et al. 2015b; Box et al. 2017).
443 However, these issues are not likely to have significantly influenced the trends observed in this study as
444 the MOD10A1 pixels are measured in the same pixel location from year to year in a location where
445 seasonal variation in solar zenith angle influence is relatively low (annual range between 28-60
446 degrees) compared to higher latitudes of the northern/southern hemisphere. MOD10A1 snow albedo is
447 produced only for cloud free pixels with full snow cover (+50 %) indicating that for pixels
448 characterized by limited full snow cover observations seasonal fitting could have influenced the
449 accuracy of the TIMESAT generated data.

450

451 *6.2 Analysis of climate variables*

452 The analysis of the effect of local scale climatic variability presented in this study made use of
453 two AWS's EYE and VN in the study area. Indeed, the use of a single or just two AWS's is not ideal
454 when comparing measurements with the spatially large-scale MOD10A1 dataset. Furthermore, given
455 the location of the AWS's on the western side of the cordillera and the presence of distinct climatic
456 gradients (e.g. Mernild et al. 2016a), the data recorded is unlikely to be fully representative of the study
457 area as a whole. However, this scarce coverage of ground observations reflects the general conditions
458 for mountainous areas of the Andes and underlines the need for remotely sensed monitoring methods.
459 Additionally, there is a need for high-resolution gridded meteorological products that work well in
460 complex terrain. A high-resolution meteorological dynamical model combined with remote sensing
461 derived products is likely to improve our understanding of the climatic variability impact on snow
462 cover dynamics. Upward trends in MAAT have higher impact the lower the altitudes (mostly
463 noticeable in the southern part of the study area) causing changes in onset and offset of snow seasons to
464 be more sensitive to even small changes in temperature (Fig. 4). Low areas with small slope gradient
465 show higher sensitivity to upward changes in MAAT in regards to snow accumulation. Especially the
466 low elevated areas on the eastern side of the cordillera show the effect of increased MAAT.

467 Above 4,600 m a.s.l., SCD shows a considerable increase in variability (Fig. 7b). This however
468 is not surprising as, in highly elevated zones where SCD can be dependent on localized terrain (slope
469 and area of topographical shadow) and weather conditions. High wind speeds, for example, often make
470 it less likely for snow cover to persist in certain areas despite high levels of solid precipitation.

471

472 *6.3 Drivers of change on snow cover variables*

473 Studies based on modeling, field measurements and remote sensing have provided insights into
474 the past, current, and future impacts of climate change on snow conditions and runoff in the Andean
475 river catchments of central Chile and Argentina (Pellicciotti et al. 2007; Apaloo et al. 2012; Delbart et
476 al. 2015; Mernild et al. 2015a, 2016a, 2016c; Ragettli et al. 2016). A number of these studies have
477 predicted that air temperatures in the central Andes will continue to increase. An increase in air
478 temperature, together with seasonal changes in precipitation patterns, will likely result in a decrease in
479 the amount of runoff from snow melt and an increase in the amount of runoff from rain (Cai et al.
480 2014; Mernild et al. 2016a, 2016c; Ragettli et al. 2016). Since the 1970's, precipitation events have
481 generally become more intense but less frequent in central Chile (Falvey and Garreaud 2007; Garreaud
482 et al. 2009). The EYE precipitation data, for example, shows a number of intense precipitation events
483 (above 200 mm m⁻¹) during the 2000–2009 period. Interestingly, none of these 'intense' events
484 occurred during the 2010–2016 period.

