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1 **Multiple Late Holocene surges of a High-Arctic tidewater**
2 **glacier system in Svalbard**

3
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16
17 **Abstract:**

18 Most large tidewater glaciers in Svalbard are known to have surged at least once in the last few
19 hundred years. However, very little information exists on the frequency, timing or magnitude
20 of surges prior to the Little Ice Age (LIA) maximum in ~1900. We investigate the sediment-
21 landform assemblages produced by multiple advances of the Nathorstbreen glacier system
22 (NGS) in order to reconstruct its Late Holocene surge history. The glacier has recently
23 undergone one of the largest surges ever observed in Svalbard, advancing ~16 km from 2008

24 to 2016. We present flow velocities and ice-marginal observations (terminus change, proglacial
25 geomorphological processes) from the later stages of this surge. A first detailed assessment of
26 the development of a glaciotectionic mud apron within the fjord during a surge is provided.
27 Geomorphological and sedimentological examination of the terrestrial moraine areas formed
28 prior to the most recent surge reveals that at least two advances were responsible for their
29 formation, based on the identification of a previously unrecognised ice-contact zone recorded
30 by the distribution of sediment facies in coastal exposures. We distinguish between an outer,
31 older advance to the distal part of the moraine system and an inner, younger advance to a
32 position ~2 km upfjord. Radiocarbon dating of shells embedded in glaciotectionic composite
33 ridges formed by the onshore bulldozing of marine mud during the outer (older) of the two
34 advances shows that it occurred at some point during the interval 700-890 cal. yr BP (i.e. ~1160
35 AD), and not during the LIA as previously assumed. We instead attribute the inner (younger)
36 advance to the LIA at ~1890. By combining these data with previous marine geological
37 investigations in inner and outer Van Keulenfjorden, we demonstrate that NGS has advanced
38 at least four times prior to the recent 2008-2016 surge: twice at ~2.7 kyr BP, at ~1160 AD, and
39 in ~1890. This represents a unique record of the timing and magnitude of Late Holocene
40 tidewater glacier surges in Svalbard.

41

42 **Keywords:** glacier surge; glacial geomorphology; glaciology; Little Ice Age; Holocene;
43 Svalbard

44

45 **1. Introduction**

46 Marine-terminating or tidewater glaciers in the High-Arctic archipelago of Svalbard have
47 undergone accelerated mass loss in recent decades (Nuth et al., 2007, 2010; Błaszczyk et al.,
48 2009; Carr et al., 2017). Most of these glaciers are known to experience flow instabilities called

49 surges (Sevestre and Benn, 2015), whereby they periodically undergo rapid advances for short
50 periods of between three to ten years, before returning to a multi-decadal quiescent phase
51 characterised by frontal thinning and retreat (Murray et al., 2003; Sund et al., 2009; Sevestre
52 et al., 2018). Tidewater glacier surges result in rapid ice mass loss to the ocean and have a
53 significant impact on the climate, oceanography, sediment budget, and geomorphology of fjord
54 systems (e.g. Elverhøi et al., 1983; Hald et al., 2001; Plassen et al., 2004; Forwick et al., 2010).

55 The current state of knowledge on surging in Svalbard is largely based on observations
56 from satellite imagery since the 1970s. Glacier surges pre-dating this are usually identified
57 from aerial photographs (1930s onwards) or historical observations/written accounts (since the
58 Little Ice Age (LIA) maximum ~1900) of characteristic surge evidence, such as widespread
59 surface crevassing and/or rapid terminus advances (e.g. Liestøl, 1969; Hagen et al., 1993;
60 Bennett et al., 1999; Ottesen et al., 2008; Flink et al., 2015). Very little is known about surge
61 behaviour prior to the LIA maximum. In terms of general glacier behaviour at this time, some
62 land-terminating glaciers in Svalbard are known to have experienced multiple episodes of ice
63 expansion during the Late Holocene (i.e. since ~4 kyr BP) in response to a decline in summer
64 insolation (Miller et al., 2017). The timings of maximum advances of tidewater glaciers during
65 the Late Holocene display a large amount of variability across Svalbard. In some areas, such
66 as inner Isfjorden, the LIA is thought to represent the Holocene maximum position (e.g. Plassen
67 et al., 2004; Ottesen and Dowdeswell, 2006; Mangerud and Landvik, 2007). Other fjords record
68 much older, more extensive tidewater advances (Hald et al., 2004; Evans and Rea, 2005;
69 Kempf et al., 2013; Flink et al., 2017; Larsen et al., 2018). Many of these tidewater glaciers are
70 inferred or known to be of surge-type, raising the question as to whether previous advances
71 (LIA maximum or earlier) were glaciodynamic surges or were in response to climate forcing
72 (e.g. Farnsworth et al., 2017; Philipps et al., 2017; Streuff et al., 2017a).

73 The best way to determine possible surging that predates observational records is by
74 investigating sediment-landform assemblages, as surge behaviour is known to produce a
75 diagnostic suite of landforms (Evans and Rea, 1999; Ottesen et al., 2017). Numerous studies
76 have investigated englacial, geomorphological and sedimentological evidence exposed at the
77 receding margins of quiescent phase surge-type glaciers in Svalbard in order to better
78 understand the processes that occur during surges (e.g. Boulton et al., 1996, 1999; Glasser et
79 al., 1998a; Bennett et al., 1999; Christoffersen et al., 2005; Larsen et al., 2006; Ottesen et al.,
80 2008, 2017; Kristensen et al., 2009a,b; Lovell et al., 2015; Sobota et al., 2016; Larsen et al.,
81 2018; Lyså et al., 2018). For tidewater glacier surges, this evidence is typically recorded on the
82 sea floor (e.g. Solheim and Pfirman, 1985; Plassen et al., 2004; Ottesen et al., 2008, 2017;
83 Forwick et al., 2010; Flink et al., 2015, 2017; Streuff et al., 2015, 2017a; Burton et al., 2016;
84 Farnsworth et al., 2017). Field observations of the active phase of glacier surges are rare (e.g.
85 Glasser et al., 1998b; Murray et al., 1998; Kristensen and Benn, 2012), but are crucial for
86 linking surge processes to the geomorphological evidence left behind. This ensures
87 interpretations based on exposed basal ice, englacial debris-rich structures, and/or sediment-
88 landform assemblages (whether in marine or terrestrial positions) are more robust.

89 In order to contribute towards closing aforementioned gaps, we aim to reconstruct the
90 history of Late Holocene advances in Van Keulenfjorden recorded within marine and terrestrial
91 sediment-landform assemblages. We do this by investigating:

- 92 (1) active geomorphological processes at the glacier margin during a recent surge;
- 93 (2) the geomorphology of the terrestrial moraine areas in the inner fjord; and
- 94 (3) sediment facies exposed within the terrestrial moraine areas.

95 We combine this information with radiocarbon dating and previous marine geological
96 investigations in order to produce a revised chronology of glacier advances.

97

98 **2. Nathorstbreen glacier system (NGS) surge history**

99 The Nathorstbreen glacier system (NGS) consists of several major tributary glaciers and
100 currently terminates as a ~5 km-wide tidewater front in Van Keulenfjorden, a ~30 km-long
101 fjord in southern Spitsbergen (77°30.55'N, 15°57.67'E) (Fig. 1). Nathorstbreen is the central
102 and longest flow-unit (~39 km in 2017) in the system and drains from the accumulation area
103 of Ljosfonn at an elevation of up to 700 m a.s.l. Radio echo sounding data from the 1980s
104 indicate that NGS was over 300 m thick along its centre-line and warm based, with a cold
105 surface layer up to 200m thick. The cold layer intersected with the bed in the terminal zone at
106 this time (Dowdeswell et al., 1984). The other major NGS flow-units confluent with
107 Nathorstbreen are Polakkbreen and Zawadzki breen to the west and Dobrowolski breen to the
108 east. The system recently experienced the largest surge in Svalbard since the 1930s surges of
109 Bråsvellbreen and Negribreen (Liestøl, 1969; Hagen et al., 1993), advancing >15 km from
110 2008 onwards and expanding onto the two moraine areas Nordre and Søre Nathorstmorenen at
111 the lateral fjord margins (Fig. 1). Prior to this, the combined front of NGS (including
112 neighbouring Liestølbreen and Doktorbreen) is thought to have advanced to more-extensive
113 downfjord positions than reached in 2016 on at least three separate occasions: once during the
114 Little Ice Age (LIA) in ~1870-80 (Ottesen et al., 2008) and twice at ~2.7 kyr BP (Kempf et al.,
115 2013).

116

117 *2.1 Recent surge*

118 The recent NGS surge is described in detail in Sund et al. (2014) up until summer 2013, with
119 the key details briefly summarised here. The combined tidewater terminus began to advance
120 from inner Van Keulenfjorden sometime after October 2008 (Sund et al., 2014) (Fig. 1),
121 signalling the onset of the frontal advance phase of the surge (stage 3 in Sund et al., 2009).
122 Prior to this, surface measurements in 2000-01 showed that the upper basins of the four main

123 flow-units active during the surge (Nathorstbreen, Dobrowolskibreen, Polakkbreen and
124 Zawadzki breen) were recording velocities of up to 1 m d^{-1} , representing an acceleration of the
125 order of 100 times' quiescent phase velocities of $0.01\text{-}0.02 \text{ m d}^{-1}$ measured in 1992 (Sund et
126 al., 2014). This indicates that the early stages of surge initiation (stage 1 in Sund et al., 2009)
127 had begun in all contributing flow-units up to eight years before the start of the tidewater frontal
128 advance. Between 2003 and 2008, Dobrowolskibreen, Polakkbreen and Zawadzki breen were
129 all observed to thin in their upper basins and thicken at lower elevations, demonstrating that
130 the downglacier propagation of mass was underway (stage 2 in Sund et al., 2009). The onset
131 of stage 2 in Dobrowolskibreen coincided with the appearance of large transverse crevasses in
132 the upper basin, which by 2006 were present all the way to the terminus of the flow-unit. By
133 2007, dense crevasse fields were also observed in the upper and middle parts of Zawadzki breen
134 and had expanded to its entire length by 2008 (Ottesen *et al.*, 2008; Sund et al., 2014). By
135 contrast, surface crevassing of both Nathorstbreen and Polakkbreen was still limited in 2008
136 and did not become widespread until after the combined tidewater front had begun to advance
137 (Sund et al., 2014).

