

1 **Controls on bedrock bedform development beneath the Uummannaq Ice Stream onset zone, West**
2 **Greenland**
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4 Timothy P. Lane^{a,b*}, David H. Roberts^a, Brice R. Rea^c, Colm Ó Cofaigh^a, Andreas Vieli^d

5 ^a Department of Geography, South Road, Durham University, UK

6 ^b Department of Geography, Liverpool John Moores University, Byrom Street, Liverpool L3 3AF, UK.

7 ^c Geography and Environment, School of Geosciences, University of Aberdeen, Meston Building, Old
8 Aberdeen, AB24 3UE, UK

9 ^d Department of Geography, University of Zurich – Irchel, Winterthurerstr. 190, CH-8057 Zurich,
10 Switzerland

11

12 *Corresponding author Timothy Lane at Department of Geography, Liverpool John Moores
13 University, Byrom Street, Liverpool L3 3AF, UK. Tel: 00447817344057; Email: t.p.lane@ljmu.ac.uk.

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15

16 **Abstract**

17 This paper investigates the controls on the formation of subglacially eroded bedrock bedforms
18 beneath the topographically confined region upstream of the Uummannaq Ice Stream (UIS). During
19 the last glacial cycle, palaeoglaciological conditions are believed to have been similar for all sites in
20 the study, characterised by thick, fast-flowing ice moving over a rigid bedrock bed. Classic bedrock
21 bedforms indicative of glacially eroded terrain were mapped, including p-forms, roches moutonnées,
22 and whalebacks. Bedform long axes and plucked face orientations display close correlation (parallel
23 and perpendicular) to palaeo-ice flow directions inferred from striae measurements. Across all sites,
24 elongation ratios (length to width) varied by an order of magnitude between 0.8:1 and 8.4:1.
25 Bedform properties (length, height, width, and long axis orientation) from four subsample areas,

26 form morphometrically distinct populations, despite their close proximity and hypothesised
27 similarity in palaeoglaciological conditions.

28

29 Variations in lithology and geological structures (e.g., joint frequency; joint dip; joint orientation;
30 bedding plane thickness; and bedding plane dip) provide lines of geological weakness, which focus
31 the glacial erosion, in turn controlling bedform geometries. Determining the relationship(s) between
32 bedding plane dip relative to palaeo-ice flow and bedform shape, relative length, amplitude, and
33 wavelength has important ramifications for understanding subglacial bed roughness, cavity
34 formation, and likely erosion style (quarrying and/or abrasion) at the ice–bed interface. This paper
35 demonstrates a direct link between bedrock bedform geometries and geological structure and
36 emphasises the need to understand bedrock bedform characteristics when reconstructing
37 palaeoglaciological conditions.

38

39 *Keywords:* ice stream; glacial erosion; bedrock bedform; abrasion; quarrying; Greenland ice sheet

40 **1. Introduction**

41

42 *1.1. Importance of glacial bedforms*

43 An understanding of past ice sheet dynamics can significantly improve our understanding of current
44 and potential future changes to ice sheet mass balance in Greenland and Antarctica under predicted
45 climate-forcing scenarios. Ice streams are the most dynamic part of ice sheets, and they exert a
46 strong influence upon mass balance (MacAyeal, 1993; Ó Cofaigh et al., 2003; Sejrup et al., 2003;
47 Stokes et al., 2005). Bed conditions are a key control upon the functioning of ice streams and
48 thereby the mass balance and health of the entire ice sheet. In recent years bedform analyses have
49 been pivotal in allowing us to understand the basal conditions operating beneath ice streams
50 onshore (Stokes and Clark, 1999; 2001; Roberts and Long, 2005; Bradwell et al., 2008; Stokes et al.,
51 2009; Phillips et al., 2010) and offshore (Canals et al., 2002; Ó Cofaigh et al., 2002; Sejrup et al.,
52 2003; Evans et al., 2009; Hogan et al., 2010; 2013). Ice stream onset to trunk zone transitions are
53 recognised as being characterised by a transition in bedform type and characteristics relating to the
54 balance between erosion and deposition (Ó Cofaigh et al., 2002; Graham et al., 2009). Onset zones
55 are often dominated by bedrock bedforms (Kleman et al., 2008) where, on average over glacial
56 cycles, sediment evacuation exceeds sediment generation as erosion is widespread and ice velocities
57 typically are increasing. Trunk regions generally are dominated by soft-beds where, over glacial
58 cycles, sediment advection of the material eroded from the upper ice stream catchment leads to
59 deposition (e.g., Ó Cofaigh et al., 2002; Graham et al., 2009).

60

61 Sedimentary bedforms have been successfully used to decode ice stream dynamics in
62 palaeoglaciated regions through detailed mapping of ice flow directions, inference of basal thermal
63 regimes, delineation of fast and slow palaeo-ice flow regions, and improvement in our
64 understanding of contemporary subglacial processes (Clark, 1993; Stokes and Clark, 2001; Ó Cofaigh

65 et al., 2002; Dowdeswell et al., 2008; King et al., 2009). However, subglacial bedrock bedforms
66 remain poorly investigated and are not fully understood because of numerous glaciological and
67 geological complexities. A growing body of research is exploring the evolution of bedrock bedforms
68 and the implications for understanding ice sheet and ice stream dynamics. Studies have been
69 undertaken on submarine and on terrestrial landforms, reporting roches moutonnées and
70 whalebacks (e.g., Gordon, 1981; Sugden et al., 1992; Roberts and Long, 2005; Roberts et al., 2010;
71 Krabbendam and Glasser, 2011), large glacially moulded bedrock ridges often controlled by bedrock
72 structure (Roberts et al., 2010), mega-scale crag-and-tail forms (Jansson et al., 2003; Ottesen et al.,
73 2008), erosional mega-grooves (Bradwell et al., 2008; Roberts et al., 2010), crag and tails, and other
74 crudely streamlined bedforms with blunt stoss and tapered lee sides (Lowe and Anderson, 2002; Ó
75 Cofaigh et al., 2002; Ó Cofaigh et al., 2013). In central west Greenland, sedimentary bedforms are
76 sparse, and areal scour dominates low elevation terrain (< 1000 m above sea level, asl) (Sugden,
77 1974; Glasser and Warren, 1990; Roberts and Long, 2005), so the majority of terrestrial glacial
78 landforms are erosional. Glacially eroded bedforms are ubiquitous in these areas, and from this
79 evidence palaeo-ice sheet basal conditions can be reconstructed.

80

81 This paper investigates bedrock bedform evolution within the fjords that feed the onset zone of the
82 Uummannaq Ice Stream (UIS). Firstly, it explores the relationship between bedrock bedforms and
83 hypothesised ice flow conditions. Secondly, it considers the influence of local geology on bedform
84 formation, and finally it considers our broader understanding of subglacial bedrock bedform genesis
85 and evolution under ice streams.

86

87 *1.2. Glacial erosion and bedrock bedforms*

88 All features of mechanical glacial erosion are formed through either abrasion or quarrying (Boulton,
89 1974; Rea and Whalley, 1994), both processes indicative of warm-based ice sliding across its bed

90 (Rea, 2007). As defined by Rea (2007), abrasion is subglacial frictional wear at the bed as debris-
91 charged basal ice slides across it. This is enhanced as ice encounters a bedrock bump, thereby
92 driving particles held in the ice toward the bed, increasing abrasive potential (Boulton, 1974).
93 Erosion is achieved either through fracture of the bedrock by large particles (> 0.01 m) or through
94 polishing by finer (< 0.01 m) material (Hindmarsh, 1996; Glasser and Bennett, 2004). Quarrying is
95 reliant upon the presence of lee side cavities beneath the glacier, which may be water- or air-filled.
96 Glacier sliding focuses overburden pressure on the bed, upstream of the cavity (Iverson, 1991;
97 Hallet, 1996). This causes crack growth and block removal, especially when basal water pressures
98 are variable (Iverson, 1991). The magnitude of plucking is heavily reliant on local effective pressure
99 (difference between overburden pressure and cavity water pressure), bedrock structure, and cavity
100 dimension (Iverson, 2012). Cavity formation is favoured in areas of low overburden pressure and
101 fluctuating subglacial water pressures, most likely in areas of thin ice (Iverson, 1991). The most
102 commonly reported small to medium features related to glacial erosion and those which will be
103 dealt with in this paper are roches moutonnées and whalebacks. Roches moutonnées are
104 characterised by a smooth, often abraded, curved stoss slope and a steep lee side formed through
105 quarrying and plucking (Lindstrom, 1988; Glasser and Bennett, 2004; Roberts and Long, 2005).
106 Quarrying of blocks from lee sides of bedrock bumps is often facilitated by subvertical joints within
107 the bedrock (Lindstrom, 1988; Sugden et al., 1992). Conversely, whalebacks display smooth, curved
108 stoss and lee sides resulting from abrasion across the entire bedform. Whalebacks are hypothesised
109 to develop when cavity formation is suppressed and when continual ice-bed coupling can be
110 achieved, restricting opportunities for cavity formation and quarrying (Evans, 1996). Previous
111 publications provide a full discussion of the processes that form roches moutonnées and whalebacks
112 (Glasser and Bennett, 2004; Roberts and Long, 2005).

113

114 Bedrock structure is well known to exert a direct control on subglacial bedform development, type,
115 and morphology (Gordon, 1981; Rea and Whalley, 1994; Roberts and Long, 2005; Dühnforth et al.,
116 2010; Roberts et al., 2010; Krabbendam and Bradwell, 2011; Krabbendam and Glasser, 2011; Hooyer
117 et al., 2012). Variability in bedding plane strike, dip, and thickness can have a strong impact on
118 bedform evolution (Roberts et al., 2010). High bedrock hardness increases abrasion resistance
119 (Krabbendam and Glasser, 2011), while joint spacing can control bedform type by facilitating or
120 resisting quarrying (Dühnforth et al., 2010; Iverson, 2012). Joint orientation can control lee side
121 plucking, irrespective of palaeo-ice flow direction (Gordon, 1981; Rea and Whalley, 1996;
122 Krabbendam and Bradwell, 2011; Hooyer et al., 2012).

123

124 Often, bedrock bedforms in areas of palaeo-ice streaming have low elongation ratios (the ratio of
125 length to width: ELRs) ($< 10:1$), in contrast to soft sediment bedforms ($> 10:1$) (Roberts and Long,
126 2005; Roberts et al., 2010). In order for bedforms to develop high elongation ratios, a smooth (on
127 the length scale of the landforms) ice-bed interface is required. On bedrock beds this situation is
128 prevented by the strength of the bedrock and lack of sediment to smooth out the bed roughness.
129 Further complexities in bedform morphology are introduced as bedform development can occur
130 during different phases of a single glacial advance-and-retreat cycle through changes in ice-bed
131 coupling (Roberts and Long, 2005). This means that bedforms can be modified by multiple ice
132 advances of varying direction, developing complex, double-plucked bedforms (Roberts and Long,
133 2005; Roberts et al., 2010).

134

135 **2. Study site**

136 *2.1. Field sites*

137 The field site lies within the Uummannaq region of central west Greenland, at 71.30° to 72.00° N
138 (Fig. 1). The region covers $\sim 25,000$ km² and is topographically constrained by two large peninsulas:

139 Svartenhuk in the north and Nuussuaq in the south (Fig. 1). Repeated glaciation throughout the
140 Quaternary has produced a series of deep, coalescent fjords that broadly run east-west (Roberts et
141 al., 2013; Lane et al., 2014). Selective linear erosion has created a high relief landscape, with
142 plateaux summits reaching up to 2000 m asl and fjords reaching 1300 m below sea level. The high-
143 level plateaux terrain fosters contemporary cold-based ice caps. The study focused on Rink-Karrat
144 and Ingia Fjords (Fig. 1). In order to compare subglacial bedforms across settings of varying
145 geological structure, two subareas were selected in both Rink-Karrat and Ingia Fjords. Within Rink-
146 Karrat Fjord, subareas KA1 and KA2 are at 200-260 m asl on the eastern flank of Karrat Island (Fig.
147 2A). They are 50 km from the present margin of Rink Isbræ and 30 km from the present margin of
148 Umiámáko Isbræ. The island is small, $\sim 28 \text{ km}^2$, and bifurcates Rink-Karrat fjord with $< 300 \text{ m}$ deep
149 water to the north and 600-700 m deep water to the south, forming a large relief (500-600 m)
150 bedrock bump within the fjord. Based upon field observations and the differences in bedform
151 morphology, KA2 was further divided into KA2i and KA2ii. Subareas IN1 and IN2 are found at 100-
152 270 m asl, on a sloping $\sim 15 \text{ km}^2$ peninsula that forms part of the Ingia Fjord wall, 18 km from the
153 present margin of Ingia Isbræ. The depth of Ingia Fjord is not known in this region. All subareas
154 exist in regions of low relief and elevation, in regions characterised by intense areal scour. Sediment
155 cover is sparse, evidenced only by a series of gravelly lateral moraines on Karrat Island (Lane et al.,
156 2014).

157

158 *2.2. Geology*

159 The Uummannaq region is characterised by three distinct bedrock types. An Archean basement
160 forms the deeply incised inner fjord system in the east, Palaeozoic–Mesozoic sediments underlie
161 Igdlorssuit Sund, and Palaeogene volcanics bound the west of the region (Fig. 1) (Pedersen and
162 Pulvertaft, 1992; Garde and Steenfelt, 1999; Henriksen et al., 2000). The sites investigated for this
163 study are underlain by an Archaen basement and form part of the Nûkavsak Formation in the

164 Archaen Rinkian belt (Kalsbeek et al., 1998). Henderson and Pulvertaft (1987a) described the
165 Nûkavsak Formation as being composed of interlayered granular semipelite, pelitic schist, and
166 metagreywacke, interpreted as a sequence of turbidites with little lithological variation through their
167 vertical and horizontal extent. In the field they were seen to display multiple subvertical and
168 subhorizontal joints with beds of variable thickness (~5-20 cm). The metagreywacke was
169 infrequently interbedded with veins of less heavily jointed quartzite up to 50 cm thick. The
170 occurrence of quartzite bands was most pronounced in IN1 and IN2.

