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1 **Exceptional retreat of Novaya Zemlya's marine-terminating** 2 **outlet glaciers between 2000 and 2013**

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10 **Abstract**

11 Novaya Zemlya (NVZ) has experienced rapid ice loss and accelerated marine-terminating glacier retreat during
12 the past two decades. However, it is unknown whether this retreat is exceptional longer-term and/or whether it
13 has persisted since 2010. Investigating this is vital, as dynamic thinning may contribute substantially to ice loss
14 from NVZ, but is not currently included in sea level rise predictions. Here, we use remotely sensed data to assess
15 controls on NVZ glacier retreat between the 1973/6 and 2015. Glaciers that terminate into lakes or the ocean
16 receded 3.5 times faster than those that terminate on land. Between 2000 and 2013, retreat rates were significantly
17 higher on marine-terminating outlet glaciers than during the previous 27 years, and we observe widespread slow-
18 down in retreat, and even advance, between 2013 and 2015. There were some common patterns in the timing of
19 glacier retreat, but the magnitude varied between individual glaciers. Rapid retreat between 2000-2013
20 corresponds to a period of significantly warmer air temperatures and reduced sea ice concentrations, and to
21 changes in the NAO and AMO. We need to assess the impact of this accelerated retreat on dynamic ice losses
22 from NVZ, to accurately quantify its future sea level rise contribution.

23 **1. Introduction**

24 Glaciers and ice caps are the main cryospheric source of global sea level rise and contributed approximately -215
25 ± 26 Gt yr^{-1} between 2003 and 2009 (Gardner et al., 2013). This ice loss is predicted to continue during the 21st
26 Century (Meier et al., 2007; Radić et al., 2014) and changes are expected to be particularly marked in the Arctic,
27 where warming of up to 8 °C is forecast (IPCC, 2013). Outside of the Greenland Ice Sheet, the Russian High
28 Arctic (RHA) accounts for approximately 20% of Arctic glacier ice (Dowdeswell and Williams, 1997; Radić et
29 al., 2014) and is, therefore, a major ice reservoir. It comprises three main archipelagos: Novaya Zemlya (glacier
30 area = 21,200 km²), Severnaya Zemlya (16,700 km²) and Franz Josef Land (12,700 km²) (Moholdt et al., 2012).
31 Between 2003 and 2009, these glaciated regions lost ice at a rate of between 9.1 Gt a⁻¹ (Moholdt et al., 2012) and
32 11 Gt a⁻¹ (Gardner et al., 2013), with over 80% of mass loss coming from Novaya Zemlya (NVZ) (Moholdt et al.,
33 2012). This much larger contribution from NVZ has been attributed to it experiencing longer melt seasons and
34 high snowmelt variability between 1995 and 2011 (Zhao et al., 2014). More recent estimates suggest that the mass
35 balance of the RHA was -6.9 ± 7.4 Gt between 2004 and 2012 (Matsuo and Heki, 2013) and that thinning rates



36 increased to $-0.40 \pm 0.09 \text{ m a}^{-1}$ between 2012/13-2014, compared to the long-term average of $-0.23 \pm 0.04 \text{ m a}^{-1}$
37 (1952 and 2014) (Melkonian et al., 2016). The RHA is, therefore, following the Arctic-wide pattern of negative
38 mass balance (Gardner et al., 2013) and glacier retreat that has been observed in Greenland (Enderlin et al., 2014;
39 McMillan et al., 2016), Svalbard (Moholdt et al., 2010a; Moholdt et al., 2010b; Nuth et al., 2010), and the
40 Canadian Arctic (Enderlin et al., 2014; McMillan et al., 2016). However, the RHA has been studied far less than
41 other Arctic regions, despite its large ice volumes. Furthermore, assessment of 21st century glacier volume loss
42 highlights the RHA as one of the largest sources of future ice loss and contribution to sea level rise, with an
43 estimated loss of 20 – 28 mm of sea level rise equivalent by 2100 (Radić et al., 2014).

44 Arctic ice loss occurs via two main mechanisms: a net increase in surface melting, relative to surface
45 accumulation, and accelerated discharge from marine-terminating outlet glaciers (e.g. Enderlin et al., 2014; van
46 den Broeke et al., 2009). These marine-terminating outlets allow ice caps to respond rapidly to climatic change,
47 both immediately through calving and frontal retreat (e.g. Blaszczyk et al., 2009; Carr et al., 2014; McNabb and
48 Hock, 2014; Moon and Joughin, 2008) and also through long-term draw down of inland ice, often referred to as
49 ‘dynamic thinning’ (e.g. Price et al., 2011; Pritchard et al., 2009). During the 2000s, widespread marine-
50 terminating glacier retreat was observed across the Arctic (e.g. Blaszczyk et al., 2009; Howat et al., 2008; McNabb
51 and Hock, 2014; Moon and Joughin, 2008; Nuth et al., 2007) and substantial retreat occurred on Novaya Zemlya
52 between 2000 and 2010 (Carr et al., 2014): retreat rates increased markedly from around 2000 on the Barents Sea
53 coast and from 2003 on the Kara Sea (Carr et al., 2014). Between 1992-2010, retreat rates on NVZ were an order
54 of magnitude higher on marine-terminating glaciers (-52.1 m a^{-1}) than on those terminating on land (-4.8 m a^{-1})
55 (Carr et al., 2014), which mirrors patterns observed on other Arctic ice masses (e.g. Dowdeswell et al., 2008;
56 Moon and Joughin, 2008; Pritchard et al., 2009; Sole et al., 2008) and was linked to changes in sea ice
57 concentrations (Carr et al., 2014). However, the pattern of frontal position changes on NVZ prior to 1992 is
58 uncertain and previous results indicate different trends, dependant on the study period: some studies suggest
59 glaciers were comparatively stable or retreating slowly between 1964 and 1993 (Zeeberg and Forman, 2001),
60 whilst others indicate large reductions in both the volume (Kotlyakov et al., 2010) and the length of the ice coast
61 (Sharov, 2005) from ~1950 to 2000. Consequently, it is difficult to contextualise the observed period of rapid
62 retreat from ~2000 until 2010 (Carr et al., 2014), and to determine if it was exceptional or part of an ongoing
63 trend. Furthermore, it is unclear whether glacier retreat has continued to accelerate after 2010, and hence further
64 increased its contribution to sea level rise, or whether it has persisted at a similar rate. This paper aims to address
65 these limitations, by extending the time series of glacier frontal position data on NVZ to include the period 1973/76
66 to 2015, which represents the limits of available satellite data.

67 Initially, surface elevation change data from NVZ suggested that there was no significant difference in thinning
68 rates between marine- and land-terminating outlet glacier catchments between 2003 and 2009 (Moholdt et al.,
69 2012). This contrasted markedly with results from Greenland (e.g. Price et al., 2011; Sole et al., 2008), but was
70 similar to the Canadian Arctic, where the vast majority of recent ice loss occurred via increased surface melting
71 (~92% of total ice loss), rather than accelerated glacier discharge (~8 %) (Gardner et al., 2011). This implied that
72 outlet glacier retreat was having a limited and/or delayed impact on inland ice, or that available data were not
73 adequately capturing surface elevation change in outlet glacier basins (Carr et al., 2014). More recent results
74 demonstrate that thinning rates on marine-terminating glaciers on the Barents Sea coast are much higher than on



75 their land-terminating neighbours, suggesting that glacier retreat and calving does promote inland, dynamic
76 thinning (Melkonian et al., 2016). However, higher melt rates also contributed to surface lowering, evidenced by
77 the concurrent increase in thinning observed on land-terminating outlets (Melkonian et al., 2016). High rates of
78 dynamic thinning have also been identified on Severnaya Zemlya, following the collapse of the Matushevich Ice
79 Shelf in 2012 (Willis et al., 2015). Here, thinning rates increased to 3–4 times above the long-term average (1984–
80 2014), following the ice-shelf collapse in summer 2012, and outlet glaciers feeding into the ice shelf accelerated
81 by up to 200% (Willis et al., 2015). The most recent evidence, therefore, suggests that NVZ and other Russian
82 High Arctic ice masses are vulnerable to dynamic thinning, following glacier retreat and/or ice-shelf collapse.
83 Consequently, it is important to understand the longer-term retreat history on NVZ, in order to evaluate its impact
84 on future dynamic thinning. Furthermore, we need to assess whether the high glacier retreat rates observed on
85 NVZ during the 2000s have continued and/or increased, as this may lead to much larger losses in the future, and
86 may indicate that a step-change in glacier behaviour occurred in ~2000.

87 In this paper we use remotely sensed data to assess glacier frontal position change for all major (>1 km wide)
88 Novaya Zemlya outlet glaciers (Fig. 1). This includes all outlets from the northern ice cap and its subsidiary ice
89 caps (Fig. 1). We were unable to find the names of these subsidiary ice masses in the literature, so we name them
90 Sub 1 and Sub 2 (Fig. 1). A total of 54 outlet glaciers were investigated, which allowed us to assess the impact of
91 different glaciological, climatic and oceanic settings on retreat. Specifically, we assessed the impact of coast
92 (Barents versus Kara Sea on the northern ice mass), ice mass (northern ice cap, Sub 1 or Sub 2), terminus type
93 (marine-, lake- and land-terminating) and latitude (Table 1). The two coasts of Novaya Zemlya are characterised
94 by very different climatic and oceanic conditions: the Barents Sea coast is influenced by water from the north
95 Atlantic (Loeng, 1991; Pfirman et al., 1994; Politova et al., 2012) and subject to Atlantic cyclonic systems
96 (Zeeberg and Forman, 2001), which results in warmer air and ocean temperatures as well as higher precipitation
97 (Przybylak and Wyszynski, 2016; Zeeberg and Forman, 2001). In contrast, the Kara Sea coast is isolated from
98 north Atlantic weather systems, by the topographic barrier of NVZ (Pavlov and Pfirman, 1995), and is subject to
99 cold, Arctic-derived water, along with much higher sea ice concentrations (Zeeberg and Forman, 2001). We
100 therefore aim to investigate whether these differing climatic and oceanic conditions lead to major differences in
101 glacier retreat between the two coasts. Glaciers identified as surge-type (Grant et al., 2009) were excluded from
102 the retreat calculations and analysis. However, frontal position data are presented separately for three glaciers that
103 were actively surging during the study period. Glacier retreat was assessed from the 1973/6 to 2015, in order to
104 provide the greatest temporal coverage possible from satellite imagery. We use these data to address the following
105 questions:

- 106 1. At multi-decadal timescales, is there a significant difference in glacier retreat rates according to: i)
107 terminus type (land-, lake- or marine-terminating); ii) coast (Barents versus Kara Sea coast); iii) ice mass
108 (northern ice mass, Sub 1 or Sub 2) and; iv) latitude?
- 109 2. Are outlet glacier retreat rates observed between 2000 and 2010 on NVZ exceptional during the past ~
110 40 years?
- 111 3. Is glacier retreat accelerating, decelerating or persisting at the same rate?
- 112 4. Can we link observed retreat to changes in external forcing (air temperatures, sea ice and/or ocean
113 temperatures)?



114 2. Methods

115 2.1. Study area

116 This paper focuses on the ice masses located on the Severny Island, which is the northern island of the Novaya
117 Zemlya archipelago (Fig. 1). The northern ice cap contains the vast majority of ice (19,841 km²) and the majority
118 of the main outlet glaciers (Fig. 1). The northern island also has two smaller ice caps, Sub 1 and Sub 2, which are
119 much smaller in area (1010 km² and 705 km² respectively) and have far fewer, smaller outlet glaciers (Sub 1 = 4;
120 Sub 2 = 5) (Fig. 1). We excluded all glaciers that have been previously identified as surge type and those smaller
121 than 1 km in width from our analysis of glacier retreat rates. However, three glaciers were observed during their
122 surge phase and are discussed separately. This resulted in a total of 54 outlet glaciers, which were located in a
123 variety of settings and hence allowed us to assess spatial controls on glacier retreat (Table 1). The impact of coast
124 could only be assessed for the main ice mass, as the glaciers on the smaller ice masses, Sub 1 and Sub 2, are
125 located on the southern ice margin so do not fall on either coast (Fig. 1).

126 2.2. Glacier frontal position

127 Outlet glacier frontal positions were acquired predominantly from Landsat imagery. These data have a spatial
128 resolution of 30 m and were obtained freely via the United States Geological Survey (USGS) Global Visualization
129 Viewer (Glovis) (<http://glovis.usgs.gov/>). The frequency of available imagery varied considerably during the
130 study period. Data were available annually from 1999 to 2015 and between 1985 and 1998, although
131 georeferencing issues during the latter time period meant that imagery needed to be re-coregistered manually
132 using stable, off-ice locations as tie-points. Prior to 1985, the only available Landsat scenes dated from 1973, and
133 these also needed to be manually georeferenced. Hexagon KH-9 imagery was used to determine frontal positions
134 in 1976 and 1977, but full coverage of the study area was not available for either year. The data resolution is 20
135 to 30 feet (~6-9 m). The earliest common date for which we have frontal positions for all glaciers is 1986, and so
136 we calculate total retreat rates for the period 1986-2015 and use these values to assess spatial variability in glacier
137 recession across the study region. All glacier frontal positions are calculated relative to 1986 (i.e. the frontal
138 position in 1986 = 0 m), to allow for direct comparison.

