

Revisiting extreme precipitation amounts over southern South America and implications for the Patagonian Icefields

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Abstract. Patagonia is thought to be one of the wettest regions on Earth, although available regional precipitation estimates vary considerably. This uncertainty complicates understanding and quantifying the observed environmental changes, such as glacier recession, biodiversity decline in fjord ecosystems and enhanced net primary production. The Patagonian Icefields, for example, are one of the largest contributors to sea-level rise outside the polar regions, and robust hydroclimatic projections are needed to understand and quantify current and future mass changes. The reported projections of precipitation from numerical modelling studies tend to overestimate those from in-situ determinations and the plausibility of these numbers have never been carefully scrutinised, despite the significance of this topic to our understanding of observed environmental changes. Here I use simple physical arguments and a linear model to test the plausibility of the current precipitation estimates and its impact on the Patagonian Icefields. The results show that environmental conditions required to sustain a mean precipitation amount exceeding $6.09 \pm 0.64 \text{ m yr}^{-1}$ are untenable according to the regional moisture flux. The revised precipitation values imply a significant reduction in surface mass balance of the Patagonian Icefields compared to previously reported values. This yields a new perspective on the response of Patagonia's glaciers to climate change and their sea-level contribution and might also help reduce uncertainties in the change of other precipitation-driven environmental phenomena.

1 Introduction

Patagonia's weather and climate are largely shaped by baroclinic eddies, which are characterized by the interaction of the planetary waves with the mean flow (Garreaud, 2009; Garreaud et al., 2013; Schneider et al., 2003; Vallis et al., 2014). The same mesoscale eddies efficiently transfer water vapor from the tropics poleward (Langhamer et al., 2018; Schneider et al., 2010; Trenberth et al., 2005), and regularly (every 9-12 days) trigger narrow filaments of water-vapor-rich bursts called atmospheric rivers. These features temporarily increase the vertical integrated water vapor content (IWV) in the Southern Hemisphere mid-latitudes by more than 200% (Durre et al., 2006; Waliser and Guan, 2017). More than half of all extreme precipitation events (above the 98th percentile) in Patagonia are associated with land-falling atmospheric rivers (Waliser and Guan, 2017). Given the tight coupling between atmospheric moisture transport and hydroclimatic response, changes in moisture transport mechanisms not only dominate the inter-annual and multi-decadal precipitation variability in Patagonia (Aravena & Luckman, 2009; Garreaud, 2007; Garreaud & Muñoz, 2005; Muñoz & Garreaud, 2005; Schneider & Gies, 2004;

Viale & Garreaud, 2015; Weidemann et al., 2013; Weidemann et al., 2018), but also dictate the fate of the ice masses in this region.

The Andes constitute an effective barrier to the impinging moist tropospheric air masses, forming one of the most extreme climatic divides found worldwide (Barrett et al., 2009; Garreaud, 2009; Garreaud et al., 2013; Rasmussen et al., 2007; Smith & Evans, 2007). The strong orographic influence on the precipitation distribution is evident from both remote sensing (Wentz et al., 1998) and terrestrial observations (Fig. 1). Despite observational uncertainty along the coast, two characteristic precipitation regions are apparent: (i) a maritime pre-cordillera region with annual precipitation exceeding 2-3 m w.e. (water equivalent), and (ii) a semi-arid rain-shadow region (< 0.5 m w.e.) east of the main ridge that extends several thousand kilometres towards the South Atlantic. However, little is known about precipitation along the main ridge and, in particular, on the Patagonian Icefields. Current estimates from firn cores (Schwikowski et al., 2006; Shiraiwa et al., 2002), discharge measurements (Escobar, 1992) and numerical modelling (Bravo et al., 2019; Lenaerts et al., 2014; Mernild et al., 2017; Schaefer et al., 2013, 2015; Weidemann et al., 2018a) suggest average annual precipitation rates of 5 to 8 m w.e. yr⁻¹, and of 7 to >10 m w.e. yr⁻¹ for the Northern and Southern Patagonian Icefield (NPI, SPI), respectively (see Table 1). Extreme precipitation rates between 15 m w.e. yr⁻¹ (Mernild et al., 2017; Schaefer et al., 2013, 2015; Schwikowski et al., 2006) and 30 m w.e. yr⁻¹ are suspected at isolated locations (Lenaerts et al., 2014). If these precipitation magnitudes are realistic, it is likely that the SPI is one of the wettest – if not *the* wettest – places on earth.

The considerable uncertainty in precipitation amounts in Patagonia not only affects our current understanding of the local hydrological cycle, but also has profound impacts on studies concerned with fjord ecosystems (Landaeta et al., 2012), biological production in water columns (Aracena et al., 2011; Vargas et al., 2018), net primary production (Jobbágy et al., 2002), glacier mass balance (Escobar, 1992; Foresta et al., 2018; Lenaerts et al., 2014; Mernild et al., 2017; Schaefer et al., 2013, 2015; Schwikowski et al., 2006; Shiraiwa et al., 2002; Willis et al., 2012) and its contribution to sea level rise (Malz et al., 2018; Marzeion et al., 2012; Rignot et al., 2003). Reducing the plausible range of precipitation rates is a key step towards improved process understanding of such systems and offers new perspectives on future changes.

Here I use simple physical scaling arguments and a linear modelling approach to test the plausibility of the current precipitation estimates in central Patagonia (45°S-52°S). In particular, I address the question of whether the water vapor flux (WVF) from the tropics to the mid-latitudes by baroclinic eddies can sustain these extreme precipitation estimates. The assessment of the hypothesis relies on three fundamental assumptions: (i) The orographically induced precipitation is proportional to the incoming WVF which acts as the major moisture resource for the precipitation system. This implies that uncertainties in the incoming WVF directly impact the precipitation estimate. (ii) The terrain forced uplift and condensation of moist air masses is assumed to be the dominant precipitation formation process in central Patagonia. (iii) The atmospheric drying ratio (DR) derived from observed isotope data is a valid measure for the cross-mountain fractionation of the WVF. Based on this

assumption, the proposed methods are constrained by the DR to accurately reproduce the fraction of the water vapor flux removed by orographic precipitation.

5 After a description of the methods (Sec. 2), the moisture transport in southern South America is explored in more detail (Sec. 3.1). The chapter begins with the analysis of the available atmospheric soundings and compares them with remote sensing products and reanalysis data. Following this, the knowledge gained from the experiments on local precipitation formation (Sec. 3.2) and its implications for the surface mass balance of the SPI (Sec. 3.3) will be discussed and critically reviewed (Section 3.4). The last section provides a conclusion of the main findings.

