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1 **Linear response of east Greenland's tidewater glaciers to ocean/atmosphere warming**

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9 **Abstract**

10 Predicting the retreat of tidewater outlet glaciers forms a major obstacle to forecasting the rate of
11 mass loss from the Greenland Ice Sheet. This reflects the challenges of modelling the highly dynamic,
12 topographically complex and data poor environment of the glacier–fjord systems that link the ice
13 sheet to the ocean. To avoid these difficulties, we investigate the extent to which tidewater glacier
14 retreat can be explained by simple variables: air temperature, meltwater runoff, ocean temperature,
15 and two simple parameterisations of ‘ocean/atmosphere’ forcing based on the combined influence
16 of runoff and ocean temperature. Over a 20-year period at 10 large tidewater outlet glaciers along
17 the east coast of Greenland, we find that ocean/atmosphere forcing can explain up to 76 % of the
18 variability in terminus position at individual glaciers and 54 % of variation in terminus position across
19 all 10 glaciers. Our findings indicate that 1) the retreat of east Greenland's tidewater glaciers is best
20 explained as a product of both oceanic and atmospheric warming and 2) despite the complexity of
21 tidewater glacier behaviour, over multi-year time scales a significant proportion of terminus position
22 change can be explained as a simple function of this forcing. These findings thus demonstrate that
23 simple parameterisations can play an important role in predicting the response of the ice sheet to
24 future climate warming.

25 **Significance statement**

26 Mass loss from the Greenland Ice Sheet is expected to be a major contributor to 21st Century sea
27 level rise, but projections retain substantial uncertainty due to the challenges of modelling the
28 retreat of the tidewater outlet glaciers that drain from the ice sheet into the ocean. Despite the
29 complexity of these glacier-fjord systems, we find that over a 20-year period much of the observed
30 tidewater glacier retreat can be explained as a predictable response to combined atmospheric and
31 oceanic warming, bringing us closer to incorporating these effects into the ice sheet models used to
32 predict sea level rise.

33 **Introduction**

34 Loss of mass from tidewater glaciers to the ocean through iceberg calving and submarine melting is a
35 major component of the mass budget of the Greenland Ice Sheet (GrIS). The contribution of this
36 *frontal ablation* to ice sheet mass balance can vary dramatically on short timescales: increased
37 frontal ablation was responsible for 39 % of GrIS mass loss from 1991-2015 (van den Broeke et al.,
38 2017), and accounted for as much as two thirds of GrIS mass loss during a phase of rapid retreat,
39 acceleration and thinning of outlet glaciers between 2000-05 (Rignot and Kanagaratnam, 2006).
40 Understanding the controls on frontal ablation is thus crucial if its contribution to the mass budget
41 of the GrIS is to be predicted by models (e.g. Nick et al., 2013).

42 Frontal ablation and tidewater glacier retreat are closely interlinked – if ice loss at the terminus is
43 more rapid than the delivery of ice from up-glacier, the terminus will retreat. A leading hypothesis
44 attributes the recent rapid retreat of many of Greenland’s tidewater glaciers to an increase in
45 submarine melting, and consequently calving, in response to oceanic warming (e.g. Straneo and
46 Heimbach, 2013). Alternatively, retreat may have been driven by increasing surface melt, with
47 meltwater runoff draining through glaciers and entering fjords at depth to form buoyant plumes
48 which enhance submarine melting at glacier termini (e.g. Jenkins, 2011, Fried et al., 2015). It has also
49 been suggested that increased surface melt and runoff may accelerate calving through
50 hydrofracturing of near-terminus crevasses (e.g. Nick et al., 2013), or by increasing basal water
51 pressure and hence basal motion (e.g. Sugiyama et al., 2011). A third hypothesis links retreat to
52 increased calving rates following a reduction in terminus buttressing by ice mélange and land-fast
53 sea ice (e.g. Christoffersen et al., 2012, Moon et al., 2015). In most cases however, it has not proven
54 possible to attribute observed variability in terminus position to a particular cause, especially when
55 multiple glaciers are considered (e.g. McFadden et al., 2011, Bevan et al., 2012, Carr et al., 2013,
56 Moon et al., 2015, Murray et al., 2015).

57 The lack of a clear relationship between observed tidewater glacier retreat and changing
58 environmental conditions presents a significant issue for modelling studies which seek to predict
59 mass loss from the GrIS under a warming climate (e.g. Goelzer et al., 2013, Fürst et al., 2015). One
60 challenge in establishing a causal relationship between environmental forcings and tidewater glacier
61 retreat is that at the scale of individual glaciers these relationships often appear highly nonlinear,
62 with feedbacks triggered as the terminus retreats across uneven bed topography obscuring the
63 forcing driving the initial retreat (e.g. Vieli et al., 2002). This difficulty is compounded by a poor
64 understanding of the oceanic forcing of these glaciers, due both to the scarcity of observations and
65 the complexities of calving and submarine melt processes at glacier termini (Straneo et al., 2013). A

66 further issue is that accurately representing these processes in ice sheet and ocean models would
67 require model resolution and a knowledge of boundary conditions that lies beyond current
68 capabilities (Benn et al., 2017).