139 Due to the lack of Landsat imagery during the 1990s, we use Synthetic Aperture Radar (SAR) Image Mode
140 Precision data during this period. The data were provided by the European Space Agency and we use European
141 Remote-sensing Satellite-1 (ERS-1) and ERS-2 products (https://earth.esa.int/web/guest/data-access/browse-data-products/-/asset_publisher/y8Qb/content/sar-precision-image-product-1477). Following Carr et al. (2013b), the
142 ERS scenes were first co-registered with ENVISAT imagery and then processed using the following steps: 1)
143 apply precise orbital state vectors; radiometric calibration; multi-look; and terrain correction. This gave an output
144 resolution of 37.5 m, which is comparable to Landsat. For each year and data type, imagery was acquired as close
145 as possible to 31st July, to minimise the impact of seasonal variability. However, this is unlikely to substantially
146 effect results, as previous studies suggest that seasonal variability in terminus position is very limited on NVZ
147 (~100 m a⁻¹) (Carr et al., 2014) and is therefore much less than the interannual and inter-decadal variability we
148 observe here. Glacier frontal position change was calculated using the box method: the terminus was repeatedly
149 digitized from successive images, within a fixed reference box and the resultant change in area is divided by the
150 reference box width, to get frontal position change (e.g. Moon and Joughin, 2008). Following previous studies
151



152 (Carr et al., 2014), we determined the frontal position errors for marine- and lake terminating outlets glaciers by
153 digitising 10 sections of rock coastline from six images, evenly spread through the time series (1976, 1986, 2000,
154 2005, 2010 and 2015) and across NVZ. The resultant error was 17.5 m, which equates to a retreat rate error of
155 1.75 m a^{-1} at the decadal time intervals discussed here. The terminus is much harder to identify on land-terminating
156 outlet glaciers due to the similarity between the debris-covered ice margins and the surrounding land, which adds
157 an additional source of error. We quantified this by re-digitising a sub-sample of six land-terminating glaciers in
158 each of the six images, which were spread across NVZ. The additional error for land-terminating glaciers was
159 66.1 m, giving a total error of 68.4m, which equates to a retreat rate error of 6.86 m a^{-1} for decadal intervals.

160 2.3. Climate and ocean data

161 Air temperature data were obtained from meteorological stations located on, and proximal to, Novaya Zemlya
162 (Fig. 1). Directly measured meteorological data are very sparse on NVZ and there are large gaps in the time series
163 for many stations. We use data from two stations, Malye Karmakuly and Im. E.K. Fedrova, as these are the closest
164 stations to the study glaciers that have a comprehensive (although still not complete) record during the study
165 period. The data were obtained from the Hydrometeorological Information, World Data Center Baseline
166 Climatological Data Sets (http://meteo.ru/english/climate/cl_data.php) and were provided at a monthly temporal
167 resolution. For each station, we calculated meteorological seasonal means (Dec-Feb, Mar-May, Jun-Aug, Sep-
168 Nov), in order to assess the timing of any changes in air temperature, as warming in certain seasons would have
169 a different impact on glacier retreat rates. Due to data gaps, particularly from 2013 onwards, we also assess
170 changes in air temperature using ERA-Interim reanalysis data ([http://www.ecmwf.int/en/research/climate-
171 reanalysis/era-interim](http://www.ecmwf.int/en/research/climate-reanalysis/era-interim)). We use temperature data from the surface (2 m elevation) and 850 m pressure level, as
172 these are likely to be a good proxy for meltwater availability (Fettweis, pers. Comm. 2017). We use the ‘monthly
173 means of daily means’ product, for all months between 1979 and 2015. As with the meteorological stations, we
174 calculate means for the meteorological seasons and annual means.

175 Sea ice data were acquired from the Nimbus-7 SMMR and DMSP SSM/I-SSMIS Passive Microwave dataset
176 (https://nsidc.org/data/docs/daac/nsidc0051_gsfsc_seaice.gd.html). The data provide information on the
177 percentage of the ocean covered by sea ice and this is measured using brightness temperatures from microwave
178 sensors. The data have a spatial resolution of $25 \times 25 \text{ km}$ and we use the monthly-averaged product. This dataset
179 was selected due to its long temporal coverage, which extends from 26 October 1978 to 31 December 2015 and
180 thus provides a consistent dataset throughout our study period. Monthly sea ice concentrations were sampled from
181 the grid squares closest to the study glaciers and were split according to coast (i.e. Barents and Kara Sea). From
182 the monthly data, we calculated seasonal means and the number of ice free months, which we define as the number
183 of months where the mean monthly sea ice cover is less than 10%.

184 Data on the North Atlantic Oscillation (NAO) were obtained from The Climatic Research Unit
185 (<https://crudata.uea.ac.uk/cru/data/nao/>) and the monthly product was used. This records the normalized pressure
186 difference between Iceland and the Azores (Hurrell, 1995). Arctic Oscillation (AO) data were acquired from the
187 Climate Prediction Centre
188 (http://www.cpc.noaa.gov/products/precip/CWlink/daily_ao_index/teleconnections.shtml). The AO is
189 characterised by winds at 55°N , which circulate anticlockwise around the Arctic (e.g. Higgins et al., 2000; Zhou



190 et al., 2001). The AO index is calculated by projecting the AO loading pattern on to the daily anomaly 1000
191 millibar height field, at 20-90°N latitude (Zhou et al., 2001). The Atlantic Multidecadal Oscillation data (AMO)
192 is a mode of variability associated with averaged, de-trended SSTs in the North Atlantic and varies over timescales
193 of 60 to 80 years (Drinkwater et al., 2013; Sutton and Hodson, 2005). Monthly data were downloaded from the
194 National Oceanic and Atmospheric Administration (<https://www.esrl.noaa.gov/psd/data/timeseries/AMO/>).

195 We use ocean temperature data from the ‘Climatological Atlas of the Nordic Seas and Northern North Atlantic’
196 (Hurrell, 1995; Korablev et al., 2014) (<https://www.nodc.noaa.gov/OC5/nordic-seas/>). The atlas compiles data
197 from over 500,000 oceanographic stations, located across the Nordic Seas, between 1900 and 2012. It provides
198 gridded climatologies of water temperature, salinity and density, at a range of depths (surface to 3500 m), for the
199 region bounded by 83.875 to 71.875 °N and 47.125°W to 57.875 °E. Here, we use data from the surface and 100m
200 depth, to capture changes in ocean temperatures at different depths: surface warming may influence glacier
201 behaviour through changes in sea ice and/or undercutting at the water-line (Benn et al., 2007), whereas warming
202 in the deeper layers can enhance sub-aqueous melting (Sutherland et al., 2013). A depth of 100 m was chosen, as
203 it is the deepest level that includes the majority of the continental shelf immediately offshore of Novaya Zemlya.
204 Further details of the data set production and error values are given in Korablev et al. (2014). We use the decadal
205 ocean temperature product to identify broad-scale changes, which is provided at the following time intervals:
206 1971-1980, 1981-1990, 1991-2000 and 2001-2012. We use the decadal product, as there are few observations
207 offshore of Novaya Zemlya during the 2000s, whereas the data coverage is much denser in the 1980s and 1990s
208 (a full inventory of the number and location of observations for each month and year is provided here:
209 <https://www.nodc.noaa.gov/OC5/nordic-seas/atlas/inventory.html>). As a result, maps of temperature changes in
210 the 2000s are produced using comparatively data few points, meaning that they may not be representative of
211 conditions in the region and that directly comparing data at a shorter temporal resolution (e.g. annual data) may
212 be inaccurate. Furthermore, the input data were measured offshore of Novaya Zemlya and not within the glacier
213 fjords. Consequently, there is uncertainty over the extent to which offshore warming is transmitted to the glacier
214 front and/or the degree of modification due to complexities in the circulation and water properties within glacial
215 fjords. We therefore use decadal-scale data to gain an overview of oceanic changes in the region, but we do not
216 attempt to use it for detailed analysis of the impact of ocean warming at the glacier front, nor for statistical testing.

217 **2.4. Statistical analysis**

218 We used a Kruksal Wallis test to investigate statistical differences in total retreat rate (1986-2015) for the different
219 categories of outlet glacier within our study population, i.e. terminus type (marine-, land- and lake-terminating),
220 coast (Barents and Kara Sea) and ice mass (northern ice cap, Sub 1 and Sub 2). The Kruksal Wallis test is a non-
221 parametric version of the one-way ANOVA (analysis of variance) test and analyses the variance using the ranks
222 of the data values, as opposed to the actual data. Consequently, it does not assume normality in the data, which is
223 required here, as Kolmogorov-Smirnov tests indicate that total retreat rate (1986-2015) is not normally distributed
224 for any of the glacier categories (e.g. terminus type). This is also the case when we test for normality at each of
225 the four time intervals discussed below (1973/6-1986, 1986-2000, 2000-2013 and 2013-2015). The Kruksal Wallis
226 test gives a p-value for the null hypothesis that two or more data samples come from the same population. As
227 such, a large p-value suggests it is likely the samples come from the same population, where as a small value



228 indicates that this is unlikely. We follow convention and use a significance value of 0.05, meaning that a p-value
229 of less than or equal to 0.05 indicates that the data samples are significantly different.

230 We assessed the influence of glacier latitude on total retreat rate (1986-2015), using simple linear regression. This
231 fits a line to the data points and gives an R^2 value and a p-value for this relationship. The R^2 value indicates how
232 well the line describes the data: if all points fell exactly on the line, the R^2 would equal 1, whereas if the points
233 were randomly distributed about the line, the R^2 would equal 0. The p-value tests the null hypothesis that the
234 regression coefficient is equal to zero, i.e. that the predictor variable (e.g. glacier catchment size) has no
235 relationship to the response variable (e.g. total glacier retreat rate). A p-value of 0.05 or less therefore indicates
236 that the null hypothesis can be rejected and that the predictor variable is related to the response variable (e.g.
237 glacier latitude is related to glacier retreat rate). The residuals for these regressions were normally distributed.
238 However, we also regressed catchment area against total retreat rate and the regression residuals were not normally
239 distributed, indicating that it is not appropriate to use regression in this case. Consequently, we used Spearman's
240 Rank Correlation Coefficient, which is non-parametric and therefore does not require the data to be normally
241 distributed. Catchments were obtained from (Moholdt et al., 2012).

242 Wilcoxon tests were used to assess significant differences in mean glacier retreat rates between four time intervals:
243 1973/6-1986, 1986-2000, 2000-2013 and 2013-2015. These intervals were chosen through manual assessment of
244 apparent breaks in the data. For each interval, data were split according to terminus type (marine, land and lake)
245 and marine-terminating glaciers were further sub-divided by coast (Barents and Kara Sea). For each category, we
246 then used the Wilcoxon test to determine whether mean retreat rates for all of the glaciers during one time period
247 (e.g. 1986-2000) were significantly different from those for another time period (e.g. 2000-2013). The Wilcoxon
248 test was selected as it is non-parametric and our retreat data are not normally distributed, and is suitable for testing
249 statistical difference between data from two time periods (Miles et al., 2013). As with the Kruksal Wallis test, a
250 p-value of less than or equal to 0.05 is taken as significant and indicates that the two time periods are significantly
251 different. We also used the Wilcoxon test to identify any significant differences in mean air temperatures and sea
252 ice conditions for the same time intervals as glacier retreat, to allow for direct comparison. For the first time
253 interval (1973/6-1986), we use air temperature data from 1976 to 1986 from the meteorological stations, but the
254 sea ice and ERA-Interim data are only available from 1979. The statistical analysis was done separately for sea
255 ice on the Barents and Kara Sea coast and using meteorological data from Malye Karmakuly and Im. E.K. Fedrova
256 (Fig. 1). ERA-Interim data was analysed as a whole, as the spatial resolution of the data does not allow us to
257 distinguish between the two coasts. In each case, we compared seasonal means for each year of a certain time
258 period, with the seasonal means for the other time period (e.g. 1976-1985 versus 2000-2012). For the sea ice data,
259 we used calendar seasons (Jan-Mar, Apr-Jun, Jul-Sep, Oct-Dec), which fits with the Arctic sea ice minima in
260 September and maxima in March. For the air temperature data, meteorological seasons (Dec-Feb, Mar-May, Jun-
261 Aug, Sep-Nov) are more appropriate. We also tested mean annual air temperatures and the number of sea-ice free
262 months.

263 In order to further investigate the temporal pattern of retreat on Novaya Zemlya, we use statistical changepoint
264 analysis (Eckley et al., 2011) We applied this to our frontal position data for marine- and lake-terminating glaciers,
265 and to the sea ice and air temperature data. Land-terminating glaciers are not included, due to the much higher
266 error margins compared to any trends, which could lead to erroneous changepoints being identified. Changepoint



267 analysis allows us to automatically identify significant changes in the time series data, and if there has been a shift
268 from one mode of behaviour to another (e.g. from slower to more rapid retreat) (Eckley et al., 2011). Formally, a
269 changepoint is a point in time where the statistical properties of prior data are different from the statistical
270 properties of subsequent data; the data between two changepoints is a segment. There are various ways that one
271 can determine when a changepoint should occur, but the most appropriate approach for our data is to consider
272 changes in regression.

273 In order to automate the process, we use the `cpt.reg` function in the R `EnvCpt` package (Killick et al., 2016) with
274 a minimum number of four data points between changes. This function uses the Pruned Exact Linear Time (PELT)
275 algorithm (Killick et al., 2012) from the changepoint package (Killick and Eckley, 2015) for fast and exact
276 detection of multiple changes. The function returns changepoint locations and estimates of the intercept and slope
277 of the regression lines between changes. We give the algorithm no information on when or how large a change
278 we might be expecting, allowing it to automatically determine statistically different parts of the data. In this way,
279 we use the analysis to determine if, and when, retreat rates change significantly on each of the marine- and lake-
280 terminating glaciers on NVZ, and whether there are any significant breaks in our sea ice and air temperature data.
281 We also apply the changepoint analysis to the number of sea ice free months, but as the data do not contain a
282 trend, we identify breaks using significant changes in the mean, rather than a change in regression. Thus, we can
283 identify any common behaviour between glaciers, the timing of any common changes, and compare this to any
284 significant changes in atmospheric temperatures and sea ice concentrations.