2 Methodology

10 DR-scaling

To provide a first assessment of the magnitude of precipitation, mean precipitation is estimated along the western slopes of the Andes (45°-52°S and 73°-76°W) using a simple DR-scaling. The DR in Patagonia, defined as the fraction of the WVF removed by orographic precipitation, is known to be the highest (~0.45-0.5) worldwide (Mayr et al., 2018; Smith and Evans, 2007). The ratio is a characteristic measure for mountain ranges and is independent of the incoming WVF. As the WVF and the DR (here we use 0.45) are known from ERA-Interim data and isotope observations (Dee et al., 2011; Langhamer et al., 2018; Mayr et al., 2018; Smith and Evans, 2007), one can estimate the mean homogeneous (uniform) precipitation amount. To add altitude-dependent precipitation variability, the amount was redistributed mass-consistently by optimizing the vertical precipitation gradient using a Newton-Raphson algorithm (Press et al., 2007). The precipitation at sea level was taken from the Global Precipitation Measurement (GPM) mission offshore of the Chilean coast (~3 m yr⁻¹). The optimization resulted in a vertical precipitation gradient of ~0.053 % m⁻¹, which represents a slightly higher lapse rate than previously reported rate of ~0.05 % m⁻¹ (Schaefer et al., 2013, 2015). This approach converts the entire specified WVF fraction into precipitation regardless of the saturation vapor deficit of the impinging air masses. However, orographic precipitation can only occur when the terrain forced uplift and cooling of air masses lead to water vapor condensation. To take this condition into account, only lower tropospheric (below 950 hPa) air masses are considered with a relative humidity equal to or exceeding 90% (Jarosch et al., 2012; Weidemann et al., 2013). The DR-scaling provides a first-order approximation but neglects heterogeneity and important processes such as airflow dynamics and cloud physics.

Linear orographic precipitation model

To account for these aspects, a set of realistic and extreme ensemble experiments has been designed using a linear orographic precipitation model (OPM), which represents many processes, such as condensation and hydrometeor conversion, using relative simple formulations for airflow dynamics and cloud physics (Garreaud et al., 2016; Jarosch et al., 2012; Smith and Barstad, 2004; Smith and Evans, 2007; Weidemann et al., 2018a). The model builds upon the original formulation of the linear

orographic precipitation model (Smith and Barstad, 2004), including a correction of the WVF downstream (Smith and Evans, 2007) and an optimization to enforce the model towards a given drying ratio. It solves two steady-state advection equations describing the condensation of water vapor by terrain forced uplift and conversion from cloud water to hydrometeors. Mountain wave theory allows for the decay of the vertical velocity caused by tilting mountain waves, and consequently constrains the water vapor condensation rate. Assuming horizontal uniform background flow and properties (e.g. atmospheric stability), the orographic precipitation can be represented by a transfer function

$$\hat{P}(k, l) = \frac{C_w i \sigma \hat{h}(k, l)}{(1 - imH_w)(1 + i\sigma\tau_c)(1 + i\sigma\tau_f)},$$

with the double Fourier transform of the terrain given by

$$\hat{h}(k, l) = (2\pi)^{-2} \iint h(x, y) e^{-i(kx+ly)} dx dy,$$

The spatial pattern of precipitation rate in physical space is obtained from an inverse Fourier transform

$$P(x, y) = \max \left[\left(\iint \hat{P}(k, l) e^{i(kx+ly)} dk dl + P_\infty \right), 0 \right],$$

followed by the truncation of negative values. The vertical wavenumber is defined by

$$m(k, l) = \left[\left(\frac{N_m^2 - \sigma^2}{\sigma^2} \right) (k^2 + l^2) \right]^{0.5}$$

where the intrinsic frequency is $\sigma = Uk + Vl$. The parsimonious model contains five parameters, an uplift sensitivity factor C_w , the moist buoyancy frequency N_m^2 , the water vapor scale height H_w , and the condensation and fallout time scales τ_c and τ_f . The total precipitation field is obtained by adding up the orographic precipitation from the orographic precipitation model and the background precipitation caused by the synoptic-scale uplift. For consistency, the latter one is calculated by removing the orographic component from the ERA-Interim precipitation field (Dee et al., 2011; Jarosch et al., 2012). To enforce the model towards a given drying ratio DR_{def} , the background precipitation is scaled by a constant, so that the calculated DR corresponds to DR_{def} . The model is solved on a 90 m SRTM dataset, resampled at 1 km resolution (Jarvis et al., 2008). The mean horizontal wind velocities (U, V) and the parameters C_w and N_m^2 are calculated from 6-hourly ERA-Interim fields (2010-2016) below the 500 hPa geopotential level off Patagonia's west coast between 48°-52°S and 75°-78°W (Fig. 1, D1) (Smith

and Barstad, 2004). On contrary to most other studies, H_w is directly derived from the incoming WVF using $H_w = WVF/(\rho q_w U)$, where q_w is the total mixing ratio and ρ the air density. The time scales of $\tau_c = \tau_f = 850$ s are fixed for all experiments, which are realistic values for Southern Andes and produce remarkable similar results to numerical models (Garreaud et al., 2016; Smith and Evans, 2007).

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The orographic precipitation model is used to conduct a suite of ensemble experiments with 40 ensemble members. The ensembles members account for the initial condition uncertainty in the wind direction and moisture content by randomly perturbing U (5%), V (5%), and H_w (10%) around its mean value. In the first experiment, it was tested whether a composite of ‘realistic’ atmospheric ambient conditions (derived from the reanalysis data), the observed drying ratio of 0.45 (Mayr et al., 2018) and WVF provides the basis to sustain the precipitation estimates of previous studies. The second experiment delivers an ‘extreme’ scenario by fixing the drying ratio at a higher value of 0.6 and setting the uplift sensitivity factor ($C_w = 0.004$) and moist stability frequency ($N_m = 0.007$) to their 98th percentile values.

Atmospheric Simulations using the Weather Research and Forecast model (WRF)

15 To analyze the influence of nonlinear flow regimes on precipitation patterns, atmospheric simulations were performed with the Weather Research and Forecast (WRF) model, version 3.8.1. The model was configured with three one-way nested domains with a horizontal resolution of 12.5 km, 2.5 km and 500 m which were centered over the Southern Patagonian Icefields. The model configuration and parameterizations used in this study are shown in Table S4. To achieve the required resolution in the inner domain, the standard terrain data was replaced by NASA Shuttle Radar Topography Mission (SRTM) data
20 (<http://www.cgiar-csi.org/data/srtm-90m-digital-elevation-database-v4-1>). Furthermore, the land use classification was updated with the ESA CCI data set (<https://www.esa-landcover-cci.org>). This way the glacier outlines could be improved significantly. The outermost domain was driven at its lateral boundaries by the ERA-Interim reanalysis dataset with a spatial resolution of $0.75^\circ \times 0.75^\circ$ in longitude/latitude and a time interval of 6 hours. With the above setup, individual events were calculated with WRF. Each simulation had a spin-up of at least 12 hours.