69 In this paper, we seek to address these challenges to improve our understanding of the retreat of
70 Greenland's tidewater glaciers on timescales relevant to predictions of mass loss over coming
71 decades. We focus our study on 10 tidewater glaciers along Greenland's east coast of varying size
72 and spanning $> 10^\circ$ of latitude (Table S1; Figure 1). Over a 20-year period (1993-2012) we compare
73 the observed pattern and rate of retreat with variability in five environmental forcings, assessing the
74 ability of these forcings to explain variability in the terminus position (P) of the study glaciers, both
75 individually and collectively. These forcings comprise near-terminus air temperature (T_A), glacier
76 meltwater runoff (Q), ocean temperature (T_O) and two parameterisations of combined
77 'ocean/atmosphere' forcing (M_1 and M_2). These ocean/atmosphere forcing parameterisations reflect
78 the theory that frontal ablation will depend not only on ocean temperature but also runoff due to its
79 role in stimulating buoyant upwelling adjacent to the terminus (e.g. Jenkins, 2011, Chauché et al.,
80 2014) and driving the renewal of warm water in the fjord (e.g. Cowton et al., 2016, Carroll et al.,
81 2017), thereby increasing the transfer of heat between the ocean and ice. To represent this
82 combined ocean/atmosphere forcing we define $M_1 = Q(T_O - T_f)$ and $M_2 = Q^{1/3}(T_O - T_f)$. In these
83 parameterisations, ocean temperature is expressed relative to the *in situ* freezing point at the
84 calving front, approximated as $T_f = -2.13$ °C based on a depth of 300 m and salinity of 34.5 psu (e.g.
85 Straneo et al., 2012). M_1 is thus a simple product of runoff and the oceanic heat available for
86 melting, while the addition of the exponent to the formulation for M_2 is based on the expectation
87 that submarine melt rate will scale linearly with temperature and with runoff raised to the power of
88 $1/3$ (Jenkins, 2011).

89 **Results**

90 Time series of variability in T_A , Q , T_O and P for each of the study glaciers are plotted in Figure 2 (see
91 also Methods). These time series, along with the two ocean/atmosphere forcings M_1 and M_2 , are
92 displayed as normalised values for each glacier in Figure 3. Glaciers are grouped into 'northern' and
93 'southern' subsets based on their location with respect to a steep latitudinal gradient in ocean
94 temperature at $\sim 69^\circ$ N, which reflects the influence of the Irminger Current (Seale et al., 2011;
95 Figure 1). Features specific to the individual glaciers (in particular, fjord and subglacial topography)
96 may modify their response to environmental forcings (e.g. Carr et al., 2013), and so the normalised
97 values are also averaged for the five southern and five northern glaciers to show the regional trends,
98 thereby emphasising the climatic signal (Figure 3f,l).

99 We begin by examining the relationship between terminus position and the environmental forcings
100 at the scale of individual glaciers. At the southern glaciers, there is a marked increase in the values of
101 the forcings and retreat of the glaciers between 2000 and 2005, with periods of relative stability
102 either side (Figures 2 and 3a-f). There are strong correlations between P and the forcings ($R^2 = 0.24-$
103 0.76 , depending on the glacier and forcing; Figure 4, Table S2) for both the individual glaciers and
104 regional trends. Because the time series involved are non-stationary, there is however an increased
105 risk of spurious correlations resulting from similar long-term trends in the forcing and response
106 variables existing over the study period (Granger and Newbold, 1974). We therefore run an Engle-
107 Granger test for cointegration (Engle and Granger, 1987), which facilitates statistical comparison
108 between two (or more) non-stationary time series showing stochastic trends (Methods). We find
109 that P is significantly cointegrated ($p < 0.05$) with Q and M_1 at all of the southern study glaciers, with
110 T_A and T_O at Mogens 3 (M3), AP Bernstorffs Glacier (AB) and Helheim Glacier (HG), and with M_2 at AB
111 and HG (Figure 4, Table S2). Cointegration indicates a temporally-constant functional relationship,
112 meaning that these results support the existence of causal relationships between P and the
113 environmental forcings. However, because the forcings demonstrate similar temporal variability to
114 each other, determining which (if any) is the key control on terminus position from this analysis
115 alone remains difficult.

116 The results are qualitatively similar at the northern glaciers, which show a brief retreat during a
117 phase of higher T_A , Q and T_O (and thus also M_1 and M_2) between ~1994-1995, then a slight re-
118 advance, before embarking on a more sustained retreat in keeping with the increase in the forcings
119 after ~2001 (Figure 3g-l). The statistical significance of these trends is however weaker at the
120 northern glaciers (Figure 4 and Table S2), with significant cointegration of P with all forcings at
121 Daugaard-Jensen Glacier (DJ) and with M_1 and M_2 at Waltershausen Glacier (WG). This may be due
122 in part to the smaller absolute variability in the time series at the northern glaciers, increasing the
123 magnitude of short-term noise relative to the long-term trends (Figures 2 and 3). Nevertheless, clear
124 similarities appear between the variability in the forcings in P when the normalised data from the
125 northern glaciers are combined to show the regional trends (Figure 3l). Correlation of the individual
126 forcings and P for the combined northern glaciers data sets give R^2 values of 0.51-0.63 (significant at
127 $p < 0.01$, Figure 4, Table S2); however, only M_1 is significantly cointegrated with P at $p < 0.05$.

128 This analysis indicates that, despite the complexities introduced by bed topography and ice
129 dynamics, over timescales of a few years or more many individual glaciers display a largely linear
130 response to environmental forcings. This is particularly apparent at the southern glaciers, where
131 both the increase in forcings and glacier retreat have been more pronounced (Figures 2 and 3).
132 However, because at this level P demonstrates strong correlations with multiple forcings, it remains

133 unclear whether this retreat has been driven primarily by warming of the atmosphere, ocean, or
134 both. To gain further insight, we therefore examine variation in glacier retreat across all 10 study
135 glaciers.