138 The first indications of terminus advance and calving were observed at the front of
139 Dobrowolskibreen in late 2007, followed by the abrupt advance of the combined terminus
140 sometime after October 2008 when velocities increased simultaneously (and by a factor of
141 three compared to 2007 values) within the Nathorstbreen and Zawadzki breen flow-units (Sund
142 et al., 2009, 2014; Sund and Eiken, 2010). By September 2009, the terminus had advanced ~8
143 km at the centre-line into the deepest part of the inner fjord (~70 m according to pre-surge
144 bathymetry), representing the largest advance in a single year during the surge (Figs 1 and 2a).
145 The highest recorded velocities throughout the duration of the surge were also during this
146 period, averaging 20 m d^{-1} from March-May 2009 and 25 m d^{-1} from May-September 2009 at
147 the front (Sund et al., 2014). All four contributing flow-units appeared to advance at a similar

148 rate at this time, resulting in the combined front maintaining its overall shape and width and
149 precluding the development of looped moraines. By 2010, the terminus began to spread
150 towards the bay in front of the former tributaries Doktorbreen and Liestølbreen (not active
151 during the surge), with velocities reducing to 12.9 m d^{-1} across the front in the second part of
152 2010 (Sund et al., 2014). The terminus continued advancing northwards and by August 2011
153 had advanced onto Nordre Nathorstmorenen, effectively closing off the bay at the front of
154 Doktorbreen and Liestølbreen. From August 2011 until August 2012, the front advanced a
155 further $\sim 1.5 \text{ km}$ along the central axis of Van Keulenfjorden, recording average velocities of
156 $\sim 5 \text{ m d}^{-1}$ in the period 2011-12 (Sund et al., 2014). By July 2012, when fieldwork for this study
157 was conducted, the front had advanced into a shallow ($\sim 20 \text{ m}$ deep according to pre-surge
158 bathymetry; Ottesen et al., 2008) and narrow part of the fjord and as a result ceased calving
159 (Figs 1 and 2b). In 2011-12 there was a reduction in both the rate of terminus advance (to ~ 1
160 m d^{-1}) and surface velocities, with velocities in winter 2012 ($\sim 2 \text{ m d}^{-1}$) about one-third of those
161 in winter 2011 ($\sim 6 \text{ m d}^{-1}$; Sund et al., 2014). Between December 2012 and January 2013, frontal
162 velocities remained $\sim 2 \text{ m d}^{-1}$ (Schellenberger et al., 2016). The front was still advancing in
163 August 2013 at the end of the period covered by the Sund et al. (2014) observations, and
164 extended $\sim 0.9 \text{ km}$ further downfjord compared to the August 2012 position. Between 2008 and
165 2013, NGS advanced a total of $\sim 15 \text{ km}$ (Fig. 1), with an additional $\sim 3 \text{ km}$ in length estimated
166 to have been lost through calving in the period 2009-2012 (Sund et al., 2014).

167

168 *2.2 Little Ice Age maximum surge*

169 There is evidence that NGS surged during the Little Ice Age (LIA) based on historical maps
170 and observations (Dunér and Nordenskiöld, 1865; Hamberg, 1905; Gripp, 1929), and marine
171 geological investigations (Ottesen et al., 2008). Photogrammetric mapping in 1898 by
172 Hamberg (1905) shows a tidewater glacier front with a large calving bay terminating $\sim 3 \text{ km}$

173 upfjord from the northern extents of Nordre and Søre Nathorstmorenen (Fig. 1). An earlier map
174 by Dunér and Nordenskiöld (1865) shows the combined glacier front in 1864 terminating at a
175 position a further ~9 km upfjord from the 1898 terminus. From this, Ottesen et al. (2008)
176 inferred that NGS advanced sometime after 1864, and by 1898 was in the early stages of retreat
177 from the maximum position reached during this advance. The maximum position reached prior
178 to 1898 was suggested to be immediately downfjord of the northern extents of Nordre and Søre
179 Nathorstmorenen (Liestøl, 1973, 1977; Ottesen et al., 2008) (Fig. 1). This implies that between
180 1864 and 1898 NGS advanced ~12 km, followed by a retreat of ~3 km (Ottesen et al., 2008).
181 Assuming an average retreat rate comparable to the ~160 m a⁻¹ that Nathorstbreen underwent
182 in the subsequent quiescent phase between 1898 and 2008 (~18 km in 110 years), this suggests
183 that the maximum position was reached ~15-20 years prior to 1898, in the late 1870s or early
184 1880s (Liestøl, 1973, 1977; Ottesen et al., 2008).

185 Several pieces of evidence suggest that the LIA advance was a surge. Firstly, the glacier
186 advanced ~12 km within a period of ~20 years, which is comparable in size and timescale to
187 the recent surge and a number of other observed surges of Svalbard glaciers (Murray et al.,
188 2003). Secondly, Hamberg (1905) mapped looped moraines on the glacier surface in 1898,
189 which are diagnostic of surges (cf. Meier and Post, 1969). Thirdly, swath bathymetry data from
190 inner Van Keulenfjorden presented by Ottesen et al. (2008) revealed a submarine landform
191 assemblage that is consistent with surging. This included: (i) glacial lineations, formed beneath
192 fast-flowing ice (e.g. King et al., 2009); (ii) a large terminal moraine located just beyond the
193 northern extents of Nordre and Søre Nathorstmorenen (Fig. 1), interpreted to be glaciotectonic
194 in origin (Ottesen et al., 2008); (iii) geometrical ridges, interpreted as crevasse squeeze ridges
195 formed by the injection of seafloor sediments into basal crevasses (e.g. Lovell et al., 2015); and
196 (iv) annual retreat moraines, marking minor winter readvances during terminus retreat in the
197 quiescent phase (e.g. Flink et al., 2015). This landform assemblage, or slight variations of it, is

198 found at the marine margins of several other known surge-type glaciers in Svalbard and is
199 suggested to be diagnostic of tidewater glacier surging (Ottesen et al., 2008, 2017; Flink et al.,
200 2015).

201

202 *2.3 Late Holocene maximum surge-like advances*

203 The large terminal moraine and debris flow lobe mapped by Ottesen et al. (2008), and assumed
204 to be of LIA age, were investigated and reinterpreted by Kempf et al. (2013) based on a
205 combination of swath bathymetry data, high-resolution seismics, and sediment cores. The
206 seismic data showed that the debris lobe actually consisted of two stacked units, which could
207 be correlated to a sediment core (JM07-014) collected from just beyond their distal margins
208 (Fig. 1). The age-depth model developed from the core indicates that both debris flow lobes
209 were deposited during a period of rapid sediment accumulation between 2.61 and 2.79 cal. kyr
210 BP, thus suggesting that these, and the terminal moraine complex, are considerably older than
211 the previously-assumed LIA age (Kempf et al., 2013). The implication is that the LIA surge
212 did not reach the crest of the terminal moraine complex, which was formed by the two ~2.7
213 kyr BP advances, as it would presumably have reworked the ridge and disturbed the debris
214 flow lobes. This was not apparent from the seismic profiles or the core. Instead, Kempf et al.
215 (2013) suggested that the LIA surge must have terminated to the east of the large moraine.
216 Kempf et al. (2013) concluded that the two advances at ~2.7 kyr BP were surge-like, and
217 estimated a time interval between deposition of the two lobes of ~100-150 years based on the
218 thickness of the acoustically stratified sediments. This is comparable to the modern NGS surge
219 return period of ~130 years.

220

221 **3. Methods**

222 The glacier margin from 2008-2017 was mapped from ASTER, Landsat (ETM+ and OLI) and
223 Sentinel-2 satellite imagery (acquired from earthexplorer.usgs.gov), apart from the 2009
224 margin, which was taken from Sund et al. (2014). Flow velocities were derived by feature-
225 tracking on TerraSAR-X (2013-2015) and Sentinel-1 (2015-2017) satellite image pairs. The
226 geomorphology of Nordre and Søre Nathorstmorenen was mapped from uncorrected 1:15,000
227 scale digital aerial photographs (acquired by the Norwegian Polar Institute (NPI) in summer
228 2011) and during fieldwork in July 2012 by adhering to the general mapping principles outlined
229 in Chandler et al. (2018). Sediment sections were cleaned before being logged as scaled two-
230 dimensional or vertical logs. Sediment facies were identified based on physical characteristics
231 (e.g. grain size range, sedimentary structures) following Evans and Benn (2004). Samples of
232 50 sandstone clasts were collected for clast shape and roundness analysis following Benn and
233 Ballantyne (1994). Clast shape data were plotted as ternary diagrams using TriPlot (Graham
234 and Midgley, 2000), clast roundness data were plotted as frequency distributions, and C_{40} , RA
235 and RWR indices were calculated (see Lukas et al., 2013). Bulk sediment samples were oven
236 (diamict) or freeze (mud) dried and dry-sieved to separate the fraction finer than 2ϕ ($250\mu\text{m}$).
237 The finer fraction was treated with hydrogen peroxide and disaggregated with a dispersing
238 agent before being analysed using a Beckman-Coulter Laser Sizer. These were plotted as grain
239 size distributions using GRADISTAT (Blott and Pye, 2001). Paired and individual bivalve
240 shells were radiocarbon dated at the ^{14}C CHRONO Centre for Climate, the Environment and
241 Chronology at Queen's University Belfast. The radiocarbon ages were calibrated using OxCal
242 4.3 (Bronk Ramsey and Lee, 2013; Bronk Ramsey, 2017) and the internationally accepted
243 Marine13 radiocarbon calibration curve (Reimer et al., 2013), using a marine reservoir
244 correction with a ΔR value of 70 ± 30 years (Mangerud et al., 2006; Mangerud and Svendsen,
245 2018). All radiocarbon ages are reported in the text as calibrated years before present (cal. yr
246 BP or cal. kyr BP). Key ages younger than 1000 cal. yr BP are also presented as years AD for

247 comparison with reported historical and modern dates. Bayesian modelling within OxCal 4.3
248 was used to produce a robust age estimation for the likely timing of an identified glacial
249 advance by constructing a simple ‘Phase’ model. This included using selected radiocarbon
250 ages of bivalve shells, which are assumed to be slightly older than (or maximum-limiting) the
251 advance, and using the ‘Boundary End’ age function within the model.