285 3. Results

286 3.1. Spatial controls on glacier retreat

287 The Kruskal Wallis test was used to identify significant differences in total retreat rate (1986-2015) for glaciers
288 located in different settings. First, terminus type was investigated. Results demonstrated that total retreat rates
289 (1986-2015) were significantly higher on lake- and marine-terminating glaciers than those terminating on land, at
290 a very high confidence interval (<0.001) (Fig. 2). Retreat rates were 3.5 times higher on glaciers terminating in
291 water (lake = -49.1 m a^{-1} and marine = -46.9 m a^{-1}) than those ending on land (-13.8 m a^{-1}) (Fig. 2). In contrast,
292 there was no significant difference between lake- and marine-terminating glaciers (Fig. 2). Next, we assessed the
293 role of coastal setting (i.e. Barents Sea versus Kara Sea) as climatic and oceanic conditions differ markedly
294 between the two coasts. When comparing glaciers with the same terminus type, there was no significant difference
295 in retreat rates between the two coasts (Fig. 2: p -value = 0.178 for marine-terminating glaciers and 1 for land-
296 terminating). Retreat rates on land-terminating glaciers were very similar on both coasts: Barents Sea = -6.5 m a^{-1}
297 and Kara Sea = -9.0 m a^{-1} (Fig. 2). For marine-terminating outlets, retreat rates were higher on the Barents Sea
298 (-55.9 m a^{-1}) than on the Kara Sea (-37.2 m a^{-1}), but the difference was not significant ($p=0.178$) (Fig. 2). Results
299 confirmed that the significant difference in total retreat rates between land- and marine-terminating glaciers
300 persists when individual coasts are considered (Fig. 2). Finally, we tested for differences in retreat rate between
301 the ice caps of Novaya Zemlya, specifically the northern ice cap, which is by far the largest, and the two smaller,
302 subsidiary ice caps Sub 1 and Sub 2. Here, we found no significant difference in retreat rates between the ice
303 masses (Fig. 2). Retreat rates were highest on Sub 2, followed by the northern ice cap, and lowest on Sub 1 (Fig.
304 2). Our results therefore demonstrate that the only significant difference in total retreat rates (1986-2015) relates



305 to glacier terminus type, with land-terminating outlets retreating 3.5 times slower than those ending in lakes or
306 the ocean (Fig. 2).

307 We used simple linear regression to assess the relationship between total retreat rate (1986-2015) and latitude, as
308 there is a strong north-south gradient in climatic conditions on NVZ, but no significant linear relationship was
309 apparent ($R^2 = 0.001$ $p = 0.819$) (Fig. 3). However, if we divide the glaciers according to terminus type, total
310 retreat rate shows a significant positive relationship for land-terminating glaciers ($R^2 = 0.363$ $p = 0.023$), although
311 the R^2 value is comparatively small (Fig. 3). This indicates that more southerly land-terminating outlets are
312 retreating more rapidly than those in the north. Conversely, total retreat rate for lake-terminating glaciers has a
313 significant inverse relationship with total retreat rate ($R^2 = 0.811$ $p = 0.014$), suggesting that glaciers at high
314 latitudes retreat more rapidly (Fig. 3). No linear relationship is apparent between latitude and total retreat rate for
315 marine-terminating glaciers and the data show considerable scatter, particularly in the north (Fig. 3). We find no
316 significant relationship between catchment area and total retreat rate ($RHO = -0.149$ $p = 0.339$), which
317 demonstrates that observed retreat patterns are not simply a function of glacier size (i.e. that larger glacier retreat
318 more, simply because they are bigger).

319 3.2. Temporal change

320 Based on an initial assessment of the temporal pattern of retreat for individual glaciers, we manually identified
321 major break points in the data and divided glacier retreat rates into four time intervals: 1973/6 to 1986, 1986 to
322 2000, 2000 to 2013 and 2013 to 2015 (Fig. 4). Data were separated according to terminus type and, in the case of
323 marine-terminating glaciers, according to coast. We then used the Wilcoxon test to evaluate the statistical
324 difference between these time periods for each category (Table 2). For land- and lake-terminating glaciers, there
325 were no significant differences in retreat rates between any of the time periods (Fig. 4; Table 2). Indeed, retreat
326 rates on lake-terminating glaciers were remarkably consistent between 1986 and 2015, both over time and between
327 glaciers (Figs. 4 & 5). For marine-terminating glaciers on the Barents Sea coast, the periods 1973/6 – 1986 and
328 1986-2000 were not significantly different from each other and mean retreat rates were comparatively low (-20.5
329 and -22.3 $m a^{-1}$ respectively). In contrast, the periods 2000-2013 and 2013-2015 were both significantly different
330 to all other time intervals (Fig. 4; Table 2). Between 2000 and 2013, retreat rates were much higher than at any
331 other time (-85.4 $m a^{-1}$). Conversely, the average frontal position change between 2013 and 2015 was positive,
332 giving a mean advance of $+11.6$ $m a^{-1}$ (Fig. 4). On the Kara Sea coast, marine terminating outlet glacier retreat
333 rates were significantly higher between 2000 and 2013 than any other time period (-64.8 $m a^{-1}$) (Fig. 4; Table 2).
334 Retreat rates reduced substantially during the period 2013-2015 (-22.7 $m a^{-1}$) and were very similar to values in
335 1973/6-1986 (-27.2 $m a^{-1}$) and 1986-2000 (-22.4 $m a^{-1}$) (Fig. 4). On both the Barents and Kara sea coasts, the
336 temporal pattern of marine-terminating outlet glacier retreat showed large variability, both between individual
337 glaciers and over time (Fig. 5).

338 Following our initial analysis, we used changepoint analysis to further assess the temporal patterns of glacier
339 retreat, by identifying the timing of significant breaks in the data. On the Barents Sea coast, five glaciers underwent
340 a significant change in retreat rate from the early 1990s onwards (Fig. 6). Of these, retreat rates on four glaciers
341 (MAK, TAI2, VEL and VIZ; see Fig. 1 for glacier locations and names) subsequently increased, whereas retreat
342 was slower on INO between 1989 and 2006. The most widespread step-change on the Barents Sea coast occurred



343 in the early 2000s, after which nine glaciers retreated more rapidly (Fig. 6). A second widespread change in glacier
344 retreat rates occurred in the mid-2000s, which was also the second changepoint for four glaciers (Fig. 6). Of these
345 eight glaciers, only VOE retreated more slowly after the mid-2000s changepoint. On the Kara Sea coast, we see
346 a broadly similar temporal pattern, with two glaciers showing a significant change in retreat rate from the early
347 1990s, and again in 2005 and 2007 (Fig. 6). In the case of MG, retreat rates were higher after each breakpoint,
348 whereas for SHU1, retreat rates were lower between the 1990s and mid-2000s. Four glaciers began to retreat more
349 rapidly from 2000 onwards, and five other glaciers showed a significant change in retreat rates beginning between
350 2005 and 2010 (Fig. 6), with VER being the only glacier to show a reduction in retreat rates after this change (Fig.
351 6). Focusing on lake-terminating glaciers, a significant change in retreat rates began between 2006 and 2008 on
352 all but one glacier, which began to retreat more rapidly from 2004 onwards (Fig. 6).

353 3.3. Climatic controls

354 At Im. E.K. Fedrova, mean annual air temperatures were significantly warmer in 2000-2012 (-3.9 °C) than in
355 1976-1985 (-6.5 °C) or 1986-1999 (-6.4 °C) (Fig. 4; Table 3). Looking at seasonal patterns, air temperatures were
356 significantly higher during spring, summer and autumn in 2000-2012, compared to 1976-1985, and in summer,
357 autumn and winter, when compared with 1986-1999 (Fig. 4; Table 3). Summer air temperatures averaged 5.1 °C
358 in 2000-2012, compared to 3.8 °C in 1986-1999 and 3.3 °C in 1976-1985 (Fig. 4). Warming was particularly
359 marked in winter, increasing from -16.1 °C (1976-1985) and -17.5 °C (1986-1999) to -12.9 °C in 2000-2012 (Fig.
360 4). Winter air temperatures then reduced to -15.9 °C for the period 2013-2015 (Fig. 4), although this change was
361 not statistically significant (Table 3). A similar change in mean annual air temperatures was evident on Malye
362 Karmakuly, where temperatures were significantly higher in 2000-2012 (-3.1 °C) than in 1976-1985 (-5.4 °C) or
363 1986-1999 (-5.0 °C) (Table 3; Fig 4). In all seasons, air temperatures were significantly higher in 2000-2012,
364 compared to 1976-1985 (Table 3), with the largest absolute increases occurring in winter (Fig. 4). However, only
365 autumn air temperatures were significantly warmer in 2000-2012 than 1986-1999 (Fig. 4; Table 3). No significant
366 differences in air temperatures were observed between 1976-1985 and 1986-1999 at either station (Table 3).

367 In the ERA-Interim reanalysis data, mean annual air temperatures increased significantly between 1986-1999 and
368 2000-2012 at both the surface and 850 m pressure level (Table 3). Winter (surface) and autumn (850 m)
369 temperatures also warmed significantly between these time intervals (Table 3). Surface air temperatures were
370 significantly warmer in 2013-2015, compared to 1986-1999, in winter and annually (Table 3). No significant
371 differences in air temperatures were observed at either height between 2000-2012 and 2013-2015 for any season
372 (Table 3). Surface air temperatures were comparable between 2000-2012 and 2013-2015 in winter and autumn,
373 and somewhat warmer in spring (+ 2.6 °C) and summer (+0.7 °C) in 2013-2015 (Fig. 4). At 850m height, winter
374 (-0.7 °C) and autumn temperatures were slightly cooler (-0.7 °C) and summer temperatures were warmer (+0.8 °C)
375 in 2013-2015 than in 2000-2012 (Fig. 4). At the regional scale, warmer surface air temperatures penetrate further
376 into the Barents Sea and the southern Kara Sea with each time step (Supp. Fig. 1). We observed a similar, although
377 less marked, northward progression of the isotherms at 850 m height (Supp. Fig. 1).

378 On the Barents Sea coast, sea ice concentrations during all seasons were significantly lower in 2000-2012 than in
379 1976-1985 or 1986-1999, as was the number of ice free months (Fig. 7; Table 4). Between 1976-1985 and 2000-
380 2012, mean winter sea ice concentrations reduced from 68% to 35%, mean spring values declined from 59% to



381 28% and mean autumn averages fell from 27% to 7 % (Fig. 7). Mean summer sea ice concentrations reduced
382 slightly, from 12% to 5 % (Fig. 7). Over the same time interval, the number of ice free months increased from 3.0
383 to 6.9 (Fig. 7). Summer sea ice concentrations on the Barents Sea coast reduced significantly between 2000-2012
384 and 2013-2015, but no significant change was observed in any other month, nor in the number of ice free months
385 (Fig. 7; Table 4). With exception of winter, sea ice concentrations were significantly lower in 2013-2015 than in
386 1976-1985 or 1986-1999 (Fig.4; Table 4). As on the Barents Sea coast, sea ice concentrations on the Kara Sea
387 were significantly lower in all seasons in 2000-2012, compared to 1976-1985 or 1986-1999 (Fig. 7; Table 4).
388 Summer mean sea ice concentrations declined from 25% in 1976-1985, to 13% in 2000-2012 (Fig. 7). Over the
389 same time interval, autumn mean concentrations reduced from 56% to 33%, spring values declined from 87% to
390 73% and winter values decreased from 87% to 79% (Fig. 7). The number of ice free months also reduced from
391 1.6 (1976-1985) to 3.0 (2000-2012) (Fig. 7). No significant differences were apparent between seasonal sea ice
392 concentrations and the number of ice free months in 2013-2015 and any other time period, with the exception of
393 summer sea ice concentrations between 1976-1985 and 2013-2015 (Table 4).

394 Focusing on the changepoint analysis, we see a significant change in air temperatures at Im. E.K. Fedrova from
395 2008 onwards, after which air temperatures increased markedly (Fig. 6). On the Barents Sea coast, we observe
396 significant breaks in summer sea ice concentrations at 2000 and 2008: before 2000, summer sea ice showed a
397 downward trend, but large interannual variability; between 2000 and 2008, there was a slight upward trend and
398 much lower variability and; from 2008 onwards, summer sea ice concentrations were much lower, and showed
399 both a downward trend and limited interannual variability (Supp. Fig. 2). From 2005 onwards, we observed much
400 lower interannual variability in spring, summer and autumn sea ice concentrations (Supp. Fig. 2). After 2005,
401 summer sea ice concentrations on the Kara Sea coast showed much smaller interannual variability and had lower
402 values (Supp. Fig. 3). The number of ice free months increased significantly on both the Kara Sea (from 2003)
403 and Barents Sea (from 2005) (Fig. 6).

404 Between 1970 and 1989, the summer and annual NAO index were largely positive, with a few years of negative
405 values (Fig. 8A). From 1989 to 1994, values were all positive, followed by strongly negative values in 1995 (Fig.
406 8A). Subsequently, the summer and annual NAO index remained weakly negative between 1999 and 2012, with
407 values becoming increasingly negative in the final five years of this period (Fig. 8A). In 2013, the NAO index
408 became strongly positive, particularly during summer, and values were also positive in 2015 and 2016 (Fig. 8A).
409 The AO index follows an overall similar pattern to the NAO until ~2000, although shifts are less distinct: the
410 index is generally negative until 1988, followed by five years of more positive values. In the 2000s, the AO index
411 fluctuates between positive and negative, and more negative summer values are observed in 2009, 2011, 2014 and
412 2015 (Fig. 8B). The AMO was generally negative from 1970 – 2000, although values fluctuated and were positive
413 around 1990 (Fig. 8C). Subsequently, the AMO entered a positive phase from 2000 onwards (Fig. 8C).