136 Any environmental control on P should also be able to explain variation in retreat rate between
137 glaciers. In particular, the absolute magnitude of retreat is consistently lower at the northern
138 compared to the southern glaciers (with the standard deviation in P at the northern glaciers just 17
139 % of that exhibited at the southern glaciers), a trend which remains true for an expanded sample of
140 32 of east Greenland's tidewater glaciers (Seale et al., 2011). When the absolute variability at all
141 glaciers is considered together, there is a significant ($p < 0.01$) correlation of P with Q (Figure 5a; $R^2 =$
142 0.40), T_O (Figure 5b; $R^2 = 0.36$) and T_A (Figure 5a; $R^2 = 0.21$). However, while T_A , Q and T_O are all
143 typically higher at the southern than the northern glaciers, the latitudinal difference in the
144 magnitude of the variability is less marked compared to that in P : the standard deviation in Q , T_O and
145 T_A at the northern glaciers is 60, 74 and 93 % respectively of the standard deviation at the southern
146 glaciers. The implication is that for a given change in these forcings, the southern glaciers have
147 responded more sensitively than the northern glaciers. Combining Q and T to create M_1 and M_2
148 increases the latitudinal gradient in the forcings to give better agreement with that observed in P ,
149 with the standard deviation in both M_1 and M_2 at the northern glaciers 36 % of that exhibited at the
150 southern glaciers. Combined with a good correlation at the glacier level (Figure 4), this helps to
151 strengthen the correlation of P with M_1 (Figure 5c; $R^2 = 0.54$), and to a lesser extent the slightly more
152 complex ocean/atmosphere forcing parameter M_2 (Figure 5d; $R^2 = 0.45$).

153 We additionally test the ability of the environmental forcings to explain only the inter-glacier
154 variability in long-term retreat rate, a property of arguably greater importance than the year-to-year
155 variability from the perspective of predicting future ice sheet mass loss. To examine this, we
156 compare the overall retreat of each glacier (defined as the difference between the mean values from
157 1993-1995 and 2010-2012) against the equivalent change in the five forcings. Again M_1 and M_2
158 provide the strongest correlation, giving R^2 values of 0.54 and 0.57 ($p < 0.01$) respectively, compared
159 to 0.41 ($p < 0.01$) for T_O (Figure 5e-h; Table S3). There is no significant correlation between the
160 magnitude of the overall change in P and T_A and Q at $p < 0.05$, with the northern glaciers again
161 showing a much smaller retreat for a given increase in the atmospheric forcing.

162 **Discussion**

163 Our findings demonstrate that the timing and magnitude of tidewater glacier retreat along
164 Greenland's east coast can be effectively explained as a combined linear response to atmospheric
165 and oceanic conditions. Whilst variation in runoff alone can explain a large proportion of glacier

166 retreat at individual glaciers (Figure 4), the sensitivity of this relationship is much stronger in
167 southeast Greenland where ocean waters are warmer (and continued to warm more rapidly over
168 the study period) compared to northeast Greenland (Figure 5a-b,f-g). It may thus be that contact
169 with warm ocean waters preconditions the southern glaciers to greater sensitivity to changes in
170 atmospheric temperature and hence runoff – if the ocean temperature is close to the *in situ* melting
171 point, this will limit the potential for submarine melting, irrespective of the vigour of runoff-driven
172 circulation. Whilst previous studies have hypothesised that regional differences in glacier stability in
173 east Greenland may be linked to the strong latitudinal ocean temperature gradient (Seale et al.,
174 2011, Walsh et al., 2012) and that a combined warming of ocean and atmosphere may provide the
175 key trigger for rapid glacier retreat (Bevan et al., 2012, Christoffersen et al., 2012), we are able to
176 demonstrate quantitatively that the combined influence of ocean and atmospheric temperature
177 provides the strongest predictor of both spatial and temporal variation in glacier terminus position
178 (Figure 5). In this way, our results agree with recent observations from the Antarctic Peninsula which
179 show that, while there has been a strong atmospheric warming trend in this region, the magnitude
180 of glacier retreat is much greater in areas where glaciers are in contact with warm Circumpolar Deep
181 Water (Cook et al., 2016). While the existence of a correlation cannot alone provide conclusive
182 evidence of a causal link, our results thus join a growing body of evidence indicating a role for both
183 oceanic and atmospheric warming in driving the retreat of marine-terminating outlet glaciers.

184 Our results suggest that variability in terminus position across the 10 study glaciers can be
185 parameterised as

$$\frac{dP}{dt} = a \frac{dM_1}{dt},$$

187 (1)

188 where t is time and $a = -0.014 \pm 0.002$ or -0.018 ± 0.006 km / (m³ s⁻¹ °C) depending on whether the
189 parameterisation is fitted to maximise agreement with the year-to-year variability (Figure 5d) or
190 overall retreat (Figure 5i) respectively (Figure S3). This simple formulation effectively captures both
191 the temporal variability in the rate of change of glacier front position and the widely differing
192 magnitude of response at the different outlet glaciers (Figure 6). Across 10 glaciers, equation (1) can
193 explain 54 % of year-to-year variability in terminus position (Figure 5d) and 54 % of variation in the
194 overall retreat rate (Figure 5i). As such, while the prediction of individual tidewater glacier behaviour
195 on timescales of a few years or less may require detailed glacier-specific knowledge of bedrock
196 topography (e.g. Howat et al., 2008, Carr et al., 2013) and high-resolution modelling of ice dynamics
197 (e.g. Todd et al., 2018), our results show that on longer timescales variation in the glaciers' terminus
198 positions can be captured with much simpler parameterisations. These parameterisations translate

199 the complex interaction of ice sheets with the atmosphere and ocean into simple yet statistically
200 strong relationships that provide a new pathway for the inclusion of tidewater glacier retreat in the
201 large-scale ice sheet models needed to predict global sea level rise (e.g. Fürst et al., 2015,
202 Aschwanden et al., 2016).

203 This quasi-linear behaviour likely reflects the complex topography and thus relatively frequent
204 occurrence of pinning points (such as lateral constrictions and submarine sills) within Greenlandic
205 glacier-fjord systems. This means that, unlike regions of West Antarctica where bed topography may
206 precondition the ice sheet to centennial scale unstable retreat (e.g. Joughin et al., 2014), change at
207 many of Greenland's tidewater glaciers may occur as series of rapid short-lived retreats which
208 collectively do not deviate far from the linear response to climate warming. Capturing the exact
209 timing and magnitude of these steps is difficult and may not be necessary if the aim is to predict ice
210 sheet mass loss on timescales of decades or longer. A good example of this can be seen when
211 comparing KG and Helheim Glacier (HG): as the forcings increased between 2000 and 2005, HG
212 retreated steadily whilst KG remained comparatively stable before undergoing a rapid ~ 5 km retreat
213 between topographic pinning points in 2004-5 (Figures 3d-e and S1). If viewed over the period 2000
214 to 2005, the retreat of KG appears sudden while the retreat of HG appears prolonged; however,
215 when considered over the full 20-year time series, both glaciers exhibited a broadly similar retreat
216 between 2000-2005 with periods of comparative stability before and after.