252

253 **4. The surging margin of the Nathorstbreen glacier system (NGS)**

254 During the recent surge, NGS advanced onto the terrestrial lateral moraine areas (Figs 1 and
255 2), providing a rare opportunity to observe ice-marginal processes during a surge and
256 investigate the geomorphological impact on the fjord and surrounding terrestrial areas.

257

258 *4.1 Frontal change and flow velocities 2013-2017*

259 The glacier front continued to advance from July 2013 until at least March 2016, showing that
260 the surge was still ongoing, albeit at a much-reduced rate of terminus change than in preceding
261 years (Figs 1 and 3). We measured flow velocities close to the front, recording an overall
262 deceleration from 2 m d⁻¹ in early 2013 to ~0.1 m d⁻¹ in late 2017, punctuated by dramatic
263 acceleration peaks in the summer months that coincide with precipitation events (Fig. 4). The
264 front advanced ~500 m in the centre of the fjord between July 2013 and July 2014. Flow
265 velocities increased abruptly in summer 2013 from ~2 m d⁻¹ in June to ~5 m d⁻¹ in July,
266 decreasing to <0.5 m d⁻¹ in winter 2013-14 (Fig. 4). Frontal velocity peaked again at 4 m d⁻¹ in
267 July 2014, before dropping to <0.5 m d⁻¹ from September 2014 through to April 2015. A data
268 gap has resulted in no identifiable velocity peak in summer 2015, but the terminus advanced
269 ~300 m from July 2014 to July 2015 (Figs 3 and 4). The front continued to advance a further
270 ~200 m from July 2015 until March 2016. Velocities reduced from ~1 m d⁻¹ in August 2015 to
271 <0.5 m d⁻¹ in winter 2015-16 (Fig. 4). A further abrupt velocity increase occurred in summer

272 2016 from $\sim 0.5 \text{ m d}^{-1}$ in May to almost $\sim 2.5 \text{ m d}^{-1}$ in July (Fig. 4). Velocities reduced to ~ 0.5
273 m d^{-1} in late 2016 and to $\sim 0.1 \text{ m d}^{-1}$ in early 2017. As of August 2017, most of the glacier front
274 had receded relative to its 2016 position (Fig. 3), but it still experienced an abrupt summer
275 speed-up in July 2017 to $\sim 1 \text{ m d}^{-1}$. By August 2017, velocities had reduced to $\sim 0.1 \text{ m d}^{-1}$,
276 comparable to pre-advance values in 2007. Together with the frontal recession, this indicates
277 that the active surge phase terminated sometime during winter 2016-17 (Figs 3 and 4). The
278 total advance during the 2008-2016 surge was $\sim 16 \text{ km}$, with $\sim 1 \text{ km}$ of the advance occurring
279 in the later stages of the surge from 2013-2016 (Fig. 1).

280

281 *4.2 Ice margin observations in July 2012*

282 When fieldwork was conducted in July 2012, the central and most-extensive part of the
283 advancing terminus had reached a position approximately level with the distinct curved, spit-
284 like arm of Nordre Nathorstmorenen (Fig. 1). Despite having advanced onto the moraine areas,
285 the ice itself could not be accessed in 2012 due to the complex, fragmented nature of the
286 margin, the extensive areas of mud, and the presence of large (up to 15 m wide) and turbulent
287 meltwater channels along the full lengths of both lateral terrestrial margins (Figs 5, 6a and 6b).
288 Chaotic crevassing was observed across the $\sim 10 \text{ km}$ of the margin that was explored, including
289 the $\sim 5 \text{ km}$ of the front terminating in the fjord. Based on satellite images and aerial photographs
290 throughout the duration of the surge and observations from a helicopter flight over the glacier
291 in March 2011, this extended along the entire length of the terrestrial margin. The chaotic
292 crevassing was evident as large (ranging from ~ 5 - 30 m high), heavily-fractured blocks with
293 multiple distinct, typically sharp, pinnacles (Figs 6a and 6c). The orientation of the blocks and
294 pinnacles varied, ranging from vertical and near-vertical, those tilted at up to $\sim 45^\circ$ (in all
295 directions), through to those that had clearly toppled over and/or broken off. The latter were
296 typically debris-covered and formed dense groups stranded within the areas of mud and shallow

297 fjord waters at the margin. Due to this, there was no clearly-defined ice cliff (as found at calving
298 margins). Instead, the margin was heavily fragmented and stepped in height in most places,
299 increasing from ~2-5 m amongst the jumble of toppled and broken-off debris-covered blocks
300 up to ~20-30 m-high clean-ice pinnacles, often over a distance of tens of metres. Refrozen
301 breccias of smaller ice fragments and blocks were commonly observed within crevasses and
302 between large blocks, and some vertical and near-vertical crevasses contained muddy debris
303 extending up to tens of metres above the fjord level (Fig. 6b). Areas of debris-rich ice were
304 observed all along SW margin (Fig. 6b). Where the front terminated directly on the moraine
305 area, the moraine surface mostly appeared undisturbed, aside from at least one location at the
306 SW margin where part of the moraine had been excavated in front of the margin.

307

308 *4.3 Mud apron*

309 *4.3.1 Description* - A large area of seafloor mud located above the waterline first appeared
310 within the fjord across most of the tidewater front in summer 2012 (Figs 3 and 6), hereafter
311 referred to using the non-genetic term mud apron (Kristensen et al., 2009a). By mid-July 2012,
312 the mud apron extended up to 500 m in length from the NE margin and had begun to encroach
313 onto the spit-like arm of Nordre Nathorstmorenen (Figs 6c and 6d). The mud apron covered ~2
314 km² across the entire front at this time, and also extended upglacier at the SW lateral margin,
315 forming a ~10-20 m-wide border adjacent to the channel (e.g. Fig. 6b). Beyond the mud apron
316 in the centre of the fjord, the water was extremely turbid and <1 m deep ~1 km downfjord from
317 the glacier front (position marked by X in Fig. 3). The pre-surge water depth in this part of the
318 fjord was ~20 m (Fig. 1). The shallow water depth was also apparent from the large number of
319 icebergs stranded in turbid water in front of the margin, in addition to those surrounded by the
320 mud apron itself. Stranded icebergs were found all across the central part of the front and
321 towards the area where the SW lateral channel emerged into the fjord (Fig. 6e). The glacier

322 was clearly no longer calving into deep water across the entire front in July 2012, and the only
323 floating ice found in inner Van Keulenfjorden were tiny berglets transported into the fjord
324 along the lateral channels. The mud apron persisted in the fjord at the advancing margin from
325 July 2012 to March 2016, by which time it almost entirely covered the spit-like arm of Nordre
326 Nathorstmorenen (Fig. 3). Throughout this period, the mud apron was present at both the SW
327 and NE margins, and continued to be most extensive at the latter, but was not visible above the
328 waterline in the central part of the front. By August 2017, the margin had started to recede,
329 exposing parts of the mud apron that had been beneath the glacier in previous years (Fig. 3).
330 The surface of the mud apron was examined where it encroached onto the spit-like arm of
331 Nordre Nathorstmorenen (Figs 6c, 6d and 6f). It was characterised by flat areas of highly
332 saturated, slurry-like mud with frequent surface pools (Figs 6d and 6f) and running water. Flow
333 structures and small (~1 m high and several metres long) transverse ridges were also visible.
334 The sediment sampled from the mud apron (see Fig. 5 for sample locations) was a clayey silt,
335 with peaks in the medium and coarse-grain silt ranges (Fig. 7) and very few larger clasts. A
336 small lobe of the mud apron located on the moraine surface in summer 2012 was observed to
337 have advanced by tens of centimetres from one day to the next relative to marker cairns,
338 confirming that it was actively flowing in front of the advancing ice margin.

339

340 *4.3.2 Interpretation: glaciotectonic remobilisation of fjord-floor sediments* - The position,
341 morphology, and sedimentary characteristics of the mud apron are consistent with a
342 continuously-failing mobile moraine formed through the bulldozing of fjord-floor sediments
343 in front of the advancing glacier (Fig. 8). The lack of clasts within the mud apron indicate it
344 has a marine rather than a subglacial origin (e.g. Boulton et al., 1996; Kristensen et al., 2009a;
345 King et al., 2016). The formation of the mud apron is best explained by the tectonic thickening
346 of fjord-floor sediments in response to glacial push. The observed low gradient of both the

347 subaerial and submarine parts of the mud apron are consistent with the oversteepening and
348 failure of fjord-floor sediments with low shear strength (Kristensen et al., 2009a). The
349 transverse ridges within the mud apron, which are typically aligned parallel to the ice margin,
350 are interpreted as minor compressional ridges formed as a result of glacial push (Boulton et al.,
351 1996; Kristensen et al., 2009a). The presence of flow structures on the mud apron surface and
352 the measured apron advance of tens of centimetres a day indicate that it was actively flowing,
353 supporting the interpretation of a continuously-elevating and -failing sediment mass.