171

172 *2.3. Palaeoglaciological background*

173 During the Last Glacial Maximum (LGM), the Uummannaq region was dominated by the UIS, a large
174 ice stream system (Roberts et al., 2013) that reached the edge of the continental shelf (Ó Cofaigh et
175 al., 2013; Dowdeswell et al., 2014). During the last glaciation, outlets draining the northern sector of
176 the Uummannaq region (Ingia, Umiámáko, and Rink Isbræ) flowed south into Igdlorssuit Sund (Fig. 1)
177 and joined ice draining from the southern sector. These branches coalesced in Igdlorssuit Sund,
178 which became the palaeo-UIS onset zone; and confluent trunk ice flowed west through the
179 Uummannaq Trough to the shelf edge at the LGM. Geomorphological and geochronological data
180 have constrained the LGM upper limit of warm-based ice to 1400-1968 m asl within 50 km of the
181 present ice sheet margin and to > 1040 m asl in the outer fjords – close to Karrat Island (Roberts et
182 al., 2013; Lane et al., 2014). Higher elevation areas remained exposed as nunataks or were covered
183 by protective, cold-based ice caps (Roberts et al., 2013; Lane et al., 2014). Following the LGM, the
184 UIS began to retreat from the shelf edge by 14.9 cal. ka BP (Ó Cofaigh et al., 2013), unzipping and
185 retreating into individual fjords. Outlet glaciers reached their present margins after 8.7 ¹⁰Be ka in
186 the south and 5 ¹⁰Be ka in the north. A more detailed discussion of the deglacial chronology can be
187 found in Roberts et al. (2013) and Lane et al. (2014). At the LGM, all study locations were therefore
188 positioned within fjords that constituted upstream branches of the palaeo-UIS. Palaeoglaciological

189 reconstructions indicate that during the LGM ice would have remained broadly topographically
190 constrained within fjords (Lane et al., 2014), with overtopping of fjord mountains up to 1400 – 1968
191 m asl.

192

193 **3. Methods**

194 Initial mapping used 1:50,000 topographic maps, 1:100,000 geological maps (Henderson and
195 Pulvertaft, 1987b), and 1:150,000 aerial photographs (~5 m resolution) (Kort and Matrikelstyrelsen).

196 This initial assessment of the broad-scale topography identified regions of areal scour containing
197 bedrock bedforms for more detailed study. These regions were then mapped onto a topographic

198 base map. Subsequently, four areas were chosen for detailed field analysis (Fig. 2) based upon their
199 accessibility and the ubiquity of bedforms within them. In each area, 40 - 50 glacial bedforms were

200 identified and mapped using previously acknowledged criteria classifying them as roches
201 moutonnées (bedrock bedforms, displaying abraded stoss slopes and middle surfaces, plucked lee

202 faces with evidence of block removal) or whalebacks (bedforms with smooth stoss slopes and lee
203 slopes, evidence for widespread abrasion, and little/no evidence of plucking from lee slopes) and

204 defined by length as macro (> 100 m), meso (100– 10 m), or micro (< 10 m) features (Glasser and
205 Warren, 1990; Glasser and Bennett, 2004; Roberts and Long, 2005). Following this, bedform

206 features were recorded (bedform long axis orientation, direction of plucked face(s), and presence
207 and direction of striae). Bedform dimensions (length, width, and height) and long axis orientation

208 were measured using a tape measure, and ELRs were subsequently calculated. Transverse
209 wavelength (transverse distance between bedform crests - TW) was measured in each area using a

210 tape measure. This was used in conjunction with the bedform length to calculate bedform density
211 (bedforms/km²). Bedform density estimates and differences between subareas were validated

212 through bedform counting within known areas (1 km²) using aerial photographs. Bedrock lithology
213 was characterised using published reports from the region (Henderson and Pulvertaft, 1987b;

214 Kalsbeek et al., 1998), and bedrock structure was characterised within each study area in the field
215 through measurement of bedding plane thickness, strike and dip, and joint density. Data from
216 bedding planes and joints were plotted on lower hemisphere stereographic projections (stereonet),
217 and striae were plotted on rose diagrams.

218

219 Apparent dip (d_a) was calculated for each subarea (Fig. 3). The d_a is a measure of bedding plane dip
220 relative to palaeo-ice flow direction, indicated in this study by striae data. Apparent dip is routinely
221 used to measure the inclination of geological beds when not seen perpendicular to bedding plane
222 strike (Lisle, 2004). However, here we use it to measure the bedding plane dip experienced by
223 palaeo-ice flow. Unless ice flow was directly parallel ($d_a = 0$) or perpendicular ($d_a = \text{true dip}$) to
224 bedding plane strike, the bedding plane dip that the ice encountered (d_a) would be somewhere
225 between 0° and the true dip.

226

227 Apparent dip is calculated with the following equation: $d_a = \arctan (\tan(d_t) * \sin (S_{xs} - S_b))$, where d_a
228 = apparent dip; d_t = true dip; S_{xs} = ice direction; and S_b = mean bedding strike. Notably, $(S_{xs} - S_b)$ was
229 converted to an absolute number. A negative apparent dip is possible, which means that palaeo-ice
230 flow direction was up-dip, not down-dip.

231

232 **4. Results**

233 An overview of bedform data is presented in Table 1, with photographs from each study area shown
234 in Figures 4 and 5. Table 2 presents structural bedrock data, and Figure 6 presents idealised
235 schematic diagrams of bedforms from each study area.

236

237 *4.1. Karrat Island subarea 1 (KA1)*

238 Bedrock in KA1 is predominantly metagreywacke. Beds are 5-30 cm thick, and bedding plane
239 geometries are tightly clustered, striking SSW with an average dip angle of 42° (Fig. 7e). Two distinct

240 joint systems strike WNW and SW (Fig. 7e), with a mean dip of 74° and 9 joints m⁻². Fifty bedforms
241 were measured in KA1: 6 meso-scale whalebacks and 44 meso-scale roches moutonnées. Bedforms
242 range from 1.9 to 23.0 m in length, 1.6 to 4.0 m in width, and 0.9 to 9.0 m in height, giving a mean
243 elongation ratio (ELR) of 2.52:1. Length:height ratios are 7.6, reflecting the relatively low relief, long
244 cross profile of these classic roches moutonnées forms. The r^2 values shown in Figure 7 demonstrate
245 weak correlations between bedform variables in KA1. Length and width show a weak positive
246 correlation, with some bedforms becoming wider as they become longer. Height appears to have no
247 correlation with bedform length ($r^2 = 0.001$), but shows a weak to moderate correlation with
248 bedform width ($r^2 = 0.486$). Thus, taller bedforms are also wider, not longer. Throughout KA1,
249 bedforms are well developed but display some fragmentation and weathering across their surface.
250 Roches moutonnées display abrasion on stoss sides and display plucking on middle and lee slopes,
251 producing south westerly facing plucked sides (e.g., Figure 4b). Whalebacks are poorly developed.
252 Transverse wavelengths are up to 20 m, with densities of 190 – 240 bedforms km⁻². Striae directions
253 are tightly clustered, indicating ice flow at 21 - 201° (Fig. 8), in broad accordance with mean bedform
254 long axis orientation.

255 **Table 1.** Key morphological features of the bedforms mapped in this study from all four subareas.

256 See Figure 2 for area location. The key highlights the units used in the table.

Subarea	Alt. range	RM (n)	WB (n)	Long axis	Striae	Min. L	Max. L	Min. W	Max. W	Min. H	Max. H
KA1	200-280	44	6	220	201	1.9	23.0	1.6	4.0	0.9	9.0
KA2i	200-300	20	0	175	265	0.8	13.2	1.5	15.0	0.7	4.4
KA2ii	200-300	30	0	185	212/268	0.8	11.1	0.9	10.0	0.6	1.7
IN1	100-270	9	41	243	244	4.1	72.0	0.7	18.9	0.2	7.6
IN2	100-250	5	25	266	225	7.5	96.0	1.0	12.0	0.9	3.3

257

Subarea	ELR	L:H ratio	H:W ratio	Bedf. dens.	TW
KA1	2.52	7.61	0.39	192-240	20
KA2i	0.81	3.63	0.27	144-168	40
KA2ii	0.88	3.15	0.26	144-168	40
IN1	4.79	21.37	0.35	160-190	12
IN2	8.42	19.84	0.45	160-190	12

258

Key			
RM (n)	<i>Number of roches moutonnées sampled</i>	Min. H	<i>Minimum bedform height in metres</i>
WB (n)	<i>Number of whalebacks sampled</i>	Max. H	<i>Maximum bedform length in metres</i>
Long axis	<i>Mean bedform long axis in degrees</i>	ELR	<i>Elongation ratio</i>
Striae	<i>Mean striae orientation in degrees</i>	L:H Ratio	<i>Ratio of bedform length to height</i>
Min. L	<i>Minimum bedform length in metres</i>	H:W Ratio	<i>Ratio of bedform height to width</i>
Max. L	<i>Maximum bedform length in metres</i>	Bedf. dens.	<i>Average bedform density per km²</i>
Min. W	<i>Minimum bedform width in metres</i>	TW	<i>Average transverse wavelength</i>
Max. W	<i>Maximum bedform length in metres</i>		

259

260 **Table 2.** Bedding features measured for each sub area, include mean strike and dip (with 1 σ standard
 261 deviations), and apparent dip.

Subarea	Bedding strike	Std. dev. (1 σ)	Bedding dip	Std. dev. (1 σ)	Joint density (m ⁻²)	Apparent dip (d _a) (°)
KA1	196.4	15.5	40.6	12.2	9	3.9
KA2i	189.9	9.3	51.0	12.8	8	50.0
KA2ii	189.9	9.3	51.0	12.8	8	24.8 ^a
KA2ii	189.9	9.3	51.0	12.8	8	50.4 ^b
IN1	237.2	9.3	23.0	3.6	5	2.9
IN2	234.8	7.1	21.7	2.5	4	3.9

262 ^aUsing striae direction 212° (see Table 1).

263 ^bUsing striae direction 268° (see Table 1).

264

265 4.2. Karrat Island subarea 2 (KA2)

266 As in KA1, bedrock in KA2 is mainly metagreywacke with some politic schist, though this was not
267 seen on bedform surfaces. Bedding planes are tightly clustered; striking SSW with a mean dip of 51°
268 (Fig. 7). Subvertical joint systems strike S and W, with 8 joints m⁻². Though the area was subdivided
269 into KA2i and KA2ii, exposed bedrock was indistinguishable across KA2. Throughout KA2, bedforms
270 are characterised by small, individual, rectilinear, meso-scale roches moutonnées (cf. Roberts and
271 Long, 2005). In comparison to KA1, bedforms display less evidence of abrasion on their stoss
272 surfaces. Plucked faces are present in KA2, but they are sloping (Fig. 4d), in contrast to the stepped
273 lee side plucked faces of KA1. Bedforms display a transverse wavelength of 40 m and densities of
274 144 - 168 bedforms km⁻². After field observations (based upon variations in striae direction and
275 number of plucked faces), it was decided that bedforms in KA2 represent two distinct bedform
276 populations. As a result KA2 was subdivided into KA2i and KA2ii (see Figure 2a). Comparison of the
277 morphometric measurements from KA1 and KA2 demonstrate that bedforms from these areas lie in
278 three distinctly different populations (Figs. 7b-d), with bedforms in KA2 shorter, wider, and higher
279 than those in KA1.

280

281 Striae on roches moutonnées in KA2i ($n = 40$) display evidence for unidirectional ice flow (85 - 265°)
282 (Figs. 2a and 8; Table 1), almost perpendicular to bedform long axis orientation (175 - 355°). Roches
283 moutonnées were short (0.8 - 13.2 m) and wide (1.5 - 15.0 m), resulting in mean ELRs of 0.8:1.
284 Plucked faces are orientated west, concordant with the reconstructed ice flow direction and bedding
285 plane and joint directions. Westerly plucked faces are glacially abraded (Fig. 4d), with faces
286 conformable to bedding plane orientation.

287

288 Bedforms within KA2ii ($n = 10$) display evidence for two distinct phases of cross-cutting ice flow, 1
289 (32 - 212°) and 2 (88 - 268°) (Fig. 8; Table 1). Mean bedform long axis orientation is 10 - 190°,

290 oblique to the striae direction 1 (32 - 212°), and perpendicular to striae direction 2 (88 - 268°).
291 Within KA2ii, the striae show a cross-cutting relationship, with the 88 - 268° set appearing
292 superimposed upon the 32 - 212° set, suggesting switching of ice direction over the subarea.
293 Bedforms were of similar length to those from KA2i (0.80 – 11.10 m), but narrower (0.90 – 10.00 m),
294 with mean ELRs of 0.88:1. Bedforms from KA2ii display abraded lee side faces and double plucked
295 faces, oriented west (268 – 277°) and south-southwest (185 - 215°), broadly concordant with striae
296 directions. As in KA2i, west plucked faces were glacially abraded and conformable with bedding
297 plane dip and strike. In contrast to this, SSW-facing plucked faces are oblique to bedding plane and
298 to primary joint orientations and display blocky, stepped faces.

299
300 Bedform length:height ratios in KA2i are 3.63, and 3.15 in KA2ii, indicating lower relief bedforms in
301 KA2ii. However, bedforms from KA2ii show far less variation in height compared to width, reaching
302 a maximum height of 1.70 m. Throughout KA2, r^2 values reveal weak correlations between bedform
303 characteristics. Bedform width has no relationship to either bedform height ($r^2 = 0.008$) or length (r^2
304 = 0.045). A weak positive relationship exists between bedform length and height, suggesting that
305 some longer bedforms are taller, not wider.