25 3 Results and Discussion

3.1 Moisture transport

Observations of IWV and WVF are sparse in South America and limits the analysis of the moisture transport to a few locations (see Fig. 2). The only available soundings for the region are Puerto Montt (41.4347°S , 73.0975°W) on the Pacific coast and Punta Arenas (53.0033°S , 70.8450°W) located at the Strait of Magellan (Durre et al., 2006). Along the coast at the latitude of
30 Puerto Montt, the average WVF in the period 1990-2017 was about $165.52 \pm 48.51 \text{ kg m}^{-1} \text{ s}^{-1}$. Land-falling atmospheric rivers temporarily amplify the WVF by more than $400 \text{ kg m}^{-1} \text{ s}^{-1}$. There is also clear evidence that enhanced atmospheric circulation

during strong El Niño events (Ocean Niño Index >1.5) increase the moisture flux over several months (see Fig. 2). The El Niño signal is less pronounced in Punta Arenas. The atmospheric soundings show opposite linear long-term WVF trends over the period 1990-2016 with a significant ($p<0.08$) decrease of $-4.46 \text{ kg m}^{-1} \text{ s}^{-1}$ (-2.70%) decade⁻¹ in Puerto Montt, and a significant ($p<0.05$) positive trend of $8.79 \text{ kg m}^{-1} \text{ s}^{-1}$ (5.11%) decade⁻¹ in Punta Arenas (see Fig. 2). However, change-point analysis shows that the observed WVF trend in Punta Arenas is not constant over time, but has shown significant abrupt shifts in the past that characterize the transition of water vapor rich and poor periods (Killick et al., 2012). A significant transition took place in 2006 which marks the beginning of a relative water vapor rich period (Fig. 2).

The ERA-Interim data, on which the analysis is based, reflects the interannual WVF variability and overall trend of the soundings but slightly overestimates the rate of change in Puerto Montt ($-4.94 \text{ kg m}^{-1} \text{ s}^{-1}$ decade⁻¹, -4.43% decade⁻¹), and underestimates the observed trend in Punta Arenas ($4.10 \text{ kg m}^{-1} \text{ s}^{-1}$ decade⁻¹, 2.70% decade⁻¹). The mean WVF at both sites is weaker than the observed moisture transport. In Puerto Montt, the WVF is about $111.54 \pm 34.40 \text{ kg m}^{-1} \text{ s}^{-1}$, which is almost 30% less than the estimate from the atmospheric sounding. The differences between observed WVF ($172.12 \pm 54.19 \text{ kg m}^{-1} \text{ s}^{-1}$) and reanalysis data ($152.08 \pm 57.08 \text{ kg m}^{-1} \text{ s}^{-1}$) is much lower in Punta Arenas. It is evident from the soundings that ERA-Interim data is too dry (according to the IWV) in the vicinity of Puerto Montt (-2.23 mm , -14.9% , $p<0.01$), and slightly too wet in the south (0.48 mm , 4.6% , $p<0.05$) (see Fig. S2 and Table S2). The comparison with atmospheric water vapor data obtained by the Special Sensor Microwave Imager/Sounder (SSMIS) over the ocean confirms the north-south pattern (Wentz et al., 1998) (see supporting information Fig. S4). While IWV differences between ERA-Interim data and SSMIS south of 45°S are on average smaller than 0.16 mm (1.1%), larger deficits are apparent north of 45°S ($<-0.8 \text{ mm}$).

Based on the comparison with the atmospheric soundings and SSMIS observation, ERA-Interim underestimates the IWV along the west coast of Patagonia (D1 in Fig. 1), where the corresponding parameters for the assessment were calculated, by less than 5%. However, comparison with the soundings suggests that the WVF in the ERA-Interim data along the west coast is weaker by 10-20% due to uncertainties in moisture advection. In the following analysis, a WVF bias of 10% is assumed and corrected accordingly.

3.2 Physical constraints on local precipitation

To obtain the plausible range of precipitation amounts in central Patagonia, the DR-scaling and the linear model are driven by the ERA-Interim data for the period 2010-2016. The DR-scaling results in a mean precipitation rate of $3.45 \pm 0.14 \text{ m yr}^{-1}$ in the Pre-Cordillera region, with maximum values of $4.89 \pm 0.97 \text{ m yr}^{-1}$ (see Fig. 1). Averaged over the SPI and NPI, this approach produces values of $5.46 \pm 1.30 \text{ m yr}^{-1}$ and $5.38 \pm 1.26 \text{ m yr}^{-1}$, respectively (Table 1). The highest peaks on the icefields receive precipitation amounts of up to $9.63 \pm 3.69 \text{ m yr}^{-1}$.

The orographic precipitation model is applied to a large domain (Fig. 1, D2) to avoid spurious numerical artefacts. The ensemble mean of the realistic experiment (DR=0.45) gives an average precipitation amount of $5.06 \pm 0.51 \text{ m yr}^{-1}$ over the SPI

and $5.38 \pm 0.59 \text{ m yr}^{-1}$ over the NPI (Table 1, Fig. 3), indicating that the WVF can sustain relatively high mean precipitation amounts in Patagonia. However, precipitation estimates are up to 38% lower than estimates from previous numerical studies (Escobar, 1992; Lenaerts et al., 2014; Mernild et al., 2017; Schaefer et al., 2013, 2015; Schwikowski et al., 2006). The highest mean amounts are found on the western slopes of the icefields ($5.93 \pm 0.60 \text{ m yr}^{-1}$) and at the southernmost end of the SPI. The eastern slopes receive considerably less precipitation ($4.04 \pm 0.42 \text{ m yr}^{-1}$). Comparison with in-situ observations from the Dirección General de Aguas (DGA, Chile) indicates that the model slightly overestimates precipitation on the leeward side by $0.29 \pm 0.37 \text{ m yr}^{-1}$ (see Table S3). Higher deviations ($1.07 \pm 1.30 \text{ m yr}^{-1}$) occur at the stations on the west side which are located at the foot of the Patagonian Icefields. The overestimation is the result of the rapid increase in model terrain elevation and the absence of nonlinear processes in the linear model (see Sec. 3.4). The maximum precipitation amount ($10.09 \pm 0.92 \text{ m yr}^{-1}$), found on the SPI plateau, is ~30-70% lower than previously simulated extremes (Lenaerts et al., 2014; Schaefer et al., 2013, 2015) and accumulation rates derived from an ice core (Shiraiwa et al., 2002). The large ensemble spread indicates that extreme precipitation is very sensitive to small uncertainties in ambient flow conditions (see Table 1). Even though the uncertainty in the background flow regime and dynamics may also be a possible origin of the extreme precipitation predicted by the mesoscale models, the responsible mechanisms explaining the significant differences remain unclear. It is likely that one reason is the model parameterization of processes. Some microphysical parameterization schemes are more ‘graupel-friendly’ than others, which can lead to strong hydrometeor formation. Since the choice of parameterization combinations can lead to very different results, each model must be examined individually. The sources are manifold and can only be speculative in the context of this study.

The extreme experiment shows higher averaged precipitation amounts of $5.99 \pm 0.59 \text{ m yr}^{-1}$ and $6.09 \pm 0.64 \text{ m yr}^{-1}$ at the SPI and NPI, respectively (Fig. 3). The combination of short time scales, large drying ratio, strong moist stability frequency, and large uplift sensitivity factor increases the total precipitation and enhances the cross-mountain fractionation. Despite the precipitation-enhancing parameter choices, the maximum precipitation ($11.58 \pm 0.98 \text{ m yr}^{-1}$) represents a reduction of up to 60% compared to other numerical studies (Lenaerts et al., 2014; Schaefer et al., 2013, 2015). In addition, the estimated maximum is presumably an overestimate itself, due to the ‘extreme’ parameter choice and to the exclusion of nonlinear effects, such as flow blocking (see Sec. 3.4), given the linear nature of the orographic model.