354

355 **5. Geomorphology of Nordre and Søre Nathorstmorenen**

356 Nordre and Søre Nathorstmorenen extend for ~15 km along both sides of inner Van
357 Keulenfjorden and cover a total area of ~40 km² (Figs 1 and 5). The moraine areas consist of
358 hummocky terrain with multiple ponds and sediment flows, networks of geometrical ridges,
359 and multi-crested composite ridge systems.

360

361 *5.1 Hummocky terrain*

362 *5.1.1 Description* - The terrain across both Nordre and Søre Nathorstmorenen is characterised
363 by hummocks and ridges interspersed with ponds, depressions and sediment flows, with a
364 typical elevation range of ~5-10 m (Figs 9a, 9b and 10a). Most of the topography consists of
365 irregular hummocks and ridges, which in places are interspersed with organised networks of
366 sharp-crested geometrical ridges (see *Section 5.2*). The ponds are widely distributed across both
367 moraine areas. At Nordre Nathorstmorenen, the densest grouping and largest ponds are located
368 in a broad corridor towards the distal margins of the hummocky terrain, whereas closer to the
369 active margin they tend to be smaller and more widely spaced (Fig. 5). The densest groupings
370 of ponds at Søre Nathorstmorenen are located at Søre Leirodden and at the fjord edge between
371 the two large outwash corridors that dissect the moraine area (Fig. 5). Sediment flows (Fig. 5),

372 exposed ice cores (Fig. 10b) and tension cracks within ridges, hummocks and the general
373 moraine surface (Fig. 10c) are found across both moraine areas. The dominant sediment facies
374 within the hummocky terrain are two diamicts with a wide range in grain sizes from mud to
375 large boulders, described in *Section 6*.

376

377 *5.1.2 Interpretation: Ice-cored terrain formed by subaerial stagnation during quiescence* – The
378 hummocky terrain records subaerial stagnation and de-icing of the glacier in a terrestrial
379 position. The melting of ice cores and degradation of the terrain surface through thermo-
380 erosional processes (cf. Etzelmüller et al., 1996) is evident across Nordre and Søre
381 Nathorstmorenen in the form of tension cracks and numerous sediment flows, indicative of
382 internal instabilities and sediment remobilisation (Lawson, 1982; Lukas et al., 2005). Complete
383 or partial melting of buried ice has also resulted in the widespread kettle topography of ponds
384 and drained depressions (akin to thermo-karst; Healy, 1975).

385

386 *5.2 Geometrical ridge networks*

387 *5.2.1 Description* - Dense groups or networks of predominantly sharp-crested ridges within
388 Nordre and Søre Nathorstmorenen are mapped as geometrical ridges (Figs 5, 9c, 9d and 10d).
389 These ridges were previously identified and described by Gripp (1929) as '*Lehmmauern*'
390 ('loam walls'; van der Meer, 2004). Individual ridges are typically 2-8 m high, 1-3 m wide and
391 ~50-100 m long. Ridge orientations are predominantly offset by 45° from, or sub-parallel to,
392 the central axis of the fjord (Fig. 5). The ridges display a variety of morphologies, ranging from
393 rounded elongate hummocks to free-standing vertical pinnacles or towers (Figs 10d-f). Sharp-
394 crested ridges and pinnacles are primarily located close to the active margin, such as in the area
395 around the spit-like arm of Nordre Nathorstmorenen (Figs 9c and 9d). The ridges become more
396 rounded in a downfjord direction towards Nordre and Søre Leirodden and towards the lateral

397 margins of the moraine areas, as also noted by Gripp (1929). As a result, geometrical ridges
398 are harder to distinguish from the general hummocky terrain in these areas. The ridges are
399 predominantly composed of diamict, described in *Section 6*.

400

401 *5.2.2 Interpretation: crevasse-squeeze ridges* – The geometrical ridge networks are interpreted
402 as crevasse-squeeze ridges, commonly observed at the submarine and terrestrial margins of
403 surge-type glaciers (e.g. Sharp, 1985; Boulton et al., 1996; Evans and Rea, 1999; Ottesen and
404 Dowdeswell, 2006; Ottesen et al., 2008, 2017; Lovell et al., 2015; Farnsworth et al., 2016;
405 Ingólfsson et al. 2016). Crevasse-squeeze ridges are formed by the injection of deformable
406 basal debris into vertical and near-vertical crevasses, as observed at the active ice margin (Fig.
407 6b). The ridges are then exposed and preserved at the margin as the glacier stagnates during
408 quiescence. This mechanism is most consistent with the formation of the sharp-crested ridges
409 and free-standing pinnacles observed in the moraine areas, and agrees with the interpretation
410 of Gripp (1929) for the same features.

411

412 *5.3 Composite ridge systems*

413 *5.3.1 Description* - Extensive areas of undulating topography with multiple linear ridge crests
414 are found at the distal margins of Nordre Nathorstmorenen, forming a sharp boundary with the
415 hummocky ice-cored terrain. These areas are identified as composite ridge systems based on
416 the following key geomorphological characteristics (cf. Lovell and Boston, 2017): their
417 comparatively smooth surface texture compared with the hummocky ice-cored terrain, the
418 orientation of ridge crests (which can be both perpendicular to and parallel with the fjord axis,
419 depending on whether the ridges are in a frontal or lateral position), and the deep channels that
420 are cut into them (Figs 5, 9a, 9b, 10g and 10h). The Nordre Nathorstmorenen composite ridge
421 systems are divided into the Nordre Leirodden and North-East (NE) systems (Figs 5, 9a and

422 9b). The Nordre Leirodden composite ridge system, which was briefly described by Gripp
423 (1929; in van der Meer, 2004: pp. 54-57), covers an area of $\sim 2 \text{ km}^2$ and extends upglacier for
424 $\sim 6 \text{ km}$ from Nordre Leirodden at the NW margin of the moraine area (Fig. 5). The surface of
425 this part of the composite ridge systems reaches heights of up to 10 m above fjord/outwash
426 plain level and is dissected by several deep inactive channels (Fig. 9a). Low-amplitude ridge
427 crests are aligned sub-perpendicularly to the fjord axis close to Nordre Leirodden, and sub-
428 parallel to the fjord axis in a lateral position (Fig. 5). These ridge crests are very subdued and
429 reach typical heights of $< 0.5 \text{ m}$, and as a result are often difficult to identify in the field (Fig.
430 10g) but stand out on aerial photographs. The surface of the Nordre Leirodden composite ridge
431 system is composed predominantly of sand and gravel, with shells and shell fragments visible
432 in places. The NE composite ridge system is separated from the Nordre Leirodden system by
433 a large, inactive outwash corridor that joins the lateral outwash at the distal margins of the
434 moraine system (Fig. 5). The NE ridge system is also characterised by an undulating smooth
435 topography, dissected by multiple outwash corridors and reaching a height of $\sim 5\text{-}10 \text{ m}$ (Figs.
436 9b and 10h). The NE system extends along the lateral margins for $\sim 6 \text{ km}$ and reaches a
437 maximum width of $\sim 1.5 \text{ km}$, in total covering $\sim 5 \text{ km}^2$. Ridge crests and linear depressions are
438 typically aligned sub-perpendicularly to the fjord axis (Figs 5 and 9b). The biggest distinction
439 between the NE and Nordre Leirodden composite ridge systems is that the surface of the NE
440 system is composed of mud, with little to no sand and gravel (and no larger clasts) (Fig. 10h).
441 The mud is shell-rich with abundant shell fragments and complete paired bivalve shells
442 embedded in the surface.

443

444 *5.3.2 Interpretation: glaciotectonic moraines formed in a proglacial position* – The Nordre
445 Nathorstmorenen composite ridge systems are interpreted as glaciotectonic proglacial
446 moraines formed by glacier advance into foreland sediments (cf. Croot, 1988; Boulton et al.,

447 1999; Benediktsson et al., 2010; Lovell and Boston, 2017). The majority of the ridge crests,
448 certainly within the Nordre Leirodden ridges, are oriented parallel to the inferred ice-contact
449 face (boundary with the hummocky ice-cored terrain). This is consistent with ridge crests
450 formed perpendicular to the inferred direction of ice push (e.g. Hart and Watts, 1997; Boulton
451 et al., 1999; Lovell and Boston, 2017) as the glacier spread laterally towards the margins. The
452 smooth surface texture of the Nordre Nathorstmorenen composite ridge systems reflects their
453 surface sediment composition of sorted sand and gravel (Nordre Leirodden ridges) and mud
454 (NE ridges). Information on the internal structure of the Nordre Leirodden ridge system can be
455 found in *Section 6.3*.

456

457 **6. Sedimentology of Nordre and Søre Nathorstmorenen**

458 The sedimentary composition of the moraine areas was investigated within a series of sections,
459 mostly located at the fjord edge (Fig. 5). Five main sediment facies and facies associations
460 (FA) were identified: (1) shell-rich diamict (diamict 1); (2) shell-poor diamict (diamict 2); (3)
461 deformed fines (mud), sand and gravel (FA1) (4) undeformed fines (mud) and sand (FA2); and
462 (5) massive sand with contorted lenses (FA3). Information on grain size distributions, clast
463 shape, and clast roundness can be found in Fig. 7. Calibrated radiocarbon ages of shells are
464 reported here in relation to the sediment facies they were sampled from, and are discussed
465 further in *Section 7.2.3*.