306

307 *4.3. Ingia subarea 1 (IN1)*

308 Bedrock in IN1 is characterised by thinly bedded (5-20 cm) metagreywacke, striking SW with an
309 average dip of 22° (Fig. 9). Well-developed joint systems strike SW and SSE, with an average dip of
310 72°, and joint density of 5 joints m⁻². In places the metagreywacke contained massive bands of
311 quartzite up to 50 cm thick. Bedforms in IN1 are dominated by poorly developed meso-scale
312 whalebacks ($n = 41$) (cf. Roberts and Long, 2005), with some meso-roches moutonnées ($n = 9$).
313 Occasional microscale whalebacks were superimposed upon the larger bedforms, forming crude
314 whaleback swales. Swales are dense groups (or swarms) of whalebacks or roches moutonnées,

315 often superimposed upon one another (cf. Roberts and Long, 2005). Bedform long axis orientations
316 are tightly clustered at 74 - 254°. Bedforms are 4.1 – 72.0 m in length, 0.7 - 18.9 m in width, and 0.2
317 - 7.6 m in height (Figs. 9a-d). Mean elongation ratio (ELR) was 4.79:1. Evidence of abrasion is
318 present on the stoss and lee sides of all whalebacks, and lee side plucking is recorded on roches
319 moutonnées. All bedforms (whalebacks and roches moutonnées) also display plucking on their
320 lateral, south eastern faces (Figs. 5c and d). As a result, the bedforms were highly asymmetrical in
321 transverse profile. Transverse wavelength is 12 m, with a mean density of 160-190 bedforms km⁻².
322 Throughout IN1 striae are consistent, indicating a NE-SW (63 -243°) ice flow direction, subparallel to
323 fjord axis/thalweg. All relationships between bedform dimensions show weak relationships in IN2,
324 with r^2 values from 0.131 (width vs. height) to 0.290 (length vs. height) (Fig. 9).

325

326 4.4. *Ingia subarea 2 (IN2)*

327 The metagreywacke found in IN2 is lithologically comparable to bedrock in IN1. It is thinly bedded,
328 with beds reaching 20 cm thick. Bedding planes strike WSW with a mean dip of 22° (Fig. 9e). Joint
329 systems are well developed and strike SW and S, with an average dip of 62° (Fig. 9e) and joint
330 density of 4 joints m⁻² (Table 1). The metagreywacke contained bands of quartzite up to 50 cm thick
331 in places.

332

333 As in IN1, bedforms are characterised by poorly developed meso-scale whalebacks ($n = 25$) (cf.
334 Roberts and Long, 2005) with occasional smaller bedforms superimposed upon them. Bedform
335 cross-profile throughout IN2 is highly asymmetrical. Rare roches moutonnées ($n = 5$) were recorded.
336 Bedform ELRs from IN2 (8.42:1) are the highest in this study. Glacial polish is evident on stoss and
337 lee positions across bedforms, and where present, plucking is focused along the lateral, southern
338 flanks. The transverse wavelength of the bedform ridges is 12 m, giving an average density of 160-
339 190 bedforms km⁻². As in IN1, at site IN2 striae show that palaeo-ice flow was 45 - 225°, comparable

340 to the mean orientation of bedform long axes of 86 - 266° (Table 1). Striae were very rare, resulting
341 in 10 measurements from IN1 and 12 from IN2. Ratios of length:height are 21.37 and 19.84 for IN1
342 and IN2, respectively; and height:width ratios are 0.35 and 0.45. Combined with the similarity in
343 long axis orientation, this demonstrates that, in contrast to the distinct populations of KA1 and KA2,
344 IN1 and IN2 represent a similar bedform population (Figs. 9b-d). Overall, r^2 values from IN2 are the
345 highest of all subareas. Relationships between length and width ($r^2 = 0.382$) and length and height
346 ($r^2 = 0.201$) remain relatively weak, but a weak-to-moderate positive correlation is found between
347 width and height ($r^2 = 0.477$).

348

349 **5. Discussion**

350 *5.1. Bedform relationship to ice flow*

351 The results appear directly analogous to those reported from other west Greenland palaeo-ice
352 stream beds (Roberts and Long, 2005; Roberts et al., 2010), with low ELRs (2.8 - 3.7:1) and high
353 bedform density (> 200 bedforms/km²). Evidence for abrasion and plucking was found in all
354 subareas, but in varying degrees of significance. It is generally thought that plucking is the more
355 efficient agent of bedrock erosion (Briner and Swanson, 1998; Dühnforth et al., 2010). However,
356 others have suggested that this is a generalisation (Krabbendam and Glasser, 2011), with erosional
357 efficiency strongly dependent on local bedrock properties and structures, meaning that small
358 variations in bedding structure result in large changes in erosion type. Owing to the absence of
359 empirical data regarding the amount of erosion from this study, conclusions into the relative
360 efficiency of either erosion method cannot be made.

361

362 Based upon regional geomorphological evidence (Lane et al., 2014), all field sites in this study would
363 have been covered by 700 - 1000 m of ice during the last glaciation. Palaeo-ice flow velocities
364 through Rink-Karrat and Ingia Fjord are assumed to have been high because of ice streaming through

365 the over-deepened Igdlorssuit and Uummannaq troughs (Ó Cofaigh et al., 2013; Roberts et al.,
366 2013). Striae show evidence for unidirectional ice flow (21 - 201°), sub-parallel to mean bedform
367 long axis (Table 1; Figure 8), and so bedforms in KA1 show a clear relationship to palaeo-ice flow
368 direction. Roches moutonnées show clear evidence of stoss side abrasion and extensive plucked
369 southwest-facing lee side faces – consistent with ice flow direction. In KA2i, unidirectional ice flow
370 (85 - 265°) was perpendicular to bedding strike, forming short rectilinear bedforms with abraded
371 stoss sides and westerly plucked lee side faces. Two clear phases of ice flow are recorded for KA2ii
372 (32 - 212° and 88 - 268°; Table 1; Figure 8). Because of the cross-cutting relationship mentioned in
373 section 4.2, all striae are assumed to represent the deglacial phase of ice flow across Karrat Island,
374 with the 32 - 212° formed first, overprinted by the 88 - 268° set as ice became topographically
375 constrained during deglaciation. The ELRs vary between KA1 and KA2i/KA2ii (2.52:1 and
376 0.80:1/0.88:1, respectively). However, given their close proximity (< 1.5 km), it is unlikely that ice
377 thickness and basal thermal conditions were significantly different to have altered bedform
378 geometry.

379

380 Bedform long axes in IN1 and IN2 are subparallel to palaeo-ice flow, suggesting that bedforms
381 represent a similar response to their palaeoglaciological conditions to bedforms in KA1. Despite a
382 uniform ice flow direction as recorded in striae over the two areas (64 - 244°), bedform long axes
383 show a small switch from 74 - 254° (IN1) to 86 - 266° (IN2). It is likely that this change in long axis is
384 because of the variation in ice flow direction associated with ice flow bending around the
385 topographic obstacle of the peninsula, as well as the bend within Ingia Fjord. As opposed to the
386 dominance of roches moutonnées in KA1 and KA2 (94 out of 100), 66 of 80 bedforms in IN1 and IN2
387 are whalebacks. Previously it has been suggested that whalebacks are formed through cavity
388 suppression beneath thick, fast-flowing ice (Evans, 1996). However, the southeast flanks of the

389 whaleback forms have undergone lateral plucking, a process requiring the presence of sub-glacial
390 cavities.

391

392 Ice flow direction has exerted the main control upon bedform formation indicated by the orientation
393 of bedform long axes and plucked faces. The multidirectional ice flow in KA2ii has caused the
394 development of dual plucked lee face orientations and bedforms with low ELRs, as reported in a
395 number of other studies (e.g., Roberts and Long, 2005; Roberts et al., 2010). Apparently, therefore,
396 bedforms in areas of unidirectional ice flow (KA1, KA2i, IN1, and IN2) display higher ELRs. This is a
397 result of a single ice flow direction promoting the development of elongate bedforms, as structural
398 weaknesses can be continuously exploited. In contrast, multidirectional ice flow is able to erode
399 variably oriented structural weaknesses with high angles, forcing elongation ratios to remain low.
400 Although this hypothesis is applicable for most areas in this study, it does not apply to KA2ii, where
401 bedforms are extremely short, with ELRs < 1:1, despite unidirectional ice flow and comparable
402 palaeo-ice thicknesses to other subareas.

403

404 *5.2. Bedform relationship to geological structure*

405 Subglacial bedform geometry is known to be strongly controlled by bedrock structure (Gordon,
406 1981; Roberts and Long, 2005; Dühnforth et al., 2010; Krabbendam and Glasser, 2011; Hooyer et al.,
407 2012). As described above, bedding plane strike, joint orientation and joint spacing vary greatly
408 between Karrat Island and Ingia (Figs. 7 and 9; Table 2), and these differences have exerted control
409 on bedform morphology.

410

411 In KA1, KA2i, and KA2ii, bedding plane and joint characteristics have had a demonstrable impact
412 upon bedform properties. In KA1 the primary joint set is subvertical and strikes WNW, orthogonal to
413 palaeo-ice flow direction (Fig. 6), facilitating widespread lee side plucking of bedding plane and joint

414 defined blocks. In KA1 bedform width is likely to have been controlled by joints sub-parallel to ice
415 flow (striking SW) and moderately dipping bedding planes striking SSW – acting as lines of weakness
416 along which ice has been able to erode. Plucked lee side development in KA2i has been facilitated
417 by extensive, well-developed, SSW-striking bedding planes and S-striking joint systems,
418 perpendicular to palaeo-ice flow direction (85 - 265°), and dipping down-ice. The orientation and
419 steep bedding plane dip is thought to have facilitated ice-bed separation in the lee sides of
420 bedforms, allowing the exploitation of bedding planes and joints subparallel to palaeo-ice flow by
421 plucking directly along metagreywacke beds. This process has generated the down-ice sloping lee
422 side plucked faces. The less extensive east to west joint system has aided plucking of blocks
423 bounded by down-ice dipping bedding surfaces, N-S and E-W trending joints. The low density of this
424 E-W joint set allowed bedforms to maintain their large width and consequently low ELRs. The wide
425 spacing of the joints provided fewer points of structural weakness at which erosion can be focused
426 to breach the laterally extensive metagreywacke beds. Thus, the low ELRs (~0.8:1) of bedforms in
427 KA2i are a direct result of bedrock structure.

428

429 Bedforms in KA2ii exhibit multiple plucked faces, recording plucking in response to ice flow from the
430 east and northeast. The striae chronology outlined above is assumed to record deglaciation across
431 Karrat Island. While bedform development is assumed to be a continuous process, the bedforms
432 observed in KA2ii are most simply related to the two striae directions observed. Though uncertain,
433 the chronology proposed above based on cross-cutting striae (see section 5.1) and on increasing
434 topographic control on ice flow during deglaciation. If the chronology is incorrect, the events would
435 be reversed, making little difference to the landform development model below. During NE-SW
436 directed ice flow SSW- and S-facing faces were plucked, with block removal facilitated by E-W and N-
437 S trending joints and by SSW-striking bedding planes. This phase of ice flow was oblique to joint
438 orientation, and the resulting plucked faces are not well developed. Subsequently, E-W ice flow

439 caused plucking of westerly oriented faces, again removing blocks bounded by the E-W and N-S
440 trending joints, but with block removal in a westerly direction. Roberts et al. (2010) reported
441 bedforms formed through bidirectional ice flow, transverse and subparallel to the bedding plane
442 strike and dip, directly comparable to KA2ii. As in KA2ii, the distinct geological and bidirectional ice
443 flows have produced short, wide bedforms with low ELRs and multiple plucked faces (Roberts et al.,
444 2010).

445

446 In IN1 and IN2, bedrock structure has exerted a strong control upon bedform type (roches
447 moutonnées/whaleback), length, width and orientation. Although little lee side plucking was
448 documented at IN1 and IN2, almost all bedforms displayed evidence for extensive block removal
449 from their steep, lateral SE faces (Figs. 5D, E). This plucking has been facilitated by exploitation of
450 the exposed ends of SW-striking bedding planes, with block removal occurring along subvertical
451 joints. Bedforms on Ingia are dominated by whalebacks (66 out of 80). As discussed above, their
452 presence and their abundance is unlikely to be a function of ice flow as ice flow characteristics
453 between Ingia and Karrat Island are similar. Instead it is due to bedrock properties. They display
454 ubiquitous stoss and lee side abrasion and little evidence of lee side plucking, despite the presence
455 of joints running transverse to palaeo-ice flow. This absence of evidence of lee side plucking is
456 interpreted to be the result of three factors. Firstly, strike of bedding planes and the majority of
457 joints are sub-parallel to palaeo-ice flow, with low transverse joint density (~ 2 joints/m² in
458 comparison to 6 transverse joints/m² in KA1 and KA2). Secondly, a number of bedforms display thick
459 (30-50 cm) bands of quartzite, interbedded with metagreywacke. In places these cap the bedforms
460 or outcrop at the bedform surface. The crystalline quartzite contains far fewer transverse structural
461 weaknesses (~ 1 joints/m²) than metagreywacke and is harder and more resistant to erosion.
462 Therefore, where present, quartzite has limited plucking creating abraded stoss and lee side forms
463 (whalebacks). Thirdly, the thin bedding planes in IN1 and IN2 have an apparent down-ice dip of 2.88

464 – 3.87° relative to palaeo-ice flow (see section 5.3 for a discussion of this). As the metagreywacke in
465 IN1 and IN2 is thinly bedded (5-15 cm), and thus less resistant to subglacial erosion than thicker
466 bedded metagreywacke, erosion removes material along the thin beds. This therefore prevents the
467 development of a classic stepped lee side face, preserving the sloped face and whaleback form.