414 At the broad spatial scale, data indicate that surface ocean temperatures have warmed in the Barents Sea over time
415 (Fig. 9). Warming was particularly marked in the area extending approximately 100 km offshore of the Barents
416 Sea coast and south of 76 °N. Here, temperatures ranged between 2 and 4 °C in 1971-1980 and reached up to 7
417 °C by 2001-2012 (Fig. 9), although it should be noted that data are much sparser for the latter period. The Kara
418 Sea also warmed over the study period, with temperatures increasing from 0-2 °C in 1971-1980, to 4-5 °C in
419 2001-2012 (Fig. 9). Although input data are comparatively sparse for 2001-2012, it appears that ocean



420 temperatures have warmed in both the Barents and Kara Seas at each time step, suggesting there may be a broad
421 scale warming trend in the region. At 100 m depth, the data suggest that warmer ocean water extends substantially
422 during the study period, on both the Barents and Kara Sea coasts (Fig.9).

423 3.4. Glacier surging

424 During the study period, we observed three glaciers surging: ANU, MAS and SER (Fig. 1). These were excluded
425 from the analysis of glacier retreat rates and are discussed separately here. ANU has previously been identified as
426 possibly surge-type, based on the presence of looped-moraine (Grant et al., 2009). Here, we identify an active
427 surge phase, on the basis of a number of characteristics identified from satellite imagery and following the
428 classification of Grant et al. (2009): rapid frontal advance, heavy crevassing and digitate terminus. High flow
429 speeds are also evident close to the terminus (Melkonian et al., 2016), which is consistent with the active phase
430 of surging. Our results show that advance began in 2008 and was ongoing in 2015, with the glacier advancing 683
431 m during this period (Fig. 10). MAS was previously confirmed as surge-type (Grant et al., 2009) and our data
432 suggest that its active phase persisted between 1989 and 2007 (Fig. 10A). The imagery indicates that surging on
433 MAS originates from the eastern limb of the glacier, which may be partially fed by the neighbouring glacier (Figs.
434 10B & C). This ice appears to have impacted on the eastern margin of the main outlet of MAS, causing glacier
435 advance and heavy crevassing on the eastern portion of its terminus (Figs. 10B & C). This explanation is consistent
436 with the lack of signs of surge type behaviour on the western margin of MAS (Figs. 10B & C) and considerable
437 visible displacement of ice and surface features on the eastern tributary (Figs. 10B & C). SRE was also confirmed
438 as a surge-type glacier by Grant et al. (2009), who suggested that glacier advance occurred between 1976/77 and
439 2001. Our results indicate that advance began somewhat later, sometime between July 1983 and July 1986, and
440 ended before August 2000 (Fig. 10A).

441 4. Discussion

442 4.1. Spatial patterns of glacier retreat

443 Our results demonstrate that retreat rates on marine terminating outlet glaciers (-46.9 m a^{-1}) were more than three
444 times higher than those on land (-13.8 m a^{-1}) between 1986 and 2015 (Fig. 2). This is consistent with previous,
445 shorter-term studies from Greenland (Moon and Joughin, 2008; Sole et al., 2008) and Svalbard (Dowdeswell et
446 al., 2008), which demonstrated an order of magnitude difference between marine- and land-terminating glaciers.
447 It also confirms that the differences in retreat rates, relating to terminus type, observed between 1992 and 2010
448 on NVZ (Carr et al., 2014) persist at multi-decadal timescales. Recent results suggest that marine-terminating
449 glacier retreat and/or ice tongue collapse can cause dynamic thinning in the RHA (Melkonian et al., 2016; Willis
450 et al., 2015), meaning that these long-term differences in retreat rates may lead to substantially higher thinning
451 rates in marine-terminating basins, at multi-decadal timescales. The Russian High Arctic is forecast to be the third
452 largest source of ice volume loss by 2100, outside of the ice sheets (Radić and Hock, 2011). However, these
453 estimates only account for surface mass balance, and not ice dynamics, meaning that they may underestimate 21st
454 Century ice loss for the RHA. Consequently, dynamic changes associated with marine-terminating outlet glacier
455 retreat on NVZ need to be taken into account, in order to accurately forecast its near-future ice loss and sea level
456 rise contribution.



457 Our data showed no significant difference in total retreat rates for marine-terminating (-46.9 m a^{-1}) and lake-
458 terminating glaciers (-49.1 m a^{-1}). This contrasts with results from Patagonia, which were obtained during a similar
459 time period (mid-1980s to 2001/11) and showed that lake-terminating outlet glaciers retreated significantly more
460 rapidly than those ending in the ocean (Sakakibara and Sugiyama, 2014). For example, marine-terminating outlets
461 retreat at an average rate of -37.8 m a^{-1} between 2000 and 2010/11, whereas lake-terminating glaciers receded at
462 -80.8 m a^{-1} (Sakakibara and Sugiyama, 2014). Lake-terminating glacier retreat on NVZ also differs from
463 Patagonia, in that retreat rates are remarkably consistent between individual glaciers and remained similar over
464 time (Figs. 4 & 5). Conversely, frontal position changes in Patagonia showed major spatial variations and retreat
465 rates on several lake-terminating glaciers changed substantially between the two halves of the study period (mid-
466 1980's – 2000 and 2000-2010/11) (Sakakibara and Sugiyama, 2014).

467 One potential explanation for the common behaviour of the lake-terminating outlet glaciers on NVZ is that retreat
468 may be dynamically controlled and sustained by a series of feedbacks, once it has begun. As observed on large
469 Greenlandic tidewater glaciers, initial retreat may bring the terminus close to floatation, leading to faster flow and
470 thinning, which promote further increases in calving and retreat (e.g. Howat et al., 2007; Hughes, 1986; Joughin
471 et al., 2004; Meier and Post, 1987; Nick et al., 2009). This has been suggested as a potential mechanism for the
472 rapid recession for Upsala Glacier in Patagonia (Sakakibara and Sugiyama, 2014) and Yakutat Glacier, Alaska
473 (Trüssel et al., 2013). However, rapid retreat was not observed on all lake-terminating glaciers in Patagonia
474 (Sakakibara and Sugiyama, 2014) and the potential for these feedbacks to develop depends on basal topography
475 (e.g. Carr et al., 2015; Porter et al., 2014; Rignot et al., 2016). Consequently, the basal topography would need to
476 be similar for each of the NVZ glaciers to explain the very similar retreat patterns, which is not implausible, but
477 perhaps unlikely. Alternatively, it may be that the proglacial lakes act as a buffer for atmospheric warming, due
478 the greater thermal conductivity of water relative to air, and so reduce variability in retreat rates. Furthermore,
479 lake-terminating glaciers are not subject to variations in sea ice and ocean temperatures, which may account for
480 their more consistent retreat rates, compared to marine-terminating glaciers (Figs. 4 & 5). In order to differentiate
481 between these two explanations, data on lake temperature changes during the study period, and lake bathymetry
482 would be required. However, neither are currently available and we highlight this as an important area for further
483 research, given the rapid recession observed on these lake-terminating glaciers.

484 For the period between 1986 and 2015, we find no significant difference in retreat rates between the Barents and
485 Kara Sea coasts (Fig. 2). This is contrary to the results of a previous, shorter-term study, which showed that retreat
486 rates on the Barents Sea coast were significantly higher than on the Kara Sea between 1992 and 2010 (Carr et al.,
487 2014) and the higher thinning rates observed on marine outlets on the Barents Sea coast (Melkonian et al., 2016).
488 Furthermore, there are substantial differences in climatic and oceanic conditions on the two coasts (Figs. 4 & 7)
489 (Pfirman et al., 1994; Politova et al., 2012; Przybylak and Wyszynski, 2016; Zeeberg and Forman, 2001), so we
490 would expect to see significant differences in outlet glacier retreat rates. This indicates that longer-term glacier
491 retreat rates on NVZ may relate to much broader, regional scale climatic change, which is supported by the
492 widespread recession of glaciers across the Arctic during the past two decades (e.g. Blaszczyk et al., 2009; Carr
493 et al., 2014; Howat and Eddy, 2011; Jensen et al., 2016; Moon and Joughin, 2008). One potential overarching
494 control on NVZ frontal positions are fluctuations in the North Atlantic Oscillation (NAO), which covaries with
495 northern hemisphere air temperatures, Arctic sea ice and North Atlantic ocean temperatures (Hurrell, 1995;



496 Hurrell et al., 2003; IPCC, 2013). More recent work has also recognised the influence of the Atlantic Multidecadal
497 Oscillation (AMO) on oceanic and atmospheric conditions in the Barents Sea, and broader north Atlantic
498 (Drinkwater et al., 2013; Oziel et al., 2016). Our data suggest that the major phases of frontal position change on
499 NVZ correspond to changes the NAO and AMO (Fig. 8; Section 4.2.): rapid retreat between 2000-2013 coincides
500 with a weakly negative NAO and positive AMO, following almost three decades characterised by a generally
501 positive NAO and negative AMO (Fig. 8). As such, these large-scale changes may overwhelm smaller-scale
502 spatial variations between the two coasts of NVZ, when retreat is considered on multi-decadal time frames.

503 Marine-terminating outlet glacier retreat rates do not show a linear relationship latitude and there is considerable
504 scatter when the two variables are regressed (Fig. 3). This may be due to the influence of fjord geometry on glacier
505 response to climatic forcing (Carr et al., 2014) and the capacity for warmer ocean waters to access the calving
506 fronts. In contrast, southerly land-terminating outlets retreat more rapidly than those in the north, which we
507 attribute to the substantial latitudinal air temperature gradient on NVZ (Zeeberg and Forman, 2001). Conversely,
508 lake-terminating glaciers retreat more rapidly at more northerly latitudes (Fig. 3), which we speculate may relate
509 to the bathymetry and basal topography of individual glaciers, but data are not currently available to confirm this.

510 **4.2. Temporal patterns**

511 Our results show that retreat rates on marine-terminating outlet glaciers on NVZ were significantly higher between
512 2000 and 2013 than during the preceding 27 years (Fig. 4). At the same time, land-terminating outlets experienced
513 much lower retreat rates and did not change significantly during the study period (Figs. 4 & 5). This is consistent
514 with studies from elsewhere in the Arctic, which identified the 2000s as a period of elevated retreat on marine-
515 terminating glaciers (e.g. Blaszczyk et al., 2009; Howat and Eddy, 2011; Jensen et al., 2016; Moon and Joughin,
516 2008) and increasing ice loss (e.g. Gardner et al., 2013; Lenaerts et al., 2013; Moholdt et al., 2012; Nuth et al.,
517 2010; Shepherd et al., 2012). As discussed above, recent evidence suggests that glacier retreat in the Russian High
518 Arctic can trigger substantial dynamic thinning and ice acceleration (Melkonian et al., 2016; Willis et al., 2015),
519 but it not currently incorporated into predictions of 21st century ice loss from the region (Radić and Hock, 2011).
520 Consequently, the period of higher retreat rates during the 2000s may have a much longer-term impact on ice
521 losses from NVZ, and this needs to be quantified and incorporated into forecasts of ice loss and sea level rise
522 prediction.

523 Within the decadal patterns of glacier retreat, we observe clusters in the timing of significant changes in marine-
524 terminating glacier retreat rates (Fig. 6). Specifically, we see breaks in the frontal position time series on both the
525 Barents and Kara Sea coasts, beginning in the early 1990s, ~2000 and the mid-2000s (Fig. 6). This demonstrates
526 some synchronicity in changes in glacier behaviour around NVZ, although it is not ubiquitous (Fig. 6). The timing
527 of these changes coincides with those observed in Greenland, where the onset of widespread retreat and
528 acceleration in south-east Greenland began in ~2000 (e.g. Howat et al., 2008; Moon and Joughin, 2008; Seale et
529 al., 2011), and occurred from the mid-2000s onwards in the north-west (e.g. Carr et al., 2013b; Howat and Eddy,
530 2011; Jensen et al., 2016; McFadden et al., 2011; Moon et al., 2012). Whilst these changes could be coincidental,
531 they may also relate to broad, regional-scale changes observed in the North Atlantic region during the 2000s
532 (Beszczyńska-Möller et al., 2012; Hanna et al., 2013; Hanna et al., 2012; Holliday et al., 2008; Sutherland et al.,
533 2013). Data demonstrate that the NAO was weakly negative from the mid-1990s until 2012, in contrast to strongly
534 positive conditions in the late 1980s and early 1990s, and the AMO was persistently positive from 2000 onwards,



535 following three decades of overall positive conditions (Fig. 8). These changes coincide with increases in glacier
536 retreat rates, sea ice decline and atmospheric warming in NVZ between 2000 and 2013 (Figs. 4 & 7).

537 Between the 1950s and mid-1990s, positive phases of the NAO were associated with the influx of warm Atlantic
538 Water into the Barents Sea (Hurrell, 1995; Loeng, 1991) and increased penetration of Atlantic cyclones and air
539 masses into the region, which lead to elevated air temperatures and precipitation (Zeeberg and Forman, 2001).
540 Conversely, negative NAO phases were associated with cooler oceanic and atmospheric conditions in the Barents
541 Sea (Zeeberg and Forman, 2001). During this period, therefore, the impact of the NAO was opposite in the Barents
542 Sea and in western portions of the Atlantic-influenced Arctic (e.g. the Labrador Sea) (Drinkwater et al., 2013;
543 Oziel et al., 2016). However, since the mid-1990s, changes in the Barents Sea and the western Atlantic Arctic
544 have been in phase, and warming and sea ice reductions have been widespread across both regions (Drinkwater
545 et al., 2013; Oziel et al., 2016). As such, increased glacier retreat rates on NVZ during the 2000s (Figs 4 & 5) may
546 have resulted from the switch to a weaker, and predominantly negative, NAO phase from the mid-1990s (Fig. 8),
547 which would promote warmer air and ocean temperatures, and reduced sea ice, as we observe in our data (Figs. 4
548 & 7). Previous studies have suggested a 3-5 year lag between NAO shifts and changes in conditions on NVZ, due
549 to the time required for Atlantic Water to transit into the Barents Sea (Belkin et al., 1998; Zeeberg and Forman,
550 2001), which is consistent with the onset of retreat in ~2000 (Figs. 4 & 8). However, it has recently been suggested
551 that the NAO's role may have reduced since the mid-1990s, and that the AMO may be the dominant influence on
552 warming in the North Atlantic (Drinkwater et al., 2013; Oziel et al., 2016). The AMO is thought to promote
553 blocking of high-pressure systems by westerly winds, which changes the wind field (Häkkinen et al., 2011). This
554 allows warm water to penetrate further into the Barents and other Nordic Seas, leading to atmospheric and oceanic
555 warming during periods with a weakly negative NAO (Häkkinen et al., 2011). As such, rapid retreat on NVZ
556 between 2000 and 2013 may have resulted from the combined effects of a weaker, more negative NAO from the
557 mid-1990s and a more positive AMO from 2000 onwards (Fig. 8). This suggests that synoptic climatic patterns
558 may be an important control on glacier retreat rates on NVZ and that the recent relationship between the NAO
559 and glacier change on NVZ contrasts with that observed during the 20th century (Zeeberg and Forman, 2001).