466

467 *6.1 Diamict 1*

468 *6.1.1 Description* - Diamict 1 is the lowermost of two diamicts identified (Figs 11a and 12).
469 Diamict 1 is shell-rich, silty to fine-sandy, well-consolidated, matrix-supported, and contains
470 occasional boulders reaching maximum a-axis lengths of 1 m. Clasts within diamict 1 are
471 predominantly sub-angular and sub-rounded (Fig. 7), with striae common. This diamict is

472 typically structureless, but does contain thin clay stringers towards the distal parts of the
473 moraine areas, in particular close to the sharp boundary with the composite ridge systems
474 (composed of fine material) in Nordre Nathorstmorenen (e.g. section NNM01; Fig. 12a).
475 Diamict 1 was found both within geometrical ridges and coastal sections cut into the
476 hummocky terrain, but is restricted to the distal parts (downfjord ~1.5-2 km) of both moraine
477 areas. Within Nordre Nathorstmorenen, diamict 1 was not identified upfjord from section
478 NNM04, and diamict 1 was the only diamict identified in the downfjord (distal) ~1.5 km of
479 Søre Nathorstmorenen. Shells in diamict 1 were sampled for dating from Søre
480 Nathorstmorenen: two single *Hiatella arctica* shells from a coastal section returned ages of
481 5720-5870 and 10380-10660 cal. yr BP, and one pair of bivalve shells embedded in the moraine
482 surface at Søre Leirodden (Fig. 11f) returned ages of 1200-1290 and 1170-1260 cal. yr BP
483 (Table 1).

484

485 *6.1.2 Interpretation: lower till* - Based on its fine-grained matrix, presence of multiple shells,
486 and predominantly sub-angular to sub-rounded and striated clasts, diamict 1 is interpreted as a
487 till derived from marine sediment (cf. Boulton et al., 1996; Ó Cofaigh and Evans, 2001). Clay
488 lenses within the diamict at section NNM01 (Fig. 12a) are interpreted as evidence of the
489 incorporation and attenuation of underlying material into the basal zone during glacier
490 overriding (e.g. Kristensen et al., 2009a). Diamict 1 is interpreted to have been transported
491 subglacially during an advance that reached Nordre and Søre Leirodden.

492

493 *6.2 Diamict 2*

494 *6.2.1 Description* - Diamict 2 is matrix-supported and overlies diamict 1 at section NNM04
495 (Figs 11a and 12c), forming a sharp, erosional contact. The two diamicts can be differentiated
496 because diamict 2 is typically friable, poorly-consolidated and shell-poor, containing fewer and

497 more fragmentary shells than diamict 1. Thin, sometimes contorted, layers of sand and
498 occasional clay stringers are also common (e.g. Figs 11b, 12c and 13a). The main difference in
499 grain size distribution of diamicts 1 and 2 is the larger peaks within the coarse silt/fine sand
500 ranges displayed by diamict 2. Clast shape samples taken from sections NNM04 and SNM02
501 and from within geometrical ridges show predominantly sub-angular to sub-rounded clasts
502 (Fig. 7). Diamict 2 is the only diamict observed close to the active margin and is not found in
503 the distal parts of the moraine areas (e.g. downfjord from NNM04 and SNM01). Within Søre
504 Nathorstmorenen, the lateral transition in surface sediment cover from diamict 2 to diamict 1
505 is indistinct and, unlike at Nordre Nathorstmorenen, the two diamicts were not observed
506 together in section. Close to the active margin, diamict 2 is massive and sand lenses/clay
507 stringers are rare, but these increase in frequency in a downfjord direction (e.g. Figs 11b, 12c
508 and 13a). A single *Hiatella arctica* shell sampled from diamict 2 at Søre Nathorstmorenen (Fig.
509 5) returned an age of 7730-7860 cal. yr BP (Table 1).

510

511 *6.2.2 Interpretation: upper till* - Diamict 2 is interpreted as a second till derived from marine
512 sediment, deposited by a separate, less-extensive and more-recent glacier advance than that
513 associated with the deposition of diamict 1. Two main factors indicate that the diamicts relate
514 to separate advances: (1) at section NNM04, diamict 2 directly overlies diamict 1 and there is
515 a sharp, erosional contact between the two (Figs 11a and 12c). This indicates that the glacier
516 overrode and likely eroded diamict 1 following its deposition, emplacing diamict 2 on top. (2)
517 The spatial distribution of diamict 2 is restricted to the zones of sharp-crested geometrical
518 ridges close to the active margin, and it is not found in the distal parts of the moraine areas.
519 This is consistent with its deposition during a second, less-extensive glacier advance than that
520 which reached Nordre and Søre Leirodden and deposited diamict 1. The sand layers and clay

521 stringers within diamict 2 are interpreted as subglacially-reworked FA1 sediments (see *Section*
522 *6.3*).

523

524 *6.3 Deformed fines, sand and gravel facies (FA1)*

525 *6.3.1 Description* - This sediment facies association is exposed within the hummocky terrain
526 of both moraine areas (e.g. sections NNM01, NNM05, SNM01 and SNM02) and within the
527 Nordre Nathorstmorenen composite ridge systems (e.g. section NNM02) (Figs 12 and 13). It
528 is characterised by poorly-sorted sand with intercalated clay stringers, layers and lenses that
529 have been subject to minor faulting (e.g. NNM01; NNM05), shearing (e.g. SNM01) and/or
530 intense folding (e.g. NNM02; SNM02); faulted fine sand layers within a clay matrix (e.g.
531 NNM01); and clastic dykes and flame structures (e.g. NNM01 and NNM02). FA1 is found in
532 three main settings within the moraine areas, described below.

533 Firstly, FA1 underlies diamict 1 in both sections NNM01 and NNM05 within Nordre
534 Nathorstmorenen. The lowermost ~1.8 m of NNM01, located within the hummocky terrain
535 close to the boundary with the Nordre Leirodden composite ridge system, consists of FA1. This
536 includes a ~1.2 m section characterised by fine sand with thin clay stringers and shells
537 overlying shale bedrock. The shale is broken up and extremely friable, and angular clasts have
538 been ripped up from it and are present within the lowermost 0.2 m of the fine sand. Thin,
539 sinuous clastic dykes with flame-like structures are evident, trending upwards and towards the
540 true left of the section. A number of small-scale minor reverse faults can be identified,
541 particularly at a height of ~1 m, where the fine sand is laminated and contains small lenses of
542 black shale of a coarse sand grain size (Fig. 12a). The fine sand is overlain by ~0.6 m of massive
543 sand interspersed with discontinuous thin, faulted fine sand lenses.

544 Secondly, FA1 is also found close to the lateral transition between diamicts 1 and 2,
545 ~1.5-2 km upfjord from the distal margins of the moraine area. Coastal exposures in this part

546 of Søre Nathorstmorenen commonly contain deformed sands with sheared clay lenses (Fig.
547 11d) overlain by up to 1-2 m of diamict 2. At section SNM01 (Fig. 13a), ~1 m of diamict 2
548 containing several thin sand layers is overlain by ~0.5 m of FA1 in the form of sheared sand
549 with clay lenses. Approximately 200 m upfjord from this location, section SNM02 (Fig. 13b)
550 shows a ~1 m-wide layer of FA1, consisting of sands and thin clay lenses, forming a downfjord-
551 verging overturned fold around a core of diamict 2. The clay lenses within the sand and the
552 thin sand layers within diamict 2 are aligned parallel to the axial surface of the fold.

553 The third main location of FA1 is within the Nordre Nathorstmorenen composite ridge
554 systems. The internal composition of the Nordre Leirodden composite ridge system was
555 investigated in coastal exposures and logged in section NNM02 (Figs 11c and 12e). In coastal
556 exposures composed primarily of deformed sand, large-scale anticlinal folding was observed
557 in several places where bedding dipped in opposite directions over distances of tens of metres.
558 Sub-vertical muddy stringers and thrust and reverse faults with small displacements are also
559 common. At section NNM02, located in the wall of a channel cut through the Nordre Leirodden
560 ridges, FA1 is characterised by folded and contorted fine and coarse sand layers forming
561 attenuated synclines, with evidence for clastic dykes and flame structures (Fig. 12e).
562 Asymmetric folds are expressed on the surface of the Nordre Leirodden composite ridge system
563 as linear stripes or low-amplitude muddy ridge crests that stand-out against the general sandy-
564 gravelly surface (Fig. 10g). Several of these fold axes strike ~290°, comparable to fold axes
565 measured in section NNM02 (Fig. 12e). The internal structure of the NE composite ridge
566 system was not investigated, but the surface consisted of mud (massive clayey silt) (Fig. 7a).
567 Four paired shells of *Hiatella arctica* embedded in the NE composite ridge system surface (Fig.
568 11e) returned ages of 750-870, 830-950, 850-960 and 950-1110 cal. yr BP (Table 1).

569

570 6.3.2 *Interpretation: glaciotectonised shallow marine and fluvial sediments* - The massive and,
571 in places, laminated sands and shell-rich muds that contain evidence for faulting, folding and
572 shearing are interpreted as shallow marine, lacustrine and fluvial sediments that have been
573 glaciotectonically deformed in proglacial (e.g. NNM02 within the Nordre Leirodden composite
574 ridge system) and submarginal/subglacial (e.g. NNM01, NNM05, SNM01 and SNM02)
575 settings. FA1 sediments overlain by diamict 1 towards the distal margin of Nordre
576 Nathorstmorenen (NNM01 and NNM05) are interpreted as glaciotectonised shallow
577 marine/lacustrine sediments due to overriding and deformation as the glacier advanced to
578 Nordre and Søre Leirodden. The clastic dykes, flame structures (interpreted as water-escape
579 features) and reverse faulting within FA1 in section NNM01 are consistent with dewatering
580 and compaction caused by overlying ice (e.g. Phillips et al., 2008). FA1 sediments found in
581 conjunction with diamict 2 at the latter's downfjord limit in Søre Nathorstmorenen (e.g.
582 SNM01 and SNM 02) are similarly interpreted as evidence for the deformation of shallow
583 marine/lacustrine sediments in an ice-marginal position. The location of these sections, ~2 km
584 upfjord from the distal extent of the moraine system at Søre Leirodden indicates that they
585 represent a second former ice-contact zone. This is consistent with a second, less-extensive
586 glacier advance that deposited diamict 2. Deformed FA1 sediments are also found within the
587 glaciotectonic composite ridge systems of Nordre Nathorstmorenen. As the glacier advanced
588 downfjord and into terrestrial positions, it is likely to have encountered shallow marine
589 sediments within the fjord, lacustrine sediments in ponds on the moraine/glacier surface, and
590 outwash sediments. These sediments were then pushed and bulldozed into a series of ridges,
591 with ridge crests typically oriented perpendicular to the direction of ice push. The style of
592 folding and strike of fold axes within NNM02 (Fig. 12e) and exposed on the surface (Fig. 10g),
593 in conjunction with the general alignment of ridge crests on the Nordre Leirodden composite

594 ridge system (Figs 5 and 9a), are consistent with ice push from the south as the glacier advanced
595 into a terrestrial position and towards the lateral margins.