217 This topographic influence accounts for some of the largest outliers in the relationship between P
218 and M_1 (Figure 5d), with a ~3-4 km discrepancy between the observed and parameterised modelled
219 terminus position briefly existing at KG due to the delayed response of this glacier to
220 ocean/atmosphere warming during 2000-2005 (Figures 3e and 6a). At KG, this discrepancy is short-
221 lived, but this observation illustrates how equation (1) is likely to be least effective at glaciers at
222 which current behaviour is particularly strongly influenced by topography: for example, looking to
223 west Greenland, Jakobshavn Isbræ may have been undergoing an unstable retreat into deeper water
224 since the loss of its floating tongue in the late 1990s (Joughin et al., 2008), whilst the stability of
225 Store Glacier to the north is attributed to the presence of an exceptionally prominent topographic
226 pinning point (Todd et al., 2018). While such glaciers will ultimately adjust to a new climatically
227 stable position, their terminus position may differ more strongly from the linear trend in the short
228 term. Nevertheless, our findings suggest that simple formulations such as equation (1) can play an
229 important role in parameterising the response to climate warming of many tidewater glaciers,
230 including major outlets such as KG, HG and DJ.

231 The efficacy of this approach is likely to be dependent on the timeframe in question. The influence
232 of topographic pinning points will be magnified on short timescales (~5 years or less), with this effect
233 reduced when retreat rates are averaged over longer timescales. Furthermore, the slow response
234 time of glaciers will modulate climatic signals by filtering out higher frequency variation – for
235 example, this may explain the muted response of the southern glaciers to the short-lived
236 cooling/warming over 2009-10 (Figure 2-3). At much longer timescales, glaciers will become less
237 sensitive to the ocean as they retreat into shallower water and onto dry land, while evolving ice
238 sheet mass balance and geometry will also likely impact upon the relationship between forcings and
239 terminus position. We therefore suggest that the relationship described in equation (1) is most
240 appropriate when considering processes on timescales of ~5-100 years, with uncertainty increasing
241 either side of this window.

242 The dependency of retreat rate on both runoff and ocean temperature points to a key role for
243 calving front processes in driving the retreat of Greenland's tidewater glaciers. The obvious link lies
244 in submarine melting: theory and modelling suggest that submarine melt rate is dependent on both
245 ocean temperature and runoff, with the latter driving buoyant plumes that increase the turbulent
246 transfer of oceanic heat to glacier calving fronts (e.g. Jenkins, 2011, Xu et al., 2013). The role of
247 submarine melting as a control on terminus position appears straightforward where glaciers are
248 relatively slow flowing and warm ocean waters are capable of inducing submarine melt rates on par
249 with ice velocity; in such circumstances undercutting by submarine melting may be the primary
250 source of frontal ablation (Bartholomaeus et al., 2013, Luckman et al., 2015), such that changes in
251 terminus position are determined by the difference between ice velocity and submarine melt rate
252 (Slater et al., 2017). The applicability of this mechanism at faster flowing glaciers is less clear
253 however, as predictions of ice front-averaged submarine melt rates fall far below terminus velocities
254 (Todd and Christoffersen, 2014, Rignot et al., 2016). Indeed, observations indicate a mechanistic
255 difference between the small scale calving in submarine melt dominated systems (Luckman et al.,
256 2015) and the massive buoyant calving of icebergs from Greenland's largest and fastest flowing
257 glaciers (James et al., 2014). Nevertheless, our findings indicate that terminus position at these large
258 and fast flowing glaciers also responds rapidly and predictably to variability in ocean/atmosphere
259 forcing.

260 We also note that the lack of improvement in the correlation between P and $M_2 (= Q^{1/3}(T_o - T_f))$
261 compared to $M_1 (= Q(T_o - T_f))$ is at odds with the dependency expected if retreat rate was a direct
262 function of submarine melt rate (Jenkins, 2011). It may be that this theoretical relationship (which is
263 yet to be validated by field data) does not reflect the reality of the relationship between T_o , Q and
264 submarine melting - for example, Slater *et al* (2016) report that the correct value for the exponent

265 may be as high as $\frac{3}{4}$ under certain circumstances. Alternatively, the apparently simple relationship
266 between P and M_1 may be the integrated result of not only submarine melting but also additional
267 factors including ice mélange / sea ice coverage (e.g. Christoffersen et al., 2012, Moon et al., 2015)
268 and hydrologically forced acceleration of ice motion (e.g. Sugiyama et al., 2011). The stronger
269 correlations between P and Q rather than T_A (Figures 4 and 5) indicate that catchment-wide melt,
270 and hence runoff, is of greater importance than local air temperatures at the terminus as a control
271 on retreat rate. While this suggests that processes driven by local surface melting (e.g. through
272 hydrofracture-driven calving) are of secondary importance, we cannot discount the possibility that
273 our results reflect a more complex mix of processes related to basal hydrology, glacier dynamics,
274 submarine melting and calving. Thus whilst our findings indicate that a combined ocean/atmosphere
275 forcing is a key control on the stability of even large, fast flowing tidewater glaciers, further research
276 is needed to identify and constrain the processes that link this forcing with frontal ablation and
277 glacier retreat.