3.3 Consequences of revised precipitation estimates on the surface mass balance of the SPI

These revised precipitation estimates have critical implications for our current understanding of the response of Patagonia’s glaciers to climate change. Recent numerical studies (Mernild et al., 2017; Schaefer et al., 2015) suggest a mean annual surface mass gain of $1.78 \pm 0.36 \text{ m}$ to $2.24 \text{ m w.e. yr}^{-1}$ for the SPI over recent decades, while surface mass balance (SMB) estimates for the NPI range between $-0.16 \pm 0.73 \text{ m w.e. yr}^{-1}$ and $0.14 \pm 0.49 \text{ m w.e. yr}^{-1}$. However, these assessments used mean precipitation rates well above (40-65%) the plausible range presented in this study.

To quantify the effect of the revised precipitation values on the SMB, we use the significant linear relation ($R^2=0.96$, $p<0.05$) between precipitation sum and annual SMB derived from Schaefer et al. (2015) (see Fig. 4), given by $SMB=1.2588 \cdot P_s - 3.9355$, where P_s is the mean solid precipitation. Taking into account the proposed solid to total precipitation ratio of 0.596 (Schaefer et al., 2015), the mean solid precipitation is 3.02 ± 0.30 m w.e. (realistic scenario) and 3.57 ± 0.35 m. w.e. (extreme scenario) for the SPI. Based on this assumption, the revised accumulation values would result in a mean SMB between 0.56 ± 0.45 m w.e. yr^{-1} (7.82 ± 6.28 km^3 yr^{-1} , extreme scenario) and -0.14 ± 0.39 m w.e. yr^{-1} (-1.95 ± 5.45 km^3 yr^{-1} , realistic scenario) on the SPI (Fig. 4). The SMB estimate from the DR-scaling is within these limits. Taking account of the recent geodetic mass balance observations (-0.941 ± 0.19 m w.e.) (Malz et al., 2018), the mean mass loss due to calving ranges between -1.5 ± 0.64 (-20.95 ± 8.94 km^3 yr^{-1}) and -0.8 ± 0.58 m. w.e. yr^{-1} (-11.18 ± 8.10 km^3 yr^{-1}). The mean mass balance and calving flows derived here are subject to approach-related uncertainties and may deviate strongly from the values of individual years. A recently published study showed that calving fluxes at Jorge Montt Glacier fluctuated between 1.16 ± 0.66 km^3 yr^{-1} and 3.81 ± 1.10 km^3 yr^{-1} in the years 2012-2018 (Bown et al., 2019). Single events cannot be represented with the approach presented here, since the mean SMB is used together with the geodetic mass balance observations which also constitutes an integrated value. Furthermore, an invariant and homogeneous liquid to solid precipitation ratio and a universal relationship between annual precipitation sums and SMB is assumed. Recently published studies indicate that the solid to liquid precipitation ratio vary locally (Bravo et al., 2019). Together with the snowdrift effect, which is not considered here, this leads to large uncertainties in the mass change estimates. However, this analysis clearly shows how sensitive the estimation of SMB and calving rates react to precipitation uncertainties.

Given the strong link between glacier SMB and the local hydrological cycle, the long-term SMB evolution scales with the strength of the WVF, which is, in turn projected to increase in a warming climate. The WVF sensitivity along Patagonia's west coast ($\sim 50^\circ S$) is on the order of $\sim 15\% K^{-1}$ ($\sim 3\% decade^{-1}$) as a result of the strengthening of the westerlies ($\sim 20\% K^{-1}$) and increase in IWV ($\sim 5\% K^{-1}$) south of $45^\circ S$. The latter is weaker than the change in global-mean IWV which scales according to the Clausius-Clapeyron relation ($7\% K^{-1}$) but is consistent with the assumption that increased latent heat flux is compensated by the sensible heat flux (Held et al., 2006; Schneider et al., 2010). The observed zonal wind trend is associated with a bias towards a more positive Southern Annular Mode (Garreaud et al., 2013; Marshall et al., 2017; Thompson & Solomon, 2002). The change of the WVF leads to stronger moisture flux convergence along the coastal zone west of the Andes main ridge. Ignoring the fact that the solid-liquid ratio changes, which appears to be a reasonable assumption since temperature changes in the lower troposphere are negligible (~ 0.01 K dec^{-1}), a mean mass gain of 0.57 ± 0.06 m w.e. per degree warming (0.11 ± 0.02 m w.e. $decade^{-1}$) is expected over the SPI. This rate is consistent with other studies (Mernild et al., 2017). Thus, although the precipitation values presented here indicate that present-day SMB of the SPI is likely not as positive as suggested by previous studies, SMB can be expected to show an increasing trend under continued warming conditions.

3.4 Limitations and nonlinearities

Given the linear nature of the approach used, the knowledge gained must be critically assessed and is only valid under certain conditions. This linear assumption requires a stably stratified atmospheric flow, more precisely given by a positive moist buoyancy frequency. During the study period from 2010 to 2016, the condition was fulfilled in more than 99% of all days. As a part of this assumption a linear mountain flow response is required, to guarantee that the airflow crosses the mountain range. To ensure a linear flow regime, the non-dimensional mountain height $\hat{H} = (N_m h_m)/U$ must be smaller than one, where h_m is the mean barrier height. Assuming a mean $h_m = 2200$ m, the conditions ($\hat{H} < 1$) is fulfilled in >82% of all considered cases (see Fig. S5). In the remaining cases ($\hat{H} \geq 1$), the Andes block the atmospheric flow, and a northerly low-level barrier jet forms along the west slope, parallel to the main ridge (Barrett et al., 2009; Falvey and Garreaud, 2007; Garreaud and Muñoz, 2005; Viale and Garreaud, 2015) (see Fig. 5). The low-level jet constitutes an effective barrier to the flow that extends upwind, greatly reducing the uplift motions and thus the condensation of water vapor along the west slopes. The shift in the vertical uplift enhances precipitation upstream of the Andes, while reducing precipitation at the slopes. The effect of blocking is clearly evident in the precipitation fields of high-resolution (500 m) atmospheric simulations of single events using the Weather Research and Forecast (WRF) model (see Fig. 6 and Table S4). Two water-vapor-rich events were chosen to illustrate the influence of the flow regime on the spatial distribution of precipitation. While the linear flow regime has a pronounced precipitation maximum on the slopes, flow blocking shifts the precipitation far upstream (600-700 km) leading to a more homogeneous pattern.