596

597 *6.4 Undeformed fines and sand (FA2)*

598 *6.4.1 Description* - FA2 is characterised by wavy laminated alternating couplets of clay-fine
599 silt grading into coarse silt-fine sand (NNM03; Figs 11g and 12b), ripple-bedded sands
600 (NNM03 and SNM03; Figs 12b and 13c), and massive clay, silt and sand (NNM03, NNM05
601 and SNM03). This facies association is found at the transition between the zones of diamicts 1
602 and 2 (i.e. where they are the predominant sediment facies) in both moraine complexes,
603 approximately 1.5 km upfjord from the distal margins. At section NNM03, located immediately
604 upfjord from the observed downfjord limit of diamict 2 within Nordre Nathorstmorenen, FA2
605 consists of a ~3 m-thick coarsening-upwards sequence of undisturbed wavy-laminated
606 alternating clay-fine silt couplets (~1 cm thick) (Fig. 11g), the uppermost ~1 m of which grades
607 into alternating silt-fine sand couplets (Fig. 12b). This sequence sits on top of bubbly, opaque
608 ice at least 0.2-0.3 m thick, and extends laterally for ~30 m. The laminated clays and silts are
609 overlain by a ~0.5 m-thick coarse sand layer with asymmetric ripples prograding in opposite
610 directions. Undeformed sorted sediments of FA2 are also found in section SNM03 in Søre
611 Nathorstmorenen, in a ~5-10 m-long exposure in a ridge located ~200 m from the fjord edge
612 (Fig. 5). Here, the lowermost 0.5 m of the exposure consists of wavy laminated alternating
613 couplets of fine and medium sand containing asymmetric ripples prograding in opposite
614 directions, interspersed with thin (<3 cm) horizontally-bedded lenses of fine gravel (Fig. 13c).

615

616 *6.4.2 Interpretation: undeformed shallow lacustrine sediments* - FA2 is consistent with
617 sediments of a shallow marine or shallow lacustrine origin. The wavy-laminated couplets
618 resemble tidally-influenced sediments (cf. Stewart, 1991), and the location of NNM03 at the

619 fjord edge suggests that the sediments may have been uplifted following deposition in the fjord.
620 However, the undeformed nature of the sediments and their height above the current fjord level
621 (up to ~3 m at NNM03) suggests that a marine origin is unlikely, as it is difficult to envisage a
622 mechanism by which uplift could have occurred with little or no disturbance of the sediments.
623 This is also the case for FA2 at SNM03, which is located ~300 m from the fjord and therefore
624 would have to have been transported a considerable distance without deformation. The absence
625 of shells within FA2 also suggests a marine origin is unlikely. Based on this, FA2 is most
626 consistent with undeformed shallow lacustrine sediments deposited *in situ* in ponds on the
627 former glacier or moraine surface that have since drained. NNM03 is underlain by glacier ice,
628 indicating deposition in a supraglacial pond before being lowered to the moraine surface. The
629 asymmetric ripples prograding in opposite directions within NNM03 and SNM03 are
630 consistent with delivery of underflows into a pond from changing sediment efflux points.

631

632 *6.5 Massive sand with contorted lenses (FA3)*

633 *6.5.1 Description* - FA3 is only observed at section SNM03, where it overlies, and has a sharp
634 contact with, the undeformed laminated sands of FA2. FA3 is characterised by a ~1.7 m-thick
635 layer of massive fine sand containing contorted coarse sand lenses and scattered gravel-sized
636 clasts (Fig. 13c). There is evidence for small rip-up clasts of medium sand within the lowermost
637 ~10-20 cm, sourced from the underlying laminated sands of FA2, and the thin lenses of coarser
638 sand within the otherwise fine sand matrix display a variety of shapes and alignments.

639

640 *6.5.2 Interpretation: subaerial sediment flow deposit* – FA3 is interpreted as a sediment flow
641 subjected to gravitational and water-sorting processes. This origin is consistent with the sharp,
642 erosional lower contact, homogenous nature of the fine sand matrix, the scattered gravel-sized
643 clasts, and the contorted coarse sand inclusions (cf. Lawson, 1982).

644

645 **7. Discussion**

646 *7.1 Formation of mud apron and glaciotectonic surge moraines*

647 The development of a mud apron during the recent surge of NGS provides a link between active
648 phase processes and the formation of surge moraines, both in submarine and terrestrial
649 positions. The emergence of the mud apron above the waterline in summer 2012 coincided
650 with the glacier front reaching a narrow and shallow (~20 m) part of inner Van Keulenfjorden
651 (Fig. 3), as noted by Sund et al. (2014). The extremely shallow water depths (<1 m) in the
652 centre of the fjord in 2012 at ~1 km from the advancing margin demonstrate that a low gradient
653 debris flow lobe extended downfjord from the subaerial part of the mud apron. Based on the
654 pre-surge bathymetry (Fig. 1), the glacier front advanced up a reverse slope for ~10 km from
655 the deepest part of inner Van Keulenfjorden in 2009 to its 2012 position. The soft marine
656 sediments were pushed upslope as the front advanced, incrementally increasing the thickness
657 of the sediment wedge over a distance of ~10 km through tectonic shortening (Fig. 8). By the
658 time the glacier front reached the top of the reverse slope, i.e. at the shallowest part of the fjord,
659 the sediment wedge was thick enough to breach the fjord surface. The advance against a reverse
660 slope also explains why the mud apron is able to attain a significant thickness despite being
661 composed of sediment with a very low shear strength and high porewater pressure, which might
662 be expected to fail continuously (Kristensen et al., 2009a). The gravitational forces acting on
663 the distal slope of the sediment wedge would be less influential than the lateral compression as
664 it advanced upslope, therefore allowing the sediments to thicken. Once the glacier reached the
665 top of the reverse slope and the bed began to slope away downfjord, gravitational processes
666 exerted a larger influence on the sediment wedge and a low-gradient debris flow lobe
667 developed through quasi-continuous slope failure (Kristensen et al., 2009a). It is interesting to
668 note that Hamberg (1905) reported that in 1898 the water near Nordre Leirodden (Fig. 5)

669 contained stranded icebergs and was too shallow for boats due to the large amount of mud
670 deposited by the glacier. These observations are consistent with a mud apron within the fjord
671 associated with the 1898 position of NGS and reflect our own experience of navigating a boat
672 across the mud apron near the ice margin in 2012.

673 The mud apron that formed during the recent NGS surge is morphologically similar to
674 large submarine terminal moraines and associated debris flow lobes observed on the seafloor
675 in front of a number of tidewater surge-type glaciers (e.g. Solheim and Pfirman, 1985; Plassen
676 et al., 2004; Ottesen et al., 2008, 2017; Kristensen et al., 2009a; Forwick et al., 2010; Flink et
677 al., 2015; Streuff et al., 2015, 2017a; Burton et al., 2016), and we therefore interpret these as
678 having a common genetic origin (cf. Kristensen et al., 2009a). That is not to say that seafloor
679 sediments would have necessarily been pushed above the waterline during the formation of
680 surge terminal moraines in these other examples — this will depend on water depth. Aside
681 from Van Keulenfjorden, to our knowledge the only other report of mud pushed above the
682 waterline at an advancing glacier margin in Svalbard is from the 2002-10 surge of
683 Comfortlessbreen, which advanced ~700 m into the shallow water of Engelskbukta (King et
684 al., 2016; Lønne, 2016). However, such processes have been inferred during past surges of
685 Sefströmbreen (Boulton et al., 1996), Paulabreen (Kristensen et al., 2009a,b) and Osbornebreen
686 (Evans and Rea, 2005; Farnsworth et al., 2017) based on terrestrial geomorphological evidence.
687 Similar to these studies, the terrestrial composite ridge system at the lateral margin of Nordre
688 Nathorstmorenen is also inferred to have formed by a process of glaciotectonic pushing of
689 marine sediments onshore in front of an advancing ice margin. The grain size distribution of
690 the surface of the muddy part of the composite ridge system is very similar to that of the modern
691 mud apron (Fig. 7). Abundant shells on the surface of the ridges, many of which were paired
692 bivalve shells (Fig. 11e), also indicate a marine origin. The internal structure of this part of the
693 composite ridge system is unknown, but we note that the surface morphology is similar to the

694 terrestrial mud aprons/glaciotectonic moraines described by Boulton et al. (1996) and
695 Kristensen et al. (2009a,b).

696 There is a significant difference in the thickness of the active mud apron in terrestrial
697 areas, which was <10 cm thick where it had started to encroach onto Nordre Nathorstmorenen
698 (e.g. Fig. 6c), and the ~8-10 m high composite ridge system. Two main processes are likely to
699 explain this difference. Firstly, the sediments within the highly saturated, slurry-like active mud
700 apron have low shear strength, facilitating its continuous failure and the development of the
701 low-gradient debris flow lobe observed in both marine and terrestrial settings. By contrast,
702 large volumes of marine mud pushed further onshore would dewater, increasing the shear
703 strength of the mud. Subjected to continued push by an advancing ice margin, this would allow
704 the marine muds to be pushed into a series of steeper and higher ridges. Drying-out of marine
705 sediments can also result in extra cohesion caused by the crystallisation of salts (cf. Boulton et
706 al., 1996). Secondly, completely or partially frozen marine mud would have higher shear
707 strength than unfrozen mud, which could contribute to the coherence of the sediment mass
708 during proglacial deformation (e.g. Etzelmüller et al., 1996; Etzelmüller and Hagen, 2005).
709 The latter might be expected to form thrust-block moraines as the frozen sediments deformed
710 in a coherent manner, forming a series of thrust slabs (e.g. Evans and England, 1991).