278 Over a 20-year period, we have observed a significant correlation between variability in glacier
279 terminus position and a simple parameterisation that combines oceanic and atmospheric forcings at
280 10 tidewater glaciers along Greenland's east coast. Our results demonstrate that while increased
281 melting and runoff in response to atmospheric warming can explain much of the temporal variability
282 in glacier terminus position, the temperature of the adjacent ocean waters is also a strong
283 determinant of the absolute magnitude of retreat. We find that even at very large and fast flowing
284 glaciers like Kangerdlugssuaq Glacier and Helheim Glacier, where the nonlinear response to climate
285 forcing has previously been emphasised, over timescales of a few years or longer, this forcing
286 dominates over site-specific effects relating to the complexities of local topography. While
287 topography remains an important factor in modulating the response of tidewater glaciers to climate,
288 our findings nevertheless suggest that simple parameterisations linking terminus retreat to runoff
289 and ocean temperature, suitable for inclusion in large-scale ice sheet models, have an important role
290 to play in modelling the response of the Greenland ice sheet to atmospheric and oceanic warming.

291

292 **Methods**

293 *Study glaciers*

294 Details of the 10 study glaciers are given in Table S1. These glaciers represent a subset of the 32
295 glaciers documented by Seale *et al* (2011), chosen to span a range of conditions along the east coast
296 of Greenland. Within each region, the largest outlet glaciers (with respect to ice velocity and

297 terminus width; Table S1) were selected. In the far northeast of Greenland, the major outlet glaciers
298 drain into substantial floating ice tongues (e.g. Wilson et al., 2017). Charting the retreat of these
299 glaciers (where changes in grounding line position rather than calving front position are likely of
300 primary importance) is not possible with the methods employed here, and so no glaciers were
301 selected in this region.

302 *Air temperature*

303 Mean summer air temperature (Figure 2a-b), T_A , is based on the May-September mean of monthly
304 temperatures from ERA-Interim global atmospheric reanalysis (Dee et al., 2011). For each glacier,
305 temperatures are extracted from the reanalysis cell in which the terminus lies. To account for
306 differing mean topography between cells, these values are adjusted to give sea level temperature
307 assuming an atmospheric lapse rate of 0.0065 °C / m.

308 *Runoff*

309 Annual mean catchment runoff, Q , for each of the 10 glaciers (Figures 1 and 2c-d) is obtained from a
310 1 km surface melting, retention and runoff model forced with ERA-Interim and 20CR reanalyses
311 (Wilton et al., 2017). Runoff due to basal melting is expected to be limited and is therefore not
312 considered. Meltwater is routed through glacial catchments using the hydraulic potential gradient
313 (Shreve, 1972) based on the ice surface and bed topography (Bamber et al., 2013). Q is predicted to
314 be greatest at KG due to its large catchment area and more melt-favourable hypsometry relative to
315 HG and DJ, which have comparable catchment areas (Figures 1 and 2c-d).

316 *Ocean temperature*

317 We seek to compare changes in glacier terminus position to a measure of ocean water temperature,
318 T_O , in the fjords adjacent to the glaciers. Because there are few *in situ* hydrographic measurements
319 from fjords, and the fjords are not well resolved in ocean circulation models, we define $T_O = T_R + c$,
320 where T_R is ocean temperature based on reanalysis values for the continental shelf and c is a
321 correction to account for temperature differences between the shelf and fjords.

322 T_R is obtained from the GLORYS2V3 1/4° ocean reanalysis product (Ferry et al., 2012). A decision
323 must be made as to where to sample these data for each glacier. Because cross-shelf troughs are
324 poorly mapped and not well resolved in the reanalysis, cells close to fjord mouths tend to be
325 unrealistically shallow (e.g. Fenty et al., 2016) and so the warmer, deeper waters (crucial to the fjord
326 heat budget) are not captured. Conversely, if the nearest cell of depth equal to that of the grounding
327 line is chosen, this can be hundreds of kilometres away from the fjord mouth on the shelf break, and
328 it is not clear that a pathway of such depth will exist between that cell and the glacier. As a

329 compromise, we opt for the nearest cell of depth > 400 m, which is deep enough to sample the
330 warmer Atlantic waters (AW) existing at depths greater than ~ 200 m whilst in many cases being
331 located on the shelf rather than beyond the shelf break (Figure 1). For simplicity and consistency
332 between glaciers, we take T_R as the annual mean temperature between 200-400 m (Figure 2e-f). This
333 falls within the likely depth range of up-fjord currents (e.g. Cowton et al., 2016), and allows key
334 inter-annual trends in AW temperature to be captured whilst reducing noise due to large seasonal
335 variations in shelf surface water temperatures which likely have limited influence on the glaciers
336 (Straneo and Heimbach, 2013).

337 To obtain values for the correction term c , we test these time series of T_R against available *in situ*
338 observations from moorings and CTD surveys in the vicinity of T1 (Holfort et al., 2008, Murray et al.,
339 2010), HG (Straneo et al., 2016), KG (Azetsu-Scott and Syvitski, 1999, Dowdeswell, 2004, Straneo et
340 al., 2012, Inall et al., 2014) and, in the absence of data from the northern study glaciers,
341 Nioghalvfjærdsbræ (NG) in the far north east of Greenland (Wilson and Straneo, 2015) (Figures 1 and
342 S2). Fitting of T_R to the observations indicates that the reanalysis data overestimate *in situ*
343 temperatures in these locations by approximately 1.5 °C (T1), 2.9 °C (HG), 3.1 °C (KG) and 0.3 °C
344 (NG). While this may in part reflect errors in the reanalysis product (which is poorly constrained by
345 observations on the shelf), significant cooling of AW is expected as it crosses the continental shelf
346 from the core of the warm currents at the shelf break to the fjords (Straneo et al., 2012). To better
347 represent the temperature of subsurface waters entering the fjords, we use these observations to
348 adjust the values of T_R derived from the reanalysis data to give T_O . For the cluster of glaciers in
349 southeast Greenland (M3, T1 and AB) we set $c = 1.5$ °C, while at HG and KG we set $c = 2.9$ °C and 3.1
350 °C respectively. For the glaciers in northeast Greenland (BG, VG, DJ, WG, HK), influenced by the same
351 cooler recirculated AW as NG (Straneo et al., 2012), we set $c = 0.3$ °C. These offsets are then used to
352 calculate the values of T_O used throughout the paper. While this adjustment is necessarily
353 approximate given the scarcity of *in situ* observations, its application enables better representation
354 of the temporal and spatial variability in the temperature of ocean water entering Greenland's
355 fjords.