Upstream precipitation can be further enhanced by microphysical processes such as the seeder-feeder mechanism and rapid warm air autoconversion. Studies have shown that these processes can lead to higher rain accumulations upstream when fronts and embedded atmospheric rivers intersect the west coast of central Chile (Garreaud et al., 2016; Massmann et al., 2017; Viale et al., 2013; Viale and Garreaud, 2015). The lifting of moist air masses upstream produces mid-tropospheric stratiform clouds (seeder) which can be strong enough to produce snow/graupel aloft and light precipitation in the pre-frontal region. If the frontal system is slowed down by blocking, low-level convergence enhances in the area of the narrow cold-frontal rainband and fuels the updrafts. The enhanced updrafts facilitate the development of low-level clouds by collision-coalescence between supercooled droplets. When the narrow cold frontal rainband propagates further east it triggers the seeder-feeder mechanism and low-tropospheric clouds are seeded by the precipitation that is formed by mid-tropospheric clouds aloft. The associated rapid transformation of cloud water into hydrometeors and increased hydrometeor sizes are absent in the approach presented. Here, the process is treated simplistic by the choice of small time scales and by constraining the synoptic-scale uplift (background precipitation). This solution most likely lead to (i) an overestimation of precipitation on the west slopes of the SPI, (ii) an underestimation of precipitation in the Pre-Cordillera zone, but (iii) satisfies the given DR_{def} constraint. Compliance with the DR criterion is the necessary condition to verify the plausibility of precipitation estimates.

4 Conclusion

The present study has shown on the basis of simple physical arguments and a linear model that it is very unlikely that the moisture flux from the Pacific will be sufficient to sustain the reported extreme mean precipitation amounts for Patagonia. While the approaches and assumptions employed in this study contain substantial uncertainties, precipitation estimates using other parameter combinations fall within the range between the realistic and the extreme scenario. Hence, this study offers a plausible range of precipitation estimates based on clearly defined assumptions: (i) the orographically induced precipitation is proportional to the incoming WVF, (ii) the terrain forced uplift and condensation of moist air masses is assumed to be the dominant precipitation formation process in central Patagonia, and (iii) the atmospheric drying ratio (DR) derived from observed isotope data is a valid measure for the cross-mountain fractionation of the WVF. The values within this range are about 40-65% lower than previously assumed. Extreme precipitation in wind-exposed regions is in the range of $11.58 \pm 0.98 \text{ m yr}^{-1}$, up to 60% lower than estimated by other numerical studies (Lenaerts et al., 2014; Schaefer et al., 2013, 2015). It should also be noted that processes such as snowdrift and nonlinear effects have not been taken into account so that the actual accumulation rates are probably still below these estimates. This result makes it very unlikely that Patagonia is the wettest place on Earth. More importantly, the drier hydroclimatic condition represents a major constraint for the Patagonian Icefields and reduces the precipitation contribution to the glacier mass balance. The missing contribution is evident in the surface mass balance. According to the results, the average SMB was between $0.56 \pm 0.45 \text{ m w.e. yr}^{-1}$ ($7.82 \pm 6.28 \text{ km}^3 \text{ yr}^{-1}$) and $-0.14 \pm 0.39 \text{ m w.e. yr}^{-1}$ ($-1.95 \pm 5.45 \text{ km}^3 \text{ yr}^{-1}$) in the last decades. On the long-term, the regional precipitation is likely to increase by $\sim 15\%$ per degree warming in response to stronger moisture flux. Assuming that the liquid to solid precipitation ratio and the relationship between annual precipitation sum and SMB are universal and valid for the next decades, the WVF changes would result in a glacier surface mass gain of about $0.57 \pm 0.06 \text{ m w.e. per degree warming}$. This positive trend contradicts the recently published geodetic mass balance observations (Malz et al., 2018), which detected quick glacier recessions in these regions. The observed retreat is significantly stronger than the gain in ice mass implying that the ice mass budget is partially decoupled from the climate signal and primarily caused by dynamic adjustments of tidewater and lake calving glaciers. The pronounced dynamic glacier response emphasizes that ice dynamic processes need to be given more prominence in order to quantify the response of the Patagonian glaciers to climate change and their contribution to future sea-level rise. While the change in ice masses is a vivid example of the response to reduced precipitation, it also opens new perspectives for future studies on environmental change in Patagonia and can also help reduce uncertainties in the quantification of other precipitation-driven environmental phenomena.

Acknowledgments, Samples, and Data

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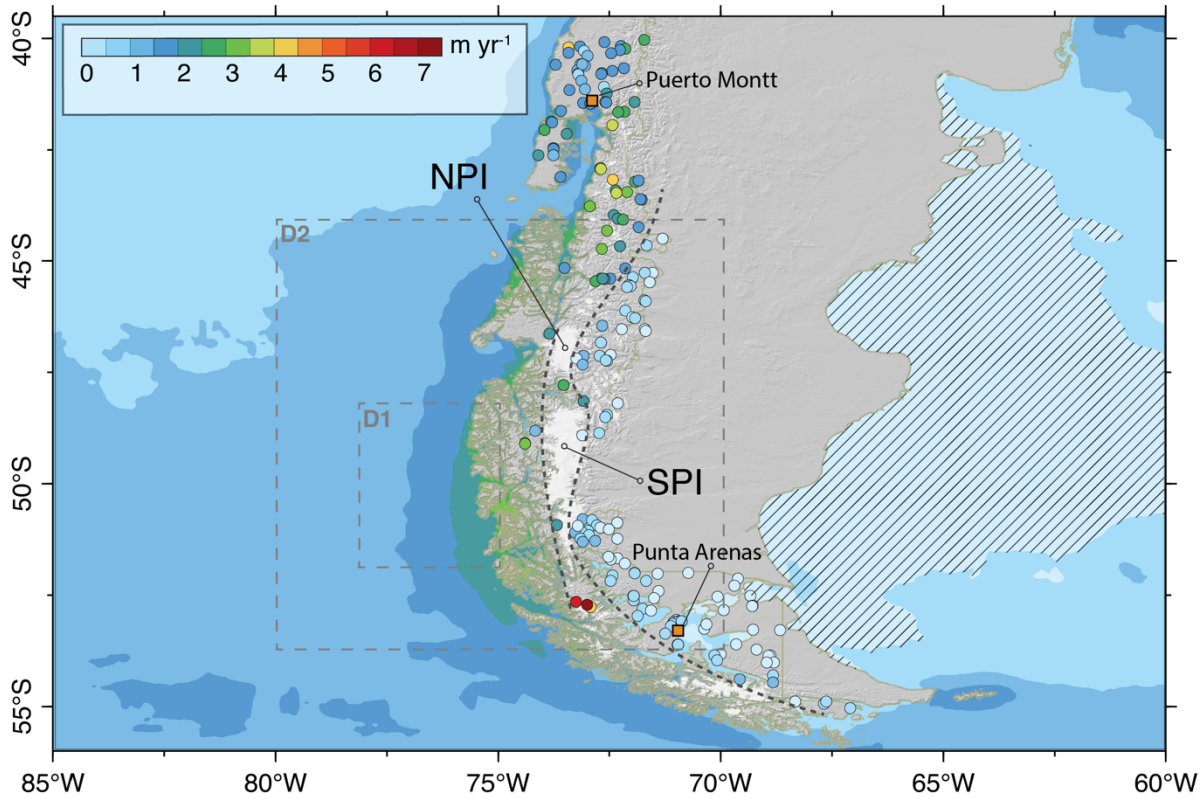
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5 **Figure 1: Precipitation climatology in southern South America. The filled circles indicate precipitation amounts measured by the observational network, established by the Dirección Meteorológica de Chile (DMC), Dirección General de Aguas (DGA), and own weather stations (see Table S1). The colour shaded areas over the ocean shows the rainfall distribution based on the Global Precipitation Measurement (GPM) satellite mission. Black dashed lines roughly delineate the maritime Pre-Cordillera range, Andes main ridge, and the semi-arid Pampa region. Also indicated are the Northern (NPI) and Southern Patagonian Icefields (SPI). The dashed area shows the semi-arid rain-shadow region. Also shown are the simulation (D2) and forcing (D1) domains.**