468

469 A similar style of bedrock plucking to that recorded on the lateral flanks of bedforms in IN1 and IN2
470 has been outlined by Krabbendam and Bradwell (2011), termed lateral plucking. In their study,
471 erosion by this process led to the development of a series of negative relief megagroove features,
472 not the positive relief bedforms observed in the present study (Krabbendam and Bradwell, 2011).
473 During lateral plucking, cavities develop on steep vertical surfaces, lateral to the main bedform
474 surface and approximately parallel to the long axis. Erosion occurs along joints and bedding planes
475 (allowing block loosening and translocation) about a vertical axis, leading to erosion perpendicular to
476 ice flow direction (Krabbendam and Bradwell, 2011). A relative absence of protuberances on
477 megagroove floors, or roches moutonnées/whaleback upper surfaces, led to the suggestion that
478 erosion was abrasion-dominated (Krabbendam and Bradwell, 2011). This is despite the variable
479 bedding dip (5-40°) reported by Krabbendam and Bradwell (2011), which is likely to have exposed
480 lines of weakness to plucking. In IN1 and IN2, lateral plucking has acted to control bedform width
481 through rock removal from lateral flanks. Bedforms in IN1 and IN2 only display this on south eastern
482 flanks where exposed bedding planes can be readily exploited. It appears that the type of plucking
483 (lee side or lateral) and the resultant landform (positive bedforms or negative grooves) are highly
484 dependent upon local geological structure (e.g., bedding plane dip, bed thickness, joint orientation,
485 joint spacing) and probably the form of the surface prior to the most recent glaciation.

486

487 The distinct offset between plucked-face orientation and palaeo-ice flow direction (which was
488 observed in KA2i, KA2ii, IN1, and IN2) has also been reported by other studies (Gordon, 1981; Rea

489 and Whalley, 1996; Krabbendam and Bradwell, 2011; Hooyer et al., 2012). Plucking is facilitated by
490 pre-glacial bedding planes and joints and by bedrock rock bridges (Hooyer et al., 2012) (Fig. 10).
491 Block removal occurs when rock bridges between joint-bounded blocks fail; meaning that plucked-
492 faced distribution preferentially follows joint orientation as opposed to ice flow direction (Hooyer et
493 al., 2012).

494

495 *5.3. Importance of apparent bedding dip relative to ice flow direction*

496 Bedding plane dip angle was outlined by Gordon (1981) as one of the factors affecting bedform
497 characteristics and was considered by Rea and Whalley (1996) in their plucking calculation. More
498 recently, Krabbendam and Bradwell (2011) discussed its impact upon lee side plucking. Over a
499 regional scale, Kelly et al. (2014) demonstrated the importance of bedding plane dip on the
500 prevalence of glacial steps, scour surfaces, and overdeepenings. From this study, the strike and dip
501 of bedding *relative* to palaeo-ice flow direction(s) (d_a) is a significant factor controlling bedform relief
502 and morphology (Table 2). We have calculated this as apparent dip, as outlined above (section 3).
503 When compared to bedform morphometry, a relationship exists between measured ELR and
504 calculated apparent dip, with high d_a correlated to low ELR. This preliminary data set provides an
505 important insight into the potential relationships between apparent dip and bedform morphometry.
506 While the interpretations are currently tentative, these relationships are discussed below.

507

508 Based on these data, likely bedform characteristics under different apparent bedding plane dip
509 configurations can be predicted (Fig. 11). When d_a is low relative to palaeo-ice flow (either up- or
510 down-ice), resultant bedforms are low in relief and amplitude, with high ELRs (e.g., IN1 and IN2) (Fig.
511 11). Low d_a , either perpendicular or parallel to flow, favours low form drag (Fig. 11), offering a
512 pathway of low resistance for basal ice. This leads to the maintenance of a bed with low relief and
513 roughness, reducing the difference in stress between stoss to lee faces, decreasing the likelihood of

514 quarrying/plucking, and promoting the development of long bedforms. This low form drag is
515 particularly likely to occur in regions of shallow dipping up-ice bedding planes. Here, bedforms
516 produced through glacial erosion will be roche moutonnée in form, with high ELRs (Fig. 11), the
517 smooth up-ice surfaces promoting low bed roughness. Bedforms in areas of shallow, down-ice
518 dipping bedding planes will be whaleback in form, also with long ELR (Fig. 11). In this setting the top
519 surfaces of bedforms have the potential to detach as a slab. In contrast, if bedding plane dip relative
520 to palaeo-ice flow is high, resultant bedforms have larger amplitudes and display low ELRs (e.g.,
521 KA2), increasing along bed stresses leading to the formation of high relief roches moutonnées
522 through widespread lee side plucking (Fig. 11). Such settings have high roughness leading to high
523 basal drag. If bedding planes dip up-ice, lee side cavity formation and subsequent plucking is
524 encouraged, producing classic roches moutonnées (Fig. 11) (Benn and Evans, 2010). Conversely, if
525 the bedding planes dip down-ice, roches moutonnées with smooth non stepped lee faces form (Fig.
526 11) (e.g., KA2i). Resultant bedforms are likely to have low ELRs and high amplitude. Variability in d_a
527 relative to ice flow only controls bedform length, so the development of high ELRs is dependent
528 upon bedform. On a smaller scale, joints are likely to act in a similar way to bedding planes. Thus
529 bedding plane and joint orientation relative to palaeo-ice flow is also able to control bed roughness
530 on the scale of a single bedform, and multiple structures orientated at high angles to palaeo-ice flow
531 will encourage quarrying. However, these finer scale effects cannot be resolved using data from this
532 study. Although the difference in dip angle has been referred to here as 'low' and 'high', in reality
533 ice flow relative to bedding plane dip and bedform properties (ELR/bedform type) lie on a
534 continuum. Increases in the dip angle are therefore hypothesised to gradually decrease bedform
535 length until a dip close to 90° is reached.

536

537 **6. Conclusion**

538 This study has investigated the controls on bedrock bedform development in two neighbouring
539 fjords in central West Greenland. These lie in the upstream reaches of the palaeo-UIS and are
540 believed to have experienced topographically constrained, fast ice flow during the LGM.
541 Reconstructions have suggested that palaeoglaciological conditions were similar for all sites in the
542 study, being characterised by thick, fast-flowing ice moving over a rigid bedrock bed. Palaeo-ice flow
543 direction has exerted a first-order control on bedform orientation, with the majority of areas
544 displaying bedform long axes sub-parallel to ice flow direction, as indicated by striae evidence.
545 Bedforms in areas that experienced multiple ice flow directions (KA2ii) display multiple plucked
546 faces, resulting in widespread bedform shortening, as reported in other studies. In contrast,
547 subareas that experienced unidirectional ice flow displayed longer bedforms. However, the
548 variability in bedform morphology cannot be explained solely by palaeo-glacier dynamics. Instead,
549 pre-glacial joint and bedding plane orientations have controlled bedform width and height through
550 plucking. Bedforms on Karrat and Ingia displayed evidence for areally extensive abrasion and
551 plucking.

552

553 Elongation ratios varied in this study between 0.81:1 and 8.42:1. Owing to the uniformity of
554 hypothesised palaeoglaciological conditions across the study areas, bedform ELR appears to be
555 controlled by the apparent dip (d_a). Shallow d_a promotes low bed roughness and high ELRs, whereas
556 steep d_a encourages high roughness and low ELRs. Whaleback bedforms were found in areas of
557 shallow, down-ice dipping bedrock, with low joint densities. This has resulted in plucking along
558 bedding planes, with some abrasion of lee sides, creating whaleback forms. The results of this
559 further highlight the need for further investigation on bedrock bedform characteristics before using
560 them to investigate palaeo-ice flow dynamics.

561

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570

571

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707

Table 1. Key morphological features of the bedforms mapped in this study from all four subareas.

See Figure 2 for area location. The key highlights the units used in the table.

Subarea	Alt. range	RM (n)	WB (n)	Long axis	Striae	Min. L	Max. L	Min. W	Max. W	Min. H	Max. H
KA1	200-280	44	6	220	201	1.9	23.0	1.6	4.0	0.9	9.0
KA2i	200-300	20	0	175	265	0.8	13.2	1.5	15.0	0.7	4.4
KA2ii	200-300	30	0	185	212/268	0.8	11.1	0.9	10.0	0.6	1.7
IN1	100-270	9	41	243	244	4.1	72.0	0.7	18.9	0.2	7.6
IN2	100-250	5	25	266	225	7.5	96.0	1.0	12.0	0.9	3.3

Subarea	ELR	L:H ratio	H:W ratio	Bedf. dens.	TW
KA1	2.52	7.61	0.39	192-240	20
KA2i	0.81	3.63	0.27	144-168	40
KA2ii	0.88	3.15	0.26	144-168	40
IN1	4.79	21.37	0.35	160-190	12
IN2	8.42	19.84	0.45	160-190	12

Key

RM (n)	<i>Number of roches moutonnées sampled</i>	Min. H	<i>Minimum bedform height in metres</i>
WB (n)	<i>Number of whalebacks sampled</i>	Max. H	<i>Maximum bedform length in metres</i>
Long axis	<i>Mean bedform long axis in degrees</i>	ELR	<i>Elongation ratio</i>
Striae	<i>Mean striae orientation in degrees</i>	L:H Ratio	<i>Ratio of bedform length to height</i>
Min. L	<i>Minimum bedform length in metres</i>	H:W Ratio	<i>Ratio of bedform height to width</i>
Max. L	<i>Maximum bedform length in metres</i>	Bedf. dens.	<i>Average bedform density per km²</i>
Min. W	<i>Minimum bedform width in metres</i>	TW	<i>Average transverse wavelength</i>
Max. W	<i>Maximum bedform length in metres</i>		

Table 2. Bedding features measured for each sub area, include mean strike and dip (with 1σ standard deviations), and apparent dip.

Subarea	Bedding strike	Std. dev. (1σ)	Bedding dip	Std. dev. (1σ)	Joint density (m^{-2})	Apparent dip (d_a) ($^\circ$)
KA1	196.4	15.5	40.6	12.2	9	3.9
KA2i	189.9	9.3	51.0	12.8	8	50.0
KA2ii	189.9	9.3	51.0	12.8	8	24.8 ^a
KA2ii	189.9	9.3	51.0	12.8	8	50.4 ^b
IN1	237.2	9.3	23.0	3.6	5	2.9
IN2	234.8	7.1	21.7	2.5	4	3.9

^aUsing striae direction 212° (see Table 1).

^bUsing striae direction 268° (see Table 1).

Fig. 1. (A) Overview topographic map of the Uummannaq region. Altitudes have been extracted from ASTER data and ocean floor bathymetry is from IBCAO data (Jakobsson et al., 2012). Broad geological zones of bedrock are shown, separated by white dashed lines. These are: (1) Palaeogene basalts; (2) Palaeozoic-Mesozoic sediments; (3) Archean basement (adapted from Henderson, 1971); (B) enlargement of Karrat Island, showing local geology; (C) enlargement of the local geology in Ingia Fjord (from Henderson and Pulvertaft, 1987a; b).

Fig. 2. Aerial photographs from (A) Karrat Island and (B) the peninsula Ingia Fjord, with dashed boxes indicating study areas. Rose diagrams show striae measurements from specific sites across Karrat Island and Ingia peninsula. Average bedform long axes and secondary axes orientations are shown in white crosses.

Fig. 3. Schematic diagram of the calculation of the apparent bedding plane dip (d_a) when palaeo-ice flow direction is taken into account. Other variables are: d_t = true dip; S_{xs} = ice direction; S_b = mean bedding strike.

Fig. 4. Photographs of bedforms from study sub-areas with long-axis orientations (grey arrows). White lines highlight bedform outlines for clarity when viewing: (A) aerial image showing locations of the photographs; (B) roche moutonnées from KA1, with abraded stoss sides and heavily plucked lee sides (left of image). Frequent striae were found across upper faces. Palaeo-ice flow right to left; (C) example of cross-cutting striae found within KA2ii; (D) short, rectilinear roche moutonnées from KA2ii. Multidirectional striae and plucked faces suggest multiple palaeo-ice flow directions.

Fig. 5. Photographs of bedforms from study sub-areas with long-axis orientations (grey arrows). White lines highlight bedform outlines for clarity when viewing: (A) aerial image showing locations of the photographs; (B) field of smooth whalebacks and occasional roche moutonnées in IN1: palaeo-ice flow from right to left of the image; (C) cross profile of clearly asymmetrical whalebacks from IN2 with palaeo-ice flow toward the camera. Little lee-side plucking is evident; (D) side view of a whaleback bedform in IN1. Gently dipping to sub horizontal bedding planes are evident, and the lee-side morphology of the bedform conforms to bedding plane dip. The bedform flank facing the camera displays clear evidence of block removal through 'lateral plucking'; (E) another view of the lateral face of a representative bedform from IN1, with lateral plucking, and the smoothed long-profile of the bedform visible.

Fig. 6. Schematic diagrams representative of the bedforms found in each sub-area of this study. Black lines represent bedding planes and grey dashed lines represent principle joint sets. This is a diagrammatic representation of bedform morphology, and bedding structures are not in precise locations.

Fig. 7. Bedform data from KA1 (black triangles) and KA2 (open squares): (A) scatter plot of width against length; (B) scatter plot of height against length; (C) scatter plot of elongation ratio (ELR) against length; (D) scatter plot of width against height; (E) bedding strike and joint dip orientation plotted in stereonet as poles to planes, with several planes to indicate dominant bedding direction.

Fig. 8. Striae data from all sub-areas plotted in rose diagrams. Note that striae in IN1 and IN2 were sparse, and poorly preserved on bedrock surfaces.

Fig. 9. Bedform data from IN1 (black triangles) and IN2 (open squares): (A) scatter plot of width against length for IN1 and IN2; (B) scatter plot of height against length for IN1 and IN2; (C) scatter plot of elongation ratio (ELR) against length for IN1 and IN2; (D) scatter plot of width against height for IN1 and IN2; (E) bedding strike and joint dip orientation plotted in stereonet as poles to planes, with several planes to indicate dominant bedding direction.