711

712 *7.2 Late Holocene surge history of NGS*

713 Ottesen et al. (2008) used historical mapping and the submarine geomorphological record to
714 infer that NGS surged to a position downfjord of Nordre and Søre Nathorstmorenen in ~1870
715 (Ottesen et al., 2008; Fig. 1). This implies that the terrestrial moraine areas formed during the
716 LIA surge. The following sections demonstrate that Nordre and Søre Nathorstmorenen are
717 actually the result of two surges based on (1) the identification of two ice-contact zones
718 recorded by the distribution of sediment facies, facies associations and terrestrial

719 geomorphology; (2) the correlation between the ice-contact zones and submarine
720 geomorphology; and (3) radiocarbon dating of shells emplaced in a terrestrial position by the
721 outer, older surge.

722

723 *7.2.1 Identification of two ice-contact zones within the terrestrial moraine areas*

724 The sediment exposures record a downfjord transition from predominantly diamict 2 to a zone
725 of FA1 sediments, followed by a zone of predominantly diamict 1 extending to the distal part
726 of the moraine areas. In Nordre Nathorstmorenen only, the diamict 1 zone transitions into the
727 composite ridge systems comprising deformed FA1 sediments. In simple terms, we interpret
728 this as evidence of transitions (ice proximal to distal) from a subglacial zone (diamict 2) to a
729 submarginal/proglacial zone (FA1 in close association with diamict 2), back to a subglacial
730 zone (diamict 1), and finally to a proglacial glaciotectonised zone (FA1) within the composite
731 ridge systems. On this basis, two ice-contact zones can be identified at the subglacial-
732 submarginal/proglacial transitions (Fig. 14), which we attribute to two advances being
733 responsible for forming the terrestrial moraine areas. The second of these advances was less
734 extensive than the first, forming the inner ice-contact zone located ~2 km upfjord from the
735 distal extent of the moraine areas.

736 The transition from diamict 2 to FA1 deformed shallow marine sediments occurs at
737 approximately the same position within Nordre and Søre Nathorstmorenen (Fig. 14), which
738 delimits the former ice-contact zone (i.e. maximum downfjord position) of the inner advance.
739 This zone also coincides with the downfjord limit of the area of sharp-crested crevasse-squeeze
740 ridges (Figs 5 and 14) — beyond this, the ridges are, in general, more-rounded and typically
741 indistinguishable from the surrounding hummocky terrain of the moraine systems. We suggest
742 this reflects the differences in relative age of the crevasse-squeeze ridges formed by the inner,

743 younger advance and those formed during the outer, older advance to Nordre and Søre
744 Leirodden.

745 The ice-contact zone of the outer advance is located at the transition from hummocky
746 ice-cored terrain to composite ridge systems within Nordre Nathorstmorenen. This ice-contact
747 zone can be correlated with the downfjord extent of Søre Nathorstmorenen on the south side
748 of the fjord, delimiting the approximate maximum position of the outer advance (Fig. 14). The
749 lateral extents of the outer advance are constrained by the contact with the composite ridge
750 systems (Nordre Nathorstmorenen) and lateral limit of the moraine area (Søre
751 Nathorstmorenen). The lateral margins of the inner advance are harder to determine, as there
752 are no obvious geomorphological features (e.g. lateral meltwater channels or composite ridge
753 systems) that coincide with the ice-contact zones within either moraine system. It is possible
754 that the glacier extended to the lateral margins of both moraine areas. However, it seems likely
755 that the inner advance not only reached a less-extensive downfjord position, but also was
756 laterally less extensive. This is certainly the case for the recent surge, which reached a less-
757 extensive downfjord position and has only impinged on the lateral moraine areas by a few
758 hundred metres (Figs 1-3). In the absence of a clearly demarcated lateral margin for the inner
759 advance, we suggest it may coincide with observed differences in meltwater pond density. In
760 Nordre Nathorstmorenen, there is an identifiable corridor immediately adjacent to the fjord and
761 glacier margin that contains fewer meltwater ponds than the outermost part of the moraine
762 system (Fig. 5). This corridor widens from ~0.2 km at section NNM04 to ~1.5 km adjacent to
763 the 2012 glacier margin (Figs 5 and 14). We propose that the zone of fewer meltwater ponds
764 may represent the footprint of the inner advance. This is based on the logic that the relative
765 abundance of meltwater ponds in the outer, older parts of the moraine system reflect the longer
766 time it has had to de-ice and thus for ponds to develop. Søre Nathorstmorenen contains a similar
767 pattern, with dense areas of meltwater ponds located towards the distal margins, but the lateral

768 contrast is indistinct. We have therefore defined an approximate lateral extent of the inner
769 advance on Søre Nathorstmorenen based on a similar distance of encroachment onto the
770 moraine areas on both sides of the fjord (Fig. 14).

771

772 *7.2.2 Correlation of inner and outer advances with submarine geomorphology*

773 Based on the terrestrial evidence, it is logical to expect two separate advances to also be
774 recorded in the submarine geomorphology. The large submarine terminal ridge and associated
775 debris flow lobe mapped by Ottesen et al. (2008; Fig. 1), or at least the ice-proximal slope of
776 the ridge (cf. Kempf et al., 2013), correlates reasonably well with the maximum position of the
777 outer advance (i.e. the distal extent of the terrestrial moraine areas). However, Ottesen et al.
778 (2008) did not identify any terminal ridges or debris flow lobes in inner Van Keulenfjorden
779 that correlate to the inner advance. A series of small (average height of ~5 m) ridges aligned
780 broadly perpendicular to the fjord axis were identified and interpreted as annual moraines
781 formed during quiescent phase recession (Ottesen et al., 2008). Two of the largest of these
782 ridges are over 10 m high, ~500 m wide and are located either side of the position of the inner
783 ice-contact zone within the terrestrial moraine areas (labelled R1 and R2 in Fig. 14). The taller
784 of the two ridges (R1) corresponds closely to the position of the glacier front in 1898 mapped
785 by Hamberg (1905), which has the shape of a calving margin. We suggest this ridge is
786 consistent with a recessional moraine formed during quiescence (cf. Flink et al., 2015). Ridge
787 R2 is located ~1 km downfjord from the 1898 margin and immediately downfjord from the
788 approximate maximum position of the inner advance, as recorded within the terrestrial moraine
789 areas. Based on an assumed convex-shaped glacier front consistent with a surging margin (e.g.
790 Figs 1-3), we suggest that R2 is the most likely candidate to record the submarine position of
791 the inner advance maximum position (Fig. 14). Although R2 does not have a debris flow lobe
792 on its distal slope and is not as large as submarine terminal moraines identified at other

793 tidewater glaciers, we note that (i) debris flow lobes are not always found at surge terminal
794 moraines (e.g. Streuff et al., 2017a); and (ii) R2 is comparable in size and morphology to the
795 terminal moraine formed at the 2004 surge maximum position of Tunabreen in Tempelfjorden
796 (cf. Flink et al., 2015). In the latter case, Flink et al. (2015) concluded that where a surge follows
797 soon after a previous surge (e.g. ~40 years at Tunabreen), the glacier will not encounter as thick
798 glaciomarine sediment, and therefore will have less material available to bulldoze into a
799 terminal moraine. In addition, the velocity data from the final stages of the recent NGS surge
800 demonstrate that the glacier front experiences pulses of rapid flow acceleration in the summer
801 months during overall deceleration, probably during enhanced precipitation events as rainfall
802 is routed directly to the bed through the heavily-crevassed terminus (cf. Sevestre et al., 2018)
803 (Fig. 4). It is possible that most of the frontal advance in the later years occurs during these
804 concentrated periods of enhanced frontal velocities. Such a pulsing effect, possibly
805 characterised by parts of the front advancing whilst other areas are almost stationary (Fig. 3),
806 may well have an impact on the size and morphology of any submarine moraines formed at the
807 margin.

808

809 *7.2.3 Revised chronology of surging*

810 Glacier surges are separated by multi-decadal periods of quiescent phase recession. The two
811 advances recorded within the moraine areas are therefore expected to be of different ages. The
812 oldest radiocarbon ages of 10380-10660, 7730-7860 and 5720-5870 cal. yr BP (Table 1) were
813 from individual shells embedded in diamicts 1 and 2 in coastal sections within Søre
814 Nathorstmorenen. These are considerably older than the two advances dated to between 2610
815 and 2790 cal. yr BP that formed the large submarine terminal moraine and debris flow lobes in
816 outer Van Keulenfjorden (Kempf et al., 2013; Fig. 1), suggesting that the shells have undergone
817 significant (e.g. multiple cycles of) remobilisation and redeposition (Lyså et al., 2018). The

818 remaining ages were from four sets of paired bivalve shells sampled from the surface of the
819 NE composite ridge system (750-870, 830-950, 850-960 and 950-1110 cal. yr BP; Table 1) and
820 one set of paired bivalve shells embedded in the moraine surface at the distal extent of Søre
821 Nathorstmorenen (1200-1290 and 1170-1260 cal. yr BP; Table 1). The NE composite ridge
822 system is interpreted to have formed by onshore bulldozing of marine mud in a proglacial
823 position during the outer, assumed older, advance. Similarly, the paired bivalve shells
824 embedded in the surface of Søre Nathorstmorenen are within the part of the moraine complex
825 formed by the outer advance (Fig. 14). We use the four, slightly younger ages from the NE
826 composite ridge system to produce a robust modelled age for the outer advance occurring
827 during the period 700-890 cal. yr BP (Fig. 15), or sometime around ~1160 AD.