485 Per-pixel correlations between SCE/SAL and the Multivariate El Niño index (MEI) show that
486 MEI has a strong and significant impact on inter-annual SCE/SAL variability in the region (Fig. 11).
487 Although El Niño events are often associated with increases in precipitation, they can also be
488 associated with increases in air temperature (Cai et al. 2014) which, together, can have a pronounced
489 altitude dependent effect on snow cover during spring and autumn. An increase in air temperature, for
490 example, causes the 0°C isotherm to ascend to higher elevations resulting in a larger proportion of
491 precipitation falling as rain as opposed to snow. Mernild et al. (2016c) observed this phenomenon for
492 the Olivares basin (33°12' S; 70°09' W) between 1979 and 2014, where precipitation has been
493 increasingly falling as rain in recent years. This change in the partitioning of precipitation over
494 mountainous areas can offset the positive effects of increased precipitation on snow accumulation, with
495 rainfall often enhancing snow and ice melt rates on glacier surfaces. The significant spatial variation in

496 correlation between MEI and SCE for one-month lag, we suspect is caused by the presence of snow
497 cover on glaciers giving negative or no correlation on short term but increasing correlations with time.

498 Reduction in SCE in this study has also been studied in modelling studies based on MERRA
499 satellite data (Mernild et al. 2016b, 2016c) which estimate that snow cover extent in the central Andes
500 has reduced $\sim 1.3\%$ per decade 2000–2014 (linear trend, for the b1 window in Fig. 1) (Mernild et al.
501 2016b). Mernild et al. (2016c) suggest that the largest decreases in snow cover have occurred within
502 the 3,000–5,000 m a.s.l. elevation range, where more than 70 % of seasonal precipitation falls as snow.
503 In comparison, the rate of SCE change observed in this study for the same period and area is
504 significantly higher, with per decade reductions equating to $\sim 2.8\%$. This difference between the results
505 presented here and in Mernild et al. (2016c) may either be an indicator of faster SCE reduction during
506 the 2000–2016 period or highlight possible SCE overestimations in the MERRA model utilized by the
507 latter. For other regions of the world snow cover reductions are well documented. In the Arctic, for
508 example, a general decrease in the amount of snow has been observed between 1999 and 2009, together
509 with reductions in maximum winter snow water equivalent, a later snow-cover onset in autumn and
510 earlier snow-free date in spring, and a decreasing snow-cover duration (Liston and Hiemstra 2011).
511 Indeed a warming trend especially during autumn has been identified at EYE (Burger et al. 2018) and
512 correspond well with the later onset of snow seasons seen in the area (Fig. 6.).

513 To sustain all year runoff, rivers of the central Andes rely on substantial contributions from snow and
514 ice-melt, and river discharge here is strongly linked to snow cover changes (Delbart et al. 2015).
515 Decreases in SCE at the magnitudes shown in this study has the potential to cause a substantial
516 redistribution in seasonal runoff for this region, where $\sim 21\%$ of river runoff originates from snow- and
517 ice- melt (increasing to $\sim 85\%$ during dry summers) (Peña and Nazarala 1987; Mernild et al. 2016a).
518 Glaciers in the central Andes are shrinking and down wasting as a consequence of climate warming and
519 changes in precipitation patterns (Masiokas et al. 2006; Bodin et al. 2010; Gacitua et al. 2015; Malmros
520 et al. 2016). Although initially increasing, ice melt runoff will begin to reduce in the future as lowest
521 elevation land ice disappears. If these ice/snow cover trends continue, runoff conditions will likely
522 change, especially during spring, dry summers and periods of drought, affecting the future
523 sustainability of freshwater resources in areas downstream of the central Andes (Peña and Nazarala
524 1987; Delbart et al. 2015; Saavedra et al. 2016; Carey et al. 2017; López-Moreno et al. 2017). Whether

525 this change in runoff will cause the low lying areas in the catchment to become wetter or drier is
526 largely determined by local topography (Polk et al. 2017, López-Moreno et al. 2017).