Fig. 10. Schematic diagram showing the development of bedforms in each sub-area, based on an idealised version of the joint and bedding systems. The schematic shows the planview of bedding structures, side profile of bedding plane dip, and planview of resultant bedforms.

Fig. 11. Model of bedform formation in regions of varied relative up- or down-ice bedding plane dip and hypothesised resulting bedforms. Black arrows represent ice flow direction. Green areas indicate surfaces dominated by glacial abrasion; red areas indicate surfaces dominated by lee-side plucking.

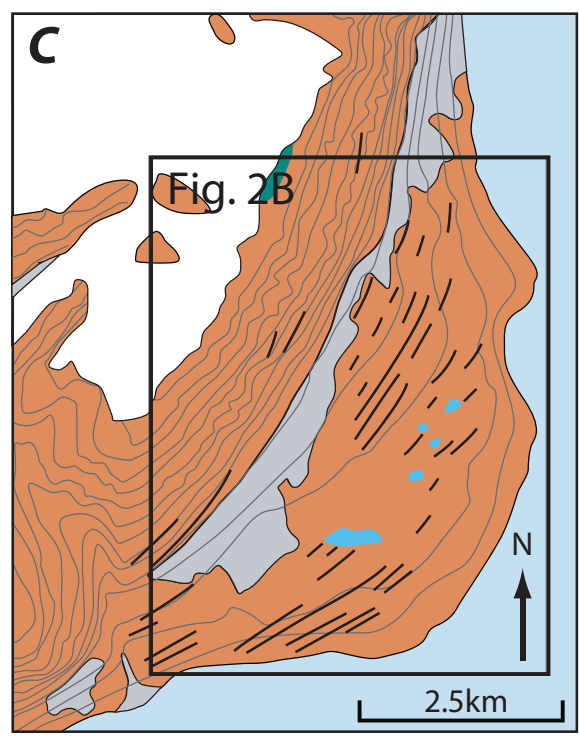
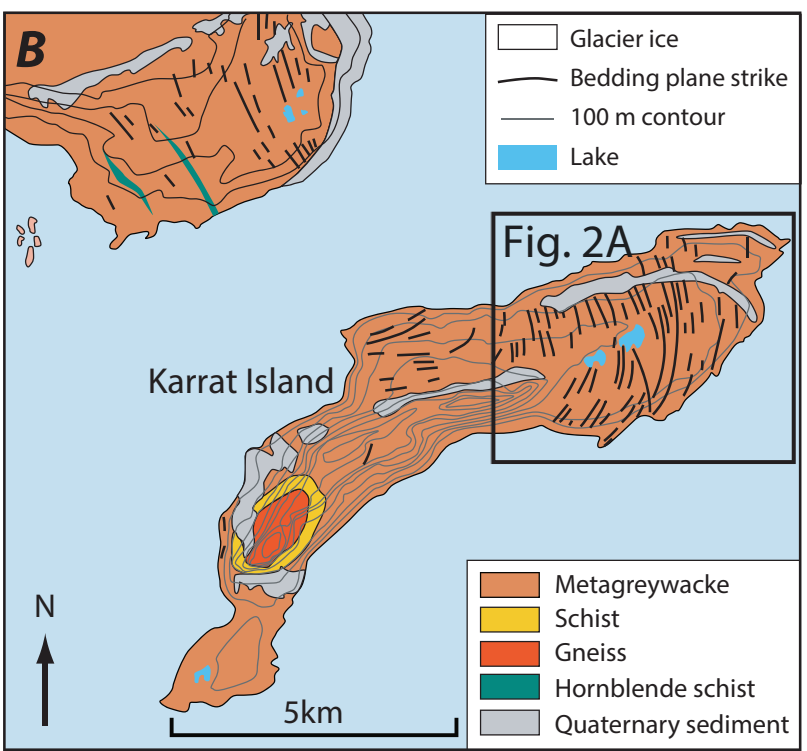
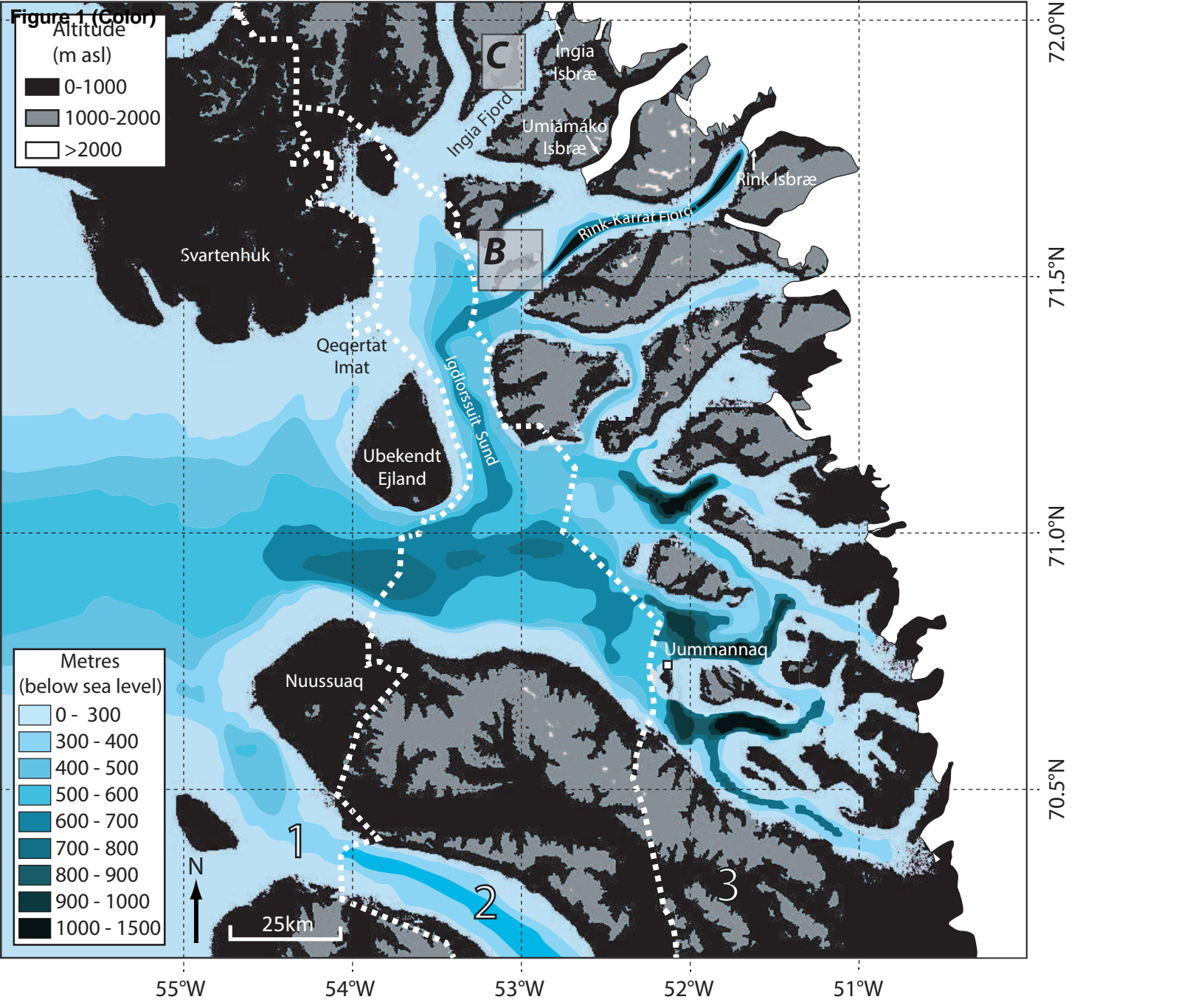


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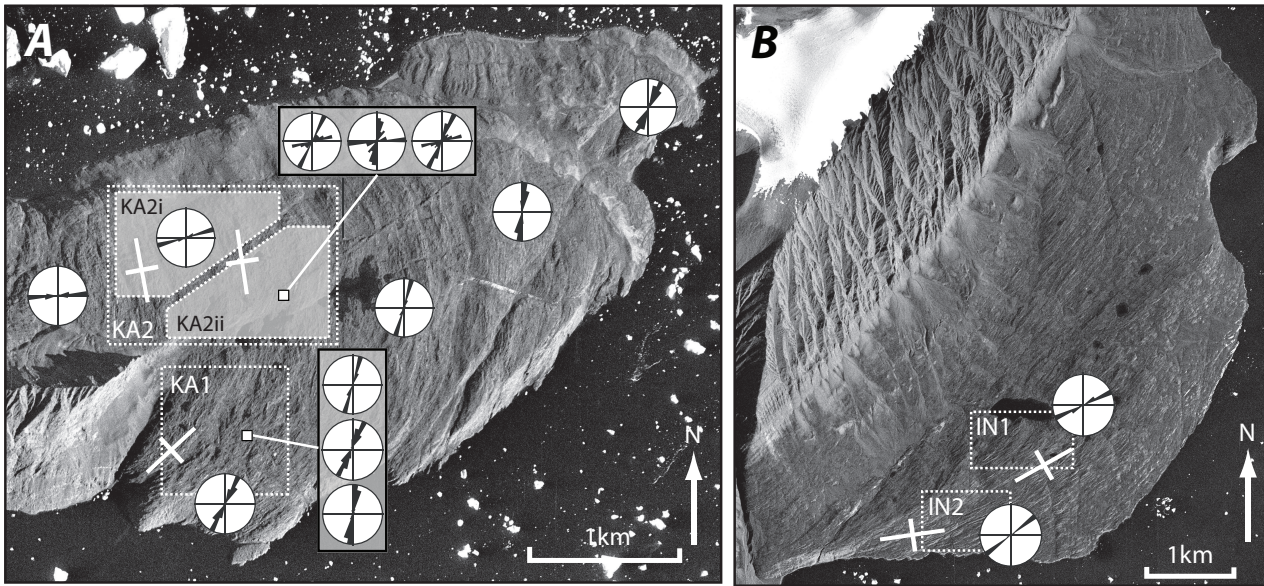


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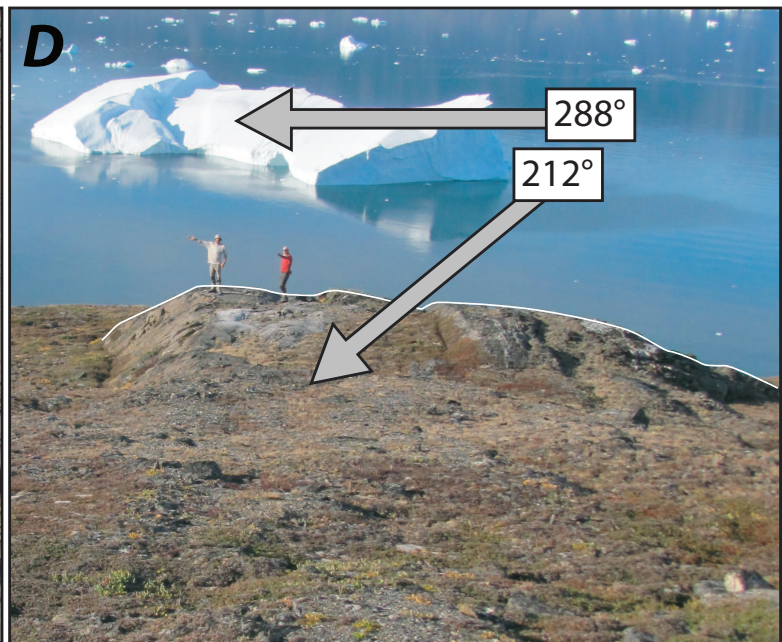
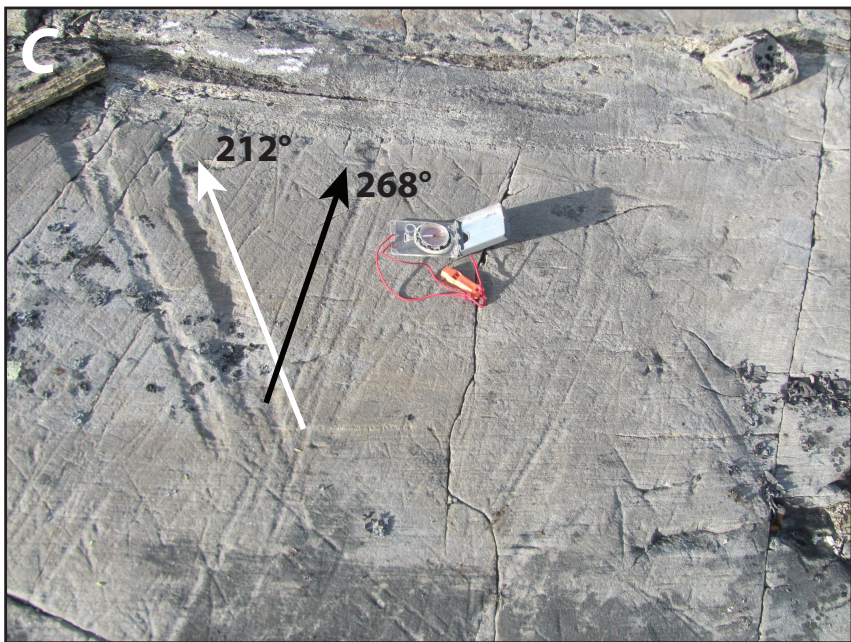
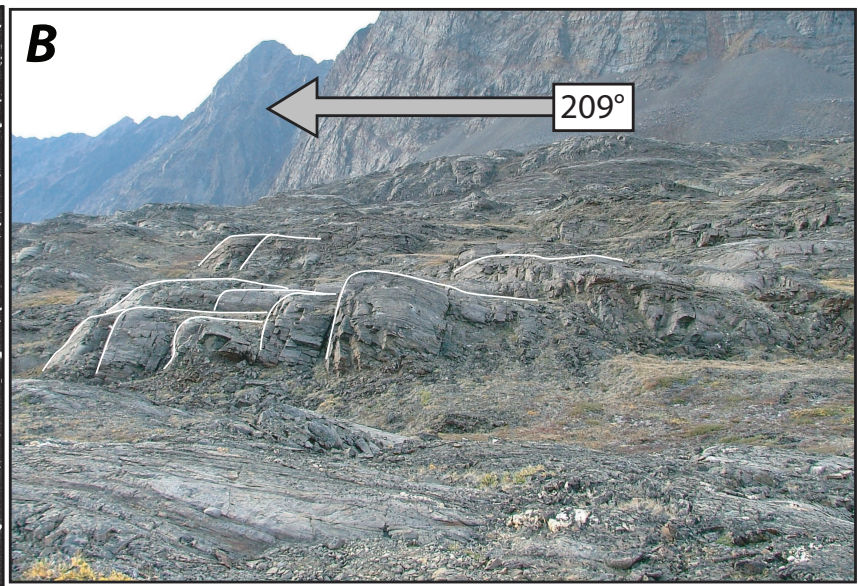
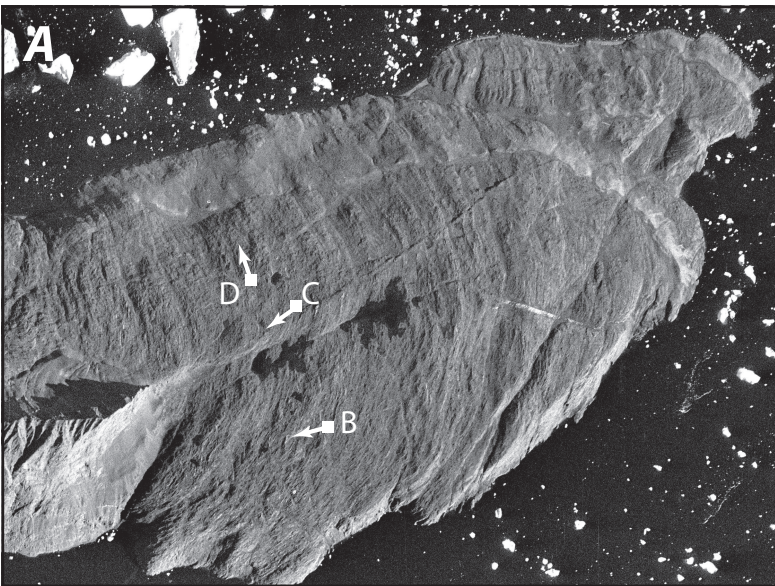