828 The inner advance corresponds closely to the 1898 glacier front (Fig. 14). Similar to
829 Ottesen et al. (2008), we interpret the 1898 position as representing the initial stages of frontal
830 recession following a late 19th century surge. However, we have identified that this surge did
831 not extend to the distal part of the moraine system as previously thought, but terminated ~2 km
832 upfjord at our proposed maximum position of the inner advance (Fig. 14). By 1898 the glacier
833 front in the centre of the fjord had calved back ~1-1.5 km from the likely surge maximum
834 position. If we assume a quiescent phase recession rate of ~160 m a⁻¹ (as recorded in the period
835 1898 to 2008), the inner advance likely reached its maximum position ~6-10 years prior to
836 1898, suggesting the LIA surge occurred ~1890. We therefore determine that NGS has surged
837 at least five times: twice between 2.61 and 2.79 cal. kyr BP (Kempf et al., 2013), at ~1160 AD,
838 in ~1890, and from 2008-2016 AD (Fig 16).

839

840 *7.3 Timings of Late Holocene tidewater glacier advances in Svalbard*

841 There are very few records of the timings of Late Holocene tidewater glacier advances in
842 Svalbard before the LIA maximum. Indeed, prior to an inferred ~1800 AD surge of

843 Kongsvegen/Kronebreen in Kongsfjorden (Liestøl, 1988), only six glacier systems have had
844 dated advances since the onset of the neoglacial period at ~4 kyr BP (Hald et al., 2004) (Fig.
845 16). The oldest dated advances since ~4 kyr BP are the two advances of NGS between 2.61
846 and 2.79 cal. kyr BP, which were interpreted as surges (Kempf et al., 2013). The Hinlopen-
847 Oslobreen glacier systems in Vaigattbogen also surged sometime prior to 2.6 cal. kyr BP,
848 although this could have occurred at any point since the early Holocene (Flink and Noormets,
849 2018). In Hornsund, southern Spitsbergen, a tidewater glacier advance to the Treskelen
850 Peninsula was dated to 1.9 ± 0.3 kyr BP using ^{10}Be cosmogenic nuclide dating (Philipps et al.,
851 2017). Philipps et al. (2017) suggested this advance was likely to be in response to regional
852 climate forcing rather than a surge, although they also noted that several of the glaciers feeding
853 into Hornsund are reported to have surged at or since the LIA. Also in southern Spitsbergen,
854 Paulabreen in Van Mijenfjorden surged ~650 yr BP (~1300 AD) (Hald et al., 2001; Larsen et
855 al., 2018). This surge formed the terrestrial glaciotectonic moraine systems of Damesmorenen,
856 Crednermorenen and Torrelmorenen through onshore bulldozing of marine mud (Kristensen et
857 al., 2009a; Larsen et al., 2018; Lyså et al., 2018). Based on the submarine geomorphological
858 record (Ottesen et al., 2008), Paulabreen surged at least twice more between the dated ~1300
859 AD surge and an inferred surge in ~1898 (Larsen et al., 2018). In northwest Spitsbergen, St.
860 Jonsfjorden and Magdalenefjorden both have tidewater glacier advances dated to before the
861 LIA (Farnsworth et al., 2017; Streuff et al., 2017a). Osbornbreen in St. Jonsfjorden advanced
862 and deposited a moraine dated to $\sim 520 \pm 70$ cal. yr BP (~1430 AD), which has been interpreted
863 as a surge based on the terrestrial and submarine geomorphological record (Evans and Rea,
864 2005; Farnsworth et al., 2017). In Magdalenefjorden, Waggonwaybreen advanced at ~300 cal.
865 yr BP (~1650 AD) (Streuff et al., 2017a). The submarine geomorphology was interpreted by
866 Streuff et al. (2017a) to be more consistent with this advance being a response to LIA cooling
867 rather than a surge. Both Paulabreen and NGS are inferred to have surged at the end of the LIA

868 maximum (Ottesen et al., 2008) and have surged in the last 15 years (Kristensen and Benn,
869 2012; Sund et al., 2014). Osbornebreen also underwent an observed surge in 1986-1988
870 (Dowdeswell et al., 1991).

871 The recent expansion in the availability of high-resolution submarine imagery has
872 helped to identify new evidence for dynamic glacier flow in both fjord (Ottesen and
873 Dowdeswell, 2006; Ottesen et al., 2008; Flink et al., 2015, 2017; Streuff et al., 2015, 2017a;
874 Burton et al., 2016; Ewertowski et al., 2016; Farnsworth et al., 2017; Allaart et al., 2018;
875 Ćwiąkała et al., 2018; Larsen et al., 2018) and open-marine settings (Ottesen et al., 2017; Flink
876 and Noormets, 2018). In the majority of these examples, there is clear geomorphological
877 evidence for surging. As more areas are explored, it seems likely that such observations will
878 increase. However, chronological control on the timing of advances is crucial in order to
879 understand tidewater glacier behaviour during the Late Holocene. From the available data, it is
880 clear that there is a great deal of variability across Svalbard. In some areas (e.g. inner Isfjorden,
881 Lomfjorden), the LIA maximum is thought to represent the most-extensive Holocene glacier
882 position (Plassen et al., 2004; Ottesen and Dowdeswell, 2006; Mangerud and Landvik, 2007;
883 Streuff et al., 2017b). By contrast, glaciers in Mohnbukta experienced a surge-type advance
884 prior to 7.7 cal. kyr BP (Flink et al., 2017), and both NGS (Kempf et al., 2013 and this study)
885 and Paulabreen (Larsen et al., 2018; Lyså et al., 2018) surged at least three times prior to the
886 LIA to more advanced positions than their LIA maximums. In terrestrial settings, Farnsworth
887 et al. (2018) established that several land-terminating glaciers in Svalbard re-advanced during
888 the late-glacial to early-Holocene period, reaching positions up to ~8 km beyond their Late
889 Holocene maximum moraines, and Miller et al. (2017) identified several episodes of land-
890 terminating glacier expansion during the Late Holocene. Understanding these variations in the
891 timings of glacier maxima during the Holocene is important in order to understand glacier

892 behaviour over longer timescales, and in particular the interplay between climatic forcing and
893 glaciodynamical (i.e. surging) influences on glacier advances.

894

895 **9. Conclusions**

896 Investigation of terrestrial and submarine sediment-landform assemblages in Van
897 Keulenfjorden, southern Spitsbergen, reveal a Late Holocene record of multiple advances of
898 the surge-type Nathorstbreen glacier system (NGS).

899 • NGS advanced ~16 km from 2008 to 2016 during its recent surge. The final years of
900 the surge (2013-2016) were characterised by year-on-year decreases in flow velocities
901 punctuated by occasional, short-lived speed-ups (e.g. fivefold increases) correlated to
902 summer precipitation events. By August 2017, NGS had started to retreat across most
903 of the front, indicating surge termination sometime in winter 2016-2017.

904 • We present the first detailed observations of the formation of a glaciotectonic mud
905 apron in the fjord during the recent surge. The mud apron emerged above the waterline
906 and began to encroach onto the lateral moraine areas in summer 2012. The mud apron
907 was caused by the bulldozing and thickening of marine sediments into a mobile,
908 continuously failing sediment wedge characterised by a low-gradient flow lobe
909 extending downfjord. These observations provide a modern analogue for the formation
910 of submarine terminal surge moraines and associated debris flow lobes, and terrestrial
911 glaciotectonic moraine systems formed by the onshore movement of marine sediments
912 during glacier surges.

913 • Investigation of the sediment-landform assemblages within the terrestrial moraine areas
914 reveals that at least two separate phases of glacier advance are recorded. This is based
915 on the identification of an additional, previously unrecognised, ice-contact zone
916 characterised by a transition from subglacial sediments to proglacially/submarginally

917 deformed sediments. We infer that this records an inner, younger advance that did not
918 extend to the distal parts of the moraine system.

919 • Radiocarbon dating of shells embedded in the surface of the glaciotectonic composite
920 ridge systems at the distal margins of the terrestrial moraine area indicate that the outer,
921 older advance occurred at ~1160 AD, rather than during the LIA as previously
922 suggested by Ottesen et al. (2008). We instead correlate the inner, younger advance to
923 the LIA and suggest it culminated in ~1890 based on the position of the calving
924 (retreating) glacier terminus mapped by Hamberg (1905) in 1898.

925 • We demonstrate that NGS has advanced at least five times in the Late Holocene: (1)
926 the recent surge advance of 2008-2016, (2) during the LIA at ~1890, (3) at ~1160 AD,
927 and (4) and (5) twice between 2.61 and 2.79 cal. kyr BP, as previously reported by
928 Kempf et al. (2013).

929 In addition to the recent 2008-2016 surge, the observed sediment-landform assemblages
930 associated with the four older advances are also consistent with surging. This work contributes
931 to the understanding of High-Arctic tidewater glacier dynamics, and in particular the frequency
932 and magnitude of surge advances, during the Late Holocene. Future work should focus on
933 combined marine and terrestrial investigations at the margins of other tidewater glaciers in
934 order to provide a more complete picture of the regional variability in Holocene glacier
935 advances in Svalbard.

936

937 **CRedit author statement**

938 **HL, DIB, SL:** Conceptualization, Methodology, Investigation, Writing – Original Draft. **DO,**

939 **AL:** Investigation, Resources, Writing – Review & Editing. **MH:** Methodology, Formal

940 Analysis, Writing – Review & Editing. **IDB:** Resources, Writing – Review & Editing. **CMB,**

941 **HS:** Writing – Review & Editing.

942

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955

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