560 Following higher retreat rates in the 2000's, our data indicate that marine-terminating glacier retreat slowed from
561 2013 onwards on both the Barents and Kara Sea coasts, with several glaciers beginning to re-advance (Figs. 4 &
562 5). Our data demonstrate that marine-terminating glaciers on NVZ have previously undergone a step-like pattern
563 of retreat, with short (1-2 year) pauses in retreat (Fig. 5). Thus, it is unclear whether this reduction in retreat rates
564 is another temporary pause, before continued retreat, or the beginning of a new phase of reduced retreat rates. One
565 possible explanation for reduced retreat rates on both coasts of NVZ are the stronger NAO values observed from
566 the late 2000s onwards: winter 2009/10 had the most negative NAO for 200 years (Delworth et al., 2016; Osborn,
567 2011) and values were strongly positive in 2013 (Fig. 8A). This is consistent with the 3 to 5 year lag required for
568 NAO-related changes in Atlantic Water inflow to reach NVZ (Zeeberg and Forman, 2001) and so we speculate
569 that reduced glacier retreat rates from 2013 onwards (Figs. 4 & 5) may relate to an increase in the influence of the
570 NAO, relative to the AMO, from the late 2000s (Fig.8). Evidence indicates that the impact of the NAO in the
571 Barents Sea is now in-phase with the western North Atlantic (Drinkwater et al., 2013; Oziel et al., 2016), and so
572 a more positive NAO could lead to cooler conditions on NVZ, and hence glacier advance. However, the
573 relationship between large-scale features, such as the NAO and AMO, ocean conditions and glacier behaviour is



574 complex (Drinkwater et al., 2013; Oziel et al., 2016) and the period of glacier advance / reduced retreat on NVZ
575 has lasted only two years. Consequently, further monitoring is required to determine whether this represents a
576 longer-term trend, or a short-term change, and to confirm its relationship to synoptic climatic patterns.

577 Despite the changes in the NAO and AMO, our data show no significant change in sea ice concentrations, nor the
578 length of the ice free season, between 2000-2012 and 2013-2015 on either the Barents Sea or Kara Sea coast
579 (Table 4; Fig. 7). Likewise, we see no significant change in winter (Jan-Mar) air temperatures at Im. K. Fedorova
580 (Table 3; Fig. 4) nor in the ERA-Interim data during any season (Table 3; Fig. 4). Although not significant, we see
581 summer warming of 0.7 °C (surface) and 0.8 °C (850 m pressure level) in the ERA-Interim data (Fig. 4), which
582 is the opposite of what we would expect if reductions air temperatures and surface melt were driving the slow-
583 down in retreat rates. As such, reduced retreat rates do not seem to be directly linked to short-term changes in sea
584 ice or air temperatures. They are also unlikely to result from changes in surface mass balance, as the response
585 time of NVZ glaciers is likely to be slow: they have long catchments (~40km), slow flow speeds (predominantly
586 <200 m a⁻¹ (Melkonian et al., 2016)) and are likely to be polythermal. Furthermore, thinning rates between 2012
587 and 2013/14 averaged 0.4 m a⁻¹ across the ice cap and reached up to 5 m a⁻¹ close to the glacier termini (Melkonian
588 et al., 2016), meaning that even a positive surface mass balance is very unlikely to deliver sufficient ice, quickly
589 enough, to promote advance and/or substantially lower retreat rates. Instead, this may be a response to oceanic
590 changes, which we cannot detect from available data, a lagged response and/or relate to more localised, glacier
591 specific factors. We suggest that the latter is unlikely, given the widespread and synchronous nature of the
592 observed reduction in retreat rates (Figs. 4 & 5). Future work should monitor retreat rates, to determine whether
593 reduced retreat is persistent, or is a short-term interruption to overall glacier retreat, and collect more extensive
594 oceanic data, to assess its impact on this change.

595 Although we observe some common behaviour, in terms of the approximate timing and general trend in retreat,
596 there is still substantial variability in the magnitude of retreat between individual marine-terminating glaciers
597 (Figs. 4 & 5). Furthermore, not all glaciers shared common changepoints and certain outlets showed a different
598 temporal pattern of retreat to the majority of the study population (Figs. 4-6). For example, INO retreated more
599 slowly between 1989 and 2006 than during the 1970s and 1980's. We attribute these differences to glacier-specific
600 factors, and, in particular, the fjord bathymetry and basal topography of individual glaciers. Previous studies have
601 highlighted the impact of fjord width on retreat rates on NVZ (Carr et al., 2014) and basal topography on marine-
602 terminating glacier behaviour elsewhere (e.g. Carr et al., 2015; Porter et al., 2014; Rignot et al., 2016). This may
603 result from the influence of fjord geometry on the stresses acting on the glacier, once it begins to retreat: as a fjord
604 widens, lateral resistive stresses will reduce and the ice must thin to conserve mass, making it more vulnerable to
605 calving (Echelmeyer et al., 1994; Raymond, 1996; van der Veen, 1998a & b), whilst retreat into progressively
606 deeper water can cause feedbacks to develop between thinning, floatation and retreat (e.g. Joughin and Alley,
607 2011; Joughin et al., 2008; Schoof, 2007). Thus, retreat into a deeper and/or wider fjord may promote higher
608 retreat rates on a given glacier, even with common climatic forcing. In addition, differences in fjord bathymetry
609 may determine whether warmer Atlantic Water can access the glacier front (Porter et al., 2014; Rignot et al.,
610 2016), which could promote further variations between glaciers. This highlights the need to collect basal
611 topographic data for NVZ outlet glaciers, which it is currently very limited, but a potentially key control on ice
612 loss rates.



613 **4.3. Climatic and oceanic controls**

614 Our data demonstrate that air temperatures were very substantially warmer between 2000 and 2012 than during
615 the preceding decades, and that sea ice concentrations were also much lower on both the Barents and Kara Sea
616 coasts during this period (Figs. 4 and 7). This is consistent with the atmospheric warming reported across the
617 Arctic during the 2000s (e.g. Carr et al., 2013a; Hanna et al., 2013; Hanna et al., 2012; Mernild et al., 2013) and
618 the well-documented decline in Arctic sea ice (Comiso et al., 2008; Kwok and Rothrock, 2009; Park et al., 2015).
619 As such, the decadal patterns of marine-terminating outlet glacier retreat correspond to decadal-scale climatic
620 change on NVZ (Figs. 4 & 7), and exceptional retreat during the 2000s coincided with significantly warmer air
621 temperatures and lower sea ice concentrations (Tables 2 & 3). Interestingly, step-changes in the air temperature
622 and sea ice data identified by the changepoint analysis did not correspond to significant changes in outlet glacier
623 retreat rates (Fig. 6), suggesting that such changes may not substantially influence retreat rates, or that the
624 relationship may be more complex, e.g. due to lags in glacier response.

625 The much lower retreat rates on land-terminating outlets (Fig. 4) may indicate an oceanic driver for retreat rates
626 on marine-terminating glaciers. Previous studies identified sea ice loss as a potentially important control on NVZ
627 retreat rates (Carr et al., 2014), which fits with observed correspondence between sea ice loss and retreat, but it is
628 unclear whether the two variables simply co-vary, or whether sea ice can drive ice loss, by extending the duration
629 of seasonally high calving rates (e.g. Amundson et al., 2010; Miles et al., 2013; Moon et al., 2015). The available
630 ocean data indicate that temperatures were substantially warmer during the 2000s (Fig. 9), which would provide
631 a plausible mechanism for widespread retreat on both coasts of NVZ (Fig 4). However, oceanic data for the 2000s
632 is sparse in the Barents and Kara Seas, compared to previous decades, so it is difficult to ascertain the magnitude
633 and spatial distribution of warming, and to link it directly with glacier retreat patterns. Lake-terminating glaciers
634 are not affected by changes in sea ice or ocean temperatures, but could be influenced by air temperatures.
635 However, despite much higher air temperatures in the 2000s, mean retreat rates on lake-terminating outlet glaciers
636 were similar for each decade of the study (Fig. 4), suggesting that the relationship is not straightforward. Instead,
637 the presence of lakes may at least partly disconnect these glaciers from climatic forcing, by buffering the effects
638 of air temperatures changes and/or by sustaining dynamic changes, following initial retreat (Sakakibara and
639 Sugiyama, 2014; Trüssel et al., 2013).

640 **4.4. Glacier Surging**

641 During the study period, we identify three actively surging glaciers, based on various lines of glaciological and
642 geomorphological evidence (Copland et al., 2003; Grant et al., 2009), including terminus advance (Fig. 10).
643 Frontal advance persisted for 18 years on ANU and 15 years on SER, respectively, whilst ANU began to advance
644 in 2008 and this continued until the end of the study period (Fig. 10). This is comparatively long for surge-type
645 glaciers, which usually undergo short active phases over timeframes of months to years (Dowdeswell et al., 1991;
646 Raymond, 1987). For comparison, surges on Tunabreen, Spitzbergen, last only ~2 years (Sevestre et al., 2015)
647 and Basin 3 on Austfonna underwent major changes in its dynamic behaviour in just a few years (Dunse et al.,
648 2015). Surges elsewhere can occur even more rapidly: the entire surge cycle of Variegated Glacier in Alaska takes
649 approximately 1-2 decades and the active phase persists for only a few months (e.g. Bindschadler et al., 1977;
650 Eisen et al., 2005; Kamb, 1987; Kamb et al., 1985; Raymond, 1987). Furthermore, the magnitude of advance on
651 these three glaciers is in the order of a few hundred meters, which is smaller than advances associated with surges



652 on Tunabreen (1.4 km) and Kongsvegen (2 km) (Sevestre et al., 2015) and much less than the many kilometres of
653 advance observed on Alaskan surge-type glaciers, such as Variegated Glacier (Bindschadler et al., 1977; Eisen et
654 al., 2005). Consequently, the active phase on NVZ appears to be long, in comparison to other regions and terminus
655 advance is more limited, which may provide insight into the mechanism(s) driving surging here and may indicate
656 that these glaciers are located towards one end of the climatic envelope required for surging in the Arctic (Sevestre
657 and Benn, 2015).

658 During the active phase of the NVZ surge glaciers, we observe large sediment plumes emanating from the glacier
659 terminus (Fig. 9), which indicates that at least part of the glacier bed is warm-based during the surge. Together
660 with the comparatively long surge interval, this supports the idea that changes in thermal regime may drive glacier
661 surging on NVZ, as hypothesised for certain Svalbard glaciers (Dunse et al., 2015; Murray et al., 2003; Sevestre
662 et al., 2015). In addition, the surge of MAS appears to have been triggered by a tributary glacier surging into its
663 lateral margin (Fig. 9). This demonstrates an alternative mechanism for surging, aside from changes in the thermal
664 regime and/or hydrology conditions of the glacier, which has not been widely observed, but will depend strongly
665 on the local glaciological and topographical setting of the glacier. The data presented here focus only on frontal
666 advance and glaciological/geomorphological evidence, whereas information on ice velocities is also an important
667 indicator of surging (Sevestre and Benn, 2015). Consequently, information on velocity and surface elevation
668 changes are needed to further investigate the surge cycle and its possible controls on NVZ. This is important, as
669 NVZ is thought to have conditions that are highly conducive to glacier surging (Sevestre and Benn, 2015), but
670 has a long surge interval. We therefore want to ensure that we can disentangle surge behaviour and the impacts of
671 climate change on NVZ.

672 5. Conclusions

673 At multi-decadal timescales, terminus type remains a major, over-arching determinant of outlet glacier retreat
674 rates on NVZ. As observed elsewhere in the Arctic, land-terminating outlets retreated far more slowly than those
675 ending in the ocean. However, we see no significant difference in retreat rates between ocean- and lake-
676 terminating glaciers, which contrasts with findings in Patagonia. Retreat rates on lake-terminating glaciers were
677 remarkably consistent between glaciers and over time, which may result from the buffering effect of lake
678 temperature and/or the impact of lake bathymetry, which could facilitate rapid retreat that is largely independent
679 of climate forcing, after an initial trigger. We cannot differentiate between these two scenarios with currently
680 available data. Retreat rates on marine-terminating glaciers were exceptional between 2000 and 2013, compared
681 to previous decades. However, retreat slowed on the vast majority of ocean-terminating glaciers from 2013
682 onwards, and several glaciers advanced, particularly on the Barents Sea coast. It is unclear whether this represents
683 a temporary pause or a longer-term change, but it should be monitored in the future, given the potential for outlet
684 glaciers to drive dynamic ice loss from NVZ. The onset of higher retreat rates coincides with a more negative,
685 weaker phase of the NAO and a more positive AMO, whilst reduced retreat rates follow stronger NAO years. This
686 suggests that synoptic atmospheric and oceanic patterns may influence NVZ glacier behaviour at decadal
687 timescales. Marine-terminating glaciers showed some common patterns in terms of the onset of rapid retreat
688 (1990s, ~2000 and mid 2000s), but showed substantial variation in the magnitude of retreat, which we attribute to
689 glacier-specific factors. Glacier retreat corresponded with decadal-scale climate patterns: between 2000-2013, air
690 temperatures were significantly warmer than the previous decades and sea ice concentrations were significantly



691 lower. Available data indicate oceanic warming, which could potentially explain why retreat rates on marine-
 692 terminating glaciers far exceed those ending on land, but data are comparatively sparse from 2000 onwards,
 693 making their relationship to glacier retreat rate difficult to evaluate. The surge phase on NVZ glaciers appears to
 694 be comparatively long, and warrants further investigation, to separate its impact on ice dynamics from that of
 695 climate-induced change and to determine the potential mechanism(s) driving these long surges. Recent results
 696 suggest that outlet glaciers can trigger dynamic losses on NVZ, but these processes are not yet included in
 697 estimates of the region's contribution to sea level rise. As such, it is vital to determine the longer-term impacts of
 698 exceptional glacier retreat during the 2000s and to monitor the near-future behaviour of these outlets.