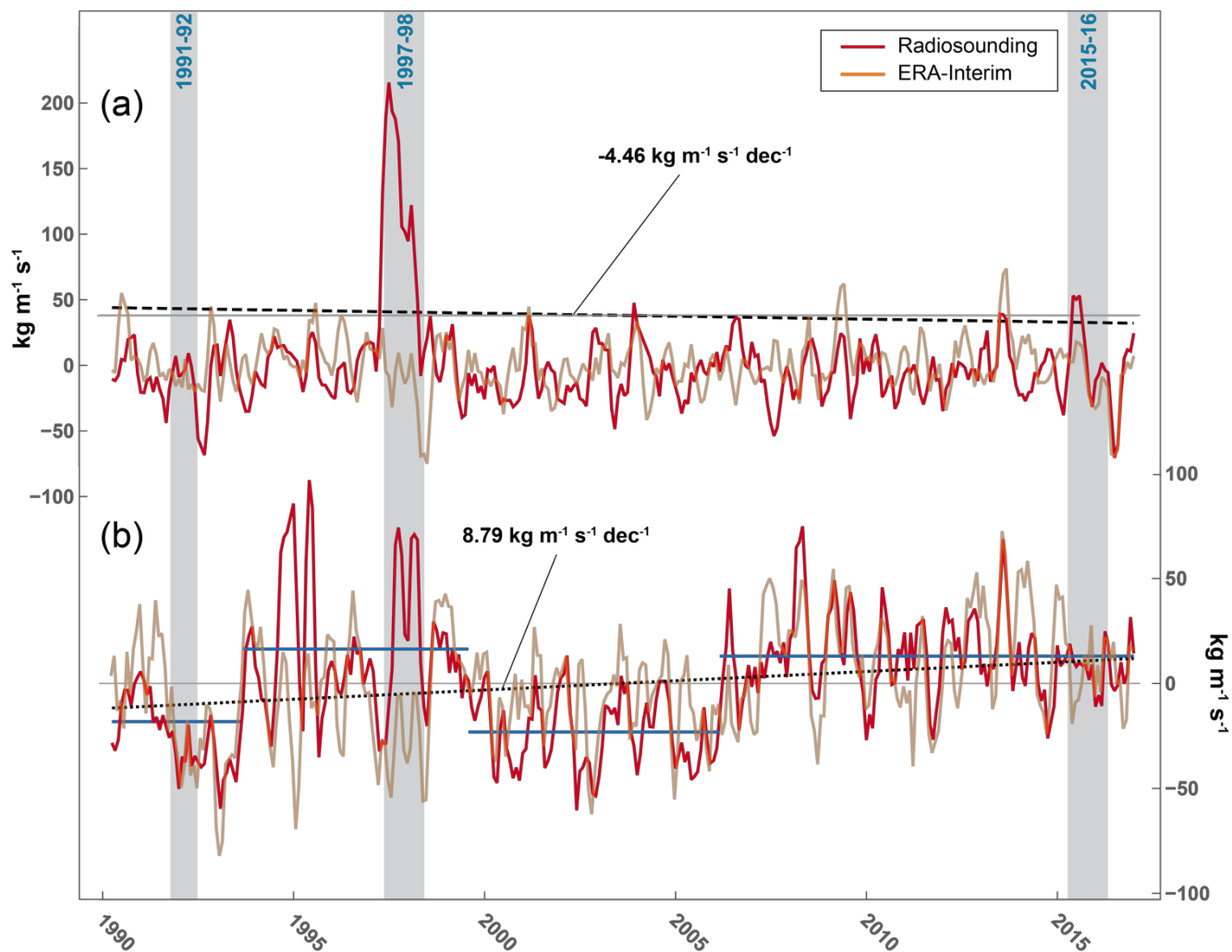


Figure 2: Monthly WVf anomalies in Puerto Montt (panel a) and Punta Arenas (panel b). Shown are the running 3-month mean WVf anomalies for the atmospheric soundings and the nearest ERA-Interim grid point from 1990-2016. The blue shaded areas indicate very strong El Niño events (ONI>1.5). The horizontal blue lines in panel b show the mean WVf over water vapor rich and poor phases.

5

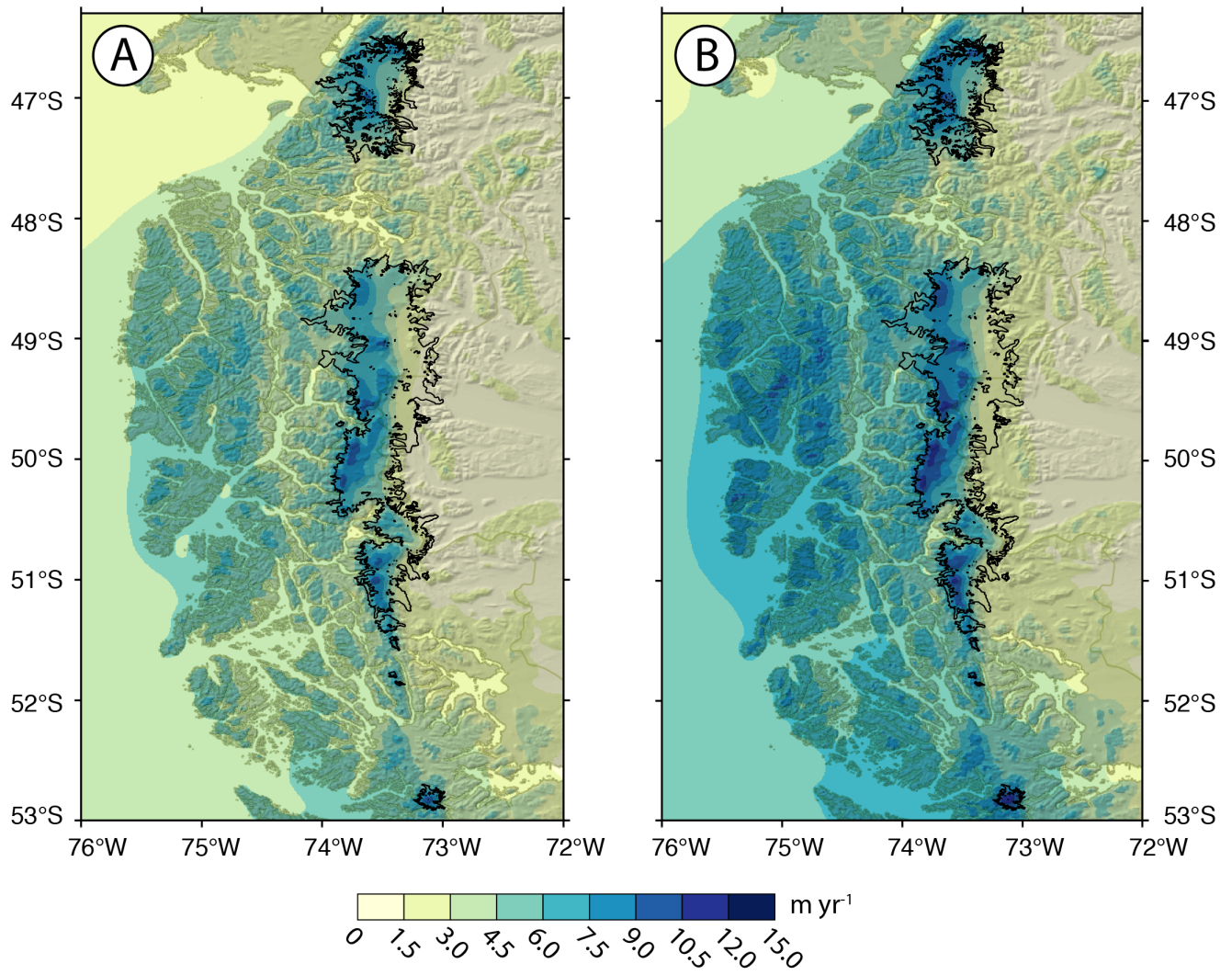
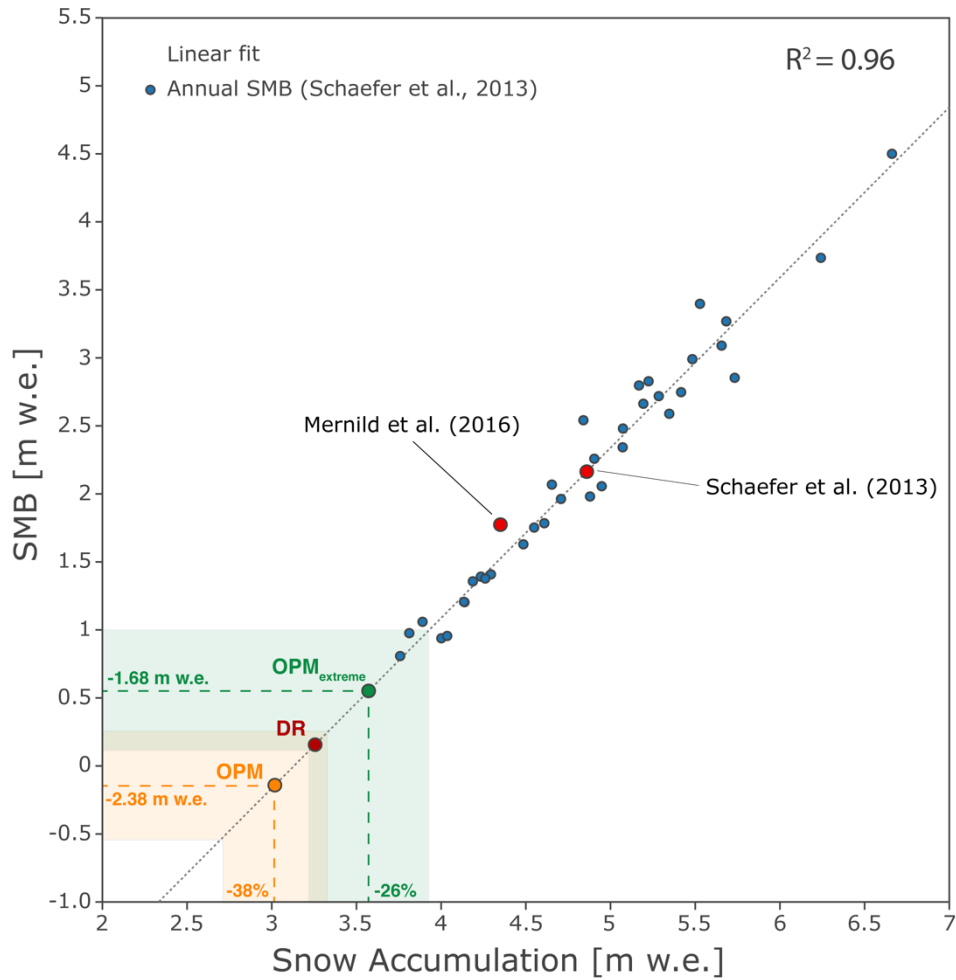
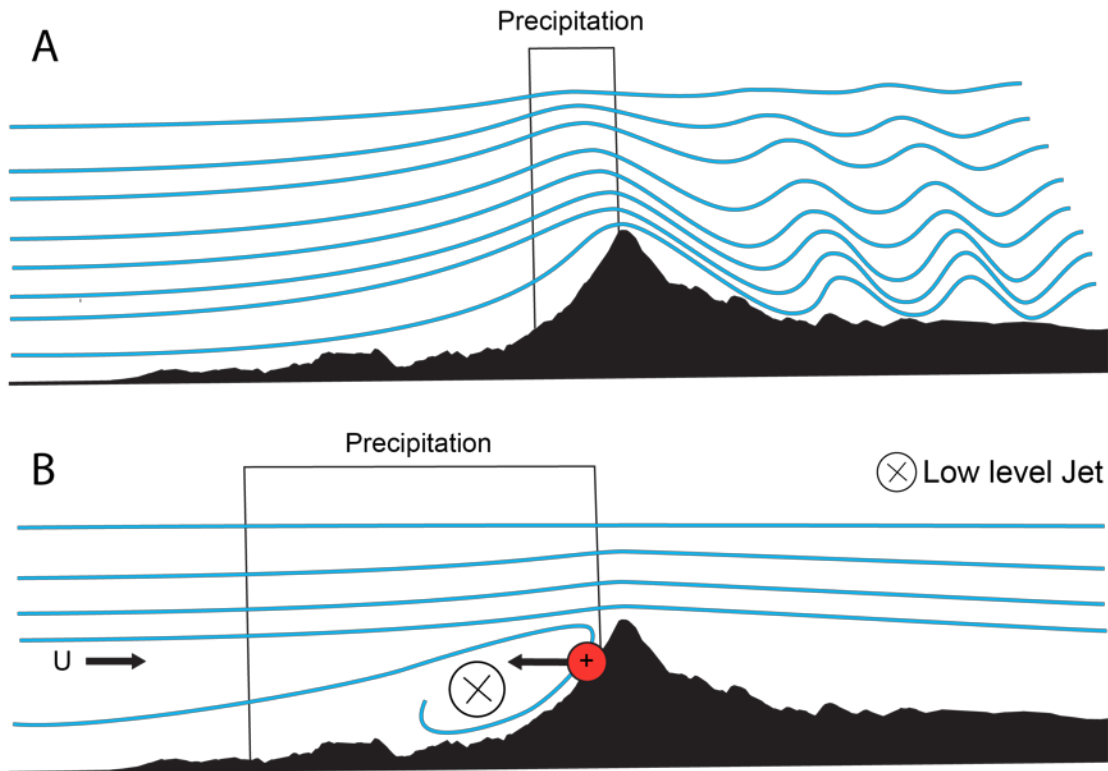


Figure 3: Results of the OPM ensemble experiments. Mean precipitation fields simulated by the OPM using (A) the ‘realistic’ (DR=0.45) parameter setup, and (B) the ‘extreme’ parameter setup using a DR of 0.6 and the 98th percentile values for the uplift sensitivity factor ($C_w=0.004$) and moist stability frequency ($N_m=0.007$).