356 *Terminus position*

357 For the period 2000-2009, width-averaged changes in glacier terminus position P (expressed as
358 distance from an arbitrary up-glacier location) are taken from Seale *et al* (2011) and based on the
359 automated classification of all available MODIS imagery. We extend this time series by manual
360 termini delineation (using the linear box method (Lea et al., 2014)) in Landsat scenes (e.g. Figure S1)
361 at approximately monthly intervals over the period 2009-2015, and where available over the period

362 1990-1999. No Landsat scenes are available during the years 1991, 1993 and 1995. At KG, HG and DJ
363 we supplement these data with terminus positions delimited using Envisat imagery by Bevan *et al*
364 (2012).

365 Because the glaciers typically undergo an annual cycle of advance and retreat, error will be
366 introduced into the mean annual position for glaciers and years where there are significant gaps in
367 the available coverage. We therefore adjust glacier lengths according to

$$368 \quad P = P_{mean} + \left(\frac{1}{2}\mu_a r\right), \quad (2)$$

369

370 where P is the adjusted mean annual terminus position, as based on P_{mean} , which is the mean of the
371 available data for each year. r is the typical annual range in terminus positions for each glacier, based
372 on the period 2010-2013 when continuous Landsat availability gives accurate near year-round
373 coverage (Table S4). Each data point is given a weighting μ based on the month within which it falls,
374 ranging linearly from 1 (October, when the termini are typically most retreated), to -1 (April, when
375 the termini are typically most advanced). The mean weighting for each year, μ_a , thus provides an
376 indication of the extent by which the available data points likely over or under estimate the true
377 mean annual terminus position. For example, the only two data points for 1995 at DJ fall in August
378 and September (when the glacier length will be close to its annual minimum). This gives $\mu_a = 0.5$, and
379 P is thus increased by $0.25 \times r (= 0.3 \text{ km})$ with respect to P_{mean} to better approximate the true annual
380 average terminus position. The difference between P_{mean} and P is shown in Figure 3 (being too small
381 to plot in Figure 2g-h) and is in most cases negligible.

382 *Statistics*

383 Statistical comparison of T_A , Q , T_O , M_1 and M_2 with P is undertaken at the level of mean annual
384 values. In Figure 5 (and Table S3) we consider data grouped from across the study glaciers, while in
385 Figures 3 and 4 (and Table S2) we relate individual glacier-specific time series of anomalies in T_A , Q ,
386 T_O , M_1 and M_2 to those in P . Because these individual time series are in general non-stationary,
387 classical linear regression may indicate a statistically significant correlation between variables in
388 instances where in fact no relationship exists (Granger and Newbold, 1974). To reduce the risk of
389 incorrectly interpreting such spurious relationships, we test for cointegration of the time series
390 (Engle and Granger, 1987), a technique that has proven valuable in examining the relationships
391 between non-stationary climate variables (e.g. Kaufmann and Stern, 2002, Mills, 2009, Beenstock et
392 al., 2012). Cointegration occurs when a relationship between two or more non-stationary time series
393 produces residuals that are themselves stationary, indicating a functional relationship that remains

394 constant in time. A more thorough description of this approach, and its application in climate
395 science, is provided by Kaufman and Stern (2002). We perform an Engle-Granger test for
396 cointegration on each of the combinations of forcing and response time series using the *egcitest*
397 function in Matlab R2016a (www.mathworks.com). Where linear regression indicates a significant
398 correlation but cointegration is not established (at $p < 0.05$), we recognise the increased risk that this
399 correlation may be spurious. All R^2 values given throughout the paper are significant at $p < 0.05$, with
400 the specific p value given in each case, and all statistical values provided in Tables S2-3.

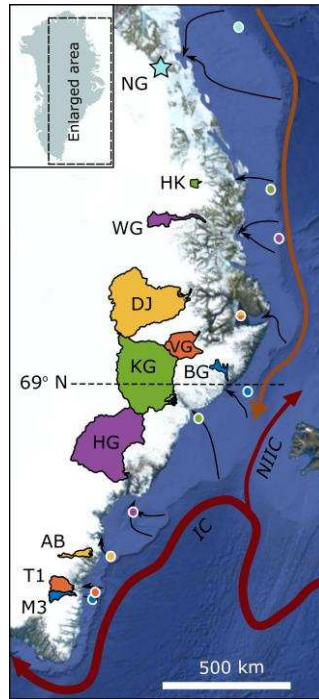
401

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408 NE/K014609/1 to PN and AS respectively and a NERC studentship to DS.