527 Directly influencing the surface energy balance, the downward trends in SAL revealed in this
528 study (Fig. 4) may possibly result in positive feedbacks in regards to snow and ice melt. This trend of
529 darkening surfaces, either from reduced snow cover or from enhanced melt conditions, is likely to be
530 reinforced by increasing air temperatures and decreasing precipitation (e.g., Mernild et al. 2016c).
531 Another positive feedback could be initiated by the accumulation of dust and debris on glacier surfaces
532 leading to more energy being absorbed and further melt conditions, especially on lower parts of
533 glaciers (Hansen and Nazarenko 2004; Oerlemans et al. 2009; Arenson et al. 2015). Minimum glacier-
534 wide albedo has shown to be a good predictor for glacier mass balances conditions for temperate
535 glaciers (López-Moreno et al. 2017; Polk et al. 2017), which suggests that glaciers in the study area
536 may have positive mass balances, at least for some of the years analyzed. Decreases in surface albedo
537 have also been observed for many other glaciated parts of the world (Box et al. 2012; Tedesco et al.
538 2013; Abermann et al. 2014; Fausto et al. 2015; Mernild et al. 2015b). The mean albedo for the
539 Greenland ice sheet ablation area (June–August), for example, declined by 22.9 % from 2000 to 2016
540 while dry snow areas only decreased by 1.2 % (Box et al. 2017). Increasing albedo values seen above
541 4.600 m a.s.l. (Fig. 5.) are likely contributed to by the increase in precipitation and the presence of dry
542 snow conditions at high altitudes (Box et al. 2017).

543 The central Andes are dominated by two distinctly different climate systems. On the western
544 side of the Cordillera the climate is influenced by oceanic atmospheric interactions, whereas on the
545 eastern side the climate can be considered continental in type (Prohaska 1976). This difference in
546 climate is highlighted in this study by the relatively weak correlation between SCE/SAL and MEI on
547 the eastern side of the Cordillera compared to the west. A higher inter-annual variability in SCE and
548 SCI and more downward trend on the eastern part may be contributed to continental climate conditions
549 (Fig. 9) as also observed in Saavedra et al. 2017.

550

551 **7. Conclusions and outlook**

552 Overall, snow cover extent (SCE) and snow albedo (SAL) decreased by 13.4 ± 4 % and 7.4 ± 2
553 %, respectively, between 2000 and 2016. SCE showed more downward trends on the eastern side of the
554 Andes Cordillera (13.9 ± 4 %), while SAL showed a uniform decline throughout the area. A seasonal

555 analysis revealed downward trends in SCE and SAL for all months of the year, with the largest
556 decreases occurring during the onset (for SCE and SAL) and at the end of the snow seasons (for SCE)
557 ($> 1\% \text{ yr}^{-1}$). SCE showed a near linear increase with elevation ($r^2=0.96, p < 0.01$), and largest relative
558 losses occurring at elevations above 4.600 m a.s.l. outside glaciated areas. Spatial analysis of the SAL
559 data revealed increasingly downward trends up to $\sim 4,600$ m a.s.l. in elevation. Above $\sim 4,600$ m a.s.l.
560 this trend is reversed, likely because of permanent or semi-permanent dry snow conditions present in
561 glacier accumulation zone. Snow cover duration (SCD) decreased on average by 43 ± 20 days
562 throughout the study area between 2000 and 2016 with largest changes occurring at elevations below
563 4.500 m a.s.l. on the eastern side and 3.500 m a.s.l. on the western side.

564 TIMESAT was unable to extract SCD for glacier areas that were covered with snow for most of
565 the year due to the lack of seasonal variation. Additionally, in situations of large variations in snow
566 conditions occurring over a very short time period the Savitzky-Golay seasonal fitting process applied
567 may introduce some errors. SCD trends for the study area indicate a shortening of the snow season
568 between 2000 and 2016. SCI trends for the included glacial areas were also shown to be downward
569 during this 16-year observation period (these being more pronounced on the eastern side of the
570 Cordillera).

571 The impact of ENSO events, which influence largescale precipitation and temperature patterns
572 in the study area, on the SCE, SCD, and SAL was shown to be evident. Per-pixel analyses revealed that
573 ENSO positively influences SCE/SAL values most strongly with a one-month time-lag. Data available
574 from the EYE meteorological station, between 2000 and 2016, reveals that the monthly SCE and SAL
575 values are primarily determined by variations in temperature, whilst montly SCI values are determined
576 mostly by precipitation. If the observed decline in SCE persist in the coming years, it will likely result
577 in a seasonal redistribution of available downstream freshwater which may cause future problems for
578 people and agriculture in the region.

579

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