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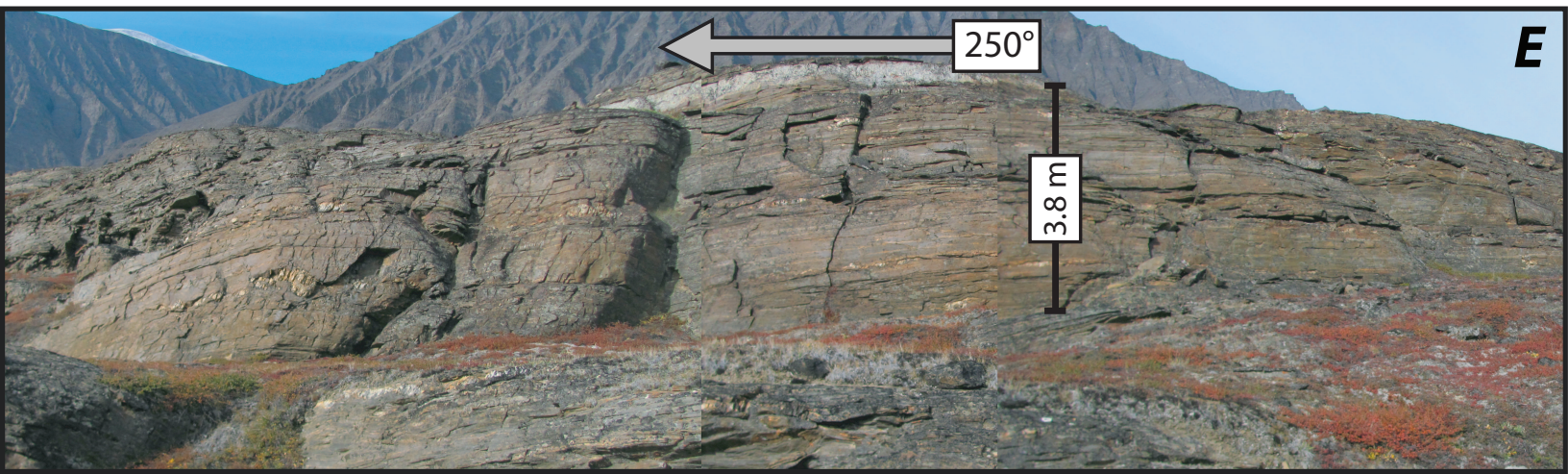
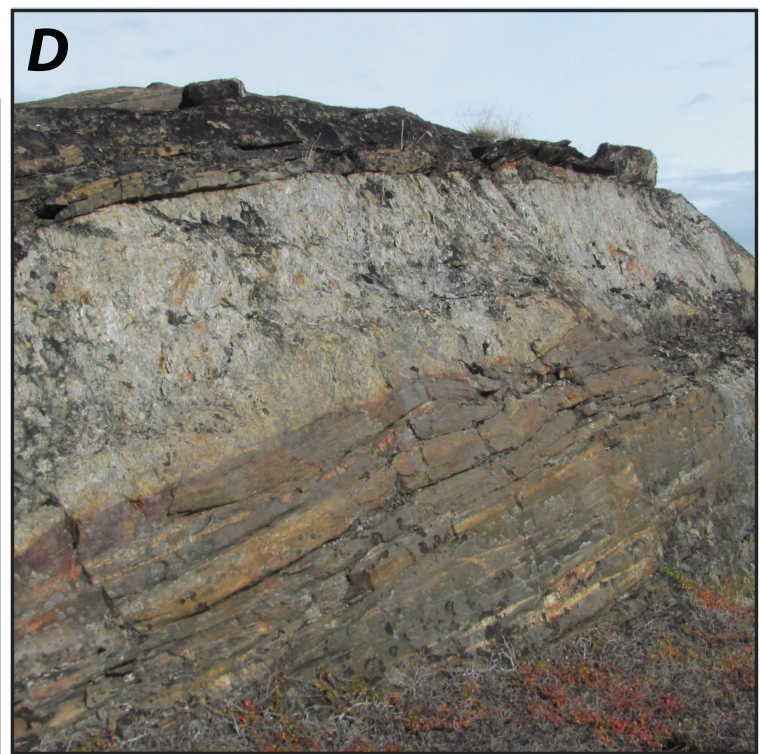
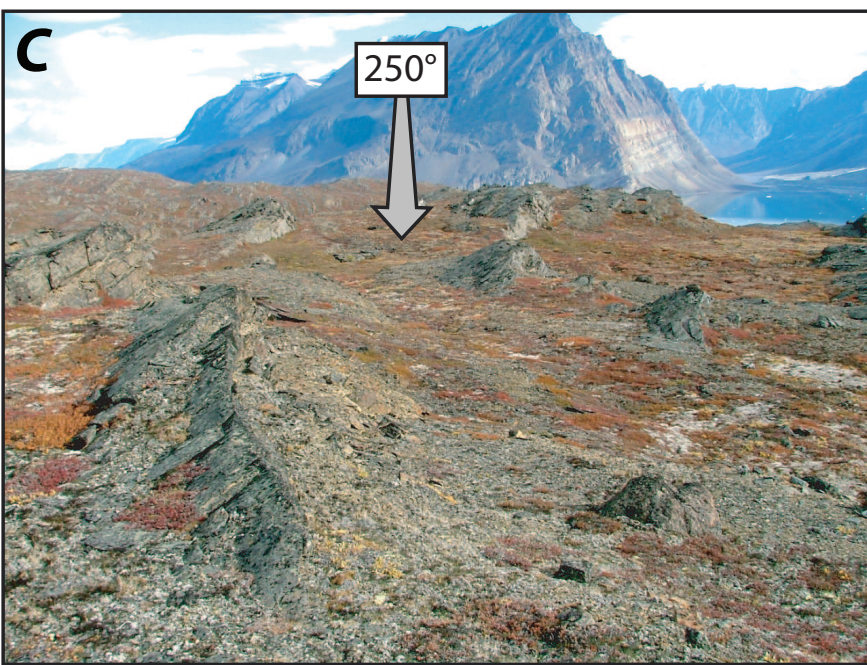
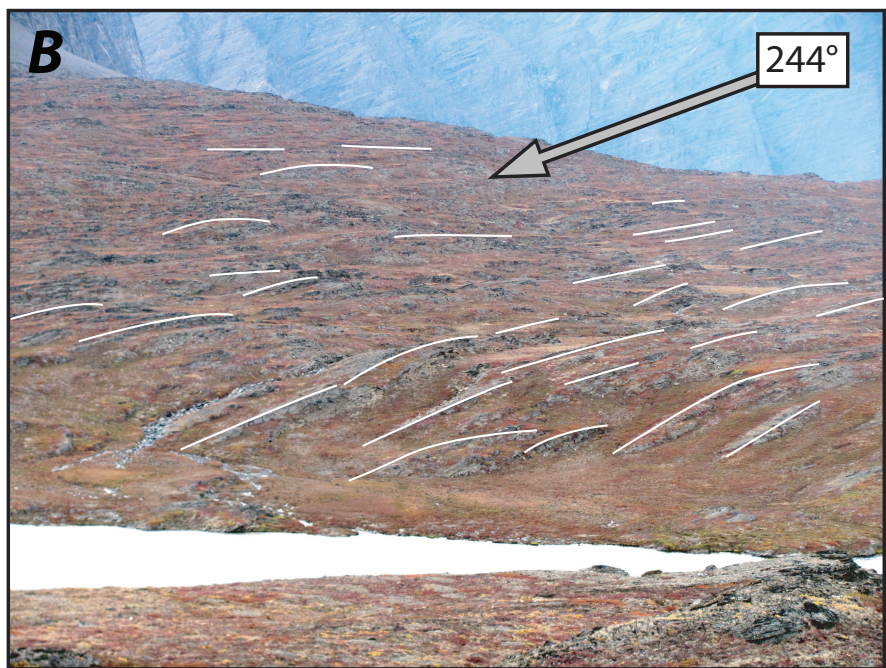
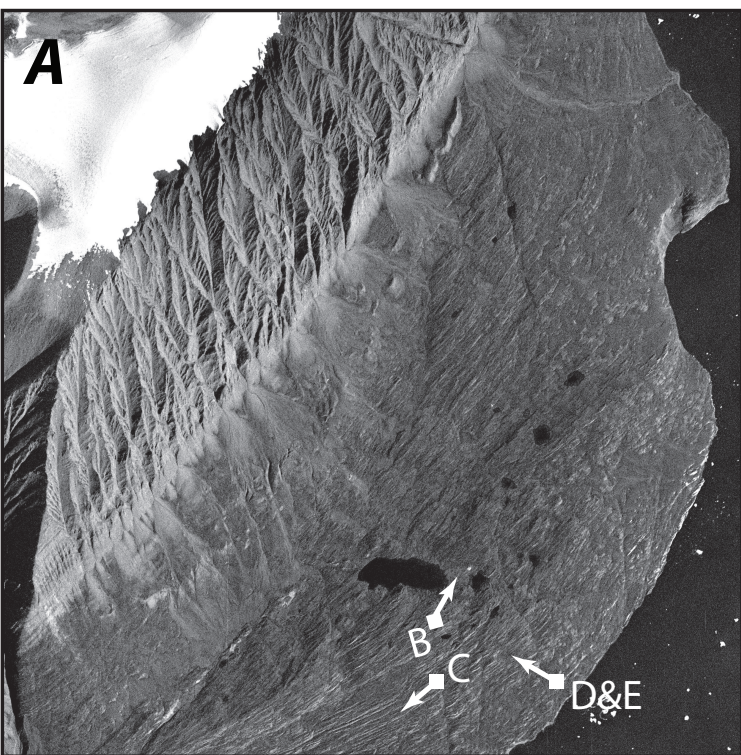


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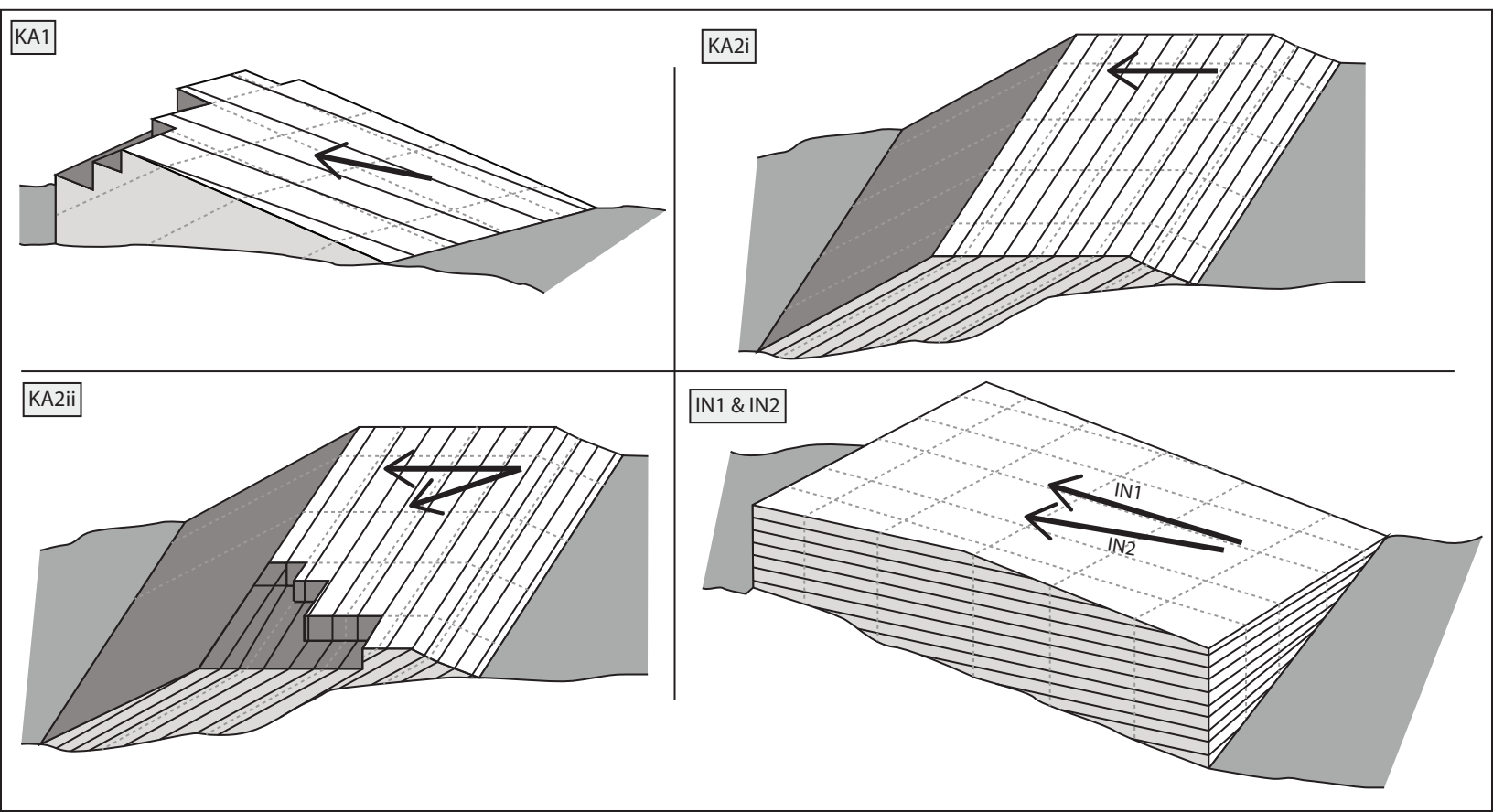


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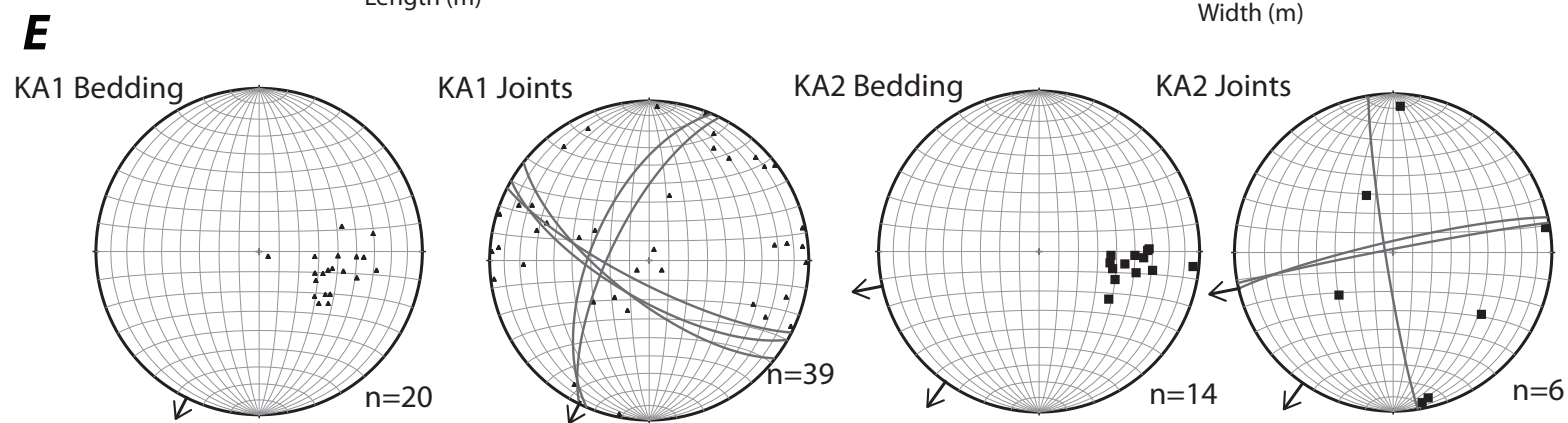
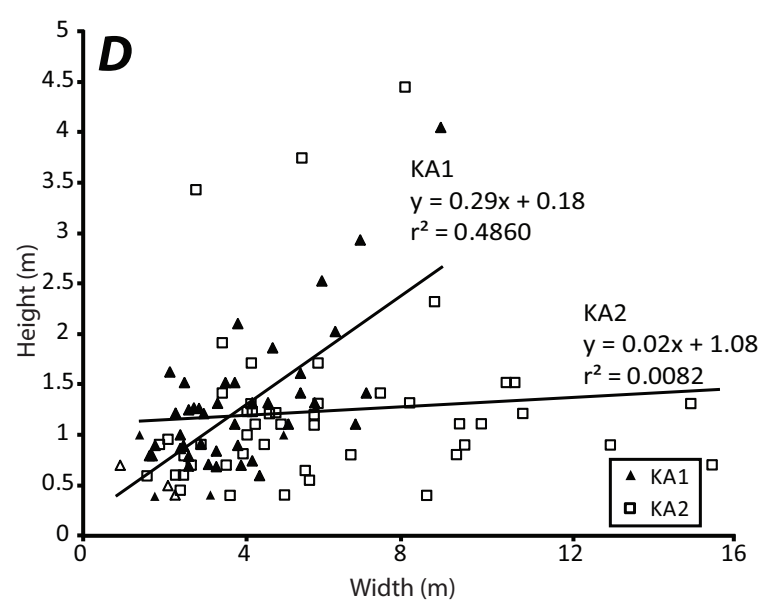
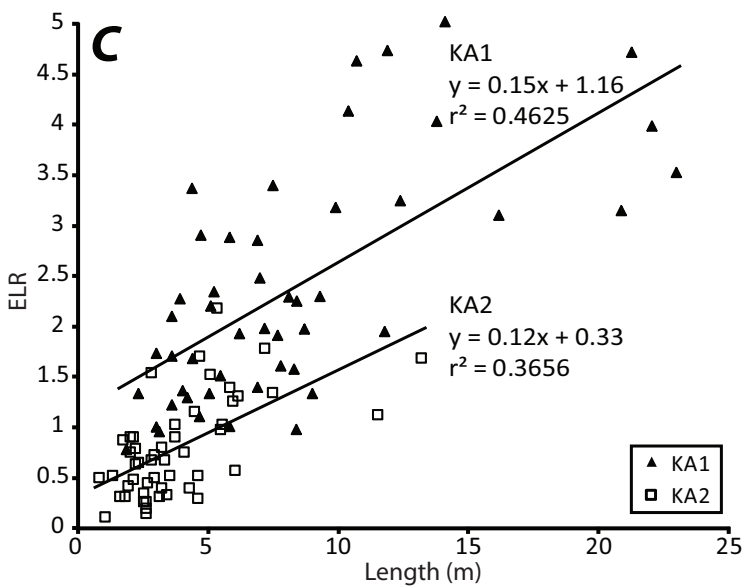
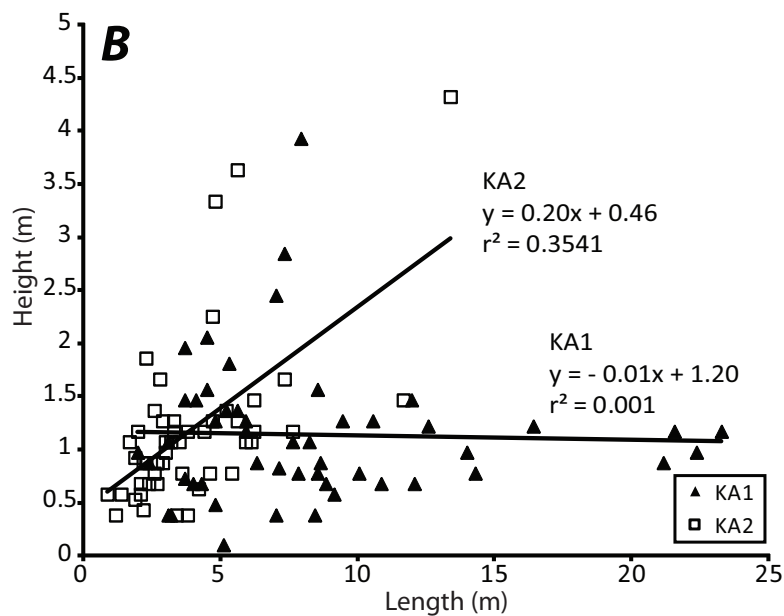
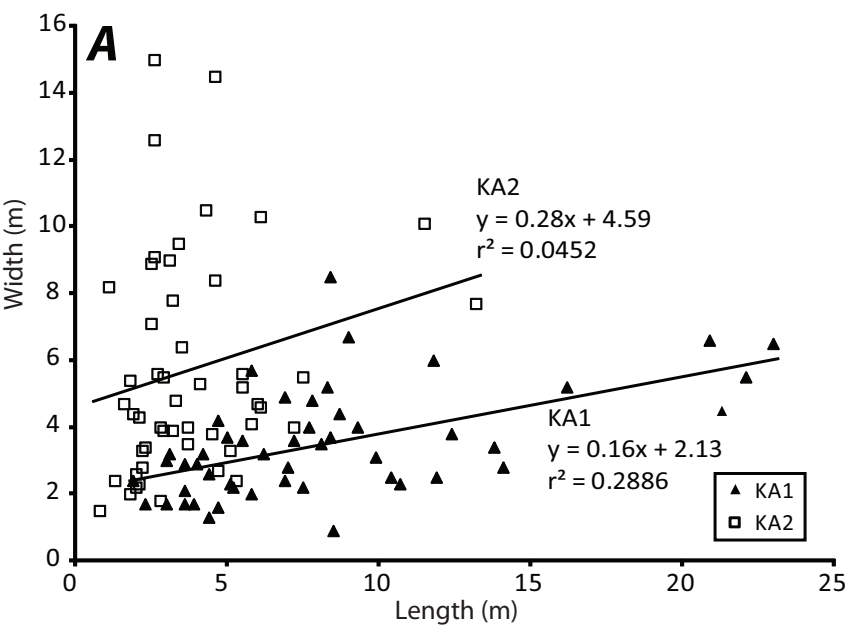


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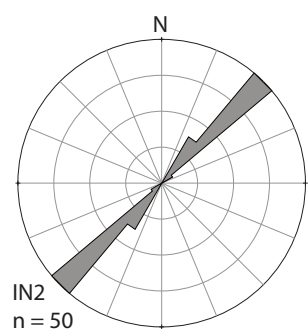
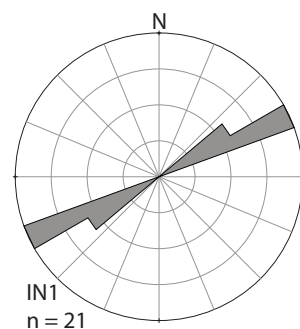
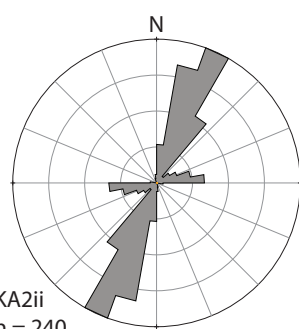
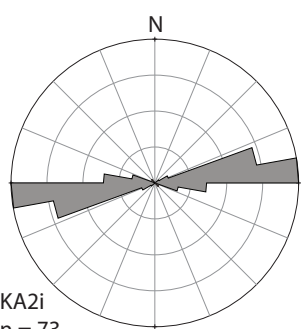
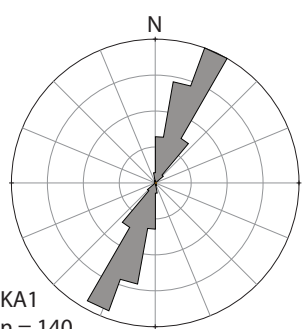


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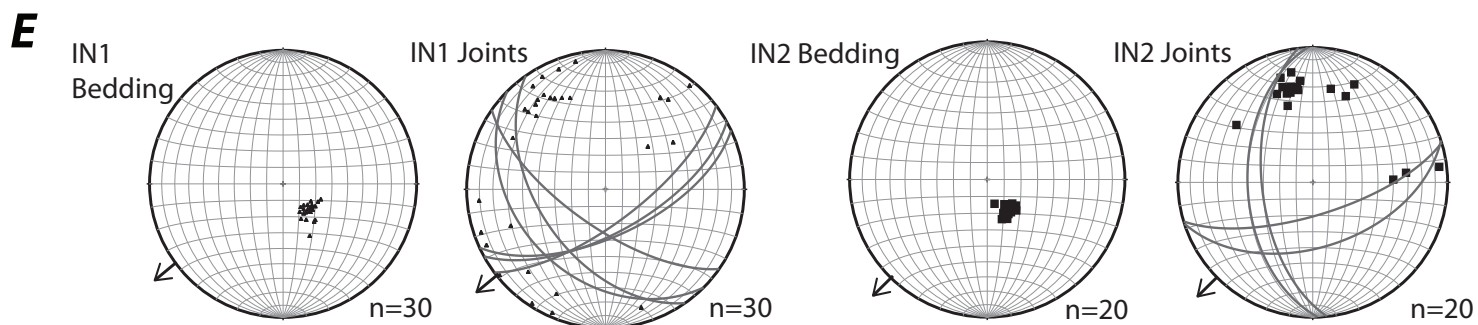
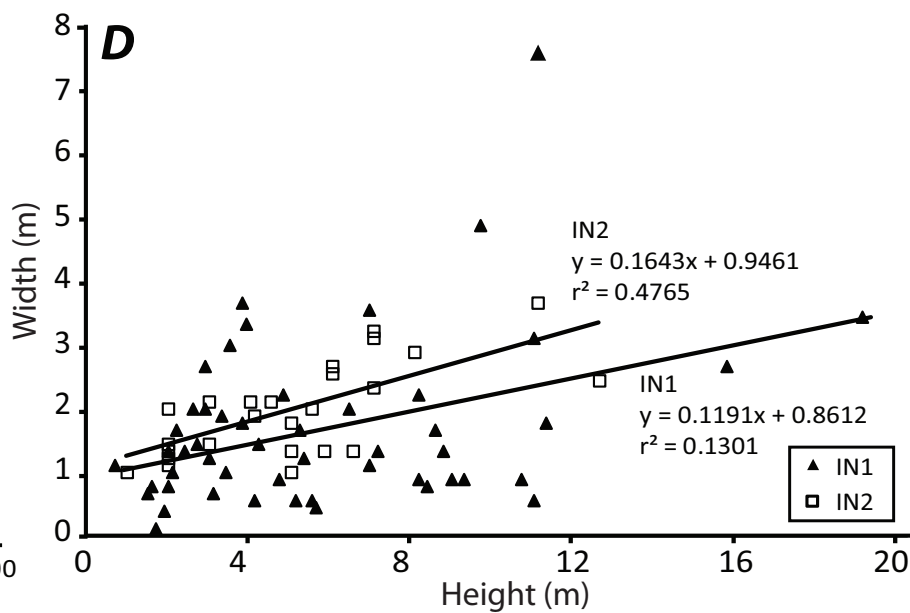
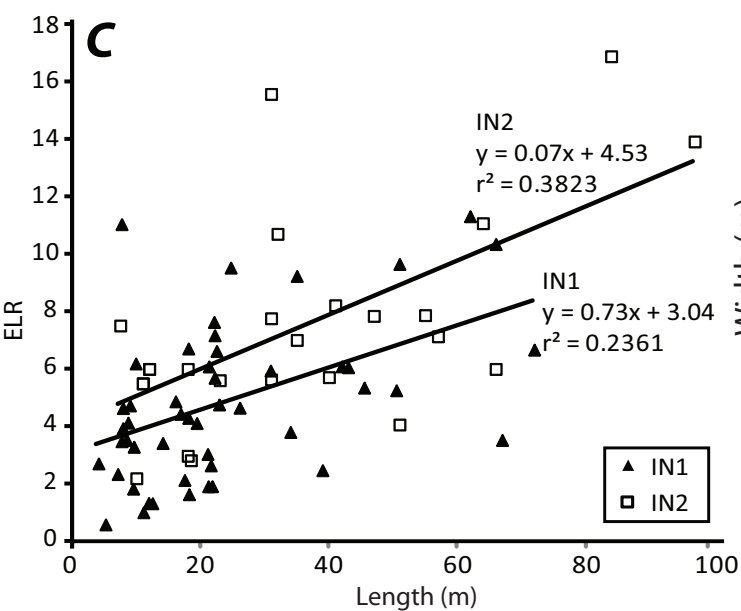
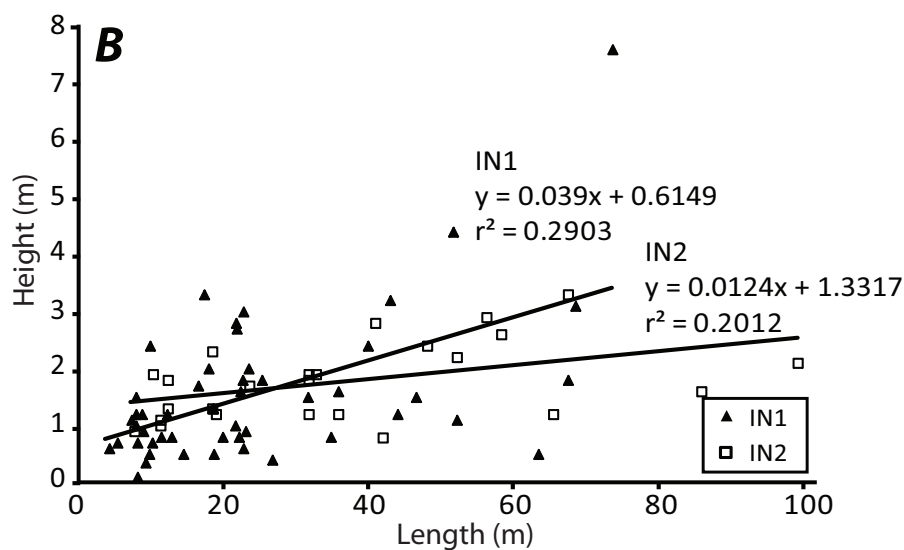
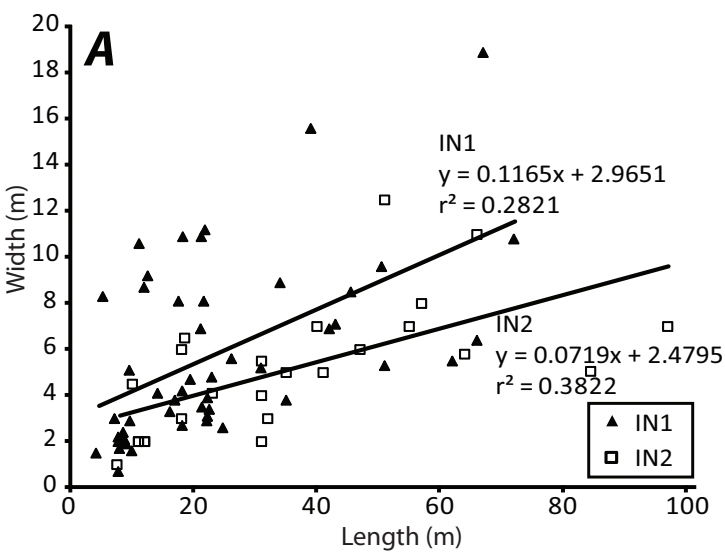
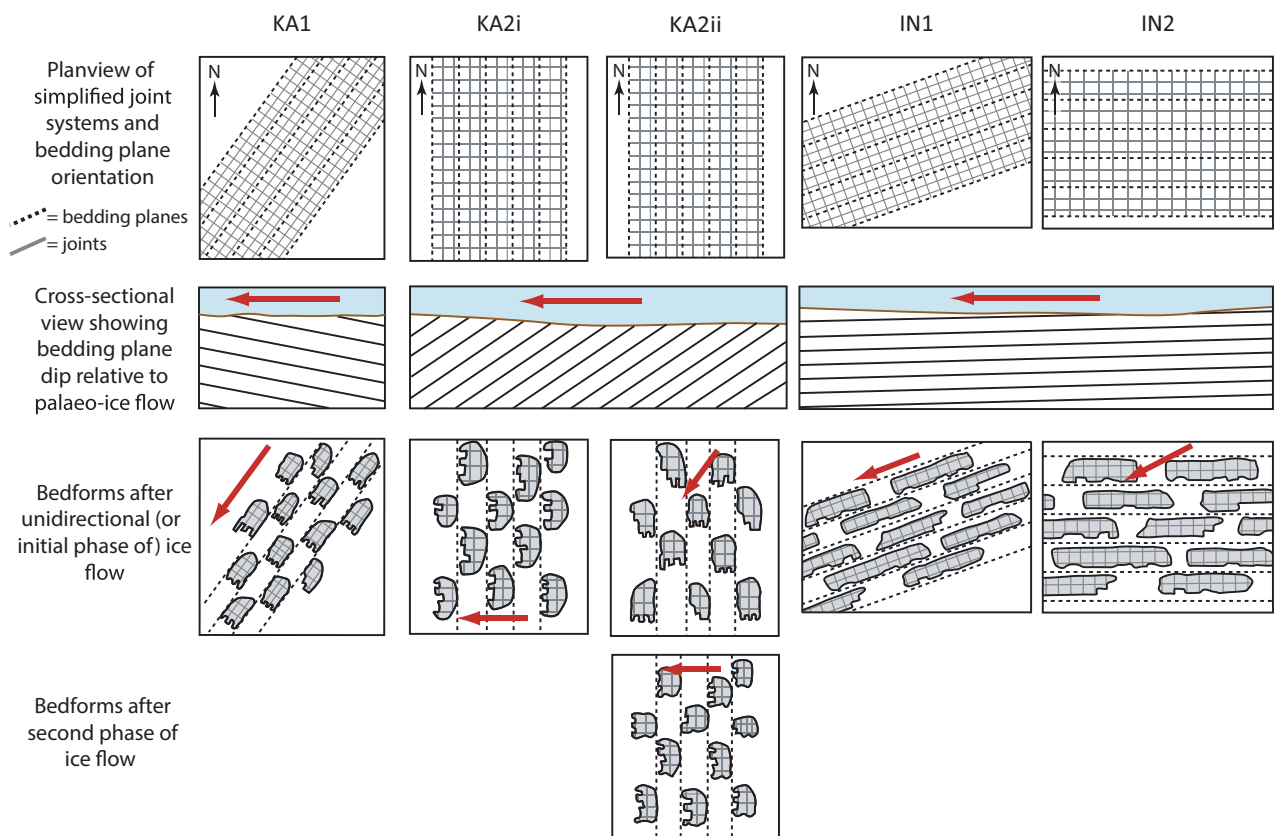


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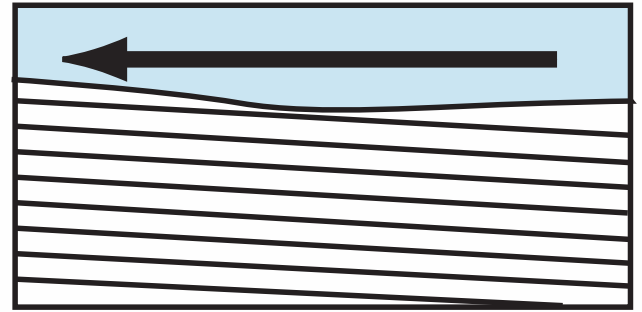
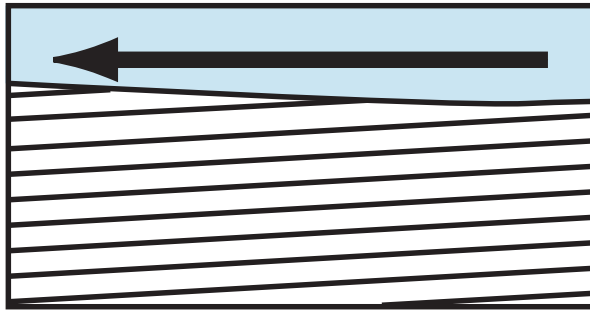


Low bedding plane dip relative to ice flow direction

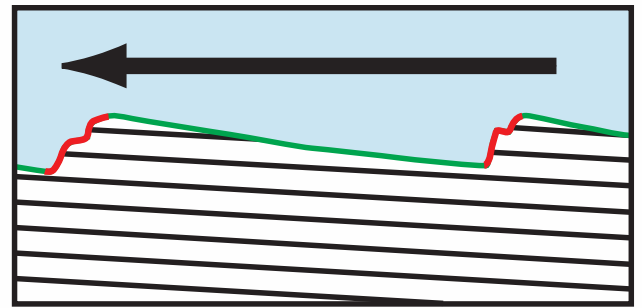
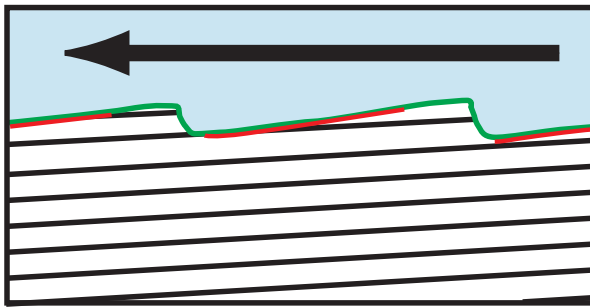
Down-ice dip

Up-ice dip

Landscape
prior to
glaciation



Bedforms
following
glaciation

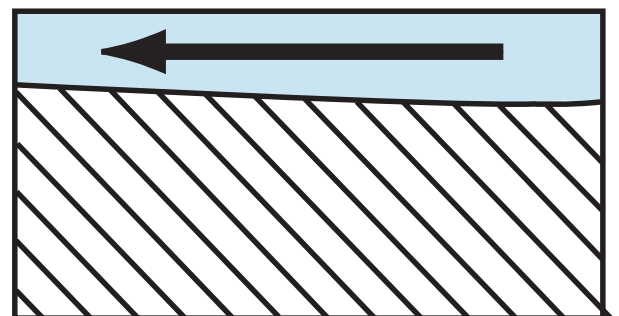
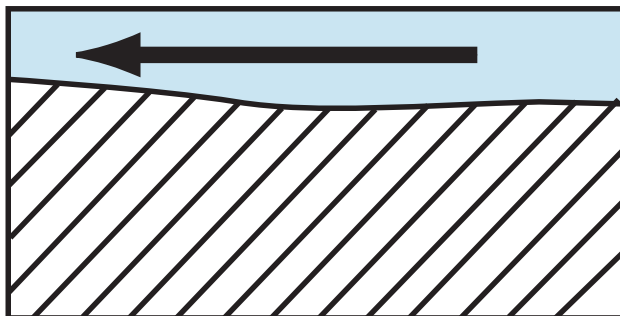


High bedding plane dip relative to ice flow direction

Down-ice dip

Up-ice dip

Landscape
prior to
glaciation



Bedforms
following
glaciation