409 **Figures**

410



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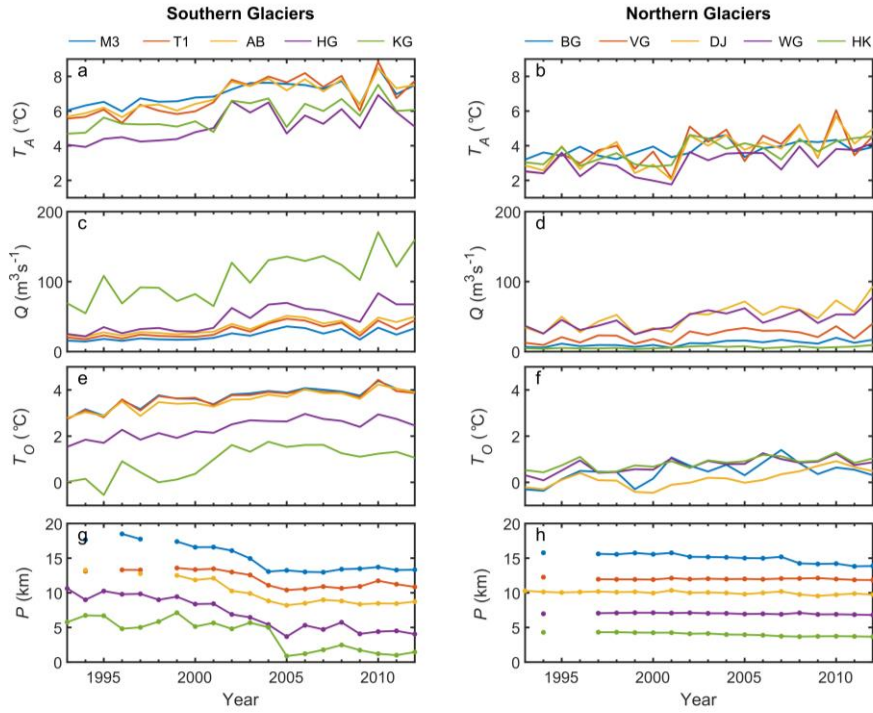
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413 Figure 1. Map showing the location of the ten study glaciers (Table S1) in east Greenland: M3 = Mogens
 414 3; T1 = Tingmjarmiut 1; AB = AP Bernstorffs Glacier; HG = Helheim Glacier; KG = Kangerdlugssuaq
 415 Glacier; BG = Borggraven; VG = Vestfjord Glacier; DJ = Daugaard-Jensen Glacier; WG = Waltershausen
 416 Glacier; HK = Heinkel Glacier. The location of Nioghalvfjærdsbræ (NG), which is referenced but does
 417 not constitute one of the study glaciers, is marked with a star. Hydrological catchments are shaded,
 418 and the divide between the northern and southern study glaciers at ~69° N is marked with the dashed
 419 line. The sample locations for ocean reanalysis temperature for the glaciers are shown as coloured
 420 circles. Also shown are the approximate locations of warm ocean currents (Straneo et al., 2012), with
 421 IC = Irminger Current and NIIC = North Iceland Irminger Current, and cross shelf troughs that may allow
 422 warm subsurface waters to access the study glaciers (black arrows; Jakobsson et al., 2012). The
 423 background image shows a satellite mosaic of Greenland with shaded sea floor bathymetry (Google
 424 Earth; Data: SIO, NOAA, U.S. Navy, NGA, GEBCO; Image: Landsat / Copernicus, IBCAO, U.S. Geological
 425 Survey).

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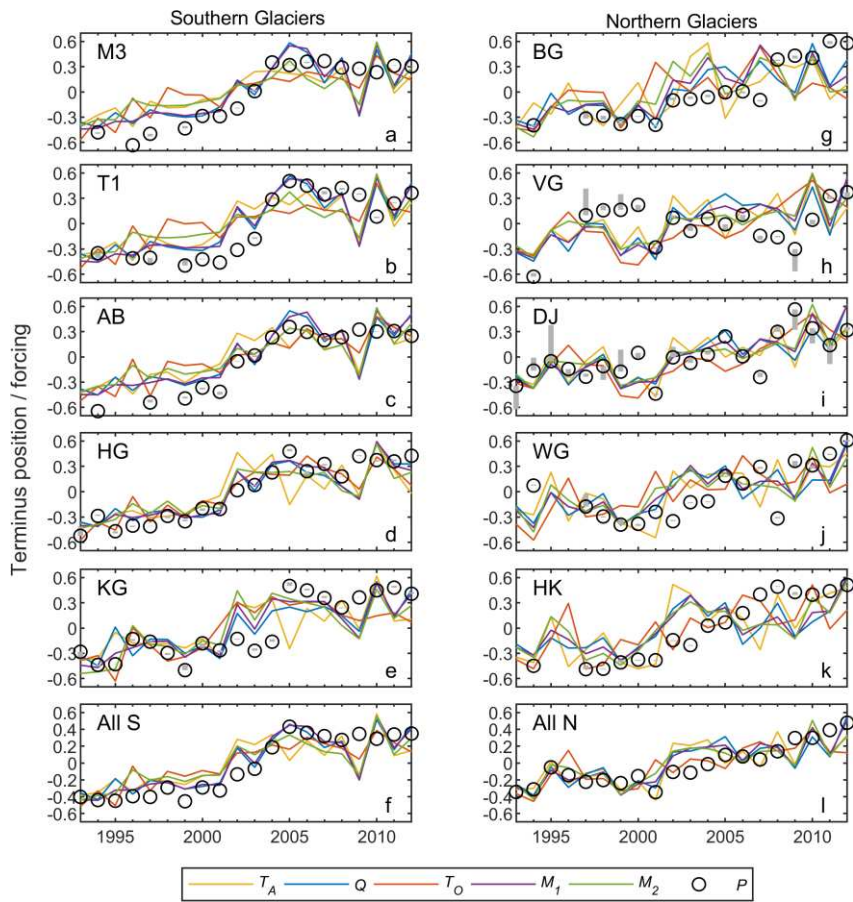
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429

430 Figure 2. Annual average values of (a-b) air temperature (T_A), (c-d) runoff (Q), (e-f) depth-averaged
 431 subsurface ocean temperature (T_O) and (g-h) glacier terminus position (P), relative to an arbitrary up-
 432 glacier location, for the 10 study glaciers (Methods; Table S1). The left and right columns show glaciers
 433 south and north of $\sim 69N$ respectively, and colours are as for Figure 1.