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- 948



| Characteristic | Category | Number of glaciers |
|----------------|-----------------------|--------------------|
| Coast | Barents Sea | 29 |
| | Kara Sea | 19 |
| Ice mass | Northern ice mass | 43 |
| | Subsidiary ice mass 1 | 4 |
| | Subsidiary ice mass 2 | 5 |
| Terminus type | Marine | 32 |
| | Lake | 6 |
| | Land | 15 |

949 **Table 1.** Number of outlet glaciers contained within each category used to assess spatial variations in retreat
 950 rate, specifically coast, ice mass and terminus type.

| | Barents Sea marine- terminating | Kara Sea marine- terminating | Land-terminating | Lake-terminating |
|---------------|---------------------------------------|---------------------------------|------------------|------------------|
| 76-86 / 86-00 | 0.440 | 0.538 | 0.982 | 0.486 |
| 76-86 / 00-13 | >0.001 | 0.018 | 0.085 | 0.686 |
| 76-86 / 13-15 | 0.008 | 0.497 | 0.945 | 0.686 |
| 86-00 / 00-13 | 0.001 | 0.008 | 0.223 | 0.886 |
| 86-00 / 13-15 | 0.001 | 0.935 | 0.909 | 0.886 |
| 00-13 / 13-15 | >0.001 | 0.009 | 0.597 | 0.686 |

951 **Table 2.** Wilcoxon test results, used to assess significant differences in retreat rates between each manually-
 952 identified time interval (1976-1986, 1986-2000, 2000-2013, 2013, 2015). Retreat rate data were tested
 953 separately for each terminus type, and marine-terminating glaciers were further sub-divided by coast. Following
 954 convention, p-values of <0.05 are considered significant and are highlighted in bold.

| Station | Time interval | Season | | | | Annual |
|-------------------|---------------|--------------|--------------|--------------|--------------|--------------|
| | | DJF | MAM | JJA | SON | |
| Im. E.K. Fedorova | 13-15 / 86-99 | 0.432 | | | | |
| Im. E.K. Fedorova | 13-15 / 76-85 | 0.937 | | | | |
| Im. E.K. Fedorova | 00-12 / 13-15 | 0.287 | | | | |
| Im. E.K. Fedorova | 00-12 / 86-99 | 0.011 | 0.643 | 0.043 | 0.008 | 0.013 |
| Im. E.K. Fedorova | 00-12 / 76-85 | 0.186 | 0.035 | 0.045 | 0.003 | 0.003 |
| Im. E.K. Fedorova | 86-99 / 76-85 | 0.188 | 0.089 | 0.704 | 0.495 | 0.828 |
| | | | | | | |
| Malye Karmakuly | 13-15 / 86-99 | | | | | |
| Malye Karmakuly | 13-15 / 76-85 | | | | | |
| Malye Karmakuly | 00-12 / 13-15 | | - | - | - | - |



| | | | | | | |
|-----------------------|---------------|--------------|--------------|--------------|--------------|------------------|
| Malye Karmakuly | 00-12 / 86-99 | 0.017 | 0.840 | 0.056 | 0.007 | 0.017 |
| Malye Karmakuly | 00-12 / 76-85 | 0.038 | 0.041 | 0.045 | 0.004 | >0.001 |
| Malye Karmakuly | 86-99 / 76-85 | 0.623 | 0.086 | 0.5977 | 0.673 | 0.212 |
| | | | | | | |
| ERA-Interim (surface) | 13-15 / 86-99 | 0.032 | 0.156 | 0.197 | 0.156 | 0.006 |
| ERA-Interim (surface) | 13-15 / 76-85 | 0.714 | 0.083 | 0.517 | 0.833 | 0.117 |
| ERA-Interim (surface) | 00-12 / 13-15 | 0.900 | 0.189 | 0.364 | 0.593 | 0.239 |
| ERA-Interim (surface) | 00-12 / 86-99 | 0.006 | 0.942 | 0.981 | 0.062 | 0.044 |
| ERA-Interim (surface) | 00-12 / 76-85 | 0.765 | 0.579 | 0.526 | 0.874 | 0.267 |
| ERA-Interim (surface) | 86-99 / 76-85 | 0.127 | 0.233 | 0.970 | 0.192 | 0.794 |
| | | | | | | |
| ERA-Interim (850 m) | 13-15 / 86-99 | 0.591 | 0.509 | 0.432 | 0.500 | 0.206 |
| ERA-Interim (850 m) | 13-15 / 76-85 | 0.548 | 0.383 | 0.833 | 0.733 | 0.383 |
| ERA-Interim (850 m) | 00-12 / 13-15 | 0.521 | 0.611 | 0.782 | 0.511 | 0.900 |
| ERA-Interim (850 m) | 00-12 / 86-99 | 0.062 | 0.752 | 0.058 | 0.041 | 0.004 |
| ERA-Interim (850 m) | 00-12 / 76-85 | 0.831 | 0.303 | 0.939 | 0.751 | 0.132 |
| ERA-Interim (850 m) | 86-99 / 76-85 | 0.149 | 0.433 | 0.433 | 0.146 | 0.576 |
| | | | | | | |

955 **Table 3.** P-values for Wilcoxon tests for significant differences in mean seasonal and mean annual air
 956 temperatures, for the periods 1976-1985, 1986-1999, 2000-2013, and 2013-2015. Following convention, p-
 957 values of <0.05 are considered significant and are highlighted in bold.

| Coast | Time interval | Season | | | | Ice-free months |
|---------|---------------|------------------|------------------|------------------|------------------|-----------------|
| | | JFM | AMJ | JAS | OND | |
| Barents | 13-15 / 86-99 | 0.003 | 0.012 | 0.003 | 0.003 | 0.003 |
| Barents | 13-15 / 76-85 | 0.067 | 0.017 | 0.017 | 0.017 | 0.017 |
| Barents | 00-12 / 13-15 | 0.704 | 0.296 | 0.039 | 0.057 | 0.086 |
| Barents | 00-12 / 86-99 | 0.002 | 0.009 | 0.019 | >0.001 | 0.001 |
| Barents | 00-12 / 76-85 | 0.006 | 0.002 | 0.002 | 0.001 | 0.002 |
| Barents | 86-99 / 76-85 | 0.279 | 0.080 | 0.218 | 0.179 | 0.213 |
| | | | | | | |
| Kara | 13-15 / 86-99 | 0.677 | 0.677 | 0.244 | 0.591 | 0.088 |
| Kara | 13-15 / 76-85 | 1 | 0.667 | 0.017 | 0.267 | 0.067 |
| Kara | 00-12 / 13-15 | 0.082 | 0.057 | 0.921 | 0.082 | 0.561 |
| Kara | 00-12 / 86-99 | >0.001 | >0.001 | >0.001 | >0.001 | 0.037 |
| Kara | 00-12 / 76-85 | >0.001 | >0.001 | >0.001 | >0.001 | 0.011 |



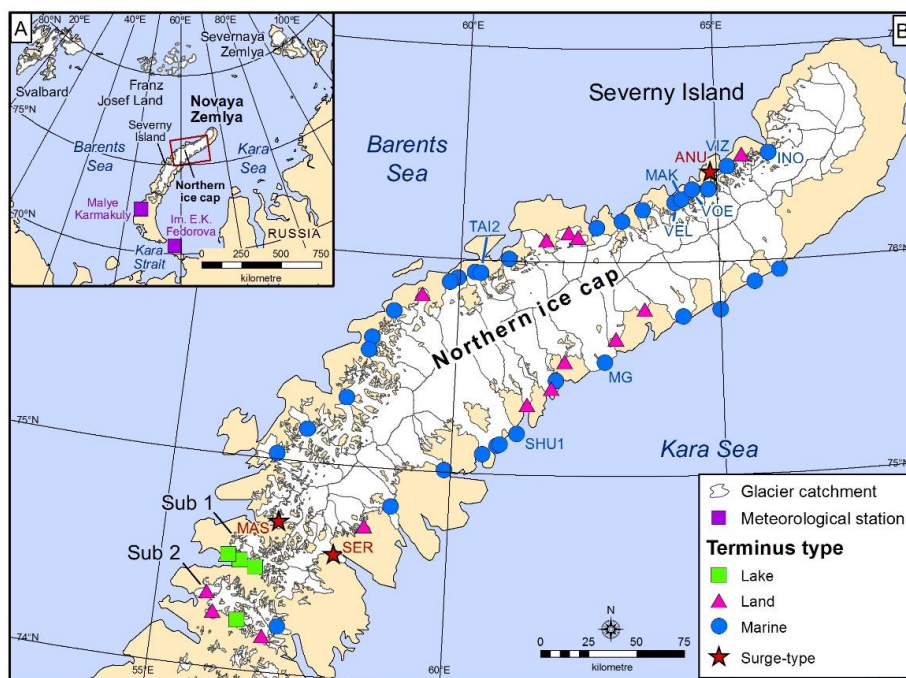
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|------|---------------|--------------|--------------|--------------|--------------|-------|
| Kara | 86-99 / 76-85 | 0.003 | 0.034 | 0.028 | 0.001 | 0.300 |
|------|---------------|--------------|--------------|--------------|--------------|-------|

958 **Table 4.** P-values for Wilcoxon tests for significant differences in mean seasonal sea ice concentrations and the
959 number of ice-free months, for the periods 1976-1985, 1986-1999 and 2000-2013. Following convention, p-
960 values of <0.05 are considered significant and are highlighted in bold.

961



962 **Figures**



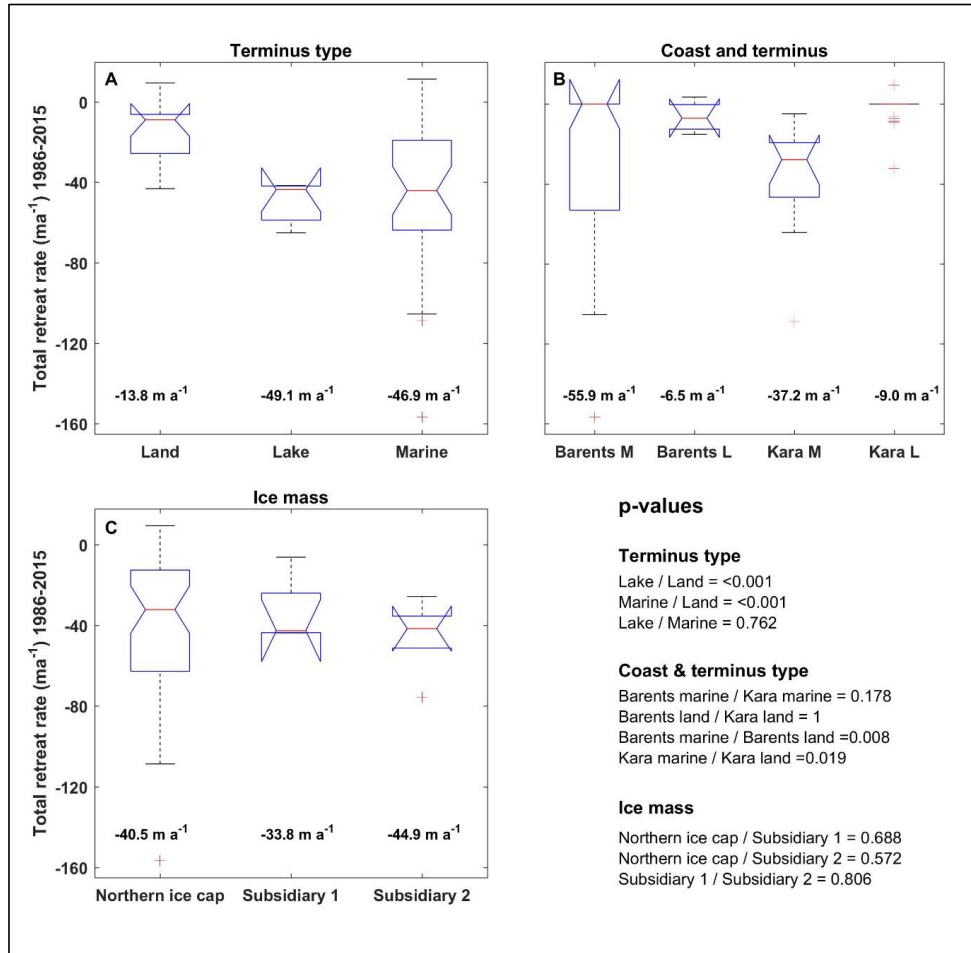
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964 **Figure 1:** Location map, showing the study area and outlet glaciers. A) Location of Novaya Zemlya, in relation
 965 to major land and water masses. Meteorological stations where air temperature data were acquired are indicated
 966 by a purple square. B) Study glacier locations and main glacier catchments (provided by G. Moholdt and available
 967 via GLIMS database). Glaciers are symbolised according to terminus type: marine terminating (blue circle); land-
 968 terminating (pink triangle); lake terminating (green square); and observed surging during the study period (red
 969 star). Glaciers observed to surge are: Anuchina (ANU), Mashigina (MAS), and Serp i Molot (SER).