5 **Figure 4: Relation between the annual specific accumulation and surface mass balance over the SPI from 1975-2000. The red and blue dots mark the multi-annual mean values using the accumulation rates from Schaefer et al. (2013) and Mernild et al. (2016), respectively. The values obtained by this study are represented by the green, orange and dark red dots. The shaded areas mark lower and upper snow accumulation and SMB bounds derived from the DR-scaling (DR), the realistic (OPM) and extreme ($OPM_{extreme}$) orographic precipitation model experiments. The percentage and numbers show the differences between the current study and the values given by Schaefer et al..**



5 **Figure 5: Schematic illustration of the interaction between the atmospheric air flow and the Andes. (A) Linear mountain flow response ($H < 1$) leads to strong uplift and precipitation along the west slopes. (B) The air flow is blocked by the topography ($H \geq 1$) and the resulting pressure gradient (indicated by the red circle) at the west slope slows down the upstream flow. The imbalance between the large-scale pressure gradient and Coriolis-force leads to a northerly low-level jet, which reduces and shifts the uplift motions upstream. This mechanism enhances precipitation in the Pre-Cordillera range, while reducing precipitation at the west slopes of the Andes.**

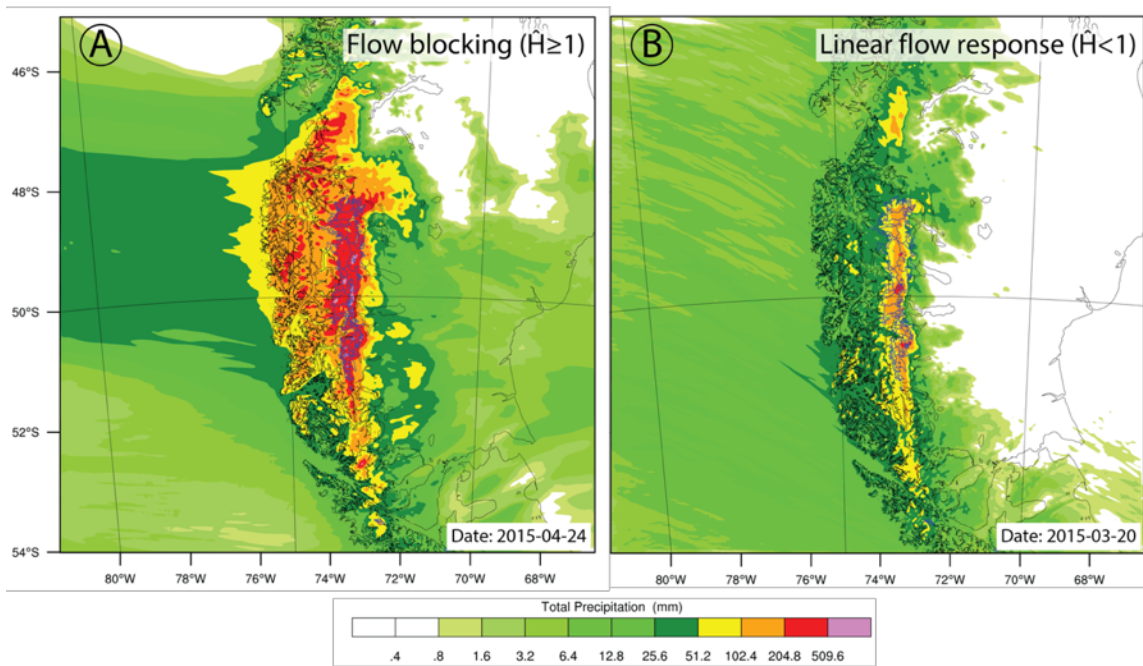


Figure 6: Total precipitation sums (3-days) over the SPI and NPI from WRF for different flow regimes. (A) Nonlinear flow response with enhanced precipitation in the Pre-Cordillera range and (B) linear flow response with strong localized precipitation along the west slopes of the Andes.

5

Table 1: Comparison of mean precipitation estimates on the SPI and NPI averaged over the entire Icefield and the western and eastern slopes. Values are given in m w.e. yr⁻¹. The local maximum values, if available, are shown in parentheses.

	SPI			NPI			Periode
	Mean	West	East	Mean	West	East	
Realistic scenario (mean)	5.06±0.51 (10.09±0.92)	5.93±0.60 (10.09±0.92)	4.06±0.42 (9.92±0.95)	5.38±0.59 (9.43±0.93)	5.83±0.64 (9.43±0.93)	4.37±0.48 (9.30±0.92)	2010-2016
> 3000 m	8.03±0.81	8.60±0.85	8.10±0.82	7.16±0.79	7.40±0.78	6.66±0.73	
2500–3000 m	6.37±0.65	6.93±0.70	6.13±0.63	6.58±0.67	6.84±0.70	5.58±0.53	
2000-2500 m	5.39±0.54	5.70±0.58	4.93±0.50	5.69±0.58	6.20±0.63	5.10±0.52	
1000-2000 m	5.29±0.54	6.13±0.62	4.26±0.44	5.58±0.62	5.77±0.64	4.81±0.53	
< 1000 m	4.26±0.44	5.43±0.56	3.04±0.32	4.81±0.54	5.77±0.64	3.05±0.35	
Extreme scenario (mean)	5.99±0.59 (11.58±0.98)	7.02±0.68 (11.58±0.98)	4.80±0.49 (11.39±0.99)	6.09±0.64 (10.37±0.96)	6.60±0.69 (10.37±0.96)	4.90±0.53 (10.12±0.95)	2010-2016
> 3000 m	8.89±0.89	9.56±0.94	8.94±0.90	7.67±0.85	7.93±0.85	7.07±0.79	
2500–3000 m	7.09±0.73	7.73±0.78	6.81±0.71	7.05±0.73	7.35±0.75	5.92±0.59	
2000-2500 m	6.08±0.61	6.46±0.65	5.55±0.57	6.16±0.64	6.75±0.69	5.48±0.57	
1000-2000 m	6.19±0.61	7.17±0.70	5.00±0.52	6.21±0.67	6.45±0.70	5.30±0.58	
< 1000 m	5.34±0.53	6.77±0.66	3.84±0.39	5.74±0.58	6.84±0.68	3.72±0.40	
DR-scaling (mean)	5.46±1.30 (8.99±3.33)			5.38±1.26 (9.63±3.69)			2010-2016
Other studies							
Schaefer et al. (2015)	8.36 (>20.0)			8.03±0.37 (>15.0)			1975-2011
Mernild et al. (2016)	8.13±0.32 (>15.0)			6.95±0.34 (>15.0)			1979-2014
Lenaerts et al. (2013)	- (>30.0)			- (>30.0)			1979-2012
Escobar et al. (1992)	7.0			6.7 (over the broad plateau)			1960-1980

