Future sea-level rise from tidewater 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, partitioned approximately equally between contributions from tidewater glaciers and ice-shelf tributary glaciers. In the RCP4.5 scenario, sea-level rise projections to 2300 are dom-

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inated by tidewater glaciers ($\sim 8\text{--}18$ mm). In this 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 $\sim 70\%$ 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. Introduction

The Antarctic Peninsula (AP) is a mountainous and heavily glaciated region, dominated by glaciers flowing directly into the sea (henceforth tidewater glaciers) and into floating ice-shelves (henceforth ice-shelf tributary glaciers). In response to the rapid warming experienced by this region over the last 50 years (*Vaughan et al.*, 2003), glaciers have contributed at an accelerated rate to global sea-level rise (SLR) in recent years (*Cook et al.*, 2005; *Wouters et al.*, 2015). In addition to an increase in near-surface air temperatures, surface waters of the surrounding ocean have warmed (*Meredith and King*, 2005). This ocean warming has been accompanied by an acceleration (*Pritchard and Vaughan*, 2007) and retreat (*Cook et al.*, 2016) of tidewater glaciers, leading to increased ice discharge to the ocean. Climatological changes have also affected ice-shelf tributary glaciers. Un-

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

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

39 *et al.*, 2015; *Borstad et al.*, 2016). A prime example of this is the progressive
 40 mechanical weakening of the remnant Larsen B Ice Shelf over the last 15
 41 years (*Borstad et al.*, 2016). The importance of these processes may, how-
 42 ever, vary for individual ice shelves.

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

64 statistical framework is introduced that builds on previous work by *Schan-*
 65 *nwell et al.* (2015), improving their statistical parameterisation by relating
 66 expected grounding line retreat to the degree of buttressing. Buttressing at
 67 the grounding line of each drainage basin is calculated by dividing the nor-
 68 mal pressure in presence of an ice shelf by the ocean pressure acting when no
 69 ice shelf is present. The combined SLR contribution over the next 300 years
 70 is computed, including for the first time the largest 235 tidewater glaciers
 71 throughout the northern AP. In addition to this, volume responses of the
 72 largest 215 ice-shelf tributary glaciers are simulated. These 450 drainage
 73 basins cover a total of 77% of the AP's area, providing a comprehensive
 74 coverage of the Antarctic Peninsula Ice Sheet (APIS).

75 **2. Data and Methods**

76 *2.1. Climate data and preprocessing*

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

88 were then compared to surface station data (Table B.2) from the AP. The
 89 remaining temperature difference between bias-corrected temperature fields
 90 and surface station data is attributed to an inaccurate height representa-
 91 tion in the temperature fields caused by the relative coarse spatial resolution
 92 of the models ($\sim 0.75^\circ$). Owing to the rugged topography of the AP, this
 93 can introduce significant temperature differences (*Jones and Lister, 2014*).
 94 To correct for this, temperature fields were shifted by a temperature-height
 95 correction factor derived for each month from every station. As most sur-
 96 face stations are clustered in the north of the AP, temperature data from
 97 automatic weather stations were additionally included to improve spatial
 98 coverage. A list of stations is provided in the appendix (Table B.2). Height
 99 correction factors were then bi-linearly interpolated and extrapolated to pro-
 100 vide an ice-sheet wide correction map for each month.
 101 The same sample of GCMs was selected for monthly ocean surface temper-
 102 ature fields which were bias-corrected against the Extended Reconstructed
 103 Sea Surface Temperature (ERSST) v4 reanalysis product (*Huang et al., 2015*)
 104 using the same methods as for the surface temperature fields. A plot of the
 105 bias for each GCM is provided in the appendix.

106 2.2. Tidewater glaciers

107 A substantial portion of the mass loss of ice sheets and near-polar glaciers
 108 comes from calving (*Rignot and Kanagaratnam, 2006; Benn et al., 2007a;*
 109 *Barrand et al., 2013b*). While the importance of iceberg calving has been
 110 recognised and a number of empirical calving laws have been proposed (*Brown*
 111 *et al., 1982; van der Veen, 1996; Benn et al., 2007b; Alley et al., 2008; Luck-*
 112 *man et al., 2015*), modelling iceberg calving remains a major source of un-

113 certainty in ice-sheet models (*O’Leary and Christoffersen, 2013*). Unlike
 114 the rest of the Antarctic Ice Sheet, the AP is located in a maritime climate,
 115 experiencing significant surface melt during the austral summer. These char-
 116 acteristics, combined with small- to medium-size calving fronts, demonstrate
 117 strong similarity to tidewater glacier systems in Alaska, Svalbard, and coastal
 118 Greenland. In the absence of a universal calving law, a scenario-type ap-
 119 proach was employed utilising three different types of calving criteria which
 120 have been used to successfully simulate calving front retreat in at least one
 121 of these regions (*Brown et al., 1982; van der Veen, 1996; Luckman et al.,*
 122 *2015*). Each calving criterion is assessed in a separate simulation.
 123 The first criterion (henceforth, water depth) relates calving rate to water
 124 depth (e.g *Brown et al., 1982*), using the updated formula from *Pelto and*
 125 *Warren (1991)*

$$V_c = 70 + 8.33D_w, \quad (1)$$

126 where V_c is the calving rate in $m\ yr^{-1}$ and D_w is the water depth in m at
 127 the calving front.
 128 The second criterion (henceforth, flotation criterion) follows *van der Veen*
 129 *(1996)* who argues that the calving front position is controlled by water depth
 130 and ice thickness, following the relationship:

$$H_c = \frac{\rho_w}{\rho_i} D_w + H_0, \quad (2)$$

131 where H_c is the critical thickness, ρ_w and ρ_i are water and ice densities,
 132 respectively, and H_0 represents the minimum thickness above the flotation
 133 thickness. Based on modelling studies from Columbia Glacier, Alaska (*van der*
 134 *Veen, 1996*), this parameter is set to 50 m in our experiments. Equation 2

135 does not provide a calving rate, but rather states that if the calving front
 136 thickness becomes less than a critical thickness H_c , the calving front becomes
 137 unstable and retreats by calving icebergs.

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

$$V_c = 0.35 \times T, \quad (3)$$

144 where V_c is in m per month and T is the ocean temperature between 20-60
 145 m in °C. Instead of ocean temperatures between 20-60 m, ocean induced
 146 calving simulations are forced by monthly ocean surface temperature pro-
 147 jections. Ocean surface temperatures do not provide a good predictor for
 148 forecasting short term calving trends as these lead frontal ablation by 1-2
 149 months (*Luckman et al.*, 2015). However, since long-term calving behaviour
 150 is investigated, use of ocean surface temperatures is justified. This is cor-
 151 roborated by a comparison over the model domain, between 1995-2004, of
 152 mean ocean surface temperatures and temperatures at 20-60 m depth, from
 153 the World Ocean Database (*Levitus et al.*, 2013). This comparison provided
 154 a mean decadal temperature difference of $0.19 \pm 0.18^\circ\text{C}$ between the two data
 155 sets. A maximum distance of 100 km between calving front and ocean pixel
 156 was selected, resulting in omission of the CSIRO GCM from further analysis.

157

158 2.3. Ice-shelf tributary glaciers

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

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

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

$$\Theta = \frac{N}{N_0}, \quad (4)$$

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

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

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

$$N_0 = \frac{1}{2}\rho gh \quad (6)$$

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

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

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

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

199 flux formula (*Schoof*, 2007, equation 29) is not valid for these cases.
200 The new parameterisation of grounding line retreat is based on the assump-
201 tion that highly buttressed drainage basins will react more to ice-shelf re-
202 moval than lightly buttressed basins. Ice flux across the grounding line is
203 computed for each drainage basin for the buttressed and the unbuttressed
204 case ($\Theta = 1$) using Schoof’s flux formula (*Schoof*, 2007). The remaining
205 input data for Schoof’s flux formula (basal drag and rheological coefficient)
206 were obtained from output of an ice-sheet model inversion (*Arthern et al.*,
207 2015).
208 Adjustment times for drainage basins are scaled to Θ . The maximum mean
209 adjustment time (for infinitesimal positive Θ) is set to 20 years, following
210 observations from Larsen A Ice Shelf (*Rott et al.*, 2014) and no mean ad-
211 justment time is allowed for $\Theta = 1$. In between these bounds, the mean
212 adjustment time is computed using Schoof’s Θ exponent:

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

213 where M is the mean adjustment time, $n=3$, and $m=1/3$.
214 As mean adjustment times are based on current observations, uncertainties
215 are associated with adjustment times derived from equation 8. To account for
216 this, we allow for uncertainty in the grounding line retreat rates within the
217 bounds of a mean adjustment time. These realisations are set by a gamma
218 distribution with shape parameters $k = M/1.5$ and $\Theta_\gamma = 1.5$. The shape
219 parameters represent greater certainty in short adjustment times and less
220 certainty over longer adjustment timescales, allowing wider spread around
221 the mean adjustment time in the latter case (Figure 1a). For each of the
222 10000 computed adjustment times, a corresponding step-response function

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

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

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

245 2.4. Model and experimental design

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

259 3. Results and Discussion

260 3.1. Sea-level rise from tidewater glaciers

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

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

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

292 The larger discrepancy in SLR between the emission scenarios can be ex-
 293 plained by the much steeper increase in ocean temperatures for the RCP8.5

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

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

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

330 3.2. Combined ice dynamical sea-level rise

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

342 The overall importance of ice-shelf tributary glaciers to SLR increases in the
 343 RCP8.5 scenario (Figure 5b). All 10 ice shelves are projected to disintegrate

344 in this simulation (Figure 4). Moreover, collapse timings of ice shelves that
 345 collapsed in the RCP4.5 occur earlier in RCP8.5. The later the forecasted
 346 ice-shelf collapse in RCP4.5, the larger is the shift in timing in the RCP8.5
 347 scenario. While there is only a 33 year shift for Larsen B North, this shift
 348 increases to 168 years for George VI North, the last ice-shelf to collapse in
 349 the RCP4.5 scenario (Figure 4).

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

364 The relative importance of each ice shelf to overall SLR can be assessed from
 365 the step size in the SLR curve triggered by individual ice-shelf collapse re-
 366 sponses. While some ice-shelf collapses result in no or only a very minor
 367 increase in sea level, there are two major steps present in the sea level curve
 368 (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.

375 Ice-shelf collapse is based on an empirical parameterisation of the physical
 376 process hypothesised as being the principal reason for ice-shelf collapse - sur-
 377 face meltwater-induced hydrofracture. A melt day threshold is likely to be
 378 a function of annual accumulation as well, but here we stick to the simplest
 379 model mainly due to the lack of observations that would allow for a more
 380 complex model. Moreover, this collapse mechanism may not be the sole pro-
 381 cess driving ice-shelf disintegration (*Shepherd et al.*, 2003; *Khazendar et al.*,
 382 2015) and thus ice-shelf collapse might be mis-forecasted. Grounding line
 383 retreat from a gradual loss of buttressing (e.g. through ice shelf thinning),
 384 where no collapse occurs, was also omitted. Moreover, bedrock topography is
 385 only taken into account for tidewater glacier retreat computations, omitting
 386 the potential of marine-ice-sheet instability (MISI, a self-sustained retreat
 387 of the grounding line on retrograde sloping bedrock) in ice-shelf tributary
 388 drainage basins. While a recent study suggests that widespread MISI is un-
 389 likely in the AP (*Ritz et al.*, 2015), there is evidence that some regions may
 390 be susceptible to this mechanism (e.g. Scar Inlet and George VI Ice Shelf)
 391 (*Farinotti et al.*, 2014; *Wouters et al.*, 2015). Despite these simplifications,
 392 the implemented grounding line retreat parameterisation predicts plausible
 393 retreat rates in agreement with theoretical considerations.

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

404

405 3.3. Uncertainty assessment

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

411 There are two main sources of uncertainty: climate (ice-shelf collapse tim-
 412 ing) and grounding line retreat parameterisation. The influence of climate
 413 variability on SLR projections is demonstrated by the difference between
 414 the two emission scenarios. In RCP8.5, projections are ~ 6 fold higher than
 415 in RCP4.5. Nonetheless, the importance of ice-shelf collapse timing in a
 416 worst-case scenario is relatively moderate. The most extreme scenario with
 417 immediate collapse of all fringing ice shelves leads to an increase of 3.7 mm
 418 (27%) in comparison to the projection from RCP8.5.

419 How much the position of the grounding line changes in response to ice-shelf
 420 collapse is of crucial importance for SLR projections from ice-shelf tributary
 421 glaciers. In the parameterisation implemented here, the mean adjustment
 422 time is scaled to buttressing and is based on available observations from
 423 Larsen A Ice Shelf. Since ice dynamical changes are still ongoing in this
 424 area, maximum adjustment time may be underestimated. Grounding line
 425 retreat rates for each basin were computed using Schoof’s flux formula. To
 426 investigate the sensitivity of the results, crucial parameters such as adjust-
 427 ment time and all input data to the flux formula were perturbed by $\pm 20\%$.
 428 Results show that by far the most important parameter is ice thickness at the
 429 grounding line. SLR projections from all other perturbed parameters vary by
 430 $< 46\%$ ($< 4.7 \pm 1.7$ mm) in comparison to the reference simulations and lie all
 431 within the reported uncertainty ranges. For perturbed ice thicknesses how-
 432 ever, SLR projections vary by up to $\sim 400\%$ (53.2 ± 16.6 mm), increasing SLR
 433 projections in RCP8.5 to 66.9 ± 25 mm, more than double the SLR projected
 434 for the combined RCP8.5 reference simulation. These results highlight the
 435 key importance of accurate estimates of ice thickness at glacier grounding
 436 lines.

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

444 leaves 71 drainage basins for the perturbed model simulation. Despite the
445 smaller number of drainage basins, change in SLR for the RCP8.5 scenario
446 is negligible ($\sim 1\%$) in comparison to the reference simulation, indicating an
447 increase in buttressing for these 70 drainage basins. Average buttressing for
448 these basins increases from 0.59 to 0.43, negating the effect of fewer drainage
449 basins modelled.

450

451 4. Conclusions

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

466 Sensitivity results show uncertainties in SLR projections remain due to calv-
467 ing, ice-shelf collapse, and grounding line retreat parameterisation. SLR

468 projections for ice-shelf tributary glaciers are highly sensitive to ice thickness
 469 and to a lesser extent ice velocity. To reduce uncertainties further in future
 470 simulations, accurate ice thickness and velocity maps are required for com-
 471 putation of buttressing and ice flux across the grounding line.
 472 The Antarctic Peninsula Ice Sheet is projected to contribute between 11 ± 2
 473 and 32 ± 16 mm to global SLR by 2300, depending on emission scenario. This
 474 corresponds to an annual contribution of 0.04 ± 0.01 mm a⁻¹ and 0.11 ± 0.05
 475 mm a⁻¹ over the next three centuries, respectively. For comparison, the
 476 SLR contribution from the entire Antarctic Ice Sheet derived from satel-
 477 lite observations between 2003-2013 was 0.25 ± 0.07 mm a⁻¹ (*Martín-Español*
 478 *et al.*, 2016). These findings underline the continued importance of ice dy-
 479 namic SLR from the AP, even though the AP comprises only 1% of the total
 480 Antarctic Ice Sheet area.

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497 **References**

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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.

733 **Appendix A. Ocean temperature bias**

734 Figure A.6: Ocean temperature bias in comparison to ERSST v4 from 1979-2005 for each
735 GCM. Dashed black line indicates multi model mean ($-0.6 \pm 0.7^\circ\text{C}$).

736 **Appendix B. GCM temperature bias**

Figure B.7: Near-surface temperature bias in comparison to ERA Interim from 1979-2005.

737 Dashed black line indicates multi model mean ($2.0 \pm 2.6^\circ\text{C}$).

738

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
739 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

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

Figure1

Figure2

Figure3

Figure4

Figure5

Figure A.6

Figure B.7