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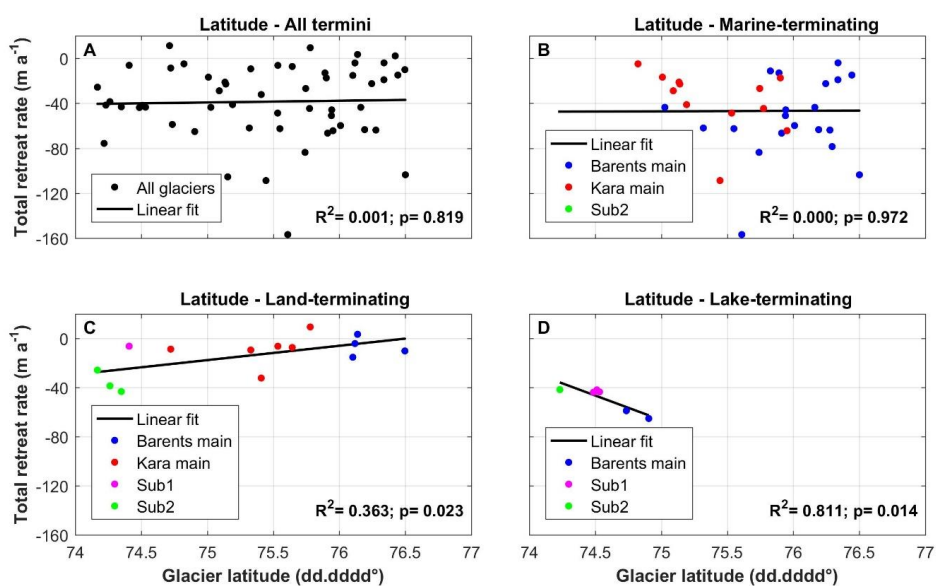


972

973 **Figure 2.** Box plots and Kruksal Wallis test results for different glacier terminus settings, for: A) terminus type;
 974 B) coast and terminus, L = land-terminating, m = marine-terminating; and C) ice mass, specifically the northern
 975 ice cap and subsidiary ice caps 1 and 2. See Figure 1 for ice cap locations. In all cases, total retreat rate (1986-
 976 2015) is used to test for significant differences between the classes. Mean total retreat rates for each class are
 977 given on each plot, below the associated box plot. For each box plot, the red central line represents the median,
 978 the blue lines the upper and lower quartile, red crosses are outliers (a value more than 1.5 times the interquartile
 979 range above / below the interquartile values) and the black lines are the whiskers, which extend from the
 980 interquartile ranges to the maximum values that are not classed as outliers. P-values for each Kruksal Wallis test
 981 are given on the right of the plot.

982

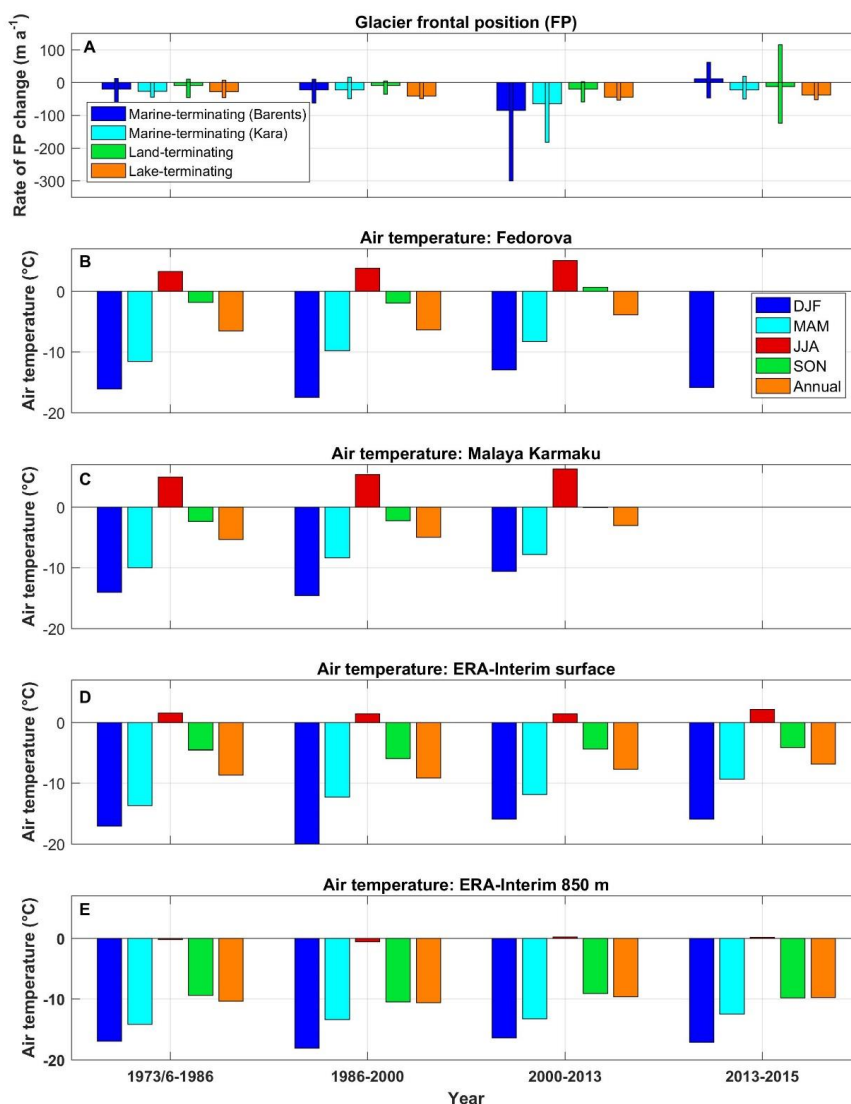
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984

985 **Figure 3.** Linear regression of total retreat rate (1986-2015) versus glacier latitude. Latitude was regressed against
 986 total glacier retreat rate for A) All outlet glaciers in the study sample; B) marine-terminating glaciers only; C)
 987 land-terminating glaciers only; D) lake-terminating glaciers only. In all cases, the linear regression line is shown,
 988 as are the associated R² and p-values. The R² value indicates how well the line describes the data and the p-value
 989 indicates the significance of the regression coefficients, i.e. the likelihood that the predictor and response variable
 990 are unrelated.

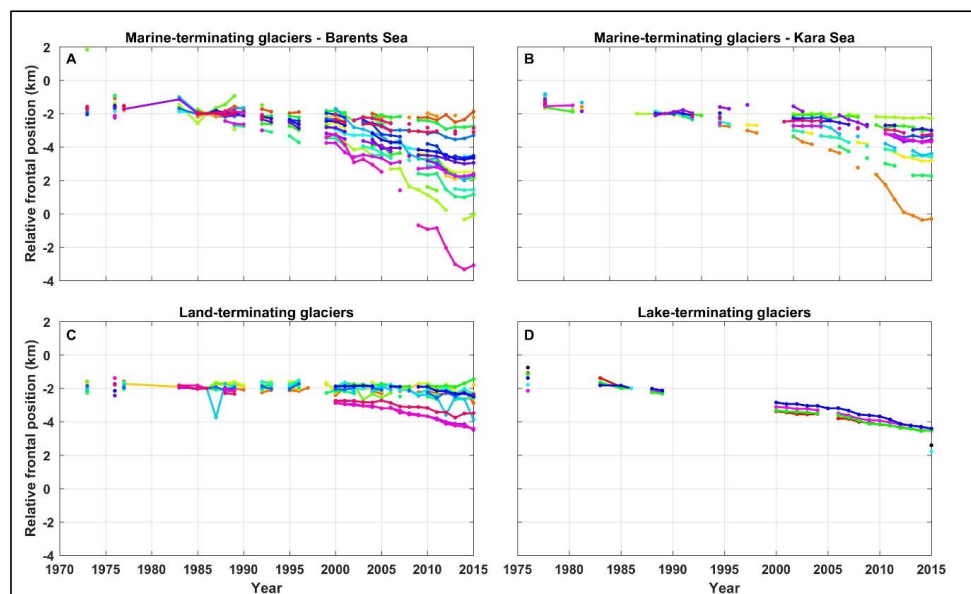
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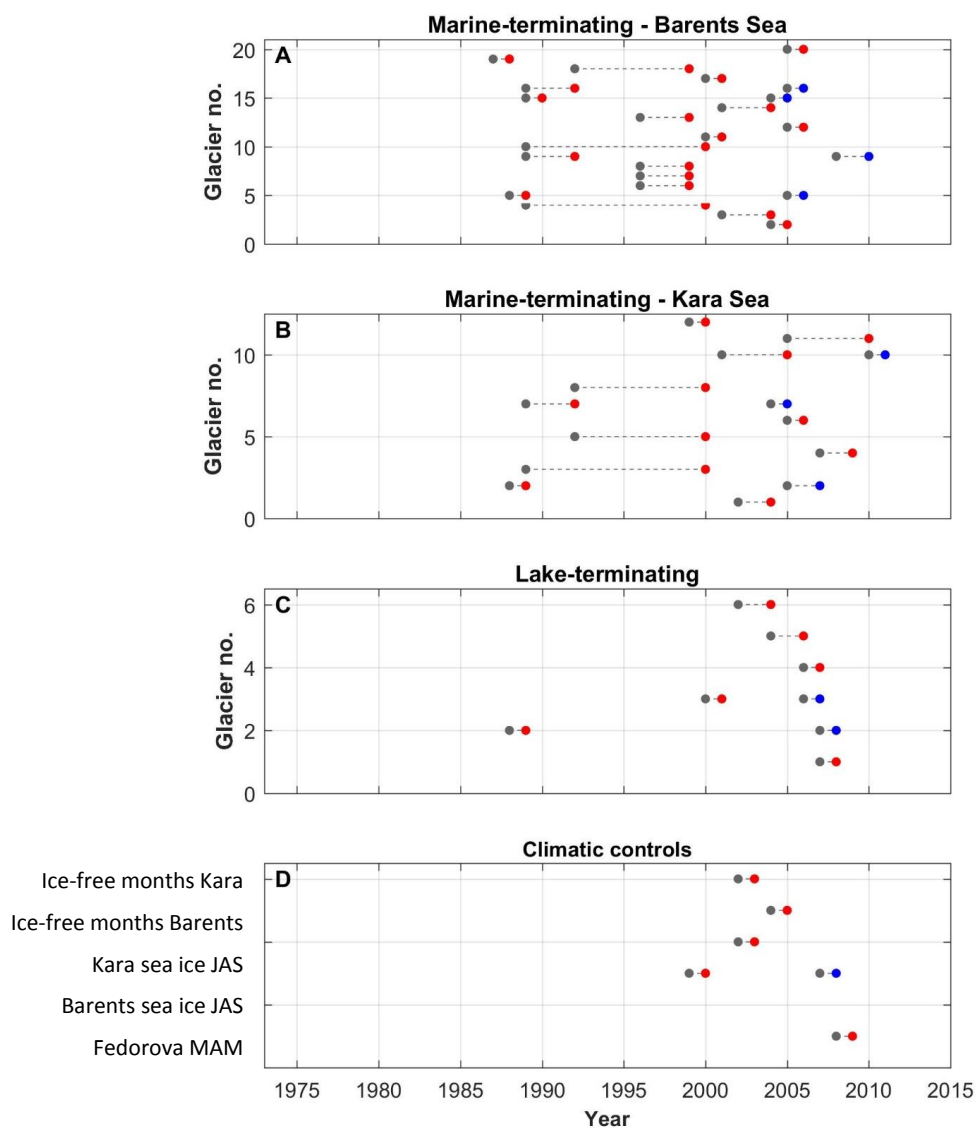
993 **Figure 4.** Mean retreat rates for Novaya Zemlya outlet glaciers, and mean air temperatures at Im. K. Fedorova
 994 and Malaya Karmaku (Fig. 1). Data are split into four time periods, based on manually identified breaks in the
 995 glacier retreat data: 1973/6-1986, 1986-2000, 2000-2013 and 2013-2015. A) Retreat rates were calculated
 996 separately for different terminus types and marine-terminating glaciers were further sub-divided into those
 997 terminating into the Barents Sea versus the Kara Sea. Wide bars represent mean values and thin bars represent the
 998 total range (i.e. minimum and maximum values) within each category. B-E) Mean seasonal air temperatures (Dec-
 999 Feb, Mar-May, Jun-Aug and Sep-Nov) and mean annual air temperatures for Im. K. Fedrova (B), Malaya
 1000 Karmaku (C), ERA-Interim surface (D) and ERA-Interim 850 m pressure level (E). Note that only mean values
 1001 for Im. K. Fedorova in Jan-Mar are calculated for 2013-2015, due to data availability.