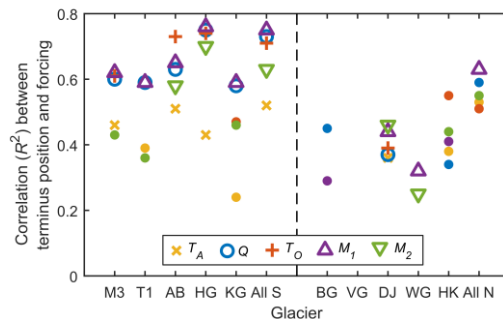
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436

437 Figure 3. Time series of normalised anomalies in air temperature (\tilde{T}_A , orange) runoff (\tilde{Q} , blue), ocean
 438 temperature (\tilde{T}_O , red), \tilde{M}_1 (purple) and \tilde{M}_2 (green) and terminus position (\tilde{P} , black circles) for each
 439 glacier. Anomalies are expressed relative to the 20-year mean, and all values are normalised with
 440 respect to the observed range at that glacier. For ease of comparison, \tilde{P} is shown inverted (i.e. positive
 441 change means retreat) and is in some cases discontinuous due to lack of observations. Vertical grey
 442 bars indicate the adjustment of P relative to P_{mean} (Methods).

443



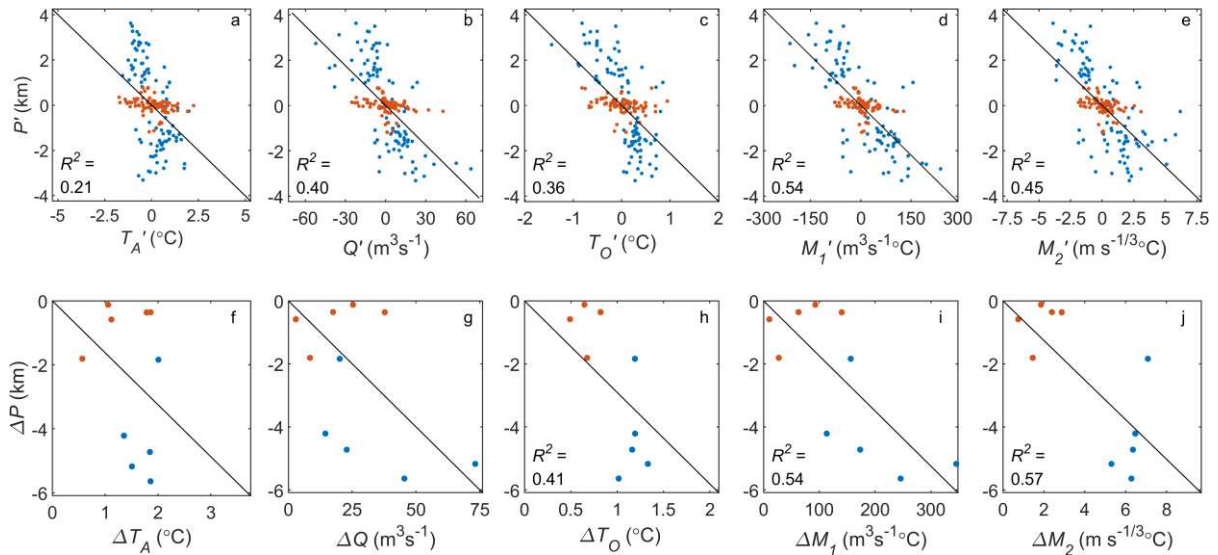
444

445 Figure 4. R^2 values for the relationship of terminus position (P) with air temperature (T_A) runoff (Q),
 446 ocean temperature (T_O) and M_1 and M_2 at each glacier and for the averaged regional southern ('All S')
 447 and northern ('All N') trends (Figure 3). Large markers show time series that are significantly

448 cointegrated at $p < 0.05$. Solid dots show instances which are correlated at $p < 0.05$, but are not
 449 cointegrated at this confidence level. No marker is shown where the time series are not significantly
 450 correlated or cointegrated. The dashed line separates the southern (left) and northern (right) glacier
 451 subsets. Statistical values are given in Table S2.

452

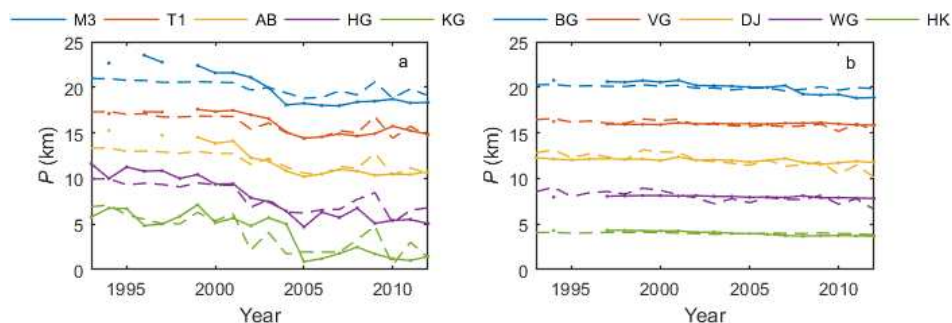
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455 Figure 5. a-d. Relationship between anomalies in terminus position (P') and (a) air temperature (T_A')
 456 (b) runoff (Q'), (c) ocean temperature (T_O'), and ocean/atmosphere forcing (d) M_1' and (e) M_2' .
 457 Anomalies are shown relative to the 20-year mean at each glacier. f-j. Relationship between overall
 458 change in terminus position (ΔP) and (f) air temperature (ΔT_A), (g) runoff (ΔQ), (h) ocean
 459 temperature (ΔT), and ocean/atmosphere forcing (i) ΔM_1 and (j) ΔM_2 . In each case, the overall
 460 change is calculated by subtracting the mean 1993-1995 value from the mean 2010-2012 value. On
 461 all plots, blue and red markers denote data from the southern and northern glaciers subsets
 462 respectively, and black lines show the best fit to all data. R^2 values (all significant at $p < 0.05$) are
 463 shown on all plots except (f) and (g), which are not significant at this level. Statistical values are given
 464 in Table S3.

465



466

467 Figure 6. Change in terminus position P at the (a) southern glaciers and (b) northern glaciers, as
 468 observed (solid lines) and parameterised based on equation (1) (dashed lines). P is shown relative to
 469 an arbitrary up-glacier location, as in Figure 2e-f.

470

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