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Future sea-level rise from tidewater glaciers and ice-shelf tributary glaciers of the Antarctic Peninsula

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Abstract

Iceberg calving and increased ice discharge from ice-shelf tributary glaciers contribute significant amounts to global sea-level rise (SLR) from the Antarctic Peninsula (AP). Owing to ongoing ice dynamical changes (collapse of buttressing ice shelves), these contributions have accelerated in recent years. As the AP is one of the fastest warming regions on Earth, further ice dynamical adjustment (increased ice discharge) is expected over the next two centuries. In this paper, the first regional SLR projection of the AP from both iceberg calving and increased ice discharge from ice-shelf tributary glaciers in response to ice-shelf collapse is presented. An ice-sheet model forced by temperature output from 13 global climate models (GCMs), in response to the high greenhouse gas emission scenario (RCP8.5), projects AP contribution to SLR of 28 ± 16 to 32 ± 16 mm by 2300 that is roughly split between tidewater glaciers and ice-shelf tributary glaciers. In the RCP4.5 scenario, sea-level rise projections to 2300 are dominated by tidewater glaciers (~8-18 mm). In this

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cooler scenario, 2.4 ± 1 mm is added to global sea levels from ice-shelf tributary drainage basins as fewer ice-shelves are projected to collapse. Sea-level projections from ice-shelf tributary glaciers are dominated by drainage basins feeding George VI Ice Shelf, accounting for ~70% of simulated SLR. Combined total ice dynamical SLR projections to 2300 from the AP vary between 11 ± 2 and 32 ± 16 mm sea-level equivalent (SLE), depending on the emission scenario used. These simulations suggest that omission of tidewater glaciers could lead to a substantial underestimation of the ice-sheet's contribution to regional SLR.

Keywords: Ice dynamics, Sea-level rise, Tidewater glaciers, Ice-shelf collapse, Ice-shelf tributary glaciers

1 1. Introduction

The Antarctic Peninsula (AP) is a mountainous and heavily glaciated 2 region, dominated by glaciers flowing directly into the sea (henceforth tide-3 water glaciers) and into floating ice-shelves (henceforth ice-shelf tributary 4 glaciers). In response to the rapid warming experienced by this region over 5 the last 50 years (Vaughan et al., 2003), glaciers have contributed at an ac-6 celerated rate to global sea-level rise (SLR) in recent years (*Cook et al.*, 2005; 7 Wouters et al., 2015). In addition to an increase in near-surface air temper-8 atures, surface waters of the surrounding ocean have warmed (Meredith and 9 King, 2005). This ocean warming has been accompanied by an acceleration 10 (Pritchard and Vaughan, 2007) and retreat (Cook et al., 2005) of tidewater 11 glaciers, leading to increased ice discharge to the ocean. 12

¹³ Climatological changes have also affected ice-shelf tributary glaciers. Un-

like tidewater glaciers, ice-shelf tributary glaciers do not flow directly into 14 the ocean, but into a floating ice-shelf. This extension of the grounded ice 15 exerts backstress (buttressing force) on the grounded glacier upstream and 16 thus restrains ice flow. If this buttressing force is reduced or removed, the 17 grounded ice upstream will speed up, thin and discharge more ice into the 18 ocean. This behaviour has been observed at several locations in the AP re-19 gion (Rott et al., 2002; Scambos et al., 2004; Rignot et al., 2004). Glaciers 20 draining into the Prince-Gustav-Channel and Larsen A embayments are still 21 adjusting to ice-shelf removal, some 20 years after ice-shelf collapse (Rott 22 et al., 2014; Scambos et al., 2014), and are contributing a significant portion 23 to the region's SLR (*McMillan et al.*, 2014). 24

Abrupt ice-shelf collapse events in the past have been linked to a combi-25 nation of atmospheric warming (Vauqhan and Doake, 1996; Scambos et al., 26 2000) and increased basal melting (Pritchard et al., 2012; Holland et al., 27 2015). Ice-shelves are thought to be structurally weakened prior to collapse 28 by i) hydrofracture of surface crevasses, and ii) basal melting at the ice-ocean 29 interface. In the latter process, warm ocean water erodes the underside of 30 the ice-shelf, thinning it and thus leaving the ice-shelf more vulnerable to 31 the process of hydrofracturing (Shepherd et al., 2003). Hydrofracture of sur-32 face crevasses occurs primarily when sufficient meltwater is available at the 33 surface of the ice-shelf and can wedge open crevasses to cause catastrophic 34 ice-shelf disintegration (Scambos et al., 2004). Recent studies suggest that 35 other ice-shelf weakening processes such as fracturing and weakening of shear 36 margins may also be important and lead to a progressive weakening of the 37 ice-shelf prior to disintegration (Khazendar et al., 2015; Borstad et al., 2016).

A prime example of this is the progressive mechanical weakening of remnant
Larsen B Ice Shelf over the last 15 years (*Borstad et al.*, 2016). The importance of these processes may however vary for individual ice-shelves.

While projections of the surface mass balance are forecasted to provide a 42 negative contribution to sea level, this is expected to be offset by sea-level 43 rise contributions from ice dynamical changes (Barrand et al., 2013a). Owing 44 to their short response times to ice dynamical perturbations, e.g. ice-shelf 45 removal, in comparison to the rest of the Antarctic Ice Sheet (Barrand et al., 46 2013a), AP glaciers are projected to play an important role in the global 47 SLR budget over the next century (Barrand et al., 2013a; Schannwell et al., 48 2015). Hitherto, ice-sheet modelling studies of the AP have focused on SLR 40 projections from ice-shelf tributary glaciers, ignoring any contributions from 50 tidewater glaciers (Barrand et al., 2013a; Schannwell et al., 2015). Given 51 the observed acceleration and retreat of most tidewater glaciers (*Cook et al.*, 52 2005; Pritchard and Vaughan, 2007), this may lead to a substantial underes-53 timation of the SLR contribution from the AP. In this paper, we present the 54 first comprehensive modelling study of SLR projections from both tidewater 55 and ice-shelf tributary glaciers of the AP. Building on the work of *Schannwell* 56 et al. (2015), ice-shelf collapse timing is not determined by thermal viability 57 limits, but is instead based on the total number of melt days - a more direct 58 and physically-based link to the process of hydrofracture. Daily instead of 59 monthly near-surface temperature projections are used to estimate timing 60 of future ice-shelf collapse events. To estimate grounding line retreat in re-61 sponse to ice-shelf removal, a new statistical framework is introduced that 62 builds on previous work by Schannwell et al. (2015), but improves upon their 63

statistical parameterisation by relating expected grounding line retreat to the 64 degree of buttressing. Buttressing for each drainage basin at the grounding 65 line is calculated by dividing the normal pressure in presence of an ice-shelf 66 by the ocean pressure acting when no ice-shelf is present. The combined SLR 67 contribution over the next 300 years is computed, including for the first time 68 the largest 235 tidewater glaciers throughout the northern AP. In addition 69 to this, volume responses of the largest 215 ice-shelf tributary glaciers are 70 simulated. These 450 drainage basins cover a total of 77% of the AP's area, 71 providing a comprehensive coverage of the Antarctic Peninsula Ice Sheet 72 (APIS). 73

74 2. Data and Methods

75 2.1. Climate data and preprocessing

In order to estimate the timing of future ice-shelf collapse events, daily 76 near-surface temperature fields from 13 GCMs from the Coupled Model In-77 tercomparison Project Phase 5 (CMIP5) (Taylor et al., 2011) were selected 78 using the Representative Concentration Pathway (RCP)4.5 (Vuuren et al., 79 2011) and RCP8.5 emission scenarios. The selection of the GCM forcings 80 are provided in Figure A.6 and follows Schannwell et al. (2015). Tempera-81 ture projection fields were bias-corrected against monthly ERA-Interim data 82 from the European Centre for Medium Range Weather Forecasts (ECMWF; 83 Dee et al., 2011) by shifting the future temperature fields by the average 84 bias for each month between the GCM and ERA-Interim temperatures over 85 the period 1979-2005 (Radić et al., 2014). The bias-corrected temperatures 86 were then compared to surface station data (Table B.2) from the AP. The 87

remaining temperature difference between bias-corrected temperature fields 88 and surface station data is attributed to an inaccurate height representa-89 tion in the temperature fields caused by the relative coarse spatial resolution 90 of the models ($\sim 0.75^{\circ}$). Owing to the rugged topography of the AP, this 91 can introduce significant temperature differences (Jones and Lister, 2014). 92 To correct for this, temperature fields were shifted by a temperature-height 93 correction factor derived for each month from every station. As most sur-94 face stations are clustered in the north of the AP, temperature data from 95 automatic weather stations were additionally included to improve spatial 96 coverage. A list of stations is provided in the appendix (Table B.2). Height 97 correction factors were then bi-linearly interpolated and extrapolated to pro-98 vide an ice-sheet wide correction map for each month. 99

The same sample of GCMs was selected for monthly ocean surface temperature fields which were bias-corrected against the Extended Reconstructed Sea Surface Temperature (ERSST) v4 reanalysis product (*Huang et al.*, 2015) using the same methods as for the surface temperature fields. A plot of the bias for each GCM is provided in the appendix.

105 2.2. Tidewater glaciers

A substantial portion of the mass loss of ice sheets and near-polar glaciers comes from calving (*Rignot and Kanagaratnam*, 2006; *Benn et al.*, 2007a; *Barrand et al.*, 2013b). While the importance of iceberg calving has been recognised and a number of empirical calving laws have been proposed (*Brown et al.*, 1982; *van der Veen*, 1996; *Benn et al.*, 2007b; *Alley et al.*, 2008; *Luckman et al.*, 2015), modelling iceberg calving remains a major source of uncertainty in ice-sheet models (*O'Leary and Christoffersen*, 2013). Unlike

the rest of the Antarctic Ice Sheet, the AP is located in a maritime climate, 113 experiencing significant surface melt during the austral summer. These char-114 acteristics, combined with small- to medium-size calving fronts, demonstrate 115 strong similarity to tidewater glacier systems in Alaska, Svalbard, and coastal 116 Greenland. In the absence of a universal calving law, a scenario-type ap-117 proach was employed utilising three different types of calving criteria which 118 have been used to successfully simulate calving front retreat in at least one 119 of these regions (Brown et al., 1982; van der Veen, 1996; Luckman et al., 120 2015). Each calving criterion is assessed in a separate simulation. 121

The first criterion (henceforth, water depth) relates calving rate to water depth (e.g *Brown et al.*, 1982), using the updated formula from *Pelto and Warren* (1991)

$$V_c = 70 + 8.33 D_w, \tag{1}$$

where V_c is the calving rate in $m yr^{-1}$ and D_w is the water depth in m at the calving front.

The second criterion (henceforth, flotation criterion) follows van der Veen (1996) who argues that the calving front position is controlled by water depth and ice thickness, following the relationship:

$$H_c = \frac{\rho_w}{\rho_i} D_w + H_0, \tag{2}$$

where H_c is the critical thickness, ρ_w and ρ_i are water and ice densities, respectively, and H_0 represents the minimum thickness above the flotation thickness. Based on modelling studies from Columbia Glacier, Alaska (*van der Veen*, 1996), this parameter is set to 50 m in our experiments. Equation 2 does not provide a calving rate, but rather states that if the calving front thickness becomes less than a critical thickness H_c , the calving front becomes unstable and retreats by calving icebergs.

Recent studies have highlighted the importance of ocean temperatures and
submarine melting to calving (e.g. *Straneo et al.*, 2010; *Luckman et al.*, 2015). *Luckman et al.* (2015) derived a linear relationship between water temperature and calving rate for 3 tidewater glaciers in Svalbard. Due to the climatic
similarities between AP glaciers and Svalbard glaciers, the linear law (henceforth, ocean criterion) was adopted, following the form:

$$V_c = 0.35 \times T,\tag{3}$$

where V_c is in m per month and T is the ocean temperature between 20-60 m 143 in °C. Instead of ocean temperatures between 20-60 m, ocean induced calv-144 ing simulations are forced by monthly ocean surface temperature projections. 145 Ocean surface temperatures do not provide a good predictor for forecasting 146 short term calving trends as these lead frontal ablation by 1-2 months (Luck-147 man et al., 2015). However, since long-term calving behaviour is investigated, 148 using ocean surface temperatures is justified. This is corroborated by a com-149 parison of mean ocean surface temperatures from the World Ocean Database 150 (Levitus et al., 2013) between 1995-2004 for the model domain with mean 151 ocean temperatures for the same period for depths between 20-60 m. This 152 results in a mean decadal temperature difference of $0.19\pm0.18^{\circ}$ C between 153 the two data sets. A maximum distance of 100 km between calving front 154 and ocean pixel was selected, resulting in omission of the CSIRO GCM from 155 further analysis. 156

157

158 2.3. Ice-shelf tributary glaciers

In order to model the ice dynamic contribution from ice-shelf tributary glaciers, two important parameters need to be estimated: i) ice-shelf collapse timing and ii) the expected grounding line retreat in response to ice-shelf removal.

Ice-shelf collapse timing is computed here according to the total number of 163 melt days in a melt year, a direct link to the physical process of hydrofracture. 164 Several studies noted that immediately prior to the collapse of Larsen B Ice 165 Shelf, the number of melt days and thus the number of observed melt ponds 166 increased dramatically (e.g. Scambos et al., 2003; van den Broeke, 2005). A 167 shelf collapse melt day threshold of 102 days was calculated based on obser-168 vational data from QuikSCAT microwave measurements over Larsen B Ice 169 Shelf (Barrand et al., 2013c), a melt day threshold similar to a range of pre-170 viously reported values (Scambos et al., 2003; van den Broeke, 2005). Future 171 melt days and ice-shelf collapse timing were computed from an ensemble of 172 13 CMIP5 GCM runs (see section Climate data and preprocessing). 173

Ice flux across the grounding line is restrained in the presence of an ice-shelf (*Schoof*, 2007). Following *Gudmundsson* (2013) the normalised buttressing factor is computed:

$$\Theta = \frac{N}{N_0},\tag{4}$$

where N is the normal pressure in presence of an ice-shelf, defined by

$$N = \vec{n}_{gl}^T \left(R \vec{n}_{gl} \right). \tag{5}$$

 N_0 is the ocean pressure acting normal to the grounding when no ice-shelf is present

$$N_0 = \frac{1}{2}\rho gh \tag{6}$$

180 The vector \vec{n}_{gl} is the unit normal to the grounding line and,

$$R = 2\eta \begin{pmatrix} 2\frac{du}{dx} + \frac{dv}{dy} & \frac{1}{2}\left(\frac{du}{dy} + \frac{dv}{dx}\right) \\ \frac{1}{2}\left(\frac{du}{dy} + \frac{dv}{dx}\right) & 2\frac{dv}{dy} + \frac{du}{dx} \end{pmatrix},$$
(7)

where η is the viscosity, $\rho = \rho_i \left(1 - \frac{\rho_i}{\rho_w}\right)$, and h is the ice thickness at the grounding line.

Defined by Equation 4, drainage basins are but tressed when 0 \leq Θ \leq 1; 183 the ice-shelf is actually pulling the grounded ice when $\Theta > 1$; and drainage 184 basins are overbuttressed when $\Theta < 0$. Overbuttressed (or $\Theta < 0$) means 185 that ice slows down as it approaches the grounding line, and mass conser-186 vation would require that ice thickens towards the grounding line $(\frac{dh}{dx} > 0)$. 187 Θ was computed for each drainage basin using velocity data from *Rignot* 188 et al. (2011), viscosity data from output of an ice-sheet model inversion of 189 surface velocity data (Arthern et al., 2015), and ice thickness data from Huss 190 and Farinotti (2014) where available and Bedmap2 (Fretwell et al., 2013) 191 elsewhere. 128 of the 215 ice-shelf tributary drainage basins are buttressed, 192 52 experience ice-shelf pulling, and 35 drainage basins are overbuttressed. 193 Basins experiencing ice-shelf pulling are characterised by narrow ice fronts 194 with strong shear margins. These basins are omitted from the analysis as 195 we do not expect any ice dynamical adjustment following ice-shelf collapse. 196 While ice dynamical changes may be expected for overbuttressed drainage 197 basins, these glaciers were also excluded from further analysis as Schoof's 198

¹⁹⁹ flux formula (*Schoof*, 2007, equation 29) is not valid for these cases.

The new parameterisation of grounding line retreat is based on the assump-200 tion that highly buttressed drainage basins will react more to ice-shelf re-201 moval than lightly buttressed basins. Ice flux across the grounding line is 202 computed for each drainage basin for the buttressed and the unbuttressed 203 case ($\Theta = 1$) using Schoof's flux formula (*Schoof*, 2007). The remaining 204 input data for Schoof's flux formula (basal drag and rheological coefficient) 205 were obtained from output of an ice-sheet model inversion (Arthern et al., 206 2015). 207

Adjustment times for drainage basins are scaled to Θ . The maximum mean adjustment time (for infinitesimal positive Θ) is set to 20 years, following observations from Larsen A Ice Shelf (*Rott et al.*, 2014) and no mean adjustment time is allowed for $\Theta = 1$. In between these bounds, the mean adjustment time is computed using Schoof's Θ exponent:

$$M \propto \Theta^{\left(\frac{n}{m+1}\right)} \tag{8}$$

where M is the mean adjustment time, n=3, and m=1/3.

As mean adjustment times are based on current observations, uncertainties 214 are associated with adjustment times derived from equation 8. To account for 215 this, we allow for uncertainty in the grounding line retreat rates within the 216 bounds of a mean adjustment time. These realisations are set by a gamma 217 distribution with shape parameters k = M/1.5 and $\Theta_{\gamma} = 1.5$. The shape 218 parameters represent greater certainty in short adjustment times and less 219 certainty over longer adjustment timescales, allowing wider spread around 220 the mean adjustment time in the latter case (Figure 1a). For each of the 221 10000 computed adjustment times, a corresponding step-response function 222

for Θ is computed (Figure 1b). This mimics the behaviour observed in the 223 Amundsen Sea Sector of West Antarctica where glaciers have been observed 224 to retreat rapidly, then remain stable, before rapid retreat commences again 225 (Favier et al., 2014). The number of steps in the function and when these 226 steps occur for each step-response function are randomly determined (Figure 227 1b). However, the maximum number of steps has to be smaller or equal to 228 the adjustment time. The grounding line retreat for each realisation is then 220 computed as follows: 230

$$\Delta x_{gl} = \sum_{M=1}^{M} \frac{(q_{bgl_M} - q_{gl})}{h_{gl}}$$
(9)

Here, q_{gl} is the unbuttressed grounding line flux and q_{bgl_M} is the buttressed flux for that year using the updated Θ value from the step-response function (Figure 1b). The retreat distance for each ice-shelf buttressed drainage basin is determined by taking the mean of the 10000 retreat realisations (see Table 1).

Grounding line retreat of >1 km is projected for 22 drainage basins. The 236 vast majority of the drainage basins are expected to show very little retreat. 237 The highest retreat rates are located at drainage basins which are strongly 238 buttressed and possess thick ice at the grounding line. The least retreat in 239 response to ice-shelf collapse is expected for the drainage basins of Larsen 240 B (Scar Inlet) and Larsen C Ice Shelf (Table 1). This is in agreement with 241 independent model simulations suggesting passive shelf ice at Larsen C Ice 242 Shelf (collapse of the shelf will not induce much grounding line retreat at 243 upstream basins (*Fürst et al.*, 2016)). 244

245 2.4. Model and experimental design

Ice dynamic contribution to SLR was simulated with the British Antarc-246 tic Survey Antarctic Peninsula Ice Sheet Model (BAS-APISM), previously 247 shown to be suitable for the unique topographic setting of the AP (Barrand 248 et al., 2013a; Schannwell et al., 2015). Our simulations comprise two exper-240 iments: i) the SLR contribution to 2300 of 235 drainage basins is computed, 250 using a range of empirically-based calving criteria. In the first simulation, 251 iceberg calving is allowed until 2100 and in the second simulation, calving is 252 permitted until 2300. Differing forcing periods for calving were applied to in-253 vestigate their influence on sea-level projections at the end of the simulation 254 period. In experiment ii) the end members of the calving simulation permit-255 ting calving until 2300 are combined with SLR projections from 215 ice-shelf 256 tributary glaciers to estimate the total ice dynamic SLR contribution for the 257 AP. Ice-shelf collapse is permitted until 2300 for all simulations. 258

259 3. Results and Discussion

260 3.1. Sea-level rise from tidewater glaciers

Simulated SLR projections from tidewater glaciers underline their crucial 261 importance to the regional sea-level budget of the AP region. For the simula-262 tion allowing calving to 2100, projections are between 3.2 ± 1.6 mm and 18.6263 mm, and for the experiment permitting calving to 2300 between 8.7 ± 2.9 and 264 18.6 mm. Uncertainty ranges $(\pm 1\sigma)$ are available for ocean criterion simula-265 tions only. Across the two experiments, differences are present in projections 266 from the ocean criterion, indicating a considerable change in ocean forcing 267 between the emission scenarios (Figure 2). 268

Differences in SLR projections are most pronounced in the simulations al-269 lowing calving to 2100 (Figure 3a). In these simulations, projections from 270 the ocean criterion are an order of magnitude smaller than projections from 271 the flotation and the water depth criteria. These two calving criteria project 272 the vast majority of their total SLR by 2300 over the next 50 years. This 273 is mainly due to the fact that a few drainage basins (e.g. Fleming Glacier, 274 Wordie Bay) rest on bedrock located well below sea level and thus are very 275 vulnerable to iceberg calving in the flotation and water depth criteria (see 276 equations 1 and 2). In contrast to the projected 18.6 and 13.7 mm by 2300 277 from the water depth and flotation criteria respectively, SLR projections us-278 ing ocean forcing are moderate, projecting 3.2 ± 1.6 mm for the RCP4.5 and 279 5.0 ± 2.3 mm for the RCP8.5 emission scenario (Figure 3a). 280

These differences in SLR projections are smaller in the simulations where 281 iceberg calving is permitted until 2300. While SLR projections from the 282 water depth and flotation criteria remain unchanged, projections from the 283 ocean criterion are an order of magnitude higher and in a very similar range 284 as the other calving criteria (Figure 3b). This means that for the water 285 depth and flotation criteria, all retreat is projected to occur prior to 2100 286 in all simulations. In contrast SLR projections from the ocean criterion are 287 small to $2050 \ (< 1 \text{ mm})$, but increase dramatically after that. The RCP8.5 288 scenario projects even marginally higher SLR than the flotation criterion at 289 13.9 ± 2.1 mm, while scenario RCP4.5 projects a SLR of 8.7 ± 2.9 mm by 2300 290 (Figure 3b). 291

The larger discrepancy in SLR between the emission scenarios can be explained by the much steeper increase in ocean temperatures for the RCP8.5

scenario in the latter two centuries of the simulation period. While there is 294 only a 1.8 ± 0.7 mm difference in the first simulation (Figure 3a), this differ-295 ence almost triples to 5.2 ± 0.8 mm in the second simulation (Figure 3b). This 296 is also reflected in the ocean temperature projections (Figure 2). In 2100, 297 the temperature difference between the scenarios is at 0.6° C, but increases 298 to 4°C by 2300. The total warming observed in the multi model mean of 299 RCP8.5 is 4.6°C (Figure 2). This ocean warming however is not spatially 300 homogeneous. Rather, there are noticeable differences between the west and 301 east coasts of the peninsula. To the west of the peninsula, warming is more 302 pronounced at 0.96°C per century, compared to 0.85°C for the eastern side 303 of the peninsula. This modelled temperature disparity between the two re-304 gions continues the pattern observed in the second half of the 20st century 305 (Meredith and King, 2005). 306

In the absence of a universal calving law, it is important to note that none 307 of our calving criteria are specifically tuned for the AP. BAS-APISM also 308 cannot simulate glacier front advance. These limitations mean that the SLR 300 numbers reported here should be understood as a first-order estimate of SLR 310 from tidewater glaciers. While surface ocean temperatures appear to be a 311 reasonable approximation of temperatures at depths between 20-60 m, un-312 certainties remain how well these modelled temperatures reproduce coastal 313 ocean temperatures. The projected 18.6 mm from the water depth criterion 314 should be interpreted as a maximum that can be expected from these 235 315 glaciers. In the simulations using this criterion, the calving front retreats at 316 each drainage basin until the bedrock on which the glacier rests is very close 317 to sea level. 318

Evaluating the suitability of calving criteria to project calving rates remains 319 difficult. Studies investigating calving behaviour of individual glaciers in dif-320 ferent environmental settings have noted that the processes controlling calv-321 ing are multi-faceted and may vary for individual glaciers (Nick et al., 2013; 322 James et al., 2014; Luckman et al., 2015). Other studies have successfully 323 reproduced calving retreat rates using simple empirical calving criteria (Vieli 324 et al., 2001; Nick and Oerlemans, 2006). An indication of the general agree-325 ment across the calving criteria is provided by the second simulation (Figure 326 3b), where Fleming and Prospect glacier, Wordie Bay, are the largest single 327 contributors to SLR regardless of the applied calving criteria, projected to 328 contribute between 1.8 - 3.4 mm to SLR by 2300. 329

330 3.2. Combined ice dynamical sea-level rise

The combined SLR projections in the RCP4.5 scenario are dominated 331 by the contributions from tidewater glaciers, accounting between 79% and 332 89% to the combined SLR. There is a very minor contribution from ice-shelf 333 tributary glaciers to 2150, and their contribution to 2300 remains small at 334 2.4 ± 1.5 mm. This relative unimportance is due to the absence of ice-shelf 335 collapse (Figure 4). In the RCP4.5 scenario, the multi model mean sug-336 gests disintegration of 50% of the 10 ice shelves (Figure 4). Only one of the 337 ice-shelf tributary glaciers of George VI Central contributes significantly to 338 SLR. This basin is responsible for 67% of the SLR projected from ice-shelf 339 tributary glaciers, demonstrated by the step in the sea level curve following 340 this shelf collapse in year 2210 (Figure 5). 341

The overall importance of ice-shelf tributary glaciers to SLR increases in the RCP8.5 scenario (Figure 5b). All 10 ice-shelves are projected to disintegrate in this simulation (Figure 4). Moreover, collapse timings of ice-shelves that collapsed in the RCP4.5 occur earlier in RCP8.5. The later the forecasted ice-shelf collapse in RCP4.5, the larger is the shift in timing in the RCP8.5 scenario. While there is only a 33 year shift for Larsen B North, this shift increases to 168 years for George VI North, the last ice-shelf to collapse in the RCP4.5 scenario (Figure 4).

The collapse of more ice-shelves results in much higher SLR projections from 350 ice-shelf tributary glaciers (Figure 4). In contrast to the RCP4.5 scenario, 351 ice-shelf tributary glaciers are as important as tidewater glaciers in this sim-352 ulation. They contribute 51.4% and 42.4% to the 26.7 ± 16.2 and 32.3 ± 16.2 353 mm projected for the combined minimum and the combined maximum, re-354 spectively (Figure 5b). These projections increase by another 6 ± 1.6 mm if 355 overbuttressed glaciers are taken into account by setting Θ for each of these 356 drainage basins to the minimum value (maximum buttressing) of all ice-shelf 357 tributary glaciers. As overbuttressed drainage basins violate the Schoof flux 358 formula, these projections should be interpreted with caution and are there-350 fore omitted from the total SLR projections. Since not all SLR projections 360 from tidewater glaciers supply uncertainty ranges, uncertainty ranges for all 361 combined SLR projections are reported as $\pm 2\sigma$ of ice-shelf tributary glacier 362 simulations. 363

The relative importance of each ice-shelf to overall SLR can be assessed from the step size in the SLR curve triggered by individual ice-shelf collapse responses. While some ice-shelf collapses result in no or only a very minor increase in sea level, there are two major steps present in the sea level curve (Figure 5b). These represent the ice-shelves that were identified as the most ³⁶⁹ crucial to overall SLR. By far the largest single contributor to SLR is George ³⁷⁰ VI Ice Shelf South followed by Larsen D Ice Shelf South. The former con-³⁷¹ tributes 7.5 ± 4.4 mm by 2300 or 54% of the total contribution from ice-shelf ³⁷² tributary glaciers, while the latter contributes 2 ± 1.6 mm by 2300 or 14% of ³⁷³ the total contribution. Combined, these ice-shelves account for 68% of the ³⁷⁴ total projected SLR from ice-shelf tributary glaciers.

Ice-shelf collapse is based on an empirical parameterisation of the physi-375 cal process hypothesised as being the principal reason for ice-shelf collapse 376 - surface meltwater-induced hydrofracture. However, this collapse mecha-377 nism may not be the sole process driving ice-shelf disintegration (Shepherd 378 et al., 2003; Khazendar et al., 2015) and thus ice-shelf collapse might be mis-379 forecasted. Grounding line retreat from a gradual loss of buttressing (e.g. 380 through ice-shelf thinning) where no collapse occurs was also omitted. More-381 over, bedrock topography is only taken into account for tidewater glacier 382 retreat computations, omitting the potential of marine-ice-sheet instability 383 (MISI), a self-sustained retreat of the grounding line on retrograde sloping 384 bedrock, in ice-shelf tributary drainage basins. While a recent study sug-385 gests that widespread MISI is unlikely in the AP (*Ritz et al.*, 2015), there 386 is evidence that some regions might be susceptible to this mechanism (e.g. 387 Scar Inlet and George VI Ice Shelf) (Farinotti et al., 2014; Wouters et al., 388 2015). Despite these simplifications, the implemented grounding line retreat 389 parameterisation predicts plausible retreat rates in agreement with theoreti-390 cal considerations. 391

³⁹² In comparison to earlier ice dynamical SLR projections from ice-shelf tribu-³⁹³ tary drainage basins by *Schannwell et al.* (2015), the projections presented

here are slightly higher for the RCP4.5 scenario and slightly lower for the 394 RCP8.5 scenario. Discrepancies in SLR between Schannwell et al. (2015) 395 and this study arise due to the improvement in grounding line retreat and 396 ice-shelf collapse parameterisations here. Unlike in the previous grounding 397 line retreat parameterisation, the new parameterisation permits estimation 398 of uncertainty ranges for each simulation. Moreover, ice-shelf collapse tim-399 ing is calibrated on observations, providing a more robust approximation for 400 future collapse estimates. 401

402

403 3.3. Uncertainty assessment

In order to test the robustness of the modelled SLR projections a suite of sensitivity experiments was performed. Since SLR projections from tidewater glaciers should be understood as a first-order estimate and the three calving criteria provide an envelope of future scenarios, the sensitivity experiments concentrate on ice-shelf tributary SLR contributions.

There are two main sources of uncertainty: climate (ice-shelf collapse tim-409 ing) and grounding line retreat parameterisation. The influence of climate 410 variability on SLR projections is demonstrated by the difference between 411 the two emission scenarios. In RCP8.5, projections are ~ 6 fold higher than 412 in RCP4.5. Nonetheless, the importance of ice-shelf collapse timing in a 413 worst-case scenario is relatively moderate. The most extreme scenario with 414 immediate collapse of all fringing ice-shelves leads to an increase of 3.7 mm 415 (27%) in comparison to the projection from RCP8.5. 416

⁴¹⁷ How much the position of the grounding line changes in response to ice-shelf ⁴¹⁸ collapse is of crucial importance for SLR projections from ice-shelf tributary

glaciers. In the parameterisation implemented here, the mean adjustment 419 time is scaled to buttressing and is based on available observations from 420 Larsen A Ice Shelf. Since ice dynamical changes are still ongoing in this 421 area, maximum adjustment time might be underestimated. Grounding line 422 retreat rates for each basin were computed using Schoof's flux formula. To 423 investigate the sensitivity of the results, key parameters such as adjustment 424 time and all input data to the flux formula were perturbed by $\pm 20\%$. Re-425 sults show that by far the most important parameter is ice thickness at the 426 grounding line. SLR projections from all other perturbed parameters vary by 427 <46% ($<4.7\pm1.7$ mm) in comparison to the reference simulations and lie all 428 within the reported uncertainty ranges. For perturbed ice thicknesses how-429 ever, SLR projections vary by up to $\sim 400\%$ (53.2 ± 16.6 mm), increasing SLR 430 projections in RCP8.5 to 66.9 ± 25 mm, more than double the SLR projected 431 for the combined RCP8.5 reference simulation. These results highlight the 432 key importance of accurate estimates of ice thickness at glacier grounding 433 lines. 434

To investigate the robustness of the results to perturbations to ice velocity, 435 the velocity map was perturbed by adding normally distributed noise ($\sigma =$ 436 1 SD of unperturbed velocity map) to the unperturbed velocity map. Ice 437 velocity was used to estimate buttressing at each drainage basin. The per-438 turbed velocity map was used to compute new Θ values for the 128 modelled 439 drainage basins. Of the 128 normally buttressed basins in the reference simu-440 lation, 26 change to being overbuttressed and 31 to being unbuttressed. This 441 leaves 71 drainage basins for the perturbed model simulation. Despite the 442 smaller number of drainage basins, change in SLR for the RCP8.5 scenario 443

is negligible ($\sim 1\%$) in comparison to the reference simulation, indicating an increase in buttressing for these 70 drainage basins. Average buttressing for these basins increases from 0.59 to 0.43, negating the effect of fewer drainage basins modelled.

448

449 4. Conclusions

This paper has presented the first comprehensive modelling study of SLR 450 projections from both tidewater and ice-shelf tributary glaciers of the AP. 451 In total, the ice dynamical response of 450 drainage basins, comprising 77%452 of the AP's area, was computed. Tidewater glaciers are an important con-453 tributor to the ice dynamic SLR projections from the AP. Omission of tide-454 water glaciers leads to an underestimation of SLR by >50%. In the RCP4.5 455 scenario, SLR projections are dominated by tidewater glaciers contributing 456 >75% of the combined SLR, while tidewater and ice-shelf tributary glaciers 457 contribute about the same to total SLR in the RCP8.5 scenario. If all ice-458 shelves disintegrate, George VI Ice Shelf is the largest single contributor, 459 accounting for 9.8 ± 5.5 mm (70%) of the total SLR projected from ice-shelf 460 tributary glaciers. This agrees well with an earlier modelling study (Schan-461 *nwell et al.*, 2015) and is consistent with present-day observations of AP 462 ice-sheet mass balance (Wouters et al., 2015). 463

464 Sensitivity results show uncertainties in SLR projections remain due to calv465 ing, ice-shelf collapse, and grounding line retreat parameterisation. SLR
466 projections for ice-shelf tributary glaciers are highly sensitive to ice thickness
467 and to a lesser extent ice velocity. To reduce uncertainties further in future

simulations, accurate ice thickness and velocity maps are required for computation of buttressing and ice flux across the grounding line.

The Antarctic Peninsula Ice Sheet is projected to contribute between 11 ± 2 470 and 32 ± 16 mm to global SLR by 2300, depending on emission scenario. This 471 corresponds to an annual contribution of 0.04 ± 0.01 mm a⁻¹ and 0.11 ± 0.05 472 mm a^{-1} over the next three centuries, respectively. For comparison, the 473 SLR contribution from the entire Antarctic Ice Sheet derived from satel-474 lite observations between 2003-2013 was 0.25 ± 0.07 mm a⁻¹ (Martín-Español 475 et al., 2016). These findings underline the continued importance of ice dy-476 namic SLR from the AP, even though the AP comprises only 1% of the total 477 Antarctic Ice Sheet area. 478

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Figure 1: (a) Gamma distributions used in grounding line retreat parameterisation for different mean adjustment times (M). (b) Sample of 100 random step-response functions for corresponding M = 20 curve in (a).

Ice-Shelf	Mean Retreat [m]	Θ	No. of basins
Larsen B	691	0.47	6
Larsen C North	405	0.40	17
Larsen C South	215	0.59	31
Larsen D North	656	0.60	16
Larsen D Central	250	0.57	11
Larsen D South	4140	0.66	20
George VI North	1960	0.52	4
George VI Central	7310	0.69	3
George VI South	10530	0.69	8
Stange	29540	0.54	1

Table 1: Ice-shelf grounding line retreat distances, mean buttressing factor (Θ) , and the number of basins for each ice-shelf entity.



Figure 2: Multi model mean ocean temperatures for the ice-sheet model domain for RCP4.5 (blue line) and RCP8.5 (red line). Shading shows $(\pm 1\sigma)$ uncertainty.



Figure 3: SLR projection from tidewater glaciers permitting calving front retreat to 2100 (a) and to 2300 (b). Shading shows $(\pm 1\sigma)$ uncertainty.



Figure 4: Multi model mean melt day projections for all ice-shelves for the RCP4.5 (solid blue line) and RCP8.5 (solid red line) scenarios. Shading shows $(\pm 1\sigma)$ uncertainty. Dashed blue lines and dashed red lines denote ice-shelf collapse timing for the RCP4.5 and RCP8.5 scenarios, respectively. Dashed black line approximates collapse threshold. Note that for Scar Inlet collapse timing for both scenarios is forecasted for the same year.



Figure 5: Combined SLR for RCP4.5 (a) and RCP8.5 (b) scenarios. Red and blue line correspond to combined minimum and combined maximum projection. Dashed blue lines approximate timing of ice-shelf collapse. Error bars are displayed where available.



725 Appendix A. Ocean temperature bias



728 Appendix B. GCM temperature bias



Figure B.7: Near-surface temperature bias in comparison to ERA Interim from 1979-2005. ⁷³⁰ Dashed black line indicates multi model mean $(2.0\pm2.6^{\circ}C)$.

Station	Type	Lat	Lon	Height (m.a.s.l)
Bellinghausen	Surface	-62.2	-58.9	16
Biscoe Island	AWS	-66.0	-66.1	20
Bonaparte Point	AWS	-64.8	-64.1	8
Cape Adams	AWS	-75.0	-62.5	25
Deception	Surface	-63.0	-60.7	8
Dismal Island	AWS	-68.1	-68.8	10
Dolleman Island	AWS	-70.6	-60.9	396
Fossil Bluff	AWS	-71.3	-68.3	66
Jubany	Surface	-62.2	-58.6	4
Kirkwood Island	AWS	-68.3	-69.0	30
Limbert	AWS	-75.9	-59.2	58
Marambio	Surface	-64.8	-64.1	198
Marsh	Surface	-62.2	-59.0	10
Racer Rock	AWS	-64.1	-61.6	17
Sky Blue	AWS	-74.8	-71.5	1556
Uranus Glacier	AWS	-71.4	-68.8	753

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Table B.2: List of weather stations used to compute the statistical lapse rate. AWS =732 Automatic Weather Station.