1002



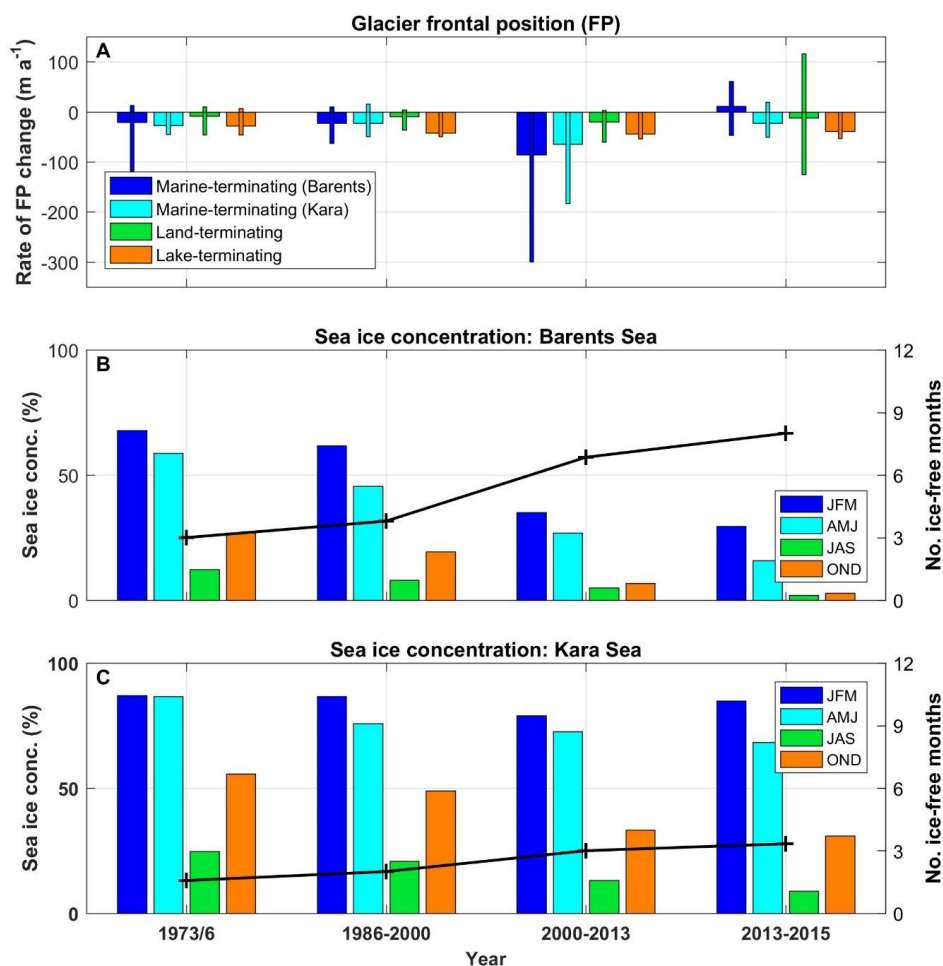
1003

1004 **Figure 5.** Relative glacier frontal position over time, from 1973 to 2015, for A) marine-terminating outlet
1005 glaciers on the Barents Sea coast; B) marine-terminating outlet glaciers on the Kara Sea coast; C) land-
1006 terminating outlet glaciers and D) Land-terminating outlet glaciers. Within each plot, frontal positions for each
1007 glacier are distinguished by different colours.



1008

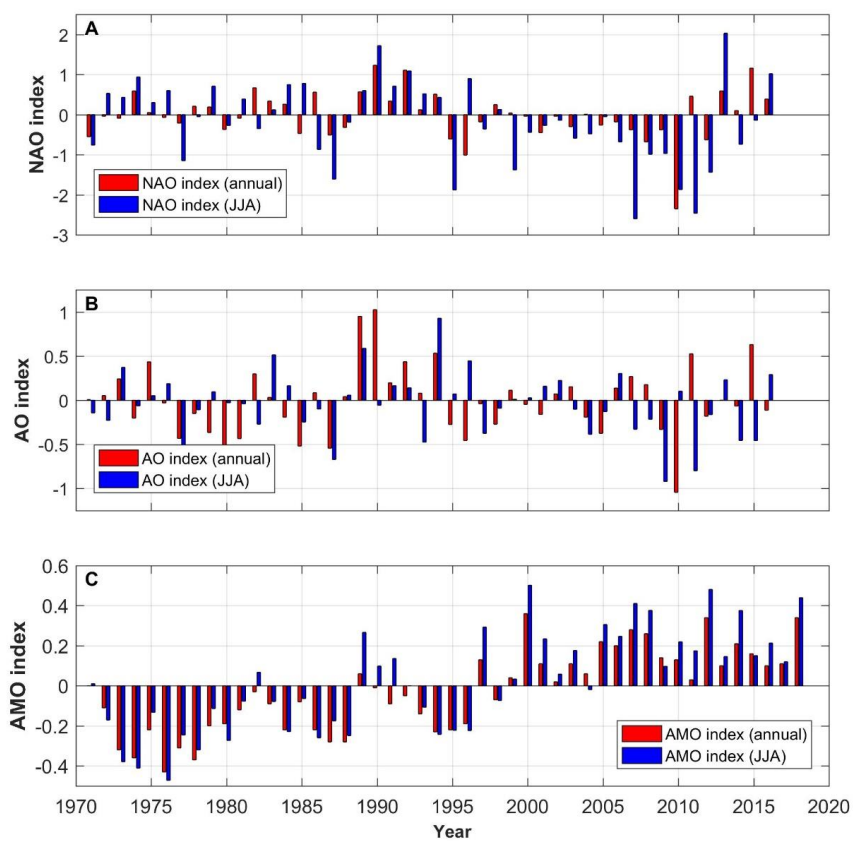
1009 **Figure 6.** Results of the changepoint analysis for glacier retreat rates and climatic controls. Red dots indicate the
 1010 start of a significantly different period in the time series data and grey dots represent the end of the previous
 1011 period, with grey dashed lines connecting the two. This is done to account for missing data: we know that the
 1012 changepoint occurred between the grey and the red dot, and that the new phase of behaviour occurred from the
 1013 red dot onwards, but not the exact timing of the change. Blue dots show the start of a second significant change
 1014 in the time series. Frontal position data were analysed separately for marine-terminating outlets on the Barents
 1015 Sea (A), Kara Sea (B) coasts and lake-terminating glaciers (C). D) Changepoint results for seasonal means in air
 1016 temperatures and sea ice, and the number of ice free months. Only climatic variables that demonstrated
 1017 changepoints are shown.



1018

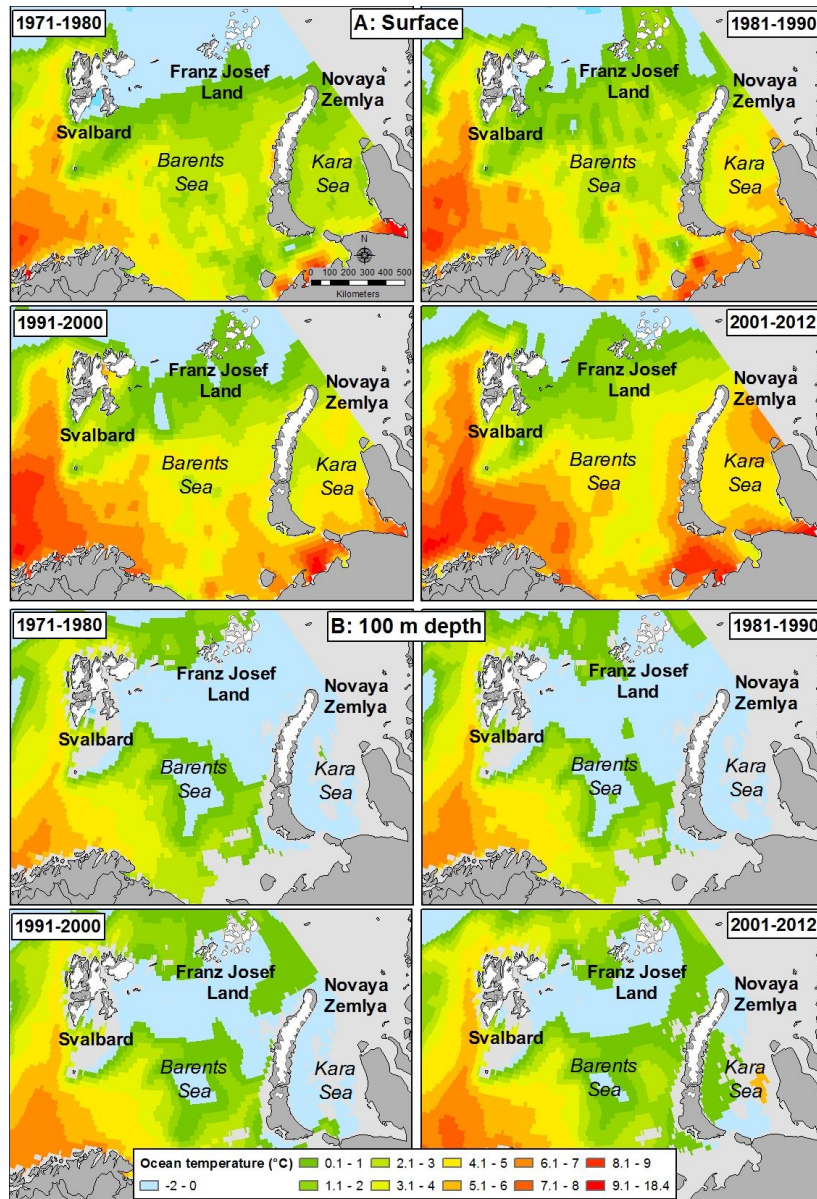
1019 **Figure 7.** Mean retreat rates for Novaya Zemlya outlet glaciers, and seasonal mean sea ice concentrations and
 1020 number of ice free months, for the Barents and Kara Sea coasts. Data are split into four time periods, based on
 1021 manually identified breaks in the glacier retreat data: 1973/6-1986, 1986-2000, 2000-2013 and 2013-2015. A)
 1022 Same as Fig. 4A. B & C) Mean seasonal sea ice concentrations (Jan-Mar, Apr-Jun, Jul-Sep and Oct-Dec) and
 1023 number of ice free months for the Barents Sea (B) and Kara Sea (C) coasts.

1024

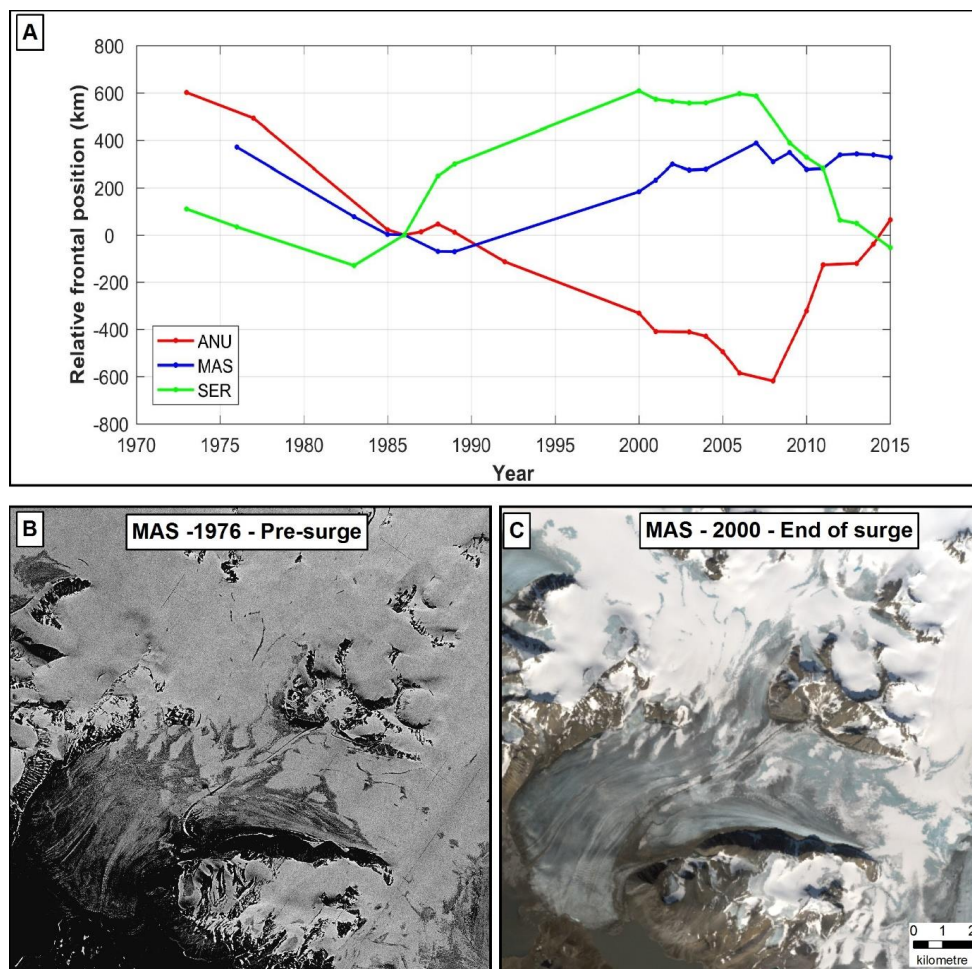


1025

1026 **Figure 8.** Time series of A) North Atlantic Oscillation (NAO); B) Arctic Oscillation (AO); and C) Atlantic
1027 Multidecadal Oscillation (AMO) for 1970 to 2016. In each case, mean annual and mean summer values are
1028 shown.



1029
 1030 **Figure 9.** Ocean temperatures from the ‘Climatological Atlas of the Nordic Seas and Northern North Atlantic’
 1031 (Korablev et al., 2014), at A) the surface and B) 100 m depth, for the following time intervals: 1971-1981, 1981-
 1032 1991, 1991-2000 and 2001-2012. These intervals were chosen, to match as closely as possible with the glacier
 1033 frontal position data and other datasets. Note that data coverage was substantially lower for 2001-2012, than
 1034 compared to other time periods. Further details on data coverage are available here:
 1035 <https://www.nodc.noaa.gov/OC5/nordic-seas/>.



1036

1037 **Figure 10.** Glaciers identified as surging during the study period, based on the surge criteria compiled by Grant
1038 et al. (2009). A) Glacier frontal position (relative to 1986) for glaciers identified as surge type: Anuchina
1039 (ANU), Mashigina (MAS), and Serp i Molot (SER). B) Pre-surge imagery of MAS. Imagery source: Hexagon,
1040 22nd July 1976. C) Imagery of MAS at the end of the surge. Imagery source: Landsat 7, 13th August